The Faroe-Shetland Basin: A regional perspective from the Palaeocene to the present day and its relationship to the opening of the North Atlantic Ocean.

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Abstract

The Faroe-Shetland Basin is located offshore NW Scotland on the SE margin of the Atlantic Ocean and comprises numerous sub-basins and intra-basin highs that are host to a number of significant hydrocarbon discoveries. The principal hydrocarbon discoveries are in Palaeocene–Eocene strata, though earlier ones are known, and therefore their existence is intimately linked to the opening and evolution of the North Atlantic from 54 Ma. The final rifting and separation of Greenland from Eurasia is commonly attributed to the arrival of a mantle plume which impacted beneath Greenland in the early Tertiary. Moreover, the ensuing plate separation is commonly described in terms of instantaneous unzipping of the North Atlantic, whereas in reality proto-plate boundaries were more diffuse during their inception, and the linked rift system, including connections with the Arctic, which we see today were not established until late Palaeogene–early Neogene time. From a regional analysis of ocean basin development, including the stratigraphic record on the adjacent continental margins, the significance of the Greenland–Iceland–Faroe Ridge, and the age and role of Iceland, we propose a dual rift model whereby North Atlantic breakup was only partial until the Oligo-Miocene, with true final breakup only being achieved when the Reykjanes and Kolbeinsey ridges became linked. As final breakup coincides with the appearance of Iceland, this model negates the need for a plume to develop the North Atlantic with rifting reliant on purely plate tectonic mechanisms, lithospheric thinning and variable de-compressive upper mantle melt along the rifts.

It is a generally accepted model that North Atlantic seafloor spreading began at 54 Ma and rifted Greenland from Eurasia and that this process continues to the present day (Pitman and Talwani 1972, Srivastava and Tapscott 1986; Fig. 1). Prior to this final rift, the North Atlantic margins had also been subjected to several earlier phases of extension between the Devonian collapse of the Caledonian Orogen and the early Tertiary break-up (Ziegler 1988, Doré et al. 1999, Roberts et al. 1999). These pre-Palaeocene rifting phases exploited the collapsed Caledonian fold belt and consequently the conjugate North Atlantic margins show similar basinward stepping rift patterns (Doré et al. 1999). The final rifting and separation of Greenland from Eurasia at 54 Ma has also been attributed to the arrival and impact of an upwelling mantle plume which impinged on the base of the lithosphere under Greenland in the early Tertiary (e.g. White 1989, White and McKenzie 1989, Smallwood et al. 1999, Smallwood and White 2002). Its modern day much reduced expression is the Iceland plume of many authors (e.g. Courtillot et al. 2003). A plume here is defined as a convective upwelling of lower mantle material due to thermal instability near the core-mantle boundary (Morgan 1971).

It is interpreted the impact of the plume head with early Tertiary rifting led to large volumes of surface and subsurface magmatic activity, including the generation of seaward dipping reflectors and the linear and abnormally thickened (up to circa 30 km) Greenland-Faroes Ridge (the GFR of Holbrook et al. 2001; the Greenland-Iceland-Faroe Ridge and GIFR of Gaina et al. 2009; Fig 1). Calculated melt volume is estimated at 5-10 x 10^6 km3 and led to the North Atlantic being referred to as the North Atlantic Igneous Province (NAIP) (White 1988, Saunders et al. 1997). Production of this melt was not uniform, with initially high production and seafloor spreading at 54 Ma declining to almost zero between 35 -25 Ma, before North Atlantic plate re-organization re-established a slow but steady spreading rate up to the present day (Fig. 2).

This instantaneous unzipping model for the North Atlantic has had a major impact on regional post Cretaceous North Atlantic plate tectonic and palaeo-geographic reconstructions, basin modeling and consequently the hydrocarbon exploration in the offshore UK, Ireland and Norway (e.g. Larsen and Watt 1985, Roberts et al. 1999, Knott et al. 1999, Holmes et al. 1999 and Carr and Scotchman 2003). The instantaneous unzipping of the North Atlantic is also partially predicated on plate tectonic modeling packages which require 100% defined plate boundaries in order to operate (e.g. Skogseid et al. 2000, Gaina et al. 2009). In reality the proto-plate boundaries were probably more diffuse during their inception and more analogous to the NE African Rift data coming out of Ethiopia (Ferguson et al. 2010, Beutal et al. 2010, Rychert et al. 2012).

As authors we use published data and interpretations to challenge the above model and suggest there is more evidence to suggest that the North Atlantic did not unzip in one event at 54 Ma, or that it was driven by a plume process. The plume engine model has also recently been challenged by others (e.g. Foulger 2002, Foulger and Anderson 2005, Foulger 2010, Lundin and Doré 2005a, 2005b, Gernigon et al. 2009 and Gaina et al. 2009). On the basis of the evidence presented in this paper, we propose that the North Atlantic opening was the result of the development of opposing rifts, one developing SW from the north Atlantic (between Greenland and Norway) and the other NE from the central Atlantic (between Ireland and SE Greenland). These changed in importance and evolved through time with plate tectonic development of the North Atlantic from the Palaeocene to the present.

Initial development and dating of the North Atlantic oceanic crust

Plate reconstructions and their development through time, like those proposed for the North Atlantic, are constrained by dated sequences of magneto-chron’s resident in the spreading oceanic crust. It is generally accepted that North Atlantic oceanic crust first began forming during Chron 24 (54.5 Ma, Fig. 2). Gaina et al. (2009) using modern gravity and magnetic data with recent regional seismic and quantitative kinematic analysis published a map of North Atlantic magneto-chron’s and their confidence in their location and recognition (Fig. 3). There are two key observations which can be made from this data. Firstly, the chron’s defining the initial margins of the oceanic crust are not well constrained along the margins of the North Atlantic (Figs. 3 and 4). For example in some areas adjacent to the proposed Continent-Ocean Boundary (COB) they exhibit a patchy magnetic pattern not typical of thru going magnetic seafloor striping (Figs. 3 and 4). Similar anomalies have been recognized in the Labrador Sea and have been attributed to highly intruded or extended continental crust (Chalmers and Pulvertaft 2001). This implies at break-up, the crust was rifting analogous to the separation of inter-digitating ‘fingers’, with areas of incipient oceanic crust adjacent to deforming and intruded continental crust. A uniform, instantaneous separation with general full production of oceanic crust along the entire length of the Atlantic rift therefore appears unrealistic. Similar observations have been observed along the active Ethiopian and Afar rifts where magmatism is confined to elongate and independent magma chambers with associated active dyke intrusions above. These active magmatic centres are separated by laterally intervening areas of tectonic quiescence (Beutal et al. 2010; Pagli et al. 2012; Rychert et al. 2012).

The first definitive and continuous magneto-chron implying extensive oceanic ridge production of basalts is Chron C21 (48 Ma), which is linked to the Aegir Ridge in the Norway Basin and to the Reykjanes Ridge in the Irminger - Iceland Basins SE of Greenland (Figs. 1 and 3). This chron does not link across the GIFR and provides the second major observation, that the GIFR and the linear interval between the East and West Jan Mayen Fracture Zones (EJMFZ and WJMFZ) both exhibit patchy magnetic patterns along their entire length. Additionally they are defined by a thickened and linear crustal signature (Holbrook et al. 2001; Gernigon et al. 2009) (Fig. 3). Further interpretation of these is discussed below.

Evolution and dating of the oceanic ridges separating the North Atlantic Province

Evolution of the North Atlantic from the Late Palaeocene to the present day involved initial spreading in the Labrador Sea, along the Reykjanes Ridge, along the Aegir Ridge and then finally along the Kolbeinsey Ridge. The Oligo-Miocene linking of the Reykjanes and Kolbeinsey Ridges resulted in the separation of the Jan Mayen micro-continent from E.Greenland.

The spreading in the Labrador Sea, in conjunction with early rifting along the Reykjanes ridge, created a triple junction to the SE of Greenland which allowed simultaneous spreading along both rift arms (Gaina et al. 2002; Lundin and Doré 2005a; Gaina et al. 2009) (Fig. 1). The Atlantic spreading arm utilized the collapsed Caledonian fold belt with its associated Mesozoic rift system (Lundin and Doré 2005a). This dual rifting involving the Labrador Sea and the Reykjanes rift continued between 54 Ma and 33 Ma (Chron 24-Chron 13) when rifting then ceased in the Labrador Sea. Abandonment of the Labrador Sea and Baffin Bay spreading ridge resulted in a North Atlantic plate re-organization and the beginning of the next phase of North Atlantic opening (Roest and Srivastava 1989, Gaina et al. 2002, Lundin and Doré 2005a, Gaina et al. 2009). This coincides with Chron 13 (33 Ma) and represents a significant change in relative motion between Greenland and Eurasia from NW-SE to NE-SW (Gaina et al. 2009). Prior to Chron 13 (33 Ma), the northeast propagation of the Reykjanes Ridge failed to penetrate any further north than the Kangerlussuaq area in E Greenland. Successive, but un-productive attempts to propagate N–NE led to the gradual peeling away of the Jan Mayen micro-continent from E Greenland (Gaina et al., 2009; Fig.5).

The eastern margin of the Jan Mayen micro-continent is defined by the first appearance of the oldest oceanic crust (Magnetic anomaly 24b, 54 Ma) which erupted from the Aegir oceanic spreading ridge (Gaina et al. 2009). The western margin is similarly defined by the first appearance of true oceanic crust represented by magneto-chron 6c/7 (24 Ma) and erupted from the currently active Kolbeinsey oceanic spreading ridge. This indicates abandonment of the Aegir Ridge and the new propagating linkage of the Kolbeinsey and Reykjanes ridges in the Oligo-Miocene, the final separation of Jan

Mayen from E Greenland, and the destruction of any vestiges of the America-Eurasia land bridge (Gaina et al. 2009; Fig.5). Onshore in E Greenland this final break-up between E Greenland and Jan Mayen is evidenced by alkaline dykes and large syenite intrusives in the Scoresby Sund-Jameson Land and Hold with Hope-Clavering Ø areas (Price et al., 1997). Abandonment of the Aegir Ridge in favour of the Kolbeinsey Ridge may have been via ridge 'jump' to the new ridge (Talwani and Eldholm 1977) or by via gradual abandonment of the Aegir Ridge (e.g. Nunns 1983; Múller et al. 2001).

The linkage between the final stages of continental rifting and oceanization is uncertain, however, the work of Van Wijk et al. (2001) offers a useful insight. They used a dynamic 2-D finite element model for the upper and lower crust and the mantle to a depth of 120 km. This was to study melt generation in a rifting environment. Initial un-stretched thickened continental crust of 38 km was extended until lithospheric breakup resulted in a zone ≈ 150 km wide and where $\beta \geq 5$. This occurred 15 million years after stretching was initiated. Outside of this zone, transition zones up to 200 km wide (where $\beta = 1.2 – 5$) were developed, including an outer crustal area where thinning factors were less than 1.2 and extension had minimal effect. The stretching factors mirrored those proposed by Reemst and Cloetingh (2000) and Skogseid et al. (2001) for transects described along the North Atlantic volcanic margin. Results showed that de-compressional partial melting took place across a 175 km wide zone and at a calculated depth of melt production between 20 and 50 km. The melt initiation began 5 million years before breakup, with maximum melt just before breakup. Melt volumes fell in the outer edges of the transition zones where lower extension rates existed. They concluded the calculated melt volumes were in agreement with melt volumes 'per unit length of margin' observed along current day volcanic margins. This therefore did not require the prerequisite of a mantle plume in the North Atlantic and lithospheric rifting alone could produce the enhanced melt volumes observed along the volcanic margins.

Additional evidence from the Afar proto-plate boundary also suggests it involves shallow de-compressive melting of the upper mantle with vertical and lateral dyke intrusion within the rift (Ferguson et al. 2010; Rychert et al. 2012). If continental rifting involves segmented de-compressive melting along the rift, the magmatic composition and volume, the eruption site timing and location and rift margin uplift will vary spatially with underlying composition of the upper mantle (see Meyer et al. 2007 for documentation of the geochemistry and spatial distribution of varying melts across the entire NAIP, both from enriched and depleted upper mantle sources and with or without crustal contamination).

The Greenland-Iceland-Faroe Ridge (GIFR)

The North Atlantic oceanic province has two anomalously thickened crustal elements: 1) the NW-SE Greenland-Iceland-Faroe ridge (up to 30+ km of basaltic crust); and, 2) the similarly orientated and thickened Vøring Spur located within the Jan Mayen fracture zone (thickened basaltic crust up to ≥ 15 km) (Fig. 1). Their linearity and non-radial orientation away from any proposed plume centre poses problems for a plume origin.

