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Late Triassic sedimentology and diagenesis of Barentsøya, Wilhelmøya and eastern Spitsbergen

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Geology
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Abstract

This study documents and discusses the sedimentology and diagenesis of the Triassic De Geerdalen Formation on eastern Svalbard. In addition to sandstone sampling, nine sedimentary sections have been measured. These include the mountains of; Klement'evfjellet, Friedrichfjellet and Smidtberget in Agardhbukta, Teistberget and Hellwaldfjellet on eastern Spitsbergen, Krefftberget, Svartnosa and Mistakodden on Barentsøya and Tumlingodden on Wilhelmøya. Based on field observations 15 facies have been defined. In turn, these have been grouped into eight facies associations that reflect environments on the marine offshore to lower shoreface (FA 1), delta front (FA 2) and delta plain (FA 3). The interpretations presented herein, is based on previous reseach in adjacent areas and cooperation with other master students. Thin section analysis and SEM investigations have been applied in order to complement sedimentological analysis, with data on sediment texture and composition, and to document factors affecting porosity in the sandstones.

The De Geerdalen Formation represents a major shallowing upward sequence, and the establishment of a paralic delta plain environment is documented in the upper part of measured sections. Recurring upwards coarsening and shallowing parasequences reflects a dynamic delta nature. Delta front facies on Barentsøya and Teistberget show a fluvial dominance upon deposition, in a proximal setting (relative to a westwards prograding distributary system) compared to visited outcrops on Spitsbergen. Towards the southern (Agardhbukta) and northern exposures (Hellwaldfjellet and Wilhelmøya) thinning of delta front sandstones (FA 2) and increased influence of basinal processes upon sedimentation is documented.

Distribution of porosity in the studied deltaic sandstones is largely controlled by diagenetic processes. Locally, early calcite cement has inhibited mechanical compaction, allowing an open sandstone fabric and intergranular porosity where cementation shows a patchy distribution. However, calcite cement is commonly extensive and the majority of porosity occurs as poorly connected molds. Other authigenic minerals include siderite, kaolinite, illite, chlorite, albite and quartz, with precipitation often associated with dissolution of unstable lithic fragments. Mechanisms affecting porosity are closely related to the immature composition of the De Geerdalen Formation sandstones.

Sammendrag

Denne oppgaven presenterer og tolker resultater fra sedimentologiske og diagenetiske undersøkelser av triassiske sedimenter i De Geerdals-formasjonen på østre Svalbard. I tillegg til innsamling av sandsteinsprøver har ni seksjoner blitt målt. Disse inkluderer; Klement'evfjellet, Friedrichfjellet og Smidtberget i Agardhbukta, Teistberget og Hellwaldfjellet på østre Spitsbergen, Krefftberget, Svartnosa og Mistakodden på Barentsøya samt Tumlingodden på Wilhelmøya. Baser på undersøkelser av blotninger har 15 facies blitt definert. Disse danner komponenter i åtte definerte facies assosiasjoner, som representerer grunnmarine miljø, delta front avsetninger og strandskråning til paralisk delta slette miljø. Tolkningen som presenteres her er basert på tidligere forskning i nærliggende områder, samt samarbeid med andre masterstudenter. Tynnslips analyser og SEM undersøkelser har blitt gjennomført med hensikt å utfylle sedimentologiske tolkninger, med data om sediment-tekstur og sammensetning, samt dokumentere faktorer som påvirket porøsitet i sandsteinene.

De Geerdals-formasjonen er kartlagt som en stor oppgrunnings-sekvens, og gjenspeiler etableringen av et paralisk delta slette miljø i øvre del av formasjonen. Gjentatte oppgrunnende og oppgrovende parasekvenser er tolket å reflektere et dynamisk delta miljø. Delta front avsetninger på Barentsøya og Teistberget viser fluvial dominans, og er tolket som proximale (i forhold til et vestlig progaderende avsetnings-system), sammenliknet med observerte blotninger på Spitsbergen. Delta front sandsteiner i sørlige (Agardhbukta) og nordlige eksponeringer (Hellwaldfjellet og Wilhelmøya) er tynnere og viser tegn til økt inflytelse av marine prosesser under avsetning.

Porøsitet i sandsteinene virker i stor grad å være styrt av diagenetiske prosesser. Tidlig kalsitt sement har hemmet mekanisk kompaksjon, og slik tillatt sedimentene å beholde en åpen tekstur hvor intergranulær porøsitet er bevart dersom sementering utviser ujevn fordeling. Kalsitt sement er ofte omfattende og mesteparten av porerommene opptrer som moldisk porøsitet. Autigene mineraler omfatter sideritt, kaolinitt, illitt, kloritt, albitt og kvarts, hvor utfelling ofte er tilknyttet ustabile litiske bergartsfragmenter. Mekanismer som påvårker porøsiteten i De Geerdals-formasjonens sandsteiner er nært tilknyttet den umodne komposisjonen.

Preface

This thesis is part of a master`s degree in geology at Department of Geology and Mineral Resources Engineering at NTNU. Professor II at the Norwegian University of Science and Technology Atle Mørk has been the main supervisor. Professor at NTNU Mai Britt Mørk and Professor at UNIS Snorre Olaussen has been co-supervisors.

The work presented herein, is part of ongoing investigations of the Upper Triassic succession on eastern Svalbard conducted by students at NTNU. Cooperation with other master students has been essential during fieldwork and later during data analysis. Chapter 3 and Chapter 4 (except section 4.4) in this thesis has been written as a collaboration with Turid Haugen and Sondre Krogh Johansen.

Acknowledgements

First and foremost I would like to thank Atle Mørk for offering me the unique opportunity to do fieldwork on eastern Svalbard, and helpful discussions. My thanks are further extended to Mai Britt Mørk for assistance during microscopy and SEM. I would also like to thank my co-supervisor Snorre Olaussen.

My field-companions in East Svalbard Triassic Research Group – NTNU is greatly acknowledged for long and joyful days in the field. Especially Sondre Krogh Johansen and Turid Haugen are thanked for cooperation and discussions during longs days of writing. Furthermore, I want to thank Jonas Enga and PhD Gareth Lord for helpful corrections. Nina Bakke, Bård Heggem and Cathinka Schaaning Forsberg are thanked for assistance in the field.

The Norwegian Petroleum Directorate, SINTEF Petroleum Research, Capricorn and UNIS is thanked for organizing, financing and logistical support. Thus enabling aspiring students to accsess outcrops in this vastly uncharted area.

Kjetil Eriksen is kindly thanked for assistance with SEM analysis and thin section preparation.

Finally, I would like to thank my friends and family for their support and patience.

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1 Introduction

1.1 Study area

This study was mainly conducted in the northern Storfjorden area on eastern Svalbard. Data from Edgeøya, Barentsøya, Wilhelmøya and eastern Spitsbergen were collected and creates a geographic north-south transect of the studied sections.

Svalbard's subaerial exposure, and excellent outcrops, is a consequence of Late Mesozoic and Cenozoic tectonics as this north-western corner of the Barents Sea was emerged (Worsley, 2008; Dallmann, 2015). The Barents Sea is enclosed by the northern Norwegian and Russian coasts to the south, Novaya Zemlya to the east, Svalbard archipelagos and Franz Josefs Land in the north and the deep Atlantic Ocean margins in the west (Fig. 1.1). Thereby defining one of the world's largest continental shelves, an epicontinental sea covering 1.3 million km^2 with average water depth of 300 m (Doré, 1995). The Barents Sea is commonly divided into a western and eastern province. The western Barents Sea consist of a complex mosaic of platforms, basins and highs, reflecting several phases of tectonic activity, while a platform morphology is assigned to the Svalbard archipelago (Faleide et al., 1984; Worsley, 2008).

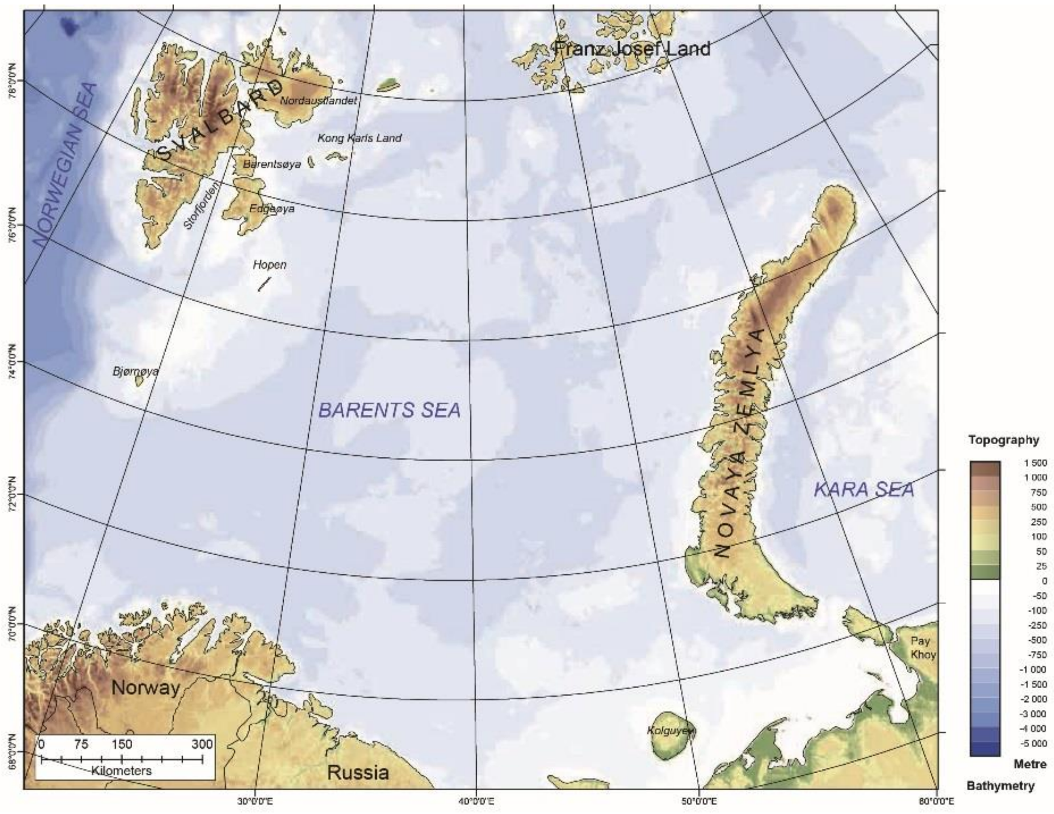

Figure 1.1: The Barents Sea is a relatively shallow sea forming part of the Arctic Ocean. From Smelror et al., 2009.

1.2 Aim

The aim of this thesis is to gain knowledge of the sedimentary and diagenetic environment of the Triassic De Geerdalen Formation on eastern Svalbard. In order to interpret the sedimentary environment results from facies analysis will be presented. Diagenetic processes and implication for porosity will also be discussed.

1.3 Previous work

The Triassic succession on Svalbard has been investigated through fieldwork expeditions since the 19th century with contributions mainly from English, Swedish, Russian, Polish and Norwegian scientists (Vigran et al., 2014). The present phase of onshore investigations was initiated in the 1970's and is engaged by the huge petroleum potential of the Barents Sea, of which the Svalbard archipelago forms excellent outcrop analogs (Worsley, 2008, Vigran et al. 2014). A comprehensive review of biostratigraphy and lithostratigraphy, collected through field work, was compiled by Vigran et al. (2014).

Recent studies have focused on the seismic expressions of clinoforms and sequence stratigraphy, often calibrated by core and outcrop data (Riis et al., 2008; Glørstad-Clark et al., 2010, 2011; Høy and Lundschien, 2011; Anell et al., 2014; Lundschien et al., 2014; Klausen et al., 2015). The nature of channel bodies have been investigated through seismic and outcrop studies on Hopen (Klausen and Mørk, 2014; Lord et al., 2014a, 2014b) and in the subsurface of the Barents Sea (Klausen et al., 2014) Among studies regarding stratigraphy and sedimentology of the eastern islands are those of Flood et al. (1971), Worsley (1973), Smith (1975), Lock et al. (1978) and Pčelina (1980, 1983). Knarud (1980) and Mørk et al. (1982) presented work regarding diagenetic and depositional environment of the Kapp Toscana Group noting the apparent differences between the Storfjorden and the Wilhelmøya subgroups. Bergan and Knarud (1993) documented regional changes in clastic mineralogy associated with this shift, while Mørk (1999) conducted a regional study on Triassic sandstone composition and provenance on the Barents Sea Shelf. Recently Mørk (2013) conducted a diagenetic study of the De Geerdalen Formation on central Spitsbergen as part of the Longyearbyen CO2 lab project. Rød et al. (2014) documented the depositional environment of the De Geerdalen Formation from Edgeøya to central Spitsbergen through facies analysis, of outcrop and core data, and geometric studies through photos and Lidar data. The work presented herein has benefited heavily from the findings of Rød et al. (2014), and forms part of ongoing investigations of the Triassic strata on eastern Svalbard conducted by NTNU students in cooperation with the NPD.

1.4 Regional geologic setting of Svalbard and the Barents Sea

The Svalbard archipelago exhibit a geological record ranging from Precambrium to recent with evidences of repeated tectonic events punctuated by quiescence and deposition of sedimentary strata. In all providing a, more or less, complete sedimentary succession through the Upper Paleozoic to Paleogene exposed across the arctic archipelago (Dallmann, 2015).

Important controls on deposition were imposed by the tectonic framework, chiefly inherited from the Caledonian and Uralian orogenies, along shelf margins and structural lineaments (Faleide et al., 1984; Doré, 1991, 1995; Gudlaugsson et al., 1998; Gee et al., 2006) and changing climatic conditions caused by northwards plate tectonic drift of the Eurasian plate (Worsley, 2008). The presently accepted lithostratigraphic subdivision of the Mesozoic is that of Mørk et al. (1999).

1.4.1 Paleozoic

The Precambrian to Silurian basement, informally referred to as "Hecla Hoek", is generally grouped into three provinces exposed in the western and northern parts of the archipelago. Consisting of gneisses, metamorphosed supracrustal and intrusive rocks it forms the substrate for deposition from Devonian to Paleogene times (Worsley, 2008; Dallmann and Elvevold, 2015).

The convergence of Laurentia and Baltica closed off the Iapetus Ocean during the Caledonian Orogeny from Early Ordovician to Early Devonian (Dallmann, 2015). Old Red Sandstones, of the Andrée Land Group, representing an arid climate were deposited in Devonian basins formed by subsequent orogenic collapse and extension (Worsley, 2008; Smelror et al., 2009; Blomeier, 2015). At present these "Old Red Sandstone" facies sediments are confined between the northwestern and eastern basement provinces in a major graben on northern Spitsbergen (Blomeier, 2015). A transition from red to grey sediments in early to mid-Devonian reflects the archipelagos northwards drift as the climate changed from the southern arid zone to equatorial latitudes (Worsley, 2008).

Late Devonian to Early Carboniferous half-graben development relates to inherited Caledonian structural lineaments with contemporaneous deposition of the fluvial and lacustrine clastics of the Billefjorden Group (Smelror et al., 2009). The group lays uncomformably on Old Red Sandstones or "Hecla Hoek" sequences (Steel and Worsley, 1984). A warm humid climate is reflected by abundant coal seams (Steel and Worsley, 1984).

A climatic change to arid conditions occurred in the Upper Mississippian, as the regionally erosive basal red beds of the Gipsdalen Group are overlain by warm-water carbonates (Stemmerik and Worsley, 2005). A late Carboniferous transgression led to the development of an extensive warm-water carbonate platform with deposition of sabkha deposits and clastics in marginal marine environments (Steel and Worsley, 1984; Smelror et al., 2009). The Gipsdalen-Bjarmeland Group transition involves an abrupt shift to cool-water carbonates, possibly caused by changing oceanographic conditions related to the developing Uralides (Stemmerik and Worsley, 2005).

The silica rich spiculithic shales of the Tempelfjorden Group reflects closure of seaway connections to the warm Tethys Ocean and the developing Uralides in the east, terminating the stable carbonate environment prevalent for almost 40 Ma (Worsley, 2008). The Permian/Triassic transition is not fully understood, but involves a hiatus in the latest Permian on highs and platforms (Mørk et al., 1999; Worsley, 2008; Vigran et al., 2014) and an abrupt change from spiculithic shales to the non-siliceous shales of the Sassendalen Group (Vigran et al., 2014).

1.4.2 Mesozoic

During the Triassic the Barents Sea region formed a large shallow epicontinental sea, bound by the continental provinces of Greenland to the west, Ellesmere Land to the northwest, Fennoscandia to the south and Siberia, the Urals, Novaya Zemlya and Taimyr to the east (Fig. 1.2).

Sea level fluctuations, paleotopography and temporal changes in sediment dispersal from the bordering hinterlands created a complex pattern of sedimentary units infilling the embayement (Mørk et al., 1982; Mørk et al., 1989; Mørk et al., 1999; Riis et al., 2008; Glørstad-Clark et al., 2010, 2011; Høy and Lundschien, 2011; Lundschien et al., 2014).

Inherited Late Paleozoic structural elements imposed controls on accommodation space development, but in general the Triassic is considered a period of relative tectonic quietness on Svalbard and in the western Barents Sea (Faleide et al., 1984; Glørstad-Clark et al., 2010). Recent work by Anell et al., (2013) and Osmundsen et al. (2014) suggest a tectonic influence on syn-sedimentary growth faulting (Edwards, 1976) on Kvalpynten, Edgeøya.

The soft shales and siltstones with subordinate sandstones of the Early to Middle Triassic Sassendalen Group ended the “Permian Chert Event” (Mørk et al., 1999; Worsley, 2008; Vigran et al., 2014) marking the onset of Mesozoic clastic deposition. The Sassendalen Group comprise three coarsening upwards sequences initiated after pronounced transgressions (Mørk et al., 1982; Mørk et al., 1989; Vigran et al., 2014). Barrier bar or deltaic sandstones on western Spitsbergen indicates coastal progradations from Greenland to the west, whereas distal shelf mudstones are found east and southwards (Mørk et al., 1982; Mørk et al., 1999; Vigran et al., 2014).

The dark, organic rich, phoshpatic shales of the Botneheia Formation on eastern Svalbard and the proximal time-equivalent Bravaisberget Formation on western Spitsbergen reflects anoxic or periodic anoxic conditions in the Middle Triassic (Mørk et al., 1982; Mørk et al., 1999; Lundschien et al., 2014), with the Steinkobbe Formation representing the same continuous depositional environment in the Barents Sea (Mørk and Elvebakk, 1999; Lundschien et al., 2014). The Botneheia Formation show TOC values up to 10% of type II/III kerogen and is a prolific hydrocarbon source rock (Mørk et al., 1982; Leith et al., 1993; Lundschien et al., 2014; Vigran et al., 2014).

The Middle Triassic to Middle Jurassic Kapp Toscana Group marks a shift in deposition by repeated deltaic and coastal progradation from various source areas with the Uralides as the dominant provenance area.

In latest Middle Jurassic to Early Cretaceous rising sea levels and crustal extension led to development of a shallow, extensive marine shelf area with deposition of the mudstone-dominated Adventdalen Group (Dypvik et al., 1991; Mørk et al., 1999; Worsley, 2008; Smelror et al., 2009). High levels of atmospheric CO2 and periodic anoxic bottom water conditions gave organic rich intervals in the basal Agardhfjellet Formation (Olaussen, 2015). On Svalbard the fluviodeltaic sandstones of the Helvetiafjellet Formation represents southerly directed deltaic build outs, overlying an unconformity caused by uplift and erosion of the northern margin (Steel and Worsley, 1984; Grundvåg, 2015).

Early Cretaceous igneous rocks displaying similar geochemistry are found over large parts of the Arctic and collectively referred to as the High Arctic Igneous Province (Nejbert et al., 2011; Senger et al., 2014). Equivalent sills and dykes of the Diabasodden Suite on Svalbard (Mørk et al., 1999) are seen to intrude the Mesozoic strata. Corfu et al. (2013) dated sills of the Diabasodden Suite from Svalbard and Franz Josefs Land to 125 Ma.

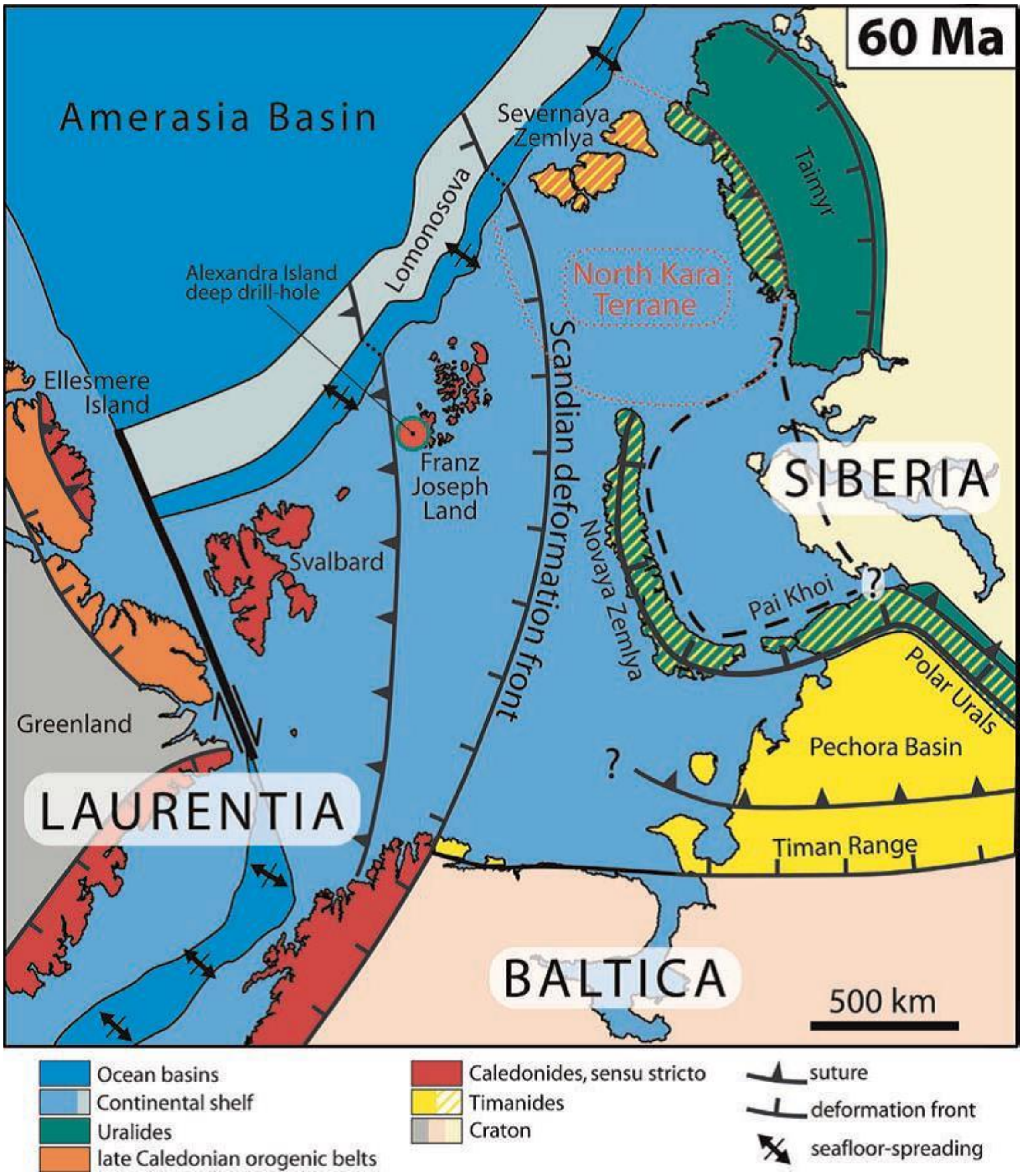

Figure 1.2: Barents Sea hinterlands and hypothetical main structural elements during initial phases of the Eurasian Basin and the Norwegian-Greenland Sea opening. From Gee et al. (2006).

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1.4.3 Cenozoic

Transpressional stresses along the western margin of the shelf commenced in the formation of the West Spitsbergen Fold- and Thrust-Belt in Paleogene (Bergh et al., 1997; Braathen et al., 1999; Smelror et al., 2009). Subsequently the Spitsbergen Central Basin developed as a foreland basin east of the orogeny and was filled by clastic sediments by Eocene (Steel and Worsley, 1984; Worsley, 2008). These sediments belonging to the Van Mijenfjorden Group lie above a regional unconformity capping Lower Cretaceous deposits (Dallmann, 2015). Eastern Svalbard, where this study is conducted, exhibit more of a platform morphology (Fig. 1.3) (Dallmann, 2015) as this area lay distal to the tectonic events in the west. The dextral strike-slip regime prevailed and resulted in the opening of the Norwegian-Greenland Sea and formation of oceanic crust from Early Eocene (Faleide et al., 1984; Smelror et al., 2009).

In Neogene and Quaternary the cooling of climate continued and resulted in repeated glaciations with alternating subsidence and uplift of the region (Mangerud et al., 1996; Dallmann, 2015). Erosion of the entire shelf area, up to 3 km in northern areas (Worsley, 2008), generated thick wedges of glacigenic sediments accumulating along the western margin (Smelror et al., 2009).

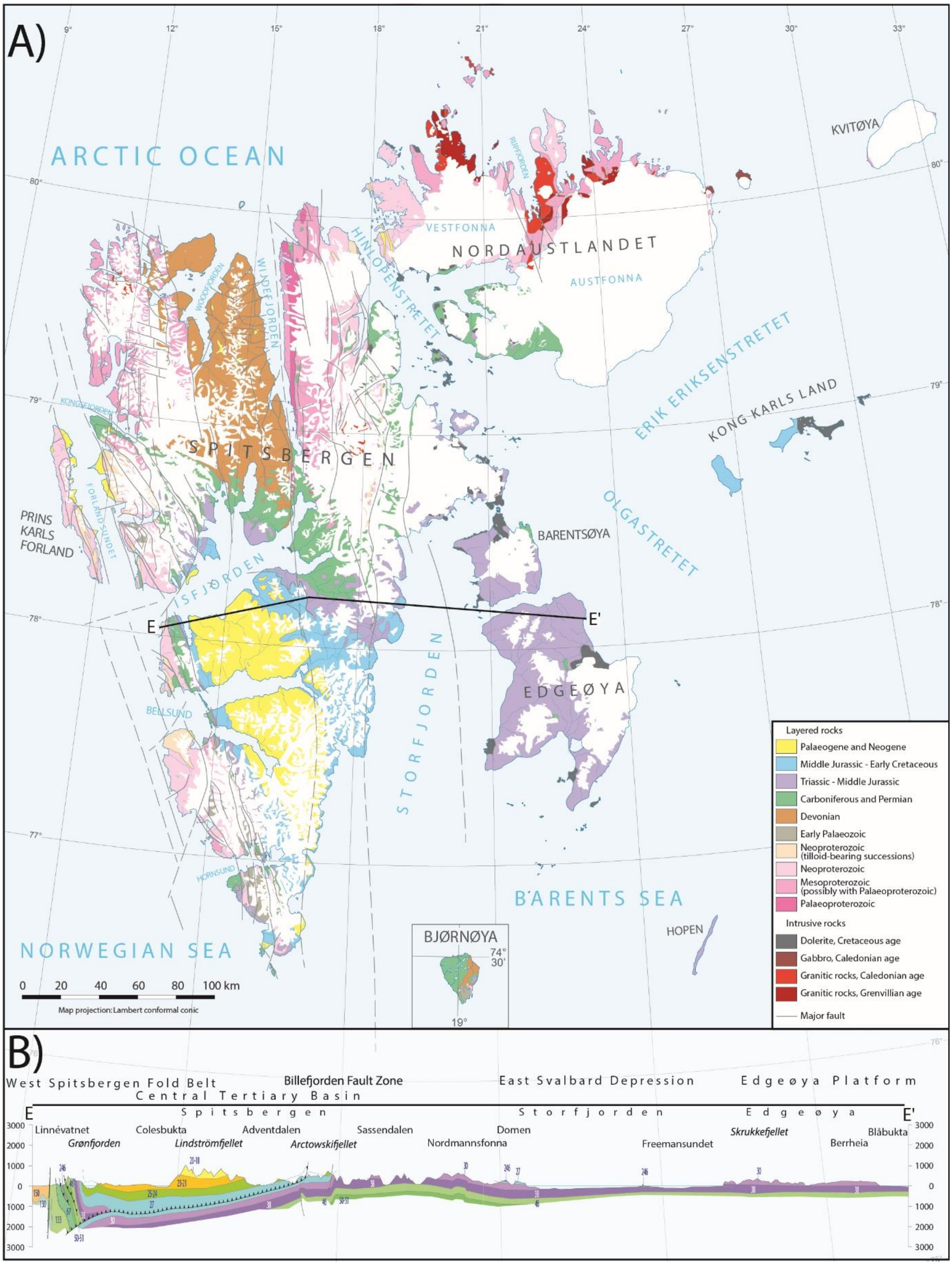


Figure 1.3: A) Bedrock geology map of the Svalbard archipelago. B) Vertically exaggerated cross-section displaying the deformed sedimentary strata in the west and platform areas in the east (modified from Dallmann, 2015).

1.5 Kapp Toscana Group

The Kapp Toscana Group reflects the establishment of a deltaic platform progressively infilling the northerly facing Triassic embayment from the south-east and marks a considerable change in sedimentation patterns from underlying units (Riis et al., 2008; Lundschien et al., 2014; Vigran et al., 2014). The group comprises Late Triassic to Middle Jurassic upwards coarsening shales to sandstones (Mørk et al., 1999). A threefold subdivision of the Kapp Toscana Group is constituted by the Ladinan to Norian Storfjorden Subgroup, Norian to Bathonian Wilhelmøya Subgroup and its offshore equivalent the Realgrunnen Subgroup (Mørk et al., 1999), marking abrupt changes in sedimentation patterns and mineralogy initiated by a regional transgressional event in the Norian (Bergan and Knarud, 1993; Mørk et al., 1999; Worsley, 2008)

In general the group is deposited in a nearshore environment and displays coastal reworking upon the deltaically introduced sediments (Mørk et al., 1982). An overview of the accepted Early Triassic to Middle Jurassic lithostratigraphy of Svalbard and the Barents Sea is given in Figure 1.4, and is that of Mørk et al. (1999) with the recently defined Hopen Member included (Lord et al., 2014).

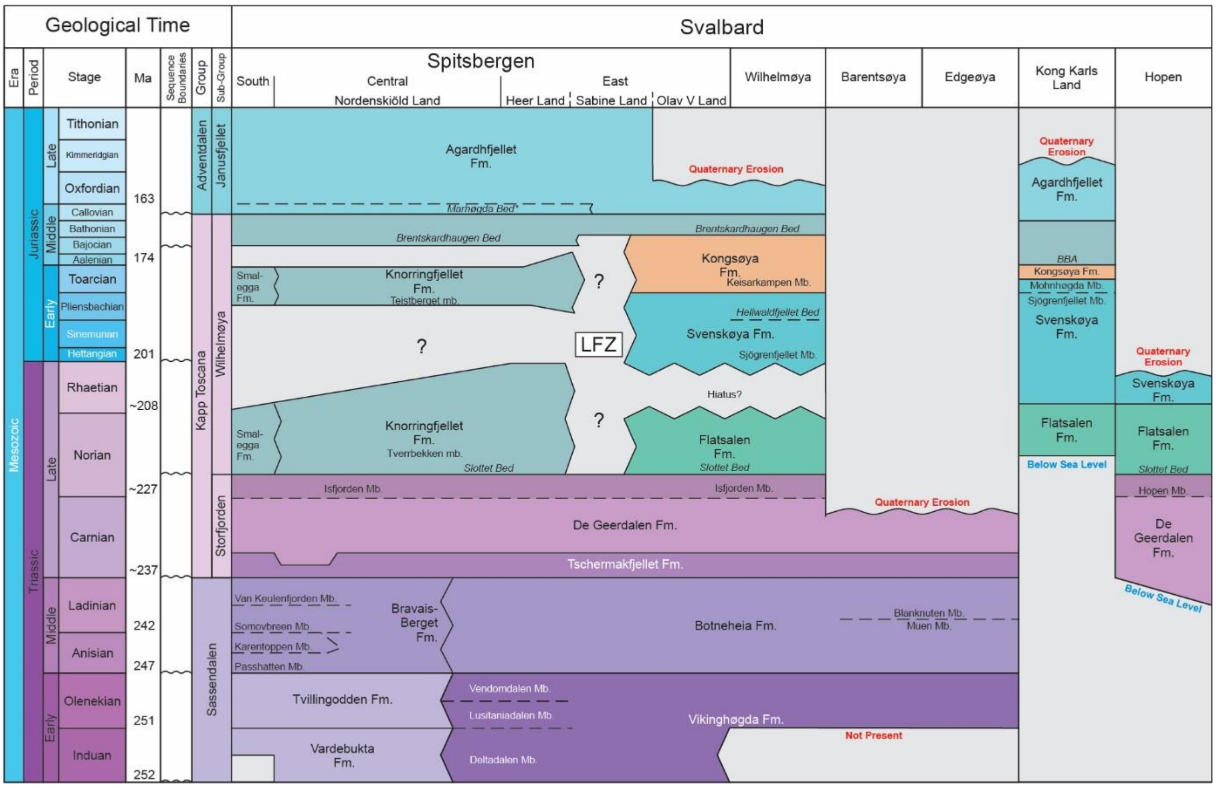


Figure 1.4: Triassic to Middle Jurassic lithostratigraphy of Svalbard and the Barents Sea. From Lord et al. (in prep).

1.5.1 Storfjorden Subgroup

The Storfjorden Subgroup comprises shallow marine and coastal deposits of claystones and immature sandstones extending vast areas of Svalbard, Bjørnøya (Skuld Formation) and the Barents Sea (Snadd Formation) of Ladinian to Norian age (Mørk et al., 1999). Together the basal Tschermakfjellet Formation and the overlying De Geerdalen formation constitute the distal equivalent of the Snadd Formation in the Barents Sea, all representing the same continuous depositional system prograding across the Barents Sea shelf (Riis et al., 2008, Lundshcien et al., 2014).

The Tschermakfjellet Formation consists of grey shales and interbedded siltstones with red weathering siderite nodules in an upwards coarsening sequence (Mørk et al., 1999). The prodeltaic shales exhibit an eastwards thickening wedge shape and a diachronous lower boundary (Mørk et al., 1999; Vigran et al., 2014).

The base of the De Geerdalen Formation is defined at the base of the first occurring prominent sandstone in the Storfjorden Subgroup (Flood et al., 1971; Lock et al., 1978; Mørk et al., 1999). The formation consists of recurring upwards coarsening successions of interbedded silt and sandstones of Carnian to early Norian age (Tozer and Parker 1968; Korčinskaja, 1982; Mørk et al., 1999; Vigran et al., 2014) were observed rhythmicity has been attributed to switching of delta lobes (Mørk et al., 1982; Mørk et al., 1999).

The upper part of the De Geerdalen Formation on western and central Spitsbergen, and Wilhelmøya is assigned to the Norian Isfjorden Member (Pčelina 1983; Mørk et al., 1999). Alternating sandstones and shales of characteristic red and green colour were deposited in a shallow marine environment with locally lagoonal conditions (Mørk et al., 1999). On Hopen an increased marine influence upon sedimentation is recorded in the uppermost parts of the De Geerdalen Formation and motivated the definition of the Hopen Member (Mørk et al., 2013; Lord et al., 2014). The member consist of marine shales and subordinate sandstones of latest Carnian to earliest Norian age and is considered time-equivalent to the Isfjorden Member on central Spitsbergen (Lord et al., 2014).

Investigaions of the De Geerdalen Formation on the westernmost outcrops on Spitsbergen points to a western sediment source (Mørk et al., 1982) and is supported by seismic studies (e.g. Glørstad-Clark, 2010). Wave reworking and redistribution of the deltaically introduced sediments is noted from central Spitsbergen (Knarud, 1980), whereas eastern areas exhibit more deltaic control upon sedimentation (Mørk et al., 1982). This development is consistent with observations by Rød et al. (2014) who documents a proximal fluvially dominated deltaic setting on Edgeøya with generation of ellipsoid sandstone body geometries. A distal setting is interpreted on central Spitsbergen with wave and tidal reworking yielding thin and laterally continuous sandstone bodies. Mud drapes and herringbone structures were reported in distributary channels on Central Spitsbergen, whereas distributary channels on Edgeøya lack tidal signatures (Rød et al., 2014).

