Integrated analysis of seismic, gravity and magnetic data on the outer Møre margin offshore mid-Norway

- to better constrain the deep basin configuration

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Abstract

Potential field data (gravity and magnetic) provide crucial information used to determine basement and deep basin configurations on the Norwegian Continental Shelf (NCS), and is underused in the exploration community. In this thesis both regional and detail studies includes the use of potential field data to better constrain the deep basin configuration on the NW Møre Margin beneath the volcanics. The work is based on the integrated SGM (seismic-gravity-magnetic) method. The main idea is based on high-pass filtering the potential field data based on the depth of investigation, before converting them to pseudohorizons that are loaded into the seismic workstation allowing for a fast joint interpretation in a sectional view. The pseudohorizon can also be draped over interpreted horizons allowing for a joint interpretation in 3D or map view. The qualitative interpretation is then tested quantitatively by XField, a new potential field modelling software. XField allows for interactive potential field modelling based directly on interpreted time sections on the seismic workstation, enabling a direct link between the quantitative interpretation and the qualitative integrated SGM method. Six horizons have been traced on a grid of 2D regional seismic lines, including Intra Miocene, Base Eocene, Top Cretaceous, Intra Cretaceous, Base Cretaceous (BCU) and Basement. In the areas of poor imaging by the seismic reflection data, the interpretation has been pushed by integrating 50 km high-pass filtered Bouguer and magnetic residual anomalies (both in section and map views) and crustal scale velocity models based on deep seismic refraction (OBS) data. The OBS velocity models are displayed in both depth and time for improved correlations to the seismic reflection profiles. The passive margin formation and development are interpreted to result from periodic extensional regimes, and do not support a passive margin where exhumed serpentinised mantle is present. The large-scale detachment-faults is however considered important in the pre-breakup configuration of the margin. A case study of the newly discovered Kolga High, identified based on 2D and new 3D seismic data combined with the complex pattern of high amplitude magnetic anomalies, and rounded positive Bouguer gravity anomalies, have been performed. 2D potential field modelling in XField suggests a non-igneous origin for the sub basalt structures on the outer Møre margin, and
a link between the magnetic anomaly responses with the thickness of the overlying basalts. The modelling also suggests that the gravity anomaly is from a deeper source, likely a deep-seated structural high at the outer margin.
Preface

This thesis is submitted to the Department of Geosciences, written for the University of Oslo in the field of Geophysics. It accounts for 60 ECTS, equivalent of one years study.

The study was conducted with supervisors Jan Inge Faleide and Sverre Planke.

The geophysical data sets used in this thesis (seismic, gravity and magnetic) were provided by TGS.
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Lastly, I will thank TGS for providing me with all the data, making this thesis possible to complete.
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1 Introduction

This thesis is focused on the mid-Norwegian margin located between 62°N to 69°N, with the main segments being the Møre and Vøring margins. Blystad et al. (1995) described the region as seismically inactive, and that it has been affected by an orogenic collapse followed by a long period of extension and rifting, culminating in continental breakup. The study area is located at the northwestern parts of the deep Cretaceous Møre Basin including the Møre Marginal high (Figure 1.1, marked in orange). The selection of the study area was based on areas where the conventional seismic data were not sufficient to give a full geological interpretation, as issues arise from blanketing effect resulting from Eocene extrusives and intrusives (Brekke, 2000).

The geological evolution of the mid-Norwegian margin can be linked to the tectonomagmatic evolution of the NE Atlantic margins. After the collapse of the Caledonian Orogeny, the margin experienced a period of c. 350 m.y. of extensional deformation and sediment basin formation until breakup in Paleocene - Eocene transition (Blystad et al., 1995; Skogseid et al., 2000; Faleide et al., 2008). The extensional regime can be described by three main rifting episodes; Permian - Early Triassic, late Mid-Jurassic - Early Cretaceous and Late Cretaceous-Early Eocene. The deep Cretaceous basins, Møre and Vøring, were formed as a result of such a long tectonic activity (Brekke, 2000).

To unravel the structures and deep basin configuration masked by the high impedance intrusions and extrusives, several different geophysical data sets will be applied using the integrated SGM (seismic-gravity-magnetic) workflow from Trulsvik et al. (2012). To validate the integrated interpretation a new potential field modelling software, XField, will be introduced.
The aim of the thesis is to unravel the deep structures and the configuration of the outer Møre Basin using the integrated SGM (seismic-gravity-magnetic) approach on selected key 2D seismic profiles; to discuss the origin and nature of the outer part in a passive margin evolution context. The benefits, limitations and uncertainties of the interactive potential field modelling software XField will also be addressed.

Figure 1.1: Regional map over the NE Atlantic modified from Funck et al. (2017), with the study area marked in orange.
2 Geological setting

The mid-Norwegian continental margin is an area offshore western Norway spanning from 62°N to 69°N in latitude (figure 2.1). This area is described as a passive rifted margin (Blystad et al., 1995). This type of margin is formed when continents break up and become separated by an opening ocean, with the oceanic and continental plates being connected. This creates a seismically inactive margin.

The mid-Norwegian continental margin consists of three different segments. The segments are divided based on NW-SE lineament patterns. The tectono-magmatic history of the margin is complex. Starting with lithospheric extension leading up to the continental plate breakup and the creation of the Norwegian-Greenland Sea. Followed by an increase in igneous activity, subsidence and maturation of the margin (Faleide et al., 2008, 2015).

In this chapter the NE Atlantic regional setting will be summarized to create the framework for the mid-Norwegian margin, followed by the geological evolution of the area. Then the tectonic, structural and stratigraphic elements forming the mid-Norwegian margin will be described.

2.1 NE Atlantic geological setting

To get a better understanding of the Norwegian continental margin, it would be of interest to look at the regional geological setting. That involves the evolution of the NE Atlantic margin.
The mid Norwegian margin is part of a larger geological system referred to as the NE Atlantic margin. Conjugate margins bounded by oceanic crust on either side, and extends approximately 3000 km from the border of the Rockall Trough and Porcupine Basin to the western Barents Sea (1.1). The margin consists of several repeated patterns of deep depocenters from Cretaceous-Cenozoic age trending in the NE-SW direction (Doré et al., 1999; Tsikalas et al., 2012).

The NE Atlantic margins geological history started at the end of the Caledonian orogeny. This was followed by approximately 350 m.y of extensional deformation, that culminated when the NE Atlantic opened at the Paleocene-Eocene transition. This extensional regime also created much of the basins that are seen in this region today. Skogseid et al. (2000) divided the geological evolution of the area into three
GEOLOGICAL SETTING

The Norwegian continental margin phases: Late Paleozoic - Early Mesozoic, Late Jurassic - Early Cretaceous and the last being Late Cretaceous - Early Tertiary.

Permian - early Triassic was a phase where the extensional regime created large half-graben basins around the Caledonian collapse. The basins were then filled with thick intra continental deposits. There is also evidence for a connection with the Arctic rift system between Norway and Greenland, as structures oriented N-S interfered with the NE trends of the Caledonian (Skogseid et al., 2000).

The Late Jurassic - Early Cretaceous rift phase is considered to be the precursor to the gradually developing continental separation in the southern part of North Atlantic. This was also the rifting phase where the deep rift basins were created in the area between the Rockall Trough to the SW Barents Sea. (Faleide et al., 2008).

Late Cretaceous - Early Tertiary was a rift phase that lasted around 20 m.y., that led to the continents separating with seafloor spreading beginning just after. A 300 km wide zone associated with lithospheric thinning and post breakup subsidence were the result of this rifting phase. There was also significant igneous activity during this rifting phase (Skogseid et al., 2000; Faleide et al., 2008).

2.2 The Norwegian continental margin

The evolution of the margin can be divided into three epochs. With the first epoch being the closing of the Iapetus Ocean during the Caledonian Orogeny in Silurian to Early Devonian time. The second epoch span from Late Devonian - Paleocene. This time interval was dominated by episodic extensional deformation and can be divided into three main rifting phases.

The Permian - early Triassic phase is considered as a reactivation of Caledonian basement trends. The middle Jurassic - early Cretaceous rifting phase was mostly a non-magmatic event, but had elements of extreme crustal thinning. The extension direction also changed during this rifting period from ENE-WSW to NW-SE, which is the direction perpendicular to the coastline of today. The last rifting phase in
late Cretaceous - early Eocene resulted in complete lithospheric separation. The margins were heavily affected by the igneous activity during breakup, resulting in a distinct imprint on margins with both extrusive and intrusive magmatism at various crustal levels like sills and seaward dipping reflector (SDR). The last epoch of the tectonic history is an ongoing seafloor spreading between to the continental plates that started in earliest Eocene up to the present day (Blystad et al., 1995; Faleide et al., 2008; Theissen-Krah et al., 2017).

The Norwegian continental margin consists of a shelf and slope, with both varying in steepness and width. The edge of the shelf and the slope are clearly defined with the water depth rapidly increasing at these locations. Compared to its conjugate margin, the Greenland margin, they have different shelf symmetries (figure 2.2). The Norwegian shelf is wide and gets narrower northwards, while the Greenland shelf is the opposite. There are also some complexity in the Norwegian shelf that is not found on the Greenland shelf, as parts of the slope is broken by large bathymetric features like the Vøringen Plateau (Tsikalas et al., 2012).

2.2.1 Møre margin

The Møre margin consists of the Møre Basin and the Møre Marginal High (figure 2.5), which reflects different tectonic events. With the Møre Basin reflecting the post-Caledonian extension and the Møre Marginal High the Early Tertiary magmatism. The margin is characterized by its narrow shelf and gentle-wide slope down the wide Møre basin (Mjelde et al., 2009; Faleide et al., 2015). The Møre margin is bounded by the Jan Mayen Lineament in the NE, the Early Eocene oceanic crust to the NW and the Møre Trøndelag Fault Complex (MTFC) to the SE (Mjelde et al., 2009).
Figure 2.2: Present day margin configuration, with the influence of different rifting periods marked. Retrieved from Faleide et al. (2015)
The Møre Basin is one of the main structural elements on the Møre margin (figure 2.3), as the Vøring Basin is on the Voring margin. The Møre Basin is bounded by the northern termination of the Tampen Spur in the south, and the Jan Mayen Lineament separates it from the Vøring Basin in the north. The Cretaceous onlap towards the mainland bound the basin in the east, while the Faroe-Shetland Escarpment along the Møre Marginal High is the western bound of the basin (Blystad et al., 1995). The formation of the basin took place during a shift in stress direction to NW-SE in the late Middle Jurassic - Early Cretaceous rifting phase (figure 2.2). As a result of the immense extension and thinning of the crust. This led to a differential subsidence in the basins creating local highs and sub-basins. By mid Cretaceous these basins were filled with a thick sequence of fine grained clastic sediments (Faleide et al., 2008). The inner flanks of the basin are steeply dipping with
the crystalline crust diminishing from < 25 km down to < 10 km (Kvarven et al., 2014; Faleide et al., 2015). The formation of the Møre Basin is described above, as the Vøring and Møre basins share the formation history (figure 2.4). But the Møre basin has been tectonically stable and experienced passive subsidence since the late Middle Jurassic - Early Cretaceous rifting period (Blystad et al., 1995).

**Figure 2.4:** Tectonostratigraphic chart for the Vøring and Møre basins, modified from Brekke, H. and Dahlgren, S. and Nyland, B. and Magnus, C. (1999).

The Møre Marginal High is described by Blystad et al. (1995) as an area situated between the Faroe Plateau and the Jan Mayen Fracture Zone, and is bounded by the transition to “normal” oceanic crust and the Faroe-Shetland escarpment in the west and east respectively. The structure is divided into two parts based on the composition beneath the top Eocene lava flow. The western part of this divide is underlain by seaward-dipping reflector sequence where the lava pile increase westwards. This is thought to be the upper part of thick oceanic crust. The eastern part
GEOLOGICAL SETTING

The Norwegian continental margin consist of a 15-100 km wide zone west of the Faroe-Shetland Escarpment, where the reflectors are subparallel to the uppermost Eocene lava and poorly continuous. Continental or transitional crust is most likely to be lying underneath this western area (Blystad et al., 1995; Breivik et al., 2006).

