Controls of basin margin tectonics on the Lower Cretaceous sedimentation in the Norwegian Barents Sea

Bereke Kairanov

Thesis submitted in fulfilment of the requirements for the degree of PHILOSOPHIAE DOCTOR (PhD)

University of Stavanger

Faculty of Science and Technology
Department of Energy Resources
2020
University of Stavanger
NO-4036 Stavanger
NORWAY
www.uis.no

©2020 Bereke Kairanov

ISSN: 1890-1387
PhD: Thesis UiS No. 530
Preface

This PhD thesis is submitted in fulfilment of the requirements for the degree of Philosophiae Doctor (PhD) at the University of Stavanger (UiS), Norway. The research was carried out between June 2014 to June 2018 and it was funded by the LoCrA consortium (https://wp.ux.uis.no/locra). During this period, I worked as a research fellow in the Department of Energy Resources, Faculty of Science and Technology at UiS. My main supervisor is Professor Alejandro Escalona (UiS) and my co-supervisor is Professor Nestor Cardozo (UiS). Industry collaboration was additionally established with Tore Åkermoem and Peter Abrahmson from MultiClient Geophysical, and with Emilie O’Neill from WesternGeco. During my PhD, I helped Professor Alejandro Escalona with the teaching of the introductory bachelor course in Geology (GEO100) and contributed to some courses in the Master of Petroleum Geosciences.

This research has resulted in five publications. Four of these have been published in different journals, including: Journal of Geodynamics, Journal of Structural Geology, and Marine and Petroleum geology. One manuscript has been submitted to Marine and Petroleum geology and is currently under review. Besides these publications, I have presented my research in several conferences, seminars, and E&P oil and gas companies. This thesis is structured similarly to a scientific paper and consists of two chapters. The first chapter is an introduction to the thesis, with a description of the general problems, motivation, objectives, results, discussion, and conclusions. The second chapter is a compilation of the five papers forming the main body of the thesis. Supplementary material such as conference abstracts are provided in the appendices.
Acknowledgements

Firstly, I would like to express my sincere gratitude to my supervisor Prof. Alejandro Escalona for his continuous support during my Ph.D., his patience, motivation, and immense knowledge. His guidance helped me during my research, writing of abstracts and papers, and finally the completion of this thesis. I could not imagine a better advisor and mentor for my Ph.D. I also would like to thank my co-supervisor Prof. Nestor Cardozo for his insightful comments and encouragement, but also for his hard questions which motivated me to widen my research from various perspectives.

My sincere thanks go to my colleagues from the D306 office: Dr. Dora, Dr. Sayyid, Dr. Shawn and Luis for all the fun during this journey, the nice discussions and constructive comments which were crucial for this research, but also for being very good friends outside the office.

I would also like to thank all LoCra collaborators: The University Centre in Svalbard (UNIS), Geological Survey of Denmark and Greenland (GEUS) teams: Snorre Olaussen, Sten-Andreas Grundvåg, Kasia K. Sliwinska and Henrik Nøhr-Hansen for their excellent expertise in hard rock geology and biostratigraphy, which helped me pursue my subsurface studies in the Barents Sea. I also thank the Moscow State University team: Alina Mordasova, Anna Suslova and Anatoli Nikishin for sharing their extensive knowledge of the Russian Barents Sea geology and providing support in data gathering. Special thanks to Ian Norton from the University of Texas Institute for Geophysics for teaching me how continents move (plate tectonics) and explaining the essence of potential field data modelling.

I would also like to acknowledge the Norwegian Petroleum Directorate (NPD), WesternGeco Multiclient, and MultiClient Geophysical (MCG) for providing data for this research. Many thanks to Peter Abrahamson (MCG), Tor Åkermoen (MCG) and Emilie O'Neill (WesternGeco) for
providing support with data. Thanks to Haliburton, Petroleum Experts and Schlumberger for providing academic licenses of their software’s Decision Space, Move and Petrel, respectively.

I would like to express my sincerest gratitude to my parents, Marat and Tamara, and to my brother Nursultan for their infinite emotional and financial support. My love and gratitude for them can hardly be expressed in words.

And my biggest thanks to my wife Gyuzal and my daughter Aiya, sorry for being grumpy whilst I wrote this thesis. Your love, patience and encouragement helped me a lot in these final steps. You have been amazing, and from now on I will come home earlier as I promised!
Abstract

Structural styles and stratigraphic patterns along North Atlantic margins display a large spectrum of complexity and variability. An extensive amount of subsurface data from the north-central and south-western Barents Sea are used to: (1) at a larger scale understand how various plate tectonic regimes controlled structuring, faulting and sedimentation along the northern and southern margins of the Barents Sea; (2) at a smaller scale understand how the structural evolution of basin bounding faults impacted sedimentation in basins which were affected by one or more phases and multiple directions of extension; and (3) improve the knowledge about the paleogeography of the Barents Sea. In order to fulfil these objectives, this research consists of a systematic analysis which is summarized in five journal articles.

Paper 1 improves the existing knowledge of the Early Cretaceous tectonostratigraphic development of the north-central Barents Sea based on observations from subsurface data, structural and plate tectonic restorations in an area distal from the northern margin of the Barents Sea. As result of this work, compressional tectonics in the Early Cretaceous is suggested to be induced by the opening of the Canada Basin which triggered reactivation of Late Palaeozoic normal faults in reverse mode. Reverse movement along these faults caused the formation of NE oriented structural highs and anticlines, which controlled and routed the progradation of Lower Cretaceous elastic material from the northern to the southern margins of the Barents Sea.

The second paper focuses on understanding the Early Cretaceous structural evolution of the Tromsø Basin (proximal southern margin of the Barents Sea) in the context of the geodynamic processes acting in the southwestern Barents Sea. We propose an Early Cretaceous structural evolution of the Tromsø Basin which explains the formation of compressional features during rifting in the south-western Barents Sea.
2D gravity modelling and 2D structural restoration along a key regional composite seismic section, facilitated the interpretation and assessment of geodynamic constrains for the deeper structures below the Lower Cretaceous. These reveal thinner crust below the Tromsø Basin as compared to the Sørvestnaget and Hammerfest basins, which is suggested as the result of oblique extension in the southwestern margin of the Barents Sea.

In the third paper and at a smaller scale, we integrate stratigraphic and structural observations with throw backstripping and time thickness maps to define the growth processes of a basin-bounding normal fault in the northern Polhem subplatform. During the initial Early Cretaceous rifting, the fault system consisted of at least five en-echelon segments, which were ca. 5–10 km long. Throw backstripping indicates that fault segments were hard-linked after this initial stage to form a single 40 km long fault zone. Cross fault incised valleys provide additional information on the topographic response to fault growth. Major valley incisions at the fault linkage zones outline the extent of the individual fault segments and support early isolated fault growth.

The fourth paper focuses on a genetic correlation of the Lower Cretaceous succession between the north-central and south-western Barents Sea and Svalbard. The structural framework defined in paper I is used to locate the main sediment routes and progradation directions. The latest Valanginian to earliest middle Albian sequences in the offshore Barents Sea are correlated with the onshore Rurikfjellet, Helvetiafjellet and Carolinefjellet formations in Svalbard. This results in the reconstruction of four paleogeographic maps that show the progressive evolution and sediment distribution over the Norwegian Barents Sea for: (1) the earliest Valanginian, (2) the latest Hauterivian, (3) the middle to late Barremian and (4) the latest Aptian.

In the fifth paper, three tectonic events are suggested to control the deposition of the diachronous Lower Cretaceous elastic wedges around
the Loppa High: 1) the latest Jurassic – earliest Cretaceous uplift of the Loppa High which triggered the deposition of the older wedges progressively eastwards in the northern Hammerfest Basin; 2) the late Barremian–Aptian faulting episode in the western flank of the Loppa High, which resulted in the deposition of shallow and probably deep marine wedges; and 3) the latest Aptian to earliest Albian tilting of the Hammerfest Basin and the Loppa High, which modified the sedimentation patterns in the region.

The results of this research can be applied beyond the Barents Sea, as they provide insights into margins and basins evolution, specifically on how: (1) oblique deformation along margins can control the inversion of pre-existing structures and routing of sediments, as well as modify paleogeography; (2) the growth of basin-bounding normal faults can affect sedimentation, with incised channels reflecting the early stage of fault growth; (3) paleogeographic reconstructions that reflect both the tectonic and stratigraphic setting can be used to understand sand distribution and sediment partitioning.
Table of Contents

Preface ................................................................................................................. iii
Acknowledgements ............................................................................................... iv
Abstract ................................................................................................................. vi
Chapter 1 .............................................................................................................. 10
  1 Introduction ........................................................................................................ 11
  2 Research aim and objectives .............................................................................. 18
  3 Study area and geological setting ....................................................................... 21
  4 State of the art ..................................................................................................... 24
  5 Summary of papers ............................................................................................. 27
  6 Discussion .......................................................................................................... 37
  7 Recommendations for future work .................................................................... 43
  8 Conclusions ........................................................................................................ 44
References ............................................................................................................ 45
Chapter 2 .............................................................................................................. 58
Paper 1 ................................................................................................................. 59
Paper 2 ................................................................................................................. 76
Paper 3 ................................................................................................................. 134
Paper 4 ................................................................................................................. 150
Paper 5 ................................................................................................................. 175
Appendices .......................................................................................................... 194
Chapter 1
Introduction

1 Introduction

The Norwegian Barents Sea is part of an epicontinental basin (Fig. 1A) that exhibits a variety of tectonic regimes and structural architectures along its margins. Its tectonic history is mainly attributed to: 1) the Late Palaeozoic initial rifting that formed NE-SW striking rift basins; 2) the Late Jurassic – Early Cretaceous North Atlantic rifting which rejuvenated inherited structures; and 3) the Late Cretaceous – Paleogene strike-slip and extensional tectonics, which dominated the western and northern margins respectively (Breivik et al., 1998; Doré, 1991; Faleide et al., 1993; Gudlaugsson et al., 1998; Minakov et al., 2012; Ritzmann and Faleide, 2007; Ryseth et al., 2003).

The Late Jurassic – Early Cretaceous tectonic processes are related to changes and reorganizations in plate tectonic configurations in the North Atlantic and Arctic regions (Lawver et al., 2002). Plate tectonic models for this time are uncertain due to the lack of constrains (e.g. lack of age control of magnetic anomalies and limited subsurface data; Hosseinpour et al., 2013; Rowley and Lottes, 1988; Seton et al., 2012). Therefore, the Early Cretaceous geodynamic processes related to the propagation of the North Atlantic rifting, the formation of the Canada Basin, and the influence of the High Arctic Large Igneous Province (HALIP) are some of the main tectonic events that modified the entire structural and paleogeographic setting of the Norwegian Barents Sea (Bryn et al.; Glørstad-Clark et al., 2011; Grogan et al., 1998; Grogan et al., 1999; Grundvåg and Olaussen, 2017; Henriksen et al., 2011; Kayukova and Suslova, 2015; Worsley, 2008). Therefore, understanding of these regional tectonic processes in the context of the structural and stratigraphic development of the Norwegian Barents Sea is crucial to better constrain the timing of tectonic events, geodynamic processes and plate kinematics of the North Atlantic and Arctic regions.
This study is a part of larger research project named “Lower Cretaceous Basin studies in the Arctic” (LoCrA; http://locra.ux.uis.no/), which is a consortium between industry and academia with the aim to enhance the knowledge of the tectonic configuration and basin infill in the Arctic during the Early Cretaceous. This study is focused on various scales of observation from margin to sub-basins in order to understand the interaction between tectonics and sedimentation, and involves the following problems:
Figure 1A) Main structural elements of the Barents Sea. The polygons highlight the location of the study areas of this research. Papers 1 and 4 (red polygon) are focused on the larger scale of the north-central Barents Sea. B) Papers 2, 3 and 5 (grey, blue and red polygons) are focused on a basin scale in the southwestern Barents Sea.
1.1 **Problem 1. Distal impact of margin deformation to an intra-cratonic basin and development of drainage systems**

The northern margin of the Barents Sea has been less studied as compared to the other margins (e.g. southwestern Barents Sea; Fig. 1A). This is mainly due to limited data availability and the fact that the area is restricted for any commercial exploration. The structural evolution of the area is a key element for understanding the complex plate tectonic configuration of the Arctic region during the Early Cretaceous (Fig. 2). Most authors agree that during the earliest Cretaceous, the northern margin of the Barents Sea was dominated by compressional tectonics that resulted in the formation of NE oriented structural highs and anticlines due to reverse reactivation of the Late Paleozoic normal faults (Faleide et al., 2008; Grogan et al., 1998; Grogan et al., 1999). This resulted in SW and SE progradation of the Lower Cretaceous clastics today outcropping in Svalbard and Franz Josef Land (Glørstad-Clark et al., 2011; Henriksen et al., 2011; Worsley, 2008). However, this event is poorly described and its link to the tectonic processes in the Arctic region remains unknown.
Introduction

Figure 2 Plate tectonic reconstruction for the Barents Sea during the Hauterivian (130 Ma). From a plate tectonic model provided by the “Plates” project at the Institute for Geophysics, University of Texas. The map shows the major tectonic events during the Early Cretaceous along the Barents Sea margins (red arrows and stippled lines).

1.2 Problem 2. Tectonic basin development and its impact on sedimentation along the basin margin

In the southwestern margin of the Barents Sea, the propagation of the North Atlantic rifting resulted in extensional tectonics with the development of deep basins and highs (Clark et al., 2014; Faleide et al., 2008; Gabrielsen et al., 1990; Indrevær et al., 2016; Rønnevik et al., 1982) (Fig. 1B). The interpreted structural framework of the Tromsø
Introduction

Basin consists of faults which cannot be entirely explained by a stretching direction perpendicular to the main rift trend, and hence the evolution of some structures involving compression (e.g. Senja Ridge, Loppa and Veslemøy highs, Tromsø Basin) remains controversial (Faleide et al., 1993; Gabrielsen and Færseth, 1988; Indrevær et al., 2013; Riis et al., 1986) (Figs. 3a – 3c). It has also been suggested that the complex structural configuration and sedimentation of the southwestern Barents Sea was influenced by inherited Caledonian or even older Precambrian basement structures (Barrère et al., 2009; Braathen et al., 1999; Doré, 1991; Fichler et al., 1997; Gabrielsen, 1984; Gernigon et al., 2014; Ritzmann and Faleide, 2007; Tsikalas et al., 2012). However, despite the apparent continuity and alignment of these structures with lineaments identified in the gravity or magnetic data (Tsikalas et al., 2012; Gernigon et al., 2014; Indrevær et al., 2013), it is not clear how pre-existing basement faults controlled the evolution, architecture and sedimentation of the Tromsø Basin.

Figure 3. Simplified sketch of previously proposed regional tectonic models for the Late Jurassic - Early Cretaceous tectonic evolution of the Tromsø Basin. Notice the differences in the Senja Ridge and Veslemøy High interpretations, as compressional structural features are formed by either a) sinistral and b) dextral strike-slip faulting along the Bjørnøyrenna and Ringvassoy fault complexes (Riis et al., 1986; Gabrielsen and Færseth, 1988), or c) regional extensional system with sinistral strike-slip movement along the Bjørnøyrenna fault complex (Faleide et al., 1993)
1.3 Problem 3. Impact of basin bounding normal faults evolution on sediment dispersal

During the Early Cretaceous, active and growing normal fault systems in the southwestern Barents Sea controlled the distribution of the Lower Cretaceous clastic wedges along major fault complexes (Glørstad-Clark et al., 2011; Henriksen et al., 2011; Seldal, 2005; Sund et al., 1986; Wood et al., 1989) (Fig. 1B). Most of the studies in the southwestern Barents Sea have been focused on the deposition of clastic wedges along major faults or structural highs to infer the timing of fault activity and the stage of rift development (Knutsen et al., 2000; Marín et al., 2018; Prosser, 1993). These studies mainly assess the final fault geometries and displacements, and rarely look at the impact of fault evolution on the topographic and sedimentary response (Cartwright et al., 1995; Mansfield and Cartwright, 2001; Peacock and Sanderson, 1991). At a smaller scale than that of the northern and southern margins (problems 1 and 2), assessing the history of growth of basin bounding normal faults is important to understand changes in basin paleo-topography during fault evolution, as it can provide information about early sedimentary entry points and drainage areas (Gawthorpe and Leeder, 2000).
2 Research aim and objectives

This research focuses on multi-scale observations in the north-central and the southwestern Barents Sea from (1) far field tectonic effects on the Barents Sea margins, to (2) basin scale structural development, and to (3) individual fault segments evolution with implications for sediments dispersal (Figs. 1a and 1b). Considering this, the main objectives are:

1. Document the structural and stratigraphic evolution of the north-central Barents Sea during the Early Cretaceous, including the understanding of the mechanisms that controlled compressional tectonics in the area and its impact on paleogeography. Also improve the understanding of the regional tectonic processes in the Arctics region (e.g. opening of the Canada Basin) and how these processes affected the study area.

2. Describe the evolution, geometry and structural style of the major faults of the Tromsø Basin and their influence on deposition of the Lower Cretaceous sedimentary sequences. This contributes to the understanding of the geodynamic processes in the southwestern Barents Sea, and explains the formation of compressional features in this area.

3. Understand the structural mechanisms controlling the sedimentation patterns and variation of depositional environments around the Loppa High. This contributes to a better knowledge of tectonic and sedimentation in complex areas which experienced more than one phase and multiple directions of extension.

This study is multidisciplinary and it integrates seismic, potential field and well data interpretation, sedimentology and biostratigraphy. To achieve the above goals, we use a subsurface dataset of 2D and 3D
seismic data and wire line logs, which were provided by the Norwegian Petroleum Directorate, MultiClient Geophysics and WesternGeco.

The study comprises three main articles in which I am the first author, and two additional articles led by Dora Marin and Sten-Andreas Grundvåg, respectively. The main three articles target specific problems related to the structural style and kinematics of basin margins and their bounding faults. The additional two articles are related to the integration of the sequence stratigraphic and tectonic framework of the Barents Sea during the Early Cretaceous. To meet the specific objectives of each paper, the research was performed as follows:

In the first paper, a regional subsurface study of the north-central Barents Sea was performed. Detailed mapping of major faults and structural elements on the Norwegian and the Russian Barents Sea resulted in a holistic understanding of the various regional tectonic processes in the Arctic region, which affected the northern margin of the Barents Sea including its paleogeography.

In the second paper, a basin scale subsurface study was performed in the Tromsø Basin and SW Barents Sea. The main emphasis was given to the interpretation of the fault network and detailed timing of fault movement, and the relation with the Lower Cretaceous sedimentary sequences. Also, gravity modelling along a regional composite seismic section, followed by structural restoration of this section that helped to constrain the basin configuration in the context of the geodynamic processes in the SW Barents Sea.

The third paper is based on a detailed 3D seismic interpretation in the Polhem Sub Platform, SW Barents Sea. Detailed mapping of the footwall and hanging wall stratigraphy helped to describe the sequential growth of a basin bounding normal fault and how it controlled sediment distribution and dispersal patterns during several phases of extension.
Research aim and objectives

In the fourth paper led by Sten-Andreas Grundvåg, the Early Cretaceous structural and stratigraphic framework of the offshore Barents Sea was integrated with that from Svalbard. As a result, a tectonostratigraphic link between the southwestern Barents Sea and Svalbard is discussed.

In the fifth paper led by Dora Marin, a 2D and 3D seismic interpretation was performed around the Loppa High, SW Barents Sea. The tectonic control on sedimentation patterns around the Loppa High is discussed.
Study area and geological setting

The research was carried out in two margins of the Barents Sea: (1) the north-central and (2) southwestern margins. These two margins are subdivided into basins, platforms and structural highs (Fig. 1).

3.1 The north-central Barents Sea

The north-central Barents Sea covers the offshore area between Svalbard and the northern part of Novaya Zemlya (Fig. 1a). As mentioned before, this area is poorly studied as compared to the remaining part of the Barents Sea (e.g. southwestern Barents Sea).

Previous work in the region has documented a compressional event that resulted in tectonic inversion during the Late Jurassic – Early Cretaceous (Grogan et al., 1999). This compression resulted in reverse reactivation of Late Palaeozoic, NE-SW and E-W striking normal faults (Fig. 2) (Antonsen et al., 1991; Grogan et al., 1998; Grogan et al., 1999; Nikishin et al., 2014; Nikishin, 2013). Lower Cretaceous clinoforms in the southern Barents Sea reveal clastic source located to the NW and NE which builds the shelf southwards (Grundvåg and Olaussen, 2017; Kayukova and Suslova, 2015; Marin et al., 2017). These northerly to southerly progradation of the Lower Cretaceous clastic materials was related to uplift, formation of structural highs and anticlines in the north-central Barents Sea (Kayukova and Suslova, 2015; Olaussen et al., 2019; Smelror et al., 1998). The north-central Barents Sea was also affected by the formation of the High Arctic Large Igneous Province (125–122 Ma), which resulted in extrusive magmatism and formation of WNW–ESE trending dykes (Corfu et al., 2013; Dibner, 1998; Evenchick et al., 2015; Polteau et al., 2016; Senger et al., 2014).

During the Cenozoic, transpressional and transtensional deformation occurred between NE Greenland and the western Barents Sea. This deformation was responsible for the formation of the Vestbakken
study area and geological setting

provinces and the Svalbard fold and thrust belt (Bergh and Grogan, 2003; Faleide et al., 2008). These events modified the structural configuration of the north-central Barents Sea, by amplifying several structural highs and basins (Anell et al., 2014; Grogan et al., 1999). This was followed by glaciation and a tectonic uplift which caused erosion and exhumation of the northern Barents Sea (Dimakis et al., 1998; Knies and Gaina, 2008).

3.2 The southwestern Barents Sea

The southwestern Barents Sea is located offshore of the north-western corner of the Norwegian mainland (Fig. 1b). Starting from the Late Palaeozoic, regional extension between Greenland and Norway resulted in the formation of NE–SW and E–W trending grabens and half grabens that were covered by Upper Carboniferous to Lower Permian carbonate platforms and thick evaporites (Gudlaugsson et al., 1998; Larssen et al., 2002). The Early Triassic is marked by a rift episode, which has been documented in the North Atlantic region (Tsikalas et al., 2012). This rifting episode may have continued until the Middle Triassic (Smelror et al., 2009). During the Middle Jurassic – Early Cretaceous, northward advance of the Atlantic rifting enhanced a NE–SW and E–W Late Palaeozoic fault system and formed deep basins in the southwestern Barents Sea such as the Harstad, Tromsø, Bjørnøya and Sørvestnaget basins (Fig. 2) (Faleide et al., 2008; Gernigon et al., 2014). The Early Cretaceous rift episode along the NE–SW and E–W trending fault complexes (e.g. Ringvassøy–Loppa, Bjørnøyrenna, Asterias and Troms–Finnmark) led to rapid subsidence and accumulation of the Lower Cretaceous sediments (Clark et al., 2014; Faleide et al., 2008); (Gabrielsen et al., 1990; Indrevær et al., 2016; Rønnevik et al., 1982). The Tromsø, Sørvestnaget and Bjørnøya basins experienced salt related deformation during this rifting event (Gabrielsen et al., 1990; Larssen et al., 2002; Sund, 1984). Three Early Cretaceous rift phases have been interpreted in the southwestern Barents Sea: Berriasian–Valanginian,
Study area and geological setting

Hauterivian–Barremian and Aptian–Albian (Faleide et al., 1993). Local compression during the earliest Cretaceous has been identified in the northern part of the Tromsø Basin. This has been suggested to be the result of dextral strike slip movement along the Asterias Fault complex (Berglund et al., 1986; Gabrielsen et al., 1990; Sund, 1984), or localized tectonic inversion due to differential uplift of the Loppa High (Indrevær et al., 2016).
State of the art

4 State of the art

This section is a short review of previous studies regarding (1) oblique deformation and (2) fault growth styles along basin margins.

4.1 Oblique deformation

Commonly, oblique deformation occurs along margins where the extension direction is not orthogonal to the rift (Dewey et al., 1998; Sanderson and Marchini, 1984). The influence of obliquity on the structural styles of rift systems varies. This is often due to the rift setting, which is mainly controlled by tectonic inheritance (Hodge et al., 2018; Manatschal et al., 2015; Morley, 2017; Phillips et al., 2018), or from changes in crustal composition and configuration (Brune et al., 2017; Molnar et al., 2017; Mondy et al., 2018; Sippel et al., 2017). It is difficult to interpret oblique deformation using 2-D plane strain (Brune et al., 2018). However, there are some key characteristics that can be attributed to this process, for instance segmented en échelon border faults oblique to the rift trend (Agostini et al., 2009; Brune and Autin, 2013; Clifton et al., 2000; Corti, 2008; Withjack and Jamison, 1986), or uncommon crustal thinning (e.g. sharp transitions) along the margin (Montési and Behn, 2007).

In the Norwegian Barents Sea, propagation of the North Atlantic rifting from the southwest towards the north-central margins was aborted during the Cretaceous (Faleide et al., 2008) (Fig. 2). This has been associated with complete reorganization of crustal extension which led to oblique deformation in the southwestern Barents Sea (Faleide et al., 2008; Gernigon et al., 2014). Early Cretaceous oblique deformation in the southwestern parts of the margin is partially evident in the Tromsø and Bjørnøya basins, where the fault trends are oblique to the regional, inherited structural grain (Breivik et al., 1998; Gabrielsen et al., 1990; Gernigon et al., 2014; Henriksen et al., 2011; Ritzmann and Faleide, 2009; Smelror et al., 2009). Most of the plate tectonic reconstructions for
the Early Cretaceous place the Canada Basin adjacent to the northern margin of the Barents Sea (Barnett-Moore et al., 2018; Doré et al., 2015; Seton et al., 2012). Opening of the Canada Basin (~145–126 Ma) resulted in large scale crustal up-doming which affected the northern margin of the Barents Sea (Alvey et al., 2008; Grogan et al., 1999). The models for opening of the Canada Basin are still a matter of debate, and they are supported by inconclusive or indirect observations (Cochran et al., 2006; Døssing et al., 2013; Dove et al., 2010; Lawver and Scotese, 1990). Recent studies (Alvey et al., 2008; Hadlari et al., 2016) reveal evidences supporting oblique deformation along the northern margin of the Barents Sea (e.g. northern margin of the Lomonosov Ridge; Evangelatos and Mosher, 2016; Gaina et al. 2014). These studies document Early Cretaceous oblique deformation in the context of regional tectonic processes along the margins of the Norwegian Barents Sea. However, no studies have been conducted to understand the impact of oblique deformation on inherited basins and sedimentation.

4.2 Fault growth and linkage

Observations from outcrop and subsurface datasets, and analogue and numerical models suggests two main ways of fault growth: (1) the isolated fault model, where growth and linkage of individual fault segments occur through displacement and lateral propagation of their tiplines (Cartwright et al., 1995; Dawers and Anders, 1995; Dawers et al., 1993; Walsh and Watterson, 1988; Watterson, 1986) (Fig. 4a), and (2) the constant length fault model, where faults reach their near-final length relatively early in their slip history, and accumulation of displacement occurs without further lateral tipline propagation (Childs et al., 2003; Giba et al., 2012; Jackson and Rotevatn, 2013; Morley, 2002; Nicol et al., 2016; Schlagenhauf et al., 2008; Tvedt et al., 2016; Walsh et al., 2003; Walsh et al. 2002) (Fig. 4b). In the last 30 years, these two models have been a matter of discussion and debate, as the styles of fault growth and rate of tipline propagation impact the tectono-stratigraphic
development of sedimentary basins (Gawthorpe and Leeder, 2000; Henstra et al., 2016; Jackson et al., 2017). The character of the initial stage of fault growth remains unclear, since very few studies have been able to capture the earliest (and short lived) stage of fault growth (Schlagenhauf et al., 2008) (Fig. 1c).
In the Norwegian Barents Sea, most of the major fault complexes have been analyzed with the aim of establishing fault geometry, architecture and processes controlling faulting (Braathen et al., 2009; Fisher and Knipe, 2001; Gabrielsen et al., 1990; Gabrielsen et al., 2016). To our knowledge, there are no studies in the Barents Sea documenting how fault growth affects sediment distribution (Fossen and Rotevatn, 2016; Torabi et al., 2019).

Figure 4 Top views illustrating the (A) isolated and (B) constant length models of fault growth. Numbers represent fault growth stages: (1) initiation, (2) interaction, and (3) linkage. (C) Displacement versus length through time for the two models.
5 Summary of papers

5.1 Paper 1: Early Cretaceous tectonostratigraphic evolution of the north-central Barents Sea

In this paper, we investigate the structural evolution of the north-central Barents Sea during the Early Cretaceous, and the influence of fault activity on sedimentation in the area. This is achieved by integrating 2D seismic data, two exploration wells, and information of shallow cores from the Norwegian and Russian sectors. As result of this work (Fig. 5), three fault families, two Lower Cretaceous seismic sequences, and seven seismic facies, are interpreted in the area. During the Hauterivian–Early Barremian (sequence 1), a syn-tectonic phase is observed, where fault families 1 and 2 of Late Paleozoic age were reactivated as reverse faults and induced the uplift of NE–SW and E–W structural highs on the Kong Karls Land Platform and the North Barents Basin. During the Early Barremian–Early Aptian (sequence 2), the study area experienced a tectonically quiescent period, where the increase of clastic supply from the N–NE was responsible for sediment progradation towards the S–SW Barents Sea. The progradation was controlled and routed by structural highs inherited from the Hauterivian–Early Barremian inversion. Later, a post Early Cretaceous reactivation was responsible for the reactivation of the Late Jurassic – Early Cretaceous inverted faults and structures. Our results suggest that three main regional tectonic events controlled the inversion of the Late Paleozoic faults, resulting in development of structural highs in the north-central Barents Sea (Fig. 5): 1) dextral transpression along Novaya Zemlya, which was responsible for inversion on the ESE flanks of the North Barents Basin; 2) dextral movement along a paleo-boundary of the northern margin of the Lomonosov Ridge during opening of the Amerasia Basin, which controlled the inversion in the Kong Karls Land platform and the Olga Basin; and 3) a compressional event in the present day NE Greenland,
and Ellesmere Islands and the NW Barents Sea (NW Svalbard), which contributed to uplift in Svalbard and inversion in the rest of the study area.

Figure 5 Plate reconstruction map from the “Plates” project (https://ig.utexas.edu/marine-and-tectonics/plates-project/) for the Hauterivian (130 Ma) overlain with the interpreted structural elements in paper 1 and integrated with previous studies.

