

Regional distribution of volcanism within the North Atlantic Igneous Province

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Abstract: An overview of the distribution of volcanic facies units was compiled over the North Atlantic region. The new maps establish the pattern of volcanism associated with breakup and the initiation of seafloor spreading over the main part of the North Atlantic Igneous Province (NAIP). The maps include new analysis of the Faroe–Shetlands region that allows for a consistent volcanic facies map to be constructed over the entire eastern margin of the North Atlantic for the first time. A key result is that the various conjugate margin segments show a number of asymmetric patterns that are interpreted to result in part from pre-existing crustal and lithospheric structures. The compilation further shows that while the lateral extent of volcanism extends equally far to the south of the Iceland hot spot as it does to the north, the volume of material emplaced to the south is nearly double of that to the north. This suggests that a possible southward deflection of the Iceland mantle plume is a long-lived phenomenon originating during or shortly after impact of the plume.



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Pre-breakup volcanism from the North Atlantic Igneous Province (NAIP) extends over a pre-drift east–west distance of approximately 2000 km from Baffin Island, Canada to Lundy Island, Bristol Channel, UK (Saunders *et al.* 1997) while breakup volcanism extends over a NE–SW distance of nearly 3000 km from Lofoten/East Greenland Ridge in the north to Cape Farewell/Edoras Bank in the south (Fig. 1). Although detailed mapping of the NAIP has been published for smaller local areas (Planke & Alvestad 1999; Planke *et al.* 2000; Berndt *et al.* 2001) and simple regional maps have been published (Eldholm & Grue 1994; Saunders *et al.* 1997; Brooks 2011), the vast area covered by volcanism is a challenge for developing consistent

mapping of various volcanic features in a systematic way (Fig. 2).

Nevertheless, mapping of the volcanic facies and understanding their emplacement environment and magmatic processes play an important role in the understanding of rifting and tectonic processes associated with the generation of large igneous provinces. Indications for subaerial v. subaqueous extrusion and indicators for water depth provide information regarding palaeogeography during breakup and initial rifted margin evolution (e.g. Wright *et al.* 2011). In addition, detailed mapping of key units is the first step towards establishing the volume of material extruded and intruded, which when combined with accurate geochronology

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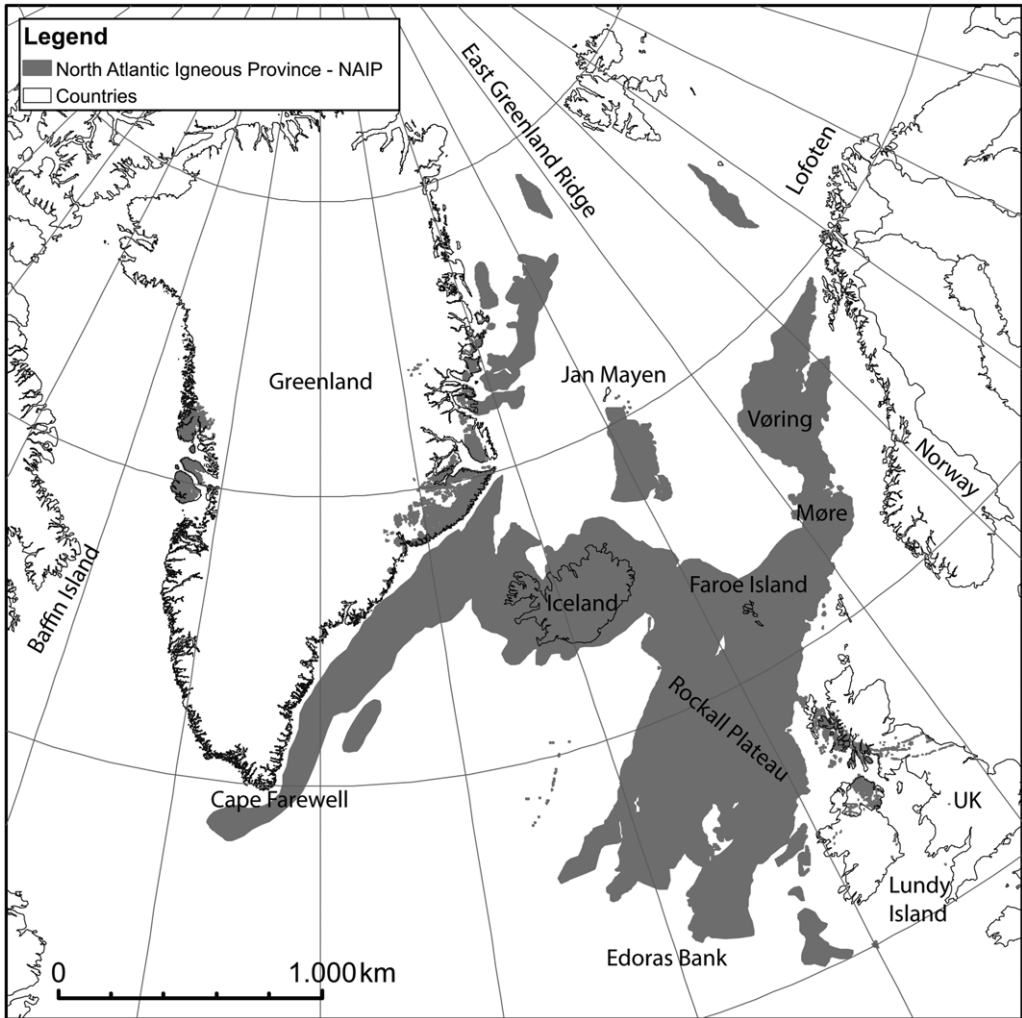


Fig. 1. Overview map of the North Atlantic showing the outline of the NAIP based on this work.

can constrain productivity rates. Together with geochemistry, the combined information places constraints on the mantle thermal and chemical structure associated with volcanic margin formation.

It is increasingly clear that NAIP volcanism is associated with a mantle thermal anomaly (e.g. Nielsen & Hopper 2002, 2004; Brown & Lesher 2014). Thus volcanism is typically interpreted in terms of mantle plume dynamics. The role of the overlying crust and lithosphere is generally disregarded. Several papers, however, have suggested that the lithosphere may play a key role in determining when and where volcanism occurs when a mantle plume impacts the base of the lithosphere (e.g. Ebinger & Sleep 1998; Nielsen *et al.* 2002; Sleep

2002). Regional mapping of volcanism and volcanic facies provides information on the patterns and distribution of volcanism that may reflect pre-existing lithospheric and crustal structure.

This contribution presents a regionally consistent interpretation of the main volcanic facies that can be mapped along the North Atlantic margins from approximately the Bight Fracture zone in the south up to the Fram Strait and SW Barents Sea margin to the north. Several areas that have not been considered in previous work are included: NE Greenland margin, the Jan Mayen microcontinent, the Greenland–Iceland–Faroe Ridge Complex (GIFRC), and Faroe–Shetlands platform. The former three regions are discussed in Blischke *et al.* (2016), Geissler *et al.* (2016) and Hjartason

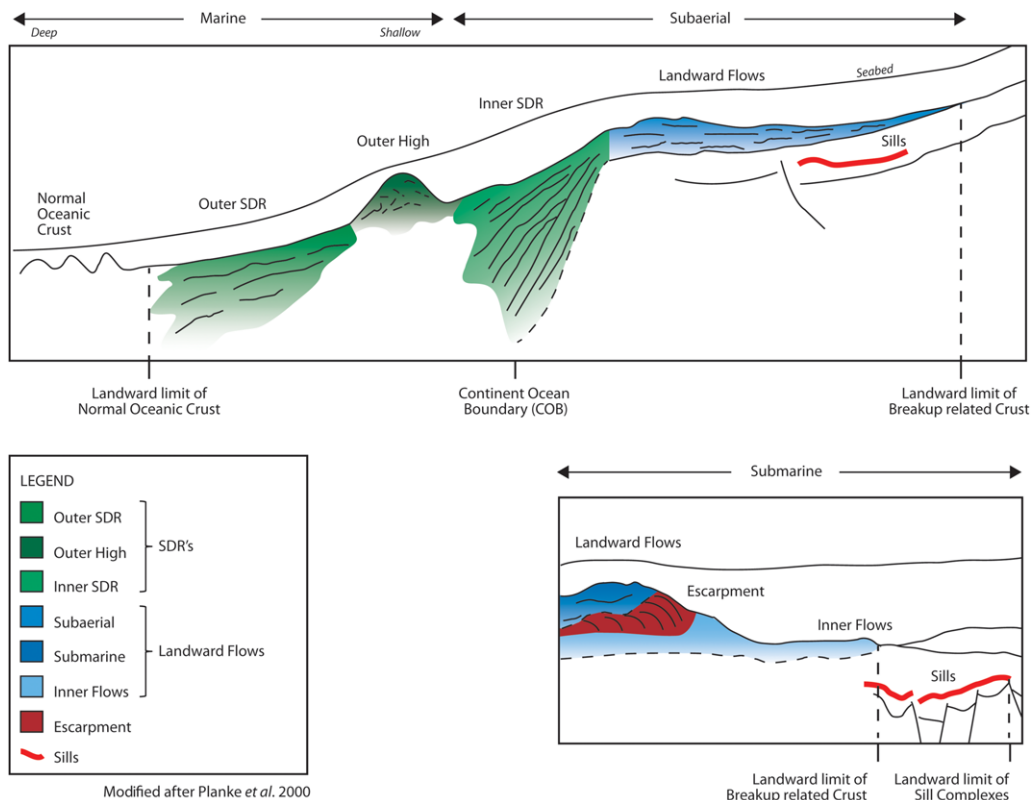


Fig. 2. Schematic model showing the different facies. The typical transition from subaerial to marine environment is illustrated in the upper panel. The lower panel shows the typical transition from landward flow to inner flow divided by an escarpment.

et al. (2017) and therefore are not discussed here. The new mapping around the Faroes is discussed in more detail and provides a regionally consistent view of the entire eastern margin from the Hatton margin up to the Lofoten margin.

