Facies Architecture and Paleogeography of the Battfjellet Formation, Rypefjellet, Spitsbergen

Masters thesis in petroleum geology

By

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Abstract

The shallow marine sandstones of the Battfjellet Fm are part of a regressive mega-sequence that represent the last stages of infilling of the Paleogene Central Basin in Spitsbergen.
The Battfjellet Fm is believed to be of Eocene age and was deposited in a foreland basin that developed in front of the West Spitsbergen Orogen, a fold-and-thrust belt that formed along the western coast of Svalbard as a response to the northward spreading of the opening of the Atlantic Ocean. The Battfjellet Fm has been interpreted to represent wave-dominated delta deposits that built out into the Central Basin in an easterly direction. The shoreline had a north-south orientation and rivers delivered sediments into the basin from a western source area. Sediments deposited by turbidite currents were deposited on the slope and basin-floor in the western parts of the basin. Hyperpycnal currents, possibly formed as a result of floods have been interpreted as important for the generation of turbidite currents although other processes such as storm-waves and tectonic movements also are capable of generating turbidity currents.

The basin-floor topography is believed to have affected the distribution of the sands deposited in this area. The sandy basin floor turbidites were only deposited in the western parts of the Central Basin where the basin has been interpreted to have been deeper. The thicknesses of sediment also reflect this trend where thicker sediment packages are found closer to the orogen. These observations point to an asymmetric infilling of the basin.

The shallow marine deposits of the Battfjellet Fm show that wave-action was important in the basin, and these deposits show a coarsening and shallowing upwards trend with mainly wave-generated deposits in the shoreface environment. These represent parasequences and the sandy sequences are capped by transgressive shales. The number of parasequences varies over short lateral distances in the study area and this has been interpreted to represent switching of delta lobes.

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1 Introduction

1.1 Purpose of study

Outcrops belonging to the Battfjellet Fm are present in the Paleogene Central Basin of Spitsbergen. In the past there have been extensive studies on the Battfjellet Fm in Van Keulenfjorden and Reindalen (e.g. Dalland, 1979; Helland-Hansen, 1990, 1992; Steel et al., 2000; Plink-Björklund et al., 2001; Mellere et al., 2002; Deibert et al., 2003; Crabaugh and Steel, 2004; Johannessen and Steel, 2005; Løseth et al., 2006; Uroza and Steel, 2008; Helland-Hansen, 2010) as well as some more recent studies on the Battfjellet Fm on the southern side of Van Mijenfjorden (Olsen, 2008; Stene, 2008). The Rypefjellet area, which has been the focus of this study, is located in an area of the Central Basin where no detailed study of the Battfjellet Fm has been carried out before although the stratigraphy of the Van Mijenfjorden Gp was studied in the area west of Rypefjellet by Croxton and Pickton (1976). Thus, the main purpose of this study has been to present a detailed paleoenvironmental and paleogeographic model of the Battfjellet Fm in this area based on detailed facies analysis and large-scale sandbody geometry.

1.2 Short introduction to the Battfjellet Fm

The Battfjellet Fm was deposited in the Central Basin, a foreland basin to the West Spitsbergen Orogen, which formed in Paleogene time due to transpressional movements as the opening of the Atlantic Ocean spread northwards. The sediments in the Battfjellet Fm are derived from this fold-and-thrust belt and have been interpreted to form a shallow-marine sequence deposited during a regressive stage of basin-filling (Steel et al., 1981).

1.3 Previous work on the Battfjellet Fm

Unlike some of the other formations in the Central Basin, the Battfjellet Fm does not contain any economically significant deposits (coal) and hence the earliest studies on the Paleogene succession in Spitsbergen only describe the Battfjellet Fm for stratigraphical purposes (e.g. Kellogg, 1975; Croxton and Pickton, 1976; Steel et al., 1981). The paleogeographical and paleoenvironmental framework of the Battfjellet Fm was established by (Helland-Hansen, 1985) while (Steel et al., 1985) synthesized the sedimentological descriptions of the deposits in the Central Basin with structural investigations of the West Spitsbergen Orogen. Over the last decade much work has been carried out on the Battfjellet Fm (e.g. Plink-Björklund et al., 2001; Mellere et al., 2002; Steel and Olsen, 2002; Deibert et al., 2003; Crabaugh and Steel, 2004; Johannessen and Steel, 2005; Løseth et al., 2006; Uroza and Steel, 2008; Helland-Hansen, 2010) the majority of this in the Van Keulenfjorden and Reindalen areas as both these areas exhibit excellent exposures of Battfjellet Fm clinothems.

1.4 Outline of chapters

The chapters in this study are set up so that the data are gradually conveyed in a constructional manner. Chapter 2 describes the methods used while doing the fieldwork when collecting the data needed to convey the paleogeographical and paleoenvironmental models. Chapter 3 sums up the extensive regional geology of Svalbard, starting with the pre-Devonian deposits and going up to the present. Chapter 4 gives detailed descriptions of the interpreted facies while chapter 5 describes how the facies are divided into different facies associations. Chapter 6 describes the large-scale sandbody geometry of the Battfjellet Fm in the study area, while chapter 7 gives a model for the depositional environment and paleogeograpgy. The results and interpretations are discussed in chapter 8 before conclusions are given in chapter 9.

2 Methods

During the summers of 2008 and 2009 the Battfjellet Fm in the Rypefjellet area north of Van Mijenfjorden (Figure 2-1) was investigated for a total of approximately 6 weeks. 23 vertical profiles were measured and together with detailed sedimentological descriptions and large-scale observations these vertical profiles form the basis for this thesis. The features that were considered important were thicknesses, grain size, colour, composition, texture, boundary types, sedimentary structures and paleocurrent measurements.

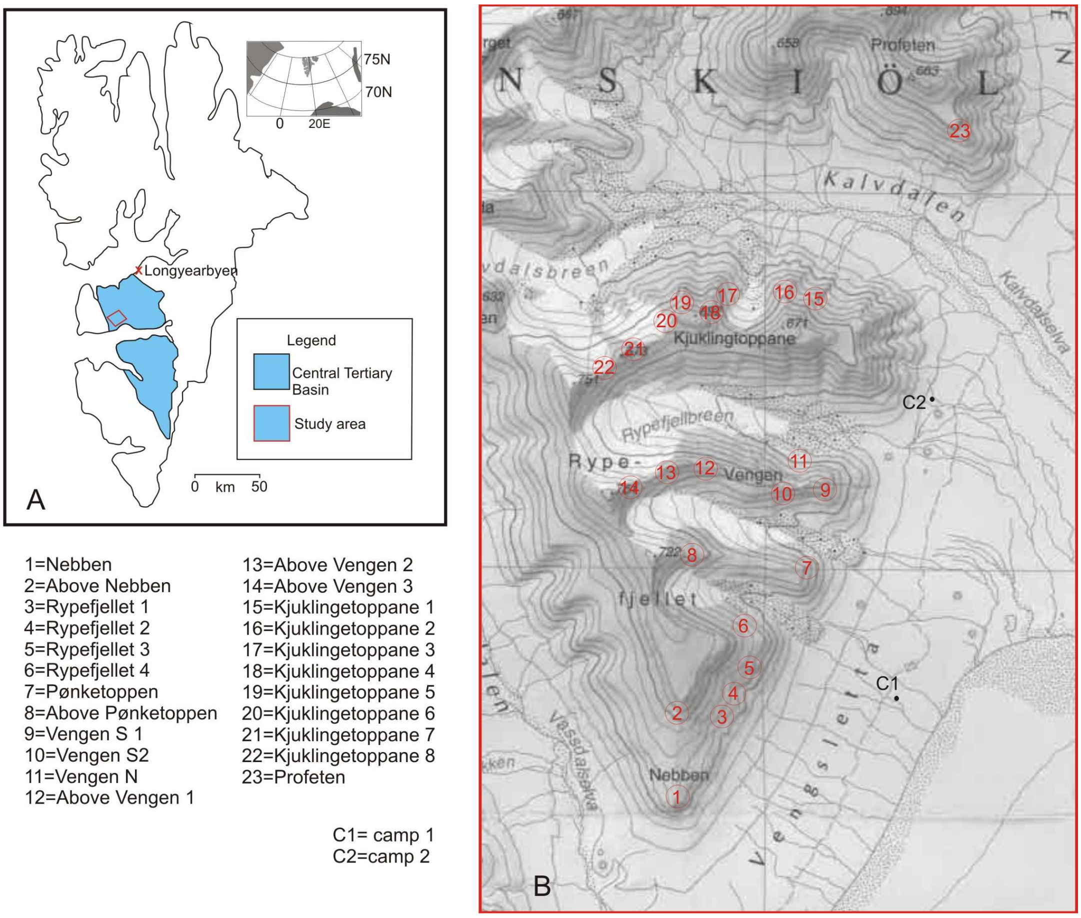


Figure 2-1 A) Map of Spitsbergen showing the location of the study area in the Central Basin and locations of logs and camps in the study area.

The outcrops were usually approached by foot and much of the time in the field was spent walking between outcrops. The camp was located in the southern part of the study area the first field season (C1, Figure 2-1) and during the second field season the camp was located in the northern part of the study area (C2, Figure 2-1). This made it possible to get a more thorough focus on one part of the study area each year. Access to a small boat during the first field season also allowed for taking panorama photographs of parts of the area and these have been important for correlation between sandbodies. Detailed photographs of part of the study area were also taken from helicopter.

3 Geologic Framework

3.1 Introduction

The Svalbard archipelago is located in the Barents Sea between 74 and 81N and 10 and 35E (Figure 3-1). Svalbard has an extensive post-Caledonian geological record including rocks ranging from all periods from the Devonian to the Paleogene. During this time period Svalbard has also moved northwards from a position near the equator to its current position (Figure 3-2), and this movement is reflected in the characteristics of the different deposits. Geologists have studied the rocks on Svalbard since the 19th century (Steel and Worsley, 1984) where the purpose of the earliest expeditions was to establish the stratigraphic framework and fossil flora and fauna of the islands. The first studies on

Figure 3-1 Map of Svalbard and surrounding areas

The Paleogene (Tertiary) successions in Spitsbergen were focused on the lower, coal-bearing strata (Steel et al., 1981) which have been successfully mined in several locations in Spitsbergen over the last century (Croxton and Pickton, 1976).

Geologically, Svalbard is an uplifted part of the Barents Shelf, an area currently explored for hydrocarbons. Onshore Svalbard there have also been several drilling projects but no significant discoveries have been made, including several non-commercial gas finds (Dalland, 1979). A better understanding of the geology of Svalbard can help to improve the current models used for petroleum exploration in similar geological settings in the Barents Sea as well as other areas of the world, which confirms that geological studies on Svalbard are still relevant.

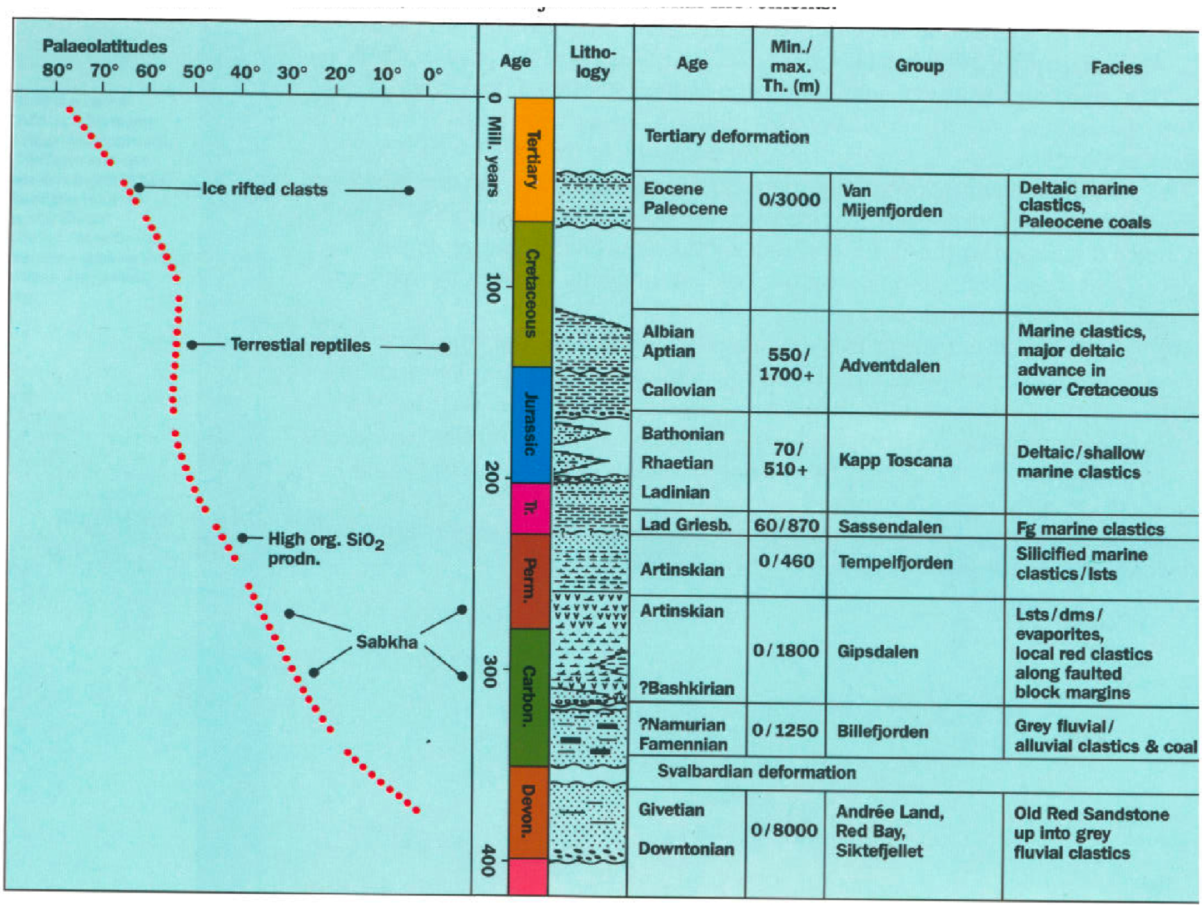


Figure 3-2 Svalbard's position through geologic time (from Worsley and Aga, 1986)

3.2 Pre-Paleogene

The pre-Devonian rocks on Svalbard are collectively named Heckla-Hoek and they appear to have undergone several deformational phases shown through intense folding and faulting. The Heckla Hoek consists of 20 lithostratigraphic groups and has a maximum aggregate thickness of 20 km (Worsley, 2008). In middle-Paleozoic time Spitsbergen, along with Europe and Greenland, was affected by the Caledonian Orogeny (Harland, 1969). This orogeny was the result of a collision between the Laurentian and

Fennoscandian plates and the main phase of deformation in Spitsbergen, which took place in Late Silurian time, has been named the Ny Friesland Orogeny (Harland, 1969; Friend et al., 1997).

In late Silurian to early Devonian time sediments were eroded from the newly formed orogen and deposited as Old Red Sandstone (Worsley and Aga, 1986). The majority of these deposits belong to the Wood Bay Fm and are of fluvial origin, deposited by northward flowing streams in a graben in northern Spitsbergen (Friend, 1965; Harland, 1969; Worsley, 2008). Around the early to mid-Devonian transition there is a change in colour in the sediments from red to grey which indicates a shift in climatic zones from the southern arid zone to the equatorial tropics (Worsley, 2008).

The Ny Friesland Orogeny was followed by two smaller tectonic episodes, the Haakanian event in the Late Silurian and the Svalbardian event in the Late Devonian (Dallmann, 1999). The Svalbardian is considered the most important of these events and it was the result of sinistral strike-slip movement between Spitsbergen and Greenland (Harland, 1969). The Svalbardian phase marks the end of the Caledonian Orogeny in Spitsbergen (Harland, 1969) although the post-Caledonian succession shows evidence of considerable tectonic movement (Steel and Worsley, 1984)

Some readjustments after the Svalbardian event such as uplift, subsidence and gravity faulting were still taking place in the Early to Middle Carboniferous. This tectonic unrest is also apparent from patchy deposits that are confined to separate basins (Harland, 1969). The post-Devonian sediments on Svalbard were also affected by various fault zones, usually directed either N-S or NW-SE, where the most important of these were the Lomfjorden/Agardbukta, Billefjorden and the Inner Hornsund and Paleo-Hornsund Fault Zones (Steel and Worsley, 1984, Figure 3-3). Thicknesses of the younger (Late Carboniferous and Permian) deposits show conditions of increasing tectonic stability (Harland, 1969). By early Permian times most of the northern areas of Svalbard had relatively stable platform conditions and remained so through the Mesozoic (Steel and Worsley, 1984).

The Carboniferous and Permian deposits in Spitsbergen have been divided into three different depositional groups. The Billefjorden Gp is the oldest of these and rests unconformably on Heckla-Hoek and Old Red Sandstone deposits. The sediments in the Billefjorden Gp have been interpreted as swamps, lakes and flood plain deposits that were eroded from graben edges and transported into the graben by alluvial fans (Gjelberg and Steel, 1981; Steel and Worsley, 1984).

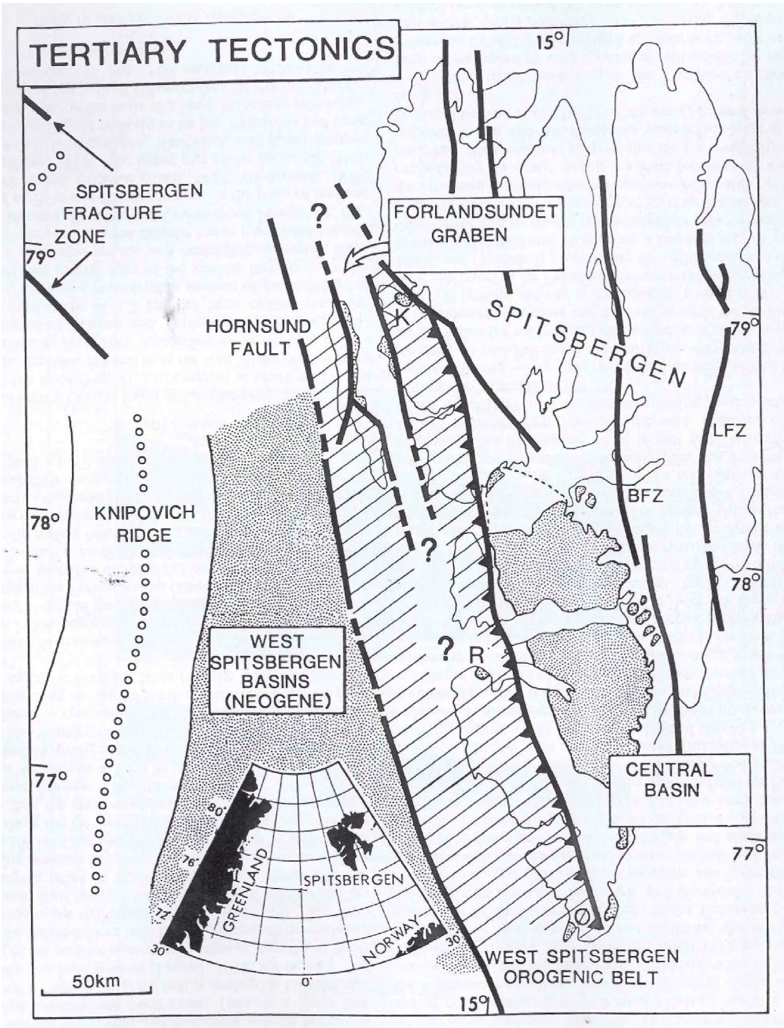


Figure 3-3 Structural lineaments in Svalbard. K: Kongsfjorden, R: Renarodden, Ø: Øyrlandet, BFZ: Billefjorden Fault Zone, LFZ: Lomfjorden/Agardbukta Fault Zone. These faults were active from post-Devonian time and were reactivated in the Paleogene (from Steel et al, 1985).

The Billefjorden Gp is overlain by the Gipsdalen Gp. Most of the Gipsdalen Gp consists of carbonate deposits that formed in a sahbka environment and these deposits reflect a change in the depositional environment that can be related to quieter tectonic conditions and a warmer and more arid climate (Steel and Worsley, 1984; Worsley, 2008). The youngest Paleozoic deposits belong the Tempelfjorden Gp and show a change to cooler- and deeper water deposition which reflects the closure of the seaway connection to the warm Tethys Ocean and the development of the Ural mountains (Worsley, 2008). The Tempelfjorden Gp contains both clastic and carbonate deposits.

Mesozoic

The tectonic stability that was established in Permian times remained through the Mesozoic with the exception of some folding and faulting at the end of the Jurassic period (Harland, 1969). At this time (Jurassic-Cretaceous boundary) there was significant vertical movement along the Billefjorden Fault Zone and possibly also the Agardbukta Lineament (Steel and Worsley, 1984). At this time regional uplift in the northern parts of Spitsbergen provided new source areas that gave great sedimentary input although the sedimentation was outpaced by basinal subsidence and eustatic sea-level rise. These mechanisms provide large-scale evidence of tectonic unrest at the end of the Mesozoic period (Steel and Worsley, 1984)

Three depositional groups are assigned to the Mesozoic Era on Spitsbergen. The oldest and lowermost of these is the Sassendalen Gp which consists of mainly marine shales with some sand- and siltstones (Steel and Worsley, 1984). The deposits in the Sassendalen Gp have been interpreted as barrier systems and delta environments, and there is also evidence of tidal influence. In the southern part of Spitsbergen the deposits contain less sand which suggests a more open-marine environment. The slightly younger Kapp Toscana Gp consists of deposits interpreted as two regressive sequences. These deposits consist of sediments that vary from offshore to continental origin, although the majority are marine or deltaic. The nature of the deposits in the Kapp Toscana Gp indicate that there was a transgression after the deposition of the Sassendalen Gp.

The uppermost of the Mesozoic groups is the Adventdalen Gp which consists of the open-marine Janusfjellet Fm, deltaic sandstones belonging to the Helvetiafjellet Fm and mixed-marine sequences in the uppermost Carolinefjellet Fm (Steel and Worsley, 1984). Uplift in the Late Cretaceous led to erosion and the uplift is believed to be more extensive in the north as the erosion into the Carolinefjellet Fm is deeper here (Steel and Worsley, 1984).

3.3 Paleogene

The uplift in Late Cretaceous marked the onset of the tectonically active Paleogene period in Spitsbergen and reflects the gradual northward movement of the opening of the Atlantic Ocean (Figure 3-4). An important product of the Atlantic opening was the development of the West Spitsbergen Orogen, which stretches 300 km along the western coast of Svalbard from Brøggerhalvøya in the north to Sørkapp in the south. The orogenic belt has been studied extensively and evidence of deformation has been collected both onshore and offshore. The sediments deposited in the Central Basin are also important for the understanding of the plate-tectonic evolution along Spitsbergen's western margin.

Eldholm (1977) based their interpretations on offshore investigations and concluded that Greenland and Svalbard were separated by a transform fault in early Paleocene time. Another of the early investigations by Eldholm et al. (1984) suggested that a transpressional regime operated in the area from early Paleocene to late Eocene time and was followed by transtension in the Oligocene. The transtensional setting was probably related to a reorganization of the plates further south and this led to the development of the present margin off Spitsbergen (Myhre and Eldholm, 1988). The fault zone that developed off the western coast of Spitsbergen has been named the Hornsund Fault Zone (Myhre et al., 1982)

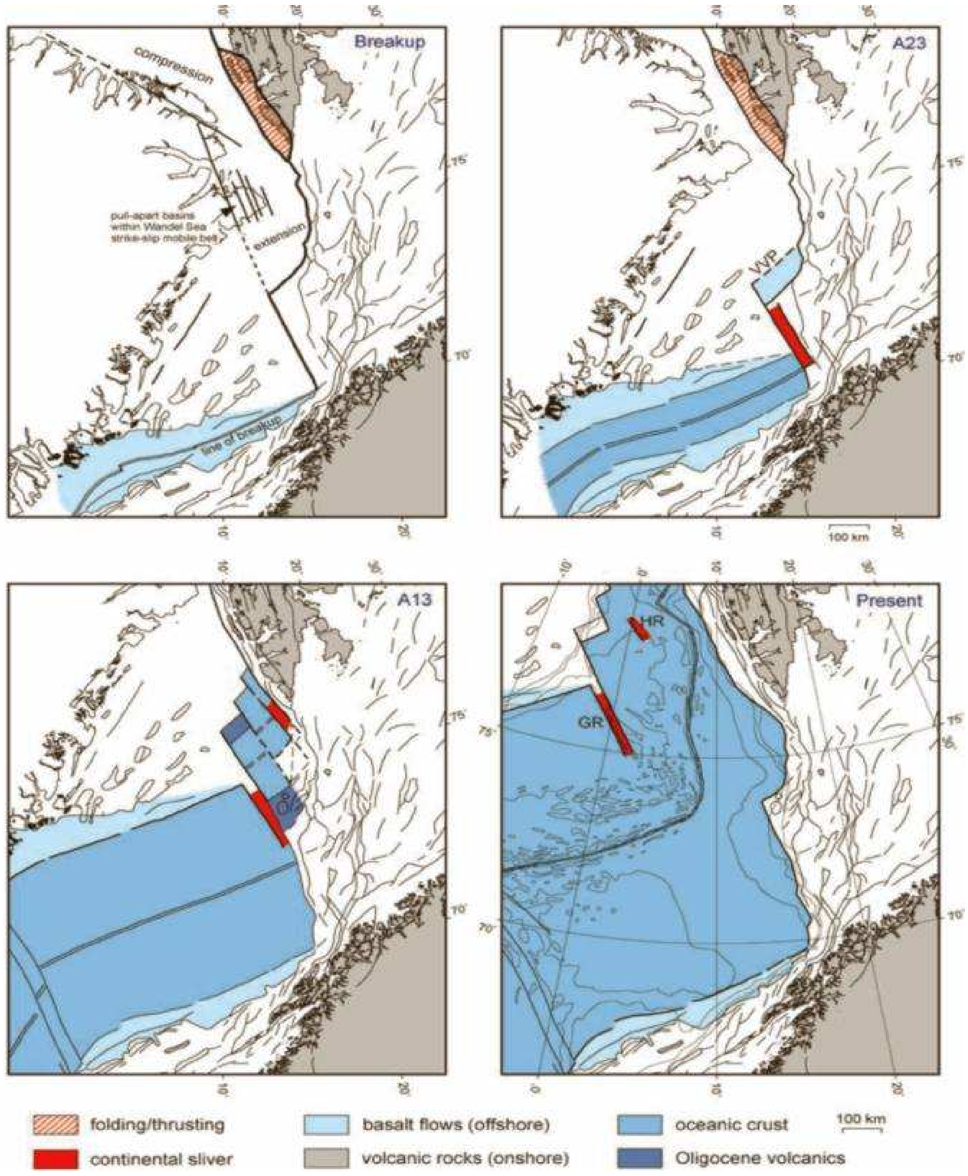


Figure 3-4 The gradual northward opening of the Atlantic Ocean (from Faleide et al., 2008).

Several models have been proposed to explain the plate-tectonic evolution in the Norwegian-Greenland Sea between Late Cretaceous and Oligocene time. Talwani and

The timing and nature of the events that led to the opening of the Greenland-Svalbard Sea and the development of the Spitsbergen fold-and-thrust belt has been discussed by several workers (e.g. Kellogg, 1975; Steel et al., 1981; Steel et al., 1985; Braathen and Bergh, 1995; Braathen et al., 1999; Bruhn and Steel, 2003) but there is still no conclusion on the exact timing of the different events. There seems to be general agreement though that a transtensional regime operated in the Early Paleocene and was followed by the main stage of deformation, the transpressional events that led to development of the fold-and-thrust belt in the period between mid/late Paleocene and Eocene time. This transpressional regime continued into the Oligocene when sea floor first was generated between Greenland and Spitsbergen. However, the timing of the onset of the dextral, transpressional strike-slip movement has later been questioned by Bruhn and Steel (2003) who indicated that this may have been taking place as early as the Late Cretaceous. Several previous workers (e.g. Kleinspehn et al., 1989; Braathen and Bergh, 1995; Braathen et al., 1999) also indicated that the transpressional movement might have started in the late Cretaceous. However, there seems to be agreement that the transpressional/compressional setting continued through the Eocene and that sea-floor spreading between Svalbard and Greenland started in the Oligocene.

