Stratigraphy and Geochemistry of the Palaeoproterozoic Dannemora inlier, north-eastern Bergslagen region, central Sweden.

PETER DAHLIN
The Palaeoproterozoic Dannemora inlier is situated in the north-eastern Bergslagen region. The inlier consists of primary and reworked volcanic deposits, stromatolitic limestone and skarn that have been subjected to upper greenschist facies metamorphism. Thicknesses of the different volcanic deposits indicate deposition within a caldera, where syn-volcanic alkali alteration was strong. The deposition was submarine and below wave base in the eastern part of the inlier, but above wave base in the central part where erosion channels together with cross-bedding occurs frequently.

The Dannemora Formation is the volcanosedimentary succession of the inlier. Two borehole profiles, a northern and a southern, cover the whole Formation and show different alteration patterns. A strong depletion of Na₂O and enrichment of K₂O dominate in the southern profile, whereas this pattern is not as evident in the northern profile. The uppermost section of the totally eight constituting the Formation, is intercalated with ore-bearing dolomitic limestone and skarn, and has experienced at least two episodes of alteration. An anticline has been established lithogeochemically from immobile element ratios and the reoccurrence of an accretionary lapilli bed.

Numerous altered sub-alkaline, calc-alkaline and basaltic dykes have been recorded in the Dannemora inlier. They are the result of mixing and fractionation of at least three magmatic sources and carry a mixed signature of subduction zone and within-plate volcanic tectonic setting.

A seismic profile across the Dannemora inlier images a strong reflector package that dips c. 50° E to the east of the inlier. This package coincides with the polyphase, E-up reverse, brittle-ductile Österbybruk deformation zone (ÖDZ). Yet another steep reflector in the Dannemora ore-field extends to a depth of more than two kilometres. This reflector might represent either a deep-seated iron deposit or a fluid-bearing fault zone.

Keywords: Dannemora, Bergslagen region, volcaniclastic rocks, sedimentology, metasomatic alteration, lithogeochemistry, reflection seismic

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"I start to think and then I sink into the paper like I was ink
When I'm writing, I'm trapped in between the lines,
I escape when I finish the rhyme..."
Rakim
This thesis is based on the following papers, which are referred to in the text by their Roman numerals.


IV **Dahlin, P.** Lithogeochemistry and alteration of the Palaeo-Proterozoic metavolcanic succession in the Dannemora area, Bergslagen region, Sweden. (*Manuscript*)
The following publication was finalised during my PhD-study but is not included in this thesis:

Personal contribution

Paper I, I conducted the field work, rock sampling, core logging, reinterpretation of the stratigraphy and wrote the paper. Interpretation of pyroclastic textures and sedimentary features were performed with support from Rodney Allen and Håkan Sjöström.

Paper II, field work, core logging, sample preparation and the ICP-MS analyses was executed by me. The interpretation of the data was made in collaboration with Åke Johansson and Ulf Bertil Andersson.

Paper III, My contribution was advising on the planning of the seismic profile, providing samples and the geological interpretation together with Karin Högdahl and Håkan Sjöström.

Paper IV, I conducted all field work, core logging and sampling. The writing benefitted considerably from the input by Karin Högdahl. The manuscript was also improved by the input from Rodney Allen and Håkan Sjöström.
### Abbreviations

<table>
<thead>
<tr>
<th>Abbreviation</th>
<th>Description</th>
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<tbody>
<tr>
<td>CDP</td>
<td>Common Depth Point</td>
</tr>
<tr>
<td>CHUR</td>
<td>Chondritic Universal Reservoir</td>
</tr>
<tr>
<td>DM</td>
<td>Depleted Mantle</td>
</tr>
<tr>
<td>FPD</td>
<td>Floatation Pumice Deposit</td>
</tr>
<tr>
<td>Ga</td>
<td>Giga anno</td>
</tr>
<tr>
<td>GDG</td>
<td>Granite-Diorite-Gabbro</td>
</tr>
<tr>
<td>GRZ</td>
<td>Gävle-Rättvik Zone</td>
</tr>
<tr>
<td>HFSE</td>
<td>High Field Strength Element</td>
</tr>
<tr>
<td>LILE</td>
<td>Large Igneous Lithophile Element</td>
</tr>
<tr>
<td>LP-HT</td>
<td>Low Pressure-High Temperature</td>
</tr>
<tr>
<td>LREE</td>
<td>Light Rare Earth Element</td>
</tr>
<tr>
<td>Ma</td>
<td>Mega anno</td>
</tr>
<tr>
<td>MORB</td>
<td>Middle Ocean Ridge Basalt</td>
</tr>
<tr>
<td>PDC</td>
<td>Pyroclastic Density Current</td>
</tr>
<tr>
<td>REE</td>
<td>Rare Earth Element</td>
</tr>
<tr>
<td>SSZ</td>
<td>Singö Shear Zone</td>
</tr>
<tr>
<td>TIB</td>
<td>Transscandinavian Igneous Belt</td>
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<tr>
<td>TWTT</td>
<td>Two-Way Travel Time</td>
</tr>
</tbody>
</table>
Contents

1. Introduction........................................................................................................13

2. Geological setting of the Bergslagen region ..................................................15
   2.1 The Dannemora inlier..............................................................................17

3. Volcanology....................................................................................................19

4. Volcanic textures: explanations and implications.........................................22
   4.1 Phenocrysts..........................................................................................22
   4.2 Glassy, juvenile fragments and devitrification......................................23
   4.3 Fiamme...............................................................................................25
   4.4 Accretionary lapilli..............................................................................26
   4.5 Flotation pumice deposit......................................................................27

5. Lithogeochemistry and isotopes..................................................................29
   5.1 Alteration.............................................................................................30
   5.2 Barium feldspar..................................................................................32
   5.3 Isotope systems and petrogenesis.......................................................32

6. Reflection seismic survey ............................................................................35

7. Summary of papers ......................................................................................37
   7.1 Dannemora Formation and its alteration (Paper I and IV).................37
   7.2 Dykes (Paper II)................................................................................39
   7.3 Reflection seismic survey (Paper III)...................................................40

8. Sammanfattning på svenska..........................................................................41

Acknowledgement ..........................................................................................43

References........................................................................................................45
1. Introduction

The Bergslagen region is a well-known ore province, in south-central Sweden. The Dannemora area is located in the eastern part of the region and represents one of the best preserved supracrustal successions. The only research made in the Dannemora area in modern time was conducted by Lager (2001), who presented a sedimentological interpretation of the succession.

