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Science and Technology

Paleosols in the Triassic De Geerdalen and Snadd formations

Jonas Enga

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Supervisor: Atle Mørk, IGB
Co-supervisor: Snorre Olaussen, UNIS

Norwegian University of Science and Technology
Department of Geology and Mineral Resources Engineering

Abstract

Parts of the Late Triassic De Geerdalen Formation and the Middle to Late Triassic Snadd Formation have been investigated through facies analysis of outcrop and core data. The purpose has been to identify paleosols, relate them to the overall depositional environment and infer about Late Triassic paleoclimatic conditions. Outcrops of the lower and middle parts of the De Geerdalen Formation were logged on Edgeøya and Hopen, whereas data from the offshore equivalent Snadd Formation were obtained from a shallow stratigraphic core from the Bjarmeland Platform in the Barents Sea.

Identification of paleosols has been based on recognition of features such as fossil roots in growth position, color variations and horizons, pedogenic slickensides and organic accumulations. They have been found to occur on top of coarsening upwards delta front sandstones, within mud dominated floodplain deposits, and on top of channel sandstones. Where they occur on top of delta front and channel deposits the paleosols can be viewed as abandonment markers, attesting to periods of non-deposition following avulsion events and termination of sediment supply. Floodplain paleosols vary in character, and some appears to be polygenetic, meaning they record a change in moisture regime. Some of the paleosols also shows features that enable classification and comparison with modern soils. Vertisols, a type of soil formed by seasonal moisture variations in swelling clays seems to be an abundant type of paleosol in the Snadd Formation. These have been identified by the presence of pedogenic slickensides, appearing as smooth, curved and striated fracture planes in mudrocks.

The paleosols attest to a dominantly humid climate with a seasonal variation in precipitation. The interpretation of a humid climate agrees with previous studies based on palynology, and supports the validity of *general circulation models* for the Late Triassic. Precipitation seasonality may have had a limiting effect on peat accumulation and possibly a contributing factor to high sediment yield in the fluvial system. Many of the features found in the paleosols appear to be comparable to modern soils on the Mississippi Delta in the Southern USA.

Samandrag

Delar av den sein triassiske De Geerdalsformasjonen og den midtre til sein triassiske Snaddformasjonen har vorte undersøkt ved facies analyse av data frå blotningar og borkjerne. Formålet har vore å identifisera paleosols og å relatera dei til avsetningsmiljøet og sein triassisk paleoklima. Blotningar av nedre og midtre del av De Geerdalsformasjonen har vorte logga på Edgeøya og Hopen, medan data frå den ekvivalente Snaddformasjonen er frå ein grunn stratigrafisk kjerne frå Bjarmeland-plattformen i Barentshavet.

Paleosols har vorte identifiserte på grunnlag av rothorisontar, fargevariasjonar og horisontar, glidespegl og akkumulasjonar av organisk materiale. Dei opptrer på toppen av oppgrovande delta front sandsteinar, i slamrike flomslette avsetningar og på toppen av kanal sandsteinar. Der paleosols opptrer over delta front og kanal avsetningar kan dei betraktast som inaktive flater, som følgje av avulsjonar og påfølgjande stopp i sedimentasjon. Flomslette paleosols viser varierte trekk, og nokre av dei ser ut til å vera polygenetiske, som betyr at dei har gjennomgått ein endring i fuktnivå. Nokre av paleosolane viser eigenskapar som mogleggjer klassifisering og samanlikning med moderne jordsmonn. Vertisols, ein jordsmonntype danna av årtidsvariasjonar i fuktinnhald i svelleleire, ser ut til å vera ein vanlig type paleosol i Snadd formasjonen. Desse har blitt identifiserte ved glidespegl, som framstår som glatte og kurva bruddflater med glidestriper i slamsteinar.

Paleosolane vitnar om eit relativt fuktig klima med årstids variasjonar i nedbør. Tolkinga av eit fuktig klima stemmer overeins med tidligare undersøkelsar basert på palynologi, og støttar gildskapen til globale klimamodellar for sein trias. Årstidsvarisjonar i nedbør kan ha hatt ein hemjande effekt på akkumulasjon av torv, og potensielt medverka til høg sediment tilførsel i elvesystema. Ein del av eigenskapane til paleosolane kan samanliknast med moderne jordsmonn på Mississippi deltaet i det sørlege USA.

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1. Introduction

1.1 Scope

The scope of this thesis is to investigate paleosols the Upper Triassic De Geerdalen and Snadd formations. The distribution and properties of the paleosols will be used to infer about Late Triassic climatic conditions, how the paleosols relate to the depositional environment and comparison with modern soils and sedimentary environments. Some attempt to classify the paleosols will also be made. The study is mainly based on sedimentological fieldwork by facies analysis of data from sedimentological logging of outcrops and core.

1.2 Study Area

The studied areas in this work include Edgeøya and Hopen, both situated on the eastern parts of the Svalbard archipelago, and a core (7430/07-U-01) from a shallow stratigraphic well drilled on the Bjarmeland Platform. Figure 1.1 shown a regional map of the area, outlining the major structural elements in the region.

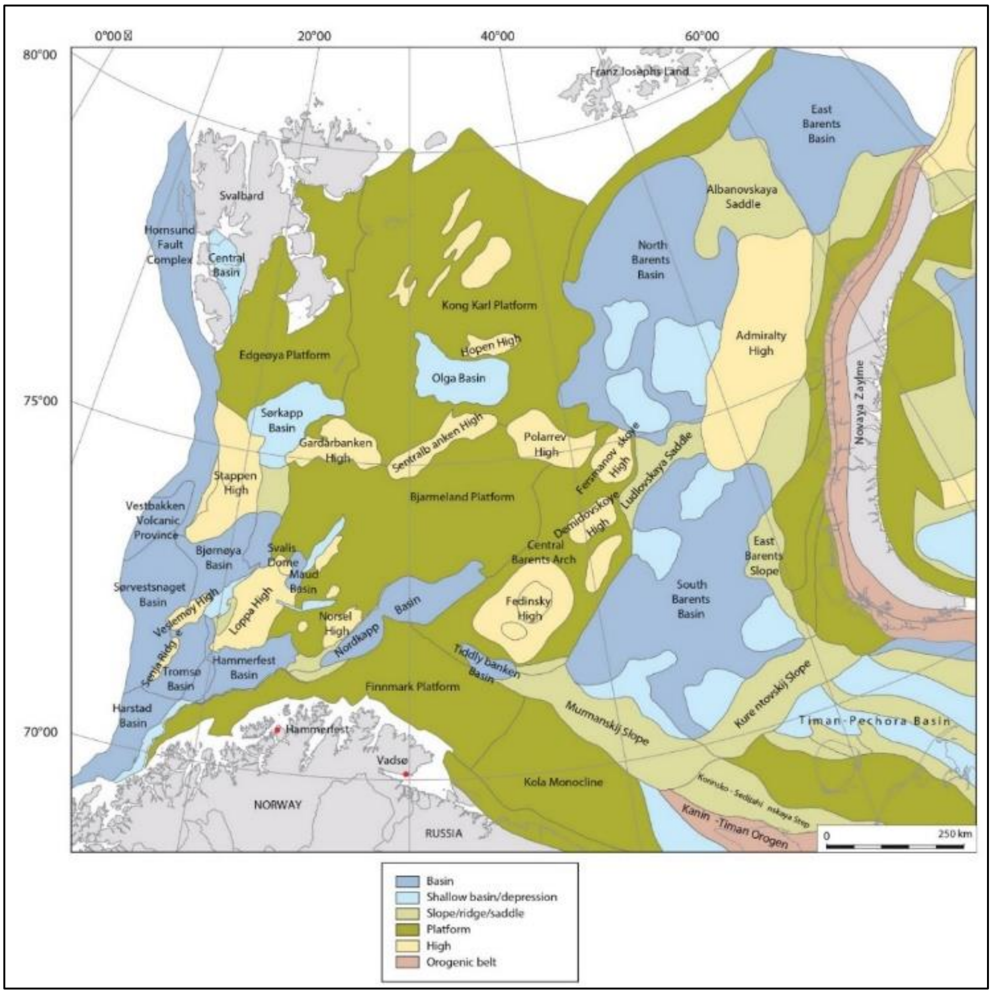

Figure 1.1: *Regional map of the Barents Sea Shelf and Svalbard, with major structural elements. From Ryseth (2014).*

The Barents Sea Self is an approximately 1.3 million km^2 wide area, forming one of the largest continental shelf areas in the world (Doré, 1995). The area is bordered in the south by the Norwegian Baltic shield, the Russian coast and Novaya Zemlya in the east and southeast, and the Atlantic Ocean in the west. The Barents Sea is commonly divided in a western and eastern province, where the Svalbard archipelago forms the uplifted northwestern corner of the western province (Worsley, 2008). The western Barents Sea area consists of various basins, platforms and highs, while the Svalbard archipelago has more of a platform appearance (Faleide et al., 1984, Worsley, 2008).

1.3 Regional Geology of Svalbard and the Barents Sea Shelf

The Svalbard and Barents Sea area has a long and complex geological history, spanning from the Precambrian to present, and with a more or less complete sedimentary succession through most of the Upper Paleozoic to the Cenozoic. The many depositional phases reflects the various tectonic events and the changing climatic conditions as a consequence of northwards plate tectonic drift (Worsley, 2008).

1.3.1 Precambrian to Paleozoic

The pre-Caledonian basement is exposed several places on the western and northern parts of Svalbard. It consists of gneisses and metamorphosed supracrustal and intrusive rocks that evolved through several orogenic episodes, and range in age from Precambrian to Early Paleozoic (Dallmann, 2015).

The Early Ordovician marks the onset of the Caledonian Orogeny, which lasted until the Early Devonian (Dallmann, 2015). This event had major influence of the structural grain of the area, especially in the western Barents Sea and Svalbard where Caledonian trends dominate the basement (Doré, 1995). Following the cessation of Caledonian compression was a period of extension and deposition of Old Red Sandstones in an arid climate, which has only been proven to be present on Spitsbergen (Worsley, 2008, Faleide et al., 1984)

In the Late Devonian the climate had changed from sub-tropical arid to tropical humid, where fluvial and lacustrine sediments belonging to the Billefjorden Group was deposited. (Worsley, 2008). As a consequence of further northwards tectonic drift the climate again became more arid, and a shallow carbonate platform developed by Mid-Carboniferous (Stemmerik and Worsley, 2005). The Middle Carboniferous to Early Permian Gipsdalen Group consists of shallow water carbonates, evaporites and minor siliciclastics (Stemmerik and Worsley, 2005). As the warm waters of the Boreal Ocean cooled in the Early Permian, the cool-water carbonates of the Bjarmeland Group was deposited (Worsley, 2008). This transition has been speculated to be related to changing circulation patterns in the Boreal Ocean, possibly as a response to the developing Uralides (Stemmerik and Worsley, 2005). The formation of the Uralides is also believed to be responsible for the shift to the silica rich spiculitic shales of the

Latest Permian Tempelfjorden Group, likely to have been triggered by closure of the connecting seaway to the warmer Tethys Ocean (Worsley, 2008).

1.3.2 Mesozoic

During the Triassic the Barents Sea was a shallow epicontinental sea, where Late Paleozoic structural elements, sea level fluctuation, and erosion and sediment input from the bordering hinterlands was the main controlling factors on the sedimentary infill patterns (Mørk et al., 1989, Riis et al., 2008, Glørstad-Clark et al., 2010). The Permian/Triassic transition is not fully understood in the region, but involves a transition from silica rich shales to non-siliceous shales, and a significantly warming of oceanic waters (Worsley, 2008).

The Early to Middle Triassic Sassendalen Group consists mainly of fine clastics, and was deposited during high subsidence and sedimentation rates (Vigran et al., 2014). On Svalbard these sediments represent repeated coastal progradations, with deltaic or barrier bar sandstones, where the main sediment source was Greenland, located to the west (Mørk et al., 1982, Vigran et al., 2014, Mørk, 1999). The Sassendalen Group in South-Western Barents Sea shows a similar development, but here the sediments were sourced from the Scandinavian Baltic Shield and the Uralides (Lundschien et al., 2014). The group shows substantial thickness variations, thinning from 700 m in Western Spitsbergen to 60-150 m in Eastern Spitsbergen, and with thicknesses exceeding 1500 m on southwestern Barents shelf (Worsley, 2008, Vigran et al., 2014).

Anoxic conditions in the Middle Triassic led to widespread deposition of highly organic rich shales over large areas, known as the Botneheia Formation on Eastern Svalbard and the equivalent Steinkobbe Formation in southwestern Barents Sea. These conditions was first established in the Barents Sea (Lundschien et al., 2014). The Bravaisberget Formation in western Spitsbergen is considered the proximal equivalent of the Botneheia Formation (Mørk et al., 1982, Mørk et al., 1999).

The Middle Triassic to Middle Jurassic Kapp Toscana Group contains several deltaic and coastal progradations from various source areas, but with the Uralides as an important sediment source. The Kapp Toscana Group will be dealt with more detailed in Section 1.4.

Rifting in the Late Jurassic to Middle Cretaceous led to submergence of the whole region, with continued deposition of siliciclastic sediments of the Adventdalen Group in open marine conditions (Worsley, 2008). Deposition of organic rich shales dominated the Late Jurassic, whereas a lowering of the sea level and more oxygenated conditions at the Jurassic-Cretaceous boundary led to deposition of grey shales, with a more condensed development on the platforms (Faleide et al., 1984, Worsley, 2008). Following was a episode of fluviodeltaic deposits, represented by the Helvetiafjellet Formation on Svalbard, while the Barents Sea saw continued deposition of marine shales and mudstones (Nemec, 1992, Mørk et al., 1999, Midtkandal et al., 2007). Late Cretaceous deposits are not present on Svalbard, as the northern areas underwent uplift and erosion at this time (Worsley, 2008). The Nygrunnen Group represents the Late Cretaceous in basin in southwestern Barents Sea, and consists mainly of claystone (Worsley et al., 1988).

1.3.3 Cenozoic

In the Paleogene the western margin of Svalbard saw the development of a fold-and-thrust belt, caused by collision with Greenland (Steel et al., 1985). The Central Basin of Spitsbergen developed as a foreland basin adjacent to the developing fold-and-thrust belt to the west, which by the Eocene was filled with clastic sediments (Steel et al., 1985, Worsley, 2008). These sediments belong to the Van Mijenfjorden Group, which lie unconformable on top of Lower Cretaceous deposits, consisting of several deltaic build outs (Steel et al., 1985). The time equivalent Sotbakken Group in the Barents Sea consists mostly of marine claystone, and is mainly present in the western basins (Worsley, 2008). After this orogenic event the Norwegian-Greenland Sea opening with associated formation of oceanic crust occurred along the western margin of Svalbard and the Barents Sea (Faleide et al., 1984).

During the Neogene and Quaternary, repeated glaciations caused erosion and alternating subsidence and uplift has been the dominant processes (Mangerud et al., 1996). The glaciations has led to removal of up to 3 km of sediments in the northern areas (Worsley, 2008).

1.4 Kapp Toscana Group

During the Triassic the Barents Sea and Svalbard area formed a wide epicontinental sea, opening towards the Panthalassa Ocean in the northwest, and gradually evolving into a paralic platform through the Late Triassic, as seen on Figure 1.2 (Riis et al., 2008).

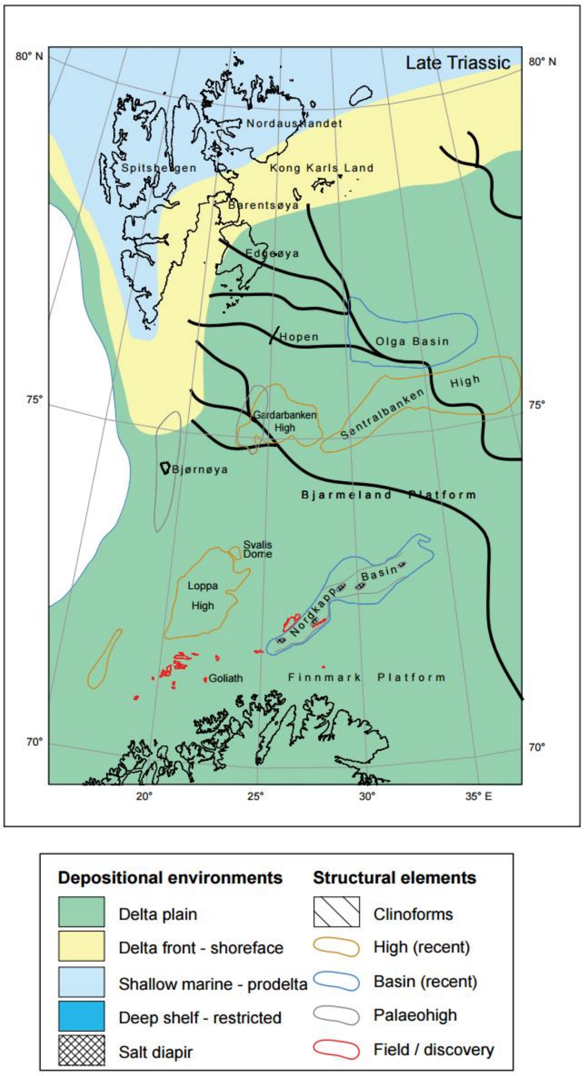

Fig 1.2: *Late Triassic paleogeographic reconstruction of the Svalbard and Barents Sea region. From Lundschien et al. (2014)*

The Triassic has generally been regarded as a period with little tectonic activity in the Northern Barents Sea and Svalbard, with some growth faulting believed to be related to prodelta deformation (Høy and Lundschien, 2011). Recent work has added the suggestion that these faults also might be linked to a tectonic stress field (Osmundsen et al., 2014).

The Late Triassic to Middle Jurassic Kapp Toscana Group is subdivided into the Ladinian to Norian Storfjorden Subgroup , the Norian to Bathonian Wilhelmøya Subgroup on Svalbard, and its offshore Barents Sea equivalent Realgrunnen Subgroup (Mørk et al., 1999). The group is exposed several places on Svalbard where it reaches a thickness of up to 475 meters, while in the Southern Barents Sea the thickness exceeds 1000 meters (Vigran et al., 2014).

The present accepted lithostratigraphy of the Triassic and Early-Middle Jurassic on Svalbard and in the Barents Sea is that of Mørk et al. (1999), which was agreed upon by the many workers on the Mesozoic succession (Vigran et al., 2014). The lithostratigraphy is shown in Figure 1.3, and is from Mørk et al. (2013) where the recently added Hopen member is included (Lord et al., 2014a)

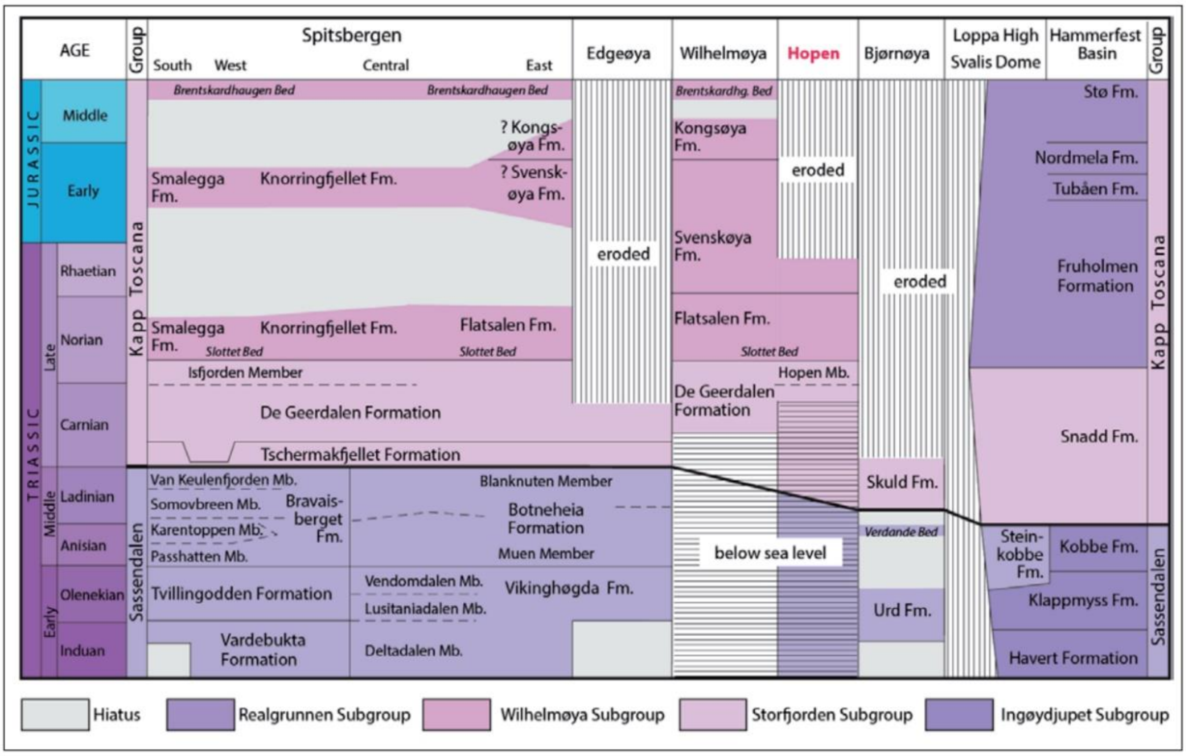

Figure 1.3: *Triassic to Middle Jurassic lithostratigraphy of Svalbard and the Barents Sea. Figure from Mørk et al. (2013).*

1.4.1 Storfjorden Subgroup

The Ladinian to Norian Storfjorden Subgroup consists of alternating sandstones and claystones that show great similarities across the whole of Svalbard and the Barents Sea shelf (Worsley et al., 1988, Lundschien et al., 2014). The Subgroup consists of the Tschermakfjellet and De Geerdalen formations on Svalbard, the Snadd Formation in the Barents sea, and the Skuld Formation on Bjørnøya (Mørk et al., 1990, Mørk et al., 1999). The sandstones of the De Geerdalen and Snadd formations are immature in regards to mineralogy and texture, and commonly classifies as arkosic or lithic arenites (Mørk, 1999, Riis et al., 2008).

The Tschermakfjellet Formation consists of grey shales with siderite nodules, marine fossils and a generally upwards increase in silt and sand, and is believed to be deposited in a prodelta environment (Vigran et al., 2014).

The De Geerdalen Formation crops out at various locations on Svalbard, where it shows important differences from west to east. The base of the Formation is defined as the first prominent sandstone in the Storfjorden Subgroup (Mørk et al., 1999). On western Spitsbergen the De Geerdalen Formation is dominated by several coarsening upward sequences, showing signs of substantial shallow marine reworking of sediments (Mørk et al., 1982, Vigran et al., 2014). The presence of repeated coarsening upwards sequences has been attributed to switching of delta lobes (Mørk et al., 1982). In Eastern and Central parts of Spitsbergen the sandstone bodies can be seen to thicken towards the east, but are generally thin and laterally continuous (Rød et al., 2014). This contrasts with Edgeøya, where sandstone bodies are thicker, interpreted to represent a more fluvial dominated setting and higher amount of accommodation space (Rød et al., 2014). The exposures on Hopen shows an even more proximal setting, where large channel bodies up to 36 m thick, but with widely varying dimensions occur in the lower and middle parts of the succession (Klausen and Mørk, 2014, Lord et al., 2014b). The Norian Isfjorden Member, with its dominantly marginal marine and lagoonal sediment is terminating the De Geerdalen Formation on Spitsbergen (Mørk et al., 1999). On Hopen the uppermost part of De Geerdalen Formation has recently been defined as the Hopen Member, considered to be time equivalent of the Isfjorden Member (Lord et al., 2014a). Abundant hummocky cross-stratified sandstones within dark mudstone suggests a shallow marine environment, whereas the underlying parts of De Geerdalen Formation on Hopen is deposited in alternating terrestrial and marine environment (Lord et al., 2014a, Klausen and Mørk, 2014).

The Snadd Formation in the Barents Sea is penetrated by shallow wells drilled by the Norwegian Petroleum Directorate and SINTEF Petroleum Research and a several exploration wells in the Southern parts (Bugge et al., 2002). Depositional elements in the Snadd Formation include shelf, marginal marine and coastal plain and channel deposits (Bugge et al., 2002, Klausen et al., 2015). The lower Ladinian part contains laminated muddy siltstone resembling the Tschermakfjellet Formation (Lundschien et al., 2014). Large scale fluvial channels in the Snadd Formation has also been described in detail from seismic data (Klausen et al., 2014), and proven to be comparable to those described on Hopen (Klausen and Mørk, 2014) .

The progradational event represented by Snadd, Tschermakfjellet and De Geerdalen formations has been interpreted as a major deltaic system, mainly sourced by sediments from the Uralides and prograding in a dominantly northwesterly direction (Riis et al., 2008, Høy and Lundschien, 2011). The progradation resulted in diachronous lithostratigraphic boundaries, visible as large scale clinoforms on seismic, that are progressively younger towards northwest and with thicknesses suggesting that the delta built into water depth of about 200-400 meters (Lundschien et al., 2014, Riis et al., 2008). The delta was fed by large scale fluvial channels with high sediment yield, that together with the relatively shallow water depth gave the deltaic system a high progradation potential, resulting in paralic and delta plain deposits reaching all the way to Svalbard (Klausen et al., 2014, Lundschien et al., 2014, Klausen et al., 2015).

1.4.2 Wilhemøya and Realgrunnen subgroups

A transgression caused a major change in the sedimentation pattern across the Barents Sea and Svalbard in the Norian, marking the onset of the Wilhelmøya and Realgrunnen subgroups (Vigran et al., 2014). Shallow marine and coastal deposits was still dominating, but with renewed sediment input from the Baltic Shield in the Barents Sea, and Greenland as a source on Svalbard (Pózer Bue and Andresen, 2013, Worsley, 2008, Mørk, 1999). The succession on Svalbard shows a markedly more condensed development, especially in eastern and central areas, than in the Barents Sea (Worsley, 2008). Sandstones in the Wilhelmøya and Realgrunnen Subgroups displays a considerably higher mineralogical maturity than the underlying Storfjorden Subgroup (Mørk, 1999). The transition from the Storfjorden Subgroup to the Realgrunnen and Wilhelmøya subgroups has been attributed to factors such as increased marine reworking caused by decreased subsidence, rejuvenation of the Baltic Shield as a sediment source in the South Western Barents Sea, and possibly a more humid climate (Mørk et al., 1982, Bergan and Knarud, 1993, Mørk, 1999, Ryseth, 2014).

1.5 Late Triassic paleoclimate

The Triassic global climate is generally regarded to have been hot, mainly due to high CO_2 levels in the atmosphere, and more or less represents a continuation of the conditions established in the Late Permian (Preto et al., 2010). An important aspect of global climates during greenhouse periods such as the Triassic, is that the tropical convergence zone is less confined, resulting in a less latitudinal dependent climate zonation (Ziegler et al., 1987, Preto et al., 2010).

In the Late Triassic the central parts of Pangea had a highly seasonal climate, dominated by a large monsoonal system, while the higher latitudes and the coastal regions had a more humid climate (Parrish, 1993, Preto et al., 2010). Within the Carnian is a humid pulse, known as the Carnian Pluvial Event, which has now been recognized over large parts of Pangea (Preto et al., 2010). The full extent and cause of this event is not fully understood, but one of the main theories is that it is related to the Wrangellia volcanic event (Arche and López-Gómez, 2014). The Carnian Pluvial Event was short lived, and the climate established after the Carnian Pluvial Event appears to have been relatively stable throughout the Triassic (Preto et al., 2010).

The Barents Sea and Svalbard`s paleolatitude in the Carnian was about 50-55^{o}N, shown on the reconstruction in Figure 1.4 (Torsvik et al., 2002, Worsley, 2008).

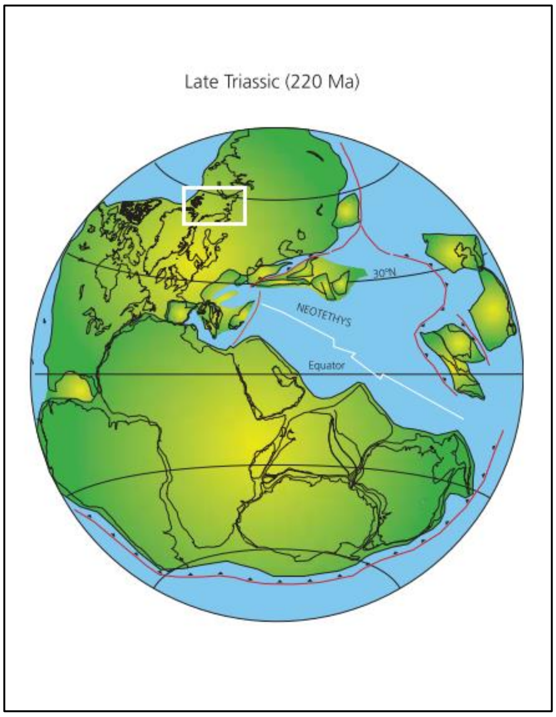

Figure 1.4: *Late Triassic paleogeographic reconstruction of Pangea. Svalbard and the Barents Sea area are outlined with a white square. From Torsvik et al. (2002).*

The Triassic climate in Svalbard and the Barents Sea has mainly been inferred from palynological studies. Hochuli and Vigran (2010) studied core material from the South-West and Central Barents Sea, and grouped palynomorphs into hygrophytic and xerophytic elements. These where then assigned to different floral phases. As hygrophytic plants require an abundance of moisture, whereas xerophytic plants are adapted to more arid conditions, variations in the ratio between these two groups should reflect climatic conditions (Hochuli and Vigran, 2010). Among their findings is a shift from xerophytic conifers (floral phase 11) to hygrophytic fern dominated assemblages (floral phase 12) in the early Carnian, interpreted to reflect a shift from arid to a humid climate. This shift also coincides with the appearance of coal in the cores. No major turnovers was found after this shift, which suggested that the humid conditions persisted throughout the Triassic (floral phase 12-20) (Hochuli and Vigran, 2010). This abrupt shift is unlikely to have been caused by plate tectonic drift alone, and is suggested to have been caused by a change in precipitation pattern, possibly related to the Carnian Pluvial Event (Hochuli and Vigran, 2010). An abundance of ferns in the Carnian succession has also been found on other places in the region, such as Spitsbergen and Kong Karls Land (N. Patterson, pers. Comm. Mars 2015). Paterson and Mangerud (2015) notes that their Upper Carnian *Leschikisporis aduncus* assemblage from Hopen is comparable to the floral phase 12 of Hochuli and Vigran (2010). Carnian palynomorphs from Siberia suggests warm humid condition prevailed there as well (Ilyina and Egorov, 2008).

Increasingly sophisticated *General Circulation Models* has made it possible to create models that are able to recreate the occurrence of climate sensitive rock facies (Sellwood and Valdes, 2006). The parts of Pangea that faced the Boreal Sea have been modeled to have a warm temperate, humid climate and warm temperate with dry summers (Sellwood and Valdes, 2006). The model from Sellwood and Valdes (2006) also suggests variations in the precipitation through the year, with rates of 4-8 mm/day during the winter, and 1-2mm/day during summer. This would add up to a mean annual precipitation between 1000 and 2000 mm. The model of Sellwood and Valdes (2006) is shown in Figure 1.5.

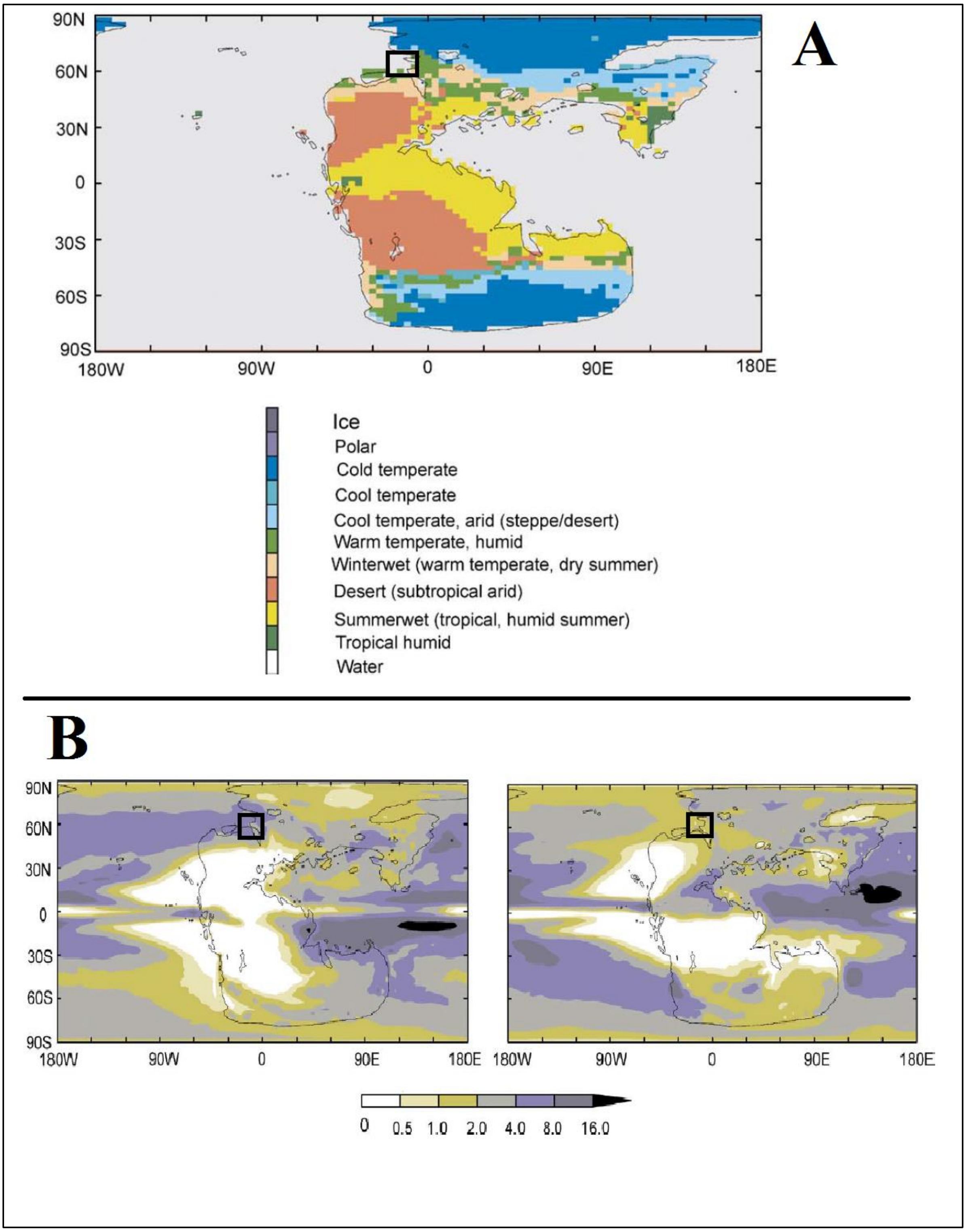

Figure 1.5: *Figures from the* general circulation model *by Sellwood and Valdes (2006). Squares outline the position of Svalbard and the Barents Sea region.* **A)** *Modelled Late Triassic paleoclimate of Pangea.* **B)** *Modelled precipitation patterns of Pangea in Late Triassic. Units is mm/day. Winter to the left and summer to the right.*

Ryseth (2014) did a study on Late Triassic to Early Jurassic core material and suggested an increase in humidity during the timespan, and that this could have influenced the sedimentation pattern. A similar development is well documented in the North Sea (Nystuen et al., 2014).

