Norwegian University of
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A Sedimentological Study of the De Geerdalen Formation with Focus on the Isfjorden Member and Palaeosols

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Abstract

In this study sedimentological depositional environments of the Upper Triassic De Geerdalen Formation in Svalbard have been investigated. Facies and facies associations of the whole formation are presented, however the main focus has been on delta top sediments and in particular palaeosols. Special attention has been paid to the Isfjorden Member, which constitutes the uppermost part of the De Geerdalen Formation. The purpose of the study has been to identify palaeosols, and relate them to the overall depositional environments. The palaeosols have been identified by three main characteristics: roots, soil horizons and soil structure. Based on field observations an attempt to classify the palaeosols has been made.

There are notable differences between brown and yellow palaeosols found in the middle and upper parts of the De Geerdalen Formation and the red and green palaeosols restricted to the Isfjorden Member. The yellow and brown palaeosols are in general immature compared to the green and red palaeosols of the Isfjorden Member. Thin sections from the Isfjorden Member on Deltaneset show excellent examples of calcrete, with clear biogenetic indicators. Distinct and alternating green and red colours might be related to fluctuations in groundwater level and reduction and oxidation of the soil profile. The palaeosols are found on floodplains, interdistributary areas and on top of proximal shoreface deposits. The number, thickness and maturity of palaeosols tend to increase upwards in the De Geerdalen Formation.

The Isfjorden Member is present on Wilhelmøya, Hahnfjella, Hellwaldfjellet, Teistberget, Klement'evfjellet, Friedrichfjellet, Schmidtberget and Deltaneset. The Isfjorden Member is not present at Krefftberget, Svartnosa and Blanknuten probably due to erosion of the upper part of the De Geerdalen Formation. The Isfjorden Member is easily recognized by coquina beds and alternating red and green mudrocks, but the lower base of the member is not clearly defined.

Field-data were collected in 2014 (5th -23th of August) and 2015 (2nd of August to 6th of September). In this thesis mainly data from 2015 is presented. The localities visited are situated on Edgeøya, Barentsøya, Wilhelmøya, Agardhdalen, and west Spitsbergen. Contemporary field work was conducted in order to collect the data. In addition XRD and thin section analyses have been performed on selected samples.

Collected data is compared with previous studies.

Sammendrag

I dette studiet har sedimentologiske avsetningsmiljøer i De Geerdalenformasjonen fra øvre trias på Svalbard blitt undersøkt. Facies og faciesassossiasjoner i hele formasjonen er presentert, men hovedfokus har vært på delta topp sedimenter, og spesielt palaeosoler. Spesiell oppmerksomhet har blitt rettet mot Isfjordenleddet som utgjør øvre del av De Geerdalenformasjonen. Målet med studiet har vært å identifisere palaeosoler og relatere dem til generelle avsetningsmiljøer. Paleosolene har blitt identifisert ved hjelp av tre hovedkjennetegn: røtter, jordhorisonter og jordstruktur. Basert på feltobservasjoner er det blitt gjort et forsøk på å klassifisere paleosolene.

Det er nevneverdig forskjeller på de brune og gule paleosolene funnet i midtre og øvre deler av De Geerdalenformasjonen og de røde og grønne paleosolene avgrenset til Isfjordenleddet. De gule og brune paleosolene er generelt umodne sammenlignet med de grønne og røde paleosolene i Isfjordenleddet. Tynnslip fra Isfjordenleddet på Deltaneset viser fremragende eksempler på calcrete, med klare biogenetiske indikatorer. Tydelig og vekslende røde og grønne fargeforandringer kan være relatert til endringer i grunnvannsnivå, som fører til reduksjon og oksidasjon av jordprofilet. Paleosolene er funnet på flomsletter, i mellomavsetningsområder, og på toppen av proksimale strandskråningsavsetninger. Antall, tykkelse og modenhet til paleosolene har en tendens til å øke oppover i De Geerdalenformasjonen.

Isfjordenleddet er til stede på Wilhelmøya, Hahnfjella, Hellwaldfjellet, Teistberget, Klement`evfjellet, Friedrichfjellet, Šmidtberget og Deltaneset. Isfjordenleddet finnes ikke på Krefftberget, Svartnosa og Blanknuten, sannsynligvis på grunn av erosjon av øvre De Geerdalenformasjonen i kvartær. Isfjordenleddet er lett gjenkjennbar på lag av skjell og vekslende røde og grønne slamsteiner, men nedre grense er ikke klart definert.

Feltdata ble samlet inn i 2014 (5.-23. august) og 2015 (2. august-6 september). I denne oppgaven er hovedsakelig data fra 2015 presentert. Lokalitetene som ble besøkt befinner seg på Edgeøya, Barentsøya, Wilhelmøya, Agardhdalen og vestre Spitsbergen. Moderne feltarbeid ble utført for å samle inn data. I tillegg har XRD- og tynnslipsanalyser blitt utført på utvalgte prøver.

Innsamlet data er sammenliknet med tidligere studier.

Acknowledgement

First of all I would like to thank my main supervisor Atle Mørk for giving me the opportunity to participate in field work in Svalbard. It has been three fantastic summers of field work with life lasting memories. Thanks also for great support, help and always taking time to proof read my drafts.

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The Norwegian Petroleum Directorate, SINTEF and UNIS are thanked for organizing and financing field work and for logistic support. Without them the field work would not have been possible. Thanks to Christian Haug Eide for useful field days on Deltaneset, and for emphasizes some of the diagnostic features of palaeosols. Maria Jensen at UNIS is thanked for taking time to discuss the interpretations of the palaeosols with me.

Thanks to Venke in the reception for making every day at UNIS starting with a smile, and to the librarian ladies at UNIS for always being helpful with finding articles and literature.

Finally, I wish to thank friends and family for great support and patience during the writing of my master thesis.

Preface

This study is a part of a master`s degree in geology organized as a cooperation between the Norwegian University of Science and Technology (NTNU) and the University Centre in Svalbard (UNIS). Main super visor for the study has been Professor II Atle Mørk at NTNU. Professor Snorre Olaussen at UNIS has been co-supervisor.

Several students from NTNU have been working on the Upper Triassic succession on Svalbard recent years. The thesis presented herein is a part of this ongoing work. Cooperation both in the field and later during data analysis has been important in this study. Chapter 6 and 7 are written together with fellow master students at NTNU Sondre Krogh Johansen and Simen Jenvin Støen. The chapters are also included in their master theses which they both submitted in June this year.

All pictures in this thesis are taken by students at NTNU the summer of 2015 if otherwise is not specified.

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1. Introduction

1.1 Purpose of Study

The scope of the study is to investigate sedimentological depositional environments of the Late Triassic De Geerdalen Formation in Svalbard. The thesis is a part of a team work with fellow master students at NTNU Sondre Krogh Johansen and Simen Jenvin Støen. Their master theses focus in on the lower reaches of the De Geerdalen Formation and sandstones. The main focus in this study is in on delta top sediments and in particular palaeosols. The Isfjorden Member constituting the upper part of the De Geerdalen Formation is studied in detail. Following topics are discussed i)Facies distribution in the De Geerdalen Formation, ii) Distribution of the Isfjorden Member, iii) Distribution of palaeosols in the De Geerdalen Formation, iii) Palaeoclimate and palaeo geographic implications in the Isfjorden Member. The data is primarily based on field observations; however petrographic thin sections and XRD-analysis are included in this work.

1.2 Study area

The archipelago of Svalbard includes all islands between 74 and 81 degrees north and 10 and 35 degrees east (Fig. 1.1). In this study the outcrops of the Northern Storfjorden area and western Spitsbergen were visited. Data from Deltaneset on central Spitsbergen, Agardhdalen on eastern Spitsbergen, Edgeøya, Barentsøya and Wilhelmøya are presented.

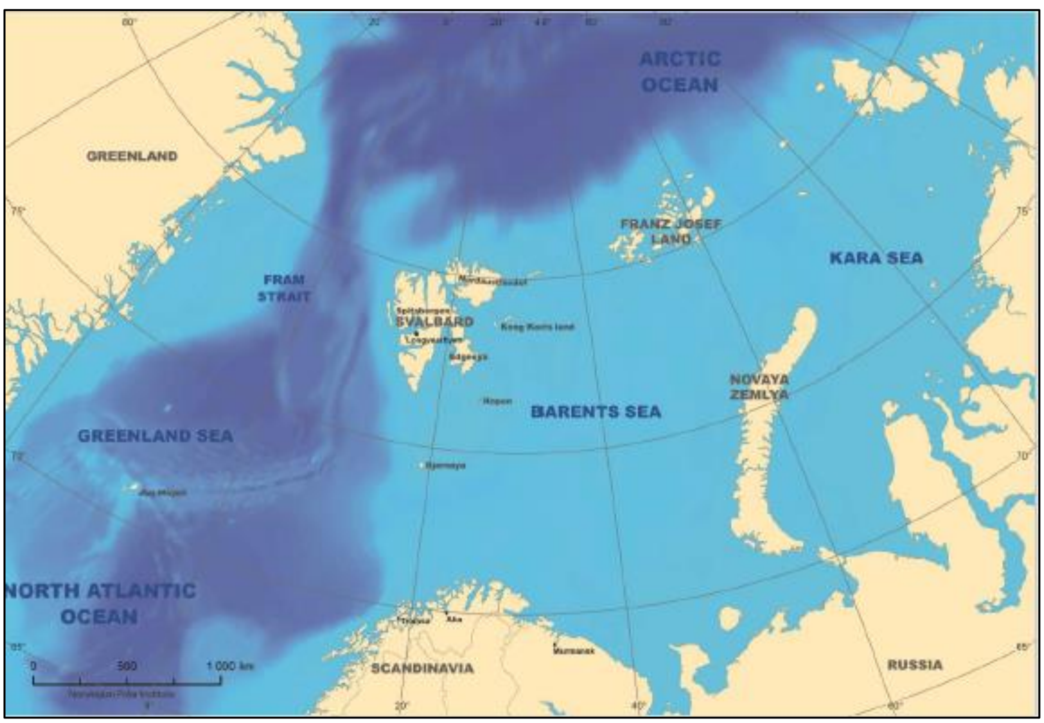


Figure 1.1: Map of Svalbard and the surrounding areas. Modified from Dallmann ed. (2015).

The position of Svalbard on the exposed north-western corner of the Barents shelf makes it a key site for understanding of the geological history of the subsurface of the Barents Sea (Worsley

2008, Dallmann ed. 2015). The Barents Sea is with an area of approximately 1.3 million km^2 one of the largest continental shelves on earth (Doré 1995). The shelf is boarded bordered by Norway and Russia in the south, Novaya Zemlya to the east, Franz Josef Land and Svalbard to the north and the Atlantic Ocean to the east (Doré 1995).

1.3 Previous work

The Triassic succession in Svalbard has since the 19th century been investigated by field work mainly by Swedish, English, Russian, Polish and Norwegian scientists (Vigran et al. 2014). Swedish scientists dominated the research on the Triassic of Svalbard in the end of the 19th century and the beginning of the 20th century. Nathorst (1910) synthesised much of this work. After the Norwegian independence in 1905 Norwegian geologists increased the contribution by mapping and sampling, even though studies and descriptions mainly were done by others. The Norwegian geologists Winsnes and Worsley (1981) produced regional geological maps from Edgeøya. Russian, Polish and British scientists carried out stratigraphic studies. Tatjana Pchelina was a great contributor to the stratigraphic work on the Triassic succession, eg. Pchelina (1980, 1983).

The outcrops in Svalbard serves as excellent analogues to the Barents Sea, and the great petroleum potential in the Barents Sea emphasized the need of more background information and led to renewed onshore activity from the 1970's (Worsley, 2008, Vigran et al. 2014). Seismic expressions of clinoforms, channel bodies and sequence stratigraphy have recently been the focus in several studies, often with additional data from cores and outcrops (Riis et al. 2008, Glørstad-Clark et al. 2010, 2011, Høy and Lundschien, 2011, Anell et al. 2014, Klausen and Mørk, 2014, Lord et al. 2014b, Lundschien et al. 2014, Klausen et al. 2014, Klausen et al. 2015).

Stratigraphy and sedimentology of the eastern islands are studied by Worsley (1973), Smith (1975) Smith et al. (1975), Lock et al. (1978) and Pchelina (1980, 1983). Diagenetic and depositional environment of the Kapp Toscana Group was investigated by Knarud (1980) and summarised by Mørk et al. (1982). They outlined the differences between the Storfjorden and the Wilhelmøya subgroups (later defined by Mørk et al. 1999). Evidence of regional changes in clastic mineralogy associated with the shift between the subgroups was presented by Bergan and Knarud (1993).

A regional study of Triassic sandstone composition and provenance in the Barents Shelf was conducted by Mørk (1999). The work of Mørk (2013) focused on diageneses in the De Geerdalen Formation on central Spitsbergen, and was presented as a part Longyearbyen CO2 lab project. The work of Rød et al. (2014) describes the depositional environment of the De Geerdalen Formation from Edgeøya to central the depositional environment of the De Geerdalen Formation from Edgeøya to central Spitsbergen based on facies analysis of outcrops and cores, and geometric studies through photos and Lidar data.

The study of Lord et al. (2014a) defined the Hopen Member which is time equivalent to the Isfjorden Member. Studies of palaeosol outcrops in the De Geerdalen Formation on Edgeøya and Hopen were conducted by Enga (2015). Core data with focus on palaeosols in the Snadd Formation is described by Stensland (2012) and Enga (2015). Work on palaeosols of the Isfjorden Member includes Knutsen (2013), Husteli (2014) and Olaussen et al. (2015).

2. Regional geology of Svalbard and the Barents Sea

The geology of Svalbard ranges from the Archean to Quaternary in age (Harland et al. 1997) (Fig. 2.1). This chapter provides an overview of the Permian to Mesozoic geology of Svalbard and the Barents Sea, with main focus on the Triassic to Middle Jurassic stratigraphy. Structural geology relevant for this study is briefly described in Chapter 2.4

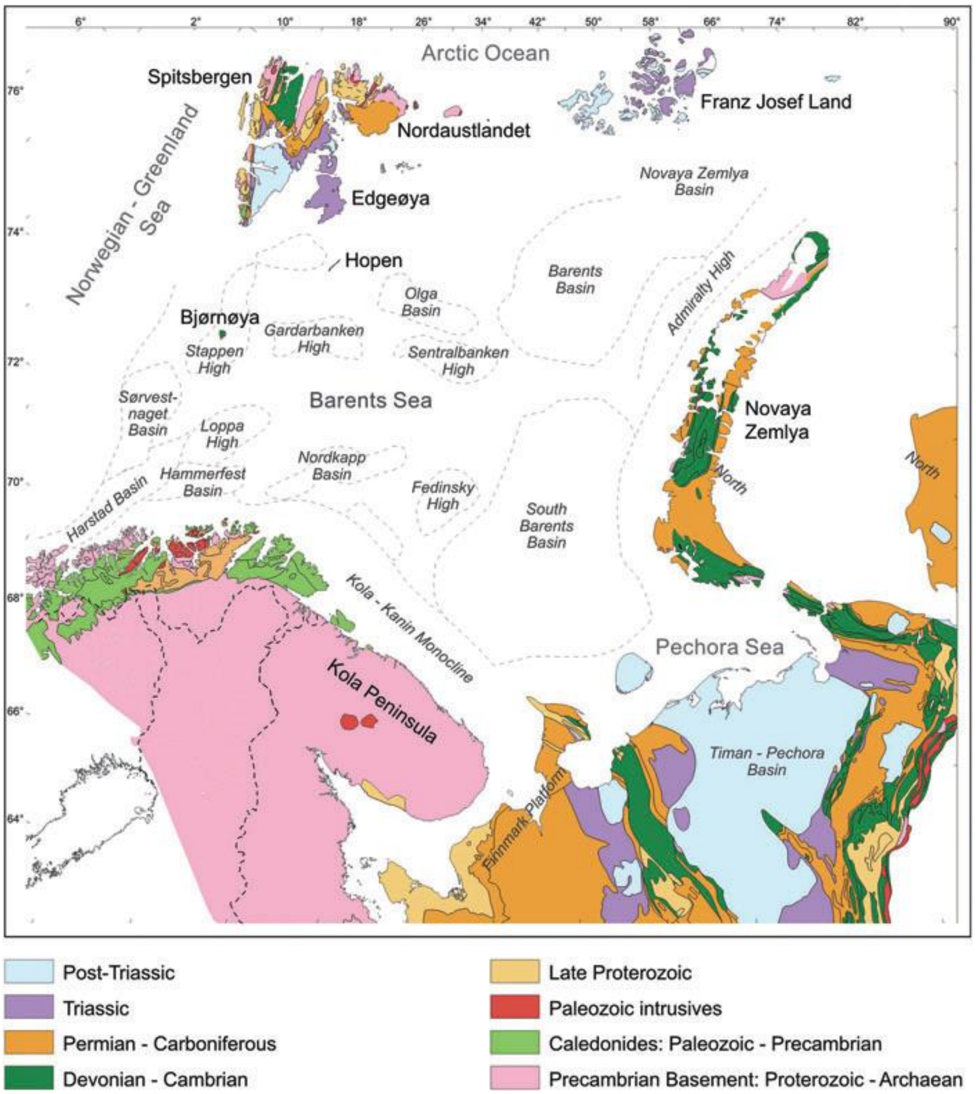


Figure 2.1: Geological map showing the high arctic position of the Svalbard archipelago and the main structural elements in the Barents Sea. Map compiled by Mørk (1999).

Svalbard has gone through significant changes in depositional environments and climatic zones, as it drifted northwards from around 40°S in the Cambrian to present day latitudes at 79˚N. Figure 2.2 shows the drifting of rates Svalbard from equatorial latitudes in the Devonian until present days (Dallmann (ed.) 2015). Relatively rapid drift occurred during the Devonian to Late

Permian, with drift in the Mesozoic being relatively slow. Spitsbergen is a result of 1,) the opening of the North Atlantic Ocean caused dextral movement between Svalbard and East Greenland and oblique compression (transpression) and the development of the West Spitsbergen fold-and thrust belt (Eldholm et al. 1987, Bergh et al. 1997, Leever et al. 2011) and 2) Neogene uplift and glaciation (Smelror et al. 2009).

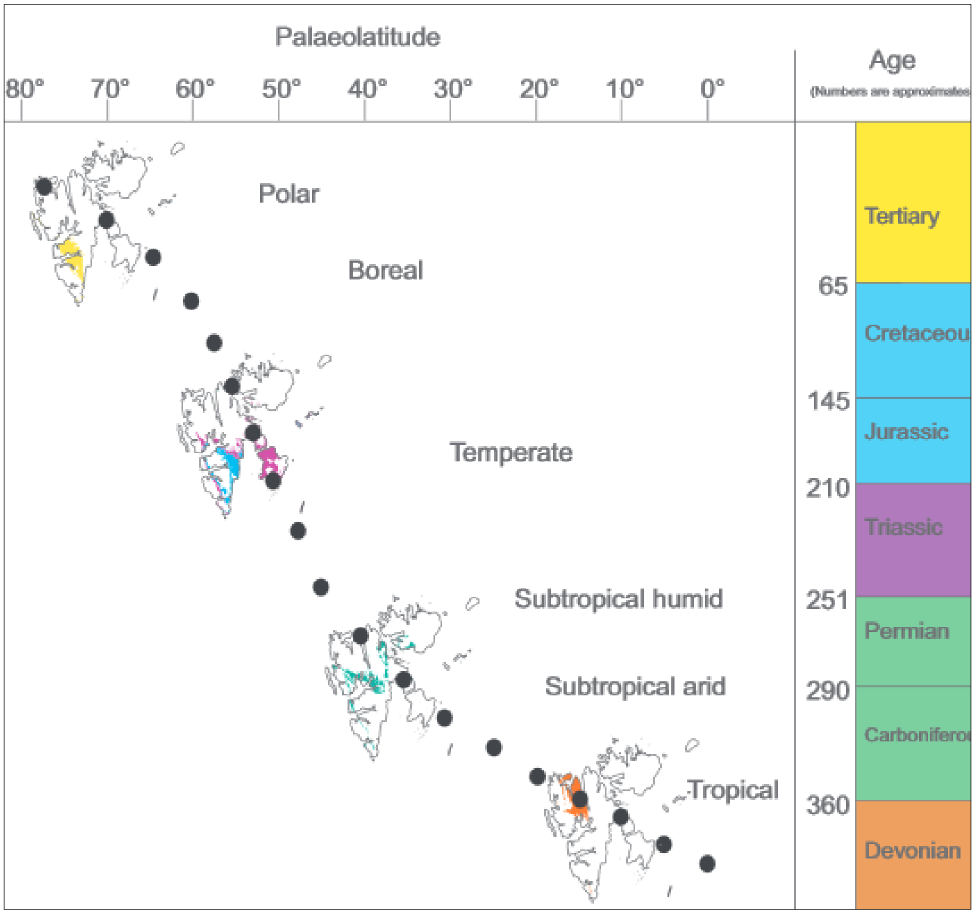


Figure 2.2: The northwards drifting of the archipelago of Svalbard from Devonian to present day latitude (Elvevold et al. 2007).

2.1 Permian

During the Late Permian the final closure of Pangea formed the supercontinent Pangea (Mørk 2015). The transition shift from warm-water carbonate ramp deposits and cool-water bioclastic carbonates and chert deposits occurred near the boundary between Lower and Upper Permian.

In Upper Permian limestone and sandstones are restricted to structural highs and the southern shelf margin (Stemmerik and Worsley 2005, Worsley 2008). The rapid change in depositional environments is probably an effect of the final closure of the seaway to the warm Tethys Ocean in the south and major plate reorganization causing increased subsidence rates in the area

(Worsley 2008). The end of Permian was marked by a major extinction event, with a loss of 90-95 % of marine species and 70 % of terrestrial species (Blomeier and Bond 2015).

2.2 Mesozoic

The formation of the Pangea supercontinent in the Permian dramatically affected the climate, fauna and sedimentological environments in the Boreal Triassic (Mørk 2015). The onset of the Triassic in Svalbard represents a dramatic change from highly cemented spiculitic shales in the latest Permian, to non-siliceous shales in the Early Triassic (Vigran et al. 2014). The mass extinction in the latest Permian provided niches that fauna did not fill before the Middle Triassic (Mørk 2015).

The Triassic sediments of Svalbard and the Barents Sea were deposited in a large embayment located on the northern margin of Pangea. During the Triassic the whole embayment was filled by sediments primarily sourced from the newly formed Ural Mountains in Siberia. An important source area was also located to the west, most likely Laurentia/Greenland (Glørstad-Cark et al. 2010, Mørk 2015). The region was bordered by the Fennoscandian Shield in the south, Laurentia in the west and present day Novaya Zemlya to the east. An open seaway existed to the northwest (Fig. 2.3) (Riis et al. 2008, Mørk 2015). The most important factors controlling the sediment infill patterns were basin subsidence, sea level fluctuations, and erosion and deposition from adjacent land and structural highs of late Palaeozoic age (Riis et al. 2008). In addition climate changes might have been an important factor. A shift from hot equatorial to subtropical climate in the Permian was replaced by dry and temperate conditions in the Triassic (Mørk 2015).

The Triassic succession forms a mega-sequence that can be divided into five second order sequences separated by maximum flooding surfaces (Riis et al. 2008, Glørstad-Clark et al. 2010, 2011, Mørk 2015, Klausen 2015). Glørstad-Clark et al. (2010, 2011) interpreted the sequences from 2D seismic data, public exploration drilling cores in the Norwegian Barents Sea, and seismic lines further to the north. In addition data collected from outcrops in Svalbard were used in the study. The uppermost second order sequence was found to correspond to the middle to uppermost parts of the Snadd Formation and the entire De Geerdalen Formation (Glørstad-Clark et al. 2010, 2011). Sediment transport was driven by progradating clinoforms with main transport direction from south-east towards north-west (Glørstad-Clark et al. 2010, 2011, Høy and Lundschien 2011.)

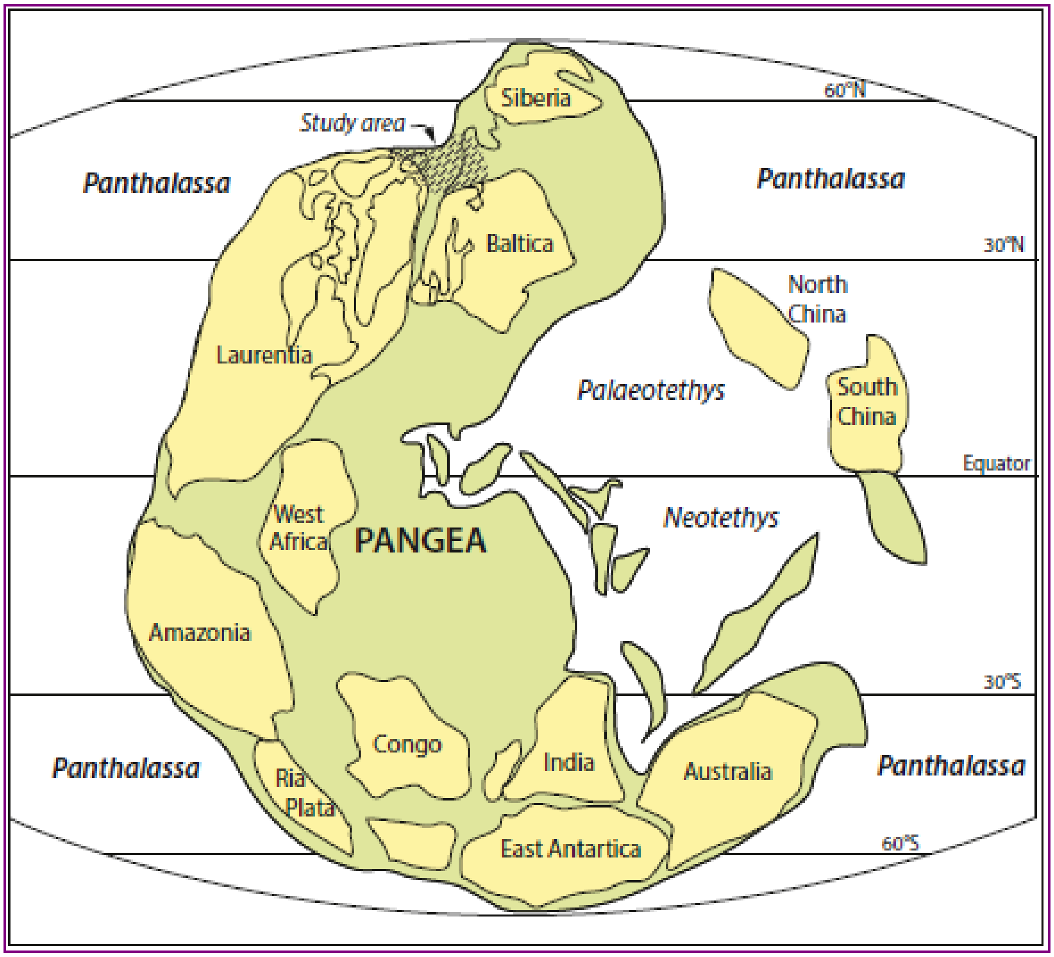


Figure 2.3: Palaeomap of the Triassic. The study area is located on the northern rim of Pangea. Map from Torsvik and Cocks (2005), modified by Vigran et al. (2014).

Seismic studies by Glørstad-Clark et al. (2010 and 2011) suggest that palaeotopography mainly formed during Late Palaeozoic rifting and uplift strongly influenced the location of the depocenters. The Stappen and palaeo-Loppa Highs acted as a barrier towards the west inhibiting major sedimentation from the east until the Ladinian. Minor clinoforms older than the Ladinian occur at the west of the Loppa High, but this is likely linked to local uplift and erosion of the Loppa High in the Early Triassic. The north western parts of the Barents Sea remained distal until the Ladinian when the accommodation space in the south was used up and sequence five prograded all the way to the north western Barents Sea and Svalbard (Glørstad-Clark et al. 2010).

The transgressive-regressive cycles in the region correspond well to sequences in the rest of the Boreal Triassic world (Mørk et al. 1989, Egorov and Mørk 2000), and even world-wide Embry 1997). This might imply that the driving force on the formation of the sequences was tectonics on a global scale, as continents rearranged (Mørk et al. 1989, Embry 2006). The Triassic in general was a tectonic quiet period, but growth faults, horst, grabens and other structures caused by syn-sedimentary processes are common on the eastern island of Edgeøya in the early Carnian

(Edwards 1976, Høy and Lundschien 2011, Osmundsen et al. 2014). The archipelago of Svalbard drifted slowly northwards during the Triassic, giving palaeolatitudes ranging from 40 degrees north at the onset of the Triassic to 60 degrees north at the end of the period (Fig. 2.2). The climate was probably arid to humid and influenced by the northward drifting of the continent (Steel and Worsley 1984).

The latest Triassic to middle Jurassic Wilhelmøya Subgroup in Svalbard and equivalent Realgrunnen Subgroup in the Barents Sea, represent lower sedimentation rates than for the units below, and extensive marine reworking in a coastal to shallow marine regime took place (Vigran et al. 2014). Marine, organic rich shales were dominating the late Jurassic forming important source rocks in the region (Henriksen et al. 2011, Leith et al. 1993). A lowering of the sea level in the Jurassic/Cretaceous transition represent a shift in sediment regime, with general more open shelf conditions and deposition of fine clastics (Bergan and Knarud 1993). The Triassic to Mid Jurassic succession in Svalbard is described in more detail in Chapter 2.3

The Mjølnir meteorite crater on the Bjarmland Platform may have caused a short, but catastrophic impact at the Jurassic-Cretaceous boundary (Smelror et al. 1999, Worsley 2008). The late Jurassic to Early Cretaceous was affected by extensive magmatic activity referred to as the Diabasodden Suite (Mørk et al. 1999). The activity is recognized by dolerite sills and dykes, and is most prominent on Franz Josefs Land and East in Svalbard (Mørk et al. 1999, Maher 2001). The intrusions penetrate host rocks from Precambrian to lower Cretaceous (Mørk et al. 1999), including the De Geerdalen Formation (Senger et al. 2014).

Southwards tilting of the shelf during the late Cretaceous caused erosion in the north and southward transport of sediments. Upper Cretaceous deposits are thus mainly missing in Svalbard (Nøttvedt et al. 1993, Mørk et al. 1999).

2.3 The Triassic to Middle Jurassic Stratigraphy of Svalbard

The nomenclature of the Triassic to Mid Jurassic succession of Svalbard is described in the Lithostratigraphic Lexicon of Svalbard (Mørk et al. 1999) and this scheme is applied herein, with revisions from Krajewski (2008) and Mørk et al. (2013) (Fig. 2.4). The succession constitutes the Lower to Mid-Triassic Sassendalen Group and the Upper Triassic to Mid-Jurassic Kapp Toscana Group (Mørk et al. 1999).

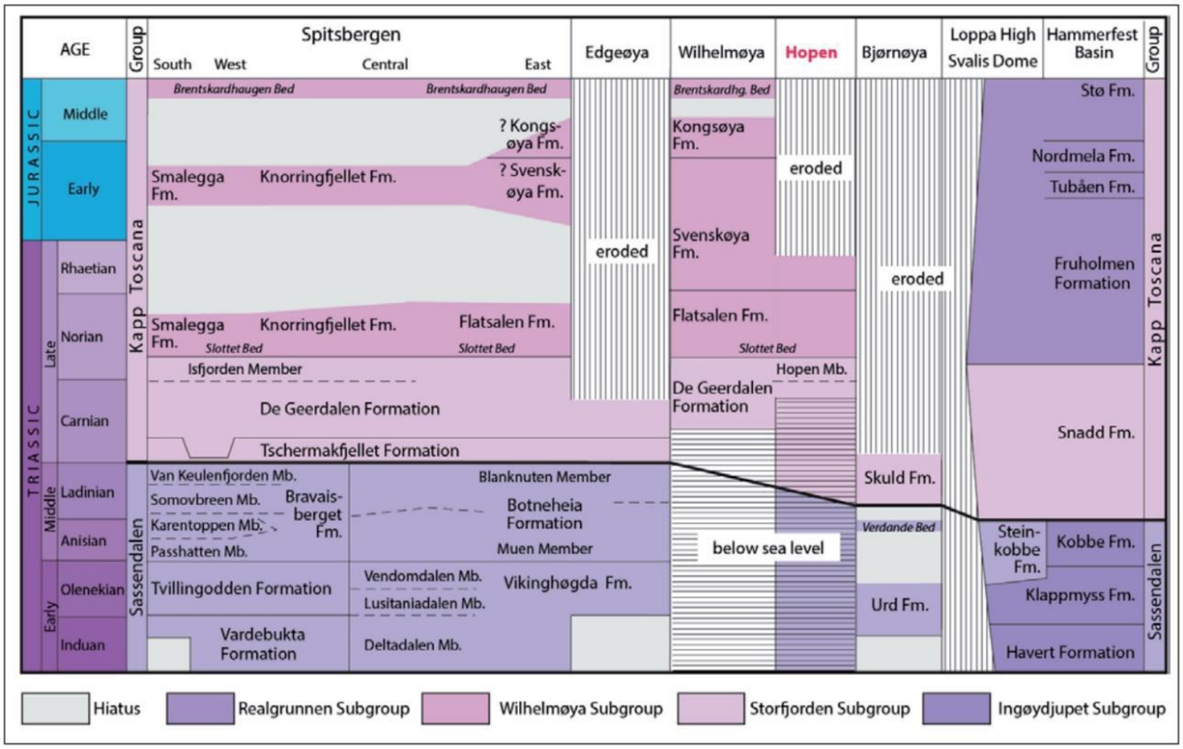


Figure 2.4: Triassic to middle Jurassic Lithostratigraphic of Svalbard and the Barents Sea (Mørk et al. 2013).

The total thickness of the Triassic succession is estimated to range from 250 meters locally to up to 1200 meters on Spitsbergen (Mørk et al. 1982). On Hopen data from the Hopen-2 drilling well made by Fina Group in 1972, estimate a total thickness of 1215 meters (Lord et al. 2014a). A seismic interpretation implies a thickness of around 2000-3000 meters on the platform areas in the western Barents Sea (Gabrielsen et al. 1990), and several kilometres in the eastern basins (Glørstad-Clark et al. 2010, 2011).

2.3.1 The Sassendalen Group

The Early to Middle Triassic Sassendalen Group was first defined by Buchan et al. (1965). The main source was probably from Greenland or small islands between Greenland and Svalbard in the west (Mørk 2015, Vigran et al. 2014). The sediments were deposited during high subsidence and sedimentation rates and consist mainly of non-siliceous fine clastics. The succession is believed to represent repeated coastal progradation from the west with deltaic or barrier sandstones (Mørk et al. 1982).

The group comprises shales, siltstones and sandstones with coastal to deltaic sediments on western Spitsbergen grading into organic rich shelf- mudstones on the eastern Svalbard (Mørk et al. 1982, Mørk et al. 1999). The basal Vardebukta Formation contains barrier bars and lagoons, while the Tvillingodden Formation consists of shallow marine bars and storm beds (Mørk et al. 1999). The Vikinghøgda Formation is the equivalent to the two formations on the central and eastern Spitsbergen (Mørk et al. 1999).

The Middle Triassic organic rich shales were defined by Buchan et al. (1965) as the Botneheia Formation. The formation was degraded to a member by Mørk et al. (1982), but later upgraded to a formation again for central and eastern areas of Svalbard (Mørk et al. 1999). The Botneheia Formation is one of the most organic rich units on Svalbard with TOC values up to 10 % (Mørk and Bjorøy 1984). The Bravaisberget Formation in western Svalbard is the proximal, coastal equivalent to the deep shelf Botneheia Formation (Krajewski et al. 2006, Riis et al. 2008, Lundschien et al. 2014). The Botneheia Formation is partly equivalent with the organic rich mudstones of the Steinkobbe Formation in the Barents Sea, first defined by Mørk and Elvebakk (1999). The formations were formed during periods of high organic production with type II/III kerogen leading to a potential prolific oil prone source rock (Mørk and Bjorøy 1984). The phosphatic rich, condensed section of the Verdande Bed on Bjørnøya is probably the uplifted and eroded equivalent to the Botneheia/Steinkobbe Formations (Mørk et al. 1990).

2.3.2 The Kapp Toscana Group

The Upper Triassic to Middle Jurassic Kapp Toscana Group (Fig. 2.4) was first defined as a formation by Buchan et al. (1965), but later upgraded to a group (Mørk et al. 1982, 1999). The sharp boundary between the Sassendalen Group and the Storfjorden Subgroup can be traced over most of the Barents Shelf and represent a shift from deep marine conditions to pro-deltaic environments. The boundary is of early Ladinian age in the southern Barents Sea, mid-Ladinian age on Bjørnøya and Carnian age in Svalbard (Riis et al. 2008, Lundschien et al. 2014). The Kapp Toscana Group is divided in two sub-groups with significant differences in sedimentological regimes: the Storfjorden Subgroup and the Wilhelmøya Subgroup (Mørk et al. 1999).

2.3.3 The Storfjorden Subgroup

The Carnian to Early Norian Storfjorden Subgroup in Svalbard consists on the Tschermakfjellet Formation and the De Geerdalen Formation (Mørk et al. 1999), and represents a major delta building out from the southeast towards the northwest (Glørstad-Clark 2010, 2011). The main sediment source had probably shifted from west to the Ural Mountains that had newly been formed in the east (Mørk 2015).

The Tschermakfjellet Formation was defined by Buchan et al. (1965) and differs from the underlying Botneheia and Bravaisberget formations by the oxic depositional environment. The formation represents pro-delta shale, deposited in advance of the De Geerdalen Formation, and has a general increase in silt and sand content upwards. Siderite nodules and marine fossils are common. The formation is easily recognized in the field by its distinct purple weathering colour (Mørk et al. 1999).

The De Geerdalen Formation overlies the Tschermakfjellet Formation. The lower base is defined as the first prominent sandstone bed. The formation contains repeated beds of shales coarsening upwards into sandstones, and depositional environments are believed to be shallow marine to deltaic (Mørk et al. 1999). The most distal and marine influenced sediments of the formation are found on central Spitsbergen (Rød et al. 2014, Vigran et al. 2014), while Hopen show more proximal sediments dominated by fluvial processes (Klausen and Mørk 2014, Lord et al. 2014a,b).

Several workers (Rønnevik et al. 1982, Mørk et al. 1989, Nøttvedt et al. 1993, Skjold et al. 1998, van Veen et al. 1993, Mørk 1999, Harstad 2016) have suggested a possible source area in the north or northeast of the De Geerdalen Formation. Later studies of seismic data and investigation of outcrops in Svalbard shows no conclusive evidence of a northerly source area (Riis et al. 2008, Glørstad-Clark et al. 2010, Rød et al. 2014, Klausen 2015).

The Snadd Formation in the Barents Sea is the equivalent to the Tschermakfjellet and the De Geerdalen formations in Svalbard (Worsley et al. 1988). The Skuld Formation represents the lowermost parts of the formations on Bjørnøya, while the upper part is eroded (Mørk et al. 1982).

2.3.1 The Isfjorden Member

The Isfjorden Member terminates the De Geerdalen Formation on Spitsbergen. It was initially defined as a 'suite' (formation) by Pchelina (1972, 1983). The Isfjorden and Hahnfjella "suita" represented the whole upper Triassic succession. Pchelina's (1983) definition of the Isfjorden "suite" represented almost the whole De Geerdalen Formation and thus a greater part of the succession than present definition of the Isfjorden Member by Mørk et al. (1999). Pchelina (1983) stated the presence of the Isfjorden "suite" on the whole archipelago of Svalbard, except Southern Spitsbergen. The presence of her defined formation included Central Spitsbergen, Hopen, Wilhelmøya, Barentsøya and Edgeøya (Pchelina 1983).

After a revision in Mørk et al. (1999) was the upper Triassic succession divided in the Tschermakfjellet and the De Geerdalen formations, and the Isfjorden "suite" was downgraded to member status belonging to the upper part of the De Geerdalen Formation. According to Mørk et al. (1999) is the thickness of the member in the range of 55 to 135 meters. The type section of the Isfjorden Member is at Storfjellet, West of Agardhdalen in the logged section of Knarud (1980).

The characteristic red and green mudstones of the Isfjorden Member displays a distinct colour not found in the underlying parts of the De Geerdalen Formation. Alternating shales with siltstone and sandstone beds is typical. Carbonate beds, phosphate nodules, plant fragments and lenses of gravel or conglomerate are common. Coquina beds in the section are common, and the base is defined as a siltstone coquina bed occurring above a thick cross bedded sandstone. A siderite bed occurring some few meters above the base of the member is common. The depositional environments are interpreted to be a shallow shelf with local lagoons (Mørk et al. 1999).

The recently defined Hopen Member is time equivalent to the present definition of the Isfjorden Member (Lord et al. 2014a). The suggested depositional setting for the Hopen Member is a marine environment with variable energy and biota. The uppermost part of the member contains offshore storm deposits and mud, while the lowermost part is characterized by low energy deposits. The lower boundary of the Hopen Member is traceable all around the island. Lord et al. (2014a) suggest a potential sequence stratigraphic boundary at the base. The Hopen Member is so far only observed on the island of Hopen on South-eastern Svalbard (Lord et al. 2014a).

2.3.4 The Wilhelmøya Subgroup

The Slottet Bed marks the onset of the Wilhelmøya Subgroup (Mørk et al. 1999). The bed represents a transgressive lag on a major flooding surface during the Norian (Riis et al. 2008 , Mørk et al. 1999). The unit is in the range of 1.5 – 11 meters, and the lower boundary is defined as the first carbonate sandstone bed with phosphatic nodules. The main lithologies in the bed are calcareous silt – and sandstones, and polymict conglomerate which locally contain phosphate (Mørk et al. 1999).

The Knorringfjellet Formation represents the entire Wilhelmøya Subgroup on western Spitsbergen. The formation contains shales, sandstones and carbonates, and represents a condensed section with long breaks in deposition (Mørk et al. 1982, 1999).

The Flatsalen Formation (Smith et al. 1975) is the lowermost formation in the Wilhelmøya Subgroup on Hopen, Wilhelmøya and east Spitsbergen (Mørk et al. 1999). The main lithology of the formation is dark grey silty shales with beds of siltstones and fine-grained sandstones. The formation shows a general upwards coarsening trend (Mørk et al. 1999).

The Svenskøya Formation (Smith et al. 1976) overlies the Flatsalen Formation and is present on Hopen, Wilhelmøya, Kong Karls Land and Olav V Land on eastern Spitsbergen. The formation is dominated by sandstones. Depositional environments are interpreted to be tidal flat, tidal channel and coastal plain deposits that grade up to wave to tidal dominated shoreline or protected bay deposits (Mørk et al. 1999).

The Kongsøya Formation (Smith et al. 1976) represents the uppermost part of Wilhelmøya Subgroup on Wilhelmøya and east Spitsbergen (Mørk et al. 1999). Main lithologies are sandstone, mudstone and conglomerate. Depositional environments are interpreted to be in a shallow marine, inner shelf setting.

Whether the Brentskardhaugen Bed belongs to Wilhelmøya Subgroup or the overlying Janusfjellet Subgroup is discussed (Mørk et al. 1999). Herein the bed is defined to terminate the Wilhelmøya Subgroup, following (Mørk et al. 1999). According to this interpretation the bed is a condensed succession. The bed is a prominent marker bed recognized by pebbly calcareous sandstones. Polymict pebbles and phosphate nodules with fossil inclusions are typical (Mørk et al. 1999).

2.4 Structural Geology

During the Cenozoic the opening of the North-Atlantic Ocean migrated northwards. As a consequence a 100-200 km wide Fold Belt developed along western Svalbard and Barents Sea. The crustal shortening of the fold belt is estimated to be around 30 km (Worsley 2008). The fold belt was formed in Paleogene when Greenland was drifting north eastwards against the Barents Shelf and at the time nearby Lomonosov Ridge (Dallmann 2015). The Triassic succession in the west is highly deformed from this episode (Mørk et al. 1982), making the reconstruction of the original Triassic succession in the area not straight forward.

Towards the east relatively un-deformed, flat lying, Mesozoic strata are present. These are however dissected by a series of structural lineaments. The most dominant being the Billefjorden

Fault Zone (Harland et al. 1974, Braathen et al. 2011). Towards the east, the Lomfjorden and Storfjorden Fault zones are present (Eiken 1985, Dallmann 2015). The extension of the Lomfjorden fault into Storfjorden is also thought to be undergoing current tectonic activity (Dallmann 2015).

The Neogene is characterized by repeated episodes of subsidence due to ice loading and isostatic uplift when the ice melted. Uplift and erosion of sediments in the range of two to three kilometres sourced the Norrland Group located in the western shelf margins (Worsley 2008, Mangerud et al. 1996).

3. Palaeosols

Studies of palaeosols constitute an important component in this study. The master thesis benefits from earlier work on palaeosols in the De Geerdalen Formation on Hopen and Edgeøya (Enga 2015) and the Snadd Formation in the Barents Sea (Stensland 2012, Enga 2015). Herein palaeosols on Edgeøya, Wilhelmøya, Barentsøya and Spitsbergen are investigated.

This chapter presents brief theory about palaeosols.

3.1 Definition of palaeosols

A palaeosol or a fossil soil is a remnant of pedogenetic (soil forming) processes that occurred on a landscape of the past (Kraus 1999, Retallack 2001). The definition of "soil" is subjected to more confusion than the definition of "fossil". Like other types of fossils, a fossil soil is the remains of a soil formed in an ancient landscape, and the soil forming processes are no longer active (Retallack 2001). Geologist also use the broader term regolith which is defined as "loose unconsolidated rock and dust that sits atop a layer of bedrock. On Earth, regolith also includes soil, which is a biologically active medium and a key component in plant growth (Rafferty 2016).

The diverse definitions of "soil" are largely influenced by different scientific fields. From farming perspective a soil is a fertile and loose ground favourable for plant growth. An engineer might define soil as loose ground that can be excavated without blasting. Biologists might require plant growth to define the ground as soil (Retallack 2001). For geologists a wider use of the term "soil" is common. Retallack (2001 p.7) define soils as "material forming the surface of a planet or similar body and altered in place from its parent material by physical, chemical or biological processes". The same definition is used in this study. With this wide definition soils include almost all landmass except rivers and lakes, and areas that recently has undergone erosion.

3.2 Factors influencing soil formation

The forming of soil is a complicated process between sediments or bedrock and the interaction with the atmosphere (Retallack 2001).

The maturity of soils is related to the degree of alternation of the bedrock or sediment. In palaeopedology a key tool to investigate degree of soil forming is study of soil horizons to investigate the layers of alternation. Solum is used as term for the most altered part of the soil, while weathered material between the solum and the underlying bedrock or sediment is termed saprolite. The solum and saprolite are used as relative description in one soil horizon. Material with similar properties and features as a saprolite in one soil horizon can thus be the solum in another soil horizon (Retallack 2001).

3.2.1 Climate

Because the development of a palaeosols is closely linked to interaction between the Earth's surface and the atmosphere they can be used in palaeoclimate reconstructions (Sheldon and Neil 2009). In fact, climate is one of the most important factors in the pedogenesis, weathering and soil forming processes (Cecil and Dulong 2003). Climate is usually defined as the average weather during a period of 30 years (Retallack 2001). Climate can be classified in several ways, depending on which factors are chosen in the classification system. The world classification of climate by Kottek et al. (2006) which is based on temperatures and seasonality is a common way to classify climate zones on Earth.

Water is the most important medium in soil forming processes. Properties of soils are thus largely influenced by moisture regime (Retallack 2001). Important factors influencing the moisture regime are rainfall, groundwater level, soil-drainage and evaporation (Retallack 2001). Cecil et al. (2003) classified climate based on number of wet months during a year (Fig. 3.1A). Wet months were defined as months where rainfall exceeded evaporation. Cecil and Dulong (2003) linked this to which type of soils are most likely to form in different climate conditions (Fig. 3.1B). The model is only valid for warm climate, which they defined as frost free most of the year except in high mountains areas (Cecil et al. 2003). As seen in the figure are entisols and Inceptisols (poorly developed soils) likely to form in a wide range of moisture regimes, while the other soil types are more dependent on number of wet months during the year (Cecil and Dulong 2003). Soil types in the De Geerdalen Formation are described in Chapter 8.2

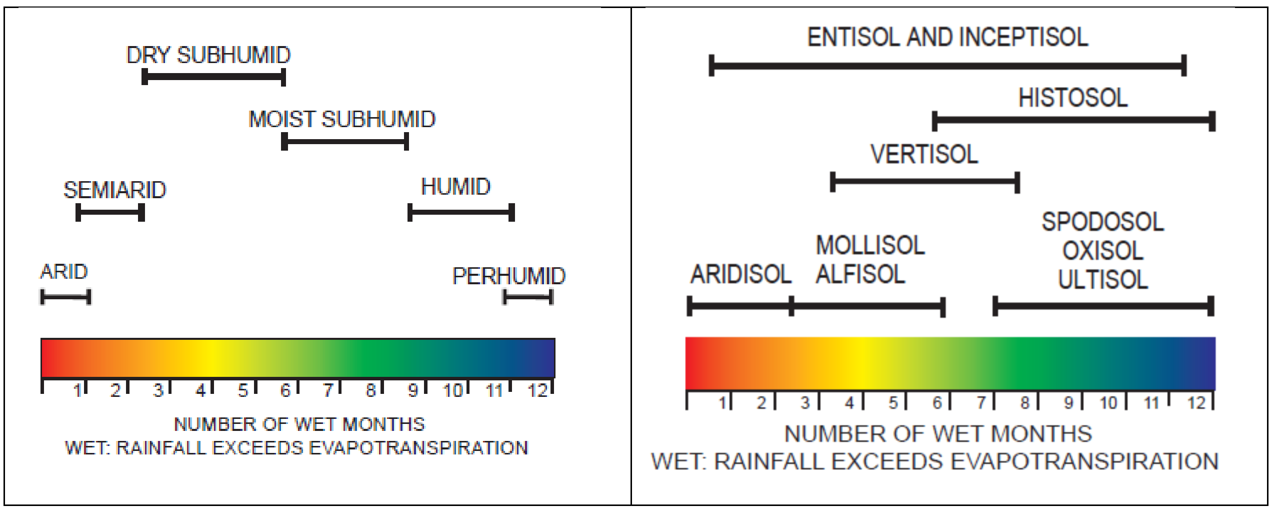


Figure 3.1: A) Classification of climate based on number of wet months (Cecil et al. 2003). B) Soil types classified according to the USDA soil taxonomy classification system and their relation to moisture regime caused by seasonal rainfall (Cecil and Dulong 2003).

3.2.2 Time

Time is also a key factor in soil forming processes. Figure 3.2 shows roughly how long time is required to create different pedogenetic features. As a general trend vertic features form already after tens of years. Mottling requires hundreds of years, and horizonation occurs after thousands of years (Wright 1992). As seen in the figure there is a lot of uncertainty linked to how long time is needed to develop pedogenetic characteristics (Wright 1992). An additional factor for the rate of soil forming processes is the temperature. If the temperature increase with 10°C the rate of chemical processes increases 2-3 times (Retallack 2001). This is mirrored in the fact that soils formed around equator tend to be better developed and more mature than soils of cold regions (Retallack 2001). Other factors that influence the rate of soil forming include properties of parent material, vegetation and sedimentation rate (Retallack 2001).

