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Geological Investigations of the southeastern Part of the Repparfjord Tectonic Window

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Abstract

The Repparfjord Tectonic Window(RTW) comprises a package of supracrustal metavolcanic and metasedimentary rocks. This thesis shows the results of structural geological and bedrock mapping of the southeastern part of the window. The mapping reveals a new stratigraphical order of the lowermost formations of the window. The metasedimentary Saltvatn group is evidentaly the lowermost unit, forming the core of a major anticline, affecting the whole window. The Saltvatn Group is followed by the highly diverse, metavolcanic and metasedimentary Holmvatn Group in the south and the metavolcanic Nussir Group in the north.

Two structure geological features, the Skifergangen Shear Zone(SSZ) and the Bratthammer Vein are both kinematical and geometrical compatible with a phase of NW-SE shortening. They both also show minor copper mineralizations associated with quartz/carbonate veins.

The Bratthammer vein shows a top to the NW thrust movement along the vein. Within the thrust plane, fault gouge with neocrystalized illite has been sampled and dated using the K-Ar dating method. Ages yielded from the analysis shows that a faulting event must have taken place at the oldest ~460 Ma., corresponding to the Caledonian Orogeny.

Preface

This thesis is written as a part of my Master of Science degree in Bedrock and Resource Geology at the Department of Geology and Mineral Resources Engineering at the Norwegian University of Science and Technology (NTNU). The work has also been a part of the project "Mineral resources in Northern Norway" (MINN) at the Geolocial Survey of Norway (NGU).

I would like to express my gratitude to my supervisor, Professor Giulio Viola for letting me do this project and for his advice, feedback and discussions throughout the master period. It has been of great value for me.

A special thanks to Espen Torgersen for the great discussions, help and companionship during our excellent period of field work in Repparfjorden. And for his support, big smile and positive attitude for the last two years.

A big thanks to my fellow students for good times and great friendships during the last five years of study.

Last but not least, I would like to thank my family for their general support and love the last 25 years.

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1 Introduction

The Repparfjord Tectonic Window (RTW) is located in the western part of Finnmark, northern Norway (Figure 1). It has previously been known as the Komagfjord Tectonic Window (Reitan, 1963) and Repparfjord-Komagfjord (Pharaoh et al. 1983). It represents a large basement culmination of Precambrian rocks within the Caledonides.

In the window large volumes of supracrustal deposits are exposed. These are predominantly sequences of clastic sedimentary and metavolcanic deposits of Paleoproterozoic age (Pharaoh et al. 1983). Two minor intrusive suites of ultra mafic to felsic composition also crop out in the RTW and cross cut the supracrustals. Above the Paleoproterozoic sequences an autochthonous Neoproterozoic sedimentary unit is well preserved.

The RTW hosts several copper deposits. The largest one is the Nussir deposit, a stratiform copper deposit spacially limited to a thin package of dolostone and its immediate footwall and hanging wall. Estimates of the deposit are currently at more than 30 million tons of copper ore (www.nussir.no).

In addition to the Nussir deposit, the Ulveryggen copper deposit is planed mined in connection with the Nussir project (Nussir.no). This is a large sandstone hosted deposit, previously mined in the 1970s. Elsewhere in the window, there are several smaller copper deposits, many of them mined in the early 1900s. Many of them are found in association with quartz-carbonate veins, which seems to be kinematically compatible and formed under a phase of NW-SE shortening(Torgersen et al. 2013).

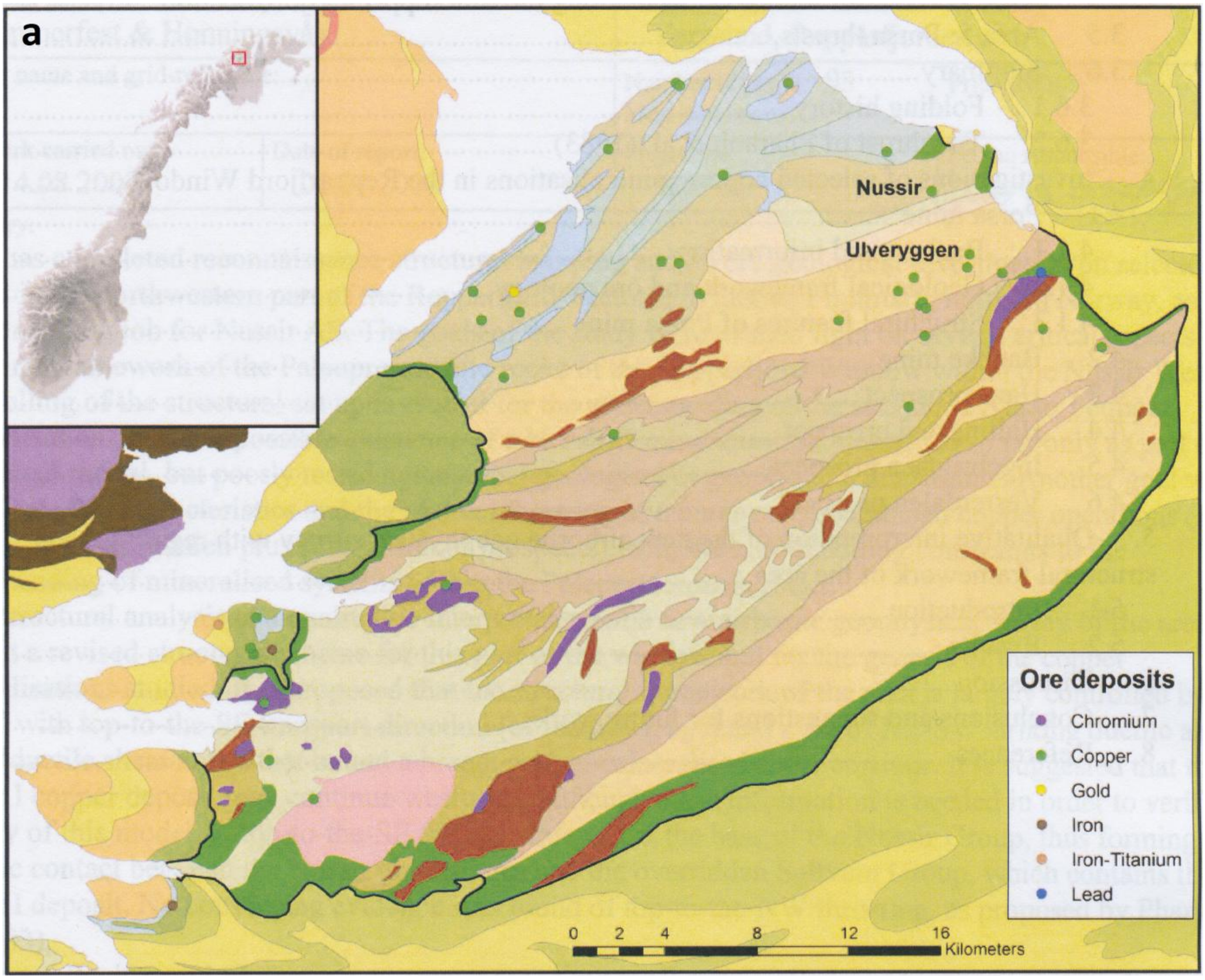


Figure 1 Location of the Repparfjord Tectonic Window in Northern Norway (a) and sketch geological map. The boundary of the window is indicated by the black line. Known ore deposits are shown (from Viola et al. 2008).

1.1 Aim of the study

This study aims primarily to create a better understanding of the geological evolution of the RTW through detailed geological mapping of key areas within the central eastern part of the window, and through detailed structural, geochronological and geochemical investigations.

Two structural geological features in the study area were paid particular attention to in order to shed light on the details of the local tectonic evolution. Their location is shown in Figure 2. The first feature is the Skifergangen shear zone, a dextral NE-SW striking ductile shear zone that deforms a package of metavolcanic rocks. The Skifergangen shear zone was mapped in detail for the first time and the results of the structural investigations and their significance is presented. Very minor copper mineralisations are associated with the shear zones and these are discussed below.

The second feature is a localised brittle-ductile thrust fault with an overall top-to-the-NW sense of shear, only 50 meters away from the Skifergangen shear zone. From the thrust fault, named the Bratthammer thrust, fault gouge was sampled for K-Ar analysis of synkinematic authigenic illite. The obtained results provide an indication of the age of at least the last faulting increment accommodated along the thrust. As in the case of the Skifergangen shear zone, the Brathammer thrust is associated with minor suphide mineralizations, whose metallogenesis will be discussed in the light of the new structural framework defined in this study.

In addition the mapping focused specifically on revealing the stratigraphy of the area. This resulted in a possible new stratigraphical order of the lower part of the Paleoproterozoic stratigraphy exposed within the window.

Geochemical major element analysis were carried out on a suite of selected metavolcanites and gabbroic dykes and compared with previous geochemical work in order to refine our understanding of the stratigraphy.

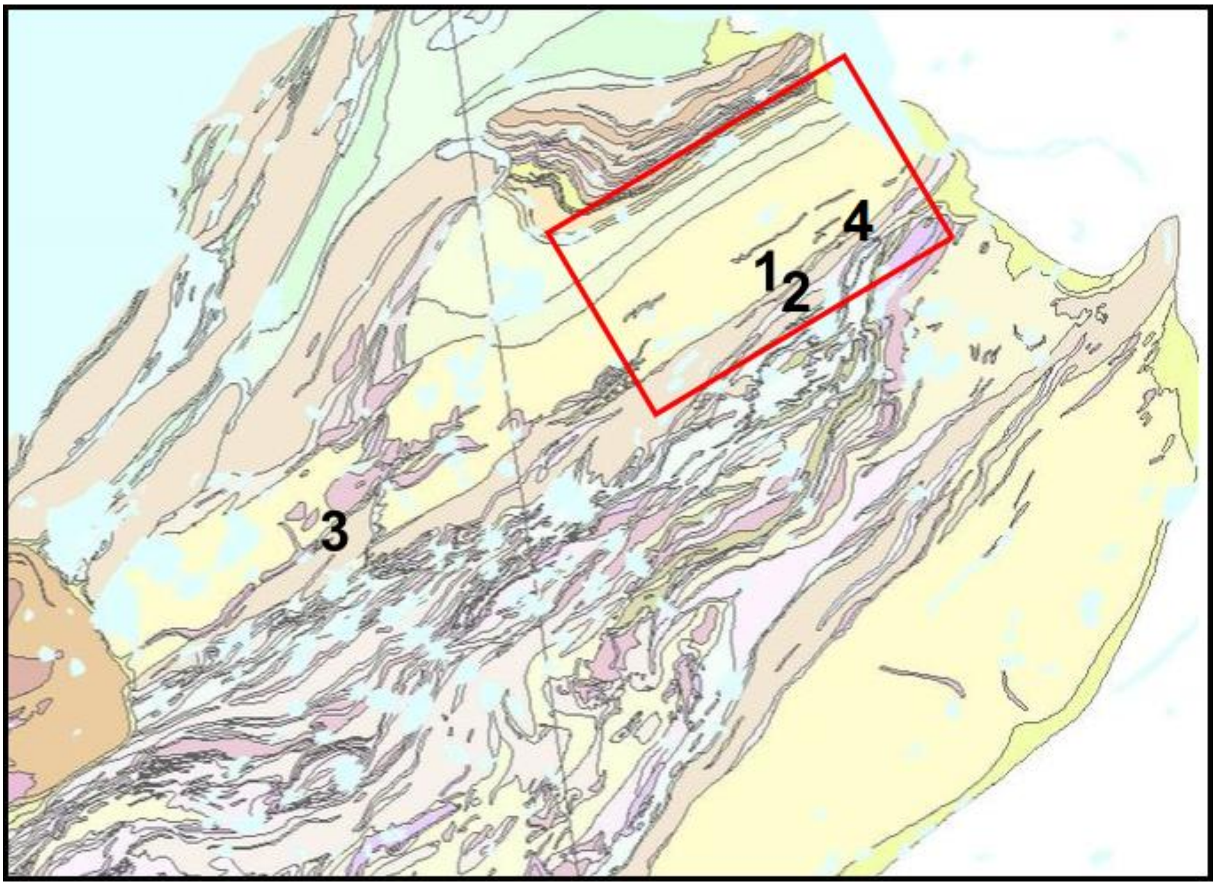


Figure 2 Key localities within RTW. The red square marks the area of the new map produced. 1: Bratthammer Vein deposit, 2: Skifergangen shear zone, 3: Key locality for the stratigraphical order, 4: Location of the apparent interfingering between the Saltvatn and Holmvatn Groups.

1.2 Previous work in the Repparfjord Tectonic Window

The first detailed mapping of the window was carried out by Reitan (1963). He subdivided the metasupracrustal rocks of the window into two groups, the Repparfjord Group and the Saltvann Group(Figure 3)

According to Reitan (op. cit.), the Repparfjord Group consists of metavolcanic and metasedimentary rocks of various origin. The lovermost formation in the Repparfjord Group is the metavolcanic Holmvann Formation, made up by a sequence of various metalavas, -tuffites and sandstones. Unconformably above it, in the east and south east of the window, lays the quartzites of the Doggelv Formation. Slates and sandstones of the Lomvann Formation rest above the Doggelv Formation. In the northern part of the window, the Kvalsund Formation overlies the Holmvann Formation and Reitan (1963) correlated the Kvalsund Formation with the Doggelv and Lomvann Formations.

As part of the Salvann Group, Reitan (1963) included the metasandstones and metaconglomerates of the Steinfjell Formation and the overlaying Djupelv and Fiskevann Formations, which are conglomerates, composed respectively of clasts of greenstone, quartz and jasper, and a dacitic vulcanite. He was not able to determine the age relationship between the two groups, and the contacts are drawn as faults on his maps. However, he proposed the possibility of the clastic sediments in the Djupelv Formation to be derived from the Holmvann Formation, due to their apparent petrographic similarities. In turn, this would imply that the Saltvann Group overlies the Holmvann Formation and the Repparfjord Group.

Pharaoh et al. (1983) reviewed the stratigraphic nomenclature of the RTW and divided Reitan's Repparfjord Group into the Holmvatn Group and the Nussir Group. Figure 3 shows the differences between Pharaoh et al. (1983) and Reitan (1963) stratigraphic subdivision of the Paleoproterozoic sequence of the window. The division of the Repparfjord Group was made on the basis of both field evidence and of the different geochemical signature of the two Holmvatn and Nussir Groups (Pharaoh et al 1983, Pharaoh & Pearce 1984). According to Pharaoh et al. (1983), the Holmvatn group is the lowermost stratigraphic unit of the window, and they proposed clear evidence of this in the area south west of Øvre Saltvatnet(Figure 2, locality 3).

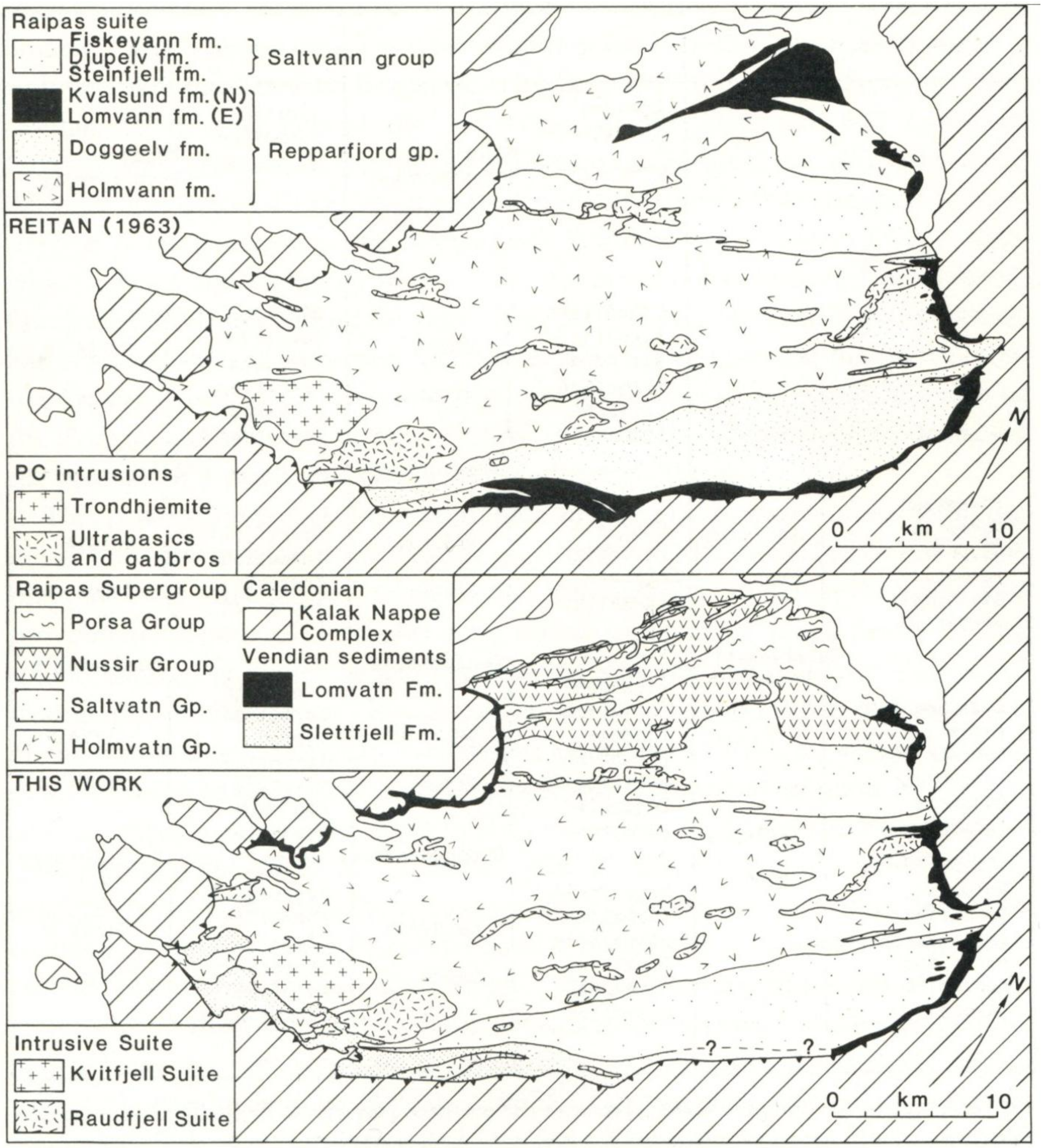


Figure 3 Simplified geological maps showing the differences in the lithostratigraphic schemes of Reitan(1963; upper) and Pharaoh et al. (1983; lower).

Jensen (1996) carried out geochemical studies of the metavolcanites, both in the RTW and the Altenes Window, farther to the SW. His work supports Pharaoh et al (1983) subdivision of the Holmvatn and Nussir Groups, as he differentiated them as belonging to two different volcanic basin sequences. The Holmvatn Group belongs to the oldest of these two basinal sequences and is characterised by a calk-alkaline signature typical of subduction-type magmatism. The

Nussir Group is a part of the younger sequence, and consists of thoelitic metavolcanites, probably related to a rift environment. Jensen (1996) argued that both sequences are probably part of an arc type environment, and most likely formed at the same convergent continental margin.

2 Regional geological framework

The Paleoproterozoic sequence exposed in the RTW crops out from within the Caledonian Kalak Nappe Complex. The window is of approximately 900 km^2 and stretches from Repparfjord in the northeast to Komagfjord in the southwest. Already in the first studies of the window, the similarities to the other basement windows of western Finnmark were pointed out(Reusch 1891).

The supracrustal rock sequences of the window make up a part of what has been called the Raipas Supergroup. The Raipas Supergroup was first recognized and decribed by Dahll (1867) in the Alta area. This term has later been used as a general descrtiptive term for the bedrock in the Altenes, Alta-Kvænangen and Repparfjord Tectonic Windows, and it describes a package of low-grade metamorphic sandstones and greenstones which seems to be fairly similar in all three windows.

Correlations have later been drawn by several authors between the tectonic windows and the supracrustal greenstone belts of Finnmarksvidda and Northern Sweden (Pharaoh et al. 1982, Pharaoh and Pearce 1984, Pharaoh and Brewer 1990, Jensen 1996) based on the lihological similarities between these units. As pointed out by Pharaoh et al. (1983) the correlation of units over such a large distance is difficult due to the lateral facies changes of the rocks. The changes can often be very significant, even over small distances. Depending on the paleoenvironmental model in use, it may even be unwise to try to make detailed correlations.

2.1 The Baltic shield

The supracrustal sequence of the RTW is believed to be of Paleoproterozoic age, and to be related to the Svecofennian domain of the Baltic shield (Pharaoh and Pearce 1984, Brewer and Pharaoh 1990).

The Baltic Shield (also known as the Fennoscandian Shield) was formed between 3.5 and 1.5 Ga, and consists of the Precambrian basement of the Fennoscandian region and northwestern Russia. To the south it stretches down to the Baltic Sea. The shield features a geochronological zonation, wherein the crust youngs from the northeast to the southwest. The shield is generally divided into three domains. The Archean domain is localized in the northeast. The Svecofennian domain in central part, and the Southwest Scandinavian domain in the southwest(Gaal and Gorbatschev 1987, Nironen 1997).