The Faroe-Shetland Escarpment is located between the Faroe Islands and the Jan Mayen Fracture Zone. It is bounded by the Møre Marginal High in the west and to the Møre Basin in the east. The structural style varies along strike, as the southern part has a lobate edge and resembles the front of a volcanic flow and/or fault front, while the northern part is more linear and reflects a interrelationship between a volcanic edge and a deeper structural setting (Blystad et al., 1995).

**Figure 2.5:** Seismic cross section over the Møre margin, modified from Blystad et al. (1995). Location of the line can be seen in figure 2.3. (M-line)

2.2.2 Vøring margin

The Vøring margin is around 500 km wide, and is composed of the Trøndelag Platform, Halten and Dønna terraces, Vøring Basin and the Vøring Marginal High from southeast to southwest respectively (figure 2.6) (Faleide et al., 2015). The margin is characterized by both extrusive and intrusive magmatism reflecting the rifting and continental breakup. The margin is considered to be the magmatic end member of
GEOLOGICAL SETTING

The Norwegian continental margin rifted passive continental margins (Ren et al., 2003; Faleide et al., 2015; Zastrozhnov et al., 2018)

The Vøring Basin is one of the main elements on the mid-Norwegian continental margin. The basin can be divided into east and west provinces by the Fles Fault Complex (figure 2.7). The eastern part of the basin comprises structures like the Træna and Rås basins and the Halten and Dønna terraces. The western province consist of Vigrid and Någrind synclines, Hel and Fenris grabens, Nyk and Utgard highs and the Gjallar Ridge. The north and south bounds of the basin is defined by two lineaments, the Bivrost lineament and the Jan Mayen lineament respectively. Another important lineament is the Surt lineament, dividing the basin in the direction that is parallel to the Jan Mayen and Bivrost lineaments (figure 2.3) (Blystad et al., 1995).

The Vøring Basin is flanked by paleo highs and platforms and the elevated mainland. The structures that make up the paleo high and platform are the Vøring Marginal High bounded by the Vøring Escarpment. The Vøring Marginal High is pre-dominantly a thick layer of basaltic Lower Eocene lavas overlying an unknown substrate (Breivik et al., 2009), while the Vøring Escarpment is the transitional terrace between the the Vøring Marginal High and the Vøring Basin (Abdelmalak et al., 2019). These same trends can also be seen along the Møre Basin, with the basin being constrained by flanking paleo highs and the elevated mainland (Brekke, H. and Dahlgren, S. and Nyland, B. and Magnus, C., 1999).

The Trøndelag Platform consists mostly of northwest dipping strata that is horizontal and bed parallel that covers an area of roughly 50 000 km$^2$ (figure 2.6) (Blystad et al., 1995). The platform has been tectonically stable since Jurassic times (Faleide et al., 2008), and Breivik et al. (2010) suggest no lower crustal bodies beneath the platform.
**Figure 2.6:** Seismic cross section over the Vøring area, modified from Blystad et al. (1995). Location of the line can be seen in figure 2.3 (F-line).

**Figure 2.7:** Seismic cross section in the southern Vøring crossing over to outer Møre margin, modified from Blystad et al. (1995). Location of the line can be seen in figure 2.3 (J-line).
2.2.3 Lineaments

The Jan Mayen lineament is the landward prolongation of the Jan Mayen Fracture Zone (JMFZ) (figure 2.3). The JMFZ has been active since the continental breakup in 55 Ma, and is a major offset in the NE Atlantic ridge (Berndt et al., 2001). It can be described as a broad feature with NW-SE oriented transverse faults in the generally N-S trending Fles Fault Complex. The lineament is defined as a sinistral shift of the basin margins including the escarpments, as well as the basin axis for the Møre and Vøring basins (Blystad et al., 1995).

The Bivrost lineament is described as a dubious lineament in Blystad et al. (1995), as Planke et al. (1991) defined the transition as an abrupt change in average depth to crystalline crustal velocities. With the velocities being around 6 km/s in the Vøring margin, and goes down to 2 km/s at the Lofoten-Vesterålen margin. As the Jan Mayen lineament, the Bivrost lineament is a landward prolongation of the Bivrost Fracture Zone. This fracture zone is important as the margins north and south of the fracture zone have very different margin physiography, breakup magmatism, structure and lithospheric stretching (Faleide et al., 2015).

The Surt lineament has the same orientation as the two other lineaments mentioned above, which is recognized as oceanic fracture zones. This is an important divide within the Vøring margin. As it divides the Hel Graben, Nyk High, Någrind Syncline, Utgard High and Træna Basin to the north from the Gjallar Ridge, Vigrid Syncline and Rås Basin to the south (Blystad et al., 1995; Ren et al., 2003). Mjelde et al. (2005) suggest that the Surt lineament acted as a Late Cretaceous - Early Tertiary adjustment features that controlled some of the magma influence in Early Tertiary rifting phase.

2.3 Stratigraphy

The stratigraphy on the mid-Norwegian margin is composed of different sequences, that have all been affected differently by the evolution of the tectonic framework in the northern North Atlantic over a time period of about 200 m.y (figure 2.8). The sedimentary environment has changed through time, as it started off in waning
orogenic forelands and incipients rift in an equatorial climate. This was followed by a stage dominated by continental rifting, where the plates drifted from equatorial to temperate climate. From this the environment changed to a passive subsiding continental margin in temperate to arctic climate as reflected by the present (Brekke et al., 2001).

The stratigraphy of the mid-Norwegian margin is well constrained by wells at the Trondelag Platform and Halten and Donna terraces, with the deepest stratigraphy being Permian to possibly Carboniferous. However in the deep basins the wells have only penetrated the Upper Cretaceous and parts of the Lower Cretaceous, so extrapolation of the stratigraphy down to the deep basin is necessary (Brekke, H. and Dahlgren, S. and Nyland, B. and Magnus, C., 1999).

The stratigraphy of the mid-Norwegian margin dates back to the Carboniferous with lacustrine and continental alluvial sedimentary settings dominating. That led to deposits characterised as red and yellow sandstones, grey siltstones and thin coal seams. This did not change until the Late Permian, where dramatic changes in the sedimentary environments occurred as the Norwegian Sea experienced transgression. Since the Norwegian Sea had open marine conditions and was between the warm arid North Sea and the humid tempered Barents Sea, it generated an environment suited for shallow marine carbonate platforms (Brekke et al., 2001).

The Triassic is characterized by deposits with continental fluvial and alluvial nature interchanged with some marine incursions from the north. As a result of the relative sea level rise in the transition from Late Triassic - Jurassic, the region changed into a shallow marine clastic shelf environment with oscillating deltas. By latest Middle Jurassic - Late Jurassic another sea level change occurred and drowned the area between Greenland and Norway. This flooding created open marine settings, and resulted in the deposition of rich organic shale layers interbedded within marine shales. In the Cretaceous the sea level had risen again, and reached its peak in Late Cretaceous. This made the facies pattern stable through all the Cretaceous period, where deep marine clays dominated the rifts and basins while the platforms and highs were dominated by shallow marine and coastal plain deposits. This was then followed by a transition from Cretaceous - Tertiary, where there was a change in
sedimentary environment due to the transition from continental rifting to passive margin setting. In Paleocene - Eocene the basin flanks, highs and domes surrounding the Vøring and Møre basins were eroded and locally deposited in shallow synclines within the basin. In the latest Cenozoic, Pliocene - Pleistocene, the glacial impact on sediment transport was compelling, with the eroded mainland as the main de-

**Figure 2.8:** Lithostratigraphic chart over the mid-Norwegian margin. Modified from Gradstein et al. (2010)
2.4 Deep structures

The mid-Norwegian margin experienced intense volcanics imprint with the Eocene breakup, which resulted in the deeper parts of the outer Møre margin being masked by the high impedance contrast between sediments and volcanics. The deep basin configuration is important to constrain when trying to understand the formation and evolution of passive margins. There exist competing models for the formation and evolution of the NE Atlantic conjugate margins. Osmundsen et al. (2016) and Peron-Pinvidic et al. (2013) introduces a general model for the formation and evolution of passive margins based on the magma-poor Iberian margin (figure 2.9). The authors suggest a continuous rift development, where the extension is controlled by large detachments faults bounding the margin into different domains; proximal, necking, distal, outer and oceanic. Where each domain has a characteristic deformation process; stretching, thinning, hyperextension, exhumation, magmatic and oceanization respectively.

Figure 2.9: Key figure illustrating the rift development for passive margin. Retrieved from Peron-Pinvidic et al. (2013)
Theissen-Krah et al. (2017) and Faleide et al. (2015) prefer the formation and evolution of the passive margin to be separated into distinct extensional phases, and argue against the continuous nature of the model from Peron-Pinvidic et al. (2013) and Osmundsen et al. (2016). The authors agree that large detachment faults may have been important for pre-breakup development of the passive margin. But they favor a model where the rift phase in Late Jurassic - Early Cretaceous accommodate for half the extension between the conjugate margins, with the rest located between Greenland and the Jan Mayen Micro Continent (JMMC) (Theissen-Krah et al., 2017). The passive margin development from breakup until now can be seen in figure 2.10. It illustrates that the line of breakup was oblique with respect to the pre-existing Mesozoic rift basins.

![Figure 2.10: Evolution of the NE Atlantic conjugate margins at three different stages; a) present, b) Chron 13 and c) Breakup. Retrieved from (Faleide et al., 2015).](image)

In figure 2.11, the conjugate margins are displayed at present configuration in figure 2.11a). The time of breakup is illustrated in 2.11b), while the 2.11c) show the pre-breakup configuration without the magmatic underplating Mjelde et al. (2016). The upper panel illustrates the the accretion of oceanic crust is not symmetric between the conjugate margins.
The main difference between the suggested models is whether the lower crustal body is related to exhumed mantle or thinned continental crust. The structures are often divided into outer and inner LCB, or proximal and distal proposed by Nirrengarten et al. (2014). These are characteristic for the margin, and the LCB’s is often characterized by having P-wave velocities above 7 km/s (Abdelmalak et al., 2017). The LCB’s have been observed below the marginal highs and the volcanic province of the margin, where the general trend is that it thickens beneath the intrabasinal highs. There are several interpretations of these LCB’s. Gernigon et al. (2004) presented five different suggestions for the formation of the LCB’s. This include a massif gabbroic complex (Mjelde et al., 1998, 1997), high velocity sills (Berndt et al., 2000), melted continental crust (Lundin and Doré, 1997), serpentinised mantle (Osmundsen et al., 2016, 2013; Osmundsen and Ebbing, 2008; Ren et al., 1998) and retrograde/high grade rocks (figure 2.13). The first model is suggested to consist of gabbroic to olivine accumulates that underplated the lower crust as a result of the volcanic event during the breakup phase (Mjelde et al., 1998, 1997). The second model arise from the complexity of the respons from the mafic lower crust, indicating that it might consist off inherited crust intruded by scattered high velocity sills (Berndt et al., 2000). The third model displays a lower crustal diapirism as a result of a deeper magma chamber inducing melted continental crust
Lundin and Doré, 1997). The fourth model accounts for a LCB that resulted from a non-magmatic environment, and is suggested in Ren et al. (1998) to consist of serpen tinised mantle. Authors like Osmundsen et al. (2016) and Peron-Pinvidic et al. (2013) support this interpretation of the LCB. The last interpretation of the LCB, comes from high-grade rocks related to a post-Caledonian metamorphic core complex (Gernigon et al., 2003).