On Hopen an even more proximal paralic setting is illustrated by distinct fluvial dominated and tidally influenced channel sandstones with west to northwest paleoflow directions (Klausen and Mørk, 2014; Lord et al., 2014b). Size and geometries of these channels are showed to be comparable to channels studied through seismic methods from the equivalent Snadd Formation (Klausen and Mørk, 2014).

These observations from outcrop studies are consistent with the regional development as mapped by seismic methods and core studies (Riis et al., 2008; Glørstad-Clark et al., 2010; Høy and Lundschien, 2011; Lundschien et al., 2014) documenting a prograding paralic platform infilling the Western Barents Sea from the southeast. Four Carnian sequences, up to 400 m thick, are recognized and the uppermost sequence outcrops on Hopen (Lundschien et al, 2014), thus the exposed channel belts on Hopen exhibit the paralic nature of this prograding system (Lord et al., 2014b). Seismic clinoforms have been traced to Kvitøya (Høy and Lundschien, 2011; Lundschien et al., 2014), and some have suggested further progradation and a possible convergence with sediment systems from eastern Greenland (Klausen et al., 2015).

Basement rocks from the Fennoscandian Shield sourced northeasterly directed clinoforms in the Early Triassic, but a growing Uralide component gradually shifted the direction of progradation towards northwest and finally towards west-northwest (Riis et al., 2008; Høy and Lundschien, 2011; Lundschien et al, 2014) establishing a paralic platform gradually infilling the western Barents Sea (Fig. 1.5).

In general, a gentle relief is assigned to the Triassic paleotopography but emergent features like the paleo-Stappen High (Worsley et al., 2001) and the paleo-Loppa high acted as barriers for sediments prograding from the southeast (Faleide et al., 1984; Glørstad-Clark, 2010). The paleo-Loppa high was emergent until early Ladinian times and subsequently developed into a major deposenter (Glørstad-Clark, 2010).

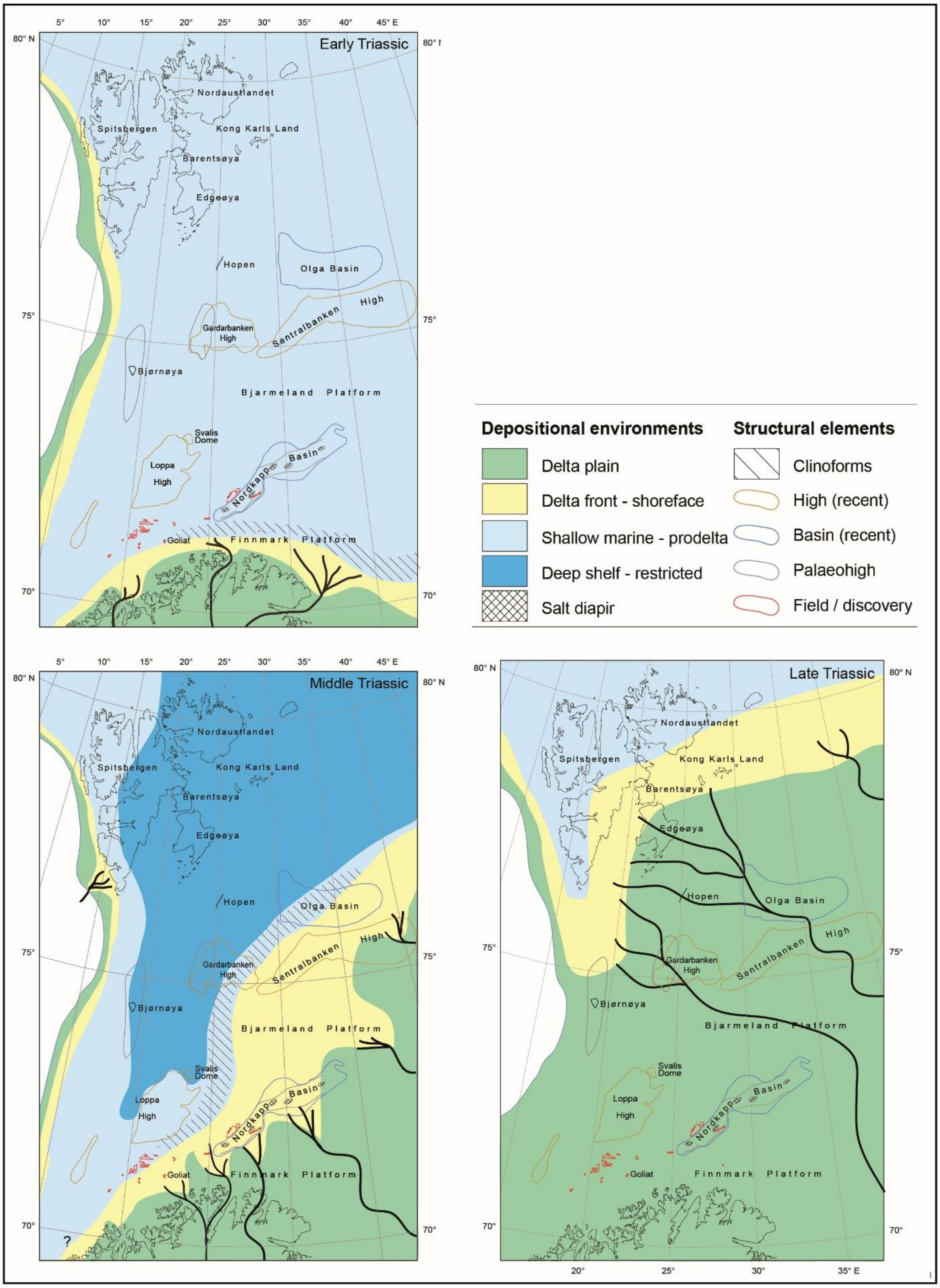


Figure 1.5: Early to Late Triassic paleogeographic reconstruction from Lundschien et al. (2014).

From Middle Triassic to Middle Jurassic times the Svalbard archipelago drifted northwards from about 45 to 60 degrees implying a shift from arid to humid climate (Steel and Worsley, 1984; Worsley 2008; Glørstad-Clark et al., 2011). Palynological studies by Hochuli and Vigran (2010) indicates a shift to humid conditions in early Carnian, which coincides with presence of coal in the studied cores. A humid climate with seasonal variations in precipitation in the Carnian is also suggested by Enga (2015).

1.5.2 Wilhelmøya Subgroup

A pronounced transgression in Norian time marks the onset for deposition of the Wilhelmøya Subgroup on Svalbard and the time-equivalent Realgrunnen Subgroup in the Barents Sea (Mørk et al., 1999; Vigran et al., 2014). The subgroup shows an increasingly condensed development towards west and northwest (Mørk et al., 1999). Uplift and erosion in Late Triassic-Early Jurassic led to the development of emerged lowlands on the shelf (Smelror et al., 2009). Deltaic to shallow marine depositional environments prevailed, but aggradation and reworking of sediments were dominant processes, as opposed to preceding progradation and subsidence (Mørk et al., 1999; Olaussen, 2015). The dominating Uralian source was largely replaced by multiple source areas e.g. the rejuvenation of the Fennoscandian Shield in the south and Greenland in the west and local emergent highs (Smelror et al., 2009; Bue and Andresen, 2013; Ryseth, 2014).

The basal Slottet Bed consists of condensed calcareous, phosphate-rich shallow marine sandstones (Mørk et al., 1999; Olaussen, 2015) and represents a condensed shelf deposit caused by the Norian transgression (Mørk et al., 1999; Worsley, 2008; Bergan and Knarud, 1993). The Norian marine transgression is tracable across Svalbard, the Barents Sea and the Sverdrup Basin and initiated deeper water conditions with deposition of the red-brown to purplish weathering shales of the Norian Flatsalen Formation (Olaussen, 2015).

Above are greenish, fine- to medium-grained sandstones grading into fine- to very-fine white sandstones with interbedded thin mudstone beds assigned to the Svenskøya Formation (Mørk et al., 1999), deposited on a tidally dominated coastal-plain (Mørk et al., 1999; Olaussen, 2015). Above the cliff-forming sandstones of the Svenskøya Formation are the shallow marine fine-grained sandstones and greyish-blue mudstones with siderite beds of the Kongsøya Formation (Mørk et al., 1999).

In western Spitsbergen the Knorringfjellet Formation constitute the condensed equivalent of the Flatsalen, Svenskøya and Kongsøya formations (Mørk et al., 1999). The subgroup is terminated by a regionally extensive layer of phosphate conglomerate of Bajocian (?) – Bathonian age referred to as the Brentskardhaugen Bed (Mørk et al., 1999).

1.5.3 Mineralogy and diagenesis of the Kapp Toscana Group

The coastal to deltaic lithic arenites of the De Geerdalen Formation are considered texturally and mineralogically immature (Flood et al., 1971; Knarud, 1980; Mørk et al., 1982, Bergan and Knarud, 1993; Mørk et al., 1999; Mørk 1999, 2013). Relatively similar detrital compositions, with a characteristic high content of plagioclase and lithic rock fragments (Fig. 1.6), are found throughout the Svalbard archipelago and in the equivalent Snadd Formation of the Barents Sea (Bergan and Knarud, 1993; Mørk, 1999; Riis et al., 2008). Provenance studies show a dominating Uralide component with possible minor input from the Timanides (Mørk, 1999; Bue and Andresen, 2013), proving consistent with seismic studies.

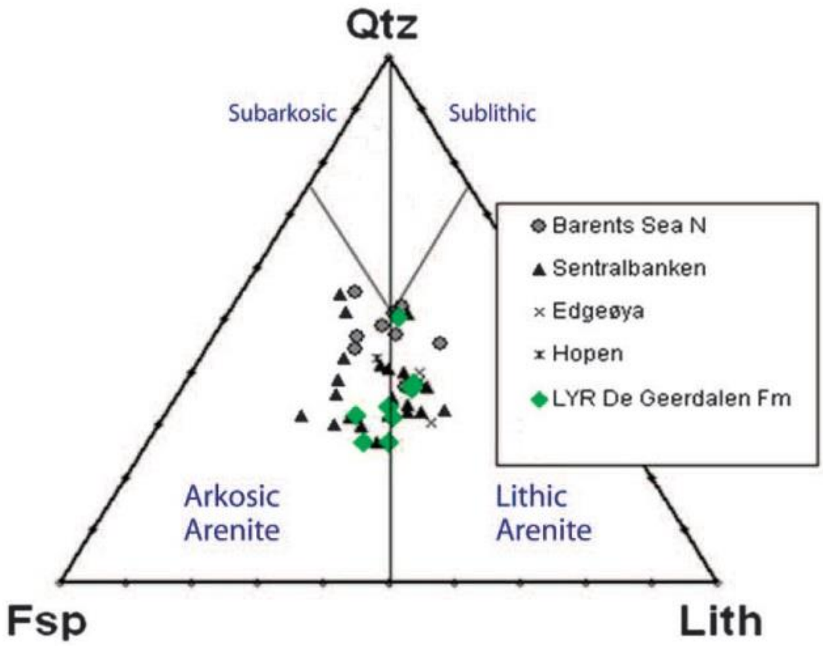


Figure 1.6: Bulk modal analysis of the immature sandstones in the De Geerdalen Formation and the equivalent Snadd Formation on Svalbard and the northern Barents Sea, from Mørk (2013).

Extensive dissolution and remineralization of labile grains has been attributed to a diagenetic environment in thermodynamic inequilibrium with restricted porewater flow (Mørk et al., 1982), with lithofacies and detrital compositions imposing controls on distribution of diagenesis

(Mørk, 2013). Early mechanical compaction was the main agent for reducing primary porosity but was locally inhibited by early carbonate cementation forming in close association to carbonate grains and biogenic fragments (Mørk et al., 1982). Magmatic intrusions, representing the Diabasodden Suite, cooled rapidly under shallow conditions and was possibly accompanied by local hydrothermal activity (Mørk, 2013).

In contrast, the overlying sandstones of the Wilhelmøya Subgroup represent an open diagenetic system with periodically high porewater flux (Mørk et al., 1982). In general, these sandstones are mineralogically and texturally mature and classified as quartz arenites (Mørk et al., 1982; Olaussen et al., 1984). Excellent primary porosity was locally reduced by extensive quartz and calcite cementation, with pressure solution and stylolite development as responsible mechanisms (Olaussen et al., 1984). Marine eogenetic minerals are characteristic components of conglomerate horizons representing condensed sections and are frequently found in the Wilhelmøya Subgroup (Mørk et al., 1999; Mørk, 2013).

2 Database and Methods

2.1 Fieldwork localities

Fieldwork was conducted in the Storfjorden area on eastern Svalbard (Fig. 2.1) from 2nd to 28th August 2015. The expedition was organized in cooperation with SINTEF Petroleum Research and the NPD, with logistical support from UNIS. In Agardhbukta complete sections of Klement'evfjellet, Smidtberget and Friedrichfjellet were measured. These sections have previously been measured by Knarud (1980) who documented the De Geerdalen Formation and the overlying Wilhelmøya Subgroup at Klement'evfjellet, Smidtberget and the adjacent Krapotkinfjellet. These logs are published in Mørk et al. (1982) and Vigran et al. (2014).

For the remainder of the field campaign, use of the vessel M/V Sigma allowed access to localities on Edgeøya, Barentsøya, Wilhelmøya and eastern Spitsbergen. On Tumlingodden, Wilhelmøya, a complete section through the De Geerdalen Formation and the Isfjorden Member was obtained. Here it is overlain by the Wilhelmøya Subgroup represented by the Flatsalen, Svenskøya and Kongsøya formations (Worsley, 1973; Vigran et al., 2014). The localities of Hellwaldfjellet, Hahnfjella and Teistberget were measured on eastern Spitsbergen and the presence of the Isfjorden Member in the upper part of the De Geerdalen Formation was confirmed. Notably a sill, belonging to the Diabasodden Suite (Mørk et al., 1999), were seen to crosscut the Isfjorden Member on both Wilhelmøya and Hellwaldfjellet.

On Barentsøya the localities of Mistakodden, Krefftberget and Svartnosa were visited. Only the presence of Tschermakfjellet Formation and the lower part of the De Geerdalen Formation is documented from here, as the strata above has been removed by Cenozoic erosion (Flood et al., 1971, Vigran et al., 2014). On Edgeøya a complete section on Muen was obtained. Here, the Vikinghøgda, Botneheia, Tschermakfjellet and De Geerdalen formations are exposed, with the upper part of the De Geerdalen Formation being eroded. On Muen, as on Barentsøya, the characteristic red beds and coquina beds of the Isfjorden Member were not found. A further description of the visited localities is provided in Johansen (in press.).

The localization of logged sections constitutes a roughly 125 km north-south transect from Wilhelmøya in the north to Agardhbukta and Muen in the south. The localities also cover roughly 60 km east-west distance from Krefftberget and Muen in the east to Friedrichfjellet in the west. Legend and sedimentary sections from the various localities are attached in Appendix A.

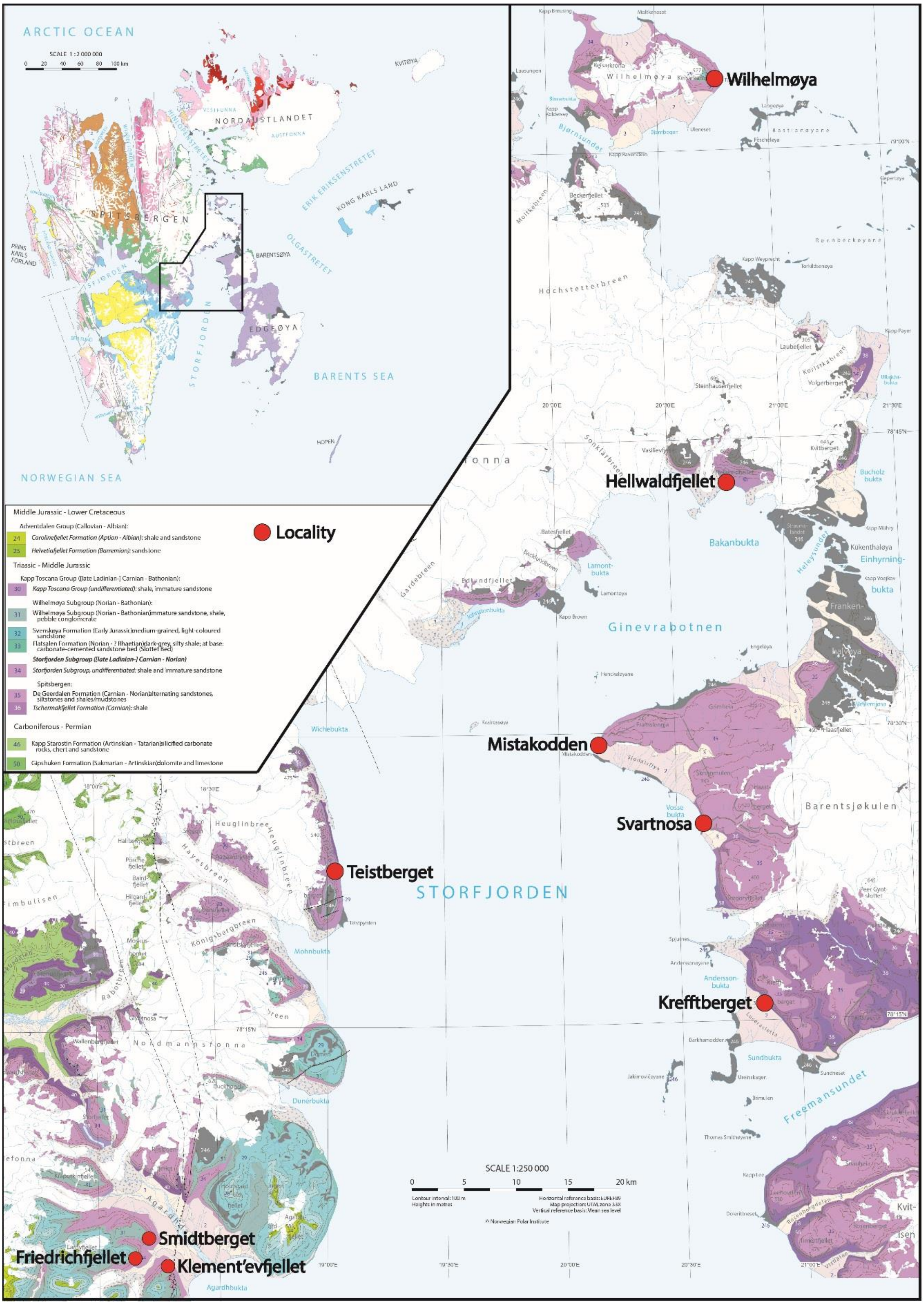


Figure 2.1: Overview of visited localities on eastern Svalbard. Modified from Dallmann (2015).

2.2 Practical work and method

2.2.1 Logging

Sections were measured with a meter stick and sedimentological logs were drawn in 1:100 scale in the field. Particular attention was given to lithology, sedimentary structures, grain size trends, mud:sand ratio, unit thicknesses and unit boundaries. Grain sizes were estimated using a handlense and a standard grain size sheet. Presence of calcite cementation was tested by the use of hydrochloric acid. GPS was used to determine coordinates and altitude at the base and top of each measured section. Upon return from the field the retrieved handwritten logs have been graphically digitized using Adobe Illustrator drawing program. Facies analysis have been conducted on logs recorded in the field, and methodology will be discussed later on.

Sections were measured, logged and sampled in teams consisting of 2 to 3 master students from NTNU/UNIS. Nina Bakke, Cathinka Schaaning Forsberg, Turid Haugen, Bård Heggem, Sondre Krogh Johansen and PhD candidate Gareth Lord participated in the fieldwork.

2.2.2 Optical Microscopy

Petrographic thin sections have been examined in order to describe authigenic minerals, with observations of sediment textures and grain-size providing a supplement to the sedimentological analysis. Thin sections of various sandstone types were selected from different stratigraphic levels and geographic locations. Analysis for authigenic textures and mineralogy was conducted using a standard petrographic microscope under plane polarized light and crossed polarized light. Visual comparators were used to estimate mean grain size, roundness and sorting. The obtained mean grain sizes were in turn used to calibrate grain sizes noted in the sedimentological logs. All thin sections were polished and porosity was stained blue by epoxy.

2.2.3 SEM

SEM electron backscatter images (BEI) are based on differences in intensity of minerals reflecting their average atomic number, and were used to interpret microstructures and textural relations. Qualitative composition and verification of mineral identification from optical microscopy was aided by energy dispersive spectra (EDS) of selected minerals.

2.2.4 Sources of error

The interpretation of depositional environment is based on observations from outcrop logging. Steep slopes and ridges were preferred when choosing logging routes in order to obtain the best possible exposures and minimize scree cover. However, scree cover is common in the sedimentological logs, and often inferred as mudstones due to their susceptibility to weathering. Thus scree cover may result in sandstone beds being underreported in the present work. Other sources of errors relate to observations and possible human misinterpretations while working in the field.

Optical microscopy was used for the petrographic analysis and potential source of errors mainly relates to the immature nature of the studied sandstones. Feldspars and lithic rock fragments are typically strongly altered and challenging to classify. Non-undulatory quartz grains may also have been misinterpreted as feldspar.

In SEM analysis possible errors are mainly related to misinterpretation of the semi-quantitative EDS spectra as they are not calibrated with standards.

3 Facies in the De Geerdalen Formation on eastern Svalbard

This chapter and Chapter 4 (except section 4.4) has been written as a collaboration between master students Turid Haugen, Sondre Krogh Johansen and Simen Jenvin Støen, who also worked together in the field. 15 facies have been described, based on field observations in eastern Svalbard. 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 is given in Table 3.1.

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 exist 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.

Perhaps the most important aspect of facies analysis is that it 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 3.1. 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 sand.

#	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), set 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 and includes marine trace fossils (*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 0.1-5 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	vf - f	-
I_1	Syn-sedimentary deformed sandstone	vf – f	Found at base or interbedded with other sand units. Plant fragments and mud clasts common. Recognized interbedded with partially to completely undisturbed sandstones.
I_2	Erosive-based sandstone lenses	vf - f	Irregular base, laterally restricted sandstone bodies. Unit width 2-4 m and unit height 0.3-1.5 m. Calcite-cemented, cone-in-cone towards the top of units. Plant fragments observed. Enclosed in mudrocks.
J	Carbonate rich sandstone	vf - f	Carbonate cemented and commonly very hard. Beds are often laterally extensive and 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) or 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. Occur within mudrocks and on top of sandstones. 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.

3.1 Facies A - Large-scale cross-stratified sandstone

Description

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. 3.1).

This facies comprises 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. 3.1E) 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. 3.1D, 3.1F) 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, caused by a dominant unidirectional current in a subaqueous environment. Mud drapes are attributed to slight changes in current velocity, possibly implemented by tidal activity or seasonal changes in stream discharge.

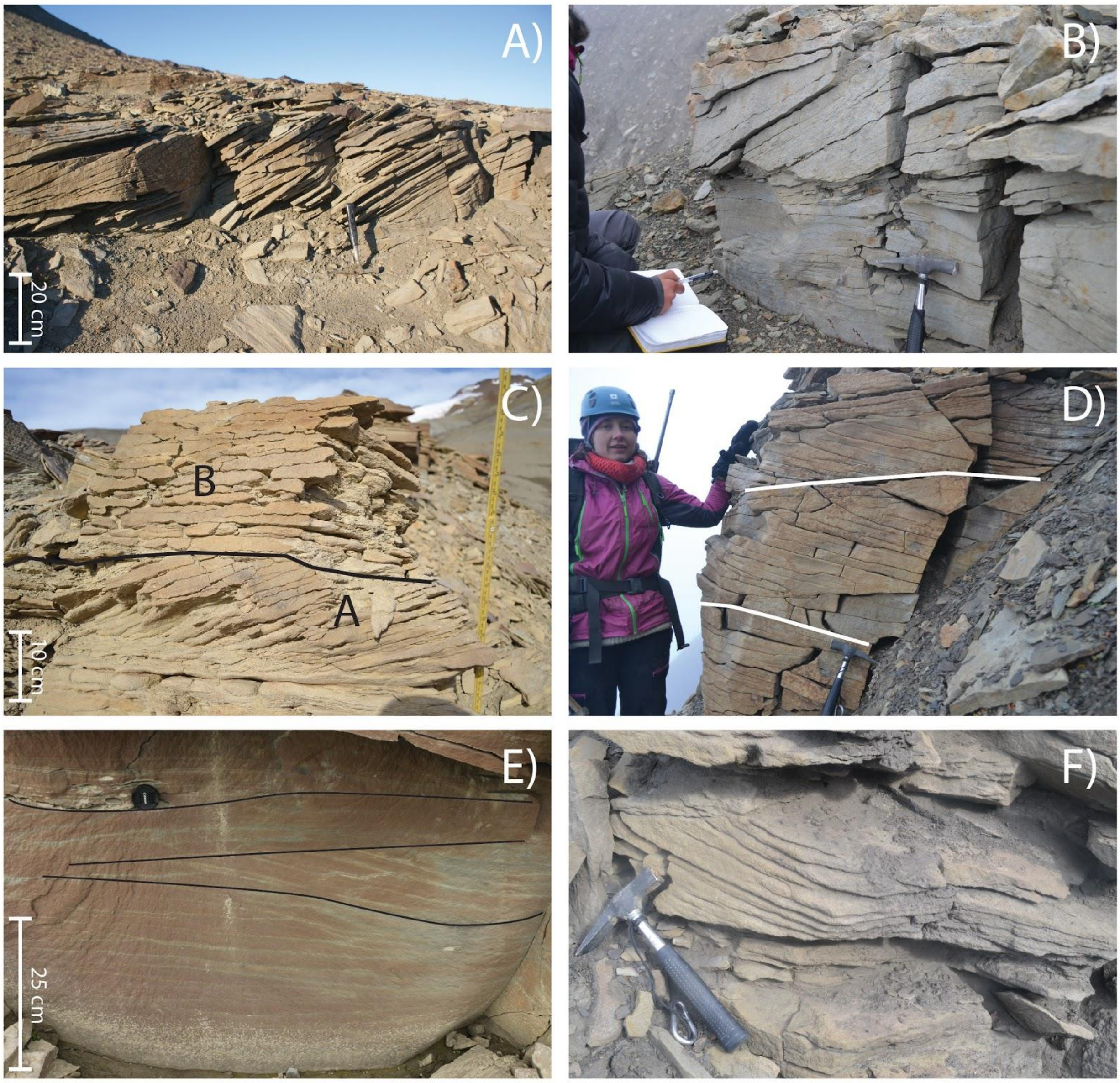


Figure 3.1: Facies A - Large-scale cross-bedded sandstone A) Tabular cross-bedded sandstone at Wilhelmøya. B) Trough cross-bedded sandstone at Šmidtberget, Agardhbukta. C) Tabular cross-bedded sandstone (facies A) overlain by small-scale asymmetric cross-bedding (facies B), Hellwaldfjellet, north-eastern 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.

3.2 Facies B - Small-scale cross-stratified sandstone

Description

This facies comprises 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. 3.2). Cementation, mainly calcite, varies from vague to pervasive resulting in differences in appearance within facies. The facies often appears as undulating, parallel wavy to straight bedding/set boundaries without apparent cross-stratification (Figs. 3.2A, 3.2B, 3.2C). 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. 3.2D). 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. 3.2A).

Figure 3.2: 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 tend 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 create 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 paleosols, when found in the upper parts of the De Geerdalen Formation, indicate that this facies commonly is associated with terrestrial 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 appears in marine environments.

3.3 Facies C - Climbing ripple cross-laminated sandstone

Description

Small-scale asymmetric climbing ripple laminated very fine to fine sandstone (Fig. 3.3). Yellow, orange and brownish colors 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).

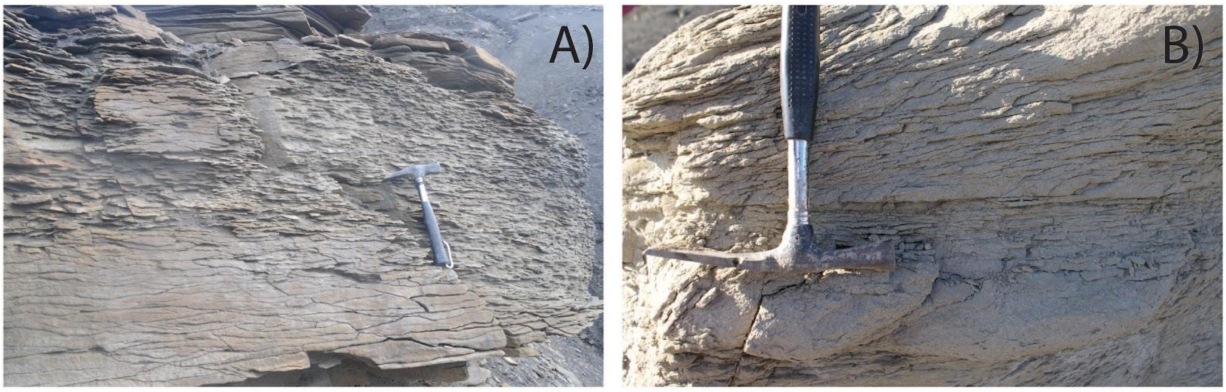


Figure 3.3: Facies C - Climbing ripple cross-laminated sandstone. A) Climbing ripple cross-lamination in the lower part of De Geerdalen Formation on Svartnosa, Barentsøya B) Climbing ripples towards the top of Wilhelmøya.

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, are favourable for this facies formation, whereas environments of low sedimentation rates and much reworking are 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 occur 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.

3.4 Facies D - Wave rippled sandstone

Description

This facies is assigned to units of very fine to fine sandstone that is ripple cross-laminated (Fig. 3.4). 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 and contain siderite concretions and nodules

Wave ripples are often seen on the 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. 3.4A, 3.4B, 3.4D). In cross-section, wave ripples are recognized by having undulating bedding, sometimes with bidirectional foresets (Figs. 3.4C, 3.4E, 3.4F). Mud drapes are also common within the facies and help 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.

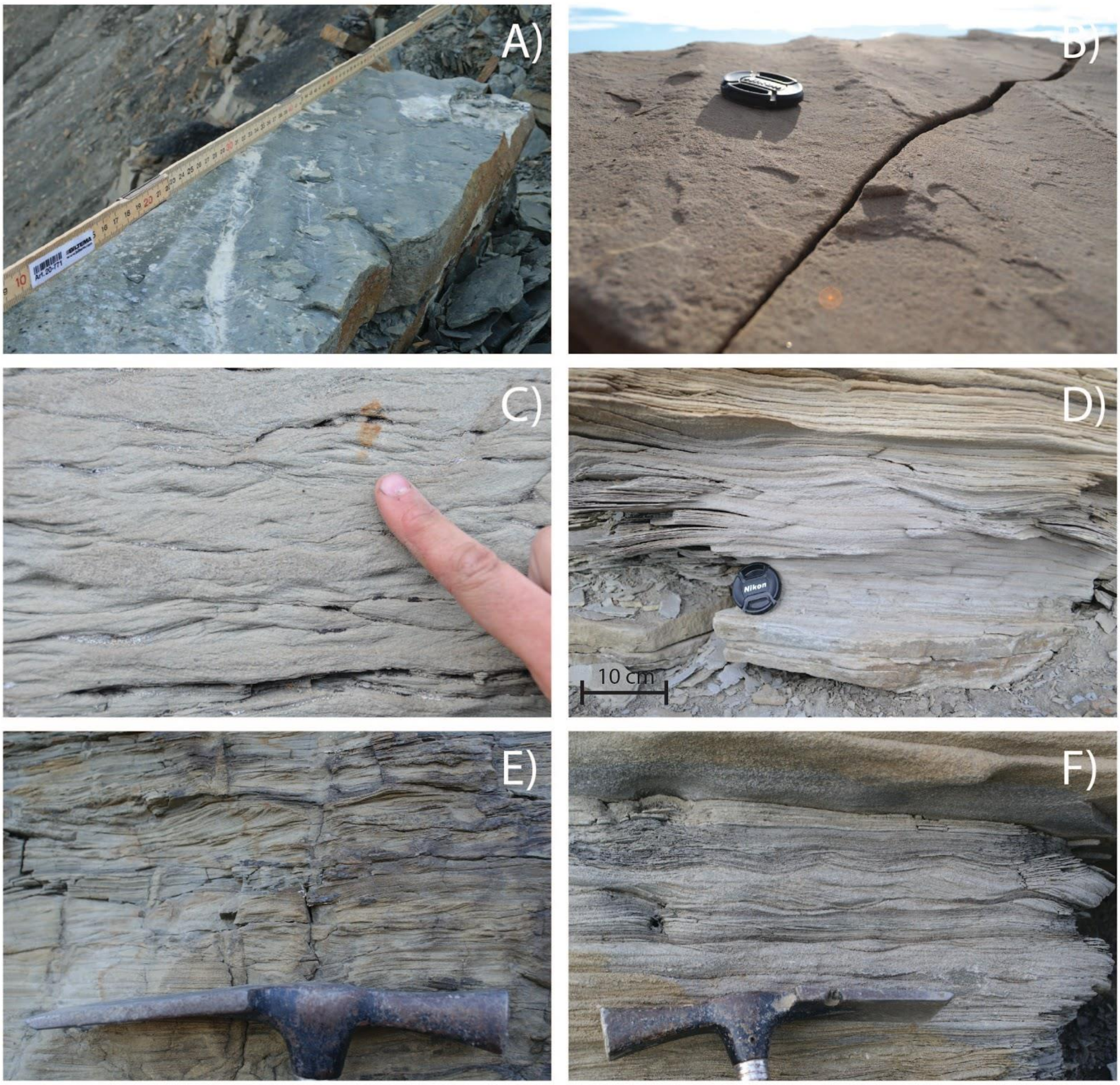


Figure 3.4: Facies D - Wave rippled sandstone A) Straight symmetrical ripples preserved in the upper part of the De Geerdalen Formation on Klement'evfjellet, Agardhbukta B) Peak-shaped ripple crests on a block of sandstone, Hellwaldfjellet, north-eastern Spitsbergen C) Bidirectional features interpreted 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 intervowen 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. 3.5A)(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. 3.5B)(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.

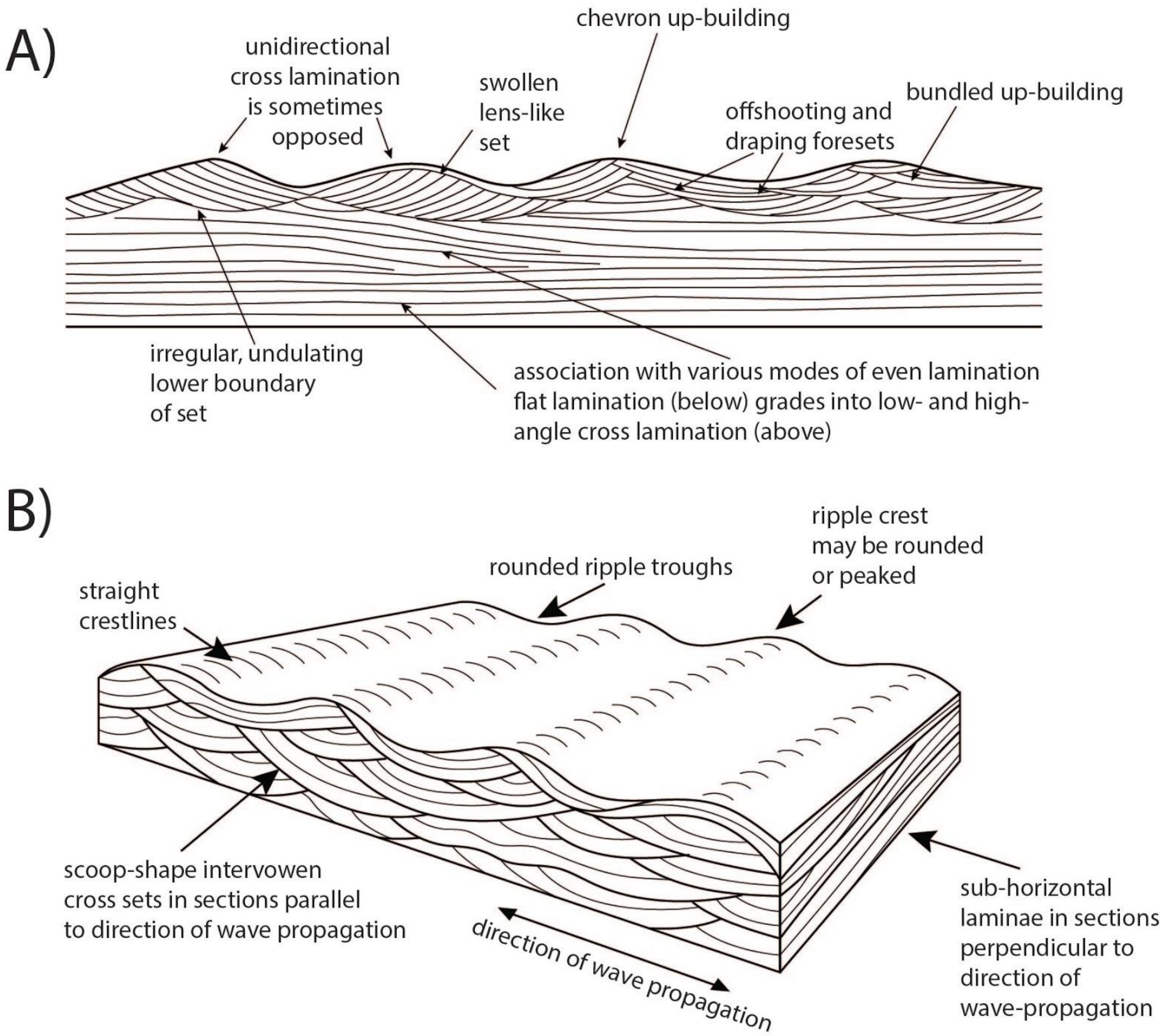


Figure 3.5: Characteristic features of ripples formed by the oscillatory movement generated by ocean waves. A) Foresets can be unidirectional and sometimes opposed, but are most commonly bidirectional (from Collinson et al. (2006) after de Raaf et al., 1977) B) Three-dimensional idealized block of lamination types that result from the bidirectional movements of water caused by waves (from Collinson et al. (2006) after Boersma, 1970)

3.5 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. 3.6). The colour is usually grey to red-brown when weathered and grey on fresh surfaces. Unit thickness is 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. Fish remains were found within low angle cross-stratified sandstone on Klement'evfjellet. This structure is commonly well developed on the Svartnosa locality (Figs. 3.6A, 3.6B, 3.6E), while in other locations it can appear more subtle and harder to recognize in the field (Figs. 3.6C, 3.6D, 3.6F).