The PhD-project resulting in this thesis was initiated by Lennart Falk, Dannemora Mineral AB and further developed by Rodney Allen, New Boliden, Håkan Sjöström, Uppsala University and Magnus Ripa at Geological Survey of Sweden (SGU). Lennart’s enthusiasm not only initiated this study, but also a new generation of ore related research in the Bergslagen region. The PhD-project was co-financed by Dannemora Mineral AB (the main part) and SGU (SGU-FoU project 60 1453/2006), and was part of a two-part project in collaboration with Luleå University, Uppsala University, New Boliden and Dannemora Mineral AB.

The aim of this study was to deduce the deposition history of the supracrustal succession, to interpret the geochemical characteristics of the volcanioclastic rocks and the isotope- and geochemistry of basaltic dykes to determine the geological evolution of the Dannemora area. The low metamorphic grade and exceptionally well preserved rocks facilitated the study. In addition, a seismic reflection investigation was conducted with the aims to localize and image major geological features such as rock contacts and deformation zones, and to constrain the geometry and possible extension at depth of the Dannemora iron ore body.

The thesis is divided into six parts. The second chapter deals with the geological setting of the Bergslagen region and the Dannemora area. Relevant aspects of volcanology, magmas, volcanic gases, eruption types and deposits are treated in chapter three. The emphasis in this chapter is the characteristics of deposits found in the Dannemora area. Chapter four explains and describes the formation of volcanic textures as a basis for the interpretation of how the deposits formed. The analytical methods and theories encountered during this study are handled in chapter five. Lithogeochemistry focuses on determining signature of magmatic processes, altera-
tation of the metavolcanic rocks and basaltic dykes. Chapter six presents a short introduction to seismic reflection. Summary and conclusions of the thesis are found in the last chapter.

Two master thesis projects developed in association with the PhD-project. One focussed on kinematics and deformation mechanisms in different minerals of a major shear zone (Björbrand) and the other dealt with geothermo-barometry of sulphides: sphalerite barometry and arsenopyrite thermometry (Åberg).

As part of the 33rd International Geological Congress 2008 four outcrops in the Dannemora area were on display during an excursion through the Bergslagen region.
2. Geological setting of the Bergslagen region

The Dannemora ore-field is located in the north-east part of the intensely mineralised Bergslagen region. The region is known for its long history of mining that dates back to the Early Middle Ages (Geijer and Magnusson, 1944). The region belongs to the Palaeoproterozoic Svecofennian orogen of the Fennoscandian Shield (Fig. 1). This orogen is bounded to the north by Archean rocks, to the west by the Transscandinavian Igneous Belt (TIB), the Palaeozoic Caledonian orogen to the northwest and in the southwest by the Sveconorwegian orogen. The accretionary and collisional Svecofennian orogeny has been divided into four stages partly overlapping both spatially and temporally (e.g. Nironen, 1997; Lahtinen et al., 2009) that occurred between 1.92 and 1.77 Ga (Lahtinen et al., 2009). The formation of volcanic arcs and accretion of microcontinents prevailed during the initial stage that was succeeded by extension of the newly formed crust, continent-continent convergence resulting in crustal thickening and finally gravitational collapse (Lahtinen et al., 2005; Korja et al., 2006; Lahtinen et al., 2009).

Fig. 1. The northern part of the Bergslagen region and Fennoscandian shield (modified after Stephens et al., 2009). Dannemora is shown with a D. The Bergslagen region is bounded by Gävle-Rättvik Zone (GRZ) to the north and Singö Shear Zone (SSZ) to the north-east.
Metaplutonic rocks with subordinate supracrustal inliers make up the Bergslagen region (cf. Stephens et al., 2007). They were originally formed during extension in a back-arc setting inboard an active continental margin. The metaplutonic rocks are c. 1.90-1.87 Ga with granitic to gabbroic compositions (GDG-suite; Stephens et al., 2007, 2009) and the inliers comprise coeval supracrustal rocks (e.g. Lundström et al., 1998; Andersson et al., 2006). The supracrustal succession is both over and underlain by elastic metasedimentary rocks. The latter dominate in the eastern parts of the region (Claesson et al., 1993; Andersson et al., 2006).

At 1.87-1.86 Ga the extension ceased and closing of the back-arc basin coincides with the convergence between the Bergslagen microcontinent in the south and the Keitele microcontinent in the north (Lahtinen et al., 2005). The shortening led to the formation of tight to isoclinal folds (F₁) with steep axial surfaces and sub-horizontal fold axes. Thus, most of the mineralisations and their host-rocks in the region are steeply dipping, but are often overprinted by both a second and a third deformation phase like for instance in the Dannemora and surrounding areas (Persson and Sjöström, 2003).

Two intrusive suites constituting 1.87-1.84 Ga granitic to gabbroic rocks in the northern and southern parts of the Bergslagen region and 1.85 to 1.75 Ga granites with associated pegmatites are subordinate to the syn-volcanic plutonic rocks (Stephens et al., 2007, 2009).

Mafic dykes in the Dannemora area are interpreted to be contemporaneous with 1.87-1.86 Ga basaltic dykes in an area to the north-east (Hermansson et al., 2008). The chemical signatures of the dykes have mainly an active continental margin and subordinated within-plate basalt affinities (Paper II).

The supracrustal inliers in the Bergslagen region have been interpreted to represent deformed and metamorphosed relics of large, rhyolitic to dacitic caldera volcanoes with associated thick and extensive ignimbrites (Allen et al., 1996). The deposition in shallow marine environment is indicated by reworked volcaniclastic deposits and frequent stromatolitic beds (Boekschoten et al., 1988; Allen et al., 1996; 2003; Paper I).

To the north and north-east the Bergslagen region is bounded by a major deformation zone (Tirén and Beckholmen, 1990; Stephens et al., 1994). The northern part of the zone, referred to as the Gävle Rättvik Zone (GRZ; Fig. 1; Högdahl et al., 2009), has been suggested to be the westwards continuation of the Singö Shear Zone (SSZ), and represents a crustal scale domain or terrane boundary (Högdahl et al., 2009, Stephens et al., 2009). This system delineates the low grade Bergslagen region from a zone of mixed tectonic units of higher grade rocks (Stålhös, 1991). The southern boundary is de-
fined by the Loftahammar tectonic belt (Stephens et al., 2009), and to the west the region is truncated by 1.80-1.65 Ga TIB granitoids (Fig. 1; e.g. Högdahl et al., 2004).