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![Figure 2.12](image)

*Figure 2.12: Interpretation of line MNR11-90277, and illustrates an interpretation of the deep structures on the Møre margin. Modified from Theissen-Krah et al. (2017).*

On the mid-Norwegian margin there are some peculiar, but prominent high amplitudes reflectors in the deeper part of the seismic section, around 7-10 s twt (Gernigon et al., 2003; Abdelmalak et al., 2017). The authors have referred to them as "T-Reflection", and is described in Abdelmalak et al. (2017) as sometimes discontinuous rough multiple reflection while attain the shape of a single smooth reflector. The "T-Reflection" has a varying geometry, contact relationship and amplitudes. Abdelmalak et al. (2017) also suggest that the "T-reflector" corresponds to the top LCB in approximate 50% of the area on the Vøring margin. Abdelmalak et al. (2017) favor a model similar to interpretation b) in figure 2.13, as the authors propose a model where the deep structures consists of preserved upper continental crust with lenses of middle- to lower crustal units affected by deep intrusions. Theissen-Krah et al. (2017) favor a model where the outer LCB corresponds to an intruded inherited lower crust and/or magmatic underplating. Other authors describes the LCB’s as an exhumed serpen tinized mantle (Osmundsen et al., 2016; Peron-Pinvidic et al., 2013; Osmundsen and Ebbing, 2008).
**Figure 2.13:** LCB models. Retrieved from Gernigon et al. (2004).
3 | Data and Method

In this chapter the different geophysical data sets will be introduced, in conjunction with the methods being used. Firstly the geophysical data types will be presented individually, followed by the integrated SGM (seismic-gravity-magnetic) method. Lastly the new interactive approach/tool for the potential field modelling, XField, will be introduced.

3.1 Data

The data used in this thesis include a cluster of 2D regional seismic lines (figure 3.1), 3D seismic lines for detail study, gravity and magnetic data provided by TGS. Published OBS transects (figures 3.1 and 5.1) was also incorporated (Kvarven et al., 2014; Mjelde et al., 2003; Raum, 2000).

The reflection seismic lines consist mostly of a selection of lines from different compilations; MNR11, GMNR94 and VMT95. The MNR11 was the seismic compilation with the best resolution with depth, compared to the other data sets. MNR11 records down to 10 twt. VMT95 and GMNR94 have similar attributes, with poor resolution with depth, and i recorded to 12 s and 14 s twt respectively. The gravity and magnetic maps have been high-pass filtered with a cut-off wavelength of 50 km. The Bouguer map have also been corrected with a density for the water layer equal to 2.2 g/cm³.
3.2 Methods

3.2.1 Reflection (MCS) data

Seismic data is the main geophysical data type for imaging the subsurface. There exist several seismic data types, examples include reflection and refraction seismic that are going to be used in this thesis. The reflection seismic mostly used, and can be expressed both in 2D and 3D. Where the 2D is often used in unexplored areas to get a quick overview, whilst 3D seismic is often used where a more detailed analysis
is wanted (Nelson et al., 1999). In this thesis there will only be regional 2D seismic lines.

The seismic data for petroleum exploration is mostly gathered from marine acquisition (figure 3.2). To acquire the seismic data, three components is needed; an input source, arrays of receivers and recording tools. The source is designed to fire a pulse that contain predefined entities such as a frequency band, maximum amplitude, phase and total energy. The response from the subsurface is then picked up by the array of receivers and recorded. Other important aspects of the acquisition is the survey planning and orientation of sources and receivers. (McQuillin et al., 1979).

The processing of seismic data is very dependent upon the field acquisition parameters, but a normal processing sequence includes deconvolution, CMP stacking and migration (Yilmaz, 2001). The goal of the processing is to enhance the subsurface image, by removing unwanted signals (noise).

The seismic section generated after processing is the one used by most geoscientists. It is used for interpretation of horizons and faults, and to further develop a geological model (Telford et al., 1990). As it is a method based on impedance contrasts between layers, there are some challenges concerning the resolution with depth below an extremely high impedance contrast. It acts as a barrier for the
seismic energy, masking the structures below. In these areas the other geophysical methods, gravity and magnetic, can act supplementary to the seismic data and give crucial information for further interpretation.

3.2.2 Seismic refraction (OBS) data

Ocean bottom seismic (OBS) data are based on the refraction method (figure 3.3), with waves travelling along surfaces, making it possible to extract the velocities of the layers. This technique is used to map deep sedimentary and crustal structures underneath areas where conventional seismic has imaging issues, like a volcanic environment with intrusions and lavas (Mjelde et al., 2003). The difference between the OBS and normal seismic is that the source and receivers are decoupled and that the receivers are located on the seabed (figure 3.3). The receivers consist of three orthogonal geophones and one hydrophone to give a 4C recording system. The benefit of this setup is that the P-wave will be recorded by the vertical geophone and the hydrophone, while the converted S-waves will be recorded by the horizontal geophones. This can be used to get longer offset reflection data and refraction data (Yilmaz, 2001). The OBS profiles used in this thesis is Profile 1 and Profile 2 in Kvarven et al. (2014), L8-96 profile from Raum (2000), Line 14-96 from Mjelde et al. (2003) and lines 2-99, 3-99 and 4-99 from Mjelde et al. (2009). In this thesis the profiles will be used to constrain the velocity of the subsurface and guide the velocity picks of the different sedimentary sequences, as well as providing constraints on LCB and Moho.
3.2.3 Gravity

Gravity is the attraction force between two objects, and is often measured as a vector (Fowler, 1990). In our case, earth science, the vertical component of the vector is the direction of interest. A gravity anomaly originate from horizontal changes in densities (figure 3.4) (Nettleton, 1971).

To obtain the gravity anomaly the modelled gravity measurement is subtracted from the observed. The gravity anomaly is dependent on the size of the mass, m, and the inverse square distance, r, to its center of mass, and is used to describe subsurface density variations. The anomaly is ambiguous as both the mass, m, and distance, r, is not measured accurately (Fowler, 1990). Another aspect of the anomaly is that it
is non-unique, in that sense a shallow body can have the same response as a smaller heavier body that is buried deeper (figure 3.5).

*Figure 3.5:* Source of the displayed gravity anomaly. Orange circle displays maximum depth of the anomaly but shallower geometries can give same response (blue and purple). Modified from Nettleton (1971)

It is possible to distinguish between shallow and deep by looking at where the energy is concentrated. With shallow anomalies having high energy content in the short wavelengths (Nettleton, 1971).

The offshore gravity data are most commonly measured by ships simultaneously with seismic lines, but other methods can be used; altimetry measurements, satellites and airborne surveys. In areas where the ship measurements are poorly covered, satellite data can be merged to create a better coverage. The gravity field can be acquired by deriving the equipotential surface (sea surface).
The processing of marine gravity profiles takes into account all the unwanted effects that affect the data set. These effects include noise, coriolis force of the moving vessel, instrument drift, diurnal variations. This can be removed by employing Eötvös correction, drift corrections, tie to absolute gravity stations, removal of reference geoid and noise to extract the free-air gravity anomaly (Nettleton, 1971).

The gravity maps generated are Free-air and Bouguer unfiltered anomaly with 400-/200-/100-/50-km high pass filters. These maps are used for identifying the presence of basins, ridges and crustal structures. The different filtered maps give different depth of investigation (Fowler, 1990). Where the > 400 km high-pass cut-off anomalies describes lower crustal terranes, uppermost mantle density variations and large bathymetric variations. The 100-400 km high-pass cut-off anomalies describes regional sedimentary basins and structural highs, large, deeply buried evaporite deposits, bathymetric features and large volcanic complexes. Lastly the <100 km high-pass cut-off anomalies describe evaporite diapirs and shallow domes, shallow fault scarps, local bathymetric features, shallow basement ridges and grabens/basins and volcanic complexes (Trulsvik et al., 2012)

For the gravity data, the anomalies will be characterized and described using a 50 km high-pass filtered Bouguer residual maps, as the main scope of the thesis is on the deep basin configuration, which includes structures that are <10 km from the surface. The descriptonal system will be based on which anomalies that coincide with each other. Anomalies will be characterized based on a descriotional system concerning numbers and letters. Positive gravity anomalies is described using a letter first, followed by a number. The letters are assigned at random, and do not correlate with any structural elements. If there are several anomalies with same response in close proximity, they receive the same lettering but different number. To distinguish between positive and negative anomalies, the positive anomalies have the letter frist followed by a number, opposite it true for the negative anomalies.

### 3.2.4 Magnetic

The earth’s magnetic field is made up of three parts; the core field generated by electric currents in the core of the earth, an external field generated from fluctuations
in radiation from the sun and the crustal field generated by magnetic susceptibility variations in the earth’s crust and upper mantle. The magnetic is measured as a total vector that is decomposed into horizontal and vertical components (3.6). The angles (inclination and declination) between the total and the decomposed vectors are important parameters for defining the magnetic response. The strength of the magnetic field, B, is given by the inverse cubed of the distance, r, from the dipole. Rocks ability to be magnetized by an external magnetic field is called magnetic susceptibility (Telford et al., 1990), which is an important material property. It is most common that volcanic rocks containing minerals such as hematite and magnetite to be magnetized, but some sedimentary rocks can be weakly magnetized by chemical or depositional processes (Nettleton, 1971).

![Diagram of magnetic field components](image)

**Figure 3.6:** Magnetic field total field (T), and the horizontal (H) and vertical (V) components. Inclination (i) is the angle between the Total and Vertical, while the Declination (d) is the angle between the Total and Horizontal. Modified from Nettleton (1971).

The acquisition of offshore magnetic data is done by either towing a magnetometer behind a ship or an airplane (Telford et al., 1990). The standard for the present day acquisition is to use airplanes, as it is much cheaper and more reliable even if the resolution is a bit lower. The magnetic data is heavily dependent on the diurnal variations, which gets stronger with latitude. A base station is then necessary to account for the diurnal and instrumental drift effects to get accurate results (Net-
As with the gravity anomaly, there are effects in the data set that need to be processed. With the magnetic anomalies the processing sequence includes editing and removal of noise, diurnal and instrumental drift corrections (Telford et al., 1990).

The magnetic maps that are generated are Total anomaly maps (Z) that are filtered using 200-/100-/50-km high-pass filter. The magnetic data are most useful for determining the presence, trends and depth of rocks commonly known to contain magnetic minerals such as mafic volcanic rocks and metamorphic terranes. The different high-pass filtered maps give, as the gravity, different depth of investigation. Where the 100-400 km high-pass cut-off magnetic anomalies describe variations in continental crystalline crustal terranes and sea floor spreading. The < 100 km high-pass cut-off magnetic anomalies describes volcanic complexes, shallow basement ridges, basement activating faults, intra-sedimentary faults and fluid seeps. In contrast to the gravity, there are no > 400 km high-pass cut-off magnetic anomalies Trulsvik et al. (2012)

As for the gravity, the magnetic data used in this thesis is the 50 km high-pass filtered map. The descriptonal system will be based on the same system used for the gravity, where to positive anomalies are assigned a letter first followed by a number and vice versa for the negative anomalies. The differentiate between the potential field anomalies, the magnetic is referred to with a capital letter while the gravity has a lower case letter.

### 3.2.5 Integrated/interactive SGM approach

Before the integrated SGM approach can begin, there needs to be a first hand interpretation of every data type (green boxes in figure 3.7). For the seismic, key horizons for the area need to be identified and interpreted confidently as far as possible. For the gravity and magnetic data, each anomaly in the study area needs to be described (shape, size, striking direction, value) and put in a descriptonal system. Then identify areas where the responses for the gravity and magnetic are positive-positive, negative-positive and negative-negative.
The seismic-gravity-magnetic (SGM) is an interpretation method that takes all the three different datasets into account on the same workstation (yellow box in figure 3.7), and commonly improves the quality and efficiency of the seismic interpretation. This is beneficial as the seismic is very expensive to produce and depth resolution is commonly limited. The other geophysical data sets then give complimentary information, such as subsurface density contrasts and magnetic susceptibility, at a lower cost. By implementing the other geophysical data sets, it is possible to better constrain the seismic interpretation. Examples of this includes constraints on basins and basement highs, geological trends that can be difficult to identify as a result of poor seismic and lithology discrimination. To make it an integrated workstation the gravity and magnetic data can be loaded as pseudo-horizons onto the seismic.

