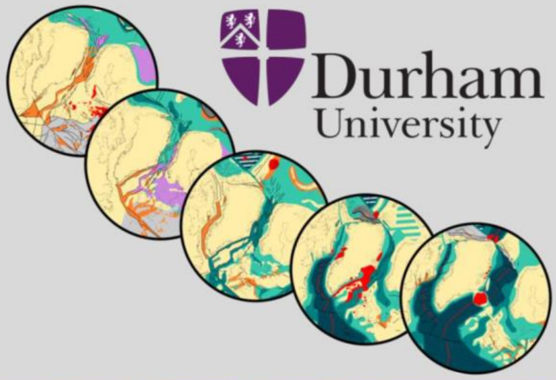
NORTH ATLANTIC 2016
DURHAM UNIVERSITY

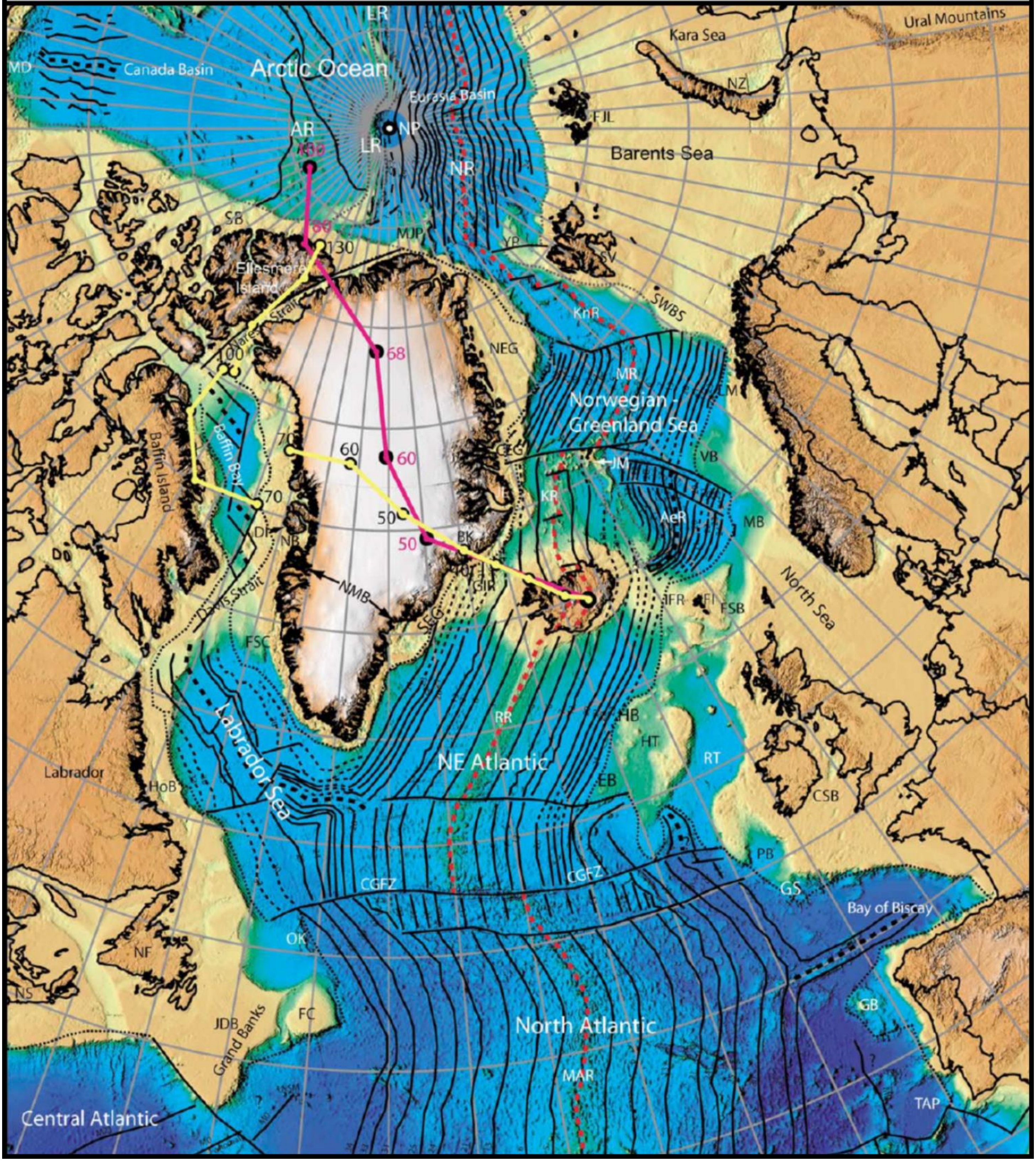
Seeking a new paradigm for the North Atlantic



Durham University

St. Chad's College, Durham University, U.K.

19th - 21st September, 2016



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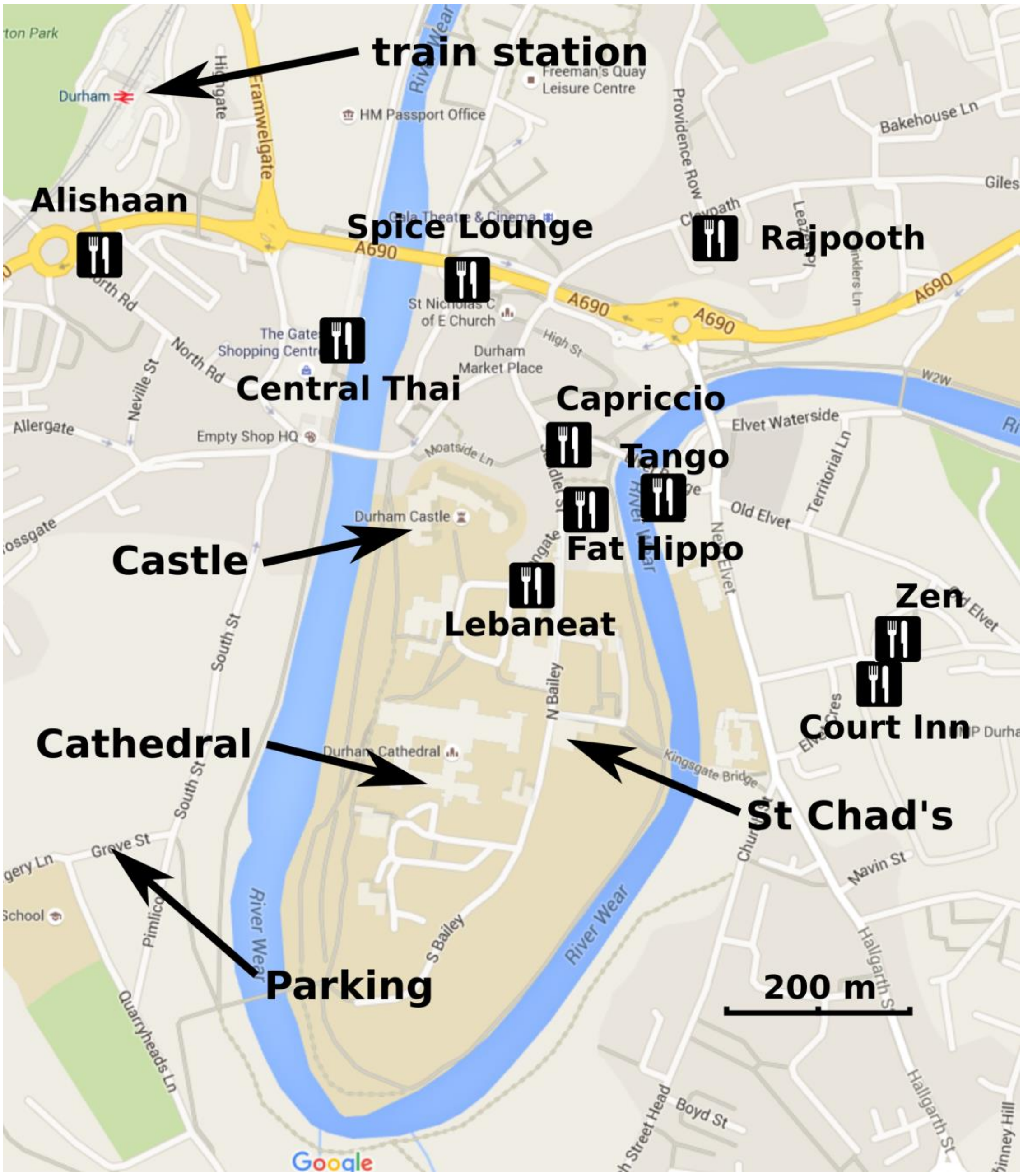

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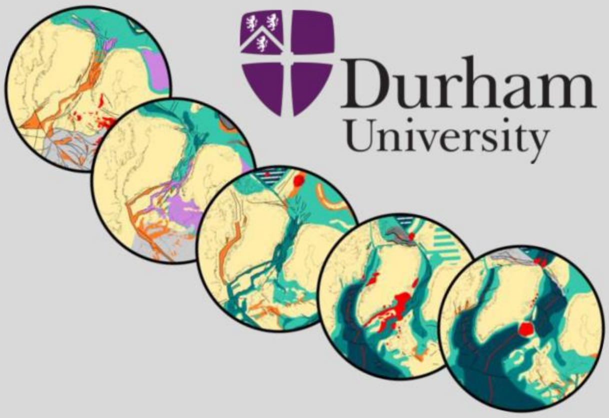

Durham University

AGENDA

Monday 19th September, 2016

8:45-9:00 Welcome, introduction

SESSION 1: RIFT PROPAGATION

9:00-9:30	Gillian Foulger	An internally consistent, holistic model for past and present tectonics and volcanism in the North Atlantic is Needed	p. 8-18
9:30-10:00	Ármann Höskuldsson	From Iceland to the Bight Transform Fault: a tale of tectonics and plate boundary formation	p. 43-44
10:00-10:30	Dieter Franke	Evidence for rift propagation towards hot-spots, including Tristan and Iceland	p. 22-24
10:30-11:00	Fernando Martinez	Reykjanes Ridge Evolution is better explained by Plate Boundary Processes than Mantle Plume flow	p. 51-54

Tea & Coffee

SESSION 2: RIFTING I

11:30-12:00	Tony Doré et al.	North Atlantic break-up: extension rate, magmatism and the driving force	p. 6-8
12:00-12:30	Laurent Gernigon	Rifting, continental breakup and post-breakup evolution of the mid-Norwegian margin	p. 26-40
12:30-13:00	Alex Peace	An evaluation of Mesozoic rift-related magmatism on the margins of the Labrador Sea: implications for rifting and passive margin asymmetry	p. 57-59

Lunch

SESSION 3: RIFTING II

14:00-14:30	David Macdonald (R. Stephenson)	An unusual rift on the Atlantic margin – the Stornoway basin	
14:30-15:00	Laurent Geoffroy	Is Iceland a volcanic passive margin C-Block?	p. 24-26
15:00-15:30	Christian Schiffer	Passive margin formation and orogenic inheritance	p. 63-66

Tea & Coffee

16:00-17:00	Discussion

19:00-	Meeting Dinner at Chad's College

TUESDAY 20TH SEPTEMBER, 2016

SESSION 4: DEEP STRUCTURE AND MAGMATISM

9:00-9:30	**MALCOLM HOLE**	**CONTROLS OF MANTLE POTENTIAL TEMPERATURE AND LITHOSPHERIC THICKNESS ON MAGMATISM IN THE NORTH ATLANTIC MAGMATIC PROVINCE**	P. 40-42
9:30-10:00	**JAMES NATLAND**	**THE PLACE OF BASALTS FROM ICELAND WITHIN THE NORTH ATLANTIC**	P. 54
10:00-10:30	**BRUCE JULIAN**	**SEISMIC TOMOGRAPHY IMAGES OF THE NORTH ATLANTIC**	P. 44-49
10:30-11:00	**KENNI PETERSEN**	**MANTLE PLUMES DRIVEN BY PLATE TECTONIC PROCESSES**	P. 63

TEA & COFFEE

SESSION 5: VERTICAL MOTIONS

11:30-12:00	**RANDELL STEPHENSON**	**A SUB-CRUSTAL PIERCING POINT FOR NORTH ATLANTIC RECONSTRUCTIONS AND TECTONIC IMPLICATIONS**	P. 66-68
12:00-12:30	**VIVI PEDERSEN**	**ISOSTATIC AND DYNAMIC SUPPORT OF HIGH PASSIVE MARGIN TOPOGRAPHY IN SOUTHERN SCANDINAVIA**	P. 59-62
12:30-13:00	**MARTYN STOKER**	**NE ATLANTIC BREAKUP: STRATIGRAPHIC CONSTRAINTS ON TECTONICS, VOLCANISM AND PALAEOGEOGRAPHIC EVOLUTION**	P. 69-73

LUNCH

SESSION6: CRUSTAL AND LITHOSPHERIC STRUCTURE

14:00-14:30	**NICK KUSZNIR**	**MAPPING CRUSTAL THICKNESS, OCT STRUCTURE AND CRUSTAL TYPE USING SATELLITE GRAVITY INVERSION: SOME ANSWERS BUT MORE QUESTIONS**	P. 49-51
14:30-15:00	**GILLIAN FOULGER & OLIVER SANFORD**	**CRUSTAL STRUCTURE BENEATH THE GREENLAND-ICELAND-FAROE RIDGE**	P. 19-21
15:00-15:30	**DAVID CORNWELL**	**LITHOSPHERIC STRUCTURE BENEATH THE FAROE ISLANDS**	P. 5-6

TEA & COFFEE

16:00-17:00	**DISCUSSION**

WEDNESDAY 21ST SEPTEMBER, 2016

SESSION7: DISCUSSION & PUBLICATION PLANNING

9:00-11:00	**DISCUSSION & PLANNING**

TEA & COFFEE

11:30-12:00	**DISCUSSION & PLANNING**
12:00-12:30	**BRIEF WORDS OF THANKS AND END OF MEETING (GILLIAN FOULGER)**

ABSTRACTS

(authors by alphabetical order)

LITHOSPHERIC STRUCTURE BENEATH THE FAROE ISLANDS

David G. Cornwell[1,*], Carmelo Sammarco[1], Nicholas Rawlinson[1], David A. Thompson[1], Richard W. England[2], Graham W. Stuart[3]

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The landmass of the Faroe Islands provides an excellent opportunity to constrain the structure of the lithosphere near the presumed ocean-continent transition (Funck et al., 2014) along the eastern North Atlantic margin. We use ambient noise and teleseismic earthquake data collected by the Faroe Islands Passive Seismic Experiment (FIPSE), a temporary 12-station broadband seismometer network, to map the major subsurface discontinuities within the crust and upper mantle beneath the Faroe Islands. Firstly, we construct a detailed structural model of the uppermost 15-20 km from a 3D shear wave velocity model derived from ambient noise and receiver function inversion to constrain lateral variations within the Faroe Islands Basalt Group (FIBG, Passey & Jolley, 2009). We correlate these layer boundaries with onshore and offshore well- and seismic-derived horizons that extend into the Faroe-Shetland Basin (e.g. Olavsdottir et al., 2016). The depth to crystalline basement beneath the FIBG is mapped at 5-10 km, with relatively low *S*-wave velocities providing evidence for sedimentary and/or hyaloclastite rocks between basalt and basement. Crustal thickness estimates of 26-33 km from *P*-wave receiver function H-κ stacking analysis are similar to Moho depth estimates from previous offshore seismic refraction/wide angle reflection experiments adjacent to the Faroe Islands (e.g. Raum, et al., 2005). Bulk crustal V_P/V_S measurements of 1.69-1.80 are consistent with a Lewisian crystalline basement modified by basaltic magmatism. A prominent positive acoustic impedance discontinuity at 18-20 km depth may indicate the uppermost extent of a high-velocity magmatically intruded lower crust, although the lack of a sharp Moho *P*-*S* conversion under the majority of the Faroe Islands hints at a layered or complex crust-mantle boundary. The lithosphere-asthenosphere boundary (LAB) beneath the Faroe Islands is constrained for the first time from analyses of *S*- wave receiver functions. The LAB occurs at 80±10 km depth (cf. Iceland and Jan Mayen LAB depths of 70-90 and 40-60 km, respectively, Kumar et al., 2005), is consistent with an ~8 % velocity reduction but is more likely to represent a gradual rather than a sharp discontinuity. A seismologically-derived LAB at 80±10 km is slightly deeper than the depth of melting of ~65 km estimated from geochemical analysis of FIBG mantle xenoliths (Hole & Millett, 2016) and the discrepancy could be explained by cooling effects and LAB deepening through vertical accretion. We further investigate whether a 'normal' mantle transition zone (MTZ) and

significant upper mantle anisotropy exist beneath the Faroe Islands using *P*-wave receiver functions and shear wave splitting analysis using teleseismic *SKS* phases, respectively, and link these to the thermal properties of the mantle. These findings, together with previous geophysical studies in the region, provide key information about the structure of the crust and upper mantle beneath the Faroe Islands, which highlight the magmatic and tectonic processes that contributed to pre-, syn- and post-continental break-up in this part of the North Atlantic Igneous Province.

References

Funck, T., Hopper, J.R. et al. 2014. Crustal structure. In: Hopper, J.R., Funck, T., Stoker, M.S., Arting, U.,Peron-Pinvidic, G., Doornenbal, H. & Gaina, C. (eds) Tectonostratigraphic Atlas of the North-East Atlantic Region. Geological Survey of Denmark and Greenland (GEUS), Copenhagen, 69–126.

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NORTH ATLANTIC BREAK-UP: EXTENSION RATE, MAGMATISM AND THE DRIVING FORCE

Tony Doré[1] and Erik Lundin[2]

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It is widely assumed in passive margin research that magma-poor margins represent slow extension rates at time of break-up, and conversely that magma-rich margins equate to fast extension. The terms are used synonymously such that, for example, "slow spreading margin" is automatically taken to mean that it will be magma-poor and include all of the phenomena associated with such margins, such as hyperextension and mantle exhumation.

However, this assumed equivalence is largely anecdotal, and based on just a few passive margins. A gradation is likely to exist between the two end-members, so confirmation of the idea could come from a systematic examination of extension rates at time of break-up, to determine whether there is a direct relationship with magmatic quantity. Such studies have not to our knowledge been done.

This situation is probably because extension rates immediately prior to break-up, and magmatic volumes, are both notoriously difficult to quantify. Extension rates at break-up are generally assumed to reflect the initial ocean floor spreading half-rates, which are in the order of 10 mm/year for a slow-spreading margin and 25 mm/year and above for a fast-spreading margin. Extension rates prior to break-up can theoretically be constrained by fault analysis and by dating of sedimentary successions or contemporaneous volcanics, but this process is imprecise. It is particularly difficult in hyperextended margins where syn-rift wedges are seldom sampled, and exhumed mantle cannot normally be dated isotopically. Linear magnetic anomalies in exhumed mantle may have been detected by deep-towed magnetometer between Iberia but are more usually absent or currently unidentified.

We will discuss these issues with respect to North-East Atlantic break-up and the breakaway of the Jan Mayen microcontinent, and by comparison to the Central and South Atlantic.

The NE Atlantic can probably be considered the type example of a magma-rich or volcanic margin, and is probably the world's best studied representative of the type. Initial spreading rates after breakup at 53 Ma were quite fast, in the order of 25 mm/year half-rate along some portions of the spreading ridge. However, immediately after break-up spreading rates slowed, reducing to 10 mm/year or less at 40 Ma. Slow to ultra-slow spreading has persisted to present day. Thus, in this instance, there is a correspondence between fast break-up and excessive magmatism, the latter of which is usually attributed to the presence of a plume. However, the NE Atlantic contains a central segment where the Aegir and Kolbeinsey Ridges overlap. The Kolbeinsey Ridge opened in the Miocene when NE Atlantic spreading rates were quite slow, and did so without excessive magmatism. A number of workers have proposed that the Kolbeinsey Ridge formed as a consequence of the "emergence" of the Iceland plume from beneath Greenland. It is thus noteworthy that the Kolbeinsey Ridge did not open in a magma-rich fashion. This may be a key area for testing the relationship between melt generation and spreading rate and, likewise, supposed plume-induced melting.

We ask two key questions:

1) Does the positive correlation between magmatism and extension rate hold, or is it possible to have (for example) a magma-rich slow-spreading margin?
2) If the correlation holds, why is this the case?

It will be shown that the answer to the second question is far from obvious. Modelling of decompressive melting at mid-ocean ridges shows a positive correlation between spreading rate and magma production, but with a maximum melt thickness (i.e. thickness of oceanic crust) of about 7 km. Thus, in areas of over-thickened crust at magmatic margins, it is widely assumed that the magmatic addition cannot be explained by rate alone, and that excess heat must have been introduced into the system by hot, rising asthenosphere, i.e. a mantle plume. In this case, given the positive correlation described above, it must also follow that the rising plume causes the extension rate to speed up, either by increasing the rate of plate separation, or by increasing lithospheric ductility, or both. However, current plate models incorporate the NE Atlantic speed-up and slow-down at time of break-up as a wider phenomenon, extending significantly beyond any conceivable plume influence.

An alternative view is that the processes at margins during break-up are different from those at mid-ocean ridges, and that the melt volume there is primarily a function of break-up rate. This idea has been argued in terms of increasing angular separation velocity and increasing magmatism with distance from the plate tectonic pole of rotation (Euler Pole), so magma-poor margins occur closer to the pole and magma-rich margins occur in distal positions. As a further alternative, recent rheological modelling work suggests that accelerated extension rates at time of break-up may be widespread and due to nonlinear decay of rift strength, a process in which both plume impingement and distance from the rotation pole could be largely incidental.

We will show that, for the examples examined, the magmatism versus rate correlation may still be valid despite initial indicators to the contrary. We will also show that, while it is quite reasonable that input of heat from a mantle plume would cause excess magma production, there is considerable ambiguity around whether it can increase extension rate and – particularly – absolute plate separation rate on a broad scale.

AN INTERNALLY CONSISTENT, HOLISTIC MODEL FOR PAST AND PRESENT TECTONICS AND VOLCANSIM IN THE NORTH ATLANTIC IS NEEDED

Gillian R. Foulger

Dept. Earth Sciences, Durham University, Durham, U.K.

The North Atlantic ocean between the Charlie-Gibbs Fracture Zone and the Knipovich Ridge has been chronically unstable from the time of continental breakup at ~ 54 Ma up to the present day. The central part of this zone, encompassing the Greenland-Iceland-Faeroe (GIF) ridge and the Jan Mayen Fracture Zone (JMFZ), has had a particularly complex history and is responsible for decoupling spreading north of the JMFZ from that south of the GIF, *e.g.*, the response to a change in direction of spreading on extinction of spreading in the Labrador Sea.

