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Sedimentology and facies distribution of the Upper Triassic De Geerdalen Formation in the Storfjorden area and Wilhelmøya, eastern Svalbard

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Abstract

This study investigates sediments of the Upper Triassic De Geerdalen Formation, as it is exposed on eastern Svalbard. The De Geerdalen Formation was deposited during the Early Carnian to Early Norian, on the distal area of a shallow embayment on the northern margin of Pangea. It includes a wide range of depositional environments, ranging from offshore shallow marine, paralic delta front and shoreface to coastal plain environments.

The aim of the study is to document and interpret the sedimentology of the formation and the spatial and temporal evolution of depositional environments, through detailed facies analysis. Fifteen facies and eight facies associations have been defined based on observations and outcrop data. These data were collected during a one-month field season in 2015 on eastern Spitsbergen, western Barentsøya and on Wilhelmøya. The interpretation presented here is based on previous research in neighbouring areas, as well as cooperation with other master students.

From the measured sections, a strong fluvial dominance is observed on the northern part of Barentsøya and interpreted to reflect a proximal position to an eastern source compared to exposures further west and north. Waves and tides become increasingly dominant towards the west, particularly in the outcrops in the Agardhbukta area and represent more distal and transitional environments compared to the succession in the eastern exposures. The succession on eastern Spitsbergen and Wilhelmøya is mostly dominated by tidal and wave processes and reflect more distal depositional environments compared to the southern and eastern part of the study area. The De Geerdalen Formation shows a typical development of multiple coarsening upwards sequences that constitute parasequences normally associated with regressive depositional systems.

Sammendrag

Denne oppgaven undersøker avsetninger av øvre trias alder tilhørende De Geerdals-formasjonen slik den er blottet på østre Svalbard. De Geerdals-formasjonen ble avsatt mellom tidlig karn og tidlig nor på den distale enden av en grunn bukt på den nordlige kanten av Pangea. Formasjonen omfatter et bredt spekter av avsetningsmiljø fra utaskjærs marine miljø og grunnemarin, paraliske delta front avsetninger til strandskråning og kystnære avsetningsmiljø.

Målet med studiet er å dokumentere og tolke sedimentologien av formasjonen, samt den romlige og tidsmessige utviklingen av avsetningsmiljøene gjennom detaljert facies analyse. Femten facies og åtte facies assosiasjoner har blitt definert med grunnlag i observasjoner av sedimentene og blotningsdata. Observasjonene ble samlet inn i løpet en felt-sesong med en måneds varighet i 2015, på Spitsbergen, vestlige deler av Barentsøya og på Wilhelmøya. Tolkningen som presenteres her er basert på tidligere forskning i nærliggende områder, samt samarbeid med andre masterstudenter.

Fra de målte seksjonene er en sterk fluvial dominans under avsetning observert på den nordlige delen av Barentsøya og tolket til å reflektere en proksimal posisjon til en østlig kilde. Bølger og tidevanns-prosesser blir stadig mer dominerende mot vest, spesielt i blotninger i området Agardhbukta og avsetningene representerer distale overgangs-miljøer sammenlignet med tilsvarende blotninger i østlige områder. Lagrekka på østlige deler av Spitsbergen og på Wilhelmøya er for det meste dominert av bølge- og tidevannsprosesser og reflekterer distale avsetningsmiljøer i forhold til den sørlige og østlige delen av studieområdet. De Geerdals-formasjonen viser en typisk utvikling av flere oppgrovende sekvenser som samlet danner parasekvenser som vanligvis assosieres med regressive avsetningssystemer.

Acknowledgements

I would like to start this master thesis by extending my sincere and heartfelt thanks to my supervisor Atle Mørk, Professor II at NTNU without whom, I would not be able to partake in such an interesting and exciting project. Throughout the last year I have benefited from his broad experience, as well as his patience and encouragement. His honest and constructive feedbacks and thorough manuscript reviews have helped improve the quality of this thesis tremendously ever since the first, early drafts.

I would also like to acknowledge my secondary supervisor at the University Centre in Svalbard (UNIS), Professor Snorre Olaussen. My understanding of the Jurassic succession in Svalbard and in arctic petroleum geology has benefited greatly from his experience on the subject. He has also helped facilitate my position as a guest master student at UNIS, making fieldwork in Svalbard and logistical support from UNIS possible.

Thanks to UNIS for logistical support and for providing me with an office space during my stay in Longyearbyen. Fieldwork on eastern Svalbard would not have been possible without the generous financial support from several institutions. Svalbard Science Forum (SSF) gave support through the Arctic Field Grant. The Norwegian Petroleum Directorate (NPD) and Capricorn AS are thanked for their generous financial support, allowing the use of the vessel M/V Sigma and enabled us to visit a great number of localities

Simen Jenvin Støen and Turid Haugen have been great partners, both if the field and in the office, and they have put their signature on this thesis in more than one way. Thank you for the great cooperation throughout this last year. Thanks are also forwarded to my friend and colleague PhD-candidate Gareth Steven Lord for great discussions, and helpful advice, both personal and scientific. A special thanks to my friends and fellow master students Bård Heggem, Cathinka Schaanning Forsberg and Nina Bakke for excellent cooperation and assistance in the field.

Last, but not least, I would like to thank my parents and my siblings for all their encouragement and support throughout my life.

Preface

This thesis is part of a master`s degree in petroleum geology at the Department of Geology and Mineral Resources Engineering (IGB) at the Norwegian University of Science and Technology (NTNU). The thesis is supported by a project work, equivalent to 15 ECTS conducted during the fall of 2015. The topic of the project was the stratigraphy of the Triassic to mid-Jurassic succession of eastern Svalbard. The main supervisor for this thesis has been Professor II Atle Mørk at NTNU. Secondary supervisor has been Professor Snorre Olaussen at the University Centre in Svalbard (UNIS).

Chapter 1 (Introduction), Chapter 2 (Regional geologic setting) and Chapter 3 (Stratigraphy of the Triassic to mid-Jurassic succession) are adapted in its entirety from the project work with only minor alterations. Chapter 4 (Methods) outlines the study area and methodology of fieldwork. Chapter 5 (Facies) and Chapter 6 (Facies Associations) are written in collaboration with fellow master students at NTNU and UNIS, Turid Haugen and Simen Jenvin Støen. The facies scheme presented here is inspired by research started in 2007-2009 and partially extended with the definition of sub-facies associations. Chapter 7 (Results and interpretations) presents the studied localities and interpretations of measured sections. Chapter 8 (Discussion) is also supported by the project work, it has been modified and greatly extended in order to incorporate elements, results and interpretations presented in this master thesis. Chapter 9 (Conclusions) summarizes the main findings from this study and includes recommendations for future work.

During the fieldwork, the students have worked closely together in teams of two or three. As this thesis and that of Simen Jenvin Støen is focusing on the part of De Geerdalen Formation underlying the Isfjorden Member, it has been natural for us to work on the same parts of the section. Many of the sections presented here is therefore also included in his thesis. Turid Haugen has been focusing mostly on the delta-top sediments including the Isfjorden Member and will submit her thesis at a later date.

All photographs are taken during fieldwork in august 2015 by students from NTNU, unless otherwise stated.

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1. Introduction

This sedimentological study investigates the lithologies seen in the Upper Triassic De Geerdalen Formation, which is outcropping over large areas on the arctic archipelago of Svalbard. Triassic rocks are subcropping in larger areas offshore in the Barents Sea, as well as outcropping onshore Svalbard and the area is also known to contain a live petroleum system (Johansen et al., 1993; Spencer et al., 2011; Henriksen et al., 2011; Lundschien et al., 2014) with both potential source and reservoir rocks (Leith et al., 1993; Riis et al., 2008; Mørk et al., 1999a; Klausen and Mørk, 2014; Lundschien et al., 2014; Vigran et al., 2014). Several fields in the Norwegian sector of the Barents Sea host hydrocarbons in reservoirs of Triassic age, sourced from Middle Triassic shales, e.g. the Goliat, Obesum and Norvarg discoveries (Ohm et al., 2008; Henriksen et al., 2011; Lundschien et al., 2014). Hence, demonstrating the importance of understanding the spatial and temporal evolution of the Triassic succession in the Barents Sea and onshore Svalbard.

1.1 Study area

The Svalbard archipelago is a located northwest in the Barents Sea at N78° E21° (Fig. 1.1). It consists of numerous smaller islands as well as the larger main islands of; Spitsbergen, Nordaustlandet, Edgeøya, Barentsøya, Bjørnøya, Wilhelmøya and Hopen.

Sparse vegetation and accessible outcrops make the archipelago a unique location to study thick successions of sedimentary rocks of ages spanning from the Late Paleozoic to recent. Since Svalbard is an exhumed corner of the Barents Sea, it acts as a window to the subsurface and is therefore an essential piece of the puzzle concerning the complex geology of the Barents Sea. For this reason, a large number of geoscientists have studied the area, resulting in a vast amount of literature on geology of the archipelago.

1.2 Purpose of study

The main focus of this thesis will be the Upper Triassic De Geerdalen Formation in north-eastern areas of Spitsbergen and on Wilhelmøya and Barentsøya. The motivation for this study emanates from the necessity to understand the spatial and temporal development of the De Geerdalen Formation in eastern areas of Svalbard.

The purpose and scope of this thesis is to;

i. Investigate sandstones in the De Geerdalen Formation on eastern Svalbard.
ii. Map and document facies, facies associations and their distribution in order to understand the sedimentology and evolution of a Late Triassic delta system.

The stratigraphic nomenclature used in this thesis follows Mørk et al. (1999a), with the addition of the later defined Muen Member of the Middle Triassic Botneheia Formation (Krajewski, 2008) and Hopen Member of the Upper Triassic De Geerdalen Formation (Mørk et al., 2013; Lord et al., 2014a), see Fig. 1.2.

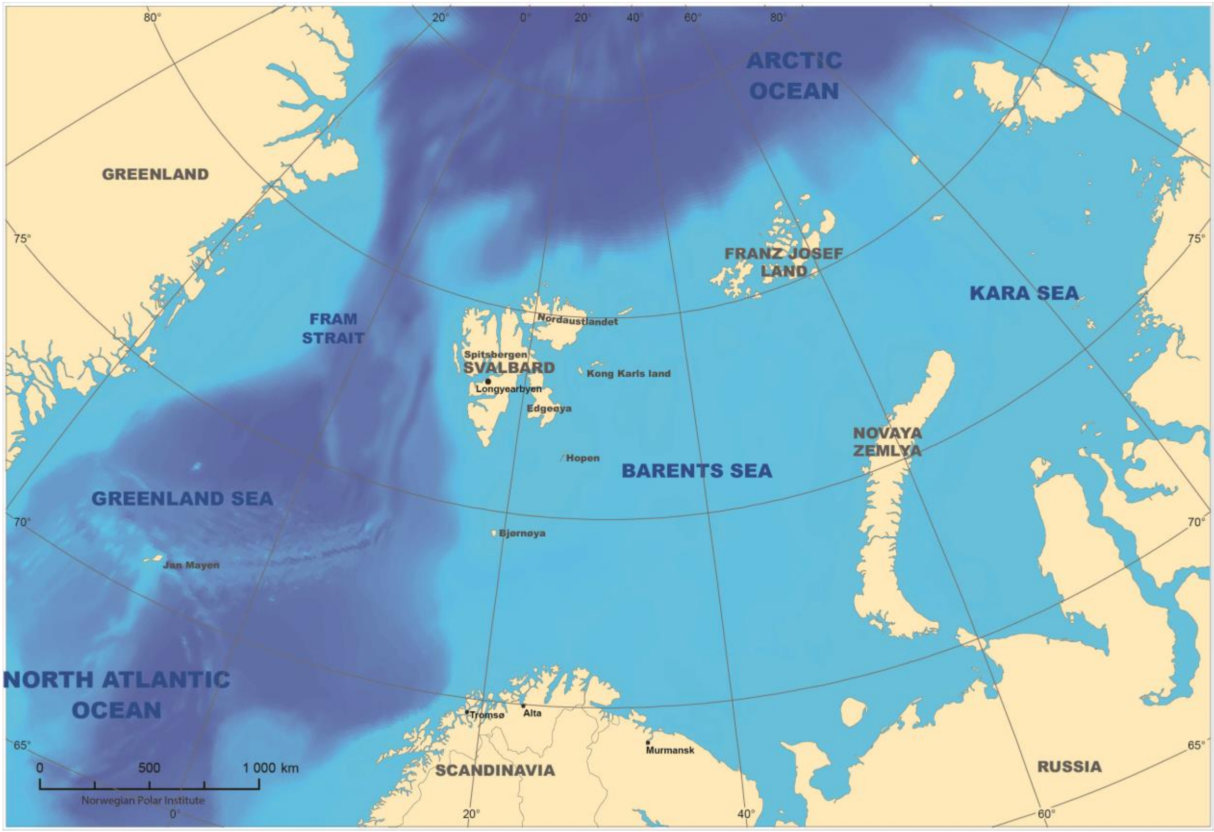

Figure 1.1: Map of the greater Barents Sea area, illustrating the high-latitude position of the Svalbard archipelago in the Norwegian Arctic. Modified from Dallmann (Ed.), 2015.

1.3 Previous research

Previous research on the Triassic succession has focused on outcrops on Spitsbergen (Buchan et al., 1965; Knarud, 1980; Rød et al., 2014), Barentsøya, Edgeøya (Flood et al., 1971; Lock et al., 1978; Winsnes and Worsley, 1981; Mørk et al., 1982) and Hopen (Mørk et al., 2013; Klausen and Mørk, 2014; Lord et al., 2014a,b; Paterson and Mangerud, 2015), with studies of cores and seismic data being conducted in the Barents Sea (Mørk et al., 1993; Leith et al., 1993; Van Veen et al., 1993; Vigran et al., 1998; Mørk and Elvebakk, 1999; Bugge et al., 2002; Riis et al., 2008; Glørstad-Clark et al., 2010, 2011; Høy and Lundschien, 2011; Lundschien et al., 2014; Anell et al., 2014b; Klausen et al., 2015). A summary of palynological and geological research of the Triassic succession on Svalbard and the Barents Sea conducted during the last 40 years is given in Vigran et al. (2014).

The first scientific contribution to geological research on Triassic rocks on the eastern islands was the early account by Falcon (1928), who participated in the 1927 Cambridge Expedition to Edgeøya. Before this, Edgeøya had been visited by B.M Kielhau in 1827 and in 1864 by A.E Nordenskiöld and the Swedish Academy of Science Expedition (Lock et al., 1978). Expeditions from the Norwegian Polar Institute resulted in several publications dealing with the stratigraphy of the Triassic and Jurassic succession (Flood et al., 1971; Worsley, 1973).

Much of the present understanding of the Upper Triassic De Geerdalen Formation, in eastern areas, stems from preliminary stratigraphic and sedimentological studies conducted by W.B. Harland and his Cambridge Expeditions between 1969 and1976 (Smith, 1975; Lock et al., 1978). Other contributions came from workers from the University of Oslo who conducted ship-based research in 1979-1984 (Knarud, 1980; Mørk et al., 1982), Tatjana Pčelina in 1972-1983 (Pčelina, 1980, 1983) and more recently in 2007 - 2009, with the addition of ongoing studies, by students from NTNU (Rød et al., 2014; Lord et al., 2014a,b; Enga, 2015; Harstad, 2016). Sections from the latter are also included in Vigran et al. (2014).

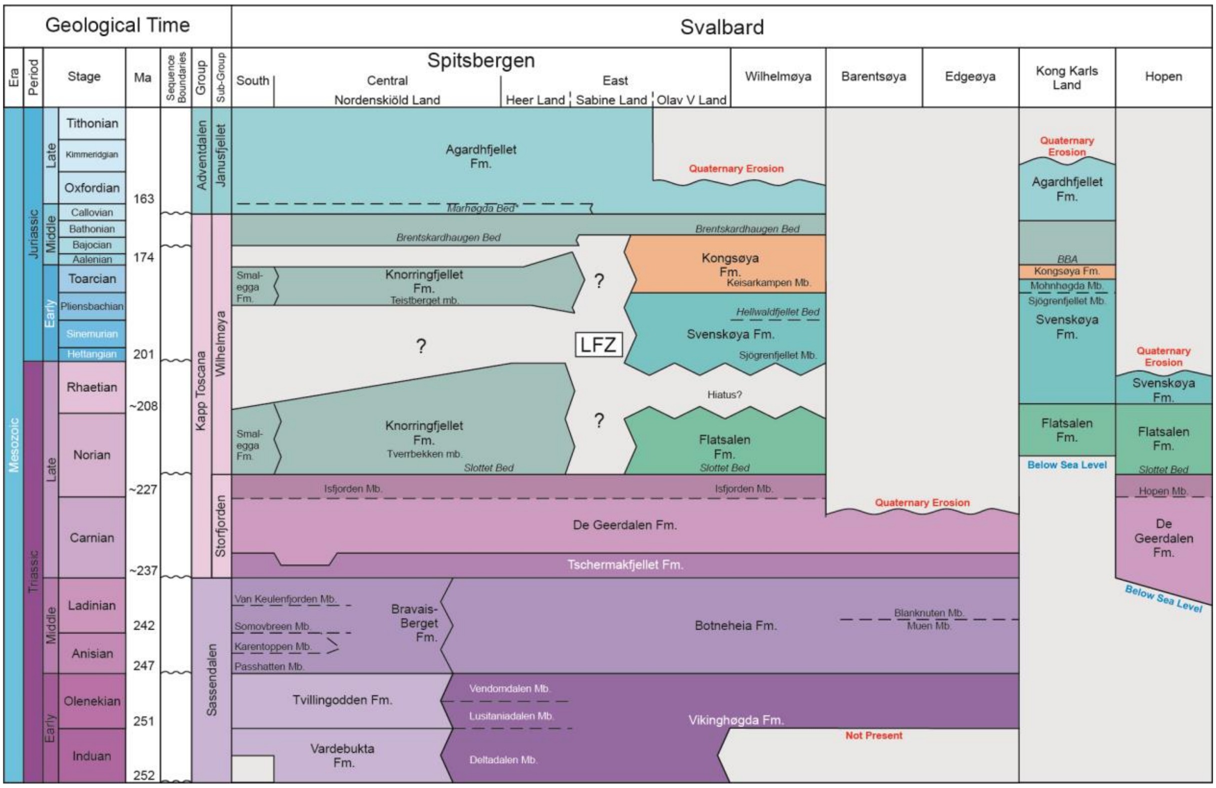

Figure 1.2: Lithostratigraphic table of the Triassic to Late Jurassic succession onshore Svalbard. From Lord et al. (in prep.)

2. Regional Geologic Setting

A review of the post-Caledonian history of Svalbard and the western Barents Sea can be found in Worsley (2008), while the tectonic evolution of the Barents Shelf has been the subject of numerous studies (Rønnevik et al., 1982; Faleide et al., 1984; Doré, 1991; Gudlaugsson et al., 1998; Faleide et al., 2008; Henriksen et al., 2011). A comprehensive overview of the main tectonic domains of Svalbard is given in Dallmann et al. (2015).

2.1 Late Paleozoic

2.1.1 Permian

By the end of the Permian the assemblage of Pangea was to a large degree at its closure and the Barents Sea region was a stable platform with deposition of primarily shallow-marine carbonates and evaporites (Doré, 1991; Stemmerik and Worsley, 2005; Blomeier, 2015). On top of platforms and highs, a hiatus occur in the latest Permian (part of the Wuchiapingian and the Changhsingian) and non-siliceous shales, deposited in basinal areas, can be found resting on top of highly cemented spiculithic shales (Mørk et al., 1999a; Vigran et al., 2014; Blomeier, 2015). The ocean waters became gradually warmer following the "Permian Chert Event" and may have been contributing to the Permian-Triassic mass extinction (Worsley, 2008; Vigran et al., 2014).

In eastern Svalbard, the only outcrops in areas of Triassic exposures are two windows of Permian rock (Fig. 2.1) at Edgeøya and the NE point of Barentsøya (Klubov, 1965; Dallmann and Elvevold, 2015). The Permian–Triassic boundary is exposed in western to central areas of Spitsbergen, on Bjørnøya (Mørk et al., 1990; Lundschien et al., 2014). It is also known from exploration wells, as well as shallow stratigraphic cores from the Svalis Dome, in the Barents Sea (Mørk and Elvebakk, 1999; Lundschien et al., 2014). On Edgeøya, there is a significant hiatus with the entire Induan unit missing, and the oldest rocks here are dated as Olenekian, indicating that the eastern area was sub-aerially exposed (Mørk et al., 1982, 1999a; Vigran et al., 2014).

Figure 2.1: Simplified geologic overview map of Svalbard. From Dallmann and Elvevold (2015).

2.2 Mesozoic

2.2.1 Triassic

In the Triassic, Svalbard and the Barents Sea region formed a large, shallow embayment extending to the Boreal part of Panthalassa towards the northwest and bounded by Siberia and the Ural Mountains to the east, Baltica to the south and Laurentia to the southwest (Figs. 2.2, 2.3, Lundschien et al., 2014).

Infill of the embayment occurred intermittently in cycles of transgressive and regressive episodes (Mørk et al., 1989; Klausen et al., 2016) from Greenland in the west (Pózer Bue and Andresen, 2013) and from Novaya Zemlya and the Ural Mountains in the east and southeast (Riis et al., 2008; Glørstad-Clark et al., 2010; Høy and Lundschien, 2011; Anell et al., 2014b). Basement rocks from the Fennoscandian Shield (Norway and the Kola Peninsula) in the south were the most dominant source of sediments in the Early Triassic. Progradation was primarily directed towards the north and northeast, while the sediment supply from the growing Ural Mountains became increasingly important during later stages and eventually progradation was turned towards the northwest, see Fig. 2.3 (Smelror et al., 2009; Høy and Lundschien, 2011; Henriksen et al., 2011; Anell et al., 2014a; Lundschien et al., 2014).

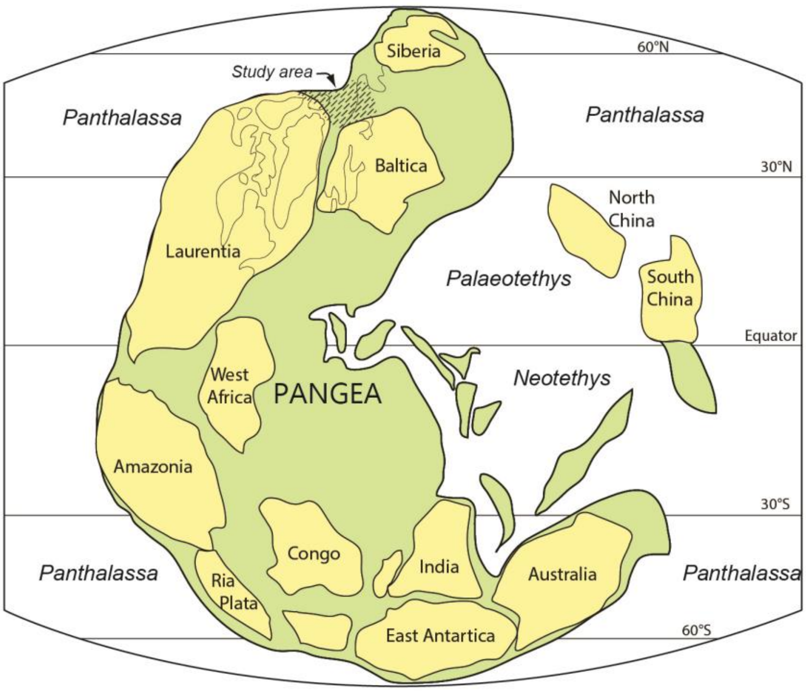

Figure 2.2: Paleogeographic reconstruction of the Pangea supercontinent (from Lundschien et al., 2014, after Torsvik and Cocks, 2005).

High resolution paleogeographical reconstructions by (Klausen et al., 2015) based on seismic interpretation of the Snadd Formation support the hypothesis of a north-westward prograding delta system. Fluvial channels directed towards the northwest and SW-NE oriented shorelines are evident from attribute analysis of seismic data from the Barents Sea (Klausen et al., 2015; Klausen et al., 2016) On Hopen, proximal fluvial channels can be seen in outcrop with paleocurrents towards northwest (Anell et al., 2014a; Klausen and Mørk, 2014; Lord et al., 2014b). Source area for the sedimentary succession on Bjørnøya is uncertain due to the limited extent of exposure here, but it has been suggested to be part of a smaller southwest to western sourced delta system (Worsley, 2008; Klausen et al., 2015).

The prograding delta system from the Ural Mountains in the south-east did not reach Svalbard before the Late Carnian and the uppermost, youngest clinoform sequences may extend past Svalbard and north of Kvitøya (Høy and Lundschien, 2011; Klausen and Mørk, 2014). Some workers have suggested that the delta system from the southeast may have extended even as far as the Sverdrup Basin in Arctic Canada (Lundschien et al., 2014; Klausen et al., 2015). A northern or north-eastern source for the Carnian in Svalbard has also previously been suggested by some workers (Mørk et al., 1982; Rønnevik et al., 1982; Nøttvedt et al., 1993; Van Veen et al., 1993; Skjold et al., 1998), but recent studies have replaced this with the south-eastern Uralide source (Riis et al., 2008; Glørstad-Clark et al., 2010, 2011; Høy and Lundschien, 2011; Anell et al., 2014b; Rød et al., 2014; Klausen et al., 2015; Klausen et al., 2016). Topography in the Early Triassic was generally flat (Høy and Lundschien, 2011), but paleo-highs existed and the paleo-Loppa High may have acted as a barrier to progradation towards the west (Van Veen et al., 1993; Glørstad-Clark et al., 2010), but would later evolve into a major depocenter in Late Triassic (Glørstad-Clark et al., 2011).

The Triassic to mid-Jurassic period has generally been considered to be relatively tectonically stable in the western Barents Sea and Svalbard (Mørk et al., 1982; Faleide et al., 1984, 2008; Gabrielsen et al., 1990; Riis et al., 2008; Glørstad-Clark et al., 2010; Henriksen et al., 2011; Høy and Lundschien, 2011; Klausen and Mørk, 2014). However, Anell et al. (2013) and Osmundsen et al. (2014) argue that minor fault activity occurred in the Triassic and re-activated deep-rooted faults that were initiated during the Paleozoic, based on interpretation of 2D seismic data and studies of growth faults at Kvalpynten, southwestern Edgeøya. A general agreement is that the Late Triassic basin evolution was dominated by regional subsidence with large volumes of sediments being deposited (Faleide et al., 1984, 2008; Riis et al., 2008; Smelror et al., 2009; Glørstad-Clark et al., 2010).

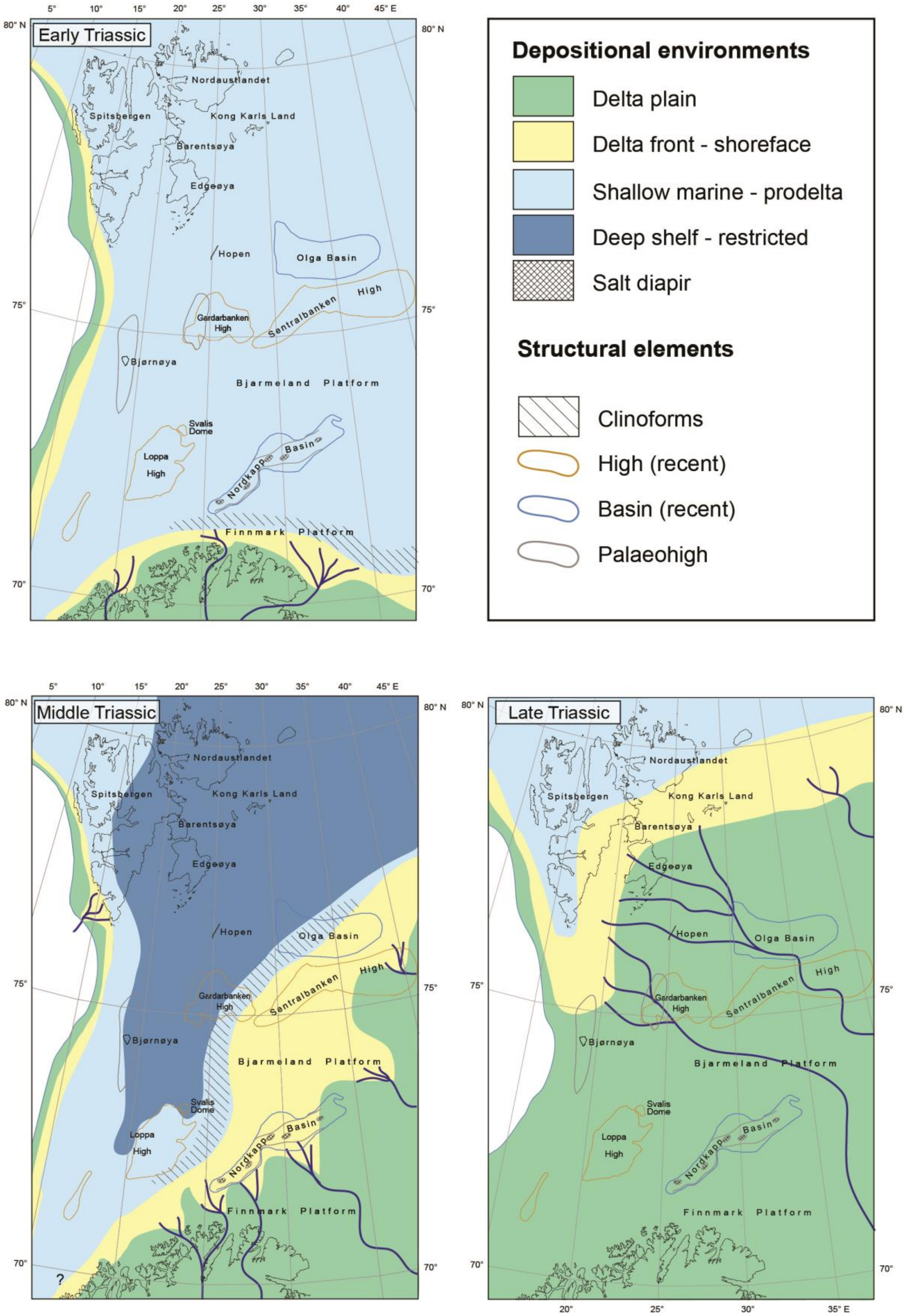

Figure 2.3: Paleogeographic reconstructions of the Triassic succession in the Barents Sea region. Important sediment sources were to the west (Greenland), south (Fennoscandian Shield) and southeast to east (Ural Mountains and Novaja Zemlya). From Lundschien et al. (2014).

During the Triassic and Jurassic, the Svalbard archipelago and the Barents Sea region shifted from a latitude around 45° in the Early Triassic to approximately 60° in mid-Jurassic time (Glørstad-Clark et al., 2011; Dallmann, 2015) and subsequently, the climate shifted from arid to humid as the area drifted towards northwards (Mørk et al., 1982; Hochuli and Vigran, 2010; Ryseth, 2014; Enga, 2015; Klausen et al., 2015). Palynological studies have suggested a prevailing humid climate on Hopen during the Late Triassic (Paterson et al., in press) and plant fossils of a humid environment have also been reported from this locality (Launis et al., 2014).

2.2.2 Jurassic

In the Early Norian, a regional transgression, accompanied with a shift from an arid to humid climate led to a major change in mineralogy from the immature plagioclase-rich quartz arenites of the Late Triassic to the more mature and coarse-grained quartz arenites of the Early Jurassic (Bergan and Knarud, 1993; Mørk, 1999; Mørk, 2013; Ryseth, 2014). On Svalbard and in larger areas of the Barents Shelf, widespread uplift and erosion occurred during the transition to the Early Jurassic (Nøttvedt et al., 1993; Smelror et al., 2009). Following the Norian flooding, the Tethyan ocean in the south were connected with the Boreal ocean to the north (Doré, 1991), and a long period of sedimentation in coastal lowlands to shallow marine shelves was initiated (Olaussen et al., 1984; Worsley, 2008; Smelror et al., 2009).

The paleogeographic setting of the Late Triassic was most likely continued in the Early Jurassic (Smelror et al., 2009; Henriksen et al., 2011). Widespread deltaic and coastal conditions were established over larger areas of the Barents Sea (Smelror et al., 2009) and Svalbard (Fig. 2.4). Shoreline regression in Rhaetian to Early Sinemurian resulted in a maximum regressive stage in the Late Pliensbachian - Early Toarcian. This coastal progradation was immediately followed by retrogradation of the coast towards the northeast (Henriksen et al., 2011; Ryseth, 2014; Larssen et al., in prep.).

The Early Jurassic Nordmela Formation in the Troms I area is interpreted to have been deposited in a coastal plains with tidal flats, while the overlying Stø Formation may have formed in a higher energy environment as interpreted from the more mature sandstones (Olaussen et al., 1984; Henriksen et al., 2011). The Early to Middle Jurassic succession is best developed in the Hammerfest Basin in the southern Barents Sea, where it forms the main reservoir unit in the Snøhvit gas field (Doré, 1995; Spencer et al., 2008) and also in the more recent Johan Castberg and Wisting oil discoveries (Klausen et al., in prep.). In the northern Barents Sea, it is either absent or thins considerably and the stratigraphy may not be complete (Doré, 1991; Lundschien et al., 2014).

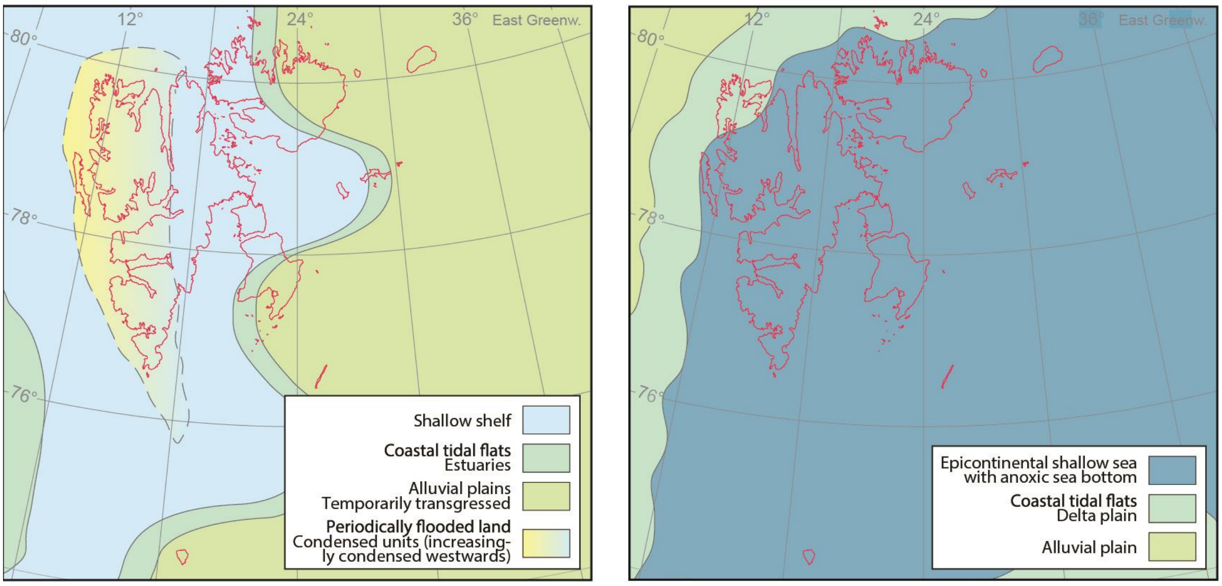

Figure 2.4: Paleogeographic reconstructions of the Jurassic succession in the Barents Sea region. **Left:** Pliensbachian, ca. 185 Ma **Right:** Kimmeridgian, ca. 155 Ma. From Olaussen (2015).

In general, the Early Jurassic to mid-Jurassic sedimentary succession on Svalbard is condensed and consists of several hiati, preventing robust bio-stratigraphic relationships to be established (Mørk et al., 1999a; Smelror et al., 2009). Sedimentary rocks from Late Triassic to Middle Jurassic are included in the Wilhelmøya Subgroup (Worsley, 1973) and on Svalbard, they are found primarily in eastern areas, as well as on Hopen and Kong Karls Land (Mørk et al., 1999a; Mørk et al., 2013).

Widespread deep shelf conditions were established in the Late Jurassic (Fig. 2.4), due to differential subsidence and uplift (Henriksen et al., 2011). Anoxic bottom water-conditions in the deeper parts of the basins, combined with high atmospheric CO_2-level, resulted in the preservation of organic rich marine mudstones (Faleide et al., 1984; Olaussen, 2015). In the southwestern Barents Sea these black shales are known as the Hekkingen Formation and form an excellent source rock for hydrocarbons. The mudrocks contains up to 20 % TOC and is dominated by marine Type II/III kerogen (Leith et al., 1993; Worsley, 2008; Henriksen et al., 2011). In eastern Svalbard, these mudrocks are included in the Agardhfjellet Formation (Mørk et al., 1999a) and can be found overlying the Brentskardhaugen Bed on Wilhelmøya and eastern Spitsbergen.

2.2.3 Cretaceous magmatism

Extensive magmatism in the Early Cretaceous resulted in the formation of a High Arctic Large Igneous Province (HALIP), which is most prominent in the area of Franz Josef Land to the east (Maher, 2001). On Svalbard, the equivalent intrusive sills and dykes are referred to as the Diabasodden Suite (Mørk et al., 1999a; Maher, 2001; Senger et al., 2014b). These dolerite intrusions penetrate several levels of the Triassic succession in central and eastern Svalbard. They are often seen intruding the soft shales of the Botneheia and Tschermakfjellet Formation, as well as the harder sandstones of the De Geerdalen Formation (Senger et al., 2013, 2014a, b). Recent U – Pb dating of samples from Svalbard and Franz Joseph Land has helped constrain the age of the Diabasodden Suite to 124.5 Ma, i.e. Early Cretaceous (Corfu et al., 2013; Senger et al., 2014b).

Regional uplift in the north of Svalbard followed a regional marine transgression in the Late Cretaceous and as a result, sediments were dominantly transported towards the south (Faleide et al., 1984; Steel and Worsley, 1984). Erosion occurred on Svalbard and no Upper Cretaceous can be found here (Nøttvedt et al., 1993; Mørk et al., 1999a).

2.3 Structural geology

The West Spitsbergen Fold-and-Thrust Belt formed due to transpressional stresses as Svalbard and the Barents Shelf slid from the north-eastern edge of Greenland to its present position east of Greenland (Harland et al., 1974; Lyberis and Manby, 1993; Braathen et al., 1999). This dextral strike-slip movement resulted from the opening of the North-Atlantic Ocean during the transition from the Paleocene to Eocene (Bergh et al., 1997; Faleide et al., 2008; Smelror et al., 2009). It is commonly subdivided into a western 'thick-skinned' thrust zone that involve metamorphic basement rocks and in an eastern 'thin-skinned' zone that have a geometry consisting of ramps and at areas. This latter forms the base of the sedimentary succession of eastern Svalbard (Bergh et al., 1997; Braathen et al., 1999; Dallmann, 2015).

To the east of this prominent mountain belt, a foreland basin developed and is referred to as the Central Tertiary Basin (CTB) in western literature, while in the Russian literature this is known as the West Spitsbergen Trough (Dallmann et al., 2015). Folding and thrusting associated with the development of the fold belt in the west have resulted in Triassic rocks being heavily deformed in this area (Buchan et al., 1965; Mørk et al., 1982), which causes complexity for paleogeographic reconstructions and stratigraphic relationships for the Triassic succession here.

2.4 Cenozoic uplift

The Barents Shelf was uplifted and large areas in the northern platform were eroded during the Neogene (Late Pliocene–Pleistocene) and approximately 2-3 km of sediments were eroded and subsequently transported to depocenters along the western margin (Smelror et al., 2009). Repeated glaciations were the primary cause for this uplift and have been instrumental in shaping the present day topography of Svalbard (Worsley, 2008).

3. Stratigraphy of the Triassic to mid-Jurassic succession

The sedimentary succession in the study area rests flat and relatively undeformed on the Edgeøya Platform. The lower boundary is defined at the Permian level and the platform has been stable since the Carboniferous (Bergsager, 1986; Gabrielsen et al., 1990). The platform dips gently towards the east and in the northern Storfjorden area and between north-eastern Spitsbergen and Wilhelmøya; it forms a gentle depression known as the East Svalbard Depression (Dallmann et al., 2015).

The Triassic succession of Svalbard is made up of the lower Sassendalen Group and the Upper Triassic to mid-Jurassic Kapp Toscana Group (Buchan et al., 1965; Mørk et al., 1982, 1999a). The total thickness of the succession is estimated to be a minimum 250 m locally, up to of a maximum of 1200 m onshore Spitsbergen (Mørk et al., 1982). The Triassic succession on Bjørnøya is in total 200 m thick and make up the youngest deposits on the island (Mørk et al., 1990).

On Hopen, the total thickness of Triassic sediments is estimated to be approximately 1215 m based on data from the Hopen-2 well, drilled by the Fina Group in 1973. The base of the Triassic is interpreted to occur at a sharp gamma ray (GR) peak in the well log at (-) 1325 m. This corresponds to a depth of 1050 m below sea level and is interpreted as representing the top of the cemented carbonates of the Permian Kapp Starostin Formation (G. Lord, pers. comm. 2015). The Upper Triassic De Geerdalen Formation is interpreted from the same log to be approximately 650 m thick in the region of Hopen (Lord et al., 2014a). This is considerably thicker than equivalent rocks on central Spitsbergen, where it thins to 200-300 m.

In the Barents Sea, the Triassic succession is thicker, with up to 2000 – 3000 m of sedimentary rocks (Worsley et al., 1988; Riis et al., 2008). Henriksen et al. (2011b) report that the total thickness of the Induan – Early Norian succession in the Norwegian sector of the Barents Sea exceeds 2500 m. The main factor governing this difference is sediment supply from the Ural Mountain and variable thickness of seismic sequences implies that the sediment supply is fluctuating throughout the Early and Middle Triassic (van Veen et al., 1993).

3.1 Lower Triassic

The Lower Triassic in western Svalbard consists of laminated shales and coastal sandstones of the basal Vardebukta Formation of Induan age and overlying Tvillingodden Formation of Olenekian age (Mørk et al., 1999a; Vigran et al., 2014). The equivalent unit in central and eastern areas of Spitsbergen and Svalbard is the Vikinghøgda Formation, defined by Mørk et al. (1999b) as grey laminated shales, with occasional silt- and sandstone beds.

The correlative units in the southwestern Barents Sea are the Havert and Klappmyss formations (Worsley et al., 1988). On Bjørnøya, the Lower Triassic is represented by the Urd Formation of Induan – Olenekian age, deposited in a shallow marine environment (Mørk et al., 1982, 1990; Vigran et al., 2014).

Prior to the opening of the North-Atlantic Ocean in the Paleocene, Svalbard was situated much closer to Greenland which was an important sediment source throughout much of the Early and Middle Triassic (Glørstad-Clark et al., 2010; Pózer Bue and Andresen, 2013). In western Svalbard, Triassic sediments reflect a more proximal position to those sediments coming from the west. Deposition was primarily distal to the sediment supply from the Ural Mountains in the south-east throughout the Lower and Middle Triassic. But the exact relationship for when the south-eastern sediment source started to dominate over the western is not well constrained (Mørk et al., 1999a; Worsley, 2008; Glørstad-Clark et al., 2010).

3.2 Middle Triassic

The Botneheia Formation, with its equivalent Bravaisberget Formation in the west, has a high content of organic matter, with Total Organic Carbon (TOC) values ranging between 1 % and 10 % (Mørk et al., 1982; Lock et al., 1978; Leith et al., 1993). The offshore equivalent of the Botneheia Formation is called the Kobbe Formation and the Steinkobbe Formation in the Barents Sea, where the latter contains the best source rock potential (Leith et al., 1993; Mørk and Elvebakk, 1999; Riis et al., 2008; Lundschien et al., 2014). It is subdivided into the lower Muen Member, consisting of fissile, black paper shales (Krajewski, 2008) and the Upper Blanknuten Member (Mørk et al., 1999a; Vigran et al., 2014). The formation is rich in Type-II/III kerogen, partly due to the presence of *Tasmanites* algae (Vigran et al., 2008). The hydrocarbon-rich shales of the Botneheia Formation were deposited in a deep marine open shelf environment with periodic oxic and anoxic bottom water conditions (Krajewski, 2008; Vigran et al., 2008, 2014). The Middle Triassic on Bjørnøya is represented by a remanié conglomerate bed with phosphate nodules called the Verdande Bed (Mørk et al., 1982; Mørk et al., 1990; Vigran et al., 2014).

3.3 Upper Triassic

3.3.1 Kapp Toscana Group

The Kapp Toscana Group was first defined by Buchan et al. (1965) as a formation and further subdivided into the lower Tschermakfjellet Member and the upper De Geerdalen Member. It was upgraded to a group status

3.3.2 Storfjorden Subgroup

The Tschermakfjellet Formation consists of a coarsening upwards succession of grey to purple, silty shales, with siderite nodules. The formation is interpreted to have been deposited in a prodelta environment and is easily mapped due to its characteristic colour (Mørk et al., 1999a; Vigran et al., 2014). The Tschermakfjellet Formation differs from the underlying Botneheia Formation, with oxic conditions becoming dominant and generally marks the initiation of sediment influx from the Ural Mountains in Svalbard. Dating of the Tschermakfjellet Formation by ammonoids has resulted in an Early Carnian age (Korčinskaja, 1982; Weitschat and Dagys, 1989). Recent palynological dating of shallow stratigraphic cores offshore Kong Karls Land also assigns the Tschermakfjellet Formation to an early Carnian age (Vigran et al., 2014; Paterson et al., 2016).

The Upper Triassic De Geerdalen Formation constitute together with the underlying Tschermakfjellet Formation, the onshore equivalent of the offshore Snadd Formation in the Barents Sea (Worsley et al., 1988; Klausen and Mørk., 2014; Klausen et al., 2014, 2015). The lower boundary is generally defined at the base of the first prominent sandstone above the Tschermakfjellet Formation (Buchan et al., 1965; Flood et al., 1971; Lock et al., 1978; Mørk et al., 1999a) and consists of several stacked coarsening upwards sequences of sandstones and shales of Carnian to Early Norian age (Tozer and Parker, 1968; Korčinskaja, 1982; Mørk et al., 1999a; Vigran et al., 2014; Paterson et al., 2016). The depositional environment is interpreted as shallow marine to deltaic in larger areas of Svalbard, with proximal fluvial processes dominating on Hopen (Klausen and Mørk, 2014; Lord et al., 2014a,b) and Edgeøya, and distal marine influences on central Spitsbergen (Vigran et al., 2014; Rød et al., 2014).

The upper part of De Geerdalen Formation in western and central Spitsbergen is referred to as the Isfjorden Member and is dominated by shales, with occasional sandstones, of alternating red and green colour. The Isfjorden Member was deposited in the Early Norian in a shallow marine, coastal environment with locally lagoonal conditions (Pčelina, 1980; Mørk et al., 1999a). The uppermost part of the formation is not exposed on the eastern islands of

Barentsøya and Edgeøya (Vigran et al., 2014), but can be seen on eastern Spitsbergen and on Wilhelmøya. The Skuld Formation on Bjørnøya is an immature sandstone succession consisting of several coarsening upwards units deposited in a shallow marine environment as interpreted from sedimentary structures and fossil fauna (Mørk et al., 1990, 1999a; Vigran et al., 2014).

On Hopen, the upper part of De Geerdalen Formation shows a distinct marine influence motivating the definition of the Hopen Member (Mørk et al., 2013; Lord et al., 2014a). The Hopen Member has recently been dated on the basis of palynology to a Late Carnian, possibly lowermost Norian age, which is younger than the uppermost rocks on Edgeøya and Barentsøya (Lord et al., 2014a; Vigran et al., 2014; Paterson and Mangerud, 2015; Paterson et al., in press).

The vertical and spatial development from the prodelta shales of the Tschermakfjellet Formation, to the shallow marine environment of De Geerdalen Formation, in central Spitsbergen and along a transect of western Edgeøya is documented in Rød et al. (2014). Syn-sedimentary growth faulting in the Upper Triassic at Kvalpynten on Edgeøya has also previously been documented by several workers (Edwards, 1976a; Lock et al., 1978; Osmundsen et al., 2014). It is suggested to have formed due to instability in the prodelta shales of the Tschermakfjellet Formation, detaching on the contact with the Botneheia Formation (Osmundsen et al., 2014).

3.4 Upper Triassic – Middle Jurassic

3.4.1 Wilhelmøya Subgroup

The Wilhelmøya Subgroup is defined at its base by the Slottet Bed, of Early Norian age (Korčinskaja, 1980; Basov et al., 1993). This bed is a well sorted and texturally mature, polymictic conglomerate that contains phosphate nodules. It is interpreted to represent a transgressive episode throughout large areas of Svalbard, Barents Shelf, Canada and Russia (Mørk et al., 1999a; Riis et al., 2008; Henriksen et al., 2011; Vigran et al., 2014).

The Flatsalen Formation is defined as the marine shales, siltstones and fine-grained sandstones occurring above the Slottet Bed. This formation corresponds to the “Basal Member”, “Transitional Member” and “Bjørnbogen Member” of Worsley (1973). The type section for the Flatsalen Formation is on Hopen, and the formation can also be seen above the Slottet Bed at Tumlingodden on Wilhelmøya and on eastern Spitsbergen (Mørk et al., 1999a). It is interpreted to have been deposited in a shallow marine environment during the Norian

(Smith et al., 1975; Korčinskaja, 1980; Mørk et al., 1999a; Anell et al., 2014a; Paterson and Mangerud, 2015).

The Svenskøya Formation was deposited during the Norian–Rhaetian (Smith et al., 1975; Pčelina, 1980; Paterson and Mangerud, 2015; Paterson et al., in press) in a proximal shallow marine environment influenced by tidal processes. The formation is composed of primarily greenish fine- to medium- grained sandstone fining upwards into white sandstones interbedded with thin mudstones (Mørk et al., 1999a). It is not present on the major eastern islands, but is exposed on eastern Spitsbergen, Wilhelmøya and Hopen (Mørk et al., 1999a, 2013; Lord et al., 2014a).

