The Jan Mayen microplate complex and the Wilson cycle

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Abstract: The opening of the North Atlantic region was one of the most important geodynamic events that shaped the present day passive margins of Europe, Greenland and North America. Although well-studied, much remains to be understood about the evolution of the North Atlantic, including the role of the Jan Mayen microplate complex. Geophysical data provide an image of the crustal structure of this microplate and enable a detailed reconstruction of the rifting and spreading history. However, the mechanisms that cause the separation of microplates between conjugate margins are still poorly understood. We assemble recent models of rifting and passive margin formation in the North Atlantic and discuss possible scenarios that may have led to the formation of the Jan Mayen microplate complex. This event was probably triggered by regional plate tectonic reorganizations rejuvenating inherited structures. The axis of rifting and continental break-up and the width of the Jan Mayen microplate complex were controlled by old Caledonian fossil subduction/suture zones. Its length is related to east–west-oriented deformation and fracture zones, possibly linked to rheological heterogeneities inherited from the pre-existing Precambrian terrane boundaries.

The North Atlantic region inspired some aspects of plate tectonic theory (Fig. 1), including the Wilson cycle, which predicts the closure of oceans, leading to continent–continent collision followed by reopening along former sutures (Wilson 1966; Dewey & Spall 1975). The North Atlantic is often considered to be a textbook example of an ocean that opened along the former sutures of at least two temporarily distinct orogenic events: the Neoproterozoic Grenvillian–Sveconorwegian and the early Paleozoic Caledonian–Variscan orogenies (Ryan & Dewey 1997; Vauchez *et al.* 1997; Bowling & Harry 2001; Thomas 2006; Misra 2016). Nevertheless, some aspects of North Atlantic geology remain enigmatic, such as the formation of the North Atlantic igneous province (Vink 1984; White & McKenzie 1989; Foulger & Anderson 2005; Meyer *et al.* 2007), the development of volcanic passive margins (Franke 2013; Geoffroy *et al.* 2015), the formation of Iceland and the development of the Jan Mayen microplate complex (JMMC), also referred to as the Jan Mayen microcontinent (Foulger *et al.* 2003; Gaina *et al.* 2009; Gernigon *et al.* 2015). The JMMC consist of both oceanic and continental crust, probably highly thinned and magmatically modified (Kuvaas & Kodaira 1997; Blischke *et al.* 2017 and references cited therein). Large parts of it remain to be studied, however. Other continental fragments have been identified in the North Atlantic region (Nemčok *et al.* 2016) and more may underlie parts of Iceland and/or the Iceland–Faroe Ridge (IFR) (Fedorova *et al.* 2005; Foulger 2006; Paquette *et al.* 2006; Gernigon *et al.* 2012; Torsvik *et al.* 2015).

Geological setting of the North Atlantic region

Following the collision of Laurentia, Baltica and Avalonia in the Ordovician and Silurian (Roberts 2003; Gee *et al.* 2008; Leslie *et al.* 2008) and subsequent gravitational extensional collapse in the late orogenic phases (Dewey 1988; Dunlap & Fossen 1998; Rey *et al.* 2001; Fossen 2010), the North Atlantic region experienced lithospheric delamination and associated uplift over a period of 30–40 Ma, followed by a long period of rifting (Andersen *et al.* 1991; Dewey *et al.* 1993). Phases

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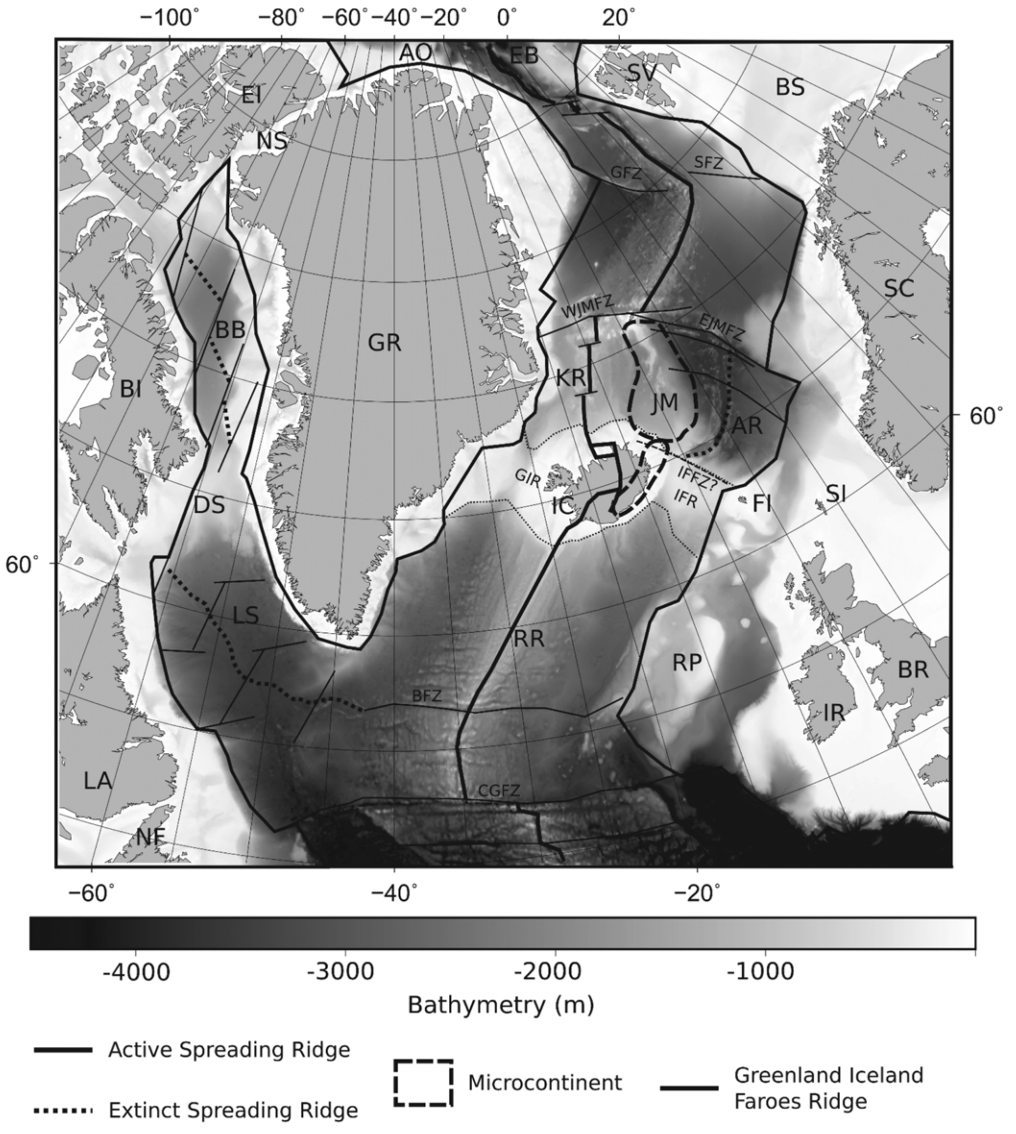


Fig. 1. Bathymetric map of the present day North Atlantic. Bathymetry from the General Bathymetric Chart of the Oceans (www.gebco.net). Major oceanic fracture zones after Doré *et al.* (2008), Mid-ocean ridges from Seton *et al.* (2012), microcontinents from Torsvik *et al.* (2015). The Greenland–Iceland–Faroe Ridge consists of the Greenland–Iceland Ridge, the Iceland Plateau and the Iceland–Faroe Ridge. The position of the Iceland–Faroe Fracture Zone is stippled, but its existence and nature is debated (see text). AO, Arctic Ocean; AR, Aegir Ridge; BB, Baffin Bay; BFZ, Bight Fracture Zone; BI, Baffin Island; BR, Britain; BS, Barents Sea; CGFZ, Charlie Gibbs Fracture Zone; DS, Davis Strait; EB, Eurasia basin; EI, Ellesmere Island; EJMFZ, East Jan Mayen Fracture Zone; FI, Faroe Islands; GIR, Greenland–Iceland Ridge; GR, Greenland; IC, Iceland; IFFZ, Iceland–Faroe Fracture Zone; IFR, Iceland–Faroe Ridge; IR, Ireland; JM, Jan Mayen; KR, Kolbeinsey Ridge; LA, Labrador; LS, Labrador Sea; NF, Newfoundland; NS, Nares Strait; RP, Rockall Plateau; RR, Reykjanes Ridge; SC, Scandinavia; SFZ, Senja Fracture Zone; SI, Shetland Islands; SV, Svalbard; WJMFZ, West Jan Mayen Fracture Zone.

of extension and cooling transitioned into continental rifting that led to final continental break-up and seafloor spreading between Greenland and Europe in the early Paleogene (Talwani & Eldholm 1977; Skogseid *et al.* 2000). During the late Mesozoic, continental break-up propagated simultaneously southwards from the Eurasia Basin and northwards from the Central Atlantic, initially into the Labrador Sea–Baffin Bay rift system and then into the North Atlantic (Srivastava 1978; Doré *et al.* 2008). Whether rifting, continental break-up and the associated magmatism were initiated by active mantle upwelling, for example a deep mantle plume (White & McKenzie 1989; Hill 1991; Nielsen *et al.* 2002; Rickers *et al.* 2013) or plate-driven processes (Nielsen *et al.* 2007; Ellis & Stoker 2014) (bottom-up or top-down views) is still under debate (van Wijk *et al.* 2001; Foulger *et al.* 2005*b*; Lundin & Doré 2005; Simon *et al.* 2009; Peace *et al.* 2017*a*).