The GIFR has been drilled by DSDP site 336 which was located on its northern flank (Fig. 1). This site penetrated Middle Eocene basalts dated 43-40 Ma (K-Ar date) at 515m below seabed (Talwani et al. 1976). The basalt grades into and is overlain by 8m of volcanic rubble (conglomerate), which in turn is overlain by 13m of thick red claystone. The latter is interpreted as a lateritic soil formed in situ by sub-aerial weathering of the basalt basement. The palaeosol is overlain by 295m of Middle Eocene to Upper Oligocene marine mudstones, the bulk of which is probably of Mid–Late Eocene age (Stoker and Varming, 2011). Micro-palaeontological evidence indicates that submergence of the GIFR at site 336, from a sub-aerial setting to a marine bathyl environment (shelf to upper slope), occurred during the Late Eocene (Talwani et al. 1976, Berggren and Schnitker 1983). However the crest of the GIFR, sited about 400m higher than the sea bed at site 336 (which itself is 463m above the palaeosol), suggests that the GIFR remained as either a continuous ridge or a string of closely spaced islands for some considerable time, possibly into the Oligo-Miocene (Talwani et al. 1976; Stoker and Varming 2011). Confirmation of this comes from evidence of the continued movement of flora and fauna from America to Eurasia during the Eocene-Miocene via the 'Thulian land bridge'. Beard (2008) records that the minute monkey-like *Teilhardina magnoliana* had migrated from China via the Bering land bridge to the coastal plain of Mississippi by 55 Ma and then further migrated to Europe by 47 Ma, as indicated by its discovery in Eocene deposits at Dormaal, Belgium (Fig. 6). Xiang et al. (2005) also noted that cornelian cherries (Cornus, cf. dogwoods) were present in North America during the Palaeocene, but had later spread to Europe and Africa by the Miocene and Denk et al. (2010) studying Miocene Icelandic oak pollen (Quercus) indicated linkage to the North American white and red oaks and their arrival in Iceland via a land bridge.

A model for the development of the GIFR (and the Jan Mayen Fracture Zone)

So what is the nature and derivation of the GIFR and when did it finally cease to be a land bridge between America and Eurasia? From Figure 3 it can be seen that the GIFR is linear and symmetric about Iceland, which led Morgan (1971) and others (e.g. White 1989; Smallwood et al. 1999; Skogseid et al. 2000) to suggest it represented the Palaeocene to recent plume track of the Iceland 'hotspot'. However, Lundin and Doré (2002, 2005b) and Foulger (2010) both note that the GIFR is not time transgressive in one direction, as the Iceland 'hotspot' has never been positioned below the Iceland-Faroes side of the GIFR (otherwise the plume head would be located under NW Scotland today). Lundin and Doré (2005b) state that lithospheric drift over a fixed mantle plume for the GIFR is untenable, or would require – in the extreme – an early plume capture at the plate boundary and subsequent plume drift to have exactly matched the lithospheric drift. This would allow the 'hotspot' to remain constantly centered on the spreading ridge. Some authors (e.g. Bott 1983; White and McKenzie 1989; Smallwood et al. 1999) suggest that the generation of oceanic crust > 7 km thick requires anomalously high asthenospheric temperatures (a mantle plume), whereas others advocate de-compressive lower temperature melt of a fertile upper mantle (Foulger and Anderson 2005; Gernigon et al. 2009).

The radial plume model also fails to explain the extreme linearity of the GIFR and a clue to its formation might be explained with reference to the recent study of the Vøring Spur, within the Jan Mayen fracture zone, by Gernigon et al. 2009 (Fig. 7). Combining Bouguer anomaly analysis with depth to MOHO estimation they note that the Vøring Spur between the East and West Jan Mayen fracture zones (JMFZ) is characterized by a Bouguer 'low', in contrast to adjacent oceanic domains, and coincides with thickened (>15 km) oceanic crust. Gernigon et al. (2009) propose that the thickened oceanic crust formed within the JMFZ during syn-rift extension across the NW-SE fractures which led in turn to lithospheric thinning and subsequent de-compressive melting of the mantle during the Mid–Late Eocene (‘overcrusting’). Similar NW-SE fracture zones have been documented along the entire North Atlantic margin (Rumph et al.1993), and specifically within the Faroe Islands (Ellis et al. 2009) and in the Kangerlussuaq area of SE Greenland (Larsen and Whitham 2005; Guarnieri 2011). Plate reconstruction of the North Atlantic prior to 55 Ma places the Faroe Islands some 50-100 km SE of Kangerlussuaq in the Late Palaeocene (Larsen et al. 1999; Skogseid et al. 2002; Larsen and Whitham 2005). This allows direct alignment of the NW-SE fracture zones described in the Faroes and Kangerlussuaq area; thus we propose that the GIFR was formed by a similar process to that described by Gernigon et al (2009) for the Vøring Spur (Fig. 8). We suggest transtensional movement across the Faroes-Kangerlussuaq lineaments during the Late Palaeocene–Early Eocene caused a zone of linear rifting, lithospheric thinning and decompressive melting of the upper mantle with increased magmatic activity to create the crustally thickened GIFR. Large Late Palaeocene–Early Eocene intrusives have also been documented in close association with the Faroes and Kangerlussuaq lineaments (Ellis et al. 2009; Guarnieri 2011). Walker et al. (2011) noted the first phase of faulting in the Faroes involved dip-slip movements along NW-SE and N-S orientated faults.

The orientation of these NW–SE lineaments probably represents inheritance from an earlier failed transient North Atlantic rift along a NW–SE axis from Baffin Island to the British Isles (Lundin and Doré 2005a). This coincided with the Early Palaeocene igneous activity (62-58 Ma) in the British NAIP, as represented by igneous centres, such as Skye, Rhum and Mull, and the NW–SE orientated dyke swarms which extend across the UK, from the Hebrides into the central North Sea and Lundy in the Bristol Channel (e.g. Brown et al. 2009; Hansen et al. 2009). However, the early Tertiary NAIP igneous activity has also been recently linked with the northern influence of the ‘Large Low Velocity Shear Velocity Province’ (previously called the ‘African Plume’) at the core-mantle boundary and located beneath central Africa (Torsvik et al. 2006; Ganerød et al 2010).

Eocene development of the Atlantic margin of Britain and Ireland

Recent re-evaluation of the Eocene sequences in the central North Atlantic, involving seismic, wells, shallow boreholes, core data, biostratigraphy and onshore geology has concluded that throughout the Eocene, the northern North Atlantic rift was not connected to the southwest North Atlantic rift (Stoker and Varming 2011; Stoker et al. 2013a, b) (Fig. 9). Eocene prograding deltaic (and fan-delta) deposits can be mapped offloading from the Munkagrunnur and Wyville Thompson ridges, and the West Shetland margin into the Faroe-Shetland Basin (Robinson 2004; Ólavsdóttir et al. 2010, 2013; Stoker and Varming 2011; Stoker et al. 2013b) (Fig. 9). To the southwest, stratigraphically equivalent rocks were deposited on the flanks of the Rockall and Hatton basins (Stoker et al. 2012) and Porcupine Basins (Moore and Shannon 1992; Fig. 9). These deposits are contemporary with similar deposits which accumulated in the North Sea (Jones and Milton 1994; Mudge and Bujak 1994) and onshore E Greenland (Larsen and Whitham 2005; Larsen et al. 2005).

On the eastern flank of the Rockall High, the inter-digitation of sub-aerial volcanic lavas and fan delta/shallow marine deposits in BGS borehole 94/3 can be tied to regionally synchronous unconformities across the greater Rockall–Hatton area, as well as in the Faroe–Shetland region and E Greenland (Stoker et al. 2012). Seismic mapping of these units in the Rockall–Hatton area indicates an archipelago of Eocene islands at or near sea level, frequently inundated, but capable of supplying sediment into the local area (Stoker et al. 2012; their Fig. 10). The intra-Eocene unconformities imply a fluctuating response to relatively small sea level rise and falls associated with the evolving North Atlantic plate tectonic regime (Shannon et al. 1993; Stoker and Varming 2011). This includes the post-depositional (latest Eocene/Oligocene) tilting of a number of the Eocene prograding units located on the flanks of major compressional folds, such as the Hatton High and the Wyville Thomson Ridge (Stoker et al. 2012).

Palaeocene–Eocene sequences offshore and onshore Southeast and East Greenland

On the opposite side of the rift and offshore SE Greenland, DSDP borehole 917 drilled 775m of crustally contaminated basalt with sub-aerial weathered horizons (Vallier et al. 1998, Fitton et al. 2000) (Fig. 1). In the borehole, the basalts are underlain by 10 cm of non-metamorphosed, quartzitic sandstone (interpreted as of fluvial origin) and of presumed Palaeocene age, which in turn is underlain by 15.9m of fine-grained metamorphosed sediments (Greenschist grade), including volcaniclastics. These are of presumed Late Cretaceous age and exhibit weak sedimentary structures suggestive of deposition from turbidity currents.

Onshore in SE Greenland, Upper Cretaceous and Early Palaeocene shallow-marine sediments in the Kangerlussauq area are overlain by Late Palaeocene shallow-marine and fluvial deposits (Fig.1). These in turn pass upwards and inter-digitate with, and ultimately inundated by, Upper Palaeocene–Lower Eocene flood basalts (Larsen and Whitham 2005). The depositional setting was strongly controlled by large syn-depositional NW–SE oblique-slip normal faults trending along the Nansen fjord and the Christian IV glacial valley, as well as the nearby Kangerlussauq fjord (Larsen and Whitham 2005; Guarnieri 2011). These faults also controlled the injection sites and the later deformation of early Eocene intrusives (Guarnieri 2011). A subsidiary NE–SW fault inland from the coast, the Sortekap fault, further constrained the Palaeocene–Early Eocene depositional system in the area (Larsen and Whitham 2005). Palaeocurrent evidence from the shallow-marine and fluvial sequences both indicate initial flow to the southeast, but then critically to the southwest in the centre of the basin (Larsen and Whitham 2005). The SE margin of the basin is not preserved.

Overlying the Lower Palaeocene sediments in Kangerlussuaq are a thick succession of Upper Palaeocene–Lower Eocene sub-aerial basalts, which were emplaced from east to west and buried a significant Late Palaeocene landscape, including deeply incised palaeo-valleys (Larsen et al. 1989; Pedersen et al. 1997). The basalts range up to 6 km thick on the coast and thin to 2–3 km inland and regionally northwards to about 1 km in the Trail Ø area (Price et al. 1997). The E Greenland sequence contains age equivalent basalts of the Beinisfjord, Malinstindur and Enni Formations, which have been described from the Faroe Islands, and which are now located between Iceland and NW Scotland (Passey and Jolley 2009) (Fig. 1). Additionally the Kangerlussuaq area has younger Eocene lava formations, the Upper Geikie, Rømer, Skrænterne and Igtertiva Formations, which are not present in the Faroe Islands (Pedersen et al. 1997; Larsen et al. 1989; Larsen et al. 1999) (Fig 10). Recent dating of the youngest Kangerlussuaq volcanic sequence, the Igtertiva Formation in the Kap Dalton area, suggests eruption into the late Early Eocene between 49–47.9 Ma (Larsen et al. 2005). The significance of this is that offshore the Faroe Islands, the equivalent rocks are represented by NE-seaward-dipping reflectors and oceanic crust, whereas in SE Greenland they comprise east-to-west prograding sub-aerial basalts. This apparent contradiction is addressed below.

In SE Greenland, Middle Eocene fluvial and shallow-marine sediments overlie the Igtertiva basalts in the Kap Dalton area and demonstrate a NE-to-SW progradational sequence, as identified by palaeocurrent data (Larsen et al. 2005). This is the opposite direction to the Early Eocene landscape developed in the Faroe-Shetland Basin, which was from S to N (Larsen et al. 2005; Stoker and Varming 2011).

