**Crustal and Thermal Heterogeneities Across the Fram Strait and the Svalbard Margin**

M.-A. Dumais\(^1\)\(^2\), L. Gernigon\(^2\), O. Olesen\(^2\), A. Lim\(^1\), S. E. Johansen\(^1\), and M. Brönner\(^2\)

\(^{1}\)Department of Geoscience and Petroleum, Norwegian University of Science and Technology, Trondheim, Norway.

\(^{2}\)Geological Survey of Norway, Trondheim, Norway

**Abstract** The lithospheric structure of the Fram Strait and the extent from the Knipovich Ridge to the Barents Sea shelf and Svalbard are poorly understood. Several multi-geophysical investigations from various campaigns since the 90s along the Western Barents Sea margin and the Northeast Greenland margin resulted in insufficient and contradicting interpretations of the crustal and upper mantle settings in the oceanic and continental domains. New airborne magnetic data across the Knipovich Ridge and west of Svalbard provide new insights, reveal the complexity of the seafloor spreading history of the Arctic Atlantic Ocean, and indicate a European-Eurasian continent-ocean boundary located ~150 km farther west than previously suggested. This new location of the continent-ocean boundary prompted to revise the existing 2-D seismic interpretations in terms of crustal domains and tectono-stratigraphic setting. This is tested using joint 2-D gravity and magnetic field modeling to derive an improved crust-mantle model of the study. One recently acquired combined 2-D controlled source electromagnetic/magneto-telluric (CSEM/MT) profile across the Mohns Ridge was also modeled with potential field data and provided new insights into the tectonic settings of the crust and the mantle thermal anomalies. This study proposes to unify the various seismic and CSEM/MT interpretations using the new aeromagnetic compilation.

**Plain Language Summary** The opening of the Fram Strait between Svalbard and Greenland is still poorly understood. Studying the seafloor spreading of the Knipovich Ridge is central to understand the development history of this key area located between the Northeast Atlantic and Arctic oceans. New high-resolution magnetic data flown above the sea reveal the complexity of the seafloor spreading history of the area. The extent of the seafloor spreading is smaller than previously suggested. The new airborne magnetic data are modeled and tested to derive an improved crust-mantle model to better understand the development of this key area in the High Arctic.

**1. Introduction**

As a link between the Atlantic and Arctic spreading systems, the Fram Strait and Svalbard are key regions to understand the geological development of the entire High Arctic. Tectonic issues such as the extent and timing of the Eurekan orogeny (Piepjohn et al., 2016) and the spreading development of the Knipovich Ridge remain problematic and questionable. The timing of the rifting, the Knipovich Ridge initiation, and the location of the continent-ocean boundary (COB) have been debated for decades leading to various interpretations and proposed locations (Breivik et al., 1999; Dumais, Gernigon, et al., 2020; Engen et al., 2006, 2008; Faleide et al., 1991; Franke et al., 2019; Gernigon et al., 2019; Libak, Mjelde, et al., 2012; Lundin & Doré, 2011; Mosar et al., 2002; Scott, 2000; Seton et al., 2012; Voss & Jokat, 2007), see review by Eagles et al. (2015). Compared to other areas in the North Atlantic, little geological, geochemical, and geophysical data have been available along the study area, due to the remoteness of the Fram Strait region. Aeromagnetic data were acquired four decades ago above the Knipovich Ridge, with a large gap remaining at the Svalbard and the Western Barents Sea margin (Olesen et al., 2010) (Figure 1b, Table 1), hampering an adequate crustal characterization of the area to further constrain the interpretation along existing regional 2-D seismic profiles. Various crustal domains have been previously derived from regional gravity data and a few 2-D refraction seismic data, for example, Breivik et al. (1999, 2003, 2005), Czuba et al. (2005), and Libak, Mjelde, et al. (2012). Ocean bottom seismic (OBS) data have been acquired in the Fram Strait across and along the Knipovich Ridge, for example, Jokat et al. (2012), and through the continental domain of the Barents Sea and the Svalbard Platform (Table 2). These data offer a constraint on the crustal geometry and expected lithology as seismic data are sensitive to the acoustic impedance contrast between the various lithologies and the associated densities. Controlled source
electromagnetic/magneto-telluric (CSEM/MT) data have also assessed the crustal thickness, crustal resistivity, and estimated the temperature of the mantle across the Knipovich Ridge (Johansen et al., 2019).

This study investigates the structural settings of the crust in the Fram Strait and its lithological heterogeneities from the oceanic domain at the Knipovich Ridge to the continental domain on the Svalbard and Western Barents Sea margins. Recently acquired by the Geological Survey of Norway (NGU), the Knipovich Ridge aeromagnetic survey 2016 (KRAS-16) (Figure 1) in the Fram Strait area have been jointly analyzed together with available regional gravity data, seismic, and CSEM/MT. It allows us to initiate an integrated interpretation of the study area to obtain an improved crust-mantle model of this key area in the Arctic. The magnetic field variations derived from the aeromagnetic data provide indications about the crustal, geophysical, and thermal properties of the area, for example, Curie point depth estimation (Ebbing et al., 2009). The aeromagnetic survey estimates the location of the COB and the timing of the continental break-up, crucial to understand the tectonic development of the area (Dumaïs, Gernigon, et al., 2020).

2. Geological Background

The study area encompasses the Fram Strait, Svalbard, Edgeøya Platform, and Nordaustlandet (Figure 1).

Onshore, bedrock fragments of Precambrian to Cenozoic age (Dallmann, 2015) are exposed on the Svalbard Archipelago. Basement deformations from the Caledonian orogeny (Holtdahl, 1926; Ohta, 1994), have been subsequently influenced by the rifting and late seafloor spreading in the Fram Strait (Skilbrei, 1992). During the Cenozoic, the Eurekan Orogeny might have merged with the West Spitsbergen Thrust and Fold Belt that developed along the west coast of Spitsbergen (Harland, 1969; Piepjohn et al., 2016; Vanvaka et al., 2019).

The Edgeøya Platform to the east (Figure 1) is mainly characterized by sub-horizontal layers of Triassic sediment successions (Dallmann, 2015). Farther east, Nordaustlandet, an island of the Svalbard archipelago, is mostly covered by glaciers. With few outcrop availabilities, the regions lithology is difficult to assess. Two basement types determined by basement outcrops from both sides of Wahlenbergfjorden, partly dividing Nordaustlandet (Dallmann, 2015; Johansson et al., 2002), and confirmed and refined from aeromagnetic data (Dumaïs & Brönner, 2020). The basement on the north shore is of pre-Caledonian origin with Mesoproterozoic and Neoproterozoic rock exposures (Lauritzen & Ohta, 1984). Caledonian granites, Grenvillian Ripijorden granites and migmatites are found on the northern tip of Nordaustlandet (Johansson et al., 2002, 2005), as well as Silurian diorites and gabbro located on the northeast of Nordaustlandet (Johansson et al., 2005). The south shore comprises a Tonian basement, composed of dolomite, sandstone, quartzite, and limestone, intruded by Jurassic-Cretaceous doleritic dikes. Cretaceous sills have also been emplaced on- and offshore Nordaustlandet (Dumaïs & Brönner, 2020; Grogan et al., 2000; Minakov et al., 2012; Polteau et al., 2016). Offshore, the Fram Strait includes the Knipovich Ridge as the physiographic expression of the active seafloor spreading. Earlier studies set the onset of the seafloor spreading in the Fram Strait between the Svalbard margin and Greenland around ~55 and 29 Ma (Ehlers & Jokat, 2009; Faleide et al., 2008; Nemčok et al., 2016; Talwani & Eldholm, 1977), while the more recent aeromagnetic data (Dumaïs, Gernigon, et al., 2020; Dumaïs, Olesen, et al., 2020) suggest a seafloor spreading initiation at c. 20 Ma, that is, after the Eurekan orogeny (Dumaïs, Gernigon, et al., 2020). Classified as an ultrastable and oblique spreading system (with seafloor spreading rates of less than 20 mm/yr), the Knipovich Ridge is a segment of the NE Atlantic-Arctic Mid-Ocean Ridge system along with the Mohns Ridge, the Molloy Transform Zone, the Lena Trough, and the Gakkel Ridge (Figure 1). At the present day, the Knipovich Ridge trends from NW-SE in the south to N-S in the north. The Knipovich Ridge is surrounded by the Vestbakken Volcanic Province, and the Hornsund Fault Zone on the eastern side, and by the Boreas and East Greenland Basins on the western side (Figure 1). On the west side, the Svalbard margin is largely covered with thick wedges of sedimentary rocks along the Svalbard and Western Barents Sea margin (Engen et al., 2006; Klitzke et al., 2015). The Fram Strait oceanic domain developed after a Late Cretaceous-Eocene rifting event between Norway and the Northeast Greenland and forms a complex system of conjugate shear margins characterized by distinct crustal, structural and magmatic properties (Faleide et al., 2008; Hamann et al., 2005; Ritzmann & Jokat, 2003; Srivastava & Roest, 1999). Pull-apart basins formed locally in the Southwest Barents Sea, for example (Breivik et al., 1998; Engen et al., 2008; Faleide et al., 1993). During the Paleocene-Eocene, the continental rifted system underwent a brief episode of compression-transpression associated with the Eurekan-Spitsbergen fold and thrust belts (Gac et al., 2020; Piepjohn et al., 2016). Northward, the Knipovich Ridge is linked through
the Molloy Transform Zone to the Gakkel Ridge located in the High Arctic region, north of Svalbard (Brozena et al., 2003; Glebovsky et al., 2006). As part of the Fram Strait region, the Hovgaard Ridge and the East Greenland Ridge, along the Greenland Fracture Zone, represent characteristic bathymetric features (Figure 1a) and may include several continental fragments preserved within the oceanic domain (Døssing & Funck, 2012; Døssing et al., 2008; Engen et al., 2008; Faleide et al., 2008; Knies & Gaina, 2008; Nemčok et al., 2016).

