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Key Points:

- Seismic reflection data are used to map the development of the Logi Ridge north of the West Jan Mayen Fracture Zone
- Lava flows and erosion products off the Logi Ridge show development throughout the Oligocene to the Middle Miocene
- Logi Ridge formation appears to be tied to dynamic topography development up to the Middle Miocene

Supporting Information:

- Supporting Information S1

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Dynamic Topography Development North of Iceland from Subaerial Exposure of the Igneous Logi Ridge, NE Atlantic

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Abstract The Logi Ridge, located north of the West Jan Mayen Fracture Zone, is E-W oriented and 140–150 km long. The seafloor surrounding the Logi Ridge is ~0.65 km shallower than the conjugate seafloor east of the Mohn's Ridge, attributed to asymmetry in the regional NE Atlantic dynamic uplift. Eight reflection seismic lines across the Logi Ridge constrain its development. Both the western and eastern parts have flat tops, indicating erosion at sea level. Three different basement types surrounding the Logi Ridge are observed: rough basement represents abyssal hills original or reactivated by later magmatic/tectonic events; smooth basement caused by basalt flows overprinting early sediments; and irregular basement formed by later basalt flows and intrusions. The surrounding sediments have two distinct units, where the unit boundary is Middle Miocene (12–14 Ma). Mass transport from the Logi Ridge appears episodically throughout Late Oligocene to Middle Miocene, when development ends. The end of erosion age can also be estimated from seamount height, or present top seamount depth. In the west there is agreement with the age constrained by the sedimentation, proving little dynamic topography change. In the east, discrepancies between the methods are explained by 0.15–0.3 km dynamic uplift after submergence. Hence, most of the regional dynamic uplift occurred before the end of the Logi Ridge development in the Middle Miocene, suggesting a causative relationship. Minor recent magmatic growth and seafloor uplift over a ~100 km wide zone southeast of the Logi Ridge may be tied to the younger dynamic uplift in the east.

1. Introduction

The North-East Atlantic is unusually shallow compared to other oceans. Cochran and Talwani (1978) showed that the residual seafloor depth is more than 2,500 m too shallow centered on Iceland, but falling to the north and south. They attributed this to a mass deficiency up to several hundred kilometers deep in the asthenosphere, caused by mantle movement in the Iceland hotspot. Dynamic topography in the NE Atlantic is unevenly distributed, and transient by nature (e.g., Hartley et al., 2011; Parnell-Turner et al., 2014; Poore et al., 2011). Presently, the whole northeast Atlantic is underlain by low shear wave velocity anomalies in the upper mantle, which is related to elevated temperature and to the dynamic topography (e.g., Rickers et al., 2013). In principle, dynamic topography can occur from short wavelength (<100 km) to long wavelength (>1,000 km), and it generally describes the effect of mantle convection on surface topography (Flament et al., 2013).

Existing dynamic topography studies in the NE Atlantic address different times and areas, primarily investigating the effect of small-scale Iceland plume pulsing. From the time of continental breakup at the Palaeocene-Eocene boundary in the Faroe-Shetland basin, interpretation of erosion from 3D seismic reflection data shows that pulses of Iceland plume material flowed laterally beneath the lithosphere, producing a transient uplift >500 m in about 1 Ma, immediately followed by a similar amount of subsidence (Champion et al., 2008; Hartley et al., 2011; Rudge et al., 2008). From a reflection seismic profile across the Reykjanes Ridge south of Iceland, Parnell-Turner et al. (2014) could constrain a continuous 55 Ma record of plume-pulsing cycles related to mantle temperature changes of up to 25–30°C. Based on the development of oceanic current-driven drift sediment deposition, Poore et al. (2011) observed similar dynamic topography

changes along the Greenland-Scotland Ridge over the last 7 Ma, also attributed to temperature fluctuations of $\sim 25^{\circ}\text{C}$. However, the longer wavelength development in the NE Atlantic is not well known, since there are few ways in which it can be investigated.

Parts of the Logi Ridge (named from Norse mythology, where Logi is the master of fire, brother to Aegir, the master of the sea) (Figure 1) have been subaerially exposed and eroded in the past, presenting an opportunity to constrain long-term dynamic topography development in a deep oceanic environment, where that would otherwise not be possible. First we date the formation of the igneous Logi Ridge from determining how mass transport tied to both construction and erosion from the Logi Ridge interacts with the surrounding sedimentation over time. By comparing these results with those from dating methods based on subsidence using parameters like seamount height and current seamount top depth, we estimate to what extent the dynamic topography has changed from the time of the Logi Ridge development ended to the present day, which also puts constraints of dynamic topography development in the past. For this study, crustal age data from seafloor spreading anomalies (Figure 2) are combined with eight seismic reflection transects over the area north of the West Jan Mayen Fracture (WJMFZ) (Figure 1).

2. Regional Geological Setting

During the Paleocene-Early Eocene, the continental breakup between Greenland and Europe was accompanied by massive magmatism that formed part of the Northeast Atlantic Igneous Province (NAIP) (e.g., Eldholm & Grue, 1994; White et al., 1987). The formation of the NAIP is believed to be caused by a long-lived mantle plume, the Iceland plume, bringing up anomalous hot mantle material from the core mantle boundary (e.g., Storey et al., 2007). There is also extensive post breakup magmatic activity in several areas in the NE Atlantic. Excess volcanism occurred along the Faeroes-Iceland-Greenland Ridge (e.g., Eldholm & Grue, 1994). Further north, the Norwegian island of Jan Mayen has an active volcano (Figure 1). Geochemical analysis shows that Jan Mayen magmatism is derived from deep, low-degree mantle melting (Trønnes et al., 1999). The Jan Mayen Platform is located across the WJMFZ adjacent to Jan Mayen, surrounding southern tip of the Mohn's Ridge. A seismic refraction line shows that it has up to 12 km thick igneous crust (Kandilarov et al., 2012). It shows geochemical similarities with the Jan Mayen magmatism (Elkins et al., 2016; Mertz et al., 2004). Also, the area north of the WJMFZ has a large seamount (Vesteris Seamount) and a large ridge (the Logi Ridge) (Figures 1 and 2). However, only the Vesteris Seamount has been dredged and dated. Results show recent magmatism with relatively unradiogenic Sr isotope ratios and highly incompatible trace element patterns, proving a depleted mantle source of non-plume origin (Haase & Devey, 1994). The Eggvin Bank located between Jan Mayen and the east coast of Greenland was created by high magmatic productivity on the northern segment of the Kolbeinsey seafloor spreading ridge (Mertz et al., 2004). A seismic refraction line across the eastern Eggvin Bank shows large variations in crustal thickness (from 8 to 13 km) (Tan et al., 2017). The basalts dredged from within or near the spreading zone of the Eggvin Bank are enriched in incompatible elements and radiogenic isotopic compositions, and excess magmatism appears to be related to an enriched mantle source rather than elevated melting temperature (Elkins et al., 2011, 2016; Haase et al., 2003) (Figure 2). Eggvin Bank tholeiites are geochemically similar to lavas in the southeastern Icelandic volcanic zone, coming from a source different from that of the mantle north of central Iceland (Mertz et al., 2004). Farther to the west, there is intrusive syenitic magmatism on the Traill Ø on the east Greenland margin, dated to 34–40 Ma (Larsen et al., 2014; Noble et al., 1988; Price et al., 1997). Traill Ø is located landward of the WJMFZ, and the syenitic intrusions are tied to the westernmost end of a large elongate magnetic high on the shelf (Traill Ø Igneous Complex [TIC]) (Figure 2).

