**Magnetotelluric evidence for massive sulphide mineralization in intruded sediments of the
outer Vøring Basin, mid-Norway**

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Abstract

A highly conductive body (0.1-0.8 $\Omega\cdot\text{m}$) is identified at mid-crustal depth (8-13 km) in the north Gjallar Ridge from magnetotelluric (MT) data and further investigated in light of other remote-sensing geophysical data (Seismic reflection, gravity, aeromagnetic). A commercial 3D controlled-source electromagnetic survey was conducted in the Vøring Basin in 2014 and, although primarily designed for hydrocarbon exploration, good quality MT data were extracted at periods ranging from 10^0 to 10^3 s. Dimensionality analysis indicates clear 1D to 2D characteristics in the MT data. 2D inversion was carried out on four profiles (totalling ~94 km) oriented perpendicular to the electromagnetic strike and one profile along strike (~45 km), using a 1D subset of the data. All inversions converged quickly to RMS values close to unity and display a very good agreement with borehole resistivity data from well 6705/10-1 located in the survey area. A striking feature on all profiles is a highly conductive (0.1-0.8 $\Omega\cdot\text{m}$) body at 8-13 km depth. To explain the prominent conductive anomaly, integration of geophysical data favours the hypothesis of electrical conduction across well-connected mineral network in pre-Cretaceous sediments. Seismic interpretation suggests a link between the conductor and intruded sedimentary successions below a detachment level and associated low-angle faults. In the Vøring Basin, low magnetic signal and temperature at the conductor's depth indicate that such thick mineral deposits could display non-magnetic behaviour while occurring well below the magnetite Curie isotherm (~585°C). Natural occurrences and magnetic properties of common iron-sulphide minerals favour a geological interpretation of mid-crustal conductivity as thick pyrrhotite deposits formed in intrusion's contact metamorphic aureoles.

Keywords

Magnetotelluric; Electrical conductivity; Igneous intrusion; Iron-sulphide; Vøring Basin; Passive margin

1. Introduction

The mid-Norwegian margin is a passive continental margin characterized by several extensional phases that occurred subsequent to the post-orogenic collapse of the Caledonian mountain chain (Doré et al. 1999; Faleide et al. 2010; Skogseid et al. 2000; Tsikalas et al. 2008; Ziegler, 1988). The main rift phases (Lundin and Doré, 2005; Brekke, 2000; Mjelde et al., 2003) include regional rifting in the late Carboniferous to early Permian (~300-280 Ma), Triassic block faulting and late Triassic-early Jurassic growth faulting. Subsidence was especially severe during the Cretaceous, giving rise to the accumulation of up to 8 km of sediments in local depocentres in the Vøring and Møre basins (Scheck-Wenderoth et al., 2007) (Figure 1). Final continental breakup occurred in the early Eocene, ~56-55 Ma, and resulted in voluminous extrusive and intrusive igneous activity in the adjacent sedimentary basin and pre-existing continental crust (Breivik et al., 2014; Eldholm and Grue, 1994; Hinz, 1981; Mjelde et al., 2007; Mutter et al., 1982; Planke et al., 2005; White and McKenzie, 1989) (Figure 1). In the Vøring Basin, the volcanic sequences emplaced during the onset of the breakup seismically mask the late Cretaceous–early Eocene and older sedimentary strata (Planke et al., 2015). This seismic imaging challenge leads to major uncertainties in the sub-basalt rift architecture and various interpretations, especially in the deeper parts of the margin (Abdelmalak et al., 2015, 2016a; Peron-Pinvidic and Osmundsen, 2016).

In the absence of deep drilling, calibration of the lithological composition of rocks at depth often comes from potential field and refraction studies. Regional 2D ocean-bottom seismometer (OBS) surveys (Mjelde et al., 2008b) coupled with regional gravity and magnetic grids (Maus et al., 2009) are currently the primary source of information on the deeper crustal

architecture of the Vøring Margin. The magnetotelluric (MT) method images the electrical conductivity distribution within the subsurface, thereby complementing the existing data sets.

