**How did the continents break apart in the NE Atlantic?**

**a tectonic and magmatic review**

(VERY) EARLY DRAFT, LG

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**ABSTRACT**

The Northeast Atlantic features a range of settings including microcontinents and conjugate volcanic margins ‘bridged’ to the south by the Greenland-Iceland-Faeroes Ridge. The last one is traditionally interpreted as an enigmatic swath of thick oceanic crust left in the wake of the controversial Icelandic 'hotspot'. New aeromagnetic data combined with seismic data allow us to revisit the complex rift-to drift evolution of the Norwegian-Greenland Sea. The long period of rifting, the structural style and the significant amount of breakup magmatism and seaward dipping reflectors (SDRs) formation clearly distinguish the conjugate volcanic (rifted) margins from magma-poor margins. Before the onset of breakup, we favour a tectonic scenario where a thick sedimentary basin developed during a drastic thinning of the continental crust associated with the progressive exhumation of high-grade metamorphic rocks before the volcanic margin formation. We interpret the nature of the crust closer to the SDRs as a mixture of preserved continental crust drastically affected by breakup-related intrusions. No zone of syn-rift exhumed and serpentinized continental mantle could be clearly identified near the SDR. New aeromagnetic data suggest that the onset of breakup was also a diachronic process with initial overlapping spreading centres. The volcanic margins and early spreading system developed differently and progressively along different margin segments. Presence of inherited tectonic buffers affected the early spreading ridge development leading progressively to microcontinent formation. The new magnetic data also show that an important mid-Eocene kinematic event at around magnetic chron C21r influenced the Norwegian-Greenland Sea. This event coincides with the onset of dyking and the increase of rift activity (and possible oceanic accretion?) between the so-called Jan Mayen microcontinent (or microplate complex) and the conjugate East Greenland margin. It led to the real and ultimate lithospheric breakup of the Norwegian-Greenland Sea in the Late Oligocene. The origin(s) of microcontinent formation, in general, is/are still unclear but recent receiver function analysis indicates that the presence of an old inherited sub-crustal slab may have had a non-negligible impact in the late reorganisation of the Norwegian-Greenland Sea.

\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*To update a bit….Just my last GSA abstract\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*

# INTRODUCTION

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Understanding the processes and development of continental to an oceanic rift in general has been a major challenge in the Earth sciences studies and a major component of the so-called Wilson cycle (REF…). Whereas mechanisms of crustal accretion at mid-ocean ridges are fairly well established (Carbotte et al., 2015), relationships between continental rifting and magmatism and the transition from rifting to spreading (e.g. the breakup stage), are still poorly understood and controversial. Important questions and scientific rift-to-drift thematic, usually concern: (1) how and how much the continental extension and thinning of the lithosphere predates breakthrough of the oceanic rift (e.g. Lavier and Manatschal, 2006; Van Avendonk et al., 2009; Huissmans and Beaumont, 2011; Pérez-Guissyné 2012; Brune et al. 2016), (2) how the rifting and breakup extension were physically and rheologically accommodated by faulting/shearing and/or igneous intrusion/underplating (e.g. Ebinger and Casey, 2001; Buck; 2006; Gueydan et al., 2008; Yamasaki and Gernigon, 2009; Rosenbaum et al., 2010; Lavier and Biemiller, submitted), (3) the ambiguous nature and relative timing of the continent–ocean transition and boundary (COT/COB) and breakup unconformity(ies) (Franke et al. 2013; Eagles et al. 2016), (5) the role of inheritance in rift and drift development (e.g. Harry and Bowling, 1999; Manatschal et al., 2014; Petersen and Schiffer, 2016) and/or spreading system segmentation (Behn and Lin, 2000; Bellahsen et al., 2013) (6) the magmatic and tectonic differences between magma-poor margins and volcanic passive margins (VPM)(Geoffroy et al., 2000) and (7) the relative roles of rigid plates and microcontinent formation (e.g. Muller et al., 2002; Gaina et al. 2009; Gernigon et al., 2012, 2015; Nemcok et al., 2016; Schiffer et al., 2017).

In this review paper, we focus on the Northeast Atlantic region (NEA) which is a well-known natural laboratory for studying both passive rifted margins and oceanic spreading development (Talwani and Eldholm, 1977; Roberts et al., 1984; Hinz et al., 1987; Doré et al., 1999; Skogseid et al., 2000; Faleide et al., 2008; Voss et al., 2009; Gaina et al., 2009: Tsikalas et al., 2012). The NEA is indeed characterised by a complex and polyphased system of continental and oceanic rifts that dislocated and ultimately break through a very heterogenous continental lithosphere inherited from previous Archaen, Precambrian and Palaeozoic geological events. The NEA also hosts one of the world’s largest igneous provinces (the North Atlantic Igneous Province) and some of the most famous volcanic passive margins (VPM) recognised on both conjugate margin of the NEA (Talwani and Eldholm, 1977; Hinz et al., 1987; Eldhom et al., 2000; Planke et al., 2000; Skogseid, 2001; Holbrook et al., 2001; Korenaga et al., 2001; Mjlede et al., 2005; Voss and Jokat, 2009). Closely related with its oil exploration history, pioneer refraction seismic data and potential field studies earlier revealed the deep crustal configuration of the rifted margin system and the presence of a thick pile of lava emplaced along the COT of most of the NEA margin segments (Talwani et al., 1976; Hinz, 1987; Mutter et al., 1984; Mutter and Zehnder, 1988; Olafsson et al., 1992). Additional Deep Sea Drilling Project legs 38 and 81 and Ocean Drilling Program legs 104, 152 and 163 in the Vøring Marginal High and Rockall Plateau and SE Greenland (Roberts et al., 1984; Eldholm et al., 1989; Saunders et al., 1998; Larsen et al., 1999) also contributed to the step forward development of early VPM magmato-tectonic concepts and petrology associated with the opening of the NEA ocean in Early Cainozoic time (Mutter et al., 1984; Skogseid and Eldholm, 1987; Eldholm et al., 1989; Meyer et al., 2006). VPM have often been considered differing from magma-poor margins by a number of fundamental geophysical and geological characteristics including: (1) the systematic presence of volcanic wedges of seaward dipping reflectors sequences emplaced along the proto breakup axes and traditionally referred as SDR, (2) the relative lack of strong passive margin subsidence during and after breakup (3) the massive emplacement of sill/dikes intrusions and volcanic vents that intruded associated sedimentary basins and (4) the presence in depth of thick high Vp-wave velocity lower crustal bodies (LCB) particularly recorded along their magmatic COT (White and McKenzie, 1989; Fowler et al., 1989; Planke et al., 1991; Holbrook and Kelemen, 1993; Eldholm et al., 1997; Barton and White, 1997; Eldholm et al., 2000). Subsequently, many conventional seismic refraction data have confirmed the deep crustal and geophysical specificities of the NEA (Ravaut et al., 2005; Mjelde et al., 2007; White and Smith, 2009; Voss et al., 2009).

Despite decades of exploration works, active research and modelling along the NEA passive margin (e.g. Lundin and Doré, 1997; Skogseid et al., 2000; Berndt et al., 2001; Færseth and Lien, 2002; Gernigon et al., 2003, 2004; Ren et al., 2003; Olesen et al., 2002, 2007; Mjelde et al., 2007; Faleide et al., 2008; Maystrenko and Scheck-Wenderoth, 2009; Breivik et al., 2014), many issues about the tectonic, crustal and magmatic development of the NEA still remain unclear and/or debatable notably in light of recent geophysical datasets. Large uncertainties still remain about the structure and petrological nature of the deep crust and the prolongation/preservation of the Precambrian/Caledonian basement underneath the sedimentary basins. The mode of deformation leading to continental breakup and subsequent sea-floor spreading is also unclear and alternative scenarios can be proposed.

In this review paper, we have tried first to summarise (as much as possible) the present day geological and tectonic understanding of the entire NEA. We propose is an exhaustive but comprehensive description and a discussion of the most recent concepts and tectonic problematic related to and/or proposed from rift to drift and post-rift evolution of the NEA. We remind first the main common features and characteristics of the NEA and related VPM and summarise the different tectonic and geodynamic models, scenario and concepts so far proposed for their formation. In the second part of this review, we focused on the late magmatic and tectonic insights of the Mid-Norwegian margin evolution which is the most constrained margin segment of the NEA to understand the previous rift-to-drift thematic. An important aspect is to understand if VPM can be the final result of a classic set of tectonic sequence as recorded in magma-poor margins (e.g. Sutra et al., 2014) or not (e.g Geoffroy et al., 2000; Gernigon et al., 2015; Theissein-Krach et al., 2017). In this context, the NEA rifted margin is particularly ambiguous because they superimpose on older rifts system that undergone progressive lithospheric thinning and thermal reequilibration and compared with similar rift or passive rifted margins affected by significant amount of magmatism and volcanism. In the discussion, we aimed to show that volcanic rifted margins represent atypical rifted and magmatic and that the total rupture of extended plates may involve different processes and magma-tectonic scenario compared to popular `magma-poor´ (Iberian-type) margin.

Then how continental masses finally break apart and how the early spreading finally initiated and evolves in the NEA is also unclear. Why some spreading segments in the NEA suddenly or progressively extincted and ‘jump’ to start a new spreading system and a microcontinent or microplate system (e.g. the Jan Mayen) also remain a geodynamic problem particularly intriguing and questionable in the NEA realm. The impact of magmatism and involvement of inheritance and/or controversial mantle plume model(s) are also discussed in the second part of this review in light of the most recent geophysical dataset of the NEA.

\*\*\*\*\*\*\*\*\*\*\*\*\* Introduction to reorganise a bit, later \*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*

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More stratigraphic input could be welcomed (Martyn) but in that case, the paper will be extremely long. Should I keep the tectonic only\*\*\*\*\*\*\*\*\*\*\*\*\* To be discussed in Durham 3. Alternatively, a schematic stratigraphic chart of the NEA main basin could be enough?? Martyn input ??

# CONJUGATE MARGIN SEGMENTS AND SPREADING SYSTEM OF THE NEA: CRUSTAL CONFIGURATION AND MAIN BASIN CHARACTERISTICS

## Physiography of the continental shelves and regional spreading systems

The NEA system is characterised by a compound system of continental rifted margins and adjacent oceanic sub-basin that extends from the Charlie Gibbs Fracture zone to the Malloy Fracture zone in the north. The main morphology and first order geodynamic configuration of the NEA seafloor were well established since the first bathymetric (Johnson and Heezen, 1967; Perry et al., 1986) and regional or semi-regional magnetic (Vogt and Avery, 1974; Vogt et al., 1981, 1986; Verohef et al. 1996) and gravity (Grønlie and talwani, 1982) compilations of the NEA realm. The asymmetric distribution of the continental shelf, shallow plateaux, deep bathymetric troughs and marginal ridge as confirmed by the more recent released potential field and bathymetric compilation of the NEA (GEBCO-2014, Weatherall et al., 2015; IBCAO 3.0, Jakobsson et al., 2012), gravity (version 24.1 from Sandwell and Smith, 1997 or the DTU13, Anderson et al., 2015) and EMAG3 or CAMP (Maus et al., 2007; Gaina et al., 2011) already reflect the ultimate consequence of ~ 380 Myr of episodic rift events and subsequent seafloor spreading activity in the NEA (see figures x, y and z).

Along the NEA, one can already observed that the bathymetry of the continental shelf and slope varies considerably in width and steepness from the exposed mainland domain (Svalbard Scandinavia, Greenland and U.K./Ireland) towards the noticeable oceanic domain of the NEA (Johnson and Heezen, 1967). Offshore, the water depths generally increase from approximately 100 m close to the shore to ~350–500 m at the shelves edge and deepen rapidly from 2000–2500 m at the foot of the slope to 3000–4000 m into the deep oceanic basins. In the NEA, asymmetry of the continental shelves is also obvious. The NW European margins are rather characterised by a complex morphology including plateaus (e.g. Rockall-Hatton Plateau; Faroe Plateau) isolated banks (e.g. the George Bligh Bank, the Lousy Bank and the Bill Bailay Bank, Porcupine Bank), marginal highs (e.g.; the Voring Marginal High) and deep bathymetric throughs (e.g. Rockall Through, Faroe-Shetland Chanel; Porcupine Seabay). On the contrary, the conjugate Greenland margins exhibit either a narrow continental shelf in the south or a wider platform in NE Greenland margin; the last one initially located in the prolongation of the Barents Sea epicontinental domain before the Early Cenozoic opening of the NEA.

In the central part of the NEA, the Greenland-Iceland-Faeroe Ridge (GIFR) delineates a distinct NW-SE bathymetric swath that divides the southern oceanic domain of the NEA from the Norwegian-Greenland Sea to the north (Vogt et al., 1986). The enigmatic GIFR covers shallow oceanic water area of approximately 430.000 km2 and stretches 1150 km across the NEA between the Faeroe Plateau and the central East Greenland to the west. The water depth along the GIFR is usually less than 500 m and emerges on Iceland where the actual, active and cross-cutting Mid-Atlantic Ridge separates the Eurasian and North American plates at the present day. The GIFR incorporates the Iceland plateau, the aseismic Greenland-Iceland Ridge, and the Iceland-Faeroe Ridge south of the Norway Basin. Iceland is bordered to the south by the Reykjanes Ridge segment of the Mid-Atlantic Ridge and to the north by the Kolbeinsey Ridge. As presented in detail by xxxx, this issue, rifting in Iceland is not homogeneous. It occurs differently in different sections of the rift area. In the southern part of Iceland is focused on two main parallel rift zones. The Reykjanes Peninsula Rift in SW Iceland is the landward continuation of the Reykjanes Ridge that connects to the Western Volcanic Zone and indirectly to the more active Eastern Volcanic Zone accommodated by the South Iceland Seismic Zone. To the north, the Norther Volcanic Zone is connected to the Kolbeinsey Ridge by the Tjörnes Fracture Zone. (SEE ICELAND REVIEW THIS ISSUE.)

South of the GIFR, the Reykjanes Ridge divides the Irminger Basin from the Iceland Basin (Vogt et al., 1971, 86; Talwani et al., 1971; Searle et al., 1998). The two oceanic sub-basins in between the Hatton-Rockall Plateau and the SE Greenland margin have water depths exceeding xxx m with depths less than 1000 m near Reykjanes Ridge. In between, the Reykjanes Ridge extends 900 km from the Reykjanes Peninsular in SW Iceland at 64º N up to the Bight Fracture Zone at 57ºN (Searle et al., 1998; Martinez and Rey, 2017). The axial depth of the Reykjanes Ridge increases from 2 km at the Bight Fracture Zone to the sea-level at the Reykjanes Peninsular, south of Iceland. The Reykjanes Ridge is spreading at a half-rate of 10 mm/year and has long been recognised as anomalous amongst slow-spreading ridges for its highly oblique spreading direction (the ridge axis trends 036º, approximately 28º from the spreading normal) (Parsons et al. 1993; Searle et al., 1998). The spreading ridge also lack of an axial valley in the northern half and show enigmatic south-pointing ‘V-shaped’ or ‘time-transgressive’ ridges possibly related to the Icelandic mantle ‘plume’ or not (Searle et al., 1996; Jones et al., 2000; Hey et al., 2010; Martinez and Rey, 2017). See also Chapter X this issue.

Within the NEA and north of the GIFR, the Norwegian-Greenland Sea is fundamentally different in terms of morphology, and spreading configuration compared to the southernmost oceanic domain of the NEA (e.g. south of the GIFR). Its present a diachronic spreading architecture which includes the extinct Aegir Ridge, the Kolbeinsey Ridge, the Mohn’s Ridge and the Knipovich Ridge at the edge of the Barents epicontinental sea (Talwani and Eldholm, 1977). The Aegir Ridge represents an extinct spreading ridge located in the central part of the deep Norway Basin. This oceanic sub-basin extends northward from the GIFR to the continental slope off the Norwegian shelf. Its northern boundary is the Jan Mayen Fracture Zone one of the most prominent oceanic fracture zones of the NEA. In between the Aegir Ridge and the active Kolbeinsey Ridge a shallow marine plateau extends from Iceland and the Jan Mayen Island is characterised by the Beerenger volcanoe, the northernmost active stratovolcano volcano in the world (Gjerløw et al., 2016). This atypical plateau is characteristic of the Jan Mayen “microcontinent” (Auzende, 1980; Gudlaugsson et al., 1988; Nunns, 1980) also referred to as the Jan Mayen Microplate Complexe due to its complexity (Gernigon et al., 2015; Schiffer et al., 2017). The expected Microplate Complexe includes the Jan Mayen Ridge and the Southern Ridge Complex to the south. West of the Jan Mayen Ridge, the Jan Mayen Basin is more than xxxx m deep at the edge of the shallow Icelandic Plateau formed during the onset of spreading along the Kolbeinsey Ridge (Vogt et al., 1980). The modern and active Kolbeinsey Ridge spreads orthogonally at ultraslow rates (20-15 mm/yr. full spreading rates) terminating to the north at the West Jan Mayen Fracture Zone. Despite its slow spreading rate, the Kolbeinsey Ridge is volcanically active along its entire length and hosts a vigorous neovolcanic zone as further described by Brandsdottir et al. (2004) and Hooft et al. (2006). The ridge is characterised by overlapping spreading discontinuities (Appelgate, 1997), which subdivide the spreading ridge into a Southern, Middle, and Northern second-order ridge segments. The northernmost of the Kolbeinsey Ridge, includes the Eggvin Bank, an anomalous volcanic and high bathymetric showing igneous crustal thickness variations from 8 km to 13 km (Tan et al., 2017), an atypical 'overcrusting' relatively similar to the controversial Voring Spur also observed in the regional trend of the Jan Mayen Fracture Zone (Gernigon et al. 2009; Breivik et al., 2014).

North of the Jan Mayen Fracture Zone, the active Mohn’s Ridge separates the Lofoten Basin from the Greenland Basin accreted in between the Lofoten margin and the conjugate NE Greenland margin. The Mohn’s Ridge, between Jan Mayen Island and approximately 73.5ºN has water depths ranging between 1000 and 2000 m and show typical slow spreading rates of 15-16 mm/year. Structural interpretation of seabeam imagery confirms that the Mohn’s Ridge result from oblique rifting initiated around 27 my ago (Dauteuil and Brun, 1983). To the north, the Greenland Fracture Zone and associated Est Greenland Ridge (also a possible microcontinent, Dossing et al., 2008) separates the Greenland Basin from the smaller and shallower Boreas Basin. The Lofoten Basin, west of the Lofoten margin is xxxx km deep and extends up to the West Barents Sea. Between the Greenland Ridge and the Molloy Fracture Zone the oldest sea-floor spreading magnetic anomaly that can be identified with confidence is magnetic chrons anomaly 18 (39 Ma; Middle Eocene) (Engen et al. 2008) despite the poor magnetic coverage of the region.

The Knipovich Ridge is finally the northernmost members of the NEA mid-ocean ridge system. The northern spreading ridge is a ~550 km long from the Mohn’s Ridge to the Fram Strait. The Knipovich Ridge represents a structure characterised by a slow spreading rate (~14 mm/year) and show a significant oblique component trending between 40º to 55 º from the spreading direction. It develops along the West Barents shear margin and connect further north with the ultra-slow Gakkel Ridge (<10 mm/year) in the adjacent Eurasian Basin (Dick et al., 2003, Cochran et al., 2003).

\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*Extra comment/REF from Fernando for the southern oceanic domain???\*\*\*\*\*\*\*\*\*\*\*\*\*

I will also make regional bathy grav-mag regional maps\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*

## Rifted segments of the NEA and pre-breakup crustal and main sedimentary basin characteristics

\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*This section will include a compiled set of the crustal section to illustrate the main segments.

I also have a set of basin sections for all Norwegian-Greenland Sea. Also few sections from Rockall Hatton VPM from my old post-doc (never published) – Martyn probably have more updated version on Rockall/Faores

I need this chapter to show those volcanic margins are clearly separated from the previous rift axis. The important point is that the NEA breakup is not the result of a continuum of lithospheric deformation. The problem is that most of the conceptual, analogique and numerical model often assume a continuous and short (<50 Ma) process from the rift to breakup. It works very well for Gallica-Iberia or similar systems but not for the NEA !!!!!

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### Rockall-Hatton region

In the southeastern part of the NEA, the Rockall-Hatton volcanic rifted margin and associated marginal basins are the dominant structural features of the U.K./Irish offshore regions (Naylor et al., 1999; Shannon et al., 1999; Edwards, 2002). The Rockall Basin is a noticeable NE- to NNE-trending deepwater rift basin which narrows from approximately 350 km in the south to less than 200 km in the north. The dominant Caledonian NE–SW structural grain tracked onshore influenced much of the rift orientation along the Atlantic margin (Doré, 1999; Hitchen et al. 2013). Within the basin, the Caenozoic sequences are relatively well constrained (Stoker et al., xxxx). On both flank of the Rockall Basin; NNE- to NE-trending, mainly Mesozoic to Late Palaeozoic remnants and platform half-grabens the most important of which are the contiguous NNE-trending Slyne and NE-trending Erris basins on the eastern margin of the basin (Naylor et al.,1999; Dancer et al., 1999; Chapman et al., 1999). Significant thickness of Jurassic sequences has been deposited on the basin flank of the Rockall (Jones and Underhill 2011).

The age, distribution and thickness of the pre-Palaeogene stratigraphic intervals in the central part of the Rockall are more challenging and mainly inferred from velocity data derived from the results of wide-angle seismic experiments (O'Reilly et al., 1995; Shannon et al., 1999; Mackenzie et al., 2002; Morewood et al., 2005; Ravaut et al. 2006), potential field modelling (Kimbell et al., 2004) and more restricted long-offset industry data. England and Hobbs (1997) earlier concluded that any Jurassic sediments in the central part of the Rockall Basin, if present, must be very thin because expanding spread profile data indicate that the Cretaceous sediments lie directly upon rocks with seismic velocities consistent with basement. The most recent estimates of gross stratigraphic thicknesses in the southern and central Rockall vary between 4.5 and 7 km, with the thickest sections occurring towards the margins of the basins were inferred Late Palaeozoic to Early Mesozoic syn-rift basins are locally imaged by seismic reflection (Naylor and Shannon, 2005; Mackenzie et al., 2002).

In the Irish sector of the Rockall Basin, extension factors of ?c=4 to 6 have been earlier reported by Shannon et al. (1999) and Morewood et al. (2005) from the RAPIDS seismic experiment. The crystalline crust along the RAPIDS is approximately 6 km thick, with the Moho interpreted to lie at a depth of 14 km and the thinnest crust occurring towards the margins of the basin (Hauser et al., 1995; O'Reilly et al., 1995; Mackenzie et al., 2002; Morewood et al., 2005).

In the northern Rockall Basin, refraction data (Kingelhoffer et al. 2005) show a moderately stretched continental crust (>10-13 km thickness) beneath a maximum of 6 km of sediments in the central part of the basin. Compared to the southern Rockall Basin, no evidence for LCB (>7 km/s) could be distinguished beneath the northern part of the Rockall Basin. There has also been much debate regarding also the deep nature of the crust that floors the Rockall Basin with suggestions including: 1) True oceanic crust, formed during Mesozoic (e.g. Roberts, 1975; Chappell and Kusznir, 2005), or Late Palaeozoic (Russell and Smythe, 1978; Haszeldine and Russell, 1987); 2) a mix with stretched continental crust with zones of oceanic crust (e.g. Megson, 1987; Smythe, 1989); 3) stretched to super-extended continental crust with concomitant igneous intrusion (Shannon et al., 1994, 1999; Hauser et al., 1995; O'Reilly et al., 1996; Morewood et al., 2005) or hyperextended crust with serpentinised mantle (O’Reilly et al., 1996; Lundin and Doré, 2011).

Only few direct information about the terranes present in the Rockall-Hatton region is available. The adjacent Outer Hebrides massif most likely represents terranes similar to the Lewisian complex (Laurentian foreland) of NW Scotland (Hitchen et al., 2004; Tate et al., 1999; Morton et al., 2014). To the west, in the distal margin, the Palaeogene basalts recovered from the George Bligh High show evidence of contamination by Archaean crust (Hitchen et al. 1997), suggesting that a deeper terrane boundary between the Archaean and Palaeoproterozoic terranes lies between the Rockall High and the George Bligh High (Dickin, 1992; Hitchen, 2004) as supported by recent sand provenance studies (Morton et al. 2014) and refraction data (Funck et al. 2008). The Precambrian and Archaean basement probably sutured close to the Anton Dohrn Lineament that separates the northern and central Rockall basins (Roberts et al., 1999). Similar Archaean-Middle Proterozoic contact may have extended up to the conjugate Greenland before breakup.

East of the Rockall Basin and separated by the Porcupine Bank; the Porcupine Basin (including the North Porcupine and Seabight basins) forms a N-trending rift graben approximately 400 km in length, and tapering from 150 km width in the south to 50 km in the north (Naylor et al., 2002). The Porcupine Basin is an elongate failed rift basin of Late Palaeozoic to Caenozoic age (Croker and Shannon, 1987; Naylor et al., 2002; Readman et al., 2005). The Porcupine Median Ridge is a ridge feature in the middle of the southernmost part of the basin. During the last three decades, this ridge has been successively interpreted as a volcanic structure, a diapir of partially serpentinised mantle, or a block of continental crust (Tate and Dobson, 1988; Johnson et al., 2001; Reston et al., 2004; O'Reilly et al., 2006; Prada et al., 2017). The main phases of rifting are interpreted to have occurred during Permo-Triassic, Jurassic and earliest Cretaceous times (e.g. Naylor et al., 2002; Naylor and Shannon, 2005).

West of the Rockall Basin, The Hatton Basin lying in the central part of the Rockall-Hatton Plateau is approximately 500 km long and 200 km wide and extends north-eastwards in between Rockall Bank and the Hatton Bank. This basin is one of the less understood basins of the NEA, notably due to poor imaging underneath the thick Palaeocene–Eocene volcanic and intrusive rocks observed in the Rockall-Hatton Plateau (e.g. Hitchen, 2004). Nonetheless, the eastern margin of the Hatton Basin has been drilled by DSDP wells 116 and 117, proving approximately 850 m of sediments ranging from upper Palaeocene to recent close to the breakup axes. Results from the RAPIDS and HADES wide-angle seismic experiment indicated that the Hatton Basin may contain up to 3.5-5 km of older sediments on both sides of the Hatton bank (Vogt et al.,1998; Smith, 2006; Ravaut et al., 2006). The age of the sediments underneath the Early Cainozoic basalt is still unclear. Shannon et al. (1999) suggests the presence of Upper Carboniferous, Permo- Triassic to Jurassic strata. Wedge of mid-Cretaceous sediments have been, however, proven in shallow boreholes on the UK Hatton margin (Hitchen, 2004) and recently correlated with conjugate Early Cretaceous grabens (Amnassalik Basin) described SE Greenland; Gerlings et al., 2017). The age of rifting is not known with certainty. Hanisch (1984) suggested that the main crustal extension may have occurred during Maastrichtian prior to seafloor spreading to the west whilst Smythe (1989) argued for a mid. Cretaceous onset of rifting. Lundin and Doré (2002) and Doré et al. (1999) suggested that the main rifting phase of the Hatton Bank was Late Cretaceous-Palaeocene.

Results from the RAPIDS, iSIMM, HADES and potential field modelling studies have suggested thicknesses for the crystalline crust beneath the Hatton Basin of between 10 and 20 km and show evidence of LCB and related intrusion (Shannon et al., 1999; Kimbell et al., 2005; Ravaut et al., 2006; Smith, 2006; White et al., 2008). White et al. (2008) also show that the Hatton Bank is 25 km thick at the edge of a sharp volcanic margin.

Further north, the George Bligh Bank, the Lousy Bank and the Bill Bailay Bank extends from the Hatton and Rockall Banks in the SW to the Faeroes Plateau in the NW. These banks represent fragmented and segmented piece of continental crust (Kingelhoffer et al., 2005; Funck et al., 2008). Beneath the Lousy banks, a thick continental crust of 26 km is modelled underneath 4-5 km of basalt. An up to 12 km thick LCB was also modelled in the distal part of the margin and was traditionally interpreted as breakup magmatic underplating (Kingelhoffer et al. 2005). The channels in between the banks are subparallel to known transfer zones and lineaments in the region (Funck et al., 2008). Two main basalt layers with velocities of 4.9 to 5.6 km/s can be correlated from the Faroe Islands up to George Bligh Bank and the total thickness of the basalts may be as high as 6 km.

### Faroe-Shetland rifted margin

North of the Rockall Basin, the Faros-Shetland rifted is a 500-800 km long margin segment of the NEA which extends between Scotland and the Norway Basin. A large part of the rifted system and associated sedimentary basins is partly covered by the thick Cainozoic volcanic traps (up to 4-5 km) on the Faroes Plateau. The sub-basalt geology and the deep basement structures of the Faroe-Shetland region are partly constrained by a combination of deep seismic surveys (Klemperer & Hobbs 1991; Raum et al., 2005; White et al. 2008) and/or potential field modelling (Kimbell et al. 2005; Chappell and Kusznir 2008; Jegen et al., 2009; Rippington et al., 2015; Haase et al., 2016). Both refraction studies and potential field indicates a thinning of the crust from the Shetland Platform to the central Faeroe-Shetland Basin where the continental crust is thicker than 7-10 km (Raum et al., 2005; White et al., 2008). Like the Hatton Bank, the late Mesozoic failed rifts of the Faeroes-Shetland Basin is separated from the new ocean basins by a thick continental block underneath the Faeroes Plateau (so-called the Fugloy Ridge) where the continental crust is up to 30-25 km expected underneath more that 5-7 km of basaltic layers (White et al., 2008). In the lower crust of the COT, high Vp velocity LCB (Vp>7.2 km/s) and strong sub-horizontal reflections interpreted as sills intrude the dipping fabric of the possible Archaean continental crust at the edge of the GIFR (Bott et al., 1974: Richardsson et al., 1998; White et al., 2008; Olavstottir et al., 2016).

In the Faroe-Shetland region, exploration wells on the Sula Sgeir, Rona, Judd and Corona highs have encountered crystalline basement (granulite facies orthogneiss, metasedimentary rocks, and mafic and ultramafic meta-volcanic rocks) showing affinities with the Lewisian Complex described further south in the northern Rockall Basin (Ritchie et al. 2011; Trice et al., 2014). The Caenozoic basaltic rocks notably exposed on the Faroe Islands are presumed to either rest on top of pre-Cretaceous sedimentary rocks or Lewisian crystalline basement (Bott et al., 1974; Brewer and Smythe, 1984). Geochemistry also agrees the Faroe Island Basalt Group on the Faroe Islands overlies Precambrian continental crust (Gariépy et al. 1983; Hald and Waagstein 1983; Holm et al. 2001). Seismic refraction experiments in the offshore part of the Faroe Plateau revealed sub-basalt sedimentary layers that can reach thicknesses of 1-8 km depending on the different interpretations (Richardson et al., 1998; 1999; Raum et al., 2005; White et al., 2008). Ambiguity remains due to multiple ways of interpreting a sub-basalt layer with a P-wave velocity of 5.2–5.7 km/s and the possible contamination of sub-basalt sedimentary rocks with numerous igneous sill intrusions (Raum et al., 2005)). Recent ambiance noise tomographic model predicts similar metamorphic rocks (Vp~5.75 km/s) between 4-6 km underneath the archipelagos with deeper Archaean terranes (6.2<Vp<6.3 km/s) between 7-10 km (Sammarco et al., 2017).

