Conodonts and depositional environment of the Middle and Upper Cambrian Alum Shale, Slemmestad, Oslo Region

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Master Thesis in Geosciences

Discipline: Geology

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01.06.2014
Forewords
This master thesis has not only made me a geologist, it has also increased my passion for geosciences. During this five year long master program, I also met my husband at the Geology building in my first weeks as a student, which I married 21.04.2012, and had a wonderful purple Amethyst-theme wedding. I also got my dearest baby girl during this education, which came to the world 03.02.13, when I was supposed to be in class. This master thesis was written in the time between August 2013 and June 2014, but the field work and preparation for the thesis started in March 2013. By having a full semester during the spring 2013, with my baby girl born February, this would never have been possible if it was not for my exceptionally supportive, understanding and helpful supervisor Hans Arne Nakrem who accommodated every class as well as examination dates. He has also been supporting and helpful and has given excellent supervision despite the long distance when I moved to Bergen with my family in December 2013. You are a wonderful person and I couldn’t have had a better supervisor!

I sincerely want to thank Johan Petter Nystuen and Krzysztof Hryniewicz for helping me with thin section analysis, and Harald Folvik and Hans Jørgen Berg for helping me with SEM-analysis at NHM, and Gunborg Bye Fjeld for helping me during heavy liquid separations. I am also very grateful for the help by Magne Høyberget during field work and for being helpful answering questions and David Bruton who showed interest and enthusiasm for this thesis. I would also like to thank Bjørn Funke for giving me some of his collected material for this research, and Berit Løken Berg for helping me with SEM-analysis at Blindern. A big thank to Salahalldin Akhavan for preparing my thin section. And a special thank to Svend Stouge for helping identify the conodonts, and to teach me a lot about conodonts.

I want to thank my supportive, helpful and positive dear friends especially Camilla Rytterager Henriksen, who have helped me babysitting, and took good care of my baby when I was at the laboratory when my husband was at work. I would never have finished this master thesis at time if it was not for your help! I am forever grateful. Of course I want to thank my fellow students, especially Christopher Kjølstad, Martin Sandbakken and Orhan Mahmic, for making these years a wonderful time. I’m going to miss all the coffee breaks and laughter at “Steinrommet”. This room, U39C at Blindern, will always have a special place in my heart. I would also like to thank my family in Bergen who always have been supportive and motivated me, and for babysitting my daughter during the weekends so I could work on my thesis.

Last but not least, I would like to thank my geologist husband, who always have been supportive, helpful and a wonderful father. Thank you for all the help and patient and for all the hours you have spent at NHM and Blindern with me so I could have been around my baby despite all the work I had to do. I could never have done this without your help and support. And so, to my dearest daughter, who I always have had a bad conscience for when not being present: From now, I will ALWAYS pay you all attention you want, and give you everything you want (yes, you can use this against me when you are a teenager).


Katarina Skagestad Kleppe
Abstract

The bituminous Cambrian and lowermost Ordovician Alum Shale from Slemmestad in the Oslo Region, Norway, is for the first time investigated for conodonts and other microfossils. Microfacies analysis is also done based on thin section analysis. This thesis is done in order to increase the understanding of the Alum Shale and the Cambrian fauna.

Nine samples were taken from limestone-rich levels ranging from the Middle Cambrian Paradoxides paradoxissimus trilobite zone to the Lower Ordovician Boeckaspis trilobite zone. The samples were dissolved in acetic acid and the acid resistant residue was studied for biogenic material using microscope and SEM. The acid resistant residue from 63µm – 500µm was heavy liquid separated in order to extract conodonts. Depositional environment interpretation was done based on microfacies analysis and microfossils present in acid resistant material.

Conodonts were present in five of the samples. Species recorded are all, except Cordylodus proavus, previously reported from age equivalent deposits in Sweden. The identified conodont species are Phakelodus tenuis, Phakelodus elongatus, Westergaardodina polymorpha, Westergaardodina ligula, Problematoconites perforatus, Trolmenia acies and Cordylodus proavus. All the conodont faunas represent the cold water realm. The presence of Cordylodus proavus may be regarded as its first occurrence in Scandinavia.

From the thin section analysis five different facies is identified, representing both high and low energy depositional conditions, with an overall upward deepening trend containing sea-level fluctuations. In one of the facies trace fossils from the ichnogenus Phacosiphon is present. Microfossils of environmental interpretation importance found in the samples are phosphatocopine ostracods, inarticulate brachiopods and fecal pellets.
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1 Introduction

1.1 General introduction

The Cambrian to lowermost Ordovician Alum Shale exposed in the village of Slemmestad (figure 1), SW of Oslo, is well known primarily for its rich fossil fauna dominated by olenid trilobites, and has been studied by several paleontologists and geologists since Brøgger in 1880. How the Alum Shale Formation was formed, as well as biostratigraphical correlation based on trilobites has been of interests for a long time. The most substantial work in this respect is the systematic treatment of trilobites by Henningsmoen (1957), which through several stages of amendments has resulted in the current accepted stratigraphical scheme (Nielsen et. al., 2014). The Alum Shale has a high concentration of organic carbon, which makes this a good source rock when exposed to right temperatures. However, the Alum Shale in the Oslo area has been exposed to too high temperatures due to Permian intrusion (Figure 1). Even though the Alum Shale at Slemmestad is not a source rock, it is indeed a source of information regarding the Cambrian fauna and depositional environment.

![Figure 1. Photo showing Cambrian Alum Shale between Precambrian basement and a Permian sill, in the village of Slemmestad.](image)

One of the faunal contributors in the Alum Shale Sea during the Cambrian was conodonts. The only conodont investigation from the Cambrian Alum Shale in Norway was done by Bruton et. al. (1988) at Nærnes beach nearby Slemmestad. Hence Norwegian Cambrian conodonts are a rather unexplored topic relative to other Cambrian faunal components like trilobites.

During a project in 2006 two pilot samples were taken from the Middle Cambrian (GIBB06) and from the Upper Cambrian (PEL06) in Slemmestad (Pers. Comm. 2014). The samples contained conodonts. The findings of conodonts in these pilot samples supported a further research on Cambrian conodonts from these deposits in Slemmestad.
Conodonts were small eel-like animals known from small phosphatic teeth-like elements from their feeding apparatus, known as conodont elements. Conodonts are widely used for biostratigraphy, and they are also used for paleoecological and biogeographical studies. They may also provide information regarding basin history, regional metamorphism and state of hydrocarbon generation.

Conodonts from Cambrian Alum Shale outside the Oslo Region are well known (Müller, 1959; Szaniawski, 1971; 1987; Bednarczyk, 1979; Andres, 1981; 1988; Borovko and Sergeyeva, 1985; Kaljo et al., 1986; Viira, et al. 1987; Müller and Hinz, 1991; 1998; Hinz, 1992; Mens et. al. 1993; 1996; Szaniawski and Bengtson 1993; 1998; Bagnoli and Stouge, 2013). The conodonts were studied for taxonomy, histology, for providing zonal schemes, and for conodont associations.

1.2 Purpose of study
The purpose of this study is to investigate if the microfossil assemblages, as well as microfacies analysis from the Alum Shale in Slemmestad may provide information regarding the depositional environment, as well as whether the conodonts found are of biostratigraphical importance. Another aim is also to investigate if there is a correlation between the different facies and conodont faunas, as well as to contribute to the understanding of the faunal composition in the Cambrian Alum Shale of this part of the Oslo area.

Samples collected during field work represent different levels primarily through the Upper Cambrian. Limestone-rich intervals were selected for sampling, thin sections were made, and the samples were dissolved in acetic acid. The acid insoluble residue was heavy liquid separated for further investigation using optical microscope and scanning electron microscope.

Hopefully, the interpretations and conclusions from this thesis may contribute to the knowledge regarding the environment during deposition of the Alum Shale in Slemmestad, and hopefully give information regarding the conodont fauna.
2 Geological background

2.1 Regional geology

The Cambrian period lasted for 55.6 million years (541-485.4 Ma), and is the first period in the Paleozoic Era (Peng et. al., 2012). This period is important in the history of life on earth, and presents one of the greatest evolutionary events in the Earth’s history; the Cambrian Explosion (Waggoner and Collins, 1994).

The Cambrian stratigraphic sequence in Norway occurs locally as allochtonous or autochtonous layers in or along the lower Caledonian nappe units (Nielsen and Schovsbo, 2011). In Oslo region the Cambrian succession is recognized with sedimentary layers of dark bituminous shale interacting with limestone layers, also known as the Alum Shale Formation (Buchardt et.al., 1997).

The paleocontinent Baltica was located at 45-60 degrees south during Cambrian time and included areas where Norway, Sweden, Denmark, Russia, and the Baltic countries are located today (Torsvik and Rehnström, 2001). As seen in figure 2, Baltica was surrounded by the Ægir Sea and The Iapetus Ocean during the Late Cambrian. The term Baltoscandia is used for the part of Baltica including Norway, Sweden and Denmark.

Figure 2. Distribution of the paleocontinents on the southern hemisphere during the Late Cambrian (Torsvik and Rehnström, 2001).
The Cambrian period is divided into global series and stages. As shown in Figure 3, the global series represents the Lower (Terreneuvian), Middle (Series 2 and 3) and Upper Cambrian (Furongian). The stages are further subdivided into trilobite zones and subzones (see Figure16, section 3.1.1). The main global series of interest for this study is the Furongian lasting from 497-485.4 Ma (Peng et al., 2012). Uppermost Middle Cambrian (series 3) and lowermost Tremadocian (earliest Ordovician) are also of interest.

![Figure 3. The Cambrian global time scale (Peng et al., 2012)](image)

2.1.1 The Alum Shale
The Alum Shale was formed on the present western and southern part of Baltica (Buchardt et al., 1997), and includes strata from Middle Cambrian (Series 3), to close to the top of the Lower Ordovician Tremadocian Series (Høyberget and Bruton, 2012). The formation is present throughout much of Baltoscandia, and the “Alum Shale Sea” covered areas from western Norway to St. Petersburg in the east and from Poland in the south to Finnmark in northern Norway at its maximum extent (Buchardt et al., 1997). The Alum Shale Formation is
the term used for the whole lithostratigraphic unit throughout Scandinavia (Nielsen and Schovsbo, 2007).

The Alum Shale Formation appears to be uniform over a large area, with sedimentation rates as low as 1mm per 1000 years (Bjørlykke, 1974). It consists of bituminous brown to black shales and mudstones with alternating limestone- and siltstone beds, and the type section is defined in the Gislövshammar-2 core, from southern Sweden (Buchardt et. al., 1997). It is finely laminated, and bioturbation is not present except from some horizons at the lower and upper part (Nielsen and Schovsbo, 2011). Trilobites are almost always absent in the shale, and it is rich in organic carbon suggesting anoxic conditions (Thickpenny, 1984). Bituminous limestone concretions (anthraconites) occur as discontinuous to semi-continuous lenses throughout the entire formation (Thickpenny, 1984).

The Alum Shale is characterized by its high content of organic matter and trace elements, mainly uranium and vanadium (Bergström and Gee, 1985). In addition it is well known for its rich fossil fauna, dominated by agnostid and olenid trilobites in the limestone rich layers (Buchardt et. al., 1997). In the Oslo Region the Furongian Alum Shale itself is usually unfossiliferous, but the anthraconite concretions can be extremely fossiliferous, dominated by olenid trilobites (Høyberget and Bruton, 2012).

The base of the Alum Shale Formation is progressively getting older when moving from the east towards the west. In the southern and western part of Baltoscandia, the Alum Shale first appears in the early Middle Cambrian, where it overlays lower Cambrian sand- and silt deposits, or lays directly on top of Precambrian continental basement (Thickpenny, 1984). In southwestern part of Sweden it first appears during middle Mid-Cambrian, while it first appears during Late Cambrian in eastern part of Sweden and Poland. In Estonia, it first appears during Tremadocian. This evolution reflects a sea level rise which with time covered large areas of the Baltic Shield and thereby led to the deposition of mud on the shelf (Nielsen and Schovsbo, 2011).

The formation of the anthraconites has been explained as the remnants of a dissolved continuous limestone bed (Bjørlykke, 1973), and as early stage concretions (Henningsmoen, 1974). According to Thickpenny (1984), the formation of the anthraconites is similar to the explanation of early formed diagenetic concretion of Raiswell (1971). This explanation suggests that the concretions is formed by nucleation on fossiliferous layers, probably on the sea floor, growing during early stages of compaction, hence not the remnants of a dissolved limestone bed. Intra-basinal heights on the shelf that penetrated the anoxic-oxic boundary in
the water column are suggested as starting points for the formation of the concretions (Thickpenny, 1984). This penetration may have allowed trilobite faunas adapted to such environment environment to colonize (Figure 4), resulting in the fossiliferous concretion despite the surrounding unfertile shale (Henningsmoen, 1957). The anthraconites consist of micritic to coarse sparitic calcite with content of pyrite (Dworatzek, 1987). The micritic and fine sparitic anthraconites consist of dark grey to black calcite with a high content of clay particles and organic material impurities. These anthraconites have no structures, but may show some lamination from the clay matrix they grew in, as relic laminations (Buchardt et al., 1997). The grain size in central parts of the concretions are commonly of arenitic grain size (Thickpenny, 1984), which include a size range from 0.0625mm – 2 mm (Encyclopedia Britannica, 2013). In thin-sections the carbonate primarily consists of rounded sand-sized grains of random orientation in a poorly laminated matrix (Thickpenny, 1984). The coarse sparitic anthraconites consist of grey to brown calcite crystals which may be up to 10cm in length, and this form of anthrachonite may account for 0% to 100% of a concretion (Buchardt et al., 1997).

Figure 4. Illustration of intra-basinal heights penetrating the anoxic-oxic boundary, allowing trilobite colonization.

The Alum Shale Formation is over- and underlain by shallow marine deposits over the entire basinal area (Thickpenny, 1984). Little variation in the lithology of these deposits may suggest that Alum Shale also is deposited in shallow water (< 200m). In shallow water, stagnation away from the open ocean may occur (Thickpenny, 1984). The constant lithology throughout the Alum Shale Formation, and the surrounding lithology, suggests that this formation was formed by shallow marine deposits (Figure 5) under such stagnating conditions (Thickpenny, 1984). This resulted in anoxic conditions favoring preservation of organic matter (Nielsen, 2004).
Slow sedimentation rates in shallow water environment, and restricted detrial supply, probably reflects the high sea level during this time (Thickpenny, 1984). Rareness of redeposited sediments reflects a gentle topography on the sea floor, and hence, the sediments have been deposited from suspension, but based on the concretions, topography on the sea floor must have been significant (Thickpenny, 1984).