Recent work on the upper part of the De Geerdalen Formation on Spitsbergen, has revealed a lack of coalbeds, abundant desiccation cracks and mature calcrete paleosols, which has led to suggestions that the climate was more variable and at times semi-arid during the Late Triassic (Husteli et al., 2015, Knutsen, 2013).

1.6 Previous work

Geological fieldwork on the Triassic succession on Svalbard has been carried out since late 19th century, with scientific contributions mainly from Swedish, Russian, Polish, British and Norwegian geologists (Vigran et al., 2014). Research activity in the last decades has primarily been motived by the petroleum potential of the Barents Sea, where the onshore exposures on Svalbard has proven to be excellent analogs to the subsurface geology in the Barents Sea (Vigran et al., 2014, Worsley, 2008). The Snadd Formation has been studied from seismic, and drill cores from deep exploration wells, and shallow stratigraphic wells.

Recent published work on the De Geerdalen and Snadd formation has focused on the nature of channel bodies, seismic expression of clinoforms and sequence stratigraphy (Glørstad-Clark et al., 2010, Høy and Lundschien, 2011, Klausen et al., 2014, Klausen and Mørk, 2014, Lord et al., 2014b, Lundschien et al., 2014, Klausen et al., 2015). Work regarding depositional environment, diagenetic evolution and regional development across Svalbard has been done by (Mørk et al., 1982), Vigran et al. (2014) and (Rød et al., 2014). Among other published work on the Triassic succession on eastern Svalbard is that of Flood et al. (1971), Smith et al. (1975), and Lock et al. (1978), who summarize the stratigraphical and sedimentological development in the area. A general model of the evolution of the Triassic shelf in the northern Barents Sea based on stratigraphic cores and seismic was made by Riis et al. (2008). Provenance and petrographic studies of the Triassic has been conducted by Mørk (1999) and Pózer Bue and Andresen (2013). Dating of the Triassic succession is mainly based on ammonoids and palynological methods, where much of the palynological work is compiled in

Vigran et al. (2014). A refined and more detailed palynological zonation of the Triassic succession on Hopen was recently made by Paterson and Mangerud (2015).

Core 7430/07-U01 from the Bjarmeland Platform was investigated by Stensland (2012), and the present work on this core has benefitted heavily from her work. Her work included an interpretation of depositional environment and diagenic evolution, based on conventional logging and petrographic analysis. The interpretation of the depositional environment was a tidal influenced delta plain with extensive early diagenesis and soil formation, and a general transition from lower delta plain to mid-upper delta plain upwards in the core. The distinction between partly oxidized and waterlogged reduced paleosols was also made, and their coexistence was attributed to their position relative to channels on the delta plain. Oxidized and mottled paleosols was interpreted to represent upper delta plain, while grey waterlogged paleosols indicated a position closer to the shoreline on the lower delta plain. In addition mineralogical description of carbonate cemented horizons in the soils where made. Several microstructures related to the carbonate suggested that much of the carbonate was precipitated as pedogenic carbonate.

Palynological analysis of the core was presented by Vigran et al. (2014). The identified palynomorph associations suggested alternating marine and terrestrial environments, assigned to the *Aulisporites astigmosus* Composite Assemblage Zone of early to middle Carnian age (Vigran et al., 2014).

2 Paleosols

As paleosols is a fundamental part of this thesis, a section with a review of soil forming processes and the geological significance of paleosols is included. Paleosols has been proven helpful in solving a diverse range of geological problems, but are for the most applied in studies of paleoenviroments (Kraus, 1999, Retallack, 2001). The main reason for their applicability is that they can record hydrological conditions and provide information on the physical, chemical and biological processes responsible for forming them (Kraus, 1999, Retallack, 2001).

2.1 Definitions and soil properties

A paleosol is simply defined as a fossil, or buried soil, but defining soil itself is more problematic, and differs widely among various scientific groups. From a biologists standpoint a soil is usually defined as a medium in which plant takes roots, while engineers commonly refers to soil as any loose surface material (Retallack, 2001). The meaning of a soil from a geological standpoint is commonly taken to be a surface of sediments or solid rock that has undergone alteration, by physical, biological or chemical processes (Retallack, 2001). All process contributing to this alteration is termed pedogenic, or soil forming processes (Wright, 1992b). Figure 2.1 shows an overview of these processes and how they interact and relate to each other. Sedimentation, plants and other organisms, and precipitation and evaporation mainly control processes at the surface, while sub-surface processes are controlled by downwards percolating surface waters, groundwater and transpiration by plants.

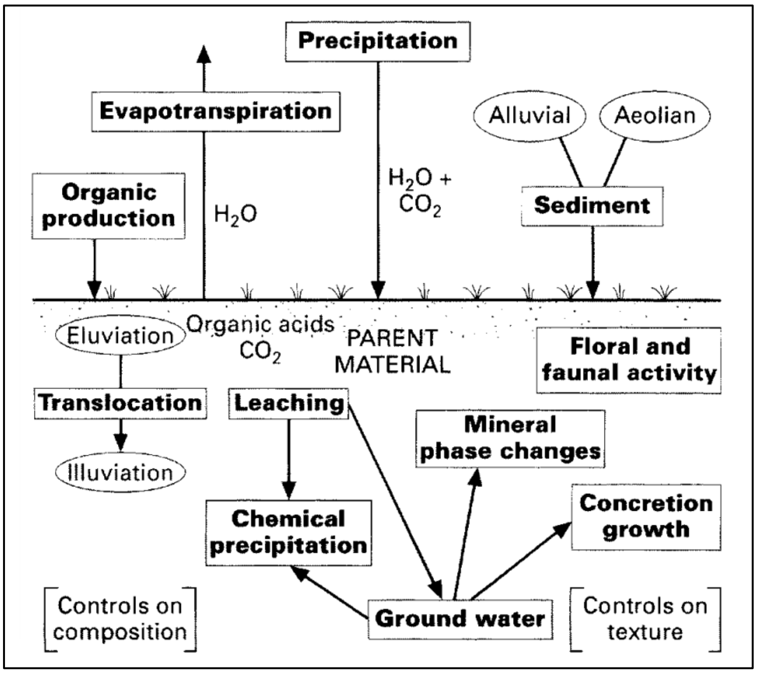

Fig 2.1: *Soil forming processes with main control on texture and composition of soils. Figure from Collinson (1996)*

Modern soils are often characterized by being highly porous and filled with open cracks and hollows (Retallack, 2001). The basic constituents of a soil are called peds, which are aggregates of soil material. The surfaces of peds, cracks and voids in soils is called cutans (Retallack, 2001). Paleosols can sometimes retain some of their original structure, where the peds and cutans can be identified, but these features are often completely wiped out by compaction and diagenesis, giving them a more massive and homogenous appearance (Retallack, 2001). Another defining property of many types of soils is horizons. Not all soils have well developed horizons, but where they are present they differ on their physical properties and dominating processes (Retallack, 2001). Common for most soil types is the disruption of plants and animals, by roots and burrowing activity. This is commonly referred to as pedoturbation, which essentially is a form of bioturbation (Wright, 1992a). In soils forming on loose sediments this activity often causes primary sedimentary structures such as bedding or lamination to be destroyed. This is referred to as destratification, and can often be one of the most noticeable features of a paleosol (Wright, 1992a).

Another distinct feature of many paleosols is mottled pattern of different colors, often variations of red brown and grey. These color variations are commonly explained by localized changes in redox conditions, which can be facilitated by e.g. a fluctuating groundwater level

(Duchaufour, 1982). Hematite is the most common mineral associated with red colors in paleosols, and is formed by oxidation of iron (Wright, 1992a). The abundance of color mottling in paleosols found in the rock record has led to the assumption that it might form in relatively short time, but no precise estimates of the time needed to form color mottling exists (see Figure 2.1) (Wright, 1992a). Diagenetic alteration of iron hydroxides to hematite can also cause reddening, meaning that red coloration or mottling is not necessarily restricted to paleosols (Wright, 1992a) .

The maturity of a soil or paleosol depends primarily on the residence time of sediments within the zone of pedogenesis, which is commonly assumed to be around 2 meters vertical depth (Wright, 1992a). This means that sedimentation rate is closely linked to the maturity of paleosols in actively aggrading systems. Assessing the time contained within a certain paleosol is not an easy task, as little information exists on how much time is needed to various features (Wright, 1992a). Figure 2.2 shows estimates of the time frame associated with some common features in paleosols, but as stated by Wright (1992a), most of them are associated with uncertainties, and also depend on other factors such as climate and properties of the parent material.

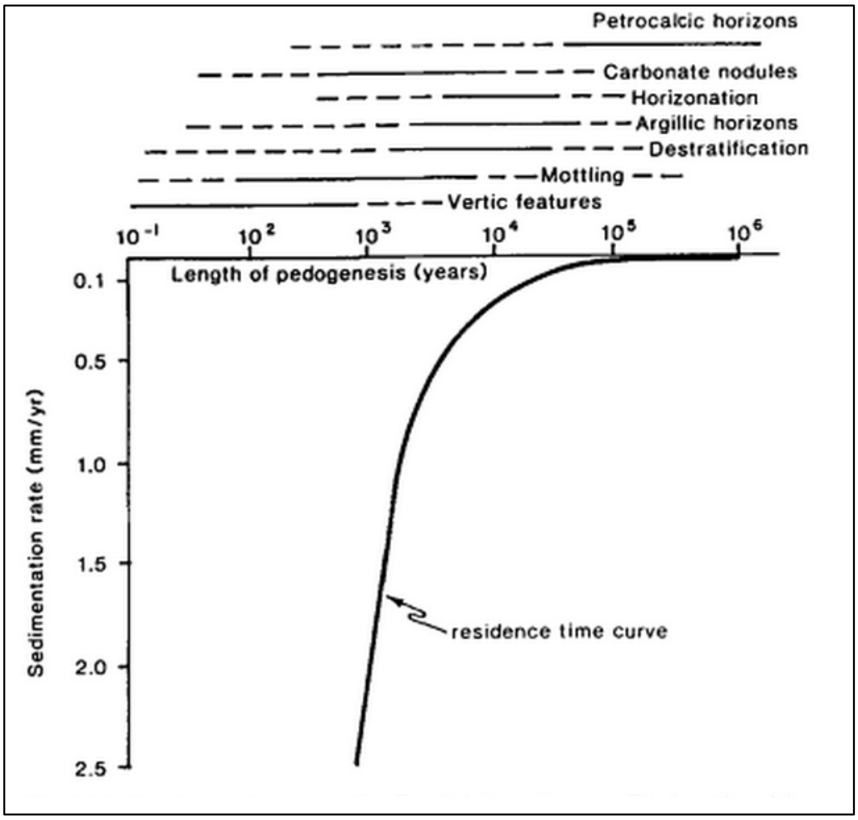

Figure 2.2: *Common features of paleosols, with the approximate time needed to form them on horizontal axis. Note that there are little information on some of the features. Vertical axis is showing sedimentation rate. From Wright (1992a).*

The environmental factor that soils are most sensitive to is the moisture regime, which is essentially determined by how wet the soil is. This is highly dependent on groundwater levels and precipitation patterns, which again varies according to the prevailing climate (Retallack, 2001). Well drained soils with low moisture content tend to be oxidized with little organic matter, whereas poorly drained soils with high moisture content is usually reduced, often with well-preserved organic matter (Retallack, 2001).

When determining paleoclimate from paleosols a uniformitarian approach is usually applied, where distribution and properties of modern soils is used as a basis for reconstructing the paleoclimate (e.g. Mack and James, 1994). Global distribution of soils vary according to temperature and precipitation regime, and dry oxidized desert soils differ widely from arctic soils with permafrost, which again differs from the thick histosols, or peats in many humid tropical areas (Retallack, 2001). Figure 2.3 shows how some of the soil orders of the USDA soil taxonomy relate to climate, expressed as number of wet months throughout the year.

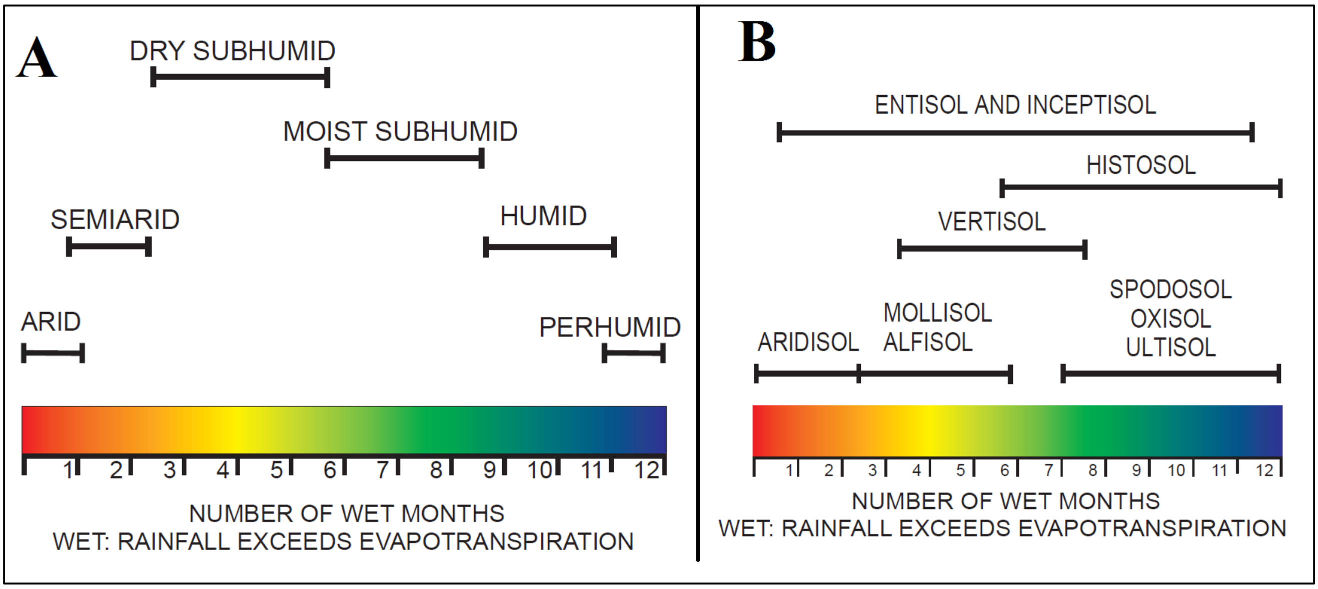

Figure 2.3: A) *Climate classification based on precipitation and evaporation seasonality. Figure from Cecil et al. (2003)* **B)** *Soil orders of the USDA soil taxonomy classification system in relation to precipitation and evaporation seasonality. Most of the soil orders are explained in Section 2.2. Figure from Cecil and Dulong (2003).*

Apart from climate, other factors also influence the character of soils. Parent material, vegetation, time period of soil formation, and landscape position are all variables that determines physical and chemical properties of the soils (Mack and James, 1994, Kraus, 1999). The hydrological conditions in environments such as alluvial and delta plains are also strongly influenced by the climate of the catchment area, and properties of the fluvial system (Kraus and Aslan, 1999). This relationship can make it difficult to ascribe distinct features of a paleosol to climatic conditions in the area they have formed, as it can be difficult to separate the effects of flooding from local precipitation. In epicontinental basins where parent material and relief tend to be relatively uniform, climate is the main controlling factor on pedogenesis (Cecil and Dulong, 2003)

2.2 Classification and soil diversity

Classifying paleosols can be problematic, and is not always as straight forward as classification of modern soil. The main reason for this is the low preservation potential of certain pedogenic features, often erased by diagenesis and compaction (Retallack, 2001). In addition parts of the soils can be subjected to erosion, resulting in preservation of only parts of the profile. After the arrival of vascular land plants in Devonian, the soil forming mechanisms can in many ways be compared to modern processes (Mack and James, 1994, Retallack, 2001).

In classifying paleosols there are two schools of thoughts. Some wants to classify them by using classification systems developed for use on modern soils, while others argue that classification systems should be designed for specific use on paleosols (Retallack, 2001). Many modern soil classification system exist, but the most used system in paleopedology is the USDA soil taxonomy classification system. The USDA soil taxonomy classification system relays on horizon recognition, and are designed to be used on modern soils. Mack et al. (1993) proposed a classification system designed specifically for paleosols, by omitting some of the criteria that are rarely preserved in paleosols. Both the Mack et al. (1993) and the USDA soil taxonomy classification are widely used by workers on paleosols. In addition, some workers have preferred the classification system proposed by Duchaufour (1982), while others have adopted the FAO-Unesco classification system (Kraus, 1999, Retallack, 2001). These classification systems will not be further mentioned here.

Some of the major soil types found both as paleosols and modern soils will be presented in the following sections. The review will focus on the type of soils that are considered relevant for this thesis. It is important to note that many paleosols has features from more than one type of soil, and classification of paleosols is usually done by assessing what feature that is most prominent (Mack et al., 1993).

2.2.1 Shrink-swell soils

Shrink-swell soils, or vertisols as they are commonly referred to, are soils that are influenced by cycles, usually seasonal, of wetting and drying (Retallack, 2001). They are usually characterized by a significant amount of clay with high swelling potential, commonly smectite. Today they are most common in semi-arid and semi-humid parts of the world, with mean annual precipitation between 180 mm and 1520 mm, but are also found in more humid regions such as Southern USA (Khitrov and Rogovneva, 2014, Retallack, 2001) . They largely occur between 50^{o}N and 45^{o}S, and are especially abundant in India, Western Australia, and Sudan (Buol et al., 2011). Common to all areas where vertisols are abundant is a dry season, but the length of the dry season is widely variable in the various areas and climatic regimes (Buol et al., 2011).

Figure 2.4: *Modern vertisol from Southern USA. Notice the deep cracks. Picture from: http://www.nrcs.usda.gov/wps/portal/nrcs/detail/soils/survey/class/maps/?cid=nrcs142p2_053611*

During the dry season these soil have deep, open cracks as seen in Figure 2.4. These cracks can sometimes be filled with material derived from the surface, forming so called clastic dikes which is common in many vertisols (Caudill et al., 1997).

A common result of extensive shrink-swell activity is homogenization of the soil profile, and consequently these soils usually have no distinct horizonation (Retallack, 2001). Evidence of vertic features in paleosols include slickensides, wedge shaped peds where they intersect, and surface microtopographic disturbances or pseudoanticlines caused by pedoturbation (Retallack, 2001, Collinson, 1996). Smectite clay however, should not be regarded as a necessary evidence for shrink-swell processes, as they commonly react to form other clay minerals during diagenesis (Driese and Foreman, 1992). Slickensides formed in vertisols are often referred to as pedogenic slickensides, to avoid confusion with similar structures formed by tectonic processes (Gray and Nickelsen, 1989). They are differentiated from tectonic slickensides by their distinct, highly curved morphology, and their association with other pedogenic features (Gray and Nickelsen, 1989).

Vertisols have been reported in ancient deposits from many different basins and geological periods (e.g. Gray and Nickelsen, 1989, Driese and Foreman, 1992, Driese and Ober, 2005, Prochnow et al., 2006, Rosenau et al., 2013b), and have even been identified in Precambrian deposits (Retallack, 2001). On the Norwegian continental shelf they are common in the Middle Triassic to Lower Jurassic Hegre and Statfjord groups in the Northern North Sea (Müller et al., 2004, Nystuen et al., 2008, Nystuen et al., 2014). The fact that vertisols are so abundant in the geological record has led to the assumption that they may form relatively rapidly compared to other types of soils (Wright, 1992a). It has been argued that a few hundred years is sufficient to form them on clay rich sediments (see Figure 2.1) (Wright, 1992a, Retallack, 2001). Unlike marine clay deposits, clays in vertisols have high bulk densities, and undergo little compaction during burial as they mostly compact before significant burial by swelling and contraction. Caudill et al. (1997) found that vertisols buried to a depth of 3.5 km, only compacted about 10%. The fact that vertisols are so dense and cohesive is probably also a contributing factor to their high preservation potential (Buol et al., 2011).

2.2.2 Organic soils

Soils rich in organic matter are called histosols, a soil order found both in the USDA soil taxonomy and in the Mack et al. (1993) classification. Highly organic rich accumulations are usually called peat. Upon burial and diagenesis, peat is transformed to coal, defined as a rock containing more than 50 weight percent and 70 volume percent carbonaceous material (McCabe, 1984). All qualities and rank of coal qualify as histosols (Mack et al., 1993), although coal is by far the most used term by geologists.

In the USDA soil taxonomy classification system the surface peat thickness must be at least 40 cm in order for a soil to classify as a histosol (Soil Survey Staff, 2014). This criteria is problematic to apply directly to paleosols, as compaction makes it difficult to assess the original thickness of the peat. The magnitude of coal compaction has been highly debated, and estimates of the compaction factor in the literature ranges from 1.2 to 30 (Large and Marshall, 2014).

Accumulation of plant debris by transportation and deposition can create allochthonous coal deposits, but most coals are believed to have formed as autochthonous accumulations, thus essentially being paleosols (Retallack, 1991). The vegetation of peatlands is usually rooted in a soil, often referred to as underclay or seathearth by coal-geologists (Hughes et al., 1992). The nature of these so called underclays found below coals has been subject to much debate. Advancement in paleosol research in recent decades has led to the understanding that most of the often muddy sediments found below thick coal seams, in fact are ancient soils (Hughes et al., 1992). Histosols often go through a progressive development, starting as an immature soil, and as more peat accumulates, and become too thick for roots to penetrate into the substratum, the peat itself acts as the substratum for the vegetation (Diessel, 1992, Retallack, 2001). Figure 2.5 illustrates this development.

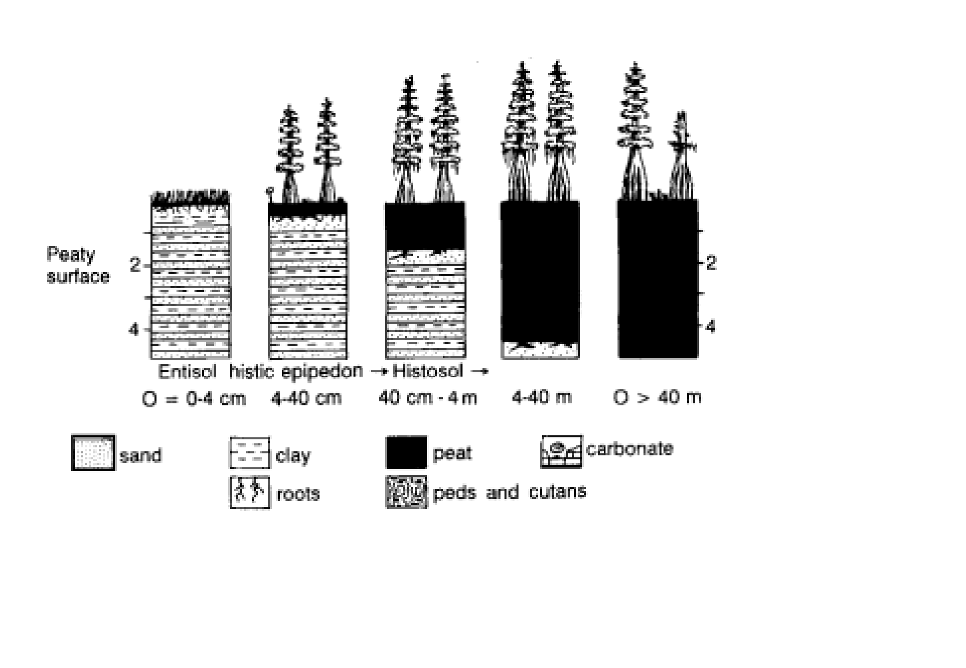

Figure 2.5: *Typical evolution of a histosol. Figure from Retallack (2001).*

In order to form histosols organic matter or peat must accumulate at a rate higher than the rate of decomposition (Diessel, 1992). This is usually the case in wetland areas, such as swamps, marshes and fens, which are characterized by being permanently or seasonally waterlogged. Permanent waterlogging causes rapid depletion of oxygen in the waters, which reduces the activity of oxygen dependent microbes that decomposes organic material (Retallack, 2001). Modern peat deposits accumulate both in humid, warm areas with high organic productivity, and in high latitude, more temperate regions, where the lower temperatures retards the decomposition of organic matter (Diessel, 1992). Seasonal dry swamps wetlands have a slow rate of peat accumulation, as oxidation during drought inhibits preservation of organic matter (Retallack, 2001, Lottes and Ziegler, 1994).

Wetlands can be fed both by groundwater and rainwater. Rainwater fed wetlands, termed ombrogenous peatland, tend to evolve into raised, or domed swamps, which means that they are topographically higher than surrounding areas (Diessel, 1992). This is an effective mechanism in excluding siliciclastic sediments as their topographic position makes flooding by rivers rare, and often form low-ash, thick, laterally extensive, and often economic coal seams (McCabe, 1984). Raised swamps require mean annual precipitation above 2500 mm, more or less evenly distributed through the year (Lottes and Ziegler, 1994, Retallack, 2001)

Groundwater fed wetlands tend to form in low lying areas, and are often called planar, or rheotrophic peatlands, and mean annual precipitation above 1300 mm is usually required to form them (Retallack, 2001). Coals from these low-lying peatlands tend to be higher in ash content than coals from raised swamps (Diessel, 1992). In order for coal to form in low lying areas, the wetlands must have a favorable position relative to sediment sources, or have formed at a time when sediment supply was low (McCabe, 1984). Most thick, low-ash economic coal deposits are believed to have formed either as precipitation fed raised swamps, or developed during periods of low sediment supply as groundwater fed rheotrophic wetlands (McCabe, 1984).

Uncertainties in the peat to coal compaction rate also makes it difficult asses the amount of time that coal deposits represent. Large and Marshall (2014) introduced an alternative approach to this question. By assessing the carbon fraction and thickness of the coal, they suggest that modern, latitudinal dependent carbon accumulation rates could yield approximate values for the time contained within coal seams, without the need to consider compaction.

2.2.3 Horizonated soils

Soils with well developed horizons are usually freely drained, meaning that surface waters percolate downwards through the soil. This process leads to downwards transport of clay, organic material and solutes which is deposited and precipitated further down in the soil.

Common horizons in modern soils includes an upper organic rich horizon, named A or O depending on the mineral fraction. The A horizon consist of organic material mixed with minerals, whereas the O horizon consists of dominantly organic matter. Further down in the profile there can be an E horizon, characterized by a high proportion of weathering resistant minerals, and removal of mainly clay and dissolved material by downwards percolating surface waters, a process termed elluviation, or translocation (Retallack, 2001). Chemical leaching or dissolution of unstable minerals also occur in this zone (Wright, 1992a). These horizons are also called the zone of leaching, and is often associated with a white to grey color, determined by the color of the often quartz rich residue (Wright, 1992a). Below the E horizon is the B horizon. This is where the elluviated material is accumulated, which makes these horizons rich in clay and mineral precipitates (Retallack, 2001). The horizon where parent material only show slight alterations, such that original features as stratification or ripples are preserved is called C horizon (Retallack, 2001). Several subordinate modifiers to further describe and classify horizons exists, but are not included here.

Soils that are dominated by eluviation of clay is in the Mack et al. (1993) classification called argillisols, which are equivalent to the ultisol and alfisol of the USDA soil taxonomy. Spodosol is another type of freely drained soil, included in both classification systems, and is characterized by eluviation of both iron oxides and organic material (Mack et al., 1993).

2.2.4 Immature soils

Both the inceptisol and entisol soil order of the US soil taxonomy are soils that are characterized by weak development of pedogenic features (Soil Survey Staff, 2014). Entisols represents very weakly developed soils, while inceptisols are weakly developed. In the classification system of (Mack et al., 1993) they are combined, and equivalent to the protosol. They form where only short time windows for soil formation exists, or when conditions are generally unfavorable for soil formation (Retallack, 2001). Immature soils usually consists of an upper organic rich A or O horizon and a lower C horizon where only slight alteration can be seen. They lack features of more evolved soils, such as an E or B horizon (Retallack, 2001). These soils forms in all climates, but tend to form more rapidly in humid, warm climates due to higher weathering rates and more abundant vegetation (Retallack, 2001). They are particularly common in alluvial and deltaic environments where active sedimentation prevents formation of mature soils (Retallack, 2001). Immature soils can form within decades, but features such as destratification may require longer time (see Figure 2.2) (Wright, 1992a).

2.2.5 Carbonate soils

Both modern soils and paleosols are often characterized by direct precipitation of minerals (Retallack, 2001). Carbonate minerals are particularly common, and are often termed pedogenic carbonate where they occur in a soil. Pedogenic carbonate are important proxies for ancient atmospheric CO_2 concentrations, and are often used to reconstruct ancient climatic conditions, especially precipitation patterns (Retallack, 2001, Retallack, 2005, Breecker et al., 2013). Siderite and calcite is the most common pedogenic carbonate minerals (Retallack, 2001, Ludvigson et al., 2013). The type of carbonate mineral that precipitates in modern soils is highly dependent on climate, and correlates well with the major climatic belts in the world (Ludvigson et al., 2013). Calcite tend to accumulate in arid regions with excess of evaporation, whereas siderite is the dominant mineral in humid regions, as shown in Figure 2.6.

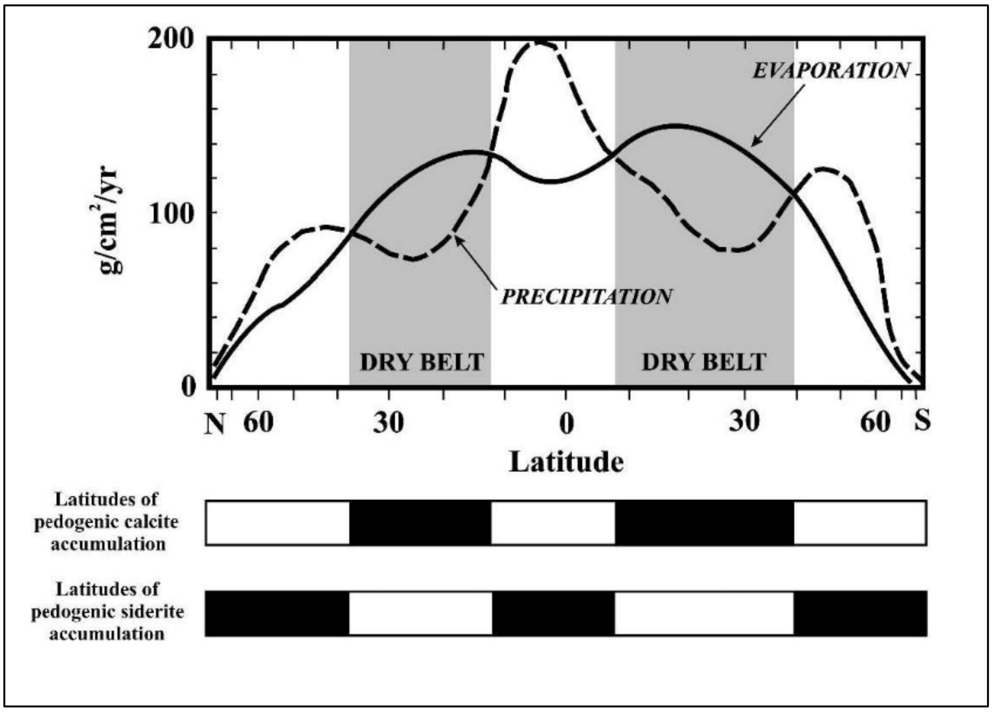

Figure 2.6: *Modern pedogenic carbonate accumulation in relation to climatic zones, illustrated with the precipitation and evaporation budget. Figure from Ludvigson et al. (2013)*

Carbonate minerals can precipitate by a number of mechanisms in many environments, so in order to assign carbonate minerals to a soil, it must be proven that the mineral formed contemporary with soil formation (Retallack, 2001). Such evidence can be cemented roots, called rhizoliths or microstructures such as *Microcodium* (Milnes, 1992). Investigation of stable isotopes can also be used, where it can be possible to infer what type of fluids that precipitated the carbonate (Cerling, 1984).

