University of Stavanger

Faculty of Science and Technology

MASTER'S THESIS

Study program/Specialization: Petroleum Geosciences Engineering	Spring Semester, 2018 Open
Writer: Ken Endre Bukta	 (Writer's signature)
Faculty supervisor: Alejandro Escalona External supervisor(s): Philip Milstead	
Title of thesis: Slørebotn Sub-basin Tectono-Stratigraphic Framework	
Credits (ECTS): 30	
Keywords:	Pages 108 Stavanger, 15.06.2018

Slørebotn Sub-basin Tectono-Stratigraphic Framework

by

Ken Endre Bukta

MSc Thesis

Presented to the Faculty of Science and Technology

The University of Stavanger

Norway

The University of Stavanger

2018

Acknowledgements

Firstly, I would like to thank Professor Escalona and Sr. Geologist Philip Milstead for their guidance throughout this study.

Secondly, I would like to thank Sr. Geologist Chris Parry for sharing his regional knowledge and the rest of the sub-surface team in Spirit Energy Norge for their pleasant reception.

Finally, this study acknowledges Ichron Ltd. for their interpreted biostratigraphic data and GeoProvider AS for providing the 2017 reprocessed 2D seismic lines.

Abstract

Slørebotn Sub-basin Tectono-Stratigraphic Framework

Ken Endre Bukta

The University of Stavanger, 2018

Supervisor(s): Alejandro Valera Escalona, Philip Milstead

The Slørebotn Sub-basin remains today as an underexplored region on the Norwegian continental shelf, in contrast to the well explored Halten Terrace and northern North Sea that have proven to be prolific hydrocarbon provinces. Therefore, a re-examination of the Slørebotn Sub-basin of the Mesozoic to Cenozoic tectono-stratigraphic evolution has been conducted in order to evaluate the play potential for the Slørebotn area. In this study, seismic reflection, well and core data have been used to define the tectono-stratigraphic framework that comprises nine sequences, ranging from: 1) Carnian-Rhaetian, consisting of arid alluvial rocks; 2) Hettangian-Toarcian, consisting of arid alluvial rocks; 3) Aalenian-Callovian, consisting of alluvial to marginal marine rocks; 4) Oxfordian-Early Ryazanian, consisting of deep marine anoxic shale that is interbedded by marine sandstones; 5) Ryazanian-Albian, consisting of open marine shales with poorly sorted slope aprons in its basal part; 6) Cenomanian, consisting of open marine shales; 7) Turonian, consisting of open marine shales and several coarse submarine fans; 8) Late Turonian-Early Maastrichtian, consisting of open marine shales, but coarse submarine fans and turbidites characterize its basal and upper part, respectively; and 9) Lower Paleocene-Early Pleistocene, consisting of slope and basinfloor turbidites at the base and deep marine Eocene turbidites. The latter was established by seismic interpretation and is to date, an untested play in the mid-Norwegian Sea. By analogy to the North Sea, the Eocene turbidite play has proven to be successful with substantial amounts of hydrocarbons discovered to date. The main difference between the

Slørebotn Sub-basin and the adjacent northern North Sea and the Halten Terrace can be observed in the Early Jurassic stratigraphic record. Although all the three reference areas experienced tectonic movement during the late Early Jurassic, the study area was clearly exposed to a greater magnitude of uplift that resulted in erosion of the entire Lower Jurassic and in some places older sedimentary rocks as well. The reason is interpreted to be related to the development of the Møre Basin in accordance to an upper-plate margin in an extensional regime and that underplating by igneous rocks is causing the excessive uplift of the Slørebotn Sub-basin.

Table of Contents

List of Figures

List of Tables

1 INTRODUCTION

The mid-Norwegian shelf has been an area of extensive oil and gas exploration, where the majority of discoveries are located on the Halten Terrace; the principal play type has primarily been Middle and Early Jurassic clastics in tilted fault blocks that are charged from the Late Jurassic Kimmeridge Clay Formation – equivalent to the Spekk Formation (Swiecicki et al., 1998). In recent years, the focus has shifted towards deep-water (> 500m) prospectivity situated within the Møre and Vøring basins. These basins are characterized by a series of structural sub-basins and highs (Fig. 1.1), which mainly formed during the Late Jurassic-Early Cretaceous extensional events (Talwani and Eldholm, 1977; Brekke and Riis, 1987; Blystad et al*.,* 1995; Grunnaleite and Gabrielsen, 1995; Jongepier et al., 1996; Doré et al., 1997b; Gabrielsen et al., 1999; Fagerland, 1990; Swiecicki et al., 1998; Brekke, 2000; Osmundsen et al., 2002; Mosar, 2003; Faleide et al., 2010). The study area is situated on a passive rift margin (Blystad et al., 1995) located along the Norwegian coastline, bounded by the Tampen Spur in the south, Jan Mayen Lineament in the north and the Møre-Trøndelag Fault Complex in the west that separates the Slørebotn Sub-basin from the deeper Møre Basin (Fig. 1.1).

1.1 Previous work

The study area is dominated by NE-SW trending structural elements that developed within the context of the North Atlantic rift system (Blystad et al., 1995). It has been suggested that the main rift episodes took place during the mid-Carboniferous, Carboniferous-Permian, Permian-Early Triassic, Late Jurassic-Early Cretaceous, mid?-Cretaceous times, followed by further rifting in the Late Cretaceous to Early Eocene that led to the breakup and onset of sea-floor spreading between NW Europe and Greenland (Skogseid et al., 1992; Skogseid et al., 2000; Osmundsen et al., 2002; Faleide et al., 2010). The mid-Cretaceous rift episode is still a matter of debate amongst some authors. For example, Zastrozhnov et al. (2018) argued that the entire mid-Norwegian margin was tectonically active during the mid-Cretaceous in contrast to Færseth and Lien. (2002), who argued that this was a time of thermal subsidence across the Møre and Vøring basins. The further renewed rifting during the Late Cretaceous-Eocene times is not obvious within the Møre Basin compared to the Vøring Basin further north (Skogseid et al., 1992; Skogseid et al., 2000; Brekke, 2000; Faleide et al., 2010). However, some reactivation along the Jan Mayen Lineament and at the Gossa High can be observed in the form of compressional features such as domes/ anticlines and inversion (Brekke and Riis, 1987) (Fig. 1.1 and Fig. 1.2).

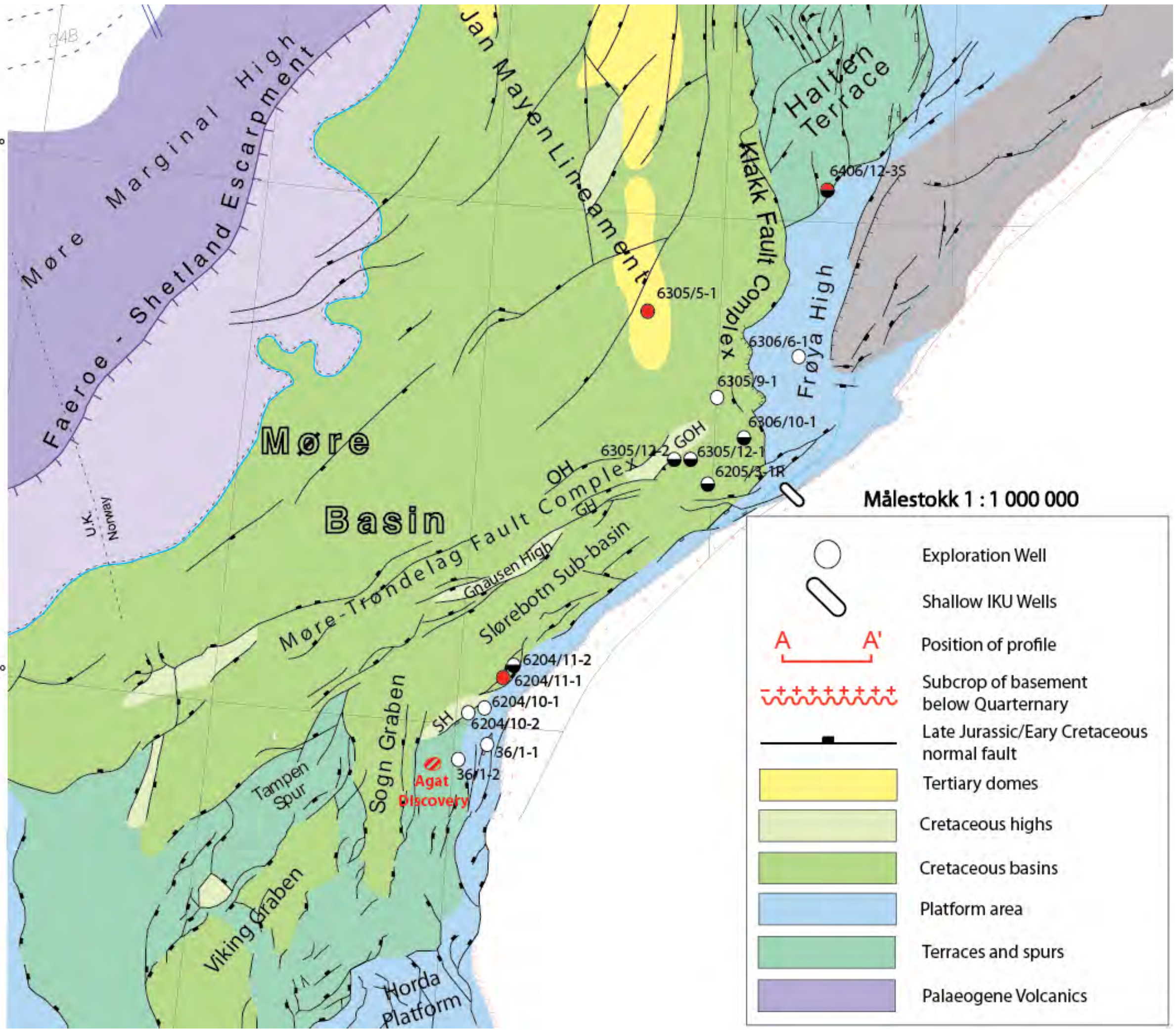


Fig. 1.1 Study area (red circle) and the structural elements that make up the Norwegian Sea. *GH=Giske High, GOH=Gossa High, OH=Ona High, and SH=Silje High (Modified from NPD, Bulletin No. 8).*

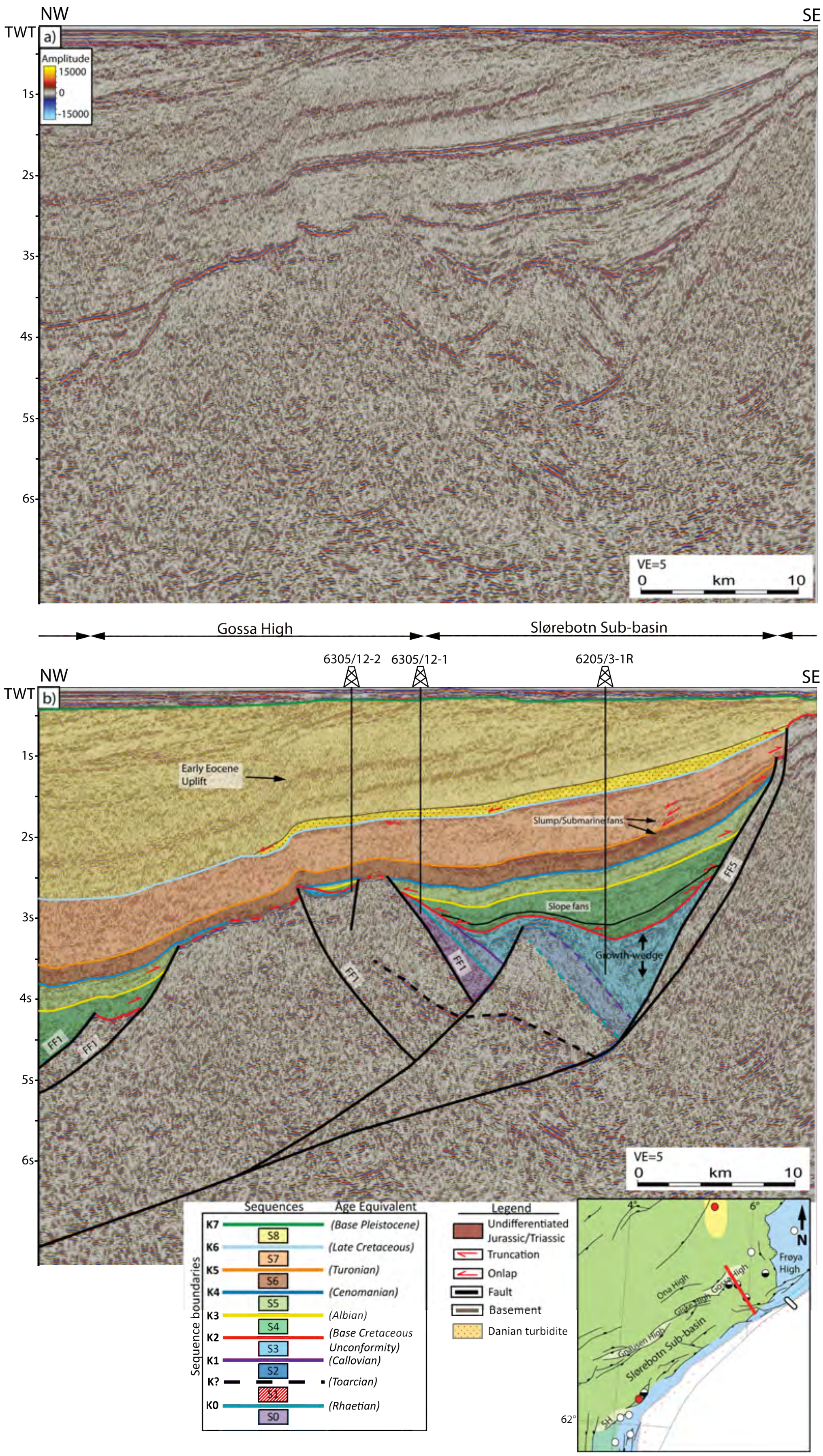


Fig. 1.2 ***Regional 2D profile crossing the Gossa High*** *a) uninterpreted and b) interpreted line. The dashed K1 and K0 are assumed to exist below well 6205/3-1R since the boundaries are proven by well 6305/12-1 in the related hanging wall. The black dashed line represents the proposed detachment of the Slørebotn Sub-basin by other authors (Jongepier et al., 1996; Osmundsen and Ebbing, 2008).*

To date, ten exploration wells and eight shallow boreholes have been drilled in the study area (Fig. 1.1) and the oldest sedimentary rocks penetrated are of Early Triassic age (Table 1.1). Several wells (i.e. 6306/6-1, 6305/12-1 and 6204/11-1) have shown a large stratigraphic unconformity separating the Early Triassic strata from the unconformably overlaying Middle Jurassic (Jongepier et al., 1996). The fact that the Jurassic and Triassic lithostratigraphic sections are so heavily eroded and deeply buried in the Slørebotn Sub-basin compared to the Halten Terrace and the northern North Sea, highlight some of the important differences and geological challenges in the study area. Jongepier et al. (1996) suggested that the reason for the geological differences is related to a phase of uplift and erosion during late Early to Middle Jurassic. While gentle flank uplift and erosion, and continued deposition took place in both the Halten Terrace and the northern North Sea, the northeastern margin of the Møre Basin, however, experienced major erosion of almost the entire Lower Jurassic. Only on the platform area, east of the Slørebotn Sub-basin (Fig. 1.1) is the Lower Jurassic strata proven by shallow IKU wells (Smelror et al., 1994). Adding to the geological problem is the renewed uplift and erosion of the Late Jurassic source rock during the Late Jurassic-Early Cretaceous extensional event, followed by thermal subsidence (Fagerland, 1990; Grunnaleite and Gabrielsen, 1995; Swiecicki et al., 1998). This resulted in a complex basin configuration and significant depths of the pre-Cretaceous strata (> 5s TWT) within the Slørebotn Sub-basin. Several authors (Fagerland, 1990; Blystad et al., 1995) have stated that there are uncertainties in the pre-Cretaceous stratigraphic interpretation due to the depth and limited wells that have encountered the pre-Cretaceous strata (Table 1.1). However, the Base Cretaceous Unconformity (BCU) is usually recognized, thus making it possible to estimate the maturity of the Late Jurassic source rock.

Table 1.1 Overview of the ten exploration wells used for this study *The information is obtained from the NPD factpages. In = informally named.*

Wells	Type	Drilling completed	Total depth (MD)	HC Formation level	Oldest rock penetrated age	Oldest penetrated Formation	Cored Formation	Core length (m)	Encountered basement	Available checkshot
6204/10-1	Exploration	1995	2709	Dry	Aptian	Agat Fm	Balder/Lisa Fms	16,6	Yes	Yes
							Kyrre Fm	36,1		
6204/10-2R	Exploration	1997	2095	Gas	Hauterivian	Åsgard Fm	Kyrre Fm	17,6	Yes	No
							Åsgard Fm	11,1		
6204/11-1	Exploration	1994	2966	HC shows	Upper Triassic	Grey Beds (In)	Kyrre Fm	43,1	No	Yes
				Gas			Intra Heather Ss	67,4		
6204/11-2	Exploration	1997	2920	Oil	Late Jurassic	Sognefjord Fm	-	-	No	No
6205/3-1R	Exploration	1990	5264	Gas/oil shows	Late Jurassic	Spekk Fm	Åsgard Fm	8,4	No	Yes
				Gas/oil shows			Spekk Fm	46,6		
6305/9-1	Exploration	2001	2655	Dry	Late Cretaceous	Springar Fm	Egga Member (In)	17,4	No	Yes
6305/12-1	Exploration	1991	4302	Weak shows	Late Triassic	Red Beds (In)	Garn Fm	0,3	No	Yes
				Weak shows			Red Beds (In)	11,1		
				Weak shows						
6305/12-2	Exploration	1993	3162	HC shows	Middle Jurassic	Undifferentiated	Lange Fm	4,3	Yes	No
				Poor shows			Brent Gp	26,9		
							Basement	2,7		
6306/6-1	Exploration	1994	1317	Dry	Late Triassic	Red Beds (In)	Rogn Fm	18,8	No	No
6306/10-1	Exploration	1990	3187	HC shows		Garn Fm	Egga Member (In)	52,7	Yes	Yes
				Gas	Middle Jurassic		Melke Fm	9,3		
							Garn Fm	8		
							Basement	3,7		

The present day structural elements (Fig. 1.1) that define the study area have been developed during several extensional periods, but did not become structurally defined until the Late Jurassic-Early Cretaceous extensional event (Blystad et al., 1995). During this time, several fault systems (i.e. Jan Mayen Lineament, Møre-Trøndelag, Klakk and Møre Margin fault complexes) (Fig. 1.1) are believed to have been highly active and fundamental to the development of the Møre and Vøring basins (Brekke and Riis, 1987; Falgerland, 1990; Blystad et al., 1995; Grunnaleite and Gabrielsen, 1995; Brekke, 2000; Osmundsen and Ebbing, 2008). The evidence of the Late Jurassic-Early Cretaceous structural development can be seen as Lower Cretaceous sediments onlap against basement highs in the study area (Fig. 1.2). The most prominent of these highs are the Gnausen, Giske and Gossa highs (Fig. 1.1) that are believed to be associated with the long-lived NE-SW trending Møre-Trøndelag Fault Complex (Brekke and Riis, 1987;

Blystad et al., 1995; Jongepier et al., 1996; Brekke, 2000; Osmundsen and Ebbing, 2008). The crests of these highs are truncated by the BCU and defined by steep extensional faults in the west, whereas the highs are delineated on their landward side by eastward dipping listric faults (Brekke and Riis, 1987; Fagerland, 1990; Blystad et al., 1995; Jongepier et al., 1996). The magnitude of extension of the Møre-Trøndelag Fault Complex (MTFC) is interpreted to decrease rapidly on the northern side of the Gossa High (Brekke and Riis, 1987), where it is believed to be intersected by the NW-SE trending Jan Mayen Lineament (Fig. 1.1 and Fig. 1.3). The impact of the Jan Mayen Lineament is still not clear, however, it has been suggested that the lineament was involved in forming a structural weak zone between the northeastern end of the Gossa High and the southern end of Frøya High (Gjelberg et al., 2005). In addition, the N-S trending Klakk Fault Complex (Blystad et al., 1995) in the north of the study area can also be seen to intersect the Jan Mayen Lineament (Jongepier et al., 1996). The result of the different fault systems interacting has developed a complex structural architecture around the Gossa High area and several authors have suggested different structural interpretations. More specifically, east for the Gossa High in the Slørebotn Sub-basin, steep pre-Cretaceous dipping reflectors in rotated fault blocks (up to 50°) can be observed (Fig. 1.2) and several attempts have been made trying to explain the origin of these rotated fault block anomalies and the detachment of the Slørebotn Sub-basin:

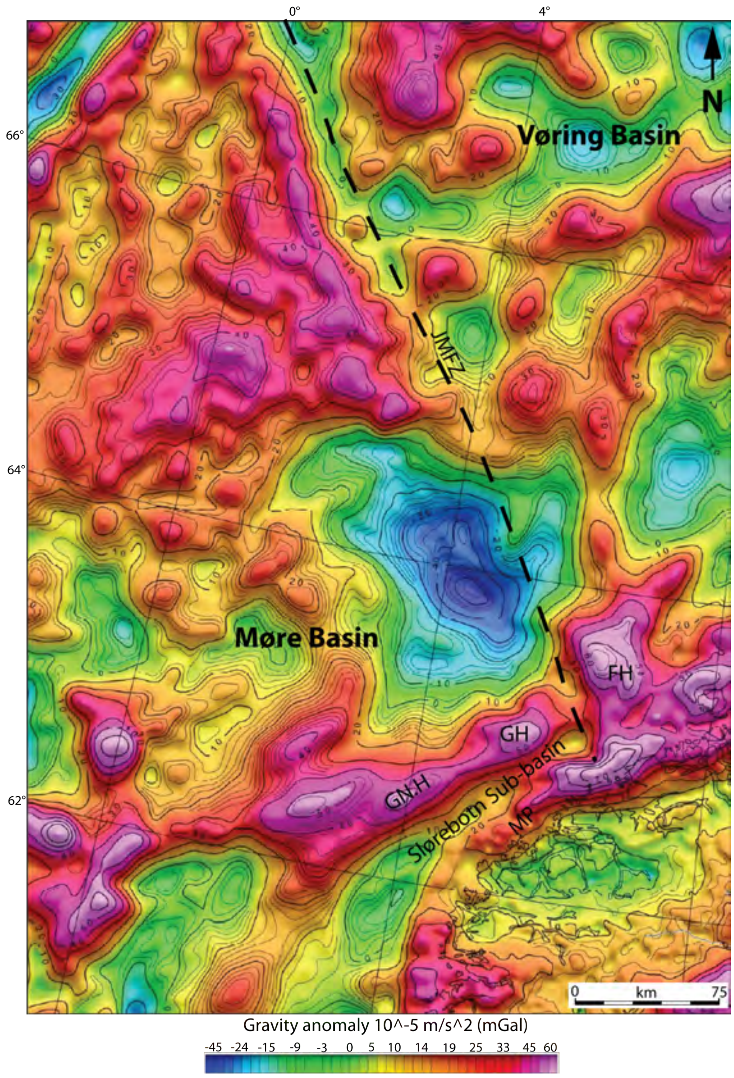


Fig. 1.3 Gravity anomaly map (free-air anomaly offshore) over the mid-Norwegian Sea area. *GH = Gossa High, GN.H = Gnausen High, FH = Frøya High, JMFZ = Jan Mayen Fracture Zone, MP = Møre platform. Modified from Olesen et al. (2010).*

- Osmundsen and Ebbing (2008) propose that the eastern Møre Margin developed according to a lower-plate in a classical extensional model (Lister et al., 1986) (Fig. 1.4a); and
- Jongepier et al. (1996) suggested that the detachment under the Slørebotn Sub-basin was related to the footwall collapse of the southeastern border fault of the Møre Basin (Fig. 1.4b).

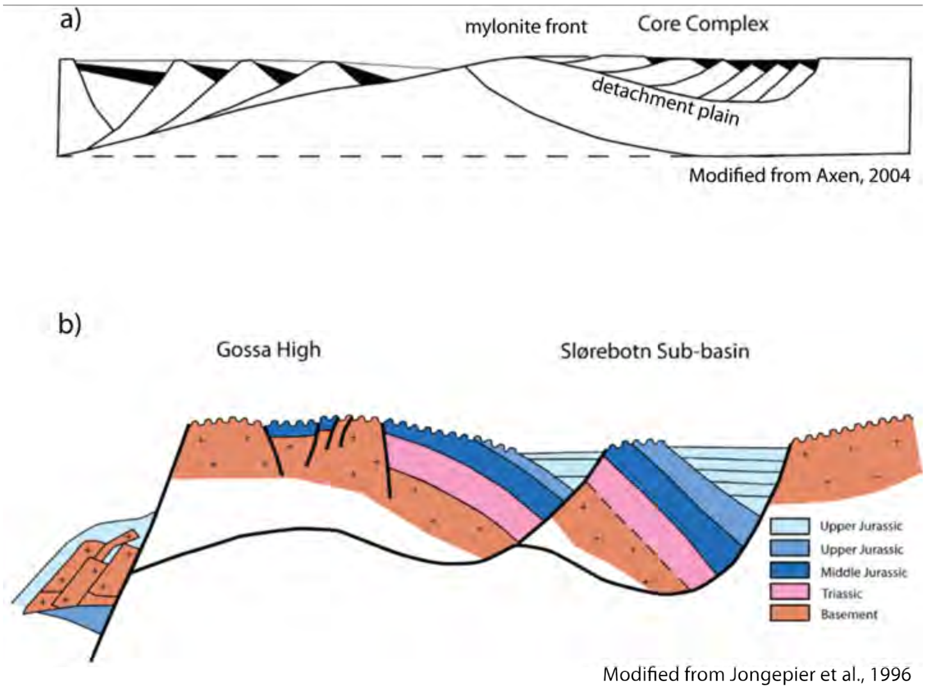


Fig. 1.4 Two potential structural interpretations have been proposed for the northeastern part of the Møre Margin: *a) Slørebotn Sub-basin is part of a core complex of a larger fault system that was fundamental to the development of the Møre Basin; and b), Slørebotn Sub-basin and Gossa High are related to a gravitational collapse of the major eastward bounding normal fault.*

Out of the ten exploration wells, six of the wells encountered hydrocarbon shows, but so far, there have not been proven any commercial discoveries. The primary targets for the wells were mainly Mesozoic strata in rotated fault blocks and structural highs (Fig. 1.1 and Fig. 1.2). It has been proposed that the reason for failure is related to the poor reservoir quality or absence of the Triassic and Jurassic targets (Jongepier et al., 1996; Brekke et al., 1999; Mørk and Johnsen, 2005). The core interpretations of the Upper Triassic and Middle Jurassic units from previous authors have shown that the reservoir quality is poor due to the immature texture and mineralogy, and that carbonate and quartz cementation has further reduced the initial reservoir properties (Jongepier et al., 1996; Mørk and Johnsen, 2005). In addition, Fagerland (1990) and Swiecicki et al. (1998) stated that there are very few potential prospects along the Slørebotn Sub-basin because the Upper Jurassic source rock is either over-mature along the sub-basin or prone to erosion on the adjacent highs (Fig. 1.2 and Fig. 1.5). While the poor reservoir quality and insufficient hydrocarbon charge are considered the main risks in the region, the top seal, however, is not (Mudge et al., 2007). Doré et al, (1997a) suggested that potential reservoir units of Mesozoic age are likely to be sealed by thick Cretaceous mudstones, which is the dominating lithology in the Møre Basin (Dalland et al., 1998; Vergara et al., 2001).

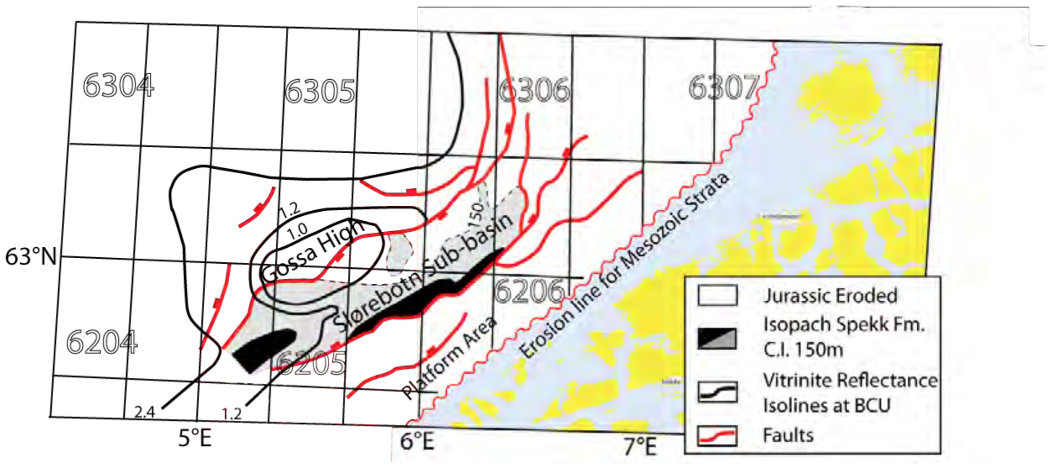


Fig. 1.5 Isopach map with maturity isolines for the Late Jurassic source rock (Spekk Formation). *The map covers the northern part of the Slørebotn Sub-basin. Modified from Fagerland (1990).*

The majority of previous work that embraces the study area has primarily focused on a regional scale interpretation of the mid-Norwegian Sea (Fagerland, 1990; Blystad et al., 1995; Swiecicki et al., 1998; Vergara et al., 2001). The more detailed work of the Slørebotn Sub-basin was focused around the Gossa High in the north within the Mesozoic intervals (Jongepier et al*.,* 1996). Therefore, a re-examination of the entire Slørebotn Sub-basin from Mesozoic to Cenozoic sequences should be conducted, in order to achieve a more detailed and comprehensive tectono-stratigraphic framework. In addition, the seismic and well data coverage has increased since previous studies (Appendix A) and the seismic imaging quality has improved during the past decades. Long-offset and wide-angle seismic acquisition combined with improved processing techniques have enhanced the seismic quality, and a more detailed interpretation as well as resolving deeper reflections of the subsurface is now possible. The increased data coverage and improved seismic quality will aid in a more detailed interpretation and therefore improve the knowledge of the distribution of potential petroleum related elements in the study area.

1.2 Objectives

The main objective for this study is to establish a tectono-stratigraphic framework to improve the current knowledge of the depositional and structural controls in the Slørebotn Sub-basin. The sequence stratigraphic framework presented in this thesis is based on Ichron (2015) interpreted biostratigraphic data that has been obtained from cores and cuttings from the exploration wells in the study area. Moreover, the framework presented in this study is further used to compare with the adjacent hydrocarbon provinces, such as the northern North Sea and Halten Terrace (Fig. 1.1), in order to evaluate the play potential of the Slørebotn Sub-basin. In addition, the interpretation of the structural evolution of the study area will be compared to the previous models (i.e. Jongepier et al., 1996; Osmundsen and Ebbing, 2008). In more specific terms, this study aims to;

- Establish a sequence stratigraphic framework for Mesozoic and Cenozoic units based on well log signature, core descriptions and biostratigraphic data;
- Describe the geometry, trajectory, and lateral variability of the main stratigraphic units that can be recognized and interpreted throughout the area by seismic reflection data;
- Establish the depositional history by seismic interpretation, seismic facies recognition, well log signature and existing core descriptions; and
- Understand the mechanisms controlling the basin infill pattern by combining structural and time-thickness maps.

2 GEOLOGICAL SETTING

The study area is situated on a passive rift margin (Blystad et al*.,* 1995) located along the Norwegian coastline, bounded by the Tampen Spur in the south and the Jan Mayen Lineament in the north (Fig. 1.1). The structural elements that define the study area can be grouped into two main trends; NE-SW and NW-SE (Blystad et al., 1995). The NW-SE trending feature is expressed as major lineaments that are believed to have originated in the Precambrian time (Brekke and Riis, 1987; Brekke, 2000). While the NE-SW trend is believed to have originated in the Late Paleozoic and been active until the final break up in the Eocene (Brekke, 2000). The majority of the structural elements that can be seen in the study area are characterized by the NE-SW trend (Fig. 1.1).

2.1 Tectonostratigraphic evoluition

2.1.1 Tectonic framework

The structural evolution of the Norwegian Sea area is relatively well known (Brekke and Riis, 1987; Blystad et al*.,* 1995; Doré et al., 1997b; Swiecicki et al., 1998; Brekke, 2000; Vergara et al., 2001; Osmundsen et al., 2002; Faleide et al., 2010). The mid-Norwegian Sea area changed from a compressional stress regime during the Caledonian Orogeny to extensional stress from the Late Devonian until the continental separation and onset of sea-floor spreading of the North Atlantic during the Eocene (Talwani and Eldholm, 1972). During this period, the area has undergone several major rift events and it has been suggested that the main Late Paleozoic to Early Mesozoic rift episodes took place during the mid-Carboniferous, Carboniferous-Permian, and Permian-Early Triassic times. These rift events formed a system of NNE-SSW trending rotated fault blocks (Blystad et al*.,* 1995; Osmundsen et al., 2002; Faleide et al., 2010).

Further crustal extension and thinning of the mid-Norwegian Sea area during the Mesozoic to Early Cenozoic led to the development of major Cretaceous basins and highs (Fig. 1.1). More specifically, the Møre and Vøring basins formed as the result of the Late Jurassic-Early Cretaceous rift event, which is evident as most of the Lower Cretaceous sequences onlap against the basement highs. Accurate dating of the tectonic movements during the Late Jurassic-Early Cretaceous in the study area is possible due to well 6205/3-1R, which targeted a Late Jurassic rotated fault block (Fig. 1.2). Three distinct tectonic events with different degrees of magnitude could be recognized within the Late Jurassic interval based on the core data and dipmeter log data (Jongepier et al., 1996); (1) during the Kimmerdigian, very gentle fault block rotation (5°) occurred; (2) followed by an increase in rotation (15-20°) during the Early Volgian times; and (3) the final phase of fault block rotation (20°) climaxed in the Middle Volgian times, forming both deep and shallow NE-SW trending normal faults (Fig. 1.1 and Fig. 1.2) (Brekke and Riis, 1987; Fagerland, 1990; Blystad et al*.,* 1995; Doré et al., 1997b; Brekke, 2000; Osmundsen et al., 2002).

The Late Cretaceous-Eocene rift episode is well documented to have affected the sedimentation and architecture of the mid-Norwegian Sea (Brekke, 2000; Martinsen et al., 2005; Faleide et al., 2010). In contrast to a potential mid-Cretaceous rift episode (Doré et al., 1997b), which was less prominent and is still a matter of debate (e.g. Zastrozhnow et al., 2018 and Færseth and Lien., 2002). The effect of the mid-Cretaceous rift episode on the Møre Basin stratigraphy is not obvious and it is therefore believed that the basin was tectonically quiet and experienced continuous subsidence due to the crustal cooling during the Cretaceous (Faleide et al., 2010). The Late Cretaceous-Eocene extensional events, however, are far more prominent and can be recognized in the Møre Basin as minor faulting, reactivation of the deeper Jurassic structures and as compressional domes along the Jan Mayen lineament (Grunnaleite and Gabrielsen, 1995; Doré et al., 1997b; Brekke, 2000; Doré et al., 2008; Osmundsen and Ebbing, 2008; Faleide et al., 2010; Ravnås et al., 2014a).

Several rift episodes that influenced the architecture of the mid-Norwegian basins have resulted in significant crustal stretching and weakening of the crust, and gravity data indicates that the crest in the western part of the Møre Basin has been significantly thinned (Fig. 1.3) (Osmundsen and Ebbing, 2008; Faleide et al., 2010). It is suggested that the Vøring and Møre basins were comparable in the pre-Cretaceous time, but developed differently thereafter (Martinsen et al., 2005). Brekke (2000) suggested that the reason for the different basin configuration (Fig. 2.1) is related to the influence of a transfer fault that separates the two basins (Fig. 1.1), the Jan Mayen Lineament, which acted as a tectonic barrier.

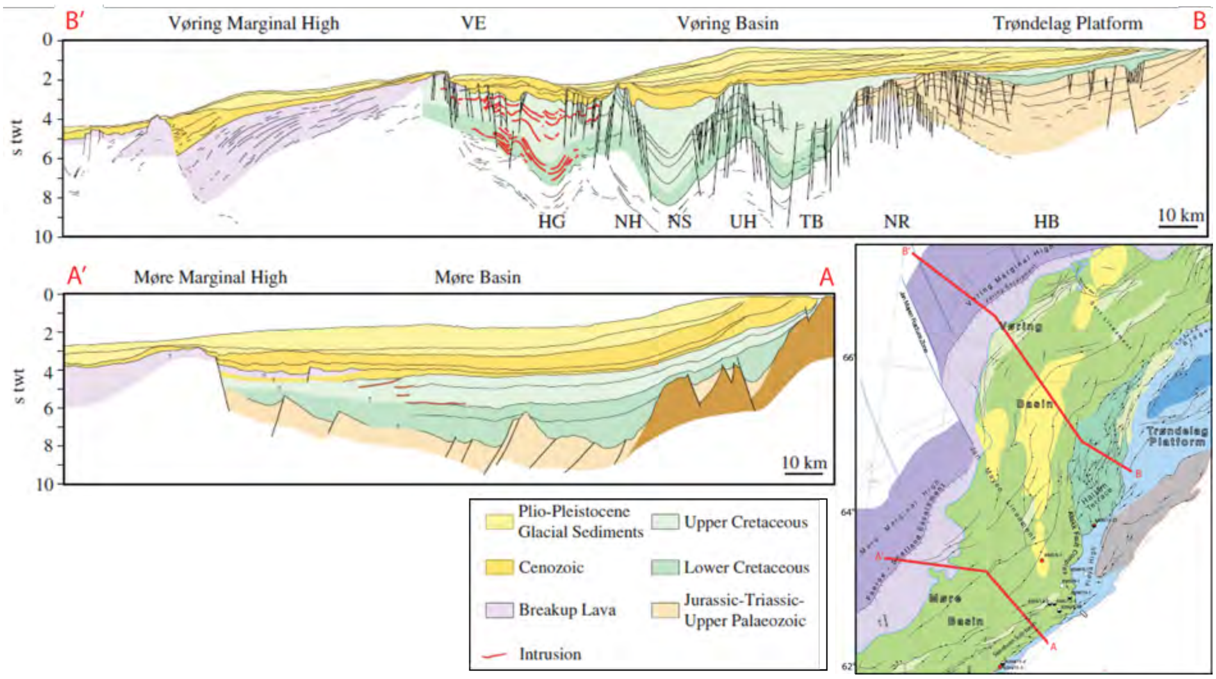


Fig. 2.1 Regional NW-SE cartoon profiles of the Møre and Vøring basins. *The Møre Basin area is characterized by a narrow shelf in the east and the area is relatively unstructured. Whereas, the Vøring Basin is characterized by a wide shelf (Trøndelag Platform) and the area is highly faulted. Modified from Faleide et al. (2010) and Tsikalas et al. (2005).*

2.1.2 Stratigraphic framework

The basins west of mid-Norway (Fig. 1.1) and east of Greenland were located adjacent to one another prior to the onset of the sea-floor spreading of the North Atlantic (Vergara et al., 2001), and it has been suggested from heavy minerals provenance analysis that the pre-Cretaceous to Cretaceous sediments were sourced from both the Fennoscandian and the East Greenland landmasses (Morton et al., 2009; Slama et al., 2011). As the continental drifting continued during the Late Cretaceous until Eocene, the Fennoscandian shield became the more dominant source of sediments of the Norwegian Sea basins (Martinsen et al., 1999; Gjelberg et al., 2001; Vergara et al., 2001; Gjelberg et al., 2005; Martinsen et al., 2005; Faleide et al., 2010).

Permian

The sediment packages associated with the pre-Triassic rift episodes are poorly resolved and have not been confirmed by wells in the study area (Brekke and Riis, 1987; Jongepier et al., 1996; Faleide et al., 2010; Ichron, 2015). Therefore, the Permian stratigraphy will not be considered in the stratigraphic correlation for this study. The reader is advised to see Bugge et al. (2002) for a more comprehensive description of the Upper Permian interval on the mid-Norwegian continental shelf.

Triassic

Rifting during the Permian continued into the earliest Triassic, where the tectonic activity gradually decreased towards the Late Triassic (Swiecicki et al., 1998). Generally, the Middle to Late Triassic sediments (i.e. red and grey beds) are seen as post-rift sequences (Jongiepier et al., 1996; Swiecicki et al., 1998) and have been penetrated by wells in the Slørebotn Sub-basin (Table 1.1). The cored Late Triassic sedimentary rocks have shown that the interval is composing of clastic sediments deposited in an arid alluvial environment (Jongepier et al., 1996; Swiecicki et al., 1998). However, the reservoir quality in this interval has only been encountered rarely (Fagerland, 1990; Jongepier et al., 1996; Swiecicki et al., 1998; Mørk and Johnsen, 2005).

Jurassic

The Jurassic was a time of high tectonic activity and the sediments are believed to have been deposited in an overall transgressive regime (Dalland et al., 1988). The following Jurassic section has been divided into three separate events as the depositional environment and magnitude of the tectonic movements differ.

Lower Jurassic

The transition from the Triassic to Jurassic coincides with a change from dominantly continental to a shallow marine environment (Faleide et al., 2010). Increasingly humid climate with a rapid change from red to grey beds marks the end of a trend, and the beginning of a regional deposition of coal-bearing sediments of the Early Jurassic Åre Formation (Swiecicki et al., 1998; Faleide et al., 2010). The Åre coals are considered to be a good source rock and have been widely encountered in the North Sea and mid-Norwegian Sea (Gjelberg et al., 1987; Swiecicki et al., 1998; Faleide et al., 2010). It has been suggested that marine conditions may have been established as early as the Late Sinemurian times in the mid-Norwegian area (Birkelund and Perch-Nielsen, 1976; Swiecicki et al., 1998), which further increased in the Pliensbachian leading to diachronous deposition of the tide-influenced Tilje Formation (Swiecicki et al., 1997). The marine transgression continued until the end of the Early Jurassic, which led to the deposition of the coarse clastics of the Tofte Formation and the open-marine shales of the Ror Formation (Fig. 2.2) (Dalland et al., 1988; Swiecicki et al., 1998). The Early Jurassic interval has not yet been encountered in the Slørebotn Sub-basin (Fagerland, 1990; Jongepier et al., 1996), only on the platform area, east of the sub-basin, has the Lower Jurassic sediments been encountered by shallow IKU wells (Smelror et al., 1994).

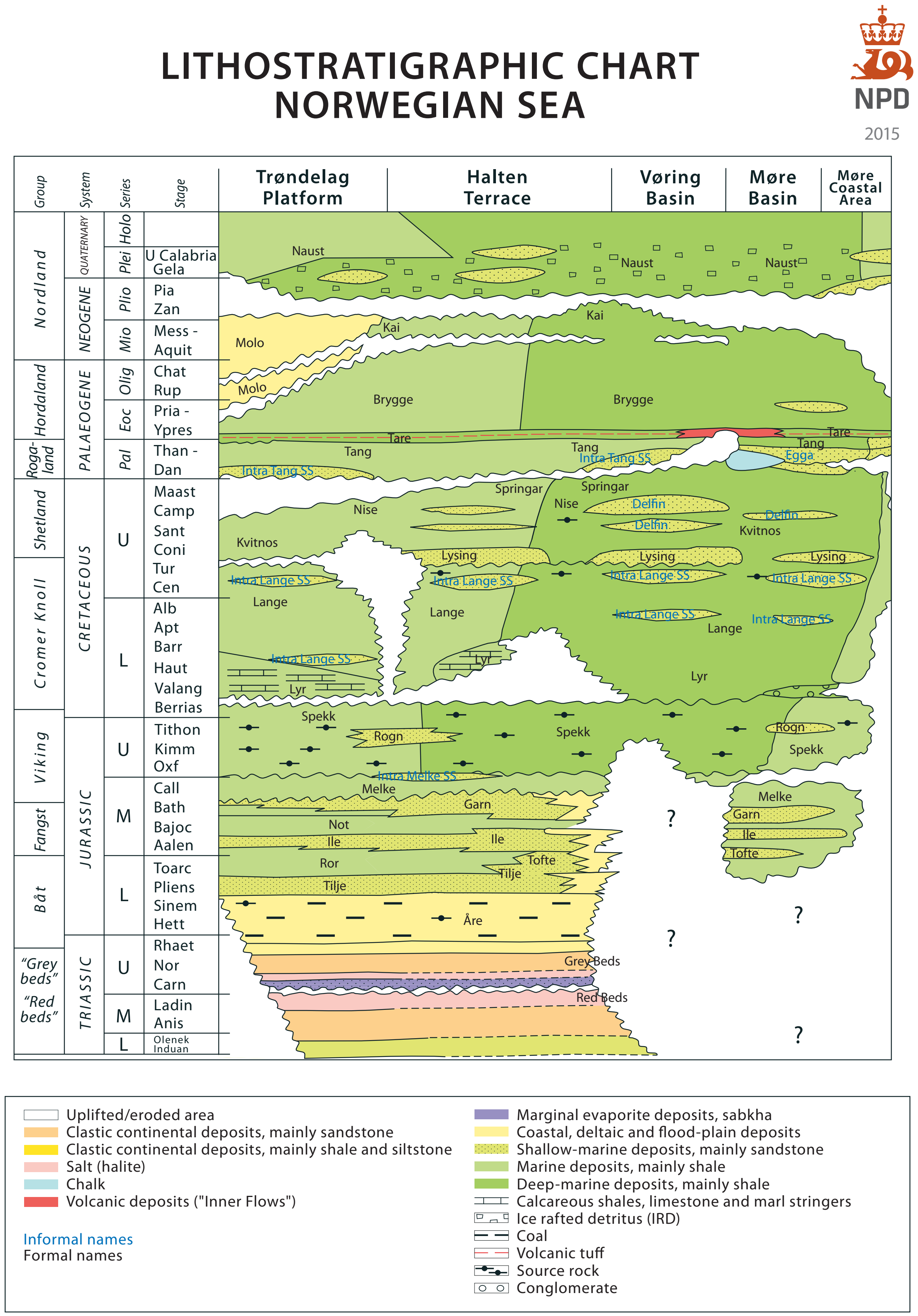


Fig. 2.2 General lithostratigraphic schemes for mid-Norway *Modified from Brekke et al. (2001).*

Middle Jurassic

The base of the Middle Jurassic is an important boundary in the proto-Atlantic rift domain that is characterized in the North Sea by regional uplift, volcanism, and northerly prograding delta systems

(Swiecicki et al., 1998; Faleide et al., 2010). In the mid-Norwegian Sea however, only minor tectonism can be observed in the seismic and well data during the Middle Jurassic time interval (Swiecicki et al., 1998).