There is no reason to suppose this chronic disequilibrium is not still ongoing today. Indeed there is evidence that it continues, *e.g.*, from the evolving configuration of the plate boundary in Iceland and on the Reykjanes ridge (pron: Rake-ya-nes, not Wreck-ya-nes), and from tectonic activity in regions away from the plate boundary in the Pliocene. Tectonic complexities include:

- Plate-boundary reorganizations and rift migrations, *e.g.*, from the Aegir ridge to the Kolbeinsey ridge;
- Continental microcontinent material of unknown extent, e.g., the Jan Mayen microcontinent, which possibly extends south beneath Iceland;
- Propagating and dying ridges, *e.g.*, on the Reykjanes ridge [e.g., Hey *et al.*, 2010], the Kolbeinsey ridge and in Iceland;
- Instability on the Reykjanes ridge in the form of evolution from boundary normal to ridge-transform staircase morphology and then to oblique spreading;
- Sagging, flexing and tilting of blocks around Britain [e.g., Stoker *et al.*, 2010];
- Stress and motion reorganizations and variable orientations over short distances, *e.g*., in Iceland;
- Tectonic reactivations (*e.g.*, of the Vøring Spur);
- Extensions (evidenced by subsidence and/or volcanism) and compressions (evidenced by uplift) at various times and places from 54 Ma to the present;
- Complex plate boundary configuration including multiple triple junctions.

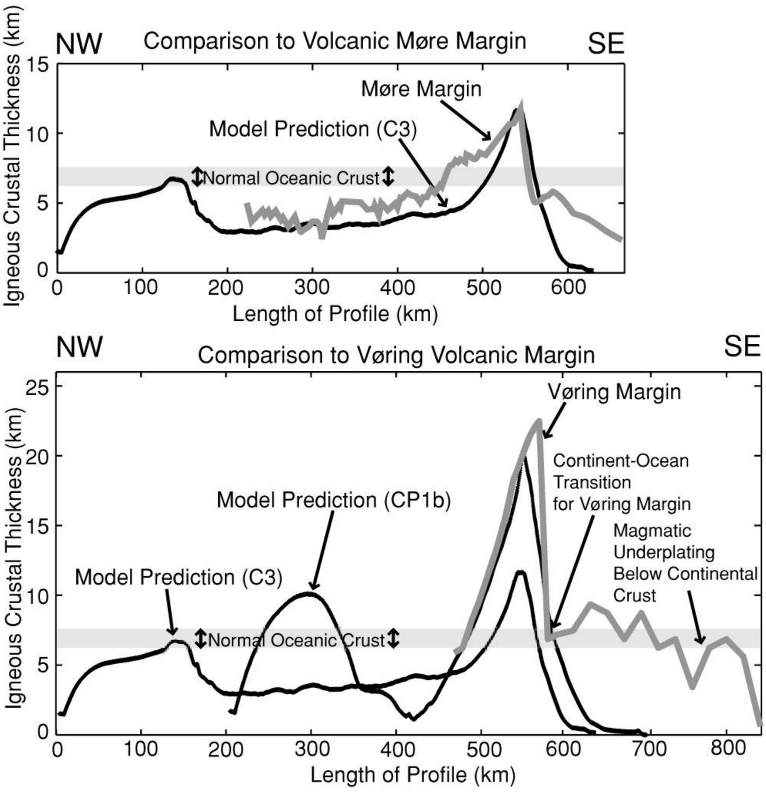

Figure 1: Comparison of numerical modeling results for melt production at the Møre and Vøring margins with observed crustal thicknesses [from Simon et al., 2009].

Continental breakup at ~ 54 Ma was accompanied by large-volume magmatism that built the volcanic margins of the North Atlantic. The volcanic rate subsequently dwindled and thinner igneous crust was created as the ocean basin formed. This basic pattern of a large-volume burst of volcanism at the onset of breakup followed by lower-volume volcanism thereafter is a natural and expected consequence of the breakup of thick lithosphere [e.g., Simon *et al.*, 2009] (*Figure 1*). An initial large burst of volcanism is likewise expected to accompany break-off of the Jan Mayen continental microplate at ~ 44 Ma. It is not necessary to appeal to additional *ad hoc*, unconnected conditions such as extraordinary compositional or temperature variations, or enhanced lateral flow of asthenosphere, to explain it.

The Jan Mayen microcontinent was rafted into the Atlantic when the Kolbeinsey ridge formed and spread. The northern part of the microcontinent is thought to be continental and its southern part likely comprises slivers of continental material interspersed by oceanic material. Its southern boundary is not well known (*Figure 2*) and continental material may continue beneath Iceland. Palinspastic reconstructions *require* a captured microplate up to ~ 210 km wide to underlie Iceland (*Figure 3*d)[Foulger, 2006]. This probably comprises both continental and oceanic crust that must be older than the oldest rocks exposed in Iceland (~ 16 Ma; *Figure 3*) [Foulger, 2006]. Geochemical data from some Icelandic basalts and (currently unconfirmed) reports of Paleozoic and Mesozoic zircons [Paquette *et al.*, 2006; Sigmarsson *et al.*, 2007] require a component of continental crust.

Such a microplate will contribute to the thickness of the Icelandic crust. Thus a map of the depth to the base of the V_S = 3.7 km/s horizon (sometimes considered to equate to the base of the crust) may give a clue to the location of this microplate (*Figure 4*). It would be interesting to know the maximum possible width of continental crust that may underlie Iceland, from palinspastic matching of Eurasia and Greenland. A related issue is that the composition of the layer beneath Iceland popularly equated to the "lower crust" ($V_P \leq 7.3$ km/s, density = ~ 3,200 kg/m^3) is not known and thus the true melt production rate is also unknown.

The formation and ascendance of the Kolbeinsey ridge and the extinction of the Aegir ridge that rafted the Jan Mayen microcontinent into the middle of the ocean was a first-order complexity of the growing basin. The conjugate, fan-shaped spreading of these two overlapping ridges resembles an overlapping ridge pair but on the scale of hundreds of kilometres rather than just a few as is the case on the Reykjanes ridge. It also resembles the current rift arrangement in south Iceland, where the Western Volcanic Zone and Eastern Volcanic Zone have a similar geometry (*Figure 3*). Overlapping ridges including both propagators and dying segments thus exist throughout the North Atlantic and on a variety of scales.

The chronically unstable configuration of the Reykjanes ridge involved initial boundary-normal spreading. A change in the direction of plate motion to boundary oblique at ~ 37 Ma was accompanied by a change to a ridge/transform staircase geometry. This immediately began to be "ironed out" by small-scale, southward growing, ridge propagators which are proposed to account for the chevron ridges of thickened crust that flank the Reykjanes ridge [e.g., Hey *et al.*, 2010]. Similar chevron ridges are observed flanking the Kolbeinsey ridge. Why a propagating ridge tip should leave in its wake thickened (or thinned?) crust is still under debate. Dynamic modeling of propagating rifts and transforms and comparisons with propagating/dying pairs on the Pacific spreading plate boundaries may contribute to this debate [e.g., Koopmann *et al.*, 2014].

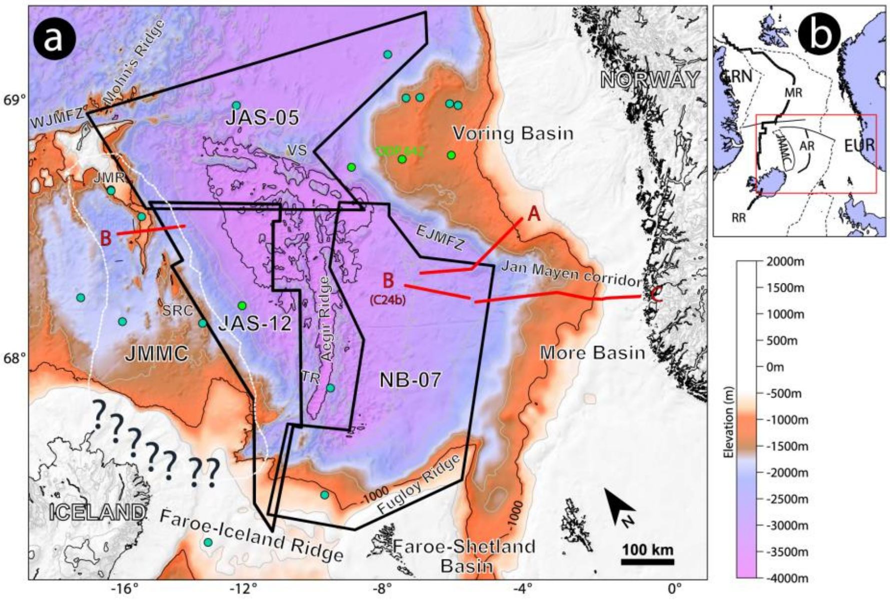

Figure 2: Main physiographic features of the Norway Basin and surroundings. EJMFZ: East Jan Mayen Fracture Zone; JMR: Jan Mayen Ridge; SRC: (Jan Mayen) Southern Ridge Complex; TR: Treitel Ridge; VS: Vøring Spur; WJMFZ: West Jan Mayen Fracture Zone. Circles indicate drillsites of Ocean Drilling Project (ODP, light green) and Deep Sea Drilling project (DSDP, sea green). White line outlines the Jan Mayen microcontinent and ??? indicates that the location of its southern boundary is unknown [modified from Gernigon et al., 2015].

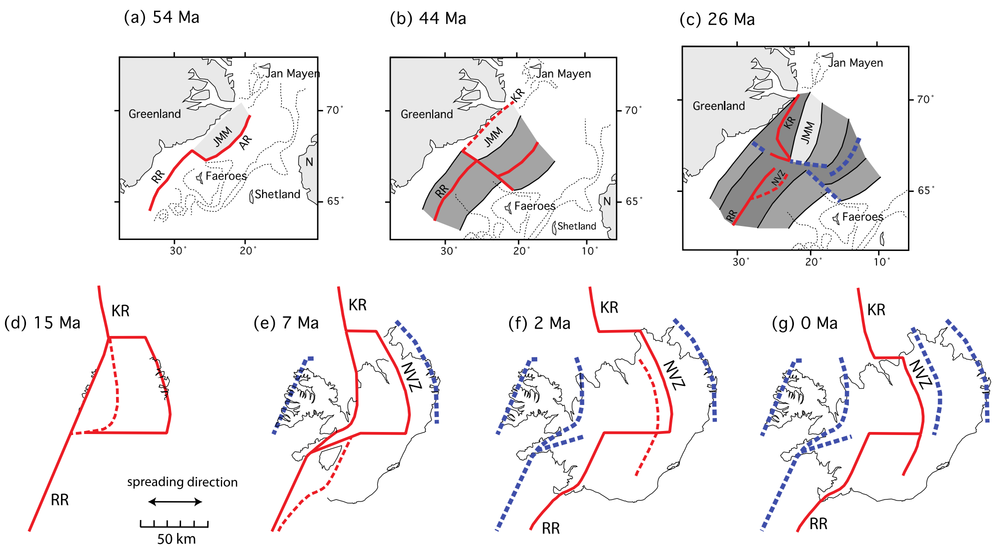

Figure 3: Tectonic evolution of the Iceland region during the past 54 Ma. Red lines–currently active plate boundaries; dashed red lines–imminent plate boundaries; dashed blue lines–extinct plate boundaries; thin lines–bathymetric contours; JMM–Jan Mayen microcontinent; KR–Kolbeinsey Ridge; N–Norway; NVZ–Northern Volcanic Zone; RR–Reykjanes Ridge, AR–Aegir Ridge [from Foulger, 2010].

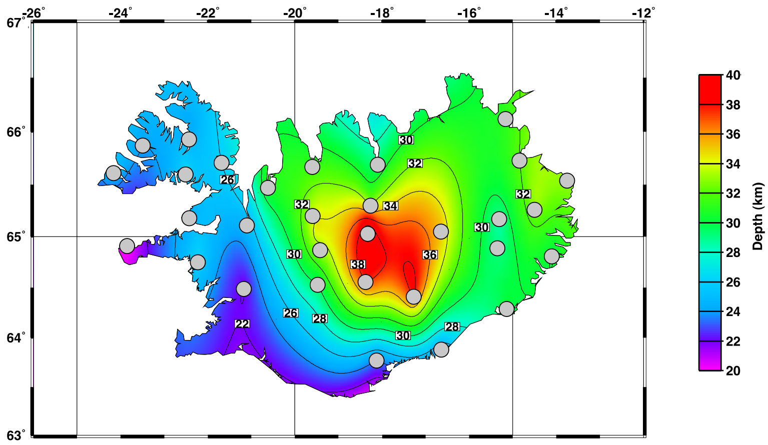

Figure 4: Contour map showing the depth to the base of the V_S = 3.7 km/s horizon, approximately equivalent to the top of Layer 4, from seismic receiver functions. Gray circles indicate seismic stations [from Foulger & Anderson, 2005; Foulger et al., 2003].

Regions close to the continent-ocean boundaries offshore Greenland, Norway and Britain exhibited complex tectonic activity throughout the lifetime of the ocean and up to the present day. These "passive" margins are, in fact, dynamic and their "amagmatic" parts are actually magmatic [Peace *et al.*, 2015]. Complexities include subsidence and uplift of blocks on a relatively small scale [Stoker, 2016; Stoker *et al.*, 2013; Stoker *et al.*, 2012; Stoker *et al.*, 2010] and volcanism away from the spreading ridge. Frequent tectonic activity is also reported reported to be widespread along the highly complex JMFZ [Gernigon *et al.*, 2012; Gernigon *et al.*, 2009b] (*Figure 5*). The complexity, extent, multiple fault branches, transtensional and transpressive features of this Fracture Zone suggest that it is a shear counterpart to the complex, diffuse array of rifts and faults that form the Icelandic extensional region.

In Iceland, subaerial observations may be made of the volcanic and tectonic activity there during the last ~ 16 Ma [e.g., Foulger & Anderson, 2005]. Contrary to oft-repeated claims, the spreading plate boundary has not migrated monotonically east but both westerly and easterly migrations have occurred. Ironically, eastward rift migration is commonly claimed to indicate a migrating plume beneath Iceland whereas westward rift migration is invoked as evidence for a plume north of it (the Aegir/Kolbeinsey rift migration). Iceland has been in a perpetual state of plate-boundary evolution/instability for its entire history and is still so at the present time. The latest major rift migration occurred and at ~ 2 Ma when the Eastern Volcanic Zone formed in south Iceland (*Figure 3*f,g).

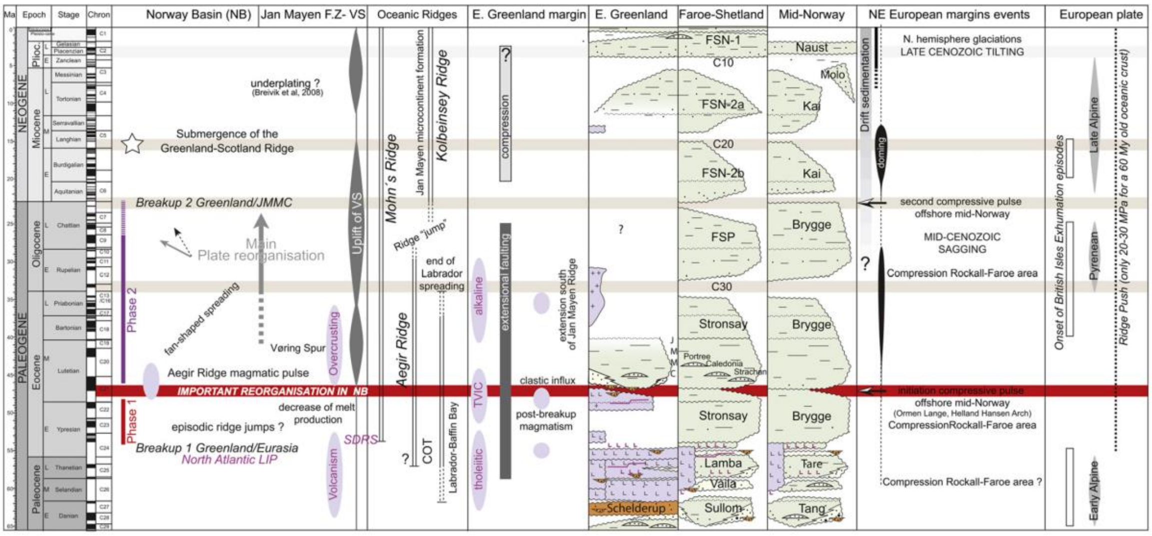

Figure 5: Chronology of tectonic events in various parts of the North Atlantic [from Gernigon et al., 2012].

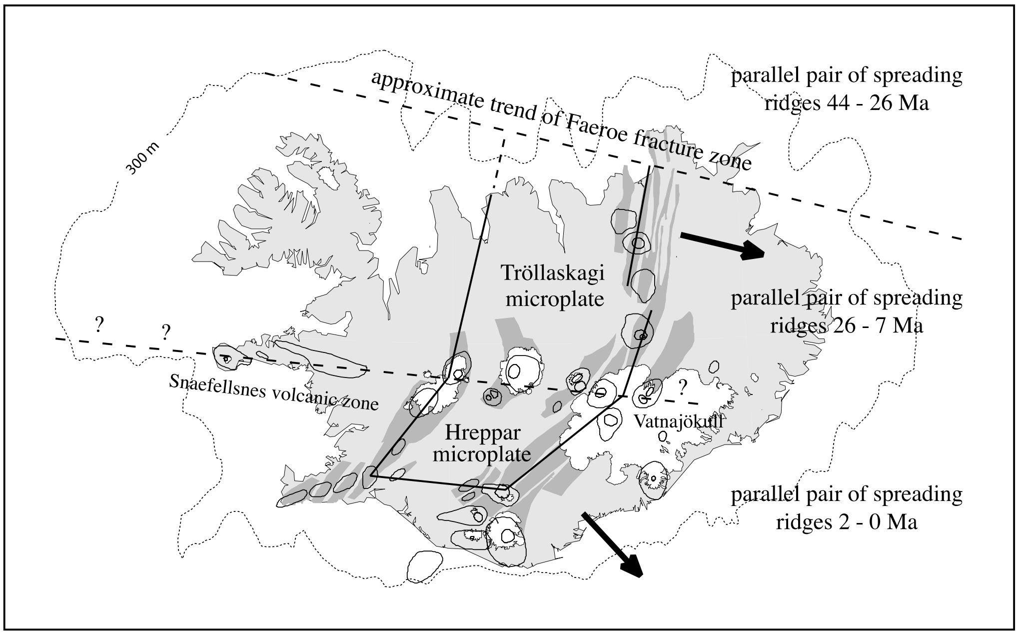

Figure 6: Simplified tectonic map of Iceland. Parallel pair spreading has migrated south from 44 Ma to present. An EW zone that may represent a long-lived composite zone of various plate-boundary elements extends from the Snaefellsnes volcanic zone across central Iceland and into Vatnajökull (lower dashed line). Solid black lines suggest locations of other boundaries of the Tröllaskagi and Hreppar microplates. Arrows show local directions of motion deduced from GPS surveying (see also Figure 7), earthquake focal mechanisms, the trends of presently active volcanic zones, and the orientations of Tertiary dikes [from Foulger et al., 2005].

An EW-trending composite volcanic zone may be discerned that extends from Snaefellsjökull in the far west of Iceland, across the centre of Iceland and beneath the icecap Vatnajökull which overlies a cluster of large volcanoes (*Figure 6*). Southeast of this, plate motion in Iceland appears to be more southerly than it is to the north as may be deduced from the orientation of the rift zones and GPS measurements (*Figure 6* and *Figure 7*). If this is so, it suggests an explanation for the major volcanism in central Iceland and beneath Vatnajökull, i.e. it indicates a component of transtension across the EW zone. This equates to a NS component of opening and would be required by rotating microplates.

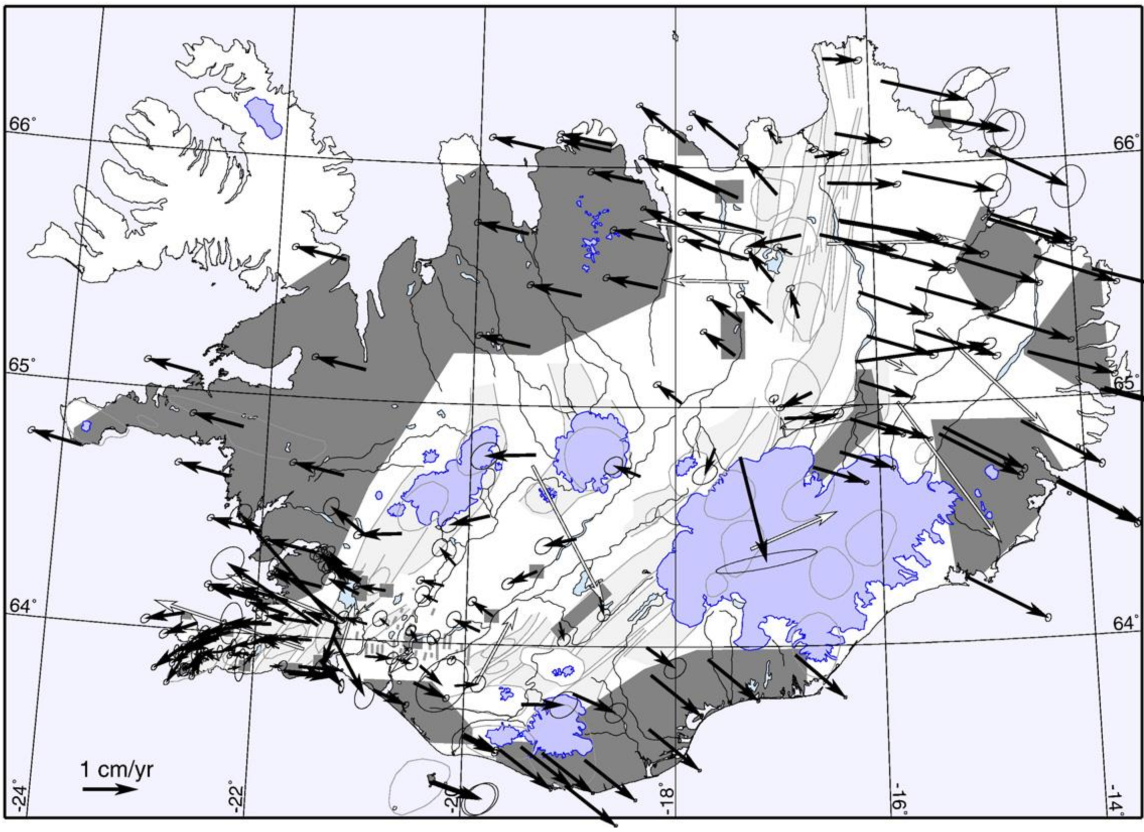

Figure 7: Vectors of horizontal velocity from GPS data between 1986 and 2002 with 1σ standard deviations. Blue: icecaps, dark grey: blocks considered by Perlt et al. [2008] to be tectonically inactive. However, it is likely that no part of Iceland is truly inactive. Insignificant velocity vectors are not filled [reproduced from Perlt et al., 2008].