The peak LP-HT amphibolite facies metamorphism occurred at 1.87 Ga in the northern part of the Bergslagen region (Andersson et al., 2006) and locally with greenschist conditions areas (e.g. Sundius, 1923; Stålhös, 1991). The southern part of the region is dominated by upper amphibolite facies migmatites and associated anatectic granites (Fig 1; Stålhös, 1969; Stephens, 2007). A second regional metamorphic pulse peaked at c 1.80 Ga (Andersson et al., 2006).

2.1 The Dannemora inlier
All rocks in the Dannemora area are metamorphosed, but for simplicity the meta-prefix is hitherto excluded from the rock names. The Dannemora inlier consists of supracrustal rocks that is surrounded by the GDG-suite (Fig. 2). The supracrustal rocks in the inlier belong to the 700-800 m thick Dannemora Formation, which consists of calc-alkaline, mainly rhyolitic to dacitic volcaniclastic rocks and dolomitic limestone (Lager, 2001; Paper I). The Formation is divided into a lower and an upper member (Paper I). The lower member is about 500-600 m thick and is sub-divided into a c. 150 m thick subunit 1 and a c. 400 m thick sub-unit 2. This member consists of thick massive ignimbrites and their reworked counterparts. The division into two sub-units is based on the occurrence of an ash-siltstone bed separating two thick ignimbrites. This member has no known mineralisations and correlates to the intensive volcanic stage of Allen et al., (1996). The upper part of the lower member has a U-Pb zircon TIMS age of 1 894 ± 4 Ma (Stephens et al., 2009). The upper member correlates to the waning volcanic stage (Allen et al., 1996) and constitutes an over 80 m thick stromatolitic limestone with intercalated ash-siltstone laminae and beds (Lager, 2001; Dahlin et al., 2012). This member hosts the iron oxide deposit with the oldest recorded mining activity dating back to late 1400’s (Tegengren, 1924).

The inlier has been affected by at least two fold phases resulting in two clearly separable cleavages (S₁ and S₂) and a dominating stretching lineation (L₂) but few meso-scale folds. Isoclinal to tight, upright km-scaled F₁-folds have been defined by the shift in stratigraphic younging directions, by geochemical data (Paper IV) and by bedding/cleavage relationships (Dahlin and
Sjöström, 2010). The F₂-folds have moderately north-west plunging fold axes and steep roughly NW – SE axial surfaces.

Fig. 2. A simplified geological map showing the Dannemora inlier (Modified after Stålhös, 1991, © Swedish Geological Survey Dnr: 30-2169/2007).
3. Volcanology

Volcanic rocks with silica > 68 wt % are classified as rhyolitic whereas the concentration in basalt is c. 45-55 wt %. The high silica content and relatively low eruption temperature (< 900°C) make granitic magmas highly viscous, which lowers the volatile diffusion rate and hinders degassing (Wallace and Anderson, 2000). Consequently, these magmas contain large amount of volatiles, which govern the explosivity and velocity of eruptions (Wallace and Anderson, 2000).

Both pressure and solubility of gases decrease in the magma on its way through the crust towards the surface. This pressure decrease results in bubble nucleation at the so-called exsolution surface, and eventually fragmentation of the magma at the fragmentation surface in the conduit (Sparks, 1976; Cashman et al., 2000). Fragmentation index, the measurement of the degree of fragmentation, is dependent on the volatile content and the magma composition (Fisher and Schmincke, 1984). On the basis of the fragmentation index and the dispersal index i.e. the areal extent of tephra, Walker (1973) classified different eruption types (Fig. 3A). Explosive plinian eruptions are thus associated with rhyolites, whereas Hawaiian eruptions are characterised by effusive activity of basalt. Addition of external water, such as ground or surface water, results in phreatomagmatic eruptions, which have an even higher fragmentation index, and consequently a more explosive nature than eruptions solely driven by magmatic volatiles.

Once an explosive eruption develops, whether a convecting or collapsing eruption column is sustained, it is governed by discharge rate, vent radius, exsolved water content and exit velocity (Fig. 3B; Wilson et al., 1980). During the course of an eruption the state of the eruption type will alternate between convecting and collapsing column. The eruption type dictates the deposition style of the ejected tephra; fall deposits will blanket a landscape whereas ignimbrites are localised to depressions (Freundt et al., 2000). As an example, at a constant vent radius of 300 m and a decrease in exit velocity from 400 m/s to 300 m/s (Fig. 3B), the eruption type will shift from a sustained convecting eruption column to a collapsing one and the formation of a pyroclastic density current (PDC). The acronym PDC is the established term
for a ground hugging current consisting of gas and clasts, i.e. juvenile, lithic
and crystal fragments (Branney and Kokelaar, 2002). Dufek et al. (2012)
concluded that the fragmentation depth also governs the eruption type; frag-
mentation at deep levels renders finer ash that hampers eruption through the
conduit and result in collapse of the eruption column.

Fig. 3. A) Volcanic eruptions types are based on fragmentation and dispersal of tephra. From left to
right as the dispersal and degree of fractionation increase, eruptive activity changes from effusive
to explosive (Modified after Walker, 1973). B) Deposit type is dependent of vent radius, exsolved
water content, exit velocity and discharge rate of the magma (Modified after Wilson et al., 1980).
Arrow shows the change of deposition type by decreased exit velocity and/or exsolved water at fixed
vent radius and magma discharge rate.

Ignimbrites are deposited by PDC’s and contain different clasts and crystal
fragments, and are generally characterised by lack of layering (e.g. Ross and
Smith, 1961; Branney and Kokelaar, 2002). Juvenile clasts are formed from
disintegrated and cooled magma, and consequently have a glassy matrix. In
contrast, lithic clasts are parts of wall rock that are ripped from e.g. the con-
duit walls. Juvenile clasts with vesicles exceeding 60 % are defined as pum-
icc and they usually have a density lower than 1 kg/dm³ i.e. they float on
water (Cashman et al., 2000). An ignimbrite can often be divided into sever-
al flow units, which represent deposition of discrete pulses. Lithic clasts are
more frequent at the base of the deposit whereas pumice size often increases
upwards, but normal grading of pumice is also common in thin flow units
(Sparks, 1976). Lithic fragments are deposited preferentially proximal to the
volcano due to their relatively high density (Walker, 1985). Thus, the use of
grading as a structure for stratigraphic younging direction in ancient ignim-
brites should be applied with caution.
Natural barriers, e.g. hills and valleys, govern the movement of PDC’s and consequently the thickness of the ignimbrites. Caldera collapse structures can host ignimbrites (so-called ponding ignimbrites) of several hundred up to thousands of metres thick (Lipman, 1984). In contrast, the thickness of deposition in valleys and other topographic lows outside a caldera are usually more modest. Pyroclastic density currents are usually accompanied by an ash plume and its deposit, the co-ignimbrite, which might be as voluminous as the primary ignimbrite (Sparks and Walker, 1977; Branney and Kokelaar, 2002). The ash is primarily derived from the fragmentation of the magma and to a minor extent through the action of abrasion of the mechanically weak pumice clasts during the eruption and in the subsequent PDC.