The new mapping highlights a number of regional asymmetries in volcanism that characterize the NAIP. We suggest that these asymmetries are the result of the pre-existing lithospheric and crustal architecture of the region, which had an important influence on determining the lateral flow of hot material and played a first-order role in determining the pattern of surface volcanism observed in the region.

Seismic facies units

The mapping is based mainly on seismic volcanostratigraphy; however, gravity and magnetic anomalies provide supplementary information about possible distribution of volcanic rocks in areas with sparse seismic data.

Seismic volcanostratigraphy is the study of the nature and geological history of volcanic rocks and their emplacement environment from seismic data. Volcanic seismic facies units are based on their shape, reflections patterns and boundary reflections (Planke *et al.* 2000; á Horni *et al.* 2014). In areas well covered by seismic data and wells, it is possible to construct detailed maps with many facies. Here, however, we focus on a simplified scheme to provide a basic regional overview of the distribution of volcanism. The scheme is based on the work of Planke *et al.* (2000) (Fig. 2). Five primary units are considered and described briefly below: landward flows, escarpments, seaward dipping reflector (SDR) sequences, sills, and intrusions and igneous centres. Along the eastern margins, a subdivision of some units is included as outlined below.

Landward flows

Landward flows are composed of both subaerial and submarine lavas. Subaerial lavas are erupted onshore and are largely the products of fissure or

vent eruptions. They are often seen to flow down-slope for several hundreds of kilometres. Submarine lavas are lavas of the same source but emplaced on, or close to, the coast in a submarine environment (Self *et al.* 1997).

Subaerial landward flows are commonly identified on seismic sections as a strong, fairly smooth, top reflector (Fig. 3a). The external shape

is sheet-like, whereas internal reflections are disrupted or hummocky and sub-parallel. The submarine landward flows typically have a rougher top reflector than the subaerial landward flows. The landward-flow unit frequently wedges out in the landward direction and terminates at a regional escarpment or merges with prograding reflections (Fig. 3b).

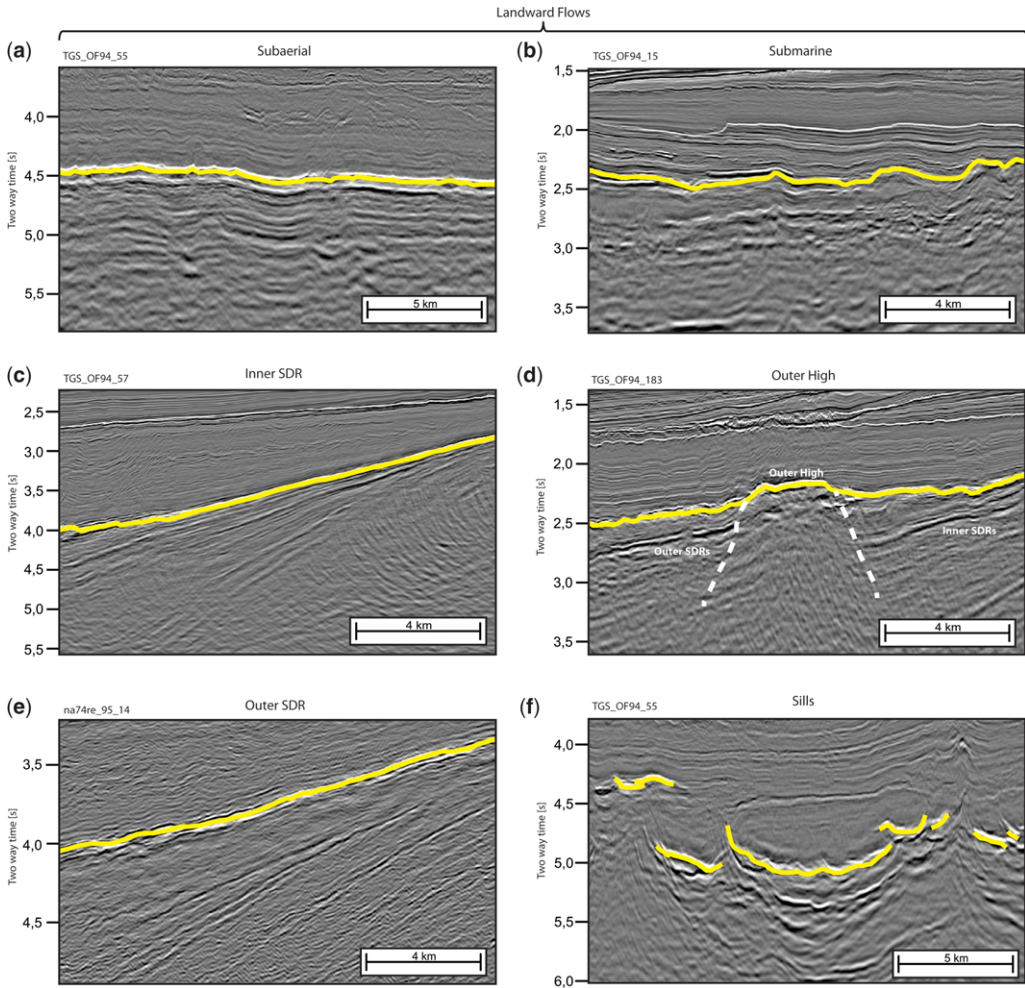


Fig. 3. Examples of seismic sections showing the different volcanic facies. The location of the profiles is not displayed for confidentiality reasons. (a) Subaerial landward flows typically are indicated by a strong, smooth and laterally continuous top reflection. (b) Submarine landward flows are also indicated by a strong top reflection, but it is rougher and laterally discontinuous. (c) Inner SDRs are observed as a wedge-shaped unit with strong internal, dipping reflections capped by a continuous, smooth top reflection. (d) An outer high is a mounded feature characterized by a fairly strong top reflector, chaotic internal reflection and is located near the seaward termination of the inner SDRs. (e) Outer SDRs are similar to inner SDRs, but are less well developed and are located seaward of an outer high (when present). The top reflection of outer SDRs trends is smooth, but is somewhat less continuous. (f) Sills emplaced in sedimentary sections are very distinct on seismic sections because of the significant acoustic impedance contrast between the intrusion and the sedimentary host rock. They frequently have a typical saucer shape.

Escarpmnts

An escarpment is a feature with steep relief. In volcanic systems, escarpments form typically from eruptive and emplacement processes. Changes in the deposition rate, the type of material erupted, or the available accumulation space are key factors that can control the development of escarpments (Smythe *et al.* 1983; Kjørboe 1999). In a recent review of the Vøring escarpment, Abdelmalak *et al.* (2015) show that major volcanic escarpments form in areas where lava flows encroach into basins. They are thus often seen in association with lava delta fronts and provide a key marker for the palaeo-shoreline (Moore *et al.* 1989).

Seaward dipping reflectors (SDRs)

One of the characteristic features of a volcanic rifted margin is presence of well-defined seaward dipping reflections below the top basalt reflection (Fig. 3c–e). SDRs have smooth to hummocky geometries and are interpreted to be subaerial lava flows erupted during an early stage of seafloor spreading. Laterally extensive continuous internal reflections are interpreted as large sheet flows, which are common in subaerial extrusions having a large magma supply that can continuously feed individual lava flows. In areas where they are particularly well developed, they have a concave-down and/or wedge appearance (e.g. Fig. 3e). SDRs form by sequential loading of older flows as rifting and seafloor spreading initiates. They are considered the most important indicator for anomalously large volcanism along rifted margins (e.g. Hinz 1981; Mutter *et al.* 1982).

Along many margins, two distinct sets of SDRs are often observed. The inner, landward set typically terminates at an outer high, followed by a second, outer set. The outer high is a mounded feature characterized by a fairly strong top reflection and chaotic internal reflections (Fig. 3d) (Planke *et al.* 2000). The outer high is interpreted to represent a transition from a subaerial to subaqueous environment. Flows erupted in shallow water can be more explosive and form hyaloclastites, preventing the formation of sheet flows that produce SDRs (e.g. Gregg & Fornari 1998). The outer SDRs are interpreted to indicate a transition from shallow to deep water, where the water pressure prevents degassing and large sheets flows can form, similar to the subaerial eruptions (Gregg & Fornari 1998; Planke *et al.* 2000; Hopper *et al.* 2003). The seismic characteristics of both SDR units are similar, however (Fig. 3c, e).