Our study area is located between the WJMFZ and the Vesteris Seamount (Figure 1). The Logi Ridge is E-W oriented and 140–150 km long, with most of the top at 400–660 m depth. However, at the western end, the Logi Ridge is reduced in size and merge with the surrounding seafloor towards the foot of the continental slope. The western Logi Ridge (WLR) is a continuous structure, while the eastern Logi Ridge (ELR) consists of an isolated seamount (Figure 1). The WLR and the surrounding area were covered obliquely by one seismic refraction and one reflection profile targeting the more near-shore areas of east Greenland north of the WJMFZ (Berger & Jokat, 2008; Geissler et al., 2017; Voss & Jokat, 2007). The velocity model shows a crustal thickness of up to 12 km under the WLR, the root somewhat wider than the surface expression, but its origin and development was not discussed (Voss & Jokat, 2007). A maximum of 1.8 km thick sedimentary cover was observed adjacent to both flanks (Voss & Jokat, 2007). The oceanic crust over the study area was formed at an average slow half-spreading rate of 8 mm/year (Gaina et al., 2016). The Logi Ridge is located

Figure 1. Bathymetry map (ETOPO 2V2) (National Geophysical Data Center, 2006) showing regional features, where the study area is indicated by a dashed circle. Early Cenozoic basalt flows or intrusions are illustrated as black areas onshore Greenland (Noble et al., 1988). Four seismic reflection profiles along the east Greenland margin (Berger & Jokat, 2008) are shown by solid black lines, tying to ODP drill site 913 marked with the white dot. The solid white arrows indicate the East Greenland Current (EGC) (Rudels et al., 2012). The inset picture shows detailed bathymetry (IBCAO) (v.3) (Jakobsson et al., 2012) of the study area and seismic reflection profiles used for this study. The western Logi Ridge is labeled WLR, and the eastern is labeled ELR. EB = Eggvin Bank; FIR = Faeroes-Iceland Ridge; GIR = Greenland-Iceland Ridge; JM = Jan Mayen; JMP = Jan Mayen Platform; KR = Kolbeinsey Ridge; LR = Logi Ridge; MR = Mohn's Ridge; NB = Norway Basin; NEGM = Northeast Greenland margin; RR = Reykjanes Ridge; TØ = Traill Ø; VP = Vøring Plateau; VS = Vøring Spur; V = Vesteris Seamount; WJMFZ = West Jan Mayen Fracture Zone.

on oceanic crust of chrons C13–C20 (~33–43 Ma) (Figure 2). Magnetic data show that the Logi Ridge primarily has a normal magnetic field cross-cutting older oceanic crust of a reversed magnetization (C18–C20: ~40–43 Ma), supporting a younger origin than the underlying oceanic crust (Figure 2).

The Logi Ridge is located along the same trend as that of the large magnetic anomaly on the shelf (TIC in Figure 2), which suggests that it could be related to the Traill Ø event (Breivik et al., 2012). However, both the bathymetric expression of the Logi Ridge and the positive magnetic anomaly on the shelf disappear towards the foot of the continental slope, so direct continuity appears to be lacking (Figures 1 and 2).

The seafloor surrounding the Logi Ridge has on average 0.65 km shallower bathymetry compared to the conjugate oceanic crust east of the Mohn's Ridge (Breivik et al., 2008). This is in agreement with the dynamic topography asymmetry estimated from gravity by Rickers et al. (2013). Their study also shows that there

Figure 2. Regional magnetic anomaly map (Verhoef et al., 1996). Seafloor spreading anomalies with ages are annotated from Gaina et al. (2009, 2017). Thick black lines show the multichannel seismic reflection lines by the Norwegian Petroleum Directorate (Sandstå et al., 2012), while the white lines indicate our own single-channel seismic reflection survey used in the study. White-filled circles show the location of the OBS positions along the Profile 1 (Tan et al., 2017). Yellow solid lines show seismic profiles along the east Greenland margin (Berger & Jokat, 2008). The northern part of the Kolbeinsey Ridge (NKR) has been dredged (red circles, Haase et al., 2003; yellow circles, Mertz et al., 2004; yellow triangles, Elkins et al., 2011), while the white triangles indicate the dredge samples from Vesteris Seamount (Haase & Devey, 1994). Thin black lines show bathymetry contours at 500 m interval (Jakobsson et al., 2012). Flow lines calculated from rotation poles by Gaina et al. (2017) are shown by white, dashed lines, traced from C5 anomaly. Most of the Logi Ridge has a normal magnetic field partly cross-cutting older oceanic crust of a reversed polarity. LR = Logi Ridge; NKR = Northern Kolbeinsey Ridge; TØ = Traill Ø; TIC = Traill Ø Igneous Complex; WJMFZ = West Jan Mayen Fracture Zone; V = Vesteris Seamount.

is good correlation between the dynamic topography and low shear wave velocities in the upper mantle between 100 and 200 km depth.

3. Data and Methods

3.1. Data Acquisition and Processing

The seismic reflection data covers the Logi Ridge and oceanic crust north of the WJMFZ. The first data set consisting of single-channel reflection profiles (Profiles 1–6) (Figures 1 and 2) was acquired during the early fall of 2011 by the R/V Håkon Mosby. The survey was a collaboration between the Department of Geosciences, University of Oslo, and the Department of Earth Science, University of Bergen, Norway. A single-channel streamer (offset channel to source: 102 m) was used as the sound receiver. Profile 1 was acquired using four equal-sized Bolt air guns with a total volume of 78.6 l, which were fired every 200 m for use with ocean bottom seismometers (OBSs). The velocity modeling of the wide-angle seismic data along Profile 1 has been published by Tan et al. (2017). A smaller source was used for the other profiles (2–6), consisting of a tuned air gun array of 12.5 l. It was fired every 50 m at a tow depth of 6 m. Processing of the seismic data (Profiles 1–6) consisted of spiking deconvolution, 5–60 Hz band-pass filtering and amplitude corrections with a geometrical exponential factor of 1.2. The second data set consists of 2D multichannel reflection seismic lines acquired in 2011 by the Norwegian Petroleum Directorate (NPD) (Sandstå et al., 2012), from which we select the two lines (NPD0004 and NPD0005) crossing the Logi Ridge. These were collected using Bolt LLX type air guns with a total volume of 67.7 l and an 8,100 m long multichannel streamer. The air

Figure 3. Cartoon of how sediment loading is corrected for to derive seamount erosion ages based on the relationship between the seamount height, seamount top depth, and normal age-dependent thermal subsidence of the oceanic crust. During the last stage of the seamount erosion, the seamount height corresponds to the depth of the surrounding basement. However, the basement has additional subsidence due to the sediment load, making the erosion age look younger. Similarly, seamount top depth will be affected by the sediment deposited after seamount submergence, making the erosion age look older if not corrected for.

gun was fired every 25 m at a source depth of 7 ± 1 m. The seismic data underwent a standard industry processing sequence including migration, and main steps of the processing scheme are available via supporting information. The depth of the seafloor (Profiles 1–6) is derived from echo-sounder recordings, while the seafloor depth of NPD0004 and NPD0005 is estimated from the seabed reflection times using a water velocity of 1,480 m/s. No high-resolution bathymetry is available over the Logi Ridge, and currently the IBCAO v.3 ship track compilation appears to be the best available (Jakobsson et al., 2012) (Figures 1 and 10).

3.2. Dating Mass Transport Off the Ridge from Sedimentary Record

The goal of the seismic interpretation is to investigate how mass transport from the Logi Ridge interacts with the regional sedimentation over time. In order to obtain that, we need to convert the time sections to depth. We used a regional empirical velocity-depth trend for the sedimentary package ($V_p = 1.8 + 0.7Z$ km/s, Z : sediment thickness in km) based on regional seismic refraction studies (Tan et al., 2017; Voss & Jokat, 2007) (Figure 2). Once the sections were converted to depth, it is possible to quantify rates of both deposition and subsidence that characterize the area. Deposition rates over time can be determined by dating a key horizon. This was done by tying it to a ODP borehole to the north of our study area by regional seismic lines (Berger & Jokat, 2008; Myhre et al., 1995) (Figure 1). Alternatively, the age can be determined by correlating sedimentary thickness below the horizon against seafloor age. Seafloor ages can be derived from Gaina et al. (2017). By doing linear regression, we can estimate the age for which the thickness is zero and thus obtain an independent dating of the horizon. Having a good inter-sedimentary dating makes it possible to calculate deposition rates above and below the horizon. These again can be used to date deposits derived from the Logi Ridge based on their position within the sedimentary section.

