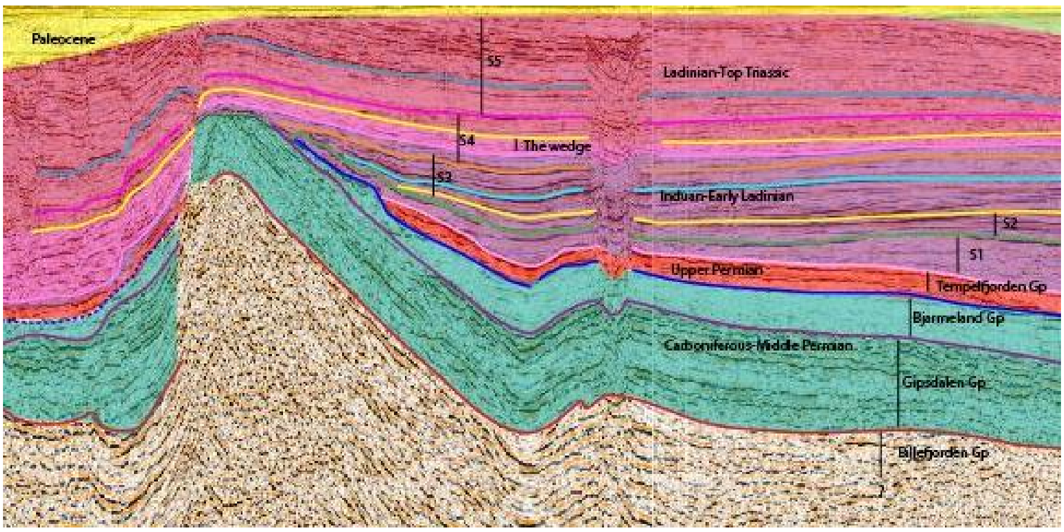
Master Thesis in Geosciences

Late Paleozoic-Triassic evolution of the paleo-Loppa High, linked to tectonic events and depositional patterns

Siri Sagvåg Bjørkesett



UNIVERSITY OF OSLO

FACULTY OF MATHEMATICS AND NATURAL SCIENCES

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Discipline: Petroleum Geology and Geophysics (PEGG)

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Abstract

Seismic 2D datasets and well ties, linked up to previous studies done in the Loppa High area are the major elements of the thesis work. To better resolve the geological history of the paleo-Loppa High area the depositional patterns, structural elements, tectonic events have been studied using seismic facies interpretation and other tools in Kingdom software. Backstripping and structural restoration have also been carried out on a profile across the high.

The main result is a geological evolution of the paleo-Loppa High starting in Carboniferous which was characterized by tectonic stable conditions. Sequences display a uniform thickness throughout the area. In late Carboniferous (Late Pennsylvanian?) the half-graben, situated at the base of the eastern flank, south on paleo-Loppa High was initiated. The upper Gipsdalen unit is thickening in the half-graben. Sediments were deposited throughout the area. In Early Permian (early Sakmarian?) movements in the half-graben continued, in addition to the initial movements on the main western boundary fault of the paleo-Loppa High. Evaporites and a carbonate platform are interpreted. Sediments were deposited throughout the area until the end of the Bjarmeland Group. In Kungurian times the first of two main tectonic events in the formation of the paleo-Loppa High structure occurred. The high was sub-aerially exposed and a barrier to sediments prograding from east to west. The Tempelfjorden Group onlapped paleo-Loppa High, while the western basin was in a state of sediment starvation. In late Permian (Guadalupian) the second main event occurred establishing the paleo-Loppa High structure. Faulting lead to tilting of the eastern flank, which exposed older sediments sub-aerially and erosional processes grinded the crest down to the nearly flat surface observed on seismic. Early-middle Triassic (Induan through early Ladinian) times were regionally a relatively calm period. This goes for the paleo-Loppa high area as well, except for three minor uplift events. Sediments deposited in this time-period onlapped the eastern flank, only draping the crest occasionally. The western basin was still in a state of sediment starvation. In middle Ladinian times the eastern side of the paleo-Loppa High was filled with sediments, and sediments started entering the western basin. Structuring of the Polheim Subplatform may have started in this time-period. The rest of Triassic times were characterized by calm and stable conditions. The former paleo-high developed into a depocenter.

Preface

This master thesis completes a two year master program in Petroleum Geology and Petroleum Geophysics at the Department of Geosciences, University of Oslo. The master thesis is associated with the PETROBAR project, and have been supervised by Professor Jan Inge Faleide.

Acknowledgements

I would like to express my special thanks to Professor Jan Inge Faleide for his guidance, constructive comments, encouragement and interesting discussions during the period of writing this thesis.

I would also like to thank Evy Glørstad-Clark for the increadeable help she has offered with the figures, the text and interesting discussions. I am also very thankful to Dr. Stephen Anthony Clark for his contribution with the BMT software, and to Professor Roy Gabrielsen for his input on the structural geology.

TGS and Fugro are acknowledged for making seismic data available.

Last but definitely not least, big thanks to my supporting partner Eskil Holtan for encouragement and care.

Siri S. Bjørkesett

Contents

1. Introduction

The Barents Sea is one of the world's largest shelf areas flanked by continental margins towards the deep NE Atlantic and Arctic oceans in the west and north, respectively. These oceanic basins have formed by seafloor spreading following continental breakup some 55 million years ago. The shelf is bounded by the mainland of Norway and Russia in the south and Novaya Zemlya in the east (*fig. 1.1*).

The Barents Shelf comprises a wide range of deep sedimentary basins that formed in response to different geological processes during a complex geological history. In some places the sedimentary succession exceeds 15 km in thickness. The eastern Barents Sea is underlain by a wide and deep sedimentary basin that extends for more than 1000 km in a north-south direction. In the western Barents Sea we find more typical rift basins formed in response to at least three major post-Caledonian rift phases (Faleide et al. 1993a,b, Gudlaugsson et al. 1998; see chapter 2 for more details). The western Barents Sea also includes characteristic positive structural elements such as the Loppa and Stappen highs, which have been uplifted and eroded during several tectonic phases affecting the area.

Seismic surveys in the Norwegian Barents Sea commenced in the 1970s, and hydrocarbon exploration drilling started in 1980. In spite of 30 years of exploration history, only few discoveries of commercial interest have been made. Important findings are a major gas discovery in the Snøhvit field and an oil discovery in the Goliat field, both located in the Hammerfest Basin.

The aim of this thesis is to determine the geological history of the paleo-Loppa High from late Carboniferous through Triassic times. The main focus will be to interpret the depositional patterns as a consequence of the tectonic evolution and sediment availability. This will lead to an increased knowledge about structural highs in the western Barents Sea and the effect they have on its surrounding basins.

A dense grid of 2D seismic profiles have been linked to selected wells and the results of previous studies, in particular the studies of the upper Paleozoic succession by Larssen et al. (2005) and the Triassic sequence stratigraphy by Glørstad-Clark et al. (2009).

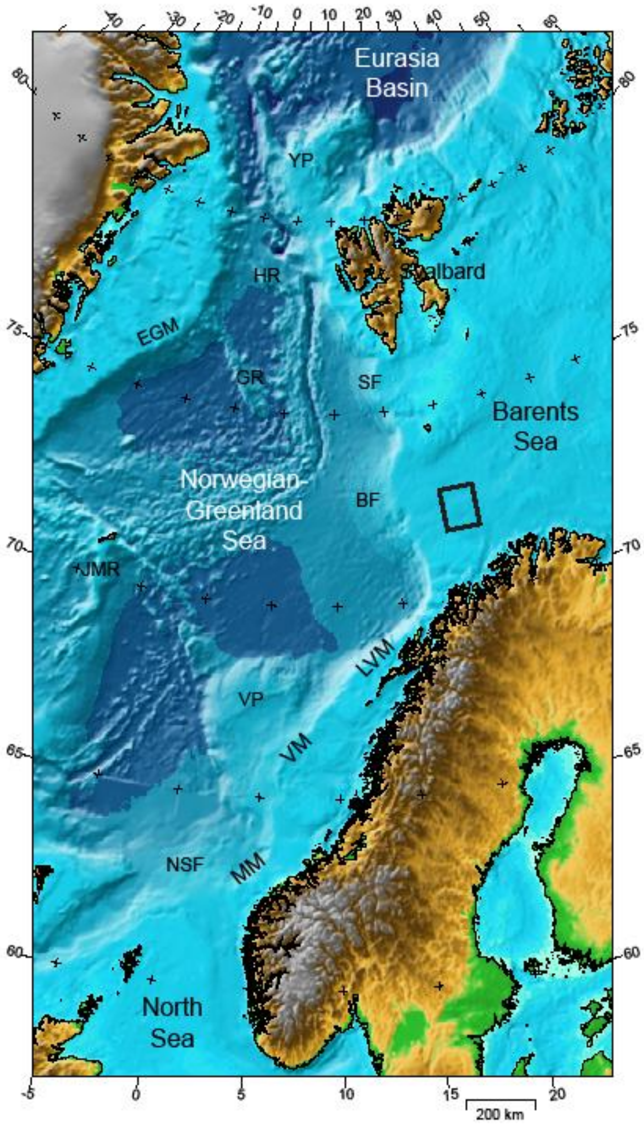


Fig. 1.1: Regional setting and location of study area in the SW Barents sea (modified from Faleide et al. 2008).

2. Geological framework

2.1 Regional setting

The Barents Sea is a wide epicontinental sea, bordered by the Svalbard archipelago and Franz Josef Land to the north, Novaya Zemlya to the east, Norwegian-Greenland Sea to the west and the coast of northern Norway and Kola Peninsula to the south (Breivik et al. 1995, Faleide et al. 1984). The continental shelf of the Barents Sea reaches about 1000 km both in north-south and east-west directions and it constitutes a relatively complete succession of sedimentary strata from late Paleozoic to Quaternary (Faleide et al. 1993a,b, Gudlaugsson et al. 1998). This is preserved in some of the worlds deepest sedimentary basins, formed as a consequence of the breakup of Pangea and the shearing and rifting between Eurasia and Greenland making the North Atlantic Ocean (Faleide et al. 1993a,b).

2.1.1 Main geological provinces

Smelror et al. (2009) roughly divides the Barents Sea Shelf into two major geological provinces. The province`s geological history differs a lot and are separated by a huge monoclinal structure located in the center of the Barents Sea. The eastern province is characterized by the complex tectonic histories of Novaya Zemlya and the Timan-Pechora Basin, and the Uralian Orogeny. The western province is mostly controlled by major post-Caledonian rifting phases as well as later rifting episodes related to the continental breakup and formation of the North Atlantic (Smelror et al. 2009).

The western Barents Sea can further be subdivided into three geological provinces separated by major fault zones (Faleide et al. 1993a,b). *Figure 2.1* shows a map of the western Barents Sea with the geological provinces marked with numbers.

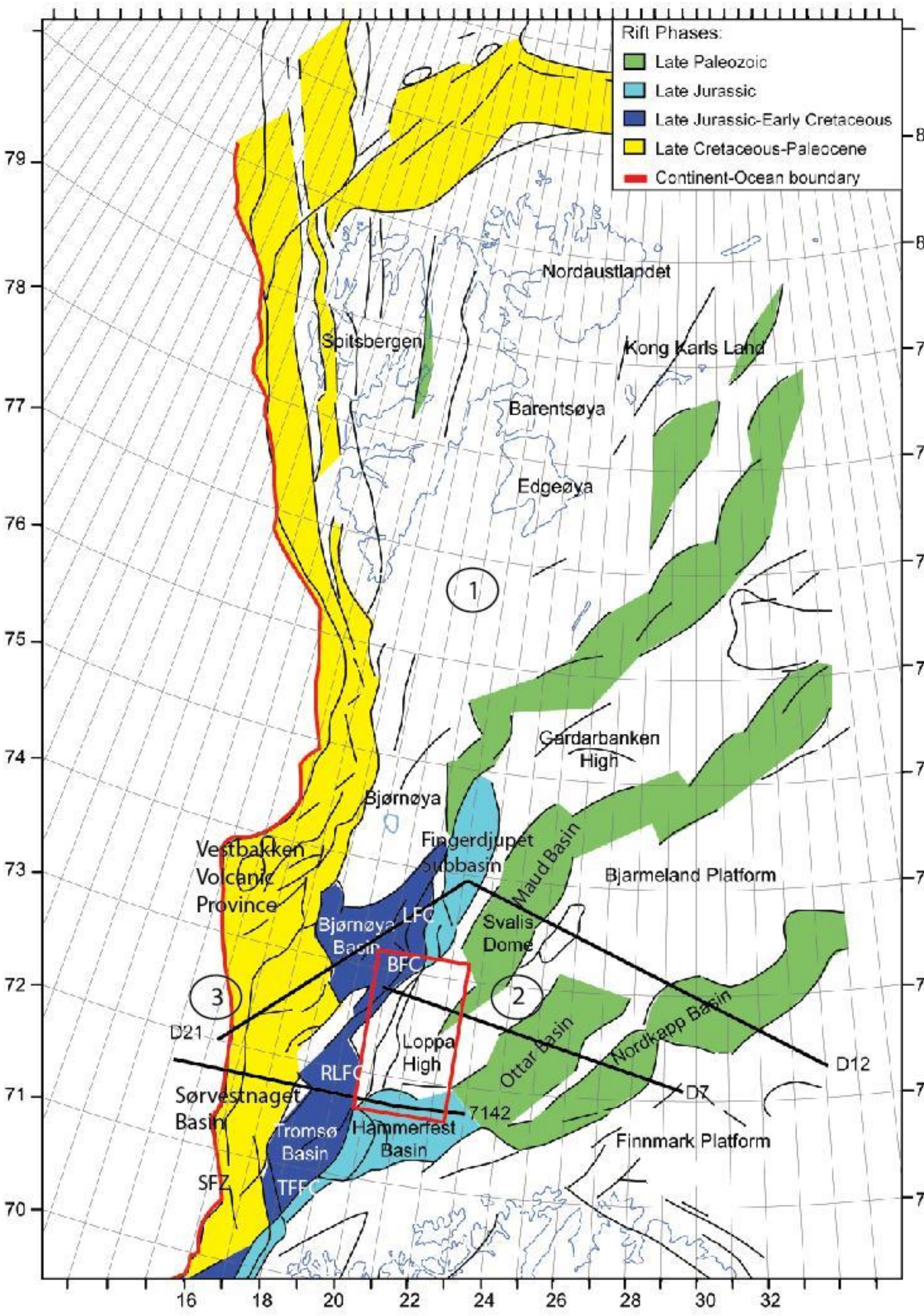


Fig. 2.1: Structural elements of the western Barents Sea. Numbers 1-3 shows the locations of three geological provinces. Black lines illustrate seismic lines in *fig. 2.2*. Red square is the location of the study area of this thesis. SFZ: Senja Fracture Zone. TFFC: Troms-Finnmark Fault Complex. RLFC: Ringvassøy-Loppa Fault Complex. BFC: Bjørnøyrenna Fault Complex. LFC: Leirdjupet Fault Complex (modified from Faleide et al. 2008 and Glørstad-Clark et al. 2009).

1. The Svalbard Platform which has acted as a stable platform since late Paleozoic times. It is covered by a relatively flat-lying succession of Upper Paleozoic and Mesozoic, mainly Triassic, sediments.

2. The basin province between the Svalbard Platform and the Norwegian coast characterized by a number of sub-basins and highs with an increasingly accentuated structural relief westwards. The basins consist of Jurassic-Cretaceous sediments and Paleocene-Eocene sediments as we move westward toward the oceanic Norwegian Greenland Sea. F*igure 2.2 A, B and C* shows three profiles of the western Barents Sea where the age of the sediments are illustrated by color.

3. The western continental margin consists of three main segments (a) a southern sheared margin along the Senja Fracture Zone; (b) a central rifted complex southwest of Bjørnøya associated with volcanism (Vestbakken Volcanic Province) and (c) a northern, initially sheared and later rifted margin along the Hornsund Fault Zone. The continent-ocean transition occurs over a narrow zone along the line of Early Tertiary breakup and the margin is covered by a thick Upper Cenozoic sedimentary wedge.

2.1.2 Structural elements of the SW Barents Sea

The extensional history of the south-western Barents Sea have created a series of rift-basins. Fault-zones divide the south-western Barents Sea into geological provinces. The Senja Fracture Zone (*fig. 2.1, SFZ*) and the Vestbakken Volcanic Province (*fig. 2.1*) marks the western limit of the continental shelf and the transition to oceanic crust. In addition the western Barents Sea holds several Jurassic-Cretaceous fault-zones that creates the boundaries of deep sedimentary basins; The Troms-Finnmark Fault Complex south of 71° N (*fig. 2.1, TFFC*), the Ringvassøy-Loppa Fault Complex (*fig. 2.1, RLFC*), Bjørnøyrenna Fault Complex (*fig. 2.1, BFC*) and Leirdjupet Fault Complex (*fig. 2.1, LFC*)(Faleide et al. 1993a,b).

The basins of the south-western Barents Sea are developed in time from oldest in east to youngest in west. In east three main basins of Late Paleozoic age are shown in *figure 2.1.* The Nordkapp Basin is of pre-Permian origin and is several kilometers deep (*fig. 2.2 C*). The Ottar Basin is situated north of the Nordkapp Basin, strikes north-east, holds two large salt domes and sediments of late Paleozoic age (*fig. 2.2 A and C*). The third main basin of Late Paleozoic age is the Maud basin which trends north-east, north-north-east, and is situated east of the Fingerdjupet Subbasin. In south it is dominated by the Svalis Dome (*fig. 2.1*)(Faleide et. al 1998).

The central part of the south-western Barents Sea holds three structures of interest of Late Jurassic age (*fig. 2.1*). The Hammerfest Basin is classified as a Late Jurassic basin (*fig. 2.2 B*), but it holds evidence of strata of pre-Permian age. In north the Hammerfest Basin bounds the Loppa High by the Asterias Fault Complex and has evidence of syn-tectonic sedimentation of pre-Permian age towards these faults (Faleide et. al 1998). The Loppa High is a result of Late Jurassic to Early Cretaceous and Late Cretaceous-Tertiary tectonism (Gabrielsen et al. 1990), but it was probably initiated by middle Carboniferous (*fig. 2.2 C*) (Faleide et. al 1998). The Fingerdjupet Subbasin is situated north of Loppa High. Late Jurassic tectonism generated the dominate fault trend, although it was formed in Early Cretaceous as a shallower part of the Bjørnøya basin (*fig. 2.2 A*)(Gabrielsen et al. 1990).

Further west two basins of Late Jurassic-Early Cretaceous age is located (*fig. 2.1*). The Tromsø basin evolved mainly due to extension in this time period, but contains salt diapirs from salt deposition in late Paleozoic times (*fig. 2.2 B*). The Bjørnøya Basin is located north of the Tromsø Basin. The two basins merged post late Paleozoic times and separated in Late Cretaceous by the Bjørnøyrenna Fault Complex. Most sediments in the Bjørnøya basin is of Early Cretaceous age (*fig. 2.2 A*) (Faleide et al. 1993a,b).

In west, just before reaching the Continent-Ocean boundary two major basins are located (fig. 2.1). The Sørvestnaget basin holds a very thick succession of Cretaceous and Tertiary sediments and it is a continuation of the Bjørnøya Basin (*fig. 2.2. B*). In north the basin is limited by the Vestbakken Volcanic Provinces lava which is of Early Tertiary age (Gabrielsen et al. 1990).

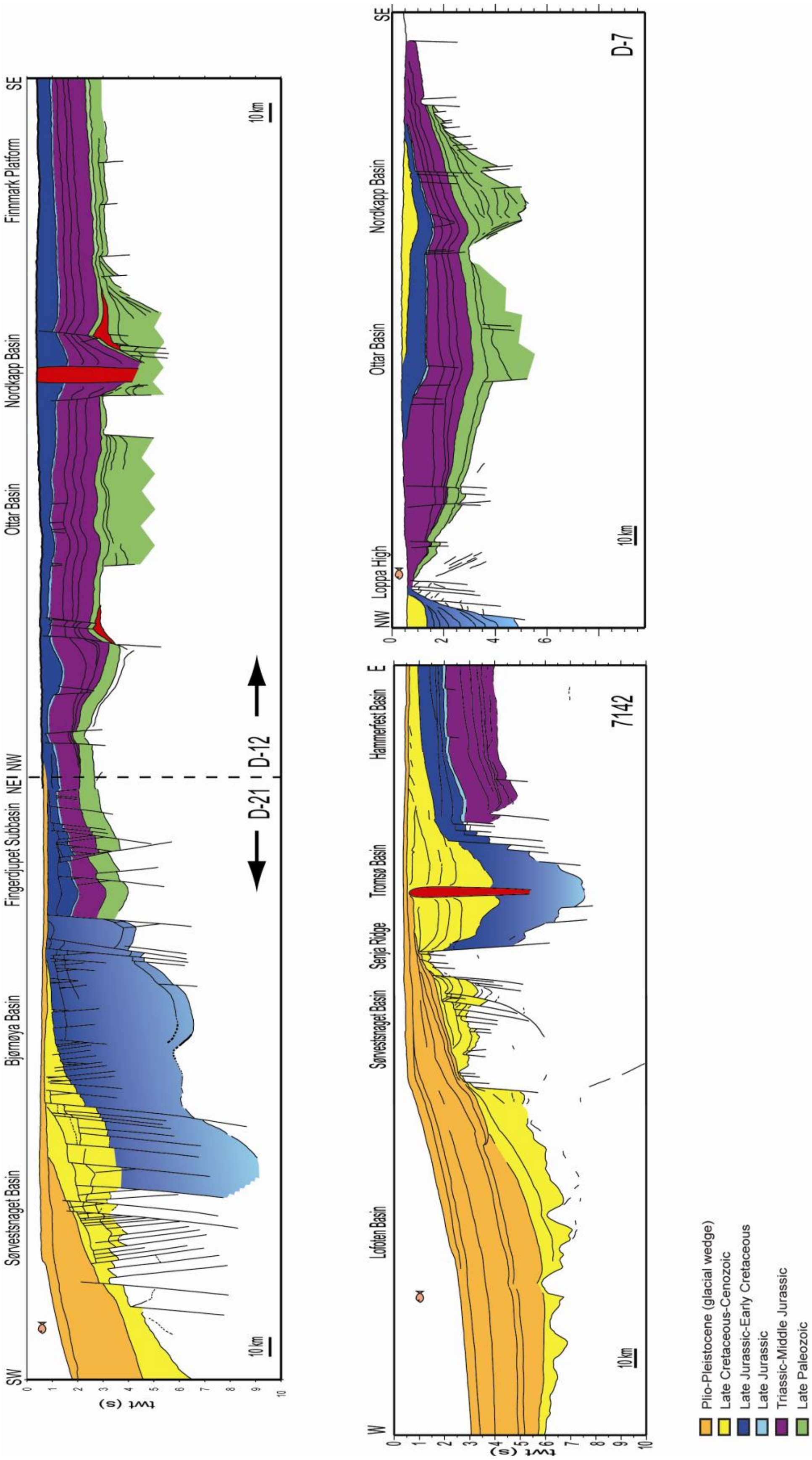


Fig. 2.2: Regional profiles across the SW Barents Sea. Locations of lines see *figure 2.1* (from Faleide et al 2009).

2.2 Regional geological evolution

The tectonic evolution of the Barents Sea starts with a major continental collision and basement formation which is followed by complex rifting events eventually leading to a continental break-up and the formation of the Norwegian-Greenland Sea and the Arctic Ocean to the north. The rifting period forms many sedimentary basins in which the most significant ones lies in the Russian part of the Barents Sea (Smelror et al. 2009).

The structural history of the western Barents Sea, Post-Caledonian basement formation, can be split into three major rift phases (Faleide et al. 1993a,b);

1. Late Devonian? to Carboniferous
2. Middle Jurassic to Early Cretaceous
3. Early Tertiary

The Barents Sea sedimentary basins were formed by a series of tectonic events since the Caledonian orogenic movements terminated in early Devonian (*fig. 2.3*). The Barents Sea continental shelf is dominated by ENE-WSE to NE-SW and NNE-SSW to NNW-SSE structural trends. The eastern part of the Barents Sea has been relatively stable since late Carboniferous, while the western part of the Barents Sea has been tectonically active from Late Paleozoic (Gabrielsen et al. 1990).

The western Barents Sea sediments is underlain by Late Silurian to Early Devonian metamorphic basement (Faleide et al. 1984, Smelror et al. 2009). The basement was formed when the Laurentian and Baltican plates collided and sutured into the Laurasian plate. The collision led to the Caledonian Orogeny (Smelror et al. 2009). Caledonian structures visible onshore have a north-eastern trend in northern Norway and a north-western trend in Svalbard. These old structures of thrust faults extend across the Barents Sea and makes up the pattern for younger extensional events (Ritsman and Faleide 2007, Dengo and Røssland 1992, Faleide et al. 1984).

Several authors claims that most known structural trends may have been established by Devonian times and some of the most important features may even be related to fracture systems formed during the Caledonian Orogeny. Most likely these old lineaments lies within the basement which was buried by younger sediments and later influenced the tectonic movements in Late Paleozoic to Cenozoic (Gabrielsen et al. 1990). Dengo and Røssland (1992) stated that Caledonian lineaments had a fundamental control in localizing the position of major north-east and north-west trending basin–bounding normal faults. It is documented in many rift basins around the world that pre-existing thrust-faults can be reactivated and influence the localization of younger extensional faults (Dengo and Røssland 1992).

Following the mountain creating processes in Late Silurian to Early Devonian is a time of exhumation and extensive erosion of hinterland in Middle Devonian to Early Carboniferous. Gradually the Caledonian Orogeny was eroded and the western Barents Sea was peneplaned in Frasnian times. This lead to accumulation of Old Red Sandstones in western Barents Sea basins (Smelror et al. 2009).

During Late Devonian-Early Carboniferous there is a change in stress regime and the area goes from compressional tectonics to extensional tectonics. Rift basins is formed on the continental shelf and filled up with continental clastics, carbonates and evaporites (Faleide et al. 1984). Late Devonian-Middle Carboniferous rift phase resulted in several interconnected rift basins filled with syn-rift deposits (Faleide et al, 1984). Stemmerik (2000) stated that sedimentation in this period is characterized by non-marine deposits in narrow, isolated half-grabens. This is documented in East Greenland, Spitsbergen and Bjørnøya (Stemmerik, 2000).

The oldest extensional event that can be mapped out in the western Barents Sea occurred in Late Devonian-Early Carboniferous and was due to the initial rifting between Norway and Greenland. This event establishes the fundamental basement architecture of half grabens and inter-basinal highs, which controls the younger basins later deposited. The basins are a result from the combined effects of sinistral strike-slip faulting in the western Barents Sea and a conjugate dextral strike-slip fault in the central Barents Sea (Dengo and Røssland. 1992). Tromsø, Bjørnøya, Nordkapp, Fingerdjupet, Maud and Ottar are basins formed at this time in addition Hammerfest Basin may also have been in initiated at this time (Gudlaugsson et al. 1998).

In regional view crustal extension ceased in Late Carboniferous and a period of basin subsidence and calm tectonic environment followed through to mid Jurassic (Dengo and Røssland 1992, Gudlaugsson et al. 1998).

The Carboniferous rifting was replaced by a quiet tectonic period in most of the Barents Sea in Middle Carboniferous. Regional subsidence and accumulation of sediments accumulated in a regional sag-basin (Gudlaugsson et al. 1998). A widespread carbonate shelf covering the areas from Sverdrup Basin to the Pechora Basin was established. Carbonates of various facies, evaporites and some clastics were deposited in layers of relatively even thickness (Faleide et al. 1984). From late Permian through Early Triassic the Ural mountain chain in east supplied the western Barents Sea with clastic sediments (Dengo and Røssland 1992). Triassic in general was a quiet tectonic period, characterized by regional subsidence leading to onlap on the local highs (Faleide et al. 1984). The calm tectonic environment followed through to mid Jurassic (Dengo and Røssland 1992, Gudlaugsson et al. 1998), interrupted only by renewed rifting in Permian-Triassic in the N-S striking structural elements of the Western Barents Sea (Gudlaugsson et al. 1998).

Late Permian-Early Triassic tectonic movements led to normal faulting, uplift, tilting and erosion. East-west crossing seismic profiles show clear evidence of syn-tectonic sedimentation. Late Permian-Early Triassic fault movements can be tracked in north-south trending structures of the Western Barents Sea, evidence of fault movement is found along the western Loppa High and as far north as the Fingerdjupet basin (Gudlaugsson et al. 1998). From Carboniferous to Permian times Loppa High experienced a total of eight identified tectonic events, two of these might be Triassic in age (Johansen et al. 1994).