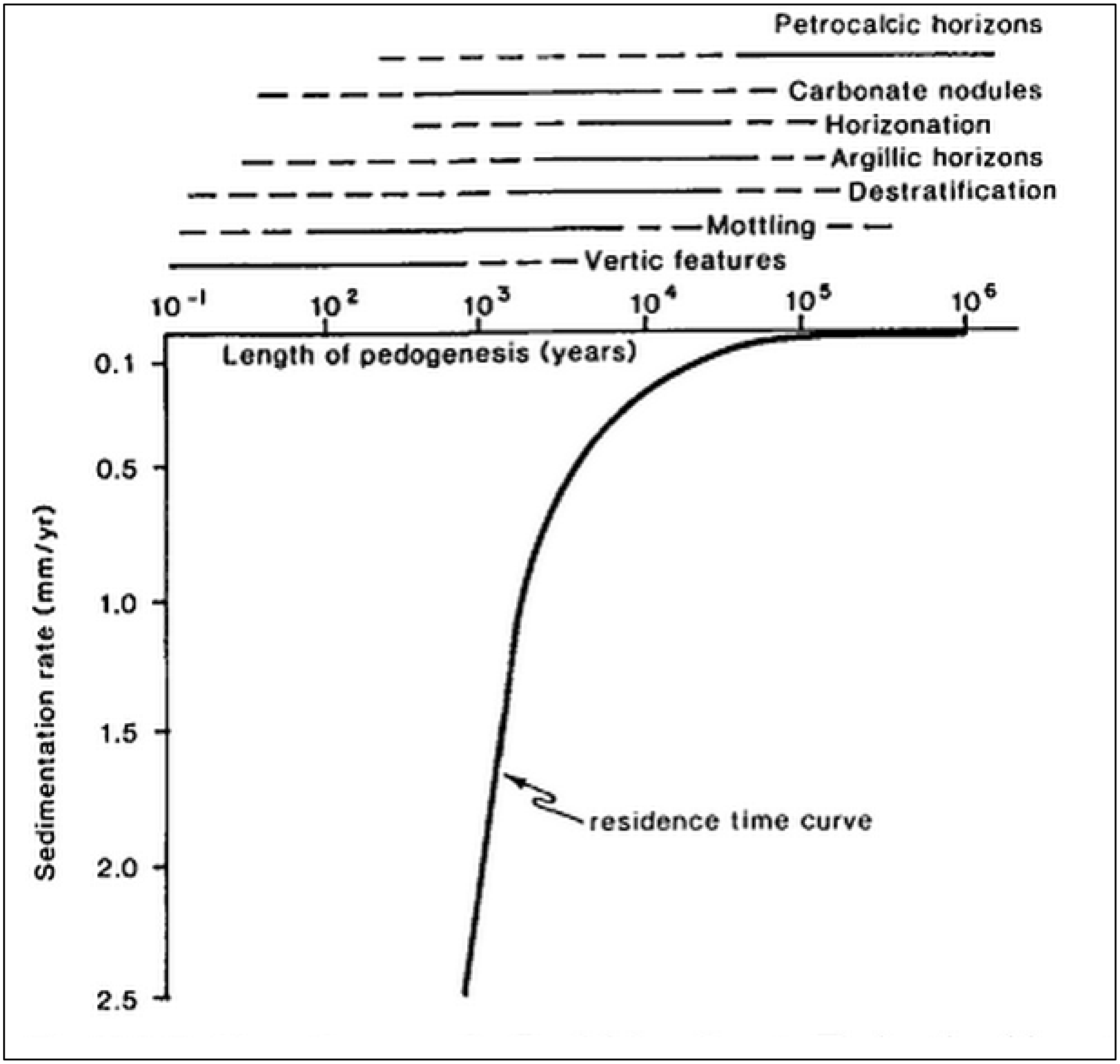


Figure 3.2: Time required developing different pedogenetic features (Wright 1992).

3.3 Palaeosols in sedimentary rocks

3.3.1 Sedimentation rate

Soils and palaeosols in sedimentary rocks are influenced by the interplay between sedimentation, erosion and non-deposition (Kraus 1999). Sedimentation rate affects the thickness and maturity of soils. High sedimentation rates might give thick and immature soils, while low sedimentation rates favour thinner and more mature soils (Wright 1992). Composite soils are stacked, overlapping soils. They develop if a soil forms on top of a buried soil, and the soil forming processes exceeds the sedimentation rate (Wright 1992). Thick composite soils points towards steady sedimentation rates and insignificant erosion (Kraus 1999). Compound soils have a thin sediment layer unaffected by pedogenetic processes between the stacked soils (Wright 1992).

They are associated with negligible erosion rates and rapid and unsteady sedimentation (Kraus 1999).

Floodplain palaeosols tend to be weakly developed due to the dynamic nature and unsteady sedimentation. If the erosion rate is greater than the soil forming processes, no palaeosols will be preserved (Wright 1992, Kraus 1999).

3.3.2 Palaeolandscape reconstruction on a local scale

Lateral changes of palaeosol-properties on a local scale are strongly dependent on topography and grain size (Kraus 1999). The paleocatena model of Kraus and Aslan (1999) is based on fluvial environments and shows how properties of palaeosols changes in a local scale from levees close to the river channel and down the slope (Fig. 3.3). A catena is a group of palaeosols which are genetic related to each other in terms of parent material and climate, but displays somehow different soil properties due to different drainage patterns and topographic elevation (Wright 1999).

In general soils formed on crevasse splays and levees are well drained because they are elevated compared to the surroundings, and consist of relatively coarse grained material. The soil, and in particular the uppermost part, are usually formed in oxidized conditions, leading to yellowish to brown colour. In Figure 3.3 this is described as Bw (coloured or structured B-horizon). Soil further down in the profile is closer to the groundwater table and thus potentially more prone to reduced conditions leading to grey colour. Groundwater saturated soils are called gleyed soils and is marked as Ag, Bg and Cg in Figure 3.3 (Kraus and Aslan 1999).

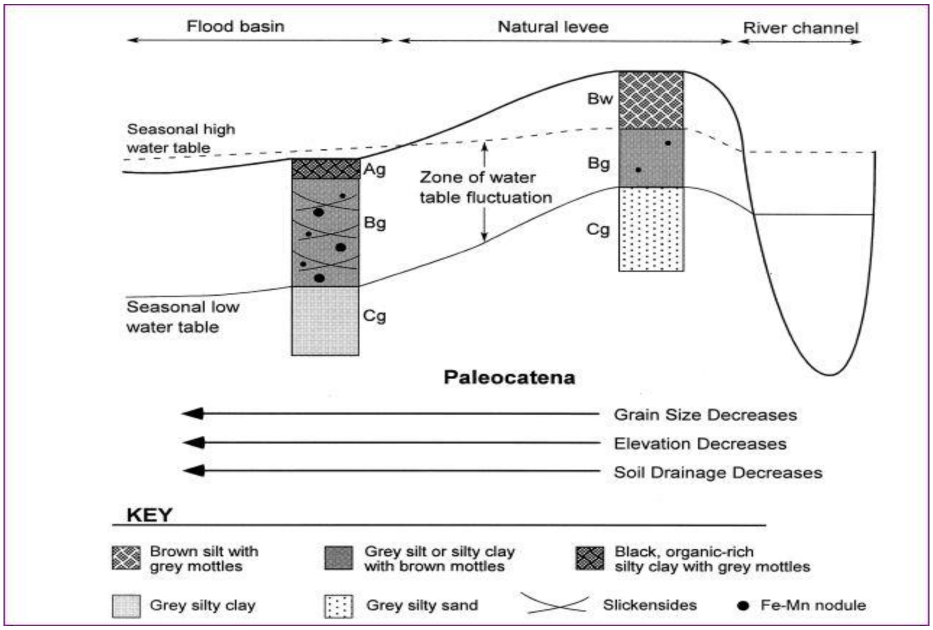


Figure 3.3: The paleocatena model. Soils formed close to channels tend to be coarser grained and formed in oxidized conditions, leading to yellow to brown colour. Soils decrease in grain size away from the active channel and more of the profile is gleyed (Kraus and Aslan 1999).

Soils tend to be less drained away from the channel because the soils consist of finer material closer to the groundwater table due to the topographic position. This favour reduced conditions and gleyed soils in both A, B and C horizons. Because of the reduced conditions organic matter can be accumulated and preserved in the A-horizon. Gleyed B and C horizons are typical for soils formed distal to channels (Kraus and Aslan 1999). On the other hand development and maturation of palaeosols tends to increase with distance from the channel, because of decrease in sediment rates away from the channel (Bown and Kraus 1987).

4. Field work localities

Nine localities all situated in the archipelago of Svalbard are presented in this study. The main field area is in the northern Storfjorden Area and Wilhelmøya, but Deltaneset on central Spitsbergen is also included (Figure 4.1).

The outcrops of Deltaneset 15 km north of Longyearbyen display excellent exposures of the Isfjorden Member and the overlying Slottet Bed of the Knorringfjellet Formation. The area has been visited by several sedimentologists (Pchelina 1983, Husteli 2014, Knutsen 2013, Olaussen et al. 2015). A 36 meter long section was logged at Deltaneset in an unnamed gully close to Wimanfjellet. The whole section belongs to the Isfjorden Member. Given a high abundance of scree cover in the Isfjorden Member is many places elsewhere in Svalbard the outcrop at this locality is exposed excellent.

Agardhdalen on eastern Spitsbergen is surrounded by mountains consisting of Upper Triassic to Middle Jurassic rocks. The upper 120 meters of the De Geerdalen Formation as well as the whole Wilhelmøya Subgroup and lower part of the Agardhfjellet Formation was measured by Knarud (1980). The log is also presented in Mørk et al. (1982) and Vigran et al. (2014). In this study three mountains of Agardhdalen were visited, and includes Klement`evfjellet, Friedrichfjellet and Šmidtberget. The entire exposed succession of the De Geerdalen Formation was logged on all three mountains.

Edgeøya east of Spitsbergen is the third largest island in Svalbard. The island mainly consists of Triassic rocks of the Sassendalen and Kapp Toscana Groups (Fig. 4.1). Due to erosion only the lower parts of the De Geerdalen is exposed, and the presence of the Isfjorden Member has not been documented on the island (Vigran et al. 2014). Blanknuten and Muen were visited during the field work, but only the log from Blanknuten is present here.

Barentsøya boarders the northern Storfjorden and is the fourth largest island in Svalbard. The upper part of the De Geerdalen Formation is also eroded on Barentsøya. Dolerites of the Diabasodden Suite is penetrating Triassic rocks several places (Mørk et al. 1999), making prominent cliffs in the mountains sides.

Wilhelmøya is the northernmost locality visited in this study. The Upper Triassic to Middle Jurassic is exposed on the island. In this study the mountain above Tumlingodden on the east side were studied, and two logs were drawn. Dolerite belonging to the Diabasodden Suite is penetrating the outcrops of the Tumlingodden section. The Wilhelmøya 15-2 log presented in this study starts out from the diabas cliff and ends in the Slottet Bed (Appendix C). The Wilhelmøya 15-1 was drawn by Johansen (2016) (Appendix D). An excursion with all participants in the field work gave the author of this thesis the opportunity to observe the whole section from base of the De Geerdalen Formation to the Slottet Bed.

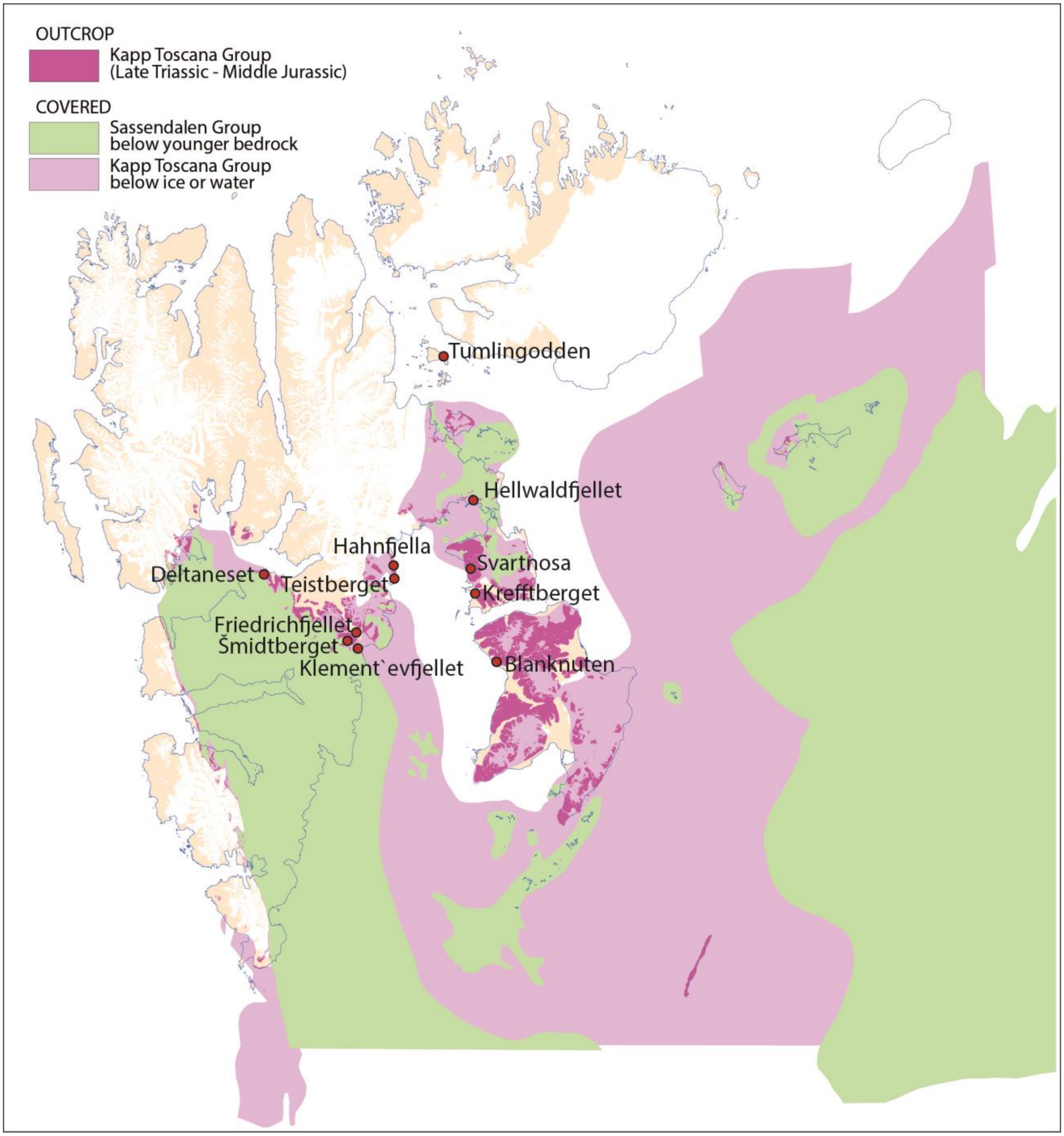


Figure 4.1: Overview over the localities visited the summer of 2015. Map from Dallmann et al. (in prep.).

5. Methods

5.1 Field work

5.1.1 Logistics

Field work took place during a period of five weeks from 2nd of August to 6th of September 2015. The expeditions were organized by SINTEF Petroleum Research and The Norwegian Petroleum Directorate. The University Centre of Svalbard (UNIS) provided field equipment.

Transport in 2014 was with MS *Stålbas* and MS *Kvitebjørn*. MS *Sigma* provided transport to all localities in 2015 except Deltaneset. The boat made it possible for the field campaign to be flexible and visit the most suitable localities every day based on localities of interest, weather, ice conditions and polar bear presence. The cabin, belonging to Longyearbyen Jeger- og Fiskerforenig, in Agardhbukta was used as base for field work for twelve days the summer 2015.

Transport to Deltaneset by Polarcirkel boats (rigid buoyancy boats) was provided by UNIS. The cabin at Deltaneset owned by UNIS functioned as base for the three day long field work.

5.1.2 Collecting of data

Contemporary field work was performed, in order to collect data from every location visited. The field work was organized in teams of two to three students. Nina Bakke, Cathinka Schaanning Forsberg, Bård Heggem, Simen Jenvin Støen, Sondre Krogh Johansen and PhD candidate Gareth Lord from NTNU and geologists from NPD participated in the fieldwork.

Visual field observations were done in accordance to Tucker (2011). Observations included sedimentary structures, organic content, bioturbation, bed thickness and presence of carbonates. 10% HCl was used to determine the present of carbonates. The logged sections were measured with meter stick, and grain size estimated by a standard grain size sheet.

Palaeosols were paid special attention. Most of the Palaeosols were buried. In order to detect the presence of palaeosols observation of adjacent outcrops and scree cover colour were performed. Palaeosols were often found to overlie channel deposits, and close to yellow, brown, red or green scree cover. A geologic hammer and a hoe were used to dig a trench to expose the palaeosols. Observation of evidences of roots, nature of the soil horizons and soil structure were paid special attention. The continuous permafrost of Svalbard defined the lower limit for how deep it was possible to dig. This often limited observations of the lower reaches of the palaeosols.

Sedimentary logs in the scale of 1:100 were drawn in the field in order to record and systemize the observations. In all approximately 1400 meters of logs are presented in this study. Outcrops and mountains were documented by standard digital cameras. Global position system (GPS) was used to get the standard UTM-coordinates at the start and end of each log. Adobe Illustrator drawing program was used to digitalize the sedimentary logs. All information collected from field observations were systemized by facies and facies association analysis (Chapter 6 and 7). In addition is field observations of palaeosols described in Chapter 8.1 and 8.2.

5.2 Laboratory analysis

XRD- and optical microscopy analysis were done on selected samples. See table 1 for information about each sample.

Table 1: Localities, stratigraphic level and brief description of selected samples chosen for laboratory analysis.

Locality/sample number	Palaeo-sol No.	Meters on log	Description	XRD	Thin section
Wilhelmøya					
Tum 15.2.10.C	5	66	Immature palaeosol on top of Distributary channel (FA 2.3)	x	
Tum 15.2.17.C	8	89	Noncalcareous red and green mudstone	x	
Hahnfjella					
Hahn 15.2.32.C	16	210	Immature palaeosol on Interdistributary area (FA 3.2)	x	
Klement`ev-fjellet					
Klem 15.1.24.C	27	195	Immature soil on flood plain on the lower delta plain (FA 3.1).	x	
Blanknuten					
Blank 12.B	37	28	Horizonated soil on top of distributary channel (FA 2.3).	x	
Deltaneset					
Sample 1	49	6.5	K-horizon Green. No mottles, no organic content. Peds.		x
Sample 8	57	17	K-horizon Purple with green mottles and green nodules.		x
Sample 11	60	26	Green bed of calcareous nodules.		x

5.2.1 XRD-analysis

Five samples were picked out for X-ray diffraction (XRD) analysis in order to quantify the mineralogy of the fine grained pedogenetic sediments. The samples are from four different locations. The sediments shows quite different visual properties, such as colour, grain size and adjacent sediments. This makes the sample set diverse in geographic location, stratigraphic level and possible type of palaeosol. The aim of the XRD-analysis is to confirm/unconfirmed that the sediments might be pedogenetic, and add more information about possible depositional environments.

The analysis was performed by Laurentius Tijhuis at the Chemical/Mineralogical Laboratory at Norwegian University of Science and Technology (NTNU). The samples were crushed to powder and run in the XRD instrument Bruker D8 advance. Identification of minerals was conducted in Diffrac.pluss.EVA. The quantification of the interpreted minerals was made in Topas software, which is based on the Rietveld method (Izumi 1992, Perkins 2011).

Interpretation of the XRD-analysis is following the suggestion of Laurentius Tijhuis to use mineral groups instead of specific minerals: Albite is plagioclase, muscovite is mica, microcline is alkali feldspar and diopside is pyroxene. Further, the mineral concentrations are rounded to the nearest percent with one decimal.

5.2.2 Optical microscopy

In total three petrographic thin sections were made from samples collected at Deltaneset in order to investigate micro-scale features in the sediments. The samples were based on field observations interpreted to have pedogenetic origin. A standard petrographic microscope with plan-polarized and cross-polarized light was used for analysis of mineralogy and autogenetic textures.

5.3 Sources of error

Visual observations and interpretation of field data is subjected to human errors. Challenges in the field like scree cover and steep terrain have also caused some limitation of the amount and quality of the collected field data. Palaeosols are commonly scree covered, and the thickness and number of them might be higher than present in this study, especially for the red and green mudstones of the Isfjorden Member. Logging and good field observations of palaeosols are time consuming, especially since most palaeosol had to be dug out for good exposure. Limit of time sometimes made it necessary to do more briefly field observations than desired.

Retallack (1988) mention three challenges regarding field observations of palaeosols: erosion of parts of the profile, overlap of soil horizons of various palaeosols and development of palaeosols followed by weathering.

The powder diffraction used in XRD-analysis is not an exact process, and several things can cause small errors in the results. For more information about errors in XRD-analysis, see Perkins (2011). However, XRD-analysis is usually adequate enough for most mineral analysis (Perkins 2011). The samples picked out for laboratory analysis is not representing any statistic valuable selection, and must be considered as case studies.

6. Facies in the De Geerdalen Formation on eastern Svalbard

This chapter has been written as collaboration between master students Turid Haugen, Simen Jenvin Støen and Sondre Krogh Johansen, who also worked together in the field. 15 facies have been described, based on field observations. Interpretations and discussion of their origin and in which depositional environments they are most likely to be found in are also included. A summary of these facies are given in table 2. The master thesis of Johansen (2016) and Støen (2016) focuses primarily on sandstones and delta front deposits. This master thesis primarily focuses on delta top sediments, and especially palaeosol. Facies O (Palaeosols) is therefore further outlined in Chapters 3 and 8 in this thesis.

The facies scheme described here is largely based on the pioneering work of Knarud (1980). His research was further extended and greatly modified by Rød et al. (2014), who described fifteen facies from central Spitsbergen and from Edgeøya. This study is complementing these works and the study area includes a number of locations not previously visited by these workers.

Facies analysis

The concept of facies was originally introduced into the geological discipline by Nicolas Steno in 1669, but its modern usage is usually attributed to Gressly (1838). Since then the term has developed and numerous interpretations exists in the geologic literature, which is summarized in Middleton (1973) and Walker (2006). Facies can be further subdivided into bio-facies, litho-facies, and micro-facies depending on the basis and focus of observations (Reading and Levell 1996, Boggs 2011, Walker 2006). Assemblages of trace fossils are commonly grouped together into ichnofacies (Pemberton et al. 1992), similar to how physical sedimentary features of sandstones are grouped into facies.

Facies analysis provides a useful foundation on which to correlate rock units, laterally as well as vertically. Spatial and temporal relationships of sedimentary rocks are most evident when seen in outcrop. Facies should therefore be described in such a way, that their corresponding rock counterparts are most easily recognized in the field (Walker 2006).

Table 2: Facies in the De Geerdalen Formation on eastern Svalbard (modified from Rød et al. 2014) cl - clay, si - silt, vf - very fine, f - fine, med - medium

#	Facies	Grain size	Description
A	Large-scale cross-stratified sandstone	f - med	Trough to tabular cross stratification, erosive base fining-upwards, set thickness are between 20 - 80 cm, while unit thickness are in the range of. 0.2 - 4 m. Rip-up clasts and plant fragments are observed and typical trace fossils include *Skolithos* and *Diplocraterion*.
B	Small-scale cross-stratified sandstone	vf - f	Asymmetric ripples (2D and 3D ripples), sets thicknesses 2-10 cm, unit thicknesses up to 1.5 m. Mud drapes and sparse bioturbation. Vague to pervasive cementation
C	Climbing ripple cross stratified sandstone	vf - f	Small-scale asymmetric climbing ripple laminated. Sharp lower contacts,while upper contacts are gradual. Unit thickness ~ 0.5 m
D	Wave rippled sandstone	vf - f	Symmetrical wave ripples, planar parallell stratification and mud flakes.Bed thicknesses 10 - 30 cm, unit thickness up to 3 m. Often observed towards top of coarsening upwards sequences. Moderately to intensely bioturbated, *Rhizocorallium* and *Skolithos*
E	Low angle cross stratified sandstone	si - f	Gently inclined stratification with wedge-shaped set boundaries. Set thickness 5 - 15 cm and unit thicknesses up to 1.5 m. Commonly bioturbated and contains plant fragments
F	Horizontally bedded sandstone	vf - f	Planar parallel stratification (PPS) and lamination (PPL). Unit thickness 30 cm to 2 m. Laminae and bed thicknesses varies within units. Parting lineation. *Skolithos*, *Diplocraterion* and *Rhizocorallium* in upper parts of units.
G	Massive, structureless sandstone	vf - med	Fractured, apparently structureless.Units are between 0.1 m to 5 m thick. Sharp erosive base, commonly enclosed by mudrocks and heterolithics. Carbonate cemented. Often bioturbated and contains plant fragments and mud flakes.
H	Hummocky and swaley cross stratified sandstone	si - f	Hummocky and swaley cross stratified. 5 to 20 cm thick sets with unit thickness up to 1 m. Moderately to intensely bioturbated, *Skolithos* and *Diplocraterion*
I	Soft sediment deformed sandstone lenses	vf - f	Irregular base, laterally restricted sandstone bodies. Unit width 2 - 4 m and unit height 0.3 - 1.5 m. Enclosed in mudrocks
J	Carbonate rich sandstone	vf - f	Carbonate cemented. Commonly very hard. Beds are often laterally extensive. Appears massive and fractured towards top. Siderite nodules and layers, and cone-in-cone structures are common. Also includes large-scale concretions.
K	Heterolithic bedding	si - vf	Alternating sand and mudrocks. Also includes wavy, lenticular and flaser bedding. Occurs at a wide range of scales from a few centimeters to tens of meters. *Skolithos* observed.
L	Coquina beds	-	Composed of fragmented bivalve shells. Red to brown colour. Lack primary sedimentary structures. Form discrete, lateral continuous layers surrounded by mudrocks.
M	Mudrocks	cl -si	Fine-grained sediments that are either laminated (shale) and non-laminated(mudrocks). Includes both terrestrial and marine mudrocks.Various degrees of bioturbation. Contains carbonate concretions.
N	Coal and coal shales	-	Organic rich intervals of coal and coal shale. Unit thickness from 1 to 20 cm thick. Locally laterally continuous Found on top of larger sandstones or sandwiched between mudrocks.
O	Paleosols	cl -si	
O_1	Brown and yellow	-	Thicknesses 0.2 to 1 m. Roots, wood fragments and organic matter are found. Occurr within mudrocks and on top of sand-stones. Commonly overlain by coal and coal shale.
O_2	Alternating red and green (Isfjorden Mb)	-	Unit thickness 0.5 to 5 m, with red and green beds between 0.2 to 1 m thick. Spherical nodules found in discrete layers. Only found within the Isfjorden Member.

6.1 Facies A - Large-scale cross-stratified sandstone

Description

Facies A includes fine to medium-grained cross-bedded sandstones. Units are often characterized by a sharp erosive base, often displaying a fining upwards trend. Sedimentary structures vary between large-scale trough- and tabular cross-stratification. Cross-bedding set thicknesses range from 20 to 80 cm, whereas stacking of sets results in unit thicknesses of 0.2 to 4 meters (Fig. 6.1). This facies comprise the coarsest sand grain size in the study area, and sandstones from this facies appear more texturally mature compared to similar facies.

Rip-up clasts (Fig. 6.1E) and plant fragments are frequently observed in the basal parts of unit. Observed colours are grey, yellow and brown, with reddish and dark colours appearing occasionally on weathered surfaces. Upper parts of sandstones may be sparsely bioturbated, whereas lower parts are essentially free of traces. Trace fossils observed within this facies are *Skolithos* and *Diplocraterion*.

Two types of cross-stratification have been distinguished based on the character of bounding surfaces. Planar bounding surfaces characterize the large-scale tabular cross-bedded sandstones, whereas curved bounding-surfaces are seen in the trough cross-bedded units.

The large-scale cross-bedded sandstones are often cemented by calcite and laterally restricted tapering into scree. Facies A is observed on all localities, but appears more prominent in the eastern localities of Barentsøya, Edgeøya and Wilhelmøya compared to Agardhbukta in the west. It occurs throughout the entire De Geerdalen Formation, but is commonly better exposed in the middle parts of the formation and in the lower parts of the Isfjorden Member. Mud draped foresets (Figs. 6.1D, 6.1F) are more commonly noticed on the western localities in the study area.

Interpretation

Formation of large-scale trough- and tabular cross-bedding is commonly assigned to the migration of 3-D and 2-D dunes, respectively, by unidirectional currents of the lower flow regime (Reineck and Singh 1980, Reading and Collinson 1996, Boggs 2011). Complexity of dune morphology is thought to increase at higher current velocities and shallower waters (Collinson et al. 2006, Boggs 2011) and stacking of co-sets represents migration of superimposed bed-forms (Reineck and Singh 1980).

Mud-drapes are not very abundant in elongate sand bars due to a lower amount of suspended sediment (Dalrymple et al. 2012). They are more common in the subtidal part of ebb channels. Structures generated by oscillatory water movements, such as wave ripples and hummocky cross-stratification (facies D and H, respectively) are more frequently found on the seaward end of outer estuary sand-bar complexes, which is more exposed to open-ocean waves (Dalrymple et al. 2012). Erosional reactivation surfaces and mud-draped foresets indicate variations in flow velocities (Reading and Collinson 1996). These features are occasionally observed, including the sections in Agardhbukta and on Teistberget, and could imply a tidal component.

Herein this facies is interpreted as representing the migration of dunes, displaying diverse morphologies, in a subaqueous environment by a dominant unidirectional current. Mud drapes are attributed to slight changes in current velocity, possibly implemented by tidal activity or seasonal changes in stream discharge.

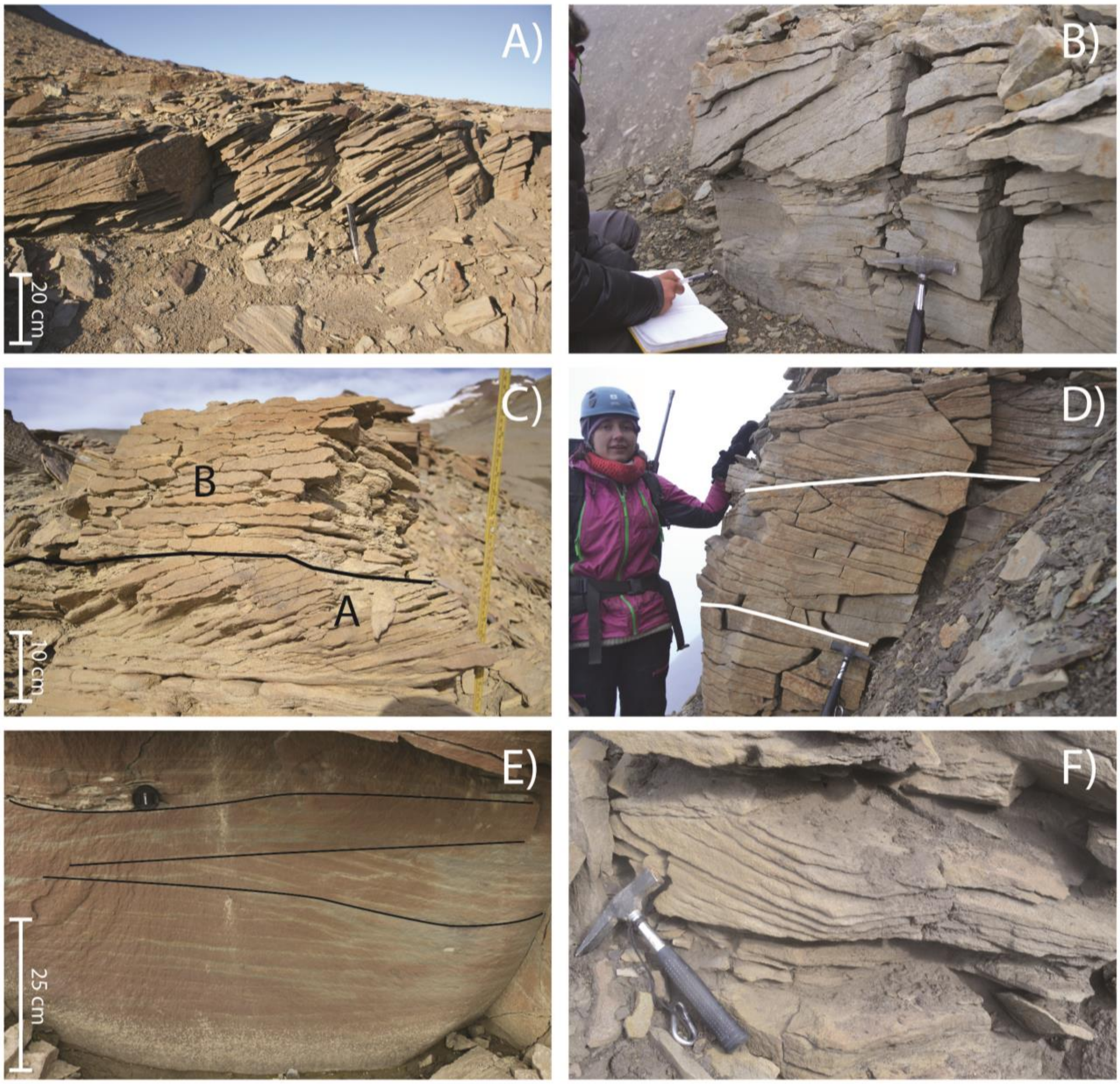


Figure 6.1: *Facies A - Large-scale cross-bedded sandstone* **A)** Tabular cross-bedded sandstone at Wilhelmøya. **B)** Trough cross-bedded sandstone at Šmidtberget, Agardhbukta. **C)** Tabular cross-bedded sandstone (facies A) overlain by small-scale asymmetric cross-bedding (facies B), Hellwaldfjellet, north-eastern Spitsbergen. **D)** Several stacked units of large-scale trough cross bedded sandstones on Šmidtberget, Agardhbukta. **E)** Large-scale cross-stratified sandstone with mud flakes to the left of the lens cap, Wilhelmøya. **F)** Mud draped foresets weather out on smaller-scale tabular cross-bedded sandstones at Svartnosa.

6.2 Facies B - Small-scale cross-stratified sandstone

Description

This facies comprise small-scale asymmetric ripple laminated very fine to fine sandstones. Ripple cross-lamination is arranged in sets of 2 to 10 cm height and stacked in units that are up to 1.5 m thick (Fig. 6.2).

Cementation, mainly calcite, varies from vague to pervasive resulting in differences in appearance within facies. The facies often appear as undulating, parallel wavy to straight bedding/set boundaries without apparent cross-stratification (Figs. 6.2A, B and C). Sparse bioturbation is occasionally observed towards the top of units. Grey, yellow, brown and reddish colours are observed. Weathering of finer material on sandstone bounding surfaces are interpreted as mud drapes (Fig. 6.2D).

Facies are commonly found overlying large-scale cross bedded sandstones (facies A) in fining upwards units. It is often interbedded horizontally bedded sandstones (facies F) and underlying mudrocks (facies M) throughout study area (Fig. 6.2A).

Figure 6.2: *Facies B - Small-scale cross bedded sandstone*. **A)** Small-scale cross-bedded sandstone above horizontally bedded sand (facies F), wave rippled sandstone (facies D) and large-scale cross-stratified sandstone. The lowermost unit represents the strongest current conditions while the units above are inferred to be deposited by a decelerating flow, Teistberget, eastern Spitsbergen **B)** Small-scale cross-bedded ripple laminated sandstone on Wilhelmøya **C)**

Small-scale ripple cross stratification on Friedrichfjellet, Agardhbukta **D)** Mud draped foresets on unidirectional current ripples. Flow direction is towards the right from left, Wilhelmøya.

Interpretation

Asymmetric ripples are formed by unidirectional currents of the lower flow regime in shallow waters (Collinson et al. 2006, Boggs 2011). Furthermore, Collinson et al. (2006) states that grain size is the dominant controlling factor on ripple size. Increasing flow velocity also tends to increase ripple size and complexity of ripple morphology (Boggs 2011). In general, co-sets of ripple lamination form as migrating ripples create net accumulation of superimposed ripples on the bed (Collinson et al. 2006).

Common depositional environments are fluvial and shallow marine, where rip-currents, longshore currents, tidal currents and breaking waves creates unidirectional currents (Reading and Collinson 1996). This facies differ from facies A in scale, but not in form or shape, and may be attributed to weaker currents and smaller grain size (Reineck and Singh 1990). Discovered plant fragments, low abundance of trace fossils and a close proximity to palaeosols, when found in the upper parts of the De Geerdalen Formation, indicates that this facies commonly is associated with terrestrial to coastal depositional environments.

This facies is interpreted to reflect a weaker current in shallow waters compared to facies A and is observed to be similar to facies C. It is often found in terrestrial, fluvial environments, but also appear prominently in marine environments.

6.3 Facies C - Climbing ripple cross-laminated sandstone

Description

Small-scale asymmetric climbing ripple laminated very fine to fine sandstone (Fig. 6.3). Yellow, orange and brownish colours observed. Facies have sharp lower contacts whereas contacts to upper facies are gradual. Facies are observed overlying large-scale cross-bedded and small-scale asymmetric cross-bedded sandstones (facies A and B) and overlain by horizontally bedded sandstones (facies F). Observations are mostly made in the lower, sandstone-rich intervals on Svartnosa, Barentsøya, but it can also be found in the uppermost part on Wilhelmøya. Noted unit thickness is about 0.5 m. It is commonly found together with facies large-scale cross-bedded (facies A) and small-scale asymmetric cross-bedded sandstones (facies B) and horizontally bedded sandstones (facies F).

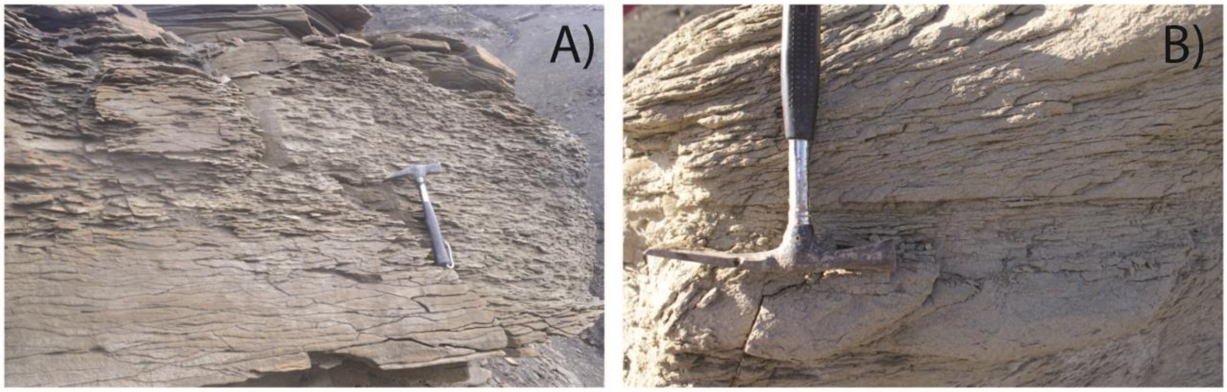


Figure 6.3: *Facies C - Climbing ripple cross-laminated sandstone.* **A)** Climbing ripple cross-lamination in the lower part of De Geerdalen Formation on Svartnosa, Barentsøya **B)** Climbing ripples towards the top of Wilhelmøya.

Interpretation

The formation of climbing-ripple cross-lamination take place as high sediment supply leads to aggradation of ripples with contemporary downstream migration, with the angle of climb reflecting rate of aggradation (Collinson et al. 2006). Environments characterized by periodic rapid deposition, especially sands from suspension, is favourable for this facies formation, whereas environments of low sedimentation rates and much reworking is not (Reineck and Singh 1980).

Fluvial floodplains, with their sub-environments crevasse splays and point bars, and seasonally flooded river deltas are environments where climbing ripple cross-laminated sandstones occur (Reading and Collinson 1996, Boggs 2011).

Herein this facies is interpreted to form under similar conditions (lower flow regime unidirectional currents in shallow waters) as facies B, but might represent seldom episodes of rapid deposition from suspension, possibly related to switches in environmental settings caused by the dynamic nature of the stream and delta front systems.

6.4 Facies D - Wave rippled sandstone

Description

This facies is assigned to units of very fine to fine sandstone that is ripple cross-laminated (Fig. 6.4). The main sedimentary structures are symmetrical wave ripples, planar parallel stratification and mud flakes. Unit thickness typically ranges from tens of cm up to ca. 3 m, while individual bed thickness ranges from 10 to 30 cm. The facies commonly have grey to red weathering colour, while fresh surfaces are usually light grey. Some of the beds are carbonate cemented with siderite concretions and nodules.

Wave ripples are often seen on top surfaces of coarsening upwards units, where the characteristic symmetric ripple form can be observed. The crests, when preserved in the rock, tend to be straight (Figs. 6.4A,B and D). In cross-section, wave ripples are recognized by having undulating bedding, sometimes with bidirectional foresets (Figs. 6.4C, E and F). Mud drapes are also common within the facies and helps tracing out foreset features. The facies are usually moderately to intensely bioturbated and trace fossils such as *Rhizocorallium* and *Skolithos* are often found associated with this facies.

Wave rippled sandstones are commonly found interbedded with heterolithic bedding (facies K), mudrocks (facies M) and hummocky cross-stratified sandstones (facies H) or overlying horizontally bedded sandstone (facies F). Wave ripples are observed throughout the entire study area, mostly in the middle part of the formation.

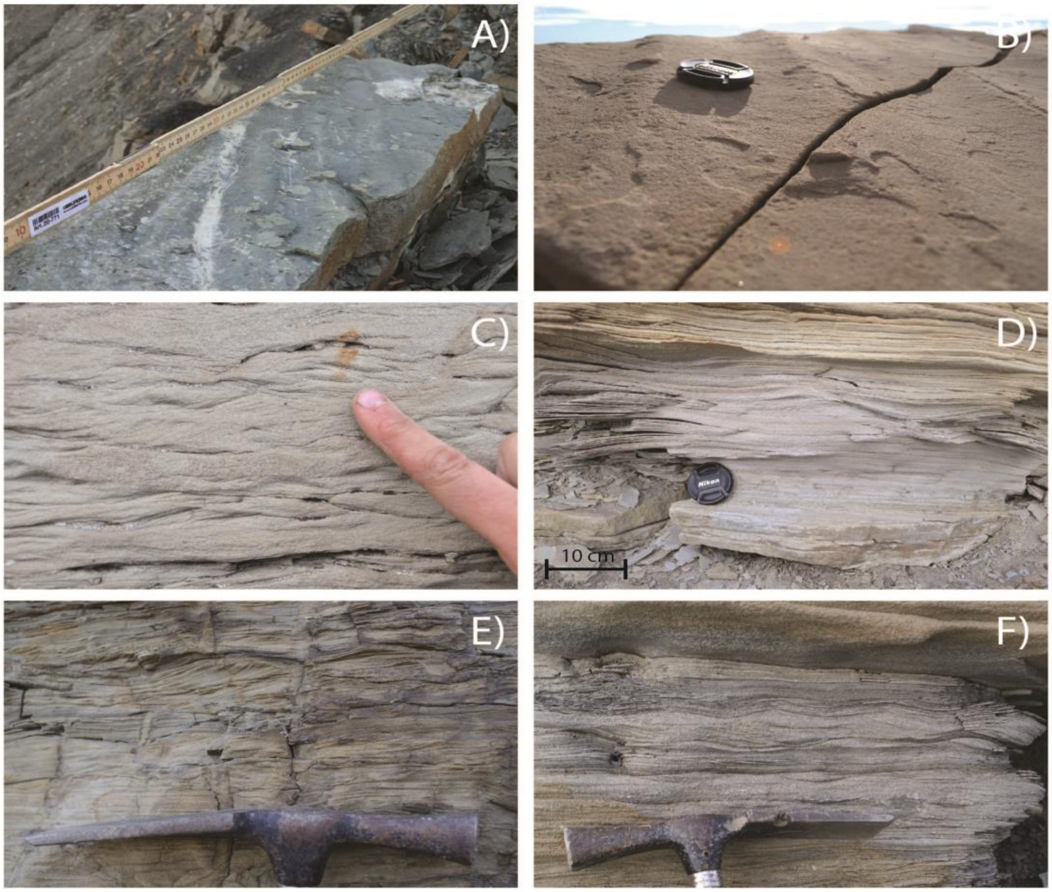


Figure 6.4: *Facies D - Wave rippled sandstone* **A)** Straight symmetrical ripples preserved in the upper part of the De Geerdalen Formation on Klement'evfjellet, Agardhbukta **B)** Peak-shaped ripple crests on a block of sandstone, Hellwaldfjellet, north-eastern Spitsbergen **C)** Bidirectional features interpreted as wave ripple lamination and with a direction of wave propagation towards left and right, Friedrichfjellet, Agardhbukta **D)** Symmetrical ripple crests seen from the side on Hellwaldfjellet, north-eastern Spitsbergen. **E)** Cross-section of interwoven sets of wave ripples, Klement'evfjellet, Agardhbukta **F)** Symmetrical ripples on Hahnfjella, eastern Spitsbergen.

Interpretation

Wave ripples are common in a wide range of sedimentary environments, but are most commonly found in shallow marine settings. They are also referred to as oscillation ripples (Boggs 2011). Wave ripples are thought to be formed by the oscillatory movement of currents in the swash zone, gradually passing into asymmetrical wave ripples and possibly dunes in the shoaling wave zone.

Wave energy is considered the most important marine process in governing coastline morphology (Wright and Coleman, 1973, Galloway 1975, Bhattacharya and Giosan 2003) and is responsible for the redistribution of sand and silt along the coast (Reading and Collinson 1996, Li et al. 2011). Waves may approach the shoreface at an oblique angle, resulting in beach-parallel longshore currents and seaward-directed rip currents (Reading and Collinson 1996). The orientation of the wave ripples alone are therefore not considered a completely reliable indicator of the direction of the palaeo-shoreline (Boggs 2011).

Criteria that was used for recognizing wave ripples were primarily the shape of ripple crest when these are preserved, lower bounding surface of sets and the three dimensional nature of set boundaries (Fig. 6.5A)(Collinson et al. 2006). Ripple crests may be either rounded or peaked where round-crested forms are most common in deeper water, while strongly peaked typically occur in shallow conditions closer to the shoreline (Fig. 6.5B)(Collinson et al. 2006). Characteristic features of wave ripples are scoop-shape interwoven cross sets in sections parallel to wave-propagation direction, while sections perpendicular to this direction consist of sub-horizontal laminae (Collinson et al. 2006).

Wave ripples are thus here inferred to represent a shallow marine environment and additional evidence, i.e. other physical sedimentary structure or marine trace fossils are necessary to make more detailed interpretations of depositional environments.

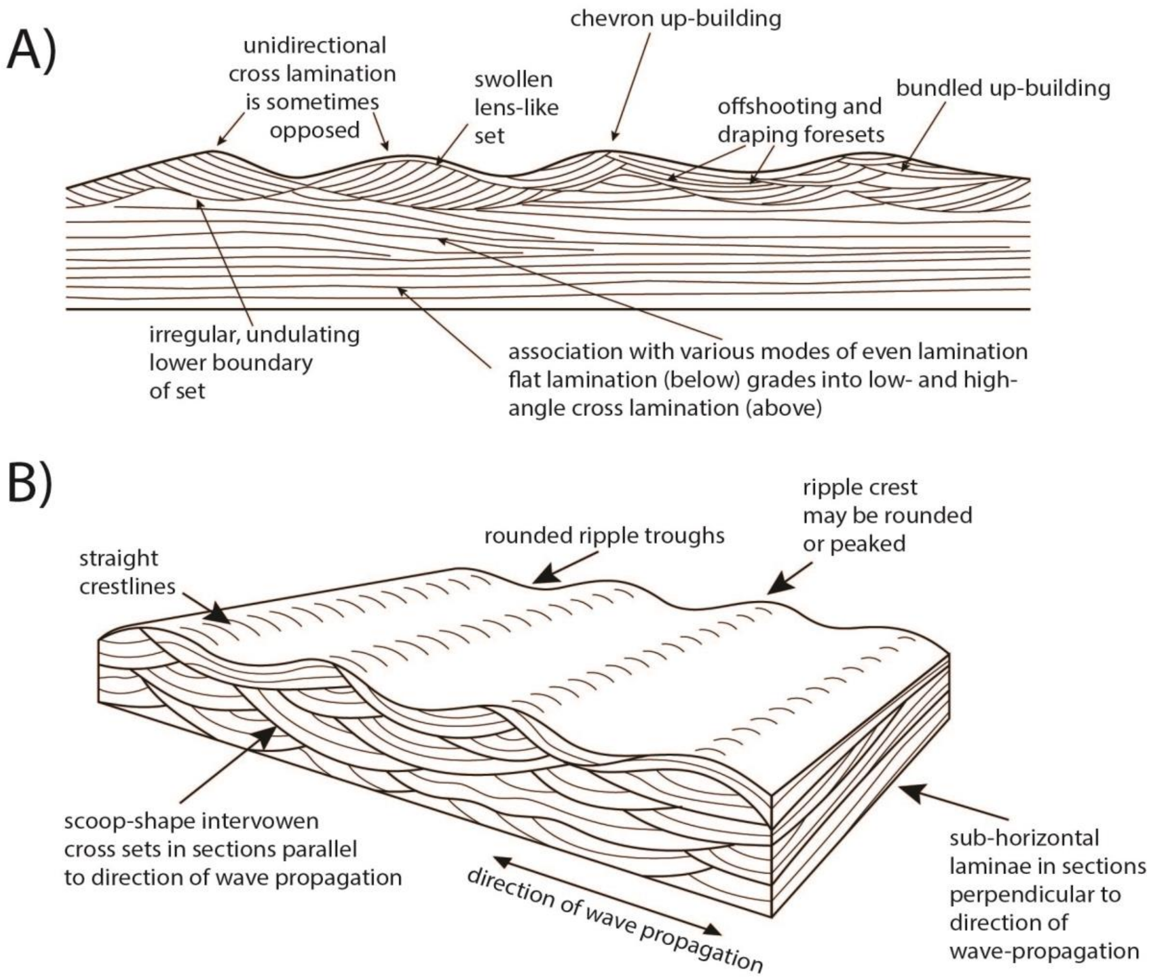


Figure 6.5: Characteristic features of ripples formed by the oscillatory movement generated by ocean waves. **A)** Foresets can be unidirectional and sometimes opposed, but are most commonly bidirectional (from Collinson et al. (2006) after de Raaf et al. 1977) **B)** Three-dimensional idealized block of lamination types that result from the bidirectional movements of water caused by waves (from Collinson et al. (2006) after Boersma1 (970)

6.5 Facies E - Low angle cross-stratified sandstone

Description

This facies consists of silty to fine sand deposited as gently inclined sets of planar parallel stratification with wedge-shaped set boundaries (Fig. 6.6).The colour is usually grey to red-brown when weathered and grey on fresh surfaces. Unit thickness are usually between tens of cm to 1.5 m, while set thickness range between 5 and 15 cm. Individual sets are composed of both beds and lamina, where the former is the most common. These sandstones are commonly bioturbated and contain plant fragments and fish remains were found within low angle cross-stratified sandstone on Klement'evfjellet. This structure is commonly well developed on the Svartnosa locality (Figs. 6.6A, 6.6B, 6.6E), while in other locations it can appear more subtle and harder to recognize in the field (Figs. 6.6C, 6.6D, 6.6F).

It is frequently found overlying or interbedded with wave rippled sandstones (facies D), horizontally bedded sandstones (facies F), large-scale cross-stratified sandstone (facies A) and facies small-scale cross-stratified sandstone (facies B).

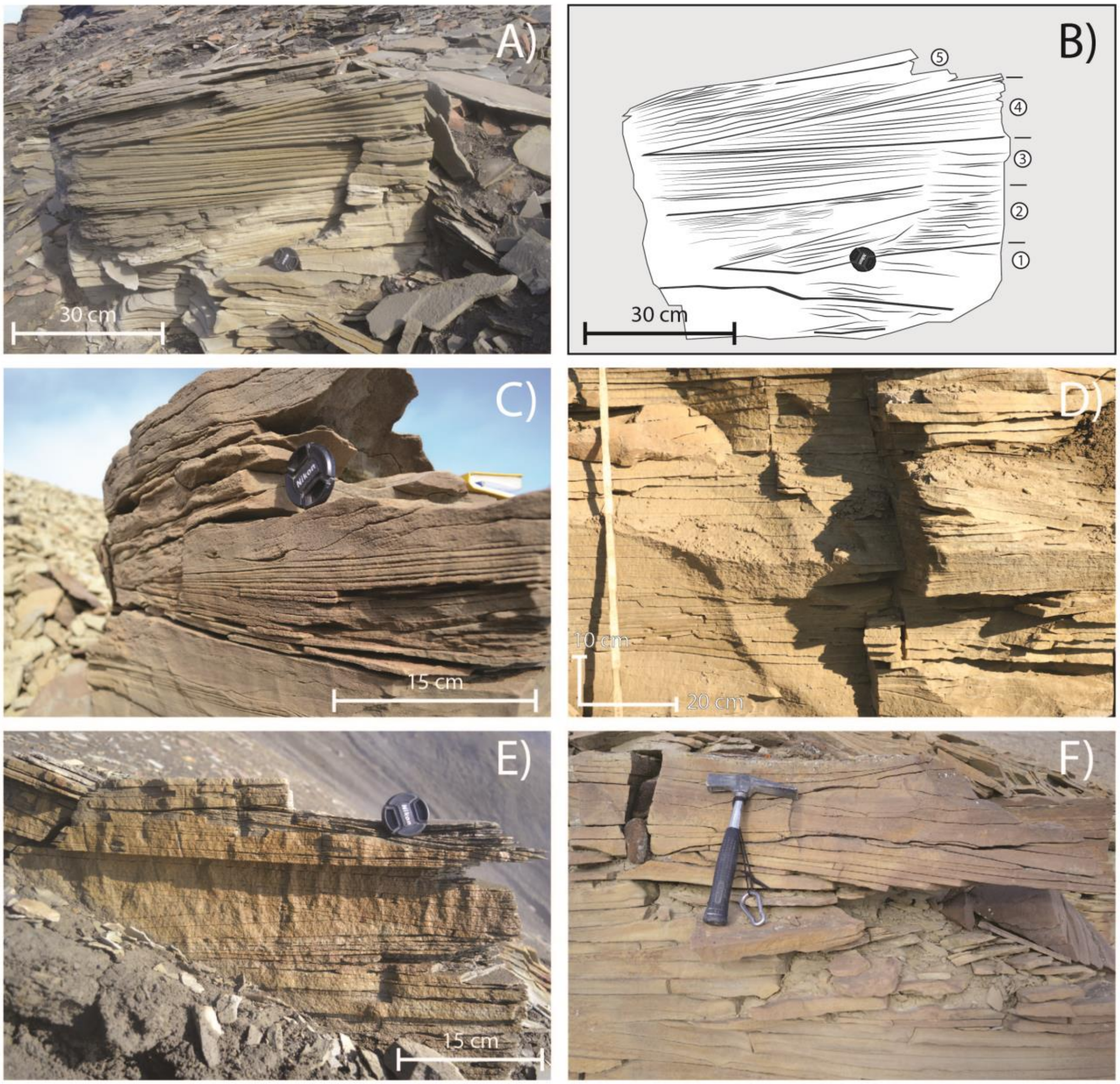


Figure 6.6: *Facies E - Low angle cross-stratified sandstone.* **A)** Low angle cross-stratification, Svartnosa, Barentsøya. **B)** Sketch of Figure A with the different cross-sets labelled. **C)** Oblique view of low angle cross stratified sandstone from the succession on Hahnfjella, eastern Spitsbergen. **D)** Low angle cross-stratification on top of planar parallel stratified (facies F) and large-scale cross stratification (facies A) on Krefftberget, Barentsøya. **E)** Low angle cross-stratification, Svartnosa Barentsøya **F)** Low angle cross-stratification on Hellwaldfjellet, north-eastern Spitsbergen.