After the development of the fold-and-thrust belt Spitsbergen has been located on an uplifted part of the Barents Shelf. After the Central Basin was filled the erosion in Spitsbergen has (see Aspelintoppen Fm description below) and the very similar summit heights of the mountains present in Svalbard today made Harland (1969) suggest there had been erosion down to a peneplan some time after the fold-and-thrust belt formed.

In the Pliocene and Pleistocene there has been some volcanic activity on Sverresfjellet, a small conical volcano in the north-western parts of Spitsbergen but today erosion by glacial ice is the most important factor for shaping and carving the landscapes of Spitsbergen. Approximately 60% of Svalbard (Worsley and Aga, 1986).

Paleogene outliers

At several locations near the Central Basin; Forlandsundet, Renarodden and Øyrlandet (Steel et al., 1981), there are small outliers with deposits believed to be time-equivalent with the oldest deposits in the Central Basin. Thus it is likely that the basin was larger and extended both south and north of the present basin-margin (Kellogg, 1975). The Paleogene outliers are not discussed further in this study.

3.3.1 The Central Basin

3.3.1.1 Mechanics of the Central Basin

The Central Basin is located on the eastern side of the West Spitsbergen Orogen (Figure 3-5) and formed simultaneously with the orogenic belt and also acted as a foreland basin (e.g. Steel et al., 1985; Helland-Hansen, 1990; Plink-Björklund et al., 2001; Mellere et al., 2002; Bruhn and Steel, 2003). Bruhn and Steel (2003) suggested that the basin first formed as a flexural depression in front of the thrust wedge in Late Cretaceous or early Paleocene time but, similar to the events concerning the development of events related to the opening of the Atlantic Ocean, the exact timing of events is not known. The majority of the Paleogene deposits on Svalbard are confined to the Central Basin.

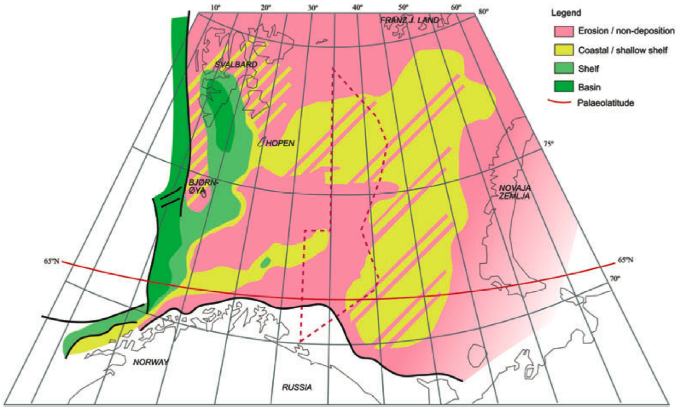


Figure 3-5 Regional uplift in Paleogene time. Note that the study area in Spitsbergen is located in a basin while there is uplift to the west along the West Spitsbergen Orogen (from Worsley, 2008)

3.3.1.2 Filling of the Central Basin

The Paleogene Central Basin in Spitsbergen is filled with up to 2.4 km thick successions of siliciclastic sediments that belong to the Van Mijenfjorden Gp. (Figure 3-6). The deposits in the Van Mijenfjorden Gp reflect the alternating transtensional and transpressional conditions that occurred along the fold-and-thrust belt (Steel et al, 1985). The most recent stratigraphic scheme divides the succession into seven different formations (Dallmann, 1999) and this is the scheme used in this study. In the past there have been various names and schemes for classifying the deposits in the Van Mijenfjorden Gp and these are shown in Figure 3-7.

Vitrinite studies by Manum and Throndsen (1978) showed that an additional 1.7 km of Paleogene deposits have been eroded from the top of the Aspelintoppen Fm in the Central basin. The ages of the different deposits in the Van Mijenfjorden Gp are not well known but are assumed to be of Paleocene-Oligocene age and deposited contemporaneously with the development of the West Spitsbergen Orogen where most of the sediments were deposited before sea-floor spreading started in this area in Oligocene time (e.g. Eldholm et al, 1984; Helland-Hansen, 1990). A dating by Manum and Throndsen (1986) gave a latest Paleocene age for the lower Gilsonryggen Mb of the Frysjaodden Fm.

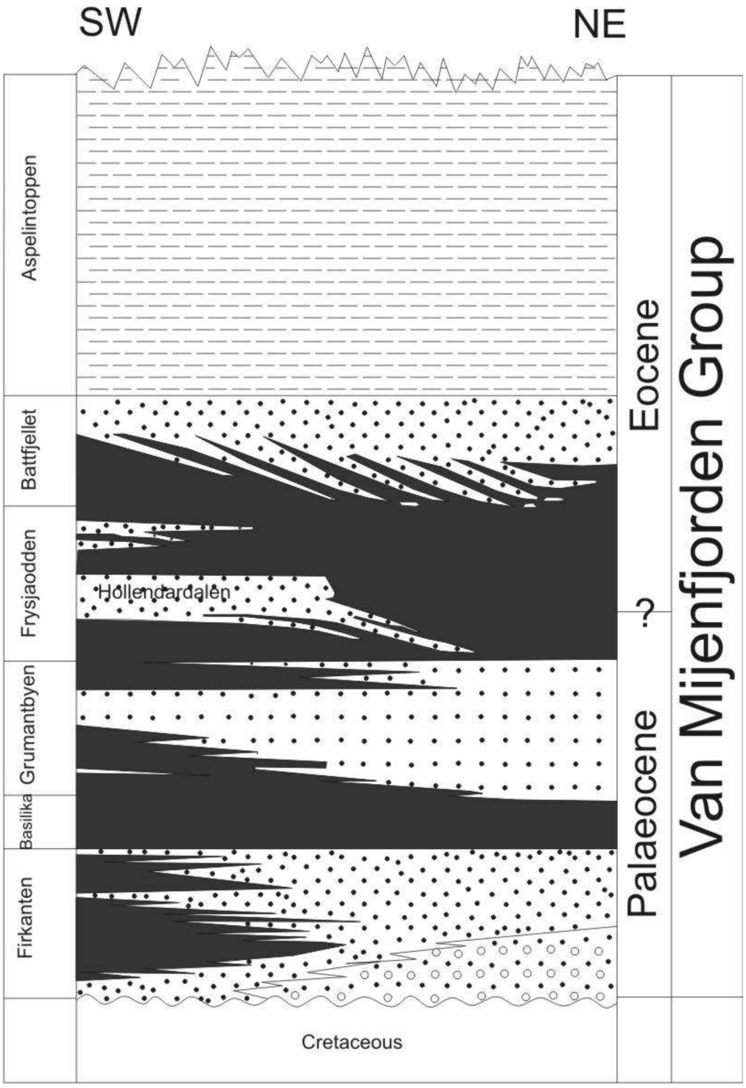


Figure 3-6 Stratigraphy of the Van Mijenfjorden Group (after Steel et al, 1985).

Central Tertiary Basin								
NATHORST 1910	HOEL 1925	LJUTKEVIC 1937	ORVIN 1940	LIVSIC 1965	LIVSIC 1967, 1974	MAJOR & NAGY 1964, 1972	DALLMANN, 1999 (present stratigraphic scheme)	
Upper coal-bearing series	Upper sandstone series	Upper coal-bearing sand-stone formation	Upper Plant-Bearing Sand-stone series	Upper sandstone formation	Storvola Formation	Aspelintoppen Formation	Aspelintoppen Formation	
Fissile sand-stone series		Upper shaley sandstone formation	Flaggy Sandstone Series	Upper transi-tional formation	Collinderodden Formation	Battfjellet Formation	Battfjellet Formation	
Upper black shale series	Middle shale series	Upper black shale formation	Upper Black Shale Series	Upper argillite formation	Frysjaodden Formation	Gilsonryggen Formation	Frysjaodden Formation	Bjørnsonfjellet Mb Gilsonryggen Mb
Green sandstone series	Lower sandstone series	Shaley green sandstone formation	Green sandstone series	Lower transitional formation	Hollendardalen Formation	Sarkofagen Formation	Hollendardalen Formation	
							Frysjaodden Fm.	Marstranderbr. Mb.
		Green sandstone formation		Green sandstone formation	Grumant(byen) Formation		Grumantbyen Formation	
Lower black shale series		Lower black shale formation	Lower Dark Shale Series	Lower argillite formation	Colesbukta Formation	Basilika Formation	Basilika Formation	
Lower light sandstone series		Lower coal-bearing sand-stone formation	Lower Light Sandstone Series	Lower coal-bearing sand-stone formation	Barentsburg Formation	Firkanten Formation	Firkanten Formation	Endalen Member Kolthoffberget Mb. Todalen Member

Figure 3-7 various nomenclatures used by different authors for the deposits in the Central Basin (modified from Dallmann, 1999).

The sediments filling the Central Basin display an overall regressive depositional trend, and show evidence of an eastward migrating depocenter and lithospheric shortening in the West Spitsbergen Orogen (Bruhn and Steel, 2003). The oldest deposits in the Van Mijenfjorden Gp were deposited mainly by sediments from the eastern side of the basin, interpreted by to be derived from a foreland bulge, while the younger sediments were derived from the orogen located to the west of the basin (Bruhn and Steel 2003). The deposits in the Van Mijenfjorden Gp are mainly wave- and tide-dominated (Steel et al, 1981).

Steel et al (1981) have divided the filling of the Central Basin into three different stages:

- **The first (transgressive) depositional cycle – Firkanten and Basilika formations**

  The lowermost of the Paleogene deposits in the Van Mijenfjorden Gp is the Firkanten Fm, which forms an angular unconformity to the underlying Cretaceous Carolinefjellet Fm. The Firkanten Fm is usually less than 200 m thick (Steel et al, 1981) and consists of three different members; the Todalen, Endalen and Kalthoffberget members respectively. The Todalen Mb has been interpreted to consist of mainly delta plain deposits (Steel et al, 1981), and is known for its economically important coal deposits. The source areas for the Todalen Mb were located on the eastern and northern margins of the basin. Tidal processes were common on the delta-plain and abandoned delta-lobes suggest the delta had a lobate shape (Steel et al, 1981).

The Endalen Mb has been interpreted as wave-dominated delta front deposits (Steel et al, 1981; Steel and Worsley, 1984; Bruhn and Steel, 2003) while the finer grained Kalthoffberget Mb, which only is present in the southern and western parts of the Central Basin, has been interpreted as lower delta-front deposits (Steel et al, 1981).

The upward change from coal-bearing delta plain deposits in the Todalen Mb into shelf and shoreline deposits reflect a deepening of sea level and a transgressive setting in the Firkanten Fm.

The Basilika Fm overlies the Firkanten Fm and shows significant thickening from 20 m in the north-eastern parts of the Central Basin to 300 m in the south and southwest of the basin (Steel et al, 1981). The deposits in the Basilika Fm are mainly black shales that tend to get sandier and siltier in the north-eastern area of the basin and towards the top of the formation. (Steel et al, 1981). Scattered, well-rounded pebbles consisting of chert and quartzite have been observed and were interpreted by Dalland (1977) as ice-rafted material. The deposits in the Basilika Fm have been interpreted to be of prodelta origin (Steel et al, 1981) and because these deposits are more distal than the Firkanten Fm the Basilika Fm forms the transgressive capping of the Firkanten-Basilika mega sequence (Steel et al, 1981).

- **The second (regressive) cycle – Grumantbyen and Hollendardalen formations**

The Grumantbyen Fm forms the lowermost unit in the second phase of infilling in the Central Basin and consists of greenish, highly bioturbated sandstones that have a thickness of 450 m in the northern areas of the Central Basin but thin out southwards (Steel et al, 1981). The Grumantbyen Fm has been interpreted as offshore deposits, probably deposited in offshore bar complexes (Dalland, 1977).

The Grumantbyen Fm is separated from the overlying Hollendardalen Fm by marine shales belonging to the Marstranderbreen Mb of the Frysjaodden Fm.

The Hollendardalen Fm consists of sandstones that are up to 150 m thick in the western part of the Central Basin, but pinch out towards the centre of the basin (Steel et al., 1981; Steel et al., 1985). The Hollendardalen Fm has been interpreted as an eastward prograding wave- and tide-dominated delta (Dalland, 1979). Coals are also occasionally present. The western source area for this formation has been related to western uplift that occurred as early as the late Paleocene, and this establishes a close relationship between the Hollendardalen Fm and the overlying formations (Steel et al, 1981).

The transition from offshore sands in Grumantbyen Fm to shallow-marine tidal deltas in the Hollendardalen Fm indicate that the second phase of infilling in the Central Basin was regressive, but it is important to remember that the sands in this cycle were derived from different basin margins (Steel et al, 1981).

- **The third (regressive) cycle – Gilsonryggen Mb, Battfjellet and Aspelintoppen formations**

The third cycle of infilling of the Central Basin is described in more detail as one of the formations in this cycle, the Battfjellet Fm, is the main focus of this study. The deposits below and above the Battfjellet Fm will also be described in detail as they are important for the understanding of the whole depositional system.

Gilsonryggen Mb of the Frysjaodden Fm

Marine silty shales of prodelta origin that belong to the Gilsonryggen Mb of the Frysjaodden Fm initiated the deposition of the third cycle of infilling in the Central Basin (Steel et al, 1981). The maximum thickness of the Gilsonryggen Mb is 900 m in the western areas of the basin, but it thins out to around 300 m in the east (Kellogg, 1975). In the western part of the Central Basin some thin sandstone wedges that thin out towards the east have been interpreted as turbidite deposits of the Bouma T_{abc} divisions (Steel et al., 1981; Steel et al., 1985). These deposits were named the Bjørnsonfjellet Mb by Steel et al (1981) and have a thickness of up to 20 m. The Bjørnsonfjellet Mb deposits deposits have a predominantly massive or structureless texture, and there also appears to be some deformation through slumping. This is significant as it indicates that deposition took place on a slope setting which suggests that there was Paleocene uplift along the western margin of Spitsbergen (Steel et al, 1981).

The Bjørnsonfjellet Mb may be of similar nature to the basin-floor deposits described by Steel et al (2000), Plink-Björklund et al (2001), Mellere et al (2002), Crabaugh and Steel (2004) at more distal locations in the basin though these deposits are not time equivalent to the Bjørnsonfjellet Mb.

In the upper parts of the Gilsonryggen Mb the sand content gradually increases, which leads to an upwards-coarsening trend. The boundary between the Gilsonryggen Mb and the overlying Battfjellet Fm has been set at the level where shales become subordinate to sands (Helland-Hansen, 1990).

Battfjellet Fm

The main focus of this study is the Battfjellet Fm, which represents an upwards-coarsening, shallow-marine sequence (Steel et al., 1981) that consists of mainly very-fine and fine-grained sandstone. The content of organic material in the Battfjellet Fm is high (Croxton and Pickton, 1976) and the thickness of the Battfjellet Fm varies from 60-200 m (Steel et al., 1981).

A striking feature of the Battfjellet Fm is the organisation into one or more superimposed, upwards-coarsening sandy sequences that are separated by finer-grained deposits (Helland-Hansen, 1990). These sequences are especially prominent in the western part of the Central Basin where they are connected to continental deposits at the top and pinch out downwards into the marine shales of the Frysjaodden Fm (Kellogg, 1975; Steel et al, 1981). These sequences can be termed clinothems (Rich, 1951) and reflect the deposition of sands in a slope setting. The term “clinothem” represents the 3D rock unit while the term “clinoform” represents the corresponding seaward-sloping surface (Rich, 1951). The clinothems show a general dip-direction towards the east, which corresponds to the inferred progradation direction of the shoreline (Steel et al, 1981; Helland-Hansen, 1990). Paleocurrent measurements from fluvial channels in the Aspelintoppen Fm and scour marks at the bases of mass-flow deposits also point to an eastward migration shoreline, while wave-crests of wave ripples, which are inferred to be parallel to the shoreline have a north-south orientation (e.g. Helland-Hansen, 1990). The clinothems disappear in the eastern part of the basin; Helland-Hansen (1990) interpreted this as evidence for an eastward shallowing the foreland basin.

The Battfjellet Fm has been interpreted as a prograding deltaic and barrier coastline with evidence of both wave and tidal processes (Steel, 1977). Wave-generated structures dominate the lower parts of the shallow-marine deposits in the Battfjellet Fm while current generated structures seem more important in the upper parts of the succession (Helland-Hansen, 1990). In some areas in the western part of the basin extensive bodies of sand have been observed in front of the clinothems and it is assumed that fluvial and gravity dispersal processes were important processes for distributing these sands on the slope and basin-floor (Helland-Hansen, 1985). Many of the recent publications on the Battfjellet Fm have been most concerned with describing how sands are transported to and deposited on the shelf, slope and basin floor (e.g. Steel et al., 2000; Plink-Björklund et al., 2001; Mellere et al., 2002; Deibert et al., 2003; Crabaugh and Steel, 2004) as well as the sequence stratigraphic interpretation of the Battfjellet Fm (e.g. Bruhn and Steel, 2003; Uroza and Steel, 2008). These publications generally acknowledge that the basin-floor sands in the Battfjellet Fm were deposited by turbidity currents. The turbidity currents were probably steady or quasi-steady and initiation by direct river effluents appears to be an important generating mechanism (e.g. Plink-Björklund et al., 2001; Mellere et al., 2002;

Plink-Björklund and Steel, 2004). These rivers formed shelf-edge deltas (Mellere et al, 2002) or shelf-deltas (Helland-Hansen, 2010) and transported sediments that accumulated on the slope and basin-floor (Steel et al, 2000; Plink-Björklund et al, 2001; Mellere et al, 2002; Plink-Björklund and Steel, 2004, Helland-Hansen, 2010). Estuaries and channel fills also show evidence of tidal movements, such as the presence of mud drapes and bi-directional currents (Plink-Björklund, 2005; Løseth et al, 2006; Uroza and Steel, 2008) but these processes seem to be of minor importance for shaping the shoreline.

The stacking pattern of the Battfjellet Fm is complex and the highly variable number of parasequences across the basin has been explained as shifting of deltaic lobes which produced a pattern of overlapping sandbodies (Helland-Hansen, 2010). The domination of wave-generated structures in the Battfjellet Fm suggests that waves dominated the depositional basin (Helland-Hansen 2010) However, the large-scale geometry of the delta deposits with lobate sand bodies and shifting and abandonment of lobes suggest a fluvially dominated system. Hence, some of the most recent works have classified the Battfjellet Fm as a fluvio-wave dominated delta (Olsen, 2008; Helland-Hansen, 2010).

Aspelintoppen Fm

The uppermost deposits in the Van Mijenfjorden Gp belong to the Aspelintoppen Fm, which has been interpreted as the continental counterpart of the underlying Battfjellet Fm and Gilsonryggen Mb. The thickness of the Aspelintoppen Fm is not known as it has been affected by at least 1700 m of erosion (Manum and Throndsen, 1978). The Aspelintoppen Fm also marks the culmination of the infilling of the Central Basin.

The boundary between the Aspelintoppen Fm and the underlying Battfjellet Fm is hard to determine precisely (Croxton and Pickton, 1976) as it commonly is scree-covered but observations from the southern side of Van Mijenfjorden reveals an abrupt and/or interfingering relationship between the Aspelintoppen and Battfjellet formations (Plink-Björklund, 2005). The Aspelintoppen Fm consists of alternating sandstones, shales, siltstones and coals where plant debris, organic matter and soft-sediment deformation structures are common (Steel et al., 1981; Steel et al., 1985). Generally, the Aspelintoppen Fm has an aggradational nature and is mud-prone with sand: shale ratio of about 0.25 (Kellogg, 1975; Steel et al, 1981; Steel and Worsley, 1984; Steel et al, 1985; Plink-Björklund, 2005). The sediments of the Aspelintoppen Fm have been interpreted to be deposited in the Central Basin from a source area on the western margin and have been interpreted to reflect fresh water shales that alternate with non-marine and marine brackish water sandstones, coals and siltstones (Steel et al., 1981). Hence, the Aspelintoppen Fm has been interpreted as an environment consisting of low-sinuosity streams that were flowing through a low-gradient depositional plain (Helland-Hansen, 1990).

4 Facies Description and Interpretation

In order to make a contribution towards a better understanding of the Battfjellet Fm, the deposits have been divided into facies based on properties such as grain-size, colour, sedimentary structures, texture, composition, fossils and bedding as well as the predicted depositional mechanism. The facies interpretation is shown in Table 4-1 below.

Facies	Dominant grain size	Main depositional features	Environment
A	Clay	Dark shales with lamination. Background sedimentation.	Offshore
B1	Very-fine	Ungraded thick sandstones, interpreted as Bouma T_A	Offshore (basin-floor fan)
B2	Very-fine	Alternating structureless and laminated strata interpreted as Bouma T_{ABC}	Offshore (basin-floor fan)
B3	Very-fine	Upward fining turbidite deposits with limited lateral extent, interpreted as	Offshore (basin-floor fan and slope)
C	Silt – very-fine	Upward coarsening sediments, dominated by storm-generated structures (HCS)	Offshore transition
D	Very-fine	Symmetrical ripples and HCS, formed by fair-weather wave action	Lower shoreface
E	Very-fine - fine	Small, erosive channel, PPL and symmetric ripple lamination. Bioturbation. Rapidly shifting currents and wave action.	Middle shoreface
F1	Fine - medium	Tabular cross-stratification from dune migration. Upward coarsening trend continued.	Upper shoreface/ foreshore
F2	Fine	Erosional base, cross-stratification, often abrupt change in colour or grain-size from underlying deposits.	Channel eroding into shoreface
G	Silt - very-fine	Organic rich deposits with roots, leaf-fossils and influence from waves. Continental origin.	Delta plain
H	Medium - coarse	Within delta-plain, contains cross-stratification and	Fluvial channel

Table 4-1 Sedimentary facies in the Rypefjellet area

The division into facies allows for integration of depositional processes and the factors that control the depositional processes. A facies should "ideally represent a distinctive rock formed under certain conditions of sedimentation and reflect a particular process, set of conditions or environments" (Reading and Levell, 1996). A facies may consist of a single bed or a group of several beds, and the concept of facies is useful when correlating between different units or predicting the presence of coal, oil or mineral ores (Reading and Levell, 1996).

4.1 Facies A – prodelta deposits/offshore shales

This facies is present on the lower part of the mountainsides in the study area (Fig 4.1A) and consists of mainly black to blue-grey coloured homogenous silty shales. In the more proximal reaches of this facies some yellow-coloured sands have also been observed. The dark, shale-rich deposits usually have a flaky appearance and consist of loose fall-out screes while the sandier parts seem more resistant to erosion (Figure 4-1 B and C). Hence, lamination is also more common in the sandier deposits. The blue-grey coloured deposits appear more iron-rich than the black sediments. Some siderite clasts have also been observed within this facies. The thickness of this facies varies from 4 m to several hundred meters in the study area.

The deposits in facies A usually surround deposits belonging to facies B where the contacts between the facies mostly appear sharp although some interfingering has been observed. The uppermost boundary of facies A is usually towards facies C where the contact has a gradual nature as more sand progressively is introduced to the depositional system. Facies A and facies C also have some common characteristics such as grain size and plane-parallel lamination but are differentiated on sand-content as well as the lack of storm-generated structures in facies A.

Figure 4-1 A) Dark shales cover the lower parts of the mountainsides at Kjuklingetoppane. B) Laminated strata in the Gilsonryggen Mb near Camp 2. C) Detailed photo of the laminated parts of the facies at same location as B). A5 notebook for scale in B) and C).

Interpretation:

The fine-grain size and monotonous appearance of the deposits in facies A together with lack of evidence of subaerial exposure or structures formed by waves or tides implies that deposition occurred in a low-energy environment below storm-wave base and thus the deposits in facies A are assigned to a prodelta setting (Reading and Collinson, 1996)and interpreted to belong to the Gilsonryggen Mb of the Frysjaodden Fm (Steel et al, 1981). The Gilsonryggen Mb represent the lowest part of the third depositional cycle of infilling in the Central Basin (Kellogg, 1975; Steel et al., 1981; Steel et al., 1985).

The fine-grained sediments in facies A are interpreted to represent background sedimentation in the offshore environment while the sands in the upper parts of the facies reflect a more turbulent depositional environment. The large-scale depositional trend in the uppermost depositional cycle in the Central Basin is regressive and thus the presence of sands in the upper parts of facies A could indicate a nearness to storm-wave base. Hence, the sands may be fallout deposits from storm-wave generated suspension deposited below storm-wave base (Helland-Hansen, 2010). Sands in the lower parts of the offshore succession are more likely to have a turbiditic, low-density hyperpycnal flow origin where the sediments are derived from river-borne suspension or unusually vigorous storms (Helland-Hansen, 2010). These beds are commonly ripple- or flat-laminated (Uroza and Steel, 2008) and are interpreted as turbidite-like “tempestites”.

Both hyperpycnal flow and storm-wave dispersal are likely depositional processes for the deposition of sands in facies A when looked at in combination with the overlying facies B and facies C (see descriptions below).

4.2 Facies B1 – sustained flow turbidites

The deposits assigned to facies B1 are usually part of the lowermost cliff-forming sandstones present in the study area. These cliffs are 10-20 m thick and usually located 3-400 m above sea level, where the altitude of the lowermost deposits seems to increase towards the northeast in the study area (see chapter 7). The deposits in facies B1 have a limited lateral extent, estimated to reach a maximum of approximately 1 km and an overall lobe shaped sandbody geometry.

Facies B1 consists of mainly thick (usually 0.5-2.0 m) ungraded or normally graded structureless or stratified (thick-laminated) beds where the grain size varies from very fine to medium sand. Most of the beds are ungraded but normal grading has been observed. The different beds appear to be stacked on top of one another (Figure 4-2 A and B) and are usually separated by flat and sharp amalgamation surfaces although thin layers of mud have been observed between the sandy beds in some locations. Siderite clasts are common near the base of the beds (Figure 4-2 C) but have been observed throughout the beds. Scour marks are present at the bases of some beds. The deposits also contain much organic material and water escape structures and soft-sediment deformation is also observed along with sparce, unidentified burrows.

Figure 4-2 A) Amalgamated Bouma T_A deposits at Kjuklingetoppane B) Detail of stacked beds at Nebben. C) Siderite clasts near bottom of flow at Kjuklingetoppane

Interpretation:

The position of facies B1 near or within the prodelta shales of facies A indicate deposition in a marine environment. There is general agreement among past works on the Battfjellet Fm (e.g. Steel et al, 2000; Plink-Björklund et al, 2001; Crabaugh and Steel,

2004) that turbidite currents were an important mechanism for dispersing sand on the shelf and basin-floor of the Central Basin. A turbidite current is a mass-gravity driven underwater current (Bouma, 1962). The thickness of the deposits in this facies suggests that they are a result of a continuous and steady supply of sand over an extended period of time which could indicate that the turbidites were generated by rivers as hyperpycnal (Plink-Björklund and Steel, 2004). The presence of river-derived structures and deposits in more proximal facies of the Battfjellet Fm (see below) could indicate that the turbidite flows were river derived, and the high content of coal and organic debris is a sign of river-derived flows. (Mellere et al, 2002).

The lack of internal structures and normal or ungraded beds indicate deposition from suspension (Romans et al., 2009), and the thickness of the beds suggests fallout from high density currents (Bouma, 1962; Lowe, 1982). The deposits in facies B1 have been interpreted to belong to the Bouma T_A interval due to their structureless and ungraded or normally graded appearance. The homogenous appearance of the deposits in this facies suggest that sediment discharge was constant over an extended period of time, pointing to deposition from steady to quasi-steady downslope decelerating turbidite currents (Plink-Björklund et al., 2001). Soft-sediment deformation indicates that sedimentation rates were high (Jones and Omoto, 2000).

The lack of the upper four Bouma-intervals could be a result of erosion of the upper strata before the overlying sediments were deposited or that a second turbidity current has overtaken the front of a first current and they were deposited on top of one another. Amalgamation surfaces observed in this facies show evidence of several flows being deposited on top of one another. The observed scours at the bases of some beds indicate that the flow was eroding into underlying deposits.