To transform the magnetic and gravity anomaly maps into pseudohorizons, it is possible to use this formulation from Trulsvik et al. (2012):

\[ \tau_{ps} = \tau_0 - z_{scale} \times a \]

(3.1)
In equation 3.1 the $\tau_{ps}$ is the pseudo value in twt for the anomalies. $\tau_0$ corresponds to at which depth in twt the equilibrium is located. $z_{scale}$ is the scaling factor for the anomaly value, a. In this thesis $\tau_0$ is set to be 5000 ms and $z_{scale}$ is set to $\frac{1}{0.025}$ for the 50km high-pass filtered residual Bouguer anomaly map. For the 50km high-pass filtered magnetic maps, $\tau_0$ was set to 4000 ms, and $z_{scale}$ was set to $\frac{1}{0.08}$.

After the pseudohorizons are loaded into the seismic workstation, the integrated interpretation can begin. In this stage the interpretation of the seismic horizons is pushed further based on the trends in the pseudohorizons, as these horizons may give indications that there should be a basement high or a sub-basin in the low resolution part of the seismic. While acting as a method for pushing the interpretation in unknown areas, the integrated interpretation can also be used as a quality control for areas with high resolution and good stratigraphic control.

After the interpretation of the seismic section is satisfactory, potential field modelling can be done to test the validity of the integrated interpretation. As it is a non-unique modelling scheme, there may be some interpretation problems concerning model testing and sensitivity analysis. Such interpretation problems includes high density contrast boundaries, the nature of the subsurface lithology, depth, geometry and physical properties of complexes that are poorly imaged by seismic data. From the potential field modelling it is only possible to obtain density and magnetic susceptibility contrasts, rather than the absolute value.

Previously the potential field modelling have been decoupled from the SGM workstation, removing the seismic interpreter from the modelling scheme. In this thesis the modelling will be incorporated into the SGM workstation more interactively (figure 3.7) using the plug-in XField. The modelling software, XField, is developed by the company ARK CLS. It is a powerful geophysical modelling tool that allows the explorationist to analyse and smoothly integrate the potential field modelling into the integrated workstation. It has the potential to reduce both risk and cost in an exploration environment, by creating geological models that are more accurate and precise than previously possible.
The program is a plug-in for both Petrel and OpendTect, the latter will be used in this thesis. The workflow in XField can be seen in figure 3.8. It has an easy to navigate user interface, where the initial setup includes three panels for displaying the gravity gradient, gravity and magnetic data in the upper portion and a model window underneath. The panels display both the forward modelled responses and the measured gravity and magnetic data. The measured data can be loaded into XField as either a file or a horizon.

To begin making a model, the seismic section at focus need to be imported into XField. The temporal and spatial model parameters are taken by default from the seismic section. These model parameters can be altered at a later time if necessary. After the seismic section is in place, the interpreted key horizons from the integrated SGM approach can be imported. Then a model can be created by splitting up the initial model-body along the interpreted horizons. The final model consists of several bodies, where each body represent a stratigraphic unit. These bodies are then given velocity, density and magnetic susceptibility values. Where the velocity values are taken from OBS profiles, and the density values are taken from established density-velocity trends. The the magnetic susceptibility values are taken from measurements on outcrop samples. XField then creates a velocity model, that will be used for depth converting the time interpreted model. This operation is hidden from the user, and may induce uncertainties. After depth converting the model, XField calculates the gravity and magnetic forward response and displays it in the same panels as the observed gravity and magnetic data. To reduce the edge effect of a rectangular model, the model-bodies can be extended to infinity. If the forward calculated and observed responses do not match, the interactive nature of the plug-in lets the user quickly change the interpretation to get a better fit. It is also possible to display the mismatch or the response of a single body to help clarify which area needs more attention.
Figure 3.8: Workflow in the XField software
4 | Results

In this chapter the different geophysical data sets will first be interpreted separately, then followed by the integrated SGM interpretation and potential field modelling. This will be done on both the regional scale and a detail study of selected anomalies. For the seismic interpretation, key horizons will be traced with confidence until reaching areas with low resolution. Then the interpretation will be pushed using the integrated SGM method. To validate the pushed interpretation, potential field modelling will be done in XField.

4.1 Seismic

In this section, the different seismic data will be interpreted. For the multichannel seismic (MCS) profiles, horizons will be traced, to make boundaries for the geological model. The Moho/LCB geometry as well as the velocities that are going to be used for the later potential field modelling, will be interpreted from the OBS data.

4.1.1 Seismic reflection (MCS)

The quality of the seismic lines varies from poor to fairly good, which was mainly a result concerning the age of the seismic data and the focus area of this thesis. The best quality data set was the MNR11 survey, with good resolution for almost the entire section with exceptions below the Eocene intrusions in the outer Møre Basin. Whilst the VMT95 had somewhat good coverage of the top sedimentary strata, the seismic survey lacked resolution with depth, making it very difficult to determine any basement trends. All the seismic surveys had different twt values, where MNR11 has 10 twt, VMT has 12-twt and GMNR94 has 14 twt. Based on the abovementioned factors, two key transects have been picked for the regional scale modelling. These key transects include the lines Transect 1 (figure 4.1) and Transect 2 (figure 4.3). From this, nine selected key horizons were traced to get stratigraphic constraints of the area and for the later modelling in XField. The selection was based on which horizons that was “area known” or had a significance concerning the regional geology, and was influenced by the interpretation from Theissen-Krah et al. (2017)
The nine key horizons that were traced includes; Seabed(pink), Intra Miocene(yellow), Base Eocene(orange), Top Cretaceous(light green), Intra Cretaceous(green), Base Cretaceous Unconformity (BCU)(dark green), the basement (blue), top basalt(brown-red) and base basalt(purple).

The first horizon interpreted on Transect 1 line was the Seabed, which was recognized as the first strong response on the seismic section and could be traced for the whole section. The Intra Miocene, Base Eocene, Top Cretaceous and Intra Cretaceous were also interpreted based on relatively strong reflectors. With Base Eocene and Top Cretaceous interpreted along strong peaks, while Intra Miocene and Middle Cretaceous were interpreted along a strong through. Even though these horizons were traced along strong reflectors, every horizon terminated towards northwest when in close proximity to intrusions. The top basalt was traced along a strong peak, while the base basalt was traced along a strong through. Both were identified based on the seismic facies change across the boundaries. The last two horizons, BCU and Basement, could only be traced with confidence southeast of the lava front. It was however possible to see some wedgelike shapes under the lava front, to further push the interpretation of the Basement with some confidence.

\[\text{Figure 4.1: Transect 1 plain seismic. Data courtesy TGS}\]
As with Transect 1, the Seabed reflector was the first horizon interpreted on Transect 2 as the first strong reflector in the section. To get the location of the Intra Miocene, Base Eocene, Top- and Intra Cretaceous, a tie line (VMT95-010) was used between Transect 1 and Transect 2. Horizons BCU and Basement could not be traced on the tie-line as the resolution was too poor in the deeper parts. Similar interpretation strategy was implemented on Transect 2 as it was with Transect 1. The horizons Intra Miocene, Base Eocene, Top- and Intra Cretaceous were confidently traced until reaching intrusions or areas close by. The volcanics were traced out of place reflectors with strong amplitudes in close proximity to a change in seismic facies. The BCU was traced along a prominent reflector that coincides with the twt gap between Intra Cretaceous and BCU on Transect 1, making it a confident interpretation. The Basement reflector was picked where it seemed like sedimentary stratigraphy units onlapped the reflector, as seen with Intra and Top Cretaceous units (figure 4.2).
4.1.2 Seismic refraction (OBS)

For the purpose of potential field modelling the OBS data are divided based on which lines lie in close proximity to the MCS line of interest. The OBS transect used for guiding velocity picks for MCS line Transect 1, is profile 2 from Kvarven.
et al. (2014) and L8-96 from Raum (2000). Both lines are subparallel with respect to the MCS line, but none of the OBS lines are further than 30 km away in the perpendicular direction. These OBS lines have been converted from depth to time, as this suits XField best. The OBS data support the interpretation of layers thinning towards the northwest. From the OBS data, the velocities for the different stratigraphic units were obtained, and can be seen in Table 4.1. From Profile 2 the Moho boundary was identified based on the sharp transition in velocity, going from around 7 km/s to 8.1 km/s. The overall trend of the Moho geometry is that it dips in the southeast direction, and it is shallows under large basement highs. The shallowest point for the Moho is located at 8.5s twt under the More Marginal high, while the deepest point lie under the deepest part of the outer More basin northwest of the Vigra high, reaching 11s twt. Both OBS profiles indicate a lower crustal body in the MCS lines, this is indicated by the high velocities at about 7+ km/s. This LCB more or less follow the Moho geometry.

There are no OBS lines parallel or subparallel to the Transect 2, but several OBS profiles intersect the MCS line at different locations. The OBS profiles of interest includes three lines from Mjelde et al. (2009); Line 2-99, Line 3-99 and Line 4-99. There is also Line 14-96 from Mjelde et al. (2003). All these transects are published depth profiles, so a conversion to time was done for the intersecting points. The lines 2-99 and 14-96 have the intersection points very close to the edge of the OBS profile, which induce uncertainties regarding the velocities and location of the LCB/Moho boundaries. As the regional geology do not change abruptly in the space between the two MCS lines, the velocities for the stratigraphic units are assumed constant. The velocities of Transect 2 can be seen in Table 4.2.

4.2 Gravity

The residual gravity map that will be used in this thesis is the Bouguer 50 km high-pass filtered map, in courtesy of TGS. This map can be seen in figure 4.5. The gravity response does not have as clear separation between the zones as the magnetic map has. A more dominating feature in the gravity map is the structural trends of the basement, so the striking direction of the anomalies can be used to some extent
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Gravity

to group the different anomalies.

The anomalies $a_1$ and $a_2$ have a clearly different striking direction than the surrounding anomalies as they strike NW-SE while the overall striking trend of the adjacent anomalies is NE-SW. The anomalies are separated by a lower value part. Anomaly $a_1$ is also more subrounded and uniform than $a_2$.

Figure 4.5: 50km high-pass filtered Bouguer residual anomaly map. Data courtesy TGS

A prominent anomaly series is the $l_1$-$l_4$ anomalies, located in the southern Vøring Basin. This series consist of four strong positive anomalies that decrease in value northwards, the anomalies also vary in shape and size. $l_1$ have a rounded shape making it difficult to find strike direction. The anomaly also intersects with the anomaly trend of the $c_1$ anomaly. $l_2$ is a subrounded anomaly that strikes in the N-S direction. Anomaly $l_3$ is the largest anomaly with a more elliptical form, also striking in the N-S direction as $l_2$. $l_4$ has a similar shape as anomaly $l_1$, making it difficult to determine a striking direction. Another observation in the $l$-series is the abrupt change in anomaly pattern in between anomalies $l_2$ and $l_3$. Elsewhere the positive $l$-series seems to be a more or less continuous strong anomaly. The negative
l-series consist of anomalies 1l-3l, where all seem to have a moderate response even though the anomalies are different shapes. Anomaly 1l is shaped like a triangle, while 2l resembles a parallelogram striking NE-SW. Anomaly 3l has a subrounded form striking in NNE-SSW.

n1-n3 is another prominent anomaly series, that is parallel to the L series separated by the negative anomalies 1n-3n. The n1 anomaly is striking in the N-S direction and is located in the southern Vøring basin. It has an elongated form and a strong response. The next anomaly in the series, n2, has a complex shape and is the anomaly that has the weakest response of the series. It seems like its almost divided in the middle, but this can be an illumination effect(?). Even though the shape is complex, the overall striking direction is NNE-SSW. The n3 anomaly has a rounded shape with a strong response, and is striking in the NE-SW direction. The negative 1n anomaly is similar in shape, response and striking direction as anomaly 2l. Anomaly 2n has an elongated form and is striking N-S, and it seems to increase in value northwards. The last anomaly, 3n, has a rectangular shape striking NE-SW. It is the weakest of the negative anomalies in the l-series.

Anomaly t1 is a stand alone anomaly where the shape resembles the letter t. Where the upper part consists of an elongated anomaly that strikes NW-SE, and that the anomaly decreases in value from southeast to northwest. The lower part, striking NNE-SSW, decreases in both value and width as it gets closer to anomaly k3.

