3D-modeling of the Skellefte District Using Electrical, Potential Field and Reflection-seismic Data

A Basis for 4D-modeling of Mineral Belts

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Cover picture: 3D local-scale model of the central Skellefte district based on resistivity, IP and Potential field data.
Abstract

As a part of a 4D modeling project, two studies with different scopes were conducted in the central Skellefte district (CSD), northern Sweden. The aim of the studies is to create a basis for a better understanding of the spatial relationship between geological structures and mineralization and to construct a 3D and 4D geology model of the area.

In the first study, we used geo-electrical data to define the geological structures at depth down to 430 m. The inversion of the resistivity and Induced Polarization (IP) data indicated a number of lithological contacts, which required further constraints prior to constructing the final 3D model. Hence we measured petrophysical properties including density, magnetic susceptibility, resistivity and IP of 154 samples, selected from drill-holes in vicinity of the resistivity/ IP profiles, to constrain the model. Forward resistivity models were then acquired using the resistivities measured on drill-cores, to test the response of different geological scenarios in 3D after inversion. The gravity and magnetic response of the resistivity/ IP models was then calculated to constrain the models down to 1.5 km depth. The models were then modified, until reaching a consistency between geo-electrical and potential field data. The result indicated the possibility of three sulphide mineralization zones within the highly conductive parts at depth ≤ 500 m. The result also helped to determine the geometry of the contact between sedimentary rocks of the Vargfors basin and volcanic rocks of the Skellefte Group.

In the second study, we tested geological models based on interpretation of reflection- seismic data using potential field data (down to 5 km depth) as well as electrical data (down to 430 m depth). The gravity and magnetic data especially benefitted the interpretation where no reflector is indicated, or poor-quality reflectors could not contribute to the understanding of major lithological contacts along the main faults and shear zones in the CSD. Moreover, the gravity and magnetic data, add significant information to reveal the spatial relationship between the Skellefte volcanics, metasedimentary rocks of the Vargfors Group and two intrusive structures of TIB gabbro-diorite and granitic rocks, which were poorly indicated on the reflection-seismic profiles.

The results further indicate that joint interpretation of the integrated geophysical techniques can provide remarkable information regarding geometry of structures, which is a base for constructing 3D and 4D geological models.
Acknowledgements

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Saman Tavakoli, October 2011, Luleå-Sweden
List of appended papers

Paper I

• 3D modelling of the central Skellefte district, Northern Sweden; an integrated model based on the electrical, potential field and petrophysical data. Submitted to” Applied Geophysics”.

Saman Tavakoli, Sten-Åke Elming, Hans Thunehed

Paper II

• 3D joint modeling of the gravity and magnetic data in the central Skellefte District in a regional scale; a model based on interpretation of reflection-seismic data. Submitted to” Applied Geophysics”.

Saman Tavakoli, Tobias E. Bauer, Sten-Åke Elming, Hans Thunehed, Pär Weihed

Paper III


Tobias E. Bauer, Saman Tavakoli, Mahdieh Dehghannejad, Maria Garcia, Pär Weihed

Other contribution not included

• Pre-1.87 Ga development of crustal domains overprinted by 1.87 Ga dextral transpression in the Palaeoproterozoic Skellefte district, Sweden – constraints from structural and geochronological investigations. Submitted to “Precambrian Research“.

Pietari Skyttä, Tobias E. Bauer, Saman Tavakoli, Tobias Hermansson, Jenny Andersson, Pär Weihed

• Skellefte mining District in 3D; results from integrated interpretation of potential field, resistivity/IP and reflection-seismic data. Extended abstract, accepted for poster presentation in 22nd ASEG conference and exhibition 2012, Brisbane, Australia.

Saman Tavakoli, Tobias E. Bauer Pietari Skyttä Sten-Åke Elming, Hans Thunehed Pär Weihed.
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**References**
1. Introduction

1.1. The aims of this study

Although the interest for 3D modeling of geological structures has increased during the recent years, it still considers as a young technique. With this study, we aim to construct a 3D geology model using geophysical and geological data, which will be a base for constructing a 4D (Time domain) model of the geological evolution of the region. Integration of a wide range of geophysical data, including resistivity, Induced Polarization (IP), gravity, magnetic, reflection-seismic and Magnetotelluric (MT) as constrains to the model provided a thorough dataset to construct a well-constrained and geologically meaningful 3D model of the central Skellefte district. Hence, the main objectives of this study are to:

1. Provide detailed geophysical information to construct a local-3D model of the geology in the central Skellefte district to determine the spatial relationship between Skellefte Group, Vargfors Group and Jörn intrusive rocks (resistivity/ IP, magnetic and gravity data).
2. Provide satisfactory constrains to create a regional-3D geology model of the central Skellefte district down to 5 km depth (gravity, magnetic and reflection-seismic data).
3. Explore potential locations for prospecting the volcanic-hosted massive sulphide (VMS) deposits and suggesting candidate locations for drilling (resistivity/ IP data).
4. Integrate the result from this study with additional geophysical data within the project to construct a final 3D model of the central Skellefte district which will be base for constructing a 4D geology model of the central Skellefte district.

1.2. Approach towards constructing a 3D geology model of the central Skellefte district

Today, 3D modeling of geological structures is a fast-growing approach to better understand the geology as it provides a well-expressed image of the subsurface geology by integrating geophysical and geological data into one model. In the central Skellefte district, the interest towards exploration at greater depths has increased during recent years, since most of the near-surface minerals have been already explored and mined, (Skyttä et al., 2011). Hence 3/4D-modeling of the geology provides an image of the geometry of the top 3 km of the Earth crust and its evolution through the geological history (Weihed et all., 2010). Consequently, the VINNOVA 4D project started in 2008 in Sweden to construct a 3/4D-GIS model of the central Skellefte district. The procedure and input data for creating such 3/4D-models of the geological structures in the central Skellefte district is presented in Fig 1. However, this study presents a part of the result from the VINNOVA 4D project through two different approaches:

- Local-scale 3D modeling of the central Skellefte district in the vicinity of the Vargfors basin which was conducted using electrical, potential field, petrophysical and drill-hole data.
- Regional-scale 3D modeling of the central Skellefte district to identify and explain the major lithological contacts and shear zones in the central Skellefte district using potential field and petrophysical data, which could not be modeled on basis of the reflection-seismic data.

1.3. General geology of the central Skellefte district

The rocks in the Skellefte mining district (Fig.2) consist of Palaeoproterozoic supracrustal and intrusive rocks formed in a volcanic arc setting (Allen et al., 1996) and metamorphosed during the Svecokarelian Orogeny (Kathol and Weihed, 2005). The dominating unit is the 1.90 – 1.88 Ga, ore-bearing Skellefte Group, which comprises mainly felsic metavolcaniclastic
and metavolcanic rocks (Allen et al., 1996; Kathol and Weihed, 2005). The overlying 1.88 – 1.86 Ga Vargfors Group mainly consists of metasedimentary rocks. The lower parts of Vargfors stratigraphy are dominated by mudstones, sandstones and monomict conglomerates formed from turbiditic currents, whereas the stratigraphic higher parts comprise polymict conglomerates formed in an alluvial fan environment (Bauer, 2010). The Vargfors Group meta-sedimentary rocks in the central Skellefte district define the distinct Vargfors syncline. The contact relationship between Skellefte Group metavolcanic rocks and Vargfors Group metasedimentary rocks range from conformable to unconformable to faulted. The massive sulphide mineralizations formed as sub-seafloor replacement in the uppermost parts of the Skellefte Group stratigraphy.

Intrusive rocks in the central Skellefte district are dominated by metatonalites to metagranites of the Jörn intrusive complex, which consist of several intrusive phases (GI-GIV; Wilson et al. 1987; Gonzáles Roldán, 2010). The faulted contact between Jörn metatonalites and Vargfors metasedimentary rocks is defined by a rim of mafic metavolcanic and metavolcaniclastic rocks (Bauer, 2010).

Structural analysis (Bauer et al., under review) and reflection-seismic investigations (Dehghannejad et al., under review) revealed a series of WNW-ESE-striking, south-dipping inverted normal faults formed during Palaeoproterozoic extension and reactivated during subsequent crustal shortening. These faults where cross-cut by north-dipping, late compressional break back faults. A set of NE-SW-striking syn-extensional faults, segments

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Fig. 1. Geological and geophysical data and modeling procedure for constructing 3D model of the central Skellefte district.
the central district into distinct fault blocks. Locally, the faults are accompanied by mafic volcanic activity. Strain was partitioned during N-S crustal shortening (c.f. Bergman Weihed 2001) resulting in low-strain domains with mainly open, asymmetric synclines and minor anticlines, and high-strain domains in the vicinity of faults with tight to isoclinal folds and local overturned strata.

The youngest major phase of intrusions comprise 1.82 – 1.78 Ga late-to-post-tectonic Revsund-type intrusive rocks, which are part of the Transscandinavian Igneous Belt (TIB). The youngest major deformation event at 1.80 Ga was partitioned into the major NE-SW-trending faults due to E-W-crustal shortening (Bergman Weihed, 2001; Weihed et al., 2002; Skyttä et al., 2010; Bauer et al., under review).

2. Methods

2.1. Principle of resistivity/ IP method

The geo-electrical methods are based on the differences in electrical conductivity of the rocks, which is a function of mineral composition, porosity, degree of water saturation and fluid composition within the rocks (Kneisel, 2006).

Based on the physics of Ohm’s law, the field measurement can be carried out through directing current flow from current electrodes to the ground. By measuring the voltage difference with two potential electrodes, the apparent resistivity \( (\rho_a) \), which indicates the distribution of the apparent resistivity of the subsurface material, can then be calculated using the current flow \( (I) \) and potential difference between the electrodes \( (\Delta \phi) \):
The parameter $K_g$ in the above equation is the geometric factor which depends on the electrode arrays. The general equation for calculating $K_g$ is as follow (Loke, 2010):

$$K_g = \frac{2\pi}{\left( \frac{1}{r_{c,p}} - \frac{1}{r_{c,p}} - \frac{1}{r_{c,p}} + \frac{1}{r_{c,p}} \right)}$$  \hspace{1cm} (2)

$r_{cp}$ in the equation indicates the distance between the current electrodes ‘c’ and the potential electrodes ‘p’. With increasing distance between electrodes, only a part of the current penetrate down to the deeper layers, which increases the difference between apparent resistivity and the true resistivity. Electrode spacing, and subsequently the depth of penetration, together with resistivity contrast between rocks are the governing parameters in variation of the apparent resistivity (Dahlin, 2001). Therefore, the field data need to be inverted or otherwise modelled in order to make interpretations based on the real resistivity. Beside resistivity, measurement of the IP is normally carried out. With IP measurements it is possible to detect the very conductive minerals even in low concentrations, while those minerals may be missed in the resistivity surveys (Loke, 2010; Leroux et al., 2007).

2.2. Magnetic and gravity methods

The magnetic method is based on determining local anomalies in the geomagnetic field, which reflects the variation in the intensity of magnetizations in different rocks (Parasnis, 1997). Although most of the rock forming minerals are basically none-magnetic, certain rock types contains magnetic minerals enough to produce magnetic anomalies (Kearey et al., 2002). However, the magnetization itself is originating from an induced part, which is the product of the magnetic susceptibility and the intensity of the Earth Magnetic Field (EMF), and from a remanent magnetization in the rocks. In explorations, the local variation of magnetic field is compared and separated from the International Geomagnetic Reference Field (IGRF).

The gravity method is based on Newton’s law of acceleration, which states a force between two point masses located at distance of $r$, is equal to $G m_1 m_2 / r^2$. In this equation $m_1$ and $m_2$ are two point masses and $r$ is the distance between them, $G$ is constant of universal gravitation, which is constant all over the Earth and is equal to $6.672 \times 10^{-11}$ (m$^3$kg$^{-1}$s$^{-2}$). Therefore, any mass unit located in vicinity of body will experience the acceleration. The unit of acceleration in SI is m s$^{-2}$ and in c.g.s is cm s$^{-2}$ (mgal). The gravity measurement is carried out both as absolute measurements of gravity in certain stations in national scale (absolute gravity) and as relative measurements (relative gravity) using a gravimeter, which has an accuracy of around 0.01 mgal (Lowrie, 1997). The gravity data needs to be corrected before interpretation, which includes correction for the latitude, topography (terrain), Bouger plate (excessive mass), free air (elevation), instrument drift and tidal effects.

2.3. Reflection-seismic method

Reflection-seismology is a technique used to find the depth to reflecting surfaces and the seismic impedances of the subsurface rocks. The principle is such that a seismic signal is generated at a known time and is reflected from the boundaries between rock layers with different seismic impedances. Geophones, recording the waves, are spread out along profiles with a critical distance from the shot-points. Within this distance, only reflected waves are recorded. However, the direct waves, which travel along the surface, are disturbing the signals, hence being considered as noise. Nevertheless, to overcome the large amount of noises, in each recording spot a group of geophones is normally used, which moves further when a whole profile is measured. The survey is normally planned in a way that reflection-seismic profiles are normal to the strike of geological structures (Lowrie, 1997).
2.4. Forward and Inverse problem of the geophysical data

The observed geophysical fields originating in the Earth’s interior, i.e. gravity, magnetic, electromagnetic, resistivity, and seismic waves are dependent of the physical properties of the rocks (Zhdanov, 2002).

The conventional approach for interpretation of geophysical data includes creating different geological models and then comparing the theoretical fields, which are computed for these models with the observed data. Numerical modeling of the geophysical data for a given model parameters is called the forward problem (Simpson and Bahr, 2005). This will enable us to predict geophysical data for various geological structures (Fig. 3).

The main aim of geophysical observation is to determine parameters (i.e. geometry, depth extent etc.) of geological structures from the geophysical data. Though, it is a challenging problem due to the complexity of the Earth’s interior (Zhdanov, 2002). Nonetheless, one can yet get an approximate image of the geological structures by a more or less simple model and try to determine the model parameters from the data. This is the inverse problem of geophysical data (Fig. 3). However, the inverse problem is more complicated compared to forward problem since there are a number of possible solutions instead of only one (Lanza and Meloni, 2006). Accordingly, the goal to solve the inverse problem is to determine the distribution of the physical property or properties that give rise to the data. However, in reality, this is not possible since (i) only a limited number of data can ever be recorded and (ii) the recorded data are not always accurate. Nevertheless, an approximate solution, which provides an estimated model, can be still found. The success of geophysical interpretations depends greatly on our skill in creating reasonable geological models, and to solve the corresponding inverse problems effectively.

One example of inverse and forward problems in geoscience is temperature variation as a function of depth beneath the Earth’s surface explained by Menke (1989). The example explains the phenomena of the increasing temperature with the depth; in other words, temperature \( T \) is related to depth \( Z \) as:

\[
T (Z) = aZ + b
\]

Fig. 3. A schematic image of forward and inverse modeling of geophysical data (Jones, 2007)
Which $a$ and $b$ are numerical constants. Considering the values for the two constants as $a = 0.1$ and $b = 25$, then we can simply explain the forward problem by evaluating the formula at any depth. The inverse problem for this example would be to determine $a$ and $b$ on basis of temperature measurements made at different depths, which i.e. are known from bore-hole data (Menke, 1989).

**Forward problem:**
Model parameters $\rightarrow$ model $\rightarrow$ prediction of data

**Inverse problem:**
Data $\rightarrow$ model $\rightarrow$ estimate of model parameters

Therefore, the inverse modeling can provide information about the unknown numerical parameters that incorporate into creating a model; whereas it is unable to provide the geometry of the model itself. Nevertheless, it is an efficient way to assess the model certainty, or to discriminate a model between several existing ones.

### 3. Data collection

#### 3.1. Resistivity / IP data

Based on prior knowledge about the geological stratigraphy of the central Skellefte district, surface geology, potential field and bore-hole data, geo-electrical data were collected along two profiles (Profiles I and II; Fig. 2) in the central part of the Skellefte district. The southern profile (Profile I; Fig. 4a), 6.8 km long, starts from the SW in the felsic volcanic rocks and continued towards NE into the sedimentary rocks of the Vargfors basin. The northern profile (Profile II; Fig. 5a), 5.6 km long, starts in the sedimentary rocks of the Vargfors basin and continued towards the NE, where it cuts the Jörn granodiorites. Both profiles were aimed to cross the major lithological contacts, particularly the contact between the Vargfors Group (sedimentary rocks) and the Skellefte Group (metavolcanic rocks). There are a number of criteria to consider when choosing an appropriate electrode configuration, depending on the noise, desired depth of investigation and the physical properties of the lithologies. In this study, a Pole-dipole electrode array is preferred, since a rather good horizontal coverage (suitable to detect vertical features) will be acquired with this configuration in comparison to a Wenner array, and also a higher signal strength is expected compared to a dipole-dipole array. With a Pole-dipole array there will be a low EM coupling and the depth of penetration will be high (Loke, 2010). The distance between the remote current electrode $C_3$ and the current electrode $C_1$ in this study is $\approx 5$ km to minimize the effect of electromagnetic coupling caused by $C_2$. In this study, there are five potential electrodes, which constitute four dipoles. Using the multi electrode array, the first potential electrode is located 200 m from $C_1$ with the dipole length of 200 m. The first two dipoles are 200 m long, while the last two are 400 m, making the total length of the dipole lines 1.4 km.