There are two major regressive-transgressive cycles recognized in the mid-Norwegian Sea during the Middle Jurassic times. The first cycle, during the Aalenian-Early Bacjocian, resulted in the deposition of the regressive tide-influenced deltaic sandstones of the Ile Formation, followed by the deposition of the transgressive shelfal mudstone of the lower Not Formation (Dalland et al., 1988; Swiecicki et al., 1998). The second cycle took place during the Late Bajocian-mid Callovian, which resulted in deposition of the regressive shoreface sandstones of the Middle Not and Garn formations (Dalland et al., 1988), followed by the deposition of the open marine claystone of the lowest Melke Formation (Fig. 2.2) (Swiecicki et al., 1998).

Several of the wells in the study area have encountered rocks of Middle Jurassic age (Table 1.1) and all wells show a large hiatus that separates the underlying Late Triassic from the overlaying Middle Jurassic strata (Jongepier et al., 1996; Mørk and Johnsen, 2005; Ichron, 2015). The cored Middle Jurassic intervals have shown that coastal plain deposits with a humid climate probably dominated the entire Slørebotn Sub-basin during the Bathonian (Jongepier et al., 1996). The Bathonian-Callovian boundary is suggested to represent a transgressive event and therefore a change in environment from dominantly continental to marginal marine (Jongepier et al., 1996). The sediments of Middle Callovian to lower Kimmeridgian age have not yet been penetrated by wells. However, based on the increased marine trend that led to the deposition of Late Jurassic deep-marine claystone, it is assumed that this was a period of further transgression (Dalland et al., 1988).

Upper Jurassic

The Late Jurassic was a time of significant tectonic activity that resulted in rotation of basement blocks and their overlaying sediments. The crests of the tilted fault blocks were exposed to erosion, thus removing significant portions of the Lower-Middle Jurassic sediments and even Upper Triassic sediments (Faleide et al., 2010). The uppermost Jurassic was a time of further transgression that led to the deposition of thick organic-rich claystones (i.e. the Spekk Formation in mid-Norway, equivalent to the Draupne/ Kimmeride clay formations in the west of the Shetlands and the North Sea, respectively) over most of the Norwegian Continental Shelf (Fig. 2.2), which continued into the Early Cretaceous (Dalland et al., 1988). The Late Jurassic-Early Cretaceous rift event formed a series of over-deepened basins that resulted in wedge-shaped syn-rift sedimentation that was only partially successful in filling the rift topography (Færseth and Ravnås, 1998). Furthermore, the deep basins were characterized by poor bottom water circulations that led to preservation of organic Late Jurassic material in an anoxic environment (Dalland et al., 1988; Faleide et al., 2010).

The Late Jurassic organic-rich claystone has proven to be an important source rock that is capable of generating substantial volumes of hydrocarbons in the mid-Norwegian area. However, across most of the study area, this interval is likely to be over-mature due to the extreme depths (>7000 meter) of the sub-basin (Fagerland, 1990; Jongepier et al., 1996; Swiecicki et al., 1998) with the exception of a narrow strip of mature source rock that was encountered by well 6205/3-1R (Fig. 1.2 and Fig. 1.5). The well proved more than 800 meters of organic rich claystones that are interbedded with deep-marine turbidite sandstones (Jongepier et al., 1996). It is worth noting that father north, on the Halten Terrace, similar sandy deposits provide the reservoir for the Draguen Field (Swiecicki et al., 1998).

Cretaceous

The extensive rift events during the Late Jurassic-Early Cretaceous caused considerable crustal extension and thinning that led to the development of major Cretaceous basins in mid-Norway, East Greenland and the SW Barents Sea (Faleide et al., 2010). These basins underwent rapid differential subsidence and uplift of highs that remained exposed during most of the Early Cretaceous (Jongepier et al., 1996; Faleide et al., 2010). The exposure resulted in the formation of a regional unconformity at the base of the Cretaceous, with the exception in the deeper parts of the basins where there may have been continuous deposition (Faleide et al., 2010). Most of the structural relief that formed has been filled in by the mid- to mid Late Cretaceous time by mainly fine-grained material (Grunnaleite and Garbrielsen, 1995; Færseth and Lien, 2002), sourced from the Norwegian mainland (Brekke, 2000; Vergara et al., 2001; Martinsen et al. 2005; Faleide et al., 2010). In addition, flat-topped marginal highs such as the Giske, Gossa and Frøya highs (e.g. Fig. 1.2), truncated by the base Cretaceous unconformity, are believed to have been local sources for Early Cretaceous slope apron fans into the basin areas (Brekke and Riis, 1987; Fagerland, 1990) and important structures for controlling the distribution of sediments in the Slørebotn Sub-basin (Jongepier et al., 1996; Gjelberg et al., 2005; Martinsen et al., 2005). Such Early Cretaceous fans have been proven by several wells (i.e. 6204/10-1, 6204/11-2 and 6205/3-1) and are believed to occupy large portions of the Slørebotn basin floor (Grunnaleite and Gabrielsen, 1995; Martinsen et al., 2005).

The Upper Cretaceous to Early Cenozoic stratigraphy in the Møre Basin is overall mud dominated (Fig. 2.1 and Fig. 2.2), with some exceptions of sandier intervals in the Turonian to Coniacian Lange and Lysing formations (Martinsen et al., 2005) as well as in the Upper Maastrichtian Springer Formation to the Early Paleocene Egga Member (Ravnås et al., 2014a).

Cenozoic

Several phases of uplift of Fennoscandia took place during the Cenozoic (Riis and Fjeldskaar, 1992; Martinsen et al., 1999; Brekke, 2000; Gjelberg et al., 2001). More specifically, at least five phases of uplift have been reported from the Late Maastrichtian to Plio-Pleistocene (Martinsen et al., 1999; Gjelberg et al., 2001). The mechanism for these uplift phases were probably related to rifting and shoulder uplift of the basin margin that took place prior to the North Atlantic break-up (Riis, 1996; Gjelberg et al. 2005).

The Late Maastrichtian-Early Paleocene phase is of interest for this study as it might have been the primary cause for coarse clastic material being sourced from the uplifted Norwegian mainland and deposited as deep-water turbidites in an otherwise fine-grained dominated Møre Basin (Dalland et al. 1988; Brekke, 2000; Gjelberg et al., 2001; Vergara et al., 2001; Ravnås et al., 2014a). It has been suggested that the turbidites were sourced from a delta that prograded westwards into the Møre Basin, which is not preserved today due to Late Tertiary uplift and erosion (Gjelberg et al., 2001).

In the study area, the base of the Paleocene represents a well-developed unconformity (Grunnaleite and Gabrielsen, 1995) that separates the overlaying Danian turbidite sand (informally named the Egga Member) from the underlying Campanian-Early Maastrichtian strata (Gjelberg et al., 2001). The base-Paleocene unconformity is also present on the Halten Terrace; however, none of the wells on the Halten Terrace have proven any sandy successions above this unconformity (Fig. 2.2). Thus, the Halten Terrace differs from the development of the Slørebotn Sub-basin since the majority of the wells have encountered this sandy interval (Gjelberg et al., 2001). The origin of this unconformity is not well understood, but it is suggested that they are deposited as the result of subaerial erosion related to a significant relative sea-level fall during an uplift of the margin (Gjelberg et al., 2001; Martinsen et al., 2005). This was caused by continued rifting of the North Atlantic during the Late Maastrichtian and Early Paleocene (Riis, 1996).

2.2 Description of the main structural elements

2.2.1 Frøya High

The Frøya High is located in the northern part of the study area (Fig. 1.1) and it is well expressed in the gravity map as an anomaly (Fig. 1.3). The high is bordered by the Klakk Fault Complex in the south that separates it from the deeper Slørebotn Sub-basin (Fig. 1.1). The BCU is interpreted to define the top of the high, which is characterized by a flat and smooth WNW dipping surface (Blystad et al., 1995). In addition, several volcanic rocks in the form of sills have been interpreted above the Frøya High in the Cenozoic stratigraphy, which is dated to 55.7 Ma (Bugge et al., 1980). Internally, the Frøya High is believed to consist mainly of basement rocks (Blystad et al., 1995), Well 6306/10-1 penetrated the basement in the southern part of the Frøya High and the core data shows that the basement comprises retrograde quartz diorite to monzonite plutonic rocks (Mørk and Johnsen, 2005).

2.2.2 Gossa High

The Gossa High is located in the northwestern part of the study area (Fig. 1.1) and is one of several basement highs that are related to the Møre-Trøndelag Fault Complex (Blystad et al., 1995; Brekke, 2000). In the gravity data, the high can be observed as a part of a NE-SW elongated anomaly belt (Fig. 1.3). The high is bounded in the north by the Jan Mayen Lineament, to the west by major normal faults that marks the beginning of the Møre Basin and to the east by highly rotated fault blocks of Late Jurassic to Late Triassic and maybe basement age (Fig. 1.2) (Blystad et al., 1995; Jongepier et al., 1996).

The crest of the high is truncated by a well-developed BCU where Early to early Late Cretaceous seismic reflectors can be observed to terminate against the high (Fig. 1.2). The basement on the Gossa High was penetrated at 3144.5 meters measured depth by well 6305/12-2 (Ichron, 2015) and core data shows that the crystalline basement comprises low-grade metamorphic greenstone (Mørk and Johnsen, 2005).

2.2.3 Gnausen, Giske and Ona highs

The less prominent Gnausen, Giske and Ona highs are all fault bounded basement highs that are related to the Møre-Trøndelag Fault Complex (Blystad et al., 1995). The highs, together with the Gossa High defines the outer margin of the study area and separates the Slørebotn Sub-basin from the deeper Møre basin in the west (Fig. 1.1) In the gravity data, the highs are seen as a part of a NE-SW elongated anomaly belt that suggests the involvement of basement (Fig. 1.3). Reflectors of Lower Cretaceous (S3) truncate against the highs, which implies that they have been structurally active prior to deposition of the Lower Cretaceous sediments.

2.2.4 Silje High

The Silje High is a NE-SW structural element that is located on the southeastern flank of the Slørebotn sub-basin, bounded in the east by the Møre platform (Fig. 1.1). The uplifted high is defined by mid Late Jurassic faults and it is belived to be related to the Møre-Trøndelag Fault Complex (Blystad et al., 1995).

2.2.5 Slørebotn Sub-basin

The Slørebotn Sub-basin is a NE-SW elongated structural feature that is bounded in the west by the Møre-Trøndelag Fault Complex, in the north by the Klakk Fault Complex, in the east by the Møre platform and in the south by the Tampen Spur (Fig. 1.1). The Slørebotn Sub-basin is composed predominantly of

Cretaceous and Cenozoic sedimentary rocks and the oldest seismic sequence boundary that can be interpreted with confidence throughout the Slørebotn Sub-basin is the Base Cretaceous Unconformity (Fagerland, 1990; Jongepier et al., 1996).

2.2.6 Møre Platform

The Møre platform is located in the eastern part of the study area, bounded by the Norwegian mainland in the east (Fig. 1.1). In the west, the platform is bordered by a major listric fault complex that separates the shallow platform from the deeper Slørebotn Sub-basin (Fig. 1.2). The crest of the platform is defined by the BCU and within the Slørebotn Sub-basin, Upper Jurassic to Early Cenozic reflectors can be observed to terminate against the platform (Fig. 1.2). In the gravity data, the platform is characterized by medium to high gravity values (Fig. 1.3).

3 DATA AND METHODOLOGY

3.1 Dataset

This study utilizes well data, 2D seismic lines and a 3D seismic cube that are provided by Spirit Energy Norge. In addition, eleven regional 2D lines that cover a length of 1275 km were reprocessed in 2017 by GeoProvider, utilizing broadband processing techniques and modern demultiple methods in an attempt to enhance the seismic imaging. The reprocessed 2D seismic coverage is limited in to the central parts of the study area (Fig. 3.1a). The total seismic reflection data covers an area of approximately 11150 km^2 (Fig. 3.1b) and the quality of the seismic images varies throughout the study area (Appendix A). Most of the 2D lines originated from different surveys and therefore, the dominate frequencies and phase varies (Appendix A). In addition, a regional velocity cube, provided by FirstGeo was used to convert the interpreted seismic horizons from time to depth.

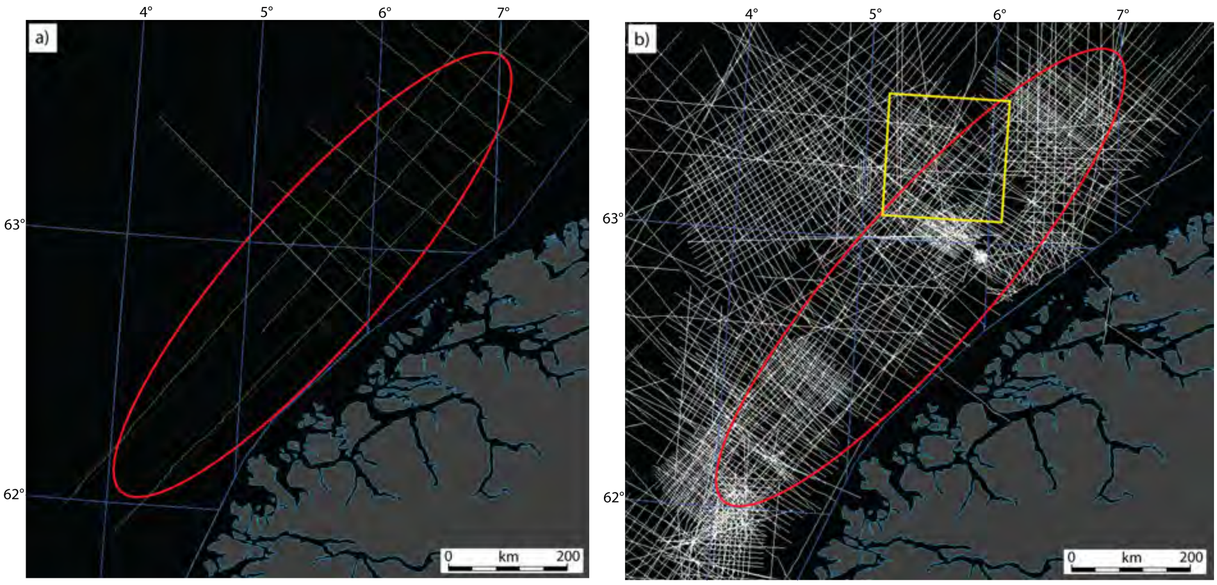


Fig. 3.1 Overview of the seismic coverage in the study area (red ellipse); *a), 2017 reprocessed 2D seismic lines (green lines); and b), conventional 2D seismic lines (white lines) and 3D seismic cube (yellow square).*

The sequence stratigraphic framework presented in this study is based on chrono-stratigraphic picks provided by Ichron (2015), for the ten exploration wells used in this study (Table 1.1). Their stratigraphic database is supported by biostratigraphic data that is collected from cores and cutting samples. The exploration wells consist of well logs such as Gamma Ray, Spontaneous Potential, Density, Neutron, Sonic, and Resistivity. Furthermore, six of the exploration wells have available checkshot data (Table 1.1) that were used for the seismic-well ties.

There are eight shallow IKU boreholes drilled close to the Norwegian mainland, on the platform east for the Slørebotn Sub-basin (Fig. 1.1), where cores were taken from the Late Triassic-Early Jurassic and Early Cretaceous intervals (Smelror et al., 1994). The exploration wells cored a total of 402.1 meters, mainly restricted to the Mesozoic intervals with the exceptions of 70.1 meters that were cut from the Lower Paleocene level (Table 1.1). Core descriptions for the exploration and shallow wells were obtained from previous studies (Smelror et al., 1994; Jongepier et al., 1996; Mørk and Johnsen, 2005) and final well reports from the public Norwegian Petroleum Directorate database were implemented in this study.

3.2 Methodology

The tectono-stratigraphic analysis of the Slørebotn Sub-basin has been carried out by utilizing both well and seismic data. The chrono-stratigraphic framework presented in this study consists of nine seismic sequence boundaries (K0-K7) that defined nine sequences (S0-S8) (Fig. 3.2). The age framework was established by tying the key seismic sequence boundaries (SB) that were recognized in the seismic data to the wells. However, only six (K2-K7) of the nine SB could be interpreted in the seismic data (Fig. 3.3). The remaining three boundaries have been either locally interpreted where possible or postulated to exist (Fig. 1.2). Furthermore, the interpreted SB consists of both genetic and depositional surfaces due to their lateral continuity, which are bounded by faulting, lap relationships, erosional contacts or the study area boundary. Time-horizons were created based on the SB interpretation, which were later converted to depth by using a regional velocity model. It should be noted that the grid spacing of the model is 1x1km and only a limited number of wells could be used for quality checking of the depth converted surfaces. Nevertheless, there were no major deviations between the depth converted surfaces and the interpreted well tops.

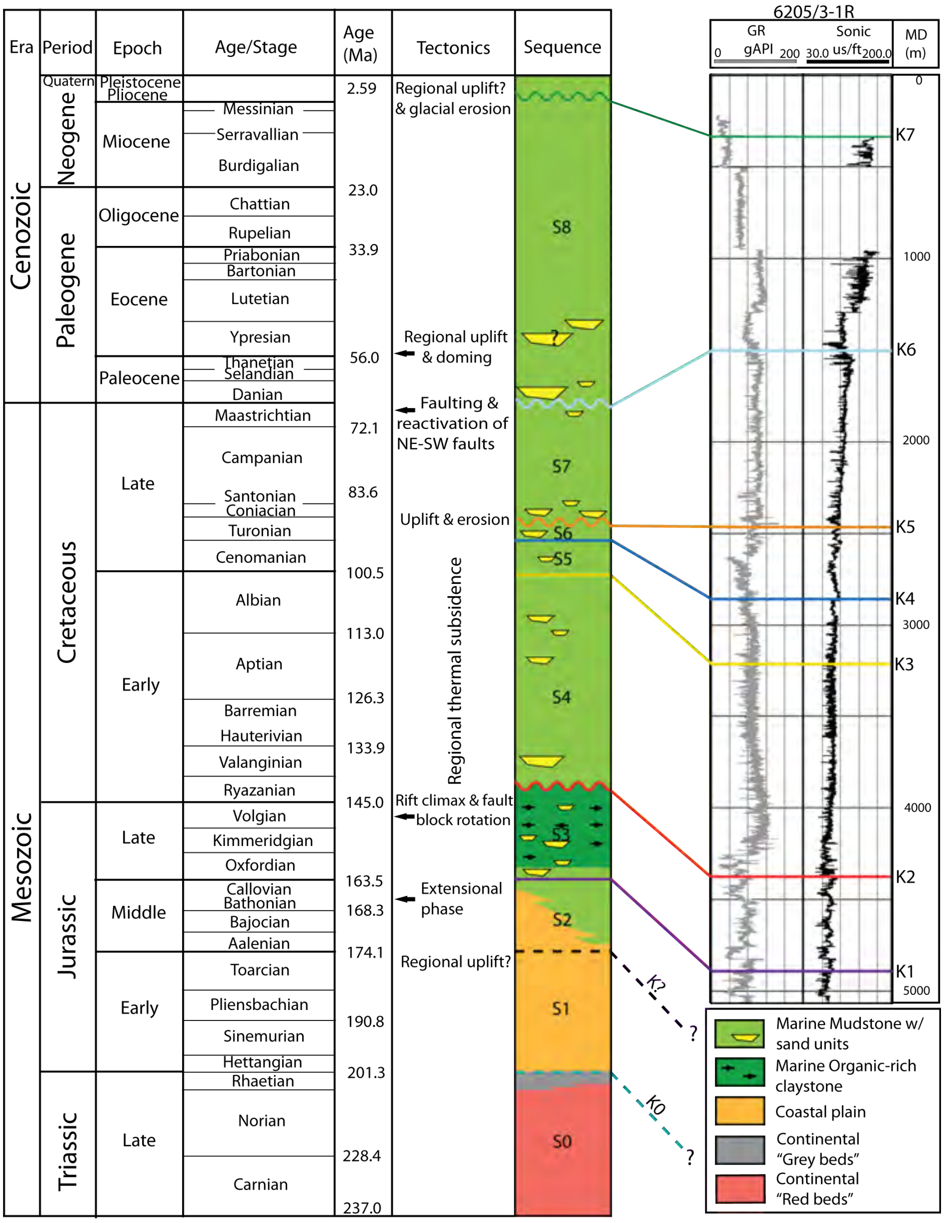


Fig. 3.2 Tectono-stratigraphic chart for the Slørebotn Sub-basin *Well 6205/3-1R is added to give a view of the general well log response for the seismic boundaries and sequences.*

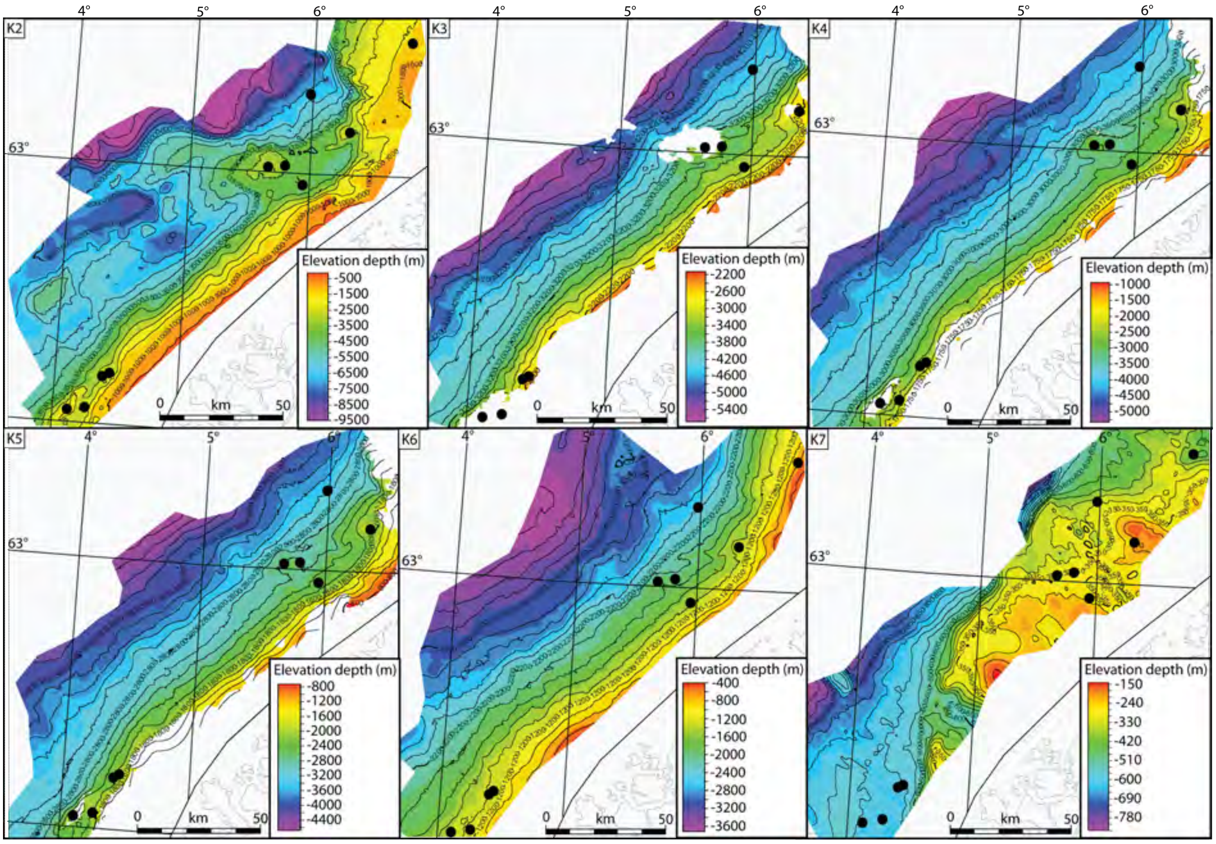


Fig. 3.3 Depth converted surfaces of the Mesozoic to Cenozoic sequence boundaries (K2-K7).

The defined sequences were used to generate six time-thickness maps (Fig. 3.4) that were further used to understand the stratigraphic extent and accommodation space created and filled through time. Out of the six sequences, five are regional (S4-S8) that extend throughout the study area and one is local (S3), located in the northern part of the Slørebotn Sub-basin.

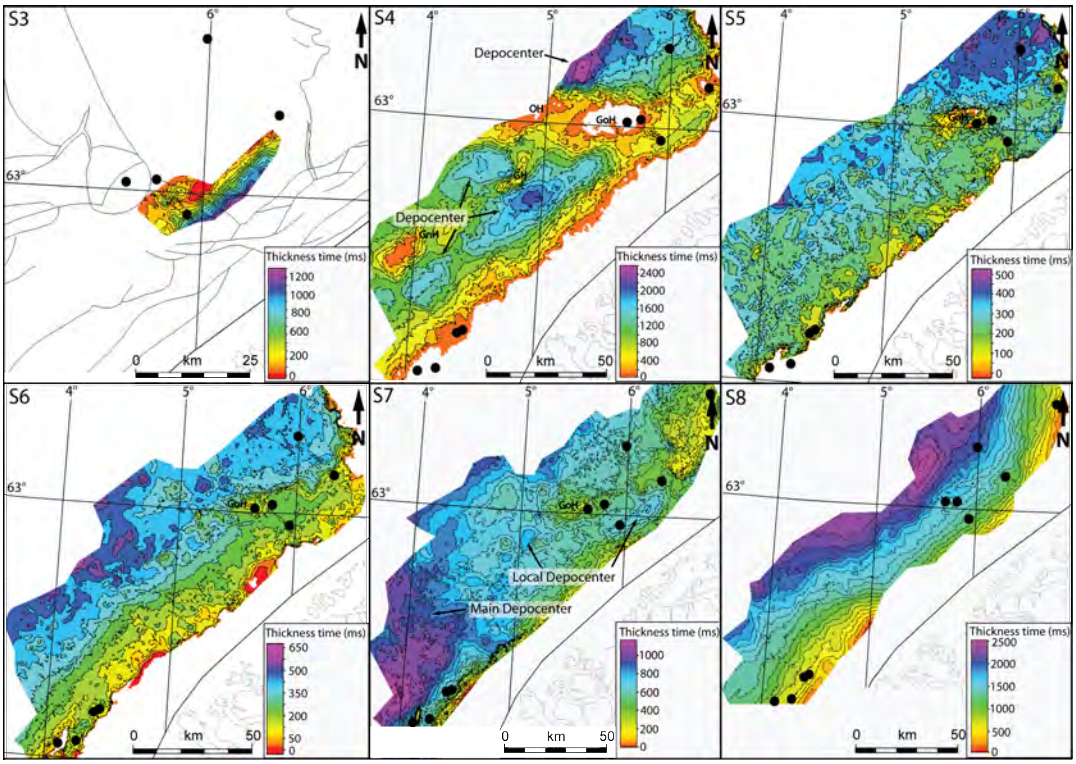


Fig. 3.4 Time-thickness maps of sequences S3 to S8. *Black dots represent well locations, see Fig. 1.1 for well labels. GoH = Gossa High, OH = Ona High.*

There are some uncertainties in the pre-Cretaceous stratigraphic analysis due to the few number of wells that encountered the pre-Cretaceous stratigraphic rocks in the study area (Fig. 3.5, Fig. 3.6 and Fig. 3.7). The structural complexity combined with areas of low seismic quality and great depths makes the interpretation difficult. The poorest seismic imaging with depth is observed in the middle and southernmost part of the Slørebotn Sub-basin (Fig. 3.8) (Appendix A), where little to no internal basement configuration is possible to interpret. Still, the seismic profiles are adequate to map the base Cretaceous unconformity and several younger stratigraphic reflections. Seismic artifacts, such as multiples, diffractions and noise are present in several of the seismic profiles (Appendix A) and if not correctly identified, may lead to incorrect interpretation.

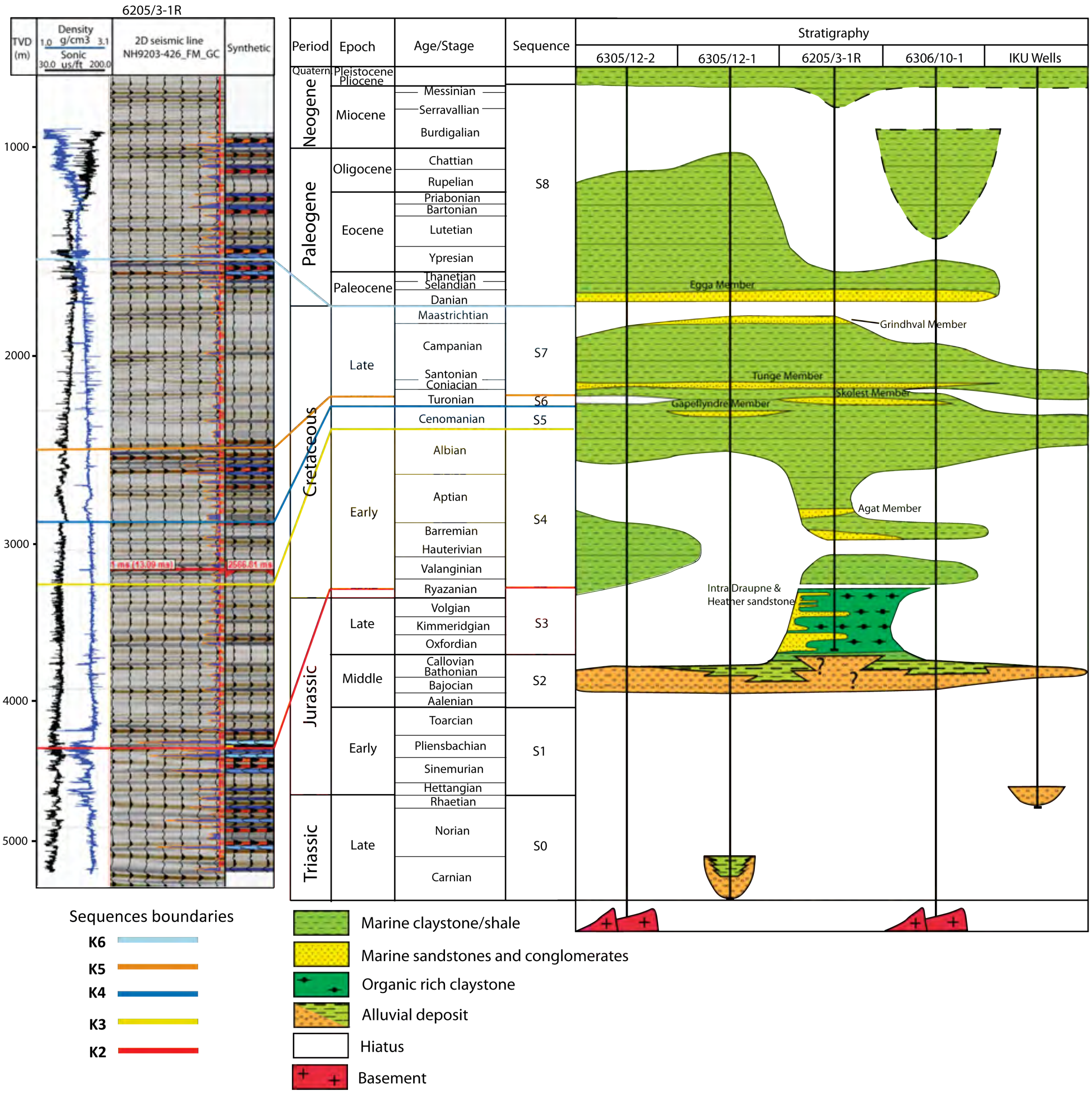


Fig. 3.5 Chrono-stratigraphic chart for the exploration wells in the Gossa High area *Well 6205/3-1R shows the well log and seismic response of the defined seismic sequence boundaries and sequences in northern part of the study area. A hard-kick is represented by a red colour. The chart (left) shows the interpreted lithology. The black dashed line in the upper part of the chart represents lithological information from the NPD factpages.*

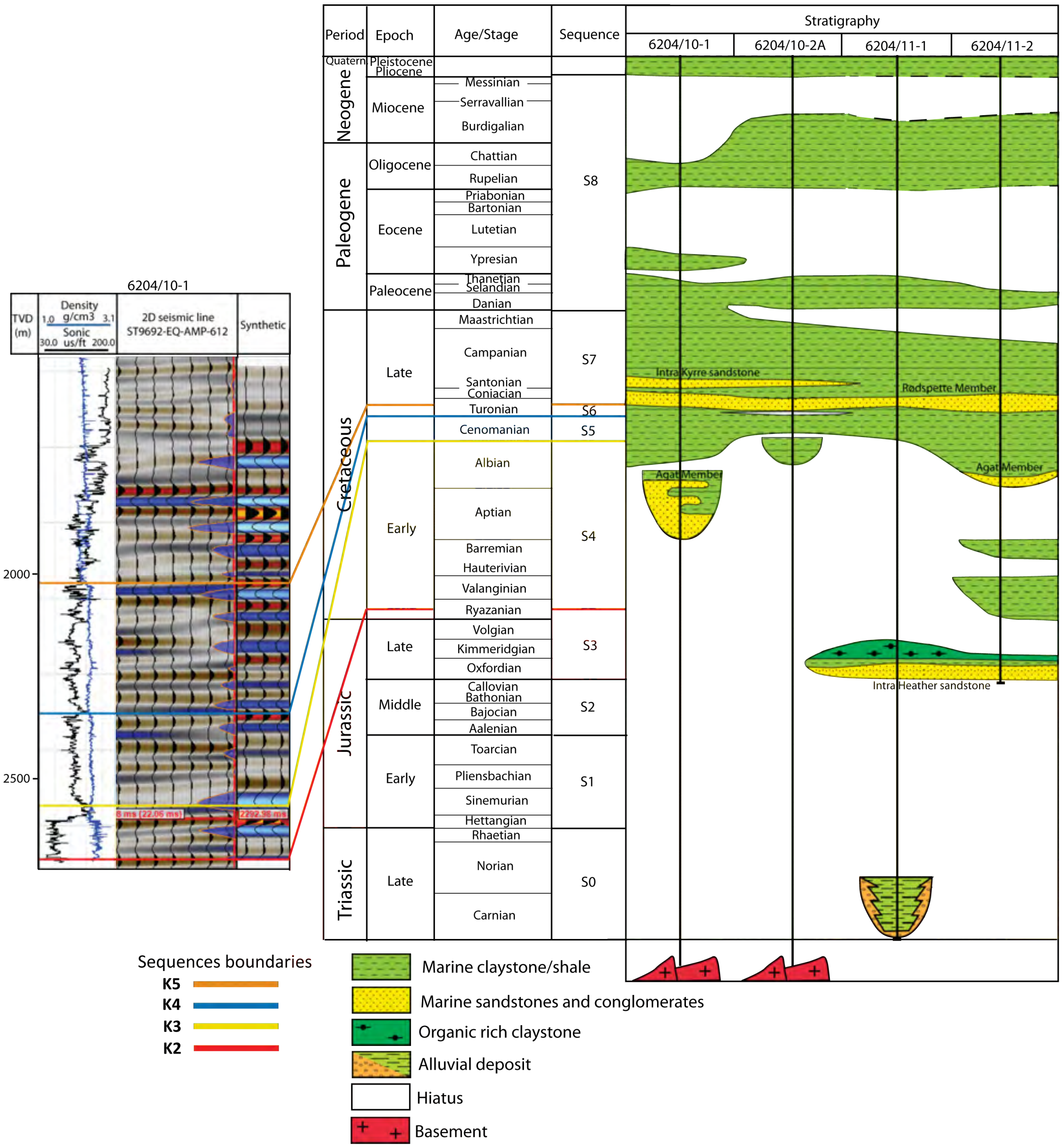


Fig. 3.6 Chrono-stratigraphic chart for the exploration wells in the southern part of the Slørebotn Sub-basin *Well 6204/10-1 shows the well log and seismic response of the defined seismic sequence boundaries and sequences in northern part of the study area. A hard-kick is represented by a blue colour. The chart (left) shows the interpreted lithology. The black dashed line in the upper part of the chart represents lithological information from the NPD factpages.*

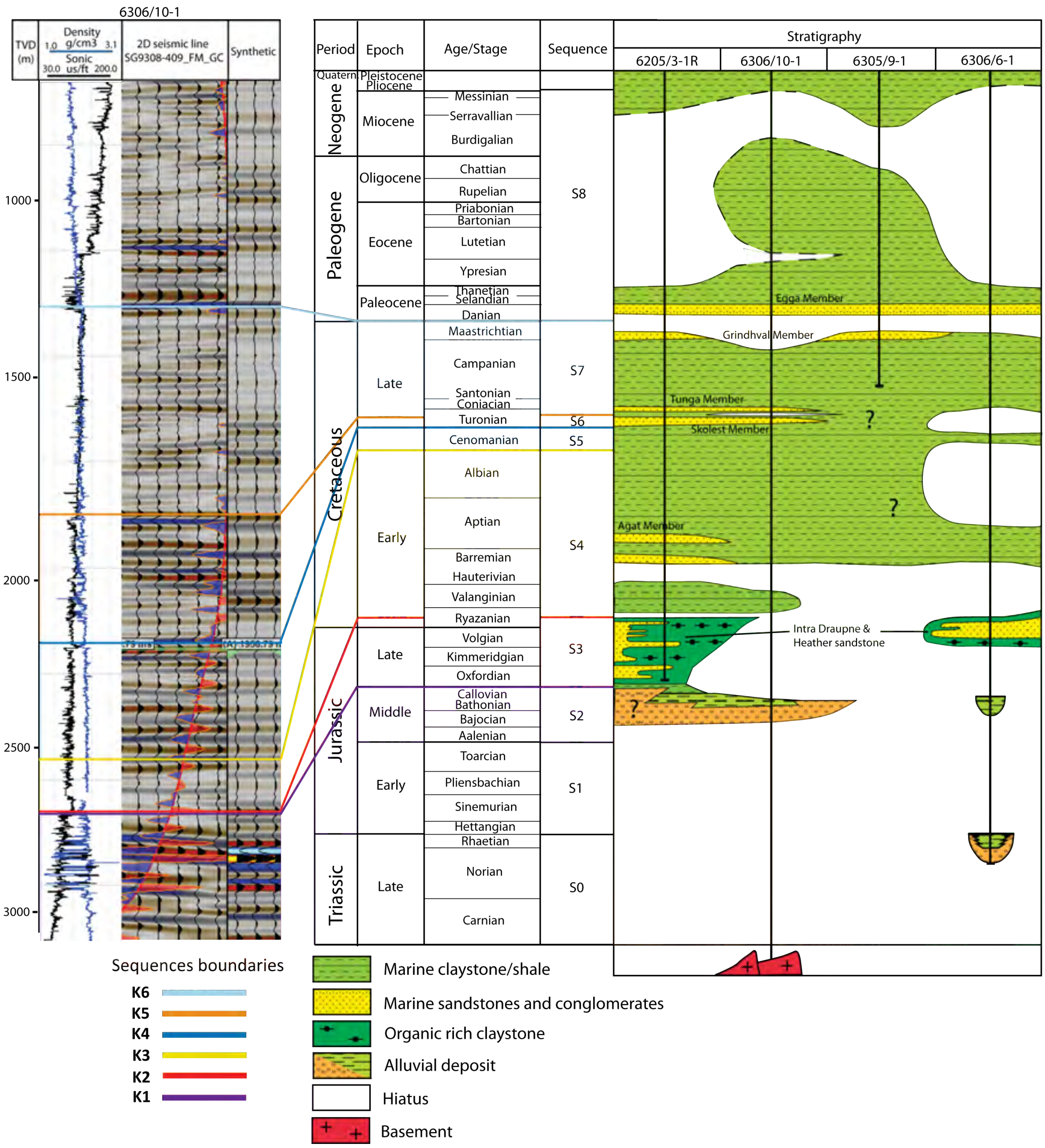


Fig. 3.7 Chrono-stratigraphic chart for the exploration wells in the northern part of the Slørebotn Sub-basin. *Well 6306/10-1 shows the well log and seismic response of the defined seismic sequence boundaries and sequences in northern part of the study area. A hard-kick is represented by a blue colour. The chart (left) shows the interpreted lithology. The black dashed line in the upper part of the chart represents lithological information from the NPD factpages.*

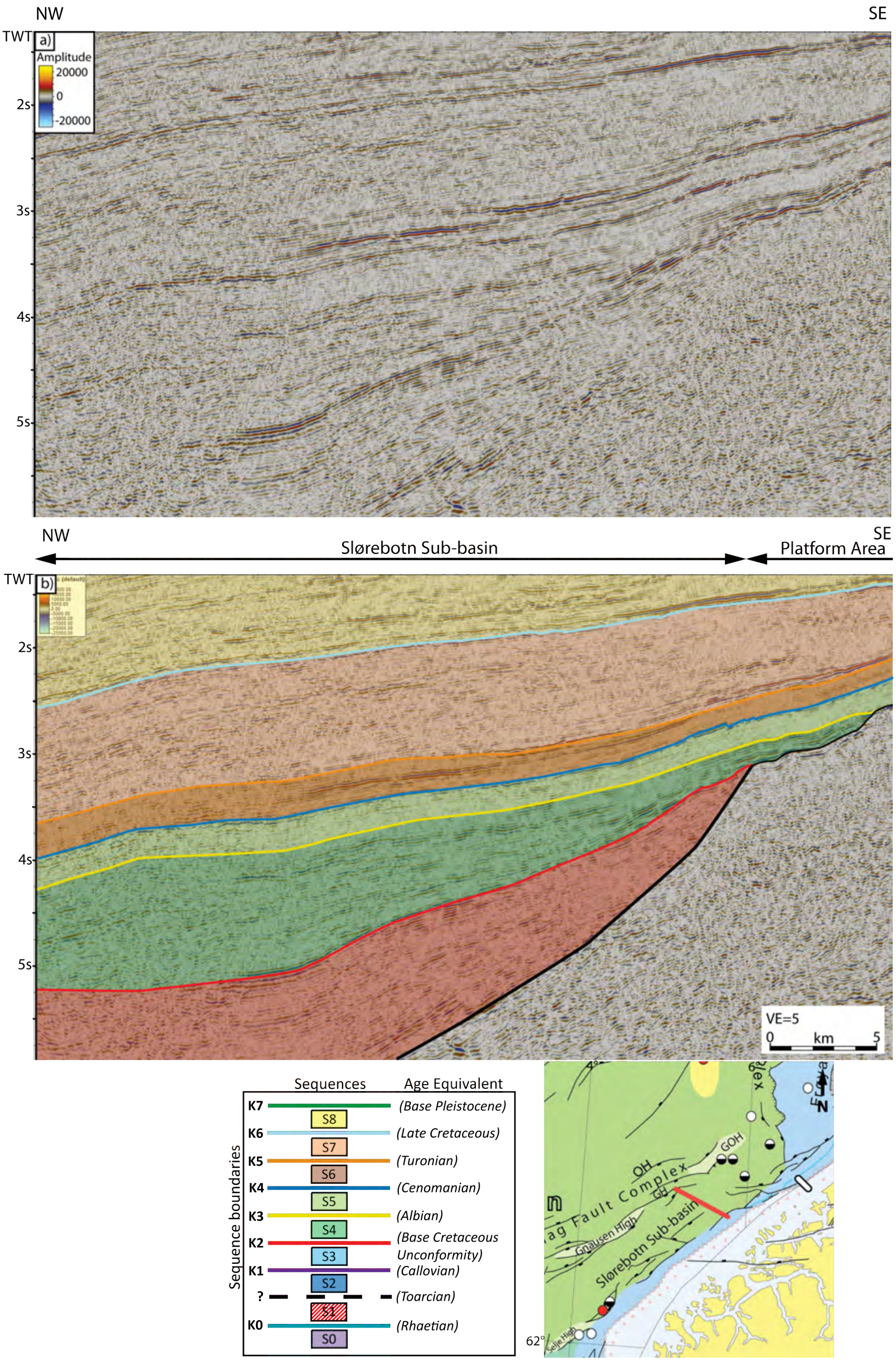


Fig. 3.8 2D seismic profile showing the general seismic imaging quality in the middle part of the Slørebotn Sub-basin *Note the heavy eroded footwall block.*

For this study, the gravity data has been used to confirm the location of basement blocks that were only partially visible in the 2D seismic profiles, as this tool is helpful for measuring lateral variation in rock densities; high-density anomalies tend to be related to basement rocks, while surface rocks such as clastics, carbonates etc. rarely exceed the density of the Earth's interior (Hinze et al., 2013).