The 'Dual Rift' model

On the basis of the above data, we propose that in Late Palaeocene-Early Eocene time two active rift systems were trying to split America and Greenland from Eurasia until 33 Ma (Chron 13) (Fig.11). The first was developing SW-to-NE along the SE margin of Greenland and along the line of the proto-Reykjanes Ridge. It penetrated into the area of Kangerlussauq and possibly into the Trail Ø area where it was unable to penetrate further northwards (Figs. 9 and 11). A second NE-to-SW propagating rift, represented by the Aegir spreading ridge, developed from the Norwegian-Greenland Sea and extended into the area N of the Faroe Islands. The two rifts were independent of each other and did not conjoin, leaving Jan Mayen firmly attached to E Greenland in the Kangerlussuaq and Hold-with-Hope area, with the Faroe region as the centre of a land bridge between East Greenland and Europe. The land bridge lasted until at least 33 Ma (and possibly until 25 Ma?). The opposed ridge system has been previously cited by others, though the timing and the ridges involved has varied (see Dewey and Windley 1988, Lundin et al. 2002, Lundin and Doré 2005a, Gaina et al. 2009). The dual ridge system can explain the difference in timing and nature of the flood basalt sequences in the Kangerlussuaq rift which were dominantly sub-aerial and emplaced E to W, as against the synchronous development of seaward dipping reflectors and early oceanic crust from the propagating Aegir rift north of the Faroe Islands (and E of Jan Mayen). The lateral by-passing of the two rift tips would induce severe fracturing in the area of the palaeo-Faroe Island location and induce anticlockwise rotation of the regional stress field and thus allow igneous material to erupt from local volcanic centres, particularly along NW–SE and NE–SW orientated fractures. The anticlockwise rotation of the regional stress field has been observed from Palaeocene-Early Eocene onshore fault analysis in the Faroe Islands (Walker et al. 2011) and later during the Oligo-Miocene separation of Jan Mayen from Greenland (Gaina et al., 2009) (Fig. 5).

Post-Palaeocene to Oligo-Miocene inversion events in the North Atlantic province

The counter-clockwise rotation of the regional stress field from 54–25 Ma could also partially explain the episodic and gradual inversion of anticlinal and domal structures in the North Atlantic throughout the Eocene, and particularly during the Oligo-Miocene (Johnson et al. 2005; Doré et al. 2008; Ritchie et al. 2008). In the Faroe–Shetland region, the disposition of the Eocene succession is folded about the axes of uplift which form the bathymetric highs of the Fugloy, Munkagrunnur and Wyville Thomson Ridges. This implies that a major phase of ocean margin structuration took place during the late Palaeogene–early Neogene (Johnson et al. 2005; Stoker et al. 2005b; Ritchie et al. 2008; Ólavsdóttir et al. 2010). The Hatton High (of the Rockall Plateau) was also folded in this interval as witnessed by the tilting of the Eocene deltaic wedges on this feature (Stoker et al. 2012).

A major consequence of Oligo–Miocene compressional deformation in the Faroe–Rockall region was the formation of the Faroe Conduit (combining the present-day Faroe-Shetland and Faroe Bank channels), which today forms the main passageway for deep-water exchange across the Greenland–Scotland Ridge (Stoker et al. 2005a, b). This early Neogene instigation of the gateway is consistent with ^{13}C, ^{18}O, taxonomic and sedimentological data from DSDP and ODP sites in the North Atlantic region, which reveal a Miocene instigation for the overflow of North Atlantic deep water into the previously isolated Tethyan (warm water) Reykjanes rift basins to the south (cf. Stoker et al. 2005a, b and references therein).

Hot fluid flow evidence coincident with the two major tectonic phases of evolution of the North Atlantic (Eocene and Oligo-Miocene plate rifting and reorganization)

Microthermometry and apatite fission track analysis (AFTA) from the Faroe-Shetland Basin and East Greenland indicates two periods of hot fluid flow through post-Palaeocene units until the present: a Late Palaeocene–Early Eocene event, and another around 20-25 Ma (Parnell et al. 1999; Scotchman et al. 2006; Parnell and Middleton 2009). The first is coincident with the early North Atlantic rifting phase, and the second with the major Oligo-Miocene plate reorganization of the North Atlantic, which led to the separation of Jan Mayen from Greenland, and North Atlantic regional uplift. Both phases are associated with the highest ocean spreading rates and therefore elevated heat flows (Fig. 2). An omni-present plume from the Paleocene to the present cannot rationally explain these two pulsed hot fluid flows in itself, or the fact that despite these fluid flushes the regional vitrinite reflectance and AFTA data suggest low thermal maturity for most of the Mesozoic–Cenozoic sequences in the Faroe-Shetland Basin and onshore E Greenland (except when in close proximity to intrusive bodies) (Surlyk et al. 1986; Stemmerick et al 1992; Scotchman et al. 2006). The AFTA and vitrinite reflectance data therefore suggest a lower regional thermal gradient more indicative of shallow melts restricted to plate boundaries and not the higher gradient expected from a large-radius mantle derived plume impacting across the base of the lithosphere.

The 'Dual Rift' Model and Palaeocene–Early Eocene sediment provenance studies

The dual rift model may also explain the dichotomy expressed in published North Atlantic Palaeocene reservoir provenance studies (e.g. Jolley and Morton 2007; Morton et al. 2012a, b). These studies involved heavy mineral and palynological analysis to identify several Scottish source areas for Palaeocene sandstones in the Faroe-Shetland Basin; they also identified a 'westerly' sourced sandstone component (the FSP3 zircon signatures of Jolley and Morton 2007 and Morton et al. 2012a, b).

The palynological data suggests a Greenland affinity, but the heavy mineral rutile/zircon data, though distinguishing a non-Scottish source, cannot distinctly type the 'western' source area. In the single rift model the zircons and other heavy minerals from SE Greenland should be observed in the Faroe-Shetland Basin which was directly SE of Kangerlussuaq in palaeo-reconstructions (e.g. Larsen et al. 1999; Skogseid et al. 2000; Larsen and Whitham 2005). In particular, the zircon age data profile of westerly FSP3 sandstones does not match surface sampling of Kangerlussuaq basement terrains (Morton et al. 2012b). The Kangerlussuaq low-RuZi sandstones have two main age groups at ~2700–2750 Ma and ~2950–3100 Ma, with a subsidiary peak at ~3200 Ma. Over 50% of the

Archaean grains are older than 2950 Ma (Whitham et al., 2004). The ~2950–3100 Ma and ~3200 Ma age peaks are either absent or very poorly developed in the FSP3 sandstones, and zircons older than 2950 Ma form only 5–11% of the Archaean population. Derivation of the FSP3 sandstones from the Kangerlussuaq area of E Greenland is therefore dismissed (Morton et al. 2012a and 2012b).

With the currently proposed synchronous dual rift model, any westerly-sourced sediments from Kangerlussauq would have entered the northerly progressing rift off SE Greenland and been unavailable to the Faroe-Shetland Basin (Figs. 9 and 11). The alternative locations for westerly derived FSP3 sandstones may have been derived from the now-rifted Jan Mayen micro-continent or other areas in NE-Central Greenland (which are presently below the ice-cap). Westerly derived America–Greenland pollen spores can, in contrast, be common to both rifts due to their less dense silt size and their potential for greater aqueous and regional wind-blown distribution.

What is the age and role of Iceland?

With the progressive dual rift model we now consider the age and role of Iceland. Plume protagonists propose that Iceland is the present signature of a northern hemisphere deep mantle plume which impacted the area of NW Greenland in the Late Cretaceous–Early Palaeocene, and then migrated via Kangerlussuaq with time and Greenland plate movement to its current Iceland position (e.g. White 1989; White and McKenzie 1989; Smallwood et al. 1999; Smallwood and White 2002). Other authors disagree and propose non-plume origins for Iceland (e.g. Foulger 2002; Foulger and Anderson 2005; Lundin and Doré 2005a, b; Gernigon et al. 2009; Gaina et al. 2009; Foulger 2010).

In the context of the dual propagating rift model and the potential final separation of America–Greenland from Eurasia in the Oligo-Miocene, we note the following pertinent facts regarding Iceland:

- The oldest dated outcrops are 17 Ma (Miocene) in NW Iceland and 13 Ma in E Iceland (Moorbath et al. 1968; Ross & Mussett 1976; Har∂arson et al. 1997).
- The time transgressive V-shaped ridges, extending up to 1000km along the Reykjanes and Kolbeinsey Ridges, are limited to Oligocene to Recent oceanic crust (Jones et al. 2002). Hey et al. 2010 and Benediktsdóttir et al. 2012 indicate that the V-shaped ridges are not symmetrical about the Reykjanes Ridge axis and this implies formation from either a 'pulsating plume' (the conventional view), or – in their preferred view – the requirement of a simple rift propagation away from central Iceland. They also note the V-shaped ridges have different geographic extent and patterns N and S of Iceland and this asymmetry is best explained by rift propagation, and not by pulses in a symmetrically radial plume.
- Iceland is dominated by tholeiite basalts (typical Mid-Ocean Ridge Basalts) and not picrites or komatiites which are more distinctive of hotter core-mantle origins (Foulger et al. 2003; Presnall 2003; Meyer et al. 2007, Søager and Holm 2011). Lower mantle picrites are predicted to have higher MgO concentrations around 15–30% (Gill et al. 1992; Lundin and Doré 2005a). Iceland basalts have +/- 10.5% Mg and are highly depleted with respect to major and trace element composition and therefore could originate from lower temperatures.
- A larger-than-expected proportion (cf 10%) of compositionally acidic (rhyolite) and intermediate (andesitic) rocks occur in Iceland (Foulger and Anderson 2005; Foulger 2010).

- Mesozoic-age zircons (representing continental crust derivation) have been described from Mount Hvitserker in NE Iceland (Paquette et al., 2006, Paquette et al. 2007) and from more widespread Icelandic locations (Bindeman et al. 2012).
- Records of elevated $^{87}Sr/^{86}SR$ and Pb ratios in rhyolites and basalts in SE Iceland are indicative of an enriched upper mantle source (Prestvik et al. 2001; Meyer et al. 2007, Søager and Holm 2011). Storey et al. (2004) also describe similar enriched Middle Miocene lavas in E Greenland.
- Tomographic studies show the low-velocity anomaly below Iceland extends only into the upper mantle (at +/- 400 km depth) and not to the core-mantle boundary. It is a vertical and cylindrical anomaly of 200-250 km diameter at depths down to 200 km, but an elongate dyke like form at greater depths and parallel to the mid-Atlantic Ridge trend (Ritsema et al. 1999; Foulger et al. 2001; Montelli et al. 2003; Hung et al. 2004). The latter suggests a closer relationship to the greater regional morphology, shallow melts and tectonics of the mid-Atlantic rift system rather than a deep mantle plume.

We note the coincidence that the oldest rocks in Iceland post-date the break-away of Jan Mayen in the Oligo-Miocene (33-24 Ma) (Fig. 5) and the dating of the V-shaped ridges beginning in the Oligocene. Similarly, at the point of final break-up, Iceland would have been positioned where the new rift intersected the zone of the Caledonian suture between Scotland and E Greenland, a point noted by Foulger and Anderson (2005). The link to the underlying Caledonian orogen is further strengthened by the fact that the volcanism in Iceland is apparently from a shallow (de-compressive) melt and from an enriched (eclogitic) upper mantle and not a deep rooted plume. Thus, from all of the above, we propose that Iceland was initiated in the Oligo-Miocene at the point where the juncture of the Reykjanes and Kolbeinsey Ridges crossed the Scotland–Greenland Caledonian orogenic trend. This implies that an earlier proto-Iceland hotspot is not required to explain earlier Palaeocene–Eocene rifting and volcanism, and regional plate tectonics can provide a simpler explanation (e.g. van Wijk et al. 2001; Foulger and Anderson 2005; Lundin and Doré 2005a; Gaina et al. 2009). The evidence for an enriched upper mantle source for some Iceland basalts and rhyolites and the Middle Miocene basalts of E Greenland (Storey et al. 2004) can also be attributed to melting in the upper mantle of buoyant subducted Iapetus oceanic crust (Foulger and Anderson 2005; Bindeman et al. 2012), from a residual Caledonian orogenic root (Ryan and Dewey 1997), or from a metasomatised upper mantle (Storey et al. 2004).

The North Atlantic Geoid?

Evidence for the plume is often invoked from the observation of the positive geoid (c.60m) covering a large sector of the North Atlantic and NW Europe and the resultant low mean Atlantic water depths (Marquart 1991; Köhler, 2004; Lundin and Doré 2005b) (Fig.12). It is interesting to note that the maximum geoid anomaly is non-radial, but quasi-linear at its maximum between Iceland and the central Atlantic between Iberia and NewFoundland. This encompasses the modern central Atlantic spreading ridges and the area of the defunct Early Palaeocene triple junction SE of Greenland, the time when both the Labrador Sea and proto-Reykjanes spreading ridges were active.

Lundin and Doré (2005a, b) noted that the geoid anomaly has a 3000–4000 km diameter while mantle tele-seismic investigations around Iceland suggest only a 1000 km radius upper mantle thermal anomaly. This also is not centered above the 200–250 km wide low velocity zone underneath Iceland. If the geoid does represent the thermal effect of a plume from the Mid-Palaeocene to the present-day, then it is also significantly larger than the extent of the North Atlantic igneous province and cannot rationally explain the non-radial and highly disparate distribution of Palaeocene–Early Eocene magmatism in space and time (e.g. the W Greenland and Baffin Bay picrites vs the BTIP more normal basalts erupted around 58–62 Ma.).

