Geological characterisation of shallow marine-to-deltaic sandstone reservoir targets: Krossfjord and Fensfjord formations, Troll Field, Norwegian North Sea

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I, Nicholas Edward Holgate, hereby declare that the work contain herein is my own and that all else is appropriately referenced.

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ABSTRACT

The sedimentological character and stratigraphic architecture of shallow-marine reservoirs are strongly controlled by the interplay of physical processes that occur at and near the shoreline (e.g. wave- vs. tide- vs. fluvial-dominated). These aspects can be further complicated by the interplay of tectonics in rift basins through fault block rotation, uplift, and subsidence. This thesis presents a subsurface case study from the Middle-to-Upper Jurassic “syn-rift” Krossfjord and Fensfjord formations, Horda Platform, offshore western Norway. The distribution, geometry, and connectivity of these sandbodies are poorly understood, as they have not been the focus of previous work. However, the formations form a significant oil and gas reservoir in the Troll and Brage fields, and a prospective reservoir in the Gjøa Field.

Analysis of core and wireline-log data from the Krossfjord and Fensfjord formations identified wave- and tide-dominated deltaic, shoreline and shelf depositional environments. The integration of biostratigraphic data enabled subdivision of the formations into ‘series’ bound by maximum flooding surfaces. The integration of 3D seismic data defined the gross stratigraphic architecture, specifically the stacking patterns of clinoform sets, and enabled further subdivision of the ‘series’. Seismic geomorphological analysis of clinoforms, calibrated using forward seismic models of outcrop analogues, aided interpretation of the shoreline process regime (e.g. relative influence of waves, tides and river-mouth processes) in the context of shoreline trajectories. Palaeogeographic reconstructions illustrate that a subaqueous delta was located over the Troll Field fronting a wave- and current-driven southerly-directed spit during Middle to Late Jurassic times. In conclusion, a robust understanding of the Krossfjord and Fensfjord formations is established in order to drive future exploration in these, and coeval, reservoirs. In addition, the novel forward seismic modelling methodology described herein has wide applications and the results are directly applicable to many other shallow-marine reservoir sandstones, for which the outcrops studied are considered to be sedimentological analogues.
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CHAPTER 1

Introduction

1.1 Rationale

Siliciclastic deltaic-to-shallow marine deposits form a significant proportion of petroleum reservoirs around the world (e.g. North Sea, Brunei, Nigeria, etc.). The reconstruction of depositional environments in shallow-marine clastic petroleum reservoirs is essential to field appraisal, development and production (Johnson and Stewart, 1985). Specifically, studies have shown how detailed knowledge of heterogeneity within shallow-marine sandstones can improve production (e.g. Jackson et al., 2009; Morad et al., 2010) and how knowledge of the shallow-marine domain can help predict where reservoir sediments may exist elsewhere on the shelf or in the deep marine domain (e.g. Bullimore et al., 2005; Hampson et al., 2009; Dixon et al., 2012). However, the sedimentological character, distribution and stratigraphic architecture of shallow-marine deposits are strongly controlled by physical processes at and near the shoreline (e.g. wave- vs. tide- vs. fluvial-dominated) complicating the ability to accurately reconstruct depositional environments. These aspects can be further confounded by the interplay of tectonics in rift basins through fault block rotation, uplift, and subsidence.

An example of a siliciclastic deltaic-to-shallow marine depositional system comes from the Middle-to-Upper Jurassic Krossfjord and Fensfjord formations of offshore Norway. These shallow-marine sandstone formations, belonging to the Viking Group, are situated below the Draupne Formation and above the Brent Group on the Horda Platform (eastern margin of the Viking Graben, northern North Sea) and overall they thin and pinch out into offshore shales of the Heather Formation towards the graben. Both formations may be classified as ‘syn-rift’ as they were deposited during the Middle-to-Late Jurassic rift event (Ravnås and Bondevik, 1997).
They are poorly understood as they have not been the focus of previous work, but they form a prospective reservoir interval in the area around the existing Troll, Brage and Gjøa fields. This study focuses on the Krossfjord and Fensfjord formations in the area of the super-giant Troll oil and gas field.

1.2 Aim

The aim of this thesis is to characterise the sedimentology and stratigraphic architecture of the Krossfjord and Fensfjord formations in the Troll Field to provide insights into the distribution and character of the reservoir sandstones within the Troll Field and elsewhere on the Horda Platform. An integrated dataset comprising core, wireline-log, biostratigraphic, 2D and 3D seismic reflection data combined with outcrop observations are used to fulfil the following objectives.

1.3 Objectives

The objectives of this thesis can be split into the following two categories:

1.3.1 Sedimentology and sequence stratigraphy of the Krossfjord and Fensfjord formations, Troll Field

- Create facies and facies association schemes using core dataset to characterise the sedimentology.
- Use biostratigraphic dataset to identify key chronostratigraphic surfaces within the formations to establish a sequence stratigraphic framework for the study area.
1.3.2 Seismic stratigraphic and clinoform architecture of the Krossfjord and Fensfjord formations, Troll Field area

- Identify and trace key sequence stratigraphic surfaces across Troll Field study area in both 3D and 2D seismic datasets.
- Examine any fault-related thickening which could provide evidence of syn-depositional fault movement during the Middle-to-Late Jurassic.
- Use outcrop analogues to characterise the seismic response of clinoformal architectures under reservoir conditions using forward seismic modelling.
- Use seismically-imaged clinoform architectures within the Troll Field seismic reflection dataset to characterise the temporal evolution of the Krossfjord and Fensfjord formations.
- Reconstruct the palaeogeographical evolution of the Troll Field area during the Middle-to-Late Jurassic utilising evidence collected during this study.

1.4 Thesis outline

This thesis comprises six chapters. Chapters 1 to 2 form the introduction and background of this study whilst chapters 3 to 5 fulfil the objectives of this project.

Chapter 2 comprises a review of literature pertinent to this study, focussing on the following key themes: (1) the tectono-stratigraphic evolution of the northern North Sea during the Middle-to-Late Jurassic rift event; (2) seismic stratigraphic techniques; and (3) forward seismic modelling.

Chapter 3 documents the sedimentological and sequence stratigraphic character of the Krossfjord and Fensfjord formations in the Troll Field. The key sedimentological features of the
formations are discussed and depositional models are presented to explain the complex distribution of depositional environments identified.

Chapter 4 combines the geometry, distribution and lithological character of clinoform-bearing outcrop analogues to the Krossfjord and Fensfjord formations with subsurface data from the Troll Field highlighted in Chapter 3, to create a suite of forward seismic models and provide calibration of seismic data from the Troll Field.

Chapter 5 uses the stratigraphic scheme established in Chapter 3 to identify the seismic stratigraphic architecture of the Krossfjord and Fensfjord formations. Reservoir characterisation is achieved through the analysis of seismically-imaged clinoforms assisted by the forward seismic modelling work conducted in Chapter 4. Observations in seismic reflection data are used to provide further explanation for features of the Krossfjord and Fensfjord formations highlighted in Chapter 3, as well as to refine the depositional models presented.

Chapter 6 concludes the thesis by summarising the main findings in the context of the project aim and objectives, and provides suggestions for future research.

1.5 Publication status

Whilst the previous section described the links between chapters, chapters 3, 4 and 5 are presented as stand-alone papers within this thesis. The author of this thesis is also the primary author for the three papers. Important contributions to these papers are provided by the co-authors and their role in each paper is described below along with the publication status.

Paper 1: Sedimentology and sequence stratigraphy of the Middle-Upper Jurassic Krossfjord and Fensfjord formations, Troll Field, northern North Sea
Nicholas E. Holgate, Christopher A.-L. Jackson, Gary J. Hampson and Tom Dreyer
Published in Petroleum Geoscience 2013, v.19; p237-258
The primary author analysed and interpreted core, wireline-log and biostratigraphic data to characterise facies, create facies associations, create a stratigraphic framework and develop depositional models. The co-authors critically appraised interpretations, contributed to ideas and reviewed the manuscript prior to submission.

**Paper 2:** Constraining uncertainty in interpretation of seismically imaged clinoforms in deltaic reservoirs, Troll Field, Norwegian North Sea: Insights from forward seismic models of outcrop analogues
Nicholas E. Holgate, Gary J. Hampson, Christopher A-L. Jackson and Steen A. Petersen
*In press at AAPG Bulletin*

The primary author conducted fieldwork, created seismic models and presented interpretations. The co-authors aided software application, critically appraised interpretations and reviewed the manuscript prior to submission.

**Paper 3:** Seismic stratigraphic analysis of the Middle-Upper Jurassic Krossfjord and Fensfjord formations, Troll oil and gas field, northern North Sea
Nicholas E. Holgate, Christopher A-L. Jackson, Gary J. Hampson and Tom Dreyer
*In review at Marine and Petroleum Geology*

The primary author interpreted and analysed seismic data to develop depositional models. The co-authors critically appraised interpretations and reviewed the manuscript prior to submission.
2.1 Regional geological framework

2.1.1 The Northern North Sea Basin

The North Sea Basin is a 170-200 km wide, 1000 km long, fault-bounded basin with mainland Norway in the east and the Shetland Platform in the west. It comprises three arms of a rift system which spread simultaneously to create the Viking Graben (with which this study is concerned), Central Graben and Moray Firth (Fig. 2.1) (Roberts et al., 1990; Rattey and Hayward, 1993). Together, the rift system represents part of the failed Mesozoic, Arctic-North Atlantic rift system (Coward et al., 2003).

The North Sea rift system was created through multiple rift events during the Permo-Triassic and the Middle to Late Jurassic. During the Permo-Triassic, extension was triggered by the break-up of the Pangean supercontinent reworking Caledonian and Variscan structures and was oriented E-W resulting in a number of N-S trending fault-bounded half grabens (Fig. 2.1a) (Ziegler, 1990a; Coward et al., 2003). During the period between the Permo-Triassic and Middle to Late Jurassic, the area was characterised by tectonic quiescence and post-rift thermal subsidence (Steel, 1993; Færseth, 1996; Færseth and Ravnås, 1998; McLeod et al., 2000). The presence of a mantle hot spot during the Aalenian to Bajocian period caused uplift creating a domal structure centred on the triple junction of the North Sea rift system (Underhill and Partington, 1993). The doming event created a source of clastic material for the Brent Group delta in the area of the Viking Graben (Richards, 1992).
The Middle to Late Jurassic rift event occurred broadly at the same time as the collapse of the thermal dome and was diachronous in both initiation and cessation across the North Sea (Fig. 2.1b, c) (Rattey and Hayward, 1993). The rift event initiated as early as the middle Bajocian, highlighted by the first fault-block rotation events (Helland-Hansen et al., 1992; Færseth and Ravnås, 1998). The diachronity in the cessation of rifting is recognised by the translation of the triple junction of the North Sea rift system from the central North Sea, during the Callovian, to the southern Møre Basin, during Volgian times, as a result of the opening of the Arctic-North Atlantic rift system (Ziegler, 1990a; Coward et al., 2003). The North Sea rift event is commonly split into two main phases: Oxfordian to early Kimmeridgian extension (Fig. 2.1b) and late Kimmeridgian to Early Cretaceous extension (Fig. 2.1c). The rift event was most widespread during a period of 10 Ma and was controlled by the pre-existing lineaments of the north-south Precambrian trend, the north-easterly Caledonide trend, and the east-south-easterly Tornquist trend (Fraser et al., 2002). The rifting is characterised as being oblique due to the NW-SE extension reactivating faults established during the N-S rifting in the Triassic (Færseth et al., 1997). Within the Viking Graben itself, a number of rift phases separated by periods of relative tectonic quiescence have been identified during the Middle to Late Jurassic rift event. Rifting formed a series of tilted half grabens east and west of the Viking Graben (Nøttvedt et al., 2000; Husmo et al., 2002). On the Norwegian side, a number of tilted half grabens occur from the relatively stable Horda Platform, with which this study is concerned, in the east to the Viking Graben in the west (Fig. 2.1b, c) (Stewart et al., 1995; Ravnås and Bondevik, 1997). The most extensive phases of rifting occurred towards the Late Jurassic resulting in the uplift and eastward tilting of normal fault blocks (Fossen et al., 2003). The diachronous cessation of rifting occurred during the Early Cretaceous with the onset of passive thermal subsidence resulting in the syn-rift topography being covered by transgressive sediments above a composite surface that is often referred to as the Base Cretaceous Unconformity (BCU) (Badley et al., 1984; Coward et al., 2003; Kyrkjebø et al., 2004).
2.1.2 Tectonic-stratigraphic evolution of the Horda Platform

The Horda Platform is a structural high located between the easterly-rotated fault blocks stepping down to the Viking Graben in the west and the uplifted Norwegian hinterland in the east, across the Øygarden Fault Complex (Fig. 2.1). During the Permo-Triassic rift event, the northern part of the Horda Platform was characterised by three active, easterly-tilted half grabens recognised through syn-rift sedimentation at the time (Fig. 2.1a) (Badley et al., 1988; Zanella and Coward, 2003). During the Middle to Late Jurassic rift event it was relatively stable with 5% stretching during the Late Jurassic compared to 30-40% within the Viking Graben (Zanella and Coward, 2003). The Troll Field, with which this study is concerned, is located in the northern tip of the Horda Platform (Fig. 2.1).

There is a complex interaction between tectonics and sedimentation in this area. Steel (1993) described the Triassic to Jurassic stratigraphy as being composed of nine ‘megasequences’ that represent individual regressive-to-transgressive, alluvial and marine clastic wedges building out from both Norwegian and Scottish hinterlands. Each megasequence is a time-stratigraphic unit of 6-18 Ma duration and its formation was governed by both eustatic sea level changes and tectonic subsidence-rate variations.

Focussing on Jurassic stratigraphy, a number of fluvio-deltaic-to-shallow marine successions including the Dunlin and Brent groups characterised the period of post-rift subsidence during uplift of the North Sea Dome in the Early to early Middle Jurassic (Fig. 2.2a). The extensively explored Brent Group (corresponding to Megasequence 7 of Steel, 1993) in particular is composed of the Broom, Rannoch, Etive and Lower Ness formations which prograded north supplied by material from the uplifted dome. The upper part of the Brent Group is formed of the retrogressive Upper Ness and Tarbert formations which retreated south in response to the diachronous onset of rifting in the late Bajocian, which deepened the basin causing a marine
incursion (Livera, 1989; Bjorlykke et al., 1992; Richards, 1992; Wehr and Brasher, 1996; Davies et al., 2000; Nøttvedt et al., 2000; Husmo et al., 2002). In total the Brent Group is c. 200 m thick, early Bajocian (MFS J24) to early Bathonian (MFS J32) in age, and encapsulates c. 6 Ma of deposition (Vollset and Doré, 1984; Richards, 1992; Partington et al., 1993; Husmo et al., 2002).

The Viking Group was deposited during the Middle to Late Jurassic rift event on the Horda Platform and is characterised by a stacked series of shallow marine-to-deltaic sandstones. It is composed of the Krossfjord and Fensfjord formations contained within Megasequence 8, with which this study is concerned, and the Sognefjord Formation contained within Megasequence 9 (Steel, 1993). Together, the three sandstone formations were sourced from the uplifted Norwegian hinterland in the east and deposited towards the subsiding Viking Graben to the west (Husmo et al., 2002) (Figs. 2.2b-e, 2.3). The sandstone formations built out during periods of tectonic quiescence, and extend across the Horda Platform and into the westwards subbasins where they interfinger with marine shales of the Heather Formation (Figs. 2.3, 2.4) (Ravnås et al., 2000). The landward incursion of the Heather Formation occurred as a series of transgressive events controlled by the basin-wide phases of fault-driven subsidence related to the Middle to Late Jurassic rift event (Fig. 2.3). Consequently, each sandstone formation of the Viking Group is bounded by a phase of tectonic activity and has a regressive-to-transgressive architecture (Steel, 1993; Færseth and Ravnås, 1998; Nøttvedt et al., 2000). In this context, the Sognefjord Formation is split into two genetic sequences separated by a maximum flooding surface associated with active rifting in the middle Oxfordian (Dreyer et al., 2005). Flooding of the North Sea Basin occurred during the late Kimmeridgian (Fig. 2.2e) and led to the widespread deposition of the Draupne Formation, a deep marine mudstone, on the Horda Platform. This in turn was capped by the Base Cretaceous Unconformity. In total, the Viking Group is c. 200 m thick and was deposited from the early Bathonian (MFS J32) to middle Ryazanian (MFS J75) (Vollset and Doré, 1984; Fraser et al., 2002).
2.1.3 Equivalent formations to the Krossfjord and Fensfjord formations during the Bathonian-Callovian in the North Sea

The Krossfjord and Fensfjord formations are time-equivalent to formations throughout the North Sea which broadly demonstrate the marine incursion due to gradual opening of the North Sea rift system. During this time, the climate in the North Sea area was relatively stable but slowly changing due to the opening connection with the northern polar ocean, which resulted in an alteration from warm humid conditions in the Late Jurassic to arid cold climates in the Early Cretaceous (Mutterlose et al., 2003). The stratigraphy in the North Sea is extremely complex with regional differences in facies development and age. This is due to a combination of the pre-existing Triassic rift topography, Mid-Jurassic doming and erosion, the Middle to Late Jurassic fault-block uplift, rotation and erosion, and local influence of movement of Permian salt in response to evolving Jurassic structures (Husmo et al., 2002).

Equivalent to the Krossfjord and Fensfjord formations on the eastern side of the North Viking Graben, in the East Shetland Basin, is the relatively thin Emerald Formation of the Humber Group which is interpreted to be a transgressive sand sheet deposited in nearshore and offshore locations (labelled 1 in Fig. 2.2b) (Johnson et al., 2005). This lies above and in more basin-marginal positions relative to the Middle Jurassic marine sandstone of the Tarbert Formation, indicating overall transgression (Stewart and Faulkner, 1991). Tectonic activity is interpreted to be absent on the shelf as indicated by the consistent thickness of the Emerald Formation. Similarly to the Krossfjord and Fensfjord formations, the Emerald Formation is overlain by the Heather and Kimmeridge Clay (Draupne equivalent) formations (Richards et al., 1993). A significant period of rift-related faulting is interpreted to have occurred in the Late Jurassic in this area (Stewart and Faulkner, 1991).
Equivalent to the Krossfjord and Fensfjord formations in the South Viking Graben are the Sleipner and Hugin formations (labelled 2 in Fig. 2.2b) (Husmo et al., 2002). The formations form part of the Vestland Group and consist of sandstones, siltstones and mudstones with extensive coal layers. The Hugin Formation represents shallow-marine and marginal-marine depositional environments which interfinger with continental deposits of the underlying Sleipner Formation and offshore marine shales of the Heather Formation above (Husmo et al., 2002; Kieft et al., 2010). These formations retreated southwards as part of the Brent Delta during deflation of the thermal dome, onset of rift-related faulting and the incursion of the Boreal Ocean (Helland-Hansen et al., 1992; Fraser et al., 2002). As a result, within the same basin the Krossfjord and Fensfjord formations in the north are prograding and the Sleipner and Hugin formations in the south are retrograding contemporaneously. This indicates that tectonics and sediment supply were the key controls on stratigraphic architecture in the Viking Graben at this time and not eustatic sea level (Husmo et al., 2002).

Further south, in the central and east provinces of the North Sea, a series of discrete, low-relief sub-basins existed, related to the gradual rifting of the Viking and Central grabens (Fig. 2.1b, c). The transgression, which influenced the retreat of the Brent Delta, continued from north to south during the late Callovian via the Utsira High. In the southern Norwegian sector, an upwards change from non-marine to marine deposition, through the Bryne and Sandnes formations of the Vestland Group and the Haugesund and Egersund formations of the Tyne and Boknefjord groups respectively, occurred due to overall transgression (labelled 3 in Fig. 2.2b). In the UK sector, the Pentland Formation, characterised by marginal-marine coastal-plain deposits, was later overlain by the shallow-marine deposits of the Fulmar Formation (labelled 4 in Fig. 2.2b). In the Danish sector, fluvial-channel sandstones of the Bryne Formation occur in the south and estuarine sandstones occur in the north (labelled 5 in Fig. 2.2b). Further west, the evolution from the coastal plain Brora Coal to the nearshore sandstone of the Beatrice Formation (labelled 6 in Fig. 2.2b) occurred due to tectonically-enhanced transgression from
the west, related to the opening of the Moray Firth Basin along the Caledonian Great Glen fault system as part of the trilete rift development (Fig. 2.1b, c) (Richards et al., 1993; Fraser et al., 2002; Husmo et al., 2002; NPD, 2013a).

Overall, sedimentation in the North Sea during the Middle to Late Jurassic appears to have been characterised by proximal depositional environments being overlain by more distal depositional environments indicating an overall transgression (e.g. Sneider et al., 1995). This was the result of active rifting opening up the North Sea area to a marine incursion from the Boreal Sea in the north and the Tethys Ocean in the south, which was previously restricted by domal uplift during the Early Jurassic (Coward et al., 2003).

### 2.1.4 Previous depositional models for the Viking Group, Troll Field

The Fensfjord and Sognefjord formations form the producing reservoir in the Troll, Brage, Fram and Gjøa fields (NPD, 2013a). The Sognefjord Formation is the primary reservoir and therefore has been the focus of previous studies. Early core- and 2D seismic-based studies on the Troll Field recognised the Sognefjord Formation as being composed of shallow-marine sandstones characterised by wave and tide-influence (Whitaker, 1984; Hellem et al., 1986). These were interpreted to be arranged with a high-energy coastal zone in the west of the Troll Field and a tidal backbasin in the east. Westward pinch-out of successive sandstone units was recognised with interpreted offshore bar deposits fronting interpreted delta-front, estuarine and barrier bar environments (Whitaker, 1984; Hellem et al., 1986; Osborne and Evans, 1987). The control on deposition was interpreted to be eustatic sea level. Following the acquisition of 3D seismic data, it was recognised that the Sognefjord Formation had a significant north-south orientation in the Troll Field, characterised by southward prograding clinoforms which were interpreted to represent a spit with a deltaic source in the north (Stewart et al., 1995). It was further postulated that the lensoid geometry of the formation was a result of local tectonic elements
focussing deposition along north-south trending faults. This work, as well as Steel (1993), suggested a much greater tectonic control over deposition in the Troll Field area than previously thought. More recently, core and seismic analysis in the Troll Field has further supported the interpretation of a spit in a high-energy coastal zone in the west and a tidal backbasin in the east (Fig. 2.5) (Dreyer et al., 2005). Most recently, detailed core and seismic analysis of clinoforms within the lower Sognefjord Formation on the Troll Field has led to the interpretation of a sand-prone subaqueous delta over the Troll Field area without a spit (Patruno et al., 2013). To date, no depositional models have been provided for the Krossfjord and Fensfjord formations in the Troll Field.

2.1.5 Petroleum geology of the Viking Group, Troll Field

The Troll Field is a super-giant oil and gas field that produces from Middle to Upper Jurassic reservoir sandstones of the Fensfjord and Sognefjord formations. The Sognefjord Formation is the primary reservoir with an oil column of 22-26 metres in the Troll West fault block and the Fensfjord Formation is the secondary reservoir with an oil column of 6-9 metres in the Troll East fault block (Fig. 2.6). Both oil columns have significant gas caps (Høye et al., 1994; Johnsen et al., 1995; NPD, 2013a). The laterally-extensive reservoir sandstones are shallow marine in origin and are regionally unique in that their burial history is limited to 1000 m sub-sea providing good reservoir quality with high porosity (<34%) and permeability (<10 Darcies) (Gray, 1987). The three relatively large traps are rotated fault-block structures where fault seal occurs through juxtaposition of reservoir and non-reservoir rocks (Fig. 2.6) (Johnsen et al., 1995; Fraser et al., 2002; NPD, 2013a). The source is provided by the overlying, regionally-extensive, organic-rich Draupne Formation (Kimmeridge Clay equivalent) that is mature to overmature (Fig. 2.3) (Gormly et al., 1994). The accumulation of gas may be linked to the overmaturity of the Draupne Formation (Fraser et al., 2002).
The Middle to Upper Jurassic reservoir sandstones of the North Sea have been intensively explored over the last c. 40 years with the discovery of major structural plays considered to be complete (Fraser et al., 2002). However, a host of authors suggest that the North Sea still maintains huge exploration potential (Johnson and Stewart, 1985; Østvedt et al., 2005), albeit with concerns about maturation (Sørensen and Tangen, 1995). Furthermore, Johnson and Fisher (1998) consider 75% of undiscovered syn-rift reserves exist within the Upper Jurassic deep-marine and shallow-marine sediments held within subtle plays. Therefore a detailed understanding of shallow-marine environments, such as those in the Troll Field during the Middle to Late Jurassic, may aid the prediction of the character, distribution and connectivity of shallow-marine and deep-marine reservoir sandbodies (e.g. Johannessen and Steel, 2005) elsewhere on the Horda Platform and towards the Viking Graben.

2.2 Sequence stratigraphy and shallow-marine successions

2.2.1 Sequence stratigraphic framework – Stratal surfaces and relationships

Sequence stratigraphy attempts to describes the genetic relationship between facies and stratal surfaces within a chronostratigraphic framework (Van Wagoner et al., 1990; Catuneanu et al., 2009). Sequence stratigraphic analysis of successions can therefore not only describe individual sedimentary successions and their genetic relationships within sedimentary basin fills but also aid the prediction and distribution of genetically related sedimentary successions on both local and regional scales (Catuneanu et al., 2009).

Sequences are described as genetically related strata bound by unconformities or their correlative conformities (Mitchum, 1977; Mitchum et al., 1977b; Van Wagoner et al., 1990; Abreu et al., 2010). The building blocks of sequences are parasequences which are conformable, genetically related successions of beds, bedsets, laminasets and laminae, bound by marine
flooding surfaces (Fig. 2.7) (Van Wagoner et al., 1990). A transgression occurs when accommodation space creation exceeds sediment supply and is commonly identified in the rock record of shoreface sediments (Coe, 2003). A flooding surface at the top of a parasequence demonstrates an abrupt increase in water depth (transgression) and is commonly associated with submarine erosion (ravinement), which may be expressed as a transgressive lag, or lack of deposition (Fig. 2.7). The ravinement surface may either be created by wave action during shoreface retreat (‘wave ravinement surface’ sensu Swift, 1968), during which a veneer of surficial sediment may be removed over large areas of the shelf, or by tidal action (‘tidal ravinement surface’ sensu Swift, 1968), where erosion occurs through tidal currents and may extend significant distances into the coastal plain through tidal channels (Cattaneo and Steel, 2003).

Successions of genetically related parasequences form parasequence sets and these can be bounded by maximum flooding surfaces, transgressive surfaces and sequence boundaries. Within the bounding surfaces described above, the stacking patterns of genetically related parasequences record the interplay between sediment supply and sea level change which control whether the shoreline progrades basinward (progradation) or retrogrades landwards (retrogradation) (Van Wagoner et al., 1990). These stacking patterns can be described by systems tracts (lowstand (LST); transgressive (TST); and highstand (HST)) which are stacked to form a depositional sequence that records one cycle of relative sea level (Fig. 2.7d-e) (Van Wagoner et al., 1990).

The highstand systems tract is bounded by the sequence boundary above and maximum flooding surface below (Fig. 2.7b, d) and is typically composed of an aggradational-progradational-degradational parasequence set (Fig. 2.7c) (Van Wagoner et al., 1990). It forms in response to sediment supply outpacing sea level rise resulting in the progradation of the shoreline (Fig. 2.7e). As a reduction in the rate of sea level rise occurs, sediment supply further
outpaces accommodation space creation causing a rapid progradation and aggradation of parasequence sets as incised river valleys form to transport sediment seaward where accommodation space is available (Fig. 2.7c, e). The resultant lowstand systems tract is bounded by a sequence boundary at the base, indicating the extent of subaerial exposure and erosion. In seismic reflection data, a sequence boundary is identified through the onlap termination of reflections above the sequence boundary and toplap and/or truncation of reflections beneath the sequence boundary (Fig. 2.7a).

The lowstand systems tract is capped by a transgressive surface, indicating the point at which sea level rise outpaces sediment supply (Fig. 2.7b, d) (Van Wagoner et al., 1990). When sea level rise outpaces sediment supply the transgressive systems tract forms, represented by a retrogradation of parasequences (Fig. 2.7c) and the landward migration of shoreline facies (Fig. 2.7e). Eventually the rate of sea level rise slows and becomes outpaced by sediment supply again. The surface across which the turnaround from transgressive systems tract to highstand systems tract occurs is termed the maximum flooding surface (Van Wagoner et al., 1990). Maximum flooding surfaces demonstrate the most landward position of the shoreline and the consequent significant decrease in sedimentation results in development of a condensed section and/or submarine unconformity (Coe, 2003). In seismic reflection data, a maximum flooding surface is identified through the recognition of downlap reflections onto the surface (Fig. 2.7a).

### 2.2.2 Sequence stratigraphic framework – Shoreline trajectories

The Krossfjord and Fensfjord formations are termed “syn-rift” as they were deposited during the Middle to Late Jurassic rift event in the northern North Sea (Steel, 1993; Stewart et al., 1995). Sedimentation during extensional tectonic events can be affected by fault growth, linkage and death controlling sediment input, distribution and accommodation. As these factors can vary locally, the evolving rift basin architecture controls the spatial and temporal distribution of
depositional environments (Gawthorpe et al., 1994; Gawthorpe and Leeder, 2000; McLeod et al., 2002). Therefore, tectonics, rather than sea level change, is the main control on the stacking patterns and geometries of sediments on a local scale (Prosser, 1993; Gawthorpe et al., 1994). Conventional applications of systems tract nomenclature assumes a regional, sinusoidal relative sea level curve, which is not considered suitable for this study given the high potential for localised variation in sediment supply and sea level during the Middle to Late Jurassic rift event. Instead, trajectory analysis, which takes into account the combined effect of sediment supply and relative sea level and can be applied descriptively at a variety of scales, is considered more appropriate for this study (e.g. Hampson et al., 2009).

Trajectory analysis studies the migration of sedimentary environments through time using geomorphological breaks-in-slope that are associated with changes in depositional processes (Helland-Hansen and Hampson, 2009). It is considered by Helland-Hansen and Hampson (2009) to have a number of advantages over a sequence stratigraphic interpretation using systems tracts including: (1) each change in shoreline migration direction is considered to be part of a continuously evolving system rather than a number of discrete systems tracts; (2) subtle changes in depositional response (e.g. within systems tracts) can be identified; (3) trajectory analysis does not anticipate the succession of depositional events implied by systems tract models; and (4) the descriptive emphasis of trajectory analysis does not involve any a priori assumptions about the type or nature of the mechanisms that drive sequence development.

Whilst trajectory analysis can occur on a variety of scales, ‘clinoforms’ (sensu Rich, 1951) are typically used to analyse shallow-marine sequences. Clinoforms (discussed in detail below) record the morphology of shorelines and shelves and their geometry can record facies belt boundaries through their breaks-in-slope (Mitchum et al., 1977a; Steel and Olsen, 2002;
Johannessen and Steel, 2005; Howell et al., 2008b; Hampson et al., 2009; Helland-Hansen and Hampson, 2009; Henriksen et al., 2009).

Four distinctive shoreline trajectory categories can be described as: (1) ascending regressive ('normal regression' of Posamentier et al., 1992b); (2) descending regressive ('forced regression' of Posamentier et al., 1992b); (3) transgressive; and (4) stationary (Fig. 2.8) (Helland-Hansen and Gjelberg, 1994; Helland-Hansen and Martinsen, 1996; Helland-Hansen and Hampson, 2009). A further subdivision of 'accretionary' and 'nonaccretionary' can be applied to both descending regressive and transgressive categories where the trajectory is either dependent or independent of sediment supply respectively (Helland-Hansen and Gjelberg, 1994).

Ascending regressive trajectories or 'normal regression' (Fig. 2.8) occurs as a result of sediment supply being greater than accommodation development leading to a relatively large number of thick parasequences being created, absence of laterally extensive erosional surfaces and a high preservation of shoreline deposits (Helland-Hansen and Martinsen, 1996; Helland-Hansen and Hampson, 2009). Stationary trajectories record shoreline position at the shelf edge and where sediment is bypassed to the basin floor (Helland-Hansen and Hampson, 2009).

Descending regressive trajectories or 'forced regression' (Fig. 2.8) occurs as a result of accommodation being reduced during relative sea-level fall, leading to thin parasequences with laterally extensive erosional surfaces on the shelf, and a limited preservation of the shoreline (Helland-Hansen and Martinsen, 1996; Plint and Nummedal, 2000; Steel and Olsen, 2002; Helland-Hansen and Hampson, 2009). A number of criteria have been proposed for the recognition of forced regressive deposits by Posamentier and Morris (2000) including: (1) the separation down depositional dip of successive shoreface deposits; (2) the presence of sharp-based shoreface deposits and similar ‘foreshortened’ stratigraphic successions; (3) the
reduction in relief of clinoforms from proximal to distal locations; (4) long distance regression of unit; (5) absence of coastal-plain facies capping proximal regressive successions; (6) a seaward dipping bounding surface at the top of the regressive succession; and (7) the increased average grain size from proximal to distal locations. However, the stratal architecture of forced regressive deposits can be highly variable due to the number of factors which influence their preservation including: (1) sea-floor gradient; (2) the ratio of sediment flux to the rate of sea level fall; (3) the continuity of sea level fall; (4) the continuity of sediment input; and (5) the relative changes in physical processes at and near the shoreline (e.g. wave- vs. tide- vs. fluvial) during sea-level fall.

A nonaccretionary transgressive trajectory (Fig. 2.8) is a result of accommodation space being created with no sediment supply leading to similar features to a forced regressive trajectory of foreshortened and/or missing facies belts, laterally extensive erosional surfaces and a low preservation potential of the shoreline system (Helland-Hansen and Martinsen, 1996; Helland-Hansen and Hampson, 2009). Transgressions can be swift and remove significant amounts of sediment to create a relatively smooth trajectory. However, the amount of sediment removed is dependent on a number of factors including: (1) the hardness of the substrate; (2) the relative energy of waves and tides; (3) the speed of transgression; (4) the amount of vegetation on the coastal plain; and (5) the grain size of the substrate (Posamentier and Morris, 2000).

An accretionary transgressive trajectory (Fig. 2.8) is the result of sediment supply outpacing accommodation space creation during transgression and has similar features to an ascending regressive trajectory with expanded facies belts, absence of laterally extensive erosional surfaces and a high preservation potential of the shoreline (Helland-Hansen and Martinsen, 1996; Helland-Hansen and Hampson, 2009). In these cases, a stepped transgressive surface may be formed through a series of offset ravinement surfaces as a result of small progradational
packages occurring during overall transgression (Helland-Hansen and Gjelberg, 1994; Cattaneo and Steel, 2003).