The structural interpretation was based on recognizing abrupt reflector terminations in the seismic data, which was thereafter divided into fault families based on the strike and timing of the faults. Furthermore, the identification of onlap relationships and growth strata has been important to understand the fault movement, as several of the faults have been reactivated through different stages in time. The combination of the structural and stratigraphic interpretations led to the possibility of distinguishing between whether the accommodation created through time was controlled by tectonics or depositional factors. This provided the basis for a discussion, regarding the gross depositional environment and sedimentary infill trends in the study area. The comparison of the structural and stratigraphic evolution to other adjacent areas such as the Halten Terrace and the northern North Se(Fig. 1.1) has been an important piece in this study for evaluating the play potential in the study area. In more specific terms, by interpreting key SB that relate to proven discoveries in the adjacent areas (e.g. Cara discovery, Agat Fm; and Ormen Lange discovery, Egga Member (Fig. 1.1)) into the study area (Fig. 3.9) or by comparing successful sequences (e.g Fenja discovery, Rogn Fm) that were deposited during the same time period and in a similar depositional environment. For this task, one well (6406/11-1S) on the Halten Terrace in the north, one well on the Ormen Lange dome (6305/5-1) in the northwest, and well 35/1-1 in the northern North Sea have been used to establish the key relationships (Fig. 3.10).

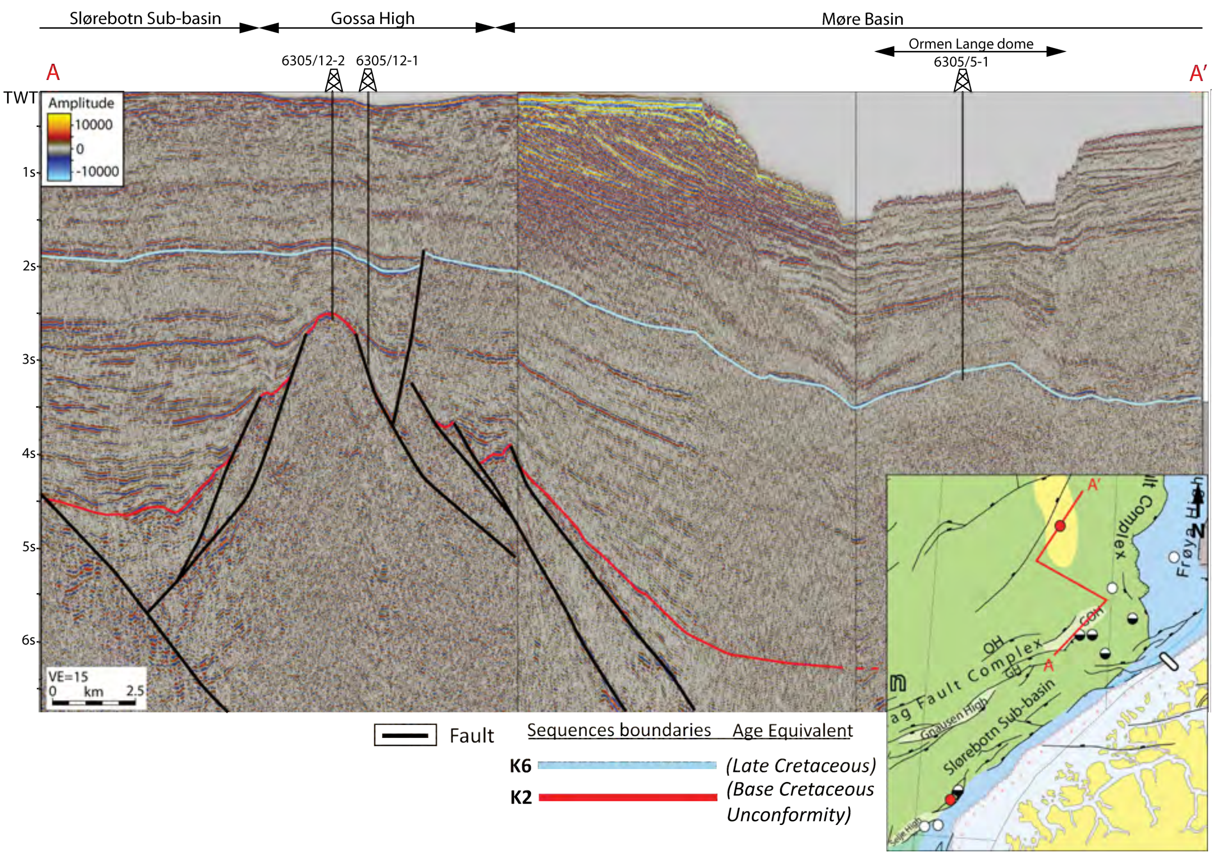


Fig. 3.9 Regional 2D profile that cross the Gossa High and the Ormen Lange dome.
Interpretation of what is believed to be the base (K6) of the main reservoir unit in the Ormen Lange field has been correlated to the study area wells.

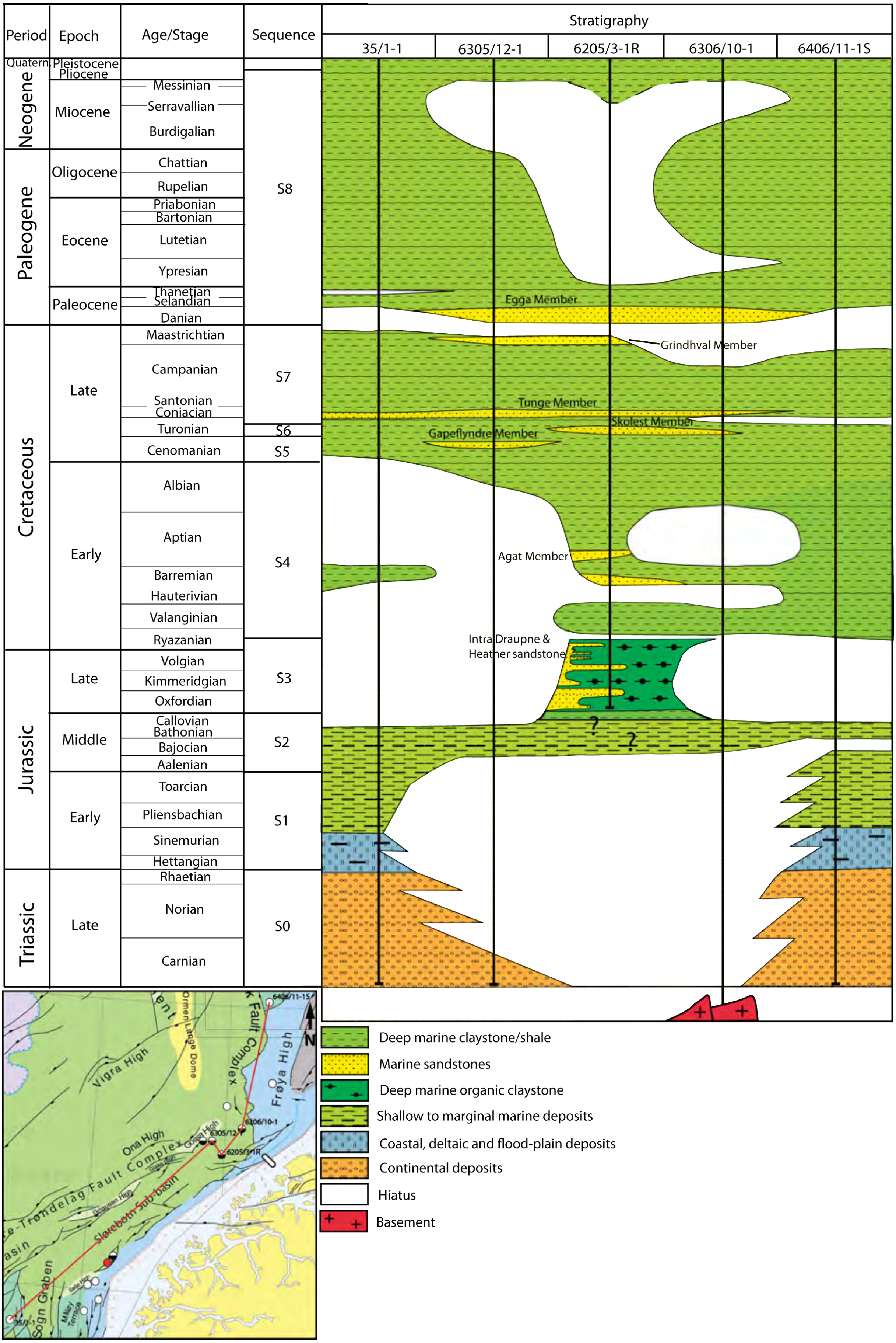


Fig. 3.10 Mega regional chronostratigraphic correlation from the Halten Terrace to the northern North Sea. *The black dashed line in the upper part of the chart represents lithological information from the NPD factpages.*

3.3 Seismic-well tie

In this study, six wells with checkshot data (Table 1.1) were used to generate the synthetic seismograms for the seismic-well ties. Well 6305/12-2 did not contain checkshot data and therefore, the checkshot data from 6305/12-1 was used as a starting point for this seismic-well tie. A zero-phase Ricker wavelet with a dominant frequency of 25 Hz was found to be a good fit for the seismic-well ties (Fig. 3.11).

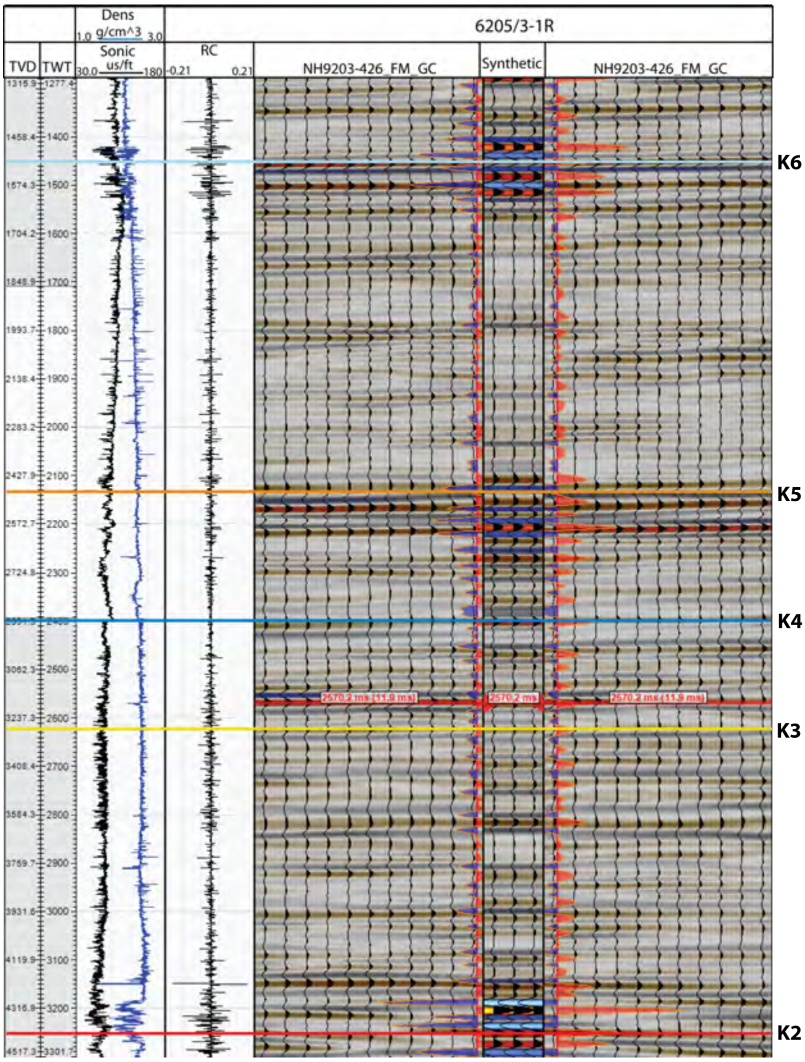


Fig. 3.11 A representative seismic-well tie displaying the match between the 2D seismic line NH9203-426 and the calculated synthetic *Zero-phase Ricker wavelet with a dominant frequency of 25 Hz was used for generating the synthetic seismogram for all six wells with checkshot data (Table 1). Red color represents a hard-kick.*

The well ties in a regional 2D interpretation study such as this are not as detailed as a field-scale interpretation project, but they are critical in order to establish an accurate age framework by tying key chrono-stratigraphic surfaces away from the areas without well control. One major challenge in this study, apart from the poor seismic imaging, has been the sparse well control over such a large area.

4 RESULTS AND OBSERVATIONS

4.1 Age framework

This paragraph will give a short summary of the relative ages of the seismic sequences that are been defined in this study and the chart in Fig. 3.2 gives a quick overview of the different sequences and their associated boundaries. The following description of the sequences will range from oldest in age to the youngest:

Sequence 0

The oldest sedimentary rocks that have been encountered are of Carnian to Norian in age, and the rocks have been assigned to sequence 0. The sequence is bounded at the top by K0, which relates to the Rhaetian, whereas the lower boundary has not been penetrated since none of the wells drilled deeper than rocks of Carnian in age.

Sequence 1

Sequence 1 corresponds to the Early Jurassic (Hettangian-Toarcian) sedimentary rocks, bounded at its basal part by K0 and top by K?. The latter has been assigned to the Toarcian, however rocks of Toarcian to Sinemurian in age have not been penetrated (Fig. 3.5, Fig. 3.6 and Fig. 3.7).

Sequence 2

Sequence 2 has been assigned to the Middle Jurassic (Aalenian-Callovian) and the unproven K? bound the sequence at the base, whereas K1 defines the top and it is characterized in the seismic data as a hard-kick (Fig. 3.7).

Sequence 3

Sequence 3 correlates to the Late Jurassic and earliest Cretaceous (Oxfordian-Early Ryazanian). The sequence is bounded below by K1 and above by K2. The latter represents a distinct hard-kick seismic reflector (Fig. 3.5 and Fig. 3.7).

Sequence 4

Sequence 4 has been assigned to the Early Cretaceous (Ryazanian-Albian) and the sequence is bounded at the base by K2 and top by K3. The latter represents a hard-kick in the seismic data (Fig. 3.5).

Sequence 5

Sequence 5 has been assigned to the early Late Cretaceous (Cenomanian) and the sequence is bounded at the base by K3 and top by K4. The latter represents a distinct hard-kick seismic reflector (Fig. 3.5, Fig. 3.6 and Fig. 3.7).

Sequence 6

Sequence 6 has been assigned to the Late Cretaceous (Turonian) and the sequence is bounded at the base by K4 and top by K5. The latter represents a soft-kick seismic reflector (Fig. 3.5, Fig. 3.6 and Fig. 3.7).

Sequence 7

Sequence 7 has been assigned to the Late Cretaceous (Late Turonian-Maastrichtian) and the sequence is bounded at the base by K5 and top by K6. The latter represents a distinct hard-kick seismic reflector (Fig. 3.5 and Fig. 3.7).

Sequence 8

Finally, the youngest sequence, sequence 8 has been assigned to the Cenozoic (Paleocene-Early Pleistocene) and the sequence is bounded at the base by K6 and top by K7. The latter represents a distinct hard-kick seismic reflector (Fig. 3.7).

4.2 Structural framework

The structural interpretation in this study led to identifying four main fault families and one prominent NW-SE trending fault. The faults are differentiated based on their timing and strike. It should be noted that only the main faults have been interpreted and that fault interpretation based on 2D seismic profiles will not lead to the precise location of the fault tips or potential fault splays.

4.2.1 Fault family 1 (FF1)

FF1 is characterized by both shallow and deep NE-SW trending normal faults that typically define the structural highs in the western part of the study area (Fig. 1.2, Fig. 4.1a, Fig. 4.2 and Fig. 4.3). FF1 intersects the K2 boundary and reflectors of S4 truncate against the fault planes of FF1, which implies that the faults were active prior to deposition of S4. It is suggested that FF1 is related to the Møre-Trøndelag Fault Complex that originated in the Caledonian orogeny and was reactivated during several tectonic episodes throughout the Mesozoic (Blystad et al., 1995; Grunnaleite and Gabrielsen, 1995; Doré et al., 1997b; Brekke, 2000).

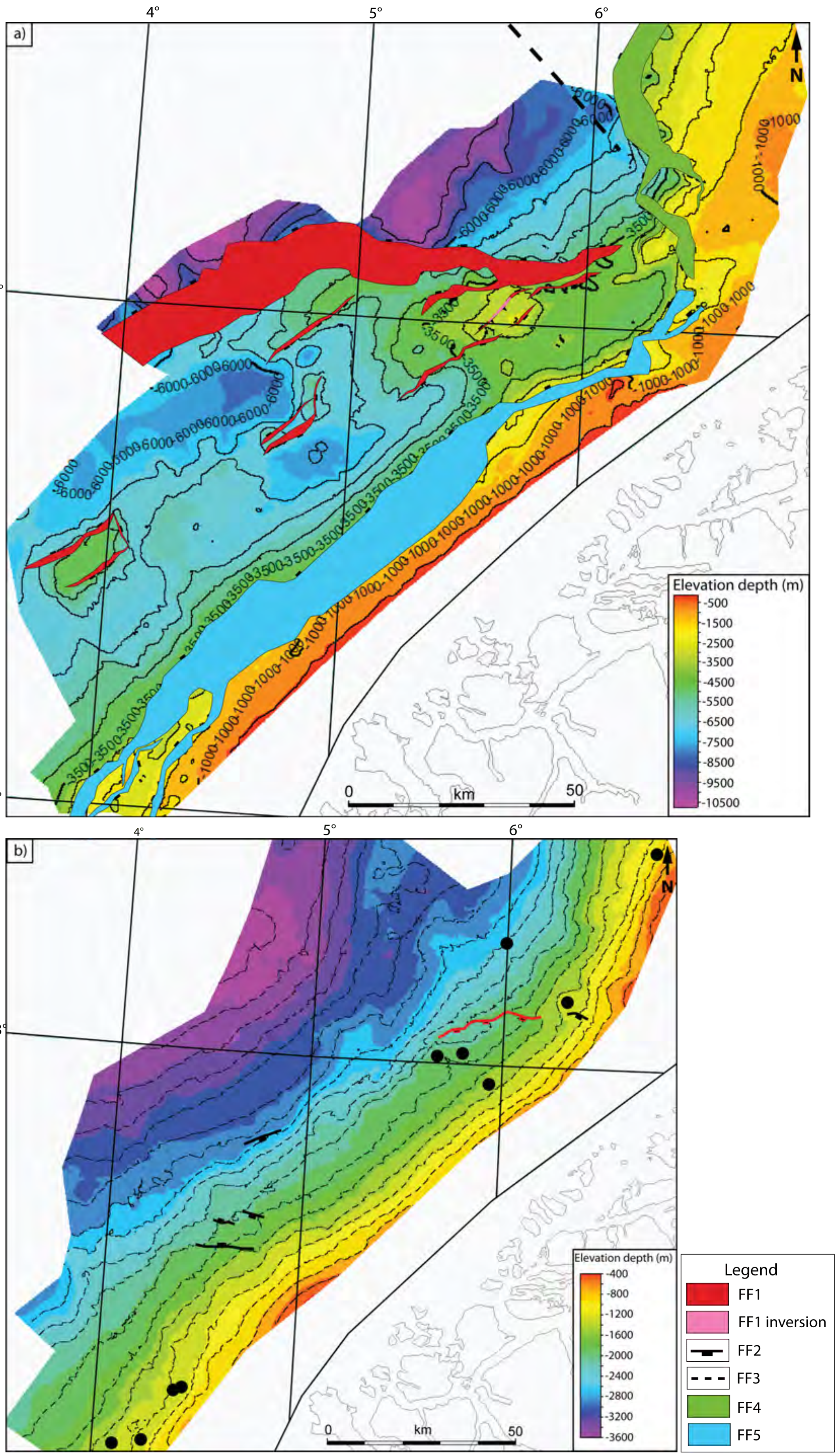


Fig. 4.1 Fault families *a) depth map of K2 surface displaying the intersecting fault families (FF1, FF3, FF4 and FF5); and b) depth map of the K6 surface displaying the intersecting FF2 and reactivated FF1.*

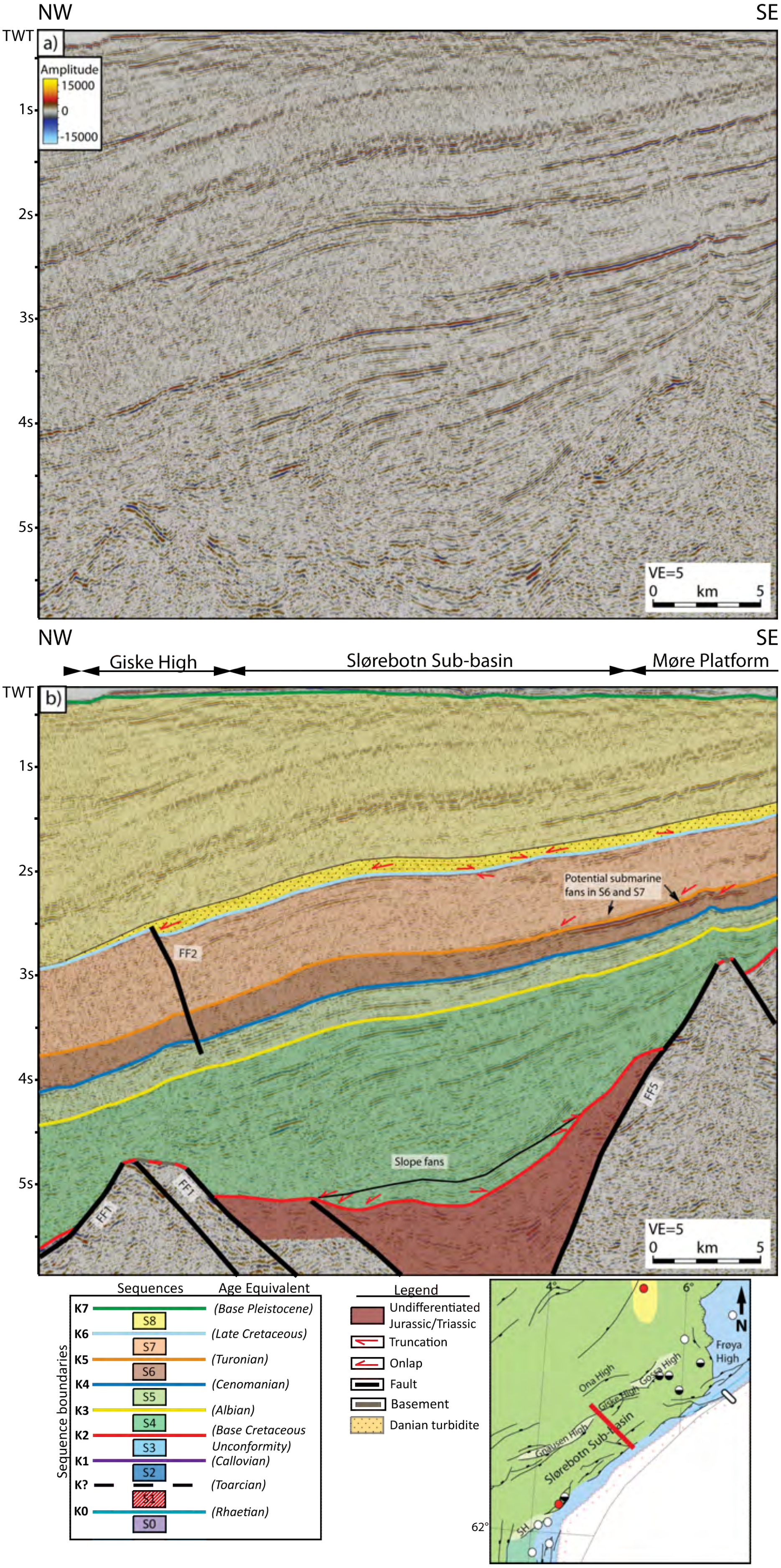


Fig. 4.2 Seismic profile crossing the Slørebotn Sub-basin and Giske High. *a) uninterpreted; and b) interpreted seismic profile. Note the inverted basin fill of the Slørebotn Sub-basin.*

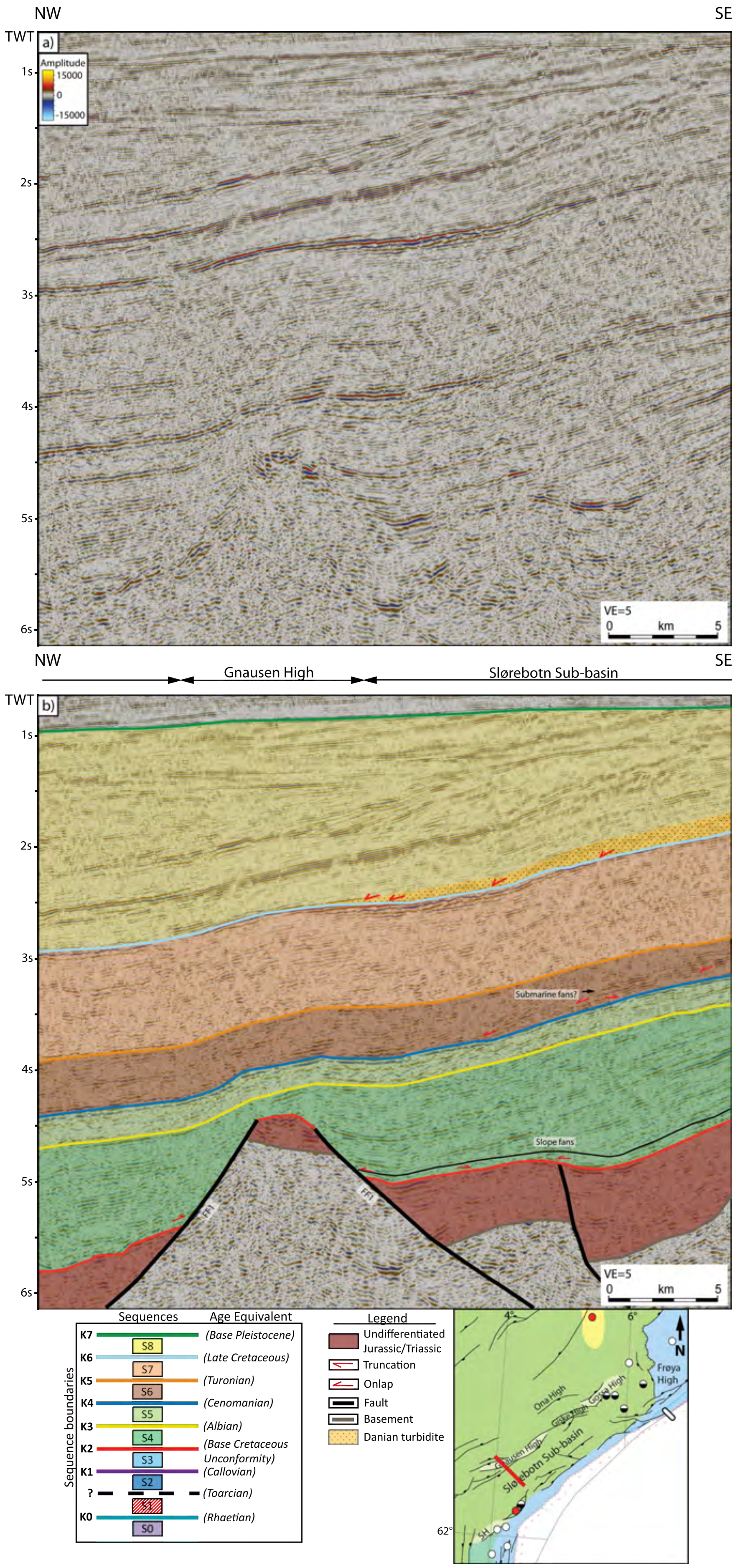
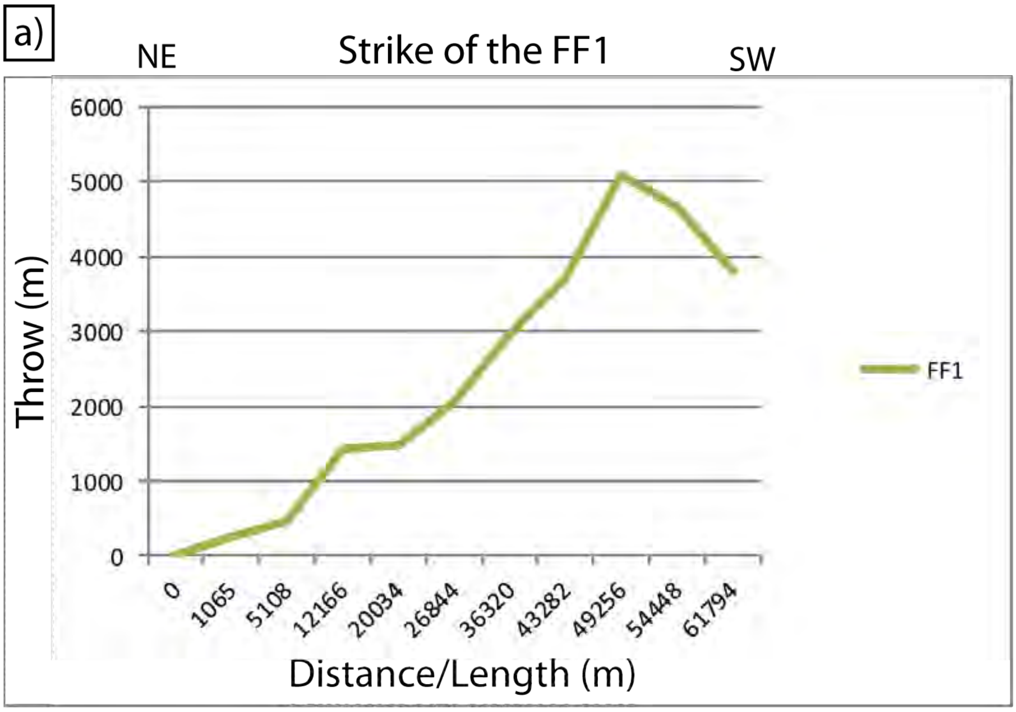


Fig. 4.3 Seismic profile crossing the Gnausen High. *a) uninterpreted; and b) interpreted seismic profile. Note the downlap of the lower S8 (Danian) reflectors above the apparently uplifted high.*

The most prominent faults of FF1 is the normal fault that borders the Gossa High in the west (Fig. 1.2). The prominent normal fault shows a slight change in the strike from the general NE-SW to ENE-WSW (Fig. 4.1a). The largest throw across this fault has been estimated to 5000 meters and in a distance of 50 km the throw decreases to zero towards the Frøya High in a ratio of 10:1 (Fig. 4.4a). In the southwest, the fault throw analysis is limited by the seismic coverage and imaging quality, but the throw of the fault is assumed to taper off because of the decreasing trend to the southwest (Fig. 4.4a). It should be noted that the K2 surface that was used for the throw analysis has been exposed to erosion during several time periods; therefore the analysis provides a minimum estimate.



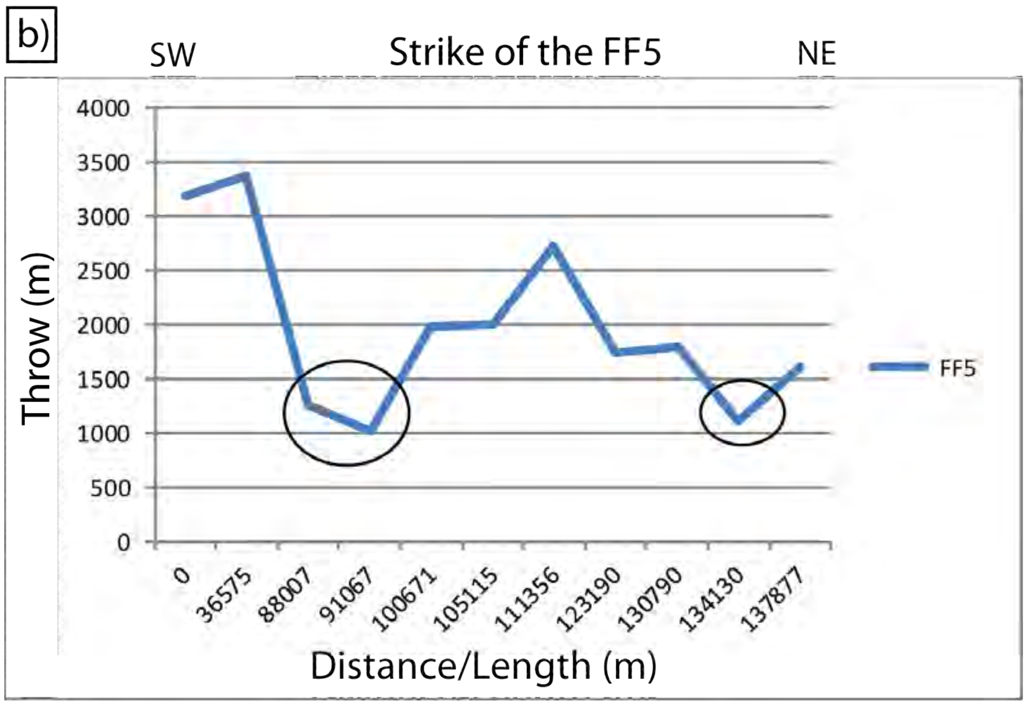


***Fig. 4.4 Throw plots for FF1 and FF5** a) throw estimation of the major normal fault that borders the Gossa High in the west; and b) throw estimation for FF5 fault complex, which borders the Møre platform in the east. The black circles represent location of potential linkage of the fault segments.*

Several of the faults assigned to FF1 shows evidence of later reactivation in the sense of both normal and reverse displacement. A particular inverted fault, which intersects the Gossa High, shows a drastic change in its displacement along strike going from normal to reverse to normal faulting again (Fig. 4.1a). The reverse reactivation caused uplift of the overlaying S6-S8 (Fig. 1.2). The reactivation of FF1 in the normal sense can be observed to intersect K6 (Fig. 4.5).

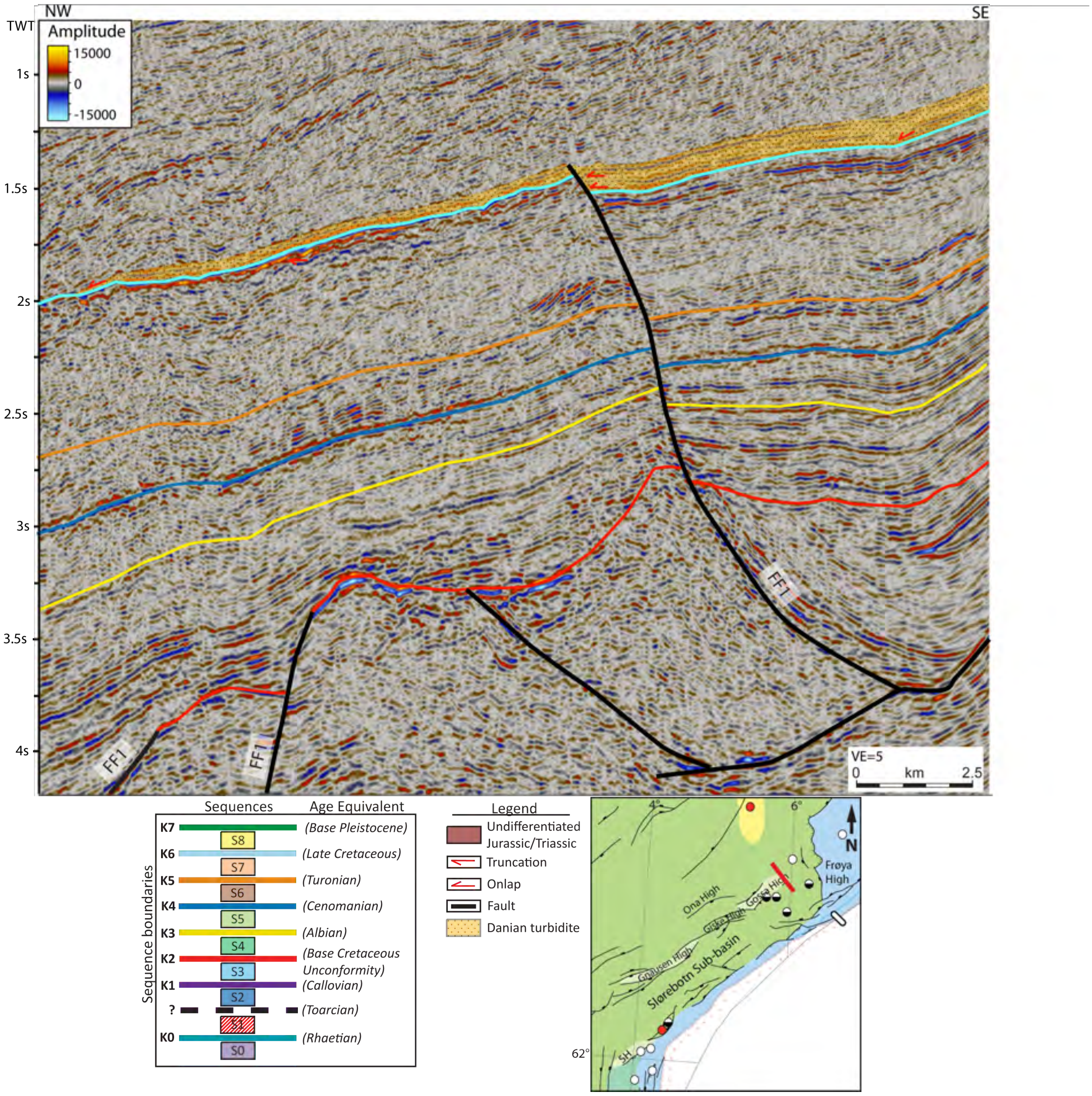


Fig. 4.5 Siesmic profile crossing the Gossa High. *Reactivation of blind faults assigned to FF1 during the Late Maastrichtian. Note the truncation of the lower S8 (Danian) against the fault plane.*

4.2.2 Fault family 2 (FF2)

FF2 represents E-W trending normal faults (Fig. 4.1b) that are typically characterized by steep dipping normal faults that intersect the surface of K6 (Fig. 4.2). Furthermore, the seismic reflectors in the lower portion of S8 can be observed to truncate against the fault planes of FF2, which suggests that the faults were active prior to deposition of S8, possibly in the Late Maastrichtian time (Fig. 4.2).

4.2.3 Fault 3 (F3)

F3 represents a NW-SE trending fault that is interpreted from the gravity data to offset large positive anomaly belts (Fig. 1.3). Previous authors (Talwani and Eldholm, 1972; Blystad et al., 1995; Brekke, 2000; Eldholm et al., 2002) interpreted F3 as an old Paleozoic transfer fault (Jan Mayen Lineament) that separates the Møre Basin from the Vøring Basin in the north (Fig. 1.1). The lineament is believed to represent a polarity change in the initial asymmetric rift basins between Norway and Greenland, with a change from upper- to lower-plate across the lineament, from the narrow Møre margin to the wider Halten Terrace, respectively (Torkse and Prestvik, 1991; Gjelberg et al., 2005). F3 is not obvious in the seismic data (Fig. 4.6); however, in a regional prospective, the presence of a transfer fault is compelling as major fault complexes (FF5) terminate against the intersection point of the lineament (Fig. 4.1a).

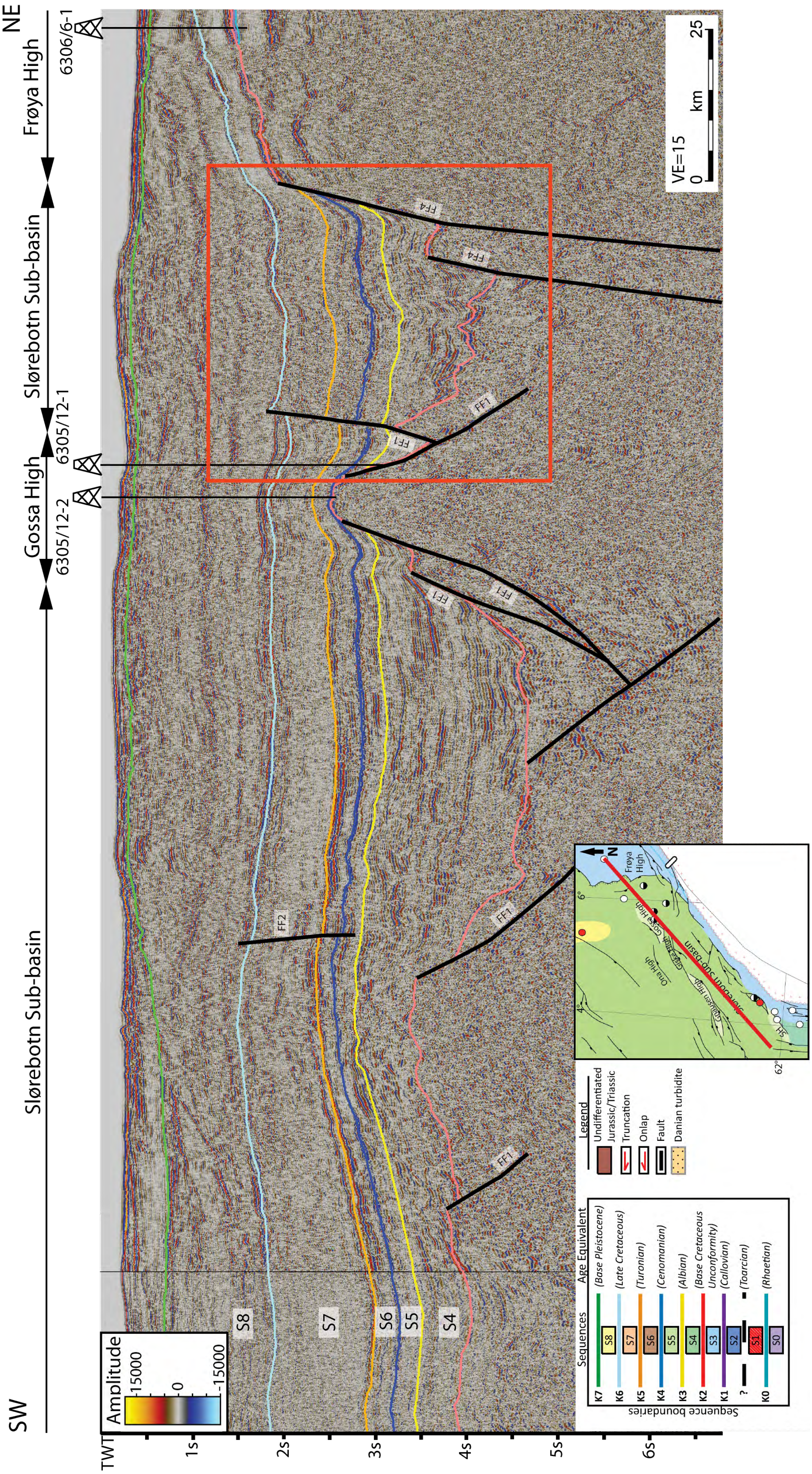


Fig. 4.6 Regional NE-SW seismic line that crosses the Frøya High, Gossa High and Slørebotn Sub-basin. *The red square represents the possible location of where F3 intersects the study area.*

4.2.4 Fault family 4 (FF4)

FF4 has been assigned to steep (67.5°) dipping NNW-SSE trending normal faults that are interpreted to border the Frøya High in the south, which then separates the high from the deeper Slørebotn Sub-basin (Fig. 4.1a and Fig. 4.6). These steep SW dipping normal faults have been interpreted to be part of the Klakk Fault Complex (Blystad et al., 1995). Furthermore, it has been suggested that the southern part of the fault complex is related to the Jan Mayen Lineament due to the abrupt change in the strike of the fault complex, which shares the similar trend as the lineament (Brekke, 2000) (Fig. 1.1 and Fig. 4.1a).

FF4 is believed to link against the faults of FF5 in the northeastern part of the Slørebotn Sub-basin (Fig. 4.1a), but because of the limited seismic coverage in that particular location, the exact linkage point was not possible to interpret. The highest estimated throw across FF4 is measured to 3800 meters at the K2 level in the northern part.

The large throw of FF4 is reflected in the seismic data where S4-S7 reflectors can be observed to truncate against the fault plane (Fig. 4.7). Hence, FF4 has played a major role in the depositional development during the Mesozoic time.

