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Crustal fragmentation, magmatism, and the diachronous opening of the Norwegian-Greenland Sea

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ABSTRACT

The Norwegian-Greenland Sea (NGS) in the NE Atlantic comprises diverse tectonic regimes and structural features including sub-oceanic basins of different ages, microcontinents and conjugate volcanic passive margins, between the Greenland-Iceland-Faroe Ridge in the south and the Arctic Ocean in the north. We summarize the tectonic evolution of the area and highlight the complexity of the conjugate volcanic and rifted margins up to lithospheric rupture in the NGS. The highly magmatic breakup in the NGS was diachronous and initiated as isolated and segmented seafloor spreading centres. The early seafloor spreading system, initiating in the Early Eocene, gradually developed into atypical propagating systems with subsequent breakup(s) following a step-by-step thinning and rupture of the lithosphere. Newly-formed spreading axes propagated initially towards local Euler poles, died out, migrated or jumped laterally, changed their propagating orientation or eventually bifurcated. With the Paleocene onset of volcanic rifting, breakup-related intrusions may have localized deformation and guided the final axis of breakup along distal regions already affected by pre-magmatic Late Cretaceous-Paleocene and older extensional phases. The final line of lithospheric breakup may have been controlled by highly oblique extension, associated plate shearing and/or melt intrusions before and during Seaward Dipping Reflectors (SDRs) formation. The Inner SDRs and accompanying volcanics formed preferentially either on thick continental ribbons and/or moderately thinned continental crust. The segmented and diachronic evolution of the NGS spreading activity is also reflected by a time delay of 1–2 Myrs expected between the emplacement of the SDRs imaged at the Møre and Vøring margins. This complex evolution was followed by several prominent changes in spreading kinematics, the first occurring in the Middle Eocene at 47 Ma–magnetic chron C21r. Inheritance and magmatism likely influenced the complex rift reorganization resulting in the final dislocation of the Jan Mayen Microplate Complex from Greenland, in the Late Oligocene/Early Miocene.

1. INTRODUCTION

The transition from continental to oceanic rifting is a fundamental process in plate tectonics (Wilson, 1966). Important issues include the mechanisms and modes of lithospheric thinning that predate the breakthrough of the oceanic rift (e.g. Lavier and Manatschal, 2006; Van Avendonk et al. 2009; Reston, 2009; Huismans and Beaumont, 2011; Brune et al. 2014; Andrés-Martinez et al. 2019), and the impact of magmatic additions during rifting and breakup (e.g. Ebinger and Casey, 2001; Buck and Karner, 2004; Ligi et al. 2018). The role of inheritance (Vauchez et al. 1997; Bowling and Harry, 2001; Beniest et al. 2018; Schiffer et al. this volume), sub-lithospheric processes including deep asthenospheric upwelling during volcanic margin formation (Skogseid, 2000; Simon et al. 2009; Petersen and Schiffer, 2016) and microcontinent formation (e.g. Müller et al. 2001; Molnar et al. 2018; Schiffer et al. 2018) are currently under scrutiny.

The definition, structure, and timing of formation of the continent – ocean 'boundaries' are also being reassessed (e.g. Eagles et al. 2016). There remains ambiguity regarding whether the early stages of seafloor spreading are a consequence of: 1) instantaneous breakup along longer sections; 2) isolated albeit coeval spreading; and/or 3) propagating rifting and spreading (Bonatti, 1985; Corti et al. 2003; VanWijk and Blackman, 2005; Nirrengarten et al. 2018). Understanding which of these prevailed can only be achieved when the lithospheric configuration of both conjugate margins and adjacent oceanic crust is fully known in space and time.

In this review paper, we focus on the Norwegian-Greenland Sea (NGS) (Fig. 1), a pertinent natural laboratory for studying both continental rifting and oceanic spreading. As part of the NE Atlantic domain, the NGS is a high-quality case example that illustrates the complexity of rifting and the breakup processes. The opening of the NGS represents the final stage of the progressive Pangea dislocation between North America, Greenland and Eurasia (Gaina et al. 2009; Oakey and Chalmer, 2012; Nirrengarten et al. 2018). This process involved a long history of rifting in the NE Atlantic region and a resulting complex mosaic of inherited basement terranes and structures (Gee et al. 2008; Gasser et al. 2014; Bingen and Viola, 2018) (Fig. 2). Ultimately, a polyphase system of continental rifts (Doré et al. 1999; Mosar et al. 2002; Tsikalas et al. 2012) facilitated rupture of the lithosphere in the Early Cenozoic (Talwani and Eldholm, 1977; Hinz et al. 1987; Skogseid and Eldholm, 1989).

A major rift to drift characteristic of the NGS is its widespread conjugate volcanic passive margins (VPMs) which are part of one of the world's largest igneous provinces (the North Atlantic Igneous Province) (Fig. 2) (White and McKenzie, 1989; Saunders et al. 1997; Meyer et al. 2007a, 2007b; Hansen et al. 2009). The main characteristics of VPMs are: (1) thick volcanic wedges of seaward dipping reflectors sequences (SDRs) emplaced along the proto-breakup axes (Hinz, 1987; Mutter et al. 1982; Eldholm et al. 1989, 2000; Planke et al. 2000; Berndt et al. 2001); (2) the massive emplacement of sill/dyke intrusions in the sedimentary basins (Planke et al. 2005); and (3) the presence of high-V_P and high density lower crust (HVLC), commonly interpreted as breakup-related underplating or intrusions (Korenaga et al. 2001; White et al. 2008; Mjelde et al. 2009; Voss et al. 2009; Breivik et al. 2014). Despite wide consensus on these geophysical observations, an understanding of the mechanism of VPMs formation and the onset of magmatic breakup remain incomplete and debated (Geoffroy, 2005; Franke et al. 2013; Tugend et al. 2018; Guan et al. in press).

In this article, we summarize what is currently understood and what is debated in terms of the geological and tectonic evolution of the NGS. First, we describe the main characteristics of the rifted sedimentary basins surrounding the proto-oceanic domain of the NGS (Fig. 2). Second, we focus on syn-magmatic and tectonic models proposed for the formation of VPMs and ultimate rupture of the lithosphere (defined as breakup in this study). Finally, we discuss how the original continental amalgamation (Baltica-Greenland) ultimately broke apart and how and when the early spreading stage culminated in the formation of a new ocean. The complex configuration of the spreading axis between atypical continental fragments, such as the Jan Mayen Microplate Complex (JMMC, Figs. 1 and 2), and the role of anomalously thick crust such as the Greenland-Iceland-Faroe Ridge (Figs. 1 and 2) are also analysed. Why and how some spreading segments in the NGS suddenly or progressively became extinct and 'jumped' or relocated to new positions to form a microcontinent or microplate complex (e.g. the JMMC) is addressed in a geodynamic approach.

2. GEODYNAMIC SETTING OF THE NORWEGIAN-GREENLAND SEA

2.1 Physiography of the continental shelves and configuration of the spreading system

The first order morphology and geodynamic configuration of the NGS seafloor are well established from pioneering geophysical investigations (Talwani and Eldholm, 1977; Vogt et al. 1986) (Fig. 1). The asymmetry of the continental shelf, shallow plateaux, deep bathymetric troughs and marginal ridges is illustrated by recent bathymetric compilations (Etopo1; Amante et al. 2009; IBCAO 3.0, Jakobsson et al. 2012), Fig. 1; Free-Air gravity compilation DTU13 (Andersen, 2012), Fig. 3). Regional up-to-date aeromagnetic surveys and compilations of the NGS (Olesen et al. 2010; Gernigon et al. 2009, 2012, 2015) highlight the configuration of the Cenozoic oceanic seafloor particularly well (Fig. 4).

The bathymetry of the NGS (Fig. 1) shows considerable variations in width and steepness of the continental shelves between the mainland and the oceanic domains. To the East, the mid-Norwegian margin, including the Møre, Vøring and Lofoten-Vesterålen segments developed in prolongation of the North Sea and Faroe-Shetland margin (Figs 1 and 2). The mid-Norwegian margin is ~350–500 km wide and displays a complex morphology including isolated, but elongated marginal highs (e.g. the Møre and Vøring Marginal highs). To the north, the Lofoten-Vesterålen margin segment is narrower (200–250 km-wide) and extends up to the Southwestern Barents Sea margin. In contrast, along the conjugate NE Greenland margin, a narrow continental shelf is present close to Jameson Land and Traill Ø (~100-1500 km), while the shelf is much wider (300–400 km) close to the Greenland Fracture Zone farther north (Figs. 1 and 2). To the south, the aseismic and partly submerged Greenland-Iceland-Faroe Ridge (GIFR) represents a distinct NW-SE bathymetric high extending from the Faroe Plateau to East Greenland (Vogt et al. 1986; Bott et al. 1983) (Figs. 1 and 2). Iceland and its exposed rift zones represent the subaerial and central part of the GIFR and this unusual feature is often interpreted to be the consequence of the Icelandic "hot-spot" (White and McKenzie, 1989; Coffin and Eldholm 1994; Smallwood et al. 1999; Skogseid et al. 2000, Jones 2002; Karson, 2016).

North of the GIFR, the oceanic domain is characterized by well-defined magnetic chrons (Fig. 4) and was formed by seafloor spreading along the extinct Aegir Ridge and the active spreading ridges Kolbeinsey, Mohn's Ridge and Knipovich (Talwani and Eldholm, 1977; Jung and Vogt, 1997). Between the extinct Aegir Ridge and the seismically active Kolbeinsey Ridge (Fig. 1), a shallow marine plateau extends from Iceland to Jan Mayen Island. The submerged part of the plateau, extending south from Jan Mayen Island is called the Jan Mayen 'microcontinent' (Auzende, 1980; Gudlaugsson et al. 1988; Nunns, 1983; Blischke et al. 2016) or the Jan Mayen Microplate Complex (JMMC) due to local complexities and uncertainties regarding its crustal affinity (Gernigon et al. 2015; Schiffer et al. 2018; Polteau et al. 2018). West of the Jan Mayen Ridge, the Jan Mayen Basin at the edge of the shallow Icelandic Plateau formed mainly during the Cenozoic and before the onset of spreading along the Kolbeinsey Ridge (Vogt et al. 1980; Kodaira, 1998). At present, the Kolbeinsey Ridge opens orthogonally at moderate/slow full-spreading rates of 20-15 mm/yr but a number of seamounts indicate vigorous intraplate magmatism (Appelgate, 1997; Brandsdottir et al. 2015; Tan et al. 2017). Between the Vøring/Lofoten and Greenland basins (Fig. 2), the Mohn's Ridge spreads obliquely at low rates of typically 15–16 mm/year (Dauteuil and Brun, 1993). The northern continuation, the Knipovich Ridge, extends to the Fram Strait and spreads slowly (~14 mm/year) and obliquely in a direction of ~40º to 55º. In the hotspot reference frames (both shallow or deep), the North Atlantic Rift spreading system is also drifting westward implying a relative "eastward" mantle flow (Carminati et al. 2009).

The oceanic crust in the NGS is segmented by several fracture zones and transform faults (Figs. 2, 3 and 4). The northern boundary of the Norway Basin is the Jan Mayen Fracture Zone, the most prominent regional oceanic fracture zone of the NGS (Figs. 1, 2, 3 and 4). Its initial development is manifested by a ~160-km-wide, west-stepping offset in the mid-Norwegian VPM (Talwani and Eldholm, 1977). The fracture zone has distinct western and eastern branches which are spatially separated by about 50 km. The western end of the East Jan Mayen

Fracture Zone strikes at a considerable angle to the present-day fracture zones. To the north, the Greenland Fracture Zone, which separates the Greenland Basin from the Boreas Basin (Fig. 1) is associated with the East Greenland Ridge, which is interpreted as a continental sliver (Døssing et al. 2008). Fracture zones and associated transfer zones have been proposed to lie along the NE Greenland and conjugate Lofoten-Vesterålen margins (Blystad et al. 1995; Tsikalas et al. 2002; Mjelde et al. 2005) but these interpretations have been questioned based on recent aeromagnetic mapping (e.g. Olesen et al. 2007; Gernigon et al. 2009, Fig. 4).

2.2 Pre-rift setting: an inherited basement mosaic

The pre-rift mosaic of inherited crust and lithosphere of the NGS was characterized by a complex pattern of structural and compositional heterogeneities long deemed likely to have influenced directly or indirectly the style of rifting, magmatism and plate breakup (Ryan and Dewey, 1997; Vauchez et al. 1997; Doré et al. 1999; Krabbendam, 2001; Bowling and Harry, 2001; Beniest et al. 2018; Schiffer et al. this volume). The oldest basement terranes surrounding the NGS (Fig. 2) include Archaean rocks, juvenile Proterozoic crust (accreted arc material) and Middle Proterozoic calc-alkaline igneous bodies (Lehtinen et al. 2009). Archaean or Middle Proterozoic material was reworked during the Late Proterozoic Grenville-Sveconorwegian orogeny at ~1.14-0.9 Ga (Lorenz et al. 2012; Slagstad et al. 2017; Bingen and Viola, 2018). In Greenland, preserved Archaean rocks (3200–2600 Ma), reworked Archaean basement (around 1900–1800 Ma ago) and juvenile Paleoproterozoic rocks (2000–1750 Ma) are also found (Henriksen et al. 2009) (Fig. 2).

The younger Caledonian-Appalachian orogenic cycle reflects the closure of the Iapetus Ocean between Laurentia and Baltica (McKerrow et al. 2000). The Scandian phase (435–425 Ma) of the Caledonian orogenic cycle comprised several levels of thrust sheets (the Lower, Middle, Upper and Uppermost Allochthons) that were transported east onto the Fennoscandian platform (Roberts et al. 2003, Gee et al. 2008; Gasser et al. 2014). In Greenland, similar foreland-propagating thrust sheets originally derived from the Laurentian margin are observed but with an opposite polarity (Higgins and Leslie 2008; Henriksen et al. 2009).

3. PRE-TERTIARY TECTONIC SETTING: MAIN RIFT AND BASIN CHARACTERISTICS

3.1 Main rifting phases

Long before seafloor spreading started in the NGS, early post-orogenic basins developed as large, intra-continental, half-graben systems, controlled by reactivated low-angle detachments (Fossen, 2010). Devonian basins are mostly recognized onshore (e.g. the Hornelen Basin; Fig. 2), but their offshore regional distribution remains unclear. Refraction data suggest that 2–3 km thick succession of Devonian may potentially exist over the Trøndleag Platform and the Halten terrace (Breivik et al. 2011). The NE Greenland margin and the Barents Sea basins initially formed by orogenic collapse or extension around Late Devonian – Early Carboniferous time (Hartz and Andressen, 1995; Fossen et al. 2010; Gernigon et al. 2014, 2018: Klitzke et al. 2019). Late Carboniferous to mid-Permian faulting events also occurred onshore East Greenland (Peacock et al. 2000). In Svalbard and on Bjørnøya Island (Figs. 1 and 2), significant rift activity is documented in the mid-Carboniferous and the mid-to-late Permian (Gudlauggson et al. 1998; Larsen et al. 2002). This Palaeozoic period coincides with deposition of massive evaporite sequences both in the Barents Sea (Nilsen et al. 1995; Gernigon et al. 2018) and NE Greenland margin (Danmarkshavn Basin) (Hamann et al. 2005; Rowan and Lindsø, 2017). The

Danmarkshavn Basin (Fig. 2) probably represented the rift axis between Norway and Greenland in the late Palaeozoic rifting stage (Gudlaugsson et al. 1998).

Subsequent extensional episodes in the proto-NGS took place during the Permian and Triassic (Tsikalas et al. 2012) with the formation of narrow and wide rifts, and low-magnitude multiple extension depocentres mostly in the platform and shallow water areas of the future conjugate rifted margins (Štolfová and Shannon, 2009; Stoker et al. 2017). Permo-Triassic extension and deposition of sedimentary successions took place along the mid-Norwegian margin (Müller et al. 2005; Bergh et al. 2007; Faerseth et al. 2012) up to the Barents Sea (Smelror et al. 2009) and the conjugate NE Greenland margin (Surlyk, 1990; Seidler et al. 2004; Hamman et al. 2005; Tsikalas et al. 2012; Guarnieri et al. 2017; Rotevatn et al. 2018). The Devonian–Triassic development of the NE Greenland Shelf resembles that of the Southwest Barents Sea, and the Danmarkshavn Basin was a likely southern continuation of the Nordkapp Basin rift (Fig. 2) (Gudlaugsson et al. 1998; Hamann et al. 2005).

Late Triassic – Early Jurassic extensional phases are manifest as rifts and basins throughout the NE Atlantic (Doré, 1999; Stoker et al. 2017; Barnett-Moore et al. 2016) and North Sea (Errat et al. 1999). Mild extensional activity occurred episodically in the Early to Mid-Jurassic locally concomitant with major peripheral or intrabasinal uplift on the NE Greenland margin (Stemmerik et al. 1998), the mid-Norwegian margin (Brekke et al. 2000; Marsh et al. 2010), and in the SW Barents Sea (Faleide et al. 2008). In the central North Sea (Fig. 2), Middle Jurassic magmatism exploited pre-existing crustal structural anisotropies established during the Caledonian Orogeny (Quirie et al. 2018). Onshore East Greenland, a Mid-Jurassic hiatus represents a complete change in the basin configuration and drainage pattern, marking the onset of a new rifting phase (Surlyk and Ineson, 2003).

The Late Jurassic – Early Cretaceous interval (160-140 Ma) marks a profound kinematic and paleogeographic change throughout the entire NE Atlantic region (Lundin and Doré, 2011; Nirrengarten et al. 2018). Onshore paleostress analysis and dating of active brittle faults (Scheiber and Viola, 2018) confirm that extension during the Jurassic rifting phase was dominantly E-W from ~201-160 Ma but changed to WNW-ESE during the Early Cretaceous (at ~125 Ma). In the Faroe-Shetland region (Fig. 2), rift initiation in the late Berriasian-Barremian was focused in the West Shetland Basin but shifted to the Faroe-Shetland Basin in the Aptian-Albian, and further intensified in the Cenomanian-Turonian (Stoker 2016).

Offshore Norway, the central axis of the Late Jurassic-Early Cretaceous rifting episode appears to be shifted seaward relative to the Permo-Triassic rift basins (Fig. 2) (Lundin and Doré, 1997; van Wijk and Cloething, 2002) but the deep parts and the age of the distal basins remain largely unknown. A limited Mid-Late Jurassic-Berriasian extensional event (Færseth and Lien, 2002) or a semi-continuous Late Jurassic to mid-Cretaceous rifting phase have both been proposed (Pascoe et al. 1999; Doré et al. 1999; Gernigon et al. 2003; Tsikalas et al. 2012; Henstra et al. 2017). At the Møre margin severe thinning and syn-tectonic sagging climaxed in the earliest Cretaceous but probably slowed and/or failed around mid-Cretaceous time (mid-late Albian?) (Gernigon et al. 2015; Theissen-Krah et al. 2017). Indications of contemporaneous faulting and fault block rotation during the Early Cretaceous – mid-Albian were also found in the Traill Ø region, NE Greenland (Fig. 2) (Surlyk and Ineson, 2003; Price and Whitham, 1997) and a decline in faulting in the Albian is indicated by observations of Albian – Cenomanian strata onlapping and covering degraded footwalls (Parson et al. 2017). However, ongoing extension prevailed and possibly migrated to the northern Vøring Basin and Lofoten-Vesterålen segments as witnessed by faulting during the Late Albian-? to Turonian (Henstra et al. 2017; Zastrozhnov et al. 2018). In summary, there was likely at least a decline in extensional deformation in the late Early Cretaceous and, perhaps locally, rifting ceased.

In the Late Cretaceous-Paleocene, a renewed phase of widespread rifting affected predominantly the distal part of the NGS rift system (Figs 2 and 5). In the Møre, Vøring and Lofoten-Vesterålen margin segment, the Late Campanian-Paleocene rifting phase is relatively well constrained by boreholes and seismic data (Ren et al. 2003; Doré et al. 1999; Tsikalas et al. 2001; Gernigon et al. 2003, 2004; Bergh et al. 2007; Henstra et al. 2017; Zastrozhnov et al. 2018). Enlargement and increased subsidence of the Faroe-Shetland Basin occurred during the Coniacian–Maastrichtian interval punctuated by episodes of uplift and contractional deformation, particularly in the Campanian (Stoker 2016), a pattern of structural activity that persisted into the Paleocene (Stoker et al. 2018). Onshore East Greenland, Price and Whitham (1997) and Parson et al. (2017) proposed that faulting was renewed between the late Campanian and Thanetian. In NE Greenland, the Late Paleozoic N-S trending rift is crosscut by Triassic NE trending faults and subsequently by N-S trending faults (Escher and Pulvertaft, 1995). This rift system is also cut by NE–SW-oriented right-lateral faults of an oblique rifting stage interpreted to have occurred in the Paleocene (Guarnieri 2015) assuming that the tensors truly reflect consistent fault-slip data. In the Traill Ø region (Figs. 1 and 2) the structural evolution is defined by an eastward stepwise migration of the area of concentrated extensional deformation and an increase in the number of active faults and decrease in the spacing between them that occurs with each rift phase (Parsons et al. 2017). Subsequent to the Devonian-Triassic extension, main rifting took place during the mid- and late Jurassic, and during the Cretaceous new deep basins started to form on the more distant part of the shelf (Hamann et al. 2005) as observed in the mid-Norwegian margin.

3.2 Faroe-Shetland margin

At the southeastern edge of the NGS, the Faroe-Shetland region (Figs. 2, 5) comprises a series of Late Palaeozoic–Palaeogene-age rift basins that have undergone a multi-phase pre-breakup rifting history. The structural framework is dominated by the NE-trending Faroe-Shetland Basin, which comprises a complex amalgam of 11 sub-basins generally separated from one another by NE-trending crystalline basement-cored structural highs (Ritchie et al. 2011a, b; Trice et al. 2014). Along its southern and southeastern margins the Faroe-Shetland Basin is separated from a suite of smaller NE-trending basins, including the West Shetland, East Solan, South Solan, West Solan and North Rona basins, by the basement-cored NE-trending Rona High and the NW-trending Judd High (Ritchie et al. 2011a). The structural architecture of the rift basins has been strongly influenced – and most probably controlled – by structural trends (NE–SW and SE–NW) that are comparable to structural fabrics observed in onshore exposures of Palaeozoic and Proterozoic basement rocks in mainland Scotland and adjacent islands (Coward 1995; Doré et al. 1997, 1999; Wilson et al. 2010).

Extension and rifting took place episodically during the Late Palaeozoic, Mesozoic and Early Cenozoic. Devono-Carboniferous basins are a relic of post-Caledonian orogenic collapse (Stoker et al. 1993; Smith and Ziska 2011), whereas Permo-Triassic, (mainly Late) Jurassic and Cretaceous basin development (Stoker et al. 1993; Quinn and Ziska 2011; Ritchie and Varming 2011; Stoker 2016; Arsenikos et al. 2018) is related to the fragmentation of Pangaea. The major rifting phase in the Faroe–Shetland region occurred in the 'Mid'-to Late Cretaceous (Figs. 2 and 5). This resulted in increased connectivity between, and a general subsidence of the sub-basins from which the Faroe-Shetland Basin acquired its larger regional expression (Larsen et al. 2010; Stoker 2016). The Cretaceous sedimentary succession preserves a rock record that is punctuated by episodes of uplift, erosion and contractional deformation, revealing a pattern of coeval extension and compression (Stoker 2016). This regime persisted into the Paleocene whereby sedimentation was controlled by a series of sag- and fault-controlled sub-basins (Dean et al. 1999; Lamers and Carmichael 1999). This has prompted speculation that the tectonostratigraphic signature represents a response to strike-slip tectonics along a developing shear margin between the Faroe-Shetland region and Central East Greenland (Geoffroy et al. 1994; Guarnieri 2015; Stoker et al. 2018).

Deep seismic surveys (Fig. 6) (Raum et al. 2005; White et al. 2008; Ólavsdóttir et al. 2017; Funck et al. 2017) and potential field studies (Kimbell et al. 2005; Rippington et al. 2015; Haase et al. 2017) indicate progressive thinning of the crust from the West Shetland Platform to the central Faroe-Shetland Basin where the continental crust is still 10–15 km thick (e.g. Smallwwod et al. 2001; Raum et al. 2005). The Faroe-Shetland Basin is separated from the oceanic basins by a thick marginal continental block underneath the Faroe Plateau and the Fugløy Ridge where the continental crust is up to 30-25 km thick beneath more than 5–7 km of basaltic layers and sediments (White et al. 2008; Fletcher et al. 2013). In the lower crust close to the continent-ocean transition, HVLC (Vp>7.2 km/s) and strong sub-horizontal reflections interpreted as sills intrude the dipping fabric of inferred Precambrian continental crust at the edge of the GIFR (Fig. 6) (Bott et al. 1983; White et al. 2008; Ólavsdóttir et al. 2017). In the offshore part of the Faroe Plateau, sub-basalt sedimentary thicknesses ranging from 1–8 km have been inferred based on seismic interpretation (Richardson et al. 1998; 1999; Raum et al. 2005; White et al. 2008). Recent ambient noise tomography suggests that metamorphic rocks (Vp ~ 5.75 km/s) with deeper (Archaean?) terranes (6.2<Vp<6.3 km/s) might lie beneath 1–2 km of undifferentiated pre-volcanic sediments and/or hyaloclastites (3.2<Vp<4.7 km/s) on the Faroe Plateau (Sammarco et al. 2017).

3.3 Mid-Norwegian margin

Five decades of petroleum exploration and drilling of more than 200 wells have revealed the overall first order structure, stratigraphy and volcanic history of the mid-Norwegian margin and rendered it the best constrained and understood part of the entire NGS (Figs. 2, 7).