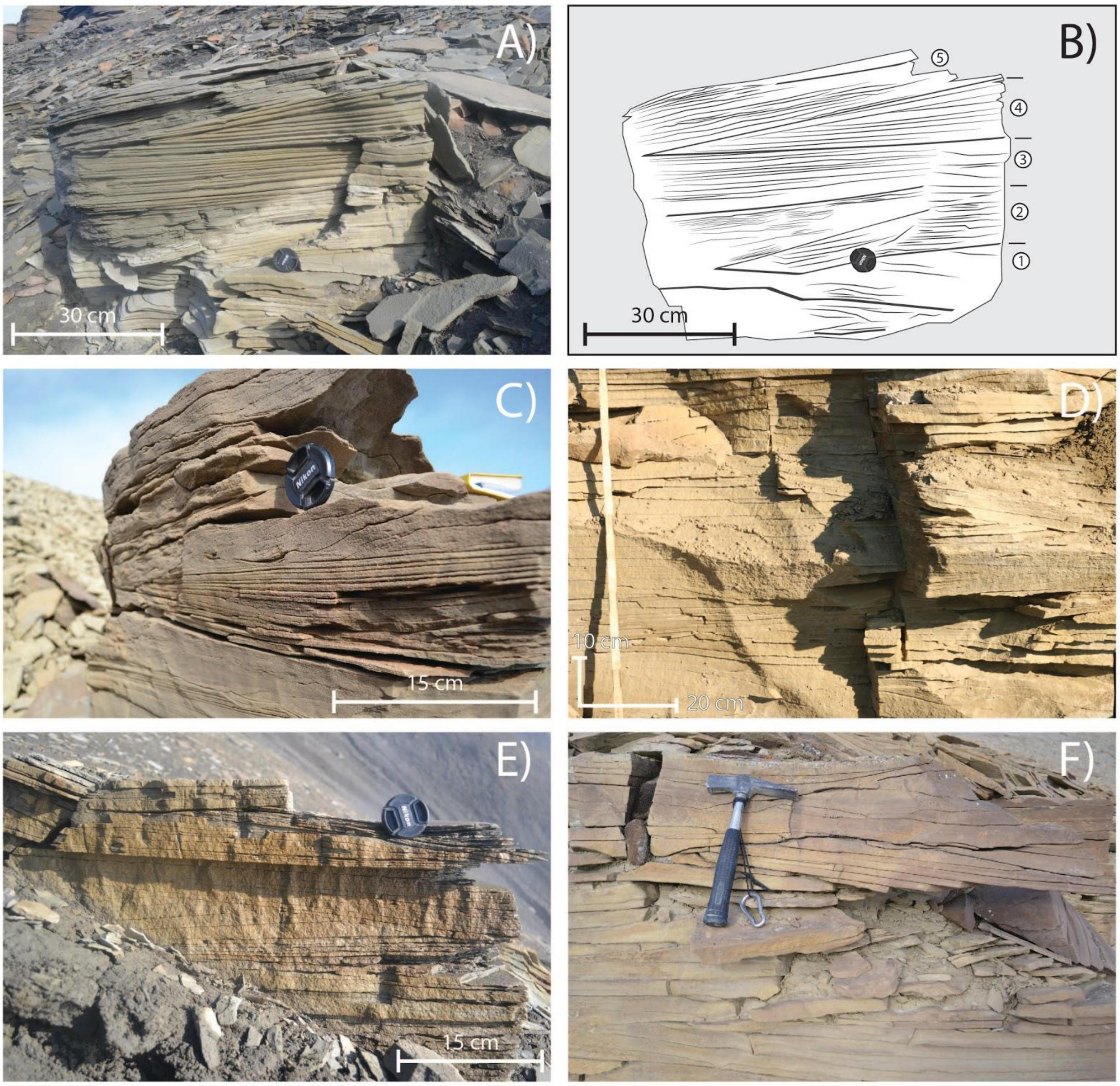


Figure 3.6: Facies E - Low angle cross-stratified sandstone. A) Low angle cross-stratification, Svartnosa, Barentsøya. B) Sketch of Figure A with the different cross-sets labelled. C) Oblique view of low angle cross stratified sandstone from the succession on Hahnfjella, eastern Spitsbergen. D) Low angle 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.

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).

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 are 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.

3.6 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. 3.7). 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. 3.7A) and cm-thick beds (Figs. 3.7B, 3.7C, 3.7D).Transitions between lamination and bedding occur within units (Fig. 3.7D). 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 occurs 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.

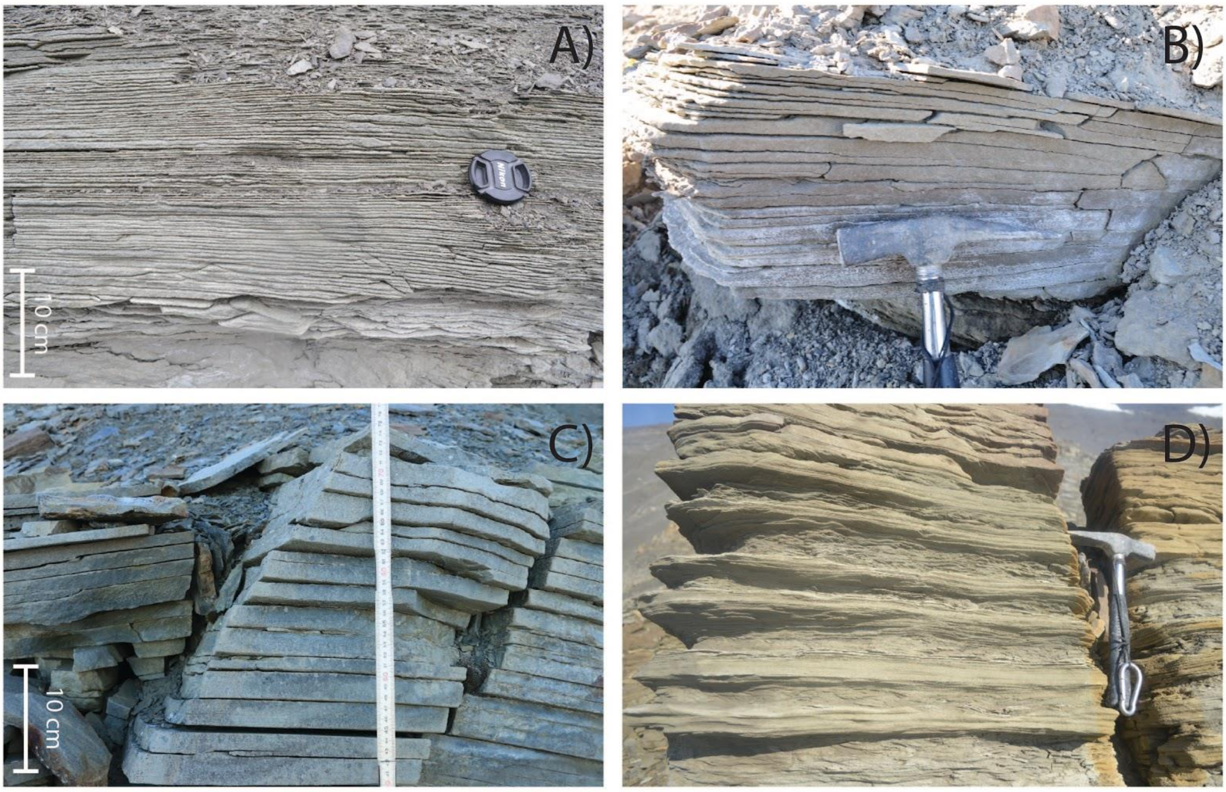


Figure 3.7: Facies F - Horizontally bedded and planar stratified sandstones. A) Decreasing bedding thickness upwards into fine laminae, Mistakodden, Barentsøya. B) Horizontally bedded sandstone on Wilhelmøya. C) Centimeter thick layers of horizontally bedded sandstone Klement'evfjellet, Agardhbukta. 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 occur in various environments, and is thus not considered a 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.

3.7 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. 3.8). 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. 3.8A), 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 3.8: 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 are 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).

3.8 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. 3.9). The concave-up part of the structure is referred to as “swales”, while the convex-up part is referred to as “hummocks” (Boggs, 2009; Boggs, 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 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. 3.9A, 3.9C, 3.9D, 3.9E). It is also common in coarsening upwards sequences on Svartnosa.

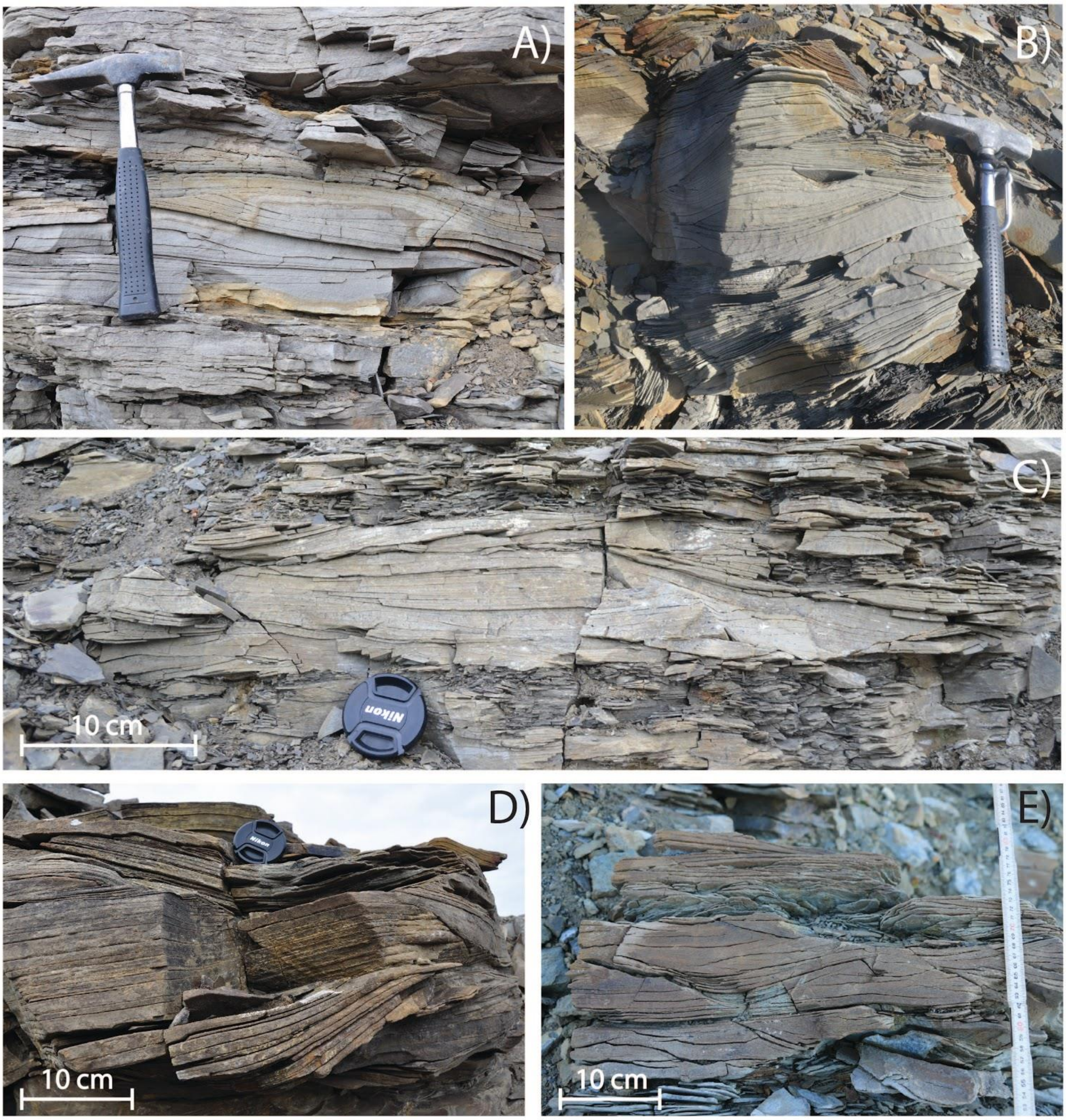


Figure 3.9: Facies H - Hummocky and swaley cross-stratified sandstone A) ”Micro-hummock” is a term applied to smaller scale hummocky cross-stratification. The small scale suggests a more proximal position close to the lower shoreface, Friedrichfjellet, Agardhbukta B) Excellent three-dimensional structure 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.

Interpretation

Hummocky cross-stratification is a sedimentary structure in sandstones that shows 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. 3.10) 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).

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).

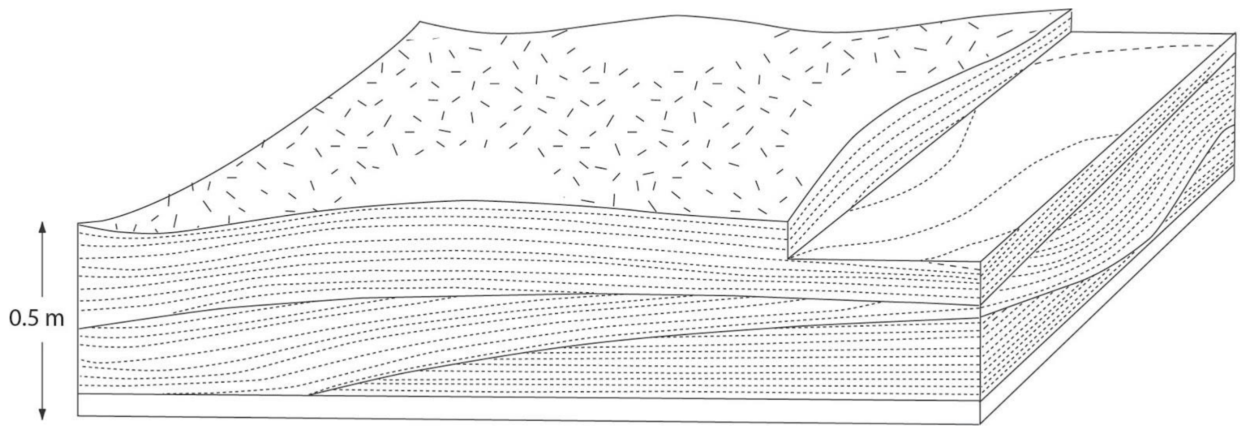


Figure 3.10: Internal geometries and bedform of hummocky cross-stratification (from Quin, 2011, after Harms et al., 1975).

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.

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 favor the preservation of swales (Walker and Plint, 1992; Hampson and Storms, 2003; Dumas and Arnott, 2006).

3.9 Facies I - Soft sediment deformed sandstones

3.9.1 Sub-Facies I_1 - syn-sedimentary deformed sandstones

Description

Very fine to fine sandstones displaying soft sediment deformation structures (Fig. 3.11). 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.

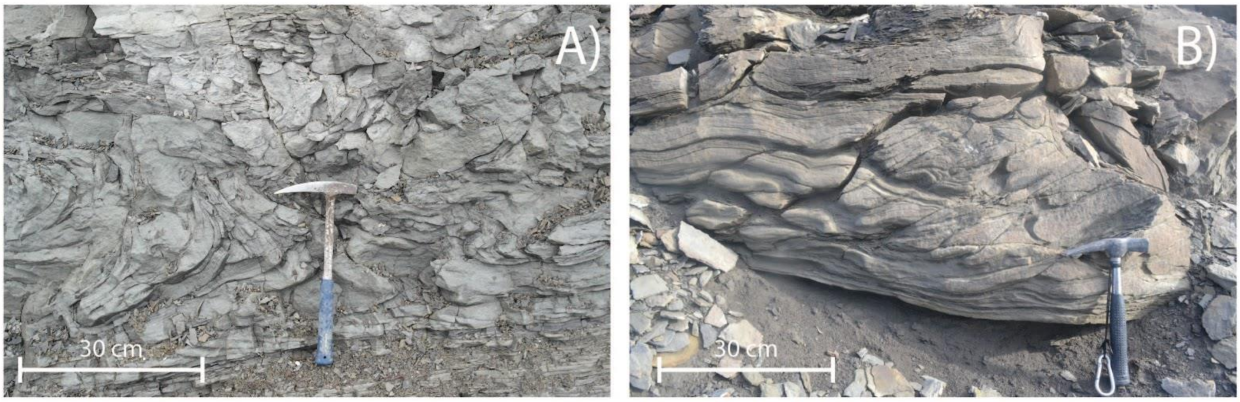


Figure 3.11, previous page**:** Facies I_1 **-** Syn-sedimentary deformed sandstones A) Heavily deformed sandstone on Mistakodden, Barentsøya. B) Characteristic features are the sharp lower contact and the disturbed internal laminations. From Svartnosa, Barentsøya.

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).

3.9.2 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. 3.12). 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 3.12: 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. 3.13) (Reineck and Singh, 1980; Bhattacharya and MacEachern, 2009).

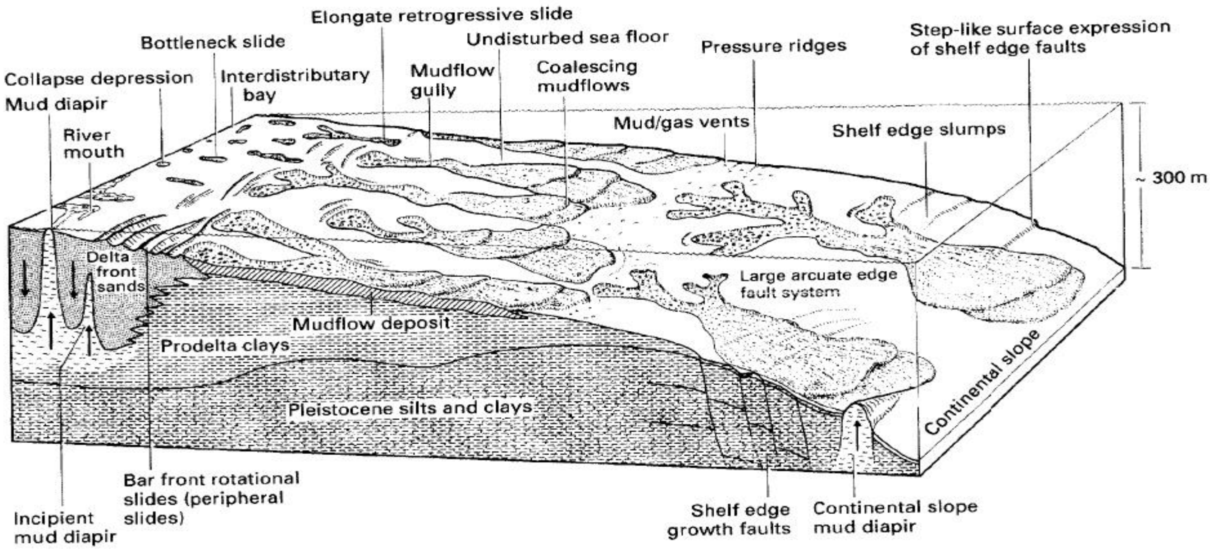


Figure 3.13: Different types of deformation caused by mass-movement of sediment supplied to the delta front and pro-delta (from Reading and Collinson, 1996).

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.

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 have 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). Favorable 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.

3.10 Facies J - Carbonate rich sandstone

Description

This facies comprises very fine to fine sandstones characterized by structures formed during diagenesis (Fig. 3.14). The sandstone units are commonly hard and heavily cemented, making observations of primary sedimentary structures difficult (Fig. 3.14C). Secondary sedimentary structures include cone-in-cone (Fig. 3.14A, 3.14B), siderite beds (Fig. 3.14F) and calcareous concretions.

A decrease in cementation to adjacent facies is noticed as they often appear less consolidated. Colour variations 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. 3.14D).

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. 3.14E, 3.14G). 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 are 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 occur as nodules within mudstones and sandstones grouped in other facies, or as nodules forming distinct layers (Fig. 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 3.14: 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) suggest 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).

3.11 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. 3.15). The thickness of mud- and sand layers ranges from 1 mm to a few centimeters. Units are up to 9 m thick (Fig. 3.15B). 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.

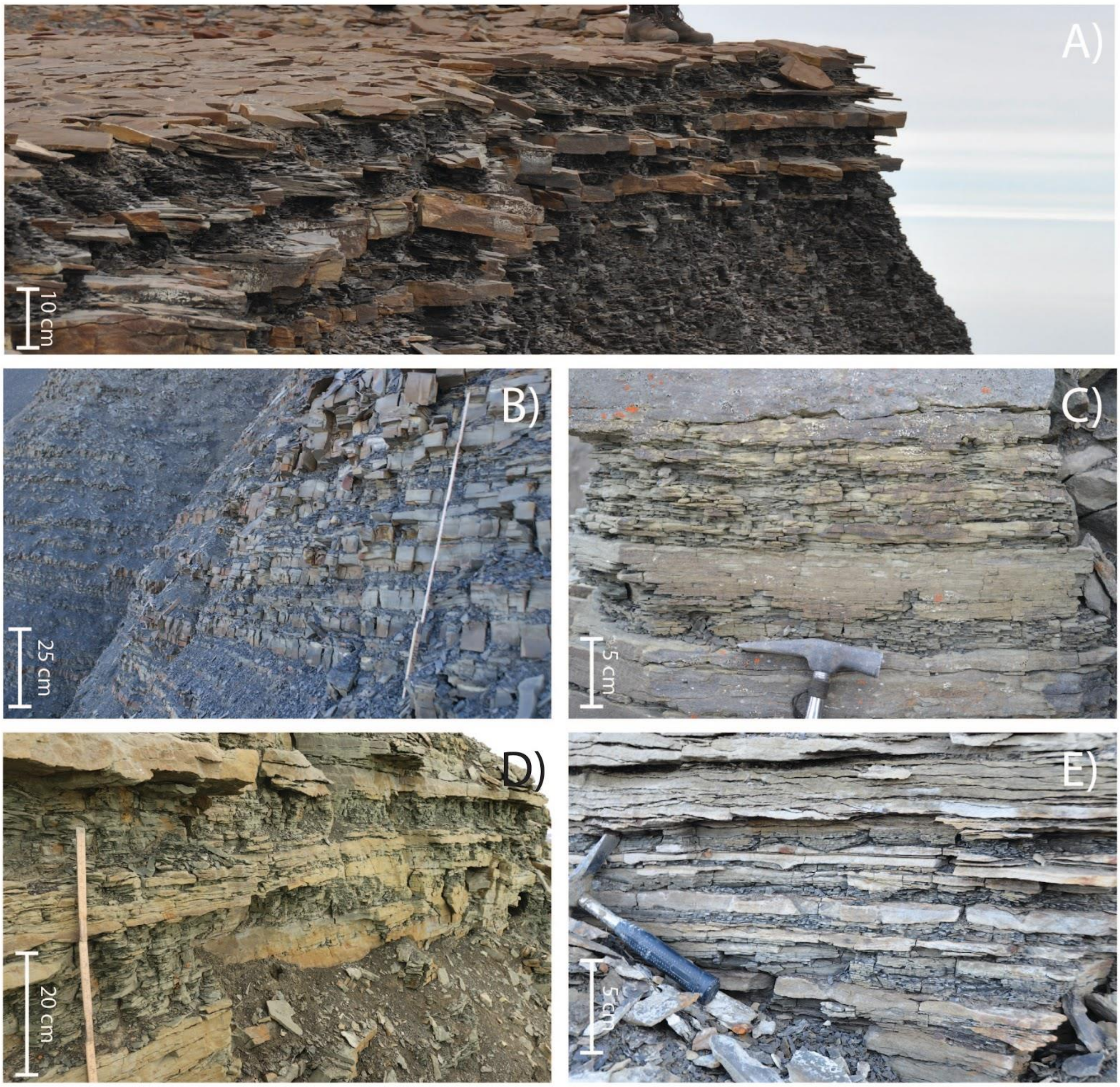


Figure 3.15: Facies K - Heterolithic bedding. Note the different scales of sandstone bed thickness.

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. 3.15, 3.16) (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).

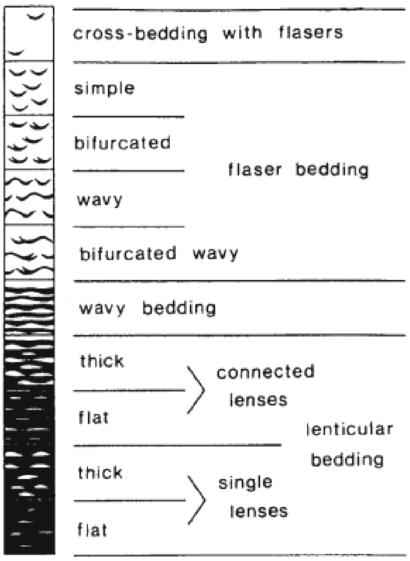


Figure 3.16: Flaser, wavy and lenticular bedding are determined based on the ratio of sand and mud. High sand-mud ratios favors the generation of flaser bedding, while wavy and lenticular bedding are characterized by an equal and low sand - mud ratio (from Reineck and Wunderlich, 1968).

The distinction between flaser, wavy and lenticular bedding is 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 contain 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 favored to be deposited and preserved in the paleo-environment (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. 3.15A, 3.15B). 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 are 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.

3.12 Facies L - Coquina beds

Description

The unit consists mainly of fragmented bivalves, lacking sedimentary structures (Fig. 3.17). 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 shales (facies N) are 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.

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 shales, 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 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.

Figure 3.17: 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.

3.13 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 are dominantly grey or black, but may also be of yellow, white or purple colour (Figs. 3.18).

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.

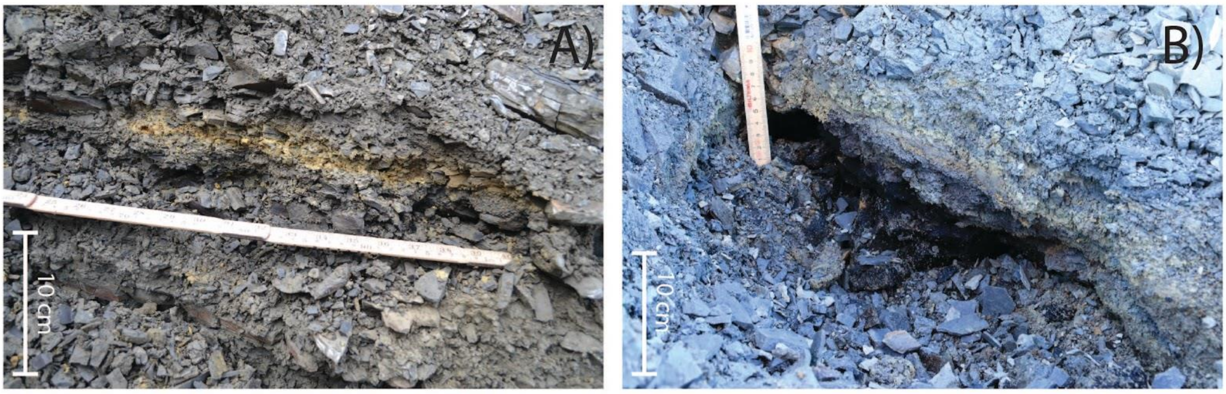


Figure 3.18: Facies M – Mudrocks. A) Grey and yellow mudrocks on Friedrichfjellet, Agardhbukta. B) Klement'evfjellet, Agardhbukta.

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 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 paleosols (Collinson, 1996; Enga, 2015). These are described in greater detail under facies O (Paleosols).

Yellow colouring of laminae and beds of mudrock (Fig. 3.18A) are 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).

3.14 Facies N - Coal and coal shale

Description

Units of coal and coal shale are from 1 to 20 cm thick (Fig. 3.19). The units often appear laterally continuous over tens of meters when examined, but scree cover is common. 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 paleosols (Fig. 3.19B), 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.

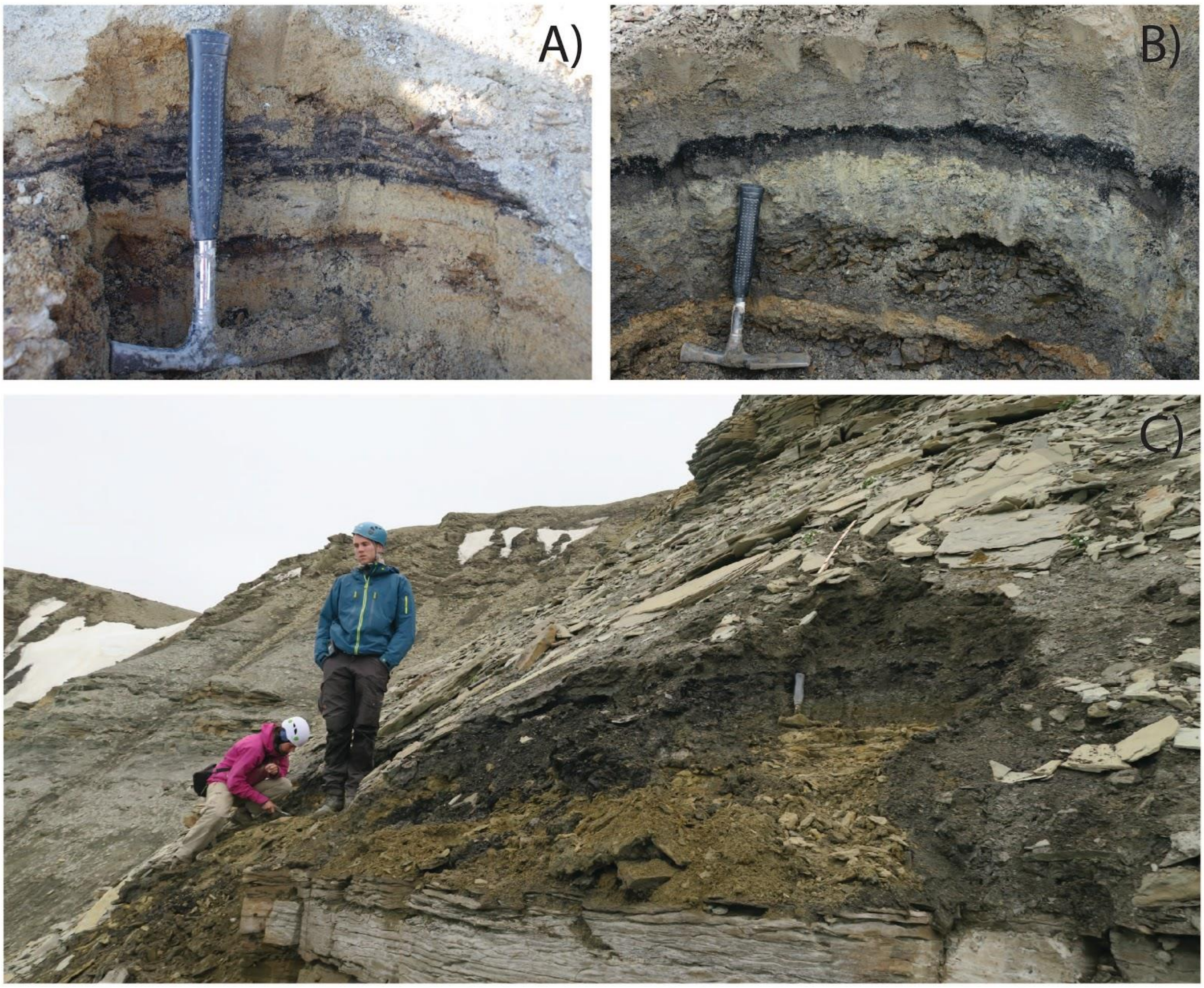


Figure 3.19: Facies N - Coal and coal shale. A) Two laterally discontinuous coal seams alternating with paleosols, Tumlingodden, Wilhelmøya. B) Coal shale overlying bleached paleosol, Šmidtberget, Agardhbukta. C) Laterally continuous coal and coal shale in the uppermost part of sediments interpreted as delta plain deposits, Blanknuten, Edgeøya.

Interpretation

According to the 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 (…)".

To form coal two requirements must be satisfied: i) The clastic sedimentation rate must be low compared to organic matter supply; ii) 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. 3.20). Swamps and mires are the most common coal forming environment and is recognized by vegetation growing in anoxic, waterlogged ground (McCabe, 1984; Retallack, 1991).

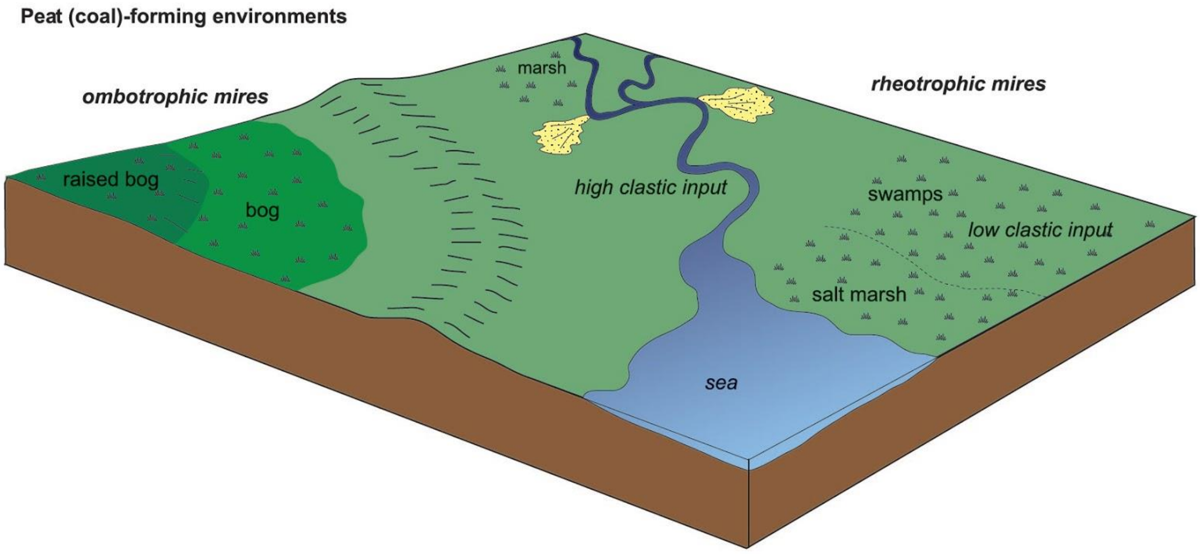


Figure 3.20: Coal-forming environments (Nichols, 2009).

Most of the coal seams found in the De Geerdalen Formation are overlying paleosols, 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 paleosols. This is substantiated by the fact that long geological time with stable conditions that favor 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 seasonal changes in precipitation or fluctuation in sediment input that prevent 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 paleoclimate with seasonal variations in precipitation (Hochuli and Vigran, 2010; Ryseth, 2014; Enga, 2015).

3.15 Facies O - Paleosols

3.15.1 Sub-facies O_1 - Brown and yellow paleosol

Description

Paleosols are found at every locality visited except Muen, but are 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, Wilhelmøya and Agardhbukta. On Wilhelmøya wood fragments up to 20 cm were found within the paleosol. Visible organic matter is found together with the paleosol on Hahnfjella, otherwise visible organic matter is sparse. The colour varies from brown to reddish brown and yellow (Fig. 3.21). At some of the outcrops a bleached yellow 10 to 50 cm thick layer occurs above a red or brown base.