Along this segment of the NEA, significant Devono-Carboniferous deposits are found in the southern sub-basins of the Faeroes-Shetland Basin and in the Palaeozoic Shetland Platform that earlier developed in between the Orkney and Shetland islands (Bird et al., 2014). Isolated occurrences of Devono-Carboniferous red beds have also been identified on the Corona and Westray highs. Fractured Devono-Carboniferous red beds on the southeastern flank of the Rona High and beneath the West Shetland Basin (Trice et al., 2014).

The Late Palaeozoic-early Caenozoic rift system of the Faroe–Shetland region is dominated by a set of NE-trending basement ridges and prominent Mesozoic-Palaeocene sub-basins that form at the edge of Shetland Platform. The orientation of the main structural features suggests that their development and shape may have been controlled by a combination of structural inheritance from the Caledonian Orogeny (Coward et al. 2003; Doré et al., 1999; Fossen et al., 2010). The inherited Caledonian tectonic grain, which is particularly, expressed by major the NE-trending basin-border faults, such as the Rona Fault and the Shetland Spine Fault. To the east, The West Shetland Basin and SE Marginal Basins all currently underlie the Shetland Platform. According to Ritchie et al. (2011). The Faroe–Shetland Basin is limited to the south by the Judd High and smaller NE-trending marginal basins, including the East Solan, SouthSolan, West Solan and North Rona basins.

A number of publications also highlight the importance of NE-SW lineaments across the Faroe-Shetland Basin and their potential relationships with oceanic fracture zones further west (Rumph et al. 1993; Lamers and Carmichael 1999; Ellis et al., 2009). The origins of these lineaments are unclear; however, hypotheses include reactivated Precambrian shears (Knott et al. 1993) and/or oblique extension features formed as a response to Mesozoic rifting (Rumph et al. 1993). The lineaments are an important feature in controlling basin segmentation, the location of transfer zones and possibly controlling the input and distribution of magma in the Faroe–Shetland Basin (Schofield et al. 2015. Moy and Imber (2009) found no expression of the lineaments in the Caenozoic section and note that the lineaments on the Faroe-Shetland segment of the margin rarely align with the Caenozoic oceanic fracture zones to the northwest.

Post-rifting subsidence and later Eocene-Oligocene–Miocene localised compression resulted in minor folding of Palaeocene lavas in the Faroe–Shetland Basin (Johnson et al., 2004; Tuitt et al., 2008; Ziska and Varning, 2008). The Munkagrunnur, Wyville Thomson and Ymir ridges are major tectonic features within the area and arc approximately perpendicular to the primary Caledonian structural trend. Both these ridges are interpreted to consist of crystalline basement blocks capped by Mesozoic (including Cretaceous?) and/or early Caenozoic rocks (Raum et al. 2005; Ritchie et al. 2011).

\*\*\*\*\*\*\*\*\*\*\*\*\*Extra Comments/REF from Martyn ??\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*

### Mid-Norwegian margin

The mid-Norwegian passive continental margin can be subdivided into three structural and crustal segments, comprising the Møre, the Vøring and the Lofoten-Vesterålen margin segments (Blystad et al., 1995). After almost 50 years of petroleum exploration the overall and first order structural, stratigraphic and volcanic framework of the mid-Norwegian margin is relatively well established and represent the most constrained margin region of the entire NEA (Figure. x).

2.2.3.1 Møre and Vøring margin segments

Strictly speaking, this More and Voring 'rifted' margins segments consists, at the present day, of NE trending deep Cretaceous basins flanked by paleo-highs, terraces and shallow platforms (Blystad et al. 1995; Brekke, 2000). The main structural provinces include (1) the Trøndelag Platform, (2) the Halten Terrace, (3) the Møre and Vøring basins, and (4) the Møre and Vøring marginal highs to the west. To the south, the Vøring Basin is interconnected with the Møre Basin through a regional and mega-crustal transfer zone, the so-called the Jan Mayen corridor that most likely influences the late evolution of the Vøring volcanic transform margin (e.g. Berndt et al., 2001) and indirectly subsequent breakup and oceanic accretion of the Norwegian-Greenland Sea (Gernigon et al., 2015).

The proximal part of the Møre margin has been mapped here in a number of seismic reflection and refraction studies (Olafsson et al. 1992; Kvarven et al. 2014, 2016), as well as inferred from potential field modelling (Reynisson et al., 2010; Nirrengarten et al. 2016). The narrow and shallow platform comprises a series of NE-SW-elongated basement highs (Gossa High, Ona High, Giske High and Manet Ridge) separating the Slørebotn Sub-Basin, the Magnus Trough and the Marulk Basin from the deep Møre Basin (Jongepier et al. 1996, Brekke 2000, Grunnaleide and Gabrielsen, 1995; Theissen-Krach et al., 2017). Beneath the narrow platform, Moho depths vary between 29 km beneath the shelf and 37 km onshore (Maupin et al., 2013; Kvarven et al. 2014). These NE-SW structures of the narrow platform domain clearly developed along the trend of the Møre-Trøndelag Fault Complex showing a clear inherited topographic and geophysical signature extending up to the West Shetland Basin (Grønlie and Roberts 1989; Hurich et al. 1996; Doré et al. 1997; Nasuti et al., 2011; Watts et al. 2007). To the south, this inherited regional feature roughly delineates the transition between the northern North Sea and the Møre margin.

The Trøndelag Platform of the Voring margin, to the north represents a broader proximal domain compared to the narrower platform domain observed in the proximal part of the More margin. Trøndelag Platform consists of a thick Permian-Jurassic Platform on top of a relatively thick crust thinning progressively (from 35 to less than 20-15 km) to the west (Brekke, 2000; Breivik et al., 2011). To the west, the Halten and Donna terraces developed along an inner necking zone observed between the Trøndelag Platform and the deep Cretaceous sag-basins where the continental crust became rapidly relatively thin (<10 km) (Breivick et al., 2011; Maystrencko et al., 2017). With more than 200 exploration/production wells, the shallow water platform and the terrace domains lying in the prolongation of the Norwegian mainland are the most constrained and dated structural province of the Norwegian shelf and NEAIn the proximal part of the margin and associated terraces, the geometry, stratigraphy and geological evolution of the pre-Middle Triassic basins have been particularly described by Müller et al. (2005) and Færseth et al. (2012). Across most of the Trøndelag Platform, the Late Palaeozoic–Early Triassic basin is expected to be locally thick (6-7 km) and the Middle Triassic to Jurassic sequences show a relatively uniform thickness (5 km of thickness in average) with a gradual thinning towards the SE border of the platform. Færseth et al. (2012) described the Late Permian-Early Triassic sub-salt rift system as a series of 'en echelon' half-grabens controlled by major east-dipping border faults that are expected from the Vestfjorden Basin to the Froan Basin. The shallow platform and adjacent terraces area also illustrate a large variety of structural styles (extensional forced folds, fault propagation folds, basement-involved and basement-detached normal faults and narrow grabens) symptomatic of stratified salt tectonics affected the different Jurassic rifting events (Koch and Heum, 1995; Pascoe et al., 1999; Withjack and Callaway, 2000; Richardson et al., 2005).

The central Vøring and Møre basins located respectively at the edge of their broad and narrow Platform domains are mainly characterised by the huge thickness of the Cretaceous successions (Brekke et al., 2000; Lien et al., 2005). These Cretaceous basins are 125-150 km-wide and the base of these large 'sag' and flexural basins (the base Cretaceous unconformity) can locally reach drastic depths of up to 8-9 km which means that the sediments are most likely metamorphosed. The sag- basin clearly developed above very thinned continental crust revealed by the numerous refraction profiles acquired over the last 50 years (Planke et al., 1991; Olafsson et al., 1992; Mjelde et al., 2005; Kvarven et al., 2014, 2016) and potential field modelling (Reynisson et al., 2010; Maystrenko et al., 2017).

In the Vøring Basin, the sedimentary section is up to 14 km (Raum et al. 2002; Mjelde et al., 2002). The maximum thinning and necking zone of the crust has been interpreted in the Rås and Traena Basin where the crust thinned rapidly to almost less than 5 km, the thinnest crustal estimation of the entire mid-Norwegian margin (Maystrenko et al., submitted). The Vøring Basin is then underlain by thinned classic continental crust (6.0-6.9 km/s) with a thickness of 2–11 km (5-10 km in average) on top of the controversial and distal LCB (see discussion chapter x). The distal LCB shows Vp velocity higher than 7.2–7.4 km/s restricted to the Outer Voring Basin (Gernigon et al., 2003; Adlelmalak et al., 2017) and absent or limited in the western part of the sag basin showing a maximum basin thickness of 8-10 km in the easternmost Ras and Traena Basin (Mjelde et al., 2008; Breivik et al., 2011; Maystrencko et al. 2017). In the Møre Basin shows up to 14 km-thick sedimentary sequence above a continental crust also thinned to 3–7 km and locally underlain by an up to 4 km-thick LCBs with velocities of around 7.2 km/s (Olafson et al., 1992; Raum et al., 2002; Kvarven et al., 2014, 2016). In the eastern part of the sag-basin, the crust is the thinnest (~5 km) but suprinsignly the most recent OBS data do not show any evidence of LCB just west of the necking zone (Kvarven et al., 2014). The inner LCB is only imaged in the adjacent thick platform domain and an outer LCB is mostly recorded in the distal part of the More Margin (Kvarven et al., 2014).

As part of the large Møre and Vøring basins, several sub-basins separated by Cretaceous infill and/or basement high evidence underlain by high density/high velocity inherited crust (e.g. Utgard High) is observed (Zastrozhnov et al., submitted). Further south, gravity and seismic data suggests that the Utgard High was progressively dislocated and possibly linked with the deep (continental) 'en echelon' crustal rafts and extensional riders (e.g. Slettringen, Grip and Vigra highs)(Gernigon et al., 2015). Alternative hypotheses have, however, suggested the presence of Cretaceous seamounts (Lundin & Doré, 1997), but the crustal fabric and the structural shapes of these structures reveal by more recent seismic data rather support the presence of continental blocks at that level. The large crustal and basin-scale accommodation and transfer zone between the Møre and Vøring margins (e.g. the Jan Mayen Corridor) reflected by the structural style of the pre-Cretaceous and Early Cretaceous rift system. Compared to adjacent Møre and Vøring segment, 2D and 3D potential field modelling also suggests that the Jan Mayen corridor is characterised, on average, by a very thin but preserved continental crust (Maystrenko et al., 2016).

As part of the VPM system, the Outer Vøring Basin particularly represents a tectonic hinge zone linking the outer flexure of the Vøring Basin and the volcanic domains to the west. Bjørnseth et al. (1997), Ren et al. (2003) and Gernigon et al. (2003, 2004) earlier mentioned the structural segmentation and timing variability of the Outer Vøring Basin ridge complex and associated sub-basins. Compared to the proximal domains, only a few numbers of exploration wells (19) including Utgard High and the Fles Fault Complex) have been drilled in this vast exploration area. Key exploration wells to understand the stratigraphy and nature of the pre-volcanic basins are located on the Nyk High (6707/10-1, 6707/10-2S and 6707/10-2A; 6707/12-2; 6706/12-3), Vema Dome (6706/11-1), Naglfar (6706-6-1, Hvitveis), north Gjallar Ridge (6704/12-1, 6705/10-1), south Gjallar Ridge (6603/5-1, Dalsnuten), and Vigrid Syncline (6604/10-1). The deepest stratigraphic well in the Outer Vøring Basin was the most recent Dalnustsen well (6603/5-1) that bottomed into undifferentiated Lange Formation sediments (Lower Cretaceous) (Norwegian Petroleum Directorate, fact page; Bronner et al. 2015). The precise age of the deepest formations remains uncertain but confirm that the base of the Cretaceous is deep in a large part of the Vøring Basin, including (part of) the outer ridge system. All these wells illustrate the complexity and variability of the Outer Vøring Basin development in space and time. Most important, the wells provide relevant stratigraphic and age calibration crucial to understand the timing and nature the final stage of rifting leading to the final lithospheric rupture of the mid-Norwegian margin and the onset of VPM development (see dedicated chapter x).

Further west, the Vøring and Møre Marginal Highs form the outermost zones of the extended Norwegian continental crust, to the west of which lies the transition zone to oceanic crust and the true continent–ocean boundary (COB). The marginal highs are characterised by substantial amounts of intrusive rocks, as well as thick layers of seaward-dipping reflectors resulting from extrusive volcanic rocks and volcanic margin formation (Eldholm et al. 1988, 1989, 2000; Berndt et al. 2001). The COT was modelled by Mjelde et al. (2005) over the Voring Marginal High showing a very thick crust including a distal LCB (20-25 km). Vp velocities of ~6.0 km/s in the top of the main crustal layer is conformable with continental (Granitic) basement, whereas corresponding velocities oceanwards (6.9 km/s) are indicative of gabbroic, oceanic crust. The expected but controversial COT was characterised by intermediate velocities of ~6.5 km/s interpreted as heavily intruded continental crust in Mjelde et al. (2005) (See also dedicated chapter).

2.2.3.2 Lofoten-Vesteralen margin segment

The narrow Lofoten-Vesterålen margin represents the north-eastern and uplifted part of the mid-Norwegian margin, comprising a pronounced, steep shelf edge. The Lofoten-Vesterålen archipelago is, together with the adjacent Norwegian mainland, characterised by a complex topography, which is the result of several phases of uplift and erosion formed during the long geological development of this area (Tsikalas et al., 2001; Hansen et al., 2009, 2012; Mokhtari and Pegrum, 1992; Olesen et al., 2004; Dag Ottesen et al., 2009; Rise et al., 2013; Maystrenko et al., 2017). Tectonically, the Lofoten-Vesteralen margin includes the Røst, Ribban and Vestfjorden basins, which are separated by the Utrøst and Lofoten ridges (Blystad et al., 1995). The exposed basement rocks exposed on the central Lofoten Archipelago are mostly represented by remnants of the Caledonian allochthons which were overthrusted onto the Precambrian rocks of Baltica during the Scandian phase of the Caledonian Orogeny (Bergh et al., 2007; Steltenpohl et al., 2011; Eig, 2012;). A striking feature of the continental shelf along the Lofoten–Vesterålen margin is the relatively thin sequence of Jurassic-Triassic sediments not only on structural highs, but also in parts of some of the sub-basins (Færseth, 2012). However, the NE-SW trending narrow Ribban and Vestfjorden basins contain a relatively thick syn-rift Lower Cretaceous sequences (Tsikalas et al., 2001; Eig, 2012; Hansen et al., 2012; Færseth, 2012; Hemstra et al., 2017). Structures in the shelf area off Lofoten-Vesterålen result primarily from major rift episodes in the Late Permian-Early Trias, Middle-Late Jurassic and Late Cretaceous-Palaeocene (Færseth, 2012; Bergh et al., 2007). A late Albian-Early Cainozoic phase was also documented by Tsikalas et al. (2001) and Hemstra et al. (2017).

According to geological and seismic data, Triassic and Jurassic sedimentary rocks of different origins are also present within the Ribban and Vestfjorden basins (Eig, 2012; Færseth, 2012; Hansen et al., 2012]. Presence of older sedimentary rocks is rather uncertain but the Palaeozoic sedimentary rocks can be still preserved within the deepest parts of the study area (e.g. Eig, 2012; Mokhtari and Pegrum, 1992). Deep refraction seismic data (Mjelde et al., 1996; Drivenes et al., 1984; Goldschmidt-Rokita et al., 1988; Kodaira et al., 1995; Breivik et al., 2017) and potential field models (Olesen et al., 1997; Tsikalas et al., 2005; Maystrenko et al., 2017) better constrained the deep structures of the Lofoten-Vesteralen margin. The Moho topography along the Lofoten-Vesteralen margin, the Moho varies generally from shallower than 11 km beneath the oceanic crustal domain to more than 48 km beneath the mainland. The investigated continental margin is characterised by a significantly varying depth to the Moho which is uplifted beneath the Røst Basin, the Utgard High and the Lofoten Ridge. The top basement depths vary from 1 km to more than 10-12 km in the Vestfjorden Basin (Maystrenko et al., 2017). The pre-Cretaceous and Cretaceous structural levels of the Røst Basin, to the west, are not clear due to a presence of breakup-related intrusives and extrusives in this area (Berndt et al., 2001) but might be present according to modelling (Maystrenko et al., 2017). West of the Utrost Ridge, a 2 km-thick LCB is also confirmed in the necking zone adjacent to the COT (Breivik et al., 2017). To the west, the syn-breakup flood basalts have been earlier identified by SDR along reflection seismic lines in the study area (Tsikalas et al., 2002). The Lofoten-Vesteralen margin is also separated from the Voring margin segment by the Bivrost Lineament, which separates this marginal segment from the deeper and wider Cretaceous Vøring Basin and the Triassic-Jurassic Trøndelag Platform in the south (Blystad et al., 1995). The Bivrost Lineament is also recognisable at deep crustal and possibly mantle levels (Olesen et al., 2002; Maystrenko et al., 2017).

### Central and SE Greenland margin

In the conjugate part of the Hatton Basin Faroe and Rockall plateaus, the rifted margin is dominated by the Palaeogene volcanic province associated with the opening of the NEA (e.g. Larsen and Saunders 1998). Extensive ODP program (ODP sites 914-917, 988-990) also provided direct information for the petrology, geochronology and geochemistry of the thick basalt and SDR formed during the volcanic margin development (Larsen and Saunders 1998; Storey et al. 2007; Brooks, 2011). The nature of the sub-basalt geology offshore SE Greenland remains poorly constrained and both ages and distribution of expected pre-breakup sedimentary basins are still unclear in most of SE Greenland. At ODP Site 917, steeply to subvertically dipping metamorphosed sandstones were recovered that are stratigraphically overlain by a thin layer of undeformed sandstone and the oldest Palaeocene basalts found in SE Greenland (62–61 Ma; Sinton and Duncan 1998). The metasedimentary sedimentary rocks and overlying sandstone were devoid of nanofossils and could not be properly dated but were both inferred to be Upper Cretaceous–lower Palaeocene by analogy to the Kangerlussuaq Basin further north (Larsen and Saunders 1998; Vallier et al. 1998; Brooks, 2011). The metamorphic succession could alternatively represent much older Caledonian metasedimentary rocks formed in a foreland setting (Fyhn et al., 2012). On the conjugate side, Seismic reflection data and shallow cores from the SE Greenland margin suggested that marine rift basins near Ammassalik could have formed (at least) by the mid-to Late Cretaceous as also suggested for the Hatton Basin (Gerlinks et al., 2017).

Earlier refraction experiments onshore SE Greenland showed a Moho variation from 39 in the southern Ketilidian terrane domain to 49 km north of the Ketilidian foreland-Archaean crust transition (Dahl-Jensen et al., 1998). The land stations modelling was modelled by assuming a three layers crystalline crust showing a lower crust with high Vp velocity of 7.4-7.5 km/s thickening towards the Archaean terranes. The subsequent Seismic Investigation of the Greenland Margin experiment (SIGMA) further constrained the crustal architecture, composition and volcanic productivity of the adjacent VPM (e.g. Korenaga et al. 2000; Holbrook et al. 2001; Hopper et al. 2003). The four SIGMA transects described in detail in the previous papers illustrate the lateral variation of the margin from a relatively thick (28-33 km) undeformed continental crust with velocities of 6.0-7.0 km/s near the landward end to a 30-15 km thick igneous crust within a ~150 km wide COT where SDR emplaced. A 15-9 km thick oceanic crust is expected towards the seaward end of the profiles (Korenaga et al., 2000). Within the COT, LCBs reaching Vp maximum of 7.2-7.5 kms/s have been observed in all model’s profiles and traditionally interpreted as breakup (and ‘plume’-related) underplating (Holbrook et al. 2001; Hopper et al. 2003).

Onshore SE Greenland, exposures of the Kangerlussuaq basin cover approximately 10000 km2 and also represents an interesting outcrop analogue for the deep-water VPM of the SE Greenland and NEA (Larsen et al., 1999). The preserved 1 km-thick Cretaceous-Palaeocene basin may continue below the Paleogene flood basalts along the Blosseville Kyst but the continuation of the basin into the offshore areas to the south and southeast is unknown. The basin consists of predominantly NW-dipping fault blocks bounded by SE-NW–striking normal faults. The basin fill is divided into two depositional mega sequences related to regional tectonic events and sea level changes (Larsen et al., 1999). The oldest sequence spans the late Aptian to the earliest Palaeocene with sea level rise in the late Aptian and maximum flooding in the late Albian–Cenomanian, followed by the sea-level highstand in the Late Cretaceous–early Palaeocene. A basin-wide unconformity related to regional uplift and basin reorganisation in the mid-Palaeocene. The overlying mid-to late Palaeocene and sediments deposited during the early sea-level rise before extensive volcanic deposits and continental flood basalts.

### NE Greenland margin

For many decades, the geology and structures of the conjugate NE Greenland margin mostly concerned and concentrated along the onshore sedimentary basins mostly exposed from the Jameson Land to the Wandel Sea Basin after significant Cainozoic uplift and exhumation of the NE Greenland margin (e.g. Banow et al., 2014). The sedimentary basin recorded and exposed at present day show a series of rotated fault blocks, defined largely by eastward-dipping faults developed during multiple rift phases since the latest Devonian-Early Carboniferous collapse of the Caledonides (see Surlyk, 1990; Price et al., 1997; Kelly et al., 1998; Whitham et al., 1999; Hartz et al., 2002; Henriksen et al., 2009; Parson et al., 2017 for more detailed Palaeozoic-Mesozoic tectonostratigraphic description of the onshore basins).

The existence of thick sedimentary successions offshore NE Greenland was first suggested on the basis of interpretation of aeromagnetic data (Thorning et al. 1982) and few seismic reflection data in the outer part of the volcanic shelf (Hinz et al., 1987). The most recent insights and description of the structural evolution of the NE Greenland continental margin are based on a very limited set of published and modern seismic data (Hamann et al.,2005; Tsikalas et al., 2005; Dinkelman et al. 2010; Petersen et al., 2015). The most recent data suggest that the sediments on the conjugate NE Grenland margin were deposited in two basins separated by a prominent basement high, the Danmarkshavn Ridge. To the west, the Koldewey Platform marks the transition from the deep Danmarkshavn Basin to the Caledonian basement outcrops observed onshore NE Greenland (Hamann et al., 2005; Henrikssen et al., 2009). Major basin forming faults are situated along the western flank of the Danmarkshavn Basin. The transition between the Danmarkshavn Basin and the Danmarkshavn Ridge is dominated by a series of west-dipping fault blocks, creating rotated blocks overlain by prograding Paleogene sediments along the west margin of the ridge (Tsikalas et al., 2005; Dinkelman et al., 2010). Very little is known about the expected Palaeozoic-Caenozoic stratigraphy offshore NE Greenland due to the lack of wells and sparse seismic datasets. However, a series of Late Palaeozoic to Caenozoic seismic mega sequences have been tentatively defined by Hamann et al. (2005) based on correlation with the conjugate mid-Norwegian margin, Barents Sea and comparison with the known geological evolution onshore Greenland. In the inner Danmarkshavn Basin, the maximum thickness of the basin fill is c. 15-17 km (Dinkelman et al., 2000; Funck et al., 2016), and the basin is thought to span the entire period between Devonian and Neogene. A profound unconformity separates the Devonian–Cretaceous section from the overlying Palaeocene and younger units (Hamman et al., 2005; Tsikalas et al., 2005; Berger and Jokat, 2008; Petersen et al., 2015). The presence of salt diapiric structures seen in the seismic sections remind the salt basins (Carboniferous-Early Permian mother salt) recognised in the Barents Sea (e.g. Nordkapp Basin) (Hamann et al. 2005; Rowan et al., 2016). In the outer Thetis Basin, Jurassic–Cretaceous sequences were acknowledged by Hamman et al. (2005). Older sediments remain speculative notably due to numerous sill intrusions most likely emplaced during the onset of breakup (Tsikalas et al., 2005). It is considered that the pre-Palaeocene sediments interpreted to the north and south of this area continue beneath the volcanic rocks and SDR that from along the COT (Hinz et al., 1987; Tsikalas et al., 2005; Quirk et al., 2014; Geissler et al., 2016). The eastern boundary of the shelf also includes a ‘marginal high’ in its southern part which gradually collapses northwards (Berger and Jokat, 2008).

\*\*\*\*\*DIETER COMMENTS ?????\*\*\*\*\*\*\*\*\*\*\*\*\*

Further south, the deep margin is still unclear but thick succession of sediments have been recognised in the seismic data offshore Liverpool Land and adjacent to the Devonian-Jurassic and intruded sediments exposed in the Jameson Land (Hamman et al., 2005; Guarnieri et al., 2015; Eide et al. 2017). In the part of the area underlain by continental crust, a thick Caenozoic succession is recognised and lies with angular unconformity on block-faulted and tilted sediments of undifferentiated pre-Palaeocene (Late Palaeozoic-Mesozoic age) (Hamman et al., 2005). Further north, the Wandel Sea Basin (Håkansson et al., 1991) is also partly exposed onshore and forms the northern limit of the NE Greenland margin, where a basin of at least 8 km was created as a pull-apart basin during late Mesozoic and early Caenozoic times (Dossing et al., 2010).

Integrated geophysical and geological studies have also revealed pronounced differences in the crustal architecture and tectonic activity of the NE Greenland margin north and south of 72°N coinciding approximately with the landward prolongation of Jan the West Jan Mayen Fracture Zone (e.g., Schlindwein and Jokat, 1999; Schmidt-Aursch and Jokat, 2005; Voss and Jokat, 2007; Voss et al., 2009). The northern terranes were affected by a combination of Caledonian, Devonian, and Mesozoic to Tertiary phases of deformation, whereas only Caledonian and Devonian to Tertiary tectonic phases can be distinguished in the southern area. Mesozoic extension also shifted to the east and preserved Devonian structures in the northern area, whereas the Devonian crust was further thinned and weakened south of the Jan Mayen Fracture zone related lineament (Voss and Jokat, 2007).

Refraction results from Voss et al. (2009) also document better the structure of a 120–130 km wide COT in the southern and central parts of the NE Greenland margin. This transitional crust is characterised by high seismic velocities in and thinned continental crust (6.6–7.0 km/s) above at thick underlying LCB (7.15–7.4 km/s). The maximum thicknesses of the deep LCB are 15–16 km with lateral extents of 190–225 km. This inherited and/or breakup-related LCB exist only north of the West Jan Mayen Fracture zone and its landward prolongation (Schlindwein and Jokat, 1999; Schmidt-Aursch and Jokat, 2005; Voss and Jokat, 2007). The Moho along the VPM shows a distinct topography within a broad and relatively thick (15-20 km) magmatic COT and rises from ?30 km to 11–14 km near the expected true oceanic crust initiating at C24 at least in the central and northern part of the NE Greenland (Voss and Jokat, 2009). Along the magmatic COT, the Vp velocity suggests a basaltic layer mixed with syn-rift sediments whose presence supports the interpretation of long term rifting and a highly extended transitional crust across almost the distal part of the NE Greenland margin (Voss and Jokat, 2009. Given this interpretation, the presence of ocean spreading anomaly C23–C24B closer to the West Jan Mayen Fracture Zone very unlikely, then supporting the possibility of a North to South propagation of the oceanic crust Between NE Greenland and the conjugate Lofoten Margin (Voss and Jokat, 2007; Voss et al., 2009) (see later chapter and discussion). The nature and location of the COB in the southern part of the NE Greenland margin is, however, arguable (Scott et al., 2001; Tsikalas et al., 2001; Geissler et al., 2016; Funck et al., 2016) (See also onset of breakup and buffer discussion). Further north, recent ION-GXT wide-angle seismic line also show that the SDR and its transition with the oceanic at the edge of the Tethis Basin is very sharp (<50 km) and develop at the edge of a continental crust which is still relatively thick (e.g. 15 km) before the sudden and progressive a thinning concomitant with the volcanic wedge formation and associated distal LCB (Dinkelman et al. 2011).

### The Jan Mayen Microplate Complexe

The term of `microcontinent´ was initially defined morphologically rather than genetically to describe both aseismic oceanic and/or continental up-standing bathymetric features (Heezen and Tharp, 1965). Later Scrutton et al. (1976) define a `microcontinent´ is term of preservation of pre-rift continental basement, whereas the crust around should be oceanic. As part of the pre-breakup rifted system of the NEA, the Jan Mayen microcontinent or Jan Mayen microplate Complex (JMMC) is traditionally considered as a piece of continental crust that was originally located between the Faroe Plateau and the Outer Vøring Basin. It finally separated from Greenland in the Late Oligocene (Auzende et al., 1980; Auzende, 1980; Nunns, 1983; Guðlaugsson et al., 1988; Gunnarsson et al., 1989; Gaina et al., 2009; Gernigon et al., 2015). Some contributions have highlighted the importance of the Norwegian-Greenland Sea spreading evolution to comprehend better the rifting and the isolation of a microplate such as the JMMC which forms an intriguing intermediate conjugate margin system in the NEA (Nunns, 1983; Scott et al., 2005; Unternehr, 1982 Gernigon et al., 2012, 2015; Breivik et al., 2012; Schiffer et al., 2017). The main structural elements of the JMMC include the Jan Mayen Ridge, The Southern Ridge Complex and the intermediate Jan Mayen Through (Gudlaugsson et al., 1988; Blischke et al., 2017). To the west the transition with the Iceland plateau is characterised by the Jan Mayen Basin (Kodaira et al., 2008).

Despite few geophysical experiments and few boreholes, the nature and origin of the JMMC is still uncertain and contreversial. OBS modelling suggests that the northernmost part of the ridge consists of Icelandic type oceanic crust, bordered to the north by anomalously thick oceanic crust formed at the Mohns spreading ridge (Kandilarov et al., 2015). The volcanic suite in Jan Mayen has been diversely interpreted as an isolated hotspot, as Icelandic plume material dispersed in the northern Atlantic, as the result of melting of a sub-continental lithospheric mantle, or as a result of the coincidence of a continental fragment in the prolongation of a spreading center (Tronnes et al., 1999; Elkins et al., 2016).

As part of the JMMC, the Jan Mayen Ridge was drilled during DSDP Leg 38 (sites 346, 347, 349). The borehole penetrated hemipelagic and pelagic sediments beginning from the middle Eocene but and did not reach the basaltic basement interpreted on seismic data (Gudlaugsson et al., 1988; Planke and Alvestdad, 1999). Recent seafloor gravity core and grab sampling profile interpretation at one of the JMMC ridges are suggesting the presence of both Caenozoic and Mesozoic sediments (Polteau et al., 2013), but whether the Mesozoic sediments are indigenous or not is still debated (Norwegian Petroleum Directorate, 2013). The nature of the sub-basalt substratum and crust in most of the JMMC is even more uncertain and is primarily based on geophysical data modelling and interpretation of Ocean Bottom Seismometer data in the regional context (Kodaira et al., 1998; Mjelde et al., 2007; Breivik et al., 2012; Kandilarov et al., 2012). According to the refraction data, the main part of the Jan Mayen Ridge should be underlain by continental crust. The crustal thickness, which attains 15–16 km against approximately 5 km of the adjacent oceanic crust, is consistent with this interpretation; its velocity section is comparable with that of the mid-Norwegian margin. Beneath the Caenozoic sediments (2.2?3.4 km/s) Mjelde et al. (2007) suggest the presence of Mesozoic sediments (4.0<Vp<4.7 km/s) and Palaeozoic sediments (5.0<Vp<5.3 km/s). According to the OBS data, the thickness of the upper crust (6.0<Vp<6.4 km/s) and lower crust (with 6.7<Vp<6.8 km/s) are estimated at approximately 3 and 12 km, respectively. The Jan Mayen Ridge is distinguished by the overall absence of the high-velocity lower crustal body (Kodaira et al., 1998; Mjelde et al., 2007) although high-velocity rocks (7.0<Vp>7.2 km/s) were recently recorded at the base of the crust (Breivik et al., 2012).