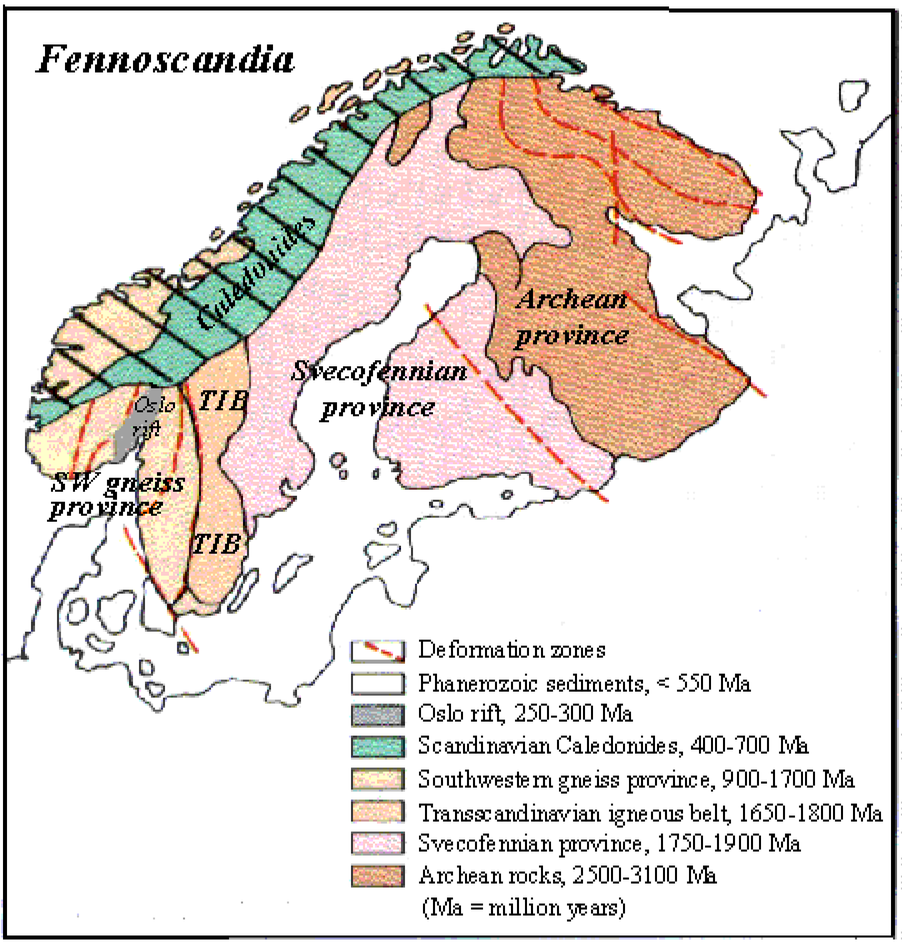


Figure 4: Simplified map of the Baltic Shield. The RTW is a northenrly exposure of the Svevofennian Domain shown. (Nrm.se 2013)

The Svecofennian (also known as Svecokarelian) Orogeny took place between 1900-1750 Ma ago (e.g. Pharaoh & Pearce 1984). Today we find rocks belonging to the Svecofennian Orogen dominating large parts of northern and central Sweden and southwestern Finland. Farther north, in Norway, the Svecofennian basement is present in the Karasjok and Kautokeino Greenstone Belts, as well as in the basement windows of Western Finnmark.

Several different models have been proposed to explain the evolution of the Svecofennian orogen. The general model starts with rifting and opening of a pre-Svecofennian ocean to the south-west of the Archean craton, creating accommodation space for sedimentary and volcanic deposits. Later convergence caused accreation of volcanic arcs onto the Archean craton. According to Nironen(1997) there were probably two different arcs colliding with the continent. The arcs, being parts of one or more microcontinents, moved north-eastward and accreted onto the craton margin at 1.91 Ga. Subduction occurred to the south-west, beneath the arc complex (Nironen 1997). At a late stage, extensive magmatism took place.

2.2 Finnmark basement

The Precambrian basement in Finnmark is exposed in the eastern and western greenstone belts of Finnmarksvidda as well as in the tectonic windows of western Finnmark. As seen in Figure 5, the windows and the basement on Finnmarksvidda can be tentatively connected by following their general trend beneath the Caledonian cover sepeating them.

The Altenes Tectonic Window (ATW) is the basement window closest to the RTW. The two windows are separated by a only 6 km wide cover of Caledonian nappes at the closest location. Jensen (1996) correlated the Holmvatn Group of the RTW with the Brattholmen group of the ATW. The groups lie in a direct continuation of strike, and shows similar lithologies and geochemical signatures. Further he correlated the Saltvatn Group with the Sagelv Group of the ATV, as these groups also are continous in strike and with similar lithologies. In the Alta-Kvænangen Tectonic Window (AKTW), Pharaoh et al. (1983) pointed out the similarities between the lower Raipas Group, with metabasaltic pillow lavas, tuffs and tuffitic sediments, overlaid by argillites and dolomites, and the Nussir and Porsa Groups of the RTW. In the AKTW, there does not seem to be an exposed equivalent of the Saltvatn and Holmvatn Groups.

On Finnmarksvidda the Karelian supracrustals sequence unconformably overlies the Archean basement((Pharaoh and Pearce 1984). The supracrustals are exposed in two greenstone belts, the KarasjokGreenstone Belt in the east, and the Kautokeino Greenstone Belt in the west. The detailed correlations are rather speculative, and will not be discussed in this work.

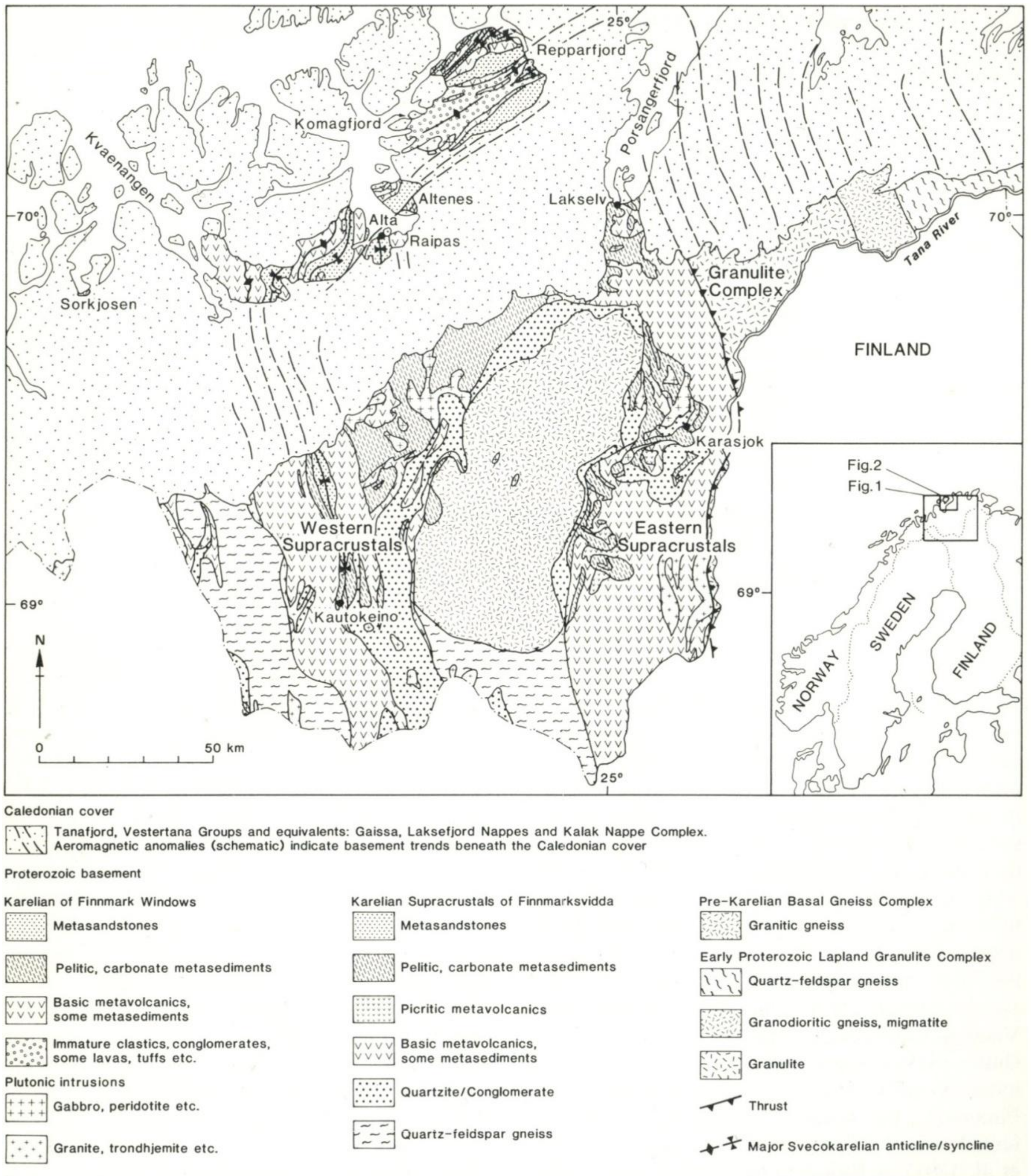


Figure 5 Precambrian basement and Caledonian cover of western Finnmark. The map indicates how the basement trends continue beneath the Caledonian cover from the tectonic windows to the Greenstone Belts of Finnmarksvidda. The map is from Pharaoh et al.(1983).

2.3 The Repparfjord Tectonic Window: the status quo

The stratigraphic order used to describe the lithologies mapped in this study follows the one suggested by Pharaoh et al. (1983). This seems to have been accepted as the official stratigraphy for the window, although, as discussed below, the new mapping presented here questions significantly the established stratigraphy.

The lowermost unit of the RTW is, according to the currently accepted stratigraphy, the highly complex and diversified Holmvatn Group. It consists of metavolcanic and clastic sedimentary deposits of various nature. The Holmvatn Group crops out exstensively in the central part of the window. Nowhere in the window is the basement exposed, and therefore there is no certainty as to what forms the basement to the paleoproterozoic sequence.

Above the Holmvatn group, and outcropping to the north of it, we find the clastic metasedimentary Saltvatn Group, consisting of sandstones, conglomeratic sandstones and conglomerates. The Saltvatn Group is described as resting unconformably above the Holmvatn Group. In the southern part of the window, at the same stratigraphic level as the formations of the Saltvatn Group, we find the Dåg'gejåkka Formation. This is a sequence of quartzitic sandstones, and even though it shows distinct differences to the Saltvatn Group, they are considered to be correlated (Pharaoh et al. 1983).

Conformable above, and to the north of the Saltvatn Group, crops out the Nussir Group. This is a group of metavolcanic rocks, with tufites, massive metabasalts, pillow lavas and lava flow deposits. According to Pharaoh et al. (1983) the total thickness of this group is more than 1700 m. Above the volcanites of the Nussir Group, we find dolomites and shales of the Porsa Group. This is the uppermost group of the Paleoproterozoic secuence of the window.

Overlying the Paleoproterozoic sequence, we find the Lomvatn Formation. The Lomvatn Formation consists of sandstones, silt and shales. The formation rests unconformably upon the Paleoproterozoic sequence, and it is found preserved as a relatively thin sequence along large parts of the margins of the window. According to Pharaoh et al. (1983), the Lomvatn Formation is autochtonus, although it has been severly deformed during the Caledonian orogeny. Based on the similarities with autochtonus foreland rocks of the Caledonian orogen throughout Norway and Sweden, the age is assumed to be of Vendian-Cambrian age (Pharaoh et al. 1983, Pharaoh 1985).

2.3.1 Geochronological constraints in the RTW

Geochronological constraints on the Paleoproterozoic sequence of the RTW are scarse. The rocks are generally believed to be of Paleoproterozoic age, but this assumption is based mostly on comparisons to the other tectonic windows and the greenstone belts farther south, in Finnmarksvidda and Sweden.

One of the very few studies with geochronological constraints on the evolution of the RTW is by Pharaoh et al. (1982). It contains K-Ar ages of hornblendes from the southeastern part of the window. Ages cluster between 1806 and 2085 Ma (Pharaoh et al. 1982) and were interpreted as the date at wich the hornblendes passed through their blocking temperatures(ca 500 °C) following the peak of metamorphism. The ages fit well with the Svecokarelian orogeny, and give a clear indication that the supracrustal sequences of the RTW experienced deformation during the Svecokarelian oregeny. The dating indicates a peak in regional metamorphism around 1840 Ma (Pharaoh et al. 1982). There are, however, no constraints on the older history.

Dallmeyer et al. (1988) did K-Ar and ^{40}Ar-^{39}Ar studies of both the basement sequence and the neoproterozoic cover unit in RTW and in the Alta-Kvænangen window. Their results show that the oldest ages recorded in the window are ~1700-1800 Ma. This is younger than the 1840 Ma of Pharaoh et al. (1983), but is still considered to be a result of Svecokarelian deformation(Dallmeyer et al. 1988).

Rejuvenation of the argon system at about 400-425 Ma. is interpreted as a result of reheating during the Caledonian Orogeny. The Caledonian thermal overprint is variable within the window, and seemingly related to the proximity to the Kalak Thrust and the Caledonian Nappes. The thermal overprint developed during the emplacement of the overlying Caledonian Kalak Nappe Complex, when the temperatures were high enough rejuvenation of the argon system. Distant to the thrust, the temperatures did not exceed the ones required for rejuvenation of the system(~300-350°C for white mica)(Dallmeyer et al. 1988).

2.3.2 Caledonian influence on the RTW

The Repparfjord window is a tectonic window. A tectonic widow is normally defined as an erosional hole through the overthrusted nappe system and into the underlying basement(Fossen and Gabrielsen 2005). The erosion of the overthrusted nappe is caused by doming of the basement, elevating it to a higher level than the surroundings and therefore forcing erosion to happen. The exact cause of the doming in the Repparfjord window is uncertain. The general theory is that the doming occurred in conjunction with the Caledonian orogeny. Possibly there could have already been a pre existing antiform forcing the thrust plates to bend upwards. Townsend (1987) launched the term Komagford antiformal stack to describe the structure forming the doming of the window.

The Caledonian Orogen in Scandinavia is represented by several nappe complexes thrust onto the Baltic shield during the closure of the Iapetus Ocean. The nappes are thrusted and deformed to a varying degree. The Kalak Nappe Complex, surrounding the Repparfjord tectonic window, is a part of the Middle Allochton of the Caledonian thrust belt. It is metamorphosed to greenschist to upper amphibolites facies and imbricated as a result of the SE directed thrusting(Dallmeyer et al. 1988, Rice 1998). As previously described, $^{40}Ar/^{39}Ar$ dating of the neoproterozoic Lomvatn Formation sediments has yielded an age of 400-420 Ma for a major tecthonothermal event. This is interpreted to be the age of the main emplacement of the Kalak Nappe Complex in the area of Repparfjorden((Dallmeyer et al. 1988).

There has been a discussion whether the window itself is a part of the Caledonian thrust sheets. Pharaoh et al. (1983) belived the window is autochtonus. Townsend (1987), Rice (1998) and other authors believed the window is allochtonus, and thrusted during the Caledonian orogeny as a part of the lower allochton. The basis for the allochtonus model is the restoration of thrust sheets. By restoring the major thrusts and imbricates it is suggested that the Repparfjord window has experienced a displacement of more than 375 km from its original location during the Caledonian orogeny. The thrusting direction seems to have been SE and later ESE/E(Townsend 1987)

3 Methods

3.1 Field work

The field work was carried out over two seasons in august 2011 and august 2012. This quite extensive amount of field work gave a great opportunity for data collection, and understanding of field relationships. Much of the work was done in collaboration with Espen Torgersen, PhD candidate at NGU.

The mapping was done with a Panasonic Toughbook borrowed from NGU. This is a convertible tablet computer specially built to withstand rough conditions. The computer is waterproof and shock resistant and has a built in GPS. The software used for the mapping is SIGMA (System for Integrated Geoscience Mapping) from the British Geological Survey (BGS). This software works as an integrated part of the ArcGIS software from ESRI.

Using this system gave great advantages as the mapping is done directly in the ArcGIS system, thus reducing the need of data processing a great deal. All the observation points made are placed directly in the map, and the database of information is already made. With the GIS it is possible to work with several layers of information directly on the screen while in field. It is very easy to switch between the different map layers or even putting several layers on top of each other by making them see-through. This can be topographical data, geophysical maps, borehole data etc. Having all this information available in the field, of course helps in the understanding of the field observations.

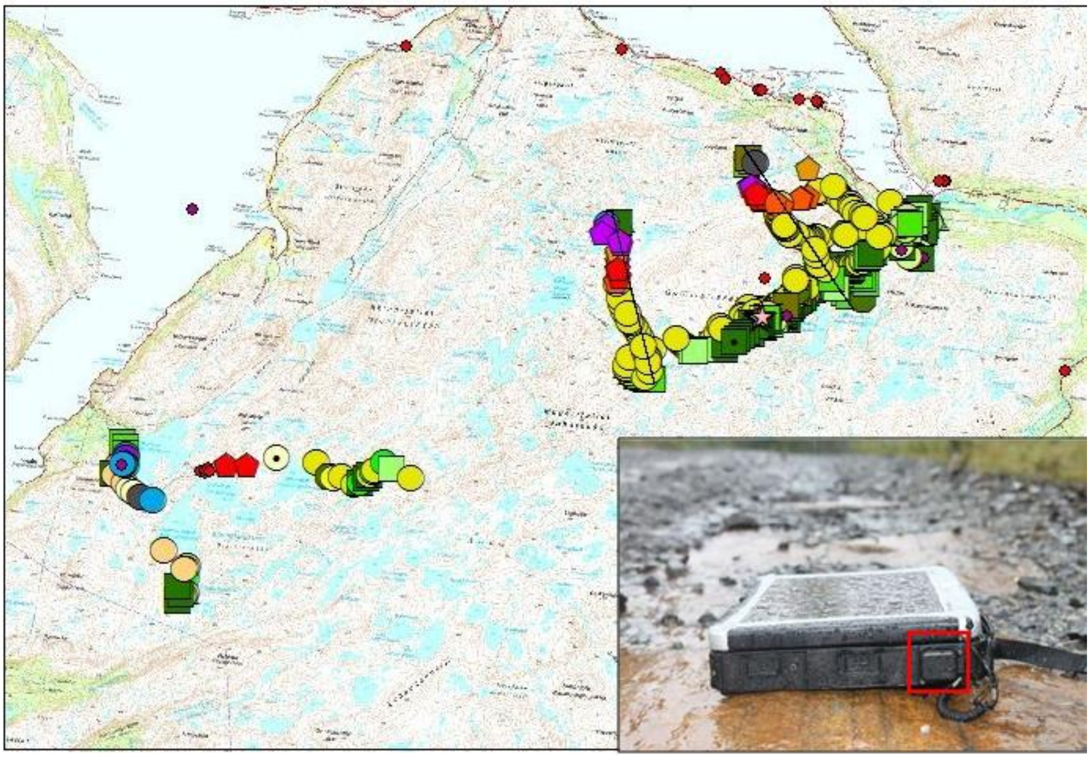


Figure 6 Map of most of the field observation points made during the field work. Different colour codes indicates the lithologies. The photo shows the toughbook. Red square marks the built in GPS(Photo from Henderson and Viola (2011))

NGU provided a great access to various maps. These were topographical maps such as the 1:50000 scale topographic map, high quality orthophotos of the eastern part of the Repparfjord window and various hillshade maps. Of these, the topographic map was by far the most used. Geophysical maps were also provided. A lot of the geophysical data was brand new, as the data was collected as a part of the NGUs MINN program(NGU.no 2013). The data included several different electromagnetic maps, magnetometric and radiometric maps. Old geological maps of the Repparfjord window have been digitalized and georeferenced. These have been the maps of Reitan(1963), (Pharaoh et al. 1983) and (Nilsson and Nilsen 1996).

The field work of the first season focused on getting to know the geology of the window and the regional geology. Detailed studies were performed in the area around Skifergangen shear zone, and samples for thin sections were collected. In the second season the focus was on several different key localities in the window. More detailed studies of the Bratthammer thrust was performed and samples for both thin sections and geochemical analysis was collected.

3.2 Optical microscopy

A total of 30 thin sections for optical microscopy studies were prepared at the thin section laboratory at NTNU by Arild Monsøy and Kjetil Eriksen. The thin sections were standard 30µm thick. Most of the thin sections were made with a polished surface in order to study the appearance of ore minerals. Some unpolished thin sections were also made. Foliated samples were oriented and cut perpendicular to their foliation and parallel to the stretching lineation in order to show the structures of interest.

3.3 K-Ar dating

The method of K-Ar dating fault gouge is an effective way of dating shallow brittle faults. Near-surface deformation of the crust will normally cause the development of brittle faults. By gaining an age of the development of deformational structures, it is possible to build a better understanding of the tectonic evolution of an area. For this study, fault gouge samples have been collected from the Bratthammer Thrust Fault in order to better understand its relation to the tectonics of the window.

Slip on brittle faults often leads to the development of cataclasite and fault gouge composed by crushed rock fragments and new grown, authigenic, synkinematic clay minerals. Particulary does illite often crystallize by retrograde hydration reactions at temperatures around 100-250°C(Viola et al. 2013).

The K-Ar age method is a radiometric method based on the radioactive decay of ^{40}K to ^{40} Ca or ^{40}Ar. ^{40}Ca is a common isotope in many minerals, and it is there for not suitable for radiometric dating. Ar, however, is a noble gas and does not bind with other atoms in the crystal lattice. It can, however, be trapped in many solid phases, and this is crucial for the method. When subjected to sufficient heat, the trapped Ar will escape and thus reset the radiometric clock(Winter 2010).

When new clay minerals, like illite crystallizes during deformation, a new radiometric clock starts, as the K now enters a new mineral phase. All the produced Ar will be trapped within the mineral. If eiter ^{40} K or ^{40}Ar is lost after the formation of illite, during an event of heating or by influence of hydrothermal fluids, the dating may not yield a meaningful age. An important assumption for the method is that no other, older K-bearing phases are present in the dated material, as their inherited ages will affect the dated ages.

The method and its application will be further discussed in chapter 6.4 K-Ar illite age dating, where the method is applied on fault gouge from the Bratthammer thrust.