The Northeast Atlantic spreading axis initially consisted of the Reykjanes Ridge and the Aegir Ridge, east of the JMMC and the Mohns Ridge further north (Talwani & Eldholm 1977; Nunns 1983, Fig. 1). Independent rotation of the JMMC resulted in the fan-shaped opening of the Norway Basin during the Eocene (Nunns 1983; Gaina *et al.* 2009; Gernigon *et al.* 2012). This reconfiguration led to a second phase of break-up and the separation of the JMMC from Greenland at about magnetic anomaly chron C7 (*c.* 24 Ma) (Vogt *et al.* 1970; Gaina *et al.* 2009; Gernigon *et al.* 2015). After a period of simultaneous rifting on both the Aegir Ridge and the complex JMMC/proto-Kolbeinsey rift/ridge system (Doré *et al.* 2008; Gaina *et al.* 2009; Gernigon *et al.* 2015), the Aegir Ridge was abandoned in the Oligocene and the spreading centre relocated to the west of the JMMC onto the Kolbeinsey Ridge.

Although the history of rifting in the North Atlantic is becoming increasingly better constrained, the mechanisms controlling the location, timing and formation of rifts, fracture zones and associated microcontinents are still poorly understood. The formation of the JMMC has been traditionally attributed to mantle plume impingement and subsequent lithospheric weakening (Müller *et al.* 2001). More recently it has been suggested to result from the breaching of lithosphere weakened as a result of pre-existing structures (e.g. Schiffer *et al.* 2015*b*). The present-day North Atlantic shows evidence for dynamic topography that may have assisted breakup processes (Schiffer & Nielsen 2016). The final separation of the JMMC is also spatially and temporally linked to enhanced magmatic activity and the subsequent formation of Iceland (Doré *et al.* 2008; Tegner *et al.* 2008; Larsen *et al.* 2013; Schiffer *et al.* 2015*b*), but it lacks the classic features of a volcanic passive margin (e.g. underplating and seaward-dipping reflectors) along its western continent–ocean boundary, conjugate to the East Greenland margin (Kodaira *et al.* 1998; Breivik *et al.* 2012; Peron-Pinvidic *et al.* 2012; Blischke *et al.* 2017). We discuss here the possible role of pre-existing structures and inheritance in the formation of the JMMC as an extension to the Wilson cycle and plate tectonic theory.

Jan Mayen microplate complex

The JMMC has a bathymetric signature stretching >500 km from north to south in the central part of the Norwegian–Greenland Sea (Fig. 1) (Gudlaugsson *et al.* 1988; Kuvaas & Kodaira 1997; Blischke *et al.* 2017). It is bordered to the north by the Jan Mayen Fracture Zone and the volcanic complex of Jan Mayen Island. To the south, it is bordered by the NE coastal shelf of Iceland, which is part of the Greenland–Iceland–Faroe Ridge (GIFR), a zone of shallow bathymetry *c.* 1100 km in length (Figs 1 & 2). The JMMC separates the Norway Basin to the east from the Iceland Plateau to the west (Vogt *et al.* 1981; Kandilarov *et al.* 2012; Blischke *et al.* 2017).

The JMMC crust has been inferred to be continental, primarily on the basis of seismic refraction data (Kodaira *et al.* 1997, 1998; Mjelde *et al.* 2007*a*; Breivik *et al.* 2012; Kandilarov *et al.* 2012). However, the crustal affinity remains uncertain for large areas of the JMMC, particularly near Iceland in the south (Breivik *et al.* 2012; Brandsdóttir *et al.* 2015) due to a lack of geophysical data and boreholes (see Gernigon *et al.* 2015; Blischke *et al.* 2017 for data coverage). Fundamentally, the distribution of oceanic v. continental crust and the nature of the deformation expected between the JMMC, Iceland and the Faroe continental block are unknown. Recent high-resolution aeromagnetic data and pre-rift reconstructions of the Norwegian–Greenland Sea show that the southern JMMC underwent extreme thinning during the first phase of break-up and, because it now has a width of *c.* 250–300 km, 400% of extension has occurred relative to its pre-drift configuration (Gernigon *et al.* 2015). It seems unlikely that this extreme extension is entirely accommodated by the thinning of continental crust. We cannot rule out the possibility that the southern JMMC consists partly of igneous crust (Gernigon *et al.* 2015) or exhumed mantle (Blischke *et al.* 2017).

An oceanic fracture zone might be present south of the JMMC between the northeastern tip of the Iceland Plateau and the Faroe Islands in the SE (i.e. the postulated Iceland–Faroe Fracture Zone; see Figs 1 & 2; Blischke *et al.* 2017). However, an oceanic fracture zone or transform requires oceanic lithosphere on both sides and, given the uncertain crustal affinity, this interpretation is speculative. A lineament exists north of the Iceland–Faroe Ridge, but magnetic and gravity potential field data do not provide conclusive evidence for a real oceanic transform or fracture zone (Fig. 3). Gernigon *et al.* (2012) showed that continuation of the magnetic chrons mapped in the Norway Basin and the high-magnetic trends observed along the IFR remain unclear, notably due to the low quality, sparse distribution of the magnetic profiles along the IFR and later igneous overprinting related to the formation of Iceland. No