In the Norwegian-Greenland Sea, the Mohns and Ægir ridges spreading initiated at 52.8 Ma (C24r) (Gaina et al., 2009; Talwani & Eldholm, 1977) and propagated progressively to the south toward the juvenile volcanic margins during the Early Eocene (Franke et al., 2019; Gernigon et al., 2019). After the extinction of the

Table 1
Description and Reference of the Aeromagnetic Data Sets Merged

| Survey | Line spacing (m) | Year        | Acquired/compiled by               | References                  |
|--------|------------------|-------------|------------------------------------|-----------------------------|
| 1      | 3,000–7,500      | 1993–2011   | Alfred Wegener Institute           | Jokat et al. (2008, 2016)   |
| 2      | 4,000            | 2008–2009   | TGS-NOPEC Geophysical Company      | Trulvik et al. (2011)       |
| 3      | 4,000–10,000     | 1969–1991   | Geological Survey of Norway        | Olesen et al. (2010)        |
| 4      | 5,500            | 2016–2018   | Geological Survey of Norway        | Dumais, Gernigon, et al. (2020) and Dumais, Olesen, et al. (2020) |

Note. The acquisition parameters are valid for the section of the data set used, for example, Olesen et al. (2010) is partially used. The location of the data sets is displayed in Figure 2.
Mid-Labrador Ridge (Labrador Basin Bay) in the Early Oligocene (33.7 Ma, C13) (Oakley & Chalmers, 2012; Roest & Srivastava, 1989; Srivastava & Roest, 1999; Suckro et al., 2013), the relative motion between Norway and Greenland changed from NNW-SSE to WNW-ESE (31–28 Ma, C12-10) (Gaina et al., 2017). From this reorganization, the ultraslow spreading Ægir Ridge became extinct after C10, causing the development of the Kolbeinsey Ridge and the detachment of Jan Mayen Microplate Complex from Greenland at ∼24 Ma (C7-6) (Blischke et al., 2017, 2022; Gernigon et al., 2015, 2019; Schiffer et al., 2019). To the north, the Knipovich Ridge initiated at c. 20 Ma (C6) (Dumais, Gernigon, et al., 2020). While earlier a failed rifting hypothesis in the Boreas Basin was disputed (Hermann & Jokat, 2013; Skogseid et al., 2000), Dumais, Gernigon, et al. (2020) interpreted the existence of an extinct rift situated at the westernmost part of the Boreas Basin close to the Northeast Greenland margin to explain the peculiar strong asymmetry of the spreading system. In their model, a ridge jump in the Fram Strait is proposed at around 18 Ma. Farther north, the Gakkel Ridge was initiated at ∼58–59 Ma (C26n-25r) followed by a seafloor spreading rate decrease from C13 (Brozena et al., 2003; Glebovsky et al., 2006; Schreider et al., 2019). A 250-km section of the Gakkel Ridge, north of Svalbard, ending in the Fram Strait, opened much later between C8 and C5 (Brozena et al., 2003; Glebovsky et al., 2006; Jokat et al., 2016). Similarly, the Molloy Ridge spreading segment initiated in the Early Miocene (10–20 Ma, C5-C6) (Engen et al., 2008; Srivastava & Tapscott, 1986).

3. Methods and Data

To further resolve the complexity of the crustal setting and achieve a consistent interpretation with various observations, we integrated existing results from various geophysical and petrophysical data and tested geological concepts into the modeling. Magnetic and gravity 2-D forward models are used to investigate, test, and improve the understanding of the crustal architecture of the study area. The gravity field is sensitive to the mass density variation affected by for example, crustal and mantle composition, isostasy, crustal flexure, and thermal variation in the lithosphere. All those mechanisms and properties are peculiar in the oceanic domain, not in isostatic equilibrium, where the mantle and crustal densities are controlled by the magma influx associated with the seafloor spreading and mantle decompression at the ridge. The magnetic field is sensitive to iron oxide content and distribution in the crust down to the Curie temperature isotherm when minerals lose their magnetization.
3.1. Aeromagnetic Data

With the acquisition of the Knipovich Ridge Aeromagnetic Survey 2016 (KRAS-16), a comprehensive magnetic data set from different campaigns, covering the Fram Strait was compiled to a continuous magnetic grid (Figure 2, Dumais, Gernigon, et al., 2020; Dumais, Olesen, et al., 2020). Details of the acquisition and processing of KRAS-16 are described in the report from Dumais, Olesen, et al. (2020). In the Fram Strait between the coast of Greenland and the Barents Seas and Svalbard margins, the original aeromagnetic data were flown with 7.5–20 km line spacing at 300 m above the sea-level in 1972 and 1973 as described in Olesen et al. (2010). Improving the coverage and the resolution of the area, the KRAS-16 survey was flown at 120 m above the sea-level with 5,500 m line spacing and 30° orientation, perpendicular to the spreading. A compilation of surveys with various resolutions and acquired over the last four decades (Jokat et al., 2008, 2016; Olesen et al., 2010; Trulsvik et al., 2011) was carefully merged to minimize the discrepancies due to different resolution caused by the line spacing (Table 1), as well as navigation, positioning, and sensor technologies. The Oasis Montaj Gridknit algorithm (Geosoft, 2013) was applied to calculate a minimal shift with the overlaps between the surveys and applied it to the grids to merge the data. Long wavelengths, with a filter cut-off at 300 km, were corrected with the magnetic field model MF7 (Maus et al., 2008) effective to map long wavelengths related to the Earth's crust magnetization. However, noise in the data remains in the surveys adjacent to KRAS-16, for example, the linear trends observed parallel to the Molloy Transform Zone are artifacts from poor leveling of the flight lines.

3.2. Gravity Data

The gravity data (Figure 3) used for latitudes below 80°N is the global gravity model Sandwell v23-1, which has a grid cell size of 1 arc-minute and an accuracy of 2 mGal with a high correlation between seafloor topography and gravity anomalies in the 12–160-km wavelength band where the sediment cover is thin (Sandwell et al., 2014). Above 80°N, the gravity data rely on the Arctic Gravity Project (ArcGP) data compilation which has 5 arc-minute resolution (Kenyon et al., 2008). The free-air anomaly data have been used for the 2-D forward calculation. The Bouguer correction could not be reliably calculated given the difference in accuracy among the coarse free-air gravity data, the high-resolution bathymetry, and the small extent of the ridge crest.

3.3. Bathymetric Data

The bathymetric data in the rift valley of the Knipovich Ridge were acquired by the Geological Institute of the Russian Academy of Sciences and the Norwegian Petroleum Directorate with a multibeam acoustic sonar between 2006 and 2010, resulting in a digital topography model with 100 m cell size (Zarayskaya, 2017). This data set is used to correlate the bathymetric highs with the magnetic anomalies (Figure 5). The regional bathymetric data outside the rift valley is a compilation with 250 m cell size from various surveys (Olesen et al., 2010, Figure 1). This regional data set is mainly used to constrain the 2D-forward models.

3.4. Seismic Data

Published seismic profiles from different deep seismic experiments were applied as structural constrains and horizon interpretation (Table 2). Profile locations are shown on Figure 1. All profiles have been interpreted for the Moho, top crustal basement, and sedimentary layers in accordance with the seismic P-wave velocities.

3.5. 2-D CSEM and MT Profile

Applied CSEM and MT profiles are based on Johansen et al. (2019) and Lim (2020). A detailed description of the data acquisition and processing can be found in Johansen et al. (2019). In their studies, Johansen et al. (2019) determined the depth of an electrical Moho. From these interpretations, the Moho and the top crustal basement have been derived from the variation in the resistivity, temperature, lithology, porosity, permeability, fluid content, and melt content was also taken into account (Johansen et al., 2019; Ni et al., 2011). Thus, the crustal interfaces interpreted from the CSEM and MT data should coincide with the seismic Moho and top crustal basement but might slightly differ as they are sensitive to different physical properties.
Figure 2.
3.6. Seismicity Data

The 3-D magnetization model was compared to the earthquakes events recorded for 11 months along 160 km of the Knipovich Ridge (75.5°–77.5°N) by the KNIPAS (Knipovich Ridge Passive Seismic Experiment) seismological network consisting of 26 broadband OBS (Meier et al., 2020). The OBS clock drift was determined using noise cross-correlation (Hannemann et al., 2014) and their position on the seafloor were determined from seismic profiles (Meier et al., 2021). The earthquakes detection was done with Lassie (Heimann et al., 2017) while the automatic picking of P and S phases was performed with PSPicker (Baillard et al., 2013) but manual re-picking was required with SEISAN (Havskov & Ottemoller, 1999; Ottemoller et al., 2011). Automag in SEISAN pick automatically the amplitudes and HYPOSAT (Schweitzer, 2001, 2018) estimated the location of the earthquakes. The depth and the magnitude of the earthquakes along the rift valley of the Knipovich Ridge were plotted against the 3-D magnetization to study the correlation between the two data sets.

However, the earthquake events recorded at lower latitude along the Knipovich Ridge were not used since their location is believed erroneous and required re-processing (Loviknes et al., 2020).

3.7. Curie Point Depth Estimation

Derived from the magnetic data, the Curie point depth (Figure 4a) was estimated using the Pycurious Python code (Mather & Delhaye, 2019) including the Bouligand et al. (2009) algorithm. The Curie point depth is the depth of the isotherm of the Curie temperature of magnetite (580°C). Above this temperature, the magnetite loses its magnetization. Therefore, the Curie point depth represents the boundary between a given magnetization and non-magnetization independently of the magnetite content. It should coincide with the bottom of the deepest causal magnetic source, which is often expected close to the crust-mantle boundary. The method determines the bottom of the deepest source of the magnetic anomaly which is interpreted as the Curie point depth. This study uses the result of Curie point depth estimation as the bottom depth of the deepest magnetic source and calls it Curie point depth for simplicity. The distribution of the depth of the magnetic sources is determined by the analysis of the power spectrum of the magnetic anomalies in the Fourier domain (Spector and Grant, 1970). The premise is that the magnetic data set covers a large enough area to contain the long wavelengths necessary to resolve the Curie point depth. Bouligand et al. (2009) algorithm is calculated over a series of windows until the data set is entirely analyzed. The Curie point depth estimation was applied to the entire aeromagnetic compilation including the high-resolution KRAS-16 data set and the adjacent aeromagnetic data sets. Given the expected values of the Curie point depth estimate across the whole data set, we used a window size of 20 km and a varying fractal parameter. The accuracy of the Curie point depth estimation from the magnetic data in the oceanic domain could be biased since the magnetization acquired during the seafloor spreading is not entirely random in all directions as expected by the methodology. However, given the observations from the magnetic data and the modeling of the spreading anomalies (Dumais, Gernigon, et al., 2020), the results appear reasonable to differentiate oceanic and continental domains qualitatively. The inversion creates a linear discrepancy in the Mohrs Ridge area, few kilometers south of KRAS-16 survey. This coincides with coarse resolution in the magnetic data compilation probably caused by insufficient overlap with the adjacent survey to the south or a profile filtered for noise reduction (Figures 1 and 2). Since the Curie point depth calculation is sensitive to the frequency content of the magnetic data to derive the depth, shallow depths as expected in the oceanic domain cannot be resolved for this linear section (Figure 4a). The adjacent survey flown in the Greenland and Boreas basins, west of the KRAS-16 survey, also caused small discrepancies in the Curie point depth with deeper values (Figure 4a) likely due to leveling issues of this adjacent survey.