Mid-Late Jurassic to Early Cretaceous renewed crustal extension occurred from Loppa High and westward. Deformation created a series of pull-apart basins, and tilted fault blocks. The north eastern structural trends inherited from Late Devonian-Early Carboniferous deformation is still current, and controls the tectonic patterns across the western Barents Sea (Dengo and Røssland 1992). A number of highs became positive features in this period as a result of faulting and differential subsidence (Faleide et al. 1993a,b) .

Middle to Late Jurassic rifting splits the Barents Sea through the Hammerfest and Bjørnøya basins, and decouples the northeastern Barents Sea from the western margin (Smelror et al. 2009). The rifting event leads to block faulting and deposition of Late Jurassic shales between the faulted blocks. At the transition from Late Jurassic to Early Cretaceous major rifting probably lead to a lowstand in sea-level, this resulted in a major erosional surface visible in the entire North Atlantic, known as the Base Cretaceous unconformity. During Early Cretaceous Bjørnøya, Tromsø and Harstad basins become deep basins and main depocenters in the western Barents Sea (Faleide et al. 1993a,b). In middle Cretaceous times the northeastern part of the Barents Sea was uplifted and eroded. This resulted in large amounts of sediments being transported into to the rapidly subsiding basins along the western margin. The uplift possibly coincides with the increased volcanic activity on Franz Josef Land, Kong Karls Land and adjacent offshore areas (Smelror et al. 2009).

Successive rifting throughout Cretaceous further enhanced the deep basins; Tromsø, Harstad and Bjørnøya on the western margin of the Barents Sea. The rifting developed into a dextral stress field along the Senja-Hornsund lineament during Paleogene. This lead to the formation of pull-apart basins in the westernmost part of the Barents Sea (e.g. Sørvestnaget Basin and Vestbakken Volcanic Province) (Smelror et al. 2009).

During Paleogene sea-floor spreading started and the final continental breakup of the North Atlantic was a fact. The Norwegian-Greenland Sea was set around the transition from Paleocene to Eocene, 55 Ma years ago (Faleide et al. 2008, Smelror et al. 2009).

Regional uplift of the Barents Sea shelf was initiated some time during the Oligocene-Miocene time interval, prior to the Plio-Pleistocene Northern Hemisphere glaciation (Faleide et al. 1996, Dimakis et al. 1998).

In Late Pliocene Pleistocene times the Barents Shelf was covered by ice in several phases. The repeated glaciations caused major uplift and erosion. The sediments were transported to the shelf margin and deposited as wedges (Faleide et al. 1996). Fans and debris-flows with glacigenic sediments (Laberg and Vorren 1996), are found with thicknesses up to 4 km. The uplift and erosion was at a maximum around Svalbard where 2-3 km of sediments were removed, while further south in the Loppa High the amount up uplift and erosion in general were less than 2 km (Dimakis et al. 1998). The glacial event is the creator of the unconformity between Mesozoic-Tertiary strata and the overlying glacial deposits (Smelror et al 2009).

Figure 2.3 summarizes the geological history of the western Barents Sea.

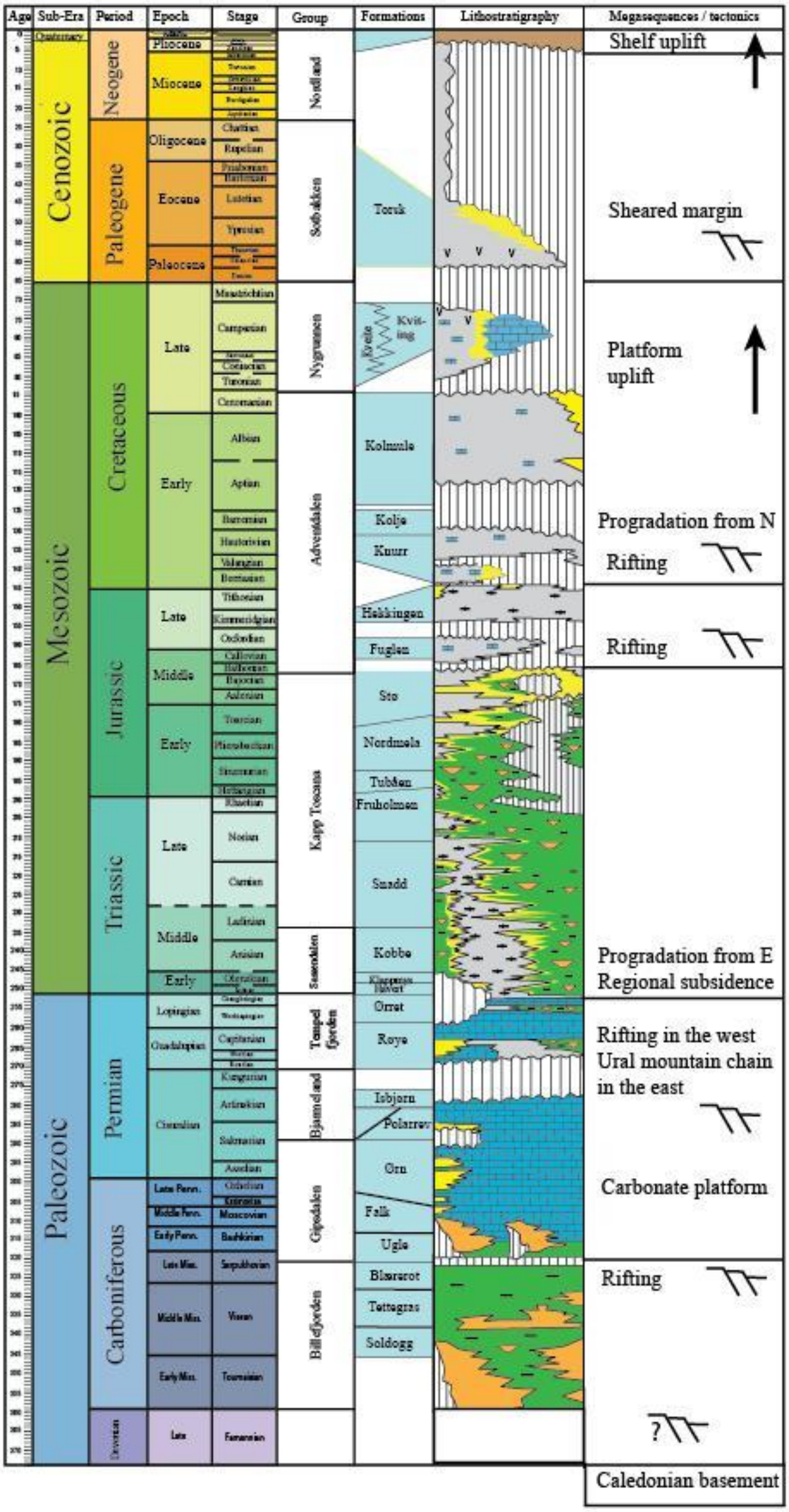


Fig. 2.3: Stratigraphic diagram of the western Barents Sea (modified from Glørstad-Clark et al. 2009)

2.3 Loppa High

Loppa high is a diamond shaped structural feature (Larssen et al. 2005), situated between 71°50’N, 20°E and 71°55’N, 22°40’E and 72°55’N, 24°10’E and 73°20’N, 23°E (Gabrielsen et al. 1990).

It incorporates the Polheim subplatform, and is according to Gabrielsen et al. (1990) bounded:

-in south by Asterias Fault complex, and on east and southeast by a monocline towards the Hammerfest Basin and the Bjarmeland Platform.

-to the west, the Loppa High is bounded by the Ringvassøy-Loppa and Bjørnøyrenna Fault Complex.

-to the north east a major salt structure, the Svalis Dome, and its associated rim syncline and the Maud Basin makes the limit of the high.

The area has undergone a complex geological history characterized by several phases of uplift and subsidence followed by tilting and erosion (Larssen et al. 2005).

Gudlaugsson et al. (1998) studied Loppa High and found a triangular or trapezoidal basement block with steeply easterly dipping flank and a fault bounded western flank. The basement has seismic reflectors that dip steeply towards east and is overlain by an Upper Paleozoic sedimentary wedge that is also dipping east and thickening eastward. At the ridge/crest there is an angular unconformity that truncates the Upper Paleozoic and basement layers, succeeded by Triassic and younger layers (Gudlaugsson et al. 1998).

The high as defined now is a result of Late Jurassic to Early Cretaceous-Tertiary tectonism (Gabrielsen et al, 1990), but the development started with mid Carboniferous rift topography, followed by tectonic uplift tilting of the flank during Late Permian and Early Triassic. The created relief was filled and draped successively by Upper Paleozoic and Lower Triassic sediments (Larssen et al. 2005). In early Cretaceous Loppa High was once again uplifted, after being a depocenter during Late Triassic and Jurassic. Tectonic activity along the Asterias Fault complex is documented by reflectors onlapping Loppa High in the Hammerfest basin in Aptian-Albian times. The erosion of Triassic sediments were a result of footwall uplift of the western boundary fault (fault A) that probably were amplified by lateral heat transport from the developing rift basins in south and west (Faleide et al. 1993a,b).

Polheim Subplatform constitutes the western part of the Loppa High area. It is situated between 72°N and 72°30’N at 20°E, where it forms the block-faulted area between the stable eastern part of the Loppa High and the Bjørnøyrenna and Ringvassøy-Loppa Fault Complexes (Gabrielsen et al. 1990).

The Polheim Subplatform is situated on the westernmost part of the Loppa High, and formed a positive, tectonically active element of Loppa High during Late Paleozoic times. In Early-Middle Triassic it was downfaulted relative to the crest of Loppa High. The platform is heavily deformed by faulting starting in Permian, and increasing in Triassic and Jurassic to Early Cretaceous. During the creation of the Ringvassøy-Loppa Fault Complex the subplatform slid westward and formed the structural pattern of rotated fault blocks (Gabrielsen et al. 1990).

3. Seismic Interpretation

2D seismic interpretation was used to create an understanding of the western side of paleo-Loppa High, which was further calibrated by existing work in the area and surrounding well bores. The 2D seismic data were provided from Fugro and TGS Nopec and well data from the Norwegian Petroleum Directorate (NPD). Lithostratigraphic and seismic stratigraphic successions from Glørstad-Clark et al. (2009) and Larssen et al. (2005) were used to correlate seismic horizons and sequences within this study. In addition, gravity measurements have been helpful to better understand the shape, depth and extension of the structural elements in the Loppa High area.

3.1 Seismic and well data

The southwestern Barents Sea is covered by a relatively dense grid of 2D reflection seismic data. Kingdom software was used as a tool to interpret the area of interest in this study. We have made a polygon around the paleo-Loppa High to limit the seismic lines relevant to this study (*fig. 3.1).*

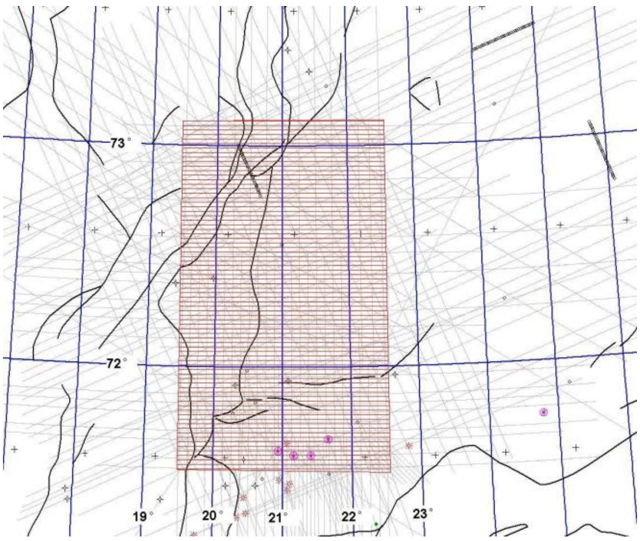


Fig. 3.1: Map of the study area showing a polygon of seismic lines relevant in this thesis

Larssen et al. (2005) published the lithostratigraphy of the Upper Paleozoic succession in the southern Barents Sea. They used 13 exploration wells and 12 IKU shallow cores, combined with seismic data to define a formal nomenclature. Two wells were used to determine Upper Paleozoic units in this paleo-Loppa High study; 7121/1-1 and 7120/2-1. They are situated on the eastern side of paleo-Loppa High and Fault Zone A. Seismic units in this study are correlated to wellbores based on the work of Larssen et al. (2005), and further correlated seismically in the study area. The Triassic seismic units were calibrated to the seismic sequence stratigraphic work of Glørstad-Clark et al. (2009). They used seismic facies analysis and time-thickness maps to reconstruct paleogeography and understand accommodation space development in the Triassic east of the paleo-Loppa High. Their work was done on 2D seismic data sets and all public data from exploration wells in the study area (Glørstad-Clark et al. 2009).

3.2 Seismic sequence stratigraphy

As a prerequisite to seismic interpretation of the paleo-Loppa High area, a framework of previously published data regarding well ties, stratigraphy and sequence stratigraphic horizons and structural elements is necessary. This knowledge is applied alongside own interpretations in a reconstruction of the geological evolution of the paleo-Loppa High concerning tectonic events and depositional patterns.

3.2.1 Stratigraphic framework

Most horizons east of the paleo-Loppa High are defined by previous workers (Larssen et al. 2005, Glørstad-Clark et al. 2009). The horizons are used as a basis for further seismic investigation west of the paleo-Loppa High in this study.

Larssen et al. (2005) defined the Upper Paleozoic lithostratigraphic nomenclature with well ties from two wells in the Loppa High area. The location of the two wells, 7120/2-1 and 7121/1-1, on a seismic profile are shown in *figure 3.2.* An overview of lithostratigraphic units in these wells is shown in *table 3.1*.

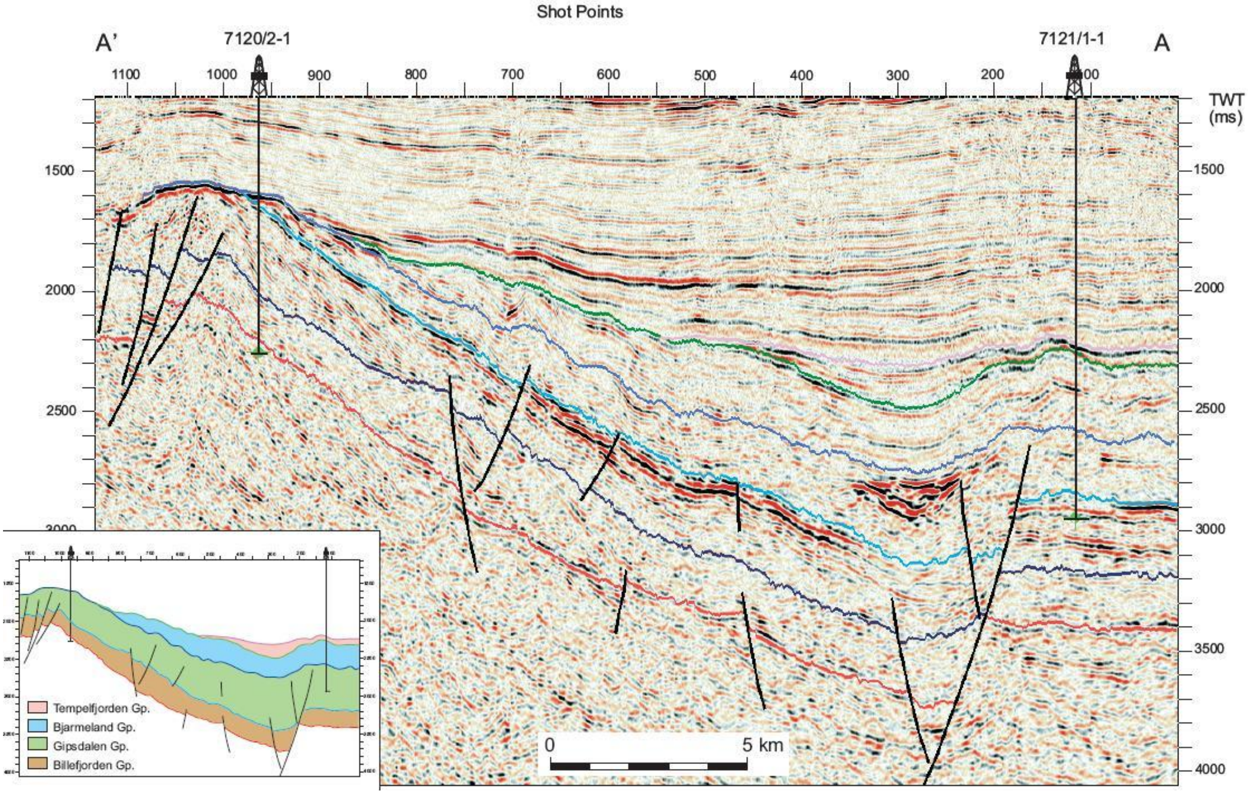


Fig. 3.2: Well 7120/2-1 and 7121/1-1 shown on a seismic profile with correlated lithostratigraphic units (from Larssen et al. 2005).

Lithostratigraphy	Well 7120/2-1	Lithostratigraphy	Well 7121/1-1 R
Top depth [m]	Lithostratigraphic unit	Top depth [m]	Lithostratigraphic unit
410	NORDLAND GP	396	NORDLAND GP
476	SOTBAKKEN GP	519	SOTBAKKEN GP
476	TORSK FM	519	TORSK FM
613	KAPP TOSCANA GP	698	KAPP TOSCANA GP
613	SNADD FM	698	FRUHOLMEN FM
1933	KOBBE FM	792	SNADD FM
1945	GIPSDALEN GP	2210	SASSENDALEN GP
1945	ØRN FM	2210	KOBBE FM
2024	FALK FM	2605	KLAPPMYSS FM
2221	UGLE FM	2786	HAVERT FM
2624	BILLEFJORDEN GP	2993	TEMPELFJORDEN GP
2624	UNDIFFERENTIATED	2993	RØYE FM
3471	BASEMENT	3502	BJARMELAND GP
		3502	ISBJØRN FM
		3586	ULV FM
		3625	ISBJØRN FM
		3700	POLARREV FM
		3990	GIPSDALEN GP
		3990	ØRN FM

Table 3.1: Lithostratigraphic units in wells 7120/2-1 and 7121/1-1 (modified from www.npd.no).

Well 7120/2-1 was drilled in 1985 on the Loppa High in the northern part of the block, offshore Northern Norway (coordinates: 71°58`57.94``N and 20°28`35.09``E). Basement was penetrated at depth 3471 m (*table 3.1*). A major unconformity (613 m) separates the Tertiary from the Triassic sediments with missing sediments in the age range Late Triassic to Early Paleocene. A second stratigraphic break was observed at 1945 m, where Middle Triassic sediments were found resting unconformable on sediments of early Permian age (the Base Ladinian Unconformity). At 1945 m a limestone with extremely high gamma ray was encountered. Below 2140 m the rocks are dominated by dolomitic limestone with thin sand- and slate- layers (www.npd.no).

Well 7121/1-1R was drilled in 1986 on the Loppa High in the northern part of the Tromsøflaket area, offshore Northern Norway (coordinates: 71°56`25.74``N and 21°4`36.52`E). Oldest penetrated age is late Carboniferous in the Ørn Formation (*table 3.1*). Triassic rocks were encountered at 698 m, unconformably underlying 178 meters of Tertiary claystone and siltstone. The Triassic sediments (2295 m thick) consist predominantly of very fine clastics with minor interbeds of sandstone, stringers of dolomite and limestone and traces of coal. The Upper Paleozoic succession (Permian - late Carboniferous Ørn Formation) was encountered at 2993 m. This section, +2007 m thick, consists predominantly of carbonates (silicified limestones, limestones, dolomitic limestones, dolomitized limestones and dolomites) with minor interbeds of chert, siltstones and anhydrites (www.npd.no).

Four major depositional units are mapped out of Late Paleozoic age in Larssen et al. (2005); the Billefjorden Group, the Gipsdalen Group, the Bjarmeland Group and the Tempelfjorden Group. The units are based on studies from exploration wells, cores, seismic interpretation and papers of previous stratigraphic and sedimentological work (see Larssen et al. 2005 for key references). A schematic view of the Upper Paleozoic lithostratigraphic sequences as interpreted by Larssen et al. (2005) is shown in *figure 3.3*, and their interpretation on a seismic profile is shown in *figure 3.2.*

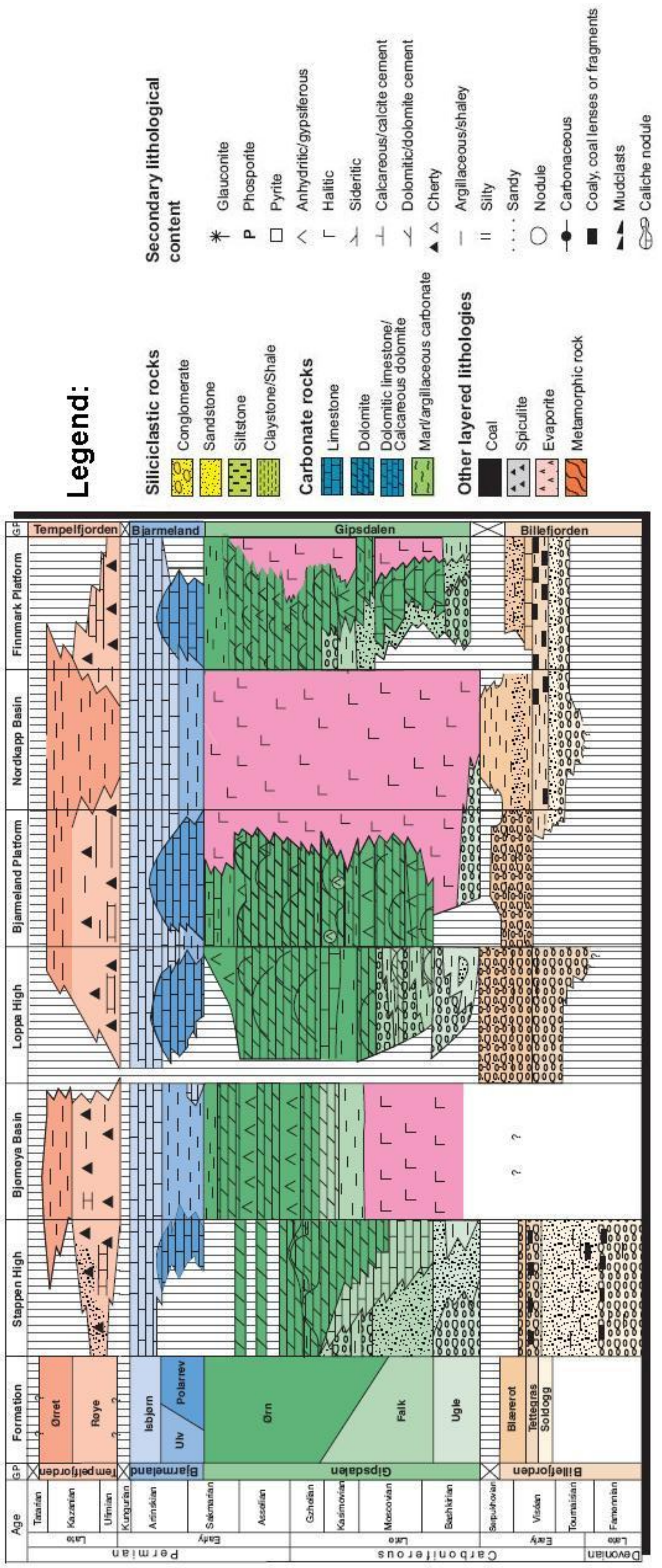


Fig. 3.3: Upper Paleozoic Lithostratigraphic units (from Larssen et al. 2005).

The Billefjorden Group is the oldest group recognized in the Loppa High area and is believed to be of Late Devonian to early Carboniferous age. It is a non-marine succession characterized by an overall transition from continental fluvial deposits, into transitional continental to marginal marine deposits. The lithologies at Loppa High consist of breccias of different colors, conglomerates, ignimbrites and other types of volcano clastics. The depositional environment is alluvial fans, proximal braided river systems and some local volcanic activity. Generally, the area is transitioning from a continental fluvial dominated environment to a sub-tropical to arid climate and marine carbonate shelf in the Gipsdalen Group (Larssen et al. 2005).

The Gipsdalen Group is of mid-Carboniferous to early Permian age, and is found at the southern part of paleo-Loppa High to be 1000 meters thick. The lower boundary of late Serpukhovian/early Bashkirian age marks a change in paleo-climate, it went from warm and humid in middle Carboniferous (Billefjorden Group) to warm and arid to semi-arid in the late Carboniferous (Gipsdalen Group). This affected the character of the deposited sediments (Larssen et al. 2005). The lithologies at Loppa High, which at the time was a low-angle ramp dipping towards the Nordkapp and eastern Hammerfest basins, are a mixture of carbonates and siliciclastics. The formation is dominated by shallow marine carbonates on the platform areas and interbedded carbonates and evaporites in the more distal ramp to basinal settings (Larssen et al. 2005). During periods with high sea level algal build-ups were developed. In general this was a time with high frequency and high amplitude eustatic sea-level change that formed stacked carbonate platforms (Smelror et al. 2009).

The Bjarmeland Group was deposited from early to middle Permian times, specifically middle Sakmarian to late Artinskian. It is thickest developed on the eastern flank of Loppa High and eastwards on the Bjarmeland Platform. It onlaps the paleo-Loppa High and consists of limestone made from cool-water fauna, and cherts. The group is not recorded on top of paleo-Loppa High. The Bjarmeland Group was deposited under a more temperate cool water carbonate environment compared to the Gipsdalen Group. There was limited siliciclastic input except in areas adjacent to sub-aerially exposed parts of Stappen High (Larssen et al. 2005).

The Tempelfjorden Group is of middle-late Permian age. The onset of deposition of this group marked the change to a platform type sedimentation in an initial development of a regional sag basin that continued to subside through the deposition of the Tempelfjorden Group sediments (Gudlaugsson et al. 1998). The group thins and is truncated at the paleo-Loppa High showing repeated tectonic uplift in late Permian (Larssen et al. 2005). The carbonate deposition was now coming to an end, and is replaced with a more siliciclastic regime (Larssen et al. 2005, Smelror et al. 2009) The lithology consists of cherts, silicified limestone and fine grained and coarse grained siliciclastics depending on where you area (Larssen et al. 2005). The depositional environment was affected by a transgression and retrogradation of the coastline. Climate became cooler, and was accompanied with a widespread sponge fauna (Larssen et al. 2005, Smelror et al. 2009).

Glørstad-Clark et al. (2009) mapped out five Triassic second order sequences shown in *figure 3.4*, which is part of the late Permian to Middle Jurassic megasequence. The second order sequences were mapped out on 2D seismic, and further tied to already published well data. The base boundary of the megasequence is situated in the upper Permian, represented by deposition of the top of the lithostratigraphic Tempelfjorden Group, which is characterized by the change from predominantly carbonates to clastic deposition. The top of the megasequence is characterized by erosion of the underlying units as a result of rifting processes in Middle-Late Jurassic. The second order sequences are characterized by stratal downlap termination representing maximum flooding surfaces. Each sequence was analyzed in terms of time-thickness maps, seismic facies analysis and paleogeographic interpretation (Glørstad-Clark et al. 2009).

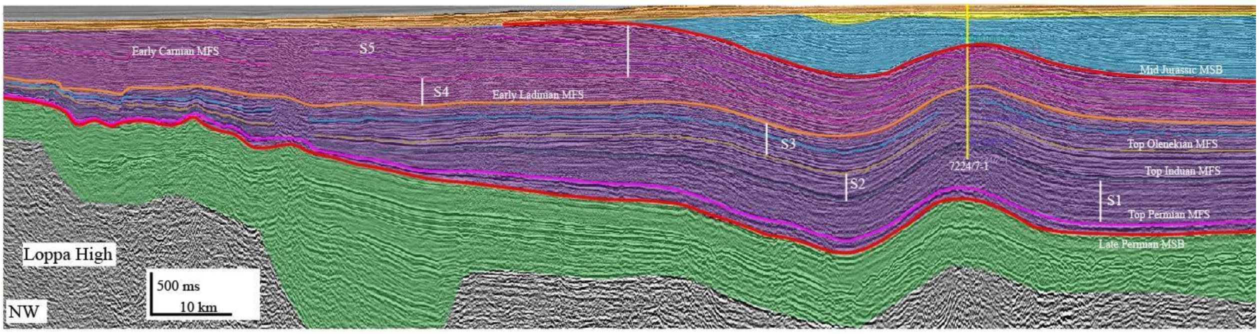


Fig. 3.4: Regional seismic profile in the SW Barents Sea with interpretation of Glørstad-Clark et al. (2009) (from Glørstad-Clark et al. 2009)

The Induan sequence (S1) is bounded at the base and top by maximum flooding surfaces that can be mapped over much of the western Barents Sea. The sediments are sourced from two areas, from the Fennoscandian Shield in the south and later from a more south-eastern source, the Urals. Gradually this prograding system was filling the main depocenter; the Bjarmeland Platform, onlapping the paleo-Loppa High to the west. S1 is pinching out towards west and accommodation space is interpreted to be higher and we have deposition in deeper water starved for sediments towards paleo-Loppa High. Sequence architecture is controlled by late Paleozoic structural elements to some degree, with the paleo-Loppa High representing a barrier to sediments prograding from the east and south. Moreover, the paleo-Loppa High has been uplifted prior to deposition of this sequence, and truncation of the upper part of this unit is also observed associated with uplift late in the sequence (Glørstad-Clark et al. 2009).