#### 3.2. Gravity and Magnetic data

Gravity data in this study are provided by the Swedish Geological Survey (SGU) and Boliden Mineral AB. These data are unevenly distributed in the central Skellefte district; however, the average spacing between gravity data close to the profiles in this study is $\approx 200$ m, while the average spacing between gravity stations in the regional scale is $\approx 800$ m. The whole gravity dataset is tied to the Swedish gravity network RG 82. The aeromagnetic data are provided by the SGU as well. The geomagnetic field is measured from an altitude of 30 m having 200 m line spacing and a station spacing of 40 m. An average intensity of 52612 nT, inclination of 76.7° and declination of 6.7° is estimated for the Earth’s magnetic field in the study area (Kathol and Weihed, 2005).

Separation of the regional anomaly from the field data is an important part of the gravity and magnetic interpretation. There are various methods used to estimate the regional fields, among which, polynomial surface, minimum curvature and finite element are the most common methods (Xu et al., 2009). In this
study, the regional gravity and magnetic fields are defined using a polynomial surface generator function (Encom Technology, 2002). Accordingly, the regional field is defined as a polynomial surface of specified order, which is defined using all data within the central part of the Skellefte district. The choice of polynomial order is directly controlled by the complexity of the field. However, polynomial orders higher than four can lead to an incorrect regional field, where the influence of sharp local variations interfere the calculation of the regional field (Encom Technology, 2002). We used a 2nd order polynomial surface for calculating of the regional field for the two profiles; however, the calculated field is manually modified to a minor extent during interpretation, this to make a better model fit. The area used for calculating the regional field is defined beyond the two profiles in order to take the effect of deep and large scale features into account. The definition of area and choice of the polynomial order was determined after testing different alternatives. During the modeling, the remanent magnetization is neglected and background density of 2670 kg/m³, which is common average density for the crustal rocks, is used.

3.3. Petrophysical data

Density and magnetic susceptibility data from rocks collected at 695 outcrops in the Kristineberg and 1038 outcrops in the central Skellefte district have been compiled in a database by SGU. We have used these data for interpretation of the gravity and magnetic data where required; however, petrophysical data in direct relation to the profiles were sparse and an accurate petrophysical characterization of the rocks cannot be obtained without a detailed geophysical description of the main geology. Moreover, electrical properties are not included in that database. Thus, we did a petrophysical study of core samples and a geological characterization of the rocks collected from drill-holes close to the two resistivity/IP profiles. Petrophysical investigations in the Skellefte district and Kristineberg were also conducted by Enmark and Niska (1983), investigations that included density and magnetic susceptibility of rock samples collected from the Gallejaur intrusion and surrounding rocks in the Skellefte district. The petrophysical data collected from the drill-cores includes density, magnetic susceptibility, resistivity and IP (Table 1).

<table>
<thead>
<tr>
<th>Rock type</th>
<th>#</th>
<th>Density (kg/m³)</th>
<th>Magnetic susceptibility K (SI)</th>
<th>Resistivity (Ω-m) at 0.4 Hz</th>
<th>IP (mrad) at 0.4 Hz</th>
</tr>
</thead>
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<tr>
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<td>2735</td>
<td>0.00032</td>
<td>11550</td>
<td>12</td>
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<td>0.00067</td>
<td>5804</td>
<td>125</td>
</tr>
</tbody>
</table>

3.4. Reflection-seismic data

Reflection-seismic data were acquired along three nearly parallel N-S trending profiles (C₁, C₂ and C₃; Fig. 2) in the central Skellefte district. Each profile is ~ 32 km long and located 3-7 km apart from the adjacent profile. The profiles were oriented perpendicular to the major geological structures in the CSD (Dehghannejad et al., under review). The shot spacing was chosen 25 m unless the spacing needed to be increased
due to lack of accessibility to roads, making a total number shot points of 3000. The receiver spacing was chosen of 25 m, except for the parts along Skellefte River in profiles C₁ and C₂, where it had to be increased.

4. Results and Discussion

In the first study, we applied geo-electrical measurements to acquire a local-scale 3D model of the CSD down to 430 m. The model was further constrained by potential field and petrophysical data, which were collected from drill-cores along or close to the two profiles down to 1.5 km. In the second study, we tested the result from interpretation of reflection-seismic data, by calculating the gravity and magnetic response of these models for three profiles in the CSD. The reflection-seismic data imaged the crust down to a depth of 4.5 km and we modeled the subsurface geology using potential field data down to 5 km at depth (Fig. 10b, c, d and Fig. 11b).

4.1. 3D local-scale model of the central Skellefte district down to 1.5 km

The interpretation of the resistivity/IP data was controlled by the resistivity determined on samples collected from drill-cores, which provided an initial model of the CSD down to 430 m depth. However, there are still a number of uncertainties within the models, which can be constrained by testing the models using additional geophysical data. Hence, we used the magnetic data to constrain and modify the previous models at shallow depths (500 m) and then the gravity data along the same profiles were modelled to test and extend the models down to 1.5 km. As a starting model for the magnetic test, the resistivity/IP models were thus used (Figs. 4b and 5b). The magnetic susceptibilities obtained from petrophysical measurement in the laboratory (Table 1) were used as representative for the different lithologies. The drill-core data with the lithological description are used to constrain the gravity and magnetic models. The result from 2D resistivity/IP inversion provided a model in which the contrast between resistivity values of the different layers enabled us to determine the geometry of the subsurface with a high resolution down to the 430 m. The IP data provided models where the small conductive areas could have been missed in the resistivity models. This was especially important when detecting high conductive zones, including the N Norrilden and two other potential ore deposits along profile (I).

Potential field modelling improved previous models obtained from resistivity and IP data, especially where the resistivity of the thin layers were masked, due to the large electrode spacing. The gravity interpretation improved the models based on the resistivity/IP and magnetic data (RIM model) by providing information about the deeper parts of the crust (1.5 km). Although the density of the metasediments and felsic volcanic rocks do not differ significantly, the resulting density model is well constrained by the RIM model. The interactive forward modelling of density and magnetic data thus improved the preliminary models based on geophysical data (Figs. 4e and 5e).

Felsic volcanic rocks (Skellefte Group)

The felsic volcanic rocks of the Skellefte Group are the most common rocks along the two studied profiles in the CSD. The locally fault-contact between the Vargfors basin sedimentary rocks and the underlying felsic volcanic rocks of the Skellefte Group is shown in Figs. 4e, 5e and Fig. 6. The felsic volcanic rocks are regularly interrupted by sequences of basalts belonging to the Skellefte Group, which in general have a thin depth extent. However, due to significant difference between density and magnetic susceptibility of the felsic rocks (2735 and 0.00032; SI) and basalts (2894 and 0.0012; SI), they can be well recognized using gravity and magnetic data. On the electrical profiles felsic volcanic rocks can be simply recognized with their low IP. The felsic volcanic rocks apparently extend beyond 1.5 km depths.

Basalt (Skellefte Group)

The Skellefte Group basalts indicate different degrees of magnetization, while they have a similar high range of densities (Table 1). Along
Fig. 4. (a) The geology around profile (I). (b) The model resistivity section after inversion. (c) The model IP section after inversion. (d) The model magnetic section after inversion and applying constrains to the model (RIM model). (e) The model gravity section after inversion and applying constrains to the model (RIMD model). (Black Line: Observed anomaly, Red line: Calculated anomaly, Blue line: Regional anomaly).
Fig. 5. (a) The geology around profile (II). (b) The model resistivity section after inversion. (c) The model IP section after inversion. (d) The model magnetic section after inversion and applying constrains to the model (RIM model). (e) The model gravity section after inversion and applying constrains to the model (RIMD model). (Black Line: Observed anomaly, Red line: Calculated anomaly, Blue line: Regional anomaly).
profile (II), they occur close to the contact with the Jörn granodiorite, which increases the magnetic field intensity. The occurrence of basalts within felsic volcanic rocks of the Skellefte Group at the Skellefte-Jörn contact is also indicated by the resistivity high and low IP along profile (II).

**Granitoid (Jörn Intrusion)**

The depth of the bottom contact of the Jörn intrusive complex exceeds 1.5 km, which is consistent with previous interpretations (i.e. 4-6 km by Wilson et al., 1987). Below the Jörn granodiorites, we suggest a rock with a higher density and magnetic susceptibility compared to granodiorite, but indicating a lower magnetic signature compared to the basalt. This is consistent with the common occurrences of tonalite (Fig. 5e) within the Jörn intrusive complex as described by Wilson et al. (1987). Therefore, parts of Jörn intrusion are probably associated with tonalitic rocks, which increased significantly the magnetic and gravity anomaly. The tonalite depth extension is estimated of ~ 1.4 km, in order to be consistent with both gravity and magnetic data.

**Sedimentary rocks (Vargfors Group)**

The sedimentary rocks along the two profiles consist of sandstone-mudstone, conglomerate and unspecified sediments. Thanks to the different electrical properties of these sedimentary rocks, the inter-sedimentary contact can be modelled with a good accuracy (Figs. 4d and 5d). However, the magnetic and gravity data did not contribute much in dividing different sedimentary rock types, since their density and magnetic susceptibility is similar. The maximum depth of sedimentary rocks in the Vargfors basin along two profiles is estimated to be ~ 700m.

**Ore deposit (sulphide mineralization)**

On basis of interpretation of geo-electrical data, three potential sulphide mineralizations were suggested along profile (I) (Fig. 4d). All three mineralizations are located at less than 500 m depth. Furthermore, one of these mineralizations is located at the contact between Skellefte Group and the Vargfors basin (Fig. 4d and Fig. 6), which according to previous geological investigations (i.e. Weihed, 2010) has a high

![Fig. 6. 3D local-scale geology model of the central Skellefte district around the Vargfors basin](image-url)
possibility of hosting VMS deposits. The IP measured on drill-core samples shows a good consistency with the IP acquired in the field, although the samples in general indicate lower IP than field data. Although potential field data did not disagree with the suggested depth and dimensions for the sulphide mineralizations on basis of the resistivity/IP data, but they also did not create any gravity or magnetic anomaly because of their small dimension and great depth extent.

4.2. 3D regional-scale model of the central Skellefte district down to 5 km based on interpretation of reflection-seismic data

A joint interpretation of the gravity and magnetic data is here performed (Fig. 10b, c and d) to test interpretation of the reflection-seismic data (Tavakoli et al., under review. b). The result of this test not only provided good constrains for the models suggested by reflection-seismic data, but also indicated to be a useful technique to provide information where no reflectors are observed or only poor quality reflectors exist. Previous investigations of reflection-seismic data in crystalline rocks showed low signal/noise ratio, which made the interpretation complicated due to i.e. strong metamorphism, alteration and folding (Malehmir and Bellefleur, 2009), which can be seen in part of the results of seismic sections in this study.

Potential field modeling along reflection-seismic profile C₁

According to the gravity model along profile C₁ (Fig. 7b), the SW dipping gabbroic-dioritic intrusion (II; Fig. 7c) is located on the boundary to the Vargfors Group mudstones-sandstones (I; Fig. 7c). The gravity and magnetic interpretations suggest a maximum depth of ~3.5 km for this gabbroic-dioritic intrusion. The local minima in magnetic and gravity anomalies at ~4 km (Fig. 7a and b) indicate intercalation of different rocks with similar petrophysical signatures to those of rhyolite (III; Fig. 7c) into the gabbroic-dioritic intrusion (II; Fig. 7c).

The model based on potential field data suggests that reflection R₁ cuts through the TIB-gabbroic-dioritic intrusion and continues through the contact between the gabbroic-dioritic intrusion and the sedimentary rocks (Fig. 7d). Furthermore, the model shows that the reflector R₂ coincides with the contact between the TIB-gabbro and Skellefte Group rhyolite (bodies II and IV; Fig. 7c). The gravity and magnetic data indicate that the reflector R₄ may be related to the contact between rhyolite and unspecified felsic volcanic rocks (IV-V; Fig. 7c and R₄; Fig. 7d). Metasedimentary rocks of the Vargfors basin (body IX; Fig. 7c) is not reflected in the gravity or magnetic data, due to the similarity in density and magnetic susceptibility with the surrounding rhyolites. At 17-17.6 km along profile C₁, an increase in the magnetic and gravity field was observed (Figs. 7a and 7b). These anomalies are interpreted to originate from a basaltic rock with higher magnetic susceptibility and density (Table 1), compared to the surrounding rhyolitic rocks, and it continuous down to ~2 km depth. The bottom contact of this basalt with the felsic volcanic rocks might be related to the reflector R₇ (R₇; Fig. 7d). At ~20.8 km, the north-dipping contact between the mafic volcanic rocks, dominated by basalt, and the Jörn granitoid (XI-XII; Fig. 7c) fits well with the potential field model and reflector R₆ (Fig. 7d). The gravity and magnetic models suggest a depth extension of ~1-1.2 km for the Jörn intrusion along C₁ (XII; Fig. 7c).

The fit between measured and calculated gravity from density/geological model along profile C₁ is reasonable, as indicated by the RMS=4.57%, whereas the fit between measured and calculated magnetic field from the model is somewhat more poor (RMS=8.68%).

Potential field modeling along reflection-seismic profile C₂

The significant gravity high, between 1-5 km along the profile C₂ (Fig. 8b) is well correlated with the expected density high of the gabbroic rocks. This rock (body II; Fig. 8c) continues down to ~3 km depth. The bottom contact of the surrounding granitoid (body I; Fig. 8c) is sub-horizontal in the model, while the contact with the adjacent sedimentary rocks of the Vargfors...
Fig. 7. Potential field modelling along profile C1 based on interpretation of reflection-seismic data
(a) Magnetic model of profile C1 (b) Gravity model of profile C1 (c) The depth section and geological structures. (d) Integration of the seismic reflectors and model based on interpretation of potential field data.

Jörn Granitoid (0.00065) Felsic volcanic - unspecified (Skellefte Group) (0.00007)
Gabbro - diorite (Transscandinavian Igneous Belt) (0.00033-0.002)
Basalt (Skellefte Group) (0.0012)
Sulphide mineralizations (0.001)
Rhyolite (Skellefte Group) (0.00015)
Sandstone - mudstone (Vargfors Group) (0.00012)
Jörn Granitoid (2760 kg/m³) Felsic volcanic - unspecified (Skellefte Group) (2600 kg/m³)
Gabbro - diorite (Transscandinavian Igneous Belt) (2910 kg/m³)
Basalt (Skellefte Group) (2850 kg/m³)
Sulphide mineralizations (3000 kg/m³)
Group (I-III; Fig. 8c) is steep (60 - 70°). At 11.6 km, close to the southern contact of the Vargfors basin (V-VIII; Fig. 8c), the gravity field increases (Fig. 8b), which cannot be explained by the density contrast between Skellefte Group rhyolites and Vargfors Group conglomerates. The gravity modeling indicated that any further extension of the Vargfors basin conglomerate results in a misfit between the measured and calculated gravity field; whereas the magnetic model did not affected considerably, indicating that Vargfors basin extends at depth down to ~ 1 km (body VIII; Fig. 8c) and overlies Skellefte Group basalt (body VII; Fig. 8c) with higher density compared to conglomerate and rhyolite. Contrasting, Dehghannejad et al. (under review) relates the reflectors R₈ and R₁₅ (Fig. 8d) to the synformal structure of the Vargfors basin, extending to a depth of ~1.7 km. To explain the sharp gravity anomaly at 13-13.5 km, the basalt has to extend through the Vargfors basin to the surface. There is, however, no field evidence to support that, also, no reflector reaches the surface in this part of profile C₂. Hence, one possibility is that the gravity data is not correct. From 17.2 km towards the northern end of the profile, the sub-horizontal reflector R₅ (Fig. 8d) might be an indication of the bottom contact between Jörn GI-type intrusive rocks and an underlying basalt which crops out at ~ 17.3 and 18.2 km (bodies X and XI; Fig. 8c). Reflector R₆ agrees well with the potential field models and corresponds to a north-dipping break-back fault, which marks the southern contact of the Jörn intrusive complex. This is also comparable with the reflector R₅ in profile C₁ (Bauer et al., 2011; Dehghannejad et al., under review). Mafic volcanic and volcaniclastic rocks that intruded along the fault (Bauer et al., under review) are believed to be the source of the magnetic high at 18.2 km. The magnetic and gravity models for the Jörn intrusive complex along C₁ suggest that intrusion-depth increases towards its center (body XI; Fig. 8c).