3.3.1 Møre and Vøring margin segments

The Møre and Vøring rifted margins segments in the prolongation of the North Sea and Faroe-Shetland margin preserve a series of N and 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/Dønna Terrace (3) the Møre and Vøring basins; and (4) the Møre and Vøring marginal highs to the west (Figs. 5 and 7). To the south, the Vøring Basin is connected to the Møre Basin through a broad regional transfer zone (e.g. Mosar et al. 2002), the so-called the Jan Mayen Corridor (Gernigon et al. 2015; Theissen-Krach et al. 2017).

The narrow, shallow platform of the Møre margin (Figs. 5 and 8) comprises a series of NE-SW-elongated basement highs, separating sub-basins from the deep Møre Basin (Jongepier et al. 1996; Grunnaleide and Gabrielsen, 1995; Theissen-Krah et al. 2017). These structures developed along the NE-SW trend of the Møre-Trøndelag Fault Complex (Figs. 2 and 7) and exhibit inherited structures that can be traced from the Faroe–Shetland region (Doré et al. 1997; Nasuti et al. 2011). This inherited regional trend roughly delineates a necking zone and an oblique transition zone between the northern North Sea and the Møre margin (Figs. 2 and 7) (e.g. Olaffson et al. 1992; Brekke, 2000).

The Trøndelag Platform of the Vøring margin is broader than the platform domain of the Møre margin (Figs. 2, 5 and 7). Across most of the Trøndelag Platform, the Upper Palaeozoic – Lower Triassic basinal succession is interpreted to be locally thick (6–7 km), whereas the Middle Triassic to Jurassic sequences show relatively uniform thickness (5 km on average) with a gradual thinning towards the SE border. In the shallow platform, Færseth et al. (2012) described the Late Permian-Early Triassic rift system from the Vestfjorden Basin to the Froan Basin (Figs. 2, 5 and 7) as a series of 'en echelon' half-grabens mostly controlled by major east-dipping border faults (Fig. 2). The adjacent terraces have a large variety of structural styles, including extensional folds, fault propagation folds, basement-involved- and basement-detached normal faults and narrow grabens. These are linked to halokinesis which was active during Jurassic-Cretaceous rifting events (e.g. Withjack and Callaway, 2000; Richardson et al. 2005).

At the edge of their respective platform domains, the 125–150 km wide central Vøring and Møre basins and the intermediate Jan Mayen Corridor are primarily characterized by huge thicknesses of Cretaceous-Cenozoic rocks (Brekke et al. 2000; Lien et al. 2005; Theissen-Krah et al. 2017)(Figs. 5, 7 and 8). The base Cretaceous unconformity locally reaches a depth of 12–13 km and thus the deepest sediments are most likely metamorphosed (Maystrenko et al. 2017c; Zastrozhnov et al. 2018). The Møre and Vøring basins include several subsidiary sag basins, separated locally by high-density basement highs (Zastrozhnov et al. 2018). In the Jan Mayen Corridor, deep 'en echelon' crustal rafts (e.g. Slettringen, Grip and Vigra highs) formed along the broad transfer zone initiated in the early stage(s) of rifting prior to the Cretaceous (Figs. 7 and 8) (Gernigon et al., 2015; Theissen-Krah et al. 2017).

The outer, distal, western parts of the Møre and Vøring basins are key tectonic hinge zones linking the flexural edge of the Cretaceous sag basins and the VPM (Figs. 5, 9). Close to the basalt a separate phase of active stretching and fault block rotation (e.g. the pre-magmatic rift climax) was initiated in the early Campanian (Gernigon et al. 2004; Fjellanger et al. 2005; Zastrozhnov et al. 2018). The remaining Paleocene 'sag' sequences are weakly faulted but contemporaneous with drastic thinning of the lithosphere and localization of the deformation towards the Paleocene-Eocene volcanic breakup axes (Ren et al. 2003; Gernigon et al. 2003, 2004) (Fig. 9). Meanwhile, extreme thinning was focused on the Hel Graben (Figs. 5 and 7) and characterized by thick Paleocene sediment accumulations (Lundin et al. 2013; Zastrozhnov et al. 2018). A sudden relocation of the rift axes to the west occurred during the magmatic rifting phase preceding the final breakup (Zastrozhnov et al. 2018). The formation of related subsidiary sag basins, such as the Vigrid and Någrind 'synclines' and the Vema – Nyk 'anticline' (Figs. 5 and 7), has been attributed to either compression/transpression (Brekke, 2000; Lundin et al. 2013) or buckling and boudinage of the crust during the Cretaceous-Paleocene extension (Zastrozhnov et al. 2018). Compared to the outer Vøring Basin, only minor Late Cretaceous-Paleocene faults are observed in the outer Møre Basin (Fig. 5). However, recent sub-basalt imaging improvement suggests that Late Cretaceous-Paleocene grabens might be present underneath the wide (50–200 km) basaltic cover to the Møre Marginal High (Manton et al. 2018).

The Moho depth beneath the Møre platform and coastal areas (Fig. 6) varies between 29 km under the shelf to 37 km onshore, as indicated by seismic refraction studies (Maupin et al. 2013; Kvarven et al. 2014). HVLC is widespread beneath the Møre platform crust (Kvarven et al. 2014, 2016). On the Trøndelag Platform and adjacent Froan Basin (Fig. 2), the Permian-

Jurassic succession caps a 30-25 km thick crystalline crust, thinning to the west to less than 20-15 km (Breivik et al. 2011). The Halten and Donna terraces (Figs. 5, 6, 7) developed along a sharp necking zone between the Trøndelag Platform and the deep Cretaceous sag-basins where the continental crust thins to less than 10–15 km (Breivik et al. 2011; Maystrenko et al. 2017c).

In the Møre Basin, Vøring Basin and intermediate Jan Mayen Corridor, the total sedimentary section locally reaches a thickness of 10–14 km (Raum et al. 2002; Mjelde et al. 2009; Maystrenko et al. 2017c). Below are rift-related, rotated crustal blocks, representing moderate-to-very thin continental crust (Olafsson et al. 1992; Mjelde et al. 2002; 2005; Reynisson et al. 2010; Kvarven et al. 2014, 2016; Nirrengarten et al. 2014) (Fig. 6). In the Møre Basin and the Jan Mayen Corridor, continental crust locally is thinned to less than 10 km and underlain in its outer parts by ~4 km-thick HVLC with V_P ~7.2-7.4 km/s (Olafsson et al. 1992; Raum et al. 2002; Mjelde et al. 2009). Kvarven et al. (2014, 2016) confirm the presence of >5–10 km thick continental crust (V_P ~ 6.1 - 6.7 km/s) beneath the central part of the main Cretaceous depocentre of the Møre Marginal High (Fig. 6). In the eastern part of the sag-basin, the crust is thinnest (~5 km) but the OBS data available do not always detect HVLC west of the necking zone (Kvarven et al. 2014).

Similarly, the Vøring Basin is underlain by continental crust (V_P ~ 6.0-6.9 km/s) thinned to 2–11 km (5–10 km on average). HVLC in the Outer Vøring Basin has velocities of V_P >7.2-7.4 km/s (Mjelde et al. 2003; 2005) but is limited (or absent) in the eastern part of the Vøring Basin (Fig. 6)(Mjelde et al. 2008; Breivik et al. 2010). The continental-ocean transition over the Vøring Marginal High was modelled by Mjelde et al. (2005) and showed a thick crust of 20–25 km underlain by a distal HVLC. Intermediate velocities of ~6.5 km/s were interpreted as heavily intruded continental crust (Mjelde et al. 2005). A layer with V_P ~6.0 km/s in the topmost crust is conformable with a continental (granitic) basement, whereas corresponding velocities oceanwards (6.9 km/s) were interpreted as gabbroic, oceanic crust (Mjelde et al. 2005).

3.3.2 The Lofoten-Vesteralen margin segment

The narrow Lofoten-Vesterålen margin segment (Figs 2, 5 and 7) is the result of several phases of rifting, uplift and erosion (Tsikalas et al. 2001; Bergh et al. 2007; Olesen et al. 2004; Tasrianto and Escalona, 2015; Maystrenko et al. 2017b). The basement rocks exposed on the central Lofoten Archipelago are mostly remnants of Caledonian allochthons overthrust onto Precambrian rocks of Baltica (Bergh et al. 2007; Steltenpohl et al. 2011). A striking feature of the continental shelf along the Lofoten–Vesterålen margin segment is the relatively thin sequence of Jurassic-Triassic sediments both on structural highs and in several of the sub-basins (Fig. 5) (Færseth, 2012). The occurrence of older sedimentary rocks is uncertain, though Palaeozoic strata might be present in the deepest parts of the area (e.g. Bergh et al. 2007). The narrow, NE-SW-trending Ribban and Vestfjorden basins (Figs. 2, 5, 7) contain thick syn-rift Lower Cretaceous sequences (Tsikalas et al. 2001; Hansen et al. 2011). In the Lofoten region, Henstra et al. (2017) present seismic evidence to distinguish Late Albian-Cenomanian from Campanian-Paleocene rift activity

Refraction seismic data (Mjelde et al. 1996; Breivik et al. 2017) and potential field models (Olesen et al. 1997; Tsikalas et al. 2005; Maystrenko et al. 2017b) show that the Moho topography along the Lofoten-Vesterålen margin varies from shallower than 11 km in the outer part of the margin domain to more than 42–48 km depth beneath the mainland (Ben Mansour et al. 2018) (Fig. 6). The depth to basement varies from < 1 km to >10–12 km in the Vestfjorden

Basin (Breivik et al. 2017; Maystrenko et al. 2017b). Pre-Cretaceous and Cretaceous structural levels in the Røst Basin to the west are generally obscured on seismic data by Early Cenozoic and breakup-related intrusives and extrusives (Berndt et al. 2001) though up to 5 km thick. Cretaceous-Paleocene sequences might be present beneath the basalts according to 3D potential field modelling (Maystrenko et al. 2017b). West of the Utrøst Ridge, a 2 km-thick HVLC is present in the necking zone adjacent to the continent-ocean transition (Fig. 6)(Breivik et al. 2017). To the west, isolated rotated blocks surrounded by minor SDRs are detected by reflection seismic lines (Tsikalas et al. 2002).

3.4 NE Greenland margin

For decades, studies of the geology and structures of the Central and NE Greenland margin concentrated along the onshore sedimentary basins, exposed from the Jameson Land to the Wandel Sea Basin due to significant Paleogene and Miocene uplifts (Dam et al. 1998). These sedimentary basins (Figs. 2, 5) contain rotated fault blocks, defined by mainly eastward-dipping faults, which formed in response to the latest Devonian-Early Carboniferous collapse of the Caledonides (Surlyk, 1990; Price and Whitham 1997; Parson et al. 2017; Rotevatn et al. 2018).

Current knowledge of the offshore structural evolution of the NE Greenland continental margin is limited, though basin evolution is thought to span the entire period between the Devonian and Neogene (Hamann et al. 2005; Tsikalas et al. 2005; Dinkelman and Keane 2010; Geissler et al. 2016; Berger and Jokat, 2009). Two major sedimentary basins, the Danmarkshavn and the Thetis basins, separated by a prominent basement high known as the Danmarkshavn Ridge, have been identified (Figs. 2, 5 and 11). Further north, the Westwind Basin (Figs. 2 and 11) formed by transtension along the margin as Greenland moved obliquely relative to the Barents Sea in the Late Cretaceous-Eocene (Granath et al. 2011). The deep Danmarkshavn Basin is bounded by the Koldewey Platform to the Caledonian basement, which crops out onshore NE Greenland (Fig. 2) (Hamann et al. 2005). Major basin-forming faults lie along the western flank of the Danmarkshavn Basin. The transition between the Danmarkshavn Basin and the Danmarkshavn Ridge comprises a series of west-dipping rotated fault blocks, overlain by prograding Paleogene sediments along the western margin of the ridge (Tsikalas et al. 2005; Dinkelman and Keane 2010). In the landward Danmarkshavn Basin, the maximum thickness of the basin fill is ca. 15–17 km (Fig. 11) (Dinkelman and Keane 2010; Granath et al. 2011; Funck et al. 2017). A major regional unconformity separates the Devonian – Cretaceous section from the overlying Paleocene and younger units (Hamman et al. 2005; Tsikalas et al. 2005; Petersen et al. 2015). By analogy with the Barents Sea region (Fig. 2), salt diapiric structures identified in seismic sections in the northern Danmarkshavn Basin have been linked to halokinetic movements of Carboniferous-Lower Permian salt (Hamann et al. 2005; Rowan and Lindsø, 2016). In the seaward Thetis Basin Jurassic – Cretaceous sequences were identified by Hamman et al. (2005) though the presence of older rocks cannot be discounted (Tsikalas et al. 2005). The Thetis Basin (Fig. 11) also appears to contain an older platformal section possibly equivalent to the western Danmarkshavn Basin (Granath et al. 2011) but due to the absence of borehole calibration this remains speculative in this frontier area. The eastern shelf also includes a distal structural high between the Jan Mayen and Greenland Fracture Zones (Hamann et al. 2005; Berger and Jokat, 2008). This structural high is made up of a series of northeast-southwest trending anticlines and domes that are interpreted to have also been affected by later Cenozoic uplift and inversion (Granath et al. 2011). East of the Greenland Escarpment (Fig. 2), the Paleocene basalts onlap a large part of the 'outer structural high', (Figs. 11 and 12), which was already a topographic elevation before the emplacement of the basalt and Inner SDRs (Granath et al. 2011).

The geology of the southern NE Greenland margin is also poorly understood due to the lack of data and the sub-basaltic imaging issue. However, a thick sedimentary succession has been recognized in seismic data offshore Liverpool Land and adjacent to Devonian-Jurassic rocks exposed in Jameson Land (Figs. 2 and 5)(Hamman et al. 2005; Guarnieri 2015; Eide et al. 2017). In the area underlain by continental crust, a thick Cenozoic succession lies unconformably on undifferentiated Late Paleozoic-Mesozoic faulted blocks (Hamman et al. 2005). Offshore Greenland, it is thought that pre-Paleocene sediments to the north and south continue beneath the volcanic rocks and SDRs that may have been deposited along the continent-ocean transition (Fig. 2) (Hinz et al. 1987; Tsikalas et al. 2005; Quirk et al. 2014; Geissler et al. 2016).

Integrated geophysical and geological studies have revealed pronounced along-strike differences in the crustal architecture of the NE Greenland margin (e.g., Schlindwein and Jokat, 1999; Schmidt-Aursch and Jokat, 2005; Voss and Jokat, 2007; Voss et al. 2009; Schiffer et al. 2016; Funck et al. 2017). The continental crust of the NE Greenland margin was affected by a combination of Caledonian, Devonian, and Mesozoic-to-Cenozoic phases of deformation. North of Jameson Land, Mesozoic extension shifted eastward of the Devonian structures (Voss et al. 2009). With respect to the conjugate margin (Fig. 6), a wide (120–130 km) but unclear continent-ocean transition is proposed in the southern and central parts of the NE Greenland margin (Voss et al. 2009). Schlindwein and Jokat (1999) modelled a ~80-km-wide, <8-km-thick HVLC in the prolongation of Kejser Franz Josef Fjord (Fig. 6). Beneath the continent-ocean transition, Voss and Jokat (2007), however, detected an HVLC body with a width of ~225 km and thickness of up to 16 km, implying major crustal asymmetry when compared to the conjugate Vøring margin (Fig. 6). The crust above the HVLC (Vp 7.15–7.4 km/s) has elevated seismic velocities (Vp 6.6–7.0 km/s). However, with receiver function modelling Schiffer et al. (2015a) detected two separate HVLC bodies in the same area, each 50–80 km wide and ~10–15 km thick. These authors interpret them as remnants of a fossil Caledonian subduction complex, including a slab of eclogitised mafic crust and an overlying wedge of serpentinised mantle peridotite. The HVLC is found up to a distance of 200 km from the earliest oceanic crust (Schiffer et al. 2015a, b). The thickness of the inherited and/or breakup-related HVLC is particularly important (up to 15-16 km) north of the West Jan Mayen Fracture zone and its landward prolongation but limited further south (Schmidt-Aursch and Jokat, 2005; Voss and Jokat, 2007, Hermann and Jokat, 2016; Abdlemalak et al., 2017). The Moho depth in the central and northern part of the NE Greenland margin (Fig. 6) varies from 30 km to 11–14 km near the earliest oceanic crust (Voss and Jokat, 2009). Farther north, long-offset seismic reflection lines also suggest the presence of two distinct and highly reflective HVLC beneath the Danmarkshavn Ridge and the Thetis Basin's "outer structural high". Close to the volcanic province, the continental crust is still relatively thick on top of the distal HVLC (10–15 km) (Dinkelman and Keane 2010; Granath et al. 2011). Seismic reflection lines (Fig. 11) show that the transition of SDRs to oceanic crust at the edge of the Thetis Basin and its "outer high" is relatively sharp (<50–75 km).

3.5 The intermediate Jan Mayen Microplate Complex (JMMC)

The JMMC (Figs. 1, 2 and 5) is traditionally considered to include a 'missing piece' of continental crust that was originally part of the pre-breakup rifted system located between the Faroe Plateau, the outer Vøring Basin and Greenland (Guðlaugsson et al. 1988; Kodaira et al. 1998; Blischke et al. 2017). The JMMC finally separated from Greenland in the earliest Miocene around magnetic anomaly chron C6b, (~22.5 Ma) (Nunns, 1983; Gernigon et al. 2012). Only the shallow part of the Jan Mayen Ridge was drilled during DSDP Leg 38 (sites

347, 349)(Fig. 1). The borehole encountered hemipelagic and pelagic sediments of Mid-Eocene and younger age, but failed to reach the basaltic basement and local SDRs observed on seismic data (Guðlaugsson et al. 1988). Recent seafloor sampling has also recovered Cenozoic and Mesozoic sediments, it remains uncertain whether or not the Mesozoic samples were in situ (Polteau et al. 2018).

Modelling of data from ocean bottom seismometers suggests that Jan Mayen Island and the northernmost part of the Jan Mayen Ridge are underlain by Icelandic-type crust, bordered to the north by anomalously thick oceanic crust (Fig. 6) (Kandilarov et al. 2015). The main part of the Jan Mayen Ridge, however, is likely underlain by continental crust, 15–16 km thick (Kodaira et al. 1998; Breivik et al. 2012). Crustal-scale velocity profiles are comparable with those of the mid-Norwegian margin. Mjelde et al. (2007) suggest the presence of Mesozoic sediments ($4.0<Vp<4.7$ km/s) and deeper Palaeozoic sediments ($5.0<Vp<5.3$ km/s) underlying the Cenozoic sediments (2.2-3.4 km/s). According to the seismic refraction data, the thicknesses of the upper crystalline crust ($6.0<Vp<6.4$ km/s) and lower crust (with $6.7<Vp<6.8$ km/s) are approximately 3 km and 12 km, respectively (Fig. 6). The central segment of the Jan Mayen Ridge is distinguished by the overall absence of HVLC (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 at the central eastern passive margin of the JMMC (Breivik et al. 2012). Brandsdóttir et al. (2015) have also reported a HVLC body at the transition from Iceland to the JMMC (Fig. 6).

In the Jan Mayen Basin, refraction data indicate the presence of highly attenuated continental crust, underlying a < 5 km deep sedimentary basin and thin basaltic layer (Kodaira et al. 1998). In contrast to the eastern margin of the JMMC where SDRs are interpreted (e.g. Planke and Alvestad, 1999), no substantial magmatic activity accompanied the formation of the western margin of the JMMC (Kodaira et al. 1998). However, widespread shallow-marine volcanic flows and intrusive sills were possibly emplaced during the Oligocene (Gudlaugsson et al. 1988; Blischke et al. 2016). The gradual thinning of the continental crust to 3 km together with the 5–10 km thickness of the adjacent oceanic crust resembles the structural architecture of non-volcanic (magma-poor?) rifted margins with no SDRs (Kodaira et al. 1998) (Fig. 6).

4. CENOZOIC TECTONO-MAGMATIC EVENTS IN THE NORWEGIAN-GREENLAND SEA

4.1 The North Atlantic Igneous Province

The North Atlantic Igneous Province formed prior to, during, and after the initiation of the breakup of the NGS (Figs. 2 and 13) (Coffin and Eldholm, 1994; Saunders et al. 1997; Meyer, 2007b; Horni et al. 2017). Exposed and submerged basaltic rocks of this large igneous province extend roughly NE – SW for more than 2000 km from Greenland to the NW European margins (Fig. 2), and cover a surface area of at least $\sim 1.3\times 10^6$ km^2. The combined volumes of extrusive and intrusive rocks could be as large as $\sim 6.6\times 10^6$ km^3 (Eldholm and Grue, 1994), depending on the interpretation of the HVLC and Icelandic-type crust. Two main periods of magmatism are often proposed within the North Atlantic Igneous Province–the first at 62-57 Ma and a second major pulse of magmatism at c. 56-54 Ma (Saunders et al. 1997; Storey et al. 2007). In a recent review, Wilkinson et al. (2017) concluded, however, that significant pauses in regional volcanism did not really take place although breaks in volcanism may have occurred locally. Radiometric and magnetostratigraphy data suggest instead a gradual and continuous process with no real or clear distinction between postulated North Atlantic Igneous Province periods 1 and 2.

4.1.1 Volcanism in the Faroe-Shetland margin

In the Faroe Islands (Figs 1 and 2), more than 3 km of subaerial flood basalt flows and minor volcaniclastic lithologies are exposed above sea level (Passey and Jolley, 2009; Millett et al. 2017). These rocks form part of the Faroe Islands Basalt Group which has a gross stratigraphic thickness of ~6.6 km but only ~3.4 km has been penetrated by the Lopra-1/1A borehole (Passey and Jolley 2009; Chalmers and Waagstein, 2006). The lava flows and Cenozoic foresets of hyaloclastite and local siliclastic sediments overlap the surrounding sedimentary basins (Ellis et al. 2002; Hardmann et al. 2017). In the Faroe Islands, all sampled lavas are dominantly tholeiitic and indicative of voluminous mantle melting (Millett et al. 2017). Isotopic evidence for continental contamination of some lava flows by Precambrian amphibolite-facies rocks has been reported (Holm et al. 2001). There is considerable debate on the chronology of the Faroe Islands Basalt Group. On the basis of palynology and biostratigraphic data alone several workers (e.g. Jolley and Bell 2002; Passey and Jolley 2008; Jolley 2009) have proposed that the extrusion of the Faroe Islands Basalt Group occurred exclusively during chron C24r (latest Thanetian–earliest Ypresian time, c. 57–54.5 Ma), which Schofield and Jolley (2013) more recently assigned to the time interval 56.1–54.9 Ma, coeval with the Paleocene–Eocene Thermal Maximum (PETM)(Fig. 13). By way of contrast, a longer stratigraphic range for the Faroe Islands Basalt Group spanning Selandian–earliest Ypresian time (chron C26r–C24r, c. 61–54.9 Ma) is implied by a combined biostratigraphic, palaeomagnetic and radiometric (K-Ar, Ar-Ar) dataset (e.g. Riisager et al. 2002; Waagstein et al. 2002; Abrahamsen 2006; Storey et al. 2007; Mudge 2015; Wilkinson et al. 2017).

4.1.2 Volcanism in the mid-Norwegian margin

On the mid-Norwegian margin, pre-, syn- and post-breakup-related volcanism is documented by numerous reflection and refraction seismic studies (Berndt et al. 2001; Breivik et al. 2006; Mjelde et al. 2009; Abdelamalak et al. 2016a, b, c; Planke et al. 2017). Subaerial basaltic rocks have been sampled by Deep-Sea Drilling Project DSDP leg 38 (sites 338, 342 and 343) and Ocean Drilling Program ODP leg 104 (sites 642, 643 and 644) (Eldholm et al. 1989). ODP Leg 104 Hole 642E was drilled at the edge of the SDRs (Figs. 7 and 9a) and encountered two main subaerial volcanic sequences, the Lower and Upper series (Meyer et al. 2007a; Abdelmalak et al. 2016a). This hole records the onset of volcanism at 56.5 – 55 Ma (equivalent of magnetic chrons C25n-C24r; Eldholm et al. 1989; Sinton et al. 1998) (Fig. 13). Micropaleontology analysis of the Upper Series supports an age of ~55-54 Ma (Abdelmalak et al. 2016a). In 2014, the Norwegian Petroleum Directorate conducted shallow drilling operations on the northernmost part of the Møre Marginal High and the resulting drill hole 6403/1-U-1 recovered 38 m of igneous rocks (lava flows, lava breccias and hyaoclatites) of exclusively tholeiitic composition (Bakke et al. 2017). No age data are yet published but palynological analysis of one sediment sample from the base of this sequence suggests an Early Eocene (Early Ypresian) age (Bakke, 2017).

4.1.2 Volcanism in East Greenland

In East Greenland, lavas were erupted in two major episodes: a Paleocene (62–57 Ma) pre-breakup episode and an Eocene (56–55 Ma) episode (Saunders et al. 1997; Hansen et al. 2009; Storey et al. 2007; Brooks 2011). Minor activity also occurred during the periods 53–51 Ma and 45–40 Ma (Tegner et al. 2008; Larsen et al. 2014). Intrusions emplaced after the C24r phase of breakup are mostly early Eocene (50–47 Ma) but also yield younger late Eocene ages of 35–36 Ma (Tegner et al. 2008). The pre-breakup lavas (e.g. Lower Basalts) exposed in Central East

Greenland are Paelocene and dated at 61.9–58.1 Ma (C26r-C25r) (Storey et al. 2007; Larsen et al. 2014). The most voluminous burst of magmatic activity is associated with the syn-breakup volcanic succession, dated at c. 56–55 Ma, which coincides with the magnetic chron C24r (Fig. 13). This succession comprises the Plateau Basalts of the Blosseville Kyst (Fig. 2) where the exposed flood basalts have a total thickness of ~7 km (Larsen et al. 1999; Brook, 2011). Latest Paleocene to earliest Eocene Plateau basalts are found on eastern Geographical Society Ø (Fig. 2), where they are up to 150 m thick (after erosion) and consist of tholeiitic lava flows with rare volcaniclastic layers (Jolley and Whitham, 2004; Larsen et al. 2014). In contrast to the mid-Norwegian margin, onshore NE Greenland was affected by a subsequent alkaline magmatic episode from 37–35 Ma, possibly in association with the late dislocation of the JMMC in Late Oligocene-Early Miocene time (Tegner et al. 1998).