The paleosols 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 paleosols are commonly overlain by coal or coal shale (facies N). Thin coal seams from 1 to 5 cm sandwiched in the paleosol are common.

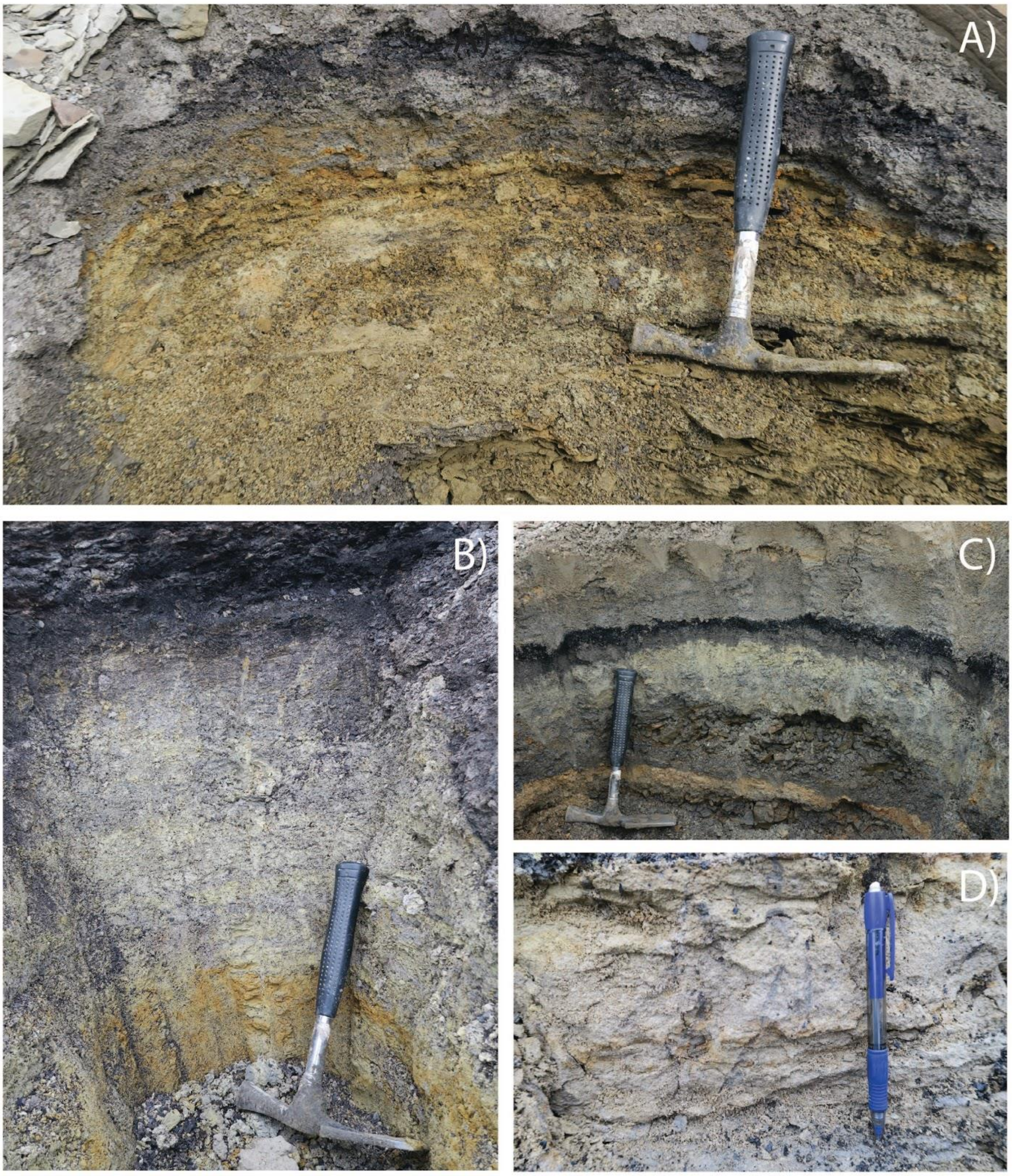


Figure 3.21: Sub-facies O_1 **-** Brown and yellow paleosol. A) Yellow, coarse grained paleosol, typical for proximal channel paleosols. The outcrop is overlying distributary channel deposits of Blanknuten, Edgeøya. B) Possible top of channel paleosol with coal shale (O-horizon) grading into a bleached layer (E horizon) and brown base (B horizon). Field observations showed irregular lines interpreted as cutans. Tumlingodden, Wilhelmøya. C) Coal seam overlying bleached layer containing coalified roots. The roots are penetrating the upper reaches of the underlying dark paleosol horizon, Šmidtberget, Agardhbukta. D) Paleosol with rootlets, Blanknuten, Edgeøya.

Soils can be classified on the background of visible soil properties. A common way to classify soils is to divide into five major horizons according to the USDA soil survey Manual. 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. 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 C horizon at the base is overlying the bedrock and is little influenced by the pedogenesis. Although this is a common classification of soil horizons, most soils show much more complexity than described here, and several methods have been used to classify soils (Boggs, 2011; Retallack, 1988). An example of a paleosol interpreted to have well developed A-, E-, B- horizons can be seen in Figure 3.21B.

Paleosols can be used to reconstruct paleolandscape. The paleocatena model from Kraus and Aslan (1999) shows how the properties of the paleosols changes in a local scale from the levees close to the river channel and down the slope (Fig. 3.22). In this model the soil horizons are termed A, B and C instead of the classification from Boggs (2011).

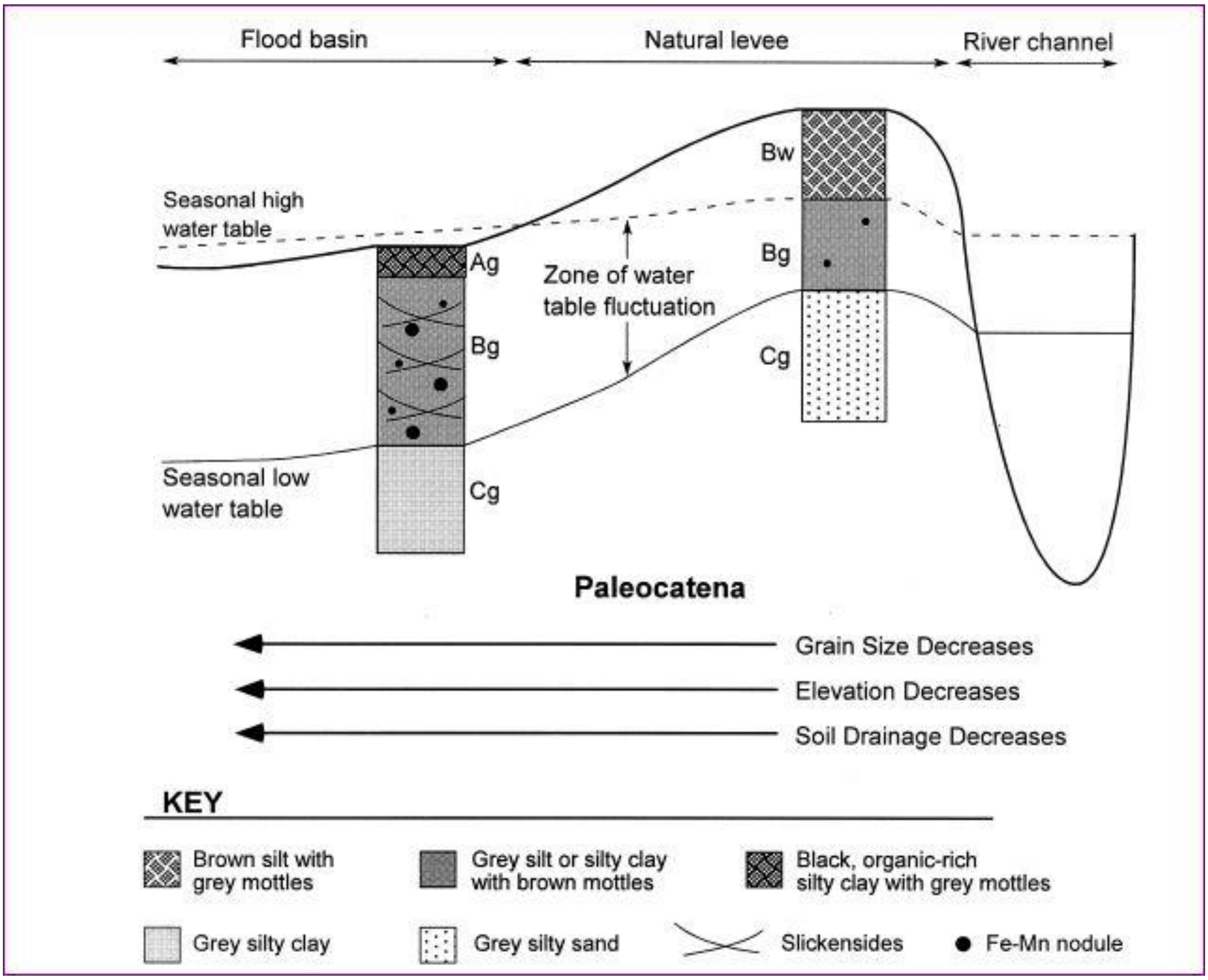


Figure 3.22: The paleo-catena model. Soils formed close to channels tend to be more coarse grained and and formed in oxidized conditions, leading to yellow to brown colour. Soils decrease in grain size away from the active channel and more of the profile is gleyed (Kraus and Aslan, 1999).

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, is usually formed under oxidized conditions, leading to yellowish to brown colour. In Figure 3.22 this is described as Bw (weathered B horizon). Soil further down in the profile is closer to the groundwater table and thus potentially more prone to reduced conditions leading to gray colour. Groundwater saturated soils are called gleyed soils and are marked as Ag, Bg and Cg in Figure 3.22 (Kraus and Aslan, 1999). An example of possible proximal channel paleosol with yellowish coarse grained material is seen on Blanknuten (Fig. 3.21A). The interpretation is supported by the observation of distributary channel deposits of the underlying unit.

Soils tend to be less drained away from the channel because the soils consist of finer material and is located closer to the groundwater table due to the topographic position. This favours 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). The development and maturation of paleosols tends to increase with distance from the channel, because of decrease in sediment accumulation away from the channel (Bown and Kraus, 1987).

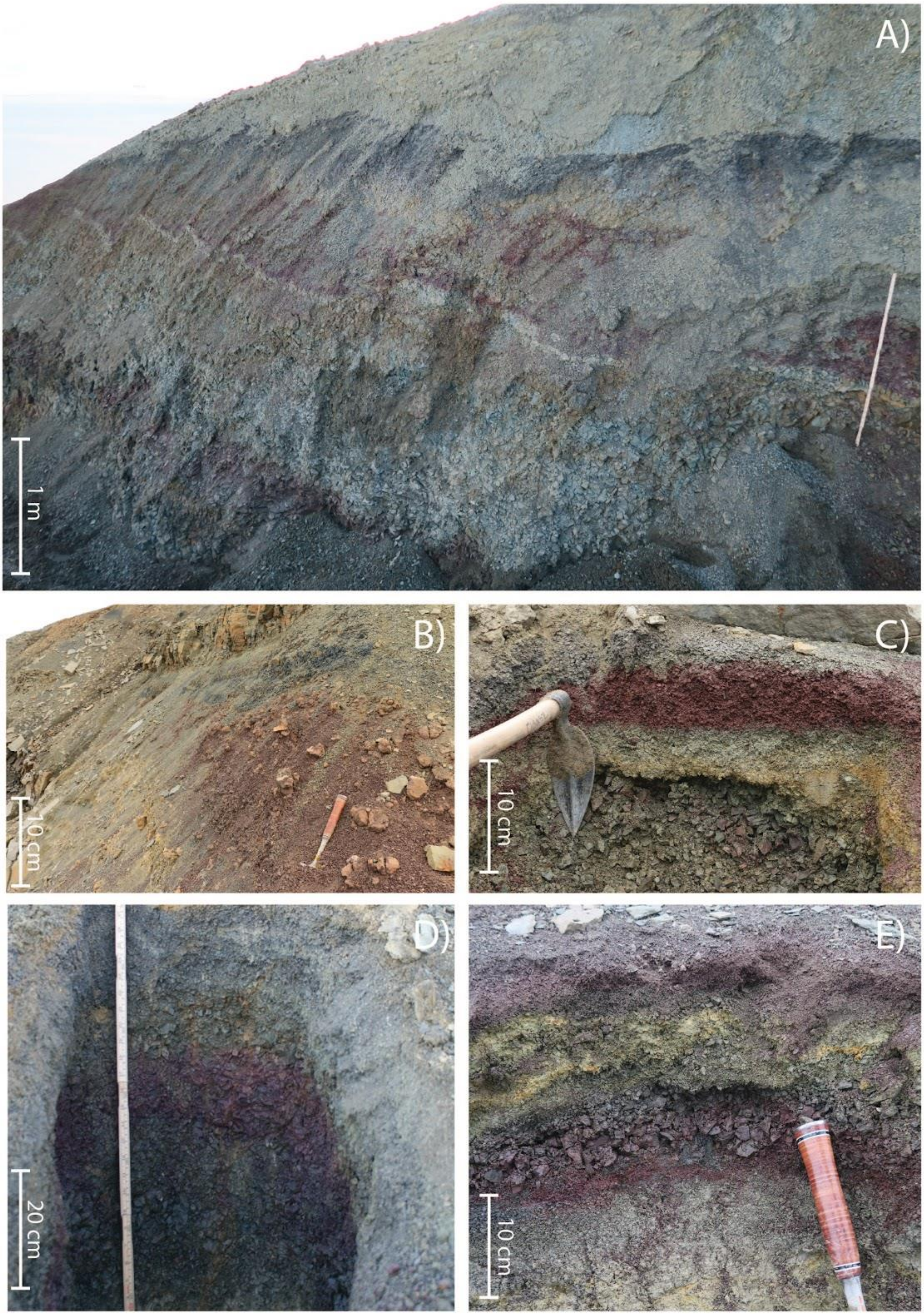
3.15.2 Sub-facies O_2 - Alternating red and green shales

Description

Alternating units of red and green shales (Fig. 23) 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.

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. Red and green shales are often found above sandstone beds.

Figure 3.23, next page**:** Facies O_2 - Alternating red and green shales. A) Overview photo of the Isfjorden Member on the slopes above the dolerite sill on Tumlingodden, Wilhelmøya B) Alternating red and green shales with larger nodules at discrete levels on top of Klement'evfjellet, Agardhbukta C) Close-up view of the uppermost layer of red and green shales, Šmidtberget, Agardhbukta. Note the presence of peds in the red shale. D) Friedrichfjellet, Agardhbukta. E) Nodules possibly originating from roots in red shale, Friedrichfjellet, Agardhbukta.


A)
1 m
B)
10 cm
C)
10 cm
D)
20 cm
E)
10 cm

Interpretation

Paleosol is a weathering product formed due to physical, biological and chemical modification during periods of subaerial exposures. Paleosols are thus not actual deposits, but instead formed by the alteration and maturation of ancient soils. Most paleosols 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). Paleosols normally represent an unconformity, because they form in periods where the landscape is degrading or stable, followed by deposition of sediments (Kraus, 1999). Three main features are important for recognition of paleosols in the field: traces of roots, soil horizons and soil structures (Retallack 1988). Observations of these features were done in the field to distinguish paleosols from adjacent sediments.

Roots are one of the most diagnostic features in paleosols because the presence of roots is evidence of subaerial exposure and plant growth (Retallack, 1988). Roots and traces of roots tend to be best preserved in former waterlogged ground. Traces of roots are recognized by irregular shaped features which are tapering and branching downwards (Figs. 3.23C, 3.23D) (Retallack, 1988). Traces of roots can also be preserved when cracks around roots are filled up with sediments, leading to the development of peds. Peds that have originated this way are usually blocky. Nodules possible originating from roots are seen in Figure 3.23E.

The second main feature for recognition of paleosols is soil horizons. Paleosols typically have sharp or erosive contact at the top and gradual boundary at the base (Retallack, 1988). This is in accordance with field observations for the brown and yellow paleosol, while the alternating red and green beds show sharper contacts both at the top and at the base. The last diagnostic feature of paleosols is the soil structure. Many paleosols appear hackly. This originates from open spaces and weaker zones surrounded by more stable aggregates in the original soil. The stable aggregates are termed peds. When buried the soil is compacted, but the original structure is preserved by a network of irregular planes termed cutans. The cutans are usually filled with sediments or mineralized by crystals (Retallack, 1988). Examples of outcrops interpreted as paleosols with cutans and peds are seen in Figure 3.23B and Figure 3.23C.

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

3.16 Ichnofacies

Ichnology, the study of trace fossils, is 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. 3.24)(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.

3.16.1 *Cruziana* ichnofacies

The *Cruziana* ichnofacies comprises 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.

3.16.2 *Skolithos* ichnofacies

Within the De Geerdalen Formation on eastern Svalbard, trace fossils such as *Skolithos* and *Diplocraterion* were observed within several facies (Figs. 3.24, 3.25A, 3.25D). *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. 25), and in beach environments (Pemberton et al., 1992; Boggs, 2011).

On Svartnosa, abundant *Skolithos* was observed concentrated within a single sandstone bed (Fig. 3.25A). Sandstones with a high abundance of vertical tubes of *Skolithos* is 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. 3.25G, 3.25H). It belongs to the *Skolithos* ichnofacies and are 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. 3.25C). 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).

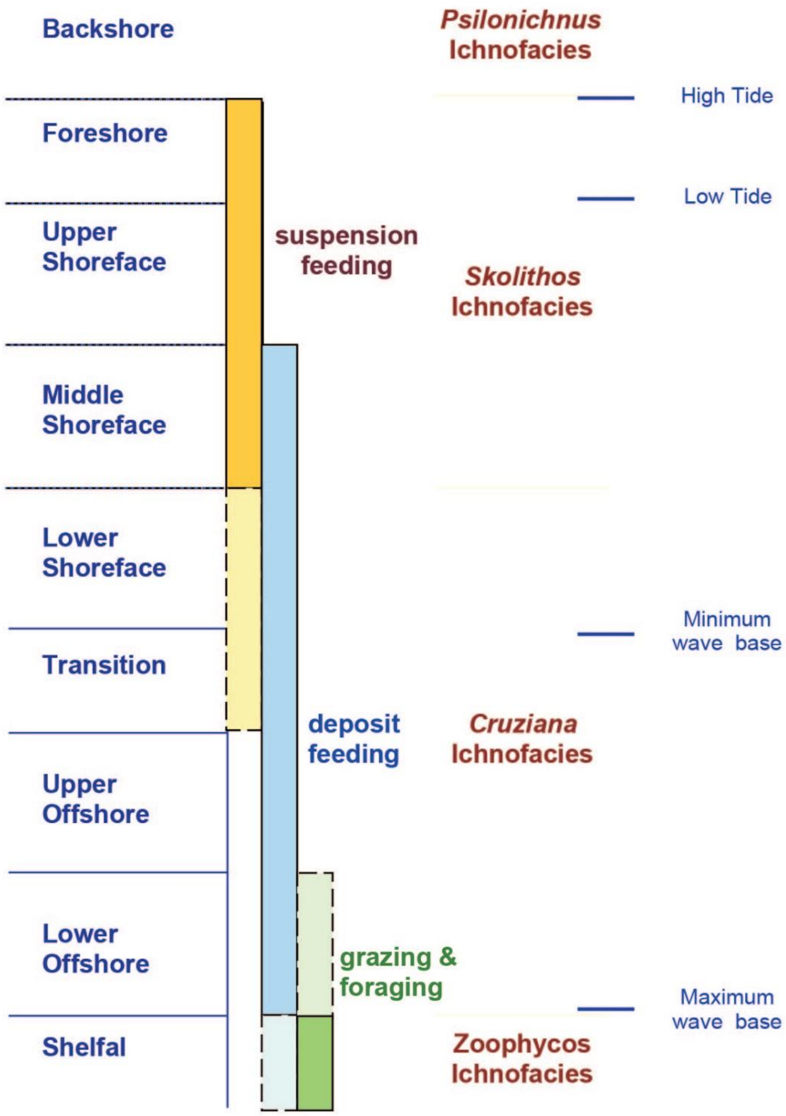


Figure 3.24: Distribution and types of ichnofacies on the shoreface. Sandstones in the De Geerdalen Formation exclusively contain trace fossils found in transitional offshore-shoreface environments, such as the *Cruziana* and *Skolithos* ichnofacies (Pemberton et al., 1992; Clifton, 2006).

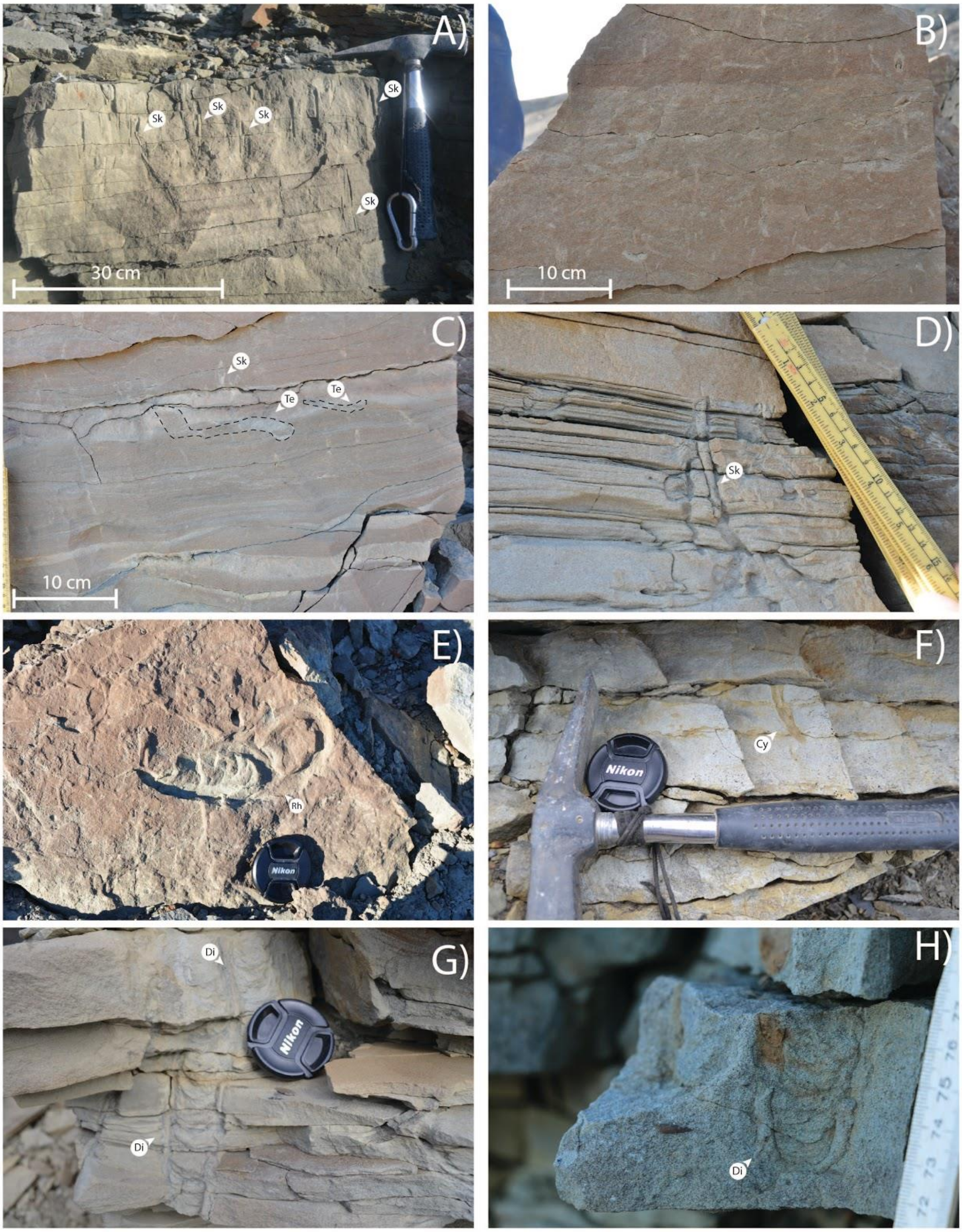


Figure 3.25: Trace fossils observed on eastern Svalbard. A) "*Skolithos* piperock" composed of vertical tubes of *Skolithos* (Sk), Svartnosa (Barentsøya). B) Bioturbated sandstone, Wilhelmøya. C) *Teichichnus* (Te) and *Skolithos* (Sk) in the upper part of De Geerdalen Formation on Wilhelmøya below the 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.

4 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 4.1) and identified based on interpretation of facies, geometries and dimensions of sandstones and other observations. The occurring facies associations are considered sub-environments in the shallow marine offshore to lower shoreface (FA 1), delta front (FA 2) and delta plain (FA 3) environments. Deltas commonly include multiple sub-environments, as illustrated by the conceptual model in Figure 4.1.

Table 4.1: Facies assocations in the De Geerdalen Formation on eastern Svalbard (modified from Rød et al., 2014). Paranthesis indicates rarely observed facies.

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 Mouth bar	A, B, C, D, E, F, I	Fluvially 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	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 (e.g. lakes and lagoons) with deposition of fine-grained material, coal, coal shales and paleosols.

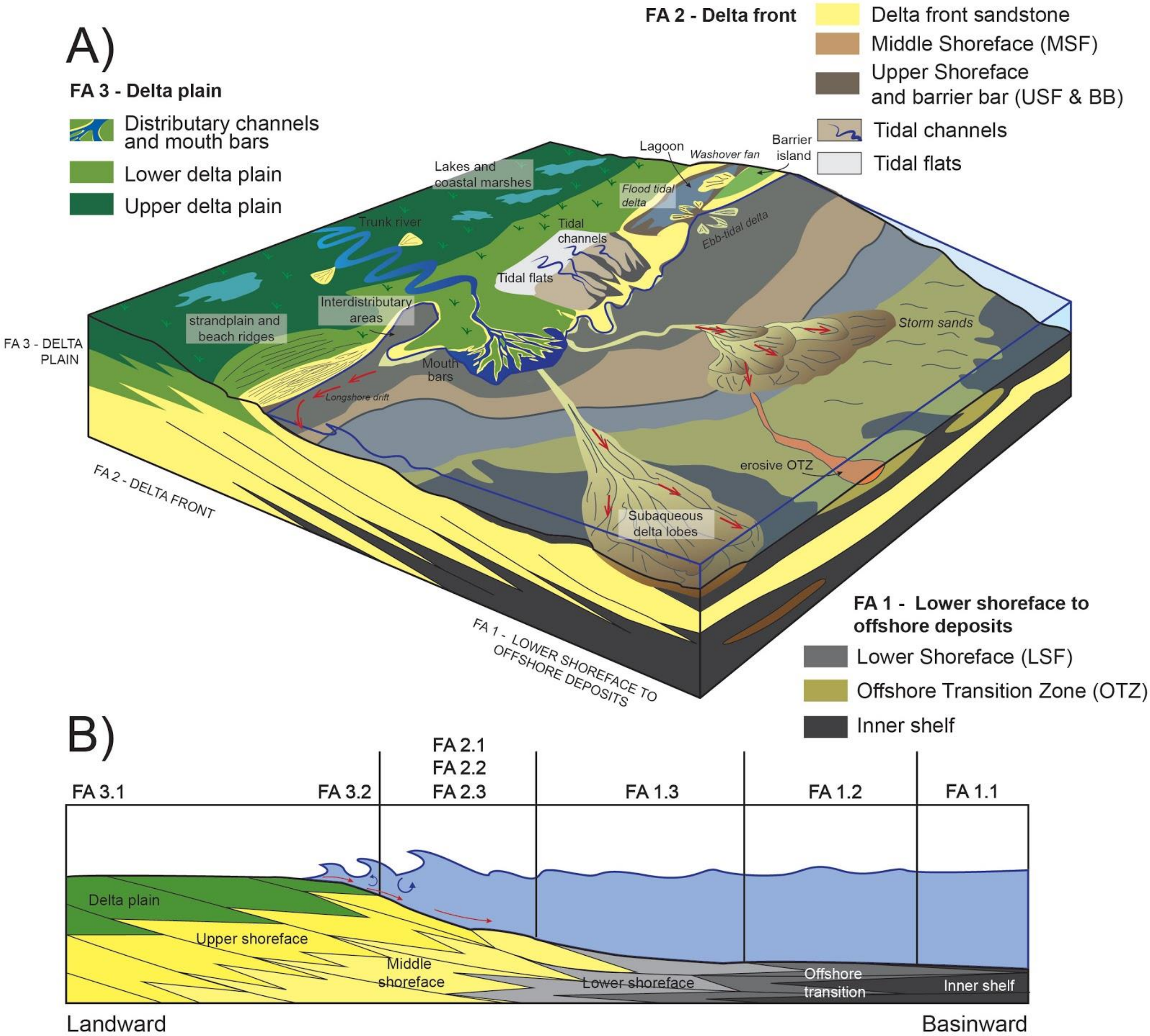


Figure 4.1: Facies associations within the De Geerdalen Formation. A) Conceptual block diagram showing the various environments and sub-environments occurring in the De Geerdalen Formation. Fluvial, wave and tidal processes are important factors in governing coastline morphology. Tidal processes generate 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 result 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.

4.1 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 decrease 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.

4.1.1 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. 4.2), and make 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, are 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 is 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 term pro-delta is different from the offshore zone, in the way that the pro-delta comprises 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 southeastern 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 4.2: 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.

4.1.2 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. 4.3). 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.

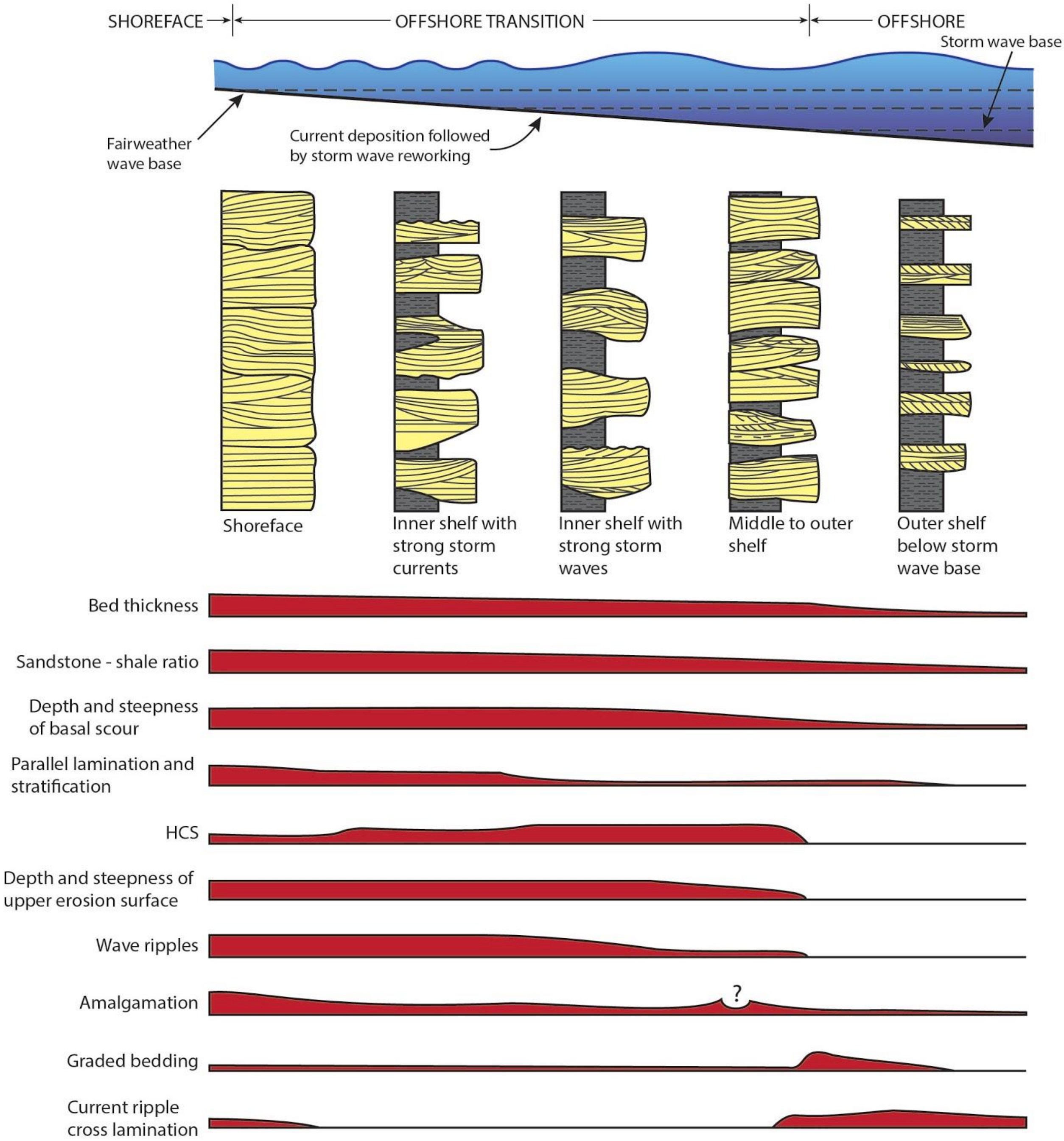


Figure 4.3: Shifts in facies occur in storm deposits as water depth and distance from the shoreface increases (Brenchley, 1985, Miall, 2000).

4.1.3 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.

4.2 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 constitute 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 mark 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.

4.2.1 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.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. 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. 4.4). 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 produces 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 result in a variety of clastic coasts where beach-ridge strandplains, chenier plains and mudflats in general have been attributed to regressive systems, while barrier island-lagoonal systems and estuaries commonly are associated with transgressive systems (Boyd et al., 1992; Dalrymple et al., 1992; Reading and Collinson, 1996). Distinguishing these sub-environments are 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).

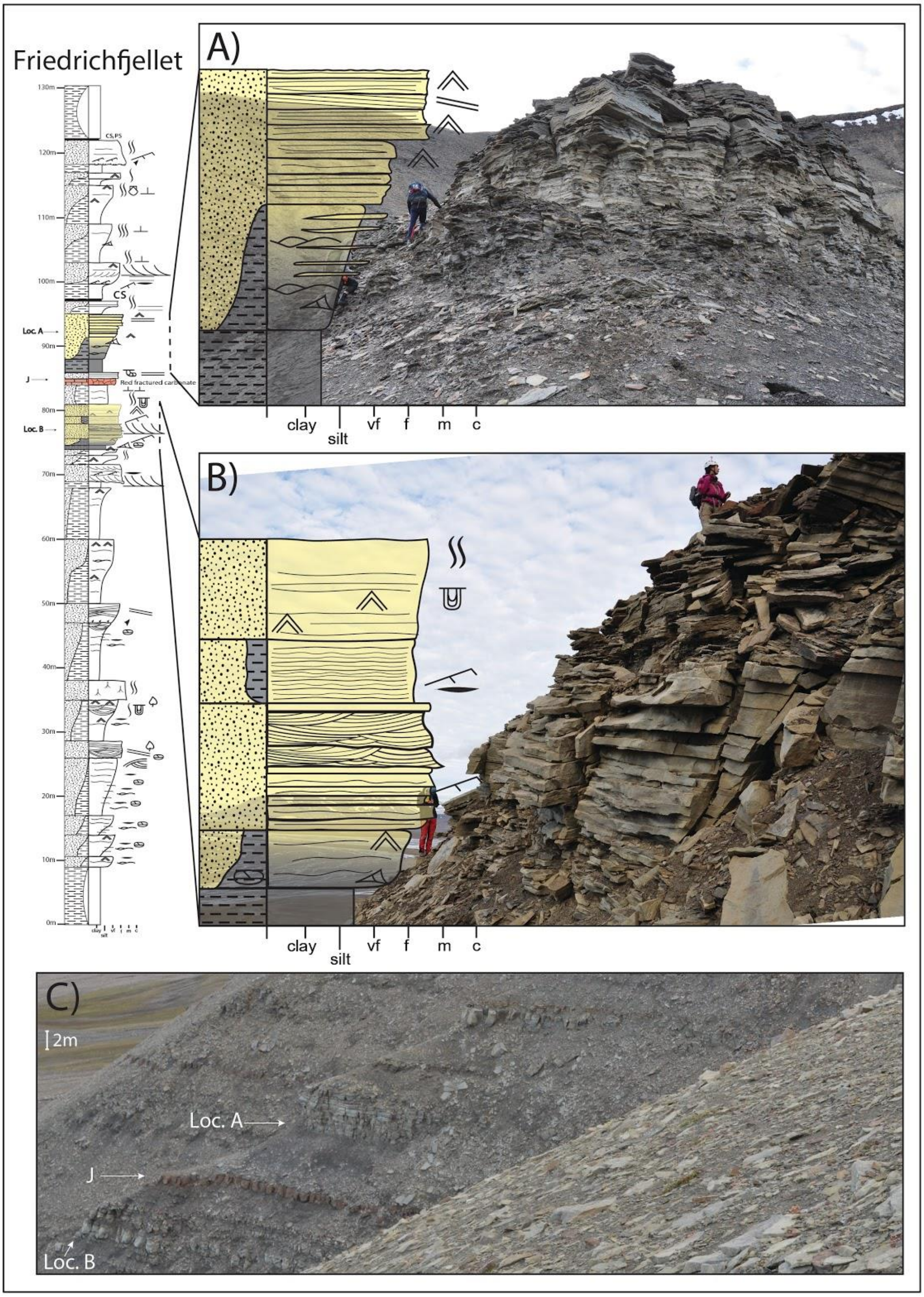


Figure 4.4, next page**:** Lower sections logged on Friedrichfjellet. A) Coarsening and shallowing upward sequence with facies indicating prevalent oscillating flow, interpreted as a prograding barrier succession. B) Interpreted barrier island sequence with large-scale cross stratified interval possibly remnant of tidal inlet or longshore bar migration. C) Overview picture. Note lateral continuity of facies J.