Figure 2. Aeromagnetic anomaly map of the study area. Magnetic striped patterns represent the oceanic domain. Round intermediate-size anomalies are observed in the Vestbakken Volcanic Province. The linear trends observed parallel to the Molloy Transform Zone are artifacts from the flying configuration. Small round high amplitude anomalies observed east of Nordaustlandet are Cretaceous sills. (COB derived from aeromagnetic data [Dumais, Gernigon, et al., 2020]: blue and white dashed line, KRAS-16 boundary: red, isochron anomalies: dotted black, 2-D profiles P: black, lineaments mapped from the magnetic data: white, ridge axis: dotted white, MoR, Molloy Ridge; VVP, Vestbakken Volcanic Province; BFZ, Billefjorden Fault Zone; SFZ, Senja Fracture Zone; HFZ, Hornsund Fault Complex Zone; GFZ, Greenland Fracture Zone; MTZ, Molloy Transform Zone; GB, Greenland Basin; BB, Boreas Basin; HR, Hovgaard Ridge; KnR, Knipovich Ridge; and MR, Mohns Ridge). The aeromagnetic survey boundaries used for the compilation in the figure inset are described in Table 1 (1- Jokat et al. 2008, 2016) which overlap with survey 3; 2- Trulsvik et al. (2011); 3- Olesen et al. (2010) which overlap with survey 1 and 4; and 4- Dumais, Gernigon, et al. (2020).
Figure 3.
3.8. Werner Deconvolution

The Werner deconvolution (Ku & Sharp, 1983; Phillips, 1997; Werner, 1955), an automated depth-to-source estimation method, was derived from the magnetic data. The depth and morphology of the magnetic top basement and the presence of intrusions are estimated by using empirical basement indicators sensitive to susceptibility variations, and approximating the geological source to a simplified geometry of features such as contacts and dikes (Goussev & Peirce, 2010). The Werner deconvolution solutions represent the depth to the top of the causal sources where specific clusters are observed (Figure 4b). Seequent Oasis Montaj software was used to calculate the Werner Deconvolution (Geosoft, 2006). A minimum and maximum window lengths of 4 and 12 km, respectively, and a window expansion increment of 500 m and window shift increment of 500 m were input as parameters. Solutions depths were constrained between 1 and 20 km from the mean sea-level. The resolution of the depth solutions depends on the resolution and accuracy of the magnetic data, but also on the profile direction. 2-D Werner deconvolution provides most accurate results when calculated perpendicular to the trend of a magnetic anomaly. Given the magnetic anomalies from the seafloor spreading are generally oriented perpendicular to the flight lines, the high-resolution KRAS-16 survey allowed reliable depth solutions, in contrary to the aeromagnetic data in the adjacent areas, covered with larger line-spacing and often filtered for noise reduction. Therefore, our depth estimation along the seismic profiles provided poor results due to the lower resolution of the areas adjacent to KRAS-16.

3.9. Modeling

A 3-D magnetization model of the recent aeromagnetic data set was calculated using an inversion method implemented in GM-SYS-3D, where a model is defined by stacked surface layers, each with a specified density and magnetization surface distribution (Geosoft, 2014; Parker & Huestis, 1974). The model was constrained by a

Figure 3. Free-air gravity anomaly map of the area. The magnetic isochrons coincide with the high free-air anomalies along the Knipovich Ridge. (COB derived from aeromagnetic data [Dumais, Gernigon, et al., 2020]: blue and white dashed line, KRAS-16 boundary: black, magnetic isochron anomalies: dotted black, 2-D profiles P: black, magnetic lineaments mapped from the magnetic data: white, ridge axis: dotted white, MoR, Molloy Ridge; BFZ, Billefjorden Fault Zone; HFZ, Hornsund Fault Zone; GFZ, Greenland Fracture Zone; MTZ, Molloy Transform Zone; GB, Greenland Basin; BB, Boreas Basin; HR, Hovgaard Ridge; KnR, Knipovich Ridge; and MR, Mohns Ridge).

Figure 4. (a) Curie point depth representing the bottom of the causal magnetic source calculated from the aeromagnetic compilation. The shallowest depth area agrees with the COB depicted from the magnetic data where the oceanic domain is inferred. The gray hatched zone is the location where the Curie point depth is assumed erroneous due to poorer data quality. (b) Werner deconvolution solutions derived from the aeromagnetic data representing the top of the causal magnetic source. The magnetic compilation is shown in shades of gray. Shallower depth solutions are observed in the oceanic domain. (KnR, Knipovich Ridge; MR, Mohns Ridge; and COB, Continent-ocean boundary).
Moho depth derived from seismic data (Funck et al., 2017), a sedimentary thickness (Engen et al., 2006), and bathymetric data (Olesen et al., 2010), where both mantle and sediment layers have a negligible magnetization. These data sets are chosen for their availability and coverage of the full area studied. The sediment thickness (Engen et al., 2006) is mainly derived from gravity and bathymetry data and calibrated for the age of the crust with magnetic isochron and for the base of the sediment layer and the Moho depth with seismic data. It provides a sufficient approximation of the location of the top of the basalt layer to provide reasonable magnetization amplitudes in the 3-D model. A lateral inversion was performed with the recommended maximum Parker-Huestis iterations and sharpening iterations of 2 and 6, respectively. A cosine roll-off filter was used with a lower high-cut limit of 4,000 m and upper high-cut limit of 2,858 m with a convergence limit of 10 nT. The reference field used was 54,125.4083 nT with an inclination of 81.10647042° and a declination of 0.02202437°. No constrains were applied to the crustal layer whereas the sediment and mantle layers were set to no magnetization. The magnetization derived from the data represents the overall magnetization of the crust without differentiation of the lithology of the upper crust, lower crust, and numerous intrusions. However, it provides preliminary insights of the type of magnetization, induced or remanent, expected for the 2-D forward models.

GM-SYS-2D has been used to produce the 2-D forward models of the studied profiles (Geosoft, 2006). The sediment and crustal layers in the models were constrained by the available seismic and EM data (Table 2). The same constants are applied to the calculated gravity and magnetic response for each model to match the observed data, an option in GM-SYS (Geosoft, 2006) allowing a common adjustment between the synthetic geometry of the models and the datum used for the observed data (Geosoft, 2013). The different interpreted layers of the

Figure 5. (a) Magnetization with volcanoes (black dots) and bathymetric highs (green outline) within the rift valley, white frames show the location of b and c and the purple frame shows the location of the KNIPAS earthquake events (Figure 6). The 100 m gridded bathymetric data (Zarayskaya, 2017) is shown in the window frames (b and c). (b) Example of high magnetization in the rift valley correlating with bathymetric highs and the presence of volcanoes. (c) Example of low magnetization in the rift valley correlating with the absence of volcanoes or bathymetric features.
crust and mantle consider similar values for density, susceptibility, and remanence to match the observed gravity and magnetic data. This ensures a certain homogeneity between the models. No rock samples, representative of the crustal lithology, were available in the Fram Strait. The initial densities were extracted from the published seismic and gravity modeling (Czuba et al., 2008; Grad & Majorowicz, 2020; Hermann & Jokat, 2013; Ljones et al., 2004; Ritzmann et al., 2002, 2004), while magnetization was taken from general publications for the oceanic domain (e.g., Dentith & Mudge, 2014; Tivey & Johnson, 1993) and regional models of the Barents Sea (Barrère et al., 2009, 2011; Marello et al., 2013). The densities were then modified to fit the gravity data and to be comparable amongst the interpreted profiles. The geometries from the seismic and CSEM/MT profiles are modified when necessary to obtain coherent and realistic density and susceptibility within the expected range for the sedimentary and crustal layers of the area. It also helps to solve the mismatch between seismic profiles often found at their intersections where the greatest modifications are found. Where profiles intersected, the horizons are modified up to 2 km to reduce the difference between the seismic interpretations. The susceptibility and remanence parameters were chosen to represent an average lithology variation between the layers from gabbro to basalts from the literature, Tivey and Johnson (1993) for the oceanic crust; Barrère et al. (2009, 2011) and Marello et al. (2013) for the continental crust. Magmatic intrusions were modeled to fit the magnetic signature. After the susceptibility and remanence magnetization were assigned, the crustal domains were delineated on each model based on the crustal thickness and magnetization. In addition, the Werner deconvolution provided a certain degree of constraints to map the top of the crustal basement and magmatic intrusions.

4. Results
4.1. Magnetic Map Description

Intermediate-size rounded anomalies of ~20–40 km-width with high amplitude above 100 nT are found in the Boreas Basin, the Svalbard margin, the Vestbakken Volcanic Province, and the Western Barents Sea Margin extending to Bjørnøya and Stappen High (Figure 2). These are associated with volcanic and magmatic activities, correlated to the presence of sills, dikes, and volcanic mounds emplaced in the continental crust (Faleide et al., 1988; Mork & Duncan, 1993; Omosanya et al., 2016). Similar intensity anomalies of 10- to 20-km width found east of Nordaustlandet and Edgeøya are also correlated to sills, dikes, and volcanic mounds emplaced in the continental crust (Faleide et al., 1988; Mork & Duncan, 1993; Omosanya et al., 2016). A 50–250 nT high-amplitude isolated linear anomalies are observed along known fracture zones such as Billefjorden and Hornsund fault zones. Prominent regional high-amplitude above 200 nT anomalies are found on the northeast coast of Greenland, on the Edgeøya Platform, on the Stappen High, and on the Loppa High. These anomalies are associated with thicker crust for the northeast coast of Greenland and Edgeøya Platform. The strong anomalies at the Stappen and Loppa highs have been explained by the presence of thick Precambrian magnetized basement (Fichler & Pastore, 2022; Gernigon et al., 2014; Marello et al., 2013). Striped linear magnetic anomalies located in the central part of the study area characterize and delineate the oceanic domain composed of basalts and gabbroic rocks. Lineaments and oceanic fracture zones perpendicular to the seafloor spreading, previously interpreted by Dissing et al. (2016) in the Boreas Basin, are also apparent in the new aeromagnetic data (Figure 2) and are reinterpreted on both side of the Knipovich Ridge (Dumais, Gernigon, et al., 2020).

The shallow Werner deconvolution solutions correlate with the oceanic domain interpreted from the magnetic data that is, shallow sources of less than 5 km depth are observed in the oceanic domain (Figure 4b). Also, in the oceanic domain, the Curie point depth estimation used as the bottom depth of the deepest magnetic source indicates shallow depths varying between 5 and 7 km, where higher temperatures in the lithosphere are expected from the magma supply feeding the spreading ridge (Figure 4a). The oceanic crust thickness is estimated to extend from less than 5 km (top) to 7 km (bottom) depth. The Curie point depth transitions to greater depths of 25 km in the continental domain, expressing a colder crust-mantle system and a thicker crust. A wide intermediate zone of 30–90 km between the shallow oceanic basement (with a Curie point depth estimate of 4–6 km) and the deep continental basement (with a Curie point depth estimate of 25–30 km) is observed along the west boundary of the KRAS-16 survey. This intermediate zone correlates with deeper Werner deconvolution solutions resolving deeper intrusions or a deeper top basement suggesting a thicker sedimentary cover (Figure 4b). Thus, the top of the magnetic sources is expected at larger depth than 7 km and the bottom layer at 12 km ±2 km depth. In the continental domain, shallower values of Curie point depth estimate of 5–15 km correlate with the presence
of sills east of Nordaustlandet and possible dikes or magmatic intrusions on the Edgeøya Platform (Dumais & Brönner, 2020; Grogan et al., 2000; Minakov et al., 2012; Polteau et al., 2016).

4.2. Gravity Map Description

Free-air gravity anomaly lows are correlated to smaller densities in sedimentary basins such as the Boreas Basin and the Greenland Basin (Figure 3). The Knipovich Ridge, the Mohns Ridge, and the Molloy Transform Zone are expressed as a narrow band associated with negative free-air gravity anomaly while the ridge flanks display free-air gravity high anomalies above 30 mGal with some degree of correlation with the seafloor topography. The magnetic isochron anomaly location correlate with the free-air gravity high anomalies found on the Knipovich Ridge flanks (Figure 3). The Hovgaard and East Greenland ridges appear as high free-air gravity high anomaly above 60 mGal. An important 100 km-wide free-air gravity high anomaly above 60 mGal is also present north-west of Bjørnøya, delimiting the Hornsund Fault Zone and investigated in the 2-D model analysis (Ch. 4.4).