4.2 'Breakup' related crustal structures and magmatism

4.2.1 Volcanostratigraphy

Seismic wedges comprising SDRs are the most recognizable magmatic structures of the NGS (Figs. 2, 5, 9, 11). SDRs mostly consist of basaltic rocks and are of much greater thickness than the average oceanic layer 2A (Mutter et al. 1982; Hinz, 1987). They are commonly part of more complex volcanic systems that reflect different paleogeographic settings on the continent-ocean transitions of the NGS. On the basis of geophysical data a standardized set of descriptive terms and nomenclature has been established to classify the seismic 'volcanostratigraphy' (e.g. Inner Flows, Landwards Flows; Inner SDRs, Outer High and Outer SDRs) in the NGS (Planke et al. 2000; Berndt et al. 2001; Abdelmalak et al. 2016b; Planke et al. 2017) (Figs. 7, 9 and 10). Planke et al. (2000) divide the SDRs into Inner and Outer packages. In this model, the Inner SDRs represent subaerally emplaced lava flows, the geometry of which being affected by the pre-existing basin architecture. The Outer SDRs are believed to represent deep marine flows emplaced after the Inner SDRs construction (Berndt et al. 2001; Planke et al. 2000; Horni et al. 2017). A mounted Outer High sometimes develops in between (Fig. 10). The volcanic Outer High may represent voluminous and shallow marine eruptive volcanoclastics and tuffs (Planke et al. 2000).

4.2.2 SDRs formation and geometry: magmatic and tectonic issues

The conjugate SDRs observed between Norway and Greenland (Figs. 2, 11) have been interpreted as emplaced either over (highly) distended continental crust (Hinz, 1981, 1987) or oceanic crust (Mutter et al. 1982). Scientific drillings provided evidence that, where sampled in the NE Atlantic, Inner SDRs magmas incorporate material from the continental lithosphere (Eldholm et al. 1989; Larsen et al. 1994; Meyer et al. 2007b). The Inner SDRs exposed onshore Southeast Greenland are also directly underlain by thick upper crustal Precambrian gneiss injected by massive mafic dykes swarms and gabbroic/alkaline plutons (Karson and Brooks 1999; Callot et al. 2001). As the Inner SDRs were earlier envisaged (Mutter et al. 1982), the crust beneath Outer SDRs may be dominantly mafic. However, in general, its origin (continental or oceanic) is uncertain (e.g. White and Smith, 2009; Geoffroy et al. 2015; Clerc et al. 2015; Patton et al. 2017; Guan et al. in press). Subsidence, flexure of individual SDRs wedges, and associated rotation of underlying dyke swarms, was likely a short-duration event (< 2.9 Ma, Lenoir et al. 2003; Geoffroy, 2005). Dense dyke swarms and the geometry of lava flows on Iceland and other parts of the GIFR (Fig. 2) also show structural similarities with SDRs (Pálmason, 1973; Walker, 1993; Bourgeois et al. 2005).

The characteristic wedge shape of the SDRs (Figs. 9, 10 and 11) might result from crustal extension and/or loading by the thick basalt pile (Geoffroy et al. 2005; Quirk et al. 2014; Buck 2017). Refraction seismics, together with recent deep seismic-reflection profiles, show that SDRs emplacement was concomitant with sharp and likely syn-magmatic crustal necking (Figs. 6 and 11) (Mjelde et al. 2005; Smith et al. 2008; Dinkelman and Keane 2010; Clerc et al. 2015; Geoffroy et al. 2015). The Inner SDRs are thus more likely to represent rollover structures controlled by continentward-dipping detachment faults (Gibson and Love, 1989; Geoffroy, 2005; Gernigon et al. 2006; Geoffroy et al. 2015). In NE Greenland, this might involve lateral outflow or passive exhumation of the ductile lower crust (Quirk et al. 2014).

4.2.3 Diachronous and segmented emplacement of SDRs: new evidence from the NGS

The fact that the SDRs acquire remnant magnetization provides information about the timing of emplacement of these large volcanic constructions. The chronology and magnetic polarity of the Lower Series at ODP Hole 642E (Fig. 9, 13) is complex due to widespread magnetization (Schönharting and Abrahamsen, 1989). However, the entire Upper Series drilled and analysed on the Vøring Marginal High (Figs. 9a and 13) shows a dominant reversed magnetic polarity correlated within reverse magnetic polarity Chron C24r (~57.1 to ~53.9 Ma)(Gradstein et al. (2012) (Schönharting and Abrahamsen, 1989; Eldholm, 1989). The low amplitude of the magnetic total field associated with the Landward Flows around and landward of ODP well 642 (Figs. 4, 7 and 9) is most probably the result of negative magnetization during the reverse C24r magnetic polarity (Fig. 13). In contrast, the main Inner SDRs wedge further west in the Vøring Marginal High (Fig. 9a) is characterized by a prominent positive magnetic signature (e.g. Berndt et al. 2001; Abdelmalak et al. 2016a, b). To explain this positif signature associated with the Inner SDRs, normal polarity and remanent magnetization are required for the uppermost part of the Inner SDRs not sampled by the ODP well. As the underlying lava flows are correlated with negative polarity chron C24r and because the flexure of the SDRs is most likely the consequence of a rapid development (< 2–3 Ma e.g. Lenoir et al. 2003), the positive magnetic anomaly associated with the Inner SDRs is here regarded as the result of late lava flows emplaced during a period of positive magnetic polarity including C24n3n (53.9-53.4 Ma), C24n2n (53.2-53.1) and/or C24n1n (53.0-52.6 Ma) (Fig. 13). Showing a positive magnetic signature, the main Inner SDRs on the Vøring Marginal High likely emplaced shortly during these normal polarities following the reverse magnetic chron C24r (57.1-53.9 Ma) (Fig. 13).

In contrast to the Vøring Marginal High, the Inner SDRs in the northern part of the Møre Marginal High have a negative magnetic signature (Fig. 4, 10) and formed landward of, and before, the first positive magnetic polarity chrons C24n1n and C24n3n and negative polarity chron C24r recorded in the oceanic domain mapped at the eastern edge of the oceanic Norway Basin (e.g. Gernigon et al. 2012, 2015) (Figs. 4, 7 and 10). This spatial configuration suggests that the Inner SDRs in the northern part of the Møre Marginal High were already emplaced during the Thanetian (late Paleocene), before chron C24r and therefore prior the emplacement of the adjacent Inner SDRs in the Vøring margin segment (Fig. 14). In the Vøring Marginal High, the Inner SDRs are principally post-C24r and initiated at least 1–2 my later, in the Ypresian (early Eocene) (Fig. 13). This indicates that the emplacement of the Inner SDRs was likely diachronous in the NGS and formed before and after the PETM event (Fig. 13).

4.2.4 Sill and dyke intrusions in sedimentary basins and deep continental crust

In the NGS, major sill complexes often reach several thousand km^2 and individual sills 50–300 m in thickness (Planke et al. 2005; Hansen and Cartwright, 2006; Schofield et al. 2017).

Svensen et al. (2004) estimated the volume of magma in the mid-Norwegian margin sill complex to be up to $0.9\text{-}2.5\times10^4$ km^3 and possibly more considering recent deep seismic data (Abdelmalak et al. 2017). The Faroe-Shetland Basin sill complex has been estimated to cover a minimum area of 22,500 km^2 (Schofield et al. 2017). Limited radiometric dates from the Faroe-Shetland Basin sill complex cluster around 55–52 Ma (Passey and Hitchen, 2011), although sills as old as Campanian (83-72 Ma) have been reported (Fitch et al. 1988; Schofield et al. 2017). Plutonic intrusions are also widespread in the northern Faroe –Shetland Basin (Passey and Hitchen 2011; McLean et al. 2017). They deformed the sediments before widespread flood basalt- and hyaloclatite emplacements in Ypresian time (~56-55 Ma) and the emplacement of the SDRs in the NGS.

On the conjugate NE Greenland margin, Granath et al. (2011); Tsikalas et al. (2012) and Reynolds et al. (2017) documented two Paleogene-aged sill complexes >10000–18000 km^2 in size. Onshore, doleritic sills and dykes as thick as 300 m are found throughout the strata of the Traill Ø region and increase in abundance towards the ocean (Parson et al. 2017). In the eastern Geographical Society Ø region (Fig. 2), dolerite sills contribute as much as 40% of the structural thickness of original Cretaceous mudstones (Price and Whitham 1997; Parson et al. 2017). In the Jameson Land Basin (Fig. 5), Eide et al. (2016) described the formation of both shallow and deep magmatic sills and their impact on host rock deformation and basin uplift. Most of the tholeiitic sills in this area were emplaced during the onset of the Early Eocene breakup (Larsen et al. 2014). Post-breakup (Early Eocene) intrusions are also common between the central East and NE Greenland margin. Some of the largest post-breakup intrusions such as the Kap Simpson and Kap Parry intrusive complexes may be part of the Traill Ø-Vøring Igneous Complex (e.g. Olesen et al. 2007; Gernigon et al. 2009).

At deeper crustal levels (Fig. 6), HVLC are traditionally thought to represent magmatic (gabbroic) intrusions and related cumulate layers emplaced within and/or underneath the pre-existing stretched continental crust during the breakup of the NGS (Eldholm et al. 2000; Mjelde et al. 2005). Continental HVLC (see Fig. 6) has P-wave velocities (and density) greater than that of average oceanic crust (>6.8-7.1 km/s)(Christensen and Mooney et al., 1995) and lower continental crust (>6.7 km/s after Holbrook et al. 2002). Velocity cut-off Vp-wave values of typically 7.1-7.3 km/s are commonly used to roughly delineate the top of such atypical velocity zones.

The integration of normal-incidence and wide-angle reflection seismic data has recently demonstrated that part of the HVLC can be seismically reflective (White et al. 2008; White and Smith, 2009; Granath et al. 2011) and/or could reveal elongated patches of high-velocity bodies interbedded within the pre-existing lower crust (Ravaut et al. 2006). The occurrence of reflection-free HVLC might indicate the presence of massive crustal intrusion and/or melt accumulations in the magmatic continent-ocean transition. The traditional 'underplating' model of HVLC implied massive mixing and accumulation of magmatic material near the base of pre-existing continental crust. This is interpreted to result in a new Moho, which is defined by differentiation of the melt into an upper gabbroic and a lower ultramafic layer which is thought to have occurred during thinning of the crust (Cox, 1993). The term 'underplating' is often used for both sill intrusion and/or basal accumulation of mafic/ultramafic material at the base of the continental crust (White et al. 2008; Thybo and Artemevia, 2013). Accumulated mafic layers up to a thickness of 10–25 km beneath pre-existing crust are realistic (Furlong and Fountain, 1986). The highest Vp of HVLC are commonly interpreted as breakup-related ultramafic magmas, richer in Mg than classic oceanic MORB (e.g. Holbrook et al. 2001). However, Meyer et al. (2007a, 2007b) note that Mg-rich magmas do not necessarily characterise high-degree or high-temperature melts and propose an origin in extensive decompression of upwelling mantle followed by later differentiation.

The HVLC along the NGS rifted margins are not only restricted to the flood basalt/SDR domain but can also extend into the oceanic domain as well as the adjacent continent, even over long distances (>100–200 km) (c.f. Funck et al. 2017) (Figs. 6, 11 and 14). Abdelmalak et al. (2017) mapped the regional distribution of HVLC beneath the NGS and noticed that the distal HVLC in Greenland is thicker than beneath the mid-Norwegian margin. Distal HVLC thins and narrows towards the Lofoten-Vesteralen segment (Fig. 6, 14). In the Mid-Norwegian margin, the top of the HVLC often coincides with a deep strong seismic reflection called the T Reflection (Figs. 5, 8, 9a) (Gernigon et al. 2003). Similar markers on top of HVLC have been observed on the JMMC and NE Greenland margin (Granath et al. 2011; Abdelmalak et al. 2017). An inherent fundamental problem with the interpretation of the HVLC (and related T Reflection(s)) is the partial overlap of seismic velocities and densities in models constituting intruded lower crust and metamorphosed and metasomatised crustal and/or metasomatised mantle rocks. Some portions of HVLC in the NGS may represent old, inherited, progressively exhumed lower continental crust (e.g. Gernigon et al. 2003, 2004; 2006; Ebbing et al. 2006; Wangen et al. 2011; Mjelde et al. 2016; Kvarven et al. 2016; Schiffer et al. 2015a,b; Abdelmalak et al. 2017). Similar high-velocity (Vp>7.0 km/s) and high-density (2.95-3.0 kg.m^{-3}) lower crust linked to high-grade metamorphic rock (e.g. eclogites, migmatites) was also interpreted in the northern North Sea (Fichler et al. 2011) and conjugate Greenland margin (Voss et al. 2009; Granath et al. 2011). In central East Greenland, Schiffer et al. (2016) suggested that the proximal HVLC bodies represent inherited metamorphosed and/or metasomatised crust or mantle, possibly associated with pre-Caledonian subduction (Schiffer et al. 2014; Schiffer et al. 2015a). Part of this material might unroof during subsequent rifting and thinning of the continental margin (Petersen and Schiffer, 2016). More controversially, highly altered serpentinised mantle below the central, distal and/or volcanic parts of the NGS conjugate rifted margins have also been favoured as explanation of the distal HVLC (Reynisson et al. 2010; Rüpke et al. 2013; Péron-Pinvidic et al. 2013). Different ages and non-magmatic interpretations of the HVLC have major implications for the thermal history, mantle temperature, crustal rheology and magmatic production because the amount of mafic material emplaced during the onset of breakup could be 20–40% less than has been hitherto thought.

4.3 Post-Eocene breakup: sea-floor spreading and geodynamic instabilities

4.3.1 Mapping the continent-ocean transition and fixing a continent-ocean 'boundary'

The process of demarcation of a continent-ocean 'boundary' is often biased and complicated due to contradicting definitions and ambiguous interpretations of sparse petrological and geophysical data (e.g. Boillot and Froitzheim, 2001; Eagles et al. 2015). On the distal edge of the continent-ocean transition, the continental oceanic "boundary" is usually defined between the outboard edge of highly attenuated, unequivocal continental crust (or potential zones of exhumed continental mantle) and the inboard edge of unequivocal oceanic crust. As recently summarized and compiled by Eagles et al. (2015) all possible interpretations of continent-ocean "boundaries" are ambiguous but the large range of interpretations simply reflects the uncertainties and limitation to accurately define a proper 'boundary'. Seismic sub-basalt imaging and the nature of rocks beneath the volcanic continent-ocean transition also add to the challenge of defining a discrete continental oceanic "boundary" in the NGS (White et al. 2008). On VPM, the magnetic signal is also dominated by volcanic constructions and/or dyke swarms in the extended continental crust (e.g. Schreckenberger, 1997; Berndt et al. 2001; Voss and Jokat, 2007), yet this method constrains pre-basaltic structures poorly (Reynisson, 2010).

Continent-ocean ‘boundaries’ in the NGS have variously been proposed to be located within the Inner SDRs (Mutter et al. 1982) or in the outer part of the Inner SDRs and/or closer or the volcanic Outer Highs or Outer SDRs (e.g. Skogseid and Eldholm, 1989). On the northern Faroe margin, seismic-refraction data suggest the continent-ocean ‘boundary’ is at the outer edge of the Inner SDRs which are underlain by intruded continental crust (Fig. 6)(White et al. 2008). On the Vøring Marginal High seismic-refraction data have been interpreted to indicate stretched continental crust and anomalously thick oceanic crust, separated by a 25-km-wide magmatic transition zone (e.g. intruded continental crust) overlain by an Inner SDRs wedge (Mjelde et al. 2005; Adbelmalak et al. 2016; Funck et al. 2017). On the outer Møre Basin, the continent-ocean 'boundary' is interpreted to be landward of magnetic chron C24n3n (Gernigon et al. 2012, 2015) near C24r (Figs 7). The oldest edge of magnetic chron C24r (57.1-53.9 Ma) is distinct south of the Vøring transform margin (Fig. 8b) where volcanism was minor with no SDRs (e.g. Berndt et al. 2001). Chron C24r marks the western limit of the ambiguous crust interpreted either as highly intruded continental crust or ‘mullioned’ continental crust/embryonic oceanic crust (Afar-type) (Gernigon et al. 2012, 2015). Near the transform margin, magnetic data suggest a possible Paleocene and related dyke swarm which coincides with a local phase of Paleocene extension in the sedimentary basin (Figs. 4, 7 and 14).

On the conjugate NE Greenland margin, the continent-ocean transition and ‘boundary’ location differs between authors. The continent-ocean boundary between 72.5°N and 74°N as proposed by Scott (2000) is located 75 km landward the position interpreted by Tsikalas et al. (2005). Ambiguous magnetic anomalies on the shelf might be related to oceanic seafloor spreading (e.g. Scott, 2000); alternatively, the origin of these anomalies may be dyke intrusion and/or shallow breakup lavas on strectched continental crust (Tsikalas et al. 2002; Voss and Jokat, 2007). Based on seismic velocities, Voss and Jokat (2007) argue for a 120 to 130-km-wide transitional crustal zone (Fig. 6). Accordingly, their alternative continent-ocean boundary is interpreted to lie about 130 km seaward from the original interpretation by Scott (2000).

Distinct SDRs are present along the NE Greenland continental slope (Hinz et al. 1991; Abdelmalak et al. 2016a, b; Geissler et al. 2016) (Figs. 2, 4, 11 and 12). Assuming the hypothesis that the Inner SDRs were at least partially emplaced over continental crust the continent-ocean ‘boundary’ might be located east of the NE Greenland continental slope (Fig. 11). Geissler et al. (2016) suggested that the continent-ocean ‘boundary’ might be located variably across the Inner or Outer SDRs or landward of the Inner SDRs. In agreement with Voss and Jokat (2007), our reconstruction suggests that the expected continent-ocean 'boundary' should be located preferentially anywhere in between the proximal bordure of the Outer SDRs and their distal limit.

4.3.2 First magnetic chrons in steady-state oceanic crust

In this study on the NGS, oceanic crust is most likely present if there are conjugate, continuous, kilometre-scale prominent magnetic striped anomalies (>400–500 nT at sea level) outside the Inner and Outer SDR domains (Figs. 2, 4 and 7). The oldest oceanic magnetic anomaly chrons recognized in the NGS commonly relate to the C24 magnetic sequences (Fig. 13 and 14). East of the NE Greenland margin, magnetic chrons C24n3n and C24n1n (Fig. 4) have been identified east of the Greenland Escarpment (Fig. 1, 2) and related SDRs (e.g. Vogt et al. 1986; Skogseid et al. 2000; Tsikalas et al. 2001; Geissler et al. 2016). In the conjugate Lofoten-Vesterålen margin, Tsikalas et al. (2001) and Olesen et al. (2007) have also found both C24n3n and C24n1n magnetic polarity chrons from the Vøring Marginal High up to the Senja Fracture Zone (Fig. 7). North of the Bivrost Lineament, some of the C24n3n and C24n1n anomalies formed above SDR wedges detected seismically (Tsikalas et al. 2001). Further north, this pair of chrons corresponds to oceanic crust as shown by a recent OBS profile across the Lofoten margin (Breivik et al. 2017). However, the presence of SDRs at that level and towards the Barents Sea Shear margin is unclear (Berndt et al. 2001). Closer to the Vøring Marginal High, the C24 magnetic sequences are indistinguable and the nature of the crust is unclear (Figs. 4, 7 and 14). Only chrons C23-C22 and younger anomalies show the characteristic signature of a continuously-striped, steady-state spreading system (Gernigon et al. 2009) (Fig. 7). The magnetic anomalies located continentwards of chron C23 in the Vøring Marginal High mostly reflect the volcanostratographic sequences imaged in seismic reflection data (e.g Berndt et al. 2001; Abdelmalak et al. 2016b) (Fig. 7 and 14) and offer little insight into the underlying substratum. The underlying crust may be heavily intruded continental crust rather than oceanic crust.

In the Norway Basin, the earliest oceanic magnetic chron (C24n1n) merges with the youngest edge of chron C24n3n at the Faroe platform shear margin where C23-C22 are the first undisputed chrons (Gernigon et al. 2012, 2015)(Fig. 7). Additional magnetic features, landward of magnetic chron C24n3n are distinct but do not show a linear pattern. These may correspond to embryonic oceanic or highly intruded continent crust (Gernigon et al. 2012). Large uncertainties remain about the pre-C24 magnetic chron identification, but Gernigon et al. (2015) have estimated slow spreading local rates during the embryonic stage of breakup (<5–10 mm/year), increasing rapidly to more than 15–25 mm/year during formation of chrons C24n3n and C24n1n. Farther south, correct identification of oceanic magnetic anomalies in the GIFR is difficult (Figs. 4 and 7). There are complications, especially due to lack of data west of the Faroes Platform, huge variations in basement depth and excess volcanism. In addition, seismic reflection seismic profiles have imaged several older, abandoned rift axes on the GIFR (Hjartason et al. 2017) and geological mapping has identified other in Iceland (Garcia et al. 2008), which further complicate the situation.

4.3.3 Oceanic Phase I: Early Eocene oceanic crust domain from C24r to C22 (57.1-48.5 Ma (Early-mid. Eocene)

From chrons C24-C22 (Figs. 15-16) steady-state seafloor spreading was locally initiated with dominantly high-spreading rates in the NGS. Between the Lofoten and NE Greenland margins, the earliest half-spreading rates varied between 15 and 25 mm/y (e.g. Gaina et al. 2009). After C24n1n, half-spreading rates of the earliest spreading phase were on average less than 25 ± 2 mm/year in the eastern part of the Norway Basin, and less than 20 ± 2 mm/year in its western part (Gernigon et al. 2015). Generally, seafloor spreading was not stable as indicated by off-axis seamounts and relics of aborted rift axes between magnetic polarity chrons C24n3n and C22n (see Gernigon et al. 2012, 2015). Instability in the oceanic rift axes might explain the spreading asymmetry of the Norway Basin during the Early to Late Ypresian. During this C24r – C22 period, and thus a few Myrs after first SDR emplacements, excess melts developed north of the East Jan Mayen Fracture Zone (Vøring Spur locality, Figs. 2, 6 and 15). This renewed phase of atypical melting is possibly associated with the long-lived leaky transform that comprises the Jan Mayen Fracture Zone and Traill Ø-Vøring Igneous Complex (Gernigon et al.

2009). Both the age and origin of this post-Early Eocene breakup magmatic event are, however, disputed (e.g. Breivik et al. 2008; Tan et al. 2017).

Tectonic instabilities during that period are also recorded in the sedimentary strata on the adjacent continental margins (Fig. 13). In the Faroe–Shetland region, this interval is characterized by an alternating succession of deltaic and shallow-marine sediments within the Faroe-Shetland Basin (Ólavsdottir et al. 2013; Stoker et al. 2018).

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

Aeromagnetic data indicate a different seafloor spreading phase (Phase II) between C21r and C10 (47.3-28.8 Ma – mid-Eocene to Late Oligocene) (Figs. 13–16). In the Norway Basin, a distinct spreading reorganization from a direction North 340° to North 8° occurred between C22n and C21 (Gernigon et al. 2008, 2012). After 47 Ma (C21r), the entire Aegir Ridge adjusted from slow to ultra-slow spreading rates (full spreading velocities decreased from 16±2 to 8±2 mm/year to the south). During Phase II, a northward-widening fan-shaped seafloor spreading pattern developed in the Norway Basin, and an almost orthogonal spreading system between the Lofoten-Vesterålen and NE Greenland margins (Figs. 16 and 17).

These variations in oceanic spreading around 47 Ma (C21r) coincide with tectonostratigraphic events in peripheral sedimentary basins of the NGS (Gernigon et al. 2014; Stoker et al. 2018) (Figs. 13 and 16). The transition from oceanic Phase I to Phase II might also coincide with the collision of NE Greenland and continental fragments in the Arctic Ocean, resulting in compression events of the Eurekan and West Spitsbergen Fold-and-Thrust Belt that are, however, poorly dated (e.g. Piepjohn et al. 2016; Jones et al. 2016). In comparison to the Norway Basin, there is no clear evidence of the C21r reorganization south of the GIFR, though this chron interval does mark the change from fragmented and discontinuous spreading (C24-C22) adjacent to the Rockall-Hatton Plateau to the first definitive and continuous magnetochron (C21) within the evolving Iceland Basin (Kimbell et al. 2005; Elliott and Parson, 2008; Stoker et al. 2012). The fragmented and discontinuous rupture along the Rockall – Faroe-Shetland continental margin has been interpreted by Stoker et al. (2012) as a rift-to-drift phase from C24 to C22 that is also supported by the sedimentary record (Fig. 13). A significant readjustment in the spreading system between SE Greenland and the Rockall-Hatton Plateau occurred later in the late Mid- to Late Eocene interval (C18/C17,~38 Ma), marked by the change from orthogonal to oblique seafloor spreading (Stoker et al. 2012) accompanied by formation of diachronous V-shaped ridges and troughs (Martinez and Hey 2017).

During the Oligocene (chron C13n period; 33.5-33.0 Ma), another major NGS reorganization occurred when the northwestern arm of the triple junction between North America, Greenland and Eurasia was abandoned (Sirastava and Tapscott, 1986; Gaina et al. 2009). A change in the opening direction between Eurasia and North America/Greenland from NNW-SSE to NW-SE may explain this, along with initiation of the West Jan Mayen Fracture Zone (Mosar et al. 2002; Talwani and Eldholm, 1977). This fracture zone, however, may have initiated earlier, closer to chrons C22/C21 (see Gernigon et al. 2009). Nonetheless, the progressive termination of slow to ultra-slow spreading along the Aegir Ridge finalized after magnetic chron C10, the last magnetic chron which is presently confirmed by modern aeromagnetic data in the Norway Basin (Gernigon et al. 2012).