The Kongsøya Formation is found above the Svenskøya Formation and consists of fine-grained, muddy sandstones and greyish-blue mudstones with siderite beds (Mørk et al., 1999a). The formation is interpreted to have been deposited during the Toarcian– Bathonian in a shallow marine to inner shelf environment (Smith et al., 1975; Pčelina, 1980; Mørk et al., 1999a). In western Spitsbergen, the Flatsalen, Svenskøya and Kongsøya formations are referred to as the Knorringfjellet Formation and represent a condensed section of these sediments (Mørk et al., 1999a).

The upper boundary of the Wilhelmøya Subgroup and the Kongsøya Formation is marked by the Brentskardhaugen Bed of Bajocian (?) – Bathonian age (Parker, 1967; Bäckström and Nagy, 1985). This thin, orange, conglomeratic to calcareous sandstone unit, locally underlies the Marhøgda Bed of Late Bathonian or Early Callovian age (Pčelina, 1980) and is composed of pebbles and phosphate nodules with fossil inclusions (Mørk et al., 1999a). The erosive base and the age of the Brentskardhaugen Bed indicate a major hiatus with the underlying Kongsøya Formation (Bäckström and Nagy, 1985; Mørk et al., 1999a; Anell et al., 2014a). In the literature, there is a discrepancy on whether the Brentskardhaugen Bed should be placed at the top of the Kapp Toscana Group (Parker, 1967; Worsley, 1973; Pčelina, 1980; Mørk et al., 1982, 1999a) or as the base of the overlying Janusfjellet Subgroup (Flood et al., 1971; Bäckström and Nagy, 1985).

The entire Subgroup is interpreted to represent a pan-Arctic regional flooding and was deposited in shallow marine to deltaic environments (Worsley, 2008). Towards the west, the interval becomes increasingly condensed (Mørk et al., 1999a). The equivalent units offshore in the southwestern Barents Sea are the Realgrunnen Subgroup, consisting of the Late Triassic to Middle Jurassic Fruholmen, Tubåen, Nordmela and Stø formations (Olaussen et al., 1984; Worsley et al., 1988).

4. Methodology

4.1 Fieldwork

Fieldwork was conducted in eastern Svalbard during a month-long field season in August 2015. A team of seven researchers were deployed in Agardhbukta on eastern Spitsbergen after a two-day expedition with M/V *Sigma* to Muen and Blanknuten on Edgeøya. In Agardhbukta, the team were accommodated in a cabin courtesy of Longyearbyen Jeger og Fiskeforening (LJFF), situated below the plains of Belemnittsletta. Over a period of two weeks, the mountains of Klement'evfjell, Friedrichfjellet and Šmidtberget were accessed by the use of rubber boats for transport across the bay. All these mountains expose outcrops of the De Geerdalen Formation. The Agardhbukta beach section is located in close proximity to the cabin, and was logged when sea-ice prevented effective transportation across the bay.

The team was picked up by the vessel M/V *Sigma* on August 17th, giving the researchers greater mobility and facilitated access to more remote outcrops of the De Geerdalen Formation and the Wilhelmøya Subgroup. The main advantages of a ship-based operation are better mobility, better polar bear safety, reduced time spent on logistics and less dependence on good weather, in contrast to being tent or helicopter-based.

A total number of 13 localities were visited during the field season and are along with measured sections further described in Chapter 7.

4.2 Methods

GPS-measurements in standard UTM-coordinates were taken at the base and top of each section, as well as at significant stratigraphic levels within the succession. An overview of these coordinates for all logs can be found in Appendix A.

The observations collected during this field season and the results presented in this master thesis, were obtained using standard sedimentological field methods. This thesis is mainly concerned with a sedimentological analysis of sandstones in the De Geerdalen Formation and hence, the main emphasis of observations has therefore been restricted to these sandstones. Mudrocks are often partly to completely covered by erosional scree and smaller plants and needs to be excavated in order to be described. Observations of the sandstones were primarily focused on features such as sedimentary structures, grain sizes, bed thicknesses, colour, stratigraphic relationships, bed contacts and textures. Sections were measured using a meter stick and grain sizes were visually estimated using a standard grain size sheet. Legend for measured sections can be found in Appendix B.

Extensive sampling of coarse-grained material for provenance and thin section analysis, as well as sampling of fine-grained material for palynological studies were conducted. These were taken in order to provide material that can be used for future projects and master theses. Thin section analysis for diagenesis is beyond the scope of this master thesis, but is partially the focus of a master thesis by Støen (in. prep.).

A total of 25 sedimentary logs were recorded from various localities on eastern Svalbard, of which 12 (logged by the author) are included in this thesis (Appendix C). The other logs are presented in several other master and doctorate theses (Haugen, in prep.; Støen, in prep., Lord et al., in prep.). All logs were recorded in field notebooks using a scale of 1:100, in order to effectively log the De Geerdalen Formation in the limited amount of time available. The start of the logs is usually from the base of the first prominent sandstone and ended at points of overlap with logs of other co-operating field groups or when the Slottet Bed was reached, which marks the onset of the Wilhelmøya Subgroup. Sections were primarily measured vertically from bottom to top, as time and exposure limitations often did not allow for measured sections to be walked out laterally.

The following workflow has been applied during the work of this thesis in order to establish and interpret depositional environments:

1. Definition of facies and facies associations based on field observations and descriptions.
2. Interpretation of facies on measured sections from each locality.
3. Interpretation of facies association and depositional environments based on groups of associated facies, observations and outcrop descriptions from the field.

4.3 Sources of error

The main sources of error in the work presented here are concerned with field observations and measurements of paleocurrents and altimetry. Due to the high latitude, the magnetic field is inclined and hence compass measurements needs to be corrected. The succession is generally extensively scree-covered, often where the underlying sediments are composed of mudrocks and shales (Hynne, 2010).

The lower boundary of the De Geerdalen Formation is defined occurring at “the base of the first prominent (metre scale to tens of metres) sandstone unit” (Mørk et al., 1999a, p. 166).

However, in the field this lower boundary is ambiguous as the lower parts of the formation are often found to be covered by scree. Determining whether a sandstone bed is “prominently occurring” can therefore prove challenging. Lock et al. (1978) defined prominent as “thicker than 1 cm”, but they also acknowledged the difficulties encountered when determining the boundary in the series of coarsening upwards cycles characterize the lower part of the formation. At some of the localities, e.g. Krefftberget and Mistakodden, locating the boundary was challenging and the logs were measured to include intervals of the underlying Botneheia and Tschermakfjellet formations. The base of the De Geerdalen Formation is therefore in this thesis taken at the first prominent sandstone, thicker than 50 cm as it occurs in the field.

5. Facies in the De Geerdalen Formation on eastern Svalbard

This chapter has been written as a collaboration between master students Turid Haugen, Simen Jenvin Støen and Sondre Krogh Johansen, who also worked together in the field. 15 facies has 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 are given in Table 1. This master thesis including that of Støen (in prep.) focuses primarily on sandstones. Facies O (Paleosols) is therefore treated in more detail by Haugen (in prep.)

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. Our study is complementing these works and the study area includes a number of locations not previously visited by these workers.

Facies analysis

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

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 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, m – medium

#	Facies	Grain size	Description
A	Large-scale cross-stratified sandstone	f – med	Trough to tabular cross stratification, erosive base fining-upwards. Set thickness is between 20 - 80 cm, while unit thickness is in the range of 0.2 - 4 m. Rip-up clasts and plant fragments are observed and typical trace fossils include *Skolithos* and *Diplocraterion*.
B	Small-scale cross-stratified sandstone	vf – f	Asymmetric ripples (2D and 3D ripples), sets thicknesses 2-10 cm, unit thicknesses up to 1.5 m. Mud drapes and sparse bioturbation. Vague to pervasive cementation
C	Climbing ripple cross-stratified sandstone	vf - f	Small-scale asymmetric climbing ripple laminated. Sharp lower contacts, while upper contacts are gradual. Unit thickness ~ 0.5 m
D	Wave rippled sandstone	vf - f	Symmetrical wave ripples, planar parallel 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 between 5 - 15 cm and unit thicknesses up to 1.5 m. Commonly bioturbated and contains plant fragments
F	Horizontally bedded sandstone	vf - f	Planar parallel stratification (PPS) and lamination (PPL). Unit thickness 30 cm to 2 m. Laminae and bed thicknesses varies within units. Parting lineation. *Skolithos*, *Diplocraterion* and *Rhizocorallium* in upper parts of units.
G	Massive, structureless sandstone	vf - med	Fractured, apparently structureless. Units are between 0.1 m to 5 m thick. Sharp erosive base, commonly enclosed by mudrocks and heterolithic units. Carbonate cemented. Often bioturbated and contains plant fragments and mud flakes.
H	Hummocky and swaley cross-stratified sandstone	si - f	Hummocky and swaley cross stratified. 5 to 20 cm thick sets with unit thickness up to 1 m. Moderately to intensely bioturbated, *Skolithos* and *Diplocraterion*.
I	Soft sediment deformed sandstone lenses	vf - f	-
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 appear 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 centimetres 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.

Facies A - Large-scale cross-stratified sandstone

Description

This facies consist of fine to medium-grained, large-scale 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. 5.1). This facies comprise the coarsest sand grain size in the study area, and sandstones from this facies appear more texturally mature compared to similar facies.

Rip-up clasts (Fig. 5.1E) 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. 5.1D, 5.1F) are more commonly noticed on the western localities in the study area.

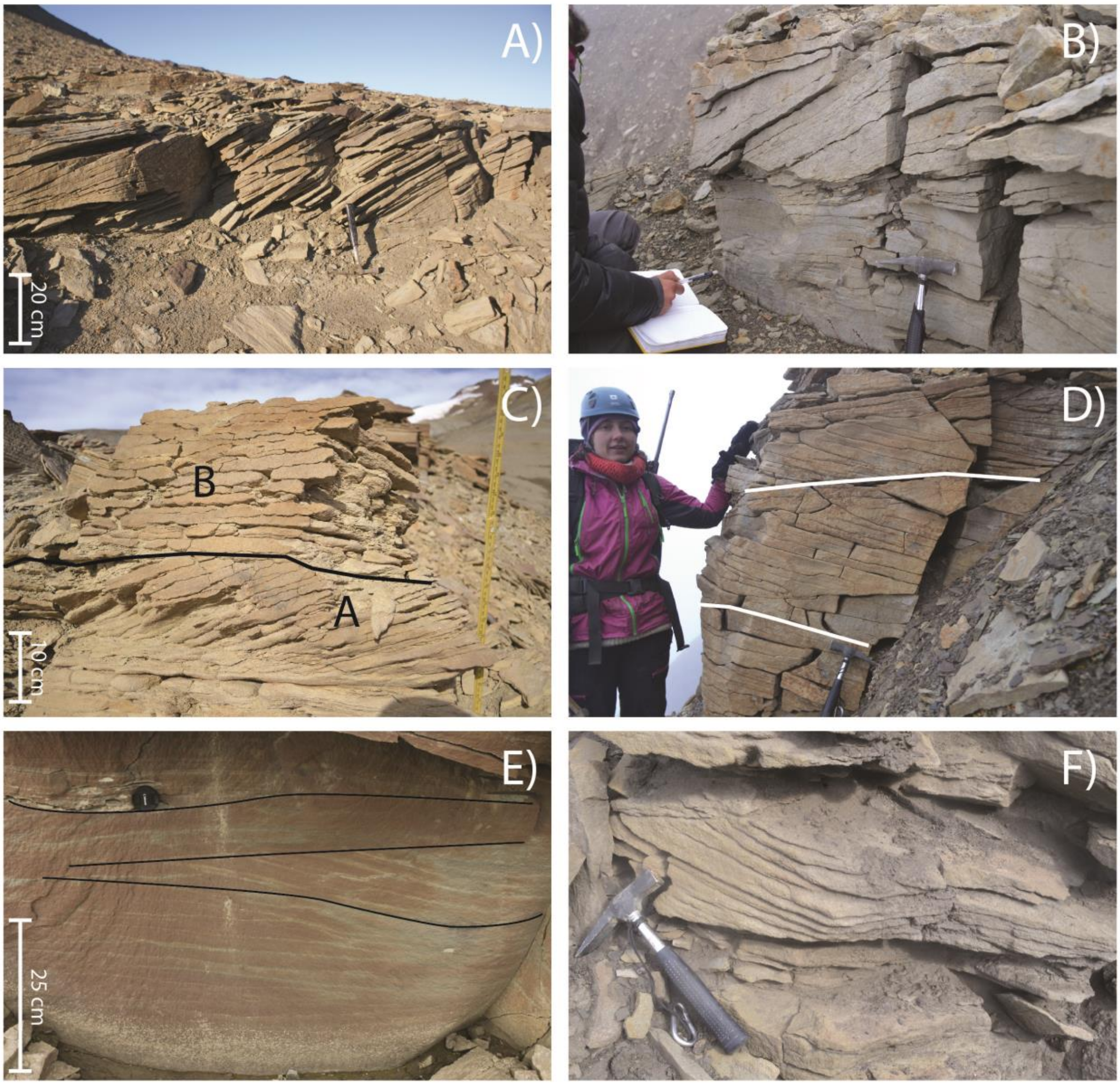

Figure 5.1: *Facies A - Large-scale cross-bedded sandstone* **A)** Tabular cross-bedded sandstone at Wilhelmøya. **B)** Trough cross-bedded sandstone at Šmidtberget, Agardhbukta. **C)** Tabular cross-bedded sandstone (facies A) overlain by small-scale asymmetric cross-bedding (facies B), Hellwaldfjellet, 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.

Interpretation

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

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

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

Facies B - Small-scale cross-stratified sandstone

Description

This facies comprise small-scale asymmetric ripple laminated very fine to fine sandstones. Ripple cross- lamination is arranged in sets of 2 to 10 cm height and stacked in units that are up to 1.5 m thick (Fig. 5.2).

Cementation, mainly calcite, varies from vague to pervasive resulting in differences in appearance within facies. The facies often appear as undulating, parallel wavy to straight bedding/set boundaries without apparent cross-stratification (Figs. 5.2A, 5.2B, 5.2C). 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 is interpreted as mud drapes (Fig. 5.2D).

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. 5.2A).

Figure 5.2: *Facies B - Small-scale cross bedded sandstone*. **A)** Small-scale cross-bedded sandstone above horizontally bedded sand (facies F), wave rippled sandstone (facies D) and large-scale cross-stratified sandstone. The lowermost unit represents the strongest current conditions while the units above are inferred to be deposited by a decelerating flow, Teistberget, eastern Spitsbergen **B)** Small-scale cross-bedded ripple laminated sandstone on Wilhelmøya **C)** Small-scale ripple cross stratification on Friedrichfjellet, Agardhbukta **D)** Mud draped foresets on unidirectional current ripples. Flow direction is towards the right from left, Wilhelmøya.

Interpretation

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

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

Facies C - Climbing ripple cross-laminated sandstone

Description

Small-scale asymmetric climbing ripple laminated very fine to fine sandstone (Fig. 5.3). Yellow, orange and brownish colours observed. Facies have sharp lower contacts whereas contacts to upper facies are gradual. Facies are observed overlying large-scale cross-bedded and small-scale asymmetric cross-bedded sandstones (facies A and B) and overlain by horizontally bedded sandstones (facies F). Observations are mostly made in the lower, sandstone-rich intervals on Svartnosa, Barentsøya, but it can also be found in the uppermost part on Wilhelmøya. Noted unit thickness is about 0.5 m.

It is commonly found together with facies large-scale cross-bedded (facies A) and small-scale asymmetric cross-bedded sandstones (facies B) and horizontally bedded sandstones (facies F).

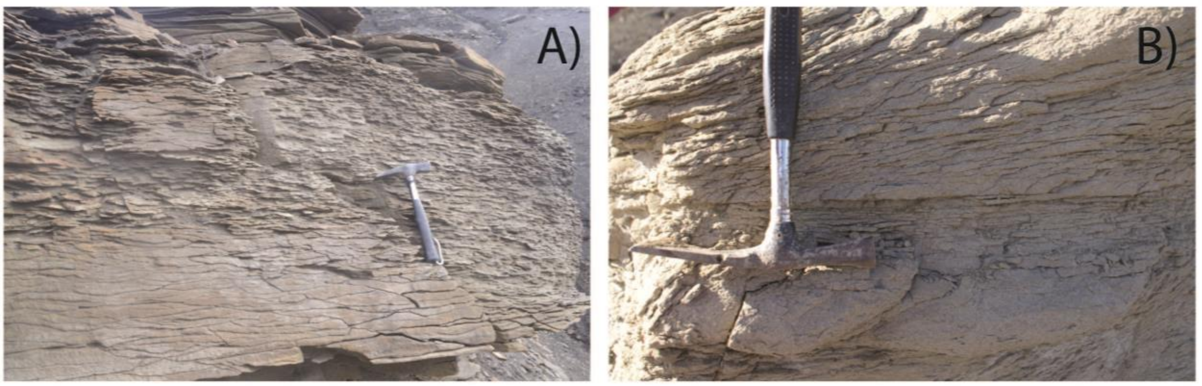

Figure 5.3: *Facies C - Climbing ripple cross-laminated sandstone.* **A)** Climbing ripple cross-lamination in the lower part of De Geerdalen Formation on Svartnosa, Barentsøya **B)** Climbing ripples towards the top of Wilhelmøya.

Interpretation

The formation of climbing-ripple cross-lamination take place as high sediment supply leads to aggradation of ripples with contemporary downstream migration, with the angle of climb reflecting rate of aggradation (Collinson et al., 2006). Environments characterized by periodic rapid deposition, especially sands from suspension, is favourable for this facies formation, whereas environments of low sedimentation rates and much reworking is not (Reineck and Singh, 1980). Fluvial floodplains, with their sub-environments crevasse splays and point bars, and seasonally flooded river deltas are environments where climbing ripple cross-laminated sandstones occur (Reading and Collinson, 1996; Boggs, 2011).

Herein this facies is interpreted to 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.

Facies D - Wave rippled sandstone

Description

This facies is assigned to units of very fine to fine sandstone that is ripple cross-laminated (Fig. 5.4). 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. 5.4A, 5.4B, 5.4D). In cross-section, wave ripples are recognized by having undulating bedding, sometimes with bidirectional foresets (Figs. 5.4C, 5.4E, 5.4F).

Mud drapes are also common within the facies and helps tracing out foreset features. The facies are usually moderately to intensely bioturbated and trace fossils such as *Rhizocorallium* and *Skolithos* are often found associated with this facies.

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

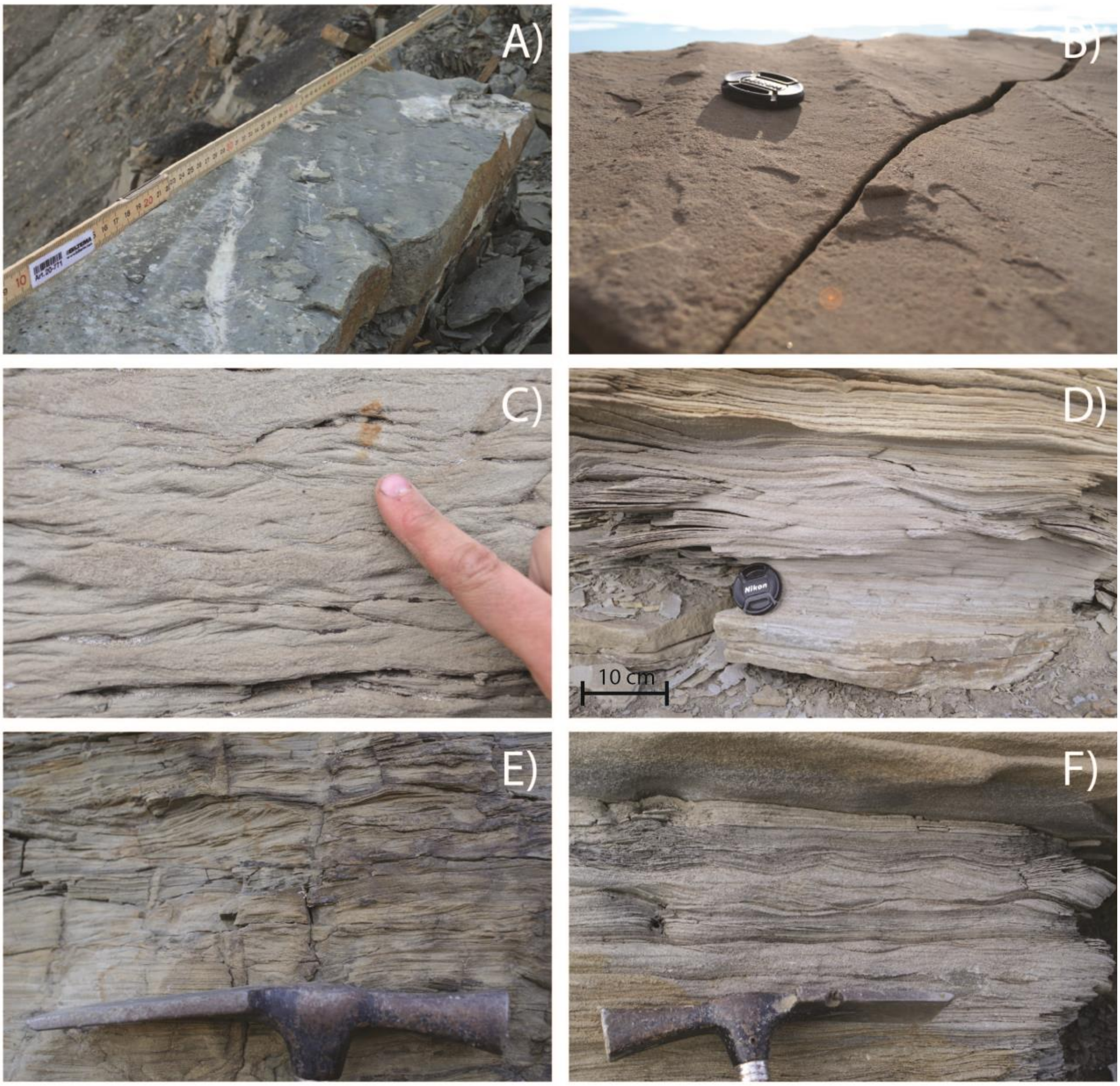

Figure 5.4: *Facies D - Wave rippled sandstone* **A)** Straight symmetrical ripples preserved in the upper part of the De Geerdalen Formation on Klement'evfjellet, Agardhbukta **B)** Peak-shaped ripple crests on a block of sandstone, Hellwaldfjellet, north-eastern Spitsbergen **C)** Bidirectional features interpreted as wave ripple lamination and with a direction of wave propagation towards left and right, Friedrichfjellet, Agardhbukta **D)** Symmetrical ripple crests seen from the side on Hellwaldfjellet, north-eastern Spitsbergen. **E)** Cross-section of interwoven sets of wave ripples, Klement'evfjellet, Agardhbukta **F)** Symmetrical ripples on Hahnfjella, eastern Spitsbergen.

Interpretation

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

Wave energy is considered the most important marine process in governing coastline

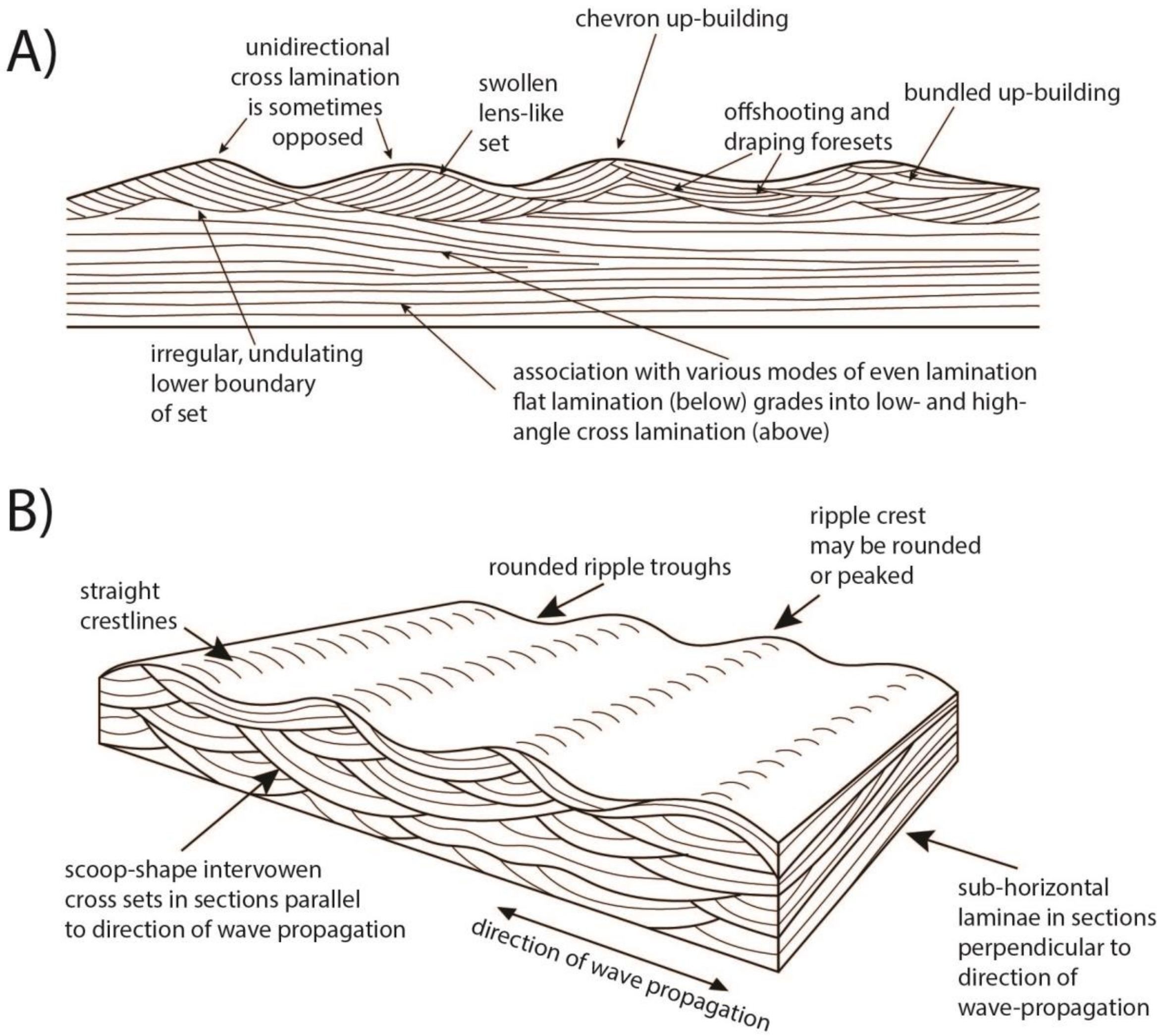

Figure 5.5: Characteristic features of ripples formed by the oscillatory movement generated by ocean waves. **A)** Foresets can be unidirectional and sometimes opposed, but are most commonly bidirectional (from Collinson et al. (2006) after de Raaf et al., 1977) **B)** Three-dimensional idealized block of lamination types that result from the bidirectional movements of water caused by waves (from Collinson et al. (2006) after Boersma, 1970)

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 paleo-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. 5.5A)(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. 5.5B)(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.

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. 5.6).The colour is usually grey to red-brown when weathered and grey on fresh surfaces. Unit thickness are usually between tens of cm to 1.5 m, while set thickness range between 5 and 15 cm. Individual sets are composed of both beds and lamina, where the former is the most common. These sandstones are commonly bioturbated and contain plant fragments and fish remains were found within low angle cross-stratified sandstone on Klement'evfjellet. This structure is commonly well developed on the Svartnosa locality (Figs. 5.6A, 5.6B, 5.6E), while in other locations it can appear more subtle and harder to recognize in the field (Figs. 5.6C, 5.6D, 5.6F).

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

Interpretation

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

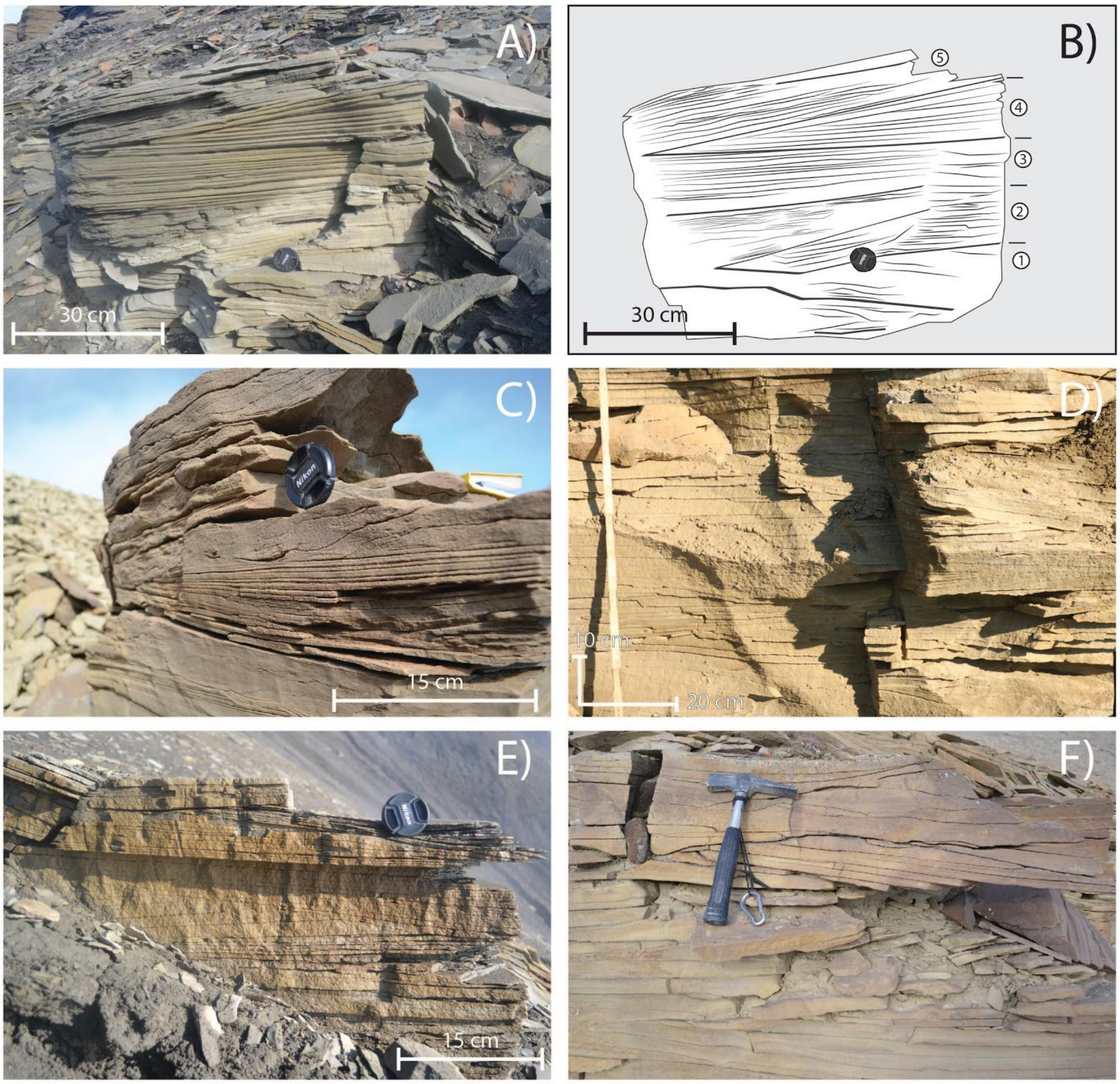

Figure 5.6: *Facies E - Low angle cross-stratified sandstone.* **A)** Low angle cross-stratification, Svartnosa, Barentsøya. **B)** Sketch of low-angle cross stratification with the individual cross-sets labelled. **C)** Oblique view of low angle cross stratified sandstone from the succession on Hahnfjella, 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.

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. 5.7). 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. 5.7A) and cm-thick beds (Figs. 5.7B, 5.7C, 5.7D).Transitions between lamination and bedding occur within units (Fig. 5.7D). Parting lineation, also known as primary current lineation (PCL), is present on upper bedding surfaces within planar parallel stratified sand intervals.

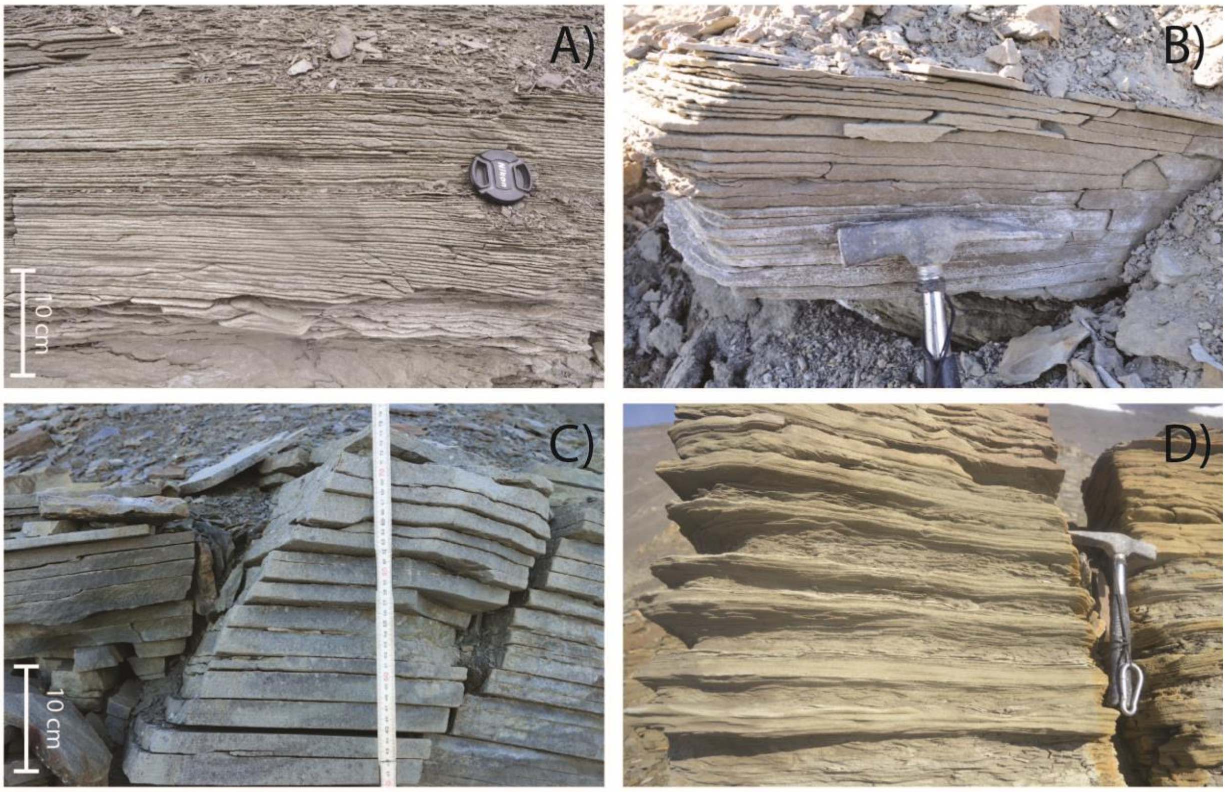

Figure 5.7: *Facies F - Horizontally bedded and planar stratified sandstones.* **A)** Decreasing bedding thickness upwards into fine laminae, Mistakodden, Barentsøya. **B)** Horizontally bedded sandstone on Wilhelmøya. **C)** Centimetre thick layers of horizontally bedded sandstone Klement’evfjellet, Agardhbukta. **D)** Rhythmic alternations of facies F and B (weathered) capped by wave rippled sandstones (facies D) with *Skolithos* traces on Svartnosa, Barentsøya.

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

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

Interpretation

Horizontally laminated bedding occurs 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.

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. 5.8). 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. 5.8A), 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 seem less abundant in our study area compared to theirs.

Figure 5.8: *Facies G - Massive, structureless sandstone.* **A)** A large block of massive, structureless sandstone as exposed on Hellwaldfjellet, north-eastern Spitsbergen. **B)** Apparently massive sandstone, Wilhelmøya.

Interpretation

Apparent lack of sedimentary structures is interpreted to be mainly the result of weathering and diagenesis, primarily calcite cementation. In some cases, the lack of primary structures can also be attributed to intense bioturbation and biogenic reworking of sediments during deposition. Liquefaction and flow of waterlogged sediments can also result in the destruction of primary sedimentary structures (Collinson et al., 2006). Massive, structureless bedding may also be the product of very rapid deposition or from liquefaction of sediment due to a sudden shock following deposition (Boggs, 2011).

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. 5.9). 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.

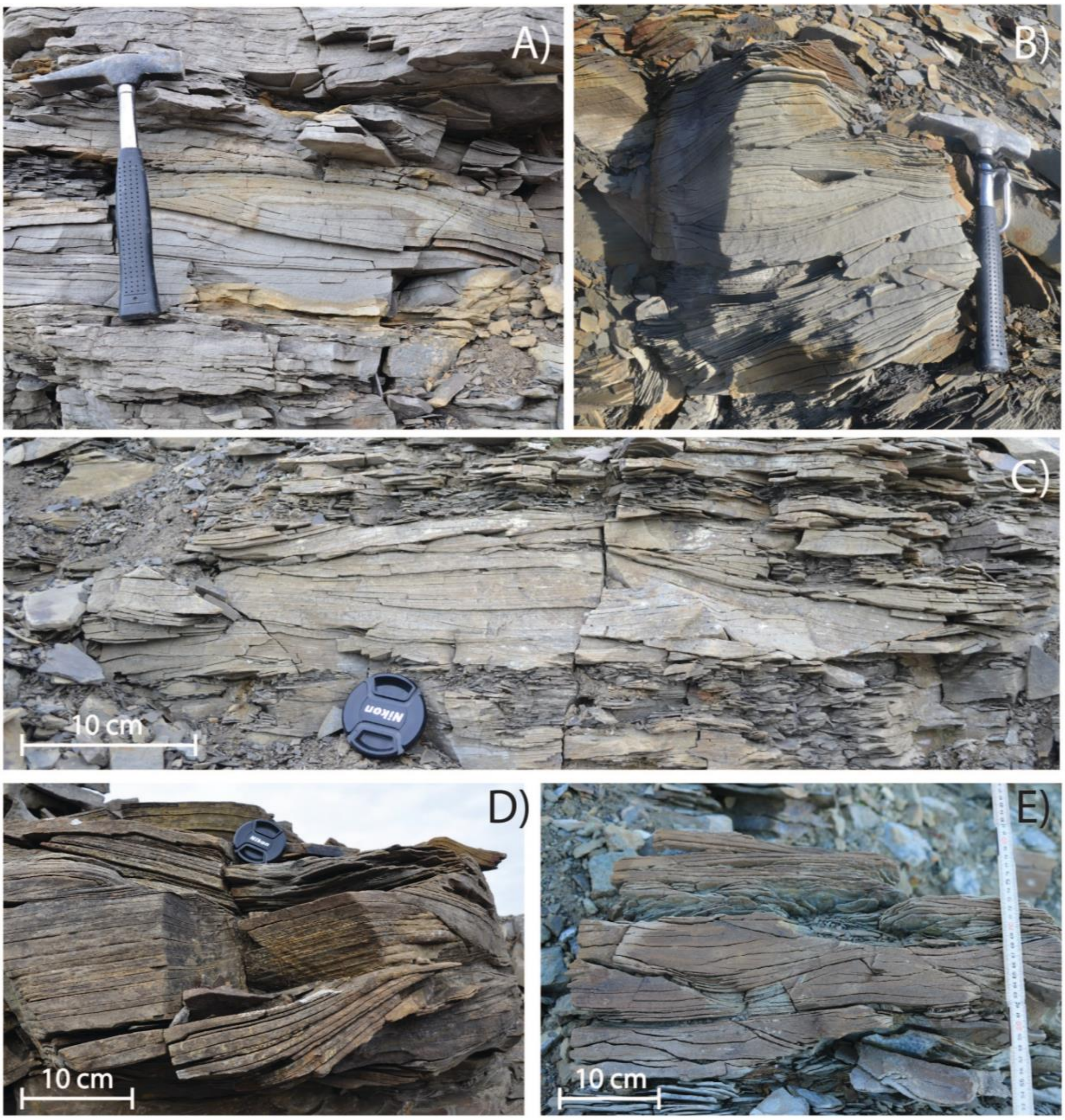

Figure 5.9: *Facies H - Hummocky and swaley cross-stratified sandstone* **A)**"Micro-hummock" is a term applied to smaller scale hummocky cross-stratification. The small scale suggests a more proximal position close to the lower shoreface, Friedrichfjellet, Agardhbukta **B)** Excellent three-dimensional structure of hummocky and swaley cross-stratification exposed on Svartnosa, Barentsøya **C)** Small-scale hummocky cross-bedding on Friedrichfjellet, Agardhbukta. **D)** Good block exposure of the internal structures in hummocky cross-bedding from Friedrichfjellet, Agardhbukta **E)** Cross-sectional view of dipping foresets, Klement'evfjellet, Agardhbukta.

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

Hummocky to swaley cross-stratified sandstones are often found below facies D (Wave rippled sandstone) and commonly together with facies F (horizontally bedded sand), facies M (Mudrocks), facies K (Heterolithic bedding) and facies E (Low angle cross-stratified sandstone). Hummocky cross-stratified sandstones are common in upwards coarsening sequences in the lower part of the De Geerdalen Formation throughout the study area. It is found frequently in the Agardhbukta outcrops (Figs. 5.9A, 5.9C, 5.9D, 5.9E). It is also common in coarsening upwards sequences on Svartnosa.

Interpretation

Hummocky cross-stratification is a sedimentary structure in sandstones that show a distinct undulating geometry of lamination. The exact formation of hummocky cross-stratification remains enigmatic despite being the subject of numerous studies (Harms et al., 1975; Harms et al., 1982; Dumas and Arnott, 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; Johnson and Baldwin, 1996; Midtgaard, 1996), forming below fair weather wave base and above, but near storm weather wave base (Dumas and Arnott, 2006). 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 (Fig. 5.10) of hummocky cross-stratified sandstones are interpreted to form by the migration of low-relief bed forms in primarily one direction due to the influence of combined wave surge and unidirectional currents (Nøttvedt and Kreisa, 1987; Walker and Plint, 1992).

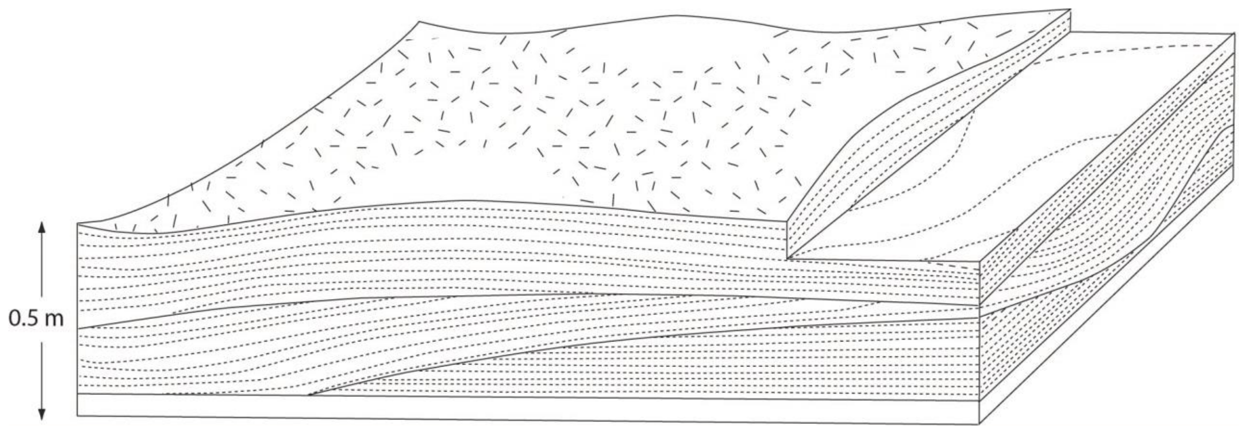

Figure 5.10: Internal geometries and bed form of hummocky cross-stratification (from Quin, 2011, after Harms et al., 1975).

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 (Fig. 5.9) (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 favour the preservation of swales (Walker and Plint, 1992; Hampson and Storms, 2003; Dumas and Arnott, 2006).

Facies I - Soft sediment deformed sandstones

Sub-Facies I_1 - syn-sedimentary deformed sandstones

Description

Very fine to fine sandstones displaying soft sediment deformation structures (Fig. 5.11). 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 (Fig. 5.11B)

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

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).

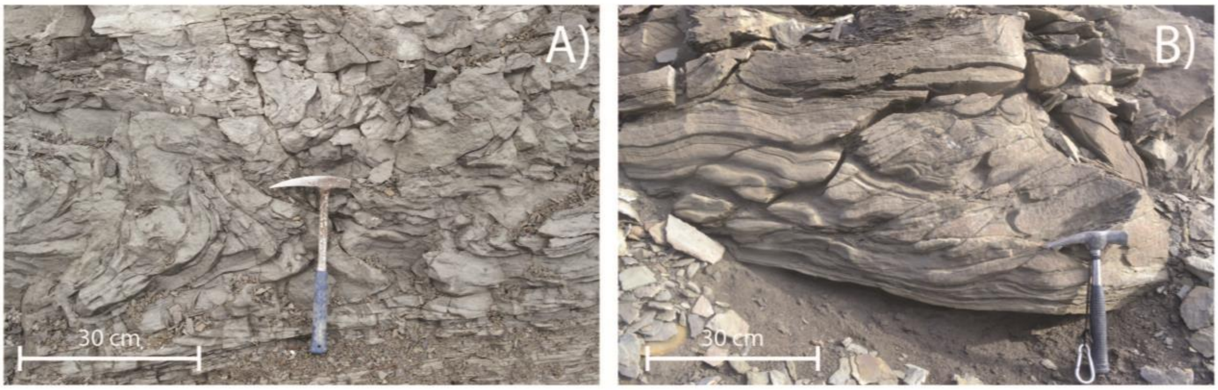

Figure 5.11: *Facies I_1 - Syn-sedimentary deformed sandstones* **A)** Heavily deformed sandstone on Mistakodden, Barentsøya. **B)** Characteristic features are the sharp lower contact and the disturbed internal laminations. From Svartnosa, Barentsøya.

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. 5.12). 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 5.12: *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. 5.13) (Reineck and Singh, 1980; Bhattacharya and MacEachern, 2009).

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

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

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

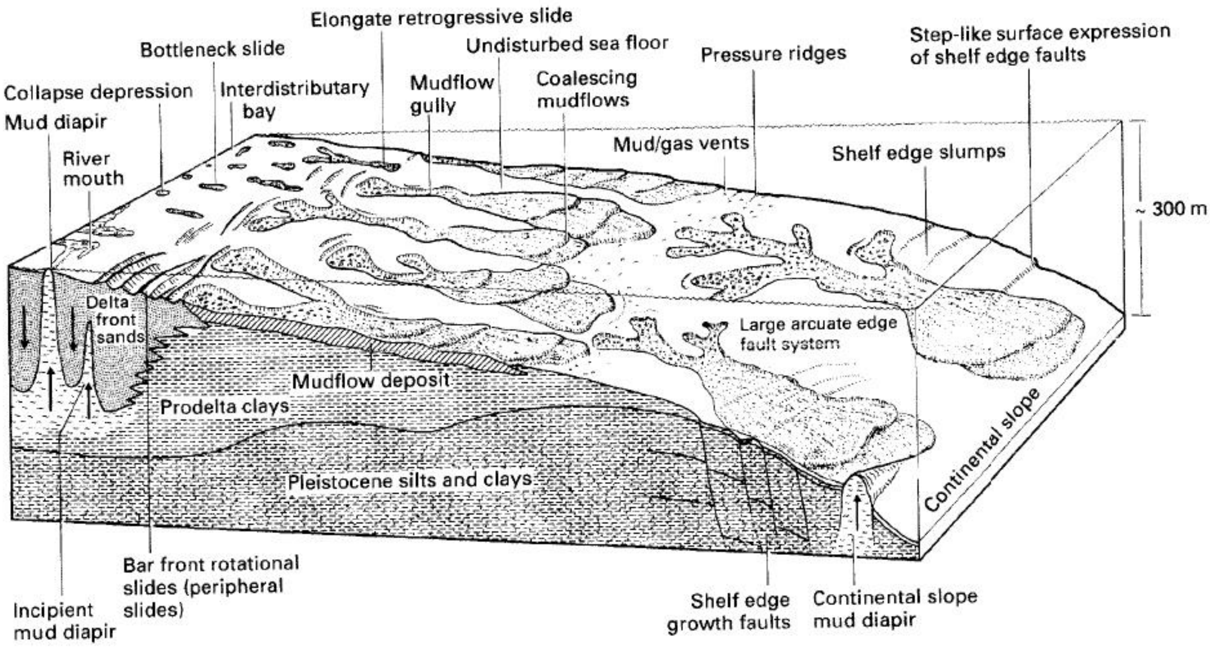

Figure 5.13: Different types of deformation caused by mass-movement of sediment supplied to the delta front and prodelta (from Reading and Collinson, 1996).

Facies J - Carbonate rich sandstone

Description

This facies comprise very fine to fine sandstones characterized by structures formed during diagenesis (Fig. 5.14). The sandstone units are commonly hard and heavily cemented, which makes observations of primary sedimentary structures difficult (Fig. 5.14C). Secondary sedimentary structures includes cone-in-cone (Fig. 5.14A, 5.14B), siderite beds (Fig. 5.14F) and calcareous concretions.

A decrease in cementation to adjacent facies is noticed as they often appear less consolidated. Colour variation between grey, brown and red are observed. Scarce to heavy bioturbation is noticed. Large calcite concretions, up to meter sized, are found on numerous locations. Facies can appear in close association with other facies forming a sandstone bench or be interbedded by mud. These sandstone benches commonly form very distinct layers that may be laterally continuous for several tens to hundreds of metres before they pinch out (Fig. 5.14D).

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. 5.14E, 5.14G). 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 is restricted to three discrete levels in Agardhbukta, where the uppermost may be the Slottet Bed. Similar units are found in the upper parts of Hellwaldfjellet and Wilhelmøya sections.

Siderite occurs as nodules within mudstones and sandstones grouped in other facies, or as nodules forming distinct layers (Fig. 5.14F). 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.

Interpretation

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

Figure 5.14: *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.