Lundin and Doré (2005a, b) have demonstrated that there is not a one-to-one association of geoids to hotspots around the earth and therefore suggested that an alternative explanation should be investigated. We would speculate that the relevant timings and distribution of the rifting zones in the North Atlantic with their associated melts could have formed the large area of thickened North Atlantic basaltic crust. In particular the greatest elevation of the geoid coincides with the area of the palaeo-triple junction and the spreading of the central Atlantic ridges into the Labrador Sea and the proto-Reykjanes Ridge (Figs 1 and 12). Synchronous lateral heat flow from both the Labrador Sea and the proto-Reykjanes Ridge would have lifted Greenland at the expense of other areas and generated the 'fossil' centre of the of the geoid (Fig.12). Transient and synchronous uplift followed by subsidence in W and E Greenland has been documented by Dam et al. 1998, who ascribed it to the impact and then the lateral spreading of a plume head; however, it could equally be ascribed to synchronous rift flank uplift in the Labrador Sea and along the Reykjanes Ridge, followed by subsidence as full rifting occurred and magma was depleted from the upper mantle and erupted onto the rift margins. The location of Greenland away from the centre of the plate tectonic-induced geoid has since been changed by plate migration of Greenland from 62 Ma to the present day. Additionally the geoid has probably been re-enforced by crustal thickening under Europe and N Africa by the Pyrenean and Alpine orogenies between 62 Ma and 14 Ma. (Figs 2 and 12).

Thus, in our opinion, it was not until the Reykjanes and Kolbeinsey ridges crossed the Scotland–Greenland Caledonian orogenic track that Iceland and its associated smaller upper mantle anomaly come into being, thereby creating the more local geoid around it. The timing and the different interactions of the varying rift zones could also explain the disparate timing of post-Palaeocene to Recent uplift and subsidence of basins and continental margins around the North Atlantic, something not possible with an omni-present and radial plume (Lundin and Doré 2005b).

CONCLUSIONS

- On the basis of regional geological and other data, we suggest that the opening of the North Atlantic between Eurasia and North America–Greenland was only partial until the Oligo-Miocene (33–25 Ma). The true final break-up occurred when the Reykjanes and Kolbeinsey ridges conjoined in the area of SE Greenland and offshore Kangerlussauq.
- Initial attempts at North Atlantic rifting involved two opposing and almost by-passing rifts. These were the proto-Reykjanes Ridge (rift) system propagating NE along the SE Greenland margin and the SW propagating Aegir Ridge between Norway and offshore E Greenland. Although rifting along the Atlantic margin of Ireland, Britain and the Faroe

Islands may have been initiated at 54 Ma, true development of continuous oceanic crust in both rifts did not develop until Chron 21 (48 Ma).

- The two initial propagating rifts failed to join in the Palaeocene–Early Eocene in the palaeo-location of Kangerlussuaq and the Faroes Islands. The Jan Mayen micro-continent was still firmly attached to Greenland and continued to form a section of a land bridge between America–Greenland and Eurasia, via the volcanic Faroe Islands and into NW Britain. The attempt of the two rifts to by-pass each other in this area caused anti-clockwise rotation of the regional stress field.
- Oligo-Miocene North Atlantic plate reorganization, including the separation of Jan Mayen from SE Greenland, destroyed any relict Thulian land bridge and initiated the mixing of cold northern Atlantic waters with warmer southern Tethyan waters.
- Oligo-Miocene plate reorganization in the North Atlantic created the final break between Greenland and Europe and coincided with the appearance of Iceland and the production of the V-shaped seabed ridges to the north and south of the island.
- The dual rift model negates the need for a plume to develop the North Atlantic– the rifting can be wholly explained by plate tectonic mechanisms, lithospheric thinning and variable de-compressive upper mantle melting along the rifts. Recent studies from the Afar rift in Ethiopia has shown that de-compressive melt generation and dyke swarm propagation are more important than plume influence in the evolution of the proto-plate boundary (Ferguson et al. 2010; Rychert et al. 2012).
- The dual rift model with de-compressive upper mantle melts more closely confined to the plate tectonically induced rifts would imply a lower and more segmented regional heat-flow from 54 Ma to the present. This is in contrast to a higher regional heat-flow evoked by a large radius mantle plume. The implied lower heat-flow with its variable timing and geographical distribution will significantly change results from regional basin modeling studies and the type and timing of hydrocarbon generation around the North Atlantic, in contrast to results from a regional plume model with elevated heat-flow.

This paper reflects the researches and views of the authors and not necessarily those of Statoil or the British Geological Survey.

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Figure Captions

Fig.1. Topography and bathymetry of the North Atlantic with active plate boundaries (black lines; extinct ridges are dashed black lines). From north to south: MR, Mohns Ridge; KR, Kolbeinsey Ridge; RR, Rekjanes Ridge; AR, extinct Aegir Ridge; LXR, Labrador Sea extinct ridge; GRN, Greenland; EUR, Eurasia; GIFR, Greenland-Iceland-Faroe Ridge; JMFZ, Jan Mayen Fracture Zone; F, Faroe Islands; FSB, Faroe-Shetland Basins; KL, Kangerlussuaq. Open circles indicate sites of Deep Sea Drilling Project (DSDP) or Ocean Drilling Program (ODP) drilling. Small white dots indicate location of recent seismicity (from Gaina et al., 2009).

Fig.2. Regional Palaeogene–Neogene tectonostratigraphic framework of the NW European Atlantic margin. Sources: Doré et al. 1999; Doré et al. 2008; Johnson et al. 2005; Ritchie et al. 2008; Statoil UK Ltd. Timescale from Gradstein et al. 2004.

Fig.3. (**a**) Magnetic data interpretations in the NW Atlantic. Background image is the free air gravity anomaly (Sandwell and Smith 1997; Forsberg and Kenyon 2004). Seaward-dipping reflectors (SDRS) are shaded transparent light grey. The black line indicates the Continent-Ocean Boundary (COB). JMMC, Jan Mayen microcontinent; EJMFZ and WJMFZ, the E and W Jan Mayen Fracture Zones respectively. (Gaina et al. 2009). (**b**) Simplified NW Atlantic magnetic interpretation (after Gaina et al. 2009).

Fig.4. Geophysical responses across the Continent-Ocean Boundary in the region between Edoras Bank and the Faroe-Iceland Ridge. (**a**) Total magnetic field (very poor data in the greyed area); (**b**) isostatically corrected Bouger gravity anomaly; (**c**) selected magnetic and gravity features. Oceanic magnetic anomaly picks indicated by labels at their northeast ends. Black lines indicate the locations of the CAM77, RAPIDS, N18 and FIRE seismic profiles (Kimbell et al. 2005).

Fig.5. Evolution of the NW Atlantic plate boundaries and kinematic evolution of the Jan Mayen micro-continent illustrated by a series of tectonic reconstructions in an absolute reference frame (after Gaina et al. 2009). JMB, Jan Mayen Block; FB, Faroes Block. Note Jan Mayen peels away from SE Greenland between 20-30 Ma therefore allowing the potential for a long lived land bridge between America and Eurasia via the Greenland-Faroe Ridge.

Fig.6. The Eocene primate *Teilhardina* with fossil locations and their relative stratigraphic age superimposed on an Eocene palaeogeographic map of the globe. A North Atlantic land bridge allowed the primate to migrate from North America to Eurasia during the Eocene (Beard 2008).

Fig.7. (**a**) Bathymetric map and main physiographic features of the Norwegian-Greenland Sea. EJMFZ, CJMFZ and WJMFZ, the east, central and west Jan Mayen Fracture zones respectively. SDRS, Seaward-dipping Reflectors; GIFR, Greenland-Iceland-Faroe Ridge. JAS-05 is the gravity-magnetic survey area designation (**b**) Gravity forward modelling and crustal model across the Vøring Spur, including gravity and density modelling. The oceanic root, observed beneath the Vøring Spur, is interpreted as a syn-rift oceanic and mafic feature ('overcrusting') formed during the Mid-Late Eocene time (Gernigon et al. 2009).

Fig.8. Interpreted fracture lineaments connecting SE Greenland (Kangerlussuaq) and the Faroe-Shetland Basin (after Lundin and Dóre 2002; see also text for additional references). The origin of the Greenland-Iceland-Faroe Ridge is interpreted to be analogous to that proposed for the thickened crust observed between the Jan Mayen Fracture zones by Gernigon et al. 2009 (see Fig. 7).

Fig. 9. Generalised palinspastic map for the Late Palaeocene to Early Eocene interval (modified after Stoker and Varming 2011). ADS, Anton Dohrn Seamount; KB, Kangerlussuaq Basin; MB, Møre Basin, NRB, North Rockall Basin; NSB, North Sea Basin; PB, Porcupine Basin; RBS, Rosemary Bank Seamount; SRB, South Rockall Basin; WTR, Wyville Thompson Ridge; VB Vøring Basin.

Fig. 10. Early Eocene (54 Ma) schematic geologic cross-section from SE Greenland (Kangerlussuaq) and across to the Faroe Islands and the Faroe-Shetland Basin (FSB). Section is balanced on the regionally correlated Colsay 1 Sandstone Formation and is based on FSB well penetrations and outcrop studies in the Faroes and SE Greenland (see text for references).

Fig. 11. Summary diagram citing evidence for two NW Atlantic active rift systems attempting to split America and Greenland from Eurasia until 33 Ma (Chron 13). One developed SW to NE along the SE margin of Greenland and formed the line of the proto-Reykjanes Ridge. It penetrated into the Kangerlussauq area where it was unable to penetrate further. A second NE to SW propagating rift, represented by the Aegir spreading ridge, developed from the Norwegian-Greenland Sea and extended into the area north of the Faroe Islands. The two rifts were independent of each other and did not conjoin, leaving Jan Mayen firmly attached to E Greenland and the Faroes as the centre of a land bridge between East Greenland and Eurasia. The land bridge lasted until at least 33 Ma (and possibly until 25 Ma?). Abreviations : KL Kangerlussuaq; JM Jan Mayen; JMFZ Jan Mayen Fracture Zone; FSB Faroe-Shetland Basin; HB Hatton Bank; RB Rockall Bank.

Fig. 12. The North Atlantic geoid anomaly (from Köhler, 2004). The geoid anomaly coincides approximately with the extent of the North Atlantic topographic-bathymetric anomaly (Sandwell and Smith, 1997) and is not centred on Iceland and is more widespread than the upper mantle low-velocity anomaly (e.g. Ritsema et al. 1999; Foulger et al. 2001).