The Olenekian sequence (S2) is characterized by a major rise in sea level marks the onset of this sequence. It is bounded at the base and top by maximum flooding surfaces. Both boundaries show seismic downlap on strong amplitude, regionally extensive surfaces. The sequence thins toward north and west, and lapping on to the paleo-Loppa High. S2 thins towards paleo-Loppa High the basal surface is a marine onlap surface. Paleo-Loppa High acted as a barrier to the sediments from east and south-east during this time period (Glørstad-Clark et al. 2009).

The Anisian - early Ladinian (S3) sequence is bounded by strong continuous amplitude surfaces at the top and base, representing maximum flooding surfaces. Clinoforms are observed close to the paleo-Loppa High, the Maud Basin and southern parts of the NW Barents Sea. Clinoforms onlap the paleo-Loppa High, while accommodation space is gradually being filled in east of the high. The Bjarmeland Platform was mostly dominated by sediment bypass at this time and sediments were deposited further west and northwest. A small depositional system is recognized on top of the paleo-Loppa High in the south, where accommodation space west of the high was too great for further progradation. Clinoform successions prograding from west to east are found close to the northern part of paleo-Loppa High, they formed as a result of erosion of locally uplifted and emerged highs (Glørstad-Clark et al. 2009).

The Ladinian-middle Carnian (*S4)* sequence is also bounded by maximum flooding surfaces of regional extent. S4 displays an easterly thinning and development of new local depocenters to the west interpreted to reflect the continuous infill of the Norwegian Barents Sea. Internal clinoforms prograde from southeast- northeast to east-west direction. Paleo-Loppa High is draped and appears to have very little structural expression during time of deposition. Towards the end of this sequence paleo-Loppa High developed into a basin, and the paleo-high was the site of shelfal and marginal marine deposits (Glørstad-Clark et al. 2009).

The Upper Triassic (S5) sequence is bounded by maximum flooding surfaces. Top of S5 is eroded in parts of the area by the Base Cretaceous Unconformity. The sequence is thickening westward and northward. The paleo-Loppa High area was still generating accommodation space, and coarse clastic material is being deposited here. The progressive infilling of the Norwegian Barents Sea was still ongoing and main depocenter moved towards west of paleo-Loppa High, Stappen High and NW Barents Sea. Water-depths at paleo-Loppa High were shallower than in underlying sequences and non marine facies with fluvial channels are present in addition to strong amplitude flooding surfaces (Glørstad-Clark et al. 2009).

The sequences described by Larssen et al. (2005) and Glørstad-Clark et al. (2009) form the basis of interpretation in this study. The horizons have been calibrated through well ties on the eastern side of paleo-Loppa High and correlated on the western side of the main fault. In addition to the existing framework, two additional horizons have been identified; latest early Ladinian; horizon a and middle Ladinian; horizon b. The horizons bound a local sequence that is important for the understanding of the development on the western side of paleo-Loppa High. They are part of sequence S4.

3.2.2 Structural elements

The Loppa High area incorporates the Polheim Subplatform in west which is bounded in east by a major fault zone that have uplifted a north south striking ridge with an eastern tilted flank, further east a monocline limits the Loppa High area. The Loppa high is surrounded and limited by a series of fault complexes, the Asterias Fault Complex in south, and Ringvassøy-Loppa and Bjørnøyrenna Fault Complexes to the west. The Loppa High area to the north is limited by the Svalis Dome, a major salt structure, its associated rim and the Maud Basin (Gabrielsen et al. 1990). *Figure 3.5* shows an outline of the Loppa High area and the surrounding structural elements. In addition, the location of three seismic lines is shown in red. The present day wide Loppa High is a Late Mesozoic structure (Gabrielsen et al. 1990), whereas the Late Paleozoic high, is defined by Glørstad-Clark et al. (2009) as *paleo-Loppa High*, it was a narrow ridge trending north-south under the western part of the present day Loppa High. The paleo-Loppa High area constitutes the Polheim Subplatform, the crest and the eastern flank of the paleo-Loppa High ridge, and a base area constituting the half-graben. The main elements of the paleo-Loppa High is shown in *figure 3.6*, the names that are framed will be used as spatial reference points, and horizon number and names are also included.

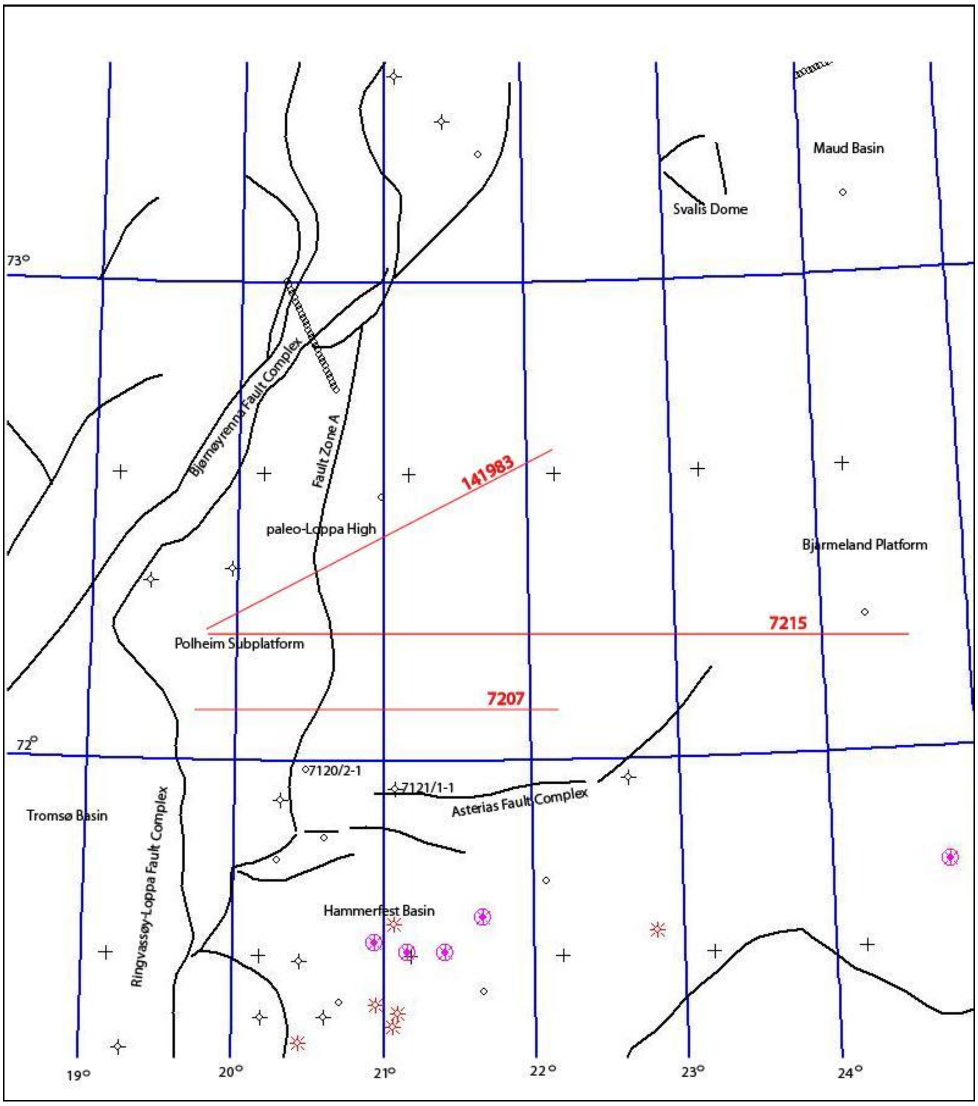


Fig. 3.5: Outline of the Loppa High area and the surrounding structural elements. Location of seismic lines, north: 141983, central: 7215 and south: 7207 marked in red.

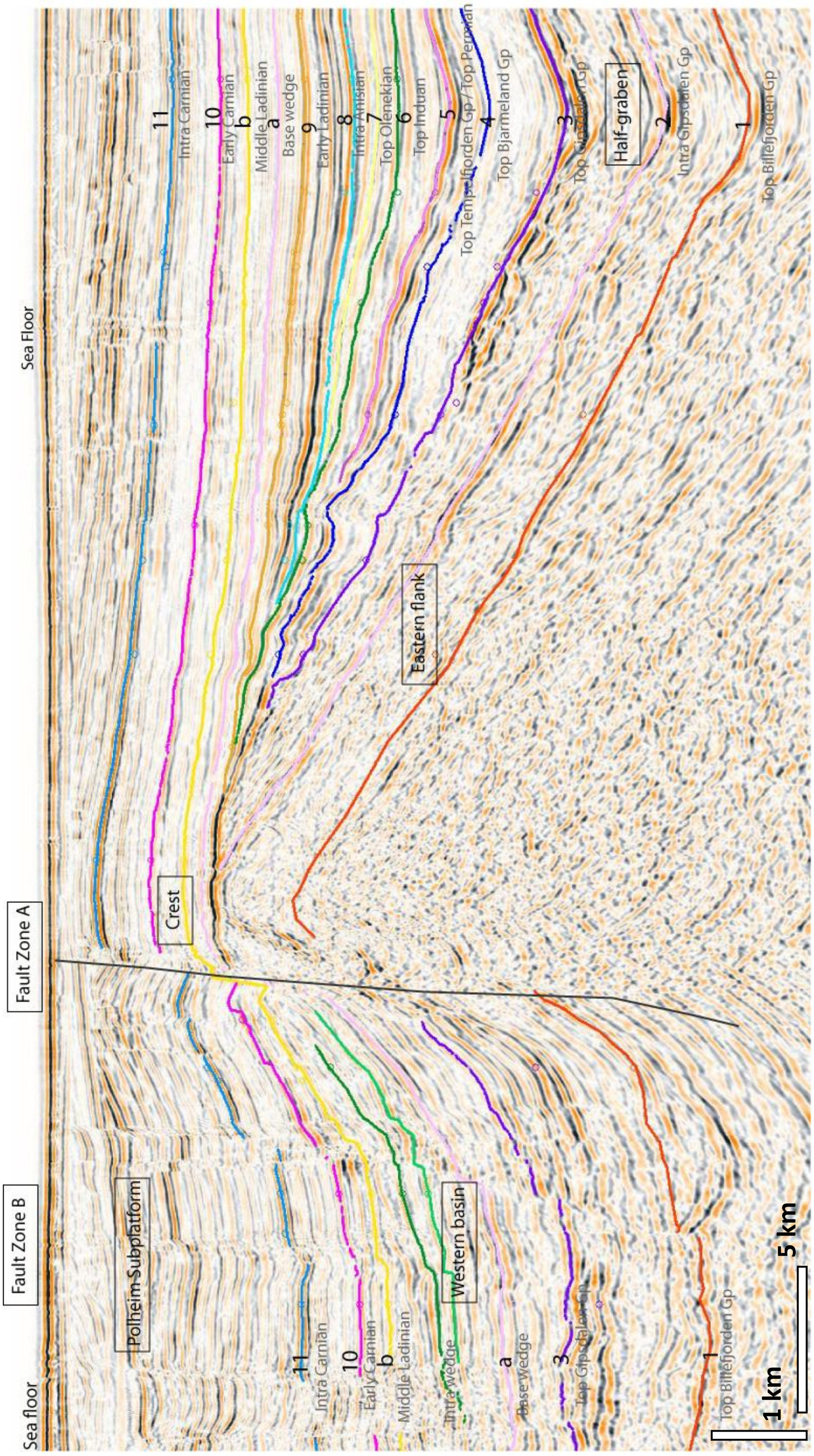


Fig. 3.6: Seismic line 7215 (squashed) showing spacial reference points in frame and a roughly outline of Fault Zone A. Horizon numbers and horizon names are situated at its given reflector.

The paleo-Loppa High area is part of an extensional regime caused by repeated stretching of the lithosphere within the north-eastern Atlantic region prior to the crustal breakup and formation of the Norwegian–Greenland Sea in Cenozoic time (Faleide et al. 1993a,b, 2008). The relief of the paleo-Loppa High is due to footwall uplift in late Permian times, but timing of the events are poorly constrained. It is classified as normal fault systems.

Polheim Subplatform and the eastern flank are separated by a major un-named fault zone. This fault zone is named Fault Zone A in this study and will be further described later in this section. Another half-graben further south in the study area is also identified, and has the same orientation as Fault Zone A. The southern half-graben is situated at the base of the eastern flank of the paleo-Loppa High. The first sign of rifting in the Loppa High area occurs in the half-graben to the south, but died out as the tectonic stress gradually moved westward to Fault Zone A. Polheim Subplatform consists of a blockfaulted area of listric faults with a detachment surface below base Triassic. The fault blocks are rotated (Gabrielsen et al. 1990). Most of the faults are synthetic to Fault Zone A, with smaller faults branches connected. The listric fault blocks on Polheim Subplatform were formed in Late Jurassic-Early Cretaceous and reactivated at later stages (Faleide et al. 1993a,b). In map view Polheim Subplatform has a small extent in the southern part of the study area (*fig. 3.5*). It expands quickly towards the central area before it shrinks northward and merges with Bjørnøyrenna Fault Complex. Before Polheim Subplatform became a subplatform it was a basin receiving sediments during late Paleozoic and Early Triassic times. In this study we will therefore use the term *the western basin* when Polheim Subplatform is discussed as a structure receiving sediments during these times. The eastern flank of paleo-Loppa High is tilted eastward, due to footwall uplift associated with Fault Zone A. On the eastern tilted flank there are several smaller faults within the Upper Paleozoic layers.

The fault zones are shown and described using seismic line 7215 in *figure 3.7*. This is a profile that is situated on the widest part of Polheim Subplatform so the faults have plenty of space and are less complex than the faults that are found north and southwards.

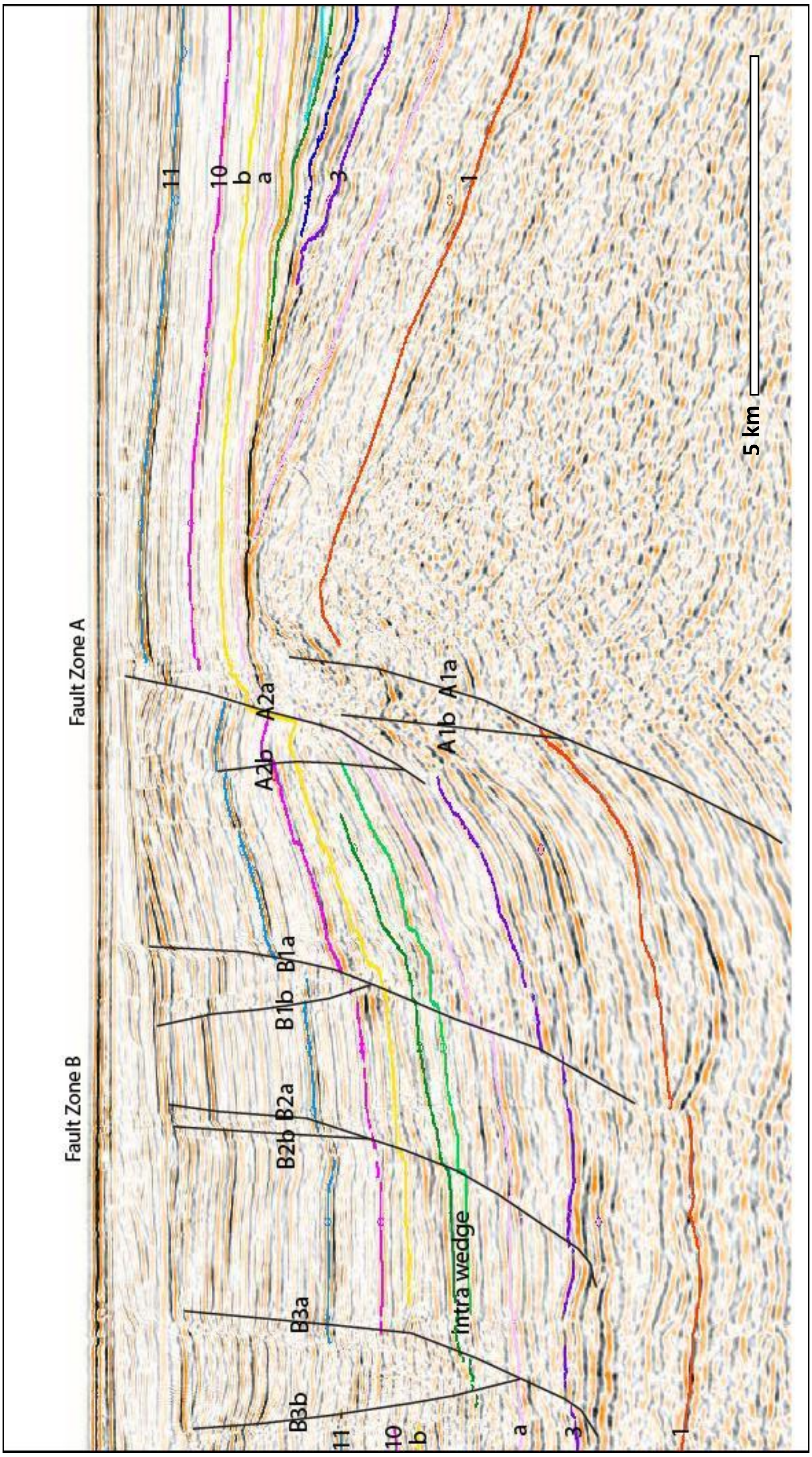


Fig. 3.7: Seismic line 7215, with approximately true geometries 1:1 illustrating the fault zones and the outline individual key faults. Stratigraphic diagram with sequences and horizons are given in *figure 3.9*.

Fault Zone A

Fault Zone A is a series of faults that constitutes the main faults in constructing the characteristic features of paleo-Loppa High. It strikes N-S and separates the crest of paleo-Loppa High and the tilted flank in east from the western basin (*fig. 3.6*). It is dominantly extensional with a top to-the west displacement, but an irregular geometry in some parts indicates a more complex structural development. Fault Zone A undulates northward in map view, and the dip of the fault plane varies. Fault Zone A consists of four main faults; A1a and A1b are connected and A2a and A2b are connected (*fig. 3.7*).

Fault A1a is a planar normal fault with normal drag. It has a top to-the west displacement, and strikes in north-south direction. The fault is the oldest in Fault Zone A, and is the main fault creating the characteristic features of paleo-Loppa High in late Paleozoic. Fault A1a has signs of being basement attached.

Fault A1b is a minor synthetic fault that branches out of fault A1a, at horizon 1. It is a much steeper normal fault with top to-the west displacement. Between the two A1 faults, part of horizon 1 is making a ramp on the hanging wall. The hanging wall is therefore split in two by fault A1b. The thickness of the unit between horizons 1 and 3 is thickening towards the fault plane, indicating syn-tectonic sedimentation.

A2a and A2b forms a younger fault system situated further up in the rock record. A2a is a listric normal fault with a detachment zone within upper Paleozoic strata. It has a top to-the west displacement. Fault A2b is a minor fault that branches out of fault A2a just above horizon 3. It stops its development at horizon 12. Fault A2b has a complex shape where it starts off by being antithetic to fault A2a, however, deeper down it turns and become nearly vertical. The unit between horizon a and b is thickening towards the fault plane of A2a and A2b. This feature is especially visible between the two green intra wedge horizons, indicating that there might have been fault activity in middle Ladinian times. Fault A2a shows signs of activity in Paleocene strata. In the upper part of the geologic record constituting the Polheim Subplatform the characteristic feature (typical of a listric fault) of a roll-over structure is present.

At a later stage a modification of the listric normal fault occurs in Fault Zone A2. The layers between fault A2a and A2b bend upwards, with a shorter wavelength than the listric fault. A compressional force reactivated the former normal listric fault and forces it to reverse. An erosional surface has removed sediments younger than Late Triassic, and Polheim Subplatform is covered by Paleocene strata. From regional knowledge we know that compressional forces have been observed in the Asterias Fault Complex in Late Jurassic-Early Cretaceous (Gabrielsen et al. 1990), and inversion tectonics observed at paleo-Loppa High could be tied to the same event.

Fault Zone B

Fault Zone B consists of three listric faults; B1a, B2a and B3a all having a top to-the west displacement. They have a detachment zone within upper Paleozoic strata, and share many of the same features as fault A2a (*fig. 3.7*). Fault Zone B makes out a tilted block faulted area also known as Polheim Subplatform, defined by Gabrielsen et al. (1990).

B1b is a minor antithetic normal fault, which branches out from the larger fault B1a at horizon b. The two B1 faults forms a V-shape and a small graben structure. B2b is nearly a vertical fault that branches out of the larger fault B2a at horizon 11. The horizontal distance between fault B2a and B2b is very small compared to the other faults discussed. B3b is a antithetic normal fault that branches out from fault B3a. The horizontal distance between the two faults is large and they form a graben structure.

The sedimentary unit between the two intra wedge horizons may increase in thickness towards fault B1a and B2a, indicating syn-tectonic sedimentation and tectonic activity in middle Ladinian times. The sedimentary unit between the intra wedge horizons is decreasing in thickness westward on the platform. This indicates that the faulting might have been largest at fault A2a, and decreasing towards the following faults, B1a, B2a and B3a. Fault B1a and B3a shows signs of activity through Paleocene strata.

Because of erosion of post Paleocene strata it is hard to date the latest tectonic activity in Fault Zone B. Sedimentary successions have a relatively constant thickness from horizon b and younger, indicating a calm tectonic period in Middle Triassic. Main activity was in Late

Jurassic-Early Cretaceous followed by minor reactivation in late Cretaceous? and Paleogene times.

During the formation of the Ringvassøy-Loppa Fault Complex the Triassic and Jurassic sediments slipped westward contributing to the formation of the tilted fault blocks in Late Jurassic-Early Cretaceous times (Gabrielsen et al. 1990).

Figure 3.8 shows three seismic lines for comparison, all with an approximately 1:1 geometry. Profile A is located to the north, with B situated in central areas, and C located to the south (location of lines shown in map view in *fig. 3.5*). Fault A1a seems to be the main fault throughout the area, the dip changes along the fault zone but it remains the most important fault concerning the construction of the characteristics of paleo-Loppa High. The slip component of fault A1a is largest in north and decreases towards south. The fault characteristic of Fault Zone B seems to be more random, but many of them share a listric feature.

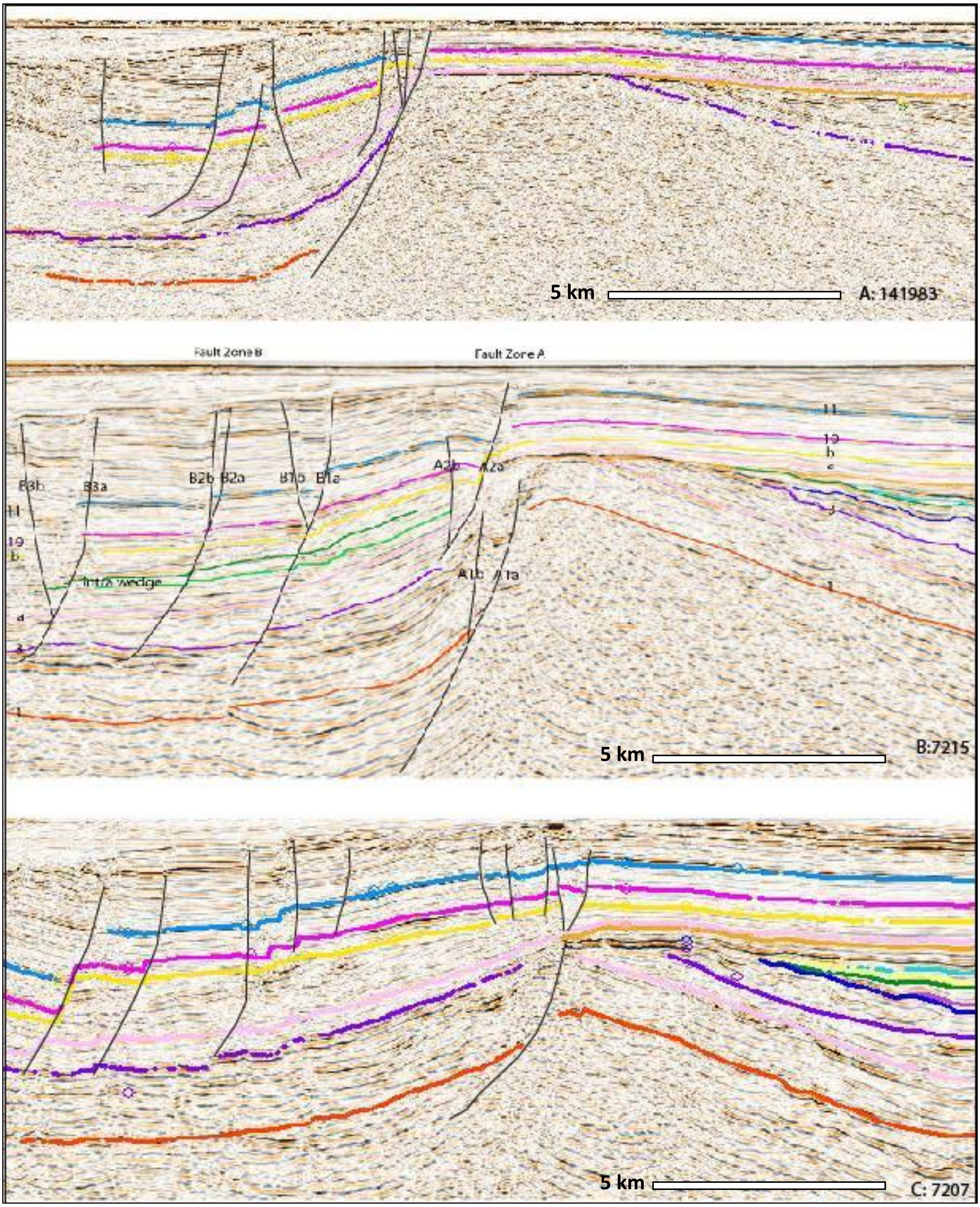


Fig. 3.8 A: Seismic line 141983, B: 7215 and C: 7207. A structural comparison of three seismic profiles in the paleo-Loppa High area. The locations of the seismic lines are shown in *figure 3.5*.

3.2.3 Sequence boundaries and seismic sequences

Table 3.2 shows the mapped horizons east of Fault Zone A of the paleo-Loppa High, listed from the top with the youngest horizon. Furthermore, it lists the horizons identified and mapped out west of the paleo-Loppa High.

Horizons defined with numbers are interpreted by Glørstad-Clark et al. (2009) and Larssen et al. (2005), while the two horizons with letters a and b represent additional horizons that are added in this study due to their relevance to the geological development on the western side of paleo-Loppa High. The names of the horizons were given by Glørstad-Clack et al. (2009) and Larssen et al. (2005). All horizons numbers and horizon names are shown on a seismic key profile in *figure 3.6* in addition to the outline of Fault Zone A, and the names used as spatial reference points when discussing the geology.

Color	Horizon name	Reflectors eastern side	Reflectors western side
Lavender Blush	Sea Floor	13	13
Dodger Blue	Intra Carnian	12	12
Magenta	Early Carnian	11	11
Goldenrod Yellow	Middle Ladinian	b	b
Light Pink	Base Wedge	a	a
Goldenrod	Early Ladinian	10	
Turquoise	Intra Anisain	9	
Light Yellow	Top Olenekian	8	
Spring Green	Top Induan	7	
Orchid / Pink	Top Tempelfjorden Gp	5 and 6	
Blue	Top Bjarmeland Gp	4	4
Violet	Top Gipsdalen Gp	3	
Light Pink	Intra Gipsdalen Gp	2	
Red	Top Billefjorden Gp	1	1

Table 3.2: Horizons mapped out east and west of Fault Zone A, given with color, horizon name and number.