2.3 Seismic stratigraphy and clinoforms

Seismic stratigraphy is the application of the above sequence stratigraphic framework to seismic reflection data (Mitchum et al., 1977a). A sequence stratigraphic interpretation using seismic data can be achieved with the understanding that seismic reflections image strata of different acoustic properties and therefore reflections may be considered as ‘chronostratigraphic’ stratal surfaces (Vail and Mitchum, 1977; Vail et al., 1977). Seismic stratigraphy also utilises seismic reflection terminations and configurations to correlate stratigraphic surfaces and interpret depositional environments in a manner similar to sequence stratigraphic techniques (Mitchum et al., 1977a).

A key component to interpreting depositional environments using seismic data and shoreline trajectory analysis is the clinoform (sensu Rich, 1951). A clinoform is a seaward-dipping, three-dimensional discontinuity surface which can help identify the following parameters in shallow-marine successions: (1) relative sea level changes; (2) rates of sediment supply; (3) calibre of sediment; (4) the proportion of wave and tide processes at shoreline and shelf; (5) morphology of shoreline; and (6) bathymetry from shoreline to shelf and slope (Driscoll and Karner, 1999; Bhattacharya, 2006; Porębski and Steel, 2006; Helland-Hansen and Hampson, 2009). The analysis of the geometry of clinoforms imaged in seismic data can therefore identify stratigraphic architectural controls and depositional environment, as discussed below.

Clinoforms vary in scale ranging from those on a continental margin which are hundreds of metres in height (e.g. Helland-Hansen, 1990) to shoreline clinoforms which can be less than 10 m in height (e.g. Bentham et al., 1991). The "clinoform break" or "rollover point" is located
between the topset and foreset and is significant in both end-members highlighted above (Pirmez et al., 1998). In shelf clinoforms, the clinoform break, called the shelf-edge break (Fig. 2.9), represents the separation of the shelf, which is subject to the prevailing basinal energy regime, and the slope, which is subject to mass gravity processes (Helland-Hansen and Hampson, 2009). In shoreline clinoforms, with which this study is concerned, the clinoform break, called the shoreline break (Fig. 2.9a), typically represents the position of the shoreline and therefore the transition zone between subaqueous and subaerial facies which can accordingly be used to conduct shoreline trajectory analysis, as described above (Helland-Hansen and Hampson, 2009).

The shape of the clinoform describes the efficiency of sediment transport across- and along-shelf (Driscoll and Karner, 1999). Sigmoidal clinoforms, composed of a gently-dipping (<1°) topset, foreset and bottomset architecture, indicate accommodation space creation is greater than or equal to sediment supply. These are classified as aggradational. Conversely, oblique clinoforms, composed of a gently-dipping topset (<0.5°), steep-dipping foreset (1-15°) and gently-dipping bottomset (<0.5°) architecture, indicate that sediment supply is outpacing accommodation space creation and are termed progradational (Driscoll and Karner, 1999). Furthermore, oblique clinoforms are typically associated with coarse-grained systems, horizontal or descending regressive trajectories, and weak reworking by hydrodynamic forces. Conversely, sigmoidal clinoforms are typically associated with fine-grained systems, ascending trajectories, and strong reworking by hydrodynamic forces (Pirmez et al., 1998; Driscoll and Karner, 1999; Cattaneo et al., 2003; Swenson et al., 2005).

Shoreline clinoform geometry and stacking patterns may vary spatially and temporally dependent on sediment calibre, proximity to source and dominant hydrodynamic regime (Driscoll and Karner, 1999). Along strike variability in the geometry of clinoforms is high when the shoreline position and the clinoform break are nearly coincident. Conversely, when the
clinoform break and the shoreline are separated by a subaqueous platform, as a result of strong hydrodynamic forces distributing sediment efficiently, the along strike variability in clinoforms is low (Nittrouer et al., 1986; Driscoll and Karner, 1999; Liu et al., 2006). In this scenario of a high energy marine setting, a “compound clinoform” geometry exists of a subaerial clinoform (<40 m height), indicating the position of the shoreline, that is laterally separated from a subaqueous clinoform (<100 m height) (Pirmez et al., 1998; Swenson et al., 2005). The total height of the compound clinoform system is usually less than 100 m and leads to shelf clinoforms down-dip, which may be similar in geometry to single shoreline clinoforms in less energetic marine settings (Helland-Hansen and Hampson, 2009).

### 2.3.1 Restrictions on the use of seismic stratigraphy

Seismic stratigraphy has enabled interpreters to understand the three-dimensional variability in sequences over large distances unlike many field-studies which are limited by outcrop orientation, outcrop accessibility and post-depositional erosion (Biddle et al., 1992). This is especially important when considering how stratal surfaces can be diachronous along strike due to variations in subsidence and sedimentation rates (Catuneau, 2002). Seismic stratigraphy is dependent on the ability to identify reflection relationships (terminations, stacking patterns etc) within seismic data. However, the ability to see these relationships is limited by the resolution of the seismic data (e.g. Widess, 1973; Meckel Jr and Nath, 1977; Neidell and Poggiagliolmi, 1977; Sheriff, 1977). This significantly constrains the interpreter’s capability to conduct seismic stratigraphic analysis because high frequency stratal units are poorly imaged, which has led to the dominance of ‘third-order’ frameworks in seismic stratigraphic studies (Catuneanu et al., 2009).

The focus of the previous section highlights how clinoforms imaged in seismic reflection data can be used to make inferences about the depositional environments and, combined with
shoreline trajectory analysis, can elucidate the evolution of depositional environments through time. However, like recognising reflection terminations, the recognition of clinoforms is dependent on the ability to identify them in seismic reflection data and the use of seismically-imaged clinoforms is often restricted by seismic resolution especially when applied to shoreline clinoforms (e.g. Rudolph et al., 1989; Adams et al., 2001; Zeng and Kerans, 2003; Bullimore et al., 2005; Fournier and Borgomano, 2007; Zeng, 2007).

### 2.4 Forward seismic modelling

Forward seismic modelling is a method by which petrophysical properties are integrated with an interpreted lithology distribution to describe seismic amplitudes. The interpreted lithology distribution is extracted from outcrop analogues where stratal geometries and facies patterns can be explicitly described in the forward model. The petrophysical properties are extracted from equivalent facies identified through core and wireline-log data from the target reservoir and integrated with the forward model to create a synthetic seismogram or forward seismic model. This method therefore bridges the gap in resolution between observed outcrop characteristics and/or well data and seismic stratigraphy. Furthermore, the method constrains lithology distributions and reservoir architectures that would otherwise be overlooked or misinterpreted in the subsurface (Fig. 2.10).

Forward seismic modelling has frequently been applied to investigate the complex seismic response of intricate carbonate stratigraphic architecture (e.g. Rudolph et al., 1989; Biddle et al., 1992; Campbell and Stafleu, 1992; Stafleu and Schlager, 1993; Stafleu et al., 1994; Bracco Gartner and Schlager, 1999; Kenter et al., 2001; Zeng and Kerans, 2003; Janson et al., 2007). One important result which occurred soon after the advent of forward seismic modelling was the recognition that not all seismic reflection patterns replicated stratal patterns visible in outcrop (e.g. Rudolph et al., 1989; Biddle et al., 1992; Campbell and Stafleu, 1992; Stafleu and Schlager,
1993; Stafleu et al., 1994). These pseudo-unconformities (sensu Schlager et al., 1991) imaged in seismic data were replicating rapid changes in dip and facies rather than corresponding to stratal surfaces.

More recently forward seismic modelling has been applied to the seismic response of deep-water systems, in order to understand stacking and connectivity between deep-water turbiditic sands (e.g. Abreu et al., 2003; Bourgeois et al., 2004; Sullivan et al., 2004; Schwab et al., 2007; Bakke et al., 2008; Stanbrook et al., 2008; Falivene et al., 2010; Bakke et al., 2011; Bakke et al., 2013). However, rarely has forward seismic modelling been applied to shallow-water systems (e.g. Helland-Hansen et al., 1994; Hodgetts and Howell, 2000; Braaksma et al., 2006; Raides et al., 2011; Johansen, 2013). Furthermore, the method has never been applied to understand the seismic response of shallow-marine to shoreface clinoforms from disparate river-, wave-, and tide-dominated end members. Considering the above discussion on the importance of determining characteristics of seismically-imaged clinoforms, forward seismic modelling is believed to be a valuable additional tool in deciphering shallow-marine siliciclastic environments.
Figure 2.1. Rifting patterns in North Sea Basin during: (a) Triassic extension; (b) Oxfordian to early Kimmeridgian extension; and (c) late Kimmeridgian to Early Cretaceous (modified after Zanella and Coward, 2003).
Figure 2.2. Palaeogeographic maps showing the distribution of depositional environments in the central and northern North Sea during the: (a) Bajocian; (b) Callovian; (c) early to late Oxfordian; (d) late Oxfordian to early Kimmeridgian; and (e) early to late Kimmeridgian (modified after Fraser et al., 2002; Husmo et al., 2002).
Figure 2.3. Stratigraphic chart showing lithostratigraphic and chronostratigraphic relationships within the Viking Group on the Horda Platform and terraces towards the Viking Graben (for location see Figure 2.2) (modified after Partington et al., 1993; Stewart et al., 1995; Fraser et al., 2002; Husmo et al., 2002).
Figure 2.4. (previous page) (a) Palaeogeographic map showing the distribution of depositional environments and active structural elements in the northern North Sea during the late Callovian (prior to MFS J46) and deposition of the Fensfjord Formation (modified after Rattey and Hayward, 1993; Færseth and Ravnås, 1998; Fraser et al., 2002; Husmo et al., 2002; Zanella and Coward, 2003; Whipp et al., 2013); and (b) schematic cross-section through the Troll Field and sub-basins towards the North Viking Graben highlighting the depositional environments and active structural elements during the deposition of the Fensfjord Formation.

Figure 2.5. Depositional model for the Sognefjord Formation in the Troll Field, northern North Sea (after Dreyer et al., 2005).
**Figure 2.6.** (a) Structural elements map for the Late Jurassic, northern North Sea highlighting producing oil and gas fields (modified after Fraser *et al.*, 2002) and (b) geoseismic profile illustrating the major rotated fault blocks and petroleum system of the northern North Sea, for location see (a) (modified after Husmo *et al.*, 2002).
Figure 2.7. Sequence stratigraphic concepts and terms including: (a) stratal termination types; (b) chronostratigraphic surfaces and systems tracts; (c) parasequence stacking patterns resulting from changing rates of coastal accommodation creation (δA) and sediment fill (δS); and an idealised stratal and sequence hierarchy with (d) depositional sequences and (e) idealised facies distributions (modified after Neal and Abreu, 2009; Abreu et al., 2010).
Figure 2.8. Shoreline trajectory classes with shoreline trajectory highlighted by bold lines (modified after Helland-Hansen and Hampson, 2009).
Figure 2.9. Simplified depositional-dip shoreline-shelf profile illustrating the facies and geometrical characteristics of clinoforms with shoreline clinoforms consisting of either a single (a) or a compound (b) clinoform set (modified after Helland-Hansen and Hampson, 2009).
Step 1: Define subsurface missing target properties (e.g. lithology distribution)

Step 2: Identify suitable outcrop analogue and create analogue model

Step 3: Create synthetic seismograms at a range of seismic frequencies

Step 4: Compare seismic model component with subsurface target to interpret

Figure 2.10. An illustrated workflow for building forward seismic models to interpret facies composition of a complex channel system, offshore Angola, using the Ainsa II turbidite system, Spain, as an outcrop analogue (modified after Bakke et al., 2008; Bakke et al., 2011).
CHAPTER 3

Sedimentology and sequence stratigraphy of the Middle-Upper Jurassic Krossfjord and Fensfjord formations, Troll Field, northern North Sea

3.1 Abstract

The Middle-to-Upper Jurassic Krossfjord and Fensfjord formations are secondary reservoir targets in the super-giant Troll oil and gas field, Horda Platform, offshore Norway. The formations comprise sandstones (c. 195 m thick) sourced from the Norwegian mainland to the east, that pinch out basinwards into offshore shales of the Heather Formation to the west. Sedimentological analysis of cores from the Troll Field has identified six facies associations, which represent wave- and tide-dominated deltaic, shoreline and shelf depositional environments. Resulting depositional models highlight the complex distribution of depositional environments and reflect spatial and temporal variations in physical processes at the shoreline, rate of sediment supply and accommodation development. These models are further complicated by the absence of coastal-plain facies, which implies that the Troll Field was fully subaqueous during deposition, that shoreline regression was forced by falling sea level, or that coastal-plain deposits were removed by transgression. Genetic sequences bounded by major flooding surfaces ('series') exhibit laterally uniform thicknesses, implying no major tectonic influence on sedimentation. The recognition of pronounced variability in facies character and stratigraphic architecture emphasise the need for a robust depositional model of the formations in order to drive future exploration in these, and coeval, reservoirs.
3.2 Introduction

The super-giant Troll oil and gas field is located on the Horda Platform on the eastern margin of the Viking Graben, northern North Sea (Fig. 3.1a) and has produced 220.7 million Sm³ (1.39 billion barrels) of oil and 391.8 billion Sm³ (13.84 trillion cubic feet) of gas during 21 years of production since 1990 (NPD, 2011). The Field is divided into the Troll West and Troll East accumulations, although pressure communication has been proven between the two accumulations (NPD, 2011). Rotated fault blocks define the traps for both accumulations (Fig. 3.1c) and the reservoir consists of shallow-marine sandstones; production to date has been from the Sognefjord Formation (Fm) (Oxfordian to Kimmeridgian/Volgian) (Fig. 3.2). The underlying Fensfjord Formation (Callovian) forms part of the reservoir and has a proven oil column of 6 - 9 metres in the northern part of Troll East (NPD, 2011). The Fensfjord Formation also forms a significant reservoir in the Brage Field (Callovian to Oxfordian), which lies 20 km to the SW of Troll (Fig. 3.1a). The Sognefjord and Fensfjord formations, together with the underlying Krossfjord Formation (Bathonian), form part of the Viking Group (Gp), which is situated above the prolific Brent Group (Fig. 3.2).

The sedimentology of the Krossfjord and Fensfjord formations is poorly understood as they have not been the focus of previous published work, despite the formations containing potentially large reserves. The formations comprise sandstones principally sourced from the Norwegian mainland to the east and pinch out basinwards into the offshore shales of the Heather Formation to the west towards the North Viking Graben (Stewart et al., 1995). The development of a detailed sedimentological and sequence stratigraphic model for the Krossfjord and Fensfjord formations is complicated by two factors. First, the sedimentological character, distribution and stratigraphic architecture of shallow-marine sandstones are strongly controlled by spatial and temporal variation in physical processes at and near the shoreline (e.g. wave- vs. tide- vs. fluvial-dominated processes) (e.g. Gani and Bhattacharya, 2007; Ainsworth et
al., 2011). Second, the geographic partitioning and the relative importance of physical processes can be further complicated in rifts due to fault block rotation, uplift and subsidence; the sedimentology and stratigraphic architecture of both the Krossfjord and Fensfjord formations may thus be anticipated to be complex because they were deposited during the Middle-to-Late Jurassic rift event (Ravnås and Bondevik, 1997).

The aims of this chapter are two-fold: (1) to produce a high-resolution sedimentological and sequence stratigraphic model for the Krossfjord and Fensfjord formations in the Troll Field; and (2) to determine the dominant shoreline processes and genetic stratigraphic relationships within and between these formations. The work reported herein will improve the understanding of syn-rift sandstone distribution in the northern North Sea, and guide future exploration and production from Krossfjord and Fensfjord reservoirs.

3.3 Geological setting and previous work

3.3.1 Regional tectonic context

The Troll Field is located on the Horda Platform, eastern flank of the North Viking Graben. The northern part of the North Sea rift basin is a fault-bounded depocentre that is 170-200 km wide and flanked by the Shetland Platform to the west and the Norwegian mainland to the east. It is the northern arm of the North Sea trilete rift system, which was initially developed during the Triassic as a continental rift system and was reactivated and significantly expanded during the Late Jurassic as a marine rift system (Roberts et al., 1990; Davies et al., 2001). It is characterised by normal faults that strike N-, NE- or NW, and which bound rotated fault blocks that are 15-50 km wide (Færseth and Ravnås, 1998). The fault blocks are arranged around a central low, known as the Viking Graben (Færseth and Ravnås, 1998). The North Viking Graben is located in
the North Sea between 59° and 61° N and represents one arm of the trilete failed rift system (Fig. 3.1a).

The basin underwent a complex tectonic evolution; an initial phase of extension occurred during the Permo-Triassic, and multiple phases of extension occurred in the Middle-to-Late Jurassic (Bajocian to Volgian). The Permo-Triassic and Middle-to-Late Jurassic phases of rifting are separated by a post-rift interval, in which two regional tectonic uplift events are identified in the Hettangian and late Toarcian-Aalenian (Steel, 1993; Færseth, 1996; Færseth and Ravnås, 1998). During the Early Jurassic the basin was characterised by tectonic quiescence and spatially uniform subsidence. The Toarcian-Aalenian event interrupted this period of quiescence and formed the North Sea thermal dome, which provided clastic sediment for the northward progradation and subsequent southwards retreat of the Middle Jurassic "Brent Delta" (Ziegler, 1990b; Underhill and Partington, 1993). The subsequent Middle-to-Late Jurassic depositional systems, rather than being sourced from the North Sea dome, were sourced from the Norwegian mainland to the east and prograded to the west. These systems were deposited across a series of north-south-trending fault blocks that formed during the Middle-to-Late Jurassic rift event (Rattey and Hayward, 1993).

The trilete Middle-to-Late Jurassic rift system formed in response to the deflation of the North Sea thermal dome (Ziegler, 1990b; Underhill and Partington, 1993). Both the initiation and cessation of rifting was diachronous across the basin; in the northern North Sea, rifting initiated in the Bajocian (Johannessen et al., 1995; Hesthammer et al., 1999). The rate of extension and fault-controlled subsidence generally increased through the Jurassic, and were greatest in the late Oxfordian to Kimmeridgian (Færseth and Ravnås, 1998). However, in detail, the Middle-to-Late Jurassic rift event can be divided into several discrete phases of basin-wide rifting and fault-related subsidence; these phases are discussed further below, so that the detailed
3.3.2 Tectonic-stratigraphic evolution of Horda Platform and Troll Field

The Troll Field and surrounding area contains three major sandstone tongues of Middle-to-Late Jurassic age. These are the Krossfjord, Fensfjord and Sognefjord formations of the Viking Group (Vollset and Doré, 1984) (Fig. 3.2), which are each 100-200 m thick near the rift margin and which pinch-out westwards into fine-grained, Heather Formation deposits in the North Viking Graben (Ravnås and Bondevik, 1997). The sandstone tongues were deposited in several transgressive-regressive cycles at the margins of a shallow sea that covered the Horda Platform (Stewart et al., 1995). The combined thickness of the sandstone tongues, and associated Heather Formation siltstones and mudstones, is up to 400 m (Husmo et al., 2002). The Heather Formation is informally split into three parts on the Horda Platform to describe its stratigraphic relationship to the three major sandstone tongues. The Heather “A” unit lies above the Brent Group and beneath the Krossfjord Formation (Bathonian), the Heather “B” unit overlies the Fensfjord Formation and underlies the Sognefjord Formation (Callovian), and the Heather “C” unit overlies the Sognefjord Formation (Oxfordian and Kimmeridgian) (Stewart et al., 1995). The Sognefjord Formation is locally directly overlain by the Draupne Formation and the Heather “C” unit is absent; in these locations the contact between the Sognefjord and Draupne formations is an angular unconformity (Stewart et al., 1995).

Deposition of the Krossfjord, Fensfjord and Sognefjord formations was driven by supply of coarse clastic material from the eastern flank of the developing North Sea rift system as a result of uplift of the Norwegian hinterland (Stewart et al., 1995; Ravnås and Bondevik, 1997). Each of the three formations has been interpreted as a 2nd order genetic sequence (sensu Galloway, 1989), which is bounded by major flooding surfaces (Fig. 3.2) (Stewart et al., 1995). Deposition
in a range of shelf-to-shoreface environments, with varying degrees of tidal and/or fluvial influence, has been interpreted for the formations (Ravnås and Bondevik, 1997; Dreyer et al., 2005).

The tectono-stratigraphic evolution of the Horda Platform and North Sea rift system resulted in development of three structural provinces: (1) the relatively stable Horda Platform in the east; (2) a number of tilted half grabens that host the Brage, Oseberg, Troll and Fram fields (Stewart et al., 1995); and (3) the deep, fault-bounded, North Viking Graben in the west (Fig. 3.1). Rifting during the Middle-to-Late Jurassic can be subdivided into three periods (Fig. 3.2) (Fraser et al., 2002).

(i) Bathonian to latest Callovian. During the initial Bathonian-to-latest Callovian period of rifting, a series of faulted terraces developed between the Viking Graben and the Horda Platform. Rotation of the normal fault blocks, which define the traps of the Oseberg and Brage fields, caused local reworking of the upper Ness and Tarbert formations (Brent Group) on the Horda Platform (Husmo et al., 2002). It was during this period of rifting that the Krossfjord and Fensfjord formations were deposited (Vollset and Doré, 1984; Steel, 1993). In the Troll Field, the Krossfjord Formation is characterised by progradation of a sand-rich delta during relatively low rates of normal faulting and fault block rotation (Ravnås et al., 2000) (Fig. 3.2). The Fensfjord Formation consists of progradationally-stacked, fine-grained sandstones, which accumulated during a period of tectonic quiescence in the middle Callovian, when the sediment supply rate was higher than the basin subsidence rate (Steel, 1993; Stewart et al., 1995; Ravnås and Bondevik, 1997; Fraser et al., 2002). At the point of maximum regression during the late Callovian, the Fensfjord Delta covered the entire Horda Platform and extended into the distal sub-basins that now host the Brage and Oseberg fields (Fig. 3.2) (Husmo et al., 2002). It has been speculated that turbidites, which were presumably sourced from collapse of the Fensfjord Delta front, were deposited westwards of the Fensfjord Delta at this time (Ravnås et al., 2000).
Landward migration of the Fensfjord Delta is attributed to fault-related subsidence outpacing sediment supply, which was coeval with footwall uplift of the western and north-western boundary faults of the Horda Platform. Fault-controlled uplift in the Brage area resulted in the creation of footwall islands, which represented an important intra-basinal sediment source (Ravnås et al., 2000; Husmo et al., 2002). A marine transgression occurred in the late Callovian and resulted in deposition of fine-grained sediments of the Heather “B” unit (Steel, 1993).

(ii) Oxfordian to Kimmeridgian. During this period the Sognefjord Formation was deposited and rifting reached its climax creating the major structural divide between the Viking Graben and the Horda Platform (Stewart et al., 1995). Increased extension caused uplift and tilting of fault blocks, resulting in fault block footwalls rising above sea level at the margins of the Horda Platform. Erosion occurred on these footwall crests, allowing older sediment of the Sognefjord Formation to be reworked and deposited (Fraser et al., 2002).

(iii) Early to middle Volgian. The final stage of rifting caused extensive faulting in the west of the Viking Graben and mild reactivation of faults on the Horda Platform; this resulted in uplift and eastward tilting of normal fault blocks (Fossen et al., 2003). Consequently, many fault blocks suffered local erosion and collapse, which resulted in Lower and Middle Jurassic strata being truncated beneath Upper Jurassic strata in several locations on the Horda Platform (Husmo et al., 2002). Deposition of the Sognefjord Formation was terminated by marine flooding, which led to deposition of deep-marine mudstones of the Draupne Formation (Fraser et al., 2002).

3.3.3 Previous depositional models for the Viking Group

Several depositional models have been published for the sandstones of the Viking Group, with a particular focus on the Sognefjord Formation. An early model interpreted the Sognefjord Formation sandstones as offshore bars, and proposed that transgressive erosion was the main
control on facies distribution (Whitaker, 1984; Hellem et al., 1986; Osborne and Evans, 1987). Eustatic sea level rise was interpreted to have been the major controlling factor on sedimentation. Subsequent models interpreted the sandstone tongues of the Viking Group to represent regressive-to-transgressive deltaic units containing a variety of shallow marine shelf-to-shoreface environments (e.g. Steel, 1993; Stewart et al., 1995). A greater tectonic influence on sedimentation was also suggested due to the recognition of variations in stratal thickness identified on seismic reflection data. The most recent model for the Oxfordian part of the Sognefjord Formation interprets the delta to have been mixed influence, with a wave-dominated spit deflecting fluvially supplied sediment towards the southwest (Dreyer et al., 2005). The spit is interpreted to have been attached to the coast in the north and bordered to the east by a tidal backbasin. These various depositional models have different implications for predicted facies distributions across the Troll Field and elsewhere on the Horda Platform and North Viking Graben.

3.4 Dataset

Thirty-two wells in the Troll Field penetrate the Fensfjord Formation, and 22 of these wells also penetrate the underlying Krossfjord Formation (NPD, 2011). Nine of these wells contain core within the interval of interest, giving a total core length of 893 m, all of which has been logged at a scale of 1:50 (Fig. 3.3 and Table 3.1). In most wells, only the upper part of the Fensfjord Formation is cored. Only one well, 31/2-4R, has complete core recovery of the Fensfjord Formation. Wells 31/5-5 and 31/2-4R have partial core recovery from the Krossfjord Formation (Table 3.1). Sedimentological facies analysis included describing grain size and shape, sorting, sedimentary structures, diagenetic features, and the nature of bedding contacts. In addition, trace and body fossils were documented, including their orientation, size, cross cutting relationships and intensity of bioturbation (cf. Bockelie and Howard, 1984; MacEachern and Bann, 2008). A biostratigraphic framework for the Fensfjord Formation was established by
Whitaker (1984) in the Brage Field. This framework has recently been extended to five cored wells in the Troll Field, through analysis of quantitative palynology and kerogen counts (Geostrat, 2011). Wireline-log data have been used to identify facies in un-cored wells. Borehole image and dipmeter data are not available for the studied wells thus palaeocurrent directions could not be reconstructed from cored or uncored sections of the wells.

Two regional 3D seismic reflection data sets that cover the Troll Field have been interpreted (North Sea exploration blocks 31/2, 31/3, 31/5 and 31/6). Troll West contains seismic survey “NH0301”, which has a coverage of c. 800 km², line spacing of 18.75 m in inline (northeast-southwest) and 12.5 m in crossline (northwest-southeast) directions, and which images to a depth of c. 3000 milliseconds two-way time (ms TWT). Troll East contains seismic survey “SG9202”, which has a coverage of c. 900 km², line spacing of 25 m in both inline (east-west) and crossline (north-south) directions, and which images to a depth of c. 2400 milliseconds two-way time. The surveys overlap in the centre of the Troll Field. Based on seismic velocity and frequency data, the seismic resolution in the interval of interest is estimated to be c. 15-25 m. Attribute data were used to understand the gross stratigraphic architecture of the intervals of interest. However, detailed analysis is complicated by the occurrence of closely spaced gas-oil and oil-water contacts that combine to produce a prominent “flat spot”, which obscures and distorts seismically resolved stratal architectures. This chapter focuses on core sedimentology with detailed seismic analysis presented in Chapter 5.

3.5 Facies analysis and wireline-log calibration

Eleven different facies (A – J) are identified in the Krossfjord and Fensfjord formation (Table 3.2, Fig. 3.4). These facies are grouped into six facies associations, which are described and interpreted below. Representative sedimentary logs of each facies association are displayed in
Figures 3.5-3.7 (full sedimentary logs in Appendix 1), and cross-plots, which illustrate the quantitative wireline-log character of the facies associations, are shown in Figure 3.8.

### 3.5.1 Facies Association 1: Offshore

Facies Association 1 (FA1) is predominantly composed of Facies A1 and A2, with minor proportions of Facies I (Table 3.2), and is cored primarily within the Heather "B" unit (Fig. 3.5b). Whereas Facies A2 is identified in all cored wells across the study area, Facies A1 is absent in wells located in the eastern part of the Troll Field.

**Description.** Facies A1 is a very fine-grained siltstone with Belemnites and *Terebellina* burrows. Facies A2 is a fine-grained siltstone with *Planolites*, *Terebellina* and *Chondrites* burrows (Fig. 3.5b). Bioturbation intensity is high (5-6 on the qualitative scale of MacEachern and Bann, 2008). Facies A1 and A2 are distinguished from one another using the prevailing ichnotaxa and grain size. Contacts between the two facies are gradational over 5 – 10 m and Facies Association 1 can occur in units that are up to 60 m thick. Very coarse- to medium-grained sandstone beds occur locally in Facies Association 1 (Facies I). Beds of Facies I have sharp bases and gradational tops, and vary in thickness from a few centimetres to one metre (e.g. at 1484 m in Fig. 3.5b). Palynofacies analysis indicates an aerobic environment with relatively high salinity, high marine species diversity and low energy (Geostrat, 2011).

**Wireline log signature.** Facies Association 1 is typified by high gamma-ray values, reflecting its high clay content (Figs. 3.5b, 3.8a). Facies A2 has lower gamma-ray values than Facies A1 due to its slightly coarser grain size. Calcite cemented horizons are easily identified and are represented by high value "spikes" in the density log and low value "spikes" in the neutron porosity log. Beds of Facies I within siltstone-dominated successions of Facies A2 are marked by pronounced decreases in neutron porosity and sonic values, increased resistivity values and a
slight decrease in gamma-ray values. The wireline log signature of Facies I is clear where thick beds of this facies are developed; in thinner beds, however, the response is weaker, and only smaller spikes in the neutron porosity and sonic logs are observed (e.g. at 1484 m in Fig. 3.5b).

*Interpretation.* Facies A is characterised by highly bioturbated, very fine-grained sediments, suggesting deposition of fine material from suspension fall-out (Howard and Reineck, 1981; Collinson and Thompson, 1989). The high concentration of siltstone and lack of primary bedding structures suggests offshore deposits were subject to extensive bioturbation below storm wave base (MacEachern and Bann, 2008), probably in a middle-to-outer shelf environment. The very dark colour of Facies A1 indicates a high organic content, which, when combined with the high bioturbation index, implies deposition in an environment ideal for benthic fauna with high nutrient and oxygen levels and normal marine salinity. This interpretation is reinforced by palynofacies analysis. Facies A1 has a higher siltstone content compared to Facies A2, which may be indicative of a deeper water environment for the former, which is reinforced by the ichnotaxa present (Pemberton *et al.*, 1992). Facies A1 coarsens upwards into Facies A2, signifying a shallowing of water depth.

The isolated, sharp based, very coarse- to medium-grained sandstone beds (Facies I) are interpreted to be the product of high-energy events that were able to transport coarse-grained material into a relatively distal environment. A number of mechanisms have been proposed for the origin of such deposits, including offshore sediment transport by major storms, floods, rip currents, river-fed hyperpycnal flows and tsunamis (Kumar and Sanders, 1976; Gruszczynski *et al.*, 1993; Mulder *et al.*, 2003). Similar interpretations have been proposed for comparable beds in the Sognefjord Formation (Dreyer *et al.*, 2005). Wells which contain thicker event beds are located in the northern part of Troll West, closer to the sediment source inferred for the Sognefjord Formation (Dreyer *et al.*, 2005).
3.5.2 Facies Association 2: Wave-dominated lower shoreface

Facies Association 2 (FA2) is identified throughout the Fensfjord Formation in the Troll Field (e.g. Fig. 3.6a) and in the Krossfjord Formation in Troll West (Fig. 3.3).

**Description.** FA2 is composed of 5-15 m thick, upward-coarsening successions of Facies B and C. Bioturbated, very fine- to medium-grained sandstone that contain rare low-angle laminations (Facies B) at the base of the association pass gradationally upwards into very fine- to medium-grained sandstone that contain rhythmically interbedded, hummocky cross-stratified and bioturbated intervals (Facies C). Sharp based, poorly sorted, structureless beds of granular- to fine-grained, light grey sandstone (Facies I), which are 0.2-0.5 m thick, occur within Facies B and C. The dominant ichnotaxa recognised in FA2 are *Skolithos* and *Ophiomorpha*. *Planolites* and *Chondrites* are also evident, in addition to rare, fully disarticulated shell fragments. Bioturbation decreases in intensity from the base to the top of the association. Palynofacies analyses indicates an aerobic environment with relatively high salinity, low marine species diversity and low energy (Geostrat, 2011).

FA2 typically overlies other coarsening upwards successions within the Fensfjord and Krossfjord formations (e.g. at 1618 m in Fig. 3.6a). The lower and upper boundaries of FA2 are sharp; the upper boundary is occasionally overlain by a dark green-to-grey, poorly sorted, matrix supported, medium- to coarse-grained sandstone (Facies J).

**Wireline log signature.** The upward transition from Facies B to C is represented on wireline logs by an upward decrease in gamma-ray and density log values, and an upward increase in neutron porosity and sonic log values (e.g. from 1618 to 1612 m in Fig. 3.6a). The abundance of mica in Facies B and C results in FA2 having overall high gamma-ray values (Serra and Serra, 2003) (e.g. at 1615 m in Fig. 3.6a; Fig. 3.8b).
Interpretation. FA2 is defined by the alternations between fair-weather suspension settling and more energetic hydrodynamic conditions. Facies B is composed of interbedded sandstone and siltstone, which is indicative of fluctuating energy levels (Bourgeois, 1980; Dott and Bourgeois, 1982; Duke, 1985). Siltstone is deposited via fair-weather suspension settling, whilst sandstone is deposited through waning, storm-generated suspension currents (Brenchley, 1985). The high bioturbation index of siltstones in Facies B suggests that they record prolonged fair-weather periods that allowed biogenic reworking of the sediment. These characteristics are typical of the "transition zone", which is above storm wave base but below fair-weather wave base (Reineck and Singh, 1973), and which is referred to here as the "distal lower shoreface" (sensu Van Wagoner et al., 1990). This zone was only disturbed by large, infrequent storm events. The low-angle, inclined laminasets and beds identified in Facies C are interpreted as hummocky cross-stratification (HCS) (e.g. at 1612.5 m in Fig. 3.6a). HCS is the result of combined flow that is formed by a unidirectional current generated by a storm, which carries sand out from the coast under the influence of high-amplitude waves. The waves then disperse the sand through oscillatory motion, depositing it as hummocks (Bourgeois, 1980; Dott and Bourgeois, 1982; Duke, 1985). Similar low-angle, inclined laminasets and beds are interpreted as HCS in the Sognefjord Formation in the Troll Field (Stewart et al., 1995; Dreyer et al., 2005). The rhythmic interbedding of bioturbated sandstone and hummocky cross-stratified sandstone reflects the alternation between fair-weather deposition and storm deposition, in water depths that lie between fair-weather wave base and storm-wave base (Walker, 1984; Brenchley, 1985). The prevalence of HCS in Facies C indicates a more proximal lower shoreface location (sensu Van Wagoner et al., 1990) compared to Facies B (Dott and Bourgeois, 1982). Thin (<10 cm), structureless beds of Facies I are interpreted to have been deposited by gravity-driven or storm-related flows (cf. Dreyer et al., 2005).
3.5.3 Facies Association 3: Wave-dominated upper shoreface and foreshore

Facies Association 3 is identified in the Fensfjord Formation throughout the Troll Field (Fig. 3.6a), except in the upper part of the formation in Troll East. It is also present in the Krossfjord Formation.