The calculated field from the gravity model of profile C₂ reasonably well fit the measured data (RMS=8.42%) and the calculated magnetic field shows a somewhat better fit (RMS=5.68%) compared to the fit for profile C₁. Potential field modeling along reflection-seismic profile C₃

Due to the crooked CDP line for the seismic profile C₃, the gravity and magnetic modeling of profile C₃ was divided into two parts (Fig. 11a). A series of SW-dipping reflectors (R₁, R₂ and R₄) in the southern part of profile C₃ (Fig. 9d), have earlier been interpreted as inverted normal faults, similar to interpretations along profile C₁ (Fig. 7d) (Dehghannejad et al., under review). The reflector R₄, which Dehghannejad et al. (under review) interpreted as a shear zone, reaches the surface at CDP 700, seems associated with south-dipping basalt (Fig. 9d). The reflector R₆ (Fig. 9d) coincides well with the suggested contact for R₃ in profile C₁, which indicates a south-dipping inverted normal fault at the contact between rhyolites and unspecified felsic volcanic rocks (III-VI; Fig. 9d). The gravity low observed at ~ 8-12 km along profile C₁, can also be observed in profile C₃ between 8.2-17.7 km (Fig. 9b), is interpreted as a south-dipping unspecified felsic volcanic rock (body VI; Fig. 9c). The reflector R₄₀, which Dehghannejad et al. (under review) correlated with the reflector R₄ in C₁ and C₂, fits well with the suggested potential field model for profile C₁, hence we suggest the reflector R₄₀ (Fig. 9d) to be the expression of a south-dipping contact between unspecified felsic volcanic rock and rhyolite to the north (bodies VI and III; Fig. 9c). Both gravity and magnetic data suggests that Gallejaur-type basalt (body XII; Fig. 9c), with high densities and high magnetic susceptibilities, underlie the mafic and felsic intrusions (bodies XIII and XIV; Fig. 9c) of the Gallejaur complex. At ~ 22 km, the magnetic high and gravity low (Figs. 9a and 9b), indicates higher magnetic susceptibility but lower density of the mafic intrusive rocks compared to the Gallejaur-type basalts (body XIII; Fig. 9c). The reflector R₄₀ could therefore be related to the contact between Gallejaur-type basalt and underlying felsic volcanic rocks (XII and III; Fig. 9c and Fig. 9d).

The fit of the data calculated from the density models to the measured data along this profile was good (RMS=3.27%) and better than the corresponding fit of the models to the magnetic data (RMS =8.67%).
Fig. 8. Potential field modelling along profile C2 based on interpretation of reflection-seismic data (a) Magnetic model of profile C2 (b) Gravity model of profile C2 (c) The depth section and geological structures (d) Integration of the seismic reflectors and model based on interpretation of potential field data.
Fig. 9. Potential field modelling along profile C3 based on interpretation of reflection-seismic data (a) Magnetic model of profile C3, (b) Gravity model of profile C3, (c) The depth section and geological structures, (d) Integration of the seismic reflectors and model based on interpretation of potential field data.
Fig. 10. Integration of reflection-seismic data and models based on potential field data down to 5 km. (a) Crooked and CDP lines for profiles C1, C2, and C3. (b) The sliced 3D model of CSD based on gravity and magnetic data cut by depth section of reflection-seismic data for profile C1. (c) The sliced 3D model of CSD based on gravity and magnetic data cut by depth section of reflection-seismic data for profile C2. (d) The sliced 3D model of CSD based on gravity and magnetic data cut by depth section of reflection-seismic data for profile C3.
Fig. 11. (a) Illustration of the CSD, surface geology and seismic profile lines which potential field modeling is conducted along these lines (b) 3D solid model of the CSD down to 5 km.
5. Conclusion

Integration of the electrical and potential field data proves efficient for modeling the often complicated geological structures. The cost effectiveness of these techniques and their good response to the lithology variation indicates that they are suitable for modeling subsurface geology; they are also capable to discover rather small ore mineralization (particularly IP), which could have been masked if modeling was performed using only potential field data.

Using the reflection-seismic data as a preliminary model for potential field modeling, only partially explained the variations in potential field data, since none-reflective areas or areas with weak reflectors did not contribute much in revealing the major lithology contacts, hence unable in providing constrains in parts of the model. Nevertheless, the capability of modeling geological contacts at great depths is undoubtedly successful when combining reflection-seismic and potential field data. However, even though previous studies indicated the effectiveness of using reflection-seismic data for detecting sulphide mineralizations (i.e. Malehmir and Bellefleur, 2009), reflection-seismic data in this study was unable to indicate the sulphide mineralization zones suggested from the resistivity/IP data (Tavakoli et al., under review. a).

The key geophysical and geological outcome of the present thesis on the basis of integrated geophysical and geological interpretation is as follow:

- The good horizontal coverage and meanwhile good precision of the data at greater depths in comparison to other electrode configurations, proved the Pole-dipole successful in detecting particularly vertical contacts in the CSD.
- Potential field modelling improved previous models obtained from resistivity and IP data, especially where the resistivity of the thin layers were masked, due to the large electrode spacing.
- Resistivity contrast between the Skellefte Group (felsic volcanic) and Vargfors Group (sedimentary) rocks indicated synformal geometry of the Vargfors basin, extended down to ~700 m depth along Profile (I). Moreover the contact between felsic volcanic rocks of Skellefte Group and Jörn granitoid along profile (II) was modeled using potential field data. The result suggested intercalation of tonalite, dipping parallel to granitoids and extending down to 1.4 km at depth.
- IP and resistivity data proposed three high conductive zones, including the Norrliden N deposit and two additional potential sulphide mineralizations along profile (I).
- Results from testing gravity and magnetic response of reflection-seismic data verified the model where good reflectors have been observed, particularly if a reflector is observed consistently on all three profiles. Elsewhere, potential field data constrained the model, indicating some of the major lithology contacts which were not revealed from the interpretation of reflection-seismic data.
- The result from regional-scale modeling indicates that the southern part of the CSD is dominated by a series of south-dipping inverted normal faults with a depth extent of ~3-5 km. In the central part of profiles, none of the diffractions $D_1$, $D_2$, $D_3$ and $D_4$ which were observed on reflection-seismic profile $C_3$ could be explained with the model based on potential field data.
- A set of north-dipping break-back faults forms the southern contact of the Jörn intrusive complex, are consistent with potential field and reflection-seismic data (reflection $R_6$), and a depth extent ranging from ~1-2 km is suggested. The potential field data indicate that Gallejaur complex forms a thin, sheet like unit with a depth extent of ~800 m, which could not be correlated to any of the reflectors along profile $C_3$. 

6. Future work

The following investigations are planned for the future work:

- In order to model the Skellefte district in more details and expand the model in CSD, integration of further geophysical and geological data are a must. The MT data from Skellefte district, which image the subsurface down to ~ 4 km at depth, can thus provide great constrains to the result of this study. There are additional data from resistivity/ IP measurements which were conducted during autumn 2010. These data include extension of the previous resistivity/ IP profiles, which not only can constrain the current models, but also provide information about possible massive sulphide mineralizations at greater depths (down to 1.5 km).
- Further geological investigations to better comprehend timing of the intrusions and formation of basins which is especially important for modeling the district in 4D.
- Drilling one of the proposed locations of sulphide mineralization, if successful, can prove the efficiency of using geo-electrical techniques in the CSD for future exploration of base metals.
- P-wave velocity of 154 samples from the drill-cores were previously measured in the geophysics laboratory at LTU, these data should be used in the interpretation of seismic data, since they provide the velocity of different rock types along the three seismic profiles.
- One of the key objectives of this project, i.e. constructing a 3/4D-model of the Skellefte district, is the basis for future exploration of base metal minerals. A Multi Criteria Analysis (MCA) of the candidate locations for ore mineralizations can therefore indicate zones with a high probability of hosting VMS deposits.

References


3D Modelling of the Central Skellefte district, Northern Sweden; an Integrated Model based on the electrical, potential field and petrophysical data

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3D Modelling of the Central Skellefte district, Northern Sweden; an Integrated Model based on the electrical, potential field and petrophysical data

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Abstract

The central Skellefte district (CSD) is a part of a major ore-bearing district in northern Sweden. Studying the depth and patterns of the contact relationship between the two major stratigraphic units of the CSD, the Skellefte Group and the Vargfors Group, is a key issue to understand the geometry and structure of the area and to guide exploration of base metals. In this study, we interpret geoelectrical data collected along two profiles and magnetic and gravity data obtained from the database of the Swedish Geological Survey (SGU) to reveal contact relationship and depth extension of the major geological structures. Petrophysical analyses of the different lithologies were conducted on samples from the database of the SGU. Electric resistivity, induced polarization (IP), magnetic susceptibility and density were determined on 154 core samples representing the different lithologies of the area. The resistivity/IP data were acquired to define structural relations down to a maximum depth of ~ 430 m. The major contact between sediments of the Vargfors basin and volcanic rocks of the Skellefte Group were outlined from the inversion of the resistivity/IP sections, suggesting a synform boundary between the Vargfors Group and Skellefte Group. The contact relationship between the felsic and mafic volcanic rocks of the Skellefte Group is also understood with the help of the resistivity/IP data. The resistivity models were tested using the magnetic data and magnetic susceptibility inferred on the resistivity bodies. The result suggests a good correlation between the initial resistivity model and the magnetic field calculated from that model. In order to map the deeper parts of the crust (down to ~ 1500 m), gravity data were compiled along the resistivity/IP profiles. The integration and interpretation of all data resulted in a 3D-model, which improved the basic understanding of the geometry of CSD. Based on previous geological investigations, the potential ore deposits are believed to be found along the volcano-sedimentary contact. This 3D model can thus be used for the base metal exploration, finding the locations of potential sulphide deposits and give a better understanding about spatial relationship between different geology contacts.

Keywords: Skellefte district, 3D-modelling, Geoelectrical method, Potential field data, Inverse modelling
1. Introduction

Located in the north of Sweden, the Skellefte district is one of the major ore producing regions in the country. Being one of the most mineralized Palaeoproterozoic arc systems in the world, the Skellefte district is also a valuable target for exploration of base metals (Weihed, 2010). The area of this study covers a part of the central Skellefte district, which comprises metavolcanic rocks of the Skellefte Group and metasedimentary rocks of the Vargfors Group. The study area is bordered to the north by the Jörn intrusive complex and the Gallejaur complex and to the south by metagreywackes of the Bothnian Supergroup (Kathol and Weihed, 2005) (Figs. 1 and 2a). The Skellefte district is known to form the boundary between a deep-marine sedimentary environment to the south (Bothnian basin), and a continental landmass to the north, the Archean continent (Lundberg 1980). In spite of being subject to numerous geological and geophysical studies (Allen et al., 1996; Kathol and Weihed, 2005; Montelius et al., 2007; Bauer et al., 2011; Bauer et al., under review), the stratigraphy of the central Skellefte district and the contact pattern of its components are not fully understood. Nevertheless, the structural evolution of the area has been studied previously, and can provide geological constraints to the new interpretations (Jessel 2001; Skyttä et al., 2010; Bauer et al., 2011; Bauer et al., under review).

Application of geoelectrical imaging for exploration has increased by the increasing interest for exploration at depth. Geoelectrical surveys have proven very efficient in mapping the often complicated geometry of the subsurface (Magnusson et al., 2010). Because of the expected significant contrast in resistivity between the sediments and the volcanic rocks in the central Skellefte district, resistivity/IP measurements were conducted using the pole-dipole configuration (Fig. 2b). The survey was conducted in a way that the variation of resistivity/IP could be imaged down to a maximum depth of ~ 430 m.

In addition, potential field data were compiled to constrain the model created based on the interpretation of the resistivity/IP data. Modelling of the magnetic field was used to further constrain the resistivity model and different resistivity bodies were given a specific magnetic susceptibility obtained by laboratory measurements on drill-core samples from bore holes close to the profiles. To extend the resistivity/IP/magnetic (RIM) model downwards, gravity data were interpreted using the RIM model as a start model which contributes to construction of final model (RIMD) which is also constrained with the density data.

This study aim at: (i) compare existing geophysical with geological data to create a local-scale 3D solid model of the inferred geology in the central Skellefte district. The local-scale geophysical model will be a start model for future 3D-geological modelling of the area on a regional-scale (ii) Construct a 3D model which forms the basis for the future exploration of potentially economical ore deposits, which are thought to be formed in the uppermost parts of Skellefte Group stratigraphy, associated with the volcano-sedimentary contact. (Allen et al., 1996; Weihed, 2010).

2. Geological settings and previous studies

The Skellefte district (Fig. 1) comprises Palaeoproterozoic metamorphosed supracrustal and associated intrusive rocks (Kathol and Weihed, 2005). The central part of the Skellefte district is composed of two major rock units, the Skellefte Group and the Vargfors Group. The Skellefte Group is the lowest stratigraphic unit in the central Skellefte district, overlain by the Vargfors Group (Allen et al., 1996; Kathol and Weihed, 2005; Bauer et al., 2011). Bauer et al. (under review) explains the presence of tonalite clasts from the Jörn intrusive complex within metasedimentary rocks of the Vargfors Group (1873+ 10 Ma; Skiöld, 1988) as an indication of the short
time span between the emplacement of the Jörn intrusive complex, its uplift and erosion, followed by the sedimentation of the Vargfors Group.

Allen et al. (1996) describes the main lithological units within the Skellefte Group as emplaced lava domes, porphyritic cryptodomes as well as volcaniclastic rocks, with mainly rhyolitic and less basaltic, andesitic and dacitic composition. Sedimentary intercalations including mudstones, volcaniclastic siltstones, sandstones, breccias, conglomerates and volcaniclastic rocks with carbonate matrix are found as well (Allen et al., 1996; Kathol and Weihed, 2005; Montelius et al., 2007). A potential field study of the spatial associations of the volcanic massive sulphides in the volcanic rocks of the Skellefte group indicated that the VMS deposits are mainly associated with the felsic volcanic rocks of the Skellefte group (Carranza and Sadeghi, 2010).

The metasedimentary rocks of the Vargfors Group mainly consist of argillites, sandstones, conglomerates and subordinate carbonate-rich mudstones to conglomerates (Dumas, 1986; Allen et al., 1996; Bauer et al., 2011). The two crosscutting fault-sets, a NW-SE-striking and a NE-SW-striking, divide the Vargfors basin into different fault-bound compartments with different internal stratigraphies and deformation patterns (Bauer et al., 2011).

The Jörn granitoids are located at the northern border of the central Skellefte district and are composed of tonalitic to granodioritic to granitic rocks (Wilson et al., 1987). According to Wilson et al. (1987), they belong to the early Svecofennian Granitoid suite, which intruded into a volcanic-arc environment on the southern

Figure 1. Simplified geological map of the Skellefte district and surroundings (modified after Kathol et al., 2005).
margin of a major Proterozoic continental crust. To the north the central Skellefte district borders to felsic volcanic and metasedimentary rocks of the Arvidsjaur Group. The existence of clasts from the Jörn intrusive complex is typical within the upper parts of Vargfors Group stratigraphy (Bauer et al., 2011; Bauer et al., under review). Bauer et al. (under review) suggests that a high-strain zone, associated with multi-phase mafic volcanic activity, forms the southern contact of the Jörn intrusive complex.

The application of geophysical techniques for exploration purposes in the Skellefte district dates back to 1920, when electromagnetic surveying was initiated, since the ore bodies were believed to be good electrical conductors. This led to a number of exploration campaigns in the area (Kathol and Weihed, 2005). Gravity surveys for iron exploration in the central Skellefte district started in 1950, following earlier surveys for bedrock mapping. Airborne magnetic surveys have been conducted in the area since 1960, while petrophysical studies of the rock in the Skellefte district started in 1970. These studies included density, magnetic susceptibility and gamma ray spectrometry measurements (Kathol and Weihed, 2005). The geological surveys of the Nordic Countries have also carried out petrophysical analyses of some 30000 samples on which density and magnetic susceptibility were determined on rocks from Sweden, Norway and Denmark (Henkel, 1991). Previous petrophysical studies in the Skellefte district have demonstrated a direct relationship between the magnetic properties and the density of the rocks (Enmark and Niska, 1983).

The gravity data in this study are provided by the Geological Survey of Sweden (SGU) and Boliden Mineral AB. These data are unevenly distributed in the central Skellefte district; however, the average spacing of the gravity data close to the profiles in this study is ~ 200 m, while the average spacing between gravity stations in a regional scale is ~ 800 m. The aeromagnetic data has been also provided by SGU. The magnetic field was measured from an altitude of 30 m with a line distance of 200 m and a station spacing of 40 m. An average intensity of 52612 nT, inclination of 76.7° and declination of 6.7° is estimated for the Earth’s magnetic field in the study area.

3. Methodology

An initial 2D model of the geological structures has been created on the basis of the surface geology map, bore-hole data and resistivity/IP field measurements. This set of data together provided enough constraints to derive a preliminary 2D model down to a depth of 430 m. This initial model was further tested and improved using resistivity/IP data from rock samples measured in the laboratory. In Section 4.2, the improved model is again tested using aeromagnetic data and magnetic susceptibilities determined of samples from different lithologies. Gravity data along the same profiles were then used with the previously tested model (RIM) as an input to extend the model down to depth of ~1500 m. A final model (RIMD) was then constructed on basis of the 2D models from the two profiles.

3.1. Apparent Resistivity/IP Measurement in the field

The geoelectrical methods are based on the differences in electrical conductivity of the rocks, which is a function of mineral composition, porosity, degree of water saturation and fluid composition within the rocks (Kneisel, 2006). Based on the physics of Ohm’s law, the field survey is carried out by directing current flow from current electrodes to the ground. By measuring the voltage difference with two potential electrodes, the apparent resistivity (ρₐ), which indicates the distribution of the apparent resistivity of the subsurface material, is then calculated using the current flow (I) and potential difference between the electrodes (Δφ):

$$\rho_a K_g \frac{\Delta \phi}{I}$$  \hspace{1cm} (1)

The parameter K_g in the above equation is the geometric factor which depends on the electrode arrays. The general equation for calculating K_g is as follow (Loke, 2010):
\[ K_g = \frac{2\pi}{\left( \frac{1}{r_{c1p1}} + \frac{1}{r_{c2p1}} + \frac{1}{r_{c3p2}} + \frac{1}{r_{c2p2}} \right)} \] (2)

\([r_{cp}]\) in the equation indicates the distance between the current electrodes ‘c’ and the potential electrodes ‘p’. With increasing distance between electrodes, only a part of the current penetrate down to the deeper layers, which increases the difference between apparent resistivity and the true resistivity. Electrode spacing, and subsequently the depth of penetration, together with resistivity contrast between rocks are the governing parameters in variation of the apparent resistivity (Dahlin, 2001). Therefore, the field data need to be inverted or otherwise modelled in order to make interpretations based on the real resistivity. Beside resistivity, measurement of the Induced Polarization (IP) is normally carried out. With IP measurements it is possible to detect the very conductive minerals even in low concentrations, while those minerals may be missed in the resistivity surveys (Loke, 2010; Leroux et al., 2007).