Precipitation of siderite in a soil is strongly favored in reducing, stagnant groundwater of permanently waterlogged, poorly drained soils (Ludvigson et al., 2013). Accumulation of calcite on the other hand, is driven by the dissolution of calcium-bearing minerals in the upper parts of the soil, and precipitated by downwards percolating surface waters. This process is often enhanced by plant transpiration, which promotes precipitation by increasing the calcium concentration (Milnes, 1992). Calcite in soils can also be precipitated by groundwater, a processes which also occur in more humid climates (Tandon and Gibling, 1994).

Caliche, or calcrete is names often used for soils and paleosols that have been cemented by calcite (Milnes, 1992). They mainly form in areas with arid to semi-arid climates, and in soils on both calcareous and non-calcareous material (Milnes, 1992). The occurrence of Holocene calcretes has been carefully mapped throughout the world, and there exists extensive data on these soils (e.g. Retallack, 2005, Royer, 1999). Global distribution of modern soils with calcite can largely be found to be confined within areas receiving less than 1000 mm mean annual precipitation (Mack and James, 1994, Royer, 1999, Retallack, 2001, Retallack, 2005, Cerling, 1984). A deviation from this pattern was found by Nordt et al. (2006), which proved that vertisols in USA contained pedogenic calcite in areas with up to 1400 mm mean annual precipitation. It has also been found that calcite precipitation in vertisols is favored by the seasonal variations in the moisture content, which leads to variation in the CO_2 concentration in the air in the soil (Breecker et al., 2013).

It appear to be fairly good correlation between the depth at which the calcic horizon occur and the amount of rainfall, where higher amounts of rainfall favors a deeper calcic horizon (Retallack, 2001). This relationship has been used to estimate precipitation patterns in paleosol studies (Kraus, 1999). The mechanism behind this correlation is relatively complex, but has partly to do with dissolution of calcite at shallow depths in humid regions (Retallack, 2005). Figure 2.7 shows this relationship, with data compiled by Retallack (2005) from soils all over the world.

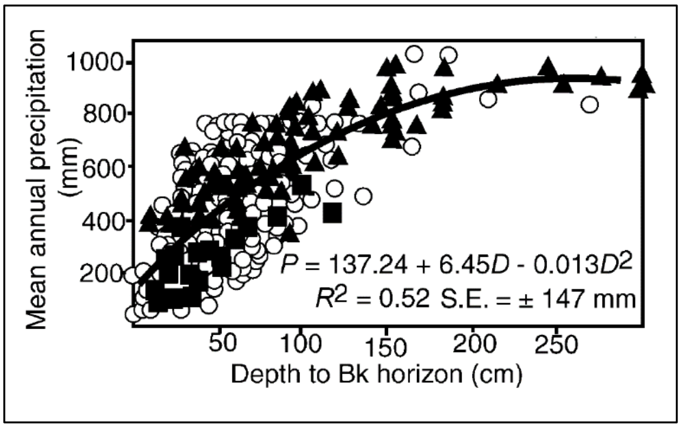

Figure 2.7: *Scatterplot of depth to calcic horizon vs mean annual precipitation, with best fit line. From Retallack (2005)*

Other minerals formed in situ in soils include gypsum and various salts, also commonly found in arid and semi-arid climates. Soils with abundant calcite, gypsum or salt are included in the aridisol order of the USDA soil taxonomy. The Mack et al. (1993) classification system separates between calcite and gypsum dominated soils, into calcisol and gypsisols, respectively.

2.3 Paleosols in sedimentary successions

Most studies of pre-Quaternary paleosols have been on alluvial deposits (e.g. (Kraus, 2002, Müller et al., 2004, Prochnow et al., 2006)), but they have been recognized within eolian, marginal marine, palustrine and deltaic deposits (Arndorff, 1993, Kraus, 1999). Paleosols in a sedimentary succession can be used to reveal the interplay between sedimentation, erosion and soil formation. Soils generally only form on landscapes that is stable and receives minimal sediment input. They will not form where erosion is ongoing, and sedimentation rates are high (Marriott and Wright, 1993).

Paleosols may be vertically stacked in different ways within a sedimentary succession. The stacking pattern and maturity of paleosols is related to the balance between sedimentation, erosion and soil formation. They can be truncated, closely stacked or cumulative depending on how these factors harmonize (Kraus, 1999). Cumulative soils are soils that form where the rate of sedimentation and soil formation balance, and are common within floodplain deposits, where small amount of sediment is deposited during floods, and incorporated in the developing soil profile (Marriott and Wright, 1993, Kraus, 1999).

Polygenetic soils are soils that undergo a change in the moisture regime through their development. Change in soil moisture regime by autocyclic processes can be facilitated by e.g. migration and avulsion of channels (Kraus and Aslan, 1993). Polygenetic paleosols has also been linked to Milancovitch cyclicity driven climate change, and longer term climatic changes in ancient cyclothems (Driese and Ober, 2005, Rosenau et al., 2013b).

Variation in soils also occur on more local scales. Catena is a concept in soil science, which is defined as a group of soil association in an area such as a valley side or a floodplain, where factors such as parent material and climate is uniform (Kraus and Aslan, 1999). The factor creating different soil types in a catena is only relative height and gradient, creating different drainage conditions, resulting in soils with different properties along the catena. The catena term has been adopted by geologists working on paleosols, where soil associations have been linked to position relative to channels in alluvial deposits (Kraus and Aslan, 1999). The main controlling factor in the paleocatena recognized by Kraus and Aslan (1999) is the topographic gradient set up by alluvial ridges on the floodplain, which together with diminishing grain size away from the ridge results in different drainage conditions. As seen in Figure 2.8 this gives rise to variations in color, amount of organic matter in topsoil usually preserved as coal and features such as nodules and slickensides in the paleosols (Kraus and Aslan, 1999).

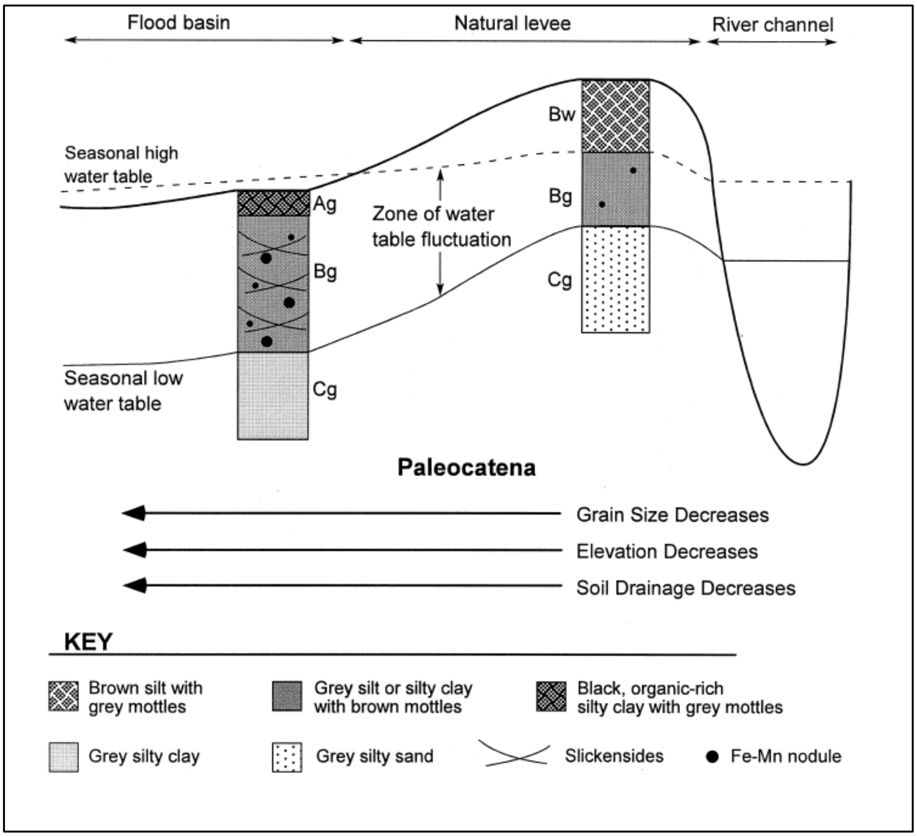

Figure 2.8: The *paleocatena model. Figure from Kraus (1999).*

A similar concept is the pedofacies relationship, introduced by Bown and Kraus (1987). This concept is based on the recognition that the sedimentation rate varies across the floodplain in response to relative position of channels, which are the main sediment supplier (Bown and Kraus, 1987). This results in lateral relations with fine grained mature cumulative paleosols in distal areas, to coarser grained, less mature composite paleosols on the levees, as seen in Figure 2.9. The pedofacies concepts relay on recognizing stages of paleosols maturity, which can then be ascribed to position on the floodplain (Collinson, 1996)

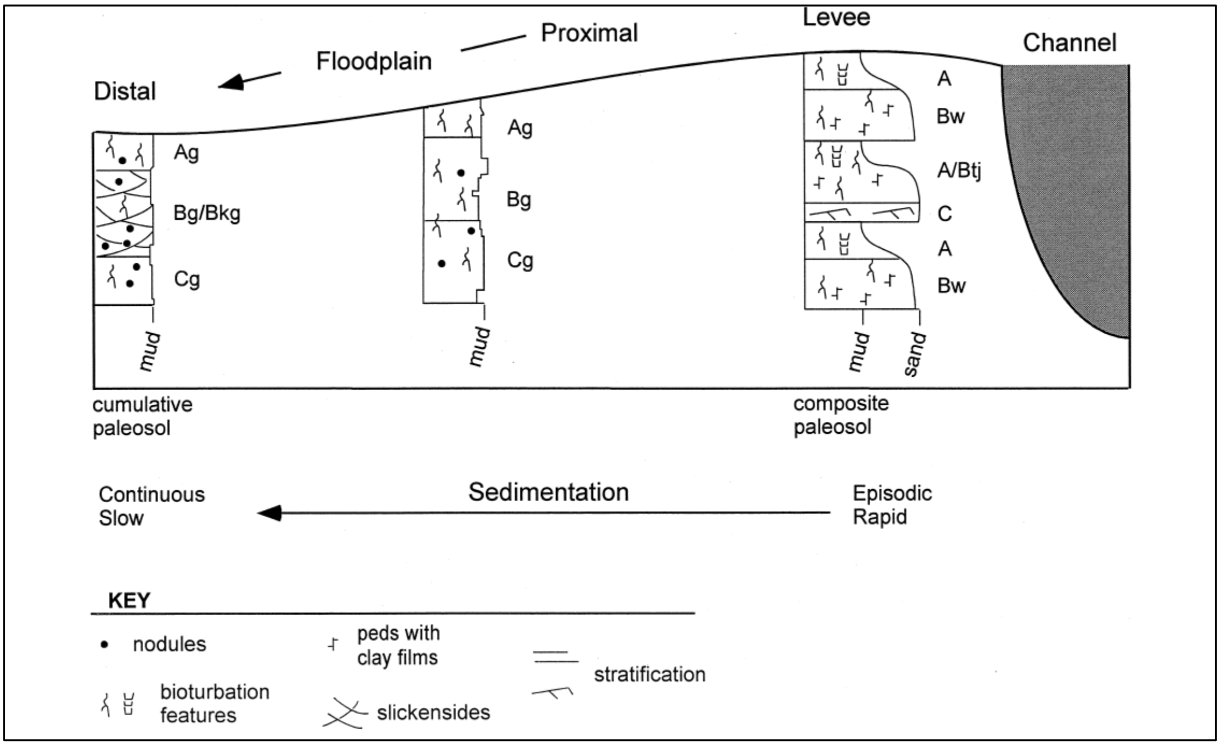

Figure 2.8: The *pedofacies relationship. Figure from Kraus (1999).*

In deltas and terminal distributive fluvial systems the downdip changes in average grain sizes and permeability of sediments, also leads to differing hydrological conditions (Hartley et al., 2013). A common pattern in modern systems is well drained soils in proximal areas, and more poorly drained soils in distal areas dominated by more fine grained deposits (Hartley et al., 2013).

Both the paleocatena and pedofacies concepts are based on fully alluvial deposits (Kraus, 1999), but as delta plain processes, at least in the upper parts, does not differ substantially from alluvial environments (Reading and Collinson, 1996), they should be considered relevant in deltaic environments as well. Similar relationships have also been found in delta plain deposits by Arndorff (1993). In this study on Early Jurassic deposits in Denmark, investigation of the lateral relations between the different paleosols, made it possible to separate between coastal swamps, levee bank paleosols, crevasse splay paleosols and backswamps (Arndorff, 1993).

3 Material

3.1Fieldwork Localities

Fieldwork was conducted summer 2014, and organized by NPD and SINTEF Petroleum Research. The vessel MS *Kvitbjørn* was used. The logging was done in cooperation with fellow master student at NTNU/UNIS, Turid Haugen. The De Geerdalen Formation was logged on Edgeøya and Hopen. Due to limited amount of time spent on each location, the logs are not complete through the whole formation.

Edgeøya is the third largest island of the Svalbard archipelago, and is situated to the east of Spitsbergen. The island mainly exposes sediments from the Triassic Sassendalen and Kapp Toscana groups (Flood et al., 1971, Vigran et al., 2014). Only the lower parts of the De Geerdalen Formation was logged here, as the upper parts are not exposed on the island (Vigran et al., 2014). On Edgeøya the localities Negerfjellet, Tjuvfjordhorga and Blanknuten was visited and logged with short logs. Location of these areas are shown in Figure 3.1.

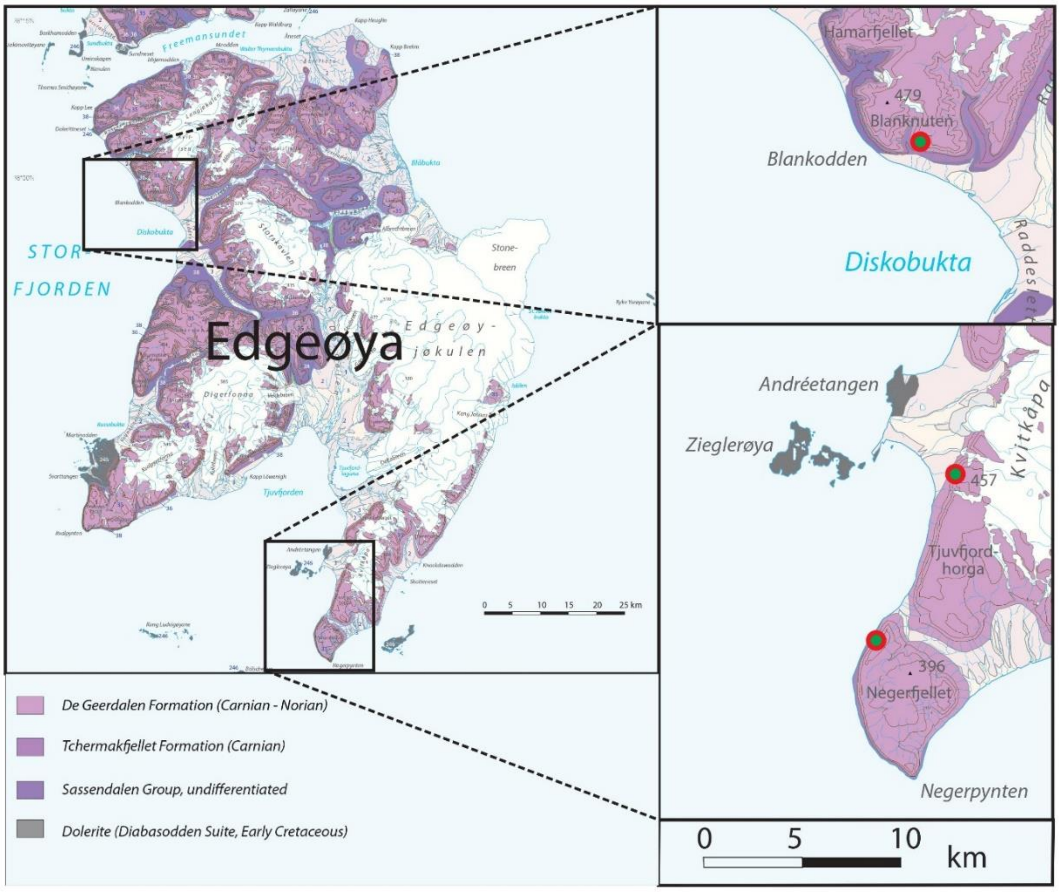

Figure 3.1: *Map showing logged localities on Edgeøya. Map from Dallmann (2015).*

A short visit to the island Hopen was also made. Hopen is a 32 km long island situated far to the southeast on Svalbard. The sedimentary strata at Hopen consists exclusively of the Kapp Toscana Group, represented by the De Geerdalen, Flatsalen and Svenskøya formations (Mørk et al., 2013). Approximately 165 m of the upper parts of the De Geerdalen Formation is exposed on the island, but the formation has been determined to be around 650 m in total thickness on Hopen from well data (Mørk et al., 2013, Lord et al., 2014a). The parts of the De Geerdalen Formation exposed on Hopen has been determined to be of Late Carnian age, possibly entering the earliest Norian (Paterson and Mangerud, 2015). Short logs along a paleosol were made to investigate changes and lateral extent of the horizon. The logging started at Russevika, and continued along the beach, at the eastern side of the base of Johan Hjortfjellet mountain, as seen on Figure 3.2.

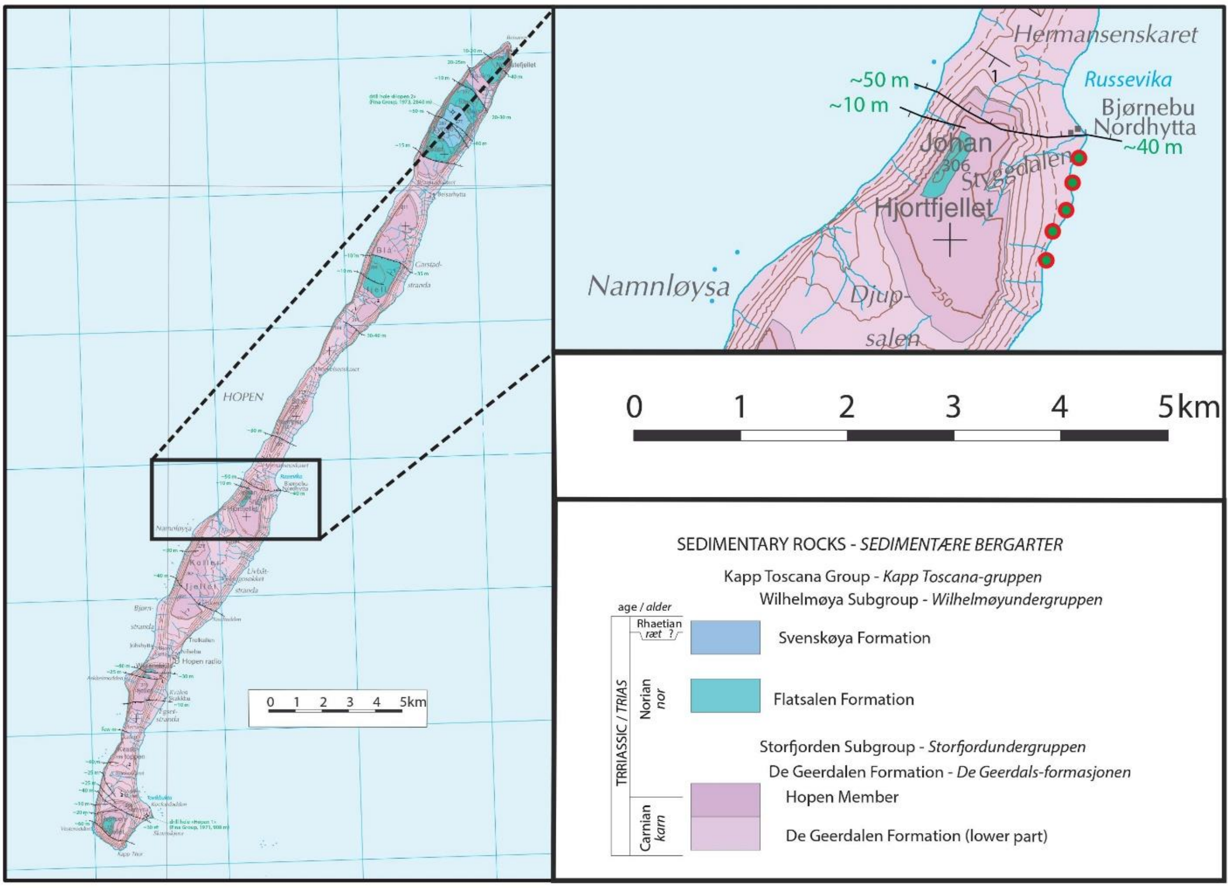

Figure 3.2: *Logged localities on Hopen. Map from Mørk et al. (2013)*

3.2 Core 7430/07-U1

Core 7430/07-U01 was drilled in 1988 by SINTEF Petroleum Research (former IKU) (Vigran et al., 2014). The core penetrated in total of 109,88 m of the Snadd Formation, and are directly overlain by Quaternary deposits (Vigran et al., 2014). The well was drilled on the Bjarmeland Platform, which is defined as the area bounded in the south and southeast by the Hammerfest and Nordkapp basins, and the Sentralbanken and Gardarbanken to the north (Gabrielsen et al., 1990). Figure 3.3 shows the position of the well.

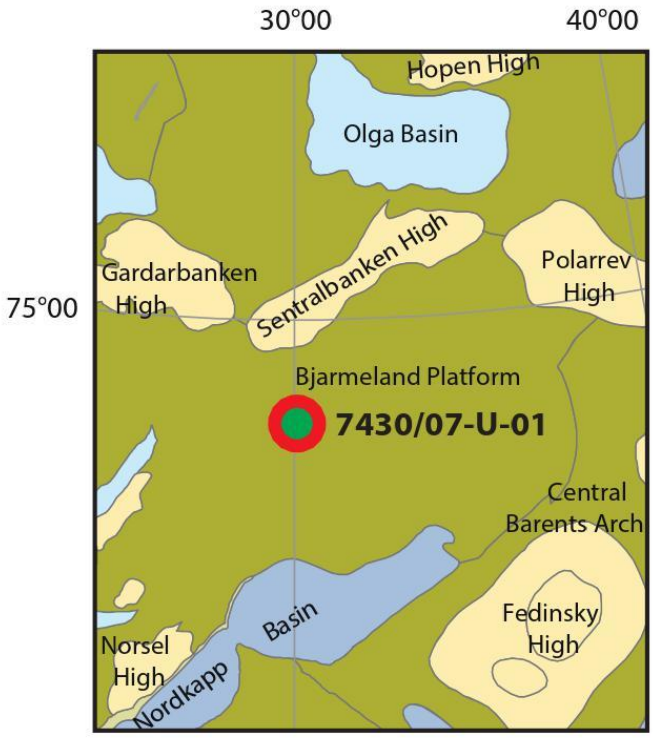

Figure 3.3: *Location of well 7430/07-U-01. Map from Ryseth (2014).*

Only logs from the paleosol horizons are presented, for the rest of the sediments the interpretations of Stensland (2012) will be referred to. The purpose of including the core was to be able to investigate some of the details preserved in the paleosols, which can be difficult to see in weathered and scree covered exposures in the field. In addition, the level of details in a core cut makes it possible to identify poorly developed paleosols, which are also likely to be hard to identify in outcrops. This allows for detailed interpretation of pedogenic processes, interplay between sedimentation, erosion and soil formation, and classification and comparison with modern soils.

4 Methods

4.1 Logging

Logging was done both in the field and of core material. Sections were measured with meter stick, and grain sizes were estimated with a standard grain size sheet. Photos have been taken using standard digital camera, though a few have been taken using a multirotor drone by Alexey Deryabin (see Deryabin and Riis (2015)). Hydrochloric acid was used to determine the presence of calcite. A standard GPS was used to determine elevation and gather the coordinates of each locality. The GPS data is included in the appendix. The handwritten logs from the field and the core logging have been redrawn using Adobe Illustrator digital drawing program. The logs from Edgeøya have been redrawn in 1:500 scale, while the logs from the core, included in the appendix is drawn in 1:100 scale.

The interpretation of the acquired data is based on facies and facies associations. The term facies or rock facies is widely used by geologists within several disciplines, and dates as far back to the 17th century, while modern scientific use of the phrase was introduced in the 19th century (Boggs, 2006). The exact definition of facies has been much debated by sedimentologists (Middleton, 1973), but can generally be said to be “a body of rock with specified characteristics”, as defined by Reading and Levell (1996). When these characteristics can be linked to specific processes, they can aid in interpreting the depositional environment. To gain a more effective environmental interpretation from facies, they are often grouped together with other facies believed to be genetically or environmentally related (Reading and Levell, 1996). A group of related facies is usually termed facies associations (Boggs, 2006).

The facies defined within the present study is mainly based on lithology, sedimentary structures, grade of bioturbation and organic material content. Facies and facies associations has been used both to separate between different types of paleosols, and to relate the paleosols to depositional elements.

4.2 Recognition and classification of paleosols

Paleosols can be recognized in a number of ways, and usually a combination of different features is necessary. Most commonly this include coal and organic matter, roots, pedogenic slickensides, bioturbation and destratification, horizonation and color variations, and mineral precipitates. Preserved roots in growth position is considered one of the best evidence of a fossil soil (Retallack, 1988) . The reason for this is that subaerial exposure, at least partially is necessary for vascular plants grow (Retallack, 2001). Still, the presence of roots alone, does not indicate prolonged exposure or weathering of any significant degree. In addition many paleosols do not have preserved roots, as the organic material constituting the root material can be degraded (Retallack, 2001). In general, well preserved roots is not present in most paleosols and should therefore not be considered a necessary evidence for a paleosol (Wright, 1992a). Another effective way of recognizing paleosols is the identification of horizons, often appearing as vertical color changes (Wright, 1992a).

Where other diagnostic features have not been found, color variations have been taken as a evidence for a paleosol. Coal is not necessarily evidence of a paleosol, but where the color of the sediments underlying the coal shows noticeable color changes towards the coal, or are highly unconsolidated they are interpreted as a paleosol. Where coal or coal shale has been found without a underlying paleosol, they have not been interpreted as a paleosol. For more thorough reviews on the principles of recognizing paleosols see Retallack (1988) and Wright (1992a).

None of the classification systems previously mentioned in has been rigorously applied, but an attempt to place some of the paleosols within the USDA soil taxonomy soil orders, and into the Mack et al. (1993) classification system will be made where diagnostic features are observed. The purpose of this is to enable easy comparison with both modern soils, and with other studies of paleosols from the literature.

4.3 Optical microscopy

A few thin section where made from selected parts of core 7430/07-U1. The purpose of this was partly to be able to calibrate some of the grain size measurements, and to investigate some of the micro scale features of the paleosols. This only forms a small part of this thesis, and results will only be briefly mentioned in Section 5.2.

4.4 TOC

In order to determine the amount of organic carbon in the coal, total organic carbon (TOC) analysis of selected coal samples from core 7430/07-U1 was done. The samples were analyzed by Robertson Geolab Nor AS in Trondheim, using a LECO CS-244 carbon and sulfur determinator. The procedure involves crushing the sample, and removing any inorganic carbon from carbonate minerals by soaking it in hydrochloric acid. The sample is then washed and dried in a vacuum. After weighing, the sample is combusted together with copper. Through infrared detection of the CO_2 produced by the combustion, the amount of carbon in the sample can be estimated. Calibration is done by measuring TOC of a known sample.

5 Results and interpretation

The following chapter will deal with the results obtained from field work, and material from core logging. The logs will be presented, together with the interpretation. The results from field work on the De Geerdalen Formation, and core logging of Snadd Formation will be treated separately. The data from the De Geerdalen considers the whole depositional environment, while the data from the Snadd Formation mostly considers the paleosols and their properties. In the discussion the results from the various parts will be synthesized and compared.

5.1 De Geerdalen Formation

The De Geerdalen Formation has been logged on Edgeøya and Hopen. Based on the logs made in the field, a description of facies and facies associations were made. Since the topic of this thesis is paleosols, the focus will be the occurrence of the paleosols and their associated facies, but in order to understand how and where they have formed, a consideration of the overall depositional environment is necessary. The facies and facies association will therefore act as a framework to be able to relate the paleosols to the environment they formed in, and what mechanism that allowed for soil formation to take place. Some of the sections presented here have been studied previously. The facies and their descriptions can therefore be seen to be similar to the work of Rød et al. (2014). The facies and facies associations will be presented first, thereafter the localities together with the field logs. As most of the data are from Edgeøya, some of the facies presented are only identified here. Only short intervals of the De Geerdalen Formation have been logged. On Hopen only a few vertical meters have been investigated. The results and interpretations presented herein should therefore not be considered as a review of the development of the De Geerdalen Formation depositional system as a whole.

5.1.1 Facies

A total of 14 facies have been identified in the De Geerdalen Formation. The criteria and parameters that have been used in grouping and description of the facies are explained in Section 4.1.

Facies A, Trough cross-stratified sandstone:

Description

Units are often erosive based, and consists mostly of fine to medium sandstone of grey color, see Figure 5.1A. This facies has been identified on both Edgeøya and Hopen. Units in this facies vary in thickness from 1 to 3 m. Thin mud drapes can be seen in places. Mud flakes are abundant especially in the lower parts of units, but often occur scattered in several intervals. No bioturbation has been observed within this facies. Some plant remains can be found between sets, and a large tree trunk has been found encased in this facies on Hopen. The sandstones in facies B often consist of more clean, quartz rich sand than the other sandstone facies. The facies is associated with horizontal stratified and asymmetric ripple laminated sandstone facies.

Discussion and interpretation

Through cross-stratification is formed by migrating 3D dunes and ripples, and is common in fluvial and paralic to shallow marine environments, but it can form anywhere where there is unidirectional currents capable of producing 3D dunes and ripples (Boggs, 2006, Reading and Collinson, 1996). This facies is here interpreted to be the result of unidirectional currents that locally were strong enough to transport large sized trees. Mud drapes may be indicative of some tidal influence, or at least variation in flow regime.

Facies B, Planar stratified sandstone

Description

This facies occur in very fine to fine sandstones, with thicknesses ranging from 0.5 to 2-3 meters, and stratification varying from thinly bedded to laminated. It is found on Edgeøya, and has roughly horizontal and parallel bedding surfaces, see Figure 5.1 B. Beds often have a sharp lower boundaries and in places show loading structures. In addition small amounts of organic material between laminas and cm sized mud clasts are found in places. Bioturbation is sparse, with only a few vertical burrows. This facies occur dominantly in sand dominated intervals, but can also be found as thin beds within more mud dominated parts of the succession as the top of small coarsening upwards units. In sand dominated intervals this facies is closely related with through cross-stratified and asymmetric ripple laminated facies.

Discussion and interpretation

Planar stratification is not a unique feature of any environments as it may form both by 3settling of particles from suspension, and transport by traction of coarser material (Boggs, 2006). The often close association with cross-stratified units implies that the planar stratification in this facies formed in response to currents. Low amount of bioturbation suggests relatively rapid burial.

Facies C, Asymmetric ripple laminated and tabular cross-stratified sandstone:

Description

This facies is comprised of fine to medium, grey sand. Bioturbation is absent in this facies, but mud clasts occasionally occur. This facies is found in close association with plane parallel and through cross-stratified sandstone, and are often found above through cross-stratified sandstone in upwards fining units on Edgeøya. Figure 5.1C shows asymmetrical ripples from Edgeøya.

Discussion and interpretation

Asymmetrical ripples are formed in lower flow regime by unidirectional currents, usually in shallow waters, but can also form in eolian environments (Boggs, 2006). They can be formed in fluvial environments, or in shallow marine environments where rip currents and breaking waves sets up unidirectional currents (Reading and Collinson, 1996). Tabular cross-stratification is formed by migrating 2D ripples and dunes (Boggs, 2006).

This facies is here interpreted to have been formed by unidirectional currents in shallow waters.

Facies D, Symmetrical rippled and symmetrical rippled laminated sandstone:

Description

Facies D is represented with very fine to fine sand with partly planar, often wavy parallel bedding. Some bioturbation is found in this facies. Ripple crests have been observed to be both straight and bifurcating. This facies is often found above intervals dominated by hummocky cross-stratification and is identified on both Edgeøya and Hopen. Figure 5.1D shows an example of this facies from Edgeøya.

Discussion and interpretation

Symmetrical ripples form by oscillatory flow by wave action, and are usually found at shallow depths above normal wave base in shallow marine environments, where wave height and fetch are largest (Reading and Collinson, 1996). It may also form in lakes and lagoons, and have also been reported from deep sea environment (Boggs, 2006). The often close association with facies F, hummocky and swaley cross-stratified sandstone suggests this facies formed in a shallow marine environment above normal wave base.