In this paper, we present new magnetotelluric (MT) data acquired in June 2014 in the north Gjallar Ridge, outer Vøring Basin (Figures 1 and 2). This survey was primarily designed for acquisition of controlled-source electromagnetic (CSEM) data (Constable 2010; Ellingsrud et al. 2002; MacGregor and Tomlinson, 2014), a technology now routinely used in hydrocarbon exploration workflows (Stefatos et al., 2014; Fanavoll et al., 2014). The CSEM set-up also allows for the acquisition of MT data when the active source is far from the seabed receiver.

In this contribution, we carried out 2D inversion of MT data along five profiles (Figure 2), totalling ~140 km. We obtained good quality MT data from periods 2 to 1500 s, which allow for the investigation of geo-electrical features in the upper continental crust (Corseri et al., 2015). To support the 2D inversion approach and the recovery of abnormally low resistivity ($< 0.8 \Omega.m$) at mid-crustal depths, we performed a dimensionality and electromagnetic strike analysis. Very low resistivity (or inversely, very high conductivity) crustal features have already been reported by Myer et al. (2013) in the outer Vøring Basin (Figure 1) and by other workers in rifted margins (Heinson et al., 2005; Jegen et al., 2016). Usually, these are interpreted as evidence for magnetite in layered mafic intrusions (Myer et al., 2013) or graphite films (Heinson et al., 2005). While a first objective of this study is to map geo-electrical features of the outer Vøring Basin, we further integrate the MT data, i.e. electrical properties, with seismic reflection and potential field data to unveil new lithological insights in the study area using the strengths of each geophysical dataset. We discuss the presence of massive, iron-sulphide deposits related to metamorphic aureoles of deep, layer-parallel igneous intrusions

in the north Gjallar Ridge and support this interpretation by reporting natural occurrences of such deposits in other large igneous provinces.

Figure 1

2. Methods and data

2.1. Magnetotelluric method

The MT method is a passive geophysical technique that images subsurface electrical conductivity (Cagniard, 1953). It involves the measurement of the natural variations of the horizontal electromagnetic fields at the surface of the earth. Depending on the periods extracted in the MT signal, usually from 10^{-3} to 10^5 seconds, the depth of investigation of MT data ranges from tens of meters to several hundreds of kilometres. In marine surveys, the highly conductive ocean layer acts as a frequency low-pass filter and one second is generally the shortest period recoverable. The impedance tensor Z_{ij} , where i and j are indices corresponding to the two orthogonal horizontal components (x and y) of the electric (E) and magnetic (H) fields, embodies the response of the conductivity of the earth and is given by:

$$\mathbf{E}_i = Z_{ij} \mathbf{H}_j$$

We describe the MT data by the apparent resistivity, ρ , and phase, φ , which relate to Z by:

$$\rho_{ij} = \frac{|Z_{ij}|^2}{\omega \mu}$$

$$\text{And } \varphi_{ij} = \arg(Z_{ij})$$

Where ω is the angular frequency and μ is the magnetic permeability of the earth.

Figure 2

2.2. MT data acquisition and analysis

The survey, acquired in 2014, was designed for 3D CSEM data acquisition to be used in hydrocarbon exploration in the uppermost 2-3 km below the seafloor. It comprises two receiver grids: the northern grid with a receiver spacing of ~1 km along four NW-SE lines and a southern grid with a receiver spacing of ~3 km along seven WNW-ESE profiles (Figure 2a). The two grids are tied by a SW-NE striking profile with a receiver spacing of 3 km. Three typical MT responses rotated along the direction of profile P3 are displayed in Figure 4. After processing using the robust multi-station processing scheme of Egbert (1997) combined with a proprietary auto-grouping algorithm (Markhus et al., 2015), the 120 stations (Figure 2) provided good quality MT signal for periods from 1 to 1500 s (Corseri et al., 2015). Subsequently, all 120 stations were used in the inversion phase.

Although the site layout allows for 3D inversion, 2D inversion is often preferable since the smaller computational requirements allow models to be more rigorously tested. However, 2D models are only reliable in regions that are geo-electrically 1D or 2D. The magnetotelluric phase tensor (Caldwell et al., 2004) is a good way to analyse electrical dimensionality because it is not affected by 3D distortion and does not assume an underlying 2D geo-electric structure. The phase tensor can be depicted graphically as an ellipse, whose major and minor axes represent the principal axes of the tensor. In a 2D case, one of the principal axes is aligned with the preferential direction of current flow, the electromagnetic strike.