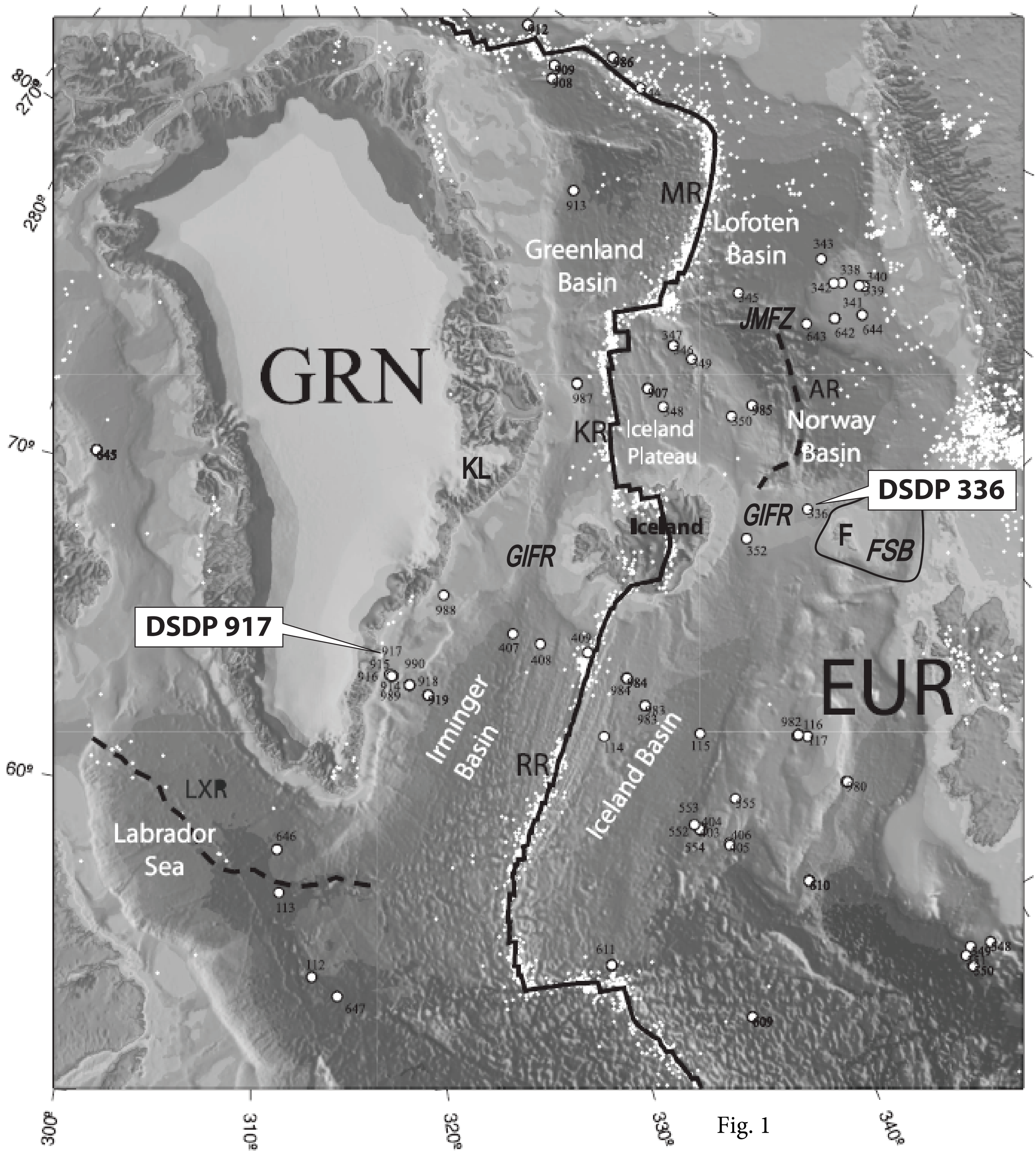
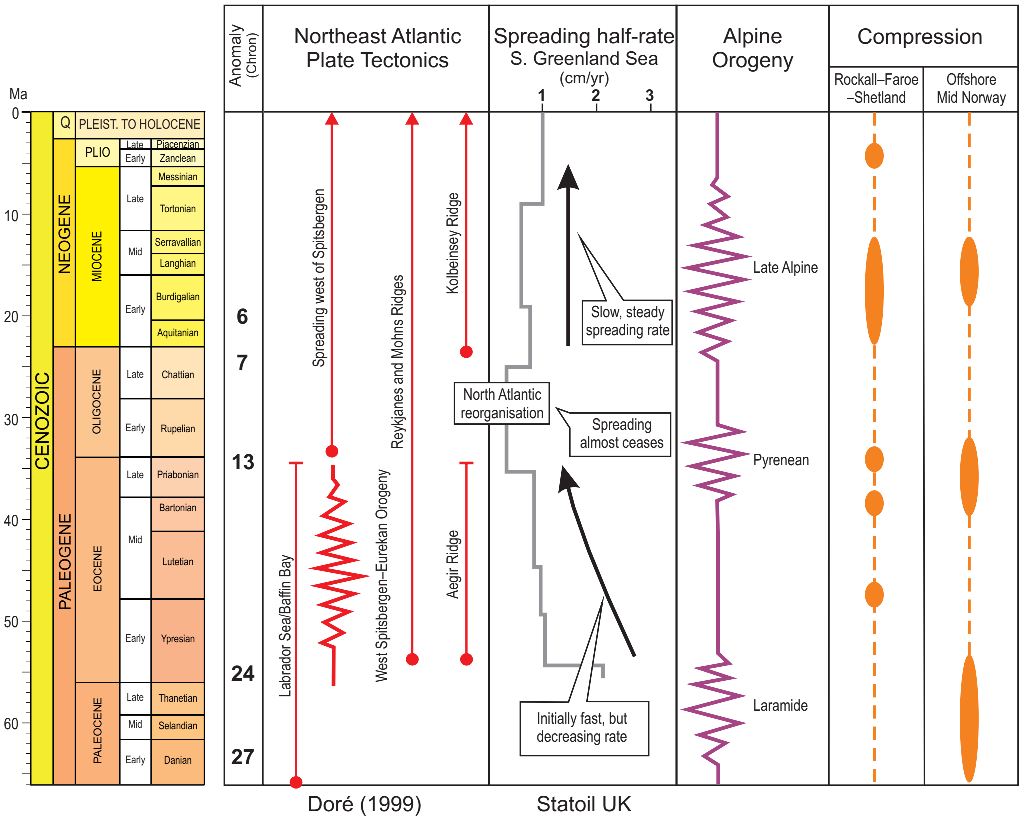

GRN
EUR
Greenland Basin
Lofoten Basin
MR
JMFZ
AR
Norway Basin
KR
Iceland Plateau
KL
Iceland
GIFR
GIFR
F
FSB
DSDP 336
DSDP 917
Irminger Basin
Iceland Basin
RR
LXR
Labrador Sea
300º
310º
320º
330º
340º
60º
70º
80º
270º
280º

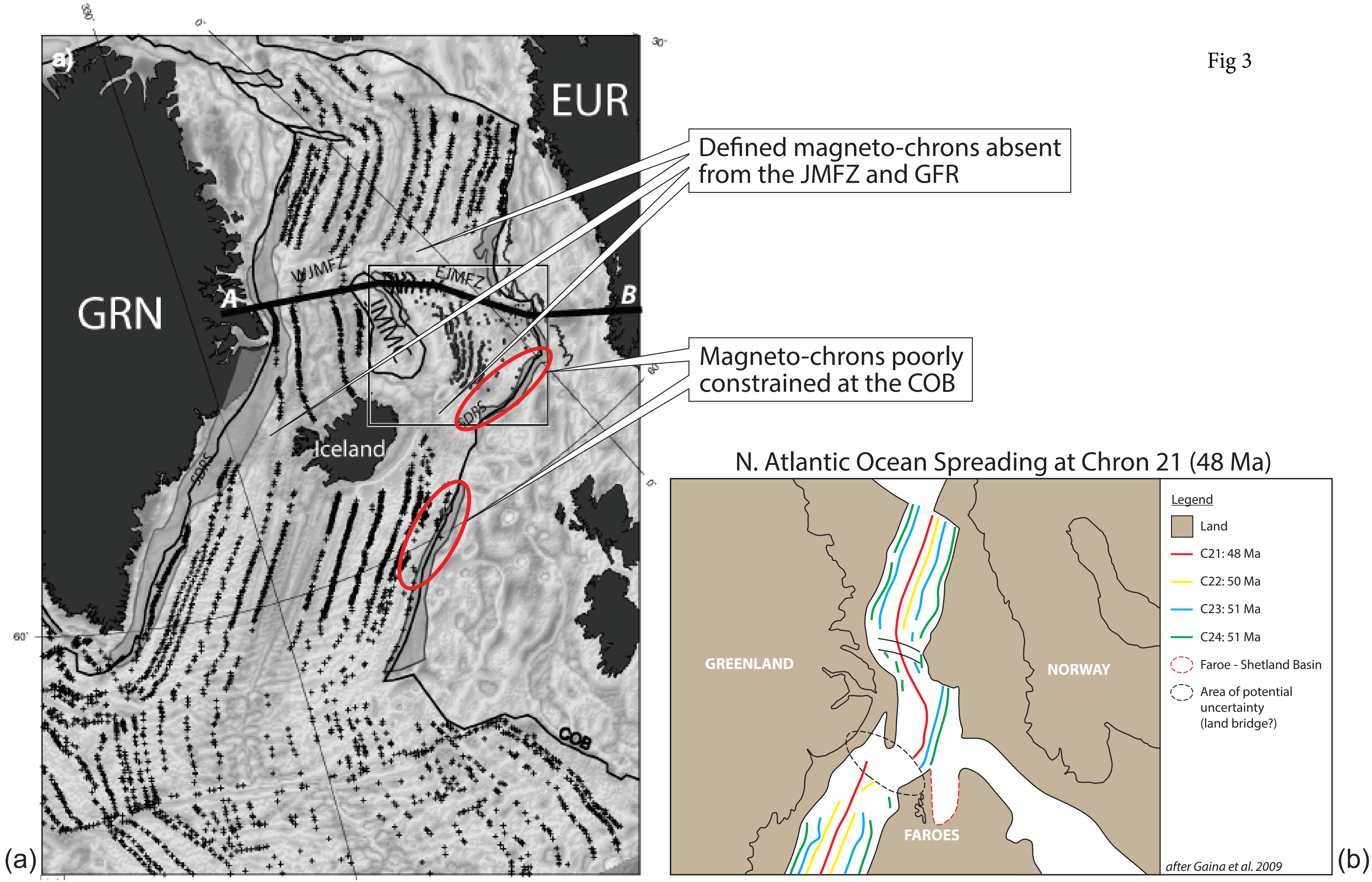
Fig. 1

Fig 2

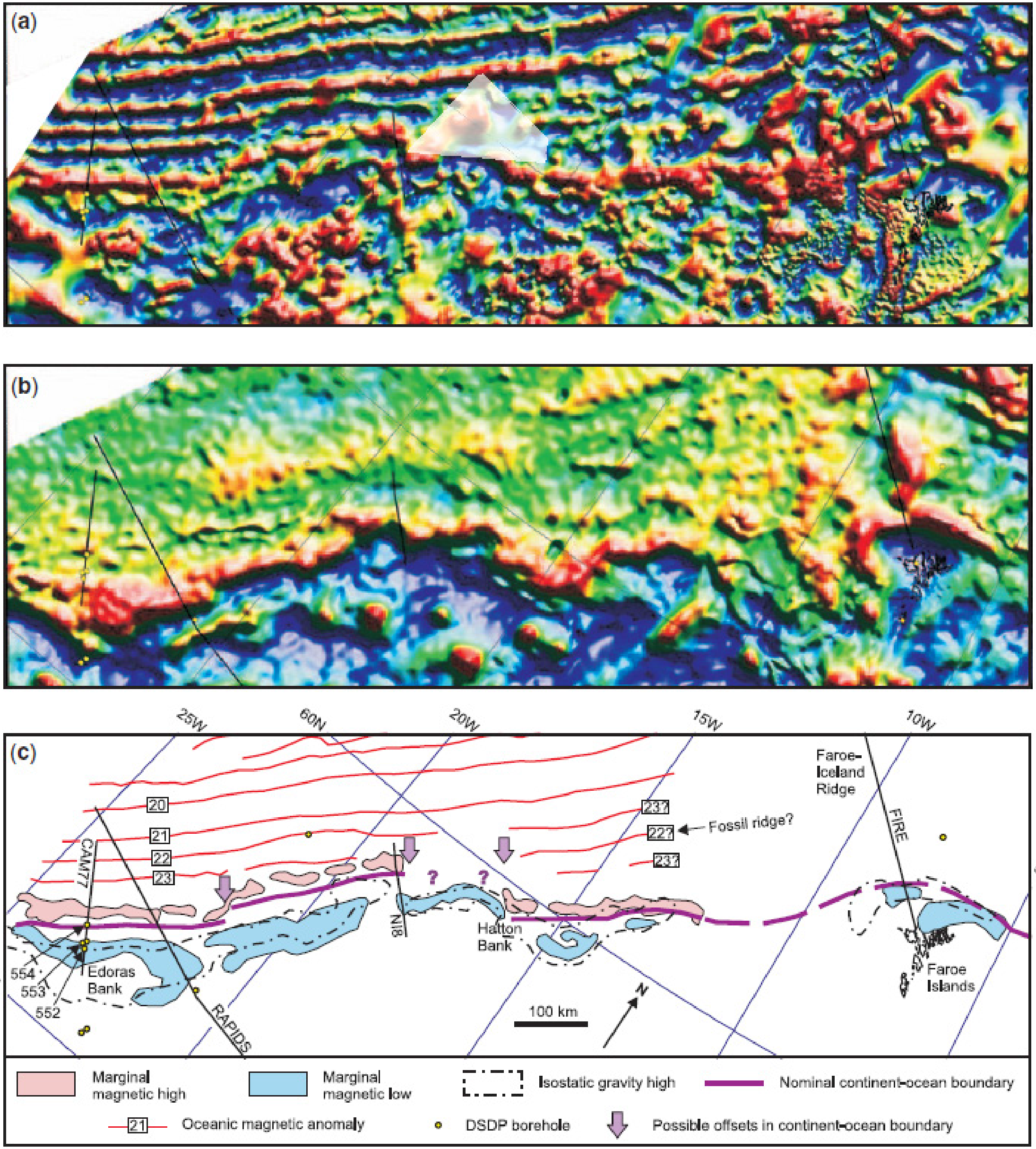


Ma
CENOZOIC
NEOGENE
PALEOGENE
Q
PLEIST. TO HOLOCENE
PLIO
MIOCENE
OLIGOCENE
EOCENE
PALEOCENE
Piacenzian
Zanclean
Messinian
Tortonian
Serravallian
Langhian
Burdigalian
Aquitanian
Chattian
Rupelian
Priabonian
Bartonian
Lutetian
Ypresian
Thanetian
Selandian
Danian
Anomaly (Chron)
Northeast Atlantic Plate Tectonics
Spreading west of Spitsbergen
Reykjanes and Mohns Ridges
Kolbeinsey Ridge
Labrador Sea/Baffin Bay
West Spitsbergen–Eurekan Orogeny
Aegir Ridge
Spreading half-rate S. Greenland Sea (cm/yr)
Slow, steady spreading rate
North Atlantic reorganisation
Spreading almost ceases
Initially fast, but decreasing rate
Alpine Orogeny
Late Alpine
Pyrenean
Laramide
Compression
Rockall–Faroe –Shetland
Offshore Mid Norway
Doré (1999)
Statoil UK