**Description.** FA3 is 2-15 m thick and is comprised of Facies D and E. Medium- to coarse-grained, well-sorted, planar laminated to trough and tabular cross-bedded sandstone (Facies D) is typically interbedded with and/or coarsens upwards into medium- to coarse-grained, apparently structureless sandstone (Facies E). Broken shell material is common within FA3, although articulated shells are occasionally present, especially in Facies E (e.g. at 1618 m in Fig. 3.6a). Overall, FA3 is poorly bioturbated. FA3 always overlies FA2 in upwards coarsening successions. Palynofacies analyses indicates an aerobic environment with relatively high salinity, low marine species diversity and high energy (Geostrat, 2011).

**Well log signature.** Gamma-ray values in FA3 are usually low due to the lack of clay and mica (Fig. 3.8c). Density values are also lower than for FA2, reflecting greater porosity in FA3 (Fig. 3.8c). Variability in the log signature of FA3 is principally due to patchy calcite cementation (e.g. at 1625.5 m in Fig. 3.6a).

**Interpretation.** Facies D lacks bioturbation and contains an abundance of planar lamination and trough cross-bedding. The well sorted character of the sandstone indicates extensive reworking, probably in a high energy marine environment above fair-weather wave base (i.e. the upper shoreface; *sensu* Van Wagoner *et al.*, 1990; see also Walker, 1984). The facies lacks evidence for tidal influence (e.g. reactivation surfaces, bi-directional current ripples; Nio and Yang, 1991). Alternation of trough cross-bedding and planar lamination can be explained by the migration of longshore bars and troughs (e.g. Nielsen and Johannessen, 2001). The bars were
dominated by unidirectional currents, which winnowed out finer grained material. Migration of the bars and superimposed three-dimensional dunes resulted in the deposition of trough cross-bedded intervals. The parallel laminations, however, formed due to the passage of weaker currents through the inter-dune trough areas, which accounts for the overall finer grain sizes (cf. Gani and Bhattacharya, 2007).

Facies E is common in the upper part of upward coarsening successions of FA3. The coarse-grained, very well sorted character of the sandstone in Facies E suggests constant wave action, which winnowed out finer grains (Hart and Plint, 1995). The structureless appearance and rare parallel laminations is consistent with deposition in a foreshore environment characterised by swash processes, although similar structures have also been documented on the crest of bar forms (Hunter et al., 1979).

The vertical stacking of Facies Associations 2 and 3 to form overall upwards coarsening successions is interpreted to represent progradation of a wave-dominated shoreface (Fig. 3.6a).

3.5.4 Facies Association 4: Wave-dominated, tide-influenced upper shoreface

Facies Association 4 (FA4) is identified in the upper part of the Fensfjord Formation in Troll East only (Fig. 3.6b).

*Description.* FA4 comprises fine- to medium-grained, planar- to trough-cross stratified sandstone (Facies F). Cross-beds in Facies F appear bi-directional (e.g. Fig. 3.4e) and siltstone drapes are identified along cross-bed foresets (e.g. at 1710 m in Fig. 3.6b). These sandstones are interbedded with units of sharp based, fining upwards, fine- to coarse-grained sandstone, which are stacked to form upwards fining units that are 0.5 m thick (Facies G). Facies G also contains abundant mica and carbonaceous fragments, and very rare parallel laminations are identified in
thin beds (<0.1 m). The dominant ichnotaxa in FA4 are *Palaeophycus*, *Planolites*, and *Skolithos*. Shell fragments occur sporadically throughout this facies association. Palynofacies analyses indicates an aerobic to dysaerobic environment with relatively low salinity, low marine species diversity and relatively low energy (Geostrat, 2011).

Successions of FA4 are 5-10 m thick and typically overlie lower shoreface deposits of FA2. FA2, in this context, is intensely bioturbated (e.g. at 1720 m in Fig. 3.6b). FA4 also occurs in association with FA5 in Troll East (Fig. 3.7a).

*Well log signature.* The upward transition from Facies F to G is represented by a decrease in gamma-ray and neutron porosity values, and an increase in density values. This is due to the coarser grain size and lower mica content of Facies G compared to Facies F. The variable grain size in beds of Facies F and G gives FA4 an overall serrated appearance on wireline logs (e.g. at 1712 m in Fig. 3.6b).

*Interpretation.* The well-sorted character and abundance of planar- to trough-cross stratified beds in Facies F suggest a relatively high-energy depositional environment above fair-weather wave base (upper shoreface *sensu* Van Wagoner *et al.*, 1990). The occurrence of apparently opposed cross-bed orientations implies episodic reversals in flow direction. Sandstone laminae are also draped with silt, occasionally displaying rhythmical silt-layer couplets (cf. Visser, 1980). The deposition of silt may have occurred during slack water periods associated with tidal currents (Nio and Yang, 1991). This facies is therefore interpreted as being deposited in an upper shoreface environment where fair-weather wave-driven processes were modulated by tidal influence (cf. Vakarelov *et al.*, 2012). Facies G consists of sharp based, coarse-grained, normally graded sandstone beds occasionally with a basal lag (e.g. at 1714.5 m in Fig. 3.6b) suggesting deposition by an energetic process that was able to erode into the underlying deposits. The units are thin (0.3-1 m) and commonly have bioturbated tops, implying that Facies
G was deposited under waning flow conditions. Whilst such deposits can be formed by a number of processes (e.g. storm-generated rip currents or gravity flows generated by river floods), the position of this facies within tide-influenced deposits suggests they could be remnants of tidal channels (cf. Israel et al., 1987).

FA4 overlies lower shoreface deposits of FA2. In this context, Facies B appears highly bioturbated, which could reflect increased sediment colonisation by burrowing organisms, promoted by tidal current action (Dashtgard et al., 2012). However, there is no direct evidence to support a wave-dominated, tide-influenced lower shoreface succession, despite Facies B and C occurring beneath wave-dominated, tide-influenced upper shoreface deposits.

### 3.5.5 Facies Association 5: Tide-dominated, wave-influenced embayment

Facies Association 5 (FA5) is identified in the middle of the Fensfjord Formation in Troll East (31/6-1; Fig. 3.7a).

**Description.** FA5 consists of fine-grained, well-sorted sandstone that contain lenses of siltstone or mudstone (Facies H). The facies association contains ripple cross-lamination, isolated ripple sets and discontinuous laminae, which occur on a millimetre scale (e.g. at 1605 m in Fig. 3.7a). The lenses of mudstone (> 1 cm thick) appear either laminated or structureless. Wavy-bedded lamina-sets show regular reversals in the direction of ripple foreset dip between beds. Planar-laminated fine-grained sandstone beds occur rarely in this facies association. Wavy bedding dominates Facies H, but lenticular bedding and flaser bedding also occur. Interbedded sandstone and siltstone occur at decimetre scale as the facies fines upwards. Syn-sedimentary micro faults are also evident. Synaeresis cracks are not identified. Bioturbation and body fossils are absent. Palynological analyses indicate brackish conditions with relatively low energy (Geostrat, 2011).
The facies association is only identified in association with FA4; thin (<3 m) successions of FA5 are typically intercalated with thicker (5-10 m) successions of FA4. Contacts between these facies association are sharp.

*Well log signature.* Gamma-ray, density and neutron porosity values are nearly uniform within FA5 (Figs. 3.7a, 3.8e). Gamma-ray and neutron porosity values appear high and density values appear low compared to other facies associations. The only variability in these logs is due to the localised occurrence of calcite cement. In these cases the gamma-ray and neutron porosity decreases and the density increases.

*Interpretation.* The interbedded rippled sandstone and mudstone layers suggest periodic fluctuations in hydrodynamic conditions, from high current velocity, capable of moving sand grains to form ripples, to slack water conditions, which allowed silt, mud and mica to settle out of suspension (Reineck and Wunderlich, 1968). The planar-laminated sandstone beds are interpreted as event beds deposited from upper flow-regime conditions, most likely during storms based on the well sorted, fine-grained character of the sandstones and the overall facies context. The upward fining trend seen in Facies H suggests that the overall hydrodynamic energy of the system was decreasing. The low bioturbation index may result from a narrow colonisation window that is related to high rates of sedimentation (e.g. Gani *et al.*, 2007) and/or brackish water conditions, and/or fluctuating salinities.

Periodic alternation of hydrodynamic energy is common to a number of tidal and estuarine environments such as subtidal flats, tidal channels, and intertidal flats (e.g. Reineck and Wunderlich, 1968). Flaser and wavy bedding have also been recognised in fluvial (Martin, 2000) and lacustrine environments (Ainsworth *et al.*, 2012). Furthermore, the thin, structureless mudstone lenses could represent fluid-mud deposits which have been identified in tidal-fluvial
channel successions, mouth bar and terminal distributary channel successions, and delta front successions (Ichaso and Dalrymple, 2009). However, FA5 is only identified in association with FA4, which is interpreted as a wave-dominated, tide-influenced shoreface; this argues against a fluvial or lacustrine environment of deposition. FA5 is therefore interpreted to document tidally-dominated deposition in a sheltered environment, such as an embayment. There is no evidence of subaerial exposure, implying deposition in a sub-tidal setting. The facies association is commonly associated with FA4, which suggests it represents shallow-water deposition above a wave-dominated, tide-influenced upper shoreface.

3.5.6 Facies Association 6: Delta-front

Facies Association 6 (FA6) is only identified in the Krossfjord Formation (well 31/5-5; Fig. 3.7b).

Description. FA6 is identified in one well only and is 30 m thick. This facies association is characterised by sharp-based, normally graded, well-sorted beds of medium- to coarse-grained sandstone (Facies I). These beds, which amalgamate to form units that are 2-4 m thick, are typically structureless although rare planar lamination is observed. Rare, (< 1 m thick) intervals of Facies I are calcite cemented. The tops of the fining upwards units consist of 0.2-0.3 m thick intervals of bioturbated, fine- to medium-grained sandstone (Facies B). Bioturbation is restricted to the top of the upwards-fining sandstone beds. No palynofacies analyses were conducted on the Krossfjord Formation.

Well log signature. FA6 is characterised by uniformly high neutron porosity values, and uniformly low density and gamma ray values (Figs. 3.7b, 3.8f). Thin intervals of calcite cement (< 1 m) cause a decrease in neutron porosity values and an increase in density values (e.g. at
1862.5 m in Fig. 3.7b). Gamma ray values are locally increased by high concentrations of carbonaceous debris (e.g. at 1875 m in Fig. 3.7b).

**Interpretation.** The fining-upwards beds of structureless and parallel-laminated sandstone are interpreted to have been deposited from high-energy, high-concentration submarine gravity-flows, which were characterised by a high rate of deposition (Lowe, 1982; Middleton, 1993). The stacked, amalgamated nature of the beds implies repeated, gravity flows (e.g. at 1875 m in Fig. 3.7b). Three mechanisms may plausibly produce the sediment gravity flows that deposited the thick-bedded, structureless sandstone of FA6. First, sustained hyperpycnal flows could be generated by the introduction of dense, sediment-laden water from rivers into the basin (Mulder et al., 2003; Plink-Björklund and Steel, 2004). Second, repeated, retrogressive failure of sand-rich, shallow-marine mouth-bars could generate turbidity currents (Olariu et al., 2010). Third, ‘sediment breaching’ (i.e. gradual retrogression of steep, subaqueous, sand-rich slopes) may generate sustained turbidity currents (Van Den Berg et al., 2002). The high-density character of the interpreted gravity flows favours a hyperpycnal flow or sediment breaching origin. Based on the coarse grain size and thickness of amalgamated beds in FA6, in combination with its close proximity to the sediment source at the palaeoshoreline, FA6 is interpreted to have been deposited in a proximal delta-front environment (cf. Mutti et al., 2000; Olariu et al., 2010).

**3.6 Sequence stratigraphic framework**

A sequence stratigraphic framework has been established for the Krossfjord and Fensfjord formations in the Troll Field, using the procedure outlined below. In each cored interval, facies associations are stacked vertically into upward-shallowing successions bounded by transgressive surfaces, which are marked by landward facies shifts (e.g. FA3 overlain by FA 2) (Figs. 3.6, 3.7). These upward-shallowing successions are equivalent to parasequences (*sensu*...)
Van Wagoner et al., 1990). Core observations indicate that parasequences typically have a wave-dominated (Facies Associations 2 and 3), mixed wave- and tide-influenced (Facies Associations 2, 4 and 5), or a fluvial-dominated character (Facies Association 6). However, the various sandstone-prone facies associations have similar wireline-log character (Fig. 3.8). As a result, parasequences can be recognised in uncored intervals and wells by vertical trends in wireline logs (e.g. intervals of upward-decreasing gamma-ray values bounded by sharp increases in gamma-ray values), but their internal facies character cannot be confidently interpreted in the absence of core.

Correlation of groups of parasequences between wells has been constrained by field-wide mapping of major seismic reflectors, and by biostratigraphic data (Figs. 3.9, 3.10). A high resolution biostratigraphic scheme for the Fensfjord Formation in the Troll Field has identified thirteen bioevents and six palynofacies associations (Geostrat, 2011). Major flooding surfaces, marked by pronounced landward facies shifts in vertical successions, have been calibrated using biostratigraphic data to identify the maximum flooding surfaces interpreted throughout the Northern North Sea in the widely used stratigraphic scheme of Partington et al. (1993) (so-called "J surfaces"). Additionally and where data permit, minor transgressive surfaces recognised in core and wireline-log data have been correlated using the Troll Field biostratigraphic scheme described above. These have not been extended or correlated to other studies beyond the Troll Field. Biostratigraphic control is absent in un-cored intervals, which broadly correspond to the lower Fensfjord Formation and Krossfjord Formation.

Sequence boundaries (sensu Van Wagoner et al., 1990) have not been interpreted, because no surfaces marked by fluvial incision or subaerial exposure have been identified in core and wireline-log data. Furthermore, palynological analysis has not conclusively identified a non-marine or marginal-marine assemblage that defines a regionally-developed surface (i.e. sequence boundary) that can be correlated across the Troll Field (Geostrat, 2011). As a
consequence, the groups of parasequences (i.e. parasequence sets; sensu Van Wagoner et al., 1990) that are correlated between regional maximum flooding surfaces (“J surfaces”) correspond to ‘genetic sequences’ (sensu Galloway, 1989). Each of the parasequence groups that are bound by regional maximum flooding surfaces is designated as a numbered ‘series’ (or “basic sequence”), following the established convention used to describe stratigraphic architecture in the Sognefjord Formation reservoir of the Troll Field (e.g. Dreyer et al., 2005).

‘Series 1’ is bounded by the J32 and J42 maximum flooding surfaces at its base and top, respectively, and consists of the Heather “A” unit, the Krossfjord Formation and the lowermost Fensfjord Formation (Figs. 3.9a, 3.10a). ‘Series 1’ contains eight parasequences, which are stacked to form a single progradational parasequence set. ‘Series 2’ is bounded by the J42 and J44 maximum flooding surfaces at its base and top, respectively, and consists of the lower-to-middle Fensfjord Formation (Figs. 3.9a, 3.10a). ‘Series 2’ contains three parasequences, the lowest of which constitutes a progradational parasequence set. The upper two parasequences are stacked within a retrogradational parasequence set, although not all parasequences can be distinguished in each well. ‘Series 3’ is bounded by the J44 and J46 maximum flooding surfaces at its base and top, respectively, and consists of the upper Fensfjord Formation and the lower Heather “B” unit (Figs. 3.9a, 3.10a). ‘Series 3’ contains a maximum of ten parasequences. The lower five parasequences are stacked to form a progradational parasequence set, although not all five parasequences can be distinguished in each well, and the upper five parasequences constitute a retrogradational parasequence set. The vertical stacking of parasequences defines an overall progradational architecture in ‘Series 1’, and a progradational-retrogradational architecture in ‘Series 2’ and ‘Series 3’ (Figs. 3.9a, 3.10a, 3.11).

Although major and minor flooding surfaces have been identified, systems-tract terminology has not been employed, since no sequence boundaries have been confidently identified as discussed above. The lack of evidence for a sequence boundary detailed above may reflect type-2 sequences where rapid subsidence exceeds the rate of sea-level fall preventing sub-aerial
exposure (Posamentier and Vail, 1988). Previous studies focusing on sequence stratigraphy in extensional rift basins have therefore placed the sequence boundary between the highstand and transgressive systems tracts (e.g. Posamentier and Allen, 1993b; Posamentier and Allen, 1993a; Howell and Flint, 1996). However, without a detailed understanding of local variations in sediment supply and timing of activity on different fault segments, it is inappropriate to use systems-tract terminology for stratigraphic sequences deposited during rifting (e.g. Gawthorpe et al., 1994). Furthermore, it is inadequate to explicitly relate the observed stratal stacking patterns to discrete portions of the eustatic sea-level curve; in the context of the rift setting studied here, it is likely that stacking patterns are driven by tectonics (i.e. relative sea-level changes) rather than eustacy.

### 3.6.1 Well correlation panels

The sequence stratigraphic framework and corresponding facies association distributions are illustrated in three well correlation panels (Figs. 3.9, 3.10 and 3.11). Figures 3.9 and 3.10 illustrate wireline-log based well correlations of the entire lower Viking Group (i.e. Heather “A” unit, Krossfjord Formation, Fensfjord Formation and Heather “B” unit). Figure 3.11 illustrates core-log based well correlations of the upper part of the Fensfjord Formation only. The following sub-sections describe the sequence stratigraphic features of each stratigraphic unit.

**Heather “A” unit and Krossfjord Formation.**

The Heather “A” unit and Krossfjord Formation are latest Bathonian in age (165 Ma) and lie above a regionally extensive transgressive surface that caps the Brent Group (MFS J32) (Partington et al., 1993). The top of the Krossfjord Formation does not correspond to a regional maximum flooding surface, but is a minor transgressive surface (Figs. 3.9, 3.10). Wells, cores and wireline logs indicate that the Krossfjord Formation is coarser-grained and has a higher sandstone content than the Fensfjord Formation.
The Krossfjord Formation and Heather “A” unit have been subdivided into seven parasequences that are c. 5-25 m thick. None of the transgressive surfaces that bound these parasequences have been correlated to regional maximum flooding surfaces (cf. Partington et al., 1993) due to lack of robust biostratigraphic age constraints.

Well correlations show thickening of offshore mudstone (Heather “A” unit) towards the north (Figs. 3.9a and 3.10a). This trend corresponds to thinning of the overlying Krossfjord Formation towards the north of the study area. Seismic data show no thickening of the Krossfjord-Heather “A” interval towards or across faults (Figs. 3.9b-c, 3.10b-c).

Overall, the Krossfjord Formation contains relatively thin parasequences (c. 5 m thick) at its base, thicker parasequences (c. 25 m thick) towards its middle of the formation and thinner parasequences (c. 10 m thick) towards its top (Figs. 3.9a and 3.10a). Two distinct facies associations are identified in core in the upper part of the Krossfjord Formation; well 31/5-5 contains coarse-grained, delta front deposits (FA6; Fig. 3.11b), whereas the most northerly well studied, 31/2-4R, contains three wave-dominated shoreface parasequences (each c. 15 m thick) bounded by local transgressive surfaces (Facies Associations 2 and 3; Fig. 3.11a).

**Fensfjord Formation**

The base of the Fensfjord Formation is defined by a minor transgressive surface that is locally associated with a c. 1 m thick transgressive lag, which is rich in bioclastic material and is calcite cemented (wells 31/2-4R and 31/5-5 in Fig. 3.11). West of the Troll Field, towards the basin centre, a tongue of Heather Formation mudstones associated with this transgressive surface separates the Krossfjord and Fensfjord formations (Husmo et al., 2002). The top of the Fensfjord Formation does not correspond to a regional maximum flooding surface, and biostratigraphic data suggest that this lithostratigraphic boundary is diachronous (Fig. 3.9a).
Biostratigraphic data indicate that two regionally significant maximum flooding surfaces, the J42 (early Callovian, 155.5 Ma) and J44 (middle Callovian, 154 Ma) surfaces, are developed within the Fensfjord Formation; these surfaces can be correlated across the Troll Field (Figs. 3.9a, 3.10a). Regional maximum flooding surface J46 (Late Callovian, 152 Ma) has not been identified within the Fensfjord Formation (Geostrat, 2011), but occurs in the overlying Heather “B” unit or lowermost part of the Sognefjord Formation (S. Patruno, pers. comm. 2011). The Fensfjord Formation can be further subdivided into thirteen parasequences, each 10-40 m thick, which are correlated across the Troll Field. The parasequences are of relatively uniform thickness laterally, and well and seismic data indicate that the Fensfjord Formation itself is isopachous (Figs. 3.9, 3.10). In Troll West, the parasequences typically comprise wave-dominated shoreface deposits (Facies Associations 2 and 3) (wells 31/2-4R, 31/2-1 and 31/5-5 in Fig. 3.11). Above maximum flooding surface J44, in an interval that contains abundant core control, wave-dominated shoreface parasequences commonly pass laterally, towards the east, into mixed, wave- and tide-influenced parasequences (Facies Associations 2, 4 and 5) (wells 31/2-3, 31/3-1, 31/6-1, 31/6-5 and 31/6-3 in Fig. 3.11); the significance of this observation is discussed below.

3.6.2 Palaeogeographic reconstructions

Three palaeogeographic maps have been reconstructed from core and wireline-log data (Fig. 3.12). The maps illustrate the limit of facies belts at maximum regression within the ‘series’ that are bounded by regional maximum flooding surfaces (Figs. 3.9a, 3.10a, 3.11). Because there are fewer well penetrations and fewer cored wells, the palaeogeographic reconstructions for the older stratigraphic intervals are less well constrained. Well data indicate no large-scale thickening across faults during the deposition of the Krossfjord and Fensfjord formations (Figs. 3.9b-c, 3.10b-c), and as a result no active structures are indicated in the reconstructions.
Because no fluvial deposits have been identified in core, no fluvial point source(s) of sediment input are shown in the reconstructions. Likewise, evidence for subaerial exposure (e.g. roots, pedogenic alteration) is also lacking in core, and therefore a subaerially exposed coastal plain is not shown in the reconstructions (Fig. 3.12).

**Series 1 – Maximum regression late Bathonian to early Callovian, below MFS J42**

A palaeogeographic reconstruction for maximum regression of the Krossfjord Formation, below the regional maximum flooding surface J42, has been compiled using core data from five wells and wireline-log data from 19 wells (Fig. 3.12c). Deposits of three environments have been recognised for this time period. Wave-dominated shoreface deposits (Facies Associations 2 and 3) are present in the northwestern part of Troll West (well 31/2-4R in Fig. 3.12c) and fluvial-dominated delta front deposits (FA5) occur in the western part of Troll West (well 31/5-5 in Fig. 3.12c). Facies associations occur as linear belts oriented NNE-SSW across the study area, as constrained by the occurrence of offshore deposits (FA1) in uncored wells west of the Troll Field (Figs. 3.9a, 3.10a).

**Series 2 – Maximum regression middle Callovian, below MFS J44**

A palaeogeographic reconstruction for maximum regression of the lower Fensfjord Formation, below the regional maximum flooding surface J44, has been compiled using core data from five wells and wireline-log data from 18 wells (Fig. 3.12b). Wave-dominated shoreface deposits (Facies Associations 2 and 3) are identified in three cored wells in Troll West (wells 31/2-3, 31/2-4R and 31/5-5 in Fig. 3.12b), and these define a facies-association belt that extends across the Troll West Field and into the previously offshore area to the north west of the Troll Field. The facies belt is, therefore, wider than its equivalent in the previous time period (Figs. 3.9a, 3.10a, 3.12c). To the east of this facies-association belt, tide-dominated, wave-influenced embayment deposits (FA5) are recognised in two cored wells in the centre of the Troll Field (wells 31/2-3 and 31/6-1 in Fig. 3.12b). The lateral extent of this facies-association belt further
to the east is not interpreted, due to the absence of core data. Facies-association belts are again orientated N-S.

**Series 3 – Maximum regression middle to late Callovian, below MFS J46**

A palaeogeographic reconstruction for maximum regression of the uppermost Fensfjord Formation, above the regional maximum flooding surface J44, has been compiled using core data from nine wells and wireline-log data from 18 wells (Fig. 3.12a). Wave-dominated shoreface deposits (Facies Associations 2 and 3) are restricted to the western half of Troll West (wells 31/2-1, 31/2-4R and 31/5-5 in Fig. 3.12a). Tide-dominated, wave-influenced embayment deposits (FA5) extend east across the centre of the Troll Field (wells 31/2-3 and 31/3-1 in Fig. 3.12a). Wave-dominated, tide-influenced shoreface deposits (FA4) occupy the entirety of Troll East (wells 31/6-1, 31/6-3, 31/6-5 and 31/6-6 in Fig. 3.12a). Offshore deposits (FA1) are identified to the west of the Troll Field in uncored wells (Figs. 3.9a, 3.10a). Facies association belts are again orientated broadly N-S, but there are lateral variations in both thickness and facies distribution (e.g. the tide-dominated, wave-influenced embayment facies-association belt thins and pinches out towards the south) (Fig. 3.12a).

The palaeogeographic reconstructions highlight temporal and spatial variations in depositional-process regime (e.g. the dominance of wave, tide and fluvial processes) during deposition of the Krossfjord and Fensfjord formations. The western part of each reconstruction consistently shows that wave-dominated shoreface deposits (Facies Associations 2 and 3) form the outer part of the sandstone tongues that pinch-out towards the west and pass basinwards into offshore deposits. The eastern part of each reconstruction indicates a more variable process regime, with significant tidal and fluvial influence. The importance of tidal influence appears to increase through time, becoming progressively more prominent from the Krossfjord Formation to the upper Fensfjord Formation (Figs. 3.11, 3.12), although it should be noted that core data from Troll East Field is only available to constrain the youngest reconstruction (Fig. 3.12a).
addition, the following features occur throughout deposition of the Krossfjord and Fensfjord formations: (1) no subaerially exposed coastal-plain deposits are identified in the study area; (2) facies-association belts are oriented NNE-SSW; and (3) facies-association belts are of similar width across the study area. These various features are addressed by depositional models presented below.

3.7 Discussion: Depositional model for Krossfjord and Fensfjord formations

The well correlations and palaeogeographic reconstructions presented above (Figs. 3.11, 3.12) highlight the complex spatial and temporal distributions of facies-association belts within the Krossfjord and Fensfjord formations of the Troll Field, albeit within a broadly similar range of shallow-marine environments. The facies associations and stratigraphic components recognised in the Krossfjord and Fensfjord formations are similar to those of the overlying Sognefjord Formation (Dreyer et al., 2005), suggesting that the same depositional model(s) may potentially be applicable to all three sandstone-prone formations in the Viking Group. In the sections below the following three key aspects of the sedimentology and stratigraphic architecture of the Krossfjord and Fensfjord formations are discussed: (1) the east-to-west variation in depositional environments across the Troll Field; (2) the vertical increase in abundance of tide-influenced deposits; and (3) the absence of coastal-plain deposits.

3.7.1 East-to-west change from mixed, wave- and tide-influenced deposits to wave-dominated deposits in the upper Fensfjord Formation

‘Series 3’ (MFS J44 – MFS J46), the upper part of the Fensfjord Formation, is characterised by pronounced partitioning between wave-dominated shoreface deposits in the west of the Troll Field and a mixed, wave- and tide-influenced environment in the east of the field (Figs. 3.11a, 3.12a). The change can be explained by spatial variation in depositional process regime (Fig.
3.13a), temporal variation in process regime (Fig. 3.13b), or a combination of the two (Fig. 3.13). In each model, the source(s) of fluvial sediment input to the shoreline is inferred to have been situated to the north of the Troll Field; this is consistent with previous interpretations (Fraser et al., 2002; Husmo et al., 2002; Dreyer et al., 2005) and is supported by the absence of fluvial deposits in core. The shoreline position is interpreted to have been consistently N-S oriented, in agreement with the linear arrangement of facies association belts (Fig. 3.12).

The first depositional model (Fig. 3.13a) illustrates spatial variation in coeval environments. Sediment was supplied from a fluvio-deltaic source in the north, and redistributed through wave-generated, southward-directed longshore currents to form a spit in Troll West. The seaward face of the spit consists of a wave-dominated shoreface. The spit protects landward areas from wave energy, which therefore appear to contain relatively stronger, and perhaps locally amplified, tidal energy in a tide-dominated embayment and back basin setting in Troll East. Similar models have been proposed for the Sognefjord Formation of the Troll Field (Dreyer et al., 2005) and the Cretaceous Frewens Delta, US Western Interior (Willis et al., 1999), both in the context of asymmetrical delta systems (Bhattacharya and Giosan, 2003). The Holocene Maguelone shoreline, southeastern France (Raynal et al., 2009), which is starved of fluvial sediment input, also displays a similar shoreline morphology with sand spits fronting lagoons. However, the area immediately landward of the spit barrier is typically composed of terrestrial vegetation and peat deposits (e.g. Nielsen and Johannessen, 2001), neither of which are identified in cored successions in the Troll Field. Furthermore, foreshore deposits in spit systems commonly contain roots (e.g. Nielsen and Johannessen, 2008), which are also lacking.

The second depositional model (Fig. 3.13b) illustrates temporal variation in environments, with early progradation of a tide-dominated, embayed shoreline to later progradation of a mixed, wave- and tide-influenced shoreface, and finally progradation of a wave-dominated shoreface. A wave-dominated, tide-influenced environment is interpreted to exist during early progradation
due to the wide-shelfal platform in front of the shoreline, which increases tidal resonance and decreases the effect of wave energy through friction (Hubbard et al., 1979; Pugh, 1987; Ainsworth et al., 2011). Tidal embayments are created through topographic restrictions such as barrier islands (e.g. Oertel, 1985). It has also been suggested that inner-shelf deltaic shorelines are irregular and contain abundant topographic restrictions, which may locally amplify tidal range (Plink-Björklund, 2012). With continued progradation, the width of the shelfal platform is decreased and wave processes become increasingly dominant, as illustrated for late progradation (Fig. 3.13b). The gradual change in depositional process regime was driven by progradation of the shallow-marine depositional system, which reflects the interplay between sediment supply, accommodation and basin physiography (Ainsworth et al., 2008). Similar depositional models have been proposed for the Holocene Mekong River Delta (Ta et al., 2002) and Holocene Song Hong (Red River) Delta (Tanabe et al., 2006).

3.7.2 Vertical increase in abundance of tide-influenced deposits

Well correlations of cored wells in the Troll Field highlight an increasing prevalence of tide-influenced deposits vertically through the Krossfjord and Fensfjord formations (Fig. 3.11), although it should be noted that the dataset is biased towards ‘Series 3’ in the upper part of the Fensfjord Formation (MFS J44 – MFS J46). The pattern of upward-increasing tidal influence could record the progressive progradation in ‘Series 1’ of parasequences (Fig. 3.9a) that are constructed via the mechanisms depicted in either depositional model (Fig 3.13). However, the same architecture cannot be invoked for ‘Series 2’ and ‘Series 3’, which each exhibit progradational-to-retrogradational stacking of parasequences (Figs. 3.9a, 3.10a, 3.11). More likely, the upward increase in tidal influence reflects the change from progradational to retrogradational stacking in ‘Series 2’ and ‘Series 3’ (Figs. 3.9a, 3.10a, 3.11), and from progradational to retrogradational stacking of the three ‘series’ (as noted by Steel, 1993; Stewart et al., 1995; Husmo et al., 2002). Numerous studies document the preferential
development and preservation of tidally-influenced deposits in net-transgressive strata (e.g. Nio and Yang, 1991; Sixsmith et al., 2008; Kieft et al., 2011).

### 3.7.3 Absence of coastal-plain deposits

Coastal-plain deposits are absent in cores from the Krossfjord and Fensfjord formations, because either: (1) palaeosols were originally developed in each parasequence and then removed by transgressive erosion (i.e. ravinement); (2) palaeosols were not developed due to forced regression during relative sea level fall; or (3) palaeosols were not developed due to construction of a broad, shallow, subaqueous platform during each parasequence-scale progradational episode.

Transgressive surfaces are identified at parasequence boundaries in both formations (Figs. 3.9a, 3.10a, 3.11) and many are associated with lags that record erosional winnowing of the substrate (e.g. Fig. 3.4i). Transgressive erosion can remove up to 20 m of the substrate (Demarest and Kraft, 1987) with such deep erosion being generally associated with channelised tidal scour. Forced regression during falling relative sea level results in the accumulation of thin, attenuated coastal-plain deposits that are more readily removed by transgressive erosion than the thicker coastal-plain intervals accumulated during normal regression (e.g. Posamentier and Morris, 2000). However, the product of such deep erosion related to channelised tidal scour is a high-relief erosional surface lined by large angular shale clasts (Cattaneo and Steel, 2003). Such features are not identified in core data and therefore argue against this explanation for the absence of coastal-plain deposits.

Forced regression during falling sea level can also result in a complete absence of coastal-plain facies. Three of the eight criteria used to identify forced regressive deposits by Posamentier and Morris (2000) are recognised in the Krossfjord and Fensfjord formations: (1) the absence of
fluvial or coastal-plain facies capping regressive successions, (2) increased average grain size in regressive deposits in a proximal to distal direction due to the cannibalisation and redistribution of older, proximal highstand sediments, and (3) the long distance of regression and anomalously thin character of parasequences due to decreased accommodation during lowering of relative sea-level. Steel (1993) previously used the first and third of these criteria to propose forced regression for the development of parasequences in the Viking Group. However, two criteria identified by Posamentier and Morris (2000) are absent: (1) sharp-based shoreface deposits indicative of erosion of shelf deposits due to the lowering of wave base during falling relative sea-level and (2) a zone of shallow-marine sediment bypass as a result of sea level fall, sometimes expressed as the detachment of lowstand shoreface deposits from their highstand precursor shorefaces. Furthermore, three criteria identified by Posamentier and Morris (2000) are indeterminable and therefore require further detailed analysis of seismic data including: (1) the progressive reduction in relief of clinoforms going from proximal to distal due to progradation into progressively shallower water, (2) seaward dipping bounding surfaces on the top of parasequences due to the lowering of sea level, and (3) the presence of “foreshortened” stratigraphic sections where the decompacted thickness of a coarsening-upwards sequence is less than the palaeowater depth. Therefore conclusive evidence of forced regression appears to be absent, but this may reflect the sparse distribution of core and well data in the Troll Field and surrounding area and the limited analysis to date of clinoform trajectories in seismic data.