If such a state existed at this latitude for most of the opening of the North Atlantic, it can explain the formation of the ridge of thickened crust that makes up the GIF ridge. The schematic map shown in *Figure 6* suggests the presence of at least four plate boundary junctions that comprise unstable triple junctions of zones of deformation. The structure blanketed by the excess volcanism of the GIF ridge is possibly the most fundamentally important element that has influenced the opening and subsequent evolution of the North Atlantic, and its core may be represented by the EW Snaefellsjökull-Vatnajökull transverse zone.

The JMFZ/Vøring Spur has functioned as a leaky transform or oblique crack during some periods. This can account for the Traill Ø volcanics, which erupted when that part of the transform was in extension. Gernigon et al. [2009a] compared it with the Azores triple junction, a complex part of the mid-Atlantic ridge where excessive volcanism occurs, building sub-areal edifices. It is possible that the GIF ridge is a similar but more profound phenomenon [Gernigon *et al.*, 2009a]. A structure that comprises a hindrance to rift-tip propagation beneath the GIF ridge would accord with the observation that, throughout the Atlantic ocean, "hot spots" function to impede and delay continental breakup and may comprise the last section to separate [Koopmann *et al.*, 2014]. Both the JMFZ and Iceland may be rejuvenated older inherited structures.

Variable directions of plate motion not only result in extensional, volcanically productive zones but must also be balanced by compressions elsewhere. If the distant plate directions for the regions north and south of Iceland are the same, these compressions must be taken up locally. Compressional features are reported from many areas along the west and east Atlantic margins and intraplate deformation may also occur *e.g.*, in Britain, the North Sea, and Europe. Magmatism tracks extension so volcanism and magmatism may be used to map the distribution of extension at various times, balanced by compression/uplift. Some reports of coupled plate boundary reorganisation events and tectonism elsewhere have already been reported, *e.g.*, vertical motions on the Lofoten-Porcupine margin correlate with extinction of spreading in the Labrador Sea.

A target of our workshop is to develop a coherent, region-wide model for the history of the North Atlantic that explains horizontal motions, vertical motions, and magmatism in an internally consistent way without appealing to *ad hoc*, unrelated events such as the onset of local convection cells that be added or subtracted with no testable consequences for the model as a whole. Plate motions are accompanied by the flow of material in the asthenosphere beneath, but it is not helpful to invoke independent mantle convection, *e.g.*, deep mantle plumes, EDGE convection or small-scale sub-lithospheric convection, to explain volcanism. Volcanism occurs where the crust is in extension and so such a model should extend our understanding of the volcanism of the JMFZ, the GIF ridge and the Reykjanes ridge and link it to events elsewhere in the North Atlantic. Extension/volcanism in one place is expected to be balanced by compression elsewhere. Compressions and extensions revealed by uplift and subsidence in the wider region, the geological history of rifting and rift migrations on land in Iceland, in a way that is consistent with geochemistry and other data. Many of the components of such a model are already in place. We hold a box of jigsaw pieces that are likely now sufficiently complete to reveal the overall picture. The time is thus ripe to assemble these pieces to make testable predictions for what remains to be observed.

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CRUSTAL STRUCTURE BENEATH THE GREENLAND-ICELAND-FAROE-SHETLAND RIDGE

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Numerous seismic studies using both passive (receiver functions) and active seismology have probed the structure and thickness of the crust beneath the Greenland-Iceland-Faroe-Shetland ridge. Receiver function and wide-angle reflection work yield results that agree to a first order. The seismological crust is ~ 30 km thick beneath the Greenland-Iceland and Iceland-Faroe ridges, and eastern Iceland, ~ 20 km beneath western Iceland, and ~ 40 km beneath central Iceland.

Beneath Iceland, the upper crust is typically 7 ± 1 km thick, heterogeneous, and with high velocity gradients. The lower crust is typically 15 - 30 ± 5 km thick and has been defined to begin where the upper-crustal velocity gradient decreases radically. This generally occurs at the V_p ~ 6.5 km/s level.

At the second order level, there are discrepancies between interpretations of the results from receiver functions and wide-angle reflection. Receiver functions suggest that the crust-mantle boundary is a transition zone ~ 5 ± 3 km thick throughout which V_p increases progressively from ~ 7.2 km/s to ~ 8.0 km/s. It may be gradational or a zone of alternating high- and low-velocity layers. There is no evidence from receiver functions for melt or exceptionally high temperatures in or near this zone. Interpretations of wide-angle reflection results as regards the depth to the base of the crust are constrained by the assumption that there is a sharp Moho beneath Iceland. Mapping of such a horizon is mostly dependent on reflections because refracted head waves are extremely rare. However, this approach suffers from the problem that there is very weak constraint on the velocity of the material beneath the reflective horizon which might thus comprise only a thin lens.

The layer with seismic velocities $V_p \leq$ ~ 7.3 km/s is 30-40 km thick beneath a region roughly in the centre of Iceland, thinning radially away from this. The location of this thick zone is slighly different from receiver-function and wide-angle reflection results (Figure 1). Receiver functions reveal a low-velocity zone ~ 10,000 km^2 in area and up to ~ 15 km thick in the lower crust beneath central Iceland. It may represent a submerged, trapped oceanic microplate. Interpretations of wide-angle reflection results place this “thick spot” beneath the Grimsvötn volcano in east-central Iceland, near the currently most active region of volcanism in the island.

A single quasi-circular region of thick crust is not expected for a long-lived site of high melt production beneath a spreading ridge, which would be expected to have built a continuous EW ridge of thickened crust. Explanations for this involving viscous flow of a hot lower crust have been suggested. However, these are at odds with interpretations of the thick crust beneath the Greenland-Iceland and Iceland-Faroe ridges as also resulting from excess magmatism.

Isostasy indicates that the density contrast between what is interpreted as the lower crust and the mantle is only ~ 90 kg/m^3 compared with ~ 300 kg/m^3 for normal oceanic crust, indicating compositional anomalies that are as yet not understood. Low attenuation and normal V_p /V_s ratios in the lower crust beneath central and southwestern Iceland, and normal uppermost mantle velocities in general, suggest that the crust and uppermost mantle are subsolidus and cooler than at equivalent depths beneath the East Pacific Rise.

Work is currently underway at Durham to study crustal structure in the region between the Faroe and Shetland islands. Reliable imaging of extrusive basalt flows using conventional seismic techniques is an outstanding problem because of high levels of scattering and attenuation of the seismic wavefield. There is progressive degradation of the seismic signal throughout layered basalt sequences, leading to ambiguous interpretations of the base of the basalt sequence and the sub-basalt geology. Velocity models developed from wide-angle refraction seismic data based on ray tracing tomography may also be unreliable owing to the fundamental assumptions of the ray-theory approximation. The current work focuses on the use of high resolution full-wavefield seismic modelling techniques that can potentially replicate the scattering and attenuation seen in acquired seismic data and accurately model the seismic wave propagation. A further issue is imaging igneous intrusions, with many intrusions within the Faroe-Shetland sill complex being below the level of seismic resolution. That may lead to inaccurate interpretations of seismic data.

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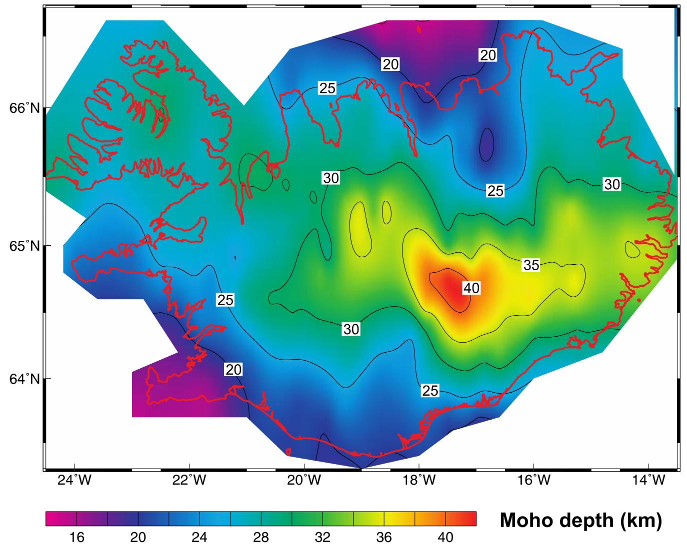

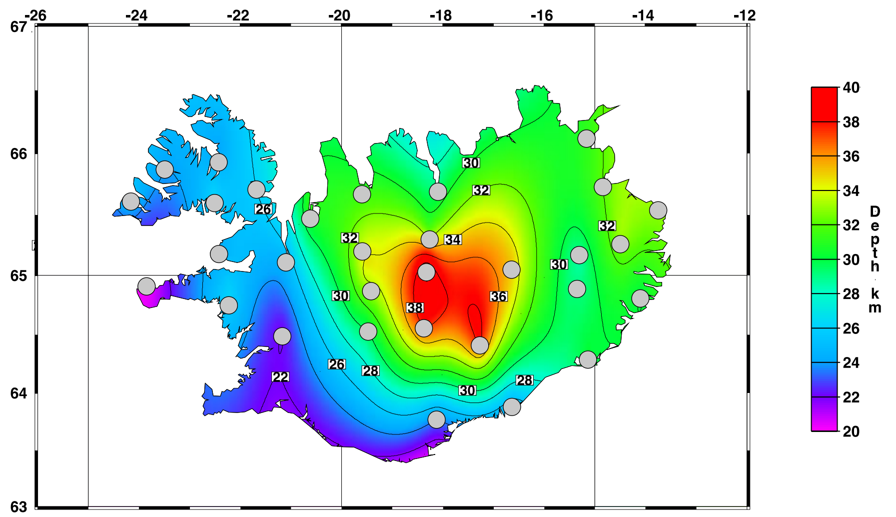

Figure 1: Top: Contour map of total crustal thickness determined using a combination of seismic profiles, receiver functions and gravity profiles [from Darbyshire et al., 1998]. Bottom: the lower crust defined as the depth to the Vs = 4.1 km s−1 horizon [from Foulger et al., 2003].

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EVIDENCE FOR RIFT PROPAGATION TOWARDS HOT-SPOTS, INCLUDING TRISTAN AND ICELAND

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Break-up of both, the North Atlantic and the South Atlantic took place in an astonishingly similar pattern. Neither ocean did initiate opening at the proposed hot-spot locations. Rather, break-up occurred in the South and North Atlantic distal from the hot-spots, with a rift propagating towards the hot-spot positions. The hot-spots at their break-up positions coincide spatially with major, likely lithosphere-scale shear zones and aborted rifts, indicating complex rifting. There is apparently a consistent delay in break-up at the proposed locations of hot-spots.

The **South Atlantic** started opening in the Early Cretaceous, at about 140 Ma (Franke, 2013; Koopmann et al., 2014), subsequent to a long phase (or phases) of continental extension starting likely during the Carboniferous–Permian (e.g. Macdonald et al., 2003; Stollhofen et al., 2000). The emplacement of the Paraná-Etendeka LIP in Brazil and Namibia peaked in late Hauterivian–early Barremian (134-129 Ma; (Peate, 1997)). Opening of the southern segment of the South Atlantic occurred from South to North and is interpreted as a successive unzipping of rift zones (e.g. Jackson et al., 2000; Nürnberg and Müller, 1991). Accordingly, SDRs were emplaced from south to north: magnetic anomaly M9 is the oldest anomaly seaward of the SDRs off Cape Town while 2000 km northwards the oldest magnetic anomaly is M0 (Koopmann et al., 2014): a difference of approximately 10 Myrs. Seafloor spreading north of the Florianópolis Fracture Zone (FZ) initiated at about 112 Ma (Heine et al., 2013; Moulin et al., 2009; Torsvik et al., 2009): an interval of more than 20 My between break-up ages of southern (~140 Ma) respectively northern (~112 Ma) segments (Franke, 2013). The Paraná-Etendeka LIP is spatially linked to this FZ.

Continental break-up in the **North Atlantic** ended a ~350-My-long predominately extensional period (Ziegler, 1988). During break-up two sublinear rift segments in the south and north of the opening oceanic basin were connected by a sinuous segment in the middle (Larsen et al., 1994). The North Atlantic LIP in East Greenland erupted at 56–55 Ma (Eldholm et al., 1989; Storey et al., 2007). The third of three episodes emplacing plateau basalts in the center of the North Atlantic LIP in East Greenland (Larsen and Watt, 1985) at around 55 Ma correlates with oceanic crust formation. This indicates episodic rifting during the two previous episodes. While it is undisputed that the southern portion of the North Atlantic opened from south to north, the northern portion is more complicated. Oceanic spreading anomalies C24A, C23 and C22, terminating against the NE Greenland continental slope between Greenland FZ and Jan Mayen FZ may indicate that spreading propagated southward towards the proposed hot-spot location, from the Greenland FZ to the proto-Jan Mayen FZ (Lundin, 2002; Voss and Jokat, 2007). Overlapping spreading centers on the Greenland–Iceland Rise persisted to 20 Ma, with northward propagation along proto-Reykjanes and proto-Kolbeinsey ridges (Lundin, 2002). The Jan Mayen FZ was an oceanic transform connecting the two ridges until abortion of the Aegir Ridge between ~33 Ma and ~22 Ma (C13n - C6B; (Gaina et al., 2009; Gernigon et al., 2009)).

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IS ICELAND A VOLCANIC PASSIVE MARGIN C-BLOCK?

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The Mio-Pliocene crust exposed to the W and E of Iceland shares many characteristics with conjugate volcanic passive margins (C-VPMs) including (1) an over-thickened mafic crust (e.g. Foulger and Julian, 2003), (2) lava wedges analogous to "seaward dipping reflectors" (e.g. Smallwood et al., 1998), (3) large syn-constructional continentward-dipping faults (Bourgeois et al., 2005), in addition to the fissural magma feeding system. The upper-crustal volcano-tectonic system, with magma storage in localized reservoirs beneath large aerial (or sub-glacial) polygenic volcanoes, from which dykes nucleate and propagate laterally outward the active rift, is identical to the 3D volcano-tectonic pattern observed at VPMs (Callot and Geoffroy, 2004; Geoffroy, 2005). In addition to these constructional and tectonic aspects, both Iceland width regarding spreading rate (Foulger, 2006), mafic-crust thickness (Foulger et al., 2005) and chemical composition (e.g. Torsvik et al., 2015) cannot solely be resolved by oceanic-type accretion within an over-productive hot-spot setting. A number of authors propose distinct interpretation to account for Iceland characteristics, involving remelting of ancient subducted lithosphere (Foulger et al., 2005) or contamination by a fragment of continental crust (Torsvik et al., 2015).

High-resolution crustal-scale seismic data from the Southern Atlantic and from the India-Seychelles break-up systems provide new illustration and constrain new numerical models of the mechanisms of break-up in presence of huge magma budget (Geoffroy et al., 2015; Guan et al., 2016). According to those observations and models, the pattern of C-VPMs and their particular crustal structure during stretching and thinning are consecutive of a combined process of large lower-crustal flow and gravitational collapse from apart a highly intruded and lava-covered central continental block (C-Block), progressively dissected with time (Geoffroy et al., 2015).

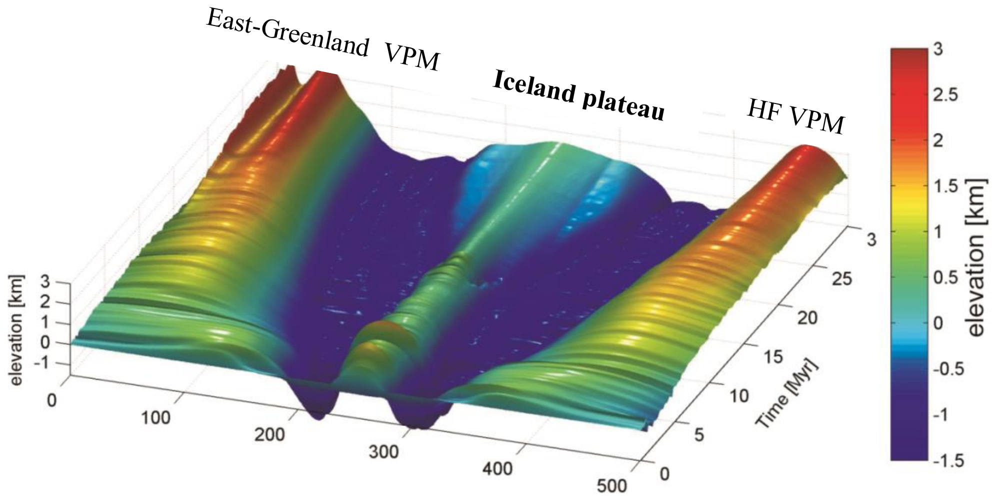

Fig. 1 Basement topography with time of C-VPMs and C-Block (here designed as the Iceland Plateau). Note the widening of C-Block with time as well as its long-term buoyancy. HF : Hatton-Faeroes. From Geoffroy et al. (2015)

The inner part of C-VPMs, is made of thinned and stretched continental crust which develop at high strain-rate during the initial stage of the breakup process. From this "purely continental" stage onward, syn-tectonic and syn-magmatic upper-crustal structures are represented by inner-SDRs which develop "seaward" with time, as the continentward syn-magmatic collapse process continues.

During the next stage, the outer part of the conjugate margin develops from apart the C-Block. This distal volcanic margin is characterized by nearly constant-in-thickness mafic crust. In the upper crustal section, outer-SDRs develop over an enigmatic lower crustal section, probably constituted by highly ductile magma-injected outward-extruded continental lower crust (which is never exhumed). The central continental block (C-Block) is thus finally comprised between outer-SDRs facing the C-Block. Covered with lavas, but remaining much more buoyant than its surroundings, the C-Block is progressively thinned from below by mantle advection whereas it is tectonically and magmatically dissected and stretched in its upper-section (Geoffroy et al., 2015; Fig. 1).

The above description of VPMs evolution would perfectly fit with the first-order tectonic history of (1) East-Greenland/Hatton-Faeroes conjugate volcanic margins, (2) known crustal structure of Greenland-Iceland (GIR) and Iceland-Faeroes (IFR) aseismic ridges and, (3), Iceland characteristics. The E-Greenland costal flexure represents the eroded basement of the earliest inner-SDR (e.g. Brooks, 2011). Away from the necked margin, outer-SDRs develop continuously across the GIR until Iceland where they are onshore exposed. As evidenced during the SIGMA experiments (Holbrook and Dahl-Jensen, 1996), the outer-SDR area extends much more to the south than initially expected, overlapping magneto-anomaly 22 along the Transect III. As outer-SDRs dip towards the central rift zones in W- and E-Iceland, Iceland is probably a highly magma-intruded dislocated C-Block, disconnected by a fault from the northern Jan Mayen micro-continent whose tectonic history differs. The amount and repartition of residual continental material in Iceland is certainly difficult to ascertain, notably because of the frequent shift of the accretion axis with time.

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RIFTING, CONTINENTAL BREAKUP AND POST-BREAKUP EVOLUTION OF THE MID-NORWEGIAN MARGIN

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1. Introduction

While mechanisms of crustal accretion at mid-ocean ridges are fairly well established, relationships between rifting and magmatism and the transition from rifting to spreading (breakup stage) are still poorly understood. In the North Atlantic, the singularity of volcanic segments of passive margins and the formation of microcontinents particularly involve and question mantle dynamics, crustal structure, inheritance and plate kinematics. Based on new potential field and seismic data, we investigated the crustal structure and the rift-to-drift evolution of the Mid-Norwegian margin and the conjugate Jan Mayen microcontinent (JMMC). Thanks to decades of geophysical investigations, the main structural elements of this volcanic and poly-rifted margin are pretty well defined. However, large uncertainties still remain about the structure and petrological nature of the deep crust. The mode of deformation leading to continental breakup and subsequent sea-floor spreading is also unclear and alternative scenarios can be proposed. Based on potential field interpretation and modelling, we first discuss the rifted margin architecture and discuss the meaning of the volcanic continent-ocean transition. Based on new aeromagnetic surveys, we also re-evaluated the post-breakup spreading evolution of the Norway Basin to better understand the conjugate system evolution after the first magmatic phase of breakup. The new data allowed us to refine the tectonic history of the Norwegian-Greenland Sea and discuss some kinematic implications of the syn- and post-breakup development of the JMMC.

2. Pre-breakup crustal architecture of the Mid-Norwegian margin

Before the first breakup development in the Early Eocene, the proto-JMMC was tied to the Mid-Norwegian margin and the East Greenland margin (Figure 1). It was then part of a complex system of sedimentary basins and polyphased rift systems that developed in the NE Atlantic after the collapse of the Caledonian Orogen (*Doré et al.,* 1999*; Skogseid et al.,* 2000*; Tsikalas et al.,* 2012). The long period of rifting (and intra-thinning cooling events), the large thickness of the pre-breakup sedimentary successions and the significant amount of breakup volcanism (seaward dipping reflectors, SDR) make the Mid-Norwegian 'volcanic rifted' margin (Figure 2a) to appear quite different from (Iberian type) magma-poor margins (e.g Boillot and Froitzheim, 2001; Sutra and Manatschal, 2012) (Figure 2b). Fundamental questions and controversial issues concern, however, our understanding of the high-velocity lower crustal bodies (Vp >7.0 km/s) underneath unambiguous crustal domains observed both in the proximal and distal parts of the Mid-Norwegian margin. In volcanic rifted margin settings, their Vp-velocities (alone) can indifferently be interpreted as pre-existing (inherited) lower continental crust, serpentinised mantle and/or breakup-related underplating (e.g. Gernigon et al., 2004; Ebbing et al., 2006; Reynisson et al., 2010; Mjelde et al., 2016). Inherited high-velocity lower crustal bodies are likely present underneath the thick crustal platform and coastal regions (e.g. Kvarven et al., 2014; Nirrengarten et al., 2014). Their geometries in depth seem to have controlled the location, geometry and steepness of the necking zone during the onset of drastic crustal thinning in Late Jurassic-Early Cretaceous time.