Given the prior knowledge of the geological stratigraphy of the region and based on the surface geology map, bore-hole data and other geophysical information, electrical measurements along two separate resistivity/IP profiles were conducted in the central Skellefte district (Fig. 2a). The southern profile (Profile I) is 6.8 km long, beginning in the SW in felsic volcanic rocks and extending towards NE into metasedimentary rocks of the Vargfors basin. The northern profile (Profile II) is 5.6 km long, starts in the metasedimentary rocks of the Vargfors basin and continues towards the NE where it cuts the Jörn granodiorites. The direction of both profiles were decided with the aim to cross the major lithological contacts, particularly the contact between the Vargfors Group and the Skellefte Group. There are a number of criteria to consider when choosing an appropriate electrode configuration, such as the noise, desired depth of investigation and the physical properties of the lithologies. In this study, a Pole-dipole electrode array was preferred, since a better horizontal coverage (suitable to detect vertical features) should be acquired with this configuration compared to a Wenner array. Furthermore, a higher signal strength will be obtained compared to a dipole-dipole array. With a pole-dipole array there will be a low EM coupling and the depth of penetration will be high (Loke, 2010). The distance between the remote electrode \( C_2 \) and the current electrode \( C_1 \) is set at 5 km to minimize the effect of electromagnetic coupling caused by \( C_2 \) (Fig. 2b). In this study, we used five potential electrodes, which constitute four dipoles. Using the multi electrode array, the first potential electrode is located 200 m from \( C_1 \) with the dipole length of 200 m. The first two dipoles are 200 m long, while the last two are 400 m, making the total length of the dipole lines 1400 m (Fig. 2b).

Due to the complicated collection of data and the high cost of field work for 3D resistivity surveys, the present survey was conducted in 2D. However, 3D information can be derived from the 2D profiles by combining results from two resistivity profiles, surface geological data and drill-hole data (Loke, 2010; Leroux et al., 2007). Although it is possible to calculate the depth of investigation using the geometric factor, there is no formula to calculate the exact depths knowing only the electrode separations (Parasnis, 1997). However, Edwards (1977) explains the relationship between the potential electrode spacing ‘\( a \)’ and depth of investigation ‘\( Z_e \)’ for different types of pole-dipole arrangement. For a pole-dipole array, considering ‘\( a \)’ as potential electrode distance (dipole length) and ‘\( n_{\text{max}} \)’ as dipole separation factor, the relationship between \( n \) and \( a \) is defined as:

\[ n_{\text{max}} = \frac{L_{c1p_{n-1}}}{L_{p_{n-1}p_n}} \] (3)

Where \( n_{\text{max}} \) is maximum dipole separation factor, \( L_{c1p_{n-1}} \) is the maximum distance between the current electrode and potential electrode, and \( L_{p_{n-1}p_n} \) is the dipole length. Since \( n_{\text{max}} \) here is 2.5 (\( P_4-P_5=400 \) m and \( P_1-P_4=1000 \) m), the largest electrode separation \( (a_{\text{max}}) \) is 400 m (Fig. 2b) and the relationship
between maximum penetration depth ($Z_{e\text{,max}}$) and electrode spacing ($a_{\text{max}}$) for $n = 2.5$ according to Edwards (1977) is $Z_{e}/a_{\text{max}} = 1.12$, it gives a maximum penetration depth of $\sim 430$ m. Fig. 3 indicates the pseudo-section of the field data for profiles (I) and (II).

### 3.2. Petrophysical Data

Density and magnetic susceptibility data from rocks collected at 695 outcrops in the Kristineberg and 1038 outcrops in the CSD have been compiled in a database by SGU. We have used these data for the interpretation of the gravity and magnetic data where required, however, petrophysical data in direct relation to the profiles were sparse and an accurate petrophysical characterization of the rocks cannot be obtained without a detailed description of the main geology. Moreover, electrical properties are not included in that database. Thus, a petrophysical study was performed on core-samples collected from drill-holes (Fig. 2a) as well as a geological characterization of the rocks. Petrophysical investigations in the Skellefte district and Kristineberg have previously been conducted by Enmark and Niska (1983), and includes density and magnetic susceptibility of rock samples from the Gallejaur intrusion and the surrounding rocks in the Skellefte district.
Figure 3. Apparent resistivity and IP pseudo-section of the raw data acquired in the field (a) The apparent resistivity pseudo-section for profile (I). (b) The IP pseudo-section for profile (I). (c) The apparent resistivity pseudo-section for profile (II). (d) The IP pseudo-section for profile (II).
3.2.1. Petrophysical measurements of the samples from drill-holes

In order to obtain a better control of the petrophysical properties of the key rock units within the CSD 154 samples were collected from 27 drill-holes located in the vicinity of the two profiles (Fig. 2a) and in the major lithological contact zones. Resistivity was measured using a resistivity meter. An alternating current (0.1 mA-0.6 mA) with frequencies of 0.1 Hz, 0.4 Hz and 4 Hz is sent through the samples and the potential difference is then measured at each frequency. The resistance at each of the frequencies was then calculated to obtain the resistivity. The resistance and IP of the samples were calculated based on 0.4 Hz for samples in the laboratory. The resistivity of the rocks is a product of several parameters including porosity, content of water, resistivity of the water, clay content and content of conductive minerals. The chemistry of electrolyte, which fills the pores in the saturated rocks, plays a crucial role in controlling the electrical conductivity of the rocks (Nover, 2005) and as a result, the resistivity determined in the laboratory is expected to be higher than the apparent resistivity measured during field work. This is mainly due to the differences in the concentration of salt in the water and also the water content. Therefore, before starting the resistivity measurement in the lab, all of the samples are soaked in water in order to fill the voids of the porous rocks with tap water, which is more resistive than natural water in the rock. The conductivity of the tap water has thus been measured to determine the influence of the porosity in the resistivity results. The conductivity of the material is the reciprocal of electrical resistivity \( \sigma =1/\rho \), where \( \rho \) = Electrical resistivity (Ω) and \( \sigma \) = Electrical conductivity (mho/cm or S/cm) and has the unit of siemens (S).

The median conductivity value of the tap water used in the laboratory at LTU was 315 µS/cm at water temperature 18.5 °C. This is close to the resistivity of water from wells in northern Sweden (Sjödahl et al., 2005).

Based on Archie’s law, the resistivity of a porous non-clayey material can be estimated by the following formula:

\[
\sigma=a \times \sigma_w \phi^m
\]

Where \( \sigma \) is electrical conductivity of the porous rock, \( \sigma_w \) is the conductivity of the brine, \( \phi \) is the bulk porosity of the rocks and \( a \) and \( m \) are tortuosity factor and cementation exponent of the rock, respectively, which are constants and depend on the nature of the rock. According to Salem and Chilingarian (1999), the cementation factor greatly depends on the shape and type of the grains. The less spheritic the samples are, the higher value of \( m \) is expected. The type of grains also controls the \( m \) considerably. A high amount of clay content, as well as existence of mixture of sediments, increases the \( m \). According to Salem and Chilingarian (1999), \( m \) can be considered 2 to 2.3 for clean sands and 1.8 to 3 for compacted sandstones. The tortuosity factor (\( a \)) is directly related to the cementation factor, that is, in a media with many dead end pores the current cannot pass through the pores, which is an indication of a high (\( a \)). In a very rough approximation, ‘\( a \)’ can be taken equal to 1 and ‘\( m \)’ to 2 (Telford et al., 1991). However, since the conductivity of the natural water in rocks in the study area is unknown (There are only water conductivity data available from drill holes in the SGU database that located > 100 km from the profile locations ranging about 5000-6000 µS/cm), the true resistivity based on Archie’s law cannot be precisely calculated. The variation of the pressure, temperature and water quality at the different depths within the cores makes it complicated to correctly quantify the resistivity of different lithologies.

Since the IP survey in the field was conducted using the time-domain method, IP values of the samples from lab, which are measured in frequency-domain, have to be converted into the time domain. In the time-domain IP, the residual voltage V(t) measured at the time t after the current is turned off is compared with the constant voltage V(c) for the time interval 0.1s-10 s. The result of this type of IP is then shown as mV/V. This ratio tends to be low and
hence is sometimes written as a percentage. There are several methods which can be used to convert frequency-domain IP data to the time-domain, though they are not widely used since the theoretical analysis of the IP effects is yet not fully understood (Telford et al., 1991). However, here a qualitative interpretation is still done using the frequency IP data from laboratory work.

To measure the density of the rocks, the samples were first soaked in water for 48 hours. The densities of the rocks were then determined on the basis of Archimedes principle and the arithmetic means, median values and standard deviations were calculated (Table 1). The magnetic susceptibility was measured using a Kappabridge (KLY-3, AGICO).

On the basis of the petrophysical properties determined on rock samples, the rocks were divided into 7 major groups (Table 1). Due to the often much skewed distribution of the data, the median value has been used to represent the magnetic susceptibility, resistivity and induced polarization (IP) of the rocks, while arithmetic means have been used for density data (Table 1). In order to simplify the classification of the rocks, samples showing high degree of alteration, intensive silicification, presence of clasts and intensive sulphidication are ignored in the petrophysical analysis.

According to Table 1, felsic volcanic rocks show the lowest density (2735 kg/m³), however, conglomerate and sandstone also demonstrate low densities, especially their median value. Basalts with high magnetic susceptibility (high K basalts) indicate lower density than low (K) basalt; however, the standard deviation for low (K) basalt is rather high. The mean density of the andesite is close to that of the sedimentary rocks. For determining the density of the Jörn granitoid which includes parts of Profile (II), we used surface petrophysical data based on 56 measurements of samples, which gave the density of 2670 kg/m³. Moreover, previous studies indicate a density of 2630-2800 kg/m³ for the Jörn granitoids, reflecting a composition varying from diorite to granite (Wilson et al., 1987).

The highest magnetic susceptibility is found among the high (K) basalts (Table 1). This is followed by low (K) basalt and sandstone from which the second one has larger standard deviation. Sulphide mineralization and andesite indicate similar behaviour in terms of magnetic susceptibility. Felsic volcanic rocks and conglomerate have the lowest magnetic susceptibilities. However, for the rocks represented by a limited number of samples, it may be expected to see wide distribution of petrophysical data. Wilson et al. (1987) explain the wide range of magnetic susceptibility for Jörn granitoids (K = 2.6-370 ×4П× 10^-5) through the heterogeneous composition of the outer zone, which experienced a secondary alteration.

High (K) basalt demonstrates the highest resistivity (Table 1) followed by andesite and low (K) basalt. Felsic volcanic rocks have a resistivity of 11550 Ωm. Sulphide mineralization and sedimentary rocks including

<table>
<thead>
<tr>
<th>Rock type</th>
<th>#</th>
<th>Density (kg/m³)</th>
<th>Magnetic susceptibility K (SI) at 0.4 Hz</th>
<th>Resistivity (Ω-m) at 0.4 Hz</th>
<th>IP (mrad) at 0.4 Hz</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td>Mean</td>
<td>σ</td>
<td>Median</td>
</tr>
<tr>
<td>Felsic volcanic rocks</td>
<td>91</td>
<td>2735</td>
<td>42</td>
<td>2730</td>
<td>0.00032</td>
</tr>
<tr>
<td>Basalt (High K)</td>
<td>10</td>
<td>2857</td>
<td>44</td>
<td>2833</td>
<td>0.04660</td>
</tr>
<tr>
<td>Basalt (Low K)</td>
<td>18</td>
<td>2894</td>
<td>53</td>
<td>2893</td>
<td>0.0012</td>
</tr>
<tr>
<td>Conglomerate</td>
<td>6</td>
<td>2802</td>
<td>85</td>
<td>2752</td>
<td>0.00012</td>
</tr>
<tr>
<td>Sandstone-mudstone</td>
<td>19</td>
<td>2805</td>
<td>67</td>
<td>2776</td>
<td>0.00092</td>
</tr>
<tr>
<td>Andesite</td>
<td>3</td>
<td>2803</td>
<td>33</td>
<td>2810</td>
<td>0.00062</td>
</tr>
<tr>
<td>Sulphide mineralization</td>
<td>7</td>
<td>3342</td>
<td>589</td>
<td>3202</td>
<td>0.00067</td>
</tr>
</tbody>
</table>
conglomerate and sandstone-mudstone indicate the lowest resistivity, as expected. In principle, it has been suggested that membrane polarization occurs in sediments containing pores which have different surface areas, and as a result the ions have different mobility for different rocks (Zadorozhnaya and Hauger., 2007). Like resistivity, it is not easy to choose a representative IP value from the measurements, not only because each measurement includes a range of values, but also because of the different resistance which was applied for each measurement (10k, 100k and 1M ohm). Zadorozhnaya and Hauger (2007) explain that amplitude of potential difference is a product of different factors namely ion mobility, diffusion coefficient, specific conductivity of solution, as well as electrical current flowing in pores, volume of pores and difference of transfer numbers. According to the IP results in Table 1, the highest IP- values are found in massive sulphide, this is followed by the high (k) basalt and sedimentary rocks. Andesite and low (k) basalt show low values of IP (25.28 and 22 mrad, respectively). Felsic volcanic rocks demonstrate lowest IP at 12.37 mrad (Table 1). Therefore, felsic volcanics can be recognized with relatively low resistivity and low IP, among the rest of studied samples.

### 3.2.2. Physical properties of the rocks

Fig. 4 is a plot of density vs. magnetic susceptibility for high (k) basalts, volcanic rocks and sedimentary rocks, respectively.

Fig. 4a indicates a poor correlation between density and magnetic susceptibility. However, all samples indicate high magnetic susceptibilities, as expected. Felsic volcanic rocks, demonstrate a wide range of distribution in the density-magnetic susceptibility plot, which do not correlate well (Fig. 4b). The density of felsic volcanic rocks is highly disseminated, which can be a reflection of their dissimilar mineral composition. There is not a coherent correlation between density

![Figure 4](image_url)

*Figure 4. Density vs. Magnetic susceptibility (a) High (k) basalt (b) Volcanic rocks (c) Sedimentary rocks.*
and magnetic susceptibility of low (K) basalt; however, they demonstrate a narrow range of magnetic susceptibility while their density is widely varied. In terms of andesite, too few samples make it difficult to find a correlation between density and magnetic susceptibility. The density-Magnetic susceptibility plot of the sedimentary rocks shows a good correlation between density and magnetic susceptibility of the sandstone-mudstone (Fig. 4c). The densities are increasing with increasing magnetic susceptibilities. Several sandstone-mudstones indicate high densities (≥2850 kg/m$^3$) which can be reflection of different mineral compositions. The conglomerates in general show limited range of magnetic susceptibility while their density is widely disseminated.

4. Interpretation of electrical and potential field data and results

4.1. Inverse and forward modelling of the electrical data

A preliminary model of the geological structures in the central Skellefte district is proposed based on DC resistivity and IP data collected along the two profiles. The field data was inverted using the RES2DINV (Loke, 2010), program and a model was created of the subsurface down to a depth of 430 m. The reliability of the field data and the interpretation is then controlled using RES3DMOD (Loke, 2010), by creating a forward resistivity model on the basis of the resistivity data obtained from laboratory measurements. After running the forward model, the inverted field data are compared to the forward model. The results were then constrained with the magnetic data and petrophysical data, running the forward and inversion models using Model Vision Pro™ (Encom Technology). The same program as for magnetic modelling is then used to model the gravity along profiles (I) and (II), targeting deeper parts of the crust as compared to the resistivity/IP and magnetic models. Such interpretation is meant to constrain the model using all available data in order to construct the 3D solid model of the geology down to 1500 m depth.

4.1.1. Inverse Modelling of the Resistivity/IP data collected along the two profiles

Profile (I)

Fig. 5 shows the resistivity/IP result after inversion for profile (I). To the SE, within the first 3400m of profile (I), there is a contrast between a high resistive rock in the deeper parts and a low resistive one in the top 260m of model (I and II; Fig. 5d). The low IP zone (I; Fig. 5d) indicates that the shallow layer can be an overburden of soil-cover overlying a high resistive rock with low IP which according to petrophysical data can be an indication of basalt (Table I; Fig. 5c and II in Fig. 5d).

The resistivity increases significantly in the shallow part of the section between 3400-3600m (Fig. 5b at 3600m) which indicates that an underlying high resistive material comes to the surface at this part. This correlates well with the surface geology where the basalt splits the felsic volcanics at the two sides (Fig. 5a). 200m beneath this high resistive basalt, there is a low resistivity zone which indicates extremely high IP (Fig. 5c and III.a; Fig. 5d). This area is close to the Norrliden N VHMS-deposit. A similar high IP value can be observed at ~4600 and 5800m along the profile which is associated with low resistivities (III.b and III.c; Fig. 5d.). These 3 conductive zones (III. a, b and c; Fig. 5d) demonstrate similar signature as the one for sulphide mineralization, which can be an indication of sulphide deposits or graphitic schist/mudstone.