The K series consist of anomalies k1-k3, and are located on the Møre Marginal high. All the anomalies are complex in the shape and have a strong anomaly response. k1 has the most complex shape, as it is composed of a subrounded part and an elongated part. Where the subrounded part of the anomaly is striking N-S while the elongated is striking NE-SW. Even though the shape is complex, the value is homogeneous across the anomaly. The k2 anomaly changes character in the southwest direction, as there are some windows with lower values (yellow color). k3 can be described as a rounded rectangular shape that strike in the NW-SE direction. This anomaly is more similar to k1 with the homogenous anomaly response, rather than the varying response of k2.
In close proximity of the positive k anomalies, there are two different strong negative anomaly sets m and k. Where m consists of anomalies 1m-4m, and k consists of 1k and 2k. The k-series strikes in the NE-SW direction, it also gets a progressively stronger anomaly response in the northeast direction reaching values around -4 mGal. The transition between 1k and 2k is characterized by an elliptical anomaly with values around 0 mGal. The m anomaly series is characterized as some of the strongest negative anomalies in the map view, reaching larger values than -4 mGal. The overall shapes of the anomalies are round to subround.

Anomalies b1, b2 and e1 are all positive anomalies with a value of 3 mGal. In the same area there are also the negative anomalies 1e and 2e. The e1 anomaly is subrounded and a bit elongated in the NE-SW direction. It is weaker than the surrounding positive anomalies. The positive anomaly b1 has a rounded shape, making it difficult to determine the striking direction. The transition between anomalies 1b and 2b is not continuous and drops to around 0 mGal before reaching anomaly b2. This anomaly has similar shape and value as b1, but is smaller in size. Anomalies e1 and b1 is separated by the negative anomaly 1e. This anomaly is characterized as a strong (-3 mGal) elongated anomaly, that strikes in the NE-SW direction. Another negative anomaly in this region is 2e. The shape of this anomaly is subrounded with a NW-SE striking direction. In close proximity the anomalies v1 and v2 are present. These are strong positive anomalies with values towards 4 mGal. Both anomalies have the same elongated subrounded shape striking NE-SW, with v2 being the larger anomaly.

The anomaly series f, consist of anomalies f1-f4 (figure 4.5). The overall strike direction of the anomaly series is NE-SW direction, even though some individual anomalies strike in a different direction. The anomaly value also seem to be constant for all anomalies in the series. Anomaly f1 is striking in the E-W direction, and consists of two moderately strong rounded positive anomalies. In contrast anomaly f2 is striking in N-S direction, and can be described as elongated. Anomalies f3 and f4 have a smooth continuous transition rather than being separate anomalies like f1 and f2. f3 is wider in the northeastern part and gets narrower further southwest before connecting to f4. The overall shape of the negative anomalies 1f-5f is triangular, but every anomaly has its own shape. 1f is an elongated anomaly striking
RESULTS

E-W with values around -3 mGal. Anomaly 2f is a rounded anomaly that have same values around -3 mGal as 1f. The 3f anomaly has elliptical shape striking E-W, with a slight increase in value towards the east. Anomaly 4f, and this anomaly has the shape of an hourglass. It strikes NE-SW and has its highest values in the southwest of around -3.5 mGal. The last anomaly in the series is 5f, that is more round in the shape compared to the other negative anomalies in the f-series, and can possibly be a prolongation of anomaly 3f.

The c1 anomaly is one of the stronger positive anomalies in the study area/AoI with values at around 4 mGal. It has an elongated form that is slightly bent leftwards. Close to anomaly c1, anomalies 1c and 1h-3h can be identified. These are all strong negative anomalies with values around -4 mGal. 1c anomaly has the shape of the number eight striking E-W. The h-anomalies are the negative anomalies closest to the coastline. 1h and 2h are similar in shape and striking direction, where both is elongated anomalies striking NE-SW. While anomaly 3h has a more round shape.

4.3 Magnetic

The magnetic anomaly map that is used in this thesis is the 50 km high-pass residual map provided by TGS (figure 4.6). Inside this region the residual anomaly pattern is complex, and can divided into six different facies.
Facies 1 has the most complex anomaly pattern of all the six different facies, and corresponds to the magnetic response of the Møre Marginal high. The pattern is chaotic, with shapes varying from subround to elongated spread across the entire facies with a mixture of strong and weak anomaly responses. The weaker responses are located in the center of the facies, while the stronger responses are located northeast and southwest of the center. The boundary between this facies and facies 3 and 5 corresponds to the Faroe Shetland Escarpment, marking a clear change in anomaly pattern from the Møre Marginal high down into the Møre Basin. A1 corresponds to a strong positive anomaly (105 nT) striking NNE-SSW, with an almost rectangular shape surrounded by three strong rounded to subrounded negative anomalies (1A-3A). Those same negative anomalies form an outline that resembles a triangle around the A1 anomaly. The positive anomalies in the J-series (J1-J3) have a shape that resembles a ridge that encircles the above mentioned negative anomalies 1A-3A. The J-series has a value of around 40 nT, and has a somewhat stronger response northwards. Other anomalies in this facies include the U and negative J series. The J-series consists of the anomalies 1J and 2J, where both have an elongated shape with values around 100 nT. The strike direction is not the same, as J1 strikes more
N-S while J2 strike NNE-SSW. The U-series is located further down in the facies compared to the rest. These anomalies include U1, 1U and 2U. As the J-series, the negative anomalies in the U series have similar shape and value but different striking direction. Both 1U and 2U have values around 80 nT, and have an elongated shape. The striking direction of 1U is NE-SW, while it is rather N-S for U2. The positive anomaly 1U, has a similar value as the J-series of around 40-50 nT, and ridge-like shape striking in the NE-SW direction. The last described anomaly series in this facies is the R-series. It is made up of the anomalies R1, R2 and 1R. Anomaly R1 is circular in its shape, and is the strongest positive anomaly in the facies with values exceeding 110 nT. Anomaly R2 consists of three different rounded to subrounded peaks, with values in the range of 30-60 nT. The negative anomaly 1R has a value around -110 nT, and encircles anomaly R1 with a shape that resembles a horseshoe.

Facies 2 is the anomaly pattern that is situated farthest to the northwest. The anomaly pattern that characterizes this region is the strong elongated alternating positive and negative anomalies striking NE-SW. These are some of the strongest anomalies in the study area, spanning from 135 nT for the positive and -120 nT for the negative. These mark the transition from continental to oceanic crust, as the anomalies are described as spreading anomalies. The spreading anomalies that can be seen include chron 24B, 24A, 23 and 22.

The region of facies 3 is located in the southwest corner of the area of interest. The anomaly pattern resembles that of facies 2, with alternating elongated anomalies striking NE-SW. Instead of the anomalies being continuous and uniform in their responses, the anomalies are connected together as subrounded shapes as seen with the S-series and anomalies D1 and D2. The anomalies in this region have values ranging from around 90 nT for the positive (S1-S4, D1-D2 and B1) and -50 nT for the negative (1S-3S, 1D-2D and 1B). An anomaly that breaks the pattern is anomaly C1, because of its low value of around 10 nT compared to the other positive anomalies that have values around 90 nT. The same can be said for the negative 1C anomaly, it has the weakest response out of the negative anomalies in the facies. Across to facies 5, the anomalies in facies 3 terminate, and the anomaly values and shapes decreases drastically.
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Magnetic

Facies 4 is located in the easterly corner of the area. The anomalies that make up the area consist of two strong positive anomalies (150 nT) and one not so strong anomaly (65 nT). These are all surrounded by strong negative anomalies with values around -120 nT. The shape of anomalies F1 and F2 has a hammer like shape, with the handle striking NW-SE, while the head is striking perpendicular in the NE-SW direction. The other strong positive anomaly is the E1. This anomaly is a polygon with an irregular shape making it difficult to see its strike direction. The last positive anomaly is the G1 which has an elliptical shape striking NW-SE. It seems like it is connected to an anomaly in facies 5 (G2), even though the anomalies have different values. The strong negative anomalies in this facies, 1F-4F, all have values of around -110 nT. The shapes of the anomalies differ quite a bit, with 1F having an elongated form striking NW-SE. While anomaly 2F has an hourglass shape, 3F has a curved shape and anomaly 4F has a rounded shape.

Facies 5 is the low value zone located in the middle of the area of interest. Here the values for the anomalies lie in the range of -20 nT for the negative anomalies to 20 nT for the positive anomalies. The shapes making up the anomaly pattern are subrounded, with some anomalies striking N-S (L1-L3 and 1G-3G) while others strike NE-SW (M1 and M2). The anomaly pattern in this region resembles the pattern found on the Vøring margin. The magnetic pattern in this region can mainly be divided into three subgroups of anomalies, the L, M and G groups. The L group consists of anomalies L1-L4 and 1L & 2L. The trend for these anomalies is that increase in strength northwards, with L3 and L4 being the strongest and L1 the weakest for the positive anomalies. The negative anomalies lie between the L and M groups, and have a very low response at around -15 nT. The orientation of the negative L anomalies is parallel to the positive L anomalies L2 and L3. The M group consists of the following anomalies; M1-M3 and 1M. The positive anomalies M1-M3 forms a horseshoe shape around the negative anomaly 1M. The shapes of the positive anomalies M1-M3 are all different, with M1 being close to round, M2 subrounded and M3 being elongated. The strength of the positive anomalies lies around the value of 10 nT, while the negative 1M anomaly is -5 nT. The last subgroup consists of the very weak negative magnetic anomalies G1-G3 as their response is around -5 nT. All the anomalies have a subrounded shape, with G3 being the largest. A small anomaly that is also present in this facies is the P1 anomaly. This anomaly has a
subrounded shape striking E-W, and gets narrower westwards. Another stand alone anomaly is the Q1 anomaly, this anomaly has a subrounded shape that strikes in the NE-SW direction. Q1 has the strongest anomaly response in this facies with values around 40 nT.

Facies 6 is located in the upper middle part of the area of interest. This region consists of three subrounded relatively strong positive anomalies (H1-H3), separated by subrounded weak negative anomalies(1H-3H). The anomaly pattern created by this facies resembles the letter “L” as H1 and H3 are striking NW-SE while H2 is striking NE-SW. The anomaly response is also different for these positive anomalies as H1 and H3 have values around 150 nT while H2 has around 50 nT. The negative anomalies have a value of around -40 nT, indicating that it is not the weakest nor the strongest anomaly response in this region.

4.4 Grav-mag correlation

As both gravity and magnetic data are supplementary to the seismic data, it is beneficial to know when the anomaly responses coincide or not as this can give important information (figures 4.7 and 4.8). There are three possible ways on how magnetic and gravity anomalies corresponds to each other; both anomalies are positive, both anomalies are negative or that they have opposite responses e.g negative gravity and positive magnetic responses. In the study area there is a mix of all the three different scenarios.
The R series coincides good with the k2 anomaly, where both the gravity and magnetic responses are overall positive with the exception of 1R. That is somewhat reflected within the k2 anomaly as R1 corresponds to the strongest part of k2. The neighbouring anomaly k1 and its corresponding magnetic anomalies A1, 1A-3A have a more complex connection. Here the positive anomalies k1 and A1 almost coincide, but A1 is a bit shifted with respect to k1. Another part of the anomaly pattern is the positive gravity anomaly k1 correlating to the negative magnetic anomalies 1A and 2A, while the negative magnetic anomaly 3A corresponds mostly to the negative gravity anomaly 1m. The anomaly that encircles the positive gravity anomaly k1 is the positive magnetic J-series. Within the J-series, the J1 anomaly corresponds to a part of the 1m anomaly, J2 corresponds to the outer positive part of the positive gravity anomaly k1, while the J3 anomaly corresponds mostly to the negative gravity anomaly 1k.
The positive anomalies in the n-series seem to coincide very well with strong magnetic positive anomalies. As n1 has relatively good overlap with the magnetic anomalies L2 and L3, even though half of L2 corresponds to a negative gravity response. The positive gravity anomaly n2 coincides with the positive magnetic anomaly H1 for most of its lower part. The last anomaly in the positive gravity series, n3, has a very good correlation with the positive magnetic anomaly Q1. The parallel positive gravity l-series also seem to have a good correlation with positive magnetic anomalies as the n-series had. Here the gravity anomaly l1 coincides with the upper portion of the positive magnetic E1 anomaly. The gravity anomaly l2 coincides also very good with the positive magnetic G-series. A similar trend for these anomalies is that both reduce in strength northwestward. Moving northwards to the positive gravity anomaly l3, the positive magnetic anomaly H2 has a relatively good fit for most of the central part of the anomaly. Further north there is also a positive-positive correlation between the magnetic anomaly H3 and the gravity anomaly l4.