4.3. 3-D Magnetization Inversion Results

At the southern tip of the Knipovich Ridge, at latitudes 73°–74°N, high-amplitude stripes of magnetization demark the continuation of the Mohns Ridge, alternating from −4 to +8 A m\(^{-1}\) associated with normal and reverse remanent magnetization (Figure 5a). Following the Knipovich Ridge rift valley northward, at latitudes 74°–76°N, the magnetization degrades from 0 to 4 A m\(^{-1}\) with rare occurrences of strong magnetization. A close analysis of the bathymetry within the rift valley draws a correlation between the presence of volcanoes and crater-shaped features, and bathymetric highs with high magnetization values above 5 A m\(^{-1}\) (Figure 5b). However, the physical extent and the intensity of the high magnetization layer may vary. Bathymetric highs and volcanoes generally occur in a rich magma supply setting. On the contrary, a low magma supply or an iron- and oxide-poor magma chamber, producing new oceanic crust poor in magnetite, would result in low magnetization (Dentith & Mudge, 2014). Crater-shaped features are rare to non-existent where the magnetization amplitude is very small, near a zero value (Figure 5c). Magmatic accretion along the spreading axis discontinued by segments of amagmatic accretion is common at oblique ultraslow spreading ridges (Dick et al., 2003). Seawater can percolate through fractures and change the thermal and chemical properties of the crust (Searle, 2013). Numerous lineaments and fracture zones are identified from the aeromagnetic data set, which are partly linked with linear features visible on the bathymetric data. Fluid circulation leads to oxidation and degradation of the well-oriented magnetic grains, possibly changing the strong remanent magnetization into weak induced magnetization (Kent & Gee, 1996). At latitudes of 76°–78°N, the central magnetic anomaly amplitude increases to reach values 5–8 A m\(^{-1}\) as expected from a normal oriented remanence. Outside the oceanic domain delimited by C6 (Figure 5), few strongly magnetized bodies are identified, but no notable linear highly magnetized bodies are observed. The lack of evidence of magnetic isochrons supports the absence of basalt layer. The intermediate rounded anomalies found east of C6 on the east flank could not be explained by neither magmatic nor amagmatic accretion. On the west flank, west of C6, no visible anomaly can support a conjugate domain from magmatic or amagmatic accretion. This observation supports the presence of a steady state oceanic crust formed by the seafloor spreading and delimited by isochrons C6 on side of the Knipovich Ridge.

The magnetization of the central magnetic anomaly C1 along the rift valley (Figure 5) and the segmentation of the seafloor spreading are compared to the seismicity data available at the northern latitude of the Knipovich Ridge (Meier et al., 2020, 2021) (Figure 6). High magnetization above 5 Am\(^{-1}\) at 76°30’N and 77°00’N correlate with shallow seismicity near the seafloor with magnitudes mainly below 2, while low magnetization near 0 are correlated with deep seismicity activity with occurrences of magnitudes above 2 (Figure 6a). Shallow earthquake events with depth less than 10 km with magnitudes below 2 observed between 76°30’N and 77°00’N are associated with a seafloor spreading segment well defined with the aeromagnetic data (Figure 6b). The magnetic signature of this segment suggests a magmatic accretion and supports the segmentation interpretation of Meier et al. (2021).

4.4. 2-D Forward Modeling

The 2-D forward models are constrained by the seismic interpreted horizons for each refraction profile available in the study area (white dotted lines in Figures 7–9). All the 2-D forward interpretation are homogenized for
modeling parameters and physical parameters to offer a unified model of the study area as shown in Table 3.

In the 2-D models, the steady state oceanic crust is characterized with a thin crust and an unambiguous striped magnetic pattern while the continental crust has a distinctive thick crust and Moho discontinuity at around 25 km depth. Between the oceanic and continental domain, a transition domain is delineated by a thickened crust compared to the oceanic domain, a deepening Moho discontinuity and the absence of an unambiguous striped magnetic pattern.

4.4.1. Profile P1 (Figure 7a)

This profile intersects the ridge and ends in the Barents Sea, south of Svalbard. Its central part coincides with a large free-air gravity anomaly which was used to estimate the COB location in earlier studies, for example, Breivik et al. (1999). The magnetic anomaly striped pattern delimit an interpreted COB (Dumais, Gernigon, et al., 2020) located about 160 km farther west on P1. This magnetic description of the COB is used for all models and data analysis of this current study. On P1, high frequencies on the magnetic signal are observed from 0 to ∼110 km which is defined as the oceanic domain. Between ∼110 and ∼270 km, the magnetic signal has a generally low background with two wide anomalies with low amplitude of 50 and 30 nT. From 300 km to the end of the profile, the magnetic signal amplitude is generally higher compared to the middle section identified as the transition zone with one prominent anomaly at ∼320 km. The central magnetic isochron C1 is vanishing, with no clear magnetic signal, and modeled with low remanent magnetization. Local intrusions or volcanic mounds east of C6 and in the Vestbakken Volcanic Province could explain the local high magnetic signature observed between the COB and the Western Barents Sea margin. Overall, the 200-km section between the COB and the Hornsund Fault Zone

---

Figure 6. (a) Profile showing the correlation between magnetization, the magnetic anomaly, the earthquake magnitudes, and earthquake depths along the rift valley of the Knipovich Ridge. Shallow earthquake events with a magnitude between 0 and 2 are observed at the highest intensity of the magnetic anomaly and magnetization. (b) Magnetization with seismicity (Meier et al., 2020, 2021). The seismic magnitude is illustrated by the size of the earthquake events and the hypocenter depth is color coded according to the color chart.
do not show clear striped magnetic pattern. This region is defined as a transition domain between the continental and the steady state oceanic crust domains. This transition zone coincides with the deepest top basement along Profile P1. A low susceptibility (0.005 SI), lower than the one assigned to the continental crust (0.012–0.038 SI) is needed to fit the observed magnetic data. One large intrusion is modeled at the necking zone between the continental domain and the transition where the Moho rises rapidly from 30 km to less than 15 km depth over a horizontal distance of less than 50 km. The Hornsund Fault Zone is also denoted with a shallow narrow elongated intrusion due to the high frequency of the related magnetic anomaly. The depth of the bottom of this causal body cannot be reliably estimated from the modeling, but the small response on the gravity signal suggests a small body in volume. East of the Hornsund Fault Zone, corresponding to a high magnetic anomaly of 200 nT, several intermediate magnetic anomalies from 110 to 130 nT are found on the Svalbard margin. These anomalies are modeled with several continental crustal layers superposed at an angle and with varying susceptibilities that are possibly related to old Caledonian nappes. There is no indication on the seismic data of a basement change in the lithology which is consistent with the small density variation between the crustal layers.

A good correlation between the observed and calculated gravity anomalies requires a variation in the crustal and mantle densities. The upper crust in the oceanic domain consists of basalt with densities of 2,700–2,750 kg m$^{-3}$ while the gabbro of the lower crust has a density of 2,900 kg m$^{-3}$ almost identical to the densities published for the ocean section of this seismic profile (Breivik et al., 2003). Densities in the transition domain delineated in the profile are 2,900 kg m$^{-3}$, within the uncertainty of the velocity-density relationship (Barton, 1986), and do
not support a clear delimitation between the upper and lower crust which is consistent with the velocity observed along the seismic transect (Breivik & Mjelde, 2001a, 2001b; Breivik et al., 2003). A 170 km-wide high gravity anomaly of 130 mGal is identified immediately west of the Hornsund Fault Zone covering half of the transition domain. No single lithological body causes this large anomaly but is caused by the crustal configuration and the lithology expected along P1. On the west side, the anomaly increases as the sediment thickness increases and the seafloor rises. The anomaly decreases abruptly with the sharp increase of the crustal thickness. The continental crust was modeled with densities of 2,900–2,990 kg m$^{-3}$ overlain by felsic-granite rocks with density of 2,600–2,750 kg m$^{-3}$ embedded by two sills imaged from the seismic data (Breivik & Mjelde, 2001b). The lower crust densities are higher than those derived from the seismic study (Breivik et al., 2005) but are within the uncertainty of the velocity-density relationship (Barton, 1986). The 2-D forward model presented in this study attempts to link both variation in susceptibility and density. A gradual density variation from a thermal mantle was attempted as suggested by Breivik et al. (2003) for the continental section of the profile but it could not explain the gravity for the full profile including both continental and oceanic section. The observed gravity did, however, require a density variation in the mantle. This was achieved with a lower density solely below the oceanic crust, suggesting that the variation of temperature or the composition in the mantle is local and does not require thermal variation over a long distance toward the transition or continental domain.

4.4.2. Profile P2 (Figure 7b)

Profile P2 intersects the Knipovich Ridge axis, the Svalbard margin, Svalbard mainland, and terminates on the Edgeöya Platform. The seismic data were acquired over several campaigns between 1998 and 2005 with seismic
The COB depicted from the magnetic striped anomalies correlates with the gradual lateral variation in the velocity of the seismic interpretation in the upper crust from about 4.5 to 5.0 km s\(^{-1}\) to 5.5–6.0 km s\(^{-1}\) (Breivik & Mjelde, 2001a). The seismic interpretation of Profile 6 intersecting at point P6 on Figure 7b observed a mantle-crust interface at 6 km depth (Hermann & Jokat, 2013) instead of 9 km (Breivik & Mjelde, 2001a; Czuba et al., 2008; Grad & Majorowicz, 2020; Ljones et al., 2004). Between 0 and ∼120 km, the present gravity model requires a thinner crust with a Moho discontinuity 2 km higher than the Moho discontinuity derived from the seismic interpretation for the oceanic domain comparable to the seismic interpretation of Hermann & Jokat (2013). The Moho discontinuity interpreted in the 2-D model follows roughly the top of the lower crust interpreted with a high density of 3,000 kg m\(^{-3}\) from the seismic data. The transition domain delimited between the COB and the Hornsund Fault Zone is demarked with a thickening of the lower crust and a deepening of the top basement between ∼160 and 250 km on the profile. Densities in the transition domain are 2,800–2,970 kg m\(^{-3}\) consistent with the seismic P-wave velocities for both the upper and lower crust (Breivik & Mjelde, 2001a, 2001b; Grad & Majorowicz, 2020). However, west of Hornsund Fault Zone, the seismic interpretations differs between 180 and 240 km (Breivik & Mjelde, 2001a, 2001b; Czuba et al., 2008; Grad & Majorowicz, 2020; Ljones et al., 2004) (Figure 7b). The free-air gravity anomaly requires a lower crust modeled in this study similar to Breivik and Mjelde (2001a) but with a more gradual thickening than the interpretation from Czuba et al. (2008) and Grad and Majorowicz (2020). Several magnetic anomalies are modeled with higher susceptibilities (0.02–0.06 S.I.) representative of dikes in the transition domain. The magnetic anomaly and free-air gravity anomaly highs located at the Hornsund Fault Zone are likely caused by the geometry of the crust. The mantle density also increases from the oceanic to the transition and continental domains. East of the Hornsund Fault Zone, the crustal properties change with higher densities and susceptibilities with a thicker crust (30 km), consistent to a continental crust. A density variation in the continental crust was also suggested by Breivik et al. (2005), the crustal 2-D forward model attempt to link both variation in susceptibility and density to the observed data. However, the densities modeled in the present study are higher than the seismic interpretations but within the uncertainty of the velocity-density relationship (Barton, 1986). Intrusions are modeled on the Edgeøya Platform where sills and dikes are expected (Dumais & Brönner, 2020; Grogan et al., 2000; Minakov et al., 2012; Polteau et al., 2016). However, the current resolution of the magnetic data does not allow us to resolve