4.2.2 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, hundreds of meters, as observed in the field. However, field data regarding geometries from outcrops are sparse. The facies association is observed in the middle parts of the Svartnosa section and at 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 rates are 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 analogs (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 depend 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 erodes the upper parts of mouth-bar sediments (Reading and Collinson, 1996).

4.2.3 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 paleosols are found at the very top of sequences. Abundant plant fragments are 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; i) laterally restricted channel sandstones displaying relatively modest dimensions (height around 2 to 3 meters, width around 10 m), ii) 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. 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, and are here described under delta front facies association (FA 2). A distributary channel is a stream carrying sediment 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 paleosol on the top indicate 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).

The outcrops show 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 supports 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).

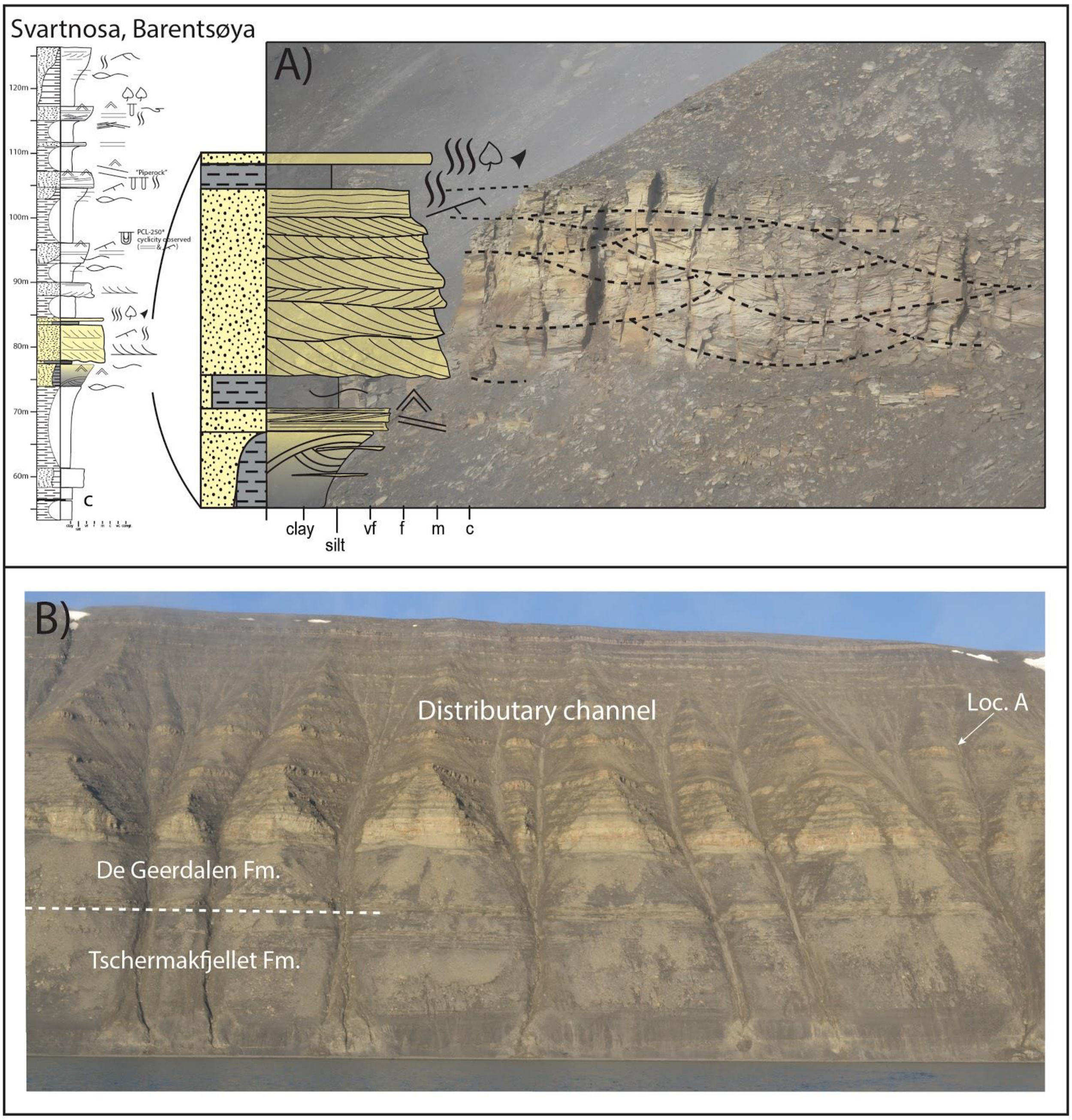


Figure 4.5: Svartnosa, Barentsøya. A) Amalgamated distributary channel deposits superimposed on lower shoreface deposits of FA 1. B) Overview photograph taken at sea level. Note the lateral continuity of sandstone bodies, and thick delta front deposits below the distributary channel.

4.3 Facies association 3 (FA 3) - Delta plain

Delta plain deposits is commonly overlying delta front sediments (FA 2), 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 is normally defined by the presence of distributary channels and the limit to the lower delta plain is taken 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 kilometers 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).

4.3.1 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 are found on Wilhelmøya (facies C). Paleosols 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 FA 2 (delta front) are 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 levels or high precipitation 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) are interpreted as rapid deposition of sediments and could possibly be formed in crevasse splays. Paleosols, 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 channels 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 shifts into interdistributary areas.

4.3.2 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 paleosols (facies O). 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).

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

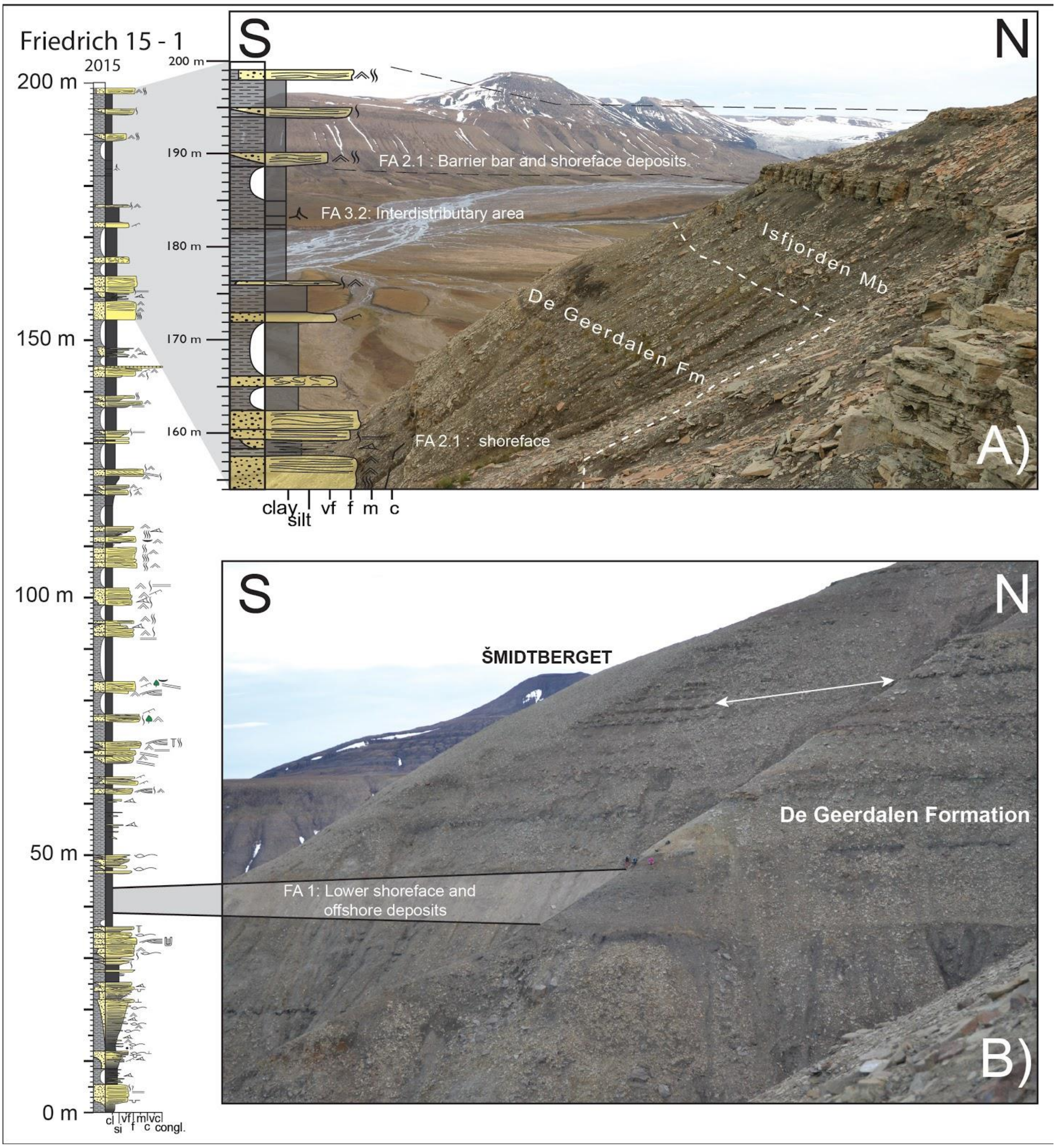


Figure 4.6: Mud-dominated facies associations on Friedrichfjellet, Agardhbukta A) Fine grained interdistributary sediments (FA 3.2). The red and green shales of the Isfjorden Member are conformably overlying the De Geerdalen Formation B) Lower shoreface to offshore deposits (FA 1).

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 favor woody vegetation and is mainly located on the upper delta plain. Swamps gradually turn 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 contain 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 disturb 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.

Paleosols indicate subaerial exposure (Boggs, 2011), and is formed only if the sedimentation rate does not exceed the rate of pedogenesis (Kraus, 1999). Paleosols are thus an indicator of little or no sedimentation, and may be one of the clearest indicators of an interdistributary regime.

Paleosols on top of distributary channels can indicate migration, diversion or abandonment of channels. Fine grained material underlying paleosols 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 paleosols, 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.

4.4 Storfjorden correlation

Defined facies (Chapter 3) and facies associations (previous sections in this chapter) has provided a framework for log interpretations (Appendix A). Furthermore, two correlation panels have been created, and correlates the western and eastern localities respectively. The panels are displayed and discussed in Chapter 6.2. The following section aims to briefly describe how the sections were correlated and discuss sources of error with correlations in the Storfjorden area.

The top Botneheia Formation and Slottet Bed are considered relatively flat surfaces and provides a datum for correlation below the Tschermakfjellet Formation and atop the De Geerdalen Formation respectively. The Slottet Bed represents an isochronous surface. The lower boundaries of the Tschermakfjellet and De Geerdalen formations are distinctly diachronous (Mørk et al., 1999; Høy and Lundschien, 2011; Lundschien et al., 2014), and considerable thickness variations of the formations (Lock et al., 1978; Mørk et al., 1999; Vigran et al., 2014) may have complicated using the base of the De Geerdalen Formation as a marker bed. Intrusive sills of the Diabasodden Suite (Mørk et al., 1999) in the upper parts of the De Geerdalen Formation are seen to elevate the Slottet Bed at Hellwaldfjellet and Wilhelmøya.

The study area is crosscut by the Lomfjorden Fault Zone on Teistberget and the “Rindedalen Structure” on Barentsøya, both roughly north-south trending lineaments of probable Cenozoic age (Lock et al., 1978). Given the lateral extent between sections and presence of structural lineaments, tectonics may have complicated correlation of sections.

On the eastern panel the Muen, Krefftberget and Mistakodden sections were correlated using the top of the Botneheia Formation as a correlative surface, and the Svartnosa section was incorporated by base De Geerdalen Formation. Excellent exposures at Svartnosa allowed for placing the formation boundary at 36 m above sea level at the measured section. Furthermore, the correlation was extended to Hellwaldfjellet and Wilhelmøya sections by an inferred base De Geerdalen Formation occurring at 156 m and 142 m above sea level, respectively, at the base of first occurring prominent sandstone. Due to abundant scree cover on the latter two localities sandstones may have been overlooked, resulting in an inaccurate placing of the base De Geerdalen Formation. Smith (1975) reported the presence of Tschermakfjellet Formation close sea level on Hellwaldfjellet. In turn the Hellwaldfjellet and Wilhelmøya sections were correlated using the Slottet Bed as datum.

On the western panel the Wilhelmøya, Hellwaldfjellet, Teistberget, Klemt’evfjellet, Smidtberget and Friedrichfjellet sections were correlated using the Slottet Bed as datum. The panels are roughly north-south oriented and may tentatively represent sections along depositional strike of the north-westerly progradading clinoform sequences mapped further to the south-east (Lundschien et al., 2014).

5 Petrographic observations

5.1 Detrital framework

The studied sandstones of the De Geerdalen Formation appear mineralogically immature, rich in lithic fragments, feldspar and monogranular quartz. The observed abundance of lithic fragments is characteristic, as well as a nearly absent matrix. Accordingly the sandstones are classified as lithic arenites, consistent with previous studies (Chapter 1).

Detrital quartz, feldspar, mica and various lithic fragments constitute the framework minerals of the sampled sandstones. Detrital monocrystalline quartz typically forms a major fraction of the detrital framework. Grains appear fairly clean, lacking major inclusions and alterations. Both undulose and uniform extinction is observed and complicates optical discrimination in regards to feldspar. Point-, long- and concavo-convex contacts to adjacent grains are observed, where the latter two typically forms due to cementation.

Feldspar occur as subangular to angular clasts displaying frequent alterations. Albite and grid twinning is occasionally observed and used to identify plagioclase and microcline respectively. Inclusions and partial dissolution preferentially occur along cleavage and twinning planes and, when present, comprise distinguishable features in regard to quartz. Grains are susceptible to alterations displaying partly dissolved and fully dissolved grains.

Both muscovite and biotite are recognized in the samples, although in low abundance. A ductile rheology is typically displayed as micas are bent around detrital grains. Grain rims are seen to be replaced by clay minerals providing expanded textures. In samples containing detrital clays micas tend to align and concentrate along bedding planes, capsuled by adjacent clays.

Lithic fragments constitute a major part of the detrital framework throughout the study area. These include rock fragments consisting of polycrystalline quartz, mica schist fragments, volcanic fragments and chert (Fig. 5.1). Polycrystalline quartz are characterized by sutured and long crystal boundaries, resembling ductile deformation fabrics found in metamorphic rocks (Fig. 5.1A). Outer rims are often irregular and susceptible to diagenetic alterations. Mica schist lithic fragments are comprised of polycrystalline mica displaying distinct foliation. Fragments are typically rounded to sub-rounded and shows low sphericity (Fig. 5.1B). A labile nature is illustrated as mica schist fragments are commonly associated with alteration and dissolution processes.

Volcanic fragments are identified by unoriented plagioclase laths in a dark matrix, where alterations provide a mottled texture (Fig. 5.1C). Chert occur as rounded clasts of quartz-aggregates and show irregular crenulate boundaries between sub-grains (Fig. 5.1D). Rock fragments show a labile nature evidenced by frequent dissolution and replacement by clay minerals. And accordingly these fragments are often associated with intragranular porosity and moldic porosity when enclosed by calcite cement.

Observed heavy minerals include zircon, apatite, tourmaline and chromium spinel.

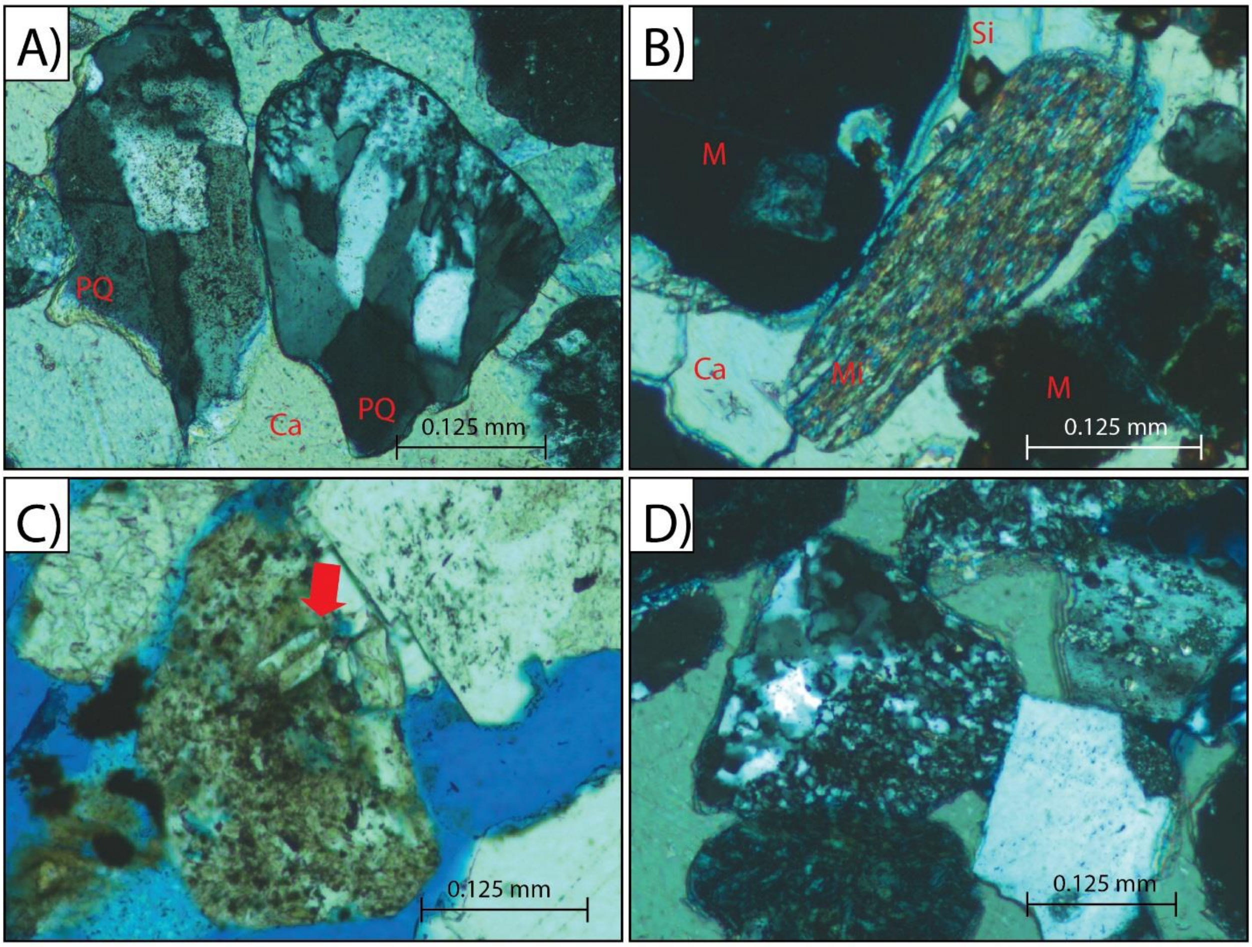


Figure 5.1: Lithic rock fragments. A) Rock fragment of polycrystalline quartz (PQ) enclosed by calcite cement (Ca). B) Mica schist lithic fragment, moldic porosity (M) and siderite (Si) enclosed by calcite cement. C) Altered volcanic fragment showing mottled texture. Red arrow pointing at plagioclase lath. D) Chert fragment. A, B, D in cross polarized light. C in plane polarized light.

5.2 Textural properties of sandstones

Sands in the De Geerdalen Formation varies from very fine to medium grained. The coarsest sand is commonly encountered in distributary channel deposits (FA 2.3). A slight decrease in mean grain size is noted from the Barentsøya localities and Teistberget, compared to the remaining localities.

Observed sorting in the sandstones of the De Geerdalen Formation varies from poor to well, with moderate sorting appearing most common (Appendix B). Samples from distributary mouth bar deposits (FA 2.2) show moderate sorting, whereas distributary channels (FA 2.3) samples varies from poor to well, with moderate sorting appearing most common. Crevasse channels (FA 3.1) were observed to be moderate to well sorted.

Rounding of clasts is seen to vary in the studied sandstones. The main controlling parameter upon roundness seem to be clast type, as no obvious roundness variations were observed in relation to different facies or facies associations. Lithic fragments, in particular volcanic fragments and mica schist, tend to be rounded to sub-rounded whereas feldspar commonly occur as sub-angular to angular clasts. Quartz, polycrystalline quartz and chert fragments may occur as rounded to angular clasts. In consequence a seemingly bimodal roundness distribution is observed, although the constituents of the rounded and angular fractions may vary.

5.3 Authigenic minerals

Authigenic minerals include calcite, siderite, quartz, kaolinite, illite, chlorite and feldspar. Distribution of authigenic minerals and porosity is controlled by diagenesis. Visable porosities were observed to range from 2 to 25 % (Appendix B). The following section will provide a description and discussion of the individual authigenic minerals, while porosity reducing and enhancing mechanisms will be discussed in Chapter 6.

Calcite

In the field presence of calcite was observed to be a common diagenetic feature in most facies and facies associations throughout the study area e.g; upper part of parasequences (FA 1.3 and 2.1), floodplain and crevasse splay (FA 3.1), distributary mouth bars (FA 2.2) and distributary channels (FA 2.3). Additionally the overriding appearance of carbonate cementation at discrete horizontal levels led to the classification of "carbonate cemented sandstone" of facies J (Chapter 3). Calcite cementation is observed to be more common in distributary channels (FA 2.3) than in distributary mouth bar (FA 2.2) sandstones.

In thin-section calcite (CaCO3) occurs as porefilling cement within the sandstones. Cementation varies from patchy to pervasive and show blocky to poikilitic textures. Where the poikilitic cement is most pervasive it often totally obliterates primary intergranular porosity and detrital grains appear to be floating or have point-contacts to adjacent grains (Figs. 5.2A, 5.2B). In these sandstones detrital quartz grains are sub-angular to sub-rounded and no quartz overgrowths were detected. These features are observed in fine- to medium-grained and medium sorted distributary channel sandstones (FA 2.3) and distributary mouth bar sediments (FA 2.2). In sandstones with slightly tighter-packed fabric poikilitic calcite engulfs, and hence post-date, quartz overgrowths.

Calcite cement is seen to replace detrital feldspar (Fig. 5.2C), and enclose feldspar pseudomorphs. Calcite cement has also been detected within cracks of detrital feldspar and enclosing oblong uncompacted micas, pointing to multiple phases of formation prior to and after mechanical compaction. A patchy distribution of poikilitic calcite cement (Fig. 5.2D) is occasionally seen in the studied thin-sections.

In very fine to fine sandstones calcite cement occur both as scattered patches in optical continuity and as replacive porefilling cement with a brownish appearance due to inclusions. In the latter, domains of blocky crystals are seen within the poikilitc areas and exhibit a drusy fabric (Fig. 5.2E). The poikilitc cement show corroded contacts to quartz and feldspar, and the domains with blocky calcite crystals are seen to partly infill voids (Fig. 5.2F).

Optical microscopy and SEM investigations of samples from facies J show abundant microcrystalline calcite. Very-fine grained siliciclastic minerals (e.g. quartz, feldspar, lithic rock fragments) are enclosed by micrite. At Klement'evfjellet an undulose laminae is noticed (Fig. 5.3A), while a massive appearance is noted at other localities. In thin-section microscopy the undulose laminae seems to result from alternating layering between micritic calcite and detrital siliciclastic clasts dispersed in micrititic calcite (Figs. 5.3B). The interbedded clastics are enclosed by micrite, and thus not clast supported. Organic material, seen as brownish inclusions, form undulose laminaes within the micrite dominated domains (Fig. 5.3C). Oblong grains of mica are straight, and does not exhibit evidence of ductile deformation (Fig. 3D). Blocky calcite crystals are seen on the rim of voids (Fig. 5.3E).

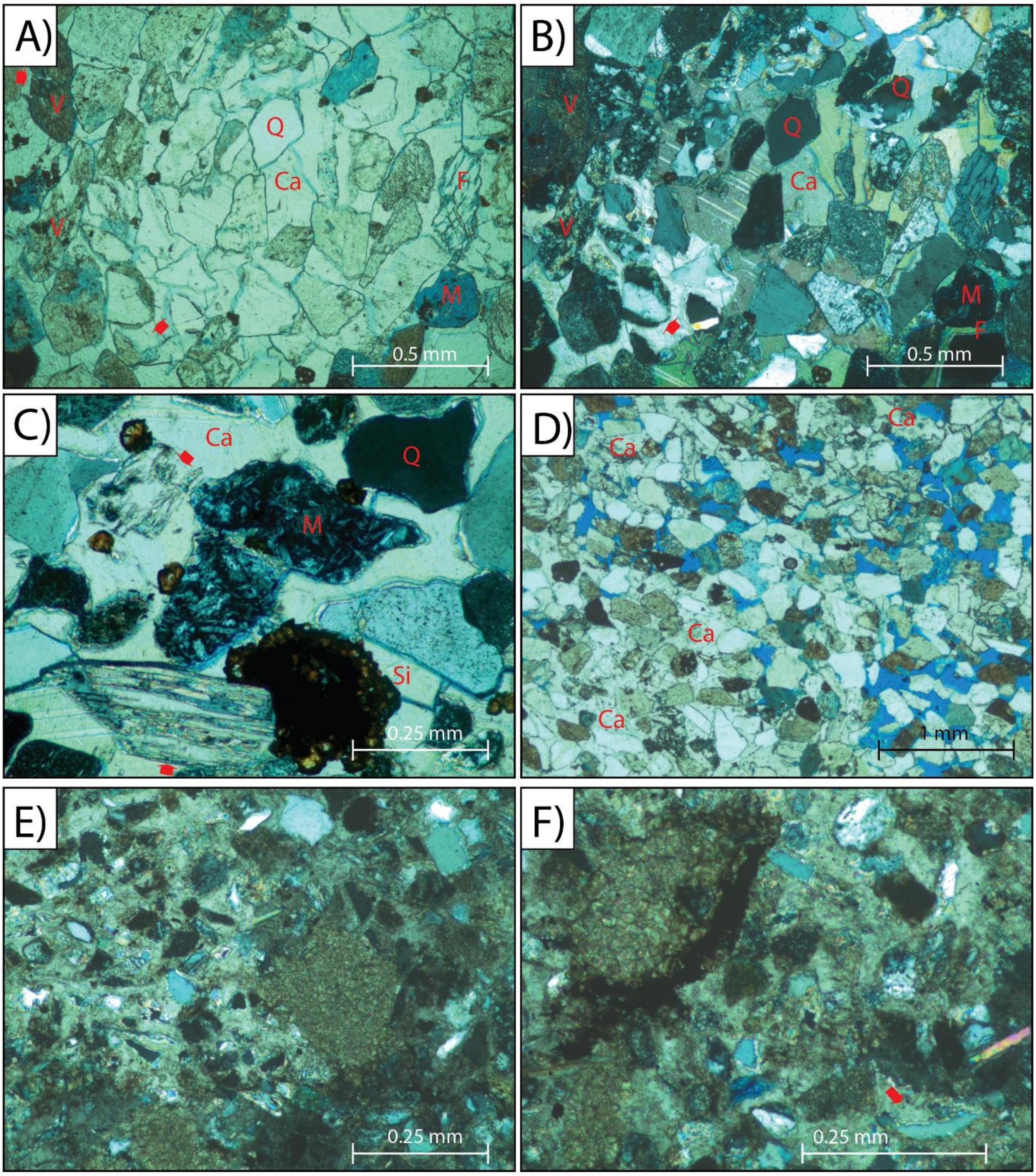


Figure 5.2: Calcite cemented sandstones. A) Calcite cemented distributary mouth bar deposit at Teistberget (162 m in log). B) Poikilitic calcite cement enclosing partly floating detrital grains. Same as Fig. 5.2A, cross polarized light. C) Calcite replacing detrital feldspar along cleavage (lower red arrow). Remnant detrital grain replaced by calcite (upper red arrow). D) Inter- and intragranular porosity (stained blue by epoxy) occur adjacent to calcite (Ca) cemented domains and illustrates a patchy distribution of calcite cement. E) Pervasive, inclusion rich, calcite cemented very-fine sandstone. F) Close-up from Fig. 5.2E, of blocky calcite infilling voids and corroded contact to detrital quartz (red arrow). A, D in plane polarized light. B, C, E, F in cross polarized light.

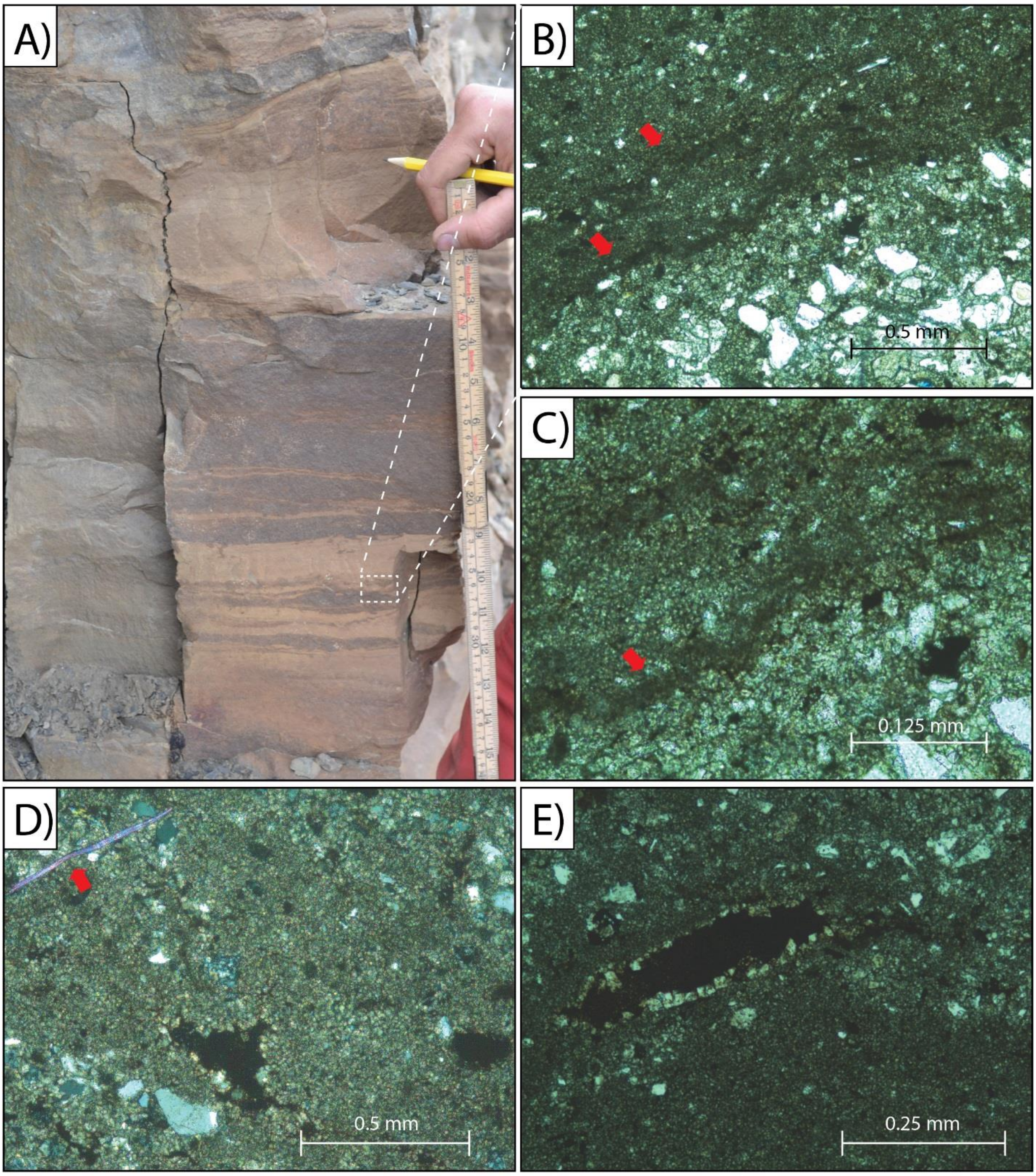


Figure 5.3: “Calcite cemented sandstones” of facies J. A) Outcrop photo from Klement’evfjellet (at 43 m in log) of undulatory “stromatolite-like” lamination. B) Detrital grain enclosed by micrite (light areas) alternates with micrite and organic matter (dark areas). C) Undulatory laminae containing organic matter (red arrow). D) Uncompacted mica (red arrow) within micrite. E) Equant crystals along rim of cavity. B, C, E in plane polarized light. E in cross polarized light.

Discussion

The overall abundance and variance in distribution and nature of calcite cementation described above suggests multiple mechanisms for calcite cementation in the De Geerdalen Formation.

In shallow marine siliciclastic sediments, eogenetic calcite cement is common and accompanies sulphate reduction and methane oxidation (Morad, 1998). Furthermore, eogenetic calcite is commonly found along parasequence boundaries and could be attributed to prescence of carbonate bioclasts and/or prolonged residence time of sediments at very shallow burial depths (Ketzer et al., 2003). Flooding surfaces are abundant in the De Geerdalen Formation and such conditions may have contributed to the abundant field-observations of calcite cementation, e.g. upper parts of parasequences.

Fine microspar crystals are typically associated with eogenetic cementation and create pore-filling cement fabrics (Worden and Burley, 2003). Microsparitic calcite cements are sensitive to recrystallization and may recrystallize into poikilitic calcite during mesogenesis (Saigal and Bjørlykke, 1987; Morad, 1998). Mesogenetic calcite can also develop even with apparent lack of pre-existing carbonate minerals. In such cases influx of source-rock derived CO_2, alkaline earth-elements from decomposing lithic fragments and adjacent mudstones and evaporites may induce precipitation (Worden and Burley, 2003). Accordingly maturation of organic material within the De Geerdalen Formation and the underlying Botneheia Formation, dissolved volcanic fragments and the heterolithic nature of the De Geerdalen Formation may have provided multiple mechanisms for mesogenetic calcite formation.

Patchy distribution of calcite cement is commonly observed, and may reflect initial patchy precipitation or it may be due to subsequent partial removal of cement during burial or outcrop exposure (Boggs, 2009). Displacive crystallization of calcite cement may locally force the detrital grains into a loosely packed fabric, but is mostly associated with calcrete profiles in arid climate (Braithwaite, 1989). Partial replacement of silica grains by calcite cement may also result in a resembling fabric (Boggs, 2009). Observations of calcite replacing detrital feldspar grains, lithic rock fragments and quartz point towards the latter mechanism imposing controls in the studied thin-sections. Replacement of detrital grains may also account for the high intergranular volumes in the sandstones with the most pervasive poikilitic cementation.

Many samples were obtained from the lower parts of distributary channels (FA 2.3). Morad et al. (2010) attributed calcite at base of such channel deposits to nucleation of calcite cement on mud and carbonate intraclasts. Observed drusy textures are seen to infill cavities, and the voids could represent dissolved fragments. Alternatively some of the drusy fabrics could have been formed due to recrystallization of micrite or carbonate grains and accordingly be termed neospar (Boggs, 2009).