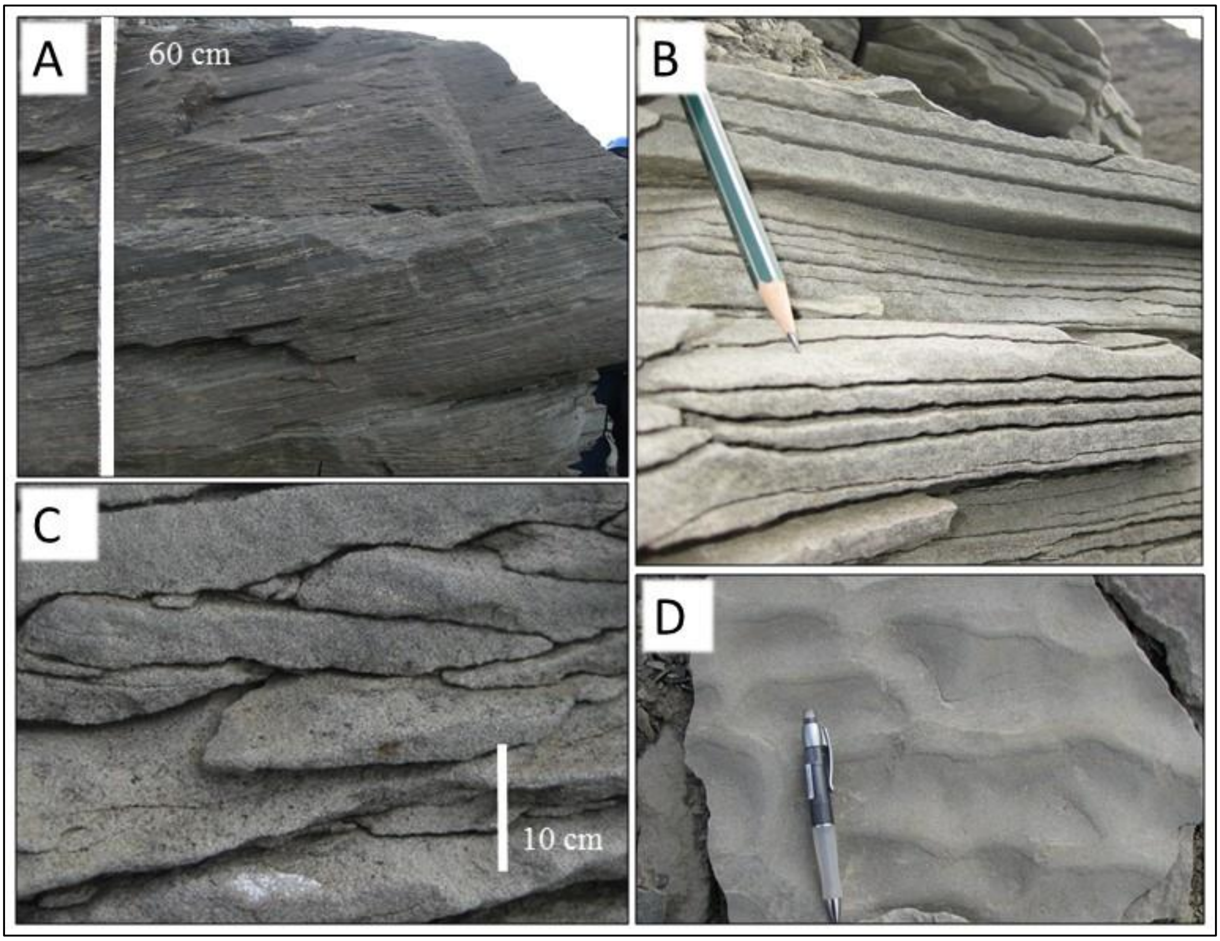

Figure 5.1: A) *Facies A, through cross-stratified sandstone. From Negerfjellet, Edgeøya.* **B)** *Facies B, planar stratified sandstone, from Tjuvfjordhorga, Edgeøya.* **C)** *Facies C, asymmetric ripple laminated and tabular cross-stratified sandstone. Here showing asymmetrical ripples in medium sand, from Negerfjellet, Edgeøya.* **D)** *Facies D, symmetrical rippled and symmetrical ripple laminated sandstone, from Tjuvfjordhorga, Edgeøya. Plan view of symmetrical ripples with slightly bifurcating ripple crests.*

Facies E, Massive structureless sandstone:

Description

These sandstones have grain size from very fine to fine sand. Facies E shows lack of any visible primary sedimentary structures, and can often be seen to be highly fractured, as shown in Figure 5.2A. Bioturbation, in places extensive and mud clasts are occasionally found. Some of the massive sandstones occur as parts of sandstone bodies with defined stratification, whereas others occur as distinct beds isolated within mudstones.

Discussion and interpretation

Units with massive or structureless bedding usually reveals some faint structures if studied carefully, or by etching and staining the rock (Boggs, 2006). Mechanisms for depositing primary structureless sand includes rapid deposition from suspension and deposition from gravity flows with high sediment concentration (Boggs, 2006). Structureless units can also be created by obliteration of primary stratification by intense bioturbation, and liquefaction by sudden events (Boggs, 2006). In fluvial environments massive sandstones can form part of channel deposits (Jones and Rust, 1983, Collinson, 1996).

As no diagnostic features have been observed in this facies, the sandstones may have been deposited by various mechanisms. Mud flakes may indicate deposition in a high-energy setting. Some of the sandstones of this facies have probably lost primary structures due to bioturbation.

Facies F, Hummocky and swaley cross-stratified sandstone:

Description

This facies include sandstones with very fine to fine sand, often silty. Both hummocky (see Figure 5.2B) and swaley cross-stratification are identified. Thickness of units vary from few dm up to 3 m. In places the sandstones have erosive base with mud clasts and consists of several amalgamated beds. Swaley cross-stratified sandstones are generally thicker and more laterally persistent than hummocky cross-stratified sands. In places soft sediment deformation can be seen. Bioturbation is common. This facies is often seen grading into facies D, symmetrical rippled sand, or facies B, planar stratified sandstone.

Discussion and interpretation

Hummocky cross stratification is believed to be formed by flow with both unidirectional and oscillatory components, near storm wave base during storm events (Boggs, 2006, Dumas and Arnott, 2006, Reading and Collinson, 1996). Swaley cross-stratification is hypothesized to form closer to normal wave base, where aggradation rates are lower, leading to preferential preservation of swales (Dumas and Arnott, 2006).

This facies is here interpreted to represent storm deposits, where hummocky cross-stratified sandstones have been deposited between normal and storm wave base, and swaley cross-stratified sandstones closer to normal wave base.

Facies G, Low angle cross-stratified sandstone:

Description

This facies consist of very fine to fine, grey sand, with units from 0,5 to 2 m thick. It is arranged in wedge shaped sets with varying thickness and dip angles, as seen in Figure 5.2C. The facies is found together with horizontal stratified and symmetrical ripple laminated sandstones, and is identified on Edgeøya.

Discussion and interpretation

Low angle cross-stratification with seaward dip can be formed by swash and backwash flow in the foreshore on beaches (Boggs, 2006).

The close association with facies D, symmetrical rippled and symmetrical rippled laminated sandstone, suggests this facies was deposited in a shallow marine environment.

Facies H, Carbonate cemented sandstone:

Description

This facies represents both concretions up to 3-4 m in diameter, and very fine sandstones with thickness up to 2 m, that are brown, and well cemented by carbonate. Cone-in-cone structures are found in some of the carbonate concretions, see Figure 5.2D. Facies H occur within less cemented sand, but are mostly found as distinct beds within mudstones.

Discussion and interpretation

Recent investigations of carbonate cemented surfaces and concretions in the Tschermakfjellet and De Geerdalen formations have been done by Tugarova and Fedyaevsky (2014), who showed that these features are the result of biochemical precipitation of carbonates by microorganisms, during very early diagenesis in the shallow marine environment. Maher et al. (in press) presents an alternative interpretation, where they argue that calcite growth occurred during shallow faulting.

This facies could represents periods of low sediment supply, allowing for extensive early cementation. If the cementation is related to shallow faulting as suggested by Maher et al. (2015), this facies probably holds no information about the depositional environment.

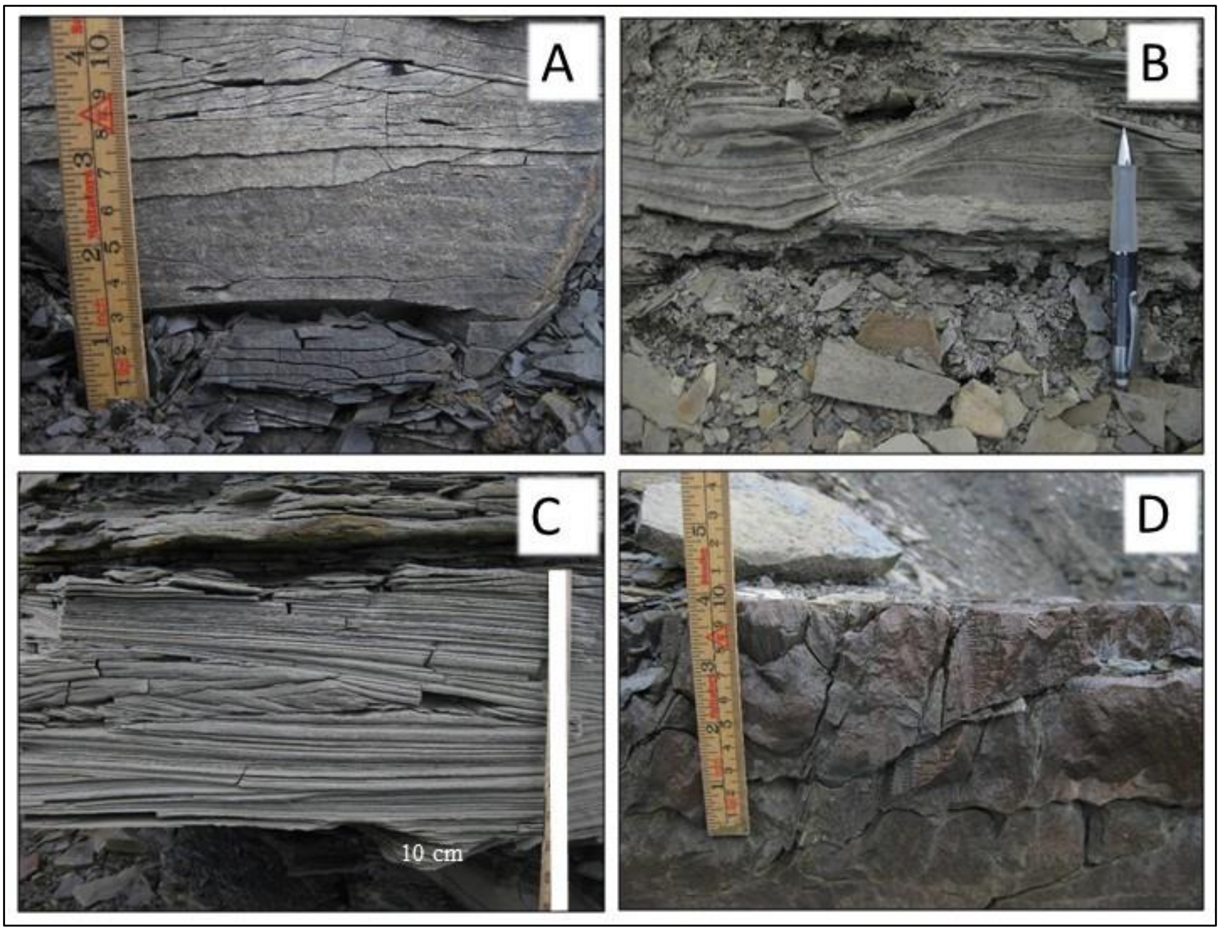

Figure 5.2: A) *Facies E, massive structureless sandstone, from Blanknuten, Edgeøya.* **B)** *Facies F, hummocky and swaley cross-stratified sandstone, here showing hummocky cross-stratification, from Negerfjellet, Edgeøya.* **C)** *Facies G, low angle cross-stratification arranged in wedge shaped sets, from Tjuvfjordhorga, Edgeøya.* **D)** *Cone-in-cone structure in facies H, carbonate cemented sandstone. Picture taken by Turid Haugen.*

Facies I, Coquina:

Description

Facies I include carbonate cemented brown and grey, massive beds, consisting chiefly of bivalve shells that are usually well preserved. The bivalves are up to 1 cm in diameter, see Figure 5.3A. Units of this facies are up to 1m thick, show laterally continuity over 100s of meters, and occur as isolated beds within mudstones. Facies I is found on Edgeøya.

Discussion and interpretation

The facies is interpreted as shallow marine shell banks, accumulated by waves or currents. The well preserved shells could indicate a low-energy setting. It may possibly also represent periods with reduced input of sediments.

Facies J, Mudstone:

Description

This facies consists of mudstones with varying silt and clay content, forming units up to 15 m thick. It is usually covered, or partly covered by scree, as seen in Figure 5.3B. Colors are mostly grey, and where any structures are visible, plane parallel lamination can be seen. In addition, few cm thin, very fine sandstone beds can be found within this facies in places. No bioturbation is observed in the mudstones.

Discussion and interpretation

Mudstones can be deposited in a number of low-energy environments, both marine and non-marine (Boggs, 2006).

The facies is here interpreted to represent deposition during low energy, by settling from suspension in standing water. As mudstones can be an ambiguous environmental indicator, this facies has mainly been interpreted based on the associated facies, such as paleosols and hummocky cross-stratified sandstones.

Facies K, Interbedded sandstone and mudstone:

Description

This facies consists of clay to very fine sand arranged in units from 1 to 5 meters thick, showing a coarsening upwards trend. Bioturbation is common and in parts extensive, with both horizontal and vertical burrows. Color of sandstones are mostly grey, but are seen to be brown in more cemented parts. Lenticular, wavy and flaser bedding are visible in places, as can be seen in Figure 5.3C. It is identified on both Edgeøya and Hopen.

Discussion and interpretation

Alternating deposition of mud and sand can be facilitated by periodic fluctuations in hydraulic conditions, often linked to tidal or fluvial currents (Dalrymple and Choi, 2007). Conditions favorable for sand deposition and preservation tend to produce flaser bedding, while wavy and lenticular bedding results from more mud prone conditions (Boggs, 2006).

Facies K is Interpreted to be deposited under varying hydrological conditions, most likely caused by tidal currents, where sand has accumulated during periods of low energy unidirectional flow, and the mud represent more stagnant conditions.

Facies L, Organic rich shale/coal-shale:

Description:

This facies occur both on Edgeøya and Hopen and consists of dark grey mud, as seen on Figure 5.3D. It is sometimes found within grey shales with a gradual transition, but are also found associated with paleosols, where they occur on top. Small lenses of coal occur in places.

Discussion and interpretation

Distinction between organic rich shale, coal shale and coal is depending on the fraction of organic material (Diessel, 1992). Organic rich shales can be deposited in swamps and marshes with high clastic sediment input, but can also be deposited from suspension in low energy environments, such as lagoons and deep marine basins, where high organic productivity close to the site of deposition gives a high organic content (Boggs, 2006).

It is here interpreted to represent both autochthonous organic accumulations in swamps and marshes, and detrital accumulations of organic rich mud.

Facies M, Paleosols

Description

Paleosols are found on both Hopen and Edgeøya. The thickness of the paleosols vary from 50 cm to approximately 2 m. Colors are varying and mainly found to be brown to reddish brown, grey and yellow. Roots with varying thickness and length can be seen in some of the paleosols. Nodules of siderite with a few cm diameter are found on Hopen. The paleosols are usually associated to layers of coal and organic rich mud, which then occur on top of the paleosols. Several thin layers of coal also occur within some of the soil profiles, as seen in Figure 5.4B. The paleosols occur both within clay rich shale, and on top of more sand rich units. In a few of the paleosols, a bleached, light yellow colored horizon around 50 cm thick can be seen. Elsewhere they mostly show grey and reddish brown dm thick horizons.

Discussion and interpretation

Paleosols are surfaces marking subaerial exposure, representing a breach in sedimentation or periods with very low sediment supply on a subaerial exposed surface (Kraus, 1999), as explained in Section 2. Thin coal seams within the paleosols suggests multiple episodes of sedimentation, plant colonization and subsequent soil formation. This facies will be further discussed in Section 5.1.3.

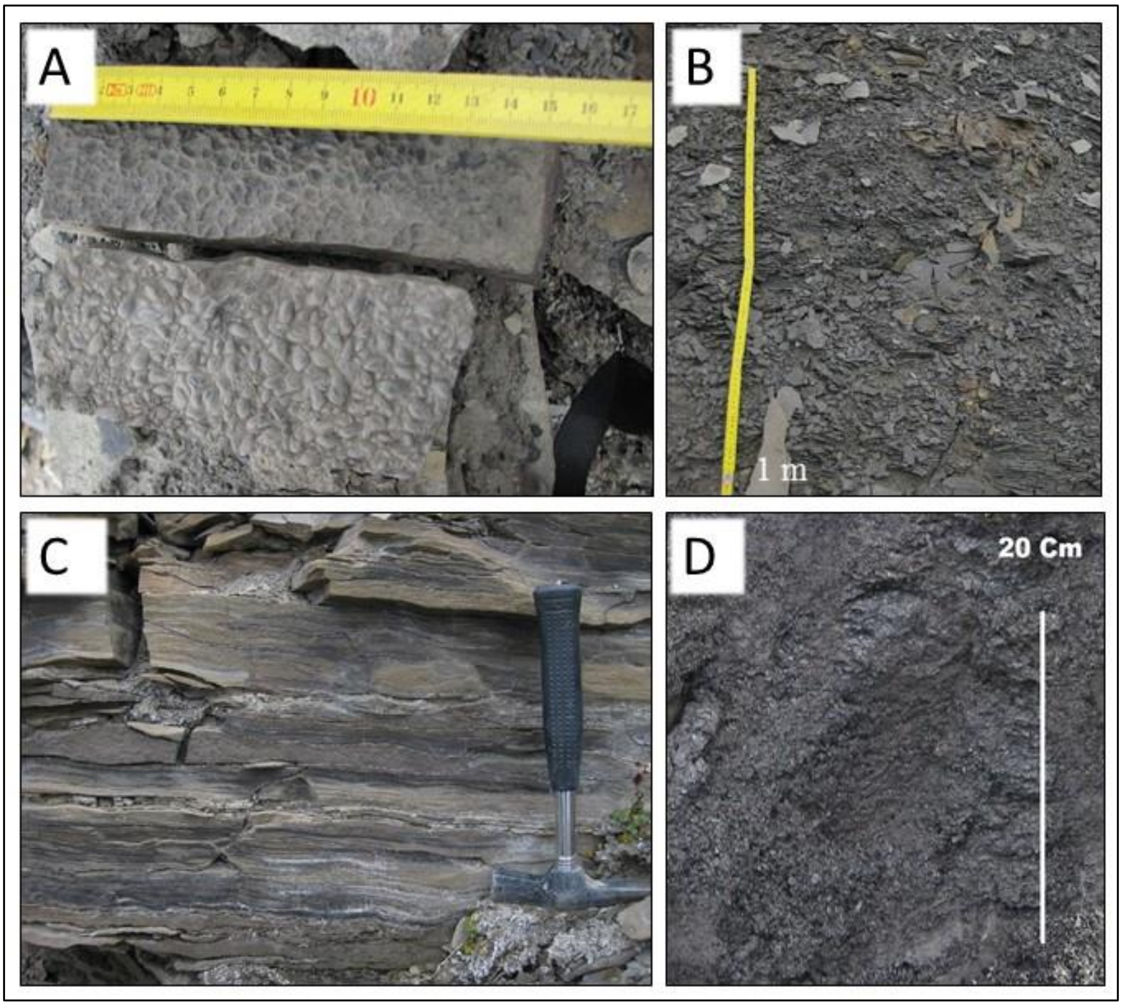

Figure 5.3: **A)** *Facies I, coquina, showing well preserved bivalve shells, from Tjuvfjordhorga, Edgeøya*. **B)** *Facies J, mudstone, partly covered. Picture from Negerfjellet, Edgeøya.* **C)** *Facies K, interbedded sandstone and mudstone, from Negerfjellet.* **D)** *Facies L, organic rich shale/coal shale, from Blanknuten.*

Facies N, Coal:

Description

The coal occur as thin layers up to 50 cm thick, but usually thinner (see Figure 5.4C). It also occur as cm sized lenses within organic rich mudstones. The coal seams generally show varying lateral continuity. On Blanknuten, seemingly in-situ tree trunks can be found directly above a coal layer, seen in Figure 5.4. It is mostly found at top of paleosols, but are also found as thin lenses and layers where there are no signs of paleosols below.

Discussion and interpretation

Peat is the precursor of coal and forms where hydrological conditions are favorable for accumulation and preservation of organic material, i.e. more or less permanent waterlogged (McCabe, 1984). Where coal occur with paleosols the seams are interpreted to be part of a O horizon of the soil as explained in Section 2.2.2. As the coals are seen to be relatively thin, they have likely developed in a short amount of time. This may be due to unfavorable conditions for peat accumulation. The fact that the peat forming environments was short lived, could be explained by limitations exerted by the sedimentary environment (Nemec, 1992), or a unfavorable climate.

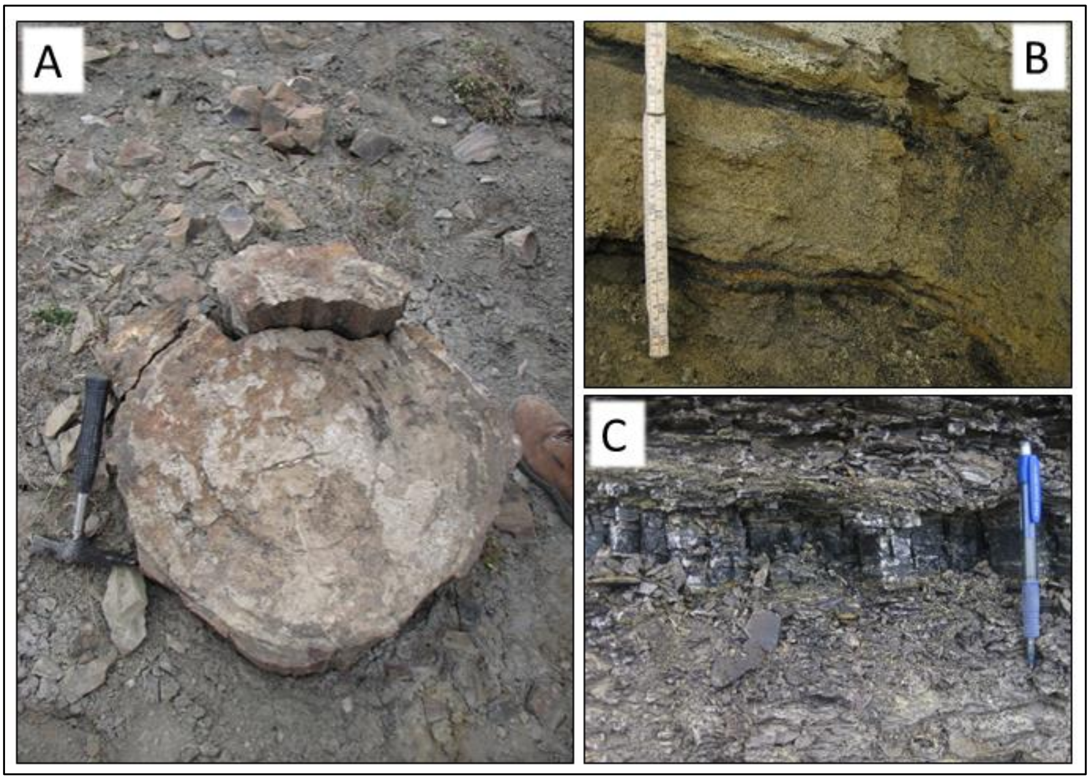

Figure 5.4: A) *Three trunk approximately 50 cm in diameter found above facies N, coal at Blanknuten, Edgeøya.* **B***) Example of a brown colored paleosol, with thin coal seams within, from Blanknuten, Edgeøya.* **C)** *Thin coal seam from Hopen.*

5.1.2 Facies associations

The 14 different facies have been grouped together into 4 facies associations. They are presented in Table 5.1with descriptions and interpretations below.

Table 5.1: *Facies associations in the De Geerdalen Formation.*

	Facies associations	Included facies
	FA1: Shallow marine to shoreface	D,F,G,H,I,J,K
	FA2: Fluvial channel	A,C,E
	FA3: Delta front	A,B,C
	FA4: Delta plain interdistributary areas	B,J,K,L,M,N

FA1, Shallow marine to shoreface:

Description

These deposits mainly comprise mudstones with thin very fine to fine grained sands, showing hummocky, swaley and low angle cross-stratification, planar stratification and symmetrical wave ripples. The sandstones in this interval consist both of decimeter thick single bed, with low lateral continuity, and thicker, partly amalgamated beds showing more lateral persistency. Siderite cemented bed and intervals, and coquina beds are also included in this facies association. Bioturbation is common, especially within parts dominated by hummocky cross-stratification and interbedded sand and mudstone. Some beds show no visible structures and appear to be extensively bioturbated. This facies association is identified on Edgeøya and can be considered equivalent to the "marine shelf to shoreface facies association" of (Rød et al., 2014). Figure 5.5 shows an example of this facies association.

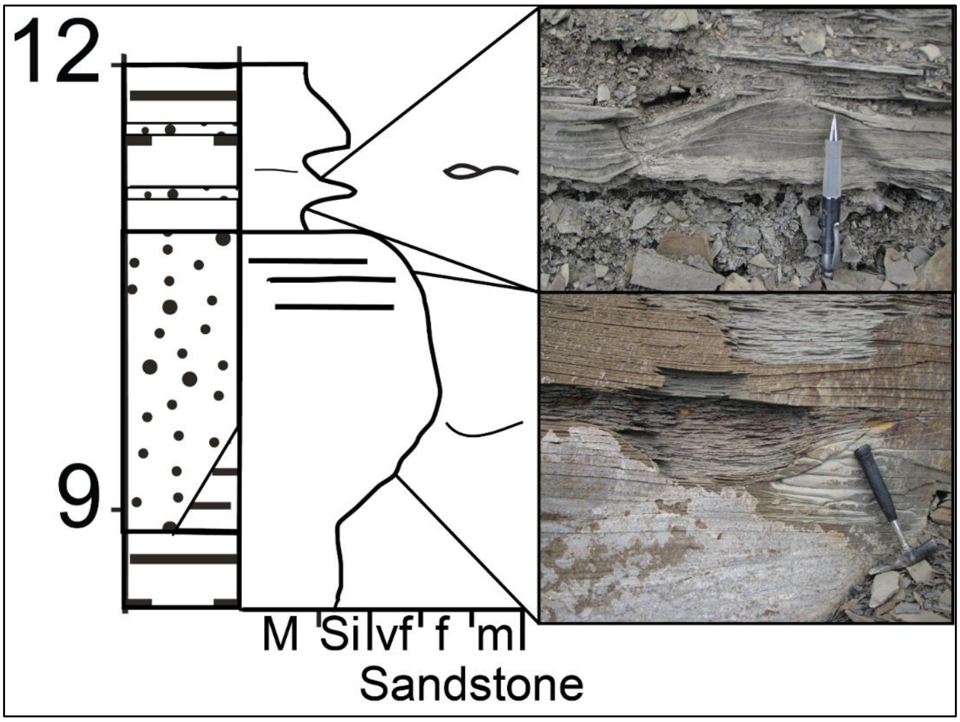

Figure 5.5: *FA1 shallow marine to shoreface. Showing a sandstone bed with swaley cross-stratification, passing upwards to planar stratification. Overlain by thin sandstone beds showing hummocky cross-stratification, encased in mudstone. From the TF 2 log.*

Discussion and interpretation

In shallow marine and shoreface environments wave and tidal energy and depth relative to wave base is the most important factors controlling the facies distribution (Reading and Collinson, 1996). The various facies in this facies association have been attributed to sub-environments according to the shoreline zonation concepts, where energy generally increases towards shallower depths relative to wave base (Reading and Collinson, 1996).

In this facies association the hummocky and swaley cross-stratified sandstones are interpreted to be deposited by storm events in the offshore transition zone. The swaley cross-stratified sandstone and the amalgamated beds are suggested to be the result of either a more proximal position towards normal wave base, or more intense storm events. Symmetrical rippled sandstones are interpreted to represent lower shoreface deposits. Planar stratified and low angle cross-stratified sands are believed to represent more proximal upper shoreface deposits. The mudstones are interpreted to represent fair-weather deposits in the offshore transition, while thicker mudstone intervals might be deposited in a shallow shelf environment below storm wave base. This facies association can represent wave influenced interdistributary bays between delta lobes, or more open marine conditions adjacent to the main delta.

FA2, Fluvial channel:

Description

The fluvial channel deposits consists of erosive based, fine to coarse grained sandstone bodies that fines upward and pinches out laterally. These sandstones usually contain less mud than the sandstones of FA 1 and FA 3. They occur within delta front and delta plain deposits, and have varying geometries. Large scale through cross-stratification with mud flakes is the dominant structure in their lower parts, with typically more planar stratification and asymmetric ripples towards the top. No clear tidal signatures or bioturbation is observed in this facies association. Fluvial channels have been identified on both Edgeøya and Hopen. On Edgeøya this facies association includes a 10 m thick and approximately 200 m wide sandstone body at Tjuvfjordhorga, and an 8 m thick sandstone with unknown width at Negerfjellet. Figure 5.6 shows an example of this facies association from Hopen.

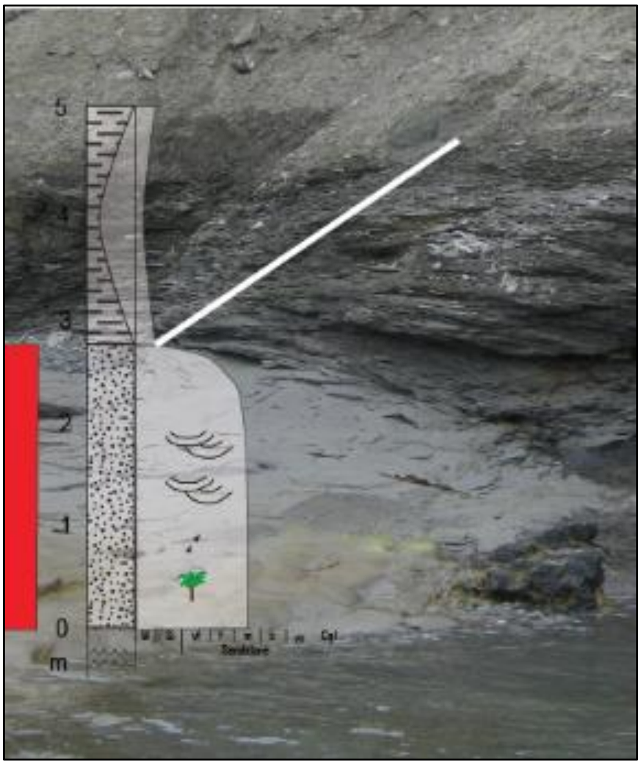

Figure 5.6: *Example of FA2, fluvial channel, from log 2, Hopen.*

Discussion and interpretation

As mentioned before the channels identified on Hopen have previously been interpreted and described in detail by Lord et al. (2014b) and Klausen and Mørk (2014), where they have been divided into fluvial distributary, fluvial trunk, tidal and estuarine channels. Channels interpreted as distributary channels have also been reported from Edgeøya (Rød et al., 2014). The channels observed on Edgeøya in this study shows geometries that are comparable with the fluvial distributary channels of Lord et al. (2014b) from Hopen. As no evidence indicative of tidal influence is observed, the channels are interpreted to be of fluvial origin.

FA3, Delta front

Description

This facies association is dominated by lateral extensive, fine grained, wave rippled and cross-stratified sandstones, with thicknesses from 3 to 6 m. Marine bioturbation, such as

Rhizocorallium occur in places. Some wave ripples are seen, often with mud drapes, but unidirectional current indicators such as through and tabular cross-stratification are dominant. The sandstones can be seen to display an upwards coarsening trend. This facies association occur on both Hopen and Edgeøya. Figure 5.7 shows an example of this facies association from Tjuvfjordhorga.

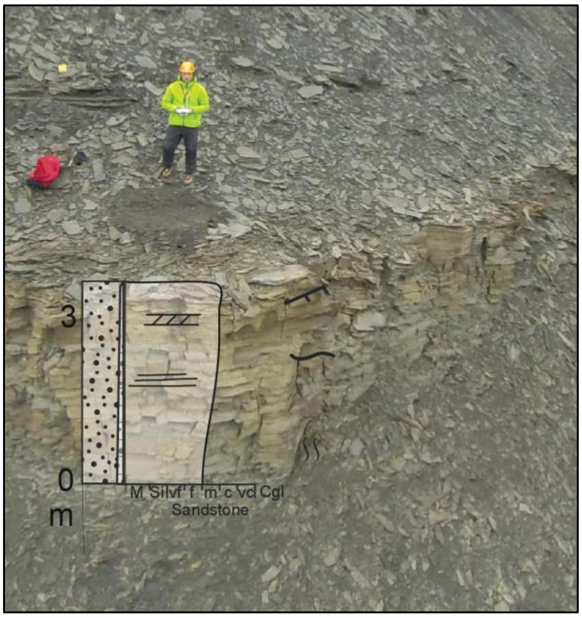

Figure 5.7: *Example of FA3, delta front, from the TF 1 log. Picture taken by Alexy Deryabin using a multirotor drone.*

Discussion and interpretation

The delta front environment is characterized by a high rate of sedimentation as fluvial currents is deaccelerated, loses competence and deposits sediments (Reading and Collinson, 1996). Coarsening-upward sequences is typical in delta front environments of prograding systems (Reading and Collinson, 1996). Mørk et al. (1982) and Rød et al. (2014)interpreted similar upwards coarsening sandstone bodies on several places on Edgeøya to be mouth bars.

The observations made here is in agreement with previous studies, and these sandstone bodies are also here interpreted to represent mouth bars, deposited where sediment-laden river waters enter the sea, on the delta front.

FA4, Delta plain, interdistributary areas

Description

These deposits are mud dominated, with thin coal and coal-shale usually associated with paleosols, in addition to usually thin, up to 2 meters thick horizontal, cross-stratified and in part massive sandstones. Included in this facies is also heterolithic intervals with interbedded sand and mudstones. This facies association has been found on Edgeøya and Hopen. An example is seen on Figure 5.8.