Sills and intrusions

Sills and intrusions are emplaced in existing host rock of any lithology. In this study, shallow

intrusions are typically observed as sills intruded into sedimentary sections (Fig. 3f) (Planke *et al.* 2000, 2005; Smallwood & Maresch 2002). These are very distinct on seismic sections because of the significant acoustic impedance contrast between the intrusion and the sedimentary host rock (Fig. 3f).

Magmatic intrusions emplaced into basaltic sections have not been identified on seismic data, but studies from exposures and wells from the Faroese area shows that they are present within the basaltic province (Varming 2009; Hansen *et al.* 2011). However, they are difficult to image on seismic sections because of the similarities in physical parameters between the host rock and the intrusions. Similarly, sills can be intruded into crystalline continental crust, although this is difficult to observe on seismic sections.

Igneous centres

The igneous centres have been divided into six subdivisions: offshore seamounts, igneous complexes, inactive calderas, active calderas, onshore inactive central volcanoes and active central volcanoes (á Horni *et al.* 2014).

Igneous centres represent a persistent volcanic eruptive or vent area that has built a complex combination of volcanic forms over time. These eruptions may be associated with faults or fissures. Igneous centres are identified on seismic sections as a mound or a bank feature with dipping sides and commonly erosional features on the top. In areas with no seismic data, it is possible to infer additional igneous centres from circular gravity anomalies (Passey & Hitchen 2011; á Horni *et al.* 2014).

Results

Figure 4 shows the regional map of main volcanic features and seismic facies. The map presents a regional overview of NAIP volcanism, including an assessment of several areas that have not been considered in detail before. In particular, new mapping of the Faroese platform is included that provides a consistent interpretation connecting the well-mapped Norwegian margin (Berndt *et al.* 2001; Planke *et al.* 2005) to the Rockall–Hatton margins and the UK and Irish margins (Elliott & Parson 2008). The map also shows several aspects of the regional distribution patterns that are discussed further below – including a comparison of the breakup-related volcanic volumes between different margin segments – and a brief discussion on the northern extent of the NAIP.

Faroe–Rockall–Hatton detailed mapping

Figure 5 illustrates the details of the main volcanic facies in the Faroese–Rockall–Hatton area and is

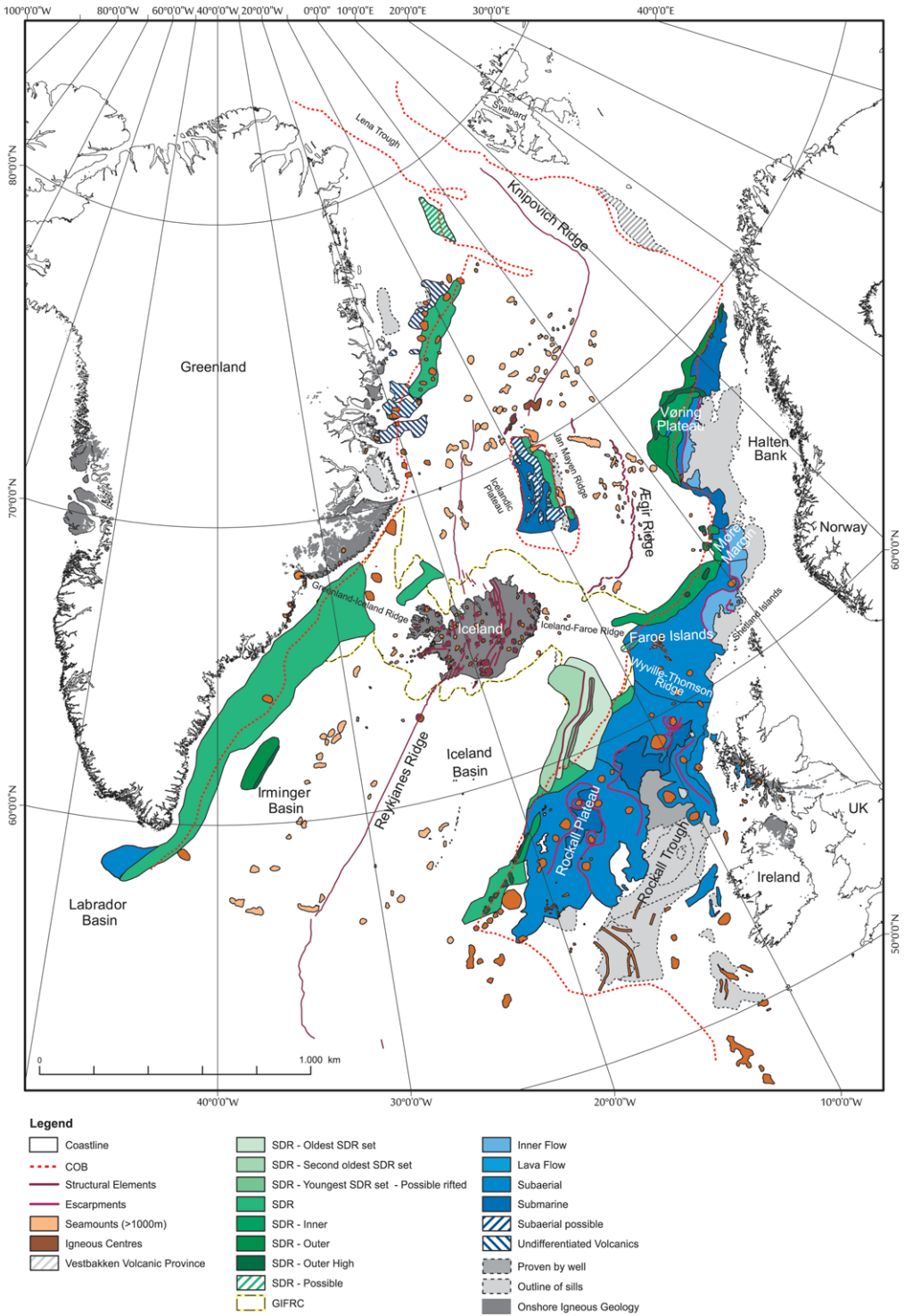


Fig. 4. Volcanic facies map of the NAIP. The COB (Funk *et al.* 2016) is shown as are the main volcanic escarpments.

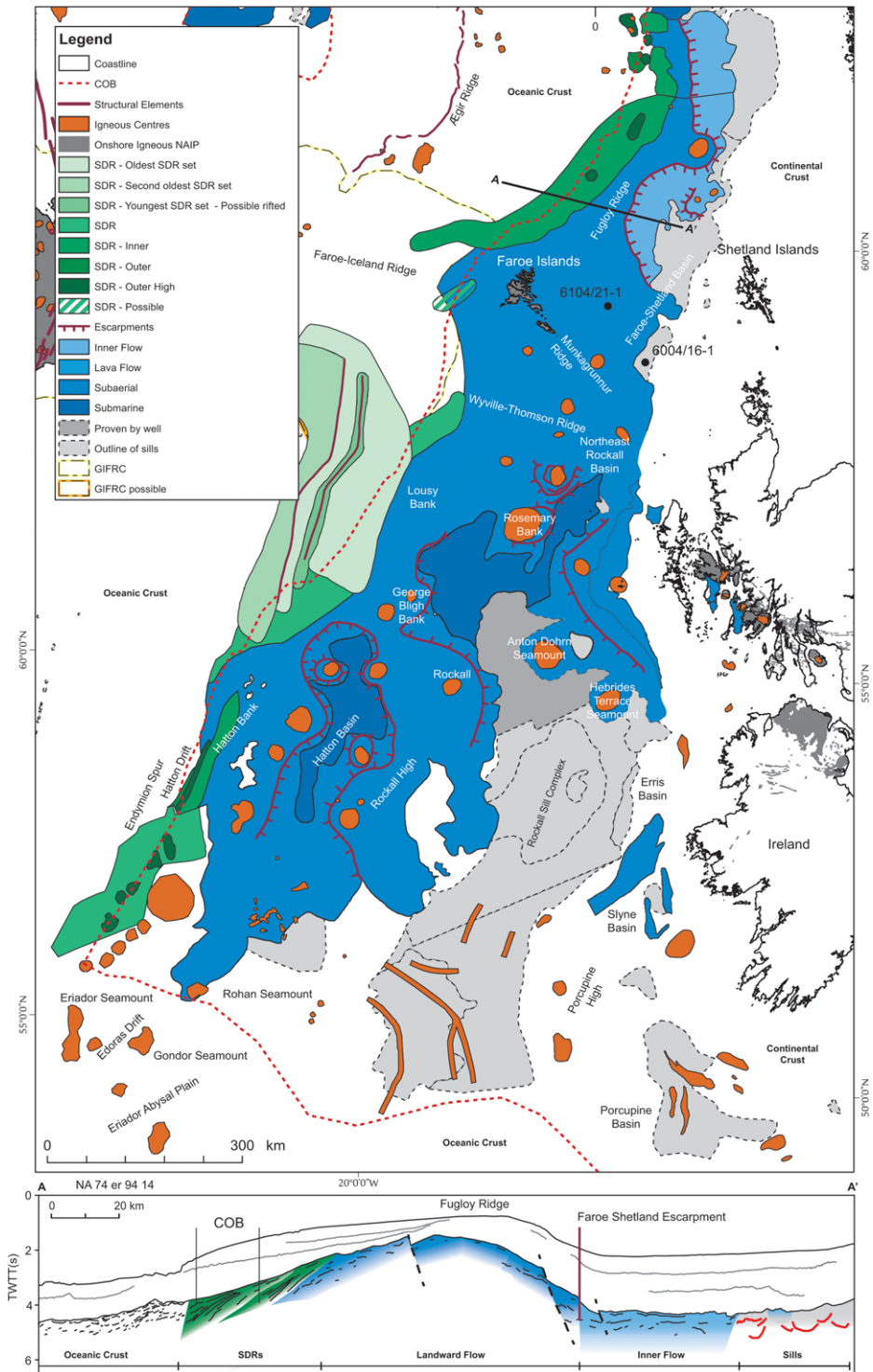


Fig. 5. Map of the volcanic facies in the Faroe-Rockall area and a cross section showing an example of the volcanic facies on a geo-seismic profile.

mainly based on a compilation of previously published maps (Á Horni *et al.* 2014) with updated seismic interpretation in several areas.