Interpretation

Low angle cross-stratification is not considered a diagnostic sedimentary structure, as it can be seen occurring in a range of depositional environments. However, the presence of bioturbation and plant fragments is interpreted as indicators of a proximal position in the shallow marine environment, specifically on the middle to upper shoreface or the beach foreshore (Reading and Collinson 1996). Low angle cross-stratified sandstones typically exhibit a gentle dip seawards when found in foreshore and backshore settings (Reading and Collinson 1996, Clifton 2006), and are herein suggested to represent shoreface deposits.

6.6 Facies F - Horizontally bedded and planar stratified sandstone

Description

This facies is assigned to units of horizontal, planar parallel lamination (PPL) or planar parallel stratification (PPS) (Fig. 6.7). The sandstones are most commonly very fine to fine, although silty and medium grained sandstones do occur. Units range between 30 cm and 2 m in thickness, with mm-thin laminae (Fig. 6.7A) and cm-thick beds (Figs. 6.7B, C and D).Transitions between lamination and bedding occur within units (Fig. 6.7D). Parting lineation, also known as primary current lineation (PCL), is present on upper bedding surfaces within planar parallel stratified sand intervals.

Stratification is seen to vary within units from lamination to bedding, roughly horizontal and parallel. Differences in mud content are also noted and appear more prominent where thinly laminated. The sandstones are most commonly grey to pale yellow, but weathers brown to red. Lower boundaries are typically sharp, while the upper are commonly more gradual. Units are often observed towards the top of sandstone benches. Bioturbation is generally absent in lower parts, but occur towards the upper parts of units. *Skolithos*, *Diplocraterion* and *Rhizocorallium* were observed within this facies on Wilhelmøya and Hahnfjella.

The facies is often found together with heterolithic bedding (facies K) and both large- and small-scale cross-stratified sandstones (facies A and B). In the latter cases, the sandstone units gradually fines upwards from cross-stratified sandstone to horizontally bedded sandstone.

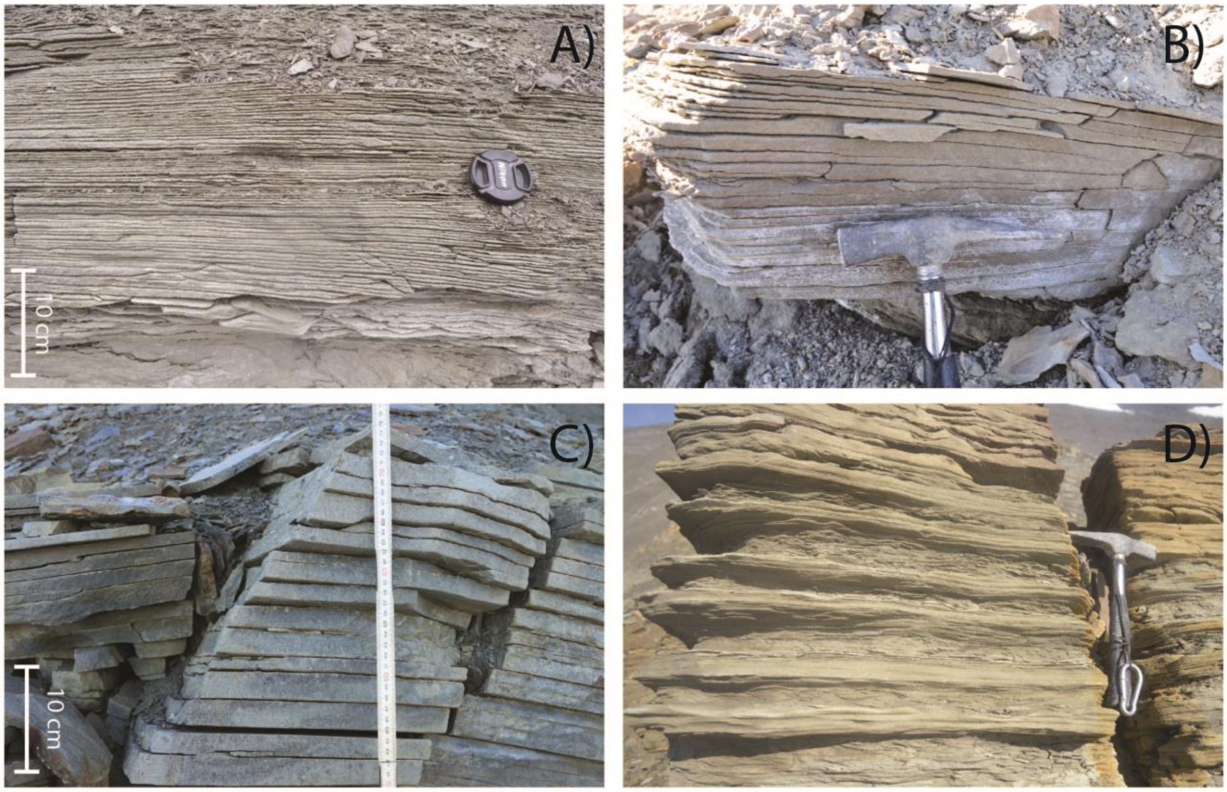


Figure 6.7: *Facies F - Horizontally bedded and planar stratified sandstones.* **A)** Decreasing bedding thickness upwards into fine laminae, Mistakodden, Barentsøya. **B)** Horizontally bedded sandstone on Wilhelmøya. **C)** Centimetre thick layers of horizontally bedded sandstone Klement'evfjellet, Agardhbukta. **D)** Rhythmic alternations of facies F and B (weathered) capped by wave rippled sandstones (facies D) with *Skolithos* traces on Svartnosa, Barentsøya.

Interpretation

Horizontally laminated bedding occurs in various environments, and is thus not considered an unique environmental indicator (Boggs 2011). Formation is assigned to settling of fines from suspension or traction of sand as bedload (Boggs 2011). The latter is referred to as "upper flow regime flat-bed" mode of transport and involves high velocity current and shallow water depth during formation (Collinson et al. 2006), resulting in coarser units than the former. Parting lineation on top surfaces of horizontally stratified sandstones normally forms in high-energy environments such as occur on the upper shoreface, foreshore and beaches.

Laminas are often defined by slight grain size variations or assembling of mica, likely representing subtle variations in depositional environment (Collinson et al. 2006). Among environments of formations are rivers and streams (Boggs 2011). Herein we have considered adjacent facies and marine indicators when trying to establish genetic origin of units.

6.7 Facies G - Massive, structureless sandstone

Description

Massive sandstones that are apparently structureless are included within this facies. They are usually blocky and occur as thick, massive units consisting of very fine to fine and sometimes medium sandstone (Fig. 6.8). Possible primary structures in this facies may be either large-scale cross-stratification or horizontal lamination, but as the name of this facies suggests, they are not easily observed in the field. The facies is commonly heavily fractured (Fig. 6.8A), a feature that can be mistaken for large-scale cross-stratification foresets.

Units are between 1 and 5 m thick, often with a sharp and erosive base. Sandstones of this facies are often enclosed by mudrocks (facies M) and heterolithic bedding (facies K) and are seen throughout the study area. Bioturbation is common. Some outcrops also have plant fragments and mud flakes. Calcite cementation is very common within this facies.

This facies is similarly defined as Rød et al. (2014) undulating fractured sandstone (facies G). However the facies seems less abundant in our study area compared to theirs.

Figure 6.8: *Facies G - Massive, structureless sandstone.* **A)** A large block of massive, structureless sandstone as exposed on Hellwaldfjellet, north-eastern Spitsbergen. **B)** Apparently massive sandstone, Wilhelmøya.

Interpretation

Apparent lack of sedimentary structures is interpreted to be mainly the result of weathering and diagenesis, primarily calcite cementation. In some cases, the lack of primary structures can also be attributed to intense bioturbation and biogenic reworking of sediments during deposition. Liquefaction and flow of waterlogged sediments can also result in the destruction of primary sedimentary structures (Collinson et al. 2006).

Massive, structureless bedding may also be the product of very rapid deposition or from liquefaction of sediment due to a sudden shock following deposition (Boggs 2011).

6.8 Facies H -Hummocky cross-stratified (HCS) to swaley cross-stratified (SCS) sandstone.

Description

This facies is defined as sandstones displaying hummocky and swaley cross-stratification. It consists of 20 cm to 1 m thick sandstone beds with grain sizes from silt to fine sand, but the best developed hummocks are primarily found in very fine to fine sands. The sandstones are characterized by cross-laminae in undulating sets (Fig. 6.9). The concave-up part of the structure is referred to as “swales”, while the convex-up part is referred to as “hummocks” (Boggs 2009, 2011). Individual laminae sets are commonly between 5 and 20 cm thick.

The sandstones are typically grey to yellow when unweathered, and display an orange to reddish brown colour where it is weathered. The beds are usually moderately to intensely bioturbated and *Skolithos* and *Diplocraterion* are observed both within units and more commonly on top surfaces where they are occur as circular holes.

Hummocky to swaley cross-stratified sandstones are often found below facies D (Wave rippled sandstone) and commonly together with facies F (horizontally bedded sand), facies M (Mudrocks), facies K (Heterolithic bedding) and facies E (Low angle cross-stratified sandstone). Hummocky cross-stratified sandstones are common in upwards coarsening sequences in the lower part of the De Geerdalen Formation throughout the study area. It is found frequently in the Agardhbukta outcrops (Figs. 6.9A, C, D, and E). It is also common in coarsening upwards sequences on Svartnosa.

Interpretation

Hummocky cross-stratification is a sedimentary structure in sandstones that show a distinct undulating geometry of lamination. The first usage of the term is generally attributed to Harms et al. (1975), while similar structures were previously known as “truncated wave ripple laminae” by Campbell (1966). The geometry and internal structures of hummocky cross-stratified sandstones (Fig. 6.10) are interpreted to form by the migration of low-relief bed forms in primarily one direction due to the influence of combined wave surge and unidirectional currents (Nøttvedt and Kreisa 1987, Walker and Plint 1992).

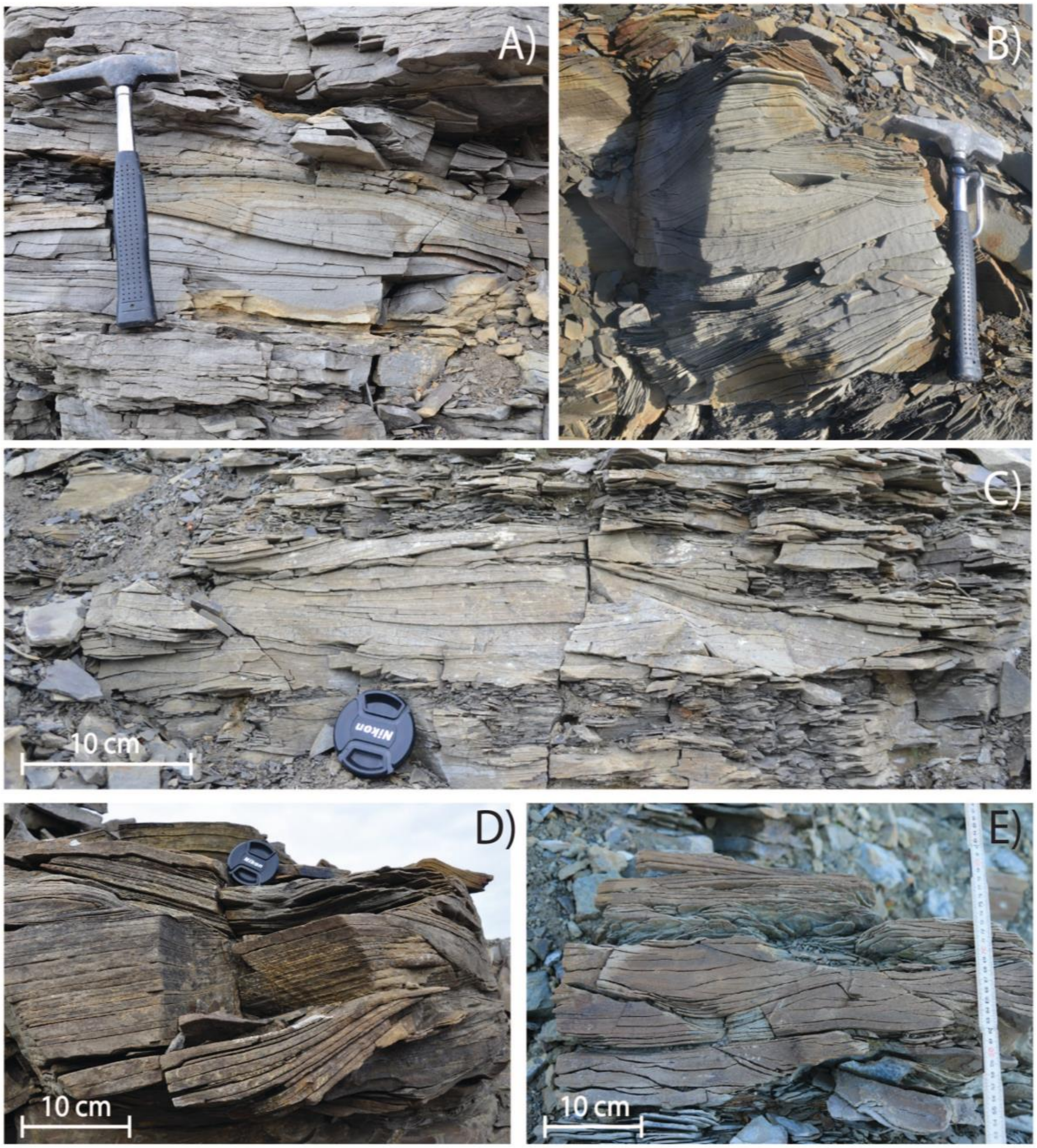


Figure 6.9: *Facies H - Hummocky and swaley cross-stratified sandstone* **A)**"Micro-hummock" is a term applied to smaller scale hummocky cross-stratification. The small scale suggests a more proximal position close to the lower shoreface, Friedrichfjellet, Agardhbukta **B)** Excellent three-dimensional structure of hummocky and swaley cross-stratification exposed on Svartnosa, Barentsøya **C)** Small-scale hummocky cross-bedding on Friedrichfjellet, Agardhbukta **D)** Another good block exposure of the structure in hummocky cross-bedding from Friedrichfjellet, Agardhbukta **E)** Klement'evfjellet, Agardhbukta.

The exact formation of hummocky cross-stratification remains enigmatic despite being the subject of numerous studies (Harms et al. 1975, Harms et al. 1982, Cheel and Leckie 1993, Dumas and Arnott 2006, Yang et al. 2006, Quin 2011). However, it is widely recognized as being characteristic of deposition in shallow marine storm-dominated inner shelf to lower shoreface settings (Harms et al. 1975, Harms et al. 1982, Cheel and Leckie 1993, Johnson and Baldwin 1996, Midtgaard 1996, Yang et al. 2006), forming below fairweather wave base and above, but near storm weather wave base (Dumas and Arnott 2006). As hummocky cross stratification is often found deposited in mud-dominated successions, it is interpreted to represent storm periods, while the shales are deposited in quiet periods between storms (Collinson et al. 2006).

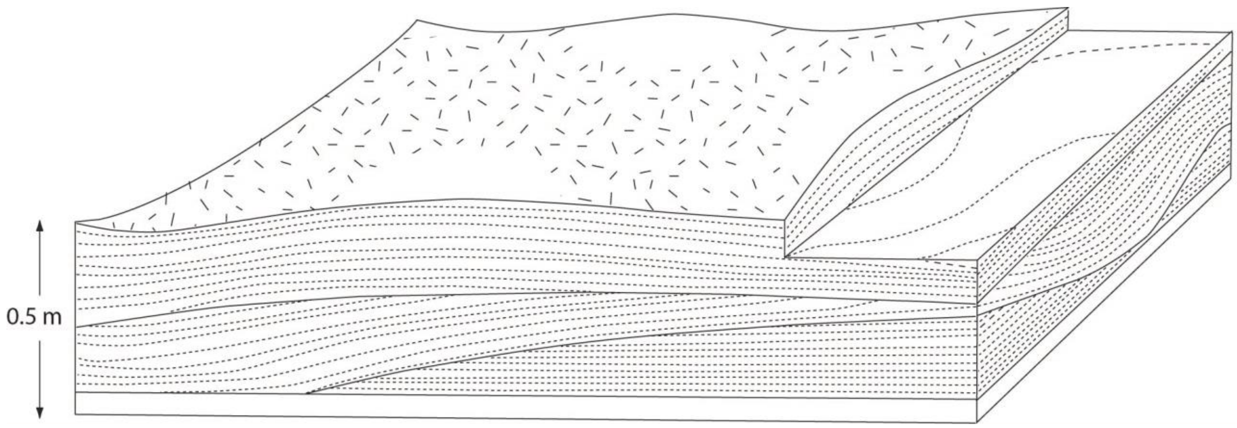


Figure 6.10: Internal geometries and bedform of hummocky cross-stratification (from Quin 2011, after Harms et al. 1975).

Some of the defining characteristics of hummocky cross-stratified sandstones include (Harms et al. 1982, Nøttvedt and Kreisa 1987, Midtgaard 1996, Dumas and Arnott 2006):

(1) Erosional lower set boundaries and low-angle dips, commonly less than 10° and rarely up to 15 °.

(2) Laminae above set boundaries are parallel or close to parallel with these

(3) Separation of laminae-sets by thin layers of mud or low angle erosional surfaces

(4) Hummocky laminae systematically thicken laterally downdip into swales

(5)

Draped scour surfaces are emphasized as important features of hummocky cross-stratification (Dott and Bourgeois 1982, Bourgeois 1983). Swaley cross-stratification forms similarly to hummocky cross-stratification, but in a more proximal setting, closer to the shoreface where lower aggradation rates favours the preservation of swales (Walker and Plint 1992, Hampson and Storms 2003, Dumas and Arnott 2006).

6.9 Facies I - Soft sediment deformed sandstones

6.9.1 Sub-Facies I_1 - syn-sedimentary deformed sandstones

Description

Very fine to fine sandstones displaying soft sediment deformation structures (Fig. 6.11). Facies often occur in the basal part of thick sandstone units, but are also found interbedded between other sandy facies. Loading structures are commonly observed on the contact to underlying beds. Interbedded units are seen to show modest thicknesses (10 to 20 cm) compared to equivalents found in the lower reaches of sandstones with thicknesses up to 1 m. Plant fragments and mud clasts are found within facies. Sandstones above and below may be partially to completely undisturbed.

This facies are found at on the Svartnosa and Mistakodden localities on Barentsøya, primarily occurring in the lower parts of the De Geerdalen Formation.

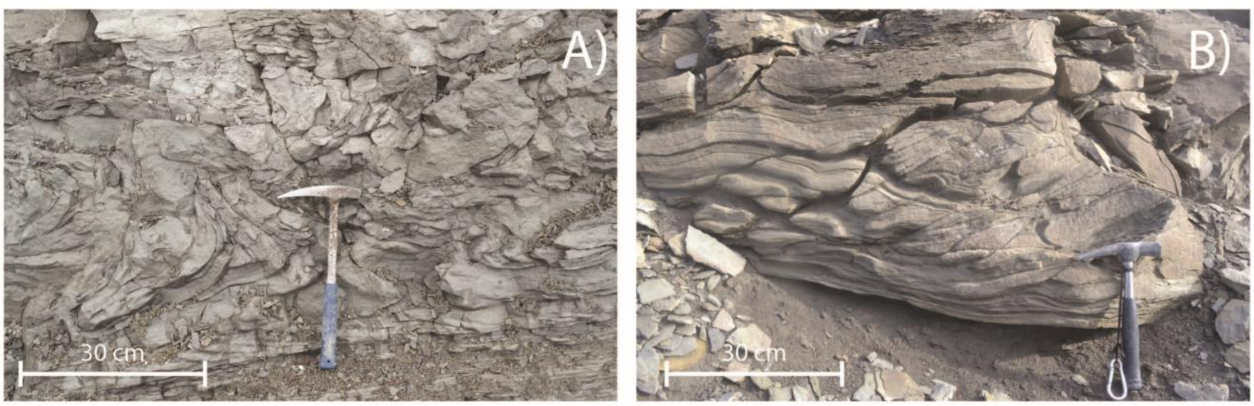


Figure 6.11: *Facies I_1 - Syn-sedimentary deformed sandstones* **A)** Heavily deformed sandstone on Mistakodden, Barentsøya. **B)** Characteristic features are the sharp lower contact and the disturbed internal laminations. From Svartnosa, Barentsøya.

Interpretation

Soft sediment deformation structures may generate from gravitational processes like downslope sliding and slumping or rapid loading of sediment (Reineck and Singh 1980). In the latter case vertical adjustments occur as sands are superimposed upon a hydroplastic mud layer (Boggs 2011). In general the genesis of soft sediment deformation structures is assigned to times prior to consolidation of the sediment (Reineck and Singh 1980).

Soft sediment deformation features are commonly found in environments with high sedimentation rates, which is often the case on the distal delta front of river-dominated deltas (Reading and Collinson 1996, Bhattacharya 2006, Bhattacharya and MacEachern 2009). For example, mass movement following deposition affects approximately 40 % of the sediment supplied to the Mississippi delta (Coleman 1981, Reading and Collinson 1996). Herein, this sub-facies is interpreted as sands being subjected to mechanical stresses imposed by rapid deposition, at times prior to consolidation of the sediment. Abundant plant fragments and mud clasts indicate a close affiliation with a terrestrial sediment source (Eide et al. 2015).

6.9.2 Sub-Facies I_2 - Erosive-based sandstone lenses

Description

This facies is comprised of irregularly based, laterally restricted, very fine to fine sandstones characterized by abundant soft sediment deformation (Fig. 6.12). Units are measured 2 to 4 m in width and heights ranging from 0.3 to 1.5 m. The irregular lamination seen within the sandstone bodies are also present in the upper parts of the underlying deformed mudrocks. The sandstone bodies are calcite cemented, showing cone-in-cone structures towards the top. Plant fragments are also found.

In the upper part of the Muen section this facies is prolific as sandstones, displaying severe soft sediment deformation, are capsuled in mudrocks (facies M) with adjacent hummocky cross-stratified sandstones (facies H), carbonate rich sandstone (facies J) and heterolithic successions (facies K). Facies I_2 is currently only recognized on the Muen locality.

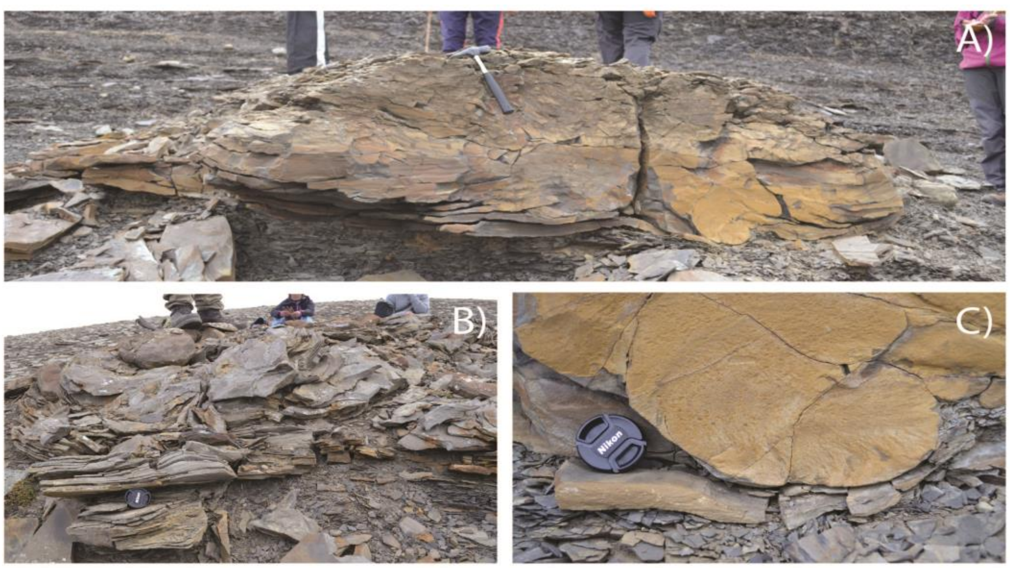


Figure 6.12: *Facies I_2 - Soft sediment deformed sandstone lenses* **A)** A soft-sediment deformed sandstone body solely capsuled in mudrock. **B)** Deformed sandstone lens. **C)** Loading structure on the sole of the sandstone body. All photographs were taken at Muen, Edgeøya.

Interpretation

As previously stated soft sediment deformation structures may generate from gravitational processes e.g. downslope sliding and slumping or rapid loading of sediment (Fig. 6.13) (Reineck and Singh 1980, Bhattacharya and MacEachern 2009).

The solitary confinement in mud and the lateral restricted nature of this sub-facies may indicate a different genetic origin than sub-facies I_1. Given the marine indicators found in adjacent facies and less abundant plant fragments, this sub-facies may be considered more distal than sub-facies I_1.

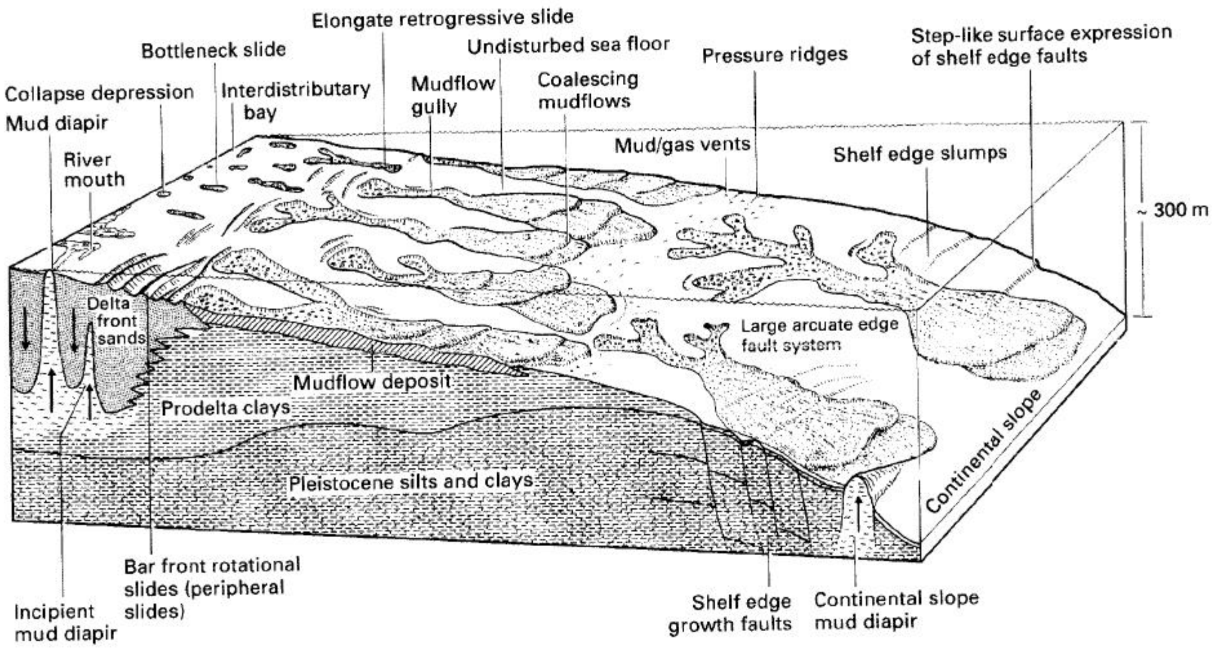


Figure 6.13: Different types of deformation caused by mass-movement of sediment supplied to the delta front and pro-delta (from Reading and Collinson 1996).

These lenticular to undulating sandstone beds corresponds well to 'erosive Offshore Transition Zone' (OTZe) sandstones as described from the Beckwith Plateau in Utah (Eide et al. 2015). These are generally characterized by erosive, undulating sandstone geometries (Eide et al. 2015), similarly to facies I_2. However, an important difference is that in this case, no observations of numerous, erosive gutter casts has been made. In contrast, beds that are tabular and laterally continuous are termed 'tabular Offshore Transition Zone' (OTZt) (Eide et al. 2015). These are generally more laterally continuous, separated by thin mudstones and normally non-erosive (Eide et al. 2015). Favourable conditions for the generation of OTZe occur when parasequences prograde into shallow waters (Eide et al. 2015). Seismic studies in the Barents Sea (Glørstad-Clark et al. 2010, Klausen et al. 2015) indicate that paleo-water depth was fairly shallow, i.e. 400 - 500 m.

However, the nature of these sandstones remains enigmatic and further studies are needed to establish the genetic origin.

6.10 Facies J - Carbonate rich sandstone

Description

This facies comprise very fine to fine sandstones characterized by structures formed during diagenesis (Fig. 6.14). The sandstone units are commonly hard and heavily cemented, which makes observations of primary sedimentary structures difficult (Fig. 6.14C). Secondary sedimentary structures includes cone-in-cone (Fig. 6.14A, 14B), siderite beds (Fig. 6.14F) and calcareous concretions.

A decrease in cementation to adjacent facies is noticed as they often appear less consolidated. Colour variation between grey, brown and red are observed. Scarce to heavy bioturbation is noticed. Large calcite concretions, up to meter sized, are found on numerous locations. Facies can appear in close association with other facies forming a sandstone bench or be interbedded by mud. These sandstone benches commonly form very distinct layers that may be laterally continuous for several tens to hundreds of metres before they pinch out (Fig. 6.14D).

Included in this facies are also pervasive carbonate cemented reddish very fine sandstones. They are characterized by undulatory bedding, occasionally ripple laminated, and with a rusty red colour (Figs. 6.14E and G). Towards the top the units are characterized by a nodular texture with frequent fractures. Thicknesses are 1 to 2 meters and occurrence is restricted to the upper parts of the formation. Occurrence of these rusty red beds is restricted to three discrete levels in Agardhbukta, where the uppermost may be the Slottet Bed. Similar units are found in the upper parts of Hellwaldfjellet and Wilhelmøya sections.

Siderite occurs as nodules within mudstones and sandstones grouped in other facies, or as nodules forming distinct layers (Fig. 6.14F). Siderite layers commonly have thicknesses of 10 to 30 cm. Siderite beds are found throughout the formation, but appear to be more prominent at the Agardhbukta localities.

Cone-in-cone structures are found on the northern localities of Hellwaldfjellet, Wilhelmøya and in the lower parts of De Geerdalen Formation on Muen. Whereas calcareous concretions, up to meter sized, are found within both mudstone and sandstone intervals throughout the study area.

Figure 6.14: *Facies J - Carbonate rich sandstone*

A) Cone-in-cone structures on Tumlingodden, Wilhelmøya. **B)** Continuous layer with cone-in-cone structures, Hellwaldfjellet, north-eastern Spitsbergen. **C)** Calcareous sandstone at Friedrichfjellet, Agardhbukta. **D)** Carbonate rich sandstone that can be laterally traced (white arrows) across the gully, Friedrichfjellet, Agardhbukta. **E)** The sandstones are usually laminated in the lower parts of units, becoming fractured towards the top, Šmidtberget, Agardhbukta. **F)** Continuous siderite beds, Friedrichfjellet, Agardhbukta. **G)** Horizontally bedded carbonate rich sandstone on Šmidtberget, Agardhbukta.

Interpretation

Formation of the carbonate cemented horizons remains a topic of debate. Recent investigations of carbonate cemented surfaces and concretions by Tugarova and Fedyaevsky (2014) suggests a genesis driven by micro-organisms and a biochemical precipitation of carbonates during very early diagenesis in a shallow marine environment. Klausen and Mørk (2014) described similar facies from the De Geerdalen Formation on Hopen and interpreted the carbonate beds as condensed sections deposited during periods of lower siliciclastic input and as representing discrete marine inundations.

Maher et al. (2016) argue for carbonate nucleation in small tensile cracks and on carbonate shell fragments during shallow faulting and seepage. Cone-in-cone structures are historically believed to result from the precipitation and growth of fibrous calcite crystals during early diagenesis (Franks 1969).

Formation of siderite predominantly occur in organic-rich brackish to meteoric pore-waters depleted of SO_4^{2-} and is commonly found in fine grained deltaic to coastal sediments (Morad 1998). Observed siderite concretions and layering might indicate a slightly higher continental influence upon marine sedimentation with organic-rich stagnant waters close to the delta front (Pettijohn et al. 1987).

6.11 Facies K - Heterolithic bedding

Description

Heterolithic bedding is here defined as thin beds of very fine to fine sandstone and siltstone alternating with mudstones (Fig. 6.15). The thickness of mud- and sand layers ranges from 1 mm to few centimetres. Units are up to 9 m thick (Fig. 6.15B). The facies includes all units with interacting mud and sand, e. g. storm deposits and lenticular-, wavy-, and flaser-bedding. Sedimentary structures preserved in the sandstones of heterolithic successions are hummocky cross-stratification and ripple cross-stratification. Bioturbation is common towards the top of units and *Skolithos* is observed within this facies.

Heterolithic bedding is found at all levels in the De Geerdalen Formation in a wide range of scales. The facies is seen throughout the study area, but type of bedding differs on the different localities. Storm deposits seem to be more dominant in the Agardhbukta sections, while flaser, lenticular and wavy bedding is more abundant in the middle to upper parts of the De Geerdalen Formation.

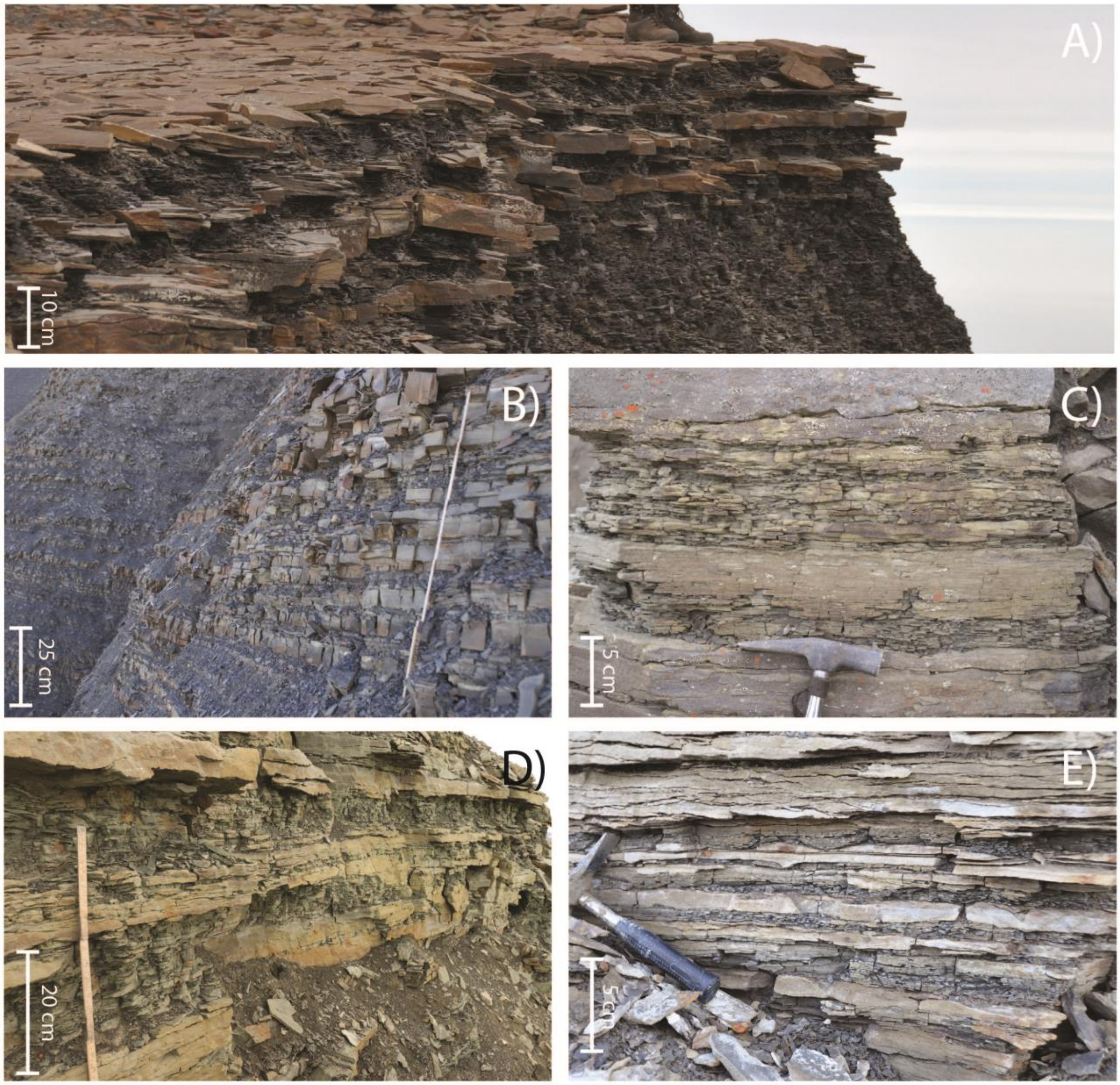


Figure 6.15: *Facies K - Heterolithic bedding.* Note the different scales of sandstone bed thickness.

A) Heterolithic bedding in the lower part of the De Geerdalen at Muen, Edgeøya. **B)** Heterolithic bedding, interpreted as storm deposited sandstones, in the lower part of the succession at Klement'evfjellet, Agardhbukta. **C)** Wavy bedding, Šmidtberget, Agardhbukta. **D)** Larger scale of alternating sand and shale on Friedrichfjellet, Agardhbukta. **E)** Heterolithic bedding overlain by small-scale cross-laminated sandstone (facies B), Friedrichfjellet, Agardhbukta.

Interpretation

Heterolithic bedding indicates alternating flow regime in an environment with both sand and mud available (Figs. 6.15, 6.16) (Davis 2012). Mud is deposited from suspension, while the sand is deposited during current or wave activity and is thus ripple laminated (Reineck and Singh 1980).

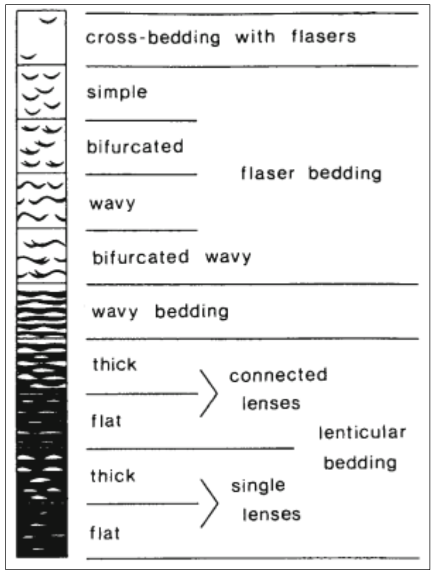


Figure 6.16: Flaser, wavy and lenticular bedding are determined based on the ratio of sand and mud. High sand-mud ratios favours the generation of flaser bedding, while wavy and lenticular bedding are characterized by an equal and low sand - mud ratio (from Reineck and Wunderlich 1968).

The distinction between flaser, wavy and lenticular bedding are based on mud/sand ratio and lateral continuity on the sand and mud layers. Wavy bedding has approximately equal amount of sand and mud. Flaser bedding contains more sand than mud, and lenticular bedding in turn has more mud than sand. The different types of heterolithic bedding reflect which grain size was favoured to be deposited and preserved in the palaeoenvironment (Reineck and Singh 1980). The preservation potential for the structures is high. Presence of vertical burrows is one of the most common biogenic signatures in tidal environments because they form beneath the surface, giving high preservation potential (Davis 2012). Flaser, wavy, and lenticular bedding is one of the most distinct indicators of tidal environments (Prothero and Schwab 1996, Boggs 2011), especially in intertidal zones, but can also occur in subtidal environments (Davis 2012).

Heterolithic bedding may form in the offshore transition zone when storm-transported sands interact with mudrocks deposited from suspension during fair weather (Figs. 6.15A and B). Associated facies in storm deposits are hummocky cross-stratified (facies H) and wave rippled sandstones (facies D) and bioturbated mudrocks (facies M). The sand to mud ratio tends to increase landwards (Walker and Plint 1992).

Heterolithic bedding is herein interpreted to have formed in tide-influenced or tide-modified delta front shoreface or delta plains (Ichaso and Dalrymple 2009, Dashtgard et al. 2012) or as storm deposits in the transition zone (Walker and Plint 1992, Johnson and Baldwin 1996). The distinguishing between the settings is based on associated facies, and the scale, thickness and lateral continuity of the sand- and mud layers. The facies could also possibly originate from other environments undergoing alternating flow regime.

6.12 Facies L - Coquina beds

Description

The unit consists mainly of fragmented bivalves, lacking sedimentary structures (Fig. 6.17). The thickness of the coquina beds is from 10 to 90 cm. All the observed units are cemented with red to brown colour, displaying orange and purple weathering colours. Coquina beds are found as discrete laterally continuous layers sandwiched between mudrocks (facies M) and locally as minor shell accumulations within sandstone bodies.

The occurrence of coquina beds is restricted to the lower parts of the Isfjorden Member, and is present at all localities where the Isfjorden Member is exposed, except on Klement'evfjellet, but this can be due to gentle slope angle and thus total scree cover on the uppermost part of the locality. Coal shale (facies N) is often found in close proximity to coquina beds. The coquina beds are more common and better developed in the northern part of the study area, especially on Wilhelmøya and Hellwaldfjellet.

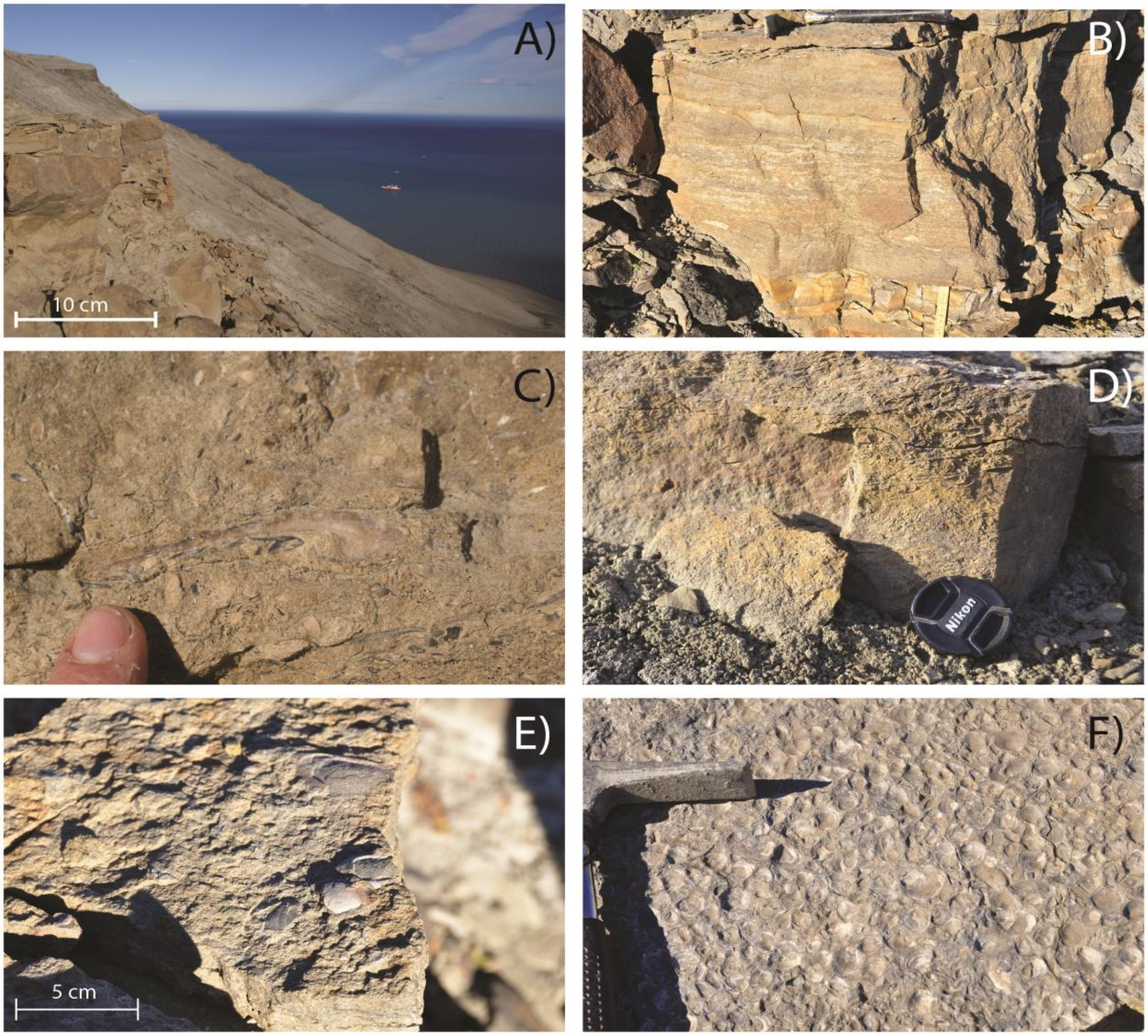


Figure 6.17: *Facies L - Coquina beds.* **A)** Laterally continuous coquina shell bank below the dolerite sill on Tumlingodden, Wilhelmøya. **B)** Coquina bed in a sandstone unit on Tumlingodden, Wilhelmøya. **C)** Bivalve shell fragment in a coquina bed on Tumlingodden, Wilhelmøya. **D)** Numerous fragmented shells in a coquina on Hellwaldfjellet, north-eastern Spitsbergen. **E)** Preserved bivalves covering the top surface of a coquina bed, Hellwaldfjellet, eastern Spitsbergen. **F)** Coquina bed cast, Hellwaldfjellet, north-eastern Spitsbergen.

Interpretation

Fragmented shells indicate high energy environment. Shell banks are often located on beaches, subjected to intense wave reworking. Massive erosion and transportation of shells can lead to concentration of shell fragments in beds where the hydrodynamic energy is small enough for deposition (Reineck and Singh 1980).

Open lakes with low siliciclastic input are dominated by chemical and biochemical processes. Most of the sedimentation is controlled by inorganic carbonate precipitation and production of shells by calcium carbonate- or silica-emitting organisms (Boggs 2011). Invertebrate remains such as bivalves, gastropods, ostracods and freshwater algae can be preserved in lacustrine settings (Boggs 2011). Plants are also commonly abundant on margins of shallow lakes and can be deposited as coal shale, especially during later stages of lake infilling (Boggs 2011).

Coquina beds could also form due to very slow rates of deposition following a major avulsion, delta lobe or distributary switching, or eustatic sea-level rise, e.g. the 'Abandonment facies association' of Reading and Collinson (1996). These conditions are common in interfluvial areas and may include limestone, coals or highly condensed horizons bioturbated by plants or animals (Reading and Collinson 1996). Abundant molluscs are found in interdistributary bay sediments in the modern Mississippi delta (Frazier 1967).

The Isfjorden Member is interpreted to be deposited in a shallow marine setting with possibly local lagoonal environments (Mørk et al. 1999). Based on associated facies, field observations also point towards a proximal shallow marine origin, and coquina beds may represent wave-reworked shallow marine shell banks accumulated by currents or waves.

6.13 Facies M - Mudrocks

Description

This facies is used to describe successions consisting of fine-grained material (clay to silt) that may be either laminated (shale) or non-laminated (mudrock), following the proposed nomenclature of Lundegard and Samuels (1980).

Mudrocks constitute the bulk of the succession at essentially every location, although they are often covered by scree, as they are more susceptible to weathering and erosion. This makes detailed field studies both difficult and time-consuming. Scree-covered areas are often inferred to be composed of mud successions.

The mudrocks facies varies greatly in thickness throughout the study area from the scale of tens of meters to a few centimetres in heterolithic successions. Laminated mudrocks are most common and may encase thin beds of silty to very fine sandstone. The colour of mudrocks is dominantly grey or black, but may also be of yellow, white or purple colour (Figs. 6.18).

In general the mudrocks are characterized by horizontal or gently undulating laminae. Load structures and irregular lamination were occasionally observed in the uppermost part of units where overlain by thick sandstone bodies. Concretions of either calcite or siderite are common and often observed in the mudstone succession at various localities.

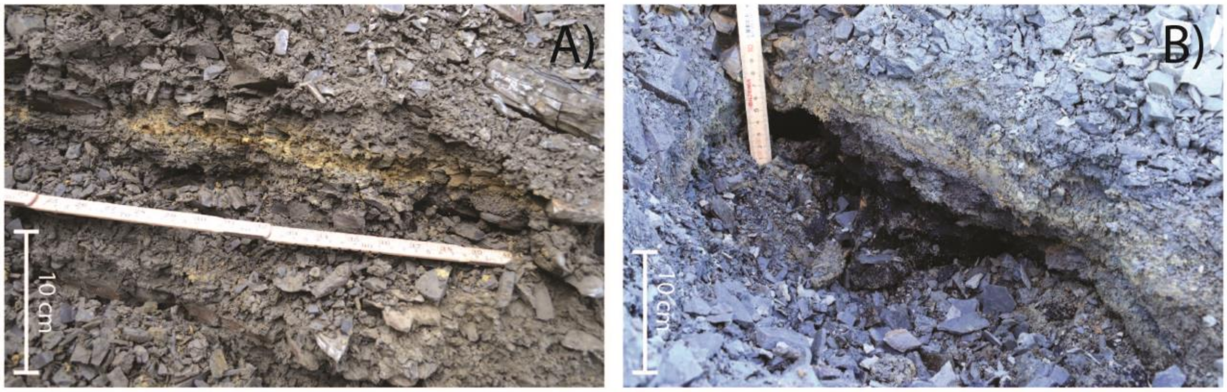


Figure 6.18: *Facies M - Mudrocks* **A)** Grey and yellow mudrocks on Friedrichfjellet, Agardhbukta. **B)** Mudrock, Klement'evfjellet, Agardhbukta.

Interpretation

Shales and mudrocks are most commonly interpreted as being deposited in low energy environments due to settling from suspension (Aplin and Macquaker 2011, Boggs 2011) or as hyperpycnal fluid muds in prodelta areas on muddy shelves (Bhattacharya and MacEachern 2009, Ichaso and Dalrymple 2009).

When bioturbated, mudrocks are, in most cases, deposited in marine offshore to marginal-marine delta front settings (Aplin and Macquaker 2011). Mudrocks are also a dominant lithology in coastal environments, such as lagoons, tidal flats, inter-distributary bays, tidal-fluvial channel deposits, mouth bars and terminal distributary channels (Bhattacharya 2006, Ichaso and Dalrymple 2009).

Marine clay and silt flocculate and settle on the seafloor, forming laminated or massive appearing layers (Ch.5, Collinson et al. 2006). Continental mudrocks on the other hand occur abundantly on the delta plain, between channels on the floodplain. In this study, these are treated separately, as they are susceptible to weathering and soil formation and are thus referred to as palaeosols (Collinson 1996, Enga 2015). These are described in greater detail under facies O (Palaeosols).

Yellow colouring of laminae and beds of mudrock (Fig. 6.18A) is interpreted to be caused by the presence of iron sulphur minerals, such as pyrite or marcasite (Boggs 2009). These minerals are commonly abundant in marine shales and may indicate reducing conditions during deposition or later during burial and diagenesis (Pettijohn et al. 1987, Boggs 2009). Reducing conditions may either be caused by anoxic, stagnant water conditions when found in marine shales or due to presence of organic matter on tidal flats (Pettijohn et al. 1987, Boggs 2009)

6.14 Facies N - Coal and coal shale

Description

Units of coal and coal shale are from 1 to 20 cm thick (Fig. 6.19). The units often appear laterally continuous over tens of meters when examined, but scree cover is common (Figs. 6.18A and C). Coal and coal shales are usually found in close proximity to the top of larger sandstones, or sandwiched in mud. Coals are distinguished from coal shales by being more solid, reflecting a higher amount of plant material.