4 Petrographical descriptions

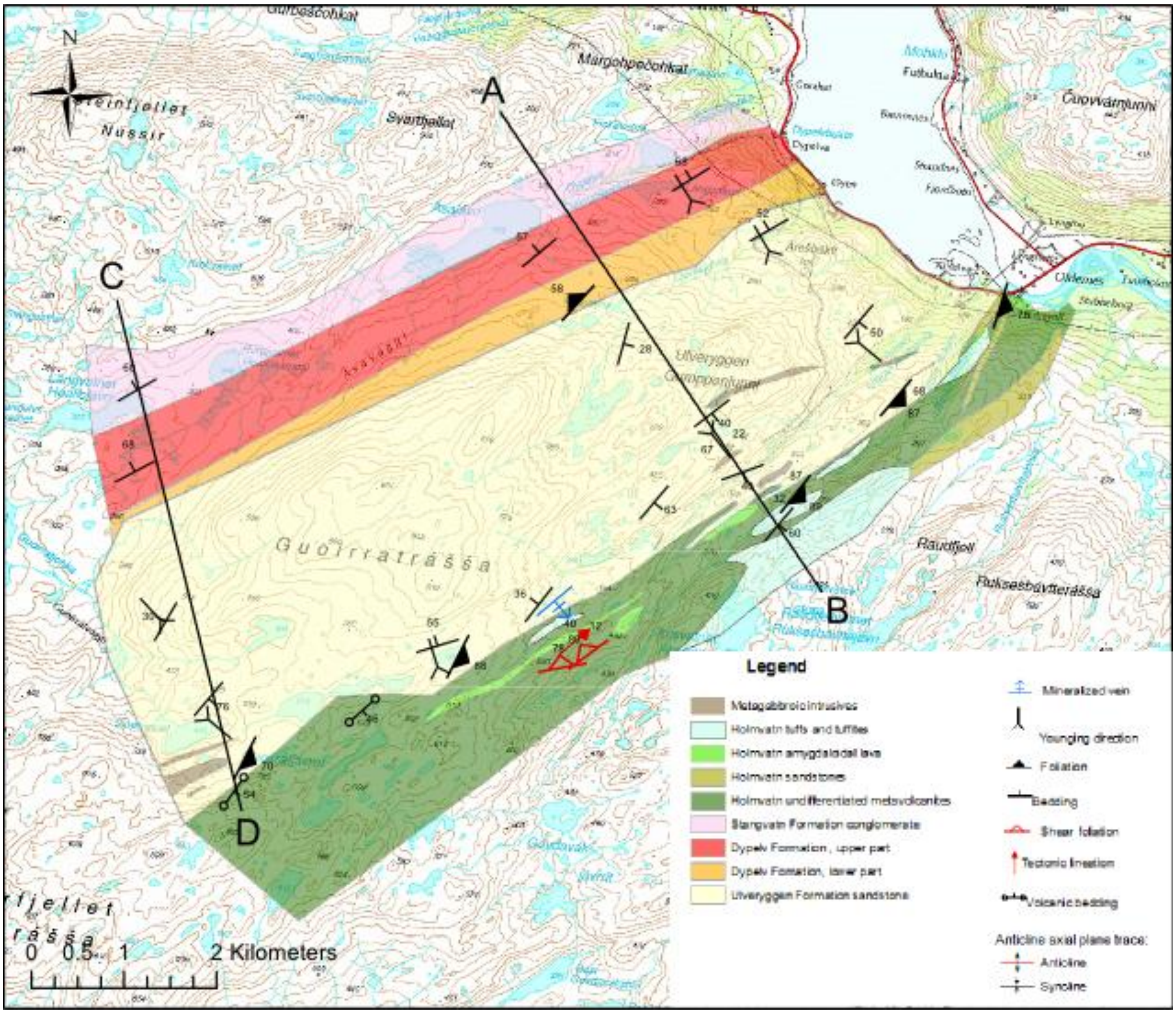


Figure 7: The map of the central eastern part of RTW. Cross section A-B and C-D are found in Error! Reference source not found. **andFigure 16. Larger version of the map is attached**

Field work mapping in the RTW have mainly focused on the central and eastern parts of the window. There crops out the lowermost Groups of the paleoproterozoic supracrustacl rock sequence of the window. In this chapter, field observations and lithological descriptions are presented from the Saltvatn Group and the Holmvatn Groups. Their field relationship is illustrated and the stratigraphy is discussed in a regional framework.

4.1 The Saltvatn Group

The Saltvatn Group was first described by Reitan (1963), as he divided the RTW stratigraphy into two groups: the Saltvatn Group, and the Repparfjord Group. The Saltvatn Group was in turn subdivided into three formations, reviewed and renamed by Pharaoh et al(1983). In stratigraphic order from the lowermost and up: The Ulveryggen Formation, the Dypelv Formaiton and the Stangvatn Formation (Figure 9). The Saltvatn Group crops out in the central part of the window, and is found all the way from the eastern to the western part. Its two upper formations, however, do not outcrop in the western part. The whole Group consists of sedimentary clastic rocks, ranging from fine grained sandstones to very coarse conglomerates. Although the sediments are deformed, sedimentary structures are well preserved, especially in the Ulveryggen Formation. The Group is folded in a major antiform, with its core in the Ulveryggen Formation, as seen in the cross sections in Figure 15 andFigure 16.

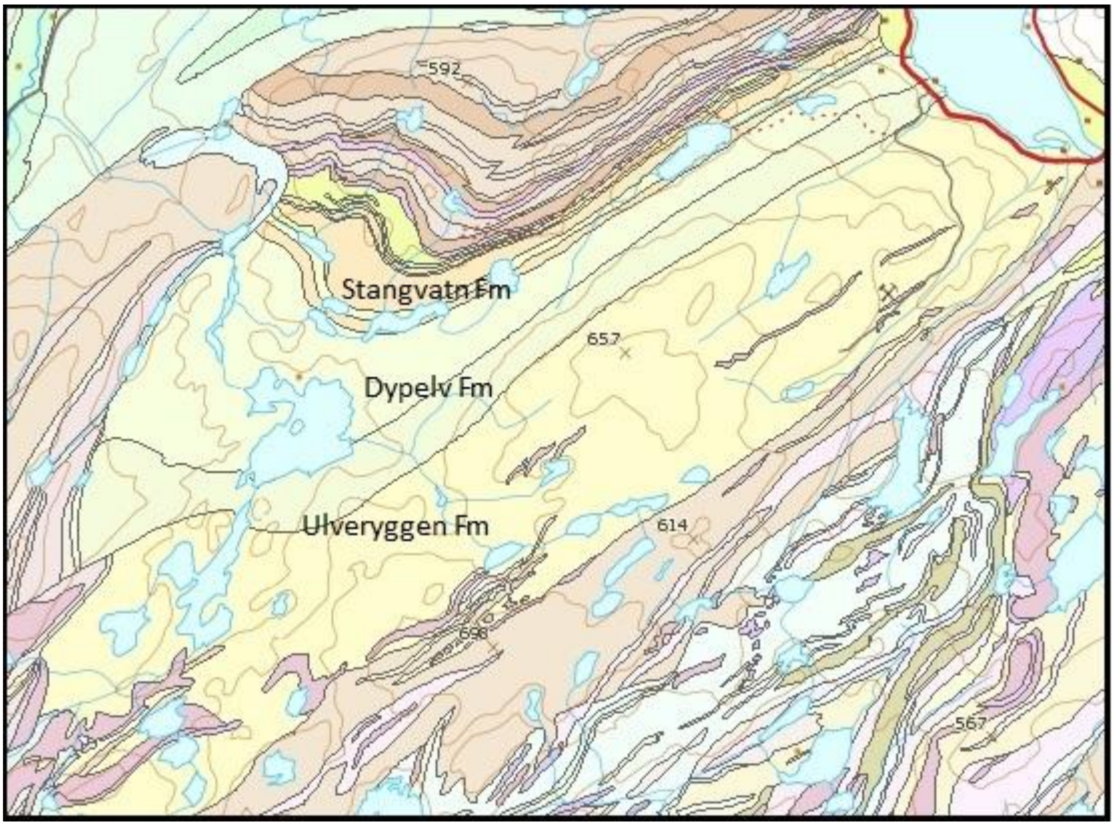


Figure 8: Map of the Saltvatn Group and its three Formations. (Modified map from: http://www.ngu.no/no/hm/Kart-og-data/)

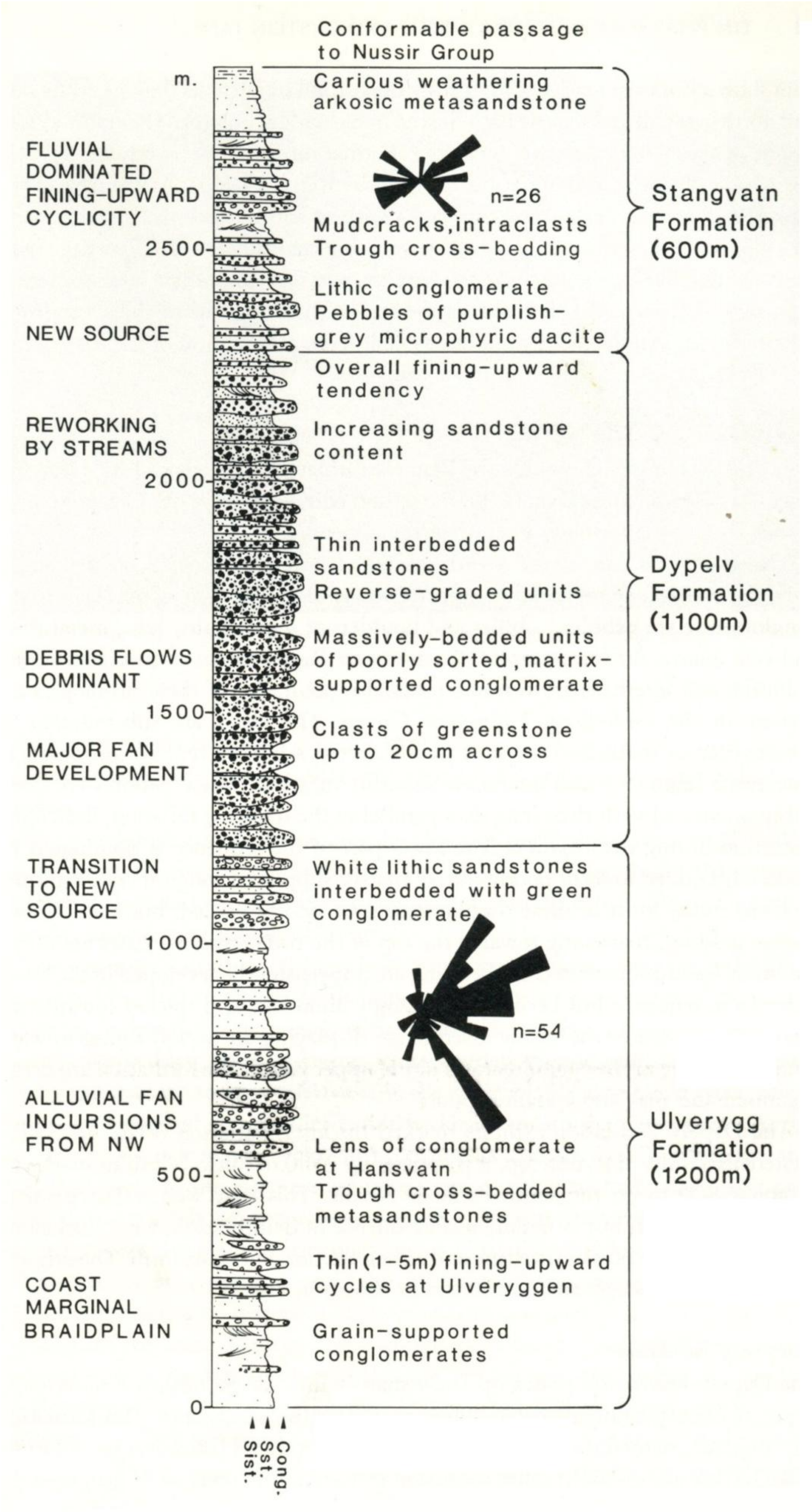


Figure 9: The lithostratigraphy of the Saltvatn Group. Individual beds are not shown to scale. Paleocurrent directions are plotted, and show a bimodal tendency, with deposition by currents flowing from west and north west. Figure from Pharaoh et al(1983)

4.1.1 The Ulveryggen Formation

The lowermost formation of the Saltvatn Group is the Ulveryggen Formation. It consists of coarse quartzitic and feldspathic sandstones and conglomerates. The total thickness of this formation is not known, with its lowermost part not exposed. Based on the exposed sequence, however, the thickness is of at least 1000 m. The primary sedimentary features of the sandstones are generally well preserved, with distinctive cross bedding and fining upwards cycles. The bedding is in many places very distinctive, as the tilted beds makes up rhythmic steps in the landscape(Figure 10A) Cross bedding often appears as trough cross bedding, with beds commonly up to 50 cm thick. Cross bedding, upward fining and a variety of other primary features confirm a normal way up throughout the area, thus excluding tectonic overturning. Channels eroded down into the underlying beds are common (Figure 10B).

Slight variations of the composition of the sandstones can be observed. Some beds are nearly quartzitic, wheras others are more arkosic. The grains are generally well rounded. The grain size varies throughout the formation, with generally more fine grained beds towards the southeast and the contact to the Holmvatn Group. Conglomeratic beds and lenses are very common throughout the entire formation.

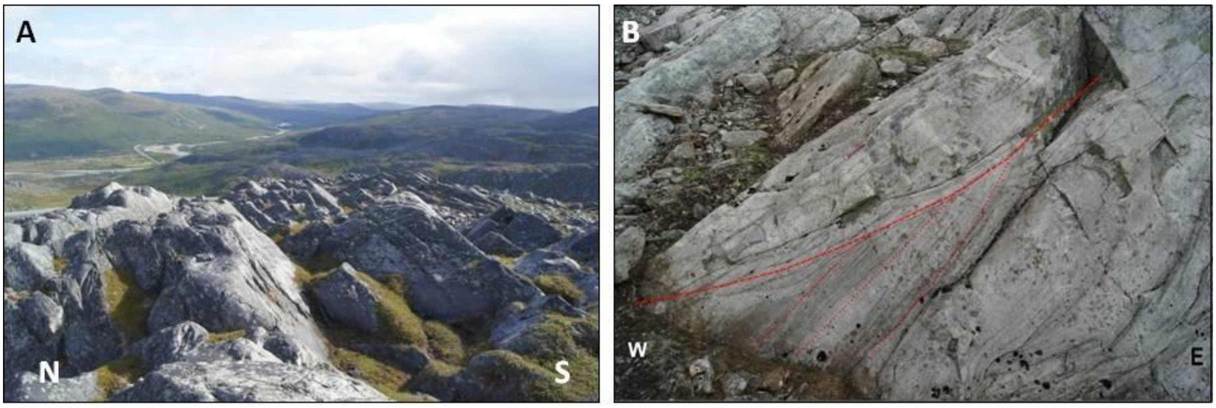


Figure 10 A) (622045.635025, 7818084.874495, UTM 34N) Well defined bedding planes within thickly bedded sandstones of the Ulveryggen Formation. B) (615589.779299, 7813488.000381, UTM 34N) Example of cross bedding within the sandstones of the Ulveryggen Formation. The red line shows a channel cutting into the underlying beds.

4.1.2 The Dypelv Formation

The Dypelv Formation directly overlies the Ulveryggen Formation. The contact between these two formations is gradational with a progressive transition from the Ulveryggen sandstones to the Dypelv conglomerate. The conglomerate consists of centimetric clasts of quartz, jasper and greenstone (Figure 11A+B). From thin section analysis it was established that there are pebbles of two different quartz types. One very fine grained grain size < 0,2 mm) and one coarser grained (grain size between 0,2 and 0,5 mm).. The clasts are generally sub-rounded to sub-angular in shape. The matrix of the conglomerate is composed of fine grained greenstone fragments in addition to quartz and feldspatihc material. Grain size varies from sandy up to 30 cm diameter large boulders. There are thick beds of poorly sorted, poorly rounded and very coarse material, interbedded with finer grained beds. Coarse sandy to conglomeratic channels are also present, with minor carbonatic horizons.

The map shows a lithological variation of the conglomerate, with the lower part richer in quartz and jasper pebbles, whereas the upper part consists mainly of pebbles of greenstone. This variation is, however, not well defined, and beds rich in both quartz and jasper may be found in the upper part of the Dypelv Formation as well. The greenstone-rich upper part of the formation is generally coarser grained, with beds of extremely coarse material (Figure 11C) The overall thickness of the formation is about 1000m, but it shows a significant increase towards west, where the thickness is more than 2500 meters.

To the east, the Dypelv conglomerates appear to be deformed more intensely. Strain is accommodated mostly in the less competent parts of the rock, wherein the weaker matrix appears as strongly foliated and bent around the more competent quartz and jasper pebbles. This can be appreciated in Figure 11B. Rotation of clasts and recrystalization in pressure shadows are commonly observed . The greenstone clasts seem to have taken up a great amount of strain, reducing the grain size, and are at times not easily differentiated from the matrix.

4.1.3 The Stangvatn Formation

Concordantly above the Dypelv Formation lays the Stangvatn Formation. It consists of a polymict conglomerate with clasts of a purple colored, volcanic rock (Figure 11D). According to Pharaoh (1983) the clasts are of dacitic composition. Grain size varies, with clasts of very different size ranging up to 10 cm in diameter. The formation contains several conglomeratic beds interbedded with fine grained, sandy to silty levels. Cross bedding is easily recognised in the finer grained beds. The thickness of the Stangvatn Formation is about 600 m, although it can vary over the strike length.. In the uppermost part of the Stangvatn Formation we find a package of a fine grained, schistose siltstone, overlaid by a dolomite of a few meters thickness. This is the copper mineralized dolomite that hosts the Nussir deposit.

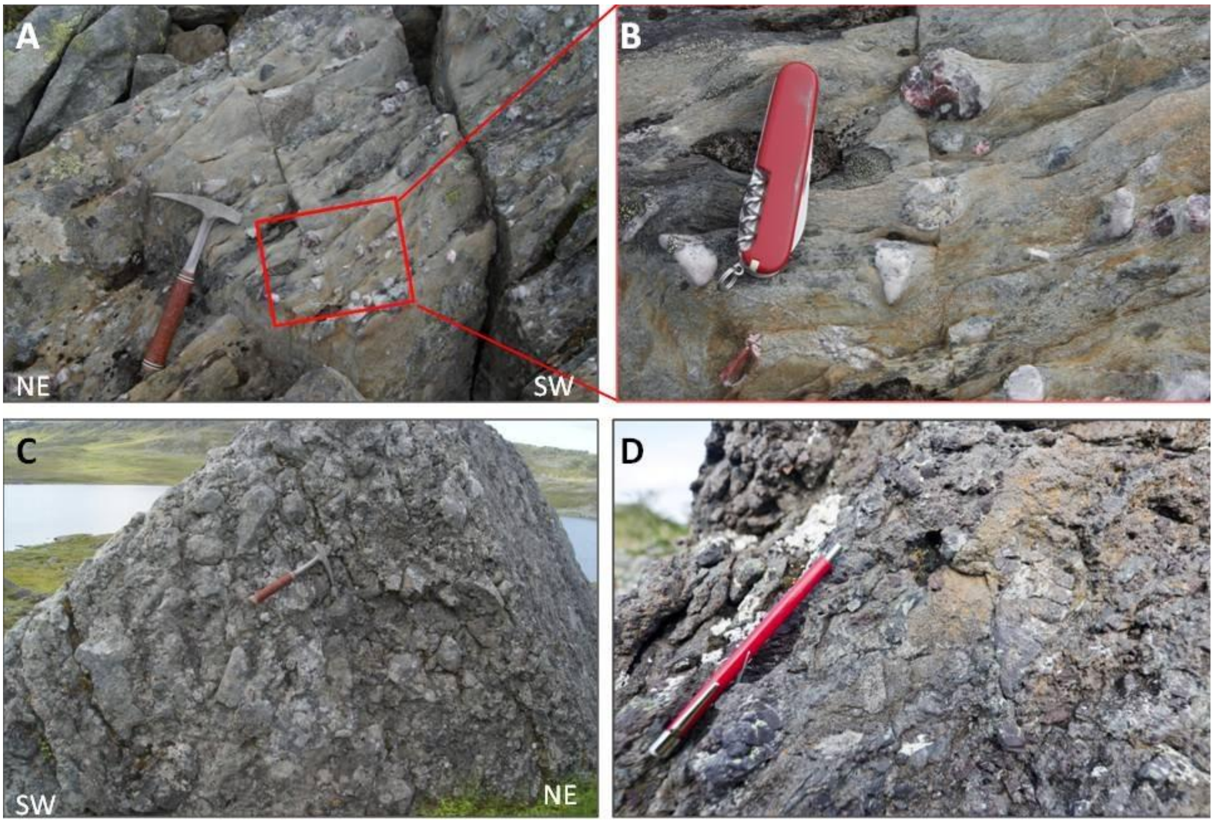
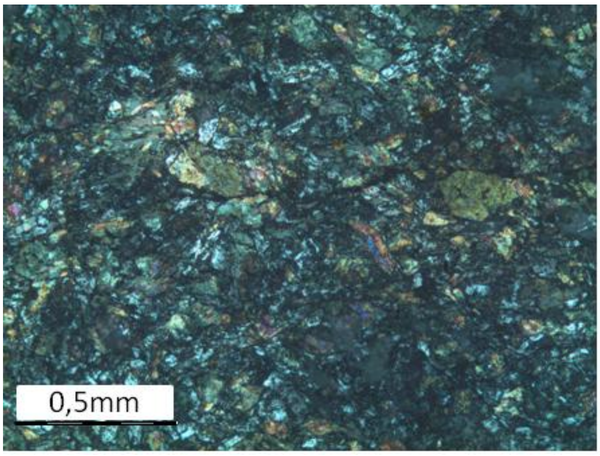


Figure 11 A)(620078.554976, 7818968.704473 UTM 34N) The Dypelv conglomerate with pebbles of quartz, jasper and greenstone within a fine grained matrix. B) Close up of the conglomerate, where it is visible how the deformation is localized to the weaker matrix and bends around the more competent pebbles. C) (619593.505247, 7819201.077134 UTM 34N) Upper part of the Dypelv conglomerate. This part of the conglomerate consists of mainly greenstone clasts. Very coarse grain size and poor sorting indicates short transport. D)(620741.384066, 7820585.023102 UTM 34N) The Stangvatn Formation conglomerate. Pebbles of dacitic composition give the purple colour of the conglomerate (Photo: Espen Torgesen).

4.2 The Holmvatn Group

The Holmvatn Group is a large and very complex package of lithologies. It outcrops extensively in the southern part of the RTW, and makes up almost half of the window's total area. The sequence of sedimentary and metavolcanic rocks is at least 3 km thick (Pharaoh et al. 1983). Pharaoh (1983) subdivided the Group into three different formations. The lowermost is the Markfjell Formation followed up stratigraphy by the Båtdalsev Formation, and the uppermost Magerfjell Formation. According to Pharaoh (1983), the Holmvatn Group forms the core of a major anticline, flanked by the overlying Saltvatn Group to the north and the Doggejohka Formation to the south. Jensen (1996) subdivided the Group further into nine formations, arranged in a homoclinal sequence dipping towards the southeast He argued that no tectonic repetition of strata was observed within the group. Interestingly enough, Jensen (1996) considered the Magerfjell Formation, the closest to the contact towards the younger Saltvatn Group, as the oldest formation in the Holmvatn Group. This requires not only an unconfomity, but also a complete overturning of the stratigraphic sequencebefore the deposition of the overlying Saltvatn Group.

This study has focussed only with the Magerfjell Formation and only in areas in close proximity to the contact to the Saltvatn Group. Generally the Magerfjell Formation consists of metavolcanic rocks. Pillow lavas, amygdoidal lavas and massive lava flows are common. Tuffitic and volcanoclastic sequences occur in between as well as beds of almost quartzitic metasandstone. Generally, the different lithological units tend not to be very continuous along strike, which contributes to the irregular map geometry. In parts of the formation, especially along the Skifergangen Shear Zone (see below), deformation makes it more or less impossible to determine the primary textures of the rocks. The metamorphism is at greenschist facies, turning most of the mafic volcanic rocks into varieties of greenstone and greenschists.(Figure 12)



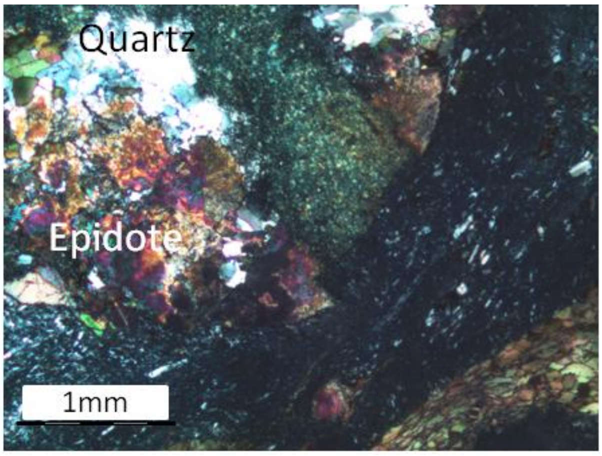


Figure 12: Metavolcanic rocks metamorphosed to greenschist facies. Growth of chlorite and epidote confirms the facies. Left photo show a fine grained groundmass of chlorite, epidote and plagioclase and possibly actinolite. Right photo shows coarser epidote in a volcanic amygdule. Fine grained chlorite to the right of it.