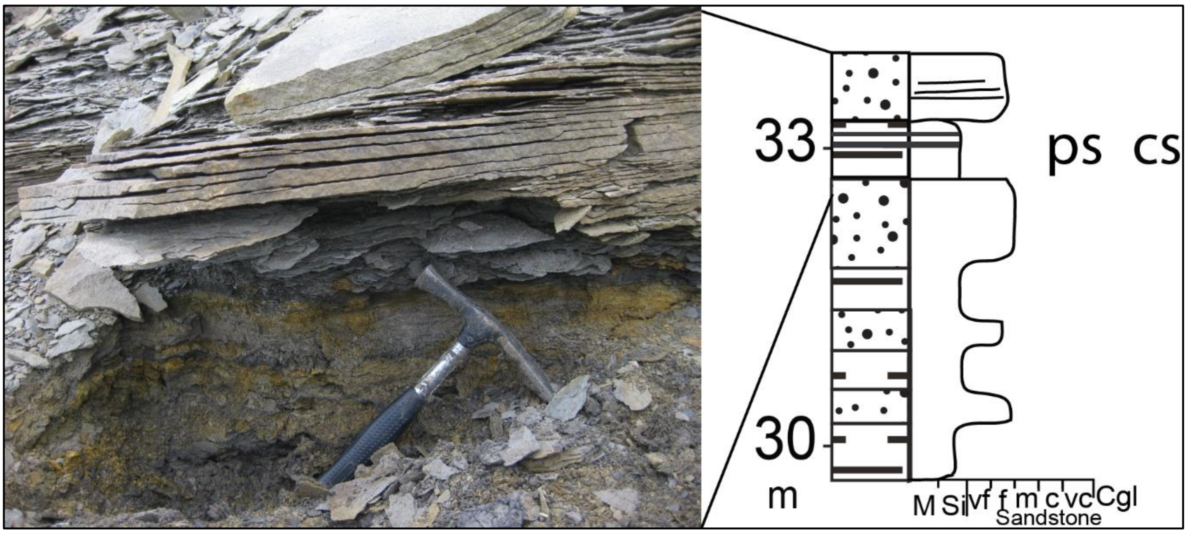

Figure 5.8: *Example of FA4, from the TF4 log. Picture shows a paleosol with thin coal shale, below a horizontal stratified sandstone interpreted to be a crevasse splay.*

Discussion and interpretation

Interdistributary areas on a delta plain is usually comprised of sub-environments such as floodplains, lakes tidal flats, marshes and swamps (Reading and Collinson, 1996). Floodplains can usually be divided into areas more proximal to channels where levees, crevasse splays and crevasse channels, and more distal areas where fine grained sediments are deposited during floods (Collinson, 1996).

The mud dominated parts within this facies association are interpreted to represent distal overbank deposits from channels on the delta plain, where the thin sandstone bed may be more proximal crevasse splays. Fine grained deposits may also represent low energy sedimentation in waterbodies such as lakes and lagoons on the delta plain, and it is possible that some of the paleosols represent the vegetated margins of tidal flats or lagoons.

5.1.3 Distribution and properties of paleosols

Edgeøya

The logs are presented below. On Tjuvfjordhorga several laterally spaced logs where made, while data from Blanknuten and Negerfjellet are confined to single short logs. A correlation panel with the logs from Tjuvfjordhorga is shown in Figure 5.9, and the single logs from Negerfjellet and Blanknuten is shown in Figure 5.10. Legend to the logs is shown in Figure 5.11.

Observations

The descriptions of paleosols from these localities are not very detailed, as they sometimes were hard to recognize, and partly covered in the steep cliffs. It was however possible to determine their overall abundance, and their relation to under and overlying sediments. The paleosols observed at Edgeøya are identified on the basis of distinct colors, presence of coal, and roots where observed. The paleosols range in thickness from 50 cm to approximately 2 m. They occur on top of or within sandstone bodies, and isolated within mud dominated sediments. Usually they are highly unconsolidated, and consist mostly of mud or sandy mud. They show no signs of any primary bedding or lamination. The colors they show are somewhat varied, and are found to be brown, yellowish brown, and dark grey. The coals and coal-shales up to 50 cm thick that occur the upper parts of some of the paleosols often show gradual transition through coal shale to grey mudstones. Not all the coals and coal-shales seem to be associated with paleosols. The paleosols at Edgeøya have been found to occur in three main ways. On top of channel sandstone bodies, on top of delta front sandstone bodies and within interdistributary delta plain deposits.

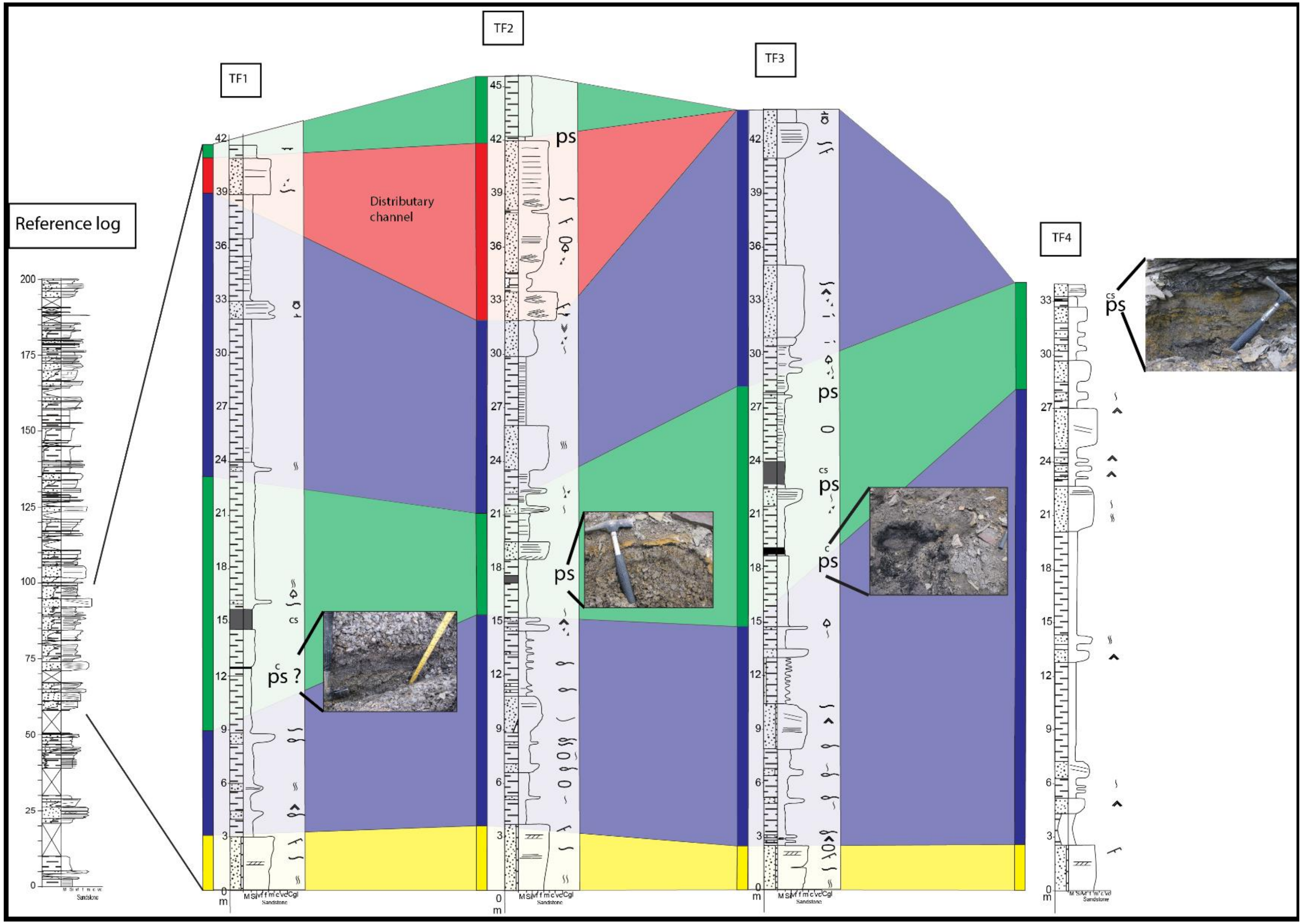

Figure 5.9: *Correlation panel with logs from Tjuvfjordhorga. Distance between each log is approximately 150 m. The pictures shows some of the paleosols. The Reference has been logged from the base of De Geerdalen Formation,and has been logged and drawn by Trond Svånå Harstad.*

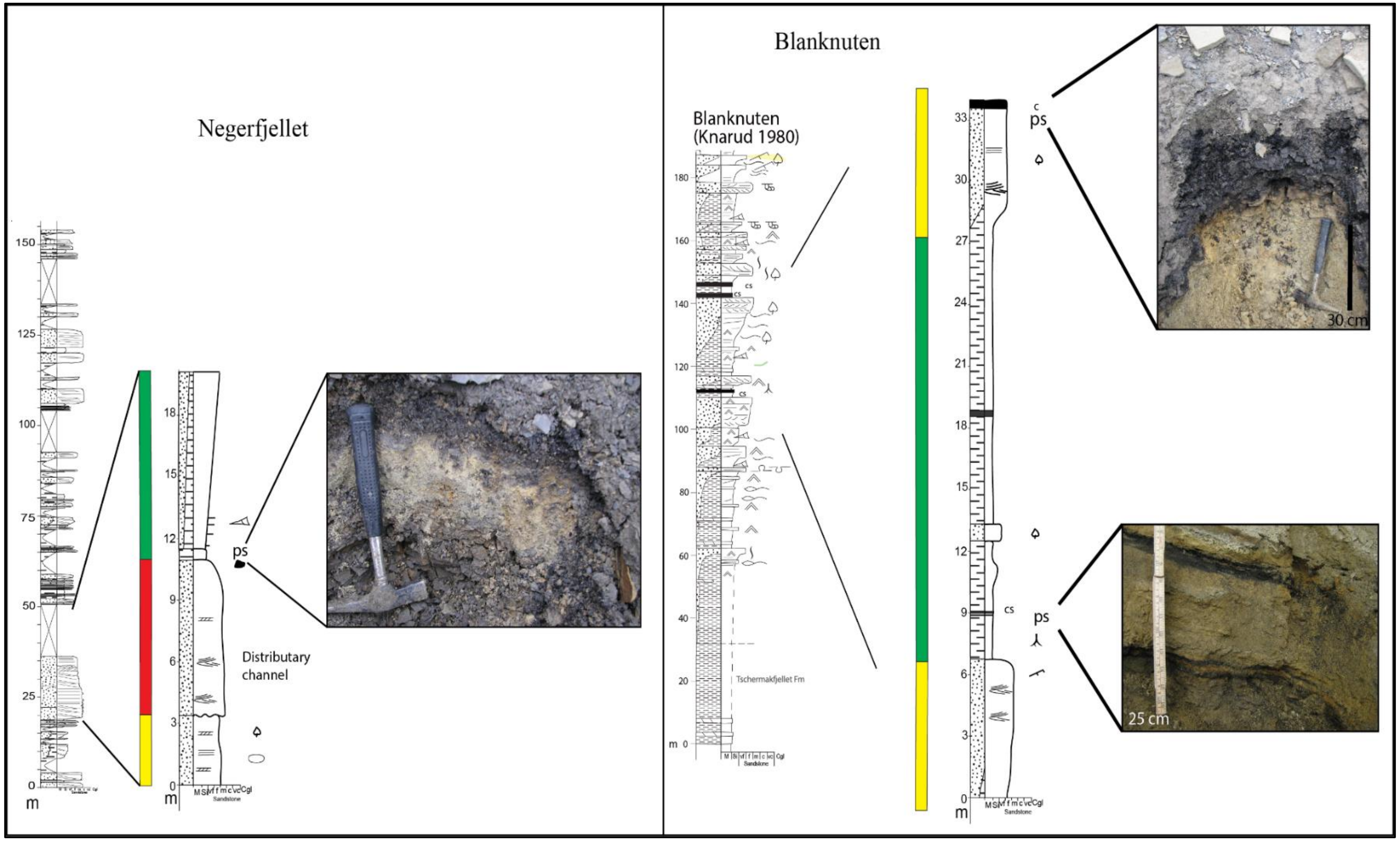

Figure 5.10: *Logs from Blanknuten and Negerfjellet. Reference log from Negerfjellet has been logged and drawn by Trond Svånå Harstad. Reference log for Blanknuten is from Rød, et al. (2014), logged by Ragnar Knarud.*

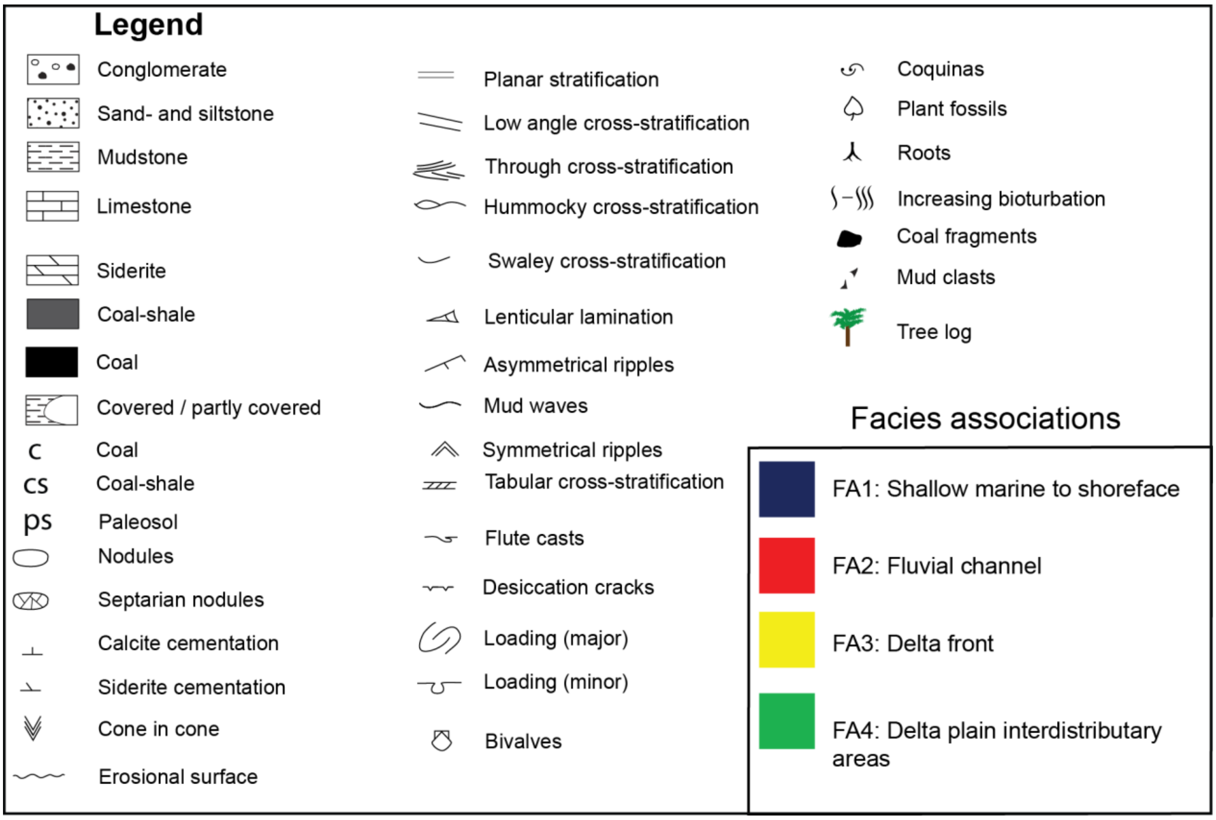

Figure 5.11: *Legend to logged sections of the De Geerdalen Formation.*

Discussion and interpretation

Paleosols associated with mud dominated delta plain deposits was found on Tjuvfjordhorga. In floodplain environments the presence and properties of paleosols are controlled by factors such as distance to channels, flooding frequency and sedimentation rate (see Section 2.3) (Kraus and Aslan, 1999). The conditions facilitating soil formation in these deposits might be periods with reduced sediment in-between floods, or a distant position to channels. Several thin coal seams occur within some of the paleosols, indicating multiple phases of plant colonization and coal formation. This is likely the result of the episodic nature of overbank deposition, where breach in sedimentation between major floods allows for plant colonization and soil formation. Delta plain interdistributary areas also include sub-environments such as tidal flats and lagoons, where paleosols might represent their vegetated margins.

Paleosol on top of channel sandstones have been identified at Tjuvfjordhorga and Negerfjellet. They are not associated to any coal seams. Such paleosols can be explained by avulsion of the channel, followed by a period of non-deposition, allowing for plant colonization and soil formation. As abandoned channels tend to be topographically elevated relative to their surroundings (Collinson, 1996), the area they occupy would receive little sediment for some time after the avulsion event. The lack of coal in these paleosols might be explained by the fact that the abandoned channel was raised above the groundwater level, inhibiting waterlogged conditions needed for accumulation of peat.

At Blanknuten paleosols were found in association with several meters thick coarsening upwards sandstone bodies interpreted as delta front deposits, where they occur on top. These paleosols also have coal on top. Where coarsening upwards sequences in deltaic environments end at sharp surfaces, they commonly mark abandonment surfaces (Reading and Collinson, 1996). Where a paleosol occur as an abandonment surface they indicate abandonment on a more landward part of the delta than abandonment surfaces indicating flooding, such as intense bioturbation (Reading and Collinson, 1996). Abandonment could involve switching of the whole delta, or it could be a more local event. A more local abandonment might be facilitated by that the distributary channel bifurcate around the old mouth bar, and deposits a new mouth bar further seaward (Diessel, 1992). This would leave the old mouth bar starved of sediments and allow for plant colonization and soil formation to take place. These paleosols are here interpreted to be coastal marshes, or at least vegetated surfaces, that evolved on sediments that were exposed due to progradation and abandonment of delta front deposits.

The distinct colors and color zonation seen in some of the paleosols could be indication of leaching, as horizonation of soil profiles typically involve downwards transportation of solids and solutes leading to color variations (Retallack, 2001). Arndorff (1993) found that paleosols showing brown colors in levee bank and crevasse splay paleosols was related to illuviation of iron, which possibly could explain some of the color seen in places. However, as no samples of these zones are analyzed here, interpretation of the processes leading to the color variation will only be speculative. Some of the colors may also be explained by oxidation of originally reduced minerals, caused my modern weathering of the outcrop.

As no diagnostic features have been confidently identified in these paleosols, they are not classified. Some of the paleosols do however appear to be relatively thin, which suggests they might fall in the category of immature paleosols belonging to the entisol or inceptisol of the USDA soil taxonomy, and the protosol of Mack et al. (1993). The paleosols with a bleached upper horizon, such as those from Negerfjellet and Blanknuten shown on Figure 5.10 could be similar to those described by Klausen and Mørk (2014), who consider them to be argillisols.

Argillisols is a soil order in the Mack et al. (1993) classification system where removal of clay from the upper horizon is the most important process, often leaving a brighter colored residue in the upper part of the soil.

Hopen

At Hopen several short, closely spaced logs of a paleosol and the underlying and overlying sediments where made. The purpose of this reveal how soil substrate, presence of coal and nodules, and thickness of the soil varies laterally. Two channel bodies occur in the logs. They correspond to the "Russevika" and Russevika Reef" channels, belonging to the lower channel zone of Lord et al. (2014b) , and their names are adapted here. Both channels are interpreted to be potential trunk rivers, based on their geometries, but the majority of the channels in this zone are interpreted to be distributary channels (Lord et al., 2014b). A picture of the paleosol, with a log is shown in Figure 5.12, while the logs showing the lateral variations are shown in Figure 5.13.

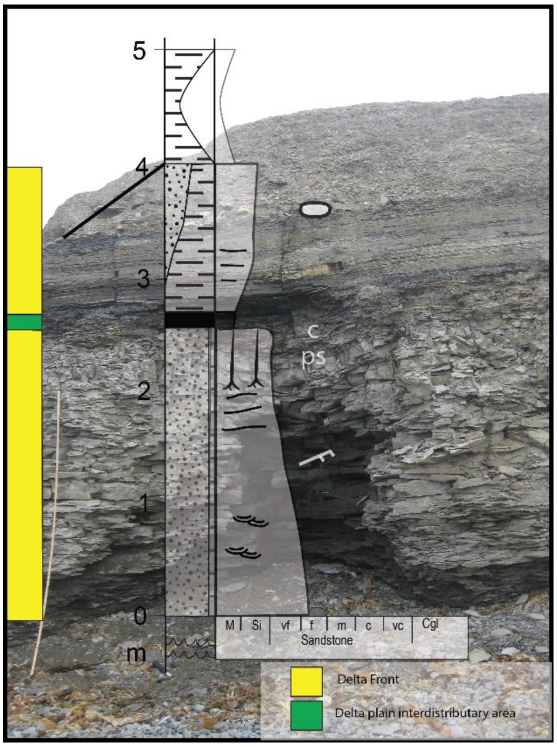

Figure 5.12: *Picture and log showing the paleosol. The log shown corresponds to log 8 on Figure 5.13.*

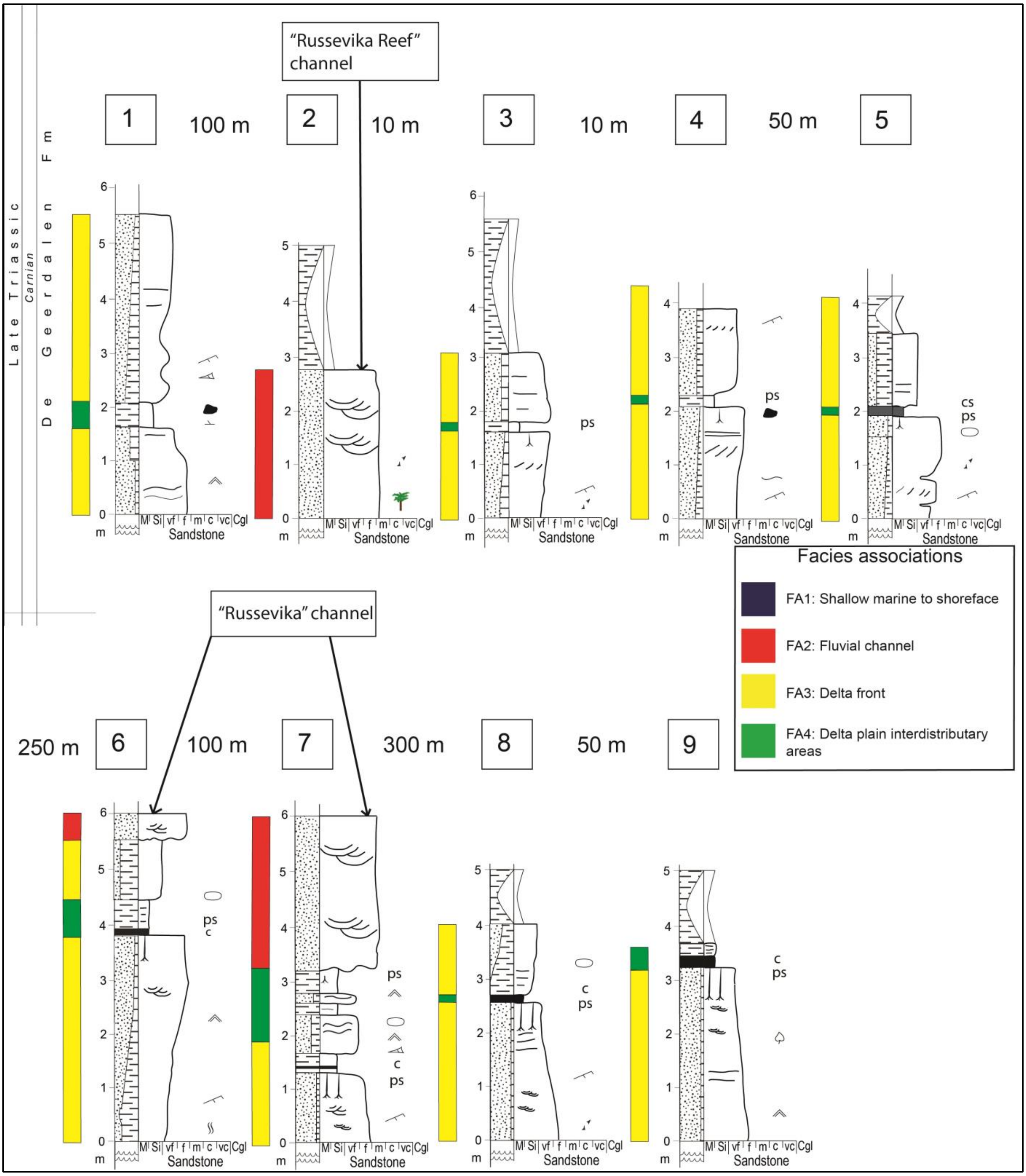

Figure 5.13: *Logs from Hopen, from south (log 1) to north (log 9), legend in Figure 5.11. Horizontal distance between each log is indicated. Base of each log is the sea level. Channel names from Lord et al. (2014b).*

Observations:

The logged section consists of a lower sand dominated unit with various cross-stratification and ripples in the lower part. This unit is continuous along the whole transect, except where it is seen to be eroded by the Russevika Reef channel to the South. Laterally it changes from being more heterolithic in the south, to more sandy and through cross-stratified to the north. Both symmetrical and asymmetrical ripples are seen, with common mud drapes in the more heterolithic parts. The paleosol occur on top of this unit.

The properties of the paleosol horizon changes noticeably across the transect. Most apparent is the change in the content of organic matter found in the upper parts of the paleosol. A clear transition from coal shale, to coal can be seen from the south on log 5, northwards to log 6. Further South only minor lenses and clasts of coal are found. The coal is locally found to be stained with a yellow color, likely to be sulphur. An abundance of well-preserved roots, arranged in vertical growth position occur almost everywhere, but both the density of roots, and the depth they penetrate varies slightly. Nodules of siderite occur occasionally at around 0.5 meters below the top of the paleosol. In one part there is two paleosols stacked on top of each other, with wave rippled heterolithic sediments in between, as seen on log 7. Figure 5.14 shows some of the features found from the paleosol. Above the paleosol a new dominantly coarsening upwards unit occurs. Where it is not covered by scree, it can be seen to have heterolithic bedding and asymmetric ripples.

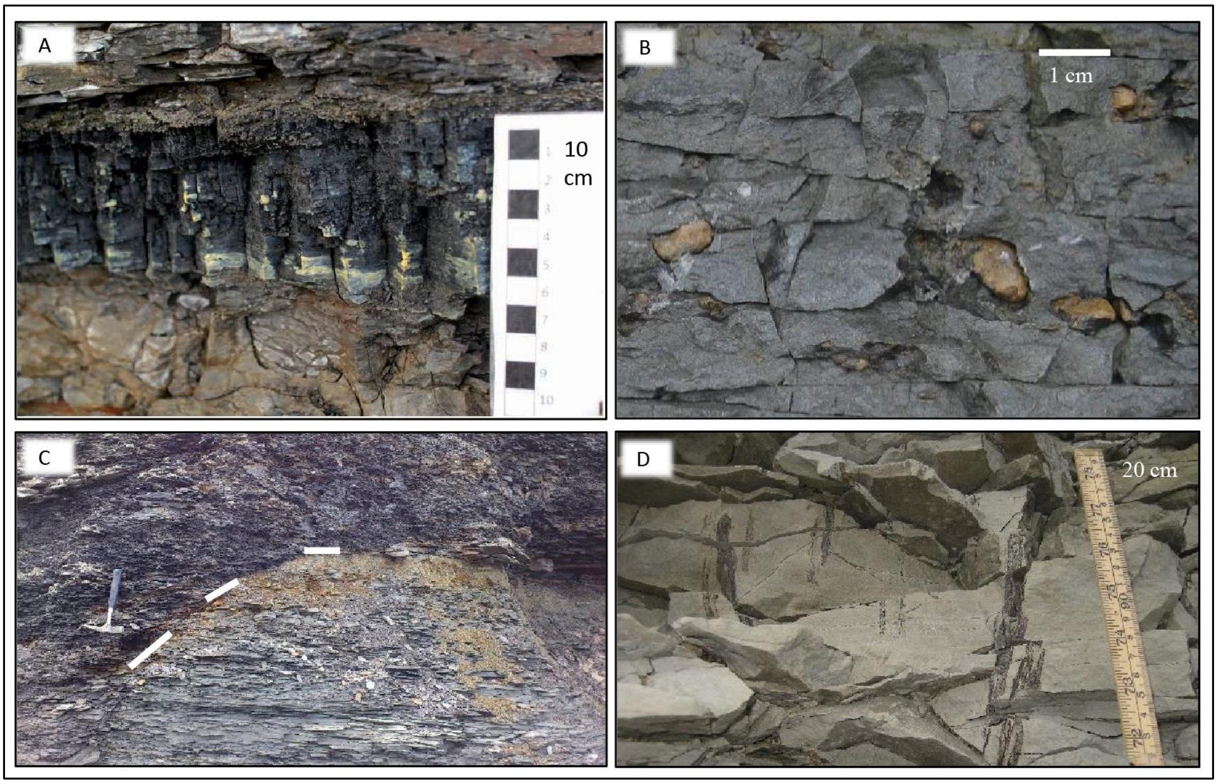

Figure 5.14: *Various features of the paleosol from Hopen* ***A)*** *Approximately 5 cm thick coal, from log7, picture from Lord et al. (2014b)* ***B)*** *Approximately 1cm sized nodules of siderite, 0.5 m from the top of the paleosol, from log 5.* ***C)*** *Truncation of the paleosol horizon by the Russevika Reef channel, between log 2 and 3. Picture taken by Turid Haugen* **D)** *Roots in growth position, from log 8. Picture taken by Turid Haugen.*

Discussion and interpretation

The lower unit is interpreted to be deposited in a delta front setting. The lateral changes from fining upward, to coarsening upward and the various current indicators could reflect variations in the energy during deposition, something that would be expected in a delta front environment, where fluvial currents interact with currents set up by tidal and wave activity (Reading and Collinson, 1996).

From south to north, coaly shale passes laterally into coal. This transition is likely to reflect the balance between sediment influx and peat accumulation in the marsh. Coal indicates more or less permanent waterlogging with minimal input of siliciclastic material, while coal-shale and carbonaceous shale is deposited where both sediments and organic material accumulate (McCabe, 1984). The coal-shale is interpreted to represent the margin of the peat forming environment, receiving more sediments than the central parts. Sulphur is considered a good indicator of presence of marine waters in the peat forming environment (Diessel, 1992), suggesting that peat formation took place close to the coast. This further suggest that the paleosol evolved as a coastal marsh at the lower delta plain, and the mechanism providing subaerial exposure of the delta front sediment is suggested to be similar to the ones mentioned for the paleosols at Blanknuten. The presence of stacked paleosol found on log 7 is likely the result of localized sediment input on the marsh. Based on their shape and position within the paleosol profile, the siderite nodules are interpreted to be of pedogenic origin. Similar nodules with comparable position within the paleosol are found in core 7430/07-U-01, where much of the siderite is proven to have originated in the soil as pedogenic carbonate (Stensland, 2012). This would indicate reducing stagnant waters, which is common in marshes and in association with coal (Ludvigson et al., 2013). The nodules does not bear any particular resemblance to the carbonate bodies in facies H associated with cone-in-cone structures, as they are much smaller in size and more spherical in shape. The Russevika Reef and Russevika channels was most likely not active at the same time as the paleosol formed, as they can be seen to partly incise the heterolithic sediments above the paleosol, meaning that both of these channels represents a later progradational stage.

The paleosol lack features that would indicate significant maturity, such as well developed horizons or thick coal. It can therefore be classified as a entisol or inceptisol after USDA soil taxonomy, and the protosol of Mack et al. (1993).

5.2 Snadd Formation, core 7430/07-U-01

Logs through some of the paleosols, and a description of their facies will be presented in this section. In addition an interpretation of the paleoclimate based on the paleosols is included. Not all of the paleosols occurring in the core are shown in the logs, but certain intervals have been picked in order to be able to show the diversity in paleosols. The paleosols have been grouped into three different types. The grouping is based on variations in organic content, thickness, grain size, and other features such as slickensides and mottling. Each type of paleosol will therefore largely have the same set of features, and could subsequently be inferred to have formed in similar environments, or by a similar set of processes. The logs start in sediments that show no sign of pedogenic modification and through the paleosol profile. In some profiles multiple episodes of sedimentation and soil formation can be seen. As mentioned in Section 1.6 an interpretation of this core has been done by Stensland (2012). An overview log from this work, also showing the intervals logged here, are shown in Figure 5.15.