The -sedimentary contact can be recognized at 5500m where the steeply dipping contact with a low resistivity separates the basalt to the SW from sandstone-mudstone to the NE (V and VI; Fig. 5d.). Bauer et al. (2011) explains this contact as a NE-dipping inverted normal fault. The sandstone continuous to 6200m on the surface where it is followed by conglomerate to the end of profile (I) (Fig. 5a and VI and VII ;Fig. 4d). The inter-sedimentary boundary (VI-VII; Fig. 5d) is well-reflected on the resistivity/IP section, where the more conductive sandstone-mudstone shares the boundary with a less conductive conglomerate (Fig. 5b and c).
**Profile (II)**

The resistivity/IP result for profile (II) is shown in Fig. 6. In general, the result from the DC resistivity and IP are consistent in profile (II) (Fig. 6b and c). To the south, the unspecified metasedimentary rocks are interpreted as a low resistive rock which shares the boundary with sandstones with higher resistivity (I and II; Fig. 6d). The transition zone between metasedimentary rocks of the Vargfors basin and felsic volcanic rocks of the Skellefte Group (II-V; Fig. 6d) is well defined in the 2D resistivity model (Fig. 6b). At 1000 m along the profile, an approximately 45° SW dipping contact, is interpreted as the contact between unspecified metasedimentary rocks and felsic volcancics to the NE, a contact which according to Bauer et al. (2011) is an inverted normal fault (II-V; Fig. 6d). At 1300 m, a layer with average resistivity and low IP (III; Fig. 6d) is inferred to origin from graphite with pyrothite minerals. However, the low IP of this layer indicates much lower value compared to the IP values listed in Table 1. At 1600 m there is a zone with low resistivity of ~ 2000 Ωm (Fig. 6c and IV; Fig. 6d). This zone underlies a shallow layer of rhyolite, which here covers a large part of the area (Fig. 6a). However, the dimension and characteristics of this zone can be better understood using potential field modelling. Between 1900-3900m from SW
along the profile, there is a succession of two common volcanic rocks within the central Skellefte district, namely felsic volcanic rocks and basalts (V-VI-V-VII; Fig. 6d), of which basalt has a significantly higher resistivity than that of felsic volcanic rocks (Fig. 6b; Table. 1). To the north, the basalt has its northern boundary in contact with granodioritic rocks of the Jörn intrusive complex (VII-IX; Fig. 6d) (Bauer et al., under review). The maximum depth to the Jörn intrusion is estimated to be 4-6 km in its central parts (Wilson et al., 1987). The geometry of the contact between the supra-crustal rocks of the Skellefte Group and the Jörn intrusive rocks is not clear on the resistivity section (Fig. 6b), probably due to the small differences in resistivity between these two rock types. However, the higher resistivity of the basalt compared to the granite, and low IP value between 3600-5600m imply that the southern contact of the Jörn intrusive complex dips towards NE. The different resistivity/IP values at the end of the profile (VIII and IX; Fig. 6d) can be an indication of an underlying layer of tonalite or basalt beneath the Jörn granodiorite.

Figure 6. (a) Surface geology and location of the resistivity/IP profile (II). (b) Resistivity depth section after inversion. (c) IP depth section after inversion. (d) Geological model based on resistivity/IP results.
4.1.2 Forward and inverse modelling of resistivity data constrained by the measured resistivities of rock samples.

Apart from interpretation of the resistivity/IP profiles acquired after the inversion, testing different geological scenarios in 3D helps to better constrain the interpretation of field data. Thus, by constraining the interpretation with the resistivities from the petrophysical analysis, forward models of the two resistivity profiles are created. The initial forward resistivity models (Figs. 7a and 8a), were divided into rectangular cells where each cell was assigned a resistivity value. The forward models were then inverted to find the optimum model consistent with the previous interpretations. In order to understand the variation of the resistivity at the bottom parts of the model, the forward models extended slightly beyond the depth of the resistivity sections down to 480m; however, without violating a good fit to the field data. The physical properties which are used during the forward modelling are not the exact data from the Table 1; instead they are result of several model tries and improving the models. Moreover, the resistivities in the forward models are in general lower than the resistivities in Table 1, due to different conductivity of the fluid in the field and in laboratory. Nevertheless, they are all within the range of standard deviation of the laboratory data.

![Forward resistivity model for profile (I)](image)

**Figure 7.** Forward resistivity model for profile (I) (a) Resistivity sections at six different depths. (b) Apparent resistivity values calculated from the forward model. (c) Theoretical apparent resistivity. (d) Inverted resistivity pseudo-section.

15
Figure 7a shows the resistivity values used to create the forward model of profile (I). Based on surface geology and the interpretation of the resistivity depth sections, each layer has been assigned a resistivity estimated on basis of laboratory measurements. The resistivities range between 500 to 25000 Ωm at six different depth levels with various thicknesses. The thickness of the layers increases with depth due to the decreasing resolution of the deeper parts of model.

Figure 7b indicates the measured apparent resistivity pseudo-section based on the model in Fig. 7a. The calculated apparent resistivity of profile (I) is shown in Fig. 7c, which indicates the pseudo-section after removing the noise from the data. The forward model was then inverted using pole-dipole array to compare with the field data (Fig. 7d). A comparison between the model based on inversion of field data (Fig. 5b) and the one based on the forward model (Fig. 7d) indicates that the two models, in general, are consistent. This indicates that the overall interpretation of the resistivity/IP field data is correct.

The forward and inverted resistivity model of profile (II) was created in a same way as profile (I). Fig. 8a shows the resistivity forward
model. The conductive sedimentary rocks in the Vargfors basin extend approximately 800m along the profile and narrow with depth (Fig. 8a; 1000 Ωm). The dominant sedimentary rocks in this part of profile (II) (unspecified sediments, II; Fig. 6d), are assigned a resistivity of ~ 1000 Ωm. Also, a layer of low resistivity (inferred as graphitic material, IV in Fig. 6d) is located between 1400-1600 m along the profile (II). The basalt (VII in Fig. 6d) is given a slightly higher resistivity (20000 Ωm) than the adjacent Jörn granitoid (18000 Ωm; VII; Fig. 6d). The resistivity of the felsic volcanic rocks (V; Fig. 6d) is assigned to be 8000 Ωm and 2500 Ωm, reflecting rhyolitic to dacitic rocks, respectively (Fig. 8a). Fig. 8b and c shows the measured apparent resistivity and calculated apparent resistivity pseudo-sections, respectively. The inverted model based on forward models is shown in Fig. 8d.

A comparison between the model based on inversion of field data (Fig. 6b) and the one based on the forward model (Fig. 8d) indicates an acceptable similarity between the two models.

4.2. Integration of resistivity and potential field data

The interpretation of the resistivity depth sections is controlled by forward and inversion models of the resistivity profiles. Resistivity was determined on rock samples as well as from inversion of the IP profiles. However, there are still a number of uncertainties within the models which can be investigated by using additional geophysical data. Hence, we use the magnetic data to constrain and modify the models for shallow depths (~ 500m). Furthermore, gravity data was modelled along the same profiles to test and extend the models down to 1500m. As starting models for the magnetic test, the resistivity/IP models were used (Figs 5b, 5c, 6b and 6c). The magnetic susceptibilities were determined by petrophysical measurement in the laboratory (Table. 1). Drill-hole data with complete lithological description were also used to constrain the gravity and magnetic models (Fig. 9).

Separation of the regional field from the field data is an important part of gravity and magnetic interpretation. There are various methods to estimate the regional fields, among those; polynomial surface, minimum curvature and finite element are the most common methods (Xu et al., 2009). In this study, the regional gravity and magnetic fields are defined using a polynomial surface generator function (Model Vision Pro™, Encom Technology). Accordingly, the regional field is defined as a polynomial surface of specified order, which is achieved using all data within the CSD. Choice of polynomial order is directly controlled by the complexity of the field. However, a polynomial order higher than four can lead to an incorrect regional field, where the influence of sharp local variations interferes with the calculation of the regional field. We have used a 2nd order polynomial surface for modeling of the regional field for the two profiles; and then manually modified the field to make a better model fit without changing the density and magnetic susceptibilities considerably. The area used for calculating the regional field is defined beyond the two profiles in order to incorporate the effect of deep and large scale features. The definition of area and choice of order of the polynomial was determined after testing different alternatives. During the modeling, the remnant magnetization is neglected and a background density of 2670 kg/m³, which is common average density for the crustal rocks, is considered.

According to the surface geology, felsic volcanic rocks are dominant in the surface of the southern 3300m of profile (I) (Fig. 10a). Although the resistivity and IP lows in this part (I in Fig. 5d) correlate well with similar resistivity and IP properties for felsic volcanic rocks, it is possible that this layer is a soil cover, which overlies a thin layer of rhyolite which shares the boundary with the basalt. The magnetic high within the southermost 3300m of profile (I) (Fig. 10d) and low magnetic susceptibility of felsic volcanics (~ 0.00033; Table. 1), suggests an underlying layer with higher magnetic susceptibilities which is consistent with the high magnetic susceptibility of basalt which underlies felsic volcanic rocks (II; Fig. 5d).
Figure 9. Location and geological description of the studied drill-holes close to the profiles. Only those drill-holes with a complete geological description and length >50 m were used.
The gravity model of this part correlates well with the suggested magnetic model since this part of the profile is associated with gravity high (Fig. 10e), which again is consistent with high density of basalts (Table 1). At 3300m, where basalts crop out at the surface, the magnetic anomaly tends to increase considerably (Fig. 10d), which marks the contact between felsic volcanic rocks and basalts (II-III; Fig. 5d). The drill-hole data (Fig. 9) and surface geology are also compatible with the suggested magnetic model. The gravity model of this magnetic basalt indicates an increase on the top of this basalt (~3300m; Fig. 10d), which supports the suggested RIM model (Fig. 10e). Between 3500-4400m, the surface geology shows felsic volcanic rock (Fig. 10a and IV; Fig. 5d); however, drill-hole data indicate that at 3600m, there is a sequence of felsic and mafic volcanics associated with the sulphide mineralization at ~200m depth (III. a; Fig. 5d and Fig. 9). The basalt in this part is supposed to have lower degree of magnetization compared to the one which surfaces at 3300m (II in Fig. 5d). The presence of magnetic basalt at 3300m next to the low magnetic basalt and massive sulphide with lower magnetic susceptibility at 3600m along the profile (I), is the main reason for the depression in the magnetic anomaly curve at 3600m (Fig. 10d). A comparison between the high IP (III. a; Fig. 5d), drill-hole NRL67103 (Fig. 9) and IP high for sulphide mineralization from laboratory data indicates that III.a in Figure. 5d is the extension of the Norrleden N sulphide mineralization. This model correlates well with the both the magnetic and the density models (Fig. 10d and e). The low resistivity zone at this part of profile (I) (IV; Fig. 5d) and magnetic and gravity low (Fig. 10d and e) are indication of felsic volcanic rocks in this part of profile (I) which continues at depth beyond 1500m. The resistivity and IP sections for profile (I) demonstrate two other highly conductive zones at 4200m and 6000m (Fig. 5c, III. b and III. c; Fig. 5d) which are already interpreted as sulphide mineralization, as they expose similar resistivity/ IP behaviour as the mineralization in the vicinity of the Norrleden N deposit (III. a; Fig. 5d). This fits well in the magnetic model of this profile at ~4200m (III. b; Fig. 5d) where there is an increase in magnetic anomaly as sulphide mineralization with higher magnetic susceptibility compared to felsic volcanics (Table. 1) comes close to the surface (III. b and IV; Fig. 5d). This is not, however, reflected well in the gravity anomaly which can be due to the small dimension of the massive sulphide (Fig. 10e). Between 4900-5600m, the dominant lithology at the surface is basalt (V; Fig. 5d), which cannot be seen in the resistivity section (Fig. 10b). However, the magnetic model suggests a thin layer of basalt which gets wider at depth eastwards (Fig. 10d and V; Fig. 5d) which is consistent with the gravity model (Fig. 10e). At 5600m, within the transition zone between basalt and sediments (V-VI; Fig. 5d), the magnetic anomaly starts to decrease (Fig. 10d), which is consistent with the high magnetic susceptibility of the basalt and low magnetic susceptibility of the adjacent sediments (Table. 1). At the resistivity section of profile (I), the conductive sandstone (VI; Fig. 5d) can be distinguished from the less conductive conglomerate (VII; Fig. 5d). Also, the IP section indicates lower IP for conglomerate to the east compared to the sandstone to the west (Fig. 10c, VI and VII; Fig. 5d). These observations are consistent with the depression in the magnetic field anomaly caused by the conglomerate (Fig. 10d). The gravity high on top of the conglomerate (VII; Fig. 5d) is assumed to be caused by the deep basalt underlying the felsic volcanics which themselves underlie the Vargfors basin. The resistivity, IP and magnetic model all indicate that the bottom of the Vargfors basin is deeper than 430m, and the gravity model suggests ~700m for the bottom most part of basin in profile (I). In general, the fit to the magnetic model for profile (I) (RMS= 3.3 %) indicates a good correlation between the set of data which are used to constrain the model. Yet, there is a slightly larger misfit for the gravity model of profile (I) (RMS= 6.2%) compared to the magnetic data, which can be due to the less control in the deeper parts.

Profile (II) starts within sedimentary rocks of the Vargfors basin (I and II; Fig. 6d). According to the surface geology, the profile begins in the sandstone-mudstone and after ~150m comes
Figure 10. (a) The geology around profile (I). (b) The model resistivity section after inversion. (c) The model IP section after inversion. (d) The model magnetic section after inversion and applying constrains to the model (RIM model). (e) The model gravity section after inversion and applying constrains to the model (RIMD model). (Black Line: Observed anomaly, Red line: Calculated anomaly, Blue line: Regional anomaly).
into sedimentary rocks of unknown origin (I-II; Fig. 6d). The resistivity section of this part indicates resistivity high from sandstone and resistivity low from unspecified sediments (Fig. 11b). A low resistivity of the unspecified sediments (II; Fig. 6d and Fig. 6b) can be an indication of argillitic to graphitic composition of these rocks or existing conglomerate which appears on the surface before start of profile (II), as they show close resistivity to that of conglomerate in profile (I). The IP section (Fig. 11c) can also specify the same geometry for this inter-sedimentary contact, where a ~ 200 m deep layer covers ~ 100m beginning of the profile (II) (I-II; Fig. 6d and Fig. 6c). The geological interpretation indicates that this part of volcano sedimentary contact is a result of SW dipping fault (Bauer et al., 2011). The magnetic model of these sediments (Fig. 11d) fits well with the low magnetic susceptibility of sediments (Table. 1). At 1000 m, the Skellefte-Vargfors contact which separates unspecified sedimentary rocks from felsic volcanic rocks (II-V; Fig. 6d) correlates with the magnetic anomaly with a local increase at the contact (Fig. 11d). Moreover, this sedimentary-volcanic contact is reflected well on the gravity anomaly, where the gravity decreases after sandstone towards NE (II-V; Fig. 6d and 1300m; Fig. 11e) due to presence of rocks with a low density (III and IV; Fig. 6d). Between 1000-2000 m, the felsic volcanic rocks are again dominant (V; Fig. 6d) which is normally represented by low resistivity and magnetic susceptibility in the major parts of two profiles (V; Fig. 6d and Figs. 11c and 11d). However, between 1100-1400m, there is an increase in the magnetic anomaly (III; Fig. 6d and Fig. 11d) which cannot be explained by the felsic volcanic rocks. The resistivity/IP sections previously suggested graphite with concentration of pyrothite. The intensive gravity low produced by this rock, suggests a rock type with low resistivity, average IP, high magnetic susceptibility and low density which can be graphite. However, after modifying the shape of this body, the fit to the gravity data is not satisfactory while there is a fairly good fit to magnetic data, which is more sensitive to the layers closer to the surface compared to gravity. Next to this layer, there is a thin layer of sandstone (IV; Fig. 6d) which is extended down to ~ 200m and is compatible with the magnetic low at 1800m (Fig. 11d). The small dimensions of sandstone make it difficult to see any variation on the gravity anomaly (IV; Fig. 6d and Fig. 11c). At 2000m, lithology changes from felsic volcanic to basalt (V-VI; Fig. 6d), continuing to ~ 2700m along the profile (II). This layer of basalt (VI; Fig. 6d), however, is a shallow layer (~ 100m at depth) with low (K) which gives a small increase to the magnetic anomaly. The profile again enters felsic volcanic rocks (VI-V; Fig. 6d) between ~2700-3000m, which reduces the magnetic anomaly (Fig. 11d). The felsic volcanic layer is indicated with a resistivity low (V; Fig. 6d and Fig. 6b). At 3000m, the IP section shows a high anomaly which is an indication of changing lithology to the basalt (V-VII; Fig. 6d). The gravity model, however, is not influenced intensely in this part due to the small depth extension of these rocks (Fig. 11e). At ~3000m, felsic volcanic rock is interrupted by a thin layer of sandstone which is an intermediate between felsic volcanics and basalt (Between V and VII; Fig. 6d). The resistivity/IP sections indicate a ~ 300m depth extent for the basalt (VII; Fig. 6d) which fits well with the observed magnetic high between ~3000-4400m (Fig. 11d). Since the Skellefte-Jörn contact appears on the surface geology at 3800m, this indicates that the contact at the depth is a transition between rocks with higher magnetic susceptibility compared to susceptibility of Jörn granite (VII; Fig. 6d). The resistivity, IP and magnetic profiles indicate the existence of basalt even after 3800m along profile (II) (VII; Fig. 6d), reflected by high resistivity and low IP, especially within the 300m of subsurface. The volcanic-Jörn contact, represented by a series of north-dipping break-back faults (Bauer et al., 2011) is associated with a layer with magnetic susceptibility higher than Jörn granitoids but lower compared to the basalt near the contact (IX; Fig. 6d). This layer also indicates a higher density compared to the Jörn granitoids. Accordingly, we interpret this layer as a tonalite which cuts both granitoid and basalt (IX; Fig. 6d) and increases gently the magnetic anomaly but gives rise to gravity anomaly considerably (Fig. 11d and 11c).
Figure 11. (a) The geology around profile (II). (b) The model resistivity section after inversion. (c) The model IP section after inversion. (d) The model magnetic section after inversion and applying constrains to the model (RIM model). (e) The model gravity section after inversion and applying constrains to the model (RIMD model). (Black Line: Observed anomaly, Red line: Calculated anomaly, Blue line: Regional anomaly).
The gravity model also indicates that the tonalite is extended to great depth near the Skellefte-Jörn contact, which causes the gravity high on top of Jörn granitoids which have lower density (Fig. 11e). Hence the Jörn granitoid (VIII; Fig. 6d) is assumed to start at 3800m and continues towards the end of profile (II) and is associated with the shallow layer of basalt and deep tonalitic rocks. According to the gravity model of profile (II), the tonalite have a density of 2904 kg/m$^3$ which is close to density of the basalts (Table. 1). In general, the fit to the magnetic model for profile (II) is better compared to the magnetic fit for profile (I) (RMS= 2.6% while the gravity model shows a poorer fit for profile (II) (RMS= 8.5%).