![Figure 9. A 2-D profile P8 with the modeled and observed data for gravity and densities (kg m\(^{-3}\), blue), and for the magnetic, susceptibility (SI, red) and magnetization (A m\(^{-1}\); inclination \(^{\circ}\), black) are shown. The bathymetric horizon is derived from Olesen et al. (2010) and shows good correlation with the CSEM/MT horizons displayed as white dashed lines (Johansen et al., 2019; Lim, 2020). The magnetic isochrons resolved are identified on the profiles. (KnR, Knipovich Ridge; MR, Mohs Ridge; and VE, Vertical exaggeration).](image-url)
accurately the depth and volume of these magnetized bodies. Two high density bodies (2,930 and 3,090 kg m\(^{-3}\)) are modeled in the lower crust agreeing with the presence of high velocity bodies (6.5–7.5 km s\(^{-1}\)) interpreted with the seismic data which were correlated with the lithosphere-asthenosphere boundary (LAB) uplift (Grad & Majorowicz, 2020).

### 4.4.3. Profile P3 (Figure 7c)

This profile crosses the Knipovich Ridge axis to the Svalbard margin. The profile is located between two magmatic segments in the rift valley. Like Profile P2, the COB demarked by the striped magnetic anomaly correlates with the gradual lateral variation in the velocity of the upper crust (Breivik & Mjelde, 2001a). The lower crust below the Knipovich Ridge requires a lower density (2,750 kg m\(^{-3}\)) than the surroundings. This was also supported by the lower seismic P-wave velocities (5.5–6.5 km s\(^{-1}\)) under the rift valley. The gravity model also requires a lower mantle density in the oceanic domain (3,280 kg m\(^{-3}\)). The profile was modeled to match Profile P6 at point “P6” as interpreted with a thinner crust by Hermann & Jokat (2013) and the observed free-air gravity anomaly (Figure 7c). The Moho was raised up by 2 km west of the rift valley improving the agreement between the modeled and observed gravity profile. This also reduces the crustal thickness difference at the intersection between Profiles P3 and P6. On the western flank, where sediments are accumulated, the crust beneath is modeled 1–2 km thicker than the seismic interpretation to fit the free-air gravity anomaly. The transition domain is demarked at 150 km where a lower susceptibility and no remanence are required for the crust and higher density for the mantle.

### Table 3

| Unit | Lithology            | Density kg m\(^{-3}\) | Susceptibility | Remanence Amplitude / Inclinaison A m\(^{-3}\) / ° |
|------|----------------------|------------------------|----------------|---------------------------------------------------|
|      | Sediment             | 2000 - 2400            | 0              | -                                                 |
|      | Sediment             | 2400 - 2570            | 0              | -                                                 |
|      | Compacted sediment   | 2500 - 2690            | Low            | -                                                 |
|      | Continental basement | 2630 - 2740            | Low            | -                                                 |
|      | Intrusion            | 2700 - 2970            | 0.01 - 0.08    | -                                                 |
|      | Intrusion            | 2800 - 2920            | Low            | 0.55 - 1.70 / +90                                |
|      | Intrusion            | 2600 - 2900            | Low            | 1.30 - 1.95 / -90                                |
|      | Basalt (normal field)| 2700 - 2750            | Low            | 0.05 - 7.50 / +90                                |
|      | Basalt (reversed field) | 2700 - 2750           | Low            | 0.05 - 5.00 / -90                                |
|      | Basalt (normal field) - Extinct ridge | 2700 | Low | 0.90-3.80 / +90 |
|      | Basalt (reversed field) - Extinct ridge | 2700 | Low | 0.60 / -90 |
|      | Continental crust     | 2880 - 3090            | 0.005 - 0.045  | -                                                 |
|      | Transition crust      | 2800 - 2970            | 0.005 - 0.01   | -                                                 |
|      | Gabbroic oceanic crust| 2900 - 2950            | 0.005 - 0.026  | -                                                 |
|      | Gabbroic oceanic crust| 2750 - 2850            | 0.005 - 0.04   | -                                                 |
|      | Mantle oceanic domain| 3270 - 3280            | 0              | -                                                 |
|      | Mantle continental domain | 3310 - 3330         | 0              | -                                                 |

Note: The 2-D forward model were constructed with a common constant to render the models comparable. A small variation in density, susceptibility, and remanence within a single geological unit is seen on individual and from profile to profile. However, there is a good correlation overall between the profile interpretations.
4.4.4. Profile P4 (Figure 8a)

This profile starts at the Hovgaard Ridge extending through the Molloy Transform Zone and terminates onshore Svalbard and is located to the north of the KRAS-16 survey. The crust below the Molloy Transform Zone was modeled with a density of 2,800 kg m\(^{-3}\) that is lower than the values beneath the Knipovich Ridge (2,900 kg m\(^{-3}\)). Two magmatic bodies are modeled between the Molloy Transform Zone and Spitsbergen. In that area, east of the Molloy Transform Zone, the observed magnetic profile has higher values than the modeled magnetic profile. This section is challenging to interpret due to the poor data quality. It is near the junction of the Knipovich Ridge and the Molloy Transform Zone, where fluids penetration in the crust and interaction with the crustal composition might be expected. The profile could be modeled with remanent magnetization diagnostic of an oceanic crust; however, no clear magnetic striped pattern is observed in the gridded data (Figure 2). A serpentinization process could explain both the reduced density of the mantle required below the Molloy Transform Zone and the magnetic high in the section between ~120 and ~190 km. The oceanic and transition crust necessitates higher susceptibilities than those on the Knipovich Ridge.

Similar to Profiles P1 and P2, a thick continental crust with high density (2,930–3,000 kg m\(^{-3}\)) is modeled. The seismic interpretation suggested a low velocity continental upper section but the gravity modeling requires higher densities (2,750 kg m\(^{-3}\)) than initially expected (2,600 kg m\(^{-3}\)) (Ritzmann et al., 2004). Similar higher densities up to 2,690 kg m\(^{-3}\) are modeled in the current study. The mantle density under the Molloy Transform Zone is reduced to 3,270 kg m\(^{-3}\), a value slightly lower than the interpretation under the Knipovich Ridge further south. The continental section of the profile can be interpreted as compacted sediments of Devonian age (2,730 kg m\(^{-3}\)) and <0.0001 SI on a basement of Precambrian age (2,920–2,950 kg m\(^{-3}\) and 0.03 SI).

4.4.5. Profile P5 (Figure 8b)

Profile P5 starts west of the Molloy Ridge, crosses north of Spitsbergen and ends offshore Nordaustlandet. The rift valley is not aligned with the highest section of the Moho. The magnetic striped pattern indicative of an oceanic crust is not observed on the gridded magnetic data (Figure 2). The data accuracy and resolution could not allow us to resolve magnetic isochrons or to determine the spreading of the Molloy Ridge. Therefore, the COB is difficult to interpret from the magnetic data as the profile is located outside the high-resolution KRAS-16 survey. The Billefjorden Fault Zone is associated with a magnetic high as previously described by Skilbrei (1992) and modeled farther south (Figure 8b). A large magmatic intrusion is modeled west of the rift valley. The thin lower crust between the Molloy Ridge and Spitsbergen is strongly magnetized (0.025 SI) with lower density than the lower crust interpreted beneath the Knipovich Ridge.

On the continental section, low crustal P-wave velocities (6.0–6.5 km s\(^{-1}\)) were interpreted on Profile P5 (Czuba et al., 2005). However, the gravity model requires higher densities (2,900–2,950 kg m\(^{-3}\)). The densities for the crustal and sediment layers in the model are similar to those found in the continental domain of profiles P1, P2, P3, and P4. The horizons from the seismic interpretation were strongly used for the gravity and magnetic model. Given the density and susceptibility comparable to Profile P4 (Figure 8a), the continental domain can be interpreted as Devonian compacted sediments on a Precambrian basement. The gravity interpretation of the mantle correlates with Profile P4 requiring low density (3,270 kg m\(^{-3}\)) beneath the Molloy Ridge and higher densities (3,330 kg m\(^{-3}\)) farther away. The gravity model slightly disagrees with the observed gravity data at the Molloy Ridge and at the shallowest point of the Moho. To fit the mismatch, the Molloy Ridge requires an increase of density or a thinner crust while the shallowest point of the mantle requires a decrease of density or a thicker crust. Modifying the geometry of the crust would move the shallowest point of the mantle toward the Molloy Ridge improving the alignment between the rift valley and the shallow Moho. The low resolutions of the gravity data and the seismic interpretations are a potential cause of misfit of the models.

4.4.6. Profile P6 (Figure 8c)

This profile starts from the Boreas Basin and intersects the Knipovich Ridge. It also intersects Profiles P2 and P3 at points “P2” and “P3,” respectively. At these points, the seismic interpretation of Profile P6 is different from Profile P2 and P3. Thus, the crustal thickness from the seismic interpretation of profile P6 is used to test and compare the 2-D forward interpretation of profiles P1 and P2. P6 is mostly located in the oceanic domain where the striped magnetic signal is found. A small portion is located in the transition domain. In the Boreas Basin, an extinct oceanic ridge has been interpreted from the KRAS-16 data set (Dumais, Gernigon, et al., 2020). The horizons from the seismic interpretation are used in the 2-D model. The densities are modified from the
seismic interpretation to be comparable to the other profiles, but the density contrasts and their interpretation are respected amongst the geological bodies. The final densities are within the uncertainty of velocity-density relationship (Barton, 1986) and representative of gabbros and basalts. The magnetization values are generally low, and the striped pattern is not clear between the current Knipovich Ridge and the extinct ridge (Figure 8c). At this location, the gridded magnetic data show a fracture zone or lineament underlying the seismic profile consistent with the lack of a magnetic striped pattern. However, the magnetic striped pattern is clearly recognized a few kilometers away from the profile. The crustal thickness and densities below the extinct ridge and the current spreading ridge are similar. On the western half of the extinct ridge, the crustal density increases. The crustal thickness also increases west of the extinct ridge. The mantle density along P6 is lower below the present-day spreading ridge like profiles P1, P2, and P3. Along profiles P2 and P3, the crustal thickness has been adjusted to Profile P6 as the crustal thickness of P6 is more representative of a slow-spreading system. For profiles P2, P3, and P6, the resulting models with a thinner crust agree with the observed gravity data and with the thin crust interpreted along the rift valley (Jokat et al., 2012). The mantle density under the extinct ridge is higher than the mantle density under the present-day ridge, perhaps indicative of a colder mantle. The mantle density increases west of the abandoned ridge but remains lower than the transition mantle densities modeled east of the Knipovich Ridge for profiles P1, P2, and P3.