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). Knarud (1980) interpreted the occurrence of siderite nodules and layers as a late diagenetic phenomenon, most likely derived from calcite or aragonite.

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).

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. 5.15). The thickness of mud- and sand layers ranges from 1 mm to few centimetres, while units are up to 9 m thick (Fig. 5.15B). 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 seems to be more dominant in the Agardhbukta sections, while lenticular-, wavy-, and flaser-bedding is more abundant in the middle to upper parts of the De Geerdalen Formation.

Interpretation

Heterolithic bedding indicates alternating flow regime in an environment with both sand and mud available (Figs. 5.15, 5.16) (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).

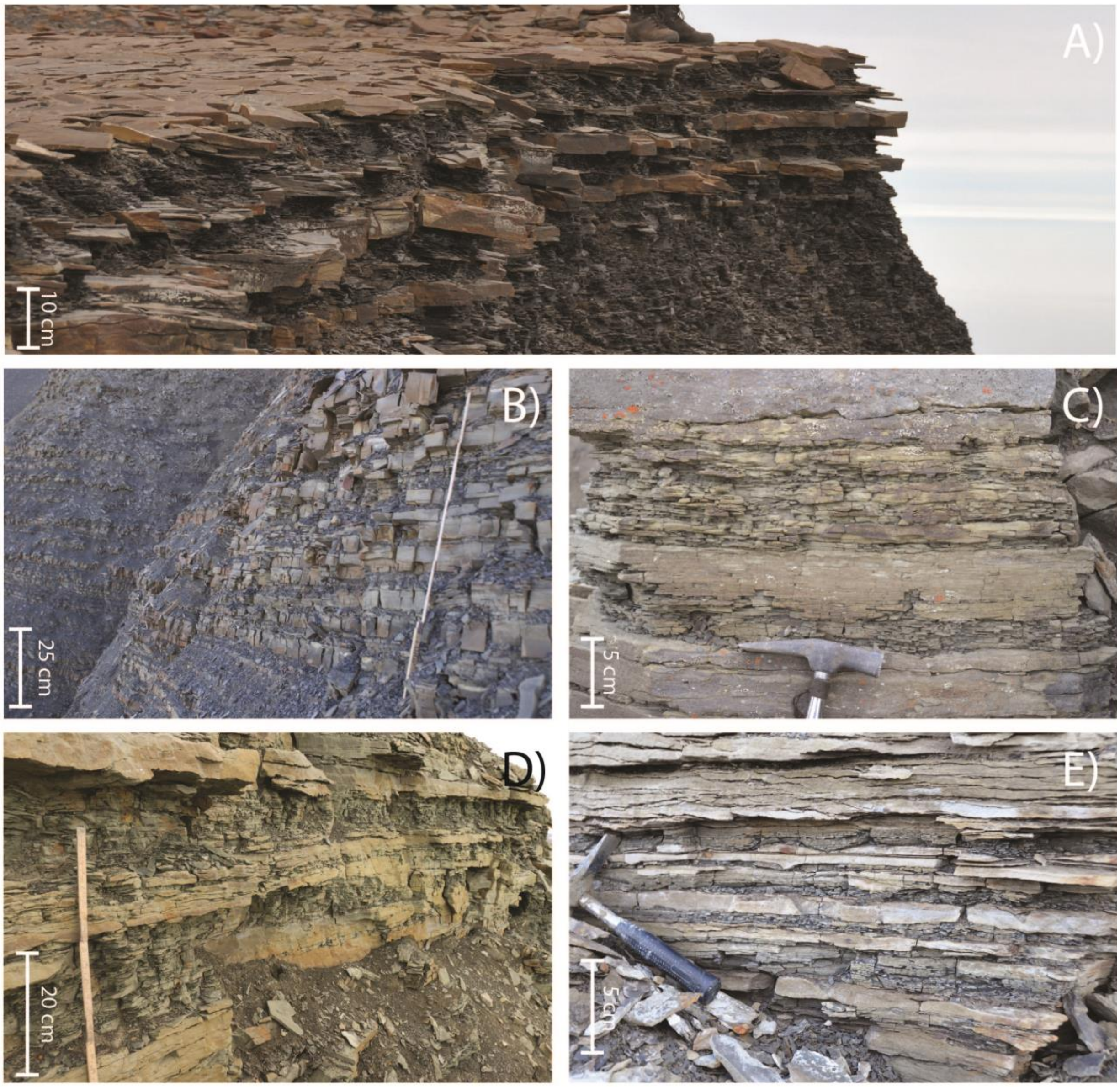

Figure 5.15: *Facies K - Heterolithic bedding*. Note the different scales of sandstone bed thickness. **A)** Heterolithic bedding in the lower part of the De Geerdalen at Muen, Edgeøya. **B)** Heterolithic bedding, interpreted as storm deposited sandstones, in the lower part of the succession at Klement'evfjellet, 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.

The distinction between flaser, wavy and lenticular bedding are based on mud/sand ratio and lateral continuity on the sand and mud layers. Wavy bedding has approximately equal amount of sand and mud. Flaser bedding contains more sand than mud, and lenticular bedding in turn has more mud than sand. The different types of heterolithic bedding reflect which grain size was favoured to be deposited and preserved in the 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).

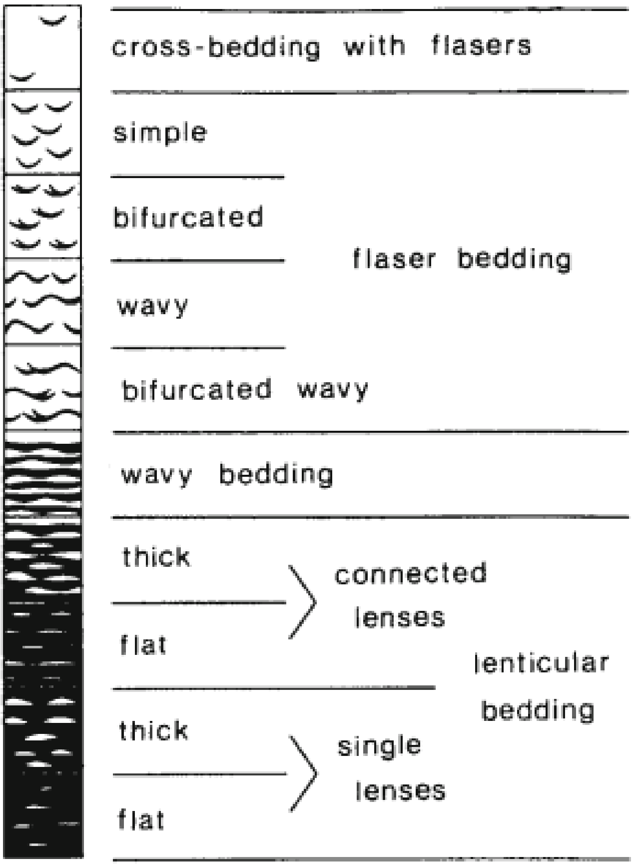

Figure 5.16: Flaser, wavy and lenticular bedding are determined based on the ratio of sand and mud. High sand-mud ratios favours the generation of flaser bedding, while wavy and lenticular bedding are characterized by an equal and low sand - mud ratio (from Reineck and Wunderlich, 1968).

Heterolithic bedding may form in the offshore transition zone when storm-transported sands interact with mudrocks deposited from suspension during fair weather (Figs. 5.15A, 5.15B). 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 is based on associated facies, and the scale, thickness and lateral continuity of the sand- and mud layers. The facies could also possibly originate from other environments undergoing alternating flow regime.

Facies L - Coquina beds

Description

The unit consists mainly of matrix-supported, fragmented bivalves, lacking sedimentary structures (Fig. 5.17). 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 in the modern Mississippi delta (Frazier, 1967).

The Isfjorden Member is interpreted to be deposited in a shallow marine setting with possibly local lagoonal environments (Mørk et al., 1999a). 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.

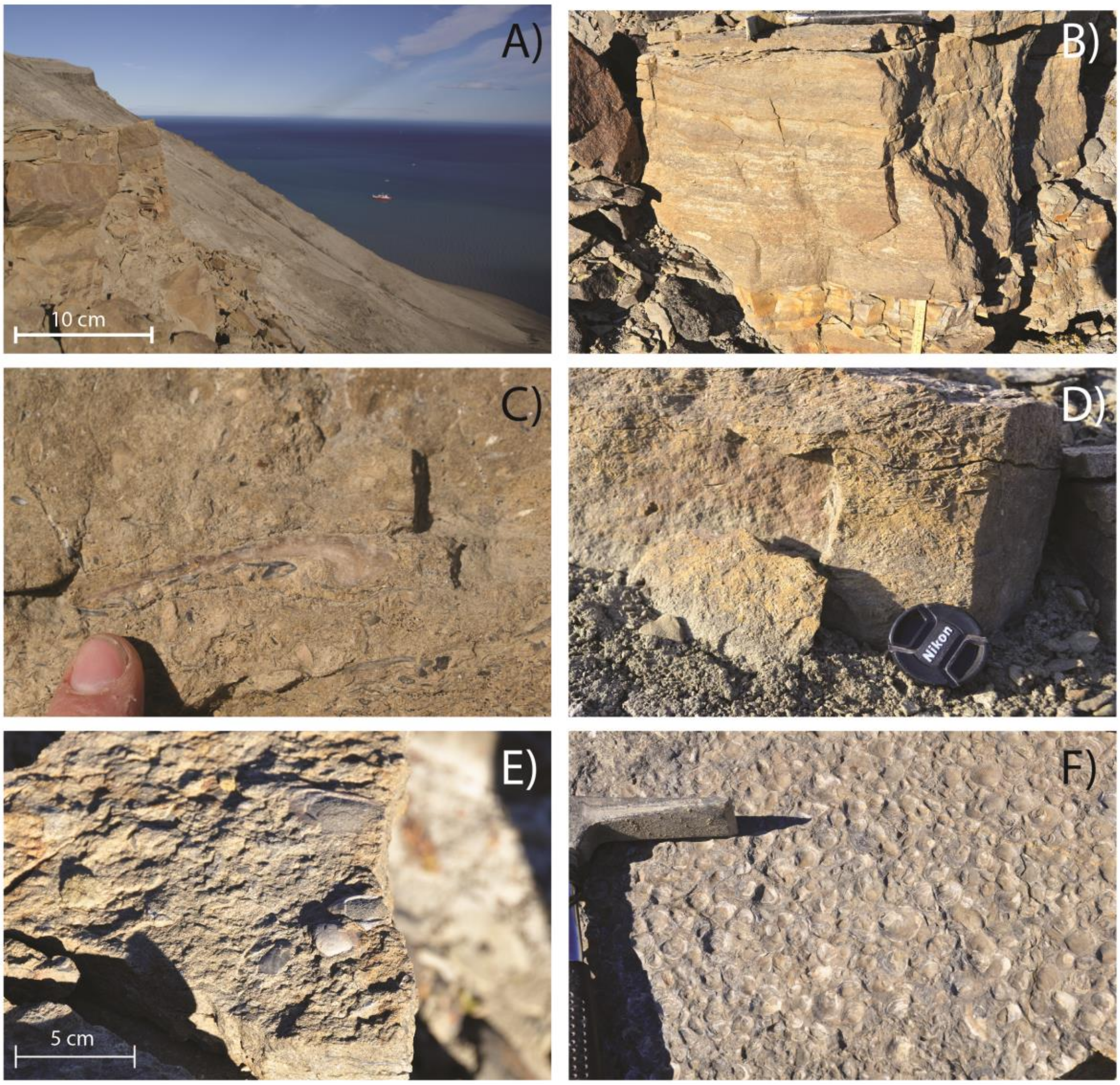

Figure 5.17: *Facies M - Coquina beds.* **A)** Laterally continuous coquina shell bank below the dolerite sill on Tumlingodden, Wilhelmøya. **B)** Coquina bed in a sandstone unit on Tumlingodden, Wilhelmøya. **C)** Bivalve shell fragment in a coquina bed on Tumlingodden, Wilhelmøya. **D)** Numerous fragmented 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.

Facies M - Mudrocks

Description

This facies is used to describe successions consisting of fine-grained material (clay to silt) that may be either laminated (shale) or non-laminated (mudrock), following the proposed nomenclature of Lundegard and Samuels (1980). Mudrocks constitute the bulk of the succession at essentially every location, although they are often covered by scree, as they are more susceptible to weathering and erosion. This makes detailed field studies both difficult and time-consuming. Scree-covered areas are often inferred to be composed of mud successions.

The mudrocks facies varies greatly in thickness throughout the study area from the scale of tens of meters to a few centimetres in heterolithic successions. Laminated mudrocks are most common and may encase thin beds of silty to very fine sandstone. The colour of mudrocks is dominantly grey or black, but may also be of yellow, white or purple colour (Figs. 5.18).

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.

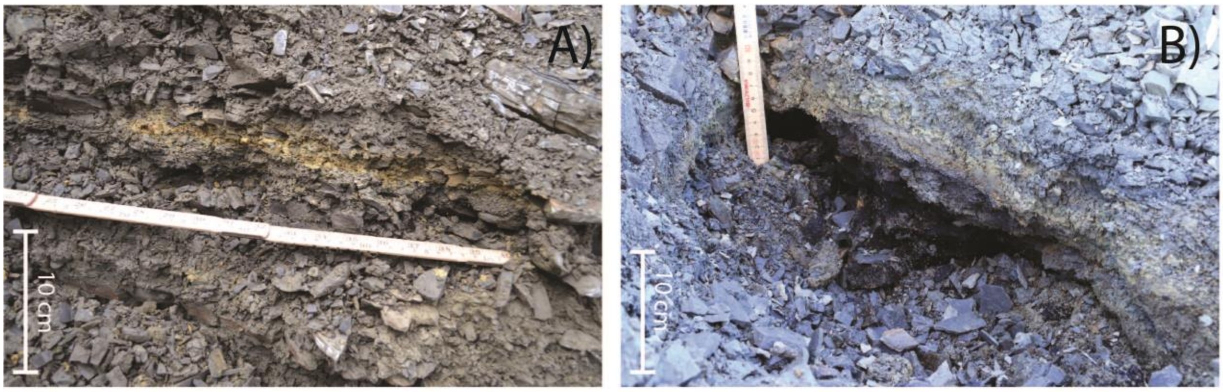

Figure 5.18: *Facies M - Mudrocks* **A)** Grey and yellow mudrocks on Friedrichfjellet, Agardhbukta. **B)** Klement'evfjellet, Agardhbukta.

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 marginal-marine delta front settings (Aplin and Macquaker, 2011). Mudrocks are also a dominant lithology in coastal environments, such as lagoons, tidal flats, interdistributary 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 (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. 5.18A) 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).

Facies N - Coal and coal shale

Description

Units of coal and coal shale are from 1 to 20 cm thick (Fig. 5.19). The units often appear laterally continuous over tens of meters when examined, but scree cover is common (Figs. 19A, 19C). 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. 5.19B), 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.

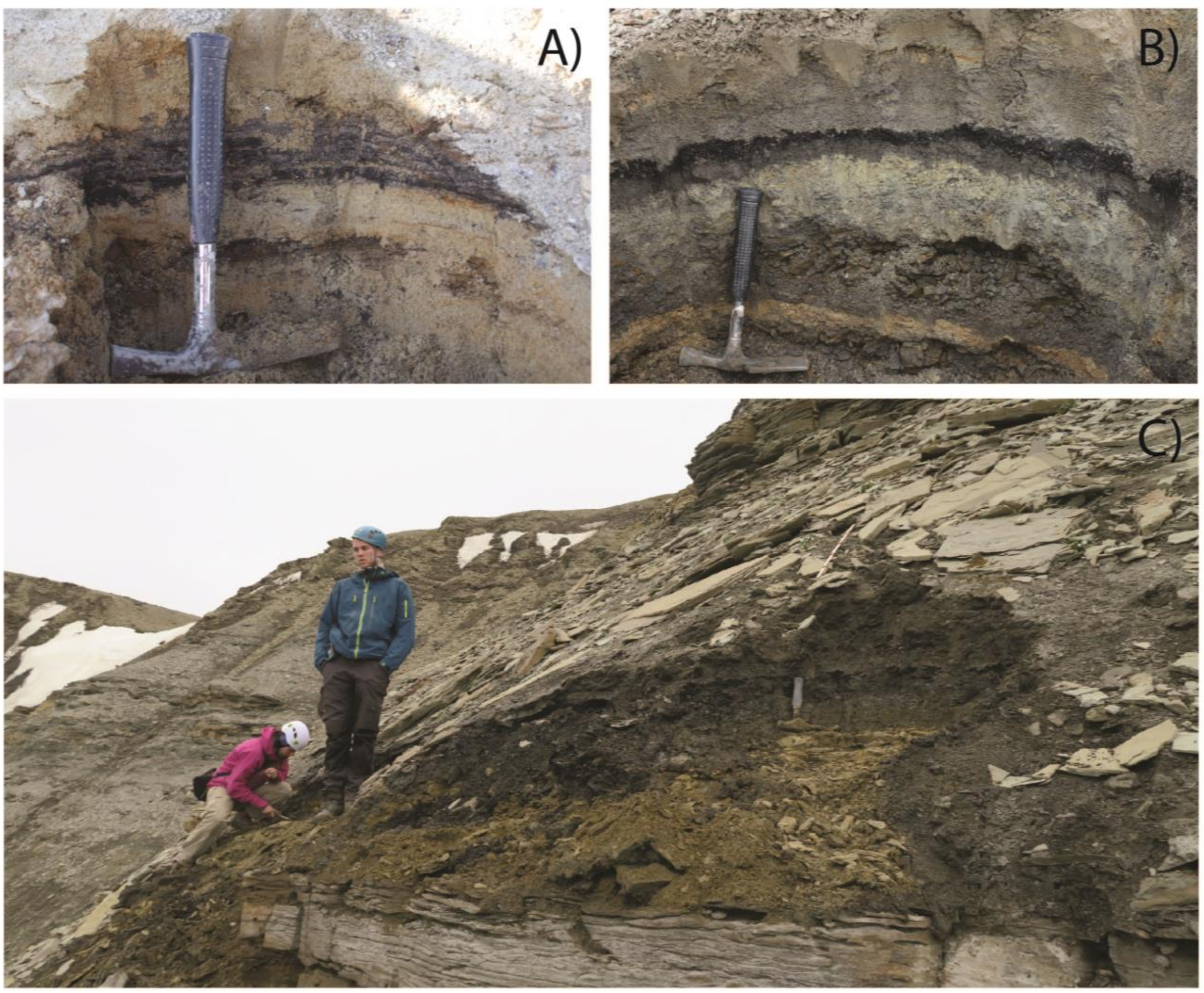

Figure 5.19: *Facies N - Coal and coal shale* **A)** Two laterally discontinuous coal seams alternating with paleosols, Tumlingodden, Wilhelmøya. **B)** Coal shale overlying bleached paleosol, Šmidtberget, Agardhbukta. **C)** Laterally continuous coal and coal shale in the uppermost part of sediments 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. 5.20). Swamps and mires are the most common coal forming environment and is recognized by vegetation growing in anoxic, waterlogged ground (McCabe, 1984; Retallack, 1991).

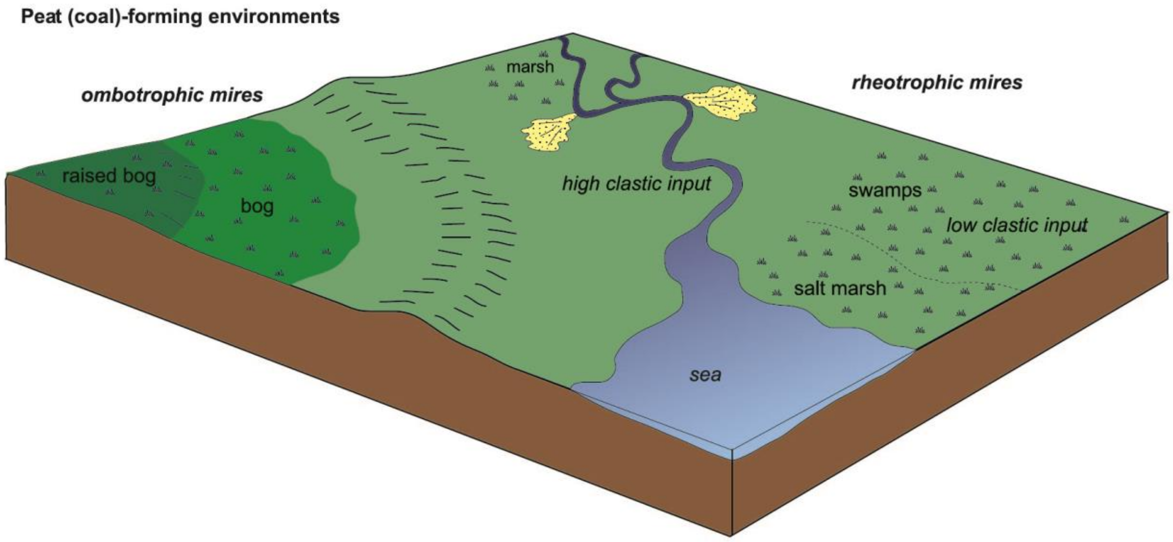

Figure 5.20: Coal-forming environments (Nichols, 2009).

Most of the coal seams found in the De Geerdalen Formation is 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 favour coal forming is required to form thick coal layers (Boggs, 2011). The coal and coal shale found in the De Geerdalen Formation are thin and discontinuous. Enga (2015) suggest 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).

Facies O - Paleosols

Sub-facies O_1 - Brown and yellow paleosol

Description

Paleosols are found at every locality visited except Muen, but is most common in the middle and upper parts of the De Geerdalen Formation under and in the Isfjorden Member (Fig. 5.21). 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. 5.21). At some of the outcrops a bleached yellow 10 to 50 cm thick layer occurs above a red or brown base (Fig. 5.21C).

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).

Interpretation

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 shows 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 Fig. 5.21B.

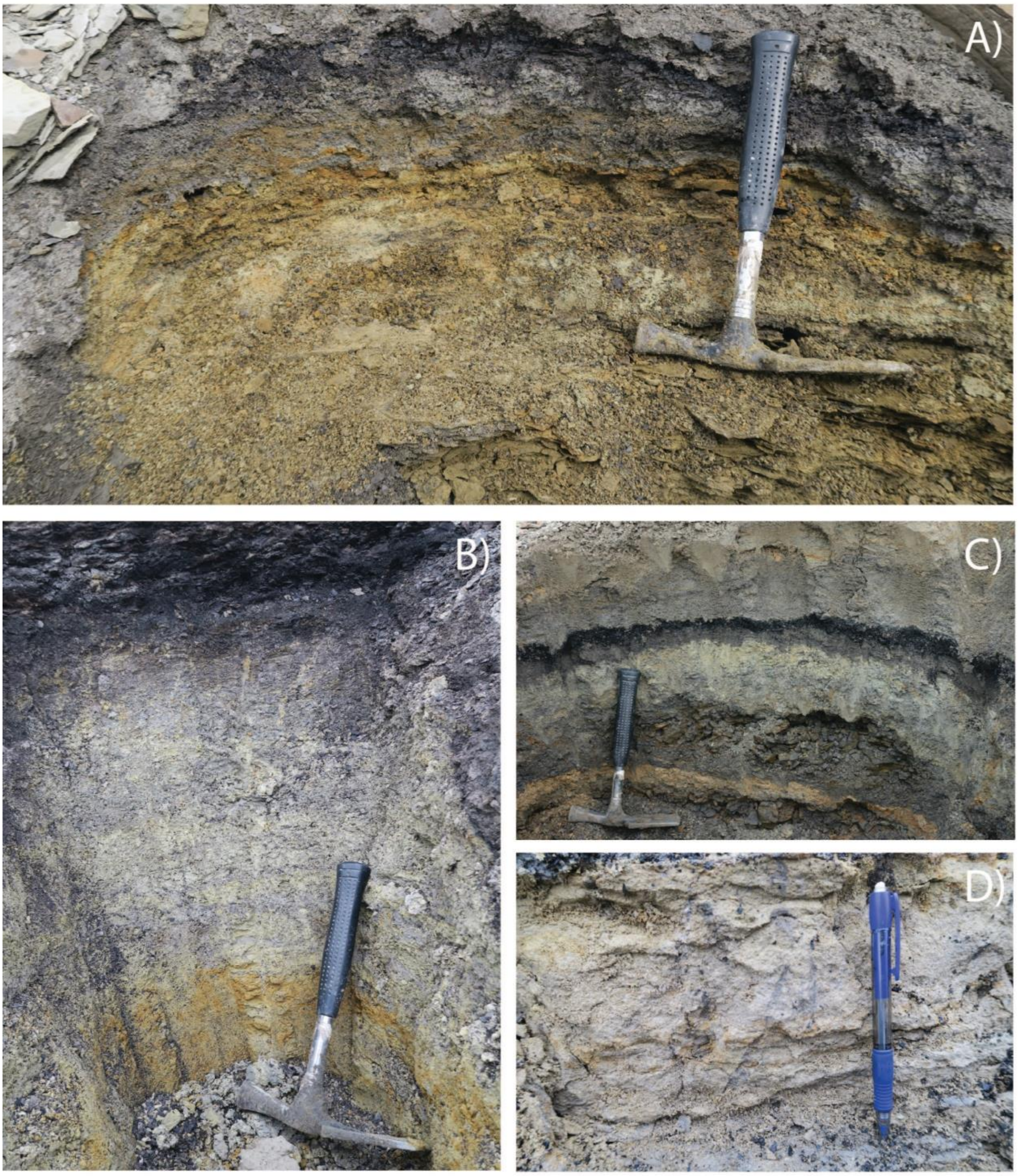

Figure 5.21: *Sub-facies O_1 - Brown and yellow paleosol.* **A)** Yellow, coarse grained paleosol, typical for proximal channel paleosols. The outcrop is overlying distributary channel deposits of Blanknuten, Edgeøya. **B)** Possible top of channel paleosol with coal shale (O-horizon) grading into a 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.

Paleosols can be used to reconstruct paleo-landscape. The paleo-catena 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. 5.22). In this model the soil horizons are termed A, B and C instead of the classification from Boggs (2011).

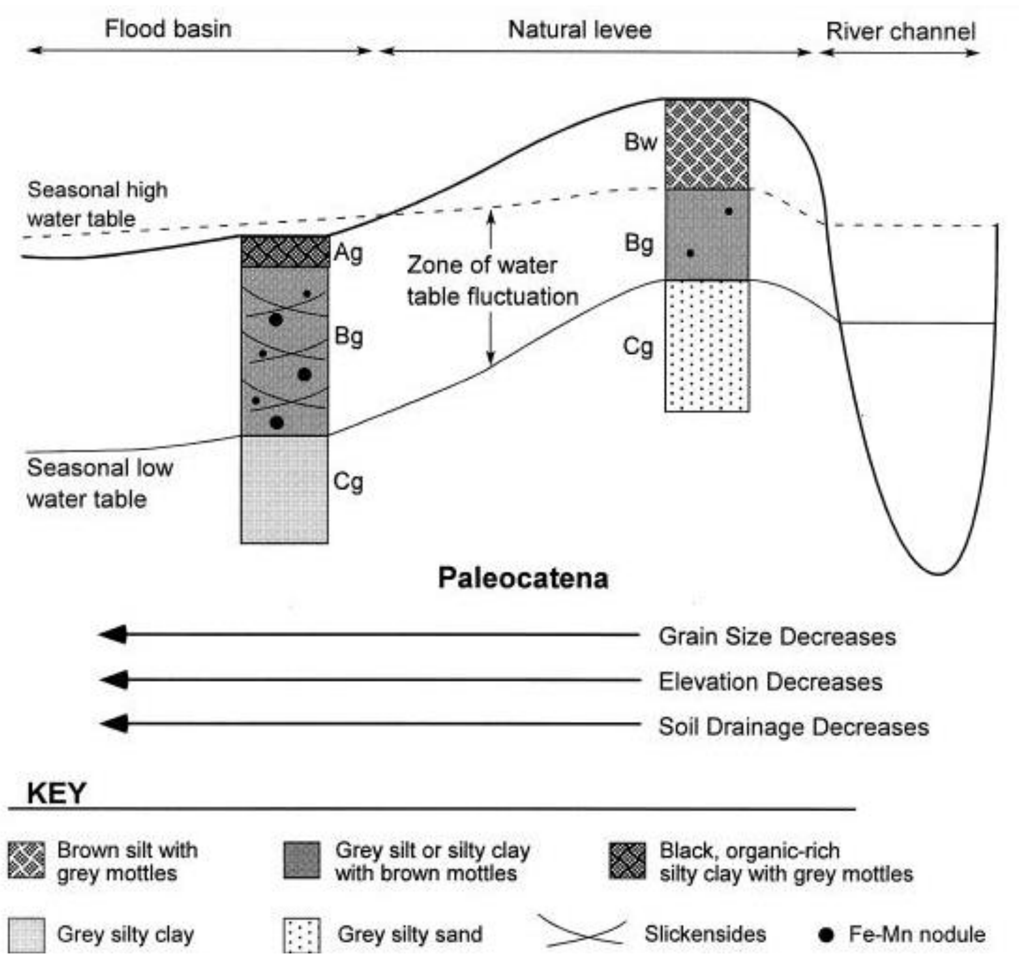

Figure 5.22: The paleo-catena model. Soils formed close to channels tend to be coarser grained and formed in oxidized conditions, leading to yellow to brown colour. Soils decrease in grain size away from the active channel and more of the profile is gleyed (Kraus and Aslan, 1999).

In general soils formed on crevasse splays and levees are well drained because they are elevated compared to the surroundings, and consist of relatively coarse grained material. The soil, and in particular the uppermost part, are usually formed in oxidized conditions, leading to yellowish to brown colour. In Fig. 5.22 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 grey colour. Groundwater saturated soils are called gleyed soils and is marked as Ag, Bg and Cg in Fig. 5.22 (Kraus and Aslan, 1999). An example of possible proximal channel paleosol with yellowish coarse grained material is seen on

Blanknuten (Fig. 5.21A). The interpretation is supported by the observation of distributary channel deposits of the underlying unit on Blanknuten

Soils tend to be less drained away from the channel because the soils consist of finer material and are located closer to the groundwater table due to the topographic position. This favour reduced conditions and gleyed soils in both A-, B- and C-horizons. Because of the reduced conditions organic matter can be accumulated and preserved in the A-horizon. Gleyed B and C horizons are typical for soils formed distal to channels (Kraus and Aslan, 1999). 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).

Sub-facies O_2 -Alternating red and green shales

Description

Alternating units of red and green shales (Fig. 5.23) is one of the most characteristic features in the Isfjorden Member (Pčelina 1983; Mørk et al. 1999a). 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.

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.

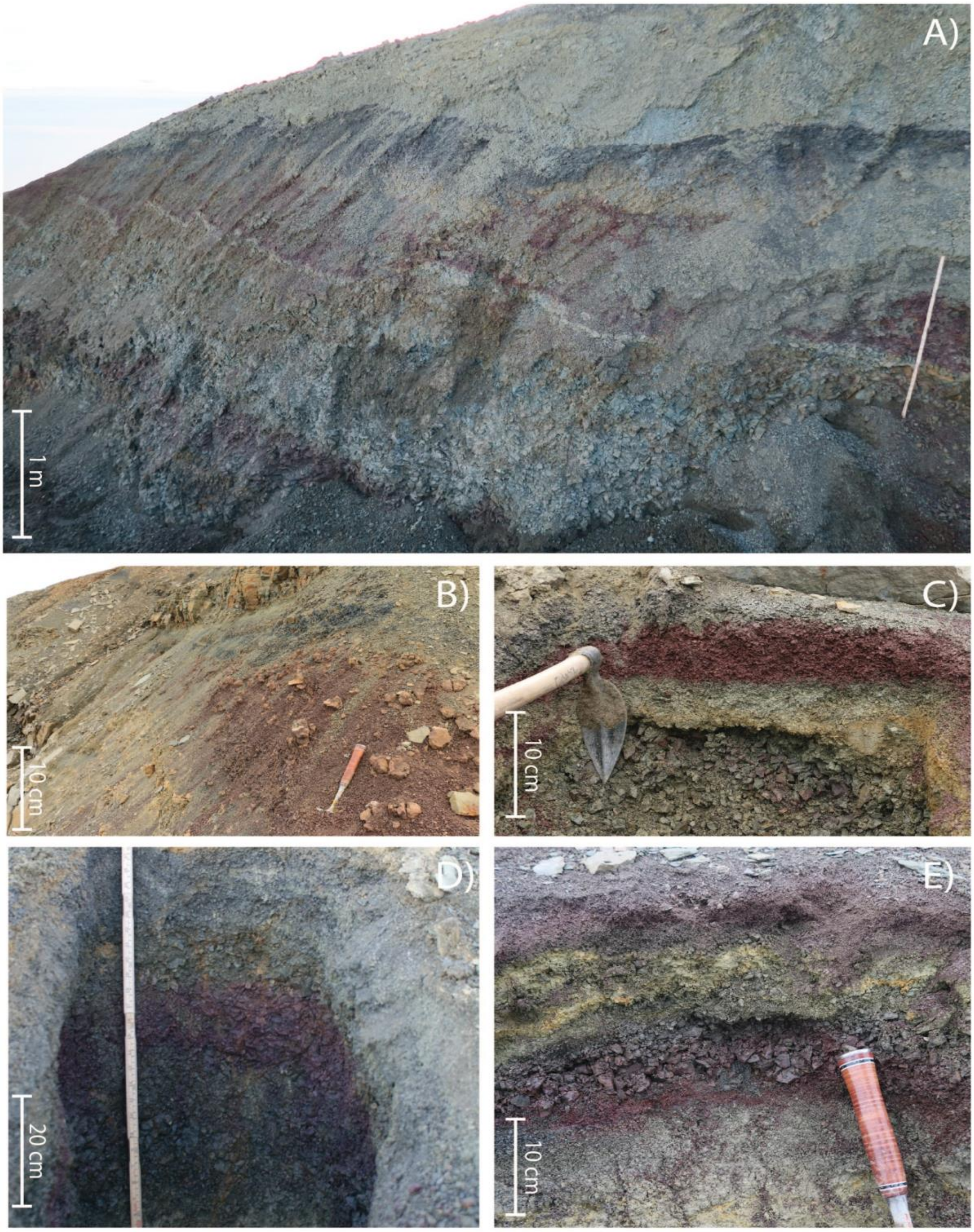

Figure 5.23: *Facies O_2 - Alternating red and green shales.* **A)** Overview photo of the Isfjorden Member on the slopes above the dolerite sill on Tumlingodden, Wilhelmøya **B)** Alternating red and green shales with larger nodules at discrete levels on top of Klement'evfjellet, Agardhbukta **C)** 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.

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 best preserved in former waterlogged ground. Traces of roots are recognized by irregular shaped features which are tapering and branching downwards (Figs. 5.23C, 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 is seen in Fig. 5.23E.

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 originate 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 Fig. 5.23B and 5.23C.

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 conditions have earlier been suggested for the Isfjorden Member (Mørk et al. 1999a, Mørk 2015).

Ichnofacies in the De Geerdalen Formation

Ichnology, the study of trace fossils, are considered a useful complementary indicator of paleo-ecological 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 paleo-bathymetry, as different trace fossils can occur in a wide range of depositional environments at different scales (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.

***Cruziana* ichnofacies**

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

***Skolithos* ichnofacies**

Within the De Geerdalen Formation on eastern Svalbard, trace fossils such as *Skolithos* and *Diplocraterion* were observed within several facies (Fig. 5.24). *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. 5.24), and in beach environments (Pemberton et al., 1992; Boggs, 2011).

On Svartnosa (Svart 15-1, ~ 106 m), abundant *Skolithos* was observed concentrated within a single sandstone bed (Fig. 5.25A). Sandstones with a high abundance of vertical tubes of *Skolithos* are commonly referred to as pipe rock (Droser, 1991). Pipe rock 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).

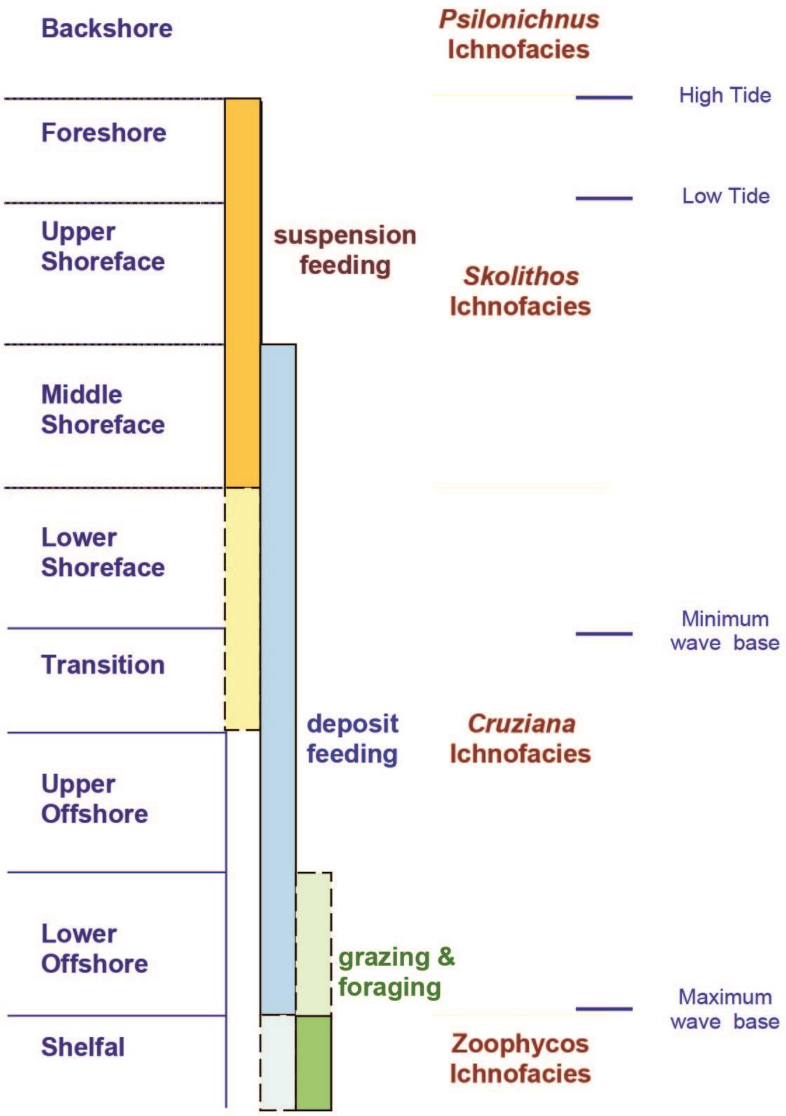

Figure 5.24: Distribution and types of ichnofacies on the shoreface. Sandstones in the De Geerdalen Formation exclusively contain marine trace fossils found in transitional offshore-shoreface environments, such as the *Cruziana* and *Skolithos* ichnofacies (Pemberton et al., 1992; Clifton, 2006).

Diplocraterion consists of vertical U-shaped burrows with spreite (Figs. 5.25G, 5.25H). It belongs to the *Skolithos* ichnofacies and is also commonly found in environments characterized by high energy waves and currents. The *Skolithos* ichnofacies commonly grade laterally seawards into the *Cruziana* ichnofacies and mixed *Skolithos*-*Cruziana* ichnofacies are known from both recent and ancient settings (Pemberton et al., 1992; Dalrymple and Choi, 2007).

In the study area, a sparse and low diversity of trace fossil assemblages is observed, except on Wilhelmøya where observations were made of *Rhizocorallium*, *Skolithos*, *Teichichnus* and *Diplocraterion* (Fig. 5.25C). 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).

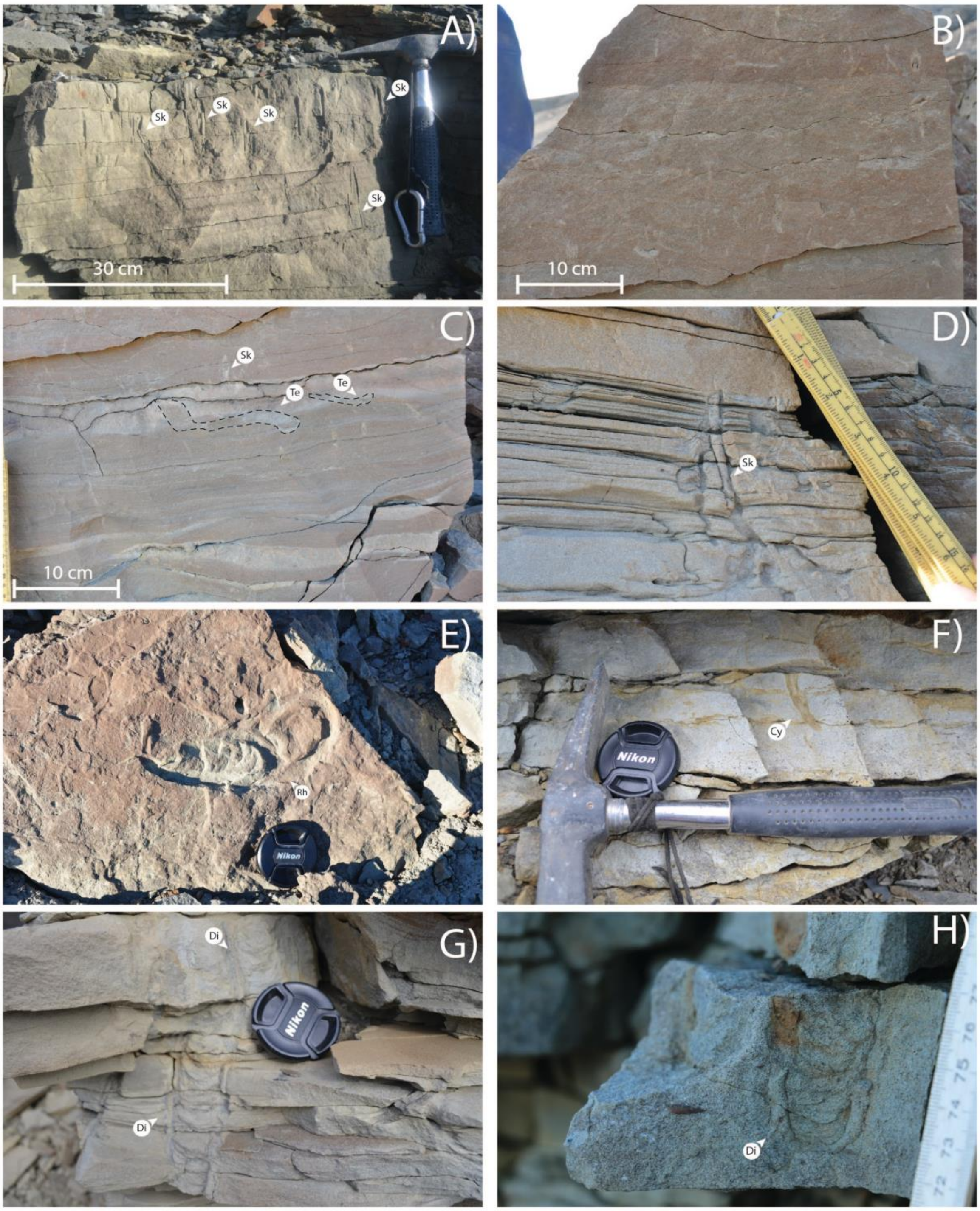

Figure 5.25: Trace fossils observed on eastern Svalbard. **A)** "*Skolithos* pipe rock" composed of vertical tubes of *Skolithos* (Sk), Svartnosa (Barentsøya). **B)** Completely bioturbated sandstone, Wilhelmøya. **C)** *Teichichnus* (Te) and *Skolithos* (Sk) in the upper part of De Geerdalen Formation 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.

6. 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 2, Fig. 6.1) and identified based on interpretation of facies, geometries and dimensions of sandstones, and other field observations. Previous studies in adjacent areas, conducted by Rød et al (2014), have served as a fundament and inspired the studies presented below.

Table 2 Facies associations in the De Geerdalen Formation on eastern Svalbard. Parenthesis indicates less prominent facies (modified from Rød et al., 2014).

Facies Association	Sub-facies association	Facies included	Characteristics
1. Lower shoreface to offshore deposits	1.1 Offshore	I, K, M	Long intervals of mudrocks interrupted by thin tabular sandstones.
	1.2 Offshore transition	D, H, I, K, M	Storm dominated sandstones interacting with mudrocks deposited from suspension.
	1.3 Lower shoreface	(B), D, F, H, K, M	Fine fair weather 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	Fluvial dominated very fine to fine sandstones showing influence of basinal processes.
	2.3 Distributary channel	A, B, (D), (G), K	Erosive-based fining upwards sandstones. Occur as laterally restricted or as laterally extensive, amalgamated channel deposits.
3. Delta plain	3.1 Floodplain with crevasse splay	(A), B, C, D, F, K, M, N, O	Delta plain deposits related to flooding of interdistributary 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.

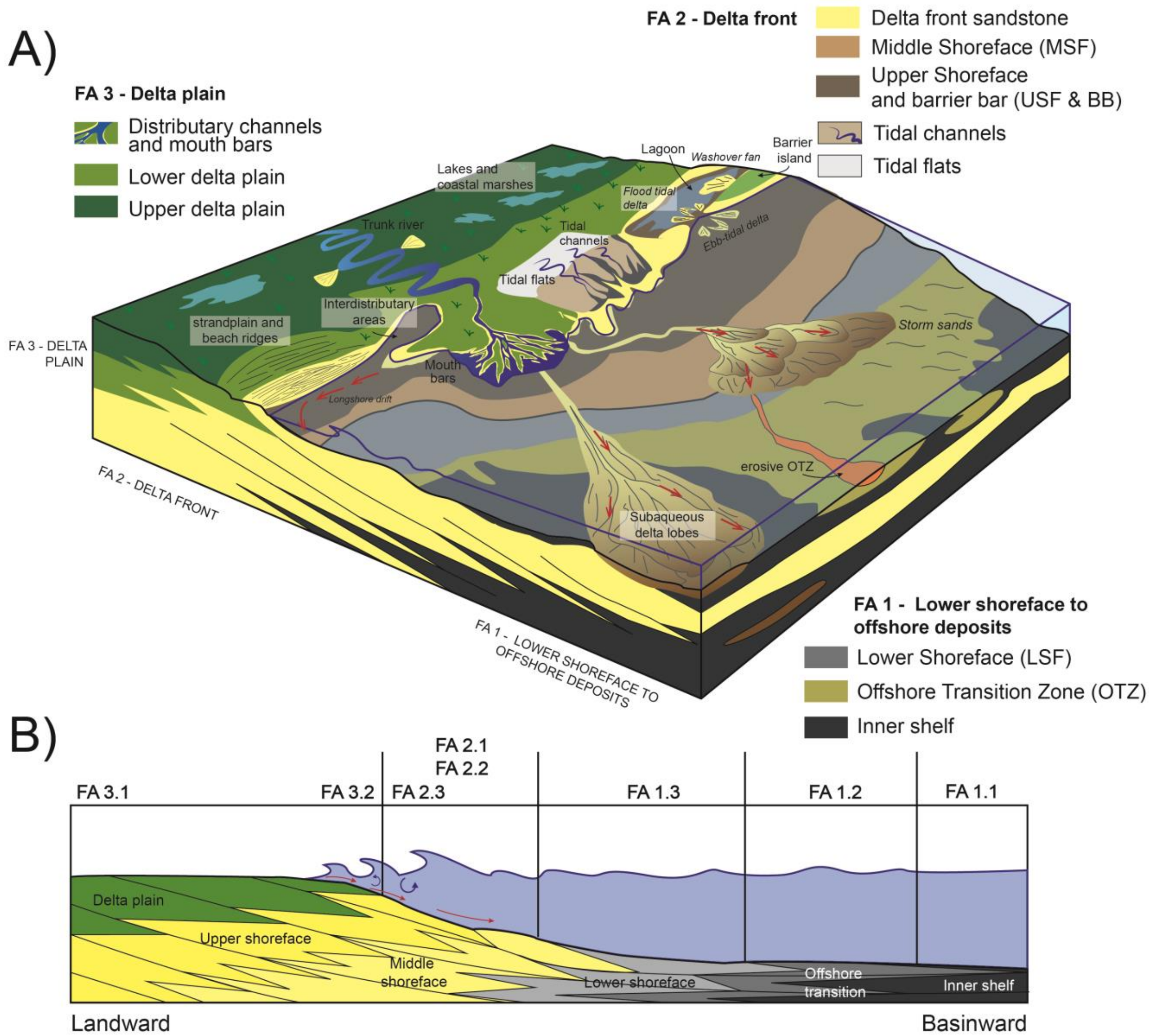

Figure 6.1: A) Conceptual block diagram showing the various environments and sub-environments occurring in the De Geerdalen Formation. Fluvial, wave and tidal processes are important factors in governing coastline morphology. Tidal processes generate tidal flats and channels and in combination with waves are important processes in barrier island complexes. Strandplains and beach ridges are mostly the product of wave energy and supplied by longshore drift. Note that the subaqueous delta lobes are not deep water fans, but results from the deposition from buoyant plumes or weak underflows off distributary mouths (drawn based on concepts from Bhattacharya and Walker, 1992; Reinson, 1992; Howell et al., 2008; Rød et al., 2014). **B)** Profile of a prograding shallow marine shoreline based on the above block diagram.

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

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

6.1.1 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. 6.2), and makes up the coarsest fraction of grain sizes seen in this facies association. Plant fragments are also found in sandstones of facies I_2.

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

Interpretation

The offshore zone is defined as shelf areas below mean stormweather base and is the site of deposition of fine mud and silt settling from suspension (Bhattacharya, 2006; Nichols, 2009). 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., 1999a) 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 (Fig. 6.1B). 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, the hummocky cross-stratified sandstones of the Muen locality interrupted by intervals of up to 20 meters of mudrock that possibly belong to the distal offshore zone (Fig. 6.3). 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.

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

Figure 6.2: *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.

6.1.2 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) is 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 stormweather 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 fair weather 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. 6.3). 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.