Oceanic basement depth increases with the square root of age ($\text{Ma}^{1/2}$) (e.g., Adam & Vidal, 2010; Stein & Stein, 1992). For the analysis of dynamic topography development, it is necessary to establish both a subsidence rate and a zero-age depth for the area. That can be done by plotting basement depth against the square root of age and then use linear regression to fit the data. The basement used should not be affected by later igneous intrusions and mass transport off the Logi Ridge. To isolate the subsidence due to cooling only, the effect of sediment loading is removed by local Airy isostatic correction (Le Douaran & Parsons, 1982):

$$B_s = B - \frac{(\rho_s - \rho_w)}{(\rho_a - \rho_w)} Z_s, \quad (1)$$

where B is the basement depth, B_s is the basement depth corrected for sediment loading, ρ_a is the density of asthenospheric mantle ($3,200 \text{ kg/m}^3$), ρ_w is the density of water ($1,030 \text{ kg/m}^3$), Z_s is the sediment thickness, and the average sediment density (ρ_s) is $1,950 \text{ kg/m}^3$, which is derived from empirical velocity-density relationships (Ludwig et al., 1970) based on a regional seismic refraction study (Voss & Jokat, 2007).

3.3. Dating Ridge Submergence from Seamount Height and Top Depth

Parts of the Logi Ridge have flat tops, indicating that it has been eroded above or close to sea level. By assuming age-dependent thermal subsidence of the oceanic crust, we can estimate the time when the Logi

Ridge was last subaerially exposed (e.g., Caplan-Auerbach et al., 2000) (Figure 3). Here we assume that sea level change during the Neogene is small (e.g., Conrad, 2013). Two somewhat different methods that both utilize that erosion of the Logi Ridge stopped after it has subsided below sealevel can be applied: The first method uses the Logi Ridge height to record the surrounding oceanic depth at the time of submergence (t) (equation (2)):

$$d_i + S\sqrt{t_s - t} = H; t = t_s - \left(\frac{H - d_i}{S} \right)^2, \quad (2)$$

where d_i is the depth of the mid-oceanic ridge at 0 Ma, t_s is the age of the underlying crust, S is the local thermal subsidence rate, and H is the height of the seamount. All these parameters are derived from our data. The zero-age depth (d_i) is a theoretical number not directly comparable to the axial depth, which for a slow-spreading ridge has a deep valley created by tectonic processes within the spreading zone. Excluding tectonism, it is useful for calculating thermal subsidence alone.

The second method uses the present depth to the top to estimate the time of the Logi Ridge submergence (equation (3)):

$$(\sqrt{t_s} - \sqrt{t_s - t})S = D; t = t_s - \left(\sqrt{t_s} - \frac{D}{S} \right)^2, \quad (3)$$

where D is the depth of the seamount top, and the other parameters as in equation (2).

For the first method, the seamount height (H) has to be corrected for the sediment loading of the surrounding seafloor up to the time when the Logi Ridge was submerged (Figure 3). The effect of the loading is to increase seafloor subsidence resulting in a seamount height above what would be expected from the seafloor subsidence with age alone, and thus bias the erosion time closer to the present. For the second method, the seamount top depth (D) must be corrected for the extra subsidence due to the sedimentation after the Logi Ridge was submerged, which would otherwise bias the erosion time to look older (Figure 3). The sediment loading corrections apply equation (1).

4. Data Analysis

Figures 4 and 5 show the seismic reflection profiles presented from west to east over the Logi Ridge. Figure 6 shows main horizons depth converted, while Figures 7, 8, and 9 show details from the seismic lines. An interpretation summary map is shown in Figure 10.

4.1. Logi Ridge Morphology

The WLR is covered by Profiles 3, 4, and 5 (Figure 1). Profile 2 crosses the ELR and the north flank of the WLR. Both Profile 6 and NPD0005 cross directly over the top of the ELR, while NPD0004 crosses the eastern edge of the ELR (Figure 1). Profile 1 lies east of the Logi Ridge. The white circle in Figure 10 indicates a seamount not seen in Profile 2 and NPD0004, proving it to be an artifact in the ship track based bathymetric map (Jakobsson et al., 2012). We also suspect that the local depression seen in the seafloor east of the Logi Ridge similarly is an artifact.

The WLR is an approximately 110 km long, steep-sided ridge for the most part standing more than 1,500 m above the surrounding seafloor (Figures 1 and 10). The top of the WLR is characterized by a summit plateau at a depth of approximately 600 m and is approximately 25–30 km wide at the base (Figure 4). The top is reasonably flat and is widest in the west at Profile 3, where it is about 5 km, but narrower at Profiles 4 (2 km) and 5 (1.5 km) in the east. The depth-converted interpretation shows that the western WLR rise ~2.6 km over the surrounding basement on the south side and ~3.5 km on the north side, while the eastern WLR rise symmetrically about 3.5 km (Figure 6). The slopes of the WLR are more or less smooth with dips of approximately 7° to 11°.

The ELR lies slightly to the south of the WLR trend and is a single, steep-sided seamount standing about 1,500 m above the surrounding seafloor, rising to 400 m below the sea surface. The base is about 30–35 km wide, while the top is flat and up to 10 km wide across the central part (NPD0005). The ELR rises approximately ~3.6 km above the surrounding basement, and the dip of the slopes is about 13° to 20° (Figure 6). There is an almost circular elevation of the seafloor surrounding the eastern part of the Logi Ridge, extending ~60 km to the east of the ELR (Figures 1 and 10).

Figure 4. Single-channel reflection seismic data from west to east (top to bottom) over the WLR. Line locations are shown on bathymetry map (top right). Blue lines outline Horizon A; the boundary between the upper and lower sedimentary unit. Green lines show the acoustic basement. An elongated mounded drift (contourite) is identified along the northern flank of the WLR. Flat tops indicate erosion at sea level. The oceanic ages at the top of each profile are from Gaina et al. (2017). Tie points between the lines are indicated on top of each seismic panel where relevant. The locations of Figures 7, 8, and 9 are indicated by dashed black boxes. Different color bars at the bottom indicate different basement morphologies.

Figure 5. Single-channel (Profiles 1 and 6) and multichannel (NPD0004 and NPD0005) reflection seismic data from west to east (top to bottom) over the ELR area. Line locations are shown on bathymetry map (top right). Black lines indicate basement faults penetrating the sedimentary sections. Blue lines outline Horizon A; the boundary between the upper and lower sedimentary unit. Green lines show the acoustic basement. The contourite is also identified along the northern flank of the ELR. The deformation of the upper unit (recent faults and a growing seamount) becomes increasingly intense next to the south side of the ELR. Flat top indicates ELR erosion at sea level. The oceanic ages at the top of each profile are from Gaina et al. (2017). Tie points between the lines are indicated on top of each seismic panel where relevant. The locations of Figures 7, 8, and 9 are indicated by dashed black boxes. Different color bars at the bottom indicate different basement morphologies.