Determination of the origin of volcanic rocks as pyroclastic or coherent in ancient metamorphosed and deformed rocks is a delicate task. The most reliable criteria that demonstrate a pyroclastic origin are the presence of fi-amme (section 4.3), accretionary lapilli (section 4.4), broken phenocrysts and bubble-wall shards (Manley, 1996). Deformation and metamorphism may obscure or even obliterate these textures; therefore studies of ancient and recent rocks require different approaches. In young pyroclastic rocks the primary devitrification products can give information about post-depositional processes (Gifkins et al., 2005a), whereas in the ancient equivalents, the mineralogy formed due to these processes might have been replaced repeatedly during prograde and/or retrograde metamorphism (Gifkins et al., 2005a). Still, in order to make a comprehensive interpretation of the geological history of metamorphosed volcanic rocks, a comparison with recent deposits is necessary.
4. Volcanic textures: explanations and implications

Characteristics in volcanic deposits provide evidence for eruption mechanism, emplacement temperature, depositional environment, stratigraphic younging direction and distance to source regions. Specific features form before eruption (e.g. phenocrysts), during deposition (e.g. accretionary lapilli) or after deposition (e.g. spherulites). In this chapter volcanic textures encountered in the Dannemora inlier are described.

4.1 Phenocrysts

Phenocrysts are mineral grains that are larger than the grain size of the matrix, and form in magmas in different parts of volcanic plumbing systems; in the magma chambers and/or during ascent through the crust prior to the eruption. The phenocrysts can be fragmented and in this sense become crystal fragments. The amount of phenocrysts and crystal fragments in a deposit can be of importance in distinguishing between different emplacement processes such as pyroclastic flow or tephra fall out, but can also be utilized for determining if the deposit was proximal or distal. Distal fall deposits normally show low abundance of crystals, in contrast to proximal deposits, but there are exceptions as PDC’s can carry crystals for long distances. The pumice fragmentation process results in crystal concentration in the ignimbrite (Sparks and Walker, 1977), and explains why relatively heavier crystals and crystal fragments form part of distally emplaced ignimbrites (Walker, 1972).

In the Dannemora area quartz is the dominating phenocryst and crystal fragment in the volcanic rocks. The quartz crystals are 0.5-5 mm; occur mostly as crystal fragments and rarely as euhedral to subhedral grains. The crystals are commonly rounded and embayed (Fig. 4A) and sometimes show polygonisation and sub-grain formation, which are secondary features caused by tectonic deformation (Passchier and Trouw, 2005). Plagioclase
phenocrysts are subordinated and Ca-rich varieties are evident from subsequent saussuritisation (Fig. 4B).

Fig. 4. A) Microphotograph of embayment in quartz. Crossed polarized light. B) Almost completely saussurised plagioclase of which only the rim is preserved. Microphotograph and crossed polarized light. Both photos are from samples collected in the Dannemora area.

Embayments in quartz phenocrysts (Fig. 4A) are very common in the volcanic rocks in the Dannemora area. It has been shown that embayments form through point corrosion of existing phenocrysts an indication of non-equilibrium in the magma chamber (Vernon, 2004). Embayments in quartz can be used to distinguish intrusive from extrusive rocks, as the former lack this feature (Vernon, 2004).

4.2 Glassy, juvenile fragments and devitrification

Glass, a common feature in volcanic rocks, is prone to devitrification, i.e. it becomes replaced by crystals (de is Latin for changed from and vitrum means glass). Devitrification of silicic glass produces different styles of crystal aggregates of cristobalite and/or feldspars collectively named spherulites (e.g. Lofgren, 1971). These are characteristic for hot pyroclastic deposits and lavas (Ross and Smith, 1961). Spherulites can have a large variety of shapes and develop at varying temperatures (Lofgren, 1974) and at different levels within an ignimbrite (Fig. 5). Textures formed by high temperature devitrification processes can be used to estimate the relative emplacement temperature and cooling history of pyroclastic deposits.
Glass rich pyroclastic deposits that are hotter than the glass transition temperature deform in a ductile manner and become welded (Fisher and Schmincke, 1984). Thus the originally clastic deposits become coherent rocks. The degree of welding depends on emplacement temperature, volatile content and compression by the overburden (Ross and Smith, 1961). The presence of spherulites indicates hot emplacement and possibly welding. Juvenile clasts i.e. pumices, are flattened into fiamme (section 4.3) in hot deposits. In low temperature deposits, the shape of the pumices might be preserved but are replaced by phyllosilicates such as clay minerals during subsequent diagenesis and metamorphism (McPhie et al., 1993).

Fig. 5. Idealized profile of an ignimbrite with defined textural zones. Spherical and pectinate shaped spherulites occur in the uppermost and thinner parts of the ignimbrite (Garth Tuff, Wales. Modified after McArthur et al., 1998).

In Dannemora, spherical spherulites with diameters of less than 1 mm (Fig. 6A) are found in the lower part of the stratigraphy in both massive pyroclastic deposits and in clasts in reworked volcanoclastic units. Pectinate spherulites are comb like and form when crystal needles grow perpendicular to the delimitations into the glassy shards (Fig. 6B). This type of spherulite grows at relatively low temperature and/or close to the upper surface of an ignimbrite (Fig. 5).
Spherulites with vugs are called lithophysae (Ross and Smith, 1961) and are formed early during cooling of lavas and densely welded pyroclastic deposits (McPhie et al., 1993). Lithophysae have been recorded from numerous ancient ignimbrite occurrences in Sweden e.g. Lake Rakkur, Norrbotten, (Lilljequist and Svensson, 1974) and Hällefors, Bergslagen (Lundström, 1995), but are absent in the Dannemora Formation, which excludes the presence of lavas and densely welded pyroclastic deposits there.

4.3 Fiamme

Fiamme are deformed juvenile clasts and the name is derived from the Italian word for flames. Juvenile clasts are mechanically weak and therefore become flattened during compaction, at either hot or cold conditions (Gifkins et al., 2005b). Flattening due to the weight of the overburden during diagenesis is referred to as cold conditions.