The area is dominated by basaltic volcanic rocks derived from many different sources, including numerous volcanic centres (Hitchen 2004; Hitchen & Johnson 2013). In terms of seismic data, the top surface of the lavas produces an easily mappable prominent reflection, which shows considerable variation in depth across the margin. The top reflection occurs at or near to the seabed on structural highs and shelves, but is at depths greater than 3 km below the seabed in the northern part of the Rockall Basin. Most of the lavas are interpreted to be subaerial (Hitchen & Johnson 2013). Thus in the Rockall Basin, significant post-early Eocene subsidence has occurred. However, it should be noted that lavas interpreted to be of submarine origin have also been described from parts of the basin (Á Horni *et al.* 2014).

The Faroe Island basalt group (FIBG) is mainly composed of landward flows, which are an expression of large-scale volcanic activity which covered an area of approximately 120 000 km² (Figs 4 & 5). The FIBG is estimated to have a total stratigraphic thickness of 6.6 km (Passey & Jolley 2009). In offshore commercial wells, wireline logs indicate that the subaerial lavas have a thickness of 795 m and the hyaloclastites a thickness of 778 m (Brugdan well – 6104/21-1) (Fig. 5). Well data further shows that the volcanic material thins towards the SE. In the Marjun well (6004/16-1), there are no subaerial lava flows, only tuff and some intrusions (Fig. 5). It has been suggested that approximately 2000 m of the FIBG have been eroded in the western part of the islands and around 200–300 m in the southern part of the islands (Waagstein 1988; Andersen *et al.* 2002; Jørgensen 2006).

SDR sequences are observed oceanward at the transition into the Norway Basin and Iceland Basin to the NW and to the SE, respectively (Fig. 4). SDRs, however, are not identified on seismic data to the west and NW where the Faroese platform connects to the Faroe–Iceland Ridge. SDRs typically form near the continent–ocean boundary (COB) and are associated with rapid subsidence as spreading progresses. The lack of well-imaged SDRs towards the Faroe–Iceland Ridge may indicate that along the ridge, little subsidence occurred as spreading was established. On the conjugate side, the main SDRs along SE Greenland become poorly defined towards the Greenland–Iceland Ridge. Both of these aseismic ridges are characterized by extremely thick igneous crust that is comparable to Iceland today (Smallwood *et al.* 1999; Holbrook *et al.* 2001). This suggests that well-developed SDRs form primarily along margin segments where the melt supply decreases as spreading

becomes established, resulting in a rapid reduction in crustal thickness.

The Faroe–Shetland Basin, which is to the east of the main landward flows, is highly intruded by sills and dykes (Figs 4 & 5). This intrusive complex is referred to as the Faroe–Shetland sill complex and its mapped extent covers an area of approximately 22 500 km² (Passey & Hitchen 2011). The sill complex, however, covers a much greater area as has been proven by well data showing that sills are found below the landward flows within the Faroese area (Á Horni *et al.* 2015). Sills are also found onshore on the Faroe Islands (Hansen *et al.* 2011).

Close to the eastward edge of the landward-flow lavas, a major continuous escarpment has been mapped crossing the border to both the Norwegian and UK areas and is referred to as the Faroe–Shetland escarpment (Figs 4 & 5) (Smythe 1983). A vertical relief of up to 1000 m is observed across the escarpment. Regional escarpments are only well mapped along the eastern margins of the North Atlantic. Along SE Greenland and around Jan Mayen, no escarpments have been mapped. Along NE Greenland, poorly developed escarpments are indicated, but show distinct differences compared to the conjugate margins – see Geissler *et al.* (2016) for a discussion. The escarpments are found between the inner flows and landward lava flows (Figs 2 & 5) and are interpreted to represent a period of volcanism associated with subsidence below sea level, generating prograding lava deltas into shallow water, thus marking a palaeoshoreline (e.g. Wright *et al.* 2011). The age of the escarpment is not known from direct dating. Stratigraphically, it is within the uppermost Beinisvörð Formation close to the break with the Malinstindur Formation of the FIBG (Smythe *et al.* 1983; Kjørboe 1999; Passey & Hitchen 2011). This break probably correlates to a volcanic hiatus observed throughout the NAIP from 58 to 56 Ma (Graham *et al.* 1998; Storey *et al.* 2007; Larsen *et al.* 2015). Thus, the escarpment may indicate a period of subsidence and submergence prior to the main flood volcanism and final breakup at 56 Ma.

Spatial distribution of volcanism

The regional volcanic facies maps present a new view of the spatial patterns of NAIP extrusive volcanism that provide insight into the development of volcanism prior to and during breakup. In this section, the spatial pattern of pre-breakup volcanism is briefly described; the volumes and production rates of breakup volcanism are then analysed. The results are compared to previous work regarding volume and productivity estimates along with early observations of asymmetric patterns.

The implications of these patterns of volcanism in terms of lithosphere plume interaction are then discussed.

Pre-breakup volcanism. Pre-breakup to breakup volcanism is best shown by the mapped landward lava flows and inner flows, which cover much of the Faroese, UK and Irish margins over a broad area. Landward flows are also observed to a lesser extent along the Greenland, Jan Mayen and Norwegian margins. It is difficult to estimate the volume of pre-breakup volcanism since generally the base of the basalt is not mappable and so thicknesses are poorly constrained.

In addition, it is difficult in these regions to separate pre-breakup volcanism from breakup volcanism. Where sampled along the Rockall–Hatton margins, dating of volcanic rocks from the landward and inner flows shows that these units include both pre-breakup and breakup volcanic rocks and that the pre-breakup phase is less extensive (Hitchen & Johnson 2013). It should be noted that onshore areas in West Greenland, East Greenland, the Faroe Islands, and parts of Scotland and Northern Ireland also include pre-breakup volcanic rocks. These areas are marked by dark grey in Figure 4. The maps show that pre-breakup volcanism affects only local areas except for the region around the Rockall Basin and the Faroes platform. Pre-breakup to breakup volcanism affected a total area of approximately $6 \times 10^5 \text{ km}^2$, of which more than half ($3.2 \times 10^5 \text{ km}^2$) is along the eastern margins south of the Wyville Thompson Ridge.

Breakup volcanism. To estimate the volumes and production rates of breakup-related volcanism, several key observations are required: the area covered by erupted basalt together with their intrusive counterparts; thickness estimates based on seismic reflection and refraction data; and estimates of timing based on both geochronology of basalt samples and interpretation of magnetic spreading anomalies.

Eldholm & Grue (1994) considered the areal coverage of flood basalts together with thickness determinations to estimate the volume of extrusive basalt of the NAIP. By assuming a distribution of high velocity lower crust based on limited refraction profiles, the total volcanic volume was constrained. Combined with available timing constraints, they then estimated the average eruptive rates for the entire province. While the distribution of extrusive rocks is fairly straightforward to constrain, a significant uncertainty arises from assumptions regarding the distribution of high velocity lower crust along-strike, discussed further below. Holbrook *et al.* (2001) and Breivik *et al.* (2009) considered igneous thickness variations interpreted along well-resolved 2D seismic refraction profiles to estimate volumes

per km along-strike. The work of Holbrook *et al.* (2001) is based on profiles on the SE Greenland margin and provides constraints on the region south of Iceland, whereas Breivik *et al.* (2009) analysed a profile along the northern Vøring margin and obtained an estimate that applies to the Vøring margin. These latter two approaches provide accurate estimates locally, but interpolating and extrapolating the crustal structure along-strike presents problems in many key areas.

