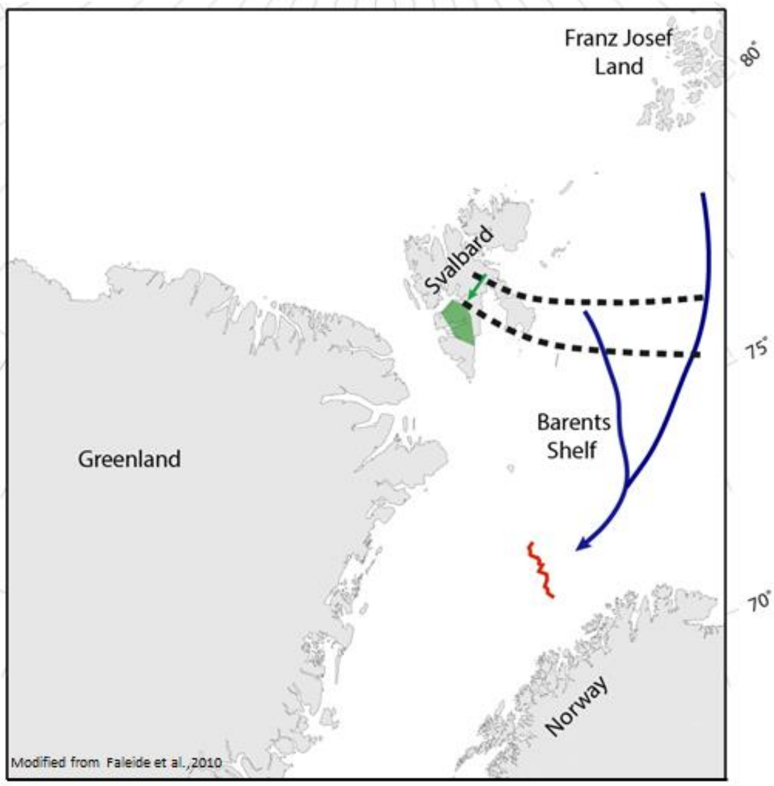
Master Thesis, Department of Geosciences

Late Paleocene – earliest Eocene prograding system in the SW Barents Sea

Bjørnar Margido Prøis



Modified from Faleide et al.,2010

UNIVERSITY OF OSLO

FACULTY OF MATHEMATICS AND NATURAL SCIENCES

Late Paleocene – earliest Eocene prograding system in the SW Barents Sea

Bjørnar Margido Prøis

Master Thesis in Geosciences

Discipline: Petroleum Geology and Petroleum Geophysics

Department of Geosciences

Faculty of Mathematics and Natural Sciences

University of Oslo

June 2015

Abstract

Two-dimensional seismic data and a regional seismic sequence analysis of the Paleocene succession in the southwestern Barents Sea are presented and discussed. The Paleogene Torsk Formation is bounded by a Cretaceous – Paleogene hiatus at the base and an erosional truncation at the top. Well data was tied to the seismic to define the base and top of the Torsk Formation.

Based on reflection terminations, three Paleocene progradational units are defined within the Torsk Formation in the Hammerfest and Tromsø basins. A regional depositional model for the prograding system is outlined and applied to describe the depositional history for the Paleocene succession in the SW Barents Sea.

The base of the Paleocene succession is an unconformity related to uplift and erosion. Subsidence of large parts of the Barents Shelf generated accommodation, and deposition initiated in Late Paleocene. The deposition in earliest Paleocene started out as widespread aggradational shelf deposits over large parts of the Barents Shelf, and was later succeeded by a Paleocene progradational system. The system prograded from ENE towards WSW, and uplifted areas on the northern Barents Shelf acted as sediment source. The geometry of the prograding system towards Loppa High shows that the area was a part of the shallow Barents Shelf during the Early Paleogene, which differs from earlier publications. The area was not a topographic or bathymetric high, but part of the wider Barents Shelf depocenter. Most of the Early Paleogene succession is missing in the Barents Sea due to Late Cenozoic uplift and erosion, but it is expected that the prograding system covered large areas with a significant thickness prior to uplift and erosion.

Abstract

Preface

This master thesis is a part of the two year master program at the University of Oslo department of Geoscience, “Petroleum Geology and Petroleum Geophysics”. The supervisors for this thesis have been Professor Jan Inge Faleide, Associate Professor Ivar Midtkandal and Professor Emeritus Johan Petter Nystuen.

Acknowledgements

First off, I would like to thank my supervisors; Associate Professor Ivar Midtkandal, Professor Jan Inge Faleide and Professor Emeritus Johan Petter Nystuen for the guidance and discussions during the work with the thesis. Thanks to TGS and Fugro for access to selected sismic 2D lines from the NBR survey, and to Senior Engineer Michel Heeremans for the preparations of the data set used in the thesis.

Thanks to all my fellow students for the discussions and collaboration through the master program. Finally, thanks to all my friends and family for support during the work with my thesis.

Preface and Acknowledgements

Table of Contents

1 Introduction

The Barents Sea is a shallow epicontinental sea located in the northwestern part of the Eurasian continental shelf (figure 1.1). The Barents Sea covers an area of approximately 1,3 million km^2 and have an average water depth of ca. 300m (Doré, 1995). The Barents Sea is bounded to the north by Franz Josef Land and Svalbard archipelagos, to the east by Novaya Zemlya, to the south by the northern coast of Norway and Russia and to the west by the Norwegian – Greenland Sea (Faleide et al., 1984). There have been several stages of tectonism since the Caledonian orogeny. The western Barents Sea was most influenced by the tectonic events during the Mesozoic and Cenozoic (Gabrielsen et al., 1990).

The main objective of this study is to analyze a Late Paleocene – earliest Eocene prograding system in the SW Barents Sea. Late Cenozoic uplift and erosion has led to limited preservation of the Paleogene succession in the Barents Sea. The main study area is the Hammerfest Basin, Tromsø Basin and the Loppa High where parts of the Paleogene succession are preserved. The studied prograding system is of mainly of Late Paleocene age but the latest deposits in the system are from the earliest Early Eocene, in this thesis it will only be referred to as Paleocene.

The data used for this work is 2D seismic lines supplemented with well data from Norwegian Petroleum Directorate (NPD). Based on seismic sequence analysis of the Paleocene succession, the infill pattern, depositional environment, progradational direction and timing of deposition will be outlined and discussed. On a more regional scale, a sediment source and mechanism of creation of accommodation will be suggested.

The importance of Loppa High relative to the depositional environment will be discussed, and is intended to increase knowledge of the complex uplift and subsidence history of the Loppa High.

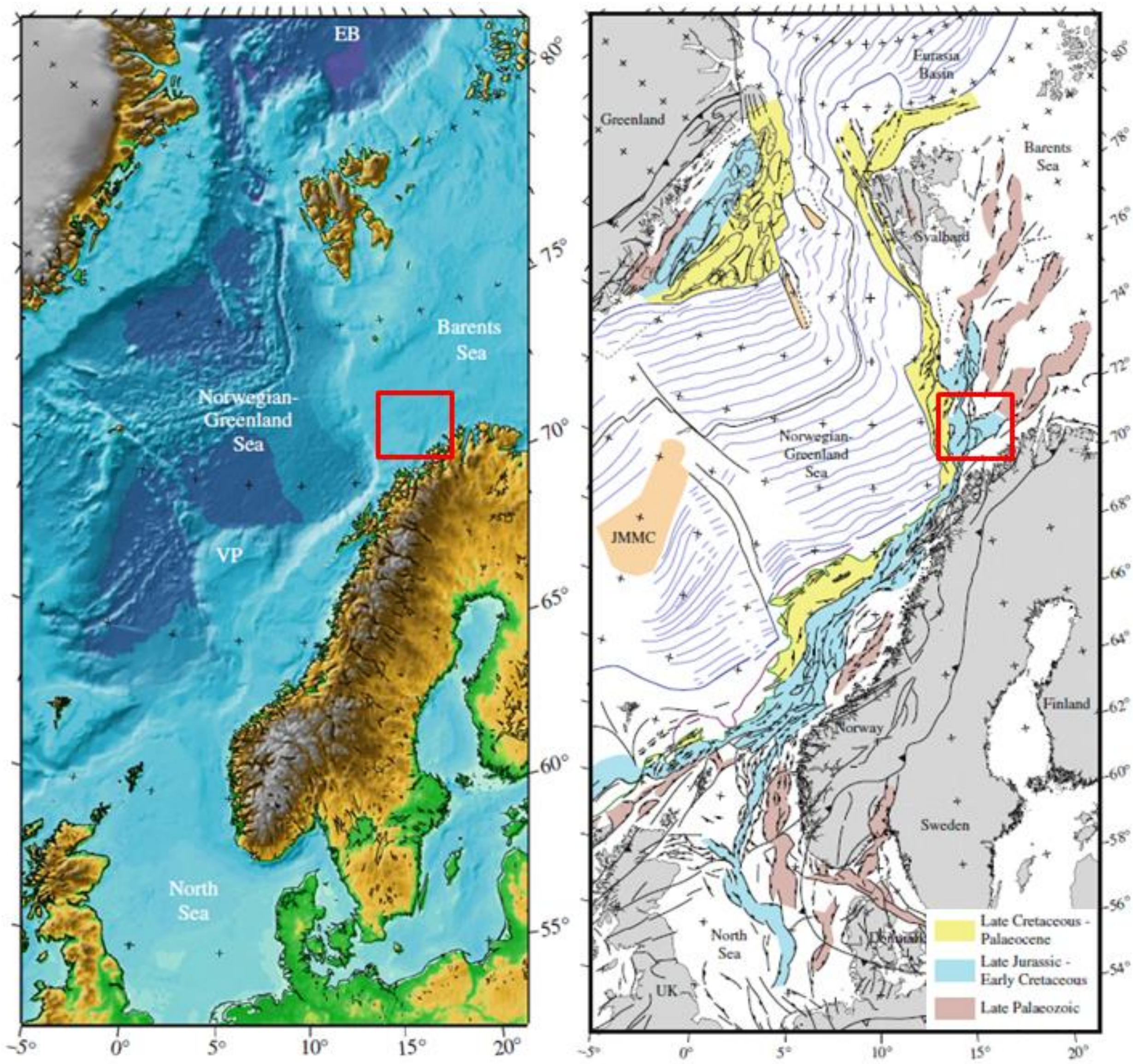


Figure 1.1: Regional settings with the location of the study area marked with a red square. The map to the right shows the structural elements in the western Barents Sea and the structural elements along the rest of the Norwegian coast. Modified from Faleide et al. (2010).

2 Geological framework

The regional geology of the Barents Sea area is generally well understood and has been the topic for several studies throughout the years (Faleide et al., 1984; Gabrielsen, 1984; Worsley et al., 1988; Gabrielsen et al., 1990; Faleide et al., 1993; Faleide et al., 1996; Gudlaugsson et al., 1998; Worsley, 2008; Smelror et al., 2009). The link between Svalbard and the Barents Sea is well known, and Svalbard can often be used as a surface analog to the subsurface deposits in the Barents Sea (Nøttvedt et al., 1993; Worsley, 2008).

The main tectonic phases which have influenced the geological settings for the Barents Shelf are the Timanian, Caledonian and Uralian orogenies, the proto-Atlantic rifting and the opening of the Euramerican Basin and northern North Atlantic Ocean (Smelror et al., 2009). The Caledonian orogeny in mid Paleozoic time represents the closure of the Iapetus Ocean. The collision between the Laurentian plate and the Baltic plate formed the Laurasian continent. The Uralian orogeny in Permian – Early Triassic represents the collision between the Laurasian continent and Western Siberia and was the final element in the fusion of the supercontinent Pangea (Doré, 1995). In Late Paleozoic and Mesozoic extensional tectonic influenced the Barents Sea as a result of breakup of continents, and in Cenozoic the opening of the Norwegian-Greenland Sea and Eurasia basin led to development of the western margin of the Barents Sea as a sheared margin. The eastern margin of the Barents Sea was mainly influenced by the Uralian orogeny, while the western Barents Sea was more influenced by the post-Caledonian rifting phases (Smelror et al., 2009).

The sedimentary cover in the Barents Sea is of late Paleozoic – Cenozoic origin and in places it exceeds 15 km in thickness (Faleide et al., 1993). The depositional environment was dependent on tectonic settings and climate. In Devonian, Carboniferous and Permian carbonate deposition dominated in large areas in the shallow epicontinental sea, while from the Triassic and onwards clastic deposition was dominant (Doré, 1995).

2.1 Regional setting

The Barents Sea can be divided into two major geological provinces, an eastern and a western, where the eastern province generally lies within the Russian borders and the western lies within the Norwegian borders. The eastern province can be divided into the South and North Barents Basin. The development of this area is linked to the development of the Uralides, Timian-Pechora and Novaya Zemlya (Worsley, 2008). The western province, which is the area of interest in this study, shows a more complex tectonic development with several basins, platforms and structural highs.

Faleide et al. (1993) divided the western Barents Sea into three main geological provinces: 1) The Svalbard Platform covered by a relatively flat laying upper Palaeozoic and Mesozoic sediment succession; 2) A basin province with several subbasins and highs between the Svalbard Platform and the Norwegian coast, where the structural relief increase westward. The preserved sediments in this area are mainly from Jurassic – Cretaceous, and westward more Paleocene – Eocene sediments are preserved; 3) The continental margin which can be subdivided into three main segments: a) a sheared margin to the south along the Senja Fracture Zone, b) a central rifted complex associated with volcanism, and c) an initially sheared and then rifted margin to the north along the Hornsund Fault Zone.

2.2 Local Settings

The area of the present study is the basin province between the Svalbard Platform and the Norwegian coast and it consists of several basins separated by intrabasinal highs (figure 2.1). The basin province can be divided to a western and eastern province separated by the Ringvassøy-Loppa Fault Complex (FC), Bjørnøyrenna FC and Leirdjupet FC, which are some of the main Jurassic – Cretaceous faults (Faleide et al., 1993). The western basins are deeper as a result of the continuous Cretaceous rifting. The basins in this area are the Harstad, Tromsø, Bjørnøya and Sørvestnaget basins and they are separated by the Senja Ridge, Veslemøy High and Stappen High, which all are intrabasinal highs (Faleide et al., 1993). The eastern basins are not as affected by the Cretaceous subsidence and are hence not as deep. The basins and highs in this area are the Finnmark Platform, Hammerfest Basin, Loppa High and Fingerdjupet Subbasin (Faleide et al., 1993).

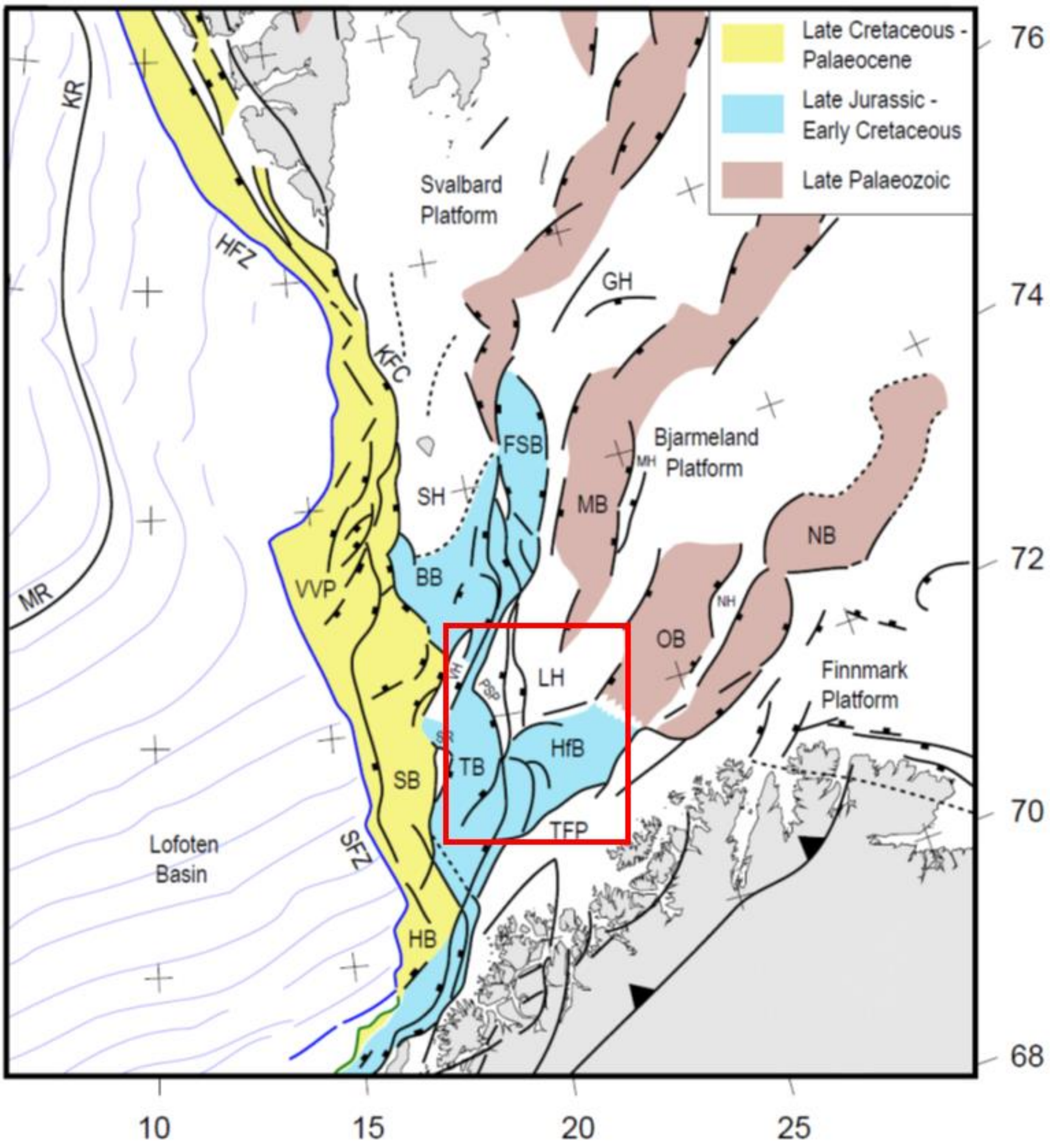


Figure 2.1: Structural elements of the western Barents Sea with the study area marked with red square. Modified from Faleide et al. (2010).

The main focus of this study is the Hammerfest and Tromsø basins where prograding units of the Paleocene succession can be found (figure 2.2). The following short descriptions of the most important structural elements in the study area are mainly based on Gabrielsen et al. (1990) and only information from elsewhere is cited:

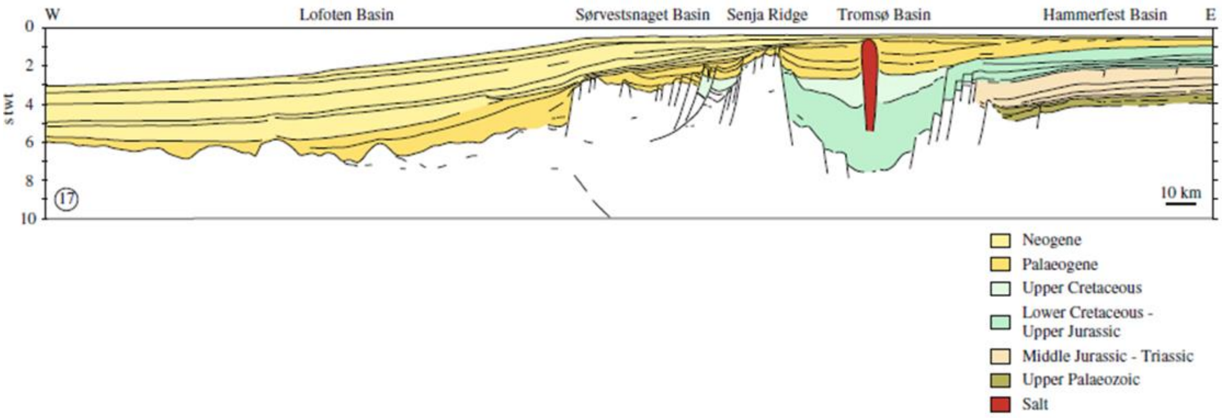


Figure 2.2: Seismic line displaying the Tromsø and Hammerfest basins with the Senja Ridge, Sørvestnaget Basin and the oceanic Lofoten Basin to the W (Faleide et al., 2010).

2.2.1 Hammerfest basin

The Hammerfest Basin is one of the Mesozoic basins not greatly affected by the Cretaceous-Tertiary subsidence. The basin has an ENE-WSW striking axis and is a relatively shallow basin. The western limit of the basin is the southern segment of the Ringvassøy-Loppa FC separating the basin from the Tromsø Basin, and to the east it borders to the Bjarmeland Platform. The northern limit is the Loppa High and the Asterias FC separated the basin and the high. To the south the Troms-Finnmark FC separates the basin from the Finnmark Platform.

The basin is a potential structural connection between the rift basins to the west and the areas to the east affected by Late Palaeozoic extension (Gudlaugson et al., 1998). The Hammerfest Basin was separated from the Finnmark Platform in Late Carboniferous, and the basin has been a depocentre from the Triassic. The present day outline was established in Middle Jurassic – Early Cretaceous.

2.2.2 Tromsø Basin

The Tromsø Basin is one of the deep Cretaceous-Tertiary basins, and it has a NNE-SSW trending axis. The basin terminates against the Troms-Finnmark FC to the southeast. To the north the Veslemøy High separates the basin from the Bjørnøya Basin. The western border is the Senja Ridge and to the east the Ringvassøy-Loppa FC separates the basin from the Hammerfest Basin and Loppa High.

The basin has a thick Cretaceous succession because of the continuous subsidence throughout Cretaceous. Salt diapirs along the axis in the south-central part of the basin indicate that the Late Paleozoic succession can be found in the basin.

2.2.3 Loppa High

Loppa High has a diamond-shaped outline and consists of an eastern platform and a crestal western and northwestern margin. The high is separated from the Hammerfest Basin to the south by the Asterias FC, and by a monocline towards the Hammerfest Basin and Bjarmeland Platform to the east and southeast. The western border of the high is the Ringvassøy-Loppa FC and Bjørnøyrenna FC, and the northeastern limit is marked by the Svalis Dome and the Maud Basin.

The area has been through several phases of uplift and subsidence, and subsequent tilting and erosion (Larssen et al., 2005). The Loppa High was formed in lates Jurassic – Early Cretaceous, and during most of the Cretaceous the Loppa High was an island and erosion probably led to deposition of deep water clastic fans in the northern Tromsø Basin.

2.2.4 Ringvassøy-Loppa Fault Complex

The southern part of the fault complex makes up the transition between the Hammerfest and Tromsø Basin, and south of this the fault complex merges with the Troms-Finnmark FC. To the north the fault complex defines the western border of the Loppa High. The regionally strike is N-S and is best defined by the westerly major faults.

The main faults are of Paleozoic and older origin but has been reactivated several times (Larssen et al., 2005). The main subsidence initiated in Middle Jurassic and ended in Early Cretaceous, and reactivation also took place in Late Cretaceous and Early Tertiary.

2.2.5 Asterias Fault Complex

The fault complex separates the Loppa High from the Hammerfest Basin. The regionally strike is east-west, and faulting with downthrow to the south dominates the fault complex.

The fault complex has been important for the development of the Hammerfest Basin since Mid Jurassic times. Depositions as young as early Eocene have been affected by faulting.

2.3 Development

Since the Caledonian orogenic movement terminated in Early Devon (Gabrielsen et al., 1990) the development of the SW Barents Sea has been affected by three major rift phases: 1) Late Devonian – Carboniferous, 2) Middle Jurassic – Early Cretaceous, and 3) Early Tertiary; each rift phase consists of several tectonic pulses (Faleide et al., 1993). These tectonic events together with changes in eustatic sea level and climate conditions have controlled the sedimentation and erosion in the area. The most important tectonic phases are displayed together with the lithostratigraphy of the Barents Sea in figure 2.3

The first of the post Caledonian rift phases, in Late Devon – Carboniferous, resulted in northeast to north striking extensional basins (Gudlaugsson et al., 1998), and the major regional fault zones was probably established during this period or earlier (Gabrielsen et al., 1990). In late Carboniferous the northeastern part had become stable and a regional carbonate platform was established, evaporites were also deposited in some basins (Gabrielsen et al., 1990). The Triassic to Early Jurassic was a tectonic quiet period (Gabrielsen et al., 1990) and sandstones deposited in the Early-Middle Jurassic are the main reservoir rock in the Barents Sea.

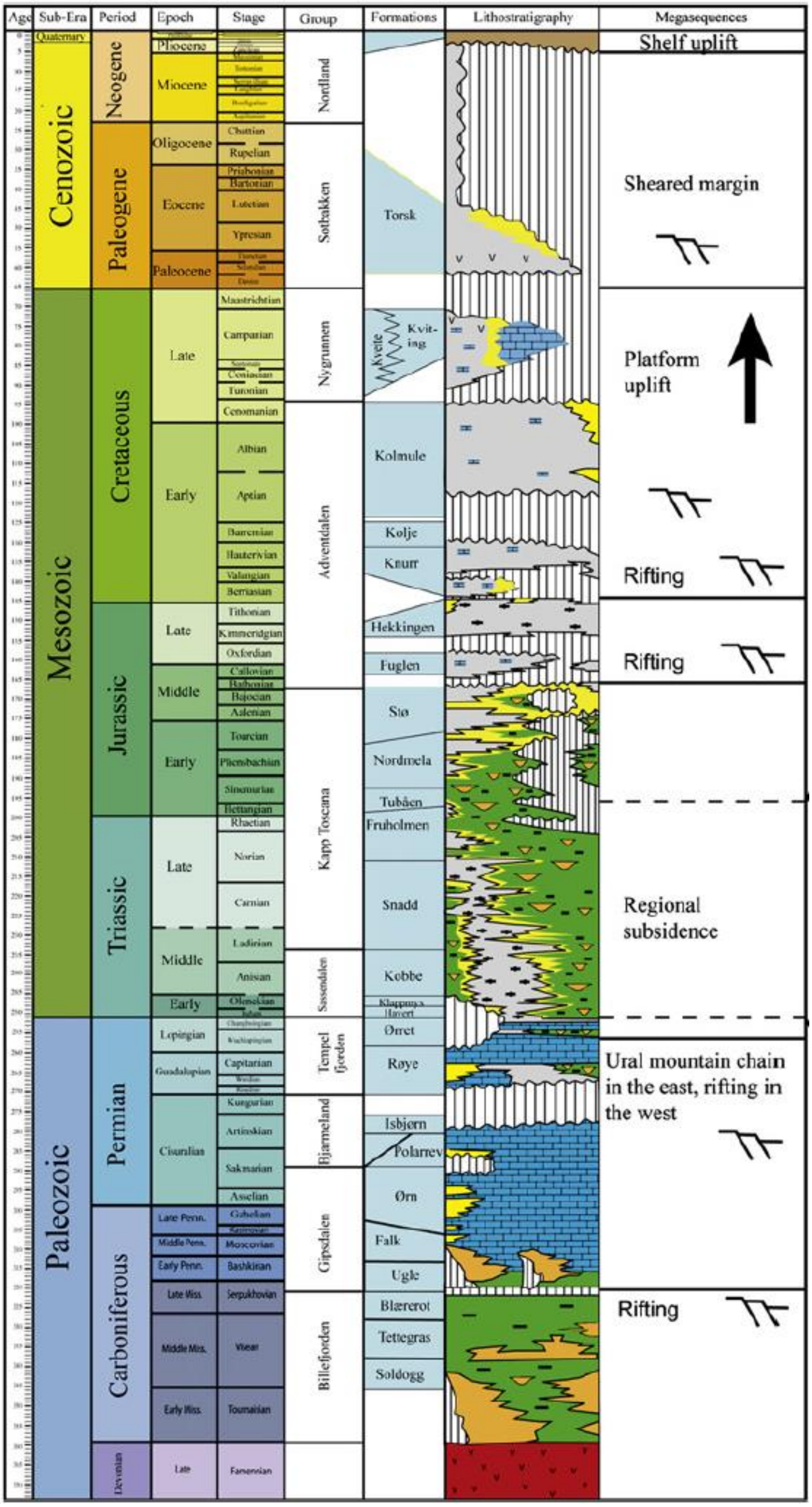


Figure 2.3: Lithostratigraphy for the western Barents Sea (Glørstad-Clark et al., 2010).