The thickness of the formation varies from less than 1m near the edge of the Baltic syncline, to over 130m in Kattegat (Buchardt et. al., 1997) (Figure 6). These variations reflect the structural differences in the southern part of Baltoscandia. In the Oslo area, the thicker part seems to correspond to the Oslo Graben. The shale decreases in thickness towards east and north in Sweden, which most likely reflects the depositional environment, while the thinner part of the formation towards the eastern part of the Baltic syncline is due to erosion. The difference in thickness of the shale throughout the formation is due to the different facies environment on the Baltic Shield, which are condensed facies and the shelf facies. The latter is typical for the areas in southern Norway among others (Buchardt et. al., 1997). On the platform, the shale is rarely over 25m in thickness, and is characterized with a high content of digenetic formed limestone (up to 50%) often as beds and the shale has abundant hiatuses. The shale near the paleoshelf on the other hand, is thicker in general, and consists of less than 10% limestone occurring primarily as concretions or lenses (Buchardt et. al., 1997).
Figure 6. Variation in thickness (m) of the Alum Shale Formation in southern Baltoscandia (modified from Buchardt et. al., 1997).

Figure 7 shows the lithostratigraphic setting of the Alum Shale Formation in the Oslo Region. The figure also includes estimated thickness as well as the shallow deposited sediments from Pre Cambrian, Lower Cambrian and Lower Ordovician.

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<th>Chronostratigraphy</th>
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<td>Cambrian (pars)</td>
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<td>Series 3</td>
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<td>late Mesoproterozoic to early Neo-proterozoic</td>
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Figure 7. Lithostratigraphic setting of the Cambrian and Lower Ordovician sediments in the Oslo Region (modified from Calner et.al., 2013).
2.1.2 Paleogeography and paleoclimate
During the Cambrian, the paleocontinents were located on the southern hemisphere, and due to the fragmentation of the Proterozoic supercontinent Rodinia, the landmasses were scattered (Waggoner and Collins, 1994). As shown in Figure 2, Baltica is estimated to have been located between 45°-60° on the southern hemisphere (Torsvik and Rehnström, 2001).

![Figure 8. Global sea level and temperature changes during Cambrian and Ordovician (Modified from Dudley, 2000).](image1)

The Cambrian world was bracketed between the late Proterozoic and the Ordovician Ice Age. The temperature was higher and more stable than today, causing retreatment of the Proterozoic ice (Waggoner and Collins, 1994). This led to higher sea levels (Figure 8), and most of the lowland areas such as Baltica were covered with shallow epicontinental seas (Waggoner and Collins, 1994), and epeiric platforms covered large areas (Figure 9)(Boggs, 2006).

![Figure 9. An epeiric platform, characteristic for flooded continental shelves (modified from Boggs, 2006).](image2)

The overall higher temperature during the Cambrian caused a higher rate of evaporation. This led to an elevated salinity in the shallow oceans which resulted in density contrasts in the water column (Jenkins et.al., 2012). This density induced layering of the water column led to
stagnation of the in the epicontinental seas. Since no oxygen rich surface water was able to descend towards the bottom, the water near the bottom became progressively more anoxic due to oxygen consuming bacteria (Bjørlykke, 2004). These conditions allowed the deposition of the Alum Shale (Figure 10).

As illustrated in Figure 11, the oxygen level during Cambrian was lower than today (Dudley, 2000), but during this time oxygen was for the first time mixed into the oceans in significant amount (Waggoner and Collins, 1994). During this period the number of oxygen-depleting bacteria was reduced, which made dissolved oxygen available to the diversity of animals. This was probably the foundation of the “Cambrian Explosion” (Waggoner and Collins, 1994).
2.1.3 Tectonics

Baltica was attached to the Proterozoic continent Rodinia in Precambrian, but was separated from this continent during late Precambrian time (Torsvik and Cocks, 2005). Baltica was a separated continent until Silurian time, when it collided with the continents Laurentia and Avalonia (Torsvik and Cocks, 2005).

The Caledonian Orogeny was initiated during the Late Ordovician (Liu et. al., 2010) as a result of the closure of the Iapetus Ocean and Tornquist Sea (Buchardt et. al., 1997). This led to deformation and folding of the shelf areas south and west of the Baltic Shield, but the deposited sediments on the shelf were practically unaffected (Buchardt et. al., 1997). The orogenic event strongly affected the Lower Paleozoic deposits in the Oslo area. This deformation, with Alum Shale working as thrust plane, led to shortening of the Lower Paleozoic sequence in the Oslo-Asker region, due to folding, faulting and thrusting (Bruton and Owen, 1982). A foreland basin was developed along the margin of the Caledonides on the Baltic Shield (Buchardt et. al., 1997). This has led to foreland-basin type structural deformations in the Oslo-Asker area (Figure 12). The Carboniferous-Permian extensional rifting of the supercontinent Pangaea led to exposure, hence erosion of the Upper Paleozoic deposits along the Baltic Shield (Buchardt et. al., 1997).

![Diagram](image)

Figure 12. Illustration of the development of the foreland basin due to the Caledonian orogenic event, with the Alum Shale working as a thrust plane (Bjørlykke, 1983).
2.2 The Oslo Region
The Oslo Region is located within a graben structure, formed during the Carboniferous-Permian extensional rifting (Neumann et. al., 2004), and is well known for its variety of rocks. The rocks present in the Oslo Region ranges from Lower Paleozoic deposits and Upper Carboniferous sediments, as well as igneous rocks of Late Carboniferous to Permian age (Ramberg et. al., 2010).

The Oslo Region extends a distance of about 200 km north and south of Oslo starting from Langesundsfjorden to the northernmost part of Mjøsa district (Figure 13). The width varies from 35 to 65 km and is bordered by major normal fault zones to the east (Neumann et. al., 2004; Ramberg et. al., 2010).

Due to the graben-structure Lower Paleozoic deposits are preserved in the Oslo Region, and the Alum Shale is common throughout the area (Buchardt et. al., 1997). Post-rifting, the Lower Paleozoic deposits were covered by erosion material from the surrounding horst area and by volcanic and magmatic rocks (Andersen, 1998). The lower Paleozoic deposits in the northern part of the Oslo Graben are strongly deformed and folded due to the Caledonian event, while the southern part is strongly affected by Permian magmatism (Buchardt et. al., 1997).
The Alum Shale has worked as a thrust plane for the lower Caledonian nappe units and is overall deformed and thermally altered (Bruton and Owen, 1982).

2.3 Local geology in the Slemmestad area
The lower Paleozoic succession in Slemmestad, which is located approximately in the middle part of the Oslo-graben, is strongly deformed and folded due to the Caledonian orogenic event. The Alum Shale Fm. in Slemmestad is exposed in several localities (Figure 14).

Figure 14. A) Geological map of the Slemmestad area (modified from NGU geological map). B) Location of Slemmestad is marked on a regional map (google maps).
3 Paleontology

During the Cambrian period, life on earth went through extreme changes from very primitive animals during the Precambrian to relatively advanced animals as well as the evolution of the first known vertebrates (Benton and Harper, 2009). Almost every metazoan phylum with hard parts, evolved during this period. This evolution of life, the "Cambrian Explosion”, is one of the greatest evolutionary events in the history of life on Earth (Waggoner and Collins, 1994).

The fossil fauna not only provides important information regarding the evolution of life, but also important information about the depositional environment including water depth, current directions, and sedimentation rates. In addition the fossil fauna can provide information on temperature, salinity, as well as the thermal maturation of the fossil hosting sediments (Armstrong and Brasier, 2005).

The fauna in the Cambrian (Figure 15) was dominated by arthropods, with trilobites as the most abundant group. Brachiopods, mollusks, echinoderms, sponges, jawless vertebrates were also a part of the Cambrian fauna (Benton and Harper, 2009).

![Figure 15. Artistic illustration of the Cambrian fauna in Burgess Shale (Pitman, 2014)](image)

The fossil fauna of the Cambrian Alum Shale is dominated by agnostid and olenid trilobites. Brachiopods, phosphatized ostracods and conodonts among other less abundant organisms are also present (Buchardt et. al., 1997; Szaniawski and Bengtson, 1998). The fauna in the Upper
Cambrian Alum Shale has pelagic organisms, which differs from the benthic fauna of Middle Cambrian and earliest Ordovician (Müller and Hinz, 1991). In the Tremadocian graptolites occur, and defines the transition between Cambrian and Ordovician with the index fossil of *Rhabdinopora flabelliforme* (Buchardt *et. al.*, 1997; Landing *et. al.*, 2000).

The Cambrian conodont fauna was dominated by protoconodonts and paraconodonts, since the euconodonts first appeared in the Late Cambrian (Armstrong and Brasier, 2005). For more details regarding Cambrian conodonts, see chapter 5.

Cambrian conodont studies have been used for stratigraphy, phylogeny and evolution, morphology, histology and function, systematic position, facies, provincialism, temperature control, geochemistry and chemoevolution (Müller and Hinz, 1991). Based on this, as well Color Alteration Index, the conodonts may provide information regarding the environmental conditions during deposition, as well as the maturation history of the surrounding sediments, which is of interest for source rock studies (Armstrong and Brasier, 2005).

This chapter presents previous work on Cambrian conodonts regarding stratigraphy and faunal studies. Due to the correlation between conodont zones and trilobite zones, trilobite groups relevant as biostratigraphic and depositional indicators are also mentioned. Conodont morphology, paleoecology and taphonomy are described in chapter 5. Microfacies analysis, as well as other microfossil groups present in the Alum Shale Fm. is presented in chapter 6.

### 3.1 Biostratigraphy

Trilobites dominated the Cambrian fauna, especially the dysoxic environments, in addition they evolved rapidly during this period. Hence they are commonly used as biostratigraphical indicators in Cambrian black shales (Buchardt *et. al.*, 1997). Cambrian conodonts are also used for biostratigraphy, but are less precise time markers relative to trilobites, but are used as biostratigraphical indicators within the trilobite series (Müller and Hinz, 1991).

Conodonts and trilobites have different hard part compositions, and will therefore have different preservation potentials in different lithologies. Hence, conodonts may be of biostratigraphical importance where trilobites have not been preserved, such as in as in Estonia (Kaljo *et. al.*, 1986; Mens *et. al.*, 1993; 1996). Conodont biostratigraphy has primarily been applied on the Cambrian-Ordovician boundary, on all continents except Africa (Müller and Hinz, 1991). Conodont research on the Cambrian – Ordovician boundary in Norway is presented by Bruton *et. al.* (1988).
3.1.1 Trilobites
The trilobite zonal – subzonal system of the Alum Shale Formation is revised several times - since Westergård (1922) established his trilobite zonation system - based on taxa from the almost complete successions in Scania (Sweden) and partly from the Furongian and Tremadocian succession in the Oslo region (Westergård, 1946; 1947; Henningsmoen, 1957; Ahlberg, 2003; Terfelt et. al., 2008; 2011; Ahlberg and Terfelt, 2012; Babcock et. al., 2012; Nielsen et. al., 2014). A trilobite zonation based on agnostids and polymerids from the Furongian Series in Scandinavia has also been suggested by Terfelt et. al. (2011), but revised in Nielsen et. al. (2014) as shown in Figure 16.
Figure 16. Trilobite zonations proposed for the Alum Shale (Modified from Nielsen et. al., 2014).

<table>
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<tr>
<th>(Super)zones</th>
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<th>Chrono-stratigraphy</th>
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</thead>
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<tr>
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<td></td>
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<tr>
<td>Henningsmoen (1957)</td>
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<tr>
<td>Nielsen et al. (2014)</td>
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<tr>
<td>Polyomilids</td>
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<tr>
<td>A. ecorne</td>
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<tr>
<td>W. scanica</td>
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<tr>
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<tr>
<td>A. holmi</td>
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<tr>
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<td>T. holmi</td>
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</tbody>
</table>

Figure 16. Trilobite zonations proposed for the Alum Shale (Modified from Nielsen et. al., 2014).
3.1.2 Conodonts
For biostratigraphical purpose Lower-, Middle-, and lower Upper Cambrian conodonts have been less studied than Upper Cambrian conodonts due to their rarity (Müller and Hinz, 1991). Paraconodonts have not been used in Scandinavia for stratigraphy, despite their abundance (Müller and Hinz, 1991). The euconodonts were not used widely for stratigraphic correlations of the Cambrian in Baltoscandia until the late 1990’s by Szaniawski and Bengtson (1998).

The first conodont zonal scheme from the Upper Cambrian of Baltica was presented by Kaljo et al. (1986). They established the C.? andresi zone and C.proavus zones based on material from the Estonian-western Russian succession. The upper Cambrian euconodont zonation from Baltica was reviewed by Szanski and Bengtson (1998) from material from Kinnekulle in southwestern Sweden, which is now the conodont zonal scheme used for the Upper Cambrian of Baltica (Figure 17). Szaniawski and Bengtson (1998) established the Proconodontus Zone with its two subzones Proconodontus transitans and P. muelleri. The upper boundary of the Proconodontus Zone is defined by the FAD of Cordyloodus? andresi.

<table>
<thead>
<tr>
<th>REGIONAL STAGE</th>
<th>NORTH AMERICA after MILLER (1988a)</th>
<th>SWEDEN</th>
<th>ESTONIA (compiled)</th>
<th>REGIONAL STAGE</th>
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<td>hiatus</td>
<td>Proconodontus andresi</td>
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<td>Hirsutodonstus hirsutus</td>
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<td>Cambroistodus minus</td>
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<td>FRANCONIAN</td>
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<td>Proconodontus</td>
<td>Pehtua</td>
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<tr>
<td></td>
<td></td>
<td>Transitans</td>
<td>uppermost part</td>
<td></td>
</tr>
</tbody>
</table>

Figure 17. Correlation of Conodont zonation of the uppermost Cambrian of Sweden with North America and Estonia (Szaniawski and Bengtson, 1998).
The Cordylodus? andresi Zone is defined by the FAD of C. andresi, and with its upper boundary defined by the FAD of C. proavus (Kaljo et. al., 1986; Szaniawski and Bengtson, 1998) which also defines the C. proavus Zone.

The C. proavus Zone is not recognized in Sweden (Szaniawski and Bengtson, 1998), but has been reported in Scandinavia, from the Oslo Region in upper part of Acerocare Zone (Bruton et. al., 1988), which corresponds to pre-Tremadocian age. According to Szaniawski and Bengtson (1998), insufficient preservation of the conodonts reported in Bruton et. al. (1988) causes some of the designations to be uncertain, and they have therefore not been regarded as certain enough for defining the boundary of the C. proavus Zone in Scandinavia.

3.2 Trilobite fauna and biofacies
Fossiliferous occurrences in black shales, as the Alum Shale - which is interpreted to have been deposited under anoxic conditions - have led to different hypotheses regarding the living conditions of the individuals (Buchardt et. al., 1997). Interpretations of the living conditions for the trilobites suggested they were allochtonous deposited (Dworatzek, 1987), or that agnostids were living near the surface attached to seaweed (Bergström, 1973). Further research has made these allegations rather doubtful due to how the assemblages are sorted and the type of specimens in them (for more detailed discussion see Buchardt et. al., 1997). Due to the assemblages and the further research on the morphology of the trilobites, it is now assumed that olenids and agnostids probably were adapted to dysoxic environment. The high dominance and low diversity also support this theory. The high abundance, high dominance and their adaption to such environments make them suitable for biostratigraphical use in black shale environments (Buchardt et. al., 1997).