An alternative interpretation is that palaeosols were never developed during each progradational episode. This interpretation implies that the area of the Troll Field was subaqueous throughout deposition of the Krossfjord and Fensfjord formations, and that a broad, shallow, subaqueous platform was repeatedly constructed during each parasequence-scale progradational episode. The drowning of a broad, shallow platform during subsequent transgression may also have promoted the development of tidal embayments behind retreating barrier islands and spits. The development of broad subaqueous platforms is common in many
modern deltas that are subject to significant wave and tidal action (e.g. Swenson et al., 2005; Plink-Björklund, 2012), such as the Yangtze Delta, East China Sea (Hori et al., 2002). The development of broad, subaqueous platforms have also been interpreted in a handful of ancient examples, including the Cretaceous Wise Gulch, Berry Gulch and Morapos sandstones, US Western Interior (Hampson et al., 2008b). Furthermore, through the detailed examination of clinoforms imaged in seismic data from the Troll Field, the development of a fully subaqueous deltaic system is proposed to explain the lack of coastal-plain facies in the stratigraphically younger Sognefjord Formation (S. Patruno, pers. comm. 2012).

3.8 Conclusions

The Middle-to-Upper Jurassic Krossfjord and Fensfjord formations are two shallow-marine sandstone tongues that form important secondary reservoirs in the super-giant Troll Field, and are prospective in surrounding areas of the Horda Platform on the eastern margin of the Viking Graben, northern North Sea. Regionally, both formations thin and pinch out into offshore shales of the Heather Formation towards the west, beyond the limit of the Horda Platform. This chapter presents the first detailed sedimentological and stratigraphic analysis of the two formations in the Troll Field, as an aid to predicting reservoir distribution and character. The main conclusions of this core- and wireline-log based analysis of the Krossfjord and Fensfjord formations in the Troll Field are summarised below.

1. Core observations indicate that mixed-influenced deltaic, shoreline and shelf environments were in existence during the Middle Jurassic in the area of the Troll Field. Palaeogeographic reconstructions of maximum regression show facies-association belts of similar width, orientation and distribution. A wide (10-20 km) belt of north-south-trending, wave-dominated shoreface deposits is present in the western part of the Troll Field. The eastern part of the field contains more irregular (0-20 km wide), north-south-trending belts of mixed wave- and tide-
influenced shoreface, tide-dominated embayment, and/or fluvial-dominated delta front deposits. The change from tide-influenced deposits in the east to wave-dominated deposits in the west can be attributed either to spatial variation in depositional process regime within an asymmetrical delta fronted by a spit, or to temporal variation in depositional process regime as the system prograded from a sheltered, inner-shelf location in the east to an exposed, outer-shelf location in the west.

2. Correlation between wells is constrained by field-wide mapping of major seismic reflectors at base-Heather “A”, top-Krossfjord and top-Fensfjord levels, combined with recognition of biostratigraphically distinctive, regional maximum flooding surfaces (J32, J42, J44, and J46) in cored wells. This framework allows recognition of three ‘series’ bounded by the regional maximum flooding surfaces. Analysis of the entire stratigraphic interval, its constituent ‘series’ and their component parasequences indicate relatively uniform thicknesses across the extent of the Troll Field, implying the absence of any major structural control on sedimentation during the Middle Jurassic in this area.

3. Coastal-plain deposits are not identified in the Krossfjord and Fensfjord formations in the Troll Field. The absence may be attributed to the development of a broad, shallow, subaqueous platform across the Troll Field during repeated parasequence-scale regressions, to transgressive erosion at the top of each parasequence, or to forced regression during falling sea level for each parasequence-scale regression. Forced regression is consistent with the thin, laterally extensive character of parasequences in the Viking Group over the Horda Platform and areas adjacent to the west.
3.9 Acknowledgements

Paul Whipp, Theresa Lloyd-Lodden, Ian Sharp, Stefano Patruno, Gavin Elliott, Howard Johnson, Peter Allison and Aruna Mannie and thanked for discussions during the course of this study and Statoil ASA for providing data. Bruce Ainsworth, John Underhill, and an anonymous reviewer are thanked for their insightful and constructive reviews and editorial comments. Partners in the Troll production licenses PL054, PL085, PL085 B & PL085 C (Petoro AS, AS Norske Shell, Statoil ASA, ConocoPhillips Norge & Total E&P Norge AS) are thanked for supporting the provision of data to undertake this study and their permission to publish the results. The views expressed in this chapter are the authors and do not necessarily represent those of Troll license partners. Thanks also to Schlumberger Limited for provision of Petrel seismic and well interpretation software via an academic software donation.
Figure 3.1. (over page) (a) Simplified map of the North Viking Graben highlighting the Horda Platform and the Troll, Brage, Fram and Gjøa fields, which host Krossfjord and Fensfjord reservoirs (modified after Færseth, 1996; Ravnås and Bondevik, 1997; Fraser et al., 2002); (b) simplified palaeoenvironmental map of the northern North Sea during the mid- to late Callovian and deposition of the Fensfjord Formation in the Troll Field (modified after Husmo et al., 2002); and (c) geoseismic profile illustrating the major fault blocks from west to east across the Viking Graben (modified after Husmo et al., 2002). Cross-sections (Figs. 3.1c, 3.2) and maps (Fig. 3.3) are located in Figure 3.1a.
**Figure 3.2.** Middle to Upper Jurassic chronostratigraphic framework for a SW-NE-oriented cross-section through the North Viking Graben and Horda Platform (Fig. 3.1a) (modified after Partington *et al.*, 1993; Stewart *et al.*, 1995; Fraser *et al.*, 2002).
**Figure 3.3.** Simplified map of the Troll Field (Fig. 3.1a) illustrating the distribution of wells available to this study, major faults, and well correlation panels (modified after Færseth, 1996; Ravnås & Bondevik, 1997; Fraser *et al.*, 2002; NPD, 2011).
Figure 3.4. (over page) Core photographs illustrating selected facies of the Krossfjord and Fensfjord formations: (a) Facies A, a fine-grained siltstone with *Terebellina* trace fossils (31/2-4R at 1498 m); (b) Facies C, a fine-grained sandstone with hummocky cross-stratification (31/6-1 at 1529 m) (NPD, 2011); (c) Facies D, a tabular cross-stratified medium- to coarse-grained sandstone (31/5-5 at 1887 m); (d) Facies F, a cross-stratified sandstone with paired mud-draped laminations (31/6-6 at 1723 m); (e) Facies F, a cross-stratified sandstone with bi-directional mud-draped laminations (31/6-6 at 1770 m) (NPD, 2011); (f) Facies G, a sharp based, normally graded, medium-grained sandstone (31/6-5 at 1714 m); (g) Facies H, a fine-grained sandstone with lenticular bedding and evidence of micro-faulting (31/6-1 at 1604.5 m); (h) Facies I, a well sorted, graded, coarse-grained to granular sandstone (31/5-5 at 1866 m) (NPD, 2011); and (i) Facies J, a very poorly sorted coarse-grained to granular sandstone with sharp base (31/2-1 at 1658 m). Photographs are located in the successions illustrated in Figures 3.5-3.7 where possible. The black and white scale bar represents 3 cm.
Figure 3.5. (over page) (a) Key to facies association sedimentary logs and (b) sedimentary log through offshore facies association from well 31/2-4R. Well location shown in Figure 3.3.

Figure 3.6. (page 92) (a) Sedimentary log through wave-dominated lower shoreface, and upper shoreface and foreshore facies associations from well 31/2-3 and (b) sedimentary log through wave-dominated, tide-influenced upper shoreface facies association from well 31/6-5, indicating parasequences bounded by local transgressive surfaces (TS). Well locations shown in Figure 3.3.

Figure 3.7. (page 93) (a) Tide-dominated, wave-influenced embayment facies association sedimentary log from well 31/6-1 and (b) delta front facies association sedimentary log from well 31/5-5 with parasequences bounded by local transgressive surfaces (TS) and basic sequences (Series-1, -2 and -3) bounded by maximum flooding surfaces (J-). Well locations shown in Figure 3.3.

Figure 3.8. (page 94) Wireline-log cross-plots of gamma ray (GR) v. sonic (DT) data and density (RHOB) v. neutron porosity (NHPI) data. The cross-plots illustrate the quantitative log character of the facies associations identified in the Krossfjord and Fensfjord formations: (a) Offshore; (b) Wave-dominated Lower Shoreface; (c) Wave-dominated Upper Shoreface and Foreshore; (d) Wave-dominated, tide-influenced Upper Shoreface; (e) Tide-dominated, wave-influenced Embayment; and (f) Delta Front. The cross-plots are compiled with data from logged core intervals. The cross-plots illustrate the absence of clear differences in quantitative log character between many facies associations, which limits their identification in un-cored wells. The wireline-logs were therefore only used to distinguish sandstone-rich facies associations from mudstone-rich facies associations.
Grain size, sedimentary structures and fossils

Bioturbation index

GR Log (API)
LithDep (m) Facies CLAY SILT SAND GR PBBL
VF F M C VC 30 110

HEATHER “B” UNIT - WELL 31/2-4R

RHOB Log (g/cm³) 2.0 2.6

NHPI Log (p.u.) 0.10 0.45

Fig. 3.4a

Chapter 3

(a) KEY

Lithology
- Sandstone
- Siltstone
- Calcite Cement
- Unconsolidated Core
- No Core
- Graded Bed
- Wood Fragment
- Carbonaceous Material
- Mica
- Calcite Nodule
- Pyrite
- Siderite
+ Glaucite

Structures
- Planar Lamination
- Tabular Cross Bedding
- Low Angle Cross Bedding
- Trough Cross Bedding
- Wavy Laminae
- Symmetrical Ripple
- Asymmetrical Ripple
- Lenticular Bedding
- Sharp Contact
- Undulatory Contact
- Gradational Contact
- Flooding Surface

Bio
- Articulated Bivalve
- Oyster
- Shell Layer
- Shell
- Shell Fragment
- Belemnite
- Chondrites
- Terebellina
- Planolites
- Burrow

(b) HEATHER “B” UNIT - WELL 31/2-4R

Grain size, sedimentary structures and fossils

Bioturbation index

GR Log (API)
LithDep (m) Facies CLAY SILT SAND GR PBBL
VF F M C VC 30 110

RHOB Log (g/cm³) 2.0 2.6

NHPI Log (p.u.) 0.10 0.45

Page 91
Figure 3.9. (over page) (a) Northwest-southeast oriented (depositional strike-oriented) well-log correlation panel across Troll Field (see Fig. 3.3 for location), using the MFS J46 as a flattened datum surface; (b) corresponding, uninterpreted composite seismic section; and (c) geoseismic interpretation along well-log correlation panel.

Figure 3.10. (page 97) (a) Northeast-southwest oriented (depositional dip-oriented) well-log correlation panel across Troll West Field (see Fig. 3.3 for location), using the MFS J46 as a flattened datum surface; (b) corresponding, uninterpreted composite seismic section; and (c) geoseismic interpretation along well-log correlation panel.

Figure 3.11. (page 98) (a) Northwest-southeast oriented (depositional strike-oriented) and (b) northeast-southwest (depositional dip-oriented) oriented core-log correlation panels across Troll Field (see Fig. 3.3 for location), using MFS J46 as a flattened datum surface.
**KEY**
- Heather Formation
- Krossfjord and Fensfjord formations
- Progradational or Retrogradational parasequence
- Stacking patterns

![Diagram](image)

**NE** 8,000 m 9,200 m **SW**

- **Series 1**
- **Series 2**
- **Series 3**

**Diagram Notes:**
- Sognefjord Fm
- Heather “B” unit
  - MFS J40
- Fensfjord Fm
  - MFS J44
  - MFS J42
- Krossfjord Fm
- Heather “A” unit
- MFS J32
- Brent Gp

**Well Logs:**
- 31/2-3
- 31/2-1
- 31/5-5

**Time (TWT ms):**
- 1450
- 1550
- 1650
- 1750
- 1850
- 1950
- 2050

**Chapter 3**
Figure 3.12. (over page) Facies-association-belt extents at maximum regression in the Troll Field as suggested by well correlation panels (Figs. 3.9a, 3.10a, 3.11) during: (a) middle to late Callovian times (between MFS J44 and MFS J46); (b) middle Callovian (between MFS J42 and MFS J44); and (c) late Bathonian to early Callovian (below MFS J42).

Figure 3.13. (page 101) Block diagrams of depositional models illustrating the lateral change from mixed, wave- and tide-influenced deposits to wave-dominated shoreface deposits from east to west across the Troll Field. Two end-member models are proposed: (a) spatial variation in depositional environments with a wave-driven spit system fronting a tide-influenced back basin; and (b) temporal variation in depositional environments with an early tide-dominated, wave-influenced embayment and shoreface developed on the inner-to-middle shelf \((t = 1)\), evolving into a wave-dominated shoreface as it progrades to the outer shelf \((t = 2)\). Coastal-plain deposits are absent in core, implying that they were either not deposited due to development of a broad subaqueous delta platform, or they were removed by later transgressive erosion; the latter is implied in the block diagrams.
(a) Series 3 - Middle to Late Callovian, below MFS J46

(b) Series 2 - Middle Callovian, below MFS J44

(c) Series 1 - Late Bathonian to Early Callovian, below MFS J42

KEY
- Offshore (FA1)
- Wave-dominated shoreface (FA2-3)
- Wave-dominated, tide-influenced shoreface (FA4)
- Tide-dominated, wave-influenced embayment (FA5)
- Delta-front (FA6)
- Cored Well
- Un-cored Well
- Well Correlation

20 km
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<th>Krossfjord Formation (m)</th>
<th>Length of core logged (m)</th>
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<td>85 (45%)</td>
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<td>228 (23%)</td>
<td>29 (0%)</td>
<td>28.15</td>
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**Table 3.1.** Inventory of core data from the lower Viking Group in the Troll Field. Unit thickness is shown in metres and the percentage of core recovered is shown in brackets.
**Facies Description**

**Thickness in core** 1-10 m

**Bioturbation and ichnotaxa**

**Wireline-log Character Interpretation**

<table>
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<tr>
<th>Facies</th>
<th>Description</th>
<th>Bioturbation and ichnotaxa</th>
<th>Wireline-log Character Interpretation</th>
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<tbody>
<tr>
<td>A1</td>
<td>Very fine-grained, dark grey to very dark green, siltstone. Moderately well sorted with sub-rounded to sub-angular grains. Packages of increased siltstone content are identified in discrete coarsening-upwards or fining-upwards sections (&lt;0.5 m). Mica and carbonaceous fragments common throughout. Pyrite and glauconite are rare. Bedding is absent.</td>
<td>BI = 5-6</td>
<td>GR = 58-80 API; DT = 60-140 µs/ft; RHOB = 2.2-3.6 g/cm³; NPHI = 0.30-0.56 p.u.</td>
</tr>
<tr>
<td>A2</td>
<td>Fine-grained, dark grey to medium grey/brown siltstone with rare fine-grained sandstone. Moderately well sorted with sub-rounded to sub-angular grains. Occasional coal fragments (1-2 mm) and glauconite. Nodular calcite cement is rare and siderite nodules are occasionally present. Bedding is largely absent due to bioturbation, apart from rare parallel laminations.</td>
<td>BI = 5-6</td>
<td>GR = 48-76 API; DT = 85-150 µs/ft; RHOB = 2.1-2.4 g/cm³; NPHI = 0.11-0.39 p.u.</td>
</tr>
<tr>
<td>B</td>
<td>Very fine-grained, heterolithic, dark grey sandstone. Moderately well sorted with sub-angular grains. Mica and carbonaceous fragments are common. Siderite is occasional and pyrite is rare. Nodular calcite cement (&lt;50 cm thick), wood and carbonaceous material are rare. Bedding is largely absent due to bioturbation, with very rare centimetre-scale low angle cross lamination.</td>
<td>BI = 5</td>
<td>GR = 65-70 API; DT = 75-160 µs/ft; RHOB = 2.2-2.7 g/cm³; NPHI = 0.35-0.56 p.u.</td>
</tr>
<tr>
<td>C</td>
<td>Coarsening-upwards very fine-to-coarse-grained, heterolithic, dark grey sandstone. Moderately well sorted with sub-angular grains. Mica and carbonaceous fragments are common. Bedding is largely absent due to bioturbation, with very rare centimetre-scale low angle cross lamination.</td>
<td>BI = 5</td>
<td>GR = 40-100 API; DT = 50-120 µs/ft; RHOB = 2.1-2.4 g/cm³; NPHI = 0.08-0.42 p.u.</td>
</tr>
<tr>
<td>D</td>
<td>Medium to coarse-grained, light grey, planar laminated to trough and tabular cross-laminated sandstone. Quartz grains are typically clear to milky, occasionally pinkish. Well sorted and may be poorly sorted in thin bed (&lt;10 cm). Bedding is rarely fully developed, with occasional shell fragments in chaotic orientation.</td>
<td>BI = 0-1</td>
<td>GR = 45-150 API; DT = 50-150 µs/ft; RHOB = 2.1-2.8 g/cm³; NPHI = 0.14-0.50 p.u.</td>
</tr>
<tr>
<td>Facies</td>
<td>Description</td>
<td>Thickness in core</td>
<td>Bioturbation and ichnotaxa</td>
</tr>
<tr>
<td>--------</td>
<td>-------------</td>
<td>------------------</td>
<td>--------------------------</td>
</tr>
<tr>
<td>F</td>
<td>Fine- to medium-grained, medium grey sandstone. Well sorted with sub-rounded grains. Mica and siltstone appear draped along laminae highlighting planar- to trough-cross stratified sandstone beds. Paired drapes are evident. Dip directions are commonly bi-directional. Thin shelled bivalves are present.</td>
<td>1-3 m</td>
<td>BI = 1-5 (Palaeophycus, Planolites, Skolithos)</td>
</tr>
<tr>
<td>G</td>
<td>Sharp based, fining upwards, fine- to coarse-grained, medium grey, moderately- to poorly-sorted sandstone. Base may contain very coarse-grained sandstone with rounded grains. Abundant mica and carbonaceous fragments, along with intense bioturbation towards the top. Very rare high angle cross-lamination to parallel lamination in thin beds (&lt;0.1 m thick) particularly near top. Only associated with Facies F.</td>
<td>0.5-1 m</td>
<td>BI = 4-5 (Palaeophycus, Planolites, Skolithos)</td>
</tr>
<tr>
<td>H</td>
<td>Fine-grained, grey sandstone interspersed throughout with lenses of siltstone and mudstone. Well-sortd with sub-rounded to sub-angular grains. Mica and carbonaceous material are common. Internal structures include ripple cross lamination, isolated ripple sets and discontinuous laminae which are evident in ripple troughs (millimetre-scale). In siltstone-rich parts lenticular and flaser bedding are evident. Wavy bedding is also present. Planar-laminated fine-grained sandstone beds occur rarely. Micro faults are common. Body and trace fossils are typically absent.</td>
<td>2-3 m</td>
<td>BI = 0</td>
</tr>
<tr>
<td>I</td>
<td>Sharp based, fining upwards, granular- to fine-grained, light grey, well sorted sandstone. Grains are sub-rounded to sub-angular, and quartz is very common. Carbonaceous material is occasional with very rare mica. Generally appears structureless with rare planar-parallel laminations defined by higher siltstone content. Body fossils are absent. Bioturbation is variable with fine-grained sandstone being slightly bioturbated and coarse-grained sandstones being apparently structureless.</td>
<td>0.1-4 m</td>
<td>BI =0-3</td>
</tr>
<tr>
<td>J</td>
<td>Dark green-to-grey, poorly sorted, matrix supported, medium- to coarse-grained sandstone. Beds are sharp based and occasionally calcite cemented. Blue- to smoky- to rose-quartz grains. Matrix contains granules and pebbles, glauconite, carbonaceous material and high concentrations of bioclastic fragments and intact shells. Beds are internally chaotic and structureless. Grain size typically fines upwards at centimetre scale. Occurs at distinct stratigraphic position marked by shifts from proximal to distal facies associations.</td>
<td>0.1-1 m</td>
<td>BI = 0</td>
</tr>
</tbody>
</table>

**Table 3.2.** Summary of facies (A-J) in the Krossfjord and Fensfjord formations. Core photographs for facies A, C, D, F, G, H, I and J are presented in Figure 3.4.
CHAPTER 4

Constraining uncertainty in interpretation of seismically imaged clinoforms in deltaic reservoirs, Troll Field, Norwegian North Sea: Insights from forward seismic models of outcrop analogues

4.1 Abstract

Forward seismic modelling of outcrop analogues has been used to characterise the seismic expression of clinoforms in different deltaic depositional environments, and thus constrain uncertainty in interpretation of intra-reservoir clinoforms imaged in seismic data from the Troll Field, Norwegian North Sea. Three outcrop analogues from the Western Cretaceous Interior Seaway, US were studied to quantify the geometry, distribution and lithological character of clinoforms in fluvial-dominated and mixed-influence deltaic deposits. Outcrop-derived geometrical data were calibrated to sedimentological and petrophysical data from the Krossfjord and Fensfjord formations in the Troll Field, and then used to create a suite of forward seismic models for comparison with real seismic reflection data from the Troll Field. Clinoforms were imaged in the forward seismic models where they were: (1) spaced wider than the tuning thickness (>10 m); (2) marked by pronounced interfingering of facies associations with different acoustic properties; and/or (3) lined by relatively thick (>50 cm) carbonate-cemented layers. However, where clinothems are thinner than the vertical resolution limit of seismic data, destructive interference occurred creating misleading geometrical relationships. Furthermore, the ability to image clinoforms is dependent on: (1) the frequency of the seismic wavelet; (2) the overburden velocity; and/or (3) the acoustic impedance contrast at the boundary between the overburden and the clinoform-bearing target. The established methodology has allowed characterisation of deltaic clinoformal architectures in reservoir seismic data from the Troll
Field, and has facilitated a more robust interpretation by bridging the critical gap in resolution between well and seismic data.

### 4.2 Introduction

Palaeoseaward-dipping stratal surfaces termed clinoforms (sensu Rich, 1951), which bound packages of rock termed clinothems (sensu Rich, 1951), are an inherent component of shallow-marine deposits. Their geometry and distribution are a reflection of depositional process regime and stratigraphic architecture (e.g. Helland-Hansen and Hampson, 2009), and clinoforms can act as barriers or baffles to flow where they are lined with mudstone, mica or carbonate cement (Howell et al., 2008a; Jackson et al., 2009; Sech et al., 2009). Interpretation of clinoform distribution and the lithological character of clinothems can therefore be essential to constrain stratigraphic architecture for exploration purposes, and to maximize production from shallow-marine reservoirs. However, whilst clinoforms can be characterised in the subsurface through core and wireline-log data, the limited distribution of wells restricts the level of detail and degree of confidence with which seismically imaged clinoformal architectures can be interpreted.

An example of siliciclastic, clinoform-bearing deltaic reservoirs are the Middle-to-Upper Jurassic Krossfjord and Fensfjord formations, Troll Field, offshore Norway (Figs. 4.1, 4.2) (Holgate et al., 2013). Within these formations, many more clinoform-related lithological breaks (vertical spacing of c. 2-4 m), defined by landward facies shifts within upwards-shallowing successions interpreted to represent transgressive surfaces, can be identified in core and wireline-log data than are “detected” (when a stratigraphic layer thickness is less than a quarter of the wavelength and therefore the thickness of the layer cannot be determined) or “resolved” (when a stratigraphic layer thickness is greater than a quarter of the wavelength and therefore discrete acoustic responses occur at the top and base of the layer) in seismic reflection data (Fig. 4.2d).
This discrepancy raises three important questions: (1) what combination of lithological attributes generates a clinoformal seismic reflection?; (2) how does clinothem thickness interact with seismic resolution to generate a seismically imaged stratigraphic architecture?; and (3) how sensitive are the seismic responses to subtle changes in the geometry and distribution of clinoforms? Furthermore, this disparity in resolution inhibits the ability to confidently interpret the seismically imaged architectures away from well control and, therefore, implement them in, for example, reservoir models. In order to bridge this critical gap in scale between stratigraphic architectures resolved in seismic reflection and well datasets, and to produce a consistent geological interpretation for these reservoir-bearing formations, forward seismic modelling of outcrop analogues is needed.

Forward seismic modelling of outcrops is a relatively novel method by which stratigraphic architectures and lithology distributions observed in outcrop can be combined with subsurface petrophysical properties, from core measurements, density and velocity logs, to model their seismic response under reservoir conditions (e.g. Rudolph et al., 1989; Biddle et al., 1992; Helland-Hansen et al., 1994; Anselmetti et al., 1997; Schwab et al., 2007; Bakke et al., 2011). The resultant forward seismic model can then be compared to seismic reflection data from the subsurface target, thus providing insights into sub-seismic sedimentological and stratigraphic complexity, which may be important for reservoir characterisation.

In this study a range of outcrops that document deltaic depositional environments similar to those interpreted in the Krossfjord and Fensfjord formations are identified, and their facies architectures are used as a basis to evaluate the sensitivity of the modelled seismic response to different lithology distributions and stratigraphic architectures (Fig. 4.1). The outcrop analogues are located in the US Cretaceous Western Interior, in Utah and Wyoming: (1) the Ferron Sandstone Member, deposited by a fluvial-dominated delta with ascending regressive shoreline trajectory (Anderson et al., 2004); (2) the Panther Tongue, deposited by a fluvial-
dominated delta with descending regressive shoreline trajectory (Enge et al., 2010a); and (3) the Chimney Rock Tongue, deposited by a mixed-influenced delta with variable shoreline trajectory (Plink-Björklund, 2008). The range of depositional environments and stratigraphic architectures represented by these analogues reflects uncertainty in the character of the subsurface reservoir, and allows the sensitivity of the modelled seismic expression to various sedimentological and stratigraphic parameters to be evaluated.

The aims of this chapter are three-fold: (1) to produce a range of forward seismic models for the three outcrop analogues highlighted above; (2) to understand how the variation in the geometry, distribution, scale, spacing of clinoform surfaces, and the lithological composition of the clinothems that they bound, affects the seismic response, and how this understanding can help differentiate between depositional environments; and (3) more specifically, to use these models, and the architectures and geometries that they contain, to provide calibration of the seismic reflection data from the Krossfjord and Fensfjord formations in the Troll Field. The work presented herein, along with the generic methodology described, is also directly applicable to many other deltaic reservoir sandstones, for which the outcrops described above may also be considered to be sedimentological analogues.

4.3 Geological setting

4.3.1 Krossfjord and Fensfjord formations, Troll Field

The Middle-to-Upper Jurassic Krossfjord and Fensfjord formations are secondary reservoir targets in the producing Troll oil and gas field located on the Horda Platform, eastern margin of the Viking Graben, northern North Sea (Fig. 4.2a, b). The formations are composed of shallow marine-to-deltaic sandstones, which were sourced from the Norwegian mainland to the east and which pinch out basinward into offshore shales of the Heather Formation to the west. In
combination, the two formations define an overall regressive-to-transgressive wedge (Fig. 4.2c; Steel, 1993; Stewart et al., 1995). This wedge comprises multiple regressive-transgressive tongues, each of which contains a single set of seismically imaged clinoforms (Fig. 4.2d).

Sedimentological analysis of the Krossfjord and Fensfjord formations, based on nine wells in the Troll Field, has identified six facies associations: (1) offshore, (2) wave-dominated lower shoreface, (3) wave-dominated upper shoreface, (4) wave-dominated, tide-influenced upper shoreface, (5) tide-dominated, wave-influenced embayment, and (6) fluvial-dominated delta front. The facies associations are stacked vertically into upward-shallowing successions (parasequences) bounded by transgressive surfaces (see Fig. 11 in Holgate et al., 2013). Palaeogeographic reconstructions of maximum regression in each formation (e.g. Fig. 4.2b) show a complex arrangement of depositional environments. A 10-20 km wide belt of north-south-trending, wave-dominated shoreface deposits is present in the western part of the Troll Field. The eastern part of the field contains more irregular, 0-20 km wide, north-south-trending belts of mixed wave- and tide-influenced shoreface, tide-influenced embayment, and/or fluvial-dominated delta front deposits. The lateral change from tide-influenced deposits in the east to wave-dominated deposits in the west can be attributed either to spatial variation in depositional process regime within an asymmetric delta fronted by a spit, or to temporal variation in depositional process regime as the system prograded from a sheltered, inner-shelf location in the east to an exposed, outer-shelf location in the west in each regressive-transgressive tongue (see Fig. 13 in Holgate et al., 2013). However, core data are sparse and the near-indistinguishable character of sandstone facies associations in wireline logs (Holgate et al., 2013) restricts the ability to discriminate between the depositional models proposed above and limits the confidence with which seismically imaged architectures can be interpreted.
### 4.3.2 Outcrop analogues from the Cretaceous Western Interior Seaway of North America

The Western Interior Seaway of North America developed in a broad, asymmetric, north-south trending, intracratonic basin, which formed in response to subduction-related subsidence and thrust-sheet loading in the Sevier Orogenic Belt along its western margin during the Late Cretaceous (Pang and Nummedal, 1995; Liu and Nummedal, 2004). A number of asymmetric, westward-thickening sediment wedges fill the basin, and are principally composed of siliciclastic sediment eroded from the Sevier Orogenic Belt. Each wedge comprises a series of basinward-thinning, regressive-to-transgressive tongues composed of coastal plain and shallow-marine strata. Both the sediment wedges and their constituent tongues formed as a result of the complex interplay between eustatic sea-level variations, laterally variable tectonic subsidence, and variations in sediment supply (Krystinik and DeJarnett, 1995). The clinoform-bearing deltaic units of the Ferron Sandstone Member, Panther Tongue, and Chimney Rock Tongue occur as tongues within the basin-fill strata (Fig. 4.3). Based on previous studies, the characteristic sedimentological facies of these units in the studied localities are summarised in Table 4.1, and form the basis for assigning rock properties taken from analogous subsurface facies to the forward seismic models. In the following section a summary is provided of the sedimentology, stratigraphy and palaeogeography of the three units with more detailed accounts given by Anderson et al. (2004), Ryer and Anderson (2004), and van den Bergh and Garrison (2004) for the Ferron Sandstone Member; by Enge et al. (2010a), and Olariu et al. (2010) for the Panther Tongue; and by Plink-Björkland (2008) for the Chimney Rock Tongue.

**Ferron Sandstone Member**

The Turonian-Coniacian Ferron Sandstone Member (Fig. 4.3b, d) is a fluvial-dominated deltaic deposit with an ascending regressive shoreline trajectory (sensu Helland-Hansen and Hampson, 2009). The unit is exposed along the Coal Cliffs, south-central Utah (Fig. 4.4b) (e.g. Anderson et
In the Coal Cliffs, the Ferron Sandstone Member forms the eastward-thinning wedge of the “Last Chance” delta system (Cotter, 1976), which consists principally of eight progradationally-to-aggradationally stacked, regressive-transgressive tongues (Kf-1 to Kf-8 of Anderson and Ryer, 2004). Regionally, the “Last Chance” delta system built out towards the northeast, although the local progradation direction of individual delta lobes was highly variable (e.g. Anderson and Ryer, 2004; Deveugle et al., 2011). The work herein is only concerned with the Kf-1 and Kf-2 tongues, which are exposed in the vicinity of Ivie Creek (Fig. 4.4b). Here, Kf-1 consists primarily of river-dominated delta-front deposits containing steeply dipping (10-15°) clinoforms defined by thin beds of mudstone and carbonaceous debris. Local progradation occurred from the southeast into an embayment (Anderson et al., 2004) (labelled 1 in Fig. 4.3b). Kf-2 consists principally of wave-modified, river-dominated delta-front deposits that contain more shallowly dipping (1-10°) clinoforms. The Kf-1 and Kf-2 tongues are separated by the Sub-A coal zone (Anderson et al., 2004).

**Panther Tongue**

The Lower Campanian Panther Tongue (Fig. 4.3c, d) is a fluvial-dominated deltaic deposit with descending regressive shoreline trajectory (sensu Helland-Hansen and Hampson, 2009), and is exposed along the eastern flank of the Wasatch Plateau and in the western part of the Book Cliffs in central Utah (Fig. 4.4c) (e.g. Young, 1955; Posamentier and Morris, 2000; Enge et al., 2010a). The Panther Tongue is composed of fluvial-dominated delta-front deposits capped by a transgressive lag (Posamentier and Morris, 2000; Hwang and Heller, 2002; Olariu et al., 2005; Howell et al., 2008a; Enge et al., 2010a; Enge et al., 2010b; Olariu et al., 2010). The proximal part of the Panther Tongue, exposed around the town of Helper, is the focus of this study (Fig. 4.4c). Here, southward-dipping (1-5°) clinoforms are defined by thin, discontinuous siltstones separating sandstone beds.
Chimney Rock Tongue

The Lower Campanian Chimney Rock Tongue (Fig. 4.3c, d) is a mixed-influenced deltaic deposit with variable shoreline trajectory (sensu Helland-Hansen and Hampson, 2009), which is exposed in a dip-oriented outcrop belt that straddles the border between Utah and Wyoming, near Flaming Gorge Reservoir (Fig. 4.4d) (Plink-Björklund, 2008). This work focuses on the lower part of the Chimney Rock Tongue, which is composed of wave-dominated delta-front deposits truncated by an incised-valley fill. These wave-dominated delta deposits are characterised by eastward-dipping (1-3°) clinoforms. Fluvial-dominated mouth-bar and delta-front deposits containing clinoforms occur locally, and distributary channels also locally erode into the tops of clinoform sets. The deposits are deeply eroded (up to 30 m) by a subaerial unconformity at the base of the overlying incised valley, which is locally marked by roots and calcite concretions.

4.4 Dataset and methodology

Forward modelling of seismic reflection data can be used to quantitatively analyse the seismic imaging of geological structures. A forward seismic model combines petrophysical properties (density and p-wave velocity) from a subsurface target with an interpreted lithology distribution and architecture from an outcrop analogue. The resulting synthetic seismic section is then compared directly to the real seismic reflection data from the subsurface target, and comparison of lithology distributions and stratigraphic architectures is undertaken. The method has previously been applied successfully to shallow-water carbonate (e.g. Rudolph et al., 1989; Biddle et al., 1992; Campbell and Stafleu, 1992; Stafleu and Schlager, 1993; Stafleu et al., 1994; Anselmetti et al., 1997) and deep-water siliciclastic depositional systems (e.g. Abreu et al., 2003; Sullivan et al., 2004; Schwab et al., 2007; Bakke et al., 2008; Falivene et al., 2010). However, forward seismic modelling has rarely been used to study shallow-marine depositional systems (e.g. Helland-Hansen et al., 1994; Hodgetts and Howell, 2000), and has never been applied at the
scale of shoreline clinoforms. The method is especially adept at resolving geological detail that occurs beneath the typical vertical resolution of seismic reflection data at typical reservoir depths, a significant problem when attempting to use seismically imaged clinoforms to characterise shallow-marine siliciclastic environments, stratigraphic architecture and lithological composition.