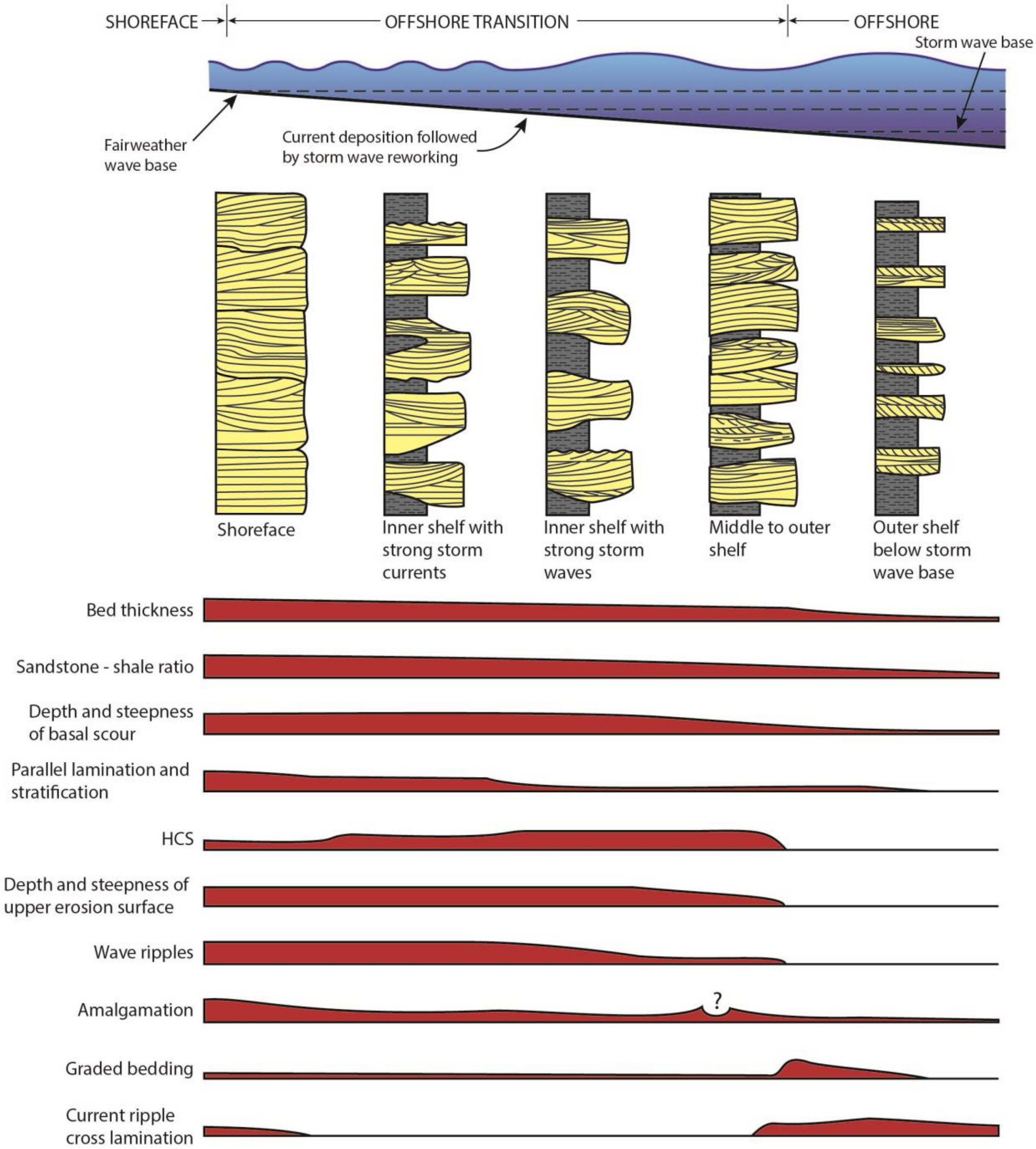

Figure 6.3: Shifts in facies occur in storm deposits as water depth and distance from the shoreface increases (Brenchley, 1985, Miall, 2000)

6.1.3 Lower shoreface (FA 1.3)

Description

The lower shoreface (Fig. 6.1A) 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 inter-finger 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 fair weather 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 fair weather.

Facies Association 2 (FA 2) - Delta front

The delta front is characterized by a relatively steep delta slope where interaction between fluvial and basinal processes constitutes the depositional framework acting upon the deltaically introduced sediments (Reading and Collinson, 1996). Progradational deltaic sequences shallow and coarsen upwards from mud dominated lower shoreface to offshore deposits (FA 1) into sand-dominated facies characterizing the delta front (FA 2) (Fig. 6.1).

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 increase in energy to the underlying FA 1. Furthermore, they are thought to represent sub-environments, within the delta system, influenced differentially by basinal and fluvial processes.

6.1.4 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 fair weather 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.6.4A). 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. 6.4B). 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 (Fig. 6.4) 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, Fig. 6.1A) exhibiting upwards-shallowing succession of sand overlying distal marine offshore to lower shoreface deposits (FA 1) and underlying proximal non-marine facies. However, the proximal non-marine 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 fair weather 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. 6.4B). 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 fair weather deposition, but are reworked by storms and during transgressions (Clifton, 2006). Distribution of the zones of the shoreline profile is largely controlled by intensity of wave energy and nature, while a tidal influence produce gradual transitions and overlap as the location of mean storm wave base and fair weather base is transient (Reading and Collinson, 1996).

The type of coast depends on controls imposed by; i) relative power of waves, tides and fluvial source; ii) sediment grain size; iii) marine sediment supply; iv) relative sea-level change (Reading and Collinson, 1996). These controls results in a variety of clastic coasts where beach-ridge strandplains, chenier plains and mudflats in general have been attributed to regressive systems, while barrier island-lagoonal systems and estuaries commonly is associated with transgressive systems (Fig. 6.1A) (Boyd et al., 1992; Dalrymple et al., 1992; Reinson, 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 sand bodies are generally characterized by linear geometries, while strand plains commonly have sheet-like sand bodies (Clifton, 2006).

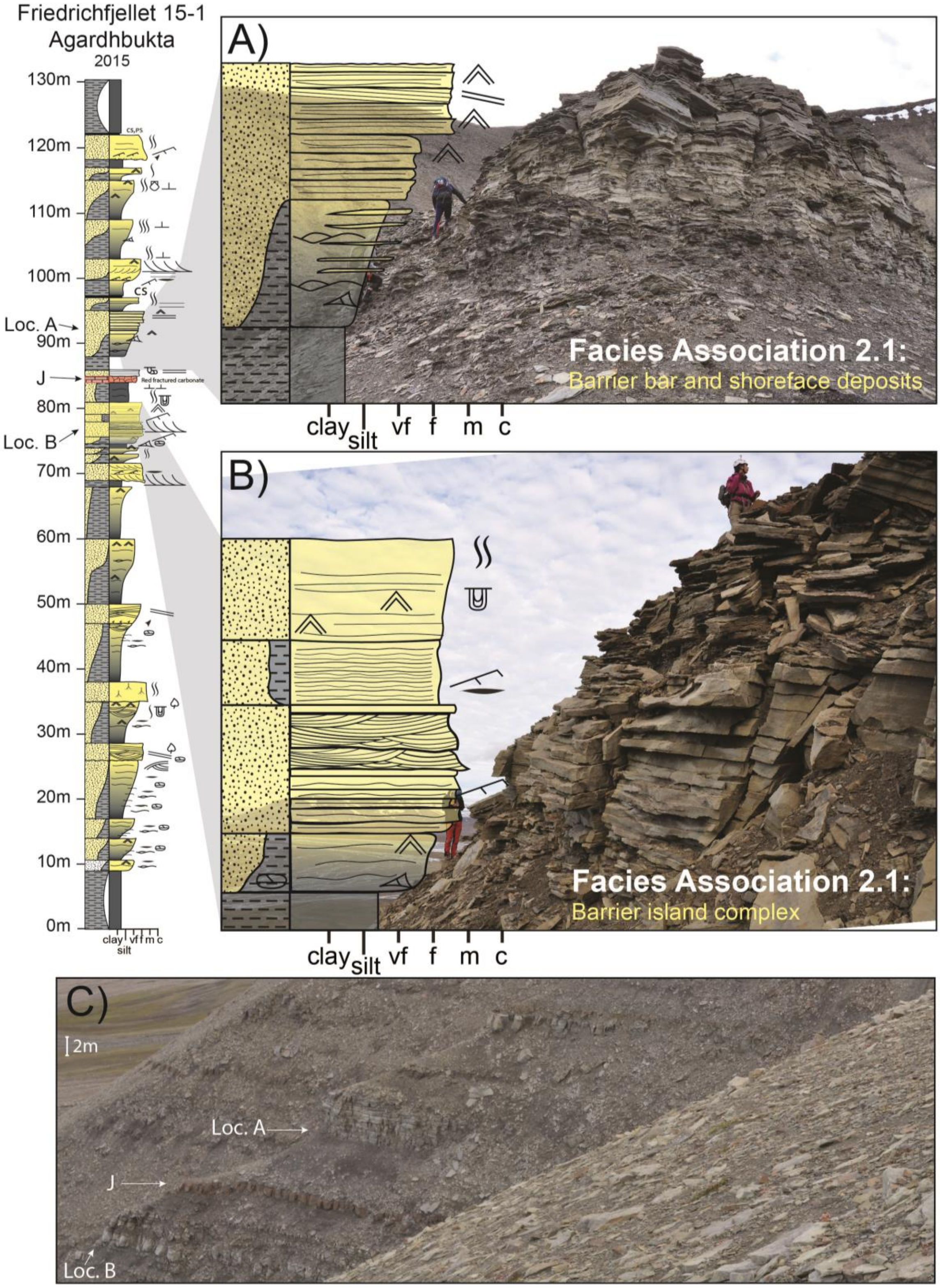

Figure 6.4: Lower sections logged on Friedrichfjellet. **A)** Coarsening and shallowing upward sequence with facies indicating prevalent oscillating flow, interpreted as a prograding barrier succession. **B)** Interpreted barrier island sequence with large-scale cross stratified interval possibly remnant of tidal inlet or longshore bar migration. **C)** Overview picture. Note the lateral continuity of facies J.

6.1.5 Distributary mouth bars (FA 2.2)

Description

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

Sandstone bodies are laterally extensive, 100s of meters, as observed in the field. However, field data regarding geometries of outcrops are sparse due to limitations in time and exposure. The facies association is mainly 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 (Fig. 6.1A) (Reading and Collinson, 1996; Olariu and Bhattacharya, 2006; Bhattacharya, 2006). An exceptionally rapid deposition rate 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 (Fig. 6.1A)(Bhattacharya, 2006). Ancient mouth bar sand bodies are shown to exhibit larger dimensions (Reynolds, 1999) than their modern analogues (Tye, 2004) and thus showing how migration and coalescence of modern bar forms created the greater ancient examples viewed in outcrops (Bhattacharya, 2006). Delta progradation is mainly achieved by coalescence of downstream migrating mouth bars (Bhattacharya, 2006). Width of distributary channels may vary spatially and temporally, roughly dictating the scale of the genetic related mouth bar. Size and shape of the mouth bar also depends on angle of plume dispersion, relative density of stream and basinal waters and processes (Bhattacharya, 2006).

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

6.1.6 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 (Fig. 6.5A). 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 is characteristic, while bioturbation is almost absent in this facies association. Sandstones are often cemented by calcite.

Based on dimensions two types of distributary channels have been documented in our area; 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. 6.5).

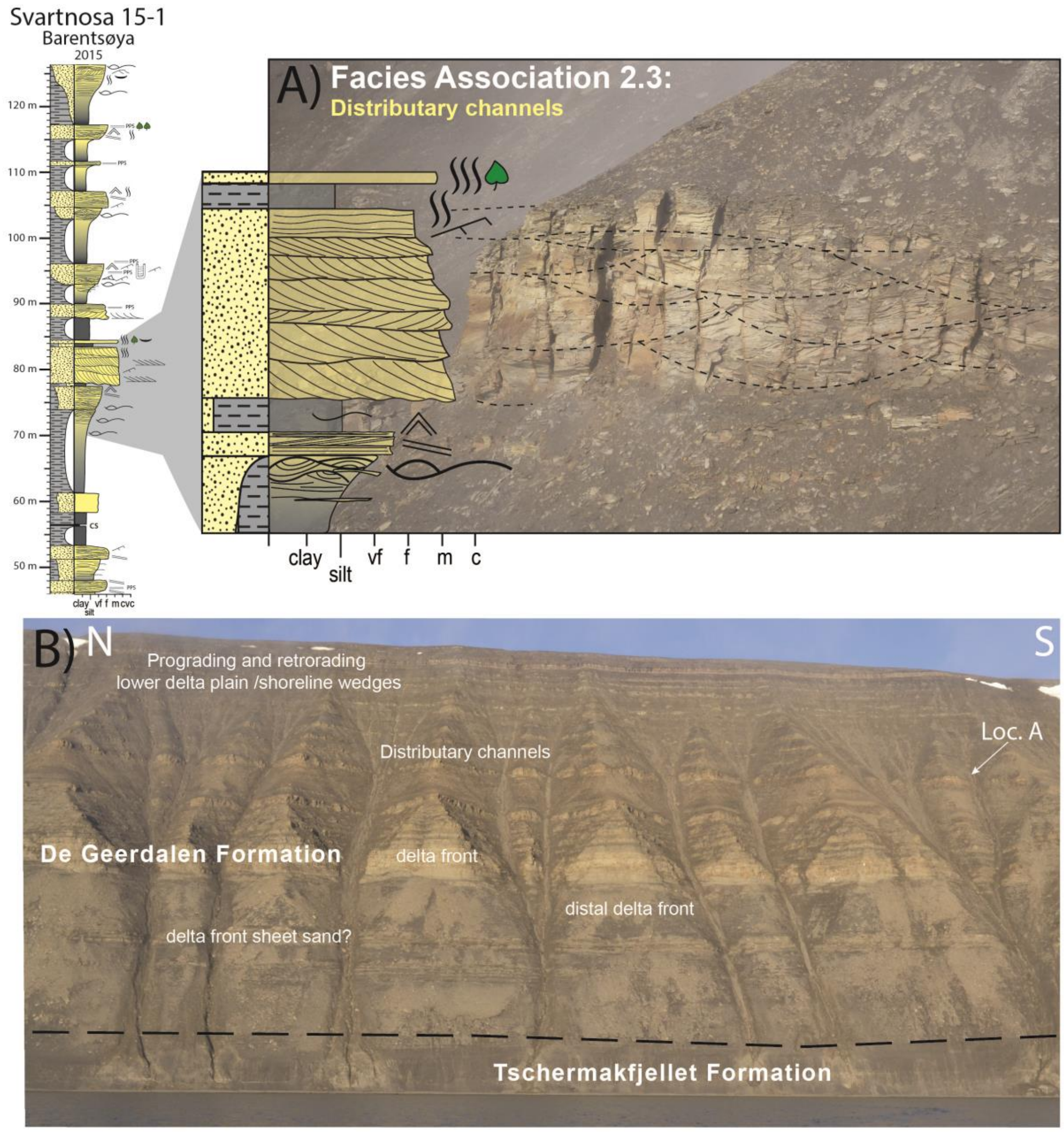

Figure 6.5: Svartnosa, Barentsøya. **A)** Amalgamated distributary channel deposits superimposed on distal delta front deposits of FA 1. **B)** Overview photograph taken at sea level. Note the lateral continuity of sandstone bodies. The channel-shaped sandstone in Fig. 6.5A was observed towards the top of the mountain on the southern side (upper right in the photograph). Width of view is approximately 400 m and the mountain is almost 200 m tall.

Distributary channels are also often found to overlie or interbedded with mouth bar deposits and 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 (Fig. 6.1A) (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 indicates abandonment of the channel (Reading and Collinson, 1996). Distributary channels differ from alluvial channels in several ways. The lower, seaward end of distributary channels is influenced by basinal processes as tidal and wave processes rework the channels into mouth bars (Bhattacharya, 2006). Hence mouth bars (FA 2.2) are more common on the distal delta front, while terminal distributary channels occur more frequently on the proximal delta front (Olariu and Bhattacharya, 2006). Distributary channels are more prone to avulsion and switching than fluvial channels due to lower slope gradient. The width to depth ratio is also smaller for distributary channels because of the relatively short lifetime and therefore limited time to migrate laterally (Reading and Collinson, 1996).

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

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

Facies Association 3 (FA 3) - Delta plain

Delta plain is commonly overlying facies association 2 - delta front, and is typically found in the uppermost part of the De Geerdalen Formation. The facies association is recognized by deposits where the marine influence is less prevalent and more closely connected to paralic and continental environments. The upper limit of modern delta plains are normally defined by the presence of distributary channels and the limit to the lower delta plain are 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 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 (Fig. 6.1A). Sub-environments include distributary channels, swamps, marshes, interdistributary bays, tidal flats and lagoons (Bhattacharya, 2006). Numerous active and inactive distributary channels are commonly found across the delta plain, separated by shallow water areas with little sedimentation and emergent areas (Reading and Collinson, 1996). Distributary channels (FA 2.3) are described under delta front deposits (FA 2) and this chapter will not provide any further description.

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

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

6.1.7 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 and coal shales (facies O and N) are found both at top of distributary channels and on floodplain deposits. Plant fragments and bioturbation in sandstone beds are common. Extensive scree cover is typical, and is often inferred as mudrocks if no sign of coarser material is seen. The thickness of floodplain deposits is in the range of 2 - 15 meters.

Interpretation

Distributary channels described under facies association 2 (Delta front) is together with floodplains one of the main signatures of the active parts on the delta plain (Bhattacharya, 2006). Flood plains are strongly related to the distributary channels since they receive most of the sediments from distributary channels. The sediment load in most rivers contains as much as 85 - 95 % mud (Schumm, 1972). The mud is primarily carried in suspension, and most of it is deposited in the channel itself, dams, in the delta front, and on floodplains related to fluctuations in water level. This leads to fine grained floodplain deposits (Bridge, 2006; Bhattacharya, 2006). Inundation is, however, not only a product of flooding of the river channels. Increased water table 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 (Fig. 6.1A). Typical features in crevasse splays are cross-lamination and small-scale cross-bedding. Floodplains are commonly exposed during low water level, leading to pedogenetic processes. In humid conditions the floodplain sediments may stay wet, and if the conditions for plant growth are good, peat may accumulate (Collinson, 1996).

Fine grained deposits alternating with thin ripple laminated sandstones displaying various degree of bioturbation close to interdistributary channels are interpreted as floodplain deposits. Climbing ripples (facies C) is interpreted as rapid deposition of sediments and could possibly be formed in crevasse splays. 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 floods.

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

6.1.8 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 (Fig. 6.1A). 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, Fig. 6.6A).

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).

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).

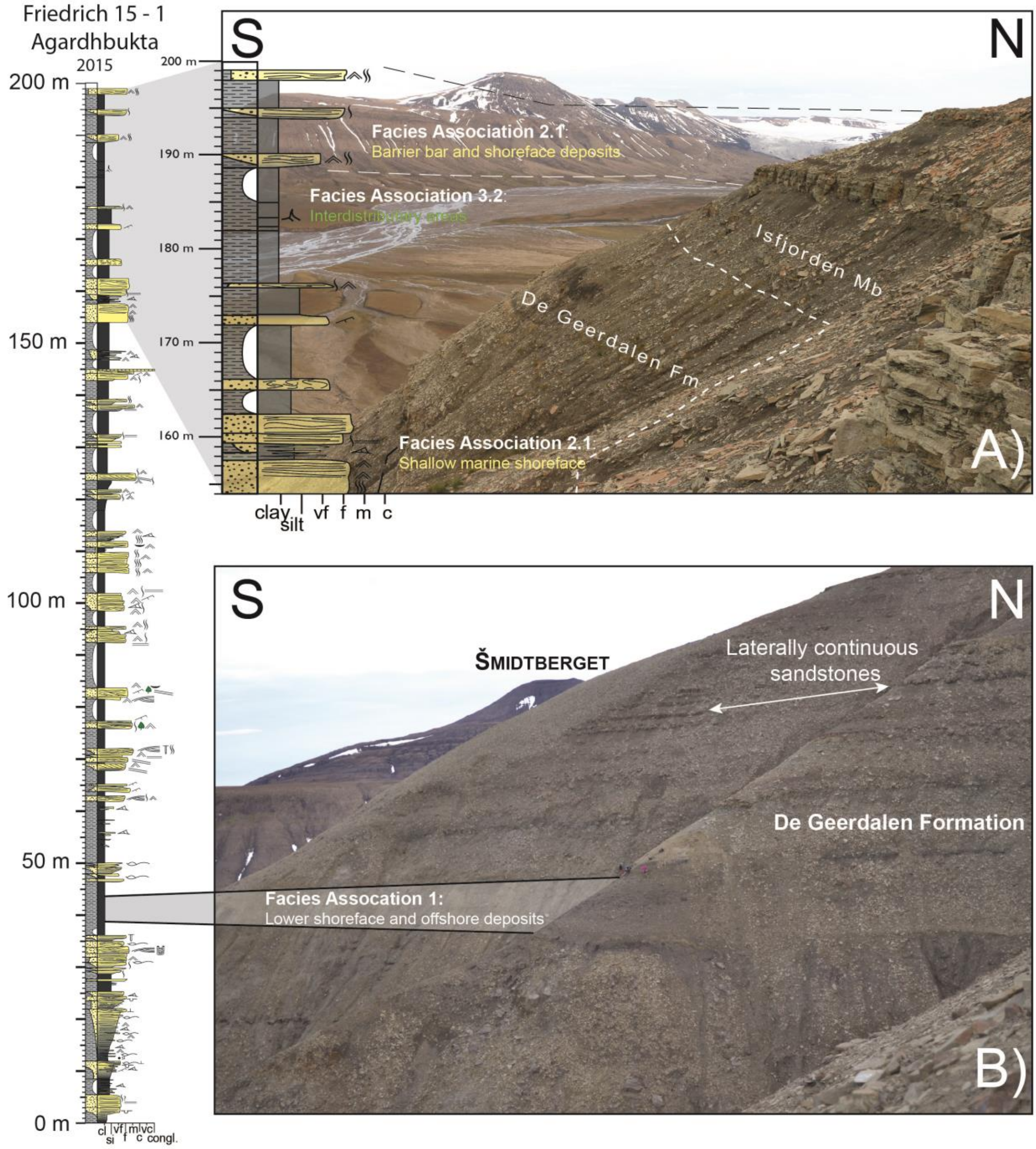

Figure 6.6: Mud-dominated facies associations on Friedrichfjellet, Agardhbukta **A)** Facies Association 3.2: Interdistributary areas. The red and green shales of the Isfjorden Member are conformably overlying the De Geerdalen Formation **B)** Facies Association 1: Lower shoreface to offshore deposits. Note the lateral continuity of the sandstones in the middle part of the mountain.

Swamps and marshes are the main peat- and coal forming environment. Swamps are freshwater sourced wetlands that favour woody vegetation and are mainly located on the upper delta plain. Swamps gradually turns into fresh, brackish or saline marshes in the seaward direction. Peats formed in saline marshes on the lower delta plain tend to have high content of impurities from terrigenous matter and sulphur (Reading and Collinson, 1996).

Coal and coal shales found in the De Geerdalen Formation appear as impure, leading to the possible interpretation of a lower delta plain setting in saline marshes. Thin coal seams and apparently limited laterally continuity may support a relatively dynamic and unstable regime. Waterbodies in upper delta plains are mainly lakes. Lower delta plains also often contains lakes, but can differentiate from upper delta plain by the presence of lagoons, estuaries and interdistributary bays. Lagoons are shallow waters roughly directed alongshore and protected by a barrier island (Reading and Collinson, 1996). Restricted water circulation causes varying salinities. Lagoons commonly accumulate pervasively bioturbated fine grained sediments deposited from quiet waters (Reading and Collinson, 1996). Bay-head deltas occurring on the landward margin of the lagoon generate small-scale facies sequences resembling those of fluvial-dominated deltas (Reading and Collinson, 1996). Interdistributary bays and their associated small-scale bay-fills frequently develop on river-dominated deltas, whereas wave-influenced deltas develop more uniform and straightened coastlines displaying more lateral continuity of correlative vertical sequences (Bhattacharya and Walker, 1992; Reading and Collinson, 1996).

In humid climates blanket bogs are formed in the margins of the lakes. In contrast, lake margins in arid climate tend to form exposed surfaces with calcretes, gypsum and halite precipitations. Common for the water bodies are shallow water depth and low energy, resulting in dominant deposition of fine materials such as clay, silt and fine sand. However, floods and diversion of distributary channels disturbs the regime and brings in coarser material resulting in a variety of features formed from crevasse splays, crevasse channels, levees and small deltas (Reading and Collinson, 1996). Long intervals of mudrocks in the upper parts of the De Geerdalen Formation are herein interpreted to represent shallow closed or semi-closed standing water on delta plain or upper delta front. Paleosols indicates 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 et al., 1999a; Mørk, 2015), but further investigations are needed to fully understand the shift.

7. Results and Interpretations

This chapter describes the sedimentology of De Geerdalen Formation from measured stratigraphical sections (Appendix A and C) and interpretations of depositional environments. For descriptions and interpretations of individual facies, the reader is referred to Chapter 5. Descriptions, interpretations and discussions regarding facies associations can be found in Chapter 6, while more general discussions regarding facies distributions and stratigraphic architecture are dealt with in Chapter 8.

7.1 Study area

The study area in eastern Svalbard is outlined in Figure 7.1. During the one-month field season, 13 locations were visited on eastern Spitsbergen, Barentsøya and Wilhelmøya. The localities were picked primarily based on previous studies (Knarud, 1980; Rød et al., 2014) focusing on areas where detailed sedimentological descriptions of the De Geerdalen Formation were lacking or where further work is necessary.

7.2 Agardhbukta, eastern Spitsbergen

Agardhbukta is an embayment on the east coast of Spitsbergen (Fig. 7.1). During the winter, the area is easily accessible by snow mobile. However during the summer, easy access is restricted from land due to rivers and streams formed by glacial meltwater, as well as plateau mountains. Hence, fieldwork in the area requires logistical support in the form of helicopter or ship.

A total of nine sections, of which four are included in this thesis, were measured from three mountains (Klement'evfjellet, Friedrichfjellet and Šmidtberget) on the southwestern side of the Agardhdalen valley. The small distance between the mountains allows for good correlations of sections and facies associations, as sandstones are observed to be continuous across the mountainside.

The main tectonic element affecting Triassic sediments in Agardhbukta is a N-S striking fault zone (Lomfjorden Fault Zone – LFZ), extending from north-eastern Ny-Friesland southwards and terminating with the Agardhbukta Fault (Dallmann et al., 2015). On Klement'evfjell, the faults were observed in a succession of shales adjacent to the mountain, as well as on the northern side of the Agardhdalen valley. The sandstone horizons on Klement'evfjellet appear to be unaffected by the Lomfjorden Fault Zone.

Figure 7.1: A) Outline of study area on eastern Svalbard, with the location of Svalbard (inset). B) Overview map of localities visited. From Lord et al. (in prep.).

7.2.1 Klement’evfjellet (Klement 15-2 and Klement 15-4)

Klement’evfjellet is a 394 m high mountain located on the southwestern side of Agardhbukta, south of Friedrichfjellet and Šmidtberget (Fig. 7.1). Situated west of the Lomfjorden Fault Zone (Dallmann et al., 2015), it is host to a series of alternating sandstones and shales belonging to the De Geerdalen Formation. The Isfjorden Member is found towards the top of the mountain, as well as the Wilhelmøya Subgroup represented by the Knorringfjellet Formation and the Brentskardhaugen Bed.

Previous work, summarized in Mørk et al. (1982), on the Triassic to Jurassic succession on Klement’evfjellet was conducted by Knarud (1980), who investigated the De Geerdalen Formation in 1978. On the eastern slope of Klement’evfjellet, the top of the Wilhelmøya Subgroup is represented by a 15 cm thick Brentskardhaugen Bed (Bäckström and Nagy, 1985). Above this is 3 m of cross-laminated sandstone with *Chondrites*, *Rhizocorallium* and *Thalassinoides* trace fossils (Bäckström and Nagy, 1985).

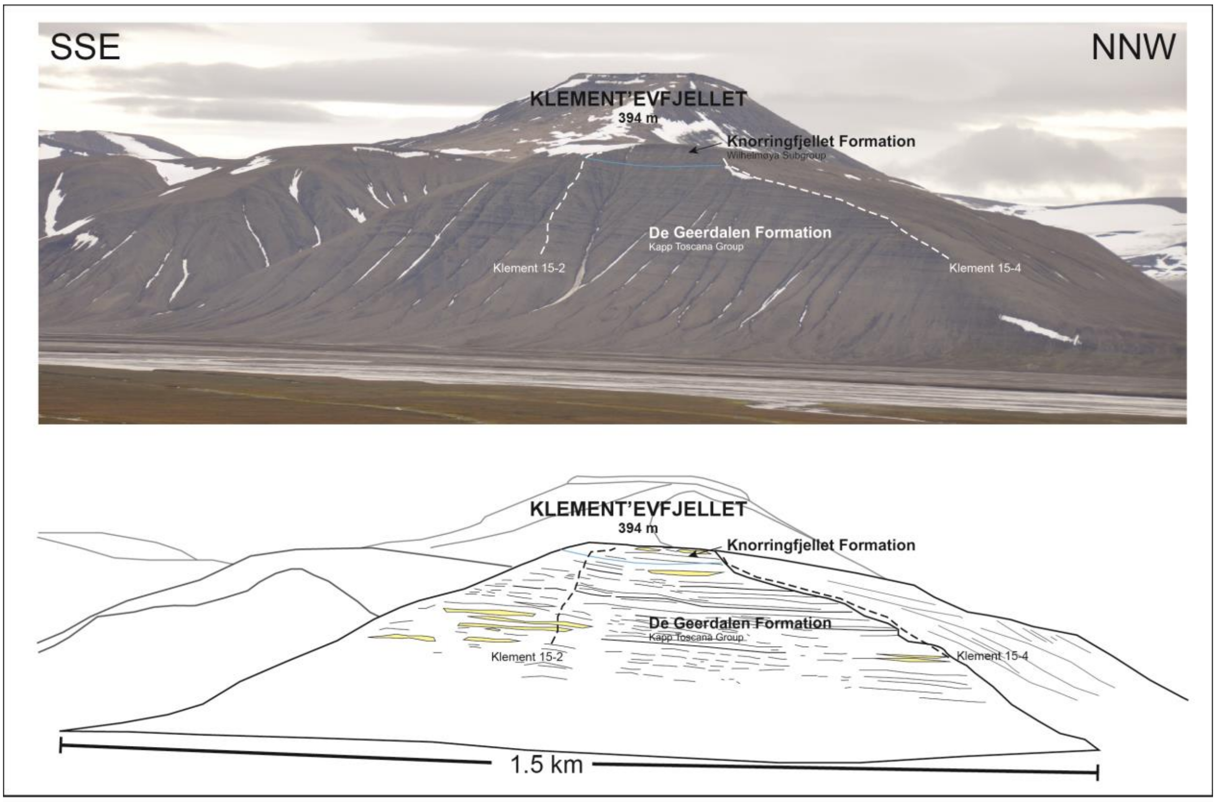

Figure 7.2: The De Geerdalen Formation on Klement’evfjellet consists of thin, horizontally continuous sandstone bodies. Log traces for Klement 15-1, 15-2 and 15-4 are indicated.

Three logs (Klement 15-1, 15-2, 15-4) were recorded from Klement’evfjellet at three different positions. Klement 15-1 is not included in this thesis, but can instead be found in Haugen (in prep.). Klement 15-2 follows a ridge in the middle part of the mountain and covers approximately 190 m of the De Geerdalen Formation, while Klement 15-4 covers 110 m long profile along the north-western side (Fig.7.2, Appendix C).

The De Geerdalen Formation on Klement’evfjellet consists primarily of three major coarsening upward sequences from shales to fine-grained sandstones (Fig. 7.3B), each in turn capped by marine shales. Each coarsening upwards sequence is characterized by increasingly proximal facies and facies associations towards the top. The geometry of sandstones bodies on Klement’evfjellet is characterized by thin, laterally continuous layers, similar to equivalent sandstones on Trehøgdene further west on central Spitsbergen (Rød et al., 2014).

Description (Klement 15-2)

0 – 31 m: This interval in Klement 15-2 covers the lowermost coarsening upwards sequence on Klement’evfjellet. The lowermost beds are partly covered and are generally heterolithic (facies K), but is also seen to consist of mudrocks (facies M, Fig. 5.18B) and interbedded sandstones with moderate bioturbation, hummocky cross stratification (facies H), small-scale ripple cross lamination (facies B) and wave ripples (facies D). The uppermost beds coarsens upwards from silty shale to very fine and fine sandstones with planar parallel stratification (facies F), wave ripples with mud drapes (facies D) and hummocky cross stratification in the lower reaches of units (facies H). Observations were also made of fish remains and plant fragments. No observations of bioturbation were made, although it does not exclude its presence.

31 – 66 m: The lowermost 11 m of the second coarsening upwards sequence consists of covered muds. Above this the succession becomes heterolithic (facies K, Fig. 5.15B) with shales (facies M) interbedded with very fine-grained hummocky cross stratified sandstone (facies H, Fig. 5.9E), plant fragments and wave ripples (facies D, Fig. 5.4E) changing into low angle cross stratified sandstones and fine-grained large-scale cross bedded sandstone (facies A). Sandstones are observed to be laterally continuous for tens of meters across the mountain slope (Fig. 7.2) and are overlain by a mud-dominated sequence with a sharp lower contact. The mud consists of brown and yellow shales, indicating the presence of paleosols.

66 – 102 m: The lower contact is similarly to the underlying sequences is also marked by scree-covered shales and mudrocks. The interval consists of multiple minor coarsening upwards units from silty shale to fine-grained sandstone forming a major coarsening and shallowing upwards sequence. This is the most sand-rich interval in the succession on Klement’evfjellet and is observed to occur at similar stratigraphic level as the successions on Friedrichfjellet and Šmidtberget. The lower minor coarsening upwards units contain moderate bioturbation, hummocky cross stratification (facies H), low-angle cross stratification (facies E) and wave ripples (facies D), increasing in abundance upwards and with mud-draped foresets. The sandstones are mostly sparse to moderately bioturbated and locally intense and contain marine trace fossils (*Diplocraterion*, Fig. 5.25H). Coarsening upwards units are usually separated by deposits of mudrock up to 4 m thick. The uppermost coarsening upwards units consist of fine sand with moderate bioturbation, ripple cross lamination (facies B), plant fragments, siderite nodules (facies J) and wave ripples towards the top (facies D, Fig. 5.4A).

102 – 152 m: This interval is mostly mud-dominated, while sandstones consists of minor coarsening upwards units to very fine sand with wave ripples (facies D), sparse to moderate bioturbation, marine trace fossils (*Skolithos)* siderite nodules and interbedded shales (facies M) and paleosols (facies O_2).

152 – 185 m: Similarly to the underlying sequence, this interval is also mud-dominated and consists of mudrocks (facies M), coal shales (facies N), and red and green shales with calcareous nodules at discrete levels (facies O, Fig. 5.23B). Interbedded sandstones are usually discontinuous laterally and consist of calcareous ripple cross stratified sandstones (facies J and B), sparse bioturbation and wave ripples (facies D). Siderite nodules are also commonly observed within both sandstones and shales.

185 – 204 m: The uppermost part of Klement 15-2 consists of a sequence of shales and fining upwards units of sand with channel geometry and large-scale cross stratification (facies A). This part of the interval was not studied in detail as it is seen to be different from the underlying De Geerdalen Formation and beyond the scope of this thesis. The succession is seen to rest conformably on the underlying sequence and at the base is a calcareous sandstone bed (facies J).

Interpretation (Klement 15-2)

0 – 31 m: The lowermost mud-dominated part of the succession is interpreted as reflecting an upward transition from a storm-dominated inner shelf environment, through the offshore transition zone and up to the lower shoreface. The coarsening upwards pattern, heterolithic nature of the deposits, presence of bioturbation and fish remains as well as wave ripples indicates that these sediments were deposited in a shallow marine shoreface environment influenced by waves and storms, likely proximal to the lower shoreface. Presence of plant fragments are also interpreted as indicating a proximal position to the delta front. Knarud (1980) interpreted similar intervals on Klement'evfjellet as distal delta front and distal banks.

31 – 66 m: The contact to the underlying sequence is interpreted as a flooding surface (Fig. 7.3B). The lowermost 11 m are interpreted as storm-dominated inner shelf, while the coarsening upwards trend indicates a shallowing of facies to a shallow marine shoreface setting. Large-scale cross stratified sandstone with wave ripples can be interpreted, similarly to Knarud (1980), as proximal subaqueous banks on the delta front. The observation of paleosols above the sandstones may indicate a transition from marine to non-marine deposits, i.e. the lower delta plain. Similarly to the succession on Hellwaldfjellet, this can be interpreted as reflecting a distal position to an active distributary area.

66 – 102 m: Similarly to the underlying sequence, the lower contact is interpreted as a marine flooding surface. The increasing content of sand upwards indicates a shallowing upwards trend. *Diplocraterion*, hummocky cross stratification and wave ripples indicates deposition in the offshore transition zone to lower shoreface. Interbedded mudrocks likely represents deposition of fine-grained material in a shallow marine shoreface or on the transition to the lower shoreface below fair weather wave base. The uppermost coarsening upwards unit is interpreted as a proximal shallow marine environment. Siderite nodules and plant fragments may result from deposition in organic rich stagnant waters close to the delta front.

102 – 152 m: This interval is interpreted as representing a paralic, shallow marine environment with variable energy regime. Interbedded mudstones with paleosols indicate periods of subaerial exposure and deposition in a low-energy environment such as on the floodplain or in an interdistributary bay. Sandstone beds with wave ripples and *Skolithos* may result from inter-fingering of higher energy shoreface sandstones.

152 – 185 m: Wave rippled sandstones are interpreted as being shallow marine in origin. Red and green shales are interpreted as paleosols and are characteristic of the Isfjorden Member and together with coal shales reflect deposition in a low energy setting, such as an interdistributary bay. Paleosols interbedded with ripple cross laminated sandstones are interpreted as floodplain with crevasse splays. Siderite nodules indicate reducing conditions during deposition and are interpreted as reflecting a restricted environment.

185 – 204 m: The uppermost part of the mountain is interpreted as the Knorringfjellet Formation. The lower carbonate cemented sandstone at the base most likely represents a potential candidate for the Slottet Bed.

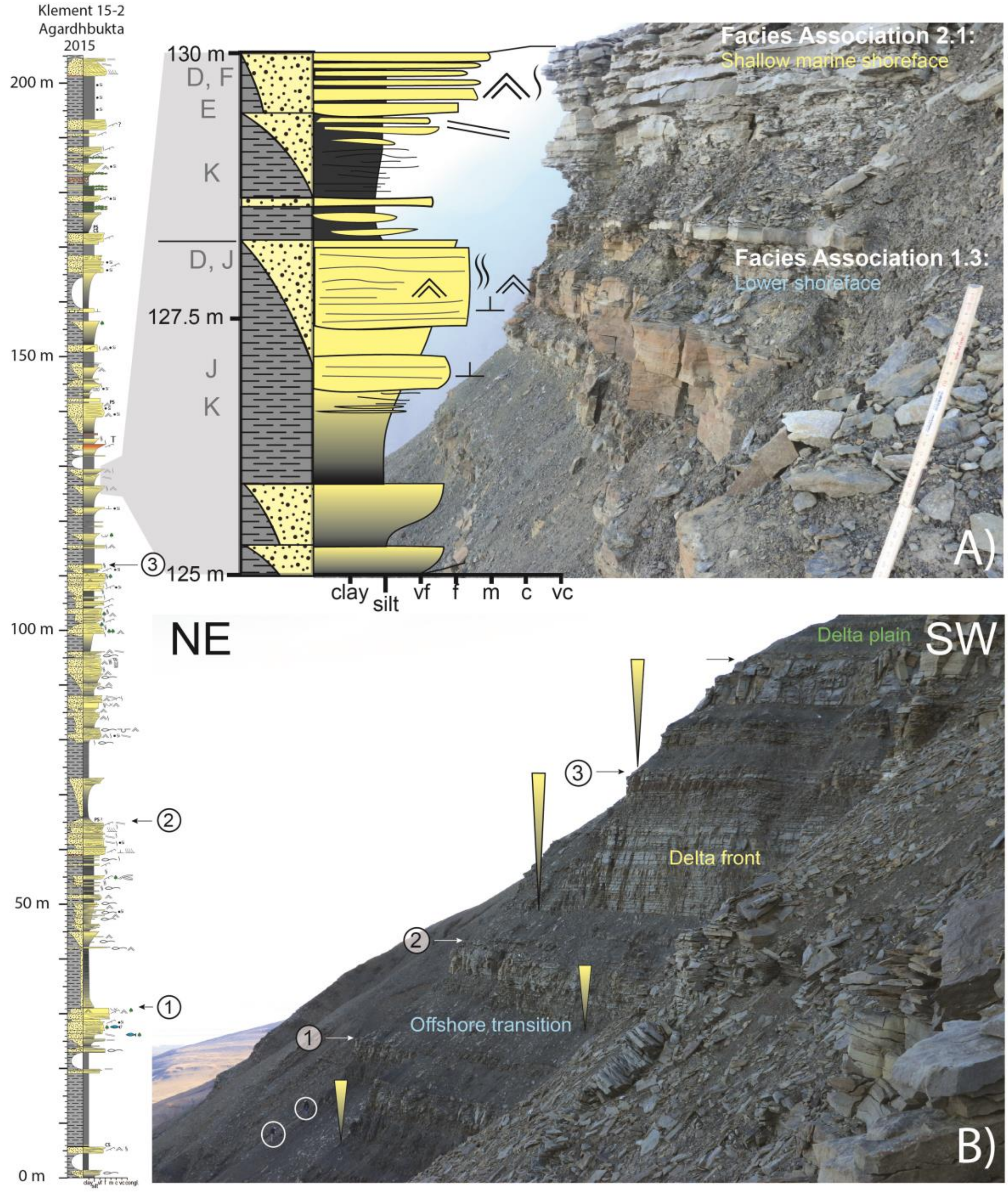

Figure 7.3: **A)** Log-outcrop correlation illustrating the transition between facies association 1.3 (lower shoreface and offshore deposits) to facies association 2.1 (barrier bar and shoreface deposits). Capital letters indicate interpreted facies. **B)** Several coarsening and shallowing upwards parasequences representing delta lobe switching. Annotated arrows indicate interpreted flooding surfaces marking the base of a new parasequence. Note geologists for scale in the lower left corner (white circles).

Description (Klement 15-4)

0 – 16 m: The lowermost beds measured in Klement 15-4 consists of horizontally bedded, fine to medium sandstone with wave ripples (facies D) and *Skolithos*. The sandstone is overlain by a 6 m minor coarsening upwards unit from silt to fine sandstone with hummocky cross stratification (facies H) and wave ripples (facies D). Above this is 8 m of covered shales.

16 – 33 m: Overlying the covered shales is a heterolithic sequence interbedded shales and sandstones with wave ripples, mud flakes and sparse to moderate bioturbation. The lowermost sandstone bed contains large-scale cross stratified sandstone with bioturbation. The rest of the interval consists of minor coarsening upwards units with small-scale ripple cross lamination and planar parallel stratification (facies F).

33 – 42 m: The heterolithic sequence is overlain by a 4 m thick coarsening upwards unit from silty shale to fine sandstone with hummocky cross stratification (facies H), wave ripples (facies D), mud flakes and sparse to moderate bioturbation. This is overlain by a 5 m thick mud-dominated unit (facies K) consisting of shales with siltstone interbeds.

42 – 88 m: The lowermost sandstone in this interval is bounded by a sharp contact to the underlying shales. The sandstone appears to be laterally continuous and uniform in thickness (Fig. 7.2), but due to scree cover adjacent to the sandstone, it is not straightforward to determine how laterally extensive the unit is. The middle part of the succession on Klement 15-4 consists of multiple minor coarsening upwards units from silty shale to fine sandstone. Sedimentary structures includes large-scale cross stratification (facies A), ripple cross lamination (facies B), wave ripples (facies D), and low-angle cross stratification (facies E). The units are heterolithic (facies K) and individual units of sandstone and shale are several decimetres thick. Bioturbation is locally moderate and the uppermost sandstones contain marine trace fossils (*Skolithos* and *Diplocraterion*). The uppermost 5 m consists of covered shales overlain by a 1 m thick unit of fine sand with low-angle cross stratification (facies E).

88 – 107 m: The lowermost 11 m consists of covered shales and siltstones. The remaining part comprises red and green shales (facies O_2) with nodular concretions weathering out from the scree. The development of this part of the succession is observed to be similar to equivalent levels in Klement 15-2.

107 –120 m: The uppermost beds in Klement 15-4 consist of multiple upwards coarsening units overlain by sandstones with large-scale cross stratification (facies A), mud flakes, plant fragments and channel geometry.

Interpretation (Klement 15-4)

0 – 16 m: Wave ripples and *Skolithos* indicate deposition in a high energy shallow marine environment. The hummocky cross-stratified sandstone and covered mudrocks is interpreted as indicating a marine flooding surface and deposition in the offshore transition zone.

16 – 33 m: Wave ripples and marine bioturbation are interpreted as indicating a shallow marine environment. Large-scale cross stratified sandstones can be interpreted as proximal shallow marine, subaqueous banks on the delta front, similarly as in Klement 15-2. The heterolithic nature of the deposits may reflect changes in wave energy and sediment supply or a tidal influence on the deposits.

33 – 42 m: This unit is correlated to a minor flooding surface interpreted in Klement 15-2. Hummocky cross stratification is interpreted as reflecting deposition in the offshore transition zone.

42 – 88 m: The sharp contact is interpreted as representing the transition to a more proximal shallow marine environment. Large-scale cross stratification, wave ripples, ripple cross lamination, lateral continuity of sandstones, as well as presence of bioturbation and marine trace fossils are interpreted as indicating a shallow marine shoreface environment. The thickness of heterolithic units indicates a fairly distal position. This part of the interval correlates to units interpreted as distal delta front and proximal banks as interpreted by Knarud (1980). The uppermost part of the interval is interpreted as consisting of shallow marine shoreface deposits.

88 – 107 m: The interpretation for this part of the succession is the same as for Klement 15-2, where equivalent units are interpreted as interdistributary bay deposits.

107 – 120 m: The uppermost part is interpreted as sandstones and shales belonging to the Knorringfjellet Formation.

7.2.2 Šmidtberget (Šmidt 15-1)

Šmidtberget is a 533 m tall trapezoidal-shaped mountain directly southwest of Friedrichfjellet (Fig. 7.1). The De Geerdalen Formation consists here, similarly to Friedrichfjellet and Klement'evfjellet, of multiple coarsening upwards sequences of sandstones and interbedded shales. The uppermost beds on Šmidtberget are generally weathered, lithologically different and much more sand-rich than the underlying sequence. It is also visually similar to, and occurs at the same stratigraphic level as the uppermost orange beds on Friedrichfjellet, 2.2 km to the northeast and Klement'evfjellet further south. It therefore most likely belongs to the Knorringfjellet Formation.

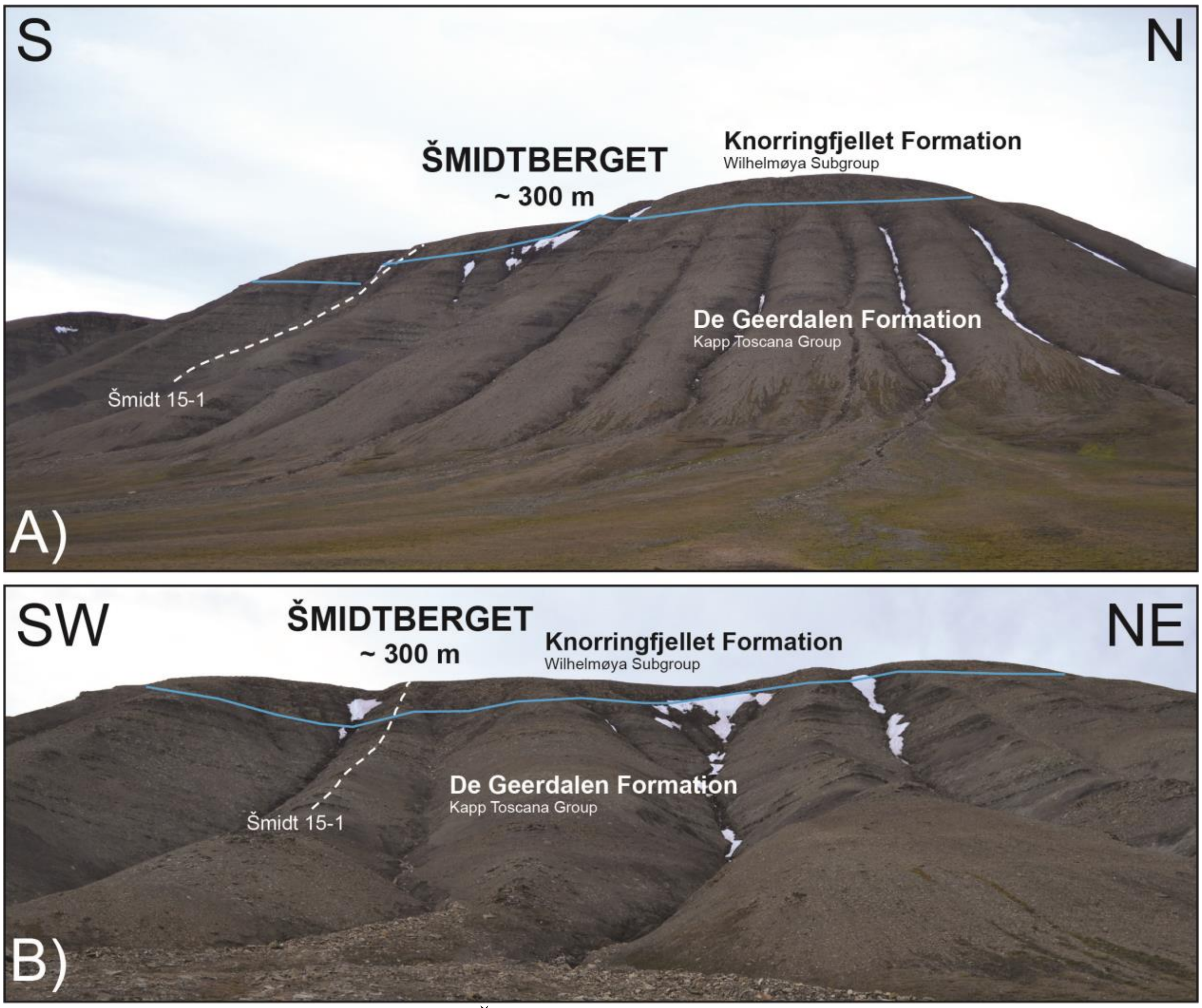

Figure 7.4: Overview photograph of Šmidtberget. **A)** The mountain exposes multiple coarsening upwards sequences of sandstones and shales of the De Geerdalen Formation, overlain by the white and orange sandstones and shales of the Knorringfjellet Formation. **B)** Outline of log trace for Šmidt 15-1. Šmidt 15-2 was measured along the ridge to the right.