The trilobite assemblages in the Alum Shale may be divided into two groups: Olenid and non-olenid trilobites based on the morphology and associated faunal elements. The non-olenids include “normal” trilobites and agnostids, and represents dysoxic to oxic environment (Figure 18). Brachiopods often occur with the non-olenids. The olenids represent dysoxic to anoxic environments as illustrated in Figure 18 (Schovsbo, 2001).
3.2.1 The olenids

The olenid trilobites can be divided in three main morphotypes: the Olenus-type, the Peltura-type and the Ctenopyge-type (Buchardt et. al., 1997).

The Olenus-type is assumed to have been a benthic living trilobite, but some of the trilobites within this group may have been nektobenthic. Within this group, the Parabolina species probably reflects higher oxygen levels than other members of this group (Buchardt et. al., 1997), based on their morphology and distribution in the basin (Bergström, 1980), and may therefore be placed within the non-olenids (Schovsbo, 2001).

The Peltura-type is based on their morphology interpreted to have lived an active swimming mode of life (Schovsbo, 2001). This group is more abundant in Middle Sweden and Öland than further south such as the Oslo area, where representatives from Ctenopyge and Sphaerophtalmus of the same age dominate (Buchardt et. al., 1997). The Ctenopyge-type is interpreted to have been pelagic, floating in the water column (Schovsbo, 2001).
3.2.2 The non-olenids
The non-olenid trilobites include agnostids and “normal trilobites”. Agnostids were small trilobites which lived enrolled (Robinson, 1972). It has been argued that they were pelagic based on the almost cosmopolite distribution of some species (Robinson, 1972). However, the agnostids in the Cambrian were restricted to black shale environments, indicating adaption to such environment and were therefore, probably benthic adapted to the bottom water environment (Nielsen, 1997). It has been stated that agnostids are comparable with ostracods (Buchhardt et al., 1997).

Brachiopods occur with the non-olenids in the Alum shale and are therefore assumed to have been adapted to similar environment (Popov and Holmer, 1994). Both orthide and phosphatic forms are included in the Cambrian brachiopods, and include several Lingula-type brachiopods (Bergström, 1980).

3.3 Conodont fauna and bioprovinces
During the Cambrian, as well as through the early Tremadocian most conodont faunas were relatively cosmopolitan. However, conodont provincialism was established during the late Tremadocian (Charpentier, 1984). Hence, most of the provincialism studies have focused on the Ordovician period, and only few reports exist regarding Cambrian conodonts faunal provincialism (Miller, 1984; Bergström, 1990).

The Upper Cambrian conodont fauna is dominated by paraconodonts and protoconodonts, which consists of a large variety of simple cone elements. In Baltica the genera Furnishina and Westergaardodina are the most abundant and comprise several species (Müller and Hinz, 1991). The group protoconodont is mostly represented by the long ranging genus Phakeledus (Bagnoli and Stouge, 2013). During the Late Cambrian diverse paraconodonts as well as the first euconodonts appear which makes this period important regarding conodont evolution (Jeong and Lee, 2000).

According to Miller (1984) the protoconodonts and paraconodonts represent the cold water realms in mid- to high latitudes, such as Scandinavia, Great Britain, Turkey, Iran, South China and deep water areas along the margins of North America, India, Kazakhstan and other low-paleolatitudes land masses.

The euconodont zonation starting from Proconodontus up to C. proavus zone is typical for the warm water realm in low latitudes, such as the Laurentian platform in North America (Miller 1980; 1984), North China (An, 1981; 1983), South China (Dong et. al., 2004), Kazakhstan
(Dubinina, 2000), Iran (Müller, 1973), Korea (Lee and Lee, 1988) and Australia (Druce and Jones, 1971).

Miller (1984) and Bergströöm (1990) suggested based on the differentiation of cold- and warm faunal realm during the Cambrian, that provincialism may have started in the Late Cambrian, and probably was the early stage of the development of the Ordovician realms that now are called the Midcontinent Realm and the North Atlantic Realm. However, according to Jeong and Lee (2000), this provincialism may not be an initial stage of the Ordovician conodont provincialism, but a separate branch in the evolution of conodonts, considering the end-Cambrian extinction.

Based on quantitative studies by Jeong and Lee (2000), conodonts exhibited provincialism on a global scale during the Late Cambrian. Faunas and associated Simpson Index (SI) values are shown in figure 19. Simpson Index (SI) reflects the number of taxa in common between two faunas, where low SI reflect high provincialism between two areas.

**Figure 19.** SI values between Sweden and other localities in Asia. Low SI values indicate strong provincialism (modified from Jeong and Lee, 2000).

### 3.4 Controlling factors for provincialism

Climate and physical barriers are the two factors controlling provincialism of conodonts, as well as for other marine organisms (Bergström, 1990). Physical barriers include emerged areas and ocean currents, while climatic factors include water temperatures and salinities. Areas with unfavorable climatic conditions may form migration barriers (Jeong and Lee, 2000). Water depth is not regarded as an important factor, based on for example the hypothesis that some conodonts were able to change position within the water column to
favorable conditions (Miller, 1984). It is suggested that water temperature was one of the most controlling factor in the distribution of conodonts (Jeong and Lee, 2000).

Another factor that may have affected the provincialism was the ecological mode of life of the conodonts, but their habitat being benthic, necto-benthic or pelagic is still not certainly known (Jeong and Lee, 2000). Miller (1984) suggested that protoconodonts, paraconodonts and early euconodonts were pelagic and cosmopolitan. This may be the reason why conodont provincialism was not strong in the Cambrian (Jeong and Lee, 2000). For more details regarding Cambrian conodonts and their mode of life, see section 5.1.3.1.

3.5 Baltic Conodonts

Baltoscandian conodonts are well known based on conodonts from the Swedish Alum Shale (Bruton et al., 1988; Müller and Hinz, 1991; Szaniawski and Bengtson, 1998; Bagnoli and Stouge, 2013). The Upper Cambrian euconodont succession in Baltica is not similar to the coeval Midcontinent euconodont succession, representing warm water realm. In northeastern Europe, the Laurentian *Eoconodontus* Biozone, with its two subzones, has not been identified (Bagnoli and Stouge, 2013). The cosmopolitan euconodont species *P. muelleri* and *E. notchpeakensis* are most common in Baltica, but *E. notchpeakensis* is extremely rare before the appearance of *C. proavus* (Bagnoli and Stouge, 2013). The presence of *E. notchpeakensis* in the *C.? andresi* Zone in Estonia and Öland, Sweden, may suggest that this zone can be correlated to the *Eoconodontus* Zone of the Midcontinent Realm as shown in Figure 20 (Bagnoli and Stouge, 2013). The *C.? andresi* Zone established by Bagnoli and Stouge (2013), is only known from the Baltoscandic region (Bagnoli and Stouge, 2013). Bagnoli and Stouge (2012) consider specimens that are assigned to *C. andresi* outside the Baltoscandic region to belong to *C.? aff. andresi*, in the *Acerocarina* superzone.
Different paraconodont associations may also reflect differences in water depth. Bagnoli and Stouge (2013) concluded with three paraconodont associations based on conodonts from the Swedish Furongian Alum Shale, associated with different lithofacies. The paraconodont associations *Furnishina*-*, *Prooneotodus-* and *Westergaardodina* association reflecting deeper marine, deep water, and shallow water environment respectively.

### 3.6 Conodonts from the Oslo Region

The only study of Cambrian conodonts in the Oslo region is done by Bruton *et. al.* (1982; 1988) from the Nærnes Beach. These conodont studies were focused on the Cambrian – Ordovician boundary and presented co-occurrence of conodonts with trilobites and early Ordovician graptolites. Samples were taken from the uppermost Cambrian (current *Acrocarna* trilobite superzone) and the lowermost Ordovician (*Boeckaspis* trilobite zone). The different species and the stratigraphic ranges of the conodonts from Nærnes Beach are shown in Figure 21.
Figure 21. Stratigraphic ranges of the conodonts at Nærsnes Beach. A = *Acerocarina* trilobite superzone, B = *Boeckaspis* trilobite zone (modified from Bruton et al., 1988).
4 Material and methods

4.1 Field work
The fieldwork of this study was done during the spring 2013. Sections of the Alum Shale Formation, spanning from the Cambrian “Series 3” into the Lower Ordovician (Tremadocian) were investigated and sampled in Slemmestad. Slemmestad is located in Røyken commune in the county of Buskerud (Figure 22).

The exposed sections at Slemmestad used for this field work include a section of Middle Cambrian, a section of the earliest part of Furongian, and a section of the upper half of Furongian which spans the Cambrian-Ordovician boundary, in addition to an entire section of the Tremadocian.

The field work was done together with supervisor Hans Arne Nakrem and Magne Høyberget. Material from six different stratigraphic levels was collected from two different areas in Slemmestad during this field work. In total, samples from nine different levels were collected in purpose of this thesis. Two of them were collected and kept in the museum collection before this fieldwork took place, and one was collected and provided for study by Bjørn Funke from a presently inaccessible locality.

The nine samples are collected from five different outcrops in Slemmestad, and are marked on the map below (Figure 22). Sample KAM1 and KAM2 are collected inside the Norcem industrial area where access requires permission.
The nine samples were collected from levels ranging from the Middle Cambrian (Series 3) representing the trilobite superzone *Paradoxides paradoxissimus* to the lowermost Ordovician (Tremadocian), representing the *Boeckaspis* trilobite zone. The samples were collected according to the well-established trilobite zones by Nielsen *et. al.* (2014) (see Figure 16, section 3.1.1). The different samples with corresponding GPS coordinates, trilobite superzones and weights are shown in Table 1.
Table 1. The different samples with corresponding coordinates, trilobite superzones and weight.

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<thead>
<tr>
<th>Sample name</th>
<th>UTM Coordinates</th>
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<td>KAM6</td>
<td>32V 584059E, 6628274N</td>
<td>Acerocarina</td>
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</tr>
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<td>KAM5</td>
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<td>Peltura</td>
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</tbody>
</table>

An improvement of the available logs on the sections used for this study would require an extensive field work. The purpose of this thesis was not to do detailed logging. Since a less comprehensive logging would not add any further details to the existing logs, no logging was done.

The exposed succession where GIBB13 was collected includes Middle Cambrian Alum Shale deposits underlain by Precambrian basement, and is overlain by a Permian sill (Figure 23). Other samples were collected from limestone beds and nodules in the alum shale (Figure 24).
Figure 23. Location of sample GIBB13. Middle Cambrian Alum Shale underlain by Precambrian basement and overlain by a Permian sill.

Figure 24. A) Limestone nodule in the *Acerocarina* superzone, upper half of the Furongian. Scale bar is 30cm. B) Limestone bed in the *Acerocarina* superzone.
The sections at Slemmestad used for this study are presented as a simplified composite profile. The lithology and biostratigraphic location of the samples within the associated trilobite superzones is presented in the log (Figure 25).

![Composite and simplified log of the sections used for this study.](image)

**Figure 25.** Composite and simplified log of the sections used for this study. The log illustrates which trilobite superzone the different samples are taken from, and which samples that is taken from beds or concretions, as well as relative size and stratigraphic order. The log is shortened, and only shows zones where samples are taken from.
4.2 Preparation of slabs and thin sections
Material from the samples were cut with a rock saw to slabs at approximately 3 x 2 x1 cm and polished with carborundum polishing paper. In total, 25 standard petrographic thin sections (30µm thickness) were made from the nine analyzed samples by Salahalldin Akhavan at the Department of Geosciences, University of Oslo. Thin sections were made both parallel and perpendicular to bedding.

4.3 Acid processing of samples
All the nine samples were processed using standard conodont procedures. The samples were, however, not crushed, but placed in 10-15% diluted acetic acid. Undissolved fractions between 63µm – 500µm were sieved and dried. The fractions <500µm were heavy liquid separated using the heavy liquid diodomethane diluted with acetone to a density of ± 3.00g/ml. The heavy liquid was stepwise thinned out to a density of ±2.75g/ml and all the fractions between were washed with acetone, dried, collected and analyzed for conodonts and other biogenic material by using a Leica microscope. Conodonts and other biogenic material were then handpicked from the samples and studied. For details regarding the acid processing, see Appendix 1.

4.4 Microscopy
Both transmitting and reflective microscopes were used for this study. A Leica DMLP transmitting light microscope at NHM was used for analysis of thin sections and to photograph relevant conodont elements in transmitted light. Photographs were taken with a digital Leica DC 300 camera mounted on the microscope. A Leica MZ16A reflective light microscope at NHM was used for analyzing conodonts and other biogenic material. A Nikon D5100 camera mounted on the reflective microscope was used to photograph the specimens. The computer software Helicon focus was used to sharpen the photographs of each specimen photographed with the reflective light microscope.

4.5 Scanning Electron Microscope (SEM)
A Hitachi 3600N-model scanning electron microscope (SEM) located at NHM was used for imaging conodonts, and other biogenic material as well as for investigation of thin sections. Photography was done using low vacuum, and the objects were not coated.

A detector in the SEM records secondary electrons that are emitted from the surface due to irradiation of primary electrons from an electron gun. The detector records more secondary electrons from faces pointing towards the detector. These faces brighten up in the resulting
image. Faces pointing away from the detector are shown as dark areas in the image. The image hence show the object as it was illuminated from an angle, giving a 3D effect.

Chemical analyses are done using the energy dispersive spectrometer (EDS) on the SEM. When atoms are irradiated by electrons, they get excited and emit X-rays with wave lengths and energies characteristic for the atom. The EDS records the energies of the X-ray photons and can thus tell what atoms that are present at the spot where the electron beam is focused. This is used for mineral identification on a mineral grain or a microfossil. For semi quantitative analyses of areas within a thin section, the electron beam is scanned over the field of interest, with the EDS continuously recording.

Imaging and chemical analyses were primarily done at low vacuum, not requiring carbon coating. For chemical analyses of carbonate rosettes, high vacuum was used and hence the samples required carbon coating. The high vacuum analyses were done at the JEOL-JSM-6460LV scanning electron microscope at the Department of Geosciences, University of Oslo.

4.6 Microfacies analysis
The thin sections were scanned using a 4000 dpi Nikon Super Coolscan 4000 slide scanner at NHM. Point counting was then done using the computer software JMicrovision. At least 400 counts in each thin section were recorded using the recursive grid function. Dunham carbonate classification was used to classify the carbonates based on point counting results. To distinguish different fossil groups as well as microstructure analysis a Leica DMLP transmitting light microscope was used, with both plain polarized and cross polarized light.
Conodonts (Figure 26) were a group of primitive jawless vertebrates, and are placed within the phylum Chordata: animals with a notochord. These animals were the first vertebrates to produce an internal mineralized skeleton, and they can be compared to the modern hagfish (Armstrong and Brasier, 2005). They are primarily known as small calcium phosphatic tooth-like elements from their feeding apparatuses, referred to as conodont elements. True conodonts, or euconodonts, evolved during the Late Cambrian and ranged to the end of the Triassic. Protoconodonts and paraconodonts are known from Cambrian and Ordovician, and are by definition not true conodonts due to different modes of growth and internal structures, and are by some authors combined in the order Protoconodontida (Armstrong and Brasier, 2005).