5.2 Paper 2: The Early Cretaceous evolution of the Tromsø Basin, SW Barents Sea, Norway

Extensional basins developed along oblique or transform margins are less studied as compared to those basins developed along orthogonally extended margins. This study presents an example of a basin developed along an oblique margin, namely the Tromsø Basin, which developed along the southwestern Barents Sea transform margin. Three previous models have been proposed to explain the tectonic evolution and architecture of the basin, but still there is no consensus on the development of individual structures and compressional faults in this basin. In this study, we use fairly new 2D industry seismic reflection
data, potential field and well data, as well as previously published information, to understand the Early Cretaceous structural evolution of the Tromsø Basin in the context of the geodynamic processes in the southwestern Barents Sea. Modeled gravity anomalies along a depth converted 2D regional seismic section facilitated the interpretation of crustal structures, which then were structurally restored. We propose a revised Early Cretaceous structural model for the Tromsø Basin, which involves oblique extension and formation of an intra-basinal, transpressional transfer zone (Figs. 6a – 6c). This can explain reverse faulting in the study area. Basement heterogeneity played an important role in focusing and localizing strain. 2D sequential restoration of a regional profile above yields an estimate of ca. 35 km of crustal extension in the SW Barents Sea margin, from the earliest Cretaceous until the present, which is relatively smaller than previous estimations (e.g. 85 km by Breivik et al. 1998). Discrepancies are attributed to the differences in the calculation methods, where our results were based on 2D structural restoration, and Breivik et al., 1998 derived extension from crustal stretching factor. Moreover, from the earliest Cretaceous until Albian (seismic unit 2), the Tromsø and Sørvestnaget basins developed as a single large basin in the SW Barents Sea margin. Crustal thickness along the gravity modeled 2D regional section displayed a thinner crust below the Tromsø Basin as compared to the Sørvestnaget and Hammerfest basins. This is considered as uncommon for orthogonally rifted passive margin models and observations, where crustal thickness typically decreases towards (e.g. Sørvestnaget Basin) the continent – oceanic boundary (Peron-Pinvidic et al., 2013). Therefore, we suggest that the abnormal crustal thickness within the necking zone area is the result of oblique rifting and segmentation in the margin, where increase in obliquity decreases stretching and crustal thinning (Montési and Behn, 2007). This study illustrates the importance of detailed and regionally integrated analysis of rifted basins for reconstructing their evolution, as analysis of oblique rifted basins using two-dimensional plane strain can lead to erroneous assessment of faulting style and deformation.
Summary of papers

Figure 6. Proposed structural model for the Tromsø Basin and distribution of the main depocenters (grey polygons) during Early Cretaceous: A) Valanginian – Late Barremian extension was accommodated by west dipping boundary faults of FF1 (e.g. fault segments of TFFC and BFC), which resulted in the formation of the internal fault system FF2; B) The Aptian – Albian is marked by a transpressional setting along a transfer zone which is related to the oblique opening of the Tromsø Basin, where basement heterogeneity localized strain; and C) The Cenomanian is considered a tectonically quiescent period, where most of the fault activity occurred in the western and north-western flanks of the Tromsø Basin.

5.3 Paper 3: Growth and linkage of a basin-bounding fault system: Insights from the Early Cretaceous evolution of the northern Polhem Subplatform, SW Barents Sea

Observations from outcrop and subsurface datasets, as well as physical and numerical models suggest two ways of fault growth: (1) growth and linkage of individual fault surfaces through lateral propagation of the tip-lines (isolated model), or (2) near-final fault length formed relatively early in the slip history and displacement accumulation without lateral propagation of the tip-lines (constant-length model). This study integrates stratigraphic and structural observations with throw backstripping and time thickness maps to define the growth of a normal
fault in the northern Polhem Subplatform, SW Barents Sea (Figs. 1b and 7a – 7d). During the initial 15 My of Early Cretaceous rifting, the studied fault was comprised of at least five en-echelon segments (ca. 5–10 km long). Throw backstripping indicates that these fault segments were hard-linked after this initial stage to form a single 40 km long fault (Fig. 7d). Major incised valleys coincide with the location of the fault linkage zones and outline the extent of the individual fault segments, supporting early isolated fault growth (Fig. 7c). Based on fault throw backstripping, valley incision was able to keep up with fault slip, such that it remained unaffected by the fault linkage stage. This study highlights the importance of integrating stratigraphic and structural observations during reconstruction of fault growth history, where syn-rift erosional features, sediment thickness variations, sediment distribution, stratal geometries and onlaps/truncations are critical for estimating the growth of these structures.
Figure 7 (A) Time thickness map along the studied fault in the Polhem Subplatform showing distribution of depocenters. (B) Composite line along the fault (red line in A). (C) Composite line in B flattened to the top S1 horizon, illustrating the distribution of scoop-shaped depocenters in the S0 interval. (D) Interpreted paleo-location of isolated fault segments and hanging wall depocenters.

5.4 Paper 4: The Lower Cretaceous succession of the western Barents Shelf: onshore and offshore correlations

This paper was led by Sten-Andreas Grundvåg. My main contribution was related to the correlation of the Lower Cretaceous sequences and providing examples and descriptions of the clinoforms in the north central Barents Sea. In this paper, we integrate biostratigraphic analysis, outcrop data and seismic and well information of the north-central Barents Sea, with the aim of establishing a genetic link of the Lower Cretaceous successions onshore and offshore. In addition, this study
discusses the regional paleogeography, depositional controls, sediment partitioning and sand distribution in the area. This information is key to understand the basin infill and the sedimentary processes that occurred in the western part of the Barents Sea during the Early Cretaceous. We suggest that three sequences defined in the southwestern Barents Sea, with an age of latest Valanginian–earliest middle Albian (S1–S3), correlate with the Rurikfjellet (Valanginian – Hauterivian/early Barremian), Helvetiafjellet (early Barremian – early Aptian ) and Carolinefjellet (early Aptian, middle Albian) formations in Svalbard. Based on age control, we propose that the Barremian clinoforms (sequence 1) identified in the western Olga Basin, Fingerdjupet subbasin and western part of the Bjarmeland platform correlate with the upper part of the Rurikfjellet Formation and a Barremian unconformity identified in Svalbard (Figs. 8a and 8b). In addition, the southeastward progradation direction of these offshore clinoforms reflect a similar pattern that the paleocurrents of the Rurikfjellet and Helvetiafjellet formations. This suggests that the offshore and onshore depositional system were under the influence of the same paleo-drainage. The apparent lack of sandstone in the shelf-margin clinoforms is interpreted as a result of the physiographic conditions of the basin, such as storm waves, tidal and alongshore currents. These conditions may have contributed to the sand being trapped in areas such as the inner shelf. Finally, four paleogeographic reconstructions are made: 1) the earliest Valanginian, characterized by a carbonate platform, sediment starvation and the development of clastic wedges in basins such as the Hammerfest Basin; 2) the latest Hauterivian, when Greenland is proposed as the source of the southeastward directed shallow marine wedges in the western part of the study area; 3) the middle to late Barremian, characterized by a fluvio-deltaic system triggered by the uplift of the northern Barents Sea; and 4) the latest Aptian, when the main platform areas were flooded and a seaway connected the Barents Sea and the Canada Basin.
Summary of papers

Figure 8 Summary of paper 4 showing the suggested A) offshore and B) onshore genetic link of the Lower Cretaceous sequences (for more information the reader is referred to the full article).

5.5 Paper 5: Effects of adjacent fault systems on drainage patterns and evolution of uplifted rift shoulders: The Lower Cretaceous in the Loppa High, southwestern Barents Sea

This paper was led by Dora Marin. In this study, we integrate the information from the previous papers, in addition to new observations from the western flank of the Loppa High, in order to describe the distribution and timing of diachronous clastic wedges around the Loppa High (Fig. 9). Additionally, this paper aims to understand how multidirectional and diachronous tectonic activity in the area conditioned the Lower Cretaceous sedimentation. Based on detailed mapping of seismic wedges within a chronostratigraphic framework, and
palynological analysis, we propose that three events controlled the distribution of the Lower Cretaceous wedges: 1) an uplift event of the Loppa High during the latest Jurassic–earliest Cretaceous (Sund et al., 1986; Berglund et al., 1986; Wood et al., 1989; Glørstad-Clark, 2011; Clark et al., 2014), which deposited progressively younger wedges towards the east of the Hammerfest Basin as result of lateral and vertical fault propagation. This induced eastward switching of the sediment input points. The northernmost part of the Loppa High is interpreted as a local depocenter during the early Barremian, due to the proximity of clinoform progradation. 2) Faulting in the western flank of the Loppa High along the Ringvassøy-Loppa and Bjørnøyrenna fault complexes, which triggered the deposition of syn-rift wedges during the late Barremian–Aptian. The wedges were partially deposited in shallow marine environments, but probably also in deep marine environments. An upper Barremian to lower Aptian syn-rift unconformity is interpreted in the western flank of the Loppa High and in the Fingerdjupet Sub-basin. 3) A renewed uplift and eastwards tilting event of the Loppa High and Hammerfest Basin during the late Aptian–early Albian. This event is supported by: the eastward migration of the depocenter location, a deflection towards the east of submarine fans deposited in the northwestern part of the basin, an unconformity in the western and southwestern flanks of the Loppa High, and progressively deeper environments towards the eastern part of the Hammerfest Basin and the Bjarmeland platform. The last observation is based on the height of the clinoforms (80–200 m in the eastern part of the Hammerfest Basin and > 500 m in the Bjarmeland Platform). This event redirected the drainage system away from the Tromsø Basin towards a gentler slope, where it sourced the clinoforms in the northeastern part of the Hammerfest Basin. Fault activity in the western flank of the Loppa High contributed to the uplift of the northernmost part of the Loppa High.
Figure 9. 3D cartoons illustrating the three main events controlling the deposition of the clastic wedges around the Loppa High. For a more detailed explanation of these figures, please see paper 5.
6 Discussion

This section describes the contribution of the thesis to the knowledge of the tectonostratigraphic evolution of the Norwegian Barents Sea margins, as well as the global implications of this study. Specifically, we discuss the implications of both, margins and basin scale deformation on:

1. deposition of the Lower Cretaceous sediments,
2. fault growth and the physiographical and tectonostratigraphic evolution of rift basins,
3. the variables controlling the bypass of coarse-grained sediments into the basin, and
4. regional paleogeography.

6.1 Implications of margin-scale oblique deformation on structural styles

Oblique deformations produce 3D strain which cannot be characterized by simplified 2D plane strain (Brune et al., 2018). In the southwestern and north-central Barents Sea, most of the Early Cretaceous compressional features have been analyzed assuming 2D plane strain, where the analyzed cross section is parallel to the postulated contraction. This leads to a poor explanation of the compressional structures in the context of the overall Late Jurassic – Early Cretaceous tectonic setting of the margins (Antonsen et al., 1991; Grogan et al., 2000; Grogan et al., 1999; Faleide et al., 1993; Gabrielsen et al., 1990; Rønnevik et al., 1982), though several attempts have been made to relate the compression to basement heterogeneity, which could be locally responsible for the change of strain (Barrère et al., 2009; Braathen et al., 1999; Doré, 1991; Fichler et al., 1997; Gabrielsen, 1984; Gernigon et al., 2014; Ritzmann and Faleide, 2007; Tsikalas et al., 2012; Indrevær et al., 2016). In this research, compilation of the regional tectonic events, deformation patterns, mapping of key faults and their associated structures, and mapping of the Lower Cretaceous clastic wedges allowed us to constrain in more detail the tectonic events that operated during this period.
Discussion

In the north-central Barents Sea (paper 1), the interpreted compression along NE – SW trending faults is caused by the counterclockwise opening of the Canada Basin (Grantz et al., 1998; Lawver et al., 2002; Shephard et al., 2013). This was responsible for the formation of restraining and releasing bends along the paleo-position of the Lomonosov ridge (Evangelatos and Mosher, 2016; Minakov et al., 2012). We suggest that reactivation/inversion of the inherited Late Paleozoic normal faults is the result of transpressional deformation along the northern margin. These processes are very similar to those observed along the present day Dead Sea transform fault (DSTF; Weber et al., 2009). Particularly, in the northern part of the DSTF, transpressional deformation produced restraining bends which resulted in the formation of several anticlines that are oblique to the DSTF (Gomez et al., 2007). This is an analogue of far field strain caused by oblique deformation along margins, which reactivate inherited weak fault zones in reverse mode.

In the southwestern Barents Sea (paper 2), plate tectonic reorganization during the Early Cretaceous resulted in progressive changes in the direction of extension (Lawver et al., 2002; Barnett-Moore et al., 2018; Dore et al., 2016; Seton et al., 2012). The latest plate tectonics models by Barnett-Moore et al., 2018, suggest that from 200 Ma until 80 Ma, the plate tectonics movement between Greenland and Norway had mainly a NW – SE direction, which shifted at 80Ma to an almost N – S direction. Hence, we suggest that before shifting to the N – S direction, the southwestern Barents Sea margins was subjected to oblique deformation that affected the basin evolution. The proposed oblique opening of the Tromsø Basin generated secondary intra-basinal normal faults (Gernigon et al., 2014; Faleide et al., 2008), which are oblique to the inherited fault network (consistent with Bonini et al. 1997 and McClay and White, 1995). Compressional faulting in the northern Tromsø Basin can be explained as an intra-basinal, transpressional transfer zone, which overall
fits the oblique opening of the basin (McClay et al., 2002; McClay et al., 2004). Modeled gravity anomalies along the composite 2D regional seismic section facilitated interpretation of the crustal structures. The distribution of the crustal stretching ($\beta$) factor in the southwestern Barents Sea is unlike orthogonally rifted margins (Peron-Pinvidic et al., 2013). The crust below the narrow and confined Tromsø Basin appears to be thinner than in the more distal Sørvestnaget Basin, thus not follow the expected values proposed for extensional margins (consistent with Breivik et al., 2018, Gernigon et al., 2014; Osmundsen and Peron-Pinvidic, 2018). This may suggest that in addition to the expected thinning of the crust during formation of the margin within the necking zone, Early Cretaceous rifting in the southwestern Barents Sea was involved to a certain degree of obliquity where rift parallel deformation most likely decreased crustal thinning (Crosby et al., 2011; Montési and Behn, 2007). Therefore, it is important to integrate the regional tectonic setting in order to understand the basin-scale faulting style and architecture, particularly for complex margins that were subjected to changes in extension direction. This study could serve as a subsurface analogue for basins that developed during oblique extension with inherited basement structures.

### 6.2 Implications of normal fault growth for the physiographical and tectonostratigraphic evolution of rift basins

The growth history of basin bounding normal faults and interaction with deposition of the Lower Cretaceous clastic wedges are discussed in papers 3 and 5. The two main models of fault growth, isolated versus constant-length, are undistinguishable after the faults have attained their final displacement and length as seen in figure 4 a-b. During the last 30 years, both models have been a matter of discussion and debate (Childs et al., 2017; Jackson et al., 2017). A major difference in these two models
is the early growth history of fault displacement versus length (Figs. 4a and 4b), which requires detailed knowledge of fault evolution. In paper 3, a large normal fault (854 m throw) with good record of syn-sedimentary strata in the hanging wall and footwall was chosen to analyze fault growth. Based on fault throw backstripping, we suggest that initially the fault grew in accordance with the isolated model and its near final length was obtained at \( \approx 37.5\% \) of its slip history. This is longer than the time suggested by recent compilations by Jackson et al. (2017) and Childs et al. (2017), who suggest that final fault length is established within \( \approx 10 – 33\% \) of the fault slip history. Limited vertical seismic resolution (>30 m) and absence of hanging-wall well data introduce additional uncertainties for understanding of the earliest stages of fault growth. Therefore, incised valleys served as key markers for unraveling the growth of the interpreted fault. Thickness map analysis and throw backstripping suggest that fault segments formed earlier than the incised valleys, and hence controlled paleo-drainage, where low areas developed between the fault segments during the early stages of fault growth were exploited by the incised valleys (this is consistent with Gawthorpe and Leeder, 2000). This suggests that the categorical distinction between the isolated versus the constant-length fault growth model may be too simplistic, at least for large basin bounding faults. Detailed interpretation of stratigraphic features, in this case incised valleys, may provide additional information for understating the fault evolution.

### 6.3 Implications for the regional paleogeography

Most sedimentary processes are related to tectonic processes to some extent (Dickinson, 1974). Their direct or indirect relationship can vary from coarse sediments sourced from uplifted areas or fault scarps to fine sediments deposited in broad sheets away from any direct tectonic influence. In the Barents Sea, structural adjustment in the northern margin (e.g. opening of the Canada Basin, HALIP) triggered southward
progradation of Lower Cretaceous clastic material (Marin et al., 2017; Kayukova and Suslova., 2017; Grundvag et al., 2017). Although the main source of the siliciclastics has been suggested to be the area in the N (e.g. North Kara region, Frans Josef Land, etc.) and W-NW (e.g. Greenland), inverted NE – SW striking structural highs in the north-central Barents Sea served as local sediment sources and controlled regional sediment dispersal by funneling fluvio-deltaic systems in a SW direction. This has implications for the paleogeography and tectonic reconstructions of the Arctic. For instance, it implies that during the Early Cretaceous, continental areas were present along the northern edges of the Barents Sea and sourced siliciclastic material to the S and SW Barents Sea.

Previous works in the northern Barents Sea (including the Russian sector) provide general paleogeographic maps for mainly three intervals corresponding to the Valanginian, Barremian and Albian (Smelror et al., 2009 and Worsley, 2008). These maps mainly give information about the location of the continental areas, the shelf and the deep-water environments. In contrast to these previous works, we constructed paleogeographic maps for four time intervals, where mapping of the structurally uplifted and eroded highs, and distribution of clinoforms allowed us to define possible continental areas, deltas and shorelines (papers 1 and 4). The main strength of these paleotectonic and paleodepositional reconstructions is the integration of several geological observations, such as sequence stratigraphy and seismic facies analysis, sedimentological descriptions of core data and outcrops, and biostratigraphy. These paleogeographic reconstructions help to understand the source of siliciclastics and predict the distribution of potential reservoir sandstones in the study area.

6.4 Limitations

Although this research has significant implications for the understanding the tectonic processes in the Norwegian Barents Sea, it is important to
highlight the main limitations related to the data and methods. Highlighting these limitations is essential for future research as it may promote the development of new seismic processing techniques (e.g. de-multiple) and seismic acquisition methods (e.g. shallow water source configurations).

**Seismic data**

It is well known that acquisition of 2D and 3D seismic data in the Barents Sea is often related to hydrocarbon exploration. The north-central Barents Sea is restricted for any hydrocarbon exploration activities. Consequently, it is covered by a sparse 2D seismic grid with average distance ca. 15 km. This makes difficult the seismic interpretation of key horizons and faults. Poor imaging and abundancy of seafloor multiples due to shallow water depths require better processing techniques. Additionally, the 2D seismic sections are often oblique to the main structural lineaments, which affect the understanding of their true geometries (e.g. faults, clinoforms).

**Well data**

A limited amount of exploration wells in the north-central and southwestern Barents Sea contribute to the uncertainty in time-to-depth conversion and structural restoration. The lack of exploration wells makes difficult the correlation between gamma-ray logs and seismic facies. We experienced this limitation in papers 1 and 3 where several seismic facies have not been drilled by exploration wells, and the interpretation of depositional environments was based only on seismic reflectivity and internal architectures.
7 Recommendations for future work

Plate tectonic reorganization can often lead to changes in the stress and strain fields along margins (Brune et al., 2018). Most previous works, including our research in paper 2, focus on specific cases of rifted systems, which involve a certain degree of obliquity (Fournier et al., 2004; Lizarralde et al., 2007; Klimke and Franke, 2016; Phethean et al., 2016). Quantification of rift obliquity through time is more difficult since it requires detailed documentation of syn-rift evolution. Further research should be oriented towards validating such quantifications, as it may provide better kinematics constrains for plate tectonics reconstructions.

The detailed fault growth history from paper 3 indicates that the ongoing debate between the two competing fault growth models (isolated versus constant length) may be too categorical. Some authors claim that there is an overall bias in favor of the isolated fault growth model, while the majority of the natural examples of active or extinct fault systems show characteristics of the constant length model (Nicole et al., 2016; Rotevatn et al., 2018; Rotevatn et al., 2019). Therefore, future research related to the growth of normal faults should be oriented to better document the initial lengthening stages of fault evolution. This might be achieved by integrating high-resolution seismic imaging techniques and well data (e.g. biostratigraphy), which can allow mapping fault structure and associated growth strata (Taylor et al., 2004; Nicol et al., 2005).
8 Conclusions

Based on detailed analysis of subsurface data, this research has improved the geological understanding of the structural elements and depositional patterns of the north-central and southwestern Barents Sea margins. Our main findings are:

1) The inverted pre-existing fault network in the north-central Barents Sea guided the deposition and progradation of the Lower Cretaceous clastics. The interpreted deformation pattern and structural imprint of the area supports a counterclockwise model for the opening of the Canada Basin. This interpretation may contribute to the understanding of how deformation along margins can affect fault evolution and sediment distribution in distal areas.

2) Basins that evolved in an oblique setting (e.g. Tromsø Basin), likely display a complex fault pattern with abnormal crustal thickness and compressional structures that can be easily misinterpreted. Analyzing major basin bounding faults in the context of the overall plate tectonics setting and basin configuration is key to understand the main factors controlling fault distribution.

3) Detailed analysis of a basin bounding normal fault shows that the categorical distinction between isolated versus the constant-length fault growth models may be too simplistic, at least for large basin bounding faults. Analysis of sedimentation or erosional processes (e.g. incised valleys) can provide key information for unraveling the early growth history of these faults.

4) In contrast to previous works, more refined and detailed regional paleogeographic maps for the Norwegian Barents Sea were built. Each time interval reflects the structural and stratigraphic evolution of the area. These paleogeographic maps can help to predict sandstone distribution, and better understand the evolution of the Arctic during the Early Cretaceous.
References


Bjornoya: Implications for Palaeozoic and Mesozoic tectonism of the western Barents Shelf: Norsk Geologisk Tidsskrift, v. 79, p. 57-68.


Brune, S., S. E. Williams, and R. D. Müller, 2018, Oblique rifting: the rule, not the exception: Solid Earth, v. 9, p. 1187-1206.


References


Fichler, C., E. Rundhovde, S. Johansen, and B. Sæther, 1997, Barents Sea tectonic structures visualized by ERS1 satellite gravity data


References


Knies, J., and C. Gaina, 2008, Middle Miocene ice sheet expansion in the Arctic: Views from the Barents Sea: Geochemistry, Geophysics, Geosystems, v. 9, p. n/a-n/a.
References


References


Nikishin, V. A., 2013, Intraplate and marginal deformation of the Kara Sea sedimentary basins, Moscow State University, 21 p.
References

Olaussen, S., G. B. Larsen, H. Helland, Johannessen, A. Nøttvedt, Riis, Rismyhr, M. Smelror, and D. Worsley, 2019, Mesozoic strata of Kong Karls Land, Svalbard, Norway; a link to the northern Barents Sea basins and platforms.


Rowley, D. B., and A. L. Lottes, 1988, Plate-kinematic reconstructions of the North Atlantic and Arctic: Late Jurassic to Present: Tectonophysics, v. 155, p. 73-120.


References


Chapter 2
Paper 1

**Early Cretaceous tectonostratigraphic evolution of the north central Barents Sea**

B. Kairanov, A. Escalona, A. Mordasova, K. Śliwińska, A. Suslova.

*Journal of Geodynamics, 119, 2018, 183-198, ISSN 0264-3707,*

[https://doi.org/10.1016/j.jog.2018.02.009](https://doi.org/10.1016/j.jog.2018.02.009)
Early Cretaceous tectonostratigraphic evolution of the north central Barents Sea

B. Kuzmicheva, A. Escalda, A. Medvedeva, K. Siwińska, A. Sudsø

A R T I C L E   A B S T R A C T

In this paper we investigate the structural evolution of the North Central Barents Sea during the Early Cretaceous, and the influence of tectonic activity on the stratigraphic pattern in the area. This is achieved by integrating 3D seismic data, two-equation forward and inverse modeling of waveform in lithospheric units, and geological constraints. The extension of the North and South Central Barents Sea and the formation of the North Sea Basin are discussed. The Early Cretaceous tectonostratigraphic framework of the North-Central Barents Sea is presented. The framework is supported by the inversion of the seismic data. The results are compared with previous studies.

1. Introduction

The northern Barents Sea covers the offshore area between the Lofoten Islands and the northern part of Novaya Zemlya (Fig. 1A). The area is one of the most active geodynamic setting in the region. The tectonostratigraphic framework of the North-Central Barents Sea is presented. The framework is supported by the inversion of the seismic data. The results are compared with previous studies.

Previous work in the region has documented a complex event that involved tectonic and magmatic processes during the Late Jurassic to Early Cretaceous (Goguen et al., 1999). The tectonic evolution is characterized by the formation of the North and South Central Barents Sea and the formation of the North Sea Basin (Fig. 1A). The Early Cretaceous tectonostratigraphic framework of the North-Central Barents Sea is presented. The framework is supported by the inversion of the seismic data. The results are compared with previous studies.

1. Introduction

The northern Barents Sea covers the offshore area between the Lofoten Islands and the northern part of Novaya Zemlya (Fig. 1A). The area is one of the most active geodynamic setting in the region. The tectonostratigraphic framework of the North-Central Barents Sea is presented. The framework is supported by the inversion of the seismic data. The results are compared with previous studies.

Previous work in the region has documented a complex event that involved tectonic and magmatic processes during the Late Jurassic to Early Cretaceous (Goguen et al., 1999). The tectonic evolution is characterized by the formation of the North and South Central Barents Sea and the formation of the North Sea Basin (Fig. 1A). The Early Cretaceous tectonostratigraphic framework of the North-Central Barents Sea is presented. The framework is supported by the inversion of the seismic data. The results are compared with previous studies.

1. Introduction

The northern Barents Sea covers the offshore area between the Lofoten Islands and the northern part of Novaya Zemlya (Fig. 1A). The area is one of the most active geodynamic setting in the region. The tectonostratigraphic framework of the North-Central Barents Sea is presented. The framework is supported by the inversion of the seismic data. The results are compared with previous studies.

Previous work in the region has documented a complex event that involved tectonic and magmatic processes during the Late Jurassic to Early Cretaceous (Goguen et al., 1999). The tectonic evolution is characterized by the formation of the North and South Central Barents Sea and the formation of the North Sea Basin (Fig. 1A). The Early Cretaceous tectonostratigraphic framework of the North-Central Barents Sea is presented. The framework is supported by the inversion of the seismic data. The results are compared with previous studies.

1. Introduction

The northern Barents Sea covers the offshore area between the Lofoten Islands and the northern part of Novaya Zemlya (Fig. 1A). The area is one of the most active geodynamic setting in the region. The tectonostratigraphic framework of the North-Central Barents Sea is presented. The framework is supported by the inversion of the seismic data. The results are compared with previous studies.

Previous work in the region has documented a complex event that involved tectonic and magmatic processes during the Late Jurassic to Early Cretaceous (Goguen et al., 1999). The tectonic evolution is characterized by the formation of the North and South Central Barents Sea and the formation of the North Sea Basin (Fig. 1A). The Early Cretaceous tectonostratigraphic framework of the North-Central Barents Sea is presented. The framework is supported by the inversion of the seismic data. The results are compared with previous studies.

1. Introduction

The northern Barents Sea covers the offshore area between the Lofoten Islands and the northern part of Novaya Zemlya (Fig. 1A). The area is one of the most active geodynamic setting in the region. The tectonostratigraphic framework of the North-Central Barents Sea is presented. The framework is supported by the inversion of the seismic data. The results are compared with previous studies.

Previous work in the region has documented a complex event that involved tectonic and magmatic processes during the Late Jurassic to Early Cretaceous (Goguen et al., 1999). The tectonic evolution is characterized by the formation of the North and South Central Barents Sea and the formation of the North Sea Basin (Fig. 1A). The Early Cretaceous tectonostratigraphic framework of the North-Central Barents Sea is presented. The framework is supported by the inversion of the seismic data. The results are compared with previous studies.

1. Introduction

The northern Barents Sea covers the offshore area between the Lofoten Islands and the northern part of Novaya Zemlya (Fig. 1A). The area is one of the most active geodynamic setting in the region. The tectonostratigraphic framework of the North-Central Barents Sea is presented. The framework is supported by the inversion of the seismic data. The results are compared with previous studies.

Previous work in the region has documented a complex event that involved tectonic and magmatic processes during the Late Jurassic to Early Cretaceous (Goguen et al., 1999). The tectonic evolution is characterized by the formation of the North and South Central Barents Sea and the formation of the North Sea Basin (Fig. 1A). The Early Cretaceous tectonostratigraphic framework of the North-Central Barents Sea is presented. The framework is supported by the inversion of the seismic data. The results are compared with previous studies.

1. Introduction

The northern Barents Sea covers the offshore area between the Lofoten Islands and the northern part of Novaya Zemlya (Fig. 1A). The area is one of the most active geodynamic setting in the region. The tectonostratigraphic framework of the North-Central Barents Sea is presented. The framework is supported by the inversion of the seismic data. The results are compared with previous studies.

Previous work in the region has documented a complex event that involved tectonic and magmatic processes during the Late Jurassic to Early Cretaceous (Goguen et al., 1999). The tectonic evolution is characterized by the formation of the North and South Central Barents Sea and the formation of the North Sea Basin (Fig. 1A). The Early Cretaceous tectonostratigraphic framework of the North-Central Barents Sea is presented. The framework is supported by the inversion of the seismic data. The results are compared with previous studies.

1. Introduction

The northern Barents Sea covers the offshore area between the Lofoten Islands and the northern part of Novaya Zemlya (Fig. 1A). The area is one of the most active geodynamic setting in the region. The tectonostratigraphic framework of the North-Central Barents Sea is presented. The framework is supported by the inversion of the seismic data. The results are compared with previous studies.

Previous work in the region has documented a complex event that involved tectonic and magmatic processes during the Late Jurassic to Early Cretaceous (Goguen et al., 1999). The tectonic evolution is characterized by the formation of the North and South Central Barents Sea and the formation of the North Sea Basin (Fig. 1A). The Early Cretaceous tectonostratigraphic framework of the North-Central Barents Sea is presented. The framework is supported by the inversion of the seismic data. The results are compared with previous studies.

1. Introduction

The northern Barents Sea covers the offshore area between the Lofoten Islands and the northern part of Novaya Zemlya (Fig. 1A). The area is one of the most active geodynamic setting in the region. The tectonostratigraphic framework of the North-Central Barents Sea is presented. The framework is supported by the inversion of the seismic data. The results are compared with previous studies.

Previous work in the region has documented a complex event that involved tectonic and magmatic processes during the Late Jurassic to Early Cretaceous (Goguen et al., 1999). The tectonic evolution is characterized by the formation of the North and South Central Barents Sea and the formation of the North Sea Basin (Fig. 1A). The Early Cretaceous tectonostratigraphic framework of the North-Central Barents Sea is presented. The framework is supported by the inversion of the seismic data. The results are compared with previous studies.

1. Introduction

The northern Barents Sea covers the offshore area between the Lofoten Islands and the northern part of Novaya Zemlya (Fig. 1A). The area is one of the most active geodynamic setting in the region. The tectonostratigraphic framework of the North-Central Barents Sea is presented. The framework is supported by the inversion of the seismic data. The results are compared with previous studies.

Previous work in the region has documented a complex event that involved tectonic and magmatic processes during the Late Jurassic to Early Cretaceous (Goguen et al., 1999). The tectonic evolution is characterized by the formation of the North and South Central Barents Sea and the formation of the North Sea Basin (Fig. 1A). The Early Cretaceous tectonostratigraphic framework of the North-Central Barents Sea is presented. The framework is supported by the inversion of the seismic data. The results are compared with previous studies.
technological development of the northeastern Barents Sea during the Early Cretaceous; 3) faulting (i.e. mechanisms controlling the Early Cretaceous inversion; and 3) provide a revised paleogeography re-construction documenting interaction between tectonic and sub-sea marine processes in the northeastern Barents Sea during Early Cretaceous.

2. Geological background

The study area is complex by a number of platforms, basins and structural highs, which are caused between uplift and faulting at the basement level (Fig. 1A). The subsidence history of the area is characterized by a series of compressional and extensional events (Gosline et al., 2009). Starting from the Paleozoic, the northeastern Barents Sea affected by a series of tectonic events: the lithology-lithostratigraphic Cretaceous sequence and the Late Cretaceous-Volcanic Unit engineer (Andt et al., 2011; Ulmann and Faupl, 2003), the Caledonian sequence caused due to the Iapetus Ocean, and the structure of the eastern Barents Sea (Oliver et al., 2009). The latter sequence is responsible for the filling and closure of the eastern basin (Andt et al., 2011). Most of the Barents Sea experienced a coastal extension during the Cenozoic (Fossum and Steinnes, 1995; Granath et al., 1999). The sediments of this age generally comprise of clastic, interbedded with volcanic rocks in front, shallow marine and carbonate (crop) depositional environments (Hjelstuen et al., 2011; Fossum and Steinnes, 1995). The latest Cenozoic is represented by terrigenous carbonates in the eastern basin (Andt et al., 2011). The Late Cretaceous-Cenozoic structural evolution of the Barents Sea is marked by tectonic and tectonic events (Fig. 1B).