Coal and coal shale is commonly associated with underlying palaeosols (Fig. 6.19B), but coal shale surrounded by grey shale is observed on Wilhelmøya and Hahnfjella. No trace fossils, but *Rhizoliths*, are found in the facies. The facies is observed at most localities, but is only seen under and in the Isfjorden Member, and not in the lowermost parts of the De Geerdalen Formation.

Figure 6.19: *Facies N - Coal and coal shale*

A) Two laterally discontinuous coal seams alternating with palaeosols, Tumlingodden, Wilhelmøya. **B)** Coal shale overlying horizonated palaeosol, Šmidtberget, Agardhbukta. **C)** Laterally continuous coal and coal shale in the uppermost part of sediments interpreted as delta plain deposits, Blanknuten, Edgeøya.

Interpretation

According to the Mack et al. (1993) classification system is peat formed in place defined as Histosol soil. The definition is regardless of thickness Mack et al (1993). The Soil Survey Staff (2014) classification system requires at least 40 cm of peat before burial. Coal must thus be considered as a type of palaeosol with this definition. According to the commonly used definition from Schopf (1956 p. 527): "Coal is a readily combustible rock containing more than 50 % by weight and more than 70 % by volume of carbonaceous material, formed from compaction or induration of variously altered plant remains similar to those of peaty deposits (…)". The interpretation of coal and coal shales in this study is based on field observation, causing some uncertainty if all outcrops included in this facies fits with the definition of Schopf (1956). In this study coal shale is considered as dark, organic rich shale with abundant coal fragments, while coal appears as continuous coal seams.

To form coal two requirements must be satisfied: 1) The clastic sedimentation rate must be low compared to organic matter supply; 2) accumulation of organic matter must be higher than the degradation rate (Talbot and Allen 1996). Coal can originate in a number of sub-environments (Fig. 6.20), but swamps and mires are the most common coal forming environment and is recognized by vegetation growing in anoxic, waterlogged ground (McCabe 1984, Retallack 1991).

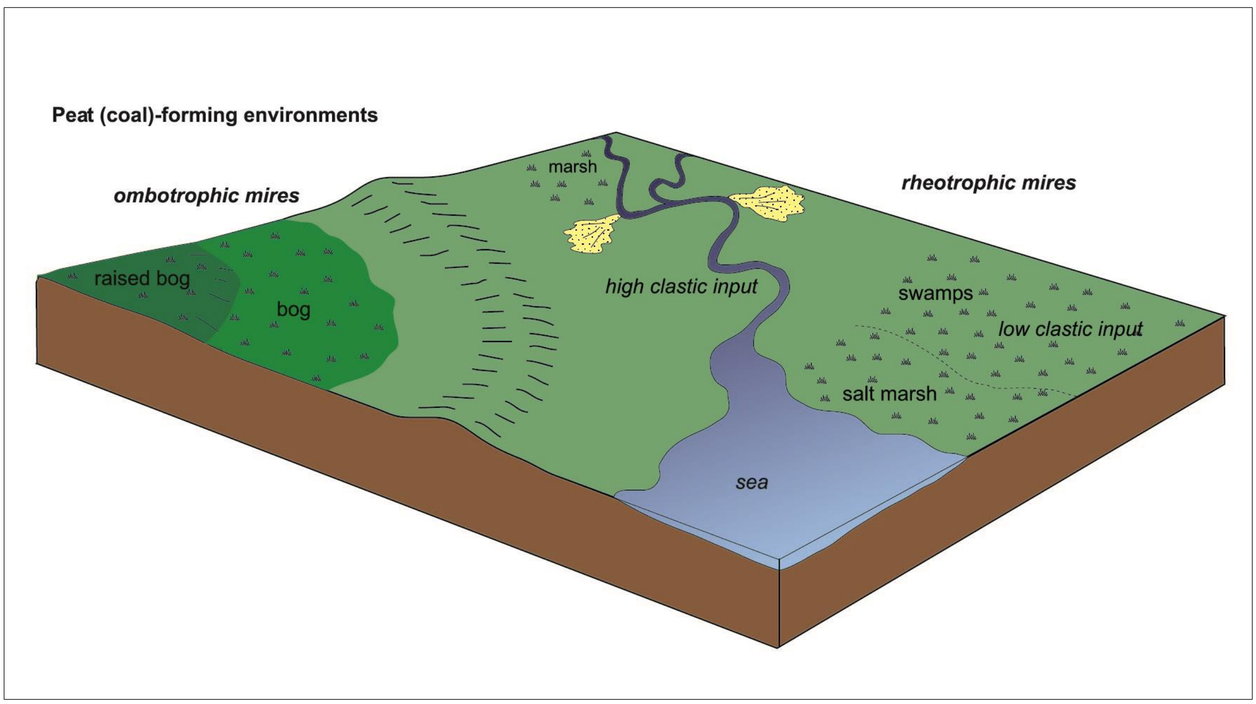


Figure 6.20: Coal-forming environments (Nichols 2009) recognized by vegetation growing under anoxic conditions in waterlogged ground.

Most of the coal seams found in the De Geerdalen Formation is overlying palaeosols, indicating that they are formed in place (histic epipedons) (Retallack 1991). Some of the coal and coal shales are lacking evidence of pedogenetic processes below. This could originate from rafted debris, but because very little coal form that way (Retallack 1991) it is more likely that this is due to poorly developed palaeosols. This is substantiated by the fact that long geological time with stable conditions that favour coal forming is required to form thick coal layers (Boggs 2011). The coal and coal shale found in the De Geerdalen Formation are thin and discontinuous. Enga (2015) suggest that lack of thick coal layers in the De Geerdalen Formation is due to seasonal changes in precipitation or fluctuation in sediment input preventing stable waterlogged ground and thus the formation of thick coal layers.

Because peats and mires require wet environments to form, they are usually found in temperate higher latitudes or in the wetter climatic belts around equator (Nichols 2009). This is consistent with the interpretation of a humid paleoclimate with seasonal variations in precipitation (Enga 2015). Coal and coal shales are here interpreted to originate from mires on a dynamic delta plain setting in a humid palaeoclimate with seasonal variations in precipitation (Hochuli and Vigran, 2010, Ryseth 2014, Enga 2015).

6.15 Facies O - Palaeosols

6.15.1 Sub-facies O_1 - Brown and yellow palaeosol

Description

This palaeosol type is found at every locality visited except Krefftberget and Deltaneset. It is most common in the middle and upper parts of the De Geerdalen Formation under and in the Isfjorden Member. The thickness is in the range of 0.2 to 1.0 meters. Roots are found on Blanknuten and Agardhbukta. On Wilhelmøya wood fragments up to 20 cm were found within the palaeosol. The colour varies from brown to reddish brown and yellow (Fig. 6.21). At some of the outcrops a 10 to 50 cm thick bleached yellow layer occurs above a brownish base (Fig. 6.21).

The palaeosols occur both in grey mudstone and on top of sandstone beds. A gradual contact at the base and sharper contact at the top is typical. The palaeosols are commonly overlain by coal or coal shale (facies N). Thin coal seams from 1 to 5 cm sandwiched in the palaeosol are common.

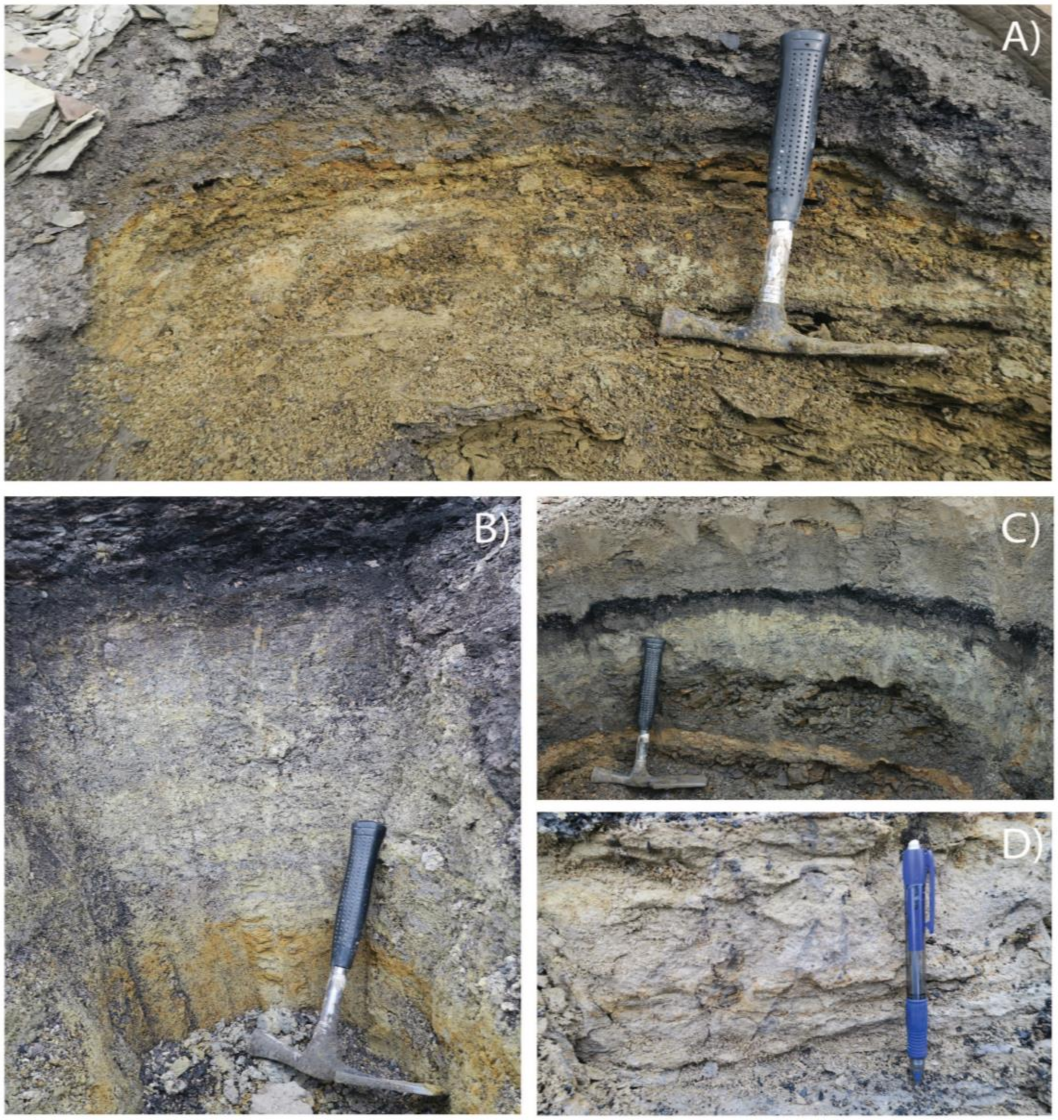


Figure 6.21: *Sub-facies O_1 - Brown and yellow palaeosol.* **A)** Yellow, coarse grained palaeosol, according to the paleocatena model typical for proximal channel palaeosols (Kraus and Aslan 1999). The outcrop is overlying distributary channel deposits of Blanknuten, Edgeøya. **B)** Possible top of channel palaeosol with coal shale (O-horizon) grading into a bleached layer (E horizon) and brown base (B horizon), Tumlingodden, Wilhelmøya. **C)** Coal shale overlying bleached layer containing coalified roots. The roots are penetrating the upper reaches of the underlying dark palaeosol horizon, Šmidtberget, Agardhbukta. **D)** Palaeosol with rootlets, Blanknuten, Edgeøya.

6.15.2 Sub-facies O_2 -Alternating red and green mudstones

Description

Alternating units of red and green mudstones (Fig. 6.22) is one of the most characteristic features in the Isfjorden Member (Pchelina 1983, Mørk et al. 1999). The Sub-facies is found at all localities where the Isfjorden Member is exposed, except Hellwaldfjellet probably due to scree cover. The unit thickness is from 0.5 to 5 meters, with individual red and green beds ranging from 0.2 to 1 meter. Contacts are relatively sharp, but often undulating. The facies often has a distinct mottled feature and peds. Spherical nodules with diameters ranging from 1 to 10 cm are seen to weather out from discrete layers. Alternating red and green mudstones are often found above sandstone beds. The red and green palaeosols at Deltaneset have thick calcareous sections. This is opposite most of the rest of De Geerdalen Formation, even though discrete layers of red and green mudstones at Friedrichfjellet, Šmidtberget and Teistberget are calcareous.

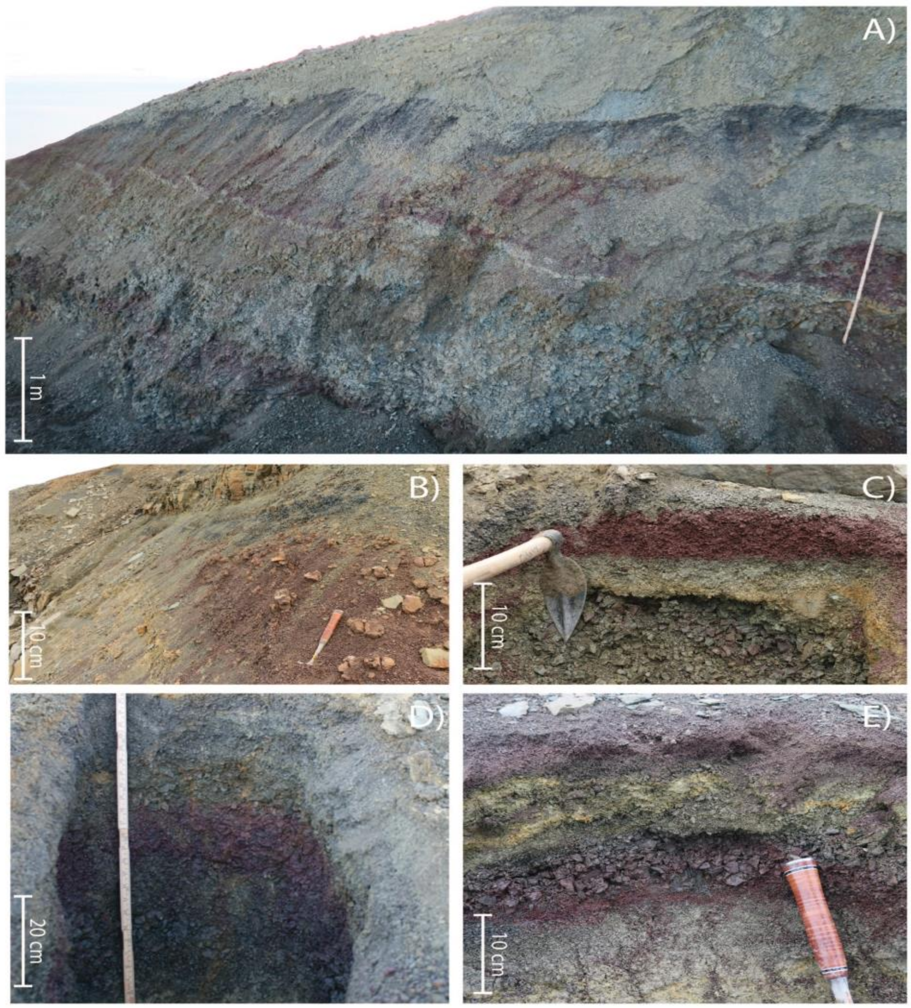


Figure 6.22: *Facies O_2 - Alternating red and green mudstone. All examples seen in this picture are noncalcareous.* **A)** Overview photo of the Isfjorden Member on the slopes above the dolerite sill on Tumlingodden, Wilhelmøya **B)** Alternating red and green mudstones with larger nodules at discrete levels on top of Klement'evfjellet, Agardhbukta **C)** Close-up view of the uppermost red and green layers, Šmidtberget, Agardhbukta. Note the presence of peds in the red shale. **D)** Red and green mudstones on Friedrichfjellet, Agardhbukta. **E)** Red mudstone with medium angular blocky structure in red shale, Friedrichfjellet, Agardhbukta.

Interpretation

Palaeosol is a weathering product formed due to physical, biological and chemical modification during periods of subaerial exposures (Boggs 2011). Five factors are important for the soil characteristics when palaeosols are formed: climate, topographic location, parent material and available time for forming of the palaeosol (Retallack 1997).

Palaeosols are thus not actual deposits, but instead formed by the alteration and maturation of ancient soils. Most palaeosols are found in continental environments (Boggs 2011), but can also form in marine strata following a relative sea level fall and subsequent subaerial exposure (Webb 1994). Palaeosols normally represent an unconformity, because they form in periods where the landscape is degrading or stable, followed by deposition of sediments (Kraus 1999).

Palaeosols are herein interpreted to have formed in a paralic and deltaic depositional setting following Enga (2015). Sub-environments are top of channel and barrier bar sandstones, floodplains and interdistributary, mud dominated areas. Restricted to possibly lagoonal environments have earlier been suggested for the Isfjorden Member (Mørk 1999, Mørk 2015).

The brown and yellow palaeosols (Facies O_1) are subdivided in Protosols, Argillisols and Vertisols. Alternating red and green mudstones (Facies O_2) are subdivided in Noncalcareous red and green mudstone and Calcrete. This is together with other field observations and interpretations of palaeosols further outlined in chapter 8.

6.16 Ichnofacies in the De Geerdalen Formation

Ichnology, the study of trace fossils, are considered a useful complementary indicator of palaeoecological conditions (Pemberton et al. 1992, Boggs 2011) and may provide unique information about depositional environments. Traces are indicative of animal behaviour and are influenced by a number of processes, such as sedimentation rate, substrate consistency, water turbidity, dissolved-oxygen content of the water, and salinity (Gingras and MacEachern 2012). Although trace fossils are commonly used in environmental interpretations, they do not give a certain estimate of palaeo-bathymetry, as different trace fossils can occur in a wide range of depositional environments at different scales (Fig. 6.23) (Pemberton et al. 1992, Boggs 2011).

Trace fossils are most easily recognized on the side of sandstone beds, but are also observed on top surfaces of beds. Trace fossils are observed to occur independently of facies and are thus here treated only as supporting observations in interpretations. Trace fossils and trace fossil assemblages from the Middle Triassic of Svalbard are described and discussed in Mørk and Bromley (2008). Some of these trace fossils are also observed to occur within the Upper Triassic succession on central and eastern Svalbard (Rød et al. 2014) and also documented here.

6.16.1 Cruziana ichnofacies

The *Cruziana* ichnofacies comprise assemblages of trace fossils commonly found at deeper waters below fair weather wave base and above storm wave base (Boggs 2011). Of the *Cruziana* ichnofacies, only *Rhizocorallium* was observed on eastern Svalbard. *Rhizocorallium* is a horizontal to slightly inclined trace that consisting U-shaped burrow with spreite structure.

6.16.2 Skolithos ichnofacies

Within the De Geerdalen Formation on eastern Svalbard, trace fossils such as *Skolithos* and *Diplocraterion* were observed within several facies (Figs. 6.23, 6.24A, 6.24D). *Skolithos* is a term applied for the trace fossil consisting of straight vertical, tube-shaped cylindrical burrows, commonly found in soft ground sand substrates (Boggs 2011). They are unbranched, most often sediment-filled and are often found perpendicular to the bedding plane (Mørk and Bromley, 2008). It is considered indicative of shallow marine, high-energy environments close to the shoreline, such as the lower, middle and upper shoreface (Fig. 6.23), and in beach environments (Pemberton et al. 1992, Boggs 2011).

On Svartnosa (Svart 15-1 ~ 106 m), abundant *Skolithos* was observed concentrated within a single sandstone bed (Fig. 6.24A). Sandstones with a high abundance of vertical tubes of *Skolithos* are commonly referred to as piperock (Droser 1991). Piperock commonly occur in a wide range of marine environments and are thus not restricted to shallow marine environments, such as beaches and intertidal settings (Droser 1991). It can also be present in deep marine settings, e.g. submarine canyons and deep sea fans (Pemberton et al. 1992).

Diplocraterion consists of vertical U-shaped burrows with spreite (Figs. 6.24G and H). It belongs to the *Skolithos* ichnofacies and is also commonly found in environments characterized by high energy waves and currents. The *Skolithos* ichnofacies commonly grade laterally seawards into the *Cruziana* ichnofacies and mixed *Skolithos-Cruziana* ichnofacies are known from both recent and ancient settings (Pemberton et al. 1992, Dalrymple and Choi 2007).

In the study area, a sparse and low diversity of trace fossil assemblages is observed, except on Wilhelmøya where observations were made of *Rhizocorallium*, *Skolithos*, *Teichichnus* and *Diplocraterion* (Fig. 6.24C). Many places, the only trace fossils observed are *Skolithos* and *Diplocraterion* and may indicate the presence of brackish-water conditions (Ichaso and Dalrymple 2009). Trace fossils are generally sparse in subtidal environments, except for vertical tubes of *Skolithos* (Dalrymple et al. 2012).

In summary the trace fossils observed in the De Geerdalen Formation are as expected to find in a delta front to shoreface setting (Hampson and Howell 2005).

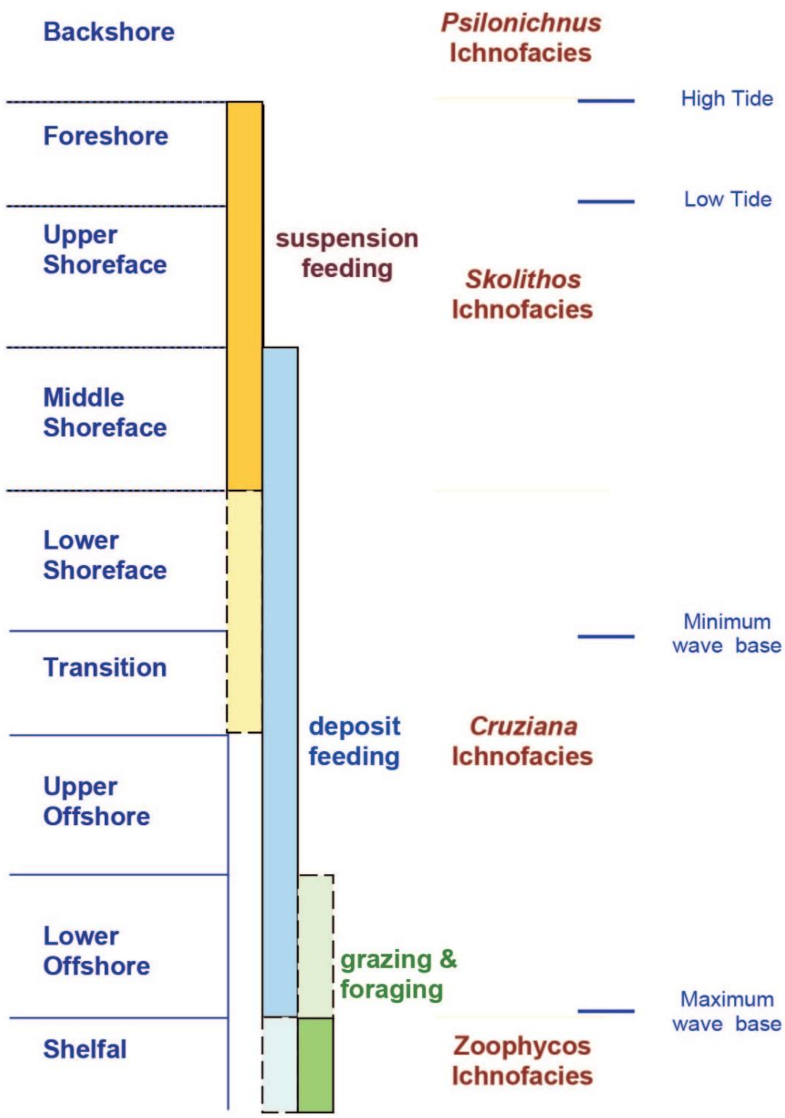


Figure 6.23: Distribution and types of ichnofacies on the shoreface. Sandstones in the De Geerdalen Formation exclusively contain trace fossils found in transitional offshore-shoreface environments, such as the *Cruziana* and *Skolithos* ichnofacies (Pemberton et al. 1992, Clifton 2006).

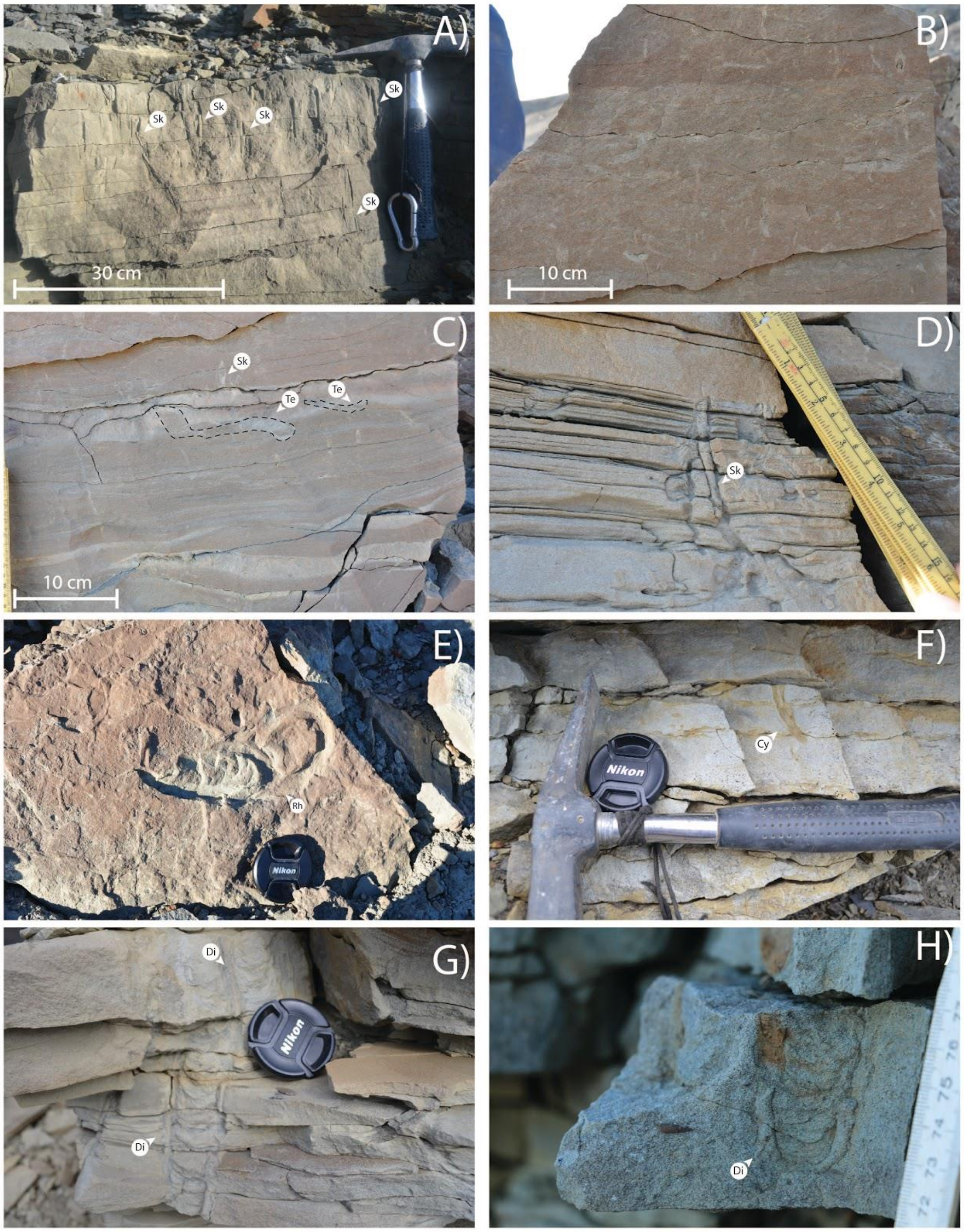


Figure 6.24: Trace fossils observed on eastern Svalbard. **A)** "*Skolithos* piperock" composed of vertical tubes of *Skolithos* (Sk), Svartnosa (Barentsøya). **B)** Bioturbated sandstone, Wilhelmøya. **C)** *Teichichnus* (Te) and *Skolithos* (Sk) in the upper part of De Geerdalen Formation on Wilhelmøya below the dolerite sill. **D)** *Skolithos*, Wilhelmøya **E)** *Rhizocorallium* (Rh), Wilhelmøya. **F)** Slightly "J-shaped" infilled tube interpreted as *Cylindrichnus* (Cy)**,** Šmidtberget, Agardhbukta. **G)** *Diplocraterion* (Di), Hahnfjella, eastern Spitsbergen. **H)** *Diplocraterion*, Klement'evfjellet, Agardhbukta.

7. Facies associations

Facies are commonly collected together into a facies association (FA), which is “a group of facies genetically related to one another and which have some environmental significance" (Collinson 1969 p. 207). Facies associations are thus inferred to be governed by a specific set of depositional controls and are more unique compared to individual facies (Reading and Levell 1996).

Facies may be stacked vertically in a preferred order, or interbedded randomly. By applying the principles of Walther's Law it is possible to predict which facies to expect when moving upwards or downwards in a vertical sequence (Reading and Levell, 1996). Eight facies associations are recognized in the study area in eastern Svalbard (Table 3) and identified based on interpretation of facies, geometries and dimensions of sandstones and other observations. Previous studies in adjacent areas, conducted by Rød et al. (2014), have served as a fundament and inspired the studies presented below.

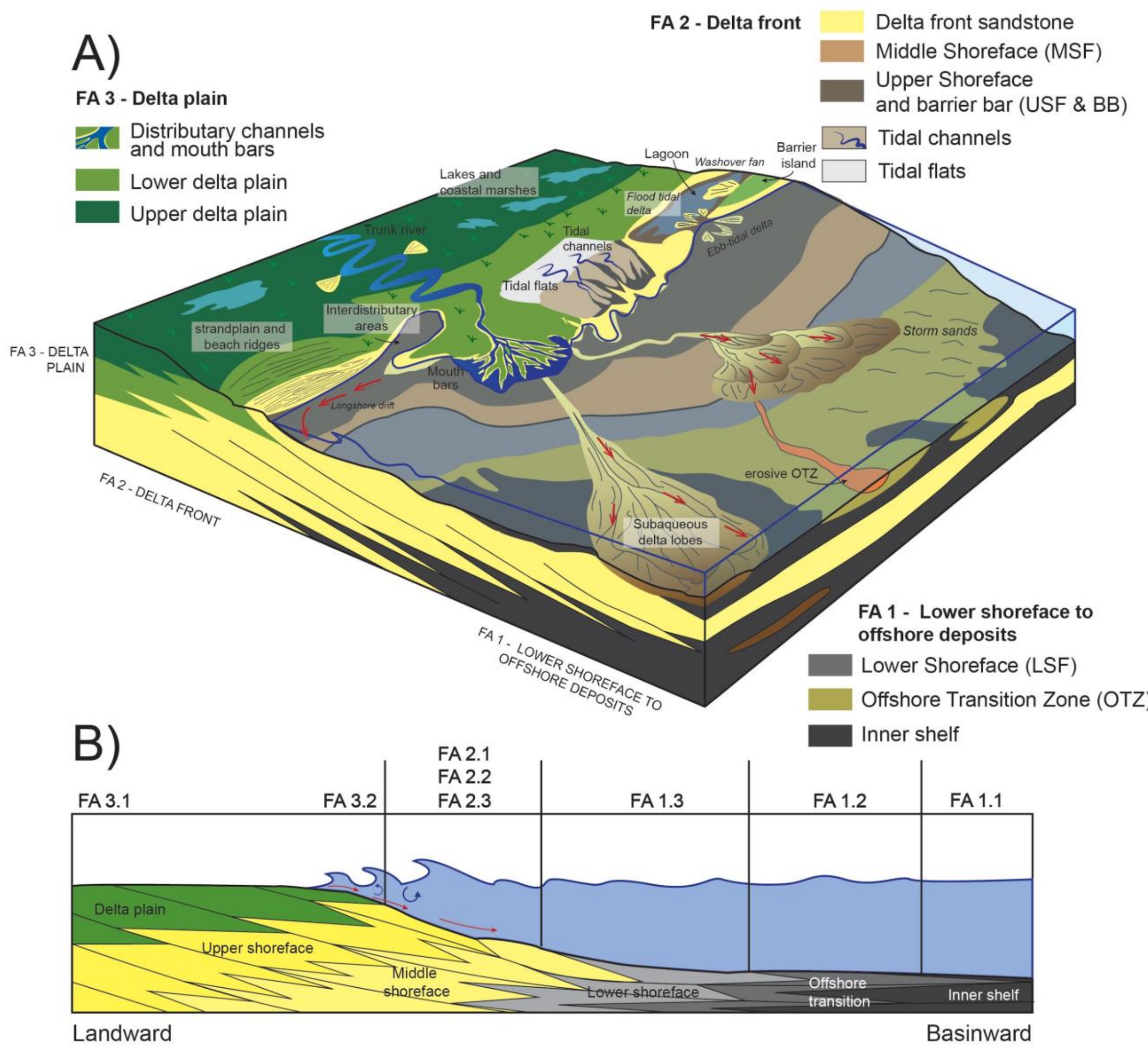


Figure 7.1: A) Conceptual block diagram showing the various environments and sub-environments occurring in the De Geerdalen Formation. Fluvial, wave and tidal processes are important factors in governing coastline morphology. Tidal processes generate tidal flats and channels and in combination with waves are important processes in barrier island complexes. Strandplains and beach ridges are mostly the product of wave energy and supplied by longshore drift. Note that the subaqueous delta lobes are not deep water fans, but results from the deposition from buoyant plumes or weak underflows off distributary mouths (drawn based on concepts from Bhattacharya and Walker 1992, Reinson 1992, Howell et al. 2008, Rød et al. 2014). **B)** Profile of a prograding shallow marine shoreline based on the above block diagram.

Table 3: Facies associations in the De Geerdalen Formation on eastern Svalbard. Parenthesis indicates less prominent facies (modified from Rød et al. 2014).

Facies Association	Sub-facies association	Facies included	Characteristics
1. Lower shoreface to offshore deposits	1.1 Offshore	I, K, M	Long intervals of mudrocks interrupted by thin tabular sandstones.
	1.2 Offshore transition	D, H, I, K, M	Storm dominated sandstones interacting with mudrocks deposited from suspension.
	1.3 Lower shoreface	(B), D, F, H, K, M	Fine fairweather sands reworked by storm events.
2. Delta front	2.1 Barrier bar and shoreface deposits	A, (B), D, E, F	Wave dominated upper shoreface to foreshore sandstones.
	2.2 Distributary mouth bar	A, B, C, D, E, F, I	Fluvial dominated very fine to fine sandstones showing influence of basinal processes.
	2.3 Distributary channel	A, B, (D), (G), K	Erosive-based fining upwards sandstones. Occur as laterally restricted or as laterally extensive, amalgamated channel deposits.
3. Delta plain	3.1 Floodplain with crevasse splay	(A), B, C, D, F, K, M, N, O	Delta plain deposits related to flooding of inter-distributary channels.
	3.2 Interdistributary areas	D, L, M, N, O	Shallow, quiet standing bodies of water with deposition of fine-grained material, coal, coal shales and palaeosols

7.1 Facies association 1 (FA 1) - marine offshore to lower shoreface deposits

Facies commonly found in offshore marine environments to lower shoreface are included in this facies association.

Generally sandstone beds decreases in bed thickness and wave ripples and planar stratification become less abundant as the water depth is gradually becoming deeper and deeper. Meanwhile, the facies become muddier and are usually more intensely bioturbated. Sedimentary structures in sandstones are as described in Johnson and Baldwin (1996) dominantly characterized by swaley and hummocky cross-stratification.

7.1.1 Offshore (FA 1.1)

Description

The offshore zone consists of facies representing the most distal portions of the delta. The facies association consists chiefly of mudrocks (facies M), that in intervals are interbedded with thin tabular sandstone bodies (facies K). Laterally restricted and soft sediment deformed sandstone lenses (facies I_2) are seen to be capsuled in mudrocks on Muen (Fig. 7.1), and makes up the coarsest fraction of grain sizes seen in this facies association. Plant fragments are also found in sandstones of facies I_2.

Sediments with an offshore marine origin are mainly seen in the lower parts of the formation, for example on Barentsøya, Edgeøya and in the Agardhbukta exposures. The dominant lithology, mudrocks of facies M, is susceptible to erosion and often found to be covered by scree.

Interpretation

The offshore zone is defined as shelf areas below mean storm weather base and is the site of deposition of fine mud and silt settling from suspension (Bhattacharya 2006, Nichols 2006). Even though it is dominated by fine-grained material, sand is also brought from the shoreline by density flows, waves and tides (Reading and Collinson 1996, Myrow et al. 2008, Boggs 2011). Bioturbation occur, and may be locally intense, but usually the offshore zone has less diversity and abundance of organisms compared to the offshore transition zone and the shoreface. The zone is often poorly oxygenated resulting in grey colour on the sediments due to partly preservation of some organic matter in the mud (Nichols 2009)

The lower boundary of the De Geerdalen Formation defined as the first prominent sandstone bed (Mørk et al. 1999) may imply that the boundary between the underlying Tschermakfjellet Formation and the De Geerdalen Formation is somewhere in the transition between the offshore zone and offshore transition zone. Lock et al (1978) mention striking variations in thickness of the Tschermakfjellet Formation. However, small-scale fluctuations in relative sea level may have moved the boundary between offshore and offshore transition zone back and forth (Nichols 2009). For example are the hummocky cross-stratified sandstones of the Muen locality interrupted by intervals of up to 20 meters of mudrock that possibly belongs to the distal offshore zone. The shales may also represent periods of fair weather. Fair weather deposits in the offshore transition zone and offshore deposits are both settling from suspension (Nichols 2009) and are thus similar in both grain size and structures. But as the sand to mud ratio tends to increase landwards it can be assumed that long intervals of mudrock belongs to the offshore zone (e.g. Muen log 194 - 202 m and 208 - 218 m).

The term pro-delta is different from the offshore zone, in the way that the pro-delta comprise a smaller area outside the delta front and is only used when a direct influence of deltaic processes is observed. The Tschermakfjellet Formation has traditionally been interpreted as the pro-delta for the south-eastern sediment source in the De Geerdalen Formation (Mørk et al. 1982, Mørk et al. 1999, Riis et al. 2008). The offshore zone is therefore here defined as the low-energy inner shelf environment characterized by mud-dominated deposits.

Figure 7.1: Facies association 1.1: Offshore deposits. Log from Muen starts at the base of the De Geerdalen Formation with attached outcrop photo from the upper parts. Note how the erosive sandstone lens (facies I_2) is capsuled in mudrocks (facies M) and creates a small topographic plateau. Geologist for scale.

7.1.2 Offshore transition (FA 1.2)

Description

Wave rippled sandstone (facies D), hummocky and swaley cross stratification (facies H), soft sediment deformed sandstones (facies I) and heterolithic bedding (facies K) are commonly found within the offshore transition zone. The facies association is generally coarsening upwards with increasing thickness of sandstone beds, attributed to increased wave activity and shallower waters due to progradation. Mudrocks are often bioturbated. This trend is seen on Klement'evfjellet and Muen. The FA is typically underlain by offshore muds (FA 1.1) and overlain by lower shoreface deposits (FA 1.3). However, fluctuations in sea level and subsequent erosion and sediment bypass may have reduced the abundance of offshore transition zone deposits on some localities (e.g. Muen).

Interpretation

The offshore transition zone extends from the boundary of the offshore zone at mean storm weather wave base up to mean fair weather wave base. It is more sand-rich compared to the offshore zone and is dominated by alternating energy conditions (Reading and Collinson 1996, Eide et al. 2015).

Storms are the main controller on sediment transport in the offshore transition zone. During fairweather conditions, fine sediments are deposited from suspension, while sand is transported and deposited during storms (Eide et al. 2015). Storm events generally erode the coast resulting in the redistribution of sand in the offshore transition zone. The amount of sand is typically higher in proximal positions compared to more distal areas. Bioturbation tends to decrease from proximal to distal setting, reflecting both the time of quiet conditions and number of organisms. Typical signatures of storm deposits are a basal lag of coarse sediments, hummocky cross stratification, wave rippled cross-lamination and burrowed intervals (Johnson and Baldwin 1996).

Wave ripples and planar parallel stratification occur more abundantly in the proximal areas of the offshore transition zone, compared to the most distal parts (Fig. 7.2). The upwards increase in sand content seen in the outcrops in Agardhbukta is interpreted as a gradual transition from distal to more proximal setting within the offshore transition zone.

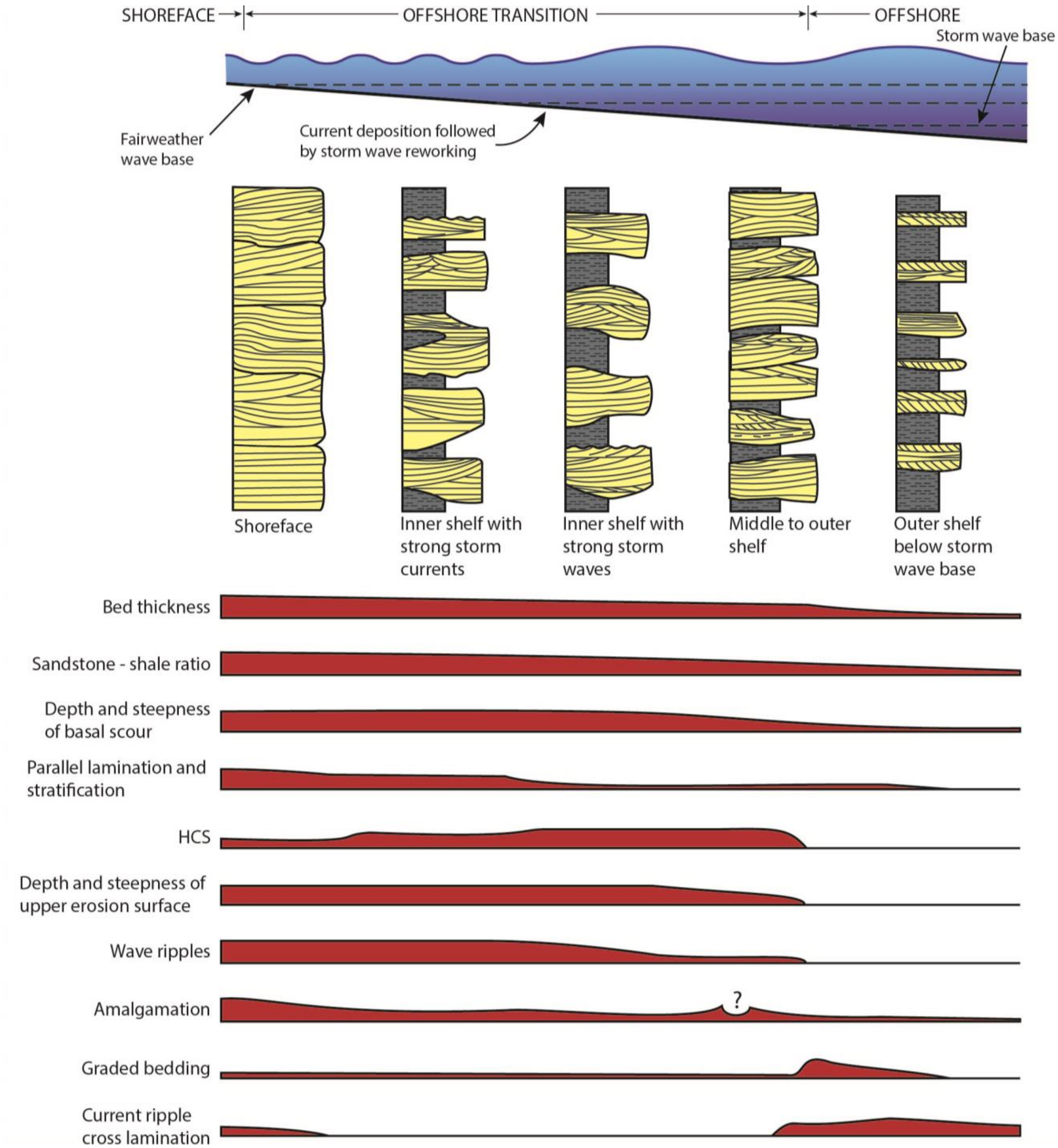


Figure 7.2: Shifts in facies occur in storm deposits as water depth and distance from the shoreface increases (Brenchley 1985, Miall 2000)

7.1.3 Lower shoreface (FA 1.3)

Description

The lower shoreface sees a decreasing mud content from the silty to very fine hummocky cross-stratified sandstones (facies H) of the offshore transition zone into rippled and planar bedded very fine sandstones (facies B, D and F) in the upper parts. Here, wave rippled sandstones (facies D), horizontally bedded sandstones (facies F) and low angle cross-stratified sandstone (facies E), are seen to interfinger although top surfaces typically show eroded wave crests. Wave ripple troughs are commonly mud draped and slight bioturbation is noticed. Sandstones are often calcite cemented, and occasionally show cone-in-cone structures as observed on Muen. Deposits belonging to this sub-facies association commonly terminate upwards coarsening parasequences assigned to the FA 1 and are mostly seen in the lower part of the De Geerdalen Formation.

Interpretation

The lower shoreface deposits comprise the sandy upper part of FA 1, and are reworked by oscillatory currents under fairweather conditions (Clifton 2006). The transition from sand to mud is in most models for clastic shorelines defined as the base of the shoreface (Clifton 2006). Wave ripples shallow into unidirectional ripples, but are reworked by storm events (Reading and Collinson 1996). Interbedded mud is indicative of alternating energy conditions (Davis 2012), and could possibly be assigned to tidal influence or periods of calm fairweather.

7.2 Facies association 2 (FA 2) - Delta front

The delta front is characterized by a relatively steep delta slope where interaction between fluvial and basinal processes constitutes the depositional framework acting upon the deltaically introduced sediments (Reading and Collinson 1996). Progradational deltaic sequences shallow and coarsen upwards from mud dominated lower shoreface to offshore deposits (FA 1) into sand-dominated facies characterizing the delta front (FA 2).

On the background of field-observations three sub-facies associations have been defined on the delta front. These sub-facies associations comprise deposits roughly from lower shoreface to foreshore and marks a substantial energy increase to the underlying FA 1. Furthermore, they are thought to represent sub-environments, within the delta system, influenced differentially by basinal and fluvial processes.

7.2.1 Barrier bar and shoreface deposits (FA 2.1)

Description

The barrier bar and shoreface facies association is found above FA 1 as coarsening upward units from silt and mud to fine sandstones characterized by structures created predominantly by oscillatory currents. It is separated from mouth bars and distributary channel deposits by less fluvial influence and distinct wave dominance upon sedimentation, although some tidal influence is recognized upon these deposits. The thickness of the FA is in the range of 2-5 meters.

Upwards increasing sand content reflects higher energy environments approaching fairweather wave base. Low angle cross-stratification (facies E), horizontal stratification (facies F) and wave rippled facies (facies D) are found in the upper reaches (Fig.7.3A). Commonly mudrocks (facies M) sharply overlie the coastal sandstones. The boundary is often erosive but sometimes wave ripple crests are preserved. Occasionally mudrocks, rootlets and coal (facies M) are preserved in the upper reaches, possibly representing lagoonal facies.

Fine to medium sand is found in large-scale trough cross-bedded intervals (facies A) in the upper part of parasequences (Fig. 7.3B). Small-scale asymmetric cross-bedded sandstones (facies B) are also found in this interval, but compose finer sand fractions with minor inclusions of intercalated mud. Tidal signatures as mud draped foresets and double mud drapes are observed in the lower part of measured sections in Agardhbukta.

A lower degree of calcite cementation is noticed compared to other sandy facies associations within the delta front environment. *Skolithos* and *Diplocraterion* are common trace fossils found within this facies association. Recurrent barrier bars terminate stacked parasequences in the lower part of measured sections. The facies association is found throughout the study area e.g. in the lower parts of measured sections in Agardhbukta and in the middle part of the Svartnosa section.

Interpretation

The barrier bar and shoreface facies association follows the basic model for open-coast clastic deposits (e.g. Clifton 2006) exhibiting upwards-shallowing succession of sand overlying distal marine offshore to lower shoreface deposits (FA 1) and underlying proximal nonmarine facies. However, the proximal nonmarine facies are commonly not preserved and erosional features in the uppermost beds are interpreted as transgressive surfaces.

The distribution of observed facies found within this facies association is governed by the zonation of the shoreline profile (Reading and Collinson 1996). On the upper shoreface fairweather waves set up longshore and onshore currents leading to migration of bars and current ripples recorded in the sedimentary record as large-scale and small-scale cross-bedded sandstones (facies A and facies B)(Fig. 7.3B). Superimposed low-angle cross-bedded sandstones (facies E) and horizontally bedded sandstones (facies F) are interpreted to record swash-backwash processes by breaking waves on the foreshore (McCubbin 1982). Foreshore deposits aggrade under fairweather deposition, but are reworked by storms and during transgressions (Clifton 2006). Distribution of the zones of the shoreline profile is largely controlled by intensity of wave energy and nature, while a tidal influence produce gradual transitions and overlap as the location of mean storm wave base and fairweather base is transient (Reading and Collinson 1996).

The type of coast depends on controls imposed by; (i) relative power of waves, tides and fluvial source; ii) sediment grain size; (iii) marine sediment supply; (iv) relative sea-level change (Reading and Collinson 1996). These controls results in a variety of clastic coasts where each-ridge strandplains, chenier plains and mudflats in general have been attributed to regressive systems, while barrier island-lagoonal systems and estuaries commonly is associated with transgressive systems (Boyd et al. 1992, Dalrymple et al. 1992, Reading and Collinson 1996). Distinguishing these sub-environments is difficult based solely on 1D data, lacking understanding on large scale geometries. Presence of lagoonal facies above sandstones indicates presence of a barrier island complex, but frequent transgressions, wave reworking and scree cover may mask such indicators used to infer about and separate sub-environments such as barrier islands and strandplains.

Barrier islands or barrier spits constitute about 15% of the seaward margin present on modern coastlines (Glaeser 1978). Modern barriers are accumulations of sand that slowly migrate landwards accompanied by rising sea levels (Clifton 2006). Wave processes dominate as the sandy barrier shelters shallow shore-parallel lagoonal waters (Reading and Collinson 1996). Depending on the tidal regime, tidal inlets and washover fans rework the upper portion of the beach face. By increasing dominance of wave relative to tidal component tidal inlets tend to migrate producing tabular extensive alongshore sand bodies (Reading and Collinson 1996). This lateral migration and reworking of the barrier island show high preservation potential and may dominate the depositional record of the barrier island (Reading and Collinson 1996). Barrier sandbodies are generally characterized by linear geometries, while strand plains commonly have sheet-like sandbodies (Clifton 2006).

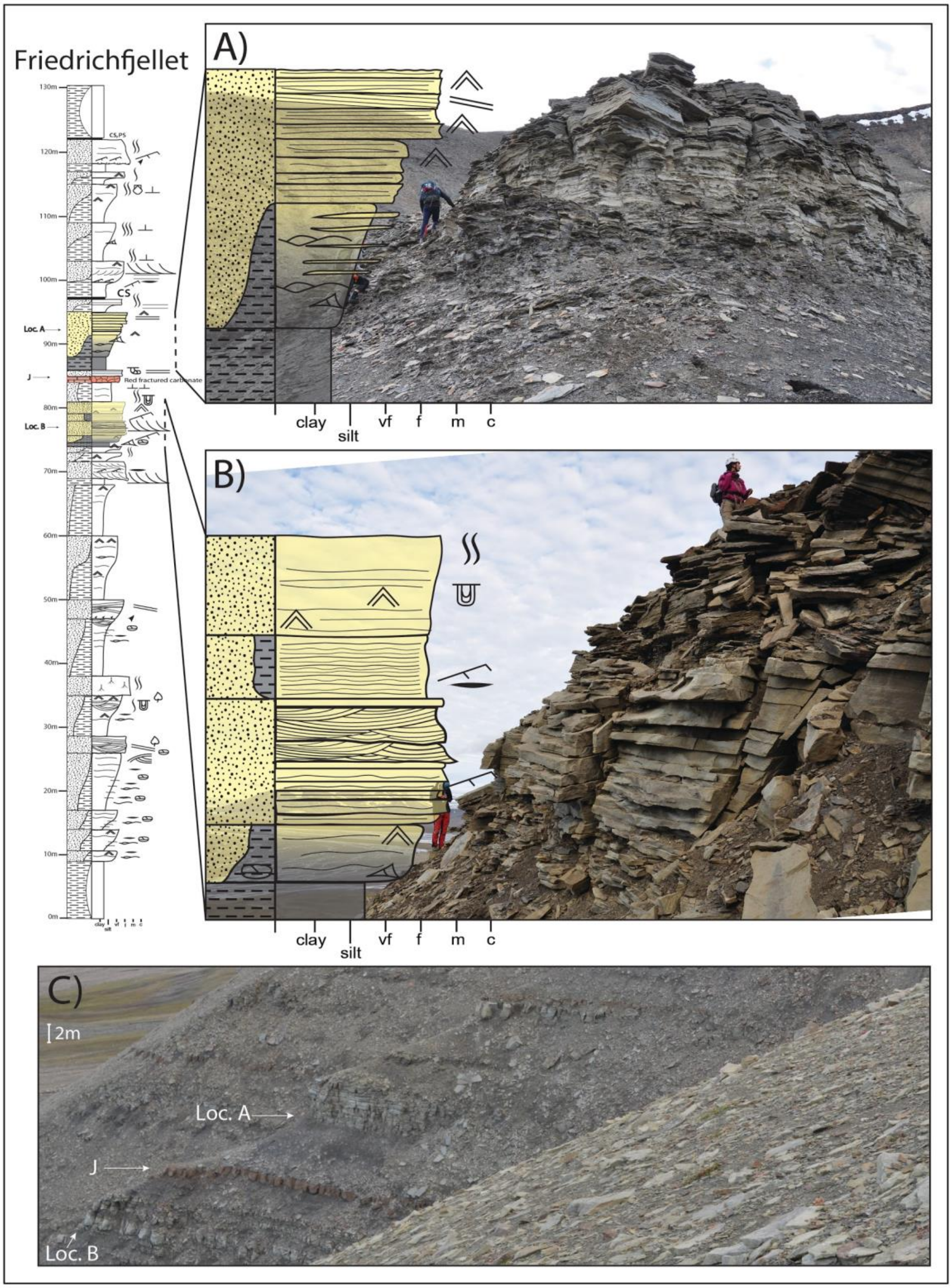


Figure 7.3: Lower sections logged on Friedrichfjellet. **A)** Coarsening and shallowing upward sequence with facies indicating prevalent oscillating flow, interpreted as a prograding barrier succession. **B)** Interpreted barrier island sequence with large-scale cross stratified interval possibly remnant of tidal inlet or longshore bar migration. **C)** Overview picture. Note lateral continuity of facies J.