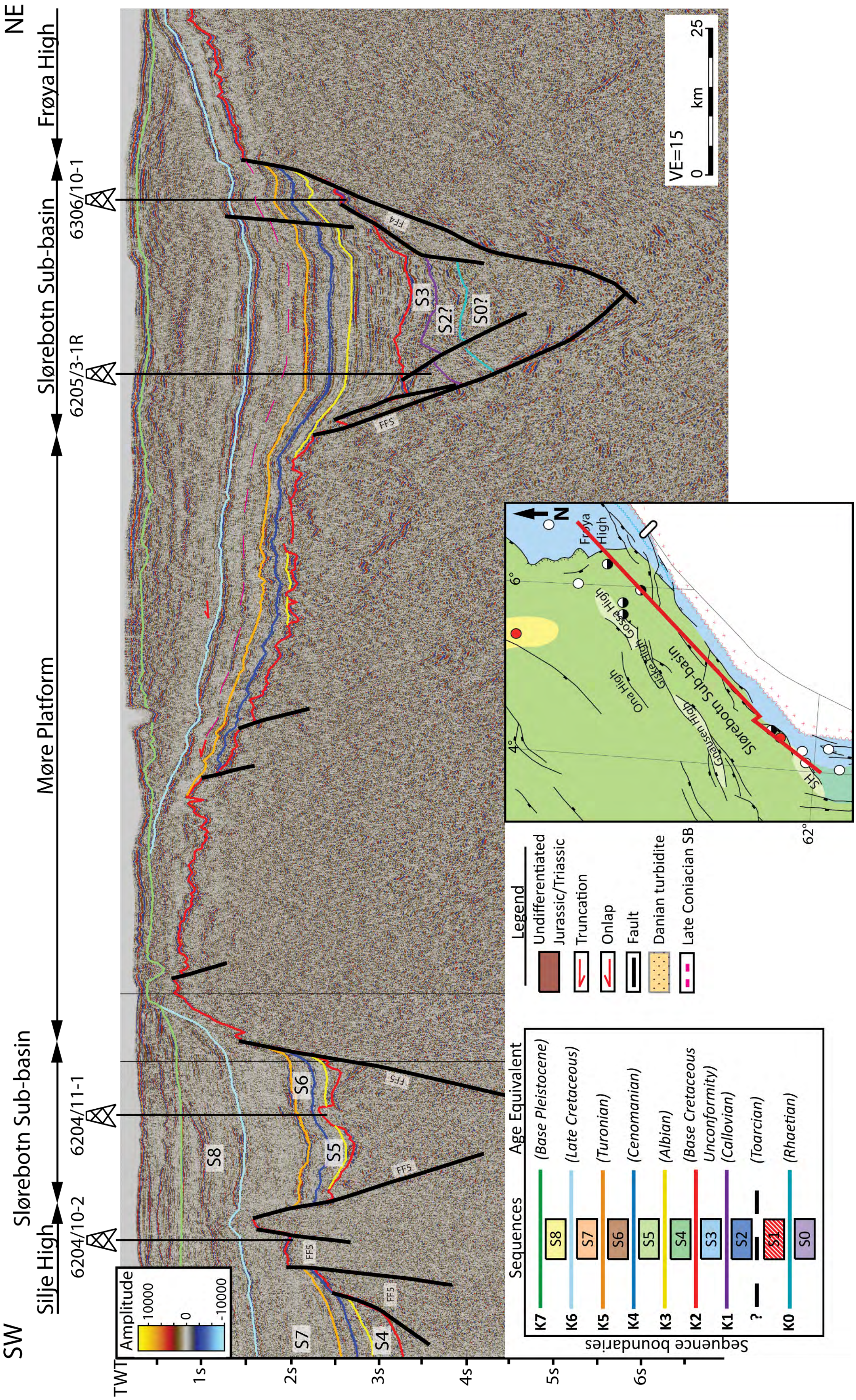


Fig. 4.7 Regional NE-SW seismic line crossing the Frøya High, Slørebotn Sub-basin, Møre platform and Silje High. *Note the internal seismic sequence in S7 that onlaps against the Møre platform.*

4.2.5 Fault family 5 (FF5)

The faults assigned to FF5 are located along the entire eastern margin of the Møre platform and the faults separates the shallow platform from the deeper Slørebotn Sub-basin in the west (Fig. 4.8). It is suggested that FF5 is part of the Møre-Trøndelag Fault Complex that originated in the Caledonian orogeny and was reactivated during several tectonic episodes throughout the Mesozoic (Blystad et al., 1995; Grunnaleite and Gabrielsen, 1995; Doré et al., 1997b; Brekke, 2000).

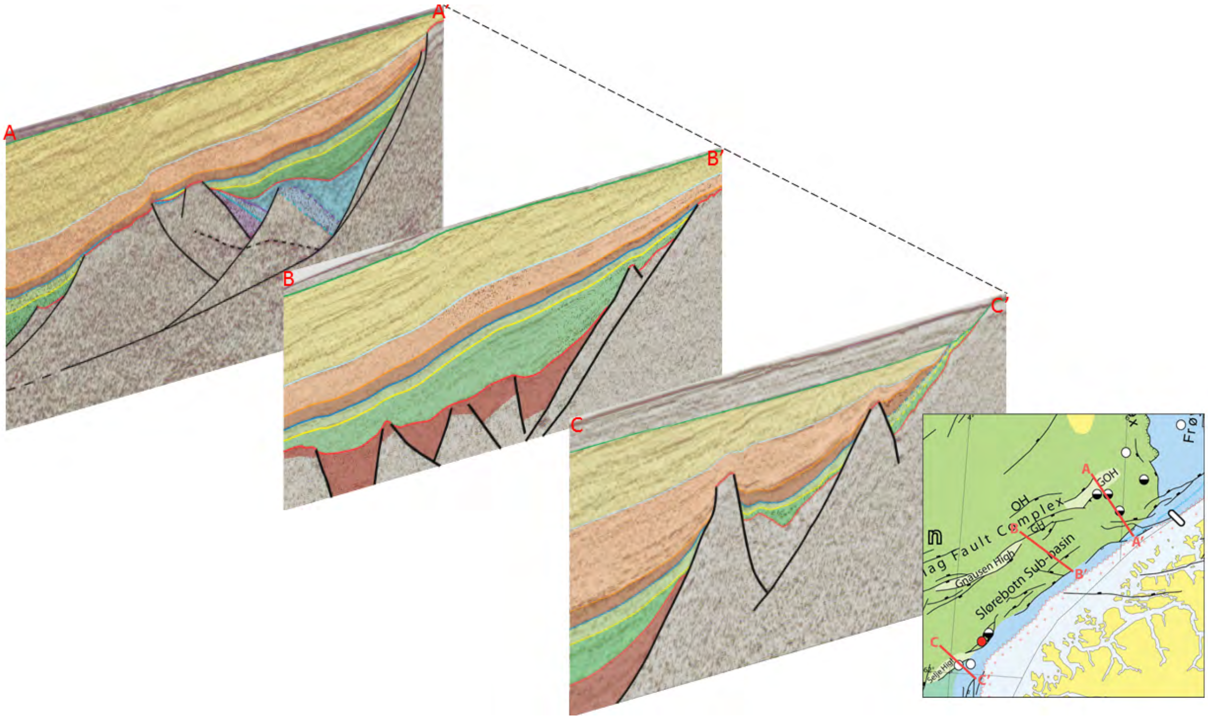


Fig. 4.8 Three NW-SE striking seismic profiles that show the general architecture of the sub-basin.

FF5 comprise a series of NE-SW listric faults with a throw that varies from 3400 to 1100 meters (Fig. 4.4b). The large variation observed in the throw plot suggests that the main faults of FF5 initially developed in the northern and southern part of the study area (Fig. 4.9) and through several stages of continued rifting the faults tips of FF5 are eventually linked. This observation is also reflected in the fault architecture of FF5, where the dip can be seen to vary along the strike, which implies that FF5 comprise a series of normal faults (Fig. 4.8). The dip of FF5 in the north is measured to approximately 59°, in the middle to 54° and 58° in the southern part.

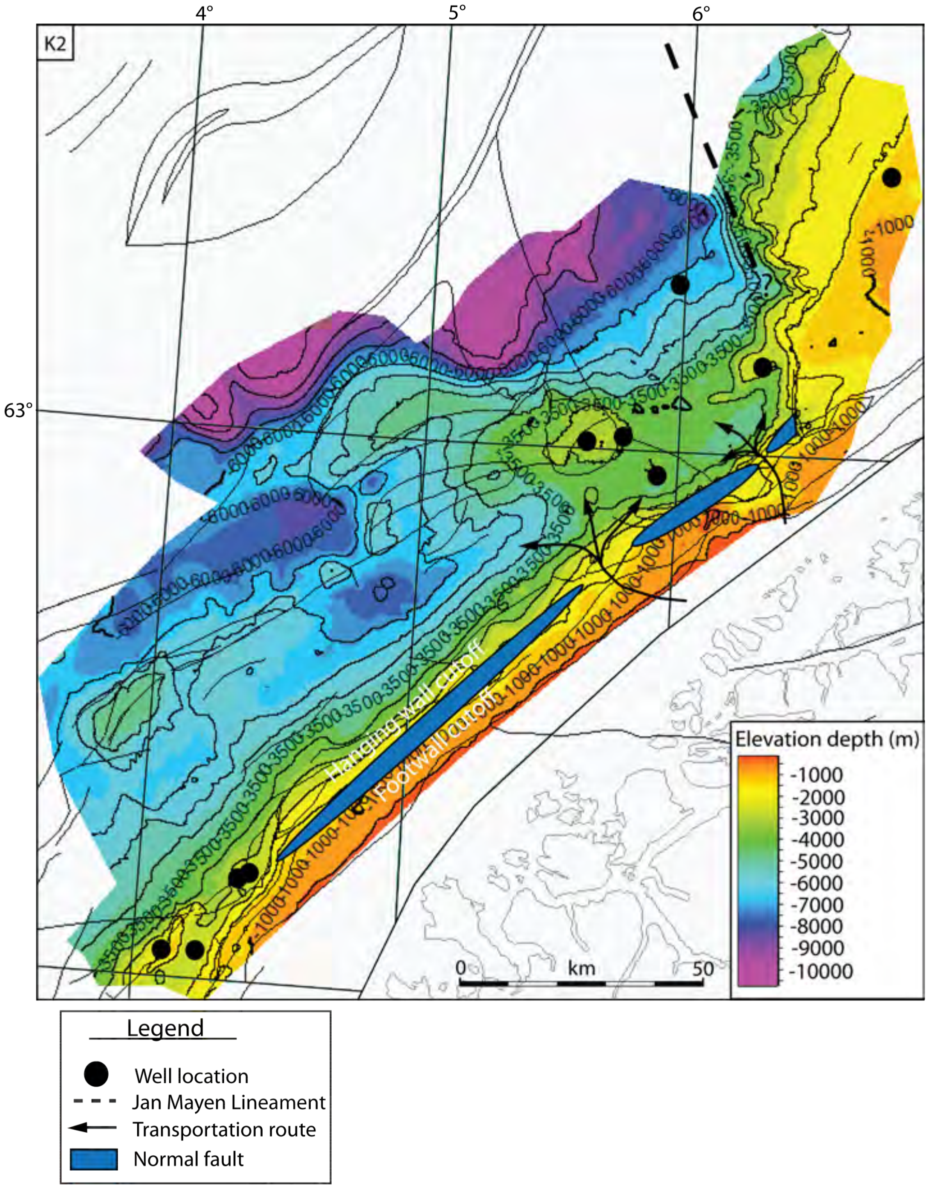


Fig. 4.9 K2 depth map with interpreted early fault location of FF5 and potential sediment routes.

FF5 is interpreted to link against F3 and FF4 in the north (Blystad et al., 1995), but due to the poor seismic coverage, the exact linkage point was not possible to pinpoint. Regarding the influence of FF5 on the study area, it is believed that the fault complex was fundamental in the development of the basin architecture and sedimentation through the Mesozoic and Cenozoic (Brekke, 2000). This is reflected in the seismic data as the entire S1-S6 and partly the S7 and S8 seismic reflectors truncate against FF5 (Fig. 1.2 and Fig. 4.7).

It should be noted that FF5 consists of several faults that split and merge along the eastern Møre margin, but the exact linkage points were not possible to interpret due to the poor 2D seismic coverage on the eastern margin area. Thus, the entire set of the fault complex has been interpreted as one major normal fault. Furthermore, the throw estimation for the middle and southern part of the study area was found to be challenging due to heavy erosion of the footwall block (Fig. 3.8).

4.3 Seismic sequences description

The following description for each seismic sequence ranges from oldest to youngest and each sequence is divided into three sub-chapters, i.e. well character, seismic character and interpretation. Although a brief interpretation for each sequence will be provided in terms of the depositional environment, tectonic evolution and comparison to the Halten Terrace and northern North Sea. These interpretations will be discussed in more detail in 5 Discussion.

4.3.1 Sequence 0 (Carnian-Rhaetian)

Well character

The upper K0 of sequence 0 (S0) was encountered by wells on the eastern flank of the Gossa High (6305/12-1), southeastern flank of the Slørebotn Sub-basin (6204/11-1) and on the inner part of the Frøya High (6306/6-1) (Fig. 3.5, Fig. 3.6 and Fig. 3.7). The base of S0 was not penetrated since none of the wells drilled deeper than Carnian rocks. The thickest S0 succession was penetrated by well 6305/12-1, which encountered more than 250 meters of S0 sedimentary rocks (Fig. 4.10 and Fig. 4.11) and core data shows that the sequence comprises conglomerates, pebbly sandstones, siltstones and mudstones that were deposited in an arid alluvial environment (Jongepier et al., 1996). The shallow IKU wells cored the uppermost part of S0 (Rhaetian) and the section comprises conglomerates and sandstones (Smelror et al., 1994). The mineral content in the cored rocks for both regions show traces of metamorphic greenstone and volcanic felsic to intermediate rocks, which are similar to the basement rocks that are recoded at the Gossa High and onshore northwest of the Møre-Trøndelag Fault Complex, respectively (Mørk and Johnsen, 2005). The cored conglomerates and sandstones on the Gossa High and Møre platform area show no reservoir quality due to the high mud content, immature texture and mineralogy of the rocks (Fagerland, 1990; Jongepier et al., 1996; Mørk and Johnsen, 2005).

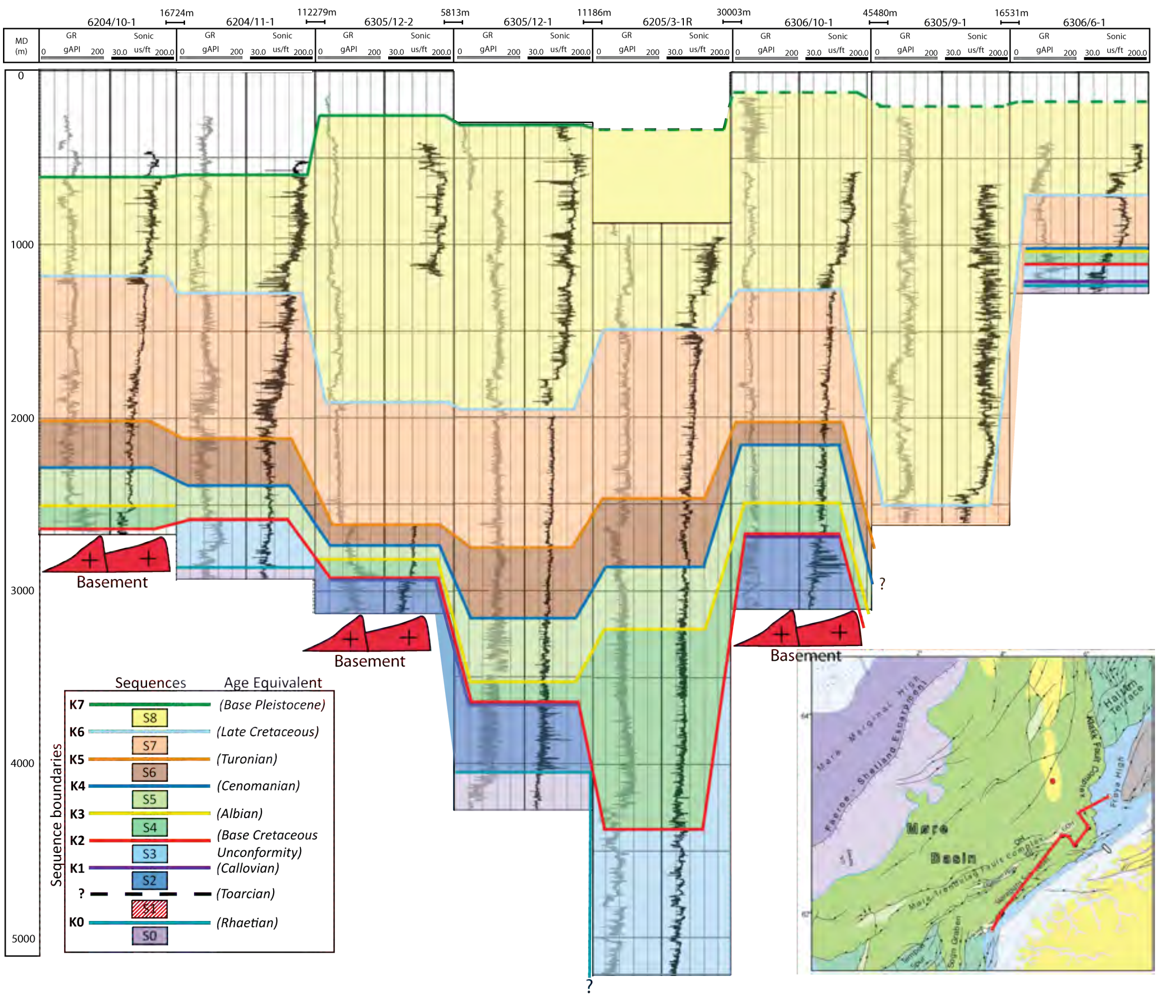


Fig. 4.10 Fig. 4.10 Well correlations between the exploration wells that contained checkshot data

The upper bounding K0 represents a large stratigraphic gap in the well logs, where the entire S1 is absent and S0 is directly overlain by sedimentary rocks of S2 or S3 (Fig. 3.5, Fig. 3.6 and Fig. 3.7). In the GR logs, S0 is characterized by a fining up-ward succession that abruptly terminates against a 5-10 meter thick low GR package in the uppermost section of the sequence (Fig. 4.11).

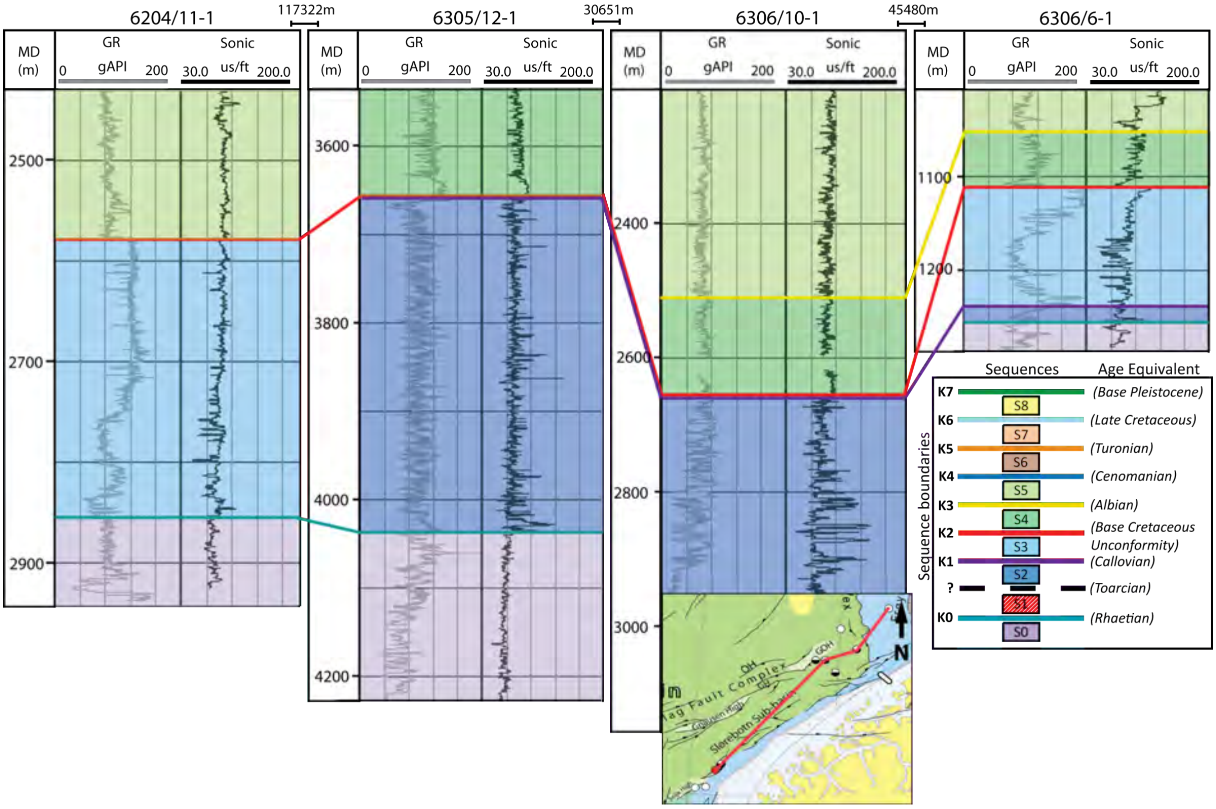


Fig. 4.11 Well correlation of S0 and S2 displaying the Early Triassic and Middle Jurassic sections in more detail.

Seismic character

Generally, S0 is too thin to be recognized in the seismic data, but on the eastern flank of the Gossa High, well 6305/12-1 proved more than 250 meters of S0 that can be locally interpreted. The sequence lies within a rotated fault block that banks against the flank of the Gossa High (Fig. 1.2). Internally, the seismic character of S0 shows a very chaotic low to medium amplitude reflector pattern.

Interpretation

S0 correlates to deposition of the Late Triassic Grey and Red beds (Ichron, 2015) that are deposited in a continental to marginal marine environment on the Halten Terrace area (Dalland et al., 1988).

S0 is hardly recognized in the seismic data and only a small number of wells have encountered the sequence (Fig. 3.5, Fig. 3.6 and Fig. 3.7). The lack of S0 and pre-S0 sedimentary rocks on the Gossa High (6305/12-2), Silje High (6204/10-2) and southern part of the Frøya High (6306/10-1) suggests a potential tectonic event that resulted in erosion and/or non-deposition on the highs (Fig. 3.5, Fig. 3.6 and Fig. 3.7). Hence, the present-day structural elements observed in the seismic data (e.g. Fig. 4.7) could have started to develop into a series of highs and lows as early as in the Early/Middle Triassic times (Jongepier et al., 1996). This is further supported by seismic interpretation of S0, where the thickness of the sequence can be interpreted to increase from well 6305/12-1 at the flank of the Gossa High towards the Slørebotn Sub-basin (Fig. 1.2).

The lithofacies of S0 predominantly reflects a continental depositional environment with arid alluvial deposits in the west (Gossa High) and east (Møre platform). The mineral composition of S0 shows the influence of being sourced from at least two different basement provinces (Mørk and Johnsen, 2005). More specifically, S0 contains traces of greenstone that has been cored on the Gossa High by well 6305/12-2 and metamorphic rocks from the Møre coast in the east (Fig. 4.12). Although S0 has not proven to consist of any reservoir properties, a continental depositional environment is likely to comprise a large variety of facies that could potentially form excellent reservoir units (Faglerland, 1990).

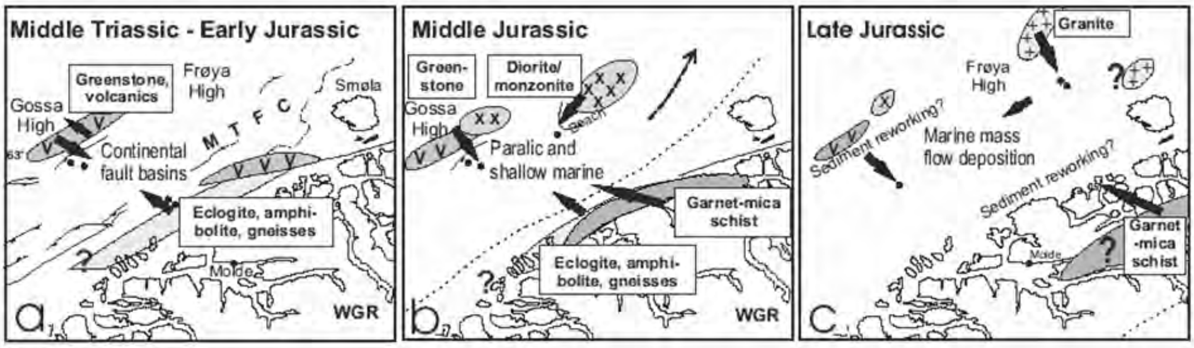


Fig. 4.12 Cartoon illustrating location of the local basement provinces (grey areas) and their characteristic rock types. *Black arrows represent sediment transport to the Slørebotn Sub-basin. MTFC, Møre-Trøndelag Fault Complex; WGR, Western Gneiss Region (Møre coast). Modified from Mørk and Johnsen. (2005).*

4.3.2 Sequence 1 (Hettangian-Toarcian)

Well character

Sequence 1 (S1) has only been penetrated by IKU shallow wells (Fig. 3.5) on the Møre platform area (Fig. 1.1 and Fig. 4.13). The cored intervals were subdivided into three units by Smelror et al. (1994); unit A (Rhaetian-Hettangian); unit B (Pliensbachian-Toarcian); and unit C (Toarcian-Aalenian). However, this study will use the Jongepier et al. (1996) re-interpretation of unit B and C, which were assigned to the Middle Jurassic (S2) due to the close assemblages of fossils found in the cored S2 rocks on the Gossa High by wells 6305/12-1 and 6305/12-2. Thus, unit A will be the only section of sedimentary rocks that corresponds to S1.

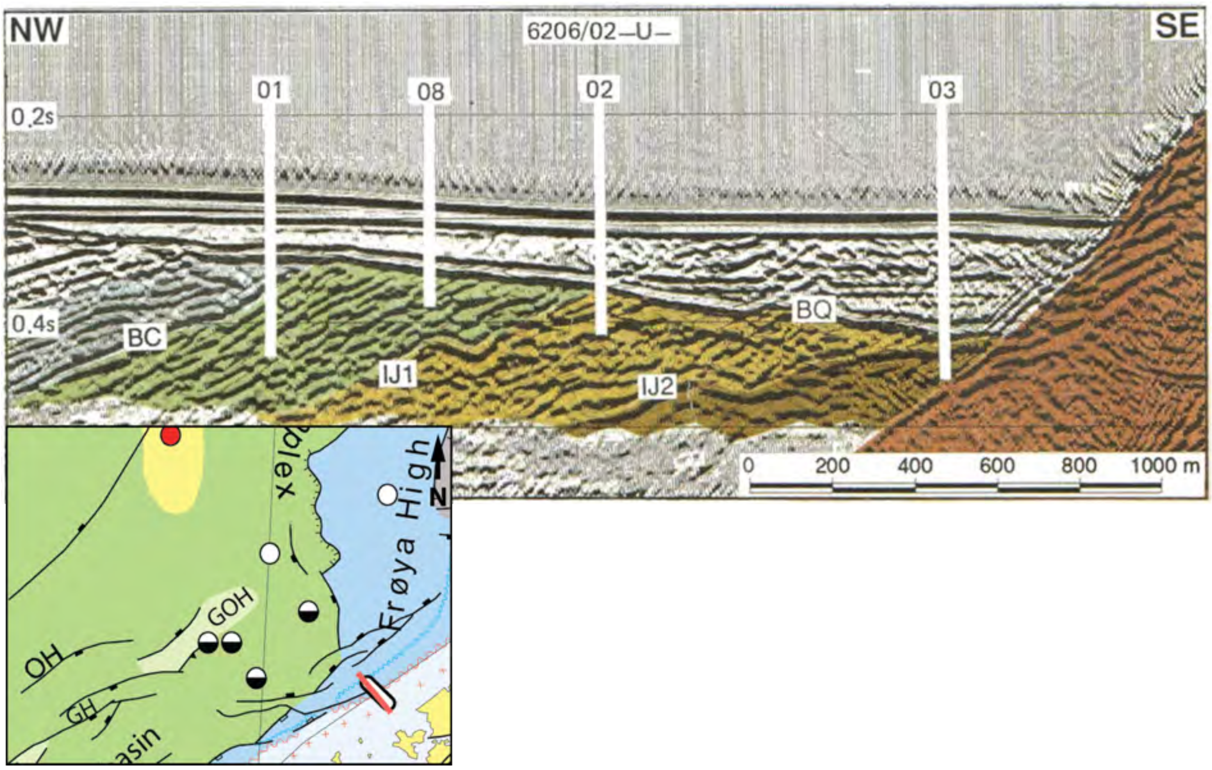


Fig. 4.13 Seismic profile across the shallow IKU core locations. *BC=Base Cretaceous reflection; IJ1 = intra Jurassic reflection 1; IJ2 = intra Jurassic reflection 2; blue colour = S4, S5 and S7; green and light brown colours = S2; Dark brown colour = lower S1 (Hettangian) and upper S0 (Rhaetian); Orange colour = basement. Modified from Smelror et al. 1994..*

Unit A comprises poorly sorted conglomerates that are interbedded with mudstones and are similar to S0, the lithofacies in S1 is interpreted as alluvial fan deposits (Smelror et al., 1994; Jongepier et al., 1996).

Seismic character

Seismically, S1 was not possible to extrapolate any further away from the well location due to the limited seismic coverage on the platform area. However, there is one 2D seismic profile located close to the shallow IKU wells and the seismic character of S1 is characterized by a chaotic acoustic signature with discontinuos reflections (Fig. 4.13) (Smelror et al., 1994).

Interpretation

In the Halten Terrace, S1 correlates to the Båt Group that comprises the Åre, Tilje, Tofte and Ror formations that are deposited in a shallow marine to deltaic envrionment (Dalland et al., 1988; Ichron, 2015).

The similar continental lithofacies found in S1 as in S2 implies that during the Upper Triassic to Hettangian times, an arid alluvial environment dominated the Møre platform area. This might be the case for other structural highs in the study area (e.g. Gossa and Frøya highs), but this must remain speculative due to the absence of S1 in the wells (Fig. 3.7). Furthermore, the lack of S1 sedimentary rocks in the wells (6306/10-1, 6305/12-2 and 6204/10-2) that drilled the structural highs suggests that a tectonic phase might have taken place in the late Early Jurassic, which caused erosion and/or non-deposition of S1 on the highs (Fig. 4.6 and Fig. 4.7). Whether S1 is present in the deeper parts of the Slørebotn Sub-basin is unknown, but it is suggested that the sub-basin developed into a low-lying structural element in the Early/Middle Triassic (Jongepier et al., 1996) and the sub-basin could still have had structural depression in the Early Jurassic.

4.3.3 Sequence 2 (Aalenian-Callovian)

Well character

Sedimentary rocks of S2 (Bajocian-Bathonian) have been encountered by wells on the Gossa High (6305/12-1 and 6305/12-2), Frøya High (6306/6-1) and shallow IKU wells on the Møre platform (Fig. 3.5, Fig. 3.6 and Fig. 3.7) (Jongepier et al., 1996). The youngest unit of S2 (Bajocian-Early Callovian) was encountered in the southern part of the Frøya High by well 6306/10-1 (Fig. 3.5).

S2 cored lithofacies from the wells 6305/12-1 and 6305/12-2 comprises alternating sandstones, siltstones, claystones, coal beds and subordinate conglomerates (Jongepier et al., 1996). Furthermore, the lithofacies in well 6305/12-2 is slightly different than the lithofacies in well 6305/12-1, in the sense that the grains are coarser and the rocks show syn-sedimentary deformation in the form of redeposited sandstone clasts and a steeply inclined limb of a fold (Jongepier et al., 1996). The lithofacies in well 6305/12-2 is interpreted as alluvial fan deposits, whereas farther east in well 6305/12-1, evidence of coastal plain deposits with a humid climate have been interpreted (Jongepier et al., 1996). Similar lithofacies found in well 6305/12-2 is described in the shallow IKU cores and the lithofacies indicate an alluvial fan depositional environment (Smelror et al., 1994). The mineral composition of S0 display similar basement signature to the basement rocks that have been penetrated on the Gossa High (6305/12-2), Frøya High (6306/10-1) and collected on the Møre coast (Mørk and Johnsen, 2005) (Fig. 4.12).

The lithological data from the youngest S2 unit encountered on the Frøya High (6306/10-1) is interpreted to represent a transgressive event, which marks a change from predominantly continental environment in the pre-Callovian to marginal marine conditions in the Early Callovian (Jongepier et al., 1996).

Different portions of S2 sedimentary rocks have been encountered on the Gossa and Frøya highs. On the Gossa High, well 6305/12-2 proved a 181.5 meter thick S2 unit that unconformably overlies metamorphic greenstones (Mørk and Johnsen, 2005). On the flank of the high, well 6305/12-1 encountered 377 meters of S2 which directly overlies S0 sedimentary rocks (Fig. 1.2, Fig. 3.7 and Fig. 4.10). The thickness of S2 on the Frøya High ranges from 15 to 288 meters in wells 6306/6-1 and 6306/10-1, respectively. In the GR logs, S2 is characterized by a heterogeneous log pattern that alternate between high and low values that are defined in the upper part by K1, which is marked by an abrupt increase in the GR and sonic logs (Fig. 4.11). The wide range of the different lithofacies found in the core data is reflected in the heterogeneous GR pattern.

Seismic character

A regional seismic interpretation of S2 was not possible as the sequence is either below the seismic resolution or it is truncated against the structural elements (Fig. 1.2 and Fig. 4.6). However, similar to S0, S2 is a part of a rotated fault block on the eastern side of the Gossa High and the sequence can be locally interpreted on the flank of the high and down to the Slørebotn Sub-basin where the thickness increases before it eventually truncate in the hanging wall against a listric fault (Fig. 1.2). The internal seismic pattern of S2 in the hanging wall is characterized by low to medium amplitude discontinuous reflectors and a similar pattern can be observed in the adjacent footwall where S2 has been postulated (Fig. 1.2). Although well 6205/3-1R only penetrated sedimentary rocks of S3 in the rotated footwall block, S2 is assumed to be present below, since the thickness of the sequence increases towards the Slørebotn Sub-basin and the listric fault is interpret to have developed after deposition of S2. In addition, the seismic signature of the fault block is characterized by a series of continueous reflectors that do not display any evidence of being an unconformity surface, which suggests that erosional forces have not removed the sequence (Fig. 1.2).

Interpretation

S2 correlates to the Fangst Group that comprises the Ile, Not, Garn and lower Melke formations (Ichron, 2015). On the Halten Terrace area, the Fangst Group represents shallow marine to coastal plain deposits (Dalland et al., 1988).

Core interpretation of S2 shows that an arid alluvial environment dominated both the western (Gossa High) and eastern (Møre platform) parts of the study area, but now with a coastal plain environment in-between, in the Slørebotn Sub-basin (Smelror et al., 1994; Jongepier et al., 1996). The cored S2 comprises alternating fine- and coarse-grained clastics, and this pattern is reflected in the GR logs as a highly heterogeneous log pattern that alternate between high and low readings (Fig. 4.10 and Fig. 4.11).

As mentioned earlier, the Slørebotn Sub-basin could have structurally formed as a low-lying area during the Early/Middle Triassic and this is further implied by the well logs and seismic data. More specifically, the thickness of S2 increases from 181.5 meters in well 6305/12-2 on the Gossa High to 377 meters in well 6305/12-1 towards the Slørebotn Sub-basin (Fig. 4.10). Seismically, the thickness of S2 is interpreted to increase down flank of the Gossa High towards the Slørebotn Sub-basin, where the sequence eventually truncates against a listric fault (Fig. 1.2). Whether S2 is present in the footwall below well 6205/3-1R is still speculative, but it is thought to be present if the assumption that the sub-basin was a low-lying structural element prior to deposition of S2 is correct. This would then mean that there are potentially thick units of S2 sedimentary rocks that have accumulated within the Slørebotn Sub-basin.

The lithological data from well 6306/10-1 on the Frøya High suggests that a rise in the relative sea-level occurred in the boundary between the Bathonian and Early Callovian, where marginal marine rocks were penetrated (Jongepier et al., 1996). The major change in the depositional environment indicates that a potential extensional event occured in the Late Bathonian. The extensional event could further explain the large thickness variations observed in the wells along the Gossa High (i.e. 6305/12-2 and 6305/12-1) (Fig. 4.10). In addition, the Gossa High is believed to have been tectonically active during the Bathonian times because of the syn-sedimentary deformation observed in the cored section of S2 in well 6305/12-2 (Jongepier et al., 1996).

The uppermost portion of the S2 (Middle to Late Callovian) stratigraphy has not been encountered in the area, but it is assumed that the depositional environment became increasingly more marine that eventually led to the deposition of Late Jurassic deep marine claystones (Dalland et al., 1988; Jongepier et al., 1996).

4.3.4 Sequence 3 (Oxfordian-Early Ryazanian)

Well character

Sequence 3 (S3) has been penetrated by wells on the Frøya High (6306/6-1) and in the northern (6205/3-1R) and southeastern (6204/11-1 and 6204/11-2) parts of the Slørebotn Sub-basin (Fig. 3.5, Fig. 3.6 and Fig. 3.7). Well 6205/3-1R targeted a rotated fault block in the Slørebotn Sub-basin (Fig. 1.2) where 870 meters of S3 was penetrated (Fig. 4.10) and several cores were taken at this interval. The core and dipmeter log data from well 6205/3-1R of S3 shows that three distinct periods of fault block rotation during; (1) the Kimmerdigian, where very gentle fault block rotation (5°) occured; (2) followed by an increase in rotation (15-20°) during the Early Volgian times; and (3) the final phase of fault block rotation (20°) climaxed in the Middle Volgian times (Jongepier et al., 1996).

The lithofacies in the cored lower section (Kimmeridgian) of S3 in well 6205/3-1R comprises alternating siltstones and mudstones with minor fine- to coarse-grained sandstone intervals that show very little tectonic disturbance (Jongepier et al., 1996). In addition, the fine-grained sedimentary rocks contain on average a TOC of 5% and the coarse-grained sandstones have typically a sharp base with mudstone rip-up clasts (Jongepier et al., 1996). The cores in the upper section (Middle Volgian to Early Ryazanian) show similar lithofacies than the underlaying section, but with slightly finer-grained sandstones. The mudstones contain high TOC levels that are interbedded with fine-grained sandstone intervals (Jongepeier et al., 1996). The sandstone intervals in the upper and lower part of S3 are interpreted as low-density turbidites and gravity flow deposits, respectively, and the reservoir qualities in these sandstones are very poor due to extensive mechanical and chemical compactions (Jongepier et al., 1996; Mørk and Johnsen, 2005).

The top part of S3 is defined by K2, which is represented in the well logs by an abrupt decrease in GR and increase in sonic values (Fig. 4.10). Internally, S3 is characterized in the GR log by alternating high (150-200 gAPI) and low values, and these low value intervals have a blocky GR signature with a sharp base that ranges from 40-50 meters in thickness (Fig. 4.14).

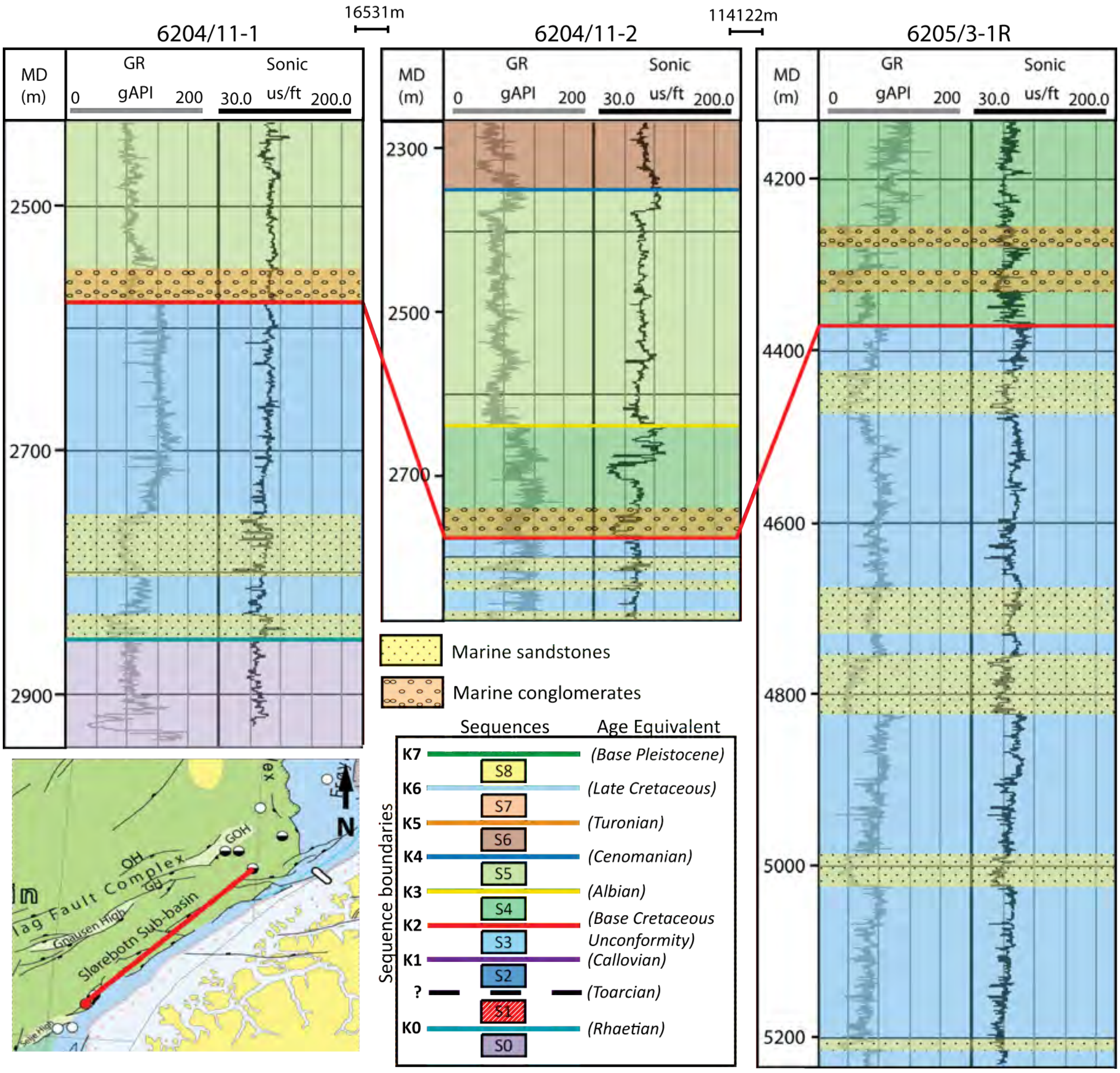
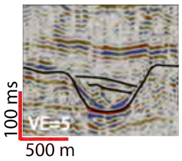
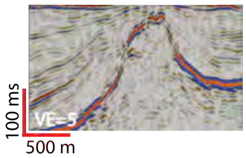
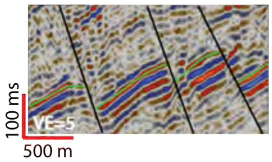
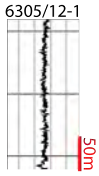
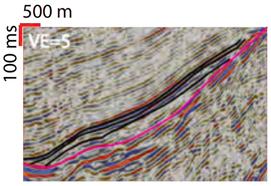
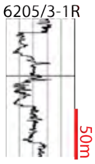
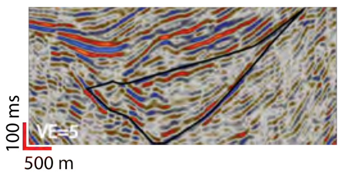
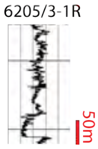
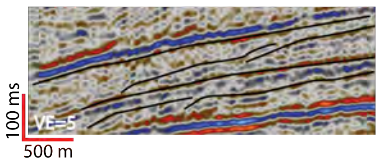
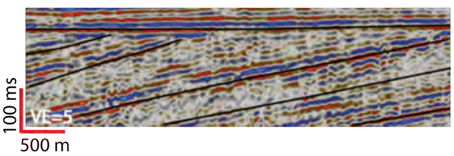
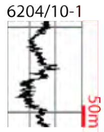
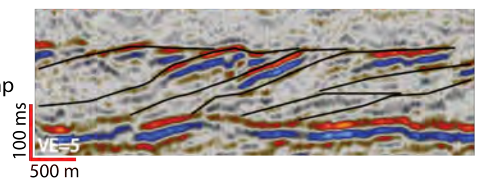
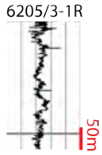


Fig. 4.14 Well correlation of S3 displaying the Late Jurassic and Early Cretaceous sections in more detail. *The low blocky GR intervals that are interbedded in the high GR of S3 is interpreted as coarse fan deposits. Core data from well 6205/3-1R shows that these intervals comprise of fine- to coarse-grained sandstone deposits.*

Seismic character

The seismic interpretation of S3 was only possible in the northern part of the Slørebotn Sub-basin where well 6205/3-1R, penetrated a rotated fault block that comprise a thick succession of S3 (Fig. 4.7). The seismic character of S3 in the rotated fault block can be divided into a lower and upper part based on the different seismic signature. The lower part is characterized by steep dipping (50°) continuous medium to high amplitude reflectors that truncate against FF5 (Fig. 1.2). Whereas, the upper part is characterized by a discontinuous to chaotic reflector pattern that displays a wedge-shaped geometry (Fig. 1.2) (Table 4.1;F5). S3 is bounded at the base and top by distinct hard-kick reflectors, K1 and K2, respectively (Fig. 1.2 and Fig. 3.5). The time-thickness map of S3 shows that the sequence has it thickest development in the eastern part of the Slørebotn Sub-basin (Fig. 3.4).