4.3.5 Oceanic Phase III: Formation of the Kolbeinsey Ridge and final JMMC dislocation from C10 to C6B (28.2-22.5 Ma)

During oceanic phase III from C10 to C6B, seafloor spreading was still active along the Mohns and Reykjanes Ridges. However, the region between the GIFR and the Jan Mayen Fracture Zone was affected by a major relocation of the spreading ridge leading to the extinction of the Aegir Ridge and the final dislocation of the JMMC from Greenland in the Late Oligocene/Early Miocene (Fig. 17). There is certainly a link between the variable spreading rates along the Aegir Ridge, together with the fan-shaped opening of the Norway Basin and the progressive detachment of the JMMC from the Greenland margin. To accommodate the unambiguous fan-shaped development of the Norway Basin (Fig. 4) it is widely accepted that at least parts of the JMMC must have rotated counterclockwise after the onset of breakup at C24r (e.g. Talwani and Eldholm, 1977; Nunns, 1983). However, there is no consensus about the initial development and precise timing of fan-shaped spreading between C20 and C7 in the Norway Basin (Talwani and Eldholm, 1977). Possible times are between chrons C13 and C5D (Unternehr, 1982), C18 and C6B (Vogt, 1986), after C20 (Nunns, 1983) or after C13 (Müller et al. 2001; Torsvik and Cock, 2011). The most recent aeromagnetic data from the Norway Basin (Fig. 4) show that significant rifting and/or intermediate spreading episode(s) between East Greenland and the southern Aegir Ridge mostly developed after chron C21r from Middle Eocene to Late Oligocene (Gernigon et al. 2012, 2015). The complete separation of the JMMC from Greenland and the development of the Kolbeinsey Ridge likely took place later, at about 22 Ma shortly before or during chron 6B in the Early Miocene (Figs. 13 and 17) (e.g. Vogt et al. 1980; Gaina et al. 2009). During spreading ridge reorganization, major magmatic production and atypical 'underplating' was also important in the GIFR region and related to complex sub-lithospheric processes (small-scale convection, astenospheric upwelling) and/or lithospheric mantle composition (Smallwood et al. 1999, Foulger et al. this issue).

5. DISCUSSION: MECHANISM OF PLATE BREAKUP AND THE DIACHRONOUS EVOLUTION OF THE NORWEGIAN-GREENLAND SEA

5.1 The role of pre-Tertiary extension in breakup of the Norwegian-Greenland Sea

5.1.1 Pre-Tertiary extension and crustal processes prior to formation of the volcanic rifted margins

The diversity of continental margins challenges genetic models for rifting. One of the key questions is the relative influence of inheritance, the thermal state and composition of the lithosphere, and the stress field (e.g. Schiffer et al. this issue). It is debated if different rifted margin types evolve similarly during the onset of breakup (Péron-Pinvidic et al. 2013) or if there are fundamental differences between volcanic and magma-poor rifting (Franke et al. 2013; Geoffroy et al. 2015; Gernigon et al. 2015; Tugend et al. 2018).

In terms of timing and lithospheric configuration, breakup of the NGS was not the result of a continuum of lithospheric deformation (e.g. without episodes of thermal cooling) initiated in Jurassic or earlier (Fig. 18). Classic 'magma-poor' rifting models or simulation commonly involve a shorter and continuous sequence of stretching, thinning and exhumation lasting tens of Myrs (< 30–50 Ma) and leading to localization of deformation before breakup (e.g. Lavier and Manatschal, 2006; Reston, 2009; Brune et al. 2014; Andrés-Martinez et al. 2019). In the

NGS, much longer periods of Late Paleozoic and Mesozoic extension are expected to have left attenuated continental crust (see Chapter 3). However, severe crustal and lithospheric deformation in the Late Jurassic predates the last rifting phase and breakup stage(s) by more than 50–70 Myrs. Accordingly, older rifted sedimentary basins in the NGS were probably under thermal relaxation before the Late Cretaceous-Paleocene renewed phase of rifting (e.g. van Wijk and Cloetingh, 2002; Fletcher et al. 2013) (Fig. 18). This is expected to result in long-term lithospheric strengthening (van Wijk and Cloetingh, 2002). A lithosphere resistant to subsequent reactivation offers an elegant explanation for why the different NGS rifting axes (e.g. Late Jurassic-Early Cretaceous and Late Cretaceous-Paleocene) migrated from the proximal to the central and then the distal domain (Fig. 18). Older failed rift basins preserved in the shallow platform areas of the NGS (e.g. Danmarkshavn Basin; Trøndelag Platform) are much further off-axis (Fig. 2). Further south, similar rift failure/migration or lateral 'jumps' of the rift axes are also known from the Rockall and Faroe-Shetland basins towards the western flanks of the Faroes and Rockall-Hatton plateaus (Guan et al. in press). Here, late rifting and VPM formation initiated on thicker isolated continental distal blocks or ribbons (White and Smith, 2009).

In the best understood 'magma-poor' rifted margin - Iberia-Newfoundland (Boillot and Froithzeim, 2001; Perez-Gussinye et al. 2001; Reston, 2009; Sibuet and Tucholke 2012; Mohn et al. 2015) - formation of oceanic crust was preceded by drastic thinning of the continental crust leading to serpentinisted mantle exhumation and denudation before breakup. In the NGS, seismic data and potential field modelling suggest that above the distal HVLC and associated crustal rafts and/or marginal plateaux there is continental crust about 5–10 km thick (e.g. Nirrengarten et al. 2014; Theissen-Krah et al. 2017; Manton et al. 2018)(Fig. 6). Structural geology and potential field modelling predict syn-rift exhumed continental mantle in the mid-Norwegian margin to be limited to localized areas (Gernigon et al. 2015; Maystrenko et al. 2017c; Zastrozhnov et al. 2018) while throughout vast areas crustal thinning may have been insufficient to allow a large zone of exhumed and heavily serpentinised mantle material to form. Alternatively, unstructured ductile and/or 'boudined' lower crust could have developed beneath the thick basin sequences observed (e.g. Clerc et al. 2015; Petersen and Schiffer, 2016; Andrés-Martinez et al. 2019). Heavily serpentinised mantle is expected to generate high magnetic susceptibility and remanence (e.g. Oufi et al. 2002). However, the low magnetization of the thinnest continental crust in the NGS (Figs. 4) calls into question the presence of a large zone of exhumed continental mantle in the central and distal parts of the rifted margin. Magnetotelluric data (Myer et al. 2013) also casts doubt on the existence of heavily serpentinised mantle in the outer Vøring Basin. In areas with clear Moho reflections at the base of the HVLC and continentward of the main volcanic province, as in the distal part of the Vøring margin, large zones of exhumed continental mantle are unlikely (Mjelde et al. 2002). There, the presence of preserved lower crust later intruded by breakup related intrusions has been favoured to explain the distal HVLC close to the volcanic margin (Gernigon et al. 2004; Abdelmalak et al. 2016a, see also chapter 4.2.4)(Fig. 19). The petrology of the subaerial andesitic lavas drilled on the Vøring Marginal High (ODP 642) also indicates continental crust beneath the Inner SDRs (Meyer et al. 2007a).

In the outer part of the mid-Norwegian margin, Late Cretaceous-Paleocene rifting and VPM developed westwards of a paleo-necking zone in the outer flexure zone of large Cretaceous sag basins already affected by cooling (Figs. 9a and 19). The Late Cretaceous-Paleocene rifting event preceding volcanic margin formation particularly affected the outer Møre and Vøring basins and the Fugløy Ridge amalgamated with the proto-JMMC and possibly the NE Greenland "outer structural high" further north (Fig. 14). They likely formed a thicker continental distal marginal plateau only mildly affected by the mid-Mesozoic thinning event(s) (Fig. 18). It seems that the NGS VPMs formed preferentially in thicker continental domains as earlier predicted by modelling studies (Petersen and Schiffer, 2016; Davis and Lavier, 2017). This indicates that there is indeed a major difference in the processes leading to end-member magma-poor and volcanic rifted margins that depends not only on the volumes of available magma but possibly on the modes of crustal thinning and periods of thermal cooling expected before SDR formation. The main difference between the NGS volcanic margin and "magma-poor" margins is also the surface expression of volcanism during and before the onset of breakup (e.g. SDR formation, Fig. 19). In so-called “magma-poor margins”, only minor volcanism is observed and it is erupted at a late stage of denudation and at different water depths (Reston, 2009).

5.2.2 Final weakening of the lithosphere by thinning and/or magmatism

Between NE Greenland and the Lofoten-northern Vøring margins, Skogseid et al. (2002), Tsikalas et al. (2002) and Lundin and Doré (2018) showed that the SDRs and the breakup axis trend obliquely to the pre-existing Paleozoic/Mesozoic rift structures (Fig. 14). Even if the breakup axis cuts obliquely across the latest rift structures of Late Cretaceous-Early Paleocene age, the line of breakup generally remains off-axis compared to the previous (failed) Palaeozoic and early-mid Mesozoic rifts domains (Fig. 2 and 20). This oblique breakup creates the final but "apparent" asymmetry of the margin that does not necessarily reflect the configuration of the pre-breakup rift system

The early breakup phase in the NGS is poorly constrained and various breakup scenarios and models have been proposed (Fig. 20). Drastic lithospheric attenuation leading to breakup is often inferred to be accommodated and controlled by large-scale crustal detachments (Lister et al. 1991; Mosar et al. 2002; Brun et al. 2018). Active shear zones are often predicted to thin significantly the attenuating crust and upper mantle, facilitate strain localization and therefore accelerate rifting before breakup (Lavier and Manatschal, 2001; Ranero and Perez-Gussinyé, 2010; Mohn et al. 2015; Petersen and Schiffer, 2016; Brune et al. 2017). Beniest et al. (2018) also showed that a continental lithosphere can break at a different location rather than at a pre-existing herogenity, without the presence of thermal anomalies and/or inherited weak zone. Combined with lateral cooling of the lithosphere, the proximal to distal migration of the conjugate rift system may have already predisposed the region to final obliquity of the breakup axis between NE Greenland and the Vøring/Lofoten margins (Fig. 20a).

Large-scale strike-slip deformation was recently proposed to have been influential in the sharp (i.e. without drastic crustal thinning) and oblique Eocene breakup of the NGS conjugate margins (Lundin and Doré, 2018) (Fig. 20b). Stoker et al. (2018) show that sediment flux and style of deformation support strike-slip motion in the breakup of the Faroe-Shetland/SE Greenland region. Dextral motion may have resulted from plate motion of ~100 km to the ENE of Greenland relative to North America and Eurasia during the Paleocene-Eocene (Nielsen et al. 1997; Guarnieri 2015; Jones et al. 2017). Thermo-mechanical feedback between shear heating coupled with temperature-dependent viscosity in continental zones of oblique extension may significantly weaken the lithosphere and facilitate rapid breakup (e.g. Minakov et al. 2013). A predisposed set of dextral transforms and large-scale shear deformation (e.g. transcontinental shear zones) may also explain why the line of breakup obliquely bisected the pre-existing Cretaceous rift system in the NGS (Skogseid et al. 2000; Lundin and Doré, 2018; Schiffer et al.

this volume) (Fig. 2). Nevertheless, the pattern of inherited structures in deep parts of the margin remains unclear and/or controversial (see the discussion about the nature of the distal HVLC).

Another possible explanation for the oblique and locally sharp breakup during the Early Eocene is magmatism (Fig. 20c). The most significant pre-, and syn-breakup phases of magmatism concentrated in the distal VPMs and were transient. The cause of breakup magmatism is still debated (e.g. mantle plume, small-scale convection, delamination of the lithosphere, fertile inherited material; White and McKenzie, 1989; Nilsen and Hooper, 2004; Foulger and Anderson 2005; Meyer et al. 2007b; Simon et al. 2009; Korenaga et al. 2004; Petersen et al. 2018; Hole and Natland, 2019), but multi-level magmatic processes likely influenced the final plate separation (Fig. 14). In any case, modelling and field examples show that lithosphere softening associated with melting mantle instabilities at the base of the thermal lithosphere (Nicolas et al., 1994; Geoffroy et al. 2007; Gac and Geoffroy, 2009) and/or crustal magma chambers (Callot et al. 2002; Doubre and Geoffroy, 2003) and/or magmatic dyke intrusion (Ebinger and Casey, 2001; Keir et al. 2006) can accommodate, localize (or relocalise) the deformation at different lithospheric levels. Magma-supported lithospheric breakup might occur at much lower stress levels than theoretically needed for faulting (e.g. Buck and Karner 2004). Narrow sub-crustal intrusions may also localize better the deformation and rapidly thin the lithosphere drastically (Callot et al. 2002; Yamasaki and Gernigon, 2009). Rapid transition from continental rifting to oceanic crust formation without mantle exhumation and melt induced weakening has also been found in moderately magmatic rifts context (e.g. South China Sea, Clift et al. 2001; Cameselle et al. 2017). As observed on exposed volcanic passive margins, high-density axial mafic dyke swarms intruding into attenuated continental crust are also expected to increase significantly down-flexure of the opening rift since subsidence increases with plate weakening and heating during the final stage of continental breakup (Klausen and Larsen, 2002; Geoffroy et al. 2015; Guan et al. in press). Magmatism thus is a potential candidate to explain the oblique rupture of the lithosphere in the NGS independently or not of pre-existing rift architecture and associated lithosphere configuration (Fig. 20c). What guides the spatial distribution of magmatic intrusions is still unclear and it may not be entirely controlled by pre-existing rifting axes and related structures. In the NE Atlantic some major Palocene-Eocene igneous centres (e.g., in Scotland) were emplaced far from the rift and proto-breakup axes (Meyer et al. 2007). Inherited lithospheric and crustal structures and material and/or deeper mantle processes (e.g. asthenospheric instabilities and heterogeneities) may have played an important role (Callot et al. 2002; Geoffroy et al. 2007; Schiffer et al. this issue).

Both processes, highly oblique spreading and magma-assisted rifting, are likely candidates to explain final and sharp rupture in the NGS (Fig. 20). However, the question remains why the onset of breakup was discontinuous in space and time and did not instantaneously affect longer margin sections. Variations in the rift-related magmatic budget and related HVLC distribution may have been influential. Early breakup areas reveal crustal domains with typically thin and narrow distal HVLC while, in contrast, delayed-breakup regions are often associated with thick and broad HVLC (Fig. 14). Yamasaki and Gernigon (200) showed that the width of the HVLC (inherited or not) has a strong influence on the localization and amount of the lithospheric deformation for any thermal condition in the uppermost mantle; a greater amount of thinning is distributed into narrower HVLC region and vice versa. It has also been suggested from the study of magma flow from the main distal dyke swarms that most of the magma feeding the SDRs came from related large Paleocene to Eocene igneous centres (Callot et al. 2001; Callot and Geoffroy, 2004). A possible explanation for early spreading segmentation is that the base of the thermal lithosphere was unstable beneath the whole area with numerous small-scale convection instabilities developing upward through vigorous thermal erosion, feeding the igneous centres (and thus dykes and lavas), and weakening the lithosphere locally and diachronically (Callot et al. 2002; Geoffroy et al. 2007). This may have equally happened under oblique extensional stress.

5.2 Development of steady state oceanic crust

5.2.1 Segmented VPMs and potential role of transfer zones

The origin of the segmentation of the VPM and associated discontinuous initial seafloor spreading cells in the NGS remains unclear. Koopman et al. (2014) suggest a causal link between magmatic volume variations and inherited rift segment boundaries. They propose that the segmentation of VPMs was influenced by pre-existing rift transfer zones that delayed rift-to-drift progression and propagation in the Atlantic Ocean. Presuming that the transfer zones formed prominent lithospheric discontinuities at the onset of rifting (e.g. Tsikalas et al. 2004), their presence may have influenced melt generation along different VPM segments of the NGS. In the north, the Bivrost Lineament fits the southern limit of a narrow distal HVLC observed between the narrow Lofoten-Vestsrålen margin and the wider NE Greenland margin (Fig. 6 and 14). The Bivrost Lineament (Fig. 7) reflects a profound and old crustal, and possibly mantle structural discontinuity expected between the Lofoten – Vesterålen and the Vøring margin (Maystrenko et al. 2017b). In the south, the Jan Mayen Corridor (Figs. 7 and 14), represents a complex Mesozoic transfer zone and coincides with a relatively amagmatic segment between the highly magmatic Møre and Vøring marginal highs (e.g. Berndt et al. 2001a, 2001b; Abdlemalak et al. 2016). Both the Bivrost Lineament and the Jan Mayen Corridor have a clear structural expression at the level of the HVLC (Figs. 7 and 14). Inheritance along the Jan Mayen Corridor is not only seen in the failed Mesozoic rift (relay zone at BCU level, Figure 7) but could indicate a direct or indirect link with older Precambrian lineaments and/or shear zones recognized or inferred onshore (Doré et al. 1997; Schiffer et al. 2018). Torske and Prestvik (1991) also show that such oblique and inherited structures strongly influence magma migration to upper crustal levels even far from the Late Cretaceous-Paleocene rift and breakup axes (e.g. the Vestbrona sill complex, Hafeez et al. 2017).

The first order segmentation of oceanic sub-basins is well preserved between the Jan Mayen and Norway Basin fracture zones (Fig. 7) but is possibly an indirect consequence of inherited structures and continental rift segmentation (Gernigon et al. 2012, 2015). It remains unclear if all oceanic fracture zones are (indirectly) inherited from pre-existing zones of weakness in the continental lithosphere (e.g. Lister et al. 1991; Taylor et al. 2009; Bellahsen et al. 2013) or simply the result of deeper thermal or compositional heterogeneities in the nascent spreading centre (e.g. Lizarralde et al. 2007). While the East Jan Mayen Fracture Zone likely is a combined (but indirect) oceanic fracture zone/continental transfer zone system, other pre-existing continental transfer or accommodation zones at the Vøring margin lack a clear relationship to potential fracture zones in the oceanic domain. Even if the early seafloor spreading likely initiated north of the Bivrost and Surt Lineament (Figs. 7 and 14), the Bivrost and Surt lineaments (or transfer zones) do not appear to have controlled nor influenced any oceanic fracture zones after breakup (Olesen et al. 1997; 2007; Maystrencko et al. 2017b)(Figs. 7, 14–17). Farther south, the structural relationships between controversial NW-SE transfer zone across the Faroe-Shetland Basin (Moy and Imber, 2009) and oceanic fractures zones inferred in the southern part of the Norway Basin or along the GIFR are also far from being well understood (Schiffer et al. 2018). The origin and role of lithosphere-scale transfer zones that result in segmented oceanic rift sections (or not) is thus a potentially fertile topic for future studies.

5.2.2 Compartemental "unzipping" of the volcanic margins

Volcanic margin formation and breakup in the NGS are typically regarded as processes that occurred almost instantaneously across the whole NGS segments at around 54 Ma (C24r) (Eldholm et al. 1989; Skogseid et al. 2000; Gaina et al. 2009). However, assuming rigid plates, and homogeneous lithosphere, an oceanic rift should, in theory, propagate over time through the extended lithosphere as a natural consequence of the relative motion of the plates about a pole of rotation (Vink, 1982). It is unlikely that either instantaneous breakup of long (>200–400 km) marginal sections or progressive opening of the oceanic basin occurred around one common pole of rotation in the NGS. Due to the complex geological history of the NGS, it is also improbable that the pre-breakup lithosphere was homogenous. Therefore, VPM development was likely a step-by-step and diachronous process (section 4.2.3). The oldest oceanic cells nucleated separately first in the northern part of the outer Møre margin and west of the Lofoten-Vesterålen margin, close to the Greenland Fracture Zone (Figs. 14 and 15). The individual nascent oceanic sub-basins show progressive development e.g. in the Norway Basin. SDRs and early spreading developed progressively towards the Faroe Plateau/GIFR and the Vøring Marginal High where time delays in the breakup are indicated (Fig. 21).

At the Vøring Marginal High, from magnetic anomalies there are locally no firm indications of undisputed oceanic accretion before magnetic chrons C23r (Figs. 4, 7 and 14). If it had occurred earlier, the Inner SDRs would overlie oceanic crust, an unlikely scenario given ODP results that show that continental crust and andesitic lava flows are likely present underneath (e.g. Meyer et al. 2007a; Abdelmalak et al. 2016a). The seismic outlines of both Inner and SDRs are often oblique to the main oceanic magnetic chrons (Fig. 7). So far, we have also no clear indicator that oceanic crust is present underneath the Outer SDRs and none of them have been drilled in the NGS. The presence of residual continental material within the mafic crust cannot be totally disregarded but so far, undisputed oceanic crust is mostly assured oceanward beyond the Outer SDRs. At the conjugate NE Greenland margin, the controversial continent-ocean ‘boundary’ and the difficult interpretation of the C24o magnetic chron at the continental slope were proposed indifferently from the inner edge of the Inner SDRs to the outer edge of the Outer SDRs (e.g. Funck et al. 2017; Geissler et al. 2016; Horni et al. 2017). If correct, this interpreted configuration and geometry would suggest that part of the Inner and Outer SDRs formed almost simultaneously or partly overlapped in space and time; thus contradicting the classic sequential model proposed by Planke et al. (2000) that the Inner SDRs formed before the late and distal Outer SDRs (e.g. Fig. 12). Alternatively, the first magnetic spreading anomalies in the oceanic domain could simply terminate progressively from north to south against the intruded continental crust and growing Outer SDRs, supporting a southward propagation of the early seafloor spreading cells in the NGS in agreement with Voss and Jokat, (2007) and Voss et al. (2009)(Figs. 14, 15 and 21). In this scenario, spreading propagated southward from the Greenland/Senja Fracture Zones at 54.2 Ma to the proto–Jan Mayen Fracture Zone at 50 Ma (e.g. Koopmann et al. 2014). It thus did not open towards the rotation poles of the Eurasian and Greenland plates that were located in northern Siberia at the time (deMets et al. 1990; Franke et al. 2000; Gaina et al. 2009).

In the Norway Basin, the kinematics were different and partly conformed with the ‘Vink' unzipping model but mainly after 47 Ma (C21r). The Eocene oceanic rift initially propagated from north to south towards calculated rotation poles situated on the GIFR (Gernigon et al.

2012, 2015). However, around C21r, the spreading configuration changed and the main extension and possibly intermediate spreading increased in the southern part of the JMMC (Figs. 16 and 17). At magnetic chron C21 time, full breakup between SE Greenland and the northern Rockall VPM (Kimbell et al. 2005; Elliot and Parson, 2008; Guan et al. in press). It may explain sudden stress release triggering the rifting pulse and relocation of the spreading axis observed between Greenland and the JMMC further north.

As pointed out by Martin (1984) and Nirrengarten et al. (2018), it appears that oceanic spreading centres preferentially grew within segments of heterogeneous and stretched continental lithosphere with or without a direct correlation with the propagation direction derived from global plate motion. In the NGS, the Vøring Marginal High, Faroe Plateau and, to some extent, the GIFR may simply have comprised (weak or strong) lithospheric barriers to volcanic rift propagation that hindered lateral growth of the volcanic margins and/or the narrow localization of nascent oceanic rifts (Figs. 14 and 21). The presence of such buffers may have favoured the complex rift-ridge overlapping that led to breaking off of the JMMC (Figs. 15–17). Some authors argue for the influence of a mantle thermal instability (Muller et al. 2001; Koptev et al. 2017) while others earlier advocated an overlapping rift configuration as a first order and simple key condition for microcontinent formation (Auzende et al. 1980; Unternehr, 1982; Molnar et al. 2018). The JMMC did not form simply by instantaneous ridge-jumping according to the global microplates classification of Li et al. (2018). Here, the dislocation is interpreted to have been progressive (i.e. a rift-overlap type of microplate) and not necessarily the result of sudden rift relocation driven by plume interaction. The observed relative magma-poor evolution of the western Jan Mayen margin (Fig. 6, Kodaira et al., 1998) would also be in conflict with a major thermal instability as origin. If a rift- and ridge-overlapping system controlled the JMMC formation, a consequence is the preservation of continental material along the Faroe-Iceland-Ridge (e.g. Gernigon et al. 2012, 2015; Ellis and Stoker, 2014). Recent work suggests that a deep inherited Caledonian slab along the trend of the GIFR may have guided and controlled the rift overlap, magmatism, and dislocation of the JMMC (Schiffer et al. 2017, Schiffer et al. this issue). The origin and crustal composition of the GIFR continue to pose major unsolved questions (Bohnhoff and Makris, 2004; Foulger, 2006; Torsvik et al. 2015) that present research challenges for the future (Foulger et al. this issue).

6. CONCLUSIONS

We draw the main following conclusions:

1. Rifting and breakup in the NGS occurred in heterogeneous continental lithospheric assemblages.
2. Palaeozoic to Cenozoic phases of rifting with intermediate cooling events preconditioned the NGS lithosphere prior to propagation of breakup. The final Late Cretaceous- Paleocene rift is commonly offset from the dominant mid-Mesozoic or older thinning axis of the NGS rifted margins. Strain hardening due to Mesozoic lithospheric cooling may explain the migration of the rift axes over time and may have partly influenced the final breakup configuration.
3. Continental breakup in the NGS was highly magmatic, with large-scale volcanism (e.g. the SDRs) during VPM formation. VPMs developed preferentially either on thick continental ribbons (e.g. the Faroe Block) or moderately thinned continental crust (e.g. the Møre Marginal Plateau). The presence of a large zone of exhumed continental mantle before

breakup and beneath the subaerial SDRs is unlikely. This is in marked contrast with magma-poor margins where extensive subcontinental mantle exhumation occurred in deep water prior to igneous crustal accretion.