4.4.7. Profile P7 (Figure 8d)

Profile P7 extends from the Knipovich Ridge and eastward almost to Bjornoya and intersects Profile P1 over the rift valley of the Knipovich Ridge. Multiple strong magnetic anomalies are modeled. The magnetic anomalies showing the greatest peak-to-peak intensity variation of 200–250 nT located to the west of the profiles are identified as magnetic isochrons (C1, C2A, and C6) associated with an oceanic crust. The magnetic isochron C2A is not fully resolved due to either the obliquity of the ridge, or the presence of a lineament identified on the magnetic gridded data (Figure 2), or the generally low magnetization of the ridge at this location, or a combination of these effects. Wider magnetic anomalies with intermediate peak-to-peak amplitude variation from 40 to 100 nT located in the transition and continental zones are identified as strong magnetized bodies with normal and reversed remanence. On the far east of the model, a magnetized body with reversed magnetization is identified east of the Hornsund Fault Zone.

The density model generally agrees with the seismic interpretation (Libak, Eide, et al., 2012). The physical properties of the crust at the intersection with profile P1 are in agreement for both profiles P1 and P7. The lower crust west of the Hornsund Fault Zone, where the crust thickens toward a continental crust, shows a higher density of 3,000 kg m$^{-3}$ compared to densities assigned to the lower crust in this profile. At the Far East, the 2-D model required a higher density of the lower continental crust compared to the seismic interpretation which shows lower resolution due to the reduced ray coverage (Libak, Eide, et al., 2012). A lower density was assigned to the mantle in the oceanic domain (3,280 kg m$^{-3}$) compared to the mantle in the transition and continental domain (3,300 kg m$^{-3}$).

4.4.8. Profile P8—CSEM and MT (Figure 9)

Profile P8 is located on the oceanic domain located at the junction between the Knipovich and Mohns ridges. The magnetic interpretation was described by Lim (2020). Densities used in the 2-D forward model are comparable to the other profiles, particularly the closest Profile P1. The central anomaly presents a small depression in the middle of the anomaly caused by the presence of two mounts in the rift valley. The eastern lobe of the central anomaly is slightly higher than the western lobe with intensities of 700 and 400 nT, respectively. The basalt layer is about 1 km in thickness with a heterogeneous magnetization. The magnetization values have slightly changed from Lim (2020) interpretation due to adjustment between the profiled data and the merged magnetic compilation, but the geometry remains unchanged. The mantle density directly under the rift valley is reduced as suggested by the mantle temperature gradient (Johansen et al., 2019). The disagreement between the modeled and observed gravity profiles is explained by the coarse resolution of the gravity data compared to the resolution and accuracy of the CSEM/MT interpretation. The geometry of the CSEM/MT interpretation requires smaller wavelengths not recorded in the gravity data. However, the gravity model requires a lateral density variation of the mantle as suggested by the MT thermal interpretation (Figure 9). This local density variation indicates the narrow nature of the thermal model. However, higher resolution gravity data are necessary to reconstruct the density gradient below the Knipovich Ridge.
4.4.9. Uncertainties of the Potential Field 2-D Models

The potential field 2-D models have inherent uncertainty due to the ambiguity or nonuniqueness of the modeling as several mass or magnetization distribution may produce the same potential field over a specific surface (Skeels, 1947). Seismic interpretation profiles were used to constrain the ambiguity, but the interpretation at the edges and the deepest layers are less resolved than the rest of a given profile. Seismic interpretation is also not unique; the response of the highest P-wave velocity rocks might be picked but does not represent the average P-wave-velocity of a geological layer or body. This uncertainty might generate uncertainties in the densities and magnetization values. The KRAS-16 aeromagnetic data set is gridded with 1 km cell size providing an estimate of the spatial resolution of the data. Due to the line spacing of the surrounding surveys the spatial resolution is often less for the profiles extending beyond the KRAS-16 boundaries with an overall accuracy less than 5 nT (Olesen et al., 2010). The gravity data have a spatial resolution of 1 arcsec and an accuracy of 2 mGal (Sandwell et al., 2014). The overall uncertainty of the models is reduced by the similarities of the studies where density and magnetization values, and the interface depths are verified by similitude of the geological layers across the different profiles.

5. Discussion

The 3-D magnetization model and the 2-D forward gravity and magnetic models illustrate the spatial variation of the crustal densities and magnetic properties along the Fram Strait surrounding the active Knipovich Ridge. Figure 10a illustrates the 2-D forward models in a 3-D perspective with the Knipovich Ridge and the COB.

5.1. Oceanic Domain

The Oceanic domain, delimited by the COB ridge-ward, comprises striped magnetized anomalies. As observed in the 3-D magnetization model and the 2-D forward models, the rift valley of the ridge varies northward from almost none to high magnetization. The 3-D magnetization model assumes a homogeneous crust from the bottom of the sediment layer to the top of the mantle without differentiation for layers 2A, 2B, and 3, the classical
division of the oceanic crust (LaFemina, 2015; Perfit, 1999). Layer 2A composition and thickness has the most influence on the magnetic anomaly (Tivey & Johnson, 1993). Layer 2A is typically thin (less than 1 km) and its magnetization values are typically higher than the other layers. However, the 3-D magnetization model provides information on the type of magnetization found along the Knipovich Ridge. The magnetization pattern along the Knipovich suggests a segmentation of the accretion process. The high magnetization areas in the north of the Knipovich Ridge correlate with the presence of volcanoes and bathymetric highs in the rift valley, suggesting the oceanic crust has different physical properties here than at the southern latitudes. Therefore, the mantle processes, the volume and composition of the magma chamber below the ridge and the cooling processes are expected to vary from south to north. An iron- or oxide-rich magma chamber at latitudes 76°N–78°N compared to the southern section of the Knipovich Ridge could explain the presence of volcanoes and higher magnetization (Figure 5b). Multiple fracture zones and the bend with the Mohns Ridge may also cause a fluid interaction with the basalt layers of the oceanic crust, de-magnetizing the basalts. Given the low magnetization at the lower latitudes (74°–76°N), an amagmatic segment is proposed to explain the low magnetization. This long amagmatic segment is comprised between two elongated bending C6 anomalies segmented by lineaments strongly visible on the magnetic data. Along this amagmatic segment very few Werner solutions are derived (Figure 4b), however, a shallow mantle is inferred by a shallow bottom depth of the deepest magnetic sources interpreted by the Curie point depth estimation.

At the Molloy Transform Zone, profiles P4 and P5 demark a narrow oceanic domain with a thin crust identified by seismic interpretation with high susceptibility values derived from the 2-D forward models. East of the Molloy Transform Zone, Profile P4 presents a section of the crust with low density values and a very high magnetic anomaly observed. Such a wide high magnetic anomaly could be caused due to a deep serpentinized peridotite; however, the aeromagnetic data quality is not optimal in this section. The process of serpentinization may affect the physical properties of the crust such as the density. The density is usually inversely proportional to the degree of serpentinization while the magnetism often increases as magnetite is produced (Dentith & Mudge, 2014). Profile P4 is near the junction of the Knipovich Ridge and the Molloy Transform Zone, where fluids penetration under the oceanic domain newly defined in light of the high-resolution KRAS-16 survey. Local reduced mantle densities are interpreted in the oceanic domain. Low mantle densities are also required under the extinct ridge identified on Profile P6 (Figure 8c). The lowest mantle densities are found under the active rift valley in Profile P8 (Figure 9), the highest resolution profile which offers a unique view of the Knipovich Ridge thermal heat variation. However, the gravity data resolution does not allow to accurately model the density variation correlated to the thermal mantle variation. The 2-D forward model (Figure 9) does not support an isostatic equilibrium thermal distribution but agrees with a thermal variation between the oceanic and continental domains as suggested by Breivik et al. (1999). The lateral variation is more discrete and narrowly concentrated under the oceanic domain newly defined in light of the high-resolution KRAS-16 survey.

Unfortunately, Profile P6 aligns with a magnetic lineament possibly associated with a fracture zone. A better image and description of the extent of the extinct ridge could be achieved by acquiring a seismic line perpendicular to the magnetic anomaly in the Boreas Basin combined with a 2-D forward modeling.

5.2. Continental Domain

The unambiguous continental domain lies roughly eastward from the Hornsund Fault Zone (Figure 7). This demarcation is seen on the 2-D forward models where the crust is thicker, with densities of 2,850–2,950 kg
m$^{-3}$ and a Moho depth of 30 km below the sea-level. The densities used in the continental crust are similar to the ones applied for gravity modeling in the original seismic interpretation (Breivik & Mjelde, 2001b; Czuba et al., 2005, 2008; Grad & Majorowicz, 2020; Ritzmann et al., 2004). The Curie point depth estimation (Figure 4a) also reaches a depth of 25–30 km depicting a colder crust with deeper bottom depths of the deepest magnetic sources compared to the oceanic domain. The free-air and magnetic maps contain low-frequency anomalies, associated with a deep basement (Figures 2 and 3). Intermediate amplitude anomalies (Figure 2) are likely associated with magmatic intrusions such as dikes or sills. Their emplacement and depth are estimated in the 2-D forward models (Figure 7). A high magnetic anomaly is found along the Hornsund Fault Zone where the basement rises to the seafloor on profile P2 (Figure 7b). On Profile P1 (Figure 7a), the magnetic anomaly is pronounced (Figure 2) and requires a magmatic intrusion (0.01 SI) along the Hornsund Fault Zone. Dikes and sills are modeled on the Edgøya platform expressing a largely intruded basement. On Profile P5 (Figure 8b), the basement is more magnetized between the Molloy Ridge and Spitsbergen and an intrusion is associated with Billefjorden Fault Zone.

The susceptibilities used in the continental and continent-ocean transition crust and the magmatic intrusion are in accordance with the values found in the continental shelf of the Barents Sea. Marello et al. (2013) have compiled susceptibilities of 0.00004–0.029 SI for the upper crust, 0.00006–0.007 SI for the lower crust and 0.003 to 0.01 SI for the continent-ocean transition crust. Barrère et al. (2009, 2011) differentiate the Caledonian nappes upper crust with susceptibilities of 0.0001–0.01 SI, the Archean-Proterozoic upper crust with 0.01–0.20 SI, the lower crust with 0.0001 SI and mafic intrusion with 0.015–0.05 SI. On Nordaustlandet, Dumais and Brönner (2020) derived a Mesoproterozoic-Neoproterozoic basement more magnetic than the Caledonian nappes basement. The continent-ocean transition crust interpreted with the 2-D forward models has susceptibilities of 0.005–0.01 SI similar to the study of Marello et al. (2013). The continental crust is interpreted with susceptibilities 0.005–0.045 SI. The highest crustal susceptibilities are modeled deep to fit the long wavelengths of the magnetic data. Generally, the continental crust along the southern section of the Knipovich Ridge (Profiles P1 and P2) is interpreted with Archean-Proterozoic and Caledonian Nappes basement in accordance to Barrère et al. (2009, 2011) and Fichler & Pastore (2022). The continental crust interpreted in profile P4 and P5 might be apparent to a Mesoproterozoic-Neoproterozoic basement with generally higher susceptibilities than those compiled for the Caledonian Nappes. Therefore, on Profiles P1 and P2, the basement susceptibility and density gradually vary across the continental domain. These variations can be associated with possible Caledonian nappes gradually extending on the margin. Profiles P4 and P5 present a different basement configuration with higher susceptibility and density in the lower crust modeled at the demarcation between the transition and the continental domains.