Three sections of the De Geerdalen Formation (Šmidt 15-1, 15-2, 15-3) were measured on the south-eastern side of the mountain, of which only Šmidt 15-1 is included within this thesis (Appendix C). The section includes 109 m of the lower to middle part of the De Geerdalen Formation. The Isfjorden Member constitutes 22 m of the uppermost part of the De Geerdalen Formation and consists primarily of red and green shales with thin sandstones. The uppermost beds measured in Šmidt 15-1 include 22 m of sandstones and shales belonging to the Knorringfjellet Formation.

Description

0 – 70 m: The lowermost succession measured in Šmidt 15-1 is generally heterolithic (facies K) and dominated by muddy lithologies (facies M) with a few interbedded sandstones. Sandstone units are commonly between 0.5 and 2 m thick and consist of fine sand with wave and current ripple lamination (facies D and B). The upper part of the interval consists of several minor coarsening upwards units from silty shale to fine sandstone, forming a major coarsening upwards sequence. Bioturbation is locally intense with *Cylindrichnus* trace fossils (Fig. 5.25F) and some beds contain shell fragments (facies J). Hummocky cross stratification (facies H), mud flakes and wave ripples (facies D) become more abundant towards the top of coarsening upwards sequences.

70 – 110 m: Sandstones in the middle part of the mountain become thicker and are observed to be laterally continuous across the mountain slope. They are horizontally bedded and fine to medium grained with planar parallel stratification, wave ripples (facies D) and in some places large-scale cross stratified (facies A, Fig. 5.1B). The succession is similarly to the underlying interval heterolithic (facies K) and sandstones in this interval are commonly planar parallel stratified (facies F) with wave ripples (facies D) on top surfaces.

110 – 130 m: The uppermost part of Šmidtberget is characterized by a heterolithic sequence of sandstones and shales. The mud-dominated parts of this interval includes the red and green shales of the Isfjorden Member (Figs. 5.21C, 5.23C), although it is observed to be thinner compared to equivalent stratigraphic levels on Klement'evfjellet and Friedrichfjellet. The sandstones are commonly horizontally bedded and observed to be carbonate-rich (Figs. 5.14E, 5.14G). Towards the top of the section, a laterally continuous, heterolithic sequence of horizontally bedded sandstones and shales were observed. The sandstones include wave ripples (facies D), wavy bedding (facies K, Fig. 5.15C), ripple cross lamination (facies B) and mud flakes. The uppermost carbonate-rich sandstone may be a potential candidate for the Slottet Bed.

130 – 150 m: Towards the top of Šmidtberget is several sandstones with large-scale cross bedding (facies A, Fig. 5.1D), mud flakes and ripple cross lamination (facies B). The sandstones have channel geometry and occur as individual bodies within the scree-covered interval. The uppermost beds were not studied in detail, but observed to be orange coloured.

Interpretation

0 – 70 m: The lowermost section consists of shallow marine shoreface deposits as indicated by presence of wave ripples and shell fragments. Mud-rich lithologies and hummocky cross stratification record deposition on the distal delta front and on the lower shoreface where sand are deposited during storms, while muds settles from suspension during calmer periods. Intense bioturbation reflect reworking by marine organisms between storm events.

70 – 110 m: Thicker and more laterally continuous sandstones indicate a more proximal depositional environment compared to underlying sequence. Sandstones are interpreted as representing a proximal shallow marine delta front environment. Large-scale trough cross bedding are interpreted as distributary mouth bars deposited in a proximal shallow marine environment.

110 – 130 m: Paleosols indicate subaerial exposure and deposition in interdistributary areas or on the floodplain, while sandstones are interpreted as representing a shallow marine environment. The red and green shales are also on Šmidtberget interpreted as a restricted depositional environment. The uppermost heterolithic unit in this interval, with wave ripples, mud flakes and ripple cross lamination are interpreted as representing a shallow marine shoreface setting. Wavy bedding is interpreted as reflecting deposition on a tidal flat where energy regime is fluctuating.

130 – 150 m: The uppermost part of the measured section on Šmidtberget is interpreted as the Knorringfjellet Formation.

7.2.3 Friedrichfjellet (Friedrich 15-2)

Friedrichfjellet is a plateau mountain on the western flank of Agardhdalen (Fig. 7.1) and the south-eastern side has a triangular-shaped slope. The mountain is located east of Šmidtberget and north of Klement'evfjellet. Three logs were measured at this location (Friedrich 15-1, 15-2, 15-3). Only Friedrich 15-2 is included in this thesis (Fig. 7.5, Appendix C), while the remaining logs are presented by Støen (in prep.) and Lord et al. (in prep.). The section was measured from the base of the mountain following a narrow gully (Fig. 7.5B), covering ca. 175 m of the lowermost part of the De Geerdalen Formation and 25 m of the uppermost part, which includes the Isfjorden Member. The De Geerdalen Formation consists of several minor coarsening upwards sequences with a major coarsening and shallowing upwards trend. The facies sequences record a gradual shallowing upwards trend of facies and facies associations. The distinct colour difference between the De Geerdalen Formation and the Isfjorden Member is particularly striking at this locality and can be distinguished when viewed from distance (Fig. 7.5A). The top of the mountain is capped by a several metres thick succession of white and orange sandstone beds, probably belonging to the Knorringfjellet Formation.

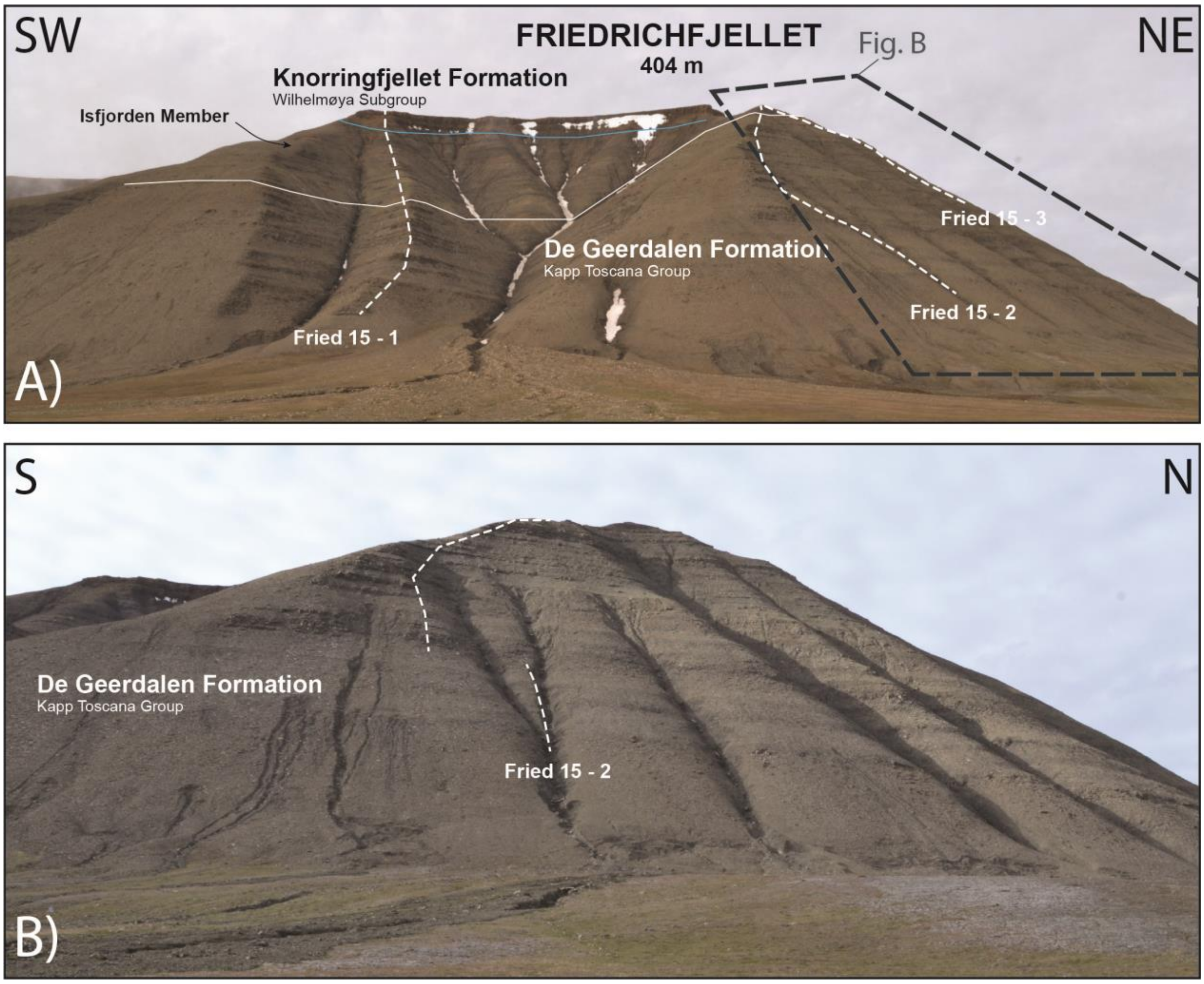

Figure 7.5: A) Annotated overview photograph of Friedrichfjellet. The De Geerdalen Formation is overlain by the Isfjorden Member (boundary indicated with white solid line) and possibly the Knorringfjellet Formation (boundary indicated with blue line). **B)** Log trace for Friedrich 15-2 on the south-eastern face of Friedrichfjellet.

Description

0 – 83.5 m: At the base of Friedrich 15-2, the succession is heterolithic and mostly consists of shales interbedded with thin very fine to fine-grained sandstone beds (facies K, Fig. 5.15D, 5.18A). The sandstones are typically 30 to 40 cm thick, ripple laminated and contain minor bioturbation and marine trace fossils (*Diplocraterion*). Sandstone beds are hummocky cross-stratified with minor loading structures at the lower contact to underlying beds (facies H) and wave ripple lamination on top surfaces of hummocks (Fig. 5.9A, 5.9D). They also contain thin beds of siderite (facies J, Fig. 5.14F).Wave ripples (facies D) and plant fragments are seen to increase in abundance upwards. In general the lower part consists of several minor coarsening upwards units that collectively form a major coarsening upwards. Towards the top of the interval, *Skolithos* trace fossils were observed together with hummocky cross stratification (facies H, Fig. 5.9C).

83.5 – 145 m: The middle part of the section is also heterolithic (facies K, Fig. 5.15E), however sandstones are thicker (1 – 2 m) and more laterally continuous compared to the underlying sequence. Sandstones consist of mostly of fine sand with small-scale ripple cross stratification (facies B, Fig. 5.2C), wave ripples (facies D, Fig. 5.4C), horizontal stratification and bedding (facies F) and low-angle cross stratification (facies E). Bioturbation is usually moderate and locally intense.

145 – 200 m: The sandstone beds become thinner towards the uppermost exposures of the mountain. The succession is heterolithic consisting of interbedded mudrocks and laterally continuous, horizontally bedded, carbonate-rich sandstones (facies J) (Figs. 5.14C, 5.14D) with sparse to moderate bioturbation and wave ripples (facies D). The top of the succession consist of thick units of mudrocks as well as the red and green shales and paleosols of the Isfjorden Member (Fig. 5.23D, E, 6.6A).

Interpretation

0 – 83.5 m: The lower section is interpreted to represent a shallowing upwards sequence from offshore, offshore transition, up to lower shoreface deposits. Hummocky cross stratification, wave ripples and heterolithic bedding are interpreted to reflect a depositional environment dominated by strong wave action. The interpretation of a shallow marine environment is also supported by the presence of bioturbation and marine trace fossils. Observations of *Diplocraterion* in the lower part of the interval and *Skolithos* in the upper part of the interval may indicate an increase in energy regime from the lower to upper shoreface (Fig. 5.24).

83.5 – 145 m: The middle section is interpreted to comprise moderate to high energy shallow marine delta front to shoreface deposits (Fig. 6.4). This interpretation is based on the observation of bioturbation, an abundance of wave ripples and heterolithic sandstones. In the most sand-rich intervals in the upper part of the section, sandstones appear to be amalgamated and laterally continuous which indicate that these sandstones could be interpreted as a prograding barrier succession.

145 – 200 m: Wave ripples and bioturbation indicate a shallow marine environment, while thick accumulations of mudrocks are interpreted as a lower energy regime with deposition in interdistributary areas. Inter-changing lithologies and facies reflect a paralic depositional environment. Laterally continuity of sandstones with wave ripples and bioturbation can indicate sand barriers in a barrier island complex (Fig. 6.6A).

7.3 Eastern Spitsbergen

7.3.1 Hahnfjella (Hahn 15-1)

Hahnfjella is a 537 m tall mountain on the eastern coast of Spitsbergen (Fig. 7.1). This locality was originally the type section for the "Hahnfjella Suite" as defined by Pčelina (1980; 1983), which is the equivalent of the lower part of the De Geerdalen Formation. The Isfjorden Member was defined originally as a suite by Pčelina (1980; 1983), before being reduced to member rank by Mørk et al. (1999a) and represent the upper part of the De Geerdalen Formation. Three sections covering the De Geerdalen Formation were followed along prominent ridges (Hahn 15-1, 15-2 and 15-3). Hahn 15-1 was measured from close to the base the cliff-forming Botneheia Formation to the top of the mountain, covering both the Tschermakfjellet Formation and the De Geerdalen Formation (Fig. 7.6, Appendix C). Only the part of the log covering the De Geerdalen Formation is presented in this thesis, as the Tschermakfjellet Formation is largely covered at this locality.

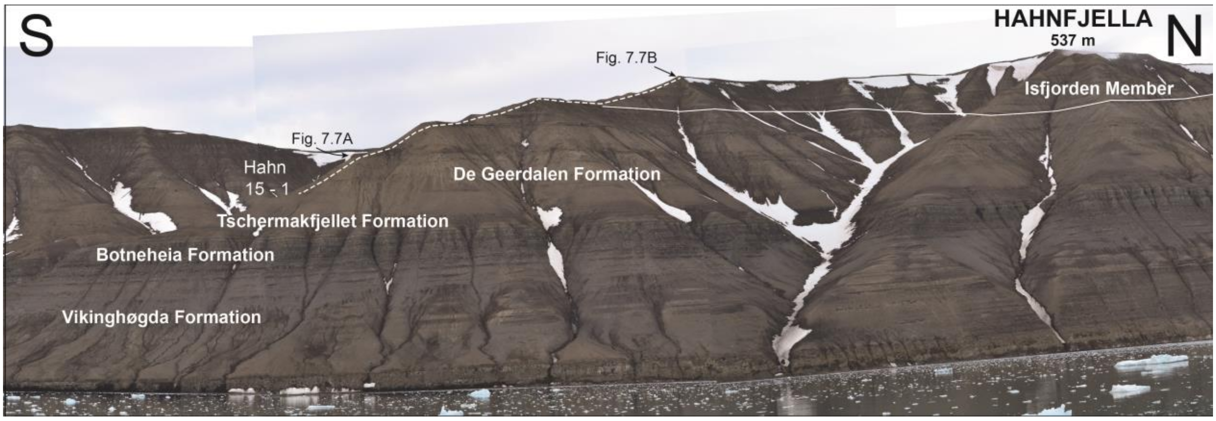

Figure 7.6: Photomosaic of Hahnfjella. The log trace for Hahn 15-1 is indicated to the left(white stippled line), covering most of the De Geerdalen Formation**.** The boundary between the De Geerdalen Formation and the Isfjorden Member is indicated by the white solid line.

The shales close to the shore belong to the Vikinghøgda Formation and are overlain by the cliff-forming Botneheia and Tschermakfjellet formations. The De Geerdalen Formation is found above the Tschermakfjellet Formation and consists of thin, but laterally extensive units of sandstone (Fig. 7.6). The section is dominated by mudrocks and shales, and only a few good outcrops of sandstones are not covered by scree (Fig. 7.7B). The lowermost part is composed of two major coarsening upwards sequences while the uppermost is dominated by minor coarsening upwards sequences.

Description

0 – 25 m: The lowermost sandstone bed measured in Hahn 15-1 consists of very fine sand with wave ripples (facies D) and intense bioturbation. The succession above is mostly scree-covered, but inferred to be mud-dominated.

25 – 48 m: The succession above the scree constitutes the first major coarsening upwards sequence. The lowermost beds consist of very fine to fine sandstone with minor bioturbation, wave ripples (facies D) and mud flakes. At 38 m is a small outcrop with 10 m of fine to medium-grained sandstones (Fig. 7.7B). The sandstones are horizontally bedded with moderate to intense bioturbation and features ripple cross lamination, wave ripples (facies D, Fig. 5.4F), low-angle and planar parallel stratification (facies E and F), and it contain abundant *Diplocraterion* (Fig. 5.25G). Tidal signatures such as mud drapes, flaser and wavy bedding are observed within the units. Loading structures is observed on the contacts between sandstone units.

48 – 103 m: This interval is covered and inferred to be shale-dominated

103 – 105 m: The covered interval is overlain by a 2 m thick sandstone unit consisting of fine sandstone with large-scale trough cross stratification grading upwards into planar-parallel stratification (facies F) and low angle cross-stratification (facies E, Fig. 5.6C). The sandstone marks the top of the second major coarsening upwards sequence.

105 – 190 m: The middle part of the mountain is mostly covered by scree. Only a few thin (< 50 cm) coarsening upwards sandstones with ripple cross lamination are exposed in the scree. They are often intensely bioturbated with marine trace fossils such as *Rhizocorallium* and *Skolithos.*

190 – 243 m: The remaining part of the succession is also scree covered, except for a few 40 – 60 cm thick coquina beds (facies L, Fig. 7.7A). At 230 m a coal bed was observed weathering out from scree.

Interpretation

0 – 25 m: The lowermost beds in Hahn 15-1 is interpreted as reflecting a shallow marine shoreface setting as indicated by wave ripples and intense bioturbation. Overlying shales are interpreted as offshore marine to marginal-marine.

25 – 48 m: The sandstones in this interval are interpreted as being deposited in a shallow marine high energy environment as indicated by wave ripples, low-angle cross stratification, *Diplocraterion* and coarse grain size. Mud drapes, flaser bedding and wavy bedding indicate a small tidal influence on the deposits.

48 – 103 m: The interval above the sandstone outcrop is mostly covered and inferred to be mud-dominated. It is interpreted as recording deposition in a low energy environment on the delta front.

103 – 105 m: The upwards change in facies from large-scale cross bedding to planar parallel stratification and low-angle cross stratification is interpreted as reflecting waning flow conditions. The mud-dominated part of the succession is interpreted as interdistributary bay deposits with locally lagoonal conditions.

105 – 190 m: The overall great amount of mud, intensity of bioturbation, and types of trace fossils are indicative of a fluctuating energy regime. During quiet periods, mud is deposited from suspension and the conditions are favourable for marine organisms and reworking of the sediments. The entire interval is interpreted as inter-fingering low-energy shallow marine and interdistributary areas to possibly coastal plain.

190 – 243 m: Coquina beds are interpreted to reflect deposition in interdistributary or lagoonal areas. The overall high amount of mud and presence of coal shales support a proximal delta plain interpretation.

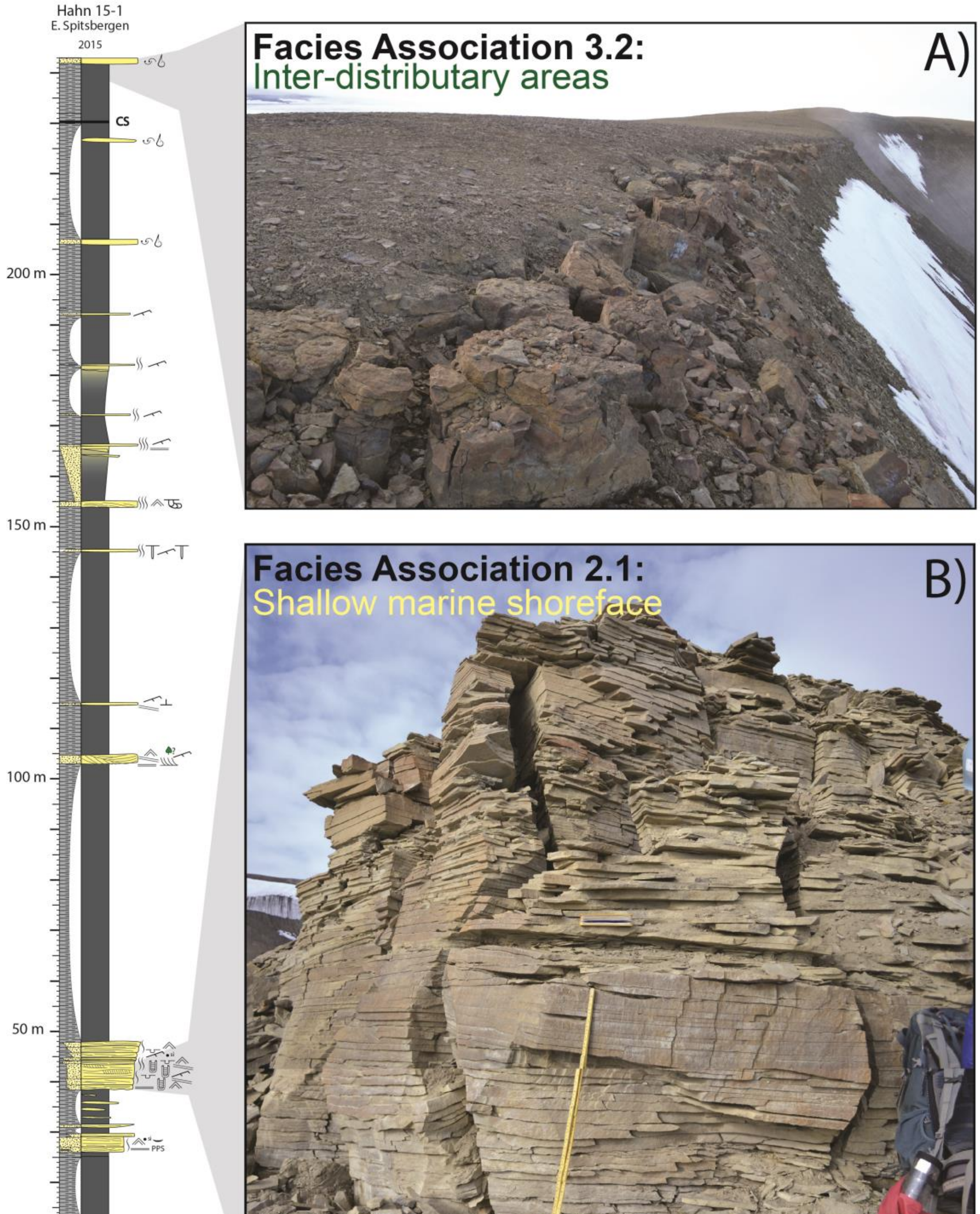

Figure 7.7: Correlation between outcrop and selected intervals in the log from Hahnfjella (Hahn 15-1) **A)** Laterally continuous siltstone bed with shell fragments close to the top of Hahnfjella. These beds occur at three distinct intervals and parts of units may be described as coquina beds (facies L). The underlying beds consist of paleosols and shales and reflect deposition in interdistributary areas (FA 3.2). **B)** Planar parallel stratified sandstones with *Diplocraterion*, wave ripples and bioturbation interpreted as Facies Association 2.1: Shallow marine shoreface deposits.

7.3.2 Teistberget (Teist 15-1)

Teistberget is a 425 m tall mountain on the western side of Storfjorden, north of Mohnbukta, easternmost in Sabine Land (Fig. 7.1). The De Geerdalen Formation on Teistberget has previously been studied by Knarud (1980) and Pčelina (1980). Due to time constraints during the fieldwork, only the lowermost 120 m of the De Geerdalen Formation was measured by Knarud (1980). The stratotype of the Teistberget Suite (Pčelina, 1980) is found on the north-eastern slope of the mountain towards the top. This unit is equivalent to the upper part of Kongsøya Formation of the Wilhelmøya Subgroup (Mørk et al., 1999a).

Two logs, Teist 15-1 covering the lower to middle part of De Geerdalen Formation and Teist 15-2 covering the upper part including the Isfjorden Member, was measured (Figs. 7.8, 7.9, Appendix C). Only Teist 15-1 is included in this thesis and the measured section covers the lowermost 216 m of the De Geerdalen Formation at the locality. Teist 15-2 continues directly from the top of Teist 15-1 and covers the upper part including the Isfjorden Member, and is presented in Haugen (in prep.).

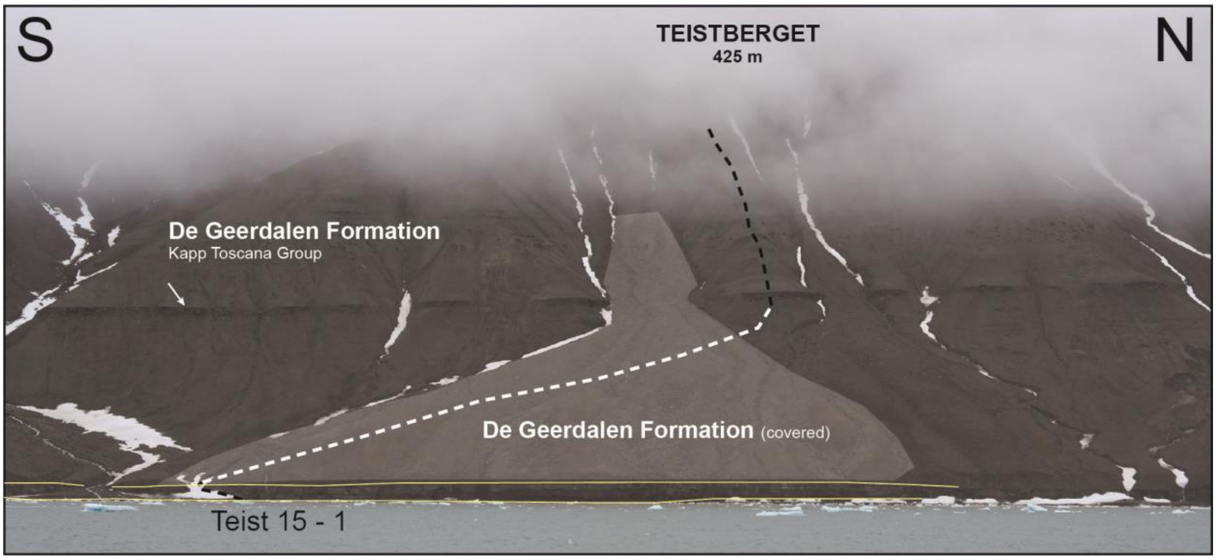

Figure 7.8: Log trace for Teist 15-1 on Teistberget, eastern Spitsbergen. Approximately 117 m of the section is covered by a colluvial fan. Two very laterally extensive outcrops of sandstone are present. The lowermost is located by the shoreline (outlined in yellow), while the upper outcrop is found above the colluvial fans in the slope and indicated by the white arrow.

Description

0 – 6 m: The lowermost beds closest to the shoreline consist of 6 m of fine-grained heterolithic sandstones and shales with large-scale trough cross-stratification (facies A, Fig. 7.9C), ripple lamination and abundant plant fragments. Foresets are draped with mud and a variation in thickness of foresets is observed (Fig. 7.9B). The sandstones are almost uniform in thickness and laterally continuous along the beach section.

6 – 123 m: The beach section is covered by a colluvial fan (Fig. 7.8). Mountain slope next to the fan are also observed to be largely scree-covered and the underlying bedrock is assumed to be mostly composed of shale.

123 – 127 m: A laterally continuous unit of sandstone with uniform thickness (ca. 4.5 m) can be seen above the colluvial fan in the middle of the slope (Fig. 7.8). It consists of two fining-upwards units from fine to very fine sandstone, with sparse bioturbation, large-scale trough cross stratification (facies A), plant fragments, loading structures and unidirectional current ripples. The uppermost part of the sandstone is low angle cross stratified (facies E) and moderately bioturbated. The entire sandstone unit is overlain by thin beds of coal shale.

127 – 148 m: Above the sandstone the succession is mostly scree-covered. At 139 m, there is 1 m of sandstone with planar parallel stratification (facies F), moderate bioturbation, wave ripples (facies D) and pyrite nodules. The succession above is covered, but seen to consist mostly of shale with occasional coal shales.

148 – 167.7 m: This interval includes laterally restricted channel-shaped fining-upward sandstones from fine to very fine sand with large-scale trough cross bedding (facies A), plant fragments and unidirectional ripple lamination (facies B, Fig. 7.9A). An odd observation regarding some of the sandstones is that although they are hard due to cementation they seem to rest unconformably upon friable and unconsolidated medium-grained sand. Primary sedimentary structures, such as parallel lamination is observed within the unconsolidated medium sand.

167.7 – 202 m: Lower part of this interval is mostly scree-covered, while the upper part include beds of fining-upwards units from fine to very fine sandstone with large-scale cross stratification (facies A), sparse to moderate bioturbation and planar parallel stratification (facies F). The units are overlain by coal shale (facies N). The upper part is also scree covered, except for a

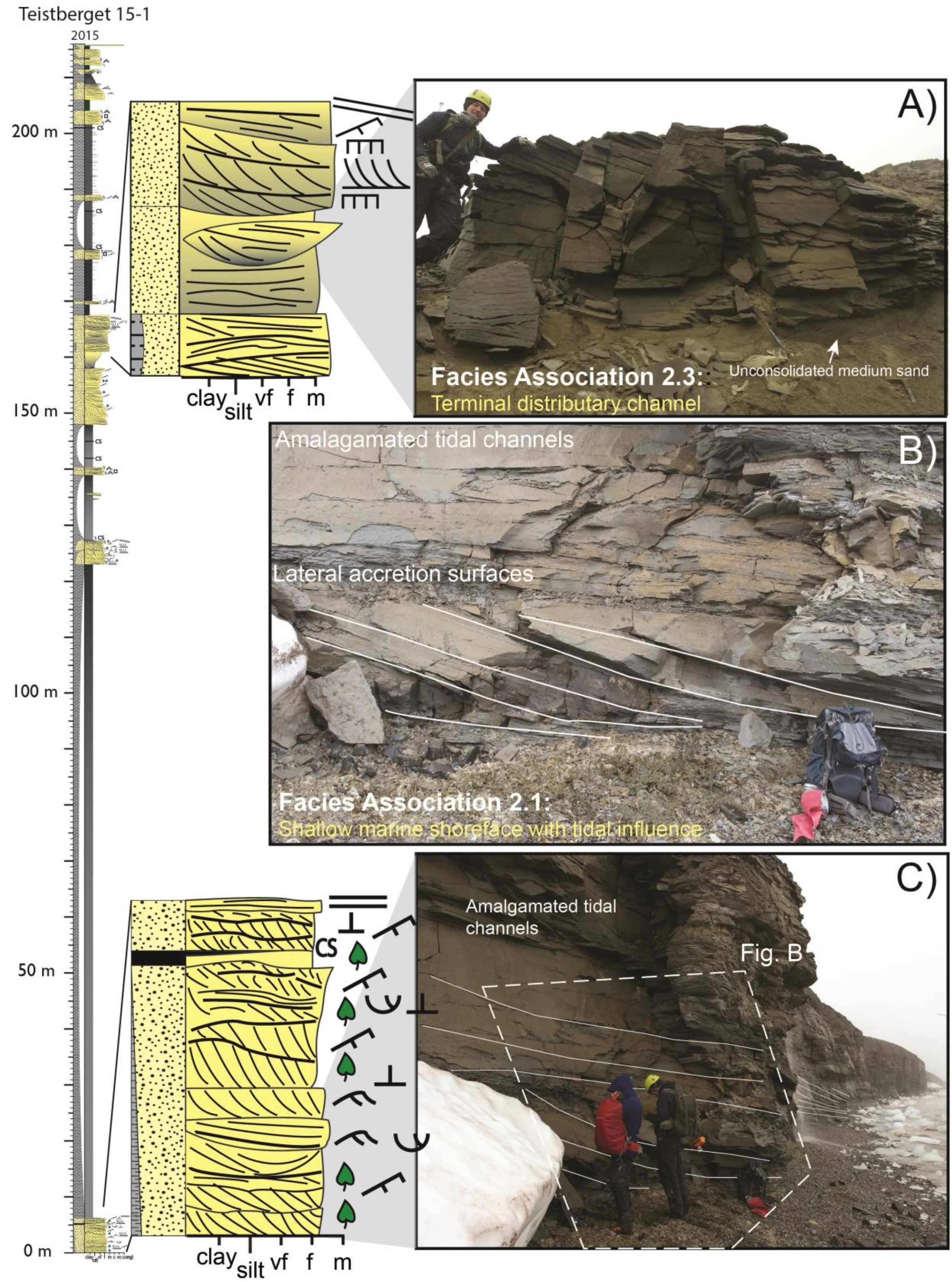

Figure 7.9: A) A large sandstone channel body resting unconformable on friable, unconsolidated medium sand. **B)** Lowermost beds closest to the beach displaying large-scale cross-stratification interpreted as lateral accreting tidal channels. These beds are interpreted as amalgamated tidal channels. **C)** The lowermost sandstones measured in the section are observed to be laterally continuous for hundreds of metres along the beach.

few interbeds of coal shale and a 1 m thick very fine-grained sandstone bed with ripple cross lamination (facies B, Fig. 5.2A), wave ripples (facies D) and bioturbation at 188.5 m.

202 – 216 m: The uppermost sandstones consist of upwards fining units from fine to very fine sand with wave ripples (facies D), plant fragments and pyrite nodules. Bioturbation is moderate in the lowermost units and less in the uppermost units. The section is slightly overlapping with the bottom of Teist 15-2.

Interpretation

0 – 6 m: Large-scale cross stratification, high amount of sand, mud drapes and abundant plant fragments indicate a shallow marine depositional environment. The lateral continuity of the beach section may indicate that the sands were redistributed along the shoreline by longshore currents. Large-scale cross stratified sandstones with mud draped foresets and variable foreset thickness are interpreted as migrating and laterally accreting tidal channels (*cf.* Dalrymple and Rhodes, 1995; Hughes, 2012).

The sandstone unit at Teistberget appears similar to another outcrop at Deltaneset on central Spitsbergen. The beds at Deltaneset are located at a similar beach section and consists of amalgamated fine-grained sandstones with large-scale trough cross stratification. The sandstones belong to the upper part of De Geerdalen Formation and are unconformably overlying a series of shales and thin sandstones with hummocky cross stratification. These deposits are interpreted to record a coarsening upwards unit from offshore marine to shoreface deposits. The amalgamated sandstones are interpreted as stacked tidally influenced channels. The beach section on Teistberget is also similar to equivalent units on Hopen (figs. 6C, 6E in Klausen and Mørk, 2014), where the sandstones have been interpreted as tidal and estuarine channels.

Below Myklegardsfjellet on the north-eastern side of Agardhbukta is another beach section exposing sandstones of the De Geerdalen Formation, which share similarities to the beach section on Teistberget. There, laterally extensive exposures of sandstones display unidirectional cross bedding (fig. 15.16 in Høy and Lundschien, 2011), abundant mud drapes and rhythmic wavy bedding. These features are common in tidally influenced environments and the cross bedded sandstones are in the study of Høy and Lundschien (2014) interpreted as fluvial or tidal channels.

6 – 123 m: The succession above the beach section is mostly covered by a colluvial fan, but inferred to consist of shales and mud deposited in an offshore marine environment.

123 – 127 m: The lateral continuity of the sandstones and presence of bioturbation may indicate that the sandstones represent a barrier island, overlain by back-barrier to lagoonal deposits. The coal shale and plant fragments above indicate a proximal position possibly on a paralic delta plain. Also, the scree-covered succession above is assumed to cover shales supporting the interpretation of back-barrier to lagoonal conditions above the barrier island. Alternatively, the sandstone can be interpreted simply as open-coast shoreface sandstones overlain by interdistributary mudrocks.

127 – 148 m: Planar parallel stratification, moderate bioturbation and wave ripples indicate a shallow marine depositional environment. Presence of pyrite and coal shales are interpreted as resulting from reducing conditions (Pettijohn et al., 1987) and may indicate deposition on a coastal environment, such as a tidal flat or lagoon. Alternatively, the deposits can be interpreted as interdistributary areas.

148 – 167.7 m: The sandstones lack marine indicators and the presence of plant fragments, as well as channel-shaped geometry is interpreted as reflecting a fluvial environment.

167.7 – 202 m: Bioturbation and wave ripples indicate a shallow marine influence on the succession. Coal shale and thick units of mud reflect a low energy environment with favourable conditions for vegetative cover. The deposits are interpreted as representing a coastal depositional environment, such as an interdistributary bay or a floodplain.

202 – 216 m: Wave ripples and moderate bioturbation indicate a shallow marine (delta front?) environment. Plant fragments and pyrite nodules may indicate reducing conditions during deposition which suggests a restricted depositional environment. In contrast to the underlying unit which is more mud-dominated, this interval contains sandstone with sedimentary structures indicating a higher energy environment.

7.3.3 Hellwaldfjellet (Hell 15-1)

Hellwaldfjellet is a 650 m tall mountain in southern Olav V Land. It is found on the northern margin of Storfjorden, directly south of Wilhelmøya (Fig. 7.1) and the highest point of the mountain is at located 677 m above the sea. Two logs were acquired at this location (Hell 15-1 and Hellwald 15-2, Appendix C). Hell 15-1 is included in this thesis and Støen (in prep.), while Hellwald 15-2 is included in Haugen (in prep.). The De Geerdalen Formation is approximately 250 m thick here and was measured from the base of the formation up to a prominent ridge formed by an igneous sill (Fig. 7.10).

Close to the shore, there is a few meters thick silty shale with ammonites and bivalves, most likely belonging to the Botneheia Formation and has been dated as Anisian (Smith, 1975). Above this, siltstones with siderite concretions and bivalves, of the Tschermakfjellet Formation occur in smaller outcrops up to the first prominent sandstone bed at the base of the De Geerdalen Formation (Smith, 1975). The De Geerdalen is also here, similarly to Wilhelmøya intruded by a system of dolerite intrusions close to the top of the mountain.

The uppermost beds on Wilhelmøya belong to the Wilhelmøya Subgroup (Worsley, 1973) and were observed during this field campaign to include a large tree trunk embedded in red sandstone. Worsley (1973) compared the “Wilhelmøya Formation” as defined from Wilhelmøya to similar deposits towards the top of Hellwaldfjellet and noted the similar development on the succession. Smith (1975) described the De Geerdalen Formation from Hellwaldfjellet and noted that the “Uleneset Member” (lower part of De Geerdalen) was slightly thicker (401m) and had a higher proportion of sand, compared to the equivalent unit on Tumlingodden, Wilhelmøya. The succession on Wilhelmøya and Hellwaldfjellet was by Smith (1975) assigned to a Norian age, based on palynological assemblages.

The De Geerdalen Formation is generally observed to include a higher amount of sandstone compared to equivalent stratigraphy on Wilhelmøya, consistent with the observations of

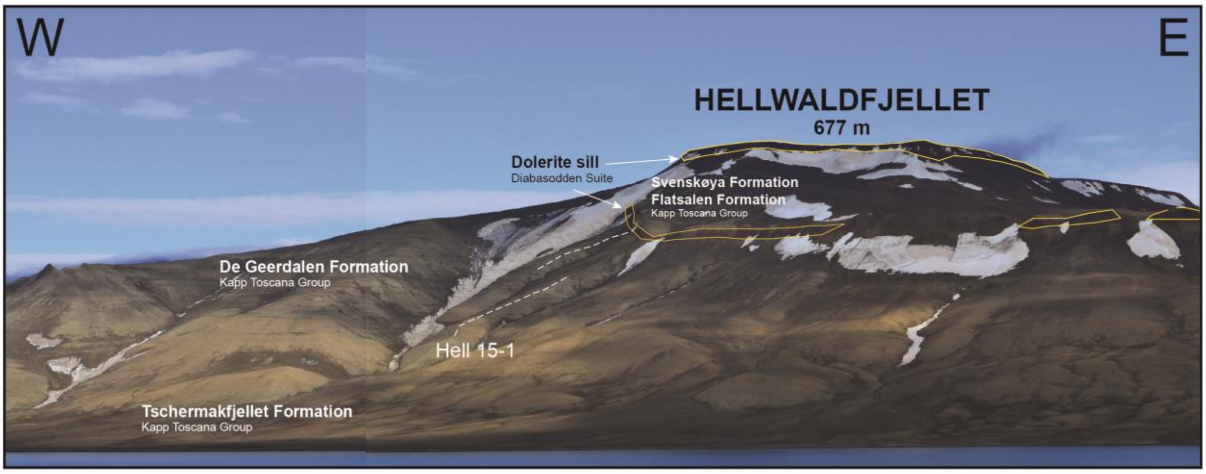

Figure 7.10: Overview photograph with annotated log trace for Hell 15-1 on Hellwaldfjellet on eastern Spitsbergen. The Tschermakfjellet Formation is overlain by the De Geerdalen Formation. The Wilhelmøya Subgroup, represented by the Flatsalen and Svenskøya formations is overlying a dolerite sill (outlined in yellow). The top of the mountain is also overlain by a dolerite sill (outlined in yellow and indicated with black arrows).

Smith (1975). At higher stratigraphic intervals the succession is mostly mud-dominated except for a few coquina beds. In general a similar development of facies is observed on Hellwaldfjellet and Wilhelmøya, where the De Geerdalen is also intruded by igneous sills.

Description

0 – 70 m: Discontinuous very fine to fine sandstones with large-scale trough cross bedding (facies A) to small-scale cross stratification (facies B, Fig. 5.1C) and wave ripples (facies D, Fig. 5.4B, 5.4D) are forming prominent benches. Most of the lower beds are horizontally bedded and stratified (facies F) while the upper part is dominantly low-angle cross stratified (facies E, Fig. 5.6E). An overall coarsening upward trend in grain sizes is observed. Sandstones are commonly extensively calcite-cemented and display cone-in-cone structures (facies J, Fig. 5.14B). Calcite cement is locally so pervasive that some of the sandstones appear structureless (facies G, Fig. 5.8A). Bioturbation is sparse in the lower part of interval, but is observed to increase upwards in the section to moderate and locally intense. At 44 m is an intensely bioturbated sandstone bed with wave ripples, plant fragments and marine trace fossils (*Skolithos*, *Rhizocorallium* and *Diplocraterion*). Areas between sandstones are commonly scree-covered and inferred to cover mud and silt.

70 – 96 m: This interval is mostly covered. Lower contact to underlying sequence is sharp and the succession is inferred to consist dominantly of fine-grained material.

96 – 103 m: Heterolithic deposits with alternating silty shale and very fine and fine sandstone with plant fragments, small-scale ripple cross stratification (facies B) moderate bioturbation and marine trace fossils (*Skolithos*). The lower part of a coarsening upwards unit contains alternating siltstones and very fine sandstones with hummocky cross stratification (facies H).

103 – 152 m: The lower part is mostly scree-covered and when excavated, observed to include paleosols and occasional discontinuous horizons of coal shale. The beds above are also largely scree-covered except for a few interbedded sandstones (0.5 to 1 m thick) with small-scale ripple cross lamination.

152 – 171 m: Lower part is covered while the upper part consists of a coarsening upwards unit from silty shale to fine sandstone with unidirectional current ripples (facies B) and marine trace fossils (*Skolithos*). It is overlain by approximately 2 m of paleosols and coal shales. At 168 m is fine sandstone with wave ripples (facies D) and another coarsening upwards unit with small scale cross bedding (facies B) that is again overlain by paleosols.

171 – 253 m: Mostly covered, except for individual benches of calcareous sandstones and possibly coquina beds (facies L, Figs. 5.17D, E). Sandstones are usually discontinuous and include ripple cross stratified (facies B) and planar parallel stratified (facies F).

253 – 270 m: Towards the top of the measured section is a dolerite sill with a thickness of approximately 17 m.

270 – 278 m: A siltstone with a thin coquina bed is found above the first dolerite sill (facies L, Fig. 5.17F), and the succession above is mostly shales.

Interpretation

0 – 70 m: Upwards facies change from horizontally bedded (facies F) to low angle cross stratification (facies E) as well as upwards increase in bioturbation is interpreted to reflect an upwards shallowing trend in facies. Intensely bioturbated sandstone with marine trace fossils and physical sedimentary structures (large and small-scale cross stratification and wave ripples) also indicate a shallow marine environment, possibly a delta front or shoreface environment. Plant fragments indicate proximity to terrestrial and paralic environments. Low-angle cross stratified sandstone is interpreted to represent upper shoreface to foreshore succession, while large-scale cross bedded sandstone is interpreted as subaqueous dunes. Abundant bioturbation in sandstones may reflect a low-energy environment with fluctuating sediment supply and favourable conditions for burrowing organisms to rework the sediment during quiet periods. Calcite cement and cone-in-cone structures are related to diagenesis. Interbedded and covered muddy successions represent deposition of fine-grained material in interdistributary areas or on the shoreface.

70 – 96 m: This interval is mostly covered. The sharp contact to underlying sequence is interpreted as representing a marine flooding surface and the succession is interpreted as recording deposition in a low energy marine environment.

96 – 103 m: Observations of moderate bioturbation, marine trace fossils, wave ripples and ripple cross lamination indicate a shallow marine environment. Hummocky cross stratification is interpreted as indicating an offshore transition or lower shoreface environment, while plant fragments indicates coastal conditions.

103 – 152 m: Coal shales and paleosols indicate subaerial exposure, possibly on a lower delta plain or in interdistributary areas. Interbedded sandstones could be interpreted as crevasse splays indicating a proximal fluvial floodplain environment. The vertical transition from marine facies to interfingering terrestrial sandstone is interpreted as being conformable and reflects a distal position to an active distributary system (Reading and Collinson, 1996).

152 – 171 m: Wave ripples and *Skolithos* is interpreted as reflecting a shallow marine environment, while paleosols indicate deposition in a paralic environment.

171 – 253 m: Covered intervals are inferred to comprise muds and shales deposited in interdistributary areas or in a floodplain environment based on observation of coquina beds and sandstones with ripple cross stratification.

253 – 270 m: The dolerite sills belong to the Diabasodden Suite.

270 – 278 m: The uppermost part of the measured section Hell 15-1 above the first dolerite sill is interpreted as interdistributary areas. The shales are interpreted as recording the transition to the overlying Flatsalen Formation.

7.4 Wilhelmøya

7.4.1 Tumlingodden (Wilhelm 15-1)

Wilhelmøya is a small island (120 km^2), situated on the southern end of Hinlopenstretet, north-east of Olav V land on Spitsbergen (Fig. 7.1). This is the most northern and remote exposure of the De Geerdalen Formation in the study area. Tumlingodden is found on the eastern point of Wilhelmøya (Fig. 7.11). Two logs of the De Geerdalen Formation were measured at this location, with Wilhelm 15-1 covering the lower to middle part of the formation included in this thesis. Tumlingodden 15-1, covering the Isfjorden Member is presented in Haugen (in prep).

Previous research

The Swedish geologist G. J. De Geer landed on Tumlingodden in 1901 to investigate the Triassic and Jurassic exposures (Buchan et al., 1965). Later work on the Triassic to mid-Jurassic exposures of the island was presented by Frebold (1930), Worsley (1973) and Smith (1975). Buchan et al. (1965) presented a log from Wilhelmøya and described the sandstones as calcareous and lens-shaped, containing plant fragments, worm tubes, small brachiopods and iron staining. They also described shell limestones in the uppermost part of the mountain, which is probably equivalent to facies L (Coquina beds, Fig. 5.17). A section covering the mid-Jurassic Wilhelmøya Subgroup has previously been measured E. P. Johannessen and S. Olaussen (Mørk et al. 1999a; Vigran et al., 2014).

The rocks exposed on the island are primarily sandstones and shales of Late Triassic to mid-Jurassic age. The Upper Triassic is represented by the Tschermakfjellet and De Geerdalen formations. It remains unclear whether the lowermost beds below De Geerdalen Formation on Wilhelmøya actually belong to the Tschermakfjellet Formation (Vigran et al., 2014). Buchan et al. (1965) assigned the lowermost beds on Wilhelmøya to the Botneheia Formation.

Fieldwork in 2015 conducted by other members of team show that the shale beds below the De Geerdalen Formation are grey with siderite nodules and they do not form the prominent cliffs typical of the cemented shales of Botneheia Formation. It is therefore reasonable to assume that the lowermost beds on Wilhelmøya belong to the Tschermakfjellet Formation.