7.2.2 Distributary mouth bars (FA 2.2)

Description

Deposits of the distributary mouth bar facies association comprise fine to medium sandstones arranged in a coarsening upwards sequence overlying mud and storm dominated facies of FA 1. Abundant soft sediment deformation in underlying shale and sandstones and climbing ripple stratified sandstones (facies C) are typically observed. Large-scale and small-scale cross-stratified sandstones (facies A and B) are common, especially in the lower parts, but often lack the characteristic erosive lower contact and mud flake conglomerate observed in the distributary channels facies association. Horizontally bedded sandstones (facies F) and low-angle cross-stratified sandstones (facies E) are found in the upper reaches, with occasional interbedded wave rippled sandstones (facies D). Abundant plant fragments and coal drapes are found in this facies association. Sparse bioturbation is observed.

Sandstone bodies are laterally extensive, 100s of meters, as observed in the field. However, field data regarding geometries from outcrops is sparse. The facies association is observed in the middle parts of the Svartnosa section and on Hellwaldfjellet.

Interpretation

A distributary mouth bar is formed near the seaward limit of the distributary channel as the expanding river flow decelerates and deposits a sandy shoal (Reading and Collinson 1996, Olariu and Bhattacharya 2006). Exceptionally rapid deposition rate is common for distributary mouth bars (Reineck and Singh 1980) and is illustrated by climbing ripple laminated sandstones and loading structures in the study area. Horizontally bedded, low-angle cross-stratified and wave rippled sandstones are found in the upper reaches, and interbedded in units, and indicate influence of basinal processes upon sedimentation. This marine affiliation is supported by marine trace fossils, where low abundance compared to FA 1 may be explained by rapid deposition. Abundant plant fragments and coal- or mud-draped foresets indicate a proximal terrestrial influence, while inclusions of heterolithic succession may be attributed to a tidal component or seasonal changes in river discharge (Reading and Collinson 1996, Dalrymple and Choi 2007).

Mouth bars are fundamental building blocks of prograding deltas and can accrete to complex bar assemblages and regional-scale lobes (Bhattacharya 2006). Ancient mouth bar sand bodies are shown to exhibit larger dimensions (Reynolds 1999) than their modern analogues (Tye 2004) and thus showing how migration and coalescence of modern bar forms created the greater ancient examples viewed in outcrops (Bhattacharya 2006). Delta progradation is mainly achieved by coalescence of downstream migrating mouth bars (Bhattacharya 2006). Width of distributary channels may vary spatially and temporally, roughly dictating the scale of the genetic related mouth bar. Size and shape of the mouth bar also depends on angle of plume dispersion, relative density of stream and basinal waters and processes (Bhattacharya 2006).

Waves straighten and elongate the mouth bar alongshore (Bhattacharya and Giosan 2003, Li et al. 2011), while tides may stabilize the distributary channel and the associated mouth bar resulting in high length-to-width ratios (Reynolds 1999). Down cutting of the associated distributary channel commonly erode the upper parts of mouth-bar sediments (Reading and Collinson 1996).

7.2.3 Distributary channels (FA 2.3)

Description

Distributary channels are typically seen as upward fining sandstone units with an erosive basal lag containing plant fragments and mud flakes. Lower reaches are dominated by trough cross-bedded intervals (facies A) while small-scale asymmetric cross-stratified sandstones (facies B) and wave rippled sandstones (facies D) are found in the upper reaches. Mud draped foresets are occasionally seen in trough cross-bedded intervals, but generally interbedded clay laminas (facies K) are found in the upper parts. Rootlets and palaeosols are found at the very top of sequences. Abundant plant fragments is characteristic, while bioturbation is almost absent in this facies association. Sandstones are often cemented by calcite.

Based on dimensions two types of distributary channels have been documented in our area; 1) laterally restricted channel sandstones displaying relatively modest dimensions (height around 2 to 3 meters, width around 10 m), 2) Amalgamated channel deposits showing extensive lateral continuity. The laterally restricted channel deposits are commonly found in the upper part of De Geerdalen Formation and in the Isfjorden Member in close proximity to floodplain deposits (FA 3). The amalgamated channel deposits are found in the middle part of the De Geerdalen Formation and consist of stacked co-sets, up to 80 cm thick, of trough- and planar cross-bedded sandstones (facies A) composing units up to 6 meters high as seen on Svartnosa (Fig. 7.4). Distributary channels are also found to overlie mouth bar deposits. Amalgamated deposits have been observed on Svartnosa.

Interpretation

Distributary channels are found on both the delta plain and on the delta front, but are here described under delta front facies association (FA 2). A distributary channel is a stream carrying sediments and water discharge from a trunk river into the sea (Olariu and Bhattacharya 2006). As the distributary channel merge with coastal waters on the delta front it becomes shallower, branches and loses its competence (Reineck and Singh 1980, Olariu and Bhattacharya 2006). Terminal distributary channels are common elements in river-dominated deltas and are the smallest channels on the distal delta plain and proximal delta front (Bhattacharya, 2006, Olariu and Bhattacharya 2006). They are closely related to distributary mouth bars (FA 2.1).

Distributary channels share many of the same features as fluvial trunk channels. Both are characterized by a predominant unidirectional flow interrupted by fluctuations in stream discharge. The base is typically erosive with a basal lag that gradually fines upwards from cross-bedded sand to ripple-laminated fine sand with alternating silt and clay. Observed rootlets or palaeosol on the top indicates abandonment of the channel (Reading and Collinson 1996). Distributary channels differ from alluvial channels in several ways. The lower, seaward end of distributary channels is influenced by basinal processes as tidal and wave processes rework the channels into mouth bars (Bhattacharya 2006). Hence mouth bars (FA 2.2) are more common on the distal delta front, while terminal distributary channels occur more frequently on the proximal delta front (Olariu and Bhattacharya 2006). Distributary channels are more prone to avulsion and switching than fluvial channels due to lower slope gradient. The width to depth ratio is also smaller for distributary channels because of the relatively short lifetime and therefore limited time to migrate laterally (Reading and Collinson 1996).

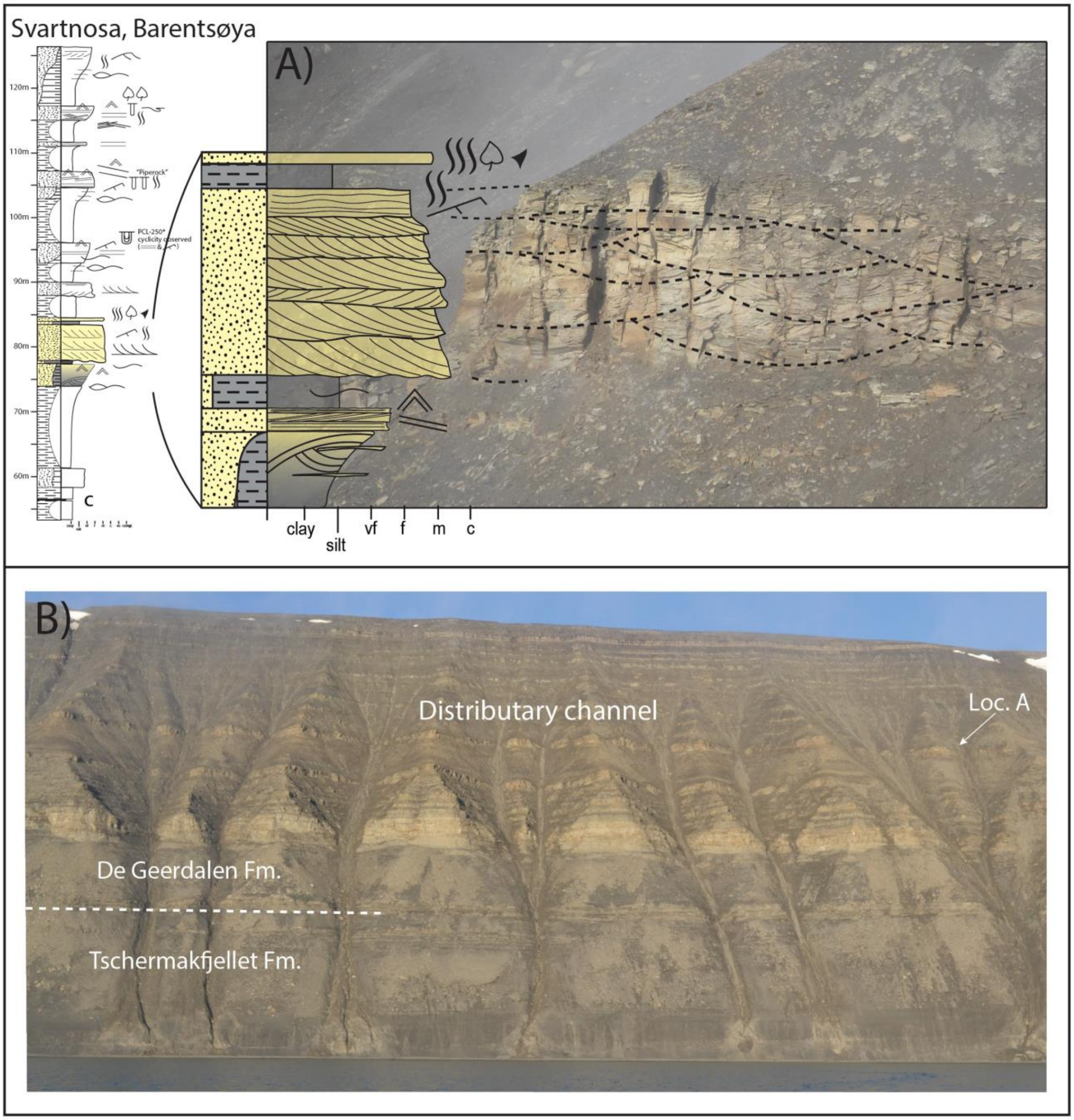


Figure 7.4: Svartnosa, Barentsøya. **A)** Amalgamated distributary channel deposits superimposed on lower shoreface deposits of FA 1. **B)** Overview photograph taken at sea level. Note the lateral continuity of sandstone bodies.

The outcrops shows many of the same features as fluvial deposits, but are interpreted as distributary channels because of lateral limitation and presence of tidal signatures. Observations of wave ripples and bioturbation on many of the outcrops support the interpretation of a marine influence on distributary channels (Reading and Collinson 1996).

The relatively modest dimensions characterizing the laterally restricted channels could possibly be explained by frequent switching and abandonment on the delta plain (Reading and Collinson 1996). The amalgamated deposits may represent periods of relatively stable base level, thereby allowing the extensive lateral migration observed at Svartnosa. Noteworthy is also this locality's position as the most proximal locality to a probable sediment source area in the south-southeast (Riis et al. 2008, Lundschien et al. 2014, Rød et al. 2014, Klausen et al. 2015)

7.3 Facies association 3 (FA 3) - Delta plain

Delta plain is commonly overlying facies association 2 - delta front, and is typically found in the uppermost part of the De Geerdalen Formation. The facies association is recognized by deposits where the marine influence is less prevalent and more closely connected to paralic and continental environments. The upper limit of modern delta plains are normally defined by the presence of distributary channels and the limit to the lower delta plain at the most landward extent of tidal influence (Bhattacharya 2006). Tidal signatures are therefore often found in sediments deposited on the lower delta plain (Bhattacharya 2006). The transition zone from fluvial to marine dominated environments in tidal dominated systems are one of the most complex depositional systems in the world due to the huge variability in terrestrial and marine processes interacting there (Bhattacharya 2006, Dalrymple 2007, Dalrymple and Choi 2007). The wide range of facies found in the FA reflects the dynamic nature of delta plains that are, at least partially, influenced by tidal processes

Delta plains are usually recognized by an assorted assemblage of sub-environments formed in brackish to non-marine conditions. Sub-environments include distributary channels, swamps, marshes, interdistributary bays, tidal flats and lagoons (Bhattacharya, 2006). Numerous active and inactive distributary channels are commonly found across the delta plain, separated by shallow water areas with little sedimentation and emergent areas (Reading and Collinson 1996). Distributary channels (FA 2.3) are described under delta front deposits (FA 2) and this chapter will not provide any further description.

The delta plain can be subdivided into lower delta plain and upper delta plain (Reading and Collinson 1996, Bhattacharya 2006). The upper delta plain is dominated by fluvial processes, and in many ways similar to alluvial environments**.** However, swamps, marshes and lakes are typically more extensive compared to alluvial environments (Reading and Collinson 1996). The upper limit of the upper delta plain is often defined at the point where the trunk river starts to be distributive (Bhattacharya 2006).

In addition to fluvial processes the lower delta plain is also often affected by basinal processes (Reading and Collinson 1996). Saline water and tide processes may penetrate the lower delta plain, but massive marine influence is inhibited by beach barrier shorelines or by a massive delta front in fair weather, although storms can cause marine water to penetrate several of kilometres inland (Reading and Collinson 1996). The limit of the lower subaerial delta plain is usually defined either at the high tide shoreline or low tide shoreline (Bhattacharya 2006).

7.3.1 Floodplain (FA 3.1)

Description

Floodplain deposits are found at all localities visited except Muen and Mistakodden. The FA is typically found close to distributary channels. They often display as mudrock (facies M) interrupted by horizontally bedded and wave- and current rippled sandstone beds (facies F, D and B). Climbing ripples (facies C) are found on Wilhelmøya. Palaeosols and coal and coal shales (facies O and N) are found both at top of distributary channels and on floodplain deposits. Plant fragments and bioturbation in sandstone beds are common. Extensive scree cover is typical, and is often inferred as mudrocks if no sign of coarser material is seen. The thickness of floodplain deposits is in the range of 2 - 15 meters.

Interpretation

Distributary channels described under facies association 2 (delta front) is together with floodplains one of the main signatures of the active parts on the delta plain (Bhattacharya 2006). Flood plains are strongly related to the distributary channels since they receive most of the sediments from distributary channels. The sediment load in most rivers contains as much as 85 - 95 % mud (Schumm 1972). The mud is primarily carried in suspension, and most of it is deposited in the channel itself, dams, in the delta front, and on floodplains related to fluctuations in water level. This leads to fine grained floodplain deposits (Bridge 2006, Bhattacharya 2006). Inundation is, however, not only a product of flooding of the river channels. Increased watertable level or high precipitation rate are also common water sources on floodplains (Collinson 1996).

The fine grained deposits on the floodplain may be interrupted by silts and sands from levees and crevasse splays. Levees are ridges that build up on both sides of channels. During floods the levees may break into a crevasse splay, leading to deposition of silts and sands in smaller lobes on the floodplain, becoming increasingly finer away from the channel. The lobes are often composed of material supplied by the former levees. Typical features in crevasse splays are cross-lamination and small-scale cross-bedding. Floodplains are commonly exposed during low water level, leading to pedogenetic processes. In humid conditions the floodplain sediments may stay wet, and if the conditions for plant growth are good, peat may accumulate (Collinson 1996).

Fine grained deposits alternating with thin ripple laminated sandstones displaying various degree of bioturbation close to interdistributary channels are interpreted as floodplain deposits. Climbing ripples (facies C) is interpreted as rapid deposition of sediments and could possibly be formed in crevasse splays. Palaeosols, coal and coal shales (facies N and O) interacting with floodplain deposits may represent periods with little sedimentation, possibly due to low frequency or magnitude of floodings.

Note that floodplains have many of the same characteristics as interdistributary areas, and the transition between the two settings is probably gradual, as parts of the floodplain distal to distributary channel are interdistributary. The dynamic nature of deltas may also have caused rapid shifts between floodplain and interdistributary areas, making the distinguishing between floodplains and interdistributary areas difficult. Nevertheless, one can assume a general trend of finer material away from distributary channels when floodplains gradually go into interdistributary areas.

7.3.2 Interdistributary areas (FA 3.2)

Description

Interdistributary and interlobe areas are herein defined as standing bodies of water such as lagoons and lakes, as well as marshes and swamps. The sub facies is recognized by less sand content compared to distributary areas and floodplain, and is dominated by mudrocks (facies M) interrupted by coal and coal shales (facies N) and palaeosols (facies O) (Fig. 7.5A). Facies successions are generally shallowing-upwards with a decreasing marine influence. Some outcrops contain thin layers of wave rippled or horizontal bedded sandstone (facies D and F).

Palaeosols (facies O) is found underlying coal or coal shales (facies N) and overlying mudrocks (facies M), distributary channels and barrier bars. On some of the outcrops the palaeosol is found overlying scree covered areas. Scree cover may be due to easily eroded fine grained material such as mudrocks (facies M).

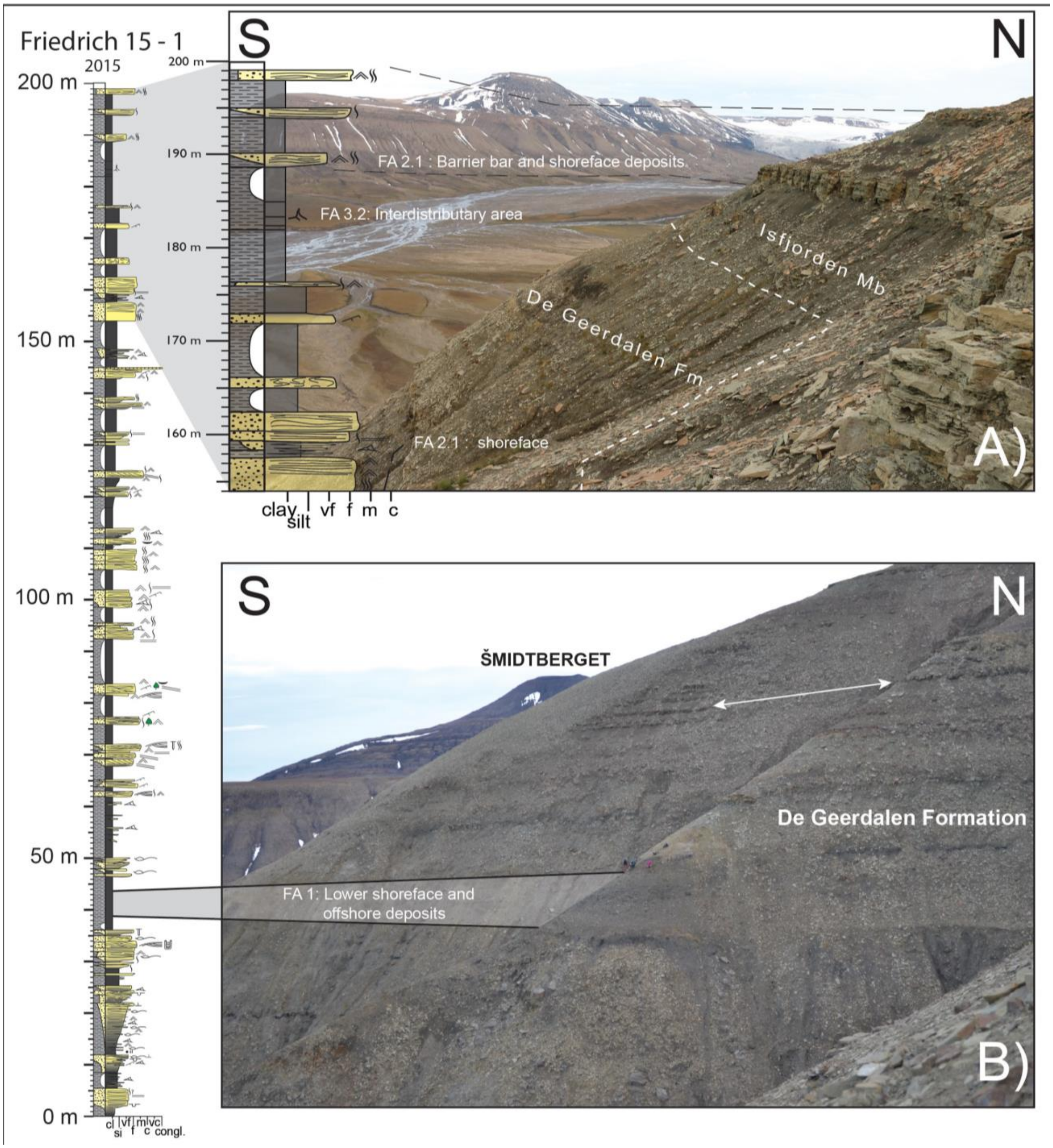


Figure 7.5: Mud-dominated facies associations on Friedrichfjellet, Agardhbukta **A)** Facies association 3.2: Interdistributary areas. The red and green shales of the Isfjorden Member are conformably overlying the De Geerdalen Formation **B)** Facies Association 1: Lower shoreface to offshore deposits.

Interpretation

Interdistributary areas constitute an important depositional element in deltaic settings. They can be either bounded by distributary channels or open to the sea. Facies in interdistributary areas are commonly less sandy compared to distributary environments and the facies successions are in general often seen as quite thin coarsening- or fining-upwards units (Bhattacharya and Walker 1992, Bhattacharya 2006).

Swamps and marshes are the main peat- and coal forming environment. Swamps are freshwater sourced wetlands that favour woody vegetation and are mainly located on the upper delta plain. Swamps gradually turns into fresh, brackish or saline marshes in the seaward direction. Peats formed in saline marshes on the lower delta plain tend to have high content of impurities from terrigenous matter and sulphur (Reading and Collinson 1996). Coal and coal shales found in the De Geerdalen Formation appear as impure, leading to the possible interpretation of a lower delta plain setting in saline marshes. Thin coal seams and apparently limited laterally continuity may support a relatively dynamic and unstable regime.

Waterbodies in upper delta plains are mainly lakes. Lower delta plains also often contains lakes, but can differentiate from upper delta plain by the presence of lagoons, estuaries and interdistributary bays. Lagoons are shallow waters roughly directed alongshore and protected by a barrier island (Reading and Collinson 1996). Restricted water circulation causes varying salinities. Lagoons commonly accumulate pervasively bioturbated fine grained sediments deposited from quiet waters (Reading and Collinson 1996). Bay-head deltas occurring on the landward margin of the lagoon generate small-scale facies sequences resembling those of fluvial-dominated deltas (Reading and Collinson 1996).

Interdistributary bays and their associated small-scale bay-fills frequently develop on river-dominated deltas, whereas wave-influenced deltas develop more uniform and straightened coastlines displaying more lateral continuity of correlative vertical sequences (Bhattacharya and Walker 1992, Reading and Collinson 1996).

In humid climates blanket bogs are formed in the margins of the lakes. In contrast, lake margins in arid climate tend to form exposed surfaces with calcretes, gypsum and halite precipitations. Common for the water bodies are shallow water depth and low energy, resulting in dominant deposition of fine materials such as clay, silt and fine sand. However, floods and diversion of distributary channels disturbs the regime and brings in coarser material resulting in a variety of features formed from crevasse splays, crevasse channels, levees and small deltas (Reading and Collinson 1996). Long intervals of mudrocks in the upper parts of the De Geerdalen Formation are herein interpreted to represent shallow closed or semiclosed standing water on delta plain or upper delta front.

Palaeosols indicates subaerial exposure (Boggs 2012), and is formed only if the sedimentation rate does not exceed the rate of pedogenesis (Kraus 1999). Palaeosols are thus an indicator of little or no sedimentation, and may be one of the clearest indicators of an interdistributary regime.

Palaeosols on top of distributary channels can indicate migration, diversion or abandonment of channels. Fine grained material underlying palaeosols may have formed in occasionally exposed standing water bodies on interdistributary areas on the delta plain. There is a significant shift from grey and yellow palaeosols, coal and coal shales to red and green beds in the Isfjorden Member. This might indicate restricted depositional environments, such as lagoons (Mørk 1999, Mørk 2015), but further investigations are needed to fully understand the shift.

8. Palaeosols in the De Geerdalen Formation

This chapter aims to present and discuss data from palaeosols in the De Geerdalen Formation. The data is collected from field observations, thin sections and XRD-analysis. The data is discussed in light of former research and theory. Note that palaeosol numbers refer to table number in Appendix A. The table provides the opportunity to compare palaeosols in a more coherent manner.

8.1 Field observations of palaeosols in the De Geerdalen Formation

Three main features are important for the recognition of palaeosols in the field: roots, soil horizons and soil structures (Retallack 1988). Observation of these features was performed in the field, and photographic interpretation and comparison, in order to distinguish palaeosols from other sediments. Observations of roots, soil horizons and soil structure are used to classify the palaeosols. This is further described in section 8.2.

8.1.1 Roots

Description

Roots in palaeosols in the De Geerdalen Formation are seen as three different features. 1) Coalified downwards branching and tapering features with a length of approximately 20 cm are seen Blanknuten and Šmidtberget (Fig.8.1 A). 2) Vertical elongated, irregular shaped calcareous nodules are observed in the Isfjorden Member at Deltaneset, Teistberget, Šmidtberget and Friedrichfjellet (Fig. 8.1 C and D). The nodules are ranging from some few cm to 10 cm. 3) Irregular coloured features with diffuse patterns in the upper soil horizons are commonly observed throughout the De Geerdalen Formation. An important observation of the features is that they are made of the same material as the rest of the palaeosol (Fig. 8.1B).

Note that many of the palaeosols in the De Geerdalen Formation does not show any evidence of roots (Appendix A)

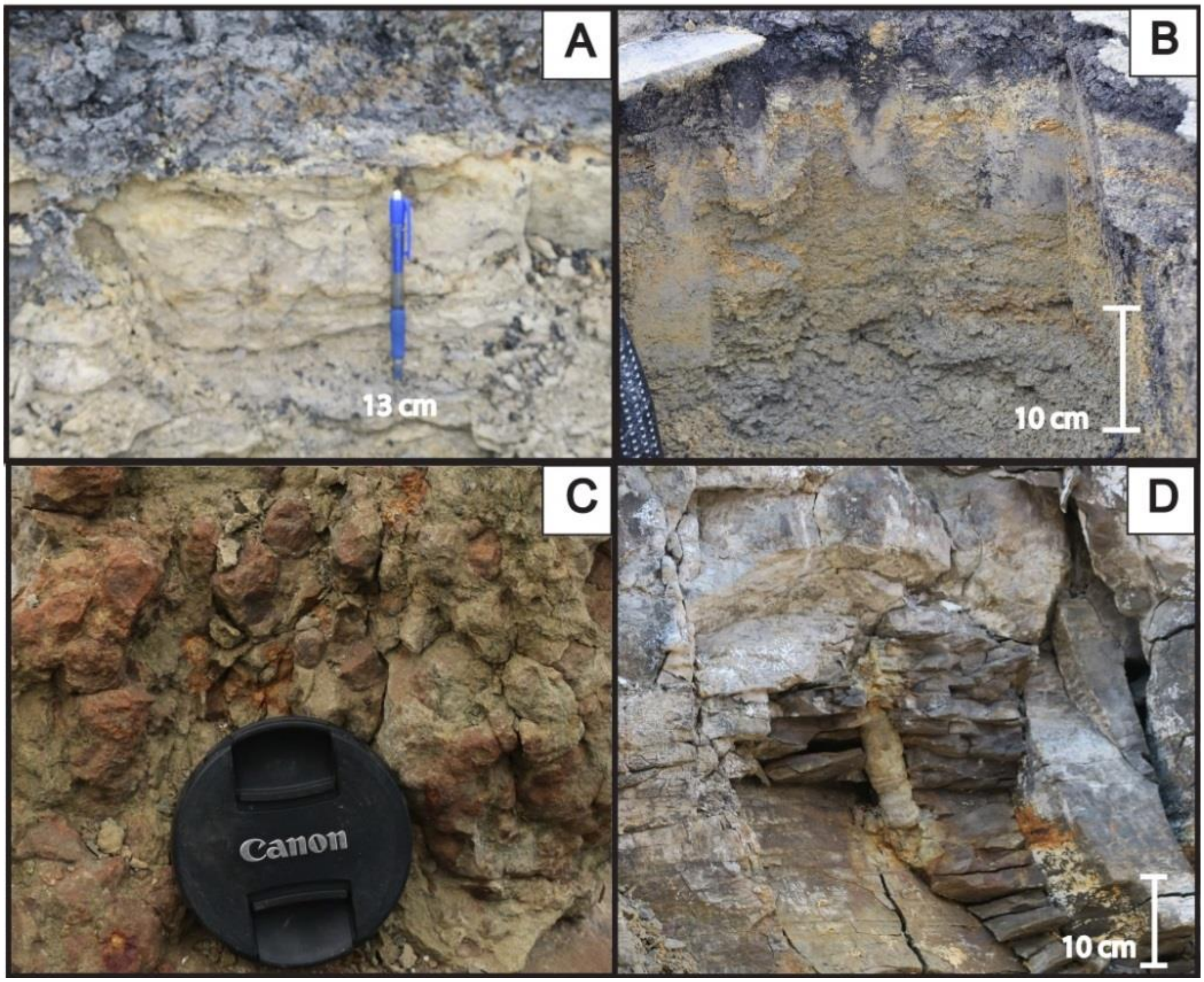


Figure 8.1: Roots. A) Coalified roots, Blanknuten. B) Irregular shaped and coloured features, Hahnfjella. C) Calcareous nodules, Šmidtberget. The camera lens is approximately five cm wide. D) Calcareous nodules hanging from calcrete bed right under the Slottet Bed, interpreted as rhizoliths. The picture is taken in Konusdalen close to the logged section at Deltaneset.

Interpretation

Roots are one of the most diagnostic features in palaeosols because their presence is evidence of subaerial exposure and plant growth, and thus meets the requirements for soils in almost all definitions (Retallack 1988). Root traces can be regarded as a trace fossil, and can be distinguished from other types of traces by the irregular shape with downwards branching and tapering (Retallack 1988). In this study the vertical branching or irregular traces are regarded as the most reliable evidence of roots in the De Geerdalen Formation.

Calcite cemented nodules that forms around roots are commonly found in red-bed succession, and especially in flood-plain deposits (Tucker 2011). Downward elongation is typical for nodules with rhizogenic origin (Tucker 2011). This is also seen in the calcareous nodules in the De Geerdalen Formation (Fig. 8.1 C and D).

Irregular colour with diffuse pattern is referred to as mottles in the literature, and are related to uneven distribution of minerals (Retallack 1997). The origin of root mottles is not fully understood, but they may originate from either 1) processes in the chemical microenvironment related to the living roots or 2) gleyed sediments resulting from anoxic conditions, caused by decay of organic matter soon after burial (Retallack 1988, Retallack 1997). With both explanations mottles can be regarded as an indirect indicator of roots, or at least plant growth (Retallack 1988).

Mottles are common in many palaeosols, but cannot be regarded as diagnostic for palaeosols (Retallack 2001). This is because uneven distribution of minerals can be caused by other processes than pedogenetic. For example are mottles in marine rocks common because of uneven distribution of minerals close to the sea floor or during burial (Retallack 2001). Some care must thus be taken in the interpretation of mottles as pedogenetic, and observations of other features associated with palaeosols are needed. Nevertheless, mottles in palaeosols in the De Geerdalen Formation are considered as most likely to have rhizogenic origin if other pedogenetic features are prominent.

Note that despite roots being a diagnostic feature, they are not required to be present in the palaeosols according to the wide definition of soils used in this study.

8.1.2 Soil horizons

Description

Soil horizons in the De Geerdalen Formation are primarily recognized by change in colour compared to surrounding sediments. The internal soil horizons in each palaeosol are different from each other in terms of colour, grain size and texture. In the brown and yellow palaeosols (Facies O_1) up to three different soil horizons are found within the individual palaeosols (Fig. 8.3). The upper horizon is often dark and organic rich. Lower horizons display greater diversity with pale grey, grey, brownish, yellowish or dark colours. Grey and pale soil horizons often have a sticky matrix that easily can be rolled like a snowball. The upper contact of the brown and yellow palaeosols (Facies O_1) is often sharp. The lower contact tends to be more gradually, but the permafrost often made it impossible to expose the lower contact. Both the upper and lower contacts in the red and green mudstones (Facies O_2) tend to be sharp.

The red and green mudstones (Facies O_2) often display only one soil horizon (red or green). On Deltaneset, Friedrichfjellet, Šmidtberget and Teistberget some of the red and green mudstones (Facies 0_2) have a strong reaction with hydrochloric acid. None of the yellow and brown palaeosols (Facies O_1) display any reaction with hydrochloric acid.

Interpretation

Soil horizons can be classified based on properties like grain size, colour, bed contacts and reaction with hydrochloric acid (Retallack 1988). A common classification system of soil horizons is that of the USDA soil survey Manual (Soil Survey Staff 2014). Classification of soil horizons in the De Geerdalen Formation in this study follows this system.

In the USDA soil survey Manual classification system the top of the soil is referred to as the O-horizon and is the surface accumulation of organic matter. The underlying A-horizon contains more or less decayed organic matter mixed with soil minerals. Many palaeosols are lacking the O-horizon making the A-horizon the surface horizon (Retallack 1988). The next layer is the E-horizon which is a bleached layer where minerals are washed out and transported downwards to the B horizon. The B-horizon thus appears more enriched in organic matter, clay and other types of material derived from the upper layers compared to over- and underlying horizons (Retallack 1988, Boggs 2011). The C-horizon is overlying the bedrock and is little influenced by pedogenetic processes. Although this is a common classification of soil horizons, most soils shows much more complexity than described here, and several methods have been used to classify soils (Boggs 2011, Retallack 1988). Calcareous soil horizons are according to this classification system defined as K-horizon (Table 4).

The complexity of soil horizons increases with time available for soil forming (Retallack 1984, 1988). Properties and rate of soil horizon forming are also influenced by other factors like climate and moisture regime (Cecil and Dulong 2003). In this study an attempt to classify soil-horizons in palaeosols of the De Geerdalen Formation has been done. See Appendix A for more detailed information about soil horizons in each palaeosol.

Table 4: Description and classification of soil horizons. Capitalized letters describes main categories of soil horizons. Main categories can be further described by sub categories represented by lowercase letters (Retallack 1988).

Category	New Term	Description	Old Term
Master Horizons	O	Surface accumulation of organic materials (peat, lignite, coal), overlying clayey or sandy part of soil	O
	A	Usually has roots and a mixture of organic and mineral matter; forms the surface of those paleosols lacking an O horizon	A
	E	Underlies an O or A horizon and appears bleached because it is lighter colored, less organic, less sesquioxidic, or less clayey than underlying material	A2
	B	Underlies an A or E horizon and appears enriched in some material compared to both underlying and overlying horizons (because it is darker colored, more organic, more sesquioxidic or more clayey) or more weathered than other horizons	B
	K	Subsurface horizon so impregnated with carbonate that it forms a massive layer (developed to stage III or more of Table 4)	K
	C	Subsurface horizon, slightly more weathered than fresh bedrock; lacks properties of other horizons, but shows mild mineral oxidation, limited accumulation of silica carbonates, soluble salts or moderate gleying	C
	R	Consolidated and unweathered bedrock	R
Gradations Between Master Horizons	AB	Horizon with some characteristics of A and B, but with A characteristics dominant	A3
	BA	As above, but with B characteristics dominant	B1
	E/B	Horizon predominantly (more than 50%) of material like B horizon, but with tongues or other inclusions of material like an E horizon	A&B
Subordinate Descriptors	a	Highly decomposed organic matter	—
	b	Buried soil horizon (used only for pedorelict horizons with paleosols; otherwise redundant)	b
	c	Concretions or nodules	cn
	e	Intermediately decomposed organic matter	—
	f	Frozen soil, with evidence of ice wedges, dikes, or layers	f
	g	Evidence of strong gleying, such as pyrite or siderite nodules	g
	h	Illuvial accumulation of organic matter	h
	i	Slightly decomposed organic matter	—
	k	Accumulation of carbonates less than for K horizon	ca
	m	Evidence of strong original induration or cementation, such as avoidance by root traces in adjacent horizons	m
	n	Evidence of accumulated sodium, such as domed columnar peds or halite casts	sa
	o	Residual accumulation of sesquioxides	—
	p	Plowing or other comparable human disturbance	p
	q	Accumulation of silica	si
	r	Weathered or soft bedrock	ox
	s	Illuvial accumulation of sesquioxides	ir
	t	Accumulation of clay	t
	v	Plinthite (in place, pedogenic laterite)	—
	w	Colored or structural B horizon	—
	x	Fragipan (a layer originally cemented by silica or clay, and avoided by roots)	x
	y	Accumulation of gypsum crystals or crystal casts	cs
	z	Accumulation of other salts or salt crystal casts	sa

8.1.3 Soil structure

Description

Observations of soil structure in the De Geerdalen Formation deals with the internal structure of the palaeosols. The most important observation is size and shape of lumps, which are classified according to table 5. Shape of lumps found in the De Geerdalen Formation includes platy, angular blocky, subangular blocky and granular (table 5). The diameter of lumps is typical in the range of 1 mm and up to 1-2 cm.

Platy structure is seen both throughout the soil profile and in the lower parts of the soil. Angular and subangular blocky structure occurs frequently both in the Brown and yellow palaeosols (Facies O_1) and the Alternating red and green mudstones of the Isfjorden Member (Facies O_2). Granular soil structure is found in palaeosol types except carbonate soils (defined in section 8.2.5). Noteworthy is that all Horizonated soils (defined in section 8.2.2) have granular soil structure.

For a more details about soil structures in each palaeosol see Appendix A.

Interpretation

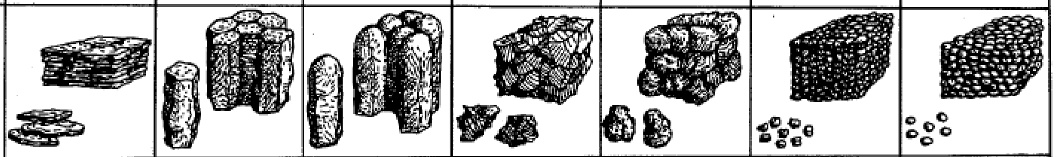
Shape and size of lumps in palaeosols are linked to modifying of parent material related to pedogenetic processes such as bioturbation by plants and animals, and wetting and drying (Retallack 2001). The modifying causes many palaeosols to appear hackly upon first sight (Retallack 1988). This structure originates from open spaces and weaker zones surrounded by more stable aggregates in the original soil. The stable aggregates are termed peds (Retallack 2001)

Platy structure indicates that relict bedding of the parent material is present in the palaeosol (table 5) (Retallack 1988). In this study palaeosols with prominent relict bedding throughout the soil profile are consider to have formed under conditions where the pedogenetic modifying of the parent material was too weak to overprint the original structure of the parent material. If relict bedding only occurs in the lower reaches of the palaeosols it is considered as the transition zone between sediments influenced by pedogenetic processes and bedrock.

Both angular and subangular blocky structure are associated with swelling and shrinking, and cracking around roots and burrows (table 5) (Retallack 1988). Palaeosols in the De Geerdalen with angular or subangular blocky structure are thus considered as influenced by cracking, possibly due to shrinking and swelling.

Granular palaeosols indicates high biological activity, and thus high soil fertility. This is characteristic for surface soils (A-horizon) on soil formed on grassland (Retallack 1988). The granular soil structure seen in all the Horizonated soil (section 8.2.2) is considered as supporting the mature nature of this soil type.

Table 5: Classification and likely causes for ped types in palaeosols (Retallack 1988).

TYPE	PLATY	PRISMATIC	COLUMNAR	ANGULAR BLOCKY	SUBANGULAR BLOCKY	GRANULAR	CRUMB
SKETCH							
DESCRIPTION	tabular and horizontal to land surface	elongate with flat top and vertical to land surface	elongate with domed top and vertical to surface	equant with sharp interlocking edges	equant with dull interlocking edges	spheroidal with slightly interlocking edges	rounded and spheroidal but not interlocking
USUAL HORIZON	E,Bs,K,C	Bt	Bn	Bt	Bt	A	A
MAIN LIKELY CAUSES	initial disruption of relict bedding; accretion of cementing material	swelling and shrinking on wetting and drying	as for prismatic, but with greater erosion by percolating water, and greater swelling of clay	cracking around roots and burrows; swelling and shrinking on wetting and drying	as for angular blocky, but with more erosion and deposition of material in cracks	active bioturbation and coating of soil with films of clay, sesquioxides and organic matter	as for granular; including fecal pellets and relict soil clasts
SIZE CLASS	very thin < 1 mm	very fine < 1 cm	very fine < 1 cm	very fine < 0.5 cm	very fine < 0.5 cm	very fine < 1 mm	very fine < 1 mm
	thin 1 to 2 mm	fine 1 to 2 cm	fine 1 to 2 cm	fine 0.5 to 1 cm	fine 0.5 to 1 cm	fine 1 to 2 mm	fine 1 to 2 mm
	medium 2 to 5 mm	medium 2 to 5 cm	medium 2 to 5 cm	medium 1 to 2 cm	medium 1 to 2 cm	medium 2 to 5 mm	medium 2 to 5 mm
	thick 5 to 10 mm	coarse 5 to 10 cm	coarse 5 to 10 cm	coarse 2 to 5 cm	coarse 2 to 5 cm	coarse 5 to 10 mm	not found
	very thick > 10 mm	very coarse > 10 cm	very coarse > 10 cm	very coarse > 5 cm	very coarse > 5 cm	very coarse > 10 mm	not found

Note that even though the internal soil structures can give an indication of conditions during paedogenesis, the structure might have been interrupted later by freezing and thawing in the active layer of the permafrost, weathering or diagenetic processes. Some care must thus be taken in the interpretations.

8.2 Classification of Palaeosols in the De Geerdalen Formation

As discussed in chapter 8.1.2 many different systems have been used in order to classify palaeosols, as those of United States Soil Taxonomy (Soil Survey Staff 2014) and the FAO (1974) classification. These systems are based on modern soil classifications. The classification of Mack et al. (1993) is specific for palaeosols, even though the system is based on modern soils. The system focuses on features with best preservation potential in the geological record, and especially properties that are easy recognizable in the field and in petrographic thin sections. The system is also simpler than the other mentioned systems, which makes the classification from field observations easier. The classification of Duchaufour (1982) is differs from the other systems by the focus on soil forming processes rather than soil properties.

Soil types relevant for this study will be described in the following paragraphs. The soil classification is mainly following on that of Mack et al. (1993) with some modification. Brown and yellow palaeosols (Facies O_1) is classified as Vertisols, Argillisols and Protosols. Red and green mudstones (Facies O_2) are classified as Noncalcareous red and green mudstones and Calcrete.

8.2.1 Dark and homogenous soils (Vertisols)

Description

This soil type is recognized by having a homogenous, dark and often organic rich matrix (Fig 8.2). The soil lack clear soil horizons, but a thin layer of coal shale is often seen in upper parts. Mottles are seen on Blanknuten, Hahnfjella and Schmidtberget. Blocky soil structure is common. Slickensides and signatures of relict bedding are not found in this soil type. The thickness is in the range of 20 to 110 cm. This soil type is most commonly observed under the Isfjorden Member. They are often found on top of Barrier bar deposits (FA2.1), Distributary channels (FA 2.3) and Floodplain (FA 3.1), but do also occur in Interdistributary areas (FA 3.2) on Hahnfjella.

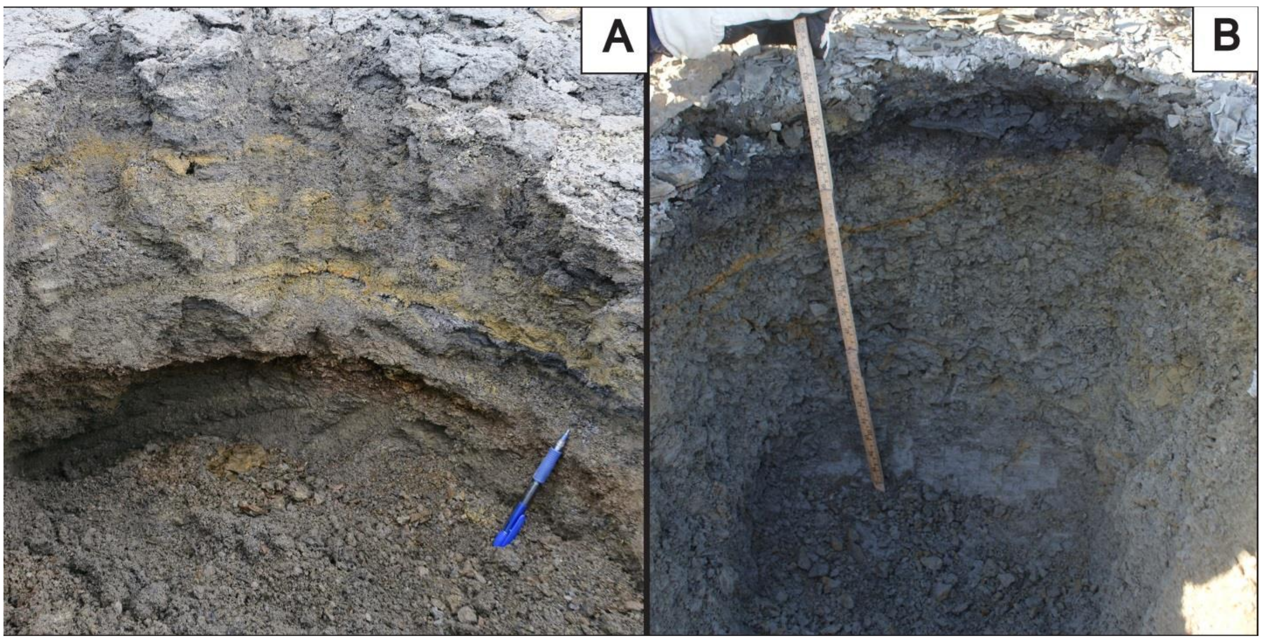


Figure 8.2: *Dark and homogenous soils.* The soil horizons are less prominent compared to the other types of palaeosols in this study. Mottles are seen in both soil profiles. A) Dark homogenous soil with very thin coal seams, Blanknuten (Appendix A, palaeosol No. 36). B) Dark homogenous soil with fine, subangular blocky soil structure, Wilhelmøya (Appendix A, palaeosol No. 2).

Interpretation

This soil type as observed in the De Geerdalen Formation does not fit entirely with any of the soils in the classification system of Mack et al. (1993) or Soil Survey Staff (1998), but a homogenous appearance due to pedoturbation is the main features of Vertisols as defined in both systems. The homogenous appearance is according to the definitions due to shrinking and swelling of expandable clay. In this study XRD-analysis was not performed on this soil type, and the presence of expandable clay is thus unknown.

In this study dark and homogenous soils are classified as Vertisols following the Mack et al. (1993) system, with the modification that additional morphological features related to shrinking and swelling does not have to be present. However, blocky structure is commonly observed in this soil type in the De Geerdalen Formation. This structure is associated with soils that have undergone shrinking and swelling (Retallack 1988) (Figure 8.2B). The question is if the blocky structure is of pedogenetic origin or formed under the diagenesis. Only if the blocky structure is pedogenetic it can be attributed to shrinking and swelling processes during soil forming.

One of the signature features of Vertisols, slickensides (Mack et al. 1993, Soil Survey Staff 2014), is not found. Enga (2015) noted the same lack of slickensides in soils in the De Geerdalen Formation oppose soils in the Snadd Formation where slickensides are abundant. Enga (2015) suggest several explanations for the lack of slickensides in the De Geerdalen Formation: 1) poor field observations 2) not preserved slickensides because of diagenetic processes different from the Barents Sea 3) weathering of near surface sediments resulting in destroyed slickensides features.

It is also possible that the homogenous appearance of this soil type in the De Geerdalen Formation was caused by something else than shrinking and swelling processes. This could for example be explained by immature soils that did not developed soil horizons. However, the lack of relict bedding and presence of mottles, coal shale and root traces points towards a more mature soil.

Today Vertisols are most commonly found in the tropical and sub-tropical zones, but have also been described from the temperate zone (Khitrov and Rogovneva 2014). Vertisols forms in areas with mean annual temperature of 180 – 1520 mm (Retallack 2001). Most Vertisols occur between 45°S and 45°N, but they are found as far north as 54° in Mordovia and Samara in Russia (Khitrov and Rogovneva 2014). The forming of Vertisol on clay or shale might require only some few hundred years (Retallack 2001).

8.2.2 Horizonated soils (Argillisols)

Observations

The horizonated soils in the De Geerdalen Formation are recognized by having at least three soil horizons, where at least one of the layers are clay rich and bleached (Figs 8.3, 9.2A and 9.3A). Mottles are commonly found in the horizonated palaeosols, but relict bedding is not observed in this type. Roots are seen on Šmidtberget. Wood fragments found in situ in the A-horizon is seen at Wilhelmøya. All the horizonated soils have granular soil structure in at least one of the soil horizons. Five palaeosols are classified as horizonated in the De Geerdalen Formation (Appendix A). The thickest horizonated palaeosol is found under the Isfjorden Member on Wilhelmøya. The rest of the horizonated palaeosols are within the member. The thickness of the horizonated palaeosols is in the range of 40 – 100 cm.

Figure 8.3: *Horizonated palaeosol on Šmidtberget.* The horizons are interpreted as A-horizon consisting of 5 cm coal shale overlying a 35 cm thick E-horizon. The horizon is bleached with root traces. The coarse reddish to yellow layer underneath is interpreted as B-horizon. Relict bedding is seen at the base, terminating the lower reaches of soil forming processes in the palaeosol (Appendix A, Palaeosol No.41).

Interpretation

The well-defined soil horizons found in this palaeosols might be due to wash out of minerals leading to bleaching of some layers and enrichment in underlying layers (illuviation). According to the Mack et al. (1993) system is soils with a layer enriched in clay due to illuviation classified as Argillisol. The thickness of the soil and soil horizons are not considered in the definition of Mack et al. (1993). The horizonated soils in the De Geerdalen Formation are placed under the Argillisol-category. Argillisols in the Mack et al. (1993) system corresponds and to Ultisols or Alfisols in the USDA soil taxonomy.

The development and complexity of soil horizons increases with time (Harden 1982, 1990). For example is transport of clay downwards by washing believed to require up to several thousands of years (Harden 1982, 1990). Roots, mottles and wood fragments emphasized the mature nature of this palaeosols (Retallack 1988). Based on this the horizonated soils are regarded as the most mature palaeosol found in this study. Lack of relict bedding is supporting long lasting soil forming conditions.

8.2.3 Immature palaeosols (Protosol)

Description

This type is mainly recognized by having yellow, greyish or brownish colours in contrast to surrounding sediments. The number and thickness of soil horizons are limited. Mottles and organic rich upper soil horizon is common, but do not appear as prominent. Relict bedding is common in parts of these palaeosols. These palaeosols are common both in and under the Isfjorden Member. The thickness is in the range of 10 to 70 cm. Examples of immature palaeosols are seen in figure 8.4.