These sands may be of a similar nature as the turbiditic sandstones of the Bjørnsonfjellet Mb of the Frysjaodden Fm located further west in the basin (Steel et al, 1981) although they are not time-equivalent. Similar to (Helland-Hansen, 1985) the thick sands within the prodelta succession have been interpreted to be part of the Battfjellet Fm in this study.

4.3 Facies B2 – surge type turbidites

Facies B2 consists of fine-grained, 12-18 cm unstructured beds interlayered with very fine-grained ripple or plane-parallel laminated sandstone beds of 2-12 cm thickness (Figure 4-3A). The silt-content is higher in the laminated beds and these beds usually contain more organic material than the unstructured beds, although all the deposits in this facies are rich in organic materials. Facies B2 is found in the lowermost sandstone cliffs in the study area where it usually has an interfingering relationship with the deposits in facies B1 or a sharp contact with the prodelta shales of facies A.

The unstructured beds are brown to grey coloured and a lens shaped geometry is apparent in some of the beds. This suggests that the deposits have a limited lateral extent, estimated somewhere between 3 and 30 m (for each bed). Some of the unstructured beds also have sole marks at the base (Figure 4-3B).

The laminated beds are white to grey coloured and contain much transported organic material (coal) with clasts up to 3 cm long. Similar to the unstructured beds the laminated beds also seem to pinch out and have a limited lateral extent.

Soft sediment deformation, often in the form of ball and pillow structures, is also common in this facies and is present in strata of up to 10 m thickness (Figure 4-3C). Currents ripples are observed in some of the deposits (Figure 4-3D) as well as scattered unidentified burrows (Figure 4-3E).

Figure 4-3 A) Variations between laminated and massive beds at Vengen (. B) Sole marks at the base of flow at the Profeten . C) Soft-sediment deformation at Rypefjellet . D) Current ripples at Rypefjellet . E) Burrow at Rypefjellet .

Interpretation:

The location of facies B2 within the marine shales and the close relationship to facies B1 suggest a marine depositional environment and that turbidite currents also were responsible for the deposition of facies B2. The alternation between structureless and laminated beds could suggest that intensity of the deposition varied with time with the laminated beds being deposited during times of lower flow intensity. The transition from ungraded, fine-grained sands to laminated, very-fine grained sand may reflect variation in the nature of the deposits such as surge-type turbidites that were short-lived and gradually lost their capacity to carry sediments (Plink-Björklund et al, 2001; Crabaugh and Steel, 2004). According to the model of Bouma (1962) the thicker (10-20 cm) beds represent suspension sedimentation from the high concentration bases of overall low-density turbidity currents, interpreted as Bouma T_A deposits, while the plane-parallel and ripple laminated deposits can be interpreted as Bouma T_B and T_C beds (Bouma, 1962). A set of ungraded and laminated beds will thus represent one flow. Lowe (1982) has interpreted this massive these thicker beds to be the result of suspension from high-density flows while the other Bouma deposits were deposited by low-density currents. The observed lamination in the deposits may also reflect varying fallout rates where the ungraded sediments were deposited during times of high fallout rates while the laminated sediments were laid down from flows with low fallout rates and traction (Plink-Björklund et al, 2001; Romans, 2009).

The high content of organic material suggests that the sediments transported by the turbidity currents were derived from land. The sparce bioturbation could suggest that sedimentation rates were high, something which also is confirmed by the presence of soft-sediment deformation (Jones and Omoto, 2000).

4.4 Facies B3 – turbidite channel deposits

This facies is usually found interlayered with or on top of facies B1 and B2 or surrounded by shales at a level between the basin-floor fans and the shoreface deposits. Facies B3 consists of deposits ranging from very fine to medium/coarse grained sandstone (Figure 4-5). The thickness of the facies is usually between 1-4 m although thicknesses up to 10 m have been observed. The deposits in facies B3 are characterized by short lateral extent with a maximum of 50-60 m. There appears to be several erosion surfaces within the facies. An upward thinning of the beds has been observed and tabular cross-stratification and structures formed by wave reworking are present in the upper parts of the deposits. Siderite clasts are also common.

Figure 4-4 Medium and coarse grained deposits together with siderite clasts in facies B3

Interpretation:

The location of this facies above or interbedded with turbidite facies, coupled with its position below shoreface deposits suggests that also this facies was deposited by turbidity currents. The position above the other turbidite facies as well as the wave-modified structures could suggest deposition at a more proximal location than the underlying turbidite facies, such as the slope. Erosion surfaces, tabular cross-stratification and the limited lateral extent of the deposits suggest that this facies was confined to a channel. The channel interpretation could suggest a close relationship to a mouth bar system. It is likely that these channels were distributing sands to the slope and basin floor. The coarser grainsize observed in this facies compared to the underlying facies B1 and B2 indicates that the coarsest sediments were deposited on the slope channels before the turbidite currents reached the basin floor.

The erosion surfaces in this facies probably represent several sedimentation units within this facies and indicate that many flows were transported through this area (Romans et al., 2009).

4.5 Facies C – offshore transition

This facies is observed in several locations in the study area and is found vertically higher than facies A and B. Facies C usually constitutes the lower parts of the second sand-forming cliffs that are present on most mountainsides in the study area. Muds and silts are the dominant grain sizes in the lower parts of the facies but the sand content gradually increases upwards and the upper parts of facies C are dominated by very-fine sandstone. Thus, an upwards-coarsening trend is apparent in the facies. The thickness of the facies is usually between 1 and 6 m.

The most common sedimentary structures are low-angle, wavy three-dimensional structures that truncate each other (Figure 4-5 A and B). Plane-parallel lamination, ball and pillow structures and other soft-sediment deformation structures are also present (Figure 4-5 C). In a few places symmetrical ripples cap the truncated, wavy three-dimensional structures. Some bioturbation has also been observed, including the bow-formed trace fossil *Glyhipichnus* (Figure 4.5D) *((*Goldring et al., 2002)

Figure 4-5 A) Typical appearance of the upper parts of facies C with large-scale low-angle, wavy three-dimensional truncating sand structures at Rypefjellet. B) Soft-sediment deformation at Kjuklingetoppane. C and D) Detail of the low-angle, wavy three-dimensional structures at locations Above Nebben and Vengen.

Interpretation

The downcutting, low-angle wavy three-dimensional beds are interpreted as hummocky cross-stratification and swaley cross-stratification. The processes that form these structures are not known but the general agreement is that storm waves are responsible for their formation (e.g. Swift et al., 1983; Duke et al., 1991). The dominance of these structures suggests a depositional environment above storm wave base. The symmetrical ripples that have been observed capping the hummocky cross-stratified deposits were probably produced by oscillatory currents and may indicate the waning stages of a storm (Olsen, 2008).

The great proportion of silts and mud in the lower part of the facies succession suggests that background sedimentation was significant during the early stages of deposition of this facies while the gradual increase of sands upwards in the facies indicate a shallower and more energetic depositional environment. This is consistent with a regressive setting. Thus the deposits in facies C are interpreted to belong to the offshore transition zone where the deposition of the sediments in this facies took place between storm wave base and fair-weather wave base (Reading and Collinson, 1996) an environment with alternations between high and low energy conditions. This means that during fair-weather sediments will settle from suspension while during storms the bottom is affected by oscillatory and shoaling waves, supplemented by storm-generated currents. The dominance of storm-dominated structures preserved in this setting is explained by the lower preservation potential of fair-weather mud and silt deposits as opposed to the sandy hummocky cross-stratification deposits from storms.

The dominance of storm-generated structures in this area show that storm-wave processes were important for reworking and redistribution of the sediments in the Battfjellet Fm. The close relationship between facies A and facies C, with the gradual increase in sand-content shows that the process was gradual and record a shallower and more energetic depositional environment. Helland-Hansen (2010) interpreted the soft-sediment deformation in the offshore transition zone to be a result of rapid deposition of storm-generated sands.

Deposits with facies characteristics suggesting deposition in facies C are also found resting on top of turbidite deposits belonging to facies B, and this suggest that the turbidite depositional process may be closely related to the storm-wave processes or storm waves may have triggered turbiditie flows.

With evidence of both current derived sediments and wave-reworked deposits in this facies it is possible that both these processes were important for forming the offshore transition zone deposits.

4.6 Facies D – lower shoreface

This facies consists of mainly very-fine and fine-grained sandstones and the facies has a thickness between 2 and 5 m in the study area. The colour of the deposits varies from grey to brownish-yellow. Facies D is dominated by plane-parallel laminations and symmetrical ripples (Figure 4-8A and C). The ripple crests are oriented approximately N-S in the study area. There are some low-angle, three dimensional trough structures similar to those observed in facies C but these are less frequent than in the offshore transition zone. Water escape structures and climbing ripples (Figure 4-8D) are observed in facies D. Facies D appears to be laterally consistent over large distances. Bioturbation is present in the form of some unidentified burrows as well as the trace fossil *Glyhipichnus* have been observed in this facies. Soft-sediment deformation in the form of ball-and-pillow structures are important in the upper reaches of this facies where deformed strata have a thickness up to 2 m (Figure 4-8 B).

Facies D is usually located on top of facies C where it continues the coarsening upwards trend seen from the previous facies. Facies C and D are distinguished on the presence of symmetrical rippled strata in facies D and that the muddy deposits are confined to the offshore transition facies.

Figure 4-6 A) vertical section of facies D at Vengen. B) Soft-sediment deformation at the location Above Nebben (photo by E. Ellens). C) Symmetrical ripples, Above Nebben. D) Climbing ripples, Above Nebben.

Interpretation

The trough shaped stratification and lamination in this facies are, similar to in Facies C, interpreted to be hummocky and swaley cross-stratification and show that storm waves played an important role in shaping this environment. However, smaller amount of storm-wave generated structures present in this facies indicates compared to the underlying facies C indicates less energetic wave-action in facies D. Less influence from storm-waves upwards in the succession is consistent with a regressive depositional environment where shallower the water depths are getting less influenced by storms.

The symmetric ripples present in facies D are interpreted to have been formed by fair-weather waves and suggest a location above fair-weather wave-base. Fair-weather wave deposits were formed as a result of fair-weather aggradation (Helland-Hansen, 2010). Plink-Björklund et al (2001) interpreted packages consisting of sandstones with wave-ripples and plane-parallel lamination to be part the most proximal areas of the delta front.

Water escape structures indicate that sedimentation rates were high, similar to those observed in previous facies. Climbing asymmetric ripples indicate deposition from unidirectional currents, such as fluvial, which suggests episodic deposition from river floods in an environment that is otherwise dominated by storm-wave processes (Hampson and Storms, 2003). It could be speculated that the hyperpycnal currents discussed in facies B that may have been river-derived also are responsible for the deposits of unidirectional current deposits found in this facies too.

4.7 Facies E – middle shoreface

This facies consists of very-fine to fine grained sandstone with thicknesses between 1 and 10 m observed in the study area and with an average around 3 m. Facies E is usually located between facies D and F (Figure 4-7 A).

The most common sedimentary structures in facies E are symmetrical ripples and small downcutting channels filled with plane-parallel and symmetric ripple-laminated deposits (Figure 4-7 B). These channels are usually 10-50 cm wide and 5-40 cm deep. In the uppermost reaches of the facies at the localities at Kjuklingetoppane there are some 40-100 cm thick strata that display intense bioturbation with *Ophiomorpha* trace fossils

(Figure 4-8 C and D). The bioturbated strata have a limited lateral extent (maximum of 40 m).

Figure 4-7 A) Vertical stacking of shoreface deposits at Kjuklingetoppane. B) Small, downcutting channel at Kjuklingetoppane. C) and D) *Ophiomorpha* trace fossils in the middle shoreface deposits at Kjuklingetoppane

Interpretation:

The presence of symmetrical ripples indicates that fair-weather waves were important for forming this facies while the lack of hummocky and swaley cross-stratification suggests a more proximal depositional environment than that of the underlying facies, and hence it is assumed that storm-waves were of little importance for generating structures, in fact storms were probably eroding rather than depositing (Helland-Hansen, 2010). Facies E continues the upward coarsening regressive trend observed in the previous facies. The location above the lower shoreface may imply that these deposits belong to the middle shoreface. Helland-Hansen (2010) interpreted the middle shoreface to be influenced by rapidly shifting fair-weather currents with scouring and vertical infilling of structures.

The presence of *Ophiomorpha* fossils in facies E suggest a shallow marine depositional environment (Hampson and Howell, 2005; Maceahern et al., 2005) although this fossil can be present in different facies and hence can not be used to specify the depositional environment in more detail.

4.8 Facies F1 – upper shoreface/foreshore

This facies consists of fine- to medium-grained sandstones in 3-10 m thick packages. Deposits belonging to facies F1 are usually located above facies E and/or below facies G and usually form the top of the uppermost sandy sequences observed in the study area.

The most common sedimentary structures in facies F1 are horizontal and low-angle lamination and tabular cross-stratification (Figure 4-8). The dip azimuths of the cross-stratifications do not show any preferred directional trends. There are also some climbing ripples present. Transported organic material (coal clasts), leaf-fossils and petrified wood are common in this facies. Some vertical burrows have also been observed.

Figure 4-8 A and B) Plane-parallel and low-angle stratification B) Tabular cross-stratification at locations Above Vengen and at Above Pønketoppen (log 2) C) Plane-parallel and low-angle stratification at Kjuklingetoppane

Interpretation:

The coarsening upwards trend observed in the previous facies continues into facies F1 and also indicates a continuation of the shallowing upwards trend interpreted from the previous facies. This means that facies F1 was deposited in a shallower environment than facies E which also is apparent from the stratigraphic position just below the delta plain deposits of facies G. The leaf fossils and petrified wood indicate a position near the shoreline, but the lack of subaerial exposure suggests that the depositional environment was marine. The sparce bioturbation observed may imply a turbulent depositional environment. The tabular cross-stratification may represent migration of dunes where the highly variable orientations of the foreset azimuths could be a result of unidirectional shifting currents in the upper shoreface (Helland-Hansen, 2010), possibly from longshore currents which can produce complex current patterns. Such complex patterns are particularly common at dissipative shorelines (Orton and Reading, 1993). Longshore currents and wave-action are processes that are important in the upper shorelface (Clifton, 2006).

4.9 Facies F2 – fluvially influenced channel in the shoreface

This facies is in close contact with facies F1 but is distinguished by the presence of an underlying erosional base that cuts down into facies F1, a limited lateral extent observed to be up to 30 m in the study area, and the transition into this facies is also often accompanied by an abrupt change in colour and/or grain-size (Figure 4-9 A and B). Facies F2 consists of fine to medium grained sandstones where the content of mud and silt appears to be very low. The most common sedimentary structure in facies F1 is tabular cross-stratification. Siderite clasts have been observed within this facies as well as some strata that contain symmetrical ripples or bi-directional cross-stratification (Figure 4-9 C). Volumetrically this facies is of minor importance compared to facies F1.

Figure 4-9 A and B) Deposits in facies F2 with showing the limited lateral extent of the facies, changes in colour from the underlying facies and tabular cross-stratification C) Bi-directional cross-stratification

Interpretation:

The erosional bases below this facies and the limited lateral extent suggest that the deposits in facies F2 were laid down in channels that erode down into the wave-dominated shoreface deposits. The cross-stratification and the generally clean appearance of the sands in this facies suggest transport and deposition by rivers, where muds and silts have been washed out and transported further seawards.

The bi-directional cross-stratification this facies could indicate that the deposits were influenced by tidal currents while the symmetrical ripples suggest waves reworked the sediments. These structures and the processes that formed them are interpreted to be of minor importance in the facies due to the small volume occupied by these structures. Thus, fluvial processes are believed to have been the most important for shaping this facies and the deposits are interpreted as the shoreward parts of a fluvial channel, and facies F2 is closely related to the fluvial channel in facies H. The sediments of facies F2 are important for the understanding of the depositional processes within the Central Basin as these channels probably supplied sediments to the Central Basin. The presence of these channels within the wave-dominated upper shoreface (facies F1) indicate a closeness to the basin. Fluvial channels will often deposit their coarsest bedload-transported sediment where they enter the basin and form mouth-bars at the point of flow expansion (Bhattacharya, 2006). Fluvial channels that supply sediment to the basin can also be filled with sediments, especially if the channel is abandoned (Bhattacharya, 2006). Estuarine fills in fluvially incised valleys have been reported in the Reindalen and Van Keulenfjorden area (Mellere et al., 2002; Plink-Björklund, 2005; Løseth et al., 2006; Plink-Björklund and Steel, 2006) but this interpretation requires both a drop in sea-level and more evidence of tidal currents to be applicable to this study and hence a tidal-fluvial channel interpretation, as observed by Uroza and Steel (2008) is probably more likely in the Rypefjellet area.

4.10 Facies G – delta plain

Facies G consists of mainly silt although successions with very-fine to fine sand are common at some localities (Figure 4-10A). The facies is found on top of facies F and is only present on the highest peaks in the study area, usually located from around 700 m above sea-level and higher. The change in grainsize from facies F to G is drastic, still, much scree-cover at the transition makes it hard to find the exact location of the boundary. The thickness of facies G is unknown but is believed to be extensive (see chapter 3).

The colour of the deposits in facies G ranges from red to grey-black. Soft-sediment deformation, leaf-fossils, root-traces and laminated strata are common sedimentary structures (Figure 4-10 B and C). Strata of ripple laminations have also been observed

(Figure 4-10 D). The deposits also contain much organic material including coal-laminations.

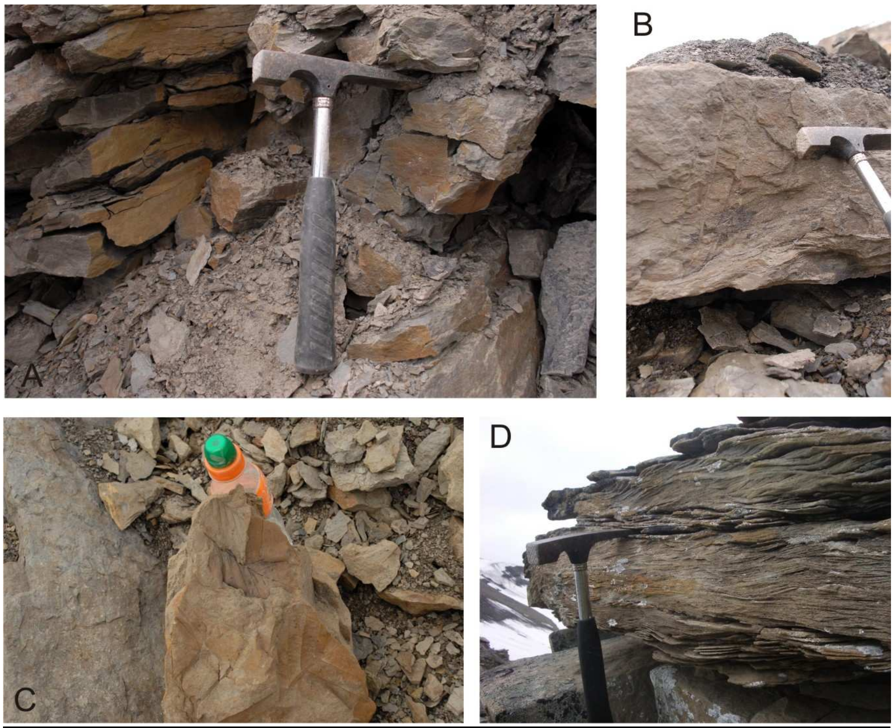


Figure 4-10 A) Typical exposure of the Aspelintoppen Fm at Kjuklingetoppane. B and C) Root traces and leaf fossils at Above Vengen. D) Ripples in the Aspelintoppen Fm at Above Pønketoppen

<u>Interpretation:</u>

The presence of rooted horizons in this facies suggests that deposition took place above sea-level. Coal-laminations are also consistent with deposition in a continental environment and hence facies G has been interpreted to represent delta plain deposits that belong to the Aspelintoppen Fm, the uppermost part of the third depositional cycle of infilling in the Central Basin (Steel et al, 1981). Dreyer and Helland-Hansen (1986) suggested that the fine-grained sediments in the Aspelintoppen Fm were deposited by suspension from overbank flooding or crevassing in a distal bay or poorly drained area of the floodplain which could explain the generation of ripples observed in this facies. The high degree of soft-sediment deformation in this facies was interpreted by Steel et al (1981) as earthquake shocking while Olsen (2008) suggested that density inversion due to the heterolithic nature of the deposits could have caused the deformation.

4.11 Facies H – fluvial channel fill

This facies is found within the delta plain deposits of facies G of the Aspelintoppen Fm and in the study area it has only been observed on the mountain Pønketoppen and on Profeten. Facies H consists of mainly pebbles and coarse sands (Figure 4-11) that form confined bodies that appear to cut down into facies G. The deposits in facies H show a fining upwards trend and the thickness of the facies in the study area was measured to be from 2 - 6 m. There are also observations of several fining upwards units stacked on top of each other. Sedimentary structures observed are plane-parallel and tabular cross-lamination and stratification and climbing ripples.

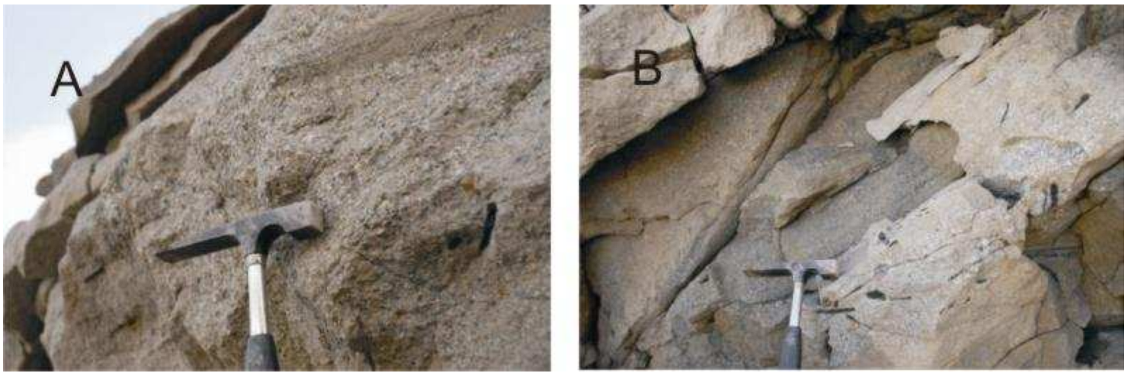


Figure 4-11 A and B) Coarse sediments and coal clasts in facies H at Above Pønketoppen (log 8)

<u>Interpretation:</u>

The coarse grain-size of the deposits in facies H indicate higher energy in this facies than in the surrounding facies G. The limited lateral extent of the sands, erosive lower boundary and the fining upwards trend with coarse sands in the base indicate that this facies was deposited in a channel. The surrounding continental deposits make it likely that the channel was fluvial. It is likely that the deposits in facies H are closely related to the underlying marine facies where the fluvial channels were feeding sediments to the delta system that later were dispersed on the slope and basin floor. Hence, it can be assumed that the fluvial channels of this facies culminate in a mouth bar structure on the delta-front although observations of this has not been made in the study area.

This facies has only been observed within the Aspelintoppen Fm in one location in the study area, and hence no trends in the transport direction can be suggested. Past workers (e.g. Helland-Hansen, 1985, Olsen, 2008) have reported a generally eastward transport-direction of the fluvial channels.

4.12 Transgressive deposits

These deposits consist of mainly of shales and have much in common with facies A. The transgressive deposits are located between sandstone deposits that belong to different facies and represent small-scale transgressions in the overall regressive setting in the Battfjellet Fm.

5 Facies Associations

5.1 Introduction

From the previous chapter it is clear that the facies are not randomly stacked together. The various facies have been grouped into facies associations where each facies association consists of several facies that are genetically related and have some environmental significance (Walker, 1984).

The position of distal facies below proximal facies, displayed by the coarsening and shallowing upwards trends, reflects a seaward migration of the shoreline. This trend is apparent in facies associations 2, 3 and 4. The location of the offshore shales of the Frysjaodden Fm underlying the Battfjellet Fm and the continental Aspelintoppen Fm above also show the overall regressive depositional setting in the Central Basin during the deposition period. Despite the overall regressive nature of the Battfjellet Fm transgressive events were common, reflected by fine-grained deposits capping the sandy coarsening upward packages. However, these transgressive deposits, which were discussed in chapter 4, are volumetrically of little importance relative to the regressive deposits.

Facies association	Facies	Environmental interpretation
FA1	B1, B2, B3	Basin-floor fans and slope-channel turbidites
FA2	A	Prodelta shales
FA3	C, D, E, F1	Wave-dominated regressive shoreface
FA4	F2, G, H	Delta plain and proximal delta front

5.2 Facies Association 1 (FA1) – basin-floor and slope turbidites

FA1 consists of facies B1, B2 and B3, and this facies association makes up the lowermost sandstone cliff in the study area and also include some scattered outcrops located between the two main sandbodies. The deposits in FA1 are over- and underlain by marine shales belonging to facies A. The whole facies association is usually 30-50 m thick and up to 1 km wide (Figure 5-1A). The facies association is usually present in the form of a lense shaped cliff-forming sandstone body situated below and detached from the other sandstone successions higher up in the Battfjellet Fm.

Facies B1 consists of structureless sandstone beds interpreted as turbidite deposits belonging to the Bouma T_A division (Bouma, 1962). These beds were most likely deposited from quasi-steady to steady decelerating turbidity flows. Facies B2 consists of alternating structureless and laminated beds and the deposits in this facies have been interpreted as turbidity deposits belonging to the Bouma T_{ABC} division (Bouma, 1962) where the lamination is caused by a decrease in the vertical flux of the depositing current and accompanying traction deposition. Facies B3 consists of sandstone beds identified by an erosional base, cross-stratification and generally a limited lateral extent. Facies B3 is usually found in the uppermost parts of the sandstone cliffs or scattered in the area between the two sandstone bodies and has been interpreted as slope channels filled with turbiditic deposits.