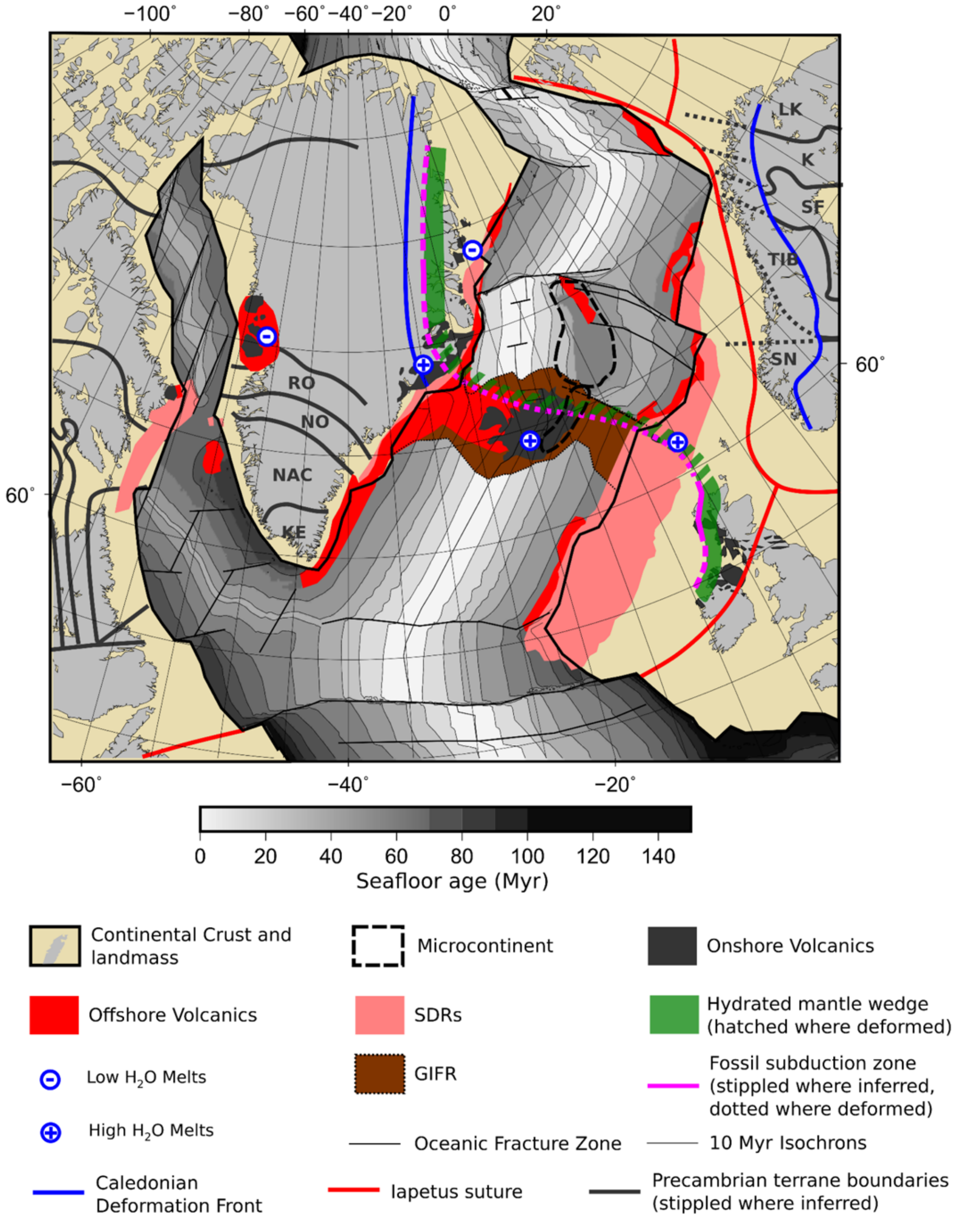


Fig. 2. Overview map of the present day North Atlantic. Seafloor age from Seton *et al.* (2012); major oceanic fracture zones after Doré *et al.* (2008); distribution of igneous rocks of the North Atlantic Igneous Province after Upton (1988), Larsen & Saunders (1998), Abdelmalak *et al.* (2012); Precambrian basement terranes after Balling (2000) and Indrevær *et al.* (2013) (Scandinavia) and St-Onge *et al.* (2009) (Greenland and northeastern Canada). Caledonian deformation front after Skogseid *et al.* (2000) and Gee *et al.* (2008). K, Karelian; KE, Ketilian Orogen; LK, Lapland–Kola; NAC, North Atlantic Craton; NO, Nagssugtoqidian Orogen; RO, Rinkian Orogen; SF, Svecofennian; SN, Sveconorwegian; TIB, Transscandinavian Igneous Belt.

magnetic chron is identified in the broad NE–SW magnetic lineations, especially west of the Faroe Platform. Additional magnetic disparities are associated with lateral variations in basement depth and possible discrete ridge jumps (e.g. Smallwood & White 2002; Hjartarson *et al.* 2017).

The IFR consists of anomalously thick crust (>20–25 km), possibly to parts associated with massive crustal underplating, which is generally attributed to increased magmatism (Staples *et al.* 1997; Richardson *et al.* 1998; Smallwood *et al.* 1999; Darbyshire *et al.* 2000; Greenhalgh & Kusznir 2007). The origin and nature of the GIFR remains controversial (McBride *et al.* 2004) because the crust shows atypical geophysical properties and differs from 'normal' continental and oceanic crust (Bott 1974; Foulger *et al.* 2003). Hjartarson *et al.* (2017) favour an oceanic origin for the IFR, but do not exclude the presence of seaward-dipping reflectors and old basement in the expected 'oceanic domain'.

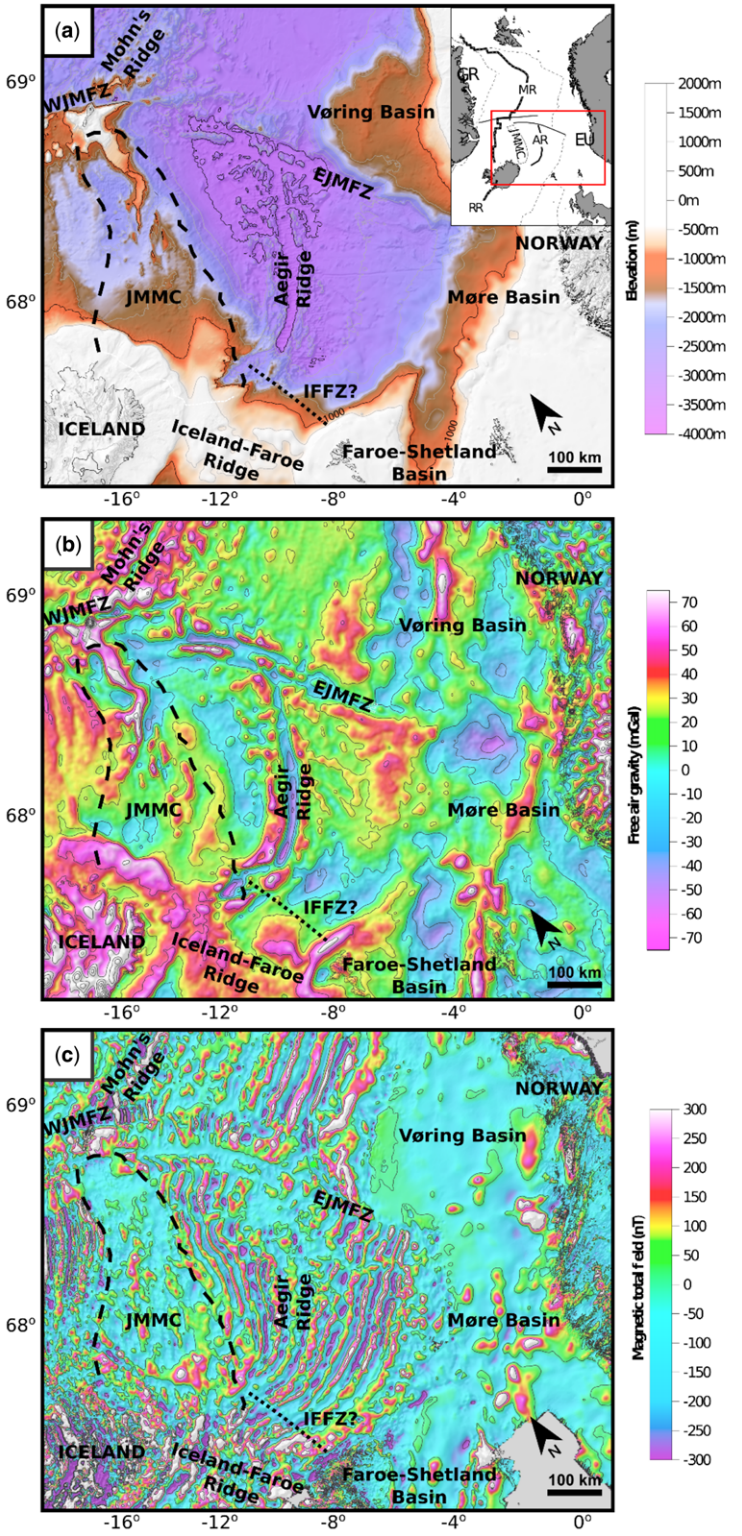
Some researchers have suggested that the excess thickness under Iceland may be partly attributed to buried continental crust, possibly extending up to the JMMC and Iceland (Fedorova *et al.* 2005; Foulger 2006). Continental zircons and geochemical analyses of lavas in SE Iceland support the presence of continental material (Paquette *et al.* 2006; Torsvik *et al.* 2015). The Aegir Ridge and the Reykjanes Ridge might never have connected during the early stage of spreading of the Norway Basin, which involved complex overlapping spreading segments along the IFR. Such overlapping spreading ridges may have preserved continental lithosphere in between (Gaina *et al.* 2009; Gernigon *et al.* 2012, 2015; Ellis & Stoker 2014). Ellis & Stoker (2014) suggested that complete continental break-up along the GIFR did not occur before the separation of the JMMC and the appearance of Iceland (first dated eruptions at *c.* 18 Ma). Gernigon *et al.* (2015) suggested earlier break-up, possibly between C22/C21 (*c.* 47 Ma) and C6 (*c.* 24 Ma) during the onset of significant rifting in the southern part of the JMMC. The continental lithosphere east of Iceland (the IFR, Fig. 1) probably did not entirely breach in the early rifting of the North Atlantic (e.g. C24r–C22, Early Eocene). To avoid further ambiguity, we refer to it as the Iceland–Faroe accommodation zone (IFAZ). Consequently, the IFAZ may characterize local continental transform margin segments, a diffuse strike-slip fault zone and/or a more complex oblique/transtensional continental rift system that initially formed along the trend of the proto IFR.