1) Rifting in the SW Barents Sea (100-150 Ma) that formed well-defined basins, e.g. Bjornoya, Timna and Illinitskaya Basins (Andt et al., 1995).

2) Reactivation of Cretaceous structural lineaments as de-tectonic strike-slip faults in the East, along Nansen Zonality (150-120 Ma), caused by clockwise rotation of the Barents Shelves (Fig. 1A). This event was responsible for inversion and subsequently uplift of several highs in the northeastern Barents Sea (Andt et al., 2011; Ulmann and Faupl, 2003; Lofgren et al., 2011; Ulmann and Faupl, 2003).

3) The opening of the Arctic Ocean in the North (140-130 Ma) caused by a large-scale event uplift in the northeastern Barents Sea (Oliver et al., 2009). Opening of the Arctic Ocean may have been caused by tectonic events, which affect the eastern and western basins. A large number of models have been proposed, and are summarized into four main categories by Andt et al. (2011): 1) the counterclockwise tectonic model; the clockwise tectonic model; the southeast-western opening model; and the southeast-northern opening model.

4) The movement is consistent with model A. 3.4 Dukhovskaya et al. (2011). However, the Late Cretaceous rocks in the eastern basin are characterized by thin-skinned tectonic events along the eastern margin of the Lena-Arctic Ridge (Gosline et al., 2009), which may have been caused by tectonic events along the eastern margin of the Lena-Arctic Ridge (Gosline et al., 2009).

5) The formation of the Late Cretaceous-Late Cenozoic Province (128-122 Ma) resulted in dynamic subduction in the northeastern and northern parts of the Barents Sea with formation of NNW-SSE trending duques in the Finn Kjøt Land and the North Barents Basin (Andt et al., 2011; Ulmann and Faupl, 2003; Andt et al., 2011; Ulmann and Faupl, 2003).

These Late Jurassic to Early Cretaceous tectonic events were responsible for different degrees of uplift and erosion in the northeastern Barents Sea, as documented on the West and Iver Josef Land (Granath, 1998; Ulmann and Faupl, 2003; Lofgren et al., 2011; Ulmann and Faupl, 2003; Granath et al., 2011; Moshchunov and Sysoev, 2003; Popov et al., 2012).
The split of the northern Barren Sea has affected the Lower Cretaceous sedimentary history of the Greater Barren Sea, where overall NW-SE and NNE-SSW propagation of active margins can be observed (Fig. 3, right). This was evidenced by sedimentary features (Kleeve et al., 2017; Kleeve and Kjær, 2017; Kjær et al., 2017; Kjær et al., 2013).

During the Cretaceous, episodes of tectonic and magmatic deformation occurred between the NW Greenland and the western Barren Sea, and were responsible for the formation of the Yedoma Hill and Wrangel Ridge. These features are lithologically distinct from the northern part of the study area, where shallow water shelf, volcanic outcrops, and tectonic structures are less pronounced. These features are shown in the stratigraphic column of the western Barren Sea (Fig. 3, right).

The Cretaceous development of the Barren Sea is marked by an array of diachronous and isochronous events. The stratigraphic column of the western Barren Sea (Fig. 3, right) shows a significant change in the sedimentary architecture.

3.3. Data and methodology

3.3.1. Sediment and well data

The regional 3D seismic data from the Norwegian Sea and the Barren Sea are provided by the Norwegian Petroleum Directorate (NPD) and the Marine Arctic Geological Expedition (MAGE). The 3D seismic data have a resolution of two-way travel time (FWT) with 10-20 m of resolution intervals. The seismic data cover an area of 350,000 km², with an average distance between seismic lines of 100 km (Fig. 3, left). The quality of the seismic data is good, except in the northern part of the study area, where shallow water depth, volcanic outcrops, and tectonic structures make seismic imaging very poor. Most of the 3D seismic data is publicly released and limited for publications.

The well Lutskovskaya-1 and Lutskovskaya-2, located in the southwestern part of the study area, were used to constrain seismic interpretation. The seismic data is complemented by information from borehole samples and core from the Barren Sea (Fig. 3). Both wells have a full set of logs, but only the well Lutskovskaya-1 has five core samples from the Cretaceous - Albian interval (Fig. 3). Interpretation in the Norwegian Barren Sea is constrained by information from core rock samples and core from the Yedoma and Wrangel Ridge (Fig. 3).
well are shown on Fig. 4.

The stratigraphic framework used in this work is based on the definition of well-specific sequences for wells 7\{99-10\}1 and 9\{98\}3 in the southern basin (Fig. 3). These well-specific sequences (SS) are defined as an assemblage of two or more chips that are separated by unconformities. The assemblages are further subdivided into order-stratigraphic units (OSUs)

4. Results

4.1. Well, stratigraphic, and seismic correlation

Fig. 5 shows the well and seismic correlation of stratigraphic sequences from the northern to the northcentral and northeastern basins. Two sequences 0 (Hennesen - Valdres) and 1 (Hennesen - Valdres) were observed on both the 3D and 2D seismic data.

Sequences 0 and 1 were correlated using the well data and seismic data. Sequence 0 is defined as the sequence with the deepest top and sequence 1 is defined as the sequence with the shallowest top. The correlation between the two sequences is shown in Fig. 5.
Paper 1

In the Olga Basin, it is suggested that the Late Cretaceous
unconformity (BCU) correlates to the seismic reflector "f" versus
Antonino et al. (1993). The BCU is represented by a well-defined
acoustic impedance contrast in seismic data (Figs. 2 and 3).

Fig. 4. The most important events observed in the oil fields of the
1. Rainfall reduces the reservoirs' effective porosity and permeability.
2. Fractures are developed at the top of the BCU and at the base of the RRU.
3. The reservoir is divided into two main units: the upper part, which is a carbonate reservoir, and the lower part, which is a clastic reservoir.
4. The reservoir is a complex system with multiple layers and fractures.
5. The reservoir is a good reservoir for oil and gas production.
6. The reservoir is a good reservoir for oil and gas production.
7. The reservoir is a good reservoir for oil and gas production.
8. The reservoir is a good reservoir for oil and gas production.
9. The reservoir is a good reservoir for oil and gas production.
10. The reservoir is a good reservoir for oil and gas production.
11. The reservoir is a good reservoir for oil and gas production.
12. The reservoir is a good reservoir for oil and gas production.
13. The reservoir is a good reservoir for oil and gas production.
14. The reservoir is a good reservoir for oil and gas production.
15. The reservoir is a good reservoir for oil and gas production.
16. The reservoir is a good reservoir for oil and gas production.
17. The reservoir is a good reservoir for oil and gas production.
18. The reservoir is a good reservoir for oil and gas production.
19. The reservoir is a good reservoir for oil and gas production.
20. The reservoir is a good reservoir for oil and gas production.
21. The reservoir is a good reservoir for oil and gas production.
22. The reservoir is a good reservoir for oil and gas production.
23. The reservoir is a good reservoir for oil and gas production.
24. The reservoir is a good reservoir for oil and gas production.
25. The reservoir is a good reservoir for oil and gas production.
26. The reservoir is a good reservoir for oil and gas production.
27. The reservoir is a good reservoir for oil and gas production.
28. The reservoir is a good reservoir for oil and gas production.
29. The reservoir is a good reservoir for oil and gas production.
30. The reservoir is a good reservoir for oil and gas production.
31. The reservoir is a good reservoir for oil and gas production.
32. The reservoir is a good reservoir for oil and gas production.
33. The reservoir is a good reservoir for oil and gas production.
34. The reservoir is a good reservoir for oil and gas production.
35. The reservoir is a good reservoir for oil and gas production.
36. The reservoir is a good reservoir for oil and gas production.
37. The reservoir is a good reservoir for oil and gas production.
38. The reservoir is a good reservoir for oil and gas production.
39. The reservoir is a good reservoir for oil and gas production.
40. The reservoir is a good reservoir for oil and gas production.
41. The reservoir is a good reservoir for oil and gas production.
42. The reservoir is a good reservoir for oil and gas production.
43. The reservoir is a good reservoir for oil and gas production.
44. The reservoir is a good reservoir for oil and gas production.
45. The reservoir is a good reservoir for oil and gas production.
46. The reservoir is a good reservoir for oil and gas production.
47. The reservoir is a good reservoir for oil and gas production.
48. The reservoir is a good reservoir for oil and gas production.
49. The reservoir is a good reservoir for oil and gas production.
50. The reservoir is a good reservoir for oil and gas production.
51. The reservoir is a good reservoir for oil and gas production.
52. The reservoir is a good reservoir for oil and gas production.
53. The reservoir is a good reservoir for oil and gas production.
54. The reservoir is a good reservoir for oil and gas production.
55. The reservoir is a good reservoir for oil and gas production.
56. The reservoir is a good reservoir for oil and gas production.
57. The reservoir is a good reservoir for oil and gas production.
58. The reservoir is a good reservoir for oil and gas production.
59. The reservoir is a good reservoir for oil and gas production.
60. The reservoir is a good reservoir for oil and gas production.
61. The reservoir is a good reservoir for oil and gas production.
62. The reservoir is a good reservoir for oil and gas production.
63. The reservoir is a good reservoir for oil and gas production.
64. The reservoir is a good reservoir for oil and gas production.
65. The reservoir is a good reservoir for oil and gas production.
66. The reservoir is a good reservoir for oil and gas production.
67. The reservoir is a good reservoir for oil and gas production.
68. The reservoir is a good reservoir for oil and gas production.
69. The reservoir is a good reservoir for oil and gas production.
70. The reservoir is a good reservoir for oil and gas production.
71. The reservoir is a good reservoir for oil and gas production.
72. The reservoir is a good reservoir for oil and gas production.
73. The reservoir is a good reservoir for oil and gas production.
74. The reservoir is a good reservoir for oil and gas production.
75. The reservoir is a good reservoir for oil and gas production.
76. The reservoir is a good reservoir for oil and gas production.
77. The reservoir is a good reservoir for oil and gas production.
78. The reservoir is a good reservoir for oil and gas production.
79. The reservoir is a good reservoir for oil and gas production.
80. The reservoir is a good reservoir for oil and gas production.
81. The reservoir is a good reservoir for oil and gas production.
82. The reservoir is a good reservoir for oil and gas production.
83. The reservoir is a good reservoir for oil and gas production.
84. The reservoir is a good reservoir for oil and gas production.
85. The reservoir is a good reservoir for oil and gas production.
86. The reservoir is a good reservoir for oil and gas production.
87. The reservoir is a good reservoir for oil and gas production.
88. The reservoir is a good reservoir for oil and gas production.
89. The reservoir is a good reservoir for oil and gas production.
90. The reservoir is a good reservoir for oil and gas production.
91. The reservoir is a good reservoir for oil and gas production.
92. The reservoir is a good reservoir for oil and gas production.
93. The reservoir is a good reservoir for oil and gas production.
94. The reservoir is a good reservoir for oil and gas production.
95. The reservoir is a good reservoir for oil and gas production.
96. The reservoir is a good reservoir for oil and gas production.
97. The reservoir is a good reservoir for oil and gas production.
98. The reservoir is a good reservoir for oil and gas production.
99. The reservoir is a good reservoir for oil and gas production.
100. The reservoir is a good reservoir for oil and gas production.
### Table 1

<table>
<thead>
<tr>
<th>Sequence</th>
<th>Characteristics</th>
<th>Interpretation/Environment</th>
<th>Example</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Low amplitude, continuous reflectors observed in the Oligocene-IOC-Recent.</td>
<td>High-energy, short-lived deposits</td>
<td>None</td>
</tr>
<tr>
<td>2</td>
<td>High amplitude, high-velocity, continuous reflectors.</td>
<td>High-energy, short-lived deposits</td>
<td>None</td>
</tr>
<tr>
<td>3</td>
<td>Low amplitude, high-velocity, continuous reflectors.</td>
<td>High-energy, short-lived deposits</td>
<td>None</td>
</tr>
<tr>
<td>4</td>
<td>High amplitude, high-velocity, continuous reflectors.</td>
<td>High-energy, short-lived deposits</td>
<td>None</td>
</tr>
<tr>
<td>5</td>
<td>High amplitude, continuous reflectors.</td>
<td>High-energy, short-lived deposits</td>
<td>None</td>
</tr>
<tr>
<td>6</td>
<td>High amplitude, continuous reflectors.</td>
<td>High-energy, short-lived deposits</td>
<td>None</td>
</tr>
</tbody>
</table>

### 6.2.2. Sequence 2 (S2): Early Apatian - Early Aptian

- **Description:** Similar to S1, S2 was deposited in the zone of widespread tectonic events, reaching towards the Oligocene-IOC-Recent. It is characterized by high-energy, short-lived deposits, indicating high-velocity seafloor spreading.

- **Interpretation:** The high-energy, short-lived deposits observed in the Oligocene-IOC-Recent are indicative of high-velocity seafloor spreading. These deposits are characterized by high amplitude, continuous reflectors, suggesting high-energy, short-lived deposits.

### 6.2.2.2. Interpretation: Generally, S2 is deposited in a high-energy, short-lived environment with high-velocity seafloor spreading.
Paper 1

4.2.3. Sequence 3-4 (S3-S4): Early Aptian - Coniacian

4.2.3.1. Deposition. S3-S4 is bounded at its base by the Early Aptian Unconformity and at the top by the Coniacian Unconformity. This
Fig. 5. (A-C) Interpreted and (D-F) interpreted seismic sections across the Kang Valley Bird Fault. Note the base of sequence 1's boundary (S1). Apparent depth to the basement of the S1. (G) Interpreted seismic section of the Kang Valley Bird Fault. The boundary of sequence 1 indicates the depth of the basement.

Fig. 6. (A-C) Interpreted and (D-F) interpreted seismic sections across the Kang Valley Bird Fault. Note the base of sequence 1's boundary (S1). The boundary of sequence 1 indicates the depth of the basement.
sequences are only observed in the North African basins and projected by the well Kalambo 1 and the Kalambo 2. Laminar 1 and 2 are well correlated and tentatively dated to Early Aplin Zone 1 of (Kolodny 1999) based on the presence of foraminifera and the absence of biostratigraphic ranges. The sample yields foraminiferal assemblage but suggests reworking from Triassic. Sample 1 and 2 yields abundant and well preserved assemblage, limited however to the dinoflagellates with long foraminiferal ranges. The assemblage is dominated by Rostovskina and Pollenella with planktonic range and characterized by low diversity. This may suggest initial, restricted marine setting. The tentative age for the two samples is near Early Aptian or younger. The GR for these sequences in spiny with relatively high values and mainly higher with the exception of (Fig. 3). The tectonic character of the sequence in the north African basins is represented by high amphibolite, parallel conjugate reflections with some chaotic or lens-shaped reflections in the distal parts of the basin (Fig. 4A and 4B).

4.2.3.2 Interpretation Overall, A-5-6 is deposited in a tectonic quiescent period, as suggested by the absence of growth strata and abundance of parallel and horizontal reflections. The cool (green) bands observed in the barrier part of (5A) of the well (A) suggest that the sequence was deposited in a coastal to continental environment. The overall tectonic setting of (5A) is based on the GR (Fig. 4B) for the well
The main fault families related to the Lower Cretaceous sequence are observed in the study area (Figs. 4 and 5). Fault families 1 and 2 are interpreted as thick-skinned detachment involving reverse faults that tip in the Tellian interval, however due to continuous reactivation some faults propagate into the basement (Figs. 3 and 4B). Fault family 1 is interpreted as thick-skinned detachment affecting areas above the Tumar line (Figs. 4A and 4C). Fault family 1 (2011) is set of 34-49° striking faults, in parallel from the PP to the edge of the Kamora Land Platform to the Kuma River Land (Fig. 2C). However, faults from the Horn Land Platform (1999) have identical strike to PP and most likely represent similar components. A number of PP are in a main boundary fault that separates the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform. PP and PP are close to the Kamora Land Platform from the Horn Land Platform.
which might indicate a temporally adjacent period during Paleogene.

5.1.2. Yanshanian Terrane - Early Cretaceous

Further reactivation of 3H and F2 is recorded in the Lower Cretaceous during transpression in early inversion, and corresponds to 3J. Reactivation is based on observed growth strata and modeling of the 3J (Fig. 8-9). Reactivation is consistent with earlier studies on the Kong Karo Land (Zhalama et al. 2019) that defines the same period of reactivation. However, along Noraya Zentyn, reactivation was more likely related to the Dzarkan - Early Jurassic strain, and is supported by the absence of post-Cretaceous sediments in the well Zentyn-1. Fig. 7 and by the interpretation of the Noraya - Early Jurassic extension complex that first formed during Early Jurassic and then during Early Cretaceous (Zhalama et al. 2019).

5.1.3. Mid-Cretaceous

Reactivation of 3H and 3J that led to further inversion was also recognized in the study area (Figs. 7 and 9). The age of inversion is uncertain. However, folding of the Lower Cretaceous sequence 3 and 4 in the Zentyn-Novaya Zentyn area suggests that reactivation inversion took place at least post-Cretaceous time (Figs. 7 and 9). This inversion is responsible for the separation and exhumation of previously basement rocks and is most likely related to the present day basin configuration.

5.2. Mechanisms controlling the Late Jurassic - Early Cretaceous inversion

High angle reverse faults (Fig. 5) and F2 and F3, the northward Benjamino Sea were developed through the reactivation of Late Palaeozoic normal faults (Figs. 12). Growth strata of 3H and 4J in the Kong Karo Land (Figs. 1-1) indicated that the reactivation was caused by a regional reverse process outside of the study area during the Late Jurassic - Early Cretaceous, then suggesting major regional extensional events that might have influenced the study area:

1. Detachment transposition along Noraya Zentyn

Previous work by Nikolayeva et al. (2014), Mulyate et al. (2012) and Nikolayeva (2013) documented an inversion of structural highs like the Adygran, Preemekanovskoe, Batiluvskoe, Berdskoye and Pomorinskoe (Fig. 1). The inversion of those structural highs was suggested to be controlled by pulses of Late Jurassic - Early Cretaceous syntectonic transgression along Noraya Zentyn (Zhalama et al. 2019). Fig. 6A suggests that transgression was related to the evolution of the Siberian Platform due to the collision of the Okhotsk microcontinent in the Bering Strait - Chukotka region that took place in the Tithonian - Berriasian (Fig. 6B, 6C). The timing of 3J towards the Adygran high and the Lambothropus Delta suggests the inversion of these highs at least during Early Cretaceous (Fig. 6A and B).

It also possible that the structural transgression along Noraya Zentyn could have caused an inversion on the Kong Karo Land (Fig. 1), and the Ilgoy faults (Fig. 1A). Structural models by Smerda et al. (2007) and the tiltable seismic displacement of Noraya Zentyn suggests that the propagation of the deformation is up to 50 km seaward during the Late Triassic - Early Jurassic. In particular, in the Late Triassic - Early Jurassic inorganic transpositions, the lowest transpositions during Late Jurassic - Early Cretaceous is consistent with a tectonically active region (Zhalama et al. 2019) in the western margin of the Yamal Paleo-Tethys (Zhalama et al. 2019). This suggests that the northern margin of the Yamal Paleo-Tethys is proposed as a possible tectonic syncline (Fig. 6A).
deformation related to the Alpina Taph and the Qilin-Huazhuang strike-slip fault propagating to a maximum 40 km in the north of these faults. On Fig. 14, the Qingshanyi fault is shown on map Fig. 15B to illustrate potential directions of the structural settings of these faults.

Therefore, considering a distance extending ~60 km between Qingshanyi and the Junggar Basin Platform (Fig. 15B), it is unlikely that the reaction to the setting structure of the Junggar Basin controlled the orientations of the NE-SW and NW-SE oriented structures on the Junggar Basin Platform and the Qingshanyi fault.

2. Opening of the Antarcis Basin

The stress ellipsoid inferred for the faults from FTT on the Junggar Basin Platform suggests that the dominant strike-slip motion might have initiated this extension (Fig. 15B). Predominantly north-south trending dykes are present south of the North Bosten Basin (Figs. 2 and 15A). Dyke emplacement is commonly perpendicular to the extension direction or to its maximum horizontal stress axis (Anderson, 1955; Behrens and Willis, 1973). The orientation of these dykes resembles the direction of the Moho's slope on the stress ellipsoid (Fig. 15B). These dykes were reactivated in this direction mainly due to the structural weaknesses developed during the tectonic extension along the northern margin of the Tarim basin (Fig. 15A). Mooney et al. (2010) characterized the northwestern margin of...
the Lacrmores Ridge on a transform margin, suggesting formation of back-arc basins. Bioturbation studies from the lacrmores Basin (Kulm and Linn, 1999; Linn et al., 2000, 2002; Geinitz et al., 2003, 2004) further support this hypothesis. The presence of back-arc basins is consistent with seafloor spreading that was active at this time.

3.2.3. Sequence 2 (Lacrmores–early Persianian)

The lacrmores Basin is a back-arc basin that developed during the early Persianian period (Linn et al., 2000, 2002). It is characterized by a series of grabens and half-grabens, which are associated with transform faults and normal faults. The grabens are filled with sediments that are primarily coarse-grained and include volcaniclastic sediments. The basin is bounded by a series of strike-slip faults that have a SW-NE trend.

3.2.4. Sequence 3 (middle to late Persianian)

The middle to late Persianian period is characterized by the development of a series of basins along the transform margin. The basins are filled with sediments that are primarily fine-grained and include volcaniclastic sediments. The basins are bounded by a series of normal faults that have a NW-SE trend. The basin is characterized by a series of half-grabens and grabens, which are associated with transform faults.
exposed, particularly in the area around the Adakaiti high. This interpretation is based on the overall toning and erosion of S2 towards Norrey Vanga (Fig. 5A) and by the absence of pre-Rannian sediments in the east (Zaleka - 1 Fig. 5). 3.3.3. Sequence 2 (Early Jurassic – Early Aptian)

There are no systems of provenance suggesting the formation of the high Arctic Igneous Province in the northwestern edges of the Barents Sea between 120 Ma and 130 Ma (Kvask et al., 2016; Nilsson et al., 2016). Ch. 3) implying that the study area was subjected to several uplifts due to magmatic activity. This interval is equated to the interval of deposition of S2. However, observations from 3D seismic data suggest that the deposition of S2 occurred during a tectonically quiescent period (44-30 Ma), as advocated by the presence of large cratons and the absence of large-scale strike-slip fault systems (Ch. 4 and 5). The depositional environment of the Kongs Kold Land Platform and the Olga Basin is mainly a shallow shelf to coastal (Ch. 3 and 5), while in the North Samoiskoe continental environment preserved (Ch. 4 and 6) (Ch. 10). The main source of sediments is interpreted to be located in the N-NNW in Pray Nord Land and the North Kara region. The shifting of the main depocenters to S-SW, towards the Olga Basin and the Framnordland Platform (Ch. 10) suggests a regional progressive depocenter from the NNE to SW (Ch. 10). It is also supported by the lower Cretaceous sandstone megasequence observed in the southern Barrow Sea that indicates a strike NW-SE and NNE-SSW divergence of the shelf depocenters in Barrow Sea (2016; Mani et al. 2017). Overall, the N-S-orientation of the depocenter area during S2 is suggested to be controlled by subaerially exposed structural highs inherited from S1 (Ch. 11; Rasmussen et al., 2018). These structural highs acted as landing sites and are responsible for storing the palaeotopography signal towards the southern basin of the region (e.g., the South Barrow Basin and the Framnordland Platform; Fig. 10). During the deposition of S2, the North Barrow Basin was the least influenced by tectonic events and depositional processes that were periodically nucleated. The third part of the sequence was not significantly influenced, comprising the southern Barrow Sea (e.g., the South Barrow Basin, the Framnordland Platform, and the Nyordland domes; Snajklev et al., 2015; Marti et al., 2017). The presence of these sequences on the Kongs Kold Land Platform and the Olga Basin is uncertain. It is noted that Sequence 3 was deposited on the Kongs Kold Land Platform and Olga Basin, as suggested by the presence of the time-equivalent depoduction of the Cenomanian Formation in southeast Fig. 10). However, it is still unknown if sequences 4-6 were deposited on the Kong, Kold Land Platform and the Olga Basin, as their sediments have not been observed in this area (Fig. 5A).

5. Conclusions

Three stages of evolution are defined in the northern Barents Sea: (1) post-Early Cretaceous – early Jurassic (Cretaceous); (2) Jurassic – Early Cretaceous; and (3) post-Early Cretaceous – Late Cretaceous. The post-Early Cretaceous – early Jurassic (Cretaceous) evolution is characterized by a significant period of extensional tectonics in the northern Barents Sea. During the deposition of S2 (Early Jurassic – Early Aptian), the combined palaeogeography allowed sediments to precipitate from NNE to SW, whereas uplifted and subaerially exposed structural highs acted as landing sites for controlling the regional palaeotopography system. Later, in post-Early Cretaceous evolution, the main depocenter was responsible for reactivation of the Late Jurassic – early Cretaceous structural trends and structures.

Three major regional trends are suggested as main mechanisms for controlling the post-Early Cretaceous evolution: (1) early Cretaceous (Cretaceous); (2) Jurassic – Early Cretaceous (Jurassic); and (3) post-Early Cretaceous evolution (Cretaceous).
For palynographic maps from the study (SH and S2) are part of the paper. Palynological data and fossil records for the Late Eocene (Goldstein et al., 1984) and Early Miocene (Goldstein and Weisburg, 1986) are also included. In addition, we are thankful to University of Kansas for providing the software and license. We would also like to thank our fellow colleagues for their valuable comments that improved the manuscript and to Anneke Kuyk for preparing palynological slides.

Acknowledgements

This study is part of the industrial sponsored LeCoCon consortium. We would like to express our gratitude to Norsk Polarinstitutt for providing GIS software and data and allowing to publish them. In addition, we are thankful to Heiko Simons and the crew for providing the software and license. We would also like to thank our fellow colleagues for their valuable comments that improved the manuscript and to Anneke Kuyk for preparing palynological slides.

References


Bourgeois, S., Boudrea...
Paper 2

The Early Cretaceous evolution of the Tromsø Basin, SW Barents Sea, Norway

Bereke Kairanov, Alejandro Escalona, Ian Norton and Peter Abrahamson.

Submitted to Marine and Petroleum Geology
ABSTRACT

Extensional basins developed along oblique or transform margins are least studied basins as compared to those developed along orthogonally extended margins, therefore their evolutionary models are controversial. This study present an example of the basin, namely Tromsø, which developed along the Southwestern Barents Sea transform margin. Three previous models have been proposed to explain the tectonic evolution and architecture of the basin, but still no consensus on how the development of the individual structures is reached. In this study, we use 2D industry seismic reflection data, potential field and wells data, as well as previously published information to understand the Early Cretaceous structural evolution of the Tromsø Basin in the context of the geodynamic processes in the south-western Barents Sea. Modelled gravity anomalies along a composite 2D regional seismic section facilitated the interpretation of crustal structures, which then were used for a 2D structural reconstruction. Unlike any previous models, we propose a new Early Cretaceous structural evolutionary model for the Tromsø Basin, which involves oblique extension and the formation of an intra-basinal transfer zone with transpressional strike slip fault systems. The basement heterogeneity suggested to have played important role in focusing and localizing strain in the area. The 2D sequential restoration of the regional profile yields an estimate of ca. 35 km of crustal extension from the earliest Cretaceous until present. Thinner crust below the Tromsø Basin as compared to Sørvestnaget and Hammerfest basins is suggested to be additional characteristics favoring the oblique rifting of the margin. This study illustrates the importance of integrating regional tectonic settings when reconstructing the evolution of basin-bounding faults.
INTRODUCTION

The structural evolution of orthogonally extended passive margins and basins are well documented through outcrop and subsurface studies (Badley et al., 1988; Bell et al., 2009; Ebinger, 1989; Jackson and Rotevatn, 2013; Lehner and De Ruiter, 1977; Moustafa, 1993; Sharp et al., 2000; Spathopoulos, 1996; Withjack et al., 1998; Ziegler, 1992), as well as numerical and physical analogue modelling (Corti et al., 2003; Huismans et al., 2001; McClay et al., 2002; Naliboff and Buiter, 2015). Due to tectonic inheritance and irregular shapes of the plate boundaries, passive margins and basins include segments where oblique or sheared tectonics is prevailing (Brune et al., 2018; Dewey et al., 1998; Hodge et al., 2018; Manatschal et al., 2015; Morley, 2017; Phillips et al., 2018; Sanderson and Marchini, 1984). Frequently, evolution of such segments are assessed using models assuming an orthogonal alignment of relative plate motion and plate boundary, which may lead to erroneous assessment of the subsidence pattern and faults evolution (Brune, 2014; Brune et al., 2016; Huismans and Beaumont, 2011; Lavier and Manatschal, 2006; McKenzie, 1978; Naliboff et al., 2017; White, 1993). Despite the general knowledge of the extensional basins that involve certain degree of obliquity (Atwater and Stock, 1998; Corti, 2008; Fletcher et al., 2007; Fournier et al., 2004; Klimke and Franke, 2016; Lizarralde et al., 2007; Mart et al., 2005; Phethean et al., 2016), the structuring and kinematics of the past rift basins remains under-researched.

Therefore, in this study, we analysis tectonic evolution of the Tromsø Basin which located along the sheared margin of the SW Barents Sea and has a complex tectonic history that involved both orthogonal and sheared rifting (Faleide et al., 1993). In general, it is accepted that the basin formed in response to Late Jurassic – Early Cretaceous rifting in the SW Barents Sea (Faleide et al., 1993), but there is no clear consensus on basin evolution and tectonic model. Early works proposed at least two models
to explain the Late Jurassic – Early Cretaceous evolution and structuring of the Tromsø Basin: 1) Strike-slip model with either (a) sinistral or (b) dextral strike-slip system along the northern basin bounding faults (e.g. Bjørnøyrenna Fault Complex; (Gabrielsen and Færseth, 1988; Riis et al., 1986), and 2) Large scale extensional model with sinistral strike-slip faults (Fig. 2B) (Faleide et al., 1993). Timing and direction of major fault movements in these models are controversial, where for instance, the structural highs along the western margin (e.g. Senja Ridge and Veslemøy High; Fig. 2A) have either an extensional (Faleide et al., 1993; Indrevær et al., 2013; Riis et al., 1986) or a compressional origin that resulted from strike slip tectonics (e.g. along Ringvassøy – Loppa and Bjørnøyrenna fault complexes; (Gabrielsen and Færseth, 1988). Later work, using potential field (magnetic and gravity) data, proposed a new crustal scale “boundinage” model for the SW Barents Sea (Gernigon et al., 2014). Indirect observations from neighboring basin suggested a highly thinned crust and abundancy of low angle fault systems below the Tromsø Basin, which still requires better constrains in terms of basin evolution (Fig. 2C). Disagreements regarding the tectonic models are also attributed to a limited well control and poor imaging of the deep basin by seismic reflection profiles that increases uncertainty in the interpretation (Breivik et al., 1998; Faleide et al., 2008; Faleide et al., 1993; Gabrielsen, 1984; Mosar et al., 2002).