Under hot conditions pumice clasts and the glassy matrix have similar rheology and the originally feathery tips of the pumice clasts are therefore preserved during welding. At cooler conditions, pumice clasts and the matrix have different rheology. Consequently deformation affects the two materials differently, and compaction results in fiamme with wedge-shaped tips (Bull and McPhie, 2007). Thus, fiamme with feathery ends are diagnostic of welding compacted pumice clasts in contrast to fiamme with wedge-shaped ends, which are typical for diagenetically and tectonically flattened pumices (Bull...
Diagenetic compaction of juvenile clasts is facilitated by devitrification and replacement to weak phyllosilicates (Fig. 7A-B and 8A; McPhie et al., 1993). Distinguishing between welding (primary) and diagenetic (secondary) compaction of pumice can be difficult, especially when possible welding textures are overprinted by later deformation (Gifkins et al., 2005b).

All recorded fiamme in the Dannemora area have been deformed due to compaction during the diagenesis and subsequent tectonic deformation, but are still easily recognized. The rocks in the upper member of the Dannemora Formation have only been affected by low temperature devitrification, shown by the sericite-replaced glassy fragments that still have their original bubble-shaped walls (Fig. 7B).

### 4.4 Accretionary lapilli

Accretionary lapilli form when vapour condensates to water droplets in suspended ash clouds. These droplets accrete ash in concentric layers resulting in spherical ash aggregates. Schumacher and Schmincke (1991) distinguished two types of accretionary lapilli; rim and core types. The rim type has coarser grains in the centre than in the rim whereas the core type has finer grains in the centre than in the rim (Fig. 8B-C). Accretionary lapilli occur in both pyroclastic fall and surge deposits. The diametre of accretionary lapilli is related to the thickness of the erupted ash cloud (Gilbert and Lane, 1994).
In the Dannemora inlier accretionary lapilli are quite common (Fig. 8B). The lapilli have a diameter of up to 10 mm, suggesting that the height of the ash cloud in which they were formed was > 4 km (Gilbert and Lane, 1994). Such a great thickness of an ash cloud requires a subaerial eruption (Schumacher and Schmincke, 1991). Consequently, the eruption was likely subaerial, even though the depositional environment was submarine as indicated by ash-siltstone layers showing normal grading and water-escape structures (Fig. 9A).

![Fig. 8. A) Phenocryst-rich fiamme with wedge-shaped ends. Pen for scale. B) Accretionary lapilli with light core and dark rim. The elliptical shape is due to tectonic deformation. C) Rim-type accretionary lapilli. The core has slightly larger grain size than the rim. Micro photo, plane polarized light. Photos are from field and from samples collected in the Dannemora area.](image)

4.5 Flotation pumice deposit

Flotation pumice deposits (FPD) contain pumice embedded in ash originating from both air-fall and attrition (wearing) of the floating pumice (Fisher and Schmincke, 1984). During submarine emplacement of PDC’s, buoyant pumice floats up to the surface of the water column, and stays afloat until it becomes water-logged and sinks. Floating pumice rafts have been reported to travel great distances from the eruption site (Shane et al., 1998). If the
pumice vesicles are filled with hot gas an under-pressure is created during cooling, which leads to faster water absorption and sinking than for cold pumices (Whitham and Sparks, 1986). Floating pumice is easily displaced by wind and waves from the site of emplacement. Thus, formation of FPD’s requires settling of the pumice shortly after the eruption or in a confined area such as a caldera that hinders the dispersion of the pumice.

A unit of phenocryst-rich pumice blocks intercalated with ash-siltstone in the Dannemora inlier is interpreted as a FPD (Fig. 9B; Paper I). This kind of deposit is an excellent marker bed in metamorphosed and folded metavolcanic rocks and also gives a reliable younging direction in the grading upwards from massive ignimbrite to the mixture of pumice and ash-siltstone.

Fig. 9. A) Layered, accretionary lapilli bearing (not visible) ash-siltstone with water escape structure (highlighted by black arrow). Pen for scale. B) Phenocryst-rich pumice blocks (grey; highlighted by black arrow) intercalated with ash-siltstone (cream coloured). Due to subsequent folding the bedding-planes are steep to vertical. Stratigraphic younging direction is towards the east i.e. downwards in the figure. Hammer for scale. Both photos are from the Dannemora area.
5. Lithogeochemistry and isotopes

Lithogeochemistry is used in order to establish the chemical composition of rocks, their magmatic affinity, tectonic setting and alteration. For very fine grained volcanic rocks lithogeochemistry have to be used for classification, which instead can be done by petrography for coarse-grained rocks.

The compatibility of elements ($K_d$) is defined by their preference to liquid or crystalline phase during processes such as partial melting and fractionation crystallisation. Elements that preferentially are concentrated in the liquid phase are called incompatible. They have $K_d < 1$ i.e. physical parameters that make them less suited to fit into minerals like olivine or pyroxene in the mantle. These elements e.g. Zr and K are incompatible and thus too big/small or have too high/low charge to fit in minerals and consequently enriched in the magma during partial melting. By contrast Ni and Cr with $K_d > 1$ are compatible in mantle rocks and remain in the crystalline phase during partial melting.

Compatibility for elements can be tested by using different elements pairs in bivariate plots. If two elements are incompatible, they should follow a linear trend with a positive slope (Fig. 10A). In the case of incompatible-compatible elements, the data points will fit a linear trend with a negative slope (Fig. 10A).

During partial melting of the mantle in a subduction zone environment, the more incompatible large igneous lithophile elements (LILE’s) such as K and Na become more enriched in the magma than the less but also incompatible high field strength elements (HFSE) e.g. Zr and Nb (Fig. 10B). The LILE’s are enriched in the mantle relative to HFSE because the former are mobile in fluids released from the sediments that overlay the oceanic crust in the subducting plate. The balance of different elemental abundances can often allude to which tectonic environment the rocks were derived from. Magmas formed in subduction zones, however, are compositionally very complex as contributions to the final magma can be derived from a multitude of sources; the subducted oceanic crust, sediments deposited on the oceanic crust, the mantle wedge and from the overlying arc crust (e.g. Becker et al., 2000). Regardless of this large variety of sources subduction related magmas
are highly enriched in LILE compared to HFSE, and display typical trough for Nb-Ta (Fig. 10B). Magmas formed in a within-plate setting that produces ocean island basalts (hot spot), lack the highly enriched hump for LILE, have a gentle negative slope for HFSE and no Nb-Ta trough (Fig. 10B). The slope of the curve reflects the decreasing incompatibility of the elements from Ba to Sr and Ba to Yb. Island arc basalts, which are formed during subduction beneath an oceanic plate, have rocks that show an enriched pattern for LILE and Th but a nearly flat trend for the HFSE (Fig. 10B).