The skew angle β is a measure of dimensionality. If the skew is zero, the ellipse is symmetric while a skew angle $>3^\circ$ or $<-3^\circ$ indicates three-dimensionality (Booker, 2014). The ellipticity is a measure of the difference in magnitudes of the maximum and minimum phases (Bibby et al., 2005); ellipticities <0.1 indicate that the underlying geological structures are 1D (Booker,

2014) and the phase tensor plots as a circle. Figure 3 shows the phase tensor ellipses for each station at periods 5.5 and 524 s, shaded by skew angle. Analysis of the phase tensor data shows that most of the MT data between 2.7 and 998 s are 1D or 2D, with low skew values and either low ellipticities or consistent strike directions. Stronger preferential current flow emerges at longer periods, as shown by the greater ellipticities of the phase tensors at 524 s. At these periods, the geo-electric strike of stations in the southern grid averages at approximately N10°E, while that of the stations in the northern grid is approximately N40°E. We used structural and potential field data to distinguish between the inherent 90° ambiguity in strike direction from the phase tensor data (Figure 3). However, the main fault orientation is not exactly corresponding to electromagnetic strike in the southern grid (Figure 2) and it is likely that more complex, deeper processes affects the preferential orientation of telluric current flows in the Gjallar Ridge. Individual data points were included in the inversions only if they had skew values $>-3^\circ$ or $<3^\circ$ and either ellipticities <0.1 or strikes within 10° of the average strike. The MT data were then rotated for inversion along profiles oriented perpendicular to strike (Figure 2). For the cross-line (P5), only data that satisfy 1D criteria (skew values $>-3^\circ$ or $<3^\circ$ and ellipticities <0.1) were included in the inversion. Most data at periods >500 s did not meet these criteria so the P5 inversion is more poorly constrain than the other inversions.

Figure 3

2.3. Additional geophysical constrains

We utilize seismic, well-logs, gravity and magnetic susceptibility data to integrate with the new resistivity data. The seismic data include the depth-converted MNR11 seismic lines, acquired by TGS and Fugro in 2011 (Figure 2a). We use resistivity log data from exploration

wildcat 6705/10-1, which is located along profile P5 (Figure 2a) and extends to 3775 m in the early Cretaceous Lange Formation. The deep induction tools offer resistivity log measurements that can be compared directly to the MT inversion results. We also use public-domain bathymetry (GEBCO, 2014), satellite gravity data (Sandwell and Smith, 2009), released magnetic data (EMAG2) (Maus et al., 2009), and local aeromagnetic data reprocessed by the Geological Survey of Norway (NGU) and TGS. We use 50 km Bouguer residual maps, which are mainly a response to deep basin or mid-crustal density variations, with typical source depths of < 10 km. Magnetic anomaly patterns reveal large tectonic trends and the distribution of oceanic crust with characteristic spreading anomalies. The wavelength and amplitude of anomalies provide information about depth to source and magnetic properties.

Figure 4

3. Results

3.1. Observation on the Gjallar Ridge MT data

Receivers 01Rx126 and 01Rx132 (Figures 2a and 3a), located along profile P3, display a drop in transverse electric (TE, with the electric field parallel to strike) apparent resistivity and increasing TE phase values for periods >200 s, which suggest low resistivity at depth. In contrast, the transverse magnetic (TM, with the magnetic field parallel to strike) apparent resistivity and phase data show little lateral variation at the different stations on profile P2 (Figure 4b). For periods <200 s, apparent resistivity and phase curves for both modes are very similar, supporting a 1D earth structure at these periods (Figure 4). The TE and TM mode split is not consistent throughout the survey. Receivers located close to the north-western boundary of the Gjallar Ridge, toward the volcanic flows, show close modes at all periods, indicating a 1D character of the shallow subsurface and at depth. On the other hand, towards

the central part and south-eastern boundary of the Gjallar Ridge, a split in apparent resistivity occurs consistently at ~200 s, with the TE mode dramatically decreasing while the TM mode increases more homogeneously. A large split in the modes occurs at the centre (01Rx126) and south-eastern boundary of the ridge (01Rx132). Jegen et al. (1998) shows that a strong anomaly in the TE mode and a marginal anomaly in the TM mode is generally indicative of a smaller scale conductive anomaly at depth.