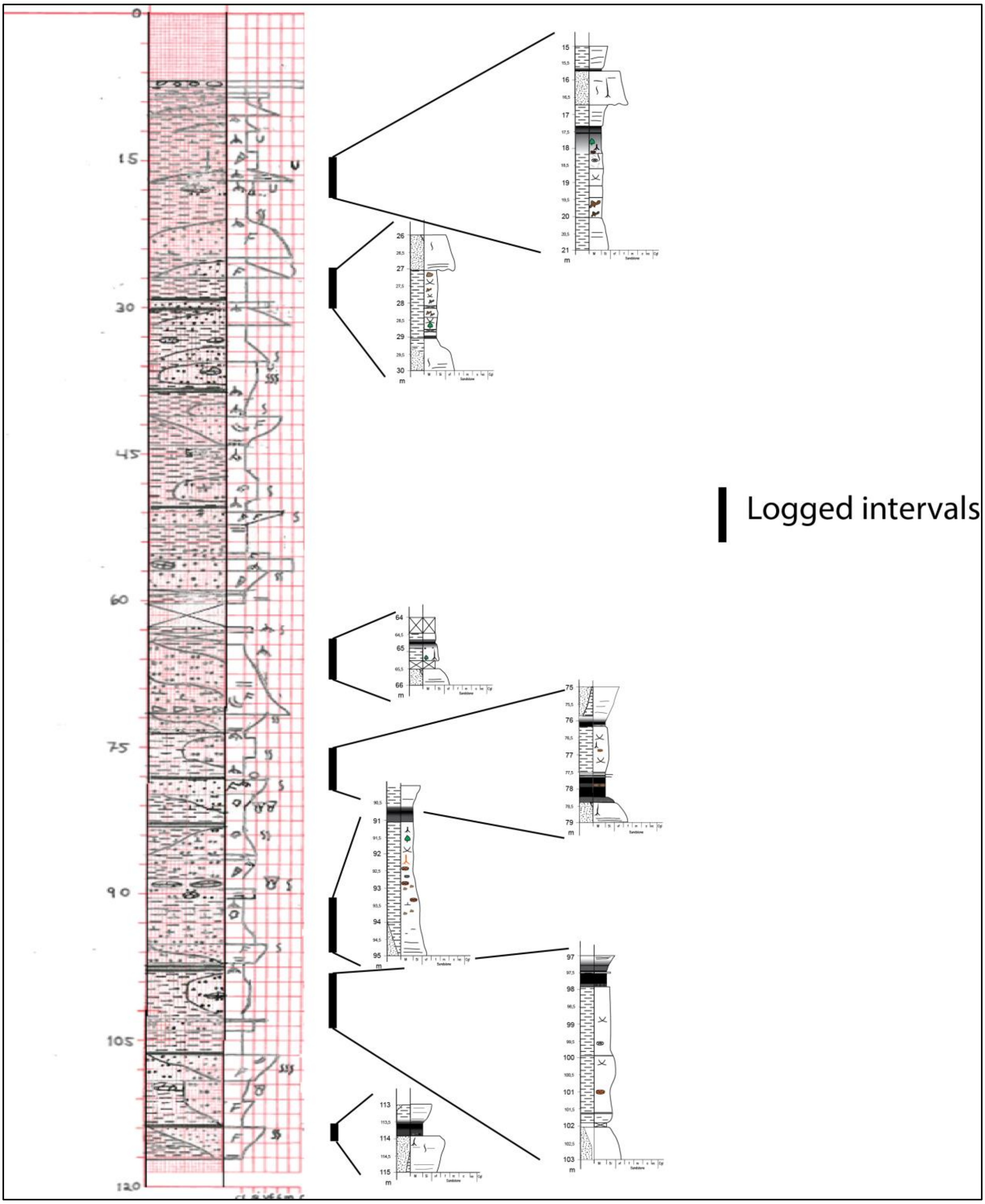

Figure 5.15: *Overview log from Stensland et al. (2013), where the logged intervals are indicated.*

5.2.1 Facies

Facies SA, Coal:

Description

The coal often shows gradual transitions to facies SC, organic rich shale/coal shale. Thickness of this facies ranges from a few cm to 70 cm. It shows a black color, with a mostly dull luster with thin bands showing more vitreous luster in places. Yellow stains indicating sulphur occur on top of some of the coals. This facies is usually found on top of one of the other paleosol facies, but some coals occur where no signs of pedogenic modification of the underlying sediments can be seen. Figure 5.17A shows a picture depicting the typical appearance of this facies. TOC analysis was done on some of the coal seams. The results are shown in Table 5.2.

Table 5.2: *TOC of selected coal seams.*

Sample depth	**% TOC**
77.20 m	34.6
78.20 m	65.3
97.30 m	63.8
97.60 m	55.5

Some of the coal accumulations appear not to be related to paleosols, meaning that features such as roots and destratification cannot be seen in the underlying sediments. This is clearly seen in a coal seam occurring at 106.60 m in the core, shown in Figure 5.16, where lamination can be seen to be well preserved below the coal.

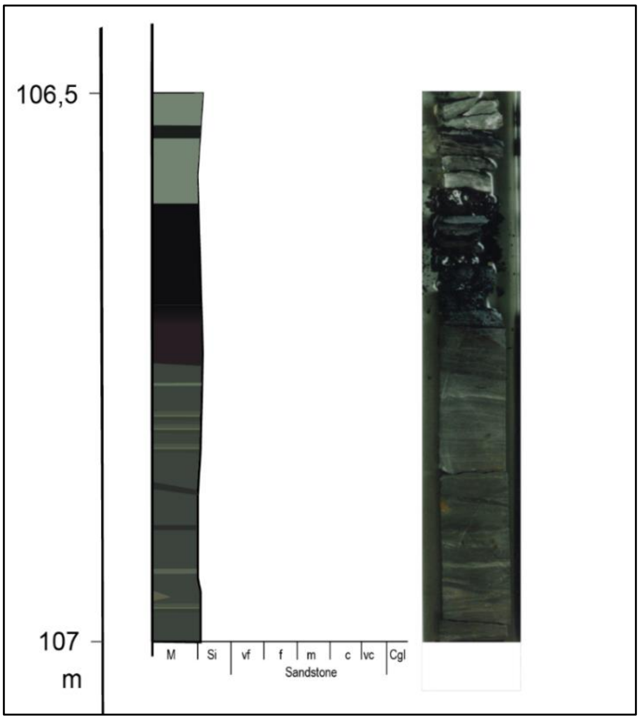

Fig 5.16: *Thin coal without any roots or other indicators of pedogenic modification of the underlying sediments.*

Discussion and interpretation

Coal deposits usually represent peat deposits formed swamps and lowlands, with little to no sediment influx, and high water table (Diessel, 1992). The term O horizon is used on surface accumulations in soils, consisting of organic material (Retallack, 2001). Coal is usually defined as a rock containing more than 50% organic carbon (McCabe, 1984). The TOC results reveal the coal generally falls within this definition, although the sample from 77.20 m shows TOC below 50%. This shows that some of the coal should be defined as coal shale, and as distinction between the two is only been based on observation, some uncertainties in this separation should be expected.

Where the coal occurs with paleosols, it is interpreted as O horizons on top of paleosols, where organic material accumulated without significant decomposing. Indications of Sulphur in some of the coals are also here interpreted to attest to influence of marine waters in the peat forming environment.

Coal can also form as floating peat in lakes or lagoons, and as detrital accumulations of organic material (Diessel, 1992). This could be the explanation for the coal seams where no roots or other indications of pedogenic modification occur below.

Facies SB, A Horizon

Description

This facies consists of mm sized clasts of coal and mud in a dark organic rich matrix. Plant remains are present in places, and this facies is not seen to be thicker than 5cm. The boundaries to the over and underlying facies can be seen to be both sharp and gradational. Figure 5.17B shows the typical appearance of this facies.

Discussion and interpretation

A coarse texture is typical for A horizons in soils, where some of the fine fraction is lost to lower horizons, and where both organic and clastic material accumulate (Wright, 1992a). These horizons can occur at the top of a soil, or be buried below a later evolved O horizon (Retallack, 2001).

This facies is interpreted to represent a stage of soil formation where both organic material and clastic sediments accumulated in the topsoil. The A horizon, where they are found below coal, could be a remnant top soil horizon, developed at an early stage of soil formation and later buried by organic accumulations.

Facies SC, Organic rich shale/coal shale

Description:

This facies consists of dark grey and dark olive grey mudstones with a high organic content. The thickness of these shales vary from a few cm to 40 cm. It often shows gradual transitions to laminated grey mudstone and coal. Thin laminas of silty mud occur within this facies in places as seen in Figure 5.17 C.

Discussion and interpretation

Organic rich shales can be formed in swamps and low lying areas receiving some input of sediment, inhibiting the formation of pure coal (Diessel, 1992). Sediment dispersal into swamps is often controlled by floods and storm events (Diessel, 1992, McCabe, 1984).

Facies SC is interpreted to be part of a O horizon, that have formed in areas closer to a sediment source than facies SA, coal.

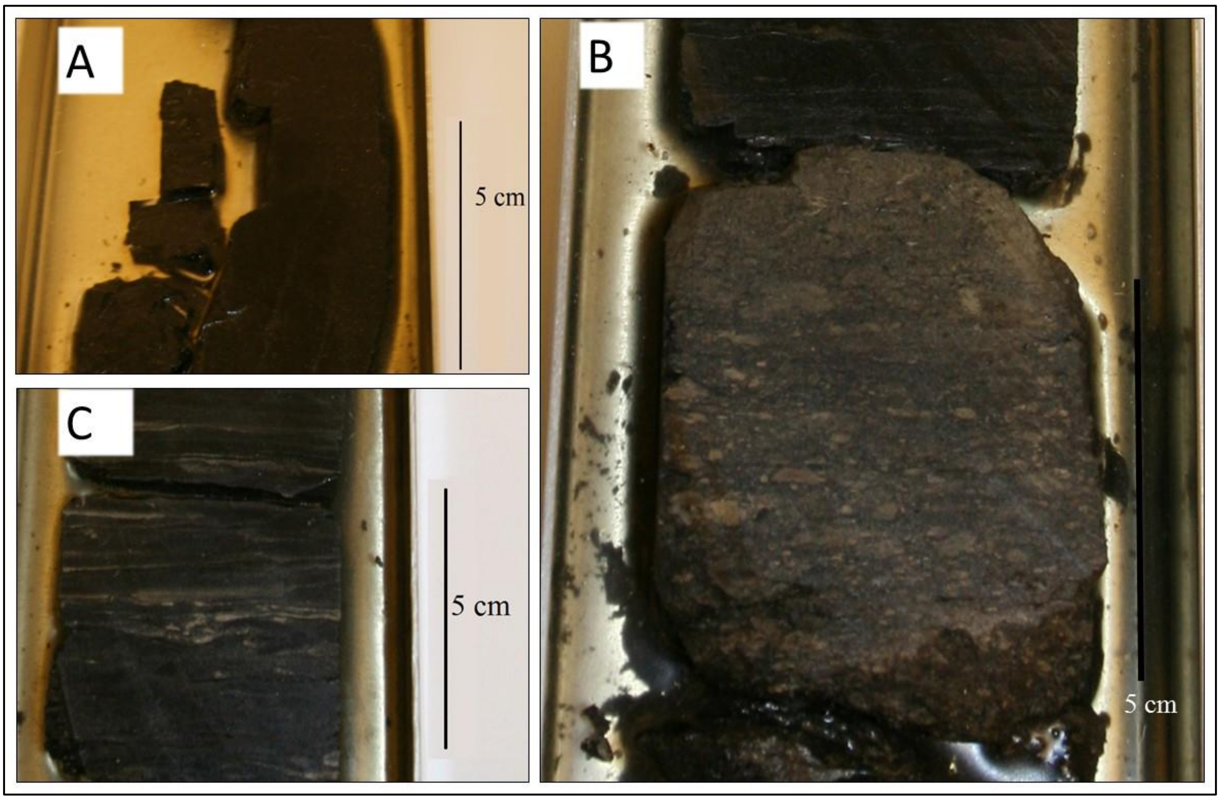

Figure 5.17: *Facies in core 7430/07-U-01.***A)** *Facies SA, coal, from 78.20 m* **B)** *Facies SB, A horizon with sharp boundary to overlying coal. From 97.80 m* **C)** *Facies SC, coal shale with thin mud laminas. From 38.30 m.*

Facies SD, Mudstone with slickensides:

Description

Facies SD shows color varying from dark to light olive grey, greenish grey in places. This facies forms unit up 3 m thick, and is the dominant facies in some of the paleosols. It contains roots in vertical growing position and plant remains are locally abundant. Roots can also often be seen to be mineralized by siderite, forming so-called rhizoliths. Nodules of both siderite and calcite occurs, though siderite is more common. This sometimes complex pattern is showed in Figure 5.18.

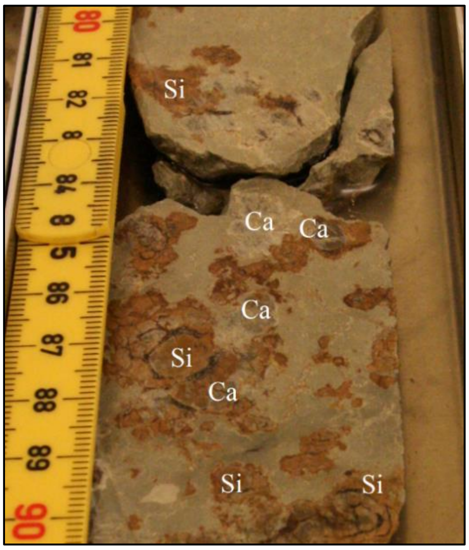

Figure 5.18: *Complex pattern of cementation in facies SD, with brown siderite and grey calcite. Ca= Calcite, Si = Siderite. From 97.80-97.90 m.*

This facies is further characterized by an abundance of slickensides. The slickensides appear as smooth striated surfaces, sometimes intersecting each other, creating wedge shaped aggregates. Their orientation appears to be random, but systematic investigation of their geometry is difficult in core cuts. They appear both in zones where the core consists of loose fragments, and in parts where the core cuts are well preserved. The slickensided planes often show high curvature, with both concave and convex surfaces. Figure 5.19 shows a number of these slickensides. This facies is also seen to frequently overlap with facies SF, mottled mudstone.

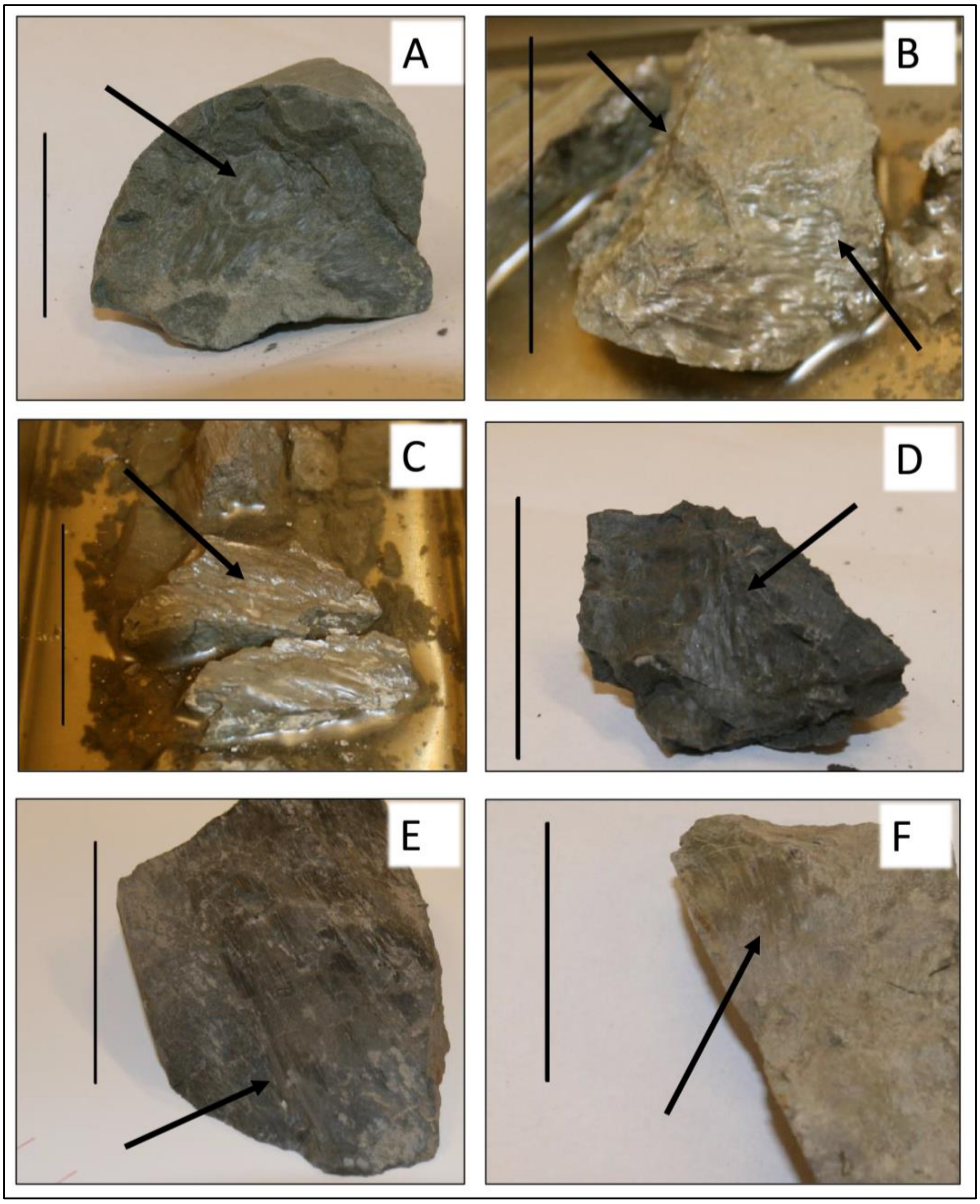

Figure 5.19: *Examples of slickensides in core 7430/07-U-01. Arrows are pointing to slickensided planes. Notice the well defined curvature and striations on some of the slickensides. Scale bars are 5cm.* **A)** *Convex shaped slickenside from 27.3m.* **B)** *Wedge shaped ped created by intersecting slickensides. From 27.3 m* **C:** *From 35.9m.* **D:** *From 39.3m* **E)** *Concave shaped slickenside, from 77m.* **F)** *Concave shaped slickenside, from 91.5m.*

The slickensides can also be seen in thin sections, where they appear as pale linear features embedded in darker mudstone. Locally, angular to sub-angular quartz and feldspar grains occur as thin linings or as thicker aggregates in the slickenside plane, as seen below in Figure 5.20.

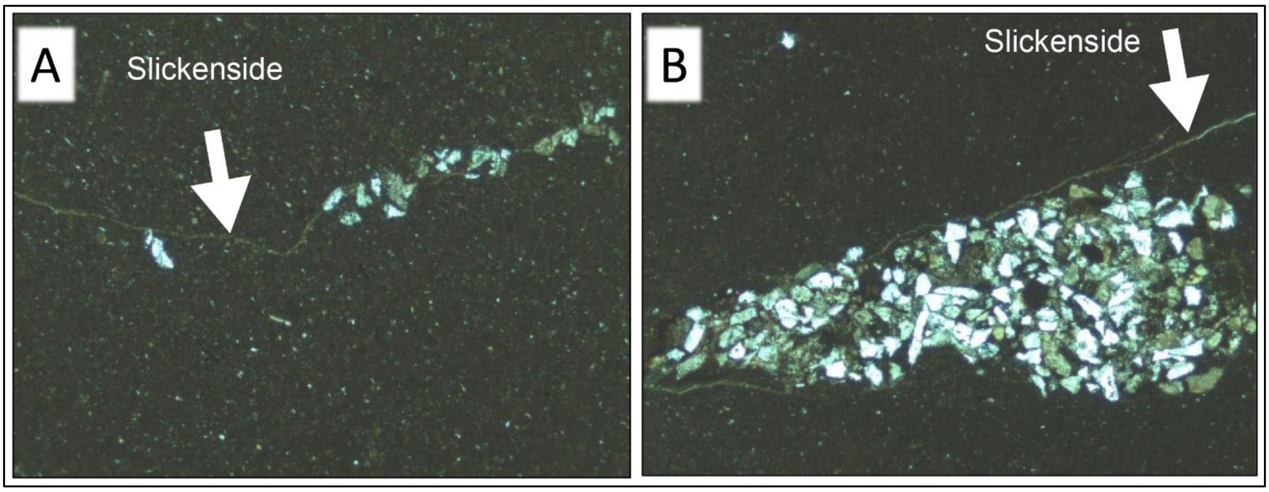

Figure 5.20: *Photomicrographs taken with plane polarized light at 12.6 m, core 7430/07-U-01. Width of both pictures is 3.75 mm* **A)** *Slickenside line with fine sand mostly consisting of quartz and feldspar.* **B)** *Aggregate of fine sand next to a slickenside.*

Discussion and interpretation

Slickensides are common features in clay rich soils that experience seasonal moisture variations, causing swelling and contraction in smectite clay (see Section 2.2.1) (Retallack, 2001). Slickensides can also form in relation to faults or other types of deformation caused by external tectonic stress, but the often curved morphology of pedogenic slickensides is hard to relate to any known tectonic deformation processes, and are best explained by contraction and expansion in unconsolidated swelling clays (Gray and Nickelsen, 1989). Precipitation of carbonate minerals is typical in clay rich B horizons which forms part of the subsoil (Retallack, 2001). Much of carbonate was proven to have originated as pedogenic carbonate by Stensland (2012), who recognized features such as *Microcodium,* cemented roots, and carbonate coated grains.

Some of the slickenside planes curve through arcs with up to 90^{o} angles in places. As they also occur in intervals with other pedogenic features such as roots and carbonate nodules and are often overlain with coal, the slickensides are likely to be of pedogenic origin. In addition they can be found in places where the core cuts are completely intact, meaning that is unlikely that they are related to e.g. fracture zones. They are thus interpreted to be related to shrink-swell phenomena caused by varying soil moisture content in unconsolidated sediments, in a subsoil B horizon.

Facies SE, Pale mudstone

Description

This facies consists of light olive grey and greenish grey colored mudstone (see Figure 5.21A), in 20 cm thick intervals. This facies is found below A or O horizons or within facies SD, mudstone with slickensides. These mudstones stands out as being brighter in color than the mudstones of facies SC and SD.

Interpretation and discussion

Bright colored horizons are common in soils that are well drained, where water percolate freely through the soil, and remove clay minerals from the upper part of the soil (Wright, 1992a, Retallack, 2001).

Facies SE is thus interpreted to be part of a E horizon, where leaching of clay and other minerals in the fine fraction has left a brighter colored quartz rich residue.

Facies SF, Mottled mudstone

Description

This facies consists of mudstones with a distinct color mottling, consisting of various colors of red, pale orange and brown, in intervals up to 1 m thick. The facies also contains some siderite nodules, and often overlaps with facies SD, mudstone with slickensides. Figure 5.21C shows how the mottling pattern of this facies typically appear in the core cuts.

Discussion and interpretation

Color pattern is most likely determined by the distribution of iron oxides and iron hydroxides, which are both common minerals in soils (Duchaufour, 1982). Hematite is commonly responsible for red coloration of sediments, particularly paleosols, and can cause a distinct red color with as low concentrations as 0.2% (Duchaufour, 1982). Red colors can also be related to later diagenetic modification, meaning that some care should be exercised in interpreting these features as exclusively pedogenic (Wright, 1992a). The fact that the color pattern only appear in zones where other indicators of pedogenic modifications such as roots and slickensides occur, should suggest that these features are pedogenic, and not formed by later burial diagenesis. This facies is interpreted to be part of a B horizon, formed under more oxidizing conditions than the other, dominantly grey mudstones.

Facies SG, Rooted and bioturbated mudstone to very fine sand.

Description

Facies SG shows little to no visible primary sedimentary structures. It shows gradual transition from facies SH stratified mudstone-very fine sand, and olive grey and brownish grey color. The facies contain abundant roots in vertical growing position, penetrating down to 50 cm below the top of paleosols. Figure 5.21 D shows a siltstone with thin roots.

Discussion and interpretation

Parts of a soil with low degree of pedogenic modification, where only faint primary structures are preserved, and typically lacks features such as mineral precipitates and abundant clay is usually termed C horizon (Retallack, 2001). Facies SG is here interpreted to represent a zone that has not undergone enough alteration to qualify as a B horizon.

Facies SH, Stratified mudstone-very fine sand:

Description

The stratification of facies SH is sometimes well defined as in horizontal laminated mud, and with color variations between the individual laminas, as seen in Figure 5.21B. In more sandy units stratification tends to be more diffuse, but is well defined in more heterolithic intervals. This facies is found in the lower parts of paleosols profiles.

Discussion and interpretation

Preserved stratification suggests that this facies has undergone little to no pedogenic modification (Retallack, 2001). This is probably due to high sedimentation rate, or subaqueous deposition in e.g. lakes.

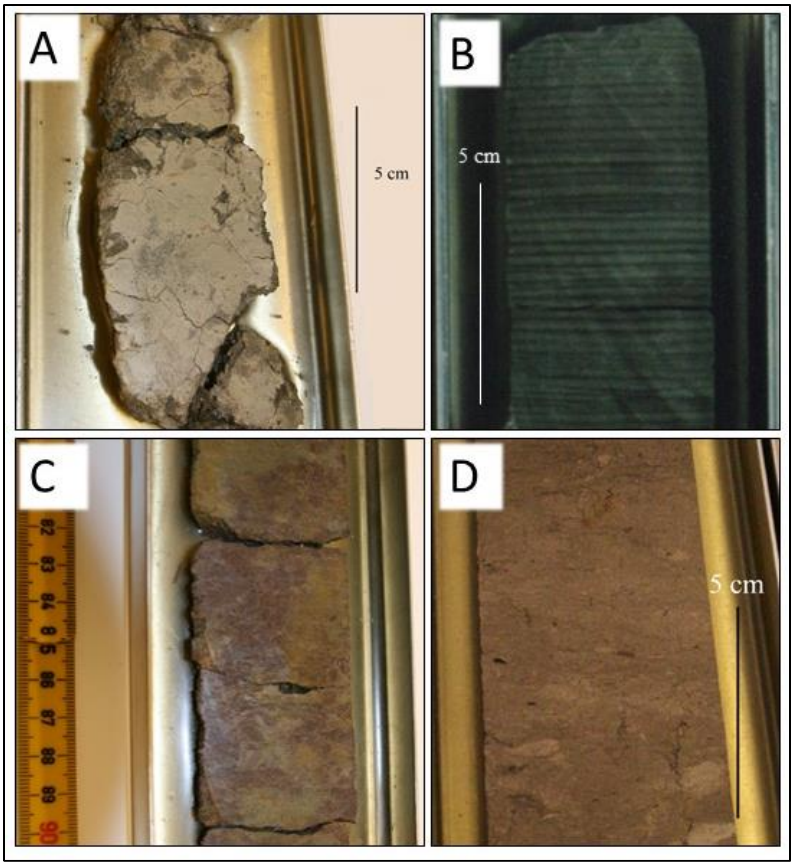

Figure 5.21: *Facies in core 7430/07-U-01* **A)** *Facies SE, pale mudstone, from 19.40 m.* **B:** *Facies SH,stratified mudstone-very fine sand. Here seen as a mudstone with well defined lamination. From 105.10 m* **C)** *Facies SF, mottled mudstone. From 27.80 m* **D)** *Facies SG, Rooted and bioturbated mudstone to very fine sand. From 64.90 m.*

5.2.2 Paleosols

A log with an example of each type of paleosol is included in this section, while the complete logs are found in the Appendix. Each paleosol of the different type have been numbered, from bottom to top of the core. This means that e.g. paleosol A1 is the stratigraphically lowest of paleosol type A. The facies that occur in each paleosol type can be seen in Table 5.3.

Table 5.3: *The different paleosol types and their facies.*

Paleosols type	**Facies**
Type A	SA,SB,SG,SH
Type B	SA, SB, SC, SD, SE,SH
Type C	SA, SB, SC, SD, SF

Type A paleosols

Description

These paleosols are characterized by being relatively thin, and show thicknesses up to around 1 m. They are commonly fining upward, but usually represent the top of overall coarsening upwards heterolithic units in the lower parts of the core. In the upper part of the core they occur on top of erosive based, upwards fining, massive sandstones. Other observed features include obliteration of primary sedimentary structures, probably caused by bioturbation, disruption by roots, or both. A gradual transition of the stratification from well defined, through more diffuse, until the sediments appear homogenous can clearly be seen. Coal up to 0.7 meters thick occurs at the top of the paleosols. No mineral precipitates have been identified in this paleosol type. Type A paleosols occur throughout the core.

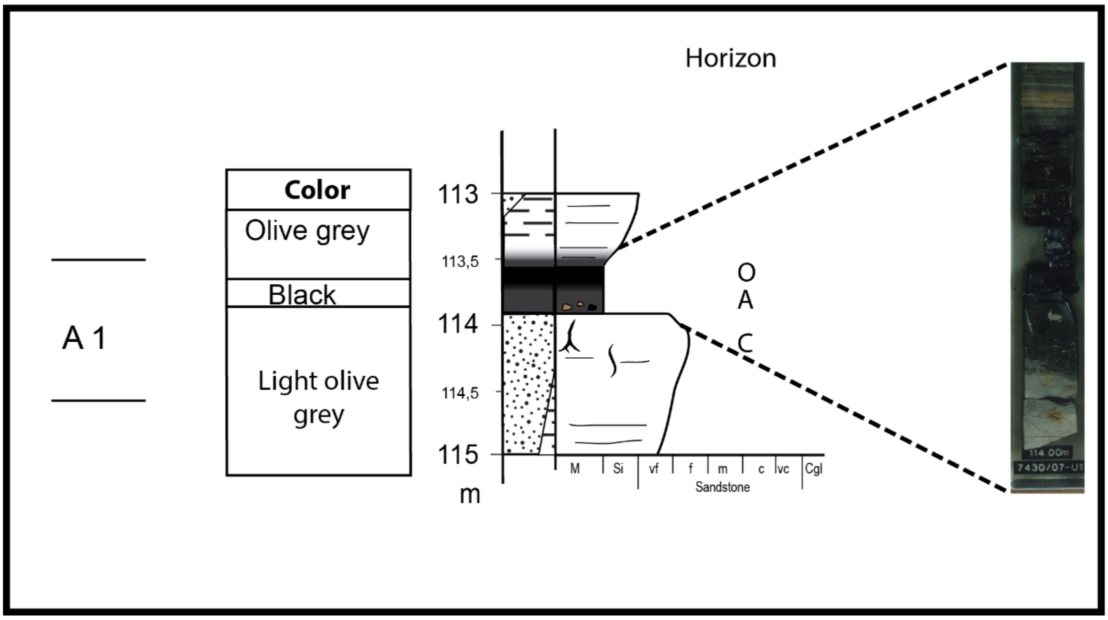

Figure 5.22: *Paleosol A1.*

Discussion and interpretation

The paleosols of type A shows poor development of horizons, and the zones that are modified by pedogenic processes are relatively thin. They have no clay rich B horizon with accumulated carbonates or other minerals, a feature that is typical for little developed soils (Retallack, 2001).They can therefore be regarded as immature paleosols, and can be classified as either entisols or inceptisol by use of the USDA soil taxonomy classification system, or as the protosol of the Mack et al. (1993) classification. Some of them do however have up to 70 cm thick coal on top, suggesting they are more evolved than the paleosols with thinner coals. These could probably also be classified as histosols.

Low maturity of paleosols can reflect short time of development, but can also be a result of poor living conditions for plants (Retallack, 2001). The presence of coal and coal shales in their upper parts suggest that they have been waterlogged, at least during some time of their evolution. This is also supported by the presence of well preserved roots. In places the distinction between roots and bioturbation is difficult. It can be explained by that the roots follow the burrows, which offer a softer substrate. Another possible explanation is organisms burrowing through the root systems. Traces of insects and other organisms burrowing into roots to feed have previously been described from Triassic paleosols (Retallack, 1988).

These paleosols are likely to have been formed in a number of sub-environments on the delta plain. In the lower part of the core they are found on top of coarsening upwards units, interpreted by Stensland (2012) to be prograding delta lobes. In the upper part they occur on top of erosive based sandstones units that was interpreted by Stensland (2012) to be minor channels and crevasse splays. Paleosols on top of crevasse splay deposits may have formed after abandonment of the channel, or they may have been formed between floods, if there were sufficient time interval between the floods.

Type B paleosols

Description

These paleosols are markedly thicker and more clay rich than group A paleosols. They are up to 3.5 m thick, and the dominating facies is mudstone with slickensides. Coal and coal shale as organic rich O horizons are up to 0.5 m thick. As some of the group A paleosols, these also define fining upwards tops of the coarsening upwards heterolithic units. They also occur on top of less defined thick units, consisting of dominantly massive and partly laminated mudstones with decimeter thin fine grained sandstones. Siderite nodules are abundant, and in places occur together with nodules of calcite, and zones that are heavily calcite cemented. The paleosols of group B occur exclusively in the lower 80 meters of the core. Figure 5.23 shows a log through paleosol B2.

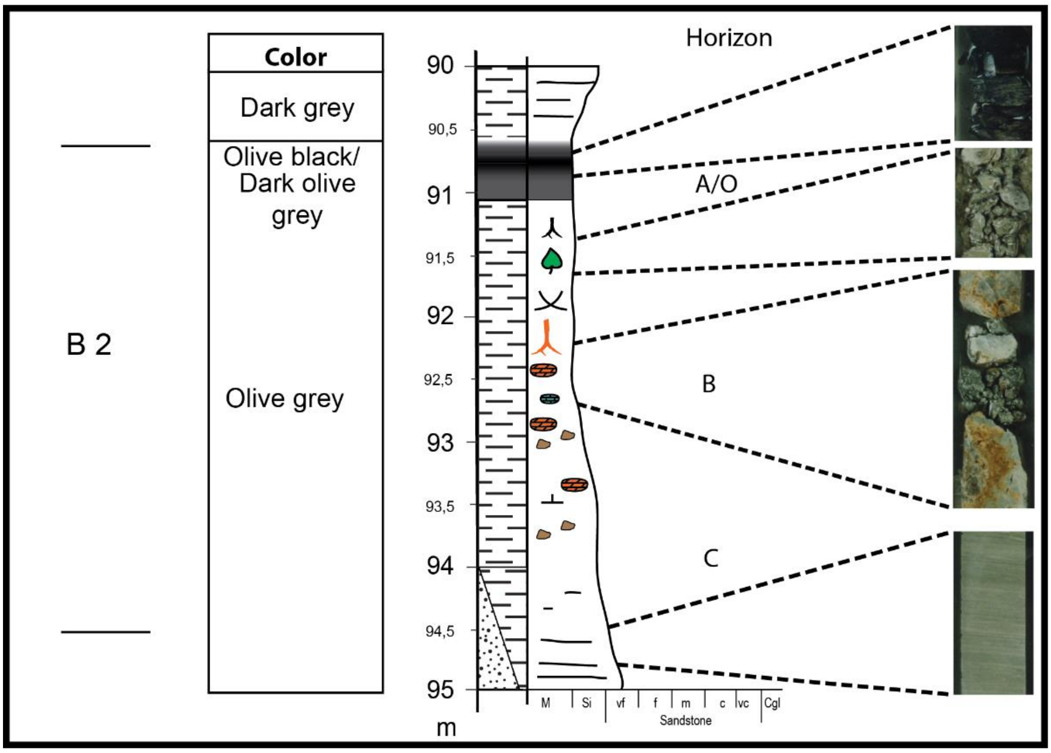

Figure 5.23: *Paleosol B2.*

Discussion and interpretation

The paleosols of Group B are dominated by mudstone with slickensides, interpreted to have been formed by a fluctuating soil moisture content. Soils with these properties belong to the vertisol soil order of USDA soil taxonomy and Mack et al. (1993). As they are also seen to have coal and coal shale in their upper parts, they could possibly also be classified as histosols.