5.1 Alteration

Elements that are released from dissolved or replaced minerals and redistributed on the centimetre or greater scale by fluids during alteration processes are considered mobile. Mobility of an element can lead to mass loss in the original rock if the element is removed or to mass gain if added to the rock (e.g. MacLean and Barrett, 1993). In contrast, immobile elements remain in the rock mass, and become relatively enriched during mass loss of mobile elements and consequently relatively depleted during mass gain (e.g. MacLean and Barrett, 1993). As ratios of immobile elements remain constant during post-magmatic alteration processes they are routinely employed to group rocks of the same original, pre-alteration composition and to discriminate rocks with different primary compositions (e.g. Barrett et al., 2008).
The isocon-method is used to show mass transport in rocks. In isocon diagrams one least altered sample is plotted against one altered sample (Fig. 11). The immobility of elements e.g. Nb, Zr, TiO_2, and Al_2O_3 is tested by plotting and fitting a line through these data points and the origin, creating an isocon. If these elements show a good fit to a straight line they are interpreted as immobile. Other mobile elements that plot above the isocon have experienced mass gain and points plotting below the isocon mass loss (Fig. 11).

![Fig. 11. Example of an isocon-plot. The data points (filled squares) plotting below the isocon have experienced mass loss and above the isocon mass gain. The isocon is defined by four elements: Nb, Al_2O_3, TiO_2 and Zr (open circles). The numbers are the factor used for scaling e.g. Ba-concentration is divided 40 and Mo-concentration is multiplied by 30. Elements plotting close to the isocon are interpreted as immobile.](image-url)

The degree and type of alteration has successively been used as an exploration tool as mineralisations are associated with wall rock alteration. Therefore alteration patterns can guide investigations towards potential prosperous regions particular in area with hidden mineralizations. This is applicable in the Bergslagen region in which volcanosedimentary successions hosts numerous base metal sulphides and iron oxide mineralizations.
5.2 Barium feldspar

Feldspar is the most common mineral group in the Earth’s crust. It consists of two main sub-groups: Plagioclase and K-feldspar. The different element positions in the feldspars have different sizes and bonding configurations. The small tetrahedral position is generally filled with small ions such as Si$^{4+}$ and Al$^{3+}$ whereas the larger position (here called position R) can commonly fit K$^+$, Na$^+$ and Ca$^{2+}$. Ions in position R can be substituted by other elements from the alkali and alkali earth metals group e.g. Ca$^{2+}$ is substituted by Ba$^{2+}$. Feldspars with more than 2% BaO are called Ba-feldspars and denoted C$_{n_{0-100}}$ after its end member celsian, and are structurally very similar to K-feldspars (Deer et al., 2001). This mineral is named after the Swedish scientist and astronomer Anders Celsius and was first described from the Långban-type manganese deposit in Jakobsberg, in the western part of the Bergslagen region by L. J. Igelström in 1867. Ba-feldspar is in turn divided into hyalophane with C$_{n_{0-40}}$ and celsian with C$_{n_{40-100}}$.

Barium feldspar is quite rare and is typically found in metasedimentary manganese deposits, metamorphosed black shales and contact metamorphosed rocks (Deer et al., 2001). Ba-feldspar was identified in the upper member of the Dannemora Formation through electron microprobe analyses. The 23 analysed grains yielded an average composition of Ab$_3$Or$_{81}$Cn$_{16}$ i.e. hyalophane.

5.3 Isotope systems and petrogenesis

Radioactive elements have an unstable configuration of neutrons and protons in their nucleus, and with time they decay into stable radiogenic isotopes. The radioactive isotope is called parent, which decays into daughter(s) isotope(s) that can be used as a petrogenetic tool.

Isotope systems such as Rb-Sr and Sm-Nd are often used in petrogenesis. The former pair belongs to the LILE group and the latter to the Rare Earth Elements (REE) group. The LILE group are known to be mobile during secondary processes such as alteration and metamorphism, whereas REE’s are generally immobile.

Partial melting of the mantle affects the two isotope systems Rb-Sr and Sm-Nd differently. Rubidium is more incompatible than Sr, thus the Rb/Sr-ratio will therefore increase in the melt, whereas the slightly lower incompatibility of Nd to Sm leads to a decrease of the Sm/Nd-ratio.
Both Nd and Sm are trace elements and depending on rock type, the amount of Nd ranges between 10 to 100 ppm and Sm 1 to 40 ppm.

The radioactive $^{147}$Sm decays to the radiogenic $^{143}$Nd. As the amount of radiogenic $^{143}$Nd increases over time, the stable $^{144}$Nd, which remains constant, is used as a reference isotope. The initial $^{143}$Nd/$^{144}$Nd-ratio provides information on the origin of the magma. Negative $^{143}$Nd/$^{144}$Nd-ratio suggests an origin from the crust and a positive ratio indicates mantle origin.

Chondritic Uniform Reservoir referred to as CHUR has been used to determine the composition of the primitive mantle. The CHUR-line reflects the evolution of the $^{143}$Nd/$^{144}$Nd-ratio of the primitive mantle with time (Fig. 12A). Partial melting of CHUR results in two evolution lines; one for the depleted residue and one for the enriched melt (Fig. 12A). As the Sm/Nd-ratio in the residue is higher than in the melt, the evolution line of the residue will be steeper than in the melt (Fig. 12A).

Since the variation of the $^{143}$Nd/$^{144}$Nd-ratio usually is very small, the connotation $\varepsilon_{\text{Nd}}$ is used:

$$\varepsilon_{\text{Nd}} = \left( \frac{^{143}\text{Nd} / ^{144}\text{Nd}}{^{143}\text{Nd} / ^{144}\text{Nd}_{\text{CHUR}}} - 1 \right) \times 10,000$$

By definition CHUR has $\varepsilon_{\text{Nd}} = 0$, whereas today’s values for the crust lies between -10 and -20 and today’s MORB has an $\varepsilon_{\text{Nd}}$-value of about +8 (Fig. 12B; DePaolo and Wasserburg, 1979). Magmas formed from depleted mantle have positive $\varepsilon$-values whereas magmas formed from enriched mantle have negative $\varepsilon$-values due to its lower Sm/Nd than the CHUR.

As CHUR probably only exist in the lower mantle and therefore the depleted mantle (DM) of DePaolo (1981) is used as a reference. Due to its depleted character, DM plot above the CHUR-line.