Figure 3.9 illustrates the sequences and horizons identified in this study and place them according to the geological timescale. *Figure 3.10* shows the sequences mapped in this study and their development throughout seismic line 7215, which is a representative profile of the paleo-Loppa High development.

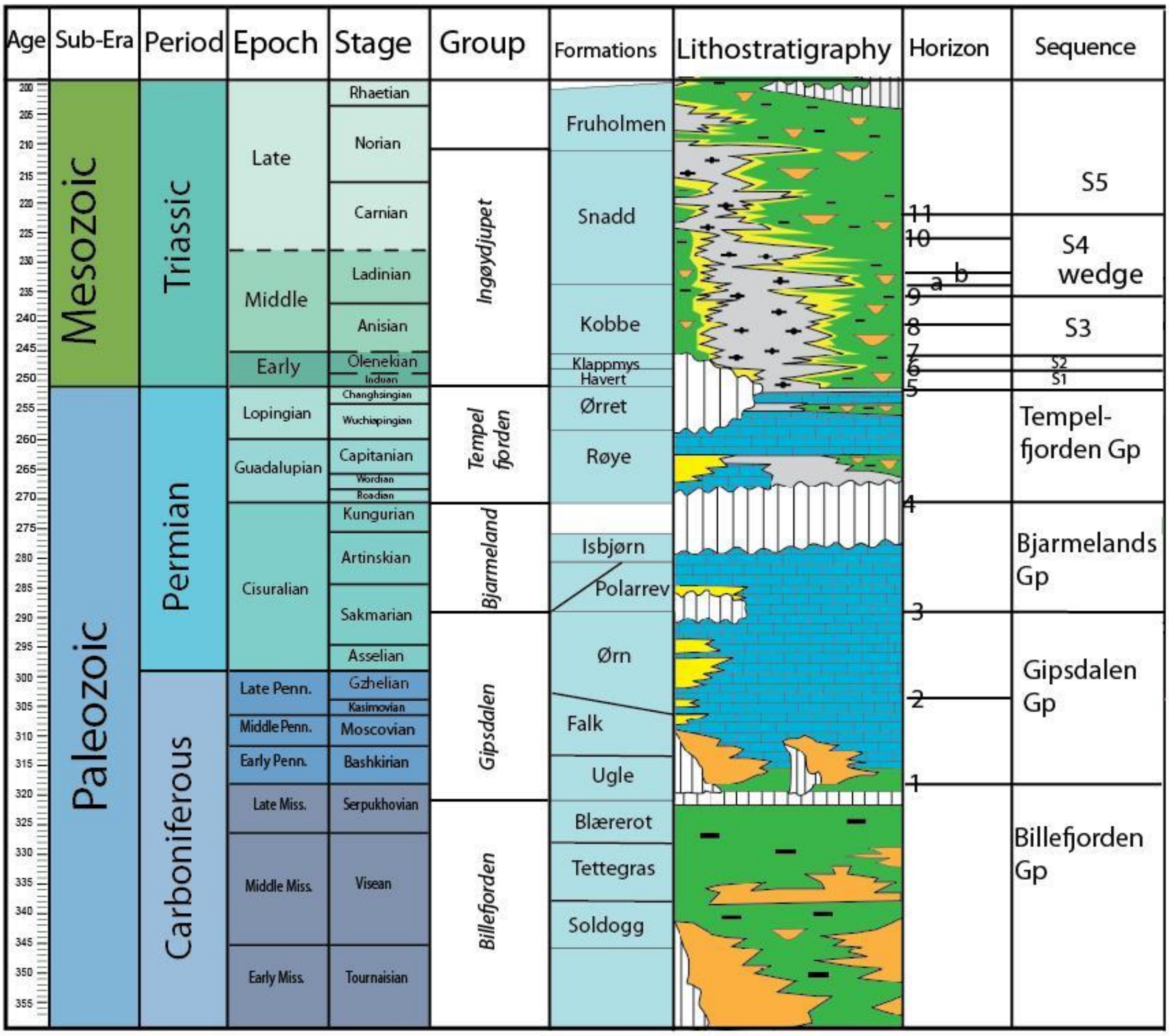


Fig. 3.9: Stratigraphic diagram with horizon numbers and sequences related to time and regional lithostratigraphy.

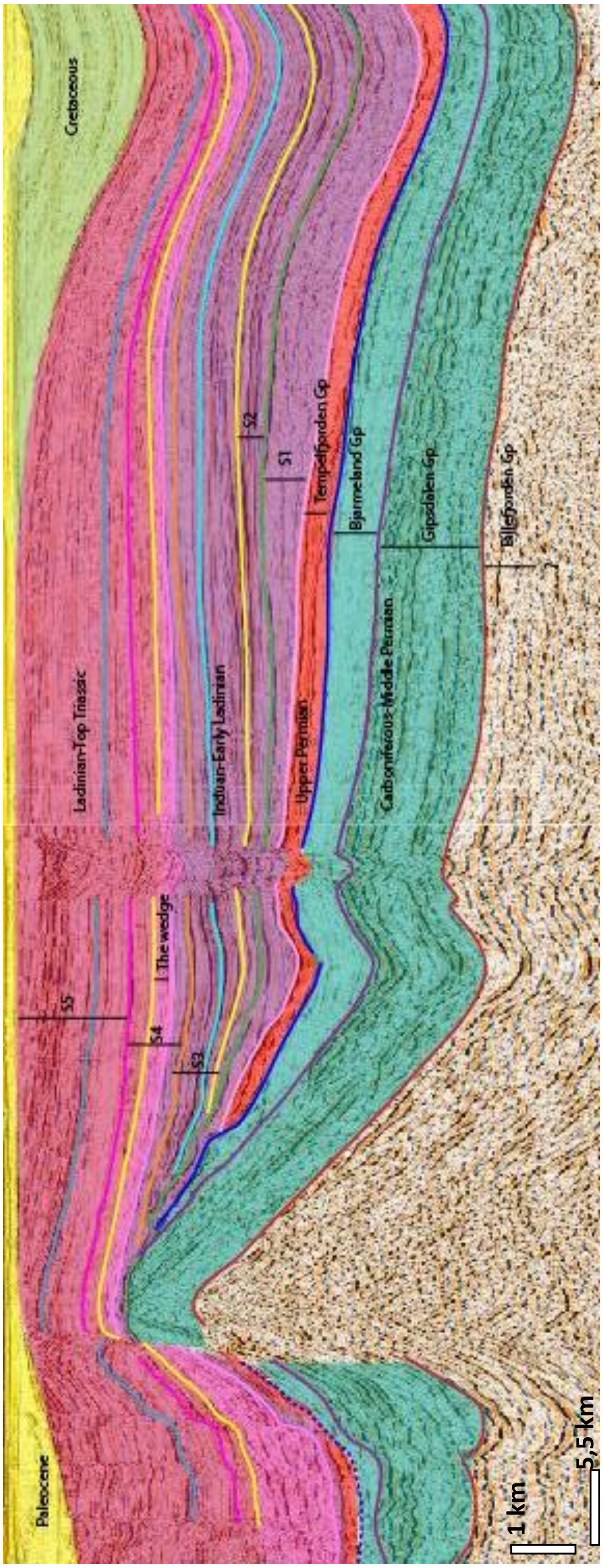


Fig. 3.10: Regional seismic profile (line 7215) in the paleo-Loppa High area with the interpreted sequences of this study. Stratigraphic diagram with sequences and horizons are given in *figure 3.9*.

The units of Upper Paleozoic are defined by Larssen et al. (2005) on lithostratigraphic criteria. Lithostratigrahic units are grouped and separated on the basis of similarities in lithology and/or lithofacies, but does not necessarily consider that these rocks accumulated over a specific time period. Seismic sequence stratigraphy combines the chronological order of the accumulation of sediments, their stratal architecture and the geometric relationships of their seismic facies to determine depositional setting and predict stratal continuity. Hence, a sequence stratigraphic package bounded by two maximum flooding surfaces could constitute of several lithostratigraphic units, depending on lateral and vertical position within the sequence. Consequently it is very important to separate the two methods. In this study we have made a deliberate choice to define and correlate the Upper Paleozoic seismic units to their lithostratigraphic units and naming convention, whereas the Triassic successions is set in a sequence stratigraphic framework with names from Glørstad-Clark et al. (2009). The reasoning behind this is due to the laterally extensive carbonates found in the Upper Paleozoic successions deposited within a defined time frame. Moreover the lithostratigraphic boundaries correlated well to the seismic facies interpreted in the Loppa High area.

3.3 Seismic interpretation procedure

The initial approach of this study was to get a basic understanding of the area of interest through seismic interpretation on paper. Several profiles were selected to identify main structural elements and important seismic horizons.

The interpretation was further carried out using Kingdom software on workstation. Horizons mapped out by Glørstad-Clark et al. (2009) and Larssen et al. (2005) were identified on the eastern side of Loppa High. In order to better correlate the horizons over Loppa High Fault Zone A we searched along the fault zone for profiles with good resolution and low fault displacement. A few profiles were found where horizons could be traced over the main fault (*fig. 3.8.C*) and then be applied as references for further correlation where the fault displacement is greater. Horizon 1 (Top Billefjorden Group) and horizon 3 (Top Gipsdalen Group) which are horizons that are eroded at the crest of paleo-Loppa High and deeply buried in the western basin were impossible to trace in this way. Seismic facies analyses were then used to correlate and identify these horizons. By creating arbitrary lines and looking at the seismic characteristic of each horizon we were able to map out key horizons throughout the study area.

3.3.1 Seismic facies analysis

Seismic facies analysis is an interpretation procedure one step beyond seismic sequence analysis (Boggs, 2006). This technique was appropriate to this study as we have zoomed in on the evolution of paleo-Loppa High, making it hard to map out the large sequence stratigraphic patterns. The data are studied for the purpose of extracting stratigraphic information. Depositional facies have characteristic features to identify them and relate them to geological factors responsible for the reflections. The characteristics of a reflector can be related to depositional features such as lithology, bed thickness, spacing and continuity. The interpretation made out of these depositional observations may have different meanings, this is when background knowledge from i.e. earlier studies, wells and outcrops are important. A reflection is thought of as a time-line that represents a time surface in three dimensions. But one should be aware of seismic artifacts such as multiples and reflected refractions.

Three steps were used to interpret seismic facies; 1. The geometry of reflections and reflection termination, 2. Reflection configuration, 3. Three-dimensional shape (Boggs, 2006).

1. The geometry of reflections and reflection termination were studied by seismic stratigraphic methods. Studying how surfaces terminate (onlap, downlap, toplap etc.) enable us to categorize seismic data into depositional packages. The depositional packages illustrate depositional environment and relative time. The reflection terminations are set into a system based on geometry and some interpretation. The terms applied in this study to describe reflection terminations follow the nomenclature in Emery and Myers (1996), and references therein.

2. The reflections also display a gross pattern that is recognizable on seismic data, referred to as reflection configuration. A depositional package can show *parallel* and *sub-parallel* layering, indicating sediments being deposited at a uniform rate on a uniformly subsiding shelf or stable basin. *Divergent* layering indicates lateral variations in rates of deposition, or progressive tilting of the sedimentary surface

during deposition. A *chaotic* pattern is result of a deformation process (Boggs 2006), but will in this study be used as a term for a pattern that displays no identifiable stratal reflections. *Clinoforms* are strata deposited by lateral outbuilding or progradation, forming gently dipping surfaces (Boggs 2006). Keep in mind that this is a local study in a limited area, and the reflection configuration defined from seismic data in the paleo-Loppa High area is not the trend in a regional view.

3. Three dimensional shape are mapped out when the seismic facies are arranged into depositional packages. Thickness variations and extent are taken into consideration. The terms used to describe shape in this study are wedge and delta.

Other important sequence stratigraphic terms to describe the sedimentary succession of this study are;

Accommodation space is the space available at a given time where sediments can accumulate. In a marine environment this is governed by sea level rise and fall and subsidence and uplift related to tectonism. Rise in sea level and subsidence increases the accommodation space, while sea-level fall and uplift decreases accommodation space (Boggs 2006).

Flooding surface are surfaces that displays an abrupt increase in water depth and separates younger from older strata. They commonly form in areas were the rate of increase in accommodation space is less than the rate of sediment input (Boggs 2006).

Angular unconformity is a surface where younger rocks rest upon an eroded surface which has been tilted or folded. The older rocks strata therefore commonly obtain a steeper angle.

The vertical resolution of the seismic data can be defined as the minimum vertical distance between two surfaces to give rise for one single reflection and is a limiting factor in seismic stratigraphy. A seismic trace have a resolution of ¼ the wavelength, in a noise-free environments, hence, the shorter the wavelength and the higher the frequency the greater vertical resolution. Reflectors spaced closer than ¼ of the wavelength will give rise to interference that will either enhance the reflection (positive interference) or decrease it (negative interference). Acoustic velocity and wavelength of a seismic signal increases with depth due to compaction and increased cementation (Emery and Myers, 1996).

Color improves the interpretability of sections, and has been used to study details in this thesis.

Squashing the seismic line is a another method that can enhance different details in the interpretation. Clinoforms often have very low dip angles, and can be hard to identify. By squashing the plot geometries like this become more visible, but it is important to keep in mind that they are exaggerated.

Flattening of an interpreted horizon removes the impact of younger strata. Although a horizon is not always deposited as a flat surface, it does give a hint of the geometries at the horizon of current interest.

3.3.2 Chronostratigraphic chart

Seismic stratigraphy is a method to interpret the relationship of depositional systems in time and space. Chronostratigraphic charts illustrate their time relationship, their relationship to surfaces of non-deposition, condensation and erosion. Non-deposition, condensation and erosion are features that have little or no thickness in the rock-record, but their significance and extent can be studied in the time dimension. The chronostratigraphic chart has time on the vertical axis and a spatial horizontal axis, on the chart the distributional facies are plotted, bounded by onlap, toplap and downlap surfaces. The remaining part of the chart represents the position and duration of non-deposition, hiatus, bypass, erosion and condensation (Emery and Myers, 1996).

Seismic line 7215 was used for the construction of the chronostratigraphic chart. It is of good quality and representative of the characteristic geology of the area. In addition general knowledge about the area has been incorporated. The horizons used in the chronostratigraphic chart are marked by colors and extended into the geological timescale, and are the same horizons used earlier in this study defined by Larssen et al. (2005) and Glørstad-Clark et al. (2009), in addition to the two horizons defined in this study. The chronostratigraphic chart shows periods of deposition versus non-deposition in relation to age and lateral extent, it does not reproduce the thickness of each sequence. The Triassic sequences are very complex and detailed, so a zoom has been created of this part to take a better look at the details.

3.3.3 Backstrippping

Seismic line 7215 was also the basis for the reconstruction of the geological history of the Loppa High area, using the software BMT (Basin Modelling Toolbox) which perform backstripping and structural restoration. The aim of the modeling was to represent the geology we see today, to reconstruct the geology at different time steps related to the interpreted horizons, and to give a simple overview of the subsidence through time. The input information was the interpreted seismic horizons in time, tied to wells 7121/1-1 and 7120/2-1. Faults-zones were mapped out and put into the model. The block-faulted Polheim Subplatform was simplified into two main faults. They both consist of several smaller antithetic and synthetic connected faults. The same applies for Fault Zone A, which was simplified into one fault. The faults on the eastern flank are two younger faults that fumbles with the sediments located between them, but is not significant respect to the major geological evolution of the paleo-Loppa High. All seismic horizons that represent the main input data to the backstripping model were in time originally, but were subsequently depth converted. Simplified lithologies of the sequences were put into the model, grouped into two major categories, represented by carbonate and clastic rocks. Paleo-water depths were included in the model to calculate the extent and amount of erosion. Rifting events are also input information to restore the subsidence. The information creates a model of the study area that restores the geometry in specific time steps tied to the interpreted horizons. Elements of uncertainty are the depth conversion, the seismic interpretation and the assignment of lithology, the amount and extent of erosion, paleo-water depths and the timing of tectonic events. In other words; all input information is based on interpretation of seismic, hence, the model is not to be viewed upon as a true reconstruction of the geological history. The aim of the model is to illustrate and further understand the consequences of the interpretations made on seismic data.

Analyzing the subsidence of the study area is important and the most complex part of this model. The total subsidence at a horizon of interest can be split into two components; the sediment and water load component and the tectonic driving force component. The sediment and water load through time constitute the isostatic subsidence. The isostatic correction is calculated using the Airy model, which is a good approximation in areas of active tectonics.

The tectonic subsidence is due to rifting. By removing the sediment and waterload of all the younger sediments the profile is decompacted and we can restore the true geometries at the time of the horizon of interest (Allen and Allen 2005). *Figure 3.11* shows the result of the modelling.

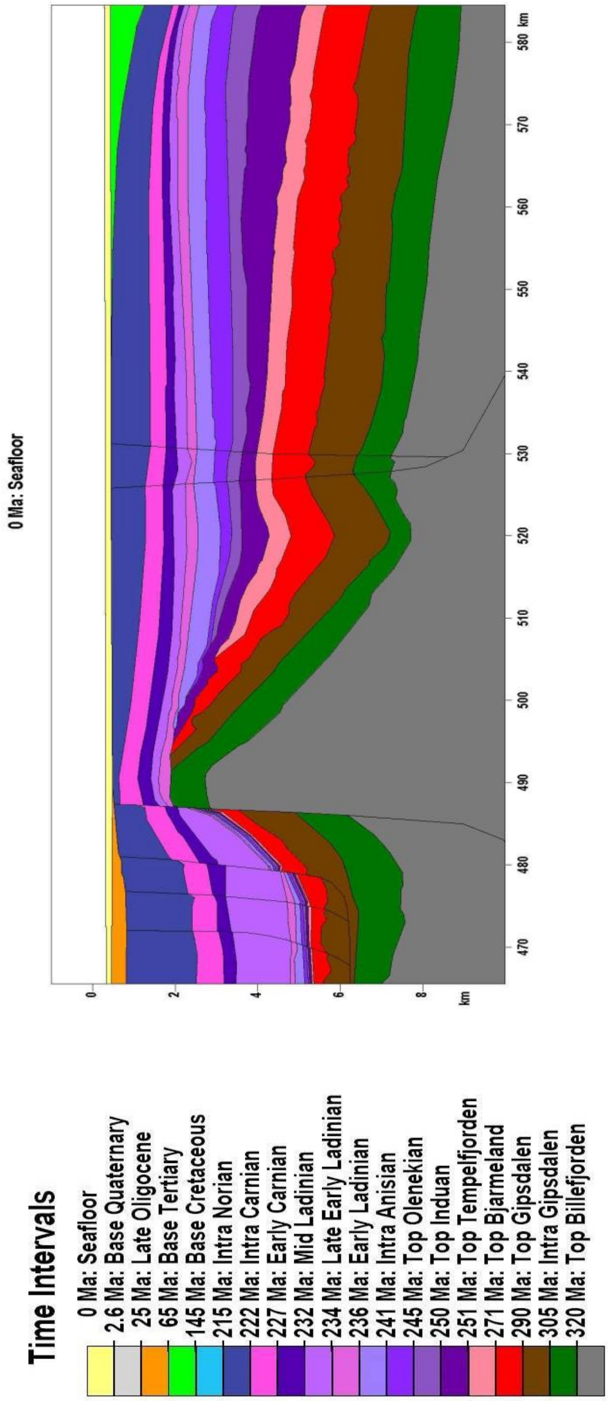


Fig. 3.11: Model of the paleo-Loppa High and legend showing color codes, age and horizon name.

3.3.4 Time-structure and time-thickness maps

Two types of maps were generated in Kingdom from the interpreted horizons. Time-structure maps were made to illustrate the structural relief/topography of selected horizons. The horizons are interpolated between each seismic line resulting in a time-structure map.

The time-thickness map illustrates the thickness in two-way travel time between two interpreted horizons. First the horizons are interpolated and made into time-structure maps, and then the thickness is calculated by subtracting the upper lying surface from the lower lying surface. The time-thickness map shows the quantity distribution of sediments within a sequence, and illustrates the main depocenters.

3.4 Sequence description and interpretation

As mentioned earlier, the work of Larssen et al. (2005) and Glørstad-Clark et al. (2009) were used as basis for the seismic mapping of the late Paleozoic and Triassic successions of the western side of paleo-Loppa High. The existing framework was not sufficient to explain the basin evolution in this part of the study area, hence, two additional horizons were identified based on seismic facies analysis. The additional horizons helped our understanding of when paleo-Loppa High became a positive structural element and when accommodation space west of the paleo-high was filled. This study is limited from horizon 1 which is the top of the Billefjorden Group in base, and an upper composite unconformity on top. The poor seismic resolution at depth of the Billefjorden Group (and partially the Gipsdalen Group) limits the possibilities of a good interpretation.

3.4.1 Billefjorden Group (Late Devonian-early Carboniferous)

The Billefjorden Group was deposited from Late Devonian to early Carboniferous. The Bjørnøya and Spitsbergen development of this sequence is dated to Viséan to early Serpukhovian age (Larssen et al. 2005). No well information exists to determine a more accurate time interval of deposition in the Loppa High area. The only well that reaches this unit in the Loppa High area is 7120/2-1 and it is not able to distinguish any formations within the group. In well 7120/2-1 it has a thickness of 847 meters and lies upon basement. The sequence is bounded on top by horizon 1, while the base boundary is not mapped out because of poor seismic resolution. The unit lies deeply buried, thus the poor seismic resolution, but we can distinguish some sub-parallel reflectors and a more chaotic character in the lower part of the group. At a much later stage the Billefjorden sequence is downfaulted west of Fault Zone A and uplifted east of Fault Zone A. Up the eastern flank of paleo-Loppa High some sub-parallel reflectors is visible, but otherwise the unit has a chaotic structure here as well eventhough it is situated more shallow.

The Billefjorden sequence is believed to have an equal thickness over the whole area, thus, indicating that it has been deposited on a relatively flat topography. At the crest of paleo-Loppa High the group is folded and then cut of at Fault Zone A. We correlate horizon 1 (top Billefjorden Group) from the eastern side of paleo-Loppa High across the faulted area and to the basin on the western side. The thickness of the overlying Gipsdalen sequence and seismic facies were considered when interpreting horizon 1 on the western side of Fault Zone A.

3.4.2 Gipsdalen Group (middle Carboniferous-early Permian)

The Gipsdalen Group was deposited from middle Carboniferous to early Permian times. At Loppa High it is dated, from well 7120/2-1, to be deposited from late Serpukhovian/early Bashkirian to early Sakmarian times. The Gipsdalen Group is split into three formations all present in the Loppa High area; Ugle, Falk and Ørn formations (Larssen et al. 2005). The sequence is bounded by horizon 1 at the base and 3 on top, and has an internal horizon 2; Intra Gipsdalen (*fig. 3.9*). Horizon 2 is a high reflectivity band especially visible on the paleo-Loppa High eastern flank. On the western side of Fault Zone A we can not find the same features and horizon 2 is therefore not interpreted there. The lower part of this unit has an equal thickness throughout the area, while the upper part of the Gipsdalen Group thickens in the half-graben and thins on the eastern flank of paleo-Loppa High. It is faulted by Fault Zone A, and the eastern flank is uplifted and tilted. The internal seismic reflectors vary from high reflectivity bands representing carbonate facies to more transparent sub-parallel to chaotic reflectors, some downlapping clinoforms on horizon 2 are observed up the eastern flank.

Onlapping geometries are observed on horizon 3, within the half graben. Horizon 2 and 3 is truncated at the crest of paleo-Loppa High. *Figure 3.12* is a seismic profile showing some of the locations where onlap, downlap and truncation have been interpreted. The interpretations in the figure are collected from several profiles.

Fig. 3.12: Seismic line 7215 showing interpreted onlap and downlaps. Stratigraphic diagram with sequences and horizons are given in *figure 3.9*.

The Gipsdalen sequence has a two-step development divided by horizon 2; Intra Gipsdalen. The lower Gipsdalen sequence is bounded by horizon 1 in base and 2 in top, while the upper Gipsdalen sequence is bounded by horizon 2 in base and 3 on top *(fig. 3.6).* The lower Gipsdalen Group has an equal thickness over the whole area east of Fault Zone A. The upper Gipsdalen Group is generally thinning from the Bjarmeland Platform towards paleo-Loppa High, has an increased thickness in the half-graben located south in the study area, and further continues to thin along the eastern flank of paleo-Loppa High. Thus, the half-graben structure is formed in a tectonic event that coincides with horizon 2. Horizon 2 is not indentified west of the main fault, but there is no reason for it not to be there. Small changes in depositional environment, topography or the poor seismic resolution due to deeper burial in the western basin may change the seismic signature of horizon 2. Tectonically, horizon 1 coincides to a rift phase generating rift basins in the western Barents Sea (Gudlaugsson et al. 1998). The top Gipsdalen Group, horizon 3, is calibrated from well-ties to be on the high reflectivity band found on the Bjarmeland Platform. It is further correlated over to the western side of paleo-Loppa High. Horizon 3 might be a sub aerial exposure surface (Larssen et al. 2005) that has been eroded.

The high amplitude belt just underneath horizon 3 has pillow-shaped features towards the Bjarmeland Platform, which is due to carbonate build-ups. The upper part of this group has parallel horizons. Asselian times are characterized by high frequency and high amplitude eustatic sea-level changes (Smelror et al. 2009). Thus, the clear layering stems from carbonate production cycles. The low frequency part of the group has a more chaotic to sub parallel appearance.

3.4.3 Bjarmeland Group (early Permian-late Permian)

Based on well 7121/1-1 south on Loppa High, the Bjarmeland Group is suggested to be deposited from middle Sakmarian age. Top of the group from various wells is set to be late Artinskian age. The Bjarmeland Group is separated into three formations; Ulv, Polarrev and Isbjørn, all drilled through by well 7121/1-1. Same well have measured the Bjarmeland Group from 3990 m to 3502 m and have a total thickness of 488 m, which is the maximum thickness measured in any wells. The Bjarmeland sequence is limited by horizon 3 in base and 4 on top (*fig. 3.9*). The unit has an equal thickness in the study area, except in the half-graben were it thickens before it thins on the eastern flank of paleo-Loppa High. The top boundary; horizon 4 is truncated at the crest and down the eastern flank of paleo-Loppa. Top Bjarmeland horizon is not identified on the western side of paleo-Loppa High, but it is likely to believe that it is there. On seismic profiles this unit has a transparent appearance (*fig. 3.12*). The Bjarmeland sequence downlaps on the eastern flank. Initiating small or no relief of paleo-Loppa High at the time of deposition. In the half-graben we see onlap while the rest of the unit towards the Bjarmeland Platform has a chaotic to sub-parallel appearance.

The Bjarmeland sequence has an equal thickness on the eastern side towards the Bjarmeland Platform. It thickens in the half-graben due to former tectonic activity enhancing the structure. The Bjarmeland Group thins on the eastern flank due to some structural expression, in addition to erosion of the top bounding horizon 4, due to a later uplift event. The Bjarmeland sequence is faulted by Fault Zone A, the eastern flank is uplifted and tilted.

The Bjarmeland Group probably draped the small paleo-Loppa High structure and was deposited as a unit over the whole area, both west and east of the main fault. The unit is thick and cut off at the crest of paleo-Loppa High indicating that it continued to be deposited on the western side of Fault Zone A. On the western side of paleo-Loppa High we have not indentified the top Bjarmeland horizon. The seismic facies separating the Bjarmeland Group and the Tempelfjorden Group are different in the basin west of Fault Zone A due to changes in sedimentation related to a tectonic event. An extensional event between the Bjarmeland and the Tempelfjorden Groups alters the area, represented by footwall uplift of the paleo-Loppa High, creating the ridge described by Gudlaugsson et al. (1998). The Tempelfjorden Group was deposited in the basin east of the paleo-Loppa High as a result of this event, lapping onto the structurally higher paleo-Loppa High. In general paleo-Loppa High became a barrier to sediments from this time period to Middle Triassic (Glørstad-Clark et al. 2009), thus hindering sediments to be transported to the western basin. This makes horizon 4, top Bjarmeland, hard to identify in the western basin, because it does not have the same appearance as horizon 4 on the eastern side of paleo-Loppa High. Although we do not find the top bounding horizon 4 in the western basin we have reasons to believe that most parts of the Bjameland sequence are present there. Because of the truncated internal reflectors at the crest of paleo-Loppa High and the transparent appearance of the unit deposited on top of the

Gipsdalen Group on both sides we correlate the western part to consist of nearly complete Bjarmeland Group aswell.