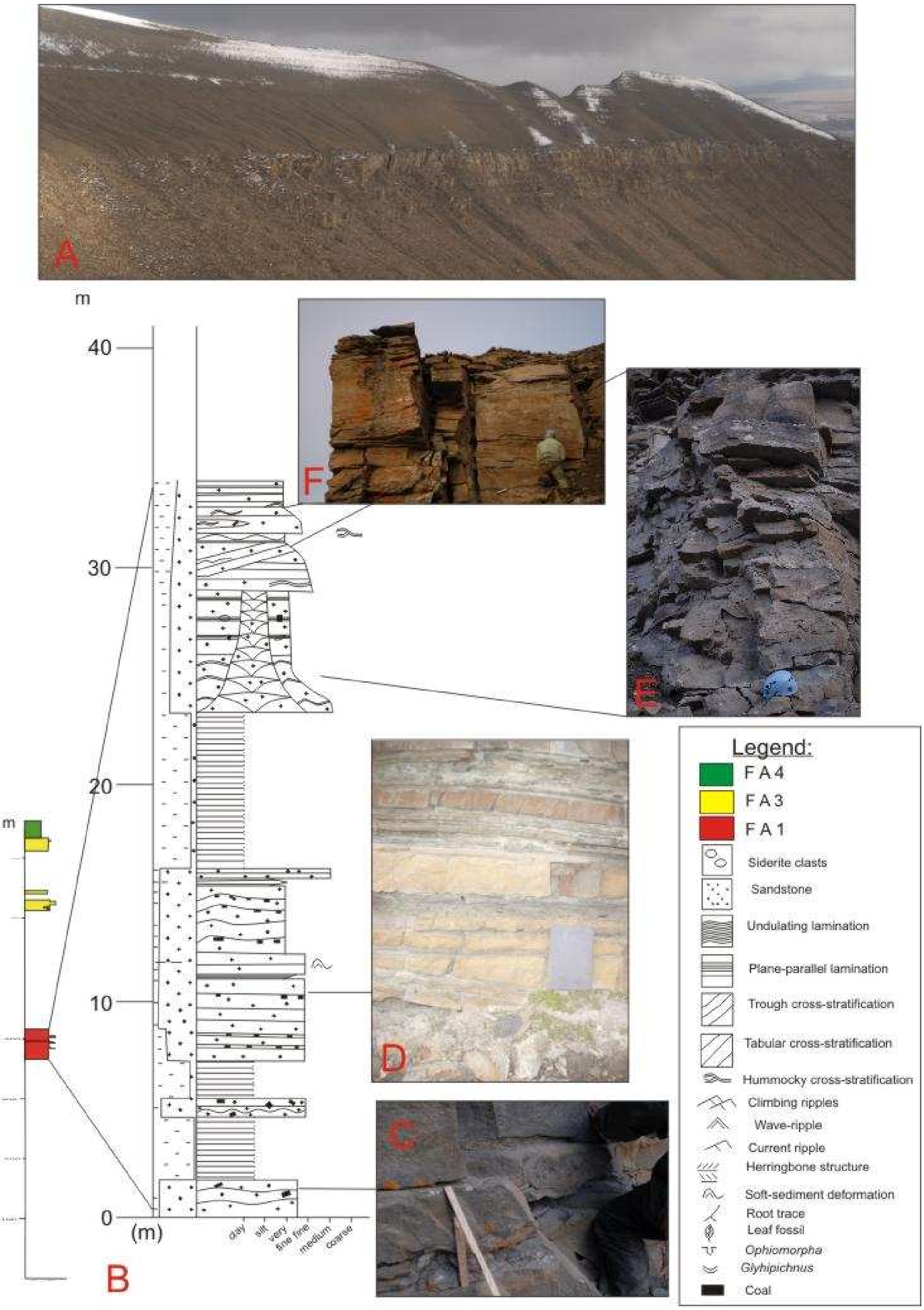


Figure 5-1A) Basin-floor fan deposits at the southern side of Vengen. B) Log-example from Vengen showing both the position of FA1 on the mountainside and a detailed log of FA1. C) Deposits belonging to facies B1 on the basin-floor D) Deposits belonging to facies B2 on the basin-floor E and F) Deposits belonging to facies B3 in the proximal areas on the basin-floor fan complex

Hence, FA1 consists of various deposits of turbiditic origin. The strong evidence of sediment gravity-flow processes, lack of wave-generated structures as well as the position within the marine shales make it reasonable to interpret this facies association as a turbidite lobe complex in the form of basin floor fans and their distributary channels. These sediments were all deposited in a prodelta setting (see below for delta interpretation). Crabaugh and Steel (2004) indicated that basin-floor fans present in the Battfjellet Fm usually were finger- to fan-shaped. Upward thickening of beds have been interpreted to show an increase in the proximity to the source area (Hampson and Storms, 2003). The scattered outcrops that are located between the main lobe-complexes and the overlying shoreface deposits are interpreted as slope-channel deposits and these were responsible for transporting the sediments down along the slope and out on the basin-floor.

5.3 Facies Association 2 (FA2) Prodelta shales

This facies association consists of facies A. FA2 is usually present on the lower slopes on the mountainsides in the study area and have been interpreted as prodelta shales. This facies association consists of only facies, A which was described, in more detail in chapter 4.

5.4 Facies Association 3 (FA3) Wave-dominated shoreface

This facies association consists of facies C, D, E and F1 where it is fully developed. The base part of this facies association usually located just below the second cliff-forming sandstone succession observed in the study area, and the facies association continued to the top of the mountains (Figure 5-2). It is mainly in this facies association that interruptions in the stacking pattern of the Battfjellet Fm in the form of marine shales is apparent, where shales capping the sandstone deposits are interpreted as small transgressions. The deposits found on top of the transgressive successions are often deposited in a more distal environment than those below.

The lowermost deposits in FA3 are deposits belonging to the offshore transition zone, facies C, which contains storm-generated structures such as hummocky cross-stratification as well as finer-grained sediments interpreted to have been deposited by background sedimentation. Above the offshore transition zone shoreface deposits belonging to facies D, E and F1 are present. The lowermost part of the shoreface is dominated by storm-wave generated structures such as hummocky- and swaley cross-stratification. In the more proximal parts of the shoreface succession the storm-generated structures are replaced by symmetrical ripples and plane-parallel lamination interpreted to have formed through the action of fair-weather waves. The transition from storm-wave generated structures to fair-weather wave generated structures reflect a gradually shallower depositional environment being less influenced by storms. Some bioturbation assigned to shallow marine environments is also observed. In the most proximal parts of the shoreface succession there are structures like small-scale troughs and channels and thick units with planar- and trough cross-stratification. The uppermost, cross-stratified deposits are interpreted to be the product of migration of sand-dunes. The sandy shoreface successions are capped by delta plain deposits that belong to facies G.

The facies succession in FA3 reflects a gradually upwards shallowing environment dominated by wave-generated structures. The facies successions in environments where the energy from waves and storms is high are often very distinctive (Hampson and Storms, 2003). Hence, the deposits in FA3 have been interpreted as being deposited in a wave-dominated shoreline based on the following evidence:

- The abundance of wave-generated structures coupled with the lack of fluvial and tidal derived structures show that the wave-energy in the basin must have been high. The sediments were derived from the river that drained the areas to the west of the shoreline, but no evidence of offshore-directed unidirectional currents are present in this facies association, suggesting that all the sediments have been reworked by waves. Thus, it is likely that sediments were transported along the shoreline by longshore currents and deposited them in environments such as barrier bars, spits, beach-ridges and strandplains

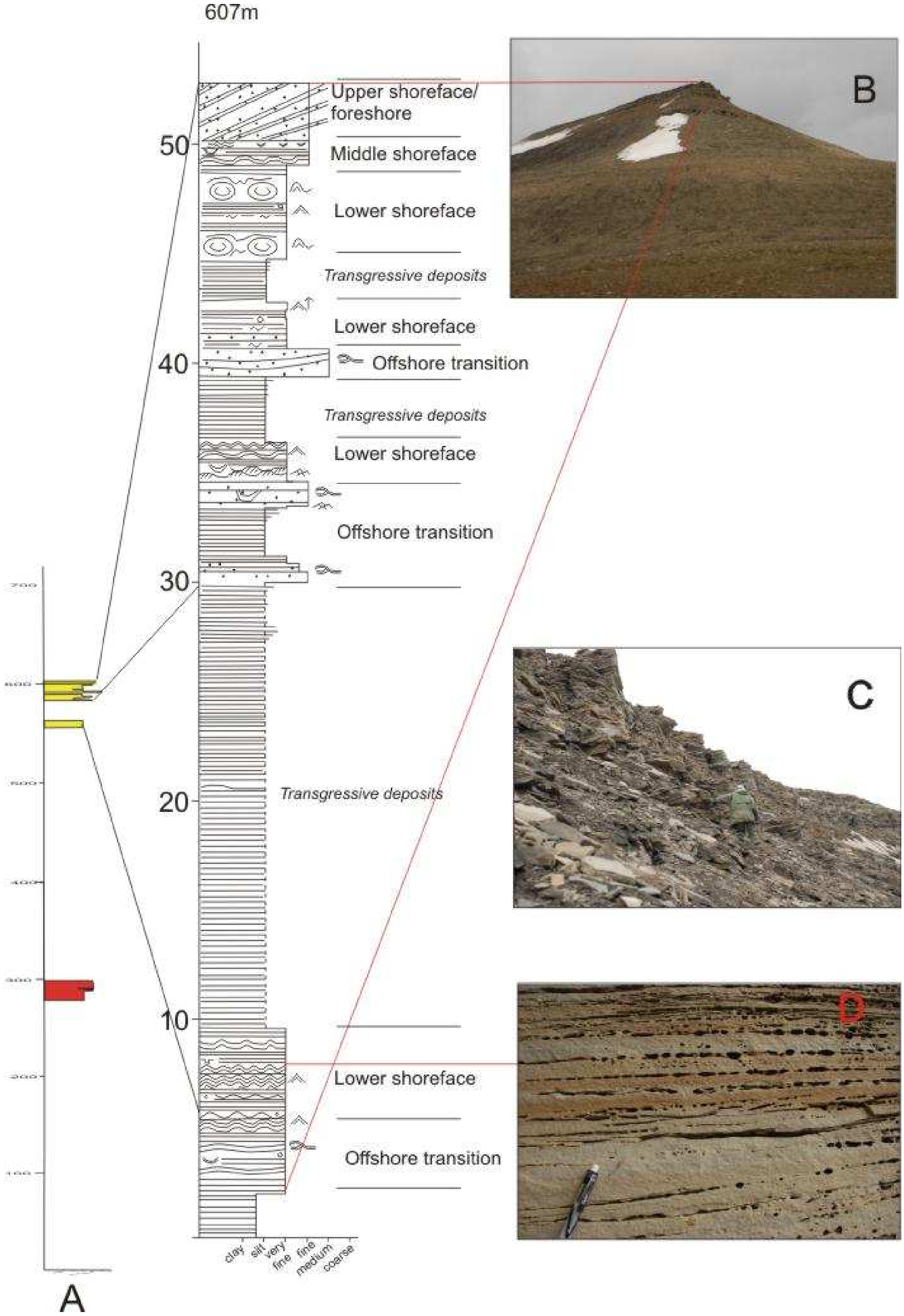


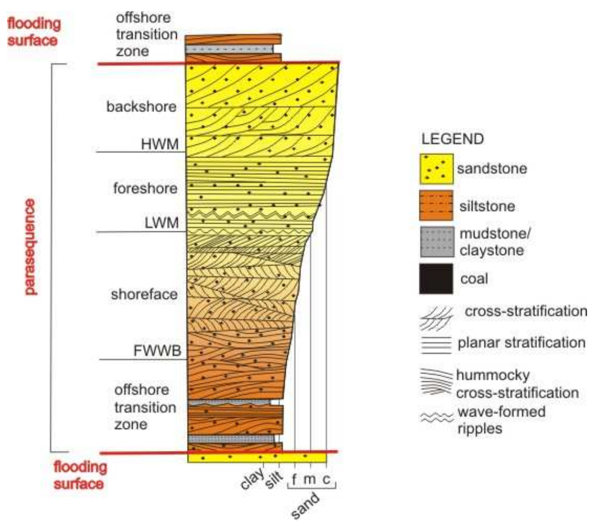
Figure 5-2 Facies association 3 A) Log example and photo of outcrops from Above Nebben. B) Outcrop example of shoreface sequence at Kjuklingetoppane. See Figure 5-1 for legend.

- Modern wave-dominated deltas such as the Grande-rivière-de-la-Balline in Hudson Bay, northern Canada (Hill et al., 2003) display a similar facies succession in the delta front environment as that observed in FA2.

- The vertical succession is similar to that of a prograding strandplain as described by Olsen et al, (1999) in the Upper Cretaceous Cliff House Fm in south-western Colorado and by Hampson and Storms (2003) in the Book Cliffs in Utah. A sequence typical of a strandplain is shown in figure 5-3 below.

No direct river effluents have been observed in FA3 and hence it can be questioned whether the facies association represents a strandplain or a wave-dominated delta as both these depositional environments produce coarsening upwards sedimentary successions (Bhattacharya and Giosan, 2003). Distinction between strandplains and wave-dominated deltas are not straightforward, as wave-dominated shorelines are complex and characterized by alongshore variability in the depositional environment. However, the stacking pattern of the shallow marine sandstones of the Battfjellet Fm suggests a deltaic depositional environment where switching of delta lobes has been important for the shaping of the Battfjellet Fm (Helland-Hansen, 2010). The wave-dominated delta interpretation of the Battfjellet Fm has also been reached by several past workers (e.g. Steel et al, 1985, Steel et al, 2000, Deibert et al, 2003; Helland-Hansen, 2010).

Figure 5-3 Typical vertical sequence of strandplain (redrawn after Coe, 2003)



5.5 Facies association 4 (FA4) – delta-plain and proximal delta front

This facies association consists of facies F2, G and H and is expressed by channels cutting down into the uppermost shoreface deposits and continues up in the continental deposits of the Aspelintoppen Fm.

The volume of facies F2 present in the shoreface is insignificant compared to facies F1 and facies G and H are only present on the highest peaks in the area. Although the volume of this facies association is small in the study area as a whole it is important as the delta deposits transport sediments from the source area to the shoreline, leading to deposition of the sand and progradation of the shoreline further into the basin.

Facies F2 and H can be regarded as different down-dip expressions of fluvial channels where facies H is located within the delta plain while facies F2 is the seaward extent of the channel as it erodes into the shoreface, probably closely related to the mouth-bar.

Thus, facies association 4 has been classified as the fluvially dominated portion of the delta, comprising the delta plain and distributary channels. Using the traditional Gilbert delta classification this facies would be the topset of the delta, while it is the “undaform” portion of the delta when using the classification of Rich (1951).

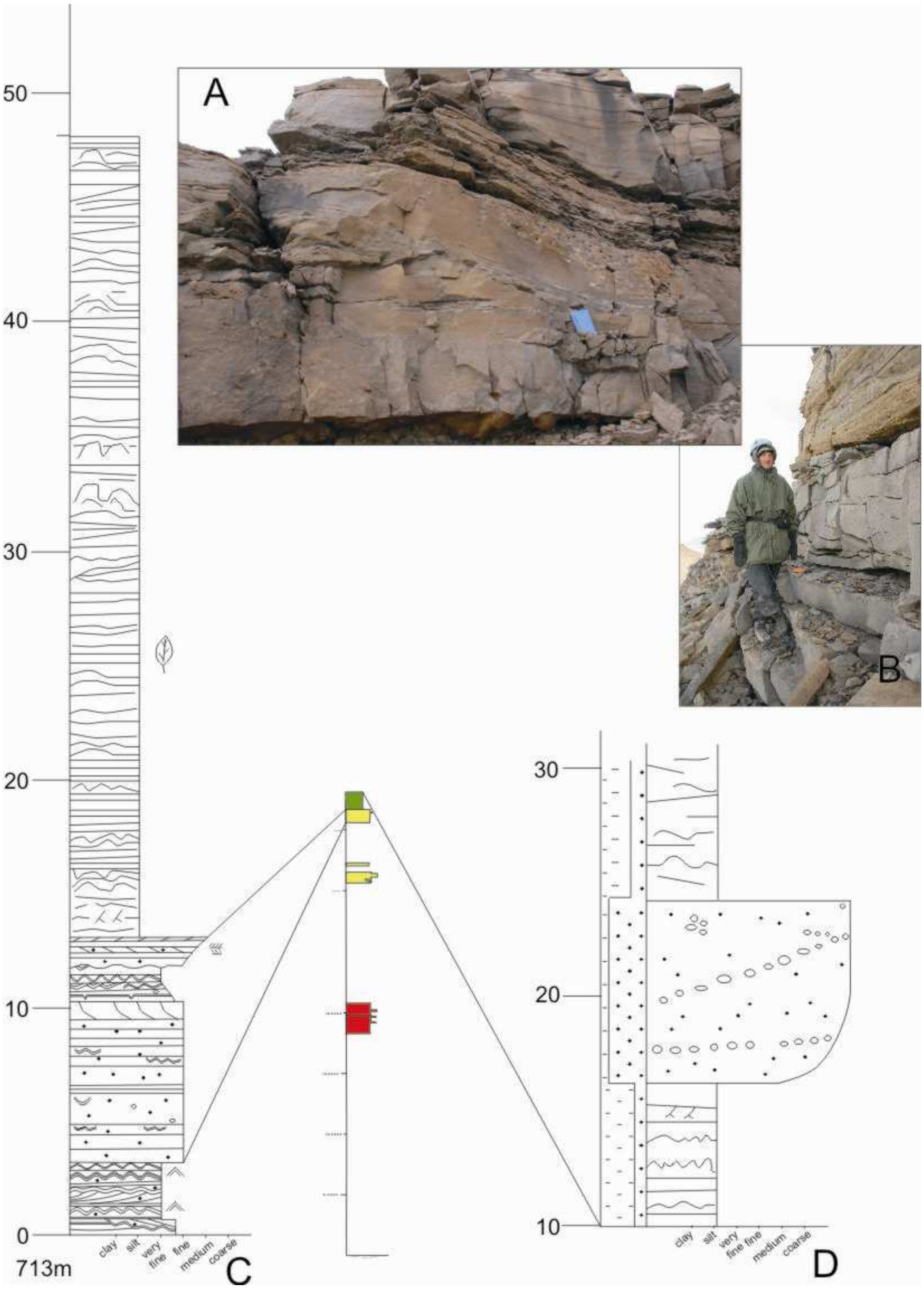


Figure 5-4 Facies association 4 A) Channel incised in the shoreface at Vengen B) Log examples of facies association 4 from Vengen. See Figure 5-1 for legend.

6 Sandbody Geometry

6.1 Introduction

The aim of this chapter is to combine the data from the logged vertical sections (facies, facies associations, elevations) with the large-scale observations of the sandbody of the Battfjellet Fm. All these data were brought together in a 3D model made in the computer programme Google SketchUp. Some snapshots from this model are shown below in Figure 6.1, showing lines along what is inferred to be strike- and dip-parallel transects. The model shows the distribution of the sands in the Battfjellet Fm and how they relate to the over- and underlying formations. To make a robust model of the sandbody it is also important to include data of the paleodrainage directions.

Correlation panels of the Battfjellet Fm have been made in the past along an east-west stretch from the southern coast of Van Mijenfjorden (Olsen, 2008, Stene, 2009), and in a southeast trending direction in Van Keulenfjorden (Steel and Olsen, 2002). The advantages of making a 3D model as opposed to a 2D transect are obvious, as a 3D model can be used to show the spatial arrangement of the deposits and it is also possible to make 2D transects from the 3D model.

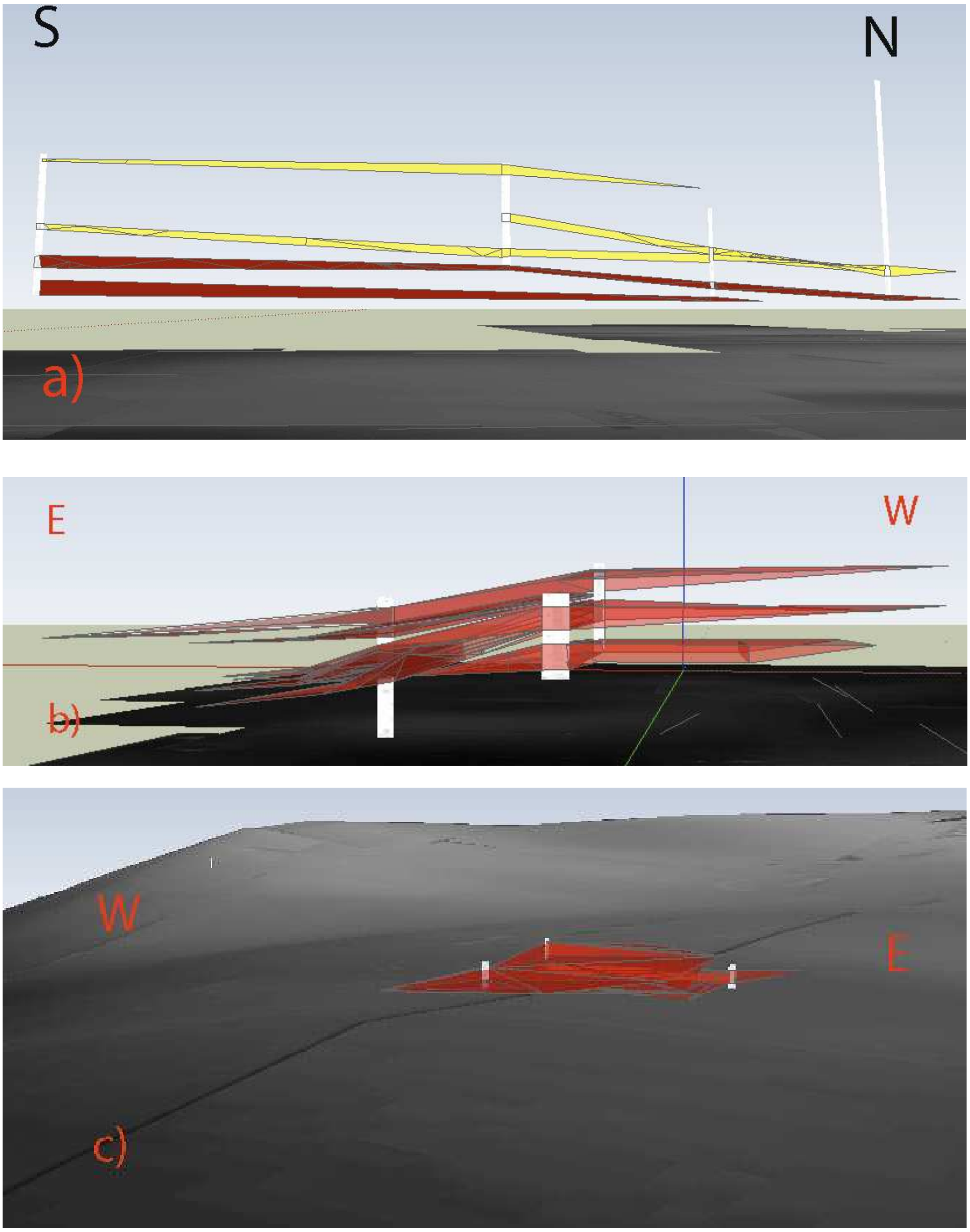


Figure 6-1 A) Part of model showing the north-south transect of Rypefjellet, showing a sloping trend towards the north. B) Transect showning the basin-floor fans at Vengen without the elevation model. C) Basin-floor fans at Vengen with topography

6.2 Methods

For both the 3D model and the selected 2D sections the various vertical logs were placed with correct vertical and horizontal distance from each other and sea-level. From the observations in the study-area it is clear that the sandy surfaces in the study area, interpreted as clinothems (Rich, 1951), slope to the east-northeast (Figure 6-2).

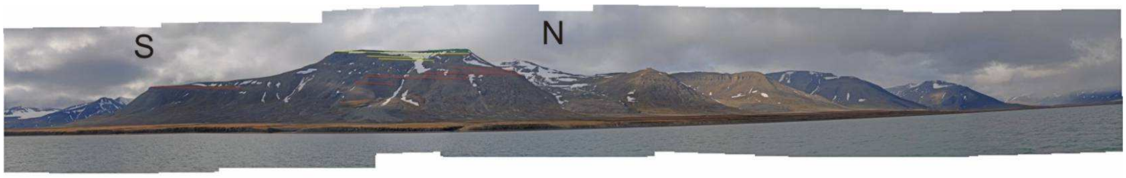


Figure 6-2 Photo showing the dipping horizons of the sands in the Battfjellet Fm on the mountainside of Rypefjellet.

6.2.1 Correlation principles

The study area has been cut by several glacial valleys and these give insight to the 3D architecture of the Battfjellet Fm. When correlating between the different sandbodies these valleys, along with glacial cover, causes problems when visually attempting to connect the different sandbodies. Hence, various sequence stratigraphic concepts have been used along with field observations and photos to correlate the sandbodies in the Battfjellet Fm. These principles are:

- Sands of shallow marine and turbidite origin thin out towards the east/northeast (Chapter 7)
- Thick sands extend further into the basin than thinner sands
- There is a continuous lateral transition from continental through shallow-marine into offshore deposits (Figure 6-3)
- Basin-floor fans are positioned as the stratigraphically lowest sands of the Battfjellet Fm deposits and are in most cases detached from the shallow marine deposits
- Individual facies gradually disappear in proximal and distal directions and are replaced by other facies as the water depths shallow or deepen
- The shallow-marine successions should be thinning in a “bottom-up” manner in proximal areas and in a “top-down” manner in distal areas (Fig 6-4)

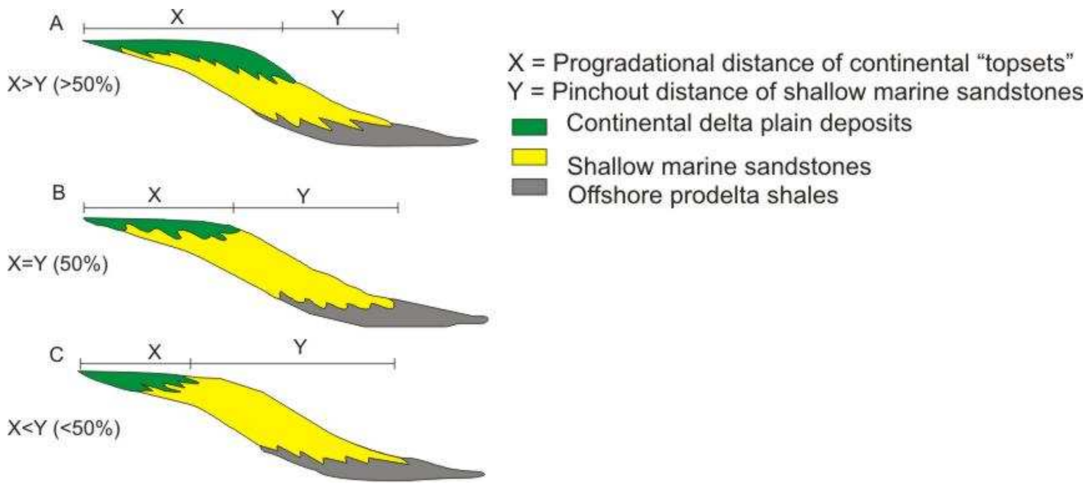


Figure 6-3 the figure shows possible scenarios for the progradation of a regressive system. The pinchout distance of the depositional system determine the probability of observing continental deposits between shallow marine systems. In the study area it is rare that continental deposits are present in-between the shallow marine deposits. Hence, it is likely that the progradational style of this system resembles that of C in the figure (after Olsen, 2008, after Helland-Hansen, 1985).

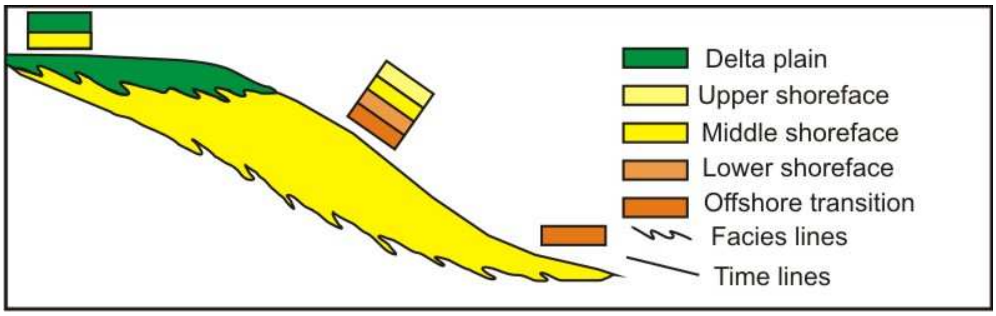


Figure 6-4 A simplified model showing the expected geometry and facies distribution of a shallow marine sandstone unit of the Battfjellet Fm (redrawn after Olsen, 2008).

The upper surface of the clinoforms represents time-lines that represent transgressions, as the deposits were flooded. This allows for a chronostratigraphic correlation (Figure 6-5; Gani and Bhattacharya, 2005). When using this type of correlation it is clear that the sediments were deposited on slopes that dipped basinwards. The top surfaces of the clinoforms represent timelines where the ending stage of sand-deposition is marked by capping from shale deposits.