4.3. Integration of the results to construct the 3D solid model

The 3D geology model based on the interpretation of the geophysical data in parts of the central Skellefte district around profiles (I) and (II) is shown in Fig. 12.

**Felsic volcanic rocks (Skellefte Group)**

The felsic volcanic rocks of the Skellefte Group are the most common rocks along the two studied profiles in the CSD. The locally faulted contact between the Vargfors basin and the underlying Skellefte Group felsic volcanic rocks is shown in Fig. 12. The felsic volcanics are regularly interrupted by sequences of basalts belonging to the Skellefte Group, which in general have thin depth extent. However, due to significant difference between density and magnetic susceptibility of the felsic rocks (2735 and 0.00032; SI) and basalts (2894 and 0.0012; SI), they can be well recognized using gravity and magnetic data. On the electrical profiles felsic volcanic rocks can be simply recognized with their low IP. The felsic volcanic rocks apparently extend to depths beyond 1500m.

**Basalt (Skellefte Group)**

The Skellefte Group basalts indicate different degree of magnetization, while they have a similar high range of densities (2894 and 0.0012; SI). Along profile (II), they occur close to the contact with the Jörn granodiorite, which increases the magnetic field intensity. The occurrence of basalts within felsic volcanic rocks of the Skellefte Group at the Skellefte-Jörn contact is also indicated by the resistivity high and low IP along the resistivity sections.

**Granitoid (Jörn intrusion)**

The depth of the bottom contact of the Jörn intrusive complex exceeds 1500m, which is consistent with previous interpretations (4-6 km by Wilson et al., 1987). Below the Jörn granodiorites, we suggest a rock with a higher density and magnetic susceptibility compared to granodiorite, but indicating a lower magnetic signature compared to the basalt. This is consistent with the common occurrence of tonalite (IX in Fig. 6d) within the Jörn intrusive complex as described by Wilson et al. (1987). Therefore, parts of Jörn intrusion are associated with tonalitic rocks, which increased significantly the magnetic and gravity anomaly. Compared to the basalt, tonalite has a lower degree of magnetization but a similar density. The tonalite depth extension is estimated of ~ 1400m, in order to make a good fit to both gravity and magnetic data.

**Sedimentary rocks (Vargfors Group)**

The sedimentary rocks along the two profiles consists of sandstone-mudstone, conglomerate and unspecified sediments. Owing to different electrical properties of these sedimentary rocks, the inter-sedimentary contact can be modelled with appropriate accuracy (VI-VII; Fig. 5d and I-II; Fig. 6d). However, the magnetic and gravity data did not contribute much in dividing different sedimentary rock types, since their density and magnetic susceptibility is similar. The maximum depth of sedimentary rocks in the Vargfors basin along two profiles is estimated to be ~ 700m.

**Ore deposit (Sulphide mineralization)**

On basis of interpretation of electrical data, three potential sulphide mineralizations were suggested along profile (I) (III. a, b, c; Fig. 5d). All three mineralizations are located at less
than 500 m depth. Furthermore, one of these mineralizations is located at the contact between Skellefte Group and the Vargfors basin (III.c; Fig. 5d), which according to previous studies has high possibility of hosting VMS deposits (Weihed, 2010). The IP measured on drill-core samples shows a good consistency with the IP acquired in the field. Although potential field data did not disagree with the suggested depth and dimensions for the sulphide mineralizations on basis of resistivity/IP data, they also did not create any gravity or magnetic anomaly because of their small dimension and great depth extent.

![Diagram of geological profile](image)

**Figure 12.** 3D model of the central Skellefte district based on the successive interpretation of geophysical and geological data.

### 5. Conclusion

A successive procedure of interpretation is carried out in a part of the central Skellefte district to understand the geometry and contact relationship of the key geological structures with the purpose of creating a 3D solid model of the geology for potential exploration activities in the future.

The result from 2D resistivity/IP field survey after the inversion provides a promising model in which the contrast between resistivity values of the different layers enabled us to understand the geometry of the subsurface with high resolution down to the 430 m. The IP result on the other hand, associated to the interpretations especially where the small conductive rocks could have been missed in the resistivity models. This was especially of great importance to map the high conductive zones, including the Norrliden N deposit and two other potential mineralizations along profile (I). Those can therefore be potential candidate locations for
drilling test in the future, especially as they are all located at shallow depths (≤500m).

Contribution of the petrophysical data to the modelling improved the models to a large extent. Having the true density, resistivity, magnetic susceptibility and IP from different rocks provided a better start model during the modelling. Testing different geological scenarios in 3D (forward resistivity modelling) constrained with the resistivity data from laboratory measurement and inverting the models, helped to verify the previous results based on the interpretation of the electrical field data.

Potential field modelling improved the previous models obtained from the resistivity and IP data, especially where the resistivity of the thin layers were masked, due to the large electrode spacing. The gravity interpretation improved the models based on the resistivity, IP and magnetic data (RIM model) by providing information about the deeper part of the crust (∼1.5 km). Although the density of the metasedimentary rocks and felsic volcanic rocks do not differ significantly, the resulting density model is well constrained by the RIM model. The interactive forward modelling of density and magnetic data thus improved the preliminary models based on electrical data.

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References


Regional scale 3D joint modeling of the gravity and magnetic data in the central Skellefte district; a model based on interpretation of reflection-seismic data

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Regional scale 3D joint modeling of the gravity and magnetic data in the central Skellefte district; a model based on interpretation of reflection-seismic data

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Abstract

The Skellefte district in northern Sweden is the most important host to volcanic-hosted massive sulfide (VMS) deposits in Sweden. Due to its economical importance, a series of geological and geophysical techniques were applied to create an image of the shallow crustal architecture and its relationship to mineral deposits down to ~ 4.5 km depth. Consequently, reflection-seismic data along three nearly parallel profiles were acquired during 2009-2010 to map the spatial relationship of geological structures at depth down to ~ 4.5 km depth. Although these seismic studies helped to understand the regional relationship between geologic units in the central Skellefte district, the reflection-seismic data did not succeed in mapping lithological contacts in the area.

In this study, we examined the model suggested on the basis of reflection-seismic data by using potential field data (down to 5 km depth) as well as constrains from electrical data (down to ~ 0.5 km depth) and surface geology. Moreover, we modeled gravity and magnetic data along the non-reflective or poorly reflective parts of the seismic profiles to identify major lithological contacts and shear zones in the central Skellefte district, which could not be modeled on basis of the reflection-seismic data. Gravity and magnetic data aid to reveal the spatial relationship between the Skellefte volcanic rocks, Vargfors Group meta-sedimentary rocks and two meta-intrusive complexes. Moreover, the result of the study suggests a ~ 4km depth extent for the tectonic contact at the southern border of the Jörn intrusive complex. To the north, both gravity and magnetic data are consistent with the previous suggested depth extent for the Gallejaur complex. Further towards NE, the north-dipping Skellefte-Jörn contact is well related with magnetic lows and gravity highs which implies that Jörn intrusive rocks have a greater thickness compared to the underlying basalt. Furthermore, this study proves previous concepts of fault geometries and fault patterns as a result of upper-crustal extension and subsequent inversion.

Keywords: 3D –modeling; potential field data; reflection-seismic data; Skellefte district, Inverse modeling; basin inversion.
1. Introduction

The Skellefte district is one of the major mining districts in Sweden and therefore it has been subject to many geological and geophysical studies during the last decades. The Skellefte district hosts ca. 80 volcanic-hosted massive sulfide (VMS) deposits, four of which are currently mined (Kathol and Weihed, 2005). Due to its economic importance, it is essential to understand the geometry of the geological structures (Weihed, 2010). Earlier studies and exploration activities concentrated mainly on shallower depths (~ 300 m) and the interest towards explorations at greater depths has increased (c.f. Dehghannejad et al., 2010). Understanding the Skellefte district in 3 and 4 (Time domain) dimensions is essential for future exploration of base metals (c.f. Bauer et al., 2010; Weihed, 2010). Thus, understanding the subsurface relationship of geologic contacts within the upper 5 km of the crust is of great importance for understanding of the geometry and spatial occurrence of ore deposits.

Previous geological and geophysical studies conducted in the central Skellefte district were in a more local scale, mainly for targeting different ore deposits (i.e. Tavakoli et al., under review; Weihed, 2010). Earlier geophysical methods in the central Skellefte district include electromagnetic (EM), gravity, magnetic, gamma ray spectrometry, resistivity, Induced Polarization (IP) and reflection-seismic studies, which were conducted in different scales mainly by Boliden Mineral AB and SGU (Kathol and Weihed., 2005) as well as Geovista AB.
Application of geoelectrical techniques in conjunction with potential field data has previously proved efficient in defining contacts between felsic volcanic-intrusive and felsic volcanic-sedimentary rocks, respectively (Tavakoli et al., under review). Dehghannejad et al. (under review) applied reflection-seismic techniques in order to define the structural framework down to a depth of ~ 4.5 km and correlated the observations to the surface geology in the Skellefte district. The results show a series of south and north-dipping reflectors, which were interpreted as major shear zones. However, in certain areas there was a lack of or only weak seismic reflectors, which did not result in any structural model. Hence, gravity and magnetic data has been used in these areas for defining lithological contacts, which has proven successful in other studies (Kadima et al., 2011; Tanfous et al., 2010). The aims of this study were to: (i) Test the gravity and magnetic response of the proposed seismic models and regional-scale geological observations (Dehghannejad et al., under review) (ii) study the spatial relationship between key lithologies at depth (iii) construct a regional 3D solid model of the central Skellefte district down to a depth of ~ 5 km based on the constrained 3D model obtained from potential field, reflection-seismic, and resistivity/IP data.

2. Geological setting and reflection-seismic data

2.1. Geological settings

Rocks in the Skellefte mining district consist of Palaeoproterozoic supracrustal and intrusive rocks formed in a volcanic arc setting (Allen et al., 1996) and metamorphosed during the Svecof Kellerian Orogeny (Kathol and Weihed, 2005). The dominating unit is the 1.90 – 1.88 Ga, ore-bearing Skellefte Group, which comprises mainly felsic meta-volcaniclastic and meta-volcanic rocks (Fig. 2; Allen et al., 1996; Kathol and Weihed, 2005). The overlying 1.88 – 1.86 Ga Vargfors Group mainly consists of meta-sedimentary rocks. The lower parts of Vargfors stratigraphy are dominated by mudstones, sandstones and monomict conglomerates formed from turbiditic currents, whereas the stratigraphic higher parts comprise polymict conglomerates formed in an alluvial fan environment (Bauer et al., under review). The Vargfors Group meta-sedimentary rocks in the central Skellefte district define the distinct Vargfors basin (Fig. 2). The contact relationship between Skellefte Group meta-volcanic rocks and Vargfors Group meta-sedimentary rocks range from conformable to unconformable to faulted (Allen et al., 1996; Bauer et al., 2011). The massive sulfide mineralizations formed as sub-seafloor replacement in the uppermost parts of Skellefte Group stratigraphy (Allen et al., 1996). Intrusive rocks in the central Skellefte district are dominated by meta-tonalites to meta-granites of the Jörn intrusive complex, which consist of several intrusive phases (GI-GIV; Wilson et al., 1987; Gonzáles Roldán, 2010). The faulted contact between Jörn meta-tonalites and Vargfors meta-sedimentary rocks is defined by a rim of mafic meta-volcanic and meta-volcaniclastic rocks (Bauer et al., under review).

Structural analysis (Bauer et al., 2011) and reflection-seismic investigations (Dehghannejad et al., under review) revealed a series of WNW-ESE-striking, south-dipping inverted normal faults that were formed during Palaeoproterozoic upper crustal extension and inverted during subsequent crustal shortening (Fig. 2). These faults are cross-cut by north-dipping, late compressional break-back faults (Bauer et al., 2011, under review). A set of NE-SW-striking syn-extensional transfer faults segments the central district into distinct fault blocks. Locally, the faults are accompanied by mafic volcanic activity. Strain was partitioned during approximately N-S crustal shortening (c.f. Bergman Weihed, 2001) resulting in low-strain domains with mainly open, asymmetric synclines and minor anticlines, and high-strain domains in the vicinity of faults with tight to isoclinal folds and local overturned strata (Bauer et al., 2011).

The youngest major phase of intrusions comprise 1.82 – 1.78 Ga late- to post-tectonic Revsund-type intrusive rocks, which are part of
Figure 2. Geological map of the CSD with the location of seismic profiles and CDP-lines of Dehghannejad et al. (under review). Modified after Bauer (2010).
the Transscandinavian Igneous Belt (TIB). The youngest major deformation event at 1.80 Ga was partitioned into the major NE-SW-trending faults due to E-W-crustal shortening (Bergman Weihed, 2001; Weihed et al., 2002; Skyttä et al., 2010; Bauer et al., 2011).

2.2 Reflection-seismic data

Reflection-seismic data were acquired along three nearly parallel N-S trending profiles (C₁, C₂ and C₃) in the central Skellefte district (Dehghannejad et al., under review) with the aim of achieving a base for a better understanding of the spatial relationship between lithologies (Fig. 2).

Profile C₁

The crust below the southern parts of profile C₁ is more reflective than in the northern parts and distinct reflectors, R₁, R₂ and R₃ (Fig. 4d), have been identified (Dehghannejad et al., under review). Dehghannejad et al. (under review) suggested that R₁ results from a high-strain zone separating meta-sedimentary rocks in south from meta-volcanic rocks in the north. The reflectors R₂ and R₃ were suggested to be the result of parallel high-strain zones. Other reflectors were also identified where R₄ was suggested to originate from a high-strain zone intersecting the surface in the vicinity of the southern border of the Vargfors basin. The reflectors R₅ and R₆ could not be correlated with the geology, since they do not reach the surface. The reflector R₇ was interpreted as the north-dipping contact between the Jörn intrusive complex in the north and volcanic rocks of the Skellefte Group to the south.

Profile C₂

The rather weak reflection R₄ at CDP 830 along profile C₂ (Fig. 5d) was interpreted to be related to R₄ in profile C₁ (Dehghannejad et al., under review). Reflector R₅ was interpreted as a north-dipping basin-bounding fault at the southern contact of the Vargfors basin, which together with the reflector R₁₅ describes a synformal shape of the Vargfors basin (Dehghannejad et al., under review). Neither the north-dipping reflector R₈ nor the south-dipping reflector R₉ appear in the neighboring profile C₂, therefore making it difficult to correlate them. Dehghannejad et al. (under review) suggest the reflectors R₈ and R₉ as their equivalent reflectors along the profile C₁; however, R₈ is a weaker reflector in profile C₂ compared to profile C₁. The sub-horizontal reflector R₅ in profile C₂ has approximately the same dip as R₅ in profile C₁, suggesting it represents the same contact.

Profile C₃

The origin of the three south-dipping reflections R₁, R₂ and R₃ in profile C₃ was interpreted by Dehghannejad et al. (under review) as a part of the same high-strain zone as in profile C₁. Furthermore, Dehghannejad et al. (under review) refers a set of reflections and diffractions between 700-1400 in profile C₃ as a series of fault blocks, separated by sub-vertical normal faults (Dehghannejad et al., under review). The south-dipping reflector R₄₀ is suggested to be related to reflection R₄ in profiles C₁ and C₂, however, this reflector does not reach the surface along the profile line of C₃, as it is cross-cut by reflection R₁₃. Nevertheless, R₄₀ probably comes to the surface at ~ CDP 1400. A series of weak reflectors (R₁₁₄) at ~ 3 km depth could not be correlated with surface geology or similar reflections along the profiles C₁ and C₂, but were suggested as a possible bottom contact of the Gallejaur complex or sub-horizontal structures within it (Dehghannejad et al., under review).