In the MT survey described by Myer et al. (2013) and located in the Vigrid syncline (Figure 1), the coast effect was ruled out to explain the split in the two modes (Key and Constable, 2011). 2D modelling has shown that the coast is likely sensed at periods longer than 4000 s in the outer Vøring Basin. This conclusion can be extended to the present work, as our survey is located even further at sea.

3.2. 2D resistivity sections

We inverted the MT data using the MARE2DEM inversion scheme, a 2D parallel adaptive finite element code that uses an unstructured triangular mesh (deGroot-Hedlin and Constable, 1990; Key and Oval, 2011). We ran inversions with starting models ranging from 1 to 1000 Ωm half-spaces with a 5% error floor for apparent resistivity data and 1.45 degrees for phase data. The final results appeared to be robust with regard to the starting model resistivity. Bathymetry, which was included in the inversion, is quasi-flat, on average 1430 m, and smoothly varying. For example, profile P3 is 49 km long and the largest bathymetry variation is 30 meters (Figure 2a).

Given the receiver separation tailored for hydrocarbon prospect delineation and the large resolution of the MT method, we chose to invert MT data along five profiles (Figure 2a), one in the northern grid, three in the southern grid and the NE-SW tie line. All inversions converged

to a RMS misfit of 1.0 after 10 to 15 iterations except for P5 (Figure 2a) which converged to a slightly higher RMS of 1.5. The very good fit to the observed data is illustrated in Figure 4. The inverted models for neighbouring profiles show consistent electrical features throughout the survey (Figure 5) and do not change significantly with different starting model half-space resistivity. We have identified four geo-electrical units, labelled R1, R2, C1 and C2 (Figure 5), with “R” standing for resistor and “C” for conductor. An extremely low resistivity anomaly stands out on all profiles, exhibiting values as low as 0.1 Ω .m at 10 km depth.

Figure 5

The shallowest feature C1 is a homogeneous and relatively flat layer that extends from the seabed to ~3 km depth with an average resistivity of 0.8 Ω .m on all profiles. Feature R1, which is most clearly shown on profile P3, is also present on all inverted profiles; its top is a flat interface while its thickness varies and its resistivity ranges from 10 to 100 Ω .m. Below R1, the highly conductive C2 is located between 8 and 13 km depth on all profiles and has resistivity ranging from 0.1 to 0.8 Ω .m. C2 is confined to the easternmost half of the profile P3 in the southern grid (Figures 1a and 5). R2 is a landward dipping feature with resistivity around 15-30 Ω .m. However, modelling tests have shown that there is little resolution below C2 so detailed interpretation of R2 is ambiguous. Testing of profile P5, where only a 1D subset was inverted (mostly <500 s), showed that there is no sensitivity below C2.

We carried out model testing, focussing on profiles P1 and P3. We imposed a lower hard bound of 1 Ω .m on the inversions and the final misfits increased significantly, showing that resistivity values <1 Ω .m are required to fit the data. We inserted sharp boundaries at the main seismic interfaces (the top Cretaceous, base Cretaceous, basement and lower crustal body) and then allowed the models to invert. C1 could be mostly bound by the sharp boundary

at the top Cretaceous but the other electrical features did not appear to directly correlate with
the seismic interfaces. We also tested the depth extent of C2 by fixing the resistivity of the
layer underlying C2, between 10–30 km and 60 km, to a minimum of 100 Ω .m. When C2 was
constrained to extend to 10–15 km depth, C2 resistivities became extremely low ($<0.1 \Omega$.m)
and the resistivity of the underlying mantle was also unreasonably low ($\sim 1 \Omega$.m). Therefore,
C2 seems to extend to at least 10–15 km depth but its precise depth extent between ~ 15 –25
km is not well resolved. Due to the smaller bandwidth of data used, all features in P5 are less
well resolved than the other profiles. Where they differ, we consider the depth extent and
resistivity of C2 to be better constrained by P1–P4 than P5. The inverted features support the
conclusions of the dimensionality analysis, with the models showing mostly horizontally
layered features in the upper crust and more lateral variation at greater depth with the
presence of C2.