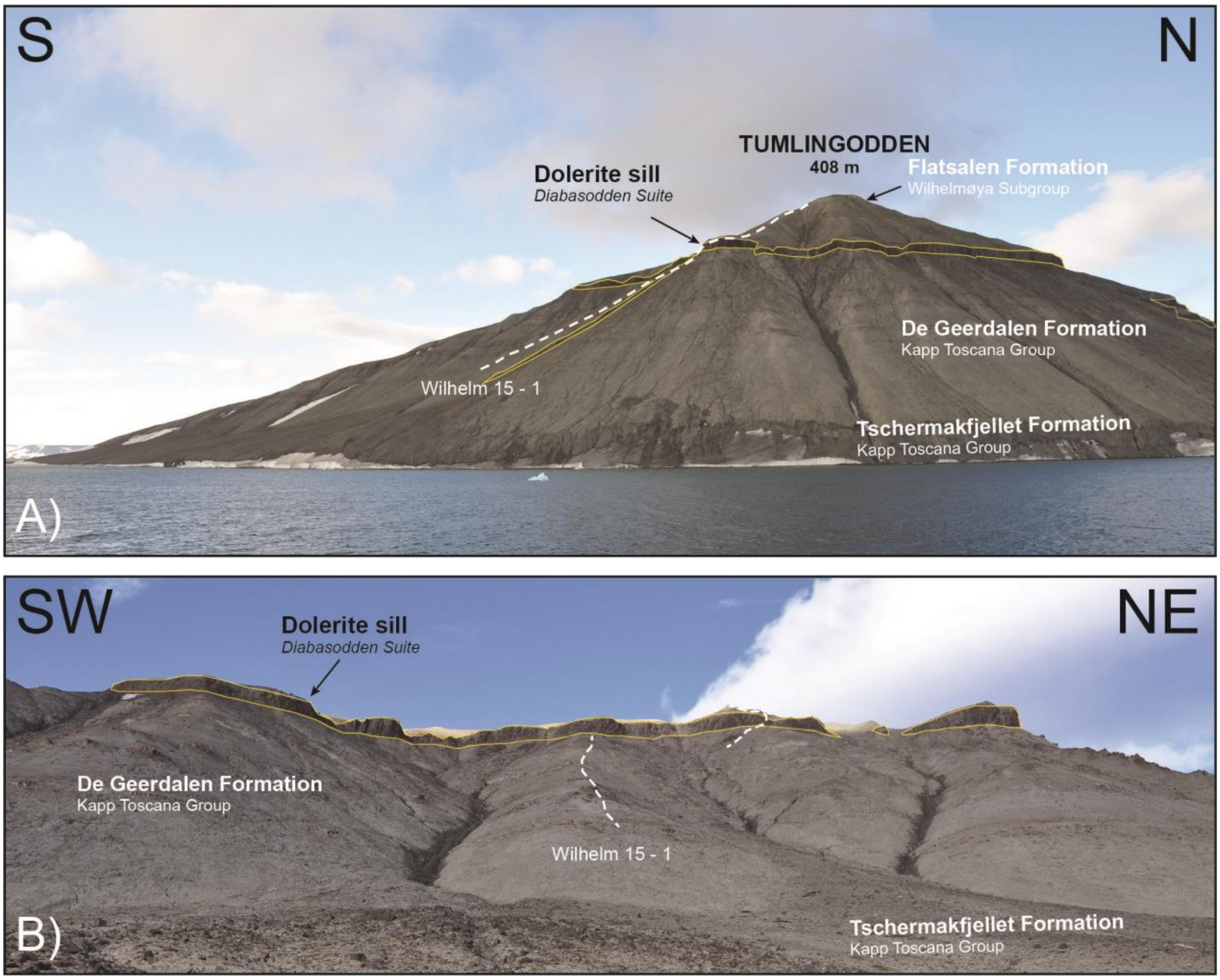

Figure 7.11: A) View towards the west of Tumlingodden on Wilhelmøya. A dolerite dike (outlined in yellow) can be seen directly adjacent to the Wilhelm 15-1 log trace, feeding the sill in the upper part of the section. B) Log trace of Wilhelm 15-1 is indicated by stippled white line. The measured section includes several sandstone bodies that form prominent benches in the scree.

A system of dolerite intrusions, most likely of the Diabasodden Suite, penetrates the succession at different levels. The thickness of the dolerite sill varies laterally from 5 m to 20 m, with an approximate average thickness of 10 m. This igneous system of dikes and sills makes accurate measurements of sandstone thickness difficult as they are displaced. Thickness measurements are therefore best considered as approximate around the intrusions.

The De Geerdalen Formation is overlain by exposures of the Wilhelmøya Subgroup represented by the Flatsalen, Svenskøya and Kongsøya formations (Mørk et al., 1999a; Vigran et al., 2014). The uppermost rocks on Wilhelmøya are another dolerite sill capped by the marine black shales of the Upper Jurassic Agardhfjellet Formation.

"The Wilhelmøya Formation"

The "Wilhelmøya Formation" was first defined by Worsley (1973), who described it from several localities in eastern Svalbard; Tumlingodden on Wilhelmøya, Hellwaldfjellet and the south-eastern island of Hopen. He assigned parts of the formation to a "Liassic" age, i.e. Rhaetian – Hettangian and noted the similarities between the succession on Wilhelmøya and on Hellwaldfjellet further south.

The lower boundary of this formation was defined by a "Basal Member", consisting of yellowish to grey sandstones overlying a bed with pebbles of sandy limestone, phosphorite and quartzite. Above this Basal Member, Worsley (1973) defined the "Bjørnbogen Member" which includes an interval of dark grey shales containing thin interbeds of plesiosaur bones in a condensed section. This unit is followed by a succession of siltstones and sandstone with occasional micaceous mud flake conglomerates, defined as the "Transitional Member" (Figs. 7.12A, B) and in our opinion equivalent to the Flatsalen Formation. The uppermost part of the Wilhelmøya Formation consist of yellowish grey sandstones and was assigned to the Tumlingodden Member (Fig. 7.12B), which corresponds to the Svenskøya Formation.

Smith (1975) considered the Wilhelmøya Formation only as a northerly extension of the De Geerdalen Formation. The remaining succession below the formation was defined as the Uleneset Member of largely Norian age, and the rank of the "Wilhelmøya Formation" was reduced to member. Smith (1975) and Pickton et al. (1979) argued that the considerable distances between localities where the "Wilhelmøya Formation" outcrops (Wilhelmøya, Hopen and Kong Karls land) prevents a robust correlation of the unit. They did not accept the boundary, as it was not easily demonstrated as a mappable horizon. The formation has since been redefined as the Wilhelmøya Subgroup (Mørk et al., 1999a)

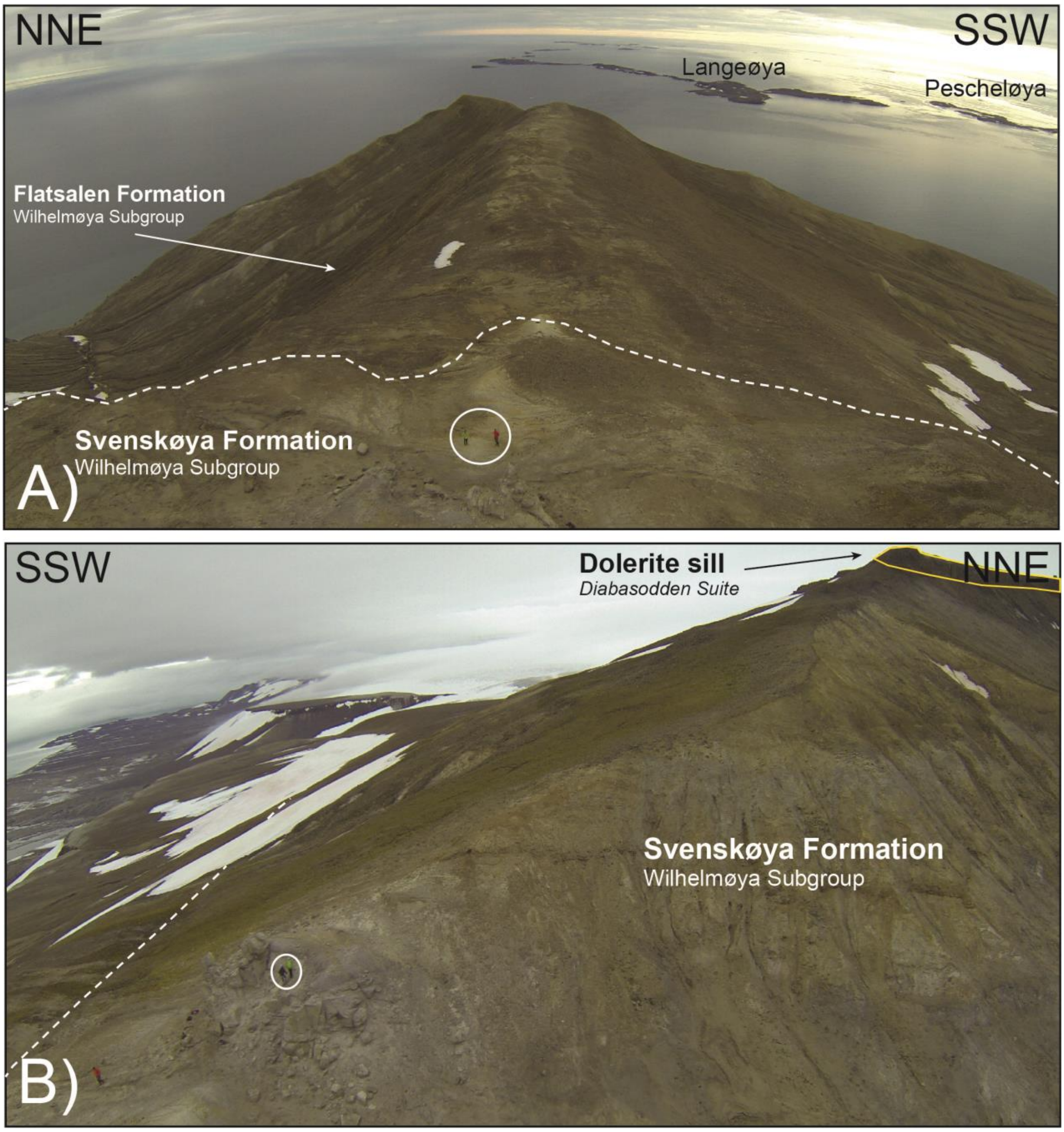

Figure 7.12: Aerial photographs of the Wilhelmøya Subgroup on the slopes above Tumlingodden, Wilhelmøya. **A)** The step-formed plateau corresponds to the Transitional Member of Worsley (1973), which consists of siltstones with interbedded sandstones that weathers purple, underlying the Tumlingodden Member. **B)** View towards the northwest of the Flatsalen and Svenskøya formations. The succession above the cliff-forming sandstone to the lower left consists of very friable fine-grained, yellowish grey sandstone and corresponds to the Tumlingodden Member of Worsley (1973). The lower boundary was defined by Worsley (1973) approximately 10 m below the sandstone cliff. Note geologists for scale (white circles). The upper dolerite sill is outlined in yellow. Aerial drone photographs courtesy of Alexey Deryabin (NPD), 2015.

Description

0 – 35 m: The lowermost bed measured in Wilhelm 15-1 is horizontally bedded and consists of cemented very fine to fine sandstone with ripple lamination (facies B) and cone-in-cone (facies J). The lower bed contact display loading structures and the succession is mostly scree-covered both above and below this sandstone bed. The sandstone bed appears to pinch out laterally over a few meters, but as the section is poorly exposed it could be the result of the scree cover. Above the scree-covered interval is a 3 m thick unit of sandstone with swaley cross stratification (facies H), low-angle cross stratification (facies E), plant fragments, minor bioturbation and marine trace fossils (*Diplocraterion*).

35 – 78 m: The middle section is also poorly exposed, except for a few small outcrops of very fine to fine sandstone. Primary sedimentary structures are large scale cross bedding (facies A, Fig. 5.1E), ripple cross lamination (facies B), wave ripples with mud drapes (facies D) and planar parallel stratification (facies F, Fig. 5.7B). The sandstones are often hard due to carbonate cement and display cone-in-cone structures (facies J, Fig. 5.14A). Their geometry is often lenticular and channel shaped with large-scale trough or tabular cross bedding. Several beds in this interval are intensely bioturbated (Fig. 5.25B) and include a diverse range of marine trace fossils (*Skolithos, Diplocraterion, Rhizocorallium* and *Teichichnus*) (Figs. 5.25C, D).

78 – 94 m: The succession above the intensely bioturbated sandstone is mostly covered silty shales. These shales are overlain by a sandstone with channel geometry and consisting of fine sand with large scale trough cross bedding fining upwards to very fine sandstone with unidirectional current ripple lamination and plant fragments. It is overlain by accumulations of coal shale and shales with large wood fragments, most likely derived from larger tree trunks. Such trunks are known to occur in the De Geerdalen Formation from Edgeøya (Enga, 2015) and Hopen (Lord et al., 2014b).

94 – 117 m: The succession above the coal shales is mostly covered and inferred to be mud-dominated (facies M) up to a coquina bed.

117 – 121 m: Above the covered interval is a laterally continuous siltstone with coquina beds (Fig. 5.17A, C). Most of the succession above is also covered, but laterally to the right is a upwards coarsening sandstone large-scale trough cross bedding (facies A) in the lower part of the unit grading upwards into small-scale cross bedding (facies B) with a rusty red colour.. The overall geometry of the sandstone is channel-shaped.

121 – 155 m: Above the channel-shaped sandstone, most of the succession is covered up to a dolerite sill.

155 – 166 m: The thickness of the dolerite sill is variable, but average thickness is approximately10 m. The sill is seen to be fed by a dike to the right of the measured section (Fig. 7.11A).

166 – 175 m: Above the sill the measured interval consists of two minor coarsening upwards units, with a coal shale horizon in between. Sedimentary structures in sandstones are small-scale ripple cross lamination (facies B, Fig. 5.2B) and possibly wave ripples (facies D). Units are usually surrounded by covering scree and also affected by the emplacement of the dolerite sill. The relation to other stratigraphic units is therefore not clear in the field.

175 – 207 m: This interval is mostly covered and inferred to be mud-dominated (facies M).

207 – 222 m: The lowermost bed in this interval is a coquina bed (Fig. 5.17B). This is overlain by laterally restricted fining upwards sandstones with large-scale tabular cross bedding (facies A, Fig. 5.1A), climbing ripple cross stratification (facies C, Fig. 5.3B), plant fragments and ripple cross lamination (facies B, Fig. 5.2D) interbedded with coal shales (facies N) and paleosols (facies O, Figs.5.19A, 5.21B). Observations of red and green shales (Fig. 5.23A) were recorded in Tumling 15-1 (Haugen, in prep), measured at a similar stratigraphic level.

222 – 237 m: The uppermost sequence measured in Wilhelm 15-1 consists of 15 m shale with occasional rootlets, overlain by a hard and cemented sandstone bed. The bed was not studied in detail, but assumed to be the Slottet Bed.

Interpretation

0 – 35 m: With the Tschermakfjellet Formation is present below these sandstones a shallow marine to transitional environment is to be expected at this level in the stratigraphy. Loading structures are interpreted as resulting from deposition of water-laden sediment on top of an unconsolidated and soft sub-stratum. In general the lowermost part of the succession is interpreted as a lower shoreface to shallow marine delta front environment. This is based on the observations of swaley cross stratification, intense bioturbation and presence of marine trace fossils (*Diplocraterion*).

35 – 78 m: A low-energy shallow marine environment is interpreted from the bioturbated intervals. This interpretation is supported by the observations of wave ripples with mud drapes, planar parallel stratification and ripple lamination. Large-scale cross stratified sandstones are interpreted as shallow marine bars.

78 – 94 m: The sandstones are interpreted as fluvial channels with crevasse splays based on field observations of large-scale cross bedding and channel geometry. The underlying and overlying deposits are interpreted as floodplain fines. Wood fragments, coal shales and plant fragments indicate presence of vegetation in overbank areas.

94 – 117 m: This covered interval is overlying a bed with coal shales and wood fragments. No observations were made that may suggest a marine influence or the presence of a flooding surface. This interval is therefore interpreted as mudrocks deposited on a floodplain and possibly also in interdistributary areas.

117 – 121 m: The coquina bed is interpreted as representing interdistributary area, while the sandstones is interpreted as a fluvial or distributary channel based on geometry and associated facies. The red colour is interpreted to reflect a content of iron oxides, most likely due to hematite grain-coatings.

121 – 155 m: This interval is mostly covered mudrocks or shales and is here interpreted as representing deposition in a low-energy, proximal to possibly paralic delta plain environment.

155 – 166 m: Dolerite sill is interpreted to belong to the Diabasodden Suite.

166 – 175 m: Interpretations for this part of the interval is challenging since the succession is disturbed by the dolerite sill. However, the presence of coal shale and ripple cross lamination can be interpreted as reflecting a proximal delta plain environment.

175 – 207 m: This interval is mostly covered shales and associated facies such as coal shale indicate low energy conditions during deposition. This interval is interpreted as representing floodplain fines.

207 – 222 m: The observations of the uppermost part of the section in Wilhelm 15-1 indicate that the rocks were deposited in a fluvial floodplain environment. Paleosols, coal shales and plant fragments are interpreted to reflect a low-energy and sub-aerially exposed environment, possibly a coastal plain / delta plain. Large-scale cross bedding represent channel-infill, while climbing ripple cross stratified sandstones are interpreted as crevasse splays. Interbedded shales, paleosols and coal shales are interpreted as indicating the formation of soils and establishment of extensive vegetation cover on the floodplain. This helps create stable conditions on overbank and hinders lateral migration of channels (Bridge, 2006; Nichols, 2009). It is important to emphasize that although vegetation is important in stabilizing and binding floodplain fines, it is the flood flow that is the major governing factor determining a rivers capability of eroding banks and transporting sediment (Bridge, 2006). Rapid subsidence also favours accumulation of fine-grained material over channel sandstones (Nichols, 2009).

222 – 237 m: Shales with rootlets are interpreted as floodplain fines, while the Slottet Bed is interpreted as a condensed shelf deposit (Mørk et al., 1999a).

7.5 Barentsøya

Barentsøya is the fourth largest island of Svalbard with a total landmass of 1288 km^2 and lie east of Spitsbergen and north of Edgeøya (Fig. 7.1). The island is dominated by flat plateau mountains in the west, while the glacier Barentsjøkulen covers approximately 560 km^2 of the eastern area of the island (Lock et al., 1978).

Barentsøya lie within the Eastern Svalbard Dolerite Belt and the northern areas of the island hosts a prominent dolerite cluster (Burov et al., 1977). These dolerites are responsible for the irregular physiographic shoreline morphology of the island and also locally increased maturities in the organic rich intervals of the Botneheia Formation (Brekke et al., 2014; Senger et al., 2014). An increase in fractures in sandstones of the De Geerdalen Formation is documented around the igneous intrusions (Senger et al., 2014).

Structural contour mapping by Lock et al. (1978) on the top of the Botneheia Formation revealed that the structure of Barentsøya and Edgeøya consists of several domes and smaller synclines. On Barentsøya a southwest-northeast trending antiform form the most prominent feature, while the narrow strait of Freemansundet between Barentsøya and Edgeøya hosts a sharply defined monocline (Lock et al., 1978).

A strong fluvial dominance in the De Geerdalen Formation has been known from the eastern areas of Barentsøya and Edgeøya for some time (Lock et al., 1978; Mørk et al., 1982), especially when compared with equivalent units further west on Spitsbergen (Rød et al., 2014). Only the lowermost part of the formation is exposed on Barentsøya. The coarsening upwards sequences in the De Geerdalen Formation on the eastern islands has in the literature been interpreted as distributary channels of a major prograding delta system (Vigran et al., 2014). The observations and interpretations presented in this thesis are consistent with these interpretations.

7.5.1 Krefftberget (Krefft 15-1)

Krefftberget is a small mountain group located in the bay of Anderssonbukta, on the southwestern part of Barentsøya (Fig. 7.1). The lowermost succession consists of a conformable sequence of shales and occasional beds of siltstone belonging to the Vikinghøgda, Botneheia and Tschermakfjellet formations. The shales are intruded by an igneous sill with approximately uniform thickness throughout the mountain slope. A 100 m thick vertical sequence of the lower part of the De Geerdalen Formation is described from this location (Fig. 7.13, Appendix C). The sandstones are poorly exposed and most of the succession is covered by scree. Therefore, only a general description is given here and interpretations are largely based on relationships to under- and over-lying units.

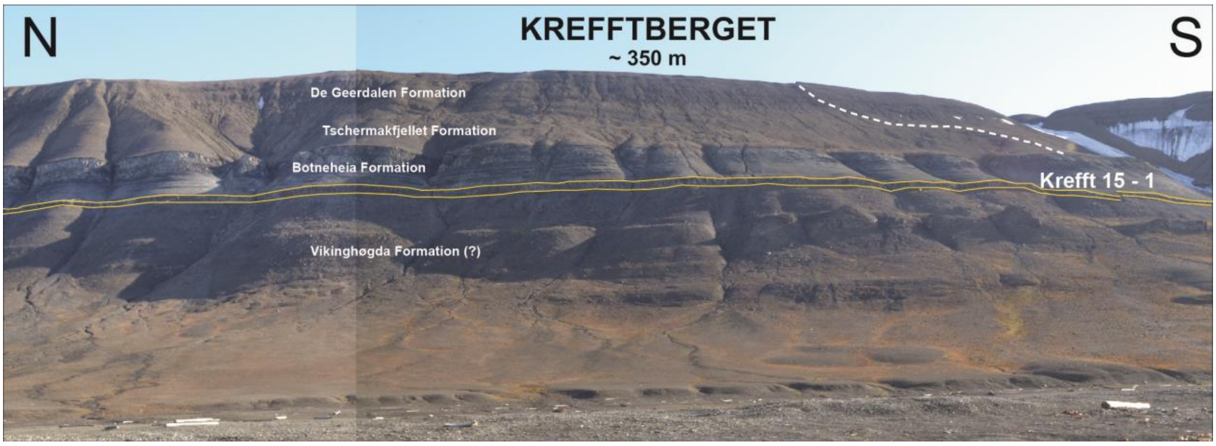

Figure 7.13: Annotated photograph of the Triassic succession on Krefftberget, southwestern Barentsøya. The lowermost beds are shales belonging to the Vikinghøgda Formation. A dolerite sill (outlined in yellow) intrudes the overlying cliff-forming Botneheia Formation at the base, while Tschermakfjellet and De Geerdalen formations are found above.

Description

0 – 2 m: The lowermost beds in Krefft 15-1 constitute cliff-forming black silty shales with phosphate nodules.

2 – 13 m: The cliff-forming black shales are in turn overlain by purple, marine and silty grey shales with weathered siderite nodules.

13 – 97.5 m: The first prominent sandstone is found at the top of the purple shales. It consists of very fine sand with horizontal bedding (facies F) and ripple lamination (facies B). Following ca. 12 m of scree is approximately 4 m thick sequence of very fine to fine sandstones with ripple lamination (facies B) and horizontal bedding (facies F). Individual beds are up to 1 m thick and contain moderate bioturbation and plant fragments. Sandstones above are generally planar parallel and low-angle cross stratified (Fig. 5.6D) with minor bioturbation and wave ripples.

97.5 – 101 m: The uppermost beds on Krefftberget consist of a 3 m thick unit of fine to medium sandstone. The lowermost part of the sandstone consists of alternating low-angle stratification and planar parallel stratification (facies E and F), while wave ripples (facies D) and ripple lamination is observed in the middle part of the unit. The uppermost part features large-scale tabular cross stratification (facies A) and ripple cross lamination (facies B). The uppermost beds include a thin unit of coal shale (facies N) with a lateral extent of a few meters.

Interpretation

0 – 2 m: The shales are black owing to a high content of organic material and represent the Botneheia Formation.

2 – 13 m: The shales are purple due to the weathering of siderite nodules and represent the Tschermakfjellet Formation.

13 – 97.5 m: Although the succession is covered by scree it is possible to interpret a shallow marine origin for the succession. This is mostly based on the observation of wave ripples and planar parallel stratification suggesting a high-energy environment. Minor to moderate bioturbation also support the interpretation of a shallow water depth. Scree-covered areas are inferred to be mostly consisting of shales and fine-grained material. These shales are interpreted to represent deposition of mud in a lower to middle shoreface environment.

97.5 – 101 m: Large-scale tabular cross stratification and presence of wave ripples indicate a high-energy environment influenced by wave energy. The presence of coal shale indicates deposition in a proximal environment, possibly on a paralic delta plain or in an interdistributary bay.

7.5.2 Svartnosa (Svart 15-1)

Svartnosa lies on the north-western side of Barentsøya, southwest of Mistakodden and directly north of Krefftberget (Fig. 7.1). The succession is the most sand-rich locality in the entire study area and features a vertical cut of a delta lobe including the prodelta, delta front and delta plain sandstones and shales (Fig. 7.14). The lowermost beds by the shore primarily consist of silty, grey shales and may belong to the Tschermakfjellet Formation. They were not studied, but samples were taken for palynology.

A 180 m long vertical section of the De Geerdalen Formation was measured along a prominent ridge (Fig. 7.14A, Appendix C). The section does not cover the entire part of the largest sandstone bodies, but the stacking patterns and development of parasequences can be seen as a coarse representation of the thicker units. The lower parts of the measured section appear to have been vertically displaced or downfaulted (Fig. 7.14). The adjacent strata appear unaffected, so the displacement is unlikely to be caused by tectonic movements, but is more probably caused by a recent to sub-recent landslide. Such landslides and displacements of larger blocks of sandstone are known from elsewhere on Svalbard (Gjelberg, 2010). A possible consequence of this displacement is that the uppermost 30 m in Svart 15-1 constitutes a repeated section. But the uppermost part of the slide is eroded so the section is considered to be more or less complete.

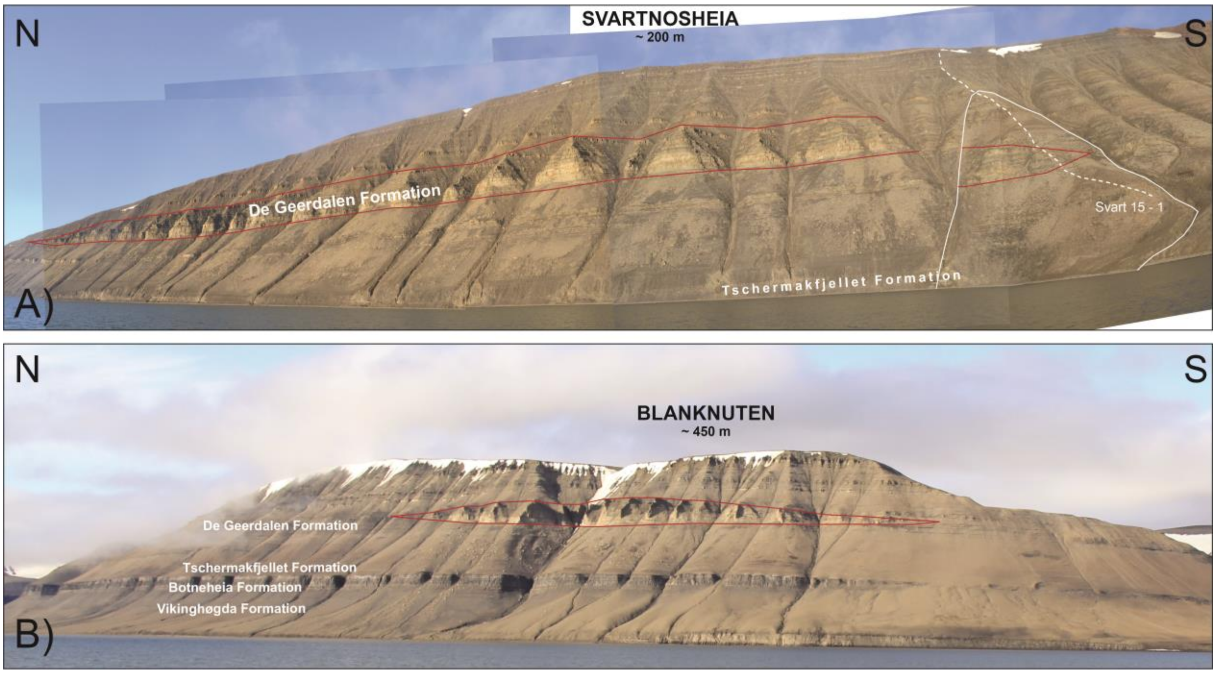

Figure 7.14: Photomosaic of sandstones on the eastern islands. Note the ellipsoidal geometry. **A)** Log trace (stippled white line) for Svart 15-1 on Svartnosa, Barentsøya. The white solid line indicates areas that have been vertically displaced. The largest sandstone cliff has a lateral extent of approximately 1 km and an estimated thickness of 20-25 m. **B)** Similar exposures of the De Geerdalen Formation on Blanknuten, Edgeøya. Photo: (B) Atle Mørk, 2012.

The sandstones on Svartnosa share many similarities with a sandstone body on Blanknuten (Rød et al., 2014). They have an ellipsoidal geometry and are observed to be laterally continuous for hundreds of meters (Figs. 6.5B, 7.14). The sandstones are cliff-forming at both locations and occur at similar stratigraphic levels.

The succession consists of five major coarsening and shallowing upwards units. These consist, in turn, of minor coarsening upwards units capped by marine shales. These are interpreted to represent minor flooding surfaces resulting from autogenic processes, i.e. smaller fluctuations in relative sea level due to switching of deltaic lobes. In general the entire succession is interpreted to represent a vertical cross-cut of a deltaic lobe and several parasequences recording the progradation of a river-dominated delta. The largest sandstone body is interpreted as amalgamated distributary channels and mouth bars, as well as delta front sandstones.

Description

0 – 6 m: The lowermost beds in section Svart 15-1 consist of large-scale cross stratified sandstone (facies A, Fig. 5.1F) with mud draped foresets, climbing ripple sandstone (facies C) and horizontal bedding (facies F).

6 – 24 m: The beds above the base of the section is mostly covered by scree and inferred to be mud-dominated.

24 – 60 m: Above the shales is a thick cliff-forming coarsening upward sequence consisting of a thick sand-rich interval of amalgamated fine to medium sandstones characterized by large-scale cross stratification and climbing ripples (facies C, Fig. 5.3A). Loading structures and strong evidence for soft-sediment deformation are found on the lower parts of sandstone units (facies I_1, Fig. 5.11B). The uppermost beds of the sequence consists of fine sand with planar parallel stratification grading into low-angle cross stratification (Figs. 5.6A, 5.6B). It is capped by mudrocks (facies M) with thin lamina of coal shale (facies N).

60 – 84 m: The succession in this interval is covered in the lower part and observed to consist of shales coarsening upwards to very fine and fine sandstone with hummocky cross stratification (facies H). The sandstones are overlain by approximately 6 m of large-scale trough cross-stratified medium-grained sandstone. Sandstones also include ripple cross stratification and plant fragments (facies A). These sandstones occur as lenticular-shaped sandstone bodies above the sandstone cliff (Fig. 6.5). Above is a bed of fine sandstone that is intensely bioturbated and contain mud flakes and plant fragments.

84 – 90 m: This interval consists of ca. 4 m of shales and a 2 m thick unit of fine sandstone with large-scale trough cross-stratification (facies A) grading into planar parallel stratification (facies E).

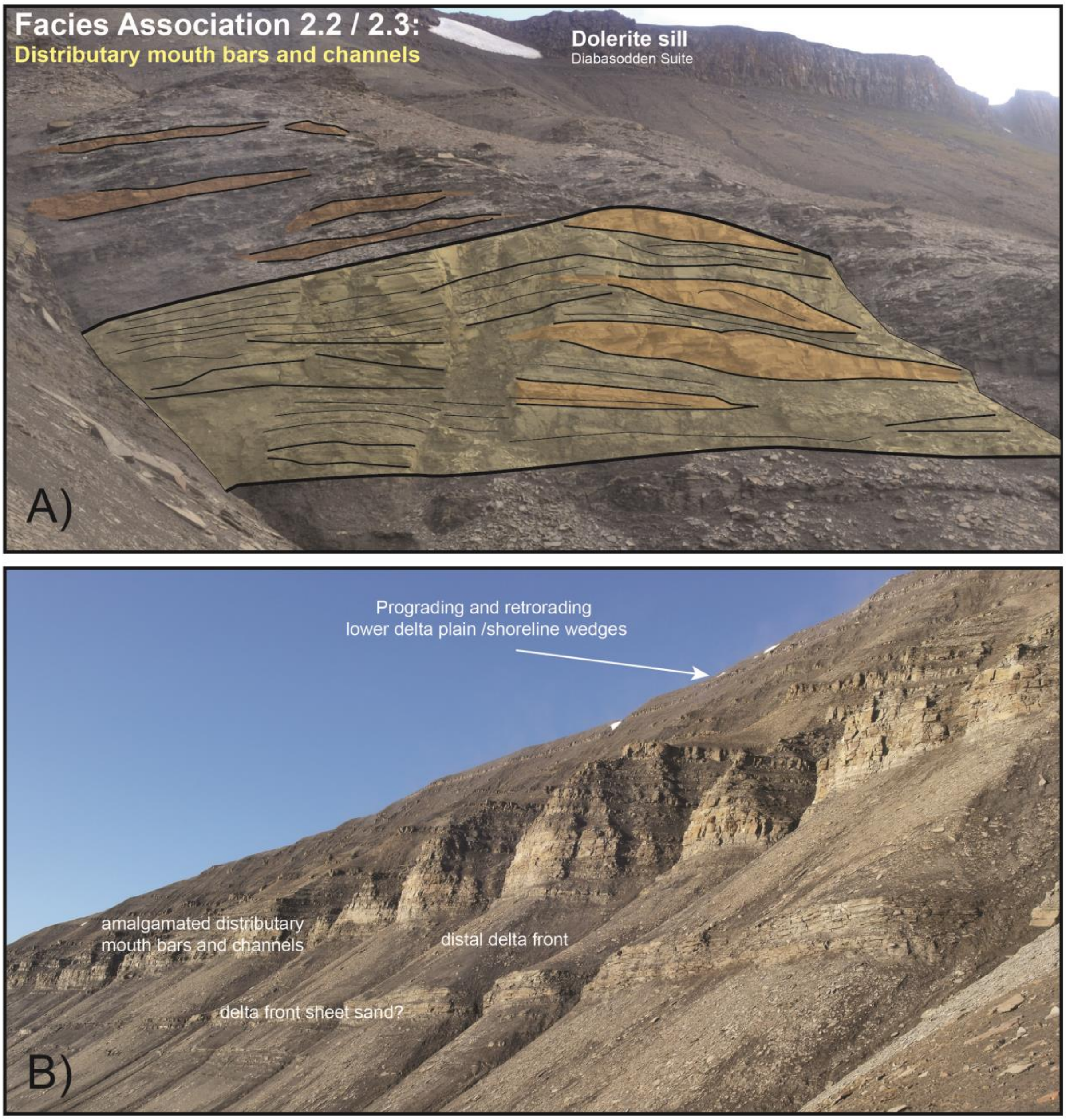

Figure 7.15: A) Facies Association 2.2 / 2.3: Amalgamated distributary channels and mouth bars indicating a fluvial dominated delta front. Orange colouring outlines individual sandy mouth bars, while yellow colouring outlines the extent of sandstones. **B)** Oblique view looking north of the sandstone rich interval at Svartnosa. The lowermost laterally continuous sandstone body below the cliff is interpreted as distal delta front and is typical on the seaward-dipping slope associated with the distal margin of distributary mouth bars. These sandstones are most likely equivalent to distal banks as interpreted by Knarud (1980).

90 – 117 m: The lower parts are characterized by heterolithic bedding where silty shales are interbedded with fine sandstone displaying ripple lamination, wave ripples, hummocky cross stratification and marine trace fossils *(Diplocraterion).* Above this is four minor coarsening upwards units from silty shale to very fine to fine sandstones with wave ripples (facies D), low angle cross-stratification (facies E, Fig. 5.6E), hummocky cross stratification (facies H), *Skolithos* 'pipe rock', intense bioturbation, plant fragments, and rhythmic alternations between small scale ripple cross lamination (facies B) and planar parallel stratification (facies F) (Fig. 5.7D).

117 – 142 m: Similar development as underlying sequence with three minor coarsening upwards units consisting of very fine to fine sandstones hummocky cross stratification (facies H, Fig. 5.9B) at the base. The uppermost units include fine sandstone with wave ripples (facies D), ripple cross lamination (facies B) and horizontal bedding (facies F).

142 – 154 m: Two minor coarsening upwards units from silty shale to very fine sandstone with hummocky cross stratification (facies H) and soft-sediment deformation (facies I_1).

154 - 182 m: The uppermost part of Svart 15-1 consist of five minor coarsening upwards sequences from silty shale (facies M) to planar parallel stratification with bioturbation (facies F). The steep slope and covering scree restricted detailed observations of the beds, but when viewed from distance (Fig. 7.14A,) the sandstones are seen to be laterally continuous for hundreds of meters across the mountain side and most likely extend beyond the exposures at Svartnosa.

Interpretation

0 – 6 m: The lowermost beds are interpreted as shallow marine delta front due to the presence of climbing ripples, interpreted to result from rapid sedimentation on the delta front. Plant fragments are also considered indicative of a proximal environment.

6 – 24 m: This interval is mostly covered, but interpreted as distal delta front shales based on relationship to under- and overlying facies.

24 – 60 m: The sandstones are interpreted as amalgamated distributary mouths bars and distributary channels, based on their large dimension and high amount of sand (Fig. 6.5, 7.15A, B). Soft – sediment deformation and loading structures indicates rapid deposition and loading of sand on the delta front. Low angle cross stratification and horizontal bedding in the upper part are interpreted as the uppermost part of a distributary mouth bar. Coal shales on the top are interpreted as indicative of delta top deposits or interdistributary areas.

60 – 84 m: The lower shale unit is interpreted to represent a flooding and an abrupt deepening of facies. Hummocky cross stratified sandstone is interpreted to constitute offshore transition to lower shoreface. The large-scale trough cross bedded sandstones is interpreted as amalgamated distributary channels based on internal bounding surfaces and their overall geometry (Figs. 6.5).

84 – 90 m: The covered shales are interpreted as interdistributary mudrocks, while sandstones are interpreted as smaller (terminal?) distributary channels.

90 – 117 m: The lower heterolithic interval and hummocky cross stratification indicates deposition in the offshore transition zone. A gradual transition through the lower shoreface to a proximal shallow shoreface setting is interpreted. This interpretation is based on the observations of marine trace fossils (*Diplocraterion)* and hummocky cross stratification in the lower part. Ripple lamination, wave ripples, abundant *Skolithos* trace fossils (Fig. 5.25A), intense bioturbation and plant fragments in the upper part indicates a minor shallowing upwards sequence from offshore transition to a shallow shoreface setting influenced by wave action. Coarsening upwards patterns are interpreted as representing repeating smaller shoreline progradations. The general heterolithic nature of the deposits is simply attributed to the alternating energy regime common on storm-dominated shelves (Fig. 6.3).

117 – 142 m: Same interpretation as underlying beds. Hummocky cross stratification indicates deposition in an offshore transitional to lower shoreface environment. Upwards increasing abundance in wave ripples and horizontal bedding indicate a shallowing upwards trend in facies to a shallow shoreface setting. The heterolithic sequence is interpreted as indicating a storm-dominated offshore environment, as no tidal indicators are observed.

142 – 154 m: This interval is interpreted as being deposited in a storm-dominated prodelta shelf environment to a lower shoreface or distal delta front setting. This interpretation is based on the presence of hummocky cross stratification and observations of soft-sediment deformation.

154 – 182 m: Lateral continuity of sandstones, combined with physical sedimentary structures and bioturbation indicates that the sandstones were deposited in a shallow marine high-energy environment. The uppermost part of the succession could thus be interpreted to represent distal responses to shoreline progradation elsewhere in the study area (Fig. 7.15B).

7.5.3 Mistakodden (Mistak 15-1)

Mistakodden is located on the north-westernmost point of Barentsøya (Fig. 7.1). It is dominated by a beach section consisting of mudrocks and shale that coarsens upwards into a more sandy succession towards the top (Figs. 7.16A, 7.16B). The beach is protected by several igneous intrusions located offshore to the west, which is more resistant to coastal wave erosion and probably responsible for the formation of the small embayment.

Sandstones belonging to De Geerdalen Formation are found further north along the beach and an approximately 50 m long section was measured (Fig. 7.16B; Appendix C). Lock et al. (1978) noted that the Botneheia Formation consisted of much sandier facies, compared to elsewhere on the eastern islands. They suggest that Mistakodden and areas north and south of Freemansundet were located within shallow areas of the basin in the Anisian, with the development of shoal-water conditions.

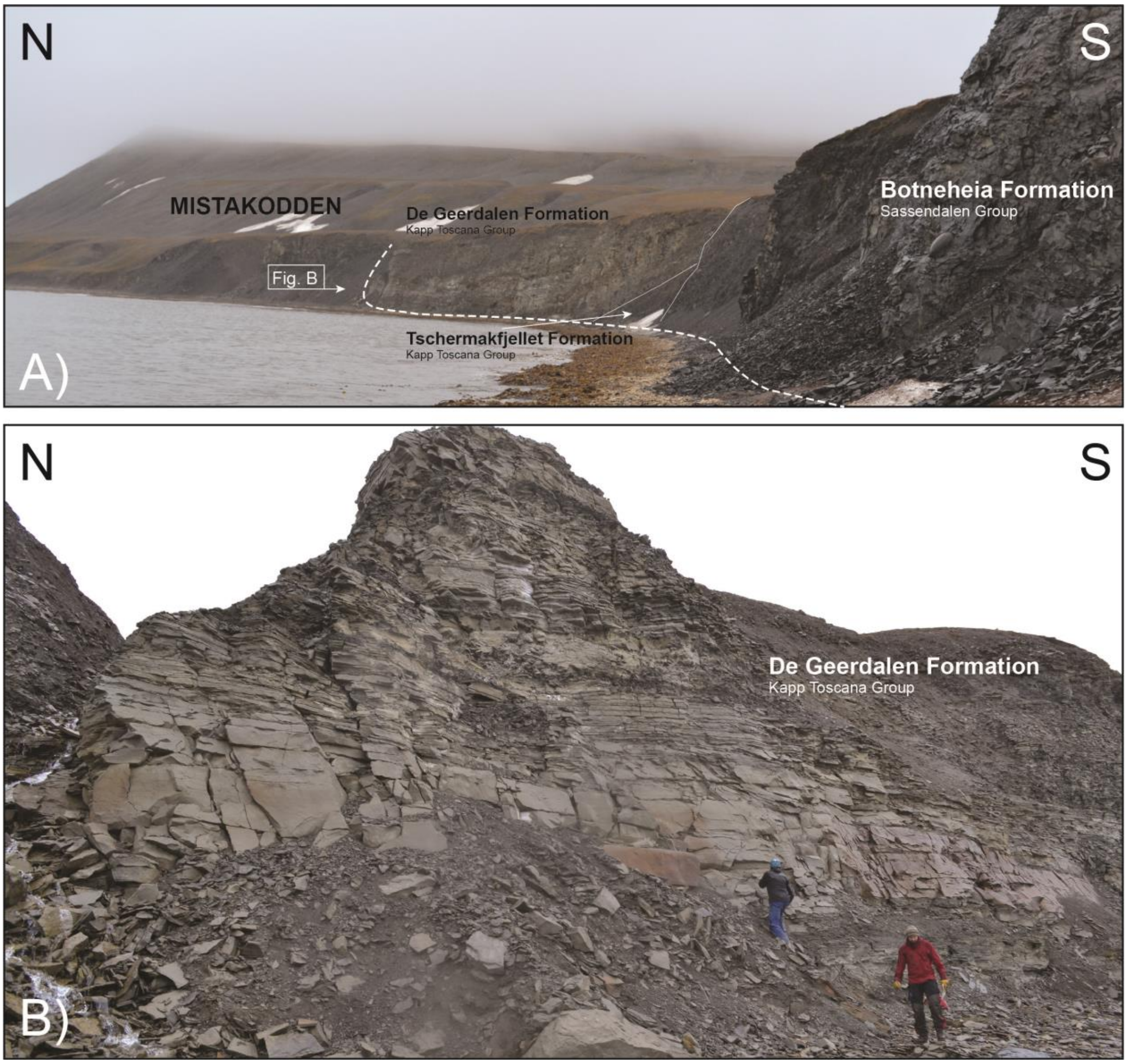

Figure 7.16: A) Overview photo looking east, with the log trace for Mistak 15-1 (white stippled line). **B)** Upper sandy unit on the far end of the beach section. Geologists for scale.

The succession on Mistakodden is characterized by complex folding and deformation of both sandstones and shales (Figs. 7.17A, 7.17B, 7.17C, 7.171D). This adds to the structural complexity of the rocks and makes accurate stratigraphic correlations difficult (Lock et al., 1978). The Tschermakfjellet Formation is apparently missing as the De Geerdalen Formation is directly overlying the shales of the Botneheia Formation (Lock et al., 1978). The beds directly below the sandstones consist of grey shales with minor siltstone beds and are here interpreted as representing the Tschermakfjellet Formation.

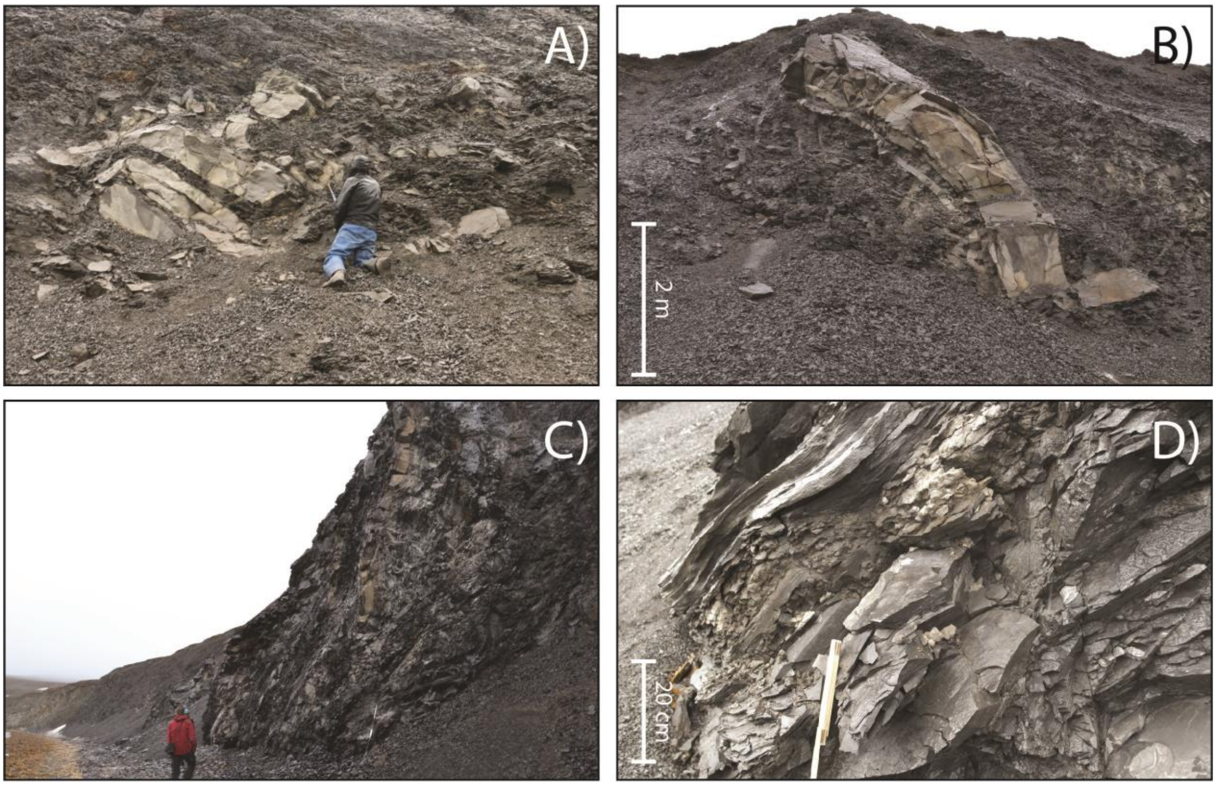

Figure 7.17: **A)** Deformed sandstone surrounded by deformed shales. **B)** Displaced sandstone surrounding the marine shales of the Botneheia Formation **C)** The shales appear to be the most affected by the deformation, being close to vertical. **D)** Phosphate nodules characteristic of the Botneheia Formation.

Several mechanisms have been proposed to explain the apparent folding and minor faulting occurring at Mistakodden: (1) Edwards (1976b) suggested that the succession on Mistakodden must have formed by syn-sedimentary growth faulting, similarly to the sediments exposed on Kvalpynten on southwestern Edgeøya (Edwards, 1976a) and Klinkhamaren on north-western Edgeøya (Osmundsen et al., 2014; Rød et al., 2014). Mistakodden is positioned directly north of these growth faults, parallel to the suggested southern paleocurrent direction (Edwards, 1976b; Osmundsen et al., 2014; Rød et al., 2014); (2) Folding and deformation is also suggested to be caused by the emplacement of the nearby igneous dolerite sills in poorly consolidated sediments (Lock et al., 1978); (3) Edgeøya and Barentsøya are located on stable platform areas (Dallmann et al., 2015) and to a limited degree affected by the West-Spitsbergen Fold-and-Thrust-Belt. Given the distal position to the western Fold-and-Thrust Belt, it seems unlikely that the folding in Triassic sediments as seen on Mistakodden is related to this tectonic activity. The most probable reason for the disturbance of sandstones and shales on Mistakodden seems to be related to the dolerites.

Description

0 – 16 m: The lowermost beds measured in Mistak 15-1 consists of deformed black shales (Fig. 7.17C) with phosphate beds (Fig. 7.17D), intense bioturbation and *Thalassinoides* trace fossils, pyrite nodules, bone fragments and ammonoid impressions.

16 – 35 m: The black shales are overlain by a thin interval consisting mostly of grey shales with thin beds of siltstone.

35 – 48 m: The shale succession is overlain by a ca. 3 m thick coarsening upwards unit from silty shale to fine sandstone with minor bioturbation and planar parallel lamination (facies F, Fig. 5.7A). This is in turn overlain by ca. 10 m of shales with smaller nodules and thin laminae of siltstone.

48 – 85.5 m: This interval covers the uppermost sandy part of the section at Mistakodden (Fig. 7.16B). The lower part is dominated by a heterolithic interval of very fine sandstones with abundant plant fragments, wave ripples (facies D), large- scale cross stratification (facies A). It is also characterized by loading structures and soft-sediment deformation features (facies I_1, Fig. 5.11A). The middle part of the sandstone succession is primarily fine, horizontally bedded sandstones with planar parallel stratification, plant fragments and ripple cross lamination with mud drapes. The uppermost part contains more abundant wave ripples and plant fragments compared to underlying units.

Interpretation

0 – 16 m: The combined observations of ammonoid imprints, phosphate beds and *Thalassinoides* suggest that the lowermost beds belong to the Botneheia Formation.

16 – 35 m: The silty shales below the sandy beach section is interpreted as representing a thin interval of the Tschermakfjellet Formation, in contrast to Lock et al. (1978) as it fine-grained lithologies and not sandstones. The grain size is also observed to be coarser than the underlying Botneheia Formation and no observations of phosphate nodules, trace fossils or fossils were made. The sediments represent deposits characteristic of outer shelf to inner shelf depositional environments.

35 – 48 m: The base of the De Geerdalen Formation is interpreted at the lower coarsening upwards unit. The overlying shales are interpreted as marine shales deposited in an offshore to lower shoreface setting.