4. The final NGS line of breakup may have been controlled by highly oblique extension and associated plate shearing and/or melt intrusions and dyking before and during Inner SDR formation.
5. Breakup of the NGS was neither instantaneous nor continuous across the entire region. Spreading ridges propagate initially towards local Euler poles, die out, migrate laterally, jump, change their propagating direction or eventually bifurcate. Spreading in our preferred model initiated in isolated cells in the northern Greenland Sea and the northern Norway Basin. The nascent "segmented ocean" grew step by step with a delay of 1–2 Myrs between emplacement of SDRs at the Møre and Vøring volcanic margins. Initial formation of oceanic crust nucleated west of the Lofoten and north of the Norway Basin in the Greenland Sea and propagated south towards the Jan Mayen Fracture Zone and the GIFR, respectively.
6. In the Norway Basin, the Eocene oceanic rift initially propagated from north to south towards a rotation pole situated on the GIFR. However, at ~47 Ma (C21r) the spreading direction changed radically as indicated by increased rifting activity in the southern part of the JMMC. At that time fan-shaped rifting and/or spreading initiated between East Greenland and the southern Aegir Ridge, while between the Lofoten-Vesterålen and NE Greenland margins near orthogonal spreading developed.
7. Progressive transition from slow to ultra-slow spreading along the Aegir Ridge in the Norway Basin arose after magnetic chron C10, the last chron currently confirmed in the Norway Basin. Complete separation of the JMMC from Greenland and the full opening of a new oceanic basin along the Kolbeinsey Ridge likely took place shortly before 22.5 Ma (chron 6B).
8. We conclude that transition from magmatic rift segments to mature seafloor spreading was a gradual process in the NGS. The early spreading system involved initial propagating segments separated by continental buffers and/or lithospheric barriers in different parts of the NGS (e.g. the Vøring Marginal High, the Faroe Plateau and part of the GIFR). The presence of such rheological heterogenities may explain the punctuated style of breakup and may have controlled the rift/drift overlap apparently required along the GIFR to explain the JMMC formation.

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LIST OF FIGURES

Figure 1: Global relief and seismicity of the Norwegian-Greenland Sea (NGS). 1 arc-minute global relief model from ETOPO 1, Amante et al. (2009). Red circles indicate earthquakes at the plate boundaries (USGS seismicity data from 1995–2018; https://earthquake.usgs.gov). The black and white contours represent the outline of the Inner and Outer SDRS after Berndt et al. (2001), Abdelmalak et al. (2016b) and Geissler et al. (2016). The yellow circle and green circles show respectively the ODP and DSDP sites. AR: Aegir Ridge; BB: Bjørnøya Basin; BK: Blosseville Kyst; EGR: East Greenland Ridge; EJMFZ: East Jan Mayen Fracture Zone; FB: Froan Basin: FIR: Faroe-Iceland Ridge; FR: Fugløy Ridge; FSB: Faroes-Shetland Basin; GE: Greenland Escarpment; GIR: Greenland-Iceland Ridge; GFZ: Greenland Fracture Zone; GØ: Geographic Ø; HaB: Hammerfest Basin; HeB: Helgeland Basin; HB: Hornelen Basin; HR: Hovgård Ridge; JL: Jameson Land; JMB: Jan Mayen Basin; JMI: Jan Mayen Island; JMMC: Jan Mayen Microplate Complex; JMR: Jan Mayen Ridge; KP: Koldewey Platform; KnR: Knipovitch Ridge; KoR: Kolbeinsey Ridge; MR; Mohns's Ridge; MMP: Møre Marginal Plateau; MTFC: Møre-Trøndelag Fault Complex; NB: Nordkapp Basin; RR: Reykjanes Ridge; SFZ: Senja Fracture Zone; SRC: Southern Ridge Complex; TB: Thetis Basin; TP: Trøndelag Platform; TØ: Traill Ø; VB:Vestfjorden Basin; VS: Vøring Spur; WB: Westwind Basin; WJMFZ: West Jan Mayen Fracture Zone.

Figure 2: Regional tectonic map of the NGS and location of the main Paleozoic to Cenozoic rift zones and sedimentary basins. Outlines of Cenozoic volcanism and SDRs after Berndt et al. (2001), Abdelmalak et al. (2016b), Geissler et al. (2016) and Horni et al. (2017). Salt diapirs and domes after Maystrenko et al. (2017a), Rowan et al. (2017) and Gernigon et al. (2018). Onshore geology modified after Bouysse (2014). Rift zones across Iceland after Einarsson et al. (2008). The location of example geological cross-sections shown in Figure 5 (Sections 1–16) and Figure 11 (sections 17–18) are indicated. Abbreviations as in Figure 1.

Figure 3: Free-air gravity anomalies of the NGS. The grid has a resolution of 1×1 km and is based on the global gravity field model DTU12 (Andersen et al. 2012). Abbreviations as in Figure 1. The location of refraction profiles A-O, shown in Figure 5 are indicated. Abbreviations, borehole and SDR outline references as in Figure 1.

Figure 4: Magnetic total field compilation of the NGS. Thick bold letters (C24-21) indicate the main magnetic anomaly chrons. The grid has a resolution of 500×500 m and includes the most recent high-resolution aeromagnetic surveys acquired by the Geological Survey of Norway between 2009 and 2014 (Details can be found in Olesen et al. (2010) and Gernigon et al. (2009, 2012, 2015). Background dataset (transparent) is the World Digital Magnetic Anomaly Map compilation from Dyment et al. (2015). Abbreviations and SDRs outline references as in Figure 1.

Figure 5: Example geological cross-sections of the conjugate volcanic rifted margins of the Norwegian-Greenland Sea. Mid-Norwegian margin sections modified after Gernigon et al. (2002); NE Greenland sections modified after Hamann et al. (2005) and Tsikalas et al. (2002); Faroe-Shetland profiles from Stoker (2016). COB: continent-ocean 'boundary'; TR: T-Reflection; SDR: Seaward dipping reflectors. Profiles (1–16) locations are shown in Figure 2.

Figure 6: Set of seismic refraction lines (A-O) across the Norwegian-Greenland Sea. SDR: Seaward dipping reflectors. Profile locations are shown in Figure 3.

Figure 7: Basin outline of the mid-Norwegian margin and configuration of the early magnetic anomaly chrons. The Base Cretaceous unconformity (BCU) highlights the structural configuration of the rift system. The seafloor spreading (gradual from C24n3n to C23n1) crosscuts obliquely the pre-existing Paleozoic/Mesozoic basins and both Inner and Outer SDRs system. Note that in the northern part of the Møre VPM, local Inner and Outer SDRs (as revealed by seismic data) are located landward of C24n3n and 24r magnetic chrons formed during the early stage of the oceanic accretion. New magnetic data and chrons interpretation after Gernigon et al. (2015). AR: Aegir Ridge; DT: Dønna Terrace; EJMFZ: East Jan Mayen Fracture Zone; FB: Froan Basin; FH: Frøya High; GH: Grip High; HB: Helgeland Basin; HG: Hel Graben; HT: Halten Terrace; JMMC: Jan Mayen Microplate Complex; MR: Mohn's Ridge; NGR: North Gjallar Ridge; NH: Nyk High; RB: Ribban Basin; RH: Røst High; RR: Rån Ridge; SGR: South Gjallar Ridge; UH: Utgard High; VB: Vestfjorden Basin; VH: Vigra High; VMH: Vøring Marginal High; VS: Vøring Syncline.

Figure 8: Regional sections in the central part of the mid-Norwegian margin. Section locations shown in Figures 4 and 7. South of the Jan Mayen Fracture Zone (section b), the first seismic evidence of seafloor spreading is identified at around C24r. COB: continent-ocean 'boundary'; HVLC: High-Velocity Lower Crust; TR: T Reflection; SDR: Seaward dipping reflectors.

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Figure 10: Example of seismic line (in two-way travel-time) from the outer Vøring Basin to the Vøring Marginal High (location shown in Figures 4 and 7). The line (a) traverses structures of the oceanic domain in the west, a major SDR complex in the centre and pre-, syn-, and post-rift successions in the east. At the edge of the outer flexure zone, the T-Reflection coincides with the top of the high-velocity lower crust (Vp > 7.0 km/s). In the eastern portion of the line (see enlarged section (b) in the lower panel), the upper Cretaceous faulted blocks of the North

Gjallar Ridge mark the onset of Late Cretaceous-Paleocene extension. The onlapping complex to the east coincides with the pinch-out of the Turonian(?)-Paleocene sedimentary successions observed at the western flexure zone of the regional sag basin. Frequently, the Campanian-Maastrichtian faulted blocks close to the basalt are truncated by the Base Tertiary Unconformity (BTU) which is itself onlapped by a younger sequence of Paleocene-Early Eocene sediments. From Maastrichtian to Early Eocene time, the deformation migrated seawards and focussed towards the growing magmatic province. A thin onlapping Early Eocene sequence (in purple) along this section coincides in time with the period of drastic magmatism including Inner SDRs formation and thinning observed further west and constrained by the ODP well 642 (e.g. Eldholm et al. 1989; Abdelmalak et al. 2016a). The ODP well 642 also drilled the transition between the Upper and Lower Series volcanics (US and LS) of the Vøring volcanic plateau. Recent re-evaluation of the well revealed that the transition between the andesite and the MORB of the overlying Inner SDRs (Horizon "K") is dated around 57–58 Ma. In time, it coincides with the C25-C24r magnetic transition (e.g. Abdelmalak et al. 2016a). The blue curve represents the total magnetic field (MagTF) along the profile.

Figure 11: Seismic section across the Møre volcanic margin. This section illustrates a volcanostratigraphic sequences emplaced during the onset of breakup. The thin Inner Flows represent an aggradational bottom set, consisting of volcaniclastic sediments mixed with pillowed basalts and lava flows (Berndt et al. 2001; Planke et al. 2000) Then, flood basalt volcanism led rapidly to effusive, subaerial volcanism, infilling the pre-existing volcanic basins, and accompanied by sill and dyke intrusions. Where the Landward Flows reached the shoreline, they fragmented in contact with water leading to foreset bedded Lava Delta development (Abdelmalak et al. 2016b). When the Inner SDRs formed and subside, eventually tuffs and (subaerial) surtseyan material erupted in shallow-water conditions and formed mound-shaped features, the so-called Outer Highs (Planke et al. 2000). Section location is shown in Figures 4 and 7. The blue curve represents the total magnetic field (MagTF) along the profile.

Figure 12: Example of Inner and Outer SDRs in the continent-ocean transition of the conjugate NE Greenland Margin (BGR data). Section location shown in Figure 4.

Figure 13: Tectonic and chronostratigraphic chart of the Norwegian-Greenland Sea spreading evolution correlated with the main Cenozoic stratigraphic and magmato-tectonic events discussed in this paper. Tectono-magmatic events in the Norwegian-Greenland Sea modified after Gernigon et al. (2012). Main compressive and tectonic phases after Doré et al. (2008) and Stoker (2016). Cenozoic stratigraphy and nomenclature after Stoker (2016); Larsen et al. (1999) and Eidvin et al. (2007). Magmatic rock ages after Tegner et al. (2008). Onset of British Isles exhumation episodes after Holford et al. (2009). Eurekan phases after Piepjohn et al. (2016). Global deep-sea oxygen, carbon isotopes and main climatic events from Zachos et al. (2001). Opening of the Eurasian Basin (C25-C24) after Berglar et al. 2016. COT: Continent-ocean transition; FSB: Faroe-Shetland Basin; GIFR: Greenland-Iceland-Faroe Ridge; NAIP: North Atlantic Igneous Province; NB: Norway Basin; JMMC: Jan Mayen Microplate Complexe; PETM: Paelocene-Eocene Thermal maximum: SDRs: Seawards Dipping Reflectors; TVIC: TraillØ-Vøring Igneous Complex; VS: Vøring Spur.

Figure 14: and Early Eocene plate reconstruction of the NGS showing the onset of breakup at C24n3n (53.98 Ma) (based on new magnetic chrons from Gernigon et al. 2015). Seafloor spreading initiated slightly after C24r at the Lofoten/NE Greenland margin and in the northern part of the Norway Basin. The map also shows the reconstruction of the top crystalline basement depth (as observed at the present day) between Norway and Greenland (compiled

after Maystrencko et al. 2017b, c and Granath et al. 2011). The thick blue line represents the extent and limit of the (distal) High-Velocity Lower Crust (HVLC) on both conjugate margins (after Abdelmalak et al. 2017). The space-overlap problem of the JMMC main elements (present day geometry, undeformed) suggests that the rifting and/or accretion must have affected the proto-JMMC after this breakup phase. Abbreviations as in Figure 1.

Figure 15: Early Eocene plate reconstruction of the NGS at C23n1n (50.6 Ma), when seafloor spreading is almost fully established north of the GIFR. The figure illustrates propagating and segmented oceanic segments, assuming that the crust underlying the SDRs is not fully oceanic. There are typically major overlaps between the present-day outline of the Jan Mayen Microplate Complex (JMMC) elements and Greenland. This may (partly) be explained by significant Paleogene extension that affected the JMMC and the corresponding conjugate areas (see inlay). Abbreviations as in Figure 1.

Figure 16: Plate reconstruction of the NGS at C21n (~47.3 Ma). In the Mid-Late Eocene increasing extension of the southern JMMC may have counter-balanced the growing fan-shaped development of the Norway Basin. This period coincides with the onset of extension and diking between East Greenland and the Proto-JMMC in front of the evolving and northward propagating Reykjanes Ridge (e.g. Larsen et al. 2014). The rift-drift overlap between the progressively dying Aegir Ridge and the newly propagating Reykjanes may indicate the presence of preserved continental crust along the Faroe-Iceland Ridge.

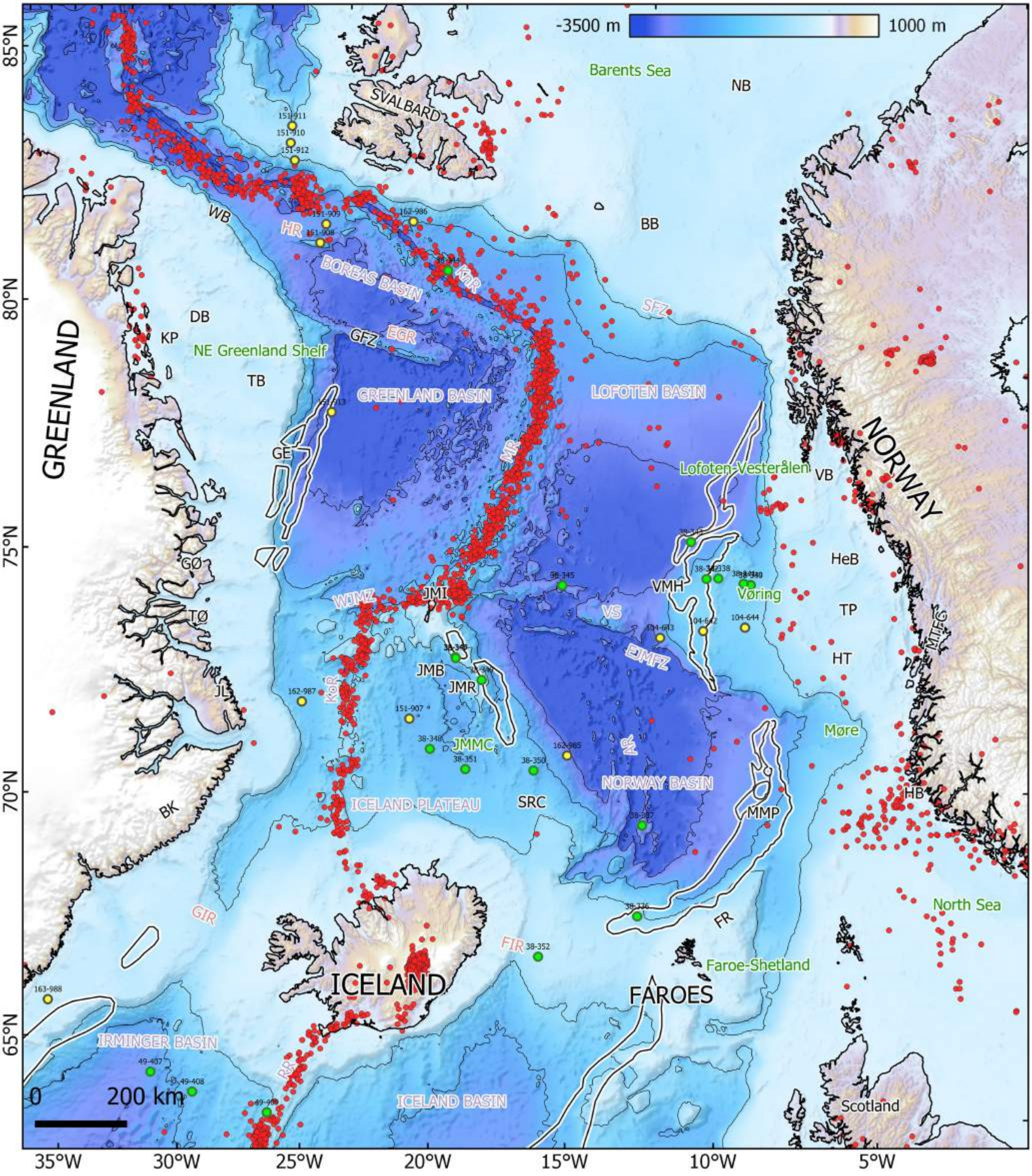
Figure 17: Plate reconstruction of the NGS at C6B (~23 Ma). This period coincides with extinction of the Aegir Ridge in the Norway Basin and the onset of spreading along the nascent Kolbeinsey Ridge between the JMMC and Greenland.

Figure 18: Regional section across the mid-Norwegian margin. Margin evolution was characterized by a first sequence of stretching (S1) and drastic thinning (T1) but the rift slow-down, or most likely failure, in the mid-Cretaceous before a second phase of rifting (S2, T2) initiated about 30–60 Myr later. After strain hardening of lithosphere, a new phase of stretching and thinning focused on the weaker western flank of a preserved marginal plateau before breakup (e.g., L-Block in the nomenclature of Guan et al. submitted).

Figure 19: Conceptual evolution of the volcanic margin formation from the Late Cretaceous rifting phase to the magmatic stage of deformation, SDR development and final lithospheric rupture (e.g. the breakup). COB: continent-ocean 'boundary'; HVLC: High-Velocity Lower Crust; TR: T Reflection; SDR: Seaward dipping reflectors.

Figure 20: Different scenarios to explain the obliquity of the final line or breakup. a) Progressive and asymmetric cooling and strain migration of the rift system. b) Strike-slip deformation of the lithospheric plate. c) Magmatic plumbing of the lithosphere. All these processes could interact.

Figure 21: Conceptual model of volcanic margin segmentation during the punctiform initiation of breakup around C24n3n (Early Ypresian ~53.98 Ma). The model considers the volcanic margins to be diachronic. In this model, the breakup and early opening of the NGS were not instantaneous but propagated in different directions.


-3500 m
1000 m
Barents Sea
NB
SVALBARD
BB
WB
HR
KnR
BOREAS BASIN
SFZ
GREENLAND
DB
KP
NE Greenland Shelf
TB
GFZ
EGR
GREENLAND BASIN
LOFOTEN BASIN
GE
MR
Lofoten-Vesterålen
VB
NORWAY
HeB
GØ
TØ
WJMZ
JMI
VMH
Vøring
VS
TP
EJMFZ
HT
MTFC
JMB
JMR
JL
KoR
Møre
JMMC
AR
NORWAY BASIN
ICELAND PLATEAU
SRC
MMP
HB
BK
North Sea
GIR
FR
FIR
Faroe-Shetland
ICELAND
FAROES
IRMINGER BASIN
RR
ICELAND BASIN
Scotland
0
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75°N
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65°N
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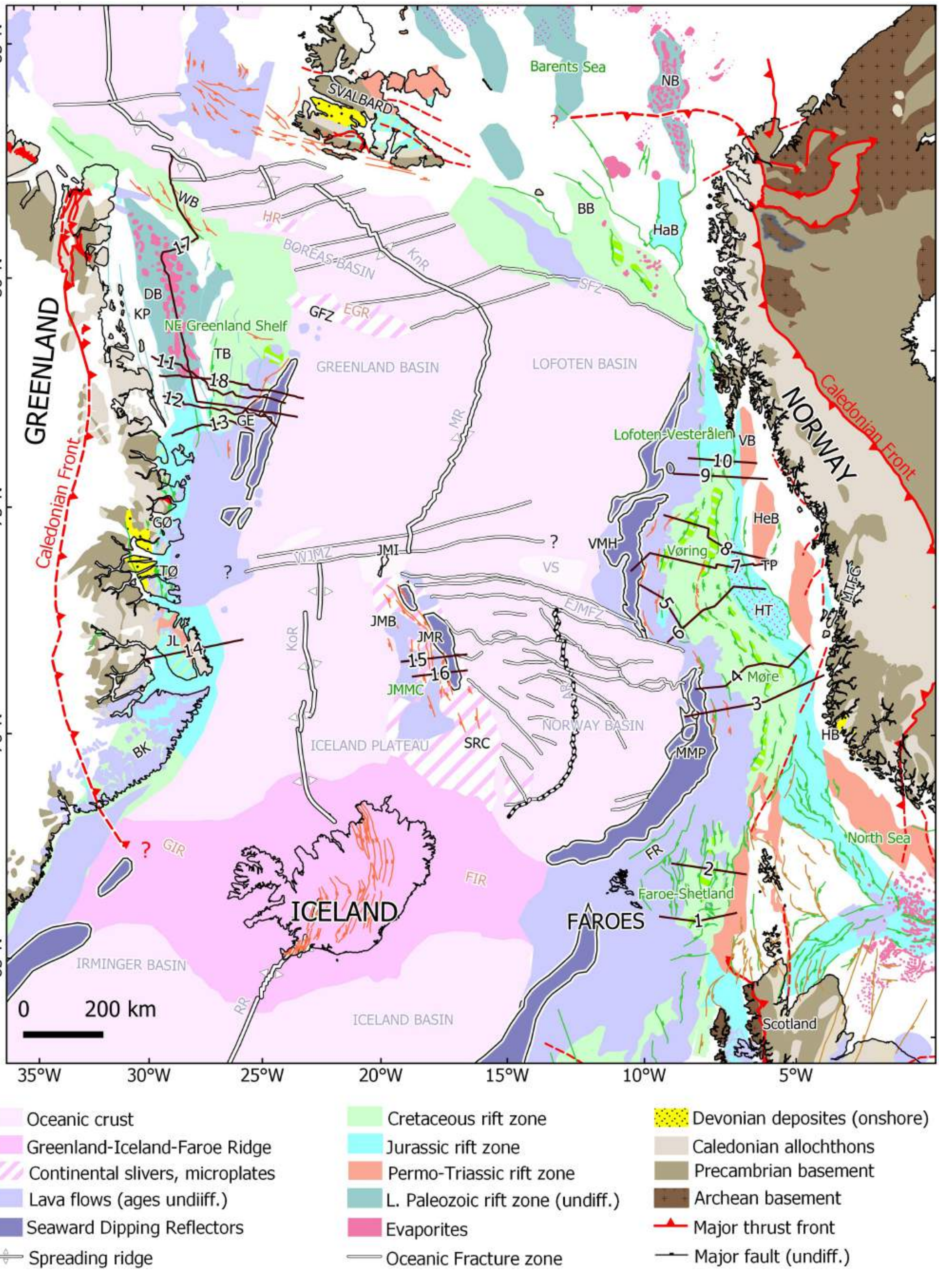
15°W
10°W
5°W

Figure 1



GREENLAND
NORWAY
ICELAND
FAROES
SVALBARD
Scotland
Barents Sea
North Sea
Caledonian Front
NE Greenland Shelf
Lofoten-Vesterålen
Vøring
Møre
Faroe-Shetland
GREENLAND BASIN
LOFOTEN BASIN
BOREAS BASIN
NORWAY BASIN
ICELAND PLATEAU
IRMINGER BASIN
ICELAND BASIN
WJMZ
EJMFZ
SFZ
GFZ
KnR
MR
KoR
RR
JMI
JMB
JMR
JMMC
SRC
VMH
VS
MMP
FR
VB
HeB
TP
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MTFG
NB
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HaB
WB
HR
EGR
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KP
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BK
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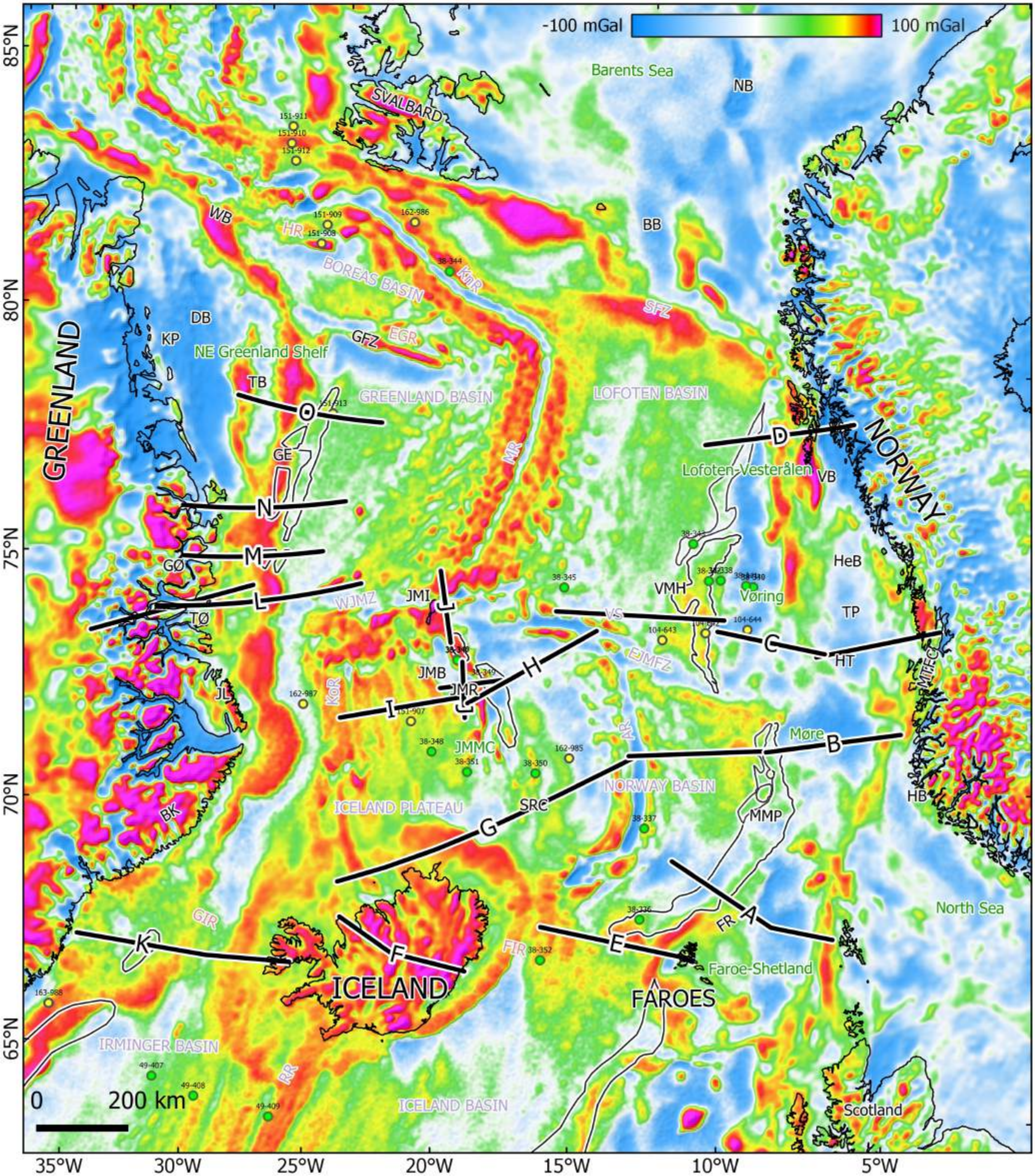
75°N
70°N
65°N
35°W
30°W
25°W
20°W
15°W
10°W
5°W
0
200 km
Oceanic crust
Greenland-Iceland-Faroe Ridge
Continental slivers, microplates
Lava flows (ages undiiff.)
Seaward Dipping Reflectors
Spreading ridge
Cretaceous rift zone
Jurassic rift zone
Permo-Triassic rift zone
L. Paleozoic rift zone (undiff.)
Evaporites
Oceanic Fracture zone
Devonian deposites (onshore)
Caledonian allochthons
Precambrian basement
Archean basement
Major thrust front
Major fault (undiff.)