The positive gravity anomalies in the f-series seems to correlate with positive magnetic anomalies in the northeastern part and deviates further southwest. This can be seen from the positive gravity anomalies f1 which coincide with the positive magnetic anomaly L4, and the positive gravity anomaly f2 coincides with the upper parts of the positive magnetic anomaly M3. Then the positive anomalies f3 and f4 deviates from the positive magnetic anomalies M3 and U1. So rather being a positive-positive connection for the whole series, the positive magnetic anomalies coincide with different negative anomalies in the f-series. Instead the positive gravity anomalies f3 and f4 coincide with the negative magnetic anomalies 1U and 2U.

In the southwestern part of study area all the positive magnetic anomalies (B1, C1, D1-D2) end near the positive gravity anomalies a1 and a2. There are no clear anomalies representing the positive gravity anomalies a1 and a2 in the magnetic maps. The positive magnetic D1 and D2 anomalies have an anticorrelation with the negative gravity anomaly 1e. The same is the case for the positive magnetic anomaly C1 and the negative anomaly 2e, even though both the magnetic and gravity anomalies have weak responses. The positive magnetic anomaly B1 coincides better than the others as there are some overlap with the positive gravity anomaly b2. A prominent positive anomaly series, S, is located to the southeast of the just
above mentioned anomalies. This positive magnetic anomaly series coincides with
the positive gravity anomalies e1, v1 and v2. Just north of the magnetic S series,
there is an anticorrelation with the positive magnetic anomaly L1 and the negative
gravity anomaly 5f. Located in the transition between the negative gravity anomalies 2f and 3f, the positive magnetic anomaly M1 is out of the ordinary as the gravity
response here is close to 0 mGal. The last one of the easily seen anomalies is the
positive magnetic F-series, where the connection with the gravity map is complex.
The positive magnetic F1 anomaly is split in the middle, with one side coinciding
with the negative gravity anomaly 1l while the other half coincides with a positive
anomaly. The positive magnetic F2 anomaly coincide a bit better as the high pos-
itive values at the edges correspond to high valued magnetic anomalies while the
center corresponds to a negative gravity anomaly.

Figure 4.8: The 50 km high-pass filtered Bouguer residual anomaly map with 50 km
high-pass filtered negative magnetic anomalies draped over(black polygons). Black and white
numbers coincide with positive and magnetic anomalies respectively. Black lines is seismic
lines in area of interest.Data courtesy TGS
RESULTS

Grav-mag correlation

The negative magnetic G-series coincides very well with the negative gravity n-series, as there is almost a full overlap when draping the magnetic anomalies over the gravity map. The same coincidence can be seen with the negative magnetic anomalies 1U & 2U with the positive gravity anomalies f3 & f4. Here the overlap is not as significant, but the shapes are similar and aligned. Another good fit between the anomaly maps are the negative magnetic anomaly 1M and the negative gravity anomalies 1f and 2f. As for the positive magnetic anomalies the same trend can be seen for the negative anomalies (2D, 1C and 1B) in the southwest corner of the study area. All the negative magnetic anomalies seems to end with the positive gravity anomalies a1 and a2. The parallel negative magnetic anomalies (1D, 2D, 1C, 1B) southwest in the study area all corresponds in a different way with the gravity map. The negative anomaly 1D coincides with the positive gravity anomaly b1, while the negative magnetic anomaly 2D coincides more or less with the upper parts of the negative gravity anomaly 1e. The negative magnetic anomaly 1C coincides mostly with the negative gravity anomaly 2e, while the negative magnetic anomaly 1B really does not coincide with anything on the gravity map. The 1B anomaly coincides with some part of the positive gravity anomalies a2 and b2 on the edges, while coinciding with a lower gravity response in the middle.

The negative magnetic anomalies 1L and 2L coincide decently with the negative gravity anomalies 4f and 3f. Even though the negative magnetic anomalies does not fully overlap the negative gravity anomalies, there seems to be a correlation. Another negative-negative correlation is between the magnetic anomaly 2F and the gravity anomaly 1c. The same can be seen for the negative gravity anomalies 2h and 3h, as they coincide with mostly the negative magnetic anomaly F1. Even though in the northeastern part of magnetic anomaly F1, the anomaly coincides with positive gravity anomalies surrounding anomaly l1. The 2F anomaly covers most of anomaly 1c, while also overlapping with the positive gravity anomaly c1. The positive gravity anomalies l3 and n2 are embraced by the negative magnetic H-series, where the gravity anomalies l3 and n2 coincides with the magnetic anomalies 1H and 3H respectively. The negative gravity anomaly 3n coincides with the last negative magnetic anomaly in the H-series, the 2H anomaly. In the narrow part of the positive gravity anomaly t1 there is an anticorrelation with the negative magnetic anomalies 1J and 2J. Another anticorrelation can be seen with the southern
part of the positive gravity anomaly c1 and the negative magnetic anomaly 4f. The last large negative magnetic anomaly response is from anomaly 3F. On the gravity maps it coincides mostly with the negative anomalies 1l and 2h, but coincides with a positive gravity anomaly that lies in between the negative 1l and 2h anomalies.

4.5 Integrated/Interactive interpretation

To further constrain the seismic interpretation of the two lines, Transect 1 and Transect 2, 50 km high-pass filtered residual anomaly maps are loaded into the workstation as pseudohorizons (figures 4.9 & 4.12).

4.5.1 Transect 1

For the seismic section Transect 1 the gravity pseudohorizon consists of anomalies a1, a2, 1f and 2k. While the magnetic pseudohorizon is made up of anomalies D2, U1 etc (figure 4.9). In the southeastern part of the section the positive magnetic D2 and positive gravity a1 anomalies are located. In the central part of the seismic section the magnetic response is close to zero, while the response from the gravity consists of two parts of the a2 anomaly with a slight dip in value in between. The northwestern part is dominated by alternating magnetic responses starting with the positive U1 anomaly. The gravity response in the northwestern part consist of the negative anomalies 1f and 2k separated by the positive anomaly X. Based on the gravity pseudohorizon and some reflection trends in the seismic, the interpretation of the key horizons is pushed towards shallower levels. Following the gravity pseudohorizon the basement trend seem to move back to deeper levels, and sedimentary strata change dip direction. With a discrepancy indicated by anomaly X.

When making the pushed interpretation, it is important to know which anomalies that have a true response, and is not caused by an edge effect or nearby anomalies. For the gravity pseudohorizon the anomalies in the southeastern part do not cross the anomalies in a good way (figures 4.5 and 4.6). From the end of the line in the southeast to the a2 anomaly, the response in the pseudohorizon should not be weighted heavily when doing the pushed interpretation (figure 4.9). The opposite is
the case for the anomalies northwest of a2, as the line more or less crosses anomalies in the center making a reliable interpretation for the gravity possible. The two D2 and 2S are crossed nicely by the seismic line, indicating that the anomalies are trustworthy in the interpretation. Northwestwards the pseudohorizon is made up of edge effect until approaching anomaly U1 and outwards where the trend changes to crossing the anomalies in the center.

**Figure 4.9:** *Grav mag pseudohorizons Transect 1. Red = 50 km high-pass filtered Bouguer, Blue = 50 km high-pass filtered magnetic. Data courtesy TGS*

The pushed interpretation of MCS line Transect 1 can be seen in figure 4.10. The interpretation of the intra Miocene reflector was based on the regional trend of the upper part of the seismic section. Base Eocene is also a reflector interpreted mostly based on seismic facies. The Cretaceous reflectors are interpreted to shallow in the northwest direction based on reflection trends in the seismic, as the reflectors identified as sills have an southeasterly dip. The top and mid-Cretaceous are interpreted to end at the base of the lava, while the BCU is interpreted to follow underneath the base of the basalt. This is in conflict with the gravity pseudohorizon which indicated more of a basin geometry in the region between anomalies a2 and X. The basement reflector is interpreted to also shallow towards the northwest. The difference is that the basement is interpreted as rotated fault blocks and not a continuous interfaces. The rotated fault block interpretation give rise for the basin geometry indicated by
the gravity pseudohorizon.

![Figure 4.10](image)

**Figure 4.10:** Pushed interpretation of Transect 1 based on pseudohorizons. Data courtesy TGS

To validate the pushed interpretation, potential field modelling was done. As XField is a layered based modelling software, each layer need to be given geophysical parameters such as velocity, density and magnetic susceptibility. These values can be seen in Table 4.1.

<table>
<thead>
<tr>
<th>Stratigraphic unit</th>
<th>Velocity [m s⁻¹]</th>
<th>Density [g cm⁻³]</th>
<th>Susceptibility</th>
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<td>Moho/Mantle</td>
<td>8100</td>
<td>3.1</td>
<td>0</td>
</tr>
</tbody>
</table>

**Table 4.1:** Velocity, density and magnetic susceptibility values for the stratigraphic units of MCS line Transect 1
The result from the potential field modelling of MCS line Transect 1 in XField can be seen in figure 4.11. In the upper panel the response from the unfiltered Bouguer anomalies is displayed, while the 50 km high-pass filtered Bouguer response is shown in the middle panel. The geological model built based on the integrated interpretation is shown in the lower panel. The forward calculated response for the model is shown as dotted lines, and the observed as smooth and continuous. There is an overall good fit between the observed and forward calculated responses, but there are areas with minor misfits. For the unfiltered response, the forward calculated response is slightly overtuned in the northwestern part. The 50 km high-pass filtered responses deviate to some extent in the southeastern part of the model, especially around the most southeasterly fault block.

**Figure 4.11:** Potential field modelling on MCS line Transect 1. Upper panel show unfiltered Bouguer response, middle panel show 50 km high-pass filtered response, lower panel show the geological model.
4.5.2 Transect 2

In the seismic section Transect 2, the gravity and magnetic pseudohorizons have higher amplitude residual anomaly response and a different pattern to what Transect 1 had. The gravity pseudohorizon consists of mostly negative anomalies in the southeast region, with the exception of the very prominent positive n1 anomaly. In between anomalies 4f and 1k, there is a varying positive response on the gravity pseudohorizon. Further northwest, anomalies k1, 2A and k2 makes up the response of the gravity pseudohorizon. Here the k1 anomaly is of significant interest, as this anomaly will be looked further into in a detail study later in the thesis. The magnetic pseudohorizon has a very low response in the southeastern part of the section. This is a combination of low magnetic responses and the scaling factor necessary to include the large anomalies in the northwest region. The large magnetic response in the northwest are from the seafloor spreading anomalies. The area of interest concerning the magnetic pseudohorizon in this section includes anomalies A1, 1A, 2A and J2. Overall in the seismic section there is a good correlation between the two pseudohorizons, especially the large positive magnetic A1 anomaly and the large positive gravity k1 anomaly.