4.2. Surrounding Basement Topography

Acoustic oceanic basement is a high-amplitude reflection in some places, while in other places it is discontinuous with an irregular pattern below (Figures 4 and 5). Based on the reflection character, the top acoustic basement has been divided into three different categories: rough, smooth, and irregular structure. The morphological categories used by Elliott and Parson (2008) and others (e.g., Geissler et al., 2017) were developed for a specific environment. The setting is different from that of the Logi Ridge in that much of the magmatism is tied to magma-rich seafloor spreading next to a subsiding continental margin. Therefore, their classification scheme is not adopted entirely, since it would imply processes that are different from what must have shaped the areas close to the Logi Ridge. Our rough basement correspond to their use to describe fault-controlled oceanic crust, abyssal hills. However, smooth basement has characteristics of both

Figure 6. Depth conversion of the seafloor (red line), main sequence boundary (Horizon A) (blue line), and top acoustic basement (green line). Left: profiles from west to east (top to bottom) around the WLR. Right: profiles from west to east (top to bottom) around the ELR. Line locations are shown on bathymetry maps. The black double-arrows show the seamount heights, measured from the seamount top to the average basement depth of the surrounding seafloor little affected by basalt flows or uplift (dotted black line).

their “smooth” and the “rubbly” basement. The origin may be similar to what they describe for the rubbly basement, hyaloclastites, and lava flows, but it does not represent original oceanic crust. Our irregular basement resembles their “transitional” but is more chaotic. It must also have a different origin, affected by later lava flows and possibly intrusions. Our categories are aimed at best describing the differences in morphology created by secondary intrusive and extrusive magmatism in a deep oceanic environment as they appear in our study.

The rough basement type is characterized by abyssal hills and basins (Figure 7). The typical height of the hills is up to 1 s TWT (~ 1 km), and the width of abyssal basins (the distance from one hill top to the next) is 25 to 30 km. These shapes and dimensions are consistent with the abyssal hills typical for the central and NE Atlantic ocean (e.g., Escartin et al., 2008). However, some of these bear evidence of rejuvenated movement caused by igneous intrusions or fault movements (marked rough basement [reactivated] in Figures 4 and 5).

The smooth basement is a sheet-like body having parallel to subparallel high-amplitude reflections (Figure 8). The reflectors are variable in continuity and can be traced from a few kilometers, up to 20–30 km. This type of basement is well developed along the southern flank of the WLR on Profiles 3 and 4, and at the southeastern flank of the ELR on Profile 2 and NPD0004 (Figure 8).

In contrast to the other types of basement, the irregular basement is not easily traced. It was identified along the northern flank of the entire Logi Ridge. Some areas can resemble abyssal hills (Figures 9a and 9d), while others have small-scale irregular, high amplitude, and chaotic reflectors (Figure 9b). High-amplitude reflections can sometimes be seen below the acoustic basement (Figure 9d).

The smooth basement is interpreted as strong submarine basalt flows (e.g., Abdelmalak et al., 2016; Planke et al., 2000) that overprinted the original rough basement as well as earlier sediments. The thickness of the reflective layer is 0.15 s (Figure 8b) to 0.3 s TWT (Figure 8a). Using an average velocity of 4 km/s for the basalt flow (Blischke et al., 2016), the thickness varies between 300 and 600 m. The irregular basement is probably affected by smaller basalt flows and intra-basement intrusions. A map summary of the distribution of the different basement types identified is shown in Figure 10.

Figure 7. Seismic sections highlighting the key characteristics of the rough basement and the overlying sedimentary pattern. Blue line outlines Horizon A; the boundary between the upper and lower sedimentary units. (a) A normal fault offsetting sedimentary strata indicates recent tectonic activity east of ELR. (b) Recently growing abyssal hill south of the main Logi Ridge, and possible basalt flows close to the WLR. Locations are indicated in Figures 4 and 5. MTD = mass transport deposit.

4.3. Interpretation of Sedimentation Processes

The sediments in the study area can be divided into two distinct units (lower and upper) (Figures 4 and 5). The lower unit is semitransparent with indistinct reflections and rests on oceanic crust, while the upper unit is characterized by a subparallel, cyclic pattern of alternating low- and high-amplitude reflections. Around the ELR, the boundary between the upper and lower sedimentary units (blue line in Figures 4 and 5) is marked by a high-amplitude and laterally continuous seismic reflection named Horizon A. However, Horizon A is indistinct close to the WLR (Figures 4 and 5).

4.3.1. Lower Sedimentary Unit

At the base of the lower unit, early sedimentation filled in rough basement topography. It has few coherent internal reflectors, indicating that changes in acoustic impedance and lithology are small. However, some high-amplitude continuous reflections were traced from the Logi Ridge to the nearby areas (Figures 7, 8, and 9). Mass transport off the ridge is expected to consist of both lava flows and erosion products. Both will deposit volcanic material in the sedimentary section, will give strong reflections due to the impedance contrast to the pelagic sediments, and may not always be easy to distinguish from each other. The strong

Figure 8. Seismic sections highlighting the key characteristics of the smooth basement and its overlying sedimentary pattern. Blue line indicates Horizon A; the boundary between the upper and lower sedimentary units. (a) Smooth acoustic basement along the southern flank of the WLR is covered by pelagic sediments and some debris flows. The black line represents a fault related to sediment collapse. (b) Smooth basement close to the ELR is covered by pelagic sediments and debris flows. Some possible intra-basement intrusions are identified below the smooth acoustic basement. Locations are indicated in Figures 4 and 5.

and continuous reflection packages in the lower part of the section are most likely dominated by lava flows deposited on early pelagic sediments due to their apparent volume and continuity with the ridge flanks. Above these in the stratigraphy, there are sheets of wavy to continuous high-amplitude subparallel reflections with an internal disrupted character (Figures 4 and 5). The change in character is also related to a reduced volume. While there still may be lava flows in these deposits, the patchy appearance away from the ridge suggests that these may not be continuous flows. Most likely, these represent gravity driven mass transport deposits (MTDs). MTDs could be related to the submarine landslides (e.g., debris flow) and eroded material produced during the seamount subaerial exposure. Disrupted reflections are observed at the foot of the WLR (Figure 8a), apparently MTDs from the WLR. At the ELR, the upper part of the lower sedimentary unit is dominated by some continuous high-amplitude reflections, increasingly disrupted close to the ELR (Figures 9c and 9d), interpreted as MTDs from that part of the Logi Ridge.

Figure 9. Seismic sections highlighting the key characteristics of the irregular basement and its overlying sedimentary pattern. (a) Irregular basement with a weak basalt flow above, and a possible volcaniclastic deposit at the foot of the WLR just below Horizon A. (b) Basement showing as discontinuous and chaotic reflectors with a weak basalt flow above. (c) Irregular basement with a possible basalt flow above. (d) Irregular basement with a strong and continuous basalt flow above. Debris flows are well developed around the ELR in the lower sedimentary unit. Contourites in the upper sedimentary unit in Figures 9a, 9b, and 9d are characterized by well stratified reflectors but also contain faults and chaotic reflectors. The contourite has several crests on Profile 6 as seen in Figure 9c. Line locations are indicated in Figures 4 and 5. MTD = mass transport deposit.

4.3.2. Upper Sedimentary Unit

Along the northern flank of the entire Logi Ridge, the upper sedimentary unit is characterized by a mounded sediment pattern. The bathymetry has a small moat next to the Logi Ridge, with the main peak 5–8 km away, before it gently slopes down to the north. Mostly, it appears to be a single structure, except on Profile 6, where it has two distinct crests (Figures 4, 5, and 9). It has the characteristics of a contourite deposit, with internal erosion and changes in the sedimentation pattern, and often an uphill migration of the crest (Rebesco et al., 2014). It must be formed by the East Greenland Current flowing south along the west side of the Greenland Basin, turning southeast north of the WJMFZ (Rudels et al., 2012) (Figure 1). The base of the drift corresponds to Horizon A (Figure 9d).