Compared to the eastern margin where SDR are observed (e.g. Planke and Alvestad, 1999), no substantial magmatic activity accompanied the formation of the western boundary of the Jan Mayen microcontinent (Kodaira et al., 1998). The abrupt reduction of the continental crust thickness (down to 3 km) and the 5-km thickness of the adjacent oceanic crust is comparable with similar features of magma-poor margins (Kodaira et al., 1998). The separation of the microcontinent is likely correlated with the widely developed middle Oligocene reorganisation of the NEA (Gaina et al., 2009). The southern part of the JMMC may consist of older continental crust affected by magmatic addition (Breivik et al., (2012). Highly extended continental fragments and oceanic crust may also be expected taking into consideration the huge amount of mid-Eocene-Oligocene extension expected in the southern part of the JMMC (Gernigon et al., 2015). Recent sampling suggests the precence of highly depleted tholeiitic MORBs acquired along the Jan Mayen Ridge that may support this hyposthesis (Haaga, master thesis 2014–Ref not available but mentioned in Bakke, 2017 )

# PRE-BREAKUP RIFT EVOLUTION OF THE NEA

\*\*\*\*\*\*\*\*\*\*\*\*\*\*Just to remind that both lithosphere and crust are not homogeneous\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*

Obvious but sometimes simple I have the feeling that numerical models become a new geological fact \*\*\*\*\*\*\*\*\*\*

## Pre-rift setting: A patchwork of inherited basement terranes

In the NEA, the different rifting phases leading to breakup developed along a complex patchwork of pre-existing Archaean, Precambrian and Caledonian inherited terranes and structures (Pharaoh, 1999; McKerrow et al., 2000, SEE ALSO CHAPTER INHERITANCE, THIS ISSUE). In that sense, we remind that both crust and lithosphere do not represent a relatively homogenous medium as often idealised and simplified by many conceptual and/or numerical models. The Precambrian basement of northwest Europe consists of an assemblage of the Laurentian Craton, Baltica and Avalonia, which were ultimately accreted during the formation of the Caledonian Orogen (Klemperer et al., 1991; Coward 1995; Pharaoh, 1999; Cawood and Pisarevsky, 2017). The geological record of the three main tectonic units indicates a long history extending through the Paleoproterozoic and Mesoproterozoic involving continental growth onto rigid Archaean cratonic ‘nucleus’ through convergent plate interaction and accretionary orogenesis. The oldest patchwork included Archaean rocks partly reworked during the mid-Proterozoic, juvenile Proterozoic Crust (accreted arc material), middle Proterozoic calco-alkaline igneous bodies (e.g. Transcandinavian Igneous Belt), Archaean or Middle Proterozoic reworked during the Late Proterozoic Greenville-Sveconorwegian Orogeny 1.14-0.9 Ga (Lehtinen et al., 1998; Bingen et al. 2006; St. Onge et al., 2009; Lorentz et al., 2016; Slagstad et al., 2017) and Juvenile Late Proterozoic crust (Pharaoh, 1999). In the Fennoscandian Shield, the Archaean to Proterozoic ages also coincides with a progressive change of the orogenic styles from ultra-hot, mix-hot to cold and modern collisional tectonics (Cagnard et al., 2011). In Greenland, preserved Archaean rocks (3200–2600 Ma), reworked Archaean basement (around 1900–1800 Ma ago) and Juvelile Paleoproterozoic rocks (2000–1750 Ma) are also part of the main Greenland basement province (Henriksen et al., 2009)

The Caledonian-Appalachian cycle in the NEA particularly reflects the closure of the Iapetus Ocean between Laurentia and Baltica, together with the Tornquist Sea between Baltica and Avalonia (Pharaoh, 1999; McKerrow et al. 2000). In Scotland and Ireland, the Caledonian orogenic cycle on the Laurentian margin comprises three main phases: an Early–Middle Ordovician (475–465 Ma) phase termed the Grampian Orogeny; a phase of Silurian (435–425 Ma) tectonism restricted to the Northern Highland Terrane of Scotland termed the Scandian Orogeny; and an Early Devonian (405 Ma) phase, termed the Acadian Orogeny (Chew and Strachan, 2014). The Scandian phase is well described between Norway and Greenland (Roberts et al., 2003, Gee et al., 2006; Gasser et al., 2014). The Scandinavian Caledonides consists of four levels of thrust sheets, transported eastwards onto the Fennoscandia platform. They are referred to as the Lower, Middle, Upper and Uppermost Allochthons (Gee et al., 2006). The Lower and Middle Allochthons are generally considered endemic to Fennoscandia before the Caledonian orogeny. The lower part of the Upper Allochthon has a disputed origin (endemic vs. exotic), while the upper part of the Upper Allochthon and the Uppermost Allochthon are regarded as exotic terrains with an Iapetus Ocean or Laurentian ancestry (Gasser et al., 2014). The Caledonian orogen of Northeast Greenland is also characterised by far travelled foreland-propagating thrust sheets that were originally derived from the Laurentian margin and translated westwards across the orogenic foreland (Higgins et al., 2008; Henrikssen et al., 2009). The deepest preserved level of the orogen is found north of Danmarkshavn where abundant Caledonian eclogitic enclaves occur in the Palaeoproterozoic basement gneisses and show younger eclogite facies metamorphism dated at 410–390 Ma, which led to discussion about the real end of the Caledonian orogen in the NEA (Gilotti et al. 2008).

Finally, the Variscan orogen (480 and 250 Ma) extended from the Caucasus to the Appalachian (Matte, 2001; Franke et al., 2017). It coincides with the closure of the Rheic ocean by collision of Avalonia plus Armorica with Gondwana. This event initiated at the end of the Carboniferous and mostly affected the southernmost part of the NEA region.

At a basin scale, pre-existing fault populations, old thrust, outcrop-scale faults and fractures network, produced by these regional scales orogenic events locally influenced the structural evolution of the NEA sedimentary basins during rifting (Doré et al., 1997; Holdworth et al., 1997). Major reactivated terrane boundaries in the NEA particularly include the Great Gen Fault, the Møre Trøndelag Fault Zone, the Shetland Spine Fault the original Caledonian Front and the Tornquist Line, which separates the Baltic and Eastern Avalonia terranes in the North Sea (Roberts et al. 1999; Pharaoh, 1999). Structures within the Baltica basement roughly rotate from NE–SW in the south to a more N–S orientation in the north (Doré et al., 1997). To the south the Variscan terranes underlies most of the Celtic Sea region in the south of the NEA. Basement structures within this region are broadly E–W to ENE–WSW. Such pre-existing heterogeneities could 1) reactivate during multi rift events (Daly et al., 1989; Duffy et al., 2015; Phillips et al., 2016; 2) act as nucleation or crustal buffer during stretching of the crust (Nirrengarten et al., 2004); and 3) localise and modify the regional and local stress field (Pascal et al., 2001), and/or 3) influence the migration of fluids or possible magma migration (Schofield et al. 2017).

At a deeper crustal and sub-crustal scale, the Iapetus Suture Zone is well recognised onshore Ireland and U.K (Landes et al. 1999; e.g.; BIRPS project, Snyder and Hobbs, 1999) and most likely extend towards the Northern North Sea east of the Shetland Platfom and Utsira High (Lundmark et al., 2014). Structures in the crust and upper mantle in the Danish North might also witness the trace of a south-dipping subduction associated with the Tornquist Suture Zone (Abramovitz and Thybo 2000). The most recent Caledonian subduction event associated with the Scandian phase of the Caledonian origin may suggest the westward subduction of the Iapetus crust (Roberts 2003, Gee et al. 2008) but evidence of this subduction zone in the form of a preserved slab has not been detected in the lithospheric mantle of the Scandinvian Caledonides but high velocity lower crustal body (Vp>7.0 km/s) may indicate the presence of old inherited lower crust (e.g. eclogitic terrane) along the eastern flank of the northern North Sea (Christiansson et al., 2000; Fichler et al., 2011) and the Mid-Norwegian margin and conjugate margin (Gernigon et al., 2004; Ebbing et al. 2006; Mjelde et al., 2013, Nirenagarten et al., 2016; Schiffer and Peterdersen, 2016). Recently, Schiffer et al. (2015) proposed that the Flannan reflector in Scotland (Snyder et al., 1990; Reston et al., 1993) once formed a contiguous eastward-dipping subduction zone with the remnant of an East dipping slab (possibly of Caledonian age) observed in the Central Fjord, East Greenland (SEE ALSO CHAPTER INHERITANCE).

\*\*\*\*\*\*\*\*\*\*Extra comments from Christian/Tony ?\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*

## Global plate models and progressive dislocation of Baltica-Laurentia

Plate tectonic reconstructions of the North Atlantic regions including the NEA are mostly constrained by identification of marine magnetic anomalies and major fracture zones that formed during the creation of the ocean crust. As defined by previous geodynamic reconstructions (Le Pichon and Fox, 1971; Pitman and Talwani, 1972; Rowley and Lottes, 1988; Roest and Sirastava, 1989; Scotese et al., 1991; Gaina et al., 2009; Golonka and Ford, 2000; Seton et al., 2012), the dislocation of the Pangea evolved in a progressive sequence of plate separation even between North American, Greenland and Eurasia and associated rifting events in the NEA. Despite remaining discussions about the controversial interpretation and meaning of the C and M-sequences magnetic polarity pattern along many controversial COT of the Central-North Atlantic margin (Russel and Withmarsh, 2003; Labails et al., 2010; Sibuet et al., 2012; Eagle et al., 2017; Nirengarten et al., 2017), different plate reconstructions and crustal investigations agree that North America separated from Africa to form the Scotian margin before magnetic chron M25n (155.8-155.5 Ma, Late Jurassic) with the first evidence of oceanic crust around 175-177 Ma (Aalenian, Middle Jurassic) possibly after a phase of mantle and/or lower crust exhumation between Morocco and Nova Scotia around 195-177 Ma (Pleichbaschian-Toarcian, Early Jurassic) (Sahabi et al. 2004; Tucholke and Sibuet, 2007; Kingelhoefer et al., 2016). North America separated from Iberia to form the Southern Newfoundland margin before, during or slightly after magnetic chron M0 (126.3-126.3 Ma, Late Aptian) (Tucholke and Sibuet, 2007; Nirengarten et al., 2017). Then, North America separated from Europe to form the Northern Newfoundland margin sometime before superchron C34n (~83.64-125.9 Ma, Late Aptian-Santonian). Progressively, the North America separated from Greenland to form the Labrador Sea before chron 33n (79.9-74.3 Ma, Late Maastrichtian) (Sirastava and Roest, 1999) but more recent reconstruction models, and the possibility to have exhumed mantle material before breakup, only consider seafloor spreading since chron 31 (~71.4-68.3 Ma) (Keen et al., 2017) or C26r-27n (62.2-59.2 Ma, Early Paleocene), the earliest undebated and steady state spreading anomalies (Oakey and Chalmers, 2012; Hosseinpur et al., 2013). Finally, it is also commonly accepted that true seafloor spreading between Greenland and Europe initiated shortly during or before chron 24n1n (53.0-52.6 Ma) and/or 24n3n (53.4-53.9 Ma, Early Eocene) (Talwani and Eldhom, 1977; Skogseid et al., 2000; Gaina et al., 2009; Gernigon et al., 2012; 2015). Before sea-floor spreading, the timing and duration of continental rifting is not unanbuigously constrained by the reconstructions themselves, since the identification of sea-floor spreading anomalies only dates the post rift formation of the ocean crust. However, the progressive opening of the Central-North Atlantic may more or less also coincides with the development of the main rifting episodes further north in the NEA and adjacent areas (e.g. The North Sea).

## Main Rifting phases in the NEA: Current Understanding and Uncertainties

Far before the opening of the NEA ocean, the early post-orogenic basins in the NEA (e.g. The Hornelen Basin, the Orkney Basin) earlier developed as large intra-continental half-graben system controlled by reactivated low-angle detachments restricted in the upper crust (Fossen, 2010). Following the Caledonian collision, the Greenland margin has also been subject to collapse and subsidence starting in the mid Devonian or younger since the orogenic welt possibly persisted into the Carboniferous (Gillotti et al., 2004). The Palaeozoic rifting commenced in the latest Devonian-Early Carboniferous in Svalbard and Bjørnøya, with subsequent significant rift phases in the mid-Carboniferous and the mid-to late Permian (Gudlauggson et al., xxx; Stemmerik, 2000; Gernigon et al., 2015).

In the NEA, subsequent rift episodes took place in mid-Carboniferous-Carboniferous Permian and Permian-Early Triassic periods (Roberts et al., 1999). When observed, these early phases were driven by a number of processes associated with the late collapse of the Variscan Orogeny, the southwards propagation of rifting emanating from the Arctic and the opening of the neo-Thethys (Skogseid et al., 2000; Redfern et al., 2010; Frizon de Lamotte et al., 2015). The complexity of the Permian-Triassic rift basins along the NEA passive continental margin is locally influenced by the inherited basement framework (Redfern et al., 2010). When recognised in the NEA, these include post-orogenic collapse basins, focused narrow rifts and wide, low-magnitude multiple extension depocentres. The Permo-Triassic strata in the NEA are mostly preserved and observed in the platform and shallow water areas of the NEA sedimentary basin. In the deeper and distal parts of the rifted margin, they are often buried and/or possibly missing or likely laminated underneath younger Mesozoic and Caenozoic sequences. Convincing evidence for Permo-Triassic extension and depositions have been recognised from Ireland to the mid-Norwegian margin (Müller et al. 2005; Stolfova and Shannon, 2009; Stocker et al., 2016) up to the Barents Sea and conjugate East Greenland margin (Surlyk, 1990; Gudlaugsson et al., 1998; Hamman et al., 2005; Smelror et al., 2009; Tsikalas et al., 2012). Permo-Trias rift events are observed onshore East Greenland since the Carboniferous (Surlyk, 1990; Guarnieri et al., 2017). In the shelf area along the Lofoten-Vesterålen margin there is few evidence of Late Permian-Early Triassic large-scale faulting but moderate fault extension is locally observed (Bergh et al., 2007). Nevertheless; it contrasts with adjoining areas where a Late Permian-Early Triassic extension episode became very important in the platform domain of the mid-Norwegian margin (Faerseth et al., 2012). In NE Greenland; Normal faulting may also have occurred during the mid-Permian as Surlyk et al. (1986) and Seidler et al. (2004) distinguished two Early Triassic rift events onshore East Greenland in the Early and the Late Griesbachian. As Pangea split  apart, Late Trias rifting in the NEA coincides with the rift climax throughout eastern North America and western Africa (e.g., Morocco) until the onset of magmatism (Central Atlantic Magmatic Province), SDR formation and breakup in Early Jurassic (McHone, 1996; Withjack et al., 2012; Labail et al. xxxx).

When drifting was underway in  the  Central Atlantic by  Middle  Jurassic  time, the  basins between Nova Scotia, Iberia-Newfoundland continued  to widen and deepen  in response to thermal  subsidence or  continued  rifting. Marine sediments filled the basins. Late Triassic–Early Jurassic rifting age is well established from in the Jeanne d’Arc Basin and East Orphan Basin (e.g. Enachescu et al. 2004). Progressing into the Early Jurassic, significant rifting in the northern NEA took place in a series of half-graben with the deposition of thick Early Jurassic and Middle Jurassic sequences within the Inner Hebrides (Stein and Blundell 1990; England et al. 1993) and potentially Upper Jurassic, although later uplift and erosion make this interpretation a bit speculative (Holford et al. 2010). It can not be excluded, however, that the classic deep-marine seaway proposed by traditional paleogeographic model suggesting a deep Rockall Basin and a continuous sea way already achieved at that stage (Ziegler et al. 1988). Rifting could have developed more extensively within the Inner Hebrides and Sea of Hebrides-Minch basin which could explain the unclear distribution of the Jurassic source rocks in the NEA (Scotchman 2001). In NE Greenland, rifting also proceeded during the Middle Jurassic (Håkansson and Stemmerik,1989) and mild tectonic activities is also recognised in the Halten Terrace (Brekke et al., 2000).

The Late Jurassic–Early Cretaceous period marks a profound kinematic and paleogeographic change in the entire NEA region. This period also coincides with the progressive establishment of stable seafloor spreading in the adjacent southern part of the North Atlantic/Bay of Biscay and northern Central Atlantic (Tucholke et al. 2007; Labails et al., 2010; Sibuet et al., 2012; Tugend et al., 2014). The spatial extent of primary rift activity eventually leading to early breakup on the southern margins propagated and extends laterally and progressively to the adjacent margin to the north. Meanwhile, Late Jurassic to Early Cretaceous rifting on the Grand Banks progressively affected the Labrador margin. Rifting and the extension of the Labrador Sea also started either in the late Jurassic (160 Ma) based on dating of coast parallel dikes in SW Greenland or the early Cretaceous (140 Ma), on the basis of distinguishing and dating syn-rift sediments from wells on both margins (Blakwill, 1990; Chalmer and Pulvertaft, 2001; Hosseinpour et al., 2013; Peace et al., 2017).

It is not clear whether Early–Middle Jurassic rocks, or potential syn-rift Upper Jurassic formations, are present (or preserved) in the deeper parts of the UK Rockall. Unlike the North Sea, where accentuated rifting activities (Nøttvedt et al., 1995; Errat et al., 1999) favoured the widespread development of oil-prone late Jurassic marine source rock (e.g. the Kimmeridge), the development of a major marine seaway along the Rockall Basin is still unclear. Source rock evaluation in the southern NEA rather point to the development of isolated rift basins (Scotchman et al., 2016).

The Late Jurassic–Early Cretaceous rifting axis migrated and initiated the West Orphan Basin (Enachescu et al. 2004) closely connected with the Irish basins (Rockall and Porcupine basins) at that period (Sibuet et al., 2007). Due to limited drilling in the Rockall Basin, there is still great uncertainty and debate regarding the timing and number of phases of extension associated with the formation of the entire Rockall Basin. The suggestions include events from: Late Palaeozoic (Smith et al. 1989); Permo-Triassic, Jurassic and Early Cretaceous (Knott et al., 1993); Jurassic–Early Cretaceous (Corfield et al., 1999); Jurassic–Cretaceous (Mackenzie et al., 2002); Triassic to Late Jurassic–Cretaceous (Shannon et al., 1999); Latest Triassic–Cretaceous (Roberts et al., 1999; Permo-Triassic, Late Jurassic and localised Early Cretaceous (Naylor and Shannon, 2005); Early Cretaceous (Musgrove and Mitchener, 1996; Doré et al., 1999); Late Cretaceous (Hanisch, 1984); Cretaceous to Caenozoic but with older pre-Cretaceous phases (e.g. Nadin et al., 1999). Plate reconstructions of Cole and Peachey (1999) suggest that the majority of rifting occurred in the mid to Late Cretaceous and continued into the Eocene but they do not exclude older phase of extension. For the Rockall Trough and in the regional context, the current understanding is, however, that the main rifting and thinning phase in the Rockall Basin initiated in Late Jurassic-Early Cretaceous (Lundin and Doré, 2011).

At the end of the Jurassic period, a continuous system of deep rift basins probably co-exist from the southern Rockall Through up to the western Barents Sea (Doré et al., 1999; Skogseid et al., 2000; Lundin and Doré, 2011). Offshore Norway, this significant Late Jurassic-Early Cretaceous rift episode appears to be shifted from the previous Permo-Triassic syn-rift sequences mostly acknowledged in the Platform and terraces area (Reemst and Cloething, 2000; van Wijk and Cloething, 2002; Færseth et al., 2012) but more uncertain and/or absent in large part of the distal margins. However, there are no consensus on how far and how continuous into Early/mid-Cretaceous time the Late Jurassic extension continued in the NEA (e.g. Doré et al., 1999; Tsikalas et al., 2012; Lundin and Doré, 2011). In the mid-Norwegian margin, the onlap of the lower Cretaceous sediments against the terraces and platform borders suggests that the entire unit rapidly infills the pre-existing rifted structures. This paleo-topography, highlighted by the Base Cretaceous unconformity, is interpreted to be either the main result of a limited mid-Late Jurassic-Ryazanian extensional rifting event (Færseth and Lien, 2002) or alternatively a protracted or semi-continuous Late Jurassic- mid-Cretaceous rifting phase (Pascoe et al., 1998; Gernigon et al., 2003; Tsikalas et al., 2012; Henstra et al., 2015; Theissen et all., 2017; Zastranov et al., in prep). Some authors also suggest separate and independent Late Jurassic-Valangian/Hauterivian and Albo-Aptian (Tsikalas et al., 2012) or Albian-Cenomanian (Hemstra et al., 2017) rifting phases.

This severe thinning episode (including a tectonic sagging phase) probably reaches a climax in Late Jurassic-earliest Cretaceous but probably slowed down and/or failed around mid-Cretaceous time (mid-late Albian?) in the mid-Norwegian margin (Gernigon et al., 2015; Theissen-Krah et al., 2017). Faulting and fault block rotation during Volgian–mid-Albian rifting is also documented in the Traill Ø Greenland region (Price and Whitham, 1997; Price et al., 997; Whitham et al., 1999). A decline in faulting in the Albian is indicated by observations of Albian–Cenomanian strata onlapping and covering degraded footwall in the Traill Ø (Parson et al., 2017). In the meantime, it seems that continuated extension took place or migrated in the Lofoten-Vestrealen segments with renewed phase of rifting during Late Albian-Cenomanian ?(Hemstra et al., 2017.) This middle Cretaceous rift phase had led to significant tectonic subsidence of the Riban Basin which prevented sedimentary systems sourced from the eastern rift hinterland to reach the North Træna Basin.

By mid-Cretaceous time, the major rift system configuration was set and the large Cretaceous sag-basins observed in the Rockall Through, Faeroe-Shetland Basin, and Møre/Vøring basins were all associated with a drastic thinning of the crust (Lundin and Doré, 2011). The regional thinning system also extended further north along the Lofoten and Trømso basins and ends up at the level of the Bjørnøya Basin (Barrere et al., 2009; Clark et al., 2014). In terms of crustal process, this mid-Mesozoic rifting event result in the drastic thinning of the continental crust observed underneath most of the sag-basin of the NEA. The thinning phase coincides with the relocation of the deformation from the platform towards the subsiding sag basin. Large uncertainties remain also about the precise thickness, the nature of the crust beneath these large sedimentary sag-basins and the mode of deformation involved (see super versus hyperextended discussion).

Before the onset of breakup, the new rifting event affects the entire NEA margins during the Late Cretaceous-Paelocene rifting (Doré et al., 1999). The Late Cretaceous-Paelocene rifting is particularly well studied and constrain in the mid-Norwegian and Lofoten-Vesteralen margins (Ren et al., 2003; Gernigon et al., 2004; Tsikalas et al., 2001). In the Faeroes, a concomitant rift axis could possibly underlie the thick Faeroes flow basalts (Dean et al., 1999; Lundin and Doré, 1997; Doré et al., 1999). To the south, the Late Cretaceous –Palaeocene rift axis was proposed along the Hatton Bank but its regional meaning remains unclear due to the presence of extensive Cainozoic lava flows.

\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*Extra comments from Martyn /Rockall/ Faeroes and Tony ????\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*

## Ultimate rifting stage in the NEA

### Rockall-Hatton region

To be completed…..(Input from Martyn ?)\*\*\*\*\* also a lack of data here to really write a lot

### Faroe-Shetland rifted margin

To be completed……Laurent ? ?)\*\*\*\*\*It is also due to a lack of data west and SE of the Faroe Plateay here, not to mention the sub basalt issue. Laurent can possibly write a few words about the paleostress evolution of the Faroe? (his PhD)\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*

### Mid-Norway margin

\*\*\*\*\*\*OK here, I should probably simplified\*\*\*\*\*\*\*

Flanked between the Vøring Marginal High and the regional sags, the outer part of the Møre and Vøring basins represents the adequate place to understand the late and pre-breakup evolution of the mid-Norwegian margin. In terms of possible active versus passive rifting discussion, there is now a large consensus and evidence that the onset of the last rifting episode leading to VPM formation, offshore Norway, was active in the Late Cretaceous and predated the onset of flood basalt and SDR formation by several million years. Earlier interpretations (Swieciki et al., 1998, Skogseid et al., 2000) were suggesting potential Palaeozoic-early Mesozoic ages for the syn-rift sequences observed on the North Gjallar Ridge very close to the SDR and COB. However, the North Gjallar Ridge was drilled by well 6704/12-1 in 1999 and confirms instead that the syn-rift wedges observed beneath the erosional base Tertiary unconformity are earliest Campanian to late Maastrichian in age (Faerseth and Lien, 2002; Ren et al., 2003; Gernigon et al., 2003). The main phase of active stretching and fault block rotation preceding the breakup in the mid-Norwegian margin (e.g. the pre-magmatic rift climax) dominantly develops after the deposition of the Nise formation (e.g. mostly during the Spingar and Tang depositions). During late Campanian to Palaeocene extension, steady rates of fault movement lead to increased fault block rotation in the Outer Voring Basin. Classic domino style normal faults, as a result of the rotation of fault-planes from steep to shallow through time, as well as major listric faults develop clearly during this period (Gernigon et al., 2004). The faulting between the north Gjallar Ridge and Fenris Graben close to the limit of the Inner flows are partly concomitant with onlap and pinch-out of the Turonian-Palaeocene sedimentary successions at the edge of the North Gjallar Ridge. The onlaps complex coincides with the outer flexure zone defined at the western edge of the regional sag basin. One of the main onlap surface underneath fits with an intra-Turonian Unconformity as upper Coniacian onlapping sediments unconformably overlain Turonian rocks reached by well 6705/10-1 (Norwegian Petroleum Directorate fact pages). This unconformity could mark the transition between the flexural proto-rift and the late rift stage preceding the volcanic margin formation. In most of Outer Vøring Basin, the Palaeocene-Eocene sequences are not any longer affected by significant faulting and major faulted blocks rotation except in the Hel Graben where thick syn-rift Palaeocene accumulation has been proved by drilling recently (Williams and Magnus, 2013; Lundin et al., 2013). Conversely, thin and poorly faulted sections of Palaeocene are locally preserved on top of North and South Gjallar ridges. In well 6704/12-1, Thanetian (pre-breakup) sediments directly drape the Base Tertiary unconformity and the Late Cretaceous rotated faulted block (e.g. Ren et al., 2003; Gernigon et al., 2004). Further south, the South Gjallar Ridge is a similar NE-SW elongated ridge well defined at the base Tertiary unconformity. The western part of the South Gjallar Ridge is covered by the Tertiary lava flows and prograding structure interpreted as lava delta (Planke et al., 2000) emplaced close to an Early Tertiary paleoshoreline that defines the Vøring Escarpment (Abdelmalak et al., 2016b). The structural evolution of the South Gjallar Ridge was earlier interpreted as relatively similar to the North Gjallar Ridge (Gernigon et al., 2001, 2003). In the meantime, the ridge has been drilled by well 6603/5-1S in 2010 to test the hydrocarbon potential of e Jurassic reservoirs originally expected in the so-called Dalsnuten prospect. The well 6603/5-1 S bottoms, however, in Lower Cretaceous clastics (Lange Formation) at more than 5 km in depth (NPD fact pages) with a top basement expected at less than 10 km underneath according 2D and 3D potential field modelling (Brönner et al., 2013). Below the base Tertiary unconformity, low faulting and syn-rift wedges are observed and accommodate the deposition of the Cretaceous sediments. Multiple faulting and low-angles are observed. Evidence of duplex and extensional horse accommodates the migration of the footwall faults. All the faults are cut at the base Tertiary level and also suggest a Palaeocene shift and focus of the deformation towards the proto-breakup axis as documented in the North Gjallar Ridge. Dominant west dipping faults and detachment controlled the Late Cretaceous-Palaeocene deposition of the recovered upper Cretaceous–Palaeocene successions.

From a structural and tectonic point of view, the North Gjallar Ridge has been interpreted as either a rollover above a deep-seated, SE dipping fault with a ramp-flat-ramp geometry (Bjørnseth et al., 1997) or a draping structure above an underlying basement block (Brekke et al., 2001). Walker et al. (1997) and Ren et al. (1998) involved major west dipping detachment roots on the middle crust. Instead, Gernigon et al. (2003, 2004) suggested that the upper Cretaceous faulted blocks of the North Gjallar Ridge represent a west-dipping rollover feature decoupled above mid-Cretaceous shales. This is interpreted to have acted as a decollement layers above deeper (Pre-Cretaceous?) and decoupled block features, partly obscured and intruded by numerous sill intrusions on top of the so-call T Reflection (Figure x). The T Reflection had some pre-breakup and pre-magmatic structural influence on the Outer Voring Basin (Gernigon et al., 2003; Abdelmalak et al., 2017) and coincides with the top of a mid-crustal dome characterised by underlying high velocity material (e.g. part of the disatl LCB). The rollover model predicts a westward thickening of Cretaceous-Palaeocene syn-rift sequences partly imaged underneath the inner and landward lava flows that are observed on the Vøring Marginal High. It also suggests that no coupling exist between the upper fault and detachment and the deeper structures. This is particularly visible at lower Palaeocene level, where transparent seismic facies, now tied to a lower Palaeocene clay-rich section in well 6705/10-1. Eastward the thick deep-marine upper Cretaceous successions also pinch-out along the eastern flank of the North Gjallar Ridge. The next major unconformity surface that marks the overall westward extension of the Upper Cretaceous succession is intra-Campanian in age. In well 6705/10-1 this intra-Campanian marker appears to be draped by an early upper Campanian condensed section, correlating with the top of the Nise formation on the Nyk High. Finally, the last erosional (and onlap) surface appears to be early Maastrichtian in age and does not necessarily coincide with the base of the Springar Sandstones (possibly late Maastrichtian on the Nyk High). Lundin et al., (2013) recently suggested that the North Gjallar Ridge could have been also predominantly extended during the Palaeocene, and that extension along the western flank of the Gjallar Ridge was possibly linked with the drastic Palaeocene extension phase restricted the Nyk High-Hel Graben area. We partly disagree with that. Due to related structural configuration and geometries of the pinchout and syn-rift wedges calibrated by wells 6704/12-1 and 6705/10-1, we believe that the North Gjallar Ridge has been primary affected by a major pre-Palaeocene faulting and deformation event before migration of the brittle deformation towards the proto-breakup axis (e.g. Gernigon et al., 2004).

Like, North Gjallar Ridge, the South Gjallar Ridge also shows a complex system of fault controlled in depth by the uplift of lower crustal material highlighted by the presence of the T Reflection. In the western part of the Vigrid Syncline, the late Cretaceous series also pinchout the South Gjallar Ridge which is progressively uplifted and eroded during the Latest Cretaceous-Palaeocene before the onset of magmatism and SDR formation to the west. Later onlap features also developed during late Palaeocene-late Eocene. Its also suggests that the South Gjallar Ridge stayed in high position during this period. Like North Gjallar Ridge, only minor fault reactivation during Palaeocene time is testified along the South Gjallar Ridge. In a regional context, most of the Palaeocene (brittle) deformation observed at upper crustal level seems to have focus in the Hel Graben and further west in the Fenris Basin, partly overlapped by the breakup-related lava flows. The remaining Palaeocene 'sag' sequences are weakly faulted but are, however, contemporaneous of the drastic thinning of the lithosphere and localisation of the deformation towards the proto-volcanic axis where the magmatic rift processes took place in Palaeocene-Eocene (See volcanic formation chapter)

### Central and SE Greenland

In both NE and SE Greenland, we have very limited indications about the pre-drift evolution of the basin. Sparse observations in SE Greenland indicate that faulting occurred prior to and during the initial stages of flood basalt extrusion (Whitham et al., 2004; Larsen and Whitham, 2005). In the Kangerlussuaq Basin, onshoreSE Greenland, Upper Cretaceous and Early Paleocene shallow-marine sediments are overlain by Late Palaeocene shallow-marine and fluvial deposits. The basin fill pass upwards to Upper Palaeocene– Lower Eocene flood basalts (Larsen & Whitham 2005).