3. Potential field and petrophysical data

Gravity and magnetic response of models based on the seismic reflectors from the three seismic profiles were calculated in order to test a possible correlation between reflection-seismic data and potential field data. Previous studies indicated the success of interpretations based on the integration of reflection-seismic and potential field data (Goleby et al., 2002; Wattanasen et al., 2006; Malehmir et al., 2007). Density and magnetic susceptibility were measured on samples from the central Skellefte district and the results were used to run the starting models. Moreover, the geologic contacts determined from surface geological mapping were used to
constrain the uppermost parts of the models. These data were especially important for constraining the model in areas where there was a lack of reflectors, or only poor quality reflectors existed (Kadima et al., 2011). Since the seismic data were acquired along crooked-line profiles, gravity and magnetic data were modeled on the common depth point (CDP) lines (Fig. 7a).

Gravity and magnetic data were provided by the Geological Survey of Sweden (SGU) and Boliden Mineral AB. The gravity data grid was created with an average station spacing of 500 m and the magnetic data grid was created from airborne magnetic measurements with 40 m station spacing acquired from 30 m altitude. The regional gravity and magnetic field is calculated using a larger grid with the Model Vision Pro™ (Encom Technology) based on 2nd polynomial surface (Figs. 3a and 3b).

3.1 Gravity and magnetic modeling

The gravity and magnetic models along the three profiles were created based on previous interpretations of the reflection-seismic data by Dehghannejad et al. (under review). The Bouger anomaly map covering the three profiles in CSD indicates that all profiles start in the SW in a rather high gravity field, pass through a low gravity part, and end in, or pass, over (profile C3) a gravity high in the NE (Fig. 3a). Similar trends as for the gravity data can be observed for the magnetic data. The profiles start in the SW in granites and sedimentary rocks with low magnetic signatures and end in mafic and felsic intrusive rocks of the Gallejaur complex and Jörn intrusive complex, which are associated with magnetic highs (Fig. 3b). Hence, the forward gravity and magnetic models along the three profiles were built utilizing the measured densities and magnetic susceptibilities (Table 1), adjusted within the range of standard deviation for each rock until an acceptable fit was acquired. The models were better defined where the seismic reflectors reach the surface, thus providing a better constrains for their geometry. In areas where no seismic reflectors could be determined, the gravity and magnetic data in most parts provided substantial information about the shape and depth of different lithologies.

3.2. Petrophysical study of drill-core samples

The petrophysical study of drill-core samples improve the understanding of the variation of lithology at depth and the modeling, since it provides the true properties of different rock types (Silvennoinen et al., 2010; Tavakoli et al., under review). In order to model the gravity

Figure 3. (a) Bouger anomaly map of CSD around the seismic profiles. (b) Magnetic anomaly map of CSD around the seismic profiles.
and magnetic response of the reflection-seismic models, we used the density and magnetic susceptibility of drill-core samples (Table 1). The mean value and standard deviation of the density was calculated for each rock type. Due to the big variations in magnetic susceptibility, we used median magnetic susceptibility as a representative value; however, the standard deviation is also calculated to demonstrate the range of magnetic susceptibility of each rock type (Table 1).

<table>
<thead>
<tr>
<th>Rock type</th>
<th>#</th>
<th>Density ((\text{kg/m}^3))</th>
<th>(k) Magnetic Susceptibility (\times 10^{-5}) SI</th>
</tr>
</thead>
<tbody>
<tr>
<td>Felsic volcanic rock</td>
<td>91</td>
<td>2735</td>
<td>2730</td>
</tr>
<tr>
<td>Basalt (High K)</td>
<td>10</td>
<td>2857</td>
<td>2833</td>
</tr>
<tr>
<td>Basalt (Low K)</td>
<td>18</td>
<td>2894</td>
<td>2893</td>
</tr>
<tr>
<td>Conglomerate</td>
<td>6</td>
<td>2802</td>
<td>2752</td>
</tr>
<tr>
<td>Mudstone-sandstone</td>
<td>19</td>
<td>2805</td>
<td>2776</td>
</tr>
<tr>
<td>Intermediate volcanic</td>
<td>3</td>
<td>2803</td>
<td>2810</td>
</tr>
</tbody>
</table>

Basalts with high and low magnetic susceptibility and high density were identified. However, felsic volcanic and sedimentary rocks (conglomerate and sandstone-mudstone succession) are in the same range of density and show low magnetic susceptibility, while the magnetic susceptibility of the mudstone-sandstone succession is somewhat higher than that of conglomerate. For both, gravity and magnetic modeling, the densities and magnetic susceptibilities were varied within the standard deviation of each rock type to acquire an acceptable fit between observed and calculated data.

### 4. Results and interpretation

Integration of the results to construct a regional 3D model of central Skellefte district.

**Profile C**

Profile C starts in the SW in Vargfors Group mudstones-sandstones and proceeds towards the north into the TIB-type gabbroic-dioritic intrusive rocks (Fig. 2). According to the gravity model of profile C (Fig. 4b), the SW-dipping gabbro-dioritic intrusion (body II; Fig. 4c) is located on the boundary to the Vargfors Group mudstones-sandstones (body I; Fig. 4c). The gravity and magnetic interpretations suggest an approximate depth of \(\sim 3.5\) km for the gabbro-dioritic intrusion. Also, the magnetic anomaly indicates that the gabbro-diorite is more magnetized compared to the adjacent meta-sedimentary rocks, which increases the magnetic anomaly at \(3.1\) km, to the east of the meta-sedimentary rocks (Fig. 4a). The local minima in magnetic and gravity anomalies at \(\sim 4\) km (Figs. 4a and 4b) indicates intercalation of different rocks with similar petrophysical signatures to those of rhyolite into the gabbroic-dioritic intrusion (body III; Fig. 4c).

The model based on potential field data suggests that reflection \(R_1\) cuts through the TIB-gabbro-dioritic intrusion and continues through the contact between the gabbroic-dioritic intrusion with the sedimentary rocks (Fig. 4d). Furthermore, the model shows that reflection \(R_2\) coincides with the contact between the TIB-gabbro and Skellefte Group rhyolite (bodies II and IV; Fig. 4c). Although, reflector \(R_3\) cannot be traced to the surface, it may represent the contact between rhyolites and unspecified felsic volcanic rocks, based on the model suggested by potential field data (bodies IV-V; Fig. 4d). At 8-14 km along C, we used a 1.5 km deep model based on electrical data constrained with potential field and drill-hole data (Tavakoli et al., under review). This model is compatible
Jörn Granitoid (0.00065) Felsic volcanic - unspecified (Skellefte Group) (0.00007)
Gabbro - diorite (Transscandinavian Igneous Belt) (0.00033-0.002)
Basalt (Skellefte Group) (0.0012)
Sulphide mineralizations (0.001)
Rhyolite (Skellefte Group) (0.00015)
Sandstone - mudstone (Vargfors Group) (0.00012)

Jörn Granitoid (2760 kg/m³)
Rhyolite (Skellefte Group) (2670 kg/m³)
Gabbro - diorite (Transscandinavian Igneous Belt) (2910 kg/m³)
Basalt (Skellefte Group) (2850 kg/m³)
Sulphide mineralizations (3000 kg/m³)

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Figure 4. Potential field modeling along profile C1 based on interpretation of reflection-seismic data.
(a) Magnetic model of profile C1. (b) Gravity model of profile C1. (c) The depth section and geological structures at depth. (d) Integration of the seismic reflectors and model based on interpretation of potential field data.
with the magnetic and gravity data. However, the gravity low at 7-12 km (Fig. 4b), with dominantly rhyolite outcropping at the surface may rather result from felsic volcanic rocks with a lower density but a slightly higher magnetic susceptibility (i.e. clastic felsic volcanic rocks) (body V; Fig. 4d). The gravity and magnetic data indicate that the reflector \( R_4 \) may be related to the contact between rhyolite and unspecified felsic volcanic rocks (V-IV; Fig. 4c and \( R_4 \); Fig. 4d). Meta-sedimentary rocks of the Vargfors basin, is not reflected in the gravity or magnetic data, due to the similarity in density and magnetic susceptibility with the surrounding rhyolitic rocks (body IX; Fig. 4c). At 17-17.6 km along profile \( C_1 \) (body X; Fig. 4c), an increase in the magnetic and gravity field was observed (Fig. 4a and 4b respectively). These anomalies are interpreted to originate from a basaltic rock with higher magnetic susceptibility and density (Table 1) compared to the surrounding rhyolitic rocks, and it continuous down to \( \sim 2 \) km depth. The bottom contact of this basalt with the felsic volcanic rocks might be related to the reflector \( R_7 \). At \( \sim 20.8 \) km, the north-dipping contact between the mafic volcanic rocks, dominated by basalt, and the Jörn granitoid (XI-XII; Fig. 4c), fits well in particular with the potential field model and reflector \( R_6 \) (Fig. 4c). The Jörn intrusive complex is divided into four distinct intrusive phases (GI – GIV; Wilson et al., 1987), each with different petrophysical properties (Kathol and Weihed, 2005). The oldest intrusive phase (GI), intersected by the profiles \( C_1 \) and \( C_2 \), is of a higher density (2660-2800 kg/m\(^3\)) and wider variation in magnetic susceptibility (30-5000×10\(^{-5}\) SI) compared to the younger phases (GII – GIV). The gravity and magnetic models suggest a depth extension of \( \sim 1-1.2 \) km for the Jörn intrusion along \( C_1 \) (body XII; Fig. 4c).

The fit between measured and calculated gravity from density/geological model along profile \( C_1 \) is reasonable, as indicated by the RMS=4.57%, whereas the fit between measured and calculated magnetic field from the model is somewhat more poor (RMS=8.68%).

Profile \( C_2 \)

Since the southern parts of profile \( C_2 \) are seismically transparent (Fig. 5d), potential field data are crucial for understanding the lithology contacts and hence modeling the subsurface in the SE parts of the study area. Felsic (body I; Fig. 5c) and mafic (body II; Fig. 5c) TIB-intrusive rocks are exposed in the beginning of the profile. The significant gravity high, between 1-5 km along the profile \( C_2 \) (Fig. 5b) is well correlated with the expected density high of the gabbroic rocks. This rock continues down to \( \sim 3 \) km depth. The bottom contact of the surrounding granitoid (body I; Fig. 5c) is sub-horizontal in the model, while the contact with the adjacent sedimentary rocks of the Vargfors Group (I-III; Fig. 5c) is steep (60 - 70°). At \( \sim 7 \) km along the profile \( C_2 \), the lithology changes from granitoid to a thin layer of Vargfors Group meta-sedimentary rocks, which crop out in a \( \sim 250 \) m wide area on the surface (body III; Fig. 5c). This part of profile is associated with a gravity low (Fig. 5b), suggesting that the meta-sedimentary rocks gets wider towards depth and continues down to \( \sim 2.6 \) km. However, there is no clear indication of the contact between felsic intrusive and meta-sedimentary rocks in the magnetic anomaly (Fig. 5a), which might be due to similar magnetic susceptibilities for the meta-sedimentary rocks and the granitoids. However, further along the profile, the magnetic field intensity increases, indicating shallow but highly magnetized basalt (body IV; Fig. 5c). At 11.6 km, close to the southern contact of the Vargfors basin (V-VIII; Fig. 5c), the gravity field increases (Fig. 5b), which cannot be explained by the density contrast between Skellefte Group rhyolites and Vargfors Group conglomerates. The gravity modeling indicated that any further extension of the Vargfors basin conglomerate results in a misfit between the measured and calculated gravity field; whereas the magnetic model did not affected considerably, indicating that Vargfors basin extends at depth down to \( \sim 1 \) km (body VIII; Fig. 5c) and overlies Skellefte Group basalt (body VII; Fig. 5c) with higher density compared to conglomerate and rhyolite. Contrasting, Dehghannejad et al. (under review) relates the reflectors \( R_9 \) and \( R_15 \) (Fig. 5d) to the synformal structure of the Vargfors basin,
Figure 5. Potential field modeling along profile C2 based on interpretation of reflection-seismic data. (a) Magnetic model of profile C2. (b) Gravity model of profile C2. (c) The depth section and geological structures at depth. (d) Integration of the seismic reflectors and model based on interpretation of potential field data.
extending to a depth of ~ 1.7 km. To explain the sharp gravity anomaly at 13-13.5 km, the basalt has to extend through the Vargfors basin to the surface. There is, however, no field evidence to support that, also, no reflector reaches the surface in this part of profile C2. Hence, one possibility is that the gravity data is not correct. From 17.2 km towards the northern end of the profile, the sub-horizontal reflector R5 (Fig. 5d) might be an indication of the bottom contact between Jörn GI-type intrusive rocks and an underlying basalt which crops out at ~ 17.3 and 18.2 km (bodies X and XI; Fig. 5c). Reflector R6 agrees well with the potential field models and corresponds to a north-dipping break-back fault, which marks the southern contact of the Jörn intrusive complex. This is also comparable with the reflector R6 in profile C1 (Bauer et al., 2011; Dehghannejad et al., under review). Mafic volcanic and volcaniclastic rocks that intruded along the fault (Bauer et al., under review) are believed to be the source of the magnetic high at 18.2 km. The magnetic and gravity models for the Jörn intrusive complex along C2 suggest that intrusion-depth increases towards its center (body XI; Fig. 5c).

The calculated field from the gravity model of profile C2 reasonably well fit the measured data (RMS=8.42%) and the calculated magnetic field shows a somewhat better fit (RMS=5.68%) compared to the fit for profile C1.

Profile C3
Due to the crooked CDP line for the seismic profile C3, the gravity and magnetic modeling of profile C3 was divided into two parts (Figs. 3a and 3b). A series of SW-dipping reflectors (R1, R2 and R3) in the southern part of profile C3, have earlier been interpreted as inverted normal faults, similar to interpretations along profile C1 (Fig. 6d; Dehghannejad et al., under review). The gravity data (Fig. 6b) suggest that the southernmost part of profile C3 (until ~5 km) is dominated by south-dipping Vargfors Group mudstones and sandstones (body I; Fig. 6c), and the contact between sandstone-mudstone and rhyolite (I-III; Fig. 6c) is consistent with reflector R1. However, the magnetic highs, located within different intervals, indicate that the meta-sedimentary rocks may be associated with minor intercalations of basalt with high magnetic susceptibilities. The slightly lower density of rhyolites compared to the mudstone-sandstone succession (Table 1), and lower concentration of basalt and decreasing their thickness along the Vargfors-Skellefte contact, decreases the Bouger anomaly eastwards (Fig. 6b). The reflector R5, which Dehghannejad et al. (under review) interpreted as a shear zone, reaches the surface at CDP 700, can be associated with a south-dipping basalt (Fig. 6d). The reflector R6 (Fig. 6d) coincides well with the suggested contact for R3 in profile C1, which indicates a south-dipping inverted normal fault at the contact between rhyolites and unspecified felsic volcanic rocks (III- VI; Fig. 6c). Local occurrence of basalts within the Skellefte Group rhyolites at 4.5-11.5 km along profile C3 may explain the local gravity anomalies but they do not show up as magnetic anomalies, which may be explained by the low magnetization of some basalt (Table 1 and body IV; Fig. 6c). The gravity low observed at ~ 8-12 km along profile C1, can also be observed in profile C3 between 8.2-17.7 km (Fig. 6b), is interpreted as a south-dipping unspecified felsic volcanic rock, with lower density than rhyolite (body VI; Fig. 6c). The reflector R40, which Dehghannejad et al. (under review) correlated with the reflector R4 in C1 and C2, fits well with the suggested potential field model for profile C1; hence we suggest the reflector R40 (Fig. 6d) to be the expression of a south-dipping contact between unspecified felsic volcanic rock and rhyolite to the north (bodies VI and III; Fig. 6c). Both gravity and magnetic data suggests that Gallejaur-type basalts (body XII; Fig. 6c), with high densities and high magnetic susceptibilities, underlies the mafic and felsic intrusions (bodies XIII and XIV; Fig. 6c) of the Gallejaur complex. At ~ 22 km, the gravity low and magnetic high (Figs. 6a and 6b), indicates lower density but higher magnetic susceptibility of mafic intrusive rocks (bodies XIII; Fig. 6c) compared to the adjacent basalts (body XII; Fig. 6c). The reflector R14 could therefore be related to the contact between the basalt and underlying felsic volcanic rocks (bodies XII and III; Fig. 6c and Fig. 6d). The joint interpretation
**Figure 6.** Potential field modeling along profile C3 based on interpretation of reflection-seismic data
(a) Magnetic model of profile C3 (b) Gravity model of profile C3 (c) The depth section and geological structures at depth (d) Integration of the seismic reflectors and model based on interpretation of potential field data.
of reflection-seismic, magnetic and gravity data suggest an approximate depth of 5-7 km for the bottommost part of Gallejaur complex, which is close to the estimated depth of 4 km by Enmark and Niska (1984) and 5-10 km by Kathol and Weihed (2005).

The fit of the data calculated from the density models to the measured data along this profile was good (RMS=3.27%) and better than the corresponding fit of the models to the magnetic data (RMS =8.67%).