The $\varepsilon_{\text{Nd}}$-values are age dependent, i.e. $\varepsilon_{\text{Nd}}$ (present) or measured and $\varepsilon_{\text{Nd}}$ (initial)-values or $\varepsilon_{\text{Nd}}$ (i)-values are different and they reflect the $^{143}$Nd/$^{144}$Nd composition in the rock when it was formed provided that the age of the rock is known. As the development curves are known for DM or CHUR model ages can be calculated that indicate when the crustal residence time for the source magma.

If $\varepsilon_{\text{Nd}}$(i) falls on the DM-line, the magma originates from a depleted mantle. When the $\varepsilon_{\text{Nd}}$(i) plot between the DM-curve and the CHUR-line; i) the source could be more primitive than DM but not as primitive as CHUR, ii) the source was a mixture of DM and old crustal material, or iii) partial melting of relatively young oceanic crust such as MORB with a short crustal residence time. Negative $\varepsilon_{\text{Nd}}$(i) is an effect of contamination older crust. The
Sm-Nd isotopic system was applied on mafic dykes in the Dannemora area to derive the magmatic evolution before emplacement (Paper II).

Fig. 12. A) Plot showing the development line for CHUR and the development lines of a partial melt and the residue. As the Sm/Nd is higher for the residue than the melt, the former will always have higher $^{143}$Nd/$^{144}$Nd-ratio than CHUR. B) Plot displaying the DM and CHUR model ages. The measured, $\epsilon_{\text{Nd}}$ is -12 in the analysed rock. From high precision dating the age of the rock is 1.75 Ga and $\epsilon_{\text{Nd}(i)}$ is calculated to +1.6. From these two points the extrapolated line the values of $T_{\text{CHUR}}$ and $T_{\text{DM}}$ are defined. The age difference between the two model ages is called the crustal residence time.
Based on Reynolds (1997), the following section is a short presentation of the basic theory the reflection seismic method.

For a seismic survey a shock wave is created by a hydraulic hammer or explosives that induce a compression wave in the country rock, a P-wave. If the induced stress is below the yield strength of the country rock, it responds in an elastic manner to the applied stress. The elastic response depends on the acoustic impedance of the country rocks, which is the product of density and seismic velocity.

A P-wave front propagates until it encounters a contrast in acoustic impedance. At an interface e.g. rock contact or a fault surface, the wave front splits into a reflected and a transmitted wave front, where the angle of incoming and reflected wave is the same. The transmitted wave front, on the other hand, leaves the interface at a different angle, depending on the difference in acoustic impedance between the geological interfaces such as lithological contacts or deformation zones. In a reflection seismic survey, geophones at the surface are used to register the arrival time and amplitude of the reflected wave. The depth to the reflecting interface i.e. the reflector can be deduced from an estimation or measurement of the P-wave velocity in the country rock. If the seismic wave impinges upon truncated reflectors or small objects relative to the size of the seismic wave length, it results in a diffracted wave i.e. radial scattering of a seismic wave front.

Below some of the terms used in paper III are explained. Two-way travel-time (TWTT) is the time required for an induced seismic wavefront to propagate from the shot points, down and to the geophones. As the ground surface is irregular, the geophones are located at different elevations relative to the reflector, which affect the TWTT. The correction to account for the topography results in a reference line (elevation datum), to which all geophone and shot points are transferred.

Stacking is the process of adding numerous seismograms of common depth point (CDP) to a comprehensive, zero-offset section. Offset is the distance from a shot point to any geophone. For a horizontal reflector, travel time increases with increasing offset. In the case of a dipping reflector, trav-
el time increases down dip and decreases up dip, compared to a horizontal reflector. Dipping reflectors give rise to varying TWTT, therefore data migration is applied to minimize this effect and to restore geometric relationships of the reflector. Two corrections can be applied depending on the orientation of the reflector: normal moveout (for horizontal reflectors) and dip moveout (for dipping reflectors), the latter also improves the lateral resolution.

A 25 km long reflection seismic profile crossing the Dannemora inlier was performed in 2010. The main objective of this investigation was to image the upper three km of the crust to provide information about the geometry of both the eastern and western boundaries of the Dannemora inlier and to possibly define iron ore deposits at depth beneath the Dannemora mine below the Dannemora mine.

The equipment used for this survey was the recording system SERCEL 408UL of Uppsala University with a capability to use 400 channels. In general, geophone spacing was, 20 m. To increase resolution, geophone spacing was reduced to 10 m over the Dannemora mine. With a hydraulic hammer as source, shot points were produced every 40 m along the profile. In this survey, strong reflectors could be expected from amphibolitic rocks, contacts between plutonic and volcanic rocks and ductile and brittle deformation zones (Juhlin and Stephens, 2006). The amphibolitic rock bodies recorded at the surface in the Dannemora area in general are too small to reflect a seismic wave if occurring at depth. Such bodies instead produce diffraction patterns.
7. Summary of papers

The following section summarizes the three papers and the manuscript that are included in this thesis. Paper I deals with the sedimentological and volcanological aspect of the rocks in the Dannemora area. The origin of dykes in the Dannemora area is treated in Paper II. Paper III is about the reflection seismic survey and the manuscript (referred to as paper IV) is about the alteration of the volcanic rocks.

7.1 Dannemora Formation and its alteration (Paper I and IV)

The sedimentological and stratigraphical conditions of the Dannemora Formation, which is a volcanosedimentary succession, have been established and new volcanic structures and textures have been described and interpreted in a geological context (Paper I). The Formation measures about 700-800 m and constitutes an upper and a lower member, of which the latter is subdivided into two sub-units. On the basis of immobile element ratios and stratigraphic position, the succession is further divided in eight sections A-H (Paper IV). The great thickness and extensive alteration, combined with erosion channels and water escape structures suggest that the volcanic material was deposited subaqueous in a caldera (Fig. 12A-B; Paper I and IV). In addition, the size of accretionary lapilli, an evidence for fall or surge deposits, indicates subaerial eruption.

Strong hydrothermal alteration is evident from the sericite dominated matrix with complete sericite replacement of plagioclases and juvenile clasts (Paper IV). In the lower part of the stratigraphy, sections A and C show different alteration patterns. This might reflect that the former was strongly altered already before the latter was deposited. The occurrence of hyalophane indicates significant enrichment of Ba in the upper member.