The approach here is to use the mapped SDR sequences to obtain an estimate of the volumes and production rates. In general, the inner SDRs mark the main part of the volcanic transition zone along a volcanic rifted margin. In terms of volumetric significance, they mark the location where the bulk of the breakup-related volcanism occurred. Previous work on volcanic rifted margins shows that the transition from rifting to full seafloor spreading is closely associated with the formation of the main, inner SDRs (e.g. Hinz 1981; Mutter *et al.* 1982). Along the North Atlantic margins, they straddle the COB and the first identifiable seafloor spreading anomalies are typically within this volcanic facies unit. For example, along both the Vøring and SE Greenland margins, magnetic anomaly C24 is well identified and within the mapped inner SDRs (Hopper *et al.* 2003; Breivik *et al.* 2009; respectively). The COB is typically located near to where the thickest new igneous crust is observed, since oceanward the entire crust is volcanic whereas continentward, only a portion of the crust is volcanic.

In addition, the inner SDRs are closely associated with the high velocity lower crust, where P-wave velocities range up to $7.5\text{--}7.6 \text{ km s}^{-1}$ (e.g. Mjelde *et al.* 2007). High velocity lower crust along volcanic rifted margins is normally interpreted as magmatic underplating, although there remains some discussion on precisely what this means and how an underplate is emplaced. As noted by White *et al.* (2008), the term 'underplate' implies that the entire high velocity body represents an igneous cumulate and thus is 100 per cent volcanic. They show, however, that along a seismic profile near the Faroe Islands, the high velocity region most likely consists of sills intruded into the continental crust. Thus the high velocity region landward of the COB may include less breakup-related igneous material than assumed by the pure underplate model.

High velocities in continental lower crust, it should be noted, are not necessarily diagnostic for magmatic underplating. For example, the lower crust beneath Lake Baikal – an intracontinental rift in central Russia – is characterized by a lower crust with velocities from 7.05 to 7.4 km s^{-1} , which ten Brink & Taylor (2002) attribute to an

unthinned part of the Siberian Platform. In the case of volcanic margins, proximity to a major and anomalous volcanic event is a key argument to support the underplating model. Recently, the suggestion that parts of the high velocity lower crust on some margins may be related to hyperextension has raised the possibility that some high velocities may represent serpentinized mantle below very thin continental crust (e.g. Osmundsen & Ebbing 2008). This complicates the interpretation of rifted margins that may have undergone a phase of protracted hyperextension prior to volcanism and final breakup as suggested for the North Atlantic (e.g. White *et al.* 2008; Lundin & Doré 2011). Lundin & Doré (2011) propose that the anomalously wide zone of high velocity lower crust below the Vøring margin can be explained if the innermost part of the margin is underlain by serpentinized mantle and only the outer part of the margin includes intruded lower crust.

Given the large uncertainties over how much of the high velocity lower crust along the Atlantic margins is related to volcanism, and, if it is indeed related, how much of the lower crust is composed of new igneous intrusions, the approach here is to not include mapping of the lower crust directly in the volume calculations. Nevertheless, the bulk of the high velocity lower crust is included in an ad hoc way since much of it falls within the assessment polygons used to estimate volcanic volume. This is explained further below in the description of the assessment polygons using an example from the SE Greenland margin.

To assess the volume and productivity of breakup volcanism, the North Atlantic is broken down into distinct conjugate margin segments to better quantify the differences between the various areas. Four distinct segment pairs are recognized: SE Greenland and Hatton–Rockall margins; East Greenland and the Faroe Islands along the GIFRC; the Jan Mayen microcontinent and the Møre margin; and NE Greenland and the Vøring–Lofoten margins. Holbrook *et al.* (2001) note a dramatic difference in igneous crustal thickness that develops over the GIFRC compared to other areas that is attributed to proximity to the Iceland plume track. They divide the North Atlantic into proximal margins and distal margins based on the influence of active upwelling from the Iceland plume. In this analysis, conjugate comparisons are focused on the distal margins for the purposes of examining transient volcanic margin formation away from the plume track.

Except for the GIFRC, which is described separately, assessment polygons are constructed for each margin segment based on the mapped inner SDRs and magnetic Chron C23n.2no (51.826 Ma; (Ogg 2012; Gaina *et al.* 2016)). The polygons constructed

are shown in Figure 6. Along each margin segment, the COB based on Funck *et al.* (2014) falls within the SDRs, consistent with the general understanding that they are closely associated with the initiation of seafloor spreading. Seaward of the COB, the entire crust up to Chron C23n.2no is newly accreted intrusive and extrusive volcanic material and the volume can be calculated directly from the area of the oceanward assessment polygon combined with the regional crustal thickness derived from wide-angle seismic data (Funck *et al.* 2016). In these areas, all of the high velocity lower crust is included in the estimate.

Landward of the COB, the total volume of the volcanic rocks is more difficult to estimate and several simplifying assumptions must be made. In general, the amount of volcanism decreases rapidly landward and the extrusive layer of SDRs thins before pinching out entirely. In general, the high velocity lower crust also thins landward. An example from SE Greenland is shown in Figure 7. This example is based on the seismic reflection and refraction profile of Hopper *et al.* (2003). The inner SDRs span a distance of over 100 km and the seawardmost extent are emplaced within unambiguous oceanic crust (Chron C24, see Hopper *et al.* (2003)). At the landward end, the SDRs thin and pinch out. A high velocity lower crust is observed that also thins and pinches out. The location of the COB along this profile is poorly constrained. The COB – based on the seismic velocity structure – is far landward, approximately 50 km further landward than the first clear spreading anomaly (Hopper *et al.* 2003). From this point and landward, the transition zone comprises both continental crust and igneous intrusive and extrusive rocks associated with breakup. The volume of crust from the seismic COB to the pinchout of the extrusive layer is $c. 930 \text{ km}^3 \text{ km}^{-1}$, of which $c. 130 \text{ km}^3 \text{ km}^{-1}$ is the extrusive basalts. The question remains how much of the high velocity lower crust is intrusive v. continental crust. Assuming a pure underplate from 7.0 km s^{-1} to the Moho, the volume of intrusives is $c. 320 \text{ km}^3 \text{ km}^{-1}$. Thus, approximately half of the transition zone crust is accreted volcanic rock. The ratio of intrusive to extrusive rock is 2.4:1, very similar to typical oceanic crust (White *et al.* 1992), but much higher than expected if the lower crust is sill intruded continental crust (e.g. White *et al.* 2008).

Based on the SE Greenland example, it is assumed that half of the volcanic transitional crust, defined as the area between the COB and the pinchout of the inner SDR, is newly accreted material associated with breakup. This approach has slightly different implications for each margin segment. However, given the previously noted uncertainties in constraining volcanic volumes, this assumption