5.3. Transition Domain: Three Possible Scenarios

The oceanic and continental domains have a clear signature on the gravity and magnetic maps and are well tested and illustrated by the 2-D forward models. The oceanic and continental domains are separated by a distinct and wide transition domain that extends for tens of kilometers (Figures 2 and 8). The Curie depth allows to estimate the extent of the transition domain on both sides of the ridge (Figure 4a) where the Curie point depth migrates from shallow (6 km below sea-level) to deep (25–20 km below sea-level). This transition is also marked by a gradual thickening of the crust and a lack of magnetic striped pattern (Figures 7 and 8). The transition domain defined in this study shares higher mantle densities compared to the adjacent oceanic domain. Similar mantle densities are also expected in the continental domain (Figures 7 and 8). The transition domain is wider at latitude 76°N and slowly narrows until latitude 80°N where it reaches its minimal extent on the eastern margin. However, such a wide and ambiguous transition zone is not unique on Earth. It is recognized along the northern Mozambican margin (Vormann & Jokat, 2021) developed during an oblique phase of continental stretching between Africa and Davie Ridge. The pull-apart passive Pará-Maranhão/Barreirinhas margin is also interpreted with a 60 km wide necking domain below the Ilha de Santana Platform, followed by a 5-km thick crust interpreted as an exhumed lower continental crust and a pro-oceanic crust before the thin but “normal” oceanic crust (Aslanian et al., 2021). About a 150 km wide and ambiguous transition zone is interpreted by Bécel et al. (2020) at the Eastern North American Volcanic Rifted Margin and described as a proto-oceanic domain emplaced before the establishment of normal, steady state oceanic crust. Exhumed lower crust was also proposed in the distal part of the Namibian volcanic margin before the onset of seafloor spreading (Geoffroy et al., 2022). In the Jequitinhonha basin, Brazil, Loureiro et al. (2018) has interpreted a 150-km wide transition zone of exhumed lower-median continental crust before the identification of a clear oceanic crust. The Pernambuca and Paraiba basins on the
Brazilian side were interpreted with more than 50 km of exhumed stretched continental crust with an abrupt thinning toward the ocean (Matos et al., 2021) similar to the configuration on profiles P1 and P2 in this study. In the Fram Strait, the presence of Mohns Ridge and the East Greenland Ridge, on the west of southernmost section of Knipovich Ridge, could explain for the more prominent continental stretching on the eastern margin. Longer seismic profiles of P1 and P2, above the western margin and Greenland Fault Zone, would confirm the nature of the crust between the ocean spreading of Knipovich and Mohns ridges. On the western margin, the transition domain comprises the East-Greenland Ridge and a large portion of the Boreas Basin.

High densities of 2,800–2,970 kg m\(^{-3}\) for the crust of the transition domain often associated with high P-wave velocity indicate an intermediate-mafic crystalline crust. The intrusions are modeled with the same average density as the host crustal layer since the gravity response does not indicate a variation. Thin high density crystalline crust is often interpreted as an oceanic crust, but in a transition domain, a proto-oceanic crust and an intruded lower continental crust or alternatively a zone of exhumed continental mantle might also exhibit high densities, for example (Bécel et al., 2020; Geoffroy et al., 2015; Jolivet et al., 2015; Sibuet et al., 2007). Thus, the interpretation of the transition domain requires more investigations beyond P-wave velocity and density analysis.

Sparse magnetized bodies in the transition domain have an estimated depth of 6–7 km below the sea-level or deeper which locate them in the upper crustal layer on the eastern margin while they can be shallower (5–7 km below sea-level) on the western margin according to the Werner deconvolution (Figure 4b). Intrusions are more numerous in the south-east margin, near the Vestbakken Volcanic Province. At that latitude, the conjugate margin is demarked by the prominent East-Greenland Ridge and a long amagmatic segment of the Knipovich Ridge characterized with low magnetization. On the north-east margin of the Knipovich Ridge, few major intrusions are found, but more prominent intrusions are located at the same latitude on the western margin in the Boreas Basin. These intrusions present in the transition zone do not resemble the signature of a steady state oceanic crust. The round shapes of the magnetic anomalies and the limited space for additional magnetic isochrons on the Northeast Greenland Margin indicate that other processes have shaped the transition zone.

The Knipovich Ridge initiated in a transtensional system, where the plate was sliding along the Hornsund Fault Zone causing pull-apart basin infilling with thick sediments. Three scenarios could provide explanation to the observations in the transition zone.

1. The transition domain could comprise an oceanic crust buried under a thick sediment layer. One may consider the numerous intrusions found on profiles P1 and P2 in the eastern transition domain have similar magnetic response to the oceanic magnetic isochrons. However, those intrusions have a rounded shape in the magnetic map and are not linear as commonly expected for oceanic seafloor spreading anomalies. Sediments have a very low-to-negligible magnetization, such as the method is still sensitive to the contrast against the strongly magnetized crust. Even at a certain depth under sediments, the process of an exhumed mantle or the serpentinization of an oceanic crust would leave indication of an oceanic crust such as some degree of linear or striped magnetic anomalies. The lack of clear magnetic stripes is the main argument against the presence of an oceanic crust hypothesis. If the transition domain was truly oceanic, it is then difficult to explain the processes behind the absence of magnetic lineation observed at present day. Moreover, no corresponding anomalies are found on the west conjugate margin where space is also not sufficient for seafloor spreading.

2. Alternatively, the transition domain could be formed locally by continental mantle exhumation and serpentinization. During the rifting before the spreading of the ridge, the continental crust was extended and drastically thinned while the rift was filled with sediments. Significant hyperextension could have initiated the progressive exhumation and denudation as of the exhumed mantle as acknowledged on the Iberian margin (Pérez-Gussinyé et al., 2001; Sutra & Manatschal, 2012; Whitmarsh et al., 1993). This scenario is consistent with the crustal structure interpreted by the seismic profiles and the crustal properties apparent in the continental domain. The crust of the transition zone could have been highly intruded during the mantle exhumation. The high-density lower crust interpreted could also fit values of exhumed serpentinized mantle density (Minshull, 2009). However, the Moho discontinuity is well defined by the refraction waves, whereas it is usually unclear in a heavily serpentinized mantle setting (Christensen, 1966, 1978, 1996; Horen et al., 1996).

3. Finally, the transition domain could represent an exhumed lower continental crust as observed in many hyperextended rift systems (Clerc et al., 2015; Sapin et al., 2021). The high densities interpreted in the transitional crust might indicate the gabброic or migmatic-eclogitic composition of a stretched lower continental crust. During the rifting but before the seafloor spreading initiation at C6, the intruded continental lower crust could
have gradually migrated and exhumed (Figure 10b) toward the proto-oceanic domain. A preexisting thick and low-viscosity lower continental crust caused by the Eurekan orogeny could explain a lateral flow of the ductile lower crust. Given the seafloor spreading initiation at C6 (20 Ma), the Eurekan deformation occurred prior to the ridge spreading (Piepjohn et al., 2016). The Eurekan deformation could have thickened and softened the crust. A rapid collapse and rifting could have led to a lateral escape of low-viscosity lower continental material. This scenario is consistent with the crustal and mantle properties interpreted in the 2-D forward models. The crustal properties of the transition domain are similar to the continental crust with higher densities in the transition domain on profiles P1, P2, and P3 (Figure 7).

In this study, the third scenario is favored given the tectonic setting in the Fram Strait before the seafloor spreading initiation of the Knipovich Ridge. The 2-D forward models illustrate the evolution of the spreading ridge from the continental to the oceanic crust (Figure 10a). On the 2-D forward model, a wide continent-ocean transition zone is interpreted on the east flank of the southern section of the Knipovich Ridge (74°N) characterized with an intruded crust and possibly exhumed mantle. However, the western flank of the southern section of the Knipovich Ridge, a much narrower continent-ocean transition zone, with less intrusion in the crust, is suggested by the aeromagnetic data (Figure 2). On the northern section of the Knipovich Ridge (78°N), a narrowing of the continent-ocean transition zone on the eastern flank and a ridge jump are interpreted from both the 2-D forward models and the aeromagnetic data. The crust is also more intruded on the eastern flank as suggested by the aeromagnetic data (Figure 2). This asymmetric opening of the Fram Strait is illustrated in Figure 10b. Numerous lineaments and important variation of the magnetization along the Knipovich Ridge are interpreted from the aeromagnetic data suggesting magmatic and amagmatic accretion and possibly mantle exhumation and fluids penetration changing the crustal composition. The structure and composition of the crust, and the properties of the mantle were extracted from the profiles across the Knipovich Ridge. However, the modeling used seismic constraints from independent interpretations and not directly from digital seismic data. This remains an inherent uncertainty to the geometries and physical properties derived from the potential field modeling in this study. The 2-D interpretations reveal several misfits between the various profile interpretations likely caused by different velocity-density models or gaps in the data (Funck et al., 2017). The interpretation highlights the necessity of a reanalysis toward uniform crustal structures with a thin oceanic crust. Nevertheless, the gravity and magnetic interpretation indicates that the Fram Strait has opened in a complex setting causing the asymmetric and heterogeneous oceanic and transition domains.

6. Conclusions

This study aimed to describe and model the crustal and mantle heterogeneities along the Knipovich Ridge and its surrounding margins. The new KRAS-16 aeromagnetic data set enclosed the oceanic and transition domains. Aeromagnetic data from previous works were compiled to provide a global overview of the transition to the continental domain. The gravity compilation, the EM interpretation and the seismic interpretation provided indications and constraints to the combined gravity and magnetic interpretation.

1. The oceanic crust is clearly demarked by the striped magnetic pattern and its location confirmed by the 2-D forward models. The COB is derived from this interpretation and confirmed by the modeling.
2. The initiation of the seafloor spreading is delineated by magnetic isochron C6 (20 Ma) and an extinct ridge is modeled in the Boreas Basin. The 2-D forward models are revised and unified accordingly suggesting the presence of a wide transition domain.
3. The magnetization in the oceanic domain is linked to the presence of volcanoes and bathymetric highs in the rift valley. Iron-oxide-rich segments are identified along the rift valley.
4. The delineation of several lineaments and the bend configuration of the Knipovich Ridge is associated with a variation in the magnetization and settings along the Knipovich Ridge. The segmentation and configuration of the ridge possibly control the seafloor spreading settings and fluid circulation influencing its composition and magnetization.
5. Mantle heterogeneities occur with an east-west lateral density variation and indicate a transition from a hotter mantle in the oceanic domain to a colder mantle underneath the older continental crust.
6. We favor the presence of a wide transition lithospheric domain comprising an exhumed lower crust or mantle. The oblique spreading constrained by the Mohns Ridge and the East Greenland Ridge may have favored a
continental stretching on the eastern margin. Compared to previous interpretations, we extend the Norwegian continental domain by up to 150 km further west in the study area, a rare occurrence in plate tectonics.