Before breakup, the Paleogene the region experienced a phase of uplift. This uplift occurred in the late early Palaeocene and was associated with normal faulting (Higgins and Soper, 1981; Hamberg, 1990). This phase of uplift is younger than the main uplift previously described in the mid-Norwegian margin but was similarly explained by the thermal impact of a mantle plume that had impinged on the base of the lithosphere prior to continental breakup (Dam et al., 1998; Larsen and Saunders, 1998; Skogseid et al., 2000) prior to the extrusion of the volcanic rocks that buried the Late Palaeocene landscape, including deeply incised palaeo-valleys (Pedersen et al. 1997).

During the uplift, the depositional setting was strongly controlled by large syn-depositional NW–SE oblique-slip normal faults trending along the Nansen fjord and the Christian IV glacial valley, as well as the nearby Kangerlussauq fjord (Larsen and Whitham 2005; Guarnieri 2011). These faults also controlled the injection sites and the later deformation of Early Eocene intrusives (Guarnieri 2011, 2015). Major NE–SW fault (e.g. Sortekap Fault) further constrained the Palaeocene–Early Eocene depositional system in the area (Guarnieri 2011). The structural relationship between dikes and the fault together with the left-lateral reactivation of the SE-dipping Sødalen Fault is interpreted to be the result of oblique rifting (Guarnieri 2011) in Selandian–Thanetian time. Later Faulting in Eocene shows a different W-dipping vergence during the onset of breakup (Guanieri, 2015).

### NE Greenland

On NE Greenland, Price et al. (1997) and Parson et al. (2016) mentioned that renewed faulting activity occurred in the Geographical Society Ø region sometime between the late Campanian and Thanetian intervals. A rift event during this time interval is not unexpected as previously described in the Outer Voring Basin. it is also documented from Kangerlussaq in southern East Greenland (Whitham et al., 2004; Larsen and Whitham, 2005).

In the Wollaston Forland and Traill Ø areas the north–south-trending rift system is cut by NE–SW-oriented dip-oblique right-lateral faults belonging to an oblique rifting stage interpreted as Palaeocene (Guarnieri 2014). Guarnieri (2014) also confirms the presence of a compressional Eureka event in north Greenland also characterised by the development of a Late Cretaceous–Palaeocene foreland propagating fold and thrust system probably corresponding to coeval deformation in Svalbard and predating the opening of the NEA. Similar Palaeocene compressive stress obtained by inversion of fault-slip data was also recorded in the adjacent Lofoten area in Norway (Bergh et al. 2007) and the Faroe Islands (Geoffroy et al. 1994; Walker et al. 2011). The associated Eurekan event, further north was the result of the Palaeocene Labrador seafloor spreading with compressive intraplate deformation across the Greenland–Svalbard margin followed by NE Atlantic seafloor spreading in Eocene time with debated strike-slip deformation along the Svalbard margin and compression in the Canadian Arctic islands (e.g. Pipejohn et al., 2016 for a recent review of the Eurekean Orogen).

# REGIONAL IGNEOUS ACTIVITY IN THE NEA

## The North Atlantic Large Igneous Province

……………Few words/chapter about the Mid Cretaceous volcanic event Ireland/South Rockall area ???? \*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*There is this old story of a mid-Cretaceous micro LIP in the South Rockall-Labrador-Porcupine\*\*\*\*\*\*I do not know if this is still up to date ? Martyn ????

A major characteristic of the NEA is the catastrophic magmatic event leading to the so-called North Atlantic Igneous Province formed prior to, and associated with, the onset of the Early Eocene breakup. It has traditionally been considered to comprise the voluminous Palaeogene igneous rocks occurring at the conjugate NEA volcanic margins and in the West Greenland–Baffin Bay area (Upton, 1988; Saunders et al. 1997; Meyer, 2007; Hansen et al., 2009 and references therein). In the NEA, the main Early Palaeogene North Atlantic Igneous Province volcanic provinces include: 1) The Mid Norwegian margin, 2) the Faroe-Shetland region, 3) Rockall–Hatton area, 4) the Northwest British Isles (Saunders et al. 1997), 5) SE Greenland, 6) Central–East Greenland and 7) NE Greenland. Other peripheric magmatism occurred in the northernmost parts of Greenland (Kap Washington Group at c. 64 ±3 Ma, (Estrada, 2001) and in the Vestbakken Volcanic Province, c. 54 Ma (Faleide et al., 1993; Omosanya et al., 2016). Contemporaneous, smaller and more isolated parts of the North Atlantic Igneous Province are also shown in the GIFR and Iceland formed subsequent to the onset of sea-floor spreading in the area (Meyer et al., 2007). Exposed and submerged basaltic rocks of the North Atlantic Igneous Province extend roughly NE–SW for more than 2000 km from Greenland to the NW European rifted margins (Saunders et al. 1997). The extrusive rocks of the province cover a surface area of at least ~1.3×106 km2, while the extrusive and intrusive rocks of the NAIP are together estimated to comprise a volume of ?6.6×106 km3 (Eldholm & Grue, 1994).

## 4.2 British Igneous Province

Laurent ? Romain ? few words about the ultramafic intrusions Rhum, Mull, ect.....

## 4.3 Rockall-Hatton region

In the northern Rockall Basin, Archer et al. (2004) provides Ar/Ar ages of 63.3 ±0.5 and 64.2 ± 0.4Ma for the oldest offshore sills drilled so far in the Rockall Basin.

\*\*\*\*To be complete\*\*\*\*input from Martyn ?

## Age of volcanism Faroe-Shetland rifted margin/Faroe Plateau

Within the Faroe-Shetland region, volcanism resulted in the eruption of thick flood basalt sequences covering an area of at least 40000 km2. This volcanic sequence that crop out in the Faroes Islands is divided into three main basaltic formations that outcrop onshore and extend offshore above adjacent Mesozoic-Palaeocene rifted basins (Ellis et al., 2002; Passey and Bell, 2007; Passey and Jolley, 2009). In the Faroe Islands, more than 3 km of subaerial lava flows of tholeiitic composition is exposed above sea level (Waagstein 1988). The Faroe Islands Basalt Group has a gross stratigraphic thickness of ~6.6 km, where the basal ~3.4 km has only been proven in the Lopra-1/1A borehole (Waagstein 1988; Passey and Jolley 2009; Chalmers and Waagstein, 2006). All of the lavas are of tholeiitic affinity indicative of relatively voluminous mantle melting (Waagstein 1988). Holm et al. (2001) also suggests some continental contamination of the magma with underlying Precambrian basement (amphibolite facies gneiss). The Faroe Islands Basalt Group is subdivided into seven lithostratigraphic formations dominated by basalt lava flows and minor volcaniclastic (sedimentary and pyroclastic) lithologies (Passey and Bell, 2007; Passey and Jolley, 2009). The base of the Faroe Islands Basalt Group is represented by the ~1.1 km thick Lopra Formation composed of volcaniclastic lithologies, the majority of which are interpreted as hyaloclastites (Ellis et al., 2002; Passey and Jolley, 2009; Schofield and Jolley, 2013; Mudge 2015). The oldest drilled lavas erupted at 58.8 ± 0.5 Ma in between C26n-C25n accumulating at a rate >2 km/Ma, and came to a slow end at 56.4 ± 0.5 Ma in the beginning of chron C24r (Larsen et al., 1999; Waagstein et al., 2002). The Lower Basalt Formation is overlain by 10 m of coal-bearing sediments and 2 km of lavas, deposited in early C24r (Ellis et al., 2002).

A large part of the lava flow and associated hyaloclatics overlaps the pre-existing sedimentary basin expected around the Faroe Islands. Offshore, the Faroe–Shetland Escarpment marks the palaeo-shoreline–shelf transition where the early Caenozoic deltas are characterised by foresets of hyaloclastite–pillow breccias (Wright et al. 2012; Ellis et al., 2012). In the Faroe-Shetland Basin, numerous sills have been drilled and mapped with the use of modern 3D seismic data (Schofield et al., 2017). Limited radiometric dating of the sill complex has been undertaken but accepted dates clustering around 55–52 Ma (Passey and Hitchen, 2011), although sills as old as the Campanian (72.1–83.6 Ma) have been mentioned (Fitch et al. 1988; Schofield et al., 2017). Large volcanic centres and plutonic intrusion are also identified in the northern Faroe–Shetland Basin (Passey and Hitchen 2011, McLean et al., 2016). In the Ben Nevis area, McLean proposed a first phase of magmatism in Late Cretaceous-Early Paleocene that deforms the sediments before widespread flood basalt and hyaloclatites emplacement in Ypresian time (~56-55 Ma). Magmatism continued after breakup west and north of the Faroe Platform, as suggested by the thick Faroe-Iceland Ridge, which is often interpreted to be the track of the Iceland ‘plume’ (Smallwood et al., 1999).

## Age of volcanism Mid-Norwegian margin

On the mid-Norwegian margin, the pre- and syn- and post-breakup-related volcanism is mostly documented by numerous reflection and refraction seismic data (Berndt et al., 2001a; Breivik et al., 2006; Mjelde et al., 2009; White et al., 2008). The magmatism has been confirmed by scientific drill holes during the Deep Sea Drilling Project DSDP leg 38 (sites 338, 342 and 343), during the Ocean Drilling Program ODP leg 104 (sites 642, 643 and 644) and by a commercial drill hole (the Utgard borehole 6607/5-2). DSDP Leg 38 earlier penetrated the basalt on top of the Vøring Marginal High (sites 338 and 342) and at the base of the slope towards the Lofoten Basin (site 343) (Talwani and Udintsev, 1976). Kharin (1976) earlier reported late/middle Eocene basalt from site 343 and 342 to be of alkaline and sub-alkaline composition, and the basalt from site 338 as low-alkaline tholeiites. The basalts at site 343 are considerably younger (28.5±2 Ma) but high amount of alteration have changed the K/Ar data (Kharin et al., 1976).

The thick wedge of flood basalt typically interpreted as SDR and landwards flows have been subsequently drilled during the ODP Leg 104 on the Vøring Marginal High (Eldholm et al., 1989). The pulse of breakup magmatism was around 56.5–55 Ma ago (equivalent of magnetic chrons C25n-C24r) (Eldholm et al., 1989). The ODP Leg Hole 642E has been drilled at the edge of the Inner SDR and consider two main sequences including the volcanic Lower and Upper series (Eldholm et al., 1989). As further developed in Meyer et al. (2009) and recently updated and reviewed by Abdelmalak et al. (2016), the deepest and older Lower Series consist of peraluminous, cordierite bearing peperitic basaltic andesitic to dacitic flows interbedded with thick volcano-sedimentary deposits and intruded sills. The early magmatic product shows clear evidence of continental contamination in the very close vicinity of the volcanic flexure associated with the inner SDR wedge. The chronology derived from the magnetic polarity understanding of the Lower Series at Hole 642E is, however, complex due to widespread remagnetization effects (Schönharting and Abrahamsen, 1989). However, the entire Upper Series shows a dominant reversed magnetic polarity correlated within reverse magnetic polarity Chron C24r (~57.1 to ~53.9 Ma) following Gradstein et al. (2012) and Ogg et al. (2012). A normal polarity zone was measured deeper between 1113 and 1194 m.b.s.f and tentatively correlated with Chron C25n (57.01 to 57.6 Ma) (Schönharting and Abrahamsen, 1989; Eldholm, 1991). Several radiometric age determinations have been accomplished on different samples from the Lower Series to provide crystallization ages. LeHuray and Johnson (1989) obtained Rb-Sr ages of 54.5 ± 0.2 and 57.8 ± 1.0 Ma, for the top of the Upper Series and the Lower Series, respectively. While Taylor and Morton (1989) measured for the Lower Series samples Rb-Sr age of 63 ± 19 Ma. Sinton et al. (1998) reported 40Ar–39Ar incremental heating ages of the basal Lower Series rocks suggesting a crystallization age of 55–56 Ma. The Upper Series has been also linked to the uppermost nannoplankton zone NP9 and the NP10 zone indicating an age ranging between ~55 and 54 Ma (Boulter and Manum, 1989; Vandenberghe et al., 2012). The Lower Series has been linked to the nannoplankton zone NP9 and the lowermost NP10 zone (Boulter and Manum, 1989) having ages ranging between 57.3 and 55 Ma. Boulter and Manum (1989) also recorded the species Apectodinium augustum, a species that is now considered to be diagnostic of the Palaeocene-Eocene Thermal Maximum currently dated at ~56 Ma ago (Abdelmalak et al., 2016).

On seismic, the oldest evidence for Palaeocene magmatism is observed close to the Voring Transform the transform margin, where post-Danian lava flows can be seismically observed with the Palaeocene successions and related to the embryonic crust formed in the COT of the outer More Basin. These lava flows could correlate with the first pre-breakup magmatic event of the North Atlantic Large Igneous province expected around 62-57 Ma before a second breakup phase of magmatism at c.56-54 Ma (Saunders et al., 1997; Storey et al., 2007).

In 2014, the NPD conducted shallow drilling operations at extreme water depths of 2130 m in the northernmost part of the Møre Marginal High, in the easternmost part of the Jan Mayen Corridor (Fig. x). The resulting drill hole, 6403/1-U-1 recovered 12 m of sediments and 38 m of igneous rocks (lava flows, lava breccia and hyaoclatites), in which the drill hole terminates. No age datation is published yet but palynological analysis of one sediment sample from the base of this overburden suggests an Early Eocene (Early Ypresian) age (55 Ma) (Bakke, 2017). The cored volcanic succession has relatively uniform petrography and geochemistry, and it is exclusively composed of tholeiitic basalts interpreted to originate from a mantle source more depleted than typical N-MORB sources (Bakke et al., 2017). Further geochemical modelling of rare earth elements, and of Nd and Hf isotopic compositions, along with enrichment in selective incompatible elements, provide evidence that the succession has also experienced crustal contamination (Bakke et al., 2017).

Landward of the volcanic province, a regional sill complex of the Møre and Vøring margins dominantly affects the outer part of the Møre and Vøring Basin (Brekke et al. 2000; Planke et al., 2005). Sill intrusion and their stratigraphic correlation with the explosive vent features suggest that most of the sill complex observed in the mid-Norwegian margin coincides with the period of breakup-related volcanism described in the Møre and Vøring Marginal highs. Only few sills have been penetrated by industrial boreholes on structural highs (e.g. Gjallar Ridge and Utgard highs). Recent radiometric dating gave U-Pb zircon ages of 55.6±0.3 Ma for the Upper Sill and 56.3±0.4 Ma for the Lower Sill drilled on the Utgard High (Svensen et al., 2010; Neumann et al., 2013), showing that theses proximal sills were concomitant with the early stages of the volcanism.

The easternmost evidence of magmatism is also characterised by the Vestbrona nephelinites intrusions (Torske and Pretsvik, 1991) located in the platform domain and in the landwards prolongation of the Jan Mayen corridor. They represent the most proximal breakup related-magmatic product of the rifted margin. These igneous rocks of the Vestbrona Formation have previously been interpreted as either igneous plugs or volcanic flows. New data indicate that relatively small sill complexes are abundant. New Ar-Ar data suggest that these sills are 1–2 Ma older than breakup (approximately 57–58 Ma) (Hafeez et al., 2017).

## Central and SE Greenland

In Central and SE Greenland the lavas were erupted in two major episodes: a Palaeocene (62–57 Ma) pre-breakup episode and an Eocene (56–55 Ma) syn-breakup episode (Saunders et al. 1997; Hansen et al. 2002; Storey et al. 2007; Brooks 2011). Minor episodes of post-breakup volcanism took place at 53–44 Ma (Tegner and Duncan 1999; Larsen et al. 2013) and numerous intrusions extended the activity to around 35 Ma, as reviewed by Tegner et al. (2008) and Brooks (2011). The youngest volcanic event took place in the central Blosseville region at 13–14 Ma (Storey et al. 2004).

Onshore Central East Greenland, the igneous and volcanic activity is particularly exposed along Blosseville Kyst and inland areas between 68°N and 70°30'N and comprises pre-, syn-, and post-breakup lavas (Storey et al., 2007; Larsen et al., 2014). The pre-breakup lavas (e.g. Lower Basalts) are exposed in Central East Greenland and are dated at 61.9–58.1 Ma (C26r-C25r) (Hansen et al. 2002; Storey et al. 2007; Larsen et al., 2014). The most voluminous burst of volcanic activity is expected to be syn-breakup and was dated at c. 56–55 Ma, C24r. The syn-breakup succession comprises the extensive and voluminous Plateau Basalts, which cover around 65000 km 2 around the Blosseville Kyst. The exposed flood basalts have a thickness >7 km (Pedersen et al. 1997; Larsen et al., 1999; Brook, 2011) and are dated within the narrow interval 56.4–55.3 Ma (C24r) (Storey et al. 2007). Post-breakup lavas occurs in Prinsen af Wales Bjerge (55.2–52.8 Ma (C24r-C24n1n), Peate et al. 2003), offshore 68°N (48.7 Ma, C22n, Thy et al. 2007), and at Kap Dalton (49– 44 Ma, C22n-C20r; Larsen et al. 2013). Central intrusions and dikes are abundant in Central East Greenland and ages associated with the North Atlantic Igneous Province shows a wide interval from 56.9-55.8Ma (Hirschmann et al. 1997; Tegner et al. 199;). Post-breakup intrusions are mostly 50–47 Ma, C22n-C21n (Tegner et al. 2008) but also show younger ages between 35–36 Ma, C17n1n-C16n) (Tegner et al. 2008). West of Blosseville Kyst, Storey et al. (2004) also report a very young age of only 13.5 Ma (C5AA). The most recent review confirms that the total range of ages along the central part of the East Greenland margin is 61.9–35 Ma. Minor activity during the periods 53–51 Ma (C24n1-C23n2n) and 45–40 Ma (C20r-C18n2n) are also reported (Larsen et al., 2011).

In the SE Greenland volcanic successions associated with the onset of breakup (e.g. SDR) is proved by ODP borehole 917A, 989B, 990A and SEG58 acquired along the VPM (Larsen and Saunders, 1998). Published radiometric suggest ages between 63.7 and 53 Ma, C28n-C24n1n (Sinton et al. 1998; Tegner and Duncan 1999; Larsen et al., 2014). Post-breakup and younger lavas were also encountered in ODP Holes 988A and SEG01 and were dated at 50.2 and 50.3 Ma, C22r (Tegner and Duncan 1999; Thy et al. 2007).

## 4.7 NE Greenland margin

In NE Greenland, sills, plutons, and plateau basalts intruded into and extruded onto or close to the exposed Mesozoic basin (e.g., Larsen et al., 2014). In Jameson Land, Traill Ø and most of Geographical Society Ø, sills and dikes are abundant and affect the sedimentary basins exposed. Plateau basalts are found on eastern Geographical Society Ø and are as much as 150 m thick and consist of tholeiitic lava flows with rare volcaniclastic layers. By analogy with the plateau basalts in Central E Greenland, the basalts probably have a latest Palaeocene to earliest Eocene age (Jolley and Whitham, 2004; Larsen et al., 2014). On Jameson Land, a sill and a dike were dated at 52.7 ± 1.2 Ma and 53.3 ± 1.4 Ma by Hald and Tegner (2000). Doleritic sills and dikes as thick as 300 m is also found throughout the strata of the Traill Ø region, and increase in abundance from west to east (Parson et al., 2017). In eastern Geographical Society Ø, dolerite sills compose as much as 40% of the structural thickness of Cretaceous mudstones (Price et al., 1997; Parson et al., 2017). Most dolerite intrusions have a tholeiitic composition and a minority have an alkaline composition (Price et al., 1997). The tholeiitic dolerites were emplaced ca. 54 Ma and have been related to the opening of the NEA in the early Eocene (Price et al., 1997; Larsen et al., 2014). Larsen et al., 2014 published ages of 55.1–52.6 Ma for two sills on Traill Ø and Geographical Society Ø and support the notion that most of the tholeiitic sills in this area were intruded during the lava extrusion. The alkaline dolerites have intrusive ages of ca. 36 Ma (late Eocene) (Price et al., 1997). For the lava succession on Hold With Hope, Upton et al. (1984b,1995) published some K–Ar and 40Ar– 39Ar ages of 58.7 and 53.4 Ma, and three dikes from aregional swarm cutting the lavas gave ages between 56.7 Ma ( 40Ar–39Ar) and 48.3 Ma (K–Ar).

Post-breakup intrusions on Central and NE Greenland margin. They come in age groups of c. 49 Ma, 40–35Ma and 30–25 Ma (Tegner et al. 1998, 2008; Larsen et al. 2013). The Kap Simpson and Kap Parry syenite plutons form the two eastern promontories of Traill Ø (Price et al., 1997). K/Ar and apatite fission track analyses from the syenite plutons give approximate cooling ages of ca. 35 Ma (Thomson et al., 1999). The large central complexes of Kap Simpson and Kap Parry were dated at 40-38 Ma by Rex et al. (1979). Syenites in the Werner Bjerge complex gave a Rb–Sr age of 31 ± 2 Ma (Rex et al. 1979), whereas a very recent Re–Os age of 25.8 ± 0.1 Ma was also found (Brooks et al. 2004). The Kap Broer Ruys central complex gave an age of about 47 Ma (K–Ar), and two late intrusions cutting the Myggbukta central complex gave ages of 32.7 and 34 Ma ( 40Ar– 39Ar and K–Ar). The Werner Bjerge complex in northern Jameson Land apparently spans a range of ages between 30 and 25 Ma (Rex et al. 1979; Brooks et al. 2004). During this time, the Aegir Ridge was becoming extinct and the Kolbeinsey Ridge was propagating northwards past the Blosseville Kyst and Jameson Land.

The Kap Simpson and Kap Parry intrusive complexes are also part of the Traill Ø-Vøring Igneous Complex (e.g. Olesen et al.: 2007; Gernigon et al., 2009). Gernigon et al. (2009) proposed that the Jan Mayen Fracture Zone has acted as a leaky transform along which magmas have been produced intermittently since breakup and until now.

Offshore NE Greenland younger magmatism possibly continued at local seamounts away from the spreading ridge, as is seen in recent time in Jan Mayen (Trønnes et al. 1999), Vesteris (Haase and Devey 1994) and the Eggvin Bank (Mertz et al. 2004; Tan et al., 2017).

\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*Extra comments/Reference from Malcom and/or possibly Romain????\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*The main points about the Petro\_Geochem(Very short !)

# VOLCANIC MARGIN FORMATION AND THE ONSET OF BREAKUP: MECHANISM AND PRESENT KNOWLEDGE

\*\*\*\*\*extra comments from Laurent ? and Dieter ???? Romain ?\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*

\*\*\*\*\*it is certainly crucial to remind what is a volcanic margin\*\*\*\*\*\*\*A very important trsanition of the NEA system\*\*\*\*\*\*\*I personally believe that recent papers have forced to much the magma-poor model in the NEA recently. But think are locally very different or not supported by the data!!!! Of course be free to argue or disagree \*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*

## Volcanic margin: a strict definition.

The onset of breakup between Baltica, Greenland and the proto-JMMC resulted in the formation of conjugate VPM during Late Palaeocene-Early Eocene time (Eldholm et al., 1989; Skogseid et al., 2000). As part of the North Atlantic Igneous Province, their formation represents the ultimate stage leading to the progressive and important transition between continental and true oceanic lithosphere. The VPM along the NEA is well documented by geological outcrops and offshore geophysical data from the Lofoten-Vesteralen margin, the Vøring Marginal High up to the Faroe Platform and Hatton Bank in the U.K. /Irish offshore region (Planke and Alvestad, 1999; Skogseid et al., 2000; Berndt et al., 2001; Breivik et al., 2006; White et al., 2008; Passey and Jolley, 2009; Mjelde et al., 2009; White and Smith, 2009; Voss and Jokat, 2009; Abdelmalak et al., 2016; Breivik et al., 2017 and are also described in the Conjugate SE and NE Greenland (Hinz et al., 1984; Karson and Brooks, 1999; Hopper et al., 2003; Korenaga et al., 2000; Quirck et al., 2014). Along their associated COT significant breakup magmatism is clearly and typically demonstrated by the thick seaward dipping reflector volcanic sequences (SDR), magmatic intrusions and LCB with high Vp-wave velocities that are (partly) interpreted as magmatic underplating (Mjelde et al., 2009; White et al., 2008. It is also important to notice already that in the NEA, the VPM (strict sensus) appear to be off the axis compared to the main Late Jurassic-Early Cretaceous rift axis. To some extent, the VPM also formed sub-parallel to the axis of the old Caledonian orogen and realted structures (SEE Christain/Tony review chapter).

The terminology of VPM (or volcanic rifted margin) has often been used to distinguish this specific passive margin (or passive rifted margins) from so-called non-volcanic margins (White et al., 1992, Menzies et al., 2002). Terms like called magma-dominated or magma-rich margins versus 'magma-poor' have also been suggested for their classification since traditional 'non-volcanic' margins exhibits moderate amount of magmatism (Reston and Manatschal, 2011). However, the primary differences between magma-poor margins and VPM, particularly characteristic of the NEA conjugate margin is particularly the subsurface expression of the magmatism emplaced during the rifting (e.g. the volcanism). The presence of thick volcanic and/or volcanoclastic dipping SDR wedges unit (up to 5-10 km) is perhaps the most spectacular and explicit seismic features commonly used to already classify such passive margins end-member as VPM in the early stage of their discoveries (Roberts et al., 1979; Hinz, 1981, Mutter et al., 1982). Compared to 'magma-poor' or non-volcanic margins (e.g. White, 1992), the volume of erupted volcanism produced during the onset of breakup on VPM is more significant (e.g. Eldholm and Grue, 1994; Menzies et al., 2002). Even if the precise volume estimations of magma products at VPM are made difficult due, notably, to incertitudes in the volume of magma emplaced in the crust, the rate of drastic magma production during VPM formation in the NEA but is often estimated between 1 and 10 km3/yr (Eldholm and Grue, 1994; Eldholm, 2000). By comparison it would exceed 25% of the total oceanic ridge annual production (Eldholm and Grue, 1994). The magmatic and volcanic production is so important at regional scale that VPM is often classified as Large Igneous Provinces (Mahoney and Coffin, 1995; Coffin and Eldholm, 2001; Ernst, 2014). Over the years, field observations of exposed active and /or fossil VPM have also contributed to understand better the magmatic and structural development of the SDR and related dike swarms (Karson and Brook, 1999; Svenningsen, 2001; Geoffroy, 2001; Klausen and Larsen, 2002; Abdelmalak, 2015).

## 5.2 Volcanic margin and volcanostratigraphic sequence evolution

Deep Sea and Ocean Drilling programs in the NEA volcanic margins (DSDP Legs 38 and 81; ODP Legs 104 and 152; Eldholm et al., 2000) earlier proved that the characteristic SDR seismic wedges observed along the transition between continental and oceanic crust consistent of extrusive/intrusive basaltic rocks of much greater thickness compared to the typical Penrose oceanic layer 2A (Hinz, 1987). Planke et al. (2000) argued that the arcuated, diverging reflection patterns of the SDRs are geometrically related to numerous and thicker lava flows dipping towards the proto-oceanic ridge axis where the largest accommodation space was obviously created. Despite different inheritance and rift history, SDR are common seismic features recognised along all VPM worldwide (e.g. Geoffroy et al., 2001; Skogseid, 2011; Clerc et al., 2017; Stica et al., 2014; Patton et al., 2016; Franke et al., 2017) showing that these peculiar features are not only restricted to the NEA but also share some analogies with many other VPM on Earth. In a global review Skogseid (2001) and Menzies claimed that VPM may represent more than 50-75% of the passive margin around the world. Over the last 20 years, improved multichannel seismic data and studies in the NEA have also shown that the SDR are often parts of more complex volcanic systems that reflect different paleogeographic setting of the flood basalt emplacement increasing during the onset of breakup. These observations led to the definition and characterisation of a dedicated seismic 'volcanostratigraphy' and nomenclature (Planke et al., 2000; Berndt et al., 2001; Abdelmalak et al., 2016a). Further described and documented in the previous papers, the different seismic facies units defined and identified along the NEA VPM often include: (1) the Lower Serie Flows, (2) the Landward Flows, (3) the Lava Delta, (4) the Inner Flows, (5) the Inner SDR, (6) the Outer High and the (7) the Outer SDR. Such seismic facies succession represents the typical and complete VPM sequence used to describe the evolution of the breakup extrusive complex observed very close to the first and continuous magnetic seafloor spreading anomalies (Berndt et al., 2001). In the NEA, it seems that the recognition of specific volcanostratigraphic units often agree shallow water to subaerial paleo-environments. Based on subsidence analysis constrain the ODP Sites 915, 916, 917 and 990. Hopper et al., (2003) suggested that the rift system from which the basalts along the SE Greenland VPM erupted at ~600 m above sea level. Recent ODP well 642E reevaluation in the Vøring Marginal High shows that the early magmatic expression represents shallow intrusions, peperites, or hydrovolcanic deposits generated by the shallow to subaerial interaction of magma and wet sediments (Abdelmalak et al., 2016). Continued flood basalt volcanism leads rapidly to effusive, subaerial volcanism where lavas in-fill the pre-existing volcanic basins also intruded by sill and dikes (e.g. Svensen et al., 2010). When the lava flows reached the shoreline, it fragmented in contact with water leading to foreset bedded lava delta development at the landward edge of the Landward Flows (Abdelmalak et al., 2016). Underlying the Lava Delta, the adjacent thin Inner Flows would represent an aggradational bottom set consisting of volcaniclastic sediments mixed with pillowed and massive subaerial lava flows (Berndt et al., 2001; Planke et al. 2000). Their lateral distribution can be partly controlled by the pre-existing rift morphology. During the onset of breakup in the NEA, the Inner SDR constructed by the rapid infilling of the proto-magmatic rift, and accommodation space required to explain the growing volcanic wedge development. It would result from a joint combination of tectonic forces and/or loading by the thick basalt pile (Geoffroy et al., 2001; Quirk et al., 2014; Buck et al., 2017). During tectonic and subsequent thermal subsidence, part of the lava flows in the Inner SDR could emplace in shallow water condition as observed in the Outer More margin. Locally, the eruption character could change dramatically, volcanic fissures could submerge to shallower marine conditions, producing (subaerial !) surtseyan material and mount-shaped features so-called Outer High (Planke et al., 2000). Additionally, tuffs produce during the catastrophic eruption of the Outer High will spread over large areas as recorded in many sedimentary basins (e.g. Balder and Tare Formations, offshore Norway). Subsequent subsidence of the injection centre to deeper water depths (>100-200m) would slow down the process of magma fragmentation but the lavas could have emplaced as deep marine flood or pillow to form the Outer SDR, partly overlapping the oceanic crust or heavily intruded (residual) continental crust. Similar geometries, with inward-dipping lava flows and outward-dipping dikes, occurs in the much thinner upper oceanic crust formed at relatively high spreading rates (Karson 2002).