In Middle Jurassic block faulting initiated again and increased from Late Jurassic to Early Cretaceous (Gabrielsen et al., 1990). Late Jurassic shales were deposited in restricted basins created between tilted fault blocks, and the hiatus between Middle Jurassic sandstones and Late Jurassic shales marks the onset of the rifting (Faleide et al., 1993). The shales were deposited in a quiet deep marine environment (Worsley et al., 1988). The organic rich Hekkingen Formation from this period is an important source rock in the Barents Sea. Several tectonic phases affected the area in Early Cretaceous and extensional faulting with large downthrow to the west resulted in development of the deep Tromsø and western Bjørnøya basins (Faleide et al., 1993). The area east of Tromsø and western Bjørnøya basins was not as affected by the large subsidence in Cretaceous, and this led to continuous sedimentation in the Tromsø Basin through most of the period, while the areas further east were transgressed in times of maximum sea level with the formation of condensed sections in the successions (Faleide et al., 1993). After a depositional break at the Cretaceous-Tertiary transition the Paleocene sequence was deposited as a sheet over most of the western Barents Sea (Worsley et al., 1988).

The western margin of the Barents Sea developed to be a sheared margin during Eocene because of seafloor spreading in the Norwegian-Greenland Sea and the Eurasia Basin. The margin consists of two large sheared segments; the southern Senja Fracture Zone and the northern Hornsund Fault Zone, and a central rifted segment; the Vestbakken Volcanic Province (Faleide et al., 2008). The Vestbakken Volcanic Province was located in a pull-apart setting and experienced rifting and volcanism in this period (Faleide et al., 1993).

Since Paleocene and the opening of the Norwegian-Greenland Sea the Barents Sea has been uplifted, and erosion has taken place. In Pliocene – Pleistocene, continental glaciation in the Barents Sea led to high erosion rates (Dimakis et al., 1998). The eroded sediments were transported to the northern and western margins and deposited in huge fans, such as the Bjørnøya and Storfjorden Fan (Faleide et al., 1996). The erosion estimate varies from 1000-1500 m in the Hammerfest Basin to more than 3000 m at the Stappen High (Faleide et al., 1993), indicating increase in uplift and erosion towards north. The uplift and erosion had big influence on the petroleum potential in the Barents Sea (Doré and Jensen, 1996), and can explain predominance of gas over oil in the Barents Sea.

2.4 Early Paleogene setting

Paleogene deposits are only preserved in a few basins in the SW Barents Sea (figure 2.4) because of Late Cenozoic uplift and erosion. Early Paleogene deposits are also found on the surface in the Paleogene Central Basin on Spitsbergen. This gives limited information of the Paleocene depositional environment.

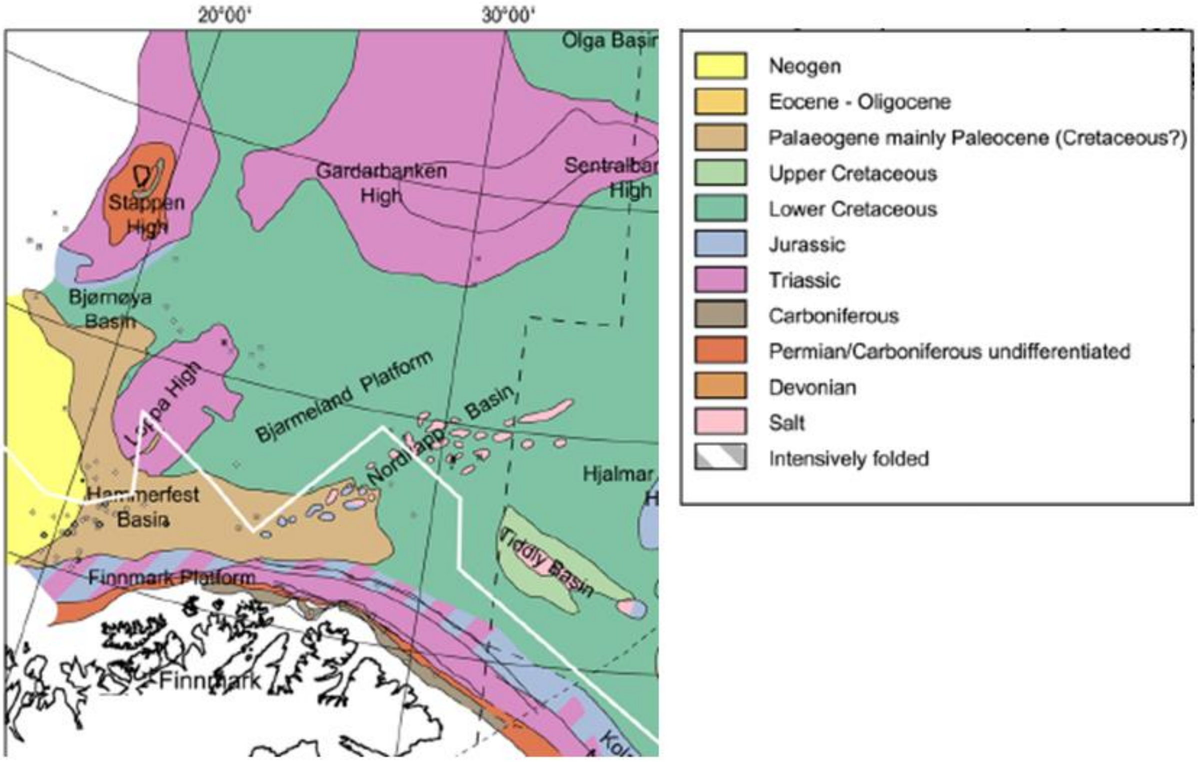


Figure 2.4: Subcrop map below Quaternary for the SW Barents Sea (Henriksen et al., 2011).

The earliest deposits of the Paleogene succession is deposited over the Cretaceous – Paleogene hiatus in the SW Barents Sea. The hiatus is related to uplift and erosion, and the Late Cretaceous succession is thin or absent over large parts of the western Barents Sea (Faleide et al., 1993; Nagy et al., 2000). This implies that the Barents Shelf must have subsided prior to the deposition started in Late Paleocene.

The Early Paleogene succession makes up the Torsk Formation of the Sotbakken Group in the SW Barents Sea. The Torsk Formation is a uniform sequence of light to medium grey claystones (Worsley et al., 1988). The Torsk Formation is interpreted to be deposited in middle to upper Bathyal environment (Nagy et al., 1997; Nagy et al., 2000).

The Early Paleogene development of the Paleogene Central Basin on Spitsbergen is described in previous papers (Steel et al., 1985; Müller and Spielhagen, 1990). Because of erosion of the Early Paleogene succession north of the study area (figure 2.4) up to the Paleogene Central Basin on Spitsbergen it is not known if the deposits from these two areas are part of the same system. It is mainly the Firkanten, Basilika and Grumantbyen formations which are relevant for this study since they are deposited approximately simultaneously with the Paleocene prograding system in the SW Barents Sea. These formations are in general deposited in a shallower environment than the deposits in the SW Barents Sea (Nøttvedt, 1985; Steel et al., 1985; Müller and Spielhagen, 1990).

The Paleogene was otherwise characterized by the seafloor spreading in the Norwegian – Greenland Sea and the Eurasia Basin, and this led to the development of the sheared western margin of the western Barents Sea (Faleide et al., 1993).

3 Data and Methods

3.1 Data

The seismic data used in this thesis are from the NBR-survey from TGS and Fugro while the well data are from NPD.

Most wells drilled in the Barents Sea are located in the Hammerfest and Tromsø basins (figure 3.1), which is the primary study area of this thesis. Hence, the interpretation of the base and top of the Torsk Formation is supported by several wells which give a greater certainty to the interpretation. The erosional unconformity at the base and top of the Torsk Formation corresponds to the “Base Tertiary” and the “upper regional unconformity”, URU, of the area (Worsley et al., 1988).

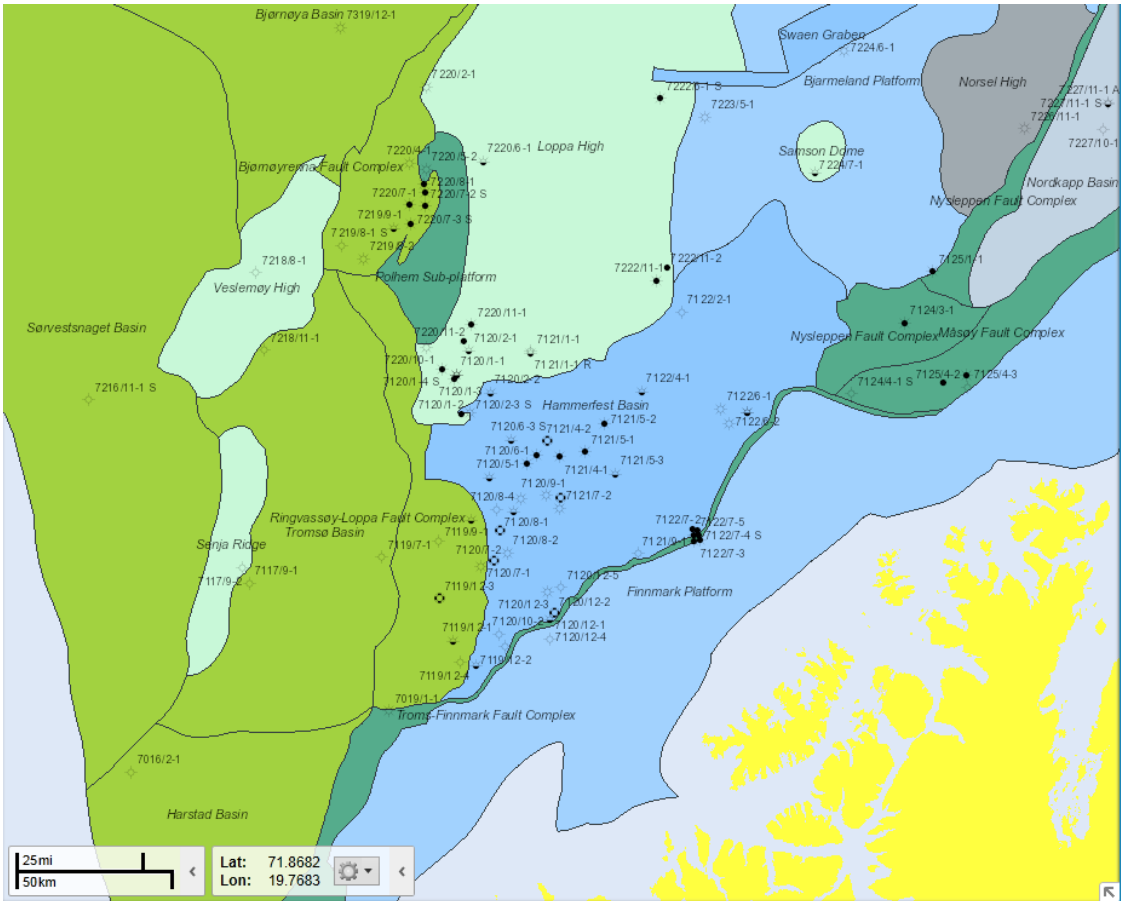


Figure 3.1: Overview of drilled wells in the southwestern Barents Sea (NPD, 2015a).

The quality of the seismic data is very good, and except for some gas chimneys, mainly along the Ringvassøy-Loppa FC, there were few imaging problems. Salt diapirs along the axis of the southern Tromsø Basin made the interpretation a little more complicated, but this was outside the main area of interest for this study. The density of the seismic 2D lines is relatively high, typically 3-10 km spacing, in both the Hammerfest and Tromsø basins, but the density is slightly higher in the northern Tromsø Basin than in the Hammerfest Basin and the southern Tromsø Basin (figure 3.2).

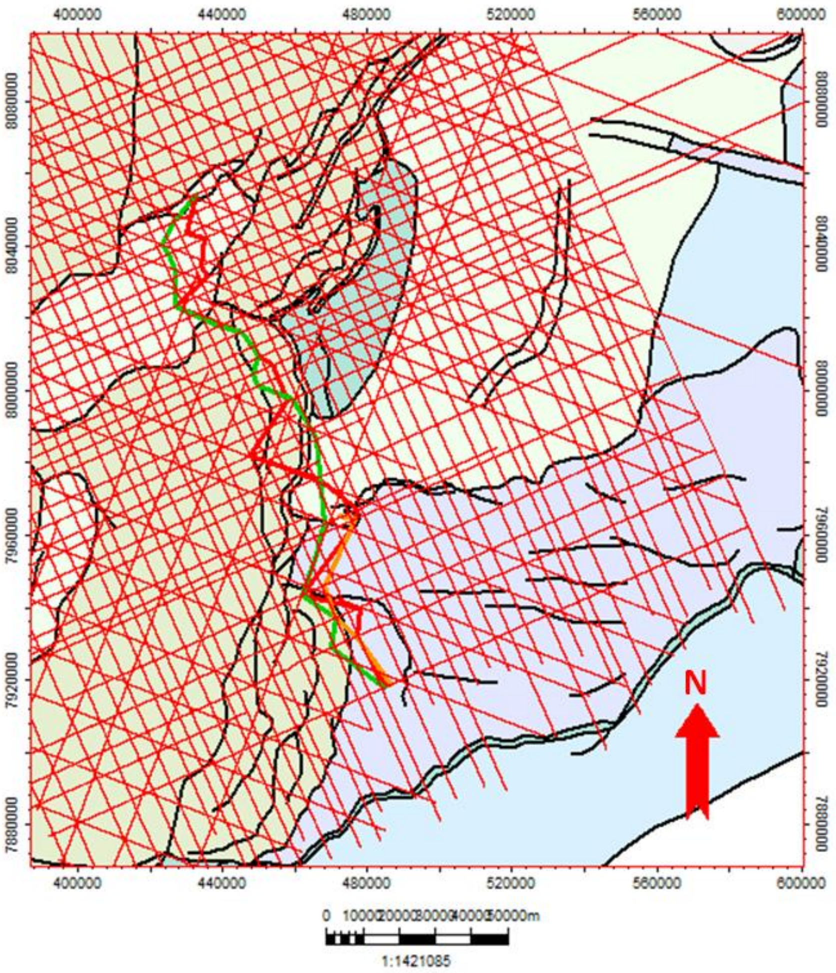


Figure 3.2: Available seismic lines covering the study area.
Map modified from NPD (2015a).

The Petrel software is used for interpretation of the 2D seismic. The Base Tertiary and URU, bounding the Paleogene succession, were interpreted on all seismic lines. The units and subunits of the Paleocene succession were interpreted on all line they were present on. In the result chapter some representative lines are used to display the results from the interpretation.

3.2 Methods

The roots of sequence stratigraphy can be traced back to the 18th century, and the history and development of sequence stratigraphy are described by Nystuen (1998). Most of the terms used in sequence stratigraphy today were introduced by Mitchum et al. (1977a), and after this the concepts and tools have been continuously refined.

Previous work on Triassic prograding systems in the Barents Sea by Glørstad-Clark et al. (2010; 2011a) gives a good overview of the applications of sequence stratigraphy. And the techniques from these studies can be applied to the study of the prograding systems from Paleocene.

The seismic sequence stratigraphy interpretation is divided into two stages, firstly seismic sequence analysis are preformed to identify stratigraphic packages and secondly seismic facies analysis of the packages are preformed (Mitchum et al., 1977b).

A stratigraphic package in the seismic data is recognized by the reflection terminations at the base and top of the package, also called baselaps and toplaps. A reflection termination can be truncated, which means it is an erosional surface, or it can be a lapout which means it has reached its depositional limit (Bertram and Milton, 1996). Different reflection terminations are illustrated in figure 3.3, and the following descriptions of the most common reflection terminations are based on Mitchum et al. (1977a) and Bertram and Milton (1996):

Downlap is when the seismic reflector lapout on an underlying seismic reflector since it has a higher dip than the underlying seismic surface. This is common for the base of prograding clinoforms and represents the transition from slope depositions to condensation or non-deposition. A downlap surface will normally represent a maximum flooding surface.

Onlap is another reflection termination against an underlying unit. In this case the lapout is a result of higher dip on the underlying seismic surface compared to the seismic reflector. Onlap can be marine or costal, where the marine onlap represent a marine hiatus or condensed interval and the costal onlap represent a change from a deposition zone to basin-margin erosion and non-deposition. Basin inversion can make an original downlap looking like an onlap; hence, the timing of deposition versus timing of basin tectonics is important to rule out.

Toplap is when the seismic reflector lapout against an overlying seismic surface with lower angle, and is an evidence of a nondepositional hiatus. Toplaps are normally associated with shallow marine deposits, but can occur in deep-marine settings.

Erosional truncation is when a seismic reflection terminates against an overlying erosional surface, and is evidence for an erosional hiatus. The erosional surface can be both marine and non-marine and represents a sequence boundary.

Structural truncation is when the seismic reflector terminates against a surface made by structural disruption, and this type of reflection does not represent a sequence boundary. Examples are faulting, gravity slides, salt flows and igneous intrusions.

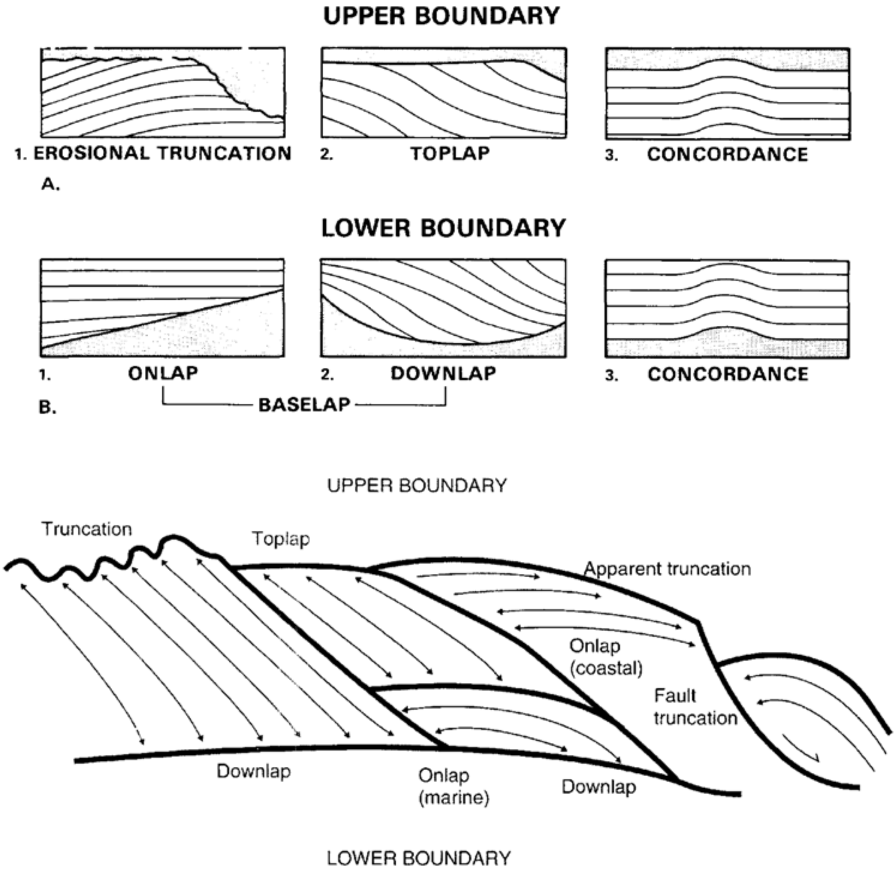


Figure 3.3: overview of different reflection terminations.
(Mitchum et al., 1977a; Bertram and Milton 1996)

During seismic facies analysis the reflection configuration, amplitude, continuity and external form within the seismic sequences are interpreted to say something about the environmental setting during deposition (Mitchum et al., 1977b).

The continuity of the reflectors is associated with continuity of the strata, and continuous reflections indicate widespread, uniformly stratified deposits (Mitchum et al., 1977b). The reflection amplitude depends on the velocity and density contrast on an interface (Mitchum et al., 1977b).

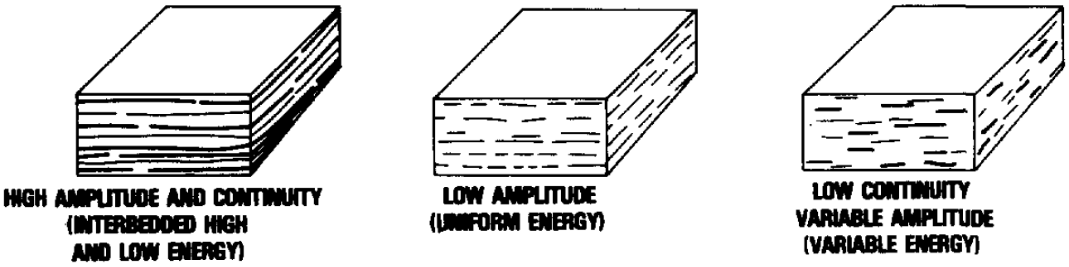


Figure 3.4: Variations of reflector amplitude and continuity. (Sangree and Widmier, 1978).

Clinoforms are the surfaces bounding clinothems and they can be formed along the shoreline by deltas or at the shelf-edge. The two types of clinoforms have different scale (figure 3.5) where, the shoreline clinoforms can be 5-10's of meters high while the shelf edge clinoforms can be 100's of meter high (Helland-Hansen and Hampson, 2009). The clinoforms observed in this study is in the scale of shelf-edge clinoforms.

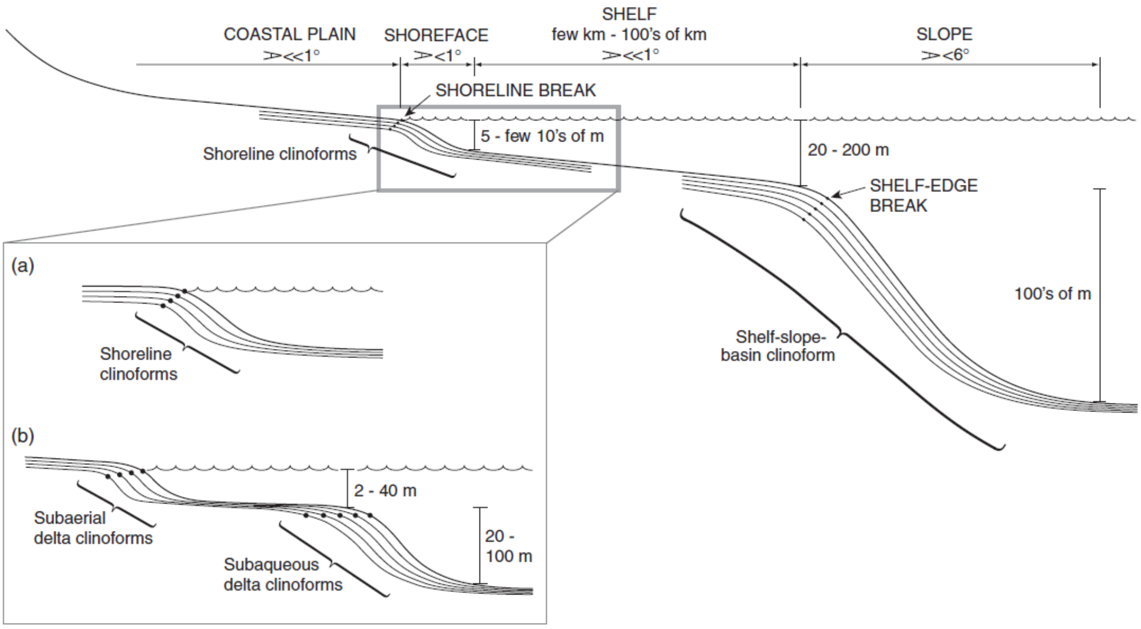


Figure 3.5: Scale of shorelines and shelf-edge clinoforms. (Helland-Hansen and Hampson, 2009).

A clinoforms consists of topset, slope and bottomset (figure 3.6). The topset normally appear flat on seismic data and the shoreline can be located anywhere along the topset (Myers and Milton, 1996). The slope is the more steeply dipping part of the clinoform developed basinward of the topset (Myers and Milton, 1996). The transition from topset to slope is marked by an increased angle of the depositions called offlap-break. The bottomset is the base of the clinoforms and are normally characterized by deep-water depositional systems (Myers and Milton, 1996).

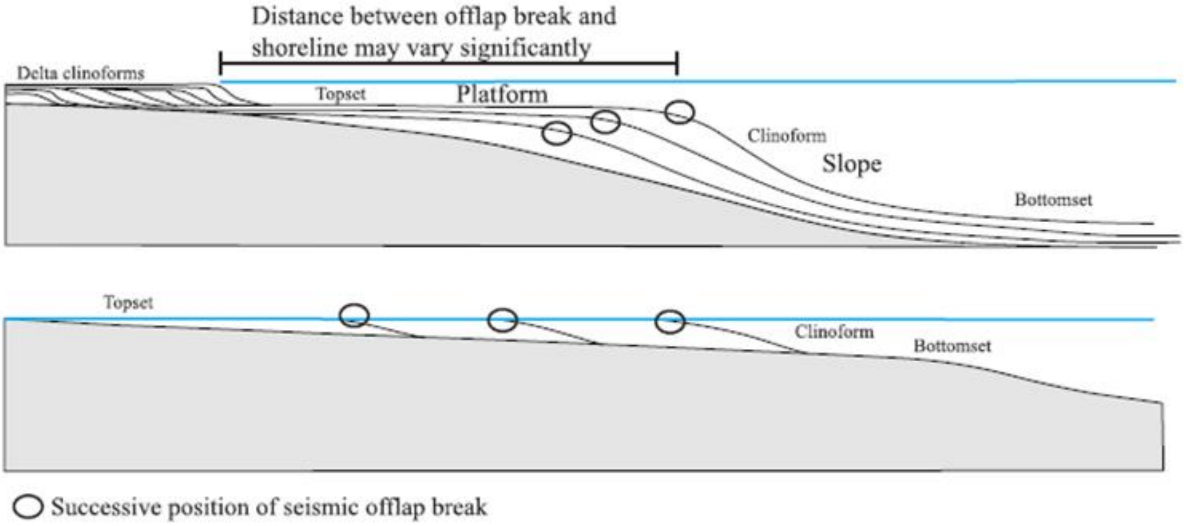


Figure 3.6: Simplified clinoforms geometry, sigmoidal clinoforms at upper profile and oblique clinoforms al lower profile (Glørstad-Clark et al., 2010).

There are two main types of clinoforms, sigmoidal and oblique, and the typical pattern of the two types of clinoforms can be seen in figures 3.6 and 3.7. Variations in depositional settings are decisive for what type of clinoform pattern is formed. Both types of clinoforms are observed in this study.

The sigmoidal progradational reflection pattern has an S-shape and is deposited in a low energy environment. The topset and bottomset are typically thin while the slope depositions are thicker (figure 3.7), the depositional angle of the slope are usually lower than 1° (Mitchum et al., 1977b). The preservation of the topset implies that sea level was rising and/or the basin was subsiding during deposition (Mitchum et al., 1977b). The sigmoidal reflections typically have high continuity and moderate to high amplitude (Sangree and Widmier, 1978).

The oblique progradational reflection pattern is typically a steep dipping slope which toplap against a nearly flat seismic reflector and downlap against an underlying unit (figure 3.7). The depositional angle of the slope is higher than for the sigmoidal pattern and can be up to 10°, and the depositional environment is typically high energy (Sangree and Widmier, 1978). The depositional conditions imply high sedimentation supply, slow/no basin subsidence and relatively stable sea level (Mitchum et al., 1977b). The reflection amplitude and continuity varies, but generally there is a decrease from upper to lover clinoform (Sangree and Widmier, 1978).