In the "calcite cemented sandstones" of facies J (Chapter 3), abundant micrite and uncompacted mica indicates an early timing of calcite cementation. In limestones, presence of micrite is commonly attributed to deposition under fairly low-energy conditions (Boggs, 2009). Whereas deposition in more agitated waters are thought to selectively remove micrite, and enhance sparry calcite cementation of larger carbonate grains (Boggs, 2009). As the observed sparry calcite is localized in close connection to cavities, a formation assigned to diagenesis is preferred in the studied sections. Blocky crystals on rims of voids is typical for meteoric phreatic cement.

The observed undulatory lamination resembles stromatolitic structures, and presence of laminated organic material and cavities are additional features characterstic of microbial formation (Boggs, 2009). These findings support observations made by Tugarova and Fedyaevsky (2014) who suggested micro-organisms and biochemical precipitation in connection to cyanobacterial mats as mechanism for calcite cementation, within siliciclastic intervals in the De Geerdalen Formation on Edgeøya. The characteristic undulatory lamination was only found in Agardhbukta, and at other localites facies J had a massive appearance. Accordingly some of the deposits may have originated by the processes regime outlined by Tugarova and Fedyaevsky (2014), and others not. Lack of evidence for significant compaction (e.g. squeezing of micas) support an early diagenetic formation, concordant with Tugarova and Fedyaevsky (2014).

Stromatolites are laminated biosedimentary structures formed largely by trapping and binding of sediments as well as chemical action of cyanobacteria in shallow marine environments (Hofmann, 1973). Growth of algal mats through sediment cover generate successive stacking of laminae (Boggs, 2009). As cyanobacteria carry out photosynthesis occurrence is restricted to shallow water depths permitting adequate light. Modern stromatolites form mainly in the shallow subtidal, intertidal and supratidal zones of the ocean and in lacustrine environments (Boggs, 2011). In turn pointing toward a shallow marine or possibly lacustrine origin of some deposits assigned to facies J (chapter 2). Occurrence of facies J is restricted to the upper parts of the De Geerdalen Formation, and the Isfjorden Member, within a paralic delta plain environment.

Siderite

Siderite was commonly observed in the field, e.g. in wave-rippled sandstones and mudstones, and discrete siderite layering was assigned to "Carbonate cemented sandstones" of facies J. In thin-section authigenic siderite occur as scattered solitary rhombs and aggregates. The rhombs are euhedral to slightly rounded and found in most thin-sections, although in modest abundance. Siderite rhombs and aggregates tend to form in close connection to unstable grains, e.g. micas showing expanded texture (Fig. 5.4A) and dissolution porosity. Siderite aggregates commonly exhibit an opaque core (Fig. 5.4B).

Authigenic siderite was also seen to enclose calcite cement and quartz in connection to organic matter (Fig. 5.4C). Authigenic siderite detected in facies J, showed a characteristic "star-like" texture of radiating blades (Fig. 5.4D).

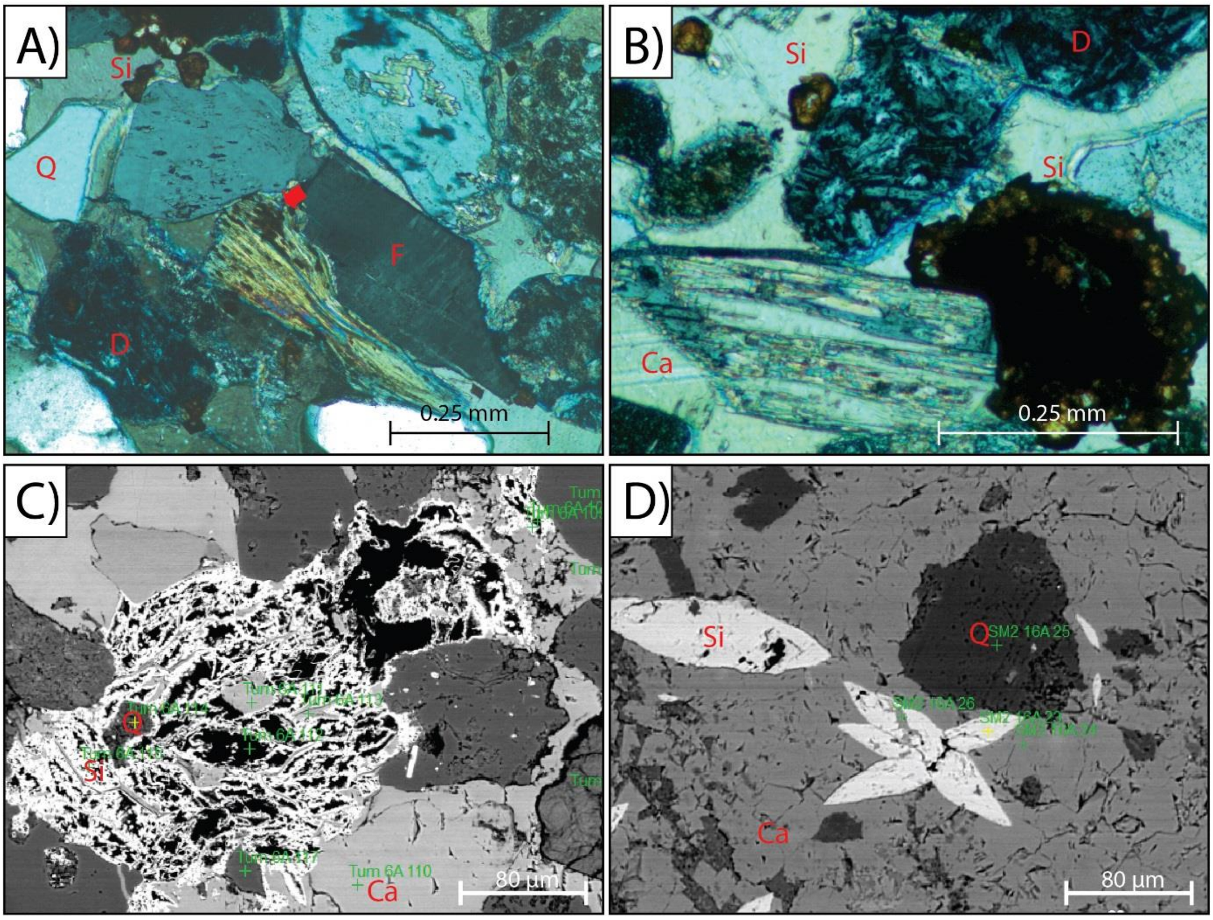


Figure 5.4: Authigenic siderite. A) Siderite rhombs dispersed in calcite cement and within expanded mica (red arrow). B) Authigenic siderite rhombs and aggregates close to dissolution porosity (D). C) Siderite (bright colour) growing in porosity and enclosing detrital quartz and calcite cement. D) Siderite making "star-like" texture within calcite aggregates in facies J. A, B in cross polarized light. C, D are SEM backscatter images.

Discussion

Formation of siderite predominantly occur in organic-rich brackish to meteoric pore-waters depleted of sulphate and is commonly found in fine grained deltaic to coastal sediments (Morad, 1998). In this study abundant siderite was noticed in interpreted crevasse sandstones occurring above organic rich floodplain sediments. According to Morad (1998) siderite preferentially forms in floodplain fines and crevasse channels in fluvial environments, and thus the abundant siderite may support a crevasse channel interpretation as opposed to a distributary channel with modest dimensions. The tendency of siderite being more abundant upwards in the formation may reflect the interpreted upwards shift from marine and delta front into a more delta plain dominated depositional style.

Oxidation of organic matter lead to increased pH and precipitation of carbonate minerals (Morad, 1998). Leaching of biotite and volcanic fragments may have provided a local source of iron, allowing precipitation of authigenic siderite. Authigenic siderite is enclosed by, and thus pre-date, calcite cement. And accordingly alterations of volcanic fragments commenced prior to calcite cementation, given the suggested association to siderite formation. Observations that siderite also engulfs calcite cement may point toward multiple stages of formation, possibly in connection to introduced organic matter to the system.

Quartz cementation

In general, quartz overgrowths are characterized by irregular edges and a poor development. Quartz cement has long and concavo-convex contacts to adjacent grains and quartz cement is seen to make triple junctions (Fig. 5.5A, 5.5B). In the fine to medium grained fraction of samples quartz cement is observed to have an uneven distribution adjacent to domains dominated by kaolinite or calcite cementation (Fig. 5.5A). In very-fine to fine grained sandstones quartz cement is more extensive and seems to have a slightly more homogenous distribution.

The original shape of the detrital grains were rarely visible, and thus detecting overgrowths were primarily based on the irregular, almost floating, outer boundary interpreted as subhedral quartz cement. Occasionally dust rims and fluid inclusions enclosed by euhedral terminated and syntaxial overgrowths were observed (Fig. 5.5C, 5.5D).

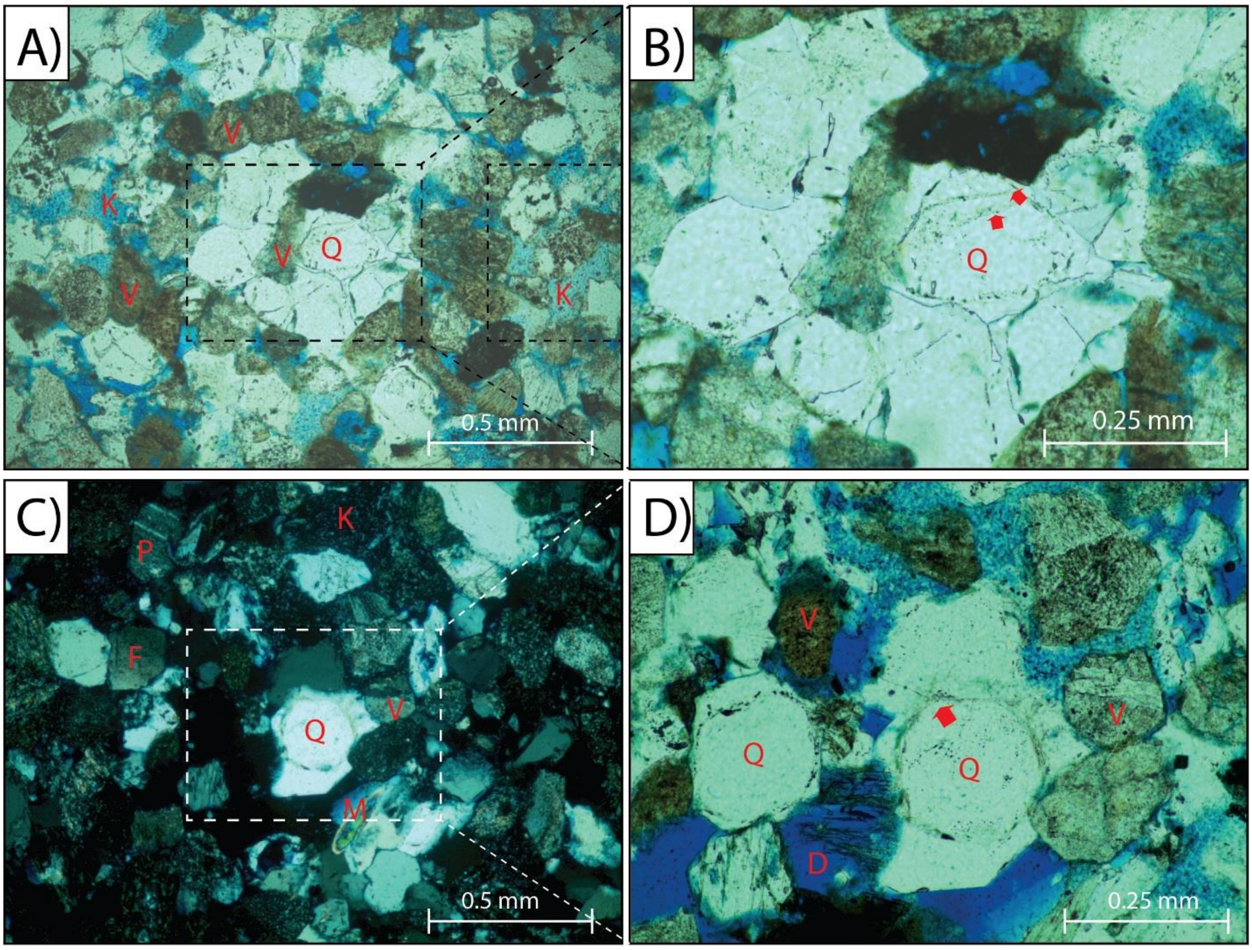


Figure 5.5: Quartz cementation in fine-grained sandstones. A) Quartz cemented domain with adjacent porefilling kaolinite cement (K). B) Quartz overgrowths making triple junctions surrounded by partly dissolved lithic fragments. Possible multiple dust rims (red arrows). Close-up from Fig. 5.5A. C) Syntaxial quartz overgrowth. D) Euhedral terminated quartz overgrowth enclosing rounded detrital quartz. Cement exhibit irregular termination towards kaolinite, volcanic fragments (V) and dissolved fragment (D). A, B, D in plane polarized light. C in cross polarized light.

Well-developed stylolites were neither observed in the field nor during thin-section inspections. However, in the very-fine and fine-grained sandstones mica is more abundant and dissolution of detrital grains was observed along the contact with micas. Note that this is a seldom phenomenon in the studied sections. Sutured contacts between detrital quartz grains were not detected. Quartz cementation is more common in samples that are not extensively calcite cemented. Quartz overgrowths are engulfed by, and thus pre-date, calcite cement.

Discussion

Quartz cementation is a function of burial depth and temperature, and it typically forms at temperatures above 70°C (Ehrenberg, 1990; Worden and Burley, 2003).

Source of silica for quartz cementation during diagenesis is much debated, although pressure solution appear to be the most important mechanism during mesogenesis (Boggs, 2009). However, the sparce abundance of observed stylolites and lack of sutured grain contacts complicates using a stylolite model (e.g. Oelkers et al., 1996) as silica source for the system. The seemingly uneven distribution of quartz cement, may reflect localized micro-environments with favorable conditions for quartz precipitation. Tentatively dissolution of lithic fragments, in particular volcanic fragments, may have provided a local source of silica (e.g. Mørk, 2013). Observations of several dust rims coating detrital grains may reflect multiple eposides of quartz overgrowth. Tentively such a multistage formation could be related to changing stability fields of dissolving grains (presuming a silica source from lithic clasts) or related to burial history.

Grain coating chlorite and illite have not been detected, and accordingly presence of clay minerals is not thought to have inhibited quartz cementation. Observations that quartz cementation is most pervasive where calcite is absent may indicate that the subsequent calcite cementation limited pore-water circulation and accordingly silica diffusion and precipitation. Corroded contacts to dissolved fragments indicate that dissolution processes continued subsequent to quartz cementation.

Clay minerals

Kaolinite occur as booklets and vermicular aggregates that fills intergranular pores or replace detrital minerals. In the polarization microscope it is not diagnostically identified, but occurrence was verified by SEM. Kaolinite is an important constituent of the studied sandstones, and distribution tend to be localized in clusters where quartz and calcite cement are absent (Fig. 5.6A).

Intergranular kaolinite occur in close connection to partly dissolved feldspars, lithic fragments and moldic porosity (Fig. 5.6B, 5.6C, 5.6D). The adjacent framework minerals to porefilling clay cement commonly exhibit corroded and irregular terminations. Kaolinite also occur in oversized pores (Fig. 5.6C). Kaolinite aggregates are found within moldic porosity and inferred to replace detrital feldspar when the outer feldspar rim is preserved.

In SEM kaolinite is seen to show vermicular and booklet textures, and occurrence in connection to illite and chlorite was seen (Fig. 5.6E). Porosity is preserved as microporosity between the booklets within aggregates. Kaolinite is also seen to replace detrital mica showing expanded textures (Fig. 5.6F).

Labile lithic fragments are common constituents in the studied thin-sections and prone to alterations. Chlorite and illite are seen to replace lithic fragments resulting in rounded to sub-rounded molds within poikilitc calcite cement (Fig. 5.7). Illite and chlorite occurring as grain-coating cement was not detected.

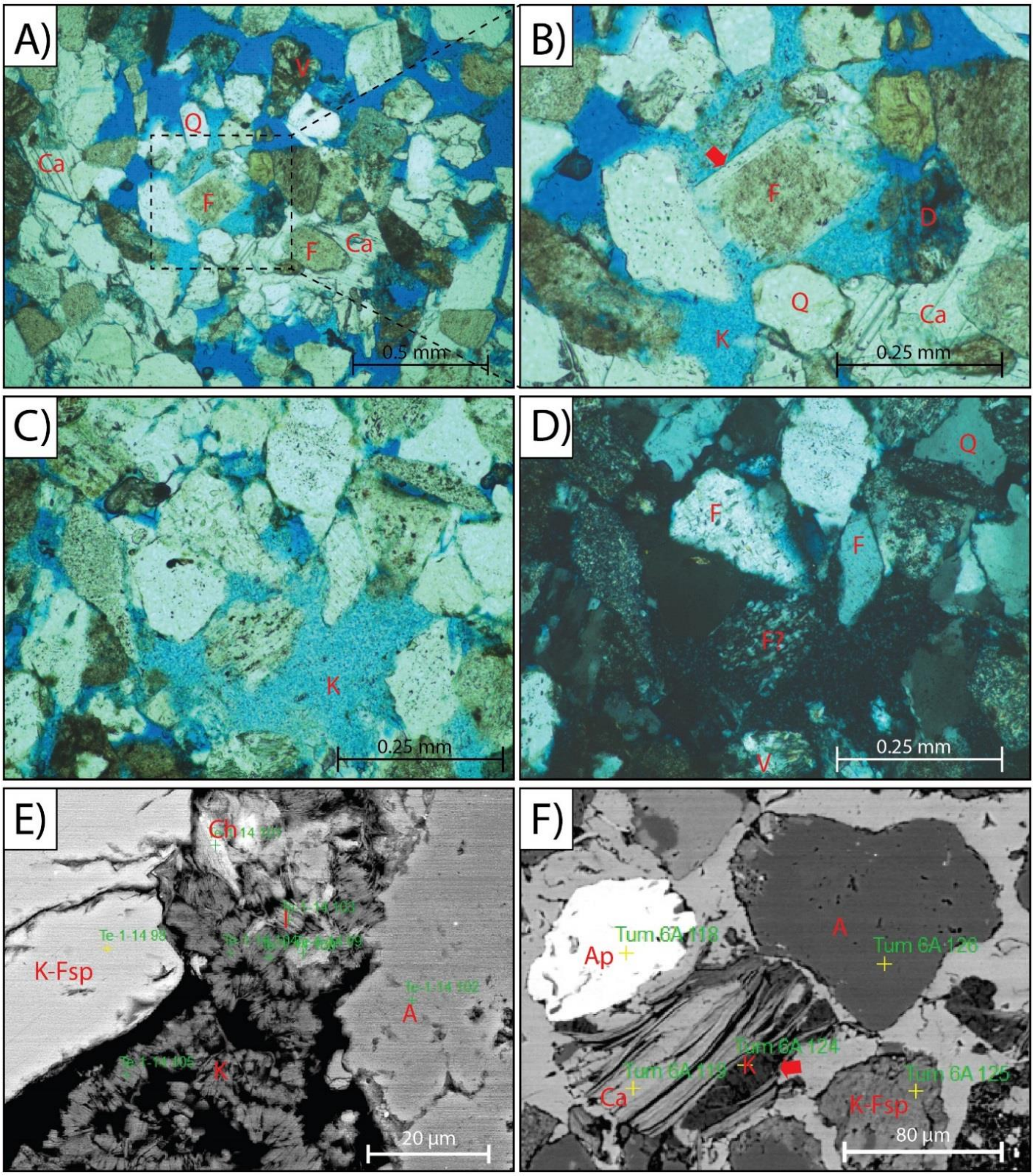


Figure 5.6: Authigenic kaolinite. A) Intergranular kaolinite infilling porosity adjacent to poikilitic calcite cement. B) Close-up of kaolinite aggregates in close connection to partly dissolved feldspar with preserved outer rim (red arrow) and dissolution porosity (D). C) Kaolinite in oversized pore. Abundant dissolution of adjacent feldspars and lithic fragments. D) Same as C, crossed polars. E) Vermicular kaolinite, illite and chlorite in pore-throat between detrital K-feldspar and albite (A). F) Kaolinite replacing mica showing expanded texture (red arrow). Mica has also been replaced by calcite cement (Ca). Apatite (Ap), albite (A) and K-feldspar are enclosed by calcite cement. A, B, C in plane polarized light. D in cross polarized light. E, F are SEM backscatter images.

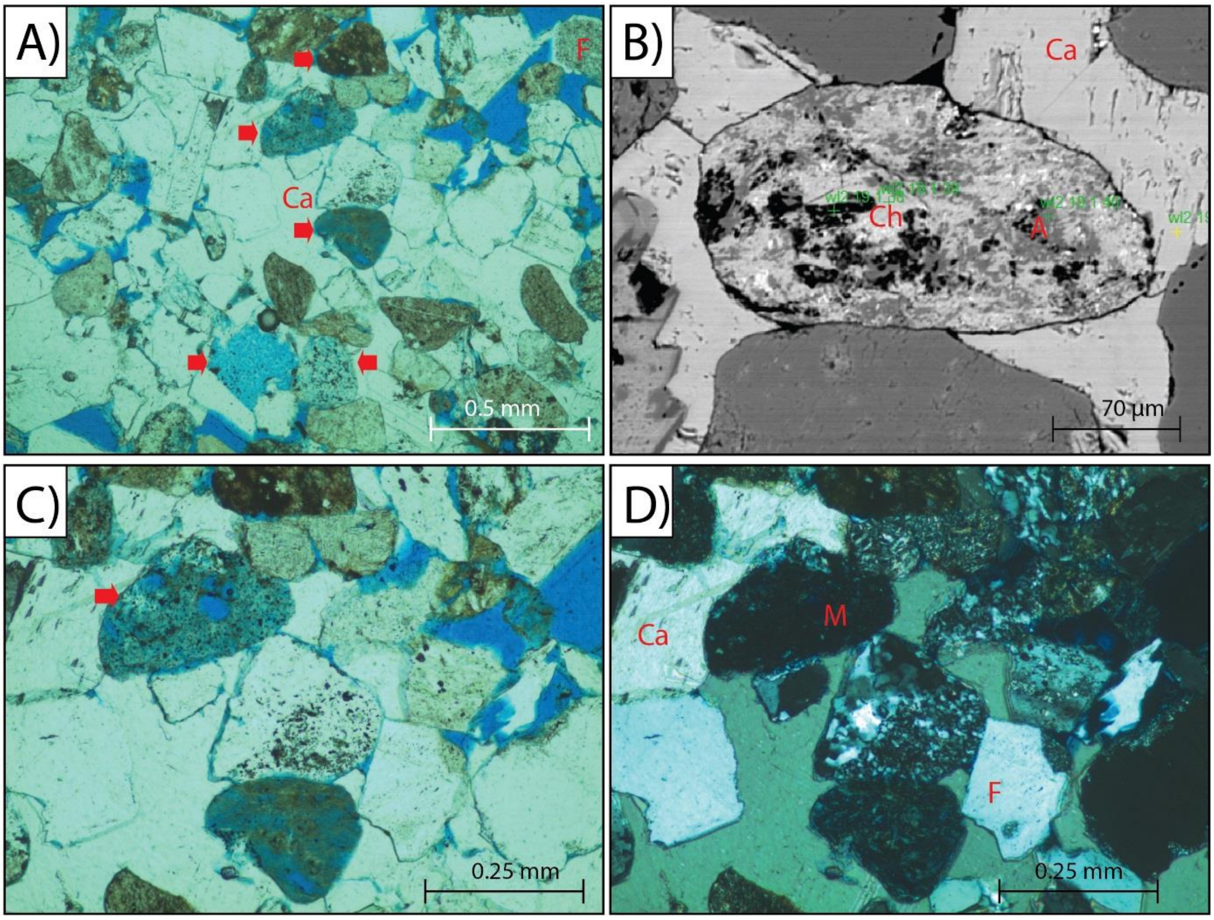


Figure 5.7: Dissolution and replacement of lithic fragments by chlorite and illite. A) Lithic fragments show various stages of dissolution and alterations (red arrows). B) SEM backscatter image showing a patchy distribution of albite (A), replacive chlorite (Ch) and dissolution porosity. C) Green chlorite alternates with porous areas and remnants of detrital grain (red arrow). D) Same as C, cross polarized light. A, C in plane polarized light.

Discussion

Kaolinite forms during weathering and eogenesis and is associated with extreme weathering profiles and humid climate with high meteoric influx (Bjørlykke, 1997; Worden and Burley, 2003). Formation of kaolinite induced by dissolution of mica and feldspars is expressed by the following weathering reaction (Bjørlykke, 1997):

$$2K(Na)AlSi_3O_8 + 2H^+ + 9H_2O = Al_2SiO_5(OH)_4 + 2H_4SiO_4 + 2K(Na^+)$$

Note that this weathering reaction is dependent on percolating water flow as cations and silica must be removed for precipitation of kaolinite (Bjørlykke, 1997). The interpreted deltaic environment is favorable for precipitation of eogenetic kaolinite (Worden and Burley, 2003).

Kaolinite may also replace mica and feldspar during mesogenesis (Worden and Burley, 2003). Dissolution of feldspar and lithic fragments are observed and may be considered a possible source of silica for precipitation of kaolinite aggregates.

Illite is an exclusively burial authigenic clay, and forms at temperatures exceeding approximatly 70°C in potassium-bearing formation waters (Worden and Burley, 2003). Several different reaction processes may take place in order to form illite, and is constrained by temperature, composition and access of K^+ (Bjørlykke, 1997). During these reactions illite and quartz or albite is formed in addition to access water, while kaolinite and K-feldspar is dissolved (Bjørlykke, 1997). The occurrence of illite in close connection to kaolinite aggregates may suggest such replacement reactions of kaolinite, with K^+ provided from dissolving K-feldspar. Mudrock beds within the formation may also provide potassium (Worden and Burley, 2003).

Authigenic chlorite may form from kaolinite if provided a source of Fe and Mg (Bjørlykke, 1997) at temperatures ranging 165 to 200°C with corresponding depths of 3.5 to 4.5 km (Worden and Burley, 2003). Accordingly the gathering of kaolinite, illite and chlorite described from intergranular pores could be explained by a local source of iron (e.g. volcanic fragments, biotite) and replacement of kaolinite during mesogenesis. Alternatively, coexistence of chlorite and kaolinite may be due to changing pore water composition during eogenesis (Worden and Burley, 2003). Chlorite replacing lithic fragments are of obvious authigenic origin. The resulting molds are enclosed by calcite cement, pointing toward replacement during mesogenesis. As the molds commonly does not exhibit connectivity to the remaining porosity the replacement of lithic fragments by clay minerals is thought to have occurred during very limited pore water percolation. Accordingly, the lithic fragments are suggested to have provided the necessary iron for chlorite precipitation.

Feldspar

Feldspar shows evidences of both dissolution and cementation processes in the studied thin-sections. The latter is characterized by overgrowths and albitization of detrital grains. Detrital feldspar commonly includes dissolution porosity, resulting in feldspar pseudomorphs when enclosed by calcite cement. Occasionally the rim of the detrital feldspar is preserved, and resisted dissolution unlike the interior of the grain (Figs. 5.8A, 5.8B).

Albitization of detrital feldspar was detected by SEM investigations, and shows partly albitized K-feldspar grains enclosed by calcite cement (Fig. 5.8C). Albite rims were also seen on K-feldspar grains (Fig 5.8D), and may represent a less developed stage of albitization. Alternatively the cases where albitization seemingly occur in a patchy and extensive distribution could simply be a primary feature of the detrital grains.

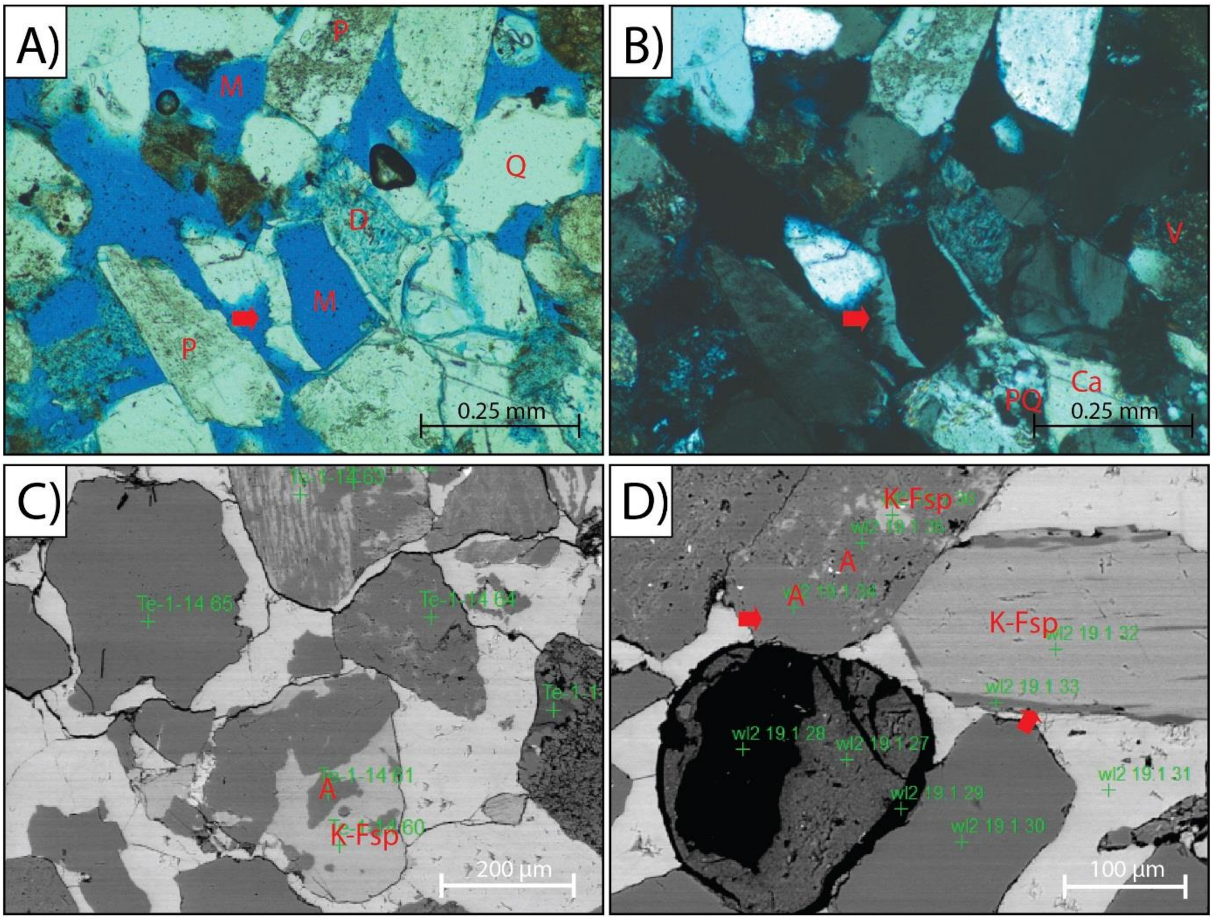


Figure 5.8: Diagenetic feldspar. A) Optical micrograph in plane polarized light of moldic porosity (M) enclosed by feldspar overgrowth adjacent to partly dissolved feldspar (D). Sericitization on plagioclase grains (P). B) Irregular outer boundary of feldspar overgrowth (red arrow). Same as A, cross polarized light. C) Detrital K-feldspars replaced by albite along rims (red arrows). D) Partly albitized K-feldspar grain. C, D are SEM backscatter images.

Discussion

Dissolution of detrital feldspar in near surface conditions is typical in humid warm conditions with meteoric water flow, and have previously been discussed in relation to clay minerals. Mesogenetic dissolution of detrital K-feldspar chiefly occur at temperatures ranging 50-150°C with inherent depths ranging 1.5 to 4.5 km (Wilkinson et al., 2001; Worden and Burley, 2003).

Feldspar cement occurs mainly as overgrowths on detrital grains. Provided adequate consentrations of dissolved silica and Na^+ and K^+ ions, feldspar overgrowth may form during eogenesis and some stages of mesogenesis (Boggs, 2009). Due to slightly different mineral chemistry and crystallography the diagenetic feldspars tend to be more thermodynamically stable in the diagenetic environment than substrate grains of detrital feldspar (Worden and Burley, 2003). In consequence, the parent grain could dissolve and leave a rind of the diagenetic overgrowth, as observed in this study. Alternatively the feldspar rims could represent the most thermally stable outer parts of zoned detrital feldspars.

Albitization occur when detrital Ca-bearing plagioclase or K-feldspar are replaced by albite (Worden and Burley, 2003). Albitization is suggested by most workers to occur at temperatures in the order of 100-150°C (Boggs, 2009). However, albitization is also reported to occur at temperatures as low as 70-100°C (e.g. Morad et al., 1990). Albitization of detrital feldspar may occur by direct replacement by albite or through intermediate stages where feldspars are replaced by e.g. calcite or anhydrite and in turn these minerals by albite (Boggs, 2009). Alternatively albitization may occur by partial dissolution of detrital feldspar and subsequent precipitation of albite in dissolution voids or along fractures (Boggs, 2009). Albite rims are seen on detrital grains enclosed by calcite, pointing toward albitization prior to calcite precipitation. As calcite is seen to replace detrital feldspar it is difficult to determine if albitization occurred by direct replacement or through intermediate stages of calcite replacement.

6 Discussion

The depositional environment interpreted from the studied sections of the De Geerdalen Formation is that of a paralic environment, with distinct deltaic signatures. Through facies analysis (Chapter 3 and 4) marine to lower shoreface, delta front and delta plain sediments have been distinguished, representing distal to proximal deltaic environments. Evidences of fluvial-, wave-, storm- and tidal influence have been observed. The overall shallowing upwards trend of the De Geerdalen Formation is noticed at all localities and is concordant with previous studies (Knarud, 1980; Mørk et al., 1982; Riis et al., 2008; Høy and Lunschien, 2011; Lundschien et al., 2014; Rød et al., 2014).

The following sections aims to discuss; (I) deltaic sequences, (II) distribution of facies and facies associations, (III) trends between localities and infer about the overall depositional environment, (IV) textural properties of sediment, V) implications of diagenesis for porosity in sandstones.

Interpretations will be based on facies models (e.g. Bhattacharya, 2006; Clifton, 2006) with data from classified facies and facies associations (Chapter 3 and 4) and logs (Appendix A). Two correlation panels (east and west of Storfjorden) have been created based on stratigraphical relationships. Discussion concerning textural properties and diagenesis of sandstones is based on petrographic observations.

6.1 Delta sequences

Modern deltas are discrete shoreline protuberances, formed where an alluvial system enters a basin and supplies sediment more rapidly than it can be redistributed by basinal processes (Bhattacharya, 2006). Thus all deltas are fundamentally regressive in nature (Bhattacharya, 2006). A regressive development concur with interpretations presented herein, as the De Geerdalen Formation shallow upwards from marine sediments (FA 1), trough delta front sands (FA 2) to delta plain and interdistributary sediments (FA3) (Fig. 6.1).

Factors controlling the morphology and facies architecture of deltas are; the proportion of wave, tide and river processes, the salinity contrast between river water and standing body of water, sedimentation rate and grain size, and the water depth into the which the river flows (Reading and Collinson, 1996; Bhattacharya, 2006). Additionally basin configuration, depth, subsidence pattern and proximity to a shelf edge may have an influence. Basinal processes may redistribute sediments to a point where the fluvial source and delta morphology no longer can be recognized. In such a case more general environmental terms (i.e. paralic, strandplain, coastal plain) may be preferable (Bhattacharya, 2006).

Modern deltas occur at a wide variety of scales ranging from continental-scale depositional systems (e.g. Mississippi delta) to bayhead deltas forming components of other depositional systems (e.g. estuarine or lagoonal systems). Furthermore, continental-scale deltas may incorporate other smaller-scale depositional systems (e.g. crevasse deltas, barriers, lagoons, strandplains), providing complex vertical facies sequences and facies architecture (Bhattacharya, 2006).

Delta complexes prograde (i.e. constructive phase) until over-extension leads to abandonment and reworking during transgression (i.e. destructive phase), resulting in cyclic deposition patterns. Traditionally the constructive phase has been associated with a fluvial character (e.g. Frazier, 1967), whereas formation of barrier islands and lagoonal systems were associated with the destructive phase (Boyd and Penland, 1988; Bhattacharya, 2006). However, Bhattacharya and Giosan (2003) showed that prograding large-scale deltas can contain depositional elements (e.g. bayhead deltas, lagoons, bays, barrier islands, strandplains) typically associated with non-deltaic coastlines or destructional phases of deltas. Thus the occurrence of barrier (island) sandstones, chiefly found in the lower parts of the De Geerdalen Formation, should not be conclusively associated with transgressive phases.