5. Discussion

The result of this study from joint interpretation of the gravity and magnetic data to test and add constraints to the interpretation of the reflection-seismic data (Dehghannejad et al., under review), provided useful information for imaging the spatial relationship between geological structures down to a depth of ~5 km (Figs. 8a and 8b). Earlier investigations with reflection-seismic data in crystalline rocks showed low signal/noise ratios, due to i.e. strong metamorphism, alteration and folding (c.f. Malehmir and Bellefleur, 2009). This might be the reason for the transparent parts in three profiles, which complicated the interpretations, particularly in the south of profile C2 and also north of profile C3. In contrary to the stacking of data along profiles C1 and C2 which were conducted on straight lines, the stacking of the data along C3 was conducted as a crooked-line. This facilities correlating reflectors with the interpretation of potential field data, although with the expense of losing signal to noise ratio, since geometrical effect due to the out-of-the plane will be increased when profiles are stacked as crooked-lines (Rodriguez-Tablante et al., 2007).

In general, the gravity and magnetic models correlate well with the previous interpretations by Dehghannejad et al. (under review). The southern part of the model is dominated by a series of south-dipping reflectors (R1-R3) interpreted as inverted normal faults (Bauer et al., under review; Dehghannejad et al., under review). However, certain interpretations by Dehghannejad et al. (under review) are conflicting with the gravity and magnetic models (this study). This model suggests that reflection R1 in profile C1 is more likely related to the contact between the mafic intrusive rocks (TIB-type) and the Vargfors Group meta-sedimentary rocks (I-II in Fig. 4c) than to a high-strain zone separating meta-sedimentary rocks to the south from Skellefte Group rocks to the north as suggested by Dehghannejad et al. (under review). Furthermore, Dehghannejad et al. (under review) suggests the reflector R2 to relate to a high-strain zone (at 600 on the CDP line; profile C3) constituting the southern contact of the Finnliden antiform. The gravity and magnetic model suggests the contact between the mafic intrusive rocks to the south (body II; Fig. 4d) and Skellefte Group rocks to the north (body IV; Fig. 4d) as the origin for reflection R2. Nevertheless, the older fault could have acted as a controlling mechanism for the younger mafic intrusion. Furthermore, this indicates that reflector R2 in profile C2 might origin from a different contact than reflector R2 in profile C1. The modeled depth extent for the meta-sedimentary rocks and TIB mafic intrusive rocks is in agreement with the basal detachment depth of 3.5-4.5 km (bodies I and II; Fig. 4c and Fig. 7b) promoted by Dehghannejad et al. (under review).

Further to the north, Dehghannejad et al. (under review) interprets the diffractions D1-D4 in profile C1 to relate to fault blocks formed in between a series of south-dipping inverted faults (Fig. 6d). They therefore suggested that D1, D2 and R10 can indicate the contact between the meta-sedimentary rocks and underlying meta-volcanic rocks of the Skellefte Group. However, gravity and magnetic data can not prove such subdivision of blocks (Figs. 6a and 6b), since rocks in this part show similar petrophysical signatures (Table 1). However, the potential field data suggests that the diffractions D3 and D4 might origin from the contact between Skellefte Group unspecified felsic volcanic rocks and rhyolites (Northern part of VI-III; Fig. 6c and Fig. 6d). Yet, due to the low amplitude or absence of gravity and magnetic anomalies, we could not test the suggested model with potential field data.
Electrical and drill-hole data have previously suggested the potential locations for VMS deposits in this area (Tavakoli et al., under review).

According to Dehghannejad et al. (under review), the reflector R₄ reaches the surface close to the southern contact of the Vargfors basin. Their interpretation, R₄ being an inverted normal fault, is in agreement with the presented model. Furthermore, this fault could explain the contact between unspecified felsic volcanic rocks of the Skellefte Group (bodies V; Fig. 4c and VI; Fig. 6c) and the underlying rhyolites (bodies IV; Fig. 4c and III; Fig. 6c). At 11.6 km along profile C₂, close to the southern contact of the Vargfors basin (V-VIII; Fig. 5c), the gravity field increases unexpectedly (Fig. 5b), which cannot be explained by the density contrast between Skellefte Group rhyolites and Vargfors Group conglomerates. A high density rock located beneath the sediments is therefore needed to explain the anomaly. Dehghannejad et al. (under review) relates the reflectors R₈ and R₁₅ (Fig. 5d) to the synformal structure of the Vargfors basin, extending to a depth of ~ 1.7 km. However, according to the (magnetic and) gravity model, with a high density basalt below the basin, the basin is somewhat more shallow and ends at a depth of ~ 1 km (Figs. 5a and 5b). Nonetheless, the potential field and reflection-seismic data might be consistent as the reflectors R₈ and R₁₅ could represent synthetic and antithetic faults, which are responsible for basin formation but exceed into Skellefte Group volcanic rocks, whereas the sedimentary basin fill is limited to the uppermost km of the crust.

**Figure 7.** Integration of reflection-seismic data and suggested model by potential field data down to 5 km. (a) Crooked-line for profiles C1, C2 and C3 and gravity/ magnetic profiles. (b) Sliced 3D model of CSD based on gravity and magnetic data cut by depth section of reflection-seismic data for profile C1. (c) The sliced 3D model of CSD based on gravity and magnetic data cut by depth section of reflection-seismic data for profile C2. (d) The sliced 3D model of CSD based on gravity and magnetic data cut by depth section of reflection-seismic data for profile C3.
The 2D-forward model by Bauer et al. (2011) resulted in a depth extent of approximately 2 km for the normal faults and approximately 800 m for the sedimentary basin fill and is therefore supporting the interpretations.

North of CDP 1300 until the end of the profiles, Reflectors R\textsubscript{7}, R\textsubscript{9} and R\textsubscript{13}–R\textsubscript{14} cannot be explained by potential field data as none of these reflectors is linked to the model suggested by gravity and magnetic data. The interpretation of reflection R\textsubscript{9}, being the contact between the Jörn granitoid and the underlying basalt, is consistent with this contact suggested by potential field data on profile C\textsubscript{1} and C\textsubscript{2} (R\textsubscript{9}; Fig. 5c and Fig. 7c). Moreover, Reflection R\textsubscript{5} which coincides with the contact between felsic volcanic rocks and basalt at the end of profiles C\textsubscript{1} and C\textsubscript{2} cannot be linked to the basalt-rhyolite contact. The gravity and magnetic data suggest ~ 3 km depth for this contact (V-X; Fig. 5c).
The north-dipping reflector \( R_{13} \) which is explained as a break-back fault formed under the later stages of crustal shortening (Bauer et al., under review; Dehghannejad et al., under review) could not be verified with the model suggested by gravity and magnetic data (Fig. 6d). Nevertheless, as this north-dipping reflector shows a similar character as other north-dipping break-back faults, we suggest that they represent identical structures.

According to both, reflection-seismic and potential field data, the Gallejaur monzonite extends down to ~ 800m depth (Fig. 6d). This is consistent with the previous interpretation of the Gallejaur complex as a thin, sheet-like unit (Enmark and Niska, 1984). The gravity and magnetic models suggest that \( R_{14} \) could be the contact between Gallejaur basalt and underlying felsic volcanic rocks which terminates at approximately 2-3 km depth (XII-III in Fig. 6c and \( R_{14} \); Fig. 6d).

The here presented model and the previous presented reflector pattern by Dehghannejad et al. (under review) give constraints on the promoted fault pattern and structural framework by Bauer et al. (2011). It is furthermore in agreement with previous suggested extension and inversion in the Skellefte district (Allen et al., 1996) and is a useful base for future reconstruction of a bigger-scale tectonic framework and 3D-geological modeling.

6. Conclusion

Based on previous reflection-seismic data, acquired during 2009-2010 and potential field data, a regional 3D geophysical model of parts of the central Skellefte district was created. The reflection-seismic data provided preliminary information for potential field modeling, which in certain parts, was consistent with the interpretation of gravity and magnetic data. In general, a good correlation between reflection-seismic data and from the gravity and magnetic models improved the model certainty. Elsewhere, were no reflections existed or weak reflections did not reach the surface, gravity and magnetic data provided valuable constrains for the model with a reasonable fit between field data and calculated data. Therefore, the geometry and depth extent of major lithological units and its correlation with high-strain zones gave constraints by using potential field data, reflection-seismic data, geological observations and physical property of the rocks.

The result of the present study confirm that (I) The southern part of the study area is dominated by a series of south-dipping inverted normal faults with a depth extend of approximately 3-5 km (II) In the central part of profiles, none of the diffractions \( D_1, D_2, D_3 \) and \( D_4 \) which were observed on \( C_3 \), could be explained with the model based on potential field data. (III) A set of north-dipping break-back faults forms the southern contact of the Jörn intrusive complex (IV) The Gallejaur complex forms a thin, sheet like unit with a depth extent of approximately 800 m (V) Previous interpretation of the bottom contact of the Jörn intrusive complex and the underlying mafic rock is consistent with the potential field and reflection-seismic data (reflection \( R_6 \)), suggesting a depth extend ranging from approximately 1 - 2 km.

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4-DIMENSIONAL GEOLOGICAL MODELLING
OF THE SKELLEFTE DISTRICT, SWEDEN

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4-DIMENSIONAL GEOLOGICAL MODELLING OF THE SKELLEFTE DISTRICT, SWEDEN

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KEY WORDS: 4D-modelling, Skellefte District, ore deposits, Palaeoproterozoic, reflection seismic, inversion modelling, structural geology

ABSTRACT:
The Skellefte mining District is located in northern Sweden and comprises Palaeoproterozoic, metamorphosed, supracrustal rocks and associated intrusive rocks, which are all host to several ore deposits. Modelling was performed in the GoCAD software platform, using structural geology and petrology as well as magnetic, gravity, seismic, electric and electromagnetic geophysical data, in order to produce robust 3-dimensional models for the upper 3000 m of the Earth's crust in selected areas within the Skellefte District. Further interpretations from the regional structures and tectonics are providing a base for restoring the geological history of the district in time, the 4th dimension.

1. INTRODUCTION

The project aim is to develop geological models that visualise the key spatial, geological, geophysical, geochemical and economic parameters of the western and central part of the Skellefte mining district, including the known ore deposits, in 3D and 4D. This is achieved by using detailed thematic geo-information gathered at scales from regional down to individual deposits. The distribution of geological units, structure and tectonics as well as magnetic, gravity, seismic, electric and electromagnetic geophysical data are utilised to produce robust 3-dimensional models for the upper 3000 m of the Earth's crust in selected areas within the Skellefte District, combining the results from new high-resolution reflection seismic recordings, and other high-resolution geophysical measurements with results from new detailed structural mapping of the same areas. Data are modelled in the GoCAD software platform (Paradigm) to construct the 3D GIS models, which are further interpreted in the 4th dimension with constraints from regional structures and tectonics.

The project is subdivided into two main parts: 1) geological investigations and 2) geophysical measurements and interpretations. The aim of the geological field work is to derive a robust structural evolutionary model which can then be utilised in combination with the geophysical data to derive a 4D evolutionary model for the Skellefte district. In the second subproject, new high resolution seismic data are collected together with new electromagnetic data from the same areas. The seismic data are processed and, if feasible, migrated, then interpreted together with the electromagnetic data and other geophysical and geological information.

2. MODELLING

2.1 Geological framework

The bedrock of the Skellefte District is composed of 1.95 – 1.85 Ga Palaeoproterozoic island-arc volcanic rocks and associated intrusive rocks (Allen et al. 1996). The dominant unit is the 1.89 – 1.88 Ga, ore-bearing, Skellefte Group, comprising mainly felsic, meta-volcanic rocks, while the above lying Vargfors Group is characterised by metasedimentary rocks (Kathol and Weihed 2005). The intrusive rocks consist of calc-alkaline, multiple phase intrusions (1.9 – 1.85 Ga). The rocks were deformed and metamorphosed during the Svecofennian Orogeny at 1.90 – 1.80 Ga (Weihed et al. 2002).

2.2 Geological interpretations

The Kristineberg area in the western part of the Skellefte District forms the first focus area of the project. So far, the focus has been on providing reliable background geological data to enable construction of a solid 3D model of the area. These pre-3D modelling steps included studies on the orientation of the magnetic properties of rocks (AMS) and determination of their age. The AMS studies indicate a two-phase crustal evolution, including early, reverse, dip-slip faulting that was overprinted by strike-slip shearing. The first, dip-slip deformation phase is also considered largely responsible for the structural geometry of the area, including the shape of the Kristineberg ore bodies. The geochronological results are still preliminary, and will after processing, give further constraints on the timing of intrusive activity and ore-formation.

The second focus area of the project comprises the Central Skellefte District where structural mapping was carried out in

* Corresponding author
the area around the Vargfors basin. Detailed mapping of structures and stratigraphy resulted in a better understanding of the geological history. A part of the Vargfors basin was restored by unfolding and unfaulting each single fault (Fig. 1). The restoration reveals the formation of fault bound compartments with varying internal structures and stratigraphy caused by the formation of a complex pattern of pre-sedimentary normal- and transfer faults. The primary geometry and different petrophysical properties of the lithologies control the later structural overprint. This modelling derived a 3D voxel-model of the deformation history and is the base for kinematic modelling in the area.

2.3 Geophysical interpretations

Inversion modelling of gravity and magnetic data has been carried out in the central Skellefte District, using petrophysical properties, measured from 220 samples from 30 drill cores in the vicinity of the seismic and resistivity profiles. The apparent resistivity and induced polarization of the shallow subsurface along two seismic profiles in the central Skellefte District were measured, using the pole-dipole method as a survey array, in order to identify geological features with a thickness of >100 meters (Fig. 2). Testing different geological scenarios in 2D and 3D is executed to create a forward model of the resistivity profiles and to check how different geological scenarios will respond after the inversion.

Four new high resolution reflection seismic profiles (85 km in total) have been acquired, two perpendicular profiles in the Kristineberg mining area and two nearly parallel profiles in the central part of the Skellefte District. A series of steeply dipping (Fig. 3; M1 and C1) to sub-horizontal reflections (Fig. 3; E1 and E2) could be identified, allowing correlation with surface geology. Some of these reflections (e.g., M1) appear in direct connection with the main ore horizon in the Kristineberg mine, which extends down to a depth of about 2 km.

2.4 Modelling techniques

A multi-scale model was constructed by integrating data from regional to outcrop scales (Fig. 4). Due to a negligible topographic relief at this scale (less than 400 m elevation difference within an area of ca. 2400 km²), it is sufficient to construct a digital elevation model (DEM) from regional scale topography data, provided by the Swedish survey (Lantmäteriet), using “triangulated irregular network” (TIN) interpolation methods in ArcGIS (ESRI). Geological surface information was collected, combined and simplified from published maps (Allen et al. 1996, Kathol et al. 2005) and own field mapping.
Structural geological measurements from own field-campaigns as well as data provided by the mining company Boliden AB and the Swedish geological survey (SGU) were integrated. To provide sub-surface depth control on the general geometry of geological bodies, seismic, gravimetric and resistivity profiles as well as drill hole logs were converted to DXF-format and imported into GOCAD (Paradigm). In the regional scale, the outlines of the main geological bodies were directly drawn onto the profiles. In areas where no profiles were available, geological modelling was carried out by using the sparse-module (Mirageoscience), which allows the import of structural field-measurements in three-dimensional orientation. After the outline digitalisation of the main geological units and bodies (map traces), the properties of the structural measurements can be transferred onto these map traces and interpolated along them. This allows construction of e.g. mean bedding planes from strike-dip readings but also construction of foliation planes and simple fold structures (Fig. 5). However, prerequisite for this procedure is a certain amount of field data.

In the deposit scale, the outlines for the main ore bodies were provided by Boliden AB in DXF-format. The models for the minor ore bodies are based on digitised level-plans and cross-sections, drawn according to drill hole data. Restoration of small-scale geological profiles (Fig.1) was operated so far manually along selected geological cross-sections, where sufficient knowledge about the deformation history is evident. Due to the complicated kinematic evolution it was needed to understand the complex behaviour of different rock types during their deformation. Thin section analyses showed the role of carbonate-rich basal layers as glide-horizons during crustal shortening. Based on these results tests on different types of kinematics scenarios are analysed at present using the 3D and 4D MOVE software (Midland Valley Exploration Ltd.) as shown in figure 6.

Figure 4. Flow chart visualizing use of primary data (sources in parentheses) and work flow towards the existing 3rd and the planned 4th dimension.

Figure 5. Modelling a folded surface from structural measurements using the Sparse-module in GoCAD. A: Creating a map-trace; B: Transferring and interpolating dip-values; C: Creating a grip-frame; D: Creating a surface

Figure 6. Modelling a small-scale inverted Vargfors basin with simulation of a ductile basal-layer in MOVE. A: primary shape of the half-graben; B: Starting point of folding during crustal shortening; C: Increased gliding along the base; D: Formation of break-back faults during progressive shortening.
3. FUTURE WORK

The project will after this field season enter a new phase, where all acquired data will be used to refine the existing 3D-model (Fig. 7) of the western and central Skellefte district. Results from these studies will be combined with the 3D-model in order to construct the 4-dimensional geological evolution (especially deformation history) of the Skellefte District. Furthermore the results can be used in the future to apply kinematic behaviours of rock packages to areas where little or no surface data is available, to be able to increase predictability of undiscovered ore deposits.

4. REFERENCES


5. ACKNOWLEDGEMENTS

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