48 – 85.5 m: This interval is interpreted as representing a fluvial dominated, shallow marine delta front environment. This interpretation is based on the observations of plant fragments and wave ripples suggesting that the sandstone were deposited in a shallow marine environment close to the delta front. Mud drapes on ripples foresets can indicate a slight tidal influence on the succession. Loading structures in the lower reaches of sandstone, soft-sediment deformation and generally a high amount of sand may indicate that the sandstones were deposited due to soft-sediment collapse due to rapid deposition close to a fluvial sediment source.

In general, the sandstones on Mistakodden features many similarities to sandstones further south on Skrukkefjellet, Klinkhamaren and Kvalpynten (Osmundsen et al., 2014). They are generally very fine to fine-grained, contain plant fragments and are dominated by small-scale cross, unidirectional cross bedding and soft sediment deformation. At Skrukkefjellet and Klinkhamaren, this has been interpreted to represent delta front and delta top (channel) sand (Hynne, 2010) and fluvially dominated delta front deposits (Glørstad-Clark, 2011; Osmundsen et al., 2014). It is not straightforward to determine whether the sandstones on Mistakodden are in fact growth faults. The high sand content of the beach section is peculiar, and sandstones appear thicker in the northern part of the beach, pinching out towards the south (Fig. 7.16). This variable thickness of sandstone bodies also occur in the growth faulted intervals on Edgeøya, further south (Osmundsen et al., 2014).

7.5.4 Farken (Fark 15-1)

Farken is a 200 m tall mountain and constitute the northern slope of Grimheia which is a mountainous area in the north-western part of Barentsøya (Figs. 7.1, 7.18). It is situated due east of Mistakodden and directly west of Frankenhalvøya and has not previously been visited by sedimentologists. The beds closest to the shoreline are mostly scree-covered, but on the slopes of the mountain sandstones and shales belonging to the De Geerdalen Formation can be accessed. A 70 m vertical section was measured covering a small interval of the De Geerdalen Formation (Fig. 7.18A, Appendix C).

Description

0 – 21.4 m: The units in this interval commonly occur as minor coarsening upwards units from very fine to fine sandstones with wave ripples (facies D), mud drapes, plant fragments, low angle cross-stratification (facies E), ripple lamination (facies B) and large–scale trough

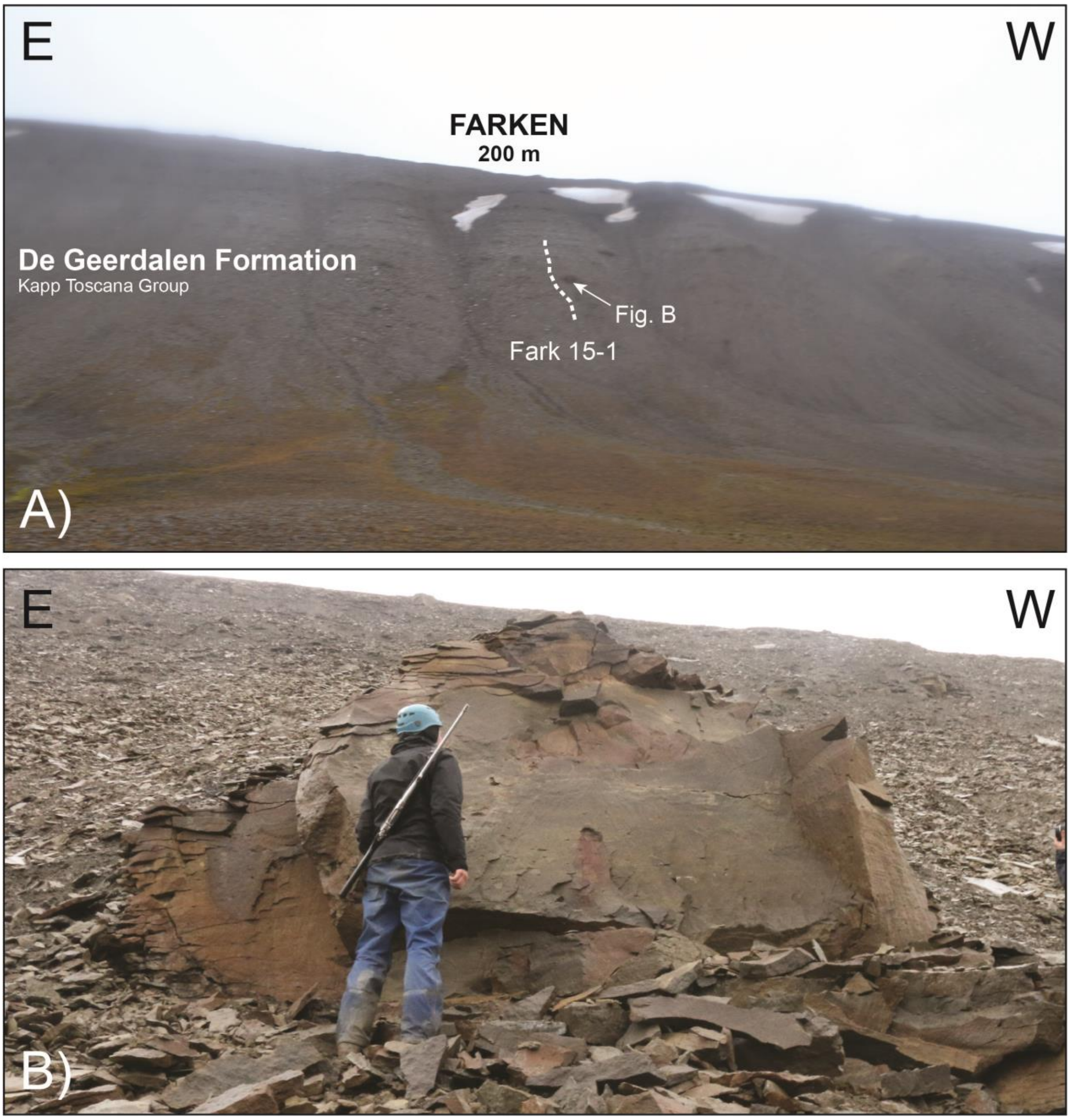

Figure 7.18: A) Log trace for Fark 15-1 on the northern side of Barentsøya. Sandstones are mostly thin and laterally continuous. The thickness of the logged section is approximately 65 m. **B)** Lenticular-shaped fining-upwards sandstone body with large-scale trough cross stratification, mud drapes and bioturbation. The sandstone becomes increasingly planar parallel stratified towards the top and is interpreted as a solitary distributary channel. Geologist for scale.

cross bedding(facies A). The sandstones are interbedded with shale and usually covered.

21.4 – 40 m: The lowermost part is covered and inferred to be mostly composed of shale. The uppermost part of this interval includes two minor coarsening upwards units from mud to very fine sand with wave ripples (facies D), low angle cross stratification (facies E) and planar parallel stratification (facies F). The second unit is capped by two thin coal beds.

40 – 67 m: Overlying the coal shales are 3 m of covered section and a ca. 4 m thick channel-shaped sandstone unit with large-scale trough cross stratification, mud drapes and bioturbation at in the lower part (Fig. 7.18B). The sandstone becomes planar bedded in the uppermost part of the unit. Above is ca. 17.5 m of scree cover, before a bed with low-angle cross stratified sandstone (facies E) is found towards the top.

Interpretation

0 – 21.4 m: Combinations of wave ripples, low angle cross-stratification and mud drapes point towards a shallow marine depositional environment influenced by tides and waves. The dimensions of the sandstones and combination of sedimentary structures such as large-scale cross stratification and ripple lamination indicate that the sandstones were deposited close to the shoreline, possibly as small distributary channels or mouth bars.

21.4 – 40 m: Coarsening upwards units with wave ripples are interpreted to reflect a shallow marine shoreface environment. Coal shales indicate a proximal or paralic depositional environment, possibly on the lower delta plain.

40 – 67 m: The sandstone is interpreted as a (terminal?) distributary channel based on the channel-shaped geometry and fining upwards trend, with large-scale trough cross bedding in the lower part. Bioturbation and close affiliation to underlying coal shales also indicate that the sandstone was deposited in a proximal coastal plain.

8. Discussion

Fifteen facies (Table 1) and eight facies associations (Table 2) have been defined for the De Geerdalen Formation in eastern Svalbard. Their distribution will be discussed in terms of main governing factors and relative influence of depositional processes (waves, tides and fluvial input). In general, the facies sequences and their stratigraphic architecture are consistent with those expected to occur in a prograding and dynamic deltaic environment influenced by both fluvial, shallow marine and paralic processes (Fig. 6.1) (Bhattacharya and Walker, 1992; Reading and Collinson, 1996; Bhattacharya, 2006). Previous sedimentological studies of the De Geerdalen Formation in eastern Svalbard (Lock et al., 1978; Knarud, 1980; Mørk et al., 1982; Klausen and Mørk, 2014; Rød et al., 2014) generally interpret the depositional environment as regressive shallow marine, river-dominated delta prograding into shallow water.

8.1 Correlation to the Barents Sea

Seismic studies of north-westward prograding clinoforms in the Barents Sea provide evidence for a significant sediment source in the southeast, probably supplied by the Uralian Orogeny (Fig. 2.3) (Riis et al., 2008; Glørstad-Clark et al., 2010; Høy and Lundschien, 2011; Lundschien et al., 2014; Klausen et al., 2015).

Mineralogical provenance studies (Mørk, 1999) and on detrital zircon (Pózer Bue and Andresen, 2013; Fleming et al., 2015) collectively indicate that major portions of Uralian-derived sands originate from the northern part of the Ural Mountains, with a minor contribution from the Timanides. Analysis of thin sections of samples from the De Geerdalen Formation and chromium spinel provenance studies suggests that sandstones in the De Geerdalen Formation are partly derived from volcanic rocks and indicates a source to the north-east or east of Svalbard (Mørk, 1999; Harstad, 2016). It has also been proposed that a significant landmass existed north of Svalbard, which is suggested to be a sediment source area for Uralian derived sandstones in the Sverdrup Basin (Embry, 1993; Anfinson et al., in press) and may have implications for sediment provenance in Svalbard. Mørk et al. (1982) found no evidence in their stratigraphic sections for a northern source. They suggest an open seaway to the north as indicated by high-energy coastal environments, as interpreted from the outcrops in the western and eastern parts of their study area.

8.2 Facies distribution

8.2.1 Agardhbukta

The position of Agardhbukta on the eastern coast of Spitsbergen make it an apt location as it constitutes a transitional area between central Spitsbergen and the localities further east, on Barentsøya and Edgeøya. It is therefore of interest to map the facies distribution here in order to understand the spatial development of the De Geerdalen Formation. Best correlation is achieved between the sections Klement 15-2 and Klement 15-4, where the lateral distance between log traces is less than 1 km (Fig. 8.1) and individual sandstone beds can be traced along the mountain slope (Fig. 7.2). Distances between the sections on Friedrichfjellet and Šmidtberget is greater compared to Klement'evfjellet. Sections Klement 15-4 and Šmidt 15-1 is separated with 3 km, while the distance between Šmidt 15-1 and Friedrich 15-1 is less with 2.2 km between the sections.

The outcrops in Agardhbukta all share a similar development in terms of vertical stacking of facies. The lower parts of the mountains are dominated by offshore marine facies associations. It is generally heterolithic and consists mostly of interchanging shales and sandstones with hummocky cross stratification and minor bioturbation. These sandstones are interpreted as characteristic of storm-dominated shelf and constitute the prodelta and most distal areas of the delta front (Fig. 6.6B).

The middle parts are characterized by more sand-rich lithologies with facies more typical of a shallow marine delta front environment. Smaller channel-shaped sandstone bodies with large-scale trough cross bedding can be interpreted as tidal channels or as distributary channels. Delta front, barrier bars and shoreface settings (Figs. 6.4, 7.3), as well as distributary mouth bars are also present in the middle part of the mountains. The largest sandstone bodies are generally observed to be laterally continuous for tens to hundreds of meters. On the right side of Klement'evfjellet and on Friedrichfjellet they are often scree covered, and thus appear more discontinuous. The great lateral continuity of sandstones in the middle part of the mountains in Agardhbukta may indicate influence by waves and redistribution of sandstone by longshore currents leading to the development of local shorelines and beaches (Rød et al., 2014). The repeated coarsening upwards pattern observed on Klement'evfjellet is more subtle at Friedrichfjellet and on Šmidtberget, but overall the facies and facies associations become gradually shallower towards the top of the mountains. Each coarsening upwards sequence is marked by mudrocks close to the base interpreted as representing marine flooding surfaces.

These flooding surfaces may on a local scale be attributed to changes in sediment supply related to major avulsions in sediment distributary patterns. On a more regional scale they may be caused by relative changes in accommodation space.

The upper part of the mountains consists of red and green shales, characteristic of the Isfjorden Member. Nodules are common in this interval and typically weather out from the shales forming distinct layers (Fig. 5.23B). These beds are easily observed on both locations on Friedrichfjellet and Klement'evfjellet. However, they appear less prominent on Šmidtberget. Paleosols, coal shales and thin sandstone beds are also commonly observed in the uppermost parts of measured sections and indicates subaerial exposure and deposition in interdistributary area (Fig. 6.6A).

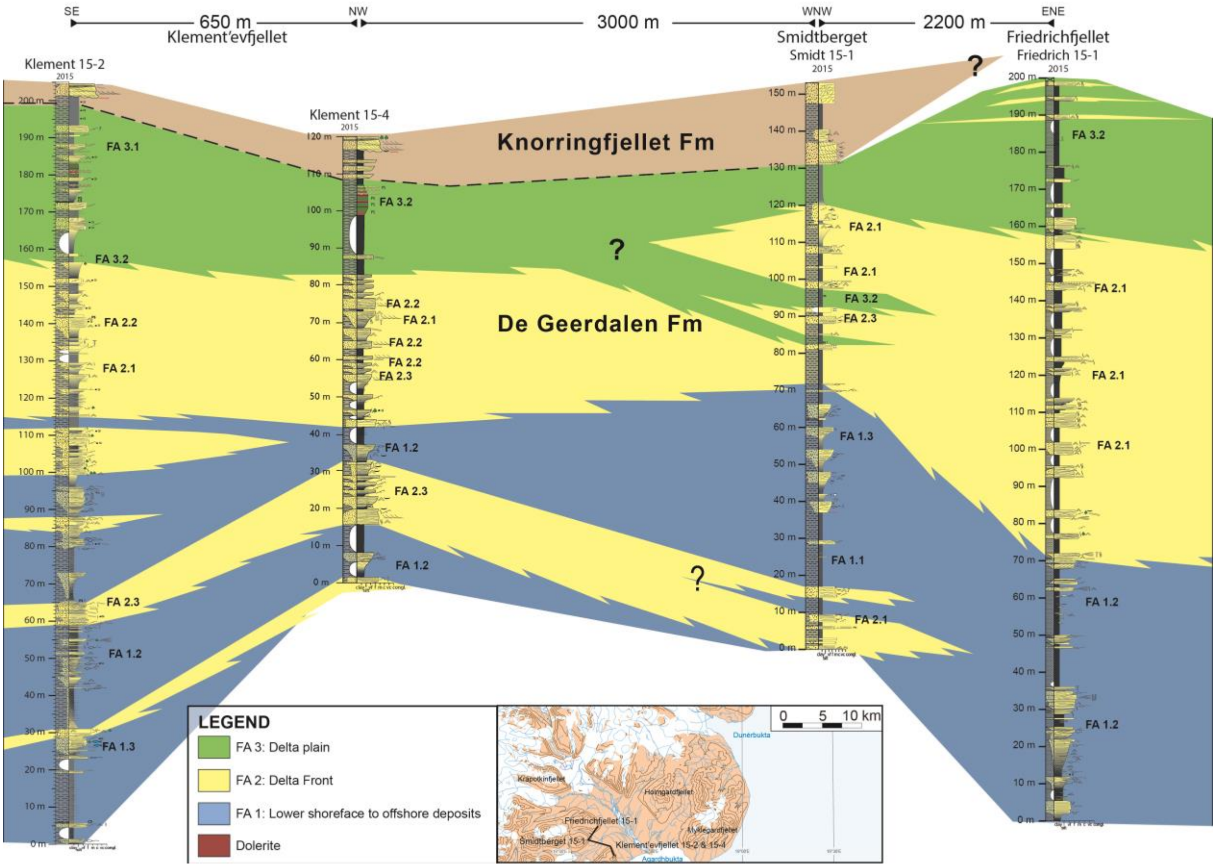

Figure 8.1: Correlation panel for the outcrops of the De Geerdalen Formation in Agardhbukta.

In summary, the succession in Agardhbukta is interpreted as a prograding shallow marine sequence. Offshore marine facies and facies associations are overlain by delta front sandstones in the middle part and delta plain facies associations in the uppermost parts of the mountains. It can be seen as distal compared to the succession on Edgeøya, Barentsøya and Hopen, but more proximal than central Spitsbergen (Rød et al., 2014). This is reflected in a higher amount of sand at higher stratigraphic levels and a more proximal suite of facies and facies associations.

8.2.2 Eastern Spitsbergen and Wilhelmøya

The southernmost sections on eastern Spitsbergen (Teist 15-1 and Hahn 15-1) are similarly to Agardhbukta closely spaced and situated on the same mountain (Fig. 7.1). This allows for good correlations between these sections. Hellwaldfjellet and Wilhelmøya are on the other hand more widely spaced and the distance between Hahnfjella and Hellwaldfjellet is 52 km, while Wilhelmøya is found 37 km north of Hellwaldfjellet. The correlation between Hellwaldfjellet and Wilhelmøya is therefore more uncertain than for the southernmost sections.

On Teistberget, the De Geerdalen Formation are generally interpreted as representing shallow marine barrier bar and shoreface deposits. A small tidal influence on the succession is observed in the lower beach section. The middle part of Teistberget also features channel-shaped sandstones indicating proximity to the delta front or to the delta plain (Fig. 7.9A). These sandstones are often found to be solitary and no observations suggesting lateral accretion or amalgamation is observed. They are therefore interpreted as distributary channels. The remaining part of the measured section on Teistberget constitutes sediments deposited in interdistributary areas and coastal plains and local accumulations of coal shale and paleosols were observed. Sandstones occur more frequently and more laterally continuous in the south compared to the northern exposures on Wilhelmøya and Hellwaldfjellet. This may reflect a more proximal position to an eastern or south-eastern sediment source consistent with the interpretations of Rød et al. (2014). On Hahnfjella, the lower part of the measured section is similar to the deposits on Teistberget and also interpreted as shallow marine (Fig. 7.7B). Sandstones are seen to be laterally continuous across the mountain slope, possibly owing to redistribution of sand by longshore currents. The uppermost part of the succession measured on Hahnfjella is mostly mud-dominated and contain thin (< 2 m thick) sandstone bodies that is intensely bioturbated and includes wave ripples. The deposits above are mostly interpreted as representing coastal plain and interdistributary areas based on the presence of coal shale and coquina beds.

The lower part of the De Geerdalen Formation at Hellwaldfjellet is more sand-rich compared to similar stratigraphic levels on Wilhelmøya and is interpreted as shallow marine. The succession becomes increasingly mud-dominated towards the top and observation of paleosols and coal shales suggests a transition to sub-aerially exposed and coastal conditions. An important observation regarding the lowermost part of the succession on Wilhelmøya is that the interpreted facies and facies association constitute shallow depositional environments. The underlying Tschermakfjellet Formation is interpreted as representing outer to inner shelf, while the lowermost part of the De Geerdalen Formation is interpreted as the offshore transition to distal delta front. The characteristic thick, heterolithic sequence with hummocky cross stratification, wave ripples as seen elsewhere in the study area is apparently missing or at least considerably thinner at this location. The uppermost part of the De Geerdalen Formation on Wilhelmøya is characterized by several channel sandstones with crevasse splays and is interpreted to indicate a proximal fluvial environment. Presence of wood fragments, paleosols and plant fragments are interpreted to suggest subaerial exposure and favourable conditions for extensive and diverse vegetation cover in a humid climate. The floodplain deposits (FA 3.1) are often overlying interdistributary muds (FA 3.2), reflecting the regressive development of the formation. The thickness of coastal and delta plain deposits are seen to increase towards the north and is thicker on Wilhelmøya and Hellwaldfjellet, compared to the exposures further south on eastern Spitsbergen. These locations are located the most distal to the suggested paleo-coastline in the southeast (e.g. Glørstad-Clark et al., 2010; Lundschien et al., 2014; Klausen et al., 2015) and hence one would expect to observe a transition from distal prodelta and delta front environments instead of a delta front to delta top environment. An explanation for the extensive and thick accumulations of delta plain and floodplain deposits in the northern part of the study area may be that the primary fluvial trunk river system and sediment source is located to the east or northeast. The sandstone deposits at Blanknuten and Svartnosa may therefore be smaller distributaries of this major system. An alternative explanation is that a secondary sediment source area is in fact located more to the east or northeast as some studies suggests (Mørk, 1999; Fleming et al., 2015; Harstad, 2016).

The north-west prograding clinform sequences mapped in the Barents Sea most likely covered Svalbard by Late Carnian time (Lundschien et al., 2014). The proximal development seen on Wilhelmøya and Hellwaldfjellet could reflect the progradation of a new clinoform sequence, resulting in the missing offshore being eroded and transported further west or north-west.

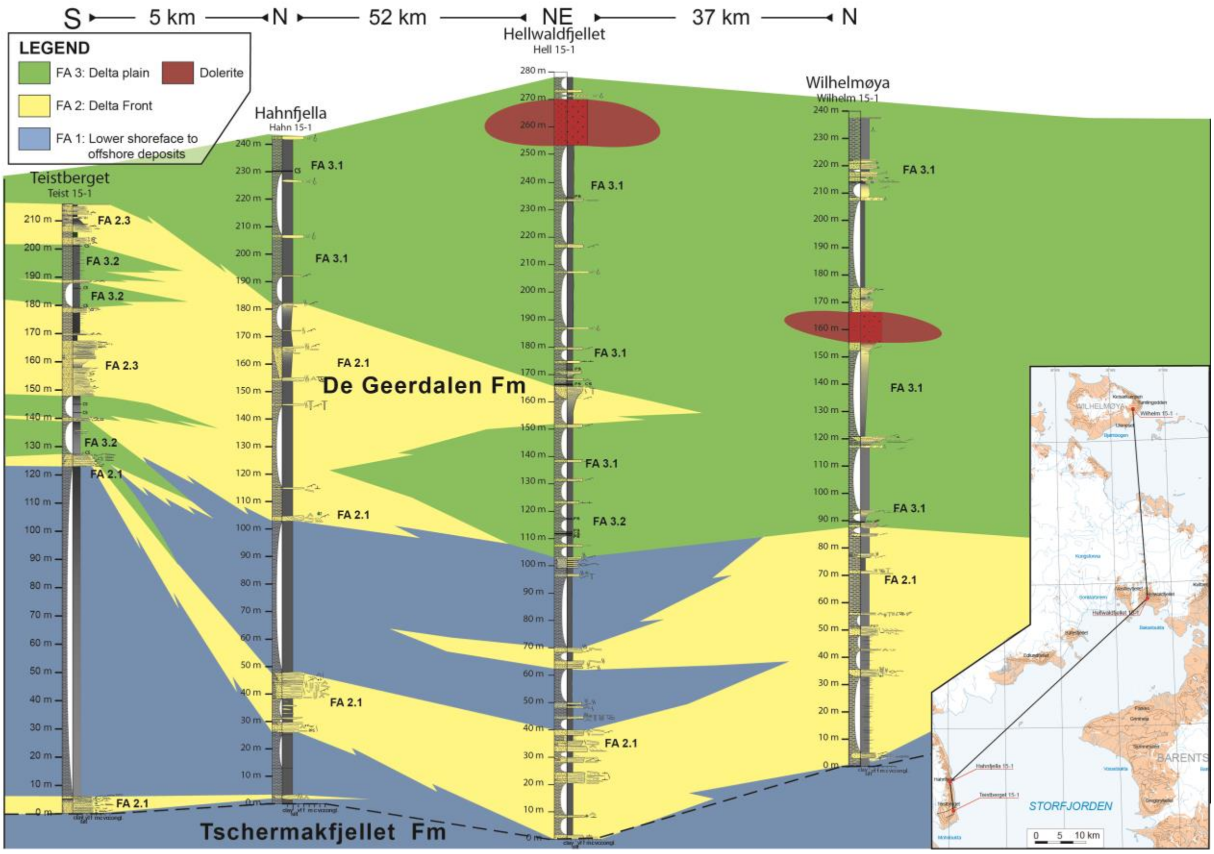

Figure 8.2: Correlation panel for the outcrops of the De Geerdalen Formation on eastern Spitsbergen and Wilhelmøya.

8.2.3 Barentsøya

Correlation between outcrops on Barentsøya is less certain compared to the outcrops in Agardhbukta, as they are more widely spaced with several kilometres between sections. The top of the Botneheia Formation provides a good time-stratigraphic surface, as it can be considered isochronous and occurs on a regional scale on Spitsbergen, Barentsøya and Edgeøya (Mørk et al., 1982). Dolerite sills and dikes are also here shown to cause considerable complexity on the sedimentary succession on parts of Barentsøya, e.g. Mistakodden and Wilhelmøya (Figs. 2.1, 7.1, 7.11B, 7.17).

Amalgamated distributary channels and mouth bars, as seen on Blanknuten (Knarud, 1980; Rød et al., 2014) and Svartnosa demonstrate the strong fluvial dominance on Barentsøya. The dimensions of individual distributary channels on Barentsøya and Edgeøya (Rød et al., 2014) are smaller, compared to the larger trunk channels that have been mapped and studied on Hopen (Klausen and Mørk, 2014; Lord et al., 2014) and in the Barents Sea (Klausen et al., 2014, 2015). This could be explained by the fact that Svalbard was situated on the distal end of the embayment which can be considered to have been largely filled during the Triassic. It was also strongly affected by paralic processes leading to reworking and redistribution of sand.

The southernmost outcrop on Krefftberget is poorly exposed, but is seen to consists of facies common in shallow marine to paralic environments such as low-angle cross stratification (facies E), horizontally bedded sandstone (facies F), offshore to interdistributary mudrocks (facies M) as well as minor bioturbation and some wave ripples (facies D). Krefftberget is located directly south of Svartnosa where the succession is interpreted as fluvial-dominated delta front sandstones with amalgamated distributary channels and mouth bars. This interpretation has also been reported from Blanknuten on Edgeøya (Mørk et al., 1982; Rød et al., 2014), where similar deposits as seen on Svartnosa have been described. Given this position between Barentsøya and Edgeøya, the entire mud-dominated part of the succession on Krefftberget could represent deposition in an interdistributary bay, while the sandstones represent inter-fingering shallow marine shoreface deposits.

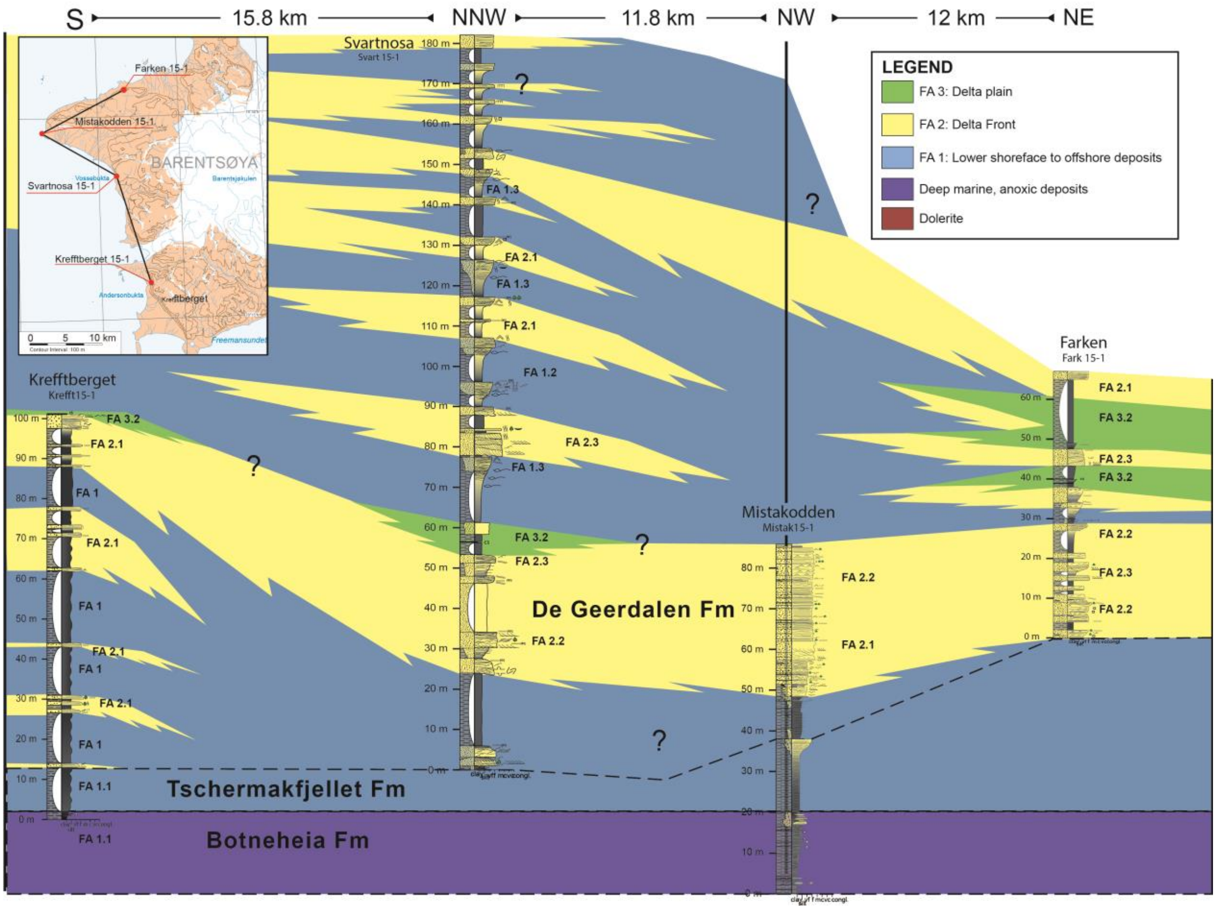

Figure 8.3: Correlation panel for the outcrops of the De Geerdalen Formation on Barentsøya.

The depositional environment for the sandy part of the succession on Mistakodden has been interpreted as a fluvial-dominated delta front. It is situated north of the growth faults on Kvalpynten (Edwards, 1976a) and Klinkhammaren (Osmundsen et al., 2014). These growth faulted sandstones are seen to thicken towards the north and towards normal fault planes (*cf.* figs. 7, 9 10, 11 in Osmundsen et al., 2014) and thinning towards the south. On Mistakodden, a similar development in thickness is observed on the northern part of the beach section (Fig. 7.16). The position of Mistakodden along a north to south trend line of growth faults may indicate that these sandstones are a part of the same system. However, outcrop data from this area is limited and further work is necessary in order to determine if these sandstones are in fact growth faults.

In contrast to exposures on Spitsbergen and on Wilhelmøya, delta plain deposits are almost absent in uppermost part of the De Geerdalen Formation on Barentsøya and Edgeøya due Quaternary uplift and erosion. Thus the succession consists mostly of lower shoreface to offshore deposits (FA 1.1, 1.2 and 1.3) overlain by delta front sandstones (FA 2.1, 2.2 and 2.3). Coal shales are present in the uppermost part of the succession on Krefftberget, indicating at least a proximity to coastal plain environments.

8.3 Delta classification and main delta processes

Deltas can be classified according to the relative control of fluvial input, wave modulation and tidal energy, following the ternary classification scheme as introduced by Galloway (1975). This classification scheme has been further extended to incorporate dominant grain size (Orton and Reading, 1993) and feeder systems (Bhattacharya and Walker, 1992; Reading and Collinson, 1996).

Some of the major challenges with this tripartite classification of deltas and also the applicability of modern analogues to ancient deltas include;

- Process dominance is often averaged over the entire delta, but dominating processes may change rapidly over short distances (Olariu, 2014).
- Most modern and ancient deltas will however normally show a mixed influence of all the processes and there has generally been a tendency to erroneously force-fit the classification into one of the three end-members (Dominguez, 1996; Bhattacharya and Giosan, 2003; Bhattacharya, 2006).
- Modern deltas are formed in the Holocene, which is characterized by a high eustatic sea level following melting of continental glaciers (Clifton, 2006). The consequence of

this is that many deltas today are mostly affected by retrogradational processes in contrast to the progradational processes interpreted in most ancient examples.

Ancient fluvial-dominated deltas are generally recognized based on the following criteria (Reading and Collinson, 1996): (i) a thick vertical succession passing from mud-dominated offshore facies, through sandy delta front facies and are overlain by mud-rich, continental delta plain facies; (ii) thicker units of sand is of limited lateral extent, since sand is mostly deposited around mouth bars; (iii) repeated upwards coarsening and shallowing successions due to repeated progradation and subsequent abandonment of individual lobes following major river avulsions. All of these requirements are fulfilled within the delta complex on eastern Svalbard, especially on Blanknuten (Mørk et al., 1982; Rød et al., 2014) and Svartnosa. Common depositional environments in fluvial-dominated deltas includes interdistributary bays, lakes or swamps, tidal flat or chenier plains, fluvial and tidal channel, levees and crevasse splays (Orton and Reading, 1993; Reading and Collinson, 1996). Fluvial-dominated deltas are also characterized primarily by deposition of bar-finger sands and are usually mud-dominated, with a low gradient delta plain (Orton and Reading, 1993; Reading and Collinson, 1996).

Wave-dominated deltas on the other hand, are characterized by vertical facies sequences grading from offshore, inner shelf mudrocks, through silty to very fine sand, to storm- and wave-dominated areas. These are further overlain by strandplain sandstones and mudrocks, and capped by delta plain to floodplain fines, paleosols and coal shales. Most of the sand is derived from within the basin and from reworked inner shelf sands supplied by longshore drift of sediments (Reading and Collinson, 1996). Wave-dominated deltas share many similarities with prograding shorelines and in ancient successions differentiation between the two types can be challenging (Hampson and Howell, 2005; Clifton, 2006). Beach, shoreface and strandplain deposits that are normally associated with prograding wave-dominated shorelines can also occur on the updrift side of deltas with asymmetric lobes (Li et al., 2011). Therefore there are considerable differences between the updrift and downdrift side of the delta mouth (Bhattacharya and Giosan, 2003). Number of distributary channels and frequency of bifurcation tend to be less in wave-dominated deltas compared to fluvial-dominated deltas, because sediments are removed by waves which decelerates progradation and channel bifurcation (Bhattacharya and Giosan, 2003; Bhattacharya and Tye, 2004; Olariu and Bhattacharya, 2006). A modern example of a delta that shows a mixed influence of dominating processes is the Danube delta on the north-western margin of the Black Sea. The southern deltaic lobe is wave-dominated with a single asymmetric distributary stream, several beach ridges and cuspate shape. The northern lobe features a morphology more characteristic of a fluvial-dominated delta with multiple distributaries and an elongate shape of the delta (Bhattacharya and Giosan, 2003; Li et al., 2011). Decrease in sediment supply and discharge following a major river avulsions resulted in a change in the delta from fluvial-dominated to wave-dominated (Olariu, 2014).

The deltaic facies sequences in the De Geerdalen Formation are dominated by wave modulation and characterized by a high fluvial supply (Glørstad-Clark, 2010; Klausen et al., 2015; Klausen et al., 2016). Fluvial processes are most evident on the eastern islands (Barentsøya and Edgeøya) where large, lenticular bodies of sandstone are interpreted as representing deltaic lobes. The uppermost part of the formation often features paleosols (Enga, 2015) and may indicate prolonged periods of subaerial exposure and establishment of widespread coastal to delta plain conditions in a restricted environment.

Heterolithic bedding, wave ripples and hummocky cross stratification are common sedimentary structures in the lower and middle part of the De Geerdalen Formation over the entire study area. They are also common at similar stratigraphic levels elsewhere on Svalbard (Rød et al., 2014). The abundance of storm and wave-generated structures is therefore considered to demonstrate a significant influence of these processes on the shallow marine succession.

Tidal processes are nevertheless recognized as an important factor both on Svalbard (Hynne, 2010; Høy and Lundschien, 2011; Klausen and Mørk, 2014; Rød et al., 2014) and in the Barents Sea (Riis et al., 2008; Klausen et al., 2015). Herringbone cross-stratification and double mud drapes are reported more frequently on central Spitsbergen compared to eastern equivalents and are interpreted to indicate a weaker fluvial component in the western exposures (Rød et al., 2014). Similarly to the Snadd Formation in the Barents Sea (Klausen et al., 2015), tidal influence is most likely preserved away from progradational distributary channels and occur in more protected settings, such as sheltered bays and interdistributary areas. In Agardhbukta, the succession on Klement'evfjell and Friedrichfjellet contain several tidal signatures such as mud drapes, wavy and flaser bedding, small-scale ripple cross-stratification and wave ripples in thinner sandstone beds. No observations of herringbone cross stratification has been made in any outcrops. Thicker sandstone units are characterized by larger scale cross stratification. These can be interpreted as tidally influenced distributary channels or channel bars (FA 2.1). On Teistberget, large-scale cross stratified sandstones with mud drapes have been interpreted as migrating and laterally accreting tidal channels.

8.4 Parasequences

The lower to middle part of De Geerdalen Formation in eastern areas of Svalbard shows a typical development of parasequences (*sensu* Van Wagoner et al., 1990), with sandstones and shales in coarsening and shallowing upwards sequences (Mørk et al., 1999a; Rød et al., 2014; Vigran et al., 2014). The top of the sandstones are usually capped by marine shales (Fig. 7.3), marking the onset of the next parasequence, resulting in a step-like topography (Knarud, 1980). Parasequences similar to the ones observed onshore Svalbard are also known from the offshore Snadd Formation in the Barents Sea (Klausen et al., 2015l). The stacked upwards-shallowing units are interpreted to form due to auto-cyclic switching of delta lobes within one major system (Riis et al., 2008). This interpretation was also suggested by (Knarud, 1980) for his facies association C1 (distal delta front). The Barents Sea embayment was likely infilled at a great pace, so that the trajectory and frequency of delta lobe switching was probably mostly governed by variable accommodation space across the basing. Anell et al. (2014b) attributes stacking patterns of clinoform sequences in the Barents Sea to variable and relative changes in accommodation and sediment supply.

The genetic stratigraphic unit as defined by Galloway (1989), proves practical for correlation of facies patterns on a larger scale in tectonically stable settings (Reading and Levell, 1996). This model defines depositional sequences as bounded by the maximum flooding surface (MFS). It was developed from the paleo-geographically stable basin of the Gulf of Mexico and has proved suitable for regions characterized by continuous, huge and even subsidence and where clastic sediment supply is the dominant control on facies patterns (Reading and Levell, 1996). In the correlation panels for the outcrops on eastern Svalbard, a flooding surface of possibly regional extent may be recognized in the lower part of the De Geerdalen Formation (Figs. 8.1, 8.2, 8.3). The dataset in this thesis is limited, concentrating on facies distributions and depositional environments at a scale that is most prone to changes in autogenic processes (Catuneanu et al., 2011; Olariu, 2014). Also, most outcrops described here are so widely spaced that detailed sequence stratigraphic correlations is difficult.

Delta lobe shifting is responsible for generating medium-scale facies sequences and coarsening upwards successions with sandy mouth bars and delta-front sediments building over muddy deeper-water prodelta facies (Reading and Collinson, 1996; Bhattacharya, 2006;

Catuneanu et al., 2009). This is common in fluvial-dominated deltas and frequent lobe shifts have been reported from the last 7000 years in the modern Mississippi River delta, occurring at 1000-year intervals (Frazier, 1967; Penland et al., 1988; Reading and Collinson, 1996). Considerable variations in morphology, facies distribution and facies associations are also common between individual, active deltaic lobes (Li et al., 2011). Lock et al. (1978) noted a significant variation in the thickness of the Tschermakfjellet Formation in eastern Svalbard. The reason for this variation is simply attributed to autogenic switching of deltaic lobes in the overlying De Geerdalen Formation (Mørk et al., 1982). Following a major avulsion of the delta, interdistributary mud- and sandstones are hence deposited directly on top of prodelta mudstones, away from the distributary channel (Reading and Collinson, 1996), e.g. as interpreted on Hellwaldfjellet.

8.5 Modern analogues

In general the De Geerdalen Formation on Barentsøya and Edgeøya show a resemblance to the modern Mississippi-Atchafalaya delta complex (Fisk et al., 1954) in terms of facies development and dominating processes. Both systems are mud-dominated and to a large degree affected by fluvial processes (table 1 and table 2 in Orton and Reading, 1993). Progradation of the Mississippi delta is governed by buoyancy-dominated flow, where stacking and amalgamation of terminal distributary mouth bars and channels results in the deposition of elongate bar-finger sands (Fig. 8.4) that are ellipsoidal in cross-section (Frazier, 1967; Olariu and Bhattacharya; 2006; Boggs, 2011). On Blanknuten (Rød et al., 2014) and Svartnosa, large ellipsoidal sandstones can be interpreted as such bar-finger sands within a large-scale deltaic lobe. Outcrop data from Svartnosa is limited as time did not permit outcrops of sandstones to be walked out laterally. It is therefore difficult to distinguish and separate between distributary channels and mouth bars as these are genetically related and distributary channels are often filled in by mouth bar deposits (Olariu and Bhattacharya, 2006).

However, it is important to emphasize that although there are similarities between the Mississippi–AAtchafalaya delta and the De Geerdalen Formation, some key differences are observed. The Mississippi is a shelf-edge delta building out into deep waters, while the estimated paleo-water depths are much lower (< 500 m) for the Barents Shelf during the Triassic (Glørstad-Clark, 2010; Klausen et al., 2015). The Mississippi River also drains almost an entire continent with a drainage area covering approximately 3,344,000 km^2 (Reading and Collinson, 1996). In comparison, the drainage area for the Kobbe and Snadd formations in the Barents Sea is estimated to be in the order of 200,000 km^2 (Klausen and Mørk, 2014).

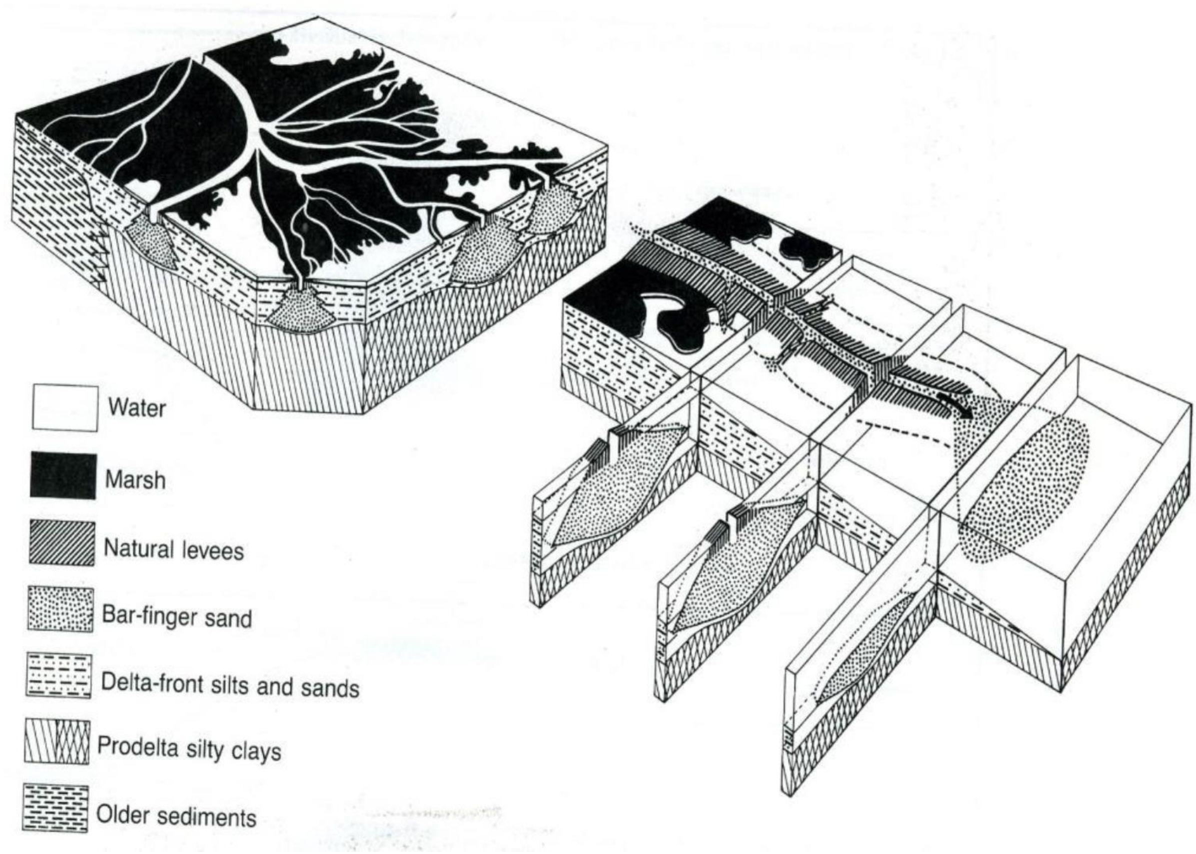

Figure 8.4: Elongate bar-finger sandstones resulting from the progradation of smaller distributary channels on the modern Mississippi delta (from Reineck, 1970, after Fisk et al., 1954).

In fact, a similar mismatch between drainage areas has also been reported for the Ferron Delta (estimated drainage area: 50,000 km^2) of the Cretaceous Western Interior Seaway, which similarly the De Geerdalen Formation, drains a mountain range and not the interior of a continent as is the case for the Mississippi, Niger, Amazon or the Nile (Bhattacharya and Tye, 2004; Olariu and Bhattacharya, 2006). Thus, these deltas are less suitable as good modern analogues for the De Geerdalen Formation.

Other fluvial-dominated deltas are the Po River Delta of north-eastern Italy (Correggiari et al., 2005) and the Ebro delta in eastern Spain. The drainage area for the Po River Delta (71,700 km^2) is closer to the estimated drainage area for the De Geerdalen Formation, and is supplied by sediments from the mountain ranges of the alpine system in central Europe. The Adriatic Sea, which is the receiving basin for sediments in the Po River, is similarly to the Barents Sea a shallow epicontinental sea. They may therefore be more applicable as analogues for the De

Geerdalen Formation. Both the modern Mississippi and Po delta are largely affected by human activity (Cencini, 1998; Bhattacharya, 2006), reducing their potential value as good modern analogues. Fluvial sedimentation and river direction has been controlled through the efforts of humans in order to inhabit and cultivate coastal areas.

The combinations of depositional controls that have been reported for the De Geerdalen are so specific that a fully analogous modern example simply does not exist. The De Geerdalen Formation was deposited as a part of a large delta system prograding into a shallow shelf and with multiple source areas. Modern fluvial-dominated deltas either occur at different scales, e.g. the Mississippi River Delta, or they are governed by different depositional controls resulting in different shoreline morphologies, such as the Danube, Po and Ebro deltas.

9. Conclusions

- The Late Triassic De Geerdalen Formation was deposited in a shallow marine environment with paralic conditions on the distal end of a large delta complex, possibly sourced from the east and south-east during the Uralian Orogeny.
- The De Geerdalen Formation primarily consists of coarsening upwards sequences forming upwards shallowing parasequences recording repeating shoreline progradations.
- The lowermost part of the formation is mostly mud-dominated and consists of facies characteristic of storm and wave-dominated shelves. Sand content increases upwards in the middle part of the formation and contain facies common in delta front and shoreface environments. The uppermost part is dominated by facies characteristic of delta top and delta plain environments.
- A proximal fluvial-dominated environment is interpreted from outcrops on Barentsøya, indicating a proximal position to an eastern or south-eastern source.
- Waves and tides become increasingly dominant towards the west, particularly in the outcrops in Agardhbukta and represent more distal and transitional environments compared to the succession in the eastern exposures.
- The succession on eastern Spitsbergen and Wilhelmøya is mostly dominated by tidal and wave processes and reflect more distal depositional environments compared to the southern and eastern part of the study area.

9.1 Recommendations for further research

The sand-rich outcrop at Svartnosa features vertical cross-cut of a delta and is worthy of more attention as the relationship to these growth faults could be further investigated. Sandstones on Mistakodden could also be related to growth faulting and further studies would resolve this question.

A great amount of samples were taken of both coarse-grained sandstones and fine-grained silts and clays. The sandstones can be used for zircon- and thin-section analysis to further investigate the diagenesis and provenance of the De Geerdalen Formation. The fine-grained material can be used for palynological studies and will improve dating of the succession.

Further fieldwork by the research group is planned in the area around Fulmardalen to the northwest of Agardhbukta. The results from this field season will add valuable information regarding the development of the De Geerdalen Formation in this area.