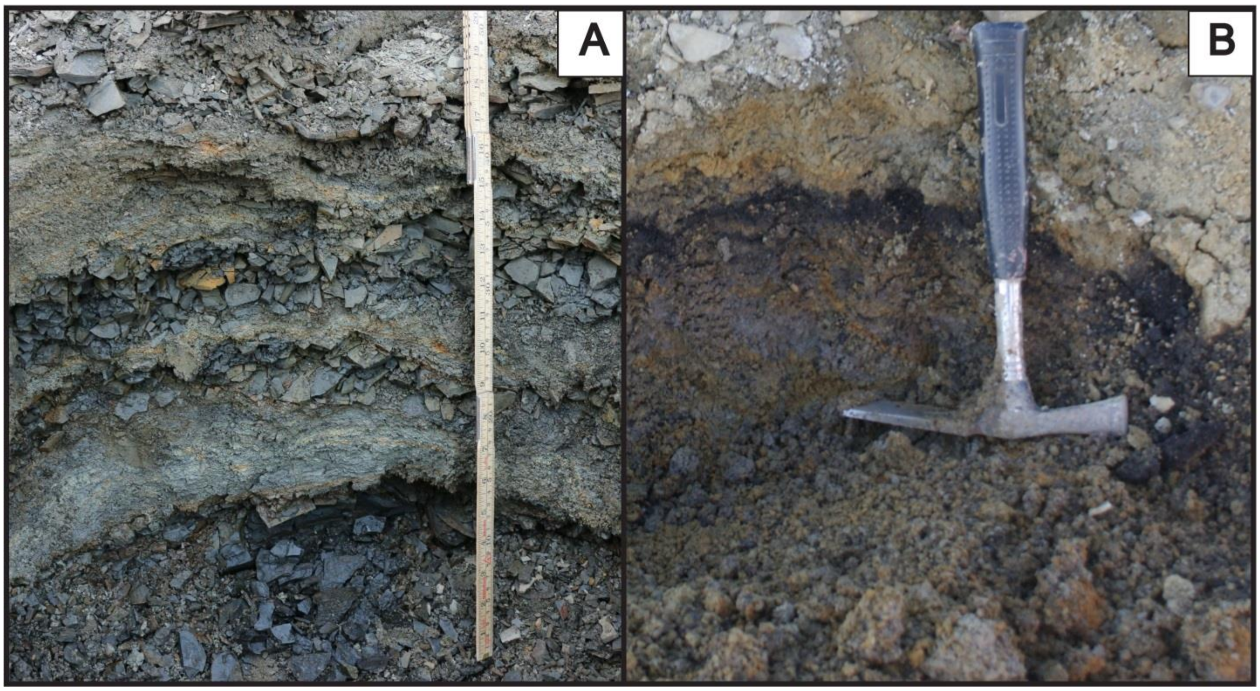


Figure 8.4: *Immature soils.* A) Horizontal orientated mottles and fine grained, sticky texture leads to the interpretation of a palaeosols. Relict bedding and weakly developed soil horizons makes the soil appearing immature, Šmidtberget (Appendix A, palaeosol No. 43). B) Organic rich shale overlying brown silty clay with mottles, Hahnfjella (Appendix A, palaeosol No. 16).

Interpretation

The lack of distinct pedogenetic features in this soil type leads to the interpretation of immature palaeosols. In the Mack et al. (1993) system is soils with weakly developed soil horizons defined as Protosols. The Protosols can have pedogenetic features characteristic for other soil types, but those are not the most prominent in the soil (Mack et al. 1993). In the system of Soil Taxonomy (Soil Survey Staff 2014), the term covers the range of characteristics of Entisol to Inceptisols. The immature palaeosols in the De Geerdalen Formation is thus placed under the Protosol category as defined in Mack et al. (1993).

The pedogenetic interpretation of the Protosols in the De Geerdalen Formation are often mainly based on changes in colour compared to surrounding sediments, and verified by minor pedogenetic features like high organic content, coal shale or weakly developed soil horizons. The distinguishing between Protosols and other sediments is thus not straight forward, since colour changes in sediments can be caused by other processes than pedogenetic. Some uncertainty about the pedogenetic origin is thus present. Coloured mudrock is for example known to form due to presence of iron sulphur minerals (Boggs 2009). This can indicate reduced conditions, e.g. because of presence of organic matter on tidal flats (Pettijohn et al. 1987, Boggs 2009).

Immature soils occur in conditions not favourable for soil forming. Reasons for this can be short time available for forming or high sedimentation rates (Retallack 2001). Even though the Protosols shows poorly pedogenetic development, they are considered as important in this study because they serve as evidence of subaerial exposure, or at least very shallow water (some few to tens of centimetres).

8.2.4 Noncalcareous red and green mudstones

Observation

The red and green mudstones are restricted to the Isfjorden Member. They are typically forming 0.3 – 2 meters thick units of alternating green and red layers, but unit thickness up to 5 m is occasionally seen (Appendix A). Discrete beds containing only red or green mudstones do occur. The thickness of these beds is typical in the range of some few to tens of cm. Typical structure of the mudstones is blocky, and mottles are common. Root traces are not observed. XRD-analyses from a noncalcareous red and green mudstone (Appendix A, palaeosol No. 8) on Wilhelmøya showed kaolinite and goethite as dominating clay minerals (Table 6).

The mudstones displays the same colour as carbonate soils, but are distinguished from them by having no reaction with hydrochloric acid. Examples of Noncalcareous red and green mudstones are seen in figure 6.22.

Interpretation

Mottles, nodules and distinct horizontal coloured layers in these sediments points towards pedogenetic origin. Alternating red and green layers might be attributed to changes in redox regime, where red colours is related to oxidising conditions and green colours occur in reduced environment. In the classification system of Mack et al. (1993) is soil that have a subsurface horizon formed under reduced conditions as the most prominent feature classified as Gleysol. In the Soil Taxonomy system (Soil Survey Staff 2014) soil horizons formed under reduced conditions are classified as gleyed sub-horizon.

In this study the palaeosols are defined as described under observations, and is not following any of the classification systems. This is done in order to underline the properties of the soil (red and green colours and noncalcareous), which are considered as both diagnostic and required features for this palaeosol type in the De Geerdalen Formation. XRD analyses and clay minerals are discussed in Chapter 8.4.

8.2.5 Carbonate soils (Calcrete)

Observation

These palaeosols have the same red and green colours as the noncalcareous green and red mudstones, but differs from the other palaeosols in this study by showing a strong reaction with hydrochloric acid. Nodules, mottles and peds are common. Deltaneset is the only locality where Carbonate soils is the most dominant palaeosol type (Fig. 8.5 A and B). On Friedrichfjellet, Šmidtberget and Teistberget only few discrete beds shows a strong reaction with hydrochloric acid, and Noncalcareous red and green mudstones are the dominating soil type in the Isfjorden Member at those localities (Fig. 8.5 C).

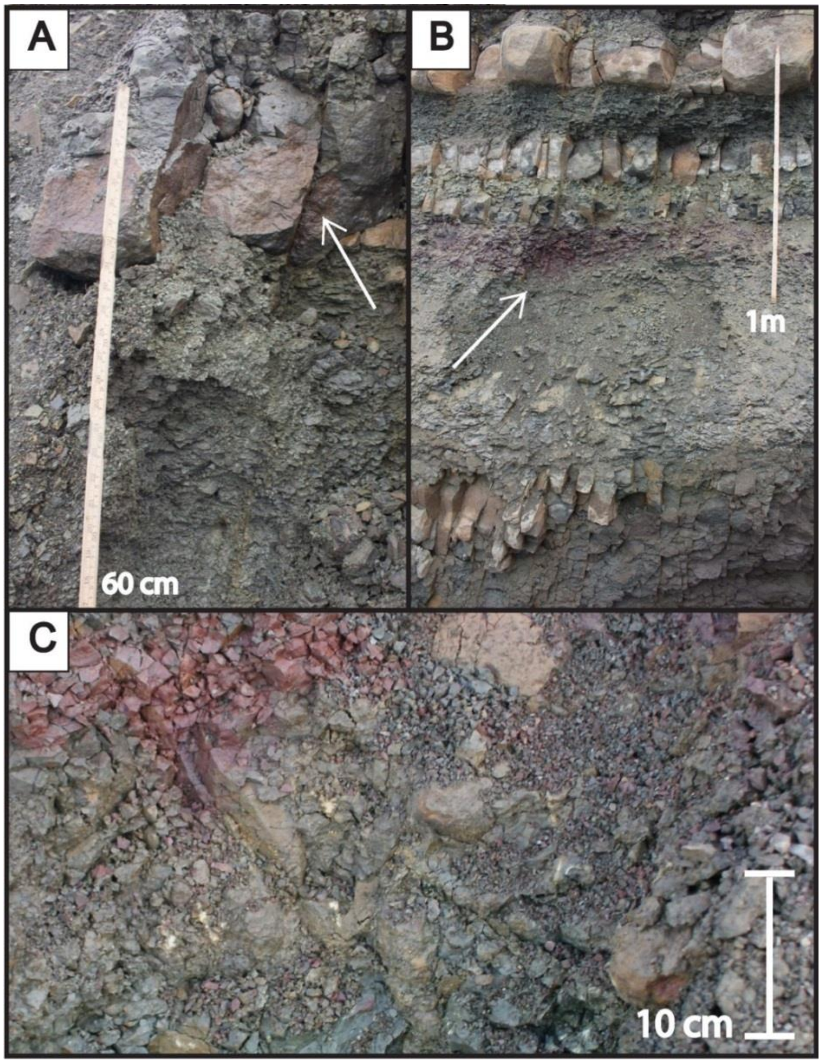


Figure 8.5: *Carbonate soils* A) Carbonate soil at Deltaneset. See Fig. 8.7 A and B for pictures of thin section made of a sample taken next to arrow (Appendix A, palaeosol No. 49). B) Carbonate soil with nodules at Deltaneset. Thin section made of a sample taken next to the arrow shows clear pedogenetic evidence (Fig. 8.8) (Appendix A, palaeosol No. 57). C) Carbonate soil with coarse angular blocky structure. Nodules might be calcified roots, Friedrichfjellet, Agardhdalen (Appendix A, palaeosol No. 31).

Interpretation

These palaeosols are interpreted as Calcrete (also termed caliche). Calcrete is according to Mack et al. (1993) recognized in the field by having calcic horizon as the most prominent feature. According to Soil Survey Staff (2014) (table 4) is calcrete classified as K-horizon: “Subsurface horizon so impregnated with carbonate that it forms a massive layer” (Retallack 1988). Thin section analysis which show alveolar features, carbonate cement and aggregates possibly cemented root casts (Fig. 8.6, 8.7, and 8.8) is consistent with a Calcrete palaeosol. This is further described in Chapter 8.3.

The forming of calcrete can involve both pedogenetic and non-pedogenetic processes (Leeder 1975, Carlisle 1983). Non-pedogenetic processes are mainly related to extensive calcium carbonate precipitation in the shallow phreatic zone (zone of ground water saturation) in semi-arid to arid climate (Wright and Tucker 1991). Pedogenetic calcretes form as a result of eluvial/illuvial processes, also known as downwards transport of material in a soil profile (Arakel and McConchie 1982). The eluvial processes involve leaching or removal of materials, while illuvial processes are deposition of the materials further down in the soil profile. Since this study only deals with soil calcrete, only this type is described in this chapter.

A definition of pedogenetic calcrete was suggested by Watts (1980) with modification of the one from Goudie (1973): “Pedogenic calcretes are terrestrial materials composed dominantly, but not exclusively, of $CaCO_3$, which occur in states ranging from nodular and powdery to highly indurated and result mainly from the displacive and/or replacive introduction of vadose carbonate into greater or lesser of soil, rock or sediment within a soil profile”.

As seen in the definition of calcrete is $CaCO_3$, or calcium carbonate, the main component. Many sources of calcium carbonate are known, and include local rocks, rainwater, dust, sea spray and biota (Cailleau et al. 2004). In semi-arid to arid climate the main source is considered to be rainwater and dust (Alonso-Zarza and Wright 2010).

The accumulation and fixing of carbonate is depended on deficit of moisture. Organisms, and in particular plants and fungi, plays an important role in the carbonate-fixing process. Fixed carbonate can to a certain degree survive seasonal rainfall with precipitation exceeding evaporation (Alonso-Zara and Wright 2010). Since the forming of calcrete is depended on relative dry conditions, the presence of them is important in palaeoclimate reconstruction.

Yaalon (1988) estimated that calcrete covers 13 % of present land surface. Australia, the driest continent on Earth, has calcrete covering 21 % of the land surface (Chen et al. 2002). Most present day calcretes are found in areas with a mean annual temperature of 16-20°C, but they also occur in cold desserts, suggesting that rainfall, rather than temperature, is the critical factor for calcrete formation. Mean annual rainfall of 100 to 500 mm is typical for calcrete formation (Goudie 1983), but calcrete can also form at areas with rainfall up to 1000 mm per year (Mack and James 1994).

Precipitation of carbonate can be divided in biogenetic and abiogenetic processes. Abiogenetic processes involve evaporation, degassing and evapotranspiration (Alonso-Zara and Wright 2010).

Biogenetic processes are driven by plants and other organisms. This includes metabolic products and calcifying of microbes, fixing of calcite by plant roots and mobilisation and precipitation of carbonates by insects, earth worms and slugs. Most calcretes form as a combination of biogenetic and abiogenetic processes.

8.3 Petrographic observations of the calcretes

Two better constrain the interpretation of calcrete soil profiles three petrographic thin sections were made from carbonate nodules in the red and green mudstones within the Isfjorden Member at Deltaneset. All thin sections show features commonly observed in fossil and modern calcretes (e.g Wright 1993, Alonso-Zara and Wright 2010). Microfeatures in calcretes can be divided in alpha microfabrics, which lack biogenetic features, and beta microfabrics who are dominated by biogenetic processes (Alonso-Zara and Wright 2010).

All the samples display both alpha-microfabrics and beta-microfabrics, but sample eight has the most prominent biogenetic features. The following paragraphs present observations and discussion of alpha- and beta-microfabrics.

The thin sections are made of sample 1, 8 and 11 from Deltaneset. See appendix A for more information about the samples, and Appendix B for the Deltaneset log.

8.3.1 Alpha-microfabrics

Alpha-microfabrics are non-biogenetic features related to supersaturation of the soil solution. This can result in carbonate replacement, recrystallization, precipitation of carbonate in pores and also non-carbonate particles in the soil. Multiple phases of calcite growth might occur (Alonso-Zara and Wright 2010). A groundmass of crystalline carbonate belongs to the alpha microfabric features and is the most prominent evidence of alpha-microfabrics seen in the thin sections in the carbonate nodules from the Isfjorden Member at Deltaneset (Fig. 8.6).

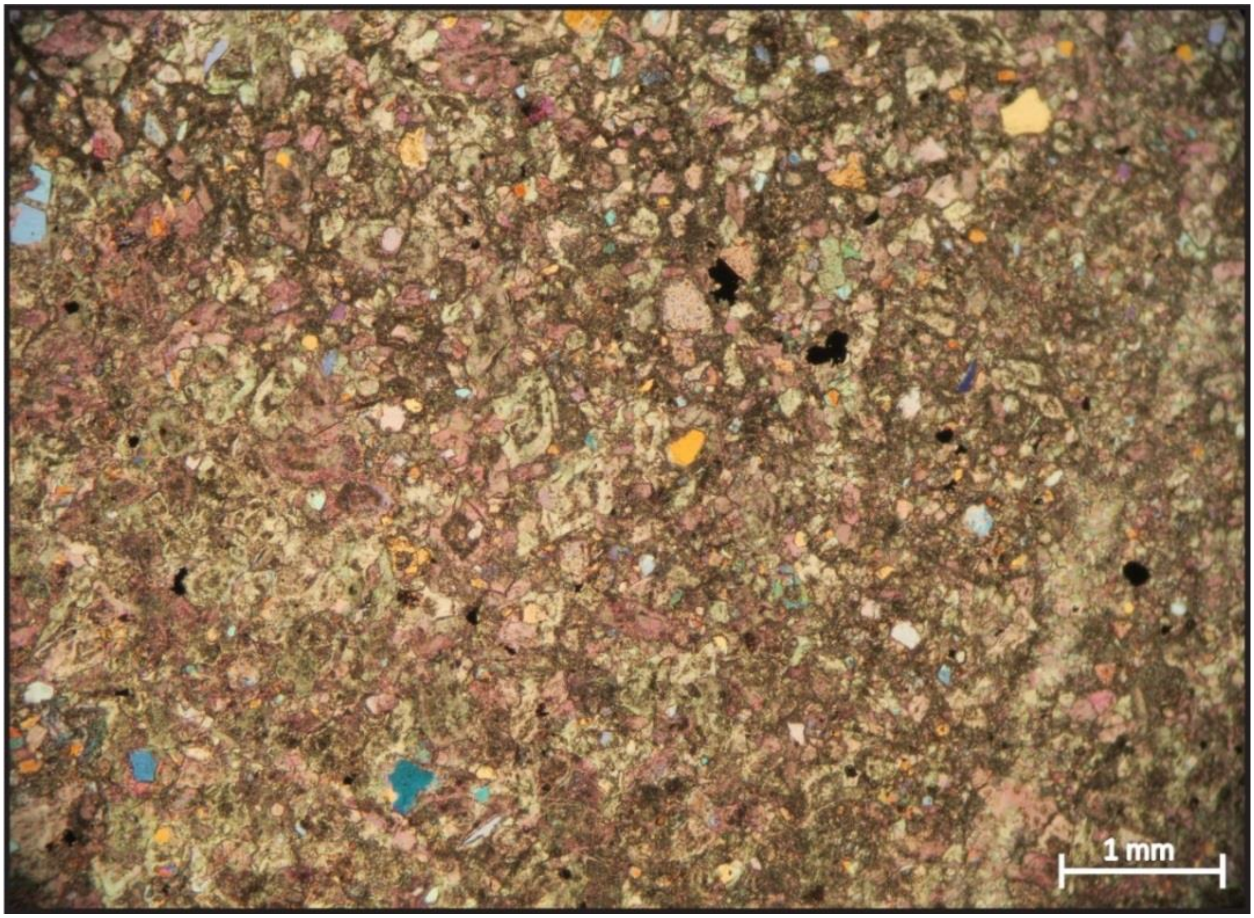


Figure: 8.6: Groundmass of crystalline carbonate is the most prominent feature of alpha-microfabrics in the samples from Deltaneset. The mass of small crystals are regarded as precipitation of carbonate due to oversaturated water. Larger crystals are probably due to infill of pore space.

8.3.2 Beta-microfabrics

A number of features are characteristic for beta microfabrics. Beta-microfabrics recognized at Deltaneset include alveolar septal structures, coated grains, calcified filaments, calcified roots, spherulites and faecal pellets.

Coated grains made of relics of the host rock, micrite or parts of alveolar-septal features are important component of beta-microfabrics. Alonso-Zarza et al. (1998) found biogenetic thin carbonate coated grains gradual to diffuse contact to surrounding matrix. Examples of coated grains from Isfjorden Member, Deltaneset are seen in Figure 8.7 D and 8.8 C.

Here also millimetre sized calcite filament consisting of straight or sinuous tubes connected to each other are seen (Figure 8.8 B). These and are believed to origin from micro-organisms, mainly fungi, but might also be related to roots and other types of microorganisms (Verrecchia et al. 1993).

Studies of pedogenetic mud aggregates in dryland river systems in the North Sea showed that it is possible to differentiate in-situ mud aggregates related to pedogenetic processes in palaeosols and reworked aggregates on the floodplain (Müller et al. 2004). Müller et al. (2004) describe aggregates with similar texture separated with calcite cement as in-situ formed aggregates. Similar observations as in-situ formed aggregates can be seen in 8.7 D and 8.8 D. In the study of Müller et al. (2004) reworked aggregates were recognized by more heterogeneous appearance. Possibly reworked aggregates are seen in Figure 8.7 C.

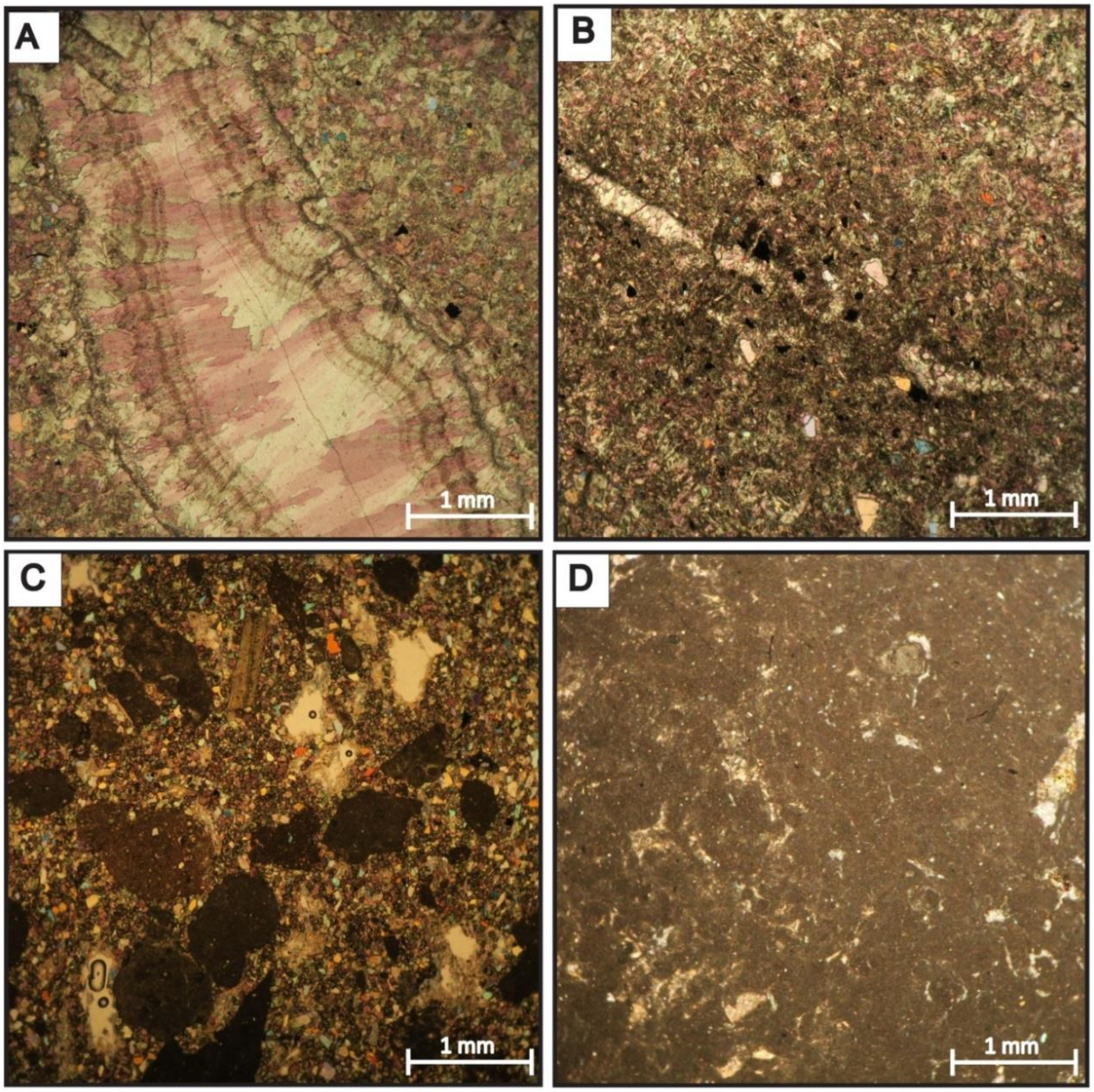


Figure 8.7: A) Calcified boarder in elongated feature, possibly a root. The larger crystals in the feature are probably infill of carbonate in pore room at a later stage. The surrounding carbonate matrix is presumably alpha-microfabrics formed as a result of supersaturation of pore water, sample 1, Deltaneset. B) Calcified infilled sample 1. C) Aggregates with similar texture (biogenetic) surrounded by calcite micrite (non-biogenetic), sample 11 D) Thin carbonate coated grains with diffuse and gradual contact to the surrounding matrix (biogenetic) Sample 11.

Alveloar-septal structures are mainly associated with fungal activity related to roots. They occur as arcuate shaped millimetre sized structures in pores, at the boarder of root traces or intercalated between micritic laminae. The septas can be formed by micritic crystals of the same size, or by needle fibre calcite. Feature interpreted as alveolar-septal are seen in Figure 8.8A, B and C.

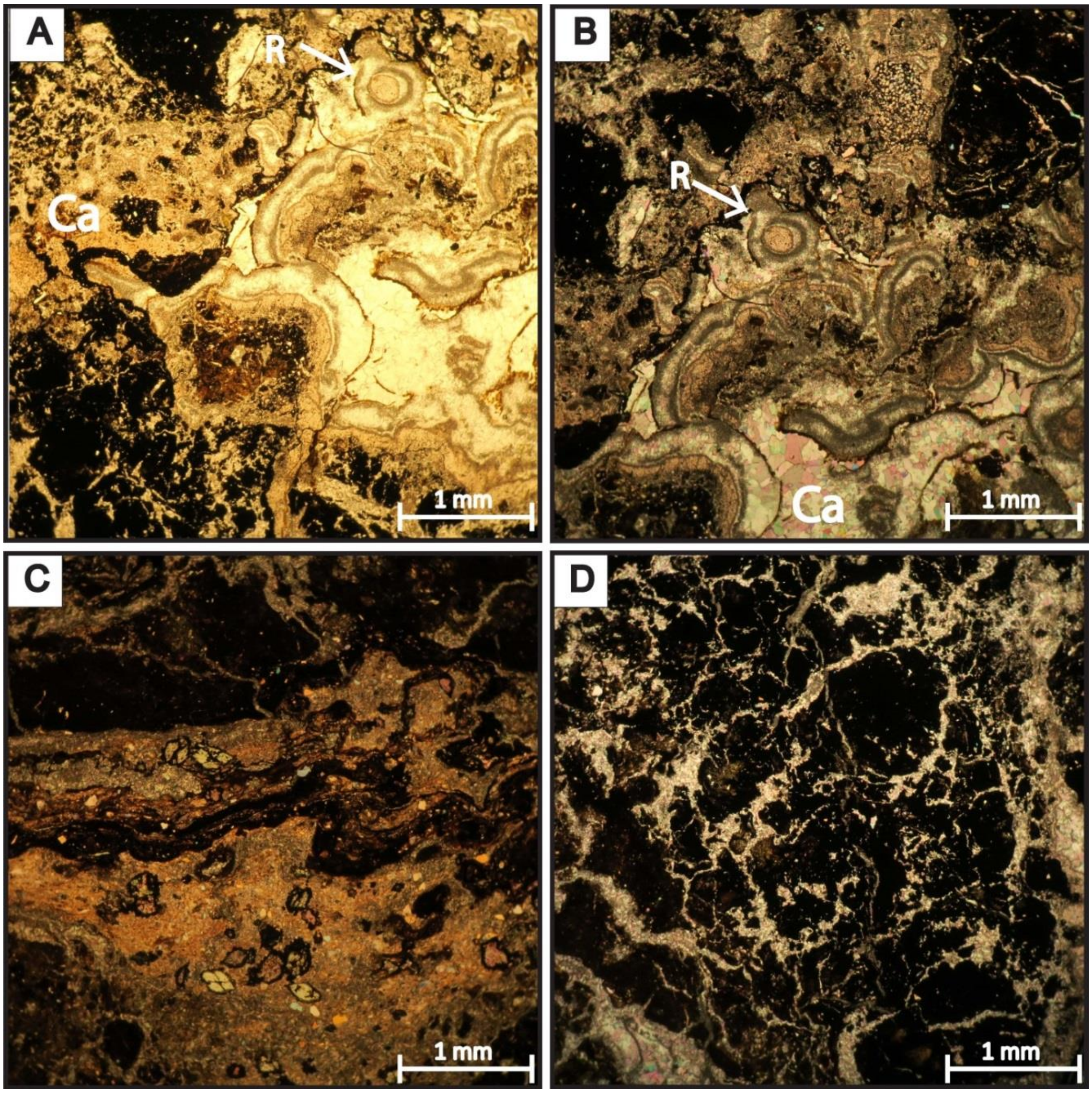


Figure 8.8: A) Alveolar features, carbonate cement and aggregates in sample 8. Possibly carbonate coating of root filaments is seen next to arrow. Plane polarized light. B) Same picture as B in cross polarized light. C) Feature interpreted as alveolar-septal coating roots. Pores are filled in with calcite micrite. Larger calcite grains are coated. Cross polarized light. D) Carbonate mudstone (micrite) aggregates (biogenetic) surrounded by calcite cemented fractures (non-biogenetic). All the pictures are from sample eight on Deltaneset.

8.4 XRD-analysis

Observation

Five samples were chosen for XRD-analyses (Table 6). The samples are taken from palaeosol No. 5, 8, 16, 27 and 37 (Appendix A) The quartz content is relatively similar in all the samples ranging from around 35 % (No. 8 and 37) to 50 % (sample 5, 16 and 27). The plagioclase and mica content differs more. Kaolinite is the dominating clay mineral in palaeosol No. 5, 8 and 37. Palaeosol No. 8 has similar amount of kaolinite and chlorite and nearly the same of goethite. In addition it is the only sample without jarosite. Sample 37 has high jarosite content compared to the other samples. Chlorite is the only clay mineral in sample three.

Table 6: Mineral composition of pedogenetic sediments based on bulk analysis in XRD. Total number not equal to 100 % in sample three is due to rounding.

Palaeosols No.	5	8	16	37	27
Locality	Wilhelmøya	Wilhelmøya	Hahnfjella	Blanknuten	Klement`evfjellet
Meters on log	66	89	210	28	195
Sample number	15.2.10.C	15.2.17.C	15.2.32C	15.12B	15.1.24C
Journal No.	160089	160090	160091	160092	160088
Mineral group	Weight %				
Quartz	53.3	39.7	50.1	34.7	49.4
Plagioclase	17.2	10.6	24.6	20.5	21.4
Mica	5.8	19.8	8.2	18.6	17.9
Jarosite	3.9	-------	3.7	14.1	5.1
Alkalie-feldspar	10.5	4.7	6.8	4.7	4.5
Pyroxene	1.8	3	1.5	2.3	1.7
Kaolinite	7.5	7.9	-------	5.1	-------
Chlorite	-------	7.9	4.3	-------	-------
Goethite	-------	6.4	-------	-------	-------
Total	100	100	99.2	100	100

Interpretation

Typical climate for forming of pedogenetic kaolinite is warm and humid climate (Sheldon and Tabor 2009), and kaolinite as dominating clay mineral is typical in areas with seasonal precipitation between 1000 and 2000 mm (Retallack 2001). Kaolinite can probably also form in soil profiles during cooler temperatures with seasonal growth during warm months (Sheldon & Tabor 2009). The finding of goethite and kaolinite as dominating clay minerals in the Non calcareous red and green mudstone (Palaeosol No. 8) is regarded as surprising keeping in mind that Calcrete is found elsewhere in the Isfjorden Member. This is further discussed in Chapter 9.4

Goethite often occurs together with kaolinite in palaeosols (Sheldon and Tabor 2009). Kaolinite and goethite as dominating clay minerals in palaeosol number eight supports the assumption that Noncalcareous red and green mudstones of the Isfjorden Member have pedogenetic origin. Dehydration and recrystallization of brown hydroxide minerals such as goethite are associated with bright red colours. Recrystallization creates coarse grains that becomes the red iron oxide hematite when weathered (Retallack 1991). From the literature especially two mechanisms have been related to the transformation of goethite to hematite. 1) Red colours are known from strongly developed palaeosols formed in dry, tropical climate. Such origin of red soils in general require a long period of soil forming conditions from some hundreds of years and up to thousands and even millions of years (Birkeland 1984). 2) The second origin of red colours in palaeosols is related to diagenetic processes (Retallack 1991). It is not possible to distinguish hematite formed by diagenetic processes from those formed on the surface (Retallack 1991).

Precipitation of both hematite and goethite is much more likely in warm climate compared to temperate climate, but goethite is preferably formed in humid climate, while hematite usually requires dry conditions (Collinson 1996).

Clay minerals are known to weather in the order kaolinite→smectite→vermiculite→chlorite (Sheldon and Tabor 2009). Kaolinte is as mentioned associated with hot and humid climate, while chlorite usually forms in dry and cool climate (Sheldon and Tabor 2009). Chlorite might also form in marine water during early diagenesis or by replacing smectite or kaolinite during burial (Bjørlykke 1997). Stensland (2012) found both detrital and autogenetic chlorite in palaeosols in the Snadd Formation. Diagenetic origin is considered as most likely for the chlorite in palaeosol number eight because there is no other evidence of cool and dry climate in the De Geerdalen Formation.

High content of quartz in sample 5, 16 and 27 might be related to well drained palaeosols where fine grained material is washed out, and coarse grained material is left behind. According to the paleocatena model from Kraus and Aslan (1999) can this be due to topographic elevation, often associated with proximal channel deposits. Palaeosol number five is found on top of a distributary channel, supporting the theory of a well-drained palaeosol, but closer examination of grain size is required to verify this theory.

Noteworthy is the absence of smectite which is often associated with poorly drained palaeosols undergoing strong seasonal precipitation (Sheldon and Tabor 2009).

9. Discussion

Following topics are discussed: I) Facies distribution in the De Geerdalen Formation II) Distribution of the Isfjorden Member III) Distribution of palaeosols in the De Geerdalen Formation IV) Palaeoclimate and palaeotopography. The theses by Johansen (2016) and Støen (2016) focus primarily on sandstone bodies and delta front deposits. Herein delta plain sediments (FA 3) and palaeosols are the core focus, and in particularly the Isfjorden Member have been given special attention due to its inherent facies differences from the lower part of the De Geerdalen Formation.

All logs referred to in this section can be found in Appendix C and D. Note that when the terms Vertisols, Argillisols, Protosols, Noncalcareous red and green mudstones and Calcrete are used they refer to the soil types as defined in this study.

9.1 Facies distribution in the De Geerdalen Formation

This section deals with the general distribution of facies and facies associations in the De Geerdalen Formation. There is an overall upwards coarsening trend regarded as shallowing of water depth when the delta built out. This trend is noted by several earlier workers (Knarud 1980, Mørk et al. 1982, Riis et al. 2008, Høy and Lundschien 2011, Lundschien et al. 2014, Rød et al. 2014, Johansen 2016, Støen 2016).

All localities visited display evidence of both fluvial-, tide- and wave- dominant processes, but these processes are unevenly distributed throughout the stratigraphy and from east to west (Appendix C). Outcrops on Barentsøya and Edgeøya hold the most prominent distributary channel deposits, both in terms of thickness and lateral continuity. In contrast, sections measured in Agardhdalen and Deltaneset are more wave dominated, with stronger tidal signatures. This is in accordance with the assumption of the progradation of a delta from southeast towards northwest (Riis et al. 2008, Glørstad-Clark 2010, Høy and Lundschien 2011).

The entirety of the De Geerdalen Formation on Wilhelmøya was logged by Johansen (2016) and Støen (2016) (see appendix), and the log overlaps with the Wilhelmøya 15-2 log presented herein. The locality features Distributary channels (FA 3.2), Floodplain deposits (FA 3.1), and Interdistributary areas (FA 3.2). Delta plain (FA 3) is considered is the most proximal facies association defined in this study. Distributary channels are associated with both Delta plain (FA3) - and Shoreface (FA 2) environment. Based on the field observations Wilhelmøya is considered the most proximal (overall) and terrestrial locality visited, which can be regarded as surprising given the northern remote location in comparison to other sections southwards.

One of the signature features of the De Geerdalen Formation is repeated upwards coarsening sequences, regarded as parasequences. This is described in several earlier studies (Mørk et al. 1999a, Rød et al. 2014, Vigran et al. 2014, Johansen 2016, Støen 2016). Parasequences have been attributed to auto-cyclic delta lobe switching within a major delta complex in the De Geerdalen (Knarud 1980) and Snadd formations (Riis et al. 2008).

Palaeosols in the De Geerdalen Formation often terminates in upwards coarsening sequences (i.e. parasequences), suggesting subaerial exposure (eg. Wilhelmøya, Svartnosa, Blanknuten, and Klement`evfjellet), (Appendix C). This is in accordance with the expectation of parasequences becoming gradually more proximal and potentially terrestrial (Johansen 2016, Støen 2016).

9.2 Distribution of the Isfjorden Member

Recognition of the Isfjorden Member is following the criteria of Mørk et al. (1999). Some of the criteria described are also characteristics for the rest of the De Geerdalen Formation, e.g. thin to thick bedded silt-and sandstone beds and carbonate beds (Mørk et al. 1999). The Isfjorden Member is thus mainly recognized by coquina beds and multi-coloured shales, and these features are considered as diagnostic characteristics of the Isfjorden Member.

The Isfjorden Member is seen at all localities where the upper part of the De Geerdalen Formation has not been eroded (Wilhelmøya, Hellwaldfjellet, Hahnfjella, Teistberget, Klement`evfjellet, Šmidtberget, Friedrichfjellet and Deltaneset). The time equivalent Hopen Member, defined by Lord et al. (2014), is not present at any of the localities, despite previous inferences that it may be present in the eastern areas of Svalbard. The transition between these two member members is likely not exposed, since the upper part of the De Geerdalen Formation on Edgeøya is eroded (Lord et al. 2014).

In this study most of the Isfjorden Member is interpreted as representing Delta plain (FA 3), which is considered as the most proximal deposits in the study area. Paralic environments ranging from shallow marine to lower delta plain with some distributary channels is suggested. Coquina beds (Facies L) is possibly representing lagoonal environments. No sedimentary structures are observed in the coquina bed. The interpretation is mainly in accordance with Mørk et al. (1999) which suggest shallow marine and possibly lagoonal environments (Mørk et al. 1999). The observation of fine grained sediments alternating with thin sandstone beds and coquina beds can support such a setting. Presence of distributary channels, coals and palaeosols suggest delta/coastal plain in a more proximal and terrestrial setting. The Isfjorden Member is considered as the upper, most proximal deposits in a large prograding delta complex.

Ryseth (2014) show that the subsidence rates decreased in the Upper Triassic in the south-west Barents Sea which gave an upward decrease in accommodation space. Similar setting is also suggested in Svalbard thus probably account for more terrestrial deposited Isfjorden Member on Spitsbergen.

9.2.1 Wilhelmøya

The presence of the Isfjorden Member on Wilhelmøya has not been conclusively presented in earlier literature or has been proposed indirectly. According to Pchelina (1983) was her defined Isfjorden “formation” present on Wilhelmøya. However this definition included a larger part of the De Geerdalen Formation. The presence of the Isfjorden Member on Wilhelmøya, as defined in Mørk et al. (1999), was thus not certain. Lord et al. (2014a) inferred the potential presence of the Hopen Member as opposed to the Isfjorden Member on Wilhelmøya, hypothesising that a lateral facies change may occur across the Lomfjorden and Storfjorden fault zones (Gareth Lord pers. comm. 2016).

The Slottet Bed is excellent exposed on Wilhelmøya (Vigran et al. 2014), and the observations in this study are in accordance with the definition in Mørk et al. (1999) and base Slottet Bed defines the upper boundary of the Isfjorden Member (Figure 9.1). Here the Isfjorden Member show several of alternating red and green mudstones and coquina beds consistent with the definition of the Isfjorden Member (Mørk et al. 1999). Thus the presence of the Isfjorden Member on Wilhelmøya can be considered verified with a high certainty in this study.

The presence of several distributary channels, coals and palaeosols might imply a slightly more proximal and terrestrial setting on Wilhelmøya, or the onset of a more northern provenance source in the Norian, not seen to the south in the Hopen area and in the west in the Deltaneset region

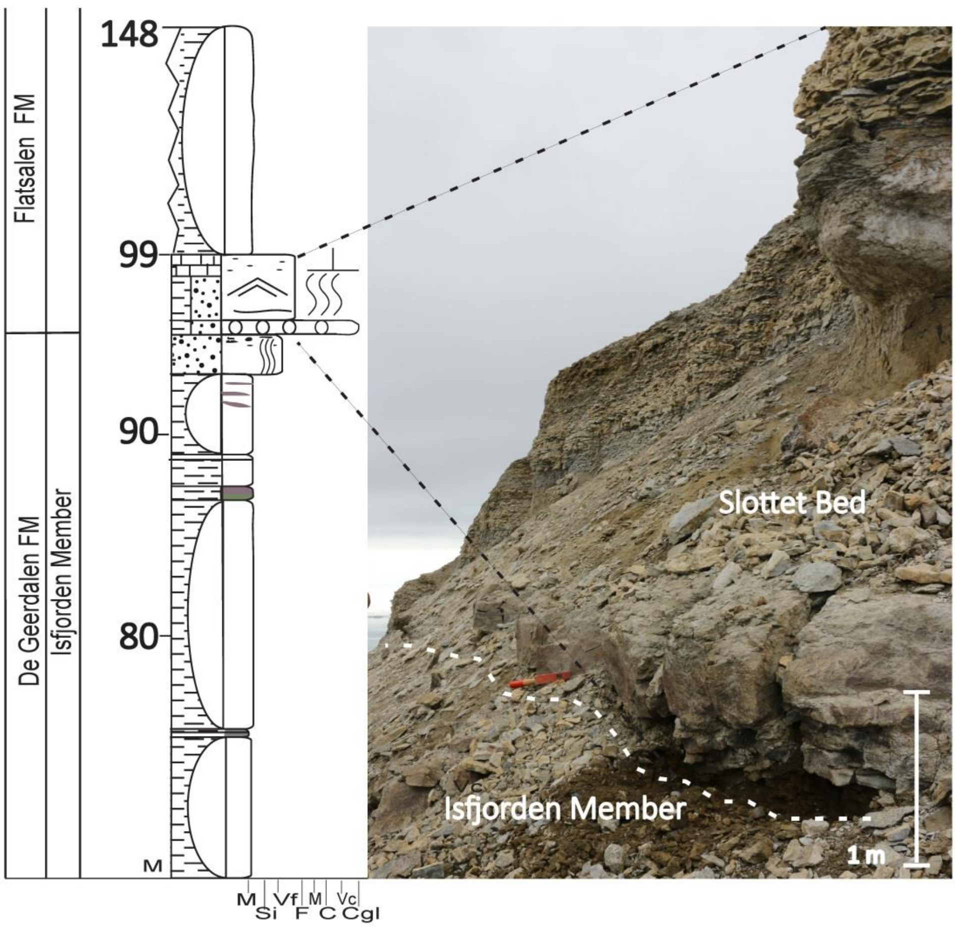


Figure 9.1: The four metre thick Slottet Bed exposed on Wilhelmøya.

9.2.2 Hellwaldfjellet, Hahnfjella and Teistberget

As the field observations on Hellwaldfjellet suffered from extensive scree cover, the interpretations in this section must be considered with some caution. However, interpretations point towards primarily delta top associations (FA 3) in these areas, with possibly three distributary channels being present. Upper shoreface deposits (FA 2.1) occur in the lower part of the section.

Two beds with shell fragments are observed on Hellwaldfjellet between 19 and 27 meters on the Hellwaldfjellet 15-2 log (Appendix C). A 50 cm thick heavily cemented sand- and siltstone bed with some shell fragments is found at the base of this unit. Weathering colour is orange, but the rock is grey in fresh cut. Thereby a 5 meter thick scree cover interval follows before a 120 cm thick sandstone unit starts at 25 meters. The unit has low angle cross bedded sandstone (Facies E) the first 40 cm and then turns into small scale cross stratified sandstone (Facies B). Shell fragments are found throughout the unit, but are most abundant from 50-80 cm, where a coquina bed is seen (Figure 9.2 B). Whether the lowermost or uppermost shell deposit is the first coquina bed is a topic open for discussion and interpretation. The lowermost outcrop only has some shell fragments, and has grey colour in fresh cut. As the lowermost outcrop does not fit completely with the definition of coquina beds in this thesis, defined as cemented sandstone units with red to orange colour consisting mainly of shell fragments, the upper outcrop is considered the first coquina bed. This also partly fits with the definition of the lower boundary of the Isfjorden Member with a coquina bed occurring over a thick large scale cross stratified sandstone unit (Mørk et al. 1999). The definition of "thick" is not accurate, but the coquina bed is overlying a 50 cm thick sandstone bed. Vertical burrows can indicate high energy environments or relative rapid sediment deposition, but still low enough energy and sediment rates that organisms could exist there. Relatively high energy environment combined with low angle cross stratified sandstone (Facies E) suggest a shallow marine shoreface setting (FA 2.1).

The Slottet Bed might be present on Hellwaldfjellet. From 114 to 118 on the Hellwaldfjellet 15-2 log carbonate rich scree with red to orange weathering colour and intense bioturbation is found. From 118 to 132 metres an intrusion is penetrating the rocks. The scree from 114 - 118 metres share many of the features with the Slottet Bed observed at Wilhelmøya: 1) carbonate rich 2) red to orange colour 3) intense bioturbated. Keeping the similarities in mind, the scree might originate from the Slottet Bed, and the bed is possibly hidden under the scree cover if the scree is not long transported. Bones are found in the Flatsalen Formation on Wilhelmøya (Vigran et al. 2014) and Hopen (Mørk et al. 2013), indicating that the upper scree belongs to the Flatsalen Formation.

The observed thickness of the Isfjorden Member on Hellwaldfjellet is 88 meters if the interpretations of the lower base of the Isfjorden Member and the Slottet Bed are correct. This would be in the range of earlier observed thickness of the member (55 to 135 meters) according to Mørk et al. (1999). None alternating red or green mudstones (Facies O_2) are seen, but this is considered as most likely due to scree cover. According to the Lithostratigraphic Lexicon of Svalbard (Mørk et al. 1999) is the De Geerdalen Formation 230 meters thick in central and eastern Spitsbergen and up to 400 metres on Barentsøya and Edgeøya. Johansen (2016) and Støen (2016) measured a total thickness of 238 meters at Wilhelmøya. They found a total thickness of

260 metres at Hellwaldfjellet, which would be within a likely range for expected thickness on Hellwaldfjellet. Based on all these observations it is likely that the end of the log at least is at the upper part of the De Geerdalen Formation, and possibly at the very top with the Slottet Bed covered by scree some meters below the dolerite sill.

On Teistberget the thickness of the exposed Isfjorden Member is 50 meters, but the lower boundary of the member as defined in Mørk et al. (1999) is not seen in the logged section (Appendix C). The first sign of the Isfjorden Member is red and green mudstones (Facies O_2) at 17 meters on the Teistberget 15-2 log. The only coquina bed in the section is seen 33 meters after the first evidence of the Isfjorden Member. Calcrete, or at least calcareous nodules are present on Teistberget (Appendix A).

105 meters of the Isfjorden Member is well exposed on Hahnfjella, the original type section for Pchelina's equivalent Isfjorden 'suite'. Almost the entirety of the Isfjorden Member is most likely present, however no evidence for the Slottet Bed was observed. The base of the Isfjorden Member at this section is a coquina bed overlying a sandstone unit which fits well with the definition in Mørk et al. (1999). Another coquina bed is also seen at 182 meters.

9.2.3 Agardhdalen

The Isfjorden Member is present at Klement`evfjellet, but unfortunately poorly exposed in the logged section in this study. The presence of the Isfjorden Member is verified by Johansen (2016) and Støen (2016) who logged neighbouring ridges west of the section of the Klement`evfjellet 15-1 log. The log of Knarud (1980) reprinted in Vigran et al. (2014) documents the exposure of the Knorringfjellet Formation starting out with the possibly exposed Slottet Bed. The Klement`evfjellet 15-1 log is situated East of the logged section of Knarud (1980), but the log stops some few meters under the correlating stratigraphic level to the Slottet Bed as described in Vigran et al. (2014).

The red and green mudstones (Facies O_2) of the Isfjorden Member are together with the Slottet Bed exposed on Friedrichfjellet and Šmidtberget. On Šmidtberget 54 meters of the Isfjorden Member is exposed. The first evidence of the Isfjorden Member on this locality is a red noncalcareous mudstone at 83 meters (Appendix D).

The first evidence of the Isfjorden Member on Friedrichfjellet is a coquina bed at 165 meters (appendix C). This fits fairly with the definition of the lower base (Mørk et al. 1999), even though the underlying sandstone is not cross stratified. The thickness of the observed Isfjorden Member in the logged section is only 36 meters. Keeping in mind that the Isfjorden is thicker at both the neighbouring Šmidtberget and all other localities it is regarded as likely that not the whole Isfjorden Member is exposed at the logged section.

9.2.4 Deltaneset

The presence of Isfjorden Member on Deltaneset is well documented in several studies (Eriksen 2012, Husteli et al. 2014, Olaussen et al. 2015). The entire logged section in this study is within the Isfjorden Member. The section differs from other localities by the thick and well developed calcrete not seen elsewhere in this study. The well-developed calcrete is verified by petrographic thin section observation (Chapter 8.3) and is regarded as evidence of at least partly terrestrial environments with stable soil forming conditions. Overall depositional environments on Deltaneset are interpreted as Interdistributary area (FA 3.2) and Shoreface (FA 2.1). In addition a 2 metre thick sandstone body which laterally pinch out after 10m is interpreted as Distributary channel (FA 2.3).

9.2.5 The lower boundary of the Isfjorden Member

The lower boundary of the Isfjorden Member at the base of a siltstone coquina bed above a thick cross bedded sandstone unit as defined in Mørk et al. (1999) is only observed on Hahnfjella. The lack of observed lower boundary is explained by 1) the lower boundary is not exposed, or 2) the lower boundary varies depending on the locality. Both explanations are regarded as likely. Keeping in mind that delta lobe switches would probably give significant lateral variations, it is considered as likely that the lower boundary is not prominent and traceable in a region scale. Lagoons might be common on a stratigraphic level in a region scale, but still individual localities may represent local depositional environments, with delta Distributary channels (FA 2.3), Delta plain (FA 3) and Delta front (FA 2) deposits between.

The question is if the Isfjorden Member deserve its status as member, or if it should be incorporated in the rest of the de Geerdalen Formation. Two arguments favour the latter suggestion 1) the Isfjorden Member probably represent the gradual development of a prograding delta 2) it might not be possible to define a traceable lower boundary of the Isfjorden Member on a regional scale. However, the Isfjorden Member is easy to recognize at different localities, especially by its multi-coloured shales and coquina beds, although pointing out the lower boundary has turned out challenging. The status of the Isfjorden Member can thus be considered reasonable as it is useful as a reference to the upper De Geerdalen Formation with easily recognized characteristics.

9.3 Distribution of palaeosols in the De Geerdalen Formation

Based on facies analyses presented in Chapter six and seven the distribution of palaeosols is discussed in light of the overall depositional environment of the De Geerdalen Formation.

Vertisols, Protosols and Argillisols (Facies O_1) are found both below and within the Isfjorden Member. In general Facies O_1 occur in a wide range of palaeoenvironments including Distributary channels (FA 2.3) (Fig 9.4D), Barrier Bars (FA 2.1) (Fig.9.4B), Floodplains (FA 3.1) (Fig. 9.2A) and Interdistributary areas (FA 3.2). The number, thickness and maturity of yellow and brown palaeosols tend to increase upwards in the De Geerdalen Formation.

Noncalcareous red and green mudstones and calcrete are restricted to the Isfjorden Member and are mostly associated with Floodplains (FA 3.1) and Interdistributary areas (FA 3.2). These palaeosol types are both thicker and more densely spaced than the other palaeosol types.

9.3.1 Distribution of Vertisols, Argillisols and Protosols (Facies O_1)

These palaeosol types are found from the middle parts of the De Geerdalen Formation and up, but are most abundant right under, and in the first half part of the Isfjorden Member. Immature palaeosols are distributed relatively evenly in the stratigraphic level of palaeosols. Argillisols are more frequently observed within the Isfjorden Member compared to underlying units (Fig. 9.2).

The maturity and complexity of yellow and brown palaeosols (Facies O_2) thus seams to increase upwards in the De Geerdalen Formation.

Conditions favourable for soil forming include subaerial exposure with minor sedimentation and erosion rates (Kraus 1999). The soil forming processes are enhanced by biologic activity causing reworking of the soil by roots, worms, and microorganism. In addition is exposure time crucial in forming of well-developed soil horizons (Wright 1992). All these requirements are expected to become more dominant upwards in a prograding delta.