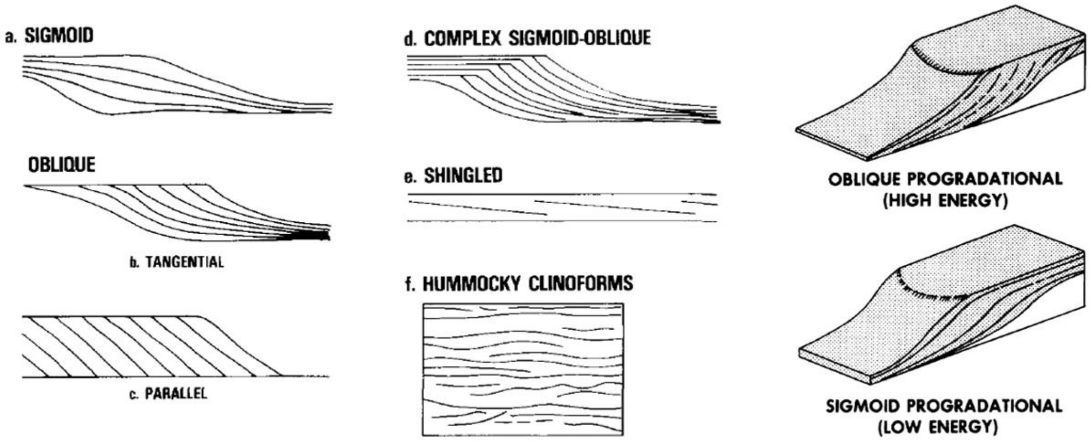


Figure 3.7: Oblique and sigmoidal clinoform pattern. (Mitchum et al., 1977b; Sangree and Widmier, 1978)

Offlap-breaks mark a change in depositional processes and products (Helland-Hansen and Hampson, 2009), in this study the offlap-breaks marks the platform-edge of a prograding system. Normally the slope and bottomset are mud-prone whereas the topset has more coarse depositions like sand (Johannessen and Steel, 2005). The width of the shelf and the proximity of the shoreline are crucial for the ability to transport sand to the shelf edge and the deepwater (Johannessen and Steel, 2005).

The platform-edge can easily be mapped out as the offlap-break, and these offlap-breaks will give information of the position and movement of the shelf-edge through time. The shelf-edge trajectory is a long-term response to changes in relative sea level and sediment supply (Helland-Hansen and Hampson, 2009). Previous study of trajectory analysis of Early Triassic prograding clinoforms in the northern Barents Sea (Anell et al., 2014) gives a good overview of the applications of trajectory analysis.

Shelf edge trajectories can be divided into three categories; ascending, flat and descending (figure 3.8). Ascending shelf-edge trajectories indicates that there was a long-term rise in the relative sea level, and the sigmoidal seismic pattern is found in this setting (Helland-Hansen and Hampson, 2009). Flat and descending shelf-edge trajectories means that there is less storage potential on the shelf and the seismic pattern in this setting is oblique (Helland-Hansen and Hampson, 2009).

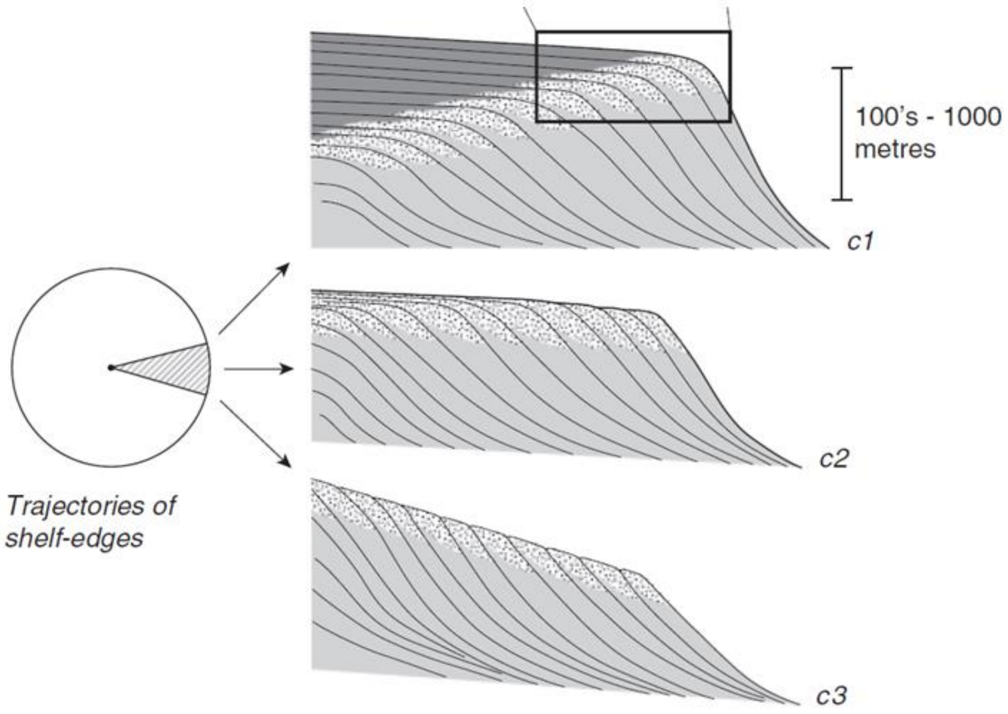


Figure 3.8: Ascending (c1), flat (c2) and descending (c3) shelf edge trajectory patterns. (Helland-Hansen and Hampson 2009)

Calculations of slope angles, thicknesses and height of clinoforms are based on an average velocity of 2100 m/s. The average velocity is based on available sonic logs from NPD (2015b) in the area. It is important to know that the velocity is only an average, and lateral and vertical variations in the velocity are not taken into account. This means that the calculations only are approximations of the actual values.

4 Results

4.1 Base Tertiary, Early Eocene and Upper Regional Unconformity

The Paleocene succession is bounded by three surfaces (Figure 4.1); the Base Tertiary (BT) at the base and the Early Eocene and the Upper Regional Unconformity (URU) at the top. The URU is a result of the uplift and erosion in late Cenozoic time and can be traced over the entire study area.

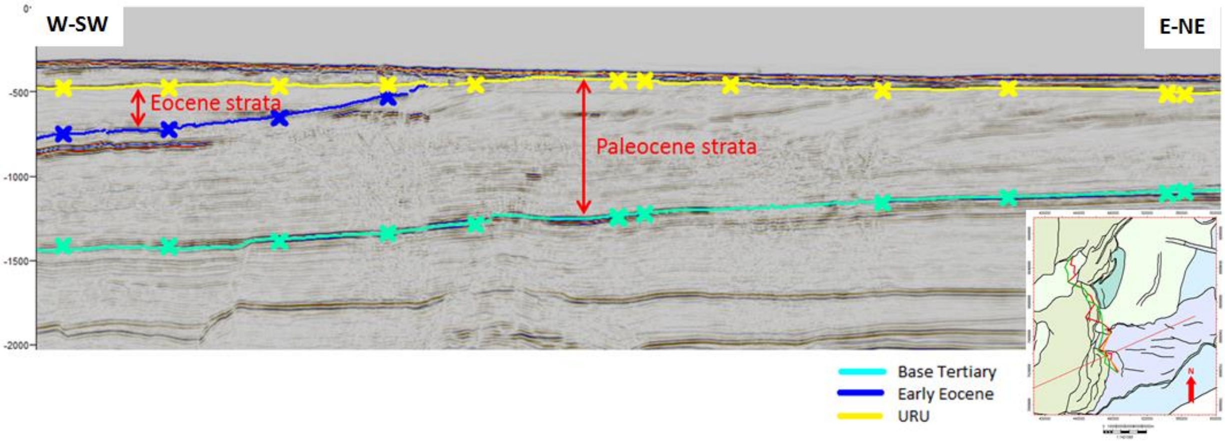


Figure 4.1: The Paleocene succession in the Hammerfest Basin and its three bounding surfaces. Depth in TWT, map modified from NPD (2015a).

The Base Tertiary reflector is a strong reflector representing a hiatus and unconformity in the Cretaceous-Paleogene transition. The surface can be traced in both the Hammerfest and Tromsø basins, but is eroded by the URU over the Finnmark Platform and most of the Loppa High. The depth to the Base Tertiary varies a lot but is in general much shallower in the Hammerfest Basin than in the Tromsø Basin. In the Hammerfest Basin, the depth varies from ca. 800 ms TWT in the east to ca. 1500 ms TWT in the west (figure 4.2). On the southwestern Loppa High, the Base Tertiary is present at shallow depths of ca. 600 ms TWT. The Paleocene succession in the Tromsø Basin is located at greater depth than the succession in the Hammerfest Basin, and the Base Tertiary are therefore found at greater depth here. In the central part of the Tromsø Basin the Base Tertiary can be found at depths up to 2500 ms TWT; in the north of the basin the reflector is found at a shallower level, typically 1500 ms TWT.

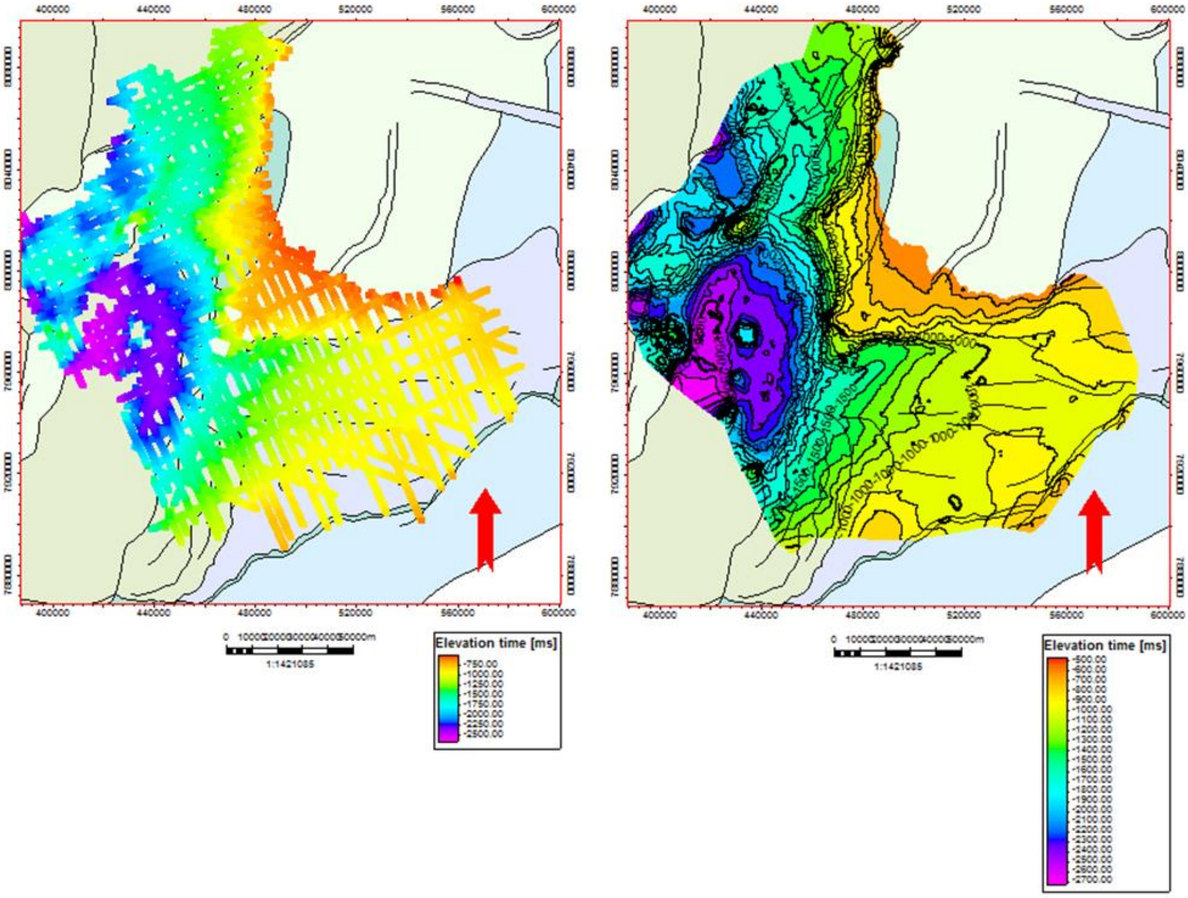


Figure 4.2: The seismic interpreted Base Tertiary to the left, and the Base Tertiary surface to the right. Map **modified from NPD (2015a).**

The Early Eocene surface is mainly found in the southern part of the Tromsø Basin, and is eroded by the URU in the transition between the Tromsø and Hammerfest basins and in the northern part of the Tromsø Basin. The surface is recognized by baselaps onto the surface from the west. These baselaps are best observed in the Hammerfest –Tromsø Basin transition and to the west in the Tromsø Basin. The average depth to the Early Eocene reflector is ca. 1000 ms TWT, but in the central part of the southern Tromsø Basin it has been mapped down to 2000 ms TWT.

The Upper Regional Unconformity (URU) is recognized by the erosional truncation of the underlying units. Cenozoic rocks are truncated in the Hammerfest and Tromsø basins, while older rocks are truncated on the Finnmark Platform and over most of the Loppa High. The URU is found at shallow depths, ca 540 ms TWT, and is overlain by sediments of glacial origin.

4.2 The Paleocene package

The Paleocene package is bounded at the top by the Early Eocene mainly in the southern Tromsø Basin, and is otherwise truncated by the URU. Most of the Paleocene is eroded close to the Loppa High and in the larger part of the Hammerfest Basin, and only the lowermost part of the succession is preserved.

Gas chimneys disturb the seismic reflections (figure 4.3) and this can be a problem during the interpretation. The Base Tertiary, Early Eocene and URU are easy to recognize across gas chimneys, and it is mainly for the interpretation of the progradational units of the Paleocene succession the gas chimneys has caused problems. Gas chimneys are mainly found along the Ringvassøy-Loppa FC, and it is few gas chimneys otherwise in the Tromsø and Hammerfest basins.

There are some faults cutting the Paleocene succession but none of them with large throw (figure 4.3), up to 200 ms TWT. Most of the observed faults are located along the Ringvassøy-Loppa FC, mainly in the area around the Polheim Sub-platform. There are also some faults along the Asterias FC, between the Hammerfest Basin and Loppa High.

In the upper part of the Paleocene package, close to the Early Eocene surface, there is a strong seismic reflector cutting the seismic stratification (figure 4.3). The reflector is strongest in the northern part of the Tromsø Basin, but is also clear in the southern part of the Tromsø Basin. The reflector is correlated to represent the Opal A to Opal CT transition (Riis and Fjeldskaar, 1992). The diagnetic horizon is shallowing to the east due to late Cenozoic uplift and erosion, and the reflector is truncated by the URU in the transition between the Tromsø and Hammerfest basins. By cutting the seismic stratification the reflector may appear as a downlap surface. But the surface is not a part of the depositional pattern, but a result of the burial history of the area. These apparent, but not real, downlap surfaces are observed in both the Tromsø and Hammerfest basins, and both within unit P3 and P4.

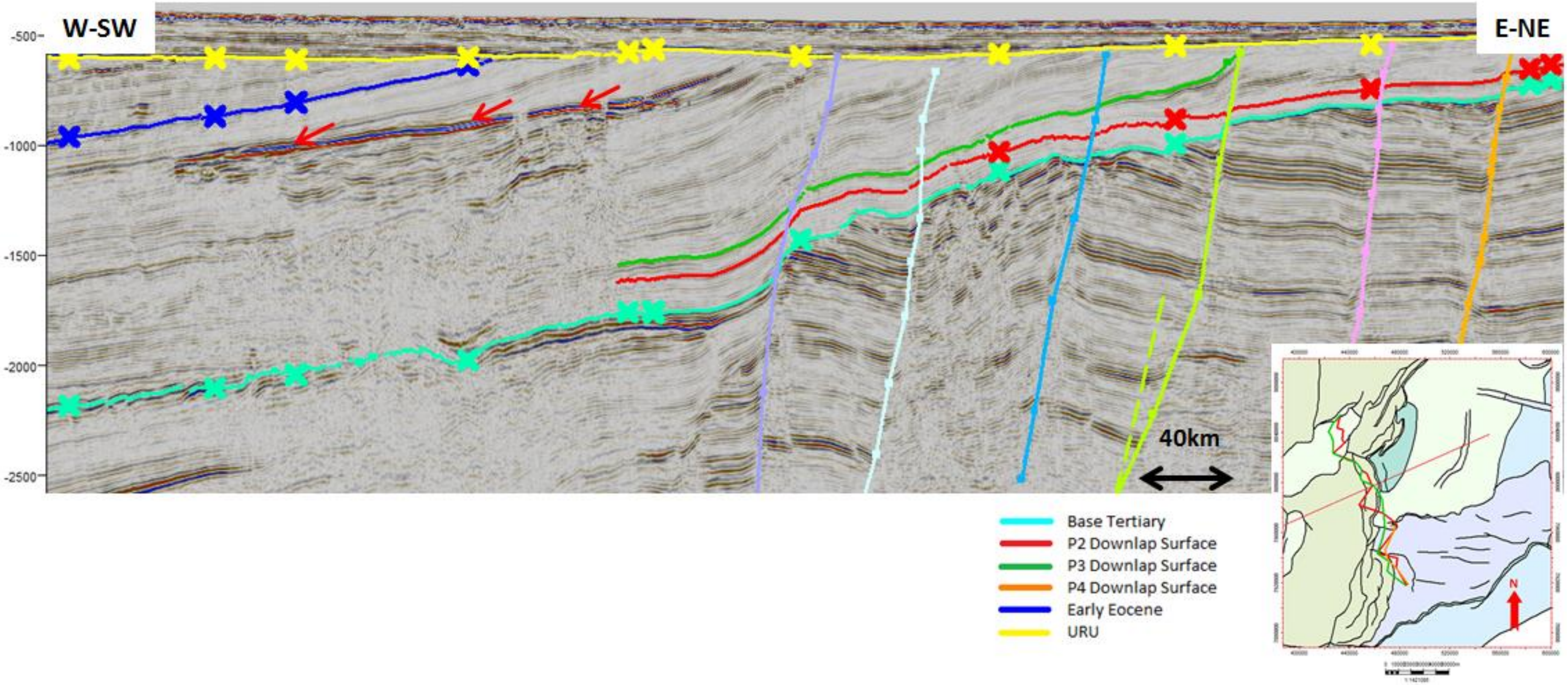


Figure 4.3: Seismic section with interpreted faults and units from the Tromsø Basin located to the WSW, over Polheim Sub-platform towards Loppa High to the ENE. Gas hinders further interpretation of the P2 and P3 downlap surfaces to the WSW. Red arrows mark the apparent downlaps towards the reflector representing the Opal A to Opal CT transition. Depth in TWT, map modified from NPD (2015a).

The Paleocene package shows two different depositional patterns, aggradational and progradational stacking. The aggradational pattern is mainly found in the west of the study area, and in the southern central Tromsø Basin the whole Paleocene package shows an aggradational pattern. To the east, a progradational system is divided into three main units based on downlap surfaces of large areal extent (figure 4.5). Figure 4.4 shows how far west each unit of the depositional systems prograded, and this is also the border between the thick aggradational succession to the west and the progradational units to the east. The progradational units are deposited over a thinner aggradational unit which was deposited as a sheet over large parts of the western Barents Sea (figure 4.5). Two of the prograding units, P2 and P3, can be found both in the Hammerfest and Tromsø basins while the last one, P4, only is found in the Hammerfest Basin. All of the three prograding units have approximately the same western progradational limit (figure 4.4).

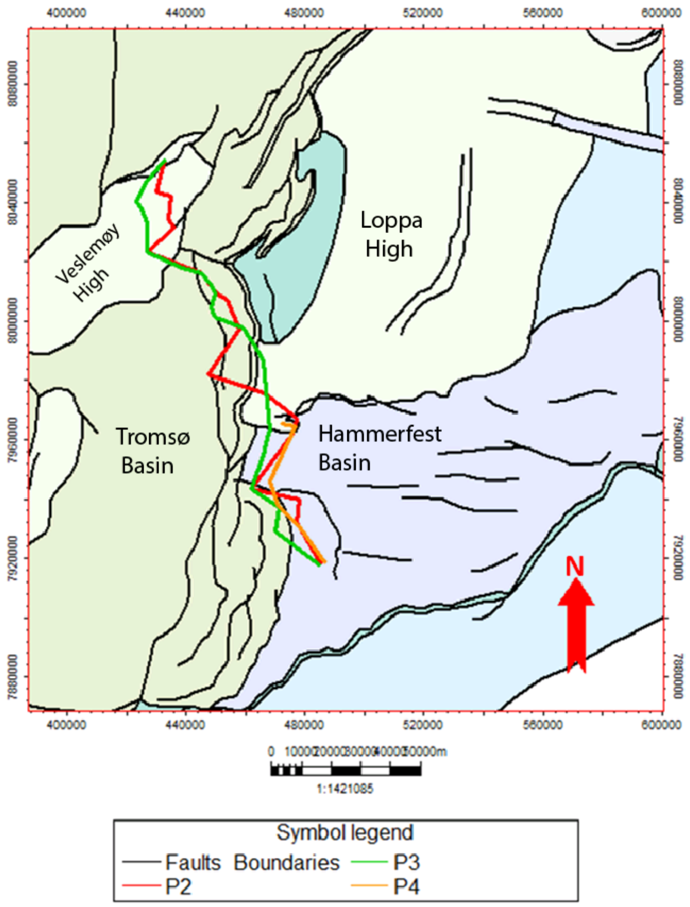


Figure 4.4: Prograditional limit for the three prograding units, the direction of progression was from the east towards west. Map modified from NPD (2015a).

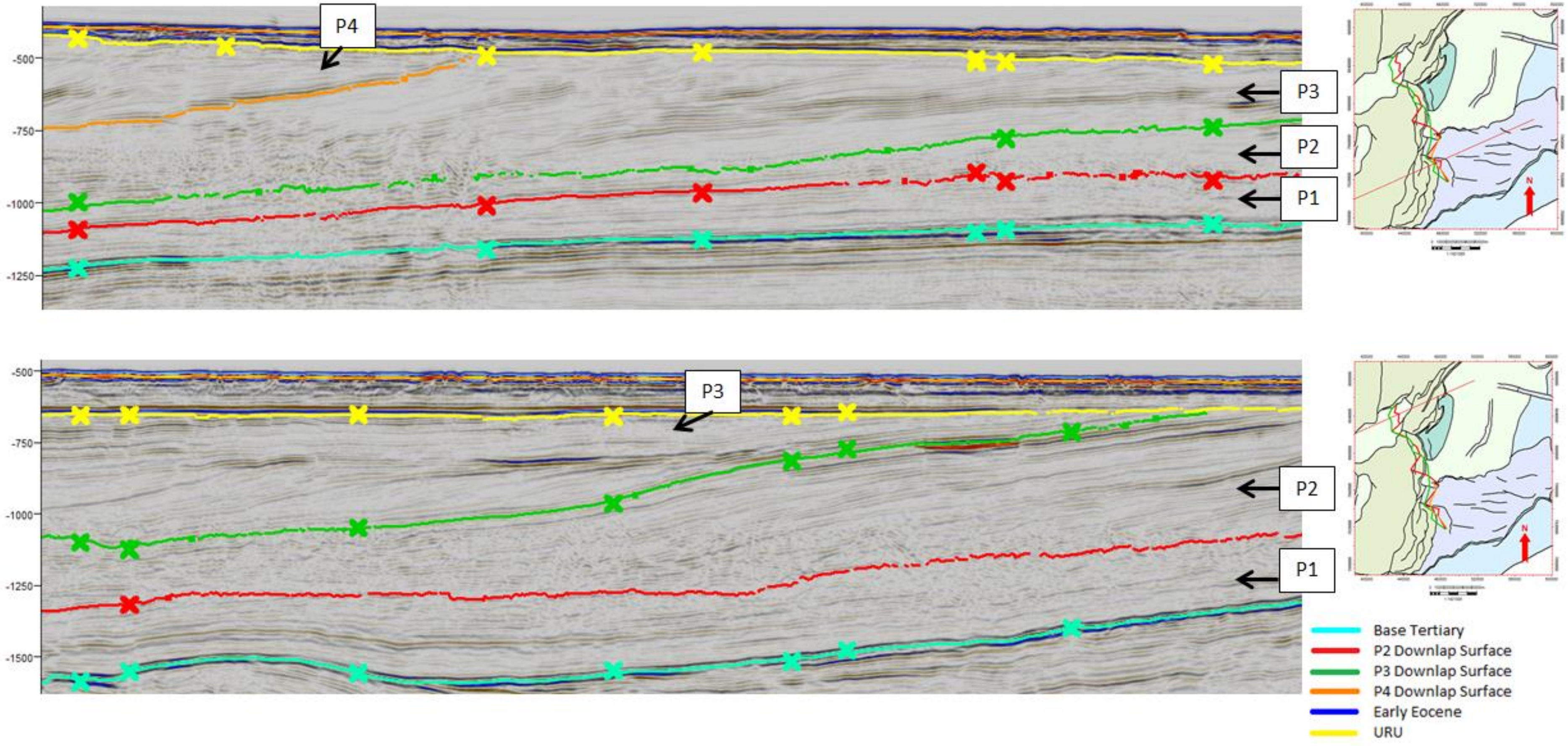


Figure 4.5: The interpreted units in the Hammerfest Basin (top) and in the northern Tromsø Basin (bottom). Depth in TWT, map modified from NPD (2015a).

The Paleocene package is eroded over large areas, and the time-thickness map in figure 4.6 shows which areas that are influenced by erosion and which areas that has the original, albeit compacted, thickness preserved. It is mainly the aggradational system under the Eocene package in the southern Tromsø Basin which has the original thickness preserved. The thickness of the progradational units increases westward both in the Hammerfest and Tromsø basins. The thickness is greatest in the northern Tromsø Basin, whereas in the southern Tromsø Basin the Paleocene succession is thinner than in the westernmost part of the Hammerfest Basin. As the time thickness map in figure 4.6 shows, the Paleocene package thins out as a result of erosion towards Loppa High and the erosional pinchout is marked in the map. Because of limited data eastwards, the erosional pinchout are not mapped out in the eastern part of the Hammerfest Basin.

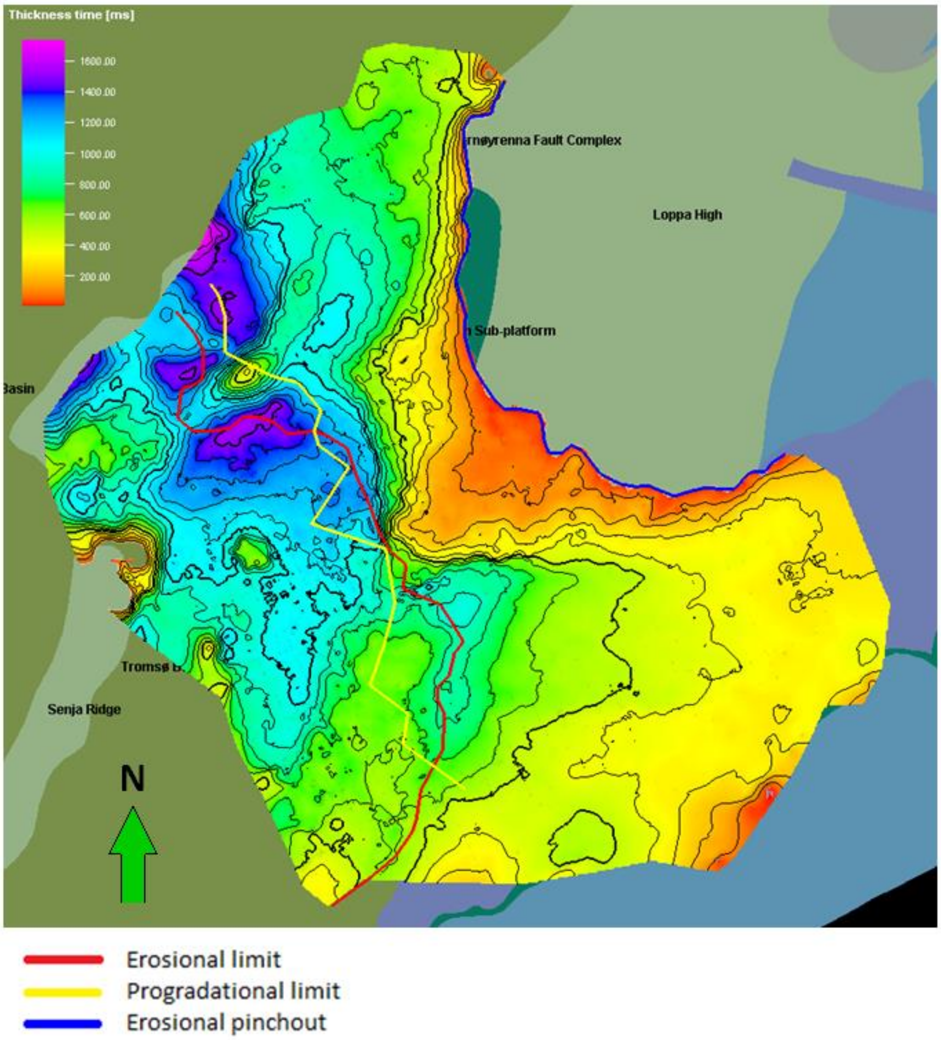


Figure 4.6: Time-thickness map for the Paleocene package. The thickness west of the red line has the original thickness preserved while the thickness east of the red line has been reduced by erosion. Map modified from NPD (2015a).