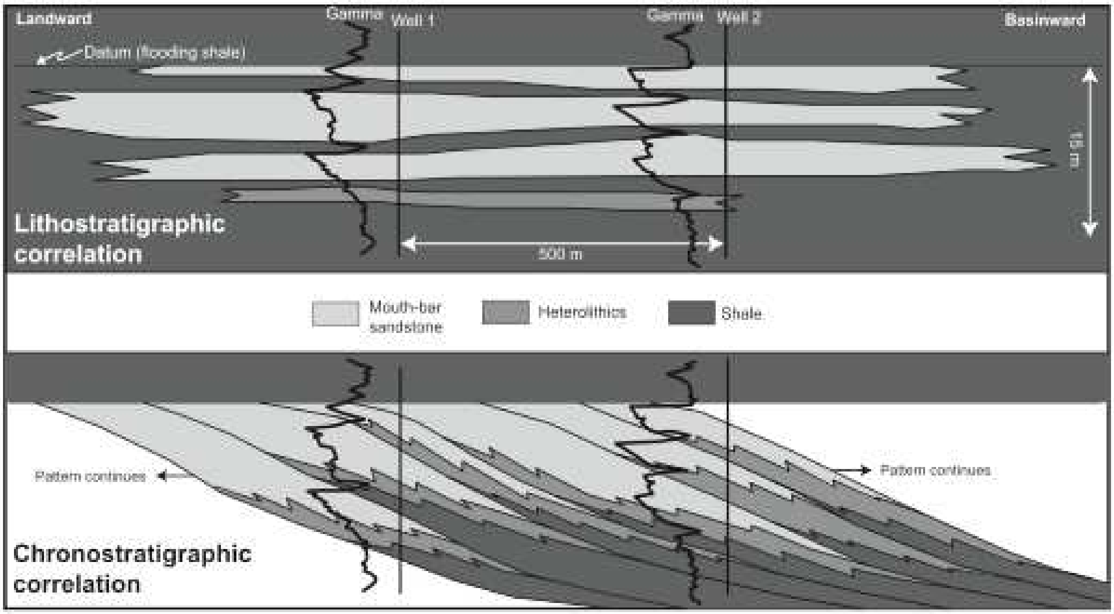


Figure 6-4 Lithostratigraphic versus chronostratigraphic correlation which shows how deposits on a slope can be interpreted very differently based on the correlation method used (from Gani and Bhattacharya, (2005), after Ainsworth and Sanlun, (1999))

6.2.2 Observations and interpretation

From the 3D model it is clear that the sandstone deposits of the Battfjellet Fm can be split into two depositional groups. The uppermost of these has been interpreted to consist of mainly shoreface deposits. When viewed from a distance these sands appear to form one continuous sandbody, but closer inspection reveals that it often consists of several (2-3) smaller, upwards coarsening sandstone tongues with thicknesses between 5 and 20 m. The various numbers of sandstone tongues in the study area are shown in Figure 6-5 below.

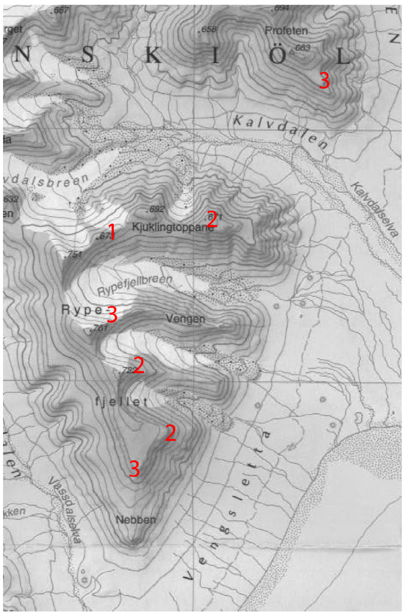


Figure 6-5 Number of shallow marine parasequences at various locations in the Rypefjellet study area (Map: Van Mijenfjorden. Norwegian Polar Institute Institute, 2008)

The shallow-marine sandstone wedges appear to form step-like cliff geometries on the mountainsides, a result of interfingering with the underlying offshore/prodelta deposits of the Frysjaodden Fm which weather more easily than the sands. The shale deposits are the product of transgressive events, flooding surfaces that occurred over short periods of time and hence can be viewed as isochronous (see Figure 6-4).

From the observations of the sandbody of the Battfjellet Fm it is clear that the number of wave-dominated, shallow-marine sandy sequences vary over short distances. These sandstone wedges have been interpreted as various parasequences and the parasequences appear to have a limited lateral extent and seem to be partially overlapping, consistent with observations made by Helland-Hansen, 2010.

The most important mechanisms that led to the superposition of sandstone wedges (clinothems) in the Battfjellet Fm were identified by Helland-Hansen (2010) as:

- Eustatic rise of sea-level
- Tectonic subsidence

- Delta-lobe switching

Where delta lobe switching was identified as the most likely mechanism at work in the Battfjellet Fm where migrating delta-lobes with limited lateral extent were deposited in the basin at different locations at different times. This type of model can explain the disparity in the number of parasequences at the various localities. This trend has also been observed at other locations in the Central Basin, and these are summarized in the figure below:

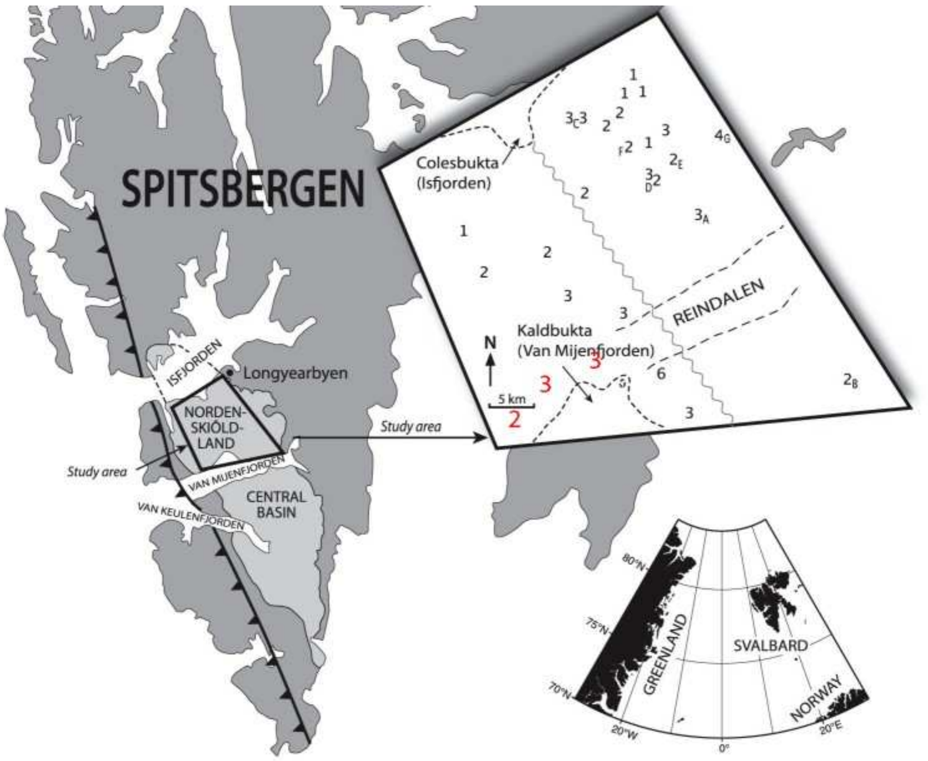


Figure 6-6 Map showing number of upwards coarsening sandstone sequences in Nordenskiöld Land. Data from the present study are marked in red (modified after Helland-Hansen, 2010).

Interfingering between the Battfjellet Fm and the Aspelintoppen Fm has not been n observed in the study area. This could be because the transition is poorly exposed in the study area or it could reflect the true nature of the boundary between the Battfjellet and Aspelintoppen formations.

The 3D model also reveals that deposits interpreted to be basin-floor fans are located in front of and stratigraphically lower than the shoreline deposits. In the study area the basin-floor fans are found at progressively higher elevations northwards. Thus, it seems likely that the basin is getting shallower in this direction. Stene (2009) observed that the distance between basin-floor fans and shoreface deposits decreased when moving away from the orogenic belt on the southern side of Van Mijenfjorden and this was interpreted as a shallowing of the basin away from the orogenic belt.

A structure-contour map of the Top Hollendardalen Fm (Dalland, 1979) shows an anticline, the Holmsenfjellet anticline, present in the northeast of study area (Fig 6-7). The Holmsenfjellet anticline has an elongate shape and Dalland (1979) interpreted the anticline to have formed as a result of compression during the development on the West Spitsbergen Orogen. This is shown by the axes of the anticline, which is inferred to be parallel to the West Spitsbergen Orogen and also constrains the development of the anticline to the main phase of folding, indicating an age not older than Late Eocene (Dalland, 1979).

The presence of the Holmsenfjellet anticline in the study area is important for understanding the depositional trends of the sediments in the Battfjellet Fm. From observations of the basin-floor fans in the study area these appear to be positioned at gradually higher stratigraphic elevations at the more basinward locations. Thus, without the knowledge of the presence of this anticline one might believe that the gradual shallowing of the basin-depth could be related to a shallowing of the sea level in the basin. The presence of some wave reworking on the uppermost sediments of the basin-floor fans does indicate a lowering of sea level, but this is most likely due to a lowering of storm-wave base during severe storms.

As the system builds out from the shoreline, depositing sands in the basin, occasional pulses of sand will be deposited on the basin-floor, but most of the time the deposition in these areas is in the form of marine shales. When the deposition of the basin-floor sands ceases the deposits will be overlain by shale (Figure 6-8). As the shoreline continues to prograde and builds out further in the basin another pulse of sand will eventually be deposited out on the slope and basin-floor. The progradational nature of the system makes the younger basin-floor fan deposit further into the basin than the older basin-floor fans while the topography of the basin-floor lets the younger fan deposit at a higher vertical level than the older basin-floor fan.

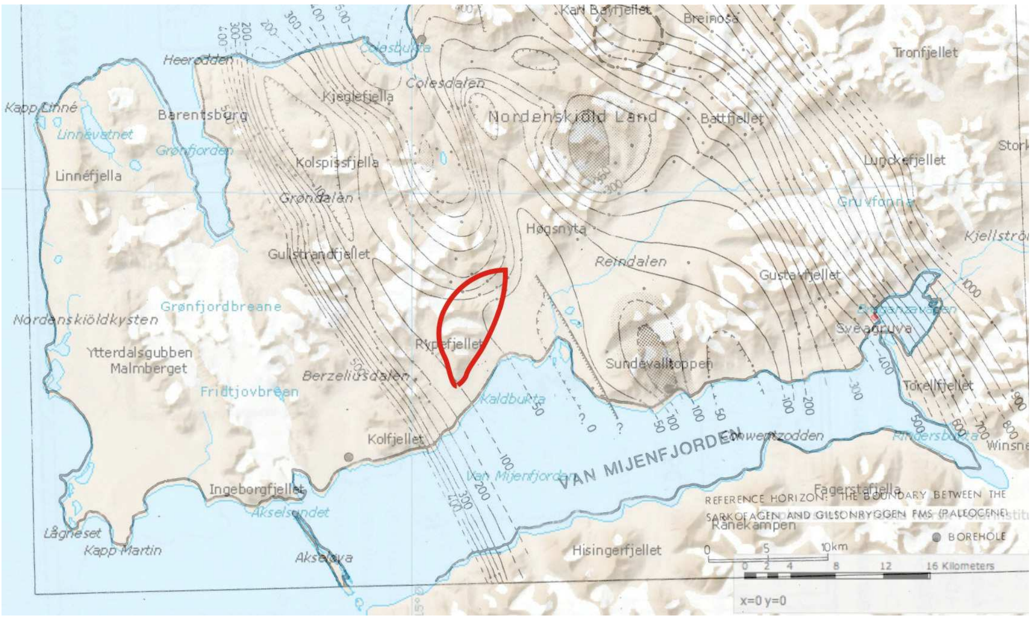


Figure 6-7 Presence of anticline on the Top Hollendardalen Fm level in the study area (Modified after Dalland, 1979 and map B10 Van Mijenfjorden, Norsk Polarinstitutt, 2008)

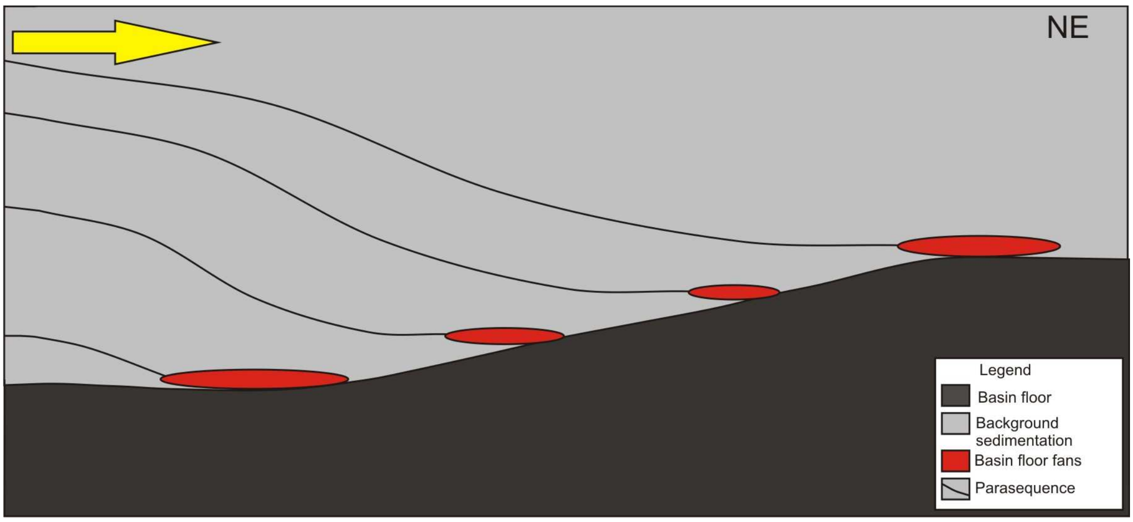


Figure 6-8 Progradation on basin-floor with anticline gives progressively higher elevation of the basin-floor fans. The arrow marks the progradation direction.

6.2.3 Sources of error

There are several possible sources of error from the correlation of the deposits in the Rypefjellet area. The elevations of the different deposits have been measured by using the barometer function of a GPS and although it was calibrated at some occasions it is sensitive to changes in air-pressure and hence the measurements will not be accurate. Still, it can be argued that these uncertainties are so small that they will not be important on a large-scale description of the sandbody geometry.

The poor visibility of the large-scale geometry of the sandbody due to the extensive scree-cover and the glacial valleys cutting down into the deposits makes it hard to connect the outcrops on different sides of the valleys. Still, sequence stratigraphic principles have been used when doing the interpretations and correlations, and as long as these are not violated the interpretation should be considered valid.

On the whole it can be argued that small mistakes in the interpretation are insignificant when describing the sandbody as long as it is valid within the sequence stratigraphic rules outlined above.

7 Paleogeography and depositional environment

7.1 Introduction

In this chapter critical aspects of the paleogeographic setting of the Battfjellet Fm are discussed based on the facies and environmental interpretations put forward in the previous chapters. The aim of this chapter is to get a better understanding of the evolution of the Central Basin and the associated environments in the early Eocene.

7.2 Depositional environment

From the Frysjaodden Fm, which underlies the Battfjellet Fm, a thickening of the deposits from 300 m in the east to 600 m in the west of the basin is observed (eg. Kellogg, 1975; Helland-Hansen, 1990, Deibert et al, 2003) and this suggests that the Central Basin was deeper in the west than in the east. Past studies of the Battfjellet Fm (e.g. Helland-Hansen, 1992; Steel et al, 2000) in Nathorst and Nordenskiöld Land identified many features that are similar throughout the formation such as facies characteristics, paleocurrent directions and the presence of clinoforms. From the interpretations and observations of the Battfjellet Fm put forward in the previous chapters it is clear that several of these characteristics are also present in the Rypefjellet study area.

7.3 Delta classification and progradational style

The Battfjellet Fm has been interpreted as an eastward prograding shoreline that gradually built into the foreland basin that formed in front of the West Spitsbergen Orogen (Kellogg, 1975; Helland-Hansen, 1992). Eastward thinning and sloping clinothems portray the depositional slope at various times during deposition, and the presence of one or more sandy, 10-30 m thick clinoforms separated by shales reflect the changing conditions in the basin over time (Kellogg, 1975; Helland-Hansen, 1990, see figure 6-6). The shales are interpreted as flooding surfaces (e.g.Helland-Hansen, 2010) giving a complex sand geometry and stacking pattern of the Battfjellet Fm. However, the clinothems are only present in the western part of the basin and were formed by thick mass-gravity deposits that in some cases also extended into basin-floor fans (Helland-Hansen, 2010). The clinoforms reflect that deposition took place on a slope that connects the shelf and basin-floor and reflects the eastward progradation of the delta along a shoreline-to-shelf bathymetric profile (Helland-Hansen, 1992). In the eastern part of the basin the mass-gravity deposits are thinner or absent, and the basin-floor deposits are dominated by shales. This has led to the interpretation that the basin was shallower further away from the orogen as there was less downwarping, and also that the distal eastern part of the basin was filled with shales simultaneously as the western areas were filled with sandy sediments.

The latitude at the time of deposition of the Battfjellet Fm was likely one where the climate was temperate with seasonal variations. Hence, changes in the depositional trends related to varying seasons were likely and would also allow for periods where storms and floods were present, while at other times of the year the conditions in the basin were quieter.

The shoreface packages observed in the Battfjellet Fm it are clearly wave-influenced and show that wave-processes were also important in the Central Basin. A wave-dominated delta will at most times produce a relatively straight shoreline with barrier bars. The presence of fluvial channel and tidal deposits in the Battfjellet Fm shows that also these processes were operating and were important for transporting sediments to the basin.

7.4 Delta size

From the observations put forward in the previous chapters, as well as from previous studies (e.g. Helland-Hansen, 2010) the lateral extent of the shallow marine sandbodies in the Battfjellet Fm have been interpreted to have a lateral extent that range from a few and up to 10 km.

A modern delta with a similar size-range is the William River delta in Canada. The delta is 9x8 km (Smith et al., 2005) and has been studied by looking at sediment samples and radar stratigraphy. The William River delta is wave-dominated and storm-wave action is important for shaping the delta (Smith et al., 2005). The delta appears to have a lobate shape and a thickness between 15 and 22 m (Smith et al., 2005), which corresponds to the inferred size of the delta lobes in the Battfjellet delta.

7.5 Paleogeography

Paleocurrent data are summarized in Figure 7-1 and these measurements have been used to reconstruct the paleogeography of the Battfjellet Fm. The structures that were measured were the dip-azimuth of current ripples and cross-stratification, orientation of wave-ripple crests and the direction of solemarks. The numbers of measurements are limited due to poor quality outcrops in many locations. Still, the measurements included in this study are expected to be of good quality and hence are believed to provide an accurate representation of the paleo-basin.

Unfortunately, the presence of the Aspelintoppen Fm is limited to the highest peaks in the study area and only one fluvial channel within the Aspelintoppen Fm has been observed in this study. This channel does not provide sufficient data to give satisfactory observations of the direction of fluvial channels but previous studies (e.g. Dreyer and Helland-Hansen, 1986; Helland-Hansen, 2010) recorded an easterly direction of fluvial channels, suggesting a shoreline oriented approximately north-south.

From the orientation of the wave-ripple crests, which are inferred to be parallel to the shoreline, it is apparent that the trend is approximately north-south. This is in accordance with the inferred orientation of the orogen and also similar to results obtained in previous studies at other locations in the basin (e.g. Helland-Hansen, 1985; Helland-Hansen, 1990; Crabaugh and Steel, 2004; Stene, 2008).

The solemark measurements at the bases of turbidites seem to parallel the orogen, but the number of measurements is very low (3) and hence not much weight should be put on these measurements. However, Crabaugh and Steel (2004) reported that basin-floor fans travelled in an along-strike direction in the later stages of basin-floor deposition on Storvola in Van Keulenfjorden which was explained by the topography of the basin-floor which was generated tectonically. The dip azimuths of the current ripples show a general trend towards the east which is consistent with an eastward migrating shoreline. The cross-stratification, which has been interpreted as the migration of sand-dunes, shows highly variable directions. This is in accordance with earlier observations by Olsen (2008), Helland-Hansen (2010) which shows that the directions of cross-stratification unfortunately does not give much information about the paleocurrent directions although it might reflect shifting currents in the depositional environment. The paleocurrent measurements carried out in this study can, based on the data from wave-ripple crests and dip-azimuths of current ripples, be confirmed as consistent with a north-south oriented coast and an eastward migrating shoreline.

The small number of river distributaries observed along the shoreline as well as the interpretation put forward previously that the delta built out into the basin in the form of lobes that later were abandoned suggests that areas of the shoreline at times received little sediment. Thus, the shoreline should be expected to be lobate and show variations in the directions of the wave-ripple crests. Still, both this and previous studies (e.g. Helland-Hansen, 1990) show very consistent that ripple crests are located in a north-south direction. Thus, it has been suggested that either the deltas were fronted by coalescing spits in a north-south direction, or that the shallowing of water in front of the deltas was so rapid that the waves did not have enough time to readjust their approach angle so that it was directed parallel to the shoreline. A possible shoreline geometry of the Battfjellet Fm at different times during the infilling of the Central Basin are shown below in Figure 7-2.

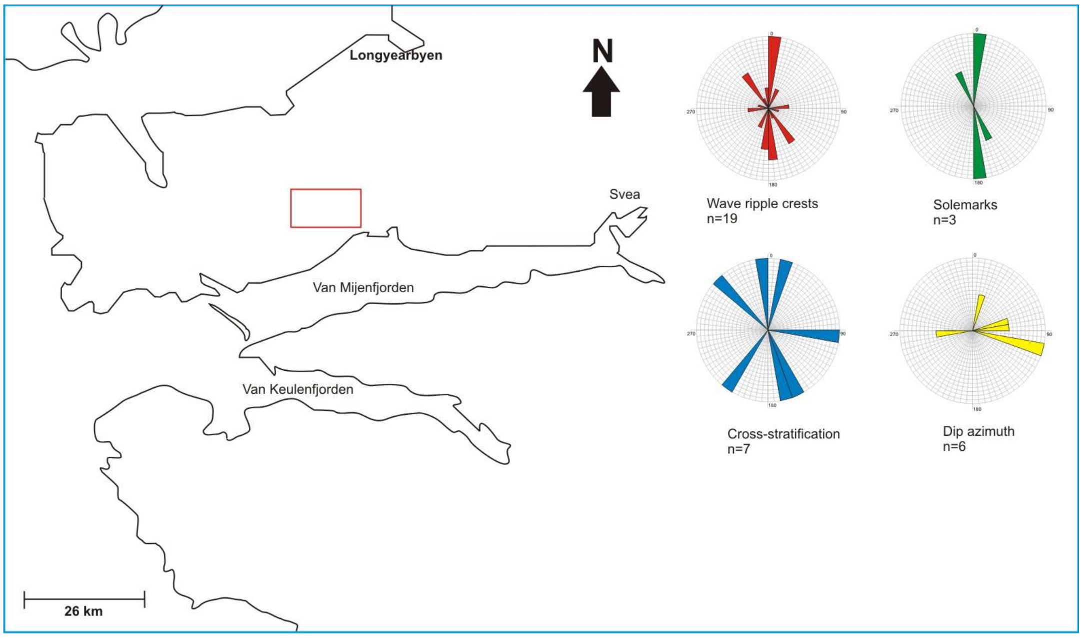


Figure 7-1 Paleocurrent data from measurements in the field are shown in the rose diagrams and the location of the study area is highlighted. The wave-ripple crests which are interpreted to be parallel to the shoreline show an N-S directed shoreline. The measurements of solemark directions of the bottom of turbidite currents are few and hence should not be put much weight on, but these show a travel direction that appears parallel to the direction of the paleoshoreline. The cross-stratification from dunes in the upper shoreface does not show any trends. The dip azimuth of current ripples appears to favour an eastward direction, consistent with the anticipated shoreline migration direction.

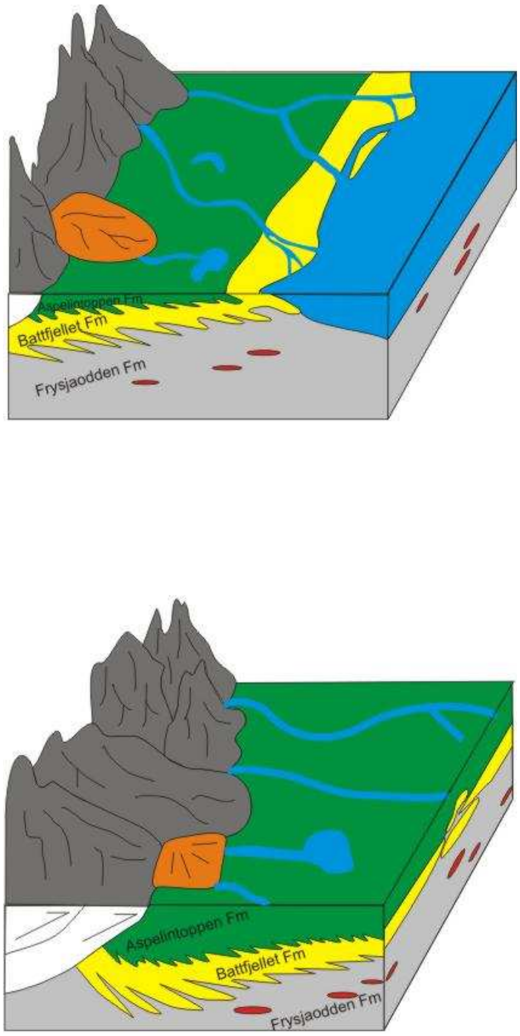


Figure 7-2 the shoreline of the Battfjellet Fm at different times during deposition. It should be noted that these figures do not show the shoreline at any particular time but rather the aim is to display what the depositional environment might have looked like at different times during the depositional process.

7.5.1 Basin formation

Thickness data from the Battfjellet, Frysjaodden and Hollendardalen formations have been used to create a thickness map shown in Figure 7-3 below. The top Battfjellet Fm was used as the uppermost measured level, as this is the uppermost formation boundary exposed in the Central Basin. Where the thickness of the Hollendardalen Fm is known this has also been included because the Hollendardalen Fm is closely related to the overlying formations and the was also deposited from sediment derived from the orogen to the west of the basin.

As the base of the Frysjaodden Fm not is exposed in the Rypefjellet study area the structure-contour map of the top Hollendardalen/base Frysjaodden (Gilsonryggen Mb) has been used as a reference horizon when determining the thicknesses. The elevation measurements were made by using the altimeter function on a GPS. The GPS was calibrated some times during the period in the field but requires a clear view to several satellites so measurements have some uncertainties associated with them.

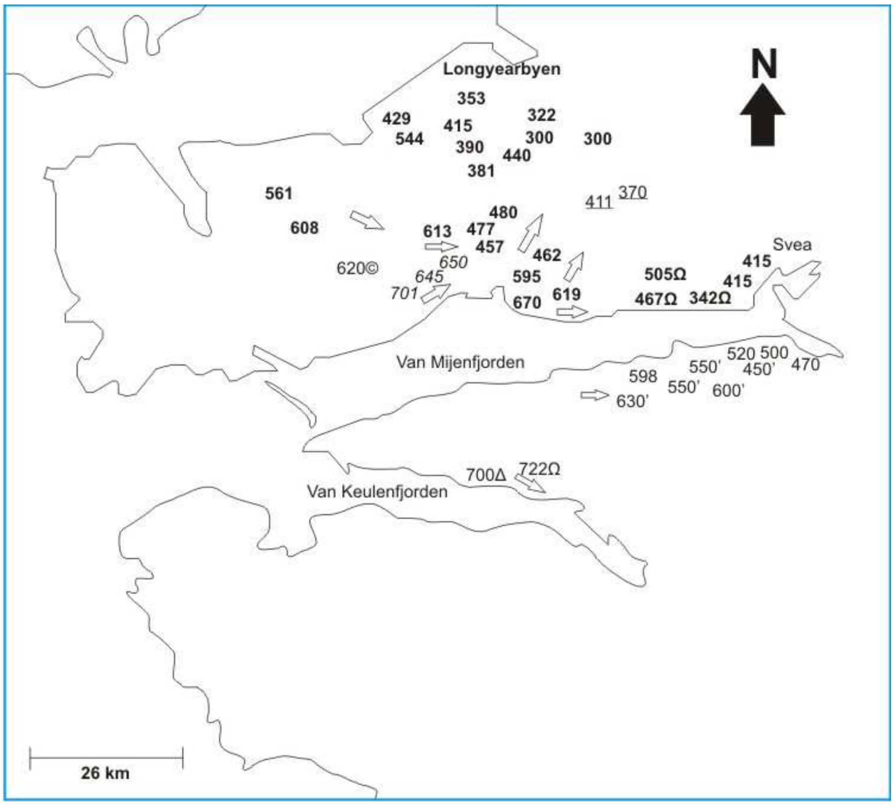


Figure 7-3 Thickness data of from the Hollendardalen, Frysjaodden and Battfjellet formations. (© is data from Croxton and Pickton, 1976, 370 from Steel et al., 1981, **595** from Helland-Hansen, 1985, Δ from wells by Store Norske Spitsbergen Kullkompani, Ω from Steel and Olsen, 2002, 450 from Olsen, 2008 and *645* from the present study). The arrows mark the apparent dip directions of clinoforms where they are present in the Central Basin (modified after Olsen, 2008).