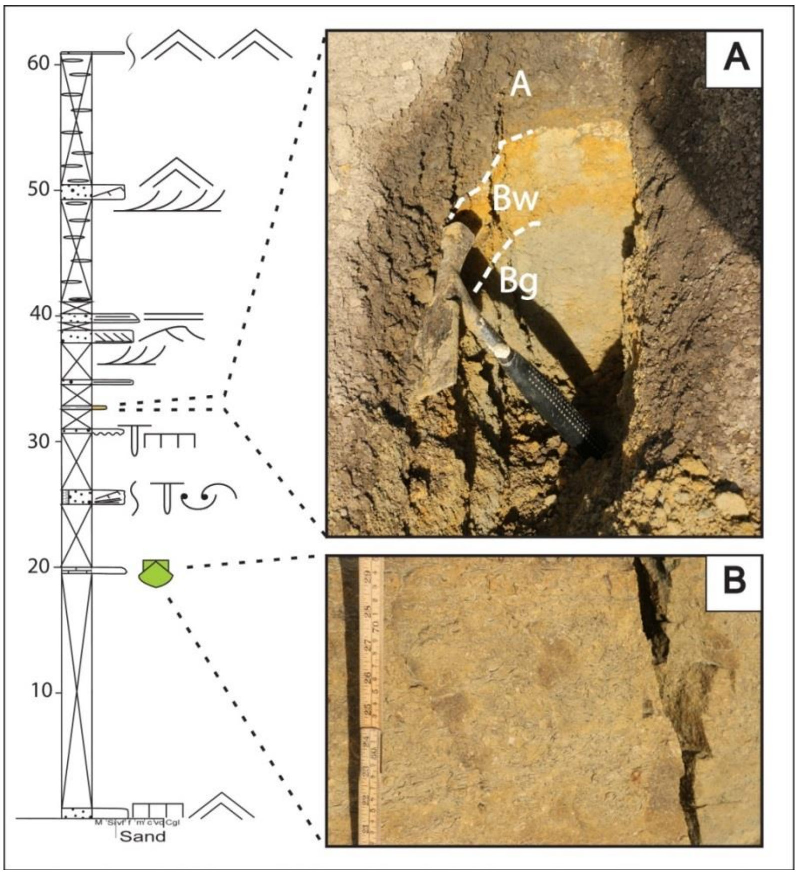


Figure 9.2: A) Example of horizonated palaeosol (No. 10) on Floodplain (FA 3.1) within the Isfjorden Member at 32.5 meters on the Hellwaldfjellet15-2 log. Soil horizons are interpreted as A-horizon (A), weathered B horizon (Bw) and gleyed B horizon (Bg). Gleyed soil imply water saturated ground (Kraus and Allan 1999). B) Coquina bed found at 25.5 meters considered as the base of the Isfjorden Member on Hellwaldfjellet.

As mentioned in Chapter 9.1 Wilhelmøya appears as the most proximal locality visited based on the overall interpretation of the outcrops. The impression of a proximal setting is supported by the presence of one of the best developed palaeosols found in this study (Fig. 9.3) (Appendix A, palaeosol No. 0). The palaeosol is defined as Argillisol, and is found 90 meters above the base of the De Geerdalen Formation in oppose to other Argillisols which occur higher in the stratigraphy at other localities (Appendix A). The palaeosol is one meter thick with well-developed soil horizons and mottles. The mature nature is supported by the presence of wood fragments with a length of up to 10 cm found in situ in the O-horizon of the palaeosol, implying growth of trees (Fig. 9.3B). Trees are considered to mainly occur in relative mature soils as smaller plants usually establish first and prepare the soil for growth of higher vegetation (Alonso-Zarza and Wright 2010). As mentioned in Chapter 8.2.2 the complexity and development of soil horizons increases with time (Retallack 1997). The bleached layer interpreted as E-horizon probably show wash out of minerals, organic matter and clay, implying long lasting conditions for soil forming. This process is considered to require up to several thousand years (Retallack 1997).

Figure 9.3. A) Argillisol at Wilhelmøya with well-developed O-, E and B- horizons. B) Wood fragments found in situ in the O-horizon in the palaeosol in figure A. Hammer for scale.

Keeping in mind the position of Wilhelmøya as the northernmost locality in this study, it would together with Deltaneset be expected as the most distal portions of the progradation of the coastline from southeast (Glørstad-Clark 2010, Lundschien et al. 2014, Klausen et al. 2015). The observations and interpretations of facies and palaeosols done in this study do not fit with the expectation of Wilhelmøya as one of the most distal localities. This can possibly be explained by an additional source in north, as suggested by several authors (Rønnevik et al. 1982, Mørk et al. 1989, Nøttvedt et al. 1993, Skjold et al. 1998, van Veen et al. 1993, Mørk 1999, Harstad 2016) but further investigations are needed to verify this.

Vertisols are as mentioned in Chapter 8.2.1 often terminating sandstone outcrops interpreted as Barrier bar (FA 2.1) and Distributary channel (FA 2.3). This is e.g. seen on Blanknuten (Figure 9.4). In addition they are often seen on Floodplain deposits (FA 3.1). Vertisols in Interdistributary areas (FA3.2) is uncommon. There is a significant shift from vertisol under the Isfjorden Member and Argillisols in the member. As described in Chapter 8.2.1 might the homogeneous appearance possibly be due to alternating moisture regime in the soil. This can possibly be due to seasonal rainfall in the De Geerdalen Formation under the Isfjorden Member as also suggested by Enga (2015). Vertisols in the De Geerdalen are often found on top of Distributary channels (FA2.3). Soils formed on top of channels are according to the Paleocatena model (Kraus and Aslan 1999) likely to be poorly developed compared to soils located further away. This might be a second for the poorly developed soil horizons in this soil type. However, high organic content found in many of these palaeosols points towards plant growth. Roots seen in figure 9.4C verify the pedogenetic origin of the soil.

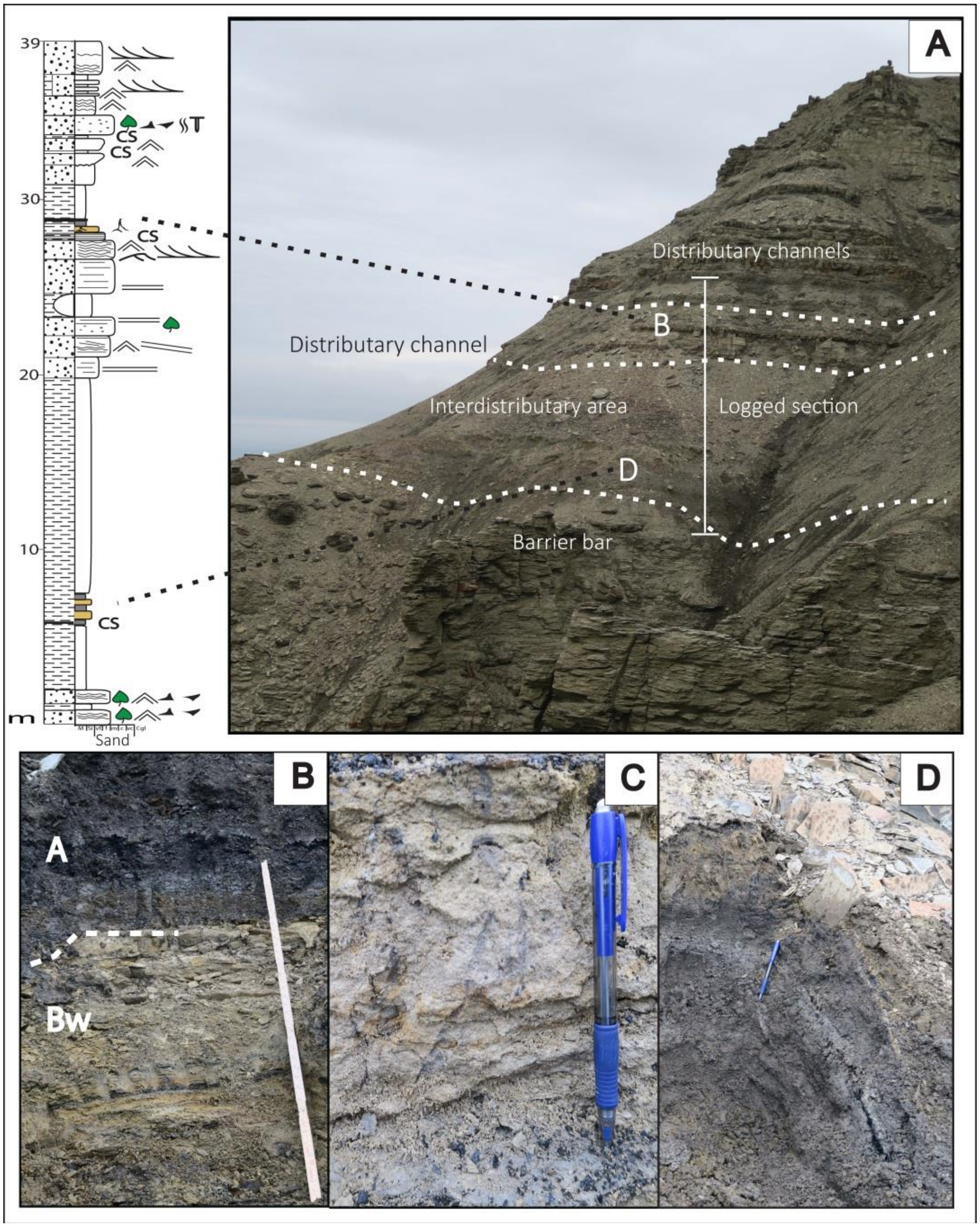


Figure 9.4: *Vertisols on Blanknuten.* **A)** The neighbour ridge of the logged section of Blanknuten 15-2. Palaeosols in picture B and Dare marked. B) Dark, homogeneous and organic rich soil interpreted as vertisol on top of distributary channel (No. 37, Appendix A). C) Coalified roots from the upper part of the Bw-horizon soil in picture B. D) Possibly vertisol with organic content and coal fragments on top of Barrier Bar deposits (No. 36, Appendix A).

Protosols are found at all levels where palaeosols occur in the De Geerdalen Formation, and all localities except Deltaneset. In this study this is considered to mirror the wide range protosols can form in (Retallack 2001). As mentioned in Chapter 3 Cecil and Dulong (2003) found that forming of immature soils are independent on moisture regime. Protosols are also the far most common soil type of the brown and yellow palaeosols. The frequent occurrence of this soil type might reflect the dynamic nature of a delta. Soil forming is not favourable if the sedimentation rates are too high or during erosion (Kraus 1999). Both scenarios are expected on a delta, possibly explaining numerous immature palaeosols.

9.3.2 Distribution of Non calcareous red and green mudstones and Calcrete (Facies O_2)

As mentioned earlier Alternating red and green mudstones (Facies O_2) are restricted to the Isfjorden Member. They are present at Wilhelmøya, Hahnfjella, Teistberget, Klement`evfjellet, Šmidtberget, Friedrichfjellet and Deltaneset (Appendix A). The colour of these palaeosols is as stated by Pchelina (1983) striking, and easily recognized in the different localities. The red and green palaeosols (Facies O_2) are in general thicker than the brown and yellow palaeosols (Facies O_1). This might be related to the higher stratigraphic level and less sediment input than for the underlying units, causing longer lasting soil forming conditions.

Calcrete is observed on Deltaneset, and probably on Šmidtberget, Friedrichfjellet and Teistberget, but is absent at the other localities. The calcrete at Deltaneset is thicker and appear as better developed than on Teistberget, Friedrichfjellet and Šmidtberget. Non calcareous red and green mudstones are seen at all localities where the Isfjorden Member is present, except Hellwaldfjellet.

9.4 Palaeoclimate and palaeo geographic implications in the Isfjorden Member

XRD analyses showed that goethite was present in the red and green palaeosol from Wilhelmøya, opposed to XRD-analyses from palaeosols displaying other colours (Table 8.1). Hematite is not present in any of the samples, but goethite transformed to hematite is considered as likely explanation of the red colours in the Isfjorden Member. The interpretation is supported by the presence of goethite and hematite in red palaeosols in the Snadd Formation in the Barents Sea (Enga 2015), and by the fact that very small amounts of hematite is needed to get a distinct red colour (Duchaufour 1982).

The Noncalcareous red and green mudstones and calcrete differs from each other by different carbonate content. Nevertheless the striking alternating red and green colours found in both palaeosol types points towards a somehow genetic relation between the two types. Alternating colours due to changes in redox conditions are described from Triassic calcrete in Denmark (Weibel 1998). The fluctuations in reducing and oxidation conditions are explained by mainly oxidising environment interrupted by locally reducing conditions. Reducing conditions might have been caused by oxygen consuming by decay of organic matter deposited together with the sediments (Weibel 1998). Stensland (2012) suggest fluctuations in ground water table related to larger scale autocyclic switching of lobes as explanation for the alternating colours in the Snadd Formation. Ground water is usually depleted in oxygen, leading to reducing conditions

(Duchaufour 1982). Low water table hence favour oxidation conditions, whereas high water table might give reducing conditions.

The Deltaneset section differs from the other localities by the thick and relatively well developed calcrete not seen elsewhere in the De Geerdalen Formation, even though calcrete probably is present in Agardhdalen and on Teistberget. Stensland (2012) also reported the presence of calcrete in the Snadd Formation in the Barents Sea. Forming of calcrete requires stable conditions with low sedimentation rates for a long time (Wright 2007), and is thus in accordance with a distal position with low sedimentation rates.

Calcrete is most common in arid to semi-arid climate (Alonso-Zarza and Wright 2010), and the presence of calcrete together with absence of coal on Deltaneset points towards dryer conditions than eastern localities and underlying units. A mean annual rainfall of 100 to 500 mm is typical for calcrete forming (Goudie 1983), but calcrete formed in areas with up to 1000 mm per year is reported (Mack and James 1994). In contrast calcrete is not observed on Wilhelmøya, but thin coal seams do occur within the Isfjorden Member there. Coal is in general associated with humid to semi-humid climate (Nichols 2009). In addition, XRD-analyses of a Noncalcareous red and green mudstone of the Isfjorden Member (Palaeosol number 5) show that kaolinite and goethite are dominating clay minerals. Both kaolinite and goethite are indicative of humid conditions. Kaolinite is usually dominating clay mineral in areas with seasonal precipitation between 1000 and 2000 mm (Retallack 2001). Sheldon and Tabor (2009) show that kaolinite soil profiles can probably also form in cooler temperate climates with seasonal growth during warm months.

It seems like there might be lateral semi regional variations in palaeomoisture conditions within the Isfjorden Member, but if the seasonal rainfall was around 1000 mm per year, the difference between the localities is not necessarily striking.

Occurrence of common calcrete soil profiles in the Isfjorden Member compared to the lower parts of the De Geerdalen Formation might suggest somewhat dryer climate in the upper part of the De Geerdalen Formation. But alternatively, and based on the fact that 1) thin coal seams are present in the Isfjorden Member on Wilhelmøya, 2) calcrete is restricted to Deltaneset, Friedrichfjellet, Šmidtberget and Teistberget and 3) red shale of the Isfjorden Member on Wilhelmøya contains goetite which gives red colour and indicates humid climate the change in climate might have been minor. This is in contrast to suggestions by Olaussen et al. (2015), which suggest a major change in climate.

Absent of calcrete on the other localities, leads to the interpretation of palaeotopography within the Isfjorden Member as driving mechanism rather than palaeoclimate. Possible explanations might be local elevated terrain with better drained palaeosols as for example on Deltaneset compared to the East Svalbard. Consistent with this concept is the overlying thin condensed Norian to Bathonian Wilhelmøya Subgroup with several lacunas on Spitsbergen. This is in contrast to the Wilhelmøya Subgroup on Wilhelmøya, Hellwaldfjellet and Hopen, and points towards a structural palaeohigh in central and west Spitsbergen which probably initiated during deposition of the Isfjorden Member. Probably detailed isotopic work on the calcretes in the De Geerdalen might better solve major changes in climate, if any, during deposition of the De Geerdalen Formation (C.f. Alonso-Zarza & Tanner 2006).

10. Conclusions

- The De Geerdalen Formation in the Storfjorden area of Svalbard was deposited in paralic shallow marine environments in the distal end of a large delta complex. Upwards coarsening units interpreted as parasequences are characteristic for the De Geerdalen Formation. The delta was probably sourced from the Uralids in the east-southeast as suggested by several authors, but there might have been an additional source in the north.
- Palaeosols are often found capping distributary channels, barrier bars, floodplains and interdistributary areas.
- The number, thickness and maturity of palaeosols tend to increase upwards in the De Geerdalen Formation suggesting slowing down of accommodation space.
- The Isfjorden Member is easy to recognize by alternating red and green mudstones and coquina beds, but the lower boundary differs depending on facies associations and locality.
- The Isfjorden Memoer is present at Wilhelmøya, Hellwaldfjellet, Hahnfjella, Teistberget, Klement'evfjellet, Friedrichfjellet, Šmidtberget and Deltaneset.
- The Isfjorden Member is not present at Krefftberget, Blanknuten and Svartnosa, probably due to erosion of the upper De Geerdalen Formation.
- There are significant differences in the measured thickness of the Isfjorden Member, but these remain within the range of 55 to 135 meters as stated in Mørk et al. (1999).
- The mode of formation for the alternating green and red mudstones common to the Isfjorden Member is possibly related to fluctuations in the water table and alternating oxidation and anoxic conditions. Red colour is assigned to oxic iron, while green colour is a result of reduced iron.
- The calcrete in the Isfjorden Member is unevenly distributed, with thicker and better developed calcrete on Deltaneset compared to Šmidtberget, Friedrichfjellet and Teistberget. No calcrete is observed at the remaining localities. This observation is regarded as a likely general trend, but the accurate distribution is considered uncertain due to scree cover and lack of petrographic thin sections.
- Facies association Interdistributary area (FA 3.2) is most frequently observed in the upper reaches of the De Geerdalen Formation within the Isfjorden Member, implying less sedimentation rates compared to sections under the Isfjorden Member.
- There is a significant change from Vertisols, Protosols and Argillisols dominating the middle parts of the De Geerdalen Formation, to the Noncalcareous red and green mudstones and Calcrete restricted to the Isfjorden Member. Calcrete might imply a minor shift in climate from warm and semi-humid with seasonal rainfall to warm and semi-arid. The formation of calcrete might also be related to better drained soil in the west, possibly due to uplift.
- The palaeosols in the De Geerdalen Formation are in general poorly developed, indicating a dynamic regime not favourable for forming of well-developed palaeosols. Palaeosols in the study area seems to be less developed in terms of thickness compared to southern Edgeøya and Hopen (Enga 2015), and the Barents Sea (Stensland 2012, Enga 2015).
- The palaeosols of the De Geerdalen Formation often lack obvious roots, indicating poor conditions for plant growth.

11. Suggestions for further work

- This study provides a brief overview of the distribution of Noncalcareous red and green mudstones and Calcrete. More detailed investigation of the distribution of these soil types together with XRD-analyses and petrographic thin sections of these palaeosols can contribute to a better understanding of the palaeoclimate of the Isfjorden Member. Samples taken during field work in this study could be used in these analyses.
- The area on central Spitsbergen between Deltaneset and Agardhdalen has not been visited in this study. Investigations of outcrops between the two localities could improve the understanding of the transition between the extensive distribution of Calcrete at Deltaneset contra mostly Noncalcareous red and green mudstones in Agardhdalen.
- Investigations of lateral variations in palaeosols and delta plain sediments could provide a better understanding of the dynamic of the prograding delta when the De Geerdalen Formation was deposited.
- Isotopic analyses on calcretes might improve the understanding of the palaeoclimate in the De Geerdalen Formation.

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Appendix A - Palaeosols in the De Geerdalen Formation

Overview of all sediments interpreted as palaeosols in the De Geerdalen Formation. The field recognition of palaeosols is based on soil horizons, soil structure and evidence of roots.Palaeosols under the Isfjorden Member are marked with light yellow background colour. Palaeosols within the Isfjorden Member have light red background colour. The third column refers to meters on logs found in Appendix C and D. The appendix where the logs are attached is indicated with letters behind the name of the logs.

	Field data				**Field observations**			
No	**Log/ sample**	**Log (m)**	**Thick-ness**	**Reaction HCl**	**Soil horizon**	**Soil structure**	**Roots**	**Classification**
	Wilhelm 15-1							
0	No sample	89	100 cm	None	**O-horizon:** 20 cm coal shale with wodd fragments up to 10 cm. **E-horizon:** 60 cm bleached pale grey layer. **Bt-horizon:** 10 cm enriched brown layer	Fine granular	Mottles	Argillisol on Distributary channel (FA 2.3).
	Wilhelm 15-2							
1	Tum 15.2.3.C	19	50 cm	None	**A-horizon:** 20 cm. High organic content. **Bt-horizon:** 30 cm. Dark brown. Enriched in clay.	Medium angular blocky	None	Vertisol on Floodplain (FA 3.1)
2	Tum 15.2.5.C	30	55 cm	None	**A-horizon:** 10 cm. Very organic rich black shale. **Bw horizon:** 40 cm: brown to yellow silty clay. **Base:** 5 cm grey, possibly gleyed.	Fine subangular blocky	None	Vertisol on Floodplain (FA 3.1)
3	No sample	39	30 cm	None	**C-horizon:** 30 cm. Weathered, brown bedrock	Medium granular	None	Protosol on Floodplain (FA 3.1)
4	No sample	42	320 cm	Not tested,	Several layers of red and green shales with two thin mottled grey and yellow clay layers.	Medium granular	None	Noncalcareous red and green mudstone in Interdistributary area (FA 3.2)
5	Tum 15.2.10C **XRD**	66	70 cm	None	Stacked palaeosol with two A horizons consisting of coal shale with root traces. **Bw-horizons**: Brown and weathered.	Fine granular	None	Protosol on top of Distributary channel (FA 2.3)
6	No sample	75	40 cm	None	Black shale with thin layer of coal shale in the middle (5 cm).	Thin platy	None	Protosol in Interdistributary area (FA 3.1)
7	Tum 15.2.11. C	87	50 cm	None	**Top 25 cm:** red Mudstone Lower 25 cm: green Mudstone	Medium angular blocky	None	Noncalcareous red and green mudstones in Inter-distributary area (FA 3.2)
8	Tum 15.2.17. C **XRD**	89	105 cm	None	**Top 75 cm:** red Mudstone **Lower 30 cm:** green Mudstone	Medium angular blocky	None	Noncalcareous red and green mudstone in Interdistributary area (FA 3.2)
9	No sample	91-93	-------	None	Red clay in scree. Not in situ.	________	None	Noncalcareous red mudstone.

No	Log/ sample	Log (m)	Thick-ness	Reaction HCl	Soil horizon	Soil structure	Roots	Classification
	Hellwald -fjellet							
10	Hell 15.2.5	32.5	40 cm	None	**A-horizon:** 15 cm. Organic rich shale. **Bw-horizon:** 10 cm. Yellow to brown silty clay with mottles. **Bg-horizon:** 10 cm. Grey clay, possibly gleyed. Sticky texture.	Fine granular	Mottles	Argillisol on Floodplain on the lower delta plain (FA 3.1).
11	No sample	112.5	30 cm	None	**Top 12 cm:** organic rich shale **Lower 18 cm:** Grey clay with yellow mottles	Fine granular	Mottles	Protosol on Floodplain (FA 3.1).
	Svart-nosa							
12	Svart 15.2.6.C	21	80 cm	None	**O-horizon:** up to 5 cm thick coal seam. **E-horizon:** 40 cm of grey/brown /yellow layered silty clay. The layer displays a slightly lighter colour upwards. Mottles **Bw-horizon:** 35 cm of brown more consolidated silty shale. Yellow/bleached irregular layer at the upper 2-3 cm.	Fine granular	Mottles	Protosol on Flood plain on the lower delta plain (FA 3.1).
	Hahn-fjella							
13 A	**No sample**	49	30 cm	None	**A-horizon:** 10 cm black, organic rich shale. **E-horizon:** 10 cm yellow to brown silt. Mottles **B-horizon:**10 cm brown silt/sand with concretions.	Fine granular	Mottles	Protosol in Interdistributary area on the lower delta plain (FA 3.2)
13 B	No sample	126	50 cm	None	Appears homogenous (no clear layering) except a five cm thick coal seam in the middle.	Fine angular blocky.	None	Vertisol in Interdistributary area (FA 3.2)
14	Hahn 15.2.24C	176	40 cm	None	**O-horizon:** 20 cm. coal shale. **B-horizon:** 30 cm. brown silt	C: Very fine granular	Mottles	Protosol in Interdistributary areas (FA 3.2)
15	Hahn 15.2.27C	185	120 cm	None	Green mudstone	Not defined	None	Noncalcareous green mudstone
16	Hahn 15.2.32. C **XRD**	210	30 cm	None	**A-Horizon:** 10 cm organic rich shale. **Bt-horizon:** 20 cm brown and grey silt with mottles	Fine granular	Mottles in Bt horizon	Protosol on Interdistributary area (FA 3.2)
17	Hahn 15.2.35C	222	20 cm	None	20 cm of organic rich shale	Thin platy	None	Vertisol in Interdistributary area (FA 3.2)
18	Hahn 15.2.36C	225	250 cm	None	Green mudstone	Coarse angular to blocky	None	Noncalcareous green mudstone in interdistributary area (FA 3.2)
19	No sample	232.5	80 cm	None	Red mudstone	Not defined	Not observed	Noncalcareous red mudstone in interdistributary area (FA 3.2)

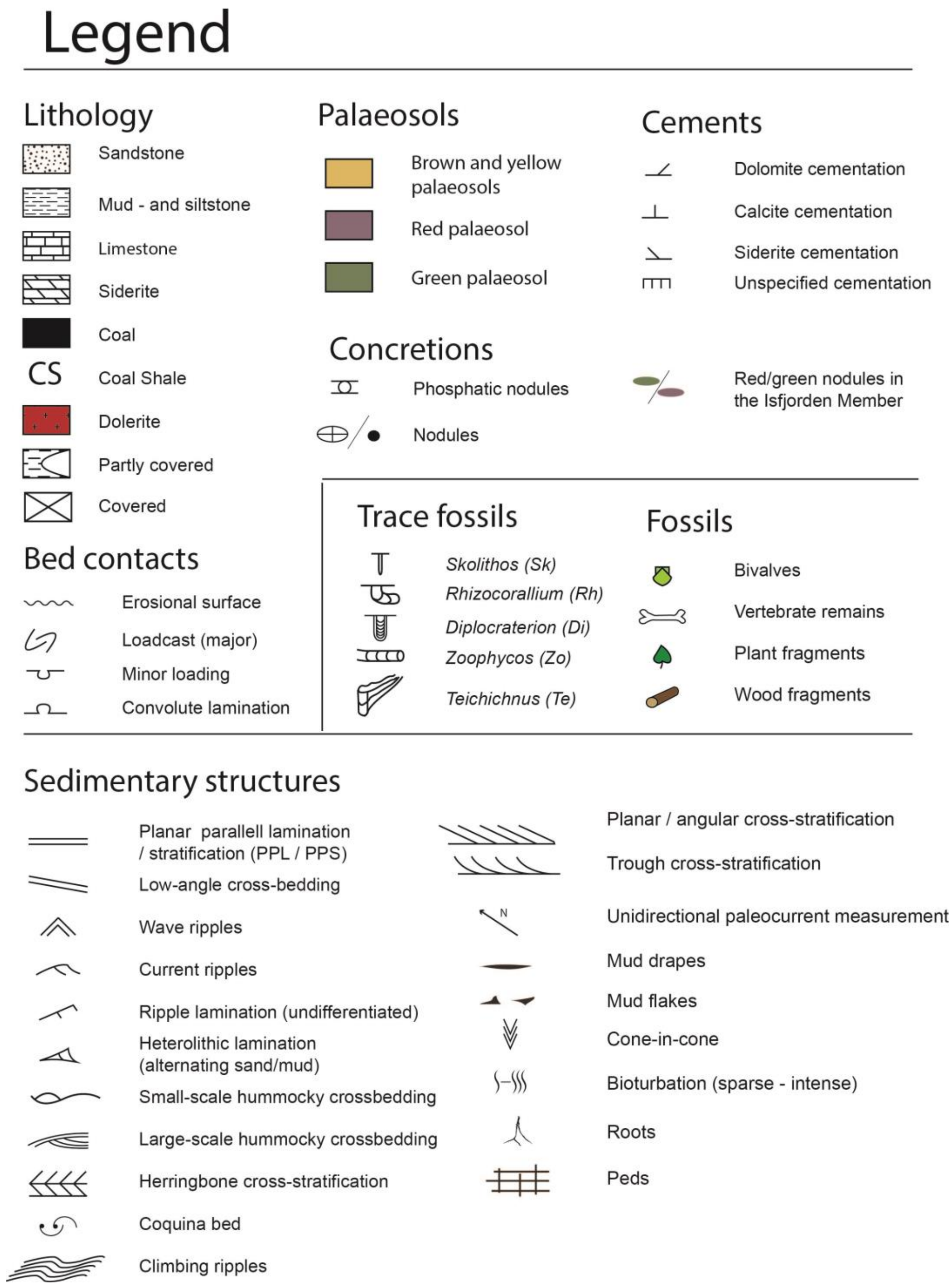
No	Log/ sample	Log (m)	Thick-ness	Reaction HCl	Soil horizon	Soil structure	Roots	Classification
20	Hahn 15.2.38B	233.8	140 cm	None	Alternating red and green mudstone	Not defined	Not observed	Noncalcareous red and green mudstone in Interdistributary area (FA 3.2)
21	No sample	244	150 cm	Scree cover	Alternating red and green mudstone. Inferred from surface colour of scree.	Scree cover	Scree cover	Noncalcareous red and green mudstone in Inter-distributary area (FA 3.2)
22	No sample	246	300 cm	Scree cover	Alternating red and green mudstone. Inferred from surface colour of scree.	Scree cover	Scree cover	Noncalcareous red and green mudstone in Inter-distributary area (FA 3.2)
23	No sample	251.5	200 cm	None	Alternating red and green mudstone	Not defined	Not obser-ved	Noncalcareous red and green mudstone in interdistributary area (FA 3.2)
	Teist-berget							
24	Teist 15.2.6.C: .	17.5	490 cm	Strong 19-22 m.	Alternating red and green mudstone (first evidence of the Isfjorden Member). From 19 to 22 meters: red nodules displaying strong reaction with hydrochloric acid.	Not defined	Nod-ules. Calci-fied roots?	Calcareous nodules from 19 to 21 meters might be calcrete. Otherwise noncalcareous red and green mudstone. (FA 3.2)
25	Teist 15.2.19. C	30.5	40 cm	None	**A-horizon:** 15 cm. Black, organic rich mudstone. **Bw horizon:** 15 cm Yellow silty mudstone with yellow and grey mottles. **Bt-horizon:** 10 cm. Grey with yellow mottles.	Very fine granular	Mottles	Argillisol on Flood plain (FA 3.1)
26	Teist 15.2.16B	62	200 cm	None	Alternating red and green mudstone	Not defined	None	Noncalcareous red and green mudstone in interdistributary area (FA 3.2)
	Klement \`evfjellet							
27	Klem 15.1.24. C **XRD**	195	12 cm	None	**Top 2 cm:** coal shale **Lower 10 cm:** brown silty shale.	Medium angular blocky	None	Protosol on flood plain on the lower delta plain (FA 3.1).
28	No sample	197	10 cm	None	Top 2 cm: coal shale Lower 8 cm: brown silty shale.	Medium angular blocky	None	Protosol on flood plain on the lower delta plain (FA 3.1).
29	Klem 15.1.31. C	240	20 cm	None	Red layer.	Coarse sub-angular blocky	None	Noncalcareous red and green Mudstone in interdistributary area (FA 3.2)

No	Log/ sample	Log (m)	Thick-ness	Reaction HCl	Soil horizon	Soil structure	Roots	Classification
	Friedrichfjellet							
30	Fred 15.2.19.C	174	30 cm	None	**Top 2 cm:** red mudstone **Middle 3 cm:** green mudstone **Lower 25 cm:** yellow silt	Mudstone: medium subangular. Yellow: Very fine granular.	None	Protosol in interdistributary area (FA 3.2)
31	Fred 15.2.20	191	20 cm	Strong	**K-horizon:** green layer.	Coarse ang-ular blocky	Mottles nodules	Calcrete on floodplain (FA 3.1).
32	Fred 15.2.22.C	195	35 cm	Strong	**K-horizon:** greenish layer.	Coarse ang-ular blocky	Mottles	Calcrete on Floodplain (FA 3.1)
	Krefft-berget							
	No palaeosol							
	Blank-nuten15-1							
33	No sample	7-8	100 cm	None	**A-horizon:** 80 cm. Grey silt clay with two yellow layers approx-imately 5 cm. **B-Horizon:** 20 cm dark silty clay. The soil appears relatively homogenous with unclear layering	A-horizon: Very fine granular	Mottles in A-horizon	Vertisol on Flood plain on the lower delta plain (3.1)
34	Blank PS.2.B	20	10 cm	None	**C-horizon:** 10 cm of yellow silty clay sandwiched between distributary channel deposits.	Very fine granular	None	Protosol on top of distributary channel (FA 2.3)
35	Blank 4.B	30	65 cm	None	**A-horizon:** 5 cm coal shale and 10 cm grey, organic rich, silty clay with mottles. **Bw-horizon:** 50 cm weathered brown sand	A-Horizon: vf granular Bw-horizon: medium platy	Mottles in A-horizon.	Protosol on floodplain (FA 3.1).
	Blank-nuten15-2							
36	Blank 7b.1 and 7.C	5	110 cm	None	Dark, organic rich and homogenous. Thin coal seams.	Very fine granular	None	Vertisol on top of Barrier bar (FA 2.1)
37	Blank 12.B (**XRD**) 12.C (coal).	28	100 cm	None	**A-Horizon:** 25 cm organic rich homogeneous mudstone. **Bw-Horizon:** 75 cm over light yellow consolidated silty sand (f-m). Roots in upper 20 cm.	A-horizon: Very fine granular Bw-horizon: Thin platy.	A: Mottles Bw: coalified roots. 20 cm	Vertisol on top of distributary channel (FA 2.3).

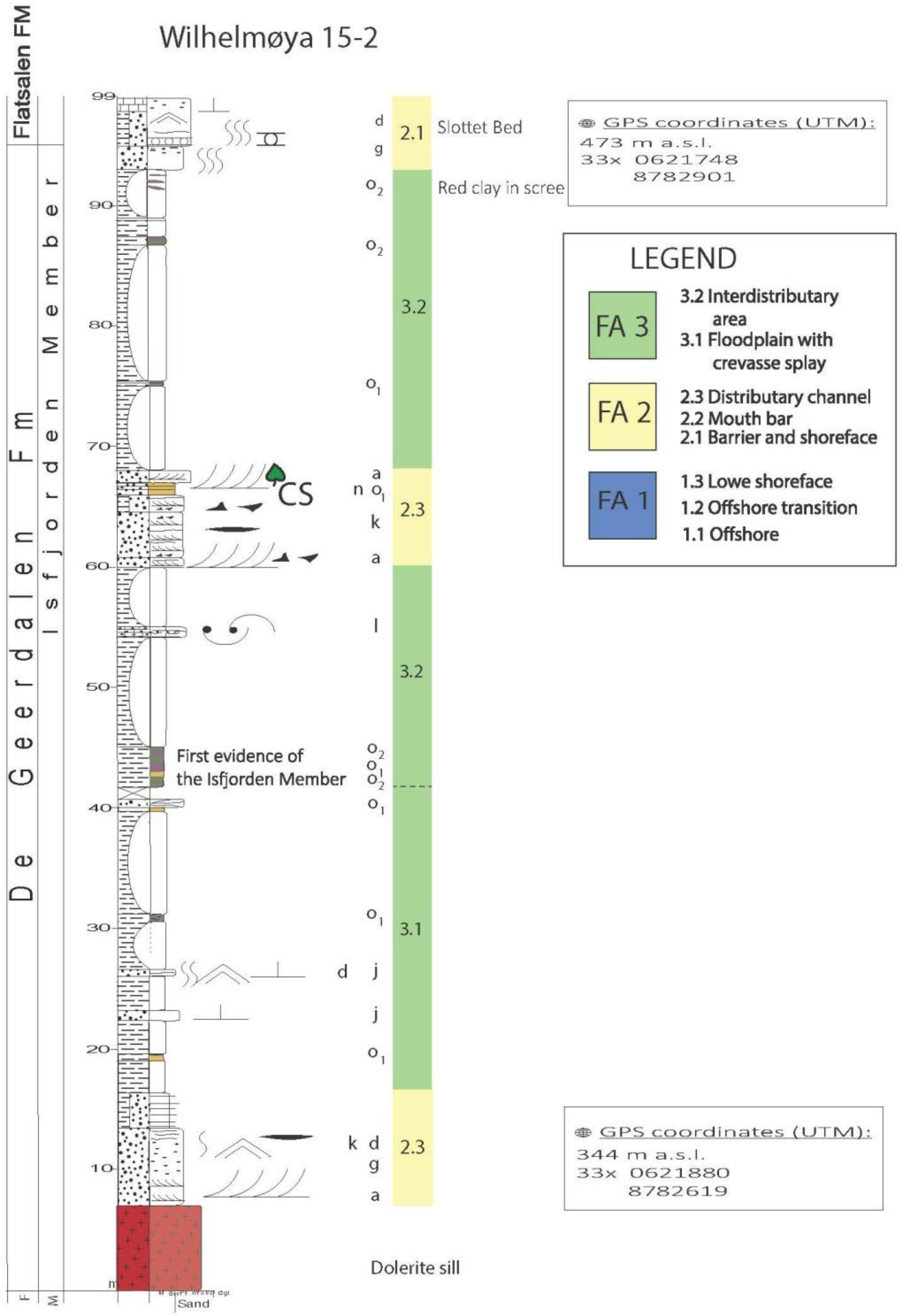
No	Log/ sample	Log (m)	Thick-ness	Reaction HCl	Soil horizon	Soil structure	Roots	Classification
	Šmidt-berget							
38	Schmidt 3C	56.5	50 cm	None	**A-horizon:** 40 cm of Organic rich dark and homogenous shale. **B-horizon:** 10 cm yellow clay	Medium angular blocky	None	Vertisol/Protosol on barrier bar (FA 2.1)
40	Schmidt $7C_1$ and $7C_2$	83.6	35 cm	None	Green shale overlying coquina bed.	Medium angular blocky	None	Noncalcareous green mudstone in Interdistributary area (FA 3.2)
41	Schmidt 8C	86.5	50 cm	None	**A-horizon:** 5 cm coal shale overlying. **E-horizon:** 35 cm bleached layer, with roots. **Bw-horizon:** coarse reddish layer. Red/orange base.	Fine granular	Coalified roots, mottles	Argillisol on floodplain (FA 3.2).
42	No sample	91.5	40 cm	None	**A-horizon:** 10 cm organic rich layer. **E-horizon:** 20 cm bleached grey and pale yellow layer. Lower boundary not exposed.	Fine granular	Mottles	Argillisol on floodplain (FA 3.2).
43	Schmidt 9C	99.7	50 cm	None	Undefined weakly developed horizons with grey to yellow colours. Relict bedding.	Fine granular	Mottles	Protosol on floodplain (FA 3.2)
44	Schmidt $10C_1$ and $10C_2$	112	30 cm	None	**Upper 10 cm:** red mudstone **Middle 10 cm:** green mudstone **Lower 5 cm:** yellow layer	Medium subangular blocky	None	Noncalcareous red and green mudstone in interdistributary area (FA 3.2)
45	No sample	115	40 cm	Not tested	**Upper 20 cm:** red mudstone **Lower 20 cm:** green mudstone Appears cemented	Medium prismatic	Nod-ules	Calcrete in interdistributary area (FA 3.2)
46	Schmidt 12C	116	60 cm	None	Green mudstone	Subangular blocky	None	Noncalcareous green mudstone in interdistributary area (FA 3.2)
47	Schmidt 14C	122.5	30 cm	None	Red mudstone	Very fine subangular blocky	None	Noncalcareous red mudstone on top of distributary channel (FA 2.3)
48	No sample	148.6	15 cm	None	**Upper 5 cm:** grey /rusty mudstone. **Lower 10 cm:** of grey/brown mudstone	Not defined	None	Protosol in interdistributary area (FA 3.2)

No	Log/ sample	Log (m)	Thick-ness	Reaction HCl	Soil horizon	Soil structure	Roots	Classification
	Delta-neset							
49	Sample 1 **Thin section**	6.5	100 cm	Strong	**K-horizon:** Green. No mottles, no organic content. Peds.	Not defined	None	Calcrete in interdistributary area (FA 3.2)
50	Sample 2	6.7	30 cm	Strong	**K-horizon:** Red with red streak. Soft. Nodules approximately 6 cm.	Not defined	Nodule s	Calcrete in interdistributary area (FA 3.2)
51	Sample 3	7.5	50 cm	Not tested	Green. Dark purple mottles. Nodules and peds	Not defined	Nodule s Mottles	Noncalcareous green mudstone in Inter-distributary area (FA 3.2)
52	Sample 4	10	25 cm	Strong	**K-horizon:** Green with mottles. Some few nodules. Peds	Not defined	Few nodules Mottles	Calcrete in interdistributary area (FA 3.2)
53	Sample 5	10.3	30 cm	Strong	**K-horizon:** Purple with light and dark mottles. Nodules. Organic matter around 2x2 mm.	Not defined	Nod-ules Mottles	Calcrete in interdistributary area (FA 3.2)
54	Sample 6	14.5	25 cm	None	Red with red streak.	Not defined	Nod-ules	Noncalcareous red mudstone in interdistributary area (FA 3.2)
55	Sample 7A	15.5	90 cm	None	Green silt/clay with black mottles and light red nodules. Peds.	Not defined	Nodule s Mottles	Noncalcareous red and green mudstone in interdistributary area (FA 3.2)
56	Sample 7B	16.5	70 cm	None	Green Mudstone	Not defined	Nodule s Mottles	Noncalcareous green mudstone in Inter-distributary area (FA 3.2)
57	Sample 8 **Thin section**	17	20 cm	Strong	**K-horizon:** Purple with green mottles and green nodules.	Not defined	Nod-ules Mottles	Calcrete in interdistributary area (FA 3.2)
58	Sample 9	17.2	20 cm	Strong	**K-horizon:** Green with black nodules. Well cemented. White streak.	Not defined	Nod-ules Mottles	Calcrete in interdistributary area (FA 3.2)
59	Sample 10	18	35 cm	Not tested	Appears well cemented. Orange. Black mottles. Chaotic inner structure. Organic fragments up to 3 cm.	Not defined	Mottles	Likely calcrete in interdistributary area (FA 3.2)
60	Sample 11**Thin section**	26	10 cm	Strong	**K-horizon**: green bed with a lot of nodules.	Not defined	Nod-ules	Calcrete, possibly in lagoon close to FA 2.1
61	Sample 12	31	40 cm	Strong	**K-horizon:** Organic fragments around 3 mm.	Not defined	Mottles	Calcrete in interdistributary area (FA 3.2)

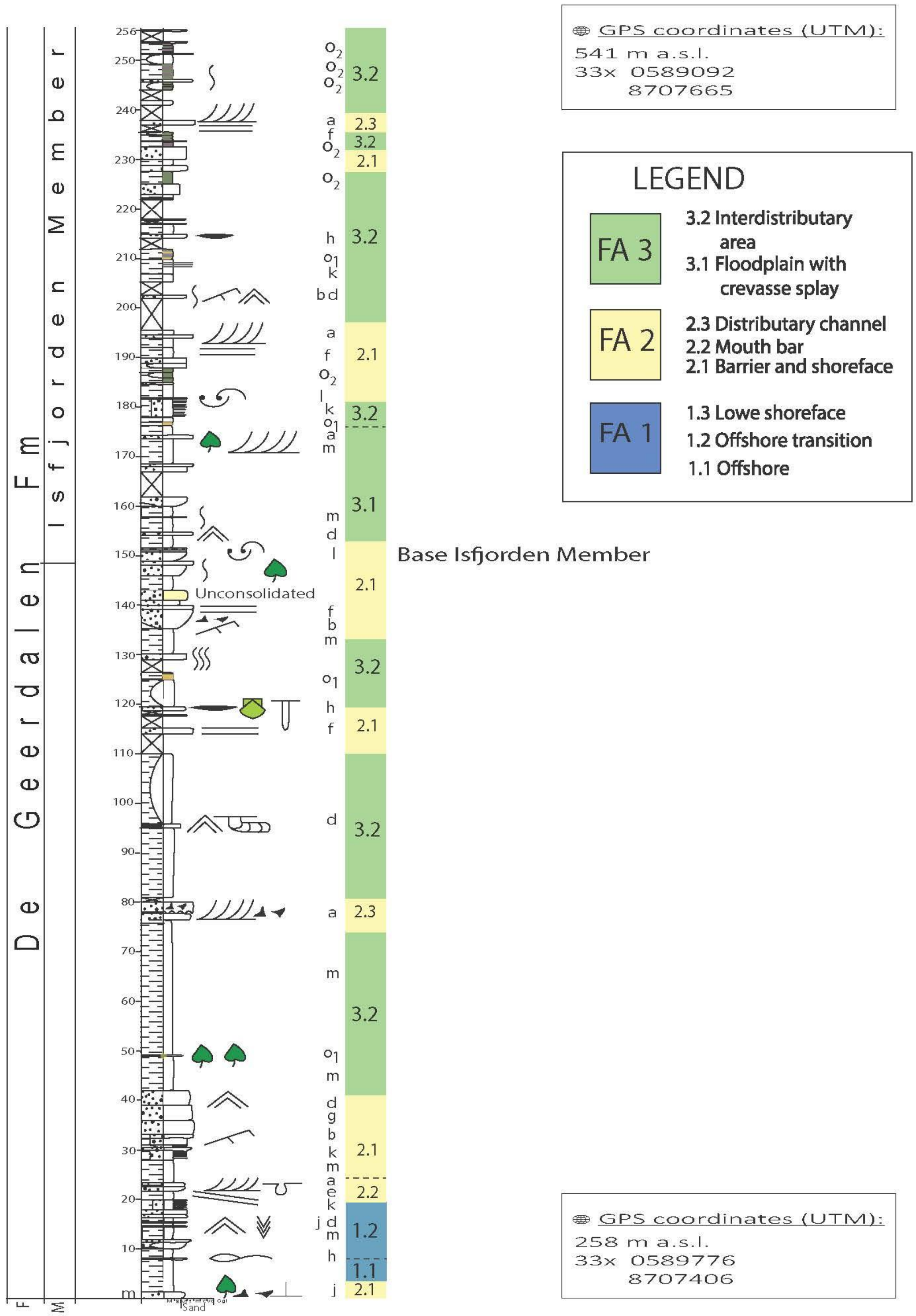
Appendix B – Legend for Measured Sections