Conodonts are the main microfossil group used for dating Paleozoic shallow marine carbonates. They are also used in paleoecological and biogeographical studies. Conodont color alteration index (CAI) is used for basin history interpretations, thermal maturation studies, and for search of hydrocarbons (Armstrong and Brasier, 2005).

The morphology, ecology and taphonomy of conodonts with focus on conodonts from Upper Cambrian Alum Shale will be briefly described in this section. Their use in biostratigraphy and faunal studies are described in the section 3.1.2 and 3.3 respectively. Due to limited information on the morphology and anatomy of Cambrian conodonts, euconodonts are used for illustrations.
5.1 Previous work
The conodont animal affinity was debated until complete fossils of conodont animals were first discovered in the Carboniferous Granton Shrimp bed in 1983, now referred to as the Granton conodonts (Briggs et al., 1983). Based on excellent preservation detailed information on the anatomy of these animals was provided, and this study, among other studies placed conodonts within the phylum Chordata (Armstrong and Brasier, 2005).

The function of the conodont elements was also debated. Pander (1856) suggested the conodont elements to have teeth function, Lindström (1974) suggested that they functioned as internal supporting organs, while Conway Morris (1976) suggested they functioned as lophoporate-supporting structures. Today, conodont elements are accepted as having a teeth function (Armstrong and Brasier, 2005).

Conodonts were first illustrated by Pander (1856), and were described as the remains of an unknown group of Paleozoic fish, and based on the teeth like shape he named the whole group “conodonts”. Hinde (1879) found a cluster of conodont elements in one of his samples from the Devonian and interpreted this cluster as an apparatus of a single specimen. Later work described each element as a separate species based on form taxonomy. Multi-element taxonomy, was first applied from the early 1960's, using different elements to reconstruct the whole apparatus for classifying a single species (Armstrong and Brasier, 2005). Walliser (1964) and Sweet and Bergström (1969) were important in the development of using the multi-element system of classifying conodonts, and this is now the system used (Armstrong and Brasier, 2005).

Several conodont classification schemes have been suggested since 1970, based on the multi-element system. The scheme proposed by Clark with others in Moore (1962), modified by Sweet (1988) and Aldridge and Smith (in Benton, 1993) is the most complete. The Conodonta in this scheme is organized based on two coniform ancestral lineages which first appeared in the Late Cambrian: the Teridontus lineage and the Proconodontus lineage. The Teridontus lineage is interpreted as being the ancestral to all familiar conodont taxa, whereas the Proconodontus lineage is impoverished (Sweet and Donoghue, 2001). The latter have been the lineage of interest regarding Cambrian - Ordovician studies of Baltica (Szaniawski and Bengtson, 1998) shown in Figure 27.
5.1.1 Conodont morphology

5.1.1.1 Soft anatomy
Due to the rareness of conodont animal fossils their anatomy is primarily based on the Granton conodont animals (Armstrong and Brasier, 2005). These conodonts are about 40mm long, eel-like and laterally compressed. The head region is distinguished with two lobe-shaped structures representing where the eyes were positioned, as well as conodont elements, representing the feeding apparatus (Briggs et. al., 1983). Notochord, chevron-shaped muscle blocks and caudal fin rays are the main structures preserved in the body, shown in Figure 28.
5.1.1.2 Conodont elements

The conodont elements represent elements from the feeding apparatus of the conodont animal. These elements are composed of calcium phosphate, and have a size range from 0.25-2mm (Armstrong and Brasier, 2005).

Most of the pre-Carboniferous euconodont elements consist of two parts, the crown and the basal body (Armstrong and Brasier, 2005; Murdock et. al., 2013). The basal body is positioned in an opening in the crown, called the basal cavity (Figure 29).
The crown in euconodonts comprises hyaline lamella tissue with growth lines and “white matter”, an internal opaque tissue commonly seen in the cusp and the cores of the serrated denticles (Szaniawski and Bengtson, 1998; Armstrong and Brasier, 2005). White matter is absent in conodonts of the order Protoconodontida, which makes this a distinguishable feature between the Proto- and Euconodontida (Szaniawski and Bengtson, 1998).

Representatives of the order Protoconodontida consists of large variety of simple cone elements (Müller and Hinz, 1998), which differs from the more complex euconodonts with more differentiated morphotypes (Szaniawski and Bengtson, 1998).

**Function**

Different morphology of the elements is interpreted as representing different function within the apparatus (Szaniawski and Bengtson, 1998).

Relatively few three-dimensional conodont apparatuses are known, and those are of younger age than Cambrian. Morphologically and functionally differences divide the elements in at least two distinct domains, the coniform taxa and the non-coniform taxa. The stereotype for all non-coniform species is the apparatus of the Silurian ozarkodinid conodonts, shown in Figure 30. The morphologically different elements in non-coniform taxa are divided in domains of paired elements representing different function within the apparatus termed the
rostral domain which comprises of paired S elements, and caudal domain comprising paired M and P elements) shown in Figure 30. The locations of the elements within the domains are interpreted from the shapes, and are not of relevance for this study. For more detailed information see Armstrong and Brasier (2005). The function of the S and M elements is interpreted as grasping the food (bar type elements), while P elements had a slicing (blade type elements) and crushing function (platform type elements) (Armstrong and Brasier, 2005). The different types of elements described above are shown in Figure 31, showing morphological terminology.

Figure 30. Conodont apparatus of an ozarkodinid conodont showing orientation and nomenclature of the different elements (Armstrong and Brasier, 2005).
No real consensus of reconstruction and description of coniform apparatuses exist. A scheme for the Silurian panderodontid conodonts was suggested based on fused clusters of elements and natural assemblages of the *Panderodus* animal by Sansom *et. al.* (1994). The apparatus
may be divided in a rostral domain containing \( q \) elements and a caudal domain containing \( p \) elements, and contains morphologically different elements within the domains. The elements were paired and lay across the midline of the animal as shown in Figure 32. It is interpreted that the \( q \) elements (rostral domain) had a grasping function, while the \( p \) elements (caudal domain) processed the food (Armstrong and Brasier, 2005).

![Figure 32. Illustration of the Phanderous unicostatus apparatus. a) Rostal view. b) Lateral view. c) Location and terminology of the elements. (Armstrong and Brasier, 2005)](image)

5.1.2 Cambrian conodonts
Most of the Cambrian conodonts belong to the protoconodonts and paraconodonts (Müller and Hinz, 1991), which by some authors are combined in the order Protoconodontida (Armstrong and Brasier, 2005). Representatives from the oldest known euconodonts (true conodonts) are from the order Proconodontida (Armstrong and Brasier, 2005).
Order Protoconodontida includes the protoconodonts, known from the Precambrian-Cambrian transition and Ordovician, and the paraconodonts, known form the Cambrian and Ordovician (Miller 1984; Armstrong and Brasier, 2005). Armstrong and Brasier (2005) describes these conodonts as “a number of weakly phosphatisized elements bearing a superficial resemblance to conodonts”. Protoconodonts and paraconodonts are by definition not true conodonts due to different modes of growth and internal structure (Armstrong and Brasier, 2005). The order Proconodontida, containing the first euconodonts, evolved in the Late Cambrian (Miller 1984; Armstrong and Brasier, 2005).

Lineages of the different Cambrian conodont orders are illustrated in figure 33.
It has been suggested that paraconodonts evolved from protoconodonts (Bengston, 1976), but this relationship has not been confirmed (Armstrong and Brasier, 2005). Protoconodonts are excluded from euconodont ancestry, while it is suggested that euconodonts are derived from paraconodonts (Murdock et al., 2013).

For discussion regarding the evolutionary relationship between proto-, para-, and euconodonts see Bengston (1983), Andres (1988) and Murdock et al. (2013).

5.1.2.1 Mode of growth
The elements of euconodonts had a centrifugal appositional mode of growth, which means that laminae in the crown and basal body are added synchronously (Murdock et al., 2013), so that the inner lamella is the oldest (Armstrong and Brasier, 2005). Hence, these elements were growing by deposition over the entire surface (Bengston, 1976). Protoconodonts and paraconodonts have a different mode of growth, with resemblance to the centripetal structure of teeth, and have different internal structure than euconodonts (Müller and Hinz, 1991). Paraconodont elements are similar to the euconodont basal body alone due to their apposition of lamella only to the proximal surface (Murdock et al., 2013). Hence, unlike euconodonts, elements of paraconodonts grew by deposition only basally (Bengston, 1976). The different modes of growth of proto-, para- and euconodonts are illustrated in Figure 34.

![Figure 34. Illustration of the different modes of growth of the A) proto-, B) para-, C) and euconodonts, showing the similarity between the paraconodonts and euconodont basal body (modified from Armstrong and Brasier, 2005).](image)

Some of the earliest euconodonts may be difficult to distinguish from paraconodonts and may only be possible under high magnification and when the preservation is good (Szaniawski and
Euconodonts are characterized by their sharp contrast between the dark basal body, and the colorless translucent crown, white matter, but there is also a morphological difference in the elements within the apparatuses (Szaniawski and Bengtson, 1998). Several incomplete clusters of paraconodont apparatuses are known and they consist of two, three or four very similar elements mainly differing in size (Andres, 1981; Szaniawski, 1987; Müller and Hinz, 1991; Szaniawski and Bengtson, 1998). However, the recognition of Cambrian paraconodont apparatuses is still unknown, and difficult to accomplish for most of the Cambrian simple cone taxa (Müller and Hinz, 1991; Bagnoli and Stouge, 2013). Within euconodont apparatuses however, the elements are strongly differentiated. This morphological difference within the apparatus was probably due to the elements developing different function (Szaniawski and Bengtson, 1998).

5.1.3 Paleoecology and Paleobiogeography

Interpretations of conodont paleoecology have favored both pelagic (Sheddon and Sweet, 1971), and nektobenthic mode of life (Barnes et. al., 1973; Barnes and Fåhraeus, 1975). Klapper and Barrick (1978) concluded based on arguments form previous publications arguing for a pelagic versus nektobenthic mode of life, that it is not possible to distinguish this based on distribution data. They concluded that the only evidence for a pelagic mode of life is conodonts in black shales, with lack of benthic organisms and bioturbation, reflecting deposition in anoxic bottom conditions (Miller, 1984).

There is not much detailed information regarding conodont paleoecology of Cambrian and earliest Ordovician age. The only detailed work is done by Miller et. al. (1981), Miller (1984), Cooper et. al. (1981), and Landing et. al. (1980) who briefly considered Cambrian-Ordovician conodont biofacies, and by Bagnoli and Stouge (2013) who studied Upper Cambrian conodonts from Sweden.

5.1.3.1 Mode of life

Conodonts mode of life have been interpreted based on functional morphology, faunal associations and facies distribution. Conodonts were exclusively marine and lived within habitats from hypersaline to bathyal, and abyssal. Conodont elements are found deposited below the calcite compensation depth (CCD), but these were probably nektonic or pelagic animals (Armstrong and Brasier, 2005). Highest numbers of conodont elements are found in limestone samples from shallow marine tropical and subtropical environments, probably reflecting they were among the dominant groups in these habitats (Armstrong and Brasier, 2005).
Most of the conodonts show facies dependence to some extent, indicating they lived close to the sea floor, as nektobenthic animals. Both the Granton animals and fossils of conodont animals from Soom Shale have characters indicating they were active nektobenthonic predators or scavengers. The coniform taxa on the other hand, are found in a wider range of facies suggesting a nektonic or pelagic mode of life (Armstrong and Brasier, 2005).

Conodonts are interpreted as macrophagous based on functional morphological studies of the feeding apparatuses, meaning that they fed on living or recently dead prey. Based on the lack of jaws, it is suggested that the conodont animal pulled chunks from the prey instead of bite, such as the modern hagfish (Armstrong and Brasier, 2005).

At various times in conodonts history, they show provincialism which is interpreted as reflecting they were sensitive to temperature. Based on this, two distinct conodont faunas were established representing the separate faunas during Ordovician at high and low latitudes. The faunas are commonly known as the North Atlantic and American Midcontinent Provinces respectively (Armstrong and Brasier, 2005). The provincialism has also been used for Late Cambrian conodonts (Jeong and Lee, 2000). For more details regarding Cambrian conodont provincialism see section 3.3.

5.1.3.2 Distribution of Cambrian conodont lineages
In this section different genera representative for each order is mentioned. The genera relevant for Baltica, and their distribution is more elaborated.

Protoconodonts includes the four genera *Protohertzia*, *Amphigeisina*, *Gapparodus* and *Phakelodus*, and is known from Precambrian-Cambrian transition. *Phakelodus* is the only cosmopolitan genus. Both *Phakelodus* and *Amphigeistina* are present in the Alum Shale Formation in Sweden, and are based on this interpreted as being pelagic. According to Miller *et. al.* (1981) *Phakelodus* occurs in facies ranging from “shallow, normal marine, to possible restricted platform-margin and shelf paleoenvironments … to deep continental slope deposits” (Miller, 1984). This genus is reported from equatorial to high paleolatitudes in western North America, Sweden (Müller, 1959) and Poland (Szaniawski, 1980). Based on the broad range of latitudes, *Phakelodus* had probably high tolerance for temperature variations, or that they only lived in preferable temperatures by changing their position in the water column. They were probably not tolerant to very shallow water or high salinity (Miller, 1984).
Paraconodonts, as a group, were probably cosmopolitan, but some genera were more widely spread than others. They are interpreted as having the same required environments as the protoconodonts, due to the occurrence in the Alum Shale in Sweden, and hence, they are interpreted as being pelagic (Miller, 1984). Most Cambrian strata that contain coniform paraconodonts contain Westergaardodina. Paraconodonts include several genera, but in Sweden the genera Westergaardodina and Furnishina are the most abundant ones (Müller and Hinz, 1991). They were probably tolerant to variations in water temperature or possibly depth, but were probably intolerant for elevated salinity. Phakelodus often occurs with paraconodonts (Miller, 1984).

Proconodonts include the four genera, Proconodontus, Eoconodontus, Cordylodus, known from North America, Asia, Europe and Australian in Cambrian deposits (Miller, 1984; Bruton et al., 1988; Müller and Hinz, 1991; Bagnoli and Stouge, 2013), and Cambrooistodus, known from North America and China. Cordylodus was cosmopolitan during the late Cambrian and Tremadocian (Miller, 1984). Genera of this lineage are found in various depositional environments, most reported from shallow platform carbonates, but they are also reported from deep continental-slope deposits, but it is possible that these have been re-deposited from shallower-water deposits (Miller, 1984). Proconodontus and Cambrooistodus are mostly found in sediments representing warm equatorial and mid-paleolatitudinal regions, and they were probably intolerant for high salinity. Proconodontus was probably cosmopolitan and intolerant for cold temperatures, and they were both probably pelagic (with some uncertainties regarding Cambrooistodus). Proconodontus is however also known from the Scandinavian Alum Shale (Müller and Hinz, 1991). Cordylodus and Eoconodontus did probably not prefer high salinity, but they were probably tolerant to it. Due to their presence in the Alum Shale they are also interpreted as being pelagic (Miller, 1984).