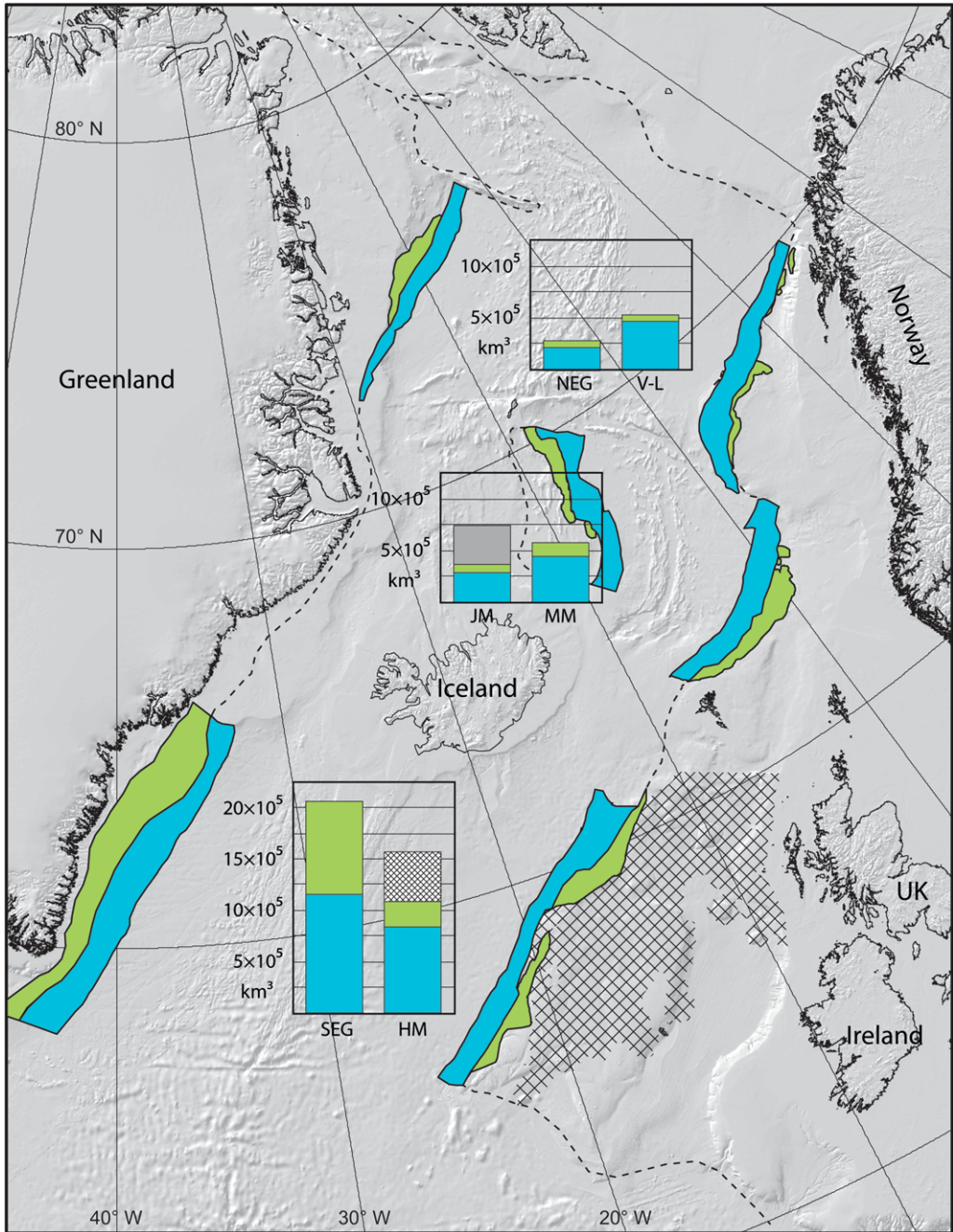


Fig. 6. Breakup-related volcanism around the North Atlantic. Blue polygons are oceanic crust from the COB (*c.* 56 Ma, dashed line) to magnetic Chron 23r (*c.* 52 Ma); green polygons are mapped SDR sequences located over continental crust. Hashed areas are landward and inner flows along the Rockall–Hatton margins south of the Wyville Thompson Ridge. Histograms show the calculated volumes of igneous accretion. The additional grey component for Jan Mayen is the estimated volume of onshore flood basalts from East Greenland. See text for details. NEG, NE Greenland margin; V-L, Vøring-Lofoten margin; JM, Jan Mayen microcontinent; MM, Møre margin; SEG, SE Greenland margin; HM, Hatton margin.

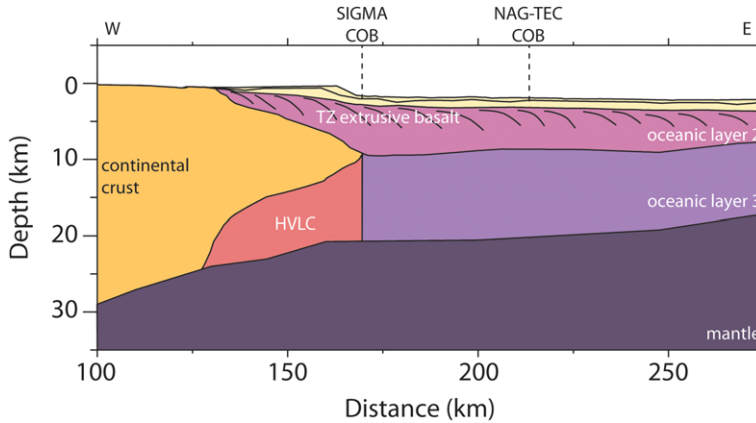


Fig. 7. Schematic cross section along the SIGMA III profile along the SE Greenland margin highlighting the key aspects of a typical volcanic margin. The volcanic transition zone (TZ) is defined as the area where both continental crust and breakup-related volcanic rocks are observed landward of the COB. The SIGMA COB is from Hopper *et al.* (2003) and the NAG-TEC COB is from the NAG-TEC Atlas (Funck *et al.* 2014). Magnetic Chron C24r is located at km 230. Oceanic layer 3 merges into the high velocity lower crust (HVLC) below continental crust. The HVLC thins and pinches out landward approximately where the extrusive basalts also pinch out. From km 125 to the SIGMA COB, about half the crust is volcanic assuming that the HVLC is magmatic underplate. See text for additional discussion.

is broadly consistent with previous estimates and provides a basis for relative comparisons between margin segments. Two consequences of this assumption are: (1) volcanic volumes associated the landward flows and other volcanic units landward of the SDRs are missing; and (2) any intrusive rocks landward of the SDR pinchout are missing. Intrusive rocks associated with the bulk of the high velocity lower crust are included in an ad hoc fashion as described above. In addition, the misplacement of the COB will lead to further errors. If placed too far landward, the volcanic volume may well be overestimated since the entire crust is assumed volcanic. In contrast, if placed too far seaward, the volcanic volume will be underestimated.

As previously discussed, landward flows are locally important, but only the Rockall–Hatton margins show significant areal coverage that potentially has a large impact on the volume estimate. Estimating volumes of these units is complicated by the fact that in most areas, the thickness of the landward flows is unconstrained except where drilled, and there is no straightforward way to estimate sill volumes. In addition, some portion of these units includes pre-breakup volcanism. On all margins except for Rockall–Hatton, these units are therefore not included. Where drilled along Rockall, Hatton and the Faroes, the thickness of landward lava flows is commonly more than 1 km (Varming 2009), and on the Outer Hebrides High, 2–2.5 km of lava flows are estimated (Stoker *et al.* 2012). For this exercise, a 1.5 km-thick layer of

basalt is included for the Rockall–Hatton margins where landward flows and sills are mapped (see Figs 4 and 6).

The issues of COB placement and high velocity lower crust potentially affect two areas of investigation. In general, the COB is poorly constrained along much of the East Greenland margin. Discrepancies between the placement of the COB based on magnetic anomaly interpretations *v.* seismic velocity structure have already been noted along the SE Greenland margin. To avoid overestimating the volcanic volume and productivity here, the COB has been placed 50 km seaward from that interpreted by Hopper *et al.* (2003). This better reconciles misfits to reconstructions based on magnetic spreading anomalies (Gaina *et al.* 2016), and takes into account the possibility that hyperextended continental crust underlies part of the Greenland margin as proposed by White *et al.* (2008). Along NE Greenland, Voss *et al.* (2009) suggest that the NE Greenland margin in the vicinity of the Jan Mayen fracture zone broke up much later than further north. A clear Chron C24 anomaly to the north becomes indistinct to the south, and the first clear anomaly in the southern area is C21. They place the COB further seaward than plate reconstructions suggest (Gaina *et al.* 2009). The entire high velocity crust in their interpretation forms transitional crust. Here, the COB is placed 60 km landward from estimates by Voss *et al.* (2009) and 25 km seaward from those by Gaina *et al.* (2009) as a compromise between the seismic evidence

and the magnetic anomaly evidence. Thus, in terms of the Voss *et al.* (2009) interpretation, the volumes here are overestimated, since a significant portion of transition zone crust is assumed 100 per cent volcanic, whereas in the Gaina *et al.* (2009) interpretation, the volume is underestimated.

The East Greenland margin north of Iceland is further complicated by the double breakup associated with the formation of the Jan Mayen microcontinent (see Blischke *et al.* 2016). High velocity lower crust and anomalously thick oceanic crust is observed off East Greenland conjugate to the Jan Mayen microcontinent (e.g. Weigel *et al.* 1995; Voss *et al.* 2009). However, the oceanic crust is associated with the second breakup event (see age grids and magnetic spreading anomalies of Gaina *et al.* (2016)). Onshore, tholeiitic lavas have been dated to 56–53 Ma and are thus clearly related to the first breakup event and separation of the eastern Jan Mayen microcontinent from the Møre margin. Additional volcanism continues until the Miocene (Larsen *et al.* 2014). Given the protracted history of volcanism and possibility that much of the high velocity lower crust is associated with the second breakup event (e.g. Voss *et al.* 2009), this region is considered separately in the volume estimation.

Finally, to allow for comparison to previous work, the volumes and productivity of the proximal region over the plume track are estimated to constrain the breakup-related volcanism of the entire province. The main complication here is that spreading anomalies are not clearly identifiable along the GIFRC. This is in part a result of repeated ridge relocation in response to the relative motion between the mid-ocean ridge spreading system and the Iceland mantle plume. In addition, along the GIFRC the location of the COB is unconstrained on the Greenland side, and poorly constrained on the Faroes side, so constructing a reasonable assessment polygon as above for the distal margin segments is not possible. Instead, a 400 km-wide plume track is assumed (e.g. Holbrook *et al.* 2001), and the initial opening rates based on plate reconstructions are used to estimate the total volume of new crust. At the time of breakup, the spreading rate between Greenland and Europe was relatively fast before decreasing to the slow rates observed today. Half rates as high as 45 mm a^{-1} have been estimated off East Greenland (Larsen & Saunders 1998) and regional reconstructions suggest values around 30 mm a^{-1} (e.g. Gaina 2014). Richardson *et al.* (1998), Smallwood *et al.* (1999) and Holbrook *et al.* (2001) show that the crustal thickness of the GIFRC is on average approximately 30 km thick. From breakup to anomaly C23n.2no, which spans 4.2 million years, $c. 3.1 \times 10^6 \text{ km}^3$ of volcanic rock was emplaced along the GIFRC. This does not include the landward flood basalts, which Larsen

et al. (1999) estimate to have a total volume of $0.25 \times 10^6 \text{ km}^3$, including both the East Greenland and Faroese lava flows. Assuming an intrusive to extrusive ratio of 2:1, an additional $0.75 \times 10^6 \text{ km}^3$ is added to the total volume for the GIFRC.