The rocks are generally rhyolitic to dacitic in composition and contain textures such as pumice clasts, fragmented crystals and Y-shaped former glass shards, indicating a pyroclastic origin (Paper I; IV). The reworked low-
er member contains welding textures such as spherulites that have been found in both primary clasts (Fig. 13A) and in lithic clasts (Fig. 13B), which indicate that this part of the deposit is partly welded (Paper I). In places a partly porous deposit is also suggested by scattered LREE data that probably is a result of their fluid mobility and high water/rock ratio (Paper IV; Bau, 1991).

![Fig. 12. A) Erosion channel in the volcaniclastic rocks. The solid lines highlight the channel and the broken line shows the bedding. B) Water-escape structure where the disrupted layer is highlighted with a solid line and the undisturbed layers with broken lines. Photo of horizontal outcrops from the Dannemora area.](image)

The flattening of pumice clasts in the upper member is ascribed to compaction from the overburden, which is evident in the wedges-shaped fiamme as feathery edged fiamme would have been produced if the compaction was due to welding (Gifkins et al., 2005b). Absence of welding in this member is also indicated by the presence of sericite replaced glass shards with preserved bubble-wall shapes (Paper IV) in contrast to spherulites in the lower member (e.g. Fig. 6A; Fig. 13A) indicating hotter deposition.

In the Dannemora area two synclines have been identified; the Bennbo and Dannemora synclines, respectively. The cores of the synclines expose the uppermost part of the stratigraphy. In the Bennbo syncline, erosion features such as erosion channels (Fig. 12A) together with cross bedding, suggest that the deposition was approximately at the wave base (Paper I). Planar bedding, normal grading and fluid-escape structures (Fig. 12B) are common in the Dannemora syncline and indicate that the deposition was submarine below wave base and thus reflects a different volcanosedimentary facies.
Drill cores from two boreholes that penetrate the whole stratigraphy have been logged. The recognition of an anticline to the west of the Dannemora syncline was based on the correlation of element ratios and the occurrence of a stratigraphic key bed (Paper IV). This implies that the two boreholes have their respective starting point in the western limb of the anticline.

7.2 Dykes (Paper II)

Calc-alkaline, sub-alkaline and basaltic dykes are very common in the Dannemora area, both within the supracrustal inlier and in the surrounding intrusive rocks (Paper II). They have an inferred age of 1.87 Ga. The geochemical signature of the dykes shows that they have a mixed subduction and within-plate affinity. The $\varepsilon_{Nd}(i)$-values of the dykes range between +0.4 and +1.6 indicating a “mildly depleted” mantle source (cf. Fig. 12B).

Combination of element concentration and isotope data indicate that they are a result of mixing of at least three magma components with similar major element compositions. Concentrations of incompatible elements were mainly controlled by mixing, whereas concentrations of compatible elements were governed by fractionation. The basaltic composition of the dykes indicates insignificant contamination from continental crust.

Three source magmas have been inferred based on their modelled concentrations of incompatible trace elements and slightly different isotope characteristics in the resulting dykes. The inferred magma one and two have similar $\varepsilon_{Nd}(i)$-value of +0.5 whereas magma three has a higher $\varepsilon_{Nd}(i)$-value of +1.6. This difference is quite small but larger than the analytical uncertainty of ± 0.5 $\varepsilon$-units. The inferred magma one is most enriched in REE and HFSE, among the three, magma two is strongly enriched in LILE’s except
Sr and less enriched in REE and HFSE than the magma one. The rocks generated from third magma have a flatter REE pattern than the rocks originating from the other two magmas. As magma one is enriched in elements that are only mobile in melts, such as HFSE, it was probably derived from melt-enriched mantle, whereas magma two is enriched in fluid enriched elements such as LILE’s. A comparison with published data from surrounding areas in the Bergslagen region, Roslagen (Johansson et al., 2012) and Avesta-Östhammar (Johansson and Hålenius, 2013) shows that magma one resemble the source for the Roslagen rocks and that magma three is similar to the source for the latter.

7.3 Reflection seismic survey (Paper III)

A seismic profile was produced across the Dannemora inlier (Paper III). A prominent steep reflector found near the Dannemora mine extends down to a depth of more than two kilometres. This reflector might represent either a fluid-bearing fault zone or a deep-seated iron deposit.

The survey also imaged a wide package of east-dipping reflectors east of the Dannemora inlier that extend down to about three kilometres. The reflectors are associated with a ductile-brittle deformation zone with east-side-up kinematics. The total vertical displacement is estimated to be about 2.5 kilometres. Malehmir et al. (2013) interpreted this strong reflector to represent a deformation structure with < 10 % anisotropy combined with the presence of amphibolitic rocks.
8. Sammanfattning på svenska

Dannemora malmfält är beläget i den nordöstra delen av Bergslagen. Gruvdrift har pågått i Dannemora så långt tillbaka som på 1400-talet (Tegengren, 1924). De metamorfa bergarterna, som är bland de bäst bevarade i Bergslagen, har mineraliseringar som ligger koncentrerade till de översta delarna av stratigrafin i en över 80 meter tjock sekvens av dolomitisk marmor, skarn och vulkaniklastiska avsättningar (Lager, 2001; Dahlin et al., 2012). I de sistnämnda påträffas sedimentära strukturer som korsskitning, erosionskanaler, vattenflyktstrukturer men även primära vulkaniska texturer såsom pisoliter, sfäruliter och pimpstenar (artikel I).


Området genomslås av en stor mängd metamorfa diabasgångar (artikel II). Genom analys av geokemin och isotoper från dessa gångar kan den tektoniska miljön fastslås till subduktionsrelaterad och att de till en mindre del har en intra-kontinentalt signatur. Gångarna härstammar från en blandning av minst tre olika ur sprungsmagmor. Om gångarna är av samma generation som de i det närliggande Forsmarksområdet intruderade för mellan 1,87 till 1,86 miljarder år sedan, vilket innebär att de är yngre än både första deformationsfasen och den metamorfa kulmen (Hermansson et al., 2008).

Genom geokemiska analyser på de metavulkaniska bergarterna fastställdes deras omvandlingsgrad (artikel IV). En antiklinal har identifierats där både sektioner med specifika elementkvoter och en bådd med pisoliter är dubblerad och därmed förekommer i båda veckbenen.

En seismisk undersökning har utförts med en profil över Dannemoraområdet (artikel III). Resultaten visar att malmkropparna kan nå ett djup på drygt två kilometer. En annan bred reflektor, som stupar 50° åt öster och kan följas ner till tre kilometers djup, är troligen kopplad till en stor plastisk och senare sprött reaktiviserad deformationszon med en kinematik som tyder på att östra blocket har rört sig tre kilometer upptåt i förhållande till det västra.
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