Close to the WJMFZ, the lower unit is thin or missing, and the upper sedimentary unit may onlap oceanic basement topography. There are no indications of volcanic or erosional material transport from the Logi Ridge in the upper unit. On the south side of the WLR, the sedimentary layers onlap the basement high and dip away from it (Figure 8a). South of the eastern WLR and the ELR, there is quite a bit of deformation over and around local basement highs, often also disrupting the seafloor. The deformation becomes increasingly intense next to the south side of the ELR (Figure 5). We outline an area affected by this disturbance south and east of the ELR (Figure 10, yellow, dashed line). One fault cutting through the entire sediment package to offset the seafloor by ~250 m shows recent movement more than 60 km east of the ELR (Figure 7a). Maximum uplift is at the fault, and seafloor subsides northwards over a distance of 50 km to a depth similar to that south of the fault. In between this profile and ELR, Profile 2 (0–25 km in Figure 4) shows minor uplift that only affected the lower parts of the upper sedimentary unit, showing somewhat older activity locally. The sedimentary pattern in these disturbed areas does not show any of the diagnostic features for contourites. Stratigraphic uplift is tied to basement highs, and unlike the contourites on the north side of the

Figure 10. The bathymetry from the IBCAO v.3 ship track compilation of the study area (Jakobsson et al., 2012). The seismic reflection profiles used for this study are shown by different line colors indicating different basement morphologies. The white dashed lines show the magnetic chron locations derived from Gaina et al. (2017). The yellow dashed line indicates the lateral extent of the recent faults and uplift close to the ELR. White and yellow solid lines show the position of the published OBS profiles AWI-20030500 (Voss & Jokar, 2007) and JM-06 Line1 (Kandilarov et al., 2012), respectively. The white circle shows a seamount not seen in Profile 2 and NPD0004, proving it an artifact in the bathymetric map. JMP = Jan Mayen Platform; NKR = Northern Kolbeinsey Ridge.

Logi Ridge, the seafloor uplift is exclusively tied to these highs. The stratigraphic disturbance also affects the lower and upper units to similar degree, which is most apparent around the seafloor-penetrating knoll south of the ELR (Figure 5). We therefore interpret this uplift and deformation to be caused by young magmatic intrusions in the underlying oceanic crust. Thus, some areas with rough basement topography appears to be reactivated in fairly recent times.

4.4. Sediment Thickness and Deposition Rates

In order to estimate regional deposition rates, different parts of the seismic profiles were tied to the seafloor age based on the magnetic grid interpretation from Gaina et al. (2017). For the lower sedimentary unit, we also need to exclude the parts of the profiles that are affected by the Logi Ridge development. That includes both the smooth and irregular basement connected to the flanks of the ridge. Also, some parts of the rough basement displaying signs of young uplift were excluded (Figures 4 and 5). For the lower sedimentary unit, sediment thickness over the rough basement generally varies between 0 and 0.5 km but can be up to 1.5 km in the basement lows (Figure 11a). The smooth basement adjacent to the south side of the Logi Ridge only has a thin layer (<0.4 km) of this sequence on top, consistent with the suggestion that older sediments are most likely present under the younger lava flows that now represent the acoustic basement. Over the irregular basement, there is no clear trend between the sediment thickness and age (Figure 11a), and these parts appear to be similarly affected. Over the rough basement, the early sedimentation preferentially fills in the bathymetric lows between the abyssal hills. The scatter of the data points is a natural consequence of the sedimentation process, and the average should be representative for the region as a whole. A linear regres-

Figure 11. The correlation of sediment thickness and basement depth with sea-floor ages derived from Gaina et al. (2017). Sampling interval is 2 km. Different colors indicate different types of basement. (a, b) Comparison of the sediment thickness of lower and upper sedimentary units with sea-floor age. Blue solid line shows the line fit for the lower unit on rough basement not reactivated, while blue dashed line represents average upper unit thickness of 0.7 km. The estimated minimum age of Horizon A is shown by a dashed red line at 12 Ma. (c) Observed basement depth along profiles against the square root of sea-floor age. Using only the undisturbed rough basement depth, the red solid line represents the best fit age-depth relationship ($d = 1.65 + 0.30 \sqrt{t}$) based on an average sediment density of $1,950 \text{ kg/m}^3$ for sediment loading corrections, while the red dash and dotted lines show the effect of varying the density by $\pm 100 \text{ kg/m}^3$.

sion between the oceanic age and lower sediment thickness for rough basement not affected by secondary igneous intrusions gives a deposition rate of 4.3 cm/ka (blue line in Figure 11b). Using the LINEST program for Excel (Morrison, 2014), the standard deviation is 0.5 cm/ka.

We can correlate our seismic profiles to seismic reflection profile AWI20030550 (Figures 1 and 2) (Berger & Jokat, 2008), where our lower and upper units corresponds to their units GB-1 and GB-2. Using a regional grid of seismic reflection lines, they tied the unit boundary age to ODP site 913 north of our study area (Figure 1). The boundary age is dated to within Middle Miocene based on an assemblage of Miocene diatoms and radiolarians (Myhre et al., 1995). However, a precise age of the unit boundary was not given. Berger and Jokat (2008) adopted an age of 15 Ma, but that is likely too old. Profile 1 shows that Horizon A could not be traced to oceanic basement younger than 13 Ma (Figures 5 and 11b), where the uncertainty of the seafloor age is unlikely to be larger than 1 Ma. Furthermore, the regression results from the lower sedimentary unit (blue line in Figure 11a) shows a zero-thickness age of $12.6 \pm 0.66 \text{ Ma}$, in good agreement with the seismic tie result. The upper unit thickness remains almost unchanged ($0.7 \pm 0.05 \text{ km}$) over oceanic age between 25 and 43 Ma but is slightly thicker (0.9 to 0.7 km) from 12 to 25 Ma next to the WJMFZ. Considering uncertainties of the seafloor oceanic ages and regression results, we suggest that the age of the boundary is 12–14 Ma. Using this age range and a thickness of the upper sedimentary unit of $0.7 \pm 0.05 \text{ km}$ gives an average deposition rate of $5.4 \pm 0.8 \text{ cm/ka}$.

Table 1*Summary of the Deposition Age of the MTDs and Basalt Flows, Erosion Age from Seamount Height and Seamount Top Depths, and Horizon A*

	Western WLR	Eastern WLR	ELR
Seafloor age	41 Ma	37 Ma	34 Ma
Basalt flows top to the base of the upper unit	0.29 km (deposition age: 18–22 Ma)	0.18 km (deposition age: 16–19 Ma)	0.57 km (deposition age: 24–29 Ma)
MTDs top to the base of the upper unit	<0.06 km (deposition age: 12–15 Ma)	<0.06 km (deposition age: 12–15 Ma)	<0.06 km (deposition age: 12–15 Ma)
Last erosion age derived from seamount height	17 Ma	13 Ma	7 Ma
Last erosion age derived from seamount top depths	16 Ma	13 Ma	6 Ma
Age of Horizon A (ridge submergence)	12–14 Ma	12–14 Ma	12–14 Ma

4.5. Logi Ridge Formation Age from Related Mass Transport Deposits

The distribution of basalt flows and MTDs around the Logi Ridge records its evolution. Using a sedimentation rate of 4.3 ± 0.5 cm/ka and the top of the lower unit at 12–14 Ma, we can estimate the deposition age of the MTDs and basalt flows. The average thickness between the basalt flows and the top of the lower unit is 0.29 km for the eastern WLR and 0.57 km for the ELR, giving ages of 18–22 and 24–29 Ma, respectively. The basalt flows on the south side of western WLR are placed higher in the stratigraphy. The average thickness between these basalt flows to the top of the lower unit is 0.18 km, giving an age of 16–19 Ma. This appears to come from the last major eruptions in this area. The seamount morphology may have been significantly affected by this episode, as the western WLR next to the flows is wider than the eastern WLR. Horizon A has high amplitude in the east and may be tied to a last minor eruption in the eastern part of the Logi Ridge. The average thickness between the MTDs and the top of the lower unit is less than 0.06 km, indicating a deposition age of 12–15 Ma. Within the coverage of seismic lines, no mass transport off the Logi Ridge was detected in the upper sedimentary section, indicating that the formation and erosion of the Logi Ridge was over by 12–14 Ma. Thus, a general conclusion from the sedimentary record is that the main igneous construction took place from the Oligocene into the Early Miocene, and including predominantly Middle Miocene erosion, the Logi Ridge appears to have developed over a time span of at least 15 Ma (Table 1).