Fig 3



a)
EUR
GRN
Iceland
A
B
WJMFZ
EJMFZ
JMM
EDRS
COB
60°
30°
0°
Defined magneto-chrons absent from the JMFZ and GFR
Magneto-chrons poorly constrained at the COB
(a)
N. Atlantic Ocean Spreading at Chron 21 (48 Ma)
GREENLAND
NORWAY
FAROES
Legend
Land
C21: 48 Ma
C22: 50 Ma
C23: 51 Ma
C24: 51 Ma
Faroe - Shetland Basin
Area of potential uncertainty (land bridge?)
after Gaina et al. 2009
(b)


(a)
(b)
(c)
25W
60N
20W
15W
10W
5W
Faroe-
Iceland
Ridge
FIRE
20
21
22
23
23?
22?
23?
Fossil ridge?
CAM77
N18
?
?
Hatton
Bank
Edoras
Bank
554
553
552
RAPIDS
Faroe
Islands
100 km
N
Marginal
magnetic high
Marginal
magnetic low
Isostatic gravity high
Nominal continent-ocean boundary
21
Oceanic magnetic anomaly
DSDP borehole
Possible offsets in continent-ocean boundary

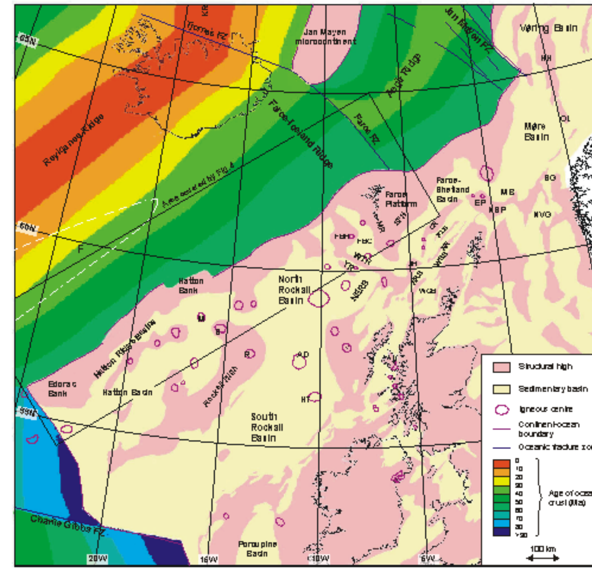

Jan Mayen
microcontinent
Vøring Basin
Reykjanes Ridge
Faroe-Iceland Ridge
Møre
Basin
Faroe
Platform
Hatton
Bank
North
Rockall
Basin
Edoras
Bank
Hatton Basin
South
Rockall
Basin
Porcupine
Basin
Charlie Gibbs FZ
Structural high
Sedimentary basin
Igneous centre
Continent-ocean
boundary
Oceanic fracture zone
Age of ocean
crust (Ma)
100 km

Fig 4

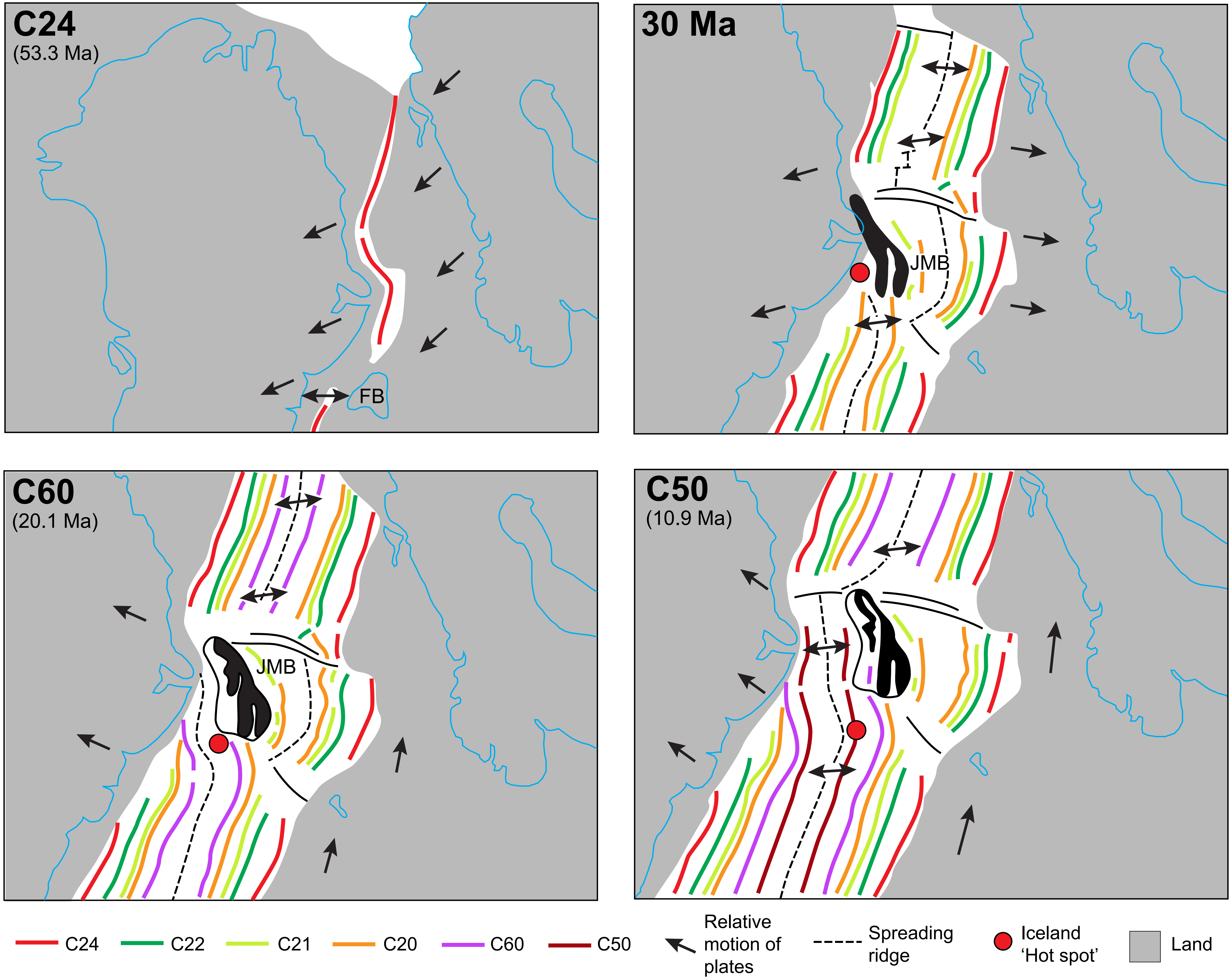

C24
(53.3 Ma)
FB
30 Ma
JMB
C60
(20.1 Ma)
JMB
C50
(10.9 Ma)
C24
C22
C21
C20
C60
C50
Relative motion of plates
Spreading ridge
Iceland 'Hot spot'
Land

Fig 6

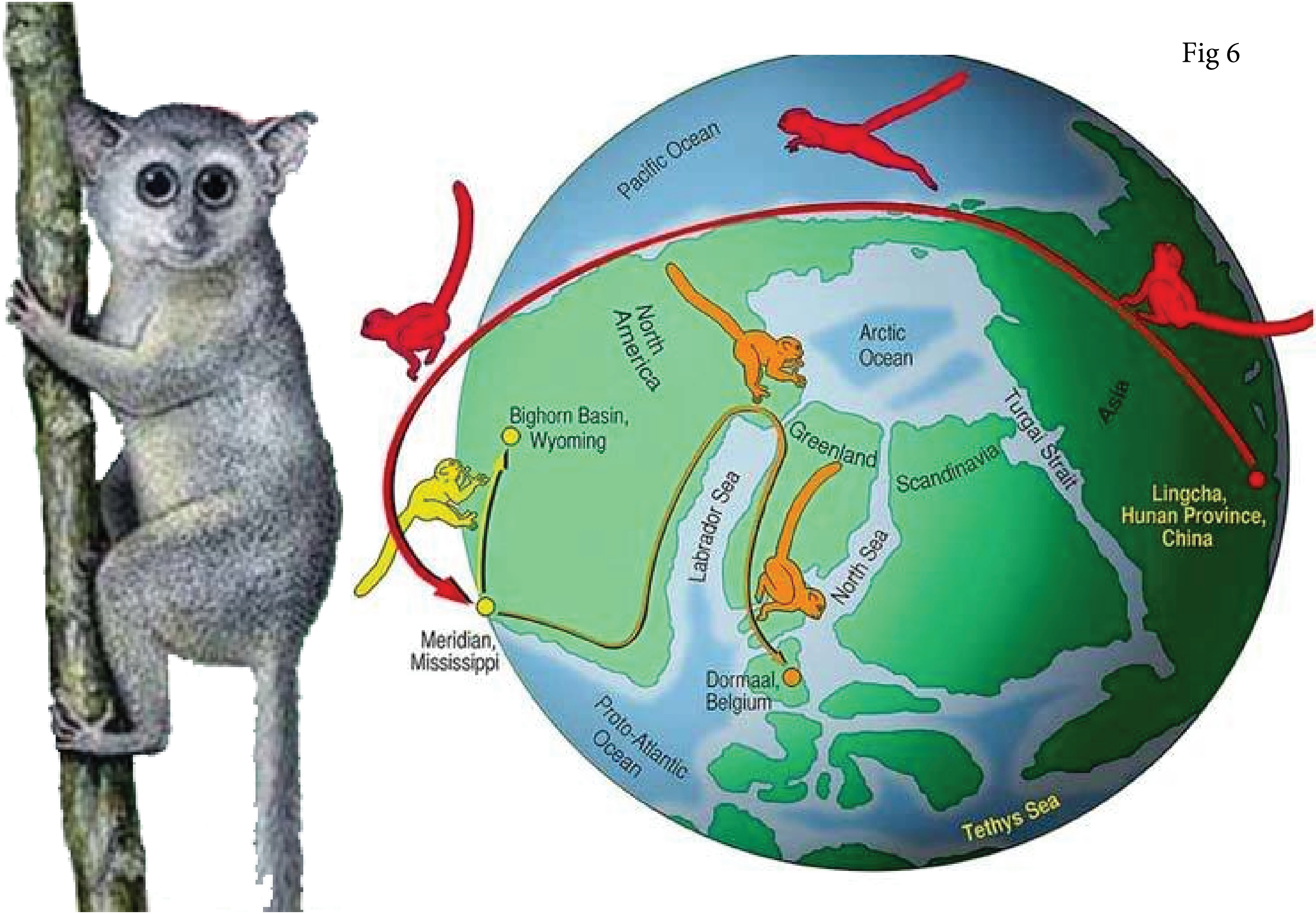
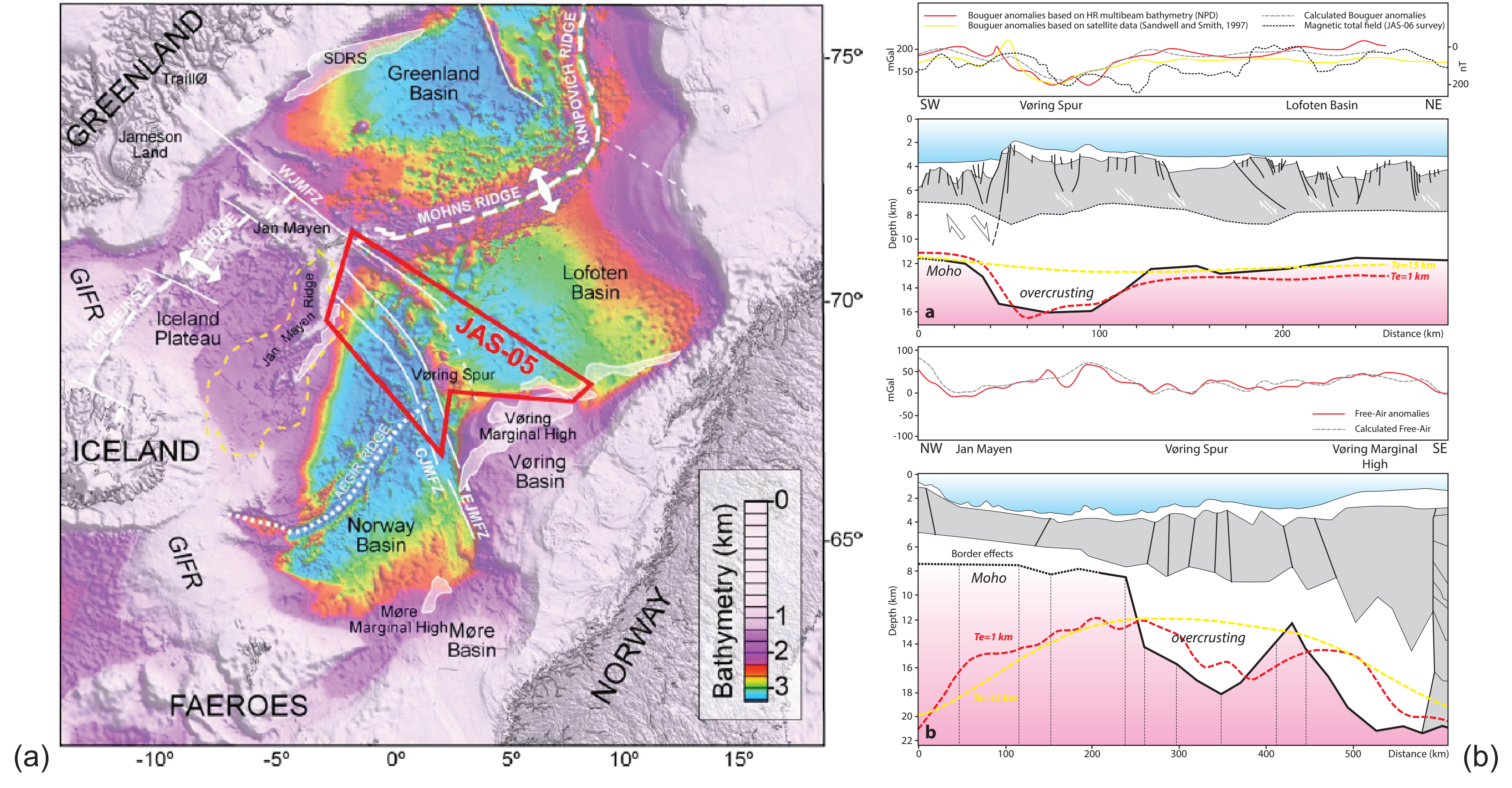