The thickness measurements from the Rypefjellet area appears to fit with the interpretation that the basin gets deeper and hence the deposits thicker closer to the orogen. The thickness data have also been incorporated into an isopach map (Figure 7-4) which shows the thickness distribution of the Hollendardalen, Frysjaodden and Battfjellet formations. From this map the contours have an oblique angle to the West Spitsbergen Orogenic Belt.

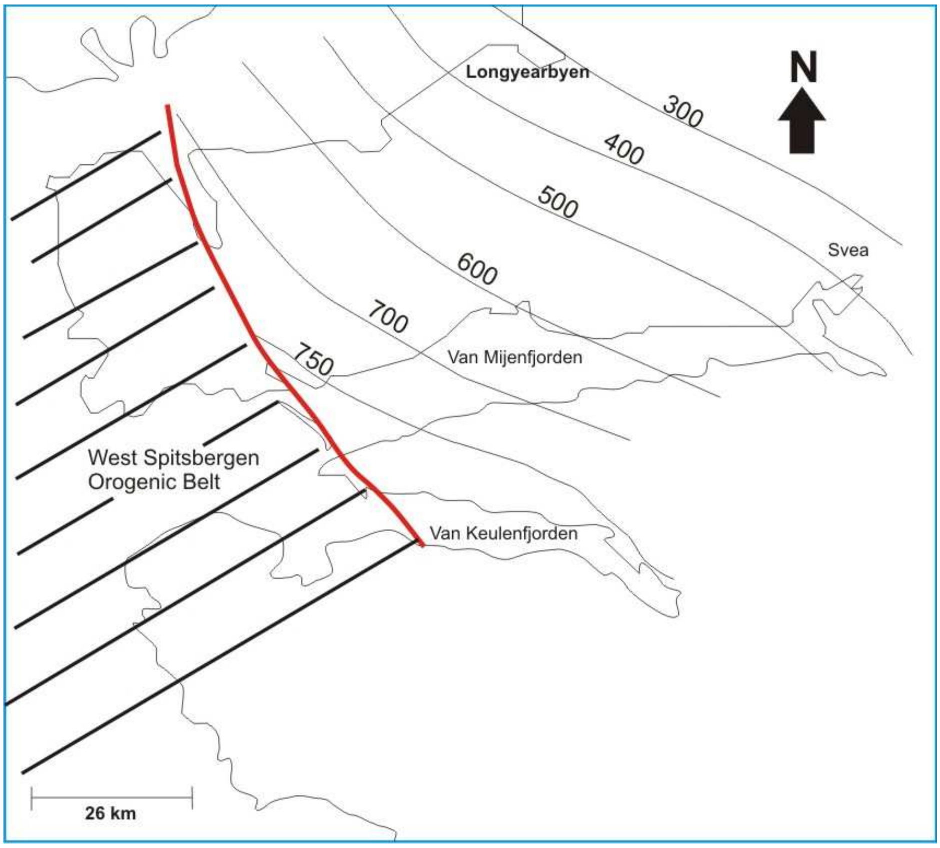


Figure 7-4 Isopach map of the Hollendardalen, Frysjaodden and Battfjellet formations in the Central Basin (modified from Olsen, 2008 and Stene, 2009).

After the deposition of the Grumantbyen Fm the water in the Central Basin was inferred to be relatively shallow (Dalland, 1979). Hence, to make room for the several hundred metres of sediments belonging to the Hollendardalen, Frysjaodden and Battfjellet formations new accommodation space would have to be generated in the Central Basin. An eustatic sea-level rise of approximately 100 m and compaction of sediments of the Firkanten and Basilika formations would not have been large enough to accommodate all the overlying sediments (Helland-Hansen, 1990). Thus, another model is needed to explain the generation of accommodation space in the Central Basin as well as the great variations in the sediment thickness between the eastern and western part of the basin. During the time of deposition the area was affected by a transpressional tectonic regime and hence thinning of the crust at this time can be eliminated (Helland-Hansen, 1990). Lithospheric loading of the crust as a result of tectonic thickening of the fold-and-thrust belt was put forward as a mechanism to explain the generation of accommodation space by Helland-Hansen (1990) and indeed, this model can explain both the generation of accommodation space in the basin and thickness variations of sediments. At the time the Battfjellet Fm was the Central Basin acted as a foreland basin to the fold-and-thrust belt. A foreland basin is an elongate trough that forms between a linear contractional belt and a stable craton (e.g. DeCelles and Giles, 1996) and such basins are generated as a result of thrust-sheet loading in the orogenic belt. Figure 7-5 displays a model of the development of the Central Basin, initiated by thrusting and uplift in the west.

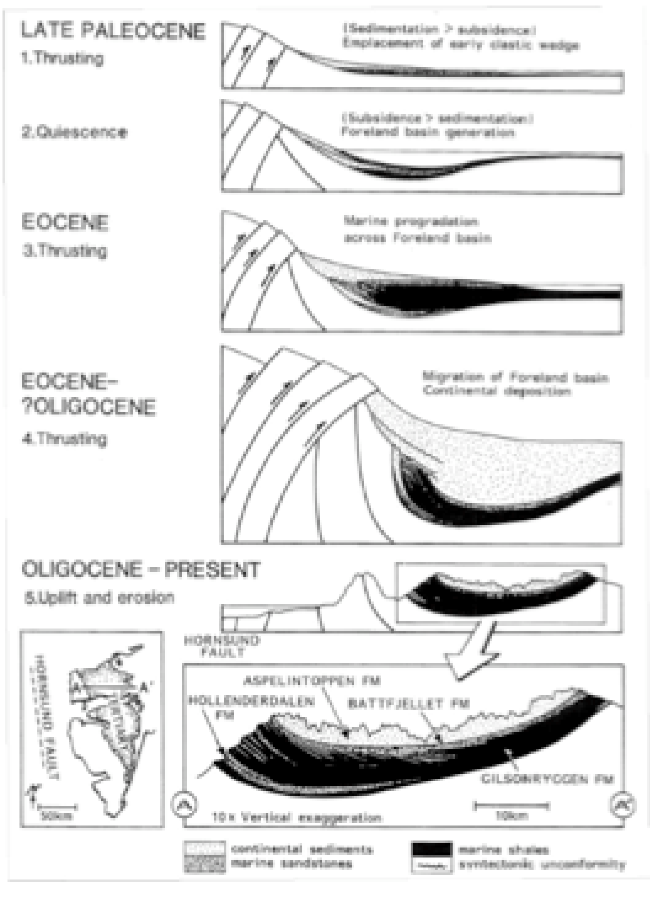


Figure 7-5 Development of the Central Basin with uplift and thrusting to the west as interpreted by Steel et al (1985) where lithospheric loading due to thickening of the crust in the orogenic belt was the main mechanism for generating accommodation space,

8 Discussion

8.1 Introduction

The previous chapters have outlined the sedimentology and depositional environment of the Battfjellet Fm in the Rypefjellet area with interpretations of facies, facies associations, sandbody geometry, depositional environment and paleogeography. This chapter will look at the problems, challenges and uncertainties associated with the presented models and interpretations.

8.2 Delta type

From the interpreted shallow-marine deposits in the study area it is clear that wave processes played an important role in shaping the shoreline in the Central Basin during the deposition of the Battfjellet Fm. The sedimentary input has been interpreted to be derived from small rivers that drained the land-area to the west of the basin, and as the majority of the sediments were river-derived, this has led to the interpretation of a deltaic shoreline (Bhattacharya and Giosan, 2003). From the study area it is clear that the number of shallow marine parasequences varies over short lateral distances which is consistent with other observations from Nordenskiöld Land (Helland-Hansen, 2010) and has been interpreted to be a result of switching delta lobes. However, this interpretation leads to a conflict between the internal and external properties of the Battfjellet Fm as switching of delta lobes usually are associated with fluvially dominated deltas. Thus, a fluvio-wave interaction delta has been proposed for the Battfjellet Fm.

Evidence of rapid sedimentation in the Rypefjellet area has been observed from frequent soft-sediment deformation and few intensely bioturbated strata. A high sedimentation rate could suggest that sediments only were transported a short distance from the source area before they were deposited which led to poor sediment sorting (Helland-Hansen, 2010) and could also explain why the fluvially influenced stacking pattern with several delta-lobes has been preserved, as new sediments were deposited before the basinal processes such as longshore currents were able to remove the fluvial characteristics of the sediments. The presence of basinwards advancing parasequences also demonstrate the progradational nature of the Battfjellet Fm delta.

A possible modern analogue for this succession has been described by Marchesini et al (2000) about 20 km south of the Po delta in the Adriatic Sea where the basin is wave-influenced but with immature sediments. Moreover, the plan-view of the Po-delta is typical of a wave-river interaction delta.

An asymmetric delta lobe would also allow for the preservation of fluvial and tidal deposits located behind spits or barriers that develop on the downdrift end of the delta, which has been observed in several modern deltas (Bhattacharya and Giosan, 2003) as well as in an ancient example from the Aberdeen Mb in the Balckhawk Fm of the Book Cliffs, USA (Charvin et al., 2010). Fluvial and wave processes are active in asymmetric deltas (Charvin et al., 2010).The requirements for asymmetric deltas, as pointed out by Bhattacharya and Giosan (2003) are strong longshore currents and high sediment input, and these deposits are often found in microtidal areas. No evidence for asymmetric delta lobes have been observed in the present study but with evidence of both wave and fluvial processes as well as high sediment input and some evidence of tides the presence of asymmetric delta lobes in the Battfjellet Fm could be likely.

8.3 Initiation and nature of turbidite currents

A large proportion of the sediments belonging to the Battfjellet Fm in the Rypefjellet area have been interpreted as deposits from turbidite currents (Bouma, 1962). A turbidite current is a mass-gravity driven underwater current. The generation of a turbidite current requires much suspended sediment, and the mechanisms that can disperse such currents have been identified by Piper and Normark (2009) as:

- Discharge from rivers or subglacial meltwater with high sediment concentration
- Slides and debris flows transformed by liquefaction and entrainment of water (earthquakes, failure)
- Oceanographic processes such as storms, tides and internal waves that lead to suspension of sediments from the coast, shelf or upper slope

Fresh, sediment-laden currents that enter a body of standing water will, when the density of the flow is a certain order higher than the water in the basin, be able to produce a hyperpycnal flow, a type of turbidity current (Mulder and Syvitski, 1995). The density difference between the fluvial and basinal waters it can lead to sediment bypass on the shoreline or in a mouth bar which will lead to sediment deposition in the offshore setting from underflows (Bhattacharya, 2006). A hyperpycnal flow is generally develops more easily in lacustrine, fresh-water basins than in marine, salt-water basins (Mulder and Syvitski, 1995), but in the same study it was also identified that hyperpycnal flows can be triggered in small and medium sized rivers during floods, which can occur several times during the year. Several small channels that eroded down into the shoreface were observed in the Rypefjellet study area and thus it is likely that many small rivers supplied sediment to the basin. The importance of floods for the generation of hyperpycnal flows were discussed by Mutti et al. (2003) where the deposition of such currents were associated with abundant hummocky cross-stratification structures. From the present study there are observations of turbidite deposits that are closely associated with hummocky cross-stratified strata which could suggest that floods played an important role in the deposition of these sediments.

From the paleogeographic interpretation of the Battfjellet Fm, the deposition in a relatively small foreland basin suggest that the water in the basin could have been brackish which would allow for a smaller density needed to generate hyperpycnal flows. (Kellogg, 1975) summed up the fauna of the Paleogene deposits in the Central Basin where there also appears to be some evidence of brackish conditions in the Battfjellet Fm deposits. Plink-Björklund and Steel (2004) identified various factors that are used to interpret hyperpycnal flow turbidites in ancient deposits, and several of these factors have been observed in the Rypefjellet area, including fluvial channels incised on the shoreface which probably were connected to the slope channels observed at deeper locations in the basin giving a connection between fluvial and turbidite deposits, presence of abundant organic material within the basin-floor fans, an abundance of thick turbidite beds and the sand-prone character of turbidite beds.

Hyperpycnal flow turbidites have been observed in the Battfjellet Fm at several locations in the Central Basin (Steel et al., 2000; Plink-Björklund et al., 2001; Mellere et al., 2002; Plink-Björklund and Steel, 2004; Petter and Steel, 2006), whereby Plink-Björklund and Steel (2004)concluded that direct river-effluents were the most important processes for generation of turbidites in the Central Basin.

Another possible mechanism for generating turbidite currents is failure on the slope including slumps and debris flows. Processes of this kind would probably require a triggering mechanism such as a tectonic movement. Steel et al (1981) suggested that frequent earthquakes were responsible for the development of soft-sediment deformation in the Battfjellet and Aspelintoppen formations, and the setting of the Central Basin as a foreland basin where the sediment source has been interpreted as a newly uplifted orogen shows that the basin was tectonically active. Hence, it can be assumed that earthquake shocks did occur. However, such catastrophic events were probably not occurring frequently enough to explain the large volume of sands deposited by turbidite currents foreland basins such as the Central Basin (Mutti, 1999; Mutti et al., 2003).

Tectonic events like these can be regarded as catastrophic and thus would probably not occur very often. The general abundance and thickness of the turbidite derived deposits within the study area suggest that deposition was relatively continuous or took place at regular intervals over an extended period of time which would require many catastrophic events.

Similarly, wave-action may also have triggered turbidite currents, particularly during storms. Prior et al. (1989) observed that storms triggered submarine landslides in the Huanghe delta as the waves caused slope failure and collapse. It was also observed that the areas which previously had experienced slope failure were more likely to experience it again relative to areas which were undisturbed. Hence, a continuous supply of sediment from many failures in the same place could possibly explain the continued supply of sands. The many storm-generated deposits on the shoreface shows that storms were important during the depositional process. However, it seems unlikely that basin-floor fans should be generated in such large quantities and volumes if triggered by storms.

The hyperpycnal flow can be favoured as the depositional mechanism from sustained flow-type turbidites (e.g. Kneller, 1995) which occur for longer than only a few hours as direct input from a river is more likely to be the source of such deposits than catastrophic events or actions by waves or tides. The great thickness of some of the turbidite deposits in the Rypefjellet area seem to require a sustained flow, and hence hyperpycnal flows have been interpreted to be the main mechanism responsible for generating turbidite currents in this area. However, this does not mean that other mechanisms did not trigger turbidite currents, and several processes may have been operating simultaneously which has been observed in the Sepik River mouth in Papua New Guinea where there is evidence of turbidite currents that are generated by hyperpycnal flows and from sediments that were stored temporarily on the shelf and slope before they turbidite current was triggered (Kineke et al., 2000).

8.4 Structural movements and effects on deposition

The sediments deposited in the Central Basin are all inferred to be of Paleogene age but from studies on the geological history of Svalbard (chapter 3) it is clear that tectonic movements took place up into Devonian/Carboniferous time, most of these related to the Caledonian orogeny. In Middle Carboniferous time there was extensive faulting in Spitsbergen, and Dalland (1979) suggested that this faulting could have produced half-grabens or grabens in the area beneath the Central Basin. Thus, these structures may have been important for controlling the sedimentation pattern in the Central Basin but will probably not be important for constratning the deposits of the Battfjellet Fm as it is one of the uppermost formations in the Central Basin and the topographic relief in marine basins will be reduced as the basin is filled (e.g. Sutcliffe and Pickering, 2009).

The presence of clinoforms only in the western part of the basin has been described earlier (e.g. Helland-Hansen, 1985) and interpreted to be a result of shallowing of the basin away from the north-south oriented orogen. Being a foreland basin it can be expected that the basin is deeper close to the orogen or even partially beneath it (Jordan, 1995) because lithospheric loading is greatest in this area.

From the sandbody geometry in the study area it is clear that the basin-floor fans were deposited at gradually shallower depths which in this study has been interpreted to be a result of basin-floor topography with the presence an anticline (Figure 6-8). Thus, it is clear that the anticline developed either before or during deposition of the basin-floor fans. Dalland (1979) interpreted the anticline to have formed during the main stage of folding in the West Spitsbergen Orogen as the axes of the anticline is parallel to the orogen.

Through lab-experiments, Alexander and Morris (1994) demonstrated that structures on the basin-floor affect deposition, and when a turbidity current meets an obstacle the velocity of the flow will decrease (Kneller, 1995). This might lead to rapid deposition as the slower moving flow is unable to keep the same amount of sediments in suspension. A structure can also lead to deflection of the flow so that it moves along a strike direction in the basin, as has been observed by Crabaugh and Steel (2004) in the Van Keulenfjorden area. The directional data from turbidite flows in the Rypefjellet area is sparce and hence it cannot be determined if longitudinal currents were moving on the basin-floor in this area.

The exact timing of the different events that led development of the fold-and-thrust belt in Spitsbergen and deposition in the Central Basin is not known, although they developed at the same time. The Battfjellet Fm is one of the uppermost formations present in the Central Basin it can be speculated that the tectonic movements associated with the development of the fold-and-thrust belt had were ending as the the Battfjellet Fm was depostied, but evidence of extensive erosion (Manum and Throndsen, 1986) in the order of 1700 m in the Aspelintoppen Fm suggest that deposition was still going on for a long time after the Battfjellet Fm had been deposited. The thickness of the eroded deposits also suggest that accommodation space was generated in the Central Basin

8.5 Future work

The study of the Rypefjellet area has added to the knowledge of the Battfjellet Fm in Central Basin, Spitsbergen as detailed facies analysis for this area has been established along with observations of variable number of parasequences over short lateral distances which is consistent with the observations of the Battfjellet Fm in other areas of the Central Basin. The presence of an anticline within the study area has also been used to explain the “climbing” geometry of the basin-floor fans in the study area.

When the details of the geology of the Battfjellet Fm has been documented over the whole area, a synthesis combining the data from the whole Central Basin would be an obvious continuation of this study. The model of the Battfjellet Fm made in Google Sketchup is very simple model; to take this a step further a better model, for instance with the use of a proper digital elevation model. It would also be possible to extend the model to include a larger area, for instance the whole Central Basin where the resulting model probably will provide a better understanding of the shoreline trajectory of the entire formation, and would also give a good visualization of the basin fill, depositional system and spatial and temporal development of the basin.

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9 Conclusions

- The Battfjellet Fm in the Rypefjellet area portrays an eastward prograding shoreline in the Central Basin foreland basin.

- The majority of the sedimentary structures show modification by waves although evidence of both tidally and fluvially derived and modified sediments have been observed.

- Basin-floor fans make up a significant volume of the sediments in the study area and have been interpreted as turbiditic deposits that most likely were the result of suspension-fallout from hyperpycnal currents although floods, waves and earthquakes could have been important generating mechanisms.

- Basin-floor topography appears to have been important for the development and depositional trend of the basin-floor fan deposits

- The internal and external properties of the delta deposits are conflicting as the former suggest a wave-dominated shoreline while the latter have properties consistent with a fluvially dominated delta. Hence the Battfjellet Fm has been interpreted as a fluvio-wave delta.

- High organic content and immature sediments suggest a short distance to the source area.

- Soft-sediment deformation and few highly bioturbated deposits could indicate a high sedimentation rate.

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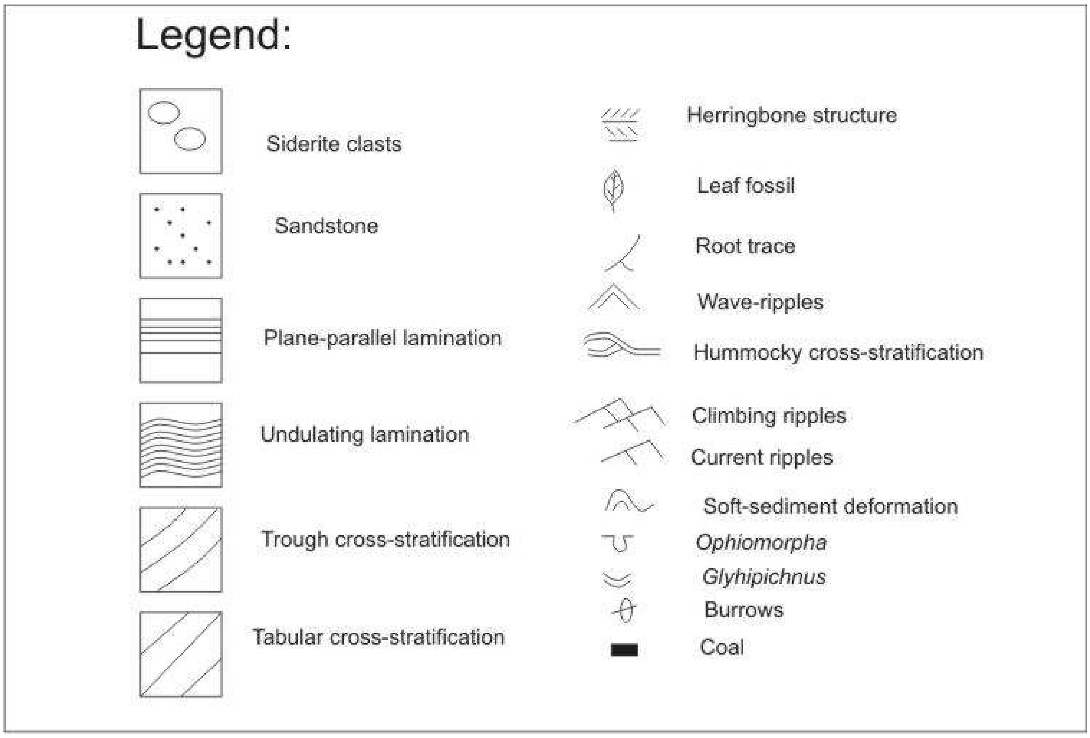
11 Appendix



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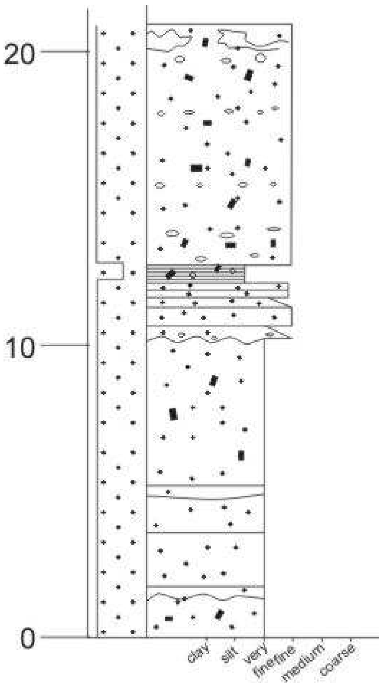

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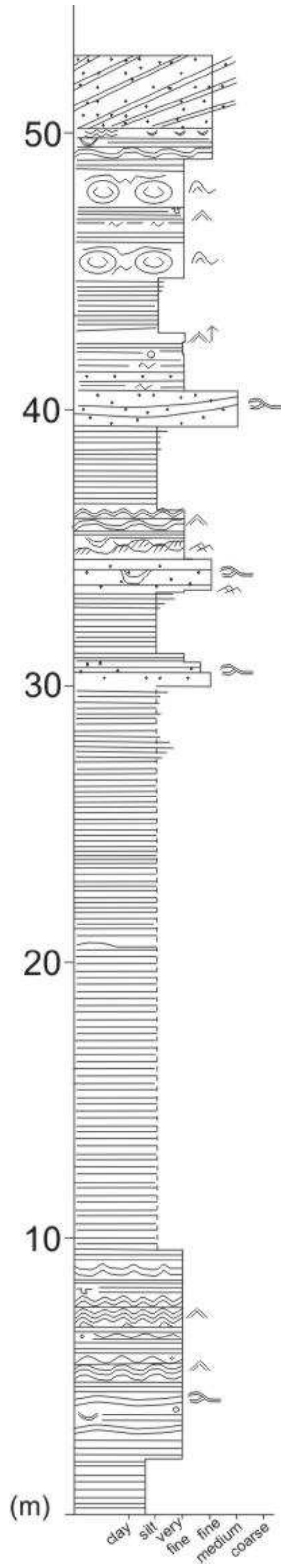

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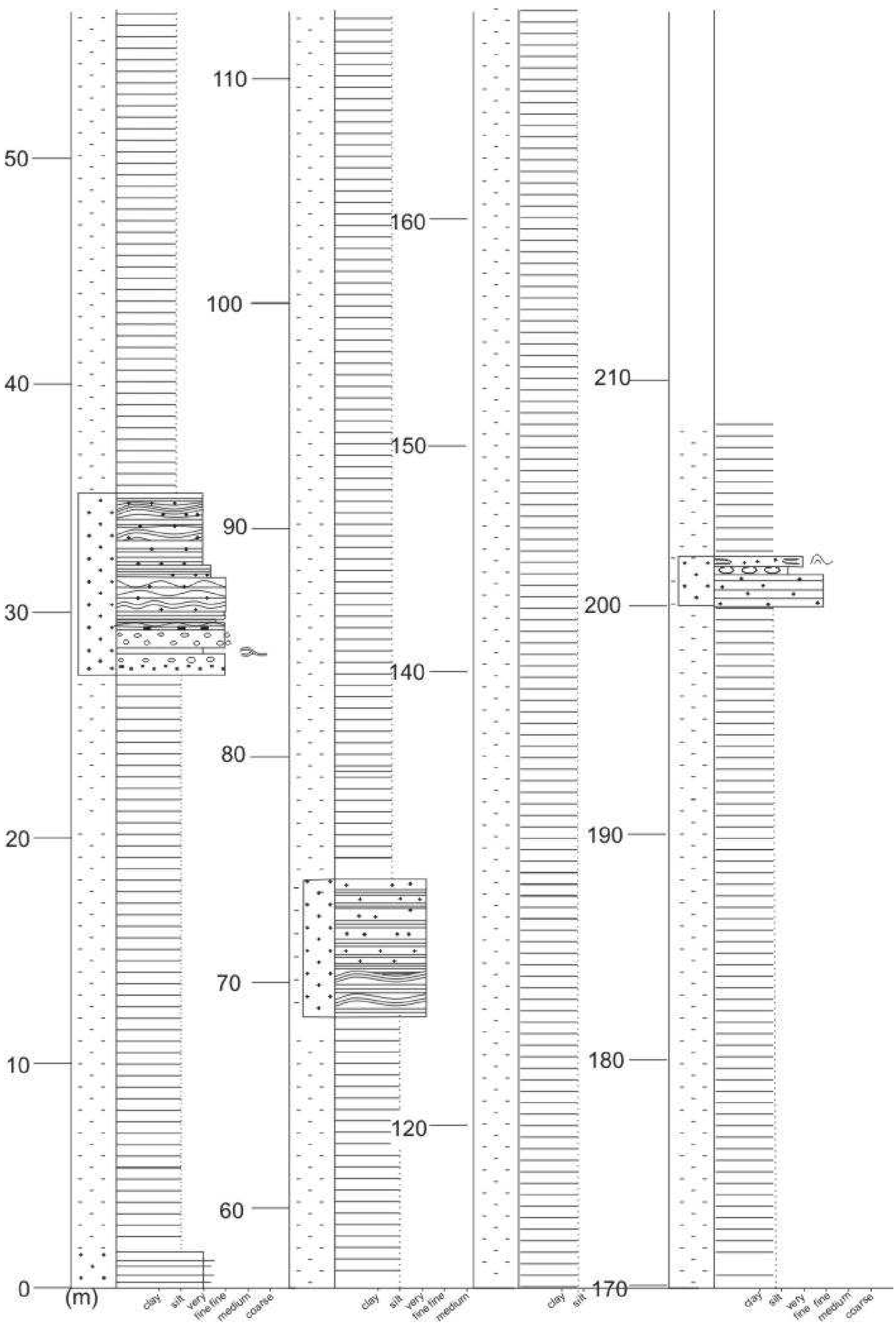