4.6. Logi Ridge Formation Age from Subsidence

A number of parameters for the subsidence-based methods need to be determined from the seismic data (equations (2) and (3)). First of all, we need to determine how the seafloor is subsiding in our study area. The correlation between sediment load corrected basement depth (excluding reactivated rough basement) and oceanic age shows that the depth of the spreading ridge at zero age is 1.65 km, and the seafloor subsidence rate is $300 \text{ m/Ma}^{1/2}$ (Figure 11c). The subsidence rate is the same as derived from the surrounding Greenland Basin (Figure 1) (Tan et al., 2018) and comparable to results from other locations ($320 \text{ m/Ma}^{1/2}$ Korenaga, 2008; $325 \pm 20 \text{ m/Ma}^{1/2}$, Crosby & McKenzie, 2009). The standard deviation of subsidence rate and zero-age depth is $40 \text{ m/Ma}^{1/2}$ and 0.2 km using Morrison (2014). However, the larger data set from the Greenland Basin from Tan et al. (2018) gives the same subsidence rate but a standard deviation of $\pm 15 \text{ m/Ma}^{1/2}$. In addition, an uncertainty of 100 kg/m^3 in the average sediment density will affect zero-age seafloor depth by 0.05 km (Figure 11c).

Both eastern and western parts of the Logi Ridge have flat tops, indicating that it was eroded at sea level (e.g., Caplan-Auerbach et al., 2000). We can then calculate the age of last subaerial exposure (t) by using equation (2) for seamount height. Three flat-topped parts formed on 41 Ma (western WLR), 37 Ma (eastern WLR), and 34 Ma (ELR) old oceanic crust were used in the calculations. The seamount height (H) is determined by the difference between the average basement depth of the surrounding seafloor and depth to the Logi Ridge top (Figure 6). The height of the western and eastern WLR is the same at 3.5 km, while ELR is slightly higher at 3.6 km. The seamount height (H) is corrected for the sediment load of the surrounding seafloor up to the time when the Logi Ridge was submerged (equation (1)), where the sediment thickness (0.8 km) is determined by the average thickness between the basement and Horizon A, and using a sedi-

ment density of $2,100 \text{ kg/m}^3$ (Voss & Jokat, 2007). The corrected heights of the western and eastern WLR are the same at 3.11 km, and 3.21 km for the ELR, which give the ages of the last subaerial exposure of the western and eastern WLR of 17 and 13 Ma, and 7 Ma for the ELR.

Alternatively, we can estimate the last erosion age based on the depth of the seamount top (equation (3)) (Figure 3). The seamount top depths (D) of the western and eastern WLR are 0.66 and 0.60 km, and 0.40 km for the ELR. The top depth (D) is corrected for the extra subsidence due to the sedimentation after the Logi Ridge was submerged (Figure 3), using a sediment thickness of the upper unit of 0.7 km and an average sediment density of $1,800 \text{ kg/m}^3$ (Voss & Jokat, 2007). After correction, the seamount top depths of the western and eastern WLR are 0.41 and 0.35 km, and 0.15 km for the ELR. That gives the last erosion age of the western and eastern WLR of 16 and 13 Ma, and 6 Ma for ELR. We note that both methods give similar results and show WLR submergence roughly in the Middle Miocene in agreement with the sedimentary record but that the ages for the ELR are considerably younger. There are a number of uncertainties to the methods that make them less precise for dating than the sediment record. However, these results together with an analysis of the associated uncertainties will be used to further discuss the development of dynamic topography of the area in the next section.

4.7. Dynamic Topography Development

Previous studies show that there is considerable dynamic uplift in our study area (Breivik et al., 2008; Hoggard et al., 2017; Rickers et al., 2013). Our zero-age basement depth of $1.65 \pm 0.2 \text{ km}$ (Figure 11c) is not necessarily the depth at the time of seafloor emplacement but will include the result of developing dynamic topography up to the present. We have no means to constrain how it changed before the time of Logi Ridge erosion at sea level. However, the development of the dynamic topography since the time of Logi Ridge submergence can be estimated by comparing the erosion age from the sedimentary record (Table 1), with the subsidence-based ages. By using the range of 12–15 Ma for the sediment constrained seamount erosion age, the dating uncertainty is incorporated into the analysis. If there is no change in dynamic uplift since last subaerial erosion time of the Logi Ridge, the erosion age should be consistent between the different methods used. On the other hand, if there is a change in dynamic topography, there will be a discrepancy.

Initial results indicate a reasonable agreement for the WLR (13–17 Ma) but not for the ELR (6–7 Ma). In order to estimate what that means in terms of dynamic topography development, we do a series of calculations over a range of zero-age depths (d_i) for the seamount height method, varied from 1.45 to 2.5 km (equation (2)). The latter is average zero-age depth of the West Atlantic (Crosby & McKenzie, 2009). A new zero-age depth is estimated by varying the parameter until the erosion ages derived from seamount height and sediments are reasonably matched (Figure 12). The amount of the post-submergence dynamic topography change can be determined from the difference between this zero-age depth and the present estimate of 1.65 km. Also, the present Logi Ridge top depth analysis is sensitive to a change in dynamic topography (equation (3)) and is similarly tested for dynamic topography change between -0.2 and 0.85 km (corresponding to the zero-age depths from 1.45 to 2.5 km).

Erosion ages derived from seamount height and from seamount top depth have similar trends, and both methods give the western WLR the oldest erosion age, followed by the eastern WLR, and with ELR having the youngest age (Figure 12). However, judging from the sedimentary record, there is no obvious erosion age difference along the Logi Ridge (Table 1). For a given erosion age, both methods show that there is an approximately 0.3 km difference between the ELR and the average of the WLR curves (Figures 12a and 12d). The best fit for the zero-age depth for the ELR is approximately 1.8–1.9 km (Figure 12a). It is slightly deeper than the 1.65 km zero-age depth, indicating a dynamic topography uplift of 0.15–0.25 km after the submergence of the ELR. For erosion age derived from seamount top depth, a reasonable fit (Figure 12d) indicates a dynamic topography growth of between 0.2 and 0.3 km since the last erosion for the ELR. For the WLR, both the erosion age derived from seamount height and seamount top depth is generally consistent with the sedimentary record (Figure 12), suggesting that there is little dynamic topography change since the Middle Miocene.

4.7.1. Sensitivity Analysis of Dynamic Topography Estimates

We tested the sensitivity of the dynamic uplift estimates to uncertainty in the parameters (Figures 13 and 14). There are two groups of parameters affecting the subsidence analysis. The first group will affect WLR and ELR profiles similarly, comprising sediment load correction, eustatic sea level changes, seafloor subsidence rate, zero-age depth, and same-sign seamount height and seafloor age variations (Figure 13), used to

Figure 12. Estimating dynamic topography development based on equations (2) and (3) with seamount height and seamount top depth corrected for sediment loading (Figure 3). The blue box outlines the estimated Middle Miocene submergence from the sedimentary record, while the red boxes (dotted line: WLR; solid line: ELR) indicate how that can be interpreted in relation to dynamic uplift change. The red double-arrow shows the dynamic topography difference between the WLR and ELR profiles. Red solid line in (a–c) shows the present-day estimate on zero-age depth (Figure 11c). (a) Erosion age calculated from seamount height depends on the zero-age depth for the theoretical subsidence curve. (b) Sensitivity of erosion age to increased sea level of 50 m. (c) Sensitivity to reducing the age difference from west to east by changing the age of seafloor under WLR and ELR by -1 and $+1$ Ma, respectively. (d) By comparing the observed depth to the top of the Logi Ridge with the predicted subsidence of the seafloor, the development of dynamic topography after submergence can be estimated. (e) Sensitivity test of erosion age to sea level increase 50 m. (f) Sensitivity to reducing the age difference from west to east by changing the age of seafloor under WLR and ELR by -1 and $+1$ Ma, respectively.

investigate absolute dynamic topography changes. The second group can affect the relative dynamic topography development estimates between WLR and ELR profiles, consisting of opposite sign seamount height, seafloor ages, and seamount top depth variations (Figure 14).