Pacific Ocean
North America
Arctic Ocean
Asia
Bighorn Basin, Wyoming
Greenland
Scandinavia
Turgai Strait
Labrador Sea
North Sea
Lingcha, Hunan Province, China
Meridian, Mississippi
Dormaal, Belgium
Proto-Atlantic Ocean
Tethys Sea

Fig 7



GREENLAND
TraillØ
Jameson Land
SDRS
Greenland Basin
KNIPOVICH RIDGE
MOHNS RIDGE
WJMFZ
Jan Mayen
RIDGE
KOLBEINSEY
GIFR
Iceland Plateau
Jan Mayen Ridge
Lofoten Basin
JAS-05
Vøring Spur
Vøring Marginal High
Vøring Basin
ICELAND
AEGIR RIDGE
CJMFZ
EJMFZ
Norway Basin
Møre Marginal High
Møre Basin
GIFR
FAEROES
NORWAY
Bathymetry (km)
0
1
2
3
-75°
-70°
-65°
-10°
-5°
0°
5°
10°
15°
(a)
Bouguer anomalies based on HR multibeam bathymetry (NPD)
Bouguer anomalies based on satellite data (Sandwell and Smith, 1997)
Calculated Bouguer anomalies
Magnetic total field (JAS-06 survey)
mGal
nT
SW
Vøring Spur
Lofoten Basin
NE
Depth (km)
Moho
overcrusting
Te=15 km
Te=1 km
a
Distance (km)
Free-Air anomalies
Calculated Free-Air
NW
Jan Mayen
Vøring Spur
Vøring Marginal High
SE
Border effects
Moho
Te=1 km
overcrusting
Te=15 km
b
Distance (km)
(b)

Fig 8

Lineament zone connects to Faroes at 54 Ma? (similar to JMFZ) (Extention between lineaments leads to decompressive melting and thickened crust between the lineaments)

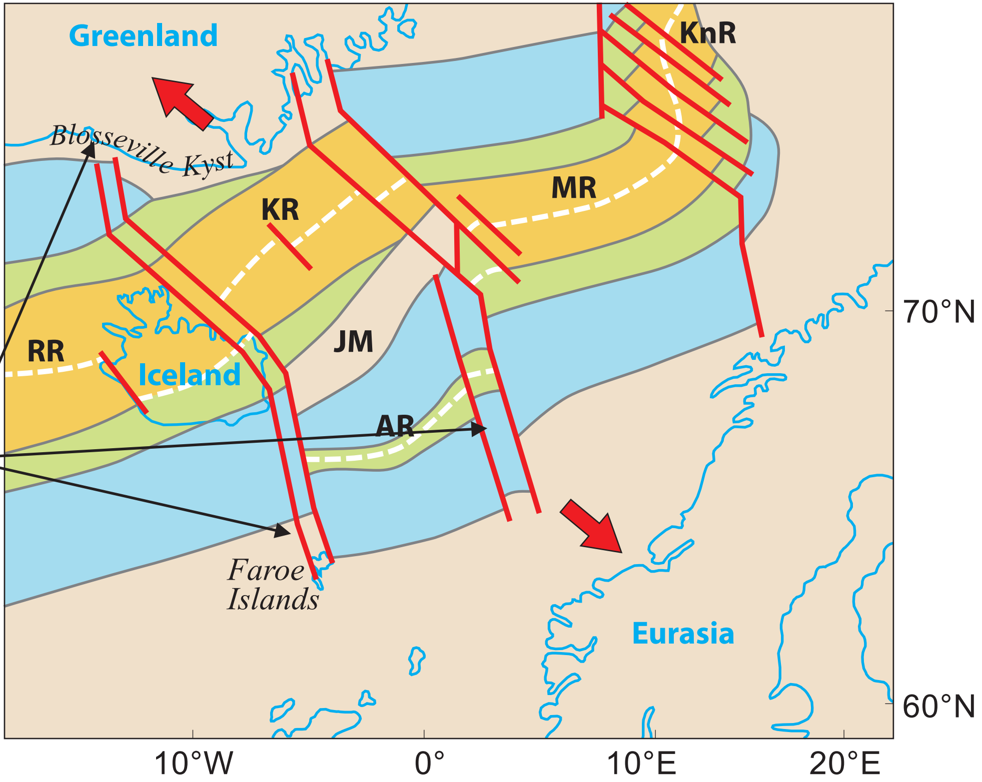
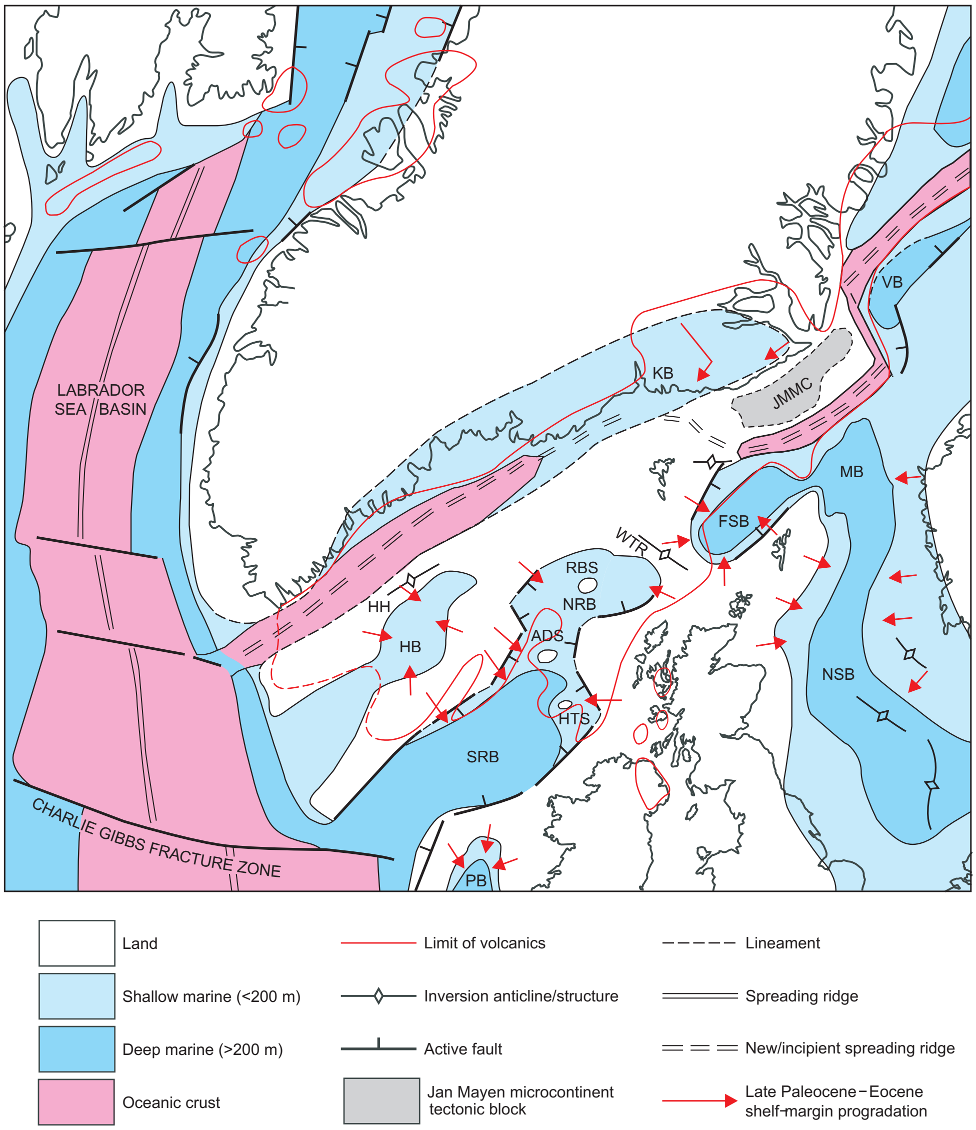
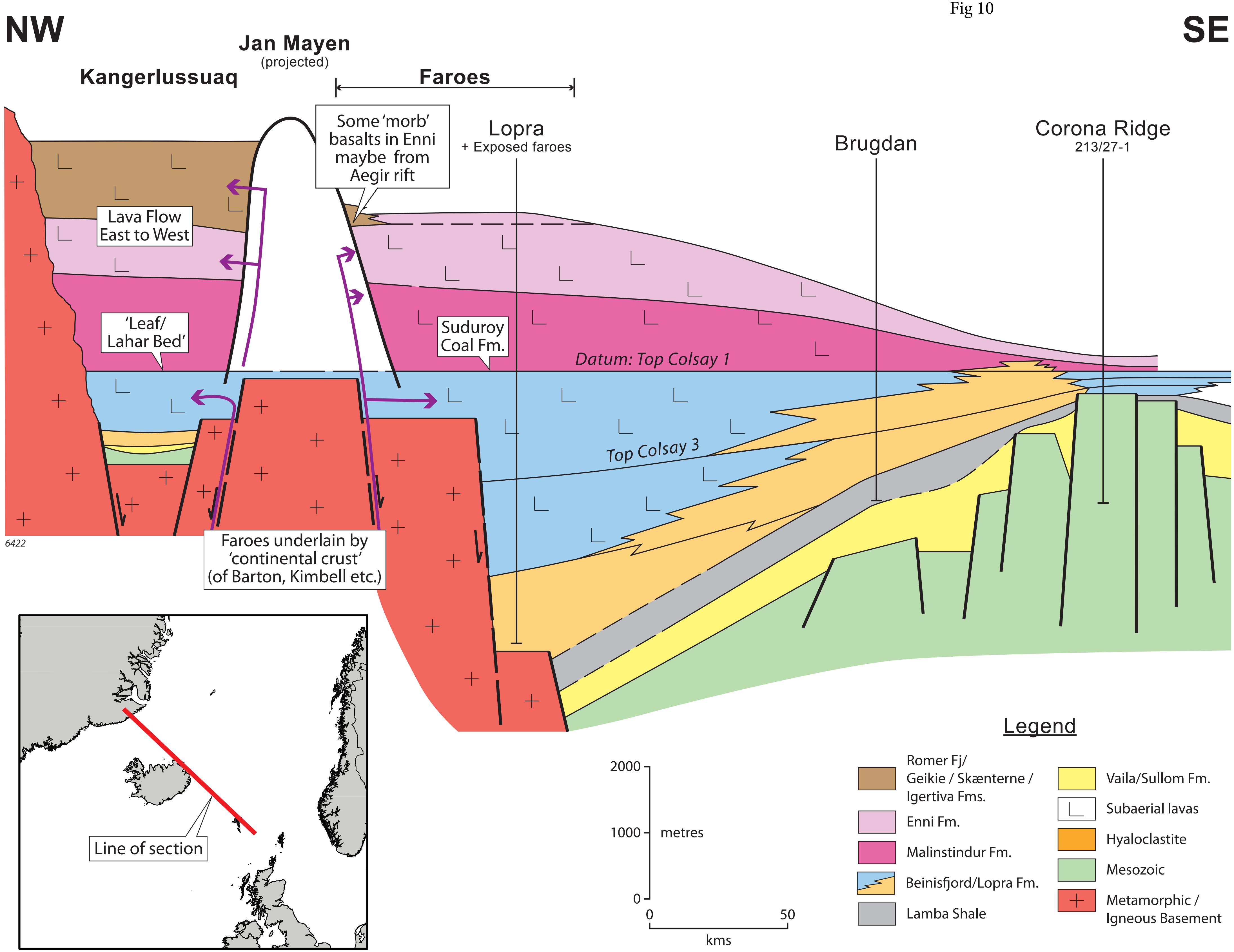