The map shows a rough division of the various metevolcanites. Further detailed mapping and classification of the lithologies would require significant work, and was not considered relevant to this study. A short description of the various lithologies found in the Magerfjell Formation follows below.

4.2.1 Metalavas

Massive and structureless volcanic rocks make up a large percentage of the Magerfjell Formation in the mapped area. These are mafic in composition and appear as massive greenstones in the field.

Pillow lavas are found in several places within the Magerfjell Formation. Pillows vary in size, but those observed in this study would normally range from 20-50 cm in diameter(Figure 13B). Chilled rims are common around the pillows. Where possible, the shape of the pillows was used to confirm the polarity of the sequence.

Amygdaloidal lavas are found to a great extent throughout the formation. Most commonly the amygdals vary between 1 and 2cm in size and are filled by epidote. Large amygdals of more than 5 cm were also observed, filled with epidote and plagioclase (Figure 13C). Commonly there is a zonation with epidote in the centre of the amygdales and plagioclase at the outer rims. The amygdaloidal lavas do sometimes also show clear flow structures. The flow structures are often seen as flow bands, and highlight the bedding within the volcanites. Higher densities of amygdales in certain zones probably represent lava flow tops (Figure 13A). The individual flows are commonly about 1-2 m thick, but range from 10cm to 5m.

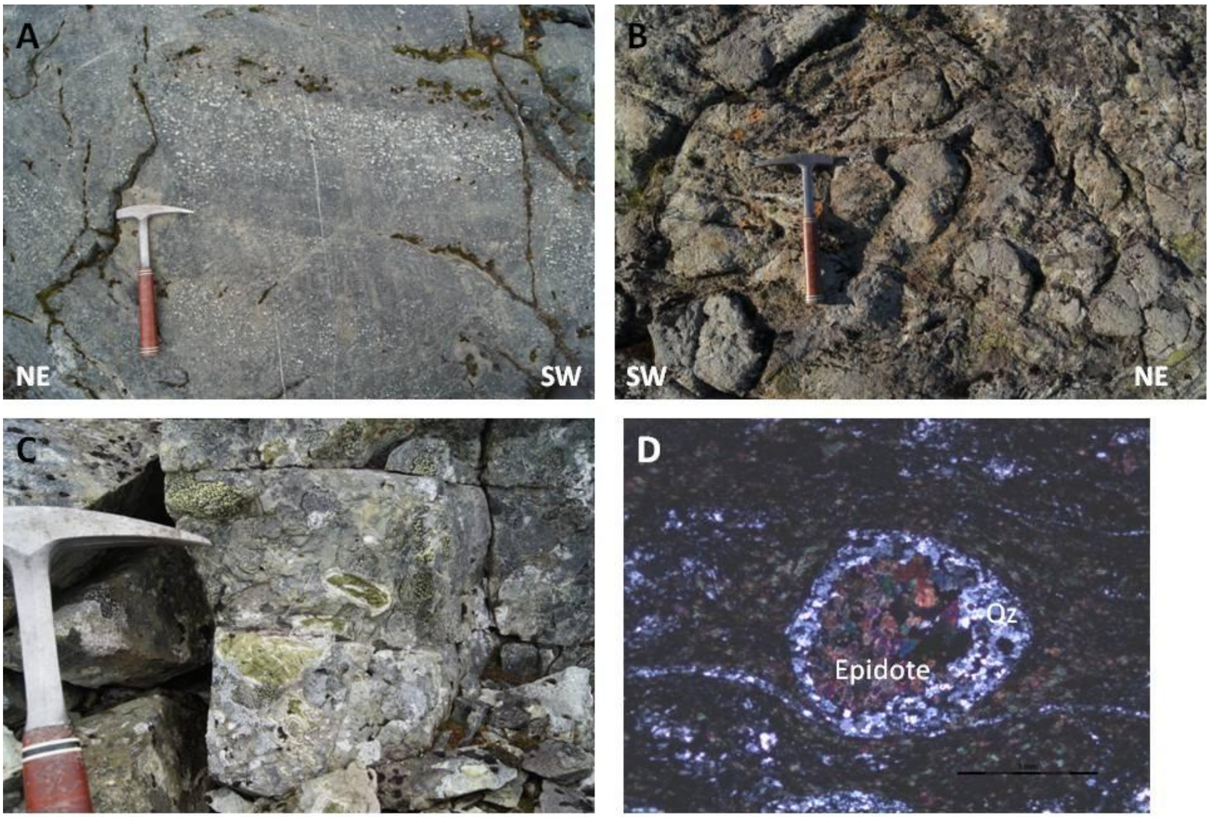


Figure 13 A)(624440.759559, 7818878.980554 UTM 34N) Amygdaloidal metabasalt. HIgher concentrations of amygdales flow tops. B: (619467.173424, 7815154.497954 UTM 34N) Pillow lava with pillows indicating the polarity. C)(617544.023935, 7814417.258852 UTM 34N). Amygdaloidal metabasaltic lava. Epidote and plagioclase fill the vesicules. D) Microphotograph of amydaloidal lava. Quartz and epidote fill the vesicules. Th ground mass is formed by fine grained chlorite, epidote, plagioclase and quartz.

4.2.2 Tuffs and tuffites

Different metavolcanic sediments occur throughout the Magerfjell Formation. Tutffitic deposits are fairly common, and appear as fine grained and normally well sorted. The composition is generally mafic to intermediate, but the overprinting deformation and metamorphism makes it hard to define the primary composition of the rocks. The thicknesses of the beds vary from a few centimeters up to several meters. In the eastern part of the formation, a distinctive package of rhythmically layered tuffites can be followed for several hundred meters. The tuffite shows cm-thin layers of volcanoclastic material of varying composition (Figure 14A). Each of the layers shows a upwards fining sequence. This package can be seen in figure. Directly above the tuffite, there is a quartzitic sandstone, with little or no volcanolclastic material.

4.2.3 Metasandstones

Sandstone beds are found throughout the formation. They are commonly only a few meters thick, and, with the exception mentioned above, do not seem to be continuous over long distances. Most of the sandstones are of pure quartzitc composition, oftenwith a greenish colour resulting from the high chlorite content in the matrix and cement. At a few localities, bedding seems to be well preserved, but no sedimentarty structures indicating the depositional environment were found.

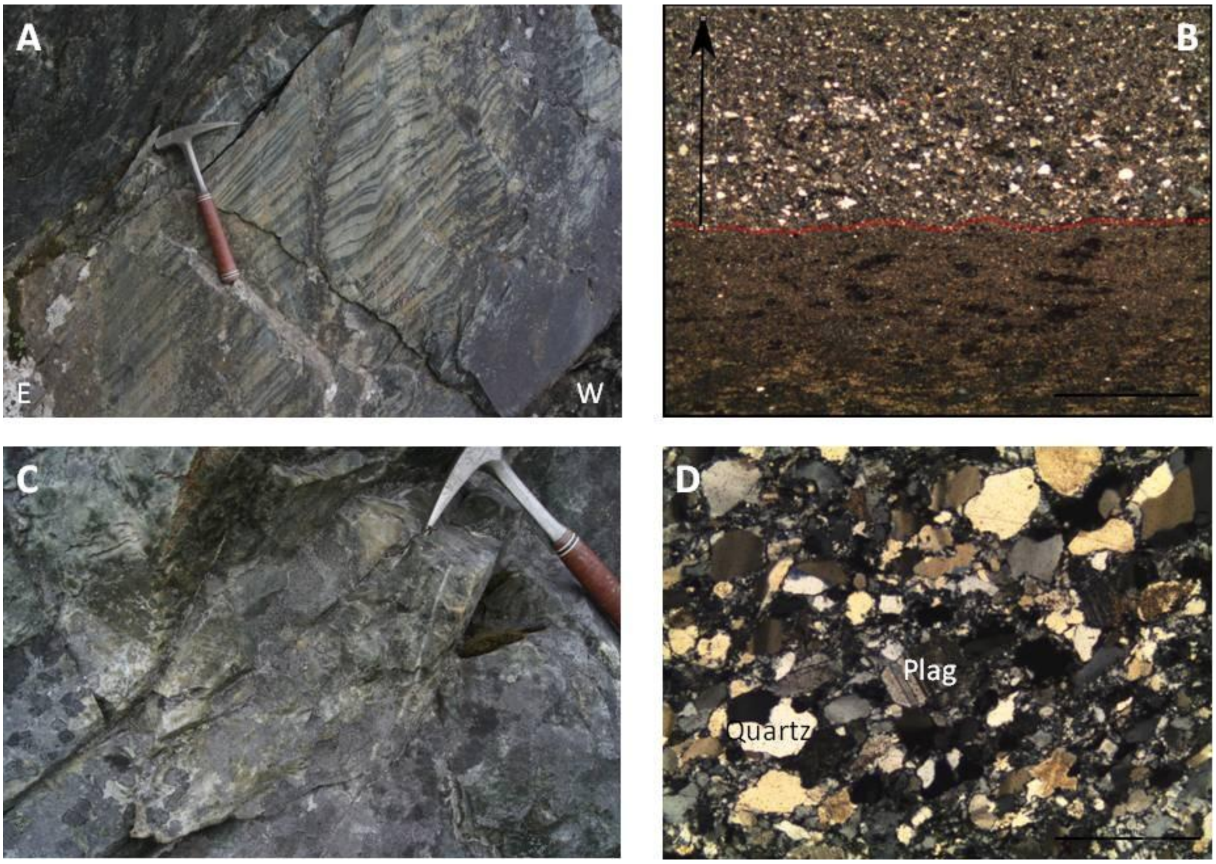


Figure 14 A)(624093.841698, 7818360.060404 UTM 34N) Laminated metatuffite. The thin layers of various compositions are interpreted as ash layers. B) Microphotograph of the metatuffite in A). The picture shows two layers of different composition. The upper layer consists of coarser quartz and feldspar, whereas the lower is very fine grained and poorer in quartz. The arrow indicates fining upwards . C)(624071.664954, 7818272.540915 UTM34N) Metasandstone of the Magerfjell Formation. The greenish colour of the sandstone is typical of the Formation. D) Microphotograph of the sandstone in C). Quartz is dominant, with subordinate plagioclase and K-feldspar.

4.3 Interpretation:

4.3.1 The Saltvatn Group

The Ulveryggen Formation shows primary features typical of a braided river system, with possible tidal influence(Pharaoh et al. 1983). The source area must have been rather silicious and constant troughout the whole time of deposition, as there are only minor variations in the composition of the sediments. The grain size variations can be explained as a result of the braided river, depositing coarser material in the channels and channel bars, and finer grained material on the flood plain. No paleocurrent directions were measured in this study, but previous work constrained a source area to the west and north based on the orientation of the trough cross beds(Figure 9, Pharaoh et al. 1983). On the basis of the relatively good sorting and rounding and the homogenous silicious composition, it is most likely that this represents a continental source area.

The Dypelv Formation represents the change to a more local source area. The very coarse and poorly sorted material supports a short transport distance. The deposition could have been from alluvial fans or debris flows. The input of volcanoclastic material may represent an increase in volcanic activity and/or a change in source area. The transition between the Ulveryggen and Dypelv Formations is gradational. This likely suggests that the change in source was also gradational, or, alternatively, that there could have been a reworking of the older, previously deposited sediments. Reworking would then cause mixing of the old and newly deposited sediment. Finer grained and better sorted material could represent a reworking of the fan material by braided streams.

The Stangvatn Formation represents yet another change in provenance. On the basis of the paleocurrent directions measured by Pharaoh et al. (1983), the source area is likely to be more southerly and dominated by felsic to intermediate volcanic rocks. The depostitional environment for the Stangvatn Formation is probably fluvial, giving the cross bedding in the finer grained beds. The repeated input of coarser material must be locally derived, most likely from alluvial fans. The source rock of the dacitic pebbles in the conglomerates has not been found within the RTW, or in other Precambrian rocks of Northern Norway (Reitan, 1963; Pharaoh, 1983). The transportation of the quite coarse conglomerate clasts cannot have been very far, so the source must have been local. The dacitic source must therefore have been eroded away completely.

The heterogeneous sedimentary rocks of the Saltvatn Group show that the depositional system was characterised by quite a rapid change of both the depositional environment and the source area over the time of deposition. There is no evidence of unconformities between the formations such that most likely the deposition was a rather continuous geological process. Jensen (1996) suggested that the volcanic rocks and the sediments are parts of basinal sequences deposited in a volcanic arc type environment. He also suggested the arc to be the source area of the sediments in the conglomerates of the Dypelv and Stangvatn Formations. The total thickness of the Group is of at least 2500 meters, which points to a continuously increasing accommodation space for the sediment deposition.

An interesting note is the change of thickness of the Dypelv and Stangvatn Formations from the west to the east. The thickness of the Dypelv Formation in the east is about 1000 m, whearas to the west, in the area of Nedre Saltvatnet and the hinge of the bend in the Nussir Group, the thickness increases to about 2500 m(Figure 8). The Stangvatn Formation shows a significant thickeningas well, but not to the same extent as the Dypelv Formation. This thickness increase is much less obvious in the Ulveryggen Formation, than in the two overlying formations. In addition to the increasing thickness, the dip direction of the beds changes from NW to NE around what is called the Nussir structure. Pharaoh et al. (1983) discussed the subject, and concluded that the thickness increase is mostly apparent and a concequence of deformation with *"(...)Thickening in the hinge of the Nussir structure and strong attenuation on its limb."* They estimated an original thickness of the Dypelv Formation of about 1100 m.

Although the beds clearly bend around the Nussir structure and the dip direction changes as a result of this fold, there is no significant change in the dip. As seen in the profiles of Figure 15 Figure 16, the formations appear homoclinal on the north western side of the large antiform of the Saltvatn Group. As there is no change in dip, nor any obvious faults and tectonic duplication of strata, the increase in thickness towards the east must be true.

A primary depositional increase in thickness in the eastern part can only be explained by having greater accommodation space there. This could be explained by a fault controlled, extensional basin. A possibility is a normal fault in the eastern part of the basin, creating a half graben, which would lead to a greater depositional space in this part of the basin than in the eastern part. A possible environment of a backarc basin in a volcanic arc system is suggested in the discussion below. Further work on the sedimentology and basin development is needed in order to confirm or disprove the existence of such a basin.

4.3.2 The Holmvatn Group

The strong deformation in parts of the Holmvatn Group makes it hard to recognize primary structures, which in turn makes the interpretation of the depositional environment very challenging. The various volcanic and sedimentary rocks are all supracrustal. Intrusive rocks are found in the Holmvatn Group, but in the studied area we only find minor metagabbroic intrusions. The volcanism in the area seems to have been spasmodic during the deposition of the Holmvatn Group. This might explain the rapid changes from tuffaceous material to coarser grained volcanoclastics and massive lavas flows. The environment of deposition could possibly have been subaqueous as we find pillow lavas in parts of the area, sporadic carbonate development and rarely preserved cross bedding, or other sedimentary structures in the sediments. This is also suggested by Pharaoh et al. (1983).

The rock assemblage consisting of a variety of volcanic rocks interbeded by sandstones and volcanoclastic rocks could indicate a volcanic arc type environment for the deposition. The thick package of volcanic material throughout the group indicates that the active arc massif must have been close. Small intra-arc basins basins give space for the deposition of the sedimentary rocks.

5 Statigraphy:

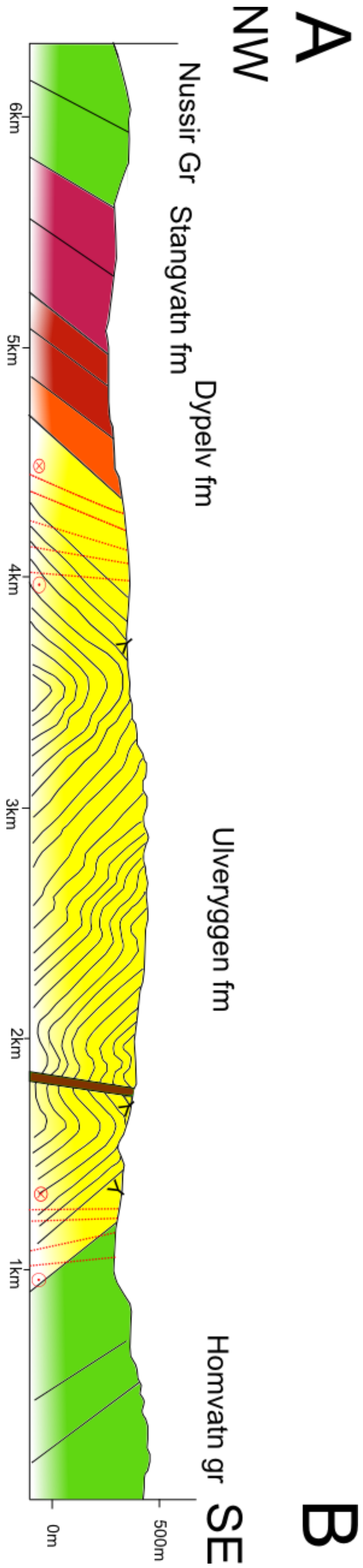


Figure 15 Easternmost cross section , The trace of the profile is drawn on the map in figure 5. The cross section clearly shows how the Ulveryggen Formation is the stratigraphicaly lowermost formation in RTW.

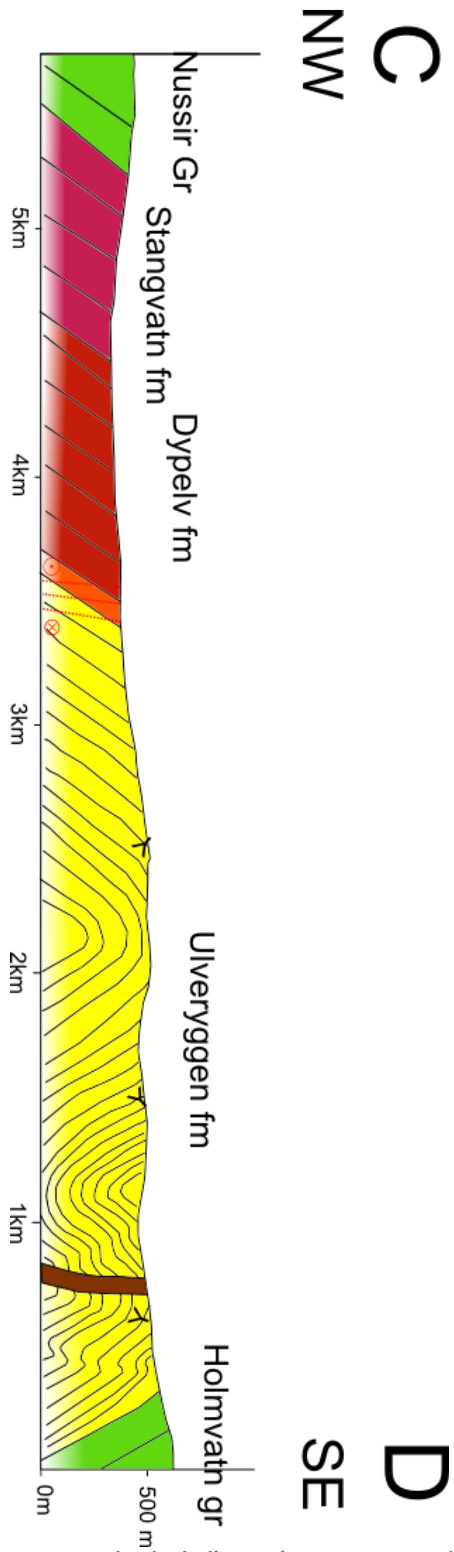


Figure 16 Westernmost cross section. The cross section is similar to the A-B cross section, but the intensity of the parasitic folds on the southeastern flank seems to be lower.

5.1.1 Evidences of a new stratigraphical order

The new cross sections of Figure 15 and Figure 16 help to constrain the stratigraphy of study area. According to this study, the Saltvatn Group is without doubt the lowermost lithostratigraphic group of the Paleoproterozoic sequence exposed within the RTW. Nowhere in the window is the base of this Group exposed, and its total thickness remains therefore unconstrained. Conformably above the Saltvatn Group lies the Holmvatn Group in the south and the Nussir Group to the north.

Both Reitan (1963) and Pharaoh et al. (1983) described the contact between the Saltvatn Group and the Holmvatn Group as being unconformable. However, whereas Reitan (1963) described the contact as being strongly tectonized, Pharaoh et al. (1983) described it as an erosional unconformity, possibly with a small angular unconformity. The contact is indeed tectonized in the eastern part of the RTW, as Reitan correctly observed. This makes it difficult to determine the correct stratigrapical order of the Holmvatn and Saltvatn Groups in this area. However, this is not the case farther west, where the contact is rather undeformed in places. Whether the Holmvatn Group is deposited conformably above the Saltvatn Group, or there is an erosional unconformity and possibly a hiatus at the contact it is not certain from this study, but there is no indication of an angular unconformity.

Folding of he supracrustal sequences in this part of the RTW also affects the appearance of the stratigraphical contact. Nowhere in the investigated area are there evidences of beds being overeturned, and with great confidence, it can be concluded that this not an issue in this part of the window. The stratigraphy is everywhere facing the correct way up.

The best evidence for the primary stratigraphic order is found in an area southwest of Øvre Saltvatn(Figure 2), where it is possible to observe directly the well-exposed contact between the Holmvatn and the Saltvatn Group. There is very little deformation of the rocks in this area, and the primary contact is well preserved The Holmvatn Group greenstones cleary overlie the metasandstones of the Saltvatn Group. As seen in Figure 17 the contact is uneven, possibly an erosional contact. The contact lies relatively flat in this area, with a shallow dip of 20-30° towards south east. The Holmvatn greenstone rests upon the Saltvatn sandstones, making the hilltops of several minor hills. As there is no overturning of the beds, this is the primary stratigraphic order, and the Holmvatn Group must therefore be younger than the Saltvatn Group, which therefore represents the oldest rock sequence exposed within the Paleoproterozoic RTW.