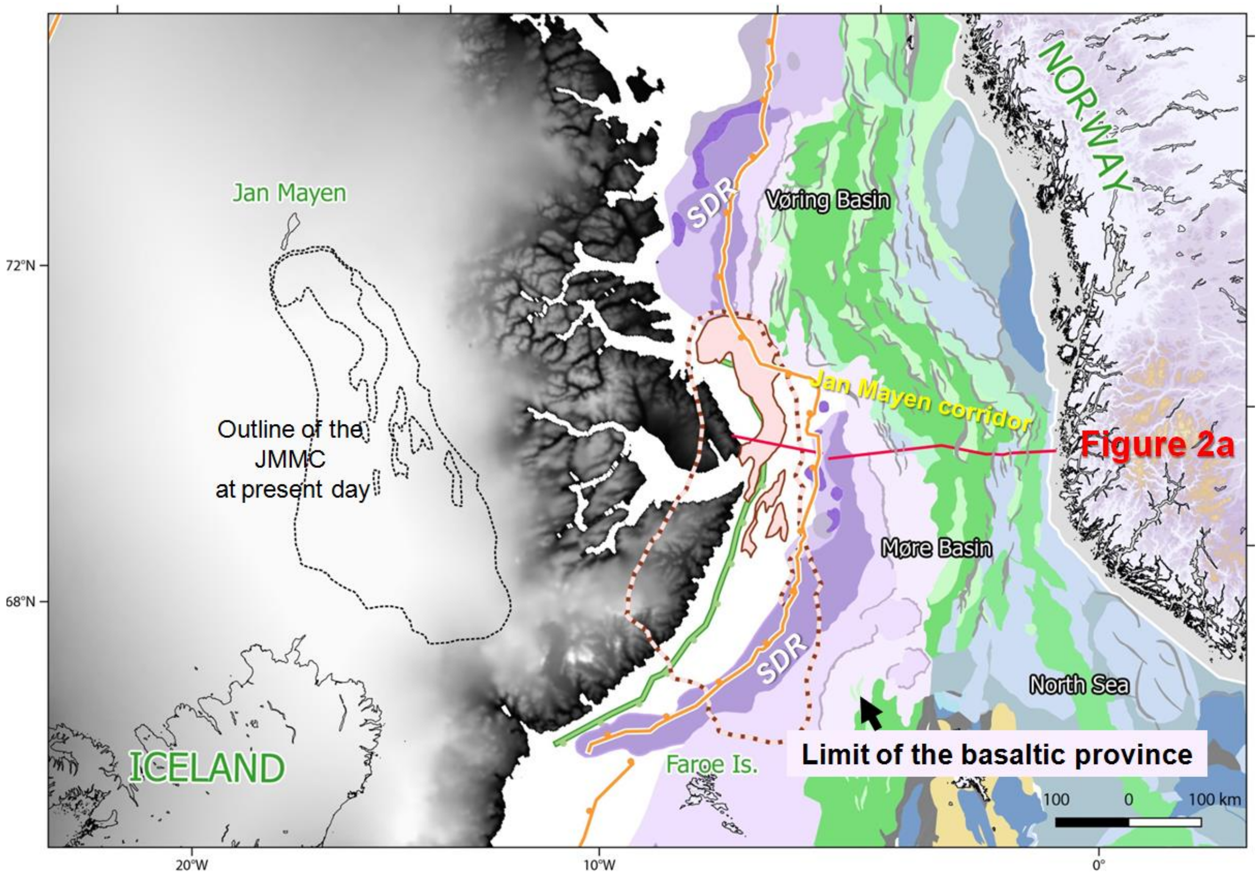

***Figure 1**: Reconstruction for magnetic chron C24n3n time showing tectonic features discussed in this study. The main physiographic elements and the outline of the Jan Mayen microcontinent (JMMC) (undeformed), as defined by the oldest magnetic anomalies recognized at present day, were restored slightly after the first phase of breakup between the European and Greenland plates. A significant overlap exists between the present-day outline of the JMMC and the original space allocated and available between Greenland and the volcanic rifted margin at magnetic chron C24n3n (~54 Ma) (dotted-pink JMMC outline).*

Presence of high-velocity/high-density material inherited from the Caledonian and/or Precambrian age is also expected underneath the Møre and Vøring basins (Figure 2a). The continental crust preserved on top of high-velocity lower crustal bodies (Vp >7.0 km/s) recognised in the central and outer parts of the rifted margin remains locally quite thick (> 5-8 km) to favour a broad zone of exhumed and denudated serpentinized mantle. Modelled Vp velocities of 7.1-7.3 km/s for the underlying lower crustal bodies would possibly require an increased process of serpentinisation that usually is symptomatic of more drastic crustal thinning ($ß_{crustal}$ >> 4) and/or complete mantle denudation (see Figure 2b). Such a crustal configuration is not so obvious and controversial in large parts of the Mid-Norwegian margin where wide crustal rafts, continental mullions and external marginal plateau have been interpreted and tested by potential field modelling (Figure 2a). We favour a "super-extended" tectonic scenario where a large part of the crust observed underneath the sedimentary basin dominantly represents both preserved inherited upper to lower continental crust. Underneath the very thick sedimentary successions, only local windows of serpentinised mantle (or alternatively exhumed and 'boudined' lower crust) could possibly have exhumed underneath the metasedimentary rocks, already highly deformed and laminated in the deepest parts of the basin (e.g. Træna Basin, Rås Basin, Vigrid Syncline, Jan Mayen corridor). Seismic data and potential field modelling suggest that the size and thickness of the preserved continental blocks/rafts (e.g. Utgard High, Vigra High, Slettringen Ridge) are much larger and thicker to be compared with allochthonous blocks often described in true 'hyper-extended' domains (e.g. the Iberian magma-poor margin, see Figure 2b). Our potential field modelling also suggests the preservation of sub-sag middle crustal material with relatively high magnetic susceptibilities associated with some of the main crustal rafts. It correlates with similar basement properties observed and modelled in their original footwalls. Closer to the continent-ocean transition, shallow to sub-aerial paleowater depths conditions before and during the onset of breakup and evidences of continental contamination of the erupted lava flows (Meyer et al., 2007; Abdelmalak et al., 2016) support the presence of continental crust beneath the basaltic sequences. The petrology of the recovered basaltic rocks does not easily agree the presence of a large zone of the exhumed and denudated mantle material lying just underneath the first basaltic layers and late Inner SDR.

Another concern with the Mid-Norwegian margin is the extremely long (>250 Ma) and complex polyphased rift history of the inherited crust described before the volcanic margin formation. In terms of timing, it is important to remind that the breakup of the Norwegian-Greenland Sea is Early Tertiary in age and does not necessarily represent the ultimate stage of a continuous and severe crustal and lithospheric thinning phase that initiated more than 90 m.y. earlier, in Late Jurassic time. Assuming that the rifted margin already experienced a drastic phase of thinning and mantle exhumation in the earliest Cretaceous time (scenario from Péron-Pinvidic and Osmundsen, 2016), it might be difficult (for us) to explain how such a mature rift system could still accommodate both active thinning and mantle denudation over a continuous minimum period of time of ~ 50-60 m.y. from Early Cretaceous to the final breakup in Latest Paleocene-Early Eocene time. By comparison with the Iberian margin and other rifted margins worldwide (whether volcanic or magma-poor end-members) breakup usually occurs rapidly after the denudation stage and/or earlier weakening of the plate (e.g. Tucholke and Sibuet, 2007; Brune et al., 2016). In the distal part of the Mid-Norwegian margin, we believe that it would be difficult to maintain a continuous thinning/exhumation phase for more than 5-30 m.y. without reaching an early breakup in mid. Cretaceous (which is not observed!). How to explain an extremely shallow exhumed mantle underneath the Vøring and Møre Marginal Highs before the Inner SDR and significant underplating emplacement is also questionable. An 'Iberian scenario' applied to the distal part of the Mid-Norwegian margin appears difficult to support.

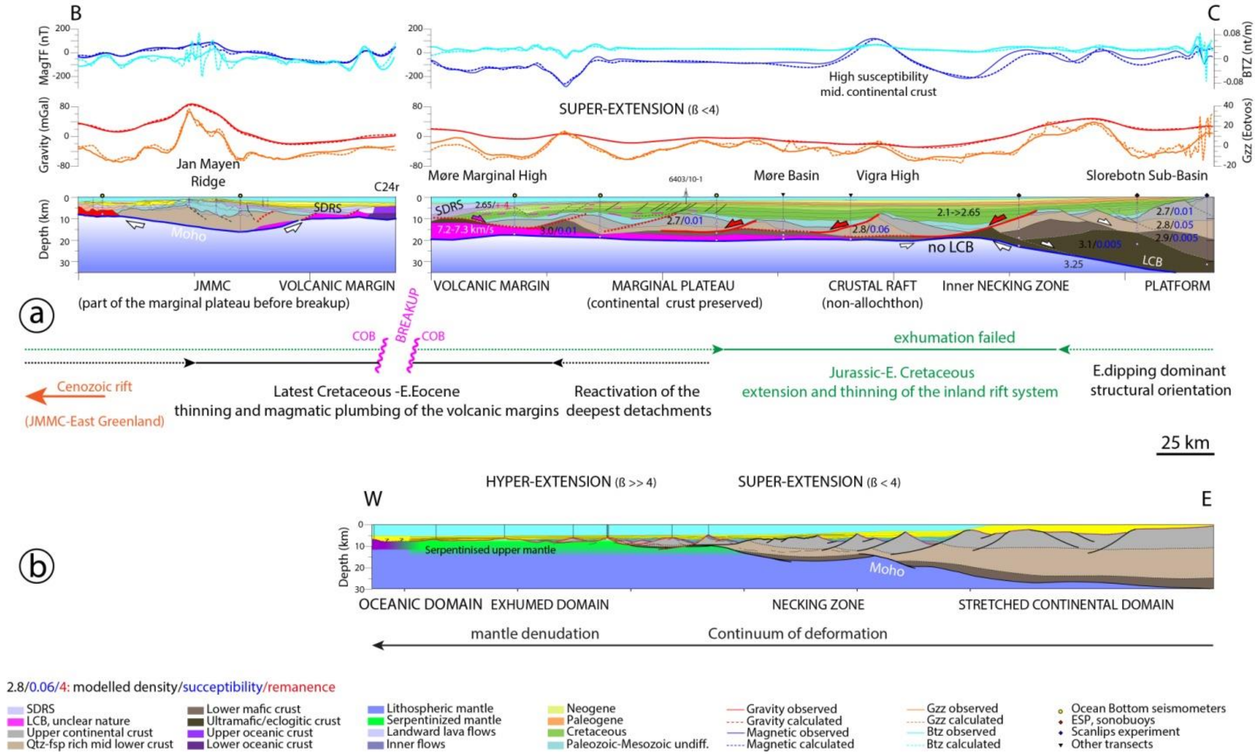

Figure 2*: a) Conjugate volcanic rifted sections between the Mid-Norwegian margin (Møre segment) and the JMMC (section a) located in Figure 1). b) Comparison (the same 1×1 scale) with the classic Iberian archetype proposed for hyper-extended and magma-poor margins (section after Sutra et al. 2013). Seismic refraction/reflection observations and potential field modelling have been used to constrain the main crustal geometries and tectono-magmatic domains, offshore Norway.*

We preferentially interpret the nature of the deep crust and lower crustal bodies closer to the SDR as a mixture of residual and inherited continental crust later affected by breakup-related intrusions (e.g. Gernigon et al., 2004, 2006; Ebbing et al., 2006; Mjelde et al., 2016) (Figures 3 and 4). A renewed phase of extension and normal faulting occurred during the latest Cretaceous and Paleocene in the outer part of the pre-existing rift axis. It most likely affected an outer crustal domain partly preserved and less thin by the previous Late Jurassic-Early Cretaceous thinning phase of the rifted margin (e.g. a conjugate necking zone?). This renewed phase of stretching preceded a second magmatic-tectonic thinning event concomitant with diachronic SDR development in the Latest Paleocene-Early Eocene period. We interpret the outer part of the Mid-Norwegian margin as the remnant of a continental marginal plateau with preserved (Meso-Paleozoic?) terraces lying, together with the JMMC, on the northern extension of the Fugløy Ridge/Faroe block, which shows thicker continental crust preserved (*White et al., 2008*). The late magmatic phase of the rifted margin development was extremely localised and may have involved independent and decoupled magmatic-tectonic processes including massive crustal dilatation by diking and crustal plumbing and/or magmatic rift localization triggered and/or pulsing with nascent underplating (Figure 3)(*e.g. Callot et al.*, 2001*; Yamasaki & Gernigon*, 2009; Geoffroy et al., 2001, 2015; Keir et al., 2013).

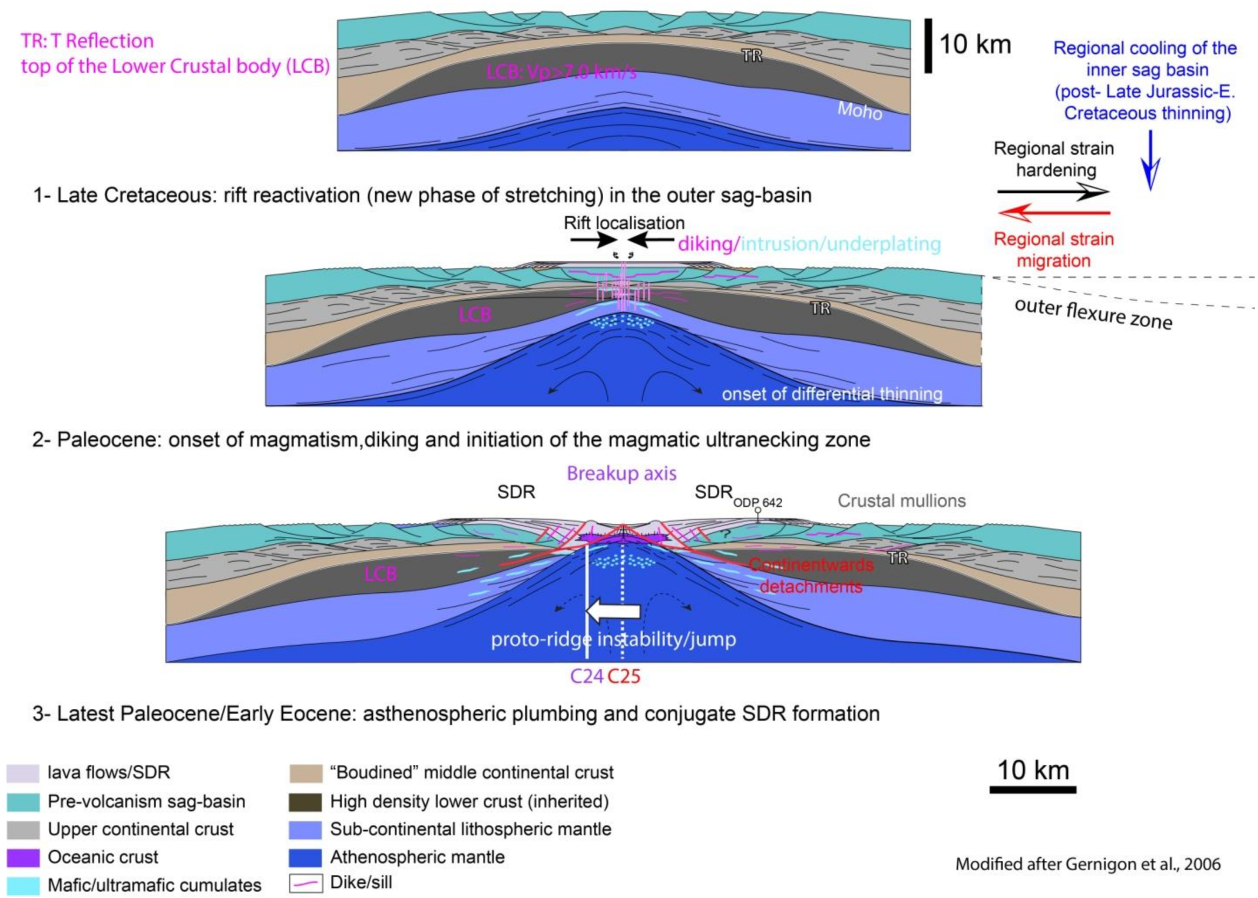

Figure 3*: Onset of breakup scenario and volcanic margin formation in the outer parts of the Mid-Norwegian margin.*

3. Post-breakup and spreading evolution

Except for a previous survey that focused on the extinct Ægir Ridge (*Jung and Vogt*, 1997), the sparse distribution and poor quality of the vintage existing magnetic profiles in the remaining part of the Norway Basin (line spacing >30-50 km) was a serious and primary impediment for accurate interpretations of the spreading history between the Mid-Norwegian margin and the conjugate JMMC (e.g. *Gaina et al.,* 2009; *Olesen et al.*, 2010*; Gernigon et al.*, 2012, 2015). Consequently, new surveys were needed to test and validate controversial tectonic models about the JMMC evolution (*e.g. Scott et al.*, 2005). Between 2009 and 2012, we have acquired and processed more than 88.000 km of new aeromagnetic data over the entire oceanic domain located between the Møre volcanic rifted margin, the Vøring volcanic transform margin and the JMMC (Figure 4). The spreading system is now fully covered by high-quality and high-resolution magnetic data (Figure 4).

The new magnetic compilation illustrates new oceanic features and documents in detail the magnetic polarity chrons of the Norway Basin (Figure 4). The new aeromagnetic compilation confirms, in particular, that fan-shaped spreading of the Norway Basin was clearly active after breakup and before the cessation of seafloor spreading and extinction of the Ægir Ridge in Early Oligocene time (~22 m.y. ago). Faster spreading rates (Figure 5) are recorded in the northern part of the Norway Basin and slower spreading rates near the Faroe-Iceland-Ridge which is located closer to the main set of rotation poles. The degree of melting is not directly related to any linear spreading rate development. Melt production is more dominant in the southern Norway Basin closer to the thick Faroe-Iceland-Ridge. A similar situation might be expected during the development of the volcanic margins.

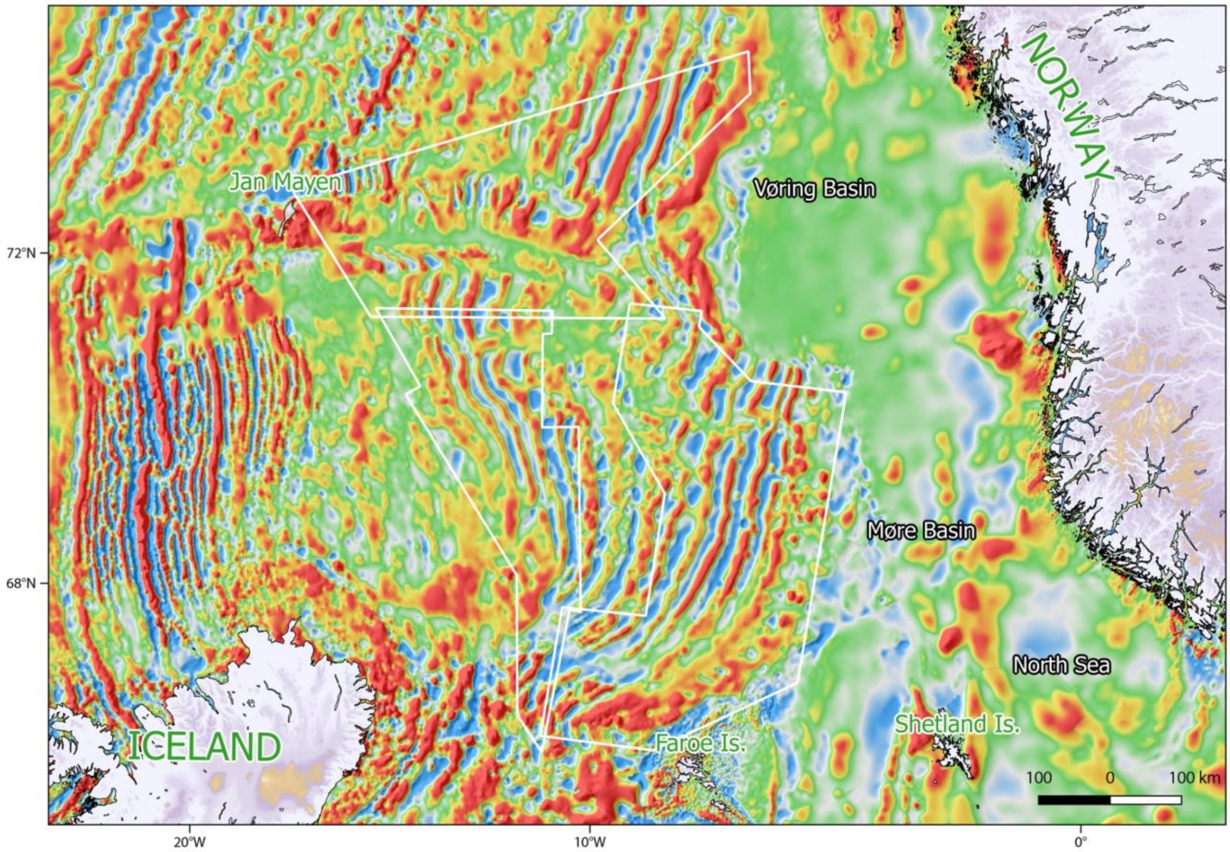

Figure 4*: New regional magnetic compilation of the Norway Basin. A total of ~88000 km of new magnetic profiles greatly improve the magnetic coverage of the Norway Basin and surrounding volcanic rifted margins. Details of the new surveys (white polygons) are further documented in Gernigon et al. (2015).*

The new magnetic data compilation (Figure 4) shows that accretion developed 'faster' on the eastern side of the Norway Basin. Both a differential spreading rate and possible episodic ridge jumps may explain the spreading asymmetry (Figure 5). Even if large uncertainties remain about pre-C24B magnetic chron picking, embryonic oceanic domain at C25 has been locally interpreted at the edge of the oldest SDR (Late Paleocene in age in the northern Møre Basin?). We estimated low spreading rates during the proto-breakup stage (<5-10 mm/year), increasing suddenly and rapidly after breakup to more than 15-25 mm/year in the early stage of steady-state seafloor spreading (Figure 5). New oceanic fracture zones have also been mapped. Some of them initiated at breakup time and some formed later in the Eocene. The JMMC was not entirely rigid during spreading of the Norway Basin, but also experienced significant stretching that increased from north to south during the fan-shaped-spreading development of the Norway Basin. The northern and southern parts of the Norway Basin, in particular, are divided by an important, small-offset fracture zone which appears quite different from the straight and classic appearance of a 'normal' fracture zone (Figure 4). This discontinuity shows a complex magnetic pattern characterized by discrete, non-continuous and oblique magnetic lows and highs interpreted as a kind of oceanic 'pseudo fault' that linked two separate oceanic spreading systems in the Norway Basin. The two segments and the 'pseudo fault' are the direct consequence of the distinct time and space development of (at least) two sub-plates bordered by the Jan Mayen Ridge and South Jan Mayen Ridge Complex (Figures 1 and 6).