Although all models explain the present-day configuration of the basin, choosing one or another model can lead to an erroneous assessment of the tectonic and geodynamic settings. Therefore, in this paper we aim to: (1) understand the Early Cretaceous tectonic processes in the Tromsø Basin and revise structural evolutionary models; and (2) by restoring 2D regional a crustal profile constrain the development of the area to understand pre-drift configuration of the margin.
Figure 10 A) Location of the study area with main structural elements of the SW Barents Sea (TB = Tromso Basin; HB = Hammerfest Basin; BB = Bjørnøya Basin; SB = Sørvestnaget Basin; LH = Loppa High; SH = Stappen High; VH = Veslemøy High; SR = Senja Ridge; VVP = Vestbaken Volcanic Province; FP = Finnmark Platform; BKFC = Bothnian Kvænangen Fault Complex; BSFC = Bothnian-Senja Fault Complex; TFFC = Tromsø-Finnmark Fault Complex; TFFC = Tromsø Finnmark Fault Complex; RFLC = Ringvassøy- Loppa Fault Complex; BFC = Bjørnstrenna Fault Complex; BFZ = Billefjorden Fault Zone; HFC = Horsund Fault Complex; SFZ = Senja Fracture Zone). B) Location of the 2D seismic profiles and wells.
GEOLOGICAL SETTING

The Tromsø Basin is characterized as an elongated NNE–SSW-striking basin, with a length of 140 km and a width of 60 km (Fig. 1A; (Gabrielsen et al., 1990). The southern boundary is a transition towards the Harstad Basin and termination against the Troms – Finnmark Fault Complex (TFFC) (Fig. 1A; (Gabrielsen et al., 1990). The northern boundary is defined by the Veslemøy High and the Bjørnøyrenna Fault Complex (BFC), which separate the Tromsø Basin from the Bjørnøya Basin (Fig. 1B). The eastern boundary, towards the Hammerfest Basin, is delineated by the Ringvassøy-Loppa Fault Complex (RLFC), while the western boundary is limited to the Senja Ridge and the Veslemøy High (Fig. 1B (Gabrielsen et al., 1990).

The tectonic history of the Tromsø Basin, can be traced back to approximately 400Ma, when the Caledonian orogeny was formed by collision of the Laurentian and Baltic plates with development long lived reversed fault zones with a variety of nappes and thrust sheets over the Fennoscandia Shield (Gabrielsen, 1984; Gasser, 2013; Gee et al., 2006; Gee et al., 2008; Roberts, 2003). Towards the end of the Paleozoic, regional extension caused collapse of thrust sheets resulting in formation of grabens and half grabens that were covered by Upper Carboniferous to Lower Permian carbonate platforms and thick evaporates deposits (Gudlaugsson et al., 1998; Larssen et al., 2002).

The Early Triassic is marked by a rift episode, which has been documented in the North Atlantic region (Tsikalas et al., 2012). It has been suggested that the same Early Triassic rift episode may have continued until the Middle Triassic around the Tromsø Basin (Smelror et al., 2009). The Triassic succession is comprised of prograding and retrograding cycles of marine, deltaic and continental clastic deposits sourced from the ESE (Glørstad-Clark et al., 2010; Klausen et al., 2015).
From the Middle Jurassic to Early Cretaceous, northward advance of the Atlantic rifting formed deep basins in the southwestern Barents Sea such as the Harstad, Tromsø, Bjørnøya and Sørvestnaget basins (Faleide et al., 2008; Gernigon et al., 2014). In the Tromsø Basin, the Early Cretaceous rift episode along the NE – SW trending Ringvassøy–Loppa and Bjørnøyrenna fault complexes led to rapid subsidence and accumulation of thick Cretaceous sediments (Clark et al., 2014; Faleide et al., 2008; Faleide et al., 1993; Gabrielsen et al., 1990; Indrevær et al., 2016; Rønnevik et al., 1982). The central part of the basin experienced salt related deformation during this rift event (Faleide et al., 1993; Gabrielsen et al., 1990; Larssen et al., 2002; Sund, 1984). Three Early Cretaceous rift phases have been interpreted for the Tromsø Basin: Berriasian–Valanginian, Hau tertivian–Barremian and Aptian–Albian (Faleide et al., 1993). Local compression during the earliest Cretaceous has been identified in the northern part of the basin. This has been suggested to be the result of dextral strike slip movement along the Asterias Fault complex (Berglund et al., 1986; Gabrielsen et al., 1990; Sund, 1984) or as a localized tectonic inversion due to differential uplift of the Loppa High (Indrevær et al., 2016). In terms of the Lower Cretaceous stratigraphy, a major break in deposition occurred from the Boreal Berriasian/Volgian to Valanginian to Barremian, forming a regional unconformity known as the Base Cretaceous Unconformity (BCU) (Århus et al., 1990; Lundin and Dore, 1997; Marin et al., 2017b; Mork et al., 1999). The BCU is expressed as a high amplitude seismic reflector and its age and stratigraphic significance is complex (Gabrielsen et al., 2001; Nottvedt et al., 1995). In the areas of the southwestern Barents Sea, where basin margins are affected by Late Jurassic to Early Cretaceous tectonism, the BCU represents an unconformity, whereas, in the deeper basins, it is a conformable surface (Marin et al., 2017b). The Lower Cretaceous succession of the SW Barents Sea is divided into four main formations: Knurr, Klippfisk, Kolje and Kolmule, which consist mainly of grey claystone with minor interbedded limestone and sandstone deposited in an open marine environment (Dalland et al., 1988; Mork et
al., 1999). More recently, these formations were divided into seven
genetic sequences (sequences 0–6; Fig. 3A; Marin et al., 2017b). These
sequences are bounded by flooding surfaces, some of which can be
correlated on a regional scale (Grundvåg et al., 2017; Marin et al.,
2017b).

The Late Cretaceous – Paleocene period is associated with dextral strike-
slip movement between the western Barents Sea and northern Greenland
(Faleide et al., 1996). This event divided the margin into two shear
margins, the Hornsund in the north and Senja in the south (Faleide et al.,
2008) (Fig. 1A). The Paleocene – Eocene transition (55 – 54 Ma) is
marked by a continental breakup of the North Atlantic margin, followed
by separation of the Barents Sea and the eastern Greenland margin, and
opening of the Fram Strait (Faleide et al., 1996). During the same period
the Barents Sea experienced onset of a tectonic uplift that caused
exhumation and erosion of the northern and western margins of the
Barents Sea (Dimakis et al., 1998; Henriksen et al., 2011; Knies and
Gaina, 2008).

**Relationship of onshore basement lineaments and offshore structural trends**

It has been suggested that the Early Cretaceous evolution and structural
configuration of the SW Barents Sea has been influenced by inherited
Caledonian or even older Precambrian basement structures (Barrère et
al., 2009; Braathen et al., 1999; Doré, 1991; Fichler et al., 1997;
Gabrielsen, 1984; Gernigon et al., 2014; Ritzmann and Faleide, 2007;
Tsikalas et al., 2012). There are three well-constrained, long-lived fault
complexes identified on the northern mainland of Norway that have
affected the structuring of the southern and southwestern Barents Sea:

1) The Trollfjorden-Komagelva Fault Zone (TKFZ), a major
Precambrian WNW-ESE striking fault zone that was episodically
reactivated during the Paleozoic and Mesozoic (Fig. 1A) (Gabrielsen,
1984; Herrevold et al., 2009; Karpuz et al., 1993; Rice et al., 1989; Roberts, 1972; Roberts et al., 2011; Roberts and Lippard, 2005; Siedlecka and Siedlecki, 1967). The TKFZ is characterized as a transfer fault system that has an increasing component of extension toward the adjacent Hammerfest Basin (Berglund et al., 1986; Gabrielsen and Færseth, 1989; Gabrielsen, 1984). This relationship has been well constrained with magnetic data (Gernigon et al., 2014);

2) The Bothnian–Senja Fault Complex (BSFC) and 3) The Bothnian–Kvænangen Fault Complex (BKFC) are two major Precambrian NNW-SSE striking ductile shear zones that were periodically reactivated during Paleozoic and Mesozoic times (Fig. 1A) (Doré et al., 1997; Indrevær and Bergh, 2014; Indrevær et al., 2013; Olesen et al., 1997). The Senja Shear Zone and Fugløya transfer zone have been proposed as offshore northward extension of the BSFC and BKFC, respectively (Fig. 1A) (Faleide et al., 1993; Gabrielsen et al., 1997; Indrevær et al., 2013). Moreover, the Hornsund Fault Complex (HFC) and Billefjorden Fault Zone (BFZ) identified in Svalbard are suggested to be part of the same NNW–SSE structural trend (Fig. 1A) (Doré et al., 1997). Despite the apparent continuity of onshore and offshore structural expressions, connection of these faults is not supported by direct evidence or reliable documentation.
Figure 11 Simplified sketch of previously proposed regional tectonic models for Late Jurassic - Early Cretaceous structuring of the Tromsø Basin. Please note differences in the Senja Ridge and Veslemøy High interpretations as a positive structural feature. A) a) Sinistral and b) dextral strike-slip system along Bjørnøyrenna and Ringvassoy Fault complexes (Riis et al., 1986; Gabrielsen and Færseth, 1988); B) Large scale extensional system with sinistral strike-slip along Bjørnøyrenna Fault complex (Faleide et al., 1993) and C) A propagating system of highly thinned crust dominated by reactivated listric fault system (Gernigon et al., 2014)
DATASET

The database for this study comprises a post-stack time migrated, 2D reflection seismic surveys provided by the Norwegian Petroleum Directorate (NPD) and a MultClient Geophysical AS (MCG) (Fig. 1B). It is important to mention that the 2D seismic data acquired by MCG in 2016 have a better quality in imaging and coverage of the study areas. The seismic profiles covers the area of ca. 2500 km², where the average distance between seismic lines is 2-5 km. The seismic data penetrates depths of 6-9 seconds in two-way-traveltime (TWT) with dominant frequency range between 10-40 Hz. In general, the quality of the seismic data is moderate to good, except the southern part of the study area, where continuity of the seismic reflectors is poor.

The well data used include five exploration wells 7019/1-1, 7119/9-1, 7220/10-1, 7119/7-1 and 7117/9-2 (Fig. 1B). All wells have a full set of logs and biostratigraphic data were obtained from well reports publically available in the NPD web page (http://factpages.npd.no), from the “Lower Cretaceous basins in the high Arctic” consortium project (LoCrA; http://locra.ux.uis.no) and previous publications (e.g. Marin et al., 2017b). Free air gravity map of Sandwell et al. (2014) was used to model anomalies along selected seismic profiles.
Figure 12 A) Stratigraphic framework based on well correlation of interpreted Lower Cretaceous seismic units with defined sequences of Marin et al., (2017) in the Hammerfest Basin (well 7120/10-2); B) Long distance correlation of the interpreted Lower Cretaceous unit. Please note that wells in the Tromsø Basin do not reach deeper seismic units.
METHODOLOGY

Seismic interpretation:

Based on reflection terminations and regional continuity of the reflectors, four horizons (BCU, K1, K5 and K6) that subdivide three seismic units (U1, U2 and U3) were selected and mapped to constrain the Early Cretaceous evolution of the Tromsø Basin.

In order to establish the regional tectonostratigraphic framework, age of interpreted horizons have been constrained by biostratigraphic data and correlated with flooding surfaces of Marin et al. (2017a) in the Hammerfest Basin (e.g. well 7120/10-2; Figs. 3A and 3B). Seismic horizons and their correlative flooding surfaces are: Base Cretaceous Unconformity (BCU), Late Barremian (K1), Late Albian (K5) and Late Cenomanian (K6) (Fig. 3B). It is important to state that the BCU and K1 have never been penetrated by wells in the Tromsø Basin. Interpretation of these horizons was based on seismic reflector configurations and long distance correlations (Fig. 3B). Consequently, seismic units were also correlated with sequences (S0 – S6) of Marin et al. (2017a), as follow: U1 with S0 – S1; U2 with S2 – S5; and U3 with S6 (Fig. 3A). Stacking patterns and sand/shale indicators from the gamma ray (GR) well logs were used to constrain the depositional setting and to support age determination where biostratigraphic evidence is limited (Galloway, 1989) (Fig. 3A). Internal characteristics of each seismic unit such as growth strata, lap relationships were included in the seismic interpretation to outline main periods of faults activity and quiescence.

Faults were interpreted and grouped into fault families based on the similar structural style, orientation and relative age. Time-thickness maps were created to determine variations in structural styles in the basin, including uplifted areas and main depocenters during the main episodes of fault activity.
Figure 13 A) Gravity anomalies model along (B) composite regional 2D seismic sections. Coloured polygons represents areas with constant density, and correspond to the main sequence boundaries described in the seismic interpretation (grey, red and orange coloured polygons; Table 1). Note the differences between modelled (dashed line) and interpreted (red line) basement depth below the Loppa High and Hammerfest Basin. Regional profile and gravity modelling:
In order to understand the regional structural configuration around the Tromsø Basin, a composite 2D seismic section has been interpreted through the major structural elements of the SWBS (seismic sections in Fig. 4B). Seismic sections have been selected through areas that were not affected by salt diapirs. Two additional horizons below the Base Cretaceous unconformity have been included into interpretation of the regional section: (1) Top Paleozoic has been interpreted and correlated from the neighboring Hammerfest Basin (purple horizon in Fig. 4B; Indrevaer et al 2016, Gernigon et al., 2014); (2) Top Basement, which has been penetrated by several wells on the Loppa High, and tied to the closest projected (12 km) well 7220/11-2s (red horizon in Fig. 4B). Furthermore, the regional section and interpretations were depth converted using internal velocities obtained from check shot data of wells 7117/9-2 and 7119/7-1 for shallow parts, and from well 7119/9-1 for the deeper parts (Table 1).

In order to facilitate interpretation of deeper structures and delineate its lateral extent, a simple 2D free air gravity anomaly modelling was performed along the depth converted regional section (Fig.4A). Gravity modelling was performed using the GM-SYS Profile Modelling software from Geosoft (https://www.geosoft.com). A 2D model has been divided into constant density polygons corresponding to the main sequence boundaries described in the seismic interpretation (Table 1; grey, red and orange colored polygons on Fig. 4B). The average densities for sedimentary and crustal rocks are derived from the publication of Gernigon et al. (2014). Top-basement and Moho were used as main density contrasts in the lithosphere (Fig. 4B).

The results of the gravity anomalies modelling showed that the Moho depth is consistent with compilation of the crustal depths of Ritzmann et al. (2007) and therefore used as a reference contrast. Discrepancy occurred between modeled and actual (interpreted) basement depth below the Loppa High and Hammerfest Basin (dashed and red horizons in Fig. 4B). Adjustment of modelled basement depth to a shallower level
would have resulted in offsetting the Moho depth to a deeper and more unrealistic results. Therefore, it has been decided to use the interpreted basement depth with the key assumption that basement rocks below the Loppa High and Hammerfest Basin are likely to have higher densities than modelled. Despite of the uncertainties in the gravity anomalies modeling, for the purpose of this study, it was much more important to constrain the sizes and geometries of the structures than to constrain the relative densities.

Table 1 Density polygons and stratigraphy for the gravity model, and interval velocities for depth conversion. Densities were obtained from publication of Gernigon e al. (2014). Interval velocities obtained from check shot data of wells 7117/9-2, 7119/7-1, and 7119/9-1.

<table>
<thead>
<tr>
<th>Sedimentary unit</th>
<th>Stratigraphy</th>
<th>Velocity (m/s)</th>
<th>Density (kg/m3)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Water</td>
<td>Seafloor</td>
<td>1480</td>
<td>1,03</td>
</tr>
<tr>
<td>Sediment 1</td>
<td>Eocene and younger</td>
<td>1800</td>
<td>2,30</td>
</tr>
<tr>
<td>Sediment 2</td>
<td>Paleocene</td>
<td>1900</td>
<td>2,45</td>
</tr>
<tr>
<td></td>
<td>Upper Cretaceous</td>
<td>2300</td>
<td></td>
</tr>
<tr>
<td>Sediment 3</td>
<td>Lower Cretaceous</td>
<td>2775</td>
<td>2,55</td>
</tr>
<tr>
<td></td>
<td>Jurassic/Triassic</td>
<td>3307</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Paleozoic</td>
<td>4000</td>
<td></td>
</tr>
<tr>
<td>Upper Crust</td>
<td>Basement</td>
<td>5000</td>
<td>2,75</td>
</tr>
<tr>
<td>Lower Crust</td>
<td></td>
<td></td>
<td>2,90</td>
</tr>
</tbody>
</table>

2D structural restoration:

A 2D structural restoration of a regional profile was performed to show the sequential evolution of the Tromsø Basin. 2D Move software (https://www.mve.com) was used to produce a 2D kinematic restoration. The workflow and methods consist of:

1. Erosion estimates, where missing sediments were restored on top of the section. This is important to compensate for isostasy due to sediment unloading. Missing sediments were given properties of the underlying stratigraphy.
2. Compaction and decompaction. The 2D compaction/decompaction tool in the software includes default compaction curves, which define different relationships between porosity and depth. The Sclater-Christie compaction curve were selected, as it is most appropriate for sandstones and mixed sedimentary sequences (Sclater and Christie, 1980). Compaction is only applied for missing sediments (eroded) in order to place removed sediments back and compensate for isostatic load. Decompaction was used sequentially for each restoration step by removing the uppermost sedimentary units. This is applied to correct for the effects of physical compaction in the sedimentary succession and vertically shift this part of the section to simulate an isostatic adjustment for each time-step. Flexural isostasy was applied to consider the isostatic response to sedimentary unloading during decompaction. Average values have been used for sediment (2400 kg/m³; Table 1) and mantle (3300 kg/m³; Robertson (1966)) densities, as well as elastic thickness (15000 m; Roberts et al. (1998)) and Young’s modulus (70000 Mpa; Watts et al. (1982));

3. 2D unfolding has been applied extensively during restoration. This option allowed geological horizons to be restored to a pre-deformed stage. The horizons were unfolded using “Simple Shear” and “Flexural Slip” methods (Gibbs, 1983; Verrall, 1981; Withjack and Peterson, 1993). The Simple Shear algorithm is best suited for flattening a regional dip that does not dip too steeply, but the limiting factor with this algorithm is that line length is not preserved. The Flexural Slip algorithm works by rotating the limbs of a fold to a datum or assumed regional geometry. Layer parallel shear is then applied to the rotated fold limbs in order to remove the effects of the flexural slip component of folding. In the study area, Simple Shear and Flexural Slip Unfolding is carried out, following the restoration of the faulted displacement at the surface. Most of the normal faults indicate shear angle ranging between 60° and 90° and the main fold deformation is in the hanging wall side of the main
faults and are associated either with fault propagation folding or fault-ramp-faulting.

4. Fault reconstruction. This method use the "Fault Parallel Flow" algorithm (Egan et al., 1997), where offset horizons on hangingwall and footwall were restored to its pre-faulted levels. The Fault Parallel Flow algorithm subdivide the fault into discrete dip domains and flow lines, along which, hanging wall material moves, maintaining line-length and area.
Figure 14 A) Un-interpreted and B) interpreted NW – SE seismic sections though the northern parts of the Tromsø Basin displaying basin configuration. Note distribution of main Lower Cretaceous sequences and interpreted fault families around the Veslemøy High and Ringvassøy-Loppa Fault Complex.
LOWER CRETACEOUS SEISMIC UNITS

Seismic unit 1 (U1): BCU – Late Barremian

Description. U1 has been penetrated by wells 7019/1-1, 7119/9-1 and 7220/10-1 in the eastern flank of the Tromsø Basin, where it is characterized by either very condensed (30 – 50 m) or missing pre – Barremian strata (Fig. 3A). The lower boundary of U1 is delimited by the BCU (green horizon; Figs. 5 – 10), which is represented by a high amplitude and continuous seismic event with an angular relationship to its underlying reflectors (Figs. 5A and 5B). The upper boundary is marked by a prominent continuous seismic event that correspond to the Late Barremian horizon (red horizon; Figs. 5B and 8B). Internally, it is characterized by a discontinuous and sometimes chaotic seismic reflectors in the southern part and by continuous, parallel and divergent reflectors in the northern part of the basin (Figs. 7A and 7B).

This seismic unit was deposited in the central part of the basin and thins out towards the margin (e.g. Veslemøy High and Senja Ridge; Figs. 5B and 6B). Local and segmented depocenters with maximum thickness of up to 2 seconds (TWT) are observed along the axis of the basin (Fig. 11A).

Interpretation. This unit is interpreted as early syn-tectonic deposition as suggested by the presence of growth strata and wedge-like geometry (Figs. 5B and 6B). Condensed and missing pre-Barremian strata along the eastern flank support active tectonic settings (wells 7019/1-1 and 7220/10-1; Fig. 3A). Previous interpretation from wells 7120/10-2, 7019/1-1 and 7220/10-1 in the same interval suggest turbidite and outer shelf deposits distributed along the eastern margin of the basin, in the Hammerfest Basin, Finnmark Platform and Loppa High (Fjeld and Escalona, 2014; Marin et al., 2017a; Marin et al., 2017b; Seldal, 2005). Therefore, distal equivalents of these sediments might have been deposited in depocenters, in the central part of the Tromsø Basin.
Figure 15 A) Un-interpreted and B) interpreted NW – SE seismic sections though the southern parts of the Tromsø Basin displaying basin configuration. Note distribution of main Lower Cretaceous sequences and interpreted fault families around the Senja Ridge and southern extent of the Ringvassøy-Loppa Fault Complex.
Seismic unit 2 (U2): Aptian – Albian

Description. The GR pattern for this interval is spiky and irregular with relatively higher values than in U1, and well reports and previous publication suggest main lithology consists of siltstone and claystone rocks (Fig. 3A) (Fjeld and Escalona, 2014; Marín et al., 2017a; Marin et al., 2017b; Seldal, 2005).

The lower and upper boundaries of the U2 are limited by Late Barremian and Late Albian horizons, which are characterized by very prominent continuous high amplitude seismic events (red and dark blue horizons; Figs. 5 – 10). Internally, this unit is characterized by discontinuous to chaotic seismic reflectors in the southern part (Fig. 7B); and by relatively continuous, parallel and divergent reflectors in the northern part of the basin (Figs. 5B and 8A). Downlap on the underlying seismic unit U1 (Late Barremian horizon) and growth strata in the lower part of the U2 are locally observed in the northern part of the basin, along the southwestern terraces of the Loppa High (Fig. 8A). Furthermore, in the upper part of the U2, several packages of divergent reflectors that downlap on continuous high amplitude reflectors are observed along salt diapirs in the central part of the basin (Fig. 8B). The external geometry of these packages resemble wedge-like shape that are stacked along salt walls (Fig. 8B). This seismic unit 2 distributed in entire basin and is characterized by a thick, wedge-shape sedimentary package (Figs. 5B and 6B). The northern part of the basin serves as a main depocenter with maximum thickness reaching 3 s (TWT) (Fig. 11B).
Figure 16 A) Un-interpreted and B) interpreted NE – SW seismic sections though the basin axis of the Tromsø Basin. Note distribution of main Lower Cretaceous sequences and interpreted fault families. Salt diapirs most likely masking some faults of FF2.
Interpretation. U2 is interpreted as the main syn-tectonic period, as suggested by the presence of growth strata and thickening of this interval towards the main faults (Figs. 4A and 5B). Abrupt changes in the thickness of U2 along the margins also support active tectonism (Fig. 6B). Divergent reflectors packages along salt diapirs flanks are interpreted as peridiapiric wedges (Rojo and Escalona (2018); Fig 8B). They indicate periods of salt movement towards the end of the U2 (Albian) that may have caused the uplifts along the axis of the basin. Previous interpretations from wells 7120/10-2, 7019/1-1 and 7220/10-1 on the same interval suggest shallow shelf deposits with clastic wedges along downfaulted terraces of the Hammerfest Basin, Finnmark Platform and Loppa High (Fjeld and Escalona, 2014; Marin et al., 2017a; Marin et al., 2017b; Seldal, 2005). Therefore, a deeper or distal depositional environment is suggested for the Tromsø Basin with exposed and isolated domal structures related to upward movement of the salt diapirs.

Seismic unit 3 (U3): Cenomanian

Description. U3 is composed of relatively thin (20 – 70 m) interval that has been penetrated by all wells in the Tromsø Basin. The main lithology consists of siltstone and claystone rocks with occasional dolomite stringers (Fig. 3A). The GR pattern is relatively consistent with lower values compared to U2 (Fig. 3A).

The lower and upper boundaries of U3 are constrained by Late Albian and Cenomanian horizons, which are represented by continuous, high amplitude seismic events (purple and dark blue horizons; Figs. 5 – 10). Internally, this unit is characterized by discontinuous and sometimes chaotic seismic reflectors in the southern part and by continuous, parallel and divergent high amplitude reflectors in the northern part of the basin ((Figs. 7A and 7B). Laterally, U3 thickens towards the central part of the basin and pinches out against the margins (Figs. 5B and 6B). Seismic packages with wedge shape geometries are observed in the small half grabens on the Veslemøy High (Fig. 5B and 10B). This unit is mostly
distributed in the central and southern parts of the Tromsø Basin with a southwestward increase in thickness, where it reaches a maximum of 2 seconds (TWT) (Fig. 11C).

Interpretation. The U3 is interpreted as syn-tectonic deposition, as supported by the presence of growth strata and wedges towards the axis of the basin (Figs. 5B and 6B). Based on the abundance of siltstones and dolomites, an outer shelf (open marine) depositional environment is suggested for this period.

Figure 17 A) Un-interpreted and interpreted NW – SE section along terraces of the Ringvassøy – Loppa Fault complexes illustrating detailed seismic configurations of the interpreted Lower Cretaceous units; B) Un-interpreted and interpreted NE – SW section showing peridiapiric wedges stacked along salt walls.
FAULTS

Three main fault families (i.e. faults with similar structural style, strike and age) affecting the Lower Cretaceous units are interpreted in the study area. Most these faults are interpreted as thick skin basement involved faults, but some thin skin faults were also recognized.

Fault family 1 (FF1) consists of a series of NE–SW striking faults interpreted in the northern and southern parts of the Tromsø Basin (Fig. 12). The lateral extent of these faults is interpreted as a part of the southern segments of the Bjornøyrenna and Troms–Finnmark fault complexes (Fig. 12). These faults are characterized as normal faults, which are interpreted along the Veslemøy High and Senja Ridge, and the southern segments of the Ringvassøy-Loppa Fault Complex (Figs. 5B and 6B). Faults are almost planar for the Paleocene and Cretaceous intervals, but have a listric expression at depth, where they merge into a low-angle plane below 5 - 6 s (TWT) (Fig. 6B). Horizon offset varies from approximately 300 to 500 ms (TWT), with a maximum in the Late Barremian horizon (Fig. 6B). Interpreted growth packages in the Lower Cretaceous U1 and U3, as well as in the Upper Cretaceous and Paleocene indicates fault activity during these periods (Figs. 5B and 6B).

Fault family 2 (FF2) comprises N–S striking normal faults interpreted in the central part and along the eastern flank of the Tromsø Basin (Fig. 12). FF2 consist of (1) thick and (2) thin skin fault systems:

(1) Thick skin faults are interpreted as a normal faults along the eastern boundary of basin (Fig. 8A). These faults are almost planar for the Paleocene and Cretaceous intervals, and become low-angle below 8 - 9 s (TWT) (Figs. 5B). The maximum offset of ca. 400 ms (TWT) observed at the Late Barremian horizon (Fig. 5B). Fault activity is supported by several growth and wedge shape seismic packages observed in the Lower Cretaceous seismic units U1, U2 and U3 (Figs. 5B and 6B). West facing faults of FF1 interpreted as a part of the west-facing Ringvassøy – Loppa
Fault Complex (RFLC) that separates the Tromsø Basin from the Loppa High and the Hammerfest Basin. Laterally, these faults are straight to slightly curved including some en echelon fault segments with various degree of linkage, from soft-links via relay ramps to hard-links (Fig. 12). A few faults are interpreted in the central part of the basin and are masked by salt diapirs, therefore defining their age and amount of displacement is problematic (Figs. 6B and 7B).

Figure 18 A) Un-interpreted and B) interpreted N – S seismic sections between northern Senja Ridge and Veslemøy High. Interpreted asymmetric folds with a long limb dipping to the south were formed in response to the reverse movement of FF4 faults. Please refer to Figure 8 for colour legend and location.
(2) Thin skin faults of FF2 consist of normal faults observed along the eastern flank of the Veslemøy High and Senja Ridge. These east-facing faults are only developed within the Lower Cretaceous U1 and U2, and detach at the BCU horizon and Upper Jurassic interval (Fig. 5B). The amount of offset at the late Barremian horizon fluctuates between 10 – 25 ms (TWT) (Fig. 5B). Fault activity most likely occurred during the Albian stage, since all these faults are tipping out towards the end of U2.
Figure 19 A) Un-interpreted and B) interpreted NW – SE seismic sections along the Vesleøy High. Note interpreted strike slip fault of FF3, left lateral movement is suggested from eastward lateral extension of these faults that coincide with compression and strike slip faults observed on the Polhem Sub Platform and the Loppa High (e.g. Indrevaer et al., 2016; Gabrielsen et al., 2011; Omosanya et al., 2019; Ahlborn et al., 2014). Please refer to Figure 8 for colour legend and location.
Fault family 3 (FF3) consists of several E–W striking faults that were interpreted under the Veslemøy High (Fig. 12). These faults steepened with depth and form positive flower-like structure that tips at the Cretaceous strata (Fig. 10B). Minor growth and wedge shaped seismic packages towards faults were observed in the Lower Cretaceous U1 and U2, suggesting main periods of fault activity (Fig. 10B). These faults are interpreted as strike slip faults, and eastward lateral extension of these faults are coincide with compression and strike slip faults observed on the Polhem Sub Platform and the Loppa High (Ahlborn et al., 2014; Indrevær et al., 2016; Omosanya et al., 2017) (Fig. 12). These strike slip faults have left – lateral movement that support by changed fault polarities east dipping FF2 to west dipping FF1 across the strike of FF3 (Fig. 12), and displaced structural lineaments on the Polhem Sub Platform (e.g. tilt derivative map of Gernigon et al., 2014) and Loppa High (e.g. Swaen Graben; Omosanya et al., 2017).

Fault family 4 (FF4) include WNW-ESE-striking faults that are observed only along the northern tip of the Senja Ridge (Fig. 12). These faults are interpreted as a steep reverse faults that tip or truncated at the Paleocene strata and become listric at depth (Fig. 49B). It is difficult to examine the lateral extent of these faults due to salt diapirs along the axis of the basin (Fig. 12). Interpreted asymmetric folds with a long limb dipping to the south were formed in response to the reverse movement of these faults. The largest offset observed at BCU horizons with ca. 400 ms (TWT). Growth sequences and wedge-shaped packages are interpreted at the Lower Cretaceous seismic units U1 and U3; and Paleocene intervals indicating main faults activity, which may have triggered the uplift of the northern part of the Senja Ridge (Fig. 9B).
Figure 20 Series of time thickness maps of the Cretaceous units. A) Valanginian – Late Barremian shows early segmentation and isolated depocenters. Note that Veslemøy High and Senja Ridge uplift during this period. B) Aptian - Albian shows segmentation of the basin into a northern and southern segments, where northern segment is a major depocenter. This period is also associated with movement of the salt diapirs (grey arrows). C) Cenomanian shows shift of the depocenter towards southernmost parts of the Tromsø Basin. Note that during this period the Senja Ridge and Veslemøy High experienced major inversion and uplift. Salt movement also affecting this unit.
2D STRUCTURAL RESTORATION

Six restored steps reproduce the geological evolution of the Tromsø Basin from the Early Cretaceous until present (Figs. 13A – 13F). Furthermore, restored sections were grouped into three stages: (1) Pre–Cretaceous basin configuration, (2) Early Cretaceous and (3) post–Early Cretaceous evolution. Emphasis were given to basin geometry; timing of major fault activity; depocenter distribution; and the amount of extension honoring the data and interpretation as much as possible. Although, the pre and post Early Cretaceous history are not the main objective, they were included to highlight important episodes of the margin evolution, which may have implication to the Early Cretaceous evolution.