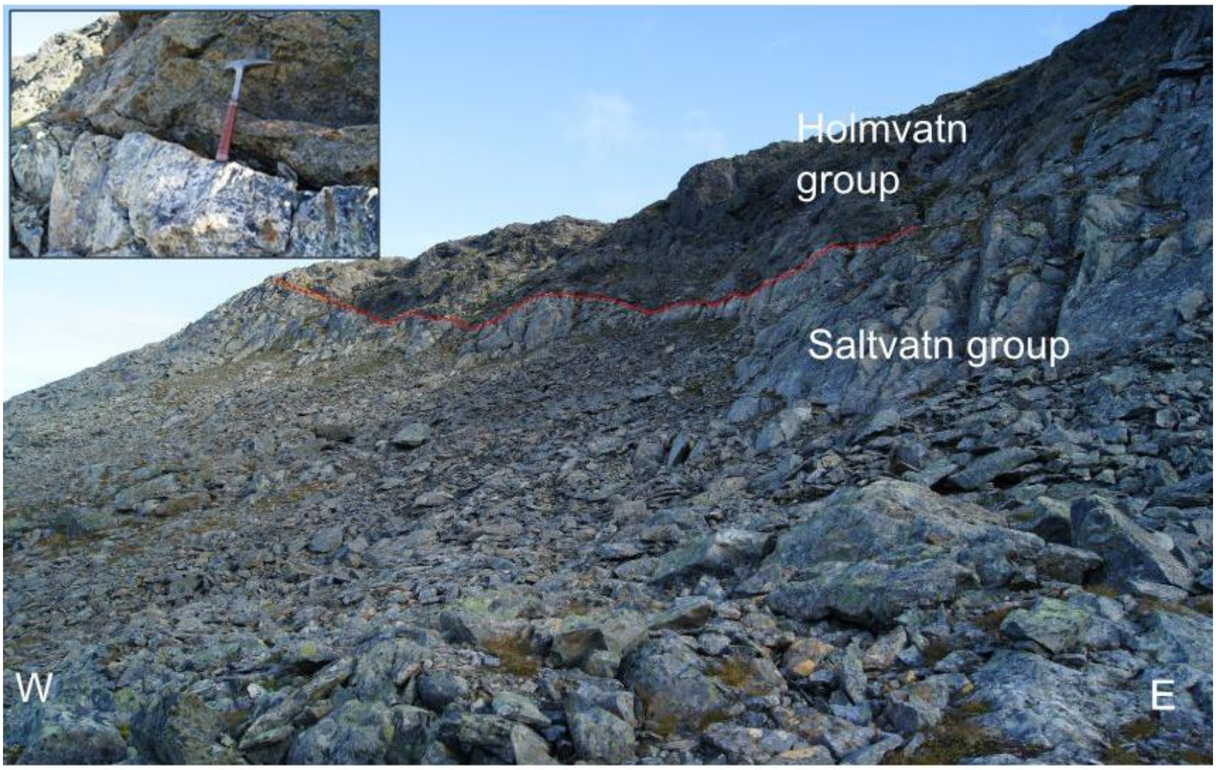


Figure 17: (607201.570112, 7810603.700221 UTM 34N). The contact between the Holmvatn Group and the Saltvatn Group. The Holmvatn Group clearly lies above the Saltvatn Group.

This area is also the key area for Pharaoh et al. (1983) conclusion about the Saltvatn Group being the younger of the two Groups. They wrote: "*It is possible to demonstrate that the Saltvatn Group overlies the Holmvatn Group in the area south-west of Øvre Saltvatnet where the contact between the two groups is folded in a major upward- facing coupled fold with wavelength of 1.5 km. Cross bedding in the Saltvatn Group is well preserved and youngs away from a primary depositional contact with the Holmvatn Group, clearly demonstrating that the latter is the older of the two Groups.*"

Their conclusion is based on the observation of the bedding in the Saltvatn Group younging away from the contact, which would therefore imply that the Saltvatn Group must have been deposited later than the Holmvatn Group. This is however not a solid argument. As the contact is folded about several fold hinges(see attached structural map), the bedding of the Saltvatn Group will be younging towards or away from the contact all depending on which side of a fold hinge the observations are made (Figure 18). This will also be dependent on the topography, as the level of erosion is controlling where the contact is exposed. Good examples of this are found at several locations along the contact, where fold hinges fold the lithological contact. The only solid evidence for the age relationship is therefore found there where the contact is exposed and the relationship between the formations on each side can be observed.

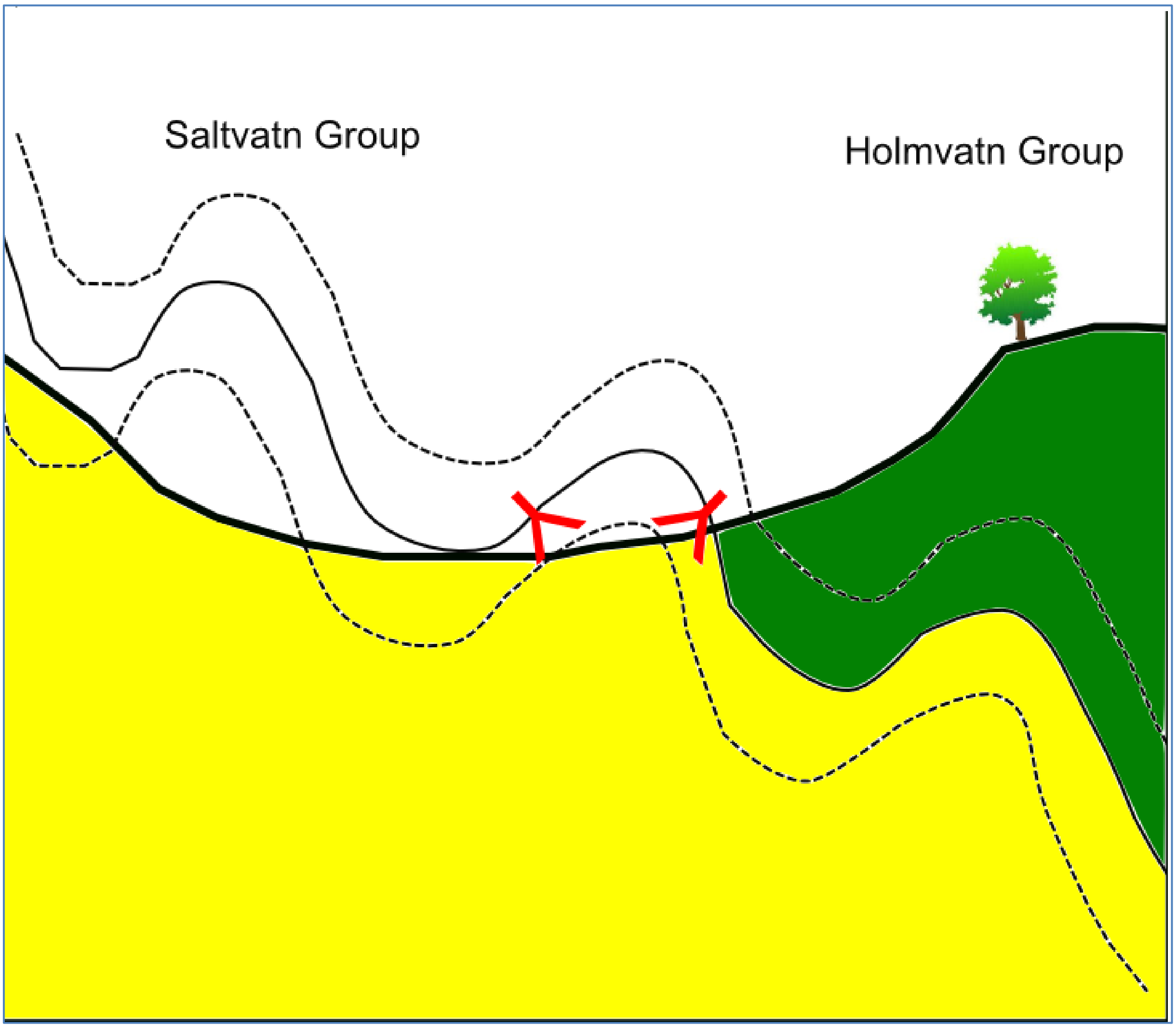


Figure 18: The contact between the Saltvatn Group and Holmvatn group is folded, and the younging direction will be either towards or away from the contact depending on what side of the fold hinge the observation is made.

At a few locations, e.g. south of the Ulveryggen mountain(Figure 2, locality 4), the Saltvatn Group and Holmvatn Group seems to be interfingering and we find tounges of the groups wedging into one another. This could be the result of primary depositonal interfingering, as the groups probably are deposited in a conformable way with no depositional hiatus. More likely, however, the tounges are a result of the cutting effect of the topography. Figure 19 shows the effect of topography in a situation where the bedding is dipping gently and we have a valley cutting down into the underlying formation. The folding will create a similar effect where the topography is at a level revealing only a down folded syncline of the overlying formation within the underlying formation. The opposite situation, with the lower formation folded up into the overlying is of course also a possibility.

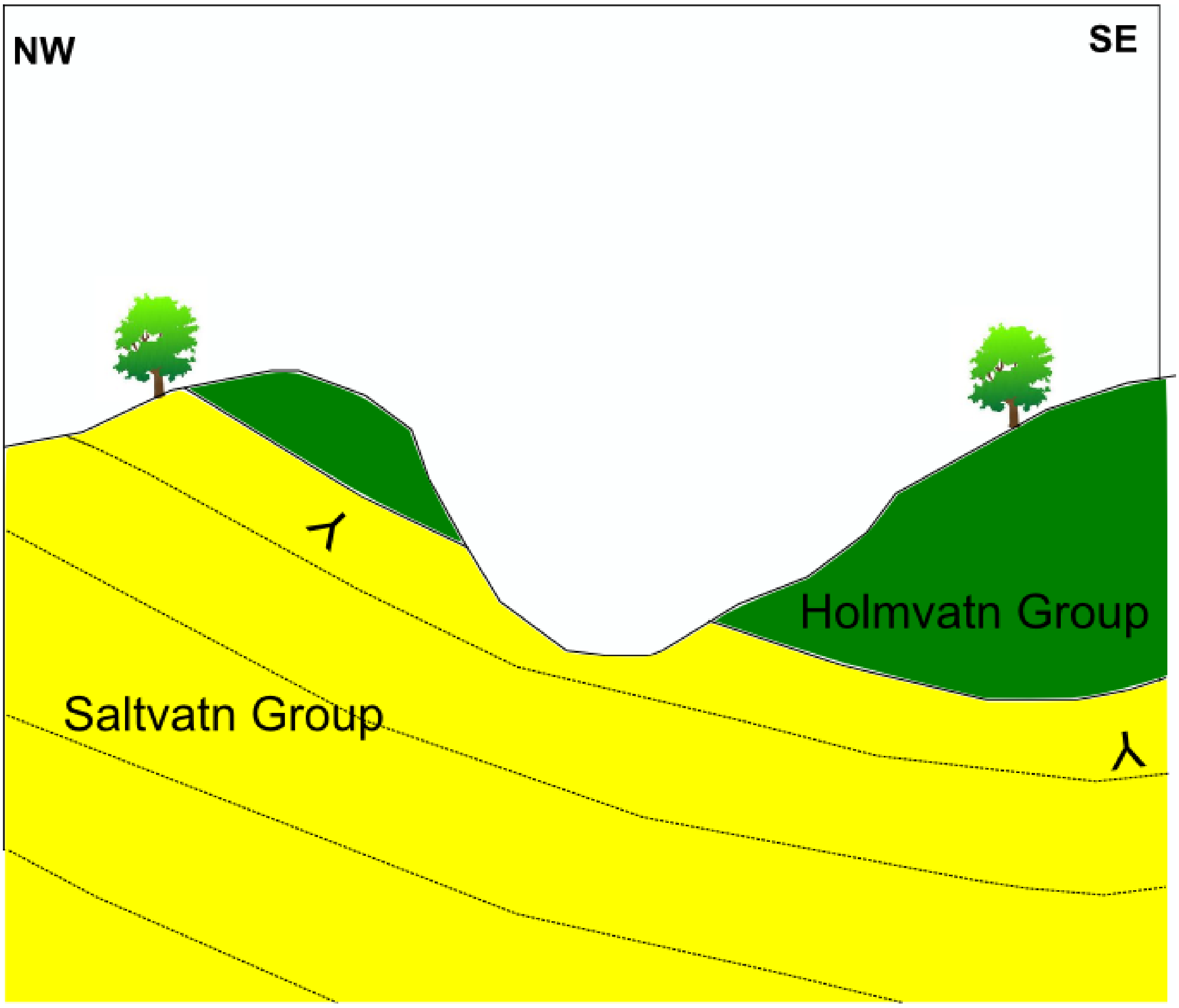


Figure 19: The effect of the topography on the apperance of the contact as seen in the area south of the Ulveryggen Mountain.

6 Structural geological results

6.1.1 Folding of the antiformal stack

The cross sections in Figure 15 and Figure 16 clearly show the folded nature of the central part of the Repparfjord Tectonic Window. The Saltvatn Group sedimentary rocks are folded by upright, open folds with a wavelength of several kilometers. As the cross section shows, there is a major antiform with its hinge in the central part of the Ulveryggen Formation. On the southeastern limb there are several minor second order folds with wavelengths of 200-700 meters, whereas the northwestern flank shows a more homoclinal style. The fold axis trends in a NE-SW (042/08) direction as shown in Figure 20. The folding has caused the development of a rather pervasive, near vertical axial planar foliation. This foliation is dominant in large parts of the studied area, although it is more penetrative in the hinge areas of the folds. It is best developed in pelitic beds in the sandstones, which also show the relationship between the axial planes and the bedding clearly. As the axial planes are always more steeply dipping than the bedding, this relationship helps confirming that the bedding is the right way up. In the Ulveryggen Formation there occur thin and elongate metagabbroic intrusions, which seems to follow the axial planar foliation. Most likely the intrusive rocks intruded along the weaker and already existing zones of axial planar foliation. The folding is most likely related to the deformation that took place during the Svecokarelian Orogeny, about 1840 Ma ago (Pharaoh et al. 1983), although this remains speculative.

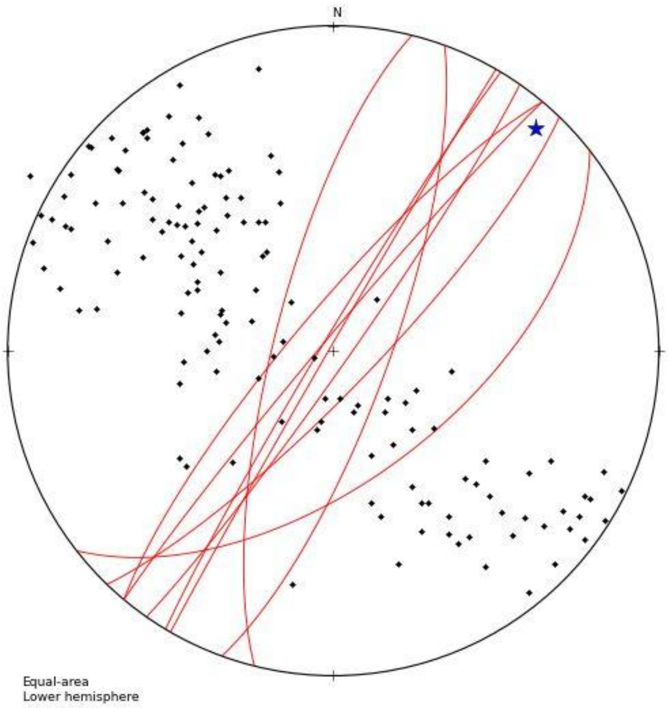


Figure 20: Structural data form the Saltvatn Group. Poles to bedding planes plotted as black dots show the cylindrical folded nature of the sandstones. The fold axis of the fold, derived from the bedding measurement, is plotted with the blue star. Axial planes are shown in red.

6.2 Skifergangen shear zone

The Skifergangen shear zone (SSZ) is located within the Holmvatn group, just south of the Ulveryggen Mountain (Figure 21). It deforms various mafic metavolcanic rocks, of both tuffitic and lavaflow origin. Locally strain is so intense that it is difficult to determine the original composition and petrography of the deformed rocks. The shear zone has not been described in detail before, and was first mentioned as a mylonitic shear zone by Viola et al. (2008). Although not recognized as a shear zone, the area has been known for its foliated appearance, from where also the name “Skifergangen” originates. In the early 1900s' minor scale exploration and mining took place in the area, and several prospecting sites can still be recognized. The main goal of the mining activity was the copper mineralizations found at several locations along the shear zone. In 1904 alone more than 50 different sites were prospectedin the area around Skifergangen(Jacobsen 1989). Prospecting activity was localised mostly along strike of the central part of the shear zone over a length of several hundred meters, and with multiple small digs. Apart from this intensive exploration, however, large-scale mining was never carried out.

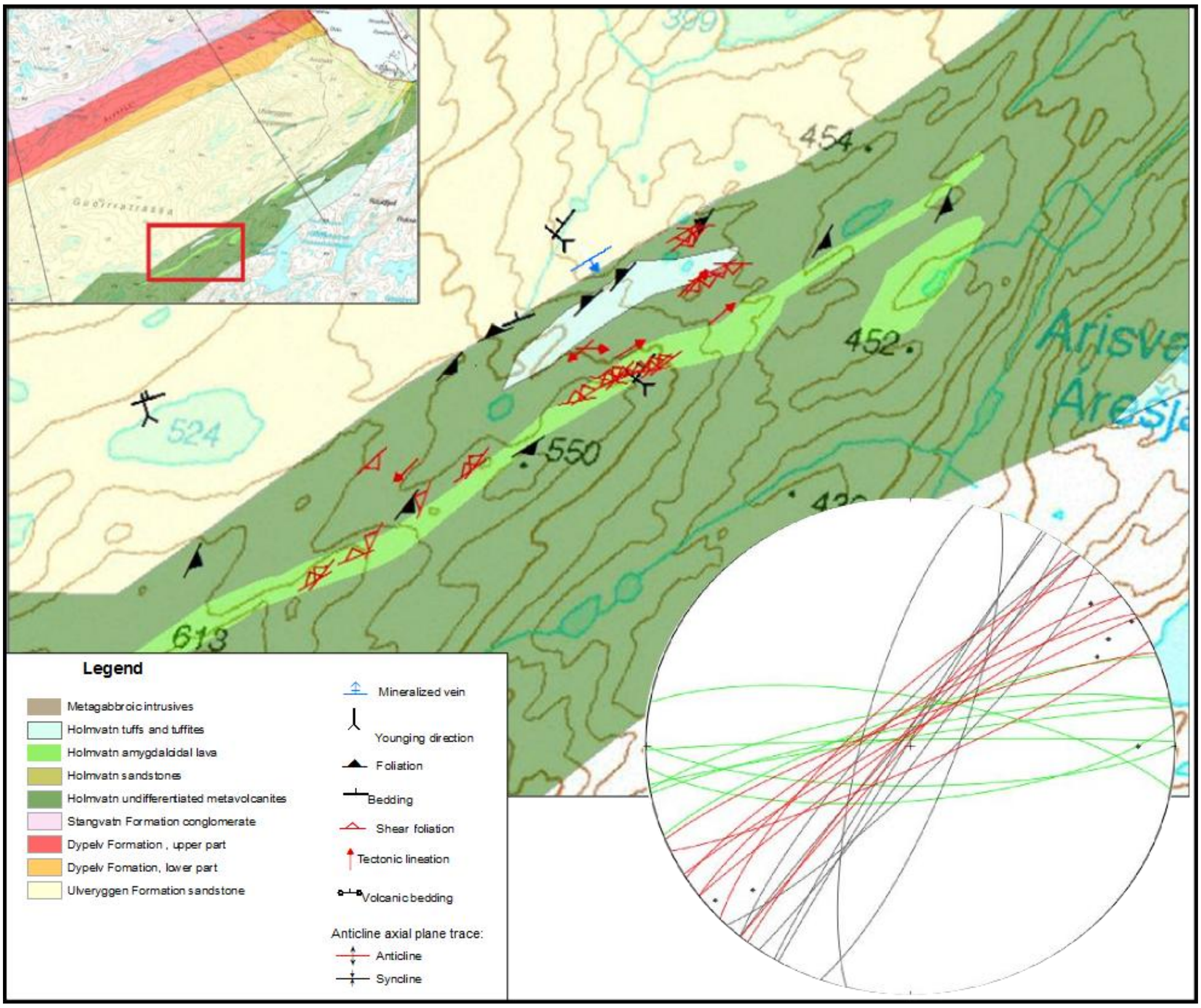


Figure 21: Location of the SSZ. The stereonet shows relevant structural data from the shear zone. Red great circles plot mylonitic foliation planes. Black great circles represent the external foliation, whose asymptotic dragging into the shear direction indicates a component of dextral shear. Green great circles plot the orientation of individual dextral shear bands, often arranged in pervasive extensional crenulation cleavage patterns (EEC). Sub-horizontal stretching lineations(black dots) indicate the strike slip nature of shear along the shear zone.

The size and boundaries of the shear zone are difficult to determine. The area is dominated by a pervasive foliation, both shear foliation and an external foliation, which was locally dragged into the shear zone. The latter is most likely axial planar to the large scale folds mentioned above. The angular difference between the shear zone foliation and the axial planar foliation is small, and varies between 10 and 30° (Figure 20) and they are therefore not easily differentiated.

6.2.1 Characteristics of the shear zone

The SSZ shows an overall ductile deformation style. The shear zone boundaries are rather irregular, giving the deformation zone a varying thickness. The shear zone architecture is partly anastomozing, and in its central part it splits up in at least two separate high strain zones with undeformed parts between. This is documented in Figure 22, which illustrates field and microtectonic observations along a NNW-SSE profile across the shear zone over a distance of ~200 m. Outside the shear zone, little or no deformation is recognised besides the axial planar cleavage, as mentioned above.

The lateral termination of the shear zone was not observed clearly, so the length of the zone remains unknown, although, based on the observations made in this study, it extends at least 1.5km in a NE-SW direction.

The stereoplot in Figure 21 shows the external foliation, the mylonitic shear foliation and the shear band foliation observed within the shear zone. The relationship between the mylonitic and the shear band foliation, is a clear indicator of a dextral sense of shear for the shear zone.

Lineations found on the mylonitic foliation planes are interpreted as stretching lineations, and their horizontal to sub-horizontal attitude (Figure 21) indicates that the movement along the shear zone was strike-slip, with no obvious indications of possible later reactivations.