5.1.4 Taphonomy
Preservation of the soft tissue from conodont animals is extremely rare (Briggs et al., 1983) and requires exceptional preservation conditions including anoxic environment and rapid burial. Hence, soft tissue from the conodont animal is only known from three locations worldwide, including the Granton conodont animals from the Carboniferous Granton Shrimp Bed (Aldridge and Theron, 1993; Briggs et al., 1983). Favorable conditions of low turbulence and rapid burial can preserve complete feeding apparatuses of 15 elements or more (Armstrong and Brassier, 2005). In addition to these criteria, preservation of partly organic
basal bodies in euconodonts requires a rapid post mortem phosphatization (Szaniawski and Bengtson, 1998).

The conodont animals that lived a nektobenthic mode of life probably settled rapidly from the water column after their death, and were hence probably deposited autochtonous (Coussens, 2002). The nektonic and pelagic conodonts, however, may have been transported laterally in the water column before deposition (Heckel and Baesemann, 1975). Conodonts from all the above mentioned modes of life may have been deposited in coprolites from predators or carnivores feeding on conodonts (Clark, 1989). Therefore the conodonts may have been transported and deposited allochtonous, despite their nektobenthic mode of life. Nektobenthic and shallow water living conodonts may be redeposited by gravity flows or turbidites to deeper marine environment (Dumoulin et al., 1996).

Conodonts are, in addition to other microfossils, used in sequence stratigraphy to interpret different tracts and surfaces as shown in Figure 35 (Emery and Myers, 1996) and may be used to reconstruct depositional environment within a basin.

Figure 35. Conodonts associated with other microfossil groups to interpret different tracts and surfaces (Emery and Myers, 1996)
Due to the composition of the conodont elements, they are more resistant to dissolution than other calcareous fossils, and may be deposited beneath Carbonate Compensation Depth (CCD).

Based on their composition, conodonts change their color when exposed to thermal maturation (heating). The color change extends from pale yellow to black or colorless, depending on the degree of burial and heating, shown in Figure 36 (Epstein et. al., 1977). This process is irreversible and the color-change therefore reflects the maximum temperature the sediments have been exposed for. This color-change is the base for the Color Alteration Index (CAI), first introduced by Epstein et. al. (1977), used to interpret thermal maturation, basin history and hence, maturation study of source rocks and the state of hydrocarbons (Armstrong and Brasier, 2005) shown in Figure 63 (Section 7.1.3).

![Color alteration index with associated color on conodont elements both experimentally produced and from field collections (Epstein et. al., 1977).](image)

Figure 36. Color alteration index with associated color on conodont elements both experimentally produced and from field collections (Epstein et. al., 1977).
5.2 Results
Conodonts were present in the acid insoluble residue due to pyritization and by their original calcium phosphatic composition. Conodonts were found in the samples GIBB13, KAM2, PEL13, KAM5 and KAM6, and were picked from the heavy mineral separated fractions. The specimens were studied and photographed by using transmitted and reflecting microscope and SEM. For more details regarding methods used see chapter 4.

Only few conodont specimens were present in the samples (Table 2 and Table 3), except from PEL13 and to some extent KAM2, which had a high content of fragmented protoconodonts, which are not of interest for biostratigraphical or faunal assemblage studies. Hence, the results are not representative for quantitative analysis. The abundance of conodonts in the five conodont bearing samples, as well as other fossils present in the acid insoluble residue is presented in Table 2. The fossil abundance including conodonts and the other microfossils extracted from the acid insoluble residue in the samples is divided in low, medium and high, corresponding to 1-5 specimens, 6-10 specimens and 11 or more specimens respectively.

Table 2. Results from the acid insoluble residue. Abundance of fossils and conodonts are presented. Unidentified objects are biogenic material biological affinity.

<table>
<thead>
<tr>
<th>Raw material</th>
<th>Insoluble sample residue</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sample</td>
<td>Weight (kg)</td>
</tr>
<tr>
<td></td>
<td></td>
</tr>
<tr>
<td>KAM7</td>
<td>5</td>
</tr>
<tr>
<td>KAM6</td>
<td>5</td>
</tr>
<tr>
<td>KAM4</td>
<td>5</td>
</tr>
<tr>
<td>KAM5</td>
<td>5</td>
</tr>
<tr>
<td>KAM1</td>
<td>5</td>
</tr>
<tr>
<td>PEL13</td>
<td>7</td>
</tr>
<tr>
<td>KAM2</td>
<td>5</td>
</tr>
<tr>
<td>KAM8</td>
<td>5</td>
</tr>
<tr>
<td>GIBB13</td>
<td>7</td>
</tr>
</tbody>
</table>
The conodonts show a CAI of 4 to 4.5, shown in Figure 37.

![Figure 37. Conodonts in reflected light with representatives from all the conodont bearing samples. A) GIBB13, PMO 221.748/2. B) KAM5, PMO 221.739/40. C,D,E) PEL13, PMO 221.742/25/44/63 respectively F) KAM2, PMO 221.746/21 upper. G,H) KAM6, PMO 221.737/1/3 respectively.](image)

The conodonts were identified together with Professor Svend Stouge from the University of Copenhagen. They were mainly identified based on the taxonomic descriptions by Müller and Hinz (1991). Morphological terminology used is based on Müller and Hinz (1991), for both simple cones (Figure 38) and Westergaardodina (Figure 39).

One of the species present, *Cordylodus proavus*, is not described by Müller and Hinz (1991). Identification of this species is based on description from Pyles and Barnes (2000). The apparatus of *Cordylodus* consists of dolabrate elements, which are elements that only have caudal processes, commonly pick shaped (Armstrong and Brasier, 2005). Morphological terminology on dolabrate elements is based on Armstrong and Brasier (2005), shown in Figure 40.

The conodont ranges in the different samples are shown in Figure 43.
Figure 38. Terminology used for simple cones, by Müller and Hinz (1991).

Figure 39. Terminology used for *Westergaardodina*, by Müller and Hinz (1991).

Figure 40. Terminology used for dolabrate elements (Modified from Armstrong and Brasier, 2005 and personal.kent.edu).
5.2.1 Conodont identification
The conodonts were identified using reflecting light microscope. Some of the conodonts are well preserved, but they are very fragile. The conodont specimens present in the samples belong to the order Protoconodontida with representatives of the genera Phakelodus, Westergaardodina, Trolmenia, Problematocconites and the order Proconodontida with representatives of the genus Cordylodus. The abundance of the different conodont species present in the samples is summarized in Table 3.

Table 3. Number of conodont species in the different samples.

<table>
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<th>Species</th>
<th>KAM7</th>
<th>KAM6</th>
<th>KAM4</th>
<th>KAM5</th>
<th>KAM1</th>
<th>PEL13</th>
<th>KAM2</th>
<th>KAM8</th>
<th>GIBB13</th>
</tr>
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<td></td>
</tr>
<tr>
<td>Problematocconites perforatus</td>
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<td>Trolmenia acies</td>
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Systematic composition
Phylum CHORDATA Bateson 1886
Class CONODONTA Eichenberg 1930
Order PARAISONODONTIDA Müller 1962
Genus Phakelodus Miller 1984

Phakelodus elongatus An 1983

Figure 44 A, D, F, H, J and K-M.

Material: 45

Occurrence: GIBB13 (4 specimens), KAM2 (25 specimens), PEL13 (13 specimens), KAM5 (2 specimens), KAM6 (1 specimen), Alum Shale Formation, Slemmestad, Norway.

Stratigraphic distribution: Middle to Upper Cambrian, Paradoxides paradoxissimus superzone to Acerocarina superzone.
Description: Slender, gently recurved elements. Rounded anterior side and keeled posterior side. The cross-section is tear-shaped at the basis. They may also occur as clusters.

*Phakelodus tenuis* Müller 1959

Figure 44 B, C, E, G and I.

Material: 47 specimens.

Occurrence: GIBB13 (1 specimens), KAM2 (12 specimens), PEL13 (32 specimens), KAM5 (2 specimens), Alum Shale Formation, Slemmestad, Norway

Stratigraphic distribution: Middle to Upper Cambrian, *Paradoxides paradoxissimus* superzone to *Peltura* superzone.

Description: slender, gently recurved elements. Rounded anterior side and posterior side. The cross-section is oval at the basis.

Genus *Westergaardodina* Müller 1959

*Westergaardodina ligula* Müller and Hinz 1991

Figure 45 D and E.

Material: 3 specimens

Occurrence: KAM5 (3 specimens), Alum Shale Formation, Slemmestad, Norway.


Description: tricuspidate elements with extremely small median projection. The anterior side is strongly convex. The posterior side is deeply excavated giving it a spoon appearance.

*Westergaardodina polymorpha* Müller and Hinz 1991

Figure 45 A-C E,F, H and I.

Material: 9 specimens

Occurrence: PEL13 (2 specimens), KAM5 (7 specimens), Alum Shale Formation, Slemmestad, Norway.

Description: gently recurved bicuspidate elements with a much larger posterior side than anterior side. The profile is rather flat and median projection is very small, or absent. The lateral projections diverge increasingly during growth. The posterior side is often enlarged in the basal part.

Genus Trolmenia Müller and Hinz 1991

Trolmenia acies Müller and Hinz 1991

Figure 46 E.

Material: 1 specimen

Occurrence: KAM5 (1 specimen), Alum Shale Formation, Slemmestad, Norway.

Stratigraphic distribution: Upper Cambrian, Peltura superzone.

Description: Slender, broadly recurved simple cone element. The anterior side has a short keel at the base. The flanks are rounded and the cusp is long and narrow.

Genus Problematoconites Müller 1959

Problematoconites perforatus Müller 1959

Figure 46 A-D.

Material: 4 specimens

Occurrence: KAM5 (4 specimens), Alum Shale Formation, Slemmestad, Norway.

Stratigraphic distribution: Upper Cambrian, Peltura superzone.

Description: Clearly recurved elements with a large basal opening, and rounded tip. Cross-section is oval at the basis.

Order Euconodonta Müller and Hinz 1991

Genus Cordylodus Pander 1856

Cordylodus proavus Müller 1959

Figure 41, Figure 42 and Figure 46 F-H.
Material: 3 specimens

Occurrence: KAM6 (3 specimens), Alum Shale Formation, Slemmestad, Norway.

Stratigraphic distribution: Upper Cambrian, Acerocarina superzone.

Description: Dolabrulate elements, with clearly recurved main denticle, and deep basal cavity. The cusp tip and denticles are filled with white matter. This species includes three different morphotype elements; rounded, compressed and twisted.

Figure 41. Cordyodus proavus specimens from the sample KAM6, representing three different morphotypes A) Compressed, B) rounded, C) "twisted". A-C) PMO 221.737/1/3/2 respectively. White matter is visible in the denticles.

Based on the well preserved specimens of this species, this species was identified using a reflected light microscope. To identify this species transmitted microscope is often used. Photograph in transmitted microscope was attempted, but the specimen broke during mounting, and photograph of the crown was hence not acquired, but white matter is clearly visible in the denticles (Figure 42).

Figure 42. Cordyodus proavus rounded element with transmitted light microscope, showing white matter in the denticles. PMO 221.737/3.
Figure 43. Stratigraphical ranges for conodonts in samples correlated with trilobite zones

<table>
<thead>
<tr>
<th>Trilobite zone</th>
<th>Log</th>
<th>Stratigraphical ranges</th>
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</thead>
<tbody>
<tr>
<td>Boeckaspis</td>
<td></td>
<td></td>
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<tr>
<td>Acerocarina</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Peltura</td>
<td></td>
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<tr>
<td>Parabolina</td>
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<tr>
<td>Paradoxides</td>
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<tr>
<td>Paradoxissimus</td>
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*Key:*
- **Phakelodus tenuis**
- **Phakelodus elongatus**
- **Westergaardodina polymorpha**
- **Westergaardodina ligula**
- **Trochenia acies**
- **Proterocrinus perforatus**
- **Cordylodus proaurus**

*Legend:* Shortening in log, dark grey limestone, Alum Shale.
Figure 44. Phakelodus elongatus: A, D, F, H, J-M. Phakelodus tenuis: B, C, E, G, I. A) KAM5, PMO 221.739/40. B) PEL13, PMO 221.742/63. C) PEL13, PMO 221.742/55. D) PEL13, PMO 221.742/40. E) PEL13, PMO 221.742/45. (F) PEL13, PMO 221.742/44. G) PEL13, PMO 221.742/25. H) GIBB13, PMO 221.748/1. I) KAM2, PMO 221.746/19. J) KAM6, PMO 221.737/5. K) KAM2, PMO 221.746/22. L-M) fragmentary clusters KAM2, PMO 221.746/56/17 respectively.
5.2.2 Conodont fauna and stratigraphic distribution

The paraconodont species are reported from older or age equivalent deposits from Sweden (see Müller and Hinz, 1991, p. 9; Bagnoli and Stouge, 2013). *Cordylodus proavus* is not certainly identified from Scandinavia, but is reported from age equivalent deposits outside Scandinavia (Miller, 1984; Mens *et al.*, 1993; 1996). The protoconodont species are reported from older and age equivalent deposits (Miller, 1984; Müller and Hinz, 1991).

All the conodont faunas are representative for the cold-water realms. The conodont faunas comprise the protoconodonts *Phakelodus tenuis* and *Phakelodus elongatus*. The succession of the protoconodonts begins in the lowermost sample, GIBB13, correlating to the *Paradoxides paradoxissimus* superzone, and includes both the species mentioned above (Table 3). *Phakelodus tenuis* has its upper range in sample KAM5, correlating to the *Peltura* superzone, while *Phakelodus elongatus* is present in all the conodont bearing samples (Table 3).

The paraconodonts is present in three of the five conodont bearing samples (Table 3). The succession of the paraconodonts begins with one unidentifiable species of the genus *Westergaardodinia* in KAM2, correlating to the *Parabolina* superzone. *Westergaardodinia polymorpha* is present in PEL13 and KAM5, both representing the *Peltura* superzone. The
fauna changes in KAM5 with its diverse fauna compared to the other samples (Table 3). Here *Westegaardodina ligula, Trolmenia acies* and *Problematokoconites perforatus* are present. The paraconodont succession has its last occurrence in sample KAM5 correlating to the *Peltura* superzone.

Only one species belongs to the order Proconodontida, *Cordylodus proavus*, and is present in KAM6, correlative to the *Acerocarina* superzone.
6 Microfacies analysis and depositional environments

6.1 Previous work
As previously mentioned, microfossils may provide information regarding the depositional environment. However, studies regarding faunal composition of the Cambrian Alum Shale in Oslo Region have mostly been devoted to trilobites (Henningsmoen, 1957; Bruton et. al., 1982; 1988; Høyberget and Bruton, 2012). Hence, information regarding other faunal elements (besides the well studied trilobites) such as brachiopods and ostracods is scarce. As mentioned, the only Cambrian conodont research from the Alum Shale in the Oslo Region is done by Bruton et. al. (1982; 1988), who in addition recorded early Ordovician graptolites and trilobites and presented co-occurrences of these fossil groups. Other microfossils in the Cambrian Alum Shale in the Oslo region are found, but no detailed description has been done. The fossils found are representatives of “small shelly fossils”. The term “small shelly fossils” is used for small originally phosphatic or secondarily phosphatized fossils recovered from Lower Paleozoic limestones by etching the rocks in acetic acid (Dzik, 1994). Published material on such microfossils is very limited from the Cambrian Alum Shale in the Oslo region, but is reported (Worsley and Nakrem, 2008).