Figure 6

4. Integrated interpretation

In this section, we interpret the geo-electrical features identified in the 2D inversion of the
MT data in light of seismic reflection, well-log, gravity and magnetic data.

In 2009, the exploration wildcat well 6705/10-1 was drilled to a depth of 3775 meters (MD
below sea level), reaching the early Cretaceous Lange Formation (Figure 6) at the
southernmost end of profile P5 (Figures 2a and 5). The borehole resistivity data provide a
direct calibration for resistivity values from the MT models. Since the MT data along profile P5
show no evidence for anisotropy, it is valid to compare them with the borehole resistivity data.
In Figure 6, the borehole resistivity log is plotted with a pseudo-well extracted from P5 at the

same location. The long period trends from the resistivity bore-hole data are clearly in agreement with the pseudo-well, while the vertical resolution of MT data is naturally much lower than the borehole measurements.

Figure 7

Figure 7 shows two seismic reflection profiles co-rendered with resistivity sections along P1 and P3. The low resistivity values of the shallow C1, on average 0.8 Ωm , indicate that this approximately flat layer has likely been infiltrated by sea water through faults and fractures and that porosity remains high in these relatively unconsolidated sediments. By integrating reflection seismic and well-log data, we can confidently interpret conductor C1 as sea water-intruded, post-tectonic Paleogene, Neogene and Quaternary sediments. This interpretation is supported by an MT model test showing that C1 and R1 can be separated by a sharp boundary inserted at the seismic interface marking the base of the Paleogene sediments. We interpret R1, with resistivity values from 10 to 100 Ωm and thickness approximately 5 km (Figure 7b), as Cretaceous sediments and possibly Jurassic sediments. These have been heavily faulted and intruded by igneous intrusions, with saucer-shaped and transgressive sill complexes readily identified on seismic data (Figure 7b; Planke et al., 2005). Low-angle detachment faults act as a geo-electrical boundary between R1 and C2 in the north Gjallar Ridge.

C2 offers resistivity ranging from 0.1 to 0.8 Ωm and lies between 8 and 13 km depth, within lower Cretaceous and/or pre-Cretaceous sediments and possibly crystalline basement (Figure 7). The implications of such low resistivity value at 10 km depth are discussed in details in the next section. A clear horizontal seismic reflection interpreted as layer-parallel igneous intrusions is located right in the middle of conductor C2 at 9 km depth (Figure 7a). Sharp boundary tests of the MT models show that the high conductivity associated with C2 can be

confined within the zone of seismically imaged sills. This allows us to envisage that a dramatic enhancement of electrical conduction occurs in the vicinity of layer-parallel igneous intrusions and its host rock. In Figure 7b, the seismic image is almost transparent at C2 depth due to the presence of shallower, saucer-shaped and transgressive sills reflecting most of the acoustic energy (Planke et al., 2015).

In figure 8, the total magnetic field shows no prominent magnetic anomalies in the Vøring Basin, while the main anomalies are caused by the transition to the oceanic domain, where break-up magmatism dominates the magnetic signature. The northern Gjallar Ridge has a positive signature on the 50 km filtered Bouguer anomaly map, while the Vigrid syncline has a negative one. These observations tend to show that the gravity field is mainly influenced by the thickness of the sediments in the studied area. From the 50 km high-pass filter Bouguer anomaly and total magnetic field maps, we can infer that the prominent conductor C2, linked to pre-Cretaceous intruded sedimentary strata, does not correlate with a density nor magnetic anomalies in the studied area (Figure 8).

Figure 7

5. Discussion

In the discussion, we focus primarily on the implications of the mapped high conductivity at mid-crustal depth (i.e. C2; Figures 5 and 7). Using an integrative approach, we exploit the strengths of the available geophysical datasets and tentatively assign a geological interpretation to the debated deep sedimentary and basement lithology in the Gjallar Ridge (Gernigon et al., 2004; Peron-Pinvidic and Osmundsen, 2016). To our knowledge this is the first time in the Vøring Margin that electrical properties of deep sedimentary, mid-crustal units have been integrated with their elastic, density, and magnetic properties.