Microplate formation

An aspect of the Wilson cycle that requires more clarification (Thomas 2006; Huerta & Harry 2012; Buiter & Torsvik 2014) is whether the locations of major, pre-existing structures can explain the formation, location and structure of microplates such as the JMMC (Schiffer *et al.* 2015*a*). Understanding the formation of continental fragments is crucial to understanding continental break-up (Lavier & Manatschal 2006; Peron-Pinvidic & Manatschal 2010). Microcontinents and continental ribbons represent one category of continental fragments produced during rifting and break-up (Lister *et al.* 1986; Peron-Pinvidic & Manatschal 2010; Tetreault & Buiter 2014).

We follow the original definition of a microcontinent Scrutton (1976) that it must contain: (1) pre-rift basement rocks; (2) crust and lithosphere of continental affinity, horizontally displaced from the original continent and surrounded by oceanic crust; and (3) a distinct morphological feature in the surrounding oceanic basins. Such a system between two pairs of conjugate margins may also include isolated fragments of oceanic crust and lithosphere that deformed together before final and definitive isolation from the conjugate continents. To make a distinction, we call such a feature a microplate complex and it can involve several sub-plates of oceanic and/or continental affinity. A true microcontinent will therefore consist of just one kind of microplate complex. The most important aspect of the present study is that such a microplate complex, like a true microcontinent, is separated from the main continental conjugate margins by two or more spreading ridges. The cause, history and processes leading to relocalization of the complex are not well understood. Suggested mechanisms include the impact of a mantle plume (Müller *et al.* 2001; Gaina *et al.* 2003; Mittelstaedt *et al.* 2008), global plate tectonic reorganization (Collier *et al.* 2008; Gaina *et al.* 2009) and ridge jumps that exploit inhomogeneities, weaknesses and rheological contrasts in the continental lithosphere after the abandonment of a previous spreading ridge (Abera *et al.* 2016; Sinha *et al.* 2016). This could be nascent or inherited underplating (Yamasaki & Gernigon 2010) and/or fossil suture zones (Petersen & Schiffer 2016). Strike-slip mechanisms under different transtensional and transpressional stress regimes have also been proposed to generate microcontinents (Nemčok *et al.* 2016). Microplates can also result from crustal fragmentation during volcanic margin formation by large-scale continent-vergent faults formed/activated by strengthening of the deep continental crust: the so-called C-block mechanism (Geoffroy *et al.* 2015).

Whittaker *et al.* (2016) proposed a model for microcontinent formation between Australia and Greater India whereby changes in the direction of plate motion caused transpression and stress build-up across large-offset fracture zones, leading


(a)
Mohn's Ridge
WJMFZ
Vøring Basin
EJMFZ
Aegir Ridge
JMMC
NORWAY
Møre Basin
IFFZ?
ICELAND
Iceland-Faroe Ridge
Faroe-Shetland Basin
100 km
N
GR
MR
AR
EU
RR
Elevation (m)
2000m
1500m
1000m
500m
0m
-500m
-1000m
-1500m
-2000m
-2500m
-3000m
-3500m
-4000m
69°
68°
-16°
-12°
-8°
-4°
0°
(b)
Free air gravity (mGal)
70
60
50
40
30
20
10
0
-10
-20
-30
-40
-50
-60
-70
(c)
Magnetic total field (nT)
300
250
200
150
100
50
0
-50
-100
-150
-200
-250
-300

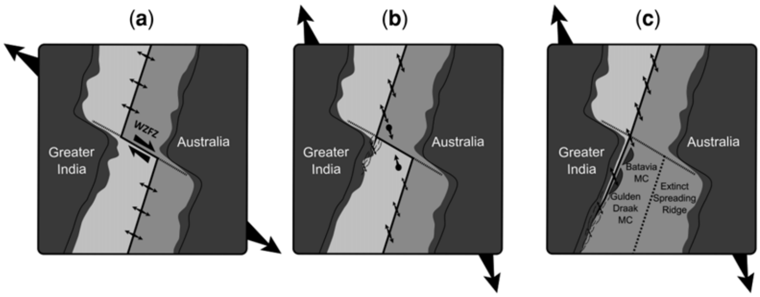


Fig. 4. Model for the formation of the Batavia and Gulden Draak microcontinents in the Indian Ocean proposed by Whittaker *et al.* (2016). (**a**) Initial seafloor spreading occurred perpendicular to the regional plate motions, including the Wallaby–Zenith Fracture Zone. (**b**) A reconfiguration of plate motions oblique to the developed spreading axes locked the fracture zone, which forced the southern spreading axis to relocate onto a new axis. (**c**) The new spreading isolates continental fragments (microcontinents) and seafloor spreading separates these from the Indian plate. Large arrows indicate plate motions. Arrows along spreading ridges indicate the spreading direction. Dots with arrows indicate the transpressional regime along the former fracture zone. WZFZ, Wallaby–Zenith Fracture Zone.

to the transfer of deformation to a less resistive locus (Fig. 4). Their proposed model is as follows. Initially, NW–SE spreading separated Australia from Greater India with transtensional or strike-slip motion along the Wallaby–Zenith Fracture Zone from 133 Ma. A plume (Kerguelen) is postulated to have been in the vicinity and may have maintained and/or enhanced crustal weakening of the SE Greater India rifted margin. Reorganizations of motion between Australia and Greater India to a NNW–SSE direction at 105 Ma resulted in transpression along the NW–SE-oriented Wallaby–Zenith Fracture Zone. As a result, the spreading centre relocated to the west along the continental margin of India, calving off the Batavia and Gulden Draak microcontinents and resulting in abandonment of the Dirck Hartog spreading ridge to the south (Fig. 4).

North Atlantic: structure and inheritance

The classic Wilson cycle model envisages closure and reopening of oceans along continental sutures. In this model, break-up is guided by lithospheric inheritance from previous orogenesis (Wilson 1966; Dewey & Spall 1975). Inheritance, rejuvenation and the control of pre-existing structures on localizing deformation occurs on various scales and styles beyond the large-scale break-up of continents (Holdsworth *et al.* 1997; Manatschal *et al.* 2015; Peace *et al.* 2017*b*). Inherited features may include variations in crustal or lithospheric thickness, structural and compositional heterogeneity across terrane boundaries, accreted terranes, sedimentary basins and/or intruded, metamorphosed and metasomatized material and fabrics. These heterogeneities may also cause thermal and rheological anomalies that vary in size, depth and degree of anisotropy and that can potentially be rejuvenated given the appropriate stresses (Krabbendam & Barr 2000; Tommasi *et al.* 2009; Manatschal *et al.* 2015; Tommasi & Vauchez 2015). Inheritance is an important control on rifting, passive margin end-member style (e.g. volcanic or non-volcanic) (Vauchez *et al.* 1997; Bowling & Harry 2001; Chenin *et al.* 2015; Manatschal *et al.* 2015; Schiffer *et al.* 2015*b*; Svartman Dias *et al.* 2015; Duretz *et al.* 2016; Petersen & Schiffer 2016), the formation of fracture zones, transform faults, transform margins (Thomas 2006; Gerya 2012; Doré *et al.* 2016), magmatism (Hansen *et al.* 2009; Whalen *et al.* 2015), compressional deformation (Sutherland *et al.* 2000; Gorczyk & Vogt 2015; Heron *et al.* 2016), the break-up of supercontinents and supercontinent cycles (Vauchez *et al.* 1997; Audet & Bürgmann 2011; Frizon de Lamotte *et al.* 2015).