Microfacies analysis also provides information regarding the depositional environment, as well as diagenesis. As for the faunal components, microfacies analysis has not been extensively used for the Cambrian Alum Shale in the Oslo Region. According to Bjørlykke (1974) limestones of the Upper Cambrian Alum Shale from this area primarily consist of grain supported limestones of trilobite exoskeletons, with matrix consisting of both limemud and sparry calcite.

6.2 Results microfacies analysis
25 thin sections from the nine conodont samples were studied using a petrographic microscope, and 20 of the thin sections were analyzed by computerized point-counting using the computer software JMicrovision. The point counting was done according to Flügel (2010) by at least 400 counts (see Appendix 2) in each thin section using the dual point counting method in a recursive grid. Classification of the samples was done according to the Dunham Classification (Figure 47). The results from the point counting are presented in Figure 48, and the raw data is presented in Appendix 2. Due to strongly differentiated laminations in KAM1, the laminations are counted separately for a depositional interpretation purpose.
6.2.1 Matrix

The matrix of the limestone nodules in the Alum Shale Formation consists primarily of fine-grained carbonate to coarse sparitic calcite, and includes high amounts of bitumen and pyrite. The matrix content is rather high in all the samples, except PEL13 and KAM5 which have lower matrix content with approximately 1% and 50% respectively. The matrix in KAM6 differs from the others with its high content of bitumen and silt sized siliciclastic particles.
The matrix of the finer grained limestone nodules is recrystallized. Trilobites and calcareous brachiopod fragments are visible in hand specimens, but in thin sections they often appear as matrix due to recrystallization and are often visible as relict structures (Figure 49). The sparitic calcite is typically present in the cemented grain supported limestones.

Figure 49. Relict structure appearing as matrix in a finer grained limestone (KAM2, PMO 221.693).

6.2.2 Grains
The grains mainly consist of trilobite fragments as well as some grains of ostracods and other unidentified fragments. The trilobite fragments are of various sizes, and are recognized by their shape and undulating light extinction under cross polarized light. Fragments have characteristic shapes which is easily recognized, as the ‘shepherd’s crooks’ (Figure 50), and the characteristic tri-lobe shapes of thorax fragments (Figure 51). Trilobite fragments are present in thin sections with increasing abundance from GIBB13, KAM4, KAM1, KAM5 to PEL13. These samples are all within the Peltura superzone except GIBB13 which is from the Paradoxides paradoxissimus superzone.
Figure 50. Trilobite fragment in sample GIBB13, showing the characteristic "shepherd's hook". PMO 221.631.

Figure 51. Trilobite fragment in sample PEL13 showing tri-lobe shape characteristic for trilobites. PMO 221.660

Well preserved ostracods are present in some of the thin sections and are recognized by their small shells which are almost of equal shape, but with slightly different size, and a characteristic overlap of one valve by the other. As the trilobites they show undulating light extinction under cross polarized light. They are easily recognized by their shape, and they are well preserved, often with blocky crystal calcite growth inside (Figure 52). This group is present in GIBB13, KAM1 and PEL13.
6.3 Facies description
Based on microfacies analysis of the 25 thin sections, five different facies (Facies 1 - 5) are identified. Four are carbonate facies, while one is siliciclastic mudstone facies (Figure 53). The different facies are described in this section, based on point counting results, fossil content, grain sizes and structures.

Due to recrystallization, some of the finer grained carbonates appear rather homogenous, and structures, if present, are only visible as relict structures. In the sparitic carbonates, structures such as cross bedding and geopetals are present. Burrows are observed in the siliciclastic mudstone facies.

Thin section analysis was done under supervision by Johan Petter Nystuen and Krzysztof Hryniewicz.
6.3.1 Neomorphized Recrystallized Limestone (Facies 1)

Facies 1 represents samples that consist of massive recrystallized fine grained limestone, with less than 10% fossil fragments. Sample GIBB13, KAM8, KAM2, KAM4 and KAM7 are all members of this facies. Due to recrystallization the grain sizes correspond to pseudospar (Flügel, 2010).

Some of the thin sections of this facies show relict laminations with an upward fining trend as shown in figure 54. The most fine-grained (< 100µm) samples within this facies are rather homogenous with respect to grain size and shape (Figure 54A), while the more recrystallized samples have larger and various grain sizes (Figure 54B). The latter also typically contains rosettes of calcium carbonate (Figure 54C) showing mosaic texture (Figure 54D). The uppermost of the samples within this facies, KAM7, shows increased siliciclastic material (primarily silt sized corroded quartz particles) relative to the others samples.

Fossil fragments in this facies are often preserved as relict structures and are strongly recrystallized, but are in some samples well preserved. The fragments, if present, are usually randomly scattered and they typically show no trend in orientation. The fossil fragments are mainly trilobite fragments. Well preserved ostracodes, possibly of the order Bradoriida, are present in GIBB13.

Figure 54. Different textures within Facies 1. A) Homogeneous texture. PMO 221.693. B) Heterogeneous grain size. PMO 221.704. C) Calcite rosette. PMO 221.705. D) Strongly recrystallized with siliciclastic material. PMO 221.703.
6.3.2 Carbonate Skeletal Pack- to Grainstone (Facies 2)  
This facies consists mainly of sparitic calcite and trilobite fragments. The only sample representative from this facies is PEL13. This facies consists of laminated skeletal-type grain supported limestone, with low amounts of carbonate mud. The samples within this facies are laminated with alternating laminas of two different matrix compositions. The laminas can be divided into packstone and grainstone. The contacts between the laminas vary from sharp to gradual, and cross lamination is present (Figure 55B).

This facies show extensive recrystallization of fossils fragments, which are of sub-centimeter in size. The larger fossil fragments are oriented parallel or sub-parallel to the bedding, whereas the smaller grains typically lack orientation. The convex side of the larger grains is mainly oriented up, showing geopetal structures (Figure 55A). Clear, blocky calcite cement fills all porosity between the fragments within the grainstone beds, and no bioturbation is present in this facies.

Since the thin sections from PEL13 are point counted regardless the different laminas, the percentage of matrix from the point counting is higher than expected for a grainstone.

![Image](image_url)

Figure 55. A) Geopetal structure in trilobite fragment from PEL13, PMO 221.660. B) Laminations and cross bedding in PEL13, PMO 221.658.

6.3.3 Carbonate Packstone (Facies 3)  
This facies consists mainly of trilobite fragments and finer grained carbonate matrix (>10%), and is classified as packstone, represented by KAM5. This facies is not cemented such as Facies 2, and the main pore filling component is the finer grained carbonate matrix. The fossil fragments are well preserved while some are partly recrystallized. The fragments are sub-
centimeter in size, and are usually equal in size. They show no trend in orientation, and no bioturbation is observed.

6.3.4 Carbonate Wacke- to Packstone (Facies 4)
This facies consists of alternating wackestone laminas and trilobite-rich packstone beds, and is represented by KAM1. Most of the grains within the packstone laminas consist of trilobite debris and possible small bivalved arthropods and circular calcareous fossils (Figure 56A) of uncertain origin, referred to as calcispheres.

The packstone beds are partly recrystallized and the fossil fragments are preferably oriented parallel/ sub-parallel to bedding, with both concave and convex side facing upwards. However, larger grains are mainly facing concave side upwards (Figure 56B). The packstone laminas contacts with underlying wackestone laminas are sharp, and they show normal grading. In addition, load casts occur in contacts with overlaying thick packstone laminas.

The wackestone laminas are dominated by calcispheres, possible ostracods. In addition, fragments of trilobites and unidentified larger elongated calcareous fragments oriented parallel to the laminas are present throughout (Figure 56A).

Figure 56. Thin section images from PMO 221.690. A) Elongated calcareous fossils. B) Larger grains facing concave side up.

6.3.5 Massive Clay-rich Mudstone (Facies 5)
This facies consists of massive mudstone made up of clay- and silt sized particles, in addition to a high content of bitumen, and is represented by KAM6. Corroded silt-sized quartz is present throughout the sample. This facies shows normal graded laminas/beds with mixed grain sizes occurring rather frequently near the top of the beds. Burrows occur, with a rather low calculated Bioturbation Index (BI) 1 according to James and Dalymple (2010) on the top of the beds. The burrows appear to be fodinichinas, with possible representatives from the ichnogenus Phycosiphon (Figure 57). No fossil fragments are present in this facies.
Figure 57. Bioturbation in sample KAM6, possibly from ichnogenus Phycosiphon. PMO 221.700

6.4 Results acid insoluble residue
In addition to conodonts, the samples contain microfossil groups found in the acetic acid insoluble residue and include inarticulate brachiopods and phosphatizised ostracods. Fragments of trilobites, in addition to other biogenic material such as fecal pellets and fragments of uncertain biological affinity or origin are also present in the acid insoluble residue. The specimens were handpicked from the heavy mineral separated fractions, and studied and photographed with microscope and SEM.

Table 2 summarizes the groups of fossils found in the different samples.

6.4.1 Inarticulate brachiopods
One inarticulate brachiopod specimen was found in the pilot sample GIBB06 (From the same level as GIBB13 from this study, see Section 1.1) (Figure 58A) and one in KAM5 (Figure 58B). Detailed taxonomic identification was not done due to bad preservation, but they are possible representatives of the order Acrotretida based on descriptions by Bruton and Harper (2000) due to circular shape and shallow pits (Figure 58B). A few additional possible inarticulate brachiopod fragments were present in GIBB13 (Figure 58C-E), however they are too fragmented to identify certainly.
6.4.2 Ostracods
One phosphatocopine ostracod (Figure 59A), and one possible phosphatocopine ostracod (Figure 59B), were found in the sample PEL13. No detailed taxonomic identification was done, but they are probably representatives of the order Bradoriida (Armstrong and Brasier, 2005).
6.4.3 Trilobites

Trilobite fragments are presented in Figure 60. Due to perimorphose trilobite fragments occur as casts of gypsum (See Appendix 4) in both KAM2 and in KAM4 (fraction >500µm). Perimorphosis occurs when a mineral or fossil is covered by a layer of secondary deposited mineral, and hence preserve the original shape as a cast (Mineralienatlas.de). Other trilobite fragments were also present due to phosphatization in sample PEL13 (fraction <500µm). No identification was done.

![Figure 60. Trilobite fragments present in the acid insoluble residue. A-D) perimorphosed trilobite fragments of thorax segments from KAM4, PMO 221.738/3/23/18/2 respectively. E) Phosphatized thorax segment from PEL13, PMO 221.745/36. F-G) Phosphatized trilobite fragments from PEL13, PMO 221.745/51 upper/49 respectively. H) Phosphatized trilobite fragment from PEL13, PMO 221.744/46.](image)

6.4.4 Bioclasts of uncertain biological affinity and origin

Other biogenic material was found in the samples KAM2, KAM5 and PEL13. A selection of biogenic material of unknown affinity is presented in Figure 61. This group includes pyritized tubes (Figure 61B-D), fragments (Figure 61I-K), and phosphatized objects of various shapes (Figure 61A, E-H, L).

Fecal pellets from samples KAM2 and PEL13, and spherical objects from samples PEL13 and KAM5 were also found (Figure 62). The spherical objects are possibly representatives of balls from the conodont genus *Westergaardodina* (Müller and Hinz (1991, plate 31, fig.10; plate 32, figs. 13-14)). They are preserved as phosphate (PEL13) or pyrite (KAM5).
Figure 62. Possible spherules from *Westergaardodina* and fecal pellets. A-D) KAM5, PMO 221.740/49/51/54/59 respectively. E-F) PEL13, PMO 221.744/20 left /20 right respectively. G) PEL13, PMO 221.745/35. H) PEL13, PMO 221.743/38. I) KAM2, PMO 221.746/60.
7 Discussion

Based on the results from this thesis and literature study, a discussion on the conodonts, as well as microfacies analysis and depositional environment are presented in this chapter.

7.1 Conodonts

Despite the few conodont specimens extracted from the different samples, the elements of *Cordyodus proavus* in sample KAM6, from the *Acerocarina* trilobite superzone is of biostratigraphical importance. Based on the few specimens from the different samples, information regarding the depositional environment and paleoecology is difficult to obtain. According to Müller and Hinz (1991), both the abundance and preservation of species may vary within very short distances in the same horizon. Whether or not a sample contains conodonts is difficult to recognize during field work. Hence, the results from the different samples may not be representative for the given horizon.

7.1.1 Stratigraphy

As mentioned, both the protoconodonts and paraconodonts are not of stratigraphical importance because they are recorded in younger and age equivalent strata from Sweden (see Müller and Hinz, 1991; Bagnoli and Stouge, 2013). Their stratigraphic range correlated with trilobite superzones are presented in Figure 63.

The occurrence of *Cordyodus proavus* in sample KAM6, corresponding to the *Acerocarina* trilobite superzone is of biostratigraphical significance. This may be regarded as first occurrence of this species in Scandinavia. The *C. proavus* zone is not recognized in Sweden (Szaniawski and Bengston, 1998), but is well known in many regions globally (Szaniawski and Bengston, 1998).

*C. proavus* has been recorded from Upper Cambrian deposits in Oslo area by Bruton *et. al.* (1988). However, Szaniawski and Bengston (1998) concluded that the identification of the specimens found by Bruton *et. al.* (1988) was uncertain due to bad preservation, and suggests that the specimens may possibly belong to *C. andresi*. No further research on Cambrian conodonts from the Oslo area is done, and the zonal scheme proposed by Szaniawski and Bengston (1998) modified from Kaljo *et. al.* (1986) has been the scheme used for Upper Cambrian conodont zonation of Baltica. The specimens of the species *C. proavus* found in this study however, support the results by Bruton *et. al.* (1988) that *C. proavus* is present in pre-Tremadocian strata in Scandinavia, and hence, may be correlated to the *C. proavus* zone from North America and Estonia.
Figure 63. Correlation of the *Cordylodus proavus* Zone of Scandinavia with North America and Estonia (modified from Szaniawski and Bengtson, 1998).

7.1.2 Fauna assemblage

The conodont faunas from this study contain typical taxa for the cold-water realms (Müller and Hinz, 1991), except the fauna in sample KAM6, which includes *Cordylodus proavus*, which in turn is a species appearing both in the warm- and cold-water realm (Miller, 1984). All faunal elements present, except *C. proavus* are recorded from Sweden in age equivalent deposits, and hence do not provide new information regarding the already associated faunal elements or paleoecology (Müller and Hinz, 1991; Bagnoli and Stouge, 2013).

The distribution of protoconodonts and paraconodonts, which are common contributors to the Alum Shale fauna, fits well with the correlation of conodonts with trilobite zones done by Müller and Hinz (1991). The proto- and paraconodonts are interpreted as reflecting the same environment and hence, discussion based on small changes within these genera in the different samples may not be regarded as reflecting changes of the environment. The absence of representatives from the expected protoconodonts and paraconodonts in some of the samples is probably due to loosing specimens during preparation, or the variability of abundance and preservation within the same horizon, rather than faunal changes. However,
different fauna associations may reflect difference in water depth, as proposed by Bagnoli and Stouge (2013).