5.1. Electrical conduction mechanism

Resistivity values of C2 (0.1-0.8 Ωm) are considered very low as mid-crustal resistivity typically spans 10-10⁴ Ωm (Bedrosian, 2007). High conductivity, inversely low resistivity, at crustal depths is generally caused by: (1) saline pore fluids, (2) partial melt or (3) well-connected conductive minerals (Bedrosian, 2007; Selway, 2014). Conductor C2 occurs between 8-13 km (Figures 5 and 7). In this case, saline pore fluids are ruled out as porosity and brine conductivity would have to be unreasonably high at this depth (Archie, 1942; Heinson et al., 2005; Storvoll et al., 2005).

Surface heat flow decreases from the Trøndelag Platform and the North Sea (Figure 1b) towards the centre of the Møre and Vøring basins (Pascal, 2015), indicating that the presence of melt at C2 depth is unlikely in the north Gjallar Ridge. Furthermore, the seismic interpretation clearly indicates that C2 is located at a significantly shallower depth than the lower crustal body (LCB) discussed by Mjelde et al. (2009) (Figure 7).

Serpentinised mantle rocks can also account for high electrical conductivity in the Earth's interior. In the Vøring Basin, low Vp/Vs-ratios, low stretching factor of the continental crust and uncertainty regarding the water budget (Gernigon et al., 2004, 2015 ; Mjelde et al., 2003) argue against the presence of exhumed serpentinised mantle. The petrology of the lava flows in borehole data 642E from the Ocean Drilling Program (ODP) shows evidence of continental contamination that cannot be easily explained by the presence of a shallow serpentinised mantle underneath the subaerial lava flows (Meyer et al., 2009; Abdelmalak et al., 2016a).

We therefore postulate that C2 is related to the presence of well-connected conductive mineral networks. Iron-sulphides (i.e. pyrite, pyrrhotite), iron-oxydes (i.e. magnetite) and

graphite are among the most conductive minerals, with resistivities ranging from 10^{-4} to 10^{-1} $\Omega\cdot\text{m}$ (Pearce et al., 2006; Telford et al., 1990).

5.2. Electrical conduction at low-angle detachment

Heinson et al. (2005) states that a similar mid-crustal conductor is likely caused by graphite films precipitated along a detachment in the Exmouth Plateau, north-west of Western Australia. Muñoz et al. (2008) also discusses plausible causes of high conductivity in detachment zone. Although the presence of low-angle detachment level is known in the north Gjallar Ridge, the conductive anomaly C2 is centred ~ 2 km deeper (Figure 7). The very dense site spacing (1 to 3 km) of our survey and the high conductivity of C2 makes the depth to its upper boundary a very robust model feature. In addition, there is no evidence for a low-angle detachment level in the Vigrid Syncline (Gernigon et al., 2004) where Myer et al. (2013) have observed another highly conductive body along a MT profile (Figures 1 and 8). We therefore consider unlikely a link between a low-angle detachment and high conductivity in the outer Vøring area.

5.3. Mineralization processes and igneous intrusions

Sediments intruded by magmatic bodies often develop contact metamorphic aureoles containing minerals that are constrained by the intrusion's thermal field (Aarnes et al., 2015). Various studies have connected the occurrence of massive magnetite and metallic sulphides with thermal aureoles (Gillett, 2003; Katz et al., 1998). Massive graphite can also precipitate in volcanic intrusions and host-rocks in the presence of carbon-bearing aqueous fluid phases (Luque et al., 1998, 2012; Ortega et al., 2010). A clear horizontal seismic reflector is interpreted as layer-parallel igneous intrusions located in the middle of conductor C2 at 9 km depth (Figure 7a), which reinforces the mineralization hypothesis.