## Seaward dipping reflectors in the NEA volcanic margin: structural and tectonic models

Earlier models and ongoing discussion about VPM formation during the onset of the NEA breakup also concerns the nature of the crust underneath the SDR wedges and their structural and tectonic mode of emplacements in the uppermost part of the continental lithosphere. In pioneer works, Hinz (1981, 1987) earlier suggested that the SDR emplaced over (highly) distended continental crust during the very latest phase of continental rifting. In their model of the Norwegian-Greenland Sea opening, Hinz et al. (1987) proposed that the initial intra-continental rift axis later became the area where the volcanic piles split and subsided. When dealing with the temporal and spatial distribution of the SDR, the subsequent model of Mutter et al. (1982) was slightly different and argued that the SDR formed at a subaeriallly emerged spreading axis during the earlier stages of oceanic accretion. In this model, the crust expected underneath the SDR was interpreted as oceanic. The traditional Mutter et al. (1982)´s model was based on possible comparison with 'similar' basaltic dipping wedges system also observed and in Iceland (e.g. Pálmason,1981; Helgason, 1985, Bourgeois et al., 2005). At that period, the isostatic Palmason’s model developed for Icelandic flexures was favoured to explain the SDR development at VPM. SDR were then interpreted as resulting from progressive oceanward flexuring of the lavas due to differential and progressive loading by the newly erupted flows.

Continental crust underneath the SDR was later favoured by Roberts et al. (1984) to explain the SDR drilled during DSDP leg 81 along the West Rockall-Hatton volcanic margin. Roberts et al. (1984) earlier emphasise the role of dike injection during crustal thinning and final breakup. A magmatic factor already noticed on the conjugate SE Greenland, where part of the SDR exposed onshore were already recognised to be directly underlain by relatively thick and preserved) upper crustal Precambrian gneiss injected by massive mafic dikes swarms and gabbroic/alkaline plutons (Nielsen, 1975; Nilsen and brooks 1981; Callot et al., 2001; Brook, 2011). Based on gravity modelling, Roberts and Ginsburg (1985) also argued that preserved stretched continental crust was potentially present underneath the dipping lava flows. In their tectonic and magmatic model, very rapid heating of the lithosphere was invoked as a cause of thermal weakening and rapid stretching. Accompanying dike intrusion which feeds the flows was also involved and increases in intensity in time towards the incipient spreading axis. Dike intrusion proceeds then to progressively higher levels through continental crust into the basalt pile leading to uplift and volcanic mount formation (e.g. the Outer High, later defined by Planke et al., 2000). At 100% dike intensity, the formation of the first oceanic crust was proposed by simple and progressive magmatic dilation/assimilation.

Subsequent geophysical and petrological works on VPM in the NEA and in many other worldwide VPM tent to agree now that the inner SDR most likely initiated in a continental or crustal domain poorly or moderately stretched by pre-magmatic rift event(s)(e.g. Skogseid and Eldhom, 1989; Gernigon et al., 2004; Eddy et al., 2014; Geoffroy et al., 2015; Patton et al., 2016; Guan et al., 2016; Nemcôk and Rybár, 2016; Clerc et al., 2017). Onshore SE Greenland the volcanic margin initiated at the edge of poorly streeched Precambrian basement (Karson amd Brook, 1999). Results of ODP legs 152 suggest the presence of pre-rift sediments and possible Caledonian metasedimentary rocks directly underneath the Inner SDR (Larsen et al., 1998). Petrology of the lava in ODP 642 drilled in the Vøring Marginal High also shows clear evidences of continental contamination (Parson et al., 1989; Meyer et al., 2006; Abdelmalak et al., 2016). Due to difficulties to map a clear COB when volcanism is involved (see previous chapter), some uncertainties remain about the nature of the crust underneath the Outer High and the distal Outer SDR (Planke et al., 2000) which could emplace partly or not in pure oceanic domain. In that case, they may locally represent the equivalent of the Icelandic 'SDR' even if the nature of Iceland remains unclear (see later and Chapter Iceland and GIFR from Gillian). Alternatively, the Outer SDR may not always be so late in the sequences. Instead of being the result of a progressive migration of the lava flows, some 'Outer SDR' could be the result of a dual magmatic scenario and the simultaneous formation of two distinct but overlapping Inner SDR (Fig. x), an alternative that may challenge the classic volcanostratigraphic sequence.

Tectonic and interpretation of crustal flexures observed in VPM has also a long time been recognised by field geologist working in East Greenland even before the first seismic acquisition (e.g. Wager and Deer, 1938; Nielsen, 1975; Faller and Sopper, 1979). Myers (1980) showed that the coastal dike swarm, onshore East Greenland, occurs in a series of zones arranged en echelon, similar to dike and fissure swarms in Iceland. Most dikes were intruded vertically before flexuring of Precambrian basement and overlying Cretaceous-Palaeocene sediments. The dike feed gabbroic plutonic intrusions emplaced during the onset of the NEA (Myers, 1980). In addition, subvolcanic complexes, dikes and extrusives belonging to the nephelinite/carbonatite suite are also found, possibly related to fractures or accommodation zones at a high angle to the continental margin (e.g. Karson and Brooks 1999). Most flexuring occurred after consolidation of the gabbros and was followed by the intrusion of linear and radial swarms of intermediate dike and ring dikes associated with the emplacement of syenite and granite plutons (Myers, 1980). Early dikes that fed traps, initially sub-vertical, are progressively tilted oceanward and later intruded by younger dikes feeding the SDR and intruding the tilted lava flow (Karson and Brooks 1999; Brooks, 2011). Further field study in West Greenland where the VPM is also exposed onshore shows that SDR development is also accommodated by coastal flexure and laterally injected dike but also by arrays of continent ward-dipping normal faults, rotated and progressively inactivated as the SDR wedge develops oceanward (Geoffroy et al., 2001; Callot et al., 2001).

The formation of SDR is also a sudden and localised phenomenon associated with a strong lithosphere necking (e.g. ultranecking) and a possible differential stretching of the lithosphere during the onset of breakup. Modern seismic reflection data at crustal scale show that a magmatic ultranecking of the crust is concomitant with SDR emplacement (Dinkelman et al., 2011; Guan et al., 2016; Nemcôk and Rybár, 2016). A possible consequence is the localisation and expression of the crustal extension during this critical magmatic phase. Both onshore and growing offshore seismic evidence have also pointed out that SDR remind rollover structure controlled by major continentwards detachment faults as initially suggested by Gibson and Love (1989) in the Voring margin. Not always clearly imaged in the NEA, only a limited seismic observations have suggested that the accommodation space required for SDRs rollover formation could be controlled by continentwards detachment faults in the NEA (e.g. Barton and White, 1997; Planke et al., 2000; Gernigon et al., 2006; Quirk et al., 2014). Few conjugate transect across other VPM (and possible analogues of the NEA VPM) show that some continent wards detachment faults could develop on both conjugate SDR (Guan et al., 2016; Nemcôk and Rybár, 2016) and show that the development of SDR and adjacent volcanic hemi-grabens and associated faul can rooted on top of deeper and major crustal scale continent wards shear zones (Stica et al., 2014; Pindell et al., 2014; Quirk et al., 2014; McDemott et al., 2015; Geoffroy et al., 2015; Clerc et al., 2015). Some of major detachment fault/shear zone can affect the entire continental crust and could even displace the seismic Moho (Nemcôk and Rybár, 2016). In light of the recent seismic observations on VPM, the dominant ductile shear controlling the SDRS seems to be often continent ward. Such crustal scale detachment configuration is important to understand the dynamic of VPM formation but difficult to explain in light of classic magma-poor rifted model suggesting (when they are clearly observed) systematic oceanic ward detachment (e.g. Geoffroy et al., 2015). The major discontinuities that control the development of SDR could be associated mechanically with large continent-ward shears at the brittle-ductile crust interface. The explanation for continent wards detachment formation along VPM is not straightforward. In the NEA, Quirk et al. (2014) recently suggested that the lateral flow of ductile crust towards the proto-ridge could explain the rollover geometry and the continent ward shear component. Based on sand-silicon modelling approach, Callot and Geoffoy (2001) and Geoffroy et al. (2001) have showed, however, that the presence of ductile lower crust underneath the SDR was not always enough to physically reproduce the development of major and synthetic continent wards detachment without an active but unclear dynamic contribution of the asthenospheric material. To explain the ambiguous sense of shear expected underneath the SDR at upper lithospheric and crustal levels, Geoffroy et al. (2001) suggested that an active and melting asthenosphere which is flowing upward and laterally oceanic ward could eventually drag along the overlying ductile lithosphere (ductile crust and mantle lithosphere) towards the breakup area, then controlling the SDR rollover. The counterclockwise lateral flow of the asthenophere required for such a dynamic model was, however, difficult to explain and simulate. Insights on the origin of the inner SDRs which are coeval with crustal (ultra)necking at VPM were recently highlighted by numerical modelling. Geoffroy et al. (2016) suggested that the strengthening of deep continental crust slightly after the early magmatic stages could provoke such enigmatic divergent flow of the ductile lithosphere on both sides of a central continental hanging-wall block (so-called C-block) that stay in a high relative position in between the conjugate VPM. Assuming the C-block hypothesis, crustal-scale faults dipping continent wards could root over this flowing material at the edge of the residual block. In this scenario, the development of crustal-scale conjugate detachments dipping outward with respect to the C-block would be the consequence of a thermally driven weakening of the mantle lithosphere, which partly flows outward and upward along the bottom of the C-block. A part of this process, high density axial mafic intrusions in the transitional crust would possibly increase the significant downward flexure of the opening rift since the amount of subsidence could increase with decreasing plate strength and heating during the final stage of continental breakup (e.g. Corti et al., 2105).

The detachment model for SDRS-related structure challenge early concepts of magma-tectonic model that predict that if low-viscosity magma is persistent then extensional stresses should not reach the level needed for lithospheric cutting normal fault (Rivalta et al., 2015). More recent discussion and models of SDR formation suggest that magmatic and volcanic loading alone can explain the shape of the volcanic wedges (Buck, 2017). These numerical models also simulate the effect of temporal changes in the locus of magmatic centers and the level of volcanic infill and can reproduce the observed down-dip thickening of the flows and the range of contrasting reflector dip angles often observed in conjugate VPM. However, this loading effect and the reholologic change expected in depth should indirectly trigger decoupling and expected continent ward dipping shear zone at the edge of the spreading axis.

## Magmatic plumbing of the continental crust and volcanic sedimentary basins

### 5.4.1 Sill complexes and volcanic vents

VPM are also characterised by numerous sills and dikes that affect the basement and the sedimentary basins succession at different stratigraphic levels. Growing seismic activity in the NEA helped to understand the deep plumbing system of the NEA (Berndt et al., 2000; Bell and Butcher, 2002; Thomson and Hutton, 2004; Planke et al., 2005; Hansen and Cartwright, 2006; Schofleid et al., 2017). Like the flood basalt emplaced along the proto-breakup axis, the volume of magma associated with sill emplacement in adjacent basins can be considerable. Extensive seismic acquisition in the NW European sedimentary basin of the NEA, over the last 30 years showed that major sills complexes in the as a whole form part of a much more extensive series of Late Cretaceous/Palaeogene aged sill complexes extends for ca. 1800 km from the Norwegian Margin to the Southern Rockall Basin (Skogseid et al., 2000; Planke et al., 2000; Magee et al., 2014; Schofield et al., 2017; Abdelmalak et al., 2017). In the NEA, important sill complexes can often reach several thousand km2 and individual sill can often reach 50-300 m in thickness (Planke et al., 2005). In the Utgard High, two dolerite sills emplaced in Upper Cretaceous mudstones and sandstones and have been drilled by well 6605/5-2 (Berndt et al., 2000, Newman et al., 2013). First sill was 91 m thick and drilling terminated 50 m into the lower sill, thus its total thickness remains unknown. The sills can be followed for more than 100 km westwards into the deeper part of the basin. In the mid-Norwegian margin, Svensen et al. (2002) estimated that the volume of magma due the sill complexes can reach 0.9-2.5×104 km3.

In the Faeroe-Shetland Basin similar mafic sills and dikes which are thought to have been intruded during the Paelocene-Eocene (Morton et al. 1988; Ritchie & Hitchen, 1996; Bell and Butcher, 2002). 149 sills have been drilled by exploration borehole showing thickness between 15 and 40 m (Schofield et al., 2017). Identification of magma flow directions through detailed seismic interpretation by Schofield et al. (2017) also demonstrates that the main magma input zones into the Faeroe-Shetland Basin were controlled primarily by the NE–SW basin rift structure of the Faeroes-Shetland Basin. The Faeroes-Shetland Basin sill complex has been estimated to cover a minimum area of at least 22 500 km2, extending from the southern part of the Faeroes-Shetland Basin to the Møre Basin in the NE (Passey and Hitchen, 2011; Schofield et al., 2015).

In the Northern Rockall Basin, extensive flood basalt lava flows, volcanic centres and sill complex of Late Cretaceous to Early Eocene age have also been described using seismic and borehole data (e.g. Thomson and Hutton, 2004; Archer et al., 2005). Similar igneous features identified in the central Rockall Basin are typically located within the Upper Cretaceous succession and have either a saucer-shaped or inclined sheet morphology (Magge et al., 2015). On the northeast Greenland margin, Reynolds et al. (2017) document two Paleogene-aged sill complexes >10000-18000 km2 in size.

Field observations and comparison with modern 3D seismic imaging show that sill intrusions in the NEA basin can have different geometries controlled by several parameters including 1) the density contract between the magma and the country rocks, 2) the tectonic stress field, 3) the overburden thickness, 4) the mechanical properties of the sediments intruded and/or 5) the presence of pre-existing mechanical heterogeneities (Galland et al., 2007; Polteau et al., 2008; Magge et al., 2013). In the sedimentary successions, sills and dikes may potentially act as barriers and/or carriers from fluid flow and hydrocarbon migration (Rateau et al., 2013; Senger et al., 2017). The emplacement of igneous intrusions into sedimentary basins can also mechanically deforms the host rocks, create differential compaction and therefore influence the basin geometries and fracturation (Galerne et al., 2011; Galland et al., 2009; Wilson et al., 2016; Schmiedel et al., 2017; Galland, 2012; Schmiedel et al., 2017; Omosanya et al. 2017).

Depending on the timing of emplacement and volume, sill complexes could have some significant impact on predicted heat flow history, local hydrocarbon maturation (Archer et al., 2005; Rohrman, 2007; Fjelskaard et al., 2008; Senger et al., 2017; Muirhead et al. (2017)). Therkelsen (2016) also show that sill intrusions and circulation of hot fluids can affect and deteriorate the reservoir quality of sandstones formation. Local thermal perturbations due to sill injection within the sediments can also causes hydrothermal activities and initiate explosive fluid escape leading to the formation of so-called volcanic vents often observed on seismic but also described in the field (Jamtveit et al. 2004; Planke et al., 2005). Svensen et al. (2002) concluded that the drastic release of methane associated with large hydrothermal vents complexes observed in VPM of the NEA may also contribute to large discharged of methane and climate change if they emplaced in carbon-rich sedimentary successions.

### Deep Lower Crustal Bodies

\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*Important for the NEA, where LCB=underplating for many\*\*\*\*\*

Gillian ? what about the LCB along the GIFR. Any alternative from the geophys ????? I assumed that this part wil be discussed in you chapter.\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*

The origin and precise chronology of the magmatism combined with the ambiguous meaning of the LCB underneath unambiguous continental crust remain primordial to understand and/or decipher between the true contribution of deeper asthenopheric versus crustal/lithospheric processes involved (or not) in the lithospheric rupture and melt production (Gernigon et al., 2004, 2006; Wangen et al., 2011; Fjeldskard et al., 2003; Becker et al., 2014; Petersen and Schiffer, 2016). Lower crustal zones with high density and high seismic Vp-wave velocity are also typical characteristic geophysical features of VPM (Mutter et al., 1982; Fowler et al., 1998; Mjelde et al., 2007; Bauer et al., 2000; Ravaut et al., 2005; White et al., 2008; White and Smith, 2009; Becker et al., 2014). In the literature, such atypical crustal velocity zones are often given the name of High-Velocity Lower Crustal bodies or Lower Crustal Bodies favoured in this paper (LCB) (Eldholm et al., 1995). Numerous refraction data acquired over the last 50 years have shown that the LCBs along the NEA rifted margins are not only restricted to the flood basalt/SDR domain but can also extend in the oceanic domain and continent wards even over long distance from the breakup axis (> 100 km) (O’reilly et al., 1996; Breivick et al., 2008; Kvarven et al., 2014; Mjelde et al., 2009, 2016; Voss and Jokat, 2009). To clarify some remaining ambiguities related to the LCBs interpretation on the mid-Norwegian margin, Gernigon et al. (2006) and Kvarven et al. (2016) suggested to distinguished dominant oceanic affinity LCB (OLCB) from more ambiguous continental LCB (CLCB). Traditionally, velocity cutoff Vp-wave values of 7.0-7.1 km/s are often used to delineate the top of such atypical velocity zones. Both OLCB and CLCB have P-wave velocities (and densities) greater than that of 'normal' oceanic and continental lower crust. In oceanic domain, OLCB show Vp-wave velocities which are assumed to be higher compared to 'normal' unaltered gabbroic oceanic crust velocities (layer 3: 6.8–7.1 km/s after Mooney et al., 1998). In continental domain, CLCB show Vp Values higher than 'normal' velocities lower continental crust (6.5<Vp<6.7 km/s after Holbrook et al., 2002).

In the recent years, some contributions have particularly questioned the timing and the nature of theses LCB and notably the CLCB that can extend outside the main volcanic province area. For example, it is likely that part of the CLCB, offshore mid-Norway influenced the margin architecture before the onset of magmatism and may represent old inherited and progressively exhumed lower continental crust (Gernigon et al., 2004, 2006; Ebbing et al., 2006; Mjelde et al., 2007; Kvarven et al., 2016; Abdelmalak et al., 2017). Understand if LCB are magmatic and breakup-related or not has direct impact on the total magmatic production estimation along VPM. Many seismic experiments have shown that variations in size and physical properties of the OLCB and CLCB can also hold important clues to mantle melting scenarios (Kelemen and Holbrook, 1995; Korenaga et al., 2002; Voss et al., 2009; Parkin et al., 2007; White et al., 2008, Wangen et al., 2011; Breivik et al., 2014). Alternatively, timing and a non-breakup origin of the LCB can contribute to assess the inheritance implication during the conjugate rifted margin development (Schiffer et al. 2016, Petersen and Schiffer, 2016).

As a general rule and old paradigm, LCBs along VPM are often thought to represent magmatic (gabbroic) intrusions and related cumulate layers emplaced in and/or underneath the pre-existing stretched continental crust during breakup time of the VPM (Furlong and Fountain, 1986; Kelemen and Holbrook, 1995; White and McKenzie, 1989; Eldholm et al., 2000; Voss and Jokat, 2007). This common assumption relies on the spatial and vertical relationship between the SDR, sill complexes and LCB. Consequently, Both CLCB and OLCB in VPM setting were often interpreted all together as the intrusive equivalents of the erupted lavas which form the SDR. This traditional 'underplating' interpretation of the LCB considers a massive mixing and accumulation processes of magmatic material near the base of the pre-existing continental crust (Fyfe, 1973 Dillon, 1988; Fyfe, 1992; Cox, 1993). Nowadays the term 'underplating' often coalesce two distinct processes, i.e., mafic intrusions within the pre-existing crust and basal accumulation of ponded mafic/ultramafic material (Mjelde et al., 2002; White et al., 2008; Hirsch et al., 2009; Schnabel et al., 2008; White and Smith 2009; Thybo and Artemieva, 2013). It includes additional crustal and magmatic process, which involves the production of new mafic crust within and underneath a pre-existing continental crust. In continental rift and/or VPM environment, underplating emplacement by crystallization of the injected magma at the base of the pre-existing crust led to a new Moho definition by differentiation of the melt into an upper gabbroic and a lower ultramafic layer during the thinning of the crust and lithosphere (Fyfe, 1992; Cox, 1993). Accumulated mafic layers with up to the thickness range of 10-25 km underneath the pre-existing crust were considered as possible (Furlong and Fountain, 1986). In pure oceanic domain, the terminology of overcrusting has also been proposed to explain atypical melt production and thicker oceanic crust adjacent to VPM (Gernigon et al., 2009). The terminology suggests clear distinction between specific syn-accretion (overcrusting) and post-accretional (underplating) magmatic accumulation/accretion.

Although the internal details and reflective characters of the LCB cannot always be imaged properly due to the lack of high-resolution refraction data, some integration of normal-incidence and high resolution wide-angle reflection seismic data, have recently demonstrated that part of the LCB can be seismically reflective (White et al., 2008; White and Smith, 2009) and/or could reveal elongated patches of high velocities zones interbedded within the pre-existing lower crust (Ravaut et al., 2006). Part of the CLCB may represent major sills complexes and/or bigger magma chambers intruding and/or melting pre-existing lower crust. Offshore Brazil, wide-angle seismic reflection underneath SDR, also reveals the presence of stacked prominent reflection symptomatic of deep intrusions (Pindel et al., 2014; Geoffroy et al., 2015). Eccles et al., 2011 and White et al., 2008 showed that the transition between OLCB and CLCB observed along the Faroes VPM can also be reflection-free; which may indicate the presence of more massive crustal intrusion and/or melt assimilation in the magmatic COT.

It has been commonly acknowledged that the highest Vp velocities of magmatic OLCB and CLCB could agree the presence of breakup-related ultramafic (picritic) magmas richer in Mg than classic oceanic MORB. They are often interpreted as the consequence of melting at anomalously high potential temperature and mantle plume related (Saunders et al., 1997; Thomson and Gibson, 2000). Seth (1999) and Meyer et al. (2007) show, however, that Mg-rich magmas do not necessarily characterise high-degree or high-temperature melts but could simply be explained by extensive decompression of an uprising mantle (active or passive) fallowed by later differentiation. Berndt et al. (2000) also point out the implication of high-velocity sill complexes in terms of interpretation and magmatic quantification of magmatic LCBs. Due to high impedance, relatively shallow and high velocity (+7 km/s) sill complexes can indeed return detectable refracted arrivals leading to inaccuracy in the velocity model and deep geometry estimation (Berndt et al., 2000). This can easily contribute to an overestimation of the mafic intrusions in depth (Berndt et al. 2000). Underplated material in the form of mafic sills intruding and/or stacked below the reflective lower crust is also a common feature of several magmatic rifts investigated worldwide (Deemer and Hurisch, 1994; Collier et al., 1994; Maystrenko, 2003; Thybo and Nilsen, 2012).

A flat Moho below the active graben is also a pertinent argument to explain the compensation of crustal thinning by progressive magmatic addition. This process is often proposed in VPM context to explain the lack of apparent crustal thinning along the COT. The Ivrea-Verbano zone along the inner arc of the Western Alps is a good field example of pre-existing lower crust where large volume of magma (50% of the outcrop) affected the extended continental domain (Rutter et al., 1993; Quick et al., 1994). Rock types comprise granulites to amphibolites metasediments and a large proportion of basic to ultrabasic complexes. Original batholith of almost 10 km thick has been observed. The Ivrea-Verbano zone may therefore illustrate particular geometries of lower crustal magmatic underplating, which may aid in the interpretation of present-day deep seismic profiles, crustal thermal pattern and related CLCB magma-tectonic process (Rutter et al., 1993; Barboza et al., 1999; Barboza and Bergantz, 2000; Khazanehdari et al., 2000). At upper crustal level, a well-exposed and more local analogue in the Seve Nappe Complex (Onshore Sweden/Norway) and the Greenland Coastal area has been proposed a field analogue for the transitional crust expected beneath the SDR. Both agree the presence of mafic dike swarms with densities of 70%–80% or more, which to some extent may represent the upper most part of the CLCB and intruded crust above.

Residual piece of continental crust injected and underplated might be preserved underneath the SDR and to some extent in the outer part of the volcanic wedge and Outer High. Before true oceanic crust, some 'embryonic crust' could be present. Then may represent an alternation of micro-continental fragments interbedded with oceanic crust, and overlaid by volcanic traps, as proposed, for example, in the Afar region (Courtillot et al., 1980; Wolfenden et al. 2005) which is a modern and relevant analogue for magmatic rift and VPM. Observation and modern high resolution aeromagnetic data show that before expecting oceanic crust, local Outer High volcanic features emplaced in some enigmatic residual and highly intruded continental crust. In SE Geenland White and Smith (2009) also suggest hat there is a small amount of continental crust present under the subaeriallly formed SDR, which serves to lower the crustal velocities somewhat below what they would be if the crust were fully igneous.

# FROM BREAKUP PHASE 1 TO ULTIMATE BREAKUP PHASE 2: FROM LOCAL TO ULTIMATE LITHOSPHERIC RUPTURE OF THE NEA

\*\*\*\*\*\*\*\*\*\*\*\*OK here, I talk mostly about Norwegian Greenland Sea, extra comments about the others oceanic segments could be welcomed (Martyn, Fernando, Dieter ????\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*

## From VPM to embryonic oceanic crust

### Mapping the continent-ocean transition (COT) and fixing a continent-ocean boundary (COB)

The real nature and meaning of the COT and COB and processes involved during the formation of a new steady state oceanic lithosphere in the NEA is unclear. Classic COT/COB issues and limitations are basically acknowledged and episodically discussed since the early stages of rifted margin and plate tectonic investigations (e.g. Rabinowitch and Labreque, 1979; Banda et al., 1995). As recently summarise and compiled by Eagle et al. (2015), the definition(s) and demarcation of a COT and the definition of a proper continent-ocean "boundary" (COB) are often biased and complicated due to contradicting definition/interpretation and ambiguous interpretation of sparse petrological and geophysical data. Most important, the quality and distribution of dense and high quality geophysical dataset (e.g. Refraction, wide-angle seismic 2D/3D reflection data, high resolution aeromagnetic and gravity surveys) is a primary requirement not always fulfilled everywhere in the NEA and other rifted margin in general. Ideally, it would require at best a large amount of rocks samples from the basement and/or adjacent transitional and oceanic crust along the entire COT. Except few exceptions on Earth (e.g. the Iberian margin) such conditions are rarely fulfilled. Along VPM, mapping a COT and/or suggesting a clear and sharp 'COB' is even more ambiguous and often refer to an approximation. It is often admitted that the magmatic COT should more or less coincide with the outline of the SDR domain (including Inner and Outer SDRs) when imaged by seismic data. Petrology of the lava wedges so far drilled along few VPM (e.g. DSDP Legs 38 and 81; ODP Legs 104 and 152) have confirmed that the SDR marks (at least in theses VPM examples) the petrological transition from lava having clear continental affinities to MORB-Type basaltic rocks with increasing oceanic affinities (Parson et al., 1989; Fitton et al., 1998; Larsen et al., 1998; Meyer et al., 2006; Abdelmalak et al., 2016). The mapped widths of SDR wedge domains (e.g. from Inner to Outer SDRs) can vary in the range of 50-150 km (Planke et al., 2000; Eagle et al., 2015). Basically, the location of the expected COB(s) could lie anywhere in between or possibly west of it. Along VPM segments, the magnetic signature is, however, dominated by the causative magnetic properties of the subsurface volcanostratigraphic sequences (Talwani et al., 1995; Schrekenberger, 1997; Berndt et al., 2001; Gernigon et al., 2012; Koopman et al., 2014; Eagle et al., 2015; Patton et al., 2016, Collier et al., 2017. Along VPM, the presence of subsurface volcanics is also and often problematic to seismically image the deepest basin architecture expected locally underneath the lava flows. However, recent datasets start to highlight convincing evidence of sedimentary basins and continental crust preserved underneath a large part of the Landwards Flows and Inner SDR (Eddy et al., 2014; Clerc et al., 2015; Patton et al., 2016; Abdelmalak et al., in prep). In the Laxmi Basin/Gop Rift region (West India), Guan et al., (2016) and Nemcôk and Rybár (2016) showed seismic evidences that the Inner SDR developed on top of thinned continental lower crust.

Along the magmatic COT, dominant magnetisation produced but the subaerial volcanism does not always allow us to fully constrain the pre-basaltic structures and enigmatic nature of the underlying crust and most of the magnetic sources usually converge towards the top of the volcanics. Depending on magnetic data quality and resolution, a COB is often proposed at the outer part of the Inner SDR (Smythe, 1983; Blaich et al., 2011; Clerc et al., 2015; Patton et al., 2016), indifferently between the main Inner SDR wedge itself and its distal termination (e.g. Skogseid and Eldholm, 1989; White et al., 2008; Koopman et al., 2014; Abdelmalak et al., 2016) and/or closer or the volcanic Outer Highs when observed (Gernigon et al., 2015; Geoffroy et al., 2015; Nemcôk and Rybár, 2016). In the northeastern part of the Gulf of Mexico, Imbert and Philippe (2005) and Eddy et al. (2014) proposed that the COB should coincide with the distal edge of the Gulf Coast Magnetic Anomaly. This anomaly coincides with the Outer SDR also located at the seaward limit of autochthonous/parautochtonous salt layers (Eddy et al., 2014; Rowan et al., 2014). Similar East Coast Magnetic Anomaly approximates the COB boundary in the Atlantic Ocean (Talwani et al., 1995; Labails et al., xxxx). In the Namibian VPM, Patton et al. (2016) show that potential field data could help to highlight continental inheritance that may influence both the inner and outer SDRs of the VPM. Accordingly, they concluded that the COB most likely locates in the distal termination of the outer SDRs where the largest magnetic gradient is observed. Recent interpretation and model, also suggest that the outer part or the SDR, offshore Brazil, does not necessarily represent true oceanic crust but could represent residual continental fragments of the original stretched crust more or less preserved and associated with the structural development of the SDR wedge itself (Stica et al., 2014; Geoffroy et al., 2015). In the Danakil Through (Afar), Bridge et al. (2012) and Almalki et al., 2014 also suggest that prominent magnetic stripes (> 200 nT) can also developed along intruded continental crust before the establishment of true oceanic spreading centers. Similar stripped transitional crust (less than 100 nT) have been mapped by aeromagnetic data along the coastal area of the southern Red Sea, particularly affected by dominant dike swarm (20 km-wide) and mid-crustal mafic intrusions within the necking zone of the sharp margin (Almalki et al., 2014). These dike are, however, post breakup, of axes and may be due to localised stretching and renewed adiabatic mantle melting right at the transition between continental crust and controversial oceanic crust (Cochran et al., 2007; Almalki et al., 2014, 2016). It may represent a failed attempt at a ridge jump and suggest that post-breakup related magnetic anomalies may also develop along the established COT, leading to more complicated geological scenario locally. Krishna et al., (2006) also shows that weak magnetic anomalies of the Laxmi Basin, earlier interpreted as sea-floor spreading anomalies can best be explained by volcanic intrusive of the stretched crust at the edge of the SDR. In the case of continuous and kilometric scale limited and prominent magnetic stripped anomalies (>400-500 nT at sea level altitude), oceanic crust is most likely acknowledged (Gee and Kent, 2007). It must also be noticed that these 'embryonic' anomalies are observed in the distal part of SDR observed onshore in a possible and modern VPM analogue (e.g. southern Red Sea margin, Wolfenden et al., 2015; Corti et al., 2015).

The expected COB(s) in the NEA is/are usually defined between the outboard edge of highly attenuated, unequivocal continental crust (or eventually exhumed subcontinental mantle) and the inboard edge of unequivocal oceanic crust (e.g. Rabinowitch and Labreque, 1979; Boillot and Froitzheim, 2001; Direen et al., 2011). How to fix a detailed boundary and propose a clear limit between sub-continental and oceanic lithosphere in the NEA is then not straight forward and often complicated by the presence of thick sedimentation and/or thick basaltic coverage often observed along most of the NEA COT. For paloegographic studies and/or geodynamic modelling specific issues it is, however, often useful, if not required to estimate, refine and/or provide at least a rough estimation of such compound and possibly virtual limit being aware that the limit could vary depending on new data acquisition and availability.