Within the central mylonitic core, there is a very pervasive shear foliation and well developed mylonitic structures. The metavolcanic sequence is deformed to very fine grained greenstone, where no primary features are observable. The mylonitic NW-SE striking foliation is the most prominent feature of the shear zone. Shear bands at a low angle to the main foliation are well developed, forming characteristic S-C' structure, locally so pervasively so as to generate proper extensional crenulation cleavages (EEC fabrics). The S-C' structures indicate consistently a dextral sense of shear (Figure 23A). Sigmaclasts confirm the dextral kinematics. The clasts are commonly formed around a core of quartz or robust plagioclase volcanoclasts and the recrystalized strain tails are normally formed by quartz or chlorite. Examples of ore minerals forming the tail are also observed(Figure 23B).

Asymmetric boudinage of quartz layers highly transposed parallel to the foliation is common within the shear zone (Figure 23C and D). Quartz levels are split into boudins and the individual boudins are rotated against the shear direction. The individual boudins are often separated by up to 10-20 cm.

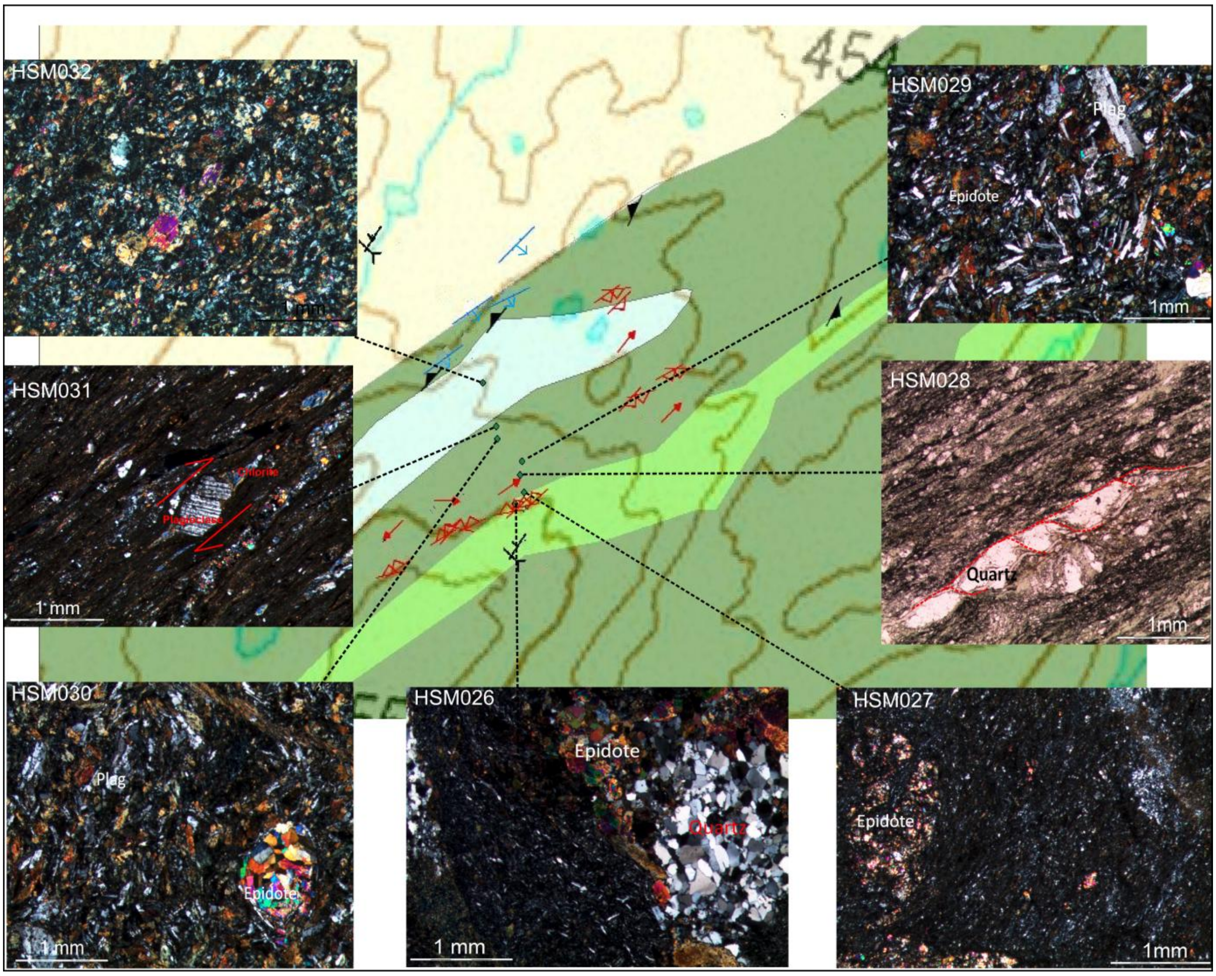


Figure 22: Microphotographs showing the varying degree of deformation across the shear zone. The samples are collected along a NNW-SSE profile across the shear zone and their sampling location is indicated on the map. All microphotograph are oriented in map view with north up. HSM026, the farthest to the south east, is a basically undeformed amygdoidal basalt. HSM027 shows a similar rock type, but the grain size is significantly reduced. HSM028 is from the central part of the shear zone and documents the pervasive mylonitic foliation with microscale shear bands giving dextral sense of shear. HSM029 and HSM030 are undeformed samples of a quite coarse-grained volcanic rock, with primary volcanic textures still visible. HSM031 is sampled from the second branch of the shear zone. The photo shows a microscopic sigmaclast. HSM032 shows a undeformed sample of a volcanoclastic, tuffitic rock.

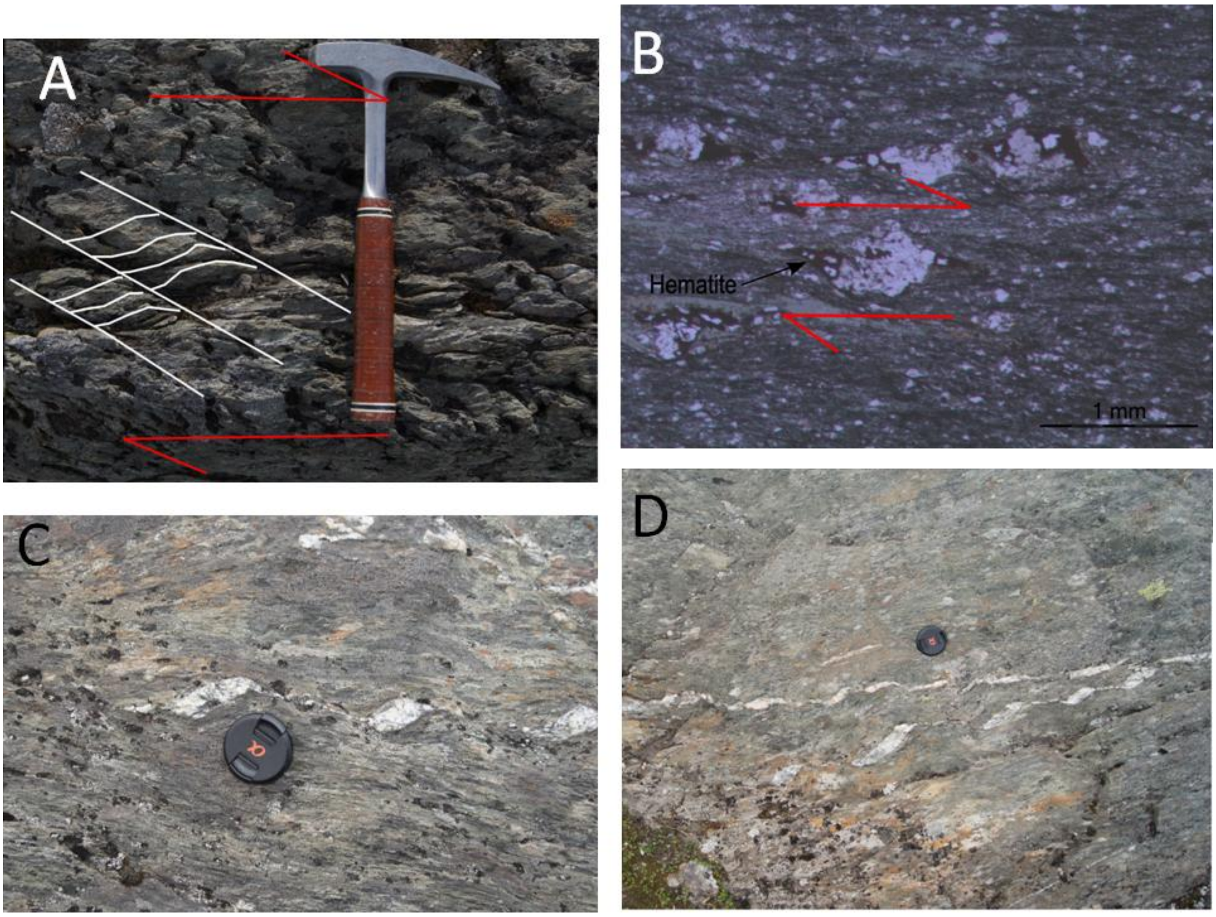


Figure 23: Structures from the shear zone. A(1063280.3649, 7882974.0524 UTM34N): Shear bands developed with a low angle to the foliation. Pervasive SC'-fabric generating a clear ECC. B): (1063104.3997, 7882855.838 UTM34N) Microphotograph showing a sigmaclast within a very fine grained, foliated mylonite. Hematite is recrystallized in the strain shadows of the clast. C and D)(1063257.2125, 7882961.094 UTM34N) Asymmetric dextral boudins formed by a relatively competent quartz vein. The now seperated boudins formed once a continous vein, and were split and rotated during continuous non coaxial dextral shear.

6.2.2 Ore mineralization

Minor mineralized veins are found at several locations throughout the SSZ. Quartz is the main mineral making up these commonly up to 20 cm wide veins. Their strike is normally parallel to the mylonitic foliation. The length varies from 10 cm to tens of meters. The ore mineralizations occur mostly in association with these veins. Chalcopyrite is the dominant ore mineral in the investigated samples. It occurs together with hematite in quartz veins (Figure 24). The mineralization can be commonly related to the shear structures, as seen in Figure 24A andB, where precipitation of quartz and ore minerals is localized into a shear band affecting the greenstones.

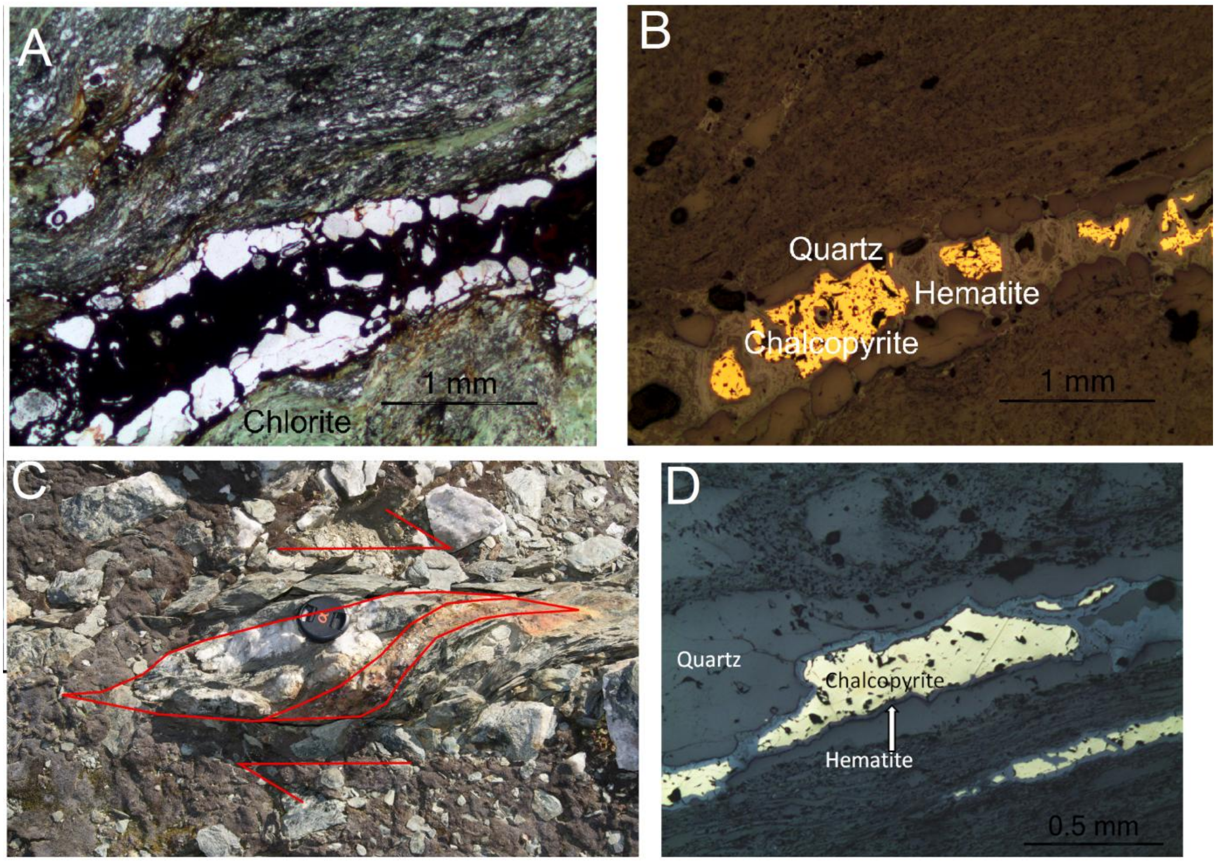


Figure 24: Ore mineralizations in Skifergangen. The minor mineralizations are associated with quartz veins. Photo A and B show the same thin section, in transmitted and reflected light, respectively. Chalcopyrite and hematite are situated within a narrow, continuous quartz vein. The vein is found at a shear band in the foliated rock. The pictures are oriented with north up. C) Transposed and deformed quartz vein forming a sigma clast. Ore mineralizations are localized in the strain shadow of the clast. D) Similar situation as in A and B. Chalcopyrite and hematite associated with a quartz vein. The differences in colour of B) and D) is due to camera settings.

6.3 Bratthammer

The Bratthammer Deposit is a relatively small (<1m) thick Cu-rich carbonate vein, located within the Magerfjell Formation of the Holmvatn Group close to the contact to the Ulveryggen Formation sandstone (**Error! Reference source not found.**). The deposit was mined in the early 1900's. The vein can be followed laterally for about 40 meters, after which it thins out in both directions. Down dip, the thickness is reduced to only 10 cm a few meters below the surface. The vein dips about 40° towards the southeast and cuts the vertical to sub vertically foliated and NE-SW striking greenstone hostrock. Mineralogically identical (although not mineralised) but smaller veins with approximately the same orientation are located in the immediate vicinity and found in both the Magerfjell Formation greenstone and in the Ulveryggen Formation sandstone.

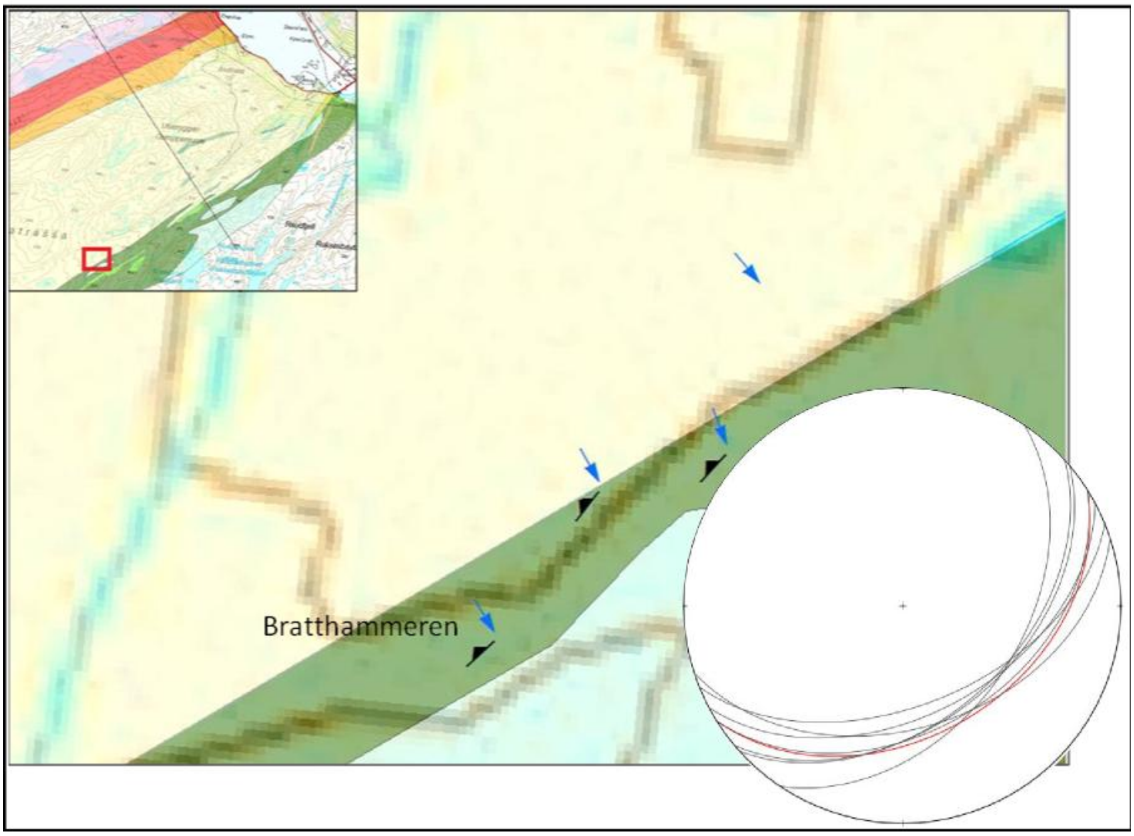


Figure 25: Map showing the location of Bratthammeren. Minor veins of similar orientation are located northeast of Bratthammeren. The stereonet shows the orientation of these veins. The Bratthammer Vein is shown in red.

6.3.1 Structural framework of the Bratthammeren vein deposit

Within the Bratthammer vein there are evidences of top to the northwest movement, which, for the present-day southeasterly dip of the vein, corresponds to a thrusting component. Small duplex imbricate structures and sigma clasts within the carbonate veins are clear kinematic indicators (Figure 26). The sense of shear as indicated by small scale indicators within the vein can be recognised only within the old, abandoned mine shaft, which gives access to sections parallel to the regional transport direction and perpendicular to the vein foliation and thus exposes the thrust.

At the thrust contact we find cataclasite and fault gouge. This is present as crushed rock fragments of various size and a very fine grained, clay rich material. The gouge is only present in very thin zones of maximum a few centimetres within the cataclastic rock. In the observed part of the vein, the thrust contact is located in the lower and more foliated part of the vein. The fault gouge is sampled for K-Ar age analysis (see below).

The Bratthammer vein has a rather pervasively foliated lower part, whereas the upper part is rather undeformed. This is also reflected in the differences in ore mineralogy, whereby in the lower, foliated part the sulphides appear along elongate bands, whereas in the upper part, massive suphide blobs are common.

Undeformed clasts of greenstones and quartz are found dispersed within the vein, and are probably derived from the greenstone hostrock, and possibly also the underlying Ulveryggen Formation sandstone.

The carbonate vein at Bratthammeren likely infilled a hybrid fracture (that is, a fracture that accommodates both a dilational and a shear component) formed in response to overall NW-SE shortening. Figure 27 proposes a conceptual scheme accounting from a mechanical perspective for the localised rock dilation and fracture infill by the vein. The largest compressive stress, σ_1, was oriented NW-SE, with the least compressive stress, σ_3, vertical to sub vertical in order to induce dilation along the southeast dipping thrust plane.

North east of Bratthammer (Figure 25) there are several minor quartz/carbonate veins up to 10-30cm in thickness, and they share approximately the same orientation as the mined larger Bratthammer vein. None of these veins can be followed laterally for more than 10 m. No indication of thrusting movement similar to the Bratthammer thrust was observed. This can, however, not be ruled out, as the veins are not well exposed along their dip direction. The veins seem to be of a simpler mineralogical composition and most of them are quartz filled. At one location, pyrite and possibly other sulphides occur together with quartz.

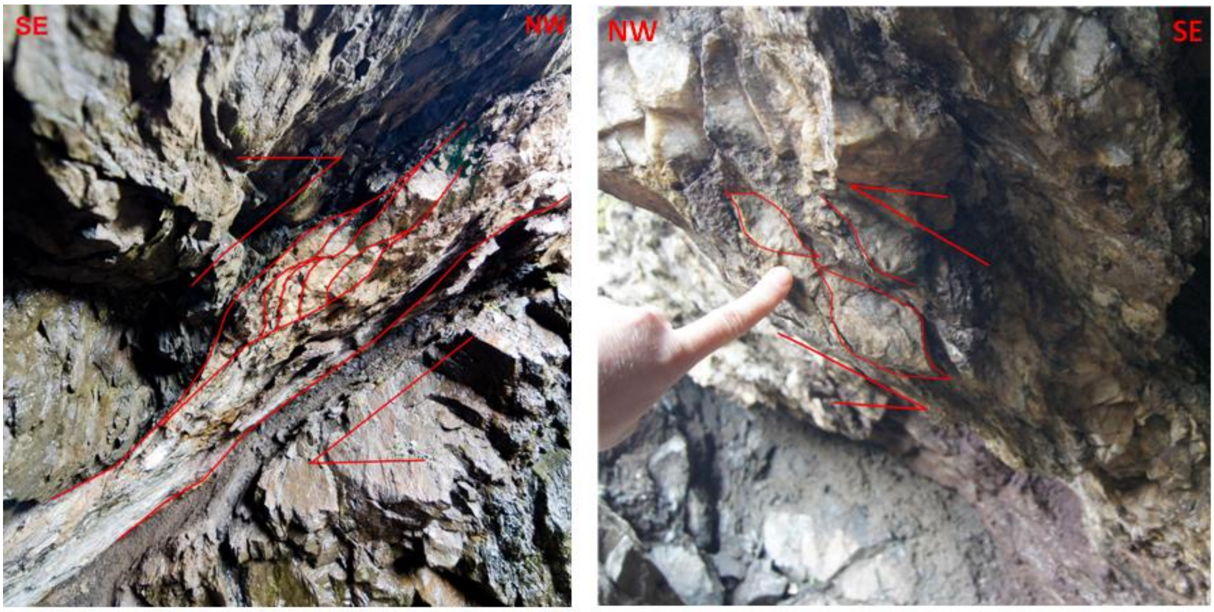


Figure 26 Kinematic indicators from the Bratthammer vein deposit, suggesting a top-to-the-NW thrust. To the left a minor duplex structure showing top to NW. The vein is there about 30 cm thick. The photo to the right shows sigma clasts with their asymmetries indicating a top to NW sense of movement.