5.4. Massive, sheet-like pyrrhotite deposit interpretation

We go further and postulate that C2 is related to the presence of a massive, sheet-like iron-sulphide deposit composed mainly of pyrrhotite, similar to the Noril'sk Ni-Cu-PGE ore deposits in the Siberian Tunguska Basin (Malitch et al., 2014; Naldrett, 1999). At Noril'sk, iron-sulphide deposits (mainly composed of pyrrhotite) occur in close association with igneous intrusions in the contact aureoles or in the intrusion itself. Some deposits can extend several tens of kilometres with thicknesses of up to 45 m and are mainly composed of pyrrhotite (90%) (Malitch et al., 2014). A 45 m thick layer with a conductivity of 10^3 S/m would yield a 4.5×10^4 S conductance (i.e. the product of conductivity and thickness). In our study, C2 is on average 5 km thick and ~ 3 S/m, giving a conductance of 1.5×10^4 S, in the same order of magnitude as the Noril'sk pyrrhotite deposits.

Figure 8

Due to the presence of minerals with high magnetic susceptibilities within the intrusions and in the host-rocks, igneous sills are typically associated with strong magnetic signatures (Morgan, 2012). This is not necessarily the case for seafloor massive sulphides (SMS) deposits which can display a weaker or missing magnetization on basalt-hosted hydrothermal sites (Szitkar et al., 2015 and reference therein). However, the processes involved in the formation of demagnetized SMS, such as alteration due to the circulation of hot hydrothermal fluids, are different from the one occurring in deep intrusions thermal aureoles.

We claim that the presence of Noril'sk type massive pyrrhotite deposits in the north Gjallar Ridge can explain the region's low magnetic response (Figure 8a). Thermo-kinematical modelling (Gernigon et al., 2006) suggests that temperatures near the top LCB, below C2, (Figure 7) are between 250 and 400°C. The Curie temperature of magnetite is $\sim 585^\circ\text{C}$, meaning

that if a thick magnetite deposits was invoked as the cause of C2, its magnetic signature should be readily seen on aeromagnetic maps. Aubourg et al. (2012) propose the existence of diagenetic magnetic windows in argillaceous sedimentary rocks during burial. They proposed that magnetite forms continuously from ~50°C to 200–250°C whereas pyrrhotite starts forming at ~200°C at the expense of magnetite with magnetite production ceasing above 250°C. For these two reasons and as opposed to Myer et al. (2013), we rule out magnetite dominated deposits as an interpretation of very high conductivity in the north Gjallar Ridge. Pyrrhotite exists in various forms with different ferromagnetic properties (Makovicky, 2006; Pearce et al., 2006). A study of the magnetic behaviour of natural pyrrhotite in the KTB ultra-deep borehole in Germany (Kontny et al., 2000) suggests that non-magnetic pyrrhotite overwhelmingly dominates at depths >8080 m or at in-situ temperatures >230°C.

5.5. Volcanogenic graphite

Fluid-deposited graphite in intruded carbonaceous host-rocks can also explain the low magnetic response in the north Gjallar Ridge. However, this type of massive volcanogenic graphite is often precipitated in thin veins and steep pipe-like bodies (Luque et al., 1998, 2012; Ortega et al., 2010). As opposed to Noril'sk stratified iron-sulphides, this type of graphite deposits (veins, vertical pipe-like body) is unlikely to provide a well-connected mineral network required to increase electrical conduction across regional distances.

Figure 9

Conclusion

In this work, we report on the results of an MT survey acquired as a by-product of a 3D controlled-source electromagnetic survey in the Vøring Basin in 2014. Dimensionality analysis

has shown that the data have 1D to 2D characteristics, which justifies the 2D modelling approach used in this work. The MT inversion results are in good agreement with a resistivity log available in the uppermost 4 km in the survey area. Inversions of all profiles display consistent and distinct geo-electrical features. All profiles image a prominent, high conductivity anomaly (0.1-0.8 $\Omega\cdot\text{m}$) in heavily intruded sedimentary strata (8-13 km). We integrate electrical resistivity properties from MT data with Bouguer anomaly, magnetic field and seismic reflection data to conclude that the high conductivity anomaly is likely caused by sheet-like, pyrrhotite-dominated deposits along layer-parallel igneous intrusions in pre-Cretaceous sediments and possibly extending to upper continental crust (Figure 9). Given the lack of a magnetic signature, we exclude magnetite deposits as a possible cause of the conductor since the sill complex, readily identified on seismic data, lies well above the Curie isotherm (585°C) of the north Gjallar Ridge (Gernigon et al. 2004). Like in the outer Vøring Basin and Exmouth Plateau, further investigation may show that very high conductivity at mid-crustal depth is a common characteristic of intruded basins.