In places on the southwestern Loppa High, mainly in depressions, parts of the Cretaceous succession are found under the Paleocene package (figure 4.7). It is also important to notice that the Paleocene package does not onlap towards Loppa High, but are deposited as a draping cover over the high. The Paleocene package is truncated at the top by the URU both over and along the southern and western flanks of the Loppa High. This shows that the thinning of the Paleocene package in the Loppa High area is a result of erosion and not a depositional feature.

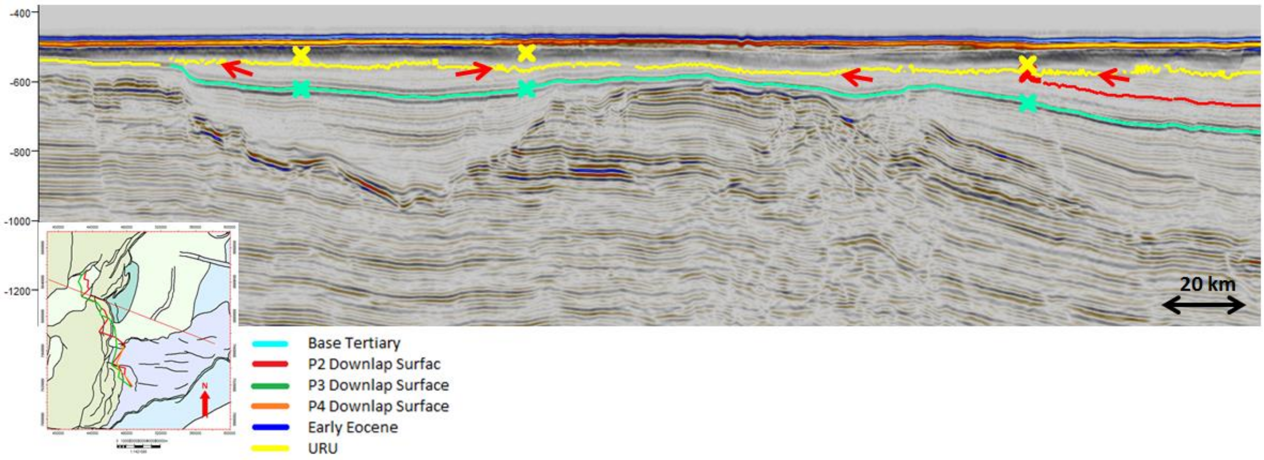


Figure 4.7: The Paleocene package up towards the southern flank of Loppa High. The infill of the depression on Loppa High is of Cretaceous origin. Depth in TWT, map modified from NPD (2015a).

4.3 Aggradational units

Two Paleocene aggradational units are observed in the area; one is located under the prograding units and is named P1 (figure 4.5), and the other is located in the Tromsø Basin west of the limit for the prograding systems.

P1 is bounded at the base by the Base Tertiary surface and at the top by the P2 unit and the URU. The unit is found in the Hammerfest Basin, parts of the Tromsø Basin and as a thin sheet over the southwestern Loppa High. The unit has a uniform thickness of ca. 130 m, but is a little thinner over the southwestern Loppa High as a result of erosion. The internal reflections are weak but are a little stronger to the west in both the Hammerfest and Tromsø basins. The internal reflections have an overall good continuity but in the eastern Hammerfest Basin the reflections are some more discontinuous. These first parallel-bedded reflections of the Paleocene succession were also observed in previous studies of the early Cenozoic in the Hammerfest and Tromsø basins (Knutsen and Vorren, 1991; Knutsen et al., 1992).

West of the progradational limit the four units cannot be distinguished, and the entire Paleocene succession shows an aggradational stacking pattern. The thickness of the Paleocene succession in the southern Tromsø Basin is ca. 1000 m, whereas in the northern part of the basin the thickness is up to 1500 m with a slight thinning westward over the Veslemøy High (figure 4.6). The internal reflections are clear and continuous. Salt diapirs through the Paleocene succession along the axis of the southern Tromsø Basin were also observed. These structures only influenced the thick aggradational deposits in the southern central Tromsø Basin, which were not the primary target in this study.

4.4 Unit P2

P2 is the lowermost of the three progradational units, it downlaps onto P1 and is bounded at the top by P3 and the URU. As seen in figure 4.8, the unit can be found in the whole Hammerfest Basin, the northern Tromsø Basin and on small parts of the southwestern Loppa High. The unit prograded from the east towards the west.

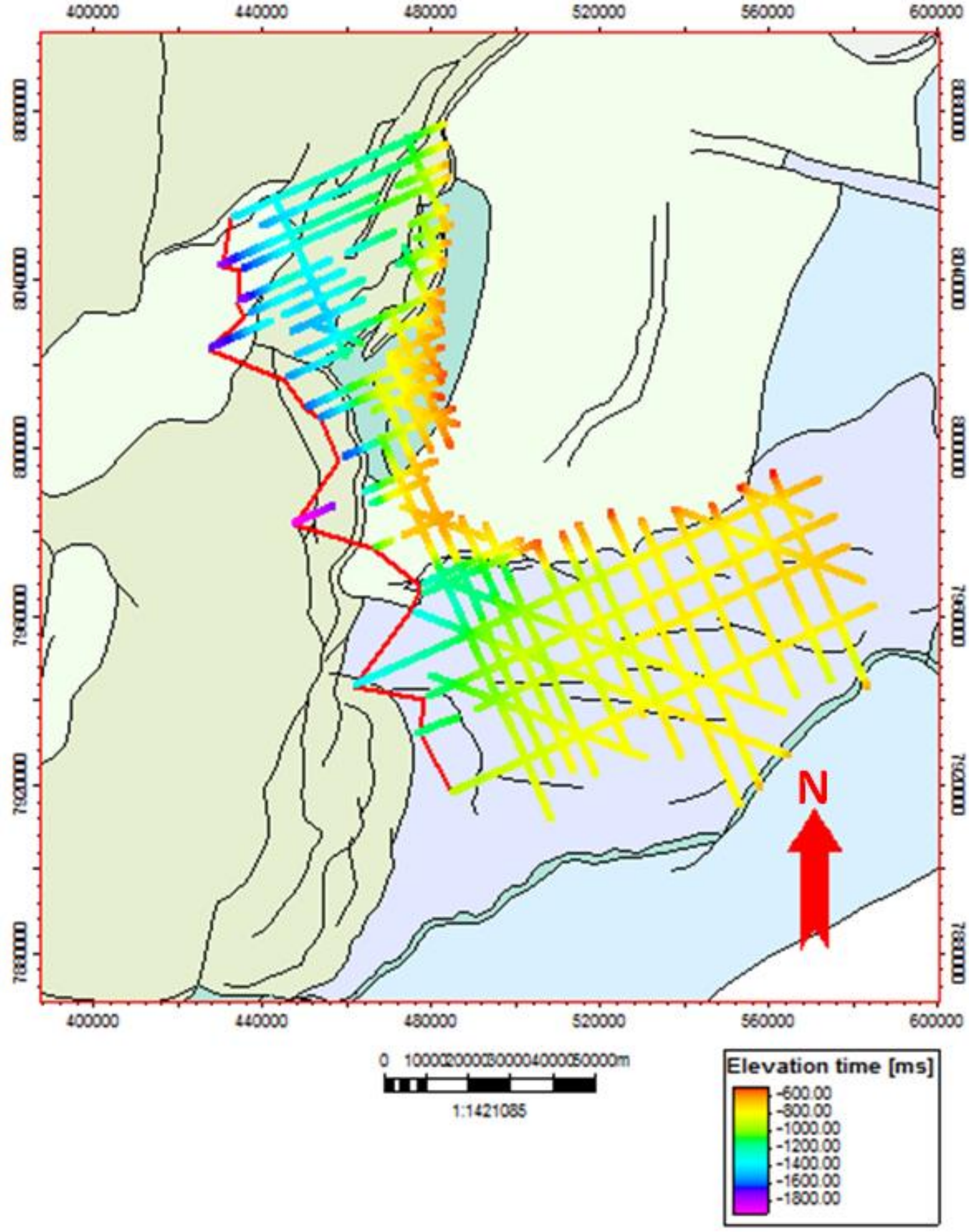


Figure 4.8: Areal extent of unit P2. Map modified from NPD (2015a).

The upper part of the unit is truncated by the URU in the eastern part of the Hammerfest Basin and in the eastern Tromsø Basin towards Loppa High (figure 4.9). This implies that the original thickness is only preserved in the western Hammerfest Basin and in the northern Tromsø Basin. The original depositional thickness has otherwise been affected by late Cenozoic erosion. The average thickness of the unit is 250 m, and the thickest part is found in the northern part of the Tromsø Basin (figure 4.9). The thickness decreases westward as an effect of the clinoform geometry, where the bottomset is thinner than the topset and slope deposition.

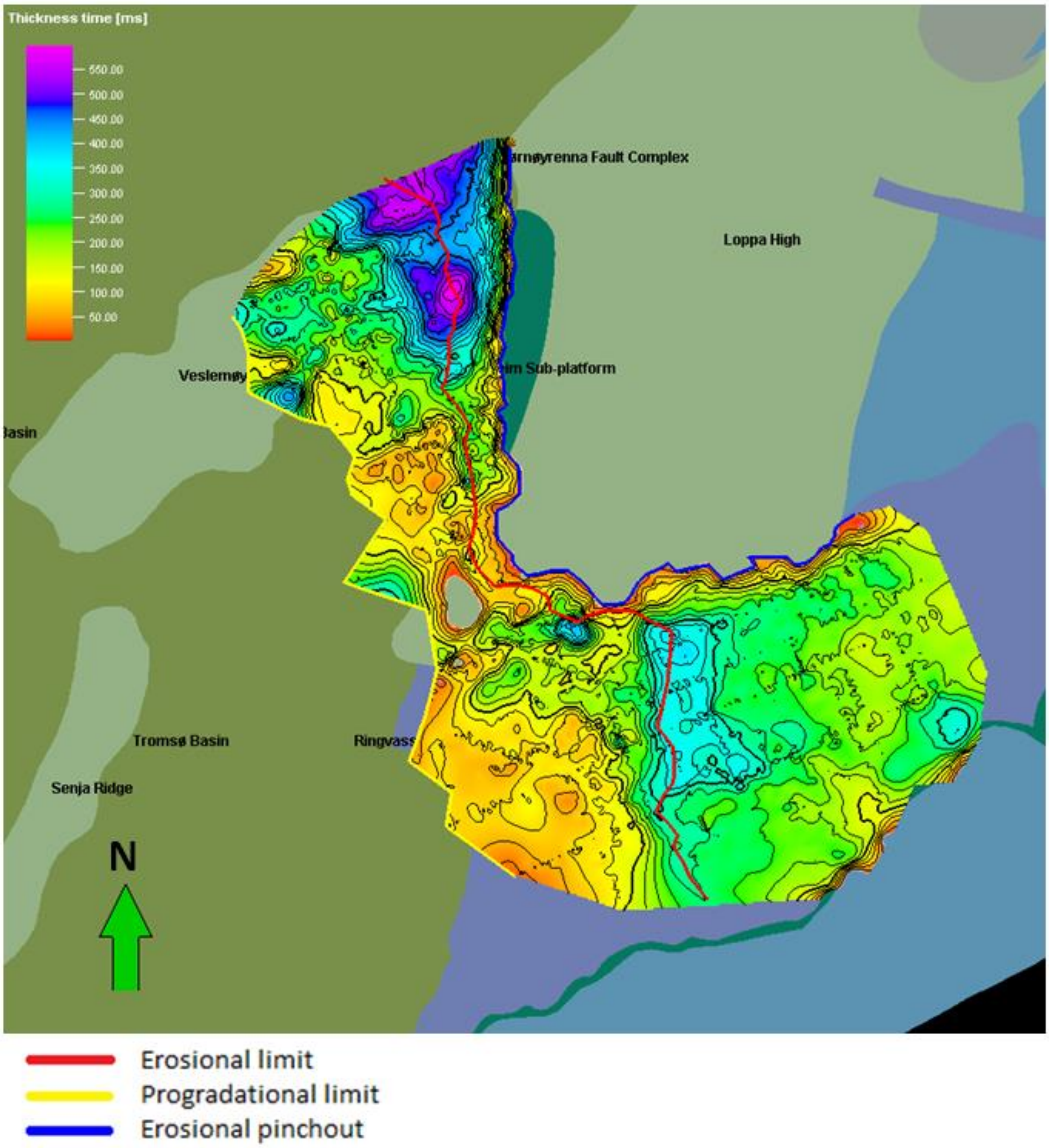


Figure 4.9: Time-thickness map for unit P2. P2 west of the red line has the original thickness preserved while the thickness east of the red line has been reduced by erosion. Map modified from NPD (2015a).

In the northern Tromsø Basin, where the preserved part of unit P2 is thicker, some seismic sections show whole clinoform profiles, whereas the erosion has removed the upper part of the clinoforms in the southern Tromsø Basin and in the Hammerfest Basin (figure 4.10). A more detailed analysis of the unit can be performed in the northern Tromsø Basin. Based on the internal reflection pattern the unit can be divided into three subunits of lower order (figure 4.11).

The lowermost subunit, P2_1, downlaps onto the top of unit P1. The subunit has a thin preserved topset, and the clinoform pattern is a mix between sigmoidal and oblique (figure 4.11). The calculated angle of the slope depositions are approximately 2,6° and the offlap-break trajectory is relatively flat (figure 4.11). The height of the clinoforms is measured to be ca. 200 m.

The second subunit, P2_2, downlaps onto the topset and slope of the subunit P2_1, and the platform-edge of the subunit did not prograde as far west as the platform-edge for the two other subunits. The subunit has a better preserved topset and a more pronounced sigmoidal reflection pattern than the first subunit (figure 4.11). The angle of the slope varies from 2,1°-2,6°, with the steepest slopes to the west. The offlap-break trajectory is ascending (figure 4.11) and the height of the clinoforms is ca. 100 m.

The last subunit, P2_3, prograded further to the west than the two first subunits. The lower part of the subunit onlaps onto the slope of subunit P2_2, whereas the upper part of the subunit is deposited over subunit P2_2. The subunit downlaps onto the top of unit P1. The unit has a clear developed topset and the reflection pattern is sigmoidal (figure 4.11). The angle of the slope deposits are approximately 2,1°. The offlap-break trajectory is ascending (figure 4.11), and the height of the clinoforms is typically 200-250 m.

In the Hammerfest Basin only the lower part of the unit is preserved (figure 4.10). The clinoform pattern cannot be destinguished in this area based on the limited part of the preserved clinoforms. The angles of the preserved part of the slope are lower than 2°, which are lower than the angles in the northern Tromsø Basin. This is probably because the angles in the northern Tromsø Basin are calculated where the slope is steepest, whereas in the Hammerfest Basin only the lower part of the slope deposits are preserved and the angle here is lower.

The amplitude and continuity of the internal reflections in all three subunits varies, but the general pattern is the same. The reflections of the topset and slope deposits have good continuity, while the bottomset has a discontinuous and chaotic pattern (figure 4.10). The amplitude of the reflections changes from weak at the lower part of the subunits to strong higher up. The amplitudes are generally strongest along the slope deposits. This pattern is clearest in the Tromsø Basin, but the same pattern is observed in the Hammerfest Basin (figure 4.10). In the western Hammerfest Basin, both the lower part of the slope and bottomset deposits are preserved and the change from continuous clear to weak and chaotic reflections is observed. In the eastern Hammerfest Basin only the bottomset of the unit is preserved and the reflections are weak and discontinuous.

In the transverse sections, NNW – SSE oriented profiles, in the Hammerfest Basin the unit changes from a chaotic pattern in the lower part to clearer and more continuous reflections higher up. In upper part with more continuous reflections, lenses which thin out towards north and south are observed (figure 4.12). The lenses are up to 20 km long and the thickness is typically up to 70 m. The lenses shape is not a result of erosion, and this means that they are depositional features. Since it is not possible to decide which subunits are present in the Hammerfest Basin, it is not possible to determine which of the subunits the lenses are part of. There are not found any lenses in the Tromsø Basin.

Interpretation:

The subunits of P2 represent sequences of lower order and they can provide information on how the sea level and depositional environment changed during deposition. The variations of the offlap-break trajectories for the three subunits and the decreasing angle of the slope from subunit P2_1 to P2_3 indicates that the depositional environment changed from high energy in the earliest depositions of the unit to lower energy at later stages. The chaotic reflections in the bottomset can be a result of gravity flows or similar depositional mechanisms. The topset and slope deposits have more continuous reflections which suggest that the deposition was continuous and widespread.

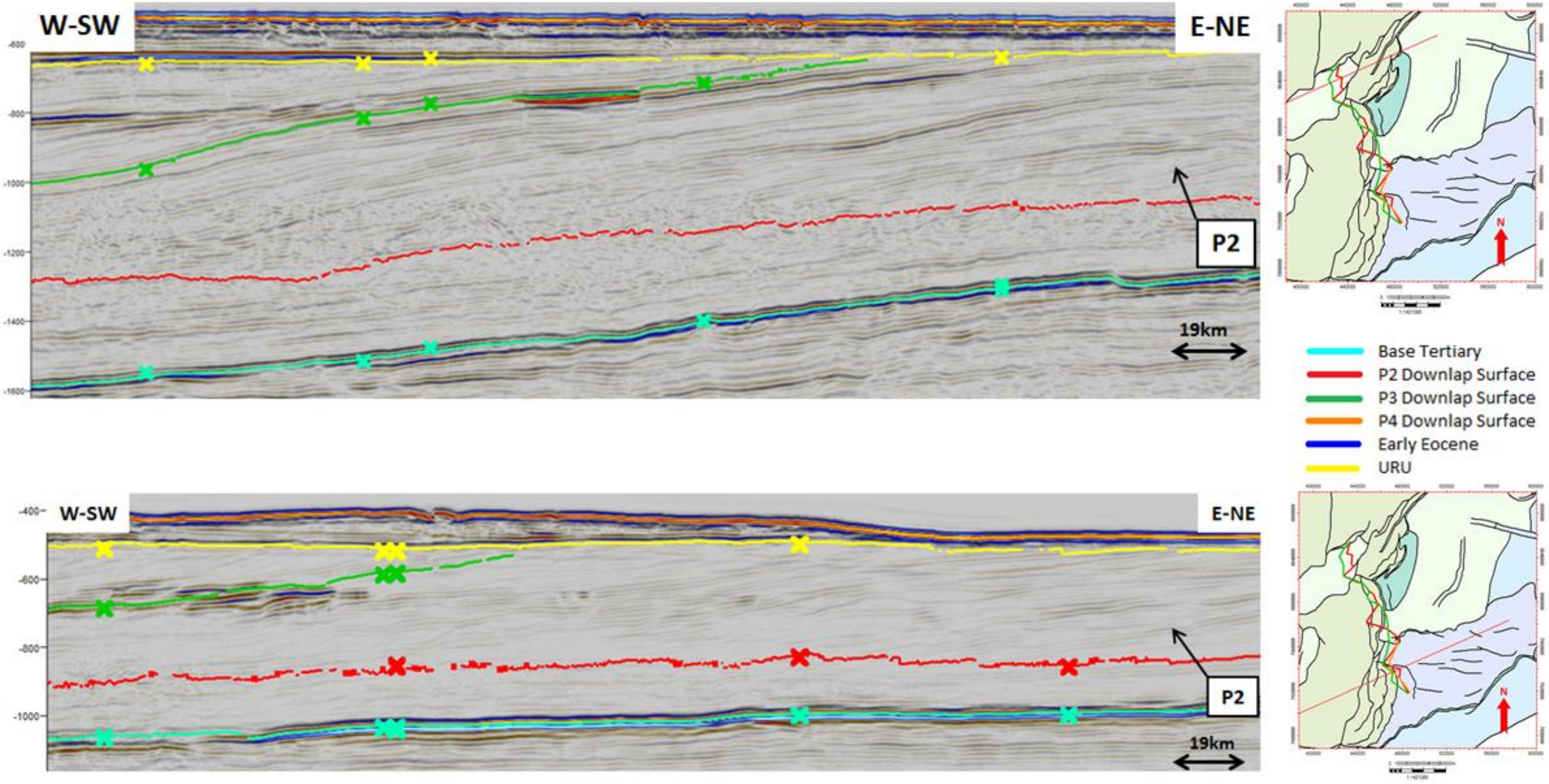


Figure 4.10: Unit P2 in the northern Tromsø Basin (top) and in the Hammerfest Basin (Bottom). Depth in TWT, map modified from NPD (2015a).

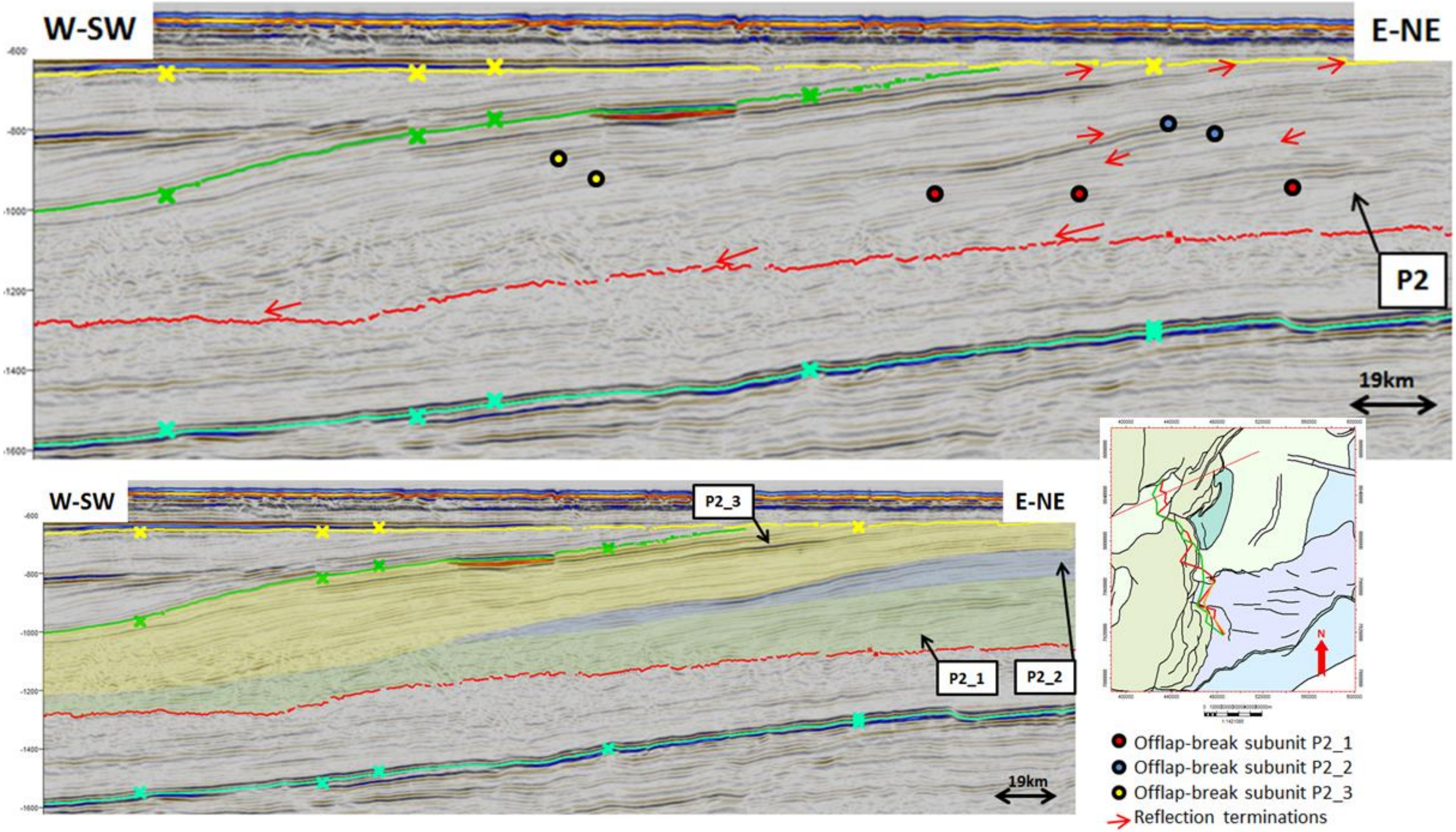


Figure 4.11: Upper seismic section show reflection terminations and offlap-breaks for unit P2 and lower figure highlights the subunits. Seismic legend same as in figure 4.10. Depth in TWT, map modified from NPD (2015a).

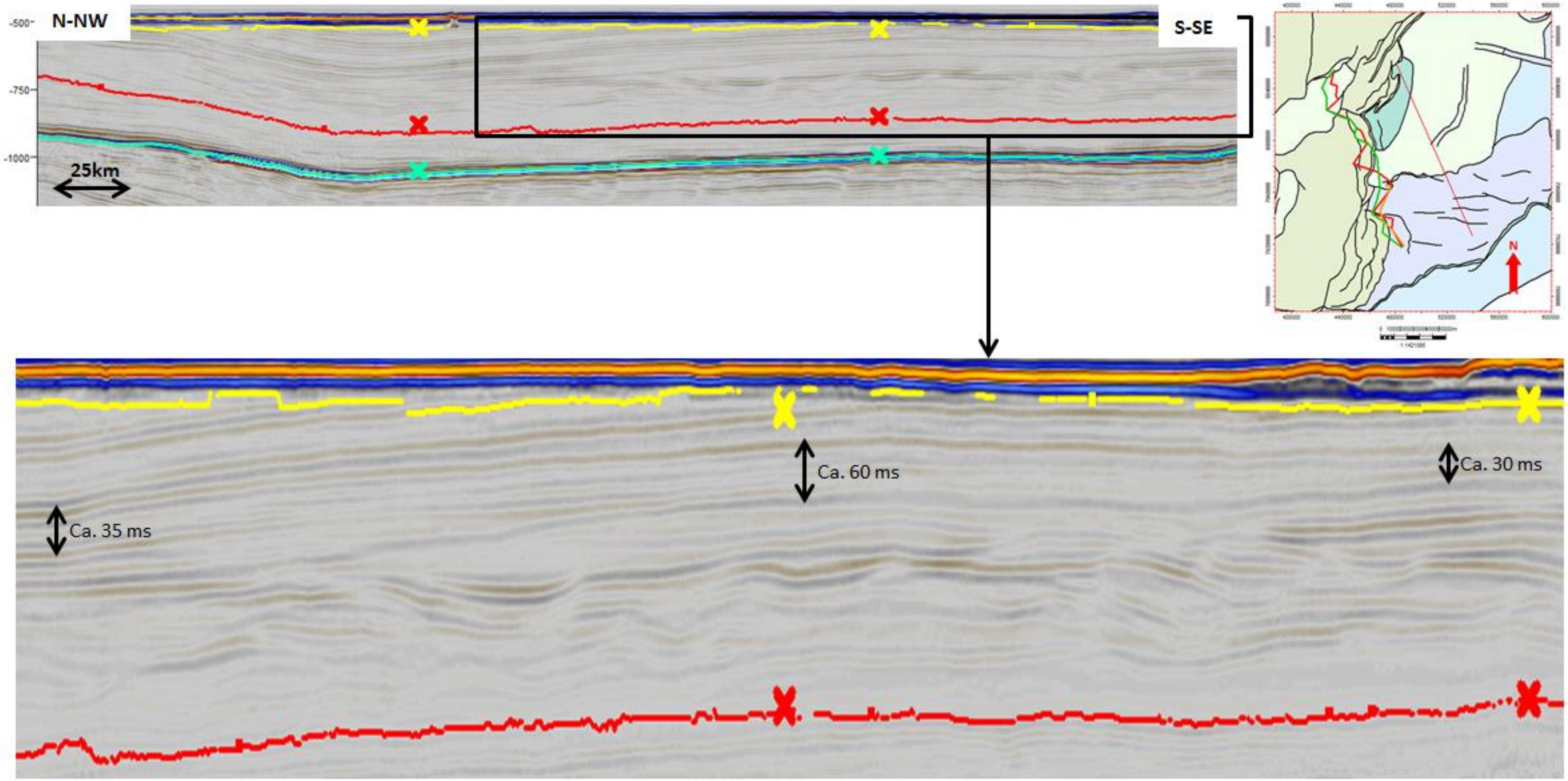


Figure 4.12: Transverse section of unit P2 in the Hammerfest Basin. Vertical scale of lower seismic section is stretched to highlight one of the lens shapes found in the area. Seismic legend same as in figure 4.10. Depth in TWT, map modified from NPD (2015a).