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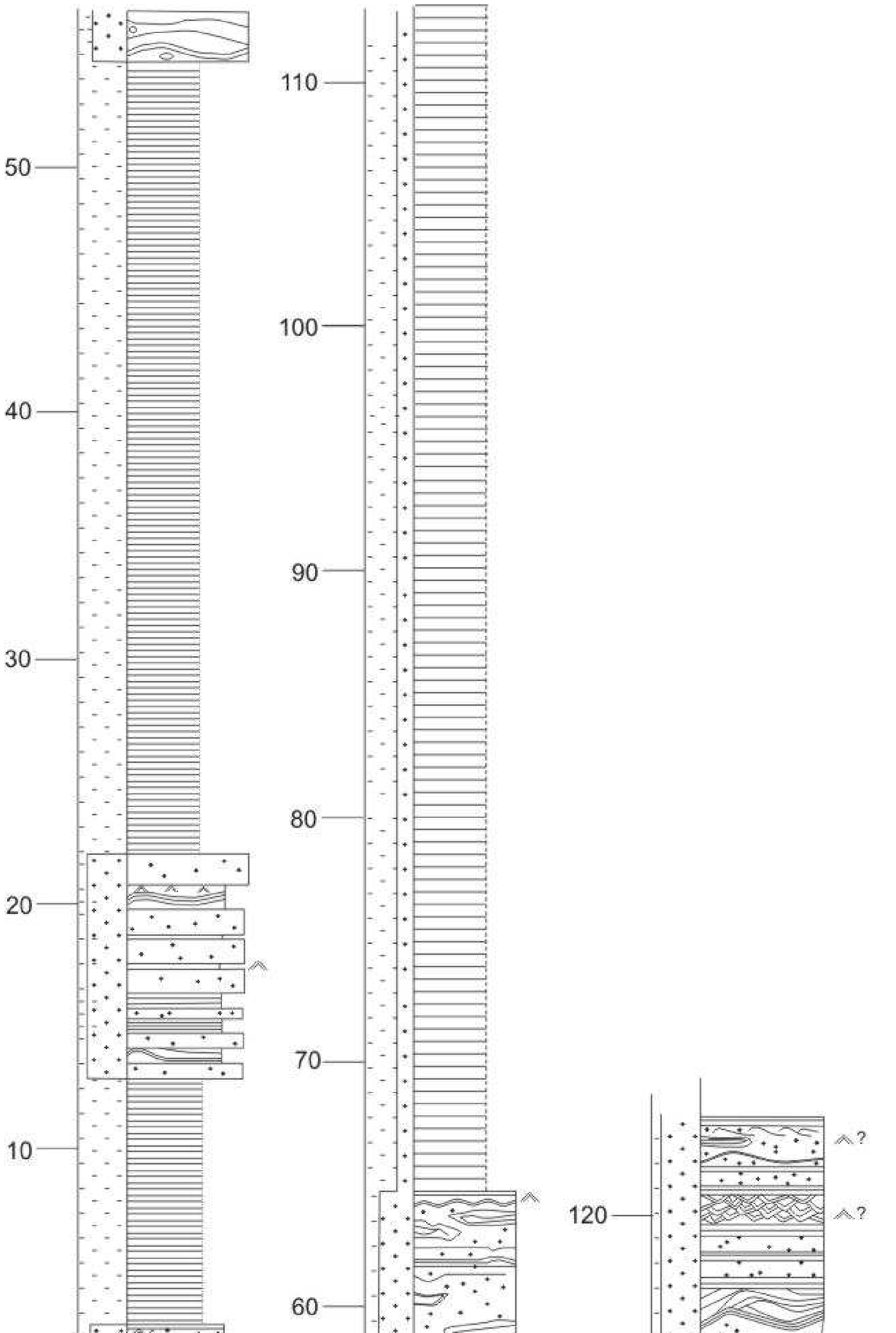

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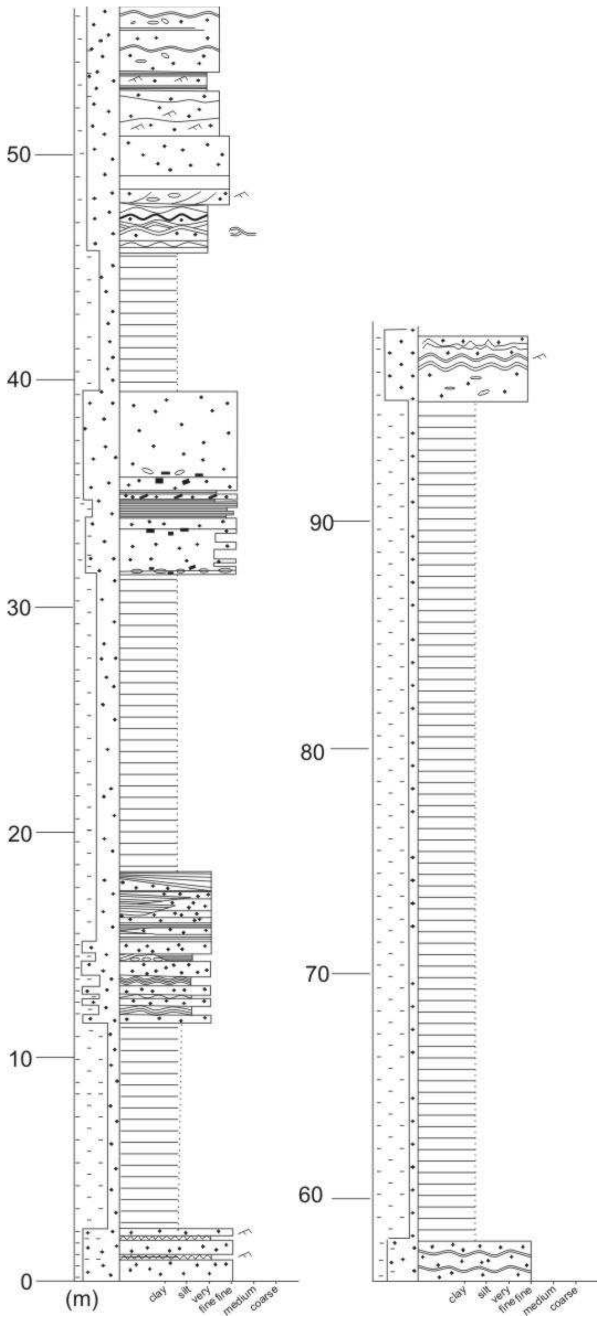

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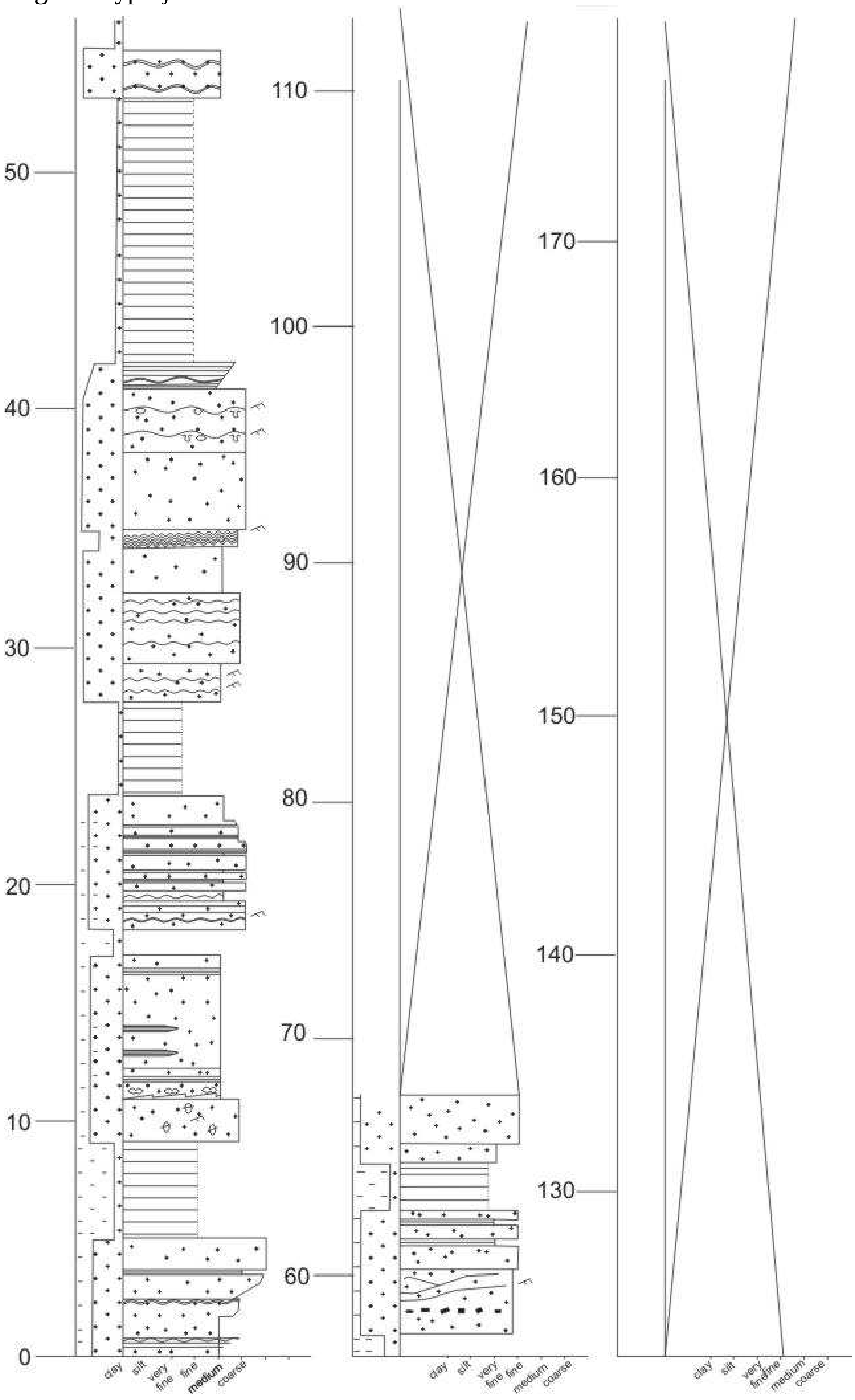

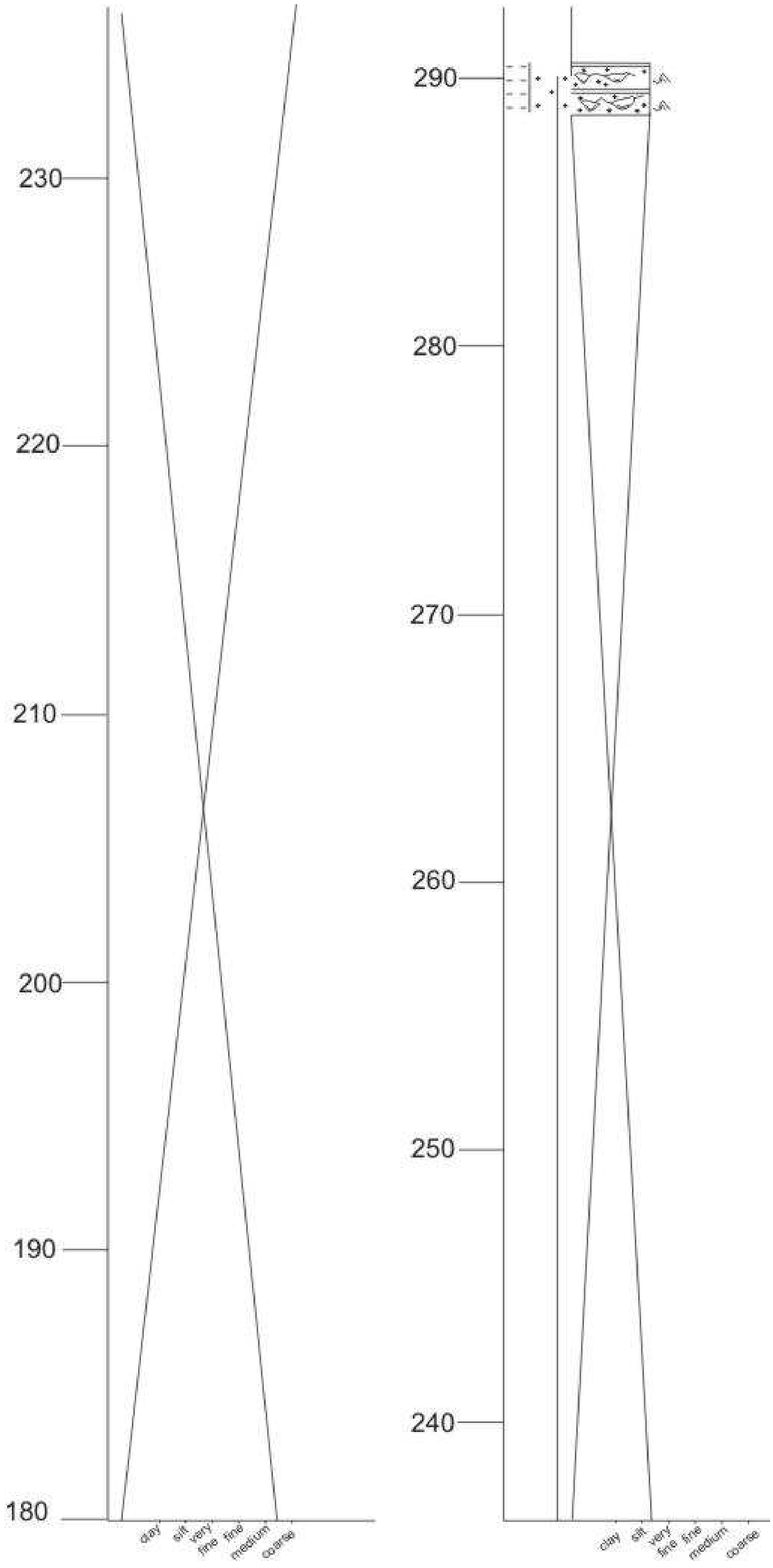

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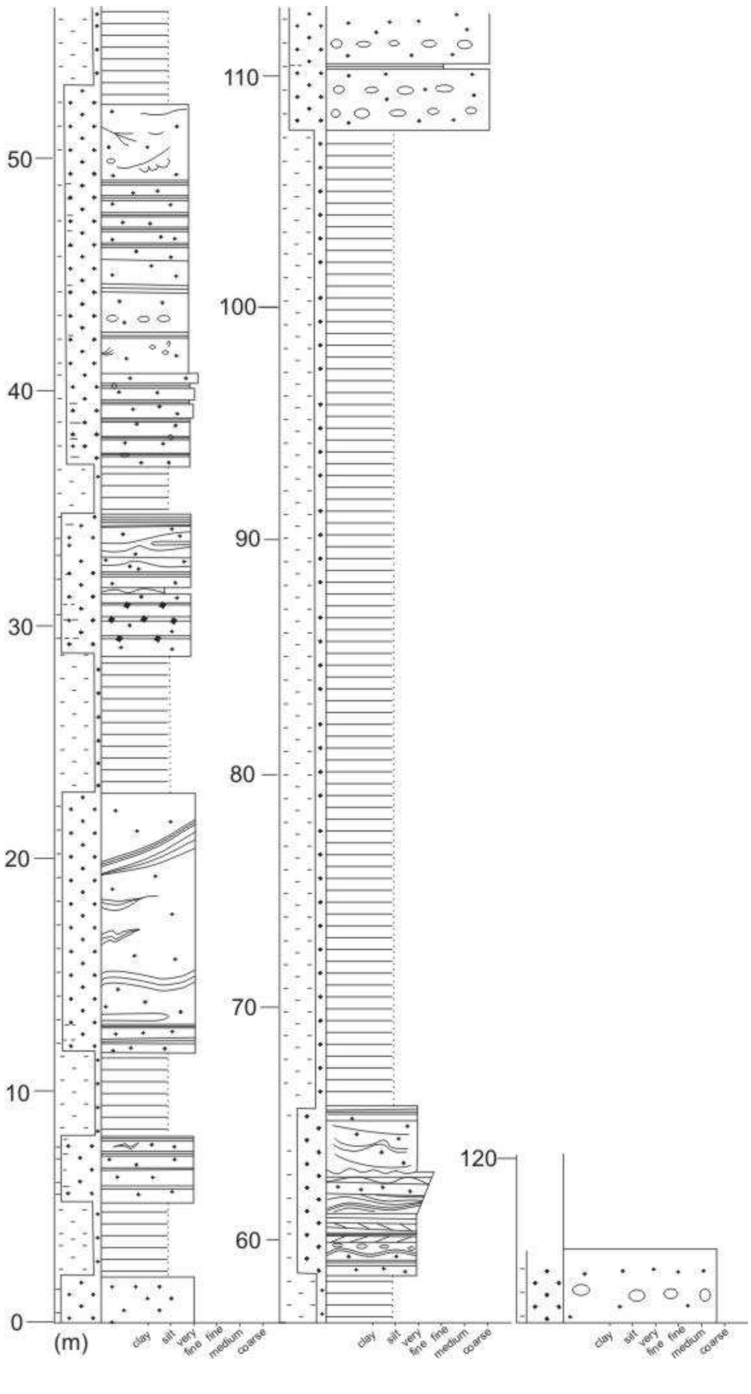

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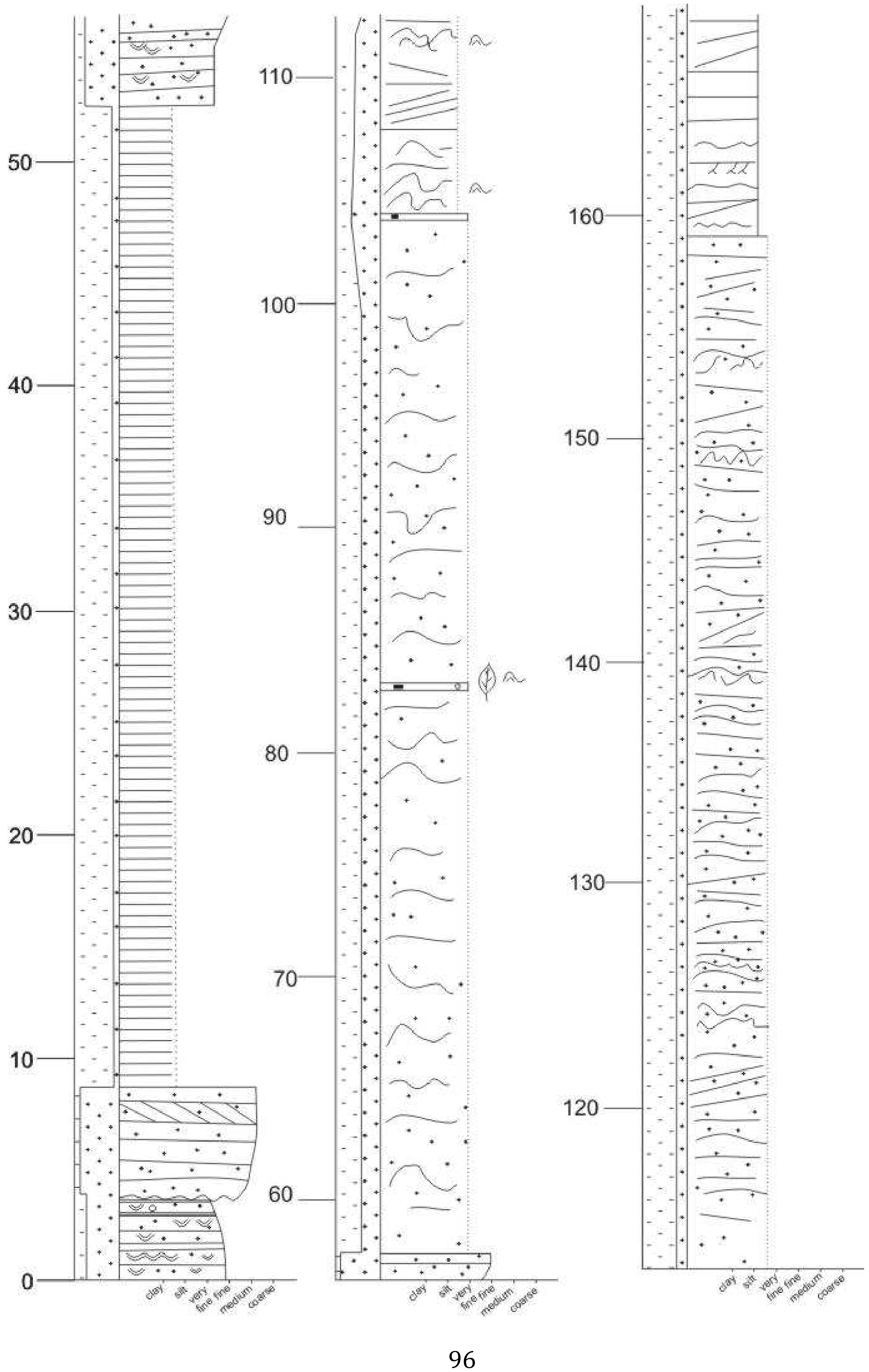

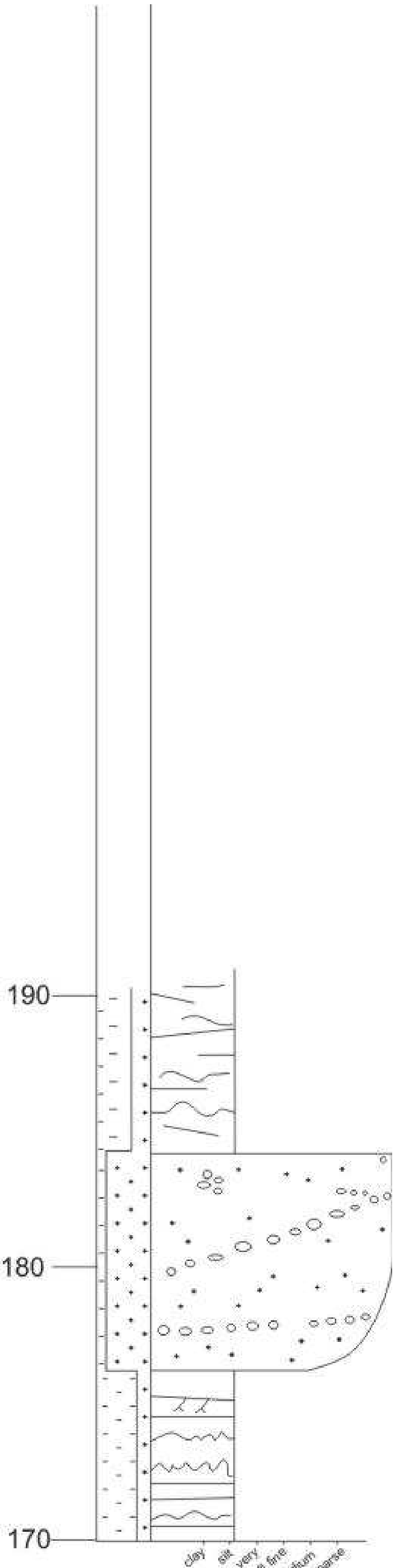
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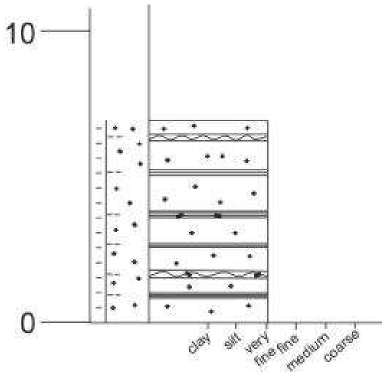

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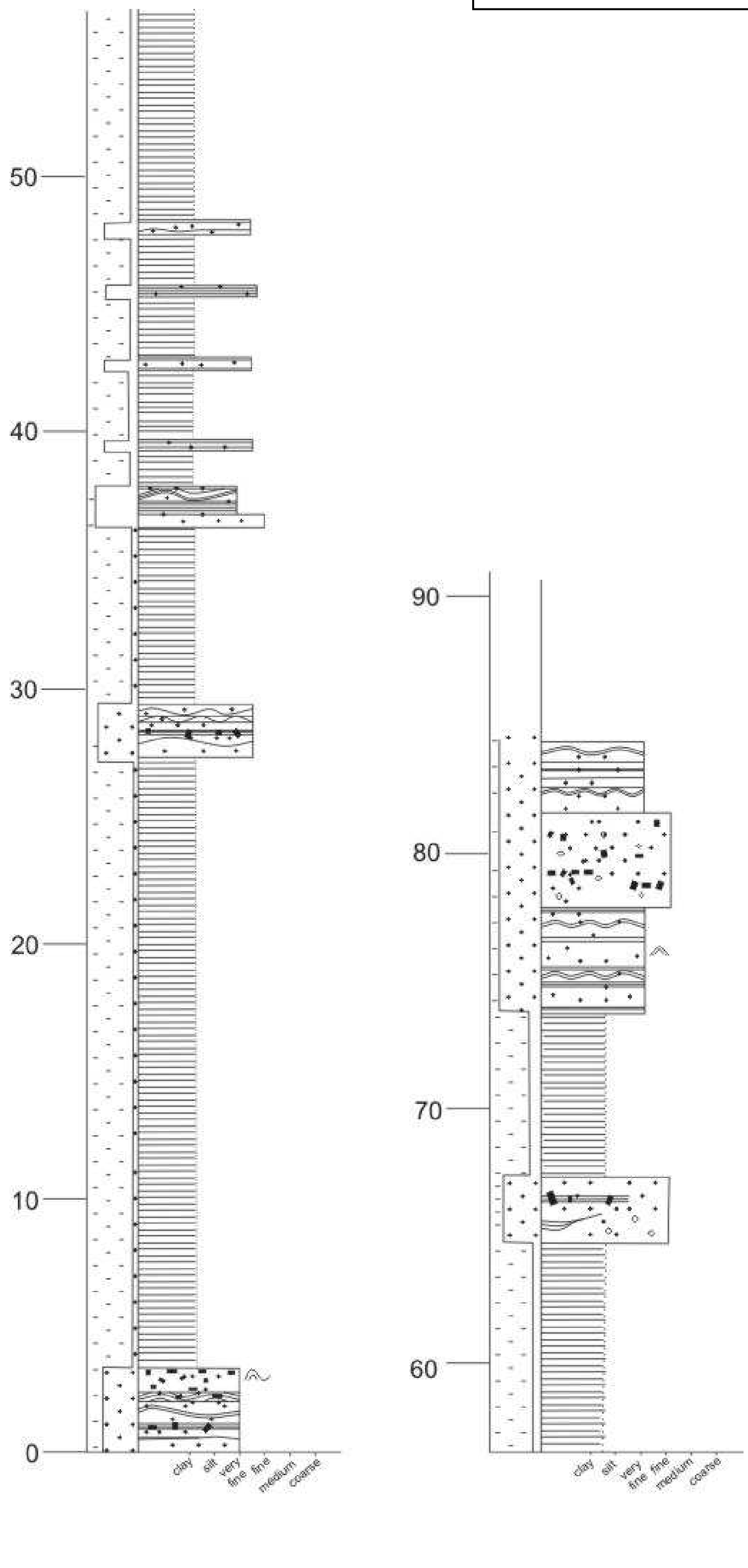