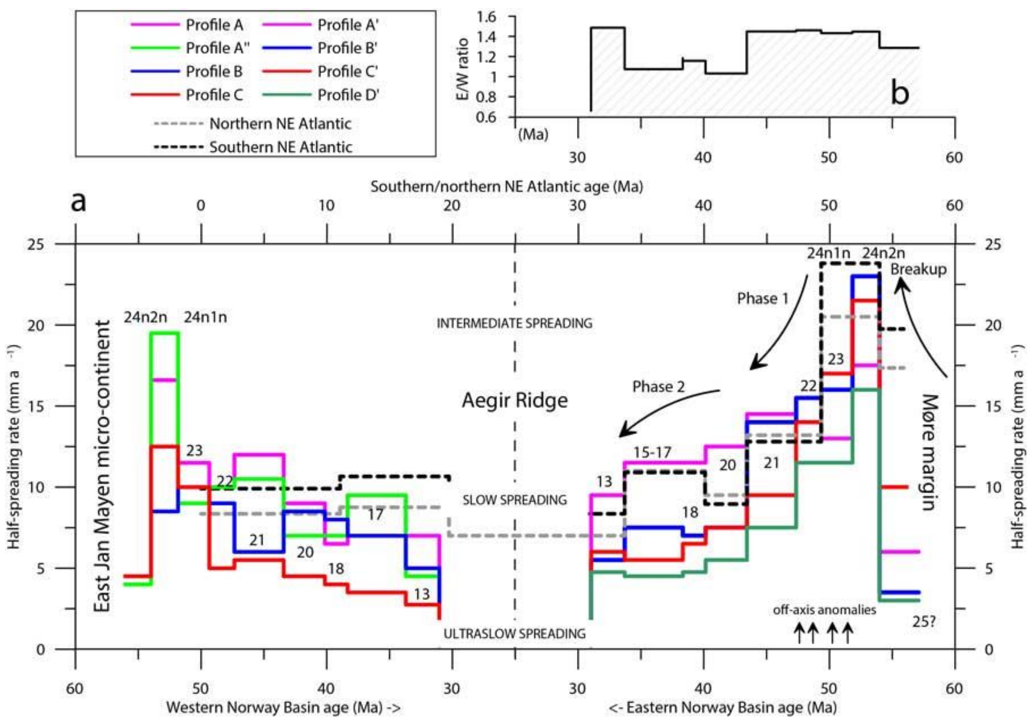

Figure 5: *a) Diagram showing the variation with time of the Norway Basin spreading system. Average spreading rates for the adjacent northern and southern NE Atlantic oceanic domain (grey and black dashed lines) from Gaina et al. (2009) are plotted for comparison. b) Plot of the spreading-rate ratio between the eastern and western parts of the Norway Basin. The diagram illustrates the general asymmetry of the Norway Basin with higher spreading rates observed in the eastern side of the spreading system.*

4. Jan Mayen 'microcontinent' formation

Located north of Iceland (at present day), the JMMC is often interpreted as an isolated continental fragment which was detached from the East Greenland margin and the Mid-Norwegian margin during the Cenozoic (T*alwani and Eldholm*, 1977*; Auzende et al.*, 1980*; Vogt et al.*, 1980*; Unternehr*, 1982*; Nunns*, 1983*; Skogseid and Eldholm*, 1987*; Guðlaugsson et al.*, 1988*; Gunnarsson et al.*, 1989*; Kodaira et al.*, 1998*; Breivik et al.*, 2012*; Kandilarov et al.*, 2012). The JMMC represents a complex intermediate conjugate system including two sets of conjugate margins. Previous interpretations based on ocean-bottom seismographs (OBS) and existing multi-channel reflection-seismic surveys on the JMMC suggested that the presumed continental basement is overlayed by Palaeozoic? to Cenozoic sediments (Kuvaas and Kodaira, 1997). The early history of the JMMC is traditionally associated with the first phase of continental breakup and many studies have considered that the most important tectonic event that influenced the Norwegian-Greenland Sea after the initial rupture occurred in the Late Oligocene. During Oligocene time, the spreading activity along the Ægir Ridge decreased until it became extinct and 'jumped' westwards to initiate the Kolbeinsey Ridge. The relocation of the spreading ridge from the aborted Ægir Ridge to the nascent Kolbeinsey Ridge resulted in the final separation of the JMMC from the Greenland Plate.

Although Müller et al.'s (2001) suggested mechanism for the ridge jump/propagation could be successfully applied to the Norwegian-Greenland Sea (explaining post-10 Ma ridge jumps towards the Iceland hotspot), the amount and timing of extension and magmatic episodes related to the formation of the JMMC is less well constrained. The nature and timing of these events and the implications for the tectonostratigraphic evolution of the sedimentary basin have been correlated with a regional kinematic model updated by means of the new aeromagnetic data. We show that no real 'jump' actually exists but instead there was probably a complex propagating and overlapping rift system leading progressively to the second phase of breakup between the JMMC and East Greenland. Despite a recent refraction survey north of Iceland (*Brandstóttir et al.,* 2015), it remains difficult to clearly identify and to fully understand the nature of the crust between the Ægir and Kolbeinsey ridges. Taking into consideration the proto-JMMC space issue when it was restored at breakup time (Figure 6), we inferred that more than 400% of post-breakup extension must have occurred in the southernmost part of the JMMC (Gernigon et al., 2015). Plate reconstruction suggests that the central and northern parts of the proto-JMMC could have been accommodated in the remaining space available between the Norwegian and Greenland continental plates at almost breakup time (Figure 1). However, the southern part of the JMMC, as observed at the present day, could not fit in the space available between Norway/Eurasia and Greenland. Whatever the nature of the crust in the proto-JMMC was (continental, oceanic?), this plate overlapping issue confirms that younger and significant deformation affected the JMMC after the first phase of breakup in the Norwegian-Greenland Sea, but before the final dislocation of the JMMC and opening of the Kolbeinsey Ridge at around C6b (22.2-21.7 Ma) (e.g. *Vogt et al.*, 1980*; Gaina et al.*, 2009). In this context, the integrity and preservation of massive continental units might have been seriously disrupted by possible oceanic accretion even before the complete formation of the Kolbeinsey Ridge and JMMC formation.

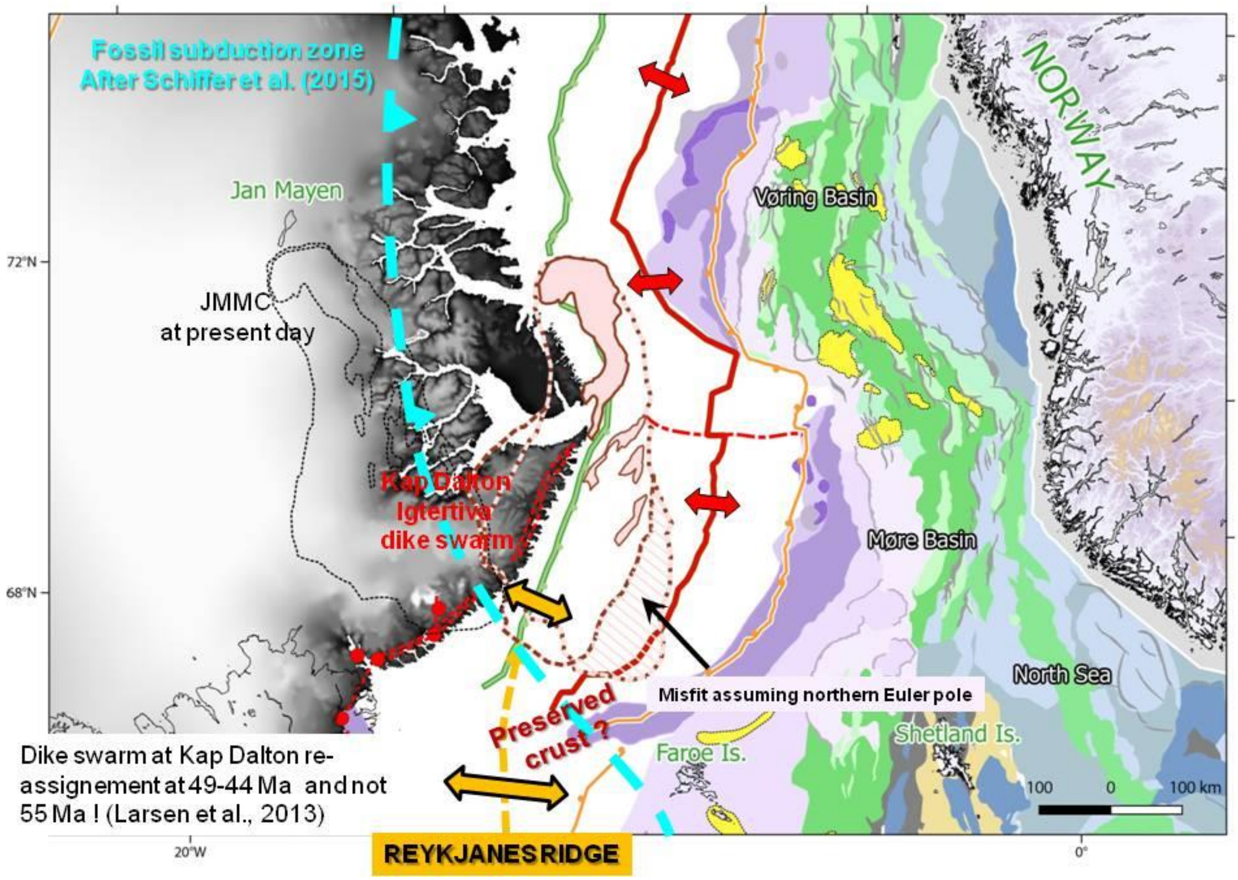

Figure 6: *Reconstruction at C21n time (old edge ~47.9 Ma) showing tectonic features and continental fragments involved in this study. This Mid-Late Eocene event coincides with an important phase of reorganization in the Norwegian-Greenland Sea. Chron C21n marks the onset of serious extension into the southern JMMC in order to counter-balance the subsequent growth and fan-shaped development of the Norway Basin spreading system. This period should coincide with the onset of extension between East Greenland and the proto-JMMC in front of the propagating Reykjanes Ridge located to the south. Rift bifurcation between the proto-JMMC and East Greenland and the second line of breakup could have been influenced by the Greenland Central Fjord-Flannan fossil subduction zone (Schiffer et al., 2015). The main physiographic elements of the JMMC (undeformed) have been restored at that stage based on the two sets of rotation poles defined between the JMMC and the Norwegian plate (fix).*

The origin and location of this rift leading to the second phase of breakup and final isolation of the ambiguous JMMC are also questionable. The action of a mantle plume, by 'thermal weakening' was proposed earlier to explain microcontinent formation (*Müller et al.*, 2001). However, in our opinion, complex rift/ridge overlap and a dual rift system geometry between the dying Ægir Ridge and nascent Kolbeinsey Ridge was probably the simplest and main tectonic process initiating the formation of the microcontinent. Natural rift overlapping can easily explain the formation of microplates without involving any mantle plume (*e.g., Auzende et al.*, 1980; Gernigon et al., 2012; Ellis and Stoker, 2014). Progressive rift migration towards a plume axis is also debatable (see http://www.mantleplumes.org/Iceland4.html). A rift overlap could have been favoured by the initial crustal configuration and the presence of inherited crustal "buffers" such as the Faroe block (Gernigon et al., 2012) and possibly similar continental terranes thought to underlie Iceland (Foulger et al., 2005). Such 'overlap' model would also involve a preserved but intensely intruded zone of continental residue between the Ægir and the Reykjanes ridges, which never really fully connected in such a scenario (Figure 6). More questionable remains, however, the final isolation stage of the JMMC. It was preceded by propagating magmatic and diking events that initiated (at least) 7-8 m.y. before the establishment of a stable oceanic accretion system between the JMMC and East Greenland (Tegner et al., 2008; *Larsen et al.*, 2013). The onset of magmatism coincides with a clear spreading reorganisation of the Norway Basin around C21n (Figure 6). Therefore, the origin and tectonic implication of magmatism (related or not to a hypothetical mantle plume; Foulger and Anderson, 2005*)* cannot be neglected and probably contributes to the breakup and indirectly to microcontinent formation.

Finally, we note that the second line of breakup between proto-JMMC and Greenland could have been influenced by the outline of the Greenland Central Fjord-Flannan fossil subduction zone recently imaged by Schiffer et al. (2015). Deep inherited orogenic structure could have played a major role in the JMMC formation and North Atlantic magmatic and rifting processes in general (e.g. Ryan and Dewey, 1997; Foulger et al., 2005).

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CONTROLS OF MANTLE POTENTIAL TEMPERATURE AND LITHOSPHERIC THICKNESS ON MAGMATISM IN THE NORTH ATLANTIC IGNEOUS PROVINCE.

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For primitive basalts that have crystallized only olivine, reverse-modelling involving the incremental addition or subtraction of equilibrium olivine to the measured bulk-rock composition allows the generation of a suite of potential parental melt compositions (Herzberg *et al.*, 2007). Comparison of parental melts compositions with primary melt compositions determined from forward models of peridotite melting (Fig. 1), allows T_P that is required to generate the melt to be estimated (Herzberg & Asimow, 2008; 2015). The initial pressure of intersection of the dry peridotite solidus (P_i) can be determined from T_P. Mantle peridotite partially melts at low melt fractions to produce melt droplets that mix during decompression melting to produce an "aggregate" or accumulated fractional melt (*AFM*). The final melting pressure (P_f), which represents the pressure at which the last drop of melt was produced, can be estimated from the melt fraction and P_i. P_m, the mean pressure of melting, represents the melt-column average pressure of melt generation for a particular location. The PRIMELT3 melting model (Herzberg & Asimow, 2008, 2015) has been applied to basalts from across the North Atlantic Igneous Province, the results of which allow constraints to be places not only on T_P during magmatism, but also on melt-fraction, P_i and importantly P_f. P_f is a particularly useful parameter because it allows inferences to be made about the depth to the base of the lithosphere at the time of magmatism, since decompression melting of peridotite must cease close to the asthenosphere-lithosphere boundary (Fig. 2).

Modelled primary magma compositions of Palaeocene basalts from the North Atlantic Igneous Province (NAIP) require melting at mantle potential temperatures (T_P) in the range 1480-1550°C (Fig.2). Modern lavas from the Icelandic rift-zones required T_P~1500°C and those from the rift-flanks T_P~1450°C (Hole & Millett, 2016; Hole, 2015; Hole *et al.*, 2015). Secular cooling of the NAIP thermal anomaly was therefore in the order of ~50°C in 61 Ma. There were systematic variations in T_P of 50-100°C from centre of the thermal anomaly to its margins at any one time, although limits on the stratigraphical distribution of T_P determinations do not rule out thermal pulsing on a timescale of millions of years. Variation in extent of melting at similar T_P was controlled by local variability in lithospheric thickness. In the west of the NAIP, lithosphere varied from ∼90 km at Disko Island to ∼65 km at Baffin Island, with similar thickness variations being evident for magmatism in the Faroe Islands, Faroe-Shetland Basin and the British Palaeogene Igneous Province (BPIP). Mean pressure of melting ≥ final pressure of melting and the two values converge for melting columns with a melting interval of < 1.5 GPa, regardless of T_P. In particular, the majority of BPIP magmas were mostly generated in the garnet-spinel transition in the upper-mantle. Calculated and observed rare earth element distributions in NAIP lavas are entirely consistent with the melting regimes derived from major element melting models. This allows a calibration of rare earth element fractionation and melting conditions that can be applied to other flood basalt provinces.

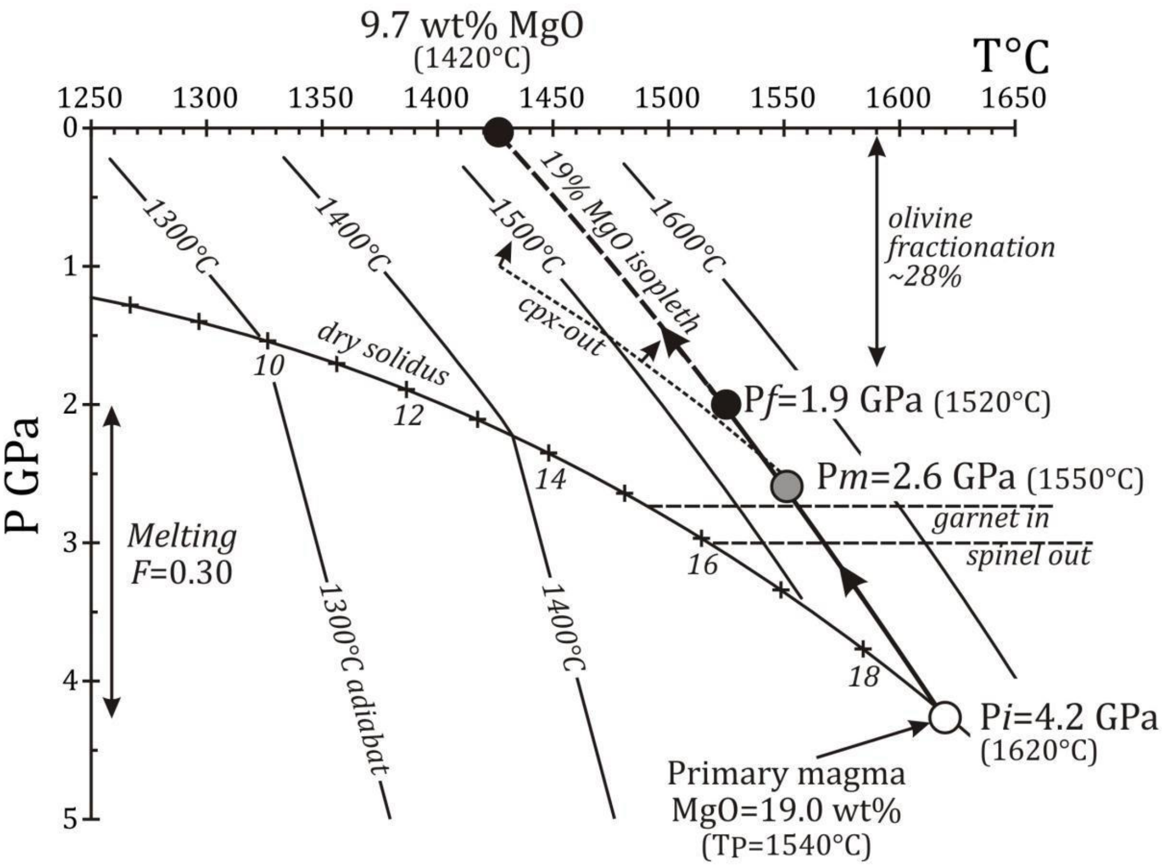

Fig. 1. Schematic pressure-temperature diagram illustrating the petrogenesis of an olivine-phyric basalt from Baffin Island (BI/CS/8) with 9.7 wt% MgO which fractionated 28% olivine and erupted with a liquidus temperature of ~ 1420°C. The primary magma to this basalt is calculated to contain 19.0 wt% MgO, requiring T_P ~ 1540°C. Pi = 4.2 GPa, and is calculated from the intersection of the dry peridotite solidus and the 1540°C adiabat. The primary magma ascended along the olivine liquidus as indicated by the arrows, which also corresponds to the 19.0 wt% MgO isopleth. The extent of melting, FAFM ~ 0.30, is calculated from phase equilibria. The final pressure of melting, Pf, is estimated to be 1.9 GPa for this sample. The mean pressure of melting Pm = 2.6 GPa at 1550°C. Note that initial melting takes place in equilibrium with garnet peridotite, but the majority of melting takes place in the spinel stability field of the upper-mantle

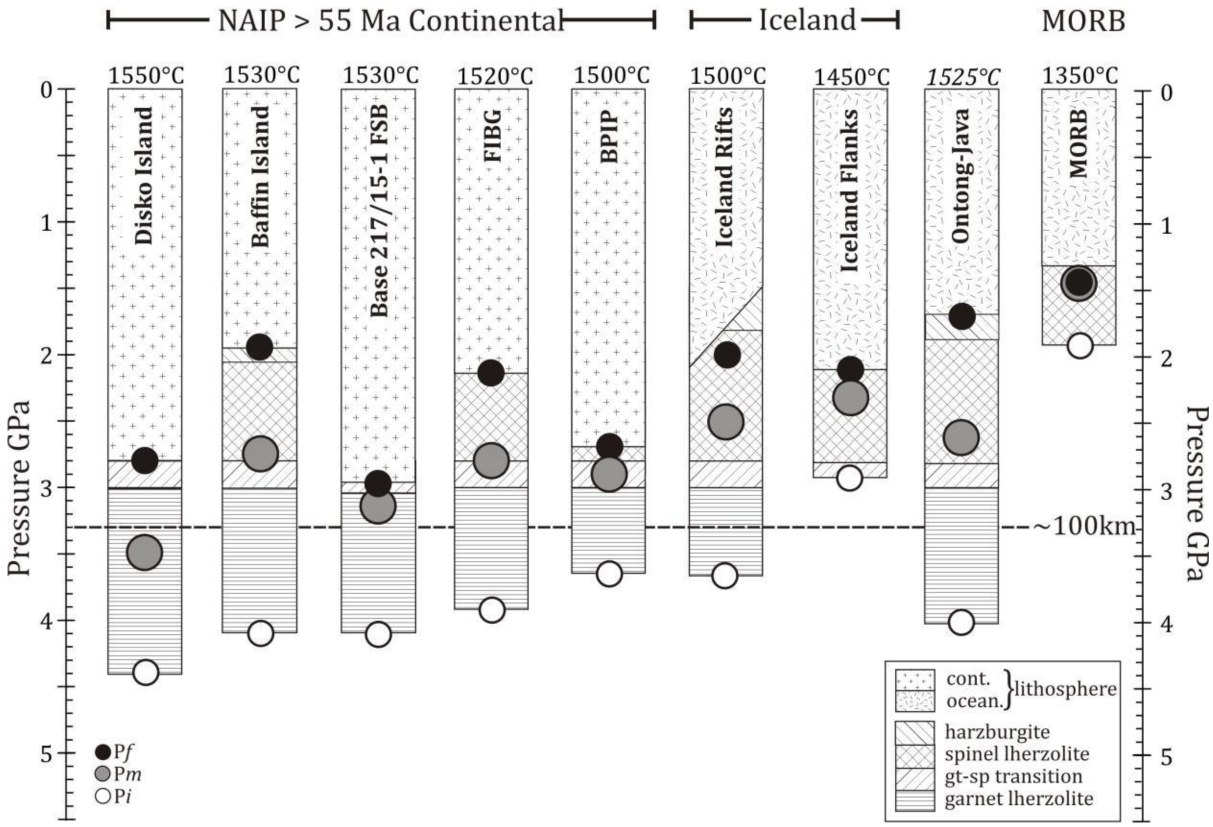

Fig. 2. Diagrammatical representations of melting columns for various locations in the NAIP, MORB from the Siqueiros Fracture Zone, and basalts from the Ontong-Java Plateau. Pi, Pf and Pm are melting column-averaged data for the location named. Melting with harzburgite as the residue is not directly correlated with pressure of melting but with extent of melting and so the harzburgite fields are schematic. Note that Siqueiros MORB, Iceland rift-flanks, BPIP and Faroe-Shetland Basin melting columns have Pf ≈ Pm.