Table 4.1 Overview of the different seismic facies that have been recognized in the study area.
See Fig.16 for the location where the seismic facies have been recognized.

Facies number	Description	Interpretation	Seismic facies	GR Pattern 0 gAPI 150
F1	Incision that cut down and terminate reflectors, whereas the internal reflectors onlap.	Canyons		Not drilled
F2	High amplitdue reflector at the top, below is characterized by noise and pull-up effect.	Sill		Not drilled, but grab samples proves the presence of volcanic rocks (Bugge et al. 1980)
F3	Highly discontinuous reflectors with a homogenous thickness.	Hydrofactured shales		6305/12-1 
F4	High to medium amplitude reflectors that onlap and downlap onto the same seismic surface (Pink line).	Slope apron/ debris flow fans		6205/3-1R 
F5	Slightly chaotic to discontinous reflectors that form a wedge geometry.	Growth wedge		6205/3-1R 
F6	Fairly continuous downlapping, low to medium amplitude reflectors.	Prograding turbidite system		6205/3-1R 
F7	Continuous low-angle, high to medium amplitude reflectors that form an oblique tangential geometry.	Prograding shelf-marging clinoforms		6204/10-1 
F8	Chaotic to discontinous, low to high amplitude reflectors.	Submarine fan/ gravitational slump		6205/3-1R 

Interpretation

S3 correlates to the Viking Group that comprise the Spekk, Melke (time-equivalent to the Draupne and Heather formations in the northern North Sea, respectively) and Rogn formations (Ichron, 2015). On the Halten Terrace area, the Viking Group mainly consists of deep-marine mudstones that are occasionally interbedded with carbonate and sandstone beds (Dalland et al., 1988).

The core and seismic data shows very different characteristics for the lower and upper part of S3. The lower part (Kimmeridgian) show evidence of very little tectonic disturbance in lithofacies (Jongepier et al., 1996) and seismically, this interval is characterized by continuous and steep dipping reflectors (Fig. 1.2). Furthermore, the lithofacies in the lower part has been interpreted to correspond to the Melke Formation (Ichron, 2015) that is deposited in an anoxic to suboxic deep marine environment, and the coarse-grained sandstone intervals (Intra Melke Formation) are interpreted as gravity flow deposits (Jongepier et al., 1996).

In contrast, the upper section (Volgian to Early Ryazanian) shows very strong tectonic control in the form of a highly chaotic wedge shaped package that banks against FF5 (Fig. 1.2). In addition, core and well data indicates that S3 was deposited during a period of substantial rifting and rotation of fault blocks (Jongepier et al., 1996). The lithofacies of the upper section has been interpreted to correspond to the Spekk Formation (Ichron, 2015) that is deposited in an anoxic deep marine environment, and the fine-grained sandstone intervals (Intra Spekk Formation) are interpreted as turbidite deposits (Jongepier et al., 1996).

The absence of S3 sedimentary rocks on the Gossa High (6305/12-1 and 6305/12-2) and the Frøya High (6306/10-1) implies that a substantial relief developed as a result of the mid Late Jurassic extensional event, which caused non-deposition on the highs and a thick accumulation of S3 in the sub-basin. Furthermore, S3 is believed to have been deposited over most parts of present-day basin areas and thus, the bounding K2 surface can be used as a minimum depth estimation for the source rock (Fagerland, 1990; Faleide et al., 2010).

4.3.5 Sequence 4 (Ryazanian-Albian)

Well character

Sequence 4 (S4) has been widely penetrated by the wells throughout the study area (Fig. 3.5, Fig. 3.6 and Fig. 3.7). It is only in the southeastern flank of the Slørebotn Sub-basin in well 6204/11-1 that the sequence is missing (Fig. 4.10). The lithofacies in S4 mainly comprises mudstones and siltstones, with the exception of a thick succession of coarse clastics at its basal part (Jongepier et al., 1996; Martinsen et al., 2005). Core data from well 6205/3-1R at the base of S4 shows that the section comprises poorly sorted conglomerates with angular clasts, interbedded with sandstones and mudstones (Fig. 4.15a) (Jongepier et al., 1996; Martinsen et al., 2005). The reservoir quality of these coarse clastics is very poor due to the immature texture and mineralogy of the rocks (Mørk and Johnsen, 2005). In addition, similar coarse clastics have been penetrated by wells 6204/10-1 and 6204/11-2 in the southeast (Fig. 3.6). Although no cores were taken, the final well-report from well 6204/10-1 describes the unit to comprise predominantly of poorly sorted conglomerates with angular clasts and layers of sandstone and claystone (NPD factpages).

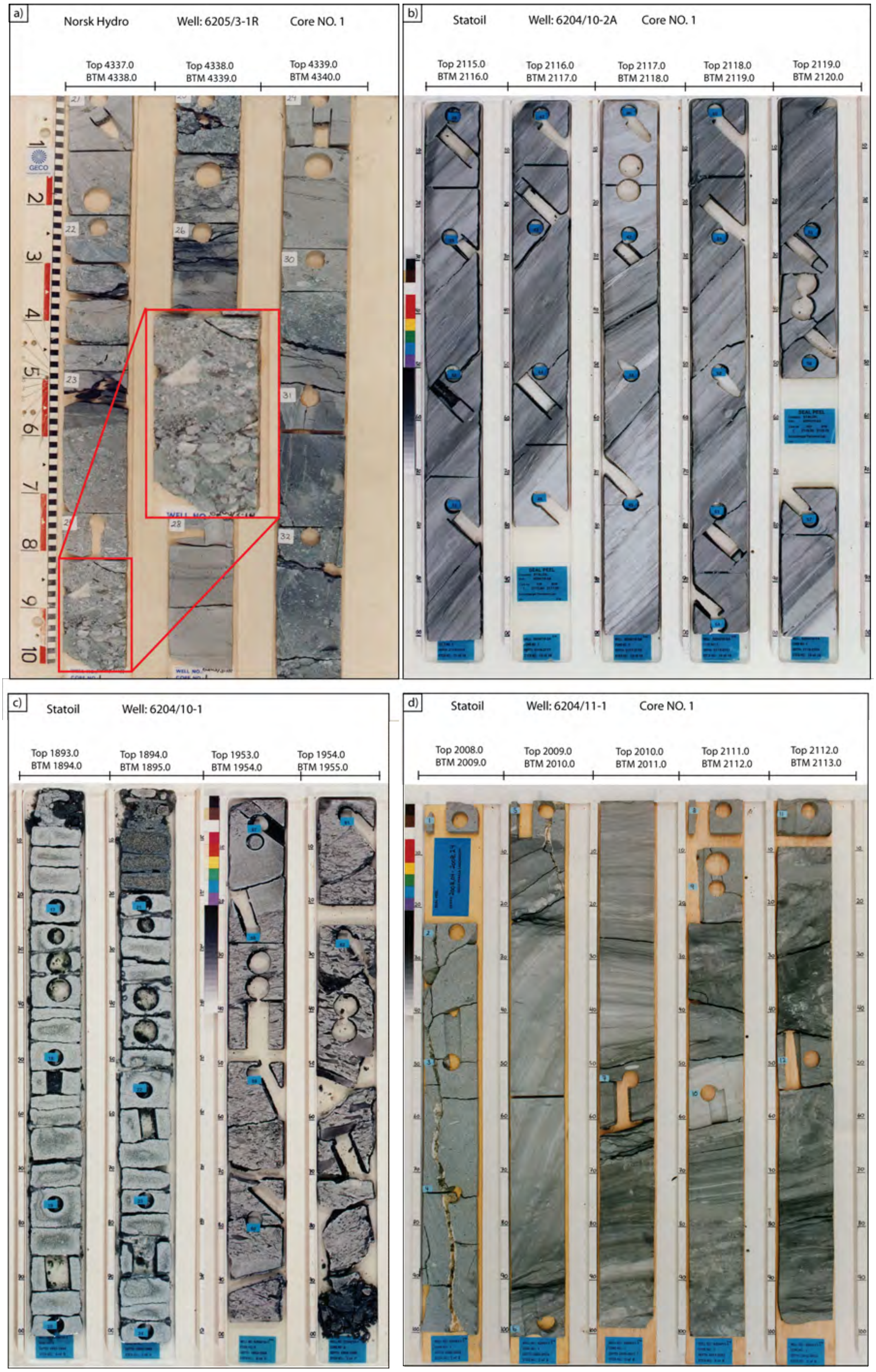


Fig. 4.15 Core photos *a) poorly sorted conglomerates penetrated by well 6205/3-1R; b), c) and d) show core photos of the sandy Rødspette Member penetrated by wells 6204/10-2A, 6204/10-1 and 6204/11-1. Modified from NPD factpages.*

In the GR logs, S4 is characterized by a low blocky interval at its basal parts (Fig. 4.14), whereas above, the sequences show a fining-upward succession (Fig. 4.10). S4 has its greatest development in the Slørebotn Sub-basin (6205/3-1R) where the entire sequence is present (Fig. 3.5). Whereas, on the flanks of the sub-basin (6305/12-1 and 6204/10-1) and on the Frøya High (6306/6-1), large gaps in the lower part of S4 (Ryazanian to Barremian) sedimentary record is interpreted to be missing (Fig. 3.5, Fig. 3.6 and Fig. 3.7).

Seismic character

In the seismic, S4 is bounded at its base by a distinct hard-kick and a moderate hard-kick at the top, K2 and K3, respectively (e.g. Fig. 3.7). The former is characterized by truncating reflectors below, whereas onlap and downlap patterns are observed above (Fig. 1.2 and Fig. 4.3). In addition, K2 is interpreted to truncate the crest of the present-day structural highs in the study area (e.g. Fig. 4.7) and several potential incised valleys are recognized at this level to cut deep down into the crest (Fig. 4.16) (Table 4.1). The onlap and downlap seismic pattern above this boundary closely resembles the shape of slope fans (Table 4.1) and similar seismic facies fans that follow the K2 topography are interpreted to occupy the majority of the Slørebotn basin floor (Fig. 1.2, Fig. 4.2, Fig. 4.3 and Fig. 4.17). The internal seismic pattern of S4 is characterized by medium to low amplitude divergent reflector pattern and the reflectors can be observed to truncate against FF1, FF4 and FF5 (Fig. 1.2, Fig. 4.6, Fig. 4.2 and Fig. 4.3). The sequence has been interpreted to have been widely distributed within the Slørebotn Sub-basin, but it is restricted by FF5 in the east, FF4 in the north and partly by FF1 in the west. Only in the northwest, at the Gossa High, is the entire S4 observed to truncate against FF1 (Fig. 1.2). These observations are reflected in the time-thickness map of S4, where the sequence has its thickest development within the Slørebotn Sub-basin, whereas the sequence thins out towards the location of the structural highs (Fig. 3.4)

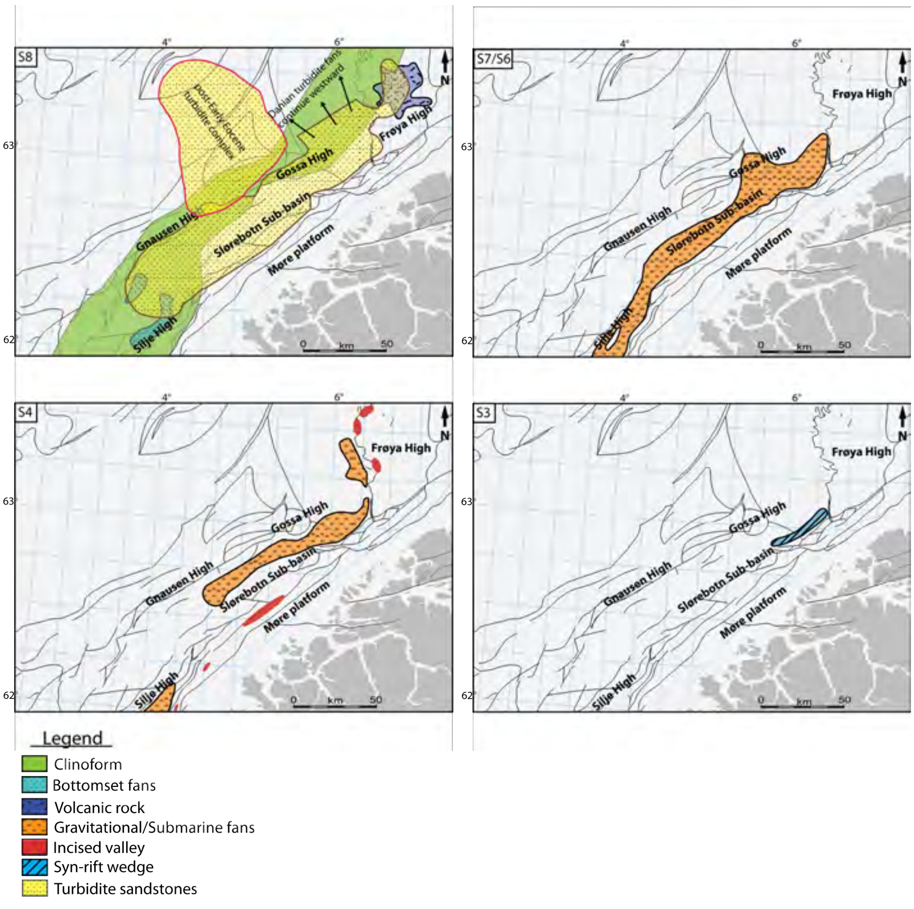


Fig. 4.16 Location of the seismic facies that are recognized within the study area.

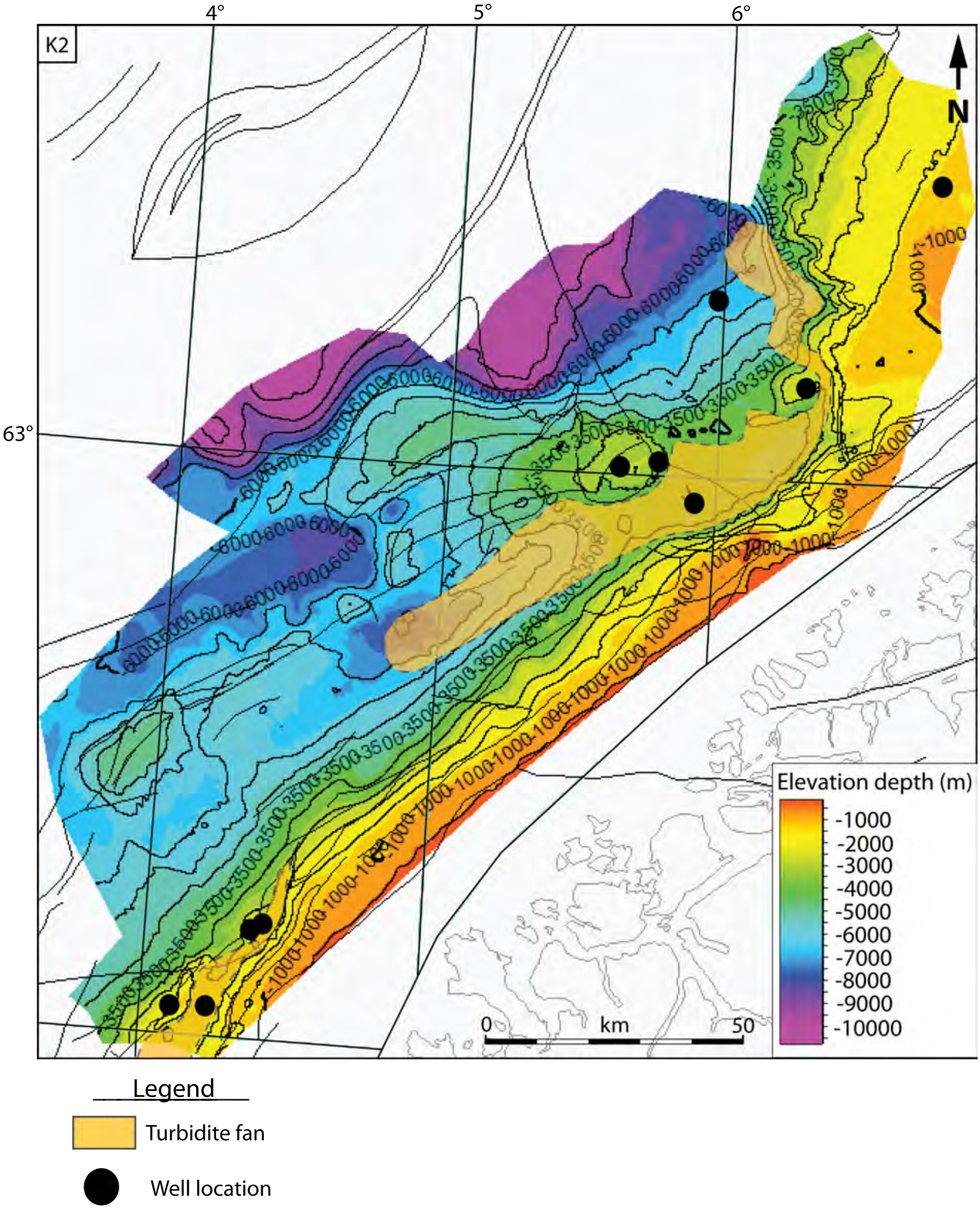


Fig. 4.17 Depth map of K2 surface with location of the interpreted debris flow deposits, time-equivalent to the Agat Member in the northern North Sea.

Interpretation

S4 correlates to the Langbarn Formation of the lower Cromer Knoll Group (Ichron, 2015), which is equivalent to the lower portion of the shallow to deep marine Lange Formation on the Halten Terrace (Dalland et al., 1988).

S4 depositional period is interpreted as a deep marine environment that was accompanied by high thermal subsidence and passive infill of the rift topography that formed during the mid Late Jurassic-Early Cretaceous (Jongepier et al., 1996; Vergara et al., 2001). This is supported by the internal divergent reflector pattern, onlapping reflectors onto the pre-existing highs (Fig. 1.2 and Fig. 4.7) and the large thickness variations across the study area (Fig. 3.4). The absence of the Ryazanian to Barremian rocks in the well logs that target the flanks of the Slørebotn Sub-basin (6305/12-1 and 6204/10-1) and the Frøya High (6306/6-1) suggests non-deposition or erosion of the initial S4 deposits. In addition, S4 seismic reflectors can be observed in the southwest to eventually over-step the structural relief that was created during the mid Late Jurassic rift episode (Fig. 4.2 and Fig. 4.3), whereas in the northwest, the sequence truncates against the Gossa High (Fig. 1.2). This implies that laterally, the subsidence rates differed considerably in the area, which was controlled by underlying structural elements (Jongepier et al., 1996).

S4 predominantly comprises open marine mudstones and siltstones (Vergara et al., 2001), which is reflected by the high GR readings in the well logs (Fig. 4.10). A notable exception from the fine-grained material is the thick coarse clastic unit that is penetrated by well 6205/3-1R in the basal part of sequence (Fig. 4.14). The core data shows that this unit comprises poorly sorted conglomerates and coarse sandstones that are interpreted as high-density slope apron fans (Jongepier et al., 1996; Swiecicki et al., 1998). Seismically, the geometry of this interval is characterized by a fan shape that can be observed to onlap the K2 in the hanging wall of FF5 (Fig. 1.2). In addition, similar seismic fan facies can be recognized in the hanging wall next to FF1, FF4 and FF5 to occupy large portions of the Slørebotn basin floor (Fig. 1.2, Fig. 4.2, Fig. 4.3 and Fig. 4.16). Regarding the reservoir potential of the slope apron deposits, the cored section in the north has shown to be quite poor due to the immature texture and mineralogy of the rocks (Mørk and Johnsen, 2005). Furthermore, in the north and middle part of the Slørebotn Sub-basin, these fans lay at depths below 5000 meters (Fig. 4.17) that are seen as unattractive for exploration due to the reduced porosity by mechanical and chemical compaction (Maast et al., 2010). In the south, however, these sandy fan deposits are interpreted to lie at an adequate depth for exploration and the sequence is interpreted to be equivalent to the Agat Member in the northern North Sea (Ichron, 2015). By analogous, similar Early Cretaceous turbidite fans of the Agat Member in the North Sea are proven reservoirs with excellent properties that can reach a thickness of up to 274 meters (Vergara et al., 2001).

4.3.6 Sequence 5 (Cenomanian)

Well character

Sequence 5 (S5) has been penetrated by all the wells in the study area (Fig. 3.5, Fig. 3.6 and Fig. 3.7), but cores were only taken by shallow IKU wells on the Møre platform area. The core data shows that S5 comprises mudstones, claystones and medium-grained sandstones that display numerous slump structures and strong bioturbation (Smelror et al., 1994).

In the well logs, S5 is generally represented as a conformable unit that does not show any stratigraphic breaks (Ichron, 2015). Only in well 6306/6-1 on the Frøya High and in well 6204/10-2A on the southeastern flank of the Slørebotn Sub-basin is the sequence interpreted to be unconformably overlain by S7 and S6 sedimentary rocks, respectively (Fig. 4.10). In the GR logs, S5 is dominated by high readings with a slightly coarsening upward trend that terminates in a distinct GR peak, i.e. the upper bounding K4 (Fig. 4.10). Furthermore, the high GR peak is interpreted in wells 6204/10-1, 6204/11-1, 6204/11-2 and 6305/12-1 as the Blodøks Formation and the thickness of the unit ranges from 15-30 meters (Ichron, 2015). Geochemical data from well 6305/12-1 shows that this interval consists of a semi-organic rich claystone with a measured TOC of 2.2% (NPD factpages).

An exception to the high GR readings that dominate S5 is a highly heterogeneous GR log pattern observed in well 6305/12-1 (Fig. 4.10). The section has been interpreted as a sandstone unit of the Gapeflyndre Member (Fig. 3.5); equivalent to the intra Lange sandstone (Dalland et al., 1988) with a gross thickness of 193 meters, average porosity of 14.4% and containing oil shows (NPD factpages).

Seismic character

S5 is bounded by K3 at the base and top by K4. The latter is characterized in the seismic data as a hard-kick reflector that has a high amplitude (Fig. 3.5 and Fig. 3.6) and shows very little evidence of tectonic disturbance (Fig. 4.3). Internally, the seismic pattern of S5 comprises parallel to minor discontinuous medium amplitude reflectors (Fig. 4.2). The distribution of the sequence is restricted by FF4 in the north (Fig. 4.6), FF5 in the east and FF1 in the northwest (Fig. 1.2). In the southwest, similar to S4, S5 can be observed to overstep the highs that are controlled by FF1 (Fig. 4.2 and Fig. 4.3). In the time-thickness map, S5 gradually increases to the northwest with a decrease above the Gossa High (Fig. 3.4).

Interpretation

S5 is assigned to the lower part of the Blålange Formation (Ichron, 2015), which correlates to the upper part of the Lange Formation on the Halten Terrace area (Dalland et al., 1988). The lithology of the Lange Formation comprises deep marine claystone deposits with occasional sandy intervals of the intra Lange (Dalland et al., 1988).

The lithofacies in the cored S5 on the platform area is interpreted as inner shelf/prodelta deposits (Smelror et al., 1994). The sequence is believed to mainly comprise open marine mudstones (Vergara et al., 2001), but a sandstone unit of the Gapeflyndre Member has been interpreted on the eastern flank of the Gossa High (Fig. 3.5). In addition, a semi-organic rich claystone of the Blodøks Formation is located in the upper part of S5, which suggests that some parts of the study area formed into a suboxic environment. The depositional period for S5 is interpreted to have taken place during minor thermal subsidence and passive infill of the remaining rift topography that formed during the mid Late Jurassic times. The internal parallel reflectors (Fig. 4.2) and the low gradual thickness increase to the west (Fig. 3.4), implies that this was a time of quiescence that was accompanied by minor subsidence farther west in the Møre Basin. Although the present-day architecture of S5 is restricted within the Slørebotn Sub-basin, the time of deposition probably occurred over the entire area since the sequence has been penetrated by wells on the Frøya (6306/10-1 and 6306/6-1) and Gossa (6305/12-2) highs (Fig. 4.10).

The fine-grained Blodøks Formation is penetrated by wells 6204/10-1, 6204/11-1, 6204/11-2 and 6305/12-1, where well 6305/12-1 shows that the claystones contain a TOC of 2.2% (NPD factpages). This implies that some parts of the study area formed as a sub-anoxic environment during the early Late Cretaceous. It is believed by some authors that an effective Cretaceous source rock is yet to be discovered in the mid-Norwegian Sea (Swiecicki et al., 1998). However, newer studies show that the hydrocarbons found in the Ormen Lange and Ellida discoveries differ from the discovered hydrocarbons (e.g. Tyrihans field) that show a strong correlation to the main Late Jurassic source rock in the mid-Norwegian Sea (Garner et al., 2017). Thus, the evidence of a potential working Cretaceous source rock in the mid-Norwegian Sea can give rise to new play types along the Slørebotn Sub-basin.

4.3.7 Sequence 6 (Turonian)

Well character

Sequence 6 (S6) has been widely penetrated in the study area, except from well 6306/6-1, where the entire sequence is seen to be missing (Fig. 3.5, Fig. 3.6 and Fig. 3.7). Furthermore, a stratigraphic gap in the upper portion of S6 (Late Turonian) is interpreted to be missing in wells 6306/10-1 and 6305/12-2; whereas in well 6204/10-2A the lower section (Early Turonian) is missing (Ichron, 2015). In the shallow IKU wells, a hiatus of approximately 2 My separates S5 from the overlaying S6, hence the upper Cenomanian and lowermost Turonian deposits are missing (Smelror et al., 2994). The lithofacies in the core comprises mainly shales with minor amounts of silts that show evidence of strong bioturbation, and intervals of medium-grained sandstones and conglomerates with pebble size clasts (Smelror, et al., 1994). In addition, a core was taken in S6 by well 6204/10-2A, but only core photos were available from the NPD webpage. Based on the photos, the cored sequence can be seen to alternate between light grey and black coloured lithotypes (Fig. 4.15b).

In the GR logs, S6 is characterized by a heterogeneous pattern that alternate between high and low readings, which differ considerably from the underlying S5 (Fig. 4.10). The over all pattern shows a fining-upward trend that ends at the upper K5. The lower GR intervals are interpreted as sandstone units of the Skolest Member and the Rødspette Member (Ichron, 2015), where the latter is equivalent to the Lysing Member (Dalland et al., 1988). These intervals occur at depths of 2440 meters in the Slørebotn Sub-basin (6205/3-1R) and as shallow as 1870-2000 meters on the flanks of the sub-basin (e.g. 6204/10-2A) (Fig. 4.10). Furthermore, the Skolest Member is interpreted to have its greatest development in the northern part of the Slørebotn Sub-basin (6205/3-1R) with a gross thickness of 208.5 meters. Whereas, the Rødspette Member is interpreted to reach a gross thickness of up to 303 meters in well 6204/11-2 and the member reaches into the overlaying S7 (Fig. 3.5and Fig. 3.6).

Seismic character

S6 is bounded at its base by a distinct hard-kick and the top by a moderate soft-kick, K4 and K5, respectively (Fig. 3.5). The latter is a fairly continuous medium to low amplitude reflector that can usually be recognized throughout the study area (Fig. 3.3). Internally, the S6 seismic reflector pattern varies from low to medium amplitude sub-parallel reflectors to discontinuous high amplitude reflectors (Fig. 1.2). The latter is interpreted to be submarine fans (Table 4.1;F8) and the pattern is recognized down slope from the Møre platfrom in the north (Fig. 1.2). The high amplitude submarine fans “light up” the seismic data in an otherwise poor reflection less sequence and the fans pinch out towards the Gossa High in the west.

The distribution of S6 is interpreted to truncate against FF4 in the north (Fig. 4.6), FF5 in the northeast (Fig. 1.2) and the lower part of S6 can be observed to downlap onto the K2 on the Gossa High (Fig. 1.2). Farther southwest, S6 fully oversteps the highs that are defined by FF1 (Fig. 4.2 and Fig. 4.3) and in the middle part of the Slørebotn Sub-basin, the sequence can be observed to overstep FF5 (Fig. 3.8 and Fig. 4.2) and reach further into the Møre platform area (Fig. 4.7). In the time-thickness map, S6 shows a similar trend as the underlying S5, with an increase in thickness to the west and decrease above the Gossa High location (Fig. 3.4).

Interpretation

S6 is assigned to the upper part of the Blålange Formation (Ichron, 2015), which represents the uppermost part of the Lange Formation of Dalland et al. (1988). In general, the lithology in the Lange Formation on the Halten Terrace comprises deep-marine claystone deposits with occasional sandy intervals of the Lysing Formation (Dalland et al., 1988).

S6 depositional period is interpreted as an open marine environment that comprises mainly mudstone deposits (Vergara et al., 2001), but sandstone members of the Skolest and Rødspette do occur frequently in the wells (Fig. 3.5, Fig. 3.6 and Fig. 3.7). The lithofacies from the cored S6 in the shallow IKU wells is interpreted to have been deposited in an outer shelf open marine environment during a relative sea-level high stand (Smelror et al., 1994). Similar to S5, S6 is believed to have been deposited over the entire study area, but has later experienced uplift and erosion on the highs during the Turonian. This is supported by the missing uppermost portion of S6 (Late Turonian) on the Gossa (6305/12-2) and Frøya (6306/10-1) highs (Fig. 3.5). In addition, at the inner part of the Frøya High, well 6306/6-1 shows that the entire S6 is missing (Fig. 3.7), which suggests that the inner part of the high experienced the greatest magnitude of uplift. In the southeast on the other hand, well 6204/10-2A did not penetrate the lower part of S6 (Early

Turonian) on the Silje High (Fig. 3.6). Hence, the high was either already elevated that prevented the initial deposits of S6 or there was uplift in the Late Cenomanian-Early Turonian that only show evidence to have affected the southern part of the study area.

A Turonian phase of uplift can also be implied by the increased influx of coarse clastics into an otherwise fine-grained dominated sub-basin (Fig. 3.5, Fig. 3.6 and Fig. 3.7). The interpreted sandy intervals occur frequently west for the Møre platform (Fig. 4.16). On a final note, similar sandy units found in S6 provide a secondary reservoir in the Smørbukk Field on the Halten Terrace (Spencer et al., 1986).

4.3.8 Sequence 7 (Late Turonian-Early Maastrichtian)

Well character

Sedimentary rocks of sequence 7 (S7) have been widely encountered in the study area (Fig. 3.5, Fig. 3.6 and Fig. 3.7) and several cores have been taken in the Rødspette Member in wells 6204/11-1 and 6204/10-1. The core photo from the latter shows a fining upward pattern with coarse angular grey clasts at the lower part of the core, whereas the upper part is characterized by finer-grained grey coloured lithotypes (Fig. 4.15c). In addition, the typical grey coloured lithotypes are interbedded by an abrupt black coloured rock at 1894.0 meters. The core is described to predominantly comprise moderately sorted medium- to coarse-grained sandstones that contain several well calcareous cemented zones (Final well-report, NPD). The core from well 6204/11-1 show slightly different characteristics with more internal layering and finer-grained grey to light grey lithotypes (Fig. 4.15d). The core is described to comprise medium dark grey, fine- to coarse-grained, moderate to poorly sorted sandstones (Final well-report, NPD).

In the GR logs, S7 is characterized by high readings that are occasionally interbedded by low GR intervals with a sharp base (Fig. 4.10). The thickness of S7 is very uniform in the wells logs, except for well 6306/6-1, where the sequence is significantly thinner. S6 is bounded at the top by K6 and the boundary is represented in the well logs by an abrupt decrease in both the GR and sonic logs (Fig. 4.10).

Seismic character

S7 is bounded at its base by K5 that represents a moderate soft-kick reflector and at the top by K6, which is characterized as a distinct hard-kick reflector (e.g. Fig. 3.5). The seismic pattern below K6 is characterized by low angle truncating reflectors, whereas above, the reflectors onlap/downlap the boundary (Fig. 1.2 and Fig. 4.5). The sequence is interpreted to cover the entire study area where only the lower portion of the sequence (Early Santonian-Late Turonian) truncates against FF4 in the north (Fig. 4.6). On the Møre platform area, the sequence is interpreted to truncate against FF5 in the north (Fig. 1.2), whereas in the middle and southern part, the sequence can be seen to overstep FF5 and eventually truncate against K7 in the innermost part of the platform area (Fig. 4.7 and Fig. 4.8). The overstepping of the present-day highs are represented in the time-thickness map as an abrupt decrease in the time-thickness values (Fig. 3.4). In addition, local depocenters can be observed within the Slørebotn Sub-basin, but the sequence has it greatest development in the southwest.

Internally, sub-parallel reflectors with low to medium amplitude to high amplitude discontinued and parallel reflectors characterize the seismic configuration of S7. The high amplitude discontinued reflectors downlap onto K5, just west for the Møre platform in the northern part of the study area (Fig. 1.2). The interval is interpreted as a slump/submarine deposits (Table 4.1;F8), which corresponds to the sandy Tunga Member that has been penetrated by wells 6205/3-1R, 6305/12-1 and 6305/12-2 (Fig. 3.5). Farther south, along the western margin of the Møre platform, similar seismic facies are observed (Fig. 1.2, Fig. 4.2, Fig. 4.3 and Fig. 4.18). The bright amplitudes decreases towards the west, which suggests that the fan deposits shale-out.

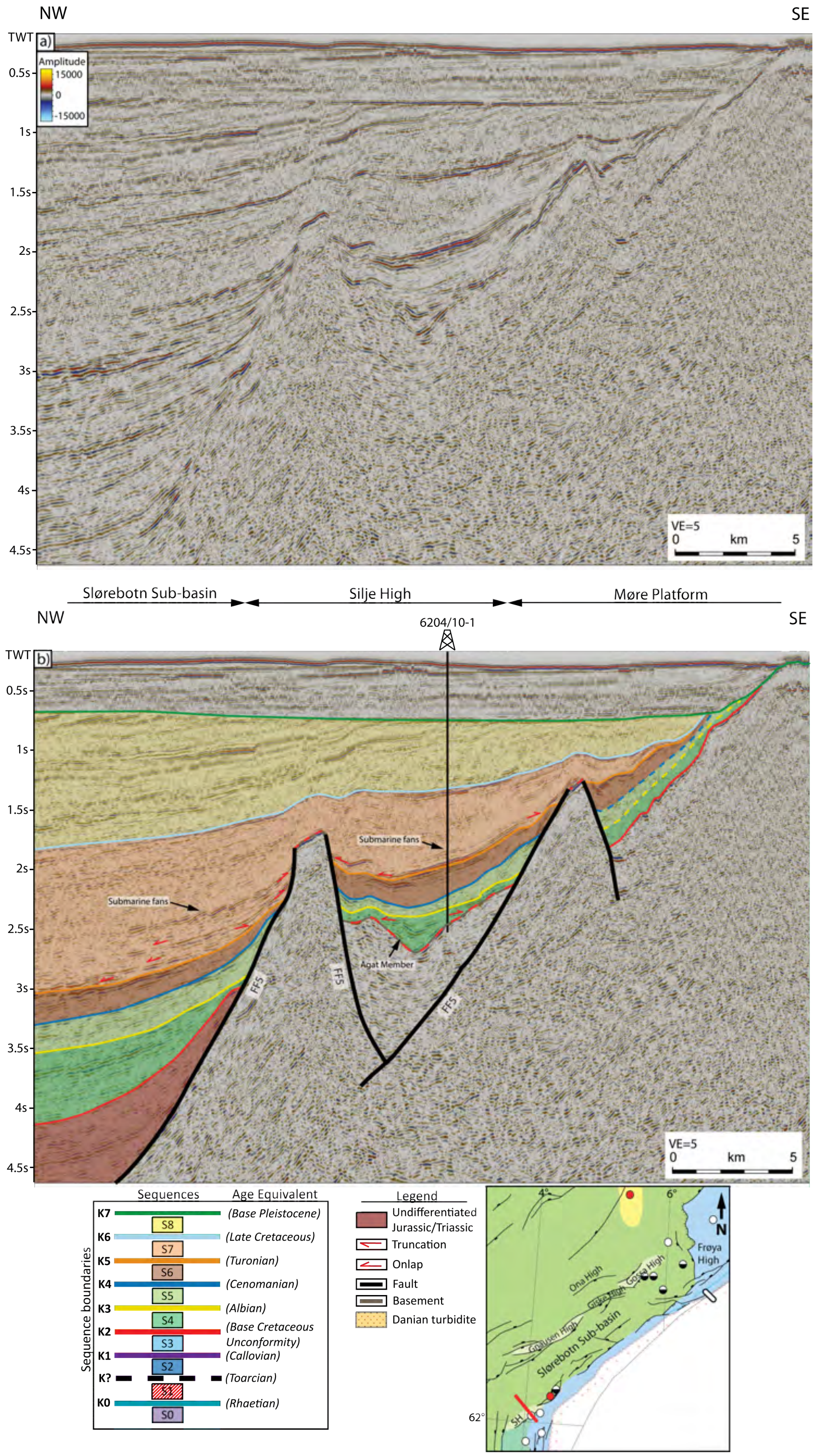


Fig. 4.18 ***2D seismic profile crossing the Silje High*** *a) uninterpreted and b) interpreted seismic profile.*

Interpretation

S7 corresponds to the Kvitnos, Nise and Springar formations that compose the Shetland Group (Ichron, 2015). On the Halten Terrace, the lithology in the Shetland Group comprises mainly open marine claystones, but marine sandstone deposits do frequently occur (Dalland et al., 1988).

S7 is interpreted to mainly comprise open deep marine claystones that are frequently interbedded by sandstone units (Fig. 3.5, Fig. 3.6 and Fig. 3.7). More specifically, four different sandstone members have been interpreted to exist within S7 (Ichron, 2015) and the members are located in different parts of the study area. In the southeastern part, the Rødspette Member, equivalent to the Lysnig Formation on the Halten Terrace (Dalland et al., 1988) and a slightly younger intra Kyrre sandstone, characterizes the base of S7 (Fig. 3.6). Whereas in the north, the base is characterized by the Tunge Member, equivalent to the intra Lange sandstone on the Halten Terrace (Fig. 3.5) (Dalland et al., 1988). In addition, the upper part of S7 in wells 6305/12-1 and 6205/3-1R comprises a sandy interval of the Grindhval Member, which differs from the predominantly fine-grained succession in the upper part of S7 (Fig. 3.5, Fig. 3.6 and Fig. 3.7). The Grindhval Member is equivalent of the Springer Formation on the Halten Terrace (Dalland et al., 1988), and the basal Tang and Springar formations in the Ormen Lange discovery, where the member comprises deep marine turbidite sandstones (Ravnås et al., 2014a).

The depositional period for the lower portion of S7 (Late Coniacian) is interpreted as syn-rift sedimentation due to the local depocenters observed in the time-thickness map (Fig. 3.4) and the internal seismic sequence (Late Coniacian) that thins out from the Slørebotn Sub-basin towards the Møre platform (Fig. 4.7). The thinning out and onlap of the internal sequence onto the Møre platform, is evidence that the platform area was elevated relative to the sub-basin during time of deposition. In addition, the increased influx of the coarse clastics in the basal part of S7 (Fig. 3.5, Fig. 3.6 and Fig. 3.7) further supports a time of tectonic activity that caused erosion and deposition of coarser material into the area. The coarse-grained material is interpreted as submarine fans that have been recognized along the western margin of the Møre platform (Fig. 4.16). The submarine fan intervals comprise a series of high to medium amplitude onlap and downlap reflectors that clearly differs from the fine-grained material in the background, which is characterized by low amplitude reflections (e.g Fig. 1.2).

In contrast, the upper part of S7 (Santonian-Campanian) shows less evidence of tectonic control on the sedimentation and the section are interpreted to have been successful in smoothening the structural relief that formed during the Late Coniacian, since the upper section can be observed in the seismic data to cover the present-day highs in the area (Fig. 4.6 and Fig. 4.7).

4.3.9 Sequence 8 (Lower Paleocene-Early Pleistocene)

Well character

Sequence 8 (S8) has been widely penetrated in the area (Fig. 4.10) and cores are taken in the Middle-Late Danian interval (Ichron, 2015). The core data from well 6205/3-1R shows that the Danian succession comprises thick amalgamated medium- to coarse-grained sandstone turbidites (Gjelberg et al., 2001; Gjelberg et al., 2005; Ravnås et al., 2014a). The core interpretation from wells 6305/9-1 and 6306/10-1 at this interval have shown that the turbidite unit contains a porosity and permeability of up to 27.5% and 2280 mD, respectively (NPD factpages). The Danian turbidite sandstone is informally named the Egga Member, and the member is present in the exploration wells in the northern part of the study area (Fig. 3.5 and Fig. 3.7), whereas in the south (Fig. 3.6), the member has not been confidently recognized (Ichron, 2015). However, cutting samples from well 6204/11-1 at the lower S8 (Danian) level show the presence of minor clean sandstone beds that are interbedded in thick claystones (NPD factpages), which could represent the distal part of the turbidite fans.

The lower boundary of S8 is marked in the well logs as a distinct decrease in the GR and sonic logs (Fig. 4.10), which corresponds to the interpreted Middle-Late Danian turbidites (Gjelberg et al., 2001; Ichron 2015). The gross thickness of the Danian sandstone interval ranges from 150 to 71.5 meters in the well logs (Fig. 4.10) and the interval directly overlies sedimentary rocks of Early Maastrichtian in age (Fig. 3.5 and Fig. 3.7), hence a large gap in the stratigraphic record is present between S8 and the underlying S7 (Ichron, 2015). Furthermore, a large stratigraphic gap is also recorded in the middle and upper part of S8 (Eocene-Miocene) in wells 6205/3-1R, 6306/6-1 and IKU wells (Fig. 3.5 and Fig. 3.7). Whereas, for the wells farther west (6305/12-1, 6305/12-2 and 6305/9-1), only the upper part of S8 (Late Oligocene-Miocene) is missing (Fig. 3.5 and Fig. 3.7). In the southeast, only portions of the middle (Eocene) and upper (Miocene) S8 are missing (Fig. 3.6).

Seismic character

The internal seismic pattern of S8 shows different characteristics and can therefore be sub-divided into a lower (Middle Danian- Early Eocene) and upper part (Middle Eocene-Early Pleistocene). The former is characterized in its lower section by high to medium amplitude parallel reflectors that downlap onto the surface of K6 (Fig. 1.2, Fig. 4.3 and Fig. 4.5). Furthermore, the seismic reflectors can be observed to truncate against FF2 and reactivated FF1 in the western part of the Sløreboth Sub-basin (Fig. 4.2 and Fig. 4.5). The seismic reflectors right above the lower section is highly discontinuous (Fig. 1.2) and the broken up reflector pattern is interpreted as hydrofactured rocks (Table 4.1;F3). The uppermost part of the lower section (Early Eocene) show evidence of doming of the unit as well as the underlying seismic reflectors above the Gossa High and in the Sløreboth Sub-basin (Fig. 1.2 and Fig. 4.2). The Early Eocene seismic reflector is characterized as a distinct hard-kick that is slightly broken-up (Fig. 4.19). Above this reflector boundary, west of the Gossa High, intervals of up to 200ms TWT of high amplitude and discontinuous reflectors onlap and downlap the seismic boundary (Fig. 4.19). The interval is interpreted as turbidite deposits because the internal seismic architecture closely resembles the sedimentary structure of a turbidite complex, e.g. stacking of basinfloor fans and channels (Fig. 4.20) (Ravnås et al., 2014a).