The results are summarized in Figure 6 and Table 1. The average productivity assumes that the duration of volcanism spans from breakup at 56 Ma to anomaly C23n.2no at 51.8 Ma. All three of the margin segments distal to the Iceland plume track show asymmetry, with one margin showing nearly double the volume of volcanic rock as the other. Given the problems with estimating the amount of volcanism over the transition zone, the plots are further broken down into the oceanic and continental areas. Significantly, the earliest oceanic crust produced shows the same volumetric conjugate asymmetry as the full margin estimate along all three conjugate segment pairs.

Conjugate asymmetry. Between SE Greenland and the Hatton Bank, asymmetry in early accretion has been noted previously (Hopper *et al.* 2003). While there remains some uncertainty over the placement of the COB along Greenland, here it is placed significantly further seaward than Hopper *et al.* (2003). This should account for the possibility that portions of the margin are underlain by hyper-extended crust as proposed by White *et al.* (2008). Despite this, there still appears to be more breakup-related volcanism along East Greenland, even when taking into account possible volumes of landward flows and sills covering the Hatton and Rockall basins. Thus, it seems clear from the analysis here that significant volcanic asymmetry between SE Greenland and the Hatton–Rockall margins is a robust observation.

North of Iceland, the sense of asymmetry is the opposite between NE Greenland and the Vøring–Lofoten margins. Here too, there are large uncertainties regarding the COB placement, especially along the Greenland margin where data are sparse. Assuming the Voss *et al.* (2009) placement of the COB, the volume along NE Greenland is potentially even less and would only make the observed asymmetry more dramatic. If instead, the COB is moved landward to where Gaina *et al.* (2009) place it, the volcanic volume would increase to approximately $0.36 \times 10^6 \text{ km}^3$, still significantly less than the Norwegian margins.

For Jan Mayen, it remains somewhat unclear how much volcanism along East Greenland should be included as part of the GIFRC or as part of the Jan Mayen microcontinent. Assuming that half of the estimated volume in Larsen *et al.* (1999) is from the Greenland side, more symmetric volcanism may be indicated (Table 1). In addition, the sense of asymmetry could be the opposite if a

Table 1. Volcanic volumes and production rates of North Atlantic margins

Margin segment	Volume (km ³)	Per cent of pair	Average productivity (km ³ km ⁻¹ Ma ⁻¹)	Previous productivity estimates ⁵
SE Greenland margin	2.06 × 10 ⁶	65%	490	576 ²
Hatton–Rockall margins	1.09 × 10 ⁶	35%	258	125 ²
Jan Mayen MC ⁶	0.38 × 10 ⁶ (0.76 × 10 ⁶)	39% (56%)	199	
Møre margin	0.59 × 10 ⁶	61% (44%)	312	
NE Greenland margin	0.29 × 10 ⁶	36%	107	
Vøring/Lofoten margins	0.53 × 10 ⁶	64%	195	
Vøring only	0.42 × 10 ⁶		366	428 ³
North of Iceland	1.79 × 10 ⁶	36%	388	
South of Iceland	3.14 × 10 ⁶	64%	748	700 ²
Greenland–Iceland–Faroe Ridge Complex ⁴	3.85 × 10 ⁶		2246	1800 ²
Total NAIP Atlantic margins	8.78 × 10 ⁶		864	830 ¹ , 800–1000 ²

¹Eldholm & Grue (1994).²Holbrook *et al.* (2001).³Breivik *et al.* (2009): note that in terms of productivity, the estimate for the Vøring is less than in Breivik *et al.* (2009). However, their estimate is based on a single profile along a relatively magma-rich portion of the Vøring margin, whereas here, the productivity has been averaged across the entire margin and includes magmatically less robust areas.⁴Includes volumes estimated by Larsen *et al.* (1999) for the onshore East Greenland basalts and the Faroe Island basalts.⁵It should be noted that productivity estimates are less easy to compare than total volumes, since different authors use different timescales. For example, Eldholm & Grue (1994) consider the volume of crust from breakup to anomaly C23, similar to here. However, they assume 3 myr for this interval compared to 4.2 myr here based on the Ogg (2012) timescale and a breakup time of 56 Ma.⁶The values in parentheses are the volumes if the Blossville Coast volcanism is included as part of the Jan Mayen volcanism.

substantial volume of early breakup volcanism occurred along East Greenland in the Jameson Land Basin. From Scoresby Sund up to the western Jan Mayen fracture zone, the areal extent of this region is approximately 27 500 km². Mathiesen *et al.* (2000) estimate that 2 km of basaltic lava flows covered the Jameson Land Basin and were subsequently eroded away. High velocity lower crust observed by Weigel *et al.* (1995) could be the intrusive counterpart to the eroded basalts. In that case, 5–6 km of intrusive rocks are unaccounted for in the volume estimates here. As noted earlier, volcanism along this part of the margin continued into the Miocene (Storey *et al.* 2004) and much of the volcanism may be younger. Thus, it remains unclear from the available data if this segment is asymmetric, and if so, which side has more volcanism.

For both margin segments north of Iceland, there are significant uncertainties for many aspects of the estimates here. Thus, while the mapping suggests possible asymmetries, it is clear that additional work and more systematic conjugate analysis are required to investigate this further. A useful exercise would be to re-analyse key conjugate seismic refraction profiles to establish a consistent interpretation of the velocity structure and implications for crustal types and COB placement.

Asymmetric spreading is known from many studies of mid-ocean ridges and back-arc spreading

systems (Hayes 1976; Stein *et al.* 1977; Barker & Hill 1980). This is thought to be a response to the migration of the lithosphere over the asthenosphere, skewing the thermal field and leading to differences in the shear traction at the base of the plate that causes the ridge to migrate. In these models, faster accretion is predicted over the cooler plate (Hayes 1976; Barker & Hill 1980). Müller *et al.* (1998) noticed that many areas of asymmetric accretion are associated with proximity to mantle plumes. They showed that the ridge axis migrates towards the warmer, plume-affected mantle resulting in a deficit of accretion on the plate over the plume and a surplus of accretion of the 'cooler' plate. Hopper *et al.* (2003) suggested that a similar thermal effect could explain asymmetric accretion between SE Greenland and the Hatton Bank since the Greenland margin is bounded by thick, and presumably cooler, Archean lithosphere. Applying this idea to the north, proximity of the Norwegian margin to the thick, cooler lithosphere of the Baltic shield could result in a similar effect, resulting in a crustal deficit along East Greenland. The kinematic modelling, however, indicates an accretional crustal deficit along the magmatically more robust margin, calling into question how important early ridge migration might be during volcanic margin formation. Along the Norwegian margins, it appears that the excess volumes are primarily expressed as anomalously thick crust along the Vøring margins

compared to NE Greenland (Funck *et al.* 2016; Haase *et al.* 2016). Nevertheless, the possibility that the magmatically rich half of the conjugate pairs seems to be the one closest to the stable cratons and thick lithosphere is intriguing and suggests pre-existing lithospheric structure may play a role in asymmetric volcanic margin development.

North–south asymmetry. The variation in volcanism away from Iceland along the North Atlantic margins has been considered previously in a number of studies. Barton & White (1995) noted an apparent symmetric distribution of estimated melt thickness away from the Faroe–Iceland Ridge by comparing profiles from Edoras Bank, Hatton Bank, the Vøring Plateau and the Lofoten Basin margin. This was explained as a decrease in asthenospheric potential temperature away from the plume centre. In contrast, Eldholm & Grue (1994) noted that the areal coverage of basaltic lavas over continental crust is greatest south of Iceland. They suggest decreased melt production to the north as a response to a propagating rift away from a plume impact site beneath central East Greenland. This mechanism also reflects decreasing temperature away from the mantle plume as heat is rapidly advected out of the system by melting after plume impact. Why the same effect does not propagate away from the plume to the south is not explained, however. Similar to Eldholm & Grue (1994), Voss *et al.* (2009) noted that the northward decrease in volcanism is more pronounced than the southward decrease in volcanism away from Iceland. They further note that to north, the volumes indicated by high velocity lower crust are very different than to south, but note the complication with the two breakup events and the later separation of Jan Mayen from East Greenland.

A comparison between the volume of volcanism north of Iceland v. south of Iceland is shown in Figure 8. The results here confirm the observations of Eldholm & Grue (1994) and Voss *et al.* (2009) that there is significant north–south asymmetry in early NAIP volcanism. This is the case for both pre-breakup volcanism, indicated by the mapped landward flows, as well as the main breakup volcanism, indicated by the SDR sequences and the development of the main part of the volcanic margins.