The two subsidence-based dating methods have different sensitivity to parameter changes (Figure 13). Estimates based on top seamount depth are independent of zero-age depth and seamount height. The methods further have opposite response to changes in eustatic sea level, sediment load corrections, and subsidence rate, and the best values should minimize the difference between the methods for both parts of the Logi Ridge. Sediment load corrections depend on the estimated density as well as the thickness, both derived from sedimentary velocity. We combine these into same-sign variations of $\pm 100 \text{ kg/m}^3$ and $\pm 0.1 \text{ km}$. Sea-level changes of up to 50 m are tested (Conrad, 2013). Seafloor age uncertainty is estimated to $\pm 1 \text{ Ma}$, which corresponds to $\pm 8 \text{ km}$ with the spreading rate of Mohn's Ridge. The subsidence rate standard deviation of $\pm 15 \text{ m/Ma}^{1/2}$ from Tan et al. (2018) was used to test this parameter. For zero-age depths, we use the standard deviation of $\pm 0.2 \text{ km}$. Uncertainty for seamount height is estimated to $\pm 0.2 \text{ km}$, due to the uncertainty in estimating the surrounding basement depth (approximately $\pm 0.20 \text{ TWT}$) (Figure 6).

For eustatic sea level change, results quickly start to deviate between the methods (Figure 13). Good overlap is obtained with no change, although a small sea level rise is possible without affecting results significantly.

Figure 13. Sensitivity test of various parameters of group 1 affecting WLR and ELR in the same direction when calculating dynamic topography changes. A positive value indicates there is dynamic uplift, while a negative value represents post-submergence dynamic topography subsidence. “Method difference” and “Difference between WLR and ELR” are calculated based on average value of dynamic topography change of WLR and ELR. The best fit between the methods occurs at values close to those of our preferred model as indicated by red boxes.

Results are not very sensitive to same-sign seafloor age changes. Sediment load corrections have opposite sensitivity, and values at or near our preferred model give good method agreement. The seamount height method estimates are more sensitive to subsidence rate uncertainty than the top depth method. The methods overlap at or slightly below our preferred value. The seamount height method is most sensitive to the parameters of seamount height and zero-age depth. The best fit to the top-depth method is for our preferred values in both cases. It is clear that the top-depth method is least sensitive to parameter changes, but the comparison with the seamount height method helps to constrain the results.

Figure 14. Sensitivity test for parameters of group 2, which affect the relative difference between the estimated dynamic topography development for WLR and ELR. A positive value indicates there is dynamic uplift, while a negative value represents post-submergence dynamic topography subsidence. “Method difference” and “difference between WLR and ELR” are calculated based on average value of dynamic topography change of WLR and ELR. The best fit between the methods occurs at values close to those of our preferred model as indicated by red box.

The second group of parameters can affect the difference in dynamic topography change between the WLR and the ELR. We run the calculations through opposite-sign variations in seamount height, seafloor age, and seamount top depth in order to estimate how robust the difference between ELR and WLR is. For seamount height and top depth variations, a good match between methods is found at or near our preferred model. Opposite-sign seafloor age variations affect both methods and have the greatest effect, where a 2 Ma reduction in age difference reduce the east-west difference by about 0.05 km (Figures 12 and 14).

In the summary of the results of the sensitivity tests, the value of the western Logi Ridge is the average of western and eastern WLR profiles. A positive value indicates dynamic uplift, while a negative value represents post-submergence dynamic topography subsidence. Assuming a method difference of less than 0.1 km, the sensitivity tests show that the change in dynamic topography for the WLR profiles since the Middle Miocene must be small, limited to -0.1 to $+0.05$ km (Figures 13 and 14). On the other hand, a reasonable estimate for the ELR indicates an uplift of 0.15 – 0.3 km since the Middle Miocene (Figures 12 and 13). The difference between the WLR and ELR profiles can only be reduced by 0.05 km within the parameter uncertainty (Figure 14). Thus, we conclude that within a reasonable uncertainty, there is a small post-Middle Miocene uplift of the ELR, while the WLR has little dynamic topography change since the Middle Miocene.

5. Discussion

5.1. Sedimentation Development Around the Logi Ridge

There is an approximately 25% increase in the average sedimentation rate from the lower to the upper sedimentary unit. Several studies argue that the high sedimentation rate at high latitudes is related to regional glacial activity and ice transport of sediments into the deep ocean (e.g., Barron & Whitman, 1981; Van Weering et al., 2008). The oldest age of glaciation in the North Atlantic is between Middle Eocene and Early Oligocene (Tripathi et al., 2008), and Greenland glaciations have taken place since 18 Ma (Thiede et al., 2011). Continuous high-amplitude reflections of the upper unit require large variation in acoustic impedance, which depends on velocity and density. It is therefore likely that they represent layers of relatively coarser sediment deposited from drifting ice. Increased sedimentation rate could also have contributions from ice-sheet erosion during glacial periods in the northern most part of the Jan Mayen microcontinent (JMMC) (Blischke et al., 2019). Ice erosion does not appear to affect the Logi Ridge, since we do not observe erosion products from it in the upper sedimentary unit.

The intensification of water exchange between the Arctic and the North Atlantic Ocean also corresponds to increasing sedimentation rate (Winkler et al., 2002). The widening of the Fram Strait facilitate transport of large amounts of sediment-laden sea ice mainly from the Eurasian continental shelves across the Fram Strait to the Northern Atlantic Ocean floor (Vinje & Finnekåsa, 1986; Wollenburg, 1993). Thus, the transport of current-related polar ocean sea ice could contribute to the increased sedimentation rate of the upper sedimentary unit.

The time of the final opening of the Fram Strait has been a subject of debate and is dependent on how it is defined. The formation of a narrow deep oceanic corridor occurred during Late Oligocene (~ 21 Ma) (Jokat et al., 2016) to Early Miocene times (20–15 Ma) (Engen et al., 2008). This is consistent with the onset of a ventilated circulation regime by 18.2 Ma in the Arctic Ocean (Jakobsson et al., 2007). However, a major change of clay mineral distribution and increased sediment accumulation rates at 11.2 Ma (Winkler et al., 2002) support a delayed effect of the widening gateway. The combination of seismic and aeromagnetic data shows that the strait had developed as a deep marine passage around Middle Miocene (14–16 Ma) (Jokat et al., 2008a, 2008b). Contourite drifts are observed north of the Fram Strait from at least 11 Ma (Mattingsdal et al., 2014). The first well developed magnetic spreading anomalies occur around 9.8 Ma (Engen et al., 2008), marking final continental separation. In Figure 15, a main opening stage of 11–16 Ma is indicated.

The base of the contourite is at the boundary between the upper and lower sedimentary unit, implying that the east Greenland current was well developed by Middle Miocene (12–14 Ma). This is comparable to an intra-Miocene unconformity (12 ± 1 Ma) at the base of a contourite along the western flanks of the Jan Mayen Ridge (Blischke et al., 2019), probably a response to the same change in oceanic currents that we see in our area. This timing is consistent with the submergence of the Greenland-Scotland Ridge (GSR) and formation of deep water north of Iceland by 12 Ma (Poore et al., 2006). It further agrees with the suggestions that the Fram Strait had become a well developed deep water passage by Middle Miocene (Jakobsson et al., 2007; Jokat et al., 2008a, 2008b).

Figure 15. Simplified chronostratigraphic summary (Cohen et al., 2013) with main magmatic buildup, dynamic uplift, seamount erosion, and contourite development for WLR, ELR, and south and southeast of ELR. Solid lines indicate the main duration of each event, while dashed lines indicate greater uncertainties and/or reduced activity. Gray lines for dynamic topography indicate lesser temporal constraints than for magmatism and erosion. The times for the breakup between Greenland and the Jan Mayen microcontinent (JMMC), as well as the Fram Strait opening and Traill Ø intrusions, are indicated by the horizontal gray bands. The red box shows the Horizon A dating with associated uncertainty.