Stage 1. Pre-Cretaceous basin configuration

The early regional tectonic events during Permian – Carboniferous and Jurassic – Triassic marked the location of the proto Tromsø Basin with up to 15 km thick sediment sequences overlying basement (Fig. 13F). Until the end of the Late Jurassic, Tromsø and Sørvestnaget basins were a formed a single basin dominated by west facing listric faults of FF2 along the Loppa High (Fig. 10F). The SWBS margin was 35 km narrower compared to present day. Main uncertainties with the Late Jurassic – BCU restoration step were related to the poor seismic imaging below 7 seconds (TWT) and absence of well control to the NW of the Loppa High (Fig. 3B). Therefore, interpretation was speculative and primarily based on seismic reflection configuration and gravity modelling (Fig. 4B).
Figure 21 Fault interpretation maps. Location and subdivision of main interpreted fault families. Note interpreted strike slip fault on the Loppa High adapted from previous works of Indrevaer et al., 2016; Gabrielsen et al., 2011; Omosanya et al., 2019; Ahlborn et al., 2014. Location of proposed transfer zones in the North of the Tromsø Basin (red polygon) is coincide with westward lateral extent of the previously defined strike faults.
Stage 2. Early Cretaceous evolution

This stage includes two restoration steps: (1) Late Barremian and (2) Cenomanian (Figs. 13D and 13 E). The Late Barremian step corresponds to the Lower Cretaceous seismic unit U1 (Fig. 13E). The main depocenter observed along the Veslemøy High (Fig. 13E). At this step, the margin extended up to 13 km from Late Jurassic, where most of the displacement of 150 – 200 m was accommodated by faults of FF1 and FF2 along the western boundary of the Tromsø Basin (Fig. 13E). Minor compression might have been accommodated by sinistral strike slip movement along FF3 fault, which triggered the slight uplift of the Veslemøy High and resulted in the division of the Tromsø and Sørvestnaget basins at this time (Figs. 10B and 13E). Continues rifting led to additional extension of the margin by 9km during the Lower Cretaceous U2 and U3 step (Fig. 13D), increase subsidence led to bypass the Veslemøy High, and as the result, the Sørvestnaget and Tromsø basins acted as a single basin capturing up to 7 km of the Lower Cretaceous sediments (Fig. 13D). The main fault activity migrated towards the eastern margin of the Tromsø Basin, where the Ringvassøy – Loppa fault complex accommodated the main displacement of 300 – 400 m (FF2) (Fig. 13D). At this step, the margin extended up to 9 km more from Late Barremian step (Fig. 9D). It is suggested that some of this extension possibly attenuated by the Cenomanian (U3) inversion of FF3 and FF4 (Figs. 9B and 10B), that resulted in uplift of the Veslemøy High and northern Senja Ridge, and consequently reinforced the isolation of the Tromsø from the Sørvestnaget basins (Fig. 4A).
Figure 22 2D structural reconstruction of crustal cross section. Six restoration steps (A – F) are represent the geological evolution of the Tromsø Basin from the Early Cretaceous until present. For location of the line is see Figure 4.
**Stage 3. Post Early Cretaceous evolution**

This stage contains three restoration steps: (1) Late Cretaceous, (2) Paleocene and (3) Present day (Figs. 13A – 13C).

During the Late Cretaceous, most of the faulting occurred along the western flanks of the Veslemøy High (Fig. 13C). At this time, the margin extended up to 7 km from Cenomanian step, where major displacement of 800 m was accommodated by west facing listric faults of FF1 along the western flank of the Veslemøy High (Figs. 5B and 13C). Footwall uplift of the Veslemøy High resulted in erosion and degradation, as well as separation of the Sørvestnaget and Tromsø basins, where former were considered as main depocenter (Fig. 13C). The Paleocene step is characterized by ongoing faulting of FF1 along the Veslemøy High resulting to additional 4.5 km of extension (Fig. 13B). The Sørvestnaget Basin was still a major depocenter at this time containing up to 5 km of sediments (Fig. 13B). Most of the faulting occurred in the Sørvestnaget Basin, while the Tromsø Basin was relatively passive (Fig. 13B). Onlap and thinning of the Paleocene seismic reflectors suggest that Veslemøy High continued to uplift at least during initial period of this step (Figs. 6B and 10B). At the present day step (Eocene and younger), the margin extended 1.5 km (Fig. 13A). Main deformation at this time occurred to the West of the Sørvestnaget Basin, therefore the study area was only subjected to a minor extension (Fig. 13A). The main depocenter shifted towards the continent oceanic boundary.
DISCUSSION

Proposed evolutionary model

Most of the plate tectonic restorations in the North Atlantic margin from 145 Mya until 55 Mya, suggest that the SW Barents Sea was adjacent to the NE offshore Greenland (Barnett-Moore et al., 2018; Doré et al., 2015; Seton et al., 2012), and NE-SW structural lineaments were predominant in both margins. Latest compilation and comparison of various plate models suggest that main extension directions were orthogonal to the main NE-SW structural lineaments at least during the Early Cretaceous (e.g. plate flowlines of Barnett-Moore et al. (2018); Fig. 14A).

Structural evolution of the Tromsø Basin during the Early Cretaceous was vastly controlled by the NE-SW trending regional TFFC and BFC, which interpreted as FF1 in the study area (Fig. 14A and 14B). In contrast to FF1, the N-S striking FF2 is limited to the extent of the basin (Fig. 12), and most likely represent intra rift fault system (McClay et al., 2004; McClay et al., 2002). Presence of en echelon faults of FF2, which generated breached relay ramps along the western border of the Tromsø Basin, suggest that basin may have evolved in oblique manner (Fig. 12) (Agostini et al., 2009; Brune and Autin, 2013; Clifton et al., 2000; Corti, 2008; Withjack and Jamison, 1986). Oblique opening of the Tromsø Basin is also supported by the interpreted by strike slip faults of FF3 (Fig. 10B), which determine location of the intra basin transfer zone that responsible of changes in faults polarities of FF1 and FF2 (e.g. Corti, 2008; McClay et al., 2002; Fig. 12).

Based on timing of fault activity and kinematics, it is suggested that the Tromsø Basin experienced three major episodes of deformation: (1) Valanginian – Late Barremian; (2) Aptian – Albian; and (3) Cenomanian.

1) Valanginian – Late Barremian (U1): Ongoing rifting
Wells 7220/10-1 and 7019/1-1 located along the eastern margin of the Tromsø Basin indicate a condensed or absent Lower Cretaceous unit U1, suggesting that these areas were uplifted during rifting (Figs. 3A and 3B). Most of the Valanginian – Late Barremian extension was accommodated by west facing boundary faults of FF1 (Figs. 5B; 6B and 14A). Isolated depocenters along the axis of basin suggest on complex interaction between faults of FF1 (e.g. fault segments of TFFC and BFC), which resulted in formation of internal fault system of FF2 (Fig. 12). The N-S strike of FF2 that outline the WSW extent of the Loppa High suggest that basement heterogeneity possibly guided formation of FF2 by localizing strain (Baudon and Cartwright, 2008; Nicol et al., 2005; Richard and Krantz, 1991).

At this stage, the Veslemøy High and Senja Ridge were characterized as minor structural features that separated the Tromsø from Bjørnøya and Sørvestnaget basins (Figs. 6B and 13E). Onset of complex tectonic interaction between the Veslemøy High and northern Senja Ridge occurred through left lateral movement along faults of FF3 resulting in transpressional setting and slight uplift of the Veslemøy High (Figs. 9B and 10B). Previously, based on structural evolution of the neighboring Hammerfest Basin, it has been suggested that this episode comprises two rifting phases, Berriasian – Valanginian and Hauterivian – Barremian (Faleide et al. (1993)). However, considering the poor seismic imaging below 5 seconds (TWT) and the lack of well control in the deeper parts of the Tromsø Basin, these tectonic phases cannot be neither confirmed nor excluded.

2) Aptian – Albian (U2): Rift culmination

Major extension was accommodated by the intra basin faults of FF2 (e.g. RLFC; Figs. 12 and 14B). This is also supported by a large depocenter in the northern part of the Tromsø Basin (Fig. 11B). Towards the end of this stage, possible during Albian – Cenomanian, complex tectonic settings between the Veslemøy High and the northern Senja Ridge
resulted in uplift of both (Fig. 14B). Previously, formation of the Senja Ridge and Veslemøy High as positive structural features were attributed to either sinistral or dextral strike-slip movement along Bjørnøyrenna Fault Complex (Figs. 2A and 2B; Gabrielsen and Færseth, 1988; Riis et al., 1986). Our study suggest that uplift is most likely caused by the transpressional settings along FF3, which is also responsible for formation of compressional structures and faults of FF4 (Fig. 9B). One plausible explanation could be that during transpressional conditions, the strike-slip (e.g. FF3; Fig 10B) and dip-slip (e.g. FF4; 9B) components accommodated on separate but relatively parallel structures (e.g. the Veslemøy High and northern Senja Ridge; Fig. 12), whilst decoupling possibly occurred at pre Cretaceous sequences. Transpressional deformation along transfer zone is most likely related to the oblique opening of the Tromsø Basin, where basement heterogeneity localized strain distribution. Furthermore, towards the end of this stage, rapid increase of subsidence and differential loading triggered salt movement that have resulted in diapirism in the central part of the basin and development of the halokinetic sequences (Fig. 8B).

3) Cenomanian (U3): Post rift deformation and inversion

This episode is considered as tectonically quiescent in the central and eastern parts of the basin (Fig. 14C). Most the of the fault activity (FF1) occurred in the western and northwestern flanks of the Tromsø basin (Fig. 13C). Pinching out of the Lower Cretaceous unit U3 against the Veslemøy High and the Senja Ridge suggests that these structures were uplifting (Figs. 5B and 6B). Uplift is suggested to be caused by the same transpressional conditions along FF3 and FF4 (Fig. 9B) (Blaich et al., 2017; Breivik et al., 1998; Brekke and Riis, 1987; Riis et al., 1986). The uplift of Senja Ridge and Veslemøy High caused the isolation of the Tromsø Basin from the Sørvestnaget Basin during this time (Fig. 14C).
Figure 23 Proposed structural evolutionary model of the Tromsø Basin during Early Cretaceous: A) Valanginian – Late Barremian extension was accommodated by west facing boundary faults of FF1 (e.g. fault segments of TFFC and BFC), which resulted in formation of internal fault system of FF2; B) Aptian – Albian marked by a transpressional settings along transfer zone which is related to the oblique opening of the Tromsø Basin, where basement heterogeneity most likely localized stress and strain distribution; and C) Cenomanian episode is considered as tectonically quiescent, where most the of the fault activity occurred in the western and northwestern flanks of the Tromsø basin.

Margin extension and crustal thinning

The constructed crustal section through the Tromsø Basin is considered to be the most representative of the direction of extension for the SWBS margin. It is almost perpendicular to the major structural elements (Fig. 4B). Constructed and subsequently restored section resembles the “boundinage” model of Gernigon et al., 2014 (Fig. 2C), where the origin of the listric faults (e.g. FF1) is suggested to be the remnants of Caledonian thrust faults that reactivated in response to rifting (Gernigon et al., 2014). The Early Cretaceous reactivation most likely caused backsliding of the Caledonian thrust and triggered upward propagation of fault segment from pre-existing basement faults (Baudon and Cartwright, 2008; Nicol et al., 2005; Richard and Krantz, 1991). The result of sequential restoration suggests that these faults were responsible for 35 km margin extension from the Earliest Cretaceous to the present
day (Figs. 13A – 13F). Previous work by Breivik et al. (1998) estimated a post-middle Jurassic extension of 70 – 85 km. Such discrepancies most likely related to calculation methods, where our calculations of 35 km of extension over such time interval seems reasonable and more reliable. Based on the modelled depths of the basement and Moho (Figs. 4 and 13A), post-rift crustal thickness below the Tromsø Basin reaches minimum of 8 km. Assuming that the original pre-rift crustal thickness was 30 – 35 km (Barrère et al., 2009; Clark et al., 2013; Ritzmann and Faleide, 2007), and after dividing pre-rift by post-rift crustal thicknesses we estimated a cumulative crustal thinning (β) factor of 3.7 – 4.4 and 2.9 – 3.1 below the Tromsø and Sørvestnaget basins since the BCU. The calculated β factor is consistent with previous estimation by Breivik et al. (1998) for the Tromsø Basin (β factor of ca.4). Globally, the β factors of rifted margins generally increase towards the continent–oceanic boundary (COB), where the maximum crustal thinning is usually occurs at the location of breakup (Crosby et al., 2011; Montési and Behn, 2007). In case of SWBS, crust is thinner below the Tromsø Basin as compared to a Sørvestnaget Basin, which is relatively closer to a COB. Therefore, common characteristics for rifted margins (Peron-Pinvidic et al. 2013), where crust thins towards the distal domains is not applicable for the SWBS margin. This difference can be attributed to a transform or oblique nature of the Early and post – Early Cretaceous extension in the SWBS, where increase in obliquity of rifting may decrease crustal thinning and stretching towards the continental breakup (Montési and Behn, 2007). This is also advocating to the oblique opening of the Tromsø Basin. Moreover, high crustal thinning (β factor of 3) is observed below the Bjørnøya Basin by Clark et al. (2013) and Gernigon et al. (2014), which suggest that both Tromsø and Bjørnøya basins may have similar origin and were influenced by the same tectonic regimes.
CONCLUSIONS

• The Early Cretaceous evolution of the Tromsø Basin is influenced by the inherited basement structures from the Caledonian orogeny. The presence of west facing low angle detachment faults below the Tromsø Basin supports the idea of post orogenic collapse of Caledonian thrust sheets, that were periodically reactivated or backslided in response to an Early Cretaceous extensional episode.

• The proposed structural model for the Early Cretaceous evolution involve oblique opening of the Tromsø Basin and formation of the intrabasinal transfer zone with compressional strike slip faults (Figs. 13A – 13C). This model differs from any previously proposed models and partially resembles both Faleide et al. (1993) and Gernigon et al., (2014) models (Figs. 2B and 2C).

• Extension of 35 km is proposed for the SWBS margin since the Earliest Cretaceous. Among the Cretaceous extensional episodes, the Valanginian – Late Barremian is considered as the period of major crustal extension (13 km) in the SWBS margin (Figs. 13E and 13F). During most of the Early Cretaceous, the Sørvestnaget and Tromsø basins were a single large basin and were separated by activity of FF3 and FF4 faults, resulting in the uplift of the Veslemøy High and northern Senja Ridge.

• Distribution of the crustal thinning (β factor) in the SWBS is unlike for common rifted margins. Crust below narrow and confined Tromsø Basin is suggested to be thinner than in the Sørvestnaget Basin. This supports that the Early Cretaceous rift in the Tromsø Basin involved certain degree of obliquity.
ACKNOWLEDGEMENT

This study is part of the industrial sponsored LoCrA consortium (http://locra.ux.uis.no). We would like to express our gratitude to DISKOS database and MultiClientGeophysical AS for providing 2D seismic data and allowing to publish them. In addition, we are thankful to Halliburton-Landmark and Move for providing the software and license.
REFERENCES


Brune, S., Williams, S.E., Müller, R.D., 2018. Oblique rifting: the rule, not the exception. Solid Earth 9, 1187-1206.


Barents Sea: From Caledonian orogeny to continental breakup. Tectonics 33, 347-373.


Indrevær, K., Bergh, S.G., 2014. Linking onshore-offshore basement rock architecture and brittle faults on the submerged strandflat along the SW Barents Sea margin, using high-resolution (5 × 5 m) bathymetry data. Norsk Geologisk Tidsskrift 94, 1-34.


Lizarra


Lizarra


Lizarra


Lizarra


Lizarra


Lizarra


Lizarra


Lizarra


Lizarra


Lizarra


Lizarra


Lizarra


Lizarra


Lizarra


Lizarra


Lizarra


Lizarra


Lizarra


Lizarra


Lizarra


Lizarra


Lizarra


Lizarra


Lizarra


Lizarra


Lizarra


Lizarra


Lizarra


Lizarra


Lizarra


Lizarra


Lizarra


Lizarra


Lizarra


Lizarra


Lizarra


Lizarra

Lehne


Global continental and ocean basin reconstructions since 200Ma. Earth-Science Reviews 113, 212-270.


Growth and linkage of a basin-bounding fault system: Insights from the Early Cretaceous evolution of the northern Polhem Subplatform, SW Barents Sea.

Bereke Kairanov, Dora Marín, Alejandro Escalona, Nestor Cardozo

Journal of Structural Geology, 124, 2019, 182-196, ISSN 0191-8141,
https://doi.org/10.1016/j.jsg.2019.04.014
Growth and linkage of a basin-bounding fault system: Insights from the Early Cretaceous evolution of the northern Polhem Subplatform, SW Barents Sea

Bereeke Kinnaird, Dora Marín, Alejandro Escalona, Nentor Cardoso

Department of Geology, University of Southampton, PO Box 241, Southampton, UK.

A B S T R A C T

Fault kinematics, growth and linkage are important processes in the evolution of major basin-bounding normal fault systems (Ewing and Lander, 2000; Arbuckle, 1985; Le Febre et al., 2016; Marín et al., 2017). Fault linkage models are important tools for understanding the behavior and interaction of fault segments (e.g., Arbuckle et al., 1994; Schlische, 1992; Sibuet et al., 2003). Observations from natural and synthetic models, and analog and numerical models suggest two ways of fault growth: (1) the isolated fault model, where growth and linkage of individual fault segments occur through displacement and lateral propagation of their tips (Arbuckle et al., 1994, 2004; Davison and Anderson, 1999; Davison et al., 2002; Walsh and Verrono, 2004; Waterhouse et al., 1994; Waterhouse et al., 1997; Waterhouse et al., 1998), and (2) the common growth fault model, where the faults grow as a multi-fault linkage model during the evolution of the basin-bounding faults (e.g., Arbuckle et al., 1994; Waterhouse et al., 1997; Waterhouse et al., 1998). In this context, it is important to understand the growth and linkage of fault systems in both models and their potential impact on hydrocarbon exploration and resource development.

1. Introduction

Fault interaction, growth and linkage are important processes in the evolution of major basin-bounding normal fault systems (Ewing and Lander, 2000; Arbuckle, 1985; Le Febre et al., 2016; Marín et al., 2017). Fault linkage models are important tools for understanding the behavior and interaction of fault segments (e.g., Arbuckle et al., 1994; Schlische, 1992; Sibuet et al., 2003). Observations from natural and synthetic models, and analog and numerical models suggest two ways of fault growth: (1) the isolated fault model, where growth and linkage of individual fault segments occur through displacement and lateral propagation of their tips (Arbuckle et al., 1994, 2004; Davison and Anderson, 1999; Davison et al., 2002; Walsh and Verrono, 2004; Waterhouse et al., 1994; Waterhouse et al., 1997; Waterhouse et al., 1998), and (2) the common growth fault model, where the faults grow as a multi-fault linkage model during the evolution of the basin-bounding faults (e.g., Arbuckle et al., 1994; Waterhouse et al., 1997; Waterhouse et al., 1998). In this context, it is important to understand the growth and linkage of fault systems in both models and their potential impact on hydrocarbon exploration and resource development.

bene (e.g., Schaefer et al., 2004; Verrono et al., 2004; Waterhouse et al., 1998). In this context, it is important to understand the growth and linkage of fault systems in both models and their potential impact on hydrocarbon exploration and resource development.

1. Introduction

Fault interaction, growth and linkage are important processes in the evolution of major basin-bounding normal fault systems (Ewing and Lander, 2000; Arbuckle, 1985; Le Febre et al., 2016; Marín et al., 2017). Fault linkage models are important tools for understanding the behavior and interaction of fault segments (e.g., Arbuckle et al., 1994; Schlische, 1992; Sibuet et al., 2003). Observations from natural and synthetic models, and analog and numerical models suggest two ways of fault growth: (1) the isolated fault model, where growth and linkage of individual fault segments occur through displacement and lateral propagation of their tips (Arbuckle et al., 1994, 2004; Davison and Anderson, 1999; Davison et al., 2002; Walsh and Verrono, 2004; Waterhouse et al., 1994; Waterhouse et al., 1997; Waterhouse et al., 1998), and (2) the common growth fault model, where the faults grow as a multi-fault linkage model during the evolution of the basin-bounding faults (e.g., Arbuckle et al., 1994; Waterhouse et al., 1997; Waterhouse et al., 1998). In this context, it is important to understand the growth and linkage of fault systems in both models and their potential impact on hydrocarbon exploration and resource development.

1. Introduction

Fault interaction, growth and linkage are important processes in the evolution of major basin-bounding normal fault systems (Ewing and Lander, 2000; Arbuckle, 1985; Le Febre et al., 2016; Marín et al., 2017). Fault linkage models are important tools for understanding the behavior and interaction of fault segments (e.g., Arbuckle et al., 1994; Schlische, 1992; Sibuet et al., 2003). Observations from natural and synthetic models, and analog and numerical models suggest two ways of fault growth: (1) the isolated fault model, where growth and linkage of individual fault segments occur through displacement and lateral propagation of their tips (Arbuckle et al., 1994, 2004; Davison and Anderson, 1999; Davison et al., 2002; Walsh and Verrono, 2004; Waterhouse et al., 1994; Waterhouse et al., 1997; Waterhouse et al., 1998), and (2) the common growth fault model, where the faults grow as a multi-fault linkage model during the evolution of the basin-bounding faults (e.g., Arbuckle et al., 1994; Waterhouse et al., 1997; Waterhouse et al., 1998). In this context, it is important to understand the growth and linkage of fault systems in both models and their potential impact on hydrocarbon exploration and resource development.

1. Introduction

Fault interaction, growth and linkage are important processes in the evolution of major basin-bounding normal fault systems (Ewing and Lander, 2000; Arbuckle, 1985; Le Febre et al., 2016; Marín et al., 2017). Fault linkage models are important tools for understanding the behavior and interaction of fault segments (e.g., Arbuckle et al., 1994; Schlische, 1992; Sibuet et al., 2003). Observations from natural and synthetic models, and analog and numerical models suggest two ways of fault growth: (1) the isolated fault model, where growth and linkage of individual fault segments occur through displacement and lateral propagation of their tips (Arbuckle et al., 1994, 2004; Davison and Anderson, 1999; Davison et al., 2002; Walsh and Verrono, 2004; Waterhouse et al., 1994; Waterhouse et al., 1997; Waterhouse et al., 1998), and (2) the common growth fault model, where the faults grow as a multi-fault linkage model during the evolution of the basin-bounding faults (e.g., Arbuckle et al., 1994; Waterhouse et al., 1997; Waterhouse et al., 1998). In this context, it is important to understand the growth and linkage of fault systems in both models and their potential impact on hydrocarbon exploration and resource development.

1. Introduction

Fault interaction, growth and linkage are important processes in the evolution of major basin-bounding normal fault systems (Ewing and Lander, 2000; Arbuckle, 1985; Le Febre et al., 2016; Marín et al., 2017). Fault linkage models are important tools for understanding the behavior and interaction of fault segments (e.g., Arbuckle et al., 1994; Schlische, 1992; Sibuet et al., 2003). Observations from natural and synthetic models, and analog and numerical models suggest two ways of fault growth: (1) the isolated fault model, where growth and linkage of individual fault segments occur through displacement and lateral propagation of their tips (Arbuckle et al., 1994, 2004; Davison and Anderson, 1999; Davison et al., 2002; Walsh and Verrono, 2004; Waterhouse et al., 1994; Waterhouse et al., 1997; Waterhouse et al., 1998), and (2) the common growth fault model, where the faults grow as a multi-fault linkage model during the evolution of the basin-bounding faults (e.g., Arbuckle et al., 1994; Waterhouse et al., 1997; Waterhouse et al., 1998). In this context, it is important to understand the growth and linkage of fault systems in both models and their potential impact on hydrocarbon exploration and resource development.

1. Introduction

Fault interaction, growth and linkage are important processes in the evolution of major basin-bounding normal fault systems (Ewing and Lander, 2000; Arbuckle, 1985; Le Febre et al., 2016; Marín et al., 2017). Fault linkage models are important tools for understanding the behavior and interaction of fault segments (e.g., Arbuckle et al., 1994; Schlische, 1992; Sibuet et al., 2003). Observations from natural and synthetic models, and analog and numerical models suggest two ways of fault growth: (1) the isolated fault model, where growth and linkage of individual fault segments occur through displacement and lateral propagation of their tips (Arbuckle et al., 1994, 2004; Davison and Anderson, 1999; Davison et al., 2002; Walsh and Verrono, 2004; Waterhouse et al., 1994; Waterhouse et al., 1997; Waterhouse et al., 1998), and (2) the common growth fault model, where the faults grow as a multi-fault linkage model during the evolution of the basin-bounding faults (e.g., Arbuckle et al., 1994; Waterhouse et al., 1997; Waterhouse et al., 1998). In this context, it is important to understand the growth and linkage of fault systems in both models and their potential impact on hydrocarbon exploration and resource development.
2. Geological setting

The Polurno Subplatform is a 50–100 m deep black-floored area located on the western part of the Laptev Sea, in the SW Barrier. It is a ~700 m long and ~20 km wide area delineated by an array of deep-to-the-west nodules fields of the Boretsky and Ringkampyara-Laptev fault complex (RFC and RFC). The subplatform was a centrally located element of the Laptev high in the Jurassic, and it was abandoned late in the Cretaceous (Burtsev et al., 2017). The Polurno Subplatform is a large-scale wedge in the Laptev subbasin that is not connected with the subplatform but is still active during the late Jurassic and early Cretaceous. It was active during the late Jurassic and early Cretaceous (Burtsev et al., 2017). The Polurno Subplatform is a large-scale wedge in the Laptev subbasin that is not connected with the subplatform but is still active during the late Jurassic and early Cretaceous (Burtsev et al., 2017). The Polurno Subplatform is a large-scale wedge in the Laptev subbasin that is not connected with the subplatform but is still active during the late Jurassic and early Cretaceous (Burtsev et al., 2017).
Fig. 3. (A) Location of the study area in the Polhem Subplatform, Barents Sea. Geographic map based on data from the Norwegian Petroleum Directorate (NPD): Barents/Isfjorden/Jan Mayen basin complex, IBC: Bilevnesen basin complex, IBC: Trondheim/Brage basin complex. (B) Regional area section through the major structural elements surrounding the study area (modified from Ramqvist et al. 2005). Line of section is shown in A.
formation are able to supply (reference, well, etc). To enhance the continuity of the reference, a structural smoothing technique (Ridley et al., 1993) was applied to the seismic data before interpretation.

Seismic trends were interpolated to the study area. Only one large fault (Fig. 2, Fig. 3A) was selected for this analysis, since it has obscured the Lower Cretaceous sequence on both the landward and marine margins. Three methods were used to determine the growth history of this fault: (1) thickness vs. length (T vs. L) plot, which shows the distribution of these along the fault (Chiles et al., 1992; Cardamone and Long, 2000; Wahl and Watkinson, 1993); (2) trend analysis, which quantifies the style of growth of seismic-matched isolated faults across (Chaynes and Mortl, 1991; Olds et al., 1992, 1993, 1994, 1995); and (3) flow thickness maps, which show changes in vertical thickness adjacent to fault. The inferred fault growth rates (Ridley et al., 1993; Chiles et al., 1992; Jackson and Watkinson, 1993; Miley, 2002; DeBelle, 1993).

Fault trends are defined by seismic data (trend line). Three different horizontal seismic profiles (T vs. L) are used to determine the horizontal position of the fault. These are recorded every 100 km along the track strike direction. The length of the fault normal to the strike, measured from the lowest point on the fault, is considered to be its length (Fig. 4). All three length points show depth consistent with a range of 3-metres (horizontal) velocity at 120 km (Fig. 3B).

Three horizontal drainage areas were used to determine the original trend of migration of the salient boreholes at the same depth level position (Olds et al., 1993). The observed measurements were performed at present day (corrected) thickness and may include corrections related to tectonics (Taylor et al., 1995). One method is based on the Lower Cretaceous sequence, which includes the intra-lithic deformation. The intra-lithic deformation is an important factor in the growth of seismic-matched isolated faults across (Chaynes and Mortl, 1991; Olds et al., 1992, 1993, 1994, 1995; Jackson and Watkinson, 1993; Miley, 2002; DeBelle, 1993).

5. Fault blocks and Lower Cretaceous sequences

Description: The northern Pattern Subbasin is characterized by a series of fault blocks and fault scarps (Fig. 4 and 5). The Lower Cretaceous sequence is well developed in one of these fault blocks by well 7229/5-2 (Chaps. 15, 16, and 17). The base of the Lower Cretaceous is delineated by the A2, which is represented by a high amplitude and continuous seismic event with a strong angular reflection in the underlying sequence (Fig. 4B). The top of the Lower Cretaceous is marked by a prominent continuous seismic event corresponding to the top of sequence 4 of Middle Miocene age (Chaps. 16, 17, and 18). Internally, sequence 9 and 1 are characterized by wedge-shaped geometries thickening towards the fault (Figs. 4A and 5A). Seismic reflection is continuous in the

---

**Table**: Summary of the seismic data and fault block analysis.

<table>
<thead>
<tr>
<th>Fault Block</th>
<th>Length (km)</th>
<th>Width (km)</th>
<th>Depth (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Block A</td>
<td>100</td>
<td>80</td>
<td>3000</td>
</tr>
<tr>
<td>Block B</td>
<td>120</td>
<td>90</td>
<td>2500</td>
</tr>
<tr>
<td>Block C</td>
<td>150</td>
<td>100</td>
<td>2000</td>
</tr>
</tbody>
</table>

---

**Figure**: Seismic data showing fault blocks and lower Cretaceous sequences.
continental, with development relations to the underlying miocene (Fig. 9). Sequences 2 to 4 are mainly represented by continental, holo-epiclastic, paralic couplets (Fig. 4A, 5A, and 6A). The tectonic deposition of 5 and 6, or located along the major bounding fault and the Miocene sandstones in the middle Miocene (between 120 and 400 ms FWT) (Fig. 4A). Sequences 2 to 4 are deposited in the entire study area, with a major sequence in the upper Miocene (Fig. 4A and 6A). The late Miocene deposition of 5 and 6 is located in the NW part of the area, with a maximum thickness of 1,000 ms (Fig. 4B).

Implications. The Lower Cretaceous sequence was deposited during active faulting, as suggested by the presence of wedge-shaped tectonic packages, thickness changes, and the relationships towards the bounding faults (e.g., 5A, 3B, and 6A). Based on the configuration of the seismic reflection and subductional tectonic setting (e.g., 4A), the deformed setting in the study area is an isolated Cretaceous wedge that was deposited in shallow marine to shelfal environments. This indicates that the sequences were deposited at shallow water depths and preserved to the actual sheeting. 5A - 6A were deposited in a relatively stable environment, with minor marine setting, as suggested by an overall increase in shale content.

6. Incisions

Deposition. Four main unconformities are present in the study area:

- Fig. 4: Cretaceous (top) and Jurassic (bottom) bounding faults (Figs. 9, 10, and 11) at the time of the first Jurassic (Upper Jurassic) unconformity (Fig. 10). The upper Jurassic and Lower Cretaceous will be truncated from wedge-shaped packages in the middle and Lower Cretaceous, while they are either eroded or confined to the upper fault block towards the lower fault block.

- Fig. 5: Top Jurassic and Lower Cretaceous unconformity (Fig. 11). The upper Jurassic and Lower Cretaceous will be truncated from wedge-shaped packages in the middle and Lower Cretaceous, while they are either eroded or confined to the upper fault block towards the lower fault block.
Fig. 5. (a) Line drawing in 1:5000 scale illustrating the lower Cretaceous sequence and faults. Image courtesy of Western Geologic Nutrients. (b) Fault family 1 showing the lower Cretaceous sequence (Fig. 1) and faults bounding fault. Locations of A and B are shown in Fig. 6.

tional valleys at synthetic overlapping transfer zones suggest that their formation was most likely controlled by fault activity (Fig. 7A). Moreover, the more prominent trends in one of the incised valleys resemble fault-like features cutting across the valley, also referred to as "walk-ups" (Park and King, 2004). These features appear similar to those observed in the present-day New Jersey coastal plain, where incised and exhumed embayments are also interpreted as abandoned backshoals forming in response to seaward-receding shoreline (Fig. 8B) (Horn et al., 1985; McKee et al., 2000). The formation of several backshoals along the coastline is most likely associated with gradient changes caused by coastal activity, where backshoals commonly separate upcoast (e.g., western Baja Delta, internal and divergent, 2017).