Fig. 3. (**a**) Bathymetry, (**b**) free air gravity and (**c**) magnetic anomaly maps of the Norway Basin, the Jan Mayen microplate complex (JMMC), Iceland, the Iceland–Faroe Ridge and surrounding conjugate margins (modified after Gernigon *et al.* 2015). The bathymetric map illustrates the special physiological nature of the JMMC, coinciding with large free air gravity anomalies. Magnetic anomalies within the boundaries of the JMMC are weak. This is in contrast with the adjacent Norway Basin, which shows clear magnetic spreading anomalies, and gravity and topographic anomalies that evidence fan-shaped spreading along the extinct Aegir Ridge. There are vague indications in the bathymetric, gravity and magnetic data for the existence of a lineament stretching from the south of the JMMC to the Faroe–Shetland Basin, possibly the Iceland–Faroe Fracture Zone (Blischke *et al.* 2017), but the data do not provide indisputable evidence for the existence and nature of this lineament. EJMFZ, East Jan Mayen Fracture Zone; IFFZ, Iceland–Faroe Fracture Zone; WJMFZ, West Jan Mayen Fracture Zone.

Precambrian orogenies

In Canada, Greenland and NW Europe, multiple suturing events have built continental lithosphere that consists of Archean to early Proterozoic cratons surrounded by younger terranes. Preserved sutures and subduction zones in the interior of the cratons have survived subsequent amalgamation, demonstrating that crustal and upper mantle heterogeneities may persist for billions of years (Balling 2000; van der Velden & Cook 2005). Terrane boundaries of any age may act as rheological boundaries that influence or control crustal deformation long after their formation and independently of subsequent plate motions. Major Precambrian terrane boundaries in the North Atlantic region are shown in Figure 2.

Multiple Precambrian suturing events have contributed to the amalgamation of the Baltic Shield in Scandinavia. The Lapland–Kola mobile belt formed by the accretion of various Archean to Paleoproterozoic terranes, including the oldest Karelian terrane (Gorbatschev & Bogdanova 1993; Bergh *et al.* 2012; Balling 2013). This was followed by the late Paleoproterozoic Svecofennian accretion, the formation of the Transscandinavian igneous belt and the Meso-Neoproterozoic Sveconorwegian orogeny (Gorbatschev & Bogdanova 1993; Bingen *et al.* 2008; Bergh *et al.* 2012; Balling 2013; Slagstad *et al.* 2017).

Precambrian terranes are also preserved in Greenland, the oldest of which are Archean in age and include the North Atlantic and Rae cratons (St-Onge *et al.* 2009). The components that together constitute the North Atlantic Craton formed from 3850 to 2550 Ma (Polat *et al.* 2014) and the Rae Craton formed from 2730 to 2900 Ma (St-Onge *et al.* 2009). Paleoproterozoic terranes in Greenland surround the North Atlantic Craton and include the Nagssugtoqidian Orogen (Van Gool *et al.* 2002), the Rinkian Orogen (Grocott & McCaffrey 2017) and the Ketilidian mobile belt (Garde *et al.* 2002).

The Precambrian terranes of NE Canada, Greenland and Scandinavia are thought to have formed as coherent mobile belts (Kerr *et al.* 1996; Wardle *et al.* 2002; St-Onge *et al.* 2009). As Greenland and North America have not undergone significant relative lateral motion or rotation, the interpretation of conjugate margins is relatively simple (Kerr *et al.* 1996; Peace *et al.* 2016). By contrast, whether Baltica has experienced rotation (Gorbatschev & Bogdanova 1993; Bergh *et al.* 2012) is currently unresolved.

Caledonian orogeny

Formation of the Ordovician to Devonian Caledonian–Appalachian Orogen preceded rifting, ocean spreading and subsequent passive margin formation of the present day North Atlantic. This Himalaya-style orogen involved at least two phases of subduction: the early eastward-dipping Grampian–Taconian event and the late westward-dipping Scandian event, which led to the assembly of part of Pangaea (Roberts 2003; Gee *et al.* 2008). The structural fabric of the crust and lithospheric mantle can be reoriented during orogenesis, resulting in fabric anisotropy that localizes subsequent deformation (Tommasi *et al.* 2009; Tommasi & Vauchez 2015).

High-velocity, lower crustal bodies (HVLCBs) are observed along many passive continental margins (Lundin & Doré 2011; Funck *et al.* 2017*a*) and have been traditionally associated with magmatic underplating or intrusions into the lower crust of passive margins during break-up (Olafsson *et al.* 1992; Eldholm & Grue 1994; Mjelde *et al.* 2007*b*; White *et al.* 2008; Thybo & Artemieva 2013). However, with improved data, alternative interpretations have been proposed, such as synrift serpentinization of the uppermost mantle under passive margins (Ren *et al.* 1998; Reynisson *et al.* 2010; Lundin & Doré 2011; Peron-Pinvidic *et al.* 2013). It has also been suggested that part of the continental HVLCBs may be remnants of inherited metamorphosed crust or hydrated meta-peridotite that existed prior to initial rifting and continental break-up (Gernigon *et al.* 2004, 2006; Fichler *et al.* 2011; Wangen *et al.* 2011; Mjelde *et al.* 2013; Nirrengarten *et al.* 2014).

Mjelde *et al.* (2013) identified a number of such 'orogenic' HVLCBs along different parts of the North Atlantic passive margins (the south and mid-Norwegian margin, the East Greenland margin, the SW Barents Sea margin and the Labrador margin), which may have higher than normal upper mantle velocities ($V_p > 8.2$ km s^{-1}). These may consist of eclogitized crust and be part of the Iapetus Suture. Petersen & Schiffer (2016) proposed a mechanism to explain the presence of old inherited HVLCBs beneath the rifted margins and concluded that they could represent preserved and subsequently deformed pre-existing subduction/suture zones that were activated during rifting and continental break-up. Eclogite in a fossil slab has a similar, but weaker, rheology than the surrounding dry olivine lithosphere (Zhang & Green 2007), whereas a fossil hydrated mantle wedge acts as an effective and dominant weak zone. Eclogites of the Bergen Arc (Norway) show softening due to fluid infiltration (Jolivet *et al.* 2005). These ultra-high-velocity HVLCBs (ultra-HVLCBs) are distributed primarily along the mid-Norwegian margin and the Scoresbysund area in East Greenland (Mjelde *et al.* 2013). This suggests that at least one fossil subduction zone may have been subject to rift-related deformation and exhumation (Petersen & Schiffer 2016).

Structures in the Central Fjord area of East Greenland (Schiffer *et al.* 2014), the Flannan reflector in northern Scotland (Snyder & Flack 1990; Warner *et al.* 1996) and the Danish North Sea (Abramovitz & Thybo 2000) have been interpreted as preserved orogenic structures of Caledonian age (i.e. fossil subduction or suture zones) (Fig. 2). Schiffer *et al.* (2015*a*) proposed that the Central Fjord structure and the Flannan reflector once formed a contiguous eastward-dipping subduction zone, possibly of Caledonian age, that may have influenced rift, magmatic and passive margin evolution in the North Atlantic (Fig. 2). Combined geophysical–petrological modelling of the Central Fjord structure suggests that it consists of a relict hydrated mantle wedge associated with a fossil subduction zone (Schiffer *et al.* 2015*b*, 2016).

The most recent Caledonian subduction event was associated with the Scandian phase, leading to the westward subduction of Iapetus crust (Roberts 2003; Gee *et al.* 2008). Evidence of this subduction zone in the form of a preserved slab has not been detected in the lithospheric mantle of the Norwegian Caledonides. However, the fossil subduction zone in the Danish North Sea detected by the Mona Lisa experiments (Abramovitz & Thybo 2000) might be the trace of this event. HVLCBs indicative of eclogite along the mid-Norwegian margin (Mjelde *et al.* 2013; Kvarven *et al.* 2016) and Norwegian North Sea (Christiansson *et al.* 2000; Fichler *et al.* 2011) might also represent deformed remnants of the Scandian subduction.

Fracture and accommodation zones

The JMMC is bound by two tectonic boundaries including the East and West JMFZs in the north and the postulated IFAZ in the south. These tectonic boundaries accommodated and allowed the non-rigid microplate to move independently from the surrounding North Atlantic oceanic domains (Gaina *et al.* 2009; Gernigon *et al.* 2012, 2015).