Figure 2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NE Greenland Shelf
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GREENLAND BASIN
LOFOTEN BASIN
GE
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Lofoten-Vesterålen
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Møre
NORWAY BASIN
ICELAND PLATEAU
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BK
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HB
GIR
FR
North Sea
FIR
ICELAND
Faroe-Shetland
FAROES
IRMINGER BASIN
RR
ICELAND BASIN
Scotland
0
200 km
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80°N
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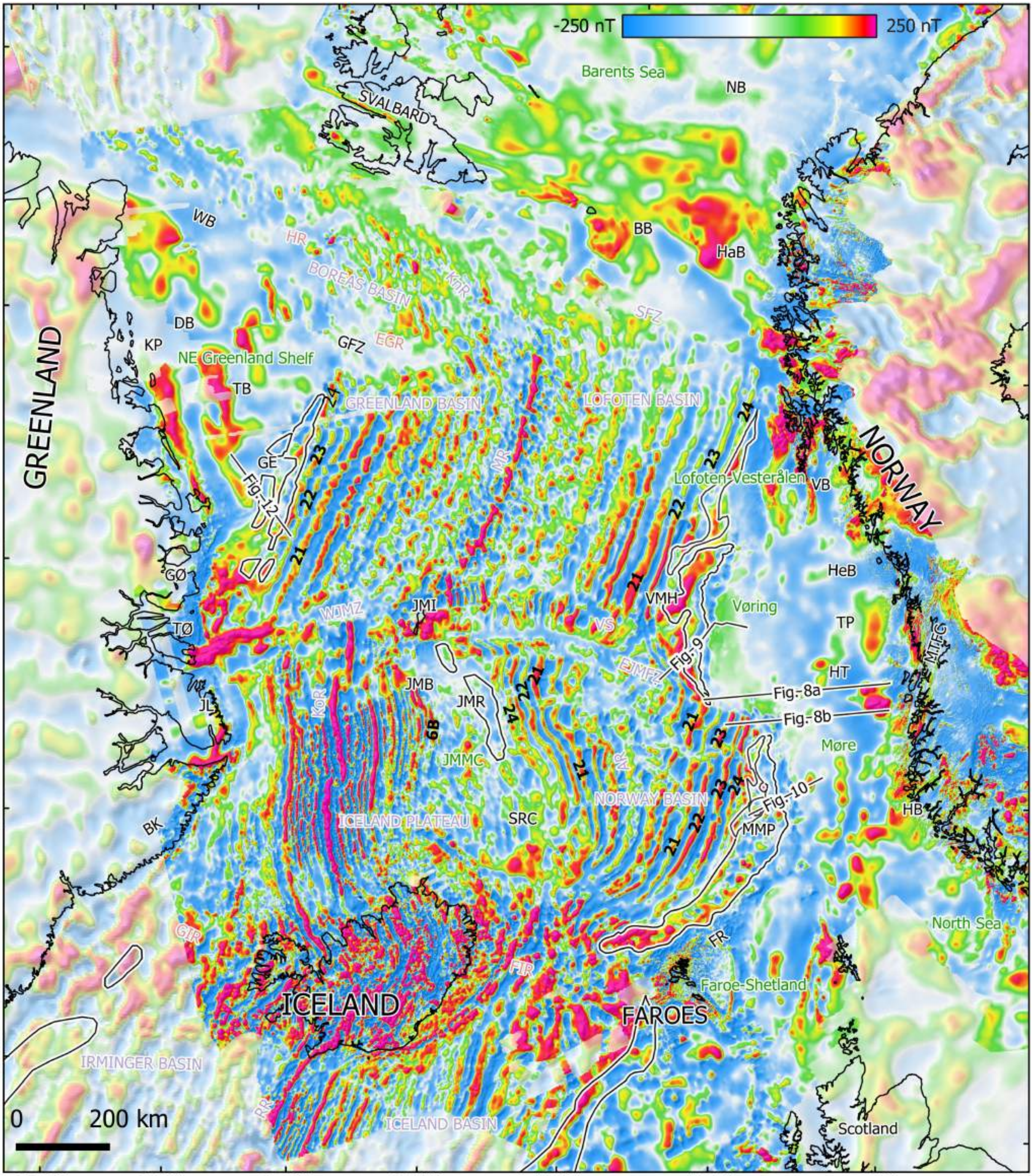
15°W
10°W
5°W

Figure 3



-250 nT
250 nT
Barents Sea
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WB
HR
BOREAS BASIN
KnR
BB
HaB
SFZ
DB
KP
GFZ
EGR
NE Greenland Shelf
GREENLAND
TB
GREENLAND BASIN
LOFOTEN BASIN
GE
Fig. 12
MR
Lofoten-Vesterålen
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NORWAY
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WJMZ
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Vøring
VS
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TP
EJMFZ
Fig. 9
HT
MTFC
JMB
JMR
Fig. 8a
Fig. 8b
JL
KoR
6B
Møre
JMMC
AR
NORWAY BASIN
Fig. 10
ICELAND PLATEAU
SRC
MMP
HB
BK
North Sea
FR
GIR
FIR
Faroe-Shetland
ICELAND
FAROES
IRMINGER BASIN
RR
ICELAND BASIN
Scotland
0
200 km
85°N
80°N
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70°N
65°N
35°W
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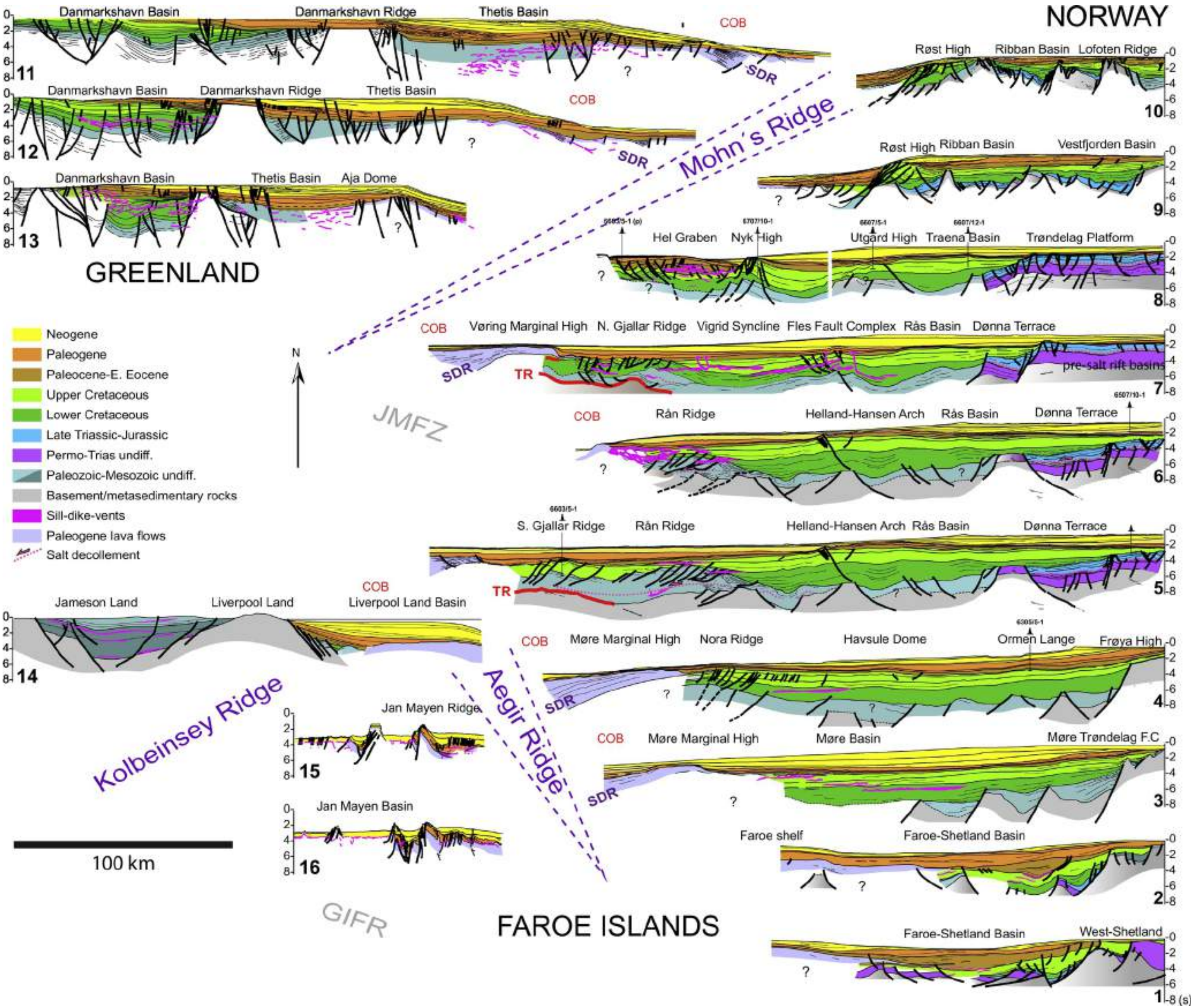
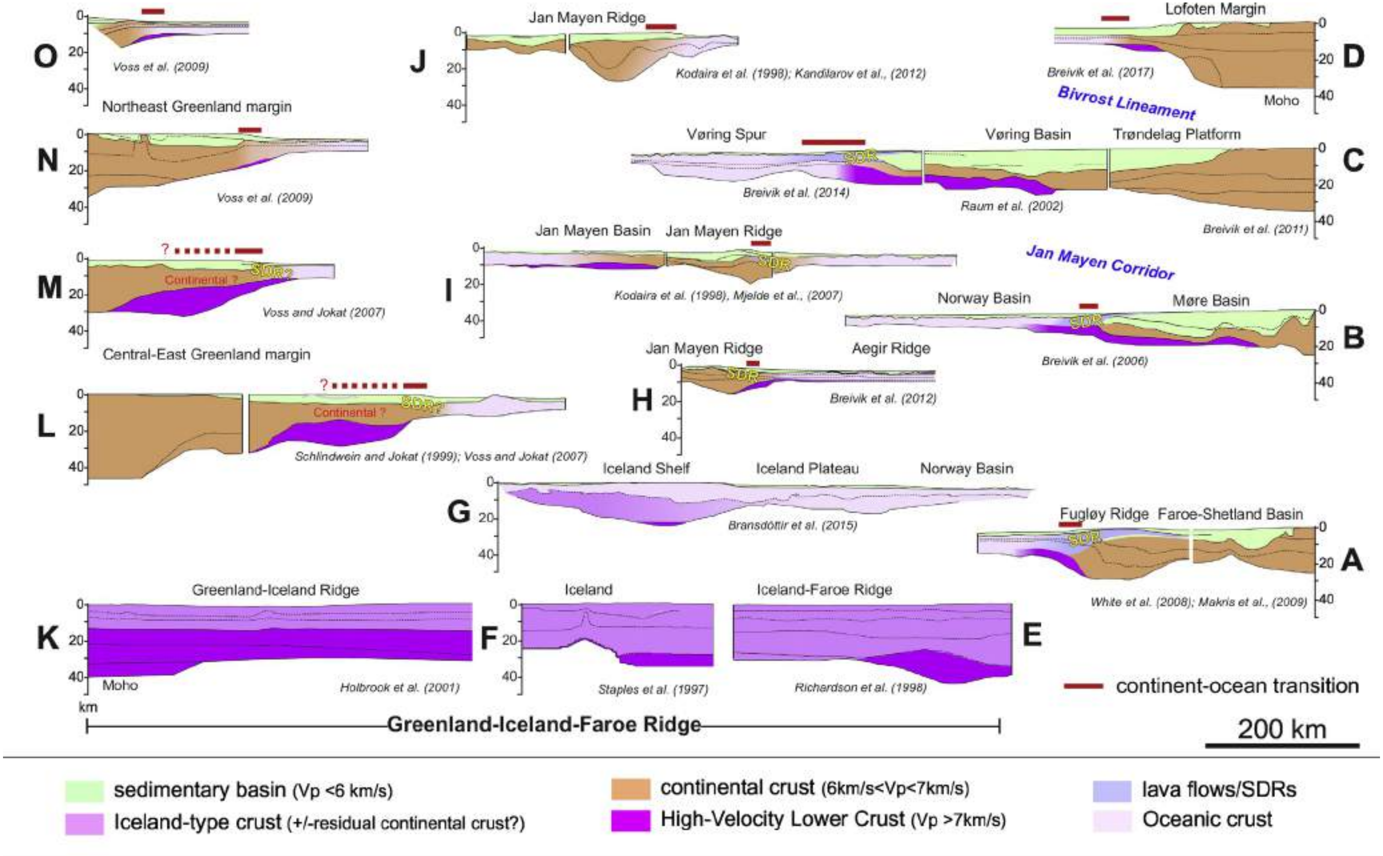
25°W
20°W
15°W
10°W
5°W

Figure 4



Danmarkshavn Basin
Danmarkshavn Ridge
Thetis Basin
COB
SDR
11
12
13
Aja Dome
GREENLAND
NORWAY
Røst High
Ribban Basin
Lofoten Ridge
10
Mohn´s Ridge
Vestfjorden Basin
9
Hel Graben
Nyk High
Utgard High
Traena Basin
Trøndelag Platform
8
Neogene
Paleogene
Paleocene-E. Eocene
Upper Cretaceous
Lower Cretaceous
Late Triassic-Jurassic
Permo-Trias undiff.
Paleozoic-Mesozoic undiff.
Basement/metasedimentary rocks
Sill-dike-vents
Paleogene lava flows
Salt decollement
N
Vøring Marginal High
N. Gjallar Ridge
Vigrid Syncline
Fles Fault Complex
Rås Basin
Dønna Terrace
TR
pre-salt rift basins
7
JMFZ
Rån Ridge
Helland-Hansen Arch
6
S. Gjallar Ridge
5
Jameson Land
Liverpool Land
Liverpool Land Basin
14
Møre Marginal High
Nora Ridge
Havsule Dome
Ormen Lange
Frøya High
4
Kolbeinsey Ridge
Jan Mayen Ridge
15
Aegir Ridge
Møre Basin
Møre Trøndelag F.C
3
Jan Mayen Basin
16
100 km
Faroe shelf
Faroe-Shetland Basin
2
GIFR
FAROE ISLANDS
West-Shetland
1
(s)

Figure 5



Northeast Greenland margin
Voss et al. (2009)
Jan Mayen Ridge
Kodaira et al. (1998); Kandilarov et al., (2012)
Lofoten Margin
Breivik et al. (2017)
Moho
Bivrost Lineament
Vøring Spur
Vøring Basin
Trøndelag Platform
SDR
Breivik et al. (2014)
Raum et al. (2002)
Breivik et al. (2011)
Jan Mayen Basin
Continental ?
SDR ?
Voss and Jokat (2007)
Kodaira et al. (1998), Mjelde et al., (2007)
Jan Mayen Corridor
Norway Basin
Møre Basin
Breivik et al. (2006)
Central-East Greenland margin
Aegir Ridge
Breivik et al. (2012)
Schlindwein and Jokat (1999); Voss and Jokat (2007)
Iceland Shelf
Iceland Plateau
Bransdóttir et al. (2015)
Fugløy Ridge
Faroe-Shetland Basin
White et al. (2008); Makris et al., (2009)
Greenland-Iceland Ridge
Iceland
Iceland-Faroe Ridge
Holbrook et al. (2001)
Staples et al. (1997)
Richardson et al. (1998)
km
continent-ocean transition
Greenland-Iceland-Faroe Ridge
200 km
sedimentary basin (Vp <6 km/s)
Iceland-type crust (+/-residual continental crust?)
continental crust (6km/s<Vp<7km/s)
High-Velocity Lower Crust (Vp >7km/s)
lava flows/SDRs
Oceanic crust

Figure 6

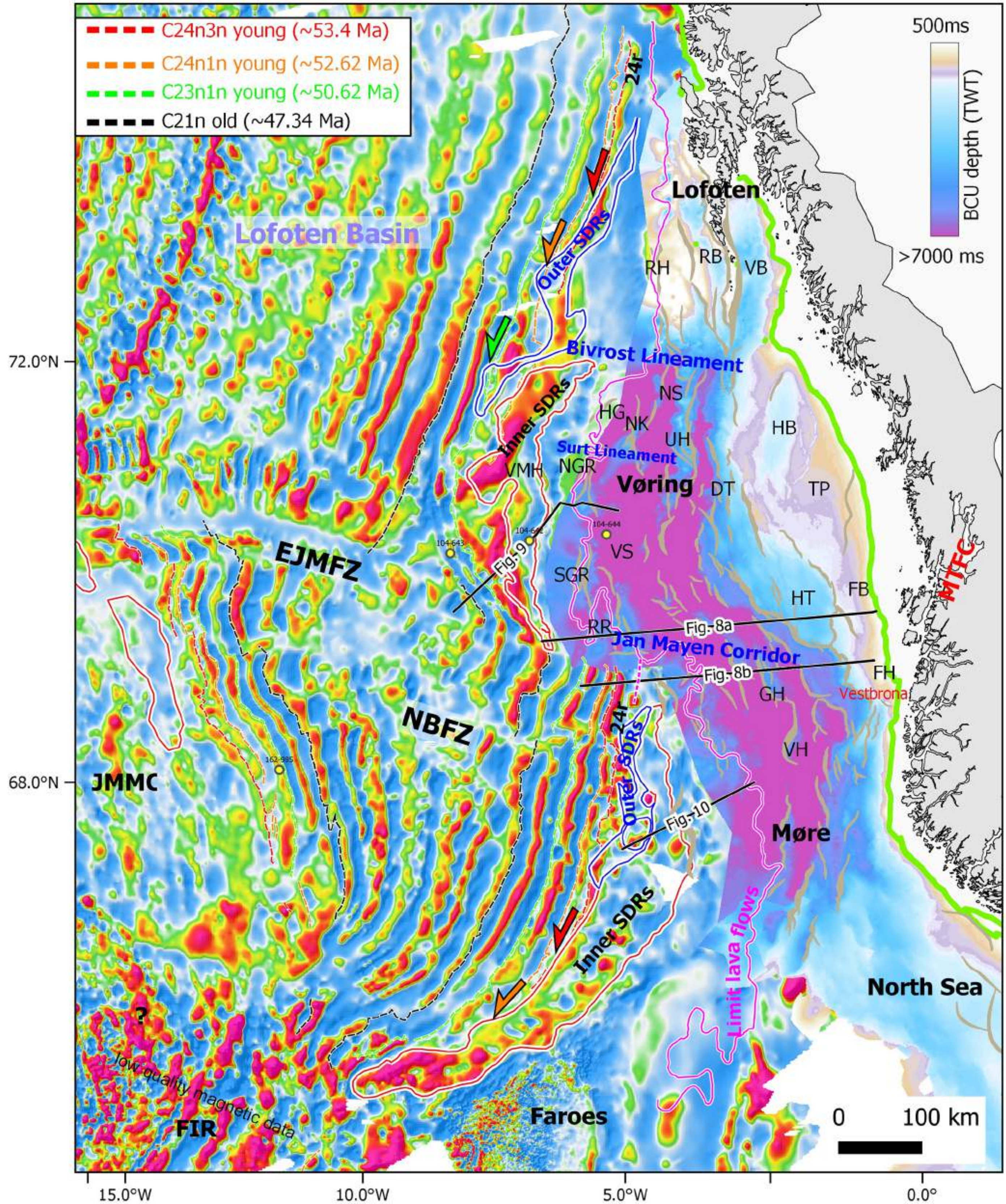

C24n3n young (~53.4 Ma)
C24n1n young (~52.62 Ma)
C23n1n young (~50.62 Ma)
C21n old (~47.34 Ma)
500ms
BCU depth (TWT)
>7000 ms
Lofoten Basin
Lofoten
Outer SDRs
RH
RB
VB
Bivrost Lineament
Inner SDRs
NS
HG
NK
UH
HB
Surt Lineament
VMH
NGR
Vøring
DT
TP
104-644
104-642
104-643
VS
EJMFZ
Fig. 9
SGR
MTFC
HT
FB
RR
Fig. 8a
Jan Mayen Corridor
Fig. 8b
FH
Vestbrona
GH
24r
NBFZ
VH
162-985
JMMC
Fig. 10
Møre
Limit lava flows
North Sea
?
low quality magnetic data
FIR
Faroes
0
100 km
72.0°N
68.0°N
15.0°W
10.0°W
5.0°W
0.0°