A four step workflow to conduct forward seismic modelling, described by Bakke et al. (2011), is utilised:

1. Define missing subsurface properties;
2. Identify and collect data from suitable outcrop analogues;
3. Assign appropriate petrophysical properties to the analogue models; and
4. Compare forward seismic model with the subsurface target.

Having identified the geological features and properties to be investigated in the models (step 1 of Bakke et al., 2011), in this case the geometry, spacing and lithological expression of reservoir-scale clinoforms, the next two steps (steps 2 and 3 of Bakke et al., 2011) are expanded upon below, to explain how they have been implemented in this work. The final step is expanded upon in the discussion section.

### 4.4.1 Identify and collect data from suitable outcrop analogues

As discussed above, three outcrop analogues have been identified that, in combination, represent similar depositional environments to those of the Troll Field reservoirs, and that contain prominent clinoforms (Fig. 4.1, Table 4.1). Specifically, six outcrop 2D sections were chosen for the Ferron Sandstone Member, and three outcrop 2D sections were chosen for both the Panther Tongue and Chimney Rock Tongue, oriented parallel (i.e. dip-oriented) and orthogonal (i.e. strike-oriented) to the mean palaeocurrent or progradation direction (Table 4.2). Detailed measured sections and photo panoramas were taken from each analogue at
particular localities in which clinoforms are well exposed (Fig. 4.4). Bedding diagrams were constructed for each photo panorama, and calibrated against measured sections (Fig. 4.5a, b; all outcrop analogue bedding diagrams and measured sections in Appendix 2). Each bedding diagram was then flattened on an approximately palaeohorizontal datum surface, and perspective effects in the original photo panorama were removed by re-scaling the thicknesses along vertical sections of c. 10 m horizontal spacing in order to match thicknesses recorded in measured sections or in laser-surveyed “pseudo logs” (Fig. 4.5c). Although less accurate than Light Detection and Ranging (LiDAR) data (e.g. collected for the Ferron Sandstone Member and Panther Tongue by Enge et al., 2010a), this approach provides sufficient accuracy (0.5-1 m) relative to the constraints provided by seismic reflection datasets that typically have a vertical resolution of >15 m at reservoir depth (Fig. 4.6).

Numerical data were extracted from the constructed bedding diagrams (Fig. 4.5c) to characterise the different outcrop analogue datasets. These include palaeocurrent data, number of clinoforms recognised (Table 4.2), mean clinothem thickness (calculated by averaging the thickness of the clinothem at intervals defined by the pseudo logs in bedding diagrams, e.g. Fig. 4.5c) (labelled $C_t$ in Fig. 4.7a, b), mean clinoform dip (calculated by averaging the dip of the entire clinoform exposed in outcrop section) (labelled $C_d$ in Fig. 4.7a, c), and clinoform foreset length (labelled $C_l$ in Fig. 4.7a, d). The data highlight characteristic differences in clinoform geometry and clinothem thickness dependent on depositional environment. The fluvial-dominated depositional environments of the Kf-1 tongue within the Ferron Sandstone Member (Fig. 4.8a, b) and the Panther Tongue (Figs. 4.5c, 4.8c) contain a greater number of clinoforms compared to wave-modified and wave-dominated depositional environments, and are short (<200 m), steep (<17°), and are typified by thin clinotoths (<2 m) (Fig. 4.7). Wave-modified deposits of the Kf-2 tongue within the Ferron Sandstone Member (Fig. 4.8d, e) and the wave-dominated deposits of the Chimney Rock Tongue (Fig. 4.8f, g) contain fewer clinoforms, which
are longer (< 400 m), shallower (< 10°), and bound thicker clinothems (< 8 m) than their fluvial-dominated counterparts (Fig. 4.7).

4.4.2 Assign appropriate petrophysical properties to the analogue models

Acoustic properties from the subsurface target need to be assigned to lithology distributions derived from the outcrop analogues. Shale proportions observed along the measured sections were used to create a $V_{\text{shale}}$ (volume of shale) log, which is used as a proxy for a gamma-ray log, in order to calibrate the outcrop sections against wireline-log data from the Troll Field (Fig. 4.6). Bulk density and p-wave velocity logs from the calibrated subsurface intervals were used to create the forward seismic models. In order to allow effective comparisons to be made between models, wireline-log data from one well (31/2-4R) were used to populate all models. This well was chosen because it contains the greatest coverage of core from the Krossfjord and Fensfjord formations, and therefore affords the greatest confidence in extracting suitable intervals for calibration with outcrop data. All intervals extracted from this well were located in the water leg of the field, so that the effect of fluid content on the p-wave velocity log can be discounted (Asquith and Krygowski, 2004).

Clinoform surfaces were extracted from the corrected bedding diagrams (Figs. 4.5c, 4.8) and imported into the forward seismic model builder (Compound Earth Simulator; Petersen, 1999). The architectures documented at outcrop are modified to enable comparison with the Troll Field reservoir. For example, the regressive-transgressive tongues studied in the outcrop analogues are thinner than those interpreted in the Troll Field, so all studied analogues are scaled (with no vertical exaggeration) to enable direct comparison of outcrop-derived and subsurface architectures (e.g. the example presented in Figure 4.5 was scaled up by a factor of 5; Fig. 4.9a). The isotropic scaling was applied to both clinoform surfaces and lithological boundaries, in order to maintain the stratigraphic and facies architectures observed at outcrop.
Several bedding diagrams were also cropped to remove the effect of present-day surface erosion (Chimney Rock Tongue CR2, CR3; Kf-2 tongue of the Ferron Sandstone Member IV1, IV2; Kf-1 tongue of the Ferron Sandstone Member JP1, JP2; Panther Tongue PT3; shown as areas with reduced colour saturation in Figure 4.8). Lithologies between clinoform surfaces were assigned petrophysical properties (gamma ray, density, p-wave velocity) from well 31/2-4R (Fig. 4.9b-d).

Petrophysical properties from the Heather “A” and “B” units, which lie respectively below and above the Krossfjord-Fensfjord interval, were used to populate underburden and overburden strata in the forward seismic models. Both of these units are assumed to be composed of flat-lying, parallel-bedded strata of uniform thickness. Furthermore, to limit the unintentional effect of the overburden on seismic reflections in the modelled clinoform sets, underburden and overburden units are assumed to have uniform petrophysical properties and are conformable above and below the modelled interval. The Heather “A” unit has a relatively high p-wave velocity response, resulting in an acoustic impedance increase at the contact between the base of the Krossfjord-Fensfjord interval and the top of the Heather “A” unit. The Heather “B” unit has a relatively low p-wave velocity response and a slightly higher acoustic impedance compared to the reservoir interval, resulting in an acoustic impedance decrease at the contact between the base of the Heather “B” unit and the top of the Krossfjord-Fensfjord interval.

The quality of the modelled seismic section is dependent on the overburden p-wave velocity and the dominant frequency (centre frequency) of the seismic wavelet. These parameters are a function of depth and can therefore be varied to understand the effect of burial depth on the modelled seismic response. The overburden p-wave velocity was calculated using sonic-log data from well 31/2-4R in the Troll Field (see Fig. 4.2b for location), and the dominant frequency of the wavelet was extracted from the Troll Field seismic reflection data, by generating frequency spectra using 100 traces surrounding the path of this well. In order to extract these two
parameters at different depths, the seismic and wireline-log data were subdivided into stratigraphic intervals defined by mapped seismic horizons, from the Nordland Group (top Pliocene) to the Statfjord Formation (Upper Triassic). Seismic reflection data quality beneath the top-Statfjord Formation horizon is poor, and therefore arbitrary horizons were created at increments of 400 m in deeper strata. Dominant frequencies and overburden velocities were calculated within the stratigraphic packages defined by the seismic horizons. Overburden p-wave velocity and the dominant frequency of the wavelet were calculated for depths in the middle of each stratigraphic interval (Fig. 4.10a-c). The resulting values are averages down the vertical borehole of well 31/2-4, and do not represent lateral variations in overburden lithology. The dominant wavelet frequencies extracted from the seismic data decrease with depth, due to attenuation of seismic energy (Fig. 4.10b). A linear regression between the dominant frequencies and overburden velocities in the seismically mapped stratigraphic intervals shows a strong correlation between the two parameters ($R^2 = 0.91$; Fig. 4.10d). Thus three pairs of values of the parameters were chosen for sensitivity testing, as outlined below, for depths that represent average values at well 31/2-4R for shallow burial (780 m), intermediate burial, as for the Krossfjord-Fensfjord reservoir interval (1640 m), and deep burial (2840 m).

After distributions of petrophysical properties (e.g. Fig. 4.9c, d) and appropriate seismic parameters (Table 4.3) were generated, elastic wave modelling was then used to create the synthetic seismic sections (cf. Petersen, 1999). This was achieved using a 2D convolution that simulates a 2D migrated, stacked seismic section, using a 2D resolution parameter applied to the reflection series (Toxopeus et al., 2003). A zero-phase Ricker wavelet was used in reverse polarity ("European" format, cf. Brown, 2004), to be consistent with the Troll Field seismic reflection data. A peak (black) represents a negative acoustic impedance boundary (decrease in acoustic impedance), and a trough (red) represents a positive acoustic impedance boundary (increase in acoustic impedance) (e.g. Fig. 4.6).
4.5 Results

Two suites of forward seismic models are presented below. The first suite (Suite A) is designed to investigate how, why and what controls seismic imaging of clinoforms, and is composed of models that test the sensitivity of the seismic response to the following variables: (1) overburden velocity and seismic frequency, (2) clinoform scale and clinothem thickness, (3) acoustic properties of overburden, and (4) thickness of carbonate-cemented layers along clinoforms. These models use lithology distributions and stratigraphic architectures from a single outcrop analogue bedding diagram (Panther Tongue section PT2; Figs. 4.9, 4.5c), in order to understand the sensitivity of the seismically imaged geometries to the various input parameters. The second suite of models (Suite B) was constructed from all the outcrop analogues (Fig. 4.8). These models highlight different seismically imaged features that can be used to help interpret depositional environments and stratigraphic architectures in seismic reflection data.

4.5.1 Sensitivity tests (Suite A)

Effects of overburden velocity and seismic frequency

Three models were generated to simulate different burial depths, and to test the impact of overburden velocity and dominant wavelet frequency variations on the seismic imaging of clinoforms (Fig. 4.10d; Table 4.3): shallow burial (780 m deep, 55 Hz centre frequency; Fig. 4.11d); intermediate burial, as for the Troll Field reservoir (1640 m deep, 49 Hz centre frequency; Fig. 4.11e); and deep burial (2840 m deep, 39 Hz centre frequency; Fig. 4.11f).

A higher wavelet-centre frequency results in a narrower wavelet, which increases the vertical resolution (Avseth et al., 2005). Consequently, a greater number of clinoforms are imaged as seismic reflections in the shallow-burial model (55 Hz) (Fig. 4.11d). Furthermore, diachronous
lithological boundaries, where seismic reflections crosscut clinoform surfaces following facies association boundaries, are also clearly imaged and associated with high-amplitude reflections. Diachronous lithological boundaries are also imaged on the intermediate-burial model (49 Hz); there are however fewer reflections and they are of lower amplitude than in the shallow-burial model (labelled 1 in Fig. 4.11d, e). Both of these models also contain a number of discontinuous reflections that appear to represent the inter-fingering between mouth bar and proximal delta front facies associations along clinoforms (labelled 3 in Fig. 4.11d, e). In contrast, in the deep-burial model (39 Hz), only gross boundaries between lithological units (e.g. from mouth bar to proximal delta front facies associations) are imaged, and the internal lithological heterogeneity, which is documented at outcrop, is not reflected in the strength, geometry or number of individual clinoform reflections (labelled 1 in Fig. 4.11f).

In all the models, reflections tend to terminate before the top or base of the clinoform. A greater number of reflections represent a thick clinothem (c. 10 m) in the north of the models, where clinothem thickness is greater than tuning thickness (labelled 2a in Fig. 4.11d, e). However, as the thickness of the clinothem decreases towards the south, reflections thin and become discontinuous (labelled 2b in Fig. 4.11d, e). Towards the south, in the shallow-burial model (55 Hz), reflections thin and may become discontinuous (labelled 3 in Fig. 4.11d). In the intermediate-burial model (49 Hz), reflections abruptly terminate, which may incorrectly suggest that clinoforms are discontinuous (labelled 3 in Fig. 4.11e). The thinning and abrupt terminations of reflections in both models are due to the downdip termination of fingers of mouth bar facies association into proximal delta front facies association. None of these subtle stratigraphic features are imaged in the deep-burial model (39 Hz) (Fig. 4.11f).

The apparent geometries of, and geometrical relationships between, clinoforms change greatly with wavelet frequencies. The geometry of clinoforms is accurately imaged at moderate-to-high centre frequencies (49 and 55 Hz; Fig. 4.11d, e). This is highlighted by the onlap termination
exhibited updip in both 55 and 49 Hz models but not in the 39 Hz model (labelled 4 in Fig. 4.11d-f). In contrast, the low centre frequency (39 Hz) model images the broad dipping clinoform geometries only, with little indication of changes in dip along the length of, or between, clinoforms (Fig. 4.11f). The steep angle (c. 10°) at which the clinoforms terminate updip (Fig. 4.11a) are not imaged, but instead, shallowly dipping toplap geometries are apparent (labelled 5 and 6 in Fig. 4.11d-f). Therefore the overall reflection geometry of the clinoforms exhibits a more sigmoidal profile (sensu Mitchum et al., 1977a) in the forward seismic model than the tangential oblique profile (sensu Mitchum et al., 1977a) observed at outcrop.

Effect of clinoform scale and clinothem thickness

A stratigraphic layer thinner than one quarter of the wavelength of the seismic wavelet will not be fully resolved in seismic reflection data (Widess, 1973). To assess the effect of clinothem thickness on the modelled seismic response, and whether the scaling applied to the forward seismic models of outcrop analogues is appropriate, three models have been created: (1) a model in which the horizontal and vertical dimensions are the same as the outcrop analogue (Fig. 4.12a, d); (2) a model with horizontal and vertical dimensions scaled to replicate clinothem thickness in the Troll Field (Fig. 4.12b, e); and (3) a model in which the dimensions are first scaled to replicate clinothem thickness in the Troll Field, and then the number of clinoforms is increased to maintain the clinothem thickness of the outcrop analogue (Fig. 4.12c, f). Note that clinoform geometry is the same in all three models.

The outcrop-scale model contains little detail apart from a thinning of the peak reflection, which represents the thinning of the mouth bar facies-association belt from south to north (labelled 1 in Fig. 4.12a, d). The number, geometry, and spacing of clinoforms, and the lithological content of clinothems are not resolved, because the height of the outcrop analogue is beneath seismic resolution (<10 m). The increase in thickness from the outcrop-scale model (c. 15 m) to the
Troll-Field-scale model (c. 70 m) results in a greater number of reflections, and improved imaging of internal features (Fig. 4.12b, e). Maintaining the outcrop-scale clinothem thickness within the Troll-Field-scale model has little effect on the resultant seismic model compared to the Troll-Field-scale model, apart from a slight dimming of amplitude (Fig. 4.12c, f), suggesting that the scaling applied to the outcrop models is realistic. Amplitude dimming occurs due to destructive interference of multiple acoustically-thin beds as shown by the presence of low-amplitude, vertically-oriented lineaments in the seismic model (labelled 1 in Fig. 4.12c, f).

**Effect of acoustic properties of overburden strata**

The acoustic-impedance properties of the overburden have a direct impact on the proportion of acoustic energy reaching the reservoir. Thin-bedded or heterogeneous overburden, characterised by multiple layers of varying acoustic impedance, can cause wave scattering and attenuation (Widmaier et al., 1996). Further effects of heterogeneous overburden include focusing of the wave field, transmission losses, surface multiples, and anelastic attenuation (Martinez, 1993; Avseth et al., 2005). It is therefore important to demonstrate the effect of variable overburden properties, calibrated for the Troll Field, on the modelled seismic response.

Three models were generated, each with a different scenario of acoustic impedance in the overburden: (1) a model with relatively heterogeneous overburden extracted from the Heather “B” unit in well 31/2-4R in the Troll Field (Fig. 4.13a, d); (2) a model with a homogeneous, acoustically fast overburden (Fig. 4.13b, e); and (3) a model in which the overburden has the same density and sonic properties as the top of the reservoir, such that there is little acoustic variation at the contact between reservoir and overburden (Fig. 4.13c, f).

As would be expected, an acoustically homogeneous overburden results in forward seismic models containing no reflections in the overburden (Fig. 4.13e, f). In fact, the homogeneous or heterogeneous composition of the overburden appears to have little effect on the resultant
seismic models at the target depth (Fig. 4.13d-f). Furthermore, the effect of any acoustic-impedance contrast between the overburden and the reservoir is only apparent within the uppermost 10 m of the reservoir interval (Fig. 4.13e, f). For example, a homogeneous overburden with acoustic properties significantly different to those of the reservoir interval results in thick, laterally continuous, high-amplitude reflections at the reservoir-overburden contact, which obscure reflections in the uppermost 10 m of the reservoir interval (Fig. 4.13b, e). Conversely, a homogeneous overburden with acoustic properties similar to those of the reservoir interval results in greater detail being imaged in the uppermost 10 m of the reservoir interval (Fig. 4.13c, f). A greater proportion of reflections in the upper part of the reservoir interval appear laterally continuous in the model with homogeneous overburden (labelled 1 and 2 in Fig. 4.13f) than in the model with heterogeneous overburden (labelled 1 and 2 in Fig. 4.13d), although the steeply dipping geometries of the updip parts of the clinoforms are not resolved in either model. The depth to which the acoustic properties of the overburden influence imaging of the reservoir interval depends on the frequency of the seismic wavelet, rather than the style of stratigraphic layering, at least in the architecturally simple overburden.

**Effect of thickness of carbonate-cemented layers along clinoforms**

Carbonate cement is common in shallow-marine and deltaic reservoirs, and typically occurs as concretionary layers along clinoforms (e.g. Morad et al., 2000; Braaksma et al., 2006; Morris et al., 2006; Hampson et al., 2008b). Cemented layers along clinoforms can act as barriers to flow, which may reduce hydrocarbon recovery (e.g. Jackson et al., 2009). It is therefore important to understand if these cements can be detected or resolved in seismic reflection data.

Three models were generated, each containing laterally continuous, carbonate-cemented layers along clinoforms. These cemented layers vary in thickness in the three models: (1) 10 cm (Fig. 4.14a, d); (2) 50 cm (Fig. 4.14b, e); and (3) 100 cm (Fig. 4.14c, f).
The overall configuration of reflections in the forward seismic models containing carbonate-cemented layers (Fig. 4.14d-f) remains unchanged compared to the corresponding model in which they are absent (Fig. 4.11e). However, the presence of thick (100 cm) cemented layers appears to increase the amplitude of reflections within the model (Fig. 4.14f), because the cemented layers have very different acoustic properties to the subjacent units (e.g. Braaksma et al., 2006). The enhanced reflectivity of the thick carbonate-cemented layers also results in steep geometries being accurately imaged in the upper parts of the clinoforms (labelled 1 in Fig. 4.14f). Conversely, the presence of thin (10 cm) cemented layers appears to have the opposite effect, and reduces the reflectivity and lateral continuity of reflections due to destructive interference (Fig. 4.14d).

4.5.2 Models of outcrop analogues (Suite B)

Fluvial-dominated deltas

Two bedding diagrams from the Kf-1 tongue of the Ferron Sandstone Member (JP1, JP2; Fig. 4.4b), and three from the Panther Tongue (PT1, PT2, PT3; Fig. 4.4c) were used to create forward seismic models of representative fluvial-dominated deltaic clinoform sets (Figs. 4.11, 4.15). Based on their regional context, the Kf-1 tongue of the Ferron Sandstone Member records regression during rising relative sea-level (i.e. with an ascending regressive shoreline trajectory) (Anderson et al., 2004), and the Panther Tongue records regression during falling relative sea-level (i.e. with a descending regressive shoreline trajectory) (Posamentier and Morris, 2000; Enge et al., 2010a). This difference in stratigraphic architecture is reflected in the local preservation of clinoform topsets in the upper part of the Kf-1 tongue (Fig. 4.15a), whereas the Panther Tongue is invariably top-truncated (Figs. 4.11a, 4.15c, d).

Within the modelled fluvial-dominated deltaic tongues, the up-scaled clinothems are consistently thinner (c. 5-10 m) than the tuning thickness (c. 10 m), resulting in fewer beds and
clinoforms being highlighted by reflections in the seismic models (Fig. 4.15). The steeply
dipping geometries (up to 17°; Fig. 4.7c), which are characteristic of fluvial-dominated delta-
front clinoforms (Driscoll and Karner, 1999), are generally imaged in the forward seismic
models. Where lithological variation is limited, reflections have low amplitude (labelled 1 in Fig.
4.15h). Conversely, where interfingering of facies associations is pronounced (Fig. 4.15a, b),
reflections have a relatively high-amplitude (Fig. 4.15e, f). High amplitudes are further
exaggerated where carbonate cementation occurs along clinoforms (Fig. 4.15e, f), which results
in localised imaging of clinoform topsets (e.g. labelled 2 in Fig. 4.15f). The cumulative effect is
that reflections are weaker in the more homogeneous, un-cemented upper part of the Panther
Tongue, where the steeply dipping upper parts of clinoforms are truncated at the top of the
tongue (Figs. 4.11e, 4.15g, h; compare with Kf-1 tongue of the Ferron Sandstone Member in Fig.
4.15e, f). Where clinothems thin below the seismic detection limit towards their toes, they
decrease in amplitude (e.g. labelled 3 in Fig. 4.15f) and become discontinuous (labelled 2a, 2b in
Fig. 4.15h).

Wave-influenced and wave-dominated deltas
Two bedding diagrams from the Kf-2 tongue of the Ferron Sandstone Member (IV1, IV2; Fig.
4.4b), and two from the Chimney Rock Tongue (CR2, CR3; Fig. 4.4d), were used to create
forward seismic models of representative wave-modified and wave-dominated deltaic
clinoform sets, respectively (Fig. 4.16). Based on its regional context, the Kf-2 tongue of the
Ferron Sandstone Member records regression during rising relative sea-level (i.e. with an
ascending regressive shoreline trajectory) (Anderson et al., 2004). The shoreline trajectory of
the studied clinoform set in the Chimney Rock Tongue is not clear, because it is top-truncated by
an incised valley (Plink-Björklund, 2008).

Within the modelled wave-modified and wave-dominated deltaic tongues, up-scaled clinothem
thickness (c. 10-20 m) is generally greater than tuning thickness (c. 10 m), and therefore all
clinoforms are represented by reflections in the forward seismic models. Similarly the shallow-dipping geometries of clinoforms (up to 10°; Fig. 4.7c), which is characteristic of wave-dominated environments (Bhattacharya, 2006), are apparent in all models. In thick (c. 20-30 m), homogeneous successions of the same facies association (e.g. labelled 1 in Fig. 4.16b-h), reflections tend to be weak and have low amplitudes. Conversely, thinner (c. 3-7 m) clinothems of the sandstone-dominated, upper shoreface facies association correspond to higher amplitude reflections, probably due to constructive interference as clinothems thin (e.g. labelled 2 in Fig. 4.16a, e; b, f; d, h). The erosional bases of channelised fluvial sandstones are defined by strong peaks (Fig. 4.16a, e; d, h). Thick sections of the channelised sandbodies can also contain isolated trough reflections, which pinch out where the sandbody thins laterally beneath seismic resolution.

4.6 Discussion

Having presented and interpreted two suites of forward seismic models, the variation in clinoform geometry, distribution, scale, and spacing, and clinothem lithological composition and how it affects their seismic expression is now discussed. An assessment is also made on how the forward seismic models constrain the character of the Krossfjord-Fensfjord reservoir interval in the Troll Field (Fig. 4.2c, d).

4.6.1 Under what conditions are clinoforms imaged, resolved and detected in seismic reflection data?

In the forward seismic models, clinoforms are imaged where they are: (1) spaced more widely than tuning thickness (c. 9.7 m in most of the cases presented); (2) marked by interfingering of facies associations with different acoustic properties; and/or (3) lined by relatively thick (>50 cm) carbonate-cemented layers.
However, clinoform geometries, and especially their termination characteristics, are obscured by strong impedance contrasts at overlying boundaries (e.g. Fig. 4.13a-e; labelled 3 in Fig. 4.16c, g), which can modify the amplitude, frequency and phase of the seismic response (e.g. Meckel Jr and Nath, 1977; Malme et al., 2005; Luo et al., 2007; He, 2008). Weak impedance contrasts result in reflections having a closer representation to the real clinoformal geometries (Fig. 4.13c, f).

Furthermore, clinoforms are less likely to be resolved where clinothems thin to the detection limit of the seismic data (quarter wavelength of the seismic wavelet). Where this occurs, they may only be represented by reflections of limited continuity and weak amplitude (e.g. labelled 1 in Fig. 4.15c, g; labelled 2a and 2b in Fig. 4.15d, h) (Meckel Jr and Nath, 1977), or by apparent onlap and downlap if they juxtapose lithologies of similar acoustic impedance (e.g. Avseth et al., 2005). This effect is exacerbated by the presence of thin (10 cm) carbonate-cemented layers along clinoforms which can cause destructive interference that results in lower continuity of reflections (e.g. Fig. 4.14a, d) (cf. Fournier and Borgomano, 2007). Furthermore, where facies associations are uniform across a clinoform surface, discrete clinoform surfaces are not imaged; instead a single reflection may represent multiple clinoforms, which under-represents the number of clinoforms present within the lithological unit (e.g. labelled 2b in Fig. 4.11e; labelled 1 in Fig. 4.15d, h). Similarly, in certain models with interfingering facies associations, reflections also occur at diachronous facies-association boundaries; these reflections may be relatively flat-lying and they do not therefore conform with clinoforms (e.g. labelled 1 in Fig. 4.15b, f), contrary to the common assumption that seismic reflections represent chronostratigraphic surfaces (Vail and Mitchum, 1977). This effect is more pronounced at low frequencies (e.g. labelled 1 in Fig. 4.11f) than at high frequencies (e.g. Fig. 4.11d), as noted in previous studies (e.g. Zeng and Kerans, 2003; Zeng, 2013).
4.6.2 What criteria allow clinoforms from different depositional environments to be diagnosed?

The lithological composition, geometry, size, and architectural context of clinoforms are governed by the interplay between wave-, tide- and river-processes (e.g. Bhattacharya, 2006), such that seismically imaged clinoforms have been used to infer depositional environment (Mitchum et al., 1977a; Berg, 1982; Bhattacharya, 2006).

In the studied outcrop analogues, fluvial-dominated delta-front clinoforms are characterised by a high degree of facies interfingering (Fig. 4.15), small spacing (Fig. 4.7b), and steep dips (up to 17°; Fig. 4.7c). In the forward seismic models, these characteristics result in reflections with more variable amplitudes (e.g. labelled 1 in Fig. 4.15a, e) and less lateral continuity (e.g. labelled 1 in Fig. 4.15c, g). Individual clinoforms are not represented by individual reflections, because most clinoforms occur beneath seismic resolution as a result of their relatively small spacing. This is compounded in the bottom set of clinoforms where both clinothem thickness and lithological contrast is drastically reduced compared to the foreset region (e.g. labelled 2 in Fig. 4.15a, e). Seismically imaged clinoforms have shallower dips than observed at outcrop as a result of bed thickness tuning (e.g. labelled 5 in Fig. 4.11e; labelled 2 in Fig. 4.15b, f), which can create apparent onlaps (labelled 6 in Fig. 4.11d, e) (Avseth et al., 2005). The clinoforms have oblique-tangential, oblique and parallel geometries, which are considered to be characteristic of fluvial-dominated delta-front clinoforms (Mitchum et al., 1977a).

In contrast to the fluvial-dominated clinoforms, wave-modified and wave-dominated delta-front clinoforms in the studied outcrop analogues are laterally continuous and exhibit shallow dips (<8°; Fig. 4.16a-d), the former reflecting the large spacing of clinoforms and the laterally extensive character of facies-association belts juxtaposed across them (Fig. 4.16a-d). The thickness of each clinothem is typically greater than the vertical seismic resolution, resulting in
laterally extensive reflections that display only minor amplitude variations (Fig. 4.16e-h). The clinoforms do not appear to exhibit the sigmoidal to shingled profile noted as characteristic in previous studies (Mitchum et al., 1977a), perhaps due to the limited lateral extent of the outcrop-based bedding diagrams containing these clinoforms.

4.6.3 What implications do the seismic models have for interpretation of Troll Field seismic reflection data?

To assess the impact of the forward seismic models on understanding the significance of seismically imaged clinoforms in the Troll Field, four comparable seismic sections from the Troll Field are presented below (Fig. 4.17) and discussed. Two sections are oriented in the regional dip-direction (SW-NE) through wells 31/2-3 and 31/5-5 (Fig. 4.17a, d). Two are oriented in regional strike-direction (SE-NW) through wells 31/2-3 and 31/5-5 (Fig. 4.17b, e) to indicate the three-dimensional aspect of clinoforms (cf. Table 4.2). Both regional dip-oriented sections are interpreted to illustrate clinoform and lithological distributions (Fig. 4.17c, f). All seismic sections have been interpreted to show the Brent Group, Krossfjord and Fensfjord formation tops, along with maximum flooding surfaces that have been correlated with the aid of core, wireline-log and biostratigraphic data (Holgate et al., 2013) (Fig. 4.2c, d). Three seismic units have been defined using the ‘series’ nomenclature applied to regressive-transgressive tongues in overlying units of the Troll Field (e.g., Dreyer et al., 2005) and extended to the Krossfjord and Fensfjord formations by Holgate et al. (2013): ‘Series 1’ is bounded by the MFS J32 and MFS J42 at its base and top respectively and comprises fluvial-dominated delta front deposits around well 31/5-5 and wave-dominated shoreface deposits around well 31/2-3; ‘Series 2’ is bounded by the MFS J42 and MFS J44 at its base and top respectively and comprises wave-dominated shoreface deposits; and ‘Series 3’ is bounded by the MFS J44 and MFS J46 at its base and top respectively and comprises wave-dominated shoreface deposits around well 31/5-5 and tide-dominated, wave-influenced embayment deposits around well 31/2-3 (Fig. 4.2b, c). Carbonate
cements have been identified throughout core data from the Troll Field (Holgate et al., 2013). Each ‘series’ or regressive-transgressive tongue contains a particular style of clinoform reflection, and is described and interpreted below using observations from forward seismic model results.

‘Series 1’, composed principally of the Krossfjord Formation, contains two distinct styles of reflection geometries. The seismic sections which intersect well 31/2-3 (Fig. 4.17a, b) contain relatively high-amplitude, discontinuous (<200 m length; <25 m height), dipping (<7°) reflections with a broadly sigmoidal profile (labelled 1 in Fig. 4.17a, b). The dip-oriented section (Fig. 4.17a) contains more laterally continuous reflections than the strike-oriented section (Fig. 4.17b). Conversely, the seismic sections which intersect well 31/5-5 (Fig. 4.17d, e) are defined by minor dipping reflections at the base (labelled 1 in Fig. 4.17d, e) and predominantly by laterally extensive, high-amplitude, flat reflections towards the top (labelled 2 in Fig. 4.17d, e). More pronounced reflection discontinuity is noted in the strike-oriented section (Fig. 4.17e).

‘Series 2’, a thin unit composed of the lower Fensfjord Formation, shows similar reflection styles in all of the seismic sections from the Troll Field (Fig. 4.17a, b, d, e), independent of orientation. The sections contain faint indications of short (<100 m in length; <10 m in height), dipping (<6°) reflections (labelled 2 in Fig. 4.17a, b, labelled 3 in Fig. 4.17d, e). Major transgressive surfaces are shown by laterally extensive, high-amplitude reflections.

‘Series 3’, composed principally of the upper Fensfjord Formation, shows similarity in reflection style between the two sets of seismic sections. The seismic section intersecting well 31/2-3 shows laterally discontinuous (<50 m in length; <5 m in height), thick reflections with strong amplitude and dip (<6°) variations (labelled 3 in Fig. 4.17a, b) with little variation between dip- and strike-oriented sections. The seismic section that intersects well 31/5-5 contains reflections that are short (<50 m in length; <10 m in height), discontinuous, thin, and steeply dipping
(<11°), with low amplitudes (labelled 4 in Fig. 4.17d, e). These are situated at the base and top of ‘Series 3’ around a laterally extensive, high-amplitude flat reflection (labelled 5 in Fig. 4.17d, e).

The sigmoidal, high-amplitude reflections within ‘Series 1’ (labelled 1 in Fig. 4.17a, b) suggest clinothem thickness is greater than the detection limit of the seismic data. This is characteristic of the wave-modified and wave-dominated environments studied in outcrop (Fig. 4.16). The sigmoidal reflections are relatively flat-lying in the strike-oriented section, which is also the case in the strike-oriented section of the wave-modified Kf-2 tongue of the Ferron Sandstone Member (e.g. Fig. 4.16b, f). Amplitude variations of these reflections are significant (e.g. labelled 1 in Fig. 4.17a, b) perhaps suggesting the juxtaposition of facies between clinoforms (e.g. Fig. 4.16a, e) or the presence of discontinuous carbonate cemented layers (e.g. Fig. 4.15b, f; labelled 1 in Fig. 4.17c) as observed in the wireline-log data (e.g. “spikes” in DT log; Fig. 4.17a, b). Dimmer reflections occur in the lower parts of the clinoform set (labelled 4 in Fig. 4.17a, b), and may represent shale-dominated prodelta deposits or down-dip thinning of clinothems (e.g. labelled 2 in Fig. 4.16c, g; labelled 4 in Fig. 4.17c). The upper parts of the clinoforms are more reflective, probably due to the juxtaposition of sand and the finer-grained lithology above (e.g. labelled 2 in Fig. 4.16b, f; labelled 1 in Fig. 4.17c). The upper parts of the clinoforms also appear to onlap onto one another in both orientations (labelled 5 in Fig. 4.17a, b). As has been argued above, these onlap geometries in the topset of clinoforms are ambiguous and could be due to the interaction with an acoustically-disparate overburden (e.g. Fig. 4.13b, e) or the degree of facies interfingering (e.g. labelled 5 in Fig. 4.11d-f). Therefore, using these geometries to conduct shoreline trajectory analysis (sensu Helland-Hansen and Hampson, 2009) may lead to erroneous interpretations about short-term changes in sea-level during progradation.