Relationships between pre-existing structures and the formation of large-scale shear and fracture zones, oceanic transforms or other accommodation/deformation zones have been proposed previously (Mohriak & Rosendahl 2003; Thomas 2006; Taylor *et al.* 2009; de Castro *et al.* 2012; Gerya 2012; Bellahsen *et al.* 2013; Gibson *et al.* 2013). The location, orientation and nature of fracture zones in the North Atlantic may be linked to lithospheric inheritance (Behn & Lin 2000). For example, the Charlie Gibbs Fracture Zone between Newfoundland and the British/Irish shelf has been linked to the location of the Iapetus Suture and the inheritance of compositional and structural weaknesses (Tate 1992; Buiter & Torsvik 2014). The Bight Fracture Zone might be linked to the Grenvillian Front, which is exposed in Labrador (Lorenz *et al.* 2012).

The IFAZ represents a complex discontinuity zone along the present day GIFR. Fragments of continental crust may be preserved along this transition zone between the Reykjanes, Aegir and Kolbeinsey ridges, together with discontinuous and/or overlapping oceanic fragments affected later by significant magmatic overprint (the Icelandic swell; Bott 1988). In the geodynamic context, it may have formed along the fossil subduction zone proposed to have existed between the East Greenland and British/Irish margins (Fig. 2). It has also been proposed that it may have comprised part of the Kangerlussuak Fjord tectonic lineament, a NW–SE-oriented lineament in East Greenland (Tegner *et al.* 2008).

Other deformation zones may correlate with Precambrian basement terrane boundaries in Scandinavia. These are overprinted by Caledonian deformation, obscuring older relationships (cf. the Caledonian deformation front in Fig. 2) and generating new orogenic fabrics (Vauchez *et al.* 1998). The westwards extrapolation of the northern Sveconorwegian suture may correlate with the East JMFZ, whereas extrapolation of the Svecofennian–Karelian suture may correspond to the formation of the Senja Fracture Zone (Doré *et al.* 1999; Fichler *et al.* 1999; Indrevær *et al.* 2013). Extrapolation of the Karelian–Lapland–Kola terrane suture converges with the complex DeGeer Fracture Zone, which marks the transition of the North Atlantic to the Arctic Ocean (Engen *et al.* 2008). These correlations suggest that Precambrian basement inheritance localizes strain during initial continental rifting. However, the exact location and grade of deformation of Precambrian sutures under the Caledonides and the highly stretched continental margins is often poorly known or not known at all. Thus any correlation is speculative and requires further work.

Iceland and magmatic evolution

Factors including the thermal state of the crust and mantle, small-scale convection, upwelling, composition, volatile content, and lithospheric and crustal structure may all have roles in magmatic evolution (King & Anderson 1998; Asimow & Langmuir 2003; Korenaga 2004; Foulger *et al.* 2005*a*; Hansen *et al.* 2009; Brown & Lesher 2014; Chenin *et al.* 2015; Hole & Millett 2016).

Inheritance may influence the amount of volcanism produced in the North Atlantic because volcanic passive margins preferentially develop in regions of heterogeneous crust where Paleozoic orogenic belts separate Precambrian terranes. Conversely, magma-poor margins often develop in the interiors of orogenic belts with either uniform Precambrian or younger Paleozoic crust (Bowling & Harry 2001). For example, the intersection of the East Greenland–Flannan fossil subduction zone with the North Atlantic rift axis correlates spatially and temporally with pre-break-up magmatism, the formation of the JMMC and the occurrence of the Iceland melt anomaly along the sub-parallel Greenland Iceland Ridge (GIR) (Schiffer *et al.* 2015*b*).

Prior to break-up (*c.* 55 Ma), magma was dominantly emplaced along and SW of the proposed East Greenland–Flannan fossil subduction zone (Fig. 2) (Ziegler 1990; Torsvik *et al.* 2002). This may partly be an effect of the south-to-north unzipping of the pre-North Atlantic lithosphere. Other processes that produce enhanced mantle melting are increased temperature, mantle composition and active asthenospheric upwelling (Brown & Lesher 2014). The zonation of areas with and without magmatism may suggest that the proposed structure is a boundary zone between lithospheric blocks of different composition and rheology that react differently to applied stresses. Different relative strengths in the crust and mantle lithosphere, for instance, could cause depth-dependent deformation, where thinning is focused in the mantle lithosphere (Huismans & Beaumont 2011). Petersen & Schiffer (2016) demonstrated that the extension of orogenic lithosphere with thickened crust (>45 km) leads to depth-dependent thinning, where the mantle lithosphere breaks earlier than the crust and, as a result, encourages pre-break-up magmatism. Indirectly, sub-continental mantle heterogeneities may encourage the localization of deformation, leading to a rapid and sudden increase in lithospheric thinning (Yamasaki & Gernigon 2010). These processes could contribute to pre-break-up adiabatic decompression melting (Petersen & Schiffer 2016).

Enhanced magmatism could also be caused by a lowered solidus due to the presence of eclogite (Foulger *et al.* 2005*a*), water (Asimow & Langmuir 2003) or carbon dioxide in the mantle (Dasgupta & Hirschmann 2006). Atypical magmatism is, surprisingly, observed along the interpolated axis of the proposed fossil subduction zone. It currently coincides with the GIFR where the igneous crustal thickness is inferred to be the greatest (Bott 1983; Smallwood *et al.* 1999; Holbrook *et al.* 2001; Mjelde & Faleide 2009; Funck *et al.* 2017*b*). However, it is unclear whether the entire thickness of Iceland-type crust (Bott 1974; Foulger *et al.* 2003) has a crustal and/or continental petrology (Foulger *et al.* 2003; Foulger & Anderson 2005).

Higher water contents have been recorded in basalts and volcanic glass in the vicinity of the fossil subduction zone (the Blosseville Kyst, East Greenland, Iceland and one sample from the Faroe Islands; see Fig. 2) than in regions further away from Iceland (West Greenland, Hold with Hope, Reykjanes Ridge) (Jamtveit *et al.* 2001; Nichols *et al.* 2002). This is consistent with a hydrated upper mantle source as a consequence of melting Caledonian subducted materials (Schiffer *et al.* 2015*a*). Water in the mantle may also contribute to enhanced melt production and thus unusually thick igneous crust (Asimow & Langmuir 2003).

The formation of the Iceland Plateau (>18 Ma) followed the extinction of the Aegir Ridge and full spreading being taken up on the Kolbeinsey Ridge (Doré *et al.* 2008). This migration of the spreading ridge was contemporaneous with far-field plate tectonic reconfigurations, the end of seafloor spreading in the Labrador–Baffin Bay system (Chalmers & Pulvertaft 2001) and a global change in the motion of the Greenland plate from SW–NE to west–east (Gaina *et al.* 2009; Abdelmalak *et al.* 2012).

Inheritance model for the formation of the JMMC

We propose a new tectonic model for the formation of the JMMC, linking the rejuvenation of old and pre-existing orogenic structures to global plate tectonic reconfigurations. In our model, a change in the orientation of the regional stress field in the Eocene rejuvenated pre-existing structures with favourable orientations. This caused the relocalization of extension and spreading ridges, resulting in the formation of a microplate between the large European and American/Greenland continental plates. Our model closely follows that of Whittaker *et al.* (2016), with the extension that a fossil subduction zone is utilized as a physical and compositional weak zone that helps to accommodate a second axis of break-up (Fig. 5).

Plate tectonic reorganizations and the rejuvenation of pre-existing structures may not be the only controls on continental break-up, but they may be the dominant controls in the case of the JMMC. In areas where no microplate formation is observed, continental break-up followed the youngest, weakest Caledonian collision zone, the west-dipping Scandian subduction in Scandinavia. This may have been better aligned with the ambient stress field during rifting and/or break-up. Following the model of Petersen & Schiffer (2016), the remnants of this subduction zone or other inherited orogenic structures may now be distributed along the mid-Norwegian margin as pre-break-up HVLCBs (Christiansson *et al.* 2000; Gernigon *et al.* 2006; Fichler *et al.* 2011; Wangen *et al.* 2011; Mjelde *et al.* 2013, 2016; Nirrengarten *et al.* 2014). The subduction zone was already deformed in the Norwegian North Sea by rifting subsequent to the Permo-Triassic and is still