5.2. Logi Ridge Origin and Dynamic Topography Development

The WLR is a fairly linear and narrow ridge. A number of mechanisms have been proposed to explain the origin of ridge-shaped seamounts. One hypothesis is that a volcanic ridge can result from lithospheric cracking. Lithospheric cracking could potentially be caused by a preexisting fracture zone or by thermal contraction of the plate (Sandwell & Fialko, 2004). Either decompression melting of mantle and/or preexisting asthenospheric melt could then rise along existing or developing zones of weakness (Batiza, 2001).

The model of Hieronymus and Bercovici (2000) involves lithospheric cracking caused by deviatoric stress in the lithosphere along the spreading axis, enhanced by the loading of the seamount itself, accessing melt in the asthenosphere caused by compositional and/or thermal anomalies. It predicts a formation age propagation from old seafloor into young seafloor (west to east for our study), perpendicular to the spreading ridge. Our age estimates do not show obvious age progression, and the Logi Ridge is oriented oblique to the spreading ridge, which is inconsistent with the model. The $\sim 40^\circ$ angle of the Logi Ridge to the plate motion flow lines also precludes that the ridge can be tied to fracture zone development as proposed by Gernigon et al. (2009) (Figure 2). Since the Logi Ridge is on the same trend as the TIC on the shelf, and the radiometric dating of the intrusion tied to the magnetic anomaly is at the Eocene/Oligocene boundary (Figure 15),

about the same time as the ridge building apparently started, a common cause is possible. However, magmatic activity occurred both in continental and oceanic lithosphere, and a process that can work in both domains is then needed.

The oceanic lithosphere cracking hypothesis is expected to be generally applicable if significant, and testable in other regions of the world. A study from the Pacific where there are a number of large ridges was aimed at such a test (Forsyth et al., 2006). The study conclude that the lack of extensional features in the seafloor and the wide distribution of magmatism are inconsistent with a lithospheric cracking model, where fissures control magma distribution (e.g., Sandwell & Fialko, 2004). They also point out that a passive lithospheric cracking model is not sufficient to produce the amount of magmatism observed through decompression melting, and instead they argue for an active process driven from below from the asthenosphere.

Our study suggests that dynamic topography in the WLR has not changed significantly since Middle Miocene submergence, while the ELR may have had approximately 0.15–0.3 km of additional uplift since then. The conclusion is based on the lack of material transport off the ELR above Horizon A, suggesting it was submerged during the Middle Miocene (Figure 15). Due to limited data coverage it cannot be precluded that there may be later transport that was not detected. If the submergence of the ELR was slightly later than the WLR, the post-Middle Miocene dynamic uplift would be less. However, the study area has average 0.65 km shallower bathymetry compared to the conjugate oceanic crust east of the Mohn's Ridge (Breivik et al., 2008), suggesting that much of the dynamic topography asymmetry had developed before the end of the Logi Ridge development. This conclusion would still be valid even if the ELR was submerged slightly later than the WLR. As there is a strong correlation between magmatism and dynamic topography worldwide (Hoggard et al., 2017), it seems that the formation of the Logi Ridge is associated with the development of the dynamic topography, implying that the Logi Ridge was formed by anomalous melting due to movement of buoyant asthenospheric material, which can affect both continental and oceanic parts. The linear orientation could then be determined first of all by the distribution of compositional and/or thermal anomalies in the asthenosphere, though cracking as a response to initial seamount loading could have played a role. Our study shows that the onset of formation of the Logi Ridge is coeval with the rifting of the Jan Mayen micro-continent off from Greenland to the south during the Late Oligocene jump of seafloor spreading to the Kolbeinsey Ridge (Figure 15). A regional igneous event resulted in flood basalts (referred as F-marker) observed in the Jan Mayen Basin and JMMC, tied to breakup of the JMMC from the central east Greenland (Blischke et al., 2016, 2019). This development has been tied to the migration of the Iceland plume to underneath the east Greenland coast, weakening the lithosphere there (Mjelde et al., 2008), coeval with the growing Iceland plume regional influence during the Oligocene (35–25 Ma) (Parnell-Turner et al., 2014). The Traill Ø intrusions were emplaced at the start of this period up to 34 Ma. Since the seafloor age under the ELR is 34 Ma, and the ridge buildup was not on-axis, the Traill Ø event may precede the Logi Ridge buildup by a few Ma (Figure 15). Most likely, both dynamic topography development and magmatism in our study area are tied to the increasing Iceland plume influence.

The continued dynamic topography rise under the easternmost part of the Logi Ridge correlates with post-Middle Miocene uplift and intrusions observed south and southeast of the ridge (Figures 10 and 15). The shallow bathymetry could potentially be caused by a shallow hot asthenospheric flow from the Iceland plume along the Kolbeinsey Ridge and past the WJMFZ (e.g., Breivik et al., 2008; Hoggard et al., 2017). However, a recent 3D density modeling study shows that the shallowest part of the present-day Iceland plume flow is turning away from the northern tip of the Kolbeinsey Ridge and eastward to the southern tip of the Mohn's Ridge (Tan et al., 2018). Thus, the shallow bathymetry around the Logi Ridge appears to be tied to a deeper asthenospheric flow.

6. Summary and Conclusions

Eight seismic reflection lines across the Logi Ridge were used to constrain the ridge formation and the dynamic topography development north of the WJMFZ. The western part of the ridge (WLR) is linear and continuous, while the eastern part (ELR) consists of a singular seamount of similar height and width. Both the WLR and ELR have flat tops, indicating that these have been eroded at the sea surface.

Three different oceanic basement characters are observed along the profiles. A rough basement resembles the original abyssal hills. However, some parts have been uplifted later by magmatic intrusions and fault activity south to southeast of the ELR. A smooth basement appears to be formed by basalt flows overprinting

early sediments on the south side of the Logi Ridge. Irregular basement corresponds to smaller basalt flows and intra-basement intrusions, seen on the north side of the ridge, and at a smaller area to the south side of the ELR. The sedimentary package has two distinct units, in parts separated by a strong reflector: Horizon A. Both by correlating the lower unit thickness with seafloor age as well as tying to ODP well 913 north of our study area, we suggest the age of Horizon A is 12–14 Ma. The average lower-unit deposition rate is 4.3 ± 0.5 cm/ka. The upper sedimentary unit has more or less uniform thickness and a deposition rate of 5.4 ± 0.8 cm/ka. Along the northern flank of the Logi Ridge, the upper sedimentary unit is dominated by a contourite deposit, which is formed by the east Greenland deep water current coming from the north, related to the opening of the Fram Strait.

The formation of the igneous Logi Ridge is dated by determining how mass transport from the ridge interacts with the surrounding sedimentation over time. Mass transport off the Logi Ridge consists of both lava flows and erosion products. Lava flows appear episodically throughout the Late Oligocene and into the Early Miocene. From the late Early Miocene to the Middle Miocene, erosional transport off the Logi Ridge dominated, and the last erosion of the Logi Ridge is in the Middle Miocene, stopping when the ridge was submerged. Deformation of the sediments up to the seafloor near the ELR, record minor recent magmatic activity to the south and southeast of the main structure. By comparing the end of erosion time estimated from the sedimentary record with those from dating methods based on subsidence using either seamount height or current top seamount depth, we estimate to what extent dynamic topography has changed since Middle Miocene submergence. It is not possible to see any significant change for the WLR within the existing uncertainty, though it appears that the ELR has 0.15–0.3 km uplift relative to the WLR since then. Since there is a ~0.65 km bathymetric asymmetry over the Mohn's Ridge (Breivik et al., 2008), being shallow in the Logi Ridge area, much of the dynamic topography had to be present by the time the Logi Ridge development ceased. This suggests a causative relationship between dynamic topography development and the magmatism, as observed worldwide (Hoggard et al., 2017). The early Logi Ridge development coincides with that of the rifting of the Jan Mayen microcontinent off east Greenland. This has been tied to the Iceland plume migrating eastwards from under Greenland towards the NE Atlantic, and the Logi Ridge development may be linked to that event by flow of asthenospheric material.

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