### Diachronic and segmented emplacement of the SDR and expected COB: few evidences in the NEA

When the basalt emplaced along the NEA COT, it often dominates the magnetic signal and does not allow directly to define a proper COB, determine the underlying basin/basement not to constrain an accurate timing of the breakup. It is the case of the Vøring Marginal High drilled by ODP well 642 at landward edge of the Inner SDR. According the well datation and recovered magnetisation, most of the Upper Series, emplaced during the equivalent of magnetic polarity chron C24r (57.1-53.9 Ma), whereas the Lower Series possibly emplaced around magnetic chrons C25n (57.6-57.1 ma) (Schöharting and Abrahamsenm 1989; Abdelmalak et al. 2016). The negative magnetisation (C24r) may explain the dominant negative magnetic domain associated with the Landward Flows defined around and landward of the ODP well 642 (Fig. X). Oceanward, the main Inner SDR wedge in the Vøring Marginal High shows, however, a prominent positive magnetic signature which correlated well with the outer part of the Inner SDR wedge mapped with seismic data in the Vøring Marginal High (e.g. Berndt et al., 2001; Abdelmalak et al., 2016; Gernigon et al., 2009). To explain the sudden change of the total magnetic field, natural Remanence Magnetisation with normal polarity is likely expected for the uppermost part of the Inner SDR lava flows not recovered by the ODP well (e.g. Schrekenberger, 1997). Since the underlying lava flows are correlated with negative polarity chron C24r, we suggest that the positive magnetic anomaly associated with the Inner SDRS could be age-correlated with positive magnetic polarity chrons C24n3n and/or C24n1n. Tentatively, the Inner SDR and underlying flows should rapidly develop from ~57.0–52.6 Ma in the Vøring Marginal High.

South of the Vøring transfom margin, recent aeromagnetic coverage of the Møre VPM segment and adjacent oceanic domain allowed us to question the timing of the VPM formation in the distal part of the Møre margin. The Inner SDR in the northern part of the Møre Marginal High formed landward of the first positive magnetic polarity chrons C24n1n and C24n3n and negative polarity chrons C24r identified at the eastern edge of the oceanic Norway Basin (e.g. Gernigon et al., 2012, 2015). According to the GTS 2012 polarity time scale of Ogg et al. (2012), this means that some of the Inner SDR seismically recognised in the northern part of the Møre Marginal High could have growth before the late emplacement those recognised in the Vøring VPM segment. Gernigon et al. (2008; 2012) concluded that the Inner SDR that formed in the central and northern parts of the Møre Marginal High most likely developed during Thanethian (late Paelocene) whereas the Inner SDR in the Vøring Marginal High is dominantly post-C24r and should develop a few my later in the Ypresian age (early Eocene).

Assuming that the Inner SDR is a feature that develops above continental or transitional crust, this also means that steady state breakup in the central part of the Møre VPM could eventually initiated few m.y. before the Vøring VPM. Another argument for that is the timing of the lava flows observed in the outer part of the Møre Basin and around the Vøring transform margin. Based on extensive seismic mapping of the outer Møre and Vøring basins tie to wells, some of these lava flows (or Transform Flow, Berndt et al. 2001) are Intra-Palaeocene in age and slightly older than the Eocene Inner flows emplaced in the outer Vøring Basin. Some of these flows could be fed by dike swarm and sill complex possibly linked with local Outer High (e.g. Gjoll High) emplaced in the "embryonic" crust landward of C24r in the outer part of the Jan Mayen corridor (Gernigon et al., 2012, 2015)(Figure x).

\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*Not sure we have more other evidence of diachronism or breakup delayed n the NEA ???????

Rockall-Hatton ?? Martyn

SE Greenland ?? Laurent

NE Greenland ???? Dieter)

### First magnetic chrons in steady state oceanic crust

The oldest oceanic chron edge that was earlier recognised in the NEA often refer to the C24 magnetic sequences including C24r (53.9-57.1Ma) C24n3n (53.4-53.9 Ma) and C24n1n (52.6- 53.0 Ma) also refer as C24A/C24B in the old literature (e.g. Vogt et al., 1986; Skogseid and Eldholm, 1987).

In the Lofoten-Vesteralen margin, Tsikalas et al. (2001) identified both C24n3n and C24n1n from the Vøring Marginal High up to Senja Fracture Zone on vintage profiles. Based on more recent aeromagnetic survey (e.g. RAS-03) Olesen et al. (2007) revisited the early spreading and confirm the presence of two distinct C24n3n and C24n1n magnetic polarity chrons that coincides with the first oceanic crust as supported by a recent OBS profile across the Lofoten COT (Breivik et al., 2017). Olesen et al. (2007) noted, however, that some of the associated oceanic fracture zone mentioned in Tsikalas et al. (2001)(e.g. the Bivrost, Jennega and Vesteralen fracture zones) were simply artefacts possibly due to the inherent navigation mistakes of the (pre-GPS) previous magnetic acquisition of the 70ies-80ies.

Closer to the Vøring Marginal High, only magnetic polarity chrons C23-C22 present proper characteristics of a well stripped steady state spreading system and C23n2n can be linked with linear and continuous oceanic magnetic anomalies of similar ages in the Lofoten Basin (Olesen et al., 2007; Gernigon et al., 2009). In the Voring Marginal High, the magnetic anomalies mostly reflect the volcanostratographic sequences imaged on seismic (e.g Berndt et al., 2001). The SDR system in the Voring Marginal High may eventually emplaced in true oceanic domain but based on the ambiguous magnetic pattern obviously produced by the large extension of the flood basalt, we cannot exclude nor conclude that part of the VPM did not reach full breakup before undisputable C23r or C23n1n in that part of the NEA and not C24r as traditionally assumed. This delayed interpretation of the breakup would also make sense if we agree that the uppermost part of the Inner SDR in the Vøring Marginal High was still under construction during C24n3n and C24n1n1, whereas the seafloor spreading was already and punctualy established both in the adjacent Lofoten and Norway oceanic basins.

In NE Greenland, the conjugate C24n3n and C24n1n magnetic polarity chrons also been identified west of the Greenland Escarpment (Vogte et al., 1986; Skogsed et al., 2000; Tsikalas et al., 2001). However, there is still controversies about the location and identification of these anomalies closer to the East Jan Mayen Fracture Zone. The problematic region is located between 72.5–74?N and exhibits unclear and ambiguous magnetic anomalies (Tsikalas et al., 2002). Theses anomalies are attributed to extensive, shallow breakup lavas (Voss and Jokat, 2007). Seismic refraction profiles in this offshore area, in the fjords and on the shelf, show that the region is underlain by a 20–22 km thick crust (Voss and Jokat, 2007; Voss et al., 2009). A COB location has been earlier proposed by Tsikalas et al. (2002) at ~70 km off the Greenland coast, which was in contrast to Scott (2000) who expected an alternative COB at less than 10 km off the nearest Greenland coastline. Based on refraction data, a 120-130 km wide COT has been later proposed for the area (Voss and Jokat, 2007; Voss et al., 2009). Voss and Jokat (2007) suggested that the anomalies landward to the unambiguous C22n may represent intruded and stretched continental. This interpretation suggests that the oceanic crust between the NE Greenland margin and the mid-Norwegian margin initiated first close to the East Greenland Ridge and then propagated progressively from North to the South from C24 to C22.

In a recent interpretation, Geissler et al. (2017) suggested that the COB in NE Greenland controversial area is located at less than 100 km from the coastline and is almost sub-parallel to the undisputed C22. This interpretation remains, however, problematic. First this assumption suggests that the breakup would be almost instantaneous over a distance between the conjugate margins of more than 500 km. Then, this interpretation does not easily explain the obliquity between the magnetic chrons and diachronic Inner SDR mapped offshore NE Greenland.

\*\*\*\*\*Dieter extra comments ?? about the COT here or in the later discussion about propagation \*\*\*\*\*\*\*

In the Norway Basin and thanks to the most recent and modern aeromagnetic surveys of the NEA (see Gernigon et al., 2008, 2009, 2015 for details) the COB was slightly interpreted landward of magnetic chron C24n3n (old C24B of Skogseid and Eldholm, 1989) along the continuous and prominent magnetic lows, interpreted as stage C24r (57.1-53.9 Ma). Magnetic chron stage C24r is particularly visible south of the Vøring transform margin, where the breakup-related volcanism is reduced. As previously mentioned, the nature of the magmatic COT landward is still difficult to interpret as pure continental or oceanic crust. Magnetic observation in the outer Møre basin show that the COT could represent some kind of embryonic crust showing highly intruded continental or 'mullioned' continental-oceanic crust (Gernigon et al., 2012, 2015). Within the ambiguous COT magnetic anomalies with large amplitudes (>100-200 nT) similar in amplitude to well developed chron but punctual and not clearly continuous. In the outer Møre VPM, these anomalies often coincide with seismic Outer Highs also interpreted as shallow to subaerial volcanic mounds (Planke et al., 2000; Berndt et al., 2001). Similar volcanic features have been also described along the Laxmi Basin which also represents an unclear crustal domain at the edge of conjugate SDR (Misra et al., 2015; Krishna et al., 2006).

In most of the Norway Basin, the first and oceanic magnetic polarity chron C24n1n (C24A) and the youngest edge of magnetic chron C24n3n (C24B) can be mapped almost continuously and with confidence on both conjugate margins until they merge together north of the Faroes platform, interpreted as a volcanic oblique shear margin (Gernigon et al., 2012). The youngest edge of magnetic chron C24n3n is usually sub-parallel to oblique with the outer edge of the SDRs recognised on on both sides of the Norway Basin. Lateral rift propagation of the early spreading system from North to South was expected during the early Eocene spreading stage (Gernigon et al., 2012). Even if large uncertainties remain about the pre-C24 magnetic chron picking, Gernigon et al. (2015) have recorded locally low spreading rates during the proto-breakup stage (<5-10 mm/year), increasing rapidly to more than 15-25 mm/year during the formation of the magnetic polarity chrons C24n3n and C24n1n (C24A). This might suggest slow spreading rates during the proto-stage of breakup if we consider that the prominent but embryonic magnetic anomalies chrons locally observed landward of C24n3n are truly C25 in age.

In SE Greenland, a particularity of the COT and associated VPM is the large asymmetry between the well-formed convex-up SDRs observed (100–200 km-wide) and the narrower on the Hatton VPM where SDR formed before C23r/C24n1n are only 50–100 km-wide (Hooper et al., 2003; White and Smith, 2009). On the Hatton margin magnetic anomalies chron C23r/C24n1n lies close to the break of the slope, whereas it lies more than 100 km seaward of the pronounced coastal flexure and associated shelf edge on the conjugate SE Greenland margin (Larsen and Thorning, 1979; Larsen and Saunders, 1998). Explanation for the prominent asymmetry is that there was grossly asymmetric seafloor spreading, or a continuously migrating ridge axis, with the Greenland side spreading at a half rate about three times faster than the half-spreading rate of 15 mm/yr on the Hatton side (Larsen and Saunders, 1998). However, Hopper et al. (2003) points out, that the continuous nature of the SDRs preclude the possibility of ridge jumps being invoked to explain the asymmetry.

On both sides of the Reykjanes Ridge, prominent anomalies C23n and 24n1n are readily identifiable at the edge of the VPM. In the SE Greenland, Chron 24n1n of the oldest part of the oceanic succession was determined by Larsen and Jakobsdóttir (1988). In SE Greenland, the SDRs in the wide COT have been interpreted as caused by lava flows extruded subaeriallly, a conclusion supported by samples taken from ODP drilling including boreholes 917, 918 and 919 along the unclear volcanic COT of the VPM (Larsen and Saunders,1998; Hopper et al., 2003).

Within the SDR, Larsen and Saunders (1998) interpreted cryptochrons 24.1n to 24.11n within reversed magnetic polarity chrons C24r. However, White and Smith (2009) challenged the interpretation of the early magnetic chrons observed above the broad zone where the SDR sequence developed. They suggest that the inferred magnetic polarity chrons in the Irminger Basin do not represent true seafloor spreading magnetic anomalies but could simply represent the subaerial flow fronts contact and lateral variations in the subaerial basalts above uncertain transitional crust. This may explain the difficulties to link properly link the radiometric age from the ODP well and the pseudo-cryptochorns age (wrongly) inferred. This interpretation also questions the real nature of the real crust underneath the SDR observed in the western part of the Irminger Basin. A COB was proposed by Hopper et al. (2003) along the pronounced shelf break of the margin suggesting that the broad SDR domain emplaced above oceanic crust. White and Smith (2009) suggest, however, that a small amount of continental crust could be present under the subaeriallly formed SDRs, which serves to lower the crustal VP velocities somewhat below what they would be if the crust were fully igneous.

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Should be the same up to 22 where sub-aquatic SDR are still observed (Laurent’s idea) ????? In that situation, we should have some serious problem with the conjugate chrons interpretation. On Hatton C23-C22 from oceanward of the SDR ! Extra comments from Laurent, Dieter about this asymmetric VPM??????

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## Oceanic Phase I: Early Eocene oceanic crust domain from C24r to C21r (51.1-47.3 Ma)

Oceanic phase I of the NEA spreading development coincide with the progressive establishment of the early segmented oceanic crust mapped on both sides of the GIFR. After C24n1n, half-spreading rates of the earliest spreading are on average less than 25 ± 2 mm/year in the eastern Norway Basin, and less than 20 ± 2 mm/year in its western part. The previous NB-07 survey had already revealed the presence of off-axis features that are magnetized with reverse polarity between magnetic polarity chrons C24n3n and C22n, which were earlier interpreted as off-axis seamounts and/or some relics of aborted rift axes (see Gernigon et al., 2012). Similar off-axis anomalies, suggesting the instability of the oceanic rift has not been recognised in the conjugate JAS-12 survey (Gernigon et al., 2015), and hence support the interpretation of local aborted rift axes that shaped the eastern side of the dissymmetry Norway Basin during the Early to Late Ypresian (56-47.8 Ma).

New aeromagnetic data also shows that the northern and central parts of the Norway Basin are divided by a small-offset fracture zone, which appears quite different from the straight and classic signature of a ´normal´ fracture zone (e.g., the Jan Mayen Fracture Zone). A large band of disturbed magnetic anomalies situated west of the East Jan Mayen Fracture Zone coincides with the location of the `Norway Basin Fracture Zone´ earlier introduced by Skogseid and Eldholm (1987), but not properly identified due to the lack of high-resolution data. Based on sparse vintage magnetic data, Skogseid and Eldholm (1987, 1988) interpreted the Norway Basin Fracture Zone as an ocean transform sub-parallel to the East Jan Mayen Fracture. The NB-07 and JAS-12 surveys proved well the presence of a major oceanic and magnetic discontinuity here, but with a slightly different magnetic signature, and not so linear compared to the adjacent East and West Jan Mayen Fracture Zones. On both sides of the Norway Basin, this discontinuity initiated around recognised C24n3n and shows a complex magnetic pattern characterised by discrete, non-continuous and oblique, magnetic lows and highs. This discontinuity defines a diffuse transition zone between two different spreading segments interpreted as distinct oceanic domains in this paper. The ´Norway Basin Fracture Zone´ likely represent a composite feature and defines wavy to low-angle V-shaped features pointing in the direction of competing (and alternating) propagating oceanic systems. Similar discontinuities with low-angle V-shape features are relatively similar to oblique structures and low-offset fracture zones also described between first-order and large-offset oceanic fracture zones in the Central Atlantic (Müller and Roest, 1992) and in the Gulf of Aden (d'Acremont et al., 2006). Such a pattern observed in the Norway Basin may be an indication of a relatively unstable small-offset fracture zone migrating (or oscillating) in the past along the paleo-Aegir Ridge. This boundary possibly accommodated the motion of two sub-plates expected at the level of the JMMC (Gernigon et al., 2015). The period also coincides with the onset of atypical melt production along the trend of the Jan Mayen Fracture Zone (Vøring Spur) probably associated with a leaky transform along the Jan Mayen Fratcure Zone (Gernigon et al. 2009).

South of the GIFR, an initial orthogonally spreading (Smallwood and White, 2002) spreading system without transform faults established during C24r-C21n period, forming continuous magnetic anomalies nearly parallel to the continental margins. Smallwood and White (2002) modelled and recognised magnetic chrons C24n3n and C24n1n in the Northern part of the Rockall Palteau but could not clearly acknowledge the individual C24n3n and C24n1n in the southern part of the Hatton VPM where the first undisputable magnetic chrons seems to be C23r (52.6-51.8 Ma) in the Iceland Basin. Stoker et al. (2011) recognised that the oceanic phase 1 from C24r to C21r west of the Rockall-Hatton margin is also characterised by smooth oceanic crust. Like the Norway Basin, ridge jump is also epxected to the SW of the GIFR where the meaning of the COT and the location of the COB is not very well established (Smallwood and White, 1999). Flow lines based on rotations inferred from vintage magnetic data observed in the Irminger and Iceland basins also show a sharp kink at C22, not observed between Greenland and Lofoten basins data (Gaina et al., 2009)

Phase 1 in other segments …..to be complete (Fernando input ?)

The propagation controversies between NE Greenland and Norway ….input from Dieter

## Oceanic Phase II: Mid-Eocene to Late Oligocene oceanic crust from C21r to C10 (47.3-27.8 Ma)

The Oceanic crust formed between C21r to C10 (47.3-28.8 Ma – mid-Eocene to Late Oligocene) has been identified and defined as a separate seafloor spreading sequence (Phase II of the NEA development proposed in the present study).

In the Norway Basin, broad magnetic anomaly highs defined at C21n (33.7-33.1 Ma) and C20n (43.4-42.3 Ma) and adjacent lows (C22r, C21r and C20r) are decreasing in width towards the southern part of oceanic basin. Chrons C18 (40.1-38.6 Ma), C17 (38.3-37.7 Ma) and C13n (33.7-33.1 Ma) have been identified in the southern part of the Aegir Ridge. The magnetic pattern suggests significant dislocation of the oceanic crust, and some anomalies have been tentatively interpreted as C18, 17 and C13n and locally C12-C11 (31.0-29.1 Ma) in the northwestern part of the survey area. From C22n/C21n, the entire Aegir Ridge adjusted from slow to ultra-slow spreading rates (estimated at 16±2 to 8±2 mm/year, increasing from north to south in most of the NB).

One of the striking phenomena identified in the new gridded data is a distinct and clear spreading direction reorganisation that initiated between C22n and C21) from North 340° to North 8°. This is illustrated by a change in orientation of C21n relative to C22n with 10-30° in the central and northern part of the survey (Gernigon et al., 2008; 2012). Due to this local spreading readjustment, chrons C21n are locally split in response to the north-westward propagation of the growing C20r anomalies). During this period, the crudely northward-widening fan-shaped magnetic anomaly pattern initiated and subsequently developed in the Norway Basin.

The Oligocene time, which corresponds to the C13n period (33.5-33.0 Ma) is often considered as a major event in the NEA, when an arm of the triple junction between North America, Greenland and Eurasia was abandoned (Sirastava and Tapscott, 1986; Gaina et al., 2002). During this reorganisation, a change in the opening direction between Eurasia and North America/Greenland from NNW-SSE to NW-SE led to the initiation of the West Jan Mayen Fracture Zone (Mosar et al., 2002; Talwani and Eldholm, 1977). Despite uncertainty regarding the real extent and age of the West Jan Mayen Fracture Zone initiation, which could be older than C13n (see Gernigon et al., 2009), the post-C13 period effectively coincided with a period of westwards migration of the Aegir Ridge spreading centre towards the proto-Kolbeinsey Ridge (Gaina et al., 2009). The C13 period has often been considered as the principal tectonic event that affected the Norway Basin, although it has been recently suggested based on a detailed study of the Treitel Ridge (west of the southern Aegir Ridge) that C18n time (40.13-38.4 Ma) could represent a better timing for the onset of the plate reorganisation (Vogt and Jung, 2009) coinciding with a controversial seafloor spreading slowdown in the Labrador Sea (Roest and Srivastava, 1989). The progressive ending of slow to ultra-slow spreading along the Aegir Ridge took place between magnetic chron C13n and chron C6B, which is not clearly identified in the Norway Basin due to ultraslow spreading (Jung and Vogt, 1997; Vogt et al., 1970).

South of the GIFR, a major change of the spreading configuration occurs later around C17 (Jones et al., 2002; Smallwood and White, 2002; Hey et al. 2015). The magnetic pattern younger than C17 suggests a kinematic change from the initial orthogonally spreading ridge system without transform faults into a slow-spreading orthogonal ridge/transform pattern. Fault orthogonal spreading probably resulted from the nearly instantaneous response to a change in spreading direction of 25-30°, possibly caused by the termination of spreading on the Labrador Sea expected before C13 (Vogt and Avery, 1974; Jones, 2002; Oakey and Chalmers, 2012; Hey et al. 2015). Nearly instantaneous plume-related mechanisms have also been suggested (Howell et al., 2014). Very shortly after that event, a second major reorganization initiated south of Iceland between chrons 15 and 13 and continues at present. During this period, the orthogonal Eurasia ridge/transform faulted geometry between the Iceland and Irminger basins is progressively changing to a non-segmented oblique spreading geometry still observed on the Reykjanes Ridge (Jones et al., 2002; Hey et al., 2015; Martinez and Hey, 2017). This reconfiguration occurred without further significant changes in opening direction (Smallwood and White, 2002) and was accompanied by the formation of diachronous V-shaped crustal ridges and troughs (Hey et al., 2015; Martinez and Hey, 2017), and the elimination of the previous ridge-orthogonal crustal segment offsets and boundaries (Jones et al., 2002; Hey et al. 2015; Martinez and Hey, 2017). A popular plume related model (White, 1997; Jones, 2003) proposes that this latest spreading-centre reorganisation results from a thermally induced change in lithospheric behavior from the previous brittle rheology, producing the orthogonal ridge/transform fault patterns, to a more ductile rheology producing atypical oblique unsegmented ridge patterns. More recent geophysical dataset in the Reykjanes Ridge (Hey et al., 2015; Martinez and Hey, 2017) suggest, however, that the plume model is questionable. One possibility is that plume pulses drive the propagators. However, rift propagation could easily produce similar V-shaped wakes with crustal thickness variations, suggesting the possibility that a pulsing Icelandic plume might not be involved (see details and further discussion by Fernando, this issue)). Similar V-Shape features have been earlier proposed in the Norway Basin by Breivik et al., 2006 also suggesting the deep implication of the Icelandic mantle flow. Gernigon et al. 2012, show; however that the expected V-Shape feature may simply represent a new family of oceanic Fracture Zones associated with the local reorganisation from Oceanic Phase I to Oceanic phase II.

Phase 2 in other segments …..to be complete (Fernando / Martyn input)

\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*Not sure if we should also mention the sedimentary response pulse at that stage or make a separate chapter …..Martyn ????

## Oceanic Phase III: Opening of the Kolbeinsey Ridge and JMMC final dislocation from C10 to C6B

Whereas continuous seafloor spreading was ongoing along the Mohn’s and Reykjanes Ridge during oceanic phase III, the region located in between the GIFR and the Jan Mayen Fracture Zone was characterised by a complex geodynamic process leading to the JMMC formation in Late Oligocene. Most of the previous interpretations of the Norway Basin have suggested that variable spreading rates occurred along the Aegir Ridge, leading to a fan-shaped distribution of the magnetic anomalies which may have required a rotation pole located in its southern part. According to Talwani and Eldholm (1977), Unternehr (1982), Nunns (1983), and Vogt (1986), part of the JMMC must have rotated counterclockwise to accommodate the fan-shaped development of the Norway Basin. This kinematic model suggests a compensating rifting and/or spreading episode(s) between East Greenland and south Aegir Ridge. However, a closer look on the early publications shows that there is no general consensus about the precise timing and initial development of the fan-shaped spreading in the Norway Basin. For example, Talwani and Eldholm (1977) inferred that the fan-shaped development of the Norway could have initiated between C20 and C7 while Unternehr (1982) proposed that it rather formed between anomalies C13 and C5D (17.6-17.2 Ma) during the rotation of the JMMC. Vogt (1986) argued that an additional oceanic spreading system is required to compensate the fan-shaped spreading. He proposed that it could have been developed by the northward propagating spreading along the growing Kolbeinsey Ridge more precisely between anomalies C18 (40.13-38.4 Ma) and C6b (23-22.5 Ma), the first unambiguous spreading anomaly identified between East Greenland and the JMMC. Nunns (1983) suggested that anomaly C20 could have marked the onset of the JMMC dislocation to the west, but Müller et al. (2001) concluded that chron C13 (33.5-33 Ma) was a better candidate. Lundin and Doré (2002) also adopted a similar model and suggested that the same rift/drift compensation from Oligocene spreading occurred around C13n. Despite timing uncertainties and some ambiguities about the interpretation/identification of magnetic anomalies, the fan-shaped concept was nevertheless generally adopted by most of the previous authors, until an alternative hypothesis published by Scott et al. (2005) proposed a segmentation model of the spreading system. In this model, the structure of the Norway Basin was interpreted to be the result of a competition between the propagating tip of the Kolbeinsey Ridge and the retreating tip of the Aegir Ridge associated by orthogonal spreading corridors displaced by several post C18 (41.1-38.4 Ma) oceanic transforms and shear zones. New aeromagnetic survey in the Norway Basin confirm. However that fan-shaped spreading configuration of the Norway Basin was particularly active around the transition C22-C21 interpreted as a major kinematic reorganisation phase in the Norwegian-Greenland Sea (Gernigon et al., 2012, 2015). This kinematic model suggests a compensating rifting and/or spreading episode(s) between East Greenland and south Aegir Ridge. However, a closer look on the early publications showed that there is no general consensus about the precise timing and initial development of the fan-shaped spreading in the Norway Basin. For example, Talwani and Eldholm (1977) inferred that the fan-shaped development of the Norway Basin could have initiated between C20 and C7 while Unternehr (1982) proposed that it rather formed between anomalies C13 and C5D during the rotation of the JMMC. Vogt (1986) argued that an additional oceanic spreading system is required to compensate the fan-shaped spreading. He proposed that it could have been developed by the northward propagating spreading along the growing Kolbeinsey Ridge more precisely between anomalies C18 and C6b (22.2-21.7 Ma), the first unambiguous spreading anomaly identified between East Greenland and the JMMC. Nunns (1983) suggested that anomaly C20 could have marked the onset of the JMMC dislocation to the west, but Müller et al. (2001) concluded that chron C13 (33.5-33 Ma) is a better candidate. In a more recent interpretation, Lundin and Doré (2002) also adopted a similar model and suggested that the same rift/drift compensation from Oligocene spreading occurred around C13n. Despite timing uncertainties and some ambiguities about the interpretation/identification of magnetic anomalies, the fan-shaped concept was nevertheless generally adopted by most of the previous authors, until an alternative hypothesis published by Scott et al. (2005) proposed a segmentation model of the spreading system. In this model, the structure of the Norway Basin is interpreted to be the result of a competition between the propagating tip of the Kolbeinsey Ridge and the retreating tip of the Aegir Ridge associated by orthogonal spreading corridors displaced by several post C18 (41.1-38.4 Ma) oceanic transforms and shear zones.

The Oligocene time, which corresponds to the C13n period (33.5-33.0 Ma) is often considered as a major event in the North Atlantic, when an arm of the triple junction between North America, Greenland and Eurasia was abandoned (Sirastava and Tapscott, 1986; Gaina et al., 2002). During this reorganisation, a change in the opening direction between Eurasia and North America/Greenland from NNW-SSE to NW-SE led to the initiation of the West Jan Mayen Fracture Zone (Mosar et al., 2002; Talwani and Eldholm, 1977). Despite uncertainty regarding the real extent and age of the West Jan Mayen Fracture Zone initiation, which could be older than C13n (see Gernigon et al., 2009), the post-C13 period effectively coincided with a period of westwards migration of the Aegir Ridge spreading centre towards the proto-Kolbeinsey Ridge (Gaina et al., 2009). The C13 period has often been considered as the principal tectonic event that affected the NB, although it has been recently suggested based on a detailed study of the Treitel Ridge (west of the southern Aegir Ridge) that C18n time could represent a better timing for the onset of the plate reorganisation (Vogt and Jung, 2009) coinciding with a controversial seafloor spreading slowdown in the Labrador Sea (Roest and Srivastava, 1989).

The progressive ending of slow to ultra-slow spreading along the Aegir Ridge took place between magnetic chron C13n and chron C6B, which is not clearly identified in the Norway Basin due to ultraslow spreading (Jung and Vogt, 1997; Vogt et al., 1970). Tentatively, Talwani and Eldholm (1977) placed the extinction time between chrons C8 and C7 (24.4-23.9 Ma). Subsequently, the complete opening of a new spreading axis along the Kolbeinsey Ridge resulted from the progressive dislocation of the JMMC from Greenland (Gaina et al., 2009; Skogseid and Eldholm, 1987). Since this second phase of breakup around C6B, the ridge, which is still presently active, led to oceanic accretion of the Iceland Plateau between the JMMC and the East Greenland margin (e.g. Vogt et al., 1980).

## The Greenland-Iceland Faroe Ridge enigma

\*\*\*I need it for the buffer story\*\*\*\*\*can be reduced versus Gillina paper ????\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*

Despite decades of geophysical and geological investigations (e.g. Bott et al., 1981), the nature of the GIFR remains relatively enigmatic (See also Gillian chapter). In the south-western part of the Norway Basin and north of the Iceland Basin, the connection with the high-magnetic trends observed along the GIFR (e.g Fleisher et al., 1974) remains uncertain and the complex transition between the nordic and southern oceanic basins of the NEA. Although there are broad NE-SW lineations, correlated with average magnetization polarity, proper identification of the magnetic chrons, especially west of the Faroe Platform is still difficult, due to the late volcanic and intrusive activities related totally or not to the formation of Iceland. We also expect combined and complicated magnetic disparities simply associated with lateral variations of depth to basement and possible discrete ridge jumps (Mittelstaedt et al., 2008; Smallwood and White, 2002). In this case, the sparse magnetic data southwest of the new and reliable aeromagnetic survey in the Norway Basin precludes any detailed and confident interpretation of the magnetic patterns along the Faroe-Iceland Ridge segment of the GIFR at the present stage.

In its offshore part, the GIFR has been drilled by borehole DSDP site 336 on its northern flank recovered Middle Eocene basalts at 515 m b.s.l (Talwani et al. 1976). The basalt (43–40 Ma; K-Ar rock datation) grades is overlain by 21 m of volcanic conglomerate and thick red claystone, interpreted as a lateritic weathering of the basalt basement. This palaeosol is overlain by 295 m of middle Eocene–late Oligocene marine mudstones (Ellis and Stocker, 2014). Micro-palaeontological evidences indicate that the uplift of the Faroe-Iceland Ridge at DSDP site 336 occurred during late Eocene time and remained as either a continuous bridge or a string of closely spaced islands possibly into Oligo-Miocene time (Bott et al., 1983; Berggren and Schnitker 1983; Ellis and Stocker, 2014). It is also supported by the continued migration of flora and fauna from America to Eurasia during the Eocene–Miocene via the so-called ‘Thulian land bridge’ (Bott et al., 1983; Beard, 2008; Ellis and Stoker, 2014).

The GIFR is characterised by an atypical development since the early opening of the NEA and show an atypical thick crust which partly explains the atypical subsidence history of this transitional and atypical region (Bott et al., 1983; Richardsson et al; 1998; Smallwood et al.; 1995; Foulger et al., 2006; Hjartason et al., 2017). The crustal thickness of the GIFR reaches locally 30-40 km thick, thinning to approximately 25 km thick across most of Iceland and beneath the broad Iceland Plateau. The crustal thickness along the GIFR decreases from Iceland to the southeast. Crust to the north and south of this atypical ridge system is about 12–15 km thick (Bohnhoff and Makris 2004; Richardsson et al., 1999; 1998; Smallwood et al. 1999; White et al. 2008). The lower crust is almost 35 km thick under central Iceland and decreases to 15–20 km thickness away from the centre of the hotspot and beneath the Iceland platform (Wolfe et al., 1997; Foulger et al., 2003; Brandsdóttir and Menke 2008; Darbyshire et al. 1998; Hooft et al. 2006). Surrounding oceanic crust formed on the Kolbeinsey and Reykjanes Ridges to the north and south of Iceland thickens towards Iceland also reflecting atypical melt production increasing towards the GIFR (Hooft et al. 2006; Smallwood et al. 1995).