Fig 9



LABRADOR SEA BASIN
CHARLIE GIBBS FRACTURE ZONE
KB
JMMC
VB
MB
FSB
WTR
RBS
NRB
HH
HB
ADS
HTS
SRB
NSB
PB
Land
Shallow marine (<200 m)
Deep marine (>200 m)
Oceanic crust
Limit of volcanics
Inversion anticline/structure
Active fault
Jan Mayen microcontinent tectonic block
Lineament
Spreading ridge
New/incipient spreading ridge
Late Paleocene–Eocene shelf-margin progradation

Fig 10



NW
SE
Jan Mayen
(projected)
Kangerlussuaq
Faroes
Some 'morb' basalts in Enni maybe from Aegir rift
Lopra
+ Exposed faroes
Brugdan
Corona Ridge
213/27-1
Lava Flow East to West
'Leaf/ Lahar Bed'
Suduroy Coal Fm.
Datum: Top Colsay 1
Top Colsay 3
Faroes underlain by 'continental crust' (of Barton, Kimbell etc.)
6422
Line of section
2000
1000
0
metres
0
50
kms
Legend
Romer Fj/ Geikie / Skænterne / Igertiva Fms.
Enni Fm.
Malinstindur Fm.
Beinisfjord/Lopra Fm.
Lamba Shale
Vaila/Sullom Fm.
Subaerial lavas
Hyaloclastite
Mesozoic
Metamorphic / Igneous Basement

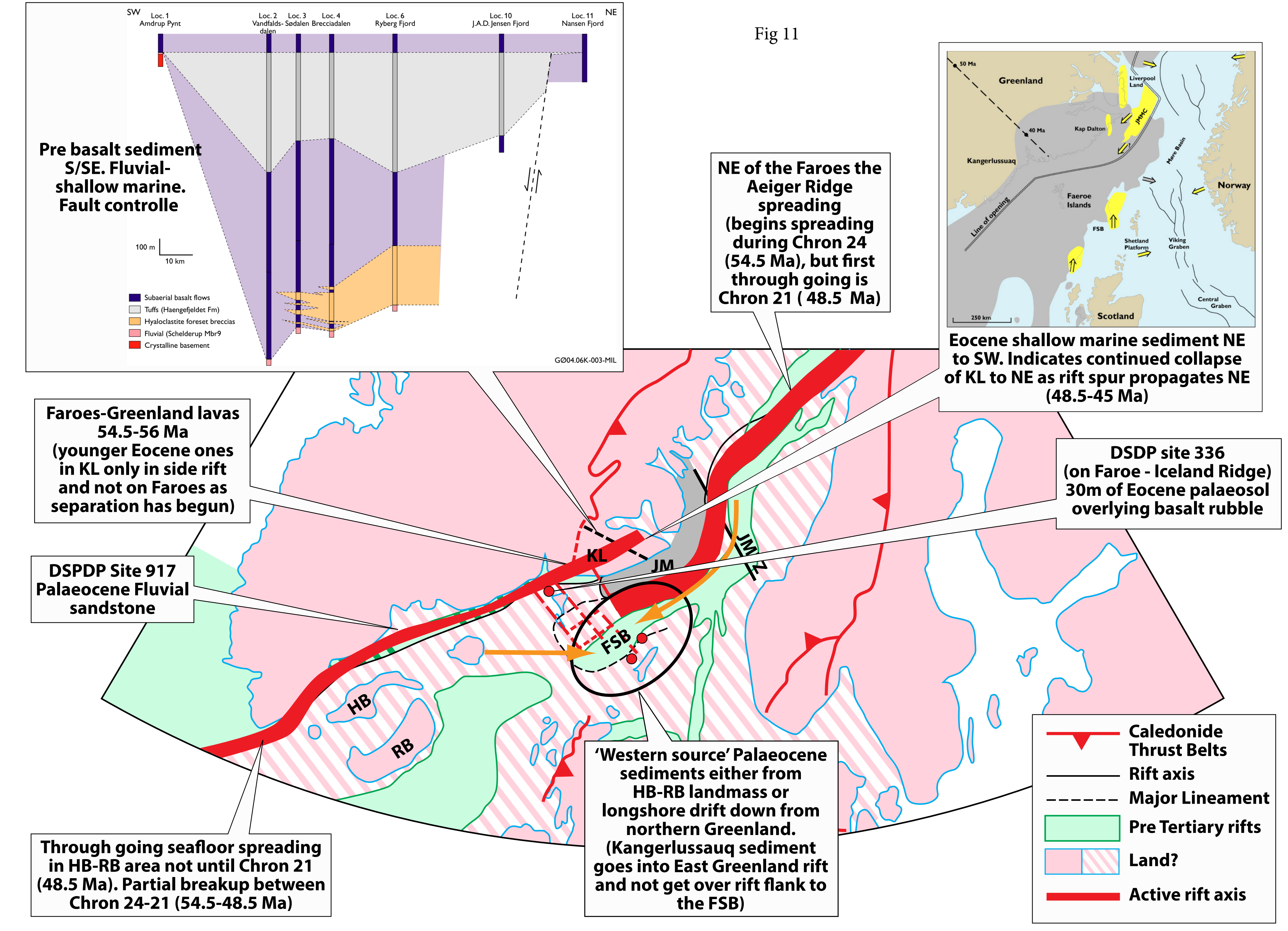
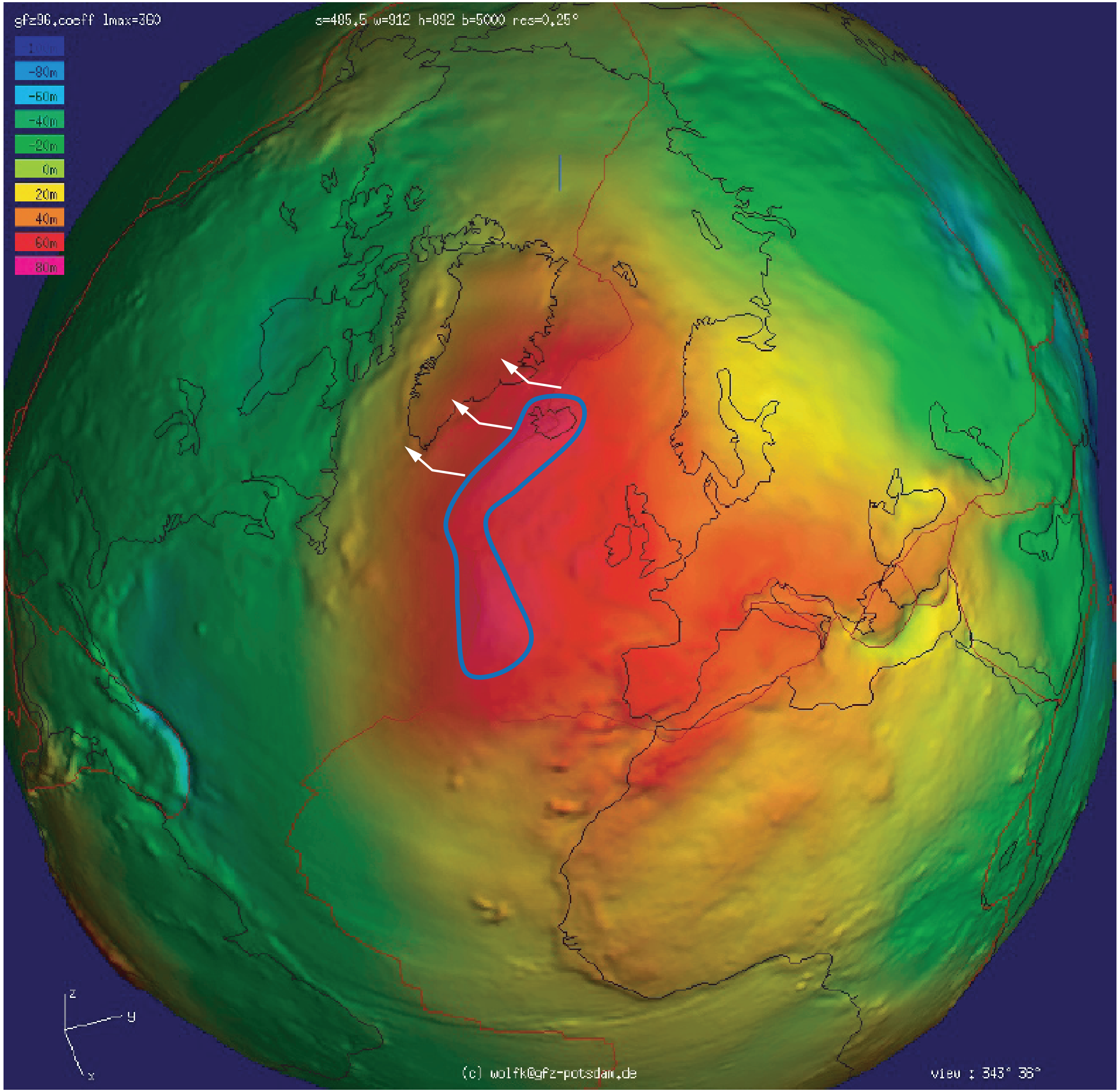

SW
NE
Loc. 1 Amdrup Pynt
Loc. 2 Vandfalds-dalen
Loc. 3 Sødalen
Loc. 4 Brecciadalen
Loc. 6 Ryberg Fjord
Loc. 10 J.A.D. Jensen Fjord
Loc. 11 Nansen Fjord
Pre basalt sediment S/SE. Fluvial-shallow marine. Fault controlle
100 m
10 km
Subaerial basalt flows
Tuffs (Haengefjeldet Fm)
Hyaloclastite foreset breccias
Fluvial (Schelderup Mbr9
Crystalline basement
GØ04.06K-003-MIL
50 Ma
40 Ma
Greenland
Liverpool Land
JMMC
Kap Dalton
Kangerlussuaq
Møre Basin
Norway
Faeroe Islands
Line of opening
FSB
Shetland Platform
Viking Graben
Central Graben
250 km
Scotland
NE of the Faroes the Aeiger Ridge spreading (begins spreading during Chron 24 (54.5 Ma), but first through going is Chron 21 (48.5 Ma)
Eocene shallow marine sediment NE to SW. Indicates continued collapse of KL to NE as rift spur propagates NE (48.5-45 Ma)
Faroes-Greenland lavas 54.5-56 Ma (younger Eocene ones in KL only in side rift and not on Faroes as separation has begun)
DSDP site 336 (on Faroe - Iceland Ridge) 30m of Eocene palaeosol overlying basalt rubble
DSPDP Site 917 Palaeocene Fluvial sandstone
KL
JM
JMFZ
FSB
HB
RB
Through going seafloor spreading in HB-RB area not until Chron 21 (48.5 Ma). Partial breakup between Chron 24-21 (54.5-48.5 Ma)
'Western source' Palaeocene sediments either from HB-RB landmass or longshore drift down from northern Greenland. (Kangerlussauq sediment goes into East Greenland rift and not get over rift flank to the FSB)
Caledonide Thrust Belts
Rift axis
Major Lineament
Pre Tertiary rifts
Land?
Active rift axis

Fig 11



Key

Relative motion of Greenland to mid-Atlantic Ridge

Area of maximum Geoid uplift

Fig 12