3.4.4 Tempelfjorden Group (middle Permian-late Permian)

The Tempelfjorden Group was deposited from middle to late Permian times and consists of two internal formations; Røye and Ørret (Larssen et al. 2005). In well 7121/1-1 only Røye Formation is present. Røye is deposited during Guadalupian times (Larssen et al. 2005). The Tempelfjorden Group is bounded by horizon 4 at the base and 5 on top (*fig. 3.9*). It thins towards the Bjarmeland Platform and has a thickness of 500-600 m in the Loppa High area measured from wells 7121/1-1 and 7120/1-1. The sequence has parallel layering and might be truncated at the crest of paleo-Loppa High. According to Larssen et al. (2005), the Tempelfjorden Group was uplifted and truncated in late Permian. The Tempelfjorden sequence is thicker in the half-graben structure and display downlaps, associated with progradation of clastic deposits, and depositional thinning and onlap to the eastern flank of paleo-Loppa (*fig. 3.12*). The base of the group is established after a major transgression (Larssen et al. 2005), and is seismically mapped out on a strong amplitude horizon; 4. Top of the group is marked by a downlap surface towards the Bjarmeland Platform and given the number 5. This horizon marks the top of Permian sediments.

The Tempelfjorden sequence thins drastically and onlaps the paleo-Loppa High eastern flank. In the western basin the group is not identified, but it is likely to believe that it is present and very thin. During deposition of the Tempelfjorden Group the paleo-Loppa High acted as a barrier to sediments and was an island interrupted by small periods of flooding. Thus, sediments with same origin as the Tempelfjorden Group have for most times not been transported to the western side of the Fault Zone A, where sediment starvation prevails in this time period. However, there is a chance there was a passage to the west somewhere else in the area, but this will be discussed in a later section.

3.4.5 Sequences S1, S2 and S3 (Induan, Olenekian, Anisian to early Ladinian)

Sequences S1, S2 and S3 will be discussed in one section due to similar geological development in respect to the paleo-Loppa High. S1 was deposited during Induan times, S2 during Olenekian and S3 from Anisian to early Ladinian (Glørstad-Clark et al. 2009). S1 is bounded by horizon 5 in base and 6 on top (*fig. 3.9*). Horizon 5 marks the beginning of Triassic times. S2 is bounded by horizon 6 in base and 7 on top (*fig. 3.9*). S3 is bounded by 7 in base and 9 on top, and has an internal horizon; 8, which represent an Intra Anisian horizon (*fig. 3.9*). All horizons in these three sequences are maximum flooding surfaces. None of the sequences are recognized in the basin west of paleo-Loppa High. The thickness of the three sequences on the eastern flank varies in north-south direction, but they all thin up the eastern flank and eventually pinch out or are truncated. A decrease in accommodation space due to a larger relief, forced the depositional pattern to change.

Paleo-Loppa High acted as a barrier to sediments during deposition of S1, S2 and S3. The sequences thickness thins dramatically up the eastern flank. The sediments onlap paleo-Loppa High and locally drape the crest in response to rise in relative sea-level. The flooding events was associated with the basal maximum flooding surfaces prior to the main deposition within the sequences. Typically, the sediments deposited at the crest were later eroded.

The sequences S1, S2 and S3 are characterized by seismic downlap east of the paleo-Loppa High, and depositional onlap up the flank (*fig. 3.13A and B*). Horizon 6; (top Induan), locally onlaps the paleo-Loppa High at the crest, other places further down the eastern flank, and other areas are characterized by a truncation of the surface. This reflects local variation in the topography of the paleo-Loppa High. The degree of how far towards the crest the reflectors onlap the high varies, but accommodation space is gradually being filled in to the east during deposition of these sequences. In response to local uplift of the paleo-Loppa High and changes in relative sea-level, some of the horizons onlap older horizons, as illustrated by horizons 7 (Top Olenekian) and 8 (Intra Anisian) lapping onto horizon 6 (Top Induan). In contrast horizons 5, 6 and 9 are onlapping the crest successively higher. Small west-east prograding systems are found close to paleo-Loppa High in all three sequences, they are steeply dipping.

Figure 3.13.B is flattened on the early Ladinian horizon, showing a clinoform unit prograding in west-east direction at the same time as a system prograded from the east to the west in late Anisian.

Fig. 3.13 A: Seismic line 7217, with interpreted onlap and downlap geometries. B: Seismic line 7245 flattened on horizon 9. Horizons 5 to 8 are truncated by horizon 9. West-east prograding clinoforms interpreted underneath horizon 9. Stratigraphic diagram with sequences and horizons are given in *figure 3.9*.

A larger prograding wedge with low, angle clinoform units prograding from the west to east is observed in the upper part of sequence 3. In general, horizon 9 were the last horizon to onlap the flank, except for a small area to the south where a small depositional system prograded from east to the crest of paleo-Loppa High. This progradational unit is interpreted to represent a delta complex, where further progradation into the basin is limited by deeper water to the west of the high. The delta complex illustrates areas where accommodation space to the east of the high is filled in, allowing delta systems to prograde to the crest of the high.

The second order sequence boundaries in Triassic are interpreted by Glørstad-Clark et al. (2009) to be maximum flooding surfaces, which are controlled by regional or global tectonics, and not a result of local tectonics related to paleo-Loppa High.

3.4.6 Sequence S4 (Ladinian-middle Carnian)

Sequence S4 was deposited from Ladinian to middle Carnian times. It is bounded in base by horizon 9 (early Ladinian) and on top by horizon 11 (middle Carnian) (*fig. 3.9*). In addition to the bounding surfaces, S4 is subdivided into four units, separated by three internal horizons. Horizon 10 at the Ladinian/Carnian boundary (Glørstad-Clark et al. 2009) and two additional horizons recognized in this study to illustrate the development west of Fault Zone A; horizons *a* and *b*. Horizon a is of latest early Ladinian age, horizon b of middle Ladinian age. They form the boundaries of the wedge, a succession defined to interpret the basin west of Fault Zone A. The wedge will be discussed in the following section, comparing and contrasting the development of this depositional unit east and west of the paleo-Loppa High.

S4 is first reviewed on the eastern side of Fault Zone A, and secondly the western side, underneath the header: the wedge.

East of fault Zone A horizon 9 is mostly onlapping the paleo-Loppa High, and is in general the youngest horizon not to drape the crest, with some exceptions. Right above horizon 9 some west-east prograding clinoforms are found, indicating yet another small uplift episode in early Ladinian. The sequence between horizons 9 and a is the first interval to drape the paleo-Loppa High throughout the area. It thins drastically towards paleo-Loppa High, as there still were some relief and accommodation space on the eastern flank and eastward towards the Bjarmeland Platform.

The sequence between horizons a and b are the interval forming the wedge west of Fault Zone A. In east the interval remains an equal, relatively thin succession transported from the Bjarmeland Platform, before it thins a little on the crest of paleo-Loppa High, indicating that some relief were still present. The sequence has parallel layering, and some gently dipping toplaps eroded by horizon b. Sediment bypass and relative shallow sea-level is interpreted for the area east of the crest of the paleo-Loppa High towards the Bjarmeland Platform.

The upper part of S4 from horizon b to horizon 11 is described throughout the area because Fault Zone A was now stable and areas both east and west of paleo-Loppa High were filled. The upper part of S4 has a nearly equal thickness, with a small increase westward. Horizon b marks the end of tectonics, it is harder to map out eastward towards the Bjarmeland Platform. The horizon becomes more and more inconsistent compared to on top of the crest of paleo-Loppa High and westward. The layering of the sequence between horizons *b* and *10* varies throughout the area with no specific pattern. Parallel, subparallel and chaotic patterns are present, and several small bodies of high amplitude. The high amplitude bodies have an u-shape and are interpreted to be small channels, the channels are found in the middle part of the study area. The sequence between horizons *10* and *11* have a more tidy appearance, with more clear reflectors, especially north in the study area. West-east prograding clinoforms are interpreted in north, while the southern area show features that can be interpreted as channels.

Glørstad-Clack et al. (2009) report of thickness increase towards Loppa High from the Bjarmeland Platform which implies that Loppa High itself have become a depocenter. Subsidence of the paleo-Loppa High area gradually makes new room for sediments.

The wedge

On the eastern side of Fault Zone A this unit is relatively thin, increasing dramatically in thickness on the western side (of Fault Zone A), representing a period of massive infilling of sediments. This depositional system has not been defined or recognized by previous workers in the area, and marks the main infilling episode of the starved accommodation space west of the paleo-high. The unit is bounded by horizons *a* at the base and *b* on top. In addition, it has several internal erosional surfaces (intra wedge) that illustrates periods of infill. However these erosional surface are hard to map out laterally due to their local nature. Thus, they were not identified as separate units within this study. The wedge was deposited in a relatively short amount of time, from latest early Ladinian to middle Ladinian. The thickness of the wedge is four times larger than the same interval on top of the paleo-Loppa High crest. Parts of the unit are characterized by sets of internal clinoforms prograding east to west, whereas other parts have sub-parallel layering or a total chaotic structure.

Horizon *a* marks the start of main sedimentation on the western side of paleo-Loppa High, it followed a time period from late Permian times to Early Ladinian with starved sedimentation patterns due to the positive structural feature represented by the paleo-Loppa High. Moreover, this indicates that accommodation space east of the paleo-high was filled in prior to this event, allowing sediments to bypass the high and deposit on the western side of the high. The event is recognized by an angular unconformity with downlapping clinoform-units prograding from east to west. The upper boundary is also an angular unconformity with toplap terminations. The sequence shows internal layering with stacked clinoform-units prograding, as a result of several phases of changes in relative sea-level. *Figure 3.14* shows some internal horizons (in green) of the wedge that illustrates the different phases of deposition. The figure is only one example of these erosional surfaces, since these erosional patterns do vary significantly laterally in the study area.

The wedge is thickening towards north, illustrating the increasing accommodation space in this area caused by a larger tectonic activity in north. The outbuilding of this sequence probably starts in south with the small delta identified by Glørstad-Clark et al. (2009), followed at a later stage by sediment input along all of the crest from south to north. Sediments bypass the top of the ridge and then enter the western basin.

The intra wedge horizons are not correlated through the basin, but on several profiles they repeat the same features.

Polheim Subplatform is set within a complex tectonic setting. Due to later tectonic activity and inversion it is challenging to determine the movements on the faults during deposition of the wedge. Faulting in Fault Zone A is the most difficult to examine at this time period. According to the fault description done in section 3.2.2 Fault Zone A2 may have been active at the time of deposition. However the sediments deposited in this zone are cut through by many faults and is therefore very hard to interpret. There is a possibility of a small thickness increase towards Fault Zone A2, indicating syn-tectonic sedimentation. The same feature is observed towards faults B1a and B2a. The sequence between the two intra wedge horizons increase in thickness towards fault B1a and B2a. It must be taken into consideration that the apparent thickness increase is very small and hard to track due to the chaotic seismic facies of the wedge and the tectonic activity affecting the area at a later stage.

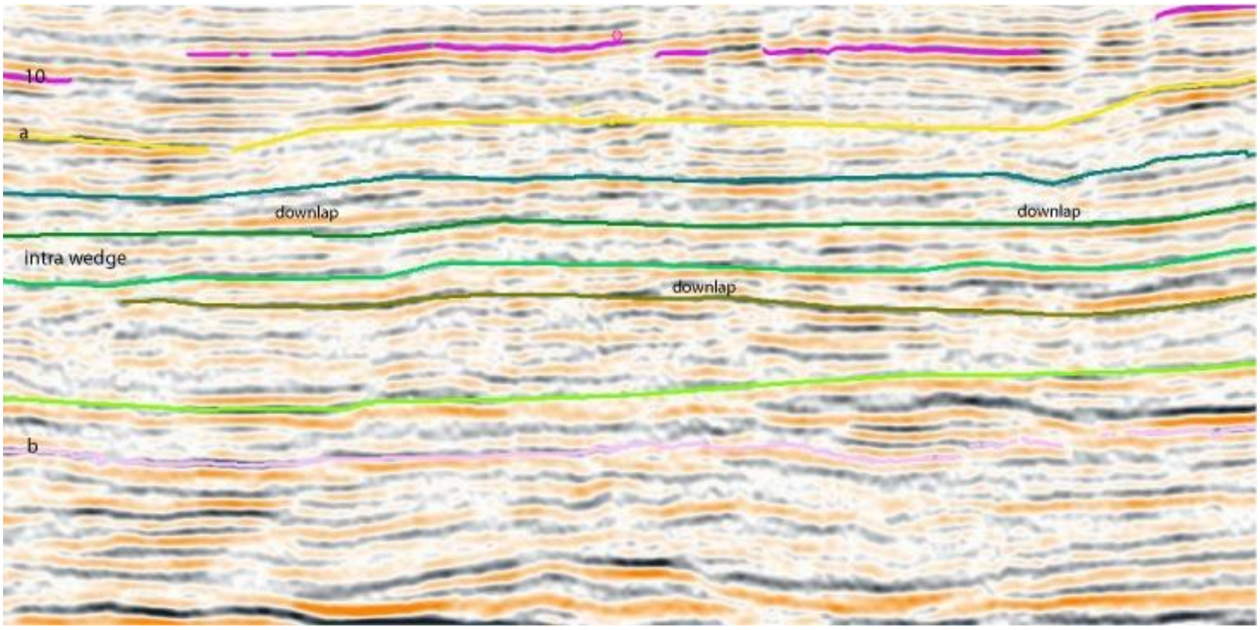


Fig. 3.14: Seismic line 141500. A close-up on the wedge, its internal composition, and downlap geometries. Stratigraphic diagram with sequences and horizons are given in *figure 3.9*.

3.4.7 Sequence S5 (upper Triassic)

Sequence S5 was deposited in upper Triassic times. It is bounded by horizon 11 at the base (*fig. 3.9*), whereas the upper boundary was eroded by a composite regional unconformity (*fig. 3.15*)(Glørstad-Clack et al. 2009). Moreover, horizon 11 was eroded on top of the paleo-Loppa High crest, north of ~72° 30’. In general, erosion to the east increased towards the north. In contrast the western side of Fault Zone A, the Polheim Subplatform, was down-faulted, and more of S5 is present. Since the upper boundary were eroded in the paleo-Loppa High area it is hard to determine any thickness variations. On larger scale Glørstad-Clark et al. (2009), report of a thickness increase westward over the paleo-Loppa High area, as this has developed into a depocenter, and accommodation space generation continued to be high.

Many clear surfaces with a large horizontal extent are observed, interpreted to be flooding surfaces. No good examples of clinoforms have been found, but clinoforms in the northern paleo-Loppa High area are reported by Glørstad-Clark et al. (2009). Many high amplitude bodies have been found especially eastward towards the Bjarmeland Platform. These could be channels, but some is also suspected be due to degradation of the seismic data as there are many younger faults interfering with the area.

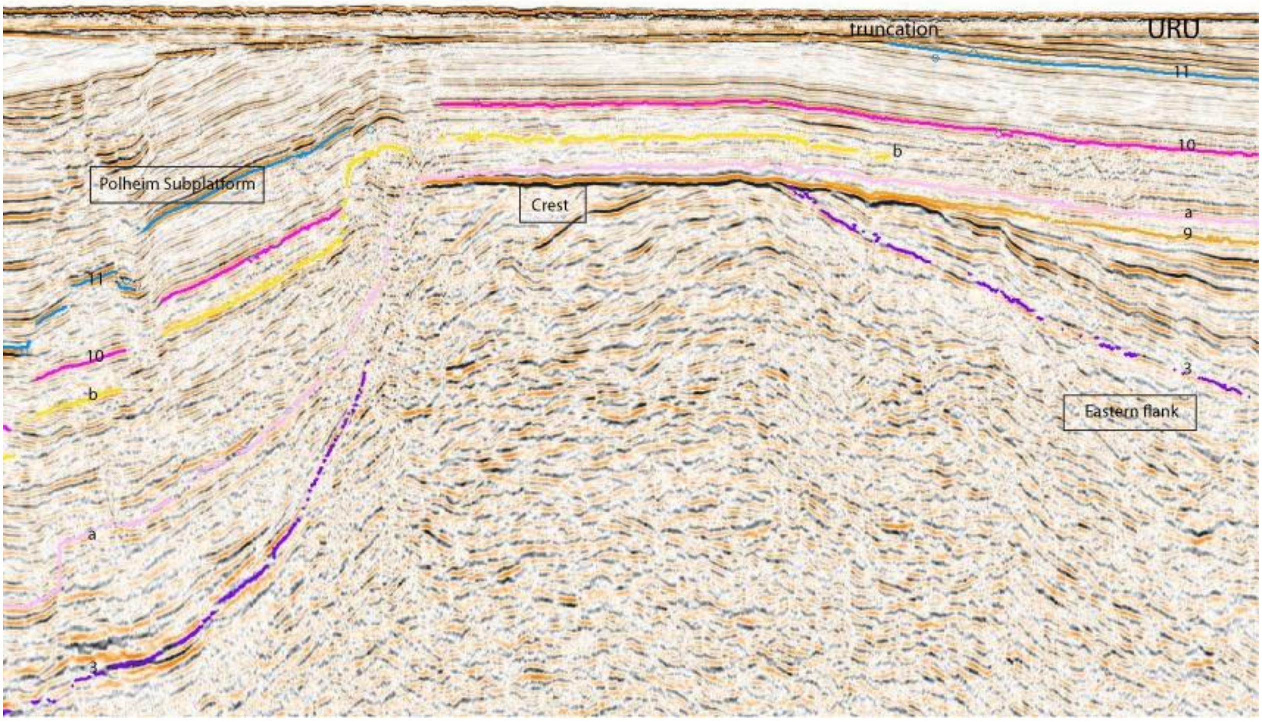


Fig. 3.15: Seismic line 141983. A composite regional unconformity truncates the intra Carnian horizon (12). Stratigraphic diagram with sequences and horizons are given in *figure 3.9*.

4. Discussion

4.1 Geological evolution of the paleo-Loppa High area

The description of the geological evolution of paleo-Loppa High is made through observations of tectonics, subsidence, uplift, erosion and deposition. The oldest horizon identified in the study area is horizon 1 (Late Mississippian). However, the outline of the geological history starts at the top of the basement, and at the first extensional event in the Western Barents Sea.

4.1.1 Late Devonian - early Carboniferous (Billefjorden Group)

The extensional geological evolution of the Western Barents Sea starts with a rift phase in Late Devonian to early Carboniferous (Gudlaugsson et al. 1998). We have not found any evidence of this rift phase in the Loppa High area. It does not mean that this part of the Barents Sea was unaffected by this tectonic event, but it is challenging to investigate this part of the succession since it is deeply buried and the seismic resolution is degraded. As seen in *figure 4.1* a non-depositional period is marked over the whole area at these times. A regional seismic stratigraphic interpretation done by Gudlaugsson et al. (1998) suggested that sediments of Carboniferous age are assigned to the lithostratigraphic Billefjorden Group, which may be resting discordant on basement (Gudlaugsson et al. 1998).

From Late Devonian to early Carboniferous the Billefjorden Group was deposited, and it is the oldest lithostratigraphic group recognized in the Loppa High area (Larssen et al. 2005). Larssen et al. (2005) assumed that the Billefjorden Group has an equal thickness in the whole area, thus it has been deposited on a relatively flat topography. Horizon 1, marking the top of the Billefjorden Group, shows no signs of paleo-Loppa High topography. The depositional environment was characterized by alluvial fans, proximal braided river systems and some local volcanic activity (Larssen et al. 2005).

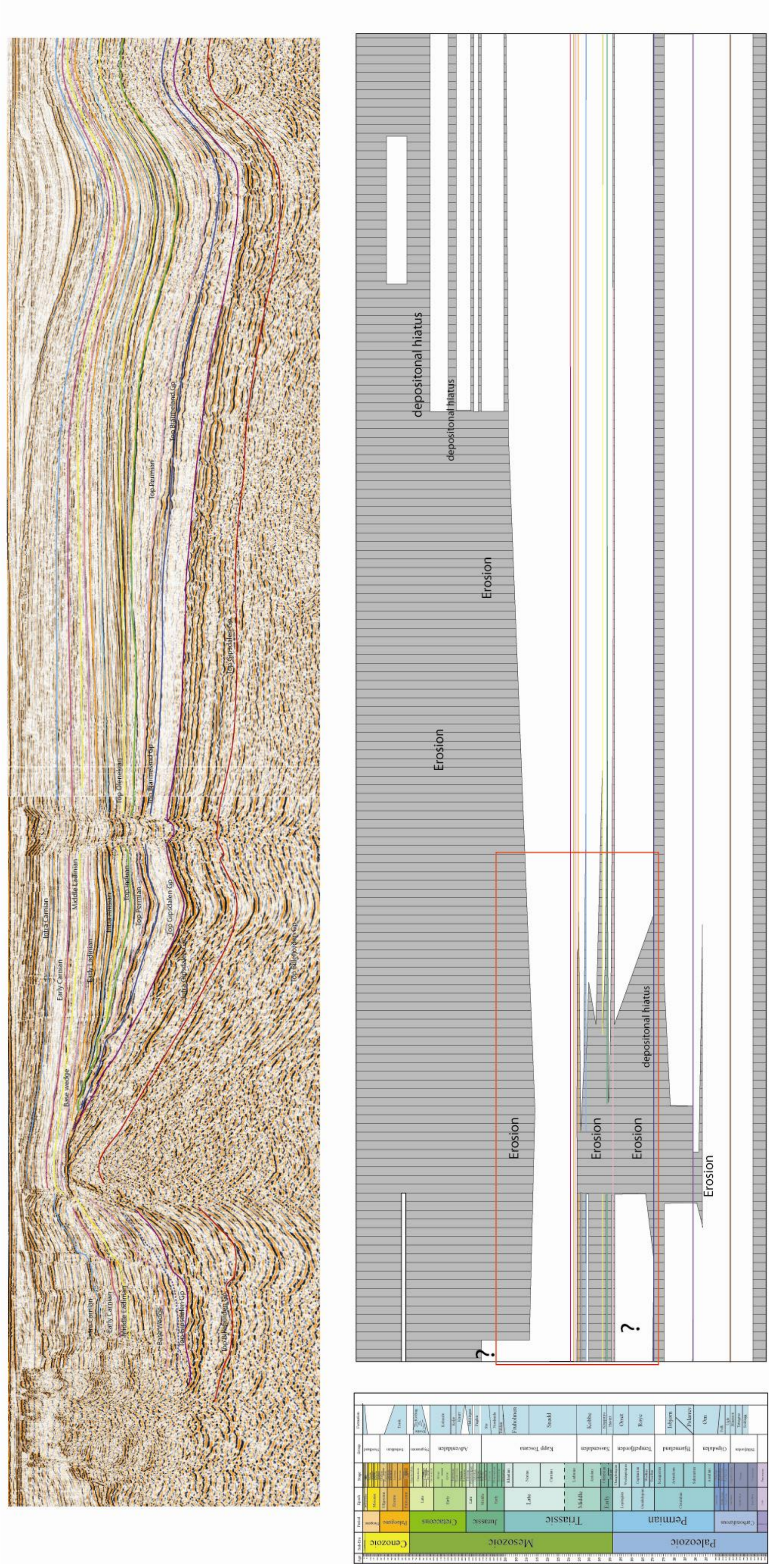


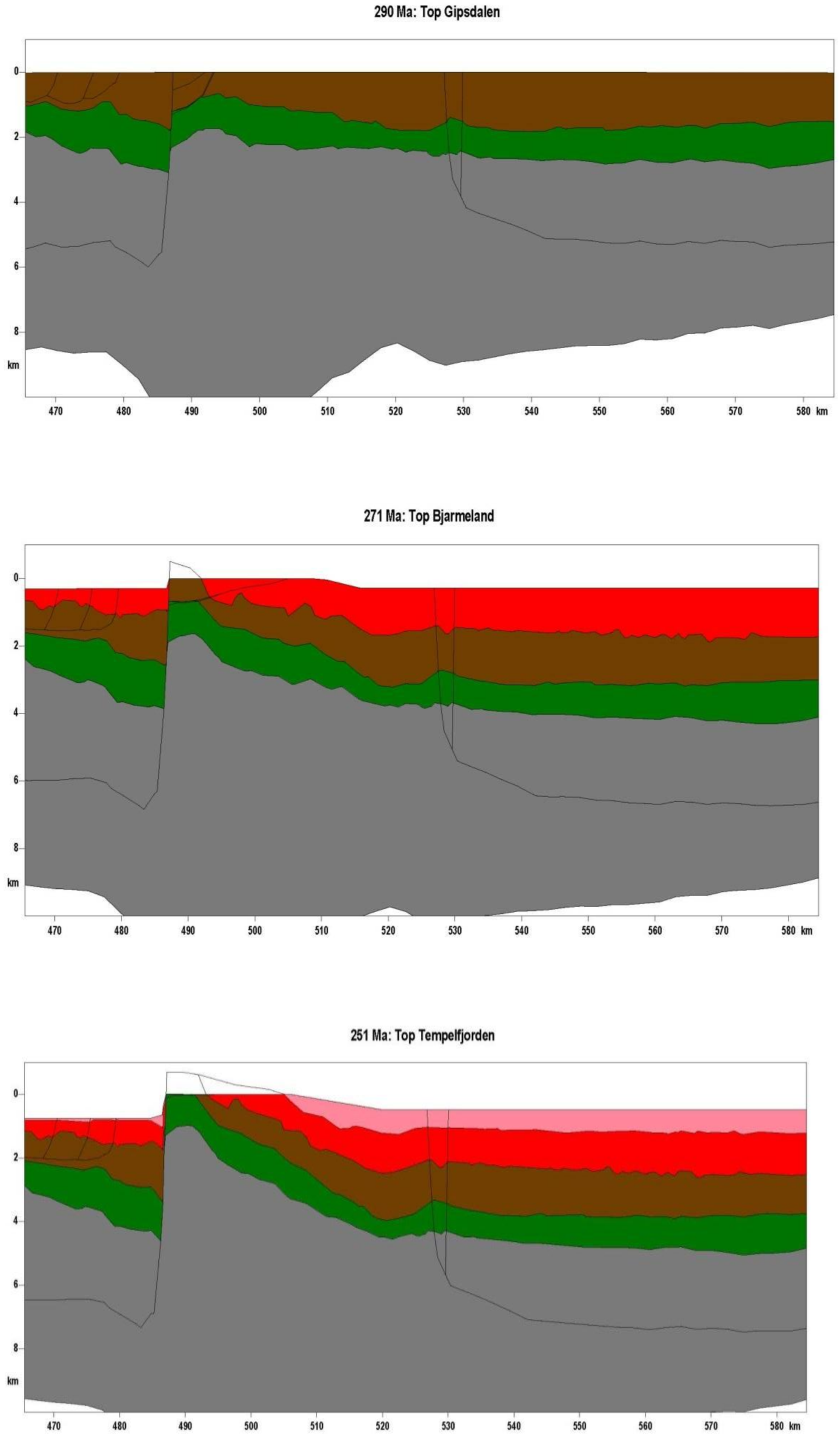
Fig. 4.1: Chronostratigraphic chart of the paleo-Loppa High area. For details on the geologic time scale see *figure 2.3*. The red box marks a zoomed area in *figure 4.8*.

4.1.2 Middle Carboniferous-late Carboniferous (lower Gipsdalen Group)

A regional extensional event in the Western Barents Sea in middle Carboniferous times coincides with the transition from Billefjorden to Gipsdalen groups. The rifting was of regional scale (Gudlaugsson et al. 1998, Larssen et al. 2005). There are no signs of tectonic activity shown in the lower Gipsdalen Group, which displays an equal thickness throughout the study area. This is an indication that the paleo-Loppa High may have been part of a relatively stable tectonic area until late Carboniferous times, whereas evidences of rifting in Late Devonian-early Carboniferous and middle Carboniferous are observed other places in the Barents Sea and on Svalbard.

4.1.3 Late Carboniferous-early Permian (upper Gipsdalen Group)

The first tectonic event identified in the Loppa High area was in late Carboniferous times, and coincides with horizon 2. It splits the Gipsdalen Group sequence development into two stages, named lower and upper Gipsdalen Group. In contrast to the lower Gipsdalen Group, which has an equal thickness throughout the area, the upper Gipsdalen Group is thinning towards the crest of paleo-Loppa High, and thickening in the half-graben. *Figure 4.2* is a model of the paleo-Loppa High area, showing three stages during late Paleozoic times. The reconstruction of middle Sakmarian time (Top Gipsdalen Group, 290 Ma) is shown in *figure 4.2 A*, it shows the lower Gipsdalen Group in green and the upper Gipsdalen Group in brown. The upper part of the Gipsdalen succession is a very complex seismic sequence and also sedimentary unit, with a mixture of coarse-grained and fine to medium grained sandstones, shales, dolomitic mudstones and anhydrite (Larssen et al. 2005). There are several uncertainties to the geological history in the upper Gipsdalen Group, but it is very likely that a series of tectonic events have contributed to the depositional patterns, also supported by Larssen et al. (2005). In the uppermost Gipsdalen Group there are signs of a tectonic event creating some paleo-Loppa High topography, but some uncertainty is linked to this proposal as no onlapping reflectors have been identified.