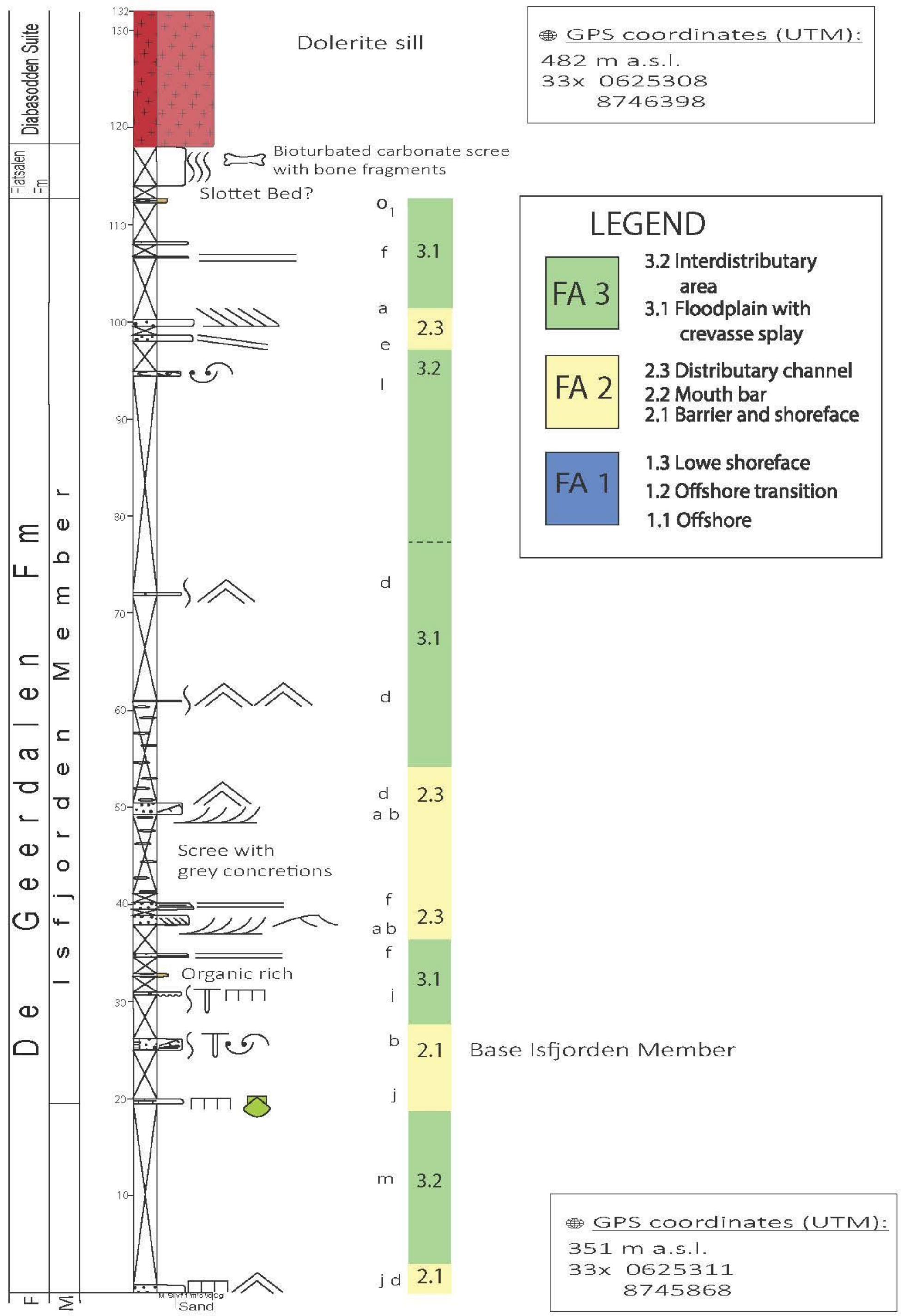
Appendix C – Measured Sections



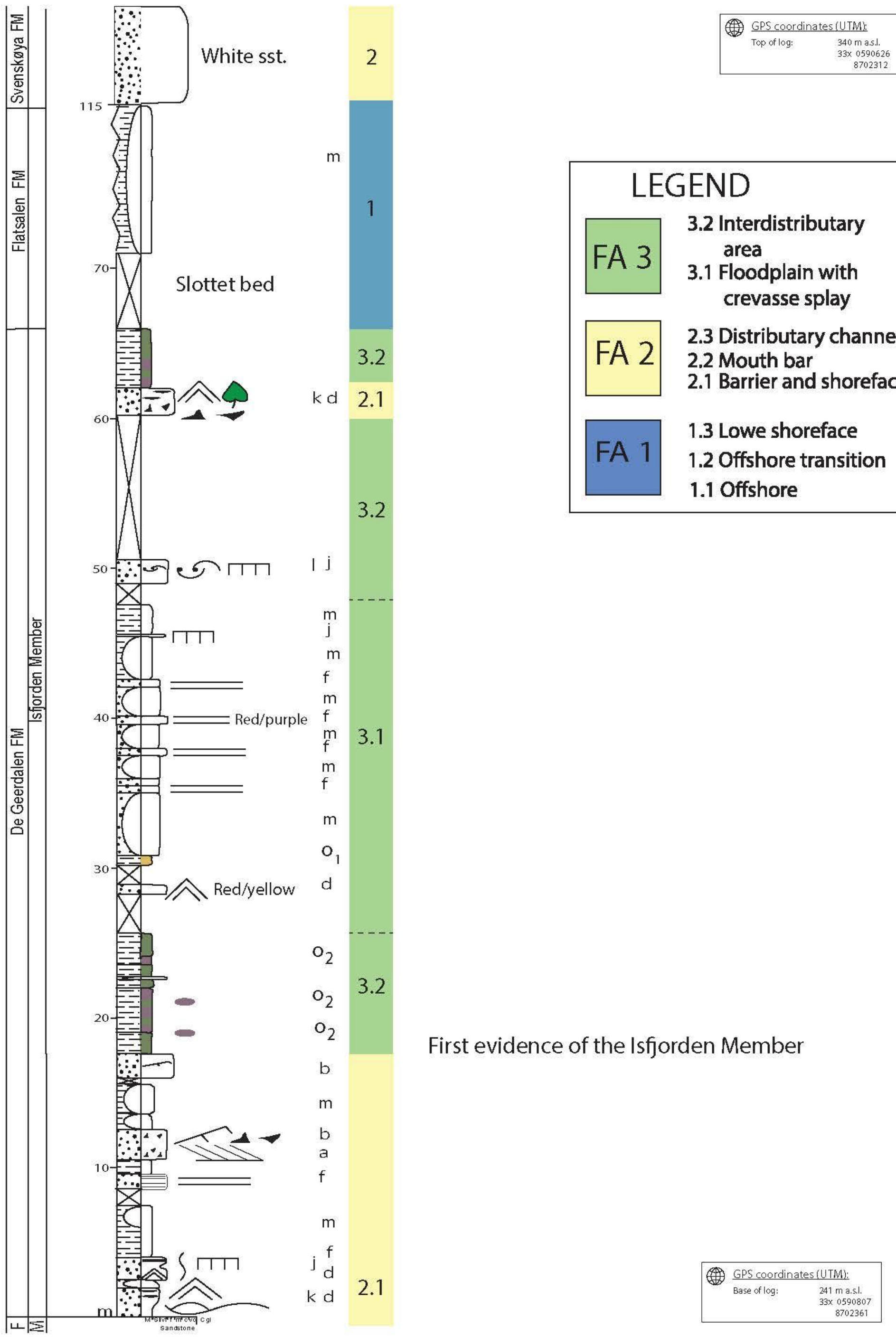
Hahnfjella 15-2



Hellwaldfjellet 15-2

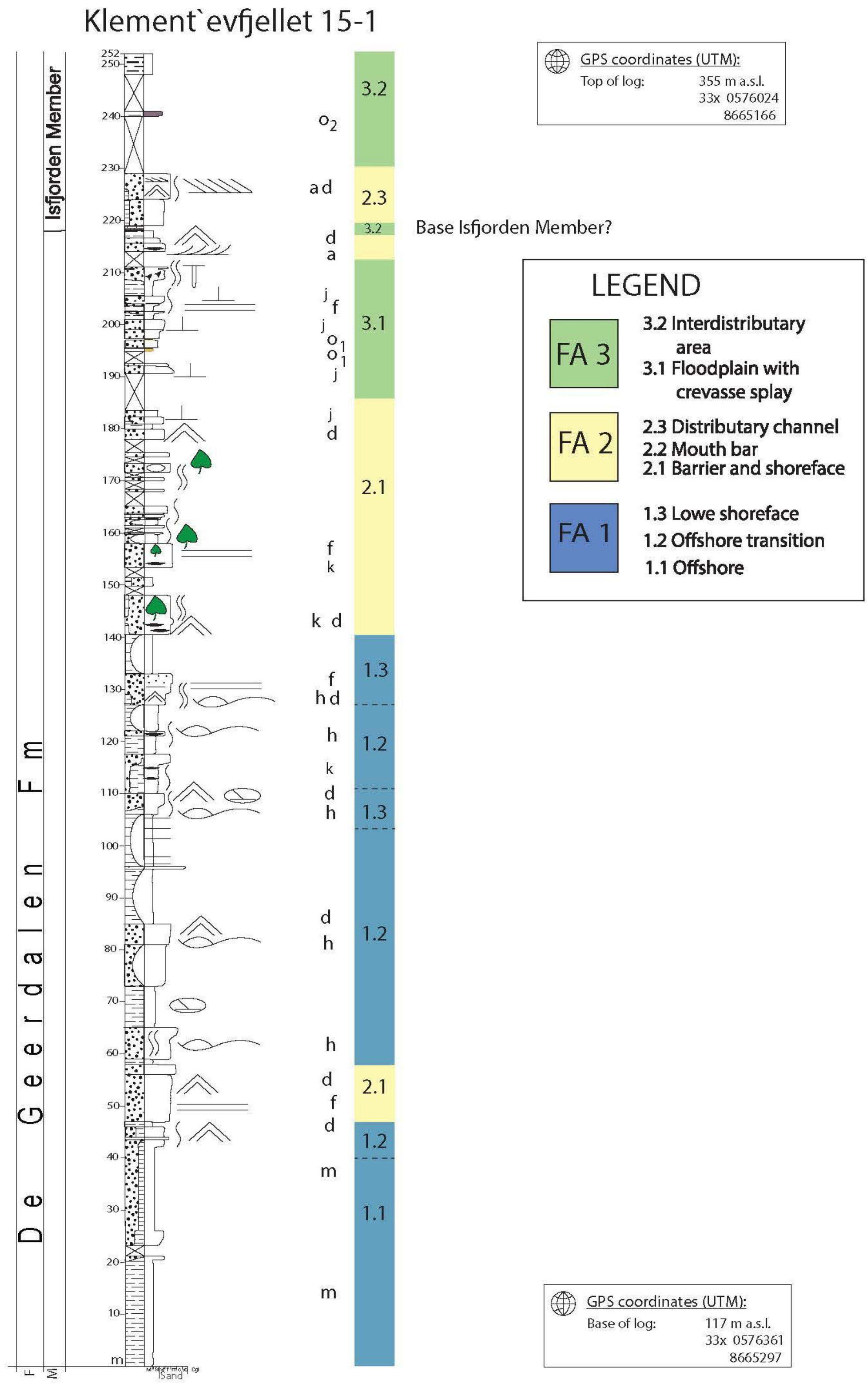


Teistberget 15-2

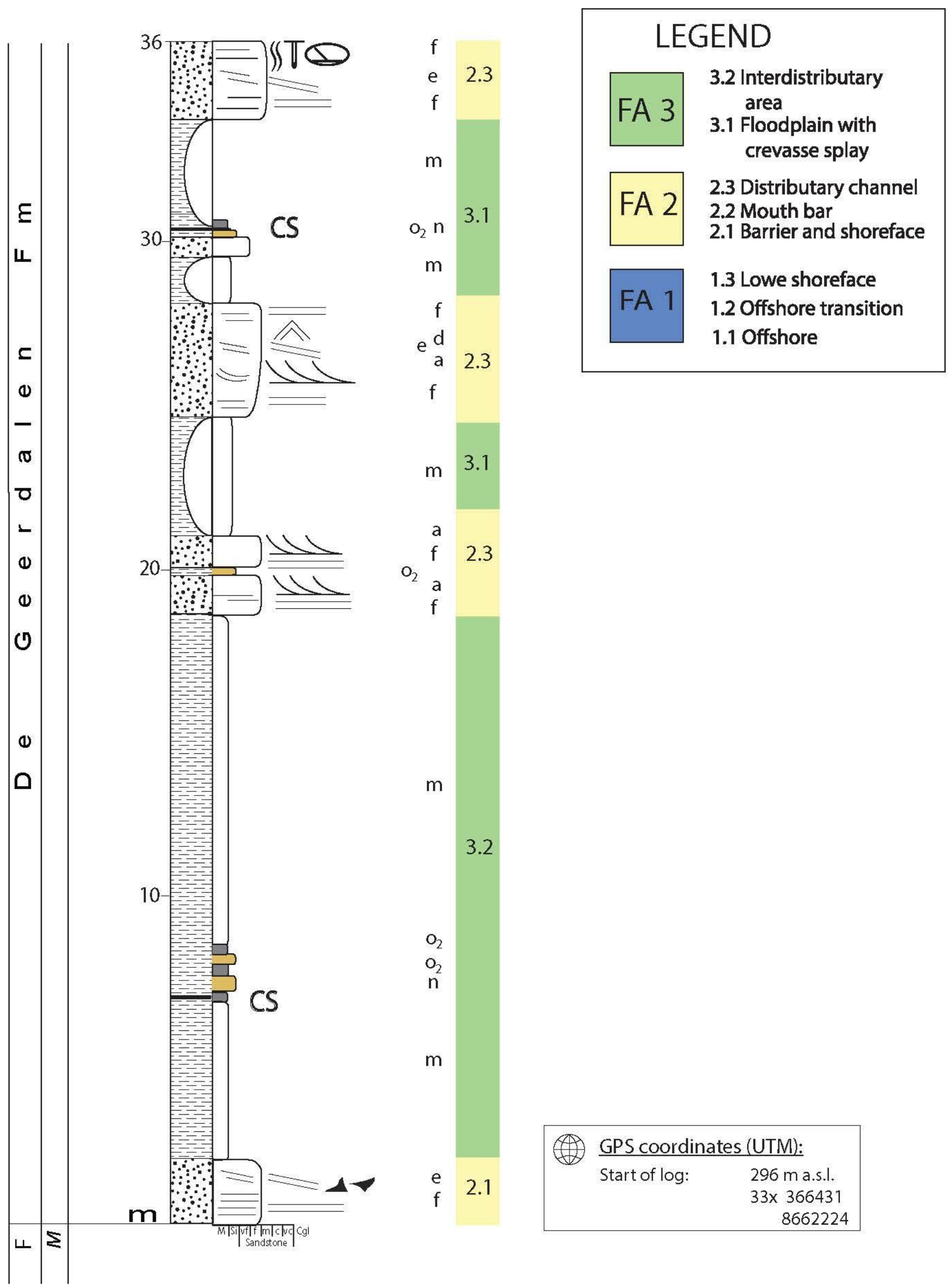


LEGEND

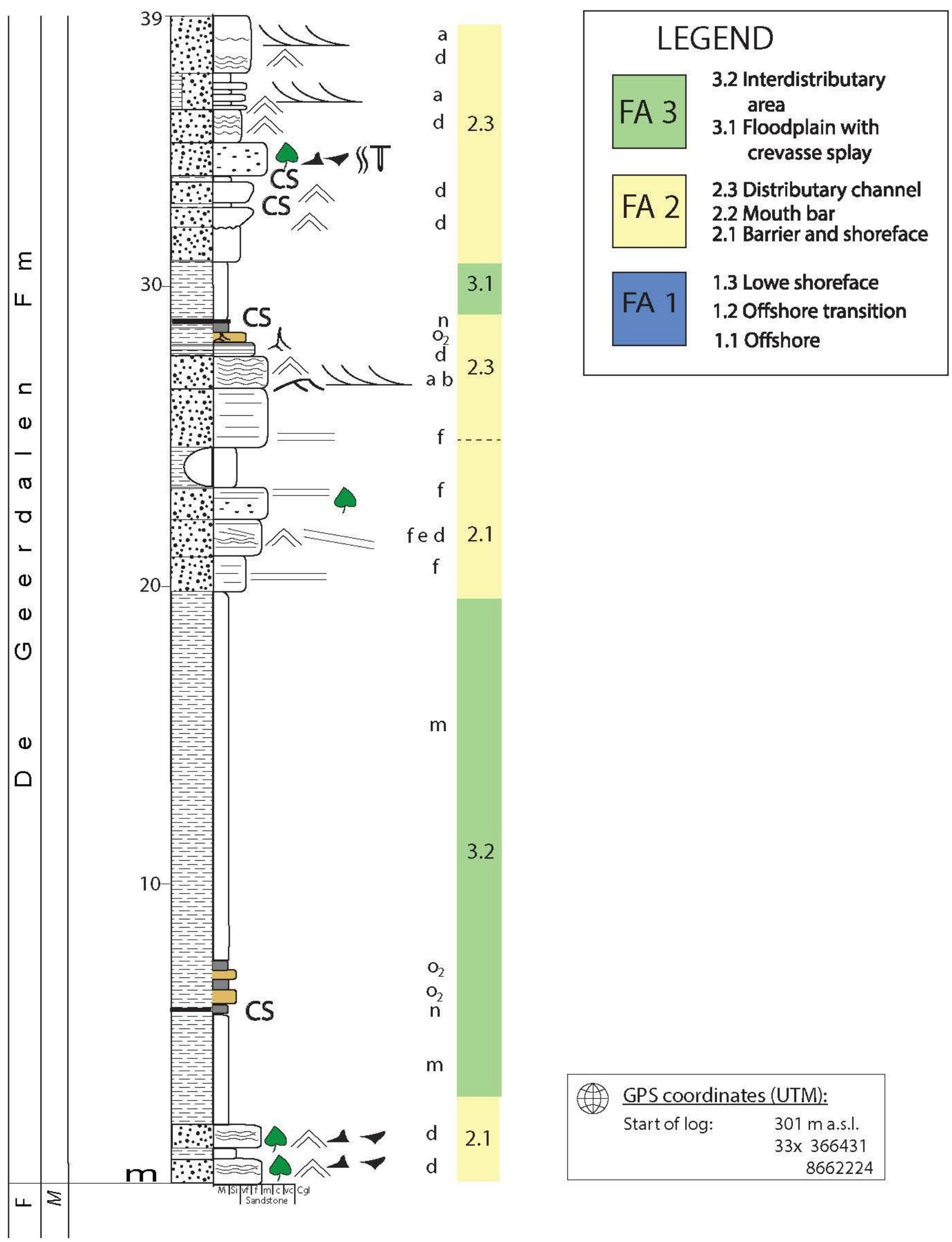
FA 3	3.2 Interdistributary area 3.1 Floodplain with crevasse splay
FA 2	2.3 Distributary channel 2.2 Mouth bar 2.1 Barrier and shoreface
FA 1	1.3 Lowe shoreface 1.2 Offshore transition 1.1 Offshore


Klement`evfjellet 15-1
GPS coordinates (UTM):
Top of log: 355 m a.s.l.
33x 0576024
8665166
Base Isfjorden Member?
LEGEND
FA 3
3.2 Interdistributary area
3.1 Floodplain with crevasse splay
FA 2
2.3 Distributary channel
2.2 Mouth bar
2.1 Barrier and shoreface
FA 1
1.3 Lowe shoreface
1.2 Offshore transition
1.1 Offshore
GPS coordinates (UTM):
Base of log: 117 m a.s.l.
33x 0576361
8665297
De Geerdalen Fm
Isfjorden Member
Sand

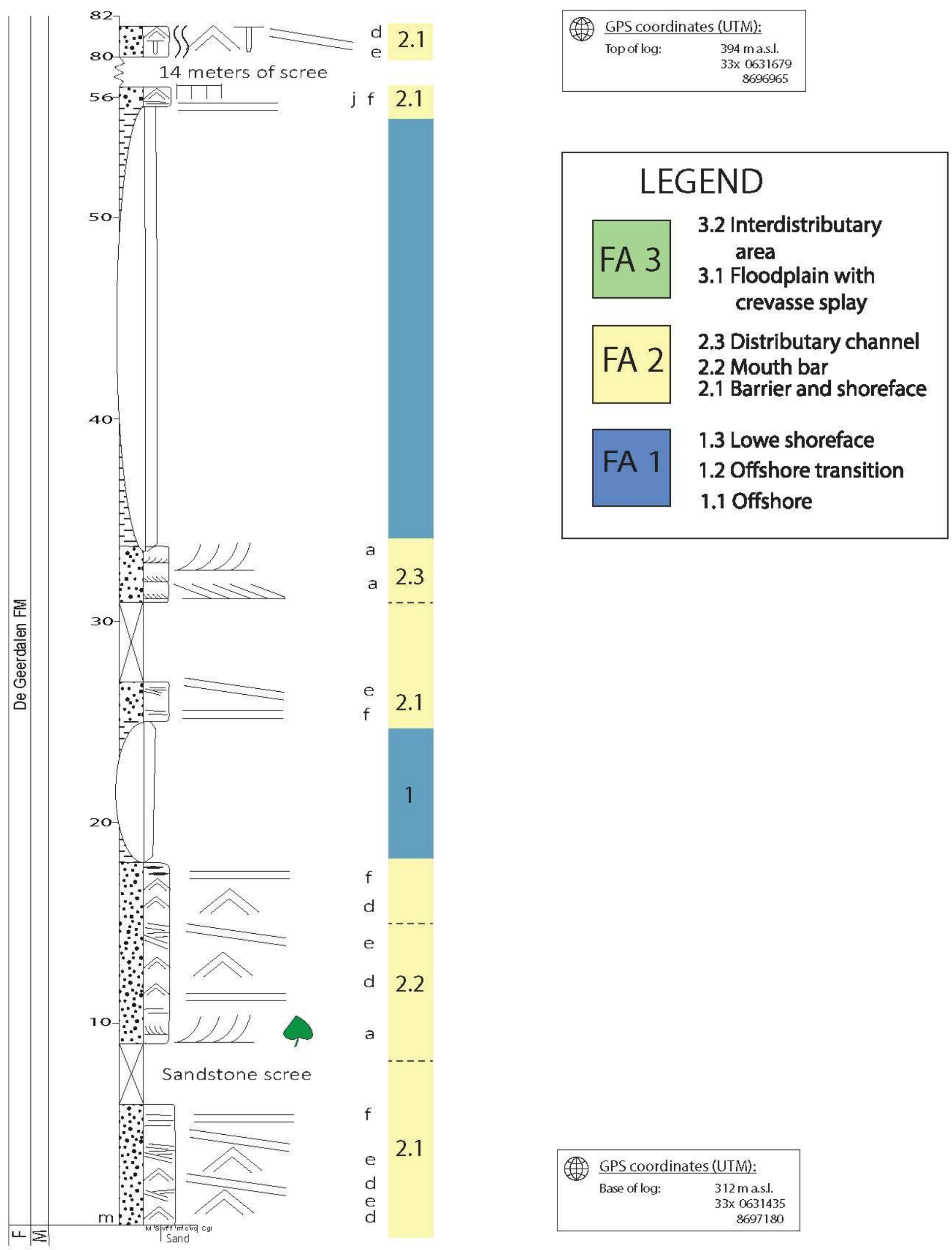
Blanknuten-15-1



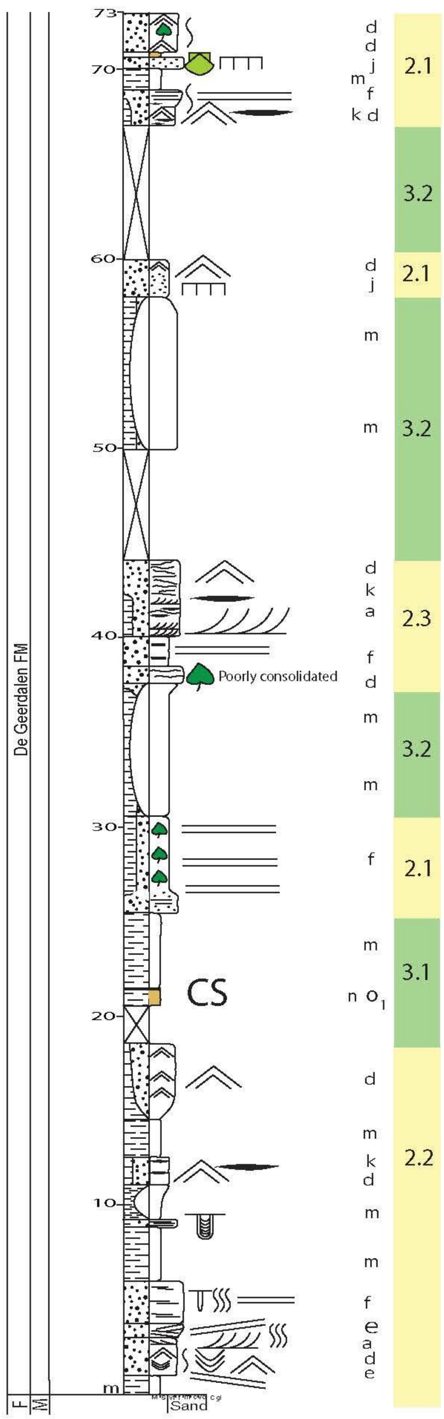
Blanknuten-15-2

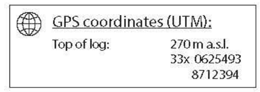


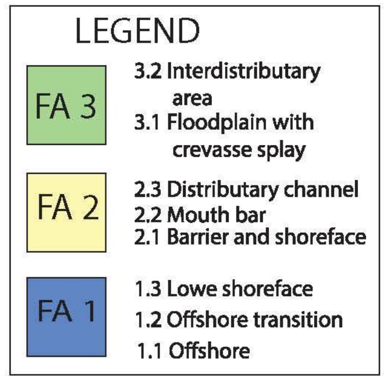
Krefftberget 15-2

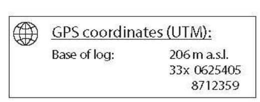


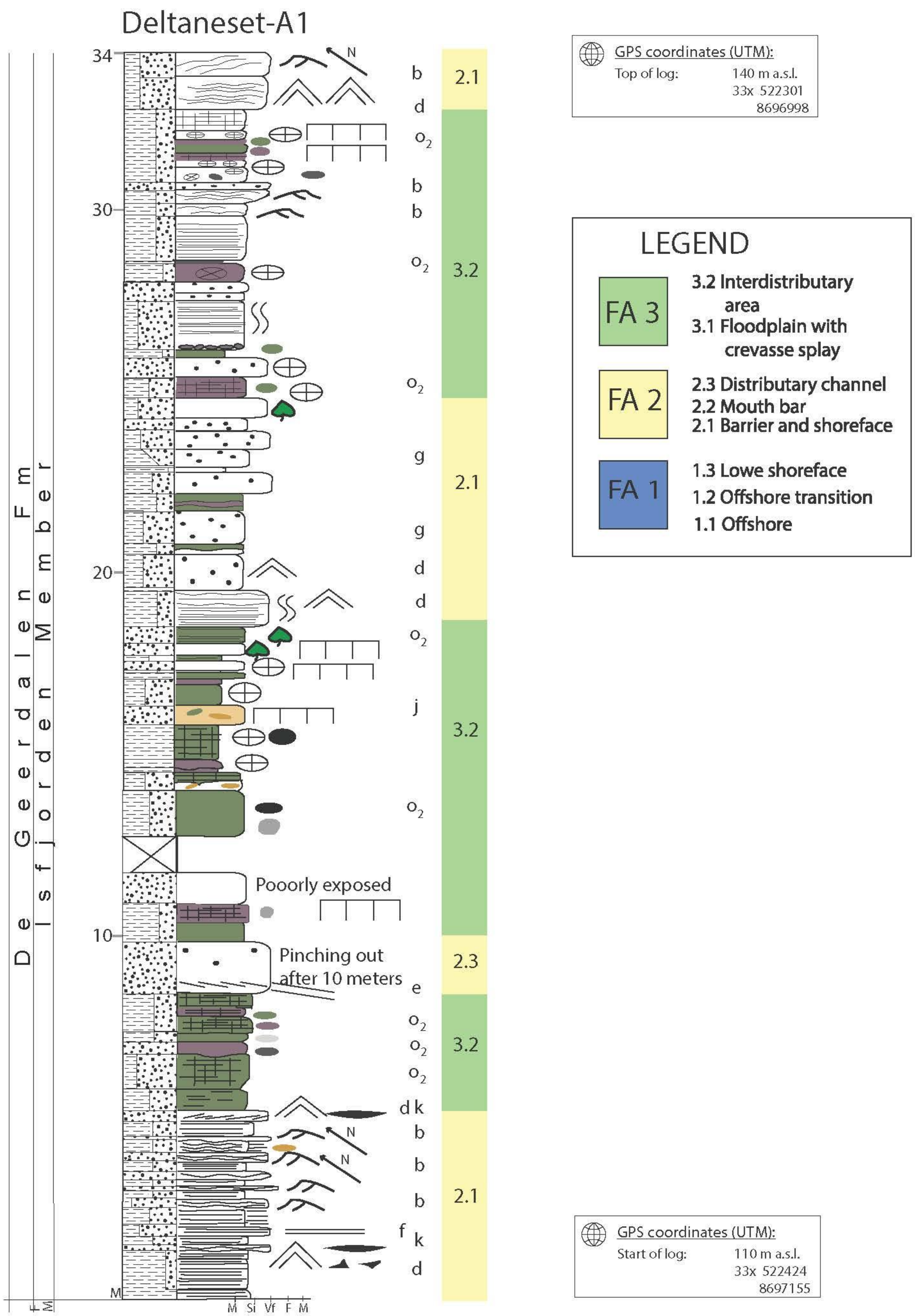
Svartnosa 15-2











Deltaneset-A1
GPS coordinates (UTM):
Top of log: 140 m a.s.l.
33x 522301
8696998
LEGEND
FA 3
3.2 Interdistributary area
3.1 Floodplain with crevasse splay
FA 2
2.3 Distributary channel
2.2 Mouth bar
2.1 Barrier and shoreface
FA 1
1.3 Lowe shoreface
1.2 Offshore transition
1.1 Offshore
De Geerdalen Fm
Isfjorden Member
34
30
20
10
M
F M
M Si Vf F M
Poorly exposed
Pinching out after 10 meters
2.1
3.2
2.1
3.2
2.3
3.2
2.1
GPS coordinates (UTM):
Start of log: 110 m a.s.l.
33x 522424
8697155

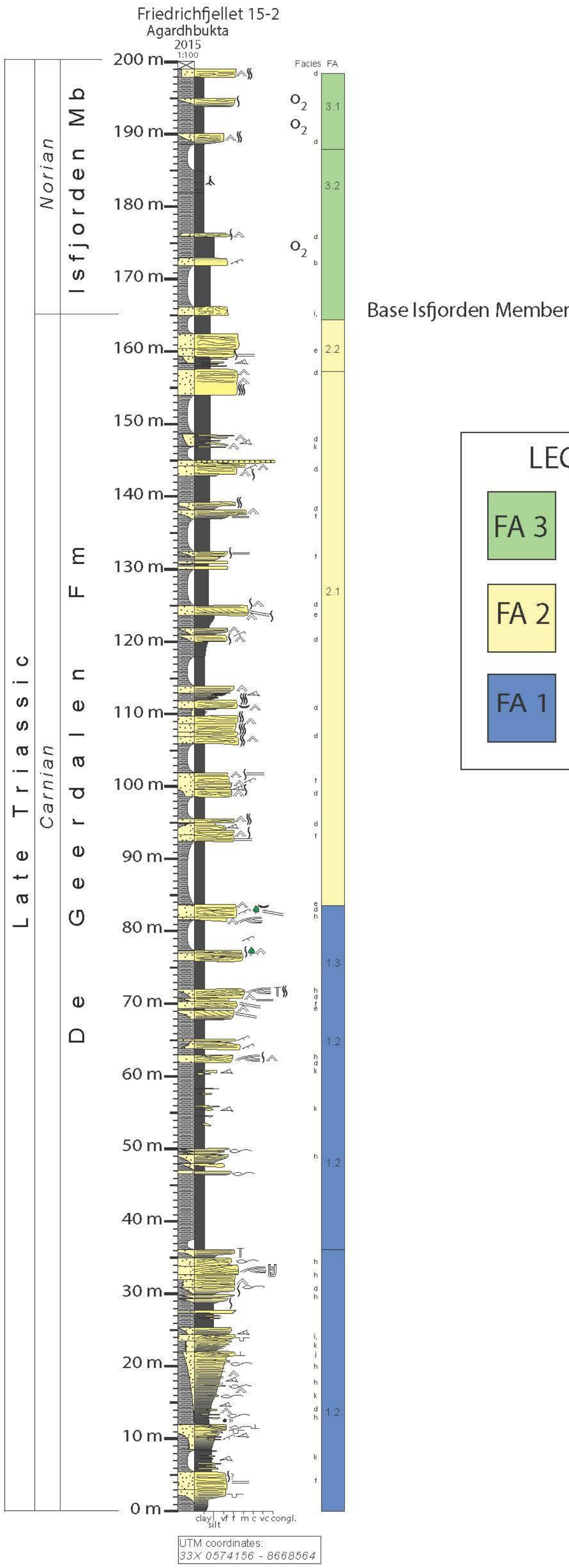
Appendix D - Sections drawn by Johansen (2016) and Støen (2016)

This appendix contains logs from thee measured sections.

Log 1: Friedrichfjellet 15-2: The section was measured by Sondre Krogh Johansen and Turid Haugen. The log is digitalized by Johansen (2016). Facies association next to the log is done in the study presented herein.

Log 2: Šmidtberget 15-2: The section was measured by Gareth Lord, Simen Jenvin Støen and Turid Haugen. The log is digitalized by Støen (2016). Facies association next to the log is done in the study presented herein.

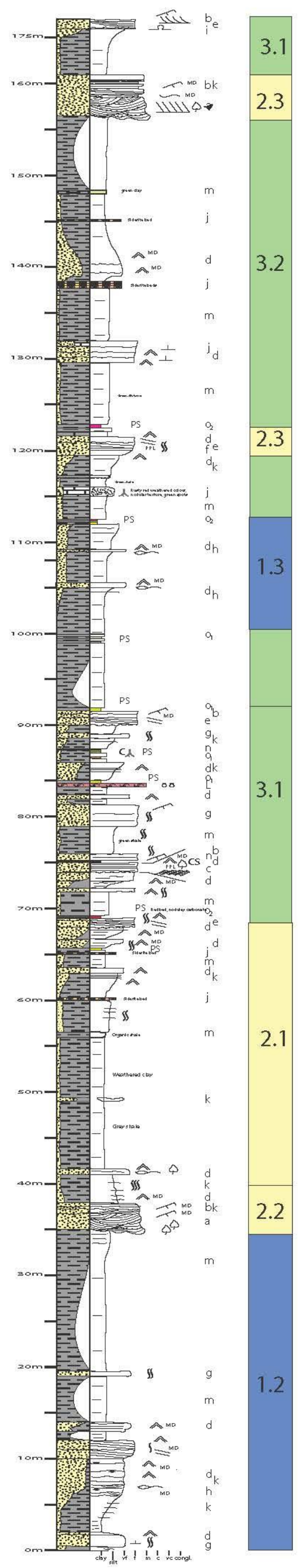
Log 3: Wilhelmøya 15-1: The section was measured by Sondre Krogh Johansen, Simen Jenvin Støen and Bård Heggem. Digitalizing and interpretations are done by Johansen (2016). The log overlaps the Wilhelmøya 15-2 log (Appendix C). The overlap starts from the dolerite sill intrusion. An excursion with all participants in the field work gave the author of this thesis the opportunity to observe the whole section from base of the De Geerdalen Formation to the Slottet Bed, including the whole measured section of the Wilhelmøya 15-1 log.



LEGEND

FA 3	3.2 Interdistributary area 3.1 Floodplain with crevasse splay
FA 2	2.3 Distributary channel 2.2 Mouth bar 2.1 Barrier and shoreface
FA 1	1.3 Lowe shoreface 1.2 Offshore transition 1.1 Offshore

Smidtberget 15-2



Base Isfjorden Member

GPS coordinates (UTM):

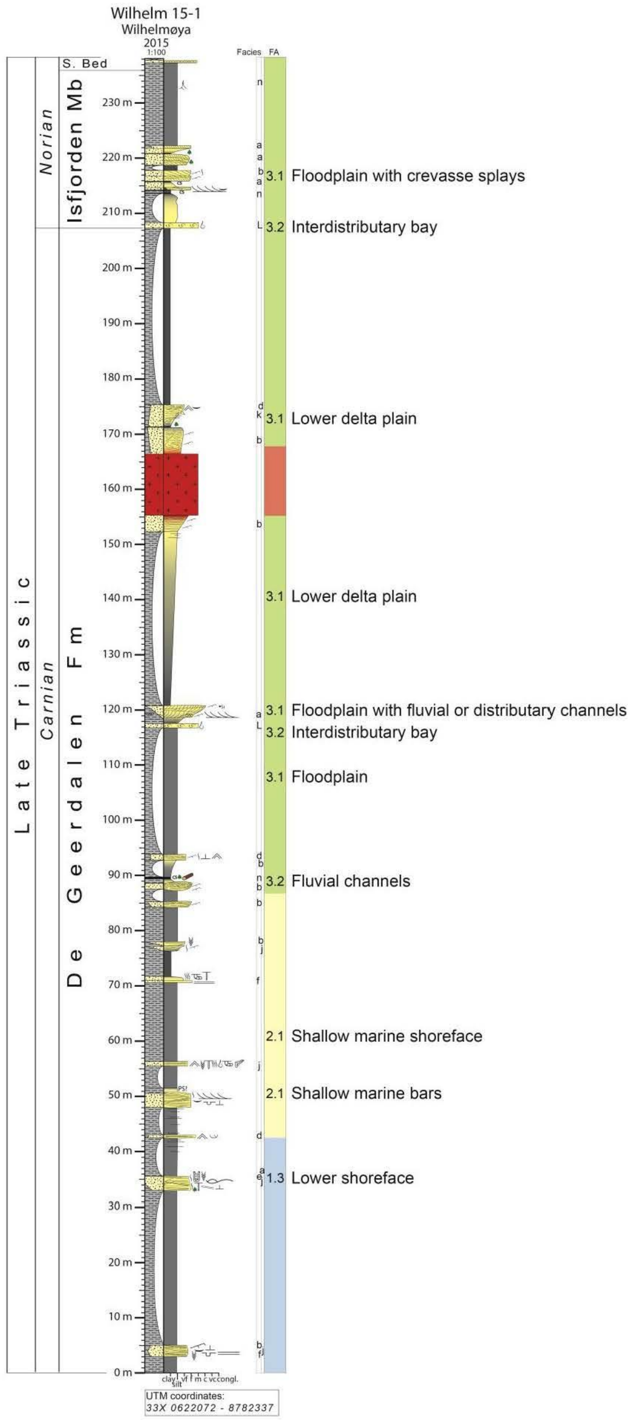
Top of log: 287 m a.s.l.
33x 0372797
8667030

LEGEND

FA 3	3.2 Interdistributary area 3.1 Floodplain with crevasse splay
FA 2	2.3 Distributary channel 2.2 Mouth bar 2.1 Barrier and shoreface
FA 1	1.3 Lowe shoreface 1.2 Offshore transition 1.1 Offshore

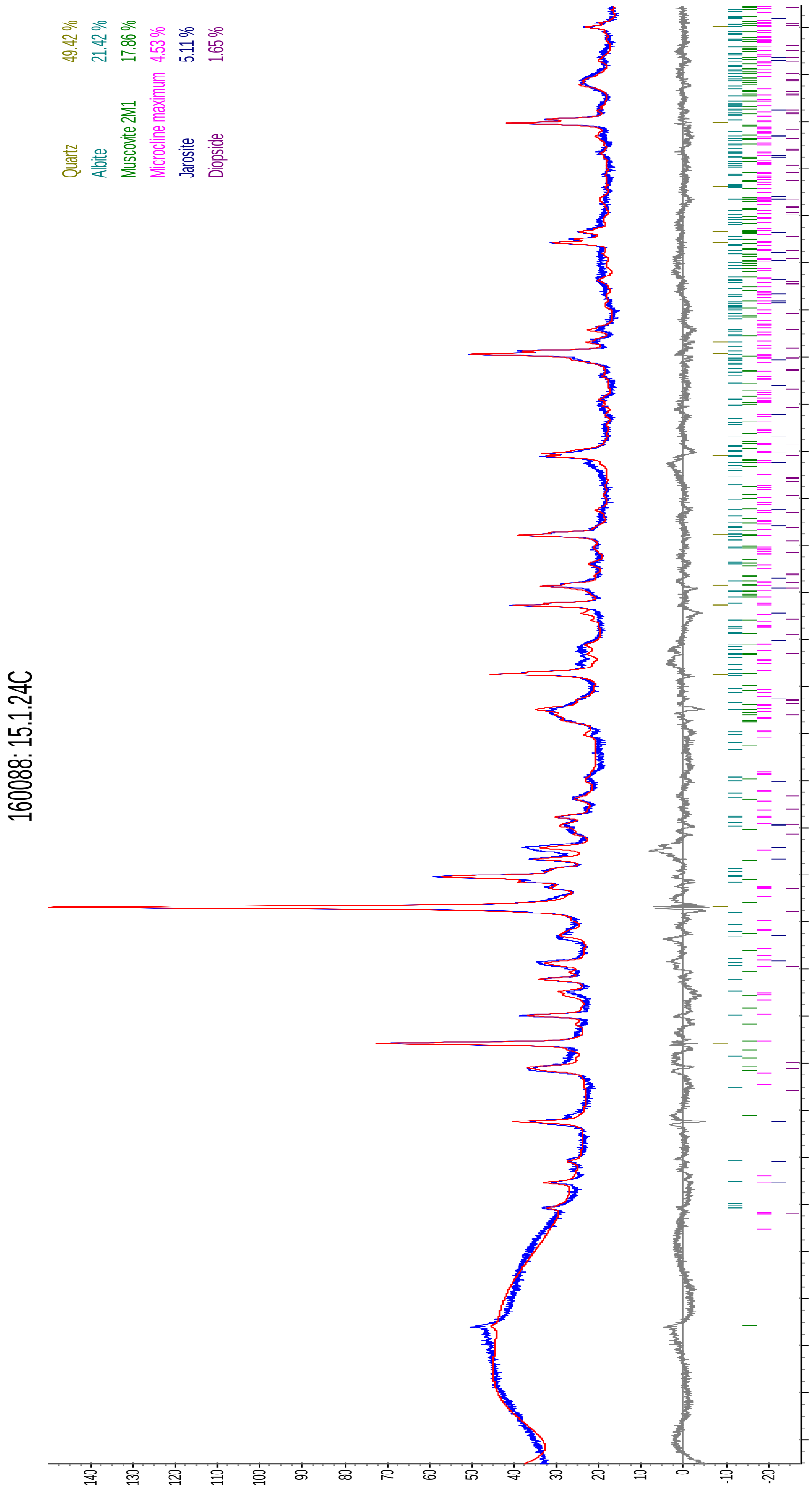
GPS coordinates (UTM):

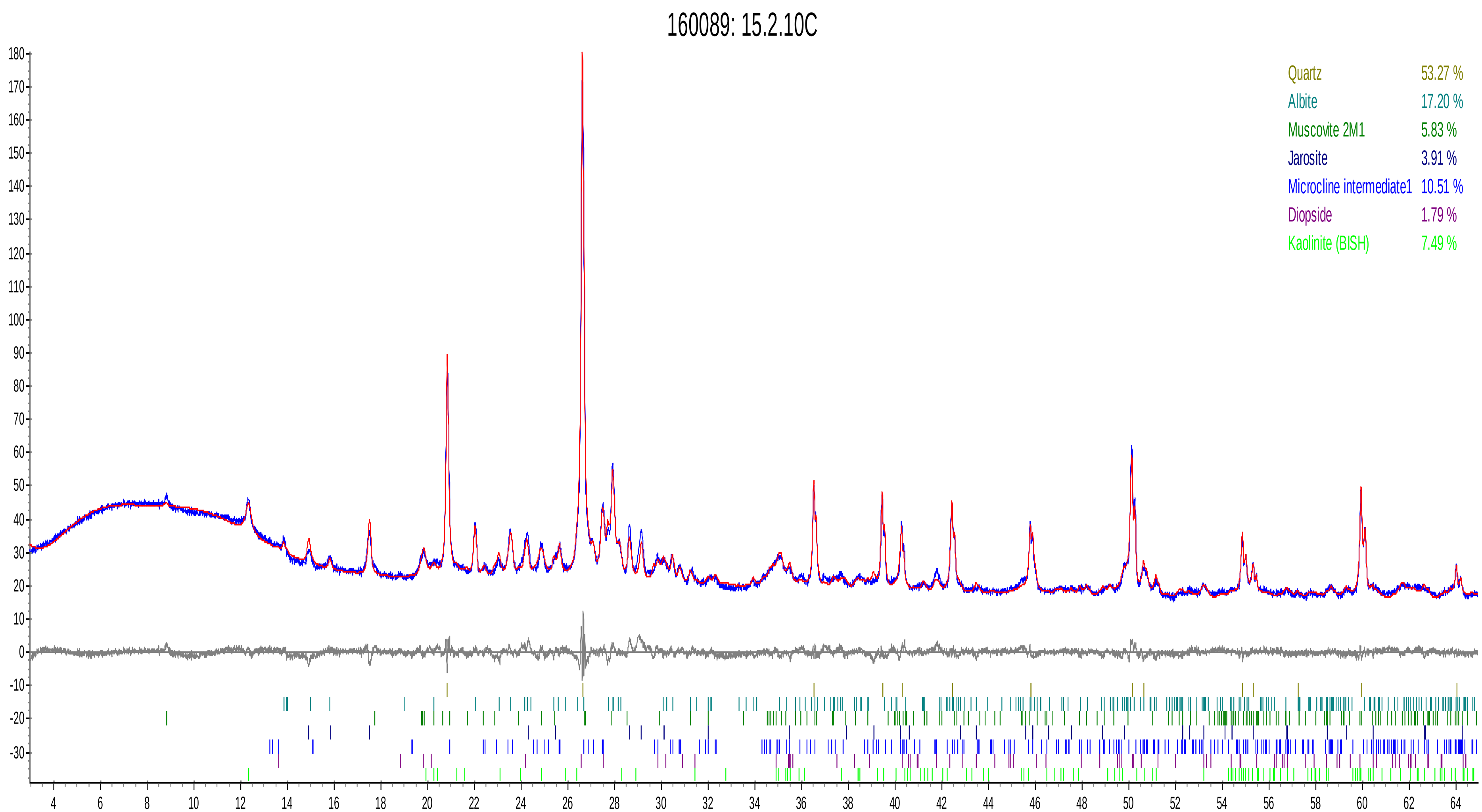
Base of log: 125 m a.s.l.
33x 0572994
8666819

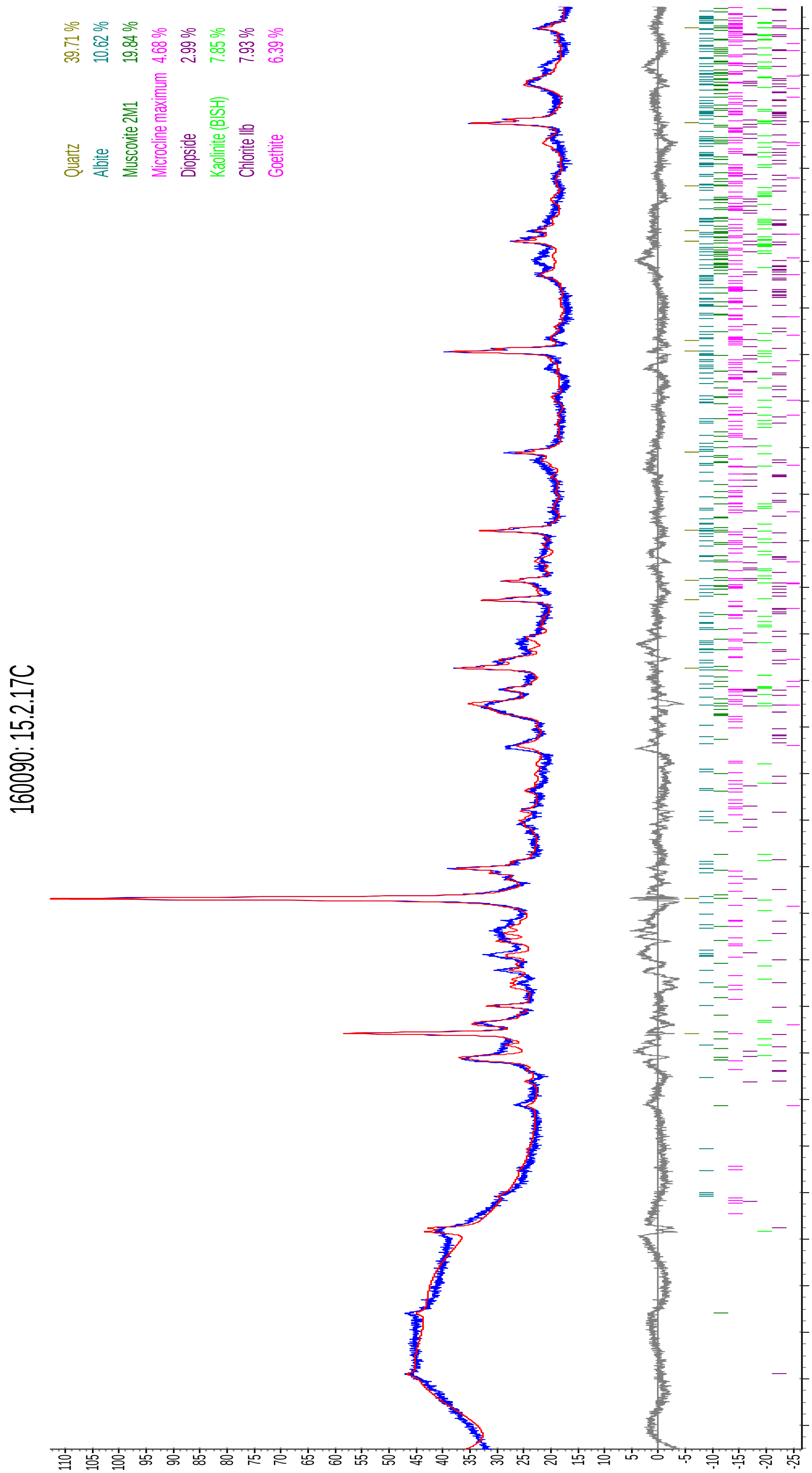


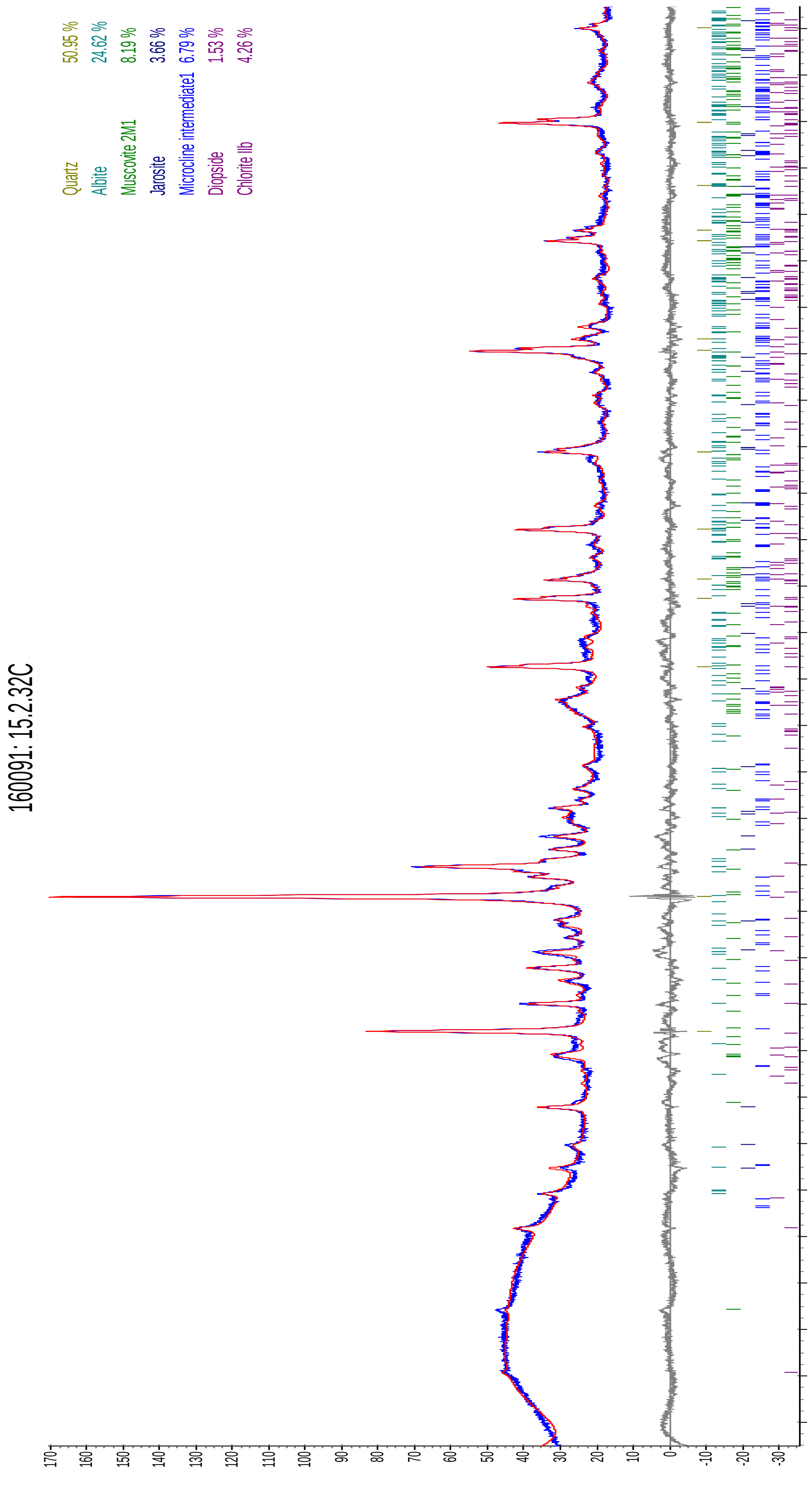
Wilhelm 15-1
Wilhelmøya
2015
1:100
Facies
FA
S. Bed
Norian
Isfjorden Mb
Late Triassic
Carnian
De Geerdalen Fm
3.1 Floodplain with crevasse splays
3.2 Interdistributary bay
3.1 Lower delta plain
3.1 Lower delta plain
3.1 Floodplain with fluvial or distributary channels
3.2 Interdistributary bay
3.1 Floodplain
3.2 Fluvial channels
2.1 Shallow marine shoreface
2.1 Shallow marine bars
1.3 Lower shoreface
UTM coordinates:
33X 0622072 - 8782337

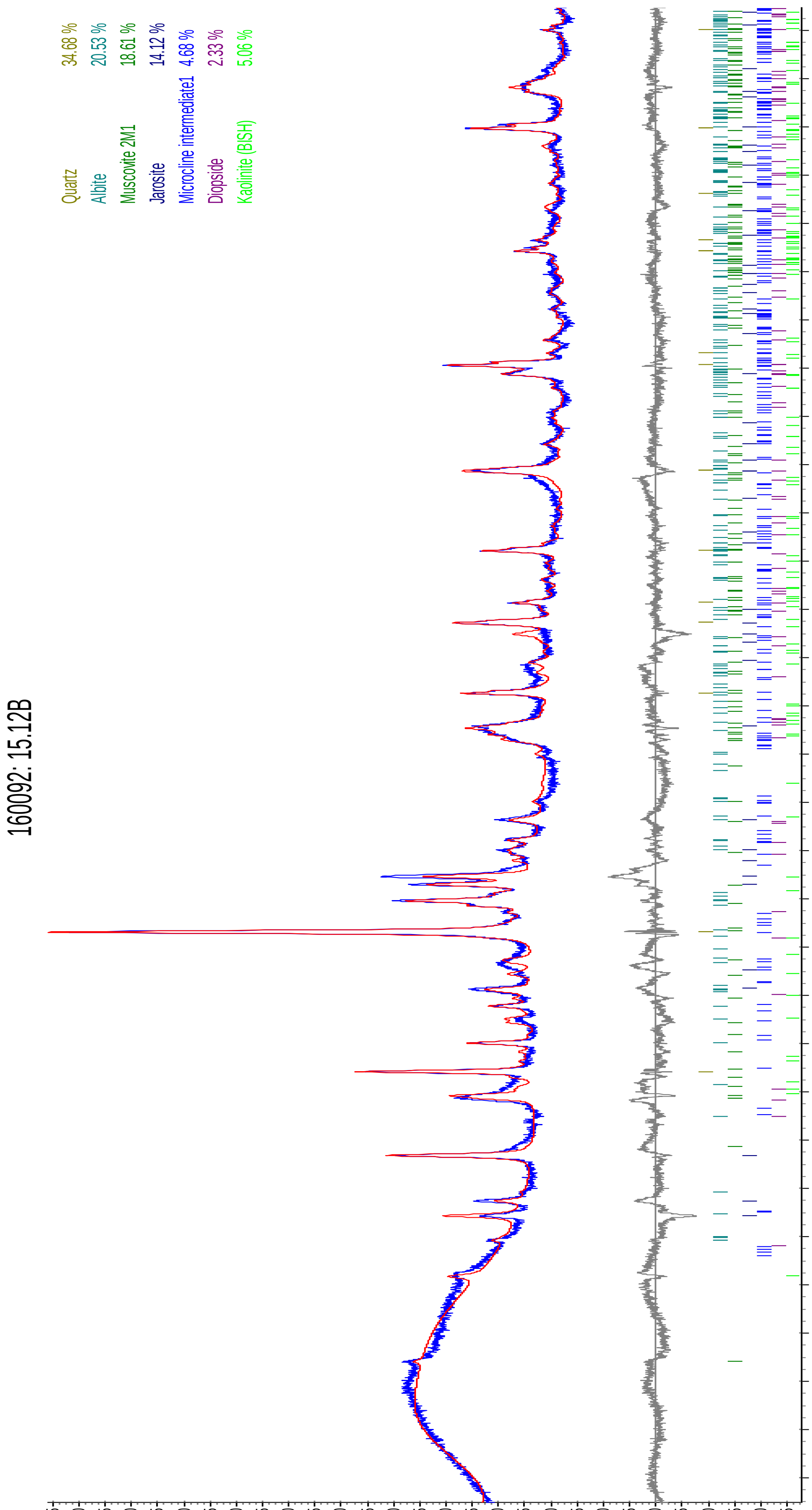
Appendix E: Results of XRD-analyses




160089: 15.2.10C
Quartz 53.27 %
Albite 17.20 %
Muscovite 2M1 5.83 %
Jarosite 3.91 %
Microcline intermediate1 10.51 %
Diopside 1.79 %
Kaolinite (BISH) 7.49 %


160090: 15.2.17C
Quartz 39.71 %
Albite 10.62 %
Muscovite 2M1 19.84 %
Microcline maximum 4.68 %
Diopside 2.99 %
Kaolinite (BISH) 7.85 %
Chlorite IIb 7.93 %
Goethite 6.39 %


160091: 15.2.32C
Quartz 50.95 %
Albite 24.62 %
Muscovite 2M1 8.19 %
Jarosite 3.66 %
Microcline intermediate1 6.79 %
Diopside 1.53 %
Chlorite IIb 4.26 %



160092: 15.12B
Quartz 34.68 %
Albite 20.53 %
Muscovite 2M1 18.61 %
Jarosite 14.12 %
Microcline intermediate1 4.68 %
Diopside 2.33 %
Kaolinite (BISH) 5.06 %