4.5 Unit P3

P3 downlaps onto P2 and is bounded at the top by P4 (only in the Hammerfest Basin) and the URU. The unit has approximately the same areal extent as P2 in the northern Tromsø Basin, but does not reach as far east towards the Loppa High. In the Hammerfest Basin the unit does not reach as far east as P2, and the areal extent of P3 can be seen in figure 4.13. The unit prograded towards west.

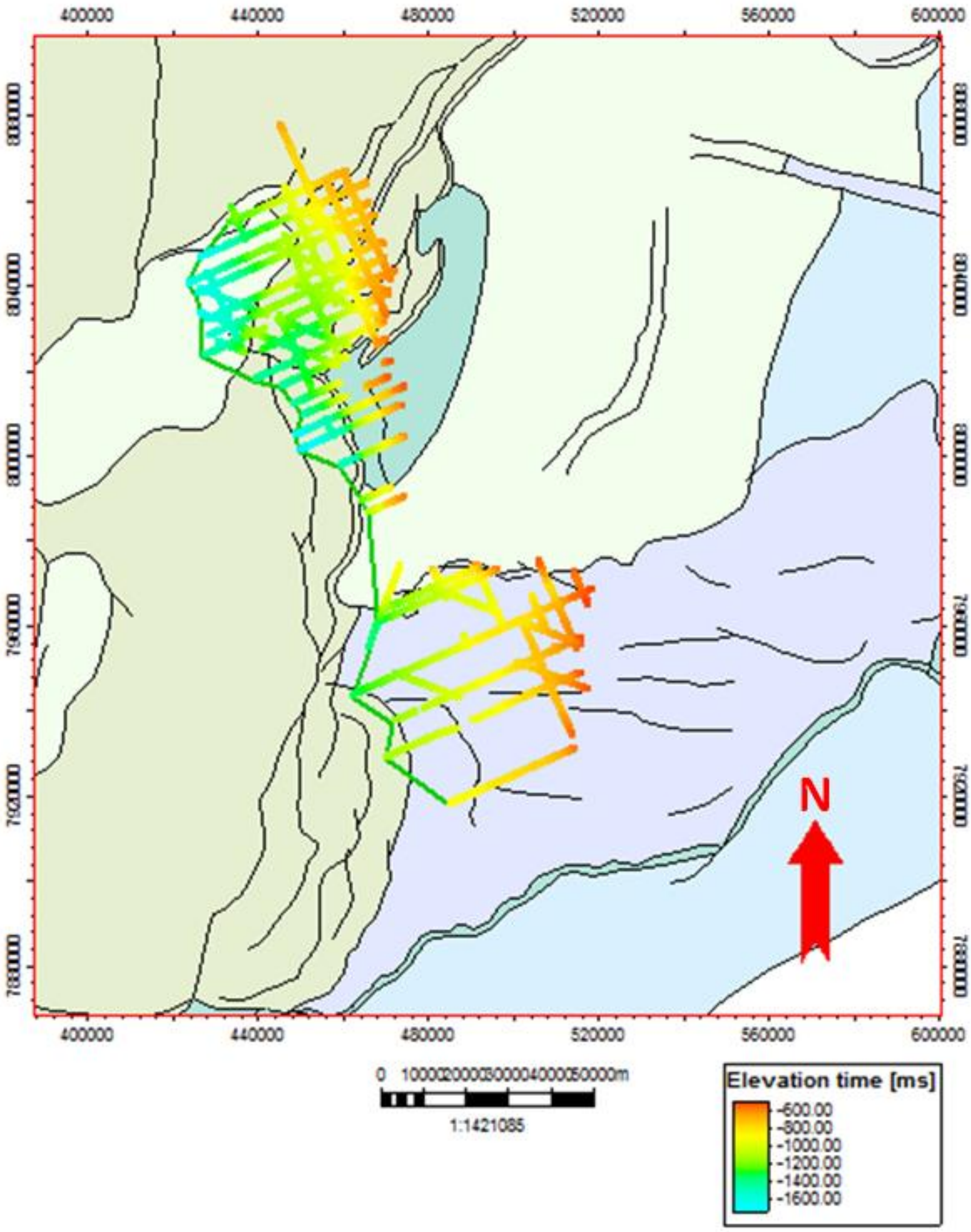


Figure 4.13: Areal extent of unit P3. Map modified from NPD (2015a).

The average thickness for the unit is ca. 300 m, and the thickest part of the unit is found in the northern Tromsø Basin (figure 4.14). The upper part of the unit is eroded in most of the area and the original thickness is only preserved under P4 in small parts of the western Hammerfest Basin. In the Hammerfest Basin the thickness decreases westward where thin bottomsets are found, whereas some more of the thicker slope depositions are preserved further east. In the northern part of the study area erosion has influenced the thickness, and the thickness increases towards the west.

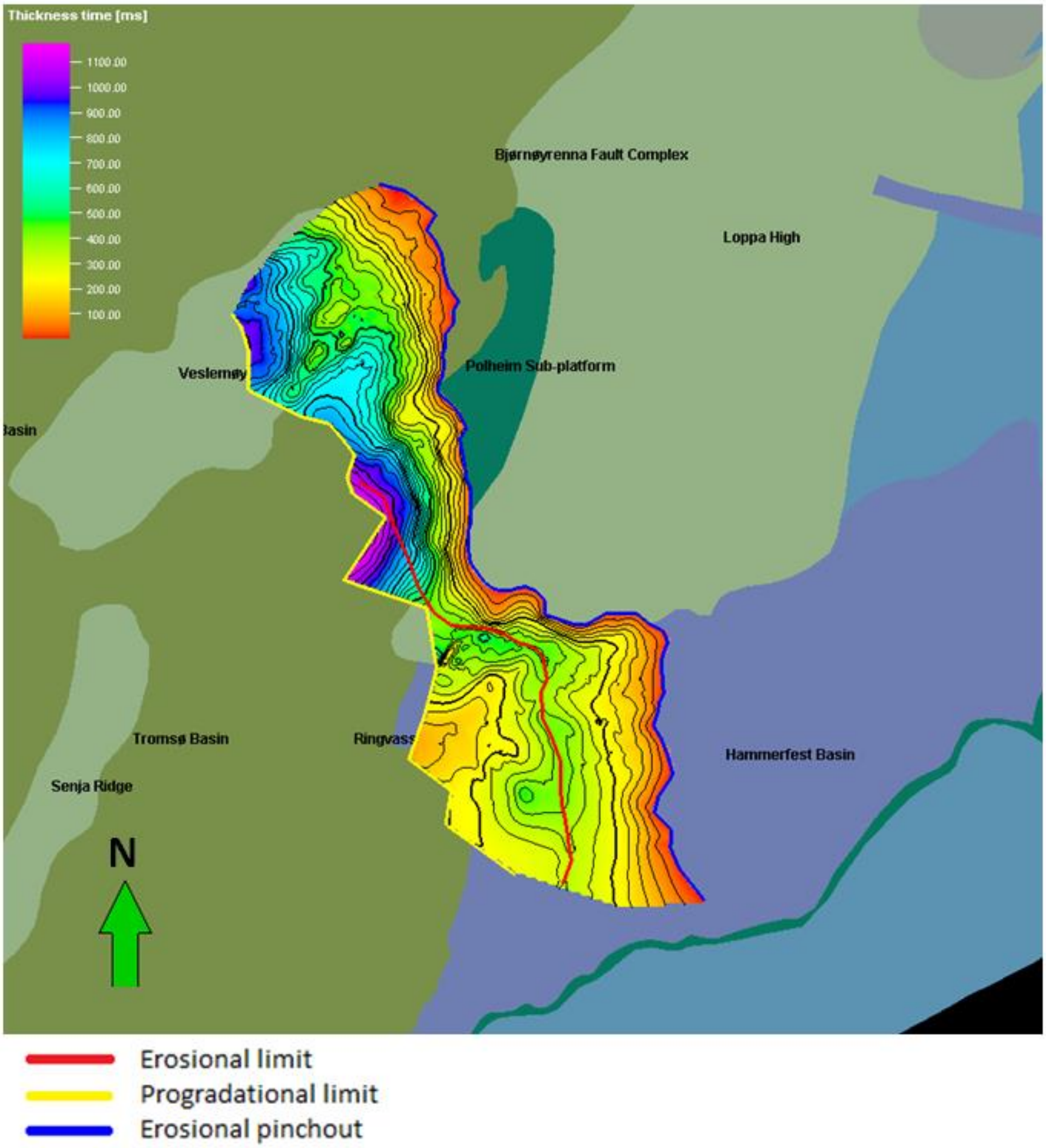


Figure 4.14: Time-thickness map for unit P3. P3 west of the red line has the original thickness preserved, whereas the thickness east of the red line has been reduced by erosion. Map modified from NPD (2015a).

Full clinoform profiles are found on several seismic lines in the northern Tromsø Basin and in one section in the Hammerfest Basin. Erosion has removed the upper part of the clinoforms elsewhere in the Tromsø and Hammerfest basins. A more detailed analysis of the unit can be performed where clinoform profiles are preserved. Since clinoform profiles are preserved in both the Tromsø and Hammerfest basins, correlation of the unit can be done between the basins. The P3 unit can be subdivided into two subunits of lower order based on the reflection pattern in the northern Tromsø Basin (figure 4.17).

The first subunit, P3_1, downlaps onto the top of unit P2, and whole successions of clinoforms from this unit can be found in both the Hammerfest Basin and the northern Tromsø Basin (figure 4.15). In the northern Tromsø Basin the clinoforms of the subunit are found on several seismic sections. Clinoforms of the subunit are only found on one seismic section in the northwestern Hammerfest Basin. The subunit is found on the easternmost preserved part of the P3 unit in the northwestern Hammerfest Basin, and the subunit is located further east in the Hammerfest Basin than in the northern Tromsø Basin. The geometrical style, slope angle and offlap-break trajectory of the clinoforms are equal in the Hammerfest and Tromsø basins (figure 4.15). The main difference is that more of the subunit is preserved in the northern Tromsø Basin than in the Hammerfest Basin. The subunit has a thin topset and the clinoform patterns are oblique (figure 4.17). The angle of the slopes are approximately 2,6° in both the Hammerfest and Tromsø basins, which are the same as for the oblique clinoforms in subunit P2_1. The offlap-break trajectory is flat (figure 4.17), and also this corresponds well with what observed for subunit P2_1. The height of the clinoforms increases westwards and goes from ca 200 m to 250 m. Along some of the NNW – SSW oriented seismic profiles in the northern Tromsø Basin the subunit thins out towards north and south over large distances. This gives the subunit a lens shape on the sections transverse of the prograding direction, the thinning northward for one of the lenses are highlighted in figure 4.18. The best preserved lens is almost complete, only partly eroded at the edges, and are at least 40 km long. Parts of other lenses are found thinning towards north or south, but because of erosion the size of them are reduced. The lenses are clearly depositional features, and each lens covers a large areal extent. These lenses are probably the same depositional structure as observed in unit P2, but the lenses of this subunit are larger.

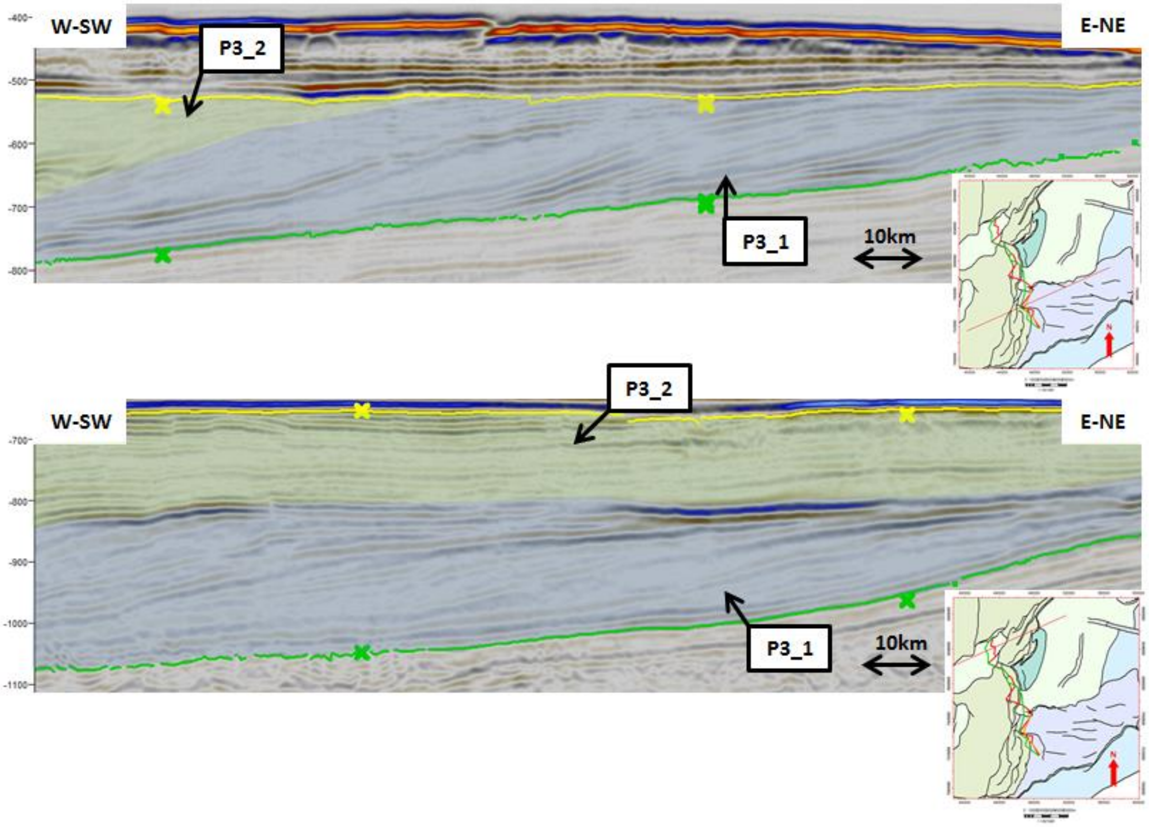


Figure 4.15: The oblique clinoforms from subunit P3_1 in the Hammerfest Basin (top) and in the northern Tromsø Basin (bottom). Depth in TWT, map modified from NPD (2015a).

The second subunit, P3_2, downlaps towards unit P2, as subunit P3_1 also did. The subunit has a clear developed topset, which are deposited over subunit P3_1. The clinoform pattern is sigmoidal and the angle of the slope is ca. 2,1°-2,5°. The offlap-break trajectory is ascending (figure 4.17), and the height of the clinoforms is typically 300 m.

Only the lower part of the unit is preserved In the Hammerfest Basin (figure 4.16), except for a small preserved part of subunit P3_1 in the northern part of the basin. The clinoform pattern is not possible to decide for the rest of the unit based on the small part of the clinoforms preserved. The angle of the preserved slope deposition increases from ca. 2° in the east to ca. 3° in the west. The eastern part with lower angles probably corresponds to subunit P3_2. The higher angles to the west could mean that there is another subunit over P3_2, but it is not possible to decide this based on the available data. Based on the thickness of unit P2 in the Hammerfest Basin compared to the height of the clinoforms in subunit P3_2 in the northern Tromsø Basin, it should be possible to find clinoform profiles from subunit P3_2 or possible younger subunits in the Hammerfest Basin. This is not found, and only the slope and bottomset depositions are observed in the area.

The continuity and amplitude of the reflections varies within the two subunits. The pattern is much the same as for the three subunits of unit P2. The reflections are weak in the lower part of each subunit but increase in amplitude higher up in the subunits. The amplitude is highest along the topset of subunit P3_1, whereas for subunit P3_2 the amplitude is strongest along the topset – slope transition (figure 4.16). The continuity is good along the topset and slope for both subunits, but the bottomset is characterized by chaotic pattern and poor continuity (figure 4.16). These amplitude and continuity patterns are best observed in the northern Tromsø Basin, but the variation from low to high amplitudes are also observed in parts of the Hammerfest Basin. Otherwise in the Hammerfest Basin, only the lower parts of the clinoforms are preserved and the reflections are generally weak and discontinuous.

Interpretation:

The subunits of P3 represent sequences of lower order, and can provide information on how the sea level and depositional environment changed during the deposition. From a relatively high energy depositional environment during deposition of subunit P3_1 the energy decreased before deposition of subunit P3_2. This change is represented by the change from oblique clinoforms in subunit P3_1 to sigmoidal clinoforms in subunit P3_2. The good reflection continuity in the slope and topset deposits indicates widespread and continuous deposition in these areas. The poor continuity and chaotic pattern in the bottomset deposits can be a result of gravity flows or less continuous deposition. The fact that subunit P3_1 is found both in the Hammerfest Basin and the northern Tromsø Basin indicates that the depositional environment has been the same in both basins.

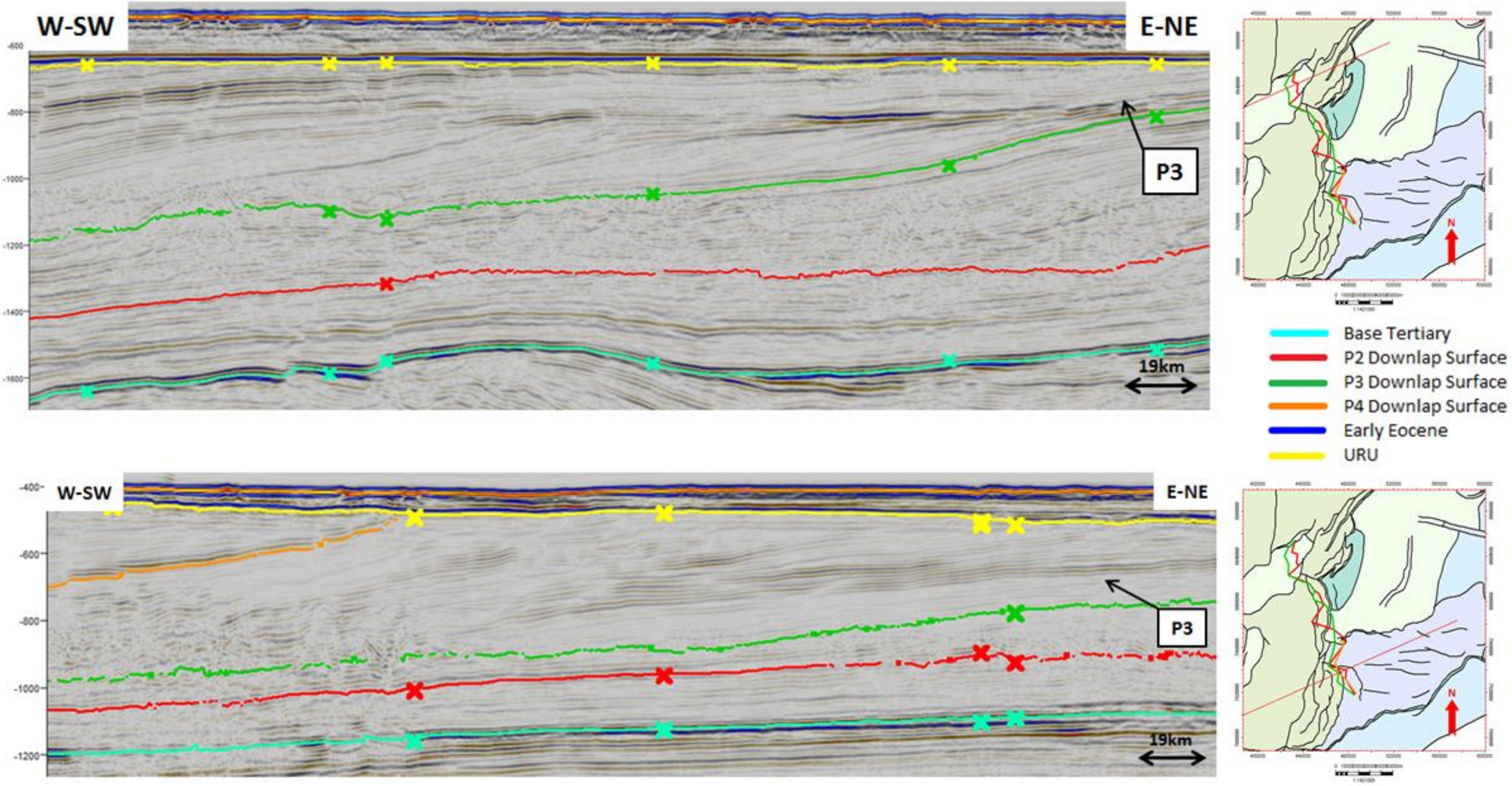


Figure 4.16: Unit P3 in the northern Tromsø Basin (top) and in the Hammerfest Basin (bottom). Depth in TWT, map modified from NPD (2015a).

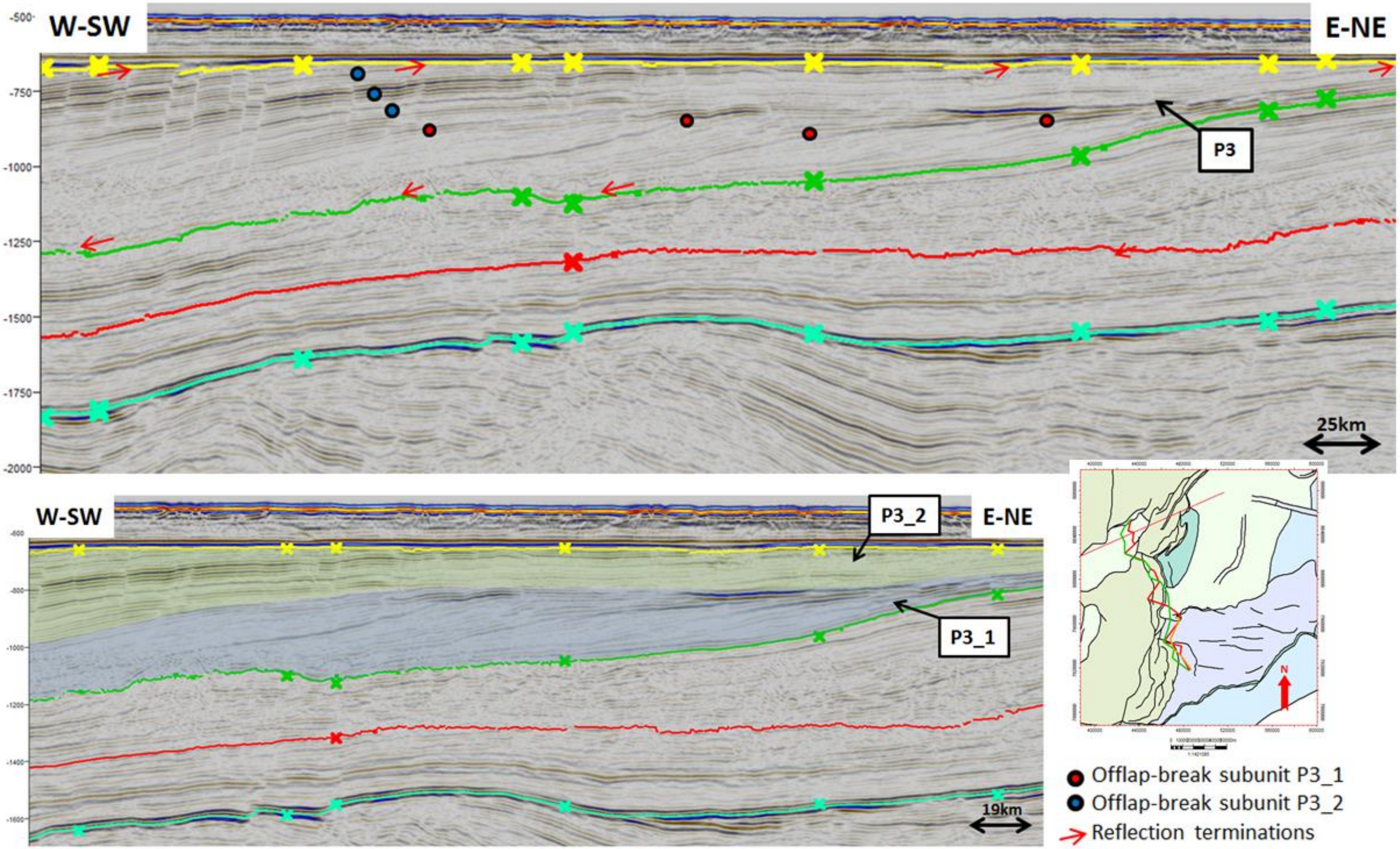


Figure 4.17: Upper seismic section show reflection terminations and offlap-breaks for unit P3 and lower section show the subunits of P3. Seismic legend same as in figure 4.16. Depth in TWT, map modified from NPD (2015a).

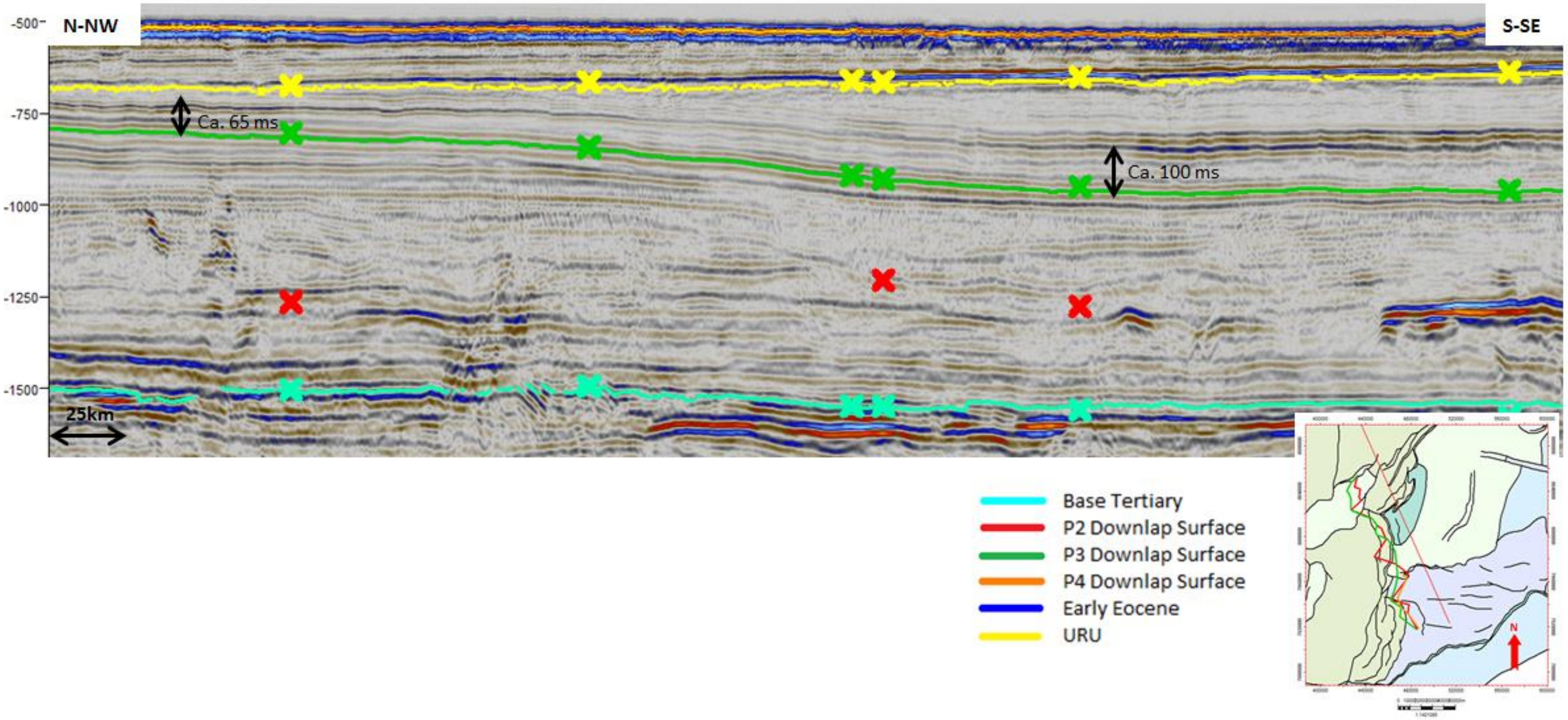


Figure 4.18: Transverse section of unit P3 in the northern Tromsø Basin. The difference in thickness making up a lens shape is highlighted. Depth in TWT, map modified from NPD (2015a).