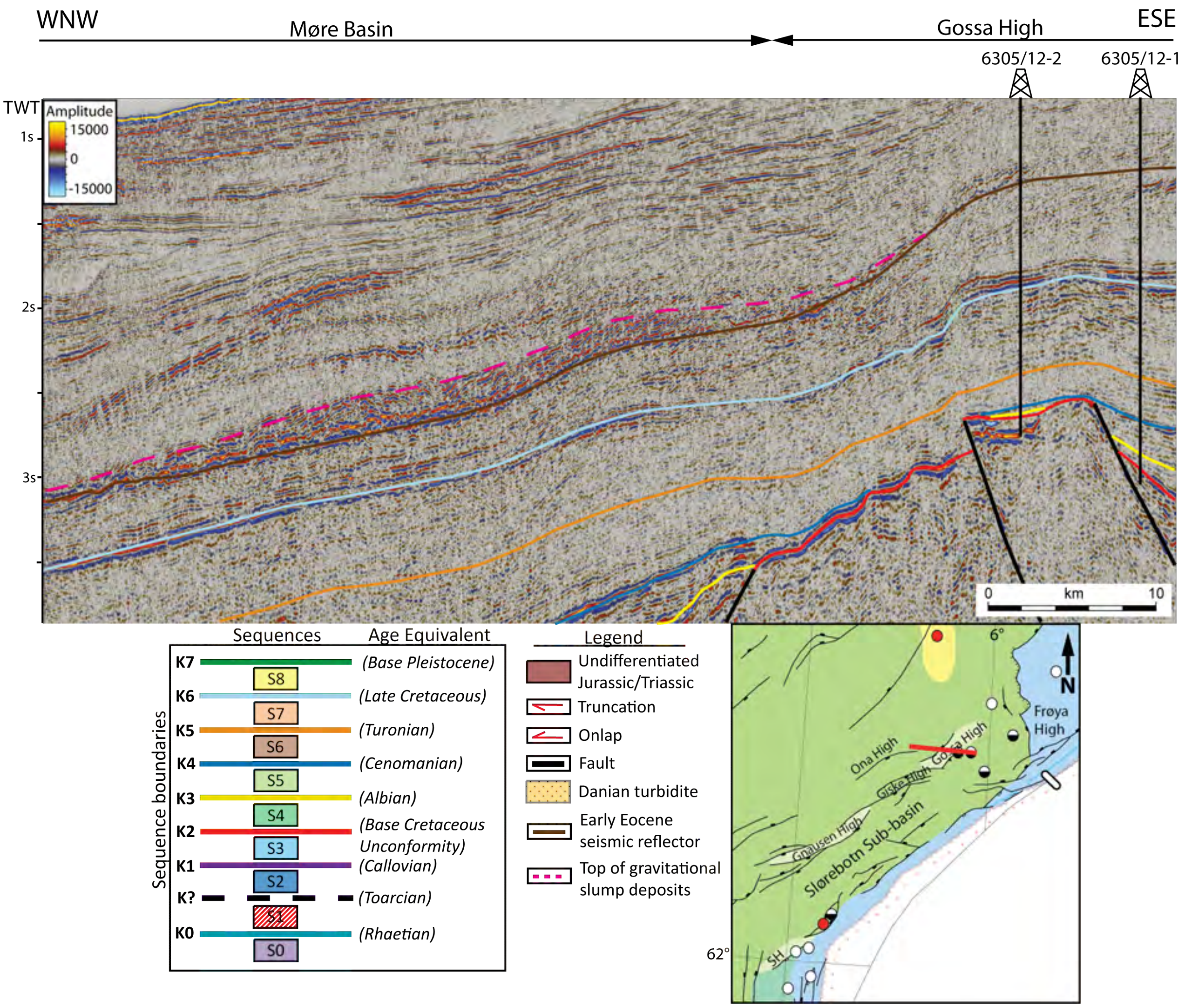


Fig. 4.19 Seismic profile showing the thick package of potential pre-Early Eocene turbidite deposits. *Note the high amplitude and highly discontinuous reflector pattern above the Early Eocene seismic boundary.*

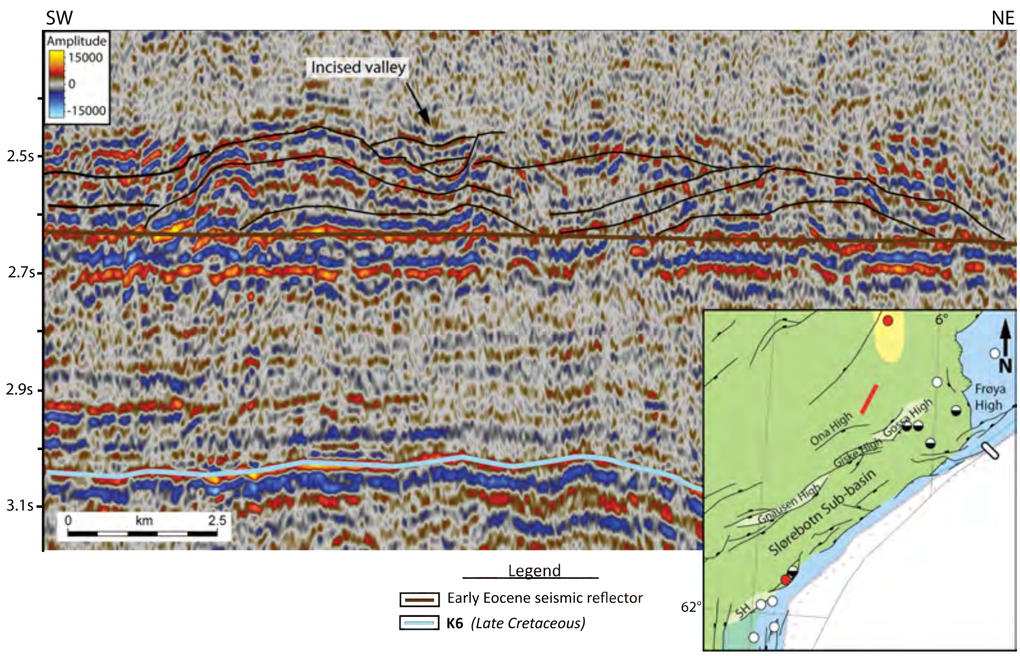


Fig. 4.20 Seismic profile that is flattened at the Early Eocene level. *The seismic architecture above the Early Eocene reflector resembles a typical turbidite complex, such as incised valleys and lens-shaped features that stack and overlap each other.*

The upper part of S8 (Middle Eocene-Early Pleistocene) is characterized by NW dipping medium amplitude reflectors that truncate the overlaying K7 (Fig. 1.2, Fig. 4.2 and Fig. 4.18). The seismic facies of the upper part has been interpreted as clinoforms (Table 4.1;F7) that have later been exposed to an erosional event that removed the topset. The clinoforms are located in the western part of the study area and several basinfloor fans are interpreted to be present in the bottomset (Fig. 4.16). On the Frøya High, several high amplitude reflectors dominate the upper part of S8. These reflectors are interpreted as volcanic sills and they occupy large parts of the high (Fig. 4.16) (Table 4.1;F2).

The time-thickness map of S8 shows a shift in the accommodation space from the Slørebotn Sub-basin in the east to the Møre Basin in the west (Fig. 3.4).

Interpretation

S8 relates to deposition of the Rogaland, Hordaland and lower Nordland groups (Ichron, 2015). The sediment compositions within these groups on the Halten Terrace are mainly dominated by deep marine claystones and siltstones with minor sandy units of the Tang and Brygge formations (Dalland et al., 1988).

The base of S8 is defined by a regional unconformity where the Middle Danian strata directly overlies rocks of Campanian to Early Maastrichtian age (Fig. 3.5, Fig. 3.6 and Fig. 3.7). It is proposed that the unconformity developed in relation to rifting along the North Atlantic spreading center during the Late Cretaceous (Gjelberg et al., 2001; Martinsen et al., 2005; Ravnås et al., 2014a). The rift phase resulted in tilting and uplift of the Norwegian margin that caused influx of coarser material that were sourced from the Norwegian mainland and into the Slørebotn Sub-basin (Sømme et al., 2009; Gjelberg et al., 2001; Gjelberg et al., 2005). The Late Cretaceous unconformity can be traced farther north to the Halten Terrace (Fig. 3.10), but there are no traces of sand above the unconformity (Swiecicki et al., 1998; Vergara et al., 2001).

The marine Danian succession has been frequently encountered by wells (e.g 6205/3-1R) in the northern part of the Slørebotn Sub-basin (Fig. 3.5 and Fig. 3.7) where it comprises up to 150 meter of medium- to coarse-grained turbidite sandstones (Gjelberg et al., 2001; Gjelberg et al., 2005; Ravnås et al., 2014a). Similar turbidite sands are found in the deeper parts of the Møre Basin, where it composes the main reservoir unit in the Ormen Lange gas discovery (Gjelberg et al., 2001; Ravnås et al., 2014a). The underlying seismic boundary (K6) of the Danian reservoir unit has been correlated from well 6305/5-1 at Ormen Lange and to the wells (e.g. 6306/12-2) in the northern part of the study area (Fig. 3.9). It is worth mentioning that the Danian turbidite succession penetrated in the Slørebotn Sub-basin are missing the Early Danian sedimentary rocks (Fig. 3.5 and Fig. 3.7) and is therefore not as well-developed compared to the complete Danian succession that has been encountered (e.g. 6305/5-1) in the Ormen Lange field (Swiecicki et al., 1998; Ichron, 2015).

A second phase of uplift is recognized to have affected the stratigraphic record up to the Early Eocene in the form of minor compressional domes above the underlying structural highs of Late Jurassic or older age (Fig. 1.2 and Fig. 4.2). The event caused a normal fault assigned to FF1 that intersects the Gossa High to change its displacement along strike (Fig. 4.1a), which resulted in a large portion of the Gossa High to uplift (Fig. 1.2). The phase can also be observed to have affected the basin fill of the Slørebotn Sub-basin, where the present-day S4-S8 strata is now elevated and forms a dome shape (Fig. 4.2). Furthermore, the uplift of the Slørebotn basin fill is reflected in the time-thickness map of S8, where the time-thickness is now significantly higher in the west compared to the Slørebotn Sub-basin in the east (Fig. 3.4).

The uplift resulted in exposure of the sedimentary units in Slørebotn Sub-basin, which is evident by the absence of Eocene sedimentary rocks in the wells (e.g. 6204/10-1 and 6205/3-1R) that targeted the sub-basin (Fig. 3.5, Fig. 3.6 and Fig. 3.7). The erosion and bypass were probably greatest in the northeast, since wells 6205/3-1R, 6306/6-1 and the IKU wells show that the entire middle and upper part of S8 (Eocene-Miocene) are missing (Fig. 3.5 and Fig. 3.7). In the southern part, the magnitude of uplift was slightly less, which is evident by the upper S8 (Oligocene) sedimentary rocks that are proven in wells 6204/10-1, 6204/10-2, 6204/11-1 and 6204/11-2 (Fig. 3.6). In the west, wells 6305/9-1, 6305/12-1 and 6305/12-2 penetrated the middle and upper part of S8 (Eocene-Oligocene), which implies that these areas were structurally lower and accumulated sediments of S8 after the uplift in the Early Eocene. In addition, seismic evidence west for the Gossa High, suggests large accumulations of post-Early Eocene turbidite deposits (Fig. 4.19). It is uncertain whether these packages comprise sandstones or mainly fine-grained material, since the material that was being eroded most likely siltstones and claystones. However, sedimentary structures have been recognized within the post-Early Eocene seismic package that suggests a turbidite complex (Fig. 4.20).

The top of S8 is defined by an unconformity surface (K7) that marks the boundary between low-angle truncating reflectors below and horizontal reflectors above (e.g. Fig. 4.18). The former reflectors are recognized as clinoforms (Table 4.1;F7) that are interpreted to dominate the majority of upper portion of the sequence (Fig. 4.16). The timing of the unconformity coincides with a glacial event that removed up to 2km of sedimentary rocks from the Norwegian mainland (Riis and Fjeldskaar, 1992; Martinsen et al., 1999), which were deposited as a wedge of glacio-marine sediments farther offshore (Swiecicki et al., 1998).

5 DISCUSSION

5.1 Timing and processes controlling the structural evolution

The work in this study has revealed some important differences in the structural evolution of the study area, where the tectonic movement and magnitude have affected the structural elements differently through certain time periods. The product of the uneven tectonic movement is reflected in the present-day geometry of the Slørebotn Sub-basin (Fig. 1.2, Fig. 4.2, Fig. 4.3 and Fig. 4.18) where the depth along the basinfloor at the K2 level varies from -3500 meters at it shallowest to more than -7000 meters in the deepest parts (Fig. 4.17). The large difference can be explained by the main fault systems (FF1, FF4 and FF5) that define the main structural elements in the area. Furthermore, the fault systems are interpreted to have had an important impact on the structural evolution through eight phases of tectonic movement during the Early/Middle Triassic, late Early Jurassic, Bathonian, mid Late Jurassic, Turonian, Late Cenomanian-Late Maastrichtian, Early Eocene, and base Pleistocene time. The tectonic evolution of the area is described below and illustrated in Fig. 5.3, Fig. 5.4 and Fig. 5.5.

5.1.1 Early/Middle Triassic rift phase

A potential Early/Middle Triassic tectonic phase is suggested based on the well data alone since there is no clear evidence of extensional movement related to this time period in the seismic data. A precise timing of the tectonic movement was not possible, since the Early to Middle Triassic strata have not been penetrated in the area (Fig. 3.5, Fig. 3.6 and Fig. 3.7). Tectonic movement during the Early/Middle Triassic is implied in wells 6305/12-2 and 6306/10-1, which penetrated S2 sedimentary rocks that directly overlie basement on the Gossa and Frøya highs (Jongepier et al., 1996; Mørk and Johnsen, 2005; Ichron, 2015). The absence of S0 and pre-S0 rocks on these highs suggests that there was a rise in the base level, which caused erosion of pre-S0 and non-deposition of S0 sedimentary rocks. In contrast, thick units of S0 are proven on the flank of the present-day low-lying Slørebotn Sub-basin by wells 6305/12-1 and 6204/11-1 (Fig. 3.5, Fig. 3.6 and Fig. 4.10). The former penetrated 250 meters of rotated S0 (Fig. 4.11), which can be seismically interpreted to thicken towards the Slørebotn Sub-basin (Fig. 1.2). In addition, the mineral content of the sequence shows evidence of metamorphic greenstone and volcanic felsic to intermediate rocks, which is similar to cored Paleozoic basement on the Gossa High and onshore on the Møre coast (Mørk and Johnsen, 2005). This implies that the fault families that bounded the Gossa High (FF1) and at least parts of the Møre platform (FF5) were exposed to erosion and acted as source areas. These findings support that the present-day structural elements may have already been starting to develop as a series of lows and highs prior to deposition of S0. Therefore, a phase of rifting is believed to have occurred in the Early or Middle Triassic times.

Regionally, a Late Permian-Early Triassic rift phase has been documented to have affected the mid-Norwegian margin (Osmundsen et al., 2002; Müller et al., 2005) in the Froan Basin and Vestfjorden Basin (Faleide et al., 2010). In addition, several authors have suggested an Early Triassic rift phase that formed a series of structural high and lows along the Møre margin (Jongepier et al., 1996; Brekke, 2000).

5.1.2 late Early Jurassic rift/uplift? and erosion

Information about the stratigraphic record of S0 in the study area is limited and the sequence has only been proven to exist on the Møre platform by shallow IKU wells (Smelror et al., 1994; Jongepier et al., 1996; Mørk and Johnsen, 2005). The absence of S1 sedimentary rocks in the well logs strongly suggests a tectonic phase in the late Early Jurassic which caused erosion of almost the entire sequence (Fig.

4.10). More specifically, in wells on the Gossa High area, the younger S2 directly overlies basement and S0 sedimentary rocks in wells 6305/12-2 and 6305/12-1, respectively (Fig. 3.5). A similar trend is also seen in the wells on the Frøya High (6306/10-1 and 6306/6-1) where again S2 directly overlies basement and S0 sedimentary rocks (Fig. 3.7). In the south, well 6204/11-1 penetrated S3 which unconformably overlies S0 and on the Silje High, well 6204/10-1 penetrated S4 sedimentary rocks that directly overlie basement (Fig. 3.6). This implies that the bounding fault systems (FF1, FF4 and FF5) that define these highs were active during the Early Jurassic times.

The proposed late Early Jurassic tectonic phase can be correlated to a Toarcian-Early Aalenian rift phase that has been documented in the northern North Sea (Hesthammer et al.,1999) and to a Pliensbachian-Early Aalenian rift phase on the Halten Terrace (Corfield and Sharp, 2000). It is therefore believed that the eastern Møre margin could have experienced similar tectonic activity during the Early Jurassic (Jongepier et al., 1996). In addition, Grunnaleite and Gabrielsen. (1995) interpreted fault movement along the Klakk Fault Complex (FF4) somewhere between the Hettangian-Pliensbachian times.

5.1.3 Bathonian rift phase

S2 sedimentary rocks are frequently penetrated by the wells in the northern part of the study area. On the Gossa High (6305/12-1 and 6305/12-2), Frøya High (6306/10-1) and Møre platform area (IKU wells), the sequence is usually seen to be overlain unconformably by S4 strata (Fig. 3.5 and Fig. 3.7) (Smelror et al., 1994; Jongepier et al., 1996). Only on the inner part of the Frøya High (6306/6-1) has S3 sedimentary rocks been proven to overlie S2 (Fig. 3.7). Core interpretation of the lower S2 (Bajocian-Bathonian) lithofacies in well 6305/12-2 and the IKU wells suggests an alluvial fan environment on the Gossa High and Møre platform with increased coastal plain deposits towards the Slørebotn Sub-basin in well 6305/12-1 (Jongepier et al., 1996). On the Frøya High, well 6306/10-1 cored the youngest portion of S2 (Bajocian-Early Callovian) and the Early Callovian lithofacies is interpreted as marginal marine deposits, which represents a raise in the relative sea-level (Jongepier et al., 1996).

Whether the erosion of S2 is a product of later exposure in the mid Late Jurassic rift phase is uncertain and can therefore not be used exclusively as evidence to establish a Bathonian rift phase. However, the drastic change from predominantly continental deposits in the Bajocian-Bathonian to marginal marine in the Early Callovian supports that a tectonic event occurred in the Bathonian. In addition, the syn-sedimentary deformation of the lithofacies found in well 6305/12-2 on the Gossa High and the thickness variations between wells 6305/12-2 and 6305/12-1 are all evidence of extensional movement.

The potential Bathonian rift phase coincides with a late Middle Jurassic extensional phase that affected the North Atlantic region as a result of increased spreading rates in the Central Atlantic in combination with a global sea-level rise (Larsen, 1987; Doré et al., 1997b; Brekke, 2000; Faleide et al., 2010).

5.1.4 mid Late Jurassic rift phase

A mid Late Jurassic rift phase is well documented in the well and seismic data. Seismically, the evidence of a rift phase be observed in the form of rotated fault blocks, growth strata in the hanging wall of active faults (Fig. 1.2) and large displacements between K2 in the hanging wall and the related footwall (Fig. 4.6 and Fig. 4.7). The rotated fault blocks in the Slørebotn Sub-basin were penetrated by well 6205/3-1R and interpretation of the core and dip-meter log data shows that the fault blocks were rotated during the Kimmeridgian (5°), Early Volgian (15-20°) and Middle Volgian (20°) times (Jongepier et al., 1996). In addition, well 6205/3-1R penetrated an 530 meter thick S3 syn-rift unit that can be observed in the seismic data as a wedge-shaped package that truncates against the fault plane of FF5 (Fig. 1.2 and Fig. 4.14).

The fact that none of the wells on the Gossa High (6305/12-1 and 6305/12-2), Frøya High (6306/10-1), Silje High (6204/10-1 and 6205/10-2A) and Møre platform (IKU wells) penetrated S3 sedimentary rocks supports that the mid Late Jurassic rift phase developed large relief between the present-day structural highs and the structurally lower Slørebotn Sub-basin (Fig. 3.5, Fig. 3.6 and Fig. 3.7). Hence, FF1, FF4 and FF5 have been interpreted to be highly active during this time period.

A mid Late Jurassic-Early Cretaceous extensional event is well documented to have affected the Norwegian shelf as a result of North Atlantic rifting (Brekke and Riis, 1987; Blystad et al*.,* 1995; Doré et al 1997b; Swiecicki et al., 1998; Brekke, 2000; Vergara et al., 2001; Osmundsen et al., 2002; Faleide et al., 2010).

5.1.5 Turonian

The fifth phase of tectonic activity is interpreted to have taken place during the Turonian time. No evidence of rifting has been observed in the seismic data and it is therefore believed that the phase was characterized by mainly uplift and vertical movement along the fault planes. The tectonic phase is interpreted to have caused noteworthy uplift of the pre-existing highs, where the Møre platform and Frøya High are interpreted to have been the elements that experience the greatest magnitude of uplift. This is evident by the absence of S6 and presence of S5 sedimentary rocks in well 6306/6-1 and shallow IKU wells on the Frøya High and Møre platform, respectively (Fig. 3.5 and Fig. 3.7). The depositional period during S5 and S6 are believed to have taken place over the entire study area, but later uplift during the Turonian caused erosion of the entire S6 on the Frøya High and Møre platform area. Further evidence of uplift can be observed in the seismic data, where an internal seismic reflector in S7 (Late Coniacian) thins out and onlap against the Møre platform (Fig. 4.7). This suggests that the bordering fault complex (FF5) along the Møre platform was reactivation during deposition of the lower S7. In addition, the present-day seismic configuration of S5 and S6 can be observed to truncate against FF5 in the northeast and FF4 in the north (Fig. 1.2 and Fig. 4.6).

The tectonic phase is interpreted to have affected FF1 that defines the Gossa, Giske and Gnausen highs to a lesser degree compared to FF4 and FF5. This is evident by the presence of the lower portion of S6 on the Gossa High, penetrated by well 6305/12-2 (Fig. 3.5). In the seismic data, S6 reflectors can be seen to overstep the structural highs that are bounded by FF1 (Fig. 1.2, Fig. 4.2 and Fig. 4.3).

The most compatible comparison can be made to previous work done by Sømme and Jackson. (2013) that interpreted a phase of uplift during the Turonian to explain the increased influx of coarse material observed in the Turonian stratigraphy in the Måløy and southern part of the Slørebotn Sub-basin area. On a more regional scale, a potential mid-Cretaceous rift episode is documented in the mid-Norwegian sea area (Doré et al., 1997b; Lundin and Doré, 1997), however the event is still a matter of debate amongst authors (e.g. Færseth and Lien., 2002; Zastrozhnow et al., 2018).

5.1.6 Late Cenomanian-Late Maastrichtian rift phase

The Late Cenomanian-Late Maastrichtian rift phase is evident in both the seismic and well data. The rift phase is interpreted to be the main reason for the development of a regional Late Cretaceous unconformity (K6), which is present over the entire study area. In the well logs, K6 is represented as a gap in the stratigraphic record where the Late Maastrichtian to Early Danian strata is missing (Fig. 3.5, Fig. 3.6 and Fig. 3.7). Seismically, K6 is characterized as a distinct hard-kick reflectors with truncating reflector below and onlap/downlap above (Fig. 1.2 and Fig. 4.5). In addition, the rift phase is interpreted to have created several normal faults that have been assigned to FF2 (Fig. 4.1b and Fig. 4.2), but the rift phase also caused reactivation of blind FF1 (Fig. 4.5). The lower S8 (Danian) seismic reflectors are interpreted to truncate against the fault planes of FF2 and reactivated FF1 (Fig. 4.2 and Fig. 4.5), which implies that the rift phase had its main development during the Late Maastrichtian. The reactivation of older FF1 is very clear in the west above the Gossa High compared to the southwest, where only minor uplift of the underlying FF1 can be observed in the terms of the downlapping lower S8 (Danian) reflectors above the Gnausen High (Fig. 4.3). Although the main phase of deformation is observed in the Late Maastrichtian, evidence of a potential Late Cenomanian tectonic event can be postulated from wells 6205/12-1, 6205/3-1R and 6305/9-1 in the northern part of the sub-basin where a sandy Grindhval Member occupy the uppermost portion of S8 (Fig. 3.5and Fig. 3.6).

The Møre platform during the rift phase experienced extensive uplift and reactivation of FF5 in the north, which is evident by the entire S7 that truncates against the fault plane (Fig. 1.2). In contrast, the middle and southern part of FF5 show evidence of only minor uplift, as S7 oversteps the faults of FF5 and truncates against K7 in the inner part of the platform area (Fig. 3.8, Fig. 4.7 and Fig. 4.18). A similar development is observed for FF4, where S7 reflectors overstep the faults (Fig. 4.6) and well data from the wells 6306/10-1 and 6306/6-1 that targeted the FF4 bounded Frøya High show evidence of the Maastrichtian stratigraphy being eroded (Fig. 3.5 and Fig. 3.7).

A Late Cretaceous rift phase is well documented to have caused uplift of the entire mid-Norwegian margin as a result of the opening of the North Atlantic (Swiecicki et al., 1998; Brekke, 2000; Gjelberg et al., 2001; Gjelberg et al., 2005; Faleide et al., 2010).

5.1.7 Early Eocene compressional phase

The Early Eocene tectonic phase is interpreted to have caused further uplift of the study area, which can be observed in the seismic data to have affected S4-S8 seismic reflectors (Fig. 1.2). The uplift severely affected the Gossa High in the sense that a fault assigned to FF1 changes its displacement from normal to reverse along the strike (Fig. 4.1a). Similar compressional deformation is not observed in the other highs (Ona, Gnausen and Giske highs) that are controlled by FF1, only minor uplift can be interpreted (Fig. 4.2 and Fig. 4.3). The Møre platform and Frøya High are interpreted to have experienced further uplift during the Early Eocene phase that limited the accommodation space for S8 sedimentary deposits over the areas (Fig. 3.4, Fig. 4.6 and Fig. 4.7). Another structural element that was strongly affected by the uplift during the Early Eocene was the Slørebotn Sub-basin. The present-day basin fill of the sub-basin was elevated to a higher level than the basements highs, which are believed to have been in a higher position prior to the Early Eocene tectonic phase (Fig. 1.2 and Fig. 4.2). The architecture of the basin fill changes from concave down in S4 to concave up from the post-S4, thus forming an inverted basin geometry. The proposed interpretation that the Slørebotn Sub-basin, Møre platform and Frøya High were the structural elements that experienced the greatest magnitude of uplift is supported by the absence of Early-Middle S8 (Eocene and Oligocene) strata in the well logs. More specifically, in the northern part of the sub-basin, well 6205/3-1R shows that the entire Eocene and Oligocene stratigraphic record are missing (Fig. 3.5). On the Frøya High (6306/6-1) and Møre platform area (IKU wells) similar development are seen in the wells as in well 6205/3-1R, where the entire Eocene and Oligocene stratigraphy are missing (Fig. 3.5 and Fig. 3.7). Whereas in the southeastern part of the sub-basin, wells 6204/11-1 and 6204/11-2 penetrated Oligocene sedimentary rocks that directly overlie sedimentary rocks of Paleocene in age (Fig. 3.6). This is also the case for wells 6204/10-1 and 6204/10-2A farther south, on the Silje High, where the majority of the Eocene sedimentary rocks are missing and rocks of Oligocene age directly overlies the Paleocene rocks (Fig. 3.6). In contrast, the wells farther west on the Gossa High (6305/12-1 and 6305/12-2) and north for the Gossa High (6305/9-1) penetrated both the Eocene and Early Oligocene sedimentary rocks (Fig. 3.5 and Fig. 3.7).

To conclude, the northern part of the Slørebotn Sub-basin, Frøya High and Møre platform area were severely uplifted during the Early Eocene that caused erosion and bypass of the Eocene and Oligocene sediments, which were deposited farther west. Hence, a shift in accommodation space changed from the Slørebotn Sub-basin to the western bounding highs (Gossa, Ona, Giske and Gnausen highs) that were once elevated relative to the Slørebotn Sub-basin during the Paleocene. The uplift of the southeastern part of the Slørebotn Sub-basin was to a lesser degree than in the north, since the Oligocene sedimentary rocks are present in the wells (e.g 6204/11-1) (Fig. 3.6).

The proposed Early Eocene compressional phase coincides with an Early Eocene rift event that is related to the final break-up between Greenland and Scandinavia (Eldholm and Thiede, 1980; Lundin and Doré, 1997; Brekke, 2000). Furthermore, locally in the Møre Basin, the rift event is expressed as compressional structures rather than extensional faulting (Grunnaleite and Gabrielsen, 1995; Brekke, 2000) and has been seen as one of the main reasons for forming several domes in the mid-Norwegian Sea area (i.e Ormen Lange dome, Helland Hansen Arch, Hedda Dome, Modgunn Arch and Naglfar Dome) (Doré et al., 2008)

5.1.8 Base Pleistocene uplift and glaciation phase

The final phase of tectonic activity is recognized at the uppermost part of S8 (base Pleistocene), where there is a dramatic change in the seismic reflector pattern. The seismic pattern below is characterized as westward dipping reflectors that truncate the bounding K7, whereas above the boundary, the seismic reflectors are uniform and horizontal (Fig. 1.2 and Fig. 4.2). In the well logs, this boundary represents a gap in the Cenozoic record where the majority of the upper S8 (Miocene) stratigraphy is absent (Fig. 3.5, Fig. 3.6 and Fig. 3.7). There is no fault activity observed in the seismic data that is related to this time period, which leads to a potential phase of uplift as the main caused for the erosion. However, uplift alone cannot entirely explain the horizontal undeviated unconformity surface that has formed (Fig. 1.2 and Fig. 4.18). Therefore, another mechanism must have been involved and several authors have described a time of major glaciation (Elsterian Ice Sheet) of the Norwegian mainland during the Plio-Pleistocene (Riss and Fjeldskaar, 1992; Martinsen et al., 1999; Ottesen et al., 2009).

To summarize the entire 5.1 sub-chapter, the eight phases of the different tectonic movements that have been recognized, affected the architecture of the study area to different degrees of deformation and thus, making the area structurally complex. The magnitude of the extension and compression are clearly greater in the northern part of the study area, where fault systems such as FF5 that bound the Møre Platform, FF4 that bound the Frøya High and FF1 that bound Gossa High are prominent features that display evidence of strong tectonic control through most of the main tectonic phases. In contrast, the structural elements in the southwest (Gnausen and Giske highs) that are controlled by FF1 show a smaller degree of deformation, and is believed to have been tectonically quiet during most of the Cretaceous period when the related Gossa High is observed to have been active. Therefore it would be reasonable to conclude that the reason why the northern part of the study area has experienced a greater magnitude of uplift compared to the south, would be because of the structural corner that is being created due to FF1, F3, FF4 and FF5 intersect each other (Fig. 4.1a).

5.1.9 Structural model

To date, there are at least two different structural models for explaining the structural evolution of the study area. The model suggested by Osmundsen and Ebbing. (2008) propose that in the classic lower-plate extensional model by Lister et al. (1986), the Slørebotn Sub-basin would represent as the core complex, Gossa High as the dome-shaped mylonite front and the eastern Møre fault complex (FF5) would act as the detachment plain (Fig. 1.4a). In addition, they further suggests that the distinct seismic reflector below the rotated fault blocks in the Slørebotn Sub-basin represents an early detachment surface in the Volgian that caused rotation of the overlaying Jurassic rocks, which were later incised by FF5 during the Cretaceous (Fig. 1.2).

The proposed model by Jongepier et al. (1996) suggests that the rotation of the fault blocks during the Late Jurassic were due to gravitational collapse of the Møre platform and Gossa High (Fig. 1.4b). Furthermore, a gravitational collapse mechanism would therefore explain the lack of compressional features observed between the Slørebotn Sub-basin and Gossa High as a result of the rotation of the fault blocks.

Although both models seem reliable in explaining the present-day structural features, there are some deviations, such as the abrupt change in the fault geometry between the supposed Slørebotn detachment and FF5 from the Jongepier et al. (1996) model (Fig. 1.4b); and that Osmundsen and Ebbing. (2008) use a lower-plate extensional model to explain the structural evolution of the study area.

The interpretation in this study supports that the study area is a small part of a larger upper-plate extensional system of Lister et al. (1986) (Fig. 5.1) and that the detachment fault is located far below the structural elements, as illustrated in the Lister et al. (1986) model. By comparing the upper-plate model to the architecture of the Møre Basin, it is possible to recognize similar characteristics such as a narrow shelf and a relatively unstructured area (Fig. 2.1). In addition, according to Lister et al. (1986) an upper-plate margin is subject to uplift due to underplating by igneous rocks (Fig. 5.1), which could then explain why the study area has experienced large magnitudes of uplift in the Early Jurassic that caused erosion of almost the entire S1 (Fig. 3.5, Fig. 3.6 and Fig. 3.7). By comparison to the adjacent Vøring Basin, the architecture of the area displays the opposite of what can be observed in the Møre Basin. The Vøring area is characterized by a wide shelf (Halten Terrace/Trøndelag Platform) and the area is highly structured with several rotated fault blocks (Fig. 2.1). By applying the Lister et al. (1986) model, the Vøring area would thus represent a lower-plate margin. In addition, a lower-plate margin would mainly experience subsidence and based on the regional well correlation to the Halten Terrace, very little evidence of erosion in S0, S1 and S2 are evident in well 6406/11-1S (Fig. 3.10). The polarity change from upper-plate to lower-plate in the Møre Basin to Vøring Basin, respectively, would have to be separated by a transfer fault (Fig. 5.2) such as the interpreted F3 (Fig. 1.3).

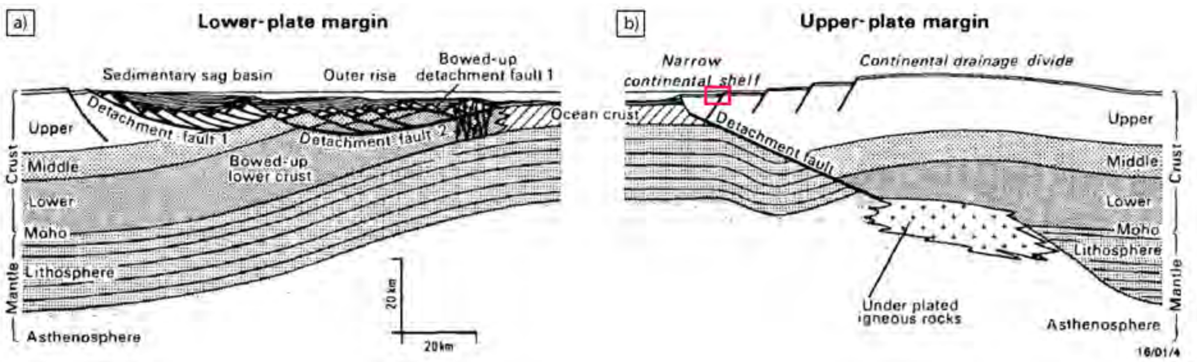


***Fig. 5.1 Detachment-fault model of passive continental margins** a) lower-plate with complex structures and tilted blocks; and b) upper-plate is relatively unstructured with uplift caused by underplating of igneous rocks. The study area (red square) would represent a small part of the upper-plate margin. Modified from Lister et al. (1986).*

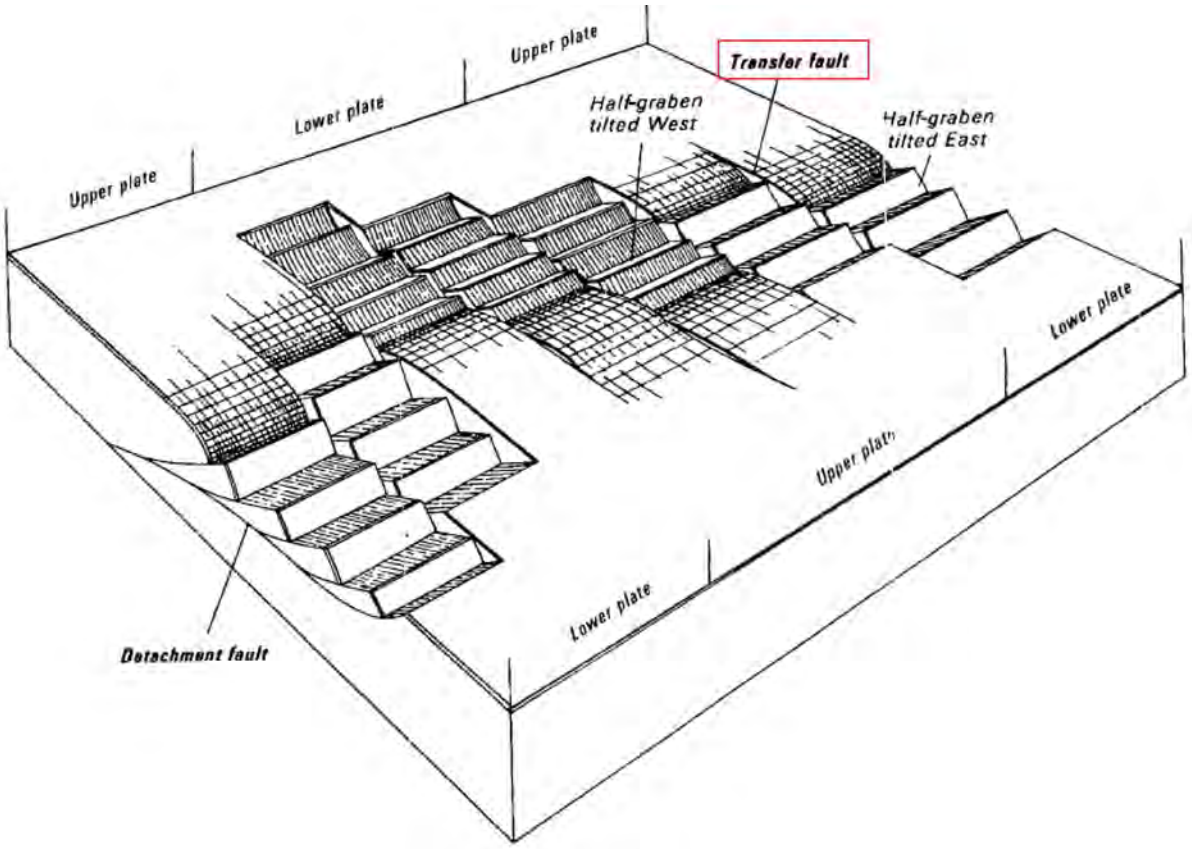
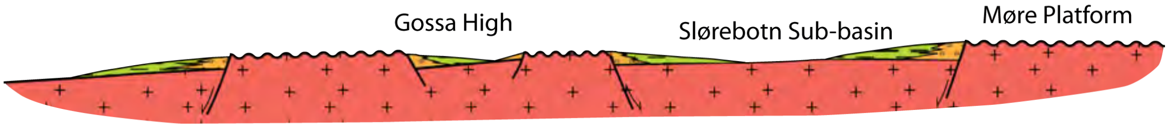


Fig. 5.2 Change from upper- to lower-plate occurs across a transfer fault (red square). *Modified from Lister et al. (1986).*

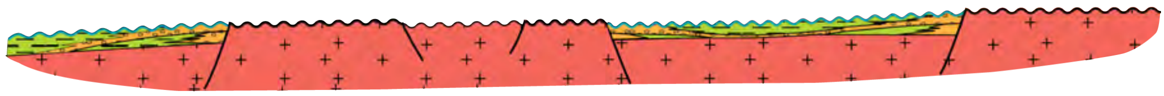
Several authors have recognized a polarity change across F3 from the Møre to Vøring basins, however there are disagreements on whether the Møre Basin is characterized as an upper- or lower-plate margin (Torkse and Prestvik, 1991; Gjelberg et al., 2005; Mosar et al., 2002a; Mosar et al., 2002b; Osmundsen and Ebbing, 2008).

The tectono-stratigraphic model that is proposed in Fig. 5.3, Fig. 5.4 and Fig. 5.5 below is only representing the evolution of the area in a small-scale. The "large-scale" evolution of the entire Møre Basin area as an upper-plate margin has not been addressed as this is out of the scope for this study. In addition, the proposed detachment surface that underlay Upper Jurassic fault blocks in the Slørebotn Sub-basin by Jongepier et al. (1996) and Osmundsen and Ebbing. (2008) have not been successfully modeled in this study. However, seen as these highly rotated fault blocks are only recognized in the northern part of the Slørebotn Sub-basin, the mechanism responsible for the deformation would therefore be a local feature that is constrained within the northern part of the sub-basin.

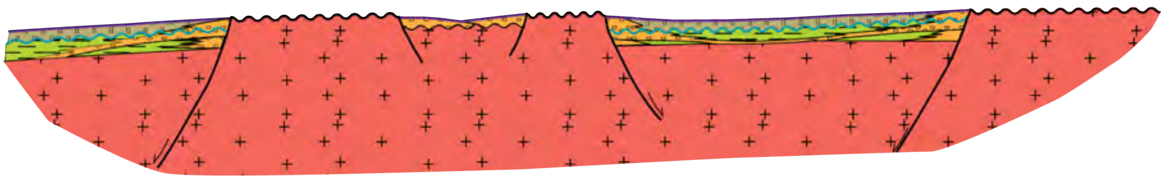
S0 (Carnian-Rhaetian):
Early/Middle Triassic extensional rifting developed
low-lying areas that were filled in by continental deposits.



S1 (Hettangian-Toarcian):
A late Early Jurassic phase of uplift caused erosion of the Early Jurassic
and portions of the Upper Triassic strata on the structural highs.



S2 (Aalenian-Callovian):
During the Bathonian, coastal plain deposits dominated the Slørebotn Sub-basin, whereas
an alluvial environment was still present at the Gossa High and Møre platform. Marginal
marine environment was first introduced in the Callovian as a result of the Bathonian rift phase.



S3 (Oxfordian-Early Ryazanian):
The main mid Late Jurassic extensional event caused rotation and erosion of the fault blocks.
Anoxic basin conditions prevailed as a result of the rift architecture.

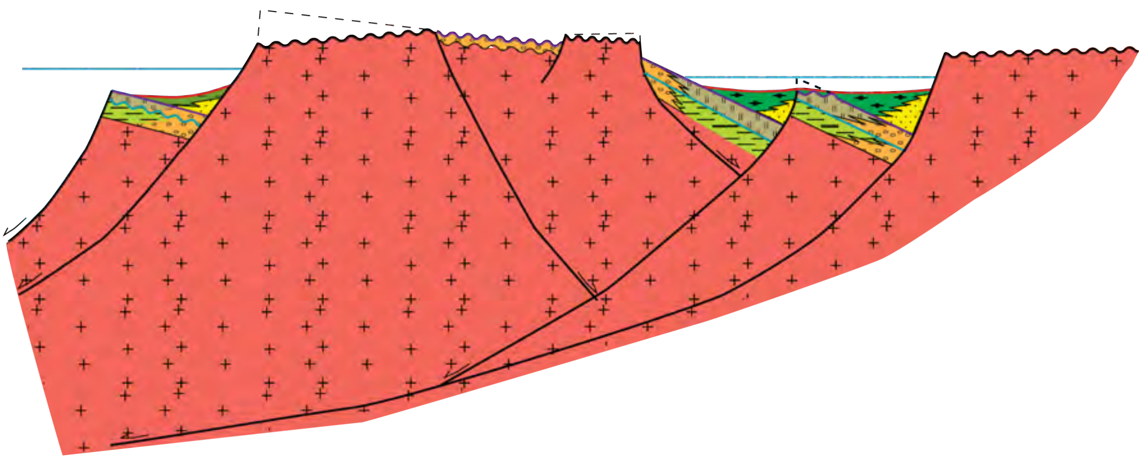
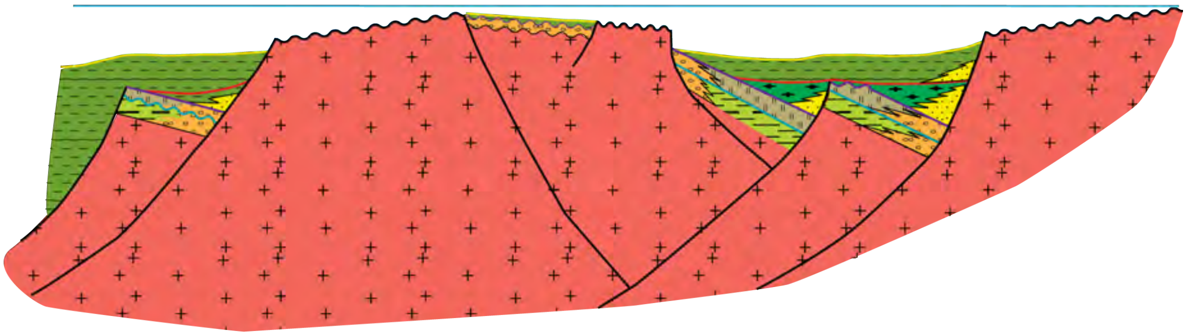
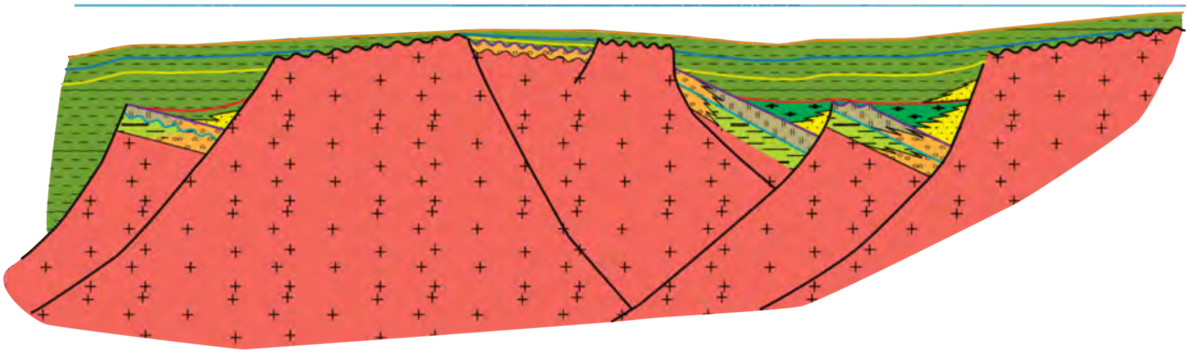


Fig. 5.3 Tectono-stratigraphic evolution for S0-S3 in the northern part of the study area. *The legend is located in Fig. 5.6*

S4 (Late Ryazanian-Late Albian):
Early Cretaceous transgression flooded the entire study area that resulted in deposition of deep marine sediments onto the highs and in the basin areas, but now in oxic conditions. The time period was not successful in smoothening the rift topography.



S5 & S6 (Cenomanian-Late Turonian):
The early Late Cretaceous was a period of quiescence and deposition of open marine sediments that eventually smoothed the relief.



Turonian uplift and erosion:
The structural highs in the northern part of the study area experienced uplift and erosion of S6 during this phase. This led to increased influx of coarse clastics into the sub-basin.