The paraconodont fauna comprises primarily *Westergaardodina* which is a common genus in the Alum Shale Formation (see Müller and Hinz, 1991). In general, more robust conodont elements, which are characteristic for the *Westergaardodinidae* in KAM5 as well as KAM2, may reflect higher energy environments. However this interpretation is more widely used for younger conodont faunas. The difference in conodont sizes between samples may be due to other factors rather than reflecting different energy settings. It is however worth noting that the samples containing larger elements are both reflecting high energy deposits (see section 6.3). However the specimen of *Trolmenia acies* present in KAM5 is rather small and fragile. This species in addition to the other paraconodont species in KAM5, makes this sample correlative with the *Westergaardodina* association proposed by Bagnoli and Stouge (2013).

Based on the small amount of species recorded from this study, conclusion regarding faunal association is uncertain, but of the three associations proposed by Bagnoli and Stouge (2013) this sample at least contains species characteristic for the *Westergardodina* association. This association reflects deposition in shallow water environments.

The only proconodont species is *C. proavus* from sample KAM6. According to Miller (1984), representatives from the order Proconodontida, including *Cordylodus* represent shallow higher energy environments, but *Cordylodus* was cosmopolitan during the Late Cambrian, and hence, lived in both faunal realms. Contrary to the proto- and paraconodonts taxa this genus probably tolerated higher salinity environments. This might explain the absence of paraconodonts in this sample. However, one single protoconodont of the genus *Phakelodod* is present in the same sample, and hence the absence of paraconodonts is probably due to variations of preservation and abundance rather than elevated salinity. As for the other samples, the few recorded conodont elements make paleoecological interpretations uncertain.

The absence of representatives from the paraconodont genus *Furnishina* may be of environmental importance. This genus, as well as *Westergaardodina*, is one of the most common and abundant genera in the Alum Shale from Sweden (Müller and Hinz, 1991; Bagnoli and Stouge, 2013). Based on the paraconodont associations proposed by Bagnoli and Stouge (2013), the absence of *Furnishina* in the samples may reflect a shallower water environment. Representatives from the paraconodont genus *Prooneotodus* which among others comprise the *Prooneotodus* association are not present either. The absence of
representatives from both *Furnishina* and *Prooneotodus*, in addition to the presence of *Westergaardodina* in most of the samples, may reflect an environment comparable to the water depth for the *Westergaardodina* association. This reflects shallow water, high energy environments during deposition, correlating with the high energy domain as shown in Figure 64.

![Storm activity diagram](image)

**Figure 64.** Depositional model for an epeiric sea (After Waters and Sando (1987) and James and Dalrymple (2010)).

Due to the low abundance of conodonts from the sample KAM6 (*Acerocarina* superzone) and the absence of conodonts from the sample KAM7 (*Boeckaspis* trilobite zone) correlation with the conodont ranges and faunas reported from the Oslo region by Bruton *et al.*, (1988) is difficult. The differences is probably due to variations in abundance and preservation, especially based on the absence of conodonts in sample KAM7, compared to the relative high conodont abundance from the same zone at Nærnses. However, conodonts from two equal genera are reported from the *Acerocarina* superzone from this study and from Nærnses; *Phakelodus* and *Cordylodus*. Both genera are present in older and age equivalent deposits from Sweden (Müller and Hinz, 1991; Bagnoli and Stouge, 2013). The genera recorded from Nærnses, includes *Eoconodontus* which lived in the same environment as *Cordylodus* (Miller, 1984), and is a common species in Baltica after the appearance of *C. proavus* (Bagnoli and Stouge, 2013) which may explain their co-occurrence.

However, despite the rather high abundance of conodonts from Nærnses, the absence of *Westergaardodina* makes it more possible that this genus may not be present due to changes in the environment.
7.1.3 Color alteration index (CAI)
The CAI of the conodonts is rather high due to the thermal alteration these deposits have been exposed for due to primarily intrusions associated with the Permian rifting. The conodonts show a CAI value of 4-4.5, which indicates a heating of 190°C to more than 300°C. The higher CAI of some of the conodonts, may be due to locally higher temperatures reflecting different distances from intrusions. The presence of a sill a few meters above the deposits containing the sample GIBB13, may have caused the slightly higher CAI of these conodonts (Figure 37). Such high temperatures cause the hydrocarbons in the Alum Shale at Slemmestad to be supermature, as shown in figure 65.

![Figure 65. Correlation of CAI with organic metamorphic facies and associated hydrocarbons (Metcalfie and Riley, 2010).](image-url)
7.2 Micro facies analysis and depositional environment

7.2.1 Matrix
The matrixes of the finer grained recrystallized carbonate rocks (Facies 1) are classified as Neomorphized Recrystallized Limestones based on texture (Flügel, 2010). Recrystallized carbonate rocks develop microspar and pseudospar with increasing crystal sizes respectively (Flügel, 2010), reflecting increasing grade of aggrading neomorphism. Such recrystallization is controlled by burial diagenesis, pressure and changes in temperature (Flügel, 2010). These recrystallized limestones may also have been affected by the high temperatures, and intrusion during the Permian rifting, also indicated by the high CAI of the conodont elements. Even though the matrix of the neomorphized recrystallized limestones is recrystallized, some of the samples show relict depositional structures. The matrix of the coarser, pack to grainstone facies (Facies 2) is of sparitic composition. Growth of such carbonate cement is favored by high pH and higher temperatures (Flügel, 2010). Sparite and pseudospar are distinguished based on texture since both are coarse grained (Flügel, 2010).

7.2.1.1 Neomorphized recrystallized limestones
The finer grained matrix in the neomorphized recrystallized samples shows pseudosparitic crystals ranging in sizes from several tenths to hundreds microns. Pseudospar is formed by continuing diagenesis of microspar (5µm – 30 µm). A diagnostic feature for neomorphized recrystallized limestones is crystals of different sizes, with mainly curved interfaces between crystals and larger crystals associated with the smaller crystals (Flügel, 2010) (Figure 54, section 6.3.1). However, based on the rather homogenous texture and smaller crystal sizes of some of the samples within Facies 1, they show an earlier stage of the aggrading neomorphic process relative to the coarser samples, and may be intermediates between microspar and pseudospar.

There are several explanations of how microspar is formed. Among these are 1) recrystallization of micrite due to aggrading neomorphism, where crystal size increase and finer crystal mosaics are replaced by coarser crystal mosaic as shown in Figure 66. 2) recrystallization of silt-sized carbonate grains and 3) one-step neomorphic process of cementation and replacement of aragonite-dominated precursors to microspar (Figure 66) (Flügel, 2010).
Recrystallized carbonate rocks may lose depositional characteristics, depending on the degree of recrystallization (Flügel, 2010). In some of the finer grained samples, relict laminas, as well as relict skeletal grains are preserved.

Relict structures of grains may be due to a gradual in situ dissolution-reprecipitation process of aragonite by calcite, so that the original skeletal grains may be preserved as these structures. (Flügel, 2010). The few skeletal grains present in some of the thin sections may be preserved due to their originally calcite composition which is more resistant than aragonite. Aragonite is metastable, hence with time, all carbonate sediments are transformed to calcite. Dissolution of aragonitic skeletal grains causing secondary porosity, and aragonite crystals are replaced by larger calcite crystals which may explain the larger crystals in some of the samples.

Figure 66. Microspar-forming processes (Flügel, 2010).
Due to the high amount of bitumen, which is not dissolved during recrystalization, the calcite crystals are surrounded by a thin layer of bitumen and are hence separated from each other, not allowing typical mosaic texture to form.

7.2.1.2 Sparite
The coarser pack- to grainstone facies (Facies 2) has sparitic texture due to early cementation. The cement is pore filling both within and between the skeletal grains, representing growth from a free substrate into the pore space (Flügel, 2010). Such crystal growth may form different types of cement based on texture. The texture of this facies represents radiaxial fibrous cement (Flügel, 2010), which is interpreted as cements of for example syndepositional marine and shallow-marine origin. Unlike the diagnostic features of neomorphized recrystallized limestones mentioned above, the boundaries between the pore filling sparite and the pore limits as well as the crystal boundaries are sharp.

The favored conditions for carbonate cementation growth are high pH and high water temperatures (Flügel, 2010). A high carbonate input in addition to an efficient fluid flow is required for cementation of carbonates at or near the sediment-water interface. The source of carbonate in burial environments is usually from the sediment (Flügel, 2010). Several factors control the cementation of carbonate and the dissolution of carbonate. Some of the main factors are the primary porosity and permeability, the composition of the pore fluids, the flow rate of water through the pores which is dependent of energy levels. In shallow marine environments water energy and the sedimentation rate are also important factors.

7.2.2 Facies interpretation

7.2.2.1 Neomorphized Recrystallized Limestones (Facies 1)
Due to recrystallization, sedimentary structures are not so easily visible in most of the samples from this facies. In terms of facies interpretation all the neomorphized recrystallized limestones are in the same category. They show some differences in grain sizes, but due to the lack of structures, as well as grains in some of the samples, the facies interpretation is based on the similarity of the thin sections. Facies interpretation of such recrystallized limestones is difficult, hence some of the interpretation of this facies is uncertain. Samples providing information on depositional environment such as relict structures, preserved skeletal grains and siliciclastic input is considered as the basis for this facies interpretation.

In some of the samples of this facies, relict laminations are slightly visible, showing normal grading. This may be recrystallized coarse suspension layers reflecting episodic high energy
events. One of the samples (KAM7) of this facies which otherwise not contain any structures or fossils, however contain higher amount of siliciclastic material, which therefore also may reflect calmer environments allowing fine grained terrigenous material to settle from the water column. Sample KAM2 shows no depositional structures due to recrystallization. However, this sample is taken from the *Parabolina* superzone which also includes high amounts of the brachiopod *Orusia lenticularis*. This species is preferably living on firmer substrates (Lehnert *et al*., 2012), which may reflect shallower water, higher energy and oxygenated bottom conditions which correlates with the non-olenid trilobite fauna (Figure 18).

The well preserved small possibly bivalved crustaceans in GIBB13 may be representatives from the arthropod order Bradoriida. These reflect oxygenated bottom conditions where benthic living organisms could live. Representatives of this order from age equivalent deposits in Australia reflect benthic mode of live (Walossek *et al*., 1993), and may suggest that the bradoriids in this sample also were benthic living organisms. This sample shows relict normal grading and skeletal grains parallel to bedding reflecting currents reaching the sea floor, or higher energy environment. This sample may reflect the shallow environments with high energy proposed by Bjørlykke (1974) as the depositional environment for Middle Cambrian deposits of the Alum shale in the Oslo area. Oxygenated bottom conditions and shallow water environments correlate with the non-olenid trilobite fauna (Figure 18), which also correlates with the high abundance of agnostid trilobites in the *Paradoxides paradoxissimus* superzone. GIBB13, however, is the only sample of this facies with bradoriids, hence oxygenated bottom conditions during deposition is not necessarily representative for the other samples within this facies based on thin section analysis. The presence of this group in some of the samples however, reflects periods of oxygenated bottom conditions.

7.2.2.2 *Carbonate Skeletal Pack- to Grainstone (Facies 2)*
The large grain sizes, low content of mud, lamination and the abundance of marine fossils and their orientation, probably reflects deposition in environments with periodic high energy events, were fine grained material was probably winnowed away. The orientation of the fossil fragments reflects an environment without any bioturbation, which may reflect a bottom environment without living organisms probably due to anoxic conditions. The larger grains are mainly facing convex side up, used as an indicator for water currents (Eklöf *et al*., 1999).
This facies shows early cementation, and the calcite cementation characteristic for this facies also requires flow rate to supply sufficient carbonate (Flügel, 2010). This flow is a result of higher energy environment during deposition, due to the syndepositional cement growth of such calcite crystals (Flügel, 2010).

Within this facies laminas are present and clearly visible, reflecting varying energy during deposition. The presence of cross lamination reflects a shallower water environment. This reflects periods where the ocean currents have reached the sea bottom and have eroded the sea floor, which causes the formation of cross laminations. However, the thin walled trilobite exoskeletons are often well preserved, especially in the most cemented layers, which in higher energy would be fragmented. This probably reflects rapid burial, and contrary to the layers with cross lamination, they were not further eroded by ocean currents reaching down to the sea floor. The few ostracods present in this facies occur in these layers. These were probably part of the pelagic fauna.

Geopetal structures which are formed when cavities in fossils are filled with calcite cement are also common in this facies, and require favorable conditions. Such geopetal structures are commonly formed in rapidly buried fossils during storm events, probably reflecting the conditions during deposition of this facies (Wieczorek, 1979; Flügel, 2010). This is in addition to cross laminations and the packstone and grainstone layers, typical for storm deposits (Flügel, 2010), suggest that this facies was deposited during high energy events.

The suggested anoxic environment due to the lack of bioturbation correlates to the olenid trilobite fauna (Figure 18), which is also supported by the trilobites in the Peltura superzone. However, based on the structures mentioned above deposition of this facies requires relatively shallow water conditions where currents in periods have reached the sea floor.

7.2.2.3 Carbonate Packstone (Facies 3)
This facies consists of high amount of trilobite skeletal grains, and finer grained recrystallized carbonate matrix as the main pore filling component in this facies. No bioturbation is present, suggesting anoxic environment. Some of the skeletal grains are partly recrystallized, and cementation occurs within few of the grains. The skeletal grains show no trend in orientation, and are well preserved. This probably excludes bed load transport or current flow, and suggests a rapid burial with no further erosion, transportation or destroying of the skeletal grains. The skeletal grains are of rather equal size, and show no sign of abrasion or bioerosion, which in addition to the lack of orientation and well preservation is characteristic
for tempestites (Flügel, 2010), which is a storm deposit. Hence, this facies was deposited during high energy events.

The lack of bioturbation, suggesting anoxic environment also correlates to the olenid trilobite fauna (Figure 18), which is also supported by the trilobites in the Peltura superzone. However, this facies was probably deposited in relatively shallow water exposed for higher energy events.

7.2.2.4 Carbonate Wacke- to Packstone (Facies 4)
This facies is characterized by clearly differentiated laminas of wacke- and packstones. Such clearly alternating layers indicate, according to Flügel (2010), environments where current flow varies considerably. However, based on the orientation of the fossil fragments, no specific sign of current is present. If currents were present during deposition the fragments would have been oriented parallel to each other and subparallel to bedding plane, and the convex side would have been facing upward (Eklöf et. al., 1999). Based on this, it may seem that these beds were deposited due to accumulation of dead organisms sinking from the water column. They would then have fallen randomly down, with the concave side dominating upward, which are seen at the base of the laminas. Their somehow parallel orientation to the bedding may also have been due to the compression of overlying sediments after deposition, and may have tilted upward standing fragments with both convex and concave side facing upward, as seen in some of the smaller fossil grains. However, very small grains may behave as small particles, hence not so dependent on which side facing upward during deposition. Hence the larger grains are more reliable for indicating currents during deposition. This interpretation, as well as the absence of bioturbation, suggests that this facies was deposited under calm, anoxic environment.