4.6 Unit P4

P4 is the uppermost prograding unit. It downlaps onto P3 and is bounded at the top by the URU and the Early Eocene surface. This unit is only found in the western Hammerfest Basin (figure 4.19) and has a small areal extent compared to the two other prograding units. This unit prograded from the east towards the west, as the two other prograding units also did.

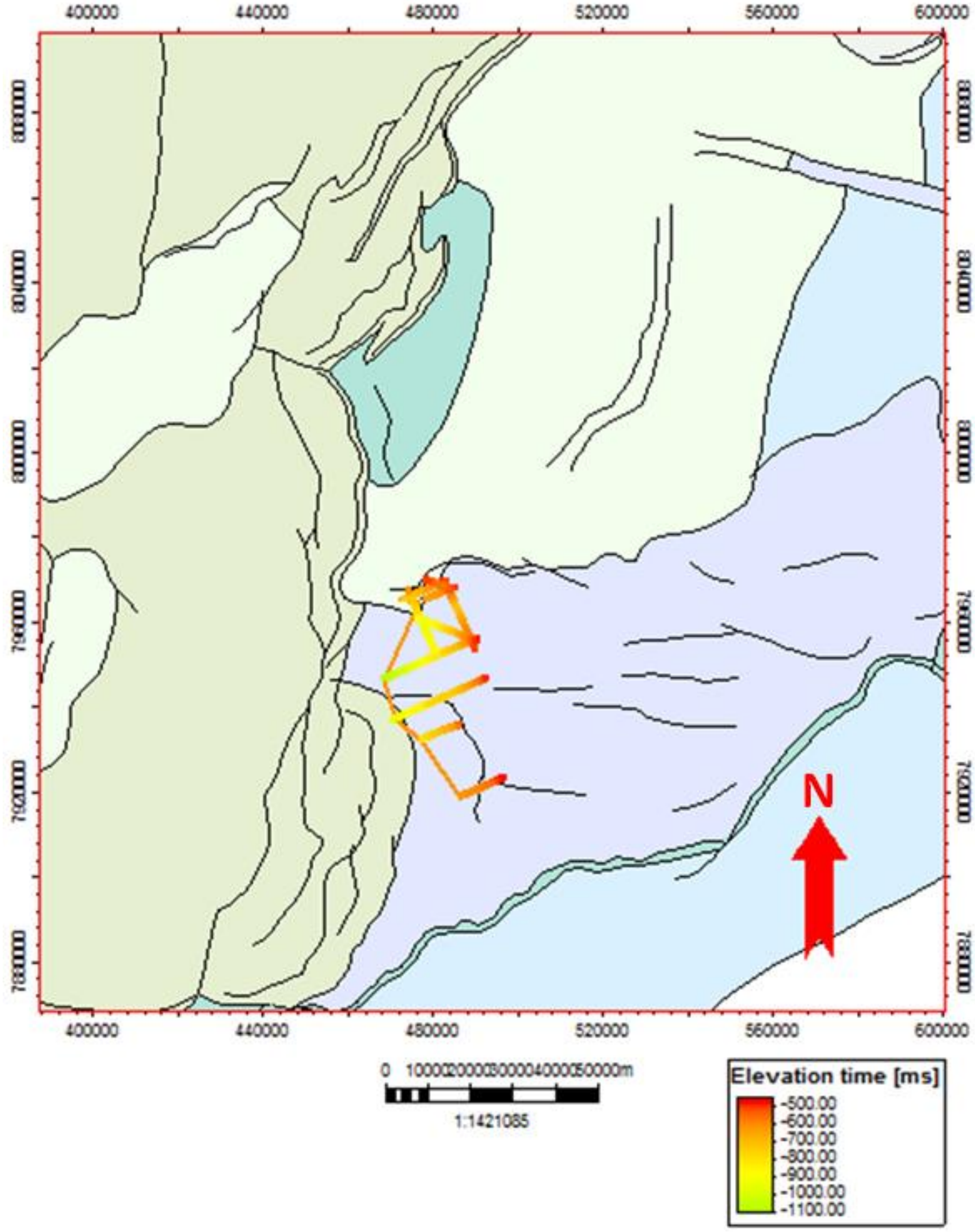


Figure 4.19: Areal extent of unit P4. Map modified from NPD (2015a).

The unit has an average thickness of 240-260 m, and there is a small decrease in thickness towards west (figure 4.20). The decrease in thickness towards west matches well with the two other units and what is expected based on clinoform geometry. The original thickness is only preserved under the Eocene package in a small part of the western Hammerfest Basin, while for the eastern part of the unit the upper part of is removed by erosion (figure 4.20).

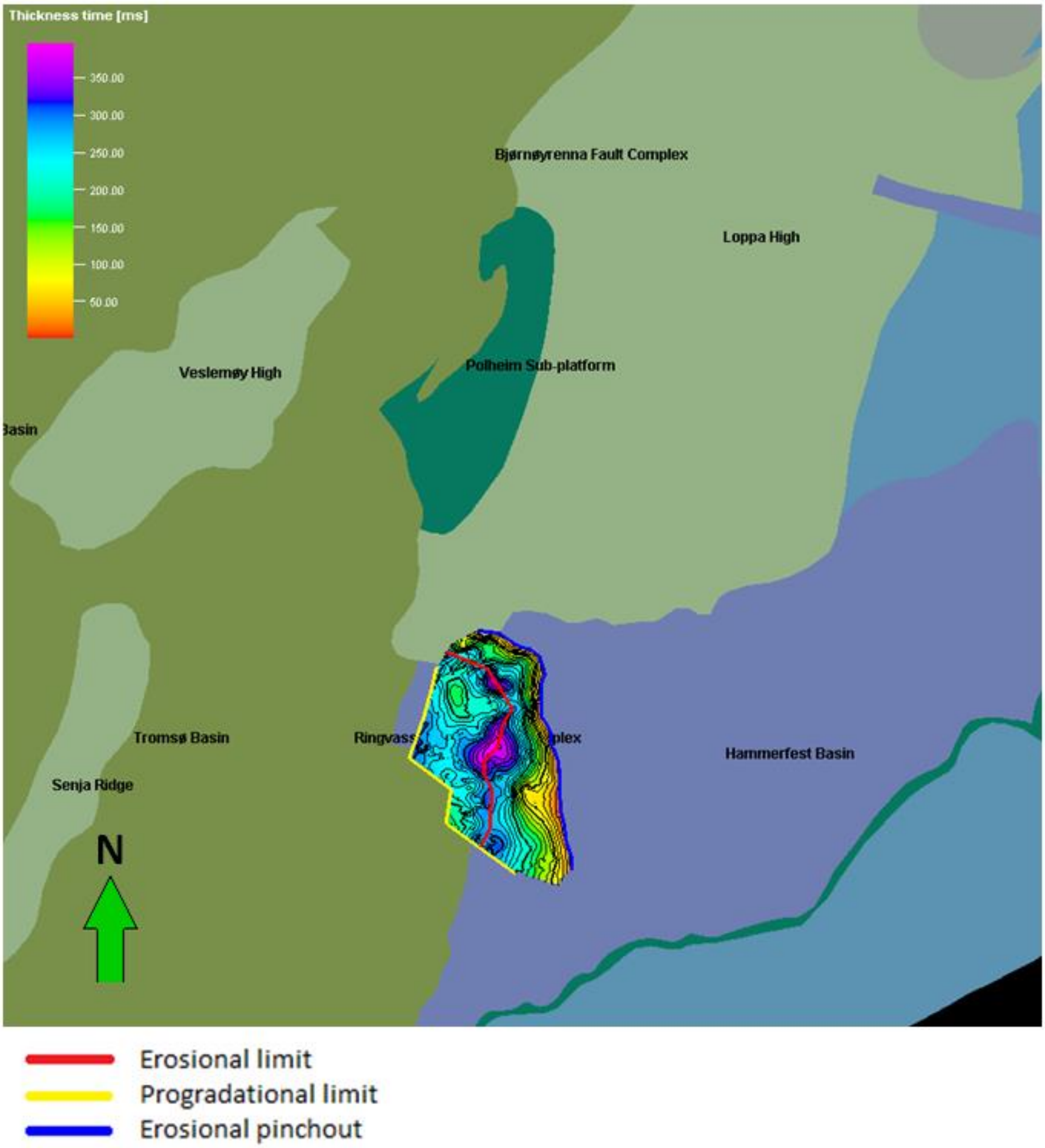


Figure 4.20: Time-thickness map for unit P4. Thickness west of red line has the original thickness preserved while the thickness east of the red line has been reduced by erosion. Map modified from NPD (2015a).

Because of erosion there are no full clinoform profiles preserved, even though the thickness of the succession is sufficient to include clinoforms of similar size within the subunits of P2 and P3. This means that only the lower part of the prograding system is preserved, and it is impossible to determine if the clinoforms are sigmoidal or oblique. The angle of the slopes varies between 2,5°-3°, which is a little steeper than for the subunits of P2 and P3.

The internal reflections have a good continuity, but the amplitude is low (figure 4.21). Since only small parts of the subunit are preserved it is hard to find a clear reflection pattern as seen for the subunits of unit P2 and P3. The reflections are also disturbed by a reflector cutting the seismic layering in some of the seismic sections (figure 4.22). This reflector is clearest on seismic sections in dip direction and is interpreted to represent the opal A to opal CT transition. In transverse sections the reflections are clear and continuous towards east and north (figure 4.23), whereas towards west and south they are more chaotic and weak.

Interpretation:

Because of the limited preservation of the unit it is not possible to say something certain about the depositional environment for the unit. The steep slope deposits suggest oblique clinoforms rather than sigmoidal clinoforms, and hence a higher energy environment. This is only assumptions and cannot be confirmed based on the limited data for the unit.

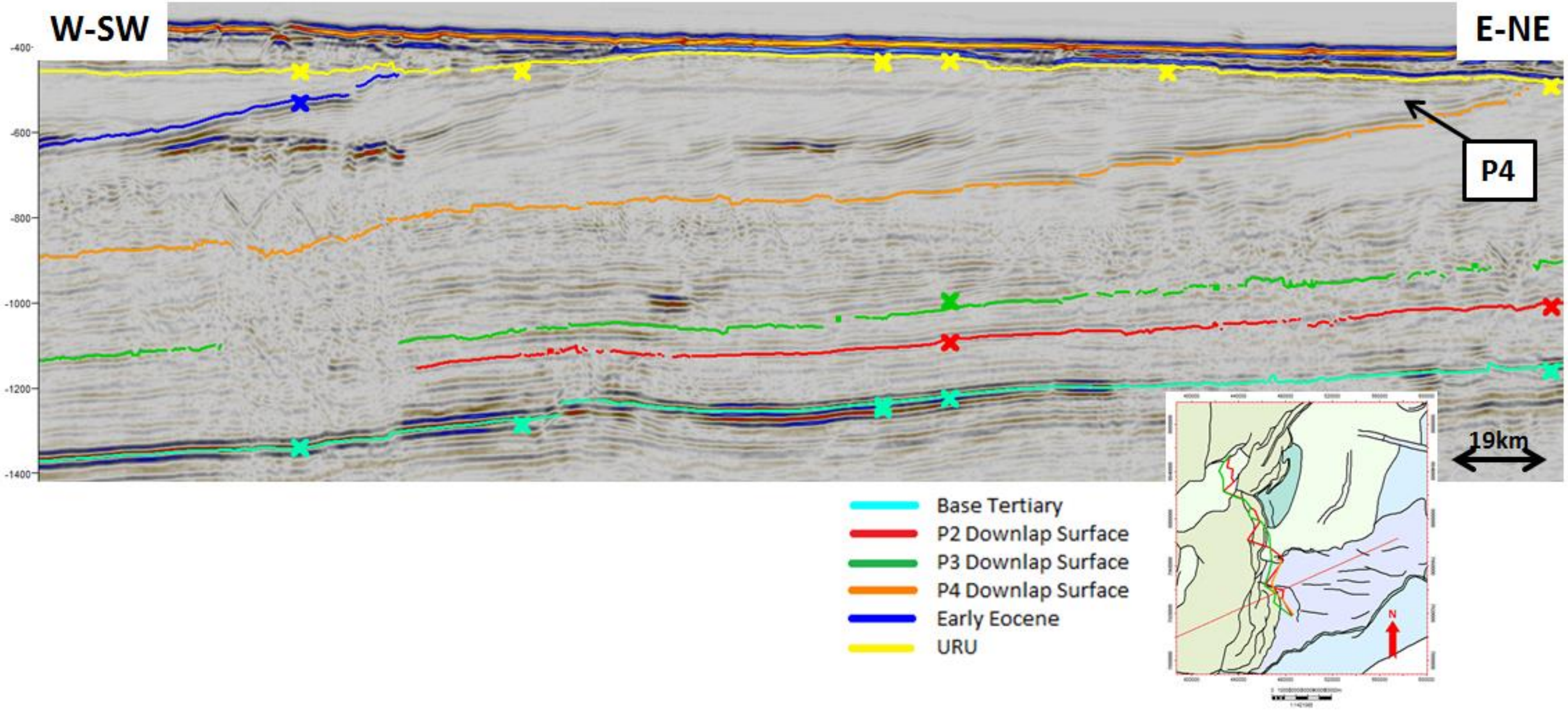


Figure 4.21: Unit P4 in the Hammerfest Basin. Depth in TWT, map modified from NPD (2015a).

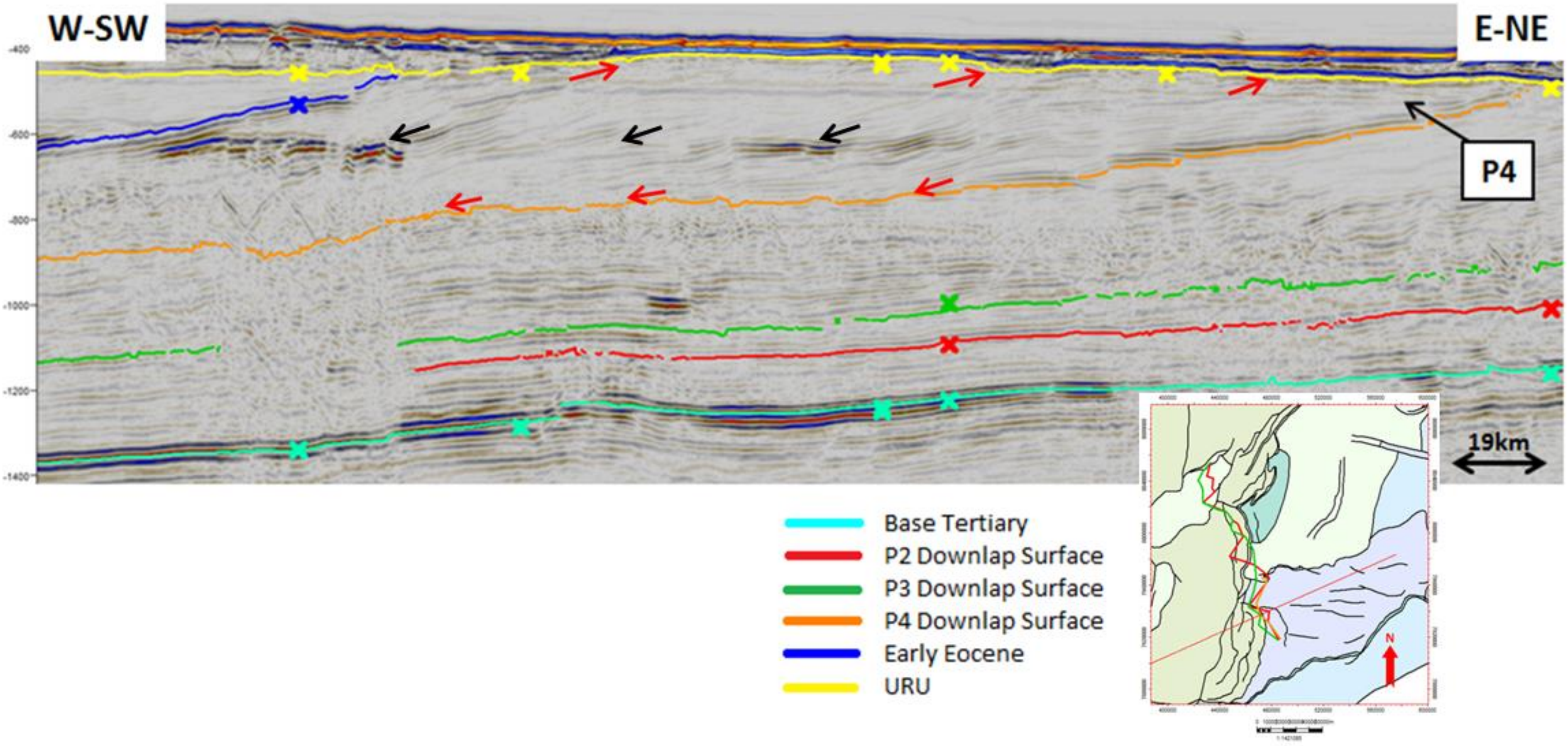


Figure 4.22: Unit P4 with reflection terminations. Black arrows shows apparent downlaps towards the reflector representing the Opal A to Opal CT. Depth in TWT, map modified from NPD (2015a).

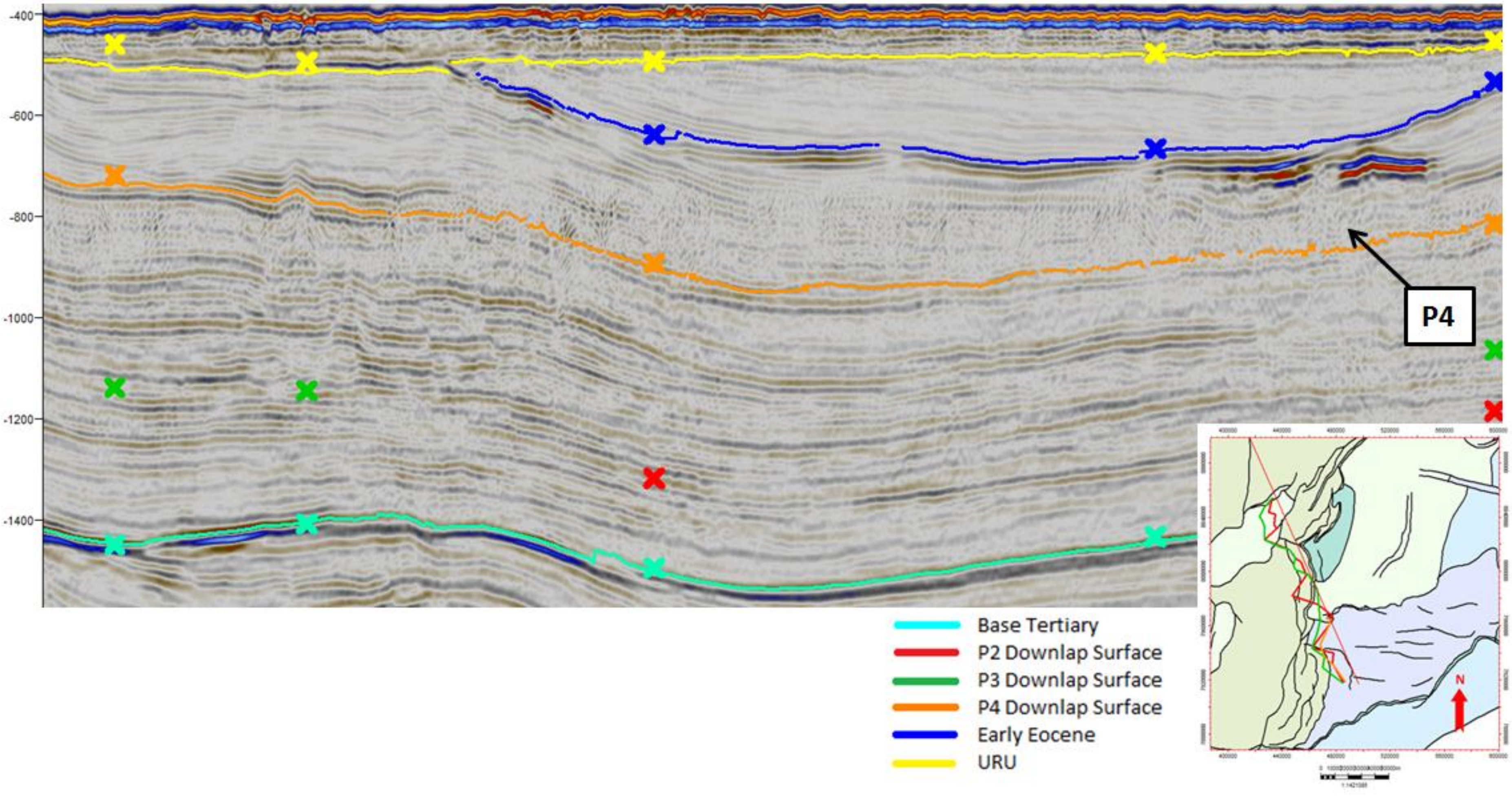


Figure 4.23: Transverse section of P4. **Depth in TWT, map modified from NPD (2015a).**

5 Discussion

5.1 Basin infill history

The seismic patterns show that the first deposits after the Cretaceous – Paleogene transition was aggradational both in the Tromsø and Hammerfest basins. Deposition was widespread and occurred as a sheet, covering large parts of the Barents Sea (Worsley et al., 1988). The aggradational infill continued throughout the whole Paleocene in the southern central Tromsø Basin. In the Hammerfest Basin and northern Tromsø Basin the seismic pattern shows that the infill changed to progradational after the deposition of the first widespread aggradational unit. The change from aggradational to progradational infill is marked by the downlap surface for unit P2 onto unit P1. At least three main successions of Paleocene progradational units are recognized over the lower aggradational P1 unit in the seismic data. These units can be subdivided further into subunits, based on internal reflections.

The first depositions for unit P2 and P3, subunits P2_1 and P3_1, show oblique clinoform pattern with a flat offlap-break trajectory (figures 4.11 and 4.17). The oblique clinoform patterns of the subunits indicate rapid infill of the basin (Mitchum et al., 1977b). Then followed deposits of sigmoidal clinoforms and ascending offlap-break trajectory for the rest of the units (figures 4.11 and 4.17), subunits P2_2, P2_3 and P3_2. This pattern indicates that the energy of the depositional environment for the two units went from relative high energy in the lower part to comparably lower energy higher up in the unit. For the last progradational unit, P4, insufficient material is preserved to compare the infill pattern with the two previous progradational units.

All three prograding units show the same progradational direction. The progradational terminations of subunit P3_1 are found further east in the Hammerfest Basin than in the northern Tromsø Basin; this shows that the front of the subunit had a SSE – NNW strike. The SSE – NNW strike is also visible along the front of the three progradational units, indicating that the direction of progradation was from ENE towards WSW (figure 5.1).

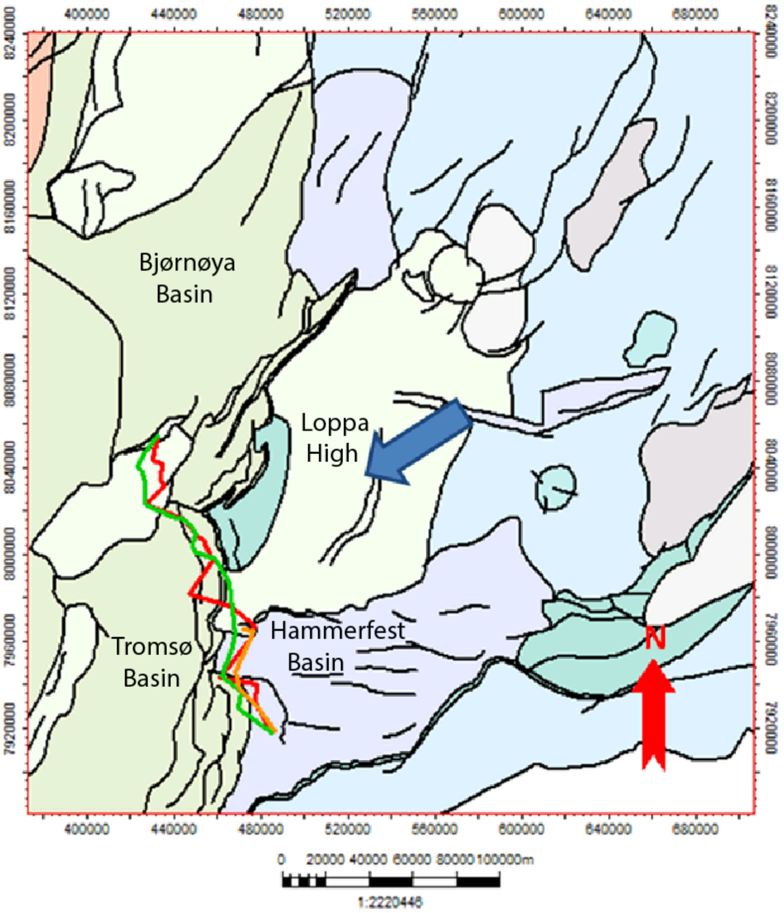


Figure 5.1: Blue arrow indicates the main direction of progradation of the three progradational units; P2 (red), P3 (green) and P4 (orange). Map modified from NPD (2015a).

Oblique clinoforms from unit P3 is found both in the Hammerfest Basin and northern Tromsø Basin (figure 4.15) presently separated by the Loppa High. It is expected that the clinoforms were deposited under the same conditions based on similarities of the geometrical style and slope angles between the clinoforms observed west and south of the Loppa High. The clinoforms from both areas are interpreted to be part of subunit P3_1, and are therefore part of the same depositional system. For the rest of the units P2 and P3, only the lower part of the clinoforms are preserved in the Hammerfest Basin and this makes correlation between the units in the Hammerfest and Tromsø basins difficult. However, similarities of slope angle and reflection pattern are observed between the units in seismic lines west and south of the Loppa High (figures 4.10 and 4.16). It is therefore assumed that the preserved part of units P2 and P3 in the Hammerfest Basin are part of the same depositional system as for the units west of Loppa High. This means that the prograding units found in the Hammerfest and Tromsø basin were part of one single, large depositional system. Glørstad-Clark et al. (2011b) observed Paleocene prograding clinoforms in the Bjørnøya Basin, located north of the Tromsø Basin. These clinoforms were probably part of the same depositional system, which suggest that the system reached further north than mapped out in this study. Since the prograding units are deposited over each other it is assumed that the succession had a significant thickness, and this will be discussed in chapter 5.7. Because of Late Cenozoic uplift and erosion it is difficult to speculate on the original size of the system, and it is not known if there were younger prograding units of the system later removed by erosion.

The lenses in the transverse sections for unit P2 in the Hammerfest Basin and subunit P3_1 in the northern Tromsø Basin (figure 4.12 and 4.18) is a result of the system infill pattern. The system front, oriented along a NNW-SSE trend, was formed by coalescing lobes building out towards WSW. The amount of deposited sediments was greatest in the central part of the lobe and less towards the sides, resulting in the observed lensoid shape geometries (figures 4.12 and 4.18).

5.2 Loppa High

Previous arguments for interpreting the Loppa High as a positive structure was onlaps in the lower part of the Paleocene succession towards the southern and western flank of the high, and draping and thinning of the Paleocene succession over the southwestern Loppa High (Knutsen and Vorren, 1991; Knutsen et al., 1992). However, no onlaps towards the southern or western flank of Loppa High are observed in this study. Regarding the draping and thinning, this study shows that units P2 and P3 are truncated by the URU towards and over the Loppa High. This implies that the thinning of the Paleocene succession in the area is a result of younger, mainly Late Cenozoic, uplift and erosion, and not a depositional feature as inferred earlier by Knutsen and Vorren (1991) and Knutsen et al. (1992).