Figure 7

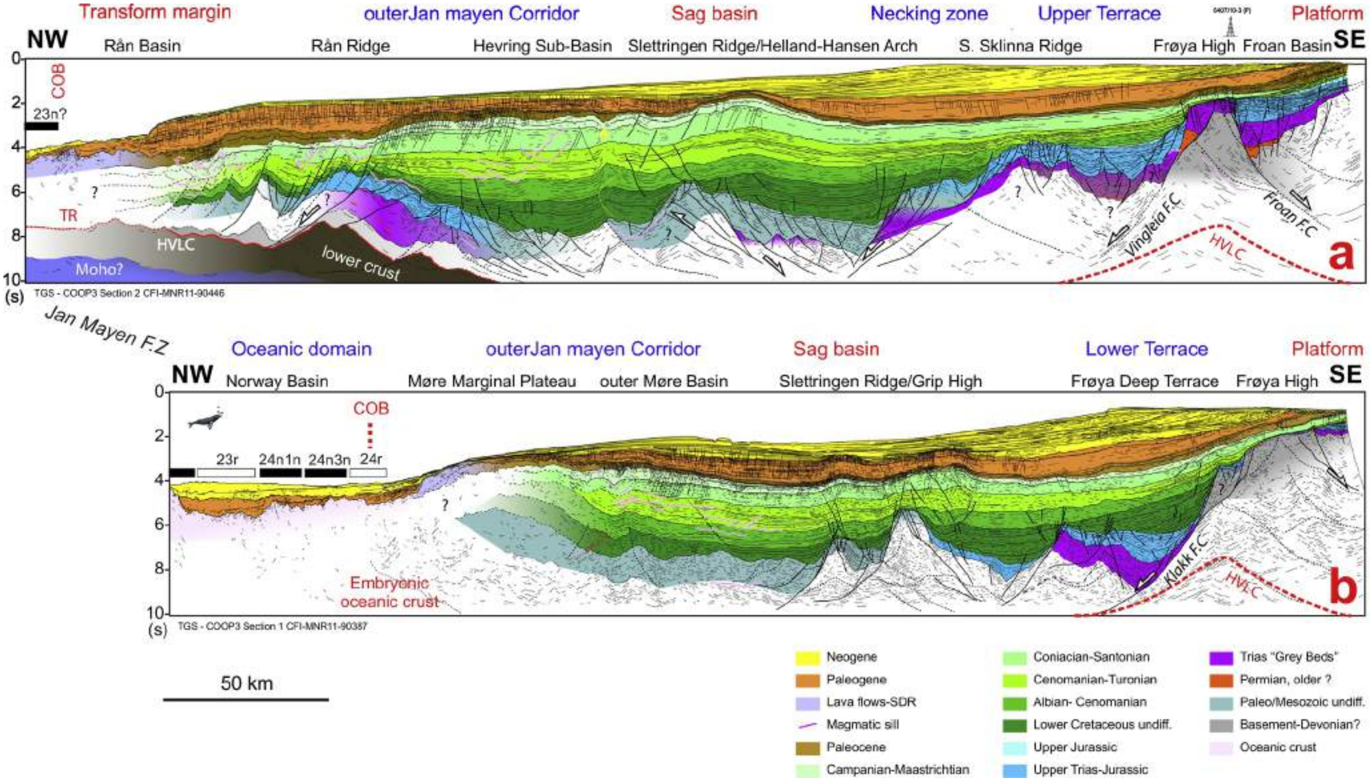
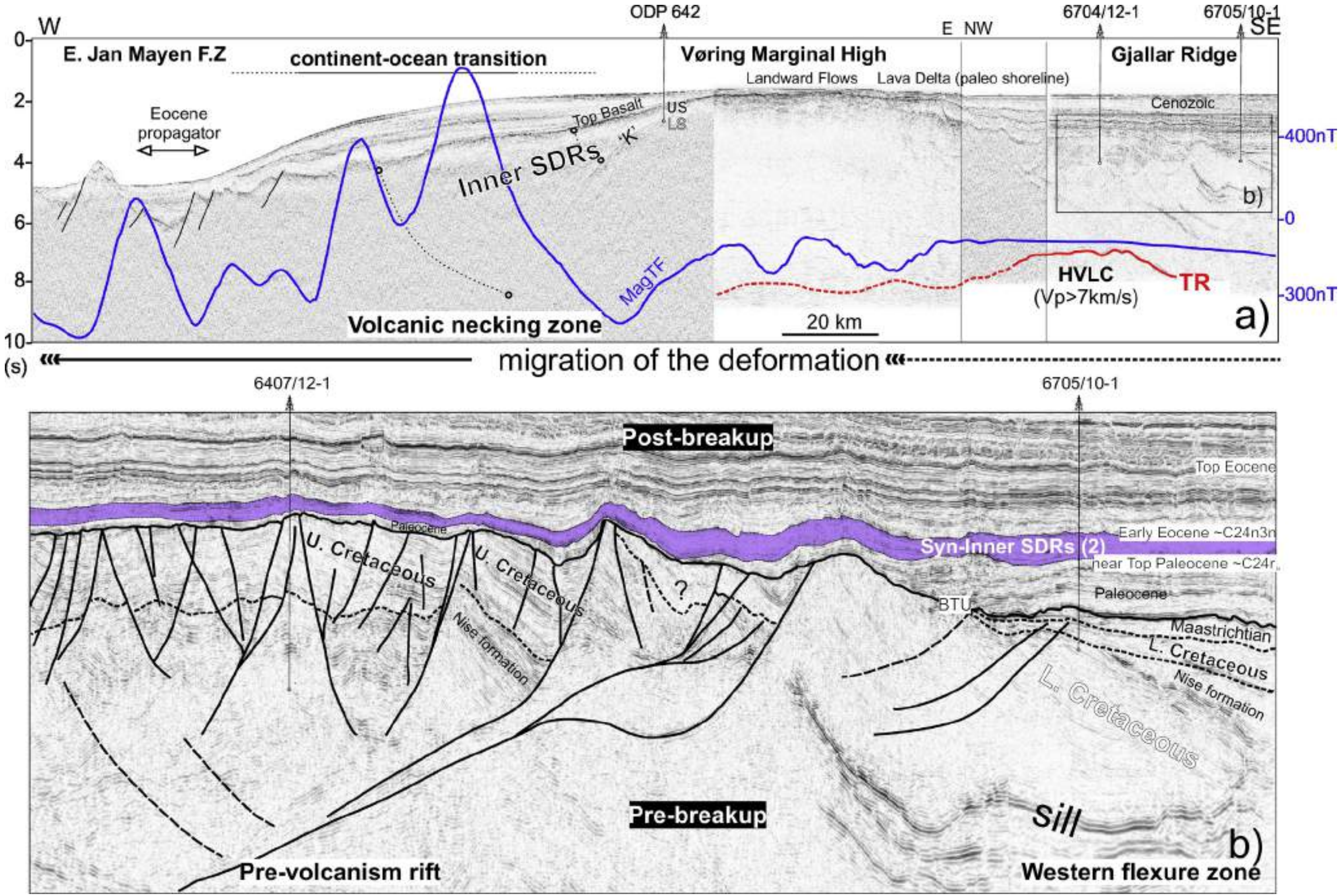

Transform margin
outerJan mayen Corridor
Sag basin
Necking zone
Upper Terrace
Platform
NW
Rån Basin
Rån Ridge
Hevring Sub-Basin
Slettringen Ridge/Helland-Hansen Arch
S. Sklinna Ridge
Frøya High
Froan Basin
SE
COB
23n?
TR
HVLC
Moho?
lower crust
Vingleia F.C
Froan F.C
a
TGS - COOP3 Section 2 CFI-MNR11-90446
Jan Mayen F.Z
Oceanic domain
outerJan mayen Corridor
Sag basin
Lower Terrace
Platform
NW
Norway Basin
Møre Marginal Plateau
outer Møre Basin
Slettringen Ridge/Grip High
Frøya Deep Terrace
Frøya High
SE
COB
23r
24n1n
24n3n
24r
Embryonic oceanic crust
Klakk F.C
HVLC
b
TGS - COOP3 Section 1 CFI-MNR11-90387
50 km
Neogene
Paleogene
Lava flows-SDR
Magmatic sill
Paleocene
Campanian-Maastrichtian
Coniacian-Santonian
Cenomanian-Turonian
Albian- Cenomanian
Lower Cretaceous undiff.
Upper Jurassic
Upper Trias-Jurassic
Trias "Grey Beds"
Permian, older ?
Paleo/Mesozoic undiff.
Basement-Devonian?
Oceanic crust

Figure 8



W
ODP 642
E NW
6704/12-1
6705/10-1
SE
E. Jan Mayen F.Z
continent-ocean transition
Vøring Marginal High
Gjallar Ridge
Landward Flows
Lava Delta (paleo shoreline)
Cenozoic
Eocene propagator
Top Basalt
US
LS
Inner SDRs
'K'
400nT
0
-300nT
b)
MagTF
HVLC
(Vp>7km/s)
TR
Volcanic necking zone
20 km
a)
0
2
4
6
8
10
(s)
migration of the deformation
6407/12-1
6705/10-1
Post-breakup
Top Eocene
Paleocene
Early Eocene ~C24n3n
Syn-Inner SDRs (2)
near Top Paleocene ~C24r
U. Cretaceous
U. Cretaceous
?
Paleocene
BTU
Maastrichtian
L. Cretaceous
Nise formation
Nise formation
L. Cretaceous
Sill
Pre-breakup
b)
Pre-volcanism rift
Western flexure zone

Figure 9

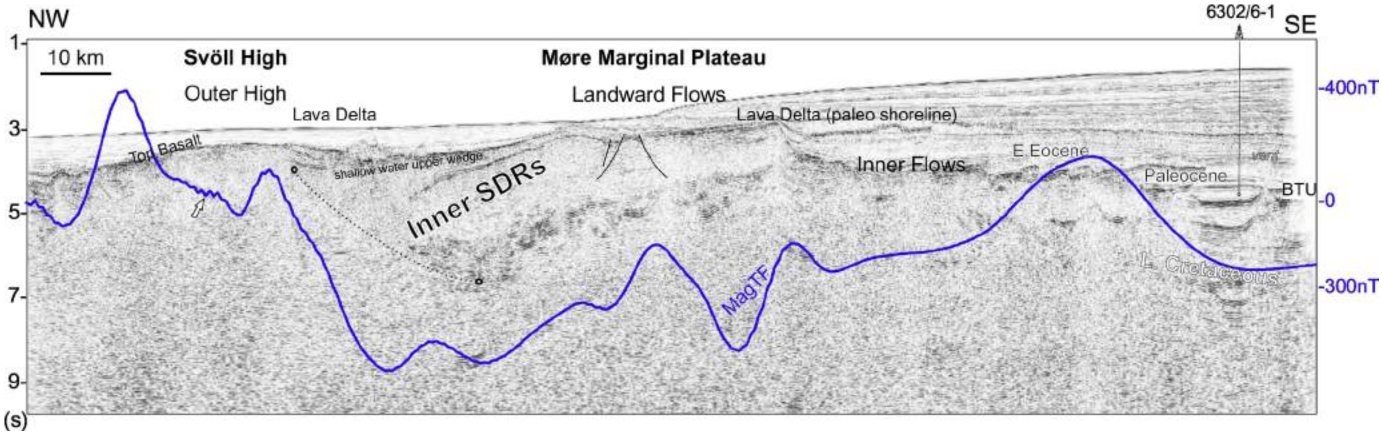

NW
SE
6302/6-1
10 km
Svöll High
Outer High
Møre Marginal Plateau
Landward Flows
Lava Delta
Lava Delta (paleo shoreline)
Top Basalt
shallow water upper wedge
Inner SDRs
Inner Flows
E Eocene
Paleocene
BTU
L. Cretaceous
MagTF
-400nT
-0
-300nT
1
3
5
7
9
(s)

Figure 10

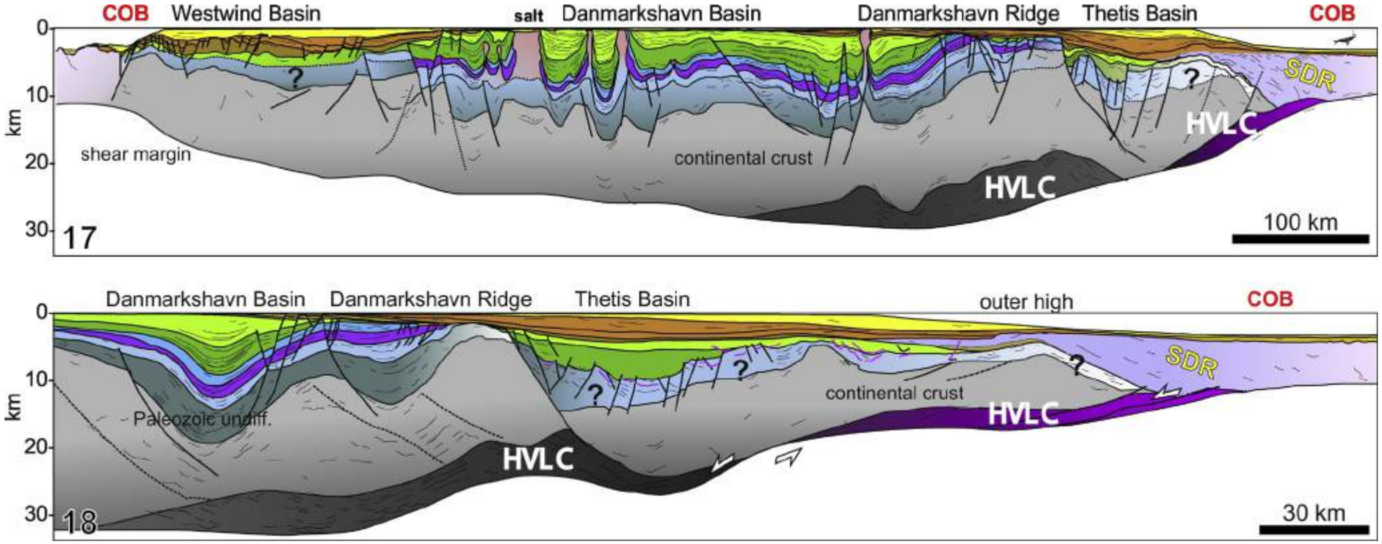
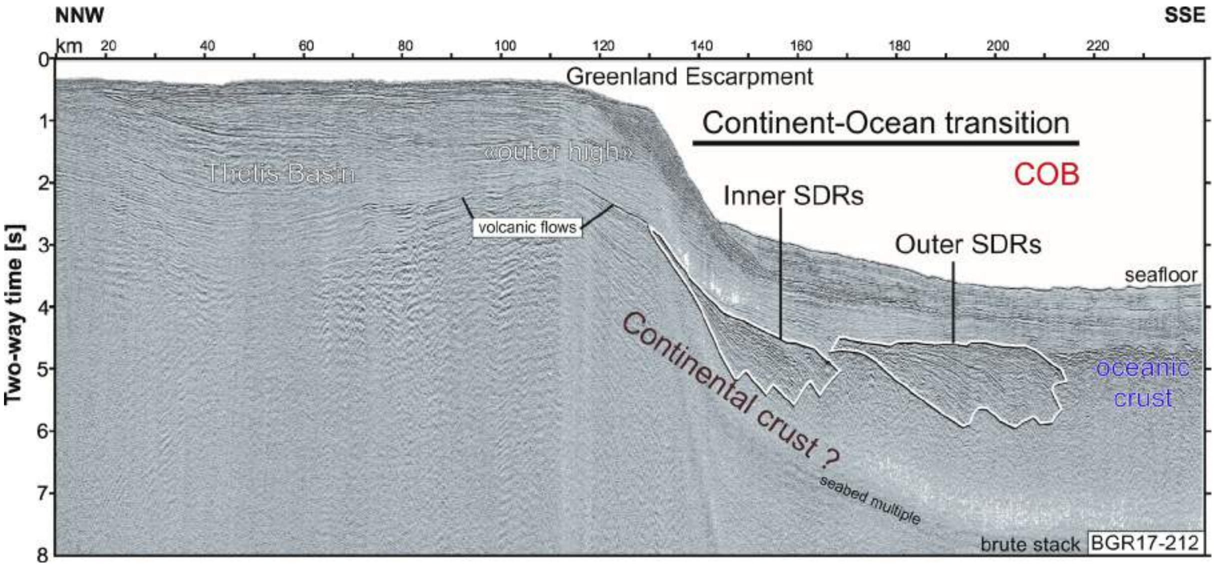

COB
Westwind Basin
salt
Danmarkshavn Basin
Danmarkshavn Ridge
Thetis Basin
COB
SDR
HVLC
shear margin
continental crust
HVLC
km
0
10
20
30
17
100 km
Danmarkshavn Basin
Danmarkshavn Ridge
Thetis Basin
outer high
COB
SDR
continental crust
HVLC
Paleozoic undiff.
HVLC
km
0
10
20
30
18
30 km

Figure 11



NNW
SSE
km
20
40
60
80
100
120
140
160
180
200
220
0
1
2
3
4
5
6
7
8
Two-way time [s]
Greenland Escarpment
Continent-Ocean transition
COB
«outer high»
Thetis Basin
Inner SDRs
volcanic flows
Outer SDRs
seafloor
oceanic crust
Continental crust ?
seabed multiple
brute stack
BGR17-212

Figure 12

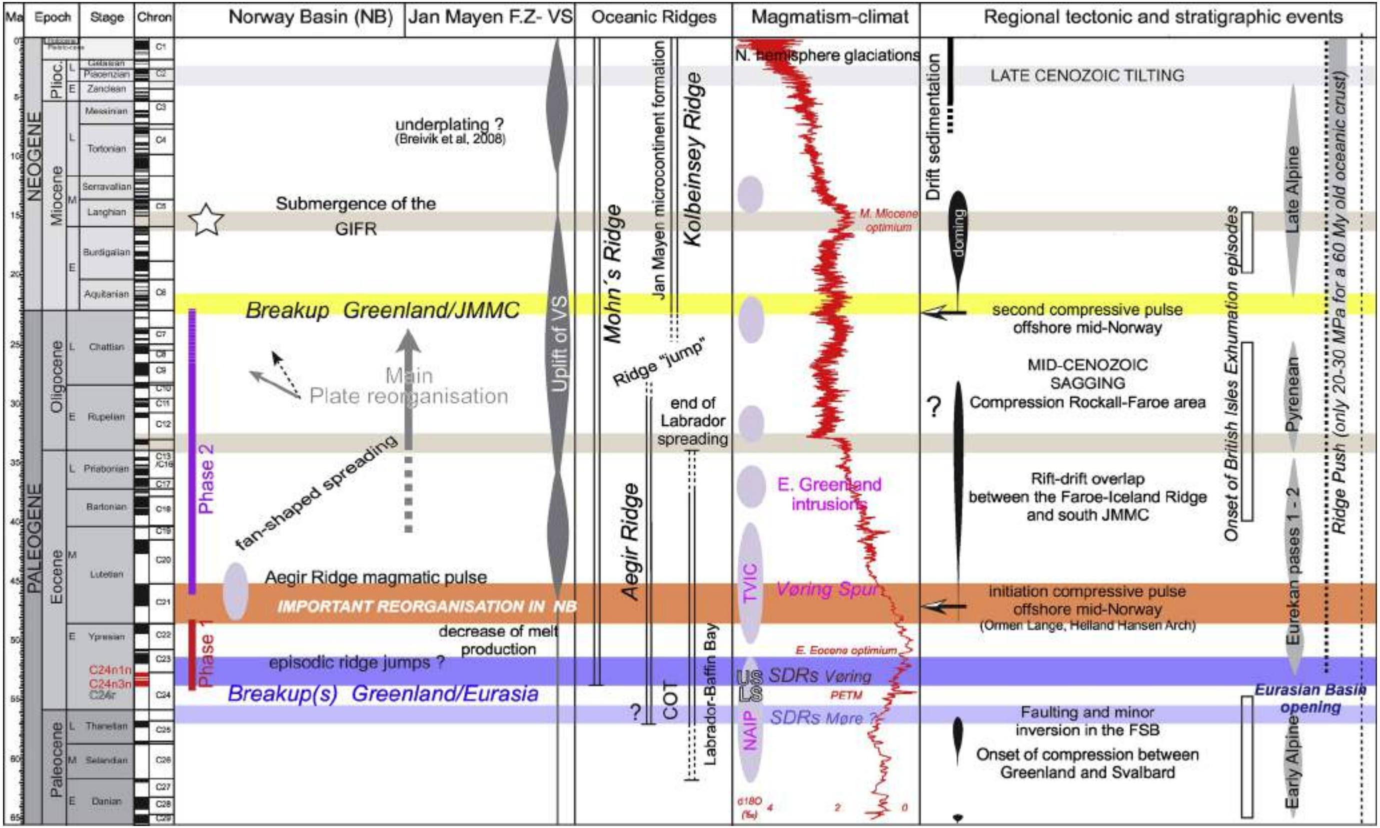
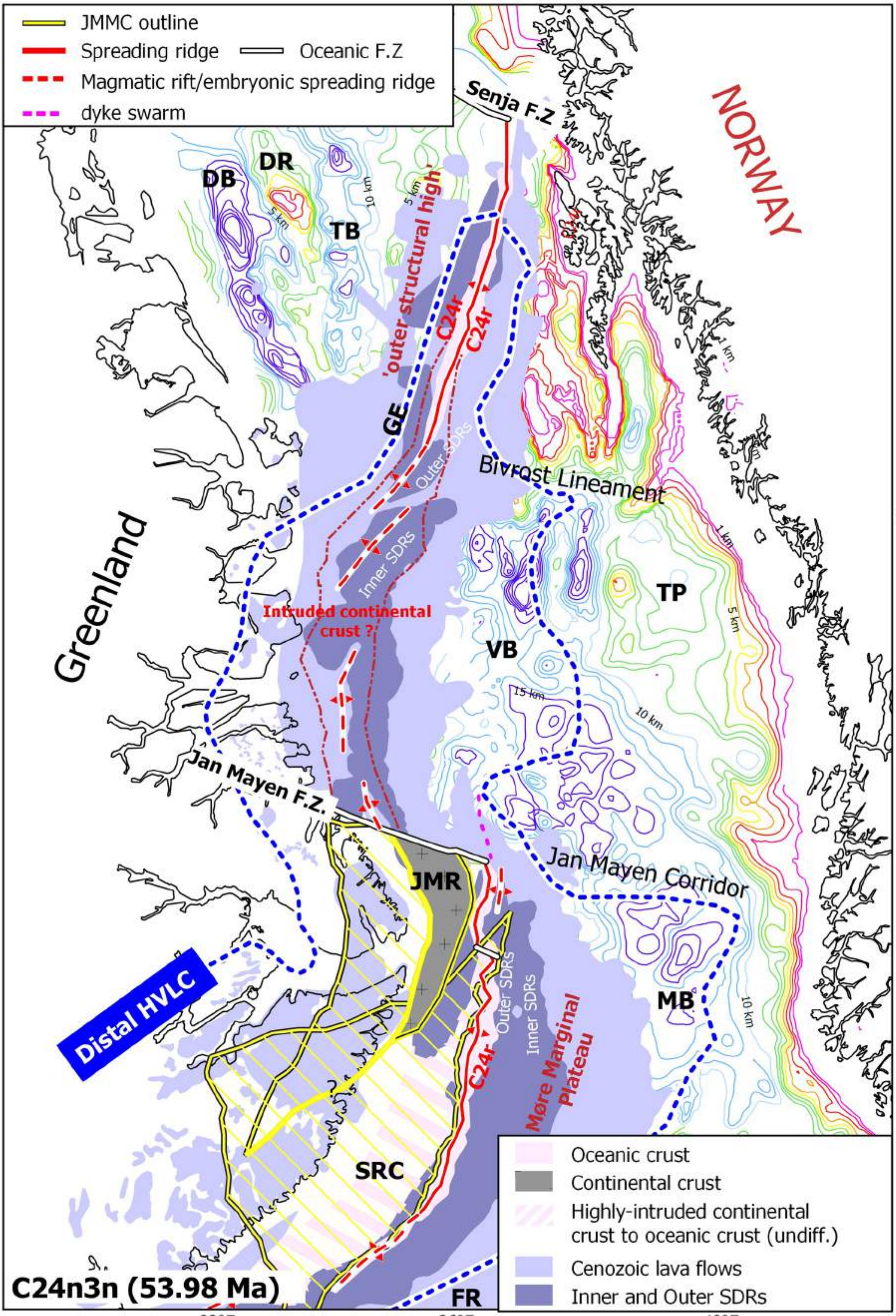

Ma
Epoch
Stage
Chron
Norway Basin (NB)
Jan Mayen F.Z- VS
Oceanic Ridges
Magmatism-climat
Regional tectonic and stratigraphic events
NEOGENE
PALEOGENE
Plioc.
Miocene
Oligocene
Eocene
Paleocene
Gelasian
Piacenzian
Zanclean
Messinian
Tortonian
Serravallian
Langhian
Burdigalian
Aquitanian
Chattian
Rupelian
Priabonian
Bartonian
Lutetian
Ypresian
C24n1n
C24n3n
C24r
Thanetian
Selandian
Danian
underplating ?
(Breivik et al, 2008)
Submergence of the
GIFR
Breakup Greenland/JMMC
Main
Plate reorganisation
fan-shaped spreading
Phase 2
Aegir Ridge magmatic pulse
IMPORTANT REORGANISATION IN NB
Phase 1
decrease of melt
production
episodic ridge jumps ?
Breakup(s) Greenland/Eurasia
Uplift of VS
Mohn's Ridge
Jan Mayen microcontinent formation
Kolbeinsey Ridge
Ridge "jump"
end of
Labrador
spreading
Aegir Ridge
COT
Labrador-Baffin Bay
N. hemisphere glaciations
M. Miocene
optimum
E. Greenland
intrusions
TVIC
Vøring Spur
E. Eocene optimium
SDRs Vøring
PETM
SDRs Møre ?
NAIP
d18O
(‰)
LATE CENOZOIC TILTING
Drift sedimentation
doming
second compressive pulse
offshore mid-Norway
MID-CENOZOIC
SAGGING
Compression Rockall-Faroe area
Rift-drift overlap
between the Faroe-Iceland Ridge
and south JMMC
initiation compressive pulse
offshore mid-Norway
(Ormen Lange, Helland Hansen Arch)
Eurasian Basin
opening
Faulting and minor
inversion in the FSB
Onset of compression between
Greenland and Svalbard
Onset of British Isles Exhumation episodes
Late Alpine
Pyrenean
Eurekan pases 1 - 2
Early Alpine
Ridge Push (only 20-30 MPa for a 60 My old oceanic crust)

Figure 13



JMMC outline
Spreading ridge
Oceanic F.Z
Magmatic rift/embryonic spreading ridge
dyke swarm
Senja F.Z
NORWAY
DB
DR
TB
'outer structural high'
C24r
C24r
GE
Outer SDRs
Bivrost Lineament
Inner SDRs
Greenland
Intruded continental crust ?
TP
VB
15 km
10 km
5 km
1 km
Jan Mayen F.Z.
Jan Mayen Corridor
JMR
Distal HVLC
Outer SDRs
Inner SDRs
Møre Marginal Plateau
MB
C24r
SRC
FR
C24n3n (53.98 Ma)
Oceanic crust
Continental crust
Highly-intruded continental crust to oceanic crust (undiff.)
Cenozoic lava flows
Inner and Outer SDRs
64°N
60°N
32°E
36°E
40°E