Relatively flat-lying, high-amplitude, laterally extensive reflections are recognised in ‘Series 1’ (labelled 2 in Fig. 4.17d, e), ‘Series 2’ (labelled 6 in Fig. 4.17a, b, d, e), and ‘Series 3’ (labelled 5 in
Fig. 4.17d, e). These flat-lying reflections are not a result of the seismic section being in strike orientation through the clinoforms as both dip and strike sections show the same reflection characteristics. Laterally continuous reflections are commonly identified in forward seismic models of wave-modified and wave-dominated deltaic units (e.g. labelled 2 in Fig. 4.16a, e, and labelled 1 in Fig. 4.16c, g). Troll Field core data are interpreted to record a wave-dominated shoreface environment (Fig. 4.17), which strengthens this interpretation in the case of ‘Series 2’ and ‘Series 3’. However, the lack of dipping reflections in ‘Series 1’ may indicate the thickness of clinothems is beneath seismic resolution and therefore only coarse lithological variations are imaged, a common feature of the fluvial-dominated Panther Tongue (e.g. labelled 1 in Fig. 4.11f; labelled 2 in Fig. 4.17f). This interpretation is reinforced in other ‘series’ by the observation of very short, thin reflections related to the larger reflection (labelled 2 in Fig. 4.17a, b; labelled 7 in Fig. 4.17d, e), which may represent a thickening of the beds to and slightly above the seismic detection limit (e.g. labelled 1 in Fig. 4.15c, g) or minor lithological heterogeneities (e.g. labelled 2 in Fig. 4.16c, g; labelled 2 in Fig. 4.17c). Furthermore, the laterally extensive reflection exhibits step-like notches resulting in steep changes in dip over a very short distance (labelled 3 in Fig. 4.17d, e; labelled 7 in Fig. 4.17a, b). Similar features are noted in forward seismic models where clinoforms apparently onlap onto one another (e.g. labelled 6 in Fig. 4.11d, e). In the example of the Panther Tongue PT2 forward seismic model in Suite A, the bed thickness was great enough to allow a visible distinction between individual clinoforms. Nevertheless, it is envisaged that thinner clinothems of more homogeneous lithological composition would limit this distinction and therefore allow a blending of the reflection to represent numerous clinoforms (e.g. labelled 7 in Fig. 4.11e; labelled 3 in Fig. 4.17f). Therefore, caution should be applied in interpreting this reflection as it could represent either numerous shallow-dipping, closely-spaced, clinoforms of limited lateral extent, or a single, laterally extensive clinoform.

Steeply dipping, low amplitude, discontinuous reflections are identified in ‘Series 3’ only (labelled 3 in Fig. 4.17a, b; labelled 4 in Fig. 4.17d, e). A very similar style of reflection is
exhibited in the Panther Tongue forward seismic model in Suite A (e.g. labelled 3 in Fig. 4.11e), as a result of thin lithological heterogeneities between clinoforms. Whilst the reflections are steeper in the Troll Field seismic section, a similar heterogeneous lithological composition can be inferred as suggested by sedimentological analysis of core data (e.g. labelled 3 in Fig. 4.17a, c). The difference in reflection thickness between the two well locations is probably related to the variation in thickness of clinothems and the heterogeneities they contain. Furthermore, similar to the forward seismic model, the reflections may represent more than one clinoform. However, this interpretation is complicated by the possible presence of thin (<50 cm) carbonate cements which creates similar discontinuous, "jittery" reflections (e.g. labelled 1 in Fig. 4.14a, d; labelled 4 in Fig. 4.17f).

4.7 Conclusions

Forward seismic modelling has been applied to understand the nature of seismically imaged clinoforms in shallow-marine and deltaic reservoirs. Comparison of a suite of models based on outcrop analogue data, which describe the geometry, distribution and lithological character of clinoforms and the clinothems that they bound in wave- and fluvial-dominated deltaic deposits, enables the seismic response of these parameters to be diagnosed. Analysis of the forward seismic models highlights the following:

- The frequency of the seismic wavelet and the overburden velocity greatly impact the level of detail resolved in the seismic models. High wavelet frequency and low overburden velocity result in partial resolution of clinoform surfaces and interfingering of facies associations with different acoustic properties. Low wavelet frequency and high overburden velocity result in resolution of only gross lithological variations.

- Clinoforms are imaged where they are: (1) spaced wider than tuning thickness; (2) marked by interfingering of facies associations with different acoustic properties; and/or (3) lined by relatively thick (>50 cm) carbonate-cemented layers. Where the
clinothem thins to the detection limit of the seismic data, they may be represented by reflections of limited continuity and weak amplitude, or by apparent onlap and downlap if they juxtapose lithologies of similar acoustic impedance. Thin (10 cm) carbonate-cemented layers along clinoforms can cause destructive interference that results in lower continuity of reflections. Clinoform geometries are also obscured by strong impedance contrasts at overlying boundaries.

- Closely spaced clinoforms, which typify the studied fluvial-dominated deltas, are unlikely to be fully resolved, with many clinothems less than a quarter of the wavelength in thickness. Conversely, widely spaced clinoforms with shallow dips, which typify the studied wave-dominated deltas, are more likely to be resolved, and associated facies composition may be inferred.

These insights have been applied to the characterisation of clinoform-bearing stratigraphic architectures in the Krossfjord and Fensfjord formations, Troll Field, offshore Norway. Forward seismic modelling thus helps to bridge the gap in spatial scales and sampling volumes between well and seismic reflection data, by constraining uncertainty in interpretation of seismically imaged clinoforms that may influence sweep and recovery during production.

### 4.8 Acknowledgements

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Dr. Michael L. Sweet, Dr. Jory A. Pacht, Frances Plants Whitehurst and an anonymous reviewer for their insightful and constructive reviews and editorial comments. Partners in the Troll production licenses PL054, PL085, PL085 B & PL085 C (Petoro AS, Statoil Petroleum AS, A/S Norske Shell, Total E&P Norge AS & ConocoPhillips Skandinavia AS) are thanked for supporting the provision of data to undertake this study and their permission to publish the results. The views expressed in this chapter are the authors and do not necessarily represent those of Troll license partners. Thanks also to Schlumberger Limited for provision of Petrel software via an academic software donation, and to Blueback Reservoir for provision of Blueback Geophysics Toolbox.
The Panther Tongue, a fluvial-dominated delta with descending regressive shoreline trajectory (Enge et al., 2010a)

The Ferron Sandstone Member, a fluvial-dominated delta with ascending regressive shoreline trajectory (Anderson et al., 2004)

The Chimney Rock Tongue, a mixed-influenced delta with variable shoreline trajectory (Plink-Bjorklund, 2008)

Krossfjord and Fensfjord formations

Figure 4.1. Tripartite classification of deltas based on depositional process regime (Galloway, 1975), highlighting the positions of the Krossfjord and Fensfjord formation reservoirs, and the potentially analogous Ferron Sandstone Member, Panther Tongue, and Chimney Rock Tongue.

Figure 4.2. (over page) (a) Simplified palaeogeographic map of the north Viking Graben highlighting the location of the Troll Field on the Horda Platform and nearby fields that host Krossfjord and Fensfjord reservoirs (after Holgate et al., 2013); (b) facies-association-belt extents at maximum regression from the Fensfjord Formation in the Troll Field during middle to late Callovian times (between maximum flooding surface (MFS) J44 and MFS J46) highlighting position of well 31/2-4R (after Holgate et al., 2013); (c) Middle to Upper Jurassic chronostratigraphic framework of studied reservoir interval for a SW-NE-oriented cross-section through the North Viking Graben and Horda Platform (Fig. 4.2a) (after Holgate et al., 2013); and (d) an interpreted seismic section (located in Fig. 4.2b) highlighting picked seismic horizons and clinoforms within the Krossfjord and Fensfjord formations. The overlaid well 31/2-3, with density (RHOB) and p-wave velocity (DT) wireline-logs, highlight the clinoform breaks recognised through core and wireline-log analysis.
Active fault
Erosion or nondeposition
Nonmarine (coastal plain and alluvial plain)
Shallow marine (foreshore and upper shoreface)
Offshore marine (lower shoreface and offshore shelf)
Volcanic rocks (tuffs, extrusives and intrusives)

Dominant source directions
Field containing Krossfjord and Fensfjord formations

KEY

Erosion or nondeposition
Nonmarine (coastal plain and alluvial plain)
Shallow marine (foreshore and upper shoreface)
Offshore marine (lower shoreface and offshore shelf)
Volcanic rocks (tuffs, extrusives and intrusives)
Active fault
Dominant source directions
Field containing Krossfjord and Fensfjord formations

Offshore (Facies Association (FA) 1)
Wave-dominated shoreface (FA2-3)
Wave-dominated, tide-influenced shoreface (FA4)
Tide-dominated, wave-influenced embayment (FA5)

Cored Well 31/2-4R (Fig. 4.6)
Figure 4.3. (over page) (a) Location map highlighting the position of the Western Interior Seaway during the Late Cretaceous (after Kauffman and Caldwell, 1993); (b) palaeoenvironmental map during middle Turonian time highlighting shoreline position during deposition of the Ferron Sandstone Member (after McGookey et al., 1972); (c) palaeoenvironmental map during early Campanian time highlighting shoreline position during deposition of the Panther Tongue and Chimney Rock Tongue (after McGookey et al., 1972); and (d) stratigraphic columns for Upper Cretaceous strata highlighting the position of: the Ferron Sandstone Member within the Mancos Shale, in the East Central Wasatch Plateau (stratigraphic column labelled 1 corresponds to location labelled 1 in Fig. 4.3b) (after Fouch et al., 1983; Lawton, 1986); the Panther Tongue within the Star Point Sandstone, in Price Canyon (stratigraphic column labelled 2 corresponds to location labelled 2 in Fig. 4.3c) (after Fouch et al., 1983; Hettinger and Kirschbaum, 2002); and the Chimney Rock Tongue within the Rock Springs Formation, in the Greater Green River Basin (stratigraphic column labelled 3 corresponds to location labelled 3 in Fig. 4.3c) (after Gill et al., 1970; Roehler, 1990; Krystinik and DeJarnett, 1995). The study focuses on small sections of two tongues in the Ferron Sandstone Member, and also the Panther Tongue, and Chimney Rock Tongue.

Figure 4.4. (page 139) (a) Map of Utah, USA, with surrounding states, highlighting locations of study areas; (b) Ferron Sandstone Member study area around Ivie Creek; (c) Panther Tongue study area around the town of Helper; and (d) Chimney Rock Tongue study area to the east of Minnies Gap. Each locality map highlights the locations of measured sections, photo panoramas and locality names pertinent to this work.
### Chapter 4

#### Mesaverde Group
- **Cody Shale**
- **Steele Shale**
- **Mancos Shale**

#### NIOBRARA SEAWAY

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<tr>
<td>89.8</td>
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#### Diagrams

- **Fig. 4.3a**: Sevier Orogenic Belt with major formations.
- **Fig. 4.3b**: Mesaverde Group map showing key formations.
- **Fig. 4.3c**: Detailed map of the Mesaverde Group in the Sevier Basin.

#### Key

- **Land**
- **Volcanic rocks (Batholith)**
- **Land, low elevation**
- **Land, high-to-moderate elevation**
- **Fanglomerate**
- **Fluvial sandstone**
- **Nonmarine and coastal plain**
- **Coal swamps**
- **Restricted marine embayment**
- **Marine sandstone and nearshore deposits**
- **Marine sandy shale**
- **Marine shale**
- **Marine calcareous mudstone**

#### Table

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#### Figures

- **Fig. 4.4a**: (a) (b) (c) (d)
- **Fig. 4.4b**: Frontier Formation.
- **Fig. 4.4c**: Ferron Sandstone Member.
- **Fig. 4.4d**: Blackhawk Formation.
Figure 4.5. (over page) (a) Photo panorama of the Panther Tongue on western cliff face of Price Canyon (sections PT1 and PT2; Fig. 4.4c); (b) bedding diagram drawn directly from the photo panorama, illustrating clinoform surfaces and facies-association distribution; and (c) bedding diagram flattened on palaeohorizontal datum surface and corrected for perspective effects by calibration against thickness measurements at outcrop.

Figure 4.6. (page 142) Measured section 15.4 (located in Figs. 4.4c, 4.5b) highlighting the facies associations identified in outcrop (Table 4.1), clinoforms identified in outcrop, and the intensity of bioturbation (using bioturbation index of Taylor and Goldring, 1993). A $V_{\text{shale}}$ (volume of shale) log is constructed from the section, and used to calibrate the petrophysical properties of facies associations in the outcrop analogue against wireline-log data from the Fensfjord Formation, Troll Field (well 31/2-4R; located in Fig. 4.2b). Sedimentary log based on core description, wireline logs including gamma ray (GR), density (RHOB) and p-wave velocity (DT), and synthetic seismogram generated by combining RHOB and DT wireline logs shown from a section of well 31/2-4R. Note the difference in vertical scale between the subsurface log and the measured section, necessitating the scaling increase of the outcrop data. Key to colours for outcrop facies associations is the same as in Figure 4.5.
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**KEY**
- **Low-angle Cross-bedding**
- **Planar Lamination**
- **Symmetrical Ripple**
- **Asymmetrical Ripple**
- **Mica**
- **Siltstone**
- **Sandstone**
- **Carbonate Cement**
- **Skolithos**
- **Palaeophycus**
- **Cylindrichnus**
- **Ophiomorpha**
- **Chondrites**
- **Carbonaceous Debris**
- **Shell**
- **Clinoform**
Figure 4.7. (a) Geometrical parameters of clinoforms and clinothems taken from bedding diagrams corrected to palaeohorizontal for each outcrop unit highlighted in Figure 4.4 and plotted in Figure 4.7b, c and d. Frequency plots of: (b) mean clinothem thickness, Ct; (c) mean angle of clinoform dip, Cd; and (d) clinoform foreset length, Cl. Ct is an isochore thickness, but closely approximates true thickness (to within 5%) for the small values of clinoform dip considered here (<17°).
**Figure 4.8. (over page)** Bedding diagrams flattened on palaeohorizontal datum surface: (a) Kf-1 tongue of the Ferron Sandstone Member at the JP2 locality (Fig. 4.4b); (b) Kf-1 tongue of the Ferron Sandstone Member at the JP1 locality (Fig. 4.4b); (c) Panther Tongue at the PT3 locality (Fig. 4.4c); (d) Kf-2 tongue of the Ferron Sandstone Member at the IV1 locality (Fig. 4.4b); (e) Kf-2 tongue of the Ferron Sandstone Member at the IV2 locality (Fig. 4.4b); (f) Chimney Rock Tongue at the CR2 locality (Fig. 4.4d); and (g) Chimney Rock Tongue at the CR3 locality (Fig. 4.4d). Bedding diagrams are at the same scale and transparent areas indicate where the bedding diagrams were modified in order to remove the effects of perspective and present-day surface erosion.

**Figure 4.9. (page 146)** (a) Scaled model of bedding diagram shown in Figure 4.5c, illustrating the distribution and geometry of facies associations and clinoforms, and associated models of petrophysical properties, based on calibration of well 31/2-4R to outcrop data (Table 4.1, Fig. 4.6): (b) Gamma ray; (c) Density; and (d) p-wave velocity property models.

**Figure 4.10. (page 147)** (a) Frequency spectrum plot of seismic reflection data between the seismic horizons of the Top Fensfjord Formation and Top Brent Group. The frequency was calculated using a 100-trace area of 3D seismic reflection data around well 31/2-4R. A smoothing process was used to clearly identify the minimum, centre and maximum frequency within the defined window of seismic data. Cross plots of: (b) wavelet centre frequency against depth of the middle of seismically mapped stratigraphic intervals near well 31/2-4R; (c) average overburden velocity against depth; and (d) wavelet centre frequency against average overburden velocity. Figure 4.10d shows a "best-fit" linear-regression line through the data points, and three sets of seismic parameters used to create seismic models have been highlighted. Data point 5 was ignored in the "best-fit" line as it lies significantly outside the linear trend, due to the effects of high gas content within the stratigraphic interval.
overburden strata modelled as laterally uniform layers with properties of Heather "B" unit

underburden strata modelled as laterally uniform layers with properties of Heather "A" unit

KEY
- - - Outcrop Outline
Mouth Bar
Proximal Delta Front
Distal Delta Front
Clinoform
KEY

Overburden:

- 1 Nordland Group (Top Pliocene) to Upper Hordaland Group (Intra-Oligocene)
- 2 Upper Hordaland Group (Intra-Oligocene) to Lower Hordaland Group (Intra-Eocene)
- 3 Lower Hordaland Group (Intra-Eocene) to Lower Rogaland Group (Base Tertiary)
- 4 Lower Rogaland Group (Base Tertiary) to Top Fensfjord Formation (upper Callovian)
- 5 Reservoir Interval: Top Fensfjord Formation (upper Callovian) to Top Brent Group (lower Bathonian)
- 6 Underburden: Top Brent Group (lower Bathonian) to Top Statfjord Formation (Lower Jurassic)
- 7 Top Statfjord Formation (Lower Jurassic) to Top Statfjord Formation +400 ms (Upper Triassic)
- 8 Top Statfjord Formation +400 ms (Upper Triassic) to Top Statfjord Formation +800 ms (Upper Triassic)
- 9 Top Statfjord Formation +800 ms (Upper Triassic) to Top Statfjord Formation +1200 ms (Upper Triassic)
- 10 Shallow Seismic Parameters
- - Reservoir Target Seismic Parameters
- - Deep Seismic Parameters
**Figure 4.11.** (over page) (a) Facies-association distribution and clinoform surfaces from corrected bedding diagram (section PT2; Fig. 4.5c), and corresponding property models for: (b) density and (c) p-wave velocity. Resultant forward seismic models for: (d) shallow burial depth (780 m) and high wavelet centre frequency (55Hz); (e) intermediate burial depth (1640 m) and moderate wavelet centre frequency (49Hz); and (f) deep burial depth (2840 m) and low wavelet centre frequency (39 Hz) (Table 4.3). Numbers refer to detailed discussion in text.

**Figure 4.12.** (page 150) (a-c) Facies-association distribution and clinoform surfaces, and (d-f) resultant forward seismic models for three scenarios: (a, d) corrected bedding diagram at outcrop scale (section PT2; Fig. 4.5c); (b, e) corrected bedding diagram scaled up by x5 to the scale of the Troll Field reservoir; and (c, f) corrected bedding diagram scaled up by x5 to reservoir scale, and with increased number of clinoforms to replicate outcrop-scale clinoform spacing. Numbers refer to detailed discussion in text.

**Figure 4.13.** (page 151) (a-c) Acoustic impedance (AI) distributions, and (d-f) resultant forward seismic models for three scenarios: (a, d) heterogeneous overburden corresponding to the Heather “B” unit, Troll Field; (b, e) homogeneous overburden with high acoustic impedance relative to the top of the reservoir interval; and (c, f) homogeneous overburden with the same acoustic impedance as the top of the reservoir interval. Numbers refer to detailed discussion in text.

**Figure 4.14.** (page 152) (a-c) Acoustic impedance (AI) distributions, and (d-f) resultant forward seismic models for three scenarios in which laterally continuous, carbonate-cemented layers occurring along clinoforms are: (a, d) 10 cm thick; (b, e) 50 cm thick; and (c, f) 100 cm thick. Numbers refer to detailed discussion in text.
Figure 4.15. (over page) (a-d) Facies-association distributions and clinoform surfaces from corrected bedding diagrams, and (e-h) resultant forward seismic models of outcrop analogues of fluvial-dominated delta-front clinoform sets: (a, e) Kf-1 tongue of the Ferron Sandstone Member at the JP1 locality (Fig. 4.4b); (b, f) Kf-1 tongue of the Ferron Sandstone Member at the JP2 locality (Fig. 4.4b); (c, g) Panther Tongue at the PT1 locality (Fig. 4.4c); and (d, h) Panther Tongue at the PT3 locality (Fig. 4.4c). The horizontal and vertical dimensions of the bedding diagrams and models are scaled up to reservoir scale, but are not vertically exaggerated. Numbers refer to detailed discussion in text.

Figure 4.16. (page 155) (a-d) Facies-association distributions and clinoform surfaces from corrected bedding diagrams, and (e-h) resultant forward seismic models of outcrop analogues of wave-modified and wave-dominated delta-front clinoform sets: (a, e) Kf-2 tongue of the Ferron Sandstone Member at the IV1 locality (Fig. 4.4b); (b, f) Kf-2 tongue of the Ferron Sandstone Member at the IV2 locality (Fig. 4.4b); (c, g) Chimney Rock Tongue at the CR2 locality (Fig. 4.4d); and (d, h) Chimney Rock Tongue at the CR3 locality (Fig. 4.4d). The horizontal and vertical dimensions of the bedding diagrams and models are scaled up to reservoir scale, but are not vertically exaggerated. Numbers refer to detailed discussion in text.
Figure 4.17. (over page) Seismic sections of the Krossfjord and Fensfjord formations from the Troll Field (located in Fig. 4.2b) with coloured facies associations identified through core analysis: (a) uninterpreted regional dip section; (b) uninterpreted regional strike section and (c) partially interpreted regional dip section through well 31/2-3, highlighting clinoform-related features numbered in the text; (d) uninterpreted regional dip section; (e) uninterpreted regional strike section and (f) partially interpreted regional dip section through well 31/5-5, highlighting clinoform-related features numbered in the text. Formation tops and maximum flooding surfaces (MFS) (Holgate et al., 2013) are highlighted (cf. Fig. 4.2c). Numbers refer to detailed discussion in text.
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<th>Description</th>
<th>Interpretation</th>
<th>Analog to Troll Field Reservoir</th>
<th>Troll Field Petrophysical Properties</th>
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<td><strong>Offshore Transition</strong></td>
<td>Laminated mudstone with very rare, thin (&lt;5 cm), sharp-based, siltstone and poorly-sorted very fine-grained sandstones.</td>
<td>Mudstone deposited through suspension fall out; rare, thin coarser units are the distal expression of major storm events.</td>
<td>Heather Formation, Troll Field</td>
<td>Well 31/2-4R: 1485-1500 m GR = 40-100 API RHOB = 2.1-2.6 g/cm³ DT = 70-150 µs/ft⁻¹</td>
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<tr>
<td><strong>Distal Delta Front</strong></td>
<td>Thin (0.5-1 m), sharp-based, very fine- to fine-grained sandstone interbedded with (c. 10%) mudstone. Parallel laminations grade into wave- and current-rippled tops with very rare low-angle cross-bedding. Sparse bioturbation.</td>
<td>Sandstone beds deposited from episodic sediment gravity flows through river flood events with mudstone interbeds due to suspension settling indicating a distal setting. Wave-rippled tops suggest minor post-depositional wave reworking.</td>
<td>Krossfjord Formation, Troll Field</td>
<td>Well 31/5-5: 1867-1869 m GR = 80-90 API RHOB = 2.1-2.3 g/cm³ DT = 70-90 µs/ft⁻¹</td>
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<tr>
<td><strong>Proximal Delta Front</strong></td>
<td>Thin to amalgamated (0.5-1 m), sharp-based, fine- to medium-grained sandstone beds. Very rare mudstone interbeds. Parallel laminations dominate with minor low-angle and trough cross-stratification. Bioturbation sparse to absent.</td>
<td>Sandstone beds deposited from episodic sediment gravity flows. Wave-rippled tops suggest post-depositional minor wave reworking. Lack of mudstone suggests proximal position to source.</td>
<td>Krossfjord Formation, Troll West Field</td>
<td>Well 31/5-5: 1865-1867 m GR = 70-80 API RHOB = 2.1-2.3 g/cm³ DT = 90-110 µs/ft⁻¹</td>
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<td><strong>(Stream) Mouth Bar</strong></td>
<td>Thick (2-10 m), sharp-based, fine- to medium-grained sandstone. Trough and rare low-angle cross-stratification. Rare organic matter. Bioturbation is rare.</td>
<td>Trough cross-stratification suggests migration of bars caused by unidirectional currents.</td>
<td>Not recognised in core data from Troll Field</td>
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<td><strong>Lower Shoreface</strong></td>
<td>Thinly (0.5-1 m) interbedded siltstone and very fine-to fine-grained sandstone. Hummocky cross-stratification with wave ripples and horizontal laminations. Siltstone and sandstone tops are intensely bioturbated.</td>
<td>Interbedded siltstone and sandstone indicate alternations between fair-weather settling and more energetic hydrodynamic conditions. HCS suggests deposition above storm wave-base.</td>
<td>Fensfjord Formation, Troll Field</td>
<td>Well 31/2-3: 1630-1635 m GR = 80-90 API RHOB = 2.4-2.6 g/cm³ DT = 90-100 µs/ft⁻¹</td>
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<td><strong>Middle Shoreface</strong></td>
<td>Thick (2-9 m), upwards coarsening successions of very fine- to fine-grained sandstone. Hummocky and swaley cross-stratification with minor current-ripple laminations. Sparse to intense bioturbation.</td>
<td>HCS and SCS successions represent amalgamated storm beds deposited above the storm wave base.</td>
<td>Fensfjord Formation, Troll Field</td>
<td>Well 31/2-3: 1628-1630 m GR = 60-80 API RHOB = 2.3-2.4 g/cm³ DT = 90-110 µs/ft⁻¹</td>
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<td>Facies Association</td>
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<td>Interpretation</td>
<td>Analog to Troll Field Reservoir</td>
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<td><strong>Upper Shoreface</strong></td>
<td>Thick (1-6 m) fine- to medium-grained sandstone. Multi-directional trough cross-stratification with minor planar laminations. Moderately- to well- sorted. Sparse bioturbation.</td>
<td>Lack of bioturbation and well-sorted character indicate extensive re-working probably in a high energy marine environment above fair-weather wave base.</td>
<td>Fensfjord Formation, Troll Field</td>
<td>Well 31/2-3: 1622-1628 m GR = 30-50 API RHOB = 2.1-2.3 g/cm⁻³ DT = 110-120 µs/ft⁻¹</td>
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<tr>
<td><strong>Foreshore</strong></td>
<td>Thin (&lt;1.5 m), fine- to medium-grained sandstone. Planar laminations with minor low-angle laminations. Top of unit both burrowed and rooted.</td>
<td>Location above wave-modified upper shoreface facies association and abundance of parallel laminations suggest upper-flow regime conditions in the swash zone.</td>
<td>Fensfjord Formation, Troll Field</td>
<td>Well 31/2-3: 1619-1622 m GR = 30-90 API RHOB = 2.1-2.2 g/cm⁻³ DT = 115-125 µs/ft⁻¹</td>
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<td><strong>Fluvial Channel Sandstone</strong></td>
<td>Medium- to coarse-grained, fining-upward sandstone with unidirectional trough- and current ripple cross-stratification. Narrow (40-200 m) channelised morphology. Bioturbation absent.</td>
<td>Trough cross-stratification suggests migration of bars and superimposed three-dimensional subaqueous dunes within distributary or fluvial channels.</td>
<td>Not recognised in core data from Troll Field</td>
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**Table 4.1.** Summary of facies associations in the Ferron Sandstone Member, Panther Tongue, and Chimney Rock Tongue (synthesised from Anderson et al., 2004; Ryer and Anderson, 2004; van den Bergh and Garrison Jr., 2004; Plink-Björklund, 2008; Enge et al., 2010b; Olariu et al., 2010).
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<th>Mean Palaeocurrent (°)</th>
<th>Depositional Orientation</th>
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<td>277</td>
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<td>131</td>
<td>130</td>
<td>Dip</td>
<td>8</td>
</tr>
<tr>
<td>Ferron Ivie Creek 2 Kf-2 (IV2)</td>
<td>110</td>
<td>30</td>
<td>028</td>
<td>111</td>
<td>Strike</td>
<td>5</td>
</tr>
<tr>
<td>Ferron Ivie Creek 3 Kf-2 (IV3)</td>
<td>170</td>
<td>40</td>
<td>038</td>
<td>152</td>
<td>Strike</td>
<td>12</td>
</tr>
<tr>
<td>Ferron Ivie Creek 4 Kf-2 (IV4)</td>
<td>160</td>
<td>35</td>
<td>131</td>
<td>179</td>
<td>Dip</td>
<td>7</td>
</tr>
<tr>
<td>Panther Tongue 1 (PT1)</td>
<td>170</td>
<td>11</td>
<td>026</td>
<td>188</td>
<td>Dip</td>
<td>17</td>
</tr>
<tr>
<td>Panther Tongue 2 (PT2)</td>
<td>260</td>
<td>15</td>
<td>025</td>
<td>205</td>
<td>Dip</td>
<td>18</td>
</tr>
<tr>
<td>Panther Tongue 3 (PT3)</td>
<td>235</td>
<td>17.5</td>
<td>097</td>
<td>185</td>
<td>Strike</td>
<td>20</td>
</tr>
<tr>
<td>Chimney Rock 1 (CR1)</td>
<td>450</td>
<td>16</td>
<td>113</td>
<td>092</td>
<td>Dip</td>
<td>6</td>
</tr>
<tr>
<td>Chimney Rock 2 (CR2)</td>
<td>190</td>
<td>21.5</td>
<td>103</td>
<td>100</td>
<td>Dip</td>
<td>5</td>
</tr>
<tr>
<td>Chimney Rock 3 (CR3)</td>
<td>275</td>
<td>25</td>
<td>095</td>
<td>078</td>
<td>Dip</td>
<td>9</td>
</tr>
</tbody>
</table>

*Table 4.2.* Statistics calculated from field measurements on each of the localities studied for the three outcrop analogues (Figs. 4.3, 4.4).
<table>
<thead>
<tr>
<th>Depth Name</th>
<th>Mid-Depth (m)</th>
<th>Frequency (Hz)</th>
<th>Averaged Overburden Velocity (m/s)</th>
<th>Detection Limit (m)</th>
</tr>
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<tr>
<td></td>
<td></td>
<td>Min</td>
<td>Centre</td>
<td>Max</td>
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<tr>
<td>Shallow</td>
<td>780</td>
<td>0</td>
<td>55</td>
<td>200</td>
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<tr>
<td>Intermediate</td>
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<tr>
<td>Deep</td>
<td>2840</td>
<td>0</td>
<td>39</td>
<td>200</td>
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</table>

**Table 4.3.** Summary of seismic parameters extracted from linear regression of wavelet centre frequency against average overburden velocity (Fig. 4.10d), used to test sensitivity of the forward seismic models to burial depth.
CHAPTER 5

Seismic stratigraphic analysis of the Middle-Upper Jurassic Krossfjord and Fensfjord formations, Troll oil and gas field, northern North Sea

5.1 Abstract

The “syn-rift” Middle-to-Late Jurassic Krossfjord and Fensfjord formations, Troll Field, northern North Sea contain a complex distribution of wave- and tide-dominated deltaic, shoreline and shelf depositional environments of varying reservoir potential. Uncertainty exists in depositional models used to explain the spatial and temporal distribution of these depositional environments and the absence of coeval coastal-plain deposits. To date, the proposed influence of growing rift-related structures on stratigraphic architectures and sedimentation patterns in the units has been poorly defined. In this study, 3D seismic data are integrated with core, biostratigraphic and wireline-log data to produce a consistent geological interpretation for the formations.

Seismic analysis has identified nine parasequences (‘series’) containing NNE-SSW-striking, delta-scale clinoforms that prograded westwards over much of the field. Quantitative analysis highlights an increase in height and dip of clinoforms from proximal to distal locations, coincident with an increase in grain size. Clinoform geometry is sigmoidal, with well-developed topsets that, based on core data, lack subaerial deposits. These geometrical and sedimentological characteristics suggest that a subaqueous delta depositional system deposited the Krossfjord and Fensfjord formations in the Troll Field. In the northeast of the field, clinoforms exhibit highly variable strike and oblique cross-sectional geometries, which suggest that sediment was supplied from here, and then redistributed through southward-directed wave and longshore current activity. Rift-related faulting is recognised to have occurred in the
western part of the Troll Field only during deposition of the youngest Fensfjord Formation ‘series’, thus challenging the notion that these units are ‘syn-rift’. Seismically imaged clinoforms in the under-explored area south of the Troll Field prograded southward, and are interpreted to represent coeval formation of a southward prograding spit that developed landward of the subaqueous delta platform present in the Troll Field. The interpreted long-lived, subaqueous delta in the Troll Field constitutes a novel type of shallow-marine reservoir.

5.2 Introduction

Clinoforms (sensu Rich, 1951) are basinward-dipping depositional surfaces that are considered to bound the building blocks of deltaic, shelfal and continental margin successions (Mitchum et al., 1977a; Cattaneo et al., 2004). Their geometry varies from oblique to sigmoidal (Sangree and Widmier, 1977), reflecting the complex interplay between accommodation, sediment supply, basin physiography, and hydrodynamics between the shoreline and shelf (Driscoll and Karner, 1999). Furthermore, their trajectory can be characterised to illustrate the interplay between sediment supply and accommodation development through time (e.g. Helland-Hansen and Martinsen, 1996; Helland-Hansen and Hampson, 2009).

Clinoform geometry, distribution, and trajectory have been successfully analysed to refine established depositional models in siliciclastic, shallow-marine settings using outcrop (e.g. Hampson, 2000; Charvin et al., 2010) and subsurface (e.g. Rasmussen, 2009) data, and to help predict the character and distribution of shallow-marine reservoirs (e.g. Bullimore et al., 2005; Jackson et al., 2009). Whilst the use of diagnostic features of clinoforms is complicated in active rift settings, in which there are temporal and spatial variations in relative sea level, accommodation creation and sediment supply throughout the rift cycle (Prosser, 1993; Ravnås and Steel, 1998), a number of studies have shown that clinoform analysis is nonetheless effective in such settings (e.g. Houseknecht et al., 2009; Glørstad-Clark et al., 2010).
Furthermore, clinoform analysis in these settings may reveal the timing of fault-driven tilting events and provide insights into the growth history of normal faults (Gawthorpe and Colella, 1990).

An example of two sand-prone, clinoform-bearing deltaic formations is provided by the Krossfjord and Fensfjord formations (Middle-to-Upper Jurassic), Troll Field, offshore Norway, northern North Sea Basin (Holgate et al., 2013). Together, the formations form a secondary oil-bearing reservoir in the super-giant Troll oil and gas field, which is located on the Horda Platform on the eastern margin of the Viking Graben, northern North Sea (Fig. 5.1). Sedimentological analysis of the formations using sparse core and wireline-log data indicate a complex spatial arrangement of wave- and tide-dominated deltaic, shoreline and shelf depositional environments (Steel, 1993; Stewart et al., 1995; Holgate et al., 2013). Two depositional models have been established to explain this complex arrangement of environments, invoking either a spatial variation in depositional regime where an asymmetrical delta exists fronted by a spit, or a temporal variation in depositional regime where the system prograded from a sheltered, inner shelf location to a more exposed outer shelf location. However, sparse distribution of core data limits the detail of these depositional models. Furthermore, the absence of coastal-plain deposits during repeated episodes of progradation remains problematic in the Krossfjord and Fensfjord formations.