The thin, truncated sheet of unit P1, and in places the lower part of unit P2, preserved over the Loppa High, shows that the Paleocene package must have covered a larger area than the high during Paleocene time. The characteristics of P2 and P3 towards the southern and western flank of Loppa High also give the impression of a larger areal extent of the Paleocene succession over the Loppa High than what is observed in the seismic data (figures 4.2 and 4.7). The units are relatively thick until they are truncated by the URU, and it is therefore natural to assume that the sedimentary succession must have had a larger areal extent over the Loppa High before younger, mainly Late Cenozoic, uplift and erosion. Oblique clinoforms from subunit P3_1 were also found south of the Loppa High in the Hammerfest Basin, and west of the Loppa High in the northern Tromsø Basin. The deposits exhibiting these clinoforms were most likely part of the same large progradational system, thus implying that the deposits from the prograding system between these two areas must have been eroded. These observations indicate that the prograding units P2, P3 and P4 covered the Loppa High area prior to the uplift and erosion. These units comprise together a significant sediment thickness that could not have been deposited in the Loppa High area if the high was a positive structure.

Based on these observations it is assumed that Loppa High was a part of the shallow Barents Shelf during Late Paleocene, and not a positive structure as earlier expected (Knutsen and Vorren, 1991; Knutsen et al., 1992; Faleide et al., 1993). Glørstad-Clark et al. (2011b) suggested that Loppa High could have been a bathymetric high affecting the submarine sediment routing during parts of the Paleogene time period. There have not been any observations supporting Loppa High as a submerged bathymetric high, but the possibility cannot be ruled out.

5.3 Sediment source

Several earlier papers have operated with the Loppa High as the main source for the Paleocene sediments prograding into the Tromsø Basin (Knutsen and Vorren, 1991; Vorren et al., 1991; Knutsen et al., 1992; Faleide et al., 1993). This was mainly based on the SSW prograding clinoforms proximal to the southwestern Loppa High. As discussed in chapter 5.1 and 5.2 the Paleocene prograding system was one large system covering the wider Loppa High area, and the Loppa High could therefore not have been the source for these sediments.

The Torsk Formation has a rather uniform lithology mainly consisting of grey claystones (Worsley et al., 1988), and it is therefore assumed that all three units have the same sediment source. Since the Torsk Formation is fine grained the sediments can have been transported over larger distances before deposition in the Hammerfest and Tromsø basins. The progradational direction for the system was from ENE towards WSW, this indicates that the source for the Paleocene sediments must be located to the east and/or north of the study area.

There is not found enough information to give a good suggestion of a sediment source to the east. The discussion will therefore mainly focus on possible sediment source areas located to the north and/or northeast of the study area.

Glørstad-Clark et al. (2011b) suggested that the sediment source could be related to Large Igneous Provinces (LIPs) from Cretaceous. There are documented two phases of Cretaceous magmatism north of the Barents Sea; the first in Early Cretaceous dated to ca. 120-125 Ma and a second in Late Cretaceous dated to ca. 80 Ma (Corfu et al., 2013). The first phase affected a wider region in the Artic, including Svalbard and Franz Josef Land in the northern Barents Sea (Maher, 2001; Corfu et al., 2013). The second event was centered north of Greenland (Corfu et al., 2013). Thermal doming related to the LIP development on Svalbard resulted in erosion of more than 1000 m in Cretaceous (Maher, 2001). Uplifted areas related to these Cretaceous LIPs to the north and northeast of the study area could have been the sediment source for the prograding Paleocene succession in the SW Barents Sea (figure 5.2).

A sediment source located to the north would probably have affected the deposits on Svalbard from the same period. The sediment source for the deposition of the mid to late Paleocene Basilika and Grumantbyen formations in the Paleogene Central Basin on Spitsbergen is expected to be located to the east and/or northeast of the basin (Figure 5.2) (Müller and Spielhagen, 1990). This implies that there were uplifted areas to the north of the study area, which could have acted as a sediment source for the Prograding system in the SW Barents Sea as well as for the Paleocene succession on Spitsbergen.

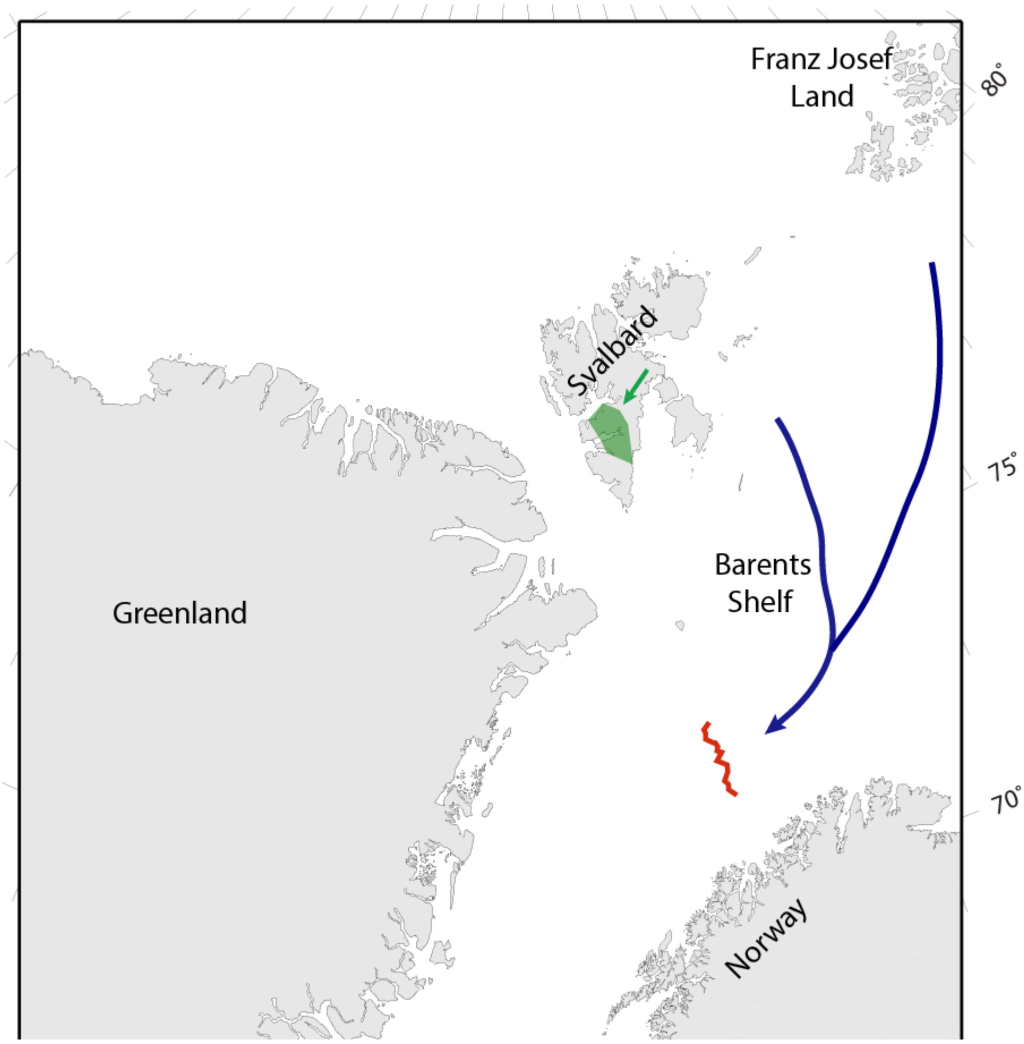


Figure 5.2: Possible sediment route (blue lines) for a sediment source to the north, red line represents the front of the prograding system. The green arrow represents the infilling direction of the Paleogene Central Basin (green field). Paleogeographic map modified from Faleide et al. (2010).

Late Cenozoic uplift and erosion has removed indicators of the Early Paleogene environment from the northern Barents Shelf. This means that there is no evidence of a sediment source located to the north in the present northern Barents Sea. However, the Lomonosov Ridge, presently separated from the Barents Shelf by the Eurasia Basin, can provide information of the Paleocene setting for the northern Barents Shelf. Reconstruction of the paleogeographic setting of the area for Early Paleogene shows that the Lomonosov Ridge probably was located at the northernmost Barents Shelf margin in Paleocene (figure 5.3). The Lomonosov Ridge separated from the shelf during late Paleocene (Backman and Moran, 2009) and was not affected by the Late Cenozoic uplift and erosion. It is therefore natural to assume that the Paleocene setting for the northern Barents Shelf is preserved on the Lomonosov Ridge.

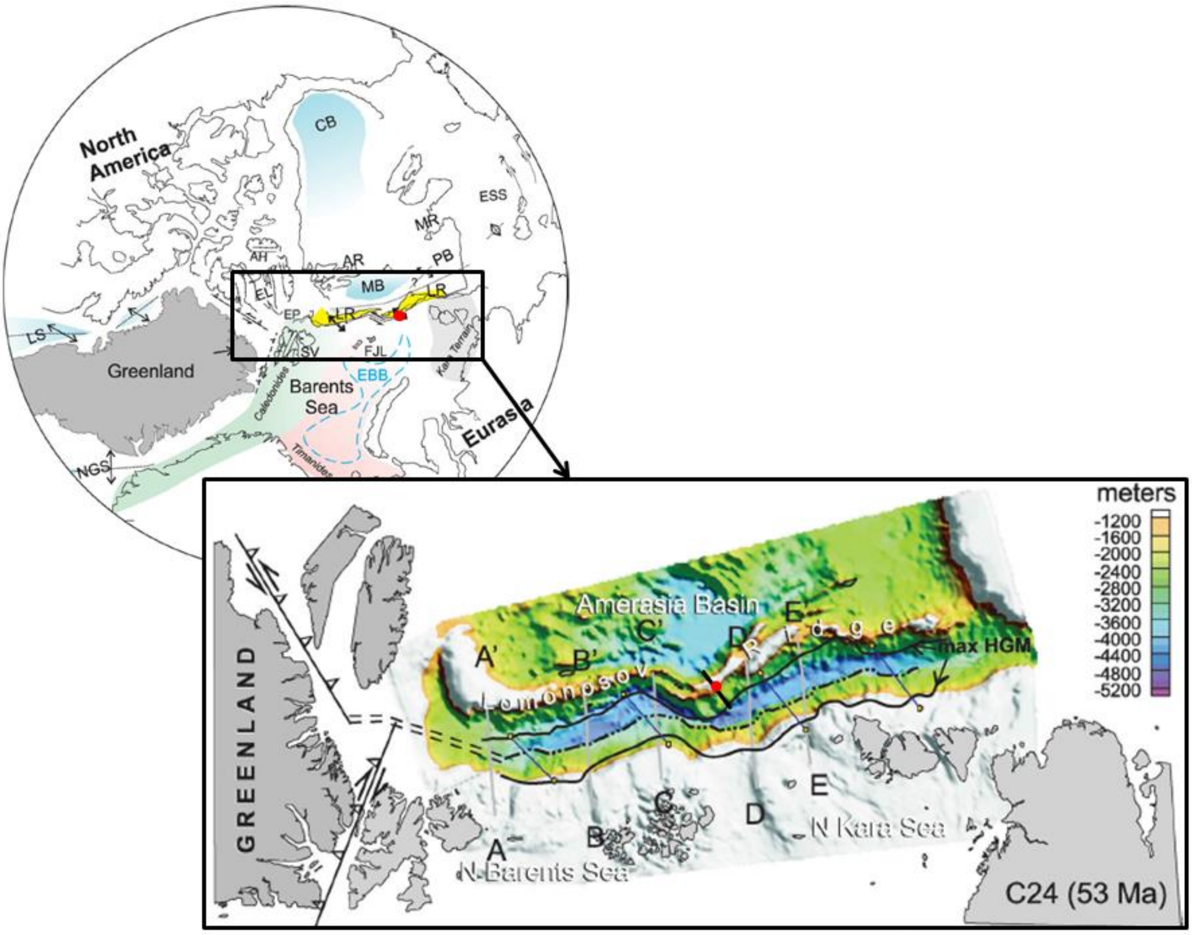


Figure 5.3: Paleogeographic map of the areas north of the Barents Shelf. Red dot marks the location of ACEX well and black line marks position of seismic line AWI 91090. Modified from Minakov et al. (2012).

There is limited seismic data coverage of the Lomonosov Ridge, but the Arctic Coring Expedition (ACEX) provides important information of the Cenozoic succession from the Arctic Ocean. The location of the ACEX well is displayed in both the seismic section in figure 5.4 and in the maps in figure 5.3. The seismic section in figure 5.4 shows a clear unconformity under the Cenozoic succession. The hiatus is dated to be from 56,2-65,5 Ma, but could have extended down to Campanian based on samples from the bottom of the ACEX core (Backman and Moran, 2009; Langinen et al., 2009). The timing of the hiatus implies that there was uplift and erosion at the ACEX borehole site, and hence likely also in the northern Barents Shelf, during deposition of the Paleocene succession in the SW Barents Sea. This provides convincing evidence for a north to northeastern located source area for the Paleocene succession in the SW Barents Sea.

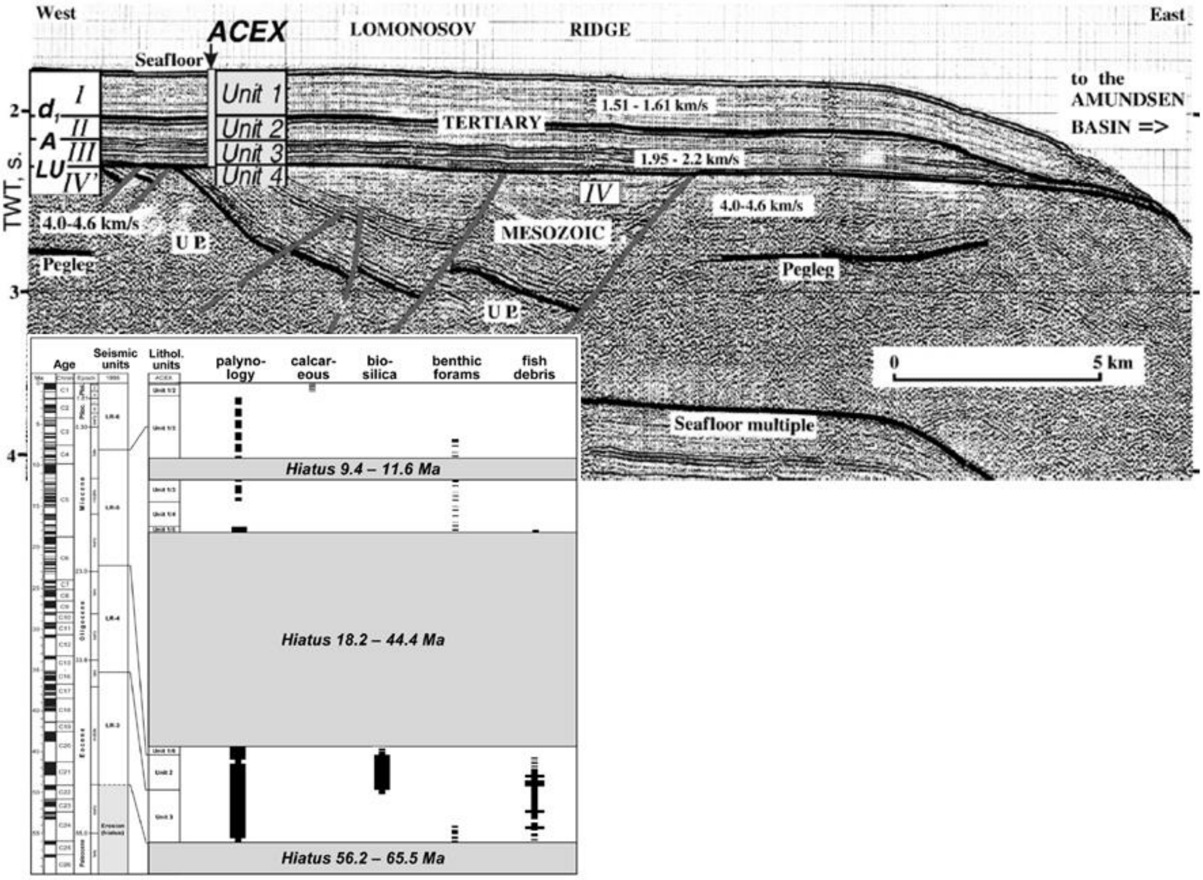


Figure 5.4: Seismic section AWI 91090 (location marked in figure 5.3) with the ACEX well location marked (Langinen et al., 2009). Chronostratigraphic chart shows age of hiatuses and deposition (Backman and Moran, 2009).

5.4 Platform-edge and shoreline position

Offlap-breaks representing the platform-edge are mapped out in several seismic profiles in the Tromsø Basin and in one profile in the Hammerfest Basin (figures 4.11 and 4.17). Since full clinoform profiles only are found for the two first prograding units, P2 and P3, it is only possible to infer the position and evolution of the platform-edge during deposition of these two units. The front of the prograding systems had a SSE-NNW strike, and it is expected that the platform-edge had the same orientation.

Even though the units are deposited over each other it is clear from the trajectory of the offlap-breaks (figures 4.11 and 4.17) that the platform-edge did move westward. The movement is expected to be the same as the direction of progradation, i.e. towards WSW. Both units show an overall westward movement of the platform-edge, and the only eastward migration is observed from subunit P2_1 to subunit P2_2 (figure 4.11). This landward migration can be explained by a high rate of increasing sea-level outpacing the sediment supply (Anell et al., 2014). The migration rate of the platform-edge was highest for the oblique clinoforms for the subunits P2_1 and P3_1, whereas the sigmoidal clinoforms in subunits P2_2, P2_3 and P3_2 suggest a slower platform-edge advance.

The limit for the lapouts for all three prograding units follows the eastern part of the Tromsø Basin rather well up to the northern part of the study area (figure 4.4). Lapouts reach over the eastern flank of Veslemøy High in the northern part of the study area. The Tromsø Basin was a deep and relatively narrow basin during deposition of the Paleocene succession and this is probably the reason for why the lapouts did not continue into the basin. The Tromsø Basin acted as a sink for the prograding system, and the abrupt change from the shallower Hammerfest Basin to the deep Tromsø Basin prevented the downlaps from further westward progradation and the depositions in the Tromsø Basin is exclusively aggradational.

The position of the shoreline is an important factor for the prograding system and has great influence on the amount and type of sediment supply (Helland-Hansen and Hampson, 2009; Johannessen and Steel, 2005). The location of the shoreline is not known for the Paleocene, but must have been located somewhere to the east and/or north of the study area. There is not found any indications of a shoreline located to the east of the study area. The Paleocene succession is missing in the eastern part of the SW Barents Sea due to erosion, so possible evidences of a shoreline located to the east could have been eroded during Late Cenozoic. To the north in the Paleogene Central Basin on Spitsbergen the Firkanten Formation is interpreted to be delta to inner-shelf deposits and the Basilika Formation and Grumantbyen Formation is interpreted to be deposited in outer- to inner-shelf deposits (Steel et al., 1985; Müller and Spielhagen, 1990). Because of Late Cenozoic uplift and erosion the Paleocene succession is missing between the Bjørnøya Basin and the Paleogene Central Basin on Spitsbergen (figure 2.4), and it is therefore not known if these deposits are from the same depositional system. The deposits could be from different depositional system, but the deposits in the Paleogene Central Basin in Spitsbergen imply a shallower environment to the north and east. The evidence of a sediment source located in the northern Barents Shelf area, as discussed in chapter 5.3, also infer a shoreline located to the north and northeast of the study area. It is therefore assumed a shoreline located to the north and northeast of the study area, but an exact position cannot be defined (figure 5.5).

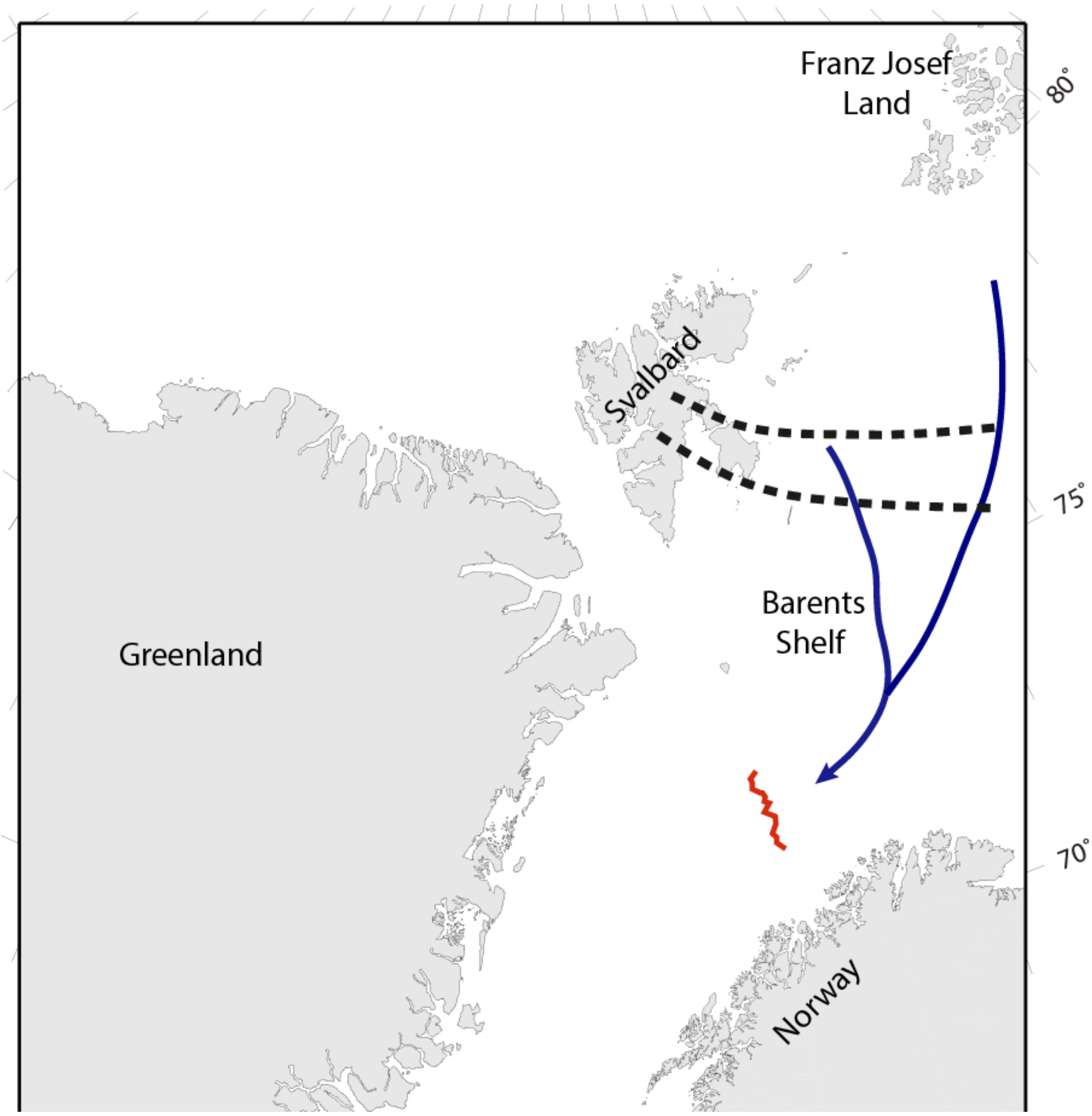


Figure 5.5: Possible shoreline locations (dotted, black lines), possible sediment rout (blue lines) and front of the prograding system (red line). Paleogeographic map modified from Faleide et al. (2010).

5.5 Correlation with biostratigraphy analysis of well 7119/9-1

Previous work on the biostratigraphy of the Torsk Formation in the western Barents Sea (Nagy et al., 1997; Nagy et al., 2000) can be used to define the paleoenvironment and depositional age. Nagy et al. (1997) divided the Paleocene succession in well 7119/9-1 into assemblages based on palynomorphs, benthic foraminiferal and diatoms. The three groups of assemblages give approximately the same results of age and depth intervals. For this correlation it is chosen to use the benthic foraminiferal assemblage, and the six defined assemblages are displayed in table 1.

Table 1: Benthic Foraminiferal assemblages, with corresponding age and depth interval, of the Torsk Formation in well 7119/9-1 (Nagy et al., 1997).

Benthic Foraminiferal assemblages	Time period	Depth interval (m)
Reticulophragmium amplectens	Early Eocene	510-710
Spiroplectammina navarroana	Late Paleocene + Early Eocene	710-890
Haplophragmoides* aff. *eggeri	Late Paleocene	890-1070
Reticulophragmium paupera	Middle late Paleocene	1080-1160
Spiroplectammina spectablis	Middle late Paleocene	1180-1400
Psammosphaera fusca	Early late Paleocene	1410-1450

Well 7119/9-1 is situated west of the progradational limit for the interpreted prograding units (figure 5.6), but the Base Tertiary and Early Eocene were interpreted at the well location. The downlap surfaces for the three prograding units were extended to the well location in order to investigate if a correlation is possible between the work of Nagy et al. (1997) and the interpreted units.

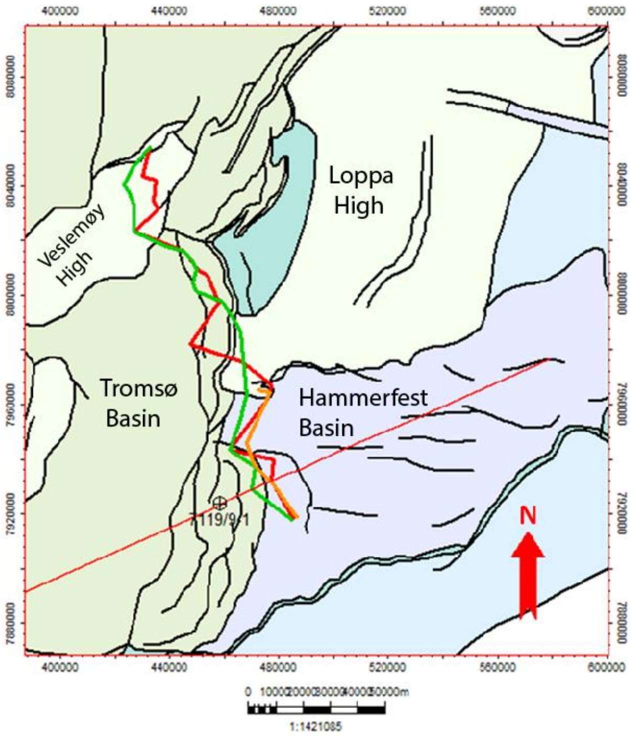


Figure 5.6: Map with location of well 7119/9-1 and the seismic line used for correlation. Map modified from NPD (2015a).

The interpreted Early Eocene fits well with the base of the *Reticulophragmium amplectens* assemblage (Early Eocene age), and the interpreted Base Tertiary fits well with the base of the *Psammosphaera fusca* assemblage (early Late Paleocene age). The correlation with the prograding units has some more uncertainties since the downlap surfaces had to be extended over to the well location. The correlation shows that the interpreted units have a good match with the assemblages (table 2).

Table 2: Correlation between interpreted units and Benthic Foraminiferal assemblages from Nagy et al. (1997) in well 7119/9-1.

Interpreted units	Corresponding Benthic Foraminiferal assemblages	Depth interval (m)
Unit P4	*Spiroplectammina navarroana*	710-890
Unit P3	*Haplophragmoides* aff. *eggeri*	890-1080
Unit P2	*Reticulophragmium paupera*	1080-1180
Unit P1	*Spiroplectammina spectablis* and *Psammosphaera fusca*	1180-1450

5.6 Water depth

The height of the observed clinoforms can give an estimate of the paleowaterdepth (Glørstad-Clark et al., 2010; Anell et al., 2014). Because a typical average velocity is used to calculate the height of the clinoforms the measurement may not be accurate. The clinoforms have also been compacted after deposition and since this is not corrected for in the data, the estimated paleowaterdepth is a minimum estimate. Biostratigraphy analysis for the Paleocene part of the Torsk Formation in the Tromsø Basin stated that the depositional environment was upper to middle bathyal (Nagy et al., 1997; Nagy et al., 2000). The height of the clinoforms varies from ca. 100 m to ca. 300 m, which means the minimum paleowaterdepth for these clinoforms matches well with the biostratigraphy analysis.