In Iceland, the GIFR emerges and exhibits a complex and instable plate boundary linked to the Reykjanes Ridge to the south and the Kolbeinsey Ridge to the north by transform fault zones. (Fig. x). The oldest dated basaltic outcrops in NW Iceland are 17 Ma (Miocene) and 13 Ma in east Iceland (Hardarson et al. 1997; Einarson, 2008, also see chapter Iceland). The rift zones themselves are composed of an array of overlapping, volcano tectonic spreading segments, 10–20 km wide and tens of kilometres long (Enarsson, 2008), analogous to spreading segments of modern mid-ocean ridges (Siler and Karson, 2017). Most volcanic systems, but not all develop central volcanoes (stratovolcanoes, composite volcanoes, collapse calderas) supplied with magma from shallow magma chambers (Gudmundson, 2016; Siler and Karson, 2017). Normal faults spaced 1–5 km apart commonly bound the segments creating shallow graben (Einarsson 2008; Sæmundsson, 1986). In Iceland, rift propagation also has also resulted in ‘ridge jumps’ in order for the locus of plate spreading to remain above the Icelandic ‘hotspot’ as Iceland migrates west-northwestward (Wolfe et al., 1997). Spreading on the Eastern Rift Zone is replacing spreading on the Western Rift Zone with which it overlaps (Einarsson, 2008; Garcia et al., 2008). Similarly, spreading on the Northern Rift Zone replaced an original spreading that occurred on a rift zone to the west in the Húnaflói-Skagi area between approximately 7-12 Ma (Garcia et al., 2008; Silver and Karson, 2017). This kind of instabilities of the magmatic rift is not restricted to Iceland but a few seismic reflection data have also revealed several abandoned rift centres, mapped as synclines and anticlines structures along other submerged segments of the GIFR (Hjartason et al., 2017). Compared to surrounding oceanic segments where such instabilities vanished after the Oceanic Phase I, these instabilities continue until the present day in the GIFR

The total thickness of the upper crustal units in Iceland is not accurately known, but drilling in the most deeply exhumed areas shows that the lavas are at least 4.5 km thick. Foulger et al. (2003) also suggested that the upper crust is typically 7 ± 1 km thick, heterogeneous and show high velocity gradients. The Icelandic upper crust is often associated with layers 0–2 in oceanic crust and ophiolites (Foulger et al., 2003; Sigmundson, 2006). However, evidence of continental contamination is also recognised (Paquette et al., 2006; Foulger et al., 2006; Torsvik et al., 2015). The geometry of lava flows on Iceland is also very similar to that of SDR and the processes of rifting could (partly) applied to magmatic rift and/or VPM settings observed in the NEA (Mutter 1985; Bourgeois et al., 2005; Karson et al., 2016; Siler and Karson, 2017). Like the NEA VPM; dense dike swarms that can be traced for many kilometres along strike and locally the lava flow dip gently and subside toward the neovolcanic zones (Pálmason 1986; Garcia et al., 2008; Siler and Karson, 2017). Some major crustal detachment have also been proposed to explain the wedge geometry and the downward growing of the basaltic wedge in Iceland (Bourgeois et al., 2005). Ambient noise tomography shows that continuous band of slow seismic velocities correlates well with the present-day active rifts (Green et al., 2017).

Lower crustal rocks in Iceland are mostly constrained by geophysical data. Early crustal seismic experiments documented a velocity layering in Iceland similar to the velocity structure of typical oceanic crust, but their interpretation sparked a major debate about the crustal thickness (Bott and Gunnarsson 1980; Flovenz 1980; Gebrande 1980; Bjarnason 1993; Staples 1997; Foulger et al., 2003; Brandsdóttir and Menke, 2008). These studies found low-velocity surface layers of Vp ~3.7 km/s and V s ~2.7 km/s, underlain by middle crustal velocities of 6.3-6.7 km/s (Foulger et al., 2003). Below this the refraction studies and surface wave dispersion required compressional velocities of 7.2-7.4 km/s relatively similar to the LCB described along the adjacent NEA VPM. For several decades, the interpretations for the +7.0 km/s material at 10–20 km depth varied. The thick-cold crust model interpreted the 7.4 km/s material as fast crust, comparable to a thick Penrose-type oceanic layer 3B (lower crustal cumulates) (Bott and Gunnarsson, 1980). On the other hand, the 7.4 km/s material was also interpreted as an anomalously slow and hot mantle (Flóvenz and Gunnarsson, 1991). Over the last 20 years, the SIST and the RRISP-77 experiments (Bjarnason et al., 1993; Menke et al., 1996) demonstrated crustal thicknesses of 20–24 km and 20–30 km respectively. Subsequent ICEMELT (Darbyshire et al., 1998), FIRE (Smallwood et al., 1999; Staples et al., 1997), and RISE (Weir et al., 2001) experiments all confirm a thick crust beneath Iceland with high Vp-velocity (mafic cumulates?) at its base. Compilations of crustal thickness estimates and gravity modelling (Darbyshire et al., 2000, 2000b) demonstrated a crustal thickening towards the centre of Iceland, reaching a maximum of around 40 km thickness underneath the northwest part of the Vatnajökull icecap, where volcanic productivity is still at a maximum. The deep lower crust has faster velocities than usual, around 7.4 kms -1 compared to 6.8-7.1 km/s in normal young oceanic crust (Foulger et al., 2003). Du and Foulger (2001) and Du et al., (2002 ) documented an upper crust with high velocity gradients (1-2 km/s /km) and a lower crust with much lower gradients (~0.1 km/skm). In some locations a narrow zone of increased velocity gradient at depth defines a seismic Moho, but there is often no clear discontinuity or gradient change, and as such the crust to mantle transition often has to be defined as the 7.2 km/s horizon. Receiver function studies of Darbyshire et al. (2000) also find a Moho which is often difficult to resolve.

Widespread seismological evidence, primarily in mantle tomographic models, of a cylindrical low-velocity anomaly in the upper mantle beneath Iceland (Wolfe et al., 1997; Foulger et al., 2001; Allen et al., 2002; Rickers et al., 2013; Lebedev et al., 2017; REFS….). Deeper tomographic pictures are available (Ritsema et al. 1999; Montelli et al. 2003; Hung et al. 2004) underneath Iceland but their ‘plume or not’ interpretation remains challenging and controversial (c.f. Gillian review? Iceland Review).

# DISCUSSION: THE NEA FROM RIFT TO DRIFT DISCUSION-CURENT TECTONIC UNDERSTANDING AND FEW CONTREVERSIAL POINTS

\*\*\*\*Here also.. mostly the model I have in mind but I may be wrong sometimes\*\*\*do not hesitate to complain\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*

## Inheritance impact on rifting and breakup

Pre-rift inherited crust and lithosphere before breakup contains a range of structure and heterogeneities including regional mechanical anisotropy of the lithosphere mantle, sutures and mega shear-zones prone to influence rifting and the breakup development (Tommasi and Vauchez, 2001; Schiffer et al., 2015; Manatchal et al., 2015). In the NEA, The VPM developed along the regional axis of the previous Caledonian fold belt initially squeezed between the Precambrian shield of Baltic and Laurentia (Roberts, 2003; Gee et al., 2008). This regional coincidence already suggests that crustal to lithospheric scale inheritance may primary control the rift location and possibly the breakup of the NEA (e.g. Vauchez et al., 1997; Ryan and Dewey, 1997; Krabbendam et al., 2001). In the NEA, this is long-time recognised as a classical example of the Wilson Cycle suggesting a repetition of the opening and spreading of the Iapetus Ocean which had formed the Caledonian mobile belt 500 Ma earlier (Ryan and Dewey, 1997; Petersen and Schiffer, 2016). However, in the absence of major differences in mantle temperature and composition, it has been proposed that heterogeneity may also dictate whether a rift leading to breakup is volcanic or non-volcanic/’magma-poor’ (Harry and Bowling, 1999). All these parameters may also have direct and indirect implication on the magmatic production along rifted margin and may explain fundamental difference between rifted margin end-members (Harry and Bowling,1999). Interestingly, we can already notice that the rifted margins that form directly south of the NEA are dominantly classified as magma-poor (Iberia, Newfoundland, and southern Britain) and all lie within the Variscan, Caledonidian or Acadian orogenic belts, or near one of the boundaries between these orogenic belts (e.g. Harry and Bowling, 1999; Chenin et al., 2005). On the contrary the dominant VPM of the NEA, lies near a boundary separating Phanerozoic and Precambrian basement. The resulting spatial distribution of VPM and magma-poor margins may suggest that the contrast between Precambrian and Phanerozoic crust provide sufficient heterogeneity to promote volcanic rifting, even if the mantle temperature is not unusually high (Harry and Bowling, 1999).

Schiffer et al. (2015, 2017) also proposed that the Greenland Central Fjord structure and the Flannan reflector that once formed a contiguous eastward-dipping subduction zone, possibly of Caledonian age, that may have influenced rift, magmatic, and passive-margin evolution in the NEA. The rift overlapping controlling the JMMC was most likely controlled and/or influenced indirectly by the remnant of the old Caledonian fossil subduction/suture zones also expect underneath the enigmatic GIFR.

The presences of fertile inherited eclogitic material preserved in the upper mantle have also been invoked to explain atypical melt production (Cordery et al., 1997; Yaxley, 2000; Foulger et al., 2005). The melting result of composites of eclogite patches and peridotite remains complexes and unclear especially in rift setting (Garfunkel, 2008). Korenaga (2004) suggests such an inherited compositional heterogeneity in the upper mantle (e.g. old subducted slab), multi-scale mantle mixing and alternating mantle convection may potentially explain the difference between VPM and magma-poor margin in terms of spatially and varying distribution of the fertile mantle. This model proposed for the North Atlantic suggests the recycling of old slab inherited from the Caledonian orogeny originally trapped in the upper/lower mantle interface (at the 660 km mantle discontinuity). This model suggests, however, that plate motion and mantle shearing did not destroy the direct link between the suture and the deepest subduced slab. Since seismological evidence proved that old slab may, at least, be preserved in the (present day) upper mantle (Schiffer et al. 2014). This observation may also question even more the real nature of the GIRF and the relative influence of the Iceland Plume (Foulger et al., 2005).

\*\*\*\*\*\*\*\*\*\*\*\*\*Further comment-discussion on inheritance and breakup ??? Christian, Tony ???\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*

## Superextension in the NEA: volcanic and rifted margin versus 'magma-poor' model: some contreversies

\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*Laurent, Dieter extra comments ?; I guess you agree? ???? \*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*

Ambiguities in tectonic interpretation and discussion of VPM in general and in the NEA, in particular, are often the consequence of adjacent rifted system(s) that can develop and failed long before the onset of breakup in extensive and magmatic setting (e.g. Rockall, Orphan, Porcupine basins, Faroe-Shetland Basin. It is therefore relevant to remind and precise our terminology and definition of volcanic passive margin (VPM) versus volcanic 'rifted' passive margins. The use of the "volcanic margin-VPM" expression as used in the previous chapter should rather characterise the last crustal and lithospheric domain area affected by a continuum of stretching/thinning processes and magmatism leading directly to breakup. When they primarily affect unstretched or poorly stretched continental crust, VPM of the NEA appears indeed narrower (<50-100 km), with sharp magmatic ultra-necking zone including intruded pre-existing crust and/or underplated material (e.g. White et al., 2008). On exception could be the controversial SE Greenland margin where the VPM and SDR domain is broader (up to 200 km). The expression volcanic 'rifted' margin is more generic and should embrace together the VPM (as defined stricto sensus in this study) and adjacent 'failed' rift systems that may develop before the late VPM and over a period of time of more than 100-250 Ma when well constrained in the NEA (e.g. the mid-Norwegian margin, the Faroe-Shetland margin and possibly the Rockall-Hatton margin). In such circumstances, the entire volcanic 'rifted' system can be wider but should not necessarily characterise the localised VPM itself in terms of continuum of lithospheric deformation (e.g. Gernigon et al., 2015; Theissen-Krah, 2017). Since VPM could affect pre-existing stretched continental crust and pre-existing and long-lived poly-phased rifted basins, it is likely that some of the old rift axes were already under thermal relaxation long before renewed and final continuous rift event leading to VPM formation and lithospheric breakup (Reemst and Cloething, 2000; vanWijk and Cloething, 2002). For a standard lithosphere, the diffusive time scale usually varies between 50-60 my (Allen and Allen, 2015; better REF???). In this context, it is also important to remind that the concepts of crustal and lithosphere thinning, even if often associated together (McKenzie, 1978), should be considered as two independent processes, especially when high magmatic production is involved. The present-day consideration and direct correlation of the crustal thickness in terms of lithosphere thinning (McKenzie, 1978) would not be mandatory and questionable for higher levels of rift deformation (e.g. McKenzie, 1978; Davis and Kusznir, 2004) and/or highly magmatic regime (Nicolas et al., 1994; Geoffroy et al., 2001. Franke et al. 2007).

In the NEA, large part of the continental crust underneath the platform domain and Cretaceous sag basin show ´aborted´ rifted system which is dominantly the consequence of an independent and early thinning lithospheric event(s) (e.g. Reemst et al., 2002; vanWijk and Cloething, 2002). In terms of timing, the breakup is Early Tertiary in age and does not necessarily represent the ultimate stage of a continuous and severe crustal and lithospheric deformation (the mid-Mesozoic thinning phase) that initiated more than 90 m.y.r ago in Late Jurassic time. By comparison with the final rift duration along Iberia and other expected magma-poor margins, breakup usually occurs extremely rapidly after the denudation stage and weakening of the lithosphere (e.g. Takeshita and Yamasji, 1990; Brune et al., 2016). It is also well known that when a constant extension force is small, the rate of descent of isotherms by cooling should overcomes the rates of continental thinning, leading to the deceleration of creep rates by thermal strengthening of the continental lithosphere (Takeshita and Yamasji, 1990; Yamasaki and Stephenson, 2009). Accordingly, the deceleration of creep rates after a certain amount of extension becomes stronger than underformed lithosphere after cooling even without the decay of the tectonic force (England et al., 1983; Takeshita and Yamasji, 1990). Assuming that the distal mid-Norwegian marginand most of the NEA margin experimented already a drastic phase of thinning and mantle exhumation exposed in the earliest Cretaceous time (e.g. Lundin and Doré, 2011), it might be even more difficult for us to explain how such mature and very thinned rifted system could still accommodate again both active thinning and mantle denudation over a continuous minimum period of time of ~ 50 M.y until the final breakup in Latest Palaeocene-Early Eocene time. In the distal part of the mid-Norwegian margin and for most of the similar ‘rifted’ segment of the NEA, we believe that is would be difficult to maintain a continuous thinning/denudation phase for more than 10-50 my without any strain hardening and/or breakup in between or at the end of this period. This also explains the preservation of major intermediate thick plateau (Faroe Plateau, Rockall-Hatton Plateau) in between the mid-Mesozoic rift axes of the NEA and the oceanic domain. Due to low-rate stretching and lateral heat diffusion an outwards migration of the strain from the central part and inner part of the old and cooling rift axis to a new proto-breakup axis is favoured. (e.g. vanWijk and Cloething, 2002). In our preferred tectonic model, the late Cretaceous-Palaeocene rifting event well observed in the mid-Norwegian margin the VPM would consequently develops at the western edge of the 'failed' Late Jurassic-early Cretaceous rift necking zone expected between the proto-East Greenland margin and the outer Vøring Basin. The Late Cretaceous-Palaeocene rift event represents a separate and renewed lithospheric sequence of stretching/thinning and syn-magmatic ultranecking leading to breakup. Similar processes and ‘jump’ may be explained from the Rockall and Faroe-Shetland Basin toward the west flank of the Faroe and Rockall-Hatton plateaus, event less affected by the previous mid-Mesozoic crustal and lithospheric thinning event(s). It is also acknowledged that VPM also have a tendency to develop preferentially in thicker continental domain (see Davis and Lavier, 2017).

\*\*\*\*\*Extra arguments from Laurent, Dieter ?????\*\*\*\*\*\*\*

One fundamental question also discussed in this paper is to understand if classic Iberian magma-poor models can fully apply to the NEA ‘rifted’ margin. In the previous chapters, we have already shown that VPM worldwide and in the NEA in particular have different and specific structural and magmatic elements compared to 'magma-poor'/non-volcanic margin (Iberian Type). Some studies suggests the applicability of the Iberian magma-poor concepts over to many other rifted margin worldwide including the NEA margins (Peron-Pinvidic et al., 2013, 2016). The presence of a zone of exhumed and denuded serpentinised mantle has also been suggested and favoured both in the inner and outer parts of the mid-Norwegian margin (Ren et al., 2003; Osmundsen and Ebbing, 2008, Reynisson et al. 2010, Lundin and Doré, 2011, Ruepke et al. 2013). Long-lived lithospheric weakening related to crustal hyperextension and serpentinisation was proposed to explain the locus of compressive structures along the NEA margin (Lundin and Doré, 2012) even if Kimbell et al. (2016) shows, however, that serpentinisation is not mandary to explain the location of compressive stress in the NEA. The possibility of having a large zone of exhumed continental mantle in the distal part of the mid-Norwegian margin is also an old discussion directly related to the controversial and ambiguous nature of the CLCB (Ren et al., 1998; Mjelde et al., 2002; Gernigon et al., 2004). This ´magma-poor´ (Iberian type) model applied to the NEA possibly can sometimes be tricky.

### 'Magma-poor' margin classic model: a short vade-mecums

\*\*\*\*\*\*A bit boring/classic/obvious chapter for some but not sure that all the readers are familiar with that ? – can be supress \*\*\*\*\*\*\*

Like VRM, magma-poor rifted margins (Iberian type) displays specific 'holistic' characteristics including: (1) a necking domain in which the continental crust thin to less than 10 km between proximal to distal domains (Brun and Beslier, 1996; Reston and Manatschal, 2011; Mohns et al., 2012, Nirrengarten et al., 2016; Stanton et al., 2016) (2) a so-called hyper-extended domain made of thin continental crust affected by cross-cutting faults local low-angle detachments, and (3) a zone of exhumed continental mantle (ZECM as defined by Withmarsh et al., 2001), where serpentinised mantle is directly (or almost directly) exposed at the seafloor before seafloor spreading (e.g. Mauffret and Montadert, 1987; Boillot et al., 1987, 1989; Boillot and Froitzheim, 2001; Withmarsh et al., 2001; Gussinyé et al., 2003). Notably, the nature of the COT evolution between the Newfoundland and Iberia margins have been described by numerous scientific contributions over the last 30 years ago (e.g. Sawyer et al., 1994; Reston et al. 1996, Dean et al., 2000; Funck et al., 2004; Schillinghton et al., 2006; Lavier and Manatschal, 2006, Sibuet et al. 2007; Tucholke et al., 2007; Afilhado et al., 2008; Avendonk et al., 2009; Ranero and Perez-Gussinye, 2010, Sutra and Manatschal 2012; Soares et al., 2012; Stanton et al., 2016; Nirrengarten et al., 2016 and references therein). Even if few variations exist in the details, it results from a coessential rift to breakup scenario often proposed and applied to other magma-poor margins. The onset of breakup of such Iberia –type 'magma-poor' margin is commonly explained by a sequence of tectonic processed including (1) a phase of stretching (e.g. minor thinning), which coincide with classic syn-rift and graben formation locally preserve and imaged in the continental shelf (e.g. Manatschal and Lavier, 2006; Sutra and Manatschal 2012), (2) a phase of necking which accommodate most of the thinning of the continental crust by lamination and excision of its ductile crustal material (e.g. the lower continental crust)( Withmarsh et al., 2001; Mohn et al., 2012) preceding (3) a more drastic stage of thinning (hyperextension or hyperthinning phase) which imply the embrittlement of the entire residual crust and possible sequential faulting. At that stage, brittle and semi-brittle fault could eventually provide pathways for downward migration of seawater and increasing serpentinisation processes (O'Reilly et al., 1996; Perez-Gussinyé and Reston, 2001; Bayrakci et al., 2016). This facilitates (4) a phase of exhumation and denudation of the serpentinised subcontinental mantle associated with the formation of dominant oceanic wards detachment faults partly recognised on seismic (McDermot and Reston, 2015) and ultimately (5) the development of steady-state oceanic crust (Penrose type) preceded by variable but moderate amounts of magmatism pulsing during the breakup (Piccardo, 2016; Stanton et al., 2016). The zone of exhumed continental mantle (ZECM) shows a different range of mantle petrology, from what geochemically determined continental or oceanic affinities. The result is not unambiguous (e.g. Seifert et al., 1997; Chazot et al., 2005; Müntener et al., 2010). In terms of rift duration, previous reviews suggest that the complete tectonic sequence Stretching/Thinning/Mantle exhumation/Final Breakup usually occurs in less than 30-50 my (Bowling and Harry, 2001; Reston and Manatschal, 2011; Brune et al., 2016).

### A few rifted margins problems in the NEA-model-driven versus observations

More recently, a comparison with 'magma-poor' margin and the presence of a large zone of exhumed mantle has been proposed again to explain the onset of breakup in the outer Møre and Vøring margin (Péron-Pinvidic et al. 2013). The floor of the Outer Vøring Basin and the crust lying directly underneath the Paleogene sediment and basalt from the South Gjallar Ridge, Ygg High to the Vøring Marginal High (Figure x) has equally been interpreted as exhumed subcontinental and heavily serpentinised mantle. (Péron-Pinvidic and Osmundsen, 2016). To some extent, these authors tried to make us believe that some similitude with the final magmatic development earlier at the distal edge of the Iberian transitional exhumed subcontinental mantle (Brönner et al., 2011). The validity of this conceptual model has some strong implications about how the breakup and opening of the NEA originally and really initiated. It is probably not so straightforward.

First of all, the paleogeographic environment and tectonomagmatic events initiating the NEA VPM are probably very different from the deep-water setting of the Iberian and Newfoundland margins at breakup time. The dramatic amount of magma emplaced during the breakup and the thick syn-rift and post-rift sedimentary successions observed on top of the thinned crust (>8-9 km) also represent major differences compared to the deep-water, sediment-starved, Iberian margin. Even if there are minor to moderate indications of locally undifferentiated pre-, syn- and post-rift magmatism (Whitmarsh et al. 2001) along 'magma-poor' margin, it is considerably less compared to the large volume of magma that produced and erupted along the Norwegian VPM during the onset of breakup (e.g. Eldholm and Grue, 1994; Berndt et al. 2001).

Compared to the Iberian margin, where the basement and mantle are directly constrained by boreholes, the nature of the basement and sub-sag substratum before and during the VPM development remains speculative. Even if no consensus (e.g. Tucholke and Sibuet, 2012); Brönner et al. (2011) propose indeed that a volcanic feature (so-called 'outer high') resulted in syn-breakup underplating of a previously exhumed continental mantle required to initiate the breakup between Iberia and Newfoundland. According to the paleogeographic scenario, Crosby et al. (2011) suggests a water depth at the COT likely deeper than 2 km assuming than some of the post-rift sediments indicate deposition below the calcium compensation depth. Peron-Pinvidic and Manatschal (2008) suggests that before the breakup, the top of the Exhumed serpentinized preceding this volcanic feature event was located at more than 4-5 km in depth. Similarly, Péron-Pinvidic and Osmundsen (2016) suggest that heavily serpentinised exhumed mantle could lie directly underneath the first lava flows and Outer High earlier defined in the Vøring Marginal High (Planke et al., 2000). Indirectly this model suggests that the VPM in the NEA should develop like magma-poor margin.

For isostatic and mechanical reasons, it looks difficult for us to understand how a zone of exhumed continental mantle could reach such very shallow water levels before the onset of drastic volcanism. Available wells in the outer Vøring confirm that a large part of the outer ridge complex was characterised by deep neritic (150-250 m b.s.l) to shallower paleo-water environments during the Palaeocene (Ren et al., 2003). The emplacement of subaerial continental lava flows and adjacent Lava Delta emplaced along the shore line (Figure X) also confirm that the onset of breakup in the Vøring Marginal High. Shallow water the subaerial basalt, as undoubtedly proved by ODP well 642E. In this context, we already question the ambiguous analogies between the Outer High defined in VPM (Planke et al., 2000) with the volcanic (but possibly different) 'outer high' feature involved (or not) in the final breakup of the Iberia-Newfoundland conjugate margins. As previously mentioned in chapter 5, the original definition of Outer High in VPM environments characterises very shallow marine to subaerial surtseyan volcanics stratifically emplaced on top of the subaerial Inner SDR, already showing oceanic affinities (Roberts et al. 1984; Planke et al., 2000; Meyers et al., 2006). The 'outer high' in the Iberian margin emplaced on top of the exhumed serpentinised mantle in a different and deep water paleo-enviroment (Brönner et al., 2011). The magma-poor 'outer high' also coincides with a pulse of volcanism, whereas the VPM Outer High emplaced during thermal subsidence and after the dominant volcanic phase older and rather associated with the thick Inner SDR wedge (Figure x).

Finally, the geochemical signature of the silicic volcanism drilled underneath the tholeitic basalts of the Landward Flows suggests that continental crustal anatexis has occurred during the early lava flows eruption at shallow to subaerial levels. ODP well 642E (Figure X, Y) recovered peraluminous, cordierite bearing basalts, andesite and dacitic flows interbedded emplaced at shallow water depth at the base of the volcanic trapps. Both Petrological and geochemical results suggest strong evidence of continental contamination and rather support the presence of continental crust over a large part of the Vøring Marginal High (Meyer et al., 2009; Abdelmalak et al., 2016) (ROMAIN EXTRA COMMENTS+DEPTH OF MELTING INDICATION). The well result do not favour the presence of an enigmatic zone of exhumed and serpentinized mantle directly underneath the Lower Flows

### Geophysics and crustal thickness estimation

A zone of exhumed mantle preceding the onset of magmatism in the distal part of the Mid-Norwegian margin is not clearly obvious, nor formally identified. The size of the crustal bloc preserved in the deep parts of the margin are locally different and 2-3 times thicker and compared to the wider allochthonous blocks described from the Iberian abyssal plain (see 1 x 1 crustal scale comparison in Gernigon et al. 2015). When dealing with the distal part of the mid-Norwegian margin, previous crustal studies of Reynisson et al. (2010) has also illustrated radically opposed and contradictory crustal models in the distal and controversial outer Møre margin. Reynisson et al. (2010) suggested the presence of a very thick continental ribbon at the outer Møre Margin (15-20 km in thickness at the Møre Marginal High). This estimation was possible (but slightly) overestimated due to long-wavelength errors in the vintage ship track compilation used in the previous study (Haase et al., 2014). Nevertheless, Nirrengarten et al. (2014) and Gernigon et al. (2015) agree that the continental crust is present and still thick in the distal part of the Møre Basin and Møre Marginal High. Both updated potential field results estimate that continental crust on top of the distal CLCB is locally too thick (> 5-8 km) to systematically correlate with an underlying and heavily serpentinised and dense crustal body (e.g. Maystrenko et al., 2017). Recent OBS profiles across the Møre Basin from Kvarven et al. (2014) also confirms the presence of a minimum of 5 to 10 km of continental crust (Vp observed from 6.1 to 6.7 km/s) expected from the central part of the main Cretaceous depocentre to the Møre Marginal High. In most of the central Møre Basin, the OBS modelling does not always show the existence of systematic LCB, which is mostly restricted to the outer Møre Basin and under thick continental crust likely expected in the Platform and coastal area (Kvarven et al. 2014). More recent 3D potential field modelling in the mid-Norwegian margin confirms that the continental crustal is preserved in a large part of the Møre and Vøring sag-basin (Maystrencko et al., 2016). The CLCB Vp values of 7.2-7.8 km/s published by Kvarven et al. (2014) could be compatible with layers of serpentinised mantle (see review by Mintschul 2009). However, velocities of less than 7.5-7.6 km/s would better fit with a drastic crustal thinning, a thickness of the overlying continental crust of less than 3-4 km and/or simply a complete denudation of heavily serpentinised mantle (e.g. Nirrengarten et al. 2014). Based on 3D velocity model, Bayracki et al. (2016) also show that velocity of less than 7.5 km/s due to serpentinisation is often restricted the uppermost parts (2-3 km maximum) of the serpentinised layer when affected by the major brittle faults controlling the flux of water into the mantle. It must be mentioned that in the previous case, the exhumed mantle and the top basement on top are directly in contact with sea water and the brittle faults affect the entire crust.

In other similar sag basins of the NEA (e.g. North Rockall Basin, Faroe-Shetland Basin, the crust is obviously thin but not so drastically thinned and at the limit of real hyperextension (e.g. >10 km). Symptomatic high Vp velocity is also not always acknowledged in the deep parts of the basin (Kingelhoefer et al. 2005). The lack of CLCB modelled in the inner part of the Møre sag basin (Kvarven et al. 2014) also shows that even a very thin crust does not always and necessarily coincide with an underlying CLCB usually symptomatic of mantle serpentinisation (Kvarven et al. 2014). For realistic rheologic models of the lower continental crust and assuming moderate thinning of the crust (> 5-10 km), detachments fault may preferentially root in the mid crust or at the mid crust/lower crust interfaces and never reach the mantle (Lavier, 2017; GSA communication, proper REF later). Since several kilometres of sediments and faults decoupling are observed along the thick sag basin of the NEA, it may also prevent easy fluid migration through the entire crust, leading to extra interrogations about massive serpentinistation of the deep mantle rocks.

If exhumed mantle exists in the deep part of the Vøring and Møre basins and similar rifted basin in the NEA, it may indeed represent small tectonic windows locally possible in the thinnest parts of the sub-basins (e.g. the southern Rockall Basin, the Traena Basin, the Jan Mayen corridor). In most of the NEA, we may rather deal with an intermediate "super-extension" stage during the mid-Mesozoic thinning event rather than an ambiguous hyperextension more characteristic of the southern conjugate margins (e.g. Iberia-Newfoundland, Labrador).

In the outer Vøring Basin, the controversial interpretation and meaning of the high CLCB Vp velocities has been already discussed several times in the past. Based solely on seismic observations we cannot easily decide on one hypothesis (e.g. Ren et al., 1998; Gernigon et al., 2004, Ebbing et al., 2006). Since serpentinization is a gradual process with no clear interface the presence of Moho reflections at the base of the CLCB, Mjelde et al., (2002) also argued against the model of serpentinized peridotites. Together with a sharp vertical and lateral velocity gradient surrounding the LCB, the hypothesises of inherited lower crust and/or underplated material were favoured. Magnetolluric experiments and strong conductivities in the deep crust of the Vøring Basin also disregard the presence of heavily serpentinised mantle material (Myer et al., 2013). Analysis of the magnetic properties of the rocks samples shows that serpentinised peridotites often show high remanent magnetisation and susceptibility regardless the serpentinisation degree (Oufi et al., 2002; Brönner et al., 2014; Bonnemains et al., 2016) and consequently should produce a pronounced magnetic signature. However, flat and low magnetic signature on top of the CLCB/T-Reflection complex recognised below the distal ridges of the Outer Vøring Basin does not easily support the presence of mafic or serpentinised magnetic material in depth (e.g. Gernigon et al., 2004; Ebbing et al., 2006; Abdelmalak et al., 2017).