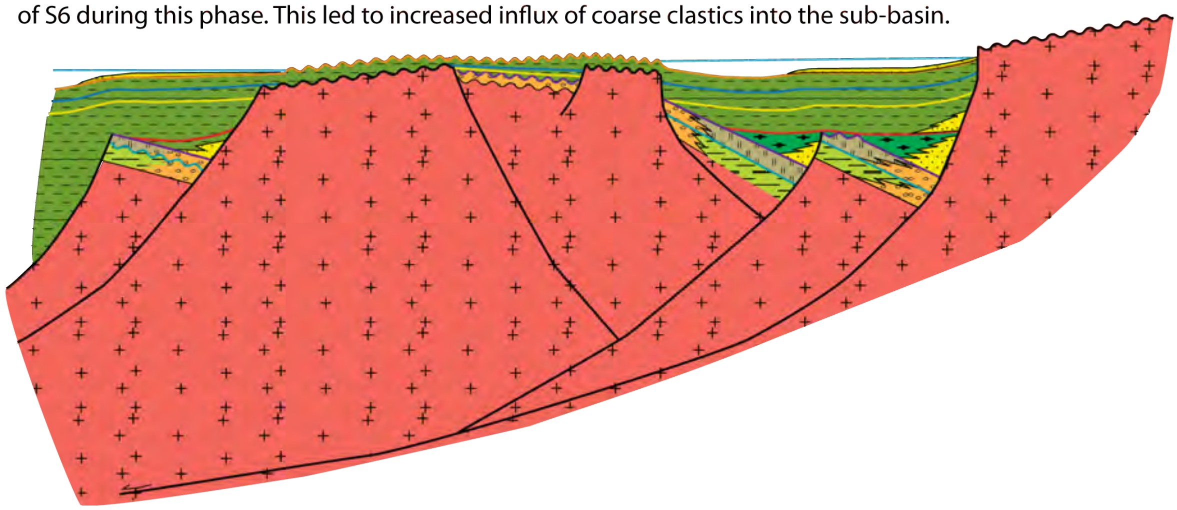
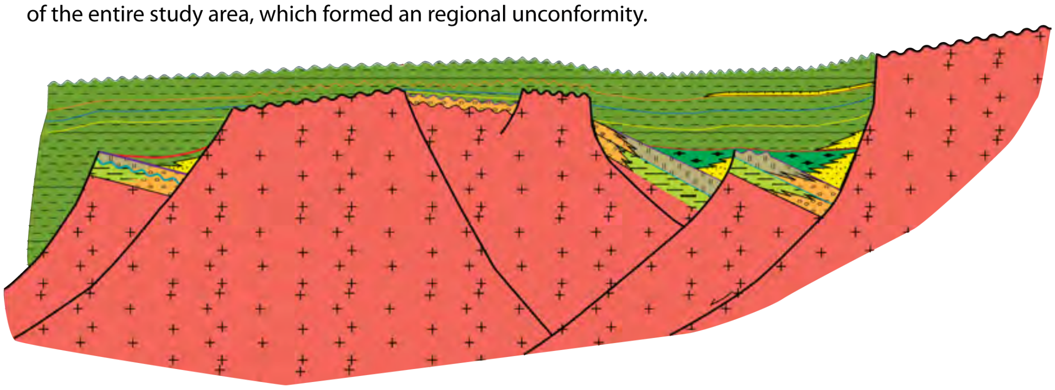
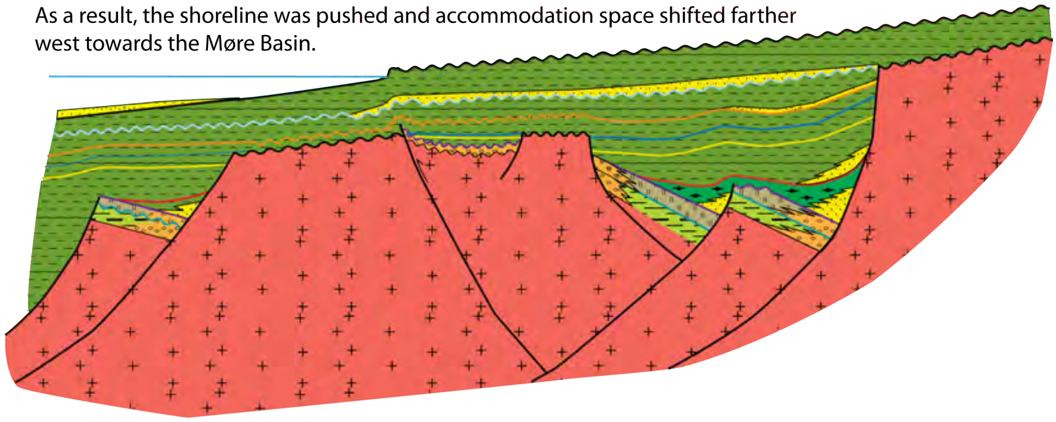


Fig. 5.4 Tectono-stratigraphic evolution for S4-S6 in the northern part of the study area. *The legend is located in Fig. 5.6*

Late Cenomanian-Late Maastrichtian rift phase:
The rift phase caused reactivation of older faults, uplift and exposure of the entire study area, which formed an regional unconformity.



S8 (Early Eocene uplift phase):
Rifting along the North Atlantic caused compression in the form of inversion of the Gossa High and inverted basin fill of the Slørebotn Sub-basin. As a result, the shoreline was pushed and accommodation space shifted farther west towards the Møre Basin.



S8 (Eocene-Lower Pleistocene):
A rise in the relative sea level in combination with glacial erosion of the Norwegian mainland resulted in large amounts of sediments being transported and deposited in the Møre area.

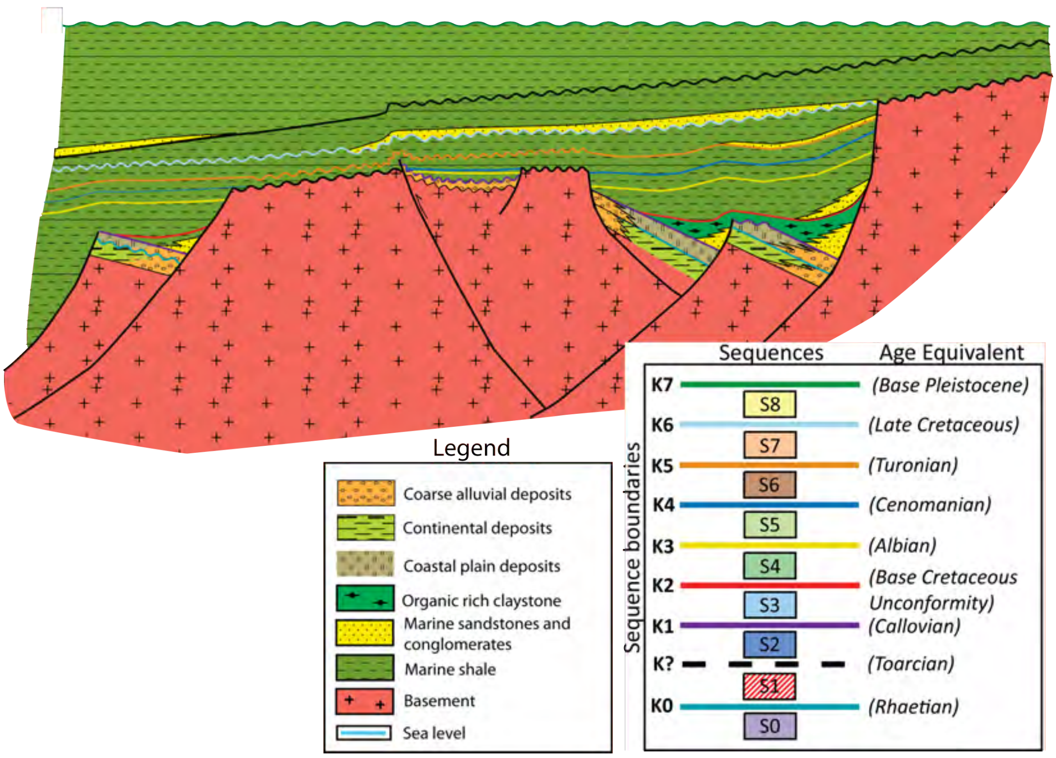


Fig. 5.5 Tectono-stratigraphic evolution for S7-S8 in the northern part of the study area.

5.1.10 Tectonic control on the deposition

The influence of the different fault families clearly affected the deposition of the Mesozoic to Cenozoic sedimentary rocks. FF1, FF4, and FF5 might have formed structural relief between the Slørebotn Sub-basin and the present-day basement highs as early as in the Early/Middle Triassic (Fig. 5.3), but it was not until the mid Late Jurassic extensional event that the area evolved into several rotated high and low-lying structural features (Fig. 4.6 and Fig. 4.7) (Jongepier et al., 1996). During this time period, the Slørebotn formed into a small and narrow Jurassic sub-basin (Fig. 5.3), whereas the structural highs had high relief that sourced and controlled the deposition of S3 syn-rift and thick S4 post-rift sediments, which is evident by the onlap reflectors of the sequences onto the highs (Fig. 4.6 and Fig. 4.7). In later stages, FF4 and FF5 is interpreted to be the most prominent fault complexes to control the sedimentations of S5, S6 and S7, in contrast to FF1, where the majority of the related faults have been quiet until the Late Maastrichtian rift event. The throw analysis of FF5 highly implies that the fault complex had its main rift centers in the northern and middle part of the study area with fault tip progradation in the NE and SW directions (Fig. 4.4b and Fig. 4.9). The importance of the potential drainage routes that this fault complex would have created in the Early Cretaceous and whether the paleovalleys were connected to the Norwegian landmasses is uncertain. However, the core data from well 6205/3-1R of the coarse-grained slope apron fans in the lower S4 implies a short distance from the source area and a connection of the mainland in the east is therefore unlikely (Jongepier et al., 1996; Gjelberg et al., 2005; Mørk and Johnsen, 2005).

A connection to the mainland was probably not established until the early Late Cretaceous when the study area experienced an increase of coarse clastics in the Slørebotn Sub-basin. Furthermore, previous authors (Sømme and Jackson, 2013) have postulated a connection to the Norwegian landmasses as a source province through a series of canyon systems along the Møre platform that fed coarse Turonian submarine fans into northern North Sea and southern part of the Slørebotn Sub-basin. These observations are in an agreement to the findings in this study, since the fine-grained and well sorted submarine fans that were cored in S6 and S7 reflects a longer transportation way compared to the poorly sorted slope apron fan cored in S4 (Fig. 4.15). In addition, serval incised valleys that originated in the earliest Cretaceous are interpreted on the Møre platform area and the continued reactivation and exposure of the platform area during the Turonian tectonic phase (5.1.5 Turonian) would result in further erosion and transportation along these potential palaeovalleys (Fig. 4.16).

There is an agreement amongst authors that a correlation between the study area and the Norwegian mainland as a catchment area was established during the Late Cretaceous-Early Paleocene (Gjelberg et al., 2001; Gjelberg et al., 2005; Sømme et al., 2009; Ravnås et al., 2014a). Uplift and erosion of the Norwegian mainland due to Late Cretaceous rifting is believed to be the main reason for the influx of the Danian turbidites into the study area. More specifically, it is proposed that the coarse Late Cretaceous-Early Paleocene sediments were transported along the present-day NE-SW orientated fjords along the coast of Norway and that their orientation favored sediment sink near the mouth of the present-day Romsdalsfjord (Fig. 5.6) (Gjelberg et at., 2005; Sømme et al., 2009). The source point of where the sediments were transported into the Møre Basin area coincides with the postulated onshore extension of F3, which suggests that the lineament created a structural weak zone for allowing input of coarse sediments into the study area during the Late Cretaceous-Early Paleocene (Fig. 5.6) (Gjelberg et al., 2001; Gjelberg et at., 2005; Sømme et al., 2009; Ravnås et al., 2014a). The NE-SW orientated fjords reflect similar trend as the interpreted NE-SW trending FF1 and FF5 in the Slørebotn Sub-basin, which implies a potential onshore extension of the faults (Fig. 5.6). The fault families are interpreted as a part of the Møre-Trøndelag Fault Complex (Blystad et al., 1995) and several authors have recognized the offshore-onshore extension of the Møre-Trøndelag Fault Complex, but the exact link remains obscure due to the lack of seismic data close to the Norwegian coastline (Gabrielsen et al., 1999; Brekke, 2000;

Gjelberg et al., 2005; Nasuti et al., 2012). To summarize, the onshore development of the Møre-Trøndelag Fault Complex (FF1 and FF5) was critical for driving sediment supply to the main source point at the mouth of the present-day Romsdalfjord. The structural weak zone created by F3 allowed coarse sediments to be transported into the study area.

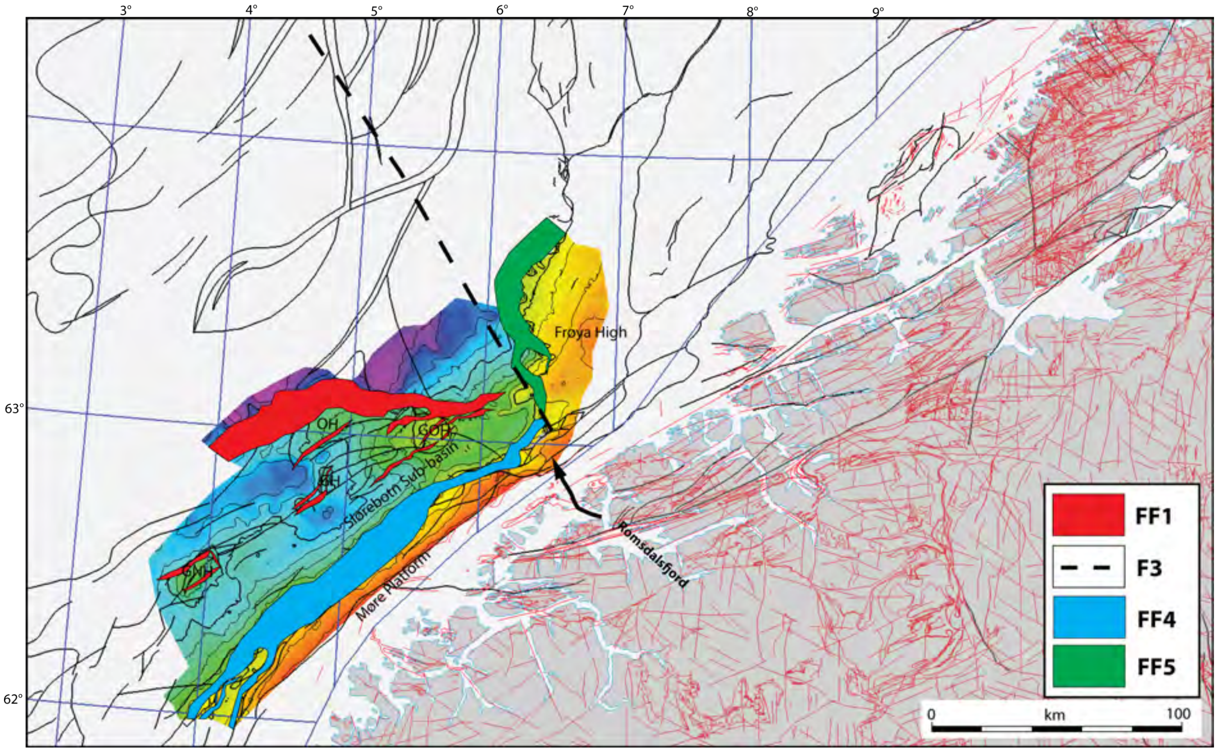


Fig. 5.6 Displaying the potential offshore-onshore extension of the fault families that are interpreted in the Slørebotn Sub-basin. *The black arrow represents the transportation route for the coarse Late Cretaceous-Early Paleocene sediments into the Møre area. The tectonic lineaments (red lines) of the Norwegian mainland (grey area) are taken from Gabrielsen et al. (2002).*

Further evidence of F3 forming a zone of structural weakness is indicated by the presence of two volcanic centers that are located in the striking zone of F3 (Fig. 5.7), but in the opposite side for each other, i.e. on the Frøya High and eastern Greenland (Torske and Prestvik, 1991). In addition, several Eocene domes are developed along the strike of F3 (Fig. 5.8), which shows the importance of this NW-SE trending transfer fault (Doré et al., 2008).

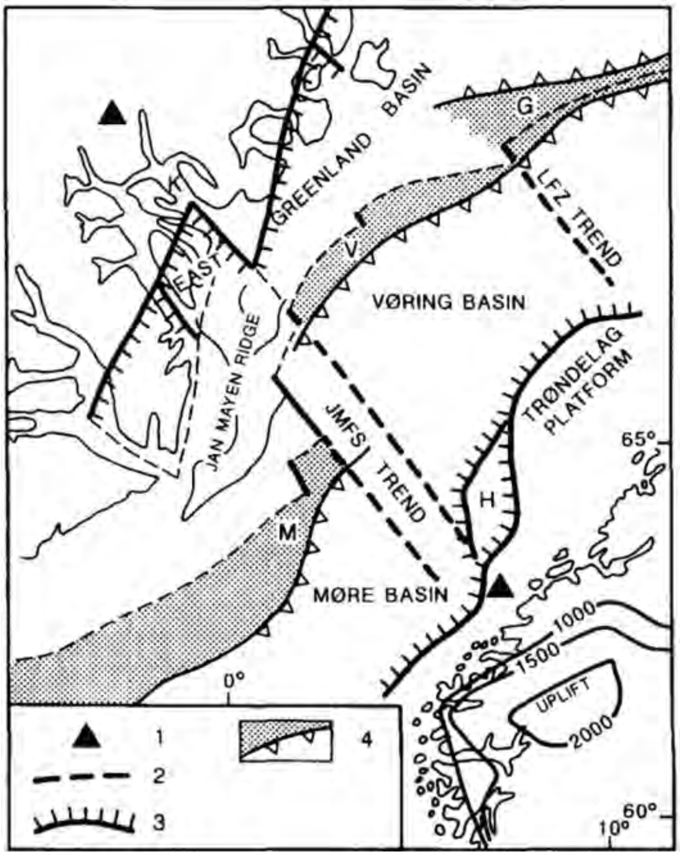


Fig. 5.7 Cartoon illustrating the strike zone of the Jan Mayen Zone (F3) between the mid-Norwegian Sea and East Greenland. *1=volcanic centers, 2=facture-zone trends in continental crust, 3=basin-margin normal faults, 4=marginal highs, JMFS=Jan Mayen Fracture Zone. Modified from Torske and Prestvik. (1991).*

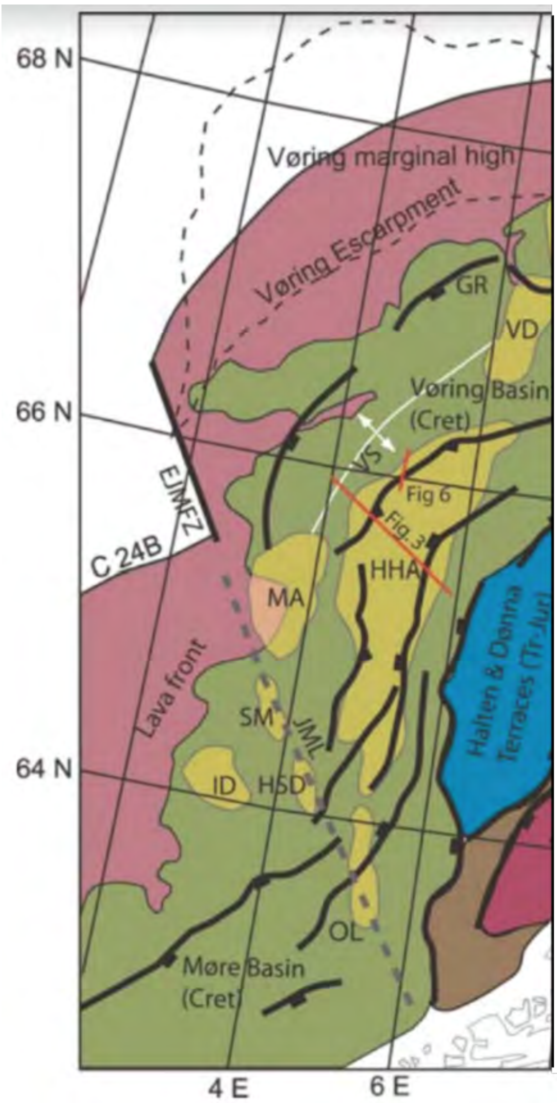


Fig. 5.8 Structural map of the mid-Norwegian Sea illustrating the inversion features along the Jan Mayen Fracture Zone. *EJMFZ=Jan Mayen Fracture Zone, ID= Isak Dome, SM=Souther Modgunn Arch, HSD=Havsule Dome, OL=Ormen Lange Dome. Modified from Doré et al. (2008).*

The effect that F3 might have had on earlier stages are evident between the northeastern end of the Gossa High and the southern end of the Frøya High (Fig. 4.6). At this location, the basement topography of the Gossa High plunges down in the north as the throw of the bounding FF1 decreases to zero (Fig. 4.1a and Fig. 4.4a). It is proposed that a suitable outcrop analogue to the study area is the onshore southern margin of the Gulf of Corinth, Greece. The region is interpreted by some authors (Zhong et al., 2018) to comprise of a system of NNE-SSW striking transfer faults that delineate several intersecting normal faults. The location of these transfer faults is characterized as a weakened zone where large river valleys occupy the area (Fig. 5.9).

Fig. 5.9 Drone picture of an interpreted transfer fault (dashed line) taken onshore southern margin of the Gulf of Corinth, Greece. *Modified from Zhong et al. (2018).*

To conclude, the location of F3 during the pre-Cretaceous was most likely characterized by a deep eroding river valley, which could explain the sudden plunge of the Gossa High basement topography in the north (Fig. 4.6). During the Cretaceous, the area was transgressed and filled in by thick S4-S7 open marine deposits and apparently, the transfer fault was reactivated during the Late Cretaceous rift episode that formed a structural weak zone for coarse Late Cretaceous-Early Paleocene sediments into the study area (Gjelberg et al., 2005; Sømme et al., 2009). The zone of structural weakness is believed to have appeared again, in the Early Eocene, which caused widespread deposition of volcanic rocks on the Frøya High (Fig. 4.16) (Torske and Prestvik, 1991). Furthermore, in the mid-Norwegian Sea, several domes are documented to have formed in the Eocene (Fig. 5.8) along the striking zone of F3 (Doré et al., 2008), which further implies the importance of the lineament during the Eocene.

5.2 Stratigraphic evolution

The overall tectono-stratigraphic framework of the study area is summarized in Fig. 3.2, and as mentioned earlier, the age framework is based on the Ichron 2015 biostratigraphic well report. The following chapter will discuss the controlling factors of the sedimentary infill and the depositional environment for each sequence that has been defined in this study.

5.2.1 Sequence 0 (Carnian-Rhaetian)

During the Late Triassic time, the study area was in a continental setting and the cored S0 lithofacies on the Gossa High (6305/12-1) and Møre Platform (shallow IKU wells) suggests an arid alluvial fan environment (Smelror et al., 1994; Jongepier et al., 1996). Well 6204/11-1 in the southeastern flank of the Slørebotn Sub-basin show similar fining upward GR log pattern in S0 as in the 6305/12-1 well on the Gossa High (Fig. 4.11). Hence, an arid alluvial depositional environment is interpreted to be the dominant depositional environment during the Early Triassic time. By comparison, to the northern North Sea in the Horda platform area (Fig. 1.1), the Rhaetian sequence is interpreted as alluvial plain deposits and increased subsidence of the area during the middle Sinemurian resulted in establishing marine conditions

(Færseth and Ravnås, 1998; Hesthammer et al., 1999; Ravnås et al., 2000; Faleide et al., 2010). The depositional environment on the Halten Terrace during the Rhaetian is interpreted to range from Coastal plain to delta plain deposits (Dalland et al., 1988; Corfield and Sharp, 2000).

The absence of pre-S0 rocks on the Gossa High (6305/12-2) and Frøya High (6306/10-1) indicates that the highs were exposed and eroded during the Early/Middle Triassic times. The mineral content of S0 shows a signature of the Gossa High basement and from the Møre coast in the east (Mørk and Johnsen, 2005). Thus, S0 was probably sourced from the west and east (Fig. 4.12) and deposition occurred down flank of the highs that accumulated a thick succession in the Slørebotn Sub-basin. This is suggested by the 250 meters thick package of S0 that was penetrated by 6305/12-1 (Fig. 4.11) and based on the seismic interpretation, the thickness of the sequence increases towards the Slørebotn Sub-basin (Fig. 1.2). The lithofacies of S0 comprises conglomerates and coarse sandstones that are texturally and mineralogy immature, which further indicates that S0 was locally sourced (Jongepier et al., 1996; Mørk and Johnsen, 2005). The reservoir potential in S0 has proven to be very poor due to the immature texture of the coarse clastic sediments, high mud content, and quartz cementation that further reduces the reservoir properties (Mørk and Johnsen, 2005). The potential of S0 as a reservoir target is low; nevertheless, a continental depositional environment is likely to have a wide distribution of facies that could potentially form excellent reservoirs (Fagerland, 1990).

5.2.2 Sequence 1 (Hettangian-Toarcian)

S1 sedimentary rocks are largely absent in the study area, and have only been encountered on the Møre platform area (Smelror et al., 1994). Similar interpretation of the arid alluvial lithofacies in S0 is interpreted for S1 due to the close assembly in the lithofacies (Smelror et al., 1994; Jongepier et al., 1996). In the northern North Sea and Halten Terrace, fluvial and increased marine conditions have been reported in the mid Early Jurassic sedimentary rocks (Dalland et al., 1988; Færseth and Ravnås, 1998; Hesthammer et al., 1999; Corfield and Sharp, 2000; Ravnås et al., 2000; Faleide et al., 2010). However, based on the fact that the overlaying S2 comprises alluvial deposits (Jongepier et al., 1996) and that the study area show clear evidence of substantial uplift during the Early Jurassic compared to the northern North Sea and Halten Terrace (Fig. 3.10). Therefore, a continental depositional environment most likely dominated the entire S1.

5.2.3 Sequence 2 (Aalenian-Callovian)

The cored S2 (Bajocian-Bathonian) on the Gossa High (6305/12-2) and Møre platform (IKU wells) mainly comprises coarse alluvial fan deposits that contains a basement signature from both the Gossa High and the Møre coast (Smelror et al., 1994; Jongepier et al., 1996; Mørk and Johnsen, 2005). On the eastern flank of the Gossa High, well 6305/12-1 penetrated S2 (Bajocian-Bathonian) coastal plain rocks (Jongepier et al., 1996) that have been interpreted in the seismic data to increase in thickness towards the Slørebotn Sub-basin (Fig. 1.2 and Fig. 3.5). In addition, the thickness of S2 in wells 6305/12-2 and 6305/12-1 changes from 181.5 meters to 377 meters, respectively (Fig. 4.11). This indicates the Slørebotn Sub-basin was structurally lower in the Middle Jurassic that accumulated thick successions of S2 sediments (Fig. 5.3).

In the Bajocian-Bathonian period, alluvial fan deposits dominated the Gossa High and Møre platform, but now with a humid coastal plain environment farther east in the Slørebotn Sub-basin (Jongepier et al., 1996). In the north, on the Frøya High, well 6306/10-1 encountered the youngest (Early Callovian) rocks of S2 that are interpreted as marginal marine deposits, which differ from the predominantly continental lithofacies in the lower S2 (Bajocian-Bathonian) rocks penetrated on the Møre platform and Gossa High (Jongepier et al., 1996). The obvious increase of the relative sea level from the Bathonian to Callovian is explained by a Bathonian extensional phase that introduced marginal marine conditions in the study area. Whether the relative sea level increased further in the Middle and Late Callovian is uncertain since sedimentary rocks of this age have not been penetrated, but it is believed that the relative sea level continued to increase, which eventually led to deposition of the Late Jurassic deep marine claystones (Dalland et al., 1998). The northern North Sea during the Middle Jurassic was dominated by deltaic to shallow marine deposits, whereas on the Halten Terrace, tidal to increasingly open-marine deposits dominate (Dalland et al., 1988; Færseth and Ravnås, 1998; Hesthammer et al., 1999; Corfield and Sharp, 2000; Ravnås et al., 2000).

To summarize, the Gossa High was still a prominent feature that sourced S2 alluvial fans together with the Møre coast area in the east during the Middle Jurassic (Fig. 4.12), but now with a humid coastal plain environment in a structurally lower Slørebotn Sub-basin (Jongepier et al., 1996). During deposition of the upper S8 (Callovian), the study area experienced a substantial increase in the relative sea level that introduced a marginal marine environment and such an environment may have dominated large portions of the study area.

5.2.4 Sequence 3 (Oxfordian-Early Ryazanian)

The raise in the relative sea level in the Middle Jurassic continued into the Late Jurassic and it is well documented that a regional transgression took place in the North Atlantic rift system during the Late Jurassic time (Doré, 1991; Faleide et al., 2010). The mid Late Jurassic rift episode caused further deepening of the Slørebotn Sub-basin and subsequent deposition of the deep marine S3 claystones. The cored S3 in well 6205/3-1R shows that the claystones contain high levels of TOC and thus the deposition took place in an anoxic environment (Jongepier et al., 1996). In the northern North Sea, similar deposition took place of the equivalent Draupne Formation in overdeepened basins with poor bottom water circulation that formed as a result of the Late Jurassic rift events (Færseth and Ravnås, 1998; Faleide et al., 2010).

The lithofacies of S3 has been widely encountered in the study area (Fig. 3.5, Fig. 3.6 and Fig. 3.7). The core and well data suggests that the organic-rich claystone is interbedded with thick intervals of coarse- to fine-grained turbidite sandstones (Fig. 4.14) (Jongepier et al., 1996). The deposition of S3 is believed to have mainly accumulated in the Slørebotn Sub-basin, while non-deposition took place on the structural highs. This is due to the lack of S3 sedimentary units on the Gossa (6305/12-1 and 6305-12-1), Frøya (6306/10-1), and Silje (6204/10-2A) highs (Fig. 3.5 and Fig. 3.6). The coarse clastic units that are interbedded in the fine-grained organic material is interpreted as gravitational flow deposits that have been laid down from the slope of rotated fault blocks during tectonic movements in the mid-Late Jurassic (Jongepier et al., 1996). Similar coarse gravitational deposits are documented in both the northern North Sea and on the Halten Terrace (Swiecicki et al., 1998; Ravnås et al., 2000). In addition, a Late Jurassic play has been proven to work on the Frøya High (Draugen Field) where shallow-marine bar sands of the Rogn Formation were found hydrocarbon bearing (Spencer et al., 1986). Farther north, on the Halten Terrace, the Pil and Bue discovery proved hydrocarbons in Late Jurassic sandstones of the intra Melke and Rogn formations (NPD factpages).

5.2.5 Sequence 4 (Ryazanian-Late Albian)

The Early Cretaceous was followed by a period of thermal subsidence and post-rift infill of the Late Jurassic rift topography, which is reflected by the divergent S4 reflector pattern that can be observed to onlap against the basement highs (Fig. 1.2, Fig. 4.3 and Fig. 4.18). S4 comprises mainly of open marine mudstones and siltstones with an exception in its basal part that comprise coarse clastic deposits that can typically be recognized in the GR logs as a low GR interval with a sharp base above K2 (Fig. 4.10

and Fig. 4.14) (Jongepier et al., 1996; Vergara et al., 2001; Martinsen et al., 2005). The core data in well 6205/3-1R in the lower S4 shows that the coarse interval comprises angular and poorly sorted conglomerates and sandstones that have been interpreted as debris flow deposits (Jongepier et al., 1996; Martinsen et al., 2005). The seismic facies signature of this coarse interval show a fan shaped geometry (Fig. 1.2) (Table 4.1;F4) and similar seismic signature has been interpreted throughout the Slørebotn Sub-basin to follow the topography of K2 (Fig. 4.2, Fig. 4.3 and Fig. 4.16). At the crest of the Frøya High and Møre platform that surround the sub-basin, several incised valleys have been interpreted to cut deep down into the basement (Fig. 4.16) (Table 4.1;F1). The location of these valleys would be important pathways for coarse clastics into the sub-basin and the interpreted fan deposits coincide with the location of the incised valleys (Fig. 4.16). In addition, the fault throw analysis of FF5 suggests two drainage pathways that could potentially have been established in the Early Cretaceous prior to the present-day hard-linkage of the fault complex (Fig. 4.4a and Fig. 4.9). This would then explain the wide distribution of the interpreted fan deposits in the northern part of the study area (Fig. 4.17). However, a connection to the Norwegian mainland as a potential source province for the coarse debris flow deposits is unlikely due to the poorly sorted and angular clasts observed in the cored debris flow unit in well 6205/3-1R (Fig. 4.15a).

The reservoir quality of the cored S4 debris flow deposits in the northern part of the Slørebotn Sub-basin (6205/3-1R) have shown to be quite poor due to the immature texture and the considerable quartz cementation (Jongepier et al., 1996; Mørk and Johnsen, 2005). Similar debris flow deposits might be expected farther north and south of well 6205/3-1R, but at these locations the fans lie at a depth below 5000 meters (Fig. 4.17), which makes them unattractive for exploration (Maast et al., 2010). In contrast, in the southeastern part of the study area, coarse debris flow deposits of the Agat Member are found on a depth of approximately 2600 meters above K2 in wells 6204/10-1 and 6204/11-2 (Fig. 3.6 and Fig. 4.10), and seismic interpretation suggests that these fans occupy large portions of the graben that is formed between the Silje High and Møre platform (Fig. 4.17 and Fig. 4.18). Based on the NPD final well-report for 6204/10-1, the Agat Member comprises predominantly of poorly sorted conglomerates with angular clasts and layers of sandstone and claystone, and this shows a very similar characteristic to that found in well 6205/3-1R.

By analogy to the northern North Sea, the Agat discovery (Fig. 1.1) has proven the Early Cretaceous turbidite play where the Agat Member comprises excellent reservoirs properties with a thickness of up to 274 meters (Vergara et al., 2001).

5.2.6 Sequence 5 (Cenomanian)

A further raise in the relative sea level occurred during the Cenomanian, which led to widespread deposition of the open marine shale-dominated S5 (Swiecicki et al., 1998; Vergara et al., 2001). Although the present-day seismic configuration of S5 can be observed to truncate against the Møre platform, Gossa High and Frøya High (Fig. 4.6 and Fig. 4.7), the deposition of S5 is interpreted as a period where the highs were submerged. This is evident by the thin section of S5 found in well 6305/12-2 on the Gossa High, in well 6306/6-1 on the Frøya High and in the shallow IKU wells (Fig. 3.6 and Fig. 4.10). Thus, a conclusion can be drawn that the structural relief that formed in the mid Late Jurassic tectonic event was successfully smoothed out during the Cenomanian.

An expectation from the predominantly open marine deposition can be observed in the boundary between S5 and S6, where a sub-anoxic shale of the Blodøks Formation have been penetrated by three wells in the south (6204/10-1, 6204/11-1 and 6204/11-2) and one well on the flank of the Gossa High (6305/12-1). The geochemical analysis from well 6305/12-1 shows that the Blodøks Formation has a of TOC of 2.2% (NPD factpages) and in the GR logs, the unit is characterized as an interval of high readings (Fig. 4.10).

This implies that at least parts of the study area in the southeast and north developed into a sub-anoxic environment. Although the Blodøks Formation has been proven to be very thin (approximately 20 meters), the exploration wells that penetrated the formation are drilled on the flank of the Slørebotn Sub-basin, and it can therefore be expected to find intervals that are better developed within the sub-basin. The importance of a Cretaceous source rock in the study area could be significant since the Late Jurassic Spekk Formation (the upper part of S3) tends to be buried too deep (Fagerland, 1990; Swiecicki et al., 1998). It is suggested that a Cretaceous source rock(s) in the mid-Norwegian Sea has played an important role in charging reservoirs (Ormen Lange and Ellida) where the Late Jurassic source rock is over-mature (Garner et al., 2017).

By comparison, to the northern North Sea, the Blodøks Formation is documented to have been deposited in anoxic conditions and the formation can reach a thickness of up to 120 meters (Surlyk et al., 2003).

5.2.7 Sequence 6 (Turonian)

Similar to S5, S6 depositional period also took place over submerged highs in an open marine environment (Swiecicki et al., 1998; Vergara et al., 2001). The main difference can be seen by the increased influx of coarser clastics (Rødspette and Skolest members) in the sequence, that were initiated by a period of uplift of the pre-existing basement highs during the Turonian (Fig. 3.5, Fig. 3.6 and Fig. 3.7). The erosional evidence of S6 in the wells on the Gossa (6305/12-2), Frøya (6306/10-1 and 6306/6-1) and Silje (6204/10-2A) highs further supports a period of uplift. The Skolest and Rødspette members are interpreted to have their thickest development in the north (6205/3-1R) and southeastern (6204/11-2) parts of the Slørebotn Sub-basin, respectively (Fig. 3.5 and Fig. 3.6). In the seismic data, these sandstone intervals are characterized as high amplitude discontinuous reflectors that are interpreted as submarine fans (Table 4.1;F8). The location of the fans along the western margin of the Møre platfrom suggests multiple entry points (Fig. 4.16), which coincides well with the interpretations of the potential sedimentation routes along FF5 and the location of the incised valleys on the Møre platform area (Fig. 4.9 and Fig. 4.16). This implies that the fault complex (FF5) along the eastern margin was not successful in developing the present-day hard-linkage and that sedimentation in the early Late Cretaceous was strongly fault controlled.

5.2.8 Sequence 7 (Late Turonian-Early Maastrichtian)

S7 is interpreted to comprise predominantly open marine shale deposits (Swiecicki et al., 1998; Vergara et al., 2001) with sandstone units of the Intra Kyrre, Rødspette and Tunge members at the base and a Grindhval Member in the uppermost part (Fig. 3.5, Fig. 3.6 and Fig. 3.7). The increased influx of coarse material at the base of S7 in an otherwise shale-prone succession is interpreted to be a product of the uplift in the Turonian (5.1.5 Turonian). The uplift caused relief between the present-day structural highs and the Slørebotn Sub-basin that accumulated coarse material. The relief was eventually smoothed in the Early Santonian, which is evident in the seismic by the upper S7 (Santonian-Maastrichtian) reflector that oversteps the Møre platform and Frøya High (Fig. 4.7).

Similar as S6, the interpreted submarine fans in S7 also suggests multiple entry points along the Møre platform (Fig. 4.16). The core interpretation from the Rødspette Member in wells 6204/11-1 and 6204/10-1 is described as fine- to coarse-grained, moderate sorted sandstones, which suggests that the sediments were transported some distance (NPD factpages), in contrast to the poorly sorted and locally derived conglomerates penetrated by well 6205/3-1R in the lower S4 (Fig. 4.15a) (Jongepier et al., 1996; Mørk and Johnsen, 2005). Hence, a source area farther east on the Møre platform, or even the Norwegian mainland along the fjords can be postulated to have sourced the coarse early Late Cretaceous sediments. The depositional mechanism for the coarse clastics is still gravitational, but now with a longer transportation way as pontential submarine fans. These observations is supported by previous work done by Sømme and Jackson. (2013) in the southern part of the Slørebotn Sub-basin. They suggests that parts of the Norwegian mainland was uplifted during the Turonian and a series of feeder canyons developed along the western Norwegian margin that fed coarse sediments into the basins in the west (Fig. 5.10).

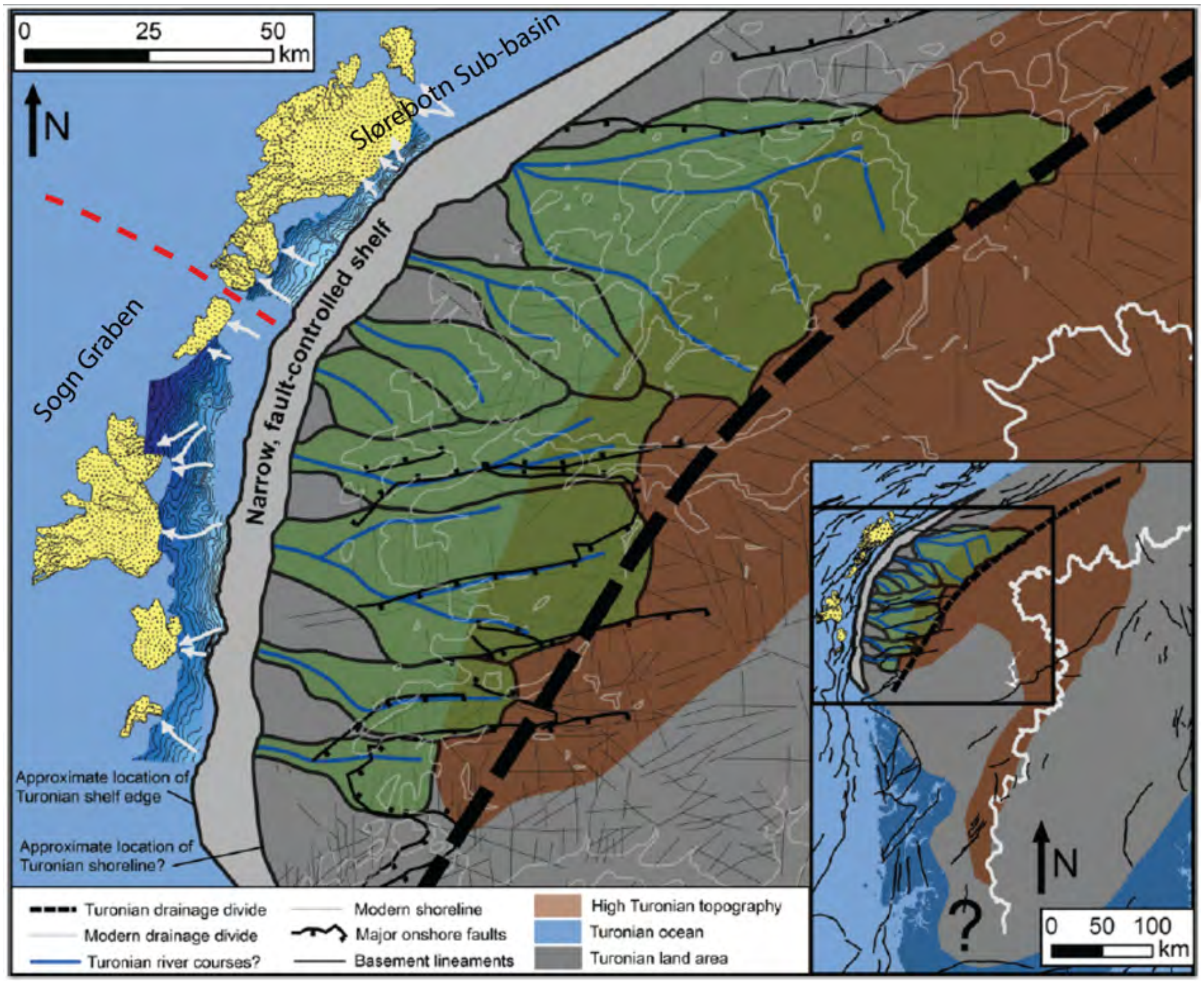


Fig. 5.10 Cartoon of a regional drainage model for the Møre-Trøndelag margin during the Turonian. *Green shapes = drainage catchments, red dashed line represents the boundary between the northern North Sea and the mid-Norwegian Sea. Modified from Sømme and Jackson. (2013).*

The northern part of the Slørebotn Sub-basin differs from the southern part in the sense that the upper portion of S7 is characterized by a sandy interval of the Grindhval Member (Fig. 3.5). The origin of this sandy member is believed in this study to be related to early tectonic movement in the Late Campanian due to the opening of the North Atlantic (Swiecicki et al., 1998; Brekke, 2000; Gjelberg et al., 2001). The Grindhval Member can be found farther west in the Ormen Lange (e.g. 6305/5-1) where it corresponds to the basal Tang and Springar formations that compose the lower part of the reservoir (Ravnås et al., 2014a).

5.2.9 Sequence 8 (Lower Paleocene-Lower Pleistocene)

The basal part of S8 is defined by an unconformity surface that formed during rift events along the North Atlantic spreading center in the Late Cretaceous, which led to uplift of the entire study area (Gjelberg et al., 2001). The unconformity is well established both in the well and seismic data (e.g Fig. 3.5 and Fig. 4.5). In the well logs, the unconformity represents a stratigraphic gap where Late Maastrichtian to lower

Danian stratigraphy is missing (Fig. 3.5, Fig. 3.6 and Fig. 3.7). Whereas in the seismic data, the unconformity surface is characterized by low angle truncating reflectors below and onlap/downlap above (Fig. 1.2).