Through Paleocene time the relative sea level raised and reached a maximum water depth in the latest Late Paleocene (Nagy et al., 1997). After the water depth maximum, there was a general shallowing during the Early Eocene (Nagy et al., 1997). According to previous correlation (chapter 5.5) the maximum water depth occurred during deposition of unit P4. An estimate of the maximum paleowaterdepth cannot be done based on the available data since there are no clinoforms preserved for this unit. The rise in relative sea level can be inferred based on the seismic data where the height of the clinoforms increases from ca. 200 m in subunit P2_1 to 300 m in subunit P3_2. The height of the clinoforms decreases from 200 m to 100 m between subunit P2_1 and P2_2, but since the clinoforms of subunit P2_2 downlaps onto the topset of subunit P2_1, the relative sea-level must have increased.

Based on the offlap-break trajectory the pattern between unit P2 and P3 are rather similar with indications of a more stable sea-level during the earliest depositions followed by rising sea-level during the later deposition for each unit. The oblique clinoforms and relative flat trajectory for the subunits P2_1 and P3_1 indicate that there was a stable or falling sea-level (Helland-Hansen and Hampson, 2009, Mitchum et al., 1977b). The depositional pattern for both units then changes to sigmoidal clinoforms for the remaining subunits, P2_2, P2_3 and P3_2. The preservation of topset indicates that these subunits were deposited during a rising relative sea-level with more storage potential on the shelf (Mitchum et al., 1977b, Sangree and Widmier, 1978). The trajectory back-step between subunit P2_1 and P2_2 (figure 4.11) can be explained by a rise in sea-level higher than the sediment supply (Anell et al., 2014).

5.7 Thickness/erosion

In this section of the discussion, the term “original thickness” refers to the depositional thickness after compaction, but unaffected by erosion. The time-thickness map for the Paleocene succession (figure 4.6) shows that it is mainly the aggradational succession in the Tromsø Basin that is unaffected by the late Cenozoic erosion. The original thickness of the progradational succession is therefore unknown, but it is possible to make an estimate of the original thickness.

Since the units are deposited over each other, a summation of their thicknesses will be used to make an assumption of the original thickness. For unit P2 and P3 full clinoform profiles are preserved in the northern Tromsø Basin and it is assumed that the thickness in this area is close to the original thickness. Figure 4.9 show that unit P2 in the northern Tromsø Basin has areas which were not affected by erosion. Several seismic sections show full clinoform profiles with a thickness of ca. 600 m preserved within the areas not affected by the erosion. Unit P3 has full clinoform profiles preserved in the same area, but the thickness has been reduced by erosion. The thickness in this area will still provide the best prediction of the original thickness, assumed to be ca. 700 m. Too much of the upper unit P4 is eroded to infer the original thickness, and the maximum thickness of ca 400 m is applied as the unit thickness. Based on this, the thickness for the three progradational units is ca. 1700 m. The thickness of the underlying aggradational P1 unit, of ca. 130 m, was added to estimate the total depositional thickness in the areas where the prograding units were deposited. This yields an original thickness of ca. 1830 m for the Paleocene succession.

The estimated thickness of the Paleocene is based on an assumption and limited data and thus represents simple calculations. Units deposited over the preserved units may have existed but later been eroded, and it has not been corrected for compaction. The thickness might have varied over the area, but this calculation is meant to give an average estimate.

The uplift and erosion in the Barents Sea has been important for the petroleum prospectivity in the area (Dorè, 1995; Dorè and Jensen, 1996; Ohm et al., 2008; Henriksen et al., 2011). Several papers have estimated the uplift and erosion in the Barents Sea (e.g. Vorren et al., 1991; Riis and Fjeldskaar, 1992; Dimakis et al., 1998; Henriksen et al., 2011), and the general trend in the western Barents Sea is an increased uplift and erosion towards east and north (figure 5.7). This erosional trend is the reason why parts of the Paleocene succession is preserved in the Hammerfest Basin, but eroded over the Loppa High and further north and east. At least two main phases of uplift and erosion have occurred after deposition of the Paleocene succession; the first is a pre-glacial Paleogene event, and the second is a Neogene event related to glaciation (Cavanagh et al., 2006; Ohm et al., 2008; Green and Duddy, 2010). The burial history is closely related to the temperature history in a depositional basin, and the maximum burial is therefore closely related to the maximum maturity. The first of the uplift and erosional events represent the maximum burial of the SW Barents Sea. This event is the Paleogene pre-glacial uplift, which is estimated to have occurred in mid-late Eocene time at ca. 40-35 Ma (Cavanagh et al., 2006; Ohm et al., 2008; Green and Duddy, 2010).

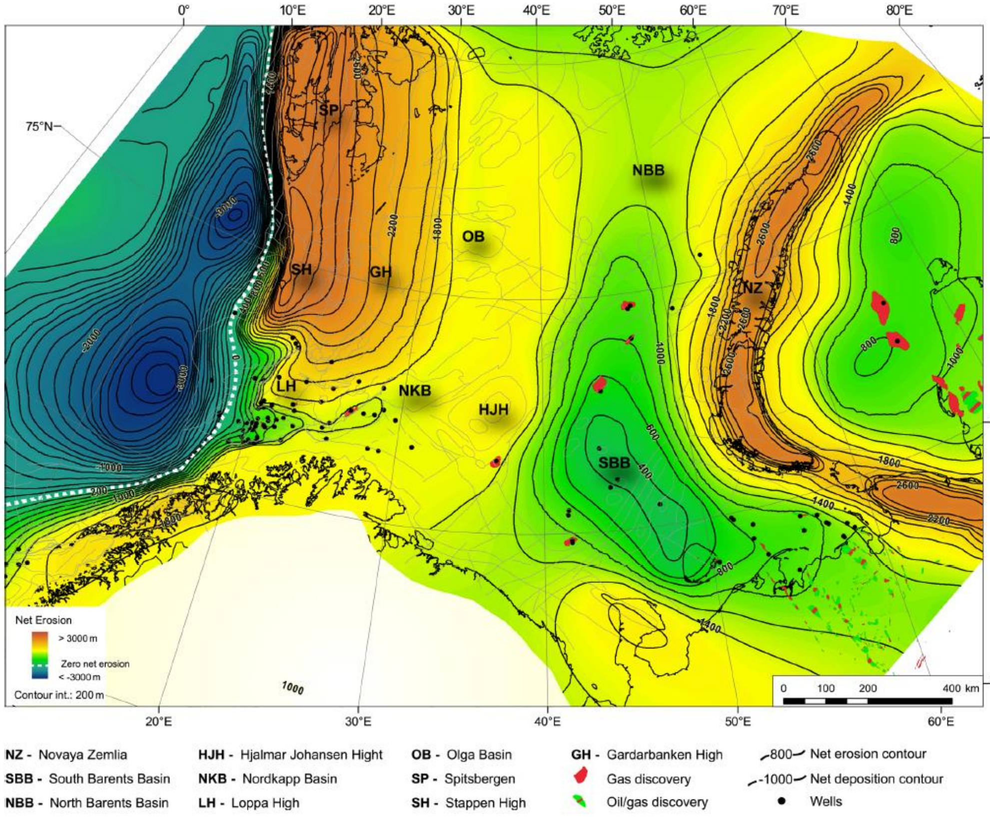


Figure 5.7: Regional map of estimated net erosion in the Barents Sea (Henriksen et al., 2011).

The estimated original thickness of the Paleocene succession is higher than the estimated erosion in some places where the Paleocene succession is completely removed at present. This implies that there should be Paleocene deposits preserved in places where they are not present, and the thickness estimate is therefore incorrect. The assumption of vertically stacking and simple addition of the thickness of the units appears to have resulted in an overestimate of the total original thickness of the Paleocene succession. Even though the thickness estimate was slightly overestimated it indicated a significant thickness, and the Paleocene prograding system probably made up the larger part of the Early Paleogene succession in the Hammerfest Basin and over the Loppa High prior to uplift and erosion.

The first of the uplift and erosion events is set to 40 Ma (Cavanagh et al., 2006; Ohm et al., 2008; Green and Duddy, 2010), implying that deposition continued into Eocene over the deposits of the Paleocene progradational system. Also the erosion of this succession had to be part of the total erosion estimate, and is a further indication of the assumptions of an overestimated thickness of the Paleocene deposits. Some of the Eocene deposits can be observed over the Paleocene succession in the Tromsø Basin (figure 4.1), but uplift and erosion has completely removed the succession from the Hammerfest Basin. It is not known how thick the Eocene cover was over the Paleocene succession. A detailed study of the Eocene succession in the area would potentially provide a better understanding of the Paleogene burial and uplift history.

5.8 Development of accommodation

Significant accommodation must have been available for the thick Paleocene succession to be deposited. Accommodation is controlled by the combination of sea-level changes and the subsidence rate (Myers and Milton 1996). Accommodation must have been generated prior to deposition of the first widespread sheet of Paleocene succession. The development of prograding clinoforms is an indication of an increase in accommodation coupled with an increase in sediment supply (Glørstad-Clark et al., 2011b).

Even though there is no paleontological evidence for a shallowing related to the Cretaceous – Paleogene hiatus (Nagy et al., 1997), there are indicators of uplift of the Barents Shelf prior to the deposition of the Paleocene succession. On Svalbard the Paleocene succession rests unconformably on Early Cretaceous deposits (Nagy et al., 2000), and in the shallow platform areas in the SW Barents Sea there are only thin deposits from Late Cretaceous (Faleide et al., 1993). The thin and absent Late Cretaceous succession is an indicator for uplift and erosion related to the Cretaceous – Paleogene hiatus.

After the Cretaceous – Paleogene transition the Barents Shelf must have subsided to make accommodation for the Paleocene succession. Reconstruction of the burial history in the Hammerfest Basin by Cavanagh et al. (2006) and Green and Duddy (2010) implies a rapid subsidence of the area. The observed fault movements are not large enough to generate the needed accommodation for the Paleocene succession, so the subsidence must be related to other mechanisms. The widespread deposition of the first part of the Paleocene succession and the size of the Paleocene prograding system indicates that large parts of the Barents Shelf subsided prior to Paleocene deposition. Subsidence of large scale like this is normally related to deeper processes, so the subsidence that created the accommodation for the Paleocene succession is expected to be related to processes in the mantle.

5.9 Age/time

The relative age of the units are given by their stratigraphic relationship. The aggradational P1 unit is oldest, followed by the progradational units P2, P3 and P4. The correlation with the biostratigraphic analysis of the Torsk Formation in well 7119/9-1 (chapter 5.5) were used to give an age estimate of the units. The age interval of depositions for each unit is displayed in table 3.

Table 3: Age estimate for the interpreted units.

Interpreted units	Age
Unit P4	Late Paleocene to Early Eocene
Unit P3	Late Paleocene
Unit P2	Middle Late Paleocene
Unit P1	Early Late Paleocene to middle Late Paleocene

The earliest sediments of the Torsk Formation in the Tromsø and Hammerfest basins were deposited in early Late Paleocene and the prograding system were mainly deposited during Late Paleocene. The exception is the uppermost prograding unit, P4, which have deposits from as young as earliest Early Eocene.

According to Nagy et al. (1997, 2000), the Cretaceous Kviting Formation, underlying the Paleocene units, is as young as Maastrichtian. Based on this the age for the Cretaceous – Paleogene hiatus, represented by Base Tertiary, are Maastrichtian – early Late Paleocene. There is no paleontological evidence of shallowing related to the Base Tertiary (Nagy et al., 1997), but it is expected that the hiatus represent by the Base Tertiary surface was related to uplift and erosion.

Well 7119/9-1 is situated in the westernmost part of the transitional area between the Hammerfest and Tromsø basins (figure 5.6), where the Paleocene sedimentary succession is aggradational. It is therefore assumed that the age of the thick aggradational unit in this area is the same as for the progradational units of the present study area, and that deposition happened simultaneously for the progradational and aggradational part of the same sedimentary system.

6. Conclusions

The seismic sequence analysis of the Paleocene succession in the Hammerfest and Tromsø basins show that the earliest infill was deposited as a sheet over the entire area. A downlap surface marked the change to a progradational infill of the area. Three progradational units were mapped out, and two of them, P2 and P3, was further subdivided based on the internal reflection pattern of the units. All three units prograded from ENE towards WSW.

The Loppa High was formed in Late Jurassic – Early Cretaceous, and has previous been expected to be a positive structure in Early Paleogene. However, this study has shown that Loppa High was a part of the depocenter and hence a part of the shallow Barents Shelf in Early Paleogene times. The Paleocene succession deposited over Loppa High, and further north and east, was eroded by younger phases of uplift and erosion. Some of the Paleocene succession is still preserved in the Hammerfest Basin, and this area is an important source for information about the Paleocene prograding system.

The interpreted Paleocene prograding units were part of a single prograding system, covering large parts of the Barents Shelf in Early Paleogene times. Deposition continued into Eocene and maximum burial occurred ca. 40 Ma. Eocene deposits of unknown thickness and extent was deposited over the Paleocene prograding system, but it is expected that the Paleocene succession made up the larger part of the Paleogene succession in the Hammerfest Basin and over Loppa High.

It is suggested that the sediment source for the prograding system in the SW Barents Sea was located on the northern Barents Shelf. This is mainly based on evidences of a hiatus in the ACEX well at the Lomonosov Ridge, which possibly represents uplift and erosion in the northernmost Barents Shelf during deposition of the prograding system in the SW Barents Sea. Also indications of a shallower environment related to Early Paleogene deposits in the Paleogene Central Basin on Spitsbergen are considered. It is suggested that the uplifted areas could be related to Cretaceous Large Igneous Provinces mainly located north of Svalbard and Franz Josef Land.

Based on biostratigraphy analysis of the Torsk Formation in the Tromsø Basin the deposition of the Paleocene succession is set to initiate in early Late Paleocene. The succession is mainly deposited in Late Paleocene time, with the exception of unit P4 which has deposits as young as earliest Early Eocene.

The Paleocene succession was deposited in a bathyal environment and the minimum estimate for the water depth is from 200-300 m. The water depth increased through Late Paleocene and reached maximum water depth during deposition of unit P4, after the maximum water depth there was a general shallowing during Early Eocene times.

Estimate of the original thickness of the Paleocene prograding system was calculated to be ca. 1800 m, but this turned out to be an overestimate. A significant thickness for the prograding system is still expected; hence a significant accommodation must have been created prior to Paleocene deposition. Since it is expected that the Barents Shelf was uplifted during the Cretaceous – Paleogene hiatus subsidence must have occurred over large parts of the Barents Shelf. The subsidence is expected to be related to mantel processes.

References

ANELL, I., MIDTKANDAL, I. & BRAATHEN, A. 2014. Trajectory analysis and inferences on geometric relationships of an Early Triassic prograding clinoform succession on the northern Barents Shelf. *Marine and Petroleum Geology*, 53, 167-179.

BACKMAN, J. & MORAN, K. 2009. Expanding the Cenozoic paleoceanograpic record in the Central Arctic Ocean: IODP Expedition 302 Synthesis. *Central European Journal of Geoscience*, 1, 157-175.

BERTRAM, G.T. & MILTON, N.J. 1996. Seismic Stratigraphy. *In*: EMERY, D. & MYERS, K.J. (eds.) *Sequence Stratigraphy. Blackwell Publishing*, 45-60.

CAVANAGH, A.J., DI PRIMO, R., SCHECK-WENDEROTH, M. & HORSFIELD, B. 2006. Severity and timing of Cenozoic exhumation in the southwestern Barents Sea. *Journal of the Geological Society,* 163, 761-774.

CORFU, F., POLTEAU, S., PLANKE, S., FALEIDE, J.I., SVENSEN, H., ZAYONCHECK, A. & STOLBOV, N. 2013. U-Pb geochronology of Cretaceous magmatism on Svalbard and Franz Josef Land, Barents Sea Large Igneous Province. *Geol. Mag.*, 150, 1127-1135.

DIMAKIS, P., BRAATHEN, B.I., FALEIDE, J.I., ELVERHØI, A. & GUDLAUGSSON, S.T. 1998. Cenozoic erosion and the preglacial uplift of the Svalbard – Barents Sea region. *Tectonphysics*, 300, 311-327.

DORÉ, A.G. 1995. Barents Sea Geology, Petroleum Resources and Commercial Potential. *Artic*, 48, 207-221.

DORÉ, A.G. & JENSEN, L.N. 1996. The impact of late Cenozoic uplift and erosion on hydrocarbon exploration: offshore Norway and some other uplifted basins. *Global and Planetary Change*, 12, 415-436.

FALEIDE,J.I., GUDLAUGSSON, S.T. & JACQUART, G. 1984. Evolution of the western Barents Sea. *Marine and Petroleum Geology*, 1, 123-150.

FALEIDE, J.I., VÅGNES, E. & GUDLAUGSSON, S.T. 1993. Late Mesozoic-Cenozoic evolution of the south-western Barents Sea in a regional rift-shear tectonic setting. *Marine and Petroleum Geology,* 10, 186-214.

FALEIDE, J.I., SOLHEIM, A., FIEDLER, A., HJELSTUEN, B.O., ANDERSEN, E.S. & VANNESTE, K. 1996. Late Cenozoic evolution of the western Barents Sea – Svalbard continental margin. *Global and Planetary Change*, 12, 53-74.

FALEIDE, J.I., TSIKALAS, F., BREIVIK, A.J., MJELDE, R., RITZMANN, O., ENGEN, Ø., WILSON, J. & ELDHOLM, O. 2008. Structure and evolution of the continental margin off Norway and the Barents Sea. *Episodes*, 31, 82-91.

FALEIDE, J.I., BJØRLYKKE, K. & GABRIELSEN, R.H. 2010. Geology of the Norwegian Continental Shelf. *In*: BJØRLYKKE, K. (ed.). *Petroleum Geoscience: From Sedimentary Environments to Rock Physics*. Springer-Verlag Berlin Heidelberg, 467-499.

GABRIELSEN, R.H. 1984. Long-lived fault zones and their influence on the tectonic development of the southwestern Barents Sea. *Journal of the Geological Society*, 141, 651-662.

GABRIELSEN, R.H., FÆRSETH, R.B., JENSEN, L.N., KALHEIM, J.E. & RIIS, F. 1990. Structural elements of the Norwegian continental shelf. Part 1: The Barents Sea Region. *Norwegian Petroleum Directorate Bulletin*, 6, 33.

GLØRSTAD-CLARK, E., FALEIDE, J.I., LUNDSCHIEN, B.A. & NYSTUEN, J.P. 2010. Triassic seismic sequence stratigraphy and paleogeography of the western Barents Sea area. *Marine and Petroleum Geology*, 27, 1448-1475.

GLØRSTAD-CLARK, E., BIRKELAND, E.P., NYSTUEN, J.P., FALEIDE, J.I. & MIDTKANDAL, I. 2011a. Triassic platform-margin deltas in the western Barents Sea. *Marine and Petroleum Geology*, 28, 1294-1314.

GLØRSTAD-CLARK, E., CLARK, S.A., FALEIDE, J.I., BJØRKSETT, S.S., GABRIELSEN, R.H. & NYSTUEN, J.P. 2011b. Basin dynamics of the Loppa High area, SW Barents Sea: A history of complex vertical movements in an epicontinental basin. *In*: GLØRSTAD-CLARK, E. (ed.). *Basin analysis in the western Barents Sea area: The interplay between accommodation space and depositional systems*. *Philosphiae Doctor Series of Dissertation, University of Oslo*, 111-180.

GREEN, P.F. & DUDDY, I.R. 2010. Synchronous exhumation events around the Artic including examples from Barents Sea and Alaska North Slope. *In:* VINING, B.A. & PICKERING, S.C. (*eds.*) *Petroleum Geology: From Mature Basins to New Frontiers – Proceedings of the 7th Petroleum Geology Conference, Geological Society of London*, 633-644.

GUDLAUGSSON, S.T., FALEIDE, J.I., JOHANSEN, S.E. & BREIVIK, A.J. 1998. Late Palaeozoic structural development of the South-western Barents Sea. *Marine and Petroleum Geology*, 15, 73-102.

HELLAND-HANSEN, W. & HAMPSON, G.J. 2009. Trajectory analysis: concepts and applications. *Basin Research*, 21, 454-483.

HENRIKSEN, E., BJØRNSETH, H.M., HALS, T.K., HEIDE, T., KIRYUKHINA, T., KLØVJAN, O.S., LARSSEN, G.B., RYSETH, A.E., RØNNING, K., SOLLID, K. & STOUPAKOVA, A. 2011. Uplift and erosion of the greater Barents Sea: impact on prospectivity and petroleum systems. *In:* SPENCER, A.M., EMBRY, A.F., GAUTIER, D.L., STOUPAKOVA, A.V. & SØRENSEN, K. (eds). *Artic Petroleum Geology. The Geological Society of London, Memoirs*, 35, 271-281.

JOHANNESSEN, E.P. & STEEL, R.J. 2005. Shelf-margin clinoforms and prediction of deepwater sands. *Basin Research*, 17, 521-550.

KNUTSEN, S.M. & VORREN, T.O. 1991. Early Cenozoic sedimentation in the Hammerfest Basin. *Marine Geology*, 101, 31-48.

KNUTSEN, S.M., SKJOLD, L.J. & SKOTT, P.H. 1992. Palaeocene and Eocene development of the Tromsø Basin – sedimentary response to rifting and early sea-floor spreading in the Barents Sea area. *Norsk Geologisk Tidsskrift*, 72, 191-207.

LANGINEN, A.E., LEBEDEVA-IVANOVA, N.N., GEE, D.G. & ZAMANSKY, Y.Y. 2009. Correlations between the Lomonosov Ridge, Marvin Spur and adjacent basins of the Arctic Ocean based on seismic data. *Tectonophysics*, 472, 309-322.

LARSSEN, G.B., ELVEBAKK, G., HENRIKSEN, L.B., KRISTENSEN, S.E., NILSSON, I., SAMUELSBERG, T.J., SVÅNÅ, T.A., STEMMERIK, L. & WORSLEY, D. 2005. Upper Palaeozoic lithostratigraphy of the southern part of the Norwegian Barents Sea. *Norges geologiske undersøkelse Bulletin,* 444, 3-45.

MAHER, H.D.J. 2001. Manifestation of the Cretaceous High Artic Large Igneous Province in Svalbard. *The Journal of Geology*, 109, 91-104.

MINAKOV, A., FALEIDE, J.I., GLEBOVSKY, V.Y. & MJELDE, R. 2012. Structure and evolution of the northern Barents-Kara Sea continental margin from integrated analysis of potential fields, bathymetry and sparse seismic data. *Geophysical Journal International*, 188, 79-102.

MITCHUM, R.M.J., VAIL, P.R. & THOMPSON, S. 1977a. Seismic Stratigraphy and Global Changes of Sea Level, Part 2: The Depositional Sequence as a Basic Unit for Stratigraphic Analysis. *In*: PAYTON, C. (ed.). *Seismic Stratigraphy: Application to Hydrocarbon Exploration*. *American Association of Petroleum Geologist Memoir*, 26, 53-62.

MITCHUM, R.M.J., VAIL, P.R. & SANGREE, J.B. 1977b. Seismic Stratigraphy and Global Changes of Sea Level, Part 6: Stratigraphic Interpretation of Seismic Reflection Pattern in Depositional Sequences. *In*: PAYTON, C. (ed.). *Seismic Stratigraphy: Application to Hydrocarbon Exploration. American Association of Petroleum Geologist Memoir*, 26, 117-133.

MÜLLER, R.D. & SPIELHAGEN, R.F. 1990. Evolution of the Central Tertiary Basin of Spitsbergen: towards a synthesis of sediment and plate tectonic history. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 80, 153-172.

MYERS, K.J. & MILTON, N.J. 1996. Concepts and Principles of Sequence Stratigraphy. *In:* EMERY, D. & MYERS, K.J. (eds.) *Sequence Stratigraphy*. *Blackwell Publishing*, 11-41.

NAGY, J., KAMINSKI, M.A., JOHNSEN, K. & MITLEHNER, A.G. 1997. Foraminiferal, palynomorph, and diatom biostratigraphy and paleoenvironments of the Torsk Formation: A reference section for the Paleocene-Eocene transition in the western Barents Sea. *In:* HASS, H.C.. & KAMINSKI, M.A. (eds.). *Contributions to the Micropaleontology and Paleoceanography of the Northern North Atlantic. Grzybowski Foundation Special Publication*, 5, 15-38.

NAGY, J., KAMINSKI, M.A., KUHNT, W. & BREMER, M.A. 2000. Agglutinated Foraminifera from Neritic to Bathyal Facies in the Palaeogene of Spitsbergen and the Barents Sea. *In:* HART, M.B., KAMINSKI, M.A. & SMART, C.W. (eds.). *Proceedings of the Fifth International Workshop on Agglutinated Foraminifera. Grzybowski Foundation Special Publication*, 7, 333-361.

NPD. 2015. FactMaps [Online]. Available: http://npd.no/en/Maps/Fact-maps/

NPD. 2015b. FactPages [online]. Available: http://factpages.npd.no/factpages/Default.aspx?culture=no

NYSTUEN, J.P. 1998. History and development of sequence stratigraphy. *In*: GRADSTEIN, F.M., SANDVIK, K.O. & MILTON, N.J. (eds.). *Sequence Stratigraphy – Concepts and Applications. Norwegian Petroleum Society, Special Publication No.8*, 31-116.

NØTTVEDT, A. 1985. Askeladden Delta Sequence (Palaeocene) on Spitsbergen – sedimentation and controls on delta formation. Polar Research, 3, 21-48.

NØTTVEDT, A., CECCHI, M., GJELBERG, J.G., KRISTENSEN, S.E., LØNØY, A., RASMUSSEN, A., RASMUSSEN, E., SKOTT, P.H. & VAN VEEN, P.M. 1993.Svalbard – Barents Sea correlation: a short review. *In:* VORREN, T.O., BERGSAGER, E., DAHL-STAMNES, Ø.A, HOLTER,E., JOHANSEN, B., LIE, E. & LUND, T.B. (eds.). *Artic Geology and Petroleum Potentional*. Amsterdam: Elsevier Science Publisher B.V , 363-375.

OHM, S.E., KARLSEN, D.A. & AUSTIN, T.J.F. 2008. Geochemically driven exploration models in uplifted areas: Examples from the Norwegian Barents Sea. *AAPG Bulletin*, 92, 1191-1223.

RIIS, F. & FJELDSKAAR, W. 1992. On the magnitude of the Late Tertiary and Quaternary erosion and its significance for the uplift of Scandinavia and the Barents Sea. *In:* LARSEN, R.M., BREKKE, H., LARSEN, B.T. & TALLERAAS, E. (eds.). *Structural and Tectonic Modelling and its Application to Petroleum Geology. NPF Special Publication*, 1, 163-185.

SANGREE, J.B. & WIDMIER, J.M. 1978. Seismic Stratigraphy and Global Changes of Sea Level, Part 9: Seismic Interpretation of Clastic Depositional Facies. *The American Association of Petroleum Geologists Bulletin*, 62, 752-771.

SMELROR, M., PETROV, O.V., LARSSEN, G.B. & WERNER, S. 2009. ATLAS: Geological History of the Barents Sea. *Norges Geologiske Undersøkelse*, 135.

STEEL, R., GJELBERG, J., HELLAND-HANSEN, W., KLEINSPEHN, K., NØTTVEDT, A. & RYE-LARSEN, M. 1985. The Tertiary strike-slip basins and orogenic belt of Spitsbergen. *The Society of Economic Paleontologists and Mineralogists*, 37, 339-359.

VORREN, T.O., RICHARDSEN, G. & KNUTSEN, S.M. 1991. Cenozoic erosion and sedimentation in the western Barents Sea. *Marine and Petroleum Geology*, 8, 317-340.

WORSLEY, D., JOHANSEN, R. & KRISTENSEN, S.E. 1988. The Mesozoic and Cenozoic succession of Tromsøflaket. *In*: DALLAND, A. WORSLEY, D. & OFSTAD, K. (eds.). *A lithostratigraphic scheme for the Mesozoic and Cenozoic succession offshore mid- and northern Norway. Norwegian Petroleum Directorate Bulletin*, 4, 42-65.

WORSLEY, D. 2008. The post – Caledonian development of Svalbard and the western Barents Sea. *Polar Research*, 27, 298-317.

References