A

B

C

Fig. 4.2: Backstripping and structural restoration of the paleo-Loppa High area, upper Paleozoic. A: Top Gipsdalen, horizon 3, middle Sakmarian. B: Top Bjarmeland, horizon 4, after tectonic event in Kungurian. C: Top Tempelfjorden, horizon 5, after tectonic event in Lopingian. For legend see *figure 3.11*.

The change in depositional pattern in the upper Gipsdalen Group reflects a two step development related to two tectonic events. The first event was responsible of initiating the half-graben structure south in the Loppa High area, coinciding with horizon 2 of late Carboniferous age. The second tectonic event initiated the formation of the paleo-Loppa High structure in the uppermost Gipsdalen Group, set to be in early Sakmarian? times. The two events of upper Gipsdalen Group will be further discussed below.

The first tectonic event (horizon 2, late Carboniferous) initiated the half-graben south in the Loppa High area, as shown in *figure 4.3* with a seismic profile of a southern line and two schematic sections. The schematic section to the left illustrates how the half-graben was initiated and the following depositional unit (named Upper Gipsdalen on *figure 4.3)* was deposited with a thickness increase in the half-graben but has an equal thickness in the rest of the area. The central and northern part of the Loppa High area, shown in *figure 4.4* and *4.5*, had little topography and was not affected by the half-graben as illustrated by the uniform thickness distribution in the area.

The second tectonic event, in early Sakmarian?, initiated the formation of the paleo-Loppa High structure. On the chronostratigraphic chart (*fig. 4.1*) the change in depositional pattern is illustrated by a non-depositional area pinching out to the east that illustrates that paleo-Loppa High had some relief in early Permian times. Fault Zone A1 was active and uplifted the paleo-Loppa High structure and tilted the eastern flank. Throughout the study area a carbonate platform entered from east in the uppermost Gipsdalen Group, shown as a high reflectivity band on seismic. The seismic facies changes to the eastern flank of the paleo-Loppa High, indicating a change in carbonate platform deposition. This change in deposition may be due to the change in topography related to the paleo-Loppa High, which may have been exposed at the time, ending the production of carbonate deposition. *Figure 4.6* shows the spatial extent of the carbonate platform. The area can be divided into three depending on their carbonate production as seen from seismic facies; south (*fig. 4.3*), central (*fig. 4.4*) and north (*fig. 4.5*). The half-graben is a local structure, only developed in the southern part of the area. An arrow in the seismic profile points to where the carbonate platform ended.

The newly uplifted paleo-Loppa High and the half-graben in south created a limited area with restricted circulation. Evaporites form in this kind of setting if evaporisation losses exceed precipitation from rain and snow (Reading 1996), and according to Larssen et al. (2005) and

Smelror et al. (2009) this is a time period when the western Barents Sea was semi-enclosed and evaporites was widespread in local basins. The high reflectivity, triangle shaped feature we see on the seismic profile in *figure 4.3*, situated right underneath horizon 3, is interpreted to represent evaporites. In the central and northern part of the area the half-graben is not developed. The initiation and formation of some topography related to the first uplift of paleo-Loppa High might have happened here as well. In the central areas of paleo-Loppa High, the carbonate platform, interpreted to be just underneath horizon 3, terminates its development further to the west (*fig. 4.4*) compared to the southern and northern areas. This might be caused by topography variations along the ridge of the newly created paleo-Loppa High. The uplifted area is smaller and the eastward dipping flank may be steeper in the central area than in the northern part of the Loppa High area (*fig. 4.5*). In south the half graben structure limits the development of the carbonate platform.

The seismic facies interpreted in upper Gipsdalen Group also has a marked change coinciding with horizon 2. Clear parallel seismic reflectors dominate the uppermost part of Gipsdalen. Asselian times are characterized by high frequency and high amplitude eustatic sea-level changes (Smelror et al. 2009). Thus, the clear layering stems from carbonate production cycles. The high frequency carbonate platform just underneath horizon 3 has pillow-shaped features towards the Bjarmeland Platform in east, which is due to carbonate build-ups. Deposits in lower Gipsdalen Group have a lower frequency and a more chaotic to sub-parallel appearance.

During early Sakmarian? the Polheim Subplatform west of Fault Zone A was downfaulted. Evidence of syn-tectonic sedimentation on Polheim Subplatform in the upper Gipsdalen Group is observed based on thickness variations towards the fault plane *(fig. 4.2 A).*

Horizon 3 marks the top of Gipsdalen Group and is correlated to middle Sakmarian times. Due to the uplift in early Sakmarian?, horizon 3 might be a sub-aerial exposure surface (Larssen et al. 2005). The upper Gipsdalen Group is thinning up the eastern flank towards the crest of paleo-Loppa High. Horizon 3 changes in character and is overlain by downlapping clinoform units in the Bjarmeland Group at the crest of paleo-Loppa High. The thinning could be a result of erosion due to the uplift. Alternatively, the thinning could be caused by facies changes associated with carbonate deposition, that could have been affected by a minor uplift of the paleo-high. Carbonate deposition may still have occurred on top of the high, but with insufficient thickness to be visible on seismic sections.

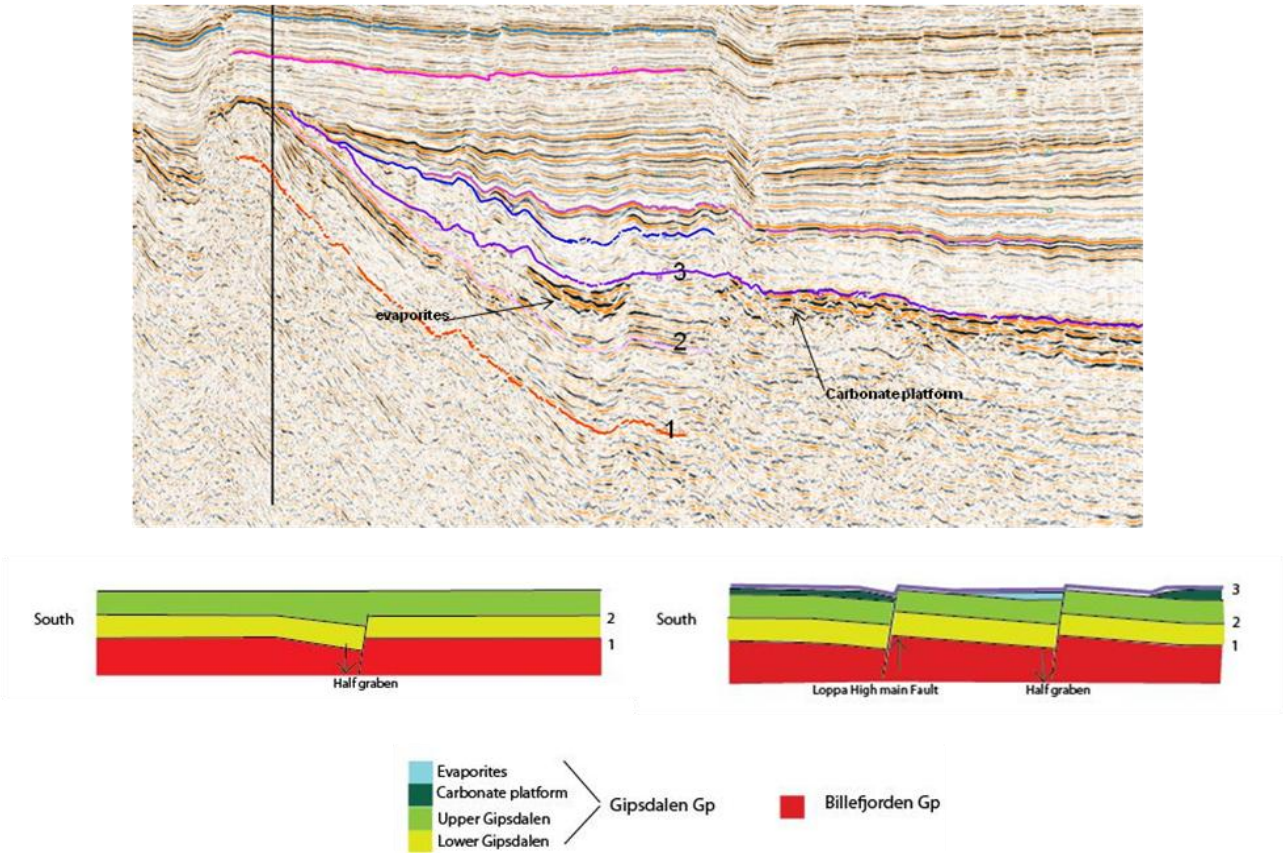


Fig. 4.3: Seismic profile in the southern part of the paleo-Loppa High area. Arrows point to the interpreted evaporites and the ending of the carbonate platform. Schematic sections are a simplified illustrations of the strata before (left) and after (right) the tectonic event in early Sakmarian?. Stratigraphic diagram with sequences and horizons are given in *figure 3.9*. Approximately locations of seismic lines are shown in *figure 4.6.*

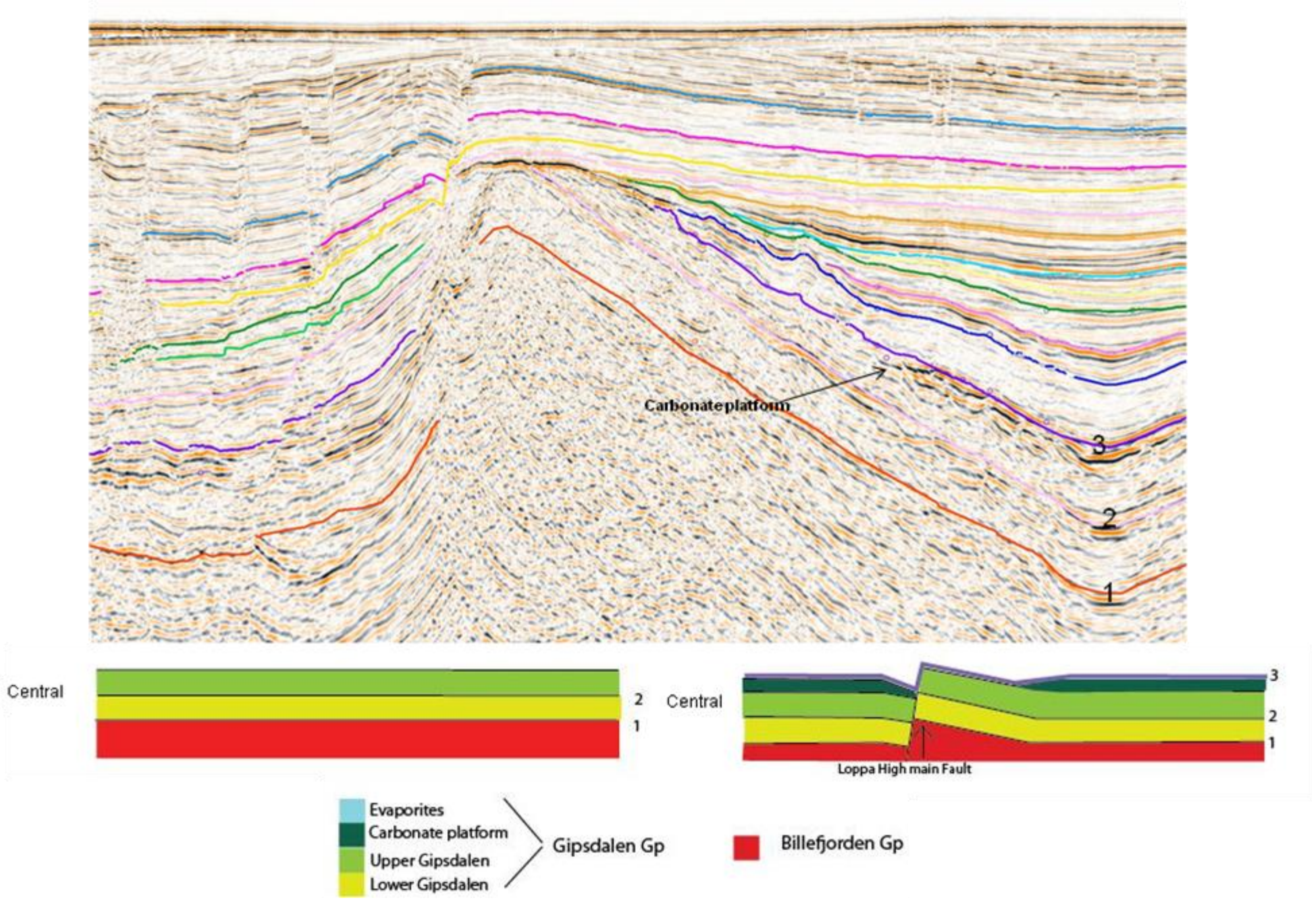


Fig. 4.4: Seismic profile in the central part of the paleo-Loppa High area. Arrow point to the interpreted ending of the carbonate platform. Schematic sections are a simplified illustrations of the strata before (left) and after (right) the tectonic event in early Sakmarian?. Stratigraphic diagram with sequences and horizons are given in *figure 3.9*. Approximately locations of seismic lines are shown in *figure 4.6.*

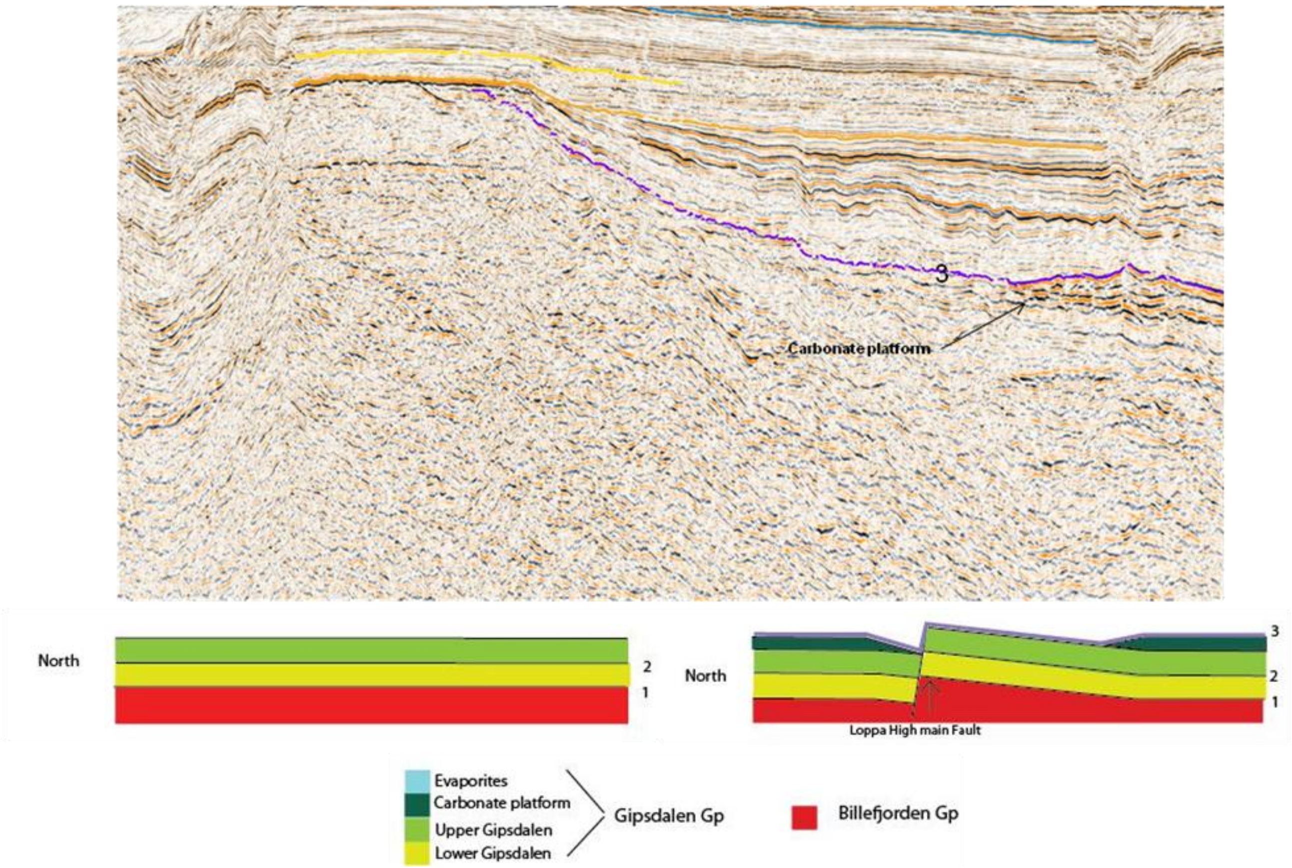


Fig. 4.5: Seismic profile in the northern part of the paleo-Loppa High area. Arrow point to the interpreted ending of the carbonate platform. Schematic sections are a simplified illustrations of the strata before (left) and after (right) the tectonic event in early Sakmarian?. Stratigraphic diagram with sequences and horizons are given in *figure 3.9*. Approximately locations of seismic lines are shown in *figure 4.6.*

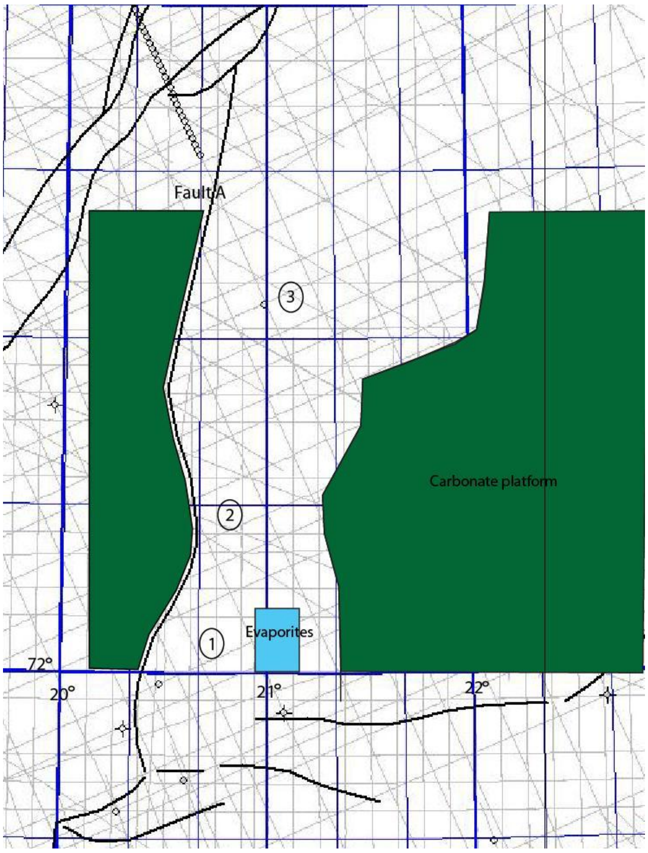


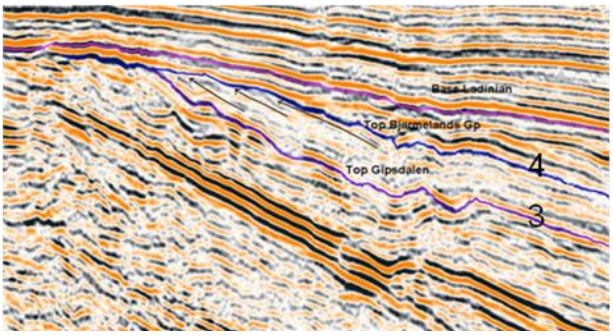
Fig. 4.6: Base map of the paleo-Loppa High area with the spatial extent of the carbonate platform of the uppermost Gipsdalen sequence. Evaporites is interpreted south in the study area. The numbers on the base map refers to the approximate locations of the seismic lines in *figures 4.4, 4.5* and *4.6.*

4.1.4 Middle Sakmarian-late Artinskian (Bjarmeland Group)

The middle Sakmarian-late Artinskian time period is dominated by cold water carbonates like crinoids, bryozoans, brachiopods and siliceous sponges. The underlying Gipsdalen Group succession is markedly different and consists of foraminifer-dominated, warm water carbonates (Larssen et al. 2005). This makes horizon 3, which separates the two sequences, a sharp seismic and lithostratigraphic boundary. Horizon 3 (middle Sakmarian) is an uneven surface due to its carbonate composition. The seismic appearance of the two sequences is also very different. Gipsdalen Group is a high reflectivity package, while Bjarmeland Group has a transparent look. It is hard to determine a single tectonic event separating the two surfaces, instead a more evenly distributed subsidence of the eastern area and uplift of the paleo-Loppa High crest is more likely during middle Sakmarian to late Artinskian times (Bjarmeland

Group). Differential subsidence gradually increased the accommodation space, south in the paleo-Loppa High area. Bjarmeland Group downlaps on horizon 3, while north in the study area the reflectors of Sakmarian age are parallel to horizon 3. This is because the magnitude of later tectonics is much greater in north, this leads to more tilting, uplift and erosion. The differential uplift in north and south is due to later tectonics (will be further discussed in 4.1.8).

The reason the Bjarmeland Group is truncated both on the horizontal and on the vertical plane is due to large tectonic events in Kungurian times directly following the deposition of this unit. Footwall uplift and rotation of the paleo-Loppa high, caused erosion of the Bjarmeland Group as seen in *figure* 4.7.



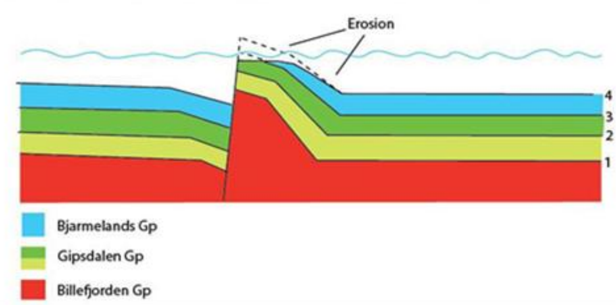


Fig. 4.7: Seismic profile with interpreted truncations in the Bjarmeland sequence. Schematic section illustrates the strata after the tectonic event in Kungurian. Stratigraphic diagram with sequences and horizons are given in *figure 3.9*.

4.1.5 Kungurian

Kungarian time was a period of major tectonic activity, which lead to a phase of non-deposition covering the whole Loppa High area coinciding with horizon 4 (*fig.4.1*). Tectonics in Fault Zone A, increased the slip component, caused footwall uplift of the crest of Loppa High, and tilted the eastern flank. The crest of Paleo-Loppa High was established as an island that existed until Middle Triassic times, with only a few drowning incidences associated with maximum flooding surfaces. The upper part of Bjarmeland Group was severely eroded as a consequence of this uplift and tilting (*fig. 4.7*). *Figure 4.2 B* shows a reconstruction of the paleo-Loppa High area after the Kungurian tectonic event (Top Bjarmeland, horizon 4, 271 Ma). The transparent area indicates the part of the succession that was currently eroded. The erosion eventually lead to the crest observed on seismic. It took about ~40 m.y., from Kungurian to Middle Triassic, to make it.

4.1.6 Guadalupian (Tempelfjorden Group)

Tempelfjorden Group was deposited during Guadalupian. Paleo-Loppa High acted as a barrier to sediments and was an island interrupted by small periods of flooding. When Paleo-Loppa High occasionally were drowned, some Tempelfjorden Group sediments were potentially deposited in the western basin, but in so small amounts that they are not recognizable on seismic. During deposition of Tempelfjorden Group the western basin experienced mostly sediment starvation in its newly formed accommodation space. As seen on *figure 4.2 C,* Tempelfjorden Group is very thin west of Fault Zone A compared to east. Tempelfjorden marks a change in depositional pattern, from a uniform thickness throughout the area in older successions, to a starved western basin and onlap on the eastern flank. The sedimentation west of Fault Zone A during Late Permian to Middle Triassic is a very uncertain part of the geological evolution due to seismic correlation problems over and around the positive structural paleo-Loppa High during this time period. On the eastern flank Tempelfjorden

Group onlaps the Bjarmeland Group and thins towards the crest of paleo-Loppa High due to the amplified relief.

4.1.7 Lopingian (uppermost Tempelfjorden Group)

The end of Permian is marked by a new major tectonic event in late Lopingian coinciding with horizon 5. This episode was associated with a non-depositional period according to Larssen et al. (2005) (*fig. 3.3*) and well 7121/1-1 (*table 3.1*). The event further uplifted the crest of paleo-Loppa High, downfaulted the western basin which increased the accommodation space. The base of the eastern flank experienced subsidence. This is the largest large tectonic event in paleo-Loppa High history, and it established paleo-Loppa High as a narrow ridge as described by Gudlaugsson et al. (1998). *Figure 4.2 C* shows the area after the second major tectonic event in Lopignian (Top Tempelfjorden Group, 251 Ma). The transparent area is the older successions that have been uplifted and further eroded. At this time the paleo-Loppa High ridge was pretty much established to what is observed on seismic.

4.1.8 Induan- early Ladinian (S1, S2 and S3)

From Early to Middle Triassic paleo-Loppa High stood up as an island and acted as a barrier to sediments prograding from east to west. Thus, the western basin was still in a state of sediment starvation. S1, S2 and S3 share much of the same development and features. No large tectonic events separate the sequences, continuous subsidence and variations in sea-level created new accommodation space. Variation in sea-level and sediment input controlled the depositional sequences bounded by maximum flooding surfaces, but sequences show a gradual infilling of accommodation space east of the paleo-Loppa High.

The chronostratigraphic chart in *figure 4.8* zooms in and summarizes the depositional and non-depositional patterns in Triassic. Starting in Induan times (sequence S1, bounded by horizon 5 and 6), a clastic system entering the paleo-Loppa High area marks the start of Triassic deposition. The chronostratigraphic chart (*fig. 4.8*) shows that the clastic system onlapped and further draped the crest of paleo-Loppa High in early Induan. The S1 sediments at the crest and the upper part of the eastern flank were later eroded and truncated due to renewed minor uplift coinciding with horizon 6(late Induan times) (Glørstad-Clark et al. 2009).

Late Induan uplift was followed by a latest Induan transgression coinciding with horizon 6. The transgression lead to a flooding event that potentially allowed some sediments to be deposited in the western basin, accompanied by a retrogradation of the clastic system to the east and south (Glørstad-Clark et a. 2009, Skjold et al. 1998). The Olenekian sequence (S2, bounded by horizon 6 and 7) marks renewed progradation into the basin as seen on *figure 4.8*, but non-deposition characterizes the eastern flank of the plaeo-Loppa High due to the uplift in late Induan. Paleo-Loppa High acted as a barrier to coarse-grained sediments prograding westward during most of Olenekian, and S2 onlaps the eastern flank. Due to this the western basin was dominated by condensed sedimentation from Induan to middle Anisian times.

The Anisian to early Ladinian, sequence S3, is bounded by horizon 7 and 9. Anisian times experienced several flooding events which are represented by several phases of retrogradation of the clastic system in response to changes in relative sea level and sediment input (Glørstad-Clark et al. 2009). *Figure 4.8* shows the deposition of sequence S3, following the maximum flooding event associated with the S2 (Olenekian) sequence. Horizon 7 (top Olenekian) marks the onset of this sequence, with a new sedimentary withdrawal on the eastern side of paleo-Loppa High although not as far east as the two previous ones (*fig. 4.8*). S3 drapes the high two times and allows sediments to be deposited in the western basin. A renewed uplift of paleo-Loppa High coinciding with horizon 8, middle Anisian, eroded S3 sediments deposited on the flank as well as older sequences sub-aerially exposed. Horizon 9 (early Ladinian) marks the upper boundary of S3 and end of Anisian times. It coincides with a tectonic event that once again uplifted paleo-Loppa High further in Early Ladinian times. The uplift caused erosion of the upper part of S3 deposited locally on the eastern flank of the high.