Down-dip, deltaic deposits are characterized by a prograding clinoform geometry (Bhattacharya, 2006). And as mentioned in the introduction seismic clinoform sequences have been traced across the western Barents Sea to Kvitøya (Riis et al., 2008; Glørstad-Clark et al., 2010; Høy and Lunschien, 2011; Lundschien et al., 2014). Combined with previous work (Knarud, 1980; Mørk et al., 1982; Rød et al., 2014), establishing the paleodepositional environment as a prograding delta.

In the lower part of the De Geerdalen Formation facies associations varies from offshore-lower shoreface deposits (FA 1) to the more proximal delta front deposits of barriers, mouth bars and distributary channels (FA 2.1, 2.2 and 2.3).

In the southwestern sections (Agardhbukta), offshore to offshore transition sediments (FA 1.1 to FA 1.2) grade upwards into lower shoreface and upper shoreface to foreshore deposits (FA 1.3 to FA 2.1), representing deposits of progressively shallower environments following Walther's Law. Following the definition from Van Wagoner et al. (1990, p. 8), a parasequence is "a succession of relatively conformable and genetically related beds or bedsets bounded by flooding surfaces or their correlative surfaces". Accordingly a facies association suddenly being overlain by a much more distal facies associations reflect a violation of Walther's law, and a flooding surface marks the onset of a new parasequence. Parasequences are commonly terminated by wave ripples and a sharp boundary to the overlying shales, marking a flooding surface. These recurring shallowing upwards successions occur on a scale of 2 to 15 meters and are found in the lower parts of the De Geerdalen Formation throughout the study area.

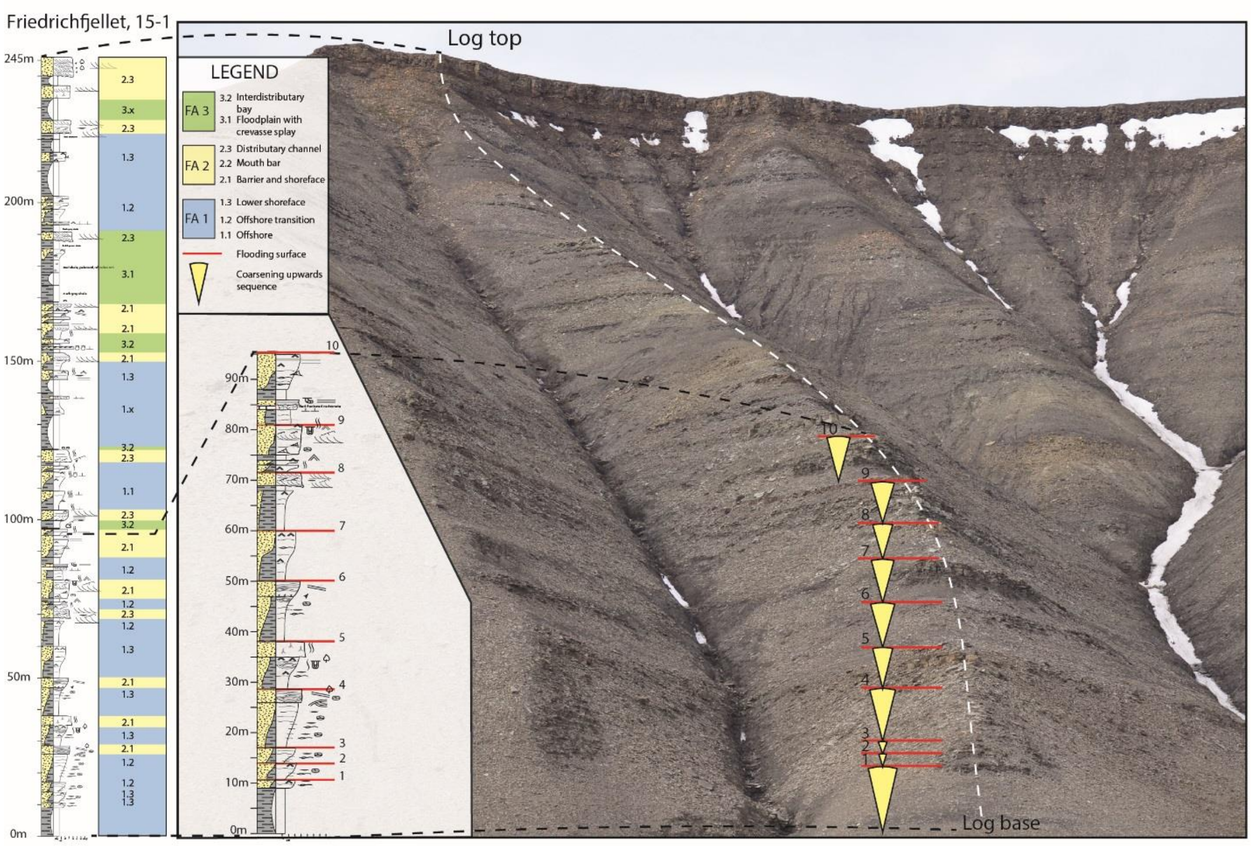


Figure 6.1: The De Geerdalen Formation measured on Friedrichfjellet shallow upwards from offshore-lower shoreface deposits (blue) to delta front (yellow) and delta plain deposits (green). Parasequences terminated by flooding surfaces (numbered for comparison with attached log) comprise small-scale CU sequences in the lower part of the section.

Deltaic coastlines are characterized by depositional regressions and transgressions. Abundant fluvial sediment supply the prograding lobes, while subsidence and marine reworking occur in abandoned parts of the delta system (Bhattacharya, 2006). Stacked parasequences, separated by flooding surfaces marking transgressive events, reflects cyclic deposition characteristic of delta sediments (Reading and Collinson, 1996). Deltaic facies sequences occur on three scales; large-scale (due to eustacy, tectonics and climate in hinterland), medium-scale (due to switching of delta lobes or distributaries within a stable depocentre) and small-scale (due to differential subsidence on delta plain) (Reading and Collinson, 1996). As mentioned in the introduction Mørk et al. (1982) attributed the occurrence of repeated coarsening upwards sequences in the De Geerdalen Formation to progradation and abandonment of minor delta lobes. The interpreted parasequences in the De Geerdalen Formation reflect a small-scale cyclicity that is characteristic, though not exclusive, to deltaic successions (Bhattacharya, 2006). Scree-cover complicated mapping of parasequences in the upper parts of the formation.

6.2 Distribution of facies and facies associations

Offshore to lower shoreface deposits (FA 1) and delta front deposits (FA 2) dominate the base of the De Geerdalen Formation, whilst delta plain and interdistributary deposits (FA 3) dominate the upper part. A tentative correlation of the eastern and western localities is given in Fig. 6.2 and Fig. 6.3 respectively. Combined with field observations and log interpretations (Appendix A), used to discuss distribution of facies and facies associations, the correlation panels have appliance as a conceptual model enabling to discuss depositional trends within this regional deltaic system.

FA 1 occurs in shallowing upwards parasequences (described above), which occasionally terminate in barrier sandstones (FA 2.1). Possibly resembling deposits seen at Botneheia on central Spitsbergen, as described by Rød et al. (2014), who also noted the extensive marine reworking upon HCS sandstones. Knarud (1980) and Mørk et al. (1982) interpreted these deposits to represent reworked delta-front sheetsand facies. Rød et al. (2014) added the suggestions that these deposits may represent reworked mouth bars (Bhattacharya, 2006) or sand sheets in the offshore area between fair-weather- and storm wave base (Willis and Gable, 2001).

A combination of the mechanisms discussed above may be responsible for the formation of FA 1 in this study. Abundant HCS sandstones and mudstones may indicate a storm dominated shallow marine environment, and an overall heterolithic expression testify to alternating energy during deposition (Reading and Collinson, 1996). A distal position to the deltaic source in the east combined with decreasing accommodation space towards west (Rød et al., 2014) may have allowed for greater sediment reworking by basinal processes. Note that FA 1.1 and FA 1.2 are mainly documented at Muen and in the Agardhbukta sections. At other localites lower shoreface deposits (FA 1.3) are commonly exposed above a scree cover interval, implying that the distal compounds of FA 1 are scree covered or non-existent.

At Barentsøya (Svartnosa and Mistakodden) mouth bar sediments (FA 2.2) sharply overlie the Tschermakfjellet Formation and marks the onset of the De Geerdalen Formaiton. The occurrence of these proximal delta front sediments clearly contrast to the distal sediments discussed from the lower part of Agardhbukta sections and illustrate the diachronous development of the De Geerdalen Formation. Interpreted mouth bar sediments (FA 2.2) show decreasing thicknesses toward the more northern localities of Hellwaldfjellet and Wilhelmøya, and Teistberget towards west, whereas no delta front deposits were interpreted as mouth bars in Agardhbukta.

Distributary channels (FA 2.3) are recognized at all localities. At Svartnosa a 6 m thick and laterally extensive amalgamated distributary channel is located. Based on dimensions, lack of heterolithic nature, abundant plant fragments and height of cross-bedding (up to 80 cm), this deposit has been interpreted as proximal. Similar proximal distributary channels are found at Hellwaldfjellet and in the middle parts of Teistberget and Wilhelmøya sections.

At base Teistberget a complex of stacked large-scale cross-bedded sandstones (facies A) is interpreted as a distributary channel (FA 2.3). Abundant mud drapes (facies K) and alterations between large- and small-scale bedforms (facies A and B) point towards an environment with fluctuating energy during deposition. The heterolithic nature of this deposit, contrasts with the proximal deposits previously described and a distal delta front setting where tidal influence is more pronounced, may explain these subtle variations in flow regime (Reading and Collinson, 1996; Dalrymple and Choi, 2007). Alternatively, the heterolithics could form due to seasonal discharge variations of the distributary (Reading and Collinson, 1996). However, the former explanation is preferred as distributary channels in Agardhbukta exhibit a heterolithic nature and occur within marine intervals pointing toward a position on the distal delta front. Additionally occurrence of proximal delta plain distributary channels are confined to eastern localities and the upper part of the De Geerdalen Formation at the western localities. Whereas interpreted distributary channels on the distal delta plain are restricted to the lower part of the stratigraphy at the western localities, implying a distal setting to the northwesterly prograding delta (e.g. Lundschien et al., 2014).

There is also a possibility that these distal distributary channels may represent tidal channels (e.g. Reading and Collinson, 1996; Dalrymple and Choi, 2007; Davis, 2012). Lateral migration of tidal channels and point bars could have resulted in large-scale epsilon cross-bedding (Dalrymple and Choi, 2007), resembling the large-scale cross-bedded sandstones previously assigned to facies A. Tidal channels may occur in various depositional settings (e.g. barrier island, tidal flat) (Reading and Collinson, 1996). The lack of observed mud cracks (at this outcrop and in the De Geerdalen Formation in general) may contradict a tidal flat as substrate for channels. On the other hand, muddy features (e.g. mudcracks, bioturbation) and the mudflat itself show low preservation potential in the geological record in contrast to tidal channels (Davis, 2012). Additionally tidal flats are most extensive in macrotidal settings (Reading and Collinson, 1996), whereas a microtidal setting has been interpreted for the De Geerdalen Formation (Knarud, 1980). Based on lack of observations of diagnostic tidal signatures (e.g. double mud drapes) and abundance of plant fragments, an incorporation into the distributary channel facies association is preferred, despite the overall heterolithic expression of these deposits. Tentatively one may speculate upon if the low-sloping gradient of the prograding system (Glørstad-Clark, 2011) and recurring transgressions have provided, at least temporally, favorable conditions for tidal modulation of deposits. Which in turn could be masked by wave- and storm action (Davis, 2012), or may simply be overlooked at outcrops.

A westward thinning of distributary channel deposits and the mentioned lack of interpreted mouth bars in Agardhbukta, should also be noted. These observations may point toward westwards decreasing accommodation space and is consistent with Rød et al. (2014).

Barrier bars (FA 2.1) comprise shoreface to foreshore sands and is distinguished from the residual delta front deposits by a profound impact of wave processes upon sedimentation (Chapter 4). These deposits are found troughout the study area. Nevertheless, their abundance varies by occurring at a stratigraphically lower position in the De Geerdalen Formation and geographically being more abundant towards west (except on Teistberget).

In Agardhbukta barrier bars terminate coarsening upwards successions (parasequences) in the lower part, while they enclose floodplain and interdistributary bay sediments (FA 3.1 and 3.2) in the upper part of measured sections. Interestingly an upwards decreasing content of mud drapes (facies K) in the barrier sandstones (FA 2.1) is observed at these localities, and may imply a decreasing tidal component upon sedimentation (Dalrymple and Choi, 2007). On Wilhelmøya, thin sheets of barrier sandstones (FA 2.1) overlie scree covered intervals and is the dominant delta front association.

At more proximal localities such as Svartnosa and Hellwaldfjellet, barrier sandstones (FA 2.1) are found above sand rich intervals dominated by deltaically introduced sandstones associated with a prograding shoreline (FA 2.2 and 2.3). The interpreted shift from dominating fluvial processes to increasing basinal processes may be related to decreasing accommodation space upwards, resulting in enhanced reworking, or abandonment and reworking during the deltas destructive phase (Boyd and Penland, 1988). Alternatively the interpreted barrier (FA 2.1) successions could represent prograding wave-influenced delta lobes (Bhattacharya and Giosan, 2003), as prograding wave-influenced deltas and shoreface successions show identical vertical facies successions in the geological record (Bhattacharya, 2006).

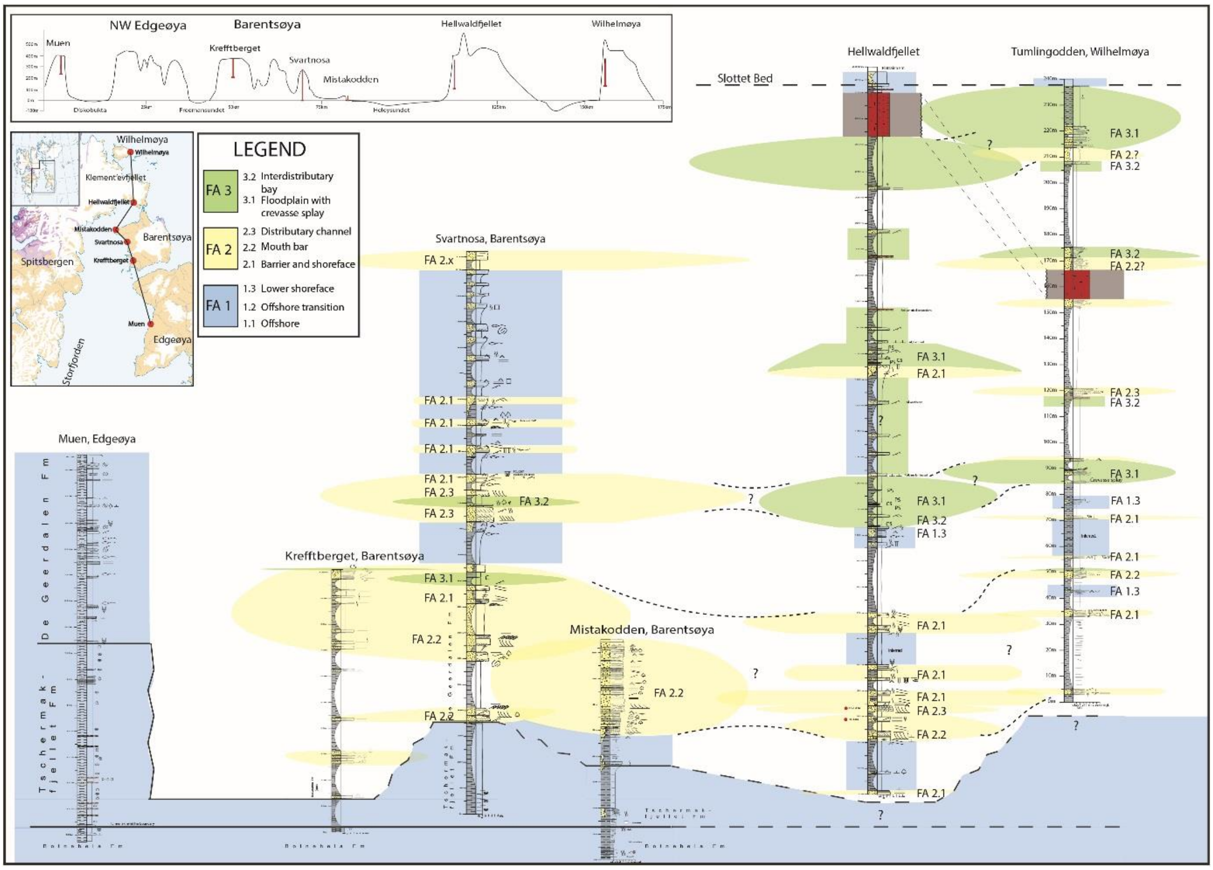


Figure 6.2: Eastern correlation panel using the top Botneheia and inferred base De Geerdalen as correlative surfaces. Topographic cross-section and log trace (red) in upper left corner.

Delta top sediments (FA 3) dominate the upper part of measured sections at all localities, except on Barentsøya. Their occurrence is abundant in, though not restricted to, the Isfjorden Member.

As mentioned in Chapter 4 sub-environments located at the delta top include floodplain and crevasse splay deposits (FA 3.1), swamps, marshes, interdistributary bays, tidal flats and lagoons (FA 3.2). Typically sediments from interdistributary areas (FA 3.2) occur stratigraphically lower in the De Geerdalen Formation than floodplain and crevasse splay deposits (FA 3.1). This trend is largely congruent with the regressive nature of deltaic deposits as the floodplain deposits usually is considered more proximal than most interdistributary deposits (e.g. lagoon, interdistributary bay).

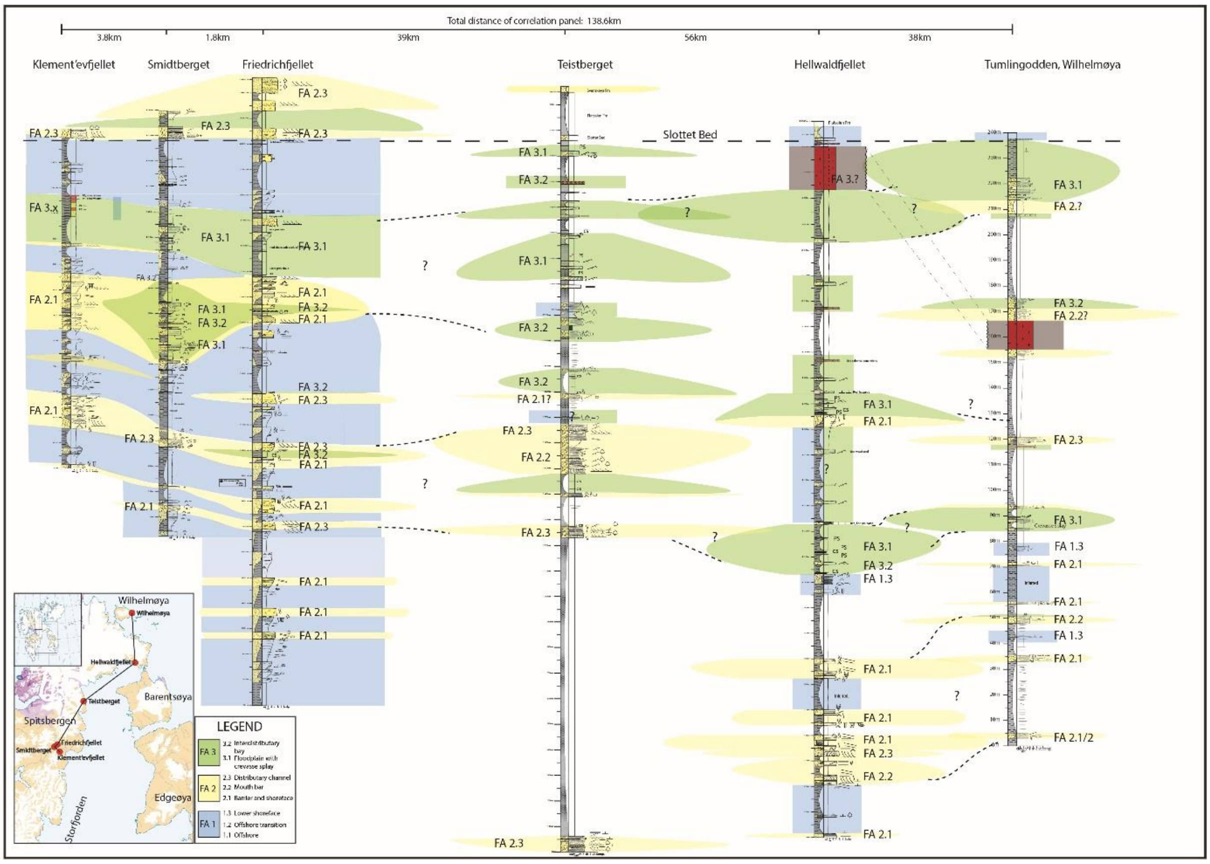


Figure 6.3: Western correlation panel using the Slottet Bed as datum.

Floodplain deposits (FA 3.1) are found lower in the stratigraphy at Hellwaldfjellet and Wilhelmøya than at localities further south. This may point to an earlier establishment of the paralic, delta plain environment towards northeast and could indicate a potential delta lobe progradation from the northeast.

6.3 Depositional environment

The idea that varying degree of fluvial, tide and wave processes result in characteristic plan view morphology and facies successions of delta deposits is widely recognized. Accordingly deltas have been classified through processes-based tripartite classification schemes (e.g. Galloway, 1975), with later modifications incorporating sediment type (Orton and Reading, 1993) and changes in sediment supply (Dalrymple et al., 1992). However, as deltas exhibit complex facies architecture and considerable variations along depositional strike (Bhattacharya, 2006), such classifications should be treated with care. Additionaly most modern and ancient deltas show a mixture of influencing processes.

Considering the scale of the deltaic system represented by the De Geerdalen Formation, the complexity of interpreted facies successions and the fact that this study is based solely on 1D-data (hence lack of plan-view morphological features) such a classification will not be attempted. However, the relative influence of modulating processes is noticed to vary among localities and will be used to infer about the paleoshoreline.

The localities on Barentsøya are inferred to represent the most proximal localities in the study area. Delta front deposits in the lower parts show prevalent fluvial signals. Great thicknesses of sandstone bodies and limited reworking by basinal processes are thought to reflect adequate accommodation space or rapid subsidence. Additionally occurrence of climbing ripple laminated sandstones and soft sediment deformation structures (facies C and I_1) are associated with rapid deposition (see Chapter 3). Similar delta front deposits are reported from Edgeøya (Rød et al., 2014), although these deposits may be considered even more proximal than those of Barentsøya due to greater thicknesses and overriding fluvial signals. As mentioned in the introduction trunk river channels are reported from Hopen (Klausen and Mørk, 2014; Lord et al., 2014b). The sandstone body previously described as a distributary channel on Svartnosa is laterally extensive over aproximatly 1 km as seen in the field, with a recorded thickness of 6 m. Whether the entire sandstone body represent a channel deposit or other sandy delta front facies associations is not certain, as deltas commonly show lateral variability in influencing processes along depositional strike (Bhattacharya, 2006). Regardless, the observed width dimensions are comparable to channelized sandstones reported from Hopen (Lord et al., 2014b). Due to the modest thicknesses and deltaic signals, as opposed to thicker and incising fluvial channels, the deposits on Svartnosa may be considered distal distributary channels to the trunk river channels observed on Hopen. However, the deposits on Hopen are younger (Riis et al., 2008; Lord et al., 2014a; Lundschien et al., 2014) and represent the most regressive stage of the De Geerdalen

Formation on Svalbard (Lord et al., 2014a). Nevertheless, the lower part of the Svartnosa section represent the most proximal delta front deposits in the study area, and an overall fluvial dominance is interpreted for the lower parts of the De Geerdalen Formation at Barentsøya. The trend of progressively more modest channel dimensions towards the northeast, observed within the study area and in regards to Edgeøya (Rød et al., 2014) and Hopen (Klausen and Mørk, 2014; Lord et al., 2014), have also been reported from the Snadd Formation of the Barents Sea (Klausen et al., 2014).

An increasing influence of wave processes is interpreted northwards to Wilhelmøya in the lower part of sections. Here, delta front sands are commonly wave modulated (FA 2.1) and show modest thicknesses. The corresponding interval at Hellwaldfjellet contain more fluvially influenced delta front sands than Wilhelmøya, but a thinning of this sandy succession is noticed compared to Svartnosa. A general westward direction of progradation is inferred from paleocurrent measurements on western Edgeøya (Knarud, 1980; Riis et al., 2008; Hynne; 2010; Høy and Lundschien, 2011; Rød et al., 2014), and in consequence may put the northern localities off-axis to the distributary systems described on Barentsøya and Edgeøya. Recent work by Klausen et al. (2016) documents regressive shoreface deposits in the Snadd Formation (southwestern Barents Sea), implying that the Carnian Barents Sea basin was at times wave-dominated. Tentatively a distal paleoposition to the main distributaries in the south, accompanied by wave action and longshore currents from the northeast could have influenced the northern deposits. Alternatively, the barrier sandstones may also represent the uppermost part of progradational delta front deposits that were reworked during the destructional phase (e.g. Boyd and Penland, 1988). Due to abundant scree cover on the northern localities dominating fluvial signals below barrier sandstones could simply be overlooked.

Delta front deposits at Teistberget are interpreted to be fluvially dominated, and thus contrast to the northern localities discussed above. Teistberget is also located 36 km west of the comparable fluvially dominated delta front deposits on Svartnosa. As the lower boundary of the De Geerdalen Formations is distinctly diachronous (Mørk et al., 1999; Høy and Lundschien, 2011; Lundschien et al., 2014) it is not suggested that these sandstones represents distal equivalents. However, given the general westwards paleoflow directions of the deltaic system and both localities fluvially dominated nature, the localities could tentatively represent a distributary system prograding westwards through the study area. Also note that a large scree covered interval in the lower part of the Teistberget section and the presence of the Storfjorden Fault Zone (Eiken, 1985) may complicate this correlation.

In turn, the fluvial dominance interpreted on Teistberget contrast to abundant barrier sandstones (FA 2.1) and heterolithic nature of deposits in the Agardhbukta sections. As previously mentioned, Rød et al. (2014) suggested restricted accommodation space and wave modification as responsible mechanisms for generating similar deposits on central Spitsbergen. Models based on modern prograding asymmetric wave-influenced deltas predicts significant river-muds in prodelta and downdrift areas, and shoreface sandstones in updrift areas (Bhattacharya and Giosan, 2003). Provided temporal longshore currents and waves from the northeast (as tentatively suggested by Klausen et al., 2016), one can speculate upon if some intervals of the Agardhbukta sections may be on the downdrift side of an asymmetric wave-dominated delta system prograding towards Teistberget. Consequently, the northern localities would be on the updrift side, of that particular distributary system, and experience wave dominance upon sedimentation. Tentatively the inferred wave influence could vary both temporal and spatial, meaning that some intervals in the De Geerdalen Formation deposits could represent a wave influenced delta, and others not.

Within the discussed paleogeographic framework, the fluvially dominated localities on Barentsøya and Teistberget would represent areas proximal to the distributary part of a westwards prograding system. Whereas localities further south and north would be positioned off-axis to the distributary system, and consequently influenced more by basinal processes. Speculatively such a paleogeographic setting would generate more wave action at the northern localities, while the southern localities may have been somewhat protected by the protruding shoreline located in the middle parts of the study area.

The development of a paralic delta plain environment is noted in the upper part of sections, and implies a shift into fluvial dominance upon sedimentation throughout the study area. This interpretation may be comparable to the more proximal deposits on Hopen (Klausen and Mørk, 2014; Lord et al., 2014b) and the Snadd Formation of the Barents Sea (Klausen et al., 2014) where a tidally influenced fluvially dominated delta plain environment is suggested. Shale is the dominant lithology in the De Geerdalen Formation on central Spitsbergen with increasing sand content towards east and northeast (Mørk et al., 1982). The high mud content characterizing the De Geerdalen Formation (Mørk et al., 1982; Hynne, 2010) and growth-faulting on Edgeøya (Edwards, 1976; Osmundsen et al., 2014) are additional features associated with a fluvially dominated delta system (Reading and Collinson, 1996).

Through facies analysis an attempt to infer about the processes regime imposing controls upon sedimentation in the De Geerdalen Formation has been attempted, and largely show a mixed influence of processes between and within localities. It is also noted that the overall deltaic system changes character with time, developing into a paralic delta plain environment.

6.4 Sediment properties

The depositional environment imply controls upon sedimentary textures and mineralogical composition (e.g. Bloch and McGowen, 1994). However, the use of textural analysis as a tool for environmental analysis have shown to produce elusive results (Boggs, 2009). Accordingly the observations made regarding sorting, grain size and rounding of clasts will be briefly discussed below, and only used to complement the sedimentological discussion.

In the distributary mouth bar deposits (FA 2.2) fluvially introduced sediments have been modulated into wave-generated structures and wave-induced current-generated structures (Chapter 4). Despite this overprinting of wave-structures the sediments seems to remain relatively immature in regards to texture. This may point toward rapid deposition at the delta front as the deposits were not subjected to wave and current activity long enough to change the maturity, consistent with interpretations in Chapter 4. Furthermore, due to abundant lithic fragments and feldspar the sandstones also appear mineralogically immature, consistent with previous studies (Bergan and Knarud, 1993; Mørk, 1999, 2013; Riis et al., 2008). Tentatively the high content, and occasionally observed angularity, of lithic clasts support the inferred limited reworking of sediments as these clasts may be prone to abrasion in high energy environments such as beaches.

Grain-size distribution reflect processes, not environments (Boggs, 2011). In river systems, variance in energy conditions and sediment supply may vary, both spatial and temporal, and accordingly impose controls upon sediment texture (Boggs, 2011). Variance in sorting observed in distributary channel sandstones (FA 2.3) may possibly reflect that dynamic nature. Crevasse sandstones (FA 3.1) generally appear well sorted, a feature that may relate to that finer grained sediments tend to be better sorted (Boggs, 2011).

The studied delta front sandstones are comprised of very fine to medium grained sand. A slight decreasing grain-size trend was noted from Barentsøya and Teistberget towards localities in the north and south. Mean grain size is believed to be a function of the size range of available material and energy of current that transport sediment (Folk, 1974). In general sediment tend to be finer in the direction of transport. As the available sediment is thought to be largely the same (i.e. Uralide source as mentioned in Chapter 1), the slight discrepancies could possibly reflect a distal position to the distributary systems. As sediment textures mainly is controlled by the sedimentary environment (Bloch and McGowen, 1994), the slight discrepancies may also reflect that samples were obtained from multiple delta front facies associations.

6.5 Diagenesis

Diagenesis comprise a broad spectrum of processes that act to modify sediments after deposition (e.g. Worden and Burley, 2003; Boggs, 2009). Diagenetic processes recognized in the sandstones of the De Geerdalen Formation include mechanical compaction, grain alterations, dissolution and cementation.

Porosity in the studied delta front sandstones seems largely to be controlled by degree and timing of calcite cementation as well as dissolution of feldspars and lithic fragments. The porosity values obtained from visual comparison range from 3 to 25 %, and includes both intergranular and intragranular porosity. The latter commonly dominates, and makes a solitary contribution to porosity in the extensively calcite cemented sandstones. Kaolinite is seen both within secondary porosity and in intergranular pores, and accordingly a component of microporosity should also be expected.

Lithic sandstones are exposed to mechanical compaction (Pittmann and Larese, 1991), and a resultant densely packed fabric has previously been reported in the De Geerdalen Formation (Mørk et al., 1982; Mørk, 2013). Early calcite cementation is often seen in the delta front sandstones and is thought to have inhibited mechanical compaction during mesogenesis. Thus, allowing preservation of a relatively open sandstone fabric, in which high porosity values are obtained where the calcite cement shows a patchy appearance. Mørk et al. (1982) suggested shell fragments and bioclasts as potential sources for the heterogenous calcite cement. Alternatively, dissolution processes may explain the phenomena (e.g. Morad, 1998; Boggs, 2009).

Dissolution processes are chiefly related to feldspars and lithic fragments, and account for the majority of visual porosity in the delta front sandstones. A warm humid climate (Hochuli and Vigran, 2010; Enga, 2015) and fluvial transport may have provided favorable conditions for chemical weathering. Bergan and Knarud (1993) attributed immaturity of sediments and minor mineral alterations in the sedimentary environment to rapid deposition and limited reworking, concordant with observations in this study. Repeated coastal progradations and tectonic uplift are typical conditions associated with telogenetic influx of meteoric water (Worden and Burley, 2003). Repeated deltaic progradations and development of a paralic delta plain reflect conditions associated with the former, whereas the latter may have occurred during Jurassic reworking (Mørk et al., 1982; Bergan and Knarud, 1993; Mørk, 2013) or even at modern outcrop exposure. Mørk (2013) also noted that dissolution porosity is commonly associated with authigenic chlorite, illite and albite. These authigenic minerals form at elevated temperatures (Bjørlykke, 1997; Wilkinson et al., 2001, Worden and Burley, 2003) with decomposing lithic fragments thought to have provided a local source of iron. Accordingly a multi-stage formation of dissolution porosity with leaching both during telogenesis and later at elevated temperatures, as suggested by Mørk (2013), seems plausible.

The detrital composition of sediments has implied controls upon diagenesis, as shown by the dissolution processes discussed above. Observations also point to slight variations in diagenesis as a function of depositional setting, with calcite cementation being more abundant in distributary channel and crevasse sandstones (FA 2.3 and 3.1) than mouth bar sediments (FA 2.2). Siderite appear to be more abundant in crevasse sandstones (FA 3.1), especially in the upper parts of the De Geerdalen Formation, and may reflect the upwards shift into a delta plain dominated depositional style.

Potential mechanisms for eogenetic calcite precipitation were mentioned in Chapter 5 and include; precipitation below the sediment surface in a shallow marine environment (e.g. Morad, 1998; Ketzer et al., 2003), calcite precipitation at base of distributary channels (e.g. Morad et al., 2010) and biochemical precipitation in association with cyanobacteria (Tugarova and Fadyaevsky, 2014). Textural observations in many sandstones and “calcite cemented sandstone” of facies J support an early diagenetic origin. Calcite cement detected within fractured feldspar and enclosing quartz cement points toward multiple phases of calcite cementation. The late calcite cement could be explained by redistribution processes (e.g. Morad, 1998; Boggs, 2009) of the eogenetic calcite in addition to coquina beds in the upper part of the formation. Other possible internal sources of calcium may be replacement of detrital feldspars (e.g. Morad, 1998) and alteration of Ca-plagioclase (e.g. Schulz et al., 1989). The calcium contribution from the latter is thought to be limited as SEM investigations mainly show albitization of detrital K-feldspar, and the mechanism itself is thought only to account for small portions of calcite cement in sandstone sequences (Morad, 1998).

The De Geerdalen Formation has previously been described as a closed isochemical system due to substantial compaction (Mørk et al., 1982). Tentatively, the heterolithic nature of the formation could have put additional restrictions on pore water flow. As discussed above cementation processes (of e.g. quartz and clay minerals) seems to largely be associated with favorable conditions found within microenvironments close to labile fragments, concordant with Mørk (2013). Thus pointing to that the detrital framework have implied an increasing control upon diagenesis during burial.

7 Conclusions

This thesis documents and discusses sedimentology and diagenesis in the Triassic De Geerdalen Formation on eastern Svalbard, based on logging of outcrops and petrographic observations. The key conclusions are summarized below.

- The De Geerdalen Formation consists of recurring small-scale upwards coarsening successions that are arranged within a major shallowing upwards sequence. The former is thought to reflect a dynamic deltaic nature (e.g. lobe switching), whereas the latter has been attributed to progradation of the overall deltaic system. The establishment of a paralic delta plain setting is seen throughout the study area. The Isfjorden Member was recognized at all localities visited, except on Barentsøya and Edgeøya, where the upper part of the De Geerdalen Formation has been removed by Cenozoic erosion.
- Distribution of facies and facies associations point toward a decreasing fluvial influence upon sedimentation distal to the fluvially dominated localities on Barentsøya and Teistberget. Towards southwest (Agardhbukta) and north (Hellwaldfjellet and Wilhelmøya) delta front sandstones are thinner and modulated by basinal processes in the lower part of the De Geerdalen Formation. The increased influence of basinal processes may reflect a more distal position to the overall deltaic distributary system.
- Mineralogical immaturity and textural properties of delta front sediments indicate limited reworking in the depositional environment. A slight decrease in mean grain-size is noticed from the fluvially dominated localities (Svartnosa and Teistberget) to the remaining localities. This trend could reflect decreasing energy distal to the distributary systems, concordant with interpretations from facies analysis. Alternatively, the trend could simply be related to samples being acquired from multiple facies associations.
- Distribution of porosity is largely controlled by diagenesis. Precipitation of calcite cement is regarded as the main porosity reducing factor.
- Early calcite cement has arrested mechanical compaction in some sandstones. Thus allowing an open sandstone fabric, with preserved intergranular porosity where cementation shows a patchy distribution. Multiple phases of formation and occurrence in all facies has been documented, and is interpreted to reflect multiple sources of calcium carbonate for calcite precipitation.