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FROM ICELAND TO THE BIGHT TRANSFORM FAULT; A TALE OF TECTONICS, VOLCANISM AND PLATE BOUNDARY REFORMATION

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In this talk we shall discuss observations made during two oceanographic missions focusing on the evolution of the Mid-Atlantic ridge south of Iceland, namely the Reykjanes ridge. First mission in 2007 was focusing on its relation to Iceland and its evolution during the past 20 Ma. The second mission was in 2013 and focused on the termination of the Reykjanes ridge and its relation to transform faults.

The Reykjanes ridge (RR) extents for about 900 km south of Iceland and is the longest straight segment of the Mid-Atlantic ridge. Our investigation shows that RR has gone through several reorganisation periods since the abandoning of the Ægir ridge. Prior it had been suggested that the plate boundary had moved twice in the Iceland region, which is the Westfjord rift (WR) and the Snæfellsnes rift (SR). However, our investigation suggest that there are two more abandoned rift system in Iceland, the Husavíkurkleif rift, active between WR and SR, and the Hvalfjordur rift, active between SR and the current rift system in Iceland.

The currently active RR plate boundary shows that the first main central volcano is to be found as the ridge extends onto the Iceland continental shelf, the volcano has the name Njordur central volcano.

Comparison study between the Bridge mission, the 2007 mission and our last mission in 2013 shows that bathymetry data can be used to estimate geological changes on the ridge in time. At about 62°10´N we observe a major seamount being created between the Bridge mission and our 2013 mission, this suggests volcanic activity in the area.

Our last mission in 2013 was more focused on the southern extremity of the RR and its connections with the Bight transform fault (BTF). Some of our data from this mission show how the RR extension to the south has decoupled the transform faults. However the scar of the faults can still be observed along the main plate boundary. In relation to the transform faults we observe several oceanic core complexes, indicating rotation of the crust and small magma activity. Further as the RR comes closer to the BTF the ridge form of the boundary changes to rift valley bounded by up to 1000 m high fault scarps. Volcanism in the rift valley is segmented and bounded by the pre-existing transform faults.

As we approach the BTF and cross it volcanism on the ocean floor is less focused and tends not to be confined within the active rift boundary. Thus we observe relatively fresh volcanic features at a distance of some 50 km away from the active plate boundary.

Lastly, bathymetry reveals that oceanic currents are strong at the floor level, strong enough to build up dunes that are up to 10 to 20 m high. This might be one of the main reasons why it has proven to be difficult to locate black smokers in the area by chemical sniffing.

SEISMIC TOMOGRAPHY IMAGES OF THE NORTH ATLANTIC

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Seismic tomography is often used to study the structure of the upper and lower mantle, and is the tool of choice for imaging structure beneath “hot spots”. A second method widely applied to study the depth extent of possible thermal anomalies is using receiver-function analysis to measure topography on the transition-zone bounding phase-change discontinuities at 410- and 660-km depths.

Many papers have been published showing tomography results for the Iceland/North Atlantic region. This region is relatively easily studied using tomography (compared with, say, Hawaii) because seismic stations can be deployed on the large landmass of Iceland and the continental areas of Greenland, Scandinavia and Europe. All studies agree that a negative wave-speed anomaly underlies the North Atlantic beneath and around Iceland, but they disagree about the depth extent and interpretation of this feature.

Several papers present images extending from the surface to the core-mantle boundary that are interpreted as hot mantle plumes [e.g., Bijwaard & Spakman, 1999; French & Romanowicz, 2015; Rickers *et al.*, 2013] (Figure 1a, b, d). Other papers have presented images interpreted as showing that the anomaly is confined to the upper mantle and does not extend through the base of the transition zone at 660 km depth and into the lower mantle [e.g., Foulger *et al.*, 2000; 2001; Hung *et al.*, 2004; Ritsema *et al.*, 2011] (Figure 1c and Figure 2). Similarly, receiver function results aimed at studying topography on the transition-zone discontinuities have variously been interpreted as showing a through-going high temperature anomaly [e.g., Shen *et al.*, 2002] or topography consistent with normal temperatures at 660 km depth [Du *et al.*, 2006].

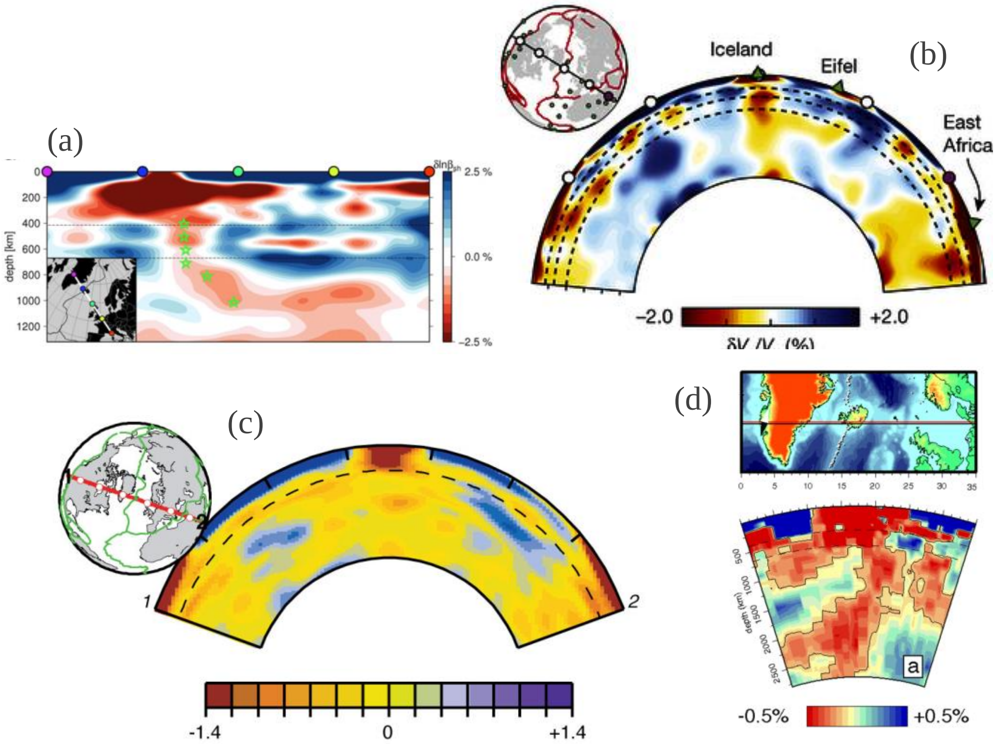

Figure 1: A choice of Icelands. Whole-mantle tomography results from (a) Rickers et al. [2013]; (b) French and Romanowicz [2015], (c) Ritsema et al. [2011], and (d) Bijwaard and Spakman [1999].

Two main problems may be highlighted:

1. The repeatability of both seismic tomography and receiver function studies is poor (Figure 3);
2. Seismic studies yield *seismic* parameters. Interpretation is often non-unique and it cannot be assumed that low wave-speeds correspond directly to high temperature.

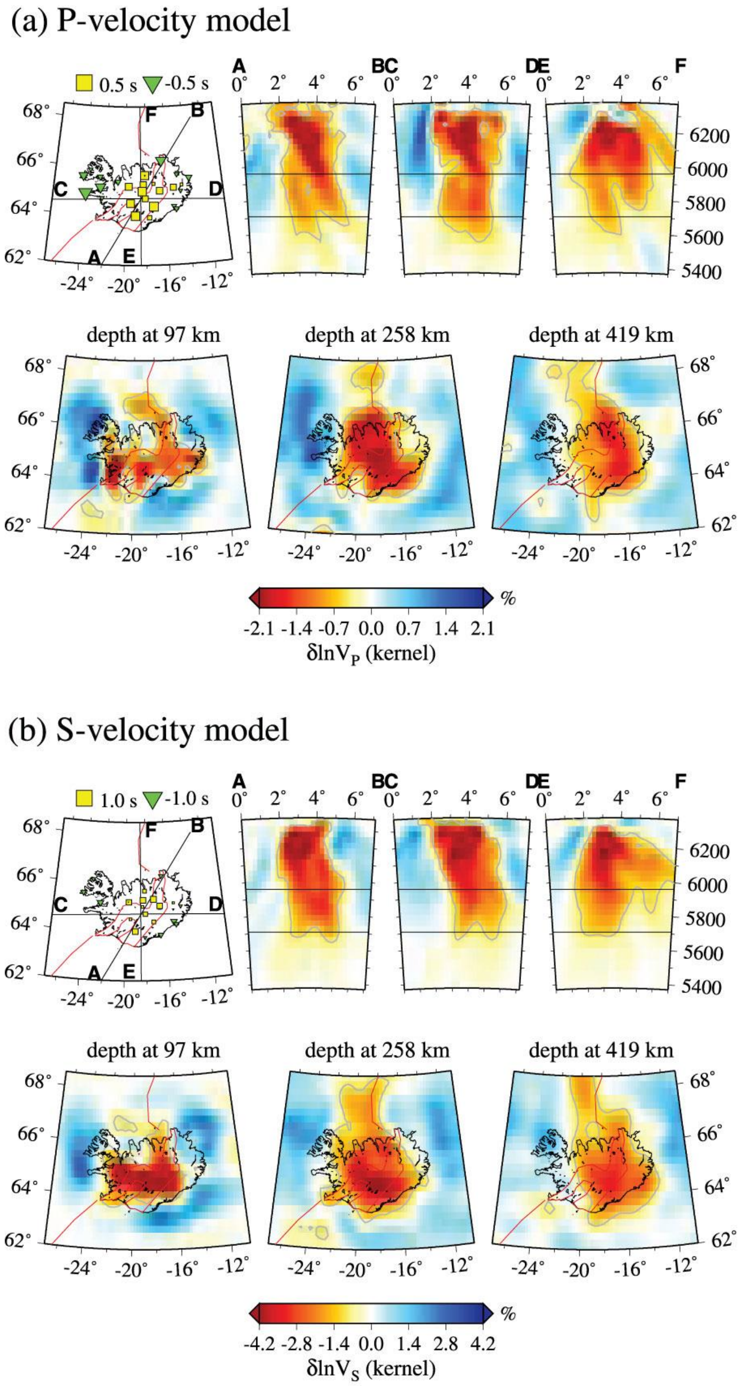

Figure 2: Morphology of the negative wave-speed anomaly beneath Iceland from teleseismic tomography [from Hung et al., 2004]

In the case of tomography, repeatability generally becomes worse with increasing depth in the mantle, because of seismometer deployments are largely limited to land masses, and seismic-ray sampling of the volume of interest is non-uniform. Examples of the limitations of seismic tomography include the facts that teleseismic tomography cannot image the three-dimensional structure of the mantle, and tomography cannot determine with certainty the strengths of calculated anomalies. Despite this, published maximum seismic anomaly strengths are often unjustifiably translated directly into physical parameters. Tomography yields seismological parameters such as wave speed and attenuation, not geological or thermal parameters. Much of the mantle is not well sampled by seismic waves, and resolution- and error-assessment methods do not express the true errors. Additional problems involve theory, correcting for the crust, the choice of background model to subtract to reveal relative anomalies, the difficulty of retrieving absolute wave speeds and choices regarding what images to select for publication.

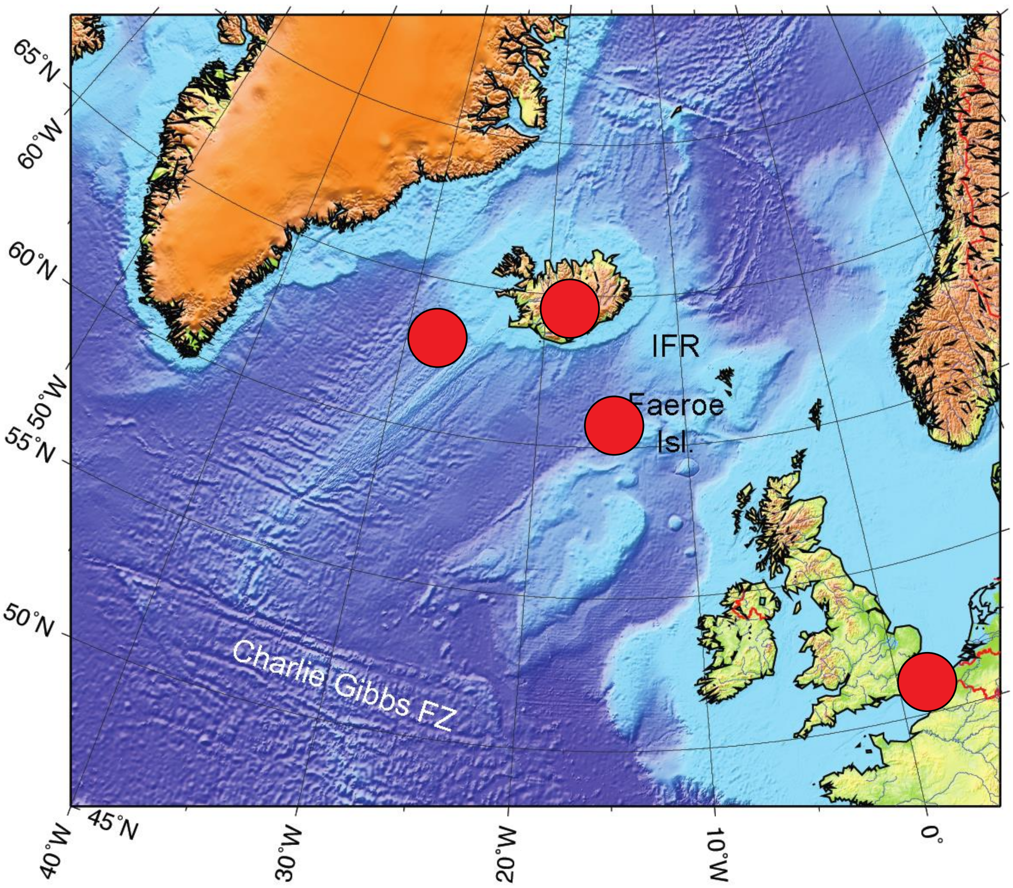

Figure 3: Locations at the core-mantle boundary where a plume erupting in Iceland is postulated, on the basis of seismic tomography and receiver functions.

In the case of receiver function studies, methods differ widely including the number of events stacked to increase signal-to-noise ratio and whether and how corrections are made for three-dimensional wave-speed variations.

At the interpretation stage, difficulty arises because seismic velocity is affected by melt content (which may be closely tied to volatile content), composition and temperature. Often these effects cannot be separated. It is commonly assumed that low wave-speeds correspond to high temperatures, perhaps simply because interpretation may be difficult unless some assumption is made. However, the slow = hot assumption has been shown to be spectacularly wrong in some of the few cases where the effects can be separated, *e.g.*, in the case of the Large Low Shear Velocity Provinces (“superplumes”) in the lower mantle, which have low wave speeds but high density and normal temperatures.

This general difficulty, where we have tools available but their strengths lie elsewhere than answering questions about the existence of deep-mantle plumes, also characterizes geochemistry. Seismology can reveal the spatial seismic structure of the mantle (up to a certain resolution), but is weak to constrain composition. In contrast, petrology and geochemistry can give insights into mantle composition, but have severely limited spatial control on magma sources. For these reasons, the two are often interpreted jointly. Nevertheless, the limitations of both methods are often underestimated. The issue of geochemistry will not be covered in detail in this presentation, however.

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MAPPING CRUSTAL THICKNESS, OCT STRUCTURE AND CRUSTAL TYPE USING SATELLITE GRAVITY

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Gravity anomaly inversion of satellite derived free-air gravity incorporating a lithosphere thermal gravity anomaly correction data now provides a useful and reliable methodology for mapping global crustal thickness in the marine domain (Chappell & Kusznir, GJI, 2008). The resulting maps of crustal thickness and continental lithosphere thinning factor may be used to determine continent-ocean boundary location, and the distribution of oceanic lithosphere, micro-continents and oceanic plateaux (e.g. Alvey et al., EPSL 2008). Crustal cross-sections using Moho depth from gravity inversion allow continent-ocean transition structure and magmatic type (magma poor, "normal" or magma rich) to be determined. Using crustal thickness and continental lithosphere thinning factor maps with superimposed shaded-relief free-air gravity anomaly, we can improve the determination of pre-breakup rifted margin conjugacy and sea-floor spreading trajectory during ocean basin formation.

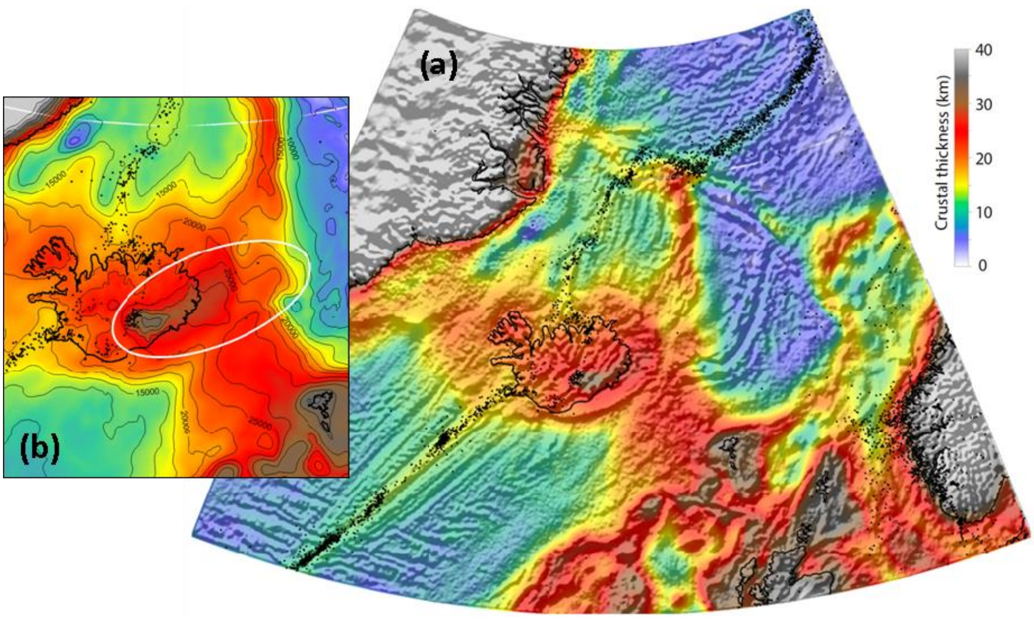

Figure 1 (a) Present day crustal thickness from gravity inversion for the NE Atlantic Ocean. (b) Higher resolution map of crustal thickness for E Iceland.

Application of these gravity inversion methods to the North Atlantic reveals a diversity of continent-ocean transition structures and compositions ranging from magma poor to magma rich. Crustal thickness mapping shows micro-continents, continental slivers, failed breakup basins and ridge jumps consistent with a complex evolution of rifting and ocean basin development. While the Moere and Voering margins are magma rich, the Aegir Ridge shows magma poor seafloor spreading in the Oligocene. The Jan Mayen micro-continent shows complex ocean-continent structure with magma-rich to the E and magma poor to the W.

Crustal thickness mapping shows large crustal thicknesses (>30 km) under SE Iceland (Torsvik et al, PNAS, 2015) extending offshore to the NE and consistent with SE Iceland being underlain by continental crust associated with a southern continuation of the Jan Mayen micro-continent. This interpretation is supported by geochemical evidence.

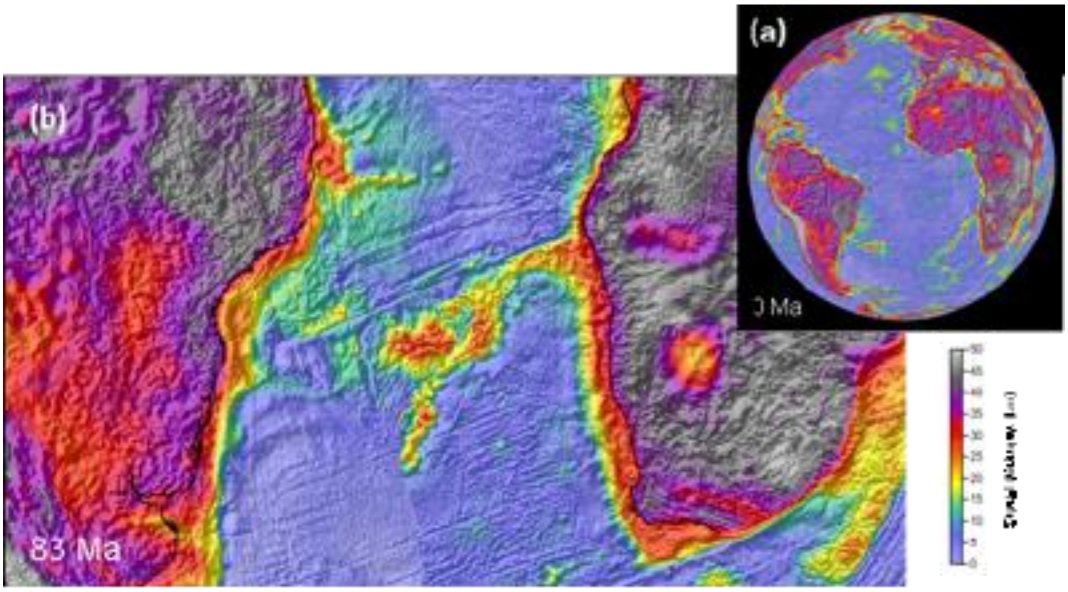

Figure 2 (a) Present day crustal thickness from gravity inversion for the Atlantic Ocean. (b) Crustal thickness restored to 83 Ma using GPlates 1.5.

Plate restoration to 83 Ma of crustal thickness derived from gravity inversion for the S Atlantic shows the Rio Grande Rise and Walvis Rise forming a single feature which is analogous to Iceland. Some continental component has been proposed for the Rio Grande Rise. Similar features with anomalously thick crust within the ocean domain with continental affinity are also observed with the Indian Ocean (Torsvik et al., Nature Geoscience, 2014) and appear to be attractors for ocean ridge jumps. Some of many questions are whether these regions clearly within the oceanic domain are underlain by lithosphere with some continental compositional component and whether the ridge jumps are attracted by rheological weaknesses controlled by compositional or thermal anomalies.

REYKJANES RIDGE EVOLUTION IS BETTER EXPLAINED BY PLATE BOUNDARY PROCESSES THAN MANTLE PLUME FLOW

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Because of its position intersecting the Iceland hot spot crustal features of the Reykjanes Ridge and flanks have been interpreted as resulting from mantle plume flow and associated thermal effects. These features include regional changes in crustal thickness associated with segmented vs. non-segmented stages of spreading (White, 1997) and ridge flank V-shaped ridges and troughs (Ito, 2001; Vogt, 1971). Here we show that plate boundary processes can explain these features and that mantle plume flow is not only not required, but that aspects of the mantle plume flow models present contradictions in accounting for thick crust at plumes that do not intersect spreading centers.