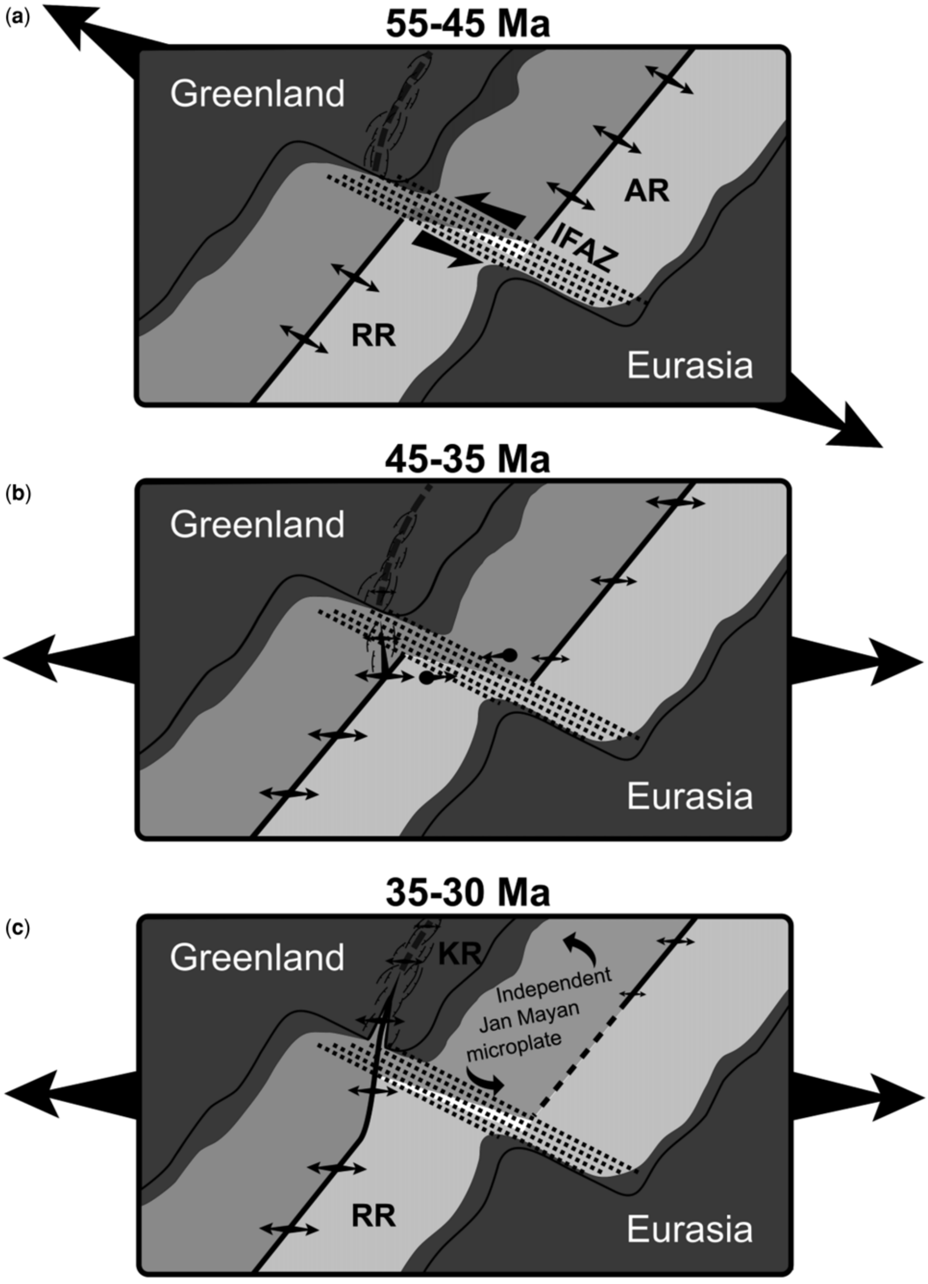


Fig. 5. Application of the model of Whittaker *et al.* (2016) to the formation of the Jan Mayen microplate complex. The original model was developed to explain microcontinent separation between Greater India and Australia. (**a**) NW–SE plate motion between Greenland and Europe with the Iceland–Faroe accommodation zone as a diffuse zone accommodating relative motion between the Reykjanes Ridge and Aegir Ridge. Continental rifting and extension occurs along the lithospheric weakness (East Greenland fossil subduction zone). (**b**) Plate tectonic reorganizations result in west–east motion between Greenland and Europe, locking up the Iceland–Faroe accommodation zone. The Reykjanes Ridge diverts towards the north following the lithospheric weakness. (**c**) Seafloor spreading develops along the Kolbeinsey Ridge, breaking the Jan Mayen microplate off from Greenland. The JMMC rotates counter-clockwise. Seafloor spreading on the Aegir Ridge is abandoned. AR, Aegir Ridge; IFAZ, Iceland–Faroe accommodation zone; KR, Kolbeinsey Ridge; RR, Reykjanes Ridge.

preserved as a large HVLCB beneath the North Sea rift (Christiansson *et al.* 2000; Fichler *et al.* 2011). A stronger, east-dipping subduction zone in East Greenland may also have been deformed, but did not accommodate break-up. Continental rifting and possible overlapping of the Reykjanes and Mohns ridge, leading to the initiation of the formation of the JMMC (Gernigon *et al.* 2012, 2015), may have been promoted by the presence of this deep-rooted weak zone.

The Caledonian and Grenvillian orogenic fabric and major associated structures are generally parallel to the NNE–SSE trend of rifting in the North Atlantic, with some exceptions, such as the opening of the Labrador Sea. Older terrane boundaries are close to perpendicular. Young Caledonian structures define the axis of rifting and continental break-up. This can be explained by the presence of deep, weak eclogite facies roots along the axis of the Caledonian Orogen and extensional collapse of the Caledonian mountain range causing earlier extension to initiate perpendicular to the axis of collision (Ryan & Dewey 1997; Rey *et al.* 2001). Precambrian structures are still preserved in stable cratons surrounded by orogens and mobile belts. Once rifting occurs, lateral weaknesses and rheological boundaries control the segmentation of the rift axis and eventually influence the formation of across-strike deformation zones of different kinds, e.g. fracture and transform zones and diffuse/oblique/transtensional rift and ridge systems.

Our suggested scenario for the formation of the JMMC complements the established Wilson cycle concept. We propose that reactivation and the petrological variation of inherited structures of different ages, coupled with changes in the regional/global stress regime, controlled microplate formation in the following sequence of events (see also Fig. 6).

(1) Early Paleocene. Rifting propagates from the Central Atlantic into the Labrador Sea–Baffin Bay rift system (Roest & Srivastava 1989; Chalmers & Pulvertaft 2001; Peace *et al.* 2016)

(2) Early Eocene (Fig. 6b). Change in the Labrador Sea–Baffin Bay spreading direction from NW–SE to west–east (Abdelmalak *et al.* 2012) and the onset of seafloor spreading in the NE Atlantic (Gaina *et al.* 2009). This was possibly related to the far-field stress field applied by the collision of Africa and Europe (Nielsen *et al.* 2007) and/or to the relocation of the postulated Iceland plume (Skogseid *et al.* 2000; Nielsen *et al.* 2002).

(3) The NW–SE stress field in the North Atlantic between Greenland and Scandinavia would have favoured deformation on deep structures associated with the Iapetus Suture on the Norwegian margin rather than the East Greenland margin with the proposed fossil subduction zone (Fig. 2). Thus initial break-up is generally parallel to and in the vicinity of the Iapetus Suture.

(4) The IFAZ forms as the southern limit of the JMMC and may be linked to the localization of strain along the proposed fossil subduction zone or other potential rheological boundaries. No continental break-up occurred between Iceland and the Faroe Islands (IFR), with underlying uninterrupted, but thinned, continental lithosphere (Ellis & Stoker 2014).

(5) Mid-late Eocene. Accelerated extension occurred in the southern part of the JMMC and local reorganization of the Norway Basin spreading system (Gernigon *et al.* 2012, 2015) developed around 47 Ma (Fig. 6c) A first phase of magmatism between Greenland and the proto-JMMC was initiated (Tegner *et al.* 2008; Larsen *et al.* 2014). In the southern JMMC, isolated spreading cells possibly developed before the steady-state development of the Kolbeinsey Ridge.

(6) Late Eocene–early Oligocene (Fig. 6c). A major plate tectonic reorganization, including a change from NW–SE to NE–SW plate motion, coincident with the abandonment of seafloor spreading along the Labrador Sea–Baffin Bay system and consequent end of the anticlockwise rotation of Greenland (Mosar *et al.* 2002; Gaina *et al.* 2009; Oakey & Chalmers 2012). This change in plate motion resulted in deformation along the fracture zones and transpression on the IFAZ.