A common feature of vertisols is clastic dykes, which are crack that are open during the dry season, and are filled with material derived from the surface of the soil (Duchaufour, 1982). The absence of clastic dikes in these vertisols could be explained by that they have been filled with the same material as the soil consisted of, making the infill indiscernible (Caudill et al., 1997). The sand seen to occur together with the slickensides shown in Figure 5.19, could possibly be related to infilling of surface material when the cracks was open to the surface. This would, however, be only speculative as very few thin sections have been analyzed, and this phenomenon could also probably be explained by other mechanisms. A much simpler explanation could be that the sand aggregates were deposited, and that their relation to the slickensides are random.

These paleosols are also interpreted to be polygenic soils, meaning that they record a change in moisture regime through their evolution. This is based on the presence of up 0.5 meters of coal on top of the slickensided horizons, which in some of the paleosols also contain nodules of calcite and siderite. These are indicators of highly contrasting moisture regimes (see Section 2.2.5). Paleosols with seemingly similar properties have been reported from Carboniferous paleosols in the USA, where they have been interpreted to evolve from initially well drained paleosols into waterlogged poorly drained paleosols, before burial or drowning (Driese and Ober, 2005, Rosenau et al., 2013a). A similar three-stage development is suggested here.

Stage 1:

At this initial stage the paleosol is suggested to represent a well drained vertisol, with formation of slickensides caused by variation in soil moisture. The calcite nodules seen in some of these paleosols where likely precipitated at this stage, as calcite requires well drained conditions in order to form in soils (Milnes, 1992). The soil was at this stage probably part of a floodplain on the delta, where variations in river discharge and precipitation caused the groundwater level to fluctuate. Extensive shrink swell activity usually result in homogenization of the soil profile (Buol et al., 2011), which in this case can explain the absence of any visible stratification in the paleosol. The paleosols was likely covered by vegetation, but accumulation of significant amounts of organic matter is unlikely, as the soil must have been dry, or at least partially dry in periods to facilitate the shrinking and swelling of the clay.

Stage 2:

At this stage the paleosol is interpreted to be water saturated and poorly drained allowing accumulation of peat and precipitation of siderite. Siderite is an indication of anoxia and reduced conditions (Ludvigson et al., 2013). At this stage the groundwater level where at, or near the surface, and sediment supply must have been at a minimum. Reducing conditions are facilitated by complete water saturation in the pores of the soil, limited flow of oxygenated groundwater, and abundant organic material together with a microbial population to decompose it (Driese and Ober, 2005). The grey colors of these paleosols is also believed to be a result of waterlogging and reducing conditions, as this is usually caused by iron in its reduced state (Retallack, 2001). Figure 5.24 shows the proposed evolution from stage 1 to stage 2.

The transition from stage 1 to stage 2 could for instance have been caused by subsidence, leaving the surface closer to the groundwater level. It could also have been caused by migration and shifting of channels, changing the local groundwater conditions. Indications of channel migration in the form of scroll bars in the Snadd Formation has been described by Klausen et al. (2014), and lateral accretion surfaces are common in channels on Hopen (Klausen and Mørk, 2014, Lord et al., 2014b), suggesting that this may be a possible explanation. Abandonment of the floodplain by an upstream avulsion could also explain this transition. As abandoned parts of deltas receive little sediments, continued subsidence would bring the surface closer to the ground water level.

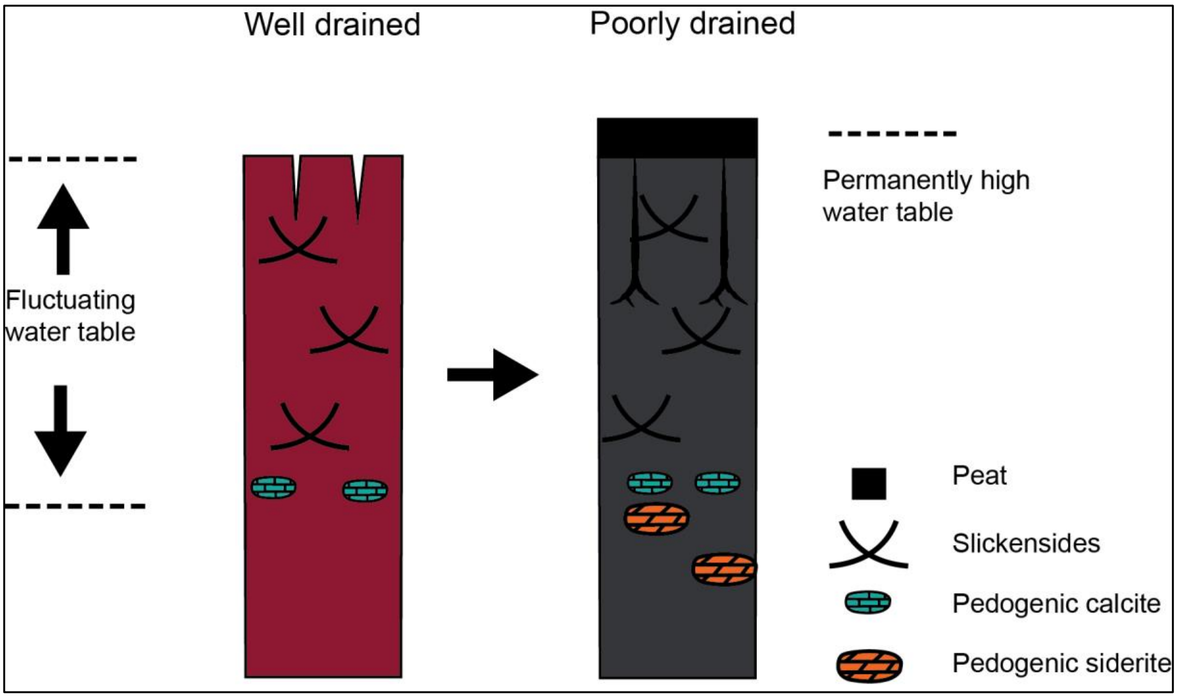

Figure: 5.24: *Transition from stage 1 to stage 2.*

Stage 3:

This stage records the termination of peat accumulation, and subsequent burial of the paleosol. Plant growth and peat termination may have been terminated in several ways. Intrusion of marine waters in the peat forming environment during transgression is a common mechanism to cease peat accumulation in near coast areas (Diessel, 1992). If the subsidence rates outpace the peat accumulation rates further from the coastline, swamps tend to evolve into lakes (Diessel, 1992). Rapid influx of sediments caused by flooding or avulsion of channels could also end peat accumulation. Gradual transition between coal, coal-shale and adjacent facies as seen in places here, typically indicates slow changes in depositional conditions (Diessel, 1992).

These paleosols are suggested to mainly have formed on the upper delta plain, at least during stage 1. This is because wetting and drying cycles would be less likely under influence of marine waters, and that vertisols in modern settings seems to be more common on the upper parts of deltas (see Section 6.3).

Type C paleosols

Description

These paleosols are distinguished from the other paleosol by the presence of facies SF, mottled mudstone. Slickensides are also present in these paleosols, also in the mottled parts, although markedly less than the type B paleosols, and where they occur they are more confined to certain parts of the profile. Only few cm of coal and coal shale are found on top. Both the C1 and C2 paleosol are overlain by thin erosive based sandstones, and the C1 paleosol is truncated by one of these sands, as seen on Figure 5.25. Several paleosols that appear to be partly eroded occur within the interval that Type C paleosols occur, which is in in the upper 30 m of the core.

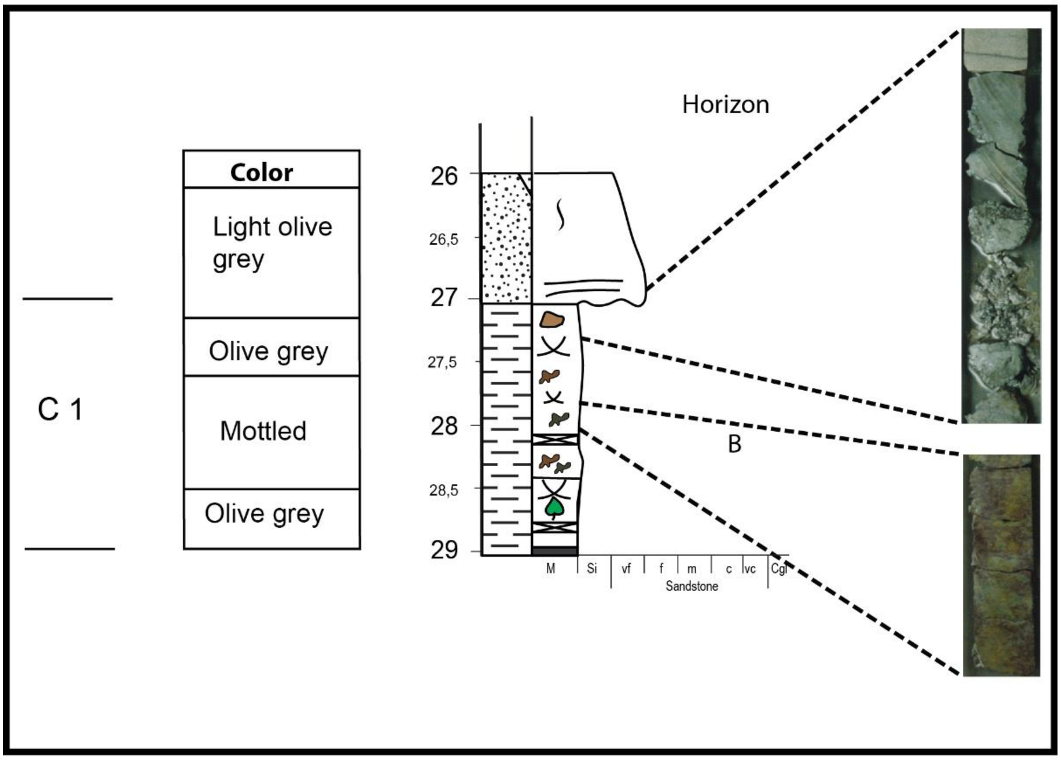

Figure 5.25: *Paleosol C1.*

Discussion and interpretation

The presence of oxidized mottles was interpreted by (Stensland, 2012) to be formed by a fluctuating water table, where varying redox potential in the groundwater led to uneven distribution of ferrous and ferric iron. Furthermore, these paleosols was interpreted to have formed on more proximal parts of the delta, based on the oxidation features, and partly supported by indication of increased freshwater component of the fluids precipitating the siderite (Stensland, 2012).

The truncation of the C1 paleosol suggests that they were not covered by thick peat layers, as peat is highly resistant to erosion (McCabe, 1984). This suggest that the environment these paleosols evolved in was unsuitable for peat accumulation, which is also evident from the oxidation features. Erosional truncation of soils is a feature that tend to occur where sedimentation is more rapid, which often is the case close to channels (Kraus and Aslan, 1999). These paleosols are therefore suggested to have formed close to a channel on the delta plain. As this type of paleosol also contain pedogenic slickensides, they could also be classified as vertisols, but as they are partly oxidized and contain little organic matter they were likely better drained than the type B paleosols.

5.2.3 Paleoclimate interpretation

The purpose of the following section is to infer about the climatic conditions based on the climate sensitive features found in the paleosols. Coal, pedogenic slickensides, pedogenic carbonate and mottles are all features that are sensitive to climate, and can be used to reconstruct the paleoclimate (e.g.Mack and James, 1994, Retallack, 2001, Ludvigson et al., 2013).

The core has been dated to early-middle Carnian (Vigran et al., 2014). The palynological zonation of Vigran et al. (2014) can not be directly correlated to the floral phases of Hochuli and Vigran (2010). However, the fact that one of the dominant palynomorphs in the core is *Leschikisporis aduncus* (Vigran et al., 2014), in addition to other palynomorphs identified as hygrophytic by Hochuli and Vigran (2010), and the abundant coal, strongly suggests that the core can be correlated to floral phase 12 or later. Which would imply that core represents the period after the humid shift in the early Carnian recognized by Hochuli and Vigran (2010).

Paleosols with pedogenic calcite together with coals and other features of waterlogged soils has been described in detail from the Carboniferous in USA and Canada (Tandon and Gibling, 1994, Driese and Ober, 2005, Rosenau et al., 2013a). Their co-existence has been attributed to contrasting climates during glacio-eustatic lowstands and highstands, as response to Milankovitch cyclicity driven glaciations and deglaciations of Gondwana (Rosenau et al., 2013b). Although climate variations at the scale of Milancovitch cycles have been identified in the Late Triassic, the ice free Triassic world did not experience eustatic variations on a similar scale as in the Paleozoic ice-house world (Preto et al., 2010). The Triassic in Svalbard and the Barents Sea is divided into several 2nd and 3rd order sequences (Mørk and Smelror, 2001), but these cycles would not be on the scales that are likely to be contained within single paleosol profiles. Therefore the polygenetic paleosols described in this study is suggested to have formed by autocyclic processes, in a climate favorable for both the formation of coal and pedogenic calcite.

Vertisols requires seasonal variations in the soil moisture in order to form. This does not necessarily mean seasonality in the precipitation evaporation budget, as this can be facilitated by seasonal flooding reflecting hinterland seasonality (Buol et al., 2011). As some of the vertisols contain calcite nodules, it is probable that the moisture variations was more pronounced, and related to the precipitation pattern, as precipitation of pedogenic calcite is highly favored in arid climates (Retallack, 2001). Likely to be relevant in this case is the work of Breecker et al. (2009) and Breecker et al. (2013) who propose that the seasonal variations in soil moisture has primary control on calcite precipitation in vertisols. Breecker et al. (2009) found that calcite solubility was at a minimum during the dry season, and that this implies that pedogenic calcite can form in relatively humid climates, but where a dry season occur. This is further confirmed by identification of vertisols with calcite nodules in places with mean annual precipitation up to 1400 mm (Nordt et al., 2006) and in the humid areas of the lower Mississippi Valley (Aslan and Autin, 1998).

The mottled type C paleosols also contain climate sensitive features. Both hematite (Stensland, 2012) and goethite (Bugge, 1989) is reported from these paleosols. These two iron oxides forms under different moisture conditions, where hematite requires relatively arid conditions and goethite precipitation is favored in humid settings, and both of these minerals is much more likely to form in warm climates, as opposed to more temperate climates (Collinson, 1996). As previously mentioned the mottles in these paleosols was interpreted by Stensland (2012) to be created by a fluctuating groundwater level causing varying redox conditions, something that also is suggestive of a seasonal climate. The oxidized type C paleosols is essentially a form of what is loosely defined as red beds (Kraus, 2002). Parrish (1993) mentions the importance of seasonality in developing red beds in fluvial and alluvial environments, and suggested they may form in areas with seasonal rainfall in both overall humid and arid climates.

The question is then weather the seasonal climate was dominantly humid or arid. As mentioned in Section 2.2.5 the depth at which the calcite occur in a soil profile can be correlated to the amount of precipitation. It’s a simple and straightforward method to use, where the only requirements is that the soil is not eroded, and that compaction of the soil is accounted for (Retallack, 2005). Paleosol B2, shown as a log in Figure 5.21, appears to be suitable for this. The top of the paleosol is recognized as a organic rich A and O horizon, whereas the top of the calcic horizon is taken to be below where no calcite nodules can be found, occurring 160 cm below the interpreted top of the profile, shown in Figure 5.26.

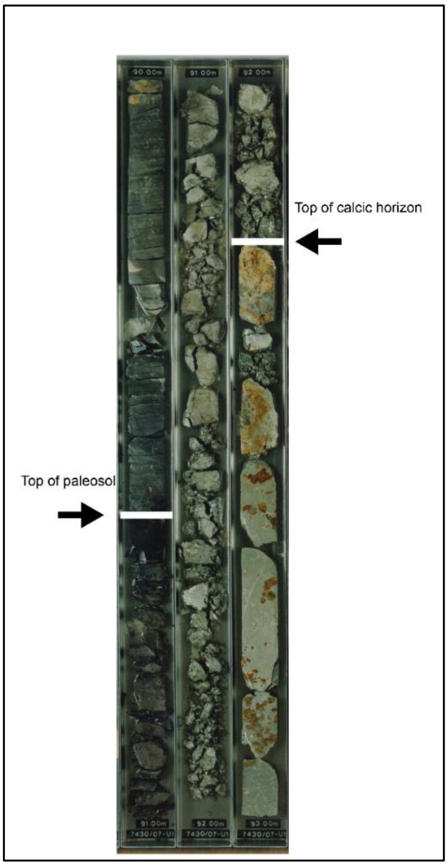

Figure 5.26: *Picture of paleosol B2 from core 7430/07-U-01, showing the interpreted top of the paleosol, and the top of the calcic horizon. From 90-93 m.*

As mentioned in Section 2.2.1 vertisols is believed to undergo significant compaction at the surface, and only minor compaction during burial. Assuming the degree of compaction to be similar to the 93% in the vertisols examined by Caudill et al. (1997), the decompacted depth to the calcic horizon would be somewhere between 1.70 and 1.80 m. From the curve presented in Retallack (2005) and shown in Figure 2.7 this suggests mean annual precipitation in the range between 800 to 1000 mm. Nordt et al. (2006) suggested that the curve from Retallack (2005) underestimates the precipitation in vertisols, and a depth of 1.70 to 1.80 m to the calcic horizons corresponds to mean annual precipitation between 1000-1400 mm, according to their adjustments. Combining the two estimates, leaves a total precipitation estimate in the range between 800 to 1400 mm. There are several uncertainties and limitations related to the use of this method, and one is that the interpreted top of the paleosol does not correspond to what was the surface of the soil as the calcite precipitated. As the B2 paleosol is interpreted to be polygenetic with peat accumulation succeeding calcite precipitation, this might be the case here. Another uncertainty is that atmospheric CO_2 levels influence the depth of calcic horizon (Kraus, 1999). Hence, the mentioned precipitation estimates should be regarded as highly uncertain. It can, however, to some degree be correlated to estimates based on clay mineralogy. Stensland (2012) interpreted the climate to have been humid, based on the kaolinite dominated clay mineral assemblage. Kaolinite is usually a dominant clay mineral in areas with a mean annual rainfall between 1000 and 2000 mm (Retallack, 2001). This, and the fact that the core most likely can be correlated to the hygrophytic dominated floral phase of Hochuli and Vigran (2010) suggests that the climate was dominantly humid.. Another strong evidence for a humid climate is that nearly all the features of the paleosols, is reported from soils in the humid areas of Louisiana in southern USA (see Section 6.3). The prevailing climate during the early-middle Carnian is therefore suggested to have been dominantly humid, but with a seasonal distribution of rainfall.

6 Discussion

6.1 Paleosols and depositional environment

The interpretation of the overall depositional environment in the investigated parts of the De Geerdalen Formation is that of a paralic environment with a strong deltaic signature. This is based on the presence of channel sandstones, coarsening upwards delta front sandstones with unidirectional current indicators, and intervals with marine influence. Indication of both wave, storm and tidal influence can be seen in the sediments. This interpretation is concordant with other studies on the De Geerdalen Formation on Edgeøya and Hopen, based on much more extensive data, and more focus on the overall depositional environment than what is presented here (e.g. Mørk et al., 1982, Klausen and Mørk, 2014, Rød et al., 2014, Lord et al., 2014b). As the logs through the successions on both Edgeøya and Hopen are far from complete, interpretation of the development or differences between the areas will not be made.

The Snadd Formation on the Bjarmeland platform has as previously mentioned, been interpreted to represent a delta plain depositional environment, with an upwards transition from tidally influenced lower delta plain to upper delta plain by Stensland (2012). Paleosols in the core appear on top of coarsening upwards delta lobe sandstones, within more mud dominated floodplain deposits, and on top of more massive sandstones, possibly representing crevasse splays or channels. Their depositional context seems to be comparable to the paleosols seen in the De Geerdalen Formation, which have been found on top of delta front sandstones, channel sandstones and within mud dominated floodplain deposits. The 110 m interval of the Snadd Formation represented by core 7430/07-U-01 can be seen to contain far more paleosols than the investigated parts of the De Geerdalen Formation. This suggest a more proximal landward position of the Bjarmeland Platform area than Svalbard, which would be concordant with the prevailing view of a northwesterly prograding coastline (Riis et al., 2008, Høy and Lundschien, 2011, Lundschien et al., 2014). However, some of the paleosols in the core are detected on the basis of subtle features that can be hard to identify in field outcrops, meaning that paleosols can easily be overlooked. Paleosols and root horizons also occur frequently in the upper parts of the De Geerdalen Formation below the Hopen Member on Hopen (Mørk et al., 2013, Klausen and Mørk, 2014).

Differences in the detail level of the data makes it difficult to directly compare the paleosols from the Snadd and the De Geerdalen formation. As previously mentioned paleosols in the field are often covered by scree and consist of highly unconsolidated material making detailed observations hard. Features such as pedogenic slickensides has not been identified in the De Geerdalen Formation. As the field logging was done prior to the core logging, these features was not anticipated to be found, meaning that they might be present, but simply overlooked. It is also possible that the slickensides is not preserved on Svalbard, due to a different burial or diagenetic evolution, or alternatively that weathering caused by recent exposure has destroyed or partially destroyed them in the near surface sediments. The mottled pattern in the type C paleosols in the Snadd Formation has not been identified either. However, it is possible that the grey and reddish brown color seen in some of the paleosols is related to differences in Iron distribution or the oxidation state of Iron, which means they could be related to a similar phenomenon as the mottles. As previously mentioned this could possibly also be related to modern weathering of the outcrop. These are only speculations, and further analyses of these horizons are needed in order to conclude how they are formed, and what they consist of.

Stensland (2012) proposed that the coexistence of paleosols showing signs of oxidation and waterlogging was controlled by channel position, and position on the delta plain. This notion is supported here, and it is suggested that the concepts mentioned earlier in Section 2.3 can be used to explain some of the variations seen in the paleosols. As these concepts are based on fully alluvial deposits, some care should be taken in applying them to deltaic and paralic environments. The upper delta plain is largely affected by similar processes as in alluvial environment, with the exception that swamps, marshes and lakes are more abundant (Reading and Collinson, 1996). The difference that has to be considered is therefore the differences in processes from the upper to the lower delta plain. The lower delta plain is affected by basinal processes, whereas the upper delta plain is by definition unaffected (Reading and Collinson, 1996). Distributary channels experience more frequent switching and avulsion than fluvial channels, and crevassing is also more frequent on the lower delta plain then on the upper delta plain, as distributary channels tend to have less confining levees (Diessel, 1992). The lower parts of delta also tend to have more fine grained material with lower permeability, in addition to lower gradients leaving the water table closer the surface and making waterlogging more widespread (Diessel, 1992, Hartley et al., 2013). This imply that interdistributary areas on a delta plain in theory may display variations in factors such as drainage state and sedimentation rate from a proximal upper delta plain position to the distal lower delta plain.

Floodplains in paralic environments are also sensitive to sea level changes, where a rising sea level may lead to increased sedimentation on the floodplain as accommodation space is created, whereas sea level fall may lead to channel incision and sediment starvation on the floodplain (Kraus, 1999). Subsequently the soil development on deltas is influenced by both position relative to channels, position relative to the coastline, and sea level variations. This implies that variations in soil properties in paralic and deltaic environments should be expected to be more complex than in alluvial environments. Still, some attempt to relate the different paleosols to sub-environment or position relative to channels and coastline will be made.

Type B and C paleosols from core 7430/07-U-01 are both suggested to have formed on the upper delta plain floodplain. Their coexistence is suggested to mainly represent position relative to channels. Their varying characters can to some degree be explained by the topographical expression of alluvial ridges, which is the main controlling factor in creating different depth to groundwater level (see Figure 2.8) (Kraus and Aslan, 1999). Pedogenic slickensides, calcite nodules and the mottling pattern is suggested to have been formed in response to a zone of fluctuating water level. The immature type A paleosols is believed to be related to crevasse splays, and are suggested to have formed closest to channels. The mottled pattern in the type C paleosols is also proposed to have formed in proximity to channels, where well oxygenated river waters led to oxidizing conditions. Channel proximity is also suggested to be the reason why they are seen to be partly eroded. The polygenetic type B paleosols are suggested to have formed further away from the channel, and to have evolved from a distal floodplain to permanently waterlogged swamp. Figure 6.1 show the suggested lateral relationship between the various features, in relation to channel distance and water table fluctuations, and is partly based on the paleocatena concept of Bown and Kraus (1987) and the pedofacies concepts of Kraus and Aslan (1999).

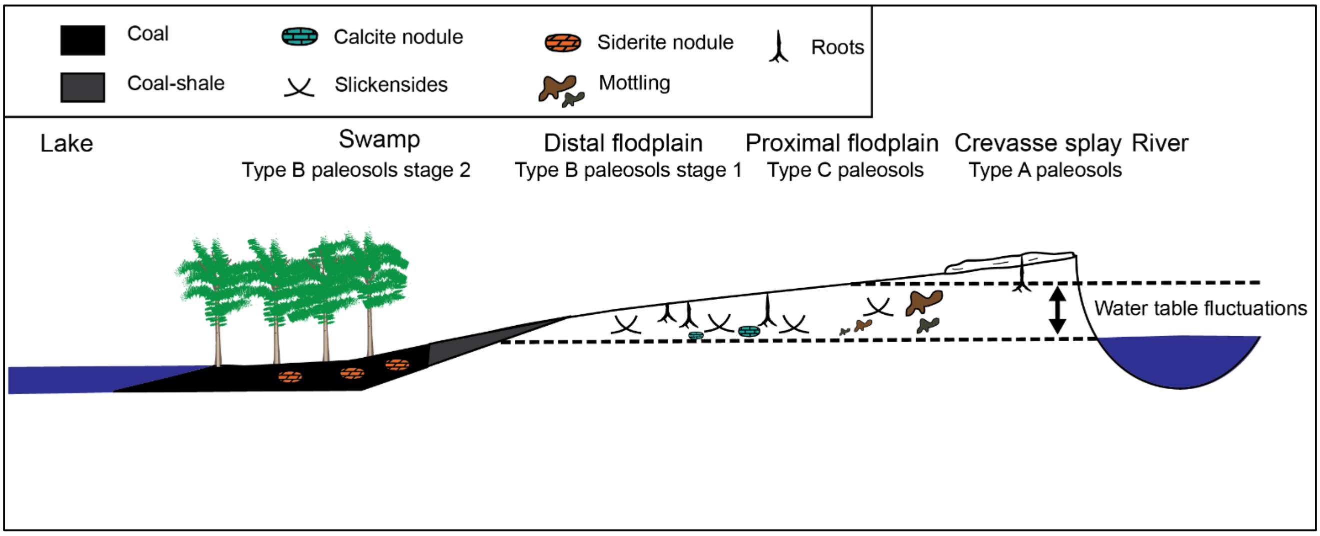

Fig 6.1: *Proposed lateral relationship between the various paleosols from core 7430/07-U-01, as a function of distance to channel. Partly based on the paleocatena*

As mentioned these variation are likely to be more complex in deltaic environments than in alluvial environment, and some of the factors such as tendency to waterlogging could also change from a proximal to distal position on the delta plain. This could for instance mean that the oxidized type C paleosols represent a more proximal position than the type B paleosols. The fact the Type C paleosols are confined to upper parts of the core could possibly support this, as that would be compatible with a prograding delta. As the position of the paleo-coastline in the Snadd and De Geerdalen depositional systems at times were positioned in the Svalbard area (e.g. Lundschien et al. 2014), paleosols on the Bjarmeland Platform may be surfaces that evolved 100s of km behind the coastline.

The type A paleosols from the Snadd Formation, is in the lower part of core 7430/7-U-01 situated on top of coarsening upwards sandstone units. These are considered to be equivalent to the paleosols found on Blanknuten and Hopen, also occurring on top of coarsening upwards sandstone units, interpreted as delta front deposits. These have been suggested to represent lower delta plain marshes and vegetated surfaces. As stated in section 5.1.3 these paleosols can be regarded as abandonment markers, where the abandonment could be either local or involving the whole delta. Delta sequences involving the whole delta are commonly divided in a prograding constructional phase, and a transgressive destruction phase following delta switching and termination of sediment supply (Reading and Collinson, 1996). As a delta complex grows, the area of inactive delta lobes increases, and modern deltas are often characterized by large areas that are not part of the active system (Diessel, 1992). As mentioned in the introduction, repeated coarsening upwards sequences in the De Geerdalen Formation has been attributed to delta switching (Mørk et al., 1982). Some of these paleosols therefore probably represents abandoned parts of the delta, where soil formation took place before the surface was drowned by the ensuing transgression. On Hopen and in core 7430/07-U-01 where these paleosols has been studied in most detail, they appear to be immature. The paleosols found on top of channel sandstones on Negerfjellet and Tjuvfjordhorga probably also represents inactive surfaces on the delta, that where established after avulsion and abandonment of the channel. These channels have been interpreted as distributary channels based on their geometries, and probably also represent the lower delta environment.

Figure 6.2 shows a schematic model of distribution of soils relative to channels and the coastline. The lower delta plain is suggested to be dominated by immature soils, as proximity to marine waters, and possibly more sediment input would inhibit more mature soils to form. Peat the form of histosols is here suggested to have formed in coastal marshes in areas removed from active channels, either in interdistributary areas or on abandoned parts of the delta. The upper delta plain is suggested to consist of histosols forming in freshwater swamps, vertisols and oxidized soils. These are suggested to show lateral transitions as displayed in Figure 6.1.

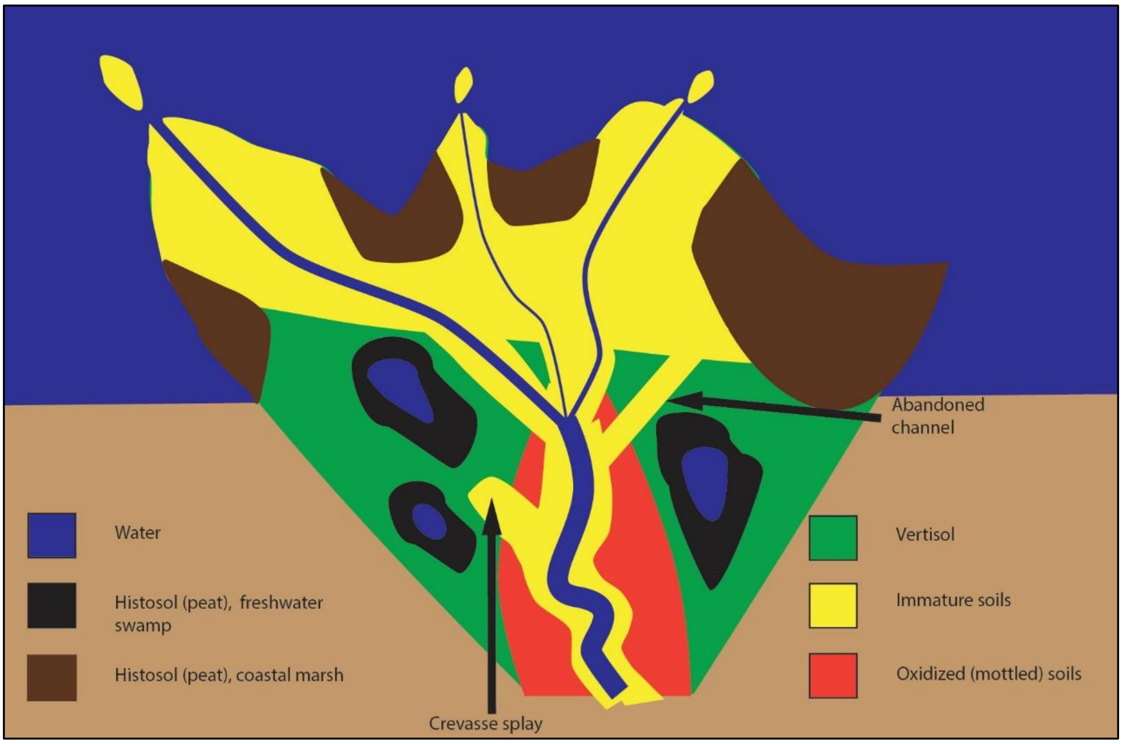

Fig 6.2: *Schematic model, with suggested distribution of soil types.*

6.1.1 Peat accumulation

Coal and coal-shale have been found to mostly occur as organic accumulation on top of paleosols. Coals that are seemingly unrelated to paleosols have also been found, and are assumed to have formed as floating peat or detrital accumulations of organic material. Such coal seams was also noted by Hynne (2010) in the De Geerdalen Formation, who also suggested floating peat as a potential explanation for their origin.

In the Snadd Formation the thickest coal is found to be approximately 70 cm, while the coals in the de Geerdalen Formation shows thicknesses up to 50 cm. This agrees with well with observations from Klausen and Mørk (2014) who mentions coal seams up to 50 cm thick in the De Geerdalen Formation on Hopen, and Lundschien et al. (2014) who reports coals close to 1 m thick from the Snadd Formation on Sentralbanken. Some of the coals show signs of marine influence, and are found in association with paleosols interpreted to represent the lower delta plain. Basin proximity in the lower delta plain leads to more saline conditions in the soil forming environment, and in modern coastal peatlands highly salt tolerant mangroves are usually the dominant plant life (Diessel, 1992). There is not much knowledge about the salt tolerance of the Carnian flora from Svalbard and the Barents Sea, but is assumed to have been low (N. Patterson, pers. Comm. Mars 2015). This means that inundation by marine waters may have been an effective way of terminating plant growth and peat accumulation. Coal related to the paleosols on the upper delta plain is suggested to represent freshwater swamps, with a sufficient distance from sediment sources to allow for peat accumulation. These could, however also represent abandoned parts of the delta. Upstream avulsion of channels is also proposed as a possible explanation for the transition from stage 1 to the peat forming stage 2 in the polygenetic type B paleosols. McCabe (1984) suggests that a possible explanation for the presence of coal within clastic sequences, is that the peat did not form contemporary with local clastic deposition. This is supported by the fact that peat forming coals in modern deltas appear to mostly form on the inactive parts (McCabe, 1984) (see Section 6.3). This could mean that much of the coal in the Snadd and De Geerdalen formations represent inactive surfaces on the delta.