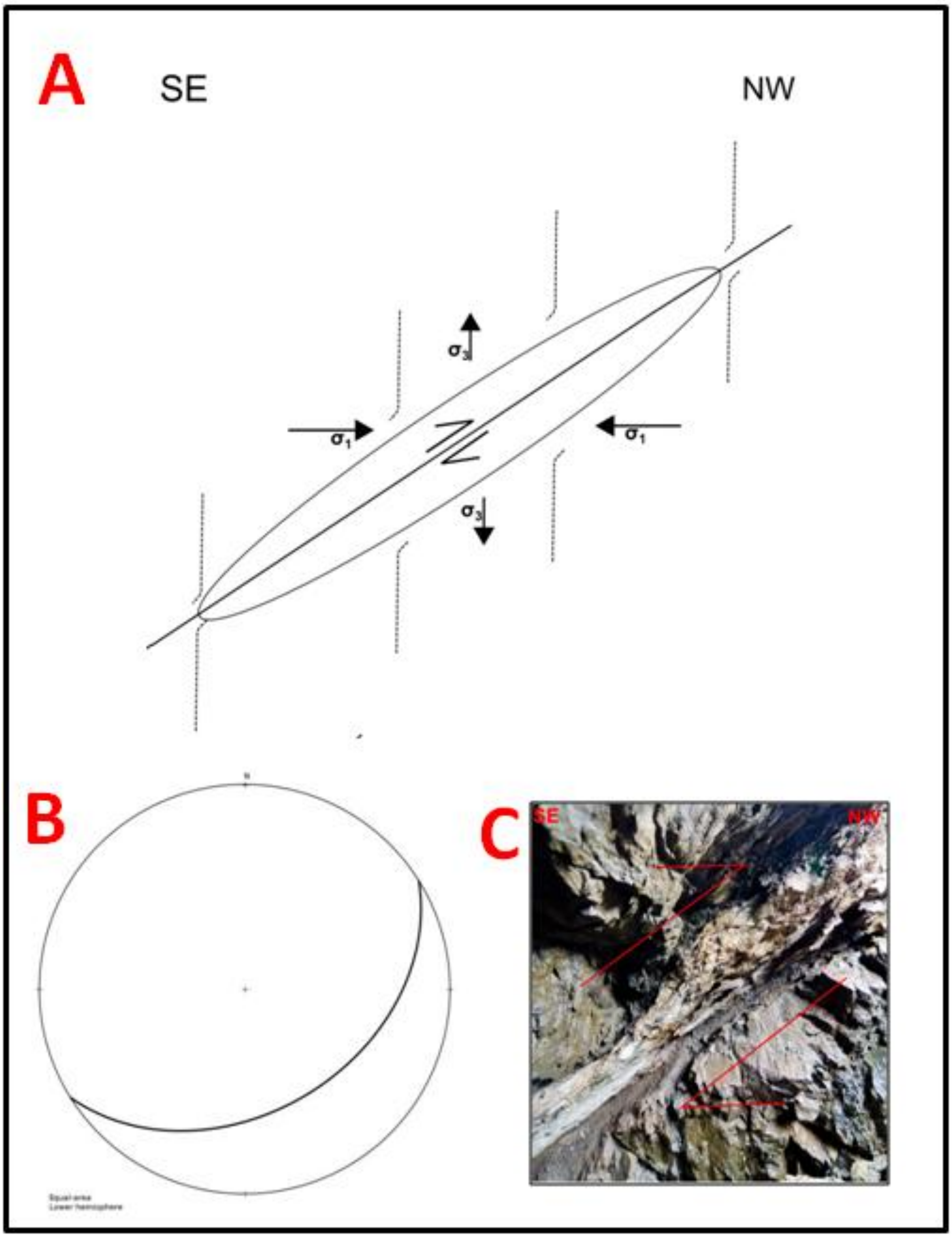


Figure 27: A) Concepual sketch of Bratthammeen vein as a result of an overall compressive stress regime. The observed minor deflection of the greenstone external foliation into the shear plane also indicates top-to-the-NW transport. B) The stereonet plots the orientation of the Bratthammeren vein. C) Close-up picture of the vein. The thickness of the vein is about 30 cm in this picture.

6.4 K-Ar illite age dating

A rapidly developing technique for the dating of relatively low-temperature faulting episodes is the characterisation and K-Ar analysis of authigenicillite crystals formed synkinematically within gouges and other brittle fault rocks. Slip on brittle faults can develop cataclasite and gouge composed of crushed rock fragments and authigenic, synkinematic clay minerals. In particular illite can crystallise by retrograde hydration reactions at formation temperatures around 100-250° C (Merriman and Kemp 1996, Haines and van der Pluijm 2008).

In order to add absolute time constraints to the deformation episode accommodated by shearing in the Brathammer area, two samples of the fault gouge from the central part of the vein were collected. Their illite was separated and dated by Dr. H. Zwingmann at the K-Ar laboratory of the CSIRO in Perth, Australia. The two samples were collected in an abandoned mine shaft, which gave accesses to deeper parts of the fault plane, with no evidence of weathering. The zone of highly deformed ultracataclasite and gouge is only a few cm thick and located just bellow and partly within the mineralized quartz-carbonate vein.

6.4.1 Results

The analytical approach used to separate, characterise and date illite follows the methodology described by Zwingmann et al. (2010). Unfortunately, due to lack of material, the only fraction analysed in the first sample (HSM_1) was the <2 µm fraction. In the second sample (HSM_116) the fractions <2 µm and 6-10 µm were analysed. XRD analysis was performed for the same fractions, and in the 2-6 µm fraction of HSM_116 in addition.

The results from the XRD analysis (Table 1) show that there is a variation in the mineralogy of the two samples. In sample HSM_1, calcite is the dominant mineral with illite polytype $2M_1$, chlorite, albite and amphibole as secondary mineral phases. In HSM_116, however, phlogopite is dominant in all fractions. In the <2 µm fraction, illite-smectite is also dominant, whereas in the coarser fractions, its content is lower. The coarser fractions have significant higher albite content.

Sample nr.	Quartz	Smectite/ vermiculite	Illite-Smectite	Illite 2M1	Chlorite	Albite	K-feldspar	Phlogopite	Amphibole	Hematite/ Goethite	Calcite	Others
HSM_1 <2µm	T	T		T	T	T			T	T	D	T-Mal
HSM_116 <2µm	<1	2	44		3	9	<1	43				
HSM_116 2-6 µm	1	8	8		11	22	6	44				
HSM_116 6-10 µm	1	6	11		7	22	5	45	<1			

Table 1: XRD data from HSM_1 and HSM_116. D=dominant mineral phase. T=traceable mineral phase. Numerical values are percentages.

The <2 µm fraction of sample HSM_1 yielded a K-Ar age of 492.2 ± 10.8 Ma. For sample HSM_116, the <2 µm fraction yielded an age of 459.6 ± 9.6 Ma and the 6-10 µm fraction an age of 575.8 ± 11.8 Ma.

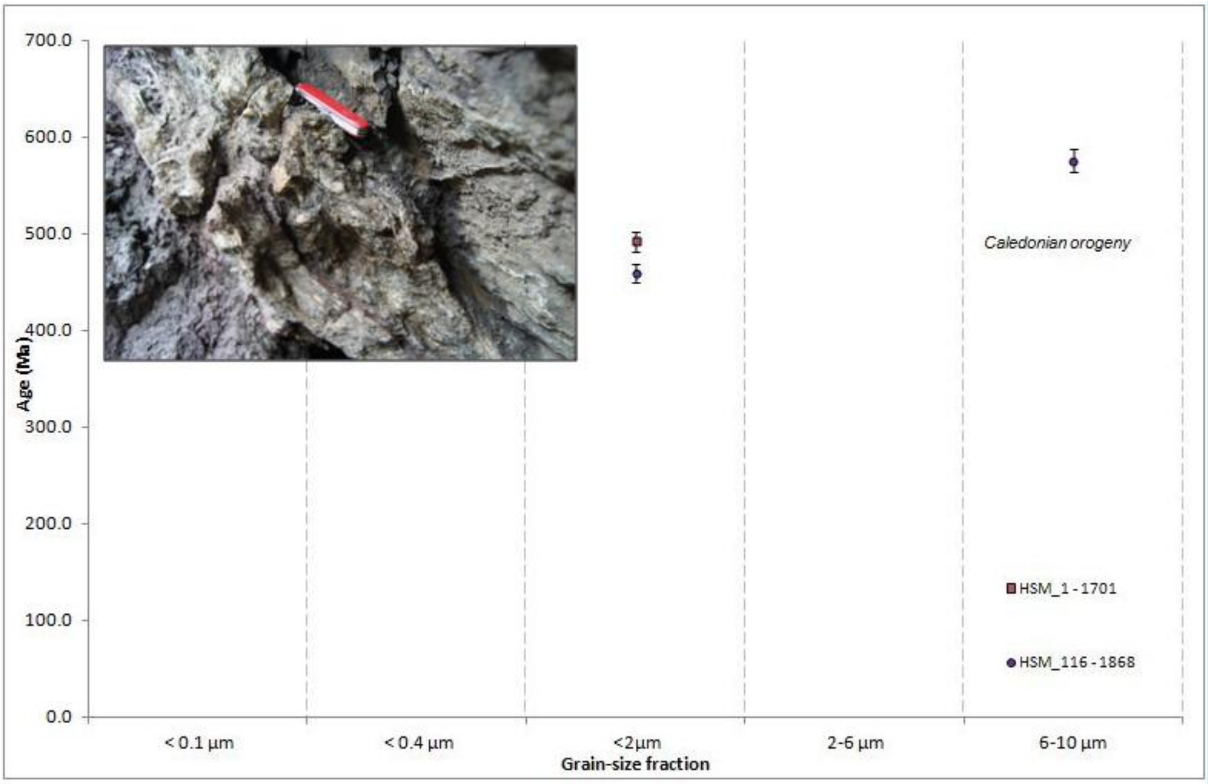


Figure 28 : K-Ar age results from the Bratthammer Thrust. The finest grain size fractions are interpreted as being the closest to the age of the actual faulting event, that is, the latest increment of strain accommodated within the fault rock. The inset shows a close up of the vein with a distinct horizon of cataclastic and fault gouge material. This is the sampling locality of HSM_116.

6.4.2 Discussion

K-Ar age vs. grain size relationships (referred to as "inclined spectra" by Pevear (1999) are often reported for synkinematic and authigenic illites separated from brittle faults (e.g.: (Solum et al. 2005, Lobens et al. 2011, Zwingmann et al. 2011, Pleuger et al. 2012, Viola et al. 2013)). Generally the age decreases with the grain size of the dated fraction. Also, it is well known that separates of illites in shales and fault rocks represent often a physical mixture between inherited detrital components and newly formed illites. As a result, the obtained ages can be devoid of geological meaning. For the K-Ar dating ages to give meaningful information, it is therefore important to consider what assumptions the method relies on. The most important assumption is that it can be reasonably assumed that the system remained close from the authigenesis of the synkinematic clays. No ^{40}K or ^{40}Ar was lost or inherited after illite formation (Zwingmann and Mancktelow 2004).

For neocrystalised illite, the ends of filamentous grains are where we find the finest size fractions, representing the most recently grown illite. Coarser size fractions represent either illite grown earlier in the synkinematic and authigenic crystallisation process, or illite derived from the host rock, that is, a detrital inherited component. They should therefore yield older ages(Zwingmann and Mancktelow 2004, Viola et al. 2013).

In the case of the Brathammer samples, the <2 µm fractions yielded younger ages than the coarser 6-10 µm fraction. Whether this is a trend that would continue and give younger ages for even finer size fractions is not certain from this study, but, comparison with many other published results suggests that it is likely the case. It seems to be evident, therefore, that ~460-490 Ma is the maximum age for the tectonic event that formed the Brathammer thrust.

The two samples are collected from the same structure. However, the <2 µm fractions of the two samples show a difference in age of approximately 30 Ma. This is a statistically significant difference, but there is no obvious indication observed in the field of the two samples being a result of different tectonic events through fault reactivation. The age difference could be a result of the different mineralogy (Table 1) of the two samples, and possibly the slightly different composition of the host rock (especially the presence and abundance of K-bearing phases) at the expense of which the gouge formed.

For any K-Ar illite age to be truly representative of the faulting age, it is crucial that only newly grown authigenic illite is present as the K-bearing phase in the analyzed fraction. Host rock contamination can affect the ages and compromise their interpretation. K-feldspar and amphibole are possible sources of contaminating potassium. As seen in the XRD-data(Table 1), the <2 µm fraction contains very low amounts of K-feldspar for HSM_116 and none for HSM_1. However, in HSM_1 there are trace amounts of amphibole. HSM_116 yields the youngest age of the two samples, and the assumption is therefore that the very low amount of K-feldspar does not affect the age in any significant way. For HSM_1, the amphibole might have increased the age to some extent. The age of 459.6 ± 9.6 Ma for the <2 µm of HSM_116 is therefore assumed to be closest to the age of the actual faulting.

This interpretation relies on the assumption that all the illite formed at the time, or after, the initiation of brittle faulting.

6.4.3 Ore mineralogy of the Bratthammer ore

The ore-vein mineralogy of The Bratthammer vein was investigated by Vokes (1956). He describes the mineralogy to be very simple, containing only pyrite and chalcopyrite and small amounts of calcite as gangue. The pyrite, he describes, appears as rounded to sub-rounded grains in a groundmass of chalcopyrite. Pyrite is the older of the two phases, and the chalcopyrite has partly digested the original pyrite. Magnetite appears as small euhedral crystals in the chalcopyrite.

Observations made from polished thin sections of the mineralized vein in Bratthammeren confirms the conclusions of Vokes(1956). Two samples are investigated, one distinctively foliated, carbonate rich sample, and one rather massive sulphide sample. Three phases are dominant in the ore mineralogy. Pyrite is the main mineral, and makes up more than half of the ore minerals. It often appears as coarse subhedral to euhedral grains. Chalcopyrite is often found associated with the pyrite as overgrowths or as separate smaller grains. It commonly appears in the outer rims of the pyrite and often the chalcopyrite sends long veinlets into the pyrite. Inclusions of magnetite are common in both the chalcopyrite and pyrite(Figure 29).

In the foliated sample, the ore minerals are concentrated along certain horizons. There seems to be a differentiation, where single bands have a higher concentration of magnetite, and others are richer in pyrite and chalcopyrite (Figure 29B). Apart from the ore minerals, the sample consists of more or less exclusively calcite.

In the massive ore, pyrite and chalocypyrite dominates the sample. Clasts of quartz are found within the ore. Fine grained calicite is present in between the sulphides.

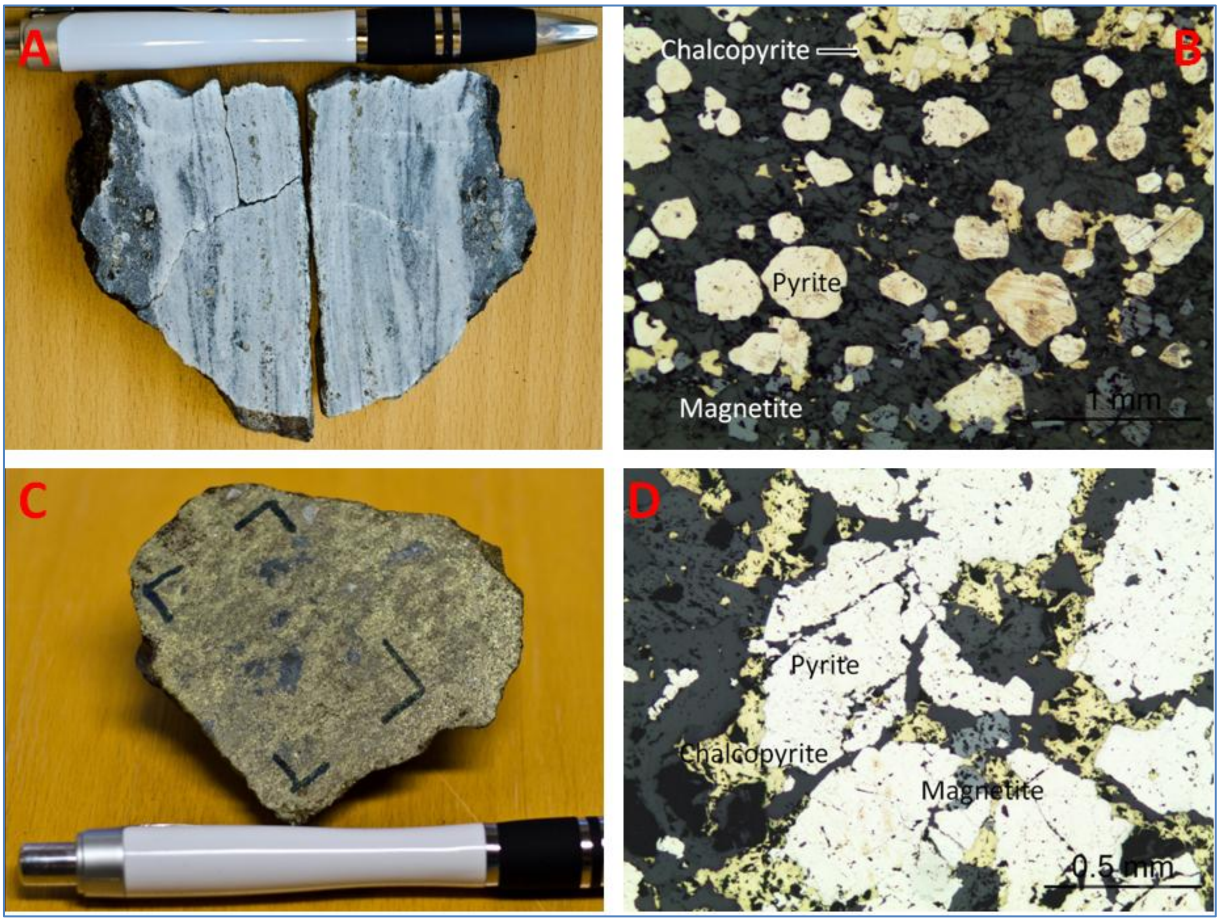


Figure 29 A) Sample from the foliated part of the Bratthammer Vein. B) Microphotograph of the sample in A). Pyrite and chalcopyrite is the dominant ore minerals, wheras magnetite often is found along certain bands. C) Sample of the massive sulphide. D) Microphotograph of the sample in C). The sample consists of mostly pyrite and chalcopyrite, with inclusions of magnetite. Chalcopyrite is commonly found along the grain boundaries of the pyrite grains.

7 Discussion:

7.1 A new model for the Saltvatn Group

By unfolding the antiform in Figure 16 we can easily see how the the stratigraphy is not as straightforward as it has previously been described in this work and by Reitan (1963) and Pharaoh et al. (1983). Figure 30 shows a version of the profile A-B, where the beds are extrapolated and connected. The Dypelv and Stangvatn Formations are located at the same stratigraphic level as the Ulveryggen Formation at the southeastern side of the antiform. Further, the Holmvatn Group and the Nussir Group are at equivalent levels.

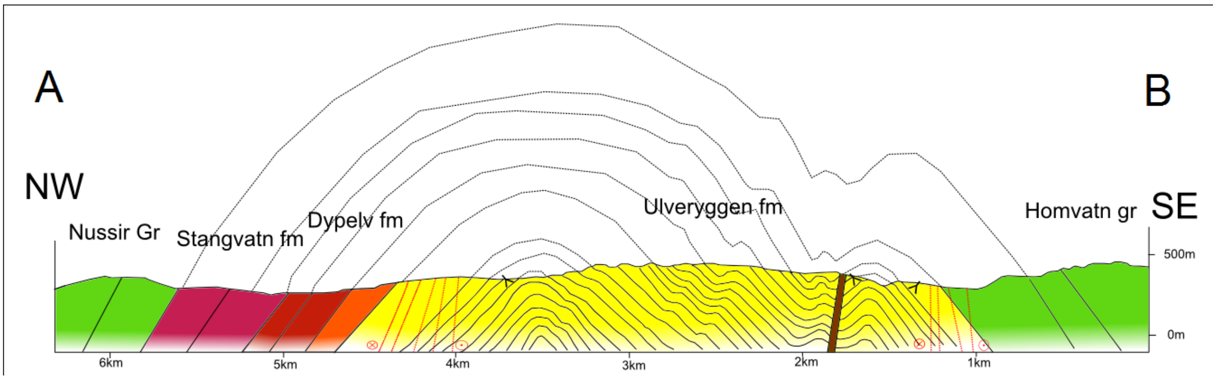


Figure 30: The profile across the Saltvatn Group, where the bedding is extrapolated. As we see, the Dypelv and Stangvatn Formations, are in the same stratigraphic level as the southeastern part of the Ulveryggen Formation. The Nussir Group and the Holmvatn Group are stratigraphical equivalent as well.

The existing depositional model for the Saltvatn Group has obvious shortcomings. As shown in Figure 30 the Dypelv and Stangvatn Formations are at similar stratigraphic level as the southeastern part of the Ulveryggen formation and thus must have been deposited at the same time. The sandstones in the Ulveryggen Formation and the conglomerates in the Dypelv and Stangvatn Formations do not show many similarities. The Ulveryggen sandstone is a homogenous package of quartz rich, fluvial deposited sand, wheras the conglomerates are party extremely coarse and very hetrogenous in composition. Both of the conglomerate formations are rich in volcanoclastic material.

A possible explanation for the different sedimentary deposits could be a model with deposition from a northeasterly direction, where the conglomerates represent a proximal depositional environment and the sandstone a more distal portion. A southwards prograding delta with braided rivers has also previously been suggested by Jensen(1996). An important objection to this model is the compositional differences between the sandstone and the conglomerates. No volcanic material is found within the sandstone, which one would expect if the sandstone were to represent the distal part of a continous depositional sequence with the conglomerates. That no volcanic material at all would be transported to this part of the basin is not very plausible.

More likely, is seems have been deposition from different sources at more or less simultaneous timing. The sandstones in the Ulveryggen Formaiton probably represent a continental provenance, wheras the conglomerates in the Dypelv and Stangvatn Formations could represent sediments derived from a volcanic arc type environment. The provenance for the conglomerates has probably changed during the time of depositon, as there is a clear compositional difference between the Dyplev and Stangvatn Formaitons.

Deposition from two directions and two different sources into the same basin would cause the sediments to meet at a point. Where they meet, some sort of interfingering of the two sedimentart packages would be expected. The contact between the Ulveryggen Formation and the Dypelv formation could very well be a result of this interfingering. The contact is not very well defined, and beds of sandstone and conglomerates are alternating. The contact between the Stangvatn conglomerate and the Ulveryggen sandstone is eroded away everywhere in the window.