(7) Locking of the IFAZ triggered continental break-up between Greenland and the proto-JMMC subsequent to continental rifting between them. This is consistent with the microplate model of Whittaker *et al.* (2016) for the Indian Ocean. Rotational rifting between Greenland and the proto-JMMC started much earlier (*c.* 48–47 Ma) than the abandonment of the Labrador Sea–Baffin Bay spreading system (*c.* 40 Ma) and break-up between Greenland and the JMMC (33–24 Ma).

(8) Ultra-slow spreading continued on the Aegir Ridge after *c.* 31 Ma (Mosar *et al.* 2002; Gaina *et al.* 2009; Gernigon *et al.* 2015), while drastic rifting and possible embryonic spreading developed south of the proto-JMMC until steady-state spreading along the Kolbeinsey Ridge was completely established at 24 Ma (Vogt *et al.* 1970; Doré *et al.* 2008; Gernigon *et al.* 2012).

(9) The Aegir Ridge was abandoned, with all plate motion accommodated by the Kolbeinsey Ridge after 24 Ma, separating the proto-JMMC

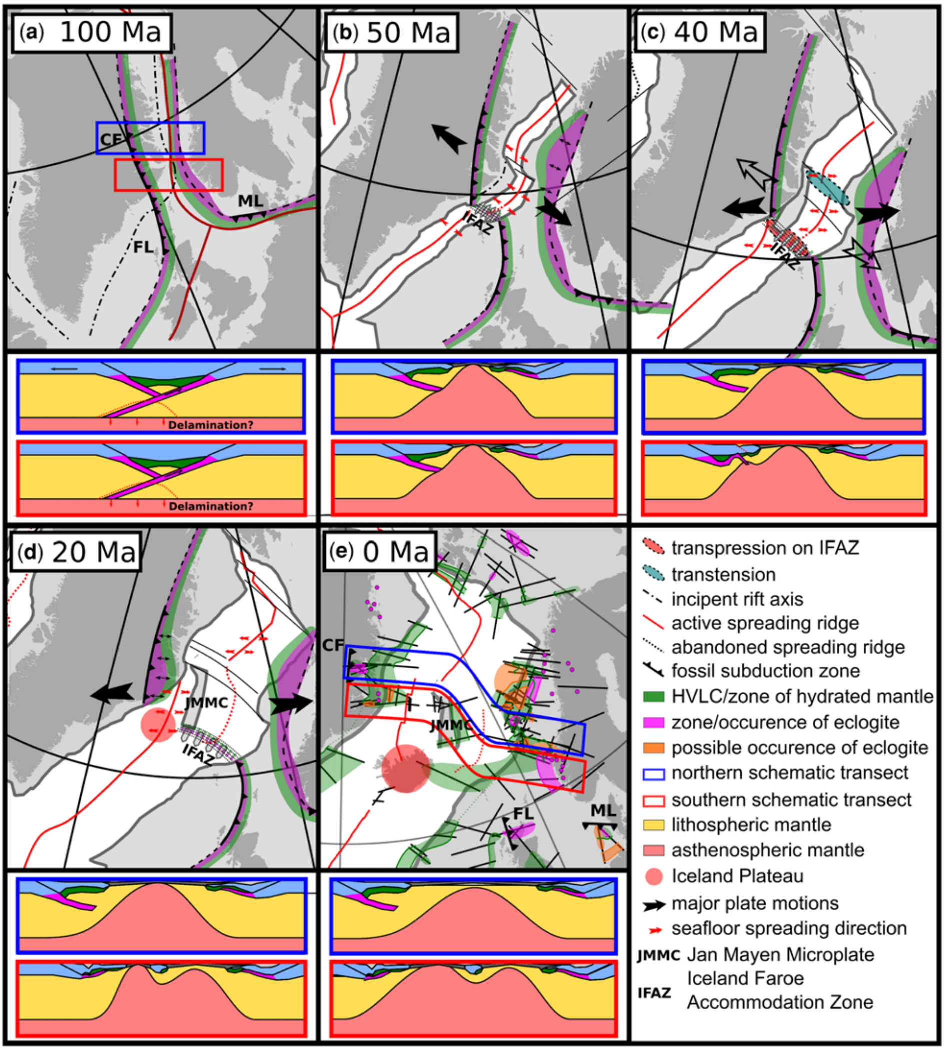


Fig. 6. Separation of the Jan Mayen microplate complex from Greenland. Palaeogeographic reconstructions from Seton *et al.* (2012). 100 Ma: the Caledonian Orogen experienced extensional collapse and multiple rift phases. Fossil subduction zones are still preserved, although possibly deformed. 50 Ma: seafloor spreading in the North Atlantic separates Greenland from Europe with NW–SE plate motions; break-up in the NE Atlantic occurs along the Iapetus Suture, which deforms. 40 Ma: plate motions change from NW–SE to west–east, which causes transpression on the Iceland–Faroe accommodation zone. The Reykjanes Ridge spreading centre develops towards the north, following lithospheric weaknesses along the East Greenland fossil subduction zone. 20 Ma: the newly formed Kolbeinsey Ridge is almost entirely developed, separating the Jan Mayen microplate complex from Greenland. The fossil subduction zone in Central East Greenland is highly deformed, whereas it is mainly preserved further north. The Aegir Ridge is successively abandoned. 0 Ma: fossil subduction zones are still preserved in East Greenland, northern Scotland and the Danish North Sea sector (CF, Central Fjord structure; FL, Flannan structure; ML, Mona Lisa structure). The fossil subduction zone has been destroyed and deformed in Norway and south-central East Greenland. It now forms high seismic velocity lower crustal bodies that are possibly eclogite high-velocity, lower crustal bodies (mapped in magenta and orange). HVLCB, high-velocity, lower crustal body; IFAZ, Iceland–Faroe accommodation zone; JMMC, Jan Mayen microplate complex.

from East Greenland (Fig. 6d). The West JMFZ, the eastern branch of which had already been established during the opening of the Norway Basin, then connected the Kolbeinsey Ridge with the Mohns Ridge north of the JMMC.

Summary

We propose a new model for the formation of a microplate complex as an extension to the established Wilson cycle concept. The new model invokes the rejuvenation of major pre-existing structures by plate-driven processes controlling both continental break-up and the formation of the JMMC.

The initial axis of continental break-up exploited the lithospheric weaknesses associated with the Iapetus Suture (Fig. 6a, b). These structures were particularly susceptible to deformation due to their preferential orientation with respect to the NW–SE to west–east-oriented extensional stress field. Fracture zones and strike-slip/oblique zones of deformation delineate the later-forming JMMC. The IFAZ represents one of these zones and may have formed along an old subduction zone. The origin of the IFAZ remains poorly defined because of poor data coverage. However, it is likely that, despite extreme thinning of the continental lithosphere, no continental break-up occurred between the present day JMMC and the Faroe Islands (e.g. Gernigon *et al.* 2015; Blischke *et al.* 2017).

Our model predicts that, following a major change in the direction of extension coeval with the abandonment of the Labrador Sea–Baffin Bay oceanic spreading and transform system, oblique deformation occurred south of the proto-JMMC and along the poorly defined IFAZ (Fig. 6c). This caused further westward relocation of the spreading centre towards a fossil subduction zone where eclogite and, especially, weak inherited serpentinite accommodated the relocation and final development of the Kolbeinsey Ridge. Complete development of the Kolbeinsey Ridge resulted in the final separation of the proto-JMMC from East Greenland (Fig. 6d) and complete break-up of the North Atlantic.

Formation of the JMMC correlates with, and can be explained by, the rejuvenation of pre-existing structures of different ages. Oblique accommodation/deformation zones, including fracture zones, defined the extent of the JMMC along the spreading axis. This model provides a simple explanation for the formation of a microplate complex involving control by both plate tectonic processes and structural inheritance.

Further work and data acquisition are required to fully understand the nature and formation of the JMMC, Iceland and the IFR. All three components are intrinsically interlinked and essential in understanding the tectonic and magmatic evolution of the entire North Atlantic. Geophysical data are lacking, especially in the south of the JMMC, offshore NW Iceland, and between Iceland and the Faroe Islands. The most fundamental, and perhaps economically important, question is the extent of continental crust underlying this region, a question that may require additional marine surveys, the re-interpretation of geochemical data and further drilling and sampling in this area.

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