The magnetic pseudohorizon does not cross that many significant anomalies in the southeast region, as the seismic line runs in between low magnetic highs until reaching the J2 anomaly. Further northwest the line crosses anomaly A1, 1A and 2A almost at the center, making the anomalies reliable in regards to the pushed interpretation. Following the 2A anomaly, the rest of the anomaly response is close to the oceanic domain and will not be a part of the study area. For the gravity pseudohorizon the seismic line crosses most anomalies close to the center (n1,k2-2, 1k, k1 and 4f), making the gravity response very reliable. The only exception for the gravity is the response between 4f and 1k as this is a positive response originating from the close by f1 and f2 anomalies.
Figure 4.12: Grav mag pseudohorizons Transect 2. Red = 50 km high-pass filtered Bouguer, Blue = 50 km high-pass filtered magnetic. Data courtesy TGS

The integrated interpretation of the MCS line Transect 2 can be seen in figure 4.13. Here the integrated interpretation was not done for the full section, as the main focus for this thesis is on the margin and not in the oceanic domain. The same trends as for the integrated interpretation of Transect 1 can be identified. Where the Cretaceous reflectors are traced to shallower depth, and that the top Cretaceous terminates at the base of the volcanics. The basement and BCU reflector seem to mirror each other. These have a different response compared to other reflectors in the section. The mid- and top-Cretaceous, Base Eocene and Intra Miocene all seem to have a more smooth variations across, with some doming in the southeastern part. The integrated interpretation correlates well with the gravity pseudohorizon, with the prominent basement high in the southeast followed by a basin that shallows towards the Møre Marginal high. The interpretation also correlates with the magnetic pseudohorizons, with strong responses at the location of the volcanics. The prominent basement high is not easily seen in the magnetic pseudohorizon.
As for the MCS line Transect 1, the integrated interpretation of the MCS line Transect 2 needs to be validated with potential field modelling. The geophysical values for the layered model are given in Table 4.2. The values are very similar to the ones used to model Transect 1, but some layers have been added.

<table>
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<tr>
<th>Stratigraphic unit</th>
<th>Velocity ( \frac{m}{s} )</th>
<th>Density ( \frac{g}{cm^3} )</th>
<th>Susceptibility</th>
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<tbody>
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<td>Water layer</td>
<td>1500</td>
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<tr>
<td>Miocene</td>
<td>2000</td>
<td>2.0</td>
<td>0</td>
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<tr>
<td>Paleocene</td>
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</tr>
</tbody>
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Table 4.2: Velocity, density and magnetic susceptibility values for the stratigraphic units of MCS line Transect 2

The potential field modelling for the MCS line Transect 2 can be seen in figure 4.14.
The unfiltered response can be seen in the upper panel, the 50 km high-pass filtered Bouguer in the middle panel while the geological model is displayed in the lower panel. There is a very good fit between the observed and forward calculated magnetic response, despite that only three layers have magnetic properties in the model (table 4.2). The unfiltered Bouguer response is slightly off in the northwestern part of the section, this is also indicated in the 50 km high-pass filtered response. There is also some misfit in the center of the 50 km high-pass filtered panel.

**Figure 4.14:** Potential field modelling on MCS line Transect 2. Upper panel show unfiltered Bouguer response, middle panel show 50 km high-pass filtered response, lower panel show the geological model.

### 4.6 Detail study

As most of the focus for the thesis has been on the regional scale, a detail study around the prominent positive gravity anomaly k1 (figure 4.5), the positive magnetic anomaly A1 and its surrounding negative anomalies 1A-3A (figure 4.6) have been conducted. These anomalies where picked based on their complex pattern compared...
to other anomalies in the study area. The seismic lines crossing the detail area consist of a zoomed section of Transect 2 (Transect 3, figure 4.15) and a 3D seismic inline (Transect 4). The horizons that was traced in this section is the same as for the regional line, with exception of the deepest reflectors as they are out of bounds of the zoomed section.

Transect 4 is an Inline from the newly aquired 3D cube from TGS, and is of very high quality with a good resolution beneath the volcanics compared to other transects. The interfaces that were traced in this section includes the seabed, top basalt, base basalt and two strong reflectors at 5-6.5s and a prominent reflector at 7-9 s (figure 4.17). As with the regional lines, the seabed is characterized as the first prominent reflector in the section. Both the top basalt and base basalt reflectors were picked based on strong reflectors in close proximity with a seismic facies change. The top basalt reflector was traced along a high amplitude peak, while the base basalt reflec-

Figure 4.15: Location of the zoomed section Transect 3. Seabed (pink), top lava (red/brown), base lava (purple), BCU (dark green). Line location seen in 4.18. Data courtesy TGS.
tor was traced along a high amplitude through. The deepest reflector is traced along an easily seen peak located below 7s in the section. It has very visible response compared to the seismic units in the same area, which are blurry and lack the necessary resolution for identification. In the area in between the base basalt reflector and the prominent deep reflector, there are two reflectors that can be identified. The two reflectors have a very similar shape and amplitude response.

**Figure 4.16:** Transect-4 seismic. Line location seen in 4.18. Data courtesy of TGS

**Figure 4.17:** Transect-4 with all traced horizons. Seabed (pink), Top lava (dark red/brown), Base lava (purple), possible BCU (pink), possible basement (light purple), LCB/T-reflection (brown). Line location seen in 4.18. Data courtesy of TGS
On the zoomed magnetic anomaly map (figure 4.18a), the anomalies A1 and 1A-3A dominate. As mentioned in the regional description, there are some weaker positive anomalies that encircles these main anomalies (J-series). Two other interesting anomalies include V1 and 1V. Where both anomalies are oval in shape and strike NE-SW, while the difference is in the polarity with V1 being positive and 1V negative. Westwards in figure 4.18a there are some strong anomalies, but these are not included in the study area as the anomalies approach the oceanic domain. The zoomed gravity map consists mainly of the k1, 1k and 1m anomalies (figure 4.18b). From the zoomed figure, it is possible to distinguish the k1 anomaly into three parts; k1-1, k1-2 and k1-3. Where k1-1 consists of the area with the maximum amplitude value, located the furthest south. Then the k1-2 anomaly has similar shape as the k1-1 anomaly but lower amplitude value. The last anomaly, k1-3, is different from the other two in both shape and amplitude value. This anomaly has also a different strike direction. While the negative anomalies 1k and 1m are unchanged.

(a) Zoomed section of the 50 km high-pass filtered magnetic maps. Data courtesy of TGS.

(b) Zoomed section of the 50 km high-pass filtered Bouguer map. Data courtesy of TGS.

Figure 4.18: Zoomed section of the 50 km high-pass filtered maps with locations of Transect 3 and 4. a) Magnetic map, b) Bouguer gravity map. Data courtesy TGS.

The anomaly maps does not have a good correlation with each other, which is illustrated in figures 4.19 and 4.21. The maximum amplitude response of k1-1 is located at the most southern point of the A1 anomaly. The k1-1 anomaly also coincides with the 2A and V1 anomalies. While the 1m anomaly corresponds to a mix of magnetic anomalies including the northern part of A1, whole 3A, and the northern part of the J-series. The weaker positive k1-3 gravity anomaly coincides with the strong
negative magnetic anomaly 1A for most of its shape, while the northeastern part coincide more with some anomalies in the J-series. There is however an area where the responses from the gravity and magnetic anomalies have a reasonably match. In this area the k1-2 anomaly coincides with the "lower" part of A1 anomaly. The maximum amplitude values for the anomalies also coincide in this area.

\textbf{Figure 4.19:} 50 km high-pass filtered magnetic map, with the 50 km high-pass filtered anomaly k1 draped over as a dotted line. Data courtesy of TGS

The 50 km high-pass filtered Bouguer and magnetic pseudohorizons of the two transects can be seen in figures 4.20 and 4.21. The pseudohorizon in Transect 3 correlates well in the region with A1 and k1-2, but the pseudohorizons seems to be shifted in the gravity and magnetic lows. The pseudohorizons are deeper in the SE direction, compared to the NW part. For Transect 4 both the gravity and magnetic pseudohorizons have large responses in the WNW. Here the gravity response is represented by the large positive k1-1 anomaly, followed by the negative 1k anomaly further ESE. While the magnetic response is defined by anomalies 2A, V1 and 1V. In the ESE part of the section, the magnetic pseudohorizon is close to its equilibrium, while
the gravity pseudohorizon has some minor amplitude changes.

Figure 4.20: Gravity (red) and magnetic (blue) pseudohorizons on seismic Transect 3 with given anomaly names. Data courtesy of TGS

Figure 4.21: Gravity (red) and magnetic (blue) pseudohorizons on seismic Transect 4 with given anomaly names. Data courtesy of TGS

The potential field modelling for the two transects can be seen in figures 4.22 and 4.23. Where the unfiltered total responses are illustrated in the upper panel, while the 50 km high-pass filtered responses are shown in the middle panel. The geological
model is displayed in the lower section of the figures. Transect 3 is a zoomed section of Transect 2, and does not go down to 10 twt, while Transect 4 is modelled for the whole line. The models for Transects 3 and 4 consist of seven and eight layers respectively. The layers in common include the water layer, Cenozoic unit, basalt unit, lava delta, Cretaceous unit, Pre-Cretaceous/Basement unit. For transect 4 an LCB/mantle unit is incorporated. The geophysical properties of each layer in the model are given in Table 4.3 and 4.4.

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**Table 4.3:** Geophysical values for Transect 3 incorporated in XField

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<tbody>
<tr>
<td>Sea</td>
<td>1500</td>
<td>2.2</td>
<td>0</td>
</tr>
<tr>
<td>Miocene</td>
<td>1900</td>
<td>1.9</td>
<td>0</td>
</tr>
<tr>
<td>Basalt</td>
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<td>2.4</td>
<td>-0.03</td>
</tr>
<tr>
<td>Lava delta</td>
<td>4000</td>
<td>2.5</td>
<td>-0.02</td>
</tr>
<tr>
<td>Cretaceous</td>
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<td>2.5</td>
<td>0</td>
</tr>
<tr>
<td>Basement</td>
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<td>2.75</td>
<td>0</td>
</tr>
<tr>
<td>LCB/T-Reflection</td>
<td>7100</td>
<td>3.1</td>
<td>0</td>
</tr>
</tbody>
</table>

**Table 4.4:** Geophysical values for Transect 4 incorporated in XField

The layers in the model assumed containing sediments (Cretaceous and Cenozoic) seem to decrease in thickness towards WNW, while the layers describing the volcanics and T-reflection/LCB have the opposite trend. The calculated response from the model is very close to the observed unfiltered Bouguer response. It does however deviate slightly in the ESE part of the modelled section. The calculated magnetic response also has a good fit with the observed magnetic total field. It deviates a bit more than the observed gravity, but not by much as the general trend of the observed data can be seen in the forward calculated response. The same can be seen in the filtered response.
Figure 4.22: XField model for Transect 3

Figure 4.23: XField model for Transect 4
As a part of the detail study, VBPR provided a basalt thickness map (figure 4.24. The figure illustrates the variation in thickness of basalts within the detail study area. Instead of the area being more or less uniform in thickness, there are localized areas with a very thick cover (dark red /brown) while there are some windows where the basalt thickness is close to zero (white).

**Figure 4.24:** Basalt thickness map, marked escarpment (dotted line). Seismic = Transect 4. Data courtesy of VBPR & TGS
5 | Discussion

5.1 Regional

To further discuss the implications of the interpreted model, the uncertainties need to be addressed. The model was built using constant velocity, density and magnetic susceptibility for each layer in the model. This creates a very simple model, where the calculated density and magnetic susceptibility variations are more dependent on the thicknesses of the layers rather than sudden density contrasts recorded by the observed data. As the XField model is built in time, it uses the primitive velocity field generated by the interval velocities to depth convert the model for calculations of the potential field anomalies. This induces uncertainties in the regarding the placement of the interfaces in the depth domain. With all these uncertainties taken into account there still is a relatively good fit between the calculated and observed response.

The mapping of the LCB and Moho for Transect 1 was constrained by the OBS data confidently, as the OBS data coverage consisted of two subparallel transects from Kvarven et al. (2014) (figure 5.1b) and Raum (2000) (figure 5.1a). This was not the case for Transect 2, as all of the available published OBS profiles had a different orientations. There were four OBS transects crossing Transect 2, but only two was applicable based on the uncertainties attached to the velocities at depth near the ends of the OBS profiles. Hence the LCB and Moho geometries was not deduced based on OBS data, but rather were interpreted as one unit based on the trends found in Transect 1.