The rift event caused increased influx of coarse clastics (informal named the Egga Member) into the Møre area from the northeast, which were sourced from the Scandinavian mainland by a series of smaller paleovalleys (Gjelberg et al., 2001; Vergara et al., 2001; Gjelberg et al., 2005; Sømme et al., 2009) and the coarse sediments can be traced from the wells in the Slørebotn Sub-basin (e.g. 6205/3-1R) to the wells (e.g. 6305/5-1) in the Ormen Lange area farther northwest (Fig. 5.11a). The initial Egga Member deposits are not encountered by the wells in the Slørebotn Sub-basin, in contrast to the Ormen Lange area, where the whole succession is present (Gjelberg et al., 2001; Gjerlberg et al., 2005; Martinsen et al., 2005; Ichron, 2015). The absence of the lowermost S8 (Danian) sedimentary rocks further supports the theory that the study area was uplifted, which caused bypass of the initial S8 (Danian) deposits. During the Late Danian times, the study area experienced an increase in the relative sea level and subsequently deposition of the slightly older Egga Member (Fig. 3.5 and Fig. 3.7). The core data from wells 6205/3-1R, 6305/9-1 and 6306/10-1 shows that the Egga Member comprises thick amalgamated coarse- to medium-grained sandstone turbidites with a porosity and permeability of up to 27.5% and 2280 mD, respectively (Gjelberg et al., 2001; NPD factpages).

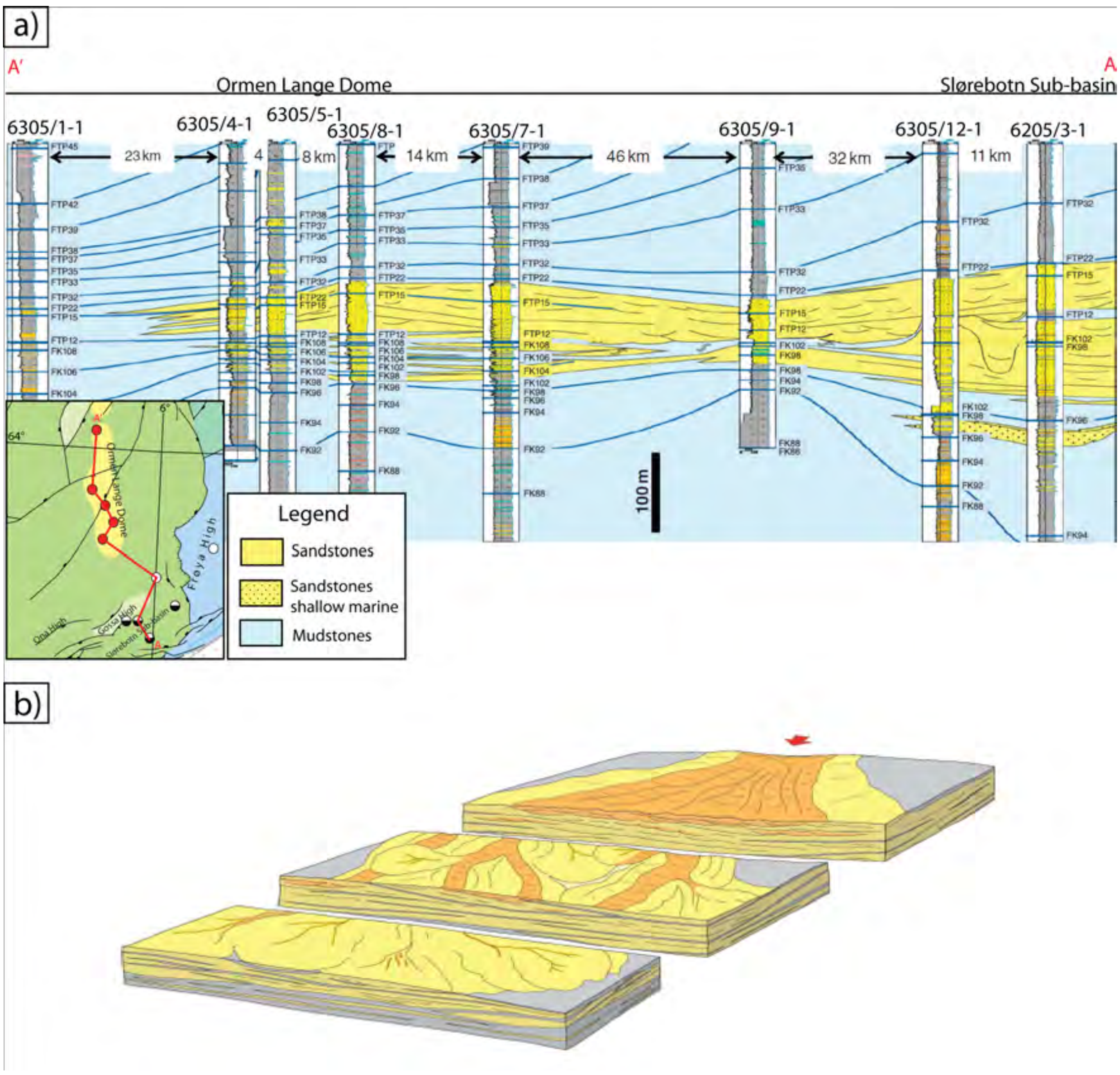


Fig. 5.11 a) well correlations of the Maastrichtian and Paleocene turbidite complex from the Slørebotn Sub-basin to the Ormen Lange dome; and b) a conceptual reservoir architecture of a channelized turbidite complex. *Modified from Ravnås et al., 2014a.*

It is suggested that the turbidites accumulated in minor depressions that developed along the basinfloor between the Norwegian mainland and the Møre Basin due to reactivations of the underlying Jurassic faults during the Late Cretaceous tectonic event (Gjelberg et al., 2005; Ravnås et al., 2014a). During this time period, the Slørebotn Sub-basin formed an intra-slope basin with the uplifted Gossa High farther west where the main depositional process of the turbidites was ponding (Gjelberg et al., 2001; Gjelberg et al., 2005; Ravnås et al., 2014a). This is in alignment with the observations in this study where the lower S8 (Danian) interval can be observed to have been strongly controlled by the Late Cenomanian-Late Maastrichtian rift topography. More specifically, the seismic reflectors of the lower S8 can be observed to truncate against FF2 and eventually spill over (Fig. 4.2 and Fig. 4.5) or downlap above structural highs in the southwest (Fig. 4.3). Hence, the lower laying FF1 that control the basement highs in the west were important for controlling and distributing the sedimentation of the lower S8 (Danian) within the Slørebotn Sub-basin (Fig. 4.16).

The Egga Member is not as clearly marked in the GR logs for the southeastern wells (e.g. 6204/10-1), compared to wells 6205/3-1R or 6306/6-1 in the north (Fig. 4.10). However, cuttings samples from well 6204/11-1 shows the presence of a thin and clean sandstone unit at the lower portion of S8 (NPD factpages) and this could possibly represent the distal part of a turbidite fan complex.

During the Early Eocene, the area was severely uplifted because of rifting along the North Atlantic that led to the final break-up between Greenland and Scandinavia (Eldholm and Thiede, 1980; Lundin and Doré, 1997; Brekke, 2000). The once low-lying Slørebotn Sub-basin was uplifted to the point that the westward bounding basement highs were now structurally below the sub-basin (Fig. 1.2 and Fig. 4.2). Furthermore, the uplift resulted in erosion and non-deposition of the middle S8 (Paleogene) sedimentary rocks in the Slørebotn Sub-basin, which is reflected in the well logs (6205/3-1R and 6306/6-1) by the absence of the middle S8 (Paleogene) sedimentary rocks (Fig. 3.5 and Fig. 3.7). In the southeast, the wells (e.g. 6204/10-2A) show evidence of uplift to a lesser extent, where only the lower Middle S8 (Eocene) is missing (Fig. 3.6). In contrast, wells 6305/9-1, 6305/12-1 and 6305/12-2 farther west penetrated the middle S8 (Paleogene) sedimentary rocks (Fig. 3.5 and Fig. 3.7). Hence, the Early Eocence uplift was greater in the northeast compared to the southeast and a shift in the accommodation space occured from the Slørebotn Sub-basin to structural highs in the west (Fig. 3.5 and Fig. 4.2 and Fig. 5.5).

Further evidence of the substantial uplift of the northeast during the Early Eocene can be observed by the thick turbidite complex that is interpreted just west of the Gossa High (Fig. 4.16 and Fig. 4.19). The package naturally thins out towards the Ormen Lange dome since this area was uplifted and formed as a structural high during the Early Eocene (Doré et al., 2008). The composition and reservoir quality of the potential turbidite complex is uncertain because none of the wells on the Gossa High (6305/12-2) and Ormen Lange field (e.g 6305/5-1) penetrated the complex, which accumulated in the depression that formed in the Møre Basin between the Gossa High and the Ormen Lange dome. The material that was being eroded during the Early Eocene would most likely come from the fine-grained dominated Upper Paleocene (Fig. 3.5, Fig. 3.6 and Fig. 3.7). Despite this, the post-Early Eocene seismic package clearly shows sedimentary structures of channels and lenses (Fig. 4.20), which are typical for a turbidite complex (Fig. 5.11b). In addition, the Norwegian mainland is believed to have been uplifted during the Early Eocene (Martinsen et al., 1999; Redfield et al., 2005). Thus, a similar scenario as for the Danian turbidites that were sourced from the Norwegian mainland (Gjelberg et al., 2005; Ravnås et al., 2014a) could therefore be expected for the interpreted post-Early Eocene turbidites.

The sedimentary history of the Eocene-Pliocene periods is generally poorly documented in the mid-Norwegian shelf, mainly because many wells often do not collect data in the shallower overburden section since the deeper targets were the objectives (Martinsen and Dreyer, 2001). In addition, it is believed that after the continent-continent separation in the Early Eocene that mainly deep-marine mudstones were deposited in the Norwegian Sea, which is based on the lack of sandstones penetrated by the wells (Lien, 2005). However, exploration wells are usually drilled on the structural highs and not low-laying areas where turbidites often accumulate. Therefore, it is interpreted in this study that thick a accumulation of coarse clastics of Eocene age is present in the Møre Basin. By analogy to the North Sea, Eocene turbidite sands are proven to be prolific hydrocarbon-bearing reservoir (e.g Frigg Field) successions (Faleide et al., 2010).

On the Frøya High, several sills have been interpreted to exist over large portions of the high in the upper part of S8 (Table 4.1;F2) (Fig. 4.16). These volcanic rocks are located relative to the strike of F3 and similar volcanic rocks both in age and position to the facture zone is found on the opposite side of the Møre margin, on Greenland (Torske and Prestvik, 1991). This implies that F3 may have acted as a zone of structural weakness that controlled the distribution of the volcanic rocks.

The uplift of the study area pushed the shoreline farther out towards the west, which resulted in the development of out-building shelf-margin clinoforms in the western part of the study area (Fig. 1.2, Fig. 4.16 and Fig. 4.18). The present-day architecture of the clinoforms shows evidence of being severely affected by erosion, where the entire topsets of the clinoforms are absent (Fig. 4.18). The cause of erosion is interpreted to be related to glaciations of the Norwegian shelf (Martinsen et al., 1999) that removed large portions of the Miocene stratigraphic record, which can be observed in all of the ten exploration wells (Fig. 3.5, Fig. 3.6 and Fig. 3.7).

5.3 Comparison to adjacent regions

A brief comparison has been made to the Halten Terrace and the northern North Sea to highlight some of the main differences. Although the areas experienced similar timing of the tectonic events, the impact and development during the tectonic movement differs.

The difference between the Halten Terrace, northern North Sea and the study area are especially evident in the Early-Middle Jurassic stratigraphic intervals (Fig. 3.10). Although all the three areas experienced tectonic movement during the late Early Jurassic, the study area clearly experienced uplift and erosion to a higher extent, where the entire Lower Jurassic and in some places older rocks are missing (Jongepier et al., 1996). On the contrary, on the Halten Terrace and northern North Sea, the majority of Early to Middle Jurassic strata tend to be preserved and form important reservoir targets (Færseth and Ravnås, 1998; Hesthammer et al., 1999; Corfield and Sharp, 2000; Ravnås et al., 2014b; Thrana et al., 2014). In addition, both the Halten Terrace and northern North Sea experienced marginal marine deposits in the Pliensbachian (Dalland et al., 1998; Folkestad et al., 2014; Ravnås et al., 2014b; Thrana et al., 2014), whereas the study area was dominated by continental deposits until the Bathonian. A marginal marine environment was not introduced before the Early Callovian as a result of a Bathonian rift episode (Jongepier et al., 1996). A possible explanation for the large differences observed in the Early and Middle Jurassic stratigraphy could be related the increased uplift due to the underplating of igneous rocks, if the assumption is correct that the study area developed as an extensional upper-plate margin (Fig. 5.1 and 5.1.9 Structural model). The increased uplift of the study area would therefore also explain the late introduction of marginal marine deposits.

The renewed rifting from the Middle Jurassic until Late Jurassic is documented in all the three areas (Halten Terrace, northern North Sea and Slørebotn Sub-basin) to have been the phase that formed the dominant architecture of the areas (Jongepier et al., 1996; Hesthammer et al., 1999; Corfield and Sharp, 2000; Marsh et al., 2010; Ravnås et al., 2014b). Subsequent to the main Late Jurassic rift event, regional deposition of the shale-prone Melke and Spekk formations (Mid-Norway), equivalent to the Draupne and Kimmeridge clay formations in the North Sea took place in all three reference areas, which was followed by thermal subsidence in the Early Cretaceous (Swiecicki et al., 1998; Faleide et al., 2010).

Renewed rifting along the North Atlantic during the Late Cretaceous caused uplift of the Halten Terrace and the study area, which resulted in forming a well-developed unconformity that can be traced over the Halten Terrace and down to the Slørebotn Sub-basin (Fig. 3.10) (Swiecicki et al., 1998; Martinsen et al., 1999; Brekke, 2000; Gjelberg et al., 2001; Vergara et al., 2001). There is a clear difference in the preserved sedimentary rocks above the unconformity, where in the study area, a thick succession of Lower Paleocene turbidite sandstones has been frequently penetrated by wells (e.g. 6205/3-1R) in the northern part of the Slørebotn Sub-basin (Fig. 3.5) (Gjelberg et al., 2001). Whereas on the Halten Terrace, no sandy development is documented above the unconformity (Fig. 3.10) (Swiecicki et al., 1998; Vergara et al., 2001; Gjelberg et al., 2005). It is suggested that the reason for the poor development of coarse clastics above the unconformity is related to the size of the shelf areas and the NE-SW orientation of the onshore fjords that favored sediment sink near the mouth of the Romsdalsfjord (Fig. 5.6) (Gjelberg et al., 2005; Sømme et al., 2009).

Following the Late Jurassic-Early Cretaceous extensional event, the northern North Sea developed as a failed rift system and was mainly in a state of thermal subsidence and post-rift infill during the Cretaceous and Cenozoic; however several phases of basin margin uplift disturbed the overall post-rift development of the study area (Nøttvedt et al., 1995; Martinsen et al., 1999). Similar to the Halten Terrace and study area, the Late Cretaceous unconformity is present in the northern North Sea (Fig. 3.10), but very little sand is found above this boundary (Martinsen et al., 1999; Vergara et al., 2001). However, farther south of the northern North Sea area, several discoveries (e.g. Cod, Frigg and Forties fields) have been made in turbidite sandstones of Early Paleocene and Eocene age (Dunn, 1975; Swiecicki et al., 1998). The Slørebotn Sub-basin area has not yet proved any thick Eocene sandstone units (Fig. 3.5, Fig. 3.6 and Fig. 3.7), but based on seismic interpretation, it is proposed that thick successions of turbidite deposits are present farther west of the sub-basin, just south of the Ormen Lange dome (Fig. 4.16 and Fig. 4.19).

6 CONCLUSIONS

This study utilizes 3D and 2D seismic data, well logs and interpreted biostraigraphic data from Ichron. (2015) to improve the knowledge of the Mesozoic to Cenozoic tectono-stratigraphic evolution of the Slørebotn Sub-basin. The main findings and contributions in this study are:

- Nine seismic sequences that are bounded by nine seismic sequence boundaries are defined. The seismic sequences are both genetic and depositional surfaces that range from Late Triassic to Late Neogene in age. Out of the nine seismic sequence boundaries, four Cretaceous and one Cenozoic boundary were possible to interpret with confidence throughout the Slørebotn Sub-basin. The interpretation revealed several shifts in the accommodation space through time in the Slørebotn Sub-basin in response to changes in the tectonic setting.
- Four fault families (FF1, FF2, FF4 and FF5) and one prominent NW-SE transfer fault (F3) have been identified to have played a major role in the structural evolution and the control on sedimentation in the Slørebotn Sub-basin throughout the Mesozoic to Cenozoic. The main tectonic events recognized to have affected the study area are:
    - an Early/Middle Triassic rift phase that might have started to already define the present-day structural elements and significant erosion of pre-S0 sedimentary rocks on the highs;
    - a late Early Jurassic rift/uplift phase that resulted in tectonic movement along FF1, FF4 and FF5, and caused erosion of almost the entire S1 sedimentary record;
    - a Bathonian rift phase that resulted in tectonic movement along FF1, FF4 and possibly FF5, but this is not certain due to the limited data. The Bathonian rift phase marks a change in the depositional environment where marginal marine conditions was firstly introduced to the Slørebotn Sub-basin;
    - a rift climax in the Slørebotn Sub-basin during the mid Late Jurassic where FF1, FF4 and FF5 were highly active. The rift event resulted in high relief and deepening of basin areas that were characterized by anoxic bottom water conditions;
    - a Turonian phase that caused uplift and vertical movement along FF4, FF5 and the northernmost part of FF1;
    - a renewed rift phase during the Late Cenomanian-Late Maastrichtian resulted in reactivation of older FF1 in the north and the development of FF2. During the rift phase F3 is believed to have formed a structural weak zone in the east that gave rise to coarse clastic input into the Slørebotn Sub-basin (Gjelberg et al., 2005);
    - an Early Eocene compressional phase caused reactivation of the fault families (FF1, FF4 and FF5), however the largest magnitude of uplift is observed in the inverted basin fill of the Slørebotn Sub-basin. The effect of F3 during the Early Eocene can be observed by the several Eocene domes along the striking zone of F3 in the Møre Basin and the numerous sill deposits at both ends of F3 on eastern Greenland and the Frøya High (Torske and Prestvik, 1991; Doré et al., 2008); and
    - a phase of uplift and glacial erosion during the Pliocene resulted in large amounts of the Miocene-Pliocene stratigraphy being eroded and deposited farther out in to the Møre Basin.
- Several coarse-grained sedimentary units are recognized in the seismic sequences and mapped within the Slørebotn Sub-basin. The main coarse-grained lithofacies that have been recognized are:
    - deep-marine gravity induced slope apron fans that have been cored by well 6205/3-1R in S3 and S4 to be a product of the mid Late Jurassic rift topography (Jongepier et al., 1996);
    - open marine medium to coarse-grained sandy submarine fans that are interpreted in S6 and

the basal part of S7.;

  - marine medium- to coarse-grained turbidite sandstones that have been cored by well 6205/3-1R at the base of S8 (Gjelberg et al., 2001; Ravnås et al., 2014a). The turbidite interval is believed to have been sourced from the Norwegian mainland through a feeder system northeast of the study area that was created by F3 (Gjelberg et al., 2005);
  - thick (up to 200 ms TWT) slope and basinfloor turbidites have been interpreted above the Early Eocene level based on the seismic facies signature just west of the Gossa High. The origin of the turbidites is believed to have been caused by uplift of the Norwegian mainland and the study area as a result of the final breakup between Greenland and Norway (Eldholm and Thiede, 1980; Lundin and Doré, 1997; Brekke, 2000). The Eocene turbidites could represent a new play type for the mid-Norwegian Sea that has not been previously recognized.
- There are important differences in Jurassic stratigraphy between the Slørebotn Sub-basin area and the adjacent Halten Terrace and northern North Sea. While the Early Jurassic sedimentary rocks tend to be preserved on the Halten Terrace and in the northern North sea (Færseth and Ravnås, 1998; Hesthammer et al., 1999; Corfield and Sharp, 2000; Ravnås et al., 2014b; Thrana et al., 2014), in the Slørebotn Sub-basin area the Early Jurassic interval and in some places also older rocks are missing (Jongepier et al., 1996). Although all three reference areas shared similar timing of the tectonic movement, it is proposed in this study that the Slørebotn Sub-basin area experienced a greater magnitude of uplift during the late Early Jurassic tectonic phase due to the characteristics of the extensional movement in an upper-plate margin framework.

On a final note

By not exploring the unknown, due to statements made by previous authors that predict the lack of hydrocarbon potential in the Slørebotn Sub-basin or any other area for that matter, they are most often basing their interpretation on limited data and their belief on the occurrence of hydrocarbons. By neglecting an area without having the sufficient data to adequately support the interpretation can be both destructive for the academic and economic development. Thus, I leave you with a final quote from Pratt. (1952), "When no man any longer believes more oil is left to be found, no more oil fields will be discovered, but so long as a single oil-finder remains with a mental vision of a new oil field to cherish, along with freedom and incentive to explore, just so long new oil fields may continue to be discovered".

7 APPENDIX

Table 7.1 Appendix A *Overview of the basic information and seismic features in the seismic survey that were used for this study.*

Survey Name	Acquisition Date	Reprocessed	Seismic quality	Hard-Kick	Seismic artifacts
FH-91	1991	-	Good	Red	Noise and multiples
FH-92	1992	-	Poor	Red	Noise and migration errors
FHM-91	1991	-	Good	Red	Noise and multiples
FRD-88	1988	-	Moderate	Blue/Red	Migration errors
FRDE-90	1990	-	Moderate/Poor	Red	Noise and multiples
GFI-85	1985	-	Poor	Blue/Red	Migration errors
GGE-91	1991	-	Moderate	Red	Noise,multiples and migration errors
GGW-91	1991	-	Poor	Red	Noise and multiples
GM1E86	1986	-	Moderate/Poor	Red	Noise
GM1|	-	-	Moderate	Red	Migration errors
GMM-94	1994	-	Poor	Blue/Red	Noise
GMNR-94	1994	-	Moderate	Red	-
GMT84	1984	-	Moderate	Red	Noise
MM90	1990	-	Moderate	Red	Multiples and migration errors
MN9105	1991	-	Poor	Red	Noise and migration errors
MN9106	1991	-	Poor	Blue	Noise
MNR04	2004	-	Poor	Red	-
MNR07	2007	-	Good/Moderate	Red	-
MV01RE-NPD-MB-92-BP-R01	1992	2001	Good/Moderate	Red	Noise
MV01RE-NPD-ML-84-BP-R01	1984	2001	Poor	Red	-
MV01RE-NPD-VRB-88-BP-R01	1984	2001	Poor	Red	-
NH8958	1989	-	Poor	Red	Noise
NH9104	1991	-	Poor	Red	Noise
NH9203	1992	-	Moderate	Red	-
NH9303	1993	-	Poor	Blue	Noise
NH9651	1996	-	Poor	Red	Noise
NH9753	1997	-	Poor	Red	Noise and migration errors
NM1-85	1985	-	Moderate	Blue	Noise and multiples
NPD-KYST-96	1996	-	Moderate/Poor	Red	Multiples and migration errors
NPD-KYST-96-PROS-1	1996	2001	Poor	Red	Multiples
NPD-KYST-96-PROS-2	1996	2002	Poor	Red	Noise and migration errors
NPD-KYST-97	1997	-	Poor	Red	-
NPD-MB-88-NPD-VRB-88_REPROS_1992	1988	1992	Moderate/Poor	Red	Noise and migration errors
NPD-MB-91	1991	-	Poor	Red	Noise
NPD-MB-92	1992	-	Moderate	Red	Migration errors
NPD-ML-84	1984	-	Poor	Red	Noise
NPD-STOR-85	1985	-	Poor	Red	Noise
OLW02	2002	-	Moderate/Poor	Red	Noise,multiples and migration errors
SEABED-PROJECT	-	-	Poor	Red	-
SG8607	1987	-	Moderate	Red	-
SG9009	2009	-	Moderate/Poor	Red	-
SG9110	1991	-	Poor	Red	Migration errors
SG9113	1991	-	Moderate/Poor	Red	-
SG9308	1993	-	Moderate	Red	Multiples
SH8805	1988	-	Moderate/Poor	Red	-
SH8906	1989	-	Poor	Red	Migration errors
SPT-MBR-94	1994	-	Moderate/Poor	Red	Migration errors
ST8503	2003	-	Moderate/Poor	Red	Multiples and migration errors
ST8629	1986	-	Moderate/Poor	Red	Multiples and migration errors
ST8703	1987	-	Poor	Red	Migration errors
ST9208	1992	-	Poor	Blue	-
ST9692-EQ-AMP	1996	-	Moderate/Poor	Red	Migration errors
ST9789-EQ-AMP	1997	-	Moderate/Poor	Red	-
VMT-95	1995	-	Poor	Red	-
WG96GNH	1996	-	Poor	Red	Noise
WG96ONH	1996	-	Moderate/Poor	Red	Noise
MN88	1988	-	Moderate/Poor	Blue	Noise
NPD-B-72	1972	-	Poor	Blue	-
NPD-FB-84	1984	-	Moderate/Poor	Blue	Migration errors
NPD_MB_88_RP92	1988	1992	Moderate/Poor	Blue	Noise
NPD_MB_92	1992	-	Moderate/Poor	Blue	-
Geoprovider NPD-ML02-72	2002	2017	Moderate	Red	Noise and migration errors
Geoprovider NPD-ML01-ML70-ML71-ML76-NR83		2017	Moderate	Blue	Noise and migration errors

8 REFERENCES

Axen, G. J. (2004). Mechanics of low-angle normal faults. In G. D. Karner (Ed), *Rheology and Deformation of the Lithosphere at Continental Margins*, pp. 46-91. New York, Columbia University Press.

Birkelund, T., & Perch-Nielsen, K. (1976). Late Palaeozoic-Mesozoic evolution of central East Greenland. *Geology of Greenland*, pp. 304-339.

Blystad, P., Brekke, H., Færseth, R., Larsen, B., Skogseid, J., Tørudbakken, B. (1995). NPD-Bulletin No 8, Structural elements of the Norwegian Continental Shelf-Part II: The Norwegian Sea Region. *The Norwegian Petroleum Directorate (NPD)*.

Brekke, H. (2000). The tectonic evolution of the Norwegian Sea continental margin, with emphasis on the Voring and More basins. In A. Nøttvedt (Ed), *Dynamics of the Norwegian Margin, 167*, pp. 327-387. Geological Society of London.

Brekke, H., Dahlgren, S., Nyland, B., Magnus, C. (1999). The prospectivity of the Vøring and Møre basins on the Norwegian Sea continental margin. In A. J. Fleet (Ed), *Petroleum Geology of Northwest Europe,* 5(1), pp. 261-274. Geological Society London, Petroleum Geology Conference series.

Brekke, H., & Riis, F. (1987). Tectonics and basin evolution of the Norwegian shelf between 62 N and 72 N. *Norsk Geologisk Tidsskrift, 67*, pp. 295-322.

Brekke, H., Sjulstad, H. I., Magnus, C., Williams, R. W. (2001). Sedimentary environments offshore Norway —an overview. In O. J. Martinsen., T. Dreyer (Ed). *Norwegian Petroleum Society Special Publications, Vol. 10*, pp. 7-37. Bergen, Elsevier Science.

Bugge, T., Prestvik, T., Rokoengen, K. (1980). Lower tertiary volcanic rocks off Kristiansund—mid Norway. *Marine Geology*, *35*(4), pp. 277-286.

Bugge, T., Ringås, J. E., Leith, D. A., Mangerud, G., Weiss, H. M., Leith, T. L. (2002). Upper Permian as a new play model on the mid-Norwegian continental shelf: Investigated by shallow stratigraphic drilling. *AAPG bulletin*, *86*(1), pp. 107-127.

Corfield, S., & Sharp, I. R. (2000). Structural style and stratigraphic architecture of fault propagation folding in extensional settings: a seismic example from the Smørbukk area, Halten Terrace, Mid-Norway. *Basin Research, 12*(3-4), pp. 329-341.

Dalland, A., Worsley, D., Ofstad, K. (1988). NPD-Bulletin No 4, A lithostratigraphic scheme for the Mesozoic and Cenozoic succesion offshore Norway north of 62° N. *The Norwegian Petroleum Directorate (NPD).*

Doré, A. G. (1991). The structural foundation and evolution of Mesozoic seaways between Europe and the Arctic. In T. J. Algeo., T. Correge., H. Falcon-Lang., P. Hesse., I. Montanez (Ed), *Palaeogeography, Palaeoclimatology, Palaeoecology, 87*(1-4), pp. 441-492.

Doré, A. G., Lundin, E. R., Birkeland, Ø., Eliassen, P. E., Jensen, L. N. (1997a). The NE Atlantic margin; implications of late Mesozoic and Cenozoic events for hydrocarbon prospectivity. *Petroleum Geoscience, 3*(2), pp. 117-131.

Doré, A. G., Lundin, E. R., Fichler, C., Olesen, O. (1997b). Patterns of basement structure and reactivation along the NE Atlantic margin. *Journal of the Geological Society, 154*(1), pp. 85-92.

Doré, A. G., Lundin, E. R., Kusznir, N. J., Pascal, C. (2008). Potential mechanisms for the genesis of Cenozoic domal structures on the NE Atlantic margin: pros, cons and some new ideas. *Geological Society, London, Special Publications, 306*(1), pp. 1-26.

Dunn, W. W. (1975). North Sea basinal area, Europe—an important oil and gas province. *Norges Geologiske Undersøkelse, 316*, pp. 69-97.

Eldholm, O., & Thiede, J. (1980). Cenozoic continental separation between Europe and Greenland. *Palaeogeography, Palaeoclimatology, Palaeoecology*, *30*, pp. 243-259.

Eldholm, O., Tsikalas, F., Faleide, J. I. (2002). Continental margin off Norway 62–75 N: Palaeogene tectono-magmatic segmentation and sedimentation. In D. W. Jolley., B. R. Bell (Ed), *The North Atlantic Igneous Province: Stratigraphy, Tectonic, Volcanic and Magmatic Processes*, *197*(1), pp. 39-68. Geological Society, London.

Fagerland, N. (1990). Mid-Norway shelf - hydrocarbon habitat in relation to tectonic elements. *Norsk Geologisk Tidsskrift, 70*, pp. 65-79.

Faleide, J. I., Bjørlykke, K., Gabrielsen, R. H. (2010). Geology of the Norwegian Continental Shelf. In K. Bjørlykke (Ed), *Petroleum Geoscience: From Sedimentary Environments to Rock Physics,* pp. 603-637. Springer Berlin Hidelberg.

Folkestad, A., Odinsen, T., Fossen, H., Pearce, M. A. (2014). Tectonic influence on the Jurassic sedimentary architecture in the northern North Sea with focus on the Brent Group. In A. W. Martinius., R. Ravnås., J. A. Howell., R. J. Steel., J. P. Wonham (Ed), *From Depositional Systems to Sedimentary Successions on the Norwegian Continental Margin*, *46*, pp. 389-416.

Færseth, R. B., & Ravnås, R. (1998). Evolution of the Oseberg fault-block in context of the northern North Sea structural framework. *Marine and Petroleum Geology, 15*(5), pp. 467-490.

Færseth, R. B., & Lien, T. (2002). Cretaceous evolution in the Norwegian Sea—a period characterized by tectonic quiescence. *Marine and Petroleum Geology, 19*(8), pp. 1005-1027.

Gabrielsen, R. H., Odinsen, T., Grunnaleite, I. (1999). Structuring of the Northern Viking Graben and the Møre Basin; the influence of basement structural grain, and the particular role of the Møre-Trøndelag Fault Complex. *Marine and Petroleum Geology, 16*(5), pp. 443-465.

Gabrielsen, R. H., Braathen, A., Dehls, J., Roberts, D. (2002). Tectonic lineaments of Norway. *Norsk Geologisk Tidsskrift, 82*(3), pp. 153-174.

Garner, L. H., Farrimond, P., Nuzzo, M. (2017). Alternative Source Rocks on the Norwegian Continental Shelf: Potential Cretaceous Sourcing in Deepwater Basins. *Poster session presented at AAPG/SEG International Conference and Exhibition, London*.

Gjelberg, J. G., Dreyer, T., Høie, A., Tjelland, T., Lilleng, T., Brooks, J., Glennie, K. (1987). Late Triassic to Mid-Jurassic sandbody development on the Barents and Mid-Norwegian shelf. *Petroleum Geology of North West Europe*, pp. 1105-1129.

Gjelberg, J. G., Enoksen, T., Kjrnes, P., Mangerud, G., Martinsen, O. J., Roe, E., Vagnes, E. (2001). The Maastrichtian and Danian depositional setting, along the eastern margin of the Møre Basin (mid-

Norwegian Shelf): implications for reservoir development of the Ormen Lange Field. In O. J. Martinsen., T. Dreyer (Ed), *Sedimentary Environments Offshore Norway - Palaeozoic to Recent, 10*, pp. 421-440. Norwegian Petroleum Society Special Publications, Elsevier.

Gjelberg, J. G., Martinsen, O. J., Charnock, M., Møller, N., Antonsen, P. (2005). The reservoir development of the Late Maastrichtian–Early Paleocene Ormen Lange gas field, Møre Basin, Mid-Norwegian Shelf. *Paper presented at the Geological Society, London, Petroleum Geology Conference series, 6,* pp. 1165-1184.

Grunnaleite, I., & Gabrielsen, R. H. (1995). Structure of the Møre Basin, mid-Norway continental margin. *Tectonophysics, 252*(1), pp. 221-251.

Hesthammer, J., Jourdan, C. A., Nielsen, P. E., Ekern, T. E., Gibbons, K. A. (1999). A tectonostratigraphic framework for the Statfjord Field, northern North Sea. *Petroleum Geoscience, 5*(3), pp. 241-256.

Hinze, W. J., Von Frese, R. R., & Saad, A. H. (2013). Gravity and magnetic exploration: Principles, practices, and applications. *Cambridge University Press*.

Ichron. (2015). Møre and Vøring Basin Stratigraphic Database. ECM 5152.

Jongepier, K., Rui, J. C., Grue, K. (1997). Triassic to early Cretaceous stratigraphic and structural development of the northeastern More Basin margin, off mid-Norway. *Oceanographic Literature Review, 7*(44), pp. 199-214.

Larsen, V. B. (1987). A synthesis of tectonically-related stratigraphy in the North Atlantic-Arctic region from Aalenian to Cenomanian time. *Norsk Geologisk Tidsskrift, 67*(4), pp. 281-293.

Lien, T. (2005). From rifting to drifting: effects on the development of deep-water hydrocarbon reservoirs in a passive margin setting, Norwegian Sea. *Norwegian Journal of Geology/Norsk Geologisk Forening, 85*(4), pp. 319-332.

Lister, G. S., Etheridge, M. A., Symonds, P. A. (1986). Detachment faulting and the evolution of passive continental margins. *Geology, 14*(3), pp. 246-250.

Maast, T. E., Jahren, J., Bjorlykke, K. (2011). Diagenetic controls on reservoir quality in Middle to Upper Jurassic sandstones in the South Viking Graben, North Sea. *AAPG bulletin*, *95*(11), pp. 1937-1958.

Marsh, N., Imber, J., Holdsworth, R. E., Brockbank, P., Ringrose, P. (2010). The structural evolution of the Halten Terrace, offshore Mid-Norway: extensional fault growth and strain localisation in a multi-layer brittle–ductile system. *Basin Research, 22*(2), pp. 195-214.

Martinsen, O. J., Bøen, F., Charnock, M. A., Mangerud, G., Nøttvedt, A. (1999). Cenozoic development of the Norwegian margin 60–64° N: sequences and sedimentary response to variable basin physiography and tectonic setting. *Paper presented at the Geological Society, London, Petroleum Geology Conference series 5,* pp. 293-304.

Martinsen, O. J., Lien, T., Jackson, C. (2005). Cretaceous and Palaeogene turbidite systems in the North Sea and Norwegian Sea Basins: source, staging area and basin physiography controls on reservoir development. *Paper presented at the Geological Society, London, Petroleum Geology Conference series, 6,* pp. 1147-1164.

Martinsen, O. J., & Dreyer, T. (2001). Sedimentary environments offshore norway—palaeozoic to recent: an introduction *Norwegian Petroleum Society Special Publications,* 10, pp. 1-5. Elsevier.

Morton, A., Hallsworth, C., Strogen, D., Whitham, A., Fanning, M. (2009). Evolution of provenance in the NE Atlantic rift: The Early–Middle Jurassic succession in the Heidrun Field, Halten Terrace, offshore Mid-Norway. *Marine and Petroleum Geology*, *26*(7), pp. 1100-1117. Elsevier.

Mosar, J. (2003). Scandinavia's North Atlantic passive margin. *Journal of Geophysical Research: Solid Earth, 108*(B8), pp. 1-18.

Mosar, J., Eide, E. A., Osmundsen, P. T., Sommaruga, A., Torsvik, T. H. (2002a). Greenland–Norway separation: a geodynamic model for the North Atlantic. *Paper presented at the Norwegian Journal of Geology, 82,* pp. 282

Mosar, J., Lewis, G., Torsvik, T. (2002b). North Atlantic sea-floor spreading rates: implications for the Tertiary development of inversion structures of the Norwegian–Greenland Sea. *Journal of the Geological Society, 159*(5), pp. 503-515.

Mudge, D., Gall, M., Holdwa, K. (2007). The Norwegian Sea-exploration in a frontier province. *Geo ExPro*, pp. 40-46.

Müller, R., Ngstuen, J. P., Eide, F., Lie, H. (2005). Late Permian to Triassic basin infill history and palaeogeography of the Mid-Norwegian shelf—East Greenland region. *Norwegian Petroleum Society Special Publications,* 12, pp. 165-189. Elsevier.

Mørk, M., & Johnsen, S. (2005). Jurassic sandstone provenance and basement erosion in the Møre margin–Froan Basin area. *Norges geologiske undersøkelse Bulletin, 443*, pp. 5-18.

Nøttvedt, A., Gabrielsen, R., Steel, R. (1995). Tectonostratigraphy and sedimentary architecture of rift basins, with reference to the northern North Sea. *Marine and Petroleum Geology, 12*(8), pp. 881-901.

Osmundsen, P. T., & Ebbing, J. (2008). Styles of extension offshore mid-Norway and implications for mechanisms of crustal thinning at passive margins. *Tectonics, 27*(6), pp. 1-25.

Osmundsen, P. T., Sommaruga, A., Skilbrei, J. R., Olesen, O. (2002). Deep structure of the Mid Norway rifted margin. *Norwegian Journal of Geology/Norsk Geologisk Forening, 82*(4), pp. 205-224.

Ottesen, D., Rise, L., Sletten Andersen, E., Bugge, T., Eidvin, T. (2009). Geological evolution of the Norwegian continental shelf between 61° N and 68° N during the last 3 million years. *Norwegian Journal of Geology/Norsk Geologisk Forening, 89*(4), pp. 251-265.

Pratt, W. E. (1952). Toward a philosophy of oil-finding. *AAPG Bulletin*, *36*(12), pp. 2231-2236.

Ravnås, R., Berge, K., Campbell, H., Harvey, C., Norton, M. J. (2014b). Halten terrace Lower and Middle Jurassic inter-rift megasequence analysis: Megasequence structure, sedimentary architecture and controlling parameters. In A. W. Martinius., R. Ravnås., J. A. Howell., R. J. Steel., J. P. Wonham (Ed), *From Depositional Systems to Sedimentary Successions on the Norwegian Continental Margin*, *46*, pp. 215-252.

Ravnås, R., Cook, A., Engenes, K., Germs, H., Grecula, M., Haga, J., Harvey, C., Maceachern, J. A. (2014a). The Ormen Lange turbidite systems: sedimentary architectures and sequence structure of sandy slope fans in a sediment-starved basin. In A. W. Martinius., R. Ravnås., J. A. Howell., R. J. Steel., J. P. Wonham (Ed), *From Depositional Systems to Sedimentary Successions on the Norwegian Continental Margin, 46*, pp. 609-646.

Ravnås, R., Nøttvedt, A., Steel, R., Windelstad, J. (2000). Syn-rift sedimentary architectures in the Northern North Sea. *Geological Society, London, Special Publications, 167*(1), pp. 133-177.

Redfield, T., Braathen, A., Gabrielsen, R., Osmundsen, P., Torsvik, T., Andriessen, P. (2005). Late Mesozoic to early Cenozoic components of vertical separation across the Møre–Trøndelag Fault Complex, Norway. *Tectonophysics, 395*(3-4), pp. 233-249.

Richardson, N. J., Underhill, J. R., Lewis, G. (2005). The role of evaporite mobility in modifying subsidence patterns during normal fault growth and linkage, Halten Terrace, Mid-Norway. *Basin Research, 17*(2), pp. 203-223.

Riis, F. (1996). Quantification of Cenozoic vertical movements of Scandinavia by correlation of morphological surfaces with offshore data. *Global and Planetary Change, 12*(1-4), pp. 331-357.

Riis, F., & Fjeldskaar, W. (1992). On the magnitude of the Late Tertiary and Quaternary erosion and its significance for the uplift of Scandinavia and the Barents Sea. *Structural and tectonic modelling and its application to petroleum geology,* 1, pp. 163-185. Elsevier Amsterdam.

Skogseid, J., Pedersen, T., Larsen, B. (1992). Vøring Basin: subsidence and tectonic evolution. *Structural and Tectonic Modelling and Its Application to Petroleum Geology, NPF Special Publications, 1*, pp. 55-82.

Skogseid, J., Planke, S., Faleide, J. I., Pedersen, T., Eldholm, O., Neverdal, F. (2000). NE Atlantic continental rifting and volcanic margin formation. *Geological Society, London, Special Publications, 167*(1), pp. 295-326.

Slama, J., Walderhaug, O., Fonneland, H., Kosler, J., Pedersen, R. B. (2011). Provenance of Neoproterozoic to upper Cretaceous sedimentary rocks, eastern Greenland: implications for recognizing the sources of sediments in the Norwegian Sea. *Sedimentary Geology*, *238*(3-4), pp. 254-267.

Smelror, M., Jacobsen, T., Rise, L., Skarbø, O., Verdenius, J., Vigran, J. (1994). Jurassic to Cretaceous stratigraphy of shallow cores on the Møre Basin Margin, Mid-Norway. *Norsk Geologisk Tidsskrift, 74*, pp. 89-107.

Spencer, A., Home, P., Wiik, V. (1986). Cretaceous stratigraphy and reservoir potential mid Norway continental shelf. *Habitat of Hydrocarbons on the Norwegian continental shelf*. pp 287-299. London.

Surlyk, F., Dons, T., Clausen, C. K., Higham, J. (2003). Upper cretaceous. In D. Evans (Ed), *Millennium Atlas: Petroleum Geology of the Central and Northern North Sea,* pp. 213-233. Geological Society of London.

Swiecicki, T., Gibbs, P., Farrow, G., Coward, M. (1998). A tectonostratigraphic framework for the Mid-Norway region. *Marine and Petroleum Geology, 15*(3), pp. 245-258.

Sømme, T. O., & Jackson, C. (2013). Source-to-sink analysis of ancient sedimentary systems using a subsurface case study from the Møre-Trøndelag area of southern Norway: Part 2–sediment dispersal and forcing mechanisms. *Basin Research, 25*(5), pp. 512-531.

Sømme, T. O., Martinsen, O. J., Thurmond, J. B. (2009). Reconstructing morphological and depositional characteristics in subsurface sedimentary systems: An example from the Maastrichtian–Danian Ormen Lange system, More Basin, Norwegian Sea. *AAPG bulletin*, *93*(10), pp. 1347-1377.

Talwani, M., & Eldholm, O. (1972). Continental margin off Norway: a geophysical study. *Geological Society of America Bulletin, 83*(12), pp. 3575-3606.

Talwani, M., & Eldholm, O. (1977). Evolution of the Norwegian-Greenland sea. *Geological Society of America Bulletin, 88*(7), pp. 969-999.

Thrana, C., Næss, A., Leary, S., Gowland, S., Brekken, M., Taylor, A. (2014). Updated depositional and stratigraphic model of the Lower Jurassic Åre Formation, Heidrun Field, Norway. In A. W. Martinius., R. Ravnås., J. A. Howell., R. J. Steel., J. P. Wonham (Ed), *From Depositional Systems to Sedimentary Successions on the Norwegian Continental Margin, 46*, pp. 253-289.

Torske, T., & Prestvik, T. (1991). Mesozoic Detachment in the Norwegian-Greenland Sea Region: Evidence from Jan Mayen Fracture Zone and Associated Alkaline Volcanics. *Geology, 19*, pp. 481-485.

Tsikalas, F., Faleide, J., Eldholm, O., Wilson, J. (2005). Late Mesozoic–Cenozoic structural and stratigraphic correlations between the conjugate mid-Norway and NE Greenland continental margins. *Paper presented at the Geological Society, London, Petroleum Geology Conference series, 6,* pp. 785-801.

Vergara, L., Wreglesworth, I., Trayfoot, M., Richardsen, G. (2001). The distribution of Cretaceous and Paleocene deep-water reservoirs in the Norwegian Sea basins. *Petroleum Geoscience, 7*(4), pp. 395-408.

Zastrozhnov, D., Gernigon, L., Gogin, I., Abdelmalak, M. M., Planke, S., Faleide, J. I., Eide, S., Myklebust, R. (2018). Cretaceous-Paleocene evolution and crustal structure of the northern Vøring Margin (offshore Mid-Norway): results from integrated geological and geophysical study. *Tectonics, 37*(2), pp. 497-528.

Zhong, X., Escalona, A., Sverdrup, E., Bukta, K. E. (2018). Impact of fault evolution in Gilbert-type fan deltas in the Evrostini area, south-central Gulf of Corinth, Greece. *Marine and Petroleum Geology, 95*, pp. 82-99.