Usually distinction between coal formed on the upper and lower delta plain is made based on their thickness, where upper delta plain coals are thickest and often more lateral extencive. (Fielding, 1985). However the opposite relationship, with thicker coal developing on the lower delta plain has also been found (Diessel, 1992). No clear pattern in thickness variations between upper and lower delta plain coals has been found in this study, and data on lateral extent of coal seams are limited.

In order for thick coal seams to form it is important that the factors favoring peat accumulation such as organic productivity and high water table are maintained over long time periods (Diessel, 1992). Klausen and Mørk (2014) suggests that the reason for absence of thick coal seams in the De Geerdalen Formation was that the coal forming environment was subjected to sediment influx and inundation of marine waters. This is supported by the fact that many of the coal seams often show gradual transitions to less organic rich facies, and has laminas of mud within them. Such features are typical where a sediment source such as a river gradually approaches the peat forming environments (Diessel, 1992). In addition to sediment influx and marine inundation the climate also could have been a factor in inhibiting formation of thick coal seams. Although not necessary inflicting the plant community, the seasonality in the precipitation might have had impact on preservation and accumulation of peat. Formation of peat in modern settings is strongly curtailed by even short periods of drought and seasons with reduced precipitation, and where temperature is not a limiting factor, a dry season of as little as one month leads to groundwater fluctuations causing oxidization of organic material

(Lottes and Ziegler, 1994). It is therefore possible that the uneven distribution of rainfall hindered the development of thick raised swamps, which usually requires precipitation rates above 2500 mm with even distribution through the year in order to form (Retallack, 2001). However, as the amount of compaction involved in the peat to coal transformation is uncertain (see Section 2.2.2), the original thickness of the peat is hard to assess. What can to some degree be estimated is the time it took to accumulate the peat. The TOC values presented in Section 5.2.1 varies between 34.6 and 65.3%. According to Large and Marshall (2014) 50 cm thick coal seams with a carbon fraction of 0.6 accumulated at 40-60^{o} latitude would require approximately 10000 to 20000 years to form. This time estimate will only serve as a minimum, as the coal seams may have hiatuses or pauses in peat accumulation within them (Large and Marshall, 2014). If the rate of peat accumulation was influenced by the precipitation seasonality, the time it took to form the peat would probably be higher. There are also some limitations to this method. It does not consider type of vegetation, which has an impact on accumulation rates (Retallack, 2001), nor does it consider the effect that high CO_2 levels in greenhouse periods such as the Triassic would have on the carbon accumulation rates.

As it is otherwise hard to estimate the amount of time that paleosols represent (see Section 2.1) it is also suggested that this time estimate, or at least the order of magnitude it represents could serve as a approximation to the time it took to form some the paleosols. The Pedogenic features such as mottling, slickensides and carbonate nodules would all be able to form within a time interval of 10^4 years (see Figure 2.2).

6.2 Late Triassic Paleoclimate on Svalbard and in the Barents Sea

Since some of the climate sensitive features has not been identified in the field, the paleoclimate interpretation made from core 7430/07-U-01 cannot be directly extrapolated to the investigated parts of the De Geerdalen Formation. The fact that coal is present on both Edgeøya and Hopen, and as previously mentioned, the late Carnian palynological assemblages from Hopen has been compared to the humid floral phases (Paterson and Mangerud, 2015) suggest a similar climate prevailed here as well.

A seasonal climate may be viewed as conflicting with the presence of a hygrophytic plant community. The plants that grew on the delta, including trees believed to have had a lifespan in the order of decades, probably had low drought tolerance (N. Patterson, pers. Comm., Mars 2015). Hochuli and Vigran (2010) also notes that they favor a warm temperate climate with minimal seasonality. The answer to this apparent contradiction might be that the plant life was restricted to low lying areas with proximity to the ground water table. This would however, imply that the vegetative cover of the delta was somehow limited. The fact that palynological samples are taken from different lithologies (Paterson and Mangerud, 2015), abundant plant fragments in the sediments, and presence of tree trunks several places is indicative of a more widespread vegetation. Launis et al. (2014) studied the Carnian flora from plant fossils collected from the De Geerdalen Formation on Hopen, and found a diverse range of species, something which also points towards a widespread vegetation. The explanation might therefore be that the dry season was rather short, and that the plant life was able to withstand the drought. Distinction between semi-arid, sub-humid and humid climates depends on the length of the dry season. A short dry season would mean the climate could be defined as either humid or moist sub-humid, after the definition of Cecil et al. (2003), displayed in Figure 2.3.

The interpretation of a humid climate, with a dry season seems to agree well with the *General Circulation Model* by Sellwood and Valdes (2006), mentioned in Section 1.5 and shown in Figure 1.5. Their model also predicts a seasonality, where it suggests that the rainfall was more concentrated in the winter season. The predicted mean annual precipitation between 1000 and 2000 mm from the model also agrees well with the estimates mentioned in Section 5.2.3. It is therefore argued that this model holds validity in its proposed paleoclimate in the Svalbard and Barents Shelf area in the Late Triassic.

The paleosol assemblage can be compared with what has been reported from the North Sea. Based on changes in alluvial architecture, paleosols and clay minerals, the Late Triassic to Early Jurassic climate of the northern North Sea region was investigated by Nystuen et al. (2014). A gradual shift from an arid climate in the Norian, to a humid climate in the Early Jurassic was argued, and attributed to northwards tectonic drift from 40^{o}N to 50^{o}N, and global climatic changes at the Triassic-Jurassic boundary. The Early Jurassic alluvial sediments of the Statfjord Group was found to be dominated by grey colored (low chroma) vertisols, with a smectite and kaolinite dominated clay mineral assemblage. Grey colored vertisols also seems to be the dominating paleosol in the floodplain deposits of the Snadd Formation. From this, it may be argued that the climatic conditions in Barents Sea in the Carnian, to some degree was comparable to that of the North Sea in Early Jurassic.

Ryseth (2014) proposed that the Triassic-Jurassic boundary in the Barents Sea saw the transition towards increasing humidity, in a similar manner as the North Sea. A similar proposition was also made by Bergan and Knarud (1993) in order to explain some of the mineralogical and sedimentological changes from the Storfjorden Subgroup to the Wilhelmøya and Realgrunnen subgroups. Even though a relatively humid climate during deposition of the Storfjorden Subgroup has been argued here, an important aspect of this climatic change could be a disappearance of the seasonal variation in precipitation to a more equable humid climate.

As noted in Section 1.5 the presence of mature calcretes, red beds, and only thin coal seams in the upper parts of the De Geerdalen Formation on Spitsbergen has led to suggestions that the climate was variable and at times semi-arid. (Knutsen, 2013, Husteli et al., 2015, Olaussen et al., 2015). Olaussen et al. (2015) also notes that similar features can be found in upper parts of the Snadd Formation in the southern Barents Sea. These parts of the succession has not been studied in the present work, but appear to be somewhat different than the early-middle Carnian parts of the Snadd Formation and the parts of the De Geerdalen Formation that has been logged. The amount of pedogenic calcite found in the Snadd Formation is relatively small, at least not enough to classify the paleosols as mature calcretes, and coal is abundant. No pedogenic calcite has been found in the De Geerdalen Formation. This might be an indicator that there was a climate shift towards a more arid climate in the latest Carnian to Norian, which would imply that the evolution of the Late Triassic climate in this region was more complex. A shift towards a cooler and more arid conditions in the latest Carnian has also been proposed to be a possible explanation for the change towards a more conifer dominated composition of the palynological assemblages seen in the Hopen Member (Lord et al., 2014a). A similar development is documented in Central Europe, where a semi-arid climate in the late Carnian followed the Carnian Pluvial Event (Preto et al., 2010). However, Hochuli and Vigran (2010) argues that the humid climate established in the early Carnian lasted throughout the Triassic. The claim that the climate was variable is, nevertheless, not unreasonable considering the scale in both time and space that the Snadd and De Geerdalen formations span. These formations represents a depositional system that covered an immense area. It also represents approximately 20 to 25 Ma of time (Ryseth, 2014). Variations in regional or local climate on such scales cannot be considered unlikely.

The Snadd and De Geerdalen deltaic system has been interpreted as a supply dominated system, where a high sediment yield and shallow water depth caused a high progradation potential (Klausen et al., 2015). A seasonal variation in precipitation might have been a contributing factor in facilitating a high sediment yield in the fluvial feeder system. The reason for this is that areas with seasonal variation in precipitation tend to have higher erosion rates and sediment yield than areas with non-seasonal climates (Reading and Levell, 1996, Cecil and Dulong, 2003, Cecil et al., 2003). The reason for this relationship is that seasonal precipitation are usually associated with more intense rainfall, causing more erosion and runoff (Hooke, 2000). In addition vegetation that serves to stabilize landscapes, are more limited where seasonality is pronounced (Hooke, 2000). However, the climate alone is not determining the magnitude of sediment yield. Sediment flux from the continents to the sea through rivers is also strongly influenced by factors such as area of drainage basin, relief, and bedrock lithology (Hooke, 2000), meaning that the role of the climate is not necessarily significant.

6.2.2 Uncertainties in paleoclimate interpretation

As mentioned in section 2.1, the main reason that paleosols are used as paleoclimate indicators is that they serve as proxies for the moisture regime. The major uncertainty in this assumption is that factors such as topography and landscape position can obscure the relationship between moisture regime and climate (Retallack, 2001). This is particularly true in low lying coastal areas and river valleys where surfaces are close to the groundwater level. Kraus (1999) states that paleosols in floodplain environments may be better suited to give information on floodplain sedimentation and hydrology, than on paleoclimate. Major discrepancies between soil moisture and climate can be seen in extensive swamplands in arid regions, such as in Sudan along the Nile River, on the Nile Delta, and in Iraq adjacent to the Eufrat and Tigris rivers (McCabe, 1984). This basically means that soils are generally wetter in alluvial valleys, deltas and coastal areas, than on other landscapes. This concept may be quite obvious, but still needs to be considered when interpreting paleoclimate from paleosols. The conclusion that can be drawn from this, is that humidity is more likely to be overestimated than underestimated when interpreting paleoclimates from paleosols in such environments. The Snadd Formation delta plain deposits has been interpreted to represent a low gradient delta plain (Klausen et al., 2015). It is therefore possible that the paleosols does not reflect the ambient climate they formed in, and that the indicators of a humid climate is merely a reflection of a low lying, low gradient, “wet” delta plain, in which case a more arid climate could be argued.

6.3 Modern analogue: The Mississippi Delta

The present day Mississippi delta located in Louisiana in southern USA is a river dominated delta, forming in a warm temperate humid climate, at 30^{o} N (Aslan and Autin, 1998, Reading and Collinson, 1996). The mean annual precipitation is around 1500 mm and is somewhat unevenly distributed with rainfall mainly concentrated in the winter and spring (Aslan and Autin, 1998). Information on soil distribution in the Mississippi Delta area has been retrieved using a map by the USDA Natural Resource Conservation Service, shown in Figure 6.3.

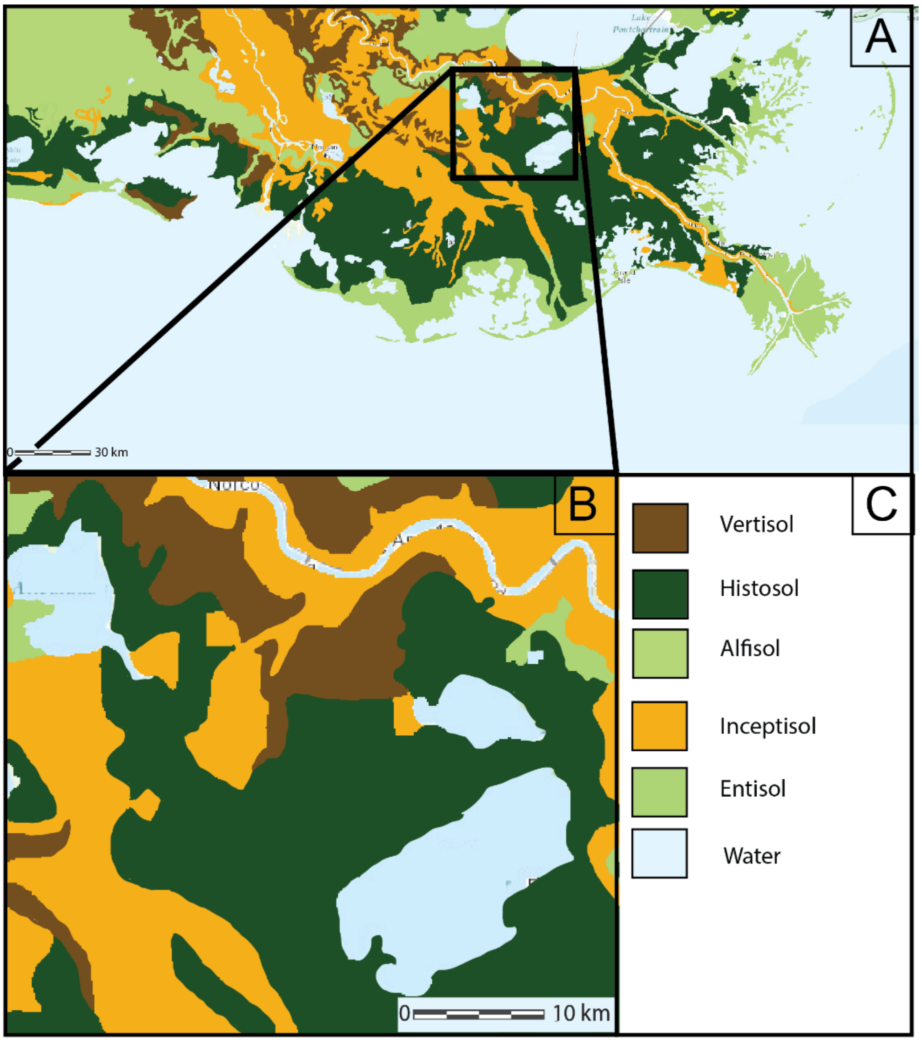

Figure 6.3: A) *Distribution of soils on the modern Mississippi delta in Louisiana, USA. Map from USDA Natural Resource Conservation Service (2010).* **B)** *Lateral variations in soil type on the upper delta plain next to the Mississippi River.* **C)** *Legend to the maps in A and B.*

As can be seen from the map in Figure 6.3 A the Mississippi delta is dominated by immature entisols on the delta front mouth bars, on the levees of distributary channels, and on barrier bars forming in front of abandoned parts of the delta. Along active channels are immature inceptisols, which can also be found along the abandoned channel systems and on lobate features extending from the channels, probably representing crevasse splays.

Peat deposits in the form of histosols are forming in small coastal basin between distributary channels, and in broader more inland positioned flood basins with freshwater forested swamps (Frazier and Osanik, 1969, Reading and Collinson, 1996). True coal forming peats however, is believed to only form on the inactive parts of the delta, as flooding rivers are diluting the peat with clastic sediments in the active parts (McCabe, 1984). At present the inactive parts of delta surface constitute approximately 75% of the delta area (Diessel, 1992). The thickest peat deposits are those accumulating in freshwater swamps on the upper delta plain (Frazier and Osanik, 1969). As seen on Figure 6.3A the lower delta plain histosols cover is extensive. Vertisols are locally present along the Mississippi River on the upper delta plain, but are more abundant further up in the Mississippi Valley, where they are also seen to locally contain calcite nodules (Aslan and Autin, 1998). Seasonal differences in precipitation, evapotranspiration and river stage are causing the moisture variations leading to creation of slickensides (Aslan and Autin, 1998). Siderite also occur as pedogenic carbonate in swamp histosols (Aslan and Autin, 1998). Soils with a mottled pattern of brown, grey and yellow colors, formed by a fluctuating water table have also been described from floodplain soils in the lower parts of the Mississippi Valley (Aslan and Autin, 1998).

Lateral variations in soil types as a function of distance to channels can also be seen. Shown in Figure 6.3B is a lateral transition from inceptisols near the river, to a band of vertisols and histosols forming along lakes. Worth noting is that the scale at which these transition takes place is over several kilometers.

It appears that the Mississippi Delta and the lower Mississippi Valley floodplain can be considered as a good analogue for the Snadd and De Geerdalen delta plain deposits, in terms of soil distribution and properties. The similarities include immature soils on exposed delta front sediments, vertisols with calcite nodules in placesand abundant histosols. The mottled soils could potentially be similar to the type C paleosols.There are several potential reasons for the similarities. Aslan and Autin (1998) mentions that the soils in the lower Mississippi Valley not necessarily reflect the climate in the area, and that the depositional and hydrological processes are more important in determining soil properties. The Mississippi River is characterized by a yearly flood where the discharge is up to four times higher during flooding, than average normal discharge (Roberts, 1997). This annual discharge variation probably amplifies groundwater and moisture variations caused by seasonality in rainfall. As mentioned in the introduction the nature of the parent material such as grainsize and mineralogy also inflicts the type of soils that will form. Meaning that similarities in e.g. the mineralogy of the sediments could be a partial reason.

Even though hydrological and depositional processes influence soil development, the climate is usually considered to be the most important factor controlling soil development (Cecil and Dulong, 2003). Worth noting is that the mean annual precipitation of 1500 mm in the area is comparable to the estimates from Section 5.2.3 and the precipitation rate proposed by the *General Circulation Model* of Sellwood and Valdes (2006). Monthly temperatures in Louisiana range from 3 to 33^{o} C with an average of about 19^{o}C (Aslan and Autin, 1998), which is also comparable to the temperature predictions of Sellwood and Valdes (2006), which range from around 10^{o} C at winter and around 28^{o} at summer. It is therefore suggested that the climate of the Mississippi Delta and the lower Mississippi Valley can be considered as a possible analogue to the early-middle Carnian climate in Svalbard and the Barents Sea. A comparison of the Late Triassic climate at 50-55^{o}N with the modern climate at 30^{o} N should not be seen as unreasonable, as the Triassic climate was warmer and less defined by latitudes than the more zonal climate today (Preto et al., 2010). Similar discrepancy between paleoclimate and modern climate in terms of latitudes has been proposed by Retallack and Alonso-Zarza (1998) who noted that Middle Triassic paleosols on Antarctica, formed at 70^{o} S paleolatitude, resembles modern soils from southern Sweden at 55-60^{o} N.

It should be mentioned that many other modern deltas exhibit features such as vertisols and histosols, meaning that the Mississippi Delta is not necessarily unique with respect to its soils. Other modern deltas associated with vertisols include the Paraná Delta in Argentina (Imbellone et al., 2009), Kızılırmak Delta in Turkey (Dengiz et al., 2012), and the Volga Delta in Russia and Kazakhstan (Khitrov and Rogovneva, 2014).

7 Conclusions

- Paleosols in the De Geerdalen Formation have been identified on top of delta front sandstones, on top of channel sandstones, and within mud dominated interdistributary delta plain deposits, in a paralic and deltaic depositional environment. Paleosols in the Snadd Formation have been found to occur in a similar depositional context.
- Paleosols on top of delta front and channel deposits can be viewed as abandonment surfaces, marking a period of non-deposition after termination of sediment supply, before the surface was drowned by transgression, or buried by renewed sedimentation. These are suggested to mainly represent immature soils, with peat accumulation in coastal swamps situated in interdistributary areas, and on inactive surfaces of the delta.
- Floodplain deposits in the Snadd Formation contain an abundance of mudrocks with pedogenic slickensides. These are suggested to have been formed by seasonal wetting and drying cycles in swelling clays on the floodplain, and can be classified as vertisols. Some of these appear to be polygenic, where authigenic processes such as channel migration and avulsions are believed to be responsible for changing the moisture regime.
- The paleosols in the early-middle Carnian parts of the Snadd Formation attest to a dominantly humid climate, but with a seasonal variation in precipitation. The interpretation of a humid climate is is concordant with other studies of the Triassic paleoclimate on Svalbard and in the Barents Sea based on palynology, and supports the validity of *General Circulation Models* for the Late Triassic.
- A seasonal variation in precipitation is likely to have inhibited the development of thick raised, precipitation fed swamps, possibly explaining the lack of thick coal seams. Seasonal variation in precipitation could also have favored a high sediment yield in the fluvial system.
- The features of the paleosols appear to be comparable to modern soils developing on the Mississippi Delta and in Mississippi River Valley. Reasons for this could be similarities in factors such as sedimentary process on the delta plain, hydrological conditions and properties of the fluvial system, and climatic conditions.

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Appendix

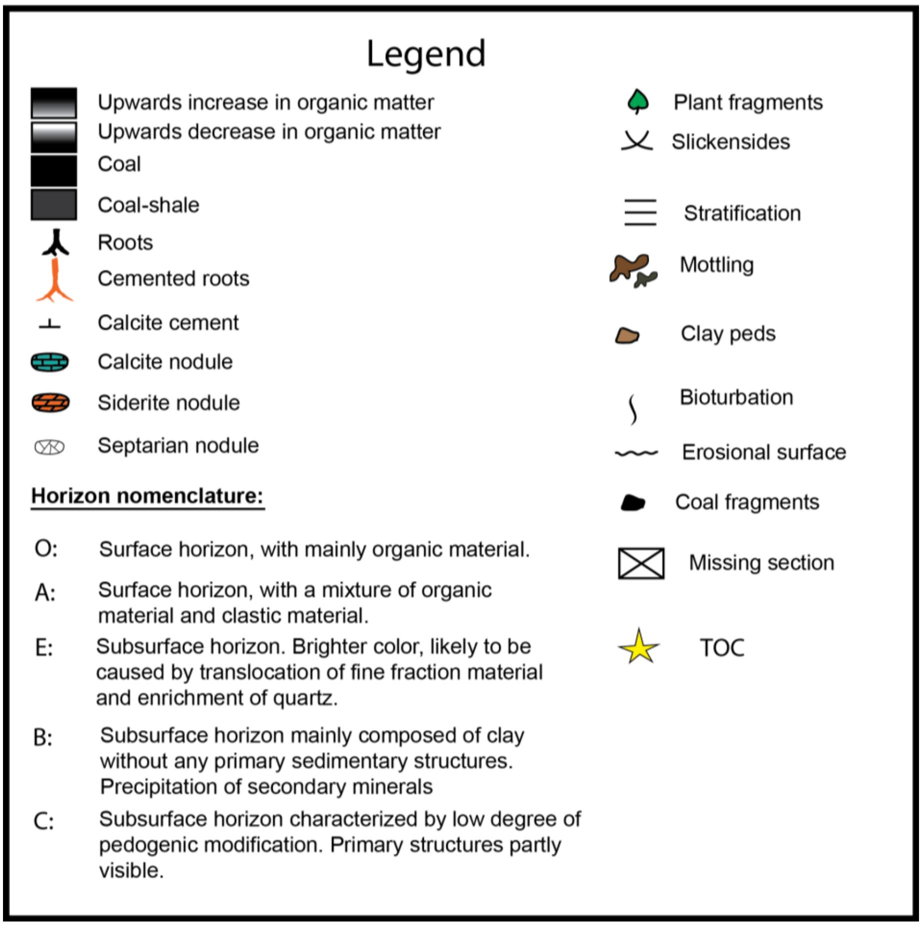
Appendix I

GPS coordinates from measured sections:

Location/Log	UTM coordinates	Elevation, start of log (m.a.s.l.)
Tjuvfjordhorga Log 1	33X E688391 N8601399	220
Log 4	33X E688505 N8601817	224
Negerfjellet	33X E685796 N8593214	160
Blanknuten	33X E645305 N8663225	276
Hopen Log 1	33X E762186 N8519053	0
Log 9	33X E762342 N8520418	0

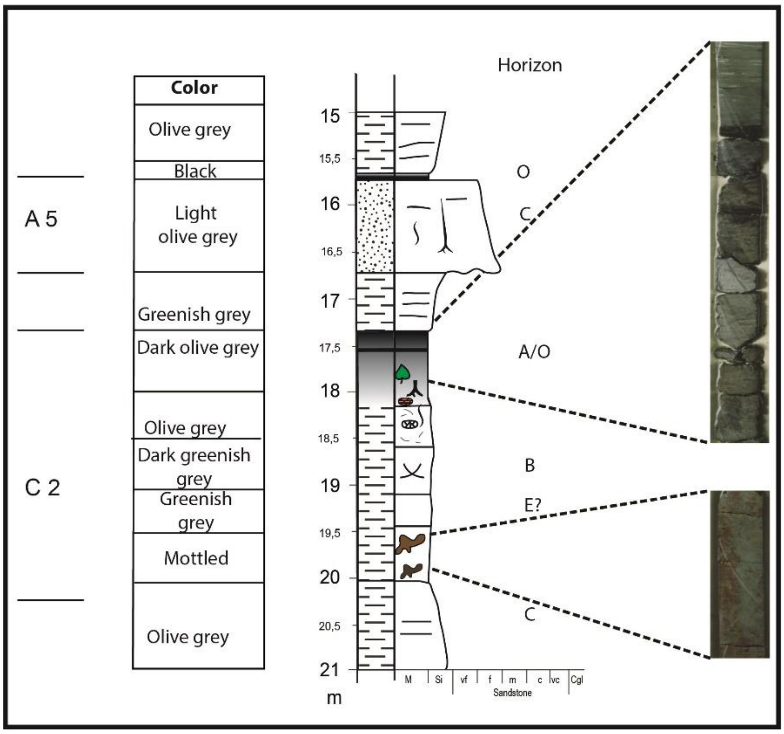
Appendix II

Legend to logs from core 7430/07-U-01:



Appendix III

Logs from core 7430/07-U-01



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Dark olive grey
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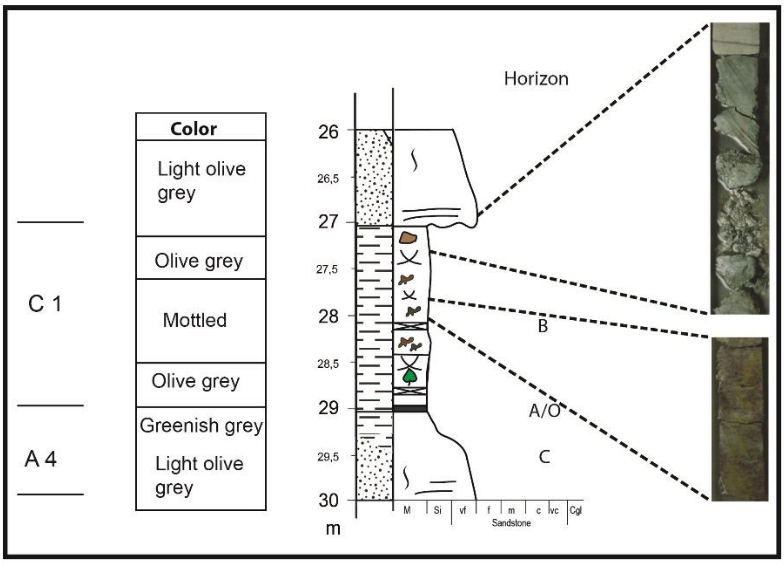

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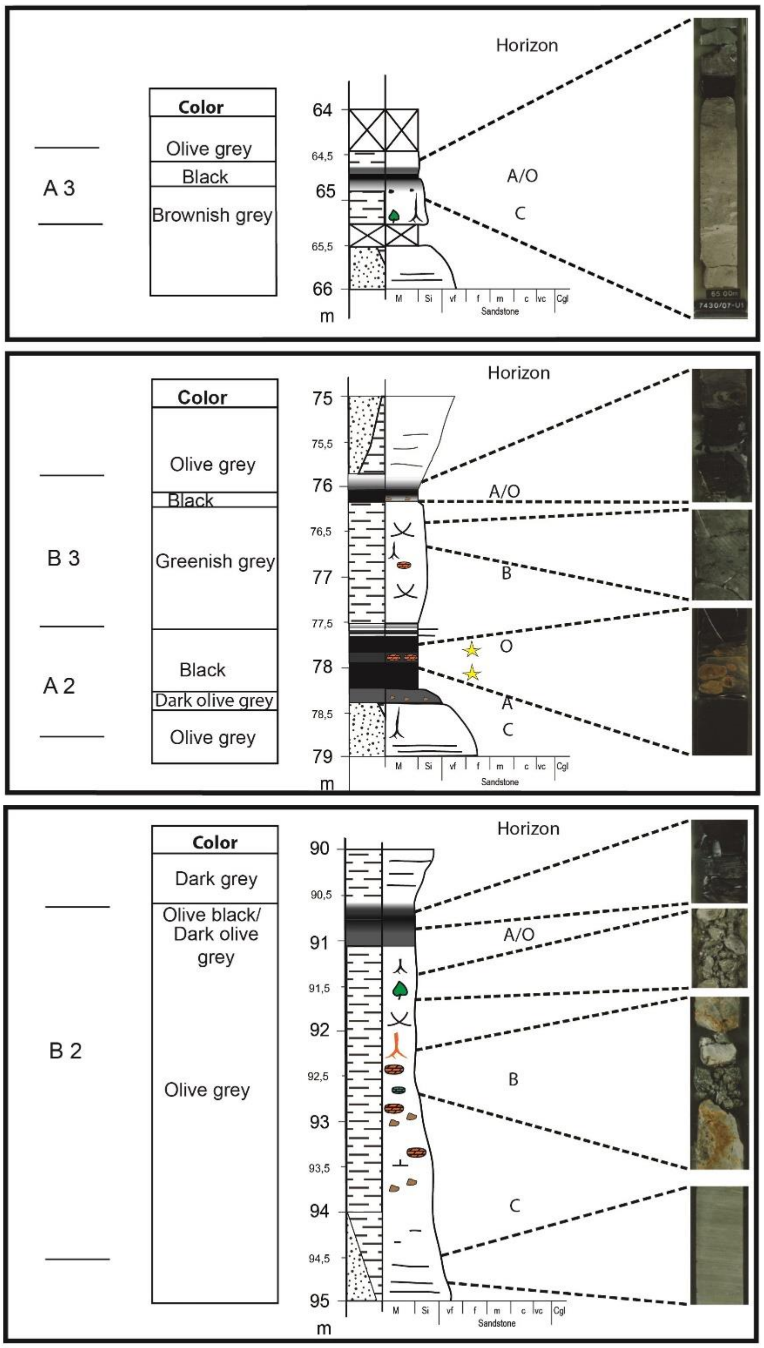
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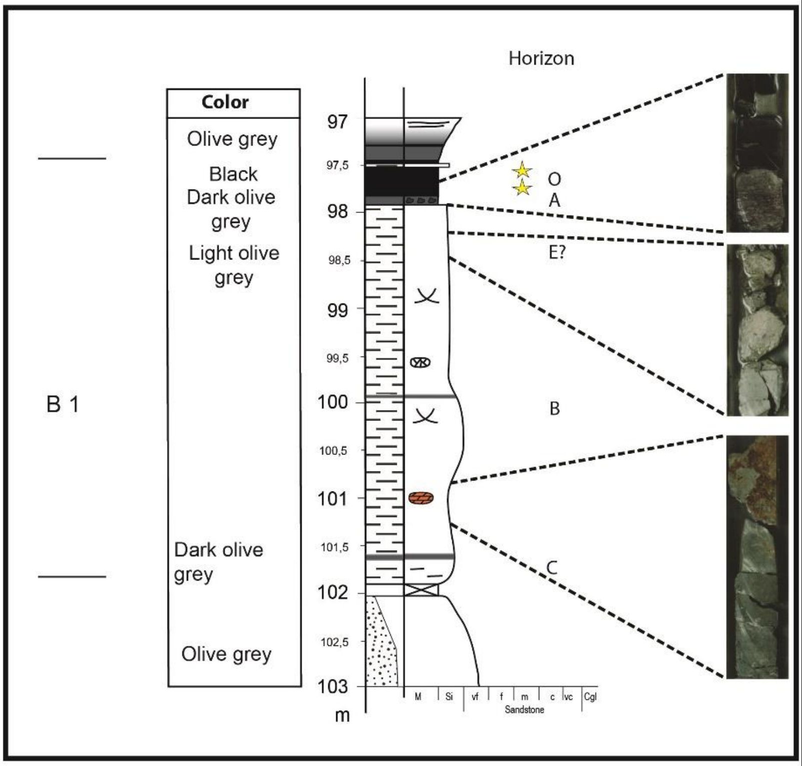
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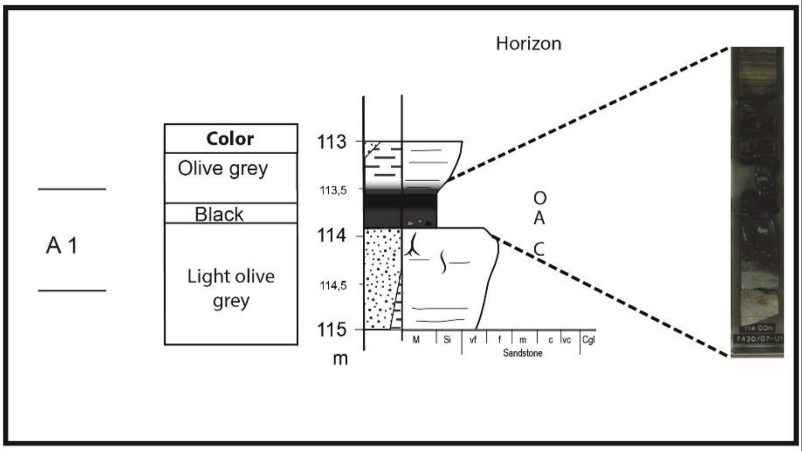
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