Figure 14

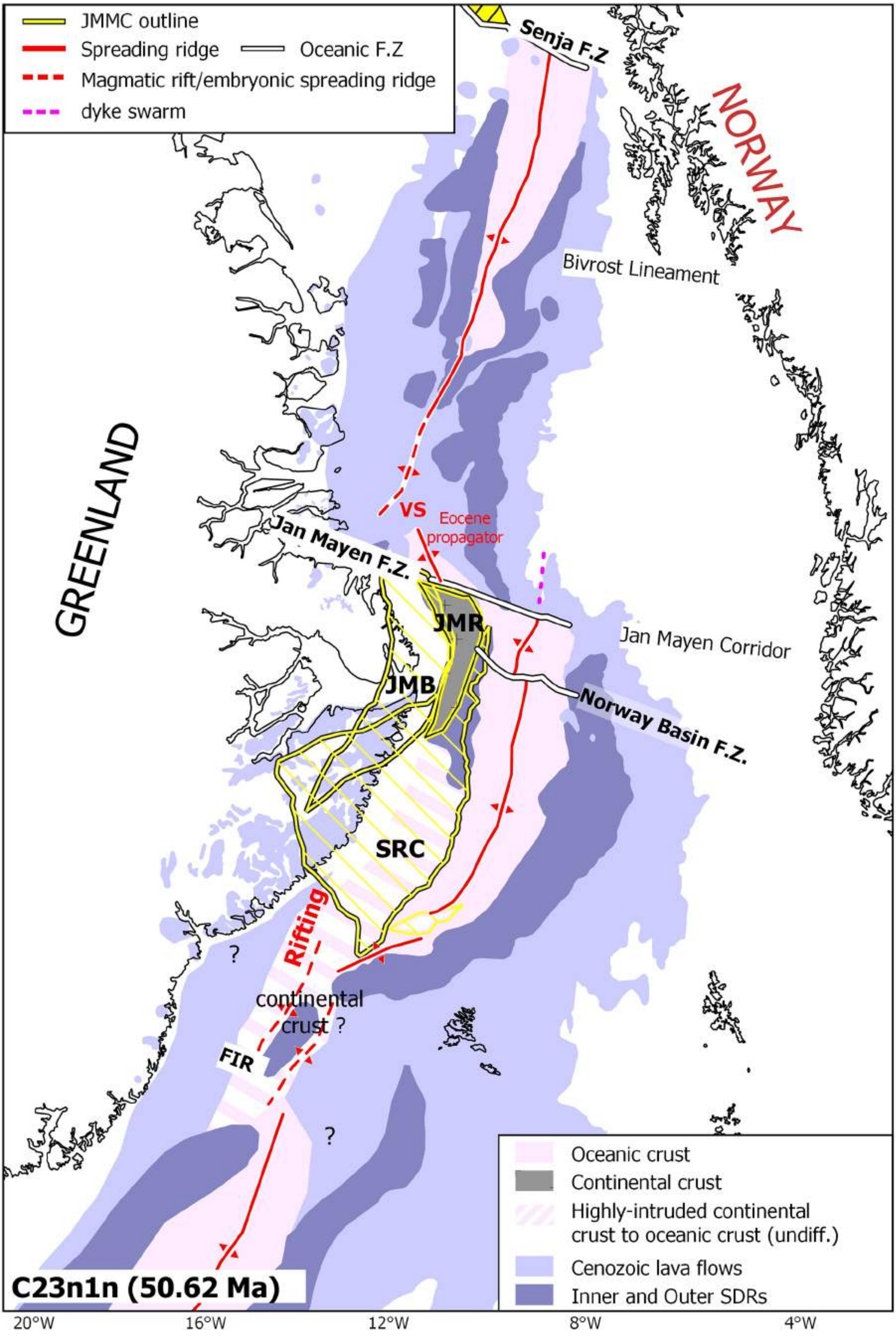

JMMC outline
Spreading ridge
Oceanic F.Z
Magmatic rift/embryonic spreading ridge
dyke swarm
Senja F.Z
NORWAY
Bivrost Lineament
GREENLAND
VS
Eocene propagator
Jan Mayen F.Z.
JMR
JMB
Jan Mayen Corridor
Norway Basin F.Z.
SRC
Rifting
continental crust ?
FIR
?
Oceanic crust
Continental crust
Highly-intruded continental crust to oceanic crust (undiff.)
Cenozoic lava flows
Inner and Outer SDRs
C23n1n (50.62 Ma)
64°N
60°N
56°N
20°W
16°W
12°W
8°W
4°W

Figure 15

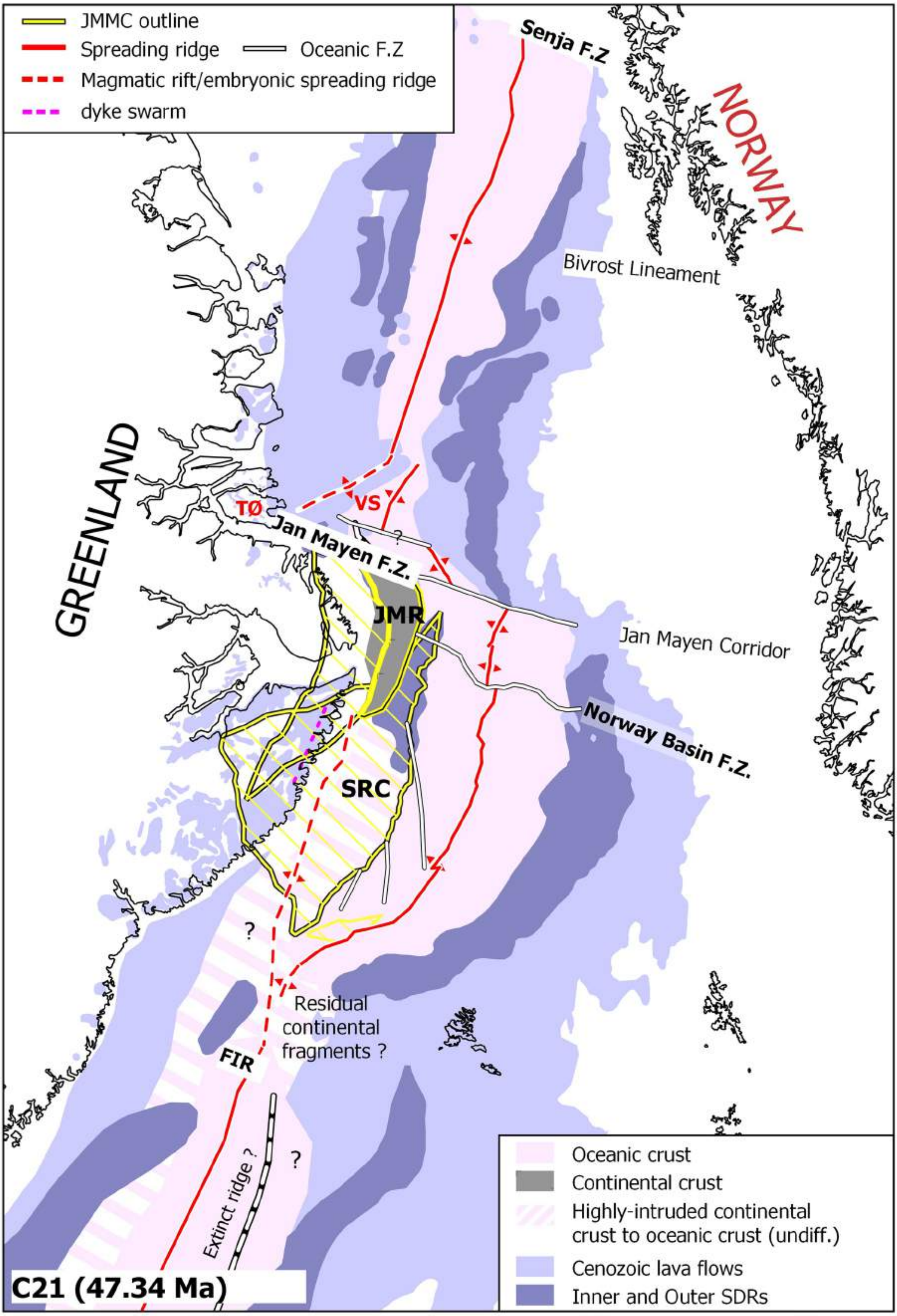
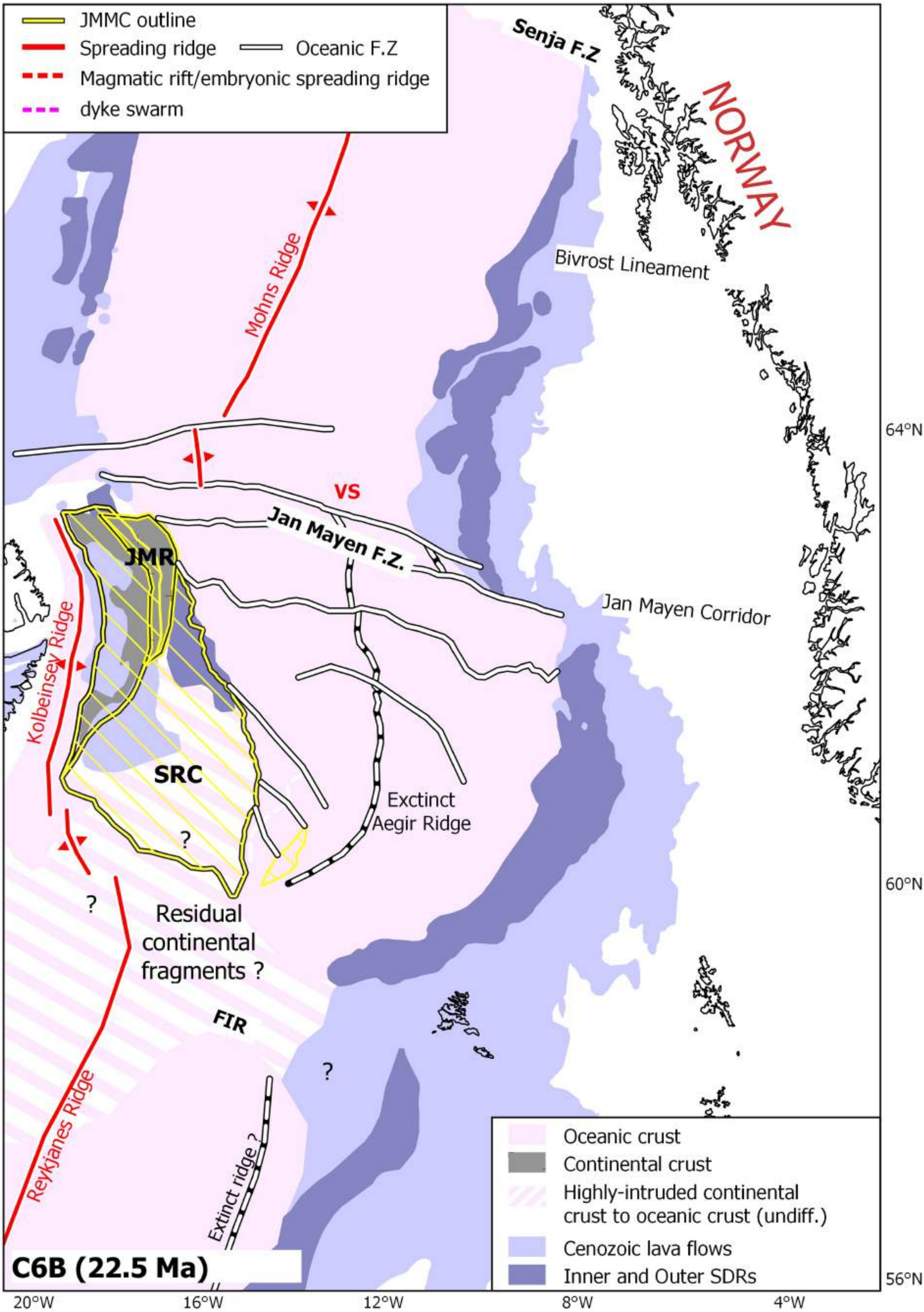

JMMC outline
Spreading ridge
Oceanic F.Z
Magmatic rift/embryonic spreading ridge
dyke swarm
Senja F.Z
NORWAY
Bivrost Lineament
GREENLAND
TØ
VS
Jan Mayen F.Z.
JMR
Jan Mayen Corridor
Norway Basin F.Z.
SRC
Residual continental fragments ?
FIR
Extinct ridge ?
64°N
60°N
56°N
20°W
16°W
12°W
8°W
4°W
Oceanic crust
Continental crust
Highly-intruded continental crust to oceanic crust (undiff.)
Cenozoic lava flows
Inner and Outer SDRs
C21 (47.34 Ma)

Figure 16



JMMC outline
Spreading ridge
Oceanic F.Z
Magmatic rift/embryonic spreading ridge
dyke swarm
Senja F.Z
NORWAY
Bivrost Lineament
Mohns Ridge
VS
Jan Mayen F.Z.
JMR
Jan Mayen Corridor
Kolbeinsey Ridge
SRC
Exctinct
Aegir Ridge
?
Residual
continental
fragments ?
FIR
Reykjanes Ridge
Extinct ridge ?
C6B (22.5 Ma)
Oceanic crust
Continental crust
Highly-intruded continental
crust to oceanic crust (undiff.)
Cenozoic lava flows
Inner and Outer SDRs
64°N
60°N
56°N
20°W
16°W
12°W
8°W
4°W

Figure 17

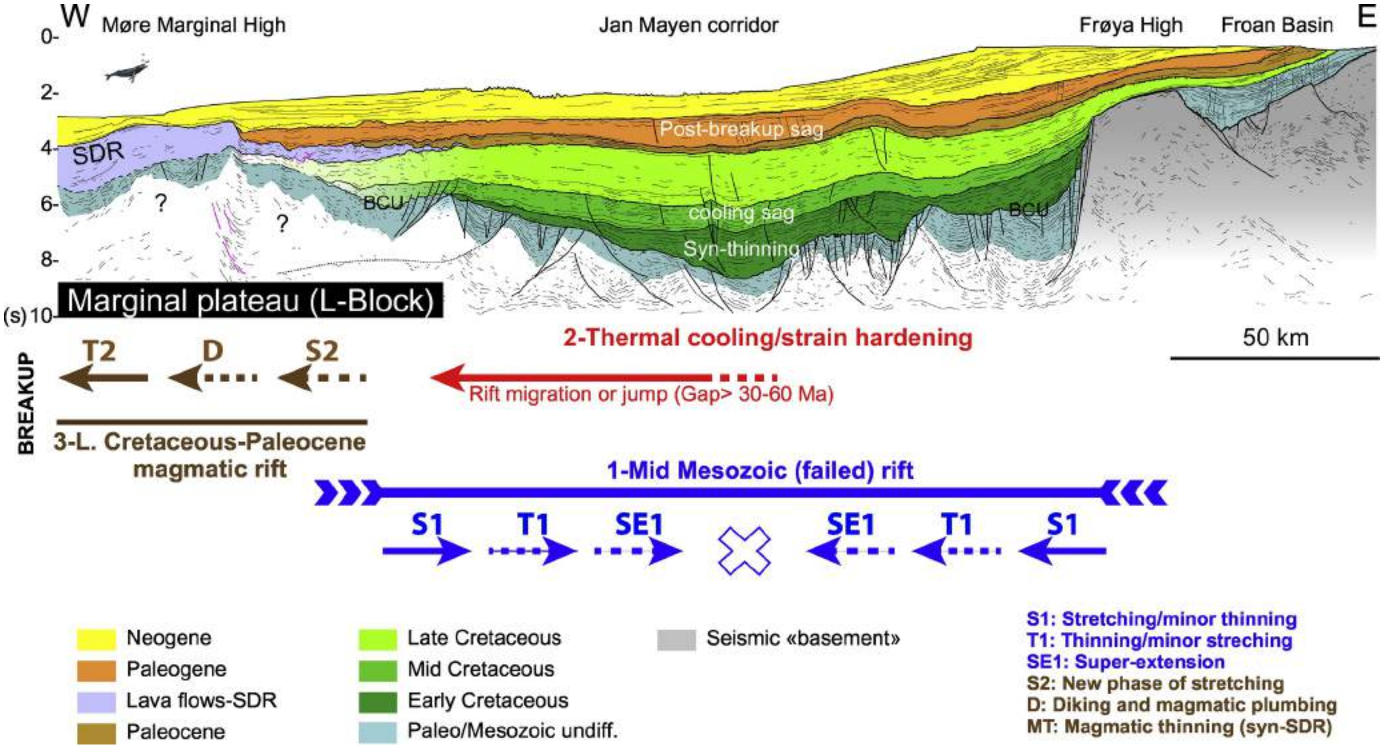

W
Møre Marginal High
Jan Mayen corridor
Frøya High
Froan Basin
E
0-
2-
4-
6-
8-
(s) 10
SDR
Post-breakup sag
BCU
cooling sag
Syn-thinning
BCU
?
?
Marginal plateau (L-Block)
50 km
2-Thermal cooling/strain hardening
Rift migration or jump (Gap> 30-60 Ma)
T2
D
S2
BREAKUP
3-L. Cretaceous-Paleocene magmatic rift
1-Mid Mesozoic (failed) rift
S1
T1
SE1
SE1
T1
S1
Neogene
Paleogene
Lava flows-SDR
Paleocene
Late Cretaceous
Mid Cretaceous
Early Cretaceous
Paleo/Mesozoic undiff.
Seismic «basement»
S1: Stretching/minor thinning
T1: Thinning/minor streching
SE1: Super-extension
S2: New phase of stretching
D: Diking and magmatic plumbing
MT: Magmatic thinning (syn-SDR)

Figure 18

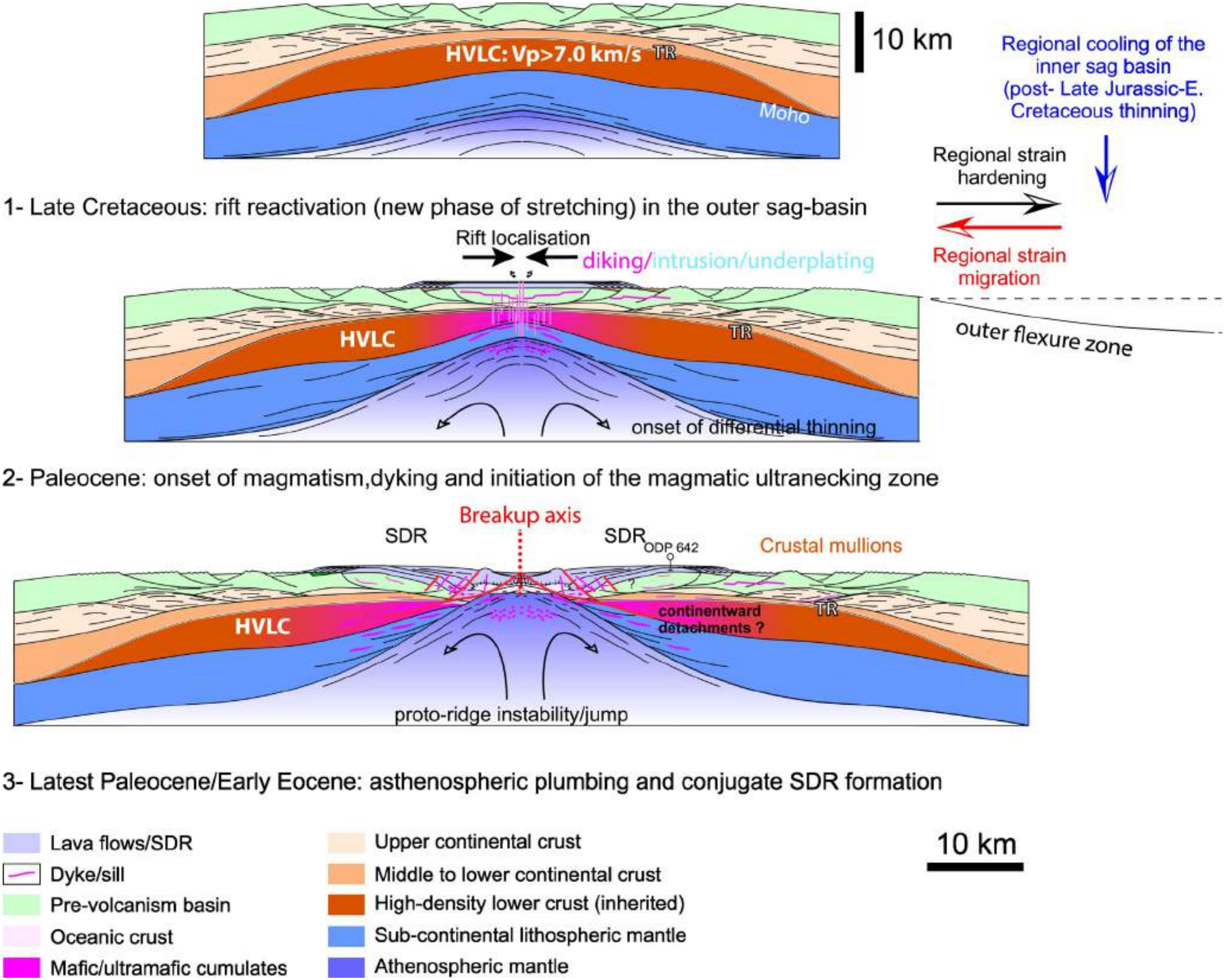

HVLC: Vp>7.0 km/s
TR
Moho
10 km
Regional cooling of the inner sag basin (post- Late Jurassic-E. Cretaceous thinning)
Regional strain hardening
Regional strain migration
1- Late Cretaceous: rift reactivation (new phase of stretching) in the outer sag-basin
Rift localisation
diking/intrusion/underplating
HVLC
TR
outer flexure zone
onset of differential thinning
2- Paleocene: onset of magmatism,dyking and initiation of the magmatic ultranecking zone
Breakup axis
SDR
SDR
ODP 642
Crustal mullions
HVLC
continentward detachments ?
TR
proto-ridge instability/jump
3- Latest Paleocene/Early Eocene: asthenospheric plumbing and conjugate SDR formation
Lava flows/SDR
Dyke/sill
Pre-volcanism basin
Oceanic crust
Mafic/ultramafic cumulates
Upper continental crust
Middle to lower continental crust
High-density lower crust (inherited)
Sub-continental lithospheric mantle
Athenospheric mantle
10 km

Figure 19

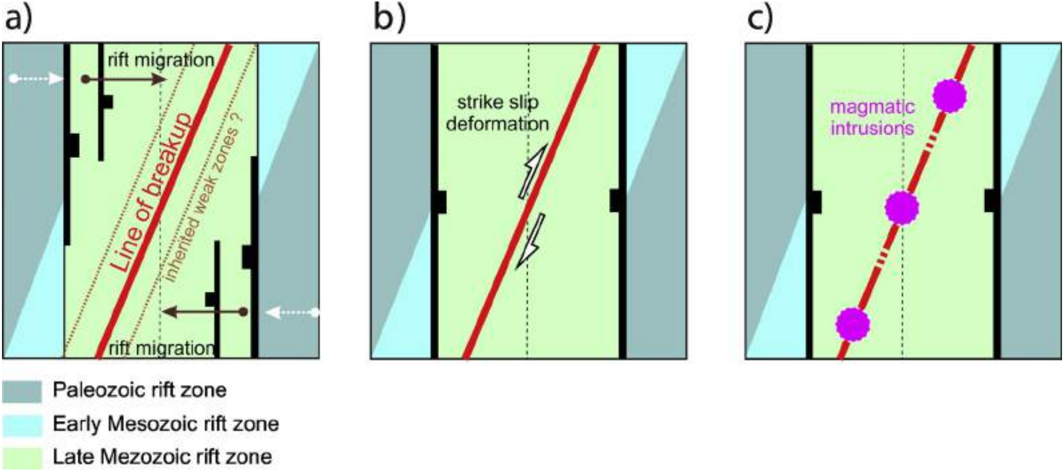

a)
rift migration
Line of breakup
inherited weak zones ?
rift migration
b)
strike slip
deformation
c)
magmatic
intrusions
Paleozoic rift zone
Early Mesozoic rift zone
Late Mezozoic rift zone

Figure 20

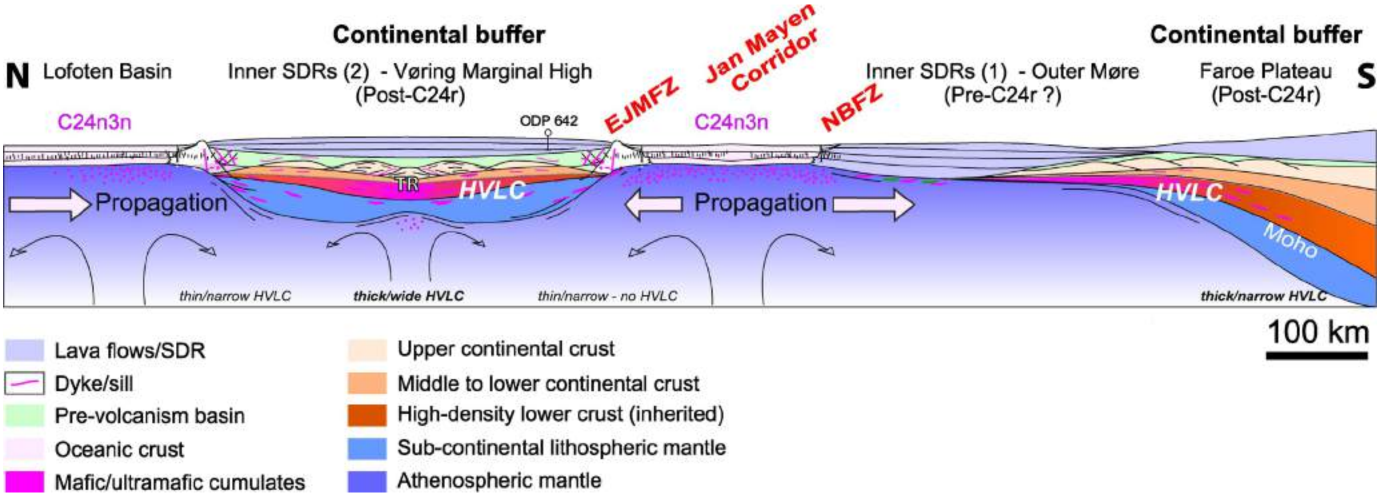

Continental buffer
Continental buffer
N
Lofoten Basin
Inner SDRs (2) - Vøring Marginal High
(Post-C24r)
Jan Mayen
Corridor
EJMFZ
NBFZ
Inner SDRs (1) - Outer Møre
(Pre-C24r ?)
Faroe Plateau
(Post-C24r)
S
C24n3n
ODP 642
C24n3n
TR
HVLC
HVLC
Propagation
Propagation
Moho
thin/narrow HVLC
thick/wide HVLC
thin/narrow - no HVLC
thick/narrow HVLC
100 km
Lava flows/SDR
Dyke/sill
Pre-volcanism basin
Oceanic crust
Mafic/ultramafic cumulates
Upper continental crust
Middle to lower continental crust
High-density lower crust (inherited)
Sub-continental lithospheric mantle
Athenospheric mantle

Figure 21