The aims of this chapter are to critically evaluate and refine existing depositional models of the Krossfjord and Fensfjord formations and to assess the degree, if any, to which proposed active growth of rift-related structures during their deposition (Ravnås and Bondevik, 1997) influenced stratigraphic architectures and sedimentation patterns, via integrated analysis of seismic, core and wireline-log data. Geomorphological analysis of the seismically-imaged clinoforms is required to refine depositional models by yielding insights into local lithological distribution and hydrodynamic regimes, which in turn will give a greater understanding of the
distribution of depositional environments. Additionally, seismic stratigraphic analysis is needed to refine sequence stratigraphic interpretations of the formations, to assess the impact of the Middle-to-Late Jurassic rift event on deposition and to guide future exploration and production strategies for Krossfjord and Fensfjord formation reservoirs. A detailed and systematic approach to the seismic analysis of shoreline clinoforms is presented using the detail afforded by a 3D seismic dataset. This seismic geomorphological method highlights the potential of using seismically-imaged clinoforms to refine depositional and sequence stratigraphic models of shallow marine-to-deltaic reservoirs.

5.3 Study area

5.3.1 Tectono-stratigraphic framework of the northern North Sea

The North Sea Basin is located between the Norwegian mainland to the east and the Shetland Platform to the west (Fig. 5.1). The basin represents part of the failed Arctic-North Atlantic rift system (Ziegler, 1990a), and is bordered by broadly N-S-striking fault systems that define 15-50 km-wide, rotated fault blocks (Færseth and Ravnås, 1998) (Fig. 5.1c). Rifting occurred in the northern North Sea Basin during the Permo-Triassic and Middle-Late Jurassic (Bajocian-to-Volgian) (Færseth and Ravnås, 1998; Nøttvedt et al., 2000). These two periods of rifting are separated by an ‘intra-rift’ interval, during which time regional tectonic uplift of the basin occurred (Steel, 1993; Færseth, 1996; Færseth and Ravnås, 1998).

The Horda Platform is separated from the Viking Graben to the west by a number of tilted half grabens that host the Brage, Oseberg, Troll and Fram fields, and is bounded to the east by the Øygarden Fault Complex (Fig. 5.1a, c) (Stewart et al., 1995). It contains a number of easterly-tilted half grabens that contain c. 3 km of pre-Jurassic syn-rift strata, c. 1 km of intra-rift Early-to-Middle Jurassic strata, and <500 m of Late-Jurassic-to-Early-Cretaceous syn-rift strata.
This study focuses on the Middle-to-Late Jurassic strata of the Viking Group. The Troll Field is located on the northern tip of the Horda Platform, which lies on the eastern flank of the North Viking Graben rift arm (Fig. 5.1).

The Viking Group is termed “syn-rift” because it was deposited during the Middle-Late Jurassic rift event (Ravnås and Bondevik, 1997). The rift event is interpreted to contain five discrete “pulses” of extension that caused progressively greater fault-related subsidence and footwall uplift, and which created a series of faulted terraces between the Viking Graben and the Horda Platform (Færseth and Ravnås, 1998; Ravnås et al., 2000). However, on the Horda Platform, which is located towards the rift margin, little rift-related deformation occurred before the Kimmeridgian (Færseth, 1996). Previous authors have suggested that each rift pulse caused backstepping of basin margin-attached clastic depositional systems, including two clastic wedges now preserved within the Viking Group (“Krossfjord-Fensfjord megasequence” and “Sognefjord megasequence” of Steel, 1993) (Færseth and Ravnås, 1998; Nøttvedt et al., 2000).

5.3.2 Stratigraphy and depositional models for the Viking Group, Troll Field

The Middle-Late Jurassic Viking Group (167.7-140.2 Ma) comprises shelfal mudstones of the Heather Formation towards the basin centre in the west, and three shallow-marine sandstone tongues of the Krossfjord, Fensfjord and Sognefjord formations towards the basin margin in the east (Vollset and Doré, 1984; Ravnås and Bondevik, 1997) (Fig. 5.1d). To describe its stratigraphic relationship of the contained sandstone tongues, the Heather Formation is informally split into three parts: the Heather “A” unit lies below the Krossfjord Formation tongue (Bathonian); the Heather “B” unit occurs between the Fensfjord and Sognefjord tongues (Callovian to Oxfordian); and the Heather “C” unit overlies the Sognefjord Formation tongue (Oxfordian to Kimmeridgian) (Stewart et al., 1995). These strata were deposited in a series of regressive-transgressive cycles that record the advance and retreat of deltaic and shallow-
marine depositional systems across the Horda Platform, with sediment being supplied from the uplifted Norwegian hinterland to the west and accommodation space being created by passive subsidence of the Horda Platform (Steel, 1993; Stewart et al., 1995; Ravnås and Bondevik, 1997; Fraser et al., 2002; Husmo et al., 2002; Sømme et al., 2013; Whipp et al., 2013).

This study is concerned with the lower Viking Group, comprising the Heather "A" unit, Krossfjord Formation, Fensfjord Formation and part of the Heather "B" unit. Within these strata, four biostratigraphically constrained, regional maximum flooding surfaces have been identified and tied to the North Sea-wide, Partington et al. (1993) "J-surface" biostratigraphic scheme: J32 (latest Bajocian, 165 Ma), J42 (early Callovian, 155.5 Ma), J44 (middle Callovian, 154 Ma) and J46 (late Callovian, 152 Ma). Based on the prevailing stratigraphic nomenclature for the Troll Field (e.g. Dreyer et al., 2005), these maximum flooding surfaces have been used to define three 'series' (corresponding to 'genetic sequences' sensu Galloway, 1989) that each contain multiple parasequences (Fig. 5.1d). 'Series 1', from J32 to J42, contains the Heather "A" unit, the Krossfjord Formation and the lowermost Fensfjord Formation; 'Series 2', from J42 to J44, contains the lower Fensfjord Formation; and 'Series 3', from J44 to J46, contains the upper Fensfjord Formation and Heather "B" unit (Holgate et al., 2013) (Fig. 5.1d).

The Sognefjord Formation, which forms the primary reservoir interval in the Troll Field, has been the focus of several sedimentological and sequence stratigraphic studies (e.g. Whitaker, 1984; Hellem et al., 1986; Osborne and Evans, 1987; Dreyer et al., 2005; Patruno et al., In Press). In contrast, the Krossfjord and Fensfjord formations have received considerably less attention. Stewart et al. (1995) and Holgate et al. (2013) indicate that these units contain a complex arrangement of depositional environments that can be mapped across the Troll Field. A wide (10–20 km), north-south-trending belt of wave-dominated shoreface deposits is present in the western part of the field, but its eastern part contains more irregular (0-20 km wide), north-south-trending belts of mixed wave- and tide-influenced shoreface, tide-dominated embayment,
and fluvial-dominated delta-front deposits. No coastal-plain deposits have been identified. The change from tide-influenced deposits in the east to wave-dominated deposits in the west have been attributed to either a temporal variation in depositional process regime, in which the system prograded from a sheltered, inner shelf location to a more exposed, outer shelf location, or a spatial variation in depositional process regime, in which an asymmetrical delta fronted by a spit protected a sheltered, inner shelf location (Holgate et al., 2013). These models are similar to those recently constructed for the more data-rich and widely studied Sognefjord Formation (Dreyer et al., 2005; Patruno et al., In Press), although earlier work suggested this unit was deposited in a series of offshore bars (Whitaker, 1984; Hellem et al., 1986; Osborne and Evans, 1987). Current uncertainty in the most applicable depositional model for the Krossfjord and Fensfjord formations, and also the Sognefjord Formation, reflects the need for a viable mechanism(s) to explain the juxtaposition of depositional environments, the absence of coastal-plain deposits, and the variation in stratigraphic architecture during the Middle-to-Late Jurassic. Furthermore, despite the formations being classically considered as "syn-rift", this work will determine the degree, if any, to which active growth of rift-related structures influenced stratigraphic architectures and sedimentation patterns.

5.4 Dataset and methodology

5.4.1 Dataset

The Troll Field study area is covered by two three-dimensional (3D) seismic surveys, NH0301 and SG9202, which overlap in the centre of the field (Fig. 5.2; Table 5.1). The seismic surveys cover c. 1595 km² and image to a depth of 2400-3000 milliseconds two-way time (ms TWT). The seismic detection limit in the interval of interest is 9.72 m with a peak frequency of 49 Hz and an overburden velocity of 1906 m/s. Three two-dimensional (2D) seismic surveys, NVGT88, SG8043 and TE90, cover an area to the southeast of the Troll Field (Fig. 5.2). The 2D seismic
surveys cover c. 2295 km² and image to a depth of 5500-6500 milliseconds two-way time (ms TWT). All seismic sections are presented in “European” reverse polarity in which a peak (black) represents a negative acoustic impedance boundary (decrease in acoustic impedance), and a trough (red) represents a positive acoustic impedance boundary (increase in acoustic impedance). Nine cored and six uncored vertical wells, containing a conventional log suite, including gamma ray, sonic, neutron and density logs, were utilised for this study (Fig. 5.2). Sedimentological analysis of 893 m of core has revealed wave- and tide-dominated deltaic, shoreline and shelf depositional environments (Holgate et al., 2013). Biostratigraphic data from the wells were used to identify the maximum flooding surfaces, which can be correlated to the North Sea-wide, ‘J surface’ stratigraphic scheme of Partington et al. (1993).

5.4.2 Seismic interpretation methodology

Five regionally interpreted maximum flooding surfaces (J32, J36, J42, J44 and J46) were identified and mapped within the seismic dataset; these surfaces were identified using core, wireline-log and biostratigraphic data (see above) and were tied to the seismic data using synthetic seismograms (Fig. 5.3). Two further candidate maximum flooding surfaces (?J33 and ?J34) have also been interpreted and mapped; these have a characteristic high-amplitude, laterally persistent seismic expression, similar to that of the biostratigraphically constrained maximum flooding surfaces described above. All interpreted maximum flooding surfaces are represented in the seismic reflection dataset as trough reflections (Table 5.2). Identification of these J-surfaces enables the adoption of the ‘series’ nomenclature established by Dreyer et al. (2005) for the Sognefjord Formation reservoir in the Troll Field, which was then applied to the Fensfjord and Krossfjord formations by Holgate et al. (2013). In this nomenclature, the numbered ‘series’ refers to a parasequence set (sensu Van Wagoner et al., 1990) and the suffix letter refers to a single parasequence (clinoform set) imaged in seismic data (Fig. 5.3).
Time-thickness maps (isochrons) for intervals bounded by the interpreted maximum flooding surfaces are used to illustrate changes in stratigraphic thickness across the Troll Field. The majority of wells in the study area penetrate footwall highs defined by the major bounding fault systems; in these locations, the Upper Jurassic section is relatively thin. In contrast, very few wells penetrate the immediate hanging walls of these fault systems within which an expanded Upper Jurassic succession is preserved (Fig. 5.2). Because of this spatial aliasing in well data, confidence in tracing seismically defined maximum flooding surfaces across the study area is slightly limited.

### 5.4.3 Quantitative analysis of clinoform dimensions and morphology

Seismically imaged clinoforms are evident throughout the Troll Field in all of the ‘series’ of the Krossfjord and Fensfjord formations. However, the majority of clinoforms appear to be near the limit of seismic resolution and, especially around heavily faulted areas, are not clearly imaged. Furthermore, sedimentological analysis of cores and well logs indicates many more lithological breaks than are imaged in seismic data (Holgate et al., 2013; Holgate et al., In Press). The apparently wide spacing of seismically imaged clinoforms may be attributed to true clinoform spacing occurring below the tuning thickness of the seismic data and/or to limited variability in lithological character and associated acoustic properties along the clinoforms (Holgate et al., In Press). Consequently, the clinoforms analysed in this study are large, widely spaced (greater than the c. 10 m detection limit of the seismic dataset), and marked by clear lithological contrasts, and they are located away from heavily faulted areas. A number (c. 20-30) of clinoforms were selected in each ‘series’ and their 3D extent mapped in the seismic data to ascertain their strike and dip (Fig. 5.4a, b). Dip-oriented seismic cross-sections of the best imaged clinoforms were depth-converted using interval velocities extracted from sonic log data from a nearby well. The cross-sections were then decompacted using published compaction curves (Sclater and Christie, 1980) for lithologies taken from nearby wells to extract accurate,
pre-compaction clinoform geometries (e.g. length, height, dip; Fig. 5.4c). Overburden lithologies utilised in the decompaction exercise were averaged from all 15 wells used in this study, as there is little variation in their lithology across the Troll Field. Tightly constrained depth-conversion and decompaction of 2D seismic lines from the area southeast of the Troll Field (Fig. 5.2) was not carried out in the absence of well data, thus the true geometry and scale of the seismic imaged clinoforms in this area are more uncertain.

Clinoform trajectory in decompacted, dip-oriented cross-sections flattened on an assumed palaeohorizontal surface was also analysed using the clinoform topset-foreset rollover point as a reference point in successive clinoforms (e.g. Helland-Hansen and Hampson, 2009). Interpretations of clinoform trajectory are subject to three key uncertainties: (1) abrupt lateral changes in lithology, which are poorly constrained by sparse well data in the study area, may have resulted in localised modification of stratal geometries due to differential post-depositional compaction (e.g. Hampson et al., 2009); (2) the chosen palaeohorizontal datum may have been affected by differential compaction and tectonic tilting (e.g. Prince and Burgess, 2013); and (3) the limited dip extent of seismically-imaged clinoform sets do not permit reconstruction of clinoform trajectory across the entire Troll Field.

5.5 Seismic-stratigraphic architecture and clinoform geometry

In the following section, the seismic-stratigraphic architecture of each parasequence and the geometry of their internal, seismically resolved clinoforms are described, principally using 3D seismic data over the Troll Field. The 2D seismic dataset to the southeast of the Troll Field is only used in the analysis of ‘Series 2’, as clinoforms are not resolved in other ‘series’ within this dataset.
5.5.1 Series 1

'Series 1' is bounded by the J32 (latest Bajocian) and J42 (early Callovian) maximum flooding surfaces, at its base and top respectively, and consists of the Heather 'A' unit, the Krossfjord Formation and the lowermost Fensfjord Formation. 'Series 1' has previously been described as a progradational parasequence set, based on wireline-log correlations (Holgate et al., 2013).

Seismic-stratigraphic analysis conducted in this study has identified five clinoform sets (parasequences), referred to as 'Series 1a' to 'Series 1e' (Fig. 5.3).

Series 1a

'Series 1a' is bounded by the J32 and ?J33 maximum flooding surfaces at its base and top, respectively, and only occurs in the eastern part of the Troll Field, where it is up to 25 ms TWT (c. 35 m) thick (Fig. 5.5a). The ?J33 surface downlaps onto the J32 surface, indicating that the point of maximum clinoform regression within 'Series 1a' lay near the centre of the field. Furthermore, basinward pinchout of 'Series 1a' had a linear geometry, striking NNE-SSW. Whilst hemipelagic sedimentation is envisaged to have occurred further west of the seismically resolved downlap line, the termination of the trough reflection representing the ?J33 surface is probably due to a lack of acoustic contrast between lithologically similar deposits above and below the surface (e.g. Holgate et al., In Press). This interpretation is reinforced by a decrease in seismic amplitude maxima, interpreted to represent the foreset of clinoforms, from east to west near the downlap point of ?J33 onto J32 (Fig. 5.5b).

Clinoforms are best imaged towards in the western part of the 'Series 1a' wedge, close to the line of maximum regression, perhaps reflecting a westward increase in clinoform height. The bottomsets of the clinoforms are marked by seismic amplitude maxima that strike NNE-SSW, parallel with the strike of the main downlap line of the 'series'. Clinoforms have relatively large heights (32-33 m), large lengths (162-220 m) and moderate-to-steep foreset dips (8-12°).
measured relative to the flattened J32 datum (Fig. 5.5c-e). Clinoform geometry appears broadly sigmoidal; however this observation is limited by the high amplitude seismic reflections of the J32 and ?J33 surfaces, which obscure clinoform reflections within the ‘series’. Clinoform trajectory appears to be close to horizontal (Fig. 5.5e).

**Series 1b**

‘Series 1b’ is bounded by the ?J33 and ?J34 maximum flooding surfaces at its base and top, respectively (Fig. 5.3). The ‘series’ occurs in the eastern and central parts of the Troll Field, where it reaches thicknesses of up to 50 ms TWT (c. 60 m) (Fig. 5.6a). The ?J34 surface downlaps onto the ?J33 surface, defining a linear, NNE-SSW-trending downlap limit, similar to ‘Series 1a’. ‘Series 1b’ is thickest and forms a wedge in the region between the downlap limits of ‘Series 1a’ and ‘Series 1b’, and clinoforms are seismically resolved in this region. Clinoform strike and amplitude maxima along clinoforms are broadly parallel to the downdip limit of ‘Series 1b’ (Fig. 5.6b-d). Clinoforms have a variable height (13-33 m), are relatively long (180-230 m), and are relatively gently dipping (6-9°) (Fig. 5.6c, d). Towards the north-east tip of the Troll Field, clinoform strike becomes more erratic, varying from ENE-WSW to NW-SE, and here clinoforms are taller (15-35 m), narrower (110-160 m) and steeper (6-13°) than in the south (Fig. 5.6c, d). Clinoform geometry is sigmoidal with well-developed topsets in the south of the Troll Field (Fig. 5.6e) but oblique in the north. Clinoform trajectory appears slightly descending towards the maximum regression of the ‘series’ (Fig. 5.6e).

**Series 1c**

‘Series 1c’ is bounded by the ?J34 and J36 maximum flooding surfaces at its base and top, respectively (Fig. 5.3), and occurs across the entirety of the Troll Field. The downlap limit of the ‘series’ is assumed to exist beyond the western boundary of the Troll Field. ‘Series 1c’ is thickest (up to 50 ms; c. 60 m) to the west of the downlap limit of ‘Series 1b’ (Fig. 5.7a). Clinoforms within ‘Series 1c’ are seismically resolved across all of the Troll Field, especially where the
‘series’ is thickest. Their strike is broadly NNE-SSW, with the greatest variability occurring in the north east of the Troll Field. Linear amplitude maxima, some of which are laterally extensive and concordant with the strike of the clinoforms, occur in the south of the Troll Field (Fig. 5.7b). Clinoforms are relatively long (130-250 m), tall (18-36 m), and steep (7-16°) (Fig. 5.7c, d). Clinoform geometry is typically sigmoidal with well developed topsets in the south of the Troll Field (Fig. 5.7e), and oblique clinoform geometries in the northeast. Clinoform trajectory appears to descend slightly towards the west (Fig. 5.7e).

**Series 1d**

The interval that contains both ‘Series 1d’ and ‘Series 1e’ is bounded by the J36 and J42 maximum flooding surfaces at its base and top, respectively (Fig. 5.3). The two ‘series’ are present across the Troll Field and thickest (up to 70 ms; c. 90 m) in the west of the Troll Field (Fig. 5.8a). ‘Series 1d’ and ‘Series 1e’ are separated by a transgressive surface. Clinoforms within ‘Series 1d’ are bound by the J36 below and a transgressive surface above, and their strike changes from ENE-WSW in the south to NNE-SSW in the north of the Troll Field. This curvate geometry is imaged in laterally extensive amplitude maxima that occur at the foreset-bottomset transition (Fig. 5.8b). Clinoform heights vary across the Troll Field (14-32 m), with lengths decreasing from east (150-300 m) to west (110-150 m), and dips increasing (from 3-9° to 8-12°) in the same direction (Fig. 5.8c, d). Clinoform geometry is generally sigmoidal, apart from oblique clinoforms located in the northeast. Clinoform trajectory appears horizontal (Fig. 5.8e).

**Series 1e**

Clinoforms within ‘Series 1e’ are bound by a transgressive surface below and the J42 above. They are seismically resolved across the Troll Field and their strike, and associated amplitude maxima, vary from NE-SW in the south to NNE-SSW in the north (Fig. 5.9b-d). Clinoform strike towards the north east of the Troll Field is more variable, from NE-SW to NNW-SSE orientations being present (Fig. 5.9c, d). Clinoform heights increase from east (22-30 m) to west (25-35 m),
and lengths shorten (from 170-260 m to 100-210 m), resulting in dips becoming steeper (from 5-8° to 7-14°) in the same direction (Fig. 5.9c, d). Clinoform geometry is sigmoidal in the south of the Troll Field and oblique in the northeast. Clinoform trajectory appears flat (Fig. 5.9e).

5.5.2 Series 2

'Series 2' is bounded by the J42 and J44 maximum flooding surfaces at its base and top, respectively, and consists of the lower-to-middle Fensfjord Formation (Fig. 5.3). 'Series 2' consists of one parasequence that contains seismically resolved clinoforms across the entirety of the Troll Field, and in the area to the southeast of the field. The 'series' is thickest to the south and west of the Troll Field (up to 60 ms or c. 80 m) (Fig. 5.10a). Clinoforms and associated amplitude maxima within 'Series 2' appear slightly curved in plan-view in the Troll Field area, and their strike generally varies from NE-SW in the south to NNE-SSW in the north (Fig. 5.10b). Clinoform strike in the north east of the Troll Field is more variable, from NE-SE to NNE-SSW (Fig. 5.10b-d). Clinoform height increases from southeast (18-29 m) to northwest (26-40 m), and their lengths shorten (from 250-350 m to 60-180 m) and dips increase (from 7-9° to 10-14°) in the same direction (Fig. 5.10c, d). Clinoform geometry is mostly sigmoidal, with steeper clinoforms appearing oblique in the northeast (Fig. 5.10e). Clinoform trajectory appears horizontal (Fig. 5.10e).

Clinoforms are also imaged in 2D seismic lines southeast of the Troll Field within 'Series 2' (Fig. 5.11), where they dip to both the southwest and southeast (Fig. 5.11a). These clinoforms are tall (30-50 ms TWT equating to c. 50-85 m using nearest well 31/6-3), long (5-12 km in N-S-oriented 2D sections and 2-7 km in E-W-oriented 2D sections equating to c. 5-14 km true dip extent), and more gently dipping (<1°) than clinoforms in the Troll Field (Fig. 5.11b, c). Clinoform geometry is shingled to sigmoidal, with well-developed topsets, and clinoform trajectory appears horizontal (Fig. 5.11b, c). In E-W-oriented sections, eastward- and westward-
dipping clinoforms are separated by a 500 m wide, 45 ms TWT (c. 75 m) tall, dome-like feature, which is present at the top of ‘Series 2’ (Fig. 5.11b).

5.5.3 Series 3

‘Series 3’ is bounded by the J44 and J46 maximum flooding surfaces at its base and top, respectively, and consists of the upper Fensfjord Formation and the lower Heather ‘B’ unit (Fig. 5.3). ‘Series 3’ contains three clinoform sets (parasequences) (Series 3a-c), and is thickest towards the west of the Troll Field (up to 90 ms; c. 120 m). The thickness of ‘Series 3’ increases from the footwall (up to 65 ms; c. 100 m) to the hangingwall (up to 90 ms; c. 120 m) of the Svartalv and Tusse fault systems developed in the west of the Troll Field (Fig. 5.12a).

Series 3a

Clinoforms within ‘Series 3a’ are seismically resolved across the entirety of the Troll Field. They strike broadly NNE-SSW (Fig. 5.12b), with greater variability occurring in the northeast, where they strike from NE-SW to NW-SE (Fig. 5.12c, d). Clinoform heights (10-44 m), lengths (60-220 m), and dips (4-17°) appear variable across the Troll Field, but the greatest heights and steepest dips occur towards the west (Fig. 5.12c, d). Clinoform geometry is sigmoidal, with steeper clinoforms appearing oblique in the northeast of the field, and clinoform trajectory appears horizontal (Fig. 5.12e).

Series 3b

Clinoforms within ‘Series 3b’ are also seismically resolved across the whole Troll Field. They are slightly curved in plan-view, such that they strike NE-SW in the southeast to NNE-SSW in the west (Fig. 5.13b-d). Clinoform height increases from east (17-28 m) to west (19-32 m), and their lengths shorten (from 120-180 m to 80-160 m) and dips increase (from 7-11° to 8-13°) in the
same direction (Fig. 5.13c, d). Clinoform geometry is sigmoidal, with steeper clinoforms appearing oblique in the northeast, and clinoform trajectory appears horizontal (Fig. 5.13e).

**Series 3c**

Clinoforms within ‘Series 3c’ are only seismically resolvable in the western half of the Troll Field, where ‘Series 3’ is thickest (Fig. 5.14a). They are curved in plan-view, changing from NE-SW in the east to NNE-SSW in the west (Fig. 5.14b-d). Clinoform height (16-28 m) and length (66-216 m) are variable across the field (Fig. 5.14c). Clinoform dip is also variable (6-16°), but is slightly steeper in clinoforms located in the immediate hangingwall of the Svartalv and Tusse fault systems (Fig. 5.14d). Clinoform geometry is sigmoidal-to-oblique and clinoform trajectory appears horizontal (Fig. 5.14e). Seismic reflection continuity and amplitude is relatively strong within this ‘series’ compared to others, probably because it lies directly below the Heather “B” shales, which have distinctly different acoustic properties to the ‘Series 3c’ sandstones (e.g. Holgate *et al.*, In Press).

**5.6 Depositional evolution of the lower Viking Group**

Block diagrams illustrating the simplified stratigraphic architecture and facies-belt distribution at maximum regression of each clinoform set (parasequence), as interpreted from integrated analysis of seismic, core, wireline-log and biostratigraphic data, are presented in Figure 5.15. The depiction of facies belts is restricted by limited well penetrations, especially in ‘Series 1’. Furthermore, quantitative wireline-log analysis does not show clear differences between various sandstone-dominated facies, such that the relative influence of wave, tidal and river-mouth processes can only be inferred from core data (Holgate *et al.*, 2013). Neither fluvial nor coastal-plain deposits have been identified in core (Holgate *et al.*, 2013), thus no fluvial sediment input points or subaerial exposure are shown in the reconstructions. Time-thickness maps of ‘Series 1’ and ‘Series 2’ (Figs. 5.5a, 5.6a, 5.7a, 5.8a, 5.9a, 5.10a) indicate no discernible
thickening across fault systems, and therefore no active rift faulting is inferred during their deposition. The time-thickness map of ‘Series 3’ (Figs. 5.12a, 5.13a, 5.14a) shows thickening across the Svartalv and Tusse fault systems with clinoform analysis suggesting the thickening is confined to ‘Series 3c’; these fault systems are therefore shown as being active only during deposition of ‘Series 3c’ (Fig. 5.15i).

5.6.1 Lateral trends within the lower Viking Group

Several lateral trends have been identified throughout the ‘series’ of the lower Viking Group. In all ‘series’, average clinoform height increases from east (21 m) to west (26 m), and their lengths shorten (from 171 m to 147 m) and dips increase (from 7° to 11°) in the same direction. Clinoform strike appears to be oriented NE-SW in the east of the Troll Field compared to NNE-SSW in the west of the Troll Field with sigmoidal clinoform geometries (Fig. 5.15). Clinoform strike in the northeast of the Troll Field commonly appears highly variable from NE-SW to NW-SE with oblique clinoformal geometries in all ‘series’ (Fig. 5.15).

The increase in clinoform height and dip from east to west is coincident with a distinct change in facies association. ‘Series 2’ to ‘Series 3c’ document both tide-influenced (FA4 of Holgate et al., 2013) and tide-dominated deposits (FA5 of Holgate et al., 2013) in the centre and east of the Troll Field (Fig. 5.15f-i). West of these deposits, throughout ‘Series 1d’ to ‘Series 3c’, are wave-dominated deposits (FA2 and FA3 of Holgate et al., 2013) (Fig. 5.15e-i). This lateral facies trend is accompanied by an overall increase in grain size from east to west.

5.6.2 Vertical trends within the lower Viking Group

Parasequence stacking from ‘Series 1a’ to ‘Series 1c’ documents punctuated progradation of shallow-marine deposits from east to west across the Troll Field (i.e. MFS J32 to J36; Fig. 5.15a-
c). Clinoform trajectory within these progradational parasequences appears slightly descending towards the maximum regression of each parasequence. ‘Series 1d’ to ‘Series 3c’ appear aggradational in parasequence stacking, and clinoform trajectory within these parasequences appears horizontal (Fig. 5.15d-i). Average clinoform length (178 m to 132 m) and height (29 m to 24 m) appear to decrease from ‘Series 1’ to ‘Series 3’, resulting in an increase in clinoform dip (from 9° to 11°).

The vertical increase in abundance of tide-influenced deposits in ‘Series 2’ and ‘Series 3’ is believed to reflect the change from progradational to retrogradational stacking, as identified in wireline-log analysis (Holgate et al., 2013). These tide-influenced facies belts, which strike parallel to the curvilinear strike orientation determined from seismic analysis of the clinoforms, appear to prograde west from ‘Series 3a’ to ‘Series 3b’, but then retreat eastward from ‘Series 3b’ to ‘Series 3c’ (Fig. 5.15g-i). Whilst, thickness maps indicate that the Tusse and Svartalv fault systems were active during deposition of ‘Series 3’, the sparse distribution of cored wells means it is not possible to ascertain whether syn-depositional movement on these structures influenced facies distributions.

5.7 Discussion

5.7.1 Appraisal of previously proposed depositional models for the Krossfjord and Fensfjord formations

Two depositional models have previously been presented to explain the pronounced partitioning between wave-dominated deposits in the west, and mixed wave- and tide-influenced deposits in the east of the Troll Field, based on sedimentological analysis of core and wireline-log data (see Figure 13 in Holgate et al., 2013). The first model infers a spatial variation in process regime between coeval depositional environments. Sediment was supplied through a
fluvial-deltaic source in the north and then reworked by southward-directed longshore currents to form a spit in the west of the Troll Field. The seaward face of the spit consisted of a wave-dominated shoreface, and a tide-influenced back-basin lay on the landward face of the spit. A similar model was proposed for the Sognefjord Formation in the Troll Field (Dreyer et al., 2005). Although this model explains the spatial variation in depositional process, it does not account for the absence of roots and other subaerial-exposure features, which are typical in spit and back-barrier deposits (e.g. Novak and Pedersen, 2000; Nielsen and Johannessen, 2001). In addition, seismic-stratigraphic analysis of clinoforms indicates a consistent NNE-SSW-oriented strike and westward dip across the Troll Field (Figs. 5.5-5.10, 5.12-5.14). The simple, consistent clinoform orientation contrasts with documented examples of spit systems, in which clinoforms within the spit dip perpendicular to shoreline strike due to the recurved spit morphology (e.g. Nielsen and Johannessen, 2009; Hein et al., 2012) and clinoforms within the back-basin dip landward (e.g. Novak and Pedersen, 2000; Braga et al., 2003). Transgressive erosion may account for the absence of roots and other subaerial-exposure features normally present on and/or landward of the spit (e.g. Storms and Swift, 2003; Nielsen and Johannessen, 2008). However, the resultant dipping-geometries representing a recurved spit morphology, used above to counter the spit system depositional model for the Troll Field, would be preserved within the buried spit platform (cf. Meistrell, 1972), and have previously been used to re-evaluate depositional models for ancient, isolated sandstone bodies within the US Cretaceous Western Interior Seaway (Nielsen and Johannessen, 2008).

The second depositional model infers temporal variation in process regime, with an initial tide-influenced environment in a shallow, sheltered embayment evolving during later progradation to a wave-dominated environment in a deeper, more exposed setting. A wide, shallow, subaqueous shelfal platform is envisaged to have existed in front of the clinoform during early progradation, which damped wave energy from the open basin through friction, and which may also have increased tidal resonance (e.g. Ainsworth et al., 2011). During progradation, the width
of the shelfal platform decreased, thereby exposing the clinoform to more wave energy. This model explains the spatial variation in depositional process, and may also explain why seismically imaged clinoforms become progressively steeper from east to west, which is consistent with a westward increase in their grain size (cf. Orton and Reading, 1993). However, this depositional model also requires elaboration to account for the absence of coastal-plain deposits in the Krossfjord and Fensfjord formations across the c. 1500 km² area of the Troll Field.

### 5.7.2 Absence of coastal-plain deposits

Two mechanisms have been previously proposed for the absence of palaeosols and coastal-plain deposits across the studied dip extent of c. 30 km in the Krossfjord and Fensfjord formations (Holgate et al., 2013), and in the overlying Sognefjord Formation (Patruno et al., In Press), Troll Field; either palaeosols were originally developed at the top of each clinoform set and then removed by transgressive erosion (ravinement), potentially assisted by forced regression, or clinoform topsets were fully subaqueous, such that palaeosols were never developed. The same two mechanisms have been proposed and vigorously debated, for numerous other shallow-marine sandstones that lack evidence of subaerial exposure and appear to be isolated on the shelf (e.g. the Shannon Sandstone in the US Cretaceous Western Interior Seaway; Suter and Clifton, 1999). A comparison of features observed in the Krossfjord and Fensfjord formations, with diagnostic features of ancient, sand-rich, top-truncated deltas and modern, fully subaqueous deltas, is presented in Table 5.3. The major discussion points are expanded upon below.

Transgressive erosion at the tops of clinoform sets (parasequences) is supported by the occurrence in core of coarse-grained lags immediately below their bounding flooding surfaces (Holgate et al., 2013). Attenuated coastal-plain deposits, consequently readily removed by
transgressive erosion, may have occurred with forced regression of each clinoform set during relative sea level fall (cf. Posamentier and Morris, 2000). This feature, along with the long-distance regression of the Viking Group, has previously been used to support a forced regression scenario during deposition of the Krossfjord and Fensfjord formations (Steel, 1993). Similar scenarios have been invoked for the absence of coastal-plain deposits within numerous other ancient (e.g. Bhattacharya and Willis, 2001; Uličný, 2001; Olariu et al., 2005; Morris et al., 2006; Vakarelov and Bhattacharya, 2009) and more recent (e.g. Knott and Hoskins, 1968; Tesson et al., 1990; Posamentier et al., 1992a; Hart and Long, 1996; Tesson et al., 2000) deltaic-to-shoreface depositional environments.

Two out of the seven characteristics of forced regressive deposits identified by Posamentier and Morris (2000) have been recognised in the Krossfjord and Fensfjord formations: (1) increased average grain size in a proximal to distal direction due to redistribution of proximal highstand sediments; and (2) the long distance of regression with anomalously thin character of parasequences due to decreased accommodation. However, long distance regression is described as “circumstantial” evidence (sensu Posamentier and Morris, 2000) and wave-dominated, tide-influenced systems have been recognised to rework sediment long distances over shallow receiving basins through subaqueous deltas due to strong storm- and tide-generated currents and without a decrease in relative sea level (e.g. Vakarelov et al., 2012). Furthermore, Holgate et al. (2013), using core data from the Krossfjord and Fensfjord formations, do not identify sharp-based shoreface deposits (e.g. Hart and Long, 1996) or a zone of shallow-marine sediment bypass indicative of forced regression (e.g. Tesson et al., 2000; Vakarelov and Bhattacharya, 2009). Additionally, seismic data analysis presented here does not identify the progressive reduction in relief of clinoforms from proximal to distal locations (e.g. Tesson et al., 1990; Morris et al., 2006) or seawards-dipping bounding surfaces on the top of parasequences (e.g. Uličný, 2001). Moreover, experimental and theoretical studies have indicated that forced regression may result in the aggradation of topset strata (e.g. Muto and
Swenson, 2005; Swenson and Muto, 2007; Prince and Burgess, 2013) further challenging the use of forced regression to account for the absence of coastal-plain deposits by the attenuation of topset strata.