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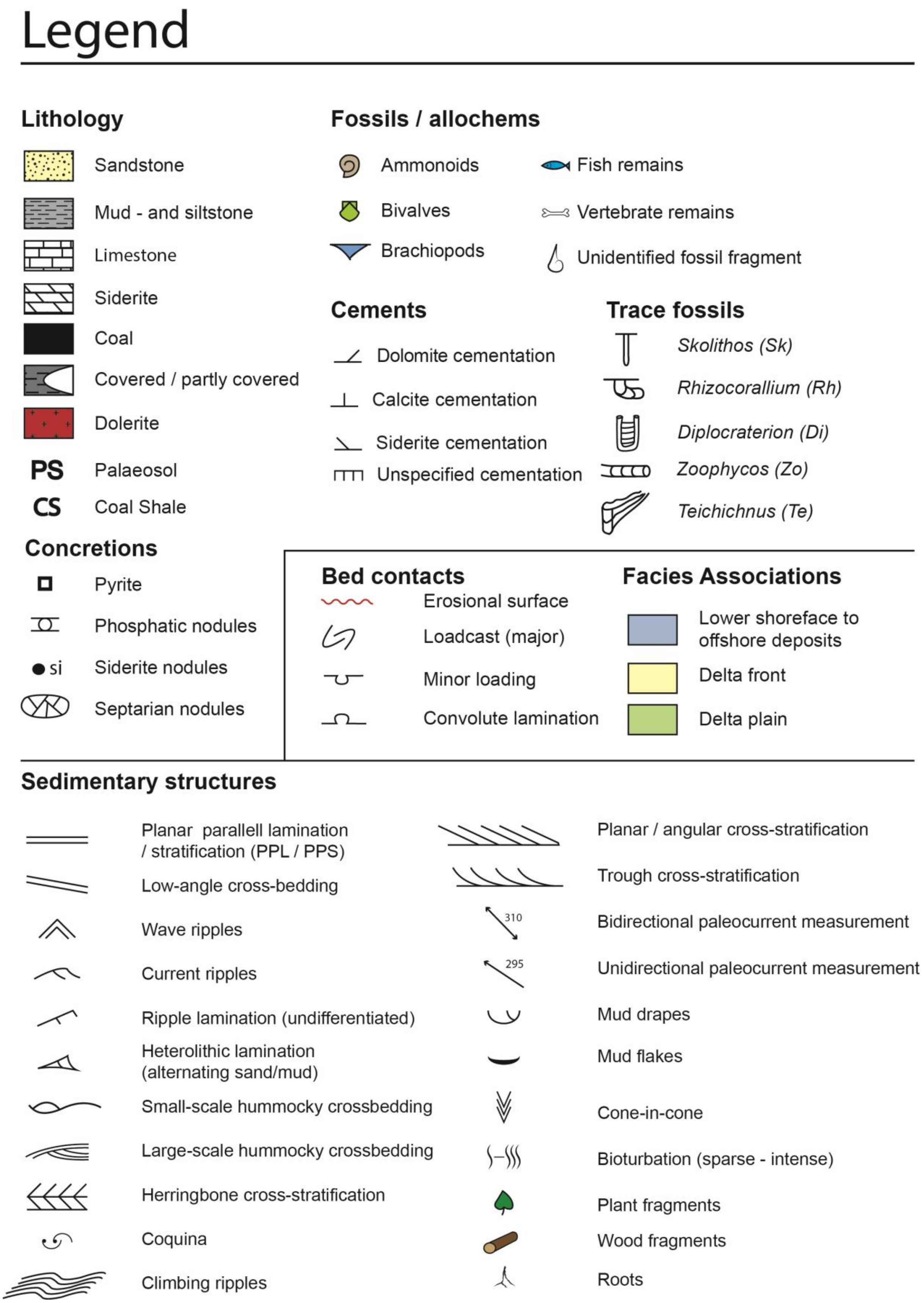
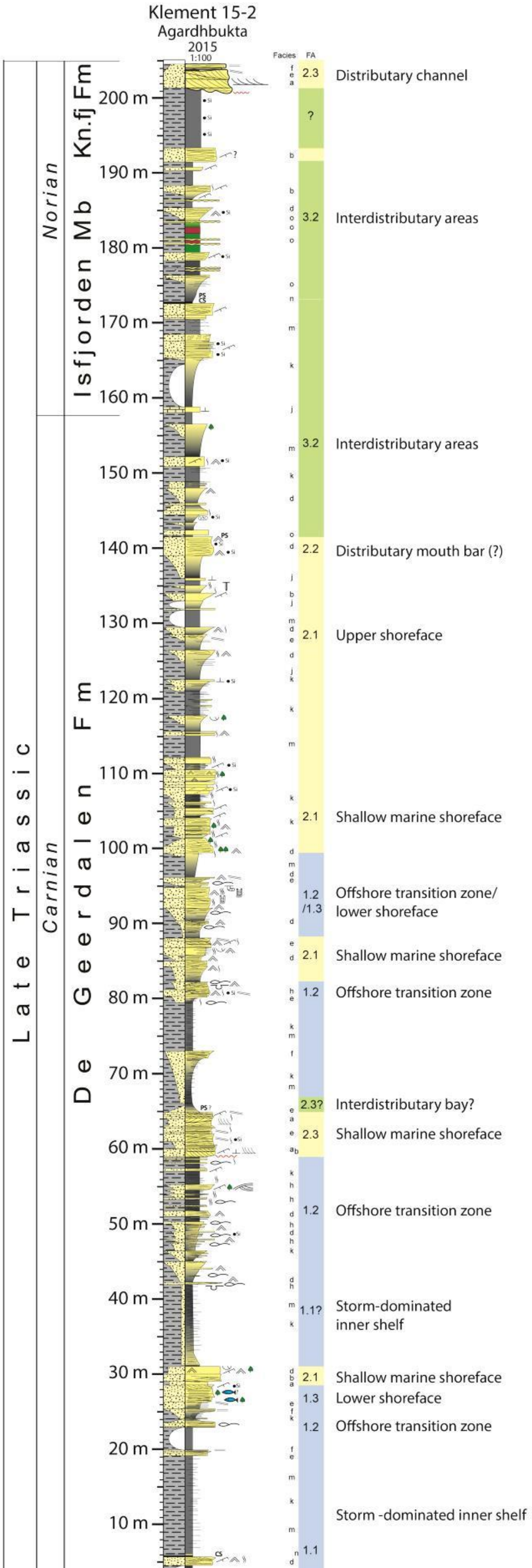
Appendix A – Overview of measured sections with UTM-coordinates

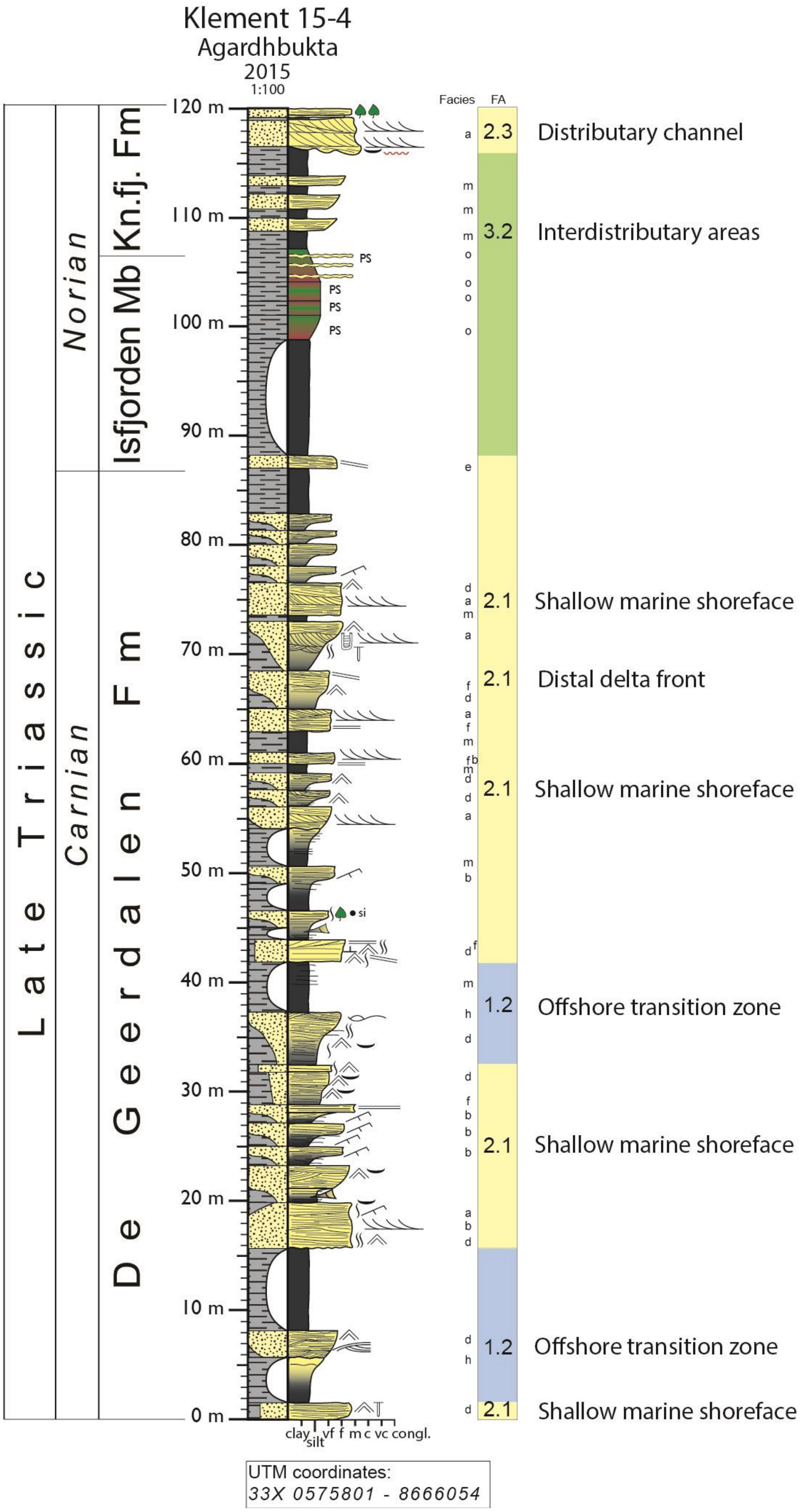
Location		Log no.	Elevation above sea-level [m a.s.l]		UTM zone base log	UTM coordinates base of log		UTM zone top log	UTM coordinates top of log	
			Base log	Top log						
Edgeøya	**Muen**	15 - 1	-	-	-	-		-		-
Eastern Spitsbergen	**Klement'evfjellet**	15 - 2	106 m	356 m	33X	0576226	8665537	33X	0575947	8665277
		15 - 4	215 m	329 m	33X	0575801	8666054	33X	0578899	8665577
	Friedrichfjellet	15 - 1	135 m	328 m	33X	0574156	8668564	33X	0573911	8668699
	Smidtberget	15 - 1	105 m	290 m	33X	0572917	8666751	33X	0572669	8666924
	Hahnfjella	15 - 1	147 m	477 m	33X	0590209	8706474	33X	0589294	8707030
	Teistberget	15 - 1	0 m		---	---	---	33X	0590803	8702355
	Hellwaldfjellet	15 - 1	156 m	460 m	33X	0625577	8745316	33X	0625714	8746169
Wilhelmøya	**Tumlingodden**	15 - 1	142 m	485 m	33X	0622072	8782337	33X	0621192	8782969
Barentsøya	**Mistakodden**	15 - 1	0 m	90 m	33X	0615227	8717614	33X	0615308	8717992
	Farken	15 - 1	109 m	195 m	33X	0625146	8723913	33X	0625189	8723839
	Svartnosa	15 - 1	34 m	253 m	33X	0625356	8711779	33X	0625663	8711925
	Krefftberget	15 - 1	203 m	359 m	33X	0631239	8697538	33X	0631535	8697777

Appendix B – Legend for Measured Sections

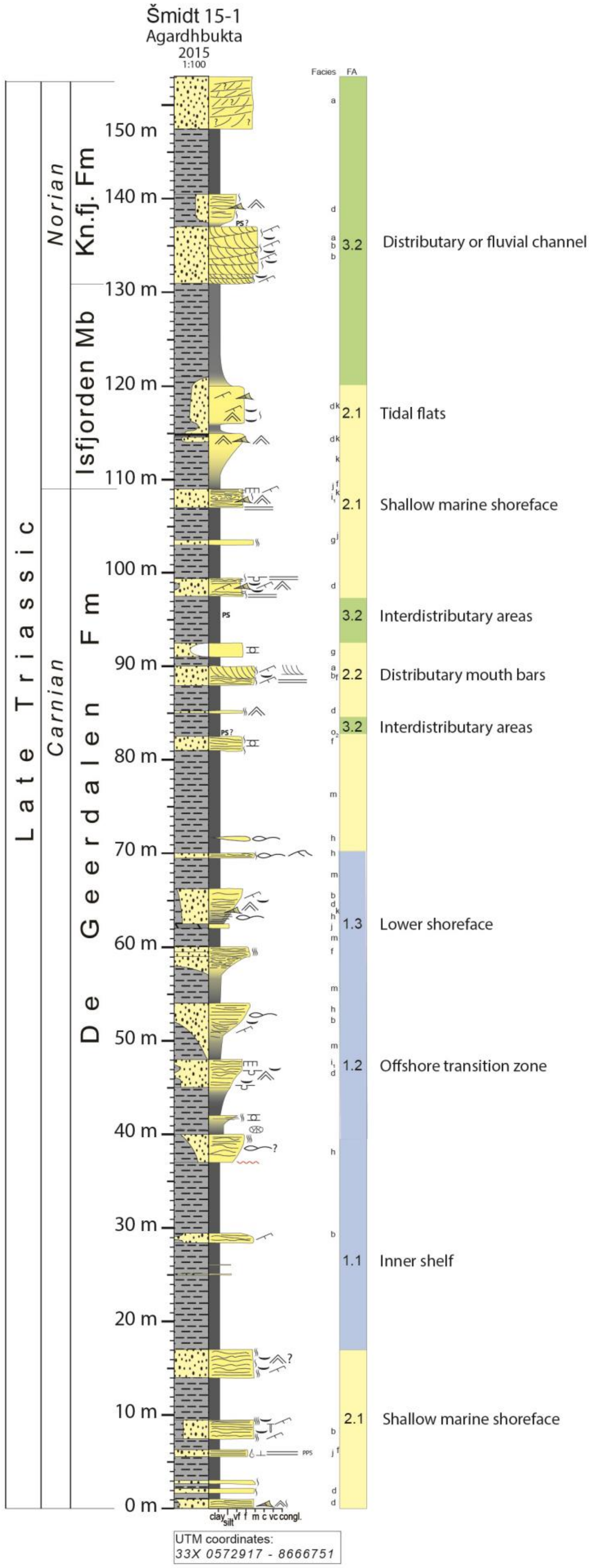


Appendix C – Measured stratigraphical sections





Klement 15-4
Agardhbukta
2015
1:100
Late Triassic
Norian
Carnian
Isfjorden Mb
Kn.fj. Fm
De Geerdalen Fm
120 m
110 m
100 m
90 m
80 m
70 m
60 m
50 m
40 m
30 m
20 m
10 m
0 m
clay silt vf f m c vc congl.
Facies
FA
PS
si
2.3 Distributary channel
3.2 Interdistributary areas
2.1 Shallow marine shoreface
2.1 Distal delta front
2.1 Shallow marine shoreface
1.2 Offshore transition zone
2.1 Shallow marine shoreface
1.2 Offshore transition zone
2.1 Shallow marine shoreface
UTM coordinates:
33X 0575801 - 8666054


Šmidt 15-1
Agardhbukta
2015
1:100
Facies
FA
Late Triassic
Carnian
Norian
De Geerdalen Fm
Isfjorden Mb
Kn.fj. Fm
150 m
140 m
130 m
120 m
110 m
100 m
90 m
80 m
70 m
60 m
50 m
40 m
30 m
20 m
10 m
0 m
clay silt vf f m c vc congl.
3.2 Distributary or fluvial channel
2.1 Tidal flats
2.1 Shallow marine shoreface
3.2 Interdistributary areas
2.2 Distributary mouth bars
3.2 Interdistributary areas
1.3 Lower shoreface
1.2 Offshore transition zone
1.1 Inner shelf
2.1 Shallow marine shoreface
UTM coordinates:
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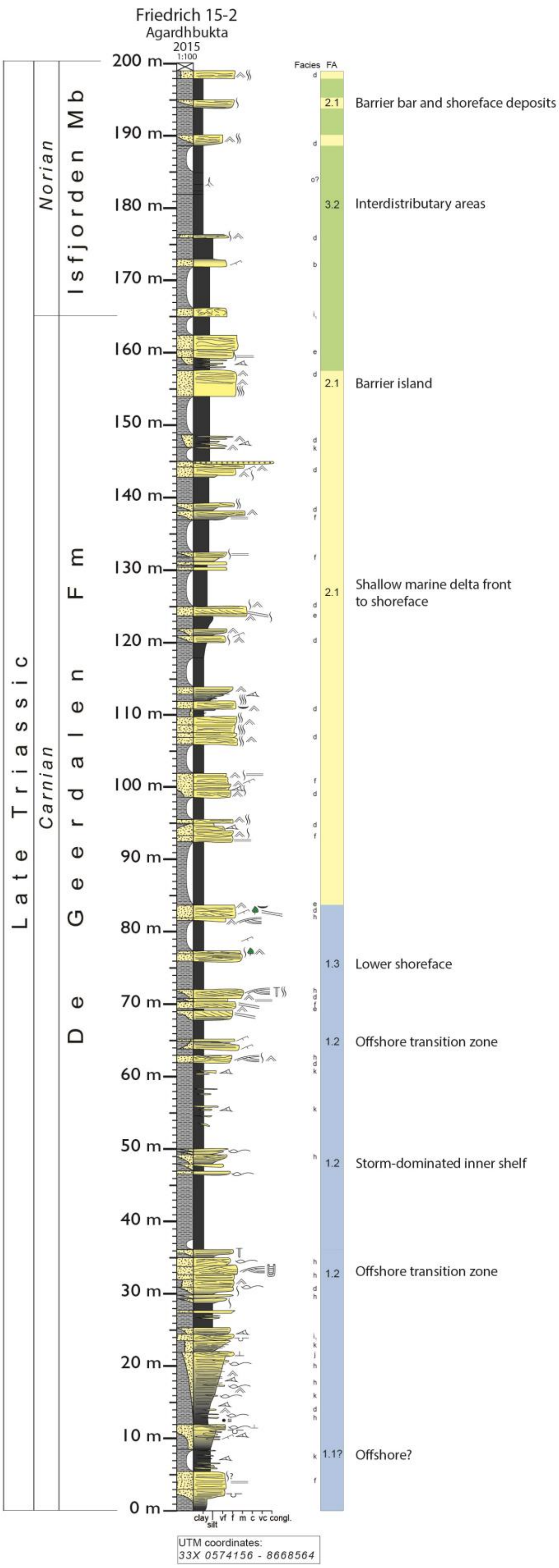
Friedrich 15-2
Agardhbukta
2015
1:100
Late Triassic
Norian
Carnian
Isfjorden Mb
De Geerdalen Fm
200 m
190 m
180 m
170 m
160 m
150 m
140 m
130 m
120 m
110 m
100 m
90 m
80 m
70 m
60 m
50 m
40 m
30 m
20 m
10 m
0 m
clay silt vf f m c vc congl.
Facies FA
2.1 Barrier bar and shoreface deposits
3.2 Interdistributary areas
2.1 Barrier island
2.1 Shallow marine delta front to shoreface
1.3 Lower shoreface
1.2 Offshore transition zone
1.2 Storm-dominated inner shelf
1.2 Offshore transition zone
1.1? Offshore?
UTM coordinates:
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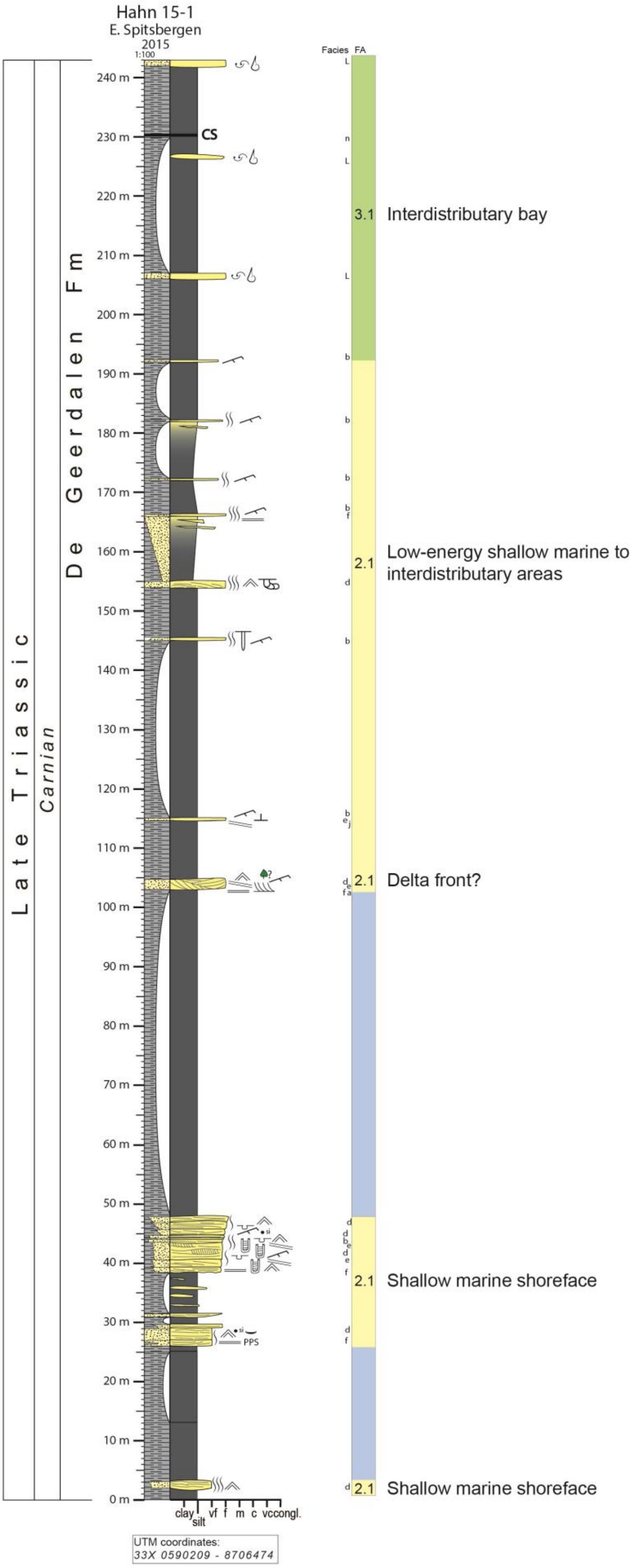
Hahn 15-1
E. Spitsbergen
2015
1:100
Late Triassic
Carnian
De Geerdalen Fm
CS
Facies FA
3.1 Interdistributary bay
2.1 Low-energy shallow marine to interdistributary areas
2.1 Delta front?
2.1 Shallow marine shoreface
2.1 Shallow marine shoreface
clay silt vf f m c vc congl.
UTM coordinates:
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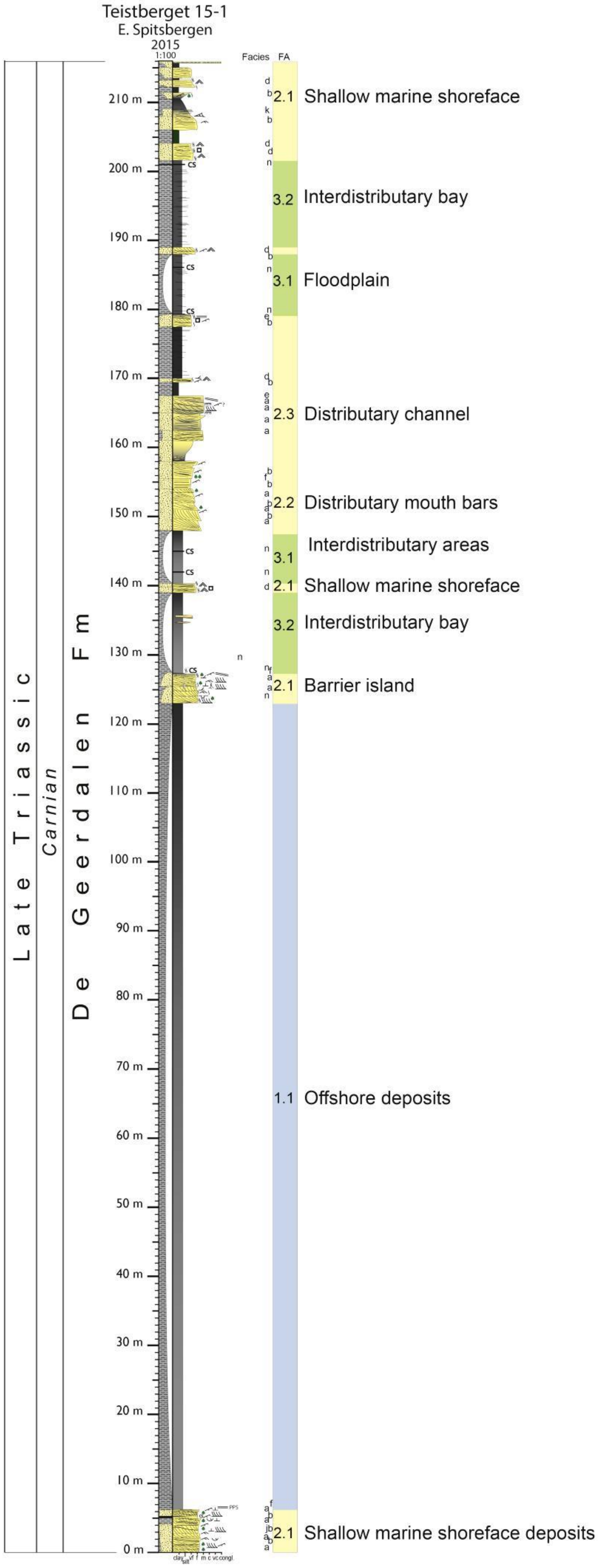
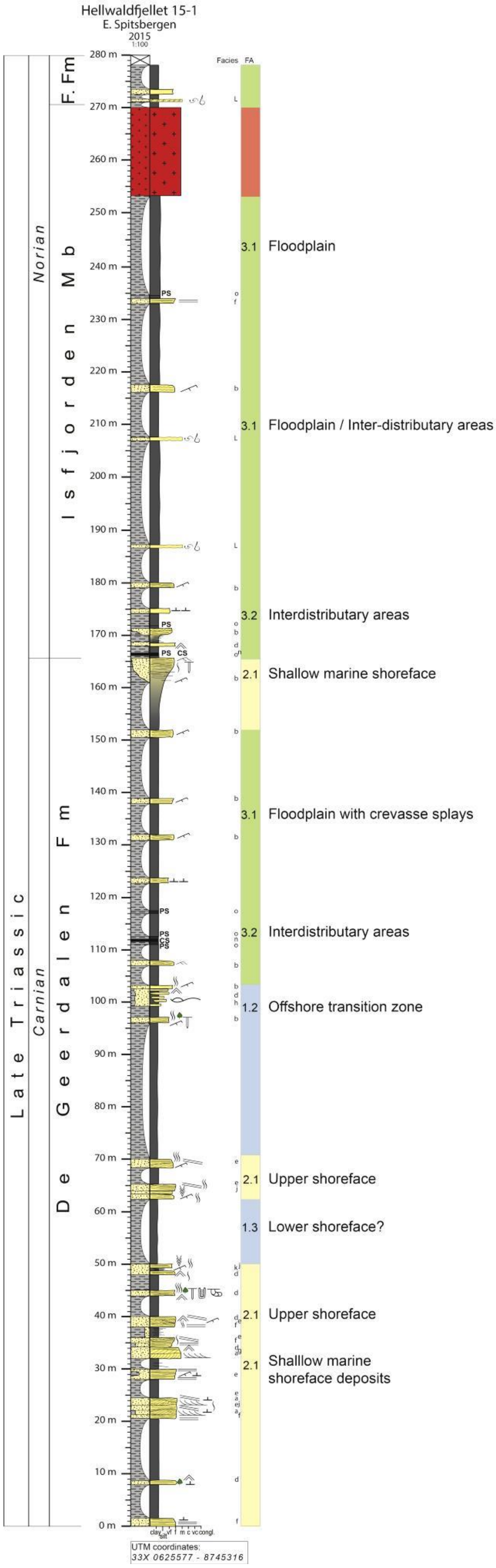
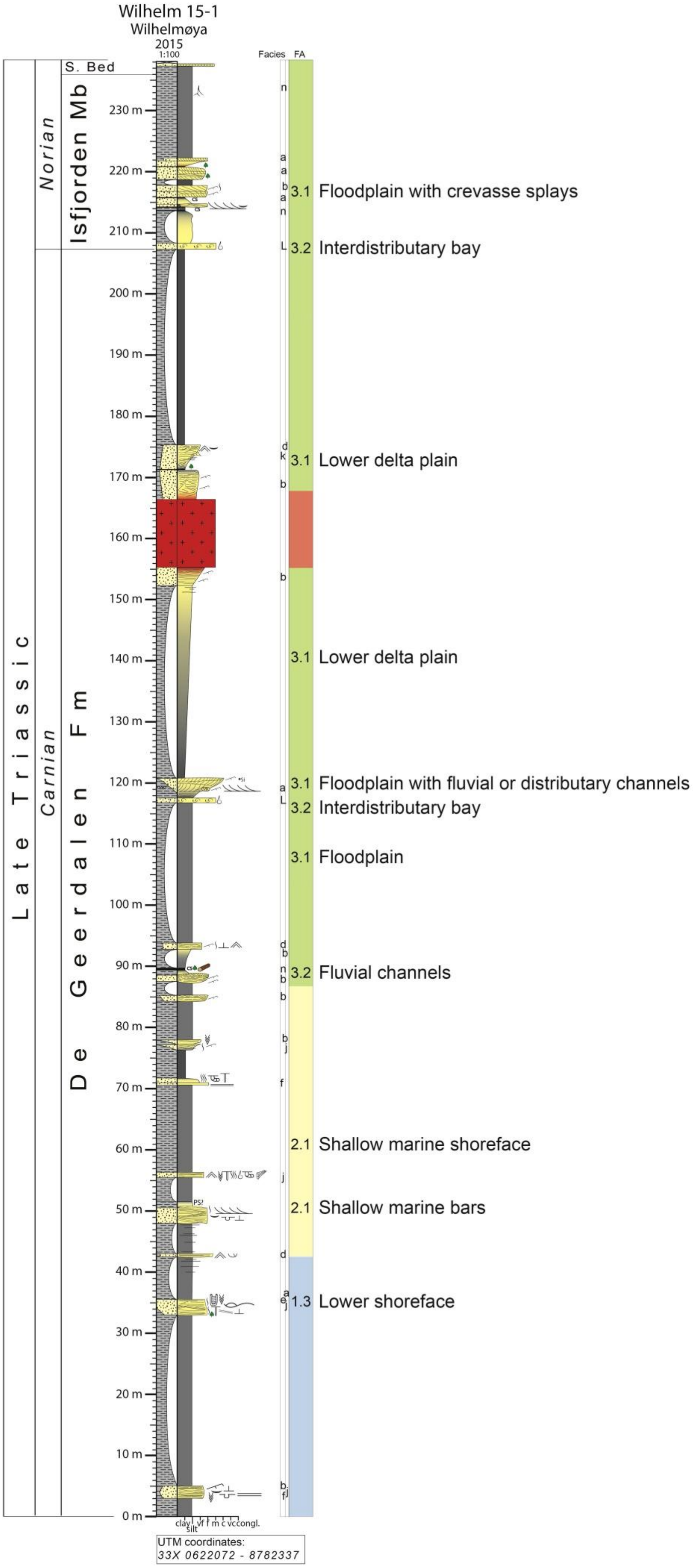
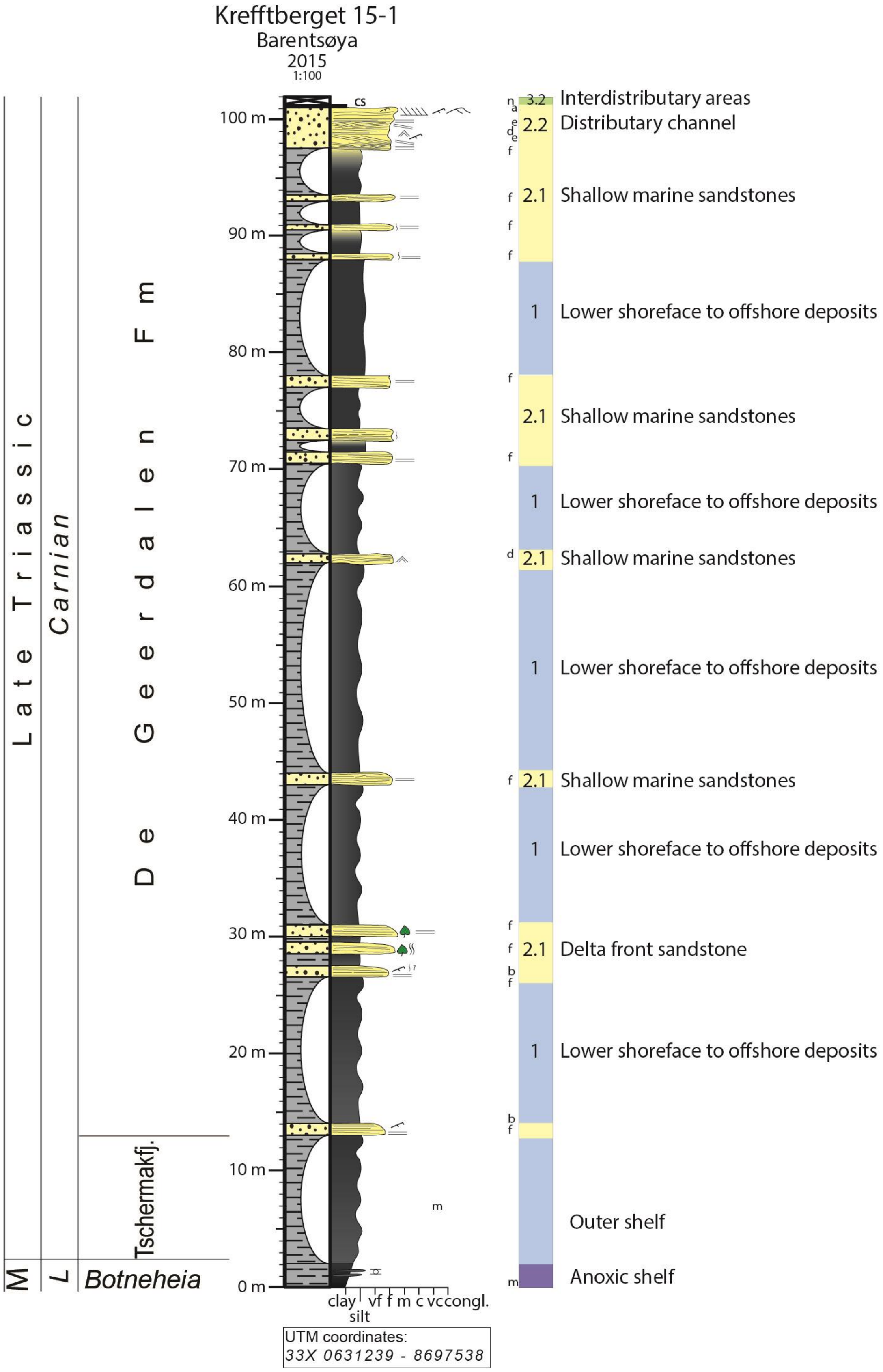


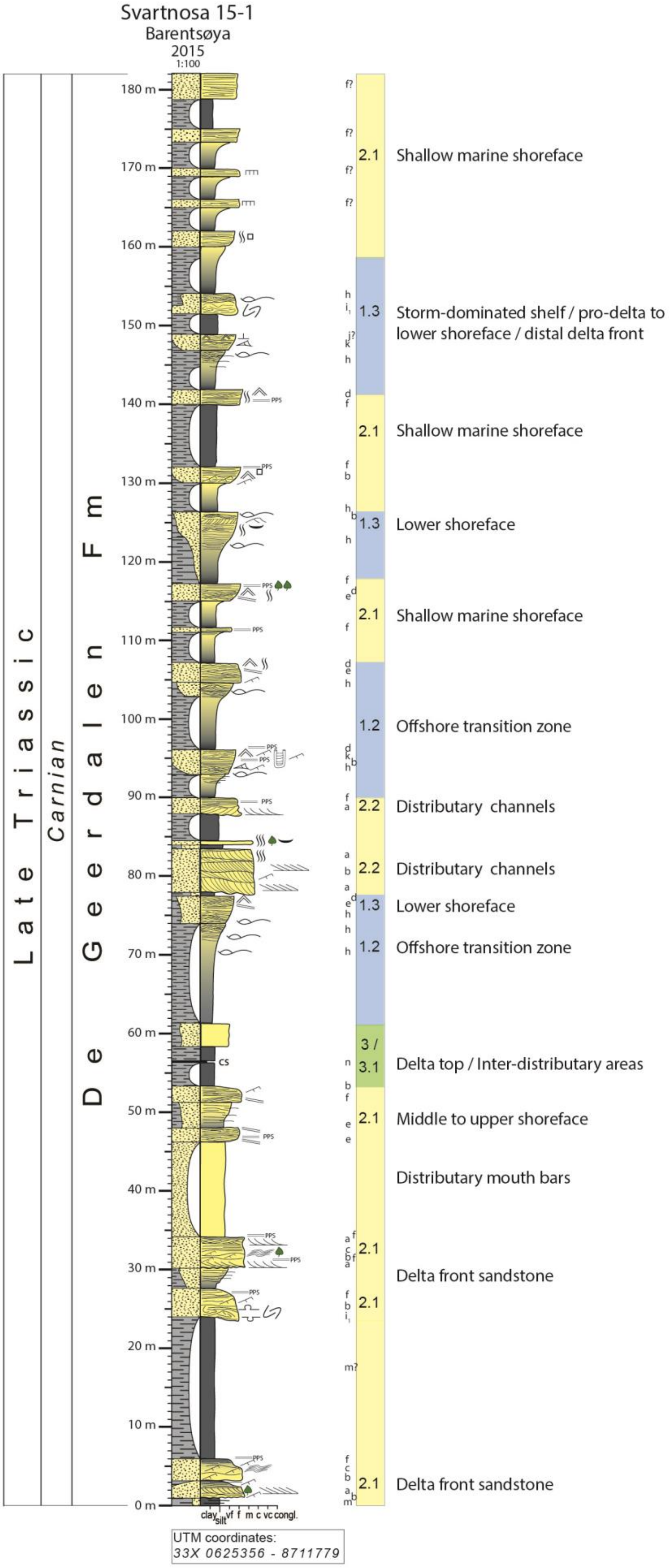
Teistberget 15-1
E. Spitsbergen
2015
1:100
Facies
FA
Late Triassic
Carnian
De Geerdalen Fm
2.1 Shallow marine shoreface
3.2 Interdistributary bay
3.1 Floodplain
2.3 Distributary channel
2.2 Distributary mouth bars
3.1 Interdistributary areas
2.1 Shallow marine shoreface
3.2 Interdistributary bay
2.1 Barrier island
1.1 Offshore deposits
2.1 Shallow marine shoreface deposits



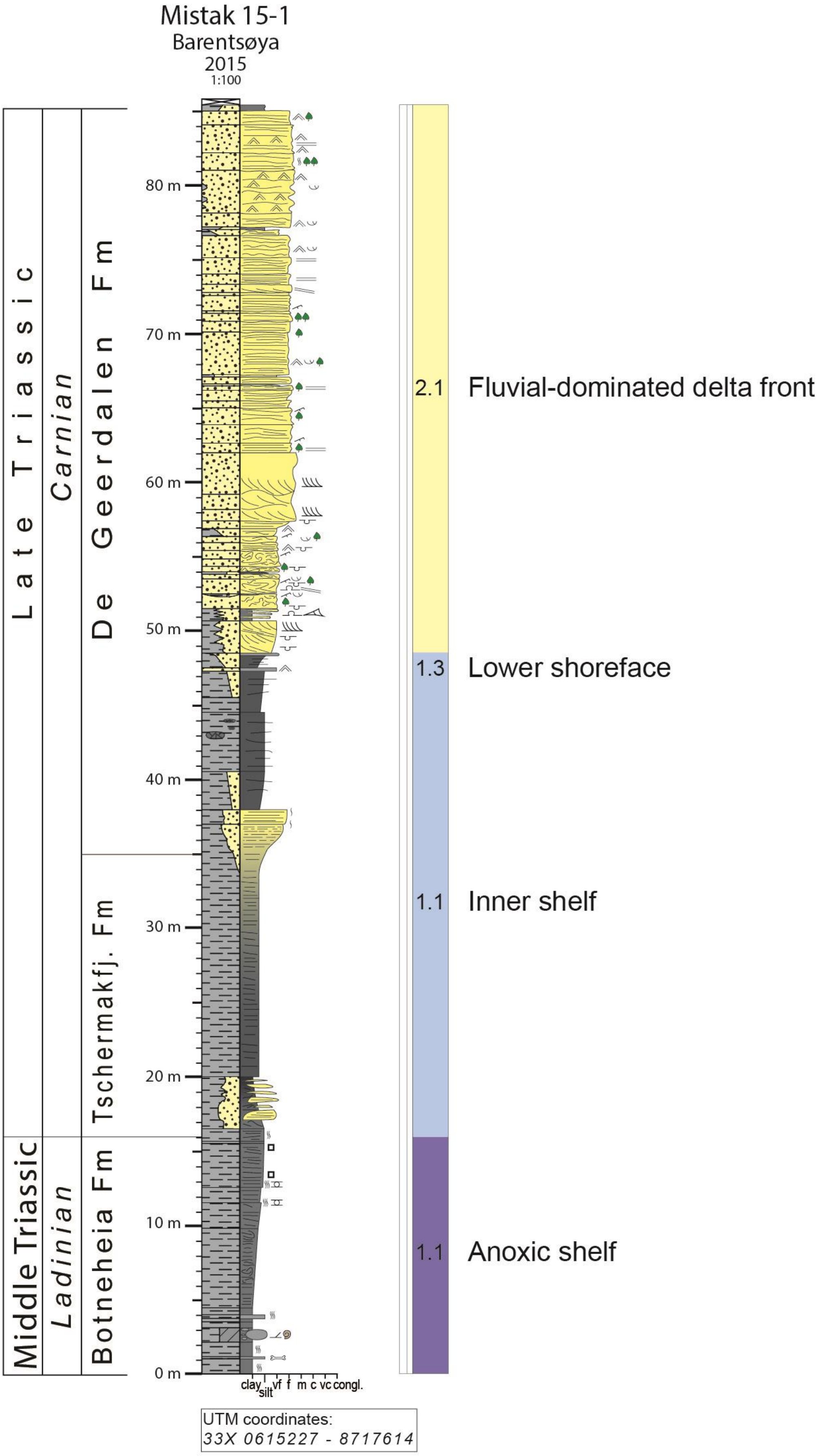
Hellwaldfjellet 15-1
E. Spitsbergen
2015
1:100
Facies
FA
280 m
270 m
260 m
250 m
240 m
230 m
220 m
210 m
200 m
190 m
180 m
170 m
160 m
150 m
140 m
130 m
120 m
110 m
100 m
90 m
80 m
70 m
60 m
50 m
40 m
30 m
20 m
10 m
0 m
Late Triassic
Norian
Carnian
F. Fm
Isfjorden Mb
De Geerdalen Fm
PS
CS
3.1 Floodplain
3.1 Floodplain / Inter-distributary areas
3.2 Interdistributary areas
2.1 Shallow marine shoreface
3.1 Floodplain with crevasse splays
3.2 Interdistributary areas
1.2 Offshore transition zone
2.1 Upper shoreface
1.3 Lower shoreface?
2.1 Upper shoreface
2.1 Shalllow marine shoreface deposits
clay silt vf f m c vc congl.
UTM coordinates:
33X 0625577 - 8745316

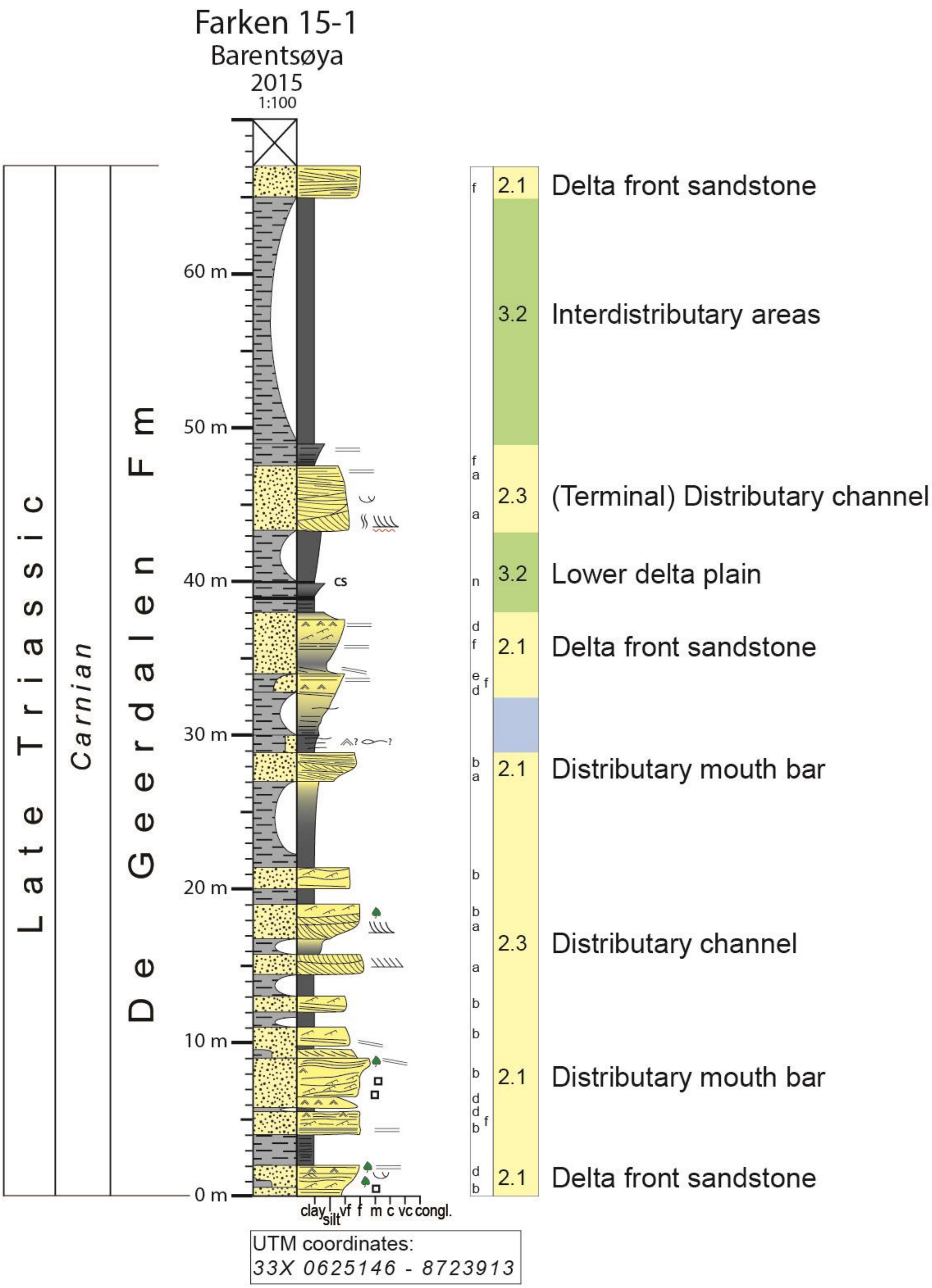

Wilhelm 15-1
Wilhelmøya
2015
1:100
Facies
FA
S. Bed
Norian
Isfjorden Mb
Late Triassic
Carnian
De Geerdalen Fm
230 m
220 m
210 m
200 m
190 m
180 m
170 m
160 m
150 m
140 m
130 m
120 m
110 m
100 m
90 m
80 m
70 m
60 m
50 m
40 m
30 m
20 m
10 m
0 m
3.1 Floodplain with crevasse splays
3.2 Interdistributary bay
3.1 Lower delta plain
3.1 Lower delta plain
3.1 Floodplain with fluvial or distributary channels
3.2 Interdistributary bay
3.1 Floodplain
3.2 Fluvial channels
2.1 Shallow marine shoreface
2.1 Shallow marine bars
1.3 Lower shoreface
clay silt vf f m c vc congl.
UTM coordinates:
33X 0622072 - 8782337


Krefftberget 15-1
Barentsøya
2015
1:100
Late Triassic
Carnian
De Geerdalen Fm
Tschermakfj.
M
L
Botneheia
100 m
90 m
80 m
70 m
60 m
50 m
40 m
30 m
20 m
10 m
0 m
cs
clay
silt
vf f m c vc congl.
UTM coordinates:
33X 0631239 - 8697538
3.2 Interdistributary areas
2.2 Distributary channel
2.1 Shallow marine sandstones
1 Lower shoreface to offshore deposits
2.1 Shallow marine sandstones
1 Lower shoreface to offshore deposits
2.1 Shallow marine sandstones
1 Lower shoreface to offshore deposits
2.1 Shallow marine sandstones
1 Lower shoreface to offshore deposits
2.1 Delta front sandstone
1 Lower shoreface to offshore deposits
Outer shelf
Anoxic shelf



Svartnosa 15-1
Barentsøya
2015
1:100
Late Triassic
Carnian
De Geerdalen Fm
180 m
170 m
160 m
150 m
140 m
130 m
120 m
110 m
100 m
90 m
80 m
70 m
60 m
50 m
40 m
30 m
20 m
10 m
0 m
clay silt vf f m c vc congl.
UTM coordinates:
33X 0625356 - 8711779
2.1 Shallow marine shoreface
1.3 Storm-dominated shelf / pro-delta to lower shoreface / distal delta front
2.1 Shallow marine shoreface
1.3 Lower shoreface
2.1 Shallow marine shoreface
1.2 Offshore transition zone
2.2 Distributary channels
2.2 Distributary channels
1.3 Lower shoreface
1.2 Offshore transition zone
3 / 3.1 Delta top / Inter-distributary areas
2.1 Middle to upper shoreface
Distributary mouth bars
2.1
Delta front sandstone
2.1
2.1 Delta front sandstone


Mistak 15-1
Barentsøya
2015
1:100
Late Triassic
Middle Triassic
Carnian
Ladinian
De Geerdalen Fm
Tschermakfj. Fm
Botneheia Fm
80 m
70 m
60 m
50 m
40 m
30 m
20 m
10 m
0 m
clay
silt
vf
f
m
c
vc
congl.
UTM coordinates:
33X 0615227 - 8717614
2.1
Fluvial-dominated delta front
1.3
Lower shoreface
1.1
Inner shelf
1.1
Anoxic shelf


Farken 15-1
Barentsøya
2015
1:100
Late Triassic
Carnian
De Geerdalen Fm
60 m
50 m
40 m
30 m
20 m
10 m
0 m
cs
clay silt vf f m c vc congl.
UTM coordinates:
33X 0625146 - 8723913
2.1 Delta front sandstone
3.2 Interdistributary areas
2.3 (Terminal) Distributary channel
3.2 Lower delta plain
2.1 Delta front sandstone
2.1 Distributary mouth bar
2.3 Distributary channel
2.1 Distributary mouth bar
2.1 Delta front sandstone