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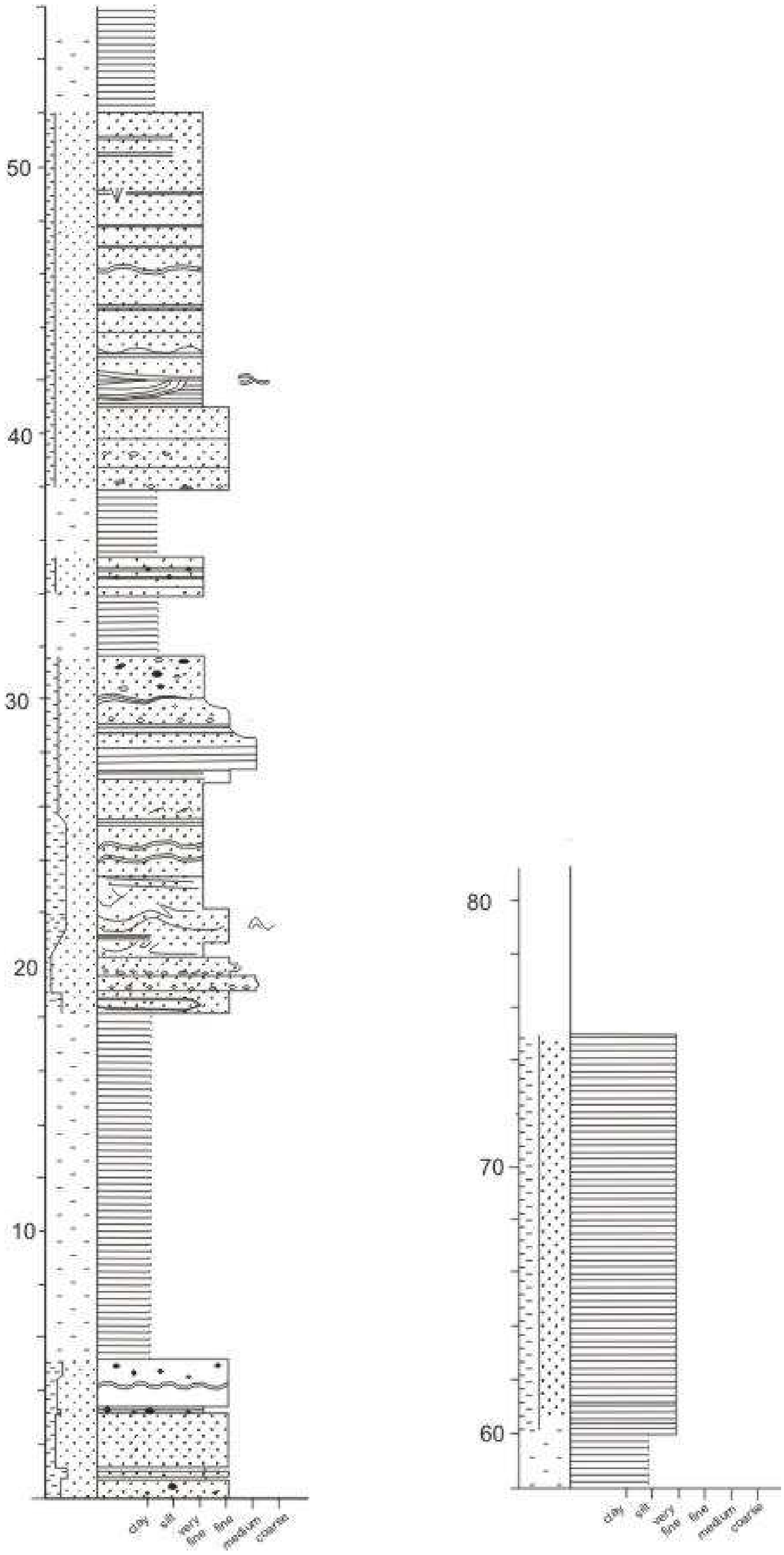

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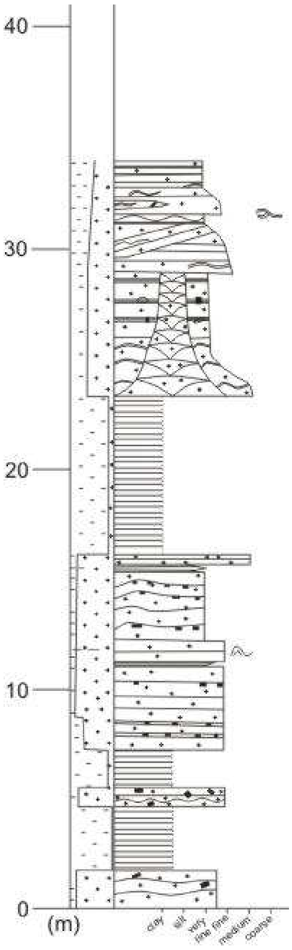

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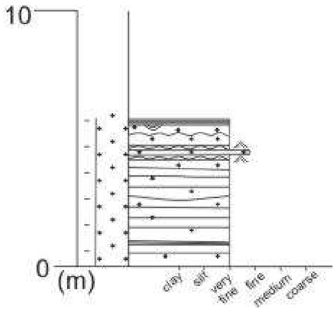

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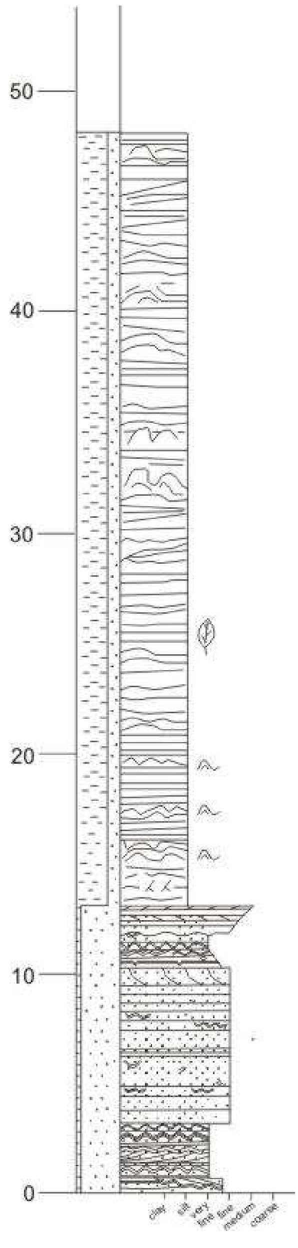

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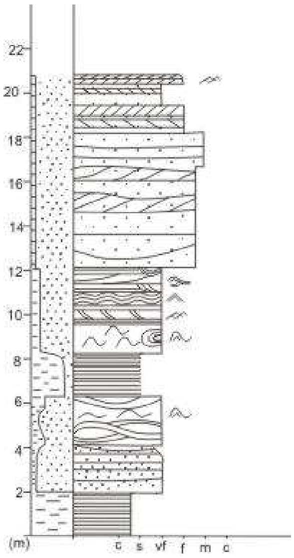

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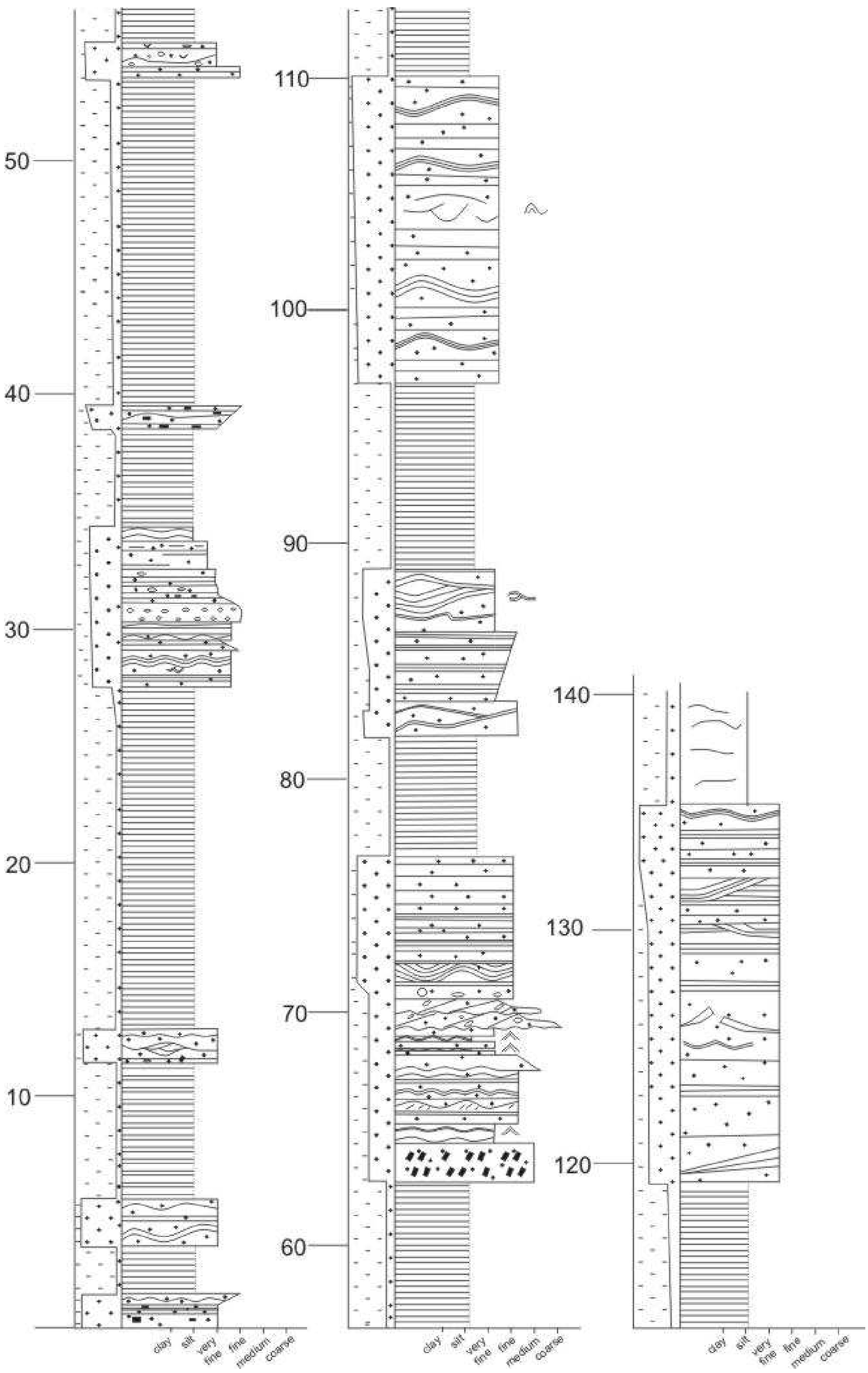

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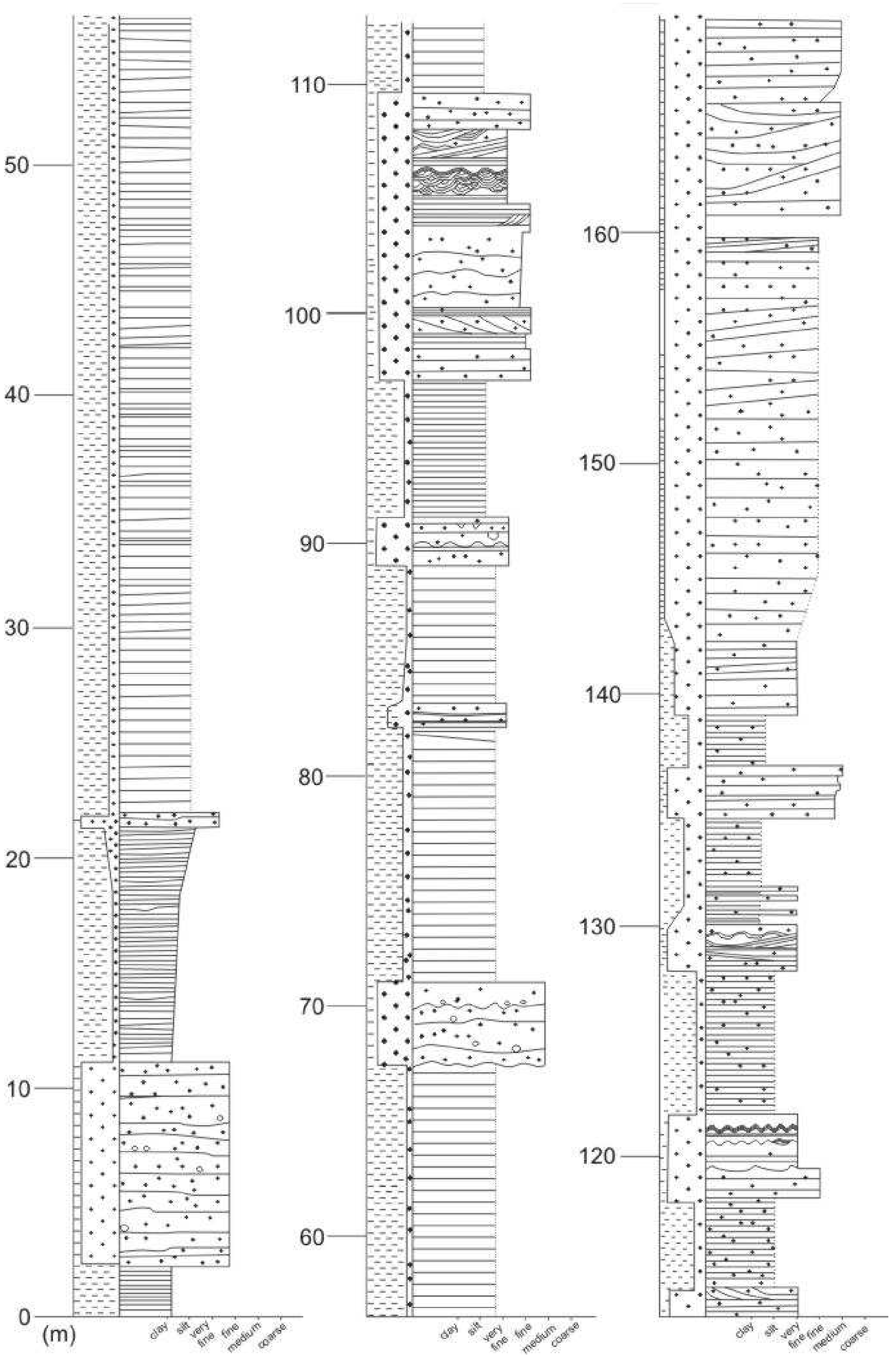

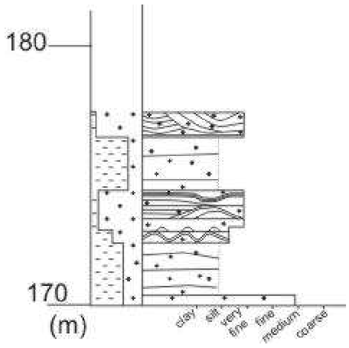
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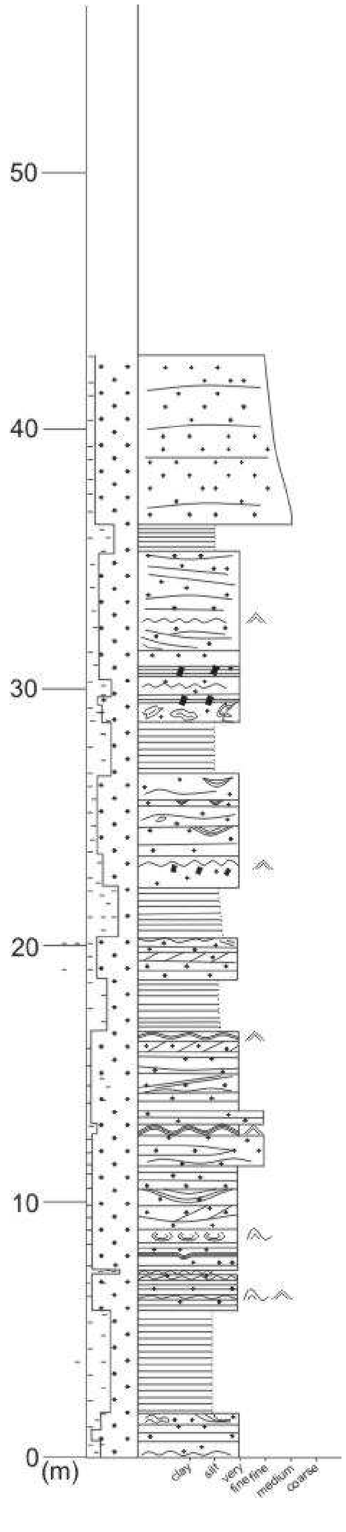

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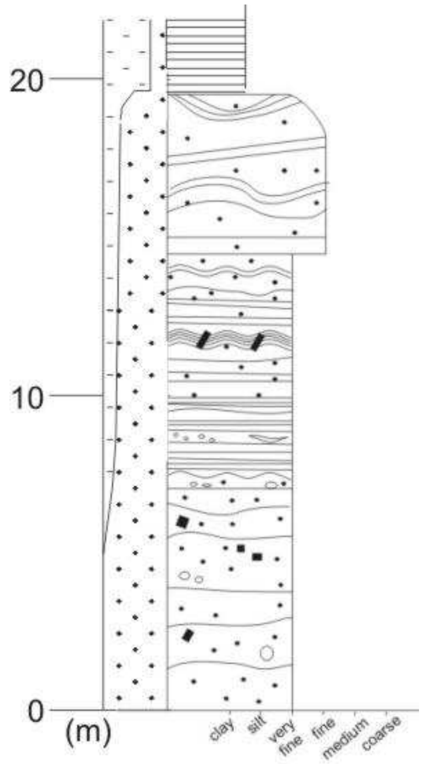

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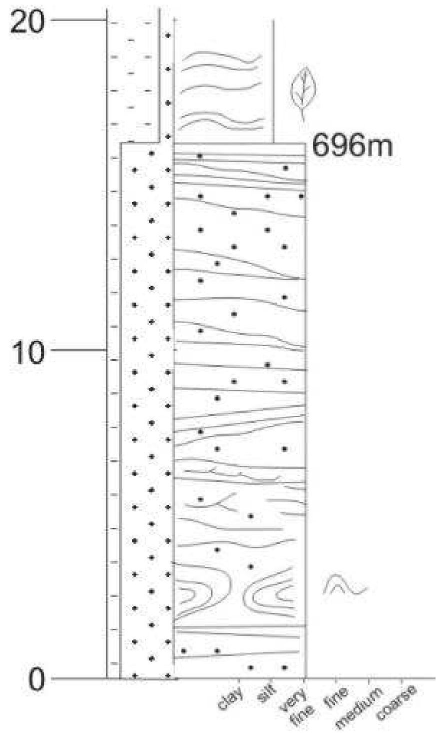

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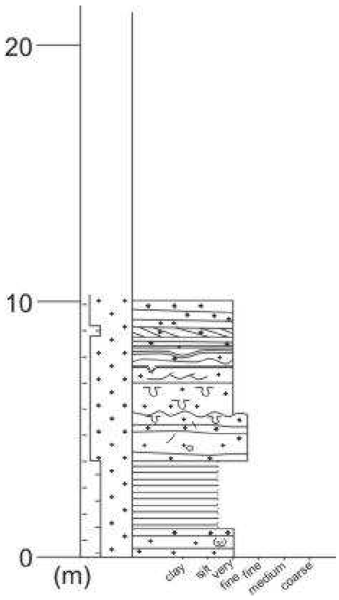

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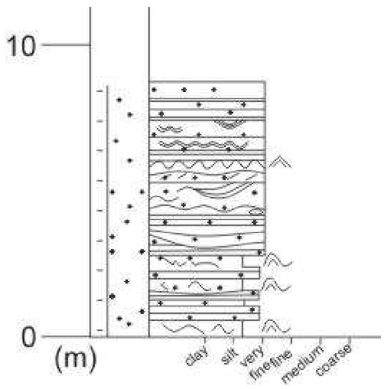
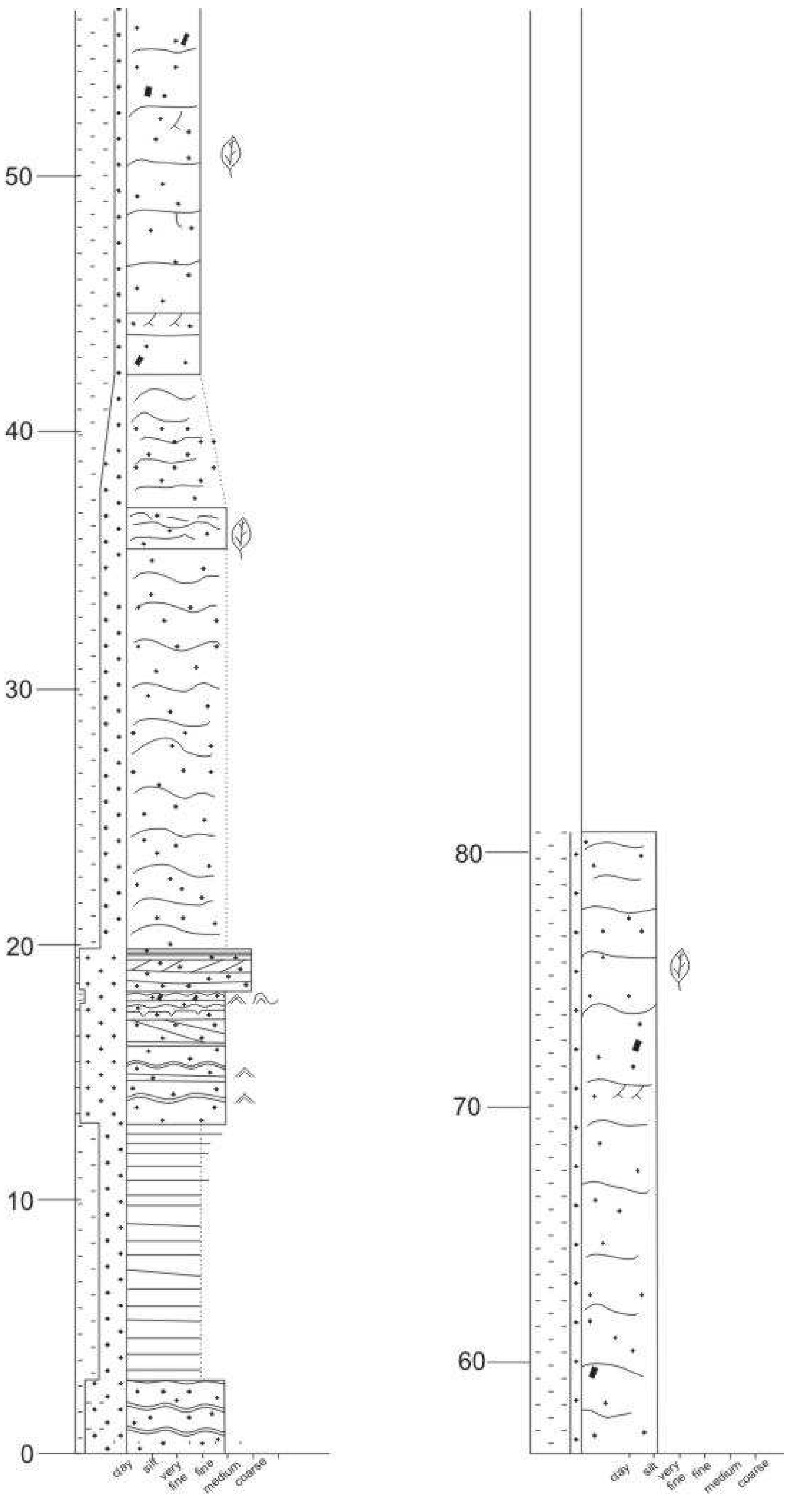
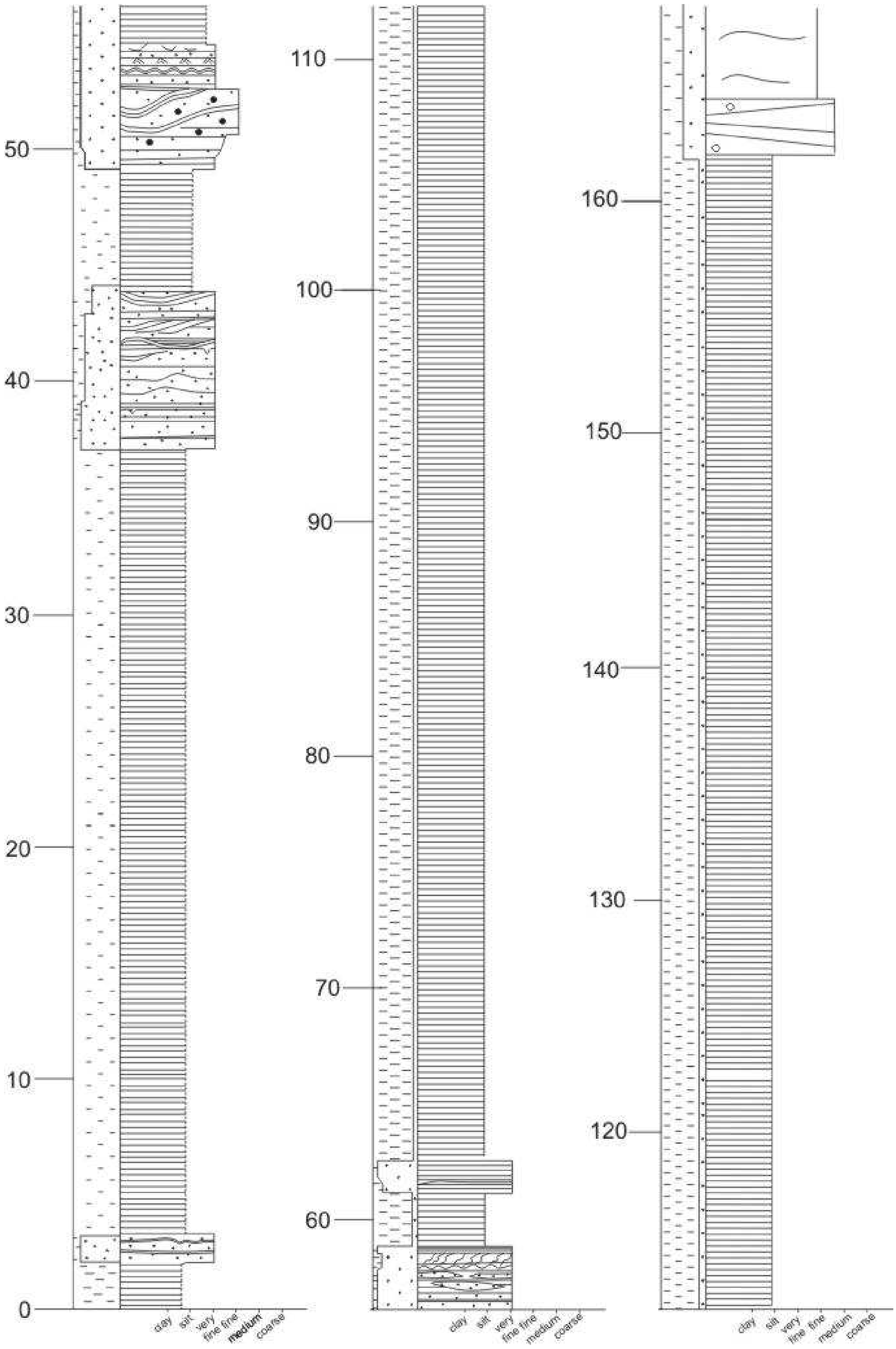
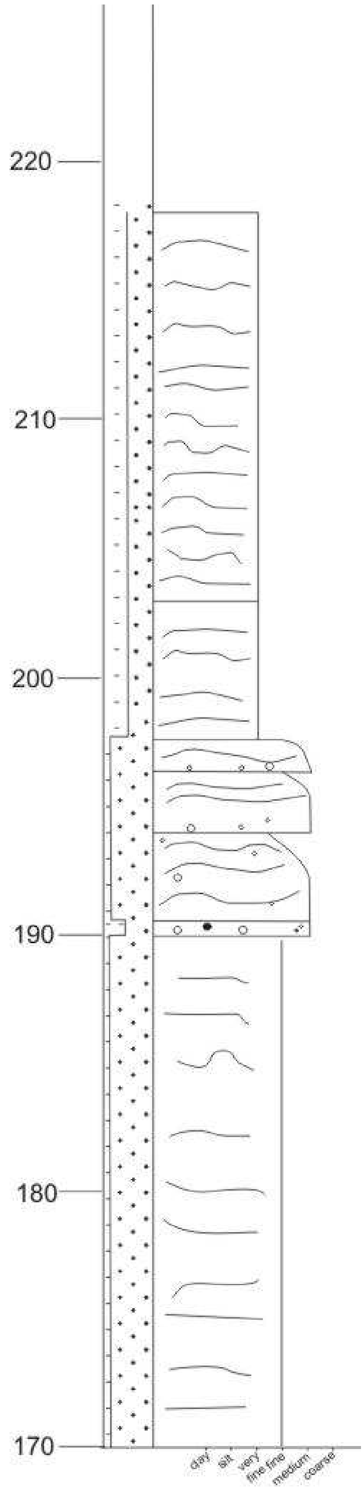


Log 22 - Kjuklingetoppane 8:



Log 23 - Profeten:


0
10
20
30
40
50
60
70
80
90
100
110
120
130
140
150
160
clay silt very fine fine medium coarse
clay silt very fine fine medium coarse
clay silt very fine fine medium coarse


220
210
200
190
180
170
clay
silt
very fine
fine
medium
coarse