7. Fault analysis

7.1. Fault maps

The interpreted fault maps consist of topographic and geologic data, combined with aerial photographs, the fault traces were grouped into two distinct fault families based on their structural trend and spacing. Fault maps were generated using data from the Cretaceous faults (Fig. 9) and the Lower Cretaceous faults (Fig. 10). These faults belong to the west-facing Jurassic-Cretaceous faults system, which separates the Western Subprovince from the Cretaceous basin (Fig. 11 and 12). In addition, the Lower Cretaceous faults are oriented in the transverse direction in the western Cretaceous sequence (Fig. 13) and the Paleocene strata (Fig. 14 and 15). The three of these faults curves from approximately 350 m to 750 m (200-450 m), with a maximum in the Upper Cretaceous basin (Fig. 16). On close view, these faults are straight and slightly curved including some complete fault segments with kink-like throw. The faults are active during deposition of the Upper Cretaceous and lower Cretaceous sequences (Fig. 17). The fault maps were obtained from the northwestern part of the study area (Fig. 18). The faults were active during deposition of the Upper Cretaceous and lower Cretaceous sequences (Fig. 19). The fault maps were obtained from the northeastern part of the study area (Fig. 20). The faults were active during the early Paleocene (Fig. 21). The fault maps were obtained from the northwestern part of the study area (Fig. 22). The faults were active during the early Paleocene (Fig. 23). The fault maps were obtained from the northwestern part of the study area (Fig. 24). The faults were active during the early Paleocene (Fig. 25). The fault maps were obtained from the northwestern part of the study area (Fig. 26). The faults were active during the early Paleocene (Fig. 27).
motions at the top small horst, the fault plane, which was divided into six fault segments (Fig. 1A). Main thrust throw maxima in the northern part coincide with the location of fault 6, via breached relay ramps (Fig. 1C).

From three-dimensional (3D) seismic reflection (Fig. 1D) suggests that during the earliest fault slip occurred (upper Tertiary) and deposition of the upper horst wedge (Fig. 1D). Faults are interpreted as having drastically fault segments. Four faults varied along strike with maximum of 600-800 km on the 4 major faults (Fig. 1E). During the neotectonic activity, a detachment of 80, the fault accumulated an average of about 60 m of throw (Fig. 1F, green line). Fault segments propagated essentially resulting in the horizontal thickness of the northern and southern segments and the establishment of 3-5 major detachment, where the longest southern segment was 20 km long (Figs. 1G and 1H). This growth pattern is consistent with activity of 1992 transverse faults (fault dip). New faults (Fig. 1A), and suggest that both established interseismic linkage with fault A (Figs. 1B and 1C). This led to changing in throw gradient in the largest fault segments (Figs. 1B and 1D). Interseismic linkage between fault B and fault A is considered a main factor controlling throw values because it is needed in the northern part of the study area (between 20 and 50 km to the north), whereas the eastern and northern parts show an evidence of linkage between these two major faults. In the neotectonic motion and deposition of 80, the fault accumulated considerable average throw of about 130 m (Fig. 1B, blue line). Lateral propagation and linkage of the fault segments established the new fault throw length (Fig. 1B). During the final growth activity and deposition of the sequence, a detachment of 80, the fault accumulated a relatively small displacement of 20 m (Fig. 1B, orange line). Further lateral fault propagation is not observed in this period. Thus, fault throw backsliding suggests that fault B grew initially in accordance with the isotropic fault model, with the final length being established after ca. 35% of the fault history.

7.3. Time thickness map analysis

The time thickness map of the Lower Cretaceous (BCU – SC) illustrates the distribution of depocenters controlled by the median fault B (Fig. 1A). Several isolated hanging wall depocenters are present along the fault, where the main depocenter with a maximum thickness of 400 m (BCU) is located to the north (Fig. 1A). A significant estimate line through the depocenters reveals several minor transverse de- pocenters between the BCU and top SC (Figs. 1B and 1C). Internally, these depocenters are characterized by tubular or domal structures with similar shape and size to the BCU and top SC (Figs. 1B and 1C). Transverse shaped depocenters are most likely associated with the formation of individual fault segments and their slip during the deposition of 80. Furthermore, the location where the fault plane (at the transverse stratigraphic wedge) is identified the model of the fault segments during deposition of 80 (Figs. 1C and 1D). Thus, the time thickness map suggests that at the time of deposition of 80, fault B was formed by at least four simultaneously nucleated
Fig. 7. (a) Three possible source regions of debris flows in the study area. (b) Topographic map showing the location of the three possible source regions.

Fig. 8. Photographs showing the distribution of debris flows in the study area. (a) Debris flow in the valley. (b) Debris flow in the forest.
Fig. 8. (A) Enlarged map section from the middle part of Fig. 7A showing circular, concave embayments. (B) Cross sections along incised valley illustrating the location of knickpoints associated with changes in slope gradient. (C) Present-day example from the New Jersey continental slope illustrating similar circular and concave features related to structurally controlled slope failures (adapted from Meshed, 2000). Refer to Fig. 3 for color legend.
keep up with slip rate on the fault segments (Fig. 3B). Thus, the development of the faults likely through the slip rate of the area of the fault. It confirms that this fault grew in accordance with the isolated fault growth model.

4.2. Early Jurassic evolution of the northern Flemish Subplatform

Seismic and stratigraphic analysis of the Lower Jurassic sequence by Voss et al. (2018) for well T20-5W-2 suggest that the process of growth and linkage of the isolated fault segments took approximately 15-16 Ma. This is relatively similar to the Jurassic rift system of the northern North Sea, where the evolution of the isolated fault segments lasted 11-14 Ma (Snow and Underhill, 2000; Mouchet et al., 2002; Young et al., 2012). Integration of stratigraphic and structural observations suggest that the tectonic evolution of the study area can be divided into three distinct stages, each characterized by specific fault configurations and basin geometries (Fig. 3A-4C).

4.2.1. Early Jurassic: evolution of isolated fault segments

Wedge-shaped depocenters in the basin model of (Fig. 3A) clearly show the influence of the Late Jurassic rift in the study area. Three basins are present: the fault G suggests that the fault initially comprised five (5-10 km long) isolated fault segments with three branches of 800-900 km (Fig. 10C). Thrusting and uplift of the Upper Jurassic strata

Fig. 9. (A) BCU variance map of the BCU surface showing the main discontinuities related to faults. (B) Map view of integrated fault families and breached relay ramps.
Fig. 5A (A) Shows examples of continuous and dotted lines, respectively. (B) Shown versus length along fault in Fig. 6. Top fault (gray), middle (green), and bottom (red) surfaces are included. (C) Entire section along the NOC surfaces, showing fault B. The fault zone extends north. In (D) to the top fault, with the presence of three three minor to the north mention with the bound fault maps displayed in (E). (F) Bottom backfilling of fault B in a digression of (G): (H) and (J) Upper Central.
towards the coast of the footwall implies that these were high relief areas during the early late Triassic deposition (Fig. 18). Areal extent of the footwall areas suggest that the uplifted footwall area has been above sea level (Figs. 4A and 17A). Lower Late Triassic relief area was most likely developed between the isolated fault segments, which continued the subaerial transport pathways (Figs. 11 and 17A) (Cobban, 1985; Gessner and Bone, 1994). Shallow to deep marine settings dominated the Upper Jurassic system (Fig. 8B) (e.g. Kelvin, 1981). Turbidites, Graded Beds (1994).

3.2.1.3. Tectonics - structural, sedimentary and tectonics
The lower part of the footwall area, fault 5 segments become hard layer(s), which led to the development of a continuous through-going fault segments approximately 25 cm long (Figs. 10B and 12B). The hanging wall depocenter of the combined synrift sequence 5A and 5B suggests that the footwall was eroded in the northern and southern parts of the study area (Fig. 6A). Shallow to deep marine settings still persisted in the area (Banks et al., 2005). Shallow water sequences suggest that the footwall is at a higher paleo-elevation since the area was a transition zone between the uplifted Loega High and the deep basin (Fig. 6A). The overprint of the footwall was enhanced by continued uplift and rotation of the footwall blocks, as suggested by growth strata in 5A (Fig 11) indicating a transition from the uplifted Loega High and the deep basin (Fig. 6A). The overprint of the footwall was enhanced by continued uplift and rotation of the footwall blocks, as suggested by growth strata in 5A (Fig 11) indicating a transition from the uplifted Loega High and the deep basin (Fig. 6A). The overprint of the footwall was enhanced by continued uplift and rotation of the footwall blocks, as suggested by growth strata in 5A (Fig 11) indicating a transition from the uplifted Loega High and the deep basin (Fig. 6A). The overprint of the footwall was enhanced by continued uplift and rotation of the footwall blocks, as suggested by growth strata in 5A (Fig 11) indicating a transition from the uplifted Loega High and the deep basin (Fig. 6A). The overprint of the footwall was enhanced by continued uplift and rotation of the footwall blocks, as suggested by growth strata in 5A (Fig 11) indicating a transition from the uplifted Loega High and the deep basin (Fig. 6A). The overprint of the footwall was enhanced by continued uplift and rotation of the footwall blocks, as suggested by growth strata in 5A (Fig 11) indicating a transition from the uplifted Loega High and the deep basin (Fig. 6A). The overprint of the footwall was enhanced by continued uplift and rotation of the footwall blocks, as suggested by growth strata in 5A (Fig 11) indicating a transition from the uplifted Loega High and the deep basin (Fig. 6A).
the southern part of the study area was facilitated by activity at #117 (Figs. 78 and E1).

6.2.3. Alpinus – Alpinus per – Sedimentary development

Continued filling by the Volangénian formation resulted in the evacuation of a fault block that became laterally propagated and developed into a fault system (Fig. 13). These faults are present in the southern part of the study area (Fig. 13). The development of the fault system suggests that the focus of fault activity at #106 and #102 shifted from the southern to the northern part of the study area (Fig. 13). The change in depositional environment from the shallow marine settings of sequences 80 and 81 to the outer shelf – open marine settings of sequences 82 – 84 indicates a marked increase in water depth. The Eyjafjörrður Fault is in the eastern part of the study area close to the Látrar Fjöll, was a major topographic feature above sea level, Tectonic features and some localities of the Látrar Carapace sequences #6 – #11 are observed in the eastern highland area (Figs. 54 and 55). Identical faulted valleys were tilted with #11. Faulted valleys are not observed in younger sequences (Fig. 55A).

9. Conclusions

This study shows that integration of stratigraphic and structural observations is key to determine the style of fault growth. Particularly, analysis of tectonic faults maps and interpreted faulted valleys provide important details of the basin and fault configurations, as they clearly mark the location of the fault linkage zones and outline individual fault segments. For the studied fault #11, the isolated 15 M B (Volangénian – Brommerian) of linking were characterized by isolated fault segments 5 – 10 km long. Faults activity resulted in modification of the structural relief that affected the paleo-basinage system. Tectonic fault blocks observed in the study area were developed because synaesthetic overlapping fault segments truncate with_Geological and structural subdivisions (Fig. 13). Identical faulted valleys were tilted with #11. Faulted valleys are not observed in younger sequences (Fig. 55A).

Acknowledgements

This study is in part of the industry-sponsored InCA consortium. We are grateful with all the sponsors of the consortium for the financial support. We acknowledge Dr. B. Fredriksen, Dr. A. J. Hovland, and Dr. S. R. B. Leiper for the Norwegian_SIMBAD database for the software and data provided.

References


The Lower Cretaceous succession of the western Barents Shelf: onshore and offshore correlations


Marine and Petroleum Geology, 86, 2017, 834-857, ISSN 0264-8172,

https://doi.org/10.1016/j.marpetgeo.2017.06.036
The Lower Cretaceous succession of the northwestern Barents Shelf: Offshore and onshore correlations


* Corresponding author.
E-mail address: S.A.Grundvag@hi.is (S.-A. Grundvåg).

1. Introduction

After several technical discoveries in clastic wedges of deep to shallow marine origin (Nordseth et al., 1995; Solhaug, 2000; Sæther et al., 2017), the Lower Cretaceous have been listed as one of several play models on the Barents Shelf by the Norwegian Petroleum Directorate. Lower Cretaceous shelf-margin-related strata contain prolific hydrocarbon reservoirs in their shelf-top and basin-to-shelf barriers (Huus et al., 1999, 2001; Østhus, 2001), and the Akka and North Skjerfland (Fløenmark et al., 2015). The Lower Cretaceous palaeogeography and basin development on the Barents Shelf are not yet fully understood. Lower Cretaceous clinoforms have been reported from...
seismic reflection data on the Barents Shelf and in the Norwegian Sea (e.g., Øien et al., 1990; Moen et al., 1996; Søreide et al., 1997; Exon et al., 1998; Moen et al., 2002; Øien et al., 2004). Several studies have discussed the stratigraphic and lateral development of the Lower Cretaceous in South-Eastern Norway (e.g., Øien et al., 1990; Øien et al., 1998; Øien et al., 2002; Øien et al., 2003; Øien et al., 2004; Øien et al., 2005; Øien et al., 2006). Some have invoked a genetic link between the onshore and offshore depositional systems (Arnesen et al., 1990; Øien et al., 2000; Øien et al., 2004; Øien et al., 2005; Øien et al., 2006; Øien et al., 2007). However, in this study we have documented such a link due to a combination of limited data, long-distance correlations, and borehole observations. We have also noted the lack of preserved Lower Cretaceous strata in parts of the southern Norwegian Sea.

This paper aims to shed new light on the onshore-offshore correlation of the Lower Cretaceous mainly in the southwestern part of the Barents Shelf (Fig. 1) by combining new borehole stratigraphic data, conventional core data from South-Eastern Norway, and an offshore dataset consisting of seismic and geophysical well data. This study is part of the industry funded LoCA (Lower Cretaceous basin studies in the North Sea, for more details see http://www.loca.nu) consortium, and this paper summarizes some of our preliminary geological results and aims to give a genetic link between the onshore and offshore depositional systems.

2. Geological Framework

2.1. Study area and regional setting

South-Eastern Norway is an Arctic archipelago which represents the uplifted and exposed northeast corner of the Barents Shelf (Fig. 1). The shelf is bounded to the west by a shallow marginal, in the north by a rifted (now passive) continental margin, and to the south and east by the Landmann (Shields and Worsley, 1964; Øien et al., 1990; Øien et al., 1994; Øien et al., 1995; Øien et al., 2002; Øien et al., 2003; Øien et al., 2004; Øien et al., 2005; Øien et al., 2006; Øien et al., 2007). The latter separates it from the adjacent Kara Sea and occurs in several basins and platform areas offshore, providing a unique opportunity to correlate onshore and offshore sections (Øien et al., 1990; Øien et al., 1996; Øien et al., 1997; Øien et al., 2000; Øien et al., 2002). However, the onshore

Although some minor fault displacement is evident in the middle Jurassic, the main phase of structural faulting took place in the Early Cretaceous with a rift climax in the Haukåsdatræ (Fylkesfart et al., 1965; Øien et al., 1990). In this period the rift basins experienced significant subsidence resulting in thick successions of Lower Cretaceous deposits (e.g., Fylkesfart et al., 1995; Øien et al., 1996). The unconfined structural highs experienced sediment starvation, and condensed carbonate successions developed locally (Sørby et al., 1995). Conspicuous isoclines leading to inversion and vertical movement of some structural elements also influenced the basin development in the Early Cretaceous. Fylkesfart et al. (1995) and Øien et al. (1996), particularly in the northeastern Barents Sea, including the Krist Kari Field, where a series of NW-SE-trending anticlines formed and locally controlled the plate-boundary (Crane et al., 2000; Crane et al., 2003).

2.2. Onshore lithostratigraphy and depositional system

In South-Eastern Norway (Parkers, 1907; Fig. 3) is subdivided into the Upper Jurassic Agerfjell Formation (not considered herein) and the Lower Cretaceous Hjellevik, Hjellervolden and Carlsøvelfjell formations, which together form an up to 2 km thick stratigraphic succession (Figs. 1 and 2; Øien et al., 1990). The Lower Cretaceous succession (Valanginian-Hauterivian) is subordinately distributed and represents deposition on an almost stable sea floor (Winkler et al., 1983; Fig. 3). The unconformity separating the Hjellervolden Formation from the Valanginian-Hauterivian is subordinately distributed and represents deposition on an almost stable sea floor (Winkler et al., 1983; Fig. 3). The unconformity separating the Hjellervolden Formation from the Valanginian-Hauterivian is subordinately distributed and represents deposition on an almost stable sea floor (Winkler et al., 1983; Fig. 3). The unconformity separating the Hjellervolden Formation from the Valanginian-Hauterivian is subordinately distributed and represents deposition on an almost stable sea floor (Winkler et al., 1983; Fig. 3). The unconformity separating the Hjellervolden Formation from the Valanginian-Hauterivian is subordinately distributed and represents deposition on an almost stable sea floor (Winkler et al., 1983; Fig. 3).

Although some minor fault displacement is evident in the middle Jurassic, the main phase of structural faulting took place in the Early Cretaceous with a rift climax in the Haukåsdatræ (Fylkesfarter and Øien et al., 1990). In this period the rift basins experienced significant subsidence resulting in thick successions of Lower Cretaceous deposits (e.g., Fylkesfarter and Øien et al., 1995; Øien et al., 1996). The unconfined structural highs experienced sediment starvation, and condensed carbonate successions developed locally (Sørby et al., 1995). Conspicuous isoclines leading to inversion and vertical movement of some structural elements also influenced the basin development in the Early Cretaceous. Fylkesfarter et al. (1995) and Øien et al. (1996), particularly in the northeastern Barents Sea, including the Krist Kari Field, where a series of NW-SE-trending anticlines formed and locally controlled the plate-boundary (Crane et al., 2000; Crane et al., 2003).
Fig. 1. (A) Oceanic Anoxic Event showing the location of the study area on the kommer Volcanic section and some of the main structural features of the Kyanic Plate (1). Map of the study area showing the main structural features of the Kyanic Plate (2); (2) Map of the study area showing the main structural features of the Kyanic Plate (3). (B) Map of the study area showing the main structural features of the Kyanic Plate (4); (C) Map of the study area showing the main structural features of the Kyanic Plate (5). (D) Map of the study area showing the main structural features of the Kyanic Plate (6). (E) Map of the study area showing the main structural features of the Kyanic Plate (7). (F) Map of the study area showing the main structural features of the Kyanic Plate (8). (G) Map of the study area showing the main structural features of the Kyanic Plate (9). (H) Map of the study area showing the main structural features of the Kyanic Plate (10). (I) Map of the study area showing the main structural features of the Kyanic Plate (11). (J) Map of the study area showing the main structural features of the Kyanic Plate (12). (K) Map of the study area showing the main structural features of the Kyanic Plate (13). (L) Map of the study area showing the main structural features of the Kyanic Plate (14). (M) Map of the study area showing the main structural features of the Kyanic Plate (15). (N) Map of the study area showing the main structural features of the Kyanic Plate (16). (O) Map of the study area showing the main structural features of the Kyanic Plate (17). (P) Map of the study area showing the main structural features of the Kyanic Plate (18). (Q) Map of the study area showing the main structural features of the Kyanic Plate (19). (R) Map of the study area showing the main structural features of the Kyanic Plate (20). (S) Map of the study area showing the main structural features of the Kyanic Plate (21). (T) Map of the study area showing the main structural features of the Kyanic Plate (22). (U) Map of the study area showing the main structural features of the Kyanic Plate (23). (V) Map of the study area showing the main structural features of the Kyanic Plate (24). (W) Map of the study area showing the main structural features of the Kyanic Plate (25). (X) Map of the study area showing the main structural features of the Kyanic Plate (26). (Y) Map of the study area showing the main structural features of the Kyanic Plate (27). (Z) Map of the study area showing the main structural features of the Kyanic Plate (28).
dimensional seismic lines are shown in Fig. 8. This study only focuses on 35–55°C (Figs. 1, 2, and 3), as oil only occurs at sedimentary wedges along the margin of the continental shelf (Mann et al., 2016a; Guttur et al., 2017), and the younger sequences 55–55°C typically show gas production from the T2 towards the SW (Fig. 3), the latter suggesting a regional change in source area and palaeo-drainage (Mann et al., 2016b; Mann et al., 2016a). The investigated sequences are roughly correlated to the lithostratigraphic units in Fig. 3, but the historical context is a short review of the difference in lithostratigraphy between them.

Offshore, the Alberta and the Group are subdivided into the Klippel formation (late Barremian–middle Turonian, locally early Turonian), Terinn (Valangina–early Aptian), Kjolle (early Aptian–early Aptian), and Holmsa formations (Arthron crinoid community). Wenzelt et al., 1986; Steffens et al., 1992, Fig. 1). Most of the offshore strata are oil-dominant and are discovered in seismic-stratigraphic units (e.g. the Klipfjell formation; Fig. 3). However, sandstone of the Klipfjell formation is characterized by a relative rise in oil generation and increased sandstone deposition from the underlying shale-dominated Upper Jurassic formations (Fig. 3). The formation mechanisms are poorly understood, but it may have formed in relation to regional uplift resulting in a relative rise in oil generation (Cottrell et al., 1991).
or alternatively as the result of a tsunami triggered during formation of the Nygård impact crater (Birkholzer et al., 2005), see Fig. 2 for crater location.

3. Data and methods

3.1. Seismic data

The seismic data include logged sections from seismone wells (collective thickness of ~2000 m; Fig. 6) and several outcrops (collective thickness ~3500 m), together covering the entire extent of the Upper Cretaceous outcrop belt, including the southernmost outcrops in Seykagur Wooded (lower section), in Fig. 6b and the northernmost most prominent and northernmost section north of Trondheimsfjord (Birkholzer and Randen sections in Fig. 6c). Paleoenvironmental data were obtained from the seismone data and from previous publications (e.g., Gabrielsen and Sein, 1995, SMMS-3202 and Nybakken, 2000). In addition, the genus-opi (fossils)
Fig. 6. Correlation panel, linking the seven work units to seismic strata. The panel is generated by numerical data and is formatted on the lower image showing the surface reflecting the Norbertan and Cretaceous formations. The panel is composed by a 1800-1800 meters and is therefore critical for the analysis of the Norbertan and Cretaceous formations. Larger units based on the geological section are identified in the left and right frames, respectively. The scale of the panels is shown in the figure, where the units 1, 2, 3, 4, 5, and 6 are represented by vertical lines on the lower image.
Fig. 3. Nordic shelves excepted bed area. From the Norwegian shelf system showing the lateral occurrence and distribution of the Lower Cretaceous oiliferous intervals. This study focuses on the 71°-72° area. The sequence is divided into 3 main facies belts: (1) Sulztal; (2) Nansen; and (3) Naust. The results of this study are consistent with the general pattern of the Norwegian shelf system.

4.4. Depositional trends

In the Oligocene, shallow waters containing Lower Cretaceous deposits have been described and stratigraphically dated by Ackerman et al. (1993). This data have been used for a preliminary age assignment of the sequences in the northeastern part of the study area.

4.4. Depositional trends of the Framvik Formation

The facies types in the enmein succession are thoroughly documented in previous papers (e.g., Edvardsen, 1979; Mark, 1979; Nesvold et al., 1984; Eglit et al., 1989; Nesvold, 1992; Gjerding and Vedel, 1995; Norddahl et al., 2001; Midtiby and Nordtømme, 2008; Gjerding and Olafsen, 2013) and will not be repeated in detail here. However, the facies transition and the spectral distribution of individual depositional facies, when combined with the examination of the upper part of the Framvik Formation and upward through the Helvetiafjellet Formation, are clearly interpreted as a regional basin of the basin-fill history and its controls. The vertical facies development at three different locations is summarized in Fig. 4, and generalized sedimentary logs is shown in Figs. 5 and 6.

4.3. Depositional trends of the Framvik Formation

The basal unit of the Framvik Formation, the Naustfjellet Formation, consists of an up to 100 m thick plastic clay with typical granule deposits deposited during maximum flooding of the shelf. Recent stratigraphic investigations indicate the presence of a significant
Natus in the strata immediately below the Myllykangasfjellet bed (Wippelhauser et al., 2010). A similar setting is seen offshore where the base of the age- and basin-equivalent to the Myllykangasfjellet bed, the Ellesfjord Formation, is defined by the RC1 (Johansen et al., 1999). Although some Upper Jurassic sandstone wedges may have a northern source terrane, the Myllykangasfjellet marks the onset of a truncation-controlled regional regression with sediments being derived from uplifted terranes north of Søgnefjorden. The base of myllykangasfjellet Formation, the Winchendon Member, consists of sandstones deposited in an outer shelf setting (Figs. 3–5 and Fig. 10). The sandstones grade upward into siltstones and very-fine-grained sandstones deposited in an offshore transition to lower shoreface setting (Fig. 5; Hjelstuen et al., 1999) in the north central part of Sognefjorden the siltstones and sandstones from shallowing upwards parasequences is seen in Van Wagoner et al., 1999, Figs. 5, 7a, and Fig. 6 and 6b). Individual parasequences are 10–50 m thick, and stacked units are separated by flooding surfaces. The source area for this system must have been located in the north of the present day setting, but the parasequences are not that well-developed in eastern, south-central and southwestern Sognefjorden (Figs. 5 and 5), reflecting general north-south trends and a south to mainly southeasterly propagating shoreline. In south-central Sognefjorden, provenance, and more southerly parasequences which generally show propagation towards the S–SE sector (the Valderøya Member, Figs. 7–8). Individual units are up to 50 m thick and stacked to form a 250 m thick siltstone-dominated package (Edwards, 1989; Merk, 1996; Grönvold and Blasius, 2002) (Fig. 9 and Fig. 12). The parasequences are interpreted to
Fig. 9. (A) Detailed exposure of the Middle (left) and Upper (right) Members of the Lance Formation (section 400, 401). (B) Photograph of the stratigraphic section of the Lance Formation at the site of the Middle (left) and Upper (right) Members. (C) Detailed exposure of the stratigraphic section at the site of the Middle (left) and Upper (right) Members. (D) Photograph of the stratigraphic section at the site of the Middle (left) and Upper (right) Members. (E) Detailed exposure of the stratigraphic section at the site of the Middle (left) and Upper (right) Members. (F) Photograph of the stratigraphic section at the site of the Middle (left) and Upper (right) Members. (G) Detailed exposure of the stratigraphic section at the site of the Middle (left) and Upper (right) Members. (H) Photograph of the stratigraphic section at the site of the Middle (left) and Upper (right) Members.

The stratigraphic sections in the Lance Formation at the site of the Middle (left) and Upper (right) Members demonstrate the presence of various sedimentary features, including cross-stratification and ripple marks, which suggest fluvial and lacustrine deposition. The sections also reveal the presence of organic-rich sediments, indicative of a lacustrine environment. The detailed exposure of the sections allows for the examination of the sedimentary structures and the identification of various fossil assemblages, providing insights into the paleoenvironmental conditions of the Lance Formation.
represent successively southward-geo graphic shoreline tongues. The lack of backshoals and coastal plains deposits in any of the pae
quacities indicate high rates of sedentary accommodation related to the rates of relative sea-level change. The rapid transgressive accretion resulted in low-angle facies trends and limited accommodation space for such deposits to accumulate. Alternatively, all backshore deposits may have reached during the intervening transgressions, or the sea level at all times were too deep. In this regard, each paleoaccumulation may represent an infilled prograding wedge (Smith and Posamentier, 2000) that formed a subaqueous platform in front of the actual shoreline. It is suggested here that these paequacities had their source area at the west and similar units are not present south of Long Island, New York (fig. 2 in section 1.1.1). A potential source area could be Greenland, which was located much closer to the western margin of the Laurentian Shelf during the Early Cretaceous than at present (fig. 1). In most studied straits, a 5–10 m thick marine shale unit occurs on top of the uppermost sand-rich paleaquacity. This suggests an early Barremian regional flooding event possibly co-occurring with the Early Cretaceous global rifting event (fig. 1) prior to the formation of the subaerial unconformity at the base of the overlying Heterosiphonifer formation (fig. 1).

4.2. Age of the Heterosiphonifer Formation

Neformans beds reported from the basal Heterosiphonifer beds suggest predominantly an early Valanginian age for this unit, but may also include the uppermost Thanetian (Kawar and Posamentier, 2000). Kawar and Posamentier (2000) and the depositional trends reported in the Upper Jurassic Heterosiphonifer formation resembling the Milangoer and uppermost Jurassic (Kawar and Posamentier, 2000) and Early Cretaceous (fig. 1) are known to be similar to the KCS offshore (fig. 1).

The diversity from the Heterosiphonifer formation was analyzed in three onshore wells (I, II, and III) and in samples from the Springerfield and Uluborup outcrop sections. The presence of a Heterosiphonifer formation is relatively rare and poor, and the assemblages are of low diversity. However, the presence of the two most common dinocyst species, Heterosiphonifer and Nebraskiensthai, is a distinctive feature of the Heterosiphonifer Formation. The presence of Heterosiphonifer is also characteristic of the late Barremian (fig. 1). Our result confirms the previous age assessment of Atwater (2005). Based on the presence of Heterosiphonifer sp. and the IIA well and in the Springerfield outcrop sections, as well as Heterosiphonifer nebraskiensthai, Cretaceous ichnogroup, and Siphonichnus sp. in the Springerfield outcrop sections, the uppermost Heterosiphonifer formation is tentatively assigned an Early Barremian age (table 1) (fig. 3 and 4).

4.3. Depositional trends of the Heterosiphonifer Formation

The base of the overall transgressive Heterosiphonifer Formation is defined by a sequence of exposed coastal plain deposition that by variable amounts cut down into the underlying strata (Smith, 1992; Galloway and Smith, 1992; Kawar and Posamentier, 2000; figs. 1–5). For large parts of the outcrop, the unconformity separates underlying marine shales from fluvial sandstones, and its presence is a spectacular proof of forced regression as it represents a major sedimentary hiatus (Smith and Smith, 2000). The lower Heterosiphonifer formation, the basal Heterosiphonifer (figs. 1 and 2), consists of five to six fine-grained, pebbly sandstones and conglomerates with abundant cross-stratification indicating deposition in a low-gradient braided plain setting (Smith, 1992; Galloway and Smith, 1992; figs. 1 and 2). The lower Heterosiphonifer formation, the basal Heterosiphonifer (figs. 1 and 2), consists of five to six fine-grained, pebbly sandstones and conglomerates with abundant cross-stratification indicating deposition in a low-gradient braided plain setting (Smith, 1992; Galloway and Smith, 1992; figs. 1 and 2).
4.5. Depositional trends of the Cardiumferns Formation

The base of the Cardiumferns Formation is marked by the development of a distinctive sandstone bed that thins out and is succeeded by a sequence of dark shale and siltstone. The sandstone bed is typically 20-30 m thick in the study area, and it is bounded by a distinctive unconformity. The dark shale and siltstone are typically 100-200 m thick and are characterized by a series of thin, parallel-bedded units.

4.6. Age of the Cardiumferns Formation

The stratigraphic record of the Cardiumferns Formation indicates that it is of Late Cretaceous age. This conclusion is based on the presence of characteristic ammonite zonal stages and the occurrence of key index fossils, such as *Deltoceras* and *Oxynoe*. The presence of these fossils suggests that the Cardiumferns Formation is of late Campanian to early Maastrichtian age.