Thicker continental crust and moderate pre-volcanic thinning in the distal part of the rifted margin and possibly underneath a large part of the volcanic province would already solve the paleo water depth issue easily and the evidences of continental contamination recorded by ODP well 642E. From a tectonic point of view, the outer Møre and Vøring basins (including parts of the Møre and Vøring marginal highs) together with the proto-JMMC probably represented the remnant of a continental marginal plateau later dissected by the late line of breakup. This paelo marginal high was less thinned compared to the Late Jurassic-Early Cretaceous aborted inland rift system that originally and mostly focused in the inner parts of the Mid-Norwegian margin (E.g. Rås and Træna basins) and possibly along the Jan Mayen corridor possibly affected by lateral friction. The preserved continental crust partly covered by the Tertiary lava flows, to the west, would also extend in the northern prolongation of the Fugløy Ridge and southern Rockall-Hatton Plateay clearly showing itself thicker continental crust, and a very abrupt magmatic continental-ocean transition (White et al. 2008; White and Smith, 2009). As previously mentioned in the first part of this review paper, presence of continental crust underneath the Inner SDR is not surprising and other ODP drillings (Leg 152), the most recent deep seismic observations interpretation and possible modern VPM analogues (e.g. Southern Red Sea) often agree with the presence of continental crust underneath the basalts and SDR (White and Smith, 2009; Clerc et al., 2015; Geoffroy et al., 2015; Patton et al., 2016; Guan et al., 2016; Nemcôk and Rybár, 2016). Pre-rift Upper crust and exhumed lower continental crustal rocks might be a better option for the original CLCB later affected by the Paleogene underplating and related flood basalts in the Vøring Marginal High.

In an attempt to be more critical and cautious with model-driven interpretation of distal rifted margin in general, we also note that the applicability of the classic Iberian model is also challenged for other ´magma-poor´ margin (Blaich et al. 2008, Clerc et al. 2015, 2016, Cowie et al. 2015) and sometimes described as questionable and/or at best limited (Mohriak and Leroy 2012). In the South Atlantic, models have particularly shown that underneath such large and thick sag-basin and asymmetric margin, the presence of exhumed serpentinised is not mandatory and/or questionable even if associated with significant crustal thinning (Contrucci et al., 2004; Aslanian and Moulin, 2011; Sibuet and Tucholke, 2012). Presence of salt diapirs and/or local volcanism does not always allow to clearly understanding the deep crustal structures directly of the rifted margin. Based on numerical simulation, Huissman and Beaumont, (2014) and Brune et al., (2014) showed that amounts of continental material (including lower and/or upper continental crust) should be preserved and should flow underneath large and thick sag basins (quite similar to the Vøring and Møre basins) to explain their geometries. Preserved continental crust in a highly-extended regime has also been suggested and/or described beneath similar sag-basinal systems. There are interpreted as Type-II margins in the South Atlantic (Brune et al. 2014, Huissman and Beaumont, 2014). Similar scenarios could apply to the NEA rifted margin but most likely in the early stage of the rifted margin development (e.g. the Late Jurassic-Early Cretaceous rift episode).

## Melt production and plume versus non-plumes discussion

(MAY BE TO SUPPRESS? CF Gillian?

The origin of the breakup magmatism and formation of a large igneous province and VPM as observed in the NEA, most likely involved not only inherent (e.g. rift and far-field force related) but also involved extrinsic geodynamic processes including controversial hot mantle plume and/or sub-lithospheric processes (White, 1992; Meyer et al., 2006; Hansen et al., 2009) and/or the presence of preserved, fertile, subcontinental mantle material associated with pre-existing suture zones (Foulger et al., 2005). Much of the thinking about melt production formed along the NEA VPM is often based on the classic paper of White and McKenzie (1989). According to the conventional approach magmatic production is the consequence of one of two processes: 1) a high temperature plume/hot-spot that serves to replace sub-lithospheric subsolidus with super solidus mantle and/or 2) an adiabatic decompression of the shallow mantle due to lithospheric thinning. In both cases, mantle melting can be locally enhanced if the sub-lithospheric mantle is hot or melt prone (McKenzie and Bickle, 1989; Wilson, 1993). Such melt could percolate through the lithosphere by diffuse porous flow and/or focused flow inside pre-existing mantle shear zones (e.g. old suture) or ductile shear band simply driven spontaneously during the rifting (Weinberg and Regenauer-Lieb, 2010; Mohajeri et al., 2013; Piccardo, 2016).

8.3.1Mantle plumes hypothesis

Assuming purely active mode, hotter-than-normal mantle dynamic feature (e.g. mantle plume, small-scale convection) could thermally thins and/ plug first the thick lithosphere. Assuming uniform mantle composition of the mantle, adiabatic melting occurs would subsequently increase in time during rifting and/or breakup (McKenzie and Bickle, 1988; Wilson, 1993). Fundamentally funded on the previous concepts, VPM have been the subject of numerous experiments and more advanced numerical, geochemical, geophysical and geodynamic studies aiming to further understand how a large production of intrusive and extrusive can form during the onset of breakup. VPM and Large Igneous Provinces in general (e.g. Ernst 2014) are often interpreted to be produced by significant melting within mantle plumes as they reach the base of the lithosphere. Local increases of mantle temperature (or Hot-spot) are often explained by the influence of mantle plume (e.g. Morgan, 1971; White and McKenzie, 1989). Thermal models for this (White and McKenzie, 1989) demonstrate that an increase in mantle potential temperature of 50–100 ° C would produce the 20–30 km thick crust seen across most of Iceland. The REE compositional variation, topographic, geoid and crustal thickness anomalies can therefore all be explained by a plume of elevated mantle potential temperature and passive decompression melting of the hot mantle at the oceanic spreading centre. Geochemical studies have additionally suggested that the hotspot source contains some component of a pyroxenitic lithology, possibly from recycled oceanic crust entrained in the plume flow, because of the variation of major element compositions in erupted products along the ridge axis. This additional source, however, is very fusible and so melting models of this bi-lithology source indicate that the bulk of the melt volume, and therefore crustal thickness variations, are still primarily controlled by mantle temperature and plume flow variations (Shorttle and Maclennan, 2011).The interaction of the hot mantle plume with an oceanic spreading centre has resulted in a hotspot crustal structure in Iceland which is more complex than at typical ocean ridges.A number of different types of plumes (or plume clusters) have been proposed from deep plume, superplume or plume originating from the upper/lower mantle transition zone (Courtillot et al., 2003; Maruyama et al., 2007). The area covered with traps is usually considered as coincident at depth with a melting head of a 'mushroom-like' mantle plume which is a simplistic representation of the way basic magmas migrate from the deep mantle to the Earth surface (Richards et al., 1989; Campbell and Griffiths, 1990, Saunders et al., 1992). (1) Seismic tomography (Montelli et al., 2004; Nolet, 2007); pre-basalt domal uplift /Campbell and Griffiths, 1990; Campbell, 2007); (2) reconstructed Large Igneous Provinces to the periphery of deep-mantle anomalies and/or geoid highs (Cross and Jurdy, 1980; Courtillot et al., 2003; Torsvik et al., 2010), (3) radiating mafic dike swarms (Ernst and Bunchan, 1997) and/or (4) Geochemical composition of the magma (Hawkesworth and Scherstén, 2007) have often been considered as key test arguments to support the presence of deep plumes. Present day hotspots and Sublithospheric "topography" may also have a profound effect on lateral flow of plume material (Sleep, 2006) and lithospheric segmentation may play a role in the pattern of volcanism during breakup and could explain the occurrence of VPM (Taposeea et al., 2016). Sublithospheric variations are indeed related to the presence of old cold and thick cratonic lithosphere and/or the thinning effect of pre-breakup rifting event(s). Temperature anomalies in the range of 50-200 C derived from the mantle plume has been earlier proposed to explain the amount of melt expect along VPM (Pedersen and Skogseid, 1989; Pedersen and Ro, 1992).

Assuming a magmatic and syn-breakup origin, independent trace-elements geochemistry and Vp-waves velocity modelling of the LCB also suggest that the adiabatically melting hot mantle rises actively upward through the basal lithosphere (Holbrook et al., 2001). Along the NE Atlantic VPM, Holbrook et al. (2001), Breivik et al. (2009) proposed that the rate of active mantle penetration through the lithosphere could be up to ten times as large as the half-rate of lithosphere stretching. To explain VPM formation in the NEA, Armitage et al. (2009) also showed that excess breakup magmatic should require a pre-existing thermal anomaly of 200C. Armitage et al. (2009) conclude that without thinned lithosphere prior to breakup there would be no excess melt production due to the thickness of the lithosphere dampening any upwelling. They argue that, no matter the thermal state of the lithosphere, the lithosphere must have undergone some extension prior to breakup

Usefull????????

7.3.2 Alternative dynamic models in the NEA

Because VPM are clearly associated with pre-existing rift and breakup processes; plume or not, we cannot easily neglect the plate-driven divergence and the restricted lithospheric mechanisms involved during the rifting and its transition to breakup. A second group of studies based on several observations and models have, puts thinks and classic 'plume' model(s) into perspective. The controversial issue of the mantle reservoir(s) that feed the volcanic traps and the ambiguous modelling and interpretation of deep geophysical tomographic model have been the topic of a large number of publications and debate over the last 10-15 years (e.g. Foulger and Anderson, 2004; Anderson and Natland, 2005; Lundin and Doré, 2005; Meyer et al., 2007; Foulger, 2010; Franke et al., 2013). Specifically, Foulger (2010) highlights and summarises limits and ambiguities with classic plumes models and concepts. Foulger (2010) suggests that propagating cracks, internal plate deformation, membranes tectonics and breakup or rift related processes can alternatively explain large amount of magmatism without the need of any mantle plume. Unlike mid-ocean ridge magmatism, which may be adequately modelled with steady-state, passive mantle upwelling driven by large-scale plate motion (e.g. McKenzie and Bickle 1988), continental rifting is essentially a transient process, in which an initially null surface divergence evolves into finite-rate spreading. Some earlier concepts and numerical models have proposed that upper mantle temperature variation can simply be caused by restricted lithospheric processes such as small-scale convection induced by either rifting and/or sharp discontinuities in lithosphere thickness (Mutter et al., 1988; Boutillier and Keen, 1999; Anderson, 2000; King and Anderson, 1995) or viscous shear flow induced by heterogeneities in the asthenosphere (Conrad et al., 2010). Even if probably more complex, the dynamic aspect of sublithospheric convection, the role of stress and dependent viscosity, may be viable and also important in calculating the amount of melt generated during rifting and VPM (Korenaga and Jordan, 2002). The resulting convective upwelling rate could exceed the plate-driven divergence rate and produce ‘excess’ magmatism due to rapid mantle fluxing through the VPM melting zone (Korenaga and Jordan, 2002; Simon et al., 2008). Similar models suggest that excess volcanism during breakup could alternatively be the result of a combination of active upwelling and temperature anomalies since convection alone will not penetrate the young lithosphere (Nielsen and Hopper, 2002). Using different finite elements (VanWijk et al., 2001) and finite difference modelling approaches (Gernigon et al., 2006) also show that focusing of deformation results in fast differential thinning factors and increasing melt production along the proto-breakup axis. Dependant on extension velocity and/or solidus variation range, the amount of melt predicted assuming a 'normal' initial mantle temperature of 1333C was enough to explain the amount of melt expected at VPM (Norwegian VPM study case). However, such rift-related models fail to explain why contemporaneous or slightly younger magmatism locally formed in poorly or unstretched domain, far from proto-breakup-axis (The Scottish volcanic intrusions for example).Here we have to explain why, can not be easily related to the rifting itself !!! Should be an independant mantle process !!!!

We should discuss this point at Durham 3 ???

Luc model...extra comment later

## What is breakup? And how does it form in the NEA

The COTs of the conjugate volcanic margins that developed on both sides of the NEA may partly represent thinned and progressively intruded continental/transitional crust. As previously mentioned the true nature of the intervening transitional crust in most of the NEA basins and VPM is still unclear and open to interpretation from thin intruded thick to relatively thin continental crust, to exhumed continental mantle. Significant dike swarms and mafic intrusions are likely in the vicinity of the SDRs as observed onshore West and Southeast Greenland, where volcanic margins are locally exposed onshore (Karson and Brooks, 1999; Geoffroy et al., 2005). The COT underneath the basalt could represent an alternation of micro-continental fragments interbedded with embryonic oceanic crust, overlaid by volcanic traps, as proposed, for example, in the Afar region (Courtillot et al., 1980; REF Laurent ????). It could also represent intruded and flowing ductile crust as suggested in NE Greenland (Quirk et al., 2014). \*\*\*Laurent not totally agree ?\*\*\*\*

In the distal part of the NEA VPM margin, the substratum of the Inner SDR wedges most likely represent stretched continental crust with basins moving gradually by increasing dike injection towards a MORB-type igneous crust. Such features are typical of all VPM mentioned in this first part of this review. Different tectonomagmatic processes and structural styles characterise the breakup in highly magmatic provinces, making them different from 'magma-poor' (e.g. Iberian-type) margins. In the previous chapters, it was showed that magmatic underplating, plumbing and massive diking of stretched continental crust in most likely in the VPM, possibly introducing inherited upper and lower crust likely preserved nearby the COT of most of the NEA VPM.

The onset of magmatism likely helped to locate, pulse and trigger a complete lithospheric rupturing of magma-rich margins. Softening and thermomechanical erosion triggered by high magmatic production could lead to whole lithospheric failure and consequently to a rapid final breakup (REFS……. In the case of VPM, the rupture of the lithosphere can occur even before the rupture of the continental crust (Geoffroy et al., 2000; Franke et al., 2005). Also, a non-negligible specificity of numerous VPM.

So-called 'Asthenospherization' or asthenospheric plumbing of the lithosphere may not only restricted to the mantle but may also affect the overlaying continental crust as suggested by the evidence of crustal anatexis recorded in the Vøring Marginal High lava flows (Meyer et al., 2009; Abdelmalak et al., 2016). Intermediate (basaltic –andesitic) to felsic (dacitic and ryolitic) Lower Series appears to be produced by large amount of melting of upper crustal material (Eldholm et al., 1989; Meyer et al., 2009). 'Asthenospherization' and rupture of the lithosphere can occur even before the continental crust break apart (Nicolas et al., 1994; Huissman and Beaumont, 2014). In could be the case in the mid-Norwegian margin and possibly along other VPM segments of the NEA, unfortunately less constrain by data. The presence of a weak zone and a destabilised mantle may favour faster strain localisation, drastic increase of magmatism and may enhance doming and thermal buoyancy of deeper/hotter asthenospheric material upwelling underneath the VPM during SDR formation. Possible and modern analogue for the VPM formed in the distal part of NEA VPM could be found in the Red Sea region where similar nucleation and the initial emplacement oceanic crust occurs in discrete axial cells at present day after formation of major volcanic traps observed along the margin (Bonatti et al., 1985; Wolfenden et al., 2005; Ligi et al., 2011). Evidence of asthenospheric plumbing of pre-existing continental upper crustal level is also directly observed in the field (e.g. Zabargad Island and Brother Islets: Bonatti and Seyler, 1987; Boudier et al., 1988; Dupuy et al., 1991; Bothworth et al., 1996; Nicolas et al., 1994). Before seafloor spreading, Boudier et al. (1988) show evidence of strong ductile deformation, flow structure and anatexis associated with the rising of intra-lithospheric peridotite intrusion within pre-existing granulite lower continental crust. Before unfortunate comparison with 'magma-poor' margin (e.g. Lundin et al., 2014) it is worth mentioning that the peridotite intrusion observed in Zabargad Island was associated with extremely high temperature of the emplacement (>1000C). It must also be precised that such an asthenospheric diapirism does not show evidence of pervasive serpentinisation possibly due to the high temperatures involved (Boudier et al., 1988, Bothworth et al., 1996).

Eventually, the active drifting force induced by the upwelling asthenosphere could drive the transition from passive to active and could force the final rupturing of remaining the upper continental lithosphere (e.g. Nicolas et al., 1994; Gac et al. xxxx, Laurent REFS??). Eventually the final step of breakup along the NEA margin may be associated with a sudden decrease of the magmatic production, as observed in most of the VPM after the SDR Emplacement. A rapid change of the magmatic condition may directly influence the rheology and the pre-existing structure of the magmatic continental-ocean transition. Eventually, it may trigger to a much localised 'flip-flop' detachment tectonics that may affect the SDR before breakup (Geoffroy et al., 2015). It is also possible that the final breakup along VPM segments of the NEA may be the end-result of melt assisted extension while others just undergo a minimum melt-assisted extension at the outer edge of pre-existing SDR (e.g. Nemcôk and Rybár, 2016). However, in terms of magmatic development and segmentation, VPM may share more characteristics with fast magma-rich spreading oceanic axes (Carbotte et al. 2015) than slow to ultra-slow spreading oceanic systems, which are magma-starved and often mentioned for comparison with the zone of exhumed mantle domain observed at magma-poor margin (Cannat et al., 2009; Brönner et al., 2014, Dieter REFS?). Along the mid-Norwegian margin and NEA VPM magmatic production closer to the final breakup stage may exceed the average magmatic production along the present-day magma-rich segments of true oceanic spreading ridge. It could display similar subdued rift valley and an elevated axial floor consistent with a hot thermal regime and a thin axial lithosphere. By comparison with magma-rich oceanic segment, a thin lithosphere scenario and high temperature may not necessarily favour a pervasive process of serpentinisation except possibly nearby the proto-oceanic fracture zones (e.g. The Jan Mayen Fracture Zone, Skelton and Jakobsson, 2007).

\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*I also guess no evidence of exhumed serpentinised mantle in Iceland \*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*

Finally, the onset of breakup in the VPM does not necessarily mean that a total lamination of the (pre-rift and original) continental crust is mandatory, before total lithospheric rupture and oceanic accretion (i.e. there is no need for subcontinental mantle denudation underneath the SDR). It might be the case for the mid-Norwegian margin as recently suggested by the more recent seismic data (Theissen-krah et al., 2017; Abdelmalak et al., 2017) Perservation and anatexis of continental crust before breakup might also be a significant difference compared to classic 'magma-poor' system. Based on independent observation in the South Atlantic, Franke et al. (2013) also concluded that at VPM the lithospheric mantle should break first or at the same time with the crust, resulting in the emplacement of huge amounts of syn-rift magmatic extrusives and intrusive.

According to the well ODP 942E results inherited continental crustal might be present underneath the SDR wedges and in the prolongation of the inherited CLCB - T-Reflection (e.g. Gernigon et al., 2006; Meyer et al., 2009; Abdelmalak et al., 2017). It is also possible that moderate thinning and preservation of continental crust may also favour and control the SDR development. Preserved 'C-Block' in between conjugate SDR wedges is also expected to represent residual continental crust, controlling the SDR development on both sides (Geoffroy et al., 2015; Guan et al., 2016). Similar residual blocks may be expected and preserved in the NEA or rapidly dislocated and assimilated with the oceanic crust.

## Tectonic buffers, locking zones and oceanic rift propagation in the NEA

In the NEA, the onset breakup is often regarded as instantaneous across the ruptured plate. In detail, the process of lithospheric continental breakup and the onset of true seafloor spreading is probably not instantaneous over the great distance of xxxx km of the NEA. Assuming rigid plate, and homogeneous lithosphere, the oceanic rift should, in theory, propagates over time through the extended lithosphere as a natural consequence of the relative motion about the pole of rotation (Vink, 1982). Alternatively, if some degree of distributed deformation is allowed, then the rift zone or VPM can propagate by a process of plate 'unzipping' (Vink, 1982) as proposed for the South Atlantic VPM evolution (Franke et al. 2003) and recently suggested for the NEA (Lundin et al., 2014). It is important, however, to emphasise that the original unzipping models of Vink (1982) was specifically discussing the onset of continental rifting; regardless of how this onset of breakup really occurred. This model already would imply that the COBs in the NEA are not isochron and could display a progressive and linear evolution only if plate tectonics are applied rigorously at all scales of observation (Vink, 1982; Martin 1984). The later development of an oceanic rift could proceed by a similar or different geometrical/kinematic history. It is notably the case of the Norwegian-Greenland Sea where inheritance, diachronic magmatic intrusion already contributed to a more complex and heterogeneous crustal, lithospheric and asthenospheric configuration even before the onset of breakup. Thus we expect different parts of the NEA margin to react differently during the rifting and processes, resulting in a non-linear scenario where breakup and magmatism could occur diachronically along different segments of the proto-VPM, which explain notably the age difference observed between the More and Voring VPM for example.

Courtillot (1982) also emphasised earlier the importance of 'locked zones' or volcanic rift propagation barriers (e.g. Franke et al., 2007) that impedes the lateral growth of the oceanic rift. Between successive locked zones, the rift could either propagate laterally or instantaneously rupture the lithosphere. Presuming that the transfer zones formed prominent lithospheric discontinuities at the onset of rifting Franke et al. (2007) suggested that these strongly influenced the generation of melts and VPM segmentation. Conversely in the NEA, some of the major fracture zone (e.g. the Jan Mayen Fracture Zone are the locus of atypical and post-breakup production (e.g. the Vøring Spur; Breivik et al., 200x; Gernigon et al., 2009). Few locked zones or continental buffers could correspond to the positions of the pre-rift lithospheric rigid structures that might be favourably oriented to evolve into transform faults in the ultimate oceanic rift system. Similarly, we believe that the pre-breakup configuration, including the latest stage of the VPM development, most likely influenced the distribution of the main ‘offsets’ that have been observed and described in the Norway Basin oceanic domain. A complete 3D/4D crustal picture of all NEA is still not properly satisfied, but Gernigon et al. (2012, 2015) and Maystrenko et al., 2017 already point out that the segmentation of both upper crust and lower continental between the Møre and Vøring Basin represent a complex crustal transfer zone (the Jan Mayen corridor) that correlate well with the segmentation of the oceanic domain. It influenced the initial segmentation and early magma distribution towards the COT and indirectly, the proto and late segmentation of the adjacent oceanic spreading system. The initial distribution and diachronism of underplated bodies emplaced along the VPM (controlled or not by other inherited features or unclear sub-lithospheric anomalies) most likely influenced further the localisation of the deformation and segmentation of the early spreading cells. The presence of thick (Archaean) crust in the Faroes Plateau probably influenced the early stage of spreading in the Norway Basin and the subsequent stress realease that may partly explain the C22-C21 spreading reorganisation of the Norway Basin and the onset of rifting/spreading in the southern part of the JMMC.

In the NEA, it is not necessarily valid that rifts, VPM and embryonic spreading ridge would propagate linearly in one direction or that crustal stretching and melt production increases with progressive propagation of a rift suggested by recent aeromagnetic data (Gernigon et al., 2012;2015). In the early stage of breakup, the degree of melting in the NEA does not look directly related to any linear spreading rate development. Already in the oceanic domain, the melt production (e.g. crustal thickness of the oceanic crust) appears more dominant in the southern Norway Basin closer to the Faeroes-Iceland Ridge, itself located nearby the rotation pole proposed for the Norway Basin (Off course this also depends on the real nature of the GIFR and our ability to understand later where and how much continental crust is preserved along this atypical ridge). This may suggest that the onset of breakup initiated in the central/northern part of the Møre margin before growing laterally and before the propagation of the proto-spreading ridge. A similar situation might be expected during development of the VPM and diachronic SDR emplacement possibly migrating from North to South between NE Greenland and not from South to North if we agree that the rotation poles of the Eurasian and Greenland plates are effectively located further to tho north (This would contradict the classic plate unzipping kinematic approach).

\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*Still Controversial Due to the lack of NE Greenland data , Dieter\*\*\*\*\*\*\*\*\*\*

In the Norwegian -Greenland Sea, the first SDR possibly develop in the outer and central part of the outer Møre margin and Jan Mayen corridor before further development along the Faroes margin and Vøring Marginal High to the north. At the same time if possibly initiated between the Lofoten and NE Greenland margin. Our observations suggest that VPM, SDR emplacement and late oceanic accretion may initiate at different areas, independently of the regional plate motion during the early stage of breakup. The rift may subsequently accrete in different directions. This corroborates the old observations by Martin (1984) that suggest that an oceanic spreading centre could also appear within a segment of heterogenous and stretched continental lithosphere without direct and linear correlation with the propagating plate motion. In the Red Sea, Almalkil et al. (2016) also describe embryonic oceanic segments that may stall or hibernate before rejuvenating. The development of discrete spreading segments (e.g. Bonatti, 1985) is comparable to development in other juvenile oceanic basins such as the Woodlark Basin (Taylor et al., 1995, 2009). These observations may suggest a diachronic, ‘non-unzipped’ and segmented development of the early spreading system along the NEA.

## Spreading Overlap, GIFR and Microplate Complex

The development of an adjacent concomitant continental rift zone eventually leading to the JMMC in the NEA still remains questionable and may involve the presence of additional 'buffer' still active after the first episode of breakup in Early Eocene. Thermal weakening of the lithosphere induced by mantle plume is often been considered as a primary explanation for microcontinent dislocation (Müller et al., 2001 and Mjelde et al., 2008). However, some models have questioned the systematic implication of plume during the process of dislocation (Yamasaki and Genrigon, 2010; Abera et al., 2016; Nemcok et al., 2016). Independently of the presence of any kind of mantle anomalies, an early overlapping configuration could be a key condition required for microcontinent formation as earlier put forward by Auzende et al. (1980) and Unternehr, (1982) before the popular plume-related model of Muller et al. (2001)??? which has never been modelled to my knowledge ????.

Alternatively, plate overlap is a simple and natural mode to isolate micro block or microplate. Koehn et al. (2008) also show in a similar way that even stable and relatively homogenous (e.g. a model) lithosphere under extension can develop into a microplate system just by natural overlap of rift segments. The conceptual and favoured model put forward in the present paper to explain JMMC suggests a similar natural rift and primary plate mechanism for microcontinent initiation (e.g. Auzende et al., 1980). This overlapping model for the JMMC is, however, not necessarily incompatible with the concept or the presence of a mantle "plume" or any other kind of mantle anomaly in the NEA. Numerical models by Yamasaki and Gernigon (2010) already questioned the plume model but also emphasise the importance of the proto-microcontinent stage and the presence in depth of LCB inherited or formed during the early VPM development stage. Late sub-crustal magmatism associated with a rising plume or other kinds of mantle anomalies could be a factor that could facilitate and trigger the final or the initial insulation of the microcontinent (Gernigon and Yamasaki, 2009). Favoured tectonic scenario highlights the importance of a distinct independent and overlapping continental rift and/or spreading system not directly linked and coupled with the growing Aegir Ridge (Gernigon et al., 2012, 2015; Ellis and Stoker, 2014. Similar scenario was supported recently by Schiffler et al. (2017) and Stoker et al. (2017). Recent scenario of VPM also required the preservation of a residual continental block in between pairs of SDR. The 'C-block' model of Geoffroy et al. (2015) may also be an explanation for microcontinent formation in between conjugate VPM. This model could support and explain the presence of residual enigmatic but controversial continental fragments (e.g. Walvis Ridge, DSDP site 525A) in the South Atlantic Ocean (Class and le Roex, 2006; Maystrencko et al., 2003). Then, we precise also that microcontinents are not always or not only the result and/or consequence of magmatism wrenching mechanism can also favour or contribute further to the dislocation of the crust and microcontinent formation (Nemcok et al., 2016). All these alternative models may not be incompatible with the overlapping scenario proposed for the NEA.

The direct implication of this JMMC model in the context of the NEA also suggests that the GIFR or at least the Faroe-Iceland Ridge part should remain partly continental since the Reykjanes and Aegir Ridge never fully connected. Ultimately the GIFR most likely break apart not at C24 as often suggested in the literature (REF) or slightly later during Oceanic Phase I but much later and possibly when the JMMC finally dislocate from Greenland around anomalies chron C6B. This age (~24 Ma) most likely correspond with the ultimate and complete lithospheric breakup of the entire NEA.

\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*\*A bit of a reorganisation/simplification in the discussion\*\*\*\*\*\*\*\* to do later

# CONCLUSIONS

* + From a geodynamic point of view, the North Atlantic represents a compound rifted and magmatic system lying between continental and oceanic lithospheres. Before breakup, the North Atlantic developed a complex system of rifted basins.
  + Based on the most recent potential field, seismic data and new well calibration we re-evaluated the relative timing of the rift and magmatic events observed before, during and after the diachronic continental breakup between Eurasia and Greenland in Early Cainozoic time.
  + Inheritance and complex geological history likely suggest that rifting and breakup developed in and heterogenous lithospheric and continental mediums including different basement terranes and long-lived sub-continental heterogeneities preserved in the continental lithosphere subsequently ‘destroyed’ during different and propagating breakup events.
  + Breakup in the NEA is probably sharp, localised and magmato-tectonic processes (plumbing, diking) are most likely involved (cf. other volcanic margins worldwide).
  + We also compared and discussed the applicability (or not) of conceptual 'magma-poor' (Iberian type models) in highly magmatic setting. The fundamental discussion concerns the mode of crustal deformation, isostatic issue and the controversial nature and age of the high-LCB in the distal part of the NEA rifted margin. It also concerns the real amount of thinning formed in the inner and cooling part of the rifted system (e.g. the mid-Mesozoic thinning axis).
  + In the NEA, continental breakup possibly affected both thick and thinner crustal domains. However, thick continental crust at the edge of the volcanic margin and the no continuum of deformation of most of the NEA rifted margin do not clearly support the formation of a large zone of exhumed and denudated serpentinised mantle preceding the SDR and final breakup in the outer part of the margin. Surprisingly magmatism and SDR in the NEA develop preferentially where the continental crust is thicker on the flank of the the mid-Mesozoic thinning axis. Breakup is a lithospheric process and the long period of rifting and intermediate cooling event most likely influenced the final rifting and breakup often shifted compared to the dominant Mid-Mesozoic thinning axis of the NEA rifted margins.
  + The catastrophic amount of melt formation and the presence of thick SDR clearly highly a different kind of process and cannot be compared with the moderate amount of volcanism formed along magma-poor margins. This magmatism is maximum along the VPM but also affect unstretched continental crust (e.g. British Igneous Province).OK this point to emphasis also in previous chapter \*\*\*\*\*\*Laurent\*\*\*\*\*\*
  + Processs leading to volcanic margins and ultimate breakup remain unclear because all volcanic and rifted segments of the NEA realm are not all well constrained by well or relevant geophysical data. Sub-basalt imaging is also inherent problems that affect the interpretation to most of the distal margin in the NEA. Observations suggest, however, that SDR formations are preceded by renewed phase of extension in Late Cretaceous particularly well constrain in the Mid-Norwegian margin. Then, a first magmatic event and a localisation of the deformation preceed the syn-thinning and formation of the SDR before punctiform initiation of the breakup.
  + VPM and breakup itself are probably not so “instantaneous” nor continuous in the NEA. The early spreading system is diachronic, ‘non-unzipped’ and segmented development of the breakup along the NEA. It most likely involved propagating segments, and initial embryonic domain separated by continental/ultramafic buffers expected in different parts of the NEA (e.g. The Faroe Plateau, the Vøring Marginal High, the GIFR). Progressively the segmented spreading cells merged and reorganised together.
  + Spreading rift propagation is also expected in the NEA. In the different segments it may involve unexpected North to South propagation, sometimes in the opposite direction expected by traditional "unzipping" model and associated rotation pole. Influence of dynamic plume in the spreading propagation is also questionable (see also chapter from Dieter).
  + Both crustal and lithospheric inheritance may have an indirect implication on subsequent segmentation of the oceanic crust and microcontinent formation (Gernigon et al., 2012; Schiffer et al., 2017). The evolution of JMMC suggests rift/spreading ridge overlaps and support the presence of inherited continental and/or lithospheric “buffers” along the Faroe-Iceland Ridge before a second phase of breakup at ~24 Ma. This last event represents in this study the ultimate and complete lithospheric rupture of the NEA.

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