The chronostratigraphic chart (*fig. 4.1*) and the zoom (*fig. 4.8*) show that the paleo-Loppa High has been eroded since early Permian. Since early Permian, the horizontal extent of the

erosion gradually increased to a maximum in early Ladinian, represented by horizon 9. This indicates that the crest was polished down over time and growing in size.
Triassic is often described as a period of calm tectonic conditions (Faleide et al. 1984), dominated by widespread subsidence creating accommodation space for prograding clastic sediments. This study supports the observations of smaller scale tectonic controls on accommodation space in the Triassic as observed by Glørstad-Clark et al. (2009). Periodic smaller scale uplifts of the paleo-Loppa High created a local source area for sediments and the high remained a positive feature from Early to Middle Triassic times. *Fig 4.9. A* have restored the geometries of paleo-Loppa High at early Induan times. It shows the infilling of clastic systems, the crest of paleo-Loppa High is a positive feature and a huge accommodation space existed west of Fault Zone A.

The west-east prograding clinoforms observed in *figure 3.12.B* overlap with east-west prograding clinoforms in late Anisian times, and proves that paleo-Loppa High have been uplifted and eroded several times. Small local systems of clinoforms prograding eastward are found in sequences S1 (Induan), S2 (Olenekian) and S3 (Anisian) along the whole eastern flank of paleo-Loppa High.

On the eastern flank the accommodation space was gradually filled in by prograding sediment systems originating from the Fennoscandian Shield in south and the Urals in south-east (Glørstad-Clark et al. 2009, Van Veen et al. 1993, Skjold et al. 1998), but these sediments never reached the basin west of paleo-Loppa High. During Induan (S1) and Olenekian (S2) times the Bjarmeland Platform was the main depocenter for sediment deposition, but during Anisian times (S3) the Bjarmeland Platform was mostly bypassed and sediments were deposited further north and west (Glørstad-Clark et al. 2009). The horizons of sequences S1, S2 and S3 onlap the eastern flank and is truncated and pinch out at different depths along the paleo-Loppa High. This illustrates the variation in topography of the paleo-Loppa High.
The paleo-waterdepth in Early-Middle Triassic times was ~300 m east of paleo-Loppa High, 0 m on the crest and ~500 m in the western basin.

The eastern flank has an asymmetric outline that affects the deposition of sediments from Induan throughout Anisian (sequences S1 S2 and S3). This indicates that the tilted eastern flank of paleo-Loppa High is split into several smaller east west striking fault-blocks, also recognized by Gudlaugsson et al. (1998). The individual movements of the fault-blocks affected the local variations in onlap and truncation illustrated by horizons 6 through 9 and the random distribution of west east prograding clinoforms. *Figure 4.10 A* is a profile in north-south direction showing small fault blocks that strikes in east-west direction, they are of a much younger age, but it is possible that they are tied to a pattern of fault blocks that were active in Early to Middle Triassic. *Figure 4.10 A* also shows how much the crest of paleo-Loppa high expands towards north. Horizon 9 is the last horizon to onlap paleo-Loppa High towards north as in south and the horizons deposited on top of paleo-Loppa High does not show any decrease in sequence thickness. These factors indicate that the northern part of the Loppa High area was faulted and further uplifted at a later stage, while the southern part did not experience the same uplift during this time period. The later uplift event of northern Loppa High is probably tied to the Late Jurassic-Early Cretaceous rifting.

Horizon 9 (early Ladinian) is the last horizon to onlap paleo-Loppa High in most places, but south in the Loppa High area, a small delta developed during the end of S3 in Late Anisian times. *Figure 4.10 B and C* shows a seismic profile and a schematic section of the delta. The delta lies on the crest of paleo-Loppa High, and shows internal downlapping clionoforms. The delta was prevented from developing any further westward because of an abrupt deepening of sea-level due to the deep basin in the west bounded by Fault Zone A. Only small amounts of sediments were transported from the delta and into the western basin. It is also restricted in northward direction because of topographic elevation and change of the outline of paleo-Loppa High.

A time-structure map of horizon 9, early Ladinian in *figure 4.11*, shows the outline of the delta and reveals the first entry point for sediments into the western basin after a long period of sediment starvation. The arrows show the sediment transport direction. The dotted arrows were possible entry points where horizon 9 is clearly draping the crest of paleo-Loppa High.

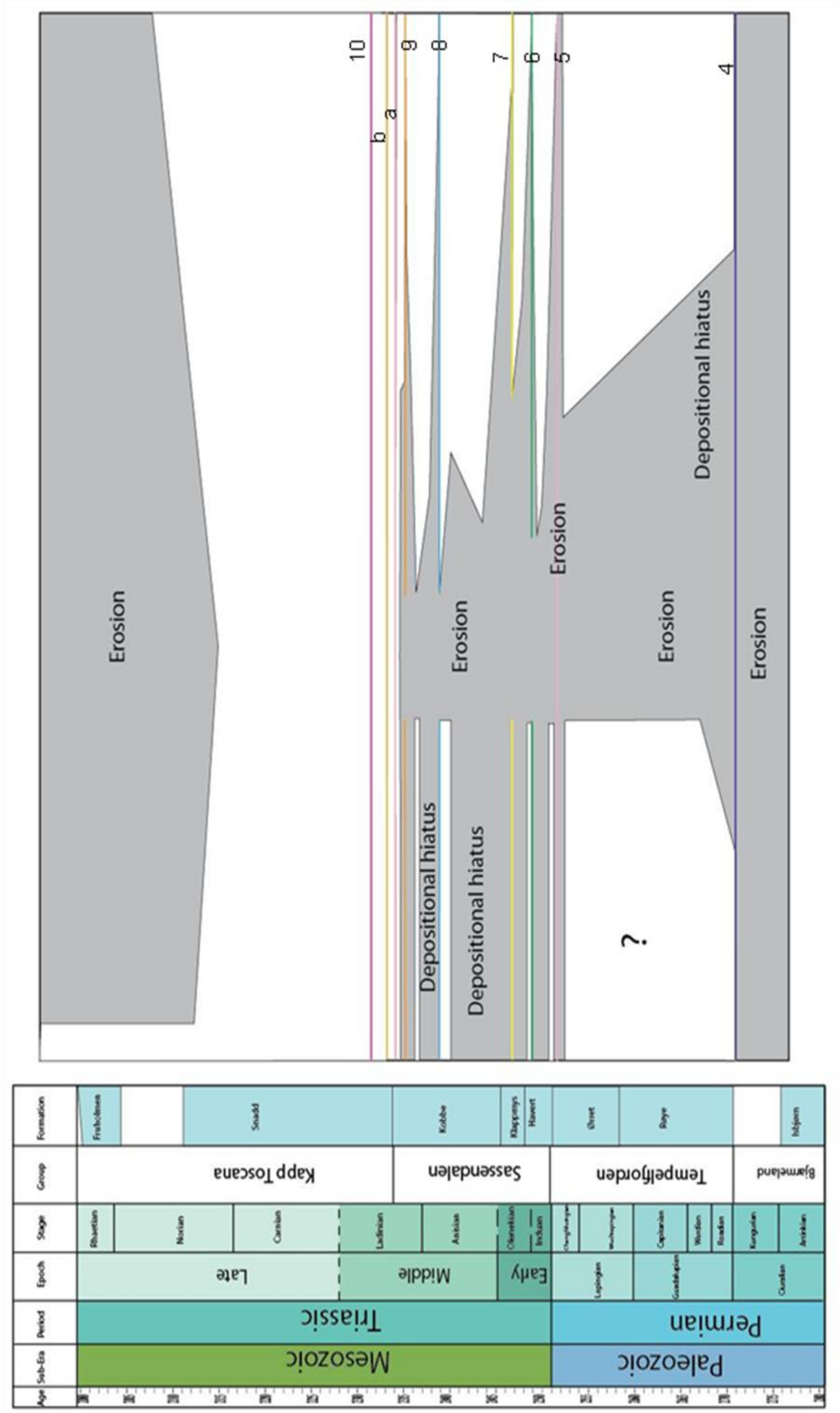


Fig. 4.8: Chronostratigraphic zoom chart of the paleo-Loppa High area from middle Permian through Triassic. Stratigraphic diagram with sequences and horizons are given in *figure 3.9*.

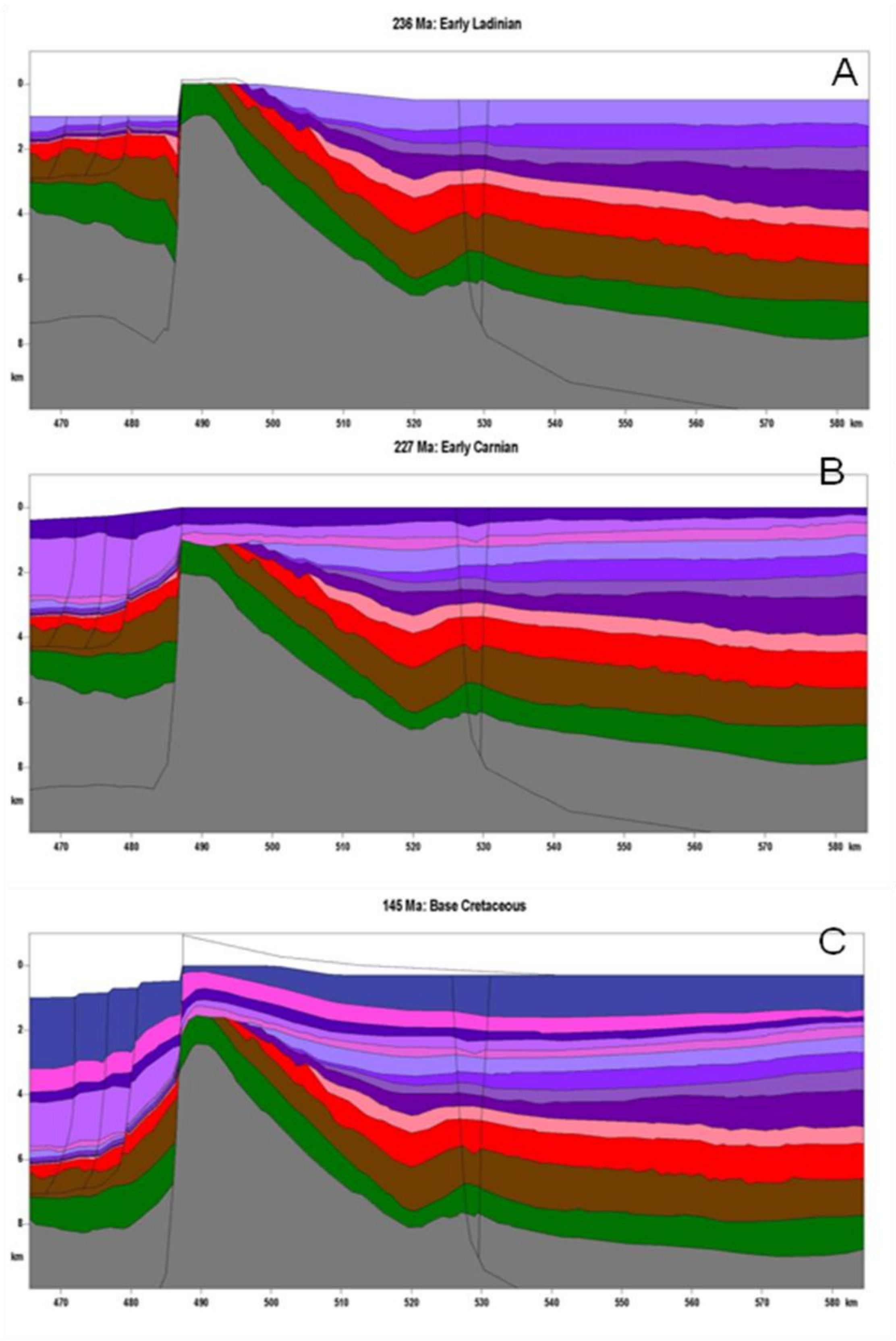


Fig. 4.9: Backstripping and structural restoration of the paleo-Loppa High area, Mesozoic. A: Early Ladinian, horizon 9. B: Early Carnian, horizon 11. C: Base Cretaceous. For legend see *figure 3.11*.

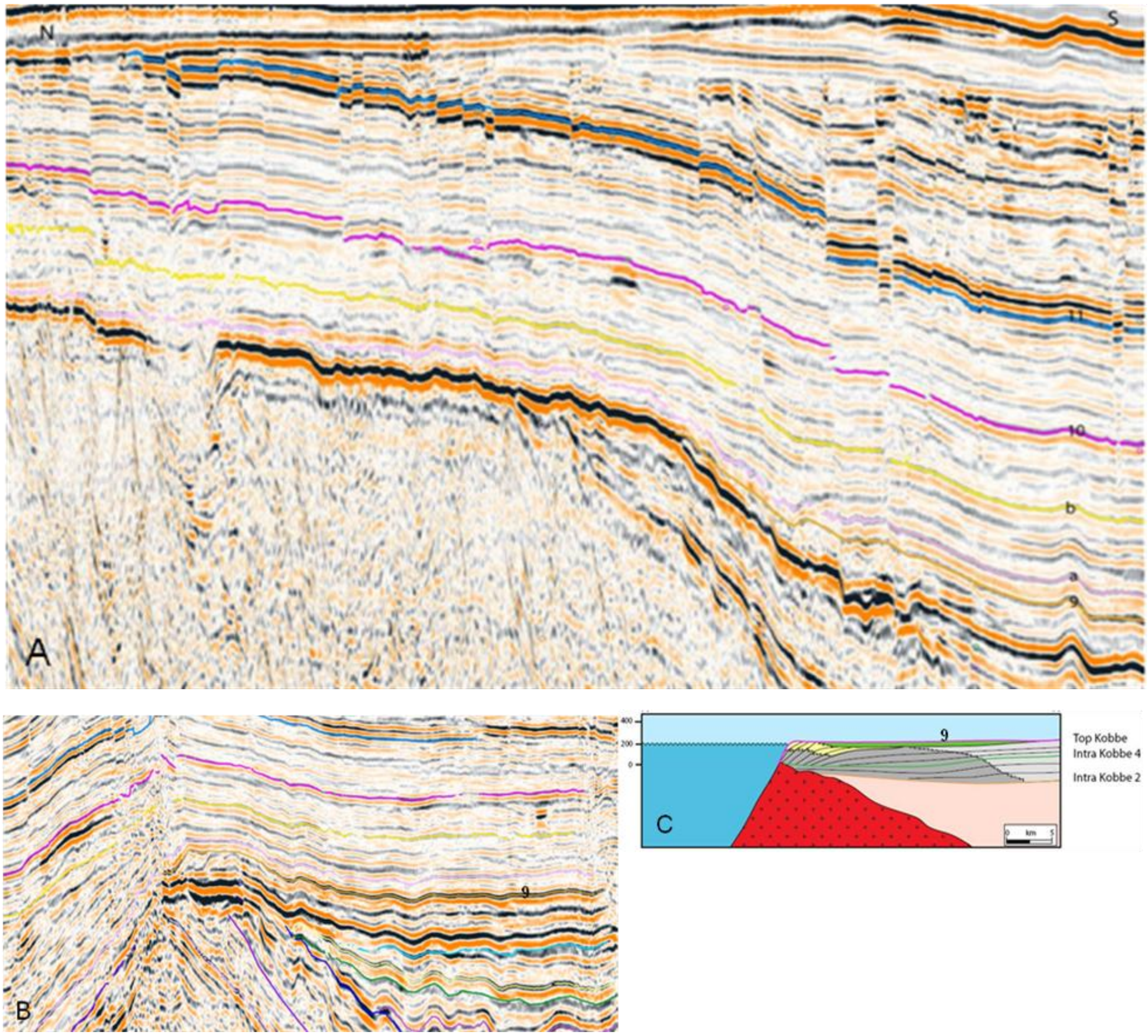


Fig. 4.10 A: Seismic profile in north-south direction showing east-west striking fault blocks and uplift of paleo-Loppa High in north. B: Seismic profile with delta on the crest of paleo-Loppa High. C: Schematic section of the delta.

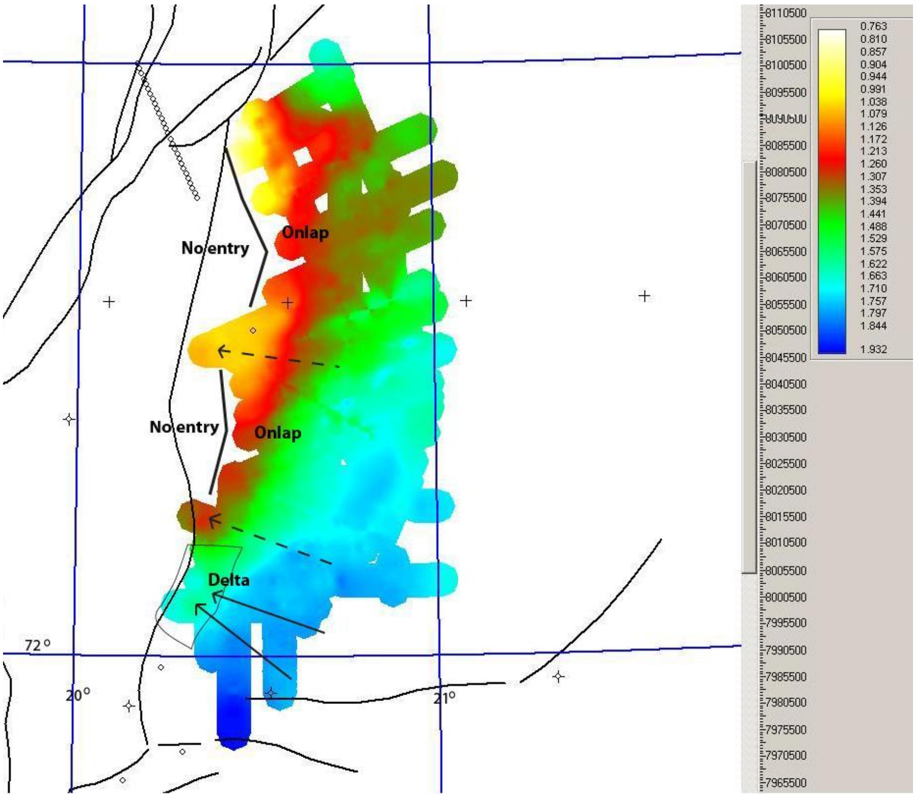


Fig. 4.11: Time-structure map of horizon 9, early Ladinian. Illustrates the outline of the delta. Arrows show the sediment transport direction and reveals the first entry point for sediments into the western basin.. Dotted arrows show possible entry points.

4.1.9 Early Ladinian-middle Carnian (S4)

The time period from early Ladinian to middle Carnian constitutes sequence S4. It is a phase of big changes in the paleo-Loppa High area. Early Ladinian is the last time period were sediments onlap on paleo-Loppa High which at the time had very little structural expression, and there were depositional hiatus on the paleo-Loppa High crest and western basin (*fig. 4.8*). Accommodation space on the eastern side of paleo-Loppa High was after a short period filled up and sediments quickly started filling up the western basin. The western basin had large room for sediments after ~40 m.y. of sediment starvation and continuous creation of accommodation space. The paleo-waterdepths of the paleo-Loppa High area at the transition from Ladinian to Carnian times range from 0 m east of the high, 0 m on the crest and ~400 m in the western basin.

From latest early Ladinian (horizon a) to middle Ladinian (horizon b) the western basin received large amount of sediments forming the characteristic wedge (*fig. 3.6*). During this period the water-depth on the crest of paleo-Loppa High was very shallow or even sub-aerially exposed and was dominated by sediment bypass. After the accommodation space in the western basin was filled in middle Ladinian, the Loppa High area continued its development as an entirety. Between horizon b and 10 the succession has a nearly equal thickness throughout the paleo-Loppa High area. The paleo-Loppa High had become a depocenter and subsidence of the whole area continuously increased the accommodation space. Glørstad-Clark et al. (2009) report of thickness increase towards Loppa High from Bjarmeland Platform, which supports that Loppa High have become a depocenter. *Figure 4.12* is a time-thickness map between horizons b and 11 that display an equal thickness throughout the Loppa High area.

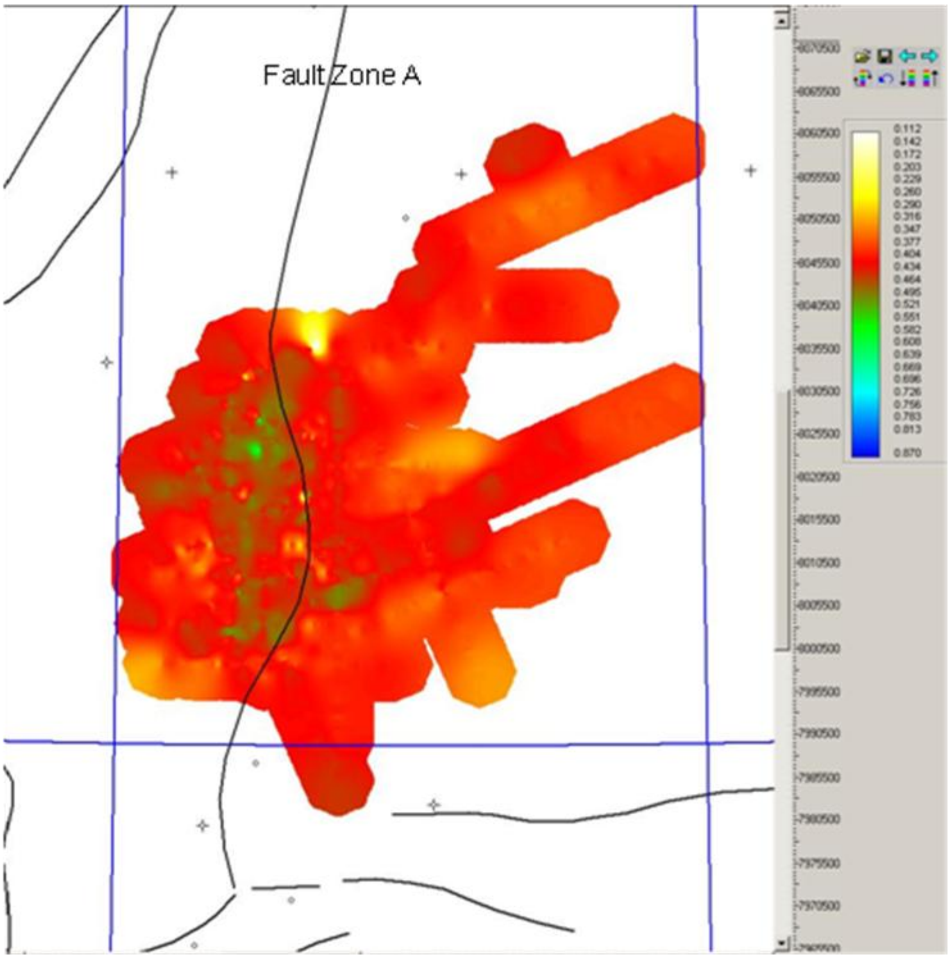


Fig. 4.12: Time-thickness map between horizons b and 11 in the paleo-Loppa High area.

The time period from middle Ladinian to middle Carnian show different depositional patterns in the northern and southern part of the study area. In north, west-east prograding clinoforms are entering the area, representing a new sediment source. The clinoforms are high and demands a minimum of 400-500 m of water-depth. The combination of the channel deposits found southward in the area reflect large lateral variations in deposition. The channel bodies as seen from seismic data are many and quite small, most likely reflecting/portraying a deltaplain (*fig. 4.13*).

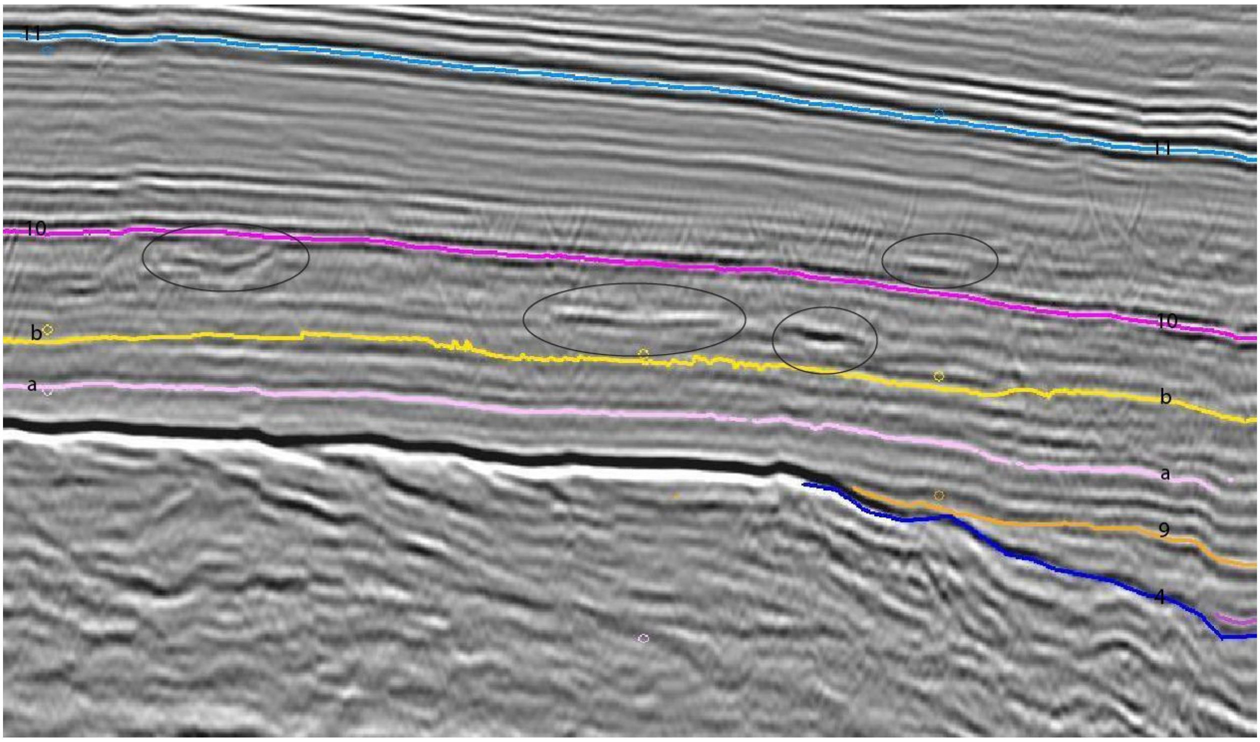


Fig. 4.13: Seismic line with channel interpretations highlighted in circles. Stratigraphic diagram with sequences and horizons are given in *figure 3.9*.

4.1.10 Latest early Ladinian-middle Ladinian (The wedge)

The wedge was deposited during sequence S4, but since it is one of the main findings in this study and not described previously, it is assigned to a different section in this thesis. The wedge is of crucial importance to understand the relief of the paleo-Loppa High and onset of sediment infill in the western basin. Horizon a, latest early Ladinian, marks the start of main sedimentation in the western basin. The sudden increased sedimentation rate in the western basin was due to the fact that sediments could bypass the area east of paleo-Loppa High, as sediment rate overcomed the rate of accommodation space generation. Accommodation space east of the high was filled in, whereas the western basin had remained underfilled since late Permian times. *Figure 4.14A* is a time thickness map showing how the thickness of the interval dramatically increases in the western basin.

As described in chapter *3.4.6 The wedge*, there is a possibility that Fault Zone A1 and faults B1a and B2a were active, and increased the accommodation space further during deposition of the wedge. Internal units within the wedge display an increase in thickness towards the faults (*fig. 4.14B*), indicating that the listric fault planes of Polheim Subplatform were activated at an earlier stage than first believed. Further investigation of this hypothesis is beyond the scope of this study. Regardless of the timing of fault movements of the listric faults on the Polheim Subplatform it resulted in a complication of the shape of the wedge and its internal composition.

The intra wedge horizons separate sequences of outbuilding into the western basin. They all show downlapping clinoforms units that build out on top of each other (*fig. 3.14*). The phases of outbuilding could be because of changes in sediment availability, changes in sea-level or movements on Fault Zone A and B. *Figure 4.14 C* is an arbitrary line in north south direction that shows how the wedge is thickening towards north, illustrating larger accommodation space.

The sediments deposited in the wedge entered the basin along the whole crest of paleo-Loppa High. However, due to less accommodation space in south, this area probably was full first and the sediments coming across the paleo-Loppa High area from east may have been directed northward before entering the wedge in the latest part of the deposition of the wedge. This may have allowed the southern area to start developing into an environment suitable for channels.