The volcanics vary in extent northwards on the margin. For Transect 1 the volcanics seems like a more continuous unit, with a larger thickness towards the oceanic domain compared to the volcanics in Transect 2. The volcanics in Transect 2 is interpreted to have different origin, as the volcanics in the northwest is postulated to be sea ward dipping basalt, while the basinward lava delta may consist of hyaloclasites. The erosional surface present in Transect two might be a significant factor for the
difference between the volcanics.

(a) L8-96 from Raum (2000)

(b) Profile 2 from Kvarven et al. (2014)

Figure 5.1: Regional velocity models across the Møre margin based on OBS data. The color-coded velocity models are converted from depth to time (s twt) and draped on top of nearly coincident seismic reflection profiles. a) based on Raum (2000); b) based on Kvarven et al. (2014).

5.1.1 Transect 1

The deep basin configuration of MCS line Transect 1 (figure 4.11), has similarities with the deep basin trends in Theissen-Krah et al. (2017) (figures 2.12 and 5.2 F). There is a good fit between the models in the southeastern part, with the exception that the model from Transect 1 is slightly deeper. This indicates that the velocities for the overlying sedimentary layers in Transect 1 are overestimated with respect to the model of Theissen-Krah et al. (2017). Further northwest there are some differences in the deep basin configuration. The basement in Theissen-Krah et al. (2017)
gradually shallows to around 7 km at the Møre marginal high, while the basement reflector in the Transect 1 have larger rotated fault blocks that have a better fit with the interpretation of Nirrengarten et al. (2014) (figure 5.2 D) and Gernigon et al. (2015) (figure 5.2 E), but do follow the overall same shallowing trend as Theissen-Krah et al. (2017) (figure 5.2 F). The models proposed by Osmundsen et al. (2016) (figure 5.2 A-C) do not fit the model for Transect 1. The post-rift sediments in figure 5.2 G) are too thin to satisfy the same post-rift sediments in figures 5.2. These models also assume that the upper crust is almost non existent, while figure 5.2 G) indicate a crust of 10 km in the same areas.

The interpreted basement highs correlate with responses in the 50 km high-pass filtered Bouguer residual anomaly map (figure 4.5), as the seismic line crosses three prominent positive anomalies and lie in close proximity to others. The LCB and Moho geometries for Transect 1 are well defined and coincide with the geometries in OBS profile 8A-96 from Raum (2000) (figure 5.1a) and Profile 2 from Kvarven et al. (2014) (5.1b). Both the LCB and Moho seem to move to shallower depth when located beneath basement highs. The LCB also seems to thin towards the southeast. The magnetic response of Profile 1, is highly influenced by the volcanics in the northwest and a weak magnetized basement (Gernigon et al., 2015).
Figure 5.2: Illustrating the different interpretations of the Møre margin. A-C) retrieved from Osmundsen et al. (2016), D retrieved from Nirrenzarten et al. (2014), E) retrieved from Gernigon et al. (2015) and F) retrieved from Theissen-Krah et al. (2017). G) Depth model of Transect 1.
5.1.2 Transect 2

Transect 2 shares much of the same deep basin configuration, and is in accordance with the model of Theissen-Krah et al. (2017), and do not support the exhumed serpentinitized mantle model in the deep basin configuration. (Osmundsen and Ebbing, 2008; Osmundsen et al., 2016; Peron-Pinvidic et al., 2013). The authors argue that the large detachment faults taper the crustal part in the distal domain, resulting in complete embrittlement of the crust allowing water to reach the mantle. They also argue that the exhumed mantel windows arise below the footwalls of these large detachment faults, and that the alloctones reduce in size NW of Vigra high as a result of an underlying NW rising detachment fault.

In Transect 2, the sediments arise to shallower levels and reduce in thickness towards the northwest approaching the Møre Marginal high. The BCU is also very shallow below the Møre Marginal high (5 km), which also indicate that some crustal remains may be present. As the OBS data coverage is limited for Transect 2, the Moho geometry was interpolated between known locations based on the Moho geometry of Transect 1. The high-pass filtered magnetic anomalies indicate that the volcanics are the main source for the anomalies, which are agreement with the interpretation of Gernigon et al. (2015).

5.1.3 Geological models for margin development

There are several models proposed on how the passive mid-Norwegian margin evolved. The main difference between the proposed models for the passive margin formation and evolution is located on the outer margin (Theissen-Krah et al., 2017; Gernigon et al., 2015; Nirrengarten et al., 2014; Peron-Pinvidic et al., 2013; Osmundsen and Ebbing, 2008).

Osmundsen et al. (2016) and Peron-Pinvidic et al. (2013) argue that there are similarities between all passive margin formations, and that the development of a passive margin is a continuous phenomenon. The authors suggest that the crustal thinning is controlled by large detachment faults, and that the margin can dived into distinct domains; proximal, necking, distal, outer and oceanic. These domains also have
processes that are characteristic for the specific domain. They argue that the large basin scale detachments have thinned the crust to the extent of complete embrittlement of the crust, allowing serpentinisation of exhumed mantle in the outer domain.

Theissen-Krah et al. (2017) and Gernigon et al. (2015) propose a passive margin evolution that is not continuous, in contrast with the suggestion of Osmundsen et al. (2016) and Peron-Pinvidic et al. (2013). The authors suggest that the deep Cretaceous basins is a result from an abandoned rift axis in the Møre basin in the Late Jurassic - Cretaceous rifting period. They argue that the Møre margin needed to accommodate for half the extension between the mid-Norwegian and Greenland conjugate margins, while the rest of the extension is suggested between the Jan Mayen Microcontinent and the east Greenland margin. This further indicates that the high velocity body beneath the outer domain is represented by magmatic underplating or preserved lower crust intruded by volcanics (Theissen-Krah et al., 2017; Gernigon et al., 2015; Mjelde et al., 2009).

Transects 1 and 2 in this thesis is overall more supportive of the model proposed by Theissen-Krah et al. (2017). As both modelled transects support the shallowing of the BCU towards the Møre Marginal high. The modelled transects also support a thick Cretaceous infill, with a rather thick pre-Cretaceous sequence. This indicates that there is still preserved crust beneath the Møre Marginal high. From the transects in this thesis the crustal thickness is interpreted to be around 10 km close to the Møre Marginal high, which does not fit with the exhumed serpentinised mantle proposed by Osmundsen et al. (2016) and Peron-Pinvidic et al. (2013). To further test/validate the proposed model a detail study at northwestern Møre margin give important insight in the discussion.

5.2 Detail study

From the results of the detail study, it becomes evident that the magnetic and gravity anomaly have a poor correlation (figure 4.19). A possible scenario for explaining the mismatch could be that the gravity and magnetic anomalies have a different origin, implying that the source depths of these anomalies are different. The detail
study also explored the nature of Kolga high, in an attempt to constrain the regional passive margin evolution.

From the potential field modelling the source of the magnetic anomalies is suggested to originate from the volcanics at around 3-4.5 s twt in figures 4.22 and 4.23. The relatively good fit for the calculated response is achieved when the volcanics are the only layers with magnetic susceptibilities. This is in accordance with Gernigon et al. (2015), as the authors suggest that the surface volcanics explain most of the magnetic response near the Møre Marginal high. The magnetic response may be explained by the basalt layer being thicker and has a slight dip, compared to the lava delta which is relatively flat and thinner (figure 5.3).

![Figure 5.3: Transect 3, with 50 km high-pass filtered magnetic pseudohorizon (blue), base basalt (purple) and top basalt (red/brown) reflectors](image)

By comparing the the 50 km high-pass filtered magnetic anomaly map (figure 5.4) with the basalt thickness map (figure 5.5), an interesting pattern occur. As the positive anomalies in the detail study area (A1 and V1) seem to correlate with the windows where the basalt thickness is very thin or close to zero. The opposite is
true for the negative anomalies (1A, 2A and 1V), as they are located in areas where the basalt is thickest, reaching thicknesses around 500-600 ms.

Figure 5.4: Mag50-3D. Data courtesy of TGS

Figure 5.5: Basalt-thickness-Mag50. Data courtesy of TGS

The potential field modelling suggest that the origin of the filtered Bouguer anomaly is located deeper than the surface volcanics (4.22, 4.23 and 5.6). This can be argued over the poor fit between the filtered Bouguer gravity map and the basalt thickness.
map. The potential field modelling suggests that the density contrast is situated at around 5-7 km. The magnitude of the filtered positive high is around 5 mGal, indicating a relatively large density contrast, suggested to be in the order of 0.3 g/cm³. In the potential field model, the geometry of the density contrast has a dome type shape, which may reflect the Miocene compressional forces (Brekke, H. and Dahlgren, S. and Nyland, B. and Magnus, C., 1999; Lundin and Doré, 2002). There is also indication of an erosional surface in the area where there is no basalt cover, correlating with k1-2 anomaly in the gravity pseudohorizon (5.6). The best fit is achieved with a dome shaped basement with density 2.85 g/cm³ overlain by non-magmatic sediments with densities 2.60 g/cm³ in the range of pre-Cretaceous to upper Cretaceous.

Figure 5.6: Transect 3, with 50 km high-pass filtered Bouguer pseudohorizon (red), base basalt (purple) and top basalt (red/brown) reflectors

5.3 Integrated SGM interpretation and XField

In an exploration environment, the integrated seismic-gravity-magnetic workflow is found to be very useful. The first hand interpretation of the different data sets, allow the interpreter to recognize important elements within each of the geophysical
Integrated SGM interpretation and XField

data sets. This makes it so that the interpreter has to take all the information into account before selecting key transects. Then the key elements can be tested qualitatively with the integrated approach, where the gravity and magnetic data are draped over the seismic as pseudohorizon. The gravity and magnetic data can then be filtered, so that the pseudohorizons reflect the depth of investigation the interpreter wants. This can help identify trends in the seismic, that previously would have been impossible/very difficult to obtain. By employing all three different data types in the same interpretation setting, it has the potential to greatly reduce risk in regards of the interpretation as it fulfill three independent data types. The method shows its importance in areas dominated by high impedance layers (salt and volcanics) where the seismic quality is poor. Even though the method is most effective in such areas, the method is still recommended to be used as it act as a quality check for the interpretation for any environment.

To validate the integrated interpretation, potential field modelling is done. This was previously disconnected from the seismic workstation (e.g GM-SYS). To further enhance the interpretation and the integrated workflow, it would be beneficial to test the gravity and magnetic data quantitatively within the same workstation. XField is a modelling software that gives the interpreter this opportunity. The benefit of having a way to quantitatively test the potential field data, is that it can greatly increase the efficiency in the interpretation workflow.

The software can be both available and practical, but the most important feature is that it gives reliable results. The calculation of the geophysical anomalies is based on the method of Talwani et al. (1959), which is commonly used for most calculations regarding potential field data. XField is different from most potential field modelling softwares as the interpreter can make to model in the time domain, directly testing the interpretation in time. Then the time model needs to be depth converted, and this operation is hidden from the user. The interactive nature of the software, makes it so the model recalculates in real-time when changes are applied. This may induce uncertainties. To combat this "hidden operation" XField has different validity checks for the model, as it does not allow for intersecting bodies or corners in time or depth domain. But before the software can be considered robust and reliable, the uncertainties in the depth conversion needs to be quantified.
6 | Conclusion

The integrated SGM interpretation allowed to give insight on the passive margin development, as well as the deep basin configuration on the northwestern part of the Møre margin. The models were validated through the new interactive potential field modelling software XField. The main conclusions can be summarized as follows:

- The interpreted transects and potential field modelling suggest that the NE Atlantic passive margin formed as an oblique opening during distinct extensional periods. The crustal architecture and deep basin configuration do not favor the exhumation of serpentinised mantle.

- The detail study around Kolga high gave new insight on the nature of the structural highs on the outer Møre margin. The magnetic response is a result of the Eocene breakup magmatism, while the density contrast is suggested to come from a deep-seated structural high. This interpretation opens for sub-basalt prospective sediments.

- By having the quantitative test quickly available for the qualitative analysis made the workflow more efficient and fluid, and it is suggested to incorporate a modelling tool like XField as a part of the integrated and interactive SGM approach.
References


REFERENCES