A backarc basin, as illustrated in

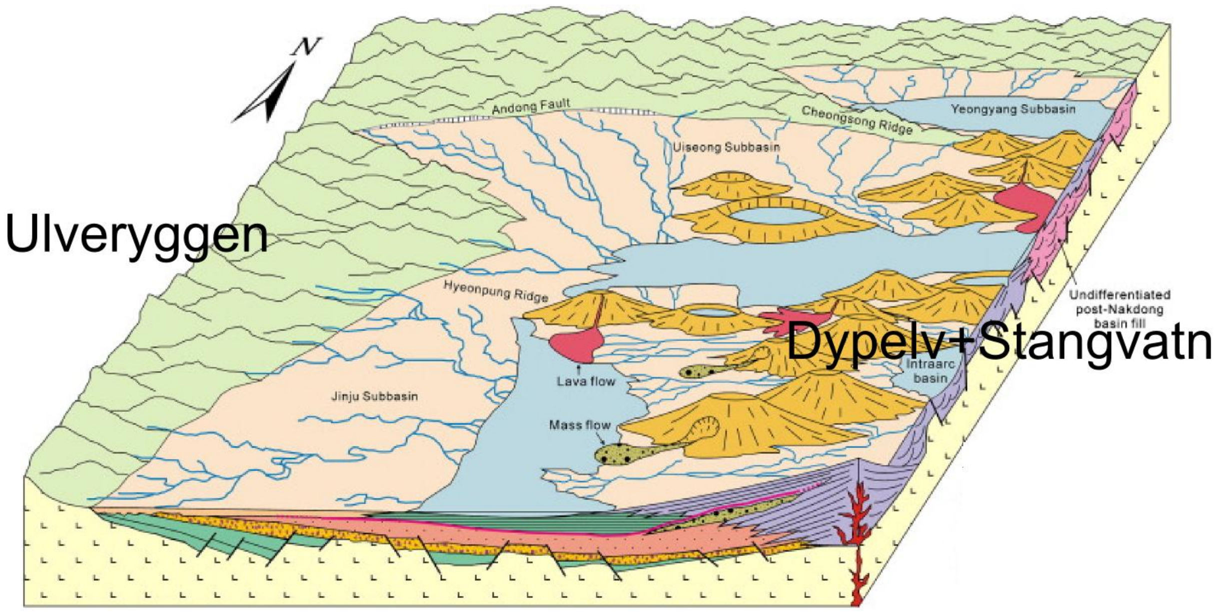


Figure 31, could be a possible model for the depositional environment, where continental derived sediments dominates the southeastern part of the basin, and sediments derived from a volcanic environment dominates in the north western part. The backarc basin model is also compatible with the environment suggested for the Holmvatn Group in chapter 4.3.2.

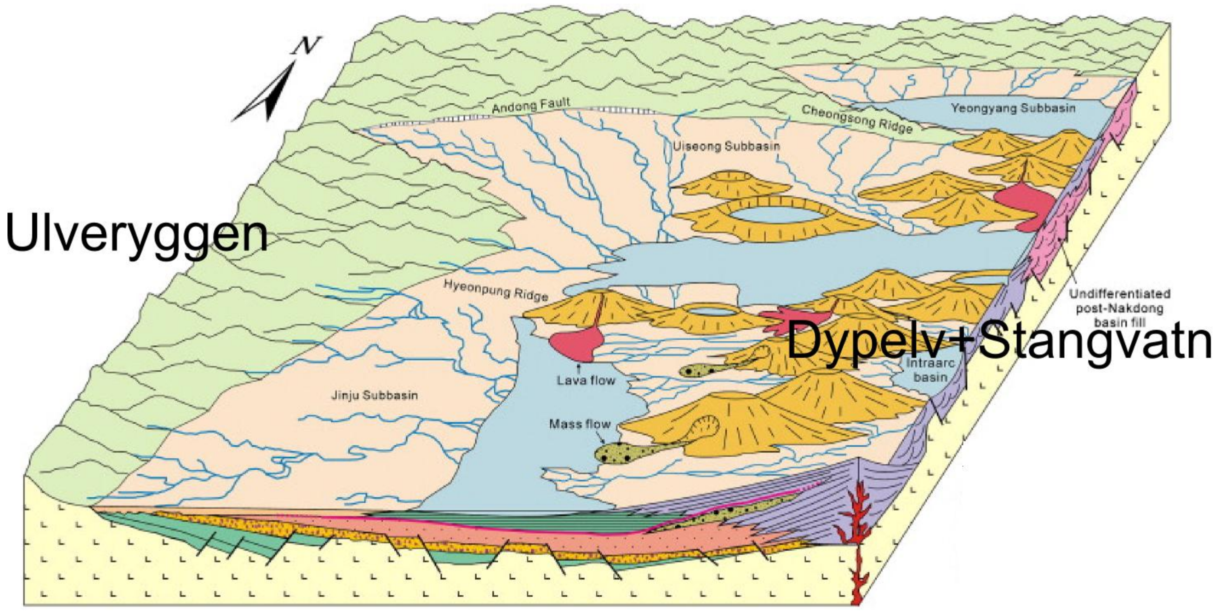


Figure 31: Possible basin model for the Saltvatn Group. A backarc basin with sedimentary input of both continental and volcanic arc provenance. Several minor, local basins with different amounts of input, can also help explaining the differences in thickness throughout the formations. Modified after Chough and Sohn (2010).

7.2 Correlating the Nussir and Holmvatn Groups

An implication of the new stratigraphy is that the Holmvatn Group and the Nussir Group now rests in the same stragigraphic position. Lithological these groups are fairly similar, composed of mostly mafic metavoclanics. Reitan(1963) correlated these two groups within his Holmvann formation. Pharaoh et al. (1983) divided the Holmvatn and Nussir Groups, and stratigraphicaly put them bellow and above the Saltvatn Formation. Based on the new mapping, this division cannot longer be made.

Jensen studied the geochemistry of the volcanic of the RTW and his conclusion is that *"The metavolcanic rocks of basin sequence II* [the Nussir Group] *consists of tholeiitic basalts. The rocks are identical in all important respects with the tholeiites of basin sequence I* [the Holmvatn group]"(Jensen 1996, p. 18). Pharaoh and Pearce (1984) comes to the same conclusion in his geochemical studies of the metavolcanics in the Holmvatn and Nussir Groups. The Nussir Group and their "upper Holmvatn Group"(closest to the contact towards the Saltvatn Group) are both tholeiitic in composition, and seems to be related to within-plate volcanism. The "lower Holmvatn Group" however, has a more calc alkaline composition(Pharaoh and Pearce 1984).

As the two groups are lithological and geochemical similar, and separated by not much more than 10 km(by measuring along the folded beds), they seem to be correlative. A further study of their geochemistry and petrology is needed in order to understand their relationship.

7.3 Folding of the Saltvatn Group: Is it really that old?

The folding of Saltvatn Group has previously been described as an old structure, formed during the Svecokarelian Orogeny at about 1840ma(Pharaoh et al. 1983). The orientation of the fold axis and the axial planes of the fold in a NE-SW orientation indicate that the folding took place in a NW-SE transpressional stress regime. This is similar to the orientation forming the Bratthammer thrust as described in chapter 6.3. The Bratthammer thrust has in this study been dated to be a Caledonian structure. As we seem to have similar oriented stress regimes, could the structures have been formed during the same deformation, and thus making the folding a Caledonian structure?

The folded Saltvatn Group is unconformably overlaid by the Lomvatn Formation at several locations in the RTW. The Lomvatn is outcropping along the contact towards the Caledonian Nappes, and has been interpreted as Neo-proterozoic of age. As the Lomvatn formation overlies the folded Saltvatn Group, this must imply that the folding took place before deposition of the Lomvatn Formation. This rules out the possibility of the folding being a Caledonian structure.

7.4 Tectonic model for the deformation

Skifergangen Shear Zone is oriented in a NE-SW direction. The dextral shear sense indicates the shear zone could have been formed under a WNW-ESE oriented compressional stress regime. As previously described, the Bratthammer thrust seems to have formed during a NW-SE compressional regime. Could the two structures possibly have formed under the same stress regime and same phase of deformation?

Ideally a simple shear zone will form at an angle of ~30 ° to the main stress field, $\sigma_{1.}$ If, however, pre existing structures, weakness zones or foliations, exists, the orientation can deviate from the ideal simple shear orientation, as the friction along existing weaknesses will be lower than the internal friction of the rock(Fossen and Gabrielsen 2005)

In the case of SSZ, the axial planar cleavage from the folding of the central part of the window probably preceded the formation of the shear zone. This has likely caused a certain obliquity in the shear zone from its ideal orientation.

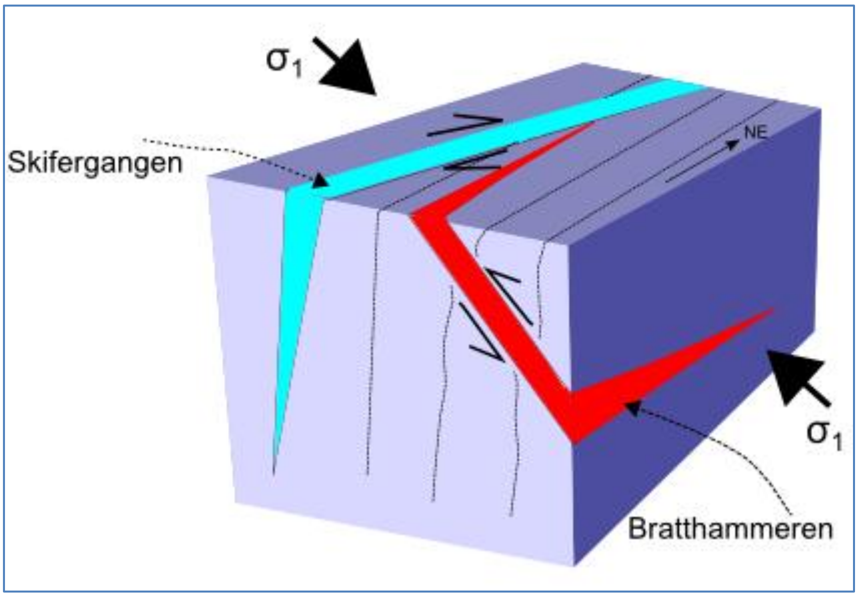


Figure 32: Block diagram showing the orientation of SSZ and Bratthammeren. The small arrows are indicating fault kinematics. Large arrows shows the orientation of the larges compressive stress, σ_1. Black, NE oriented lines illustrates the axial planar foliation

The block diagram in Figure 32 illustrates how the SSZ and Bratthammeren could have formed within the same stress field. Subhorizontal NW-SE orientation of the largest compressive stresses would cause both the thrust and vein opening at Bratthammeren, and can also explain the dextral kinematics at Skifergangen.

Age constraints from the K-Ar dating at Bratthammeren, as described above, gives Caledonian age for the Bratthammer thrust movement. If SSZ is formed under the same deformational regime, this must imply a Caledonian age also for Skifergangen.

Structures from elsewhere in the window has been described by Torgersen et al. (2013). They point out several copper mineralized quartz-carbonate veins around the window which are geometrical and kinematicaly compatible with a phase of NW-SE compression. Three kinematicaly distinctive groups of veins are found(Figure 33). Northwest and southeast dipping veins, related to thrust faults(similar to Bratthammeren), E-W to NE-SW striking veins showing dextral strike slip shearing(similar to SSZ), and NW-SE striking, sub-vertical veins.

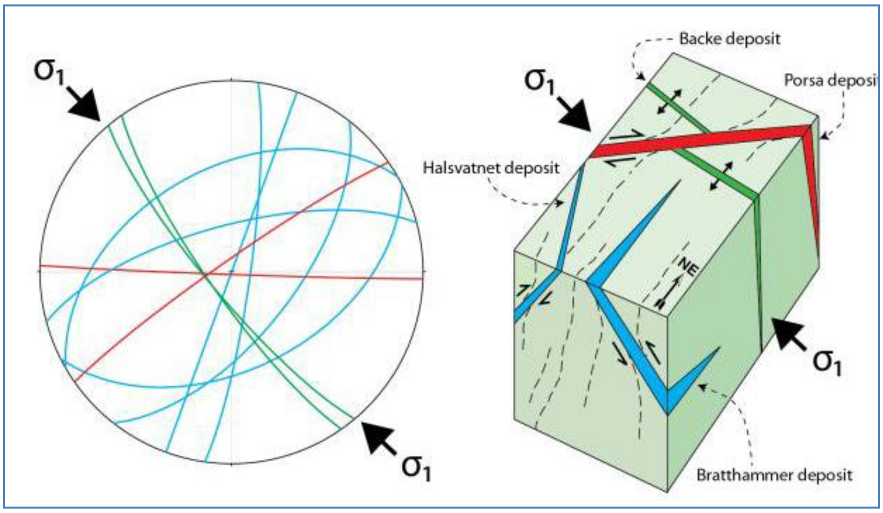


Figure 33: Figure from Torgersen et al. 2013. Compiliation of structural measuremens from Cu-bearing veins in the RTW. The stereonet show the orientation of the mineralized veins. The color codes are similar in both figures.

7.4.1 Ore mineralization

The Bratthammer vein formed as a result of the NW-SE compression. Fluid overpressure probably caused the fracturing of the greenstone hostrock. The subsequent drop in fluid pressure led to precipitation of the quartz and carbonate in the vein together with the ore metals.

The ore mineralogy and petrography in the Bratthammer Vein suggests at least two phases of ore precipitation. The pyrite is the oldest phase with chalcopyrite as a later. The chalcopyrite is partly digesting the pyrite. Whether the precipitation of the ore happened simultaneously with the vein opening or at a later stage is not known at this point. The timing of the thrusting event in relation to the vein opening is not know either. Possibly could some of the ore minerals have precipitated during the event of thrusting as well, if this happened at a later stage. Re-Os dating of the sulphides from the Bratthammer vein in progress will help establish the age of the vein formation.

Mineralizations in Skifergangen are most likely related to fluid flow along the mylonitic shear zone. Precipitation of ore has taken place in pressure shadows within the shear zone and at events of sudden pore pressure drop following fracture or vein openings.

The ore mineralogy in the mineralizations in the SSZ is somewhat different from the mineralogy of the Bratthammer Vein. Only chalocopyrite together with hematite has been observed with certainty. As chalcopyrite is the later mineral in the Bratthammer Vein, a possibility is that the mineralization in SSZ happened later than the pyrite precipitation in the Bratthammer Vein, and at the same time as the precipitation of chalcopyrite. This is although rather speculative.

8 Conclusion

A new straigraphy is established for the lower part of the RTW. The metasedimentary Saltvatn Group is the lowermost unit of the window. Following conformably above it to the north is the metavolcanic Nussir Group and at the south the metavolcanic and metasedimentary Holmvatn group.

The mapping has revealed the folded nature of the Saltvatn Group, and shows that the conglomeatic formations of the Dypelv and Stangvatn Formation are correlative with the upper part of the Ulveryggen Sandstone Formation.

The Bratthammer Vein and the Skifergangen Shear Zone are both compatible with a NW-SE compressional regime. Minor ore mineralizations have taken place related to the deformation events.

K-Ar dating of neocrystalized illite from fault gouge formed during top to the NW thrusting at the Bratthammer Vein has yielded ages at about ~460 Ma and is related to Caledonian deformation.

9 References

Brewer, T. S. and T. C. Pharaoh (1990). Early Precambrian basic rocks of the Baltic Shield. Early Precambrian Basic Magmatism. R. P. Hall and D. J. Hughes, Springer Netherlands: 273-293.
Chough, S. K. and Y. K. Sohn (2010). "Tectonic and sedimentary evolution of a Cretaceous continental arc–backarc system in the Korean peninsula: New view." Earth-Science Reviews **101**(3–4): 225-249.

Dahll, T. (1867). "Om Finnmarkens geologi: Forehandlinger i Videnskabs-Selskabet i Kristiania."

Dallmeyer, R. D., J. G. Mitchell, T. C. Pharaoh, A. Reuter and A. Andresen (1988). "K-Ar and Ar-40/Ar-39 Whole-Rock Ages of Slate Phyllite from Allochthonous Basement and Cover in the Tectonic Windows of Finnmark, Norway - Evaluating the Extent and Timing of Caledonian Tectonothermal Activity." Geological Society of America Bulletin **100**(9): 1493-1501.

Fossen, H. and R. Gabrielsen (2005). Strukturgeologi. Bergen, Fagbokforlaget.

Gaal, G. and R. Gorbatschev (1987). "An Outline of the Precambrian Evolution of the Baltic Shield." Precambrian Research **35**: 15-52.

Haines, S. H. and B. A. van der Pluijm (2008). " Clay quantification and Ar-Ar dating of synthetic and natural gouge – Application to the Miocene Sierra Mazatán detachment fault, Sonora, Mexico." Journal of Structural Geology **30**: 525-538.

Henderson, I., H. C and G. Viola (2011). "Implementering av digital berggrunnskartlegging ved Norges Geologiske Undersøkelse - NGU." Mineralproduksjon **1**: 83-101.

Jacobsen, R. (1989). Kvalsund i Finnmark: bygdebok om Kvalsunds folk og forhold, Kvalsund kommune.

Jensen, P. (1996). "The Altenes and Repparfjord tectonic windows, Finnmark, northern Norway: Remnants of a Paleoproterozoic Andean-type margin at the rim of the Baltic Shieldt."

Lobens, S., F. A. Bense, K. Wemmer, I. Dunkl, C. H. Costa, P. Layer and S. Siegesmund (2011). "Exhumation and uplift of the Sierras Pampeanas: preliminary implications from K-Ar fault gouge dating and low-T thermochronology in the Sierra de Comechingones (Argentina)." International Journal of Earth Sciences **100**(2-3): 671-694.

Merriman, R. J. and S. J. Kemp (1996). "Clay minerals andsedimentary basin maturity." Mineral. Soc. Bull **111**: 7-8.

NGU.no. (2013). from http://www.ngu.no/no/tm/Om-NGU/Prosjekter/Mineralressurser-i-Nord-Norge-MINN/.

Nilsson, L. P. and K. Nilsen (1996). VARGSUND Berggrunnskart 1935 4, 1:50000, Norges geologiske undersøkelse.

Nironen, N. (1997). "The Svecofennian Orogen: A tectonic model." Precambrian Research **86**: 21-33.

Pevear, D. R. (1999). "Illite and hydrocarbon exploration." Proceedings of the National Academy of Sciences of the United States of America **96**(7): 3440-3446.

Pharaoh, T., D. Ramsay and O. Jansen (1983). "Stratigraphy and Structure of the Northern Part of the Repparfjord-Komagfjord Window, Finnmark, Northern Norway." Norges Geologiske Undersokelse Bulletin(68): 1-&.

Pharaoh, T. C. (1985). The Stratigraphy and Sedimentology of Autochthonous Metasediments in the Repparfjord-Komagfjord Tectonic Window, West Finnmark: 1985 Pp347-357 In: The Caledonide Orogen - Scandinavia and Related Areas. Eds: Gee, DG & Sturt, BA, John Wiley & Sons.

Pharaoh, T. C. and T. S. Brewer (1990). "Spatial and Temporal Diversity of Early Proterozoic Volcanic Sequences - Comparisons between the Baltic and Laurentian Shields." Precambrian Research **47**(3-4): 169-189.

Pharaoh, T. C., R. M. Macintyre and D. M. Ramsay (1982). "K-Ar Age-Determinations on the Raipas Suite in the Komagfjord Window, Northern Norway." Norsk Geologisk Tidsskrift **62**(1): 51-57.

Pharaoh, T. C. and J. A. Pearce (1984). "Geochemical Evidence for the Geotectonic Setting of Early Proterozoic Metavolcanic Sequences in Lapland." Precambrian Research **25**(1-3): 283-308.

Pleuger, J., N. Mancktelow, H. Zwingmann and M. Manser (2012). "K-Ar dating of synkinematic clay gouges from Neoalpine faults of the Central, Western and Eastern Alps." Tectonophysics **550**: 1-16.

Reitan, P. (1963). "The geolgy of the Komagfjord tectonic window of the Raipas suite, Finnmark, Norway."

Reusch, H. (1891). Det nordlige Norges geologi.

Rice, A. H. N. (1998). "Stretching lineations and structural evolution of the Kalak Nappe Complex (Middle Allochthon) in the Repparfjord-Faegfjord area, Finnmark, northern Norway." Norsk Geologisk Tidsskrift **78**(4): 277-289.

Solum, J. G., B. A. van der Pluijm and D. R. Peacor (2005). "Neocrystallization, fabrics and age of clay minerals from an exposure of the Moab Fault, Utah." Journal of Structural Geology **27**(9): 1563-1576.

Torgersen, E., G. Viola, J. S. Sandstad and H. Stein (2013). "Structural constraints on the formation of Cu-rich mesothermal vein deposits in the Repparfjord Tectonic Window, Northern Norway."

Townsend, C. (1987). "Thrust Transport Directions and Thrust Sheet Restoration in the Caledonides of Finnmark, North Norway." Journal of Structural Geology **9**(3): 345-352.

Viola, G., J. S. Sandstad, L. P. Nilsson and B. Heincke (2008). "Structural and ore geological studies in the northwestern part of the Repparfjord Window, Kvalsund, Finnmark, Norway." NGU Report **2008.029**.

Viola, G., H. Zwingmann, J. Mattila and A. Kapyaho (2013). "K-Ar illite age constraints on the Proterozoic formation and reactivation history of a brittle fault in Fennoscandia." Terra Nova **25**(3): 236-244.

Vokes, F. M. (1956). "Some copper sulphide parageneses from the Raipas Formation of Northern Norway." Norges Geologiske Undersøkelse: 74-111.

Winter, J. D. N. (2010). An Introduction to Igneous and Metamorphic Petrology, Prentice Hall PTR.

Zwingmann, H., R. Han and J. H. Ree (2011). "Cretaceous reactivation of the Deokpori Thrust, Taebaeksan Basin, South Korea, constrained by K-Ar dating of clayey fault gouge." Tectonics **30**.

Zwingmann, H. and N. Mancktelow (2004). "Timing of Alpine fault gouges." Earth and Planetary Science Letters **223**(3-4): 415-425.

Zwingmann, Z., H., N. Mancktelow, M. Antognini and R. Lucchini (2010). "Dating of shallow faults: New constraints from the AlpTransit tunnel site (Switzerland)." Geology **38**(6): 487-490.

Appendix 1

K-Ar data from Bratthammeren:

Sample nr.	***K [%]***	***Rad. 40Ar [mol/g]***	***Rad. 40Ar [%]***	***Age [Ma]***	***Error [Ma]***	***Error [%]***
HSM_1 < 0.1 µm						
HSM_1 < 0.4 µm						
HSM_1 <2µm	0.60	5.9184E-10	61.51	**492.2**	10.8	2.2
HSM_1 2-6 µm						
HSM_1 6-10 µm						
HSM_116 < 0.1µm						
HSM_116 <0.4µm						
HSM_116 <2µm	2.960	2.6872E-09	83.02	**459.6**	9.6	2.1
HSM_116 2-6 µm						
HSM_116 6-10 µm	3.33	3.9178E-09	90.08	**575.8**	11.8	2.1