The material that formed the packstone beds has probably been transported basin ward from shallower environments by suspension flows during storm events. The material was deposited below storm wave base, and created these sharp basal contacts with underlying beds in calm anoxic environments. The wackestone layers between the packstone laminae are reflecting periods of calm environments with less input of skeletal material, where pelagic organisms have randomly fallen down from the water column.

The possible small bivalved arthropods may be representatives from the order Archaeocopida due to the calcareous carapace, common within this order (Armstrong an Brasier, 2009). They
are also interpreted as having had a pelagic mode of life due to the presence in Cambrian pelagic limestones (Lehmann and Hillmer, 1983).

The mode of life for calcispheres is difficult to elaborate due to their unknown origin (Berkyová and Munnecke, 2010). However, they are often present in pelagic sediments (Flügel, 2010), hence representing a pelagic mode of life in such deposits, which may be the depositional setting for this facies.

There is a possibility that some of the calcispheres, especially in the packstone beds with high amount of trilobite fragments, may be trilobite legs in cross section. Trilobite legs cut longitudinally may then represent some of the larger elongated grains parallel to the bedding plane (Figure 56A). This may also reflect calm environment resulting in fossil grains parallel to the bedding plane, but with no systematic orientation relative to each other.

The high content of fine grained sediments, both within the wacke- and packstone reflects deposition in calm environments, allowing fine grained material to settle. The size of the calcispheres and other fragments, also suggests particles settling from the water column in calm environments. Transportation of such larger grains relative to the fine grained matrix would have required energy levels resulting in winnowing of the finer grained particles.

The calm environment, probably due to higher sea level, and the anoxic bottom conditions correlate to the olenid trilobite fauna (Figure 18) also supported by the trilobites in the Peltura superzone.

7.2.2.5 Massive Clay-rich Mudstone (Facies 5)

This facies consists of massive clay-rich mudstone with corroded silt sized quartz particles and high content of bitumen. No fossil grains were recorded in this facies, but bioturbation is present at the top of the beds. Traces of graded bedding are present, but partly destroyed probably due to bioturbation resulting in mixed grain sizes. The burrows are probably fodinichinas, with possibly representatives from the ichnogenus Phycosiphon.

The silt sized quartz grains are corroded, reflecting abrasion during longer transportation, which compare well with the more distal part of the shelf. Input of such terrigenous material into carbonate environments may be due to for example fluvial transport or eolian transport (Flügel, 2010). The latter is most probable in this setting due to the geological setting, but the sediments may have been deposited more proximal, and transported with low energy currents more distally before settling.
The fine grained sediments could have been deposited by suspension or by low energy currents. Despite the mixed grain sizes, a normal grading trend is visible, reflecting a decrease in energy during deposition, typical for deposition by currents. *Phycosiphon* ichnogenus is common in fine grained sediments and representatives from this genus indicate that the conditions were oxygenated during deposition, and that the substrate probably was firm (MacQuaker *et al.*, 2007). This ichnogenus is euryhaline (McIlroy, 2004), and is hence present in environments with variable salinity. This ichnogenus is often a part of the *Cruziana* ichnofacies, with ichnogeneras characteristic for low energy off-shore marine settings such as shelves or epeiric seas (Gibert and Martinell, 1999; Hastois, 2013). *Phycosiphon* is often used as an indicator for location within basins, because it is interpreted as reflecting the distal part of shelves, which also correlates with *Cruziana* distribution on the shelf, as shown in figure 67. This correlates with the depositional setting as well as the fine grained material in this facies, reflecting low energy environments. However, distal parts of *Cruziana* ichnofacies, often overlaps with the *Zoophycos* ichnofacies, characteristic for anoxic deposition and sediments containing high content of organic matter. Despite the deeper location of *Zoophycos* ichnofacies (Figure 67), this ichnofacies is common in shallower oxygen restricted environments and are used as indicator for anoxic depositional environments (James and Dalrylumpe, 2010). Hence, it is possible that these two ichnofacies in periods have overlapped each other, resulting in this massive clay-rich mudstone facies.

Despite the fine grained material reflecting calm environment probably due to higher sea level, the presence of *Phycosiphon* and the trilobites in the *Acerocarina* superzone correlating with the non-olenid trilobite fauna probably supports the dysoxic environment during the latest Cambrian (Terfelt *et al.*, 2013).
7.2.3 Acid insoluble residue
Biogenic material was recovered from the acid insoluble residue due to non-carbonate composition, and may also provide information regarding depositional environment. A discussion regarding the different biogenic material of importance for environmental interpretations is presented in this section. Some of the biogenic material of uncertain biological origin is also discussed.

7.2.3.1 Brachiopods
Brachiopod fragments were present in the lowermost sample, GIBB13, and one specimen in KAM5. The brachiopod fragments in GIBB13 are fragmented, while the one specimen from KAM5 is well preserved. Such small phosphatic inarticulate brachiopods present in GIBB13, from the Middle Cambrian (*Paradoxides paradoxissimus* superzone) are also found in Middle Cambrian deposits from for example Ritland (Bruton and Harper, 2000), and the brachiopod fragments from the current study may belong to the order Acrotretida, based on description by Bruton and Harper (2000). This order was common in a wide range of marine environments during the Cambrian (Bassett et. al., 1999), and hence do not provide information regarding special environmental settings such as depth or energy. However, autochthonous brachiopods represent oxygenated bottom conditions during deposition, and therefore represent the same environment as for the non-olenid trilobites (Figure 18) and may therefore explain the presence of brachiopods within this sample, representing the non-olenid fauna.

However, the possible brachiopod fragments from this sample are all fractured, probably representing high energy shallow water environments resulting in breakage. This
interpretation fits well with the shallow, high energy environment suggested by Bjørlykke (1974) as the environment these sediments were deposited in during the Middle Cambrian in Oslo region.

The brachiopod shell from sample KAM5 on the other hand, is well preserved, and shows no sign of breakage. This may be explained by deposition of tempestites, where fossils often are well preserved (Flügel, 2010). The presence of only one valve also support this theory rather than a faunal component which would have contained more brachiopod specimens if a faunal community was rapidly buried in situ. This explanation suggests unfavorable bottom conditions for brachiopods which may be explained by anoxic bottom conditions, as proposed by Bjørlykke (1974) for these deposits in Upper Cambrian in Oslo region, and correlates with the olenid trilobite fauna (Figure 18).

7.2.3.2 Ostracods
The phosphatocopine ostracod possibly of the order Bradoriida present in sample PEL13 is well preserved with both valves.

Mode of life for phosphatocopids (benthic, necto-benthic, pelagic) lacks undisputed evidence (Vannier and Walossek, 1998), but their occurrence in black mudstones favor a nektonic or pelagic mode of life due to their dependence of oxygenated conditions, which may have been the mode of life of the phosphatocopine present in this sample. This specimen would probably have become disarticulated during high energy or transportation. Based on the occurrence in the olenid trilobite fauna, reflecting unfavorable bottom conditions due to anoxic environments also support the possibility for this specimen as a part of the pelagic fauna, deposited during calmer periods and then possibly rapid buried, allowing preservation of both valves (Figure 59A).

Another possible phosphatocopine was also found in this sample. However this specimen is uncertain due to preservation. The fractures on the shell may be due to post-burial deformation. However, the shape, size and the trace of a midline between two valves, makes this specimen possibly a phosphatocopine rather than for example a phosphatic ball from Westergaardodina.

7.2.3.3 Trilobites
Based on the knowledge of the different trilobite fauna associated with the different samples, the trilobite fragments do not provide additional information regarding the oxygen conditions near the sea floor. However, they may reflect chemical components available resulting in for
example post-deposition precipitation of gypsum, but are not of relevance for this study. The casts of the perimorphosed trilobite fragments from sample KAM2 and KAM4 both representatives from Facies 1, however, include well preserved casts of the trilobite exoskeleton, probably reflecting rapid burial, or calm environment resulting in low grade of, or no disarticulation of the thorax segments.

7.2.3.4 Biogenic material of uncertain origin
The spheric balls from the sample KAM5 are pyritized and rather large. These balls are probably representatives of pyritized balls typical for *Westergaardodina* conodonts as shown in Müller and Hinz (1991, plate 31, fig. 10; plate 32, figs. 13-14). In KAM5, larger pyritized *Westergaardodinas* are present, which make it possible that these balls originate from these conodonts. Likewise, the possible phosphatic balls from the sample PEL13 are smaller, and then correlate with both size and composition with the smaller *Westergaardodinas* from the same sample. This suggests that these balls may have a conodont origin.

Based on the consistency of fecal pellets, exceptional preservation is required to preserve such soft material. High energy or transportation would dissolve the pellets, which makes them indicators for calm anoxic bottom environment (Robbins *et. al.*, 1985), deposited from organisms living higher in the water column. However, according to Friis (1994) fecal pellets may be preserved in higher energy environments if they are deposited as aggregates of mud and buried rapidly, instead of settling down from suspension.
8 Conclusions

Based on microfacies analysis and microfossil study of the Alum Shale ranging in time from the Middle Cambrian *Paradoxides paradoxissimus* trilobite zone to the Early Ordovician *Boeckaspis* trilobite zone environmental interpretations are presented. An overall upward deepening trend is recognized based on increasing amount of fine grained particles, from shallow water with oxygenated bottom conditions inhabited by benthic living organisms in the Middle Cambrian to the earlier parts of the Late Cambrian deposits. Primarily anoxic bottom conditions without benthic living organisms existed during the later part of the Late Cambrian. However, alternating shallower high energy episodes and deeper calm environment conditions during the Late Cambrian probably reflects sea-level fluctuations.

One type of trace fossil was found, probably a species of the ichnogenus *Phycosiphon*, supporting the latest Cambrian dysoxic environment.

All the conodont faunas are typical cold water realm forms. By comparing the paraconodont species from this study with the paraconodont associations proposed by Bagnoli and Stouge (2013), the absence of the common conodont species of the genus *Furnishina* and the dominance of species from the genus *Westergaardodina* may reflect that the sea level was relatively low during deposition of the paraconodont bearing sediments. This is supported by their occurrence in shallower marine high energy deposits as interpreted from the microfacies analysis.

The presence of the species *Cordylodus proavus* in the *Acerocarina* superzone, which represents Upper Cambrian strata, may be regarded as the first occurrence in Scandinavia. Based on this correlation with the global *C. proavus* zone can be done.

The conodonts show a color alteration index (CAI) of 4 – 4.5. This implies that the sediments were exposed to temperatures from 190°C to more than 300°C, which can primarily be explained by the high temperatures associated with Permian intrusion common in the Slemmestad area. These high temperatures have made this otherwise potential source rock supramature.

**Further research**

Further research on condonts from the Cambrian Alum Shale at Slemmestad should include larger samples so that quantitative analyses can be done. This should be done in order to see if there is a correlation between conodont associations and lithology and hence environment.
9 References


**Web references**


Loucks R.G., Kerans C. and Janson X. 2003. Introduction to Carbonate Environments, Facies and Facies Tracts. Accessed 29.05.2014 from [http://www.beg.utexas.edu/lmod/_IOL-CM01/cm01-step03.htm](http://www.beg.utexas.edu/lmod/_IOL-CM01/cm01-step03.htm)


Appendix

Appendix 1 Preparation of samples.

Table 4. Acid preparation summary of samples. Additional samples were collected, but did not provide any further information.

<table>
<thead>
<tr>
<th>Code</th>
<th>Locality</th>
<th>Zone</th>
<th>Weight before (kg)</th>
<th>Weight After (kg)</th>
<th>Dissolved (g)</th>
<th>Fraction weight ≤5-500 μm (g)</th>
<th>Fraction weight &gt; 500 μm (g)</th>
<th>Initiated</th>
<th>Finished</th>
<th>Comments</th>
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<td>Peltrea led</td>
<td>7.00</td>
<td>0.29</td>
<td>6.71</td>
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<td>31 jan</td>
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<td></td>
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<td>GSST13</td>
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<td>1.55</td>
<td>5.45</td>
<td>285.00</td>
<td>31 jan</td>
<td>Light grey fraction</td>
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<td></td>
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<td>KAM1</td>
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<td>1.30</td>
<td>3.70</td>
<td>70.00</td>
<td>21.16</td>
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<td></td>
</tr>
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<td>0.58</td>
<td>4.44</td>
<td>50.28</td>
<td>11.93</td>
<td>11 aug</td>
<td>Black, root appearing fraction</td>
<td></td>
</tr>
<tr>
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<td>Selmerstad</td>
<td>Parabola heris</td>
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<td>1.03</td>
<td>3.97</td>
<td>42.43</td>
<td>17.79</td>
<td>13 mai</td>
<td>06 jun</td>
<td>Black, root appearing fraction</td>
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<tr>
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<td>Peltrea scab. wester</td>
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<td>4.96</td>
<td>67.21</td>
<td>24.47</td>
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<td>NERASNES A</td>
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<td>Acerocar ecos</td>
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<td>0.96</td>
<td>3.04</td>
<td>126.53</td>
<td>35.47</td>
<td>13 dec</td>
<td>07 feb 14</td>
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<td>Nersnes, beach, sodde B</td>
<td>Acerocar ligata</td>
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<td>4.07</td>
<td>330.41</td>
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<td>07 feb 14</td>
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</table>

*See Brun et al. 1988

**Finished late due to low amount of residue. Was acid prepared a few additional rounds in august.

Table 5. Summary of heavy liquid separation.

<table>
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<th>Sample</th>
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<th>Fluid density start (g/cm3)</th>
<th>Fluid density finish (g/cm3)</th>
<th>Finished</th>
<th>Number of fractions</th>
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<td>2.97</td>
<td>2.46</td>
<td>17oct</td>
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<tr>
<td>KAM4</td>
<td>22aug</td>
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<td>2.20</td>
<td>22aug</td>
<td>13</td>
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<td>KAM5</td>
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<td>3.05</td>
<td>2.83</td>
<td>03.sep</td>
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<td>&gt;2.85</td>
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<td>05.sep</td>
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## Appendix 2 Raw data from thin section counting

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KAM7, PMO 221.702

Spectra 1 = calcite, Spectra 2 = Quartz, Spectra 3 = Pyrite

Figure 68. BSE image of KAM7

Figure 69. Qualitative SEM EDS spectra from KAM 7.

Spectra 1 = calcite, Spectra 2 = Quartz, Spectra 3 = Pyrite
Spectra 1 and 2 = Pyrite, Spectra 3 = Dolomite, Spectra 4 = Calcite

Figure 70. BSE image of KAM7

Figure 71. Qualitative SEM EDS spectra from KAM 7.

Spectra 1 and 2 = Pyrite, Spectra 3 = Dolomite, Spectra 4 = Calcite
Figure 72. BSE image of KAM8.

Figure 73. Qualitative SEM EDS spectra from KAM 8.

Spectra 1= Calcite, Spectra 2=Quartz, Spectra 3=Muscovite, Spectra 4 = Pyrite
Appendix 4 Evidence of gypsum perimorphosis

Trilobite from KAM4, PMO 221.738/8.

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Figure 70. BSE image of KAM7.

Figure 71. Qualitative SEM EDS spectra from KAM7.

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Figure 744. BSE image of KAM4.

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