- Dissolution processes are considered the main mechanism for enhancing porosity, and is attributed to the immature composition of the sandstones. The resultant secondary porosity often occurs as molds, capsuled in calcite cement, and is poorly connected. Furthermore, dissolution of unstable fragments provided favorable conditions for precipitation of authigenic mineral phases (e.g. quartz and clay minerals) in microenvironments with a local source of ions.

7.1 Suggestions for further work

- Polycrystalline quartz fragments have received little attention in this study. Detailed investigations may provide insight in regards to a metamorphic source area with implications for provenance.
- For future provenance studies the detrital composition of sandstones should be quantified and compared to previous studies on Svalbard (Riis et al., 2008, Mørk, 2013) and the Barents Sea (Mørk, 1999). As seen in this study, diagenesis may alter the detrital composition, with implications for the obtained bulk modal composition of sandstones.
- Outcrop exposure on the Svartnosa locality is excellent. Future investigations could collect lateral data in order to better understand the spatial and temporal evolution of delta front sandstones in the De Geerdalen Formation.

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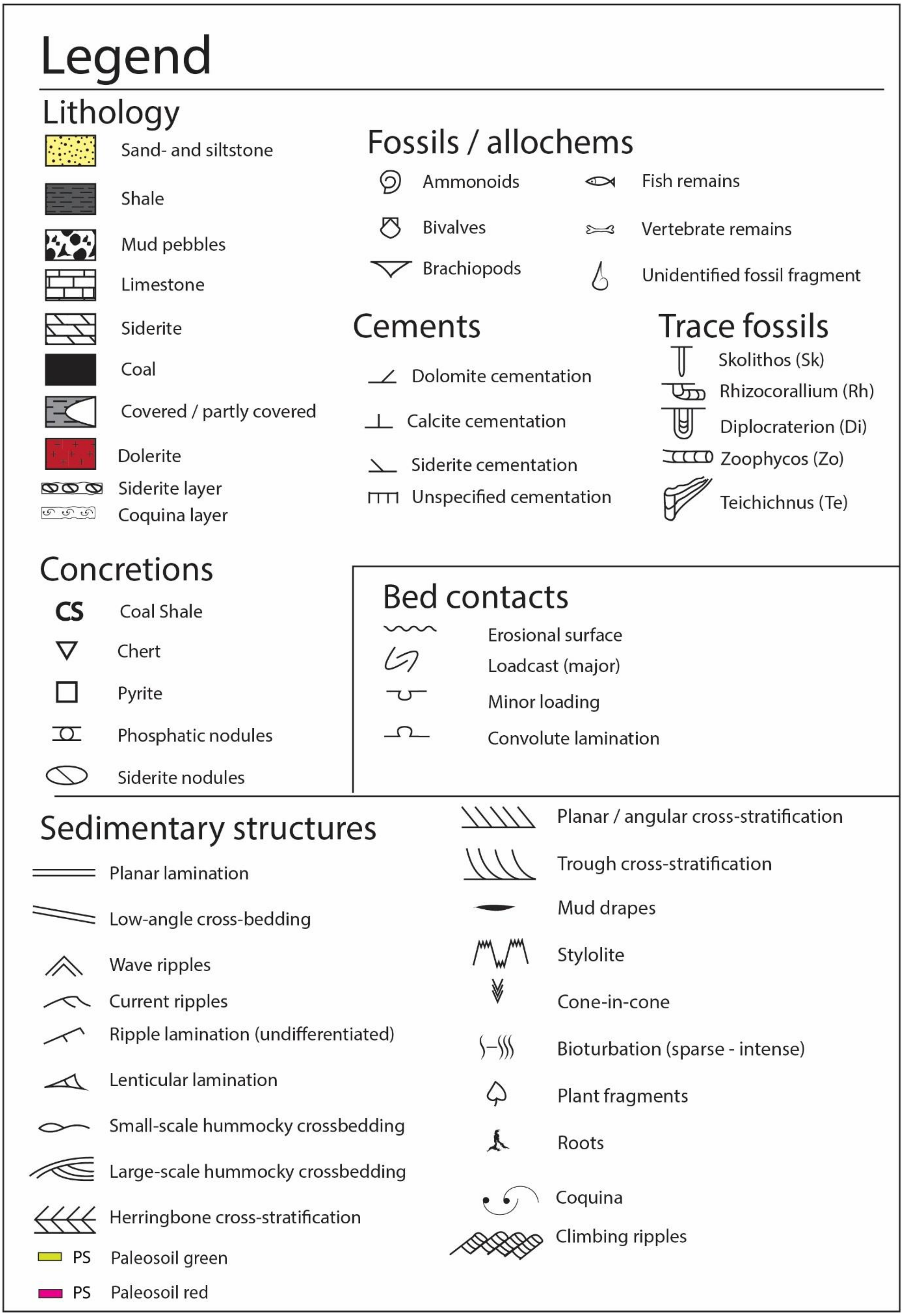
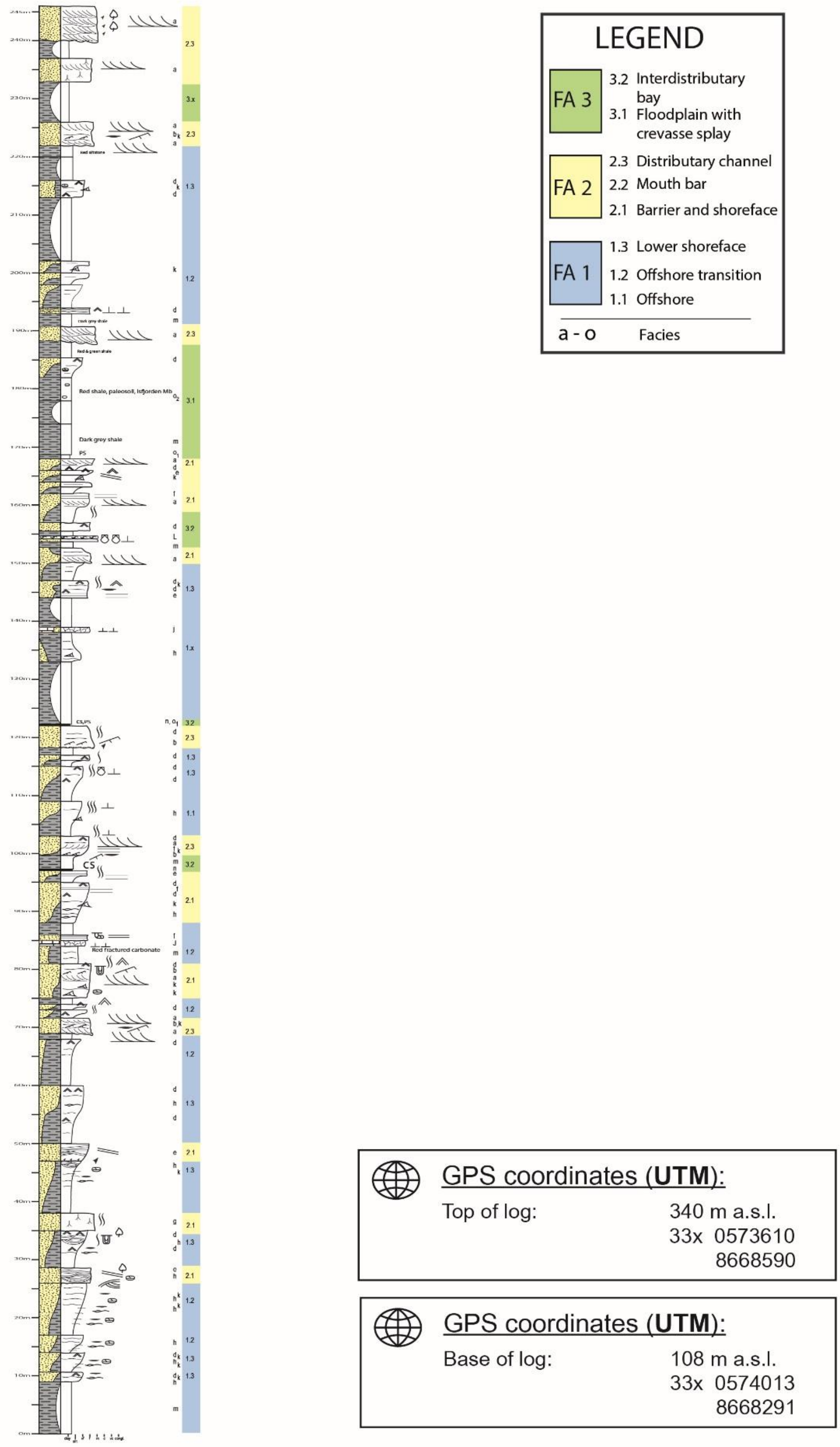
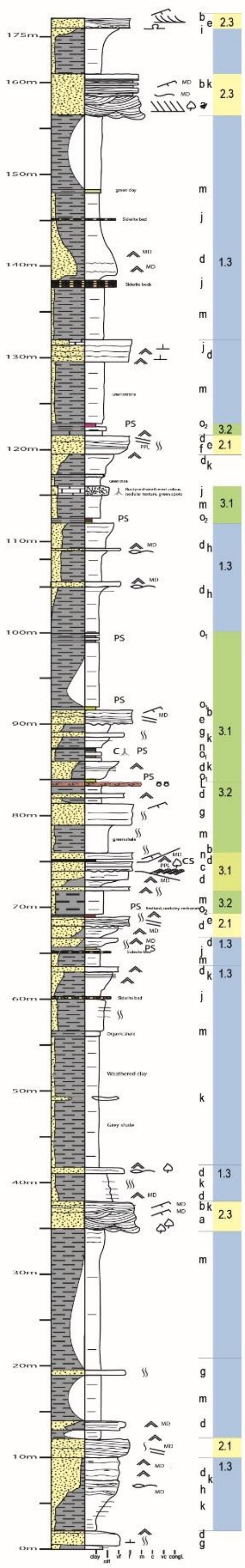
Appendix A


Legend
Lithology
Sand- and siltstone
Shale
Mud pebbles
Limestone
Siderite
Coal
Covered / partly covered
Dolerite
Siderite layer
Coquina layer
Fossils / allochems
Ammonoids
Bivalves
Brachiopods
Fish remains
Vertebrate remains
Unidentified fossil fragment
Cements
Dolomite cementation
Calcite cementation
Siderite cementation
Unspecified cementation
Trace fossils
Skolithos (Sk)
Rhizocorallium (Rh)
Diplocraterion (Di)
Zoophycos (Zo)
Teichichnus (Te)
Concretions
CS
Coal Shale
Chert
Pyrite
Phosphatic nodules
Siderite nodules
Bed contacts
Erosional surface
Loadcast (major)
Minor loading
Convolute lamination
Sedimentary structures
Planar lamination
Low-angle cross-bedding
Wave ripples
Current ripples
Ripple lamination (undifferentiated)
Lenticular lamination
Small-scale hummocky crossbedding
Large-scale hummocky crossbedding
Herringbone cross-stratification
PS
Paleosoil green
PS
Paleosoil red
Planar / angular cross-stratification
Trough cross-stratification
Mud drapes
Stylolite
Cone-in-cone
Bioturbation (sparse - intense)
Plant fragments
Roots
Coquina
Climbing ripples

Friedrichfjellet, Agardhbukta



Smidtberget, Agardhbukta



LEGEND

FA 3	3.2	Interdistributary bay
	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	Lower shoreface
	1.2	Offshore transition
	1.1	Offshore
a - o		Facies

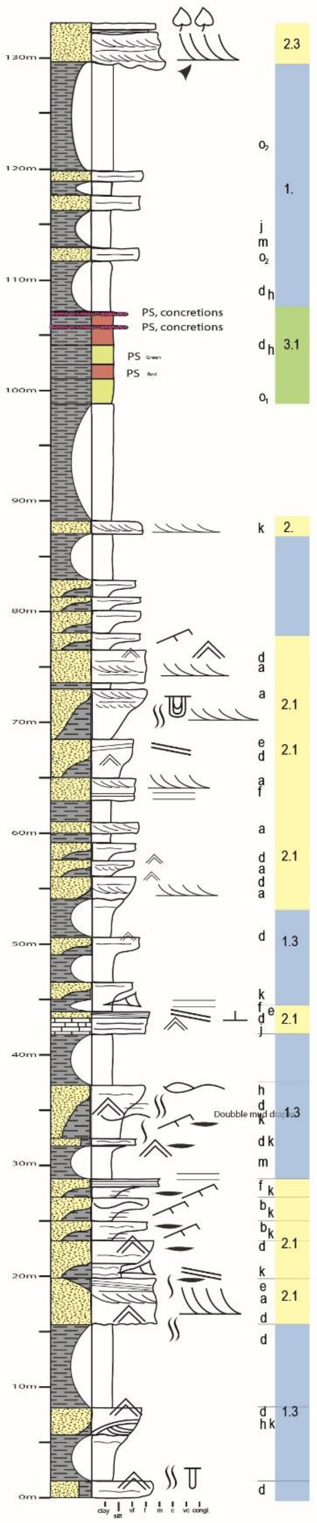
GPS coordinates (UTM):

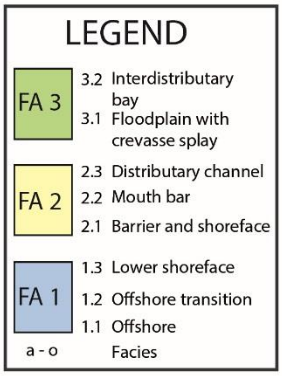
Top of log: 287 m a.s.l.
33x 0372797
8667030

GPS coordinates (UTM):

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

Klement'evfjellet, Agardhbukta





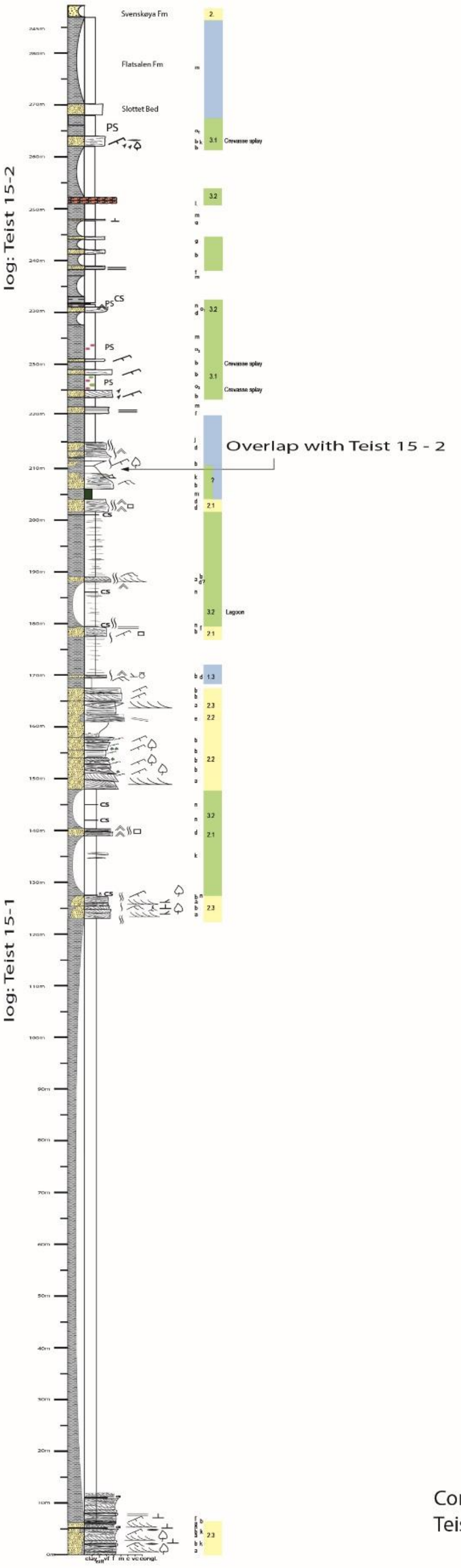
GPS coordinates (UTM):

Top of log	329 m a.s.l.
	33x 0575899
	8665577

GPS coordinates (UTM):

Base of log:	215 m a.s.l.
	33x 0575801
	8666054

Teistberget, Spitsbergen



LEGEND

FA 3	3.2	Interdistributary bay
	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	Lower shoreface
	1.2	Offshore transition
	1.1	Offshore
a - o		Facies

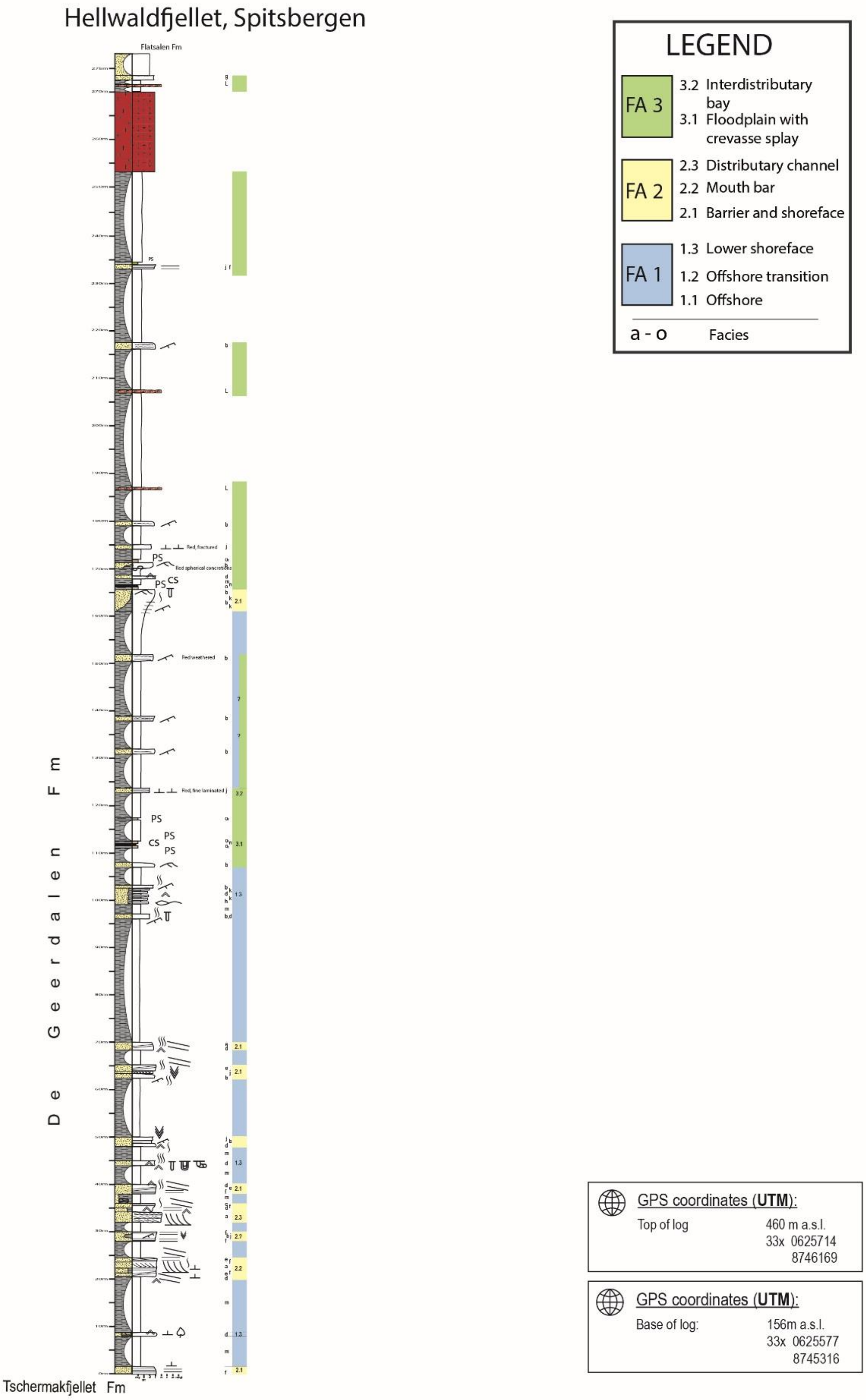
GPS coordinates (UTM):

Top of log: 340m a.s.l.
33x 0590626
8702312

GPS coordinates (UTM):

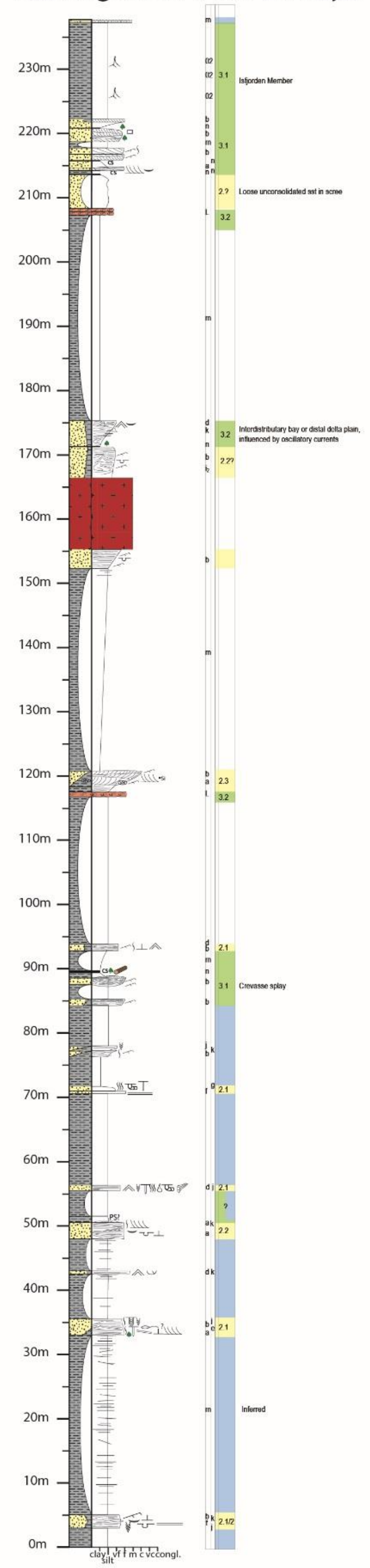
Base of log: 0 m a.s.l.
No GPS
point from base

Composite log from Sondre Krogh Johansen
Teist 15-1 and Turid Haugen, Teist 15-2



Hellwaldfjellet, Spitsbergen
Flatsalen Fm
De Geerdalen Fm
Tschermakfjellet Fm
PS
CS
LEGEND
FA 3
3.2 Interdistributary bay
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 Lower shoreface
1.2 Offshore transition
1.1 Offshore
a - o Facies
GPS coordinates (UTM):
Top of log
460 m a.s.l.
33x 0625714
8746169
GPS coordinates (UTM):
Base of log:
156m a.s.l.
33x 0625577
8745316

Tumlingodden, Wilhelmøya



LEGEND

FA 3	3.2	Interdistributary bay
	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	Lower shoreface
	1.2	Offshore transition
	1.1	Offshore
a - o		Facies

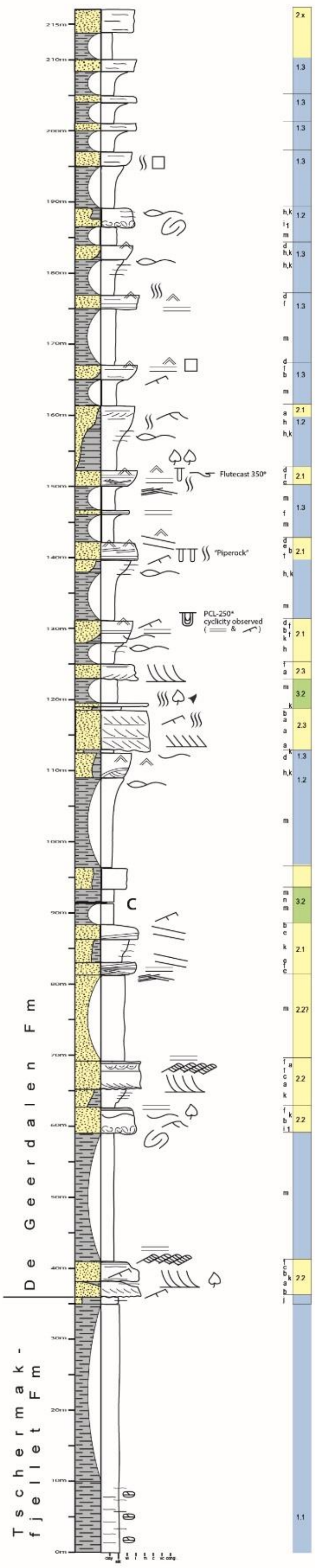
GPS coordinates (UTM):

Top of log: 485 m a.s.l.
33x 0621192
8782969

GPS coordinates (UTM):

Base of log: 142 m a.s.l.
33x 0622072
8782337

Svartnosa, Barentsøya



LEGEND

FA 3
3.2 Interdistributary bay
3.1 Floodplain with crevasse splay

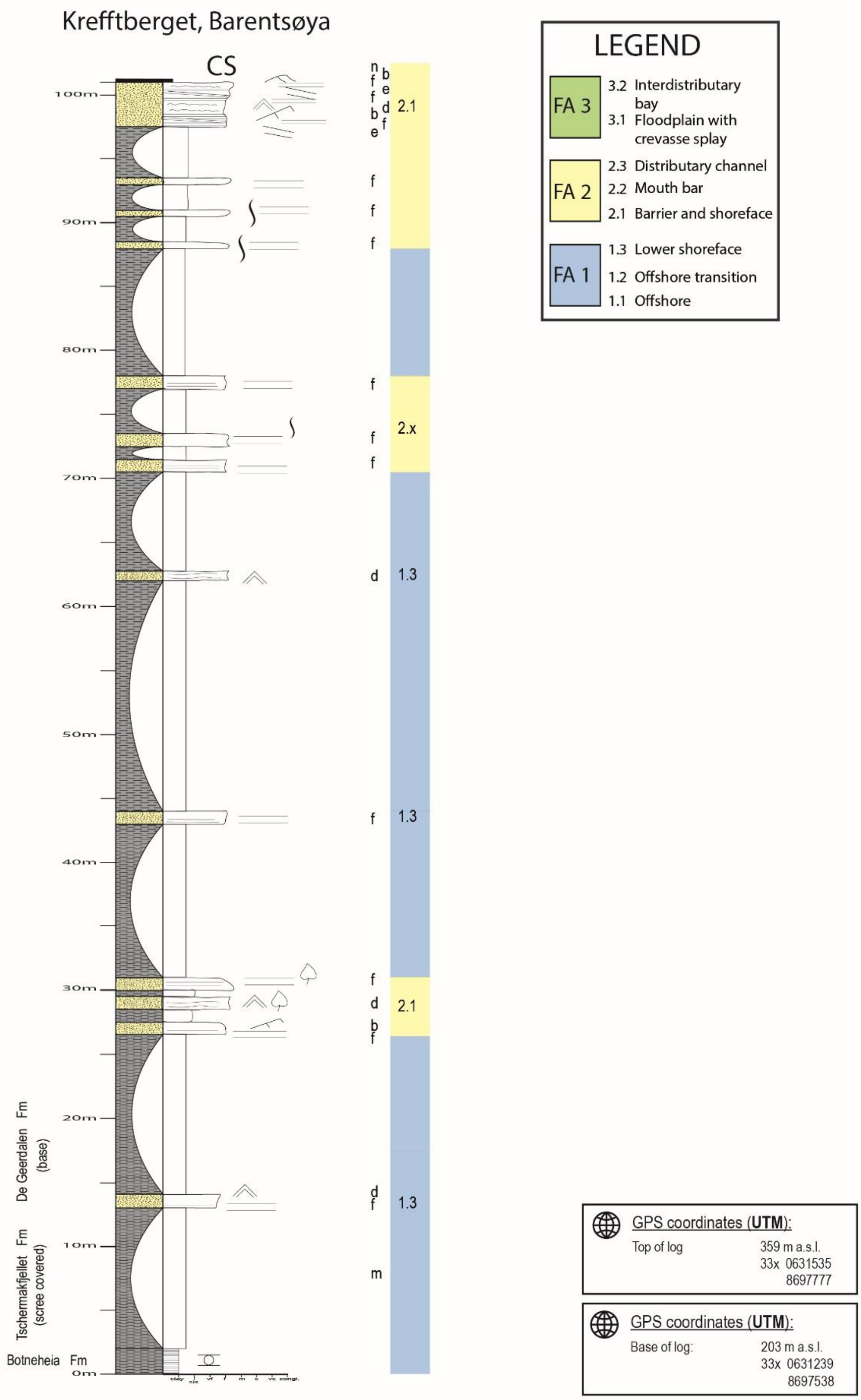
FA 2
2.3 Distributary channel
2.2 Mouth bar
2.1 Barrier and shoreface

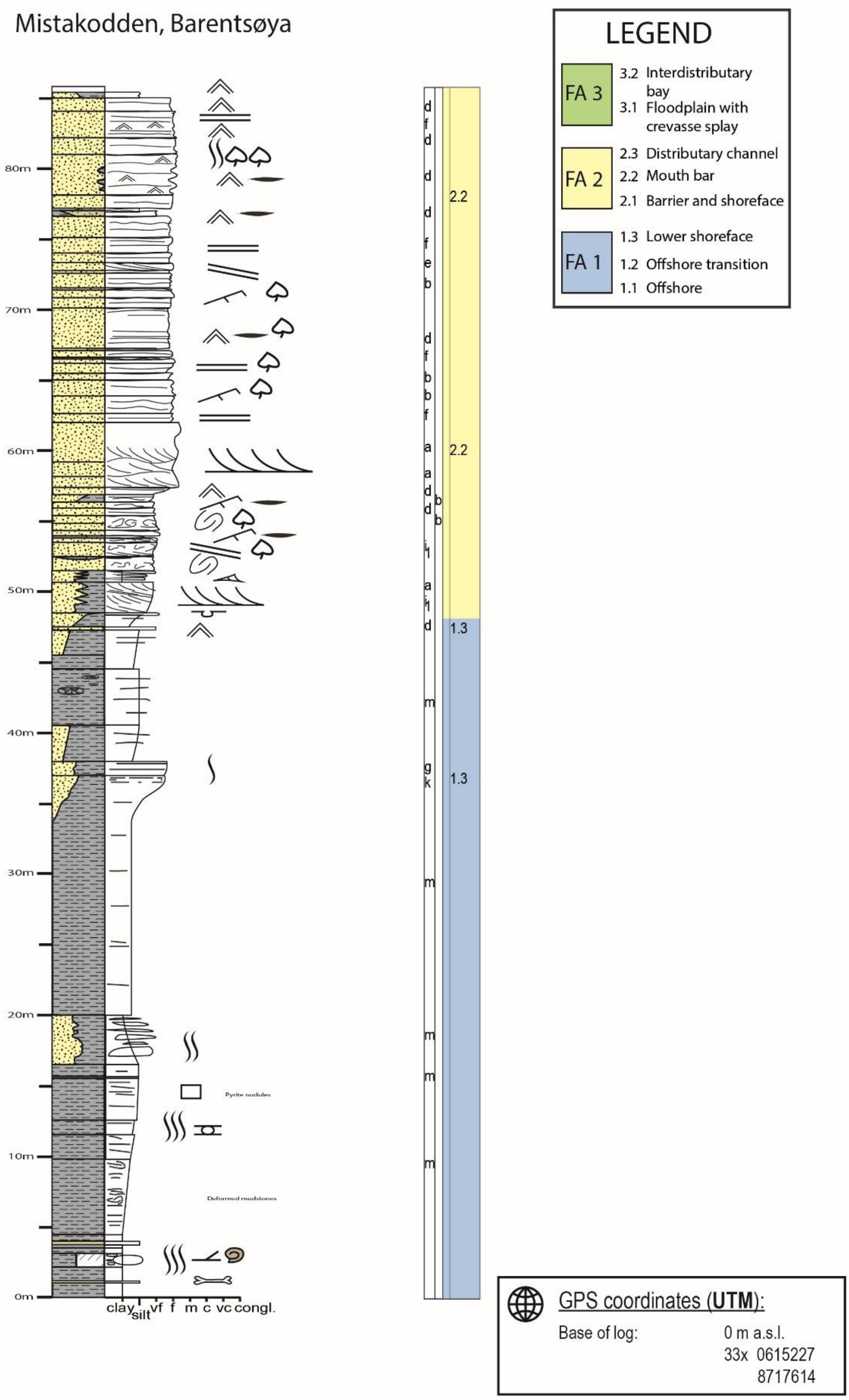
FA 1
1.3 Lower shoreface
1.2 Offshore transition
1.1 Offshore

a - o Facies

GPS coordinates (**UTM**):
Top of log 253 m a.s.l.
33x 0625663
8711925

GPS coordinates (**UTM**):
Base De Geerdalen FM: 34 m a.s.l.
33x 0625356
8711779


Krefftberget, Barentsøya
CS
100m
90m
80m
70m
60m
50m
40m
30m
20m
10m
0m
De Geerdalen Fm (base)
Tschermakfjellet Fm (scree covered)
Botneheia Fm
2.1
2.x
1.3
1.3
2.1
1.3
LEGEND
FA 3
3.2 Interdistributary bay
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 Lower shoreface
1.2 Offshore transition
1.1 Offshore
GPS coordinates (UTM):
Top of log
359 m a.s.l.
33x 0631535
8697777
GPS coordinates (UTM):
Base of log:
203 m a.s.l.
33x 0631239
8697538



Mistakodden, Barentsøya
LEGEND
FA 3
3.2 Interdistributary bay
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 Lower shoreface
1.2 Offshore transition
1.1 Offshore
80m
70m
60m
50m
40m
30m
20m
10m
0m
clay silt vf f m c vc congl.
Pyrite nodules
Deformed mudstones
2.2
2.2
1.3
1.3
GPS coordinates (UTM):
Base of log:
0 m a.s.l.
33x 0615227
8717614

Appendix B

ID	Location	Altitude in log (m)	FA	F	Grain Size	Sorting	Estimated porosity %
TE-2-4A	Teistberget	224	3.1	B	155	Moderately sorted	20
TE-1-16A	Teistberget	165	2.3	A	275	Well sorted	4
TE-1-15A	Teistberget	164	2.3	A	155	Moderately sorted	15
TE-1-14A	Teistberget	162	2.2	E	310	Moderately sorted	8
TE-1-5A	Teistberget	123	2.3	A	150	Moderately sorted	2
TE-1-3A	Teistberget	4	2.3	A	275	Poorly sorted	10
SV15-16A	Svartnosa	84	3.2	G	220	Moderately sorted	15
SV15-13A	Svartnosa	81	2.3	A	230	Well sorted	10
SV15-11A	Svartnosa	78	2.3	A	220	Moderately sorted	15
HE-1-40A	Hellwaldfjellet	234	3.1	B	150	Moderately sorted	5
HE-1-38A	Hellwaldfjellet	217	3.1	B	155	Well sorted	5
HE-1-36A	Hellwaldfjellet	187	3.1	L	-	-	-
HE-1-13A	Hellwaldfjellet	34	2.3	F	310	Well sorted	10
HE-1-11A	Hellwaldfjellet	32	2.3	A	230	Moderately sorted	20
HE-1-8A	Hellwaldfjellet	28	2.2	F	160	Moderately sorted	12
KL-4-10B	Klement'evfjellet	130	2.3	A	230	Well sorted	8
KL-4-5B	Klement'evfjellet	43	2.1	J	-	-	-
SM-2-17A	Smidtberget	157	2.3	A	180	Poor - moderate	8
SM-2-16A	Smidtberget	132	1.2	J	-	-	-
SM-2-11A	Smidtberget	115	3.1	J	-	-	-
SM-2-1A	Smidtberget	36	2.3	A	230	Moderately sorted	3
MK-1-15A	Mistakodden	82	2.2	D	200	Moderately sorted	13
MK-1-13A	Mistakodden	71	2.2	B	180	Moderately sorted	15
TUM-2-9A	Wilhelmøya	217	3.1	B	230	Well sorted	10
TUM-2-6A	Wilhelmøya	208	2.x	G	155	Well sorted	5
WI-2-19.3A	Wilhelmøya	120	2.3	A	230	Moderately sorted	25
WI-2-19.1A	Wilhelmøya	118	2.3	A	222	Well sorted	20