Data Availability Statement

The KRAS-16 aeromagnetic data are available through the search engines of the Geological Survey of Norway public repository (https://geo.ngu.no/geoscienceportalopen listed as KRAS-16) and EPOS-N Portal (https://epos-no.uib.no:444/view/project listed as Aeromagnetic survey data, Knipovich Ridge).

Acknowledgments

We are thankful to the European Plate Observing System—Norway; the Norwegian Petroleum Directorate; and the Geological Survey of Norway to help funding the project. We thank Novatem, Inc. for the data acquisition, and our colleagues from AWI (Wilfried Jokat) and TGS (Reidun Myklebust) for providing aeromagnetic data from adjacent areas. We thank Wilfried Jokat and an anonymous reviewer for their constructive comments on the manuscript. We are grateful to Richard Saltus and Carmen Gaina for their valuable suggestions on previous versions of the manuscript.

References

Aslanian, D., Gallais, F., Afifhado, A., Schuurpe, P., Moulin, M., Evain, M., et al. (2021). Deep structure of the Pará–Maranhão/Barreirinhas passive margin in the equatorial Atlantic (NE Brazil). Journal of South American Earth Sciences, 110, 103322. https://doi.org/10.1016/j.jseaes.2021.103322

Baillard, C., Crawford, W. C., Ballu, V., Hibert, C., & Mangene, A. (2013). An automatic Kurtosis-based P- and S-phase picker designed for local seismic networks. Bulletin of the Seismological Society of America, 104(1), 394–409. https://doi.org/10.1785/0120120347

Barrière, C., Ebbing, J., & Gernigon, L. (2009). Offshore prolongation of Caledonian structures and basement characterisation in the western Barents Sea from geophysical modelling. Tectonophysics, 470(1–2), 71–88. https://doi.org/10.1016/j.tecto.2008.07.012

Barrière, C., Ebbing, J., & Gernigon, L. (2011). 3-D density and magnetic crustal characterization of the southwestern Barents Shelf: Implications for the offshore prolongation of the Norwegian Caledonides. Geophysical Journal International, 184(3), 1147–1166. https://doi.org/10.1111/j.1365-246X.2010.04888.x

Barton, P. J. (1986). The relationship between seismic velocity and density in the continental crust—A useful constraint? Geophysical Journal International, 87(1), 195–208. https://doi.org/10.1111/j.1365-246X.1986.tb04553.x

Bécel, A., Davis, J. K., Shuck, B. D., Van Avendonk, H. J. A., & Gibson, J. C. (2020). Evidence for a prolonged continental breakup resulting from slow extension rates at the Eastern North American volcanic rifted margin. Journal of Geophysical Research: Solid Earth, 125(9), e2020JB002993. https://doi.org/10.1029/2020JB002993

Blischke, A., Brandsdóttir, B., Stoker, M. S., Gaina, C., Erlendsson, Ó., Tegner, C., et al. (2022). Seismic volcanostратigraphy: The key to resolving the Jan Mayen microcontinent and Iceland Plateau Rift evolution. Geochemistry, Geophysics, Geosystems, 23(4), e2021GC009948. https://doi.org/10.1029/2021GC009948

Blischke, A., Gaina, C., Hopper, J. R., Perón-Pinvidic, G., Brandsdóttir, B., Guarnieri, P., et al. (2017). The Jan Mayen microcontinent: An update of its architecture, structural development and role during the transition from the Åegir Ridge to the mid-oceanic Kolbeinsey Ridge. Geological Society, London, Special Publications, 447(1), 299–337. https://doi.org/10.1144/sp447.5

Bouligand, C., Glen, J. M. G., & Blakey, R. J. (2009). Mapping Curie temperature depth in the western United States with a fractal model for crustal magnetization. Journal of Geophysical Research, 114(B11). https://doi.org/10.1029/2009jb0060494

Breivik, A., Faleide, J. I., & Gudlaugsson, S. T. (1998). Southwestern Barents Sea margin: Late Mesozoic sedimentary basins and crustal extension. Tectonophysics, 293(1–2), 21–44. https://doi.org/10.1016/s0040-1951(98)00073-0

Breivik, A., Mjelde, R., Grogan, P., Shimamura, H., Murai, Y., & Nishimura, Y. (2003). Crustal structure and transform margin development south of Svalbard based on ocean bottom seismometer data. Tectonophysics, 369(1–2), 37–70. https://doi.org/10.1016/s0040-1951(03)00131-8

Breivik, A., Mjelde, R., Grogan, P., Shimamura, H., Murai, Y., & Nishimura, Y. (2005). Caledonide development offshore–onshore Svalbard based on ocean bottom seismometer, conventional seismic, and potential field data. Tectonophysics, 401(1–2), 79–117. https://doi.org/10.1016/j.tecto.2005.03.009

Breivik, A. J., & Mjelde, R. (2001a). Obs-98 survey: Final report oceanic profiles. Report. University of Bergen.

Breivik, A. J., & Mjelde, R. (2001b). Obs-98 survey: Final report western continental profiles. Report. University of Bergen.

Breivik, A. J., Verhoef, J., & Faleide, J. I. (1999). Effect of thermal contrasts on gravity modeling at passive margins: Results from the western Barents Sea. Journal of Geophysical Research, 104(B7), 15293–15311. https://doi.org/10.1029/99jb00022

Brozena, J. M., Childers, V. A., Lawver, L. A., Gahagan, L. M., Forsberg, R., Eldholm, O., et al. (2003). New aerogeophysical study of the Eurasia Basin and Lomonosov Ridge: Implications for basin development. Geology, 31(9), 825–828. https://doi.org/10.1130/g19528.1

Christensen, N. I. (1966). Elasticity of ultrabasic rocks. Journal of Geophysical Research, 71(24), 5921–5931. https://doi.org/10.1029/jj071i024p05921

Christensen, N. I. (1978). Ophiolites, seismic velocities and oceanic crustal structure. Tectonophysics, 47(1), 131–157. https://doi.org/10.1016/0040-1951(78)90155-5

Christensen, N. I. (1996). Poisson’s ratio and seismic seismology. Journal of Geophysical Research, 101(B2), 3139–3156. https://doi.org/10.1029/95JB03446

Christensen, N. I., & Mooney, W. D. (1995). Seismic velocity structure and composition of the continental crust: A global view. Journal of Geophysical Research, 100(B6), 9761–9788. https://doi.org/10.1029/95JB0259

Clerc, C., Jolivet, L., & Ringenbach, J.-C. (2015). Ductile extensional shear zones in the lower crust of a passive margin. Earth and Planetary Science Letters, 411, 1–7. https://doi.org/10.1016/j.epsl.2015.08.038

Czuba, W., Grad, M., Guterch, A., Mažajžički, M., Malinowski, M., Mjelde, R., et al. (2008). Seismic crustal structure along the deep transect Horstef05, Svalbard. Polish Polar Research, 29(3), 279–290.

Czuba, W., Ritzmann, O., Nishimura, Y., Grad, M., Mjelde, R., Guterch, A., & Jokat, W. (2005). Crustal structure of northern Spitsbergen along the deep seismic transect between the Molloy Deep and Nordaustlandet. Geophysical Journal International, 161(2), 347–364. https://doi.org/10.1111/j.1365-246X.2005.02593.x

Dallmann, W. K. (2015). Geosciences Atlas of Svalbard.

Dentith, M., & Mudge, S. T. (2014). Geophysics for the mineral exploration geoscientist. Cambridge University Press.

Dick, H. J. B., Lin, J., & Schouten, H. (2003). An ultralow-spreading class of ocean ridge. Nature, 426(6965), 405–412. https://doi.org/10.1038/nature02128
Srivastava, S. P., & Tapscott, C. R. (1986). Plate kinematics of the North Atlantic. In P. R. Vogt & B. E. Tucholke (Eds.), *The Western North Atlantic Region*. Geological Society of America.

Suckro, S. K., Gohl, K., Funck, T., Heyde, I., Schreckenberger, B., Gerlings, J., & Damm, V. (2013). The Davis Strait crust—A transform margin between two oceanic basins. *Geophysical Journal International, 193*(1), 78–97. https://doi.org/10.1093/gji/ggs126

Satra, E., & Manatschal, G. (2012). How does the continental crust thin in a hyperextended rifted margin? Insights from the Iberia margin. *Geology, 40*(2), 139–142. https://doi.org/10.1130/g32786.1

Talwani, M., & Eldholm, O. (1977). Evolution of the Norwegian-Greenland Sea. *The Geological Society of America Bulletin, 88*(7), 969–999. https://doi.org/10.1130/0016-7606(1977)88<969:eotns>2.0.co;2

Tivey, M. A., & Johnson, H. P. (1993). Variations in oceanic crustal structure and implications for the fine-scale magnetic anomaly signal. *Geophysical Research Letters, 20*(17), 1879–1882. https://doi.org/10.1029/93GL01485

Trulsvik, M., Myklebust, R., Polteau, S., & Planke, S. (2011). *Geophysical Atlas of the East Greenland Basin: Integrated seismic, gravity and magnetic interpretation*. Volcanic Basin Petroleum Research AS, TGS-NOPIC Geophysical Company.

Vamvaka, A., Pross, J., Monien, P., Piepjohn, K., Estrada, S., Lisker, F., & Spiegel, C. (2019). Exhuming the top end of North America: Episodic evolution of the Eurekan Belt and its potential relationships to North Atlantic plate tectonics and Arctic climate change. *Tectonics, 38*(12), 4207–4228. https://doi.org/10.1029/2019tc005621

Vormann, M., & Jokat, W. (2021). Crustal variability along the rifted/sheared East African margin: A review. *Geo-Marine Letters, 41*(2), 19. https://doi.org/10.1007/s00367-021-00690-y

Vorren, T. O., Richardson, G., Knutsen, S. M., & Henriksen, E. (1991). Cenozoic erosion and sedimentation in the western Barents Sea. *Marine and Petroleum Geology, 8*(3), 317–340. https://doi.org/10.1016/0264-8172(91)90086-g

Voss, M., & Jokat, W. (2007). Continent-ocean transition and voluminous magmatic underplating derived from P-wave velocity modelling of the East Greenland continental margin. *Geophysical Journal International, 170*(2), 580–604. https://doi.org/10.1111/j.1365-246X.2007.03438.x

Werner, S. (1955). Interpretation of magnetic anomalies as sheet-like bodies. *Sveriges Geologiska Undersökning, Series C, Årsbok, 43*(6).

Whitmarsh, R. B., Pinheiro, L. M., Miles, P. R., Recq, M., & Sibuet, J.-C. (1993). Thin crust at the western Iberia Ocean- Continent transition and ophiolites. *Tectonics, 12*(5), 1230–1239. https://doi.org/10.1029/93tc00039

Zarasyskaya, Y. A. (2017). Segmentation and seismicity of the ultraslow Knipovich and Gakkel mid-ocean ridges. *Geotectonics, 51*(2), 163–175. https://doi.org/10.1134/a0016852117010095