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583

584 Figure 1: (A) Regional location map of the Mid-Norwegian margin based on GEBCO bathymetry
585 topography grid (GEBCO, 2014). Main physiographic elements are indicated: AR: Ægir Rigde; JMMC:
586 Jan Mayen micro-continent; JMFZ: Jan Mayen Fault Zone; MMH: Møre Marginal High; VMH: Vøring
587 Marginal High. (B) Structural elements map of the Mid-Norwegian margin. VS: Vigrid Syncline. Location
588 of the survey area in the north Gjallar Ridge (red square) in Figure 2. The solid blue line, denoted M
589 (2013), is the MT profile presented by Myer et al. (2013).

590

591 Figure 2: (A) Layout of the electromagnetic (EM) survey and bathymetry. The black squares represent
 592 the receiver on the seabed. The red lines indicate the profiles chosen for 2D inversion. The solid blue
 593 lines are two depth-converted seismic reflection profiles: the northernmost seismic line is MNR11-
 594 90659; the southernmost is MNR11-90643. (B) Structural elements and faults map in the survey area.
 595 The overall orientation of the faults and Gjallar Ridge indicates a strong SW-NE trend. Same legend as
 596 Figure 1.

597

598 Figure 3: Phase tensors shaded by skew at (A) $T=5.5 \text{ s}$ and (B) $T=523.9 \text{ s}$ for all receivers. The
 599 magnetotelluric phase tensor is depicted as an ellipse whose principal axes indicate the
 600 electromagnetic strike with a 90° ambiguity (Caldwell et al., 2004). The light grey solid lines represent
 601 the main faults in the north Gjallar Ridge (Figure 2b).

Figure 4: (A) Rotated apparent resistivity and phase data of three receivers located along profile P3 (Figure 2a). The response of the final inversion model of P3 is indicated by solid lines. (B) Pseudo-sections of the observed and modelled MT data along P3. The three black rectangles in the first quadrant indicate the locations of the receivers displayed in Figure 3a.

Figure 5: Final 2D MT resistivity models along profiles P1, P2, P3, P4 and P5 (Map inserted to the bottom right of this figure and Figure 2a). Each receiver is represented by a white triangle. Only a 1D subset of the data was inverted along P5. The main geo-electrical elements are denoted as C1, C2, R1 and R2.

Figure 6: Comparison between resistivity logs from exploration well 6705/10-1 (Figure 2) and resistivity pseudo-well derived from the MT inversion along P5 at the same location. The vertical axis is depth (in meters). Note that the gas discovery "Asterix" is associated with high resistivity in the Shetland Group on the zoomed image to the right. Data courtesy of the Norwegian Petroleum Directorate, accessed via the DISKOS database.

618

619 Figure 7: (A) Resistivity section along P1 co-rendered with seismic line MNR11-90659. (B) Resistivity
 620 sections along P3 co-rendered with seismic line MNR11-90643. White triangles indicate EM receivers.
 621 Black arrows point toward sills. Black lines indicate major faults. TC: Top Cretaceous; BCU: Base
 622 Cretaceous Unconformity; LCB: Lower Crustal Body. For location of profiles refer to Figure 2a.

623

624 Figure 8: (A) Total magnetic field in the outer Vøring area. The red lines represent the MT profiles
 625 whereas the dark blue lines show the location of two MNR11 seismic lines. The MT profile discussed
 626 by Myer et al. (2013) is also displayed. (B) 50 km high-pass filtered Bouguer gravity anomaly map in
 627 the outer Vøring area.

Figure 9: Schematic illustration showing the main stratigraphic units and sills in the north Gjallar Ridge. The black solid lines represent detachment faults. LCB: Lower crustal body.