The Reykjanes Ridge flanks exhibit three distinct phases of seafloor spreading (Smallwood and White, 2002). In the first phase following continental breakup around anomaly 24 (~55 Myr), early seafloor spreading occurred on a long linear ridge accreting unsegmented crust. At about anomaly 17 (~37 Myr) a change in opening direction led to an abrupt fragmentation of the axis into a stair-step configuration forming the segmented crust of the second phase. In the third phase, the stair-step configuration diachronously evolved back to its linear configuration from north to south accreting unsegmented crust once again. The unsegmented seafloor spreading phases are associated with crustal thicknesses 2-3 km greater than the segmented phases and have been interpreted as reflecting underlying mantle plume temperature changes associated with regional plume advances and withdrawals (White, 1997). The changes in crustal thickness, however, can be more directly explained by the changes in plate boundary segmentation itself. This is because segmentation directly affects mantle advection with plate-driven mantle upwelling significantly decreasing toward offset ridge segment ends (Phipps Morgan and Forsyth, 1988). Numerical calculations using nominal values for the Reykjanes Ridge segmented phase of spreading predict about a 30% decrease in mantle vertical advection compared to a single long linear ridge. Since mantle melting rate is proportional to the total volumetric rate of vertical advection this translates directly to a 30% decrease in crustal thickness during the segmented spreading phases without any change in mantle temperature.

V-shaped crustal ridges and troughs on Reykjanes Ridge flanks have been interpreted as resulting from thermal pulses embedded within radially expanding mantle plume flow (Ito, 2001; Vogt, 1971). In this "pulsing plume" model as radially expanding annular thermal perturbations intersect the Reykjanes Ridge they result in crustal thickness changes forming the V-shaped patterns. However, another process that produces crustal thickness variations on slow spreading mid-ocean ridges is buoyant or "active" mantle upwelling (Bonatti et al., 2003; Forsyth, 1992). In active mantle upwelling, buoyancy forces associated with mantle depletion, melt retention and thermal advection may lead to mantle rising faster than predicted by passive plate-driven mantle advection (Scott and Stevenson, 1989). Episodes of buoyant mantle upwelling have been proposed to lead to crustal thickness variations on slow spreading ridge flanks of about 2 km (Bonatti et al., 2003; Pariso et al., 1995), which is the same magnitude as the Reykjanes Ridge V-shaped ridges and troughs. We propose that on a long linear ridge where the underlying mantle has systematic gradients in composition (water content; Nichols et al., 2002) buoyant upwelling instabilities can propagate and lead to the formation of the V-shaped crustal ridges.

The pulsing plume model directly relates formation of the V-shaped ridges to mantle plume flow and thus the geometry of the ridges indicates flow velocity, which is at least ten times the spreading rate (Vogt, 1971). Mantle upwelling velocities in the narrow plume stem must therefore be several times faster due to radial spreading (Ito, 2001). These geometric properties of the pulsing plume model imply extremely rapid mantle upwelling in the plume stem and huge resulting crustal thicknesses (hundreds of km) if mantle material is allowed to upwell above the solidus (Ito, 2001). Therefore pulsing plume models necessitate a highly viscous “dehydration layer” to laterally deflect the plume before it rises above the solidus (Ito, 2001). In this model, it is only the component of plate spreading that allows a small proportion of mantle plume material to rise above the solidus and to melt by decompression. Due to its higher than normal temperature this results in the excess melting responsible for Iceland and the V-shaped ridges. Yet, the dehydration layer envisioned in this model should be a property of all residual oceanic lithosphere where melt has been extracted (Hirth and Kohlstedt, 1996; Phipps Morgan, 1997). Such a model would therefore preclude melting in the absence of a spreading center over the plume, and contradicts the existence of intra-plate plumes such as proposed at Hawaii. These contradictions in the pulsing plume model are resolved if the V-shaped ridges reflect along-axis propagation of buoyant upwelling instabilities as these instabilities do not require extremely rapid mantle flow as in plume models as it is only the locus of buoyant upwelling that propagates along axis not mantle material.

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THE PLACE OF BASALTS FROM ICELAND WITHIN THE NORTH ATLANTIC

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Although Icelandic geochemical effects have been traced along the Reykjanes and Kolbeinsey Ridges (RR and KR), these only faintly hint at the broad and extraordinary diversity of lavas on Iceland from about the edge of the Icelandic shelves inland. Iceland differs from those adjacent ridges in three principal ways: 1) it has over 30 active or recently active central volcanoes; 2) these are generally centers for eruption of silicic lavas and tuffs (icelandites, dacites and rhyolites), which comprise over 21% of exposures in eastern Iceland; and 3) an array of parental basaltic lavas exists ranging from depleted picritic to enriched ferropicritic compositions, the latter having high TiO_2 and total iron as FeO_T. The latter produce ferrobasalts with >5% TiO_2 by crystallization differentiation, thence enriched silicic lavas. Only depleted and weakly differentiated basalts, but no picrites, occur on RR and KR. Field studies by G.P.L Walker in the 1960’s established that central volcanoes with ferrobasalts and silicic differentiates form wherever subaerial eruption of tuffs or subglacial hyaloclastites create concentrated zones of low crustal density; these become loci of mafic magma collection and differentiation. However, the high proportion of silicic lavas in Iceland far exceeds proportions of similar rocks seen in normal ocean crust (0.5-1%), and cannot be produced even from ~4X thicker but depleted (N-MORB) crust if all such differentiates could be concentrated near the top. Most geochemical and isotopic effects usually attributed to mantle sources can be accomplished by modest (1-5%) mixing of silicic melts into basalts. The problem of Icelandic petrology compared with most of the Mid-Atlantic Ridge thus reduces to explaining the origin of basalts with high Ti and Fe, plus their intimate association with abundant silicic lavas. Similar associations occur in the Deccan, Kerguelen and Karoo flood basalt provinces, and to a lesser degree in the Galapagos Islands, but not in intraoceanic western Pacific plateaus (greater Ontong Java and Shatsky Rise). For all but the latter two, involvement of continental crust, and presumably therefore delaminated subcontinental lower crust and upper mantle, is documented and implicated in contributing to the geochemical signature of basalts. It was also an important factor in proto-Icelandic magmatism in Scotland and Greenland. A holistic approach to Icelandic petrogenesis combining lower crustal and upper mantle partial melting, magma mixing, assimilation, and differentiation, utilizing a diversity of potential gabbroic, eclogitic, pyroxenitic and ultramafic sources, should be explored before attributing anything at Iceland to the lower mantle.

OBSERVATIONS AND MEASUREMENTS FROM THE NORTH ATLANTIC MARGINS EXPLAINED WITHOUT NEOGENE TECTONIC UPLIFT

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A new view of the North Atlantic calls for reflection on some older, classical ideas that have long held sway. Observations and measurements related to long-term landscape evolution onshore and sediment accumulation and burial offshore are used as indicators of Neogene tectonic uplift of the passive margins surrounding the North Atlantic (Japsen and Chalmers, 2000; Lidmar-Bergström et al., 2000). However, here we argue that these observations can be explained more simply via alternative mechanisms.

Low-temperature thermochronology data contain thermal information only. Cooling histories derived from thermochronological measurements are a function of 1) models of sub-grain scale dynamics, and 2) models of the geothermal gradient as driven by rock exhumation via tectonics or erosion. The chain of inference relating such cooling histories to active surface uplift can include multiple solutions, and we outline two potential cases below in which, contrary to the conclusion of previous studies, Neogene uplift is in fact not necessary to explain the inferred thermal history.

1) Deep low-enthalpy geothermal wells in the eastern North Sea provide accurate present-day temperature information via continuous temperature logging many years after drilling as well as excellent fission track and vitrinite reflectance (VR) data (Green, 2002). This area contains an almost unbroken sequence of Palaeozoic and Mesozoic sediments (Sørensen, 1986; Vejbæk, 1989), signaling a relatively straight forward Palaeozoic and Mesozoic subsidence evolution. The age of the base Quaternary sub-crop increases from SW to NE, and a long-lasting hiatus since the Late Cretaceous, in principle, leaves plenty of time for Cenozoic subsidence followed by Neogene uplift (Japsen et al., 2007). We note, however, that this interpretation does not make use of present-day temperature data and converts VR values to maximum temperatures independently of prior to modelling of the AFT data. Joint interpretation of all data yields no tectonic movements after the Late Cretaceous compressions, which caused inversion of sedimentary basins in Europe (Nielsen et al., 2005, 2007). We present new model results to support discussions.

2) In the West Greenland magmatic province (Dam, 2002) the presence of marine Paleocene sediments at a present-day elevation of c. 1200 m above sea level in the Nuussuaq peninsula, as well as the occurrence of hyaloclastics at elevations of up to c. 1600 m above msl level point to large differential vertical movements. Using VR and AFT data from the well Gro-3 Japsen et al. (2005) separated the vertical movements into a Paleogene and a Neogene uplift component. But alternative and simpler explanations can also be invoked. Using the other wells in the area as well (Ataa-1, Gant-1, Umiivik-1), the occurrence of flood basalt magmatism and differential erosion and flexural isostasy provides a full explanation of the differential vertical movements. We present new model results to support discussions.

The long-standing Davisian viewpoint, as applied to the North Atlantic margins and elsewhere, reads high-elevation low-relief landscapes as products of erosion to sea level followed by uplift to their present elevation. Anderson (2002) however demonstrated how transport-limited weathering produces diffusive smoothing of mountain summits over 10^5-10^6 years timescales. Chemical and mechanical weathering of rock mass produce debris and regolith cover that is transported downslope in streams and via different creep processes. The Aarhus surface processes group has devised state-of-the-art computational experiments that build on Anderson's foundational work and provide a new physics-based understanding of *in situ* formation of low-relief at high elevations (Andersen et al. 2015; Egholm et al., 2015). Moreover, the computational model predictions are being tested via a comprehensive set of field studies in Scandinavia and Greenland that employ cosmogenic nuclides to quantify rates of sediment production and transport (Knudsen et al., 2015). The fundamental conclusion from these recent studies is that high-elevation, low-relief terrain fringing the North Atlantic margins are due to a combination of slow weathering processes and fast glacial erosion. Both of these processes have no direct relation to sea level, and the low-relief weathering landscapes can therefore not be used to infer vertical surface displacement, which is otherwise key to classical ideas of peneplanation and Neogene uplift (e.g. Japsen and Chalmers, 2000; Lidmar-Bergström et al., 2000).

Studies of the geological history of the North Atlantic realm suggest a number of mechanisms that are capable of producing vertical movements of the rock column, many of which have been active since the Palaeozoic as well as more recently: the Caledonide Orogeny and previous orogenies, continental break-up, magmatism, erosion and sedimentation, and dynamics of the convecting mantle all have the potential to produce vertical movements over vast areas. Yet, a new mechanism to produce rock column vertical movements during the Neogene is simply not required to explain observations.

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AN EVALUATION OF MESOZOIC RIFT-RELATED MAGMATISM ON THE MARGINS OF THE LABRADOR SEA: IMPLICATIONS FOR RIFTING AND PASSIVE MARGIN ASYMMETRY

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The Labrador Sea is a small (~900 km wide) ocean basin separating southwest Greenland from Labrador, Canada. It opened following a series of rifting events that began as early as the Late Triassic or Jurassic, culminating in a brief period of seafloor spreading commencing by polarity chron 27 (C27; Danian) and ending by C13 (Eocene-Oligocene boundary). Rift-related magmatism has been documented on both conjugate margins of the Labrador Sea. In southwest Greenland this magmatism formed a major coast-parallel dyke swarm as well as other smaller dykes and intrusions. Evidence for rift-related magmatism on the conjugate Labrador margin is limited to igneous lithologies found in deep offshore exploration wells, mostly belonging to the Alexis Formation, along with a postulated Early Cretaceous nephelinite dyke swarm (ca. 142 Ma) that crops out onshore, near Makkovik, Labrador. Our field observations of this Early Cretaceous nephelinite suite lead us to conclude that the early rift-related magmatism exposed around Makkovik is volumetrically and spatially limited compared to the contemporaneous magmatism on the conjugate southwest Greenland margin. This asymmetry in the spatial extent of the exposed onshore magmatism is consistent with other observations of asymmetry between the conjugate margins of the Labrador Sea, including the total sediment thickness in offshore basins, the crustal structure, and the bathymetric profile of the shelf width. We propose that the magmatic and structural asymmetry observed between these two conjugate margins is consistent with an early rifting phase dominated by simple shear rather than pure shear deformation. In such a setting Labrador would be the lower plate margin to the southwest Greenland upper plate.

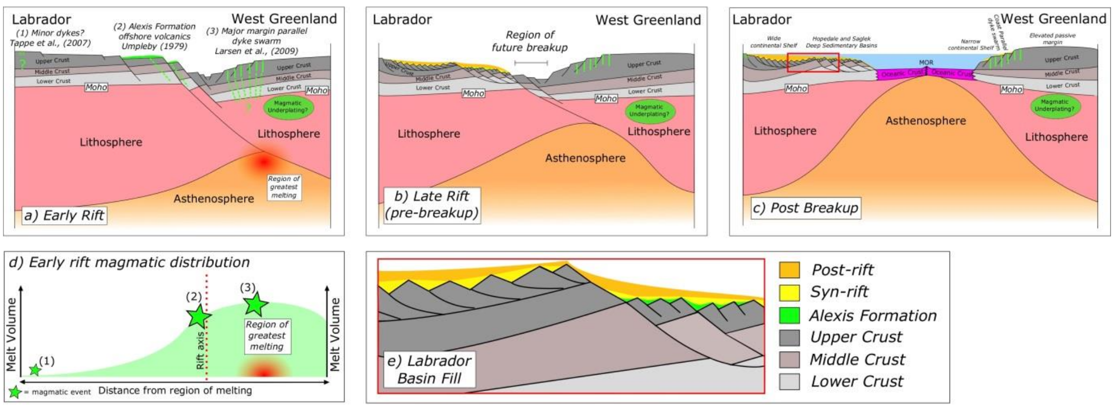

Figure 1. *A) Conceptual model of early continental rifting prior to the opening of the Labrador Sea under a simple shear rifting regime. B) Conceptual model of late continental rifting. C) Schematic depiction of the post-breakup (present) architecture of the conjugate passive margins of the Labrador Sea showing the preserved architecture from the early simple shear rifting modified from Lister et al.*

(1986), including the wide and narrow continental shelves for the Labrador and Greenland margins, respectively, the deep sedimentary basins offshore Labrador, and the minimal offshore sedimentary cover and elevated passive margin on the Greenland side of the rift. D) The theoretical distribution of melt volumes against proximity to the rift axis where the region of melting is offset from the rift axis due to the simple shear–type early rifting.

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ISOSTATIC AND DYNAMIC SUPPORT OF HIGH PASSIVE MARGIN TOPOGRAPHY IN SOUTHERN SCANDINAVIA

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Substantial controversy surrounds the origin, age, and recent evolution of high topography along the North Atlantic margins, with suggested age of formation ranging from early Palaeozoic Caledonian orogenesis[1] to Neogene uplift of a Mesozoic peneplain[2-4] (Fig. 1)**.** We focus on the well-documented elevated passive margin in southern Scandinavia, and quantify the relative contributions of crustal isostasy and dynamic topography in controlling topography. Consistent with previous work[5-9], we find that most topography is compensated by the crustal structure (Fig. 2a). Specifically, more than 1000 m high topography has existed in the region since the formation of the current crustal structure, suggesting a topographic age related to ~400 Myr old orogenesis. In contrast to the main part of Scandinavia, the western-most part of southern Norway shows negative predictions of isostatically compensated topography (Fig. 2b-c), which indicates that the current topography in this area is poorly explained by the crustal structure. We suggest that that dynamic support from the mantle may explain this deviation as dynamic uplift rejuvenated existing topography and adjacent regions within the last ~10 Myr (Fig. 2d). This idea is in agreement with a number of seismic tomography studies that predict low velocities and the presence of hot asthenosphere below the lithosphere in southern Norway[10-14]. Recent dynamic uplift can, combined with a general sea level fall[15], explain observations that have traditionally been interpreted in favour of a peneplain uplift model[16-18]. We conclude that the high topography along the Scandinavian margin cannot represent remnants of a peneplain uplifted within the last ~20 Myr. Topography must have been high since the formation of the current crustal structure, most likely dating back to the Caledonian orogeny at ~400 Ma.

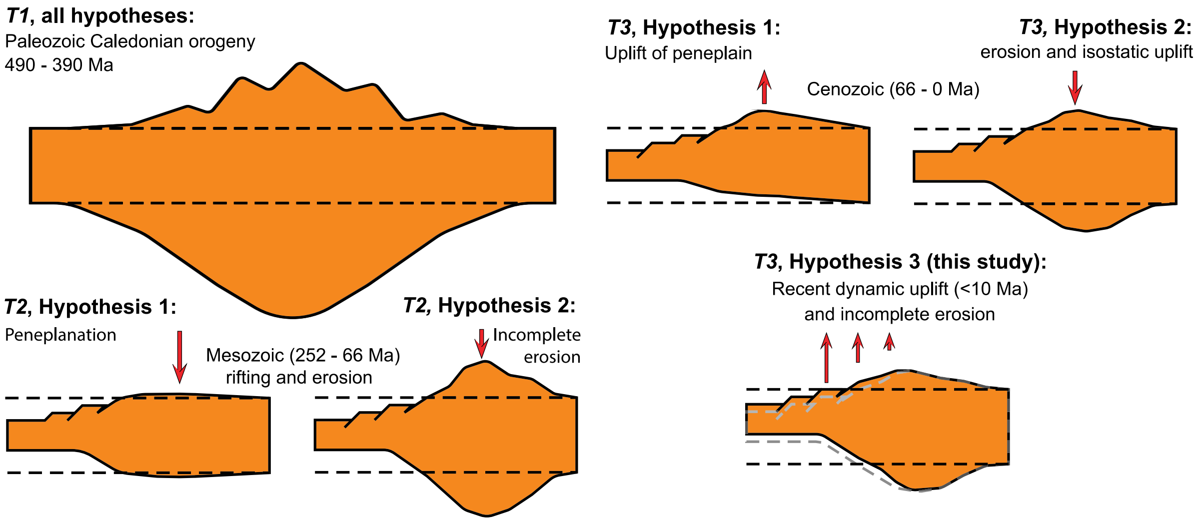

Figure 1: ***Schematic models for post-Caledonian geodynamic evolution in western Scandinavia.*** *Existing hypotheses (Hypothesis 1 and Hypothesis 2) for the geodynamic evolution of western Scandinavia since the Caledonian orogeny illustrated by crust and topography structure at three snapshots in time (T1, T2, T3), and the hypothesis proposed in this study (Hypothesis 3). Dashed black lines represent a reference crustal thickness with zero compensated topography. Arrows indicate changes in surface elevation. Note figure is not to scale.*

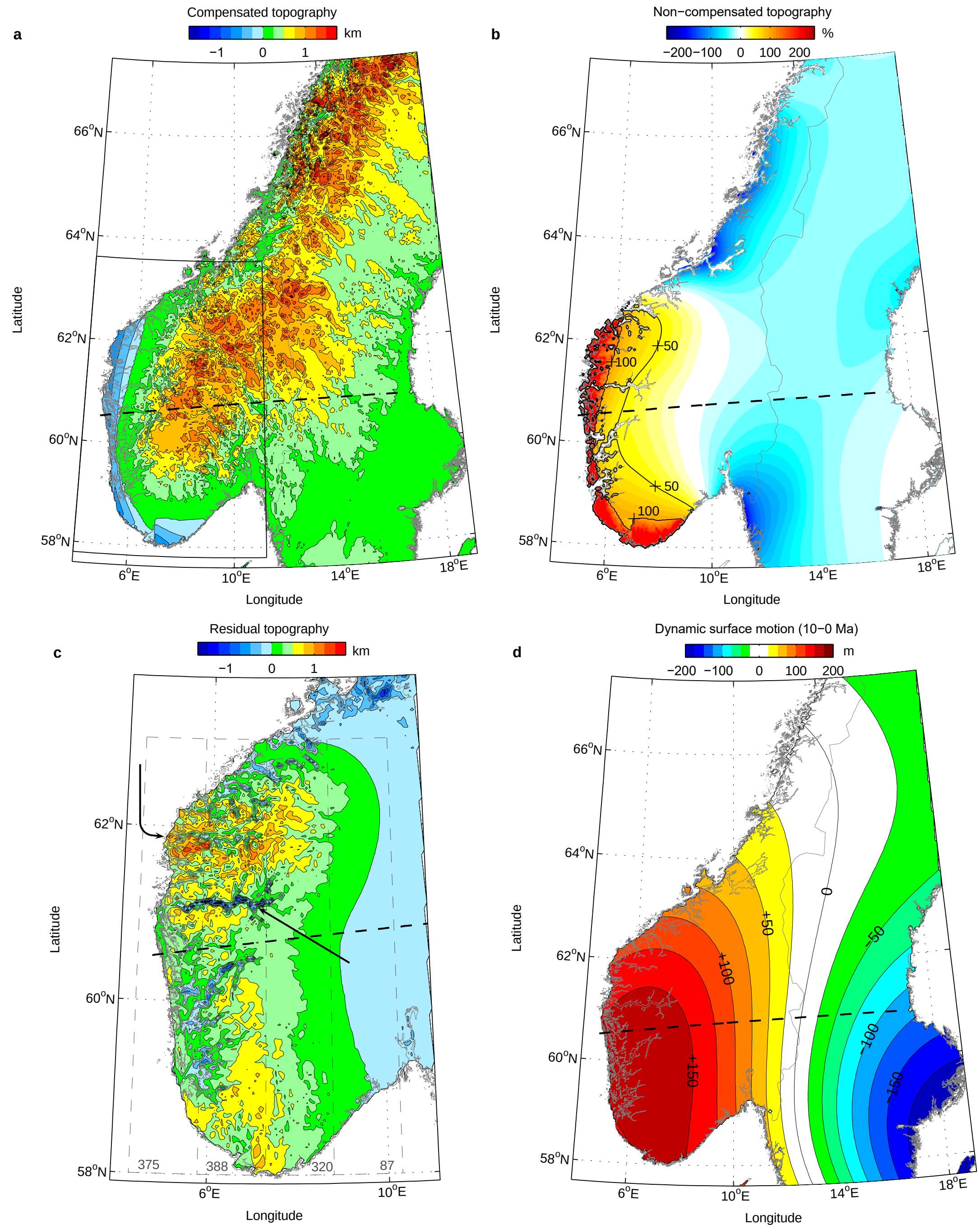

Figure 2: Regional isostatic, residual, and dynamic topography. a, Topography compensated by the crust. Regions with negative predicted topography have been corrected for water load. b, Degree of topography not compensated by the crust. Contour lines represent 50% and 100% non-compensation. c, Residual topography (real topography minus isostatically compensated topography from a). Positive values represent regions where the present topography is higher than expected from the crustal structure, and negative values regions where the topography is lower than expected. Averaged values [m] are given for transects outlined by dashed gray lines. d, Dynamic topography change since 10 Ma (TX2007V1; Ref. 19, 20). Positive values indicate rock uplift whereas negative values correspond to subsidence.