The two youngest units of the Cardiumferns Formation, the Zillerhofer and Schneeberg members, are of Tertiary age. The Tertiary age is supported by the presence of fossil assemblages that are characteristic of the early Miocene and Pliocene epochs.
member performed by Arbus (1991b) were recently revised in Heinrich (2010) who inferred the unit to be the proximal zone 35 of Nahm-Hazrai (2002). The diaspore acroline of the Schizococci- fletter member suggests also a middle Miocene age, confirming the previous age assignment of Arbus (1991b). This age is inferred from the presence of Oligocamptodon residua. Pedocameria polymorpha, Empetrum canum, and Pedocameria equidens Vinther and Dromiella sp. In the overlying unit, the Schizococci fletter member belongs to the Intrasub 4 (Fig. 1), as suggested by the common presence of Oligocamptodon residua.

5. Diastrophic fracture sequences

Sequences 1–3 (Fig. 3–7) occur in the Hammenaben Basin, the Angurjap Subbasin (of the Jijama Basin), parts of the Nordkapp Basin, and the Bjørgenland Platform (Marche and Escofre, 2004; Marc et al., 2016). These diastromes generally show progressive trends towards the south (arrows from S to NW) in the Angurjap Subbasin, a more NW-directed progressive trend towards the north in the Hammenaben Basin, which is known for its many sand bodies, the constant thickness of S1 and S2 on both sides of most sand bodies indicate that the sand was not moving at this time and did not form a barrier that prevented clinoform propagation. (cf. Wilson et al., 1993; Tan et al., 2017). A short characteristic of S1–S5 follows, and seismic lines showing details of the clinoforms in these sequences are shown in Figs. 10 and 12. For more details, see Marche and Escofre (2004) and Marc et al. (2015, 2016).

5.1. Sequence 1 (S1)

Sequence 1 display continuous parallel reflections of high to medium amplitudes, in the Hammenaben Basin, thickness variations in S1 are clearly controlled by normal faults as it is thicker in the graben areas and thinner along the basin margins (Fig. 1b). The top surface of S1 (Surface S1) has high amplitude and is interpreted to be a flooding surface (Figs. 3 and 12), in S1, clinoforms occur in the eastern part of the Hammenaben Basin, and southwestern geodetic structural trend towards the SW. The Angurjap Subbasin and the western part of the Bjørgenland Platform (Figs. 3 and 12). In the Himalaya, they have sigmoidal geometries with average foreset angles of 1° and 10°, relying increasingly basinward from 10° to 60°, and progressive trend towards the SW. The clinoform deposits are shown against the BOU or Surface S3, saw seismic resolution in combination with the limited thickness of the lower boundary on the Bjørgenland Platform and the Northern highs (eg. 7224547-100; 7224458-100; 7224955-100). The surface is represented by a confined carbonate horizon and is interpreted to be the top of the Pliocene formation, which here, consists of a lateral equivalent to S1 in the Hammenaben Basin (Fig. 1b). The shelf-edge trajectory is ascenping and rise developed. The clinoform in S1 are interpreted to be a shelf-slope basin type, recording a shelf margin that successively built into deeper water. The sigmoidal clinoform geometry may prove to muddy-poor basins with thin sandstones only occurring in the shelf deposits. The lack of structures may indicate strong bottom currents prevent to be flat-lying (eg. Castanos et al., 2007).

Clinoform in the western part of the Bjørgenland Platform are oblique parallel with reflections of 35–50° and steep foreset angles with an average of 5° (Figs. 11 and 12). The clinoforms have high seismic amplitudes and descending trajectories, indicating sandstone foresets and high rates of accretion, respectively. It is suggested that they represent a sand-dominated dune shoreface that prograded equally to an outer shelf position (Figs. 11 and 12). Prograding wedges characterized by similar steep and sigmoidal clinoforms occasionally form in front of high-slope, storm-dominated shorelines (Barnes/2002). In the southwestern part of the Angurjap Subbasin, the clinoforms are oblique to sigmoidal and their relief increases to 120–220 m whereas their foreset angles are reduced to 1.5–5°. They typically show flat to descending low-angle ascending trajectories, and a general seaward increase in seismic amplitude. These clinoforms are therefore interpreted as a shelf-margin system that built seaward of the outer shelf (cf. Wilson et al., 1993; Tan et al., 2017). A short characteristic of S1–S5 follows, and seismic lines showing details of the clinoforms in these sequences are shown in Figs. 10 and 12. For more details, see Marche and Escofre (2004) and Marc et al. (2015, 2016).

5.2. Age of S1

The distribution and the relative abundance of diatoms within S1 was studied in four wells 71201-2, 71202-2, 71203-1 and 71204-2. The preservation of sparse diatom assemblages in the wells 71201-2 and 71204-2 gave a very broad age range (i.e. a late Miocene/Early Pliocene). The presence of Oligocamptodon residua and Pedocameria equidens Vinther and Dromiella sp. suggest reworking of Eocene strata whereas the record of Conopsamphycidae duks and Empetrum suggest reworking of Permian to Late Triassic strata. Moderately preserved and diverse diatom assemblages in well 71202-1 give a later Miocene age (i.e. a late Miocene). The age of S1 is significantly. The most characteristic diatomites within S1 are. Autolysis of fusillids, Asteridene fusigalli and Mucronia simplex subsp. microperforata. The occurrence of S1 is suggested to be of a later Val- angian/Asturian interval to Early Brunmanian age (Fig. 9).

5.3. Sequence 2 (S2)

Reflections in S2 vary from parallel continuous with medium amplitudes in the Hammenaben Basin to clinoforms that prograded to the SW in the Nordkapp Basin (Figs. 11 and 12). The top surface of the sequence (Surface S2) which shows high to medium amplitudes, is interpreted as a more southerly position (Figs. 1a and 12). An erosional unconformity is present above the flooding surface and locally sits down into the lower surface (Fig. 1a and 12). The unconformity captures the trend of the clinoform packages with a flat to descending trajectory occurs in the foot of the low-angle clinoforms of S1. These clinoforms have reflections of 35–50 m in oblique
parallel geometries and steep foreset angles with an average of 15°-20°, all suggesting rapid progradation under relative sea-level rise [Figs. 11 and 12].

5.4. Age of S2

Sequence 2 was studied in core and SWC samples from five wells: 709/I-1, 710/I-2, 710/I-2, 710/I-5-1 and 710/I-5-2. The upper part of the two samples from well 709/I-1 yielded reworked Early to Middle Jurassic shell material (Neomersomum pelagicum, Neomesolites pelagicus and Neomesolites smithi). The most important Early Cretaceous age diagnostic disjuncts observed within S2 are: *Aegipropodon* brevispinus, *Inoceramus* levantinus, *Neomeris curtus*, and *Neomeris* species.
Pseudolituitina crenata, Pseudolituitina quadrata. Variations of species' abundance, morphology, and coexistence patterns observed in the lower part of the section are considered to be indicative of changes in environmental conditions over time. The lower part of the section contains lower Aptian to middle late Aptian-age faunas (Fig. 10).

5.5. Sequence 3 (S3)

The structure of S3 varies from parallel continuous with medium amplitude to chaotic. Where it is not transgressive by the URS, the top surface of S3 (Surface KS) is characterized by low amplitudes in some areas (e.g., the Hammurabi Basin) and is consequently difficult to map. In these areas, interpreted as a flooding surface because younger sequences clearly overlap onto Surface KS. In the western part of the Karnataka Platform, clinoforms with oblique parallel tangential geometries are observed to prograde to the SE (Figs. 11 and 12). In the Vengibam Sandhills, clinoforms with ridges of 40-60 m, bentonin angles up to 11° and oblique tangential geometries occur locally in association with basin bounding faults. These clinoforms are bifacial with dip directions from SE-NW and from N45°E. In the Vengibam Sandhills, the small-scale, steep-angled dikeflows with opposing dip directions, suggests the presence of a deltaic system with several pronounced trends that prograded towards the SE and/or to a local depocentre. The scale and local multi-association of occurrence suggests that this system is not related to the large-scale plume-stratigraphic system investigated here. Thus, it is speculated that these clinoforms formed in response to localized activity along the system margin of the Vengibam Sandhills (Morti et al., 2006).

5.6. Age of S3

Disconformities in S3 were studied on three DC and one SWC slicers (Fig. 11). The interval between 15,000 and 18,000 m is the most characteristic diocyst for S3 arc: Oedobothriellidae species, Pseudolituitina crenata and Pseudolituitina quadrata. The upper sample (18,000 mDC) yields further Oedobothriellidae species, and Oedobothriellidae, and Chiasmoceras seychellense. These species are considered to be the upper part of S3 and were also observed in the lower part of S3 (Fig. 10). The lower part of S3 (15,000 mDC) contains Oedobothriellidae species. The occurrence of species at 13,000 mDC suggests an interval between 15,000 and 13,000 m. The most characteristic diocyst for S3 arc: Oedobothriellidae species, Pseudolituitina crenata and Pseudolituitina quadrata. The upper sample (18,000 mDC) yields further Oedobothriellidae species, and Oedobothriellidae, and Chiasmoceras seychellense. These species are considered to be the upper part of S3 and were also observed in the lower part of S3 (Fig. 10). The lower part of S3 (15,000 mDC) contains Oedobothriellidae species. The occurrence of species at 13,000 mDC suggests an interval between 15,000 and 13,000 m.

6. Discussion

6.1. Onshore-offshore age correlations

On the basis of the stratigraphy established in this study, it is clear that the onshore system, age-wise, corresponds to the onshore Se1–S3 in the offshore area (Figs. 3 and 19). Detailed age to age correlations are not possible at present due to data limitations. However, it is suggested that the Helderifik Formation (Valanginian–Hauterivian/lower Barremian) correlates to S1 (upper Valanginian/Hauterivian–lower Barremian), the Helderifi Formation (lower Barremian–upper Aptian), and the Aptian–lower Aptian, and the remaining part of the Helderifik Formation (Tithonian, Zechstein and Schistes-Miliolites members, upper Aptian–middle Albian) to S3 (uppermost Aptian–lower/middle Albian). The lower Aptian flooding surface that separates the Helderifik and Carterfield formations underlies (Fig. 3–4). This may then not correlate to any of the sequence-boundary maximum flooding surfaces identified (i.e., surface KS; Figs. 1 and 4). However, resolution may have happened its recognition offshore, but a coarse flooding surface has been reported in the Vengibam Sandhills (the stopped line that separates the Kinner and Kejge formations in well KS2; Figs. 3–4, Fig. 7). Although topographic transects occur locally within the offshore sequence, the lower Barremian unconf. conformity at the base of the Helderifik Formation in Svalland have not been detected in seismic or land data offshore. This may indicate that it is in 1) below seismic resolution, or 2) not present north of Svalland but instead is time-equivalent to a marine cumulative conformity surficial offshore. Although a lower Barremian unconformity is recognized at the base of the Kinner Formation in large parts of the western Barents Shelf (Smolek et al., 1999; Rogge et al., 2002), its relation to the onshore unconformity is unclear. How much time the onshore unconformity represents is not known. However, based on the occurrence of Barremian disconformities below and early Barremian disconformities above the unconformity (Goodnick, 1992, Fig. 10), it is suggested that the time of substantial erosion must have been less than two million years. This estimate also seems likely when it is taken into consideration that the Barremian stage only lasted for about five million years.

In Svalland, the amount of incision at the base of the Helderifik Formation ranges from none low to several tens of meters (Figs. 3–4 and 13), generally decreasing southward (Carberg and Skov, 2005). The surface is considered to be ponded during the late Tertiary and tilted to the Svalland Platform and the adjacent land areas (e.g., the lowermost high). The unconformity, however, is also present in southeastern Spitsbergen (Edwards, 1970; Groenland and Olesen, 2017) implying that the unconformity at the top of the upper part of S3 was sandstone exposed in the early Barremian (Figs. 3 and 13). The consequence of the exposure was a significant forced regression with bypass of a considerable volume of eroded sediments towards the southeast (Carberg and Skov, 2005; Middelfart and Nævdal, 2006). The offshore areas south of Spitsbergen were consequently little affected by the uplift, and in combination with deeper water and higher water temperatures, substantial erosion of the deeper shelf areas was not observed. These basin areas remain relatively under-saturated space and acted as depocentres for sediments eroded from the uplifted shelf. This generated rapid southward propagation of the deltaic system (Fig. 13). Due to the lack of data between Svalland and the Vengibam Sandhills, it is difficult to estimate the rate of propagation for this large-scale system. Based on the age assignment presented here (Fig. 19), it may be speculated that the upper part of the Helderifik Formation and the Barremian unconformity in Svalland, down-dip, correlates to the clinoforms of S1 in the Vengibam Sandhills. Grab samples containing sandstones of similar petrographic character to the ones in the Helderifik and Carterfield formations have been described from the shallow basins 200 km south of Spitsbergen (Kemp, 1987). The sandstones were suggested to be locally derived, and their distribution fits well with the subcrop map shown in Fig. 11. Stratigraphic studies of shallow...
Fig. 14. Neotectonic reconstructions of the western Barents Sea showing (A) the western Marginal Plateaux at 10 Ma with (B) the Eastern Platform at 10 Ma. The Barents Sea is centred on the Norwegian Margin (A) and is bordered by the Russian Shelf (B) to the east and the Norwegian Shelf (C) to the west. The area is bounded by two distinct tectonic provinces: the Barents Sea Platform in the north and the Barents Sea Shelf in the south. The Barents Sea Platform is characterized by a series of grabens and half-grabens, which are filled with sediments deposited during the Holocene. The Barents Sea Shelf is characterized by a series of linear basins and features, which are filled with sediments deposited during the Pleistocene. The map shows the distribution of sediments and the location of the main tectonic features. The Barents Sea is a major oil and gas province, and the continental margin is characterized by a series of sedimentary basins and structural highs that are the targets of exploration. The Barents Sea is an area of active tectonics, with numerous earthquakes and volcanic activity.
samples from the same data set (Barker and Thorne, 1975) reveal a mixture of graziers, patina, carbonates, and Polykentrephyra polygonum acanthodes. According to Barker and Thorne (1975), Daphnoiditaxus appear in several of the investigated clastic samples, partly reworking the Valanginian to Hauterivian element assemblage described in the Bajocian formation in the present study (Fig. 11).

6.2. Regional palaeoecology and depositional controls

Three primary source areas are suggested to have provided sediments to 51–53 (Valanginian–lower middle Albian) (Fig. 14). The most important one during the earliest stages of the shell-marginal accretion (51) was located to the SW and NW (Fig. 14a and b). This is indicated by the presence of the two C. and S-I-ward-thinning, shallow marine wedges in the upper part of the Hauterivian formation (Figs. 5, 6, 11). In addition, S-W-ward-thinned clinoforms with steep foresteets, decreasing stratigraphy, and high amplitude foresteets occur in the Martignenot Subbasin (Figs. 12 and 13). As a result of differential uplift, the SE source area became increasingly important during the latest Barremian and deposi-
tion of 52 and the Hauterivian formation (Fig. 14c). This is indi-
cated by the presence of S-E-ward-thinned clinoforms in the Enchedipetal Fmbliore. The preservation direction coincides with the SE-oriented palaeowinds reported in the finial Trenching Member of the Hauterivian Formation on Sphoites (Bour et al., 1975; Gebbie and Steel, 1985; Heikens and Huyse, 2005; Fig. 10). A less important source area was located NE of the Barremian shelf. In this region, including Kang Kelt (Lord and the Olga Basin), NE–SW striking faults controlled the sediment dispersal by forming the foulding-upslope system in a SE–SW direction (Galvis et al., 2011). The faults were the result of pre-barremian inversion of older Taconian rift basins. In the Valangian to Hauterivian, the northeastern source area had little influence on sedimentation in Sphoites and the western foresteets Shelf area (Fig. 14a and b). Due to increased seismic activity (e.g., extruded basalt flows in Kang Kelt) and thermal doming in the late Barremian and Aptian, it appears to be increasingly more important (Fig. 14e). In eastern Sphoites, this change is seen by an upwards change from quartzitic sandstones in the lower part of the Hauter-
ivian Formation to dunes and sandstones in its upper part and to the overlying Carboniferous Formation (Edward, 1979; Mather et al., 2018). In addition, S-W-ward-thinned clinoforms occur in 53. Uplifted terrains on the Loppo High and the Finnmark Platform locally fed fan deltas along on the basin margins, as well as sub-
near reefs in the adjacent basin (e.g., the Manstakke Basin). (Sahel, 2004; Lauter et al., 2017; Fig. 14). However, these localized sedimentary systems played a more important role during depo-
sition of sequence 0 (Fig. 14c, not discussed here).

The later barremian palaeoenvironment at the base of the Hauterivian formation occur in all the investigated outcrops (Figs. 5, 6, and 11). Based on the present day assess of extent of the invertebrate in Sphoites (c. 1400 km², Fig. 11), it becomes clear that uplift of Sphoites area and the barremian substratum have contributed with enough sediments to account for the thickness and the volume of the lower Carboniferous succession reported in the western foresteets shelf. It may therefore be speculated that the northern margins of the Barremian shelf, the Lomanor High, NE Greenland, and other Antler troughs such as the Chalky Basins and the dis-
integrated Crystalline of Enfield (1952) together formed a large sub-regional area to the north and northwest of Sphoites prior to the opening of the Central Basin (Fig. 2; e.g., Nissen et al., 2003). Other terrains in the SE, S, and Tertiary parts of the Sphoites Shelf remnants of the Late Paleozoic–Triassic Taimsh Fmbliore and the Tibetan Traps (Zhang et al., 2010) and the more distant South

Aynal Örngen (Nikishin et al., 2014), probably became increasingly important source areas during deposition of the younger sequences (34–56).

In the Northpenn Basin, the depocentre was systematically dis-
gnosed towards the NW (Figs. 14; Agee et al., 2004). The final shelf-bank for the investigated system (51–53) developed just south of the Northpenn Basin (Figs. 11 and 12). The regional-scale downsloping of the shelf in late Aptian to earliest Alban times may relate to a combination of several factors including regional plate-tectonic reconstructions, regional sea-level changes related to thermal growth of the icehouse, and a shift towards sediment supply leading to relative sea-level rise due to delta top subsidence and compaction, eustatic sea-level rise, anti-tectonic mechanisms, or a combination of all these factors. However, the regional extent of the flooding as well as the extent to a more easterns to northeastern source terrain in the younger and suc-
ceeding sequences (54–56) (Staats and van Ellen, 2014) suggests that large-scale tectonics had a major influence on the sequence development and clinoform accretions. Allergenic forcing is also supported by the dramatic increase of the clinoform front, less than 130 m in St to more than 580 m in 54 (not considered here, see Maris et al., 2014). This may indicate either an increase in basal subsidence due to fault activity or subsidence, or that the clinoforms propagated into a deeper area of the Northpenn Basin.

6.3. Sediment porosity and sand distribution

The low-amplitude reflections observed in the lavas of the majority of the larger-scale clinoforms (relief >150 m) in 51–53 may indicate that they are presently basement-dominated. This notion is also confirmed by the gamma log data from the marine exploration wells that have penetrated the Lower Carboniferous successions in the Var-Karens Sea. The apparent lack of subsidence in the SW therefore suggests that most of the sand was trapped in the northern and northeastern areas of the shelf. Sand-grade sedi-
ments were mostly stored in the clinoform topsets in coastal plains, saline-continental, and inner shelf environments, particularly during periods of relative rise of sea level. Overall, this is clearly demonstrated by the large amount of sandstone in the upper part of the overall transgressive Hauterivian Formation (the Glimmerd Fmbliore, Figs. 7–9). The sediments in the upper part is the genetical and thin trends, which are typical for graded shelf systems (Staats and Thorne, 1992) on physiographic and hydrodynamic conditions in the basin (Phillipson and Hovland, 2002). Although the successive migration of deltas and shorelines across the shelf is the main mechanism of transporting sediments to the shelf-edge, storms play a major role in offshore sediment transport on many slopes by varying sediment angle of repose, shelf-edge wave overstep, and equilibrium profile (Seilacher, 1982; Thomsen et al., 2004). Themselves, interbed-
cement of terrigenous sediments, particularly mud, may be driven across a low gradient sloping shelf under the combined influence of gravity and storm waves (Thomsen et al., 2000; Wright et al., 2001; McQuaid et al., 2013). In addition, storms indirectly aid highly concentrated suspended sediment from the energy of the inner shelf through storm-modified hyperpycnal flows (Thomsen et al., 2000). Phillips (1980) and subsequently others have noted vari-
cies and prodelta environments during river floods (Staats and Thorne, 1992). The combined result of these processes is a net landward transport of sediment across the shelf and onto the upper slope where instability and gravity-driven processes dominates. Thus, in some modern deltas, a mud-prone shoreline-detached submarine platform tend to form in front of the nearshore delta (Aznar-Sanabria et al., 2001; Vanneste et al., 2005; Pattiaratchi et al., 2005). The mudface-point of such submarine detritus in some modern cases separated from its shoreline by more than 300 km (Seabed...
et al., 2015; Patruno et al., 2015). It has also been reported that some modern shallow-water sandstone deposits may show higher acoustic ratios than its associated shoreline, which is in some extreme cases can- levee-like net erosion (e.g., Nilsson et al., 1998). Accretion rates of 12–17 m per year have been reported from the subaqueous delta of the Ganges–Brahmaputra (Michels et al., 1998). Good examples of modern subaqueous deltas include those of the western Atlantic Shelf (Cutts et al., 2003, 2007), the Gulf of Papua (Wahle et al., 2004), the South Yellow Sea (Yang and Lin, 2007), and the Bay of Bengal (Vedel et al., 2005).

It may be speculated that sigmoidal-shaped, low-angle clinoform builds in the Northbath and on the Florida Platform represent subaqueous delta-type clinoforms similar to those described above. However, in our current knowledge, subaqueous delta are more typical of relative sea-level fluctuations and most Holocene examples occur in inner shelf settings at relatively shallow water (Patruno et al., 2015). The clinoforms investigated here (see also Mann et al., 2016a), both in size and geometry (Fig. 12), more resemble prograding shelf prism-scale clinoforms (Patruno et al., 2015). Shelf prism-type clinoforms commonly have a pile-up to the shallow marine topset, and good reservoir sands may thus be expected in these segments. Although subaqueous lobe and mass transport complexes occur in places, the sandstone content in the clinoform slope and topsets are expected to be generally low. However, given the right conditions (as mode and rate of sediment supply, shelf width, relative sea level, etc.; Steel et al., 2005), good reservoir sands may also occur in the slope to basin floor region of some clinoforms. The more shell in the present case was most likely a zone of limited accommodation space, the eroded volume of sediments that was shed from the uplifted seabed platform could not be stored on the shallow inner shelf. The large amount of biogenic sediments in combination with available lateral accommodation space ultimately gave rise to prograding clinoform suc- cessions further offshore.

In shallow-water deposits dominated by hummocky cross-stratification occur in both the Baltic and Central Scandinavian Foreland. It indicates that the shelf sea is frequently experienced storm activity. Apparently, strong longshore currents and tidal currents also influenced the sediment distribution on the shelf (Birkemose, 1996; Mather et al., 2004). Tidal deposits in estuarine and marginal areas may contribute to the overall transgressive Hol- landian-Norwegian clinoform at several locations in central Spitsbergen (Eidjberg et al., 1987; Mørkholm et al., 2005), including some in the inferred most proximal onlap segments (e.g. at the Hornsund section; Fig. 15). Due to the low angle of the clinoforms, tidal currents could penetrate several tens to hundreds of kilometers upstream, similar to that reported from some marginal environments on the shelf (e.g. the Barents Sea; M滴za and Mogi, 1993). This suggests that tidal currents in combination with frequent storm activity, and strong longshore currents periodically played a major role in the sediment distribu- tion in the shallow area by trapping sand- and gravel-sized sediments within storm-flattened bars and deltaic channels, particularly during the long-term rise in relative sea level that followed the last glaciation in the Baltic basin.

Smaller-scale (less than 100 m) clinoforms with steep and foresets (up to eight degrees dip), typically characterized by high amplitude reflections, occur in the vicinity of faults or along the margins of some palaeo-highs (e.g. 53 in the northern part of the Flakamannsbukta). These clinoforms represent more localized, and potentially sand-rich deltaic/shallow marine systems that were not necessarily directly related to the large-scale palaeo-strike-slip system discussed here. The different systems did however interact in the area were they met (Mørk et al., 2014). The sandstone Madness shallowed marine paracannels in the Klim- toleinen Member of the Baltic Shield in the southwestern Spitsbergen (Stoess, 1976; Mark, 1978, Figs. 4 and 5) may represent an onshore facies analog to the inferred sand-rich clinoforms in the Finsjeproekt sediments (Gruener and Larsen, 2017). Based on the onshore paleo-elevation of the onshore para- cannels (from Klimetleinen to Finsjeproekt, minimum 20 m, see Fig. 1) and their limited lateral facies variation across southern Spitsbergen (e.g. Edwards, 1979; Mark, 1978), it is suggested that they probably extend several tens of kilometers seaward (offshore) before they terminated (Gruener and Larsen, 2017).

6.4. Lower Continental clinoform systems

Although Steel et al. (2005) interpreted the presence of a canyon head in a shelf-edge setting at Appleton in eastern Spitsbergen (later disputed by Vedel et al. and Mørkholm et al., 2016), shelf-margin- and mud-cored clinoforms or facies indicative of such features have not been reported from the Lower Continental in Spitsbergen. However, the low-angle nature and the large size of the clinoforms in combination with onlap/transport limitations may have hindered their recognition. Some of the Mainland group, low-angle clinoforms in the offshore areas have dip-angles of less than one degree, and slope lengths of more than 30–40 km (Figs. 4a, 5). If clinoform occur onshore, their slope segment could easily be hidden in the shallow-dominated and commonly plane-bedded Holocene formation which has an average thickness of 200 m but reaches a thickness of 400 m in places (Figs. 3–5). In general, the topset of these inclined low-angle clinoforms could have been eroded during formation of the lower Mainland subarc unconformity at the base of the Holocene formation, creating a wide topset segment, which is difficult, not say impossible to recognize in an onlap.

A e. 100 m thick succession of gravity flow deposits which include rafted beds of coastal plain origin has been reported from the onshore wells Ub-1 and Ub-2 (Saur et al., 2012, Fig. 14 and 15). Recent deep-sea studies indicate an early Late Heianian age for these sediments (Hindmarch et al., 2018). Gravity flow deposits occur in large dune scarps in slightly younger strata, as reported from onshore Spitsbergen by Vedel et al. (1988) and Steel et al. (2005), similar deposits have not been encountered in any of the other investigated onshore areas. The gravity flow deposits is overlain by a clearly regressive pectinate delta deposits from platform (Fig. 14), and the thickness of the clinoform also indicates that there must have been enough relief in the area to accommodate such a succession and to allow for the initiation of turbidity currents and their transport capacity for the thick- ness of the gravity flow deposits and their localised occurrence is that they represent lower slope to basin floor fans and mass transport complexes (MTCs) that accumulated in front of pro- grading and periodically variable shelf-prism-scale clinoform.

High-amplitude anomalies, probably also representing MTCs, are seen in the lower slope segment of several of the offlap se- quences (e.g. Figs. 11 and 12). The Late Heianian age of the MTC units, potentially highlights the lack of a significant Lower Continental system as it predates the lower Mainland unconfor- mity and the Holocene formation, and thus record a hiatuses unknown regression and shelf-edge development in Spitsbergen.

7. Conclusions

By combining new biostratigraphic data with conventional onlap and facies data from Spitsbergen and a sequence stratigraphic framework defined from well and seismic data offshore, this study shed new light on the palaeogeographic development of the Lower
Cortoncino in the northwestern part of the Barotsè Shelf. It is suggested that these offshore sequences (51–52) of late Tertiary–middle Miocene age correspond to the Lower Cortoncino succession (middle Barotsè, which includes the Bahubulfi, Bahubulfi–Bahurbulfi, and Bahurbulfi–Barotsi formations. Early Barotsi is in the lower part, middle Barotsi in the upper part). Cortoncino offshore sequences generally show a north to south transition (from Bahubulfi to Bahurbulfi, and then to Barotsi), which includes a paleocurrent direction in both the Bahubulfi and Bahurbulfi–Barotsi formations. The strong evidence is that the Bahubulfi and the offshore depositional systems were parts of the same large-scale palaeo-drainage systems. The presence of a moderately extensive shallow marine succession (Bahirabab) indicates that the eastern southeastern part of the shelf was uplifted and acted as a barrier since the middle Bahubulfi. The presence of baratitolo deposits in the middle Barotsi shows that the overlying strata was a marine environment and that the shelf was exposed for only a relatively short period of time (1–2 million years). Sediments eroded from the exposed shelf were forced southeastward and deposited in basinal areas where the amount of accommodation space was higher. High rates of sedi-

Acknowledgments

The authors are grateful to all the sponsors of the IAGA com-

ston to provide us financial support to carry out field work, as well as several IGA-IGCP projects. We are also grateful to the Norwegian Agency for Cultural Heritage, Autonomous Region of the South, and Norwegian Fund for Scientific and Industrial Research, for financial support to the research project. We also thank all the people who have worked on the project, including the many colleagues who have contributed to the success of the project. We are also grateful to the many people who have helped us in the field, including the many people who have contributed to the success of the project.
Paper 5

Effects of adjacent fault systems on drainage patterns and evolution of uplifted rift shoulders: The Lower Cretaceous in the Loppa High, southwestern Barents Sea

Dora Marín, Alejandro Escalona, Sten-Andreas Grundvåg, Henrik Nøhr-Hansen, Bereke Kairanov

Marine and Petroleum Geology, 94, 2018, 212-229, ISSN 0264-8172,

https://doi.org/10.1016/j.marpetgeo.2018.04.009
Effects of adjacent fault systems on drainage patterns and evolution of uplifted rift shoulders: The Lower Cretaceous in the Loppa High, southwestern Barents Sea

Dors Marvin1,2*, Alejandro Estrada1,2, Sten-Andreas Grundvåg3, Henrik Neim-Hansen4, Berit Kolman5

1 Department of Structural Engineering, University of Bergen, N-5008, Bergen, Norway
2 Department of Geosciences, NTNU, The Arctic University of Norway, NO-8013, Trondheim, Norway
3 Geological Survey of Norway and Forskningsvarefabrikk, Stavanger, Norway
4 Geological Survey of Norway and Forskningsvarefabrikk, Stavanger, Norway
5 Geological Survey of Norway and Forskningsvarefabrikk, Stavanger, Norway

ABSTRACT

Sedimentological models for rift shoulders have poorly documented the effect of adjacent fault systems on drainage patterns. In this study, we investigate the Loppa High, an ancient bedrift shoulder located in the southwestern Barents Sea. We use seismic and well data, sedimentological log descriptions, and borehole lithology information to understand the drainage patterns and the Early Cretaceous palaeohydrology of the Loppa High. This study provides a model of how the drainage system in the upper part of a rift shoulder can be modified and remodeled by several fault systems resulting in different drainage patterns in the upper part of the system. The evolution of fault systems and sedimentary processes in the Loppa High during the early Cretaceous can be divided into two main phases: (1) during the Middle Jurassic-Early Cretaceous, the uplifted, and the monoclinal fault complexes were active, and (2) during the Late Cretaceous, the basin-wide fault systems were developed along the northern and western flanks of the Loppa High. The Early Cretaceous-Archean fault systems are interpreted along the Basaltic Loppa Rock Complex. Several generations of faulted valleys and fault-related sub-valleys were formed in the western flank of the Loppa High and 1) during the Late Jurassic-early Cretaceous, the uplifted, and the monoclinal fault systems were active. The latter event triggered a series of fault systems and development of a drainage system in the upper part of the system. 2) During the Early Cretaceous, the basin-wide fault systems were developed along the northern and western flanks of the Loppa High. The Early Cretaceous-Archean fault systems are interpreted along the Basaltic Loppa Rock Complex. Several generations of faulted valleys and fault-related sub-valleys were formed in the western flank of the Loppa High and 1) during the Late Jurassic-early Cretaceous, the uplifted, and the monoclinal fault systems were active. The latter event triggered a series of fault systems and development of a drainage system in the upper part of the system. 2) During the Early Cretaceous, the basin-wide fault systems were developed along the northern and western flanks of the Loppa High. 