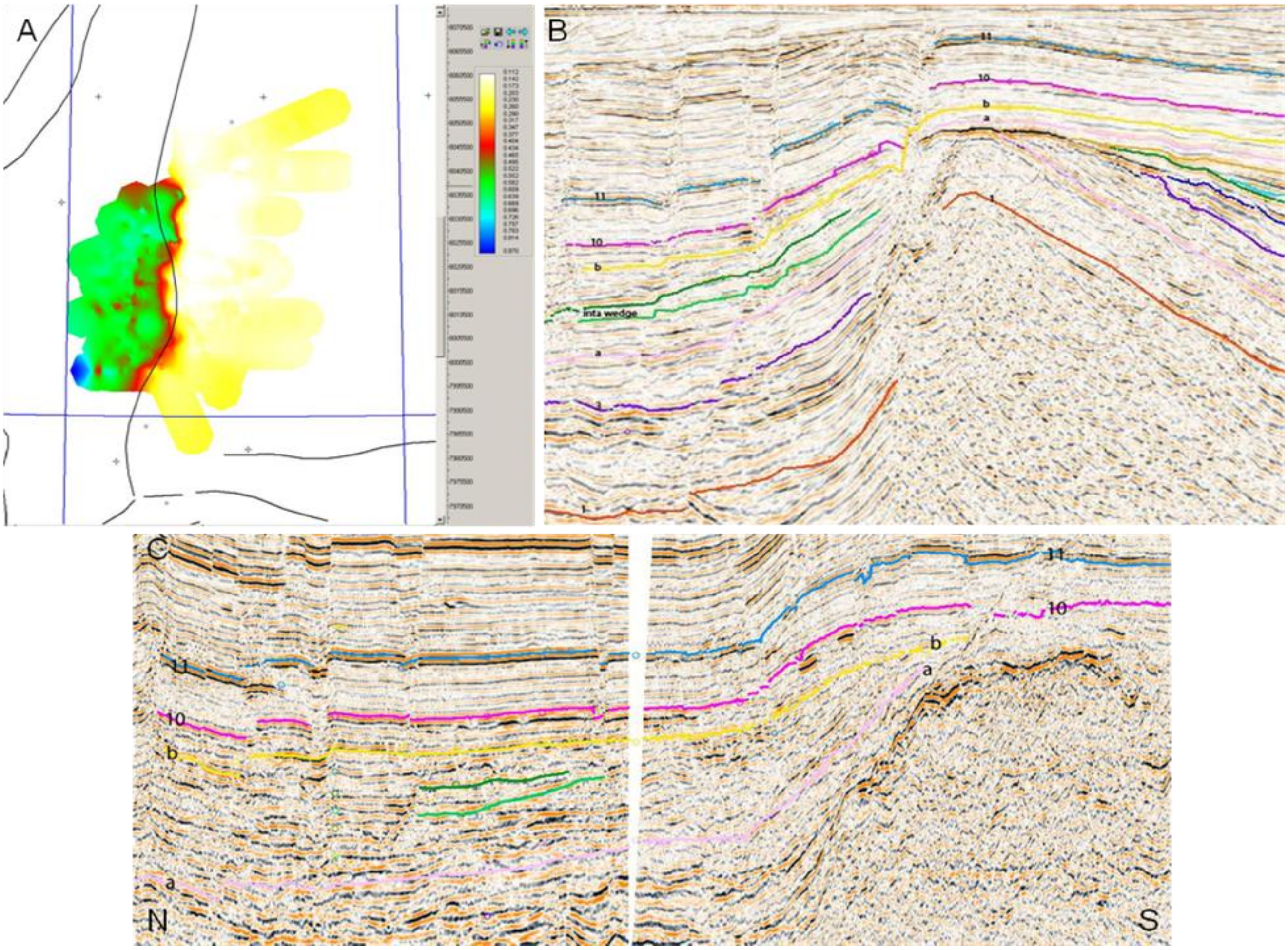


Fig. 4.14 A: Time-thickness map of the wedge. B: Seismic profile of the wedge showing layers and clinoforms. C: Arbitrary Line north-south direction showing thickness increase, and faults cutting the profile. Stratigraphic diagram with sequences and horizons are given in *figure 3.9*.

4.1.11 Middle Carnian- Norian (S5)

The deposits from Early Ladinian to Norian constitutes sequence S5. From Norian until today the chronostratigraphic chart shows non-deposition in most parts of the area (*fig. 4.1*). Tectonics and isostatic uplift events, in addition to glacial erosion are some of the factors contributing to the erosion (*fig. 3.15*). The erosion is bigger above paleo-Loppa High due to renewed uplift of the high.

Figure 4.9.C have restored the geometries of paleo-Loppa High at base Cretaceous time. Renewed fault activity have further uplifted the paleo-Loppa High structure, it was sub-aerially exposed and eroded. The Polehim Subplatform was downfaulted towards the Ringvassøy-Loppa Fault Complex in Triassic and Jurassic times making new space for sediment deposition.

Figure 4.15 is a summary of the geological history described above with emphasis on the tectonic evolution in the paleo-Loppa High area. The tectonic events are marked according to time and what area they affect. The tectonic events are moving westward in time which corresponds well to the regional western Barents Sea development (Faleide et al. 1993a,b, 2008).

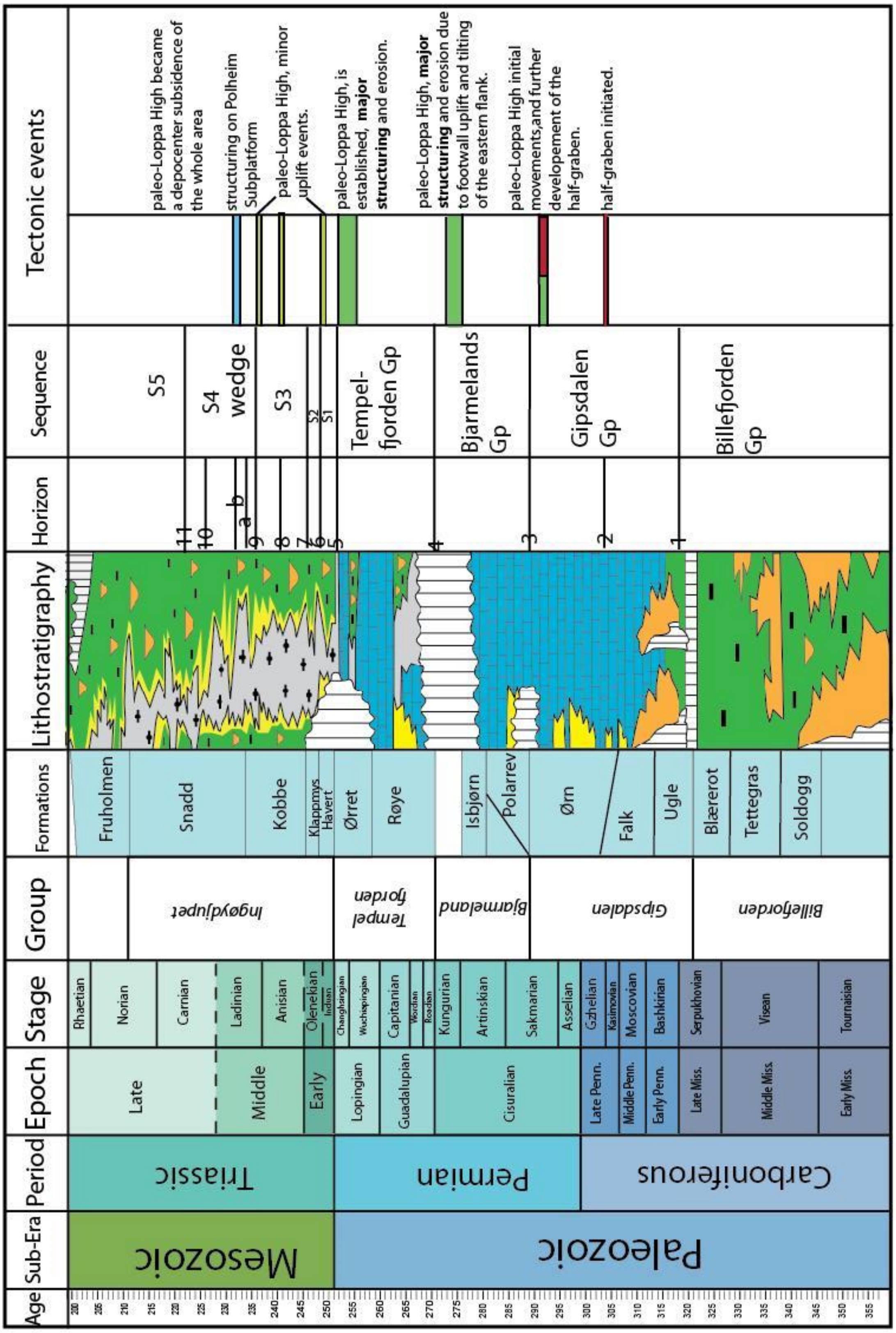


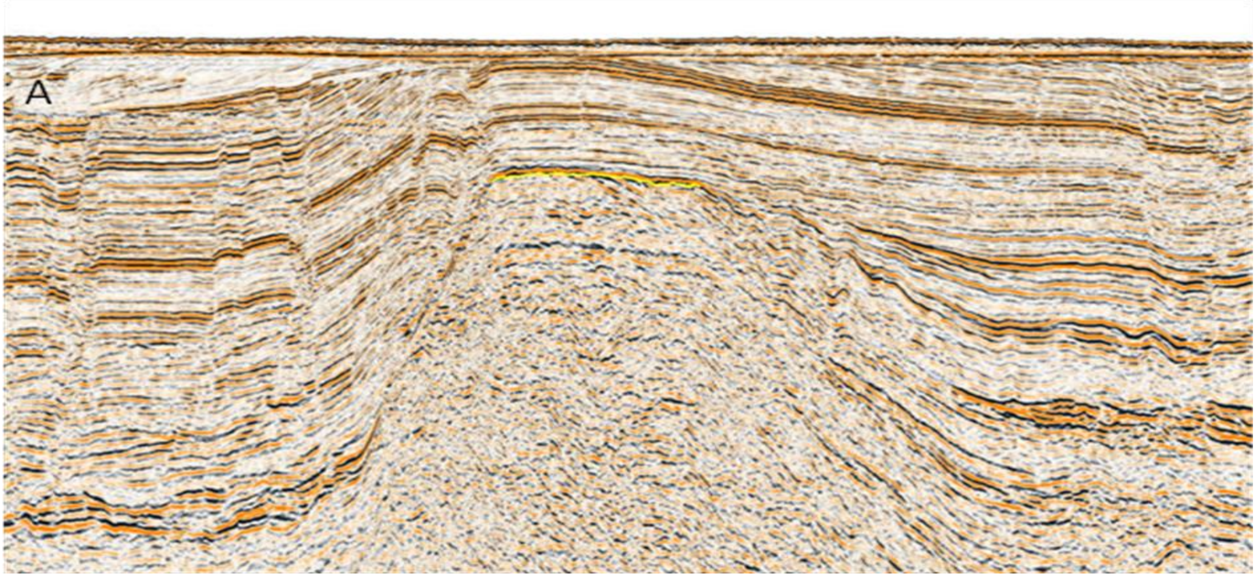
Fig. 4.15: Stratigraphic diagram with tectonic events (modified from Glørstad-Clark et al. 2009).

4.2 Shape of Fault Zone A and crest of paleo-Loppa High

Fault Zone A1 is the main fault creating the characteristic features of paleo-Loppa High. The fault zone was described in older publications (Faleide et al. 1993a,b), but recent seismic coverage has improved significantly both concerning quality and number of surveys. The crest of paleo-Loppa High constitutes the truncated Upper Paleozoic layers that make up a fairly flat surface. The crest is a high reflectivity surface, as a result of middle Triassic clastics that unconformably rests upon early Permian carbonates, also known as the Base Ladinian Unconformity (www.npd.no). The crest was mapped out in the study area, giving rise to a new interpretation of Fault Zone A1 (*fig. 4.16*).

Figure 4.16A shows a seismic profile with the interpretation of the paleo-Loppa High crest. The yellow line displays the mapped area. *Figure 4.16B* is a map of the interpreted seismic lines of the crest. Red line marks the new fault interpretation, while the black line is the old fault interpretation by Faleide et al. (1993a,b). The new fault line is more undulating in the southern part of the area, before it straightens out towards north. The deviation between the old and new fault line is largest in the northern part where the old fault interpretation is set to turn westward, while the new interpretation is fairly straight. The surface of the crest expands northward and seen from the time vs. color legend it dips in a southern direction. The crest could be divided into two blocks that have a few individual properties; the northern straight fault line and the southern undulating fault line. The crest has a larger surface in north and is situated at a shallower depth. The southern part lies deeper and the high have a narrower crest.

If the dip and height differences were formed at the initiation of paleo-Loppa High, the sediments deposited would drape the crest at and earlier stage in south than in north. However as shown in *figure 3.12* and *figure 4.10A* horizon 9 in general is the last horizon to onlap the crest in the whole area. From this we assume that the dip and height differences were formed at a younger stage. There are several factors that could contribute to this differential uplift. Paleo-Loppa High comes closer to the Bjørnøyrenna Fault Complex towards north. Major Late Jurassic-Early Cretaceous faulting , associated with this caused uplift of areas east of the fault complex. At a more regional scale the western Barents Sea experienced a north-south tilting due to uplift in north, both in Cretaceous and Cenozoic times (Faleide et al. 1993a,b, 1996).



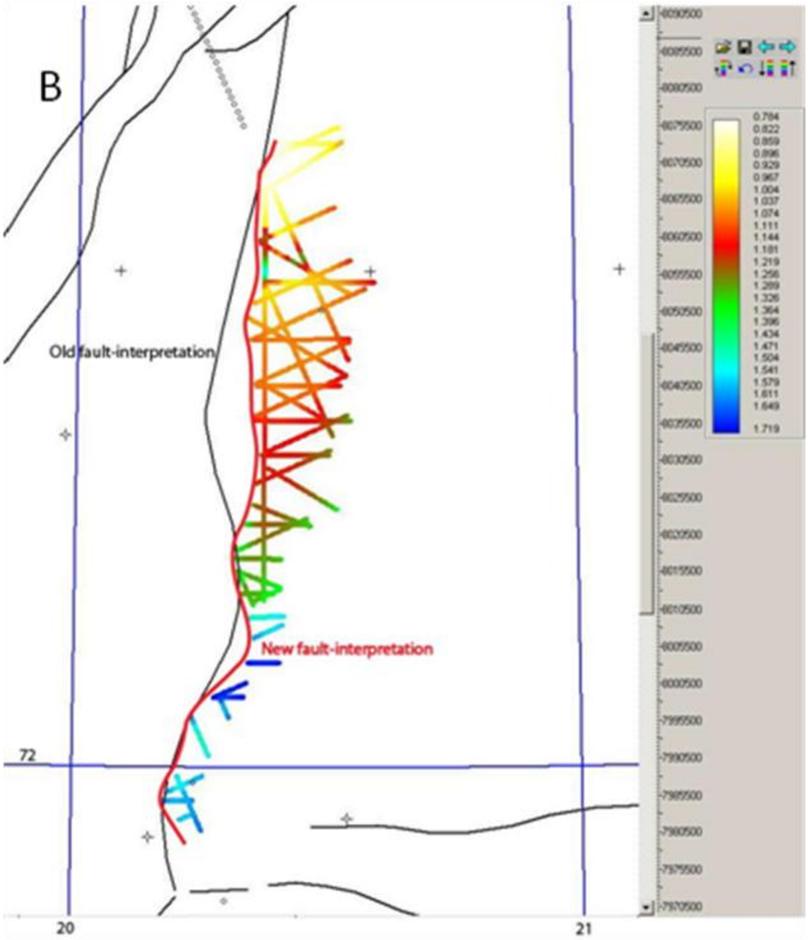


Fig. 4.16 A: Seismic profile with the interpretation of the paleo-Loppa High crest. The yellow line displays the mapped area.

*B: M*ap of the interpreted seismic lines of the crest. Red line marks the new fault interpretation, while the black line is the old fault interpretation.

4.3 Fault blocks

The eastern flank of the paleo-Loppa High has clear evidence that it is composed of several smaller fault blocks striking east to west (*fig. 4.10A*). Could the western side of paleo-Loppa High also have this division?

There are series of rotated fault blocks synthetic to Fault Zone A that strikes in north-south direction west of paleo-Loppa High. There are also several smaller faults that divide the area striking in east-west direction on this side of the fault. The wedge bounded by horizon a and b have a number of internal horizons that are hard to map out in north-south direction. An explanation could be that the underlying sequences have several smaller internal faults that make the substrate uneven and gives room for syn-tectonic sedimentation. Alternatively the faults are younger and cut trough the wedge after deposition and clutter up the depositional patterns making the internal horizons hard to track. *Figure 4.14 C* shows an arbitrary line in north-south direction. Two green horizons is interpreted in the wedge bounded by horizon a and b. The internal horizons are cut off by two faults cutting the profile apparently in an east west plane.

4. 4 Filling of the western basin

The proposed composition of the basin west of the paleo-Loppa High is quite reliable. However as there are no wells drilled deep enough west of Fault Zone A there will always be an uncertainty related to the western basin. The biggest uncertainty is related to the unit between horizon 3 and a, which in our interpretation covers middle Sakmarian to latest early Ladinian, a time period of 55 m.y. We have reasons to believe that most part of this unit consist of the Bjarmeland Group and condensed deposits of sequences Tempelfjorden, S1, S2 and S3, but we have not been able to identify horizon 4 (top Bjarmeland) or any of the other horizons 5-9 in the western basin. No wells are drilled deep enough in this area to confirm the lithology. A list of alternative entry points or sedimentation sources that could fill up the western basin is discussed below.

4.4.1 Alternative transportation routes

Paleo-Loppa High could have been uplifted and developed into a barrier to sediments already during deposition of the Bjarmeland Group. The Bjarmeland Group would then be even thinner in the western basin. The sediments could originate from the same source area as the sediments deposited on the eastern flank of Paleo-Loppa High, but they would have had to find a way around the barrier, and sequences S1, S2 and S3 would represent a thicker part of the unit. An investigation of the seismic south and north of the ridge have been performed for an alternative way in.

- The Hammerfest basin is located south of the study area. According to Faleide et al. (1993a,b) the sediments deposited in Late Permian-Early Triassic were prevented from entering the Polheim Subplatform by The Ringvassøy-Loppa Faultcomplex which at this time represented a barrier to sediments prograding from southeast and east (*fig. 2.2 B*).

- A lower-lying area between the two crest-structures *(chap. 4.2*) could allow sediment-transportation into the western basin at an earlier stage than originally believed. Sediments may have been able to pass through this opening and settle in the western basin.

- It is also possible that paleo-Loppa High were drowned at different times than what we have built our model on. This could lead to a different composition and Tempelfjorden, S1, S2 and S3 could be thicker than expected.

The sediment lithology of S1, S2 and S3 is very fine grained. Clinoforms towards Paleo-Loppa High suggests deep water starved for sediments, and thinning towards north and west (Glørstad-Clark et al. 2009). Thus, if any sediments reached the western basin through a northern, southern or lower lying passage the thickness of the unit is probably very small and hard to identify.

4.4.2 Alternative sediment sources

The unit between horizon 3 and *a* in the western basin could have a different source than the sediments east of Fault Zone A. Listed underneath is alternative sediment sources;

- The small island Bjørnøya is located north of paleo-Loppa High, situated on Stappen High. Loppa High and Stappen High are two structures that have a lot of similarities regarding depositional environment and tectonic history.

The source to sediments in the western basin located between horizon 3 and *a* could be Stappen High. Worsley et al. (2001) found two truncated surfaces on Bjørnøya and a following time gap of sediments. During the time-period between the transition from Bjarmeland to Tempelfjorden Group in Late Permian times, the Stappen High could have been sub-aerially exposed and eroded. The eroded sediments from Stappen High could end up in the western basin. The unit between horizon 3 and *a* on the western basin matches the timing of the erosional events on Stappen High, but there are no clear evidence of north south prograding sediments in the western basin. The seismic is very poor so it is not unlikely that there are sediments from this source. The thickness of the unit between horizon 3 and a is northward thickening. It would be interesting to further investigate the eroded sediments transport direction on Stappen High.

- Parts of the unit between horizon 3 and *a* could be locally derived from the fault zone and the crest of paleo-Loppa High. However due to the eastern tilt of the paleo-Loppa High flank it is unlikely that significant amounts of sediments have been transported westward.

- Another alternative sediment source is Greenland. *Figure 4.17A* shows the paleogeographic reconstruction of the Barents Sea in Artinskian time. Greenland was not situated far away from the paleo-Loppa High area. In theory sediments from Greenland could find its way into the sub-basin west of paleo-Loppa High, but it is left to be proven that clastic systems originating in Greenland have prograded to these areas.

In Tatarian (*fig. 4.17B*) (Late Permian) shallow marine sandstones along basin margins are found in East Greenland, but these are only local systems (Stemmerik 2000).

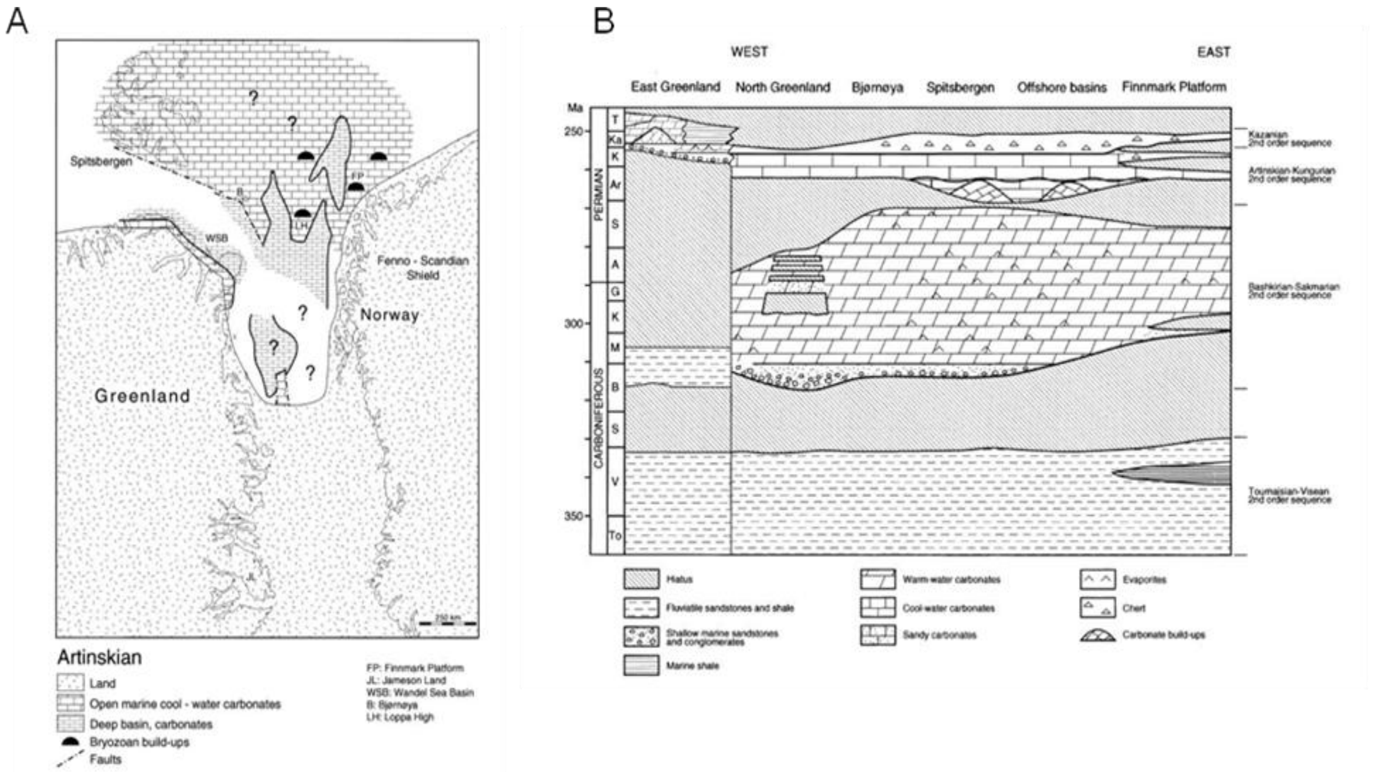


Fig. 4.17 A: Map of the paleogeographic reconstruction of the Artinskian (from Stemmerik 2000). B: Time distribution of depositional sequences (from Stemmerik 2000).

4.5 Sediments eroded from the crest of paleo-Loppa High

Due to footwall uplift and rotation of the paleo-Loppa High, several formations were eroded from the crest of the high. The crest of paleo-Loppa High was sub-aerially exposed from Kungurian to Anisian times. Erosion removed the top and grinded it down to the smooth surface we observe on seismic data. It is likely that most of the eroded sediments were transported eastward as the topography of paleo-Loppa High was tilted in this way.

Evidence of sediments originating from paleo-Loppa High is found in sequences S1, S2, and S3. However the west-east prograding clinoform systems found are small and does not explain where the majority of the sediments eroded were transported. It is a significant amount of sediments that have been removed from the paleo-Loppa High, but compared to the sediment influx from south and south-east it is possible that the locally eroded sediments have disappeared in these volumes, and that the seismic resolution is not good enough for us to distinguish the two sources. Another possibility is that the locally eroded sediments could have been transported away from the paleo-Loppa High area and been deposited elsewhere. A river or submarine currents could have transported the sediments over a significant distance. Hammerfest Basin has been considered as a possible area of deposition but no evidence has been found of the eroded sediments. A further search of the locally eroded sediments is beyond the scope of this study.

5. Summary

This thesis describes the geological history of the paleo-Loppa High with emphasis on tectonic events and depositional patterns from late Carboniferous through Triassic times. Seismic facies have been interpreted, and linked to previously studied and mapped data from the basin east of paleo-Loppa High and further correlated to seismic sections west of paleo-Loppa High. Backstripping and structural restoration have been performed to enhance the understanding of the area at a given time-step. Depositional patterns and structural geometries revealed in the backstripping reflect tectonic activity and intervals of relative tectonic quiescence, the context of these factors was important to correlate the succession over to the western side of paleo-Loppa High and to construct the geological history of the area that is reliable.

The late Carboniferous to Triassic geological history can be divided into 8 stages related to the tectonic events and the depositional patterns of the following succession:

1. Most of the Carboniferous was characterized by tectonic stable conditions within the study area. Sequences of uniform thickness were deposited throughout the area until late Carboniferous, horizon 2.
2. In late Carboniferous (Late Pennsylvanian?) the half-graben, situated at the base of the eastern flank, south on paleo-Loppa High was initiated. The upper Gipsdalen unit is a carbonate succession that is thickening in the half-graben. Sediments were deposited throughout the area.
3. In Early Permian (early Sakmarian?) movements in the half-graben continued, in addition to the initial movements on the paleo-Loppa High Fault Zone A1. Evaporites were found in the half-graben and a carbonate platform developed. Sediments were deposited throughout the area until the end of the Bjarmeland Group.
4. In Kungurian times the first of two main tectonic events in the formation of the paleo-Loppa High structure occurred. The high were sub-aerially exposed and a barrier to sediments prograding from east. The depositional geometries changed, and the

Tempelfjorden Group onlapped paleo-Loppa High. The western basin was in a state of sediment starvation.

5. In late Permian (Guadalupian) the second main event occurred establishing the paleo-Loppa High structure. This event is probably the largest of the two main events structuring the paleo-Loppa High. Faulting lead to tilting of the eastern flank, which exposed older sediments sub-aerially and erosional processes grinded the crest down to the nearly flat surface observed on seismic.
6. Early-middle Triassic (Induan through early Ladinian) times were regionally a relatively calm period period. This goes for the paleo-Loppa high area as well, except for three minor uplift events. Sediments deposited in this time-period onlapped the eastern flank, only draping the crest occasionally. The western basin was still in a state of sediment starvation. The depositional patterns of this timeperiod was influenced by the paleo-waterdepths that ranged from ~300 m east of the high, 0 m on the crest and ~500 m in the western basin.
7. In middle Ladinian times the eastern side of the paleo-Loppa High was filled with sediments, and sediments started entering the western basin. The western basin was filled in a short amount of time, approximately 4 m.y. Structuring of the Polheim Subplatform may have started in this time-period. The paleo-waterdepth during the deposition of the wedge was 0 m east of the high and on the crest and ~400 m in the western basin.
8. The rest of Triassic times were characterized by calm and stable conditions. The former paleo-high developed into a depocenter.

The post-Triassic evolution of the Loppa High area was complex, involving several phases of faulting, uplift and erosion.

Correlating the local geological history of the Loppa High area with the regional geology we find that the late Paleozoic-early Mesozoic evolution of the paleo-Loppa High was linked both to the basin formation and infilling of the central and eastern Barents Sea, and to regional extension within the NE Atlantic region where rift basins formed between Norway and Greenland.

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