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MANTLE PLUMES DRIVEN BY PLATE TECTONIC PROCESSES

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The North Atlantic Large Ignous Province (NALIP) and the suture of the Caledonian Orogeny are spatially correlated. This correlation has led several workers to suggest a causal relation between the Paleozoic mountain-building events and the excess magmatism associated with Paleogene rifting and breakup. Furthermore, it has been argued that such a relation would imply that LIP-formation in the North Atlantic was driven by plate tectonic processes rather than the buoyant rise of plumes from the mantle.

Here we present a quantitative model that demonstrates that LIP-formation can be a direct consequence of orogenic crustal thickening followed by rifting and breakup. We employ a high-resolution, two-dimensional thermomechanical rift model coupled with a petrological model (Perple_x). Our model assumes a pre-rift/breakup thermal steady state with a region of thickened crust, representing the Caledonides. The P/T-conditions at the base of the crust produce mineral assemblages that are negatively buoyant relative to the adjacent mantle. As shown in earlier works (e.g. Jull & Kelemen; 2001), this can potentially lead to convective instability and delamination of the lower crust. The modest temperatures and stresses and the high strength of the mantle lithosphere prevent such instability from developing initially, and it is therefore possible for thickened crust to remain stable during protracted periods of tectonic quiescence. Upon extension and rifting, stresses increase and the lithosphere is mechanically weakened sufficiently for the dense lower crust to drive a rapid event of delamination. This produces a phase of rapid (~0.1 Myr) magmatism. Subsequently, the delaminated material continues to sink and crosses the transition zone. If the mantle below the transition zone has higher potential temperature than above (due to convective isolation), the dense crust sinking into the lower mantle induces a mass-conserving return-flow of hot and chemically anomalous lower mantle material that essentially rises to the lithosphere as plumes. Such plumes produce a second phase of increased and protracted (10s of Myr) melt productivity during breakup and the following stage of seafloor spreading. Our model may reconcile some of the apparently contrasting explanations for the formation of the NALIP: Magmatism was caused by plumes that were again caused by plate tectonics.

PASSIVE MARGIN FORMATION AND OROGENIC INHERITANCE

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Rifts often develop along suture zones between previously collided continents, as part of the Wilson cycle. The present-day margins of the Central and North Atlantic are a manifestation of at least two complete Wilson cycles: The assembly and break-up of the supercontinents of Rodinia and Pangaea. Most recently, the North Atlantic formed where Pangaea broke apart along Caledonian and Variscan sutures (Thomas, 2006). Remnant structures and lineaments of earlier mountain-building events are generally strike-parallel and mimic the present-day North Atlantic margins, implying that rift-localization and the plate tectonic evolution is controlled by such.ancestral structures of the older orogens (Williams, 1995; Buiter and Torsvik, 2014; Schiffer et al., 2015b).

Dipping upper mantle structures in East Greenland (Schiffer et al., 2014; Schiffer et al., 2015a) and Scotland (Snyder and Flack, 1990; Warner et al., 1996; Morgan et al., 2000), imaged by seismological techniques, have been interpreted as fossil subduction zones and the seismic signature indicates the presence of eclogite and serpentinite. We speculate that this inherited orogenic material may impose a rheological control upon post-orogenic extension and we use thermo-mechanical modelling to explore such effects. Our model includes the following features: 1) Crustal thickness anomalies, 2) Eclogitised mafic crust emplaced in the mantle lithosphere, and 3) Hydrated mantle peridotite (serpentinite) formed in a pre-rift subduction setting (Fig. 1, *middle panel*).

Our models indicate that the inherited structures control the location and the structural and magmatic evolution of the rift. Rifting of thin initial crust allows for relatively large amounts of serpentinite to be preserved within the uppermost mantle (Fig. 1, *upper panel*). This facilitates rapid continental breakup and serpentinite exhumation. Magmatism does not occur before continental breakup. Rifts in thicker crust (Fig. 1, *lower panel*) preserve little or no serpentinite and thinning is more focused in the mantle lithosphere, rather than in the crust. Continental breakup is therefore preceded by magmatism.

This implies that pre-rift orogenic properties may determine whether magma-poor or magma-rich conjugate margins are formed. Our models show that inherited orogenic eclogite and serpentinite are deformed and partially emplaced either as dipping structures within the lithospheric mantle or at the base of the thinned continental crust. The former is consistent with dipping sub-Moho reflectors often observed in passive margins. The latter provides an alternative interpretation of 'lower crustal bodies' which are often regarded as igneous bodies. An additional implication of our models is that serpentinite, often observed seismically or exposed at the sea floor of passive margins, was formed prior to rifting in addition to possible syn-rift, fault-driven hydrothermal processes. Whether lower crustal and serpentinite bodies are produced previously or during rifting is of relevance for the estimation of thinning-factors of the pre-existing crust. The results are published in Petersen and Schiffer (2016).

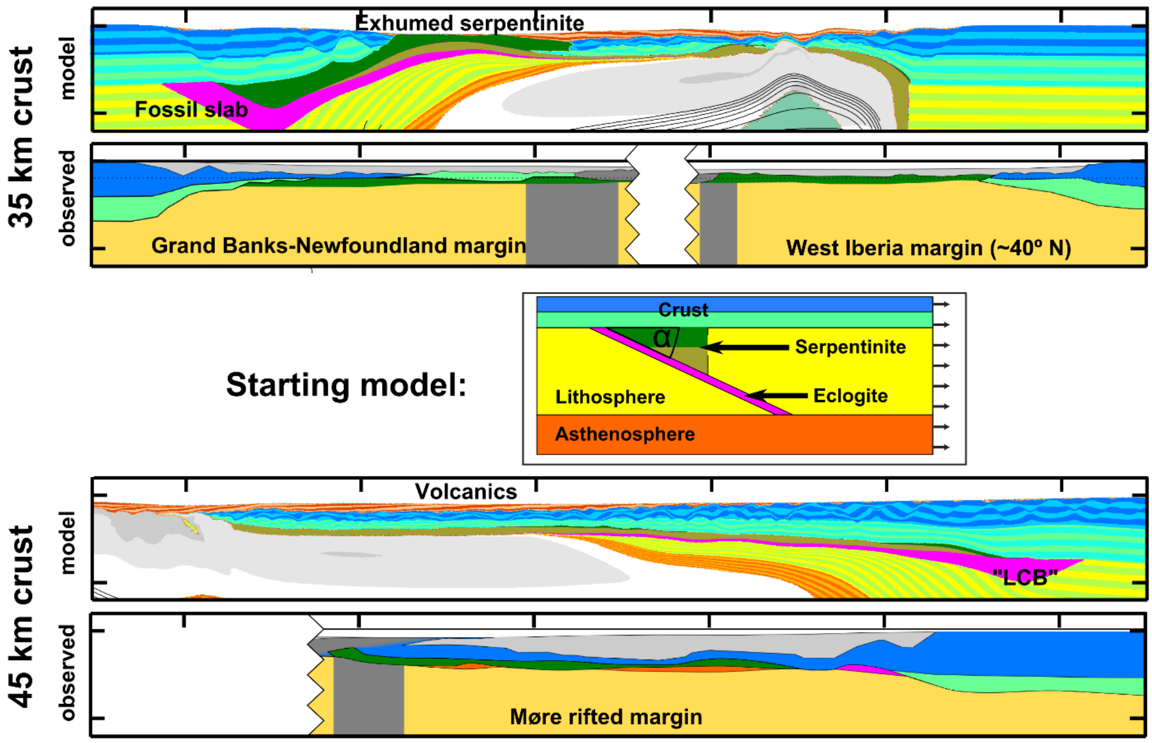

Figure 1: The starting model (middle panel) has three variable parameters, which are tested during the numerical simulations (crustal thickness, dip angle of the fossil eclogite slab and the rate of extension). Two results from the numerical simulations and possible observed analogues for a 35 km thick crust (upper panel) and 45 km thick crust (lower panel), could act as end member models for a magma-poor or magma-rich passive margin type, respectively.

The presented models are able to account for many observed features of the passive continental margins in the North Atlantic. However, it remains to be discussed and tested whether such a model is able to reproduce the large observed complexity and combination of single features, which are observed on both sides of the conjugate margin pairs. Our aim is to model the observed passive margin structure as exact as possible based on constructed key transects (e.g. Fig. 2). The key will be to test different combinations of pre-rift crustal and upper mantle properties, such as crustal thickness, lithospheric thickness, the number and orientation of fossil orogenic structures, as well the rate of extension and variations of all these properties, justified by geophysical and geological observations and the known geodynamic history of the region.

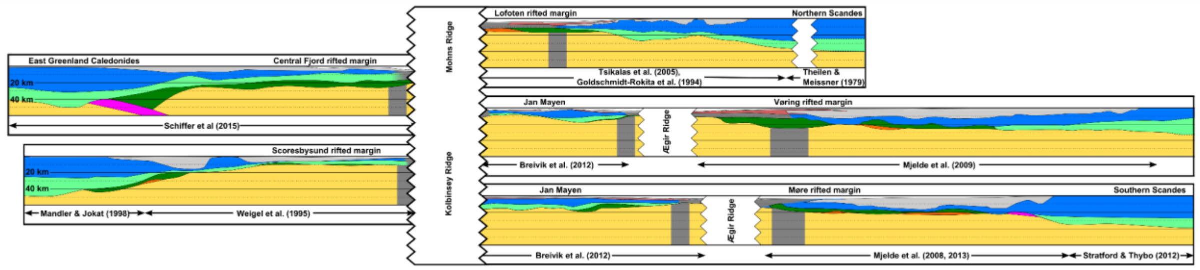

Figure 2: Examples of crustal profiles across passive continental margins of East Greenland (left) and Norway (right) including Jan Mayen. A fossil subduction zone is preserved in East Greenland, whereas at other places "lower crustal bodies" of different kind are distributed along the margins. Given the results from the numerical simulations, another pre-existing subduction zone may have existed on the Norwegian side, now heavily deformed after rifting and continental break-up.

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A SUB-CRUSTAL PIERCING POINT FOR NORTH ATLANTIC RECONSTRUCTIONS AND TECTONIC IMPLICATIONS

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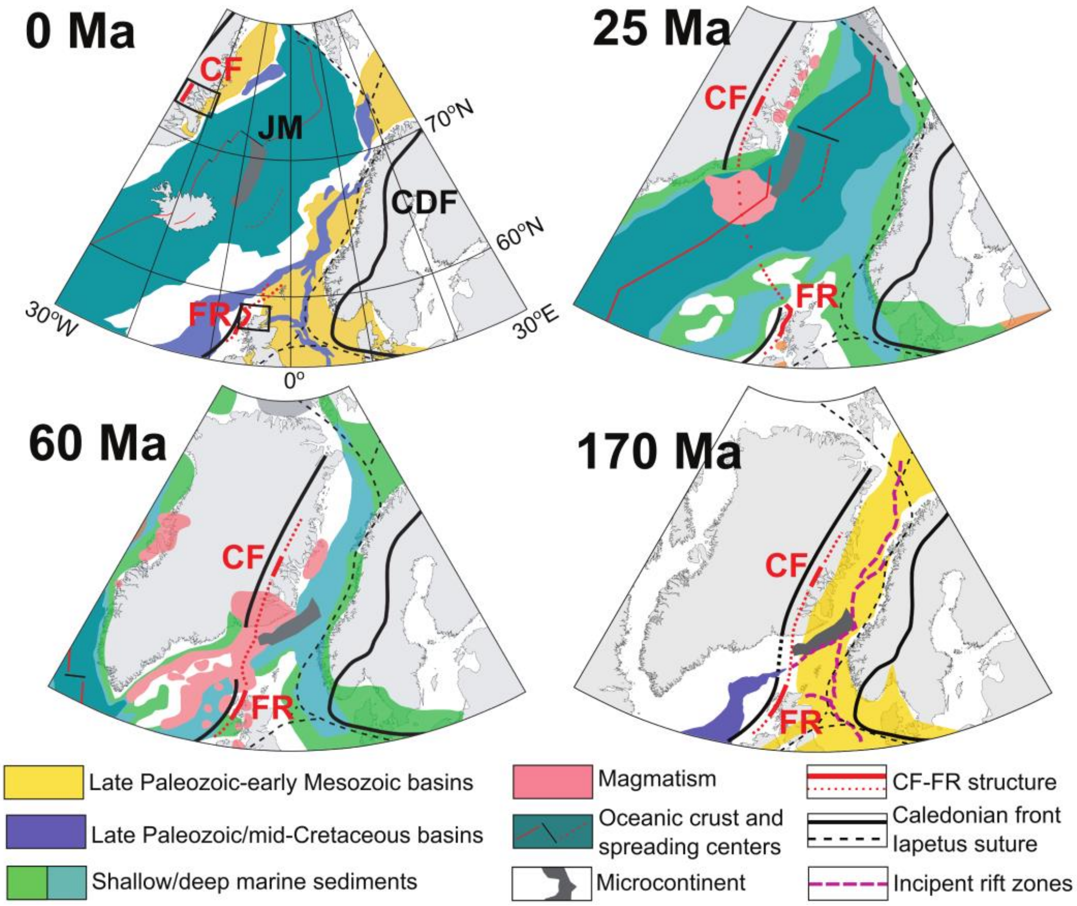
Plate tectonic reconstructions are usually constrained by the correlation of lineaments of surface geology and crustal structures. This procedure is, however, largely dependent on and complicated by assumptions on crustal structure and thinning and the identification of the continent-ocean transition. However, there are two geophysically and geometrically similar upper mantle structures in the North Atlantic on opposite margins and suggest that these represent remnants of the same Caledonian collision event. These are the Central Fjord mantle structure (Schiffer et al., 2014) and the Flannan Reflector (e.g. Chadwick and Pharaoh, 1998). The identification of the same lithosphere-scale structural lineament on conjugate margins provides a sub-crustal piercing point and hence a precise tie for North Atlantic plate reconstructions. The correlated structure coincides with the location of some major tectonic events of North Atlantic post-Caledonian evolution such as the formation of sedimentary basins, all on the hanging wall of the structure; the occurrence of the Iceland Melt Anomaly and the separation of the Jan Mayen microcontinent. This inherited orogenic structure within the pre-North Atlantic lithosphere likely played a significant role in the control of North Atlantic tectonic processes. The discovery and documentation of the structure representing the sub-crustal piercing point for North Atlantic reconstructions and ideas around it are published Schiffer et al. (2015).

The figure below shows the present day North Atlantic and palaeogeographic reconstructions at 25 Ma, 60 Ma and 170 Ma (after Skogseid et al., 2000; Torsvik et al., 2002). CF – Central Fjord; FR – Flannan reflector; JM – Jan Mayen; CDF – Caledonian Deformation Front. Black boxes at 0 Ma outline the data areas around the CF and the FR. Solid (observed) and stippled (interpolated) red lines are where the correlated sub-crustal structure intersects the Moho (Chadwick and Pharaoh, 1998; Schiffer et al., 2014). Late Palaeozoic-Early Mesozoic basins form east of this line, on the hangingwall of the sub-crustal structure (170 Ma); Cenozoic magmatism is largely over and west of it, on its footwall (60 Ma); Iceland forms above it (25 Ma).



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NE ATLANTIC BREAKUP: STRATIGRAPHIC CONSTRAINTS ON TECTONICS, VOLCANISM AND PALAEOGEOGRAPHIC EVOLUTION

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The purpose of this brief talk is to present observational stratigraphic data derived from the Late Mesozoic–Cenozoic rock record of the NE Atlantic region in general, and the continental margin offshore NW Britain (including the Faroe Islands) in particular, with a view to highlighting several paradigms that continue to influence putative models of NE Atlantic breakup, but which are largely without any robust foundation. Arguably, these 'traditional' paradigms have held back our understanding of NE Atlantic evolution for up to several decades.

- **Paradigm #1**

  A continuous rift system has existed between the NW British/Irish margin and SE Greenland margin since the Permian – the Ziegler legacy (e.g. Ziegler 1988; Knott et al. 1993; Roberts et al. 1999; Torsvik et al. 2002; Coward et al. 2003; Pharoah et al, 2010). This is typically displayed as a Faroe–Shetland–Rockall rift zone, as well as an inferred SE Greenland–Hatton rift.

  Comment

  This long-held view has resulted in palinspastic reconstructions that display a structural framework between NW Britain/Ireland and SE Greenland that is wholly unjustified by the observed geological record, and arguably presents a misleading view of intraplate deformation and palaeostress state prior to plate breakup. On most reconstructions, the depiction of these rift zones largely mimics the current bathymetric expression of the conjugate NW British/Irish/Faroese and SE Greenland continental margins; however, this morphological expression is largely a late-stage response to Cenozoic subsidence and 'passive margin' tectonics (Johnson et al. 2005; Naylor & Shannon 2005; Praeg et al. 2005; Stoker et al. 2005, 2010, 2012, 2013; Ellis & Stoker 2014). The observed stratigraphy across these conjugate margins provides no evidence for a substantive and continuous rift system in this part of the proto-NE Atlantic until the Late Cretaceous (Doré et al. 1999; Stoker 2016; Stoker et al. 2016).

- **Paradigm #2**

  The entire thickness (+6 km) of the Faroe Islands Basalt Group (FIBG) was erupted in latest Thanetian (Late Paleocene)–earliest Ypresian (Early Eocene) time (BP T-sequences T40–45), about 56–54 Ma (e.g. Jolley 1997, 2009; Jolley et al. 2002, 2012; Passey & Jolley 2009; Schofield & Jolley 2013). This correlates with the so-called syn-rift NAIP2 volcanic phase of Saunders et al. (1997), and thus, by definition, implies that the FIBG was wholly linked to the onset of seafloor spreading.

Comment

This view of the timing of the FIBG volcanism is based solely upon a biostratigraphic assemblage of terrestrial pollen/flora that the proponents of paradigm #2 link specifically to the PETM. However, an independent appraisal of the biostratigraphic data (Unpubl. BGS data) suggests that many of the biomarkers cited by the proponents of paradigm #2 are generic Paleocene pollen/spore flora that is deciduous and temperate in character and not characteristic of the PETM or the Paleocene–Eocene transition. Moreover, the biostratigraphic age assignment contrasts markedly with radiometric age dates, magnetostratigraphic data and the stratigraphic relationship between the sedimentary and volcanic formations in the Faroe–Shetland region, all of which support a more prolonged period of volcanism, i.e. Selandian (Mid-Paleocene)–earliest Ypresian (Early Eocene) (BP T-sequences T22–45), about 62–54 Ma (e.g. Waagstein et al. 2002; Mudge 2015; Unpubl. BGS data/Faroe-Shetland Consortium). This alternative viewpoint incorporates both the pre-rift NAIP1 and syn-rift NAIP2 volcanic phases of Saunders et al. (1997). Despite this conflict in interpretations, the biostratigraphic framework remains a strong influence in the scientific literature, and within industry. Although the disagreement between biostratigraphers and radiometric age daters (in particular) has the appearance of a minor, parochial academic dispute, its lack of resolution has held back stratigraphic investigations and understanding of the North Atlantic Igneous Province for almost 20 years, and has a particular bearing on paradigm #3 (below).

- **Paradigm #3**

The association of the FIBG (as defined in paradigm #2) with an unconformity at the Paleocene/Eocene boundary, along the southern flank of the Faroe-Shetland Basin, is widely regarded (in some academic circles) as unequivocal evidence for transient uplift linked to (Iceland) mantle plume processes (e.g. Smallwood & Gill 2002; Shaw Champion et al. 2008; Hartley et al. 2011).

Comment

The conflicting interpretations of the chronology of the FIBG (described above) strongly question its association with a single unconformity at the Paleocene/Eocene boundary. Whilst the existence of this unconformity is unequivocal, it is not a unique surface; it is just one of numerous Cretaceous–Cenozoic unconformities that have been identified within this part of the Faroe–Shetland region, which preserve a 90 My record of pre-, syn- and post-breakup differential uplift (e.g. Stoker et al. 2005, 2013; Stoker 2016). The significance of Cretaceous, Eocene, Oligo-Miocene and Early Pliocene differential movements in the same area as the postulated plume-related uplifts cannot be ignored. In the pre-breakup Cretaceous period, the tectonic development of the Faroe–Shetland region was characterised by a pattern of coeval extension and compression consistent with intra-plate strike-slip tectonic activity (Stoker 2016). The Cenozoic post-breakup vertical movements are on a scale (several hundreds of metres) comparable to the supposed Paleocene syn-breakup plume-driven uplifts calculated by Shaw Champion et al. (2008) and Hartley et al. (2011), yet any mantle thermal effect after this time would have shifted hundreds of kilometres to the NW, relative to the Faroe–Shetland region, as part of the seafloor-spreading process. It may be no coincidence that the southern flank of the Faroe-Shetland Basin is marked by the intersection of two long-lived fault structures: the NE-trending Rona Fault, and the NW-trending Judd Fault, the intersection, of which, produces a marked offset in the basement structure of the continental margin. The preservation of a Cretaceous–Cenozoic record of differential uplift in an area where inherited orogenic structures persist raises the possibility that the tectonic process(es) responsible for these movements were controlled by both regional-scale sources of stress and plate boundary forces, and thus do not necessarily require the involvement of a mantle plume (e.g. Holford et al. 2008, 2016).

On the basis of these differing viewpoints, it is clear that the evolution of the NE Atlantic region, particularly the conjugate margins of NW Britain/Ireland and SE Greenland, remains enigmatic. Despite the pioneering work of Ziegler (1988), in his own words he described his palinspastic reconstructions (paradigm #1) as, 'generalised.....they may have serious shortcomings....and should be regarded as working hypotheses'. Unfortunately, most subsequent authors that have considered the post-Caledonian evolution of the NE Atlantic region have largely continued to rely on these maps, thereby perpetuating all of their inherent limitations. Regarding paradigms #2 and #3, these are examples where the focus has been on one area and/or method and/or episode with little regard for geological context and no consideration of the wider consequences. In order to advance our understanding of the NE Atlantic region it requires to be analysed as a full rift system that is underpinned by observational evidence that can be verified within a multidisciplinary framework (e.g. Hopper et al. 2014).

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