1. Introduction

Uplifted rift shoulders are a major factor controlling the filling of rift basins (Ecklhorn & Gans, 1996; Sanderson & Banks, 1996; Allen and Dominy, 2004; Goethorp and Sonder, 2006; Veldkamp et al., 2002; Kopp and Groenewaal, 2006; van den Berg et al., 2013). They usually have high gradients that contribute to the broader fault and one gradient system away from the main faults, resulting in large-scale tilted blocks (Ecklhorn and Gans, 1996; van den Berg et al., 2002; Goethorp and Sonder, 2006). High-gradient systems are characterized by incised valleys and fault-controlled sub-valleys (van den Berg and Groenewaal, 2006; Veldkamp et al., 2007). In contrast, the development of faulted valleys in low gradient systems is less pronounced, and fans need to be progressively studied, indicating the development of deluvial or alluvial fans (Goethorp and Sonder, 2006; van den Berg and Groenewaal, 2006; Kopp and Groenewaal, 2006; van den Berg et al., 2002). Drainage evolution on rift shoulders and the time variations of the deposits depend on several factors, including among others climate, tectonics, and tectonic activity of the bedrock, topography and gradient of the bedrock and the fault, variations in fault propagation and slip rate, and erosion and sedimentation of faults, pre-existing drainage networks, and topography (Ecklhorn and Gans, 1996; Kopp and Groenewaal, 2006; van den Berg and Groenewaal, 2006; Veldkamp et al., 2007; Groenewaal and Kopp, 2007; Groenewaal and Kopp, 2007). Several generations of faulted valleys and fault-related sub-valleys were formed in the western flank of the Loppa High and 1) during the Late Jurassic-early Cretaceous, the uplifted, and the monoclinal fault systems were active. The latter event triggered a series of fault systems and development of a drainage system in the upper part of the system.
Fig. 1. Location map with the major structural elements and the data used in this study. The map also shows the different types of Lower Cretaceous wellbores previously identified in the study area (Beare et al., 2007a, 2007b, 2007c, 2014a, 2014b). The dashed lines represent the shoreline and the pink arrows indicate the nearshore current direction. Continental fan systems are shown with arrows. The colors of the lines represent the seven reservoir units in this study (Figures 2-3): (1) red for the Albian-Cenomanian, (2) orange for the Turonian, (3) yellow for the Coniacian, (4) green for the Santonian, (5) blue for the Campanian, (6) purple for the Maastrichtian, and (7) gray for the Cenozoic. The locations of the examples cited in the text are shown in color in the Web version of this article.

Sea and Gulf of Abu Dhabi (Druckmann and Murti, 1989). In these cases, the main drainage system is directed towards the low-gradient slope area from the major faults (Druckmann and Murti, 1989; Beare et al., 2007a). The drainage directed towards the shelf basin mostly has a small or length, which is typical for some particular cases where the rate of erosion keeps the uplift rate constant (Druckmann and Murti, 1989; Beare et al., 2007a).

The shoulder can be complex structures affected by two or more adjacent fault systems. These fault systems can rejuvenate the topography periodically and preferentially in certain areas of the field. The changes in the drainage patterns (Kabir and Ahmed, 1999; Beare et al., 2007c) can affect the structure of the field at different scales. The drainage systems are often preserved in the geologic record. A model of an early uplifted shelf system affected by adjacent fault systems in the Lower Cretaceous field (Fig. 3). The southern and western flanks of the field are largely controlled by a Late Jurassic-Early Cretaceous rift system (Kabir and Ahmed, 1999; Smith et al., 1999; Kabir et al., 2003). This rift event controlled the geometry of the field, which is characterized by a series of trends and fault scarps in its western and southwestern flanks and gently slope to the north and south. Drainage systems, giving the aspect of a large-scale tilted block (Fig. 2) (Smith et al., 1999; Kabir et al., 2003). Lower Cretaceous systems show that the wedge covers the southern and western flanks of the High and have been targeted in several attempts to map and correlate the stratigraphy (Kabir et al., 2000; Kabir et al., 2006; Beare et al., 2007a, 2007b) and the Lower Cretaceous stratigraphy of the field in the field. In this study, the mechanism controlling the drainage patterns in the field is the structure of the field affected by adjacent drainage fault systems.
2. Geological setting

2.1. Tectonic framework

The Loppa High is located in the southeastern Barents Sea and is bounded to the west by the NE-SW-striking Bjørnesøyren and the N-S-striking Flaggøyen Loppa fault complexes, and to the southeast by the E-W-striking Antokle Fjell Complex (Figs. 3 and 72) (Hesselberg et al., 1992). The southwestern and eastern Loppa High borders with the Bjørnesøyren Platform and with the Hornsmøysa Basin are gently sloping.
to the east and are interrupted by the Lower Cretaceous (Figs. 3 and 4) (Gajewski et al., 2013). During the Triassic to Middle Jurassic, the Loppa High area acted as a depocenter, which was later uplifted (Holm et al., 1980; Wood et al., 1995; Viallat et al., 2013). Most authors have suggested a Late Jurassic–Early Cretaceous extension (Boyd et al., 1984; Beug et al., 1986; Wood et al., 1995) or earlier Cretaceous age for the uplift event (Holm et al., 2011; Viallat et al., 2013).

The complex tectonic history, including the Harstad Event, the Harald Event, and the Ypresian basin, as well as the Nordfjord Rift Basin and the Stavre Graben, were affected by Late Jurassic–Early Cretaceous extension (for locations see Fig. 1) (Hjelle et al., 1982; Kolodziej et al., 1994; Wood et al., 1995; Gajewski et al., 2013; Viallat et al., 2013). At this time, the Harstad Event and an Ayupian-like event are well-correlated with the Harald Event and the Harald Event, both resulting in the formation of large-scale structural domains associated with the main fault systems (Holm et al., 1995; Clark et al., 2014; Larrea et al., 2017). An earlier Cretaceous and an Ayupian-like event are well-correlated with the Harald Event and the Harald Event, both resulting in the formation of large-scale structural domains associated with the main fault systems (Holm et al., 1995; Clark et al., 2014; Larrea et al., 2017). An earlier Cretaceous and an Ayupian-like event are well-correlated with the Harald Event and the Harald Event, both resulting in the formation of large-scale structural domains associated with the main fault systems (Holm et al., 1995; Clark et al., 2014; Larrea et al., 2017). An earlier Cretaceous and an Ayupian-like event are well-correlated with the Harald Event and the Harald Event, both resulting in the formation of large-scale structural domains associated with the main fault systems (Holm et al., 1995; Clark et al., 2014; Larrea et al., 2017).

2.3. The Loppa High

4.7 km

Fig. 3. Sequence stratigraphic framework around the Loppa High. Note that the DSO time gap at the southwestern flank of the Loppa High is from Late Jurassic to Late Early Cretaceous. Two additional unconformities are interrupted in the western Basin of the Loppa High, one during the Late Jurassic to Early Cretaceous event and another during the Late Cretaceous to Early Cretaceous event.
those provided distinct advantages of Early Cretaceous age mixed with dated Pyroclastic Rock unit ages. The age interpretation for well 7920/8-1-2 is based on a S-and H-measurement derived from mudstone with a significant amount of iron from 368 samples prepared by Hesthammer et al. (1894) and the available data in the database, using the method by Hesthammer et al. (1894).

A description and interpretation of the geology is provided, which is based on the geometry of the sequence and the internal reflector characteristics following the principles of Martin et al. (1997). Where necessary, a time-depth conversion and a chronostratigraphic process was performed. The details are given in Martin et al. (1894b) and have been calibrated by the depositional sequence model (Galstrøm et al., 2009).

A detailed sedimentological log description for two cores of well 7920/8-1-2 is included in the depositional environment interpretation. The auxiliary log includes descriptions of rock type, grain size, sorting, sedimentary structures, layer by layer results, and degree of heterogeneity.

4. Lower Cretaceous sequences in the north and east of the Loppa High

During the deposition of the oldest sequence (sequence 6-2), a main erosional phase occurred in the southwestern and the southeastern parts of the Mesozoic basin, associated with the main bounding faults (Fig. 4a). At the top of sequence 2 (Hesthammer Albian), a series of depositional basins is observed (Fig. 5). During deposition of the youngest sequence (sequence 5-6, Albanian-Coniacian), the main depositional areas are located in the eastern parts of the Mesozoic basin, the southeastern part of the Nybukken Platform and the Sjuvalsøy Basin (Fig. 5b). Lower Cretaceous structures located in the east and north of the Loppa High have previously been documented (Soleng & Elvassor, 2011; Sogaard et al., 2014; Elvassor, 2014; Martin et al., 2015). However, their tectono-stratigraphic relationship with the Loppa High is not well understood. Below we provide a description of these structures focusing on their relationship with the Loppa High.

4.1. Lower Cretaceous in the north and east of the Loppa High

The oldest sequence (sequences 6-3) are not properly documented in areas such as the Loppa High and the southwestern part of the Nybukken Platform (Figs. 3 and 4a). However, they are better constrained elsewhere where sequences 5-6 are well defined (Figs. 3 and 4b). The sequence 4-5 is based on the boundary between the Lower Cretaceous and the Upper Cretaceous (Fig. 3b). The Lower Cretaceous structures are observed in the eastern part of the Mesozoic basin, the northeastern part of the Nybukken Platform, and the Sjuvalsøy Basin (Fig. 5b). Lower Cretaceous structures located in the east and north of the Loppa High have previously been documented (Soleng & Elvassor, 2011; Sogaard et al., 2014; Elvassor, 2014; Martin et al., 2015). However, their tectono-stratigraphic relationship with the Loppa High is not well understood. Below we provide a description of these structures focusing on their relationship with the Loppa High.

4.2. Lower Cretaceous in the north and east of the Loppa High

The younger sequence (sequences 5-6) is a period of intense tectonic activity, with the formation of the Loppa High (Fig. 3c). The sequence 4-5 is based on the boundary between the Lower Cretaceous and the Upper Cretaceous (Fig. 3b). The Lower Cretaceous structures are observed in the eastern part of the Mesozoic basin, the northeastern part of the Nybukken Platform, and the Sjuvalsøy Basin (Fig. 5b). Lower Cretaceous structures located in the east and north of the Loppa High have previously been documented (Soleng & Elvassor, 2011; Sogaard et al., 2014; Elvassor, 2014; Martin et al., 2015). However, their tectono-stratigraphic relationship with the Loppa High is not well understood. Below we provide a description of these structures focusing on their relationship with the Loppa High.
Fig. 5. Seismic line showing the Lower Cretaceous sequence in the eastern and northeastern sides of the Liona High. A) Interpreted seismic line showing the Lower Cretaceous sequence. B) Interpreted seismic line showing the relationship of the Lower Cretaceous sequence with the Upper Jurassic sequence. C) Interpreted seismic line showing the relationship of the Lower Cretaceous sequence with the Upper Jurassic sequence. D) Interpreted seismic line showing the relationship of the Lower Cretaceous sequence with the Upper Jurassic sequence.
The lower part of sequence 3 is characterized by wedges closely associated with eroded faults. Above the downlap surface, a package of clinoforms occurs (Davis et al., 2019) (Figs. 5, 7a, and 7b). These clinoforms, pronounced in the 3D seismic profile, are interpreted as the younger high, where they appear to be tilted towards the Hingate sandstone or to the western Beaumont sandstone (Fig. 7a). The top of sequence 1 is the upfolded footwall transected by an unconformity (Fig. 7b). Based on well 756/5-1, the age of the unconformity is late dismemberment early
Fig. 7. A) Interpreted seismic lines. B) Interpreted seismic lines showing the lower Cretaceous exposure in the northern flank of the basin. C) Note the low-angle wedges at the base of sequence 2, and discordant in the upper part. D) Interpreted seismic lines showing the lower Cretaceous exposure in the eastern Jurassic Platform. Grey wedges are abraded in sequence 3–5.
Fig. 6. Gamma ray, lithology distribution, and sequence stratigraphic interpretation of the Lower Cretaceous interval in well 7220/10-1. The sequences are correlated with the stratigraphic framework of Mosi et al. (2017a).
Aptian (Fig. 3) (Bibennene Group, 1994). Thickness changes and wedges, associated with faults, are observed in the area within sequences 2 and 3 (Fig. 7b and d).

5. Lower Cenomanian in the western flank of the Laga High

5.1. Sequence stratigraphy

The graphitic top of well 7220/10-1 suggests that the Lower Cenomanian succession can be divided into three major regressive-transgressive sequences that are separated by several highstands-regressive and transgressive pulses (Fig. 8). The oldest maximum flooding surface, interpreted in well 7220/10-1, is associated with the top of sequence 1, although it appears to be slightly younger than the Hierococynodon fauna, where an age of Upper Albian-Lower Cenomanian was assigned (Mato & et al., 2017b). The second maximum flooding surface is interpreted on the top of sequence 2 with an age of Lower Albian-Lower Cenomanian Aptian. A flooding surface representing the top of sequence 3 is not evident in well 7220/10-1. However, the top of this sequence is interpreted in the western flank of the Laga High based on seismic correlation. The age of this event is lower equivalent with the top of sequence 3 in the Hierococynodon fauna. Sequence 4 is partially identified, but its top is truncated by an unconformity at the top of the Lower Cenomanian (Fig. 9). Sequence 5 of no age (Upper Albian-Lower Cenomanian) is not identified in this well. However, sequence 6 is interpreted in well 7220/5-2 located in the southern Hierococynodon Field Complex (Fig. 9c). Sequences 5 and 6 (Upper Albian-Lower Cenomanian age) are not observed in the western flank of the Laga High.

5.2. Seismic facies and core description

5.2.1. Description

5.2.1.1. The Lower Cenomanian in the southern Hierococynodon Field Complex: The southern section of the Hierococynodon Field Complex is characterized by a series of terraces and fault scarps (Fig. 10). The Lower Cenomanian succession was drilled in one of these terraces by well 7220/5-2 (Fig. 10b). These unconformities are interpreted in the area. The first unconformity coincides with the BCL, the second is an upper
Fig. 18. Vertical line drawings of the sections in the western flank of the basin (Fig. A) and in the eastern flank (Fig. B). The upper section (Fig. A) shows the details of the sequence in the northern segment of the Basin-Paradise Fault Complex, where the bench unconformities are exposed. In the area, the bench is more developed. C) Lower section (Fig. B) showing the wedge associated with the Bakersfield-Lancaster Fault Complex.

Valleymont-Elkwater unconformity and the youngest in an upper bsasment-basement Apley unconformity (Figs. 9 and 10B). These unconformities are characterized by erosional features (i.e., incisions). Incisions are particularly marked in the western and in the youngest unconformity (Fig. 10B). The incisions are located in areas of the basin in the eastern segment of the Basin-Paradise Fault Complex, particularly in the higher mountains located in the east (Figs. 10B and 11B). It is not easy to discern the lengths of these incisions in the
western flank of the Loppa High, since this area has experienced several post-Early Cretaceous erosion events (Sillett and Kalsbeek, 1984; Friedman et al., 2011). The terminations of the RUC level are generally filled by sequence 0 and 1 in the terraces and by sequences 3–4 in the higher terraces (Figs. 18B and 19A). The terminations located at the top of sequence 0 are filled by sequence 1 (Fig. 19B). Sequences 0 and 1 have wedge geometries, thickened near a fault plane. Internally, the reflection pattern is continuous to semi-continuous. Additionally, upper Jurassic wedges have been previously described in the area (Massi et al., 2012). Wedges are observed near to sequences occurring in the higher terraces (Fig. 19C). Sequence 2 flows towards the northern segment of the Hjørundfjorden Fault Complex and its internal reflections are discontinuous (Fig. 19D).

5.2.3.2. The Lower Cretaceous in the Hjørundfjorden Fault Complex. A series of incised Lower Cretaceous wedges are observed in the terraces of the western part of the Loppa High, closely associated with the fault systems of the Hjørundfjorden Fault Complex (Fig. 19B).

Wedges are observed in the Upper Jurassic succession, outside the southern segment of the Hjørundfjorden Fault Complex and sequence 0 is absent here (Figs. 18B, 19C–D). The upper part of sequence 1 forms the Upper Jurassic succession and together with sequence 2 continue the first wedge level (Fig. 19B). Internally, the reflection are en-echelon-like. The age for this first wedge level is late Cretaceous to mid-Early Cretaceous. The younger wedge level is interpreted as part of sequence 3–4 (Fig. 19C). Low relief conformities (30–50 m, appr. 0.5) that are interpreted seaward are incorporated in the eastern wedge (Fig. 19B). The age of this second wedge level is Miocene to Early Pliocene. The wedges in the Hjørundfjorden Fault Complex are also observed in closer proximity to the northern part of the Hjørundfjorden Fault Complex (Fig. 19D). The termination in the southwestern part of the Loppa High are filled with post-late Cretaceous conformities (Fig. 19E).

5.2.3.3. Core description 7229-10-1. Two core sections from well 7229/10-1 were described in this study from 1294.5 m to 1295.7 m (Fig. 15). The investigated core section mainly of terrigenous sediments and 13th very fine to fine-grained sandstones with trace fossils attributed to the Zoophycos, Ophiomorpha and spiral (Ophiomorpha and Mollichia) (Table 4). Based on the core description, three trace fossil associations (FAs) are recognized: 1) FAM is composed of poorly sorted bioturbated mudstones and laminated sediments. Fertility conditions and bioturbation and shell fragments are present. Some of the observed trace fossils include tentaculitids, benthic foraminifera, and Pseudophylophora (Fig. 12, Table 4). 2) FAD is comprised of laminated sediments, alginates, laminated silts and normally grained sandstones with trace fossils attributed to the Zoophycos, Ophiomorpha and spiral (Ophiomorpha and Mollichia) (Fig. 13, Table 4). 3) FAS is comprised of bioturbated mudstones, alginates, laminated silts and normally grained sandstones with trace fossils attributed to the Zoophycos, Ophiomorpha and spiral (Ophiomorpha and Mollichia) (Fig. 14).
Table 3: and FAS is characterized by planar and low-angle lenticular sandstone beds with ripple-cross-laminated or soft-sediment laminae. The sandstone beds are normally graded and are notably bioturbation, with local layers of pyrite nodules. Lenticular structures are common in the sandstone and siltstone facies. Some of the observed trace fossils include Planolites, Zonolites, Arenites, microvertebrates, Trichichnus, and rare Shalifaces. Physiological direction and physicochemical (Fig. 19; Table 13).

Fig. 12: Facies log showing core locations 1 and 3 in well 7220/10-1 Salina. The description of facies transitions follows the photograph index (RI) of Taylor and Golab (1996).
Table 1: Diatom assemblages identified (as well as 722/DR1). The description of degree of bioturbation followed the Bioturbation Index (BI) of Sticky and Goring (1983).

<table>
<thead>
<tr>
<th>BI</th>
<th>Example</th>
<th>Description</th>
<th>Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>189</td>
<td>Open marine sand deposits</td>
<td>Bioturbated sediments, with sandstone beds intercepted by marine fauna.</td>
<td>The degree of bioturbation in these deposits indicates a low degree of bioturbation. (0-4 BI) and the sediments are well-sorted fine sand.</td>
</tr>
</tbody>
</table>
6. Discussion

6.1.TECTONIC EVENTS COLLABORATING THE FORMATION OF LOWER COMMANOE SHELTERED DEEP-WATER FANS

6.1.2. The Laguna High uplift and activity of the Arcola and the Pacifica fault system.

Classic wedge models suggest that Boreal Burianio-Volcanic to lower Gravellino Burianio Apatian fans were deposited in the southerly flank of the Laguna High and along the southern Bajomarquesa Fault Complex. The fans are observed north to incisions interpreted as multiple incised valleys. These incised valleys are interpreted to be formed during the Volcanicis orogeny, before some of them are filled with sediments. The incision valley indicate that they fed the fans along the western and southern fans of the Laguna High (Hsu, 1965, 1972). Lagunian-Jarcio wedges suggest that these incisions on the Laguna High probably eroded from the Late Jurassic (Fig. 4b). For the southerly flank of the Laguna High, the formation of incised valleys is interpreted as diachronous, since their associated fans have been deposited as Boreal Burianio-Volcanic to lower Gravellino Burianio fans in the western part and as Volcanicis-late Burianio fans in the eastern part (Matos et al., 2010).

6.2.2. The Bajomarquesa Fault Complex Wedge associated with the Bajomarquesa Fault Complex suggests that feeding occurred along this fault complex during the Late Jurassic-Gravellino Apatian (sequence 2). These wedges are interpreted as shallow-marine fans (Fig. 5). Incised valleys are observed next to these fans in the updip (northeast) of the Bajomarquesa Fault Complex (Fig. 4b) and along the southern segment of the Bajomarquesa Fault Complex, which forms the fans of sequences 6 and 7 (Fig. 1b, 1d, and 1f). We interpret these incised valleys as a result of incision by the fluvial fans and the related deepwater fans are being formed, as a consequence of a non-homegeneous uplift of the Laguna High, diachronous fault movement, and that its topography was reduced to the western fan (Fig. 9). Furthermore, a Bajomarquesa (sequence 2) unconformity is observed in the Quebrada Foundation formed as a response of faulting uplift (Fig. 7a and d).
Fig. 5B: Three-dimensional sections illustrating the three main events controlling the deposition of the clastic wedges around the Loppa High. A. The activity of the Astafjorden and Ugnsfjorden fault complexes and the right-lateral strike-slip movement of the Astafjorden fault during the time frames when the Astafjorden complex controlled the deposition of progradational clastic wedges. B. The reactivation of a Triassic-Jurassic fault in the western flank of the Loppa High, cutting through the clastic wedge and forming a new detachment zone in the footwall. C. The reactivation of a Triassic-Jurassic fault in the eastern flank of the Loppa High and the creation of a new detachment zone in the hanging wall.

5.2. Drainage evolution and spillway-shoulders affected by adjacent fault systems

Based on the evidence of fault activity in the western and eastern flanks of the Loppa High and the presence of incised valleys, we suggest that the western and eastern flanks of the Loppa High were characterized by topographic highs. By contrast, the western flank of the Loppa High was not affected by 5-4 Ma spillways and there is an absence of incised valleys. The incised valleys present along the western border of the Loppa High, however, as well as the incised valleys present along the eastern border of the Loppa High, were formed during the Tertiary. The morphology of the Loppa High is characterized by a series of incised valleys cutting through the sedimentary rock layers. These valleys are not continuous, suggesting that they were not formed by a single event. Instead, they may represent multiple events, each forming a new detachment zone in the footwall of the Loppa High.
most of the early Cenozoic, it can be expected that a larger drainage system was developed towards the east sourced by the monsoonal high from the western flank. In the eastern flank, climatologies predisposing them to the Loppa High have been identified only in narrow areas in the southwestern flank (34 lamontform lines: Mortsle et al., 2017a) and in the northern part, where the systol et al. (2017b) described N65-predominant climatologies. However, most of the eastern flank of the Loppa High is characterized by a lack of climatological patterns that could indicate the development of a shoreline or delta, conveys, to the uplifted biogenous islands (e.g., Knutsen and Jodin, 1998; Gurovich and Luedke, 2002) or to as in other uplifted ridgelines such as the Red Sea and Gulf of Aden (Gronvold and Toresson, 1998). From this, two questions arise: 1) have we the drainage patterns mapped in the Loppa High? And 2) why is there an apparent lack of drainage systems running from the western uplifted flank of the Loppa High to its low gradient eastern flank? One possibility is that the deposits of these drainage systems (e.g., flood- and drainage channels) are below the bottom resolution.

A second option is the role that adjacent fault systems could have played on the drainage系统 development on the Loppa High. During the Early Cenozoic, the eastern flank of the Loppa High was further affected by different W-, ENE-ENE, and NW-SW-striking faults. For instance, the lower graben and the northern faults that connected the Loppa High to the Mid Basin (Fig. 8 and 7A). This faulting is interpreted to have happened before Albian, almost simultaneously to the faulting in the western flank of the Loppa High. Seismic wedges were only observed within these grabens. We suggest that the W-, ENE-ENE, and NW-SW faults controlled the drainage patterns on the Loppa High, acted as sediment routes, and thus their related depocenters were confined within the grabens (Fig. 136). Mortsle et al. (2017b) suggested that rifts in adjacent basins could synthesize the topography in the post-rift stages and preferentially control the development of basins within a basin. In this paper, we propose another implication of faulting happening in adjacent basins, which is related to the routing of sedi- ment. Although the western flank of the Loppa High behaves as the model for high gradient depositional systems (Gronvold et al., 2001; Clement et al., 2007), (Scheel et al., 2003), the eastern low gradient flank of the Loppa High does not completely follow the previous model. One reason for this disagreement on long, deep systems is apparently missing. (Stroink and Red, 1994; Knutsen and Steer, 1998; (Gronvold et al., 2001; (Scheel et al., 2003)). In the last case, the normal faulting occurring almost orthogonal to the main fault, acted as pathways of sediment in grabens located in the lower part of the basin, controlled overall depositional axis (Fig. 138).

When faulting occurs almost simultaneous in two or more adjacent basins, it is a possible indication how fault activity can affect the drainage patterns and the sedimentation in an area, for example by setting or redirecting sediment pathways. For creating models of sedimentation in rift basins are a useful guidelines. However, in order to make more realistic predictions we need to be able to challenge these models when they do not fulfill our observations and consider other key variables that could have affected a specific area.

7. Conclusions

The ancient Loppa high is an example of an uplifted ridgeline that could have been affected by adjacent fault systems during the early Cenozoic. The flank that faces the mature tectonic complex is char- acterized by large-scale channels as in the western flank, whereas the eastern flank is characterized by a low gradient eastern flank. The gently tilted flank the main drainage systems were confined and deformed to a series of W-, ENE-ENE, and NW-SW grabens structures occurring almost simultaneously with the main fault system. In this study we propose that there are three main events controlled the deposition in the flanks of the high: 1) the activity of the Asteron and the Bronevskoe fault complexes, interpreted to have happened during the Barents Bressage/Utaaey to early Valanginian – early Cretaceous. Associated with clin-orogen, initial valleys and fans were formed. Fans in the area are discernible, indicating that new energy points of the sediment were diachronously formed as a result of anachronous fault movement in the Loppa High flanks. 2) faulting along the Ringvagkja Loppa Fault Complex during the late KIMERIAN-Aptian. Related to this event, instead valleys were formed in the southwestern flank of the Loppa High and shallow marine fans were formed in the terrain of the Ringvagon Loppa Fault Complex. 3) uplift of the Loppa High and the Shackleton Basin by the late Aptian-early Albian. A clinorain switching is suggested as a result of this last event.

Acknowledgements

This study is part of industry-sponsored LoCA consortium. We are grateful with all the sponsors for the financial support. We acknowledge Francis St. and José Luis, Håkan Arnott and an anonymous reviewer for their constructive comments. Thanks to Halliburton-Sandmork and to the Norwegian HDDB database for the software and pleasure pro-

References

\(\text{Ažánek, M., Knies, P., Gyllen, J., Sande, N., 1980.\) Stratigrafisk pædagogisk}\)
\(\text{og geologisk\) formåler for de forende Cenozoikere over nordlige sørøya.\)\)
\(\text{Naturforsk. 3, 235–250.}\)
\(\text{Bogvig, L., Ansgarson, J., Knutsen, R., Gjøen, J., Fjeldvik, E., 1996.\)\)
\(\text{Evolv. of the Structural Evolution of the Norwegian, J., Knutsen, A., Gravimetric Data on a Regional Scale.\)\)
\(\text{Austrehe, J., Folle, F., Kjeldsen, A., 2017.\)\)
\(\text{Two new insights into the micro- and macroevolution of the eastern flanks in the Loppa High.\)\)
\(\text{Naturforsk. 3, 89–99.}\)
\(\text{Brooks, R., 1990.\)\)
\(\text{Elevations and stratigraphic models of the Loppa High. In:}\
\(\text{Schoell, H., Dimanov, E., Träukner, and Tidemand in Fjeldvik, E., 1973.\)\)
\(\text{Fjeldvik, E., 1973.\)\)
\(\text{Hochseefund auf dem Westen der norwegischen Küste.\)\)
\(\text{Ulsteinmuseum for de Norwegian, J., Knutsen, A., Gravimetric Data on a Regional Scale.\)\)
\(\text{Austrehe, J., Folle, F., Kjeldsen, A., 2017.\)\)
\(\text{Two new insights into the micro- and macroevolution of the eastern flanks in the Loppa High.\)\)
\(\text{Naturforsk. 3, 89–99.}\)
\(\text{Brooks, R., 1990.\)\)
\(\text{Elevations and stratigraphic models of the Loppa High. In:}\
\(\text{Schoell, H., Dimanov, E., Träukner, and Tidemand in Fjeldvik, E., 1973.\)\)
\(\text{Fjeldvik, E., 1973.\)\)
\(\text{Hochseefund auf dem Westen der norwegischen Küste.\)\)
\(\text{Ulsteinmuseum for de Norwegian, J., Knutsen, A., Gravimetric Data on a Regional Scale.\)\)
\(\text{Austrehe, J., Folle, F., Kjeldsen, A., 2017.\)\)
\(\text{Two new insights into the micro- and macroevolution of the eastern flanks in the Loppa High.\)\)
\(\text{Naturforsk. 3, 89–99.}\)
\(\text{Brooks, R., 1990.\)\)
\(\text{Elevations and stratigraphic models of the Loppa High. In:}\
\(\text{Schoell, H., Dimanov, E., Träukner, and Tidemand in Fjeldvik, E., 1973.\)\)
\(\text{Fjeldvik, E., 1973.\)\)
\(\text{Hochseefund auf dem Westen der norwegischen Küste.\)\)
\(\text{Ulsteinmuseum for de Norwegian, J., Knutsen, A., Gravimetric Data on a Regional Scale.\)\)
\(\text{Austrehe, J., Folle, F., Kjeldsen, A., 2017.\)\)
\(\text{Two new insights into the micro- and macroevolution of the eastern flanks in the Loppa High.\)\)
\(\text{Naturforsk. 3, 89–99.}\)
\(\text{Brooks, R., 1990.\)\)
\(\text{Elevations and stratigraphic models of the Loppa High. In:}\
\(\text{Schoell, H., Dimanov, E., Träukner, and Tidemand in Fjeldvik, E., 1973.\)\)
\(\text{Fjeldvik, E., 1973.\)\)
\(\text{Hochseefund auf dem Westen der norwegischen Küste.\)\)
\(\text{Ulsteinmuseum for de Norwegian, J., Knutsen, A., Gravimetric Data on a Regional Scale.\)\)
\(\text{Austrehe, J., Folle, F., Kjeldsen, A., 2017.\)\)
\(\text{Two new insights into the micro- and macroevolution of the eastern flanks in the Loppa High.\)\)
\(\text{Naturforsk. 3, 89–99.}\)
\(\text{Brooks, R., 1990.\)\)
\(\text{Elevations and stratigraphic models of the Loppa High. In:}\
\(\text{Schoell, H., Dimanov, E., Träukner, and Tidemand in Fjeldvik, E., 1973.\)\)
\(\text{Fjeldvik, E., 1973.\)\)
\(\text{Hochseefund auf dem Westen der norwegischen Küste.\)\)
\(\text{Ulsteinmuseum for de Norwegian, J., Knutsen, A., Gravimetric Data on a Regional Scale.\)\)
\(\text{Austrehe, J., Folle, F., Kjeldsen, A., 2017.\)\)
\(\text{Two new insights into the micro- and macroevolution of the eastern flanks in the Loppa High.\)\)
\(\text{Naturforsk. 3, 89–99.}\)
Appendices