It is well established that the present day influence of the Iceland plume is highly asymmetric (Howell *et al.* 2014). A strong influence to the south along the Reykjanes Ridge is well documented whereas the influence to north is significantly reduced (e.g. Vogt 1971; Ito 2001; Delorey *et al.* 2007). This is typically attributed to a deflection of the Iceland mantle plume to the south, causing stronger plume–ridge interaction. A key unanswered question is when this stronger influence to the south began, and why there is a deflection to

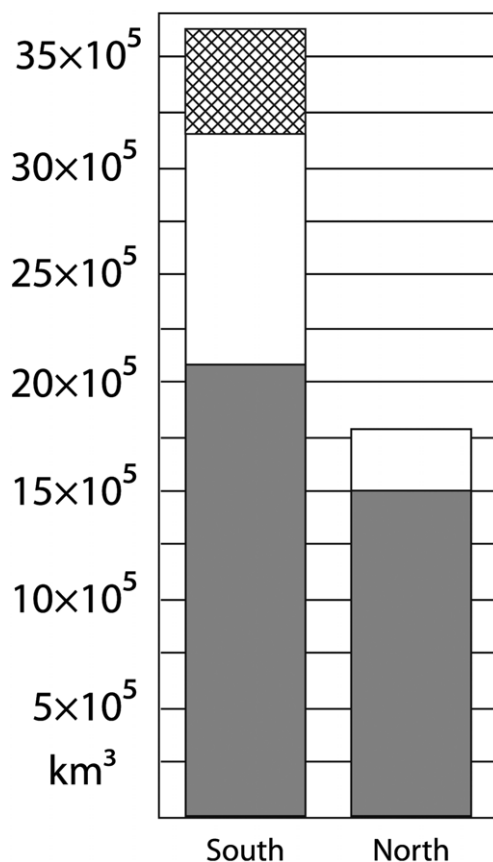


Fig. 8. Histogram summarizing the total volume of material erupted at breakup to the north and to the south of the GIFRC.

south. Based on the patterns of both the pre-breakup and breakup volcanism, it is suggested that this southward-biased influence has existed for the entire history of the NAIP and that this pattern probably reflects pre-existing lithosphere structure.

Following Nielsen *et al.* (2002), it is assumed that the Iceland plume impinged beneath central Greenland in the Paleocene and is preferentially channelled into thin spots. It is well documented that the proto-North Atlantic experienced a long period of extension from the Late Palaeozoic and throughout the Mesozoic. Figure 9 shows the Cretaceous stratigraphic distribution (Stoker *et al.* 2016) reconstructed to 80 Ma, which highlights the pattern of lithospheric thinning prior to the arrival of the plume beneath Greenland. To the north, rifting follows the Iapetus suture but splits into two distinct branches towards the south, one along the Rockall–Hatton region, and one towards the North Sea. Closer proximity of the Rockall–Hatton region to

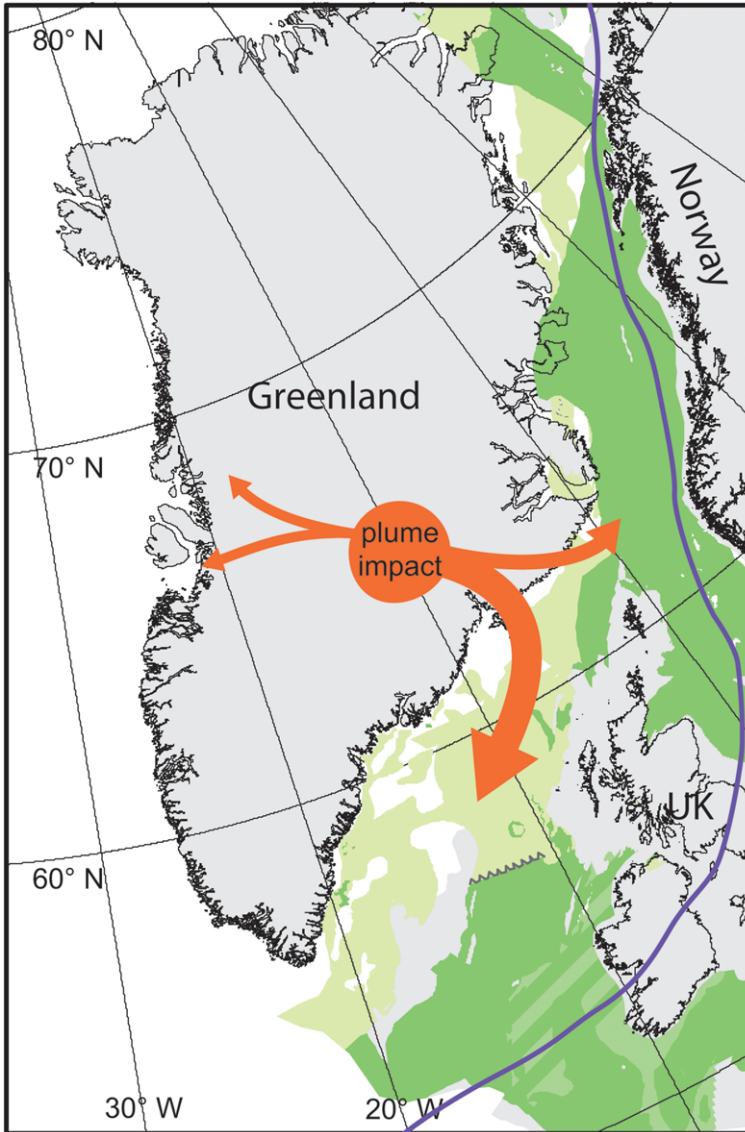


Fig. 9. Distribution of Cretaceous basins reconstructed at 80 Ma, approximately 20 myr prior to plume impact in the Paleocene. Dark green are areas with proven Cretaceous, light green are areas of inferred Cretaceous. Dark blue line is the Iapetus suture. Plume impact location proposed by Nielsen *et al.* (2002) is shown, and arrows indicate lateral flow into lithospheric thin spots. Material is preferentially channelled into the Rockall–Hatton margins as a consequence of the pre-existing lithospheric configuration.

the site of the impacting plume probably led to it being the preferred region for lateral flow of plume material.

The northern limit of NAIP volcanism. In the discussion so far, one feature mapped to the north has been ignored. Along the SW Barents margin, a volcanic unit referred to as the Vestbakken Volcanic

Province is shown on most maps of the NAIP (e.g. Abdelmalak *et al.* 2015). Faleide *et al.* (1988) interpret seismic reflection data as showing evidence for a volcanic basement along the shear margin. Volcanism associated with the shear margin is further indicated in exploration well 7316/5-1, where sills intruded into Eocene sediments were encountered (<http://factpages.npd.no> [Last accessed 26 June

2017]). The sills themselves, however, have not been dated. At another exploration well, 7216/11-1S, tuffs and other volcanic debris are found in Upper Cretaceous through Eocene sediments (Ryseth *et al.* 2003). Additional evidence for volcanism comes from basalts recovered in shallow cores from south of Bjørnøya (Mørk & Duncan 1993). These were dated to Late Pliocene and thus are unrelated to earlier breakup events of early shear margin evolution.

Eldholm *et al.* (2002) note that there is no evidence for SDR units along the shear margin. Berndt *et al.* (2001) show a strong decrease in magmatism to the north along the Norwegian margins. This is also the case along the NE Greenland margin from the Jan Mayen fracture zone towards the East Greenland Ridge (Geissler *et al.* 2016). Further north between the East Greenland Ridge and the Hovgaard Ridge, poorly developed SDRs (or better lava flows) may be identified (Geissler *et al.* in press). These are fairly minor compared to typical development of SDRs in the very magma-rich areas such as the SE Greenland and Vøring margins. In addition, much of the Boreas Basin is interpreted to be underlain by very thin crust with indications for serpentinized mantle. Within the interpreted oceanic areas, the crustal thickness of the region shows normal to thin crust north of Lofoten and the East Greenland Ridge. Overall, the evidence for significant Paleocene to Eocene volcanism is generally lacking in this region. Thus, we question if the Vestbakken Volcanic Province should be considered part of the NAIP. Geochemical data showing an Icelandic signature would be needed to firmly establish a connection between the NAIP and Vestbakken volcanic rocks, but no data exist to determine this.

Conclusion

New mapping of the NAIP establishes the pattern of volcanism associated with breakup and the initiation of seafloor spreading of the region. The new map provides a consistent view of the distribution of volcanism along the eastern margins and presents some of the first maps available for the entire East Greenland margin and Jan Mayen microcontinent. A comparison of conjugate margin pairs shows that the NE Greenland/Vøring–Lofoten margins and the SE Greenland/Rockall–Hatton margins are highly asymmetric, with the bulk of the volume of new igneous crust emplaced on one side. An explanation for this observation remains elusive, but we note that in both cases, the relatively magma-rich margin is in close proximity to thicker lithosphere associated with the stable cratons. Finally, the amount of pre-breakup and breakup volcanism is strongly

asymmetric from south to north, with significantly more volcanism south of the GIFRC. Today, the Iceland plume shows a much stronger interaction to the south along the Reykjanes Ridge compared to the Kolbeinsey Ridge to the north. We suggest that this southward bias has persisted since pre-breakup. After impact of the Iceland plume in the Paleocene, lateral flow of ponded plume material was preferentially channelled along the proto-Hatton margin, resulting in the bulk of the magmatism being emplaced to the south. The patterns of volcanism thus show that pre-existing lithospheric structure plays a first-order role in the development of the North Atlantic volcanic margins. Finally, we note that the northern limit of the NAIP appears to be the northernmost Lofoten margin and NE Greenland margin up to the East Greenland Ridge. The Vestbakken Volcanic Province and its NE Greenland counterpart are suggested to be related to local tectonics associated with shear margin development and thus we question if it is part of the NAIP.

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