Whilst a forced regression scenario is discounted, transgressive erosion may still account for the absence of coastal-plain deposits (e.g. Swift, 1968; Cattaneo and Steel, 2003; Catuneanu et al., 2009; Helland-Hansen and Hampson, 2009). Transgressive erosion is commonly identified as removing up to 10 m of the substrate (e.g. Demarest and Kraft, 1987; Kraft et al., 1987; Saito, 1994), but has been documented to remove up to 40 m of the substrate due to extreme wave and longshore drift conditions (e.g. Leckie, 1994). Furthermore, transgressive erosion without forced regression has been identified in studies on ancient shoreline and deltaic successions (e.g. Ryer, 1977; Bhattacharya and Walker, 1991), especially where regressive shoreline trajectories are horizontal, indicating a relative sea level stillstand and limiting significant aggradation of topset strata (e.g. Hampson and Howell, 2005). However, if coastal-plain deposits were allowed to develop during normal regression, transgressive erosion would result in truncated clinoform topsets (e.g. Hampson and Storms, 2003; Hampson and Howell, 2005; Porębski and Steel, 2006), which is not recognised in the seismically imaged clinoforms of the Krossfjord and Fensfjord formations, most of which are sigmoidal in geometry and appear to have well-developed topsets (e.g. Figs. 5.5e, 5.6e, 5.7e, 5.8e, 5.9e, 5.10e, 5.12e, 5.13e, 5.14e).

An alternative interpretation is that the absence of palaeosols and coastal-plain deposits may reflect the construction of a broad, subaqueous platform during progradation of each clinoform set in the Krossfjord and Fensfjord formations (Holgate et al., 2013). In deltaic successions, the clinoform break, located between the topset and foreset of clinoforms, may represent the position of the shoreline (called the "shoreline break") and therefore the transition zone between subaqueous and subaerial facies (Posamentier et al., 1988; Pirmez et al., 1998; Van Wagoner et al., 2003). With high-energy marine conditions, governed by the spatial and
temporal interplay between sediment supply and basin hydrodynamics, the position of the shoreline break may become laterally separated by a subaqueous platform (Pirmez et al., 1998). This scenario exhibits a subaerial clinoform, composed of shoreface and coastal-plain facies (<40 m in height), laterally separated from a deeper subaqueous clinoform, typically composed of mud- or sand-prone subaqueous facies (<100 m in height) (Pirmez et al., 1998; Driscoll and Karner, 1999; Swenson et al., 2005; Helland-Hansen and Hampson, 2009; Mitchell et al., 2012). In this arrangement, the clinoform break in the subaerial clinoform is closely related to the shoreline position above the fairweather wave base whereas the clinoform break in the subaqueous clinoform is situated below the fairweather wave base and at highly variable distances (up to tens of km) from the shoreline (Kuehl et al., 1997; Pirmez et al., 1998; Hernández-Molina et al., 2000). The resulting clinoform geometry of a subaerial and subaqueous delta, called a “compound clinoform”, may reflect two-stage progradation, for example in the Amazon subaqueous delta (Nittrouer et al., 1986), or synchronous progradation of the subaerial and subaqueous delta, for example in the Yangtze River Delta system (Liu et al., 2006).

Deposition of the Krossfjord and Fensfjord formations within a subaqueous delta explains both the absence of subaerial exposure and the widespread occurrence of well-developed clinoform topsets in seismic data (e.g. Figs. 5.5e, 5.6e, 5.7e, 5.8e, 5.9e, 5.10e, 5.12e, 5.13e, 5.14e). These features have also been identified and attributed to ancient examples of subaqueous deltaic deposition in the Cretaceous Wise Gulch, Berry Gulch and Morapos sandstones, US Western Interior (Hampson et al., 2008a) and the Down Cliff Clay Member, Southern UK (Hampson et al., 2014). Furthermore, subaqueous delta systems develop in areas subject to significant wave and/or tide energy, which enables sediment reworking and transport over long distances both offshore and alongshore (e.g. Swenson et al., 2005; Plink-Björklund, 2012). Evidence of wave and tidal processes is ubiquitous in core data from the Krossfjord and Fensfjord formations (Holgate et al., 2013).
5.7.3 Combined subaqueous delta and spit depositional model for the Krossfjord and Fensfjord formations

The discussion above suggests that whilst the mechanism of transgressive erosion to explain the absence of coastal plain deposits cannot be unequivocally discounted, the majority of features identified in the Krossfjord and Fensfjord formations concur with deposition in a constantly subaqueous environment (Table 5.3). Below a new depositional model for the Krossfjord and Fensfjord formations is presented, which includes two key depositional elements: (1) a subaqueous delta in the Troll Field; and (2) a coeval spit system in the area to the southeast of the Troll Field (Fig. 5.16).

Subaqueous delta

Each clinoform set or parasequence (‘series’) in the Troll Field shares the following characteristics: (1) a sand-rich composition; (2) clinoform strike is broadly NNE-SSW with greater variation in the northeast corner of the Troll Field only; (3) clinoform dip and height increase from east to west; and (4) grain size increases from east to west. In the following subsection, each of these characteristics is used to critically assess the validity of a subaqueous deltaic depositional model.

The majority of the clinoforms within the Krossfjord and Fensfjord formations are sand-rich (Holgate et al., 2013), due to high volumes of sand-grade material sourced from an uplifted rift shoulder in the east during the Middle-to-Late Jurassic (Sømme et al., 2013). Redistribution of sediment, through single or multiple fluvial source(s), is inferred to have occurred via strong, southward-directed tidal and wave-driven currents (Fig. 5.16), which are identified as being essential for the formation of compound clinoforms with subaerial and subaqueous foresets (e.g. Swenson et al., 2005; Liu et al., 2006). However, many modern deltas with compound
clinoform geometries are composed of mud-rich subaqueous clinoforms, within which the subaqueous clinoform break is situated below the fairweather wave base (e.g. the Amazon subaqueous delta; Kuehl et al., 1986a and the Ganges-Brahmaputra subaqueous delta; Kuehl et al., 1997). The principal facies associations identified within the Krossfjord and Fensfjord formations are very fine- to medium-grained, hummocky cross-stratified sandstone (Facies Association 2 of Holgate et al., 2013) and medium-grained, cross-bedded sandstone of varying wave- and tide-influence (Facies Associations 3 and 4 of Holgate et al., 2013). The hummocky cross-stratified sandstones correspond to a depositional setting beneath the fairweather wave base and therefore beneath the subaqueous clinoform break. The coarser-grained, cross-bedded sandstones, which are more prevalent in the west of the Troll Field, may have been deposited through strong, unidirectional tide- and wave-driven currents, causing high bed shear stresses, which can also extend beyond the subaqueous clinoform break (Mitchell et al., 2012). Analogous modern subaqueous clinoforms composed of sand-grade sediment have been identified in the Late Holocene prograding shallow-marine wedge off the south and south-eastern coasts of the Iberian Peninsula (Hernández-Molina et al., 2000) and the early Tortonian carbonate platform of Menorca, Balearic Islands (Pomar et al., 2002).

The strike of clinoforms within the Krossfjord and Fensfjord formations is broadly parallel to north-south-trending facies belts determined from analysis of core and wireline log data (Figs. 5.5c, 5.6c, 5.7c, 5.8c, 5.9c, 5.10c, 5.12c, 5.13c, 5.14c, 5.15) (Holgate et al., 2013). The uniform strike has an extent of c. 50 km north to south across the Troll Field (Fig. 5.15), although the true strike-extent is indeterminate due to the geographically-limited dataset (Fig. 5.2). The parallel strike is consistent with modern subaqueous deltas, which are sculpted by high-energy, hydrodynamic currents that distribute sediment across the shelf, as identified in mud-rich, Late Holocene Gargano subaqueous delta, Adriatic shelf (Cattaneo et al., 2003) and the sand-rich, Holocene subaqueous shallow-marine deposits in the SE Iberian Peninsula (Fernández-Salas et al., 2009; Ortega-Sánchez et al., 2014). Whilst the strike-extent of subaqueous clinoform-bearing
bodies is large in mud-rich subaqueous deltas (e.g. 100’s of km; Kuehl et al., 1986b), sand-grade subaqueous deltas have a more analogous strike-extent to the Krossfjord and Fensfjord formations (e.g. 10’s of km; Hernández-Molina et al., 2000). The greater variability in clinoform strike and prevalence of oblique clinoforms in the north east corner of the Troll Field is consistent with proximity to a point source of sediment input (Driscoll and Karner, 1999), as interpreted also for the Sognefjord Formation (Dreyer et al., 2005; Patruno et al., In Press).

Clinoforms within the Krossfjord and Fensfjord formations have steep dips (3-17°) with relatively short dip-extents (<350 m) (Figs. 5.5d, 5.6d, 5.7d, 5.8d, 5.9d, 5.10d, 5.12d, 5.13d, 5.14d, 5.15). Conversely, most modern, mud-prone subaqueous deltaic clinoforms are shallowly dipping (<1°), such as those recorded in the subaqueous Mekong Delta, Southern Vietnam (Unverricht et al., 2013) and the Po Delta system (Correggiari et al., 2005), with greater dip extents (e.g. c. 12 km in the Yellow River subaqueous delta; Liu et al., 2013). A rare example of steeper clinoform dips (2-3°) is also observed in the ancient subaqueous Down Cliff Clay Member (Hampson et al., 2014). The lesser dip extent and greater dip of clinoforms within the Krossfjord and Fensfjord formations may be attributed to their coarse grain size (e.g. Orton and Reading, 1993). This is reinforced by the analogous sand-prone, Late Holocene prograding shallow-marine wedge which contains steeply-dipping clinoforms (>5°), calculated from high-resolution seismic profiles offshore Cabo de Gata, Almeria (Hernández-Molina et al., 1995; Hernández-Molina et al., 2000). Furthermore, the westward increase in grain size and clinoform dip is consistent with a progressive change from predominantly tidal reworking in shallow bathymetry in the east, to predominantly wave reworking in deeper bathymetry in the west (Figs. 5.15, 5.16). Similar deepening of bathymetry has been interpreted to cause steepening of modern, sand-prone subaqueous clinoforms in the analogous Late Holocene prograding shallow-marine wedge off the south and south-eastern coasts of the Iberian Peninsula (Hernández-Molina et al., 2000). Additionally, increasing subaqueous foreset dips have been modelled to be a result of increasing sediment supply through time (Swenson et al., 2005),
perhaps reflecting the increased wave-driven current activity in the west of the Troll Field as interpreted from core data (Fig. 5.15).

Modern subaqueous deltas in high-energy marine settings display a “compound clinoform” architecture, in which a subaerial clinoform (subaerial delta) and subaqueous clinoform (subaqueous delta) coexist (Swenson et al., 2005). Subaerial clinoform deposits are not identified in the Krossfjord and Fensfjord formations, either because they are too small or are formed of facies with too little acoustic variability to be seismically imaged (e.g. Holgate et al., In Press), or because the width of the subaqueous platform between the subaerial and subaqueous deltas, which can be tens of kilometres (e.g. Kuehl et al., 1997), is greater than the width of the study area. The latter explanation is favoured, as no evidence for subaerial exposure is found in core or wireline-log data in the Troll Field, and the subaerial delta is therefore assumed to lie beyond the eastern limit of the study area (Fig. 5.16).

Spit system

Clinoforms in ‘Series 2’ in the area to the southeast of the Troll Field define a finger-like body that prograded to the south but with flanks that dipped to the west and east (Figs. 5.11, 5.16d). The clinoform-bearing body is a minimum of 30 km in length, and is located at a distance of c. 40 km from the fluvial sediment point source interpreted for the subaqueous delta in the Troll Field. Seismic data image a mounded feature separating the eastward- and westward-dipping clinoforms on the flanks of the sediment body (Fig. 5.11b). Coeval landward-dipping and seaward-dipping clinoforms have been recognised in prograding delta lobes (e.g. Gani and Bhattacharya, 2005) and in prograding spit and barrier systems (e.g. Nielsen et al., 1988). These two interpretations are not mutually exclusive, as oblique wave approach may generate asymmetric wave-dominated deltas (Bhattacharya, 2006) containing spits that extend for up to tens of kilometres (e.g. Sahalin Spit of the modern Danube Delta; Dan et al., 2011). In this context, the mounded feature separating eastward- and westward-dipping clinoforms in the
area to the southeast of the Troll Field (Fig. 5.11b) may be a partially subaerial spit ridge that
developed on top of a large subaqueous spit platform (Fig. 5.16a, b, d) (cf. Meistrell, 1972; 
Nielsen et al., 1988). Recent and modern spits are coarse-grained and contain clinoforms with
steep dips (25-30° in Lyngså spit system; Nielsen et al., 1988) than those observed in ‘Series 2’
(<1°; Fig. 5.11b, c). It is assumed that the latter were predominantly fine-grained, consistent
with their gentle dips and the distance from the potential sediment source to the northeast of
the Troll Field. The seismic response of the mounded feature on top of the clinoform-bearing
body appears similar to cross-sections of beach ridges identified in seismic data (e.g. Jackson et
al., 2010). Beach ridges are a common surface feature on subaerially exposed spits (Nielsen and
Johannessen, 2009), although their occurrence in the area to the southeast of the Troll Field
cannot be proven in the absence of well data. Spit development is only interpreted in ‘Series 2’,
implying either locally strong longshore sediment transport during deposition of this ‘series’, or
a change in the angle of wave approach and associated shoreline morphology during spit
development (Ashton and Murray, 2006). These mechanisms may reflect rift-related changes in
sediment routing and/or nearshore bathymetry outside of the immediate study area during
deposition of ‘Series 2’.

The establishment of a spit system to the southeast of the Troll Field is not inconsistent with the
coeval development of a subaqueous delta in the Troll Field (Fig. 5.16), because both spit and
subaqueous delta development are dependent on high wave and current energy parallel to the
coastline (Swenson et al., 2005; Nielsen and Johannessen, 2009). It is possible that the spit was
fed from a second point source of sediment that was located south of the postulated point
source for the subaqueous “Troll Delta” (e.g. Fig. 5.16a). In this interpretation, the spit is
genetically unrelated to the subaqueous delta. An alternative explanation is provided by the
Holocene Sf. Gheorghe mouth and the modern Sacalin (termed “Sahalin” by Dan et al., 2011) Spit
of the Danube Delta (Bhattacharya and Giosan, 2003; Giosan et al., 2005), in which asymmetric
development of a spit in the subaerial delta due to longshore sediment transport is replicated in
the subaqueous platform in front of the delta (e.g. Fig. 5.16b). An offset in the plan-view location of the spit and genetically related subaqueous delta may be developed under process regimes in which tides are an important component (e.g. Nittouer et al., 1986), as interpreted in the Troll Field (Holgate et al., 2013).

5.7.4 Is the lower Viking Group on the Troll field a “syn-rift” succession?

Although the Horda Platform was a relatively stable block (Færseth, 1996), the Krossfjord and Fensfjord formations have been referred to as “syn-rift” because they were thought to have been deposited during Middle-to-Late Jurassic rifting (Ravnås and Bondevik, 1997). Based on analysis of well data only, Holgate et al. (2013) suggested only a very limited tectonic control on sedimentation during the deposition of the Krossfjord and Fensfjord formations in the Troll Field area. Detailed seismic-stratigraphic analysis undertaken in this study concurs with this interpretation as no tectonic influence is evident throughout deposition of ‘Series 1a’ to ‘Series 3b’. Minor stratal thickening of ‘Series 3c’ is identified in the hangingwalls of the Svartalv and Tusse fault systems, central and western part of the Troll Field (Fig. 5.14a), implying increased accommodation due to normal fault growth during the latest Callovian (Fig. 5.15). The timing of this rift pulse is consistent with previous tectono-stratigraphic interpretations of the area (Færseth and Ravnås, 1998; Ravnås et al., 2000), and with the interpretation that deposition of the Heather “B” unit shales records abandonment of the Krossfjord-Fensfjord depositional system in response to increased, fault-driven tectonic subsidence (Steel, 1993). Whilst there is an absence of seismically resolvable onlap into the hangingwall dipslope and lack of evidence for erosion of the footwall crest, the thickening of the ‘series’ across the fault implies that fault growth occurred during the earliest ‘rift initiation’ (sensu Prosser, 1993) phase.

Minor thickness variations over faults are also identified in the overlying, Oxfordian-aged Sognefjord Formation, but a structural control on sedimentation in the Troll Field is not
identified (Patruno et al., In Press). Furthermore, a subaqueous delta model is also invoked to explain clinoform geometries and sedimentological character in the Sognefjord Formation, which are strikingly similar to those described herein for the Krossfjord and Fensfjord formations. The similarities in sedimentological style between the Krossfjord and Fensfjord formations and the Sognefjord Formation therefore suggest a consistent depositional regime in a long-lived (c. 20 Myr from Bathonian to Oxfordian) subaqueous delta in the area encompassing the Troll Field. The long-lived subaqueous delta forming the reservoir sandstones of the Troll Field is a novel shallow-marine reservoir type. Its occurrence is attributed to the interplay between sediment supply and basinal hydrodynamics on the relatively tectonically quiescent Horda Platform during this time, and proximity to a localised source of abundant coarse-grained sediment supply on the uplifted rift shoulder (Sømme et al., 2013).

5.8 Conclusions

Seismic-stratigraphic analysis of the Middle-to-Late Jurassic Krossfjord and Fensfjord formations in the Troll Field, Norwegian North Sea, indicates nine progradational clinoform sets (parasequences) that are stacked into three previously interpreted parasequence sets. Each clinoform set contains sigmoidal clinoforms that are 60-350 m long, 11-46 m thick, and dip towards the west at 3-17°. Clinoform dip increases from east to west, coincident with an increase in grain size. Clinoforms are generally gently curvilinear in plan-view, and consistently strike from NNE-SSW in the east to N-S in the west of the Troll Field, but exhibit highly variable strike and oblique cross-sectional geometry in the northeast of the field. Clinoforms have well-developed topsets that lack evidence for subaerial exposure. Clinoform trajectory analysis in dip-oriented cross-sections of limited extent indicates a near-horizontal, progradational trajectory in each clinoform set. Sigmoidal-to-shingled clinoforms of greater length (5-14 km), greater thickness (c. 50-85 m) and shallower dip (<1°) define a finger-like sediment body that
prograded to the south with westward- and eastward-dipping flanks in the area to the southeast of the Troll Field.

These geometrical characteristics are consistent with recently published interpretations based on core and wireline-log data that the Krossfjord and Fensfjord formations in the Troll Field were deposited as part of a long-lived, wave- and tide-influenced, subaqueous delta complex. It is envisaged that coarse-grained sediment was supplied from a fluvio-deltaic point source in the northeast of the Troll Field and redistributed through southward-directed longshore currents. The finger-like, southward-prograding clinoform-bearing body lying to the southeast of the Troll Field is interpreted as a spit developed landward and alongshore of the subaqueous "Troll Delta". The coeval development of a spit and sand-prone subaqueous delta requires two distinct sediment-input points and/or alongshore redistribution of large volumes of sand by waves and tides. Time-thickness maps over the Troll Field indicate that active rifting influenced stratigraphic architecture only during deposition of the uppermost clinoform set, in the west of the field.

The analysis presented above highlights the complex configuration of shallow-marine stratal geometries and sandstone distributions due to the interplay between sediment supply and basinal hydrodynamics. The interpreted sand-rich, subaqueous delta fronting a spit is a novel reservoir type which may often be overlooked in studies of ancient shallow-marine reservoir units. The sand-prone subaqueous delta system persisted during deposition of the overlying Sognefjord Formation, indicating that it was a long-lived feature in the area encompassing the Troll Field throughout the Middle-to-Late Jurassic. The identification of a southward-prograding spit in the Fensfjord Formation and the southward extension of the subaqueous delta beyond the limits of the study area indicate the potential for Middle-to-Late Jurassic reservoir sandstones to exist in unexplored areas to the south of the Troll Field.
5.9 Acknowledgements

Paul Whipp, Stefano Patruno, Aruna Mannie, Rebecca Bell, Thilo Wrona, Matt Lewis and Clara Rodriguez are thanked for discussions during the course of this study. Statoil ASA and TGS are thanked for providing data. Partners in the Troll production licenses PL054, PL085, PL085 B, and PL085 C (Petoro AS, Statoil Petroleum AS, A/S Norske Shell, Total E&P Norge AS, and ConocoPhillips Skandinavia AS) are thanked for supporting the provision of data to undertake this study and their permission to publish the results. The views expressed in this chapter are the authors and do not necessarily represent those of the Troll license partners. Thanks also to Schlumberger Limited for provision of Petrel software via an academic software donation, Midland Valley for provision of Move software and Blueback Reservoir for provision of Blueback Geophysics Toolbox.
Figure 5.1. (over page) (a) Simplified map of the North Viking Graben, North Sea Basin, highlighting the Horda Platform, and the Troll, Brage, Fram and Gjøa fields, which host Krossfjord and Fensfjord reservoirs (modified after Færseth, 1996; Ravnås and Bondevik, 1997; Fraser et al., 2002); (b) Simplified palaeoenvironmental map of the northern North Sea during mid- to late Callovian deposition of the Fensfjord Formation in the Troll Field (modified after Husmo et al., 2002); (c) Geoseismic profile illustrating the major fault blocks from west to east across the Viking Graben (see a for location) (modified after Husmo et al., 2002); and (d) Middle to Upper Jurassic chronostratigraphic framework for a SW-NE-oriented cross-section through the North Viking Graben and Horda Platform (see a for location) (modified after Partington et al., 1993; Stewart et al., 1995; Fraser et al., 2002).

Figure 5.2. (page 196) Simplified map of the Troll Field area (Fig. 5.1a) illustrating the distribution of 3D seismic cubes, 2D seismic lines and wells used by this study (modified after NPD, 2013b).

Figure 5.3. (page 197) Seismic-well ties for (a) well 31/2-1 and (b) well 31/6-8 (see Fig. 5.2 for location). Seismic data is Reverse “European” Polarity, whereby a trough (red) reflection represents an increase in acoustic impedance and a peak (black) reflection represents a decrease in acoustic impedance.

Figure 5.4. (page 198) Example of the extensive quantitative analysis conducted on seismically imaged clinoforms in the Troll Field including: (a) Troll Field map illustrating the number of clinoforms mapped in their 3D extent per series (in this example within ‘Series 2’); (b) 3D seismic section illustrating a single clinoform mapped within ‘Series 2’ and used to identify clinoform strike direction (see a for location); and (c) definition of clinoform geometrical parameters (length, height, dip) measured from a decompacted, dip-oriented seismic cross-section flattened on an underlying palaeohorizontal datum (see a for location).
Figure 5.1c: Map showing the distribution of geological formations and faults in the North Sea region. The map highlights the major sediment input points, erosion/non-deposition areas, and major normal faults.

Figure 5.1d: Cross-section figures illustrating the stratigraphic sequence from the Quaternary to the Upper Jurassic. The section includes the Troll Field (Upper Jurassic) and shows the migration of sedimentary units from west to east.

Table 5.1: Age and stage of the geological strata in the North Sea region. The table lists the age in Ma and stage of each stratum, along with the corresponding formation names.
Synthetic Seismogram and Seismic Cross-Section for Well 31/2-1
Seismic to West  Seismic to East
Pythonect
Figure 5.5. (over page) Seismic-stratigraphic architecture of ‘Series 1a’: (a) time-thickness map of ‘Series 1a’, from the J32 to J33 maximum flooding surfaces; (b) maximum amplitude map extracted from window near the base of ‘Series 1a’, at 10 ms above flattened J32 surface; (c, d) Troll Field maps showing strike and dip direction of selected clinoforms within ‘Series 1a’ (black). Clinoforms selected for decompaction are coloured to show (c) their maximum height, and (d) their dip; and (e) uninterpreted (upper) and interpreted (lower) seismic cross-section oriented approximately perpendicular to clinoform strike (see a for location), flattened on the J32 surface, and highlighting downlap of the J33 surface onto the J32 surface and clinoforms within ‘Series 1a’.

Figure 5.6. (page 201) Seismic-stratigraphic architecture of ‘Series 1b’: (a) time-thickness map of ‘Series 1b’ from the J33 to J34 maximum flooding surfaces; (b) maximum amplitude map extracted from window within ‘Series 1b’, at 10 ms above flattened J33 surface; (c, d) Troll Field maps showing strike and dip direction of selected clinoforms within ‘Series 1b’ (black). Clinoforms selected for decompaction are coloured to show (c) their maximum height; and (d) their dip; and (e) uninterpreted (upper) and interpreted (lower) seismic cross-section oriented approximately perpendicular to clinoform strike (see a for location), flattened on the J33 surface.

Figure 5.7. (page 202) Seismic-stratigraphic architecture of ‘Series 1c’: (a) time-thickness map of ‘Series 1c’ from the J34 to J36 maximum flooding surfaces. J-surface data points show well control for the seismically mapped J36 surface; (b) maximum amplitude map extracted from window within ‘Series 1c’ at 10 ms above flattened J34 surface; (c, d) Troll Field maps showing strike and dip direction of selected clinoforms within ‘Series 1c’ (black). Clinoforms selected for decompaction are coloured to show (c) their maximum height; and (d) their dip; and (e) uninterpreted (upper) and interpreted (lower) seismic cross-section oriented approximately perpendicular to clinoform strike (see a for location), flattened on the J34 surface.
Chapter 5
Figure 5.8. (over page) Seismic-stratigraphic architecture of 'Series 1d': (a) time-thickness map of 'Series 1d' and 'Series 1e' from the J36 to J42 maximum flooding surfaces. J-surface data points show well control for the seismically mapped J42 surface; (b) maximum amplitude map extracted from window within 'Series 1d' at 10 ms above flattened J36 surface; (c, d) Troll Field maps showing strike and dip direction of selected clinoforms within 'Series 1d' (black). Clinoforms selected for decompaction are coloured to show (c) their maximum height; and (d) their dip; and (e) uninterpreted (upper) and interpreted (lower) seismic cross-section oriented approximately perpendicular to clinoform strike (see a for location), flattened on the J36 surface.

Figure 5.9. (page 205) Seismic-stratigraphic architecture of 'Series 1e': (a) time-thickness map of 'Series 1d' and 'Series 1e' from the J36 to J42 maximum flooding surfaces. J-surface data points show well control for the seismically mapped J42 surface; (b) maximum amplitude map extracted from window within 'Series 1e' at 40 ms above flattened J36 surface; (c, d) Troll Field maps showing strike and dip direction of selected clinoforms within 'Series 1e' (black). Clinoforms selected for decompaction are coloured to show (c) their maximum height; and (d) their dip; and (e) uninterpreted (upper) and interpreted (lower) seismic cross-section oriented approximately perpendicular to clinoform strike (see a for location), flattened on the J36 surface.
Figure 5.10. (over page) Seismic-stratigraphic architecture of ‘Series 2’: (a) time-thickness map of ‘Series 2’ from the J42 to J44 maximum flooding surfaces. J-surface data points show well control for the seismically mapped J44 surface; (b) maximum amplitude map extracted from window within ‘Series 2’ at 10 ms above flattened J42 surface; (c, d) Troll Field maps showing strike and dip direction of selected clinoforms within ‘Series 2’ (black). Clinoforms selected for decompaction are coloured to show (c) their maximum height; and (d) their dip; and (e) uninterpreted (upper) and interpreted (lower) seismic cross section oriented approximately perpendicular to clinoform strike (see a for location), flattened on the J42 surface.

Figure 5.11. (page 208) Seismic-stratigraphic architecture of ‘Series 2’ in area to southeast of the Troll Field: (a) map of the Troll Field and surrounding area showing apparent dip direction of selected clinoforms traced along 2D seismic lines and (b, c) uninterpreted (upper) and interpreted (lower) 2D seismic cross-sections (see a for location), flattened on the J42 surface, highlighting (b) clinoforms dipping both east and west on either flank of a domed feature; and (c) southerly dipping clinoforms. Data courtesy of TGS.
Figure 5.12. (over page) Seismic-stratigraphic architecture of ‘Series 3a’: (a) time-thickness map of ‘Series 3’ from the J44 to J46 maximum flooding surfaces. J-surface data points show well control for the seismically mapped J46 surface; (b) maximum amplitude map extracted from window within ‘Series 3a’ at 10 ms above flattened J44 surface; (c, d) Troll Field maps showing strike and dip direction of selected clinoforms within ‘Series 3a’ (black). Clinoforms selected for decompaction are coloured to show (c) their maximum height; and (d) their dip; and (e) uninterpreted (upper) and interpreted (lower) seismic cross-section oriented approximately perpendicular to clinoform strike (see a for location), flattened on J44 surface.

Figure 5.13. (page 211) Seismic-stratigraphic architecture of ‘Series 3b’: (a) time-thickness map of ‘Series 3’ from the J44 to J46 maximum flooding surfaces. J-surface data points show well control for the seismically mapped J46 surface; (b) maximum amplitude map extracted from window within ‘Series 3b’ at 40 ms above flattened J44 surface; (c, d) Troll Field maps showing strike and dip direction of selected clinoforms within ‘Series 3b’ (black). Clinoforms selected for decompaction are coloured to show (c) their maximum height; and (d) their dip; and (e) uninterpreted (upper) and interpreted (lower) seismic cross-section oriented approximately perpendicular to clinoform strike (see a for location), flattened on J44 surface.

Figure 5.14. (page 212) Seismic-stratigraphic architecture of ‘Series 3c’: (a) time-thickness map of ‘Series 3’ from the J44 to J46 maximum flooding surfaces. J-surface data points show well control for the seismically mapped J46 surface; (b) maximum amplitude map extracted from window within ‘Series 3c’ at 70 ms above flattened J44 surface; (c, d) Troll Field maps showing strike and dip direction of selected clinoforms within ‘Series 3c’ (black). Clinoforms selected for decompaction are coloured to show (c) their maximum height; and (d) their dip; and (e) uninterpreted (upper) and interpreted (lower) seismic cross-section oriented approximately perpendicular to clinoform strike (see a for location), flattened on J44 surface.
Figure 5.15. (over page) Block diagrams illustrating temporal evolution of stratal architecture and depositional environments across the Troll Field during the deposition of: (a) ‘Series 1a’ (Fig. 5.5), between J32 and ?J33 maximum flooding surfaces (?early Bathonian); (b) ‘Series 1b’ (Fig. 5.6), between ?J33 and ?J34 maximum flooding surfaces (?middle Bathonian); (c) ‘Series 1c’ (Fig. 5.7), between ?J34 and J36 maximum flooding surfaces (?late Bathonian); (d) ‘Series 1d’ (Fig. 5.8), and (e) ‘Series 1e’ (Fig. 5.9), between J36 and J42 maximum flooding surfaces (early Callovian); (f) ‘Series 2’ (Fig. 5.10), between J42 and J44 maximum flooding surfaces (middle Callovian); (g) ‘Series 3a’ (Fig. 5.12), (h) ‘Series 3b’ (Fig. 5.13), and (i) ‘Series 3c’ (Fig. 5.14), between J44 and J46 maximum flooding surfaces (middle-late Callovian).

Figure 5.16. (page 215) Interpretive palaeogeographic maps of the Troll Field and adjacent areas. Two depositional models are proposed with development of a wave-dominated, tide-influenced subaqueous delta during early progradation of ‘Series 2’ ($t = 1$) and a wave-dominated subaqueous delta during late progradation of ‘Series 2’ ($t = 2$) with: (a) coeval spit growth from a separate source south of the Troll Field and (b) coeval spit growth from the same northern source. Idealised cross-sections of the westward-prograding subaqueous delta of the Troll Field (c), and the southward-prograding spit southeast of the Troll Field (d) relate to both depositional scenarios. Maps and cross-sections highlight the 3D configuration of stratal geometries and depositional environments.
KEY
- Un-cored well
- Wave-dominated sandstone
- Wave-dominated, tide-influenced sandstone
- Tide-dominated, wave-influenced embayment sandstone
- Delta-front sandstone
- Offshore

Core facies at maximum regression:
- Wave-dominated sandstone
- Wave-dominated, tide-influenced sandstone
- Tide-dominated, wave-influenced embayment sandstone
- Delta-front sandstone
- Offshore

Maximum flooding J-surfaces:
- MFS J32
- MFS ?J33
- MFS ?J34
- MFS J36
- MFS J42
- MFS J44

Clinoform (1:10000 scale):
- Inactive normal fault
- Potentially active normal fault
- Clinoform strike

N 20 km

VE TUSV TR

Chapter 5
### 3D Seismic Dataset

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**Table 5.1.** Summary of seismic surveys used in this study
Table 5.2. Detailed description of the seismic reflections mapped for the purpose of this study.
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<th>Observed characteristics of the Krossfjord and Fensfjord formations</th>
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<th>Modern subaqueous deltas</th>
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<td>Absence of coastal plain facies</td>
<td>✓ (e.g. Ryer, 1977)</td>
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<td>Evidence for high-energy, wave- and tide-driven hydrodynamics (sedimentary structures/sorting)</td>
<td>✓ (e.g. Bhattacharya and Walker, 1991)</td>
<td>✓ (e.g. Hernández-Molina et al., 2000)</td>
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<tr>
<td>Transgressive lag deposits</td>
<td>✓ (e.g. Bhattacharya and Walker, 1991)</td>
<td>✓ (e.g. Cattaneo et al., 2003)</td>
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<td>High-angle clinoform foreset dip (&gt;1°)</td>
<td>✓ (e.g. Uličný, 2001)</td>
<td>✓ (e.g. Hernández-Molina et al., 1995; Hernández-Molina et al., 2000)</td>
</tr>
<tr>
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<td>Presence of clinoform topsets</td>
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<td>✓ (e.g. Liu et al., 2013)</td>
</tr>
<tr>
<td>Shore detached clinoform break</td>
<td>✓ (e.g. Hampson and Howell, 2005)</td>
<td>✓ (e.g. Pirmez et al., 1998)</td>
</tr>
<tr>
<td>Long-distance regression</td>
<td>✓ (typically with forced regression; Posamentier and Morris, 2000)</td>
<td>✓ (e.g. Swenson et al., 2005)</td>
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</tbody>
</table>

**Table 5.3.** Comparison of characteristics observed in the Krossfjord and Fensfjord formations, Troll Field, with their presence (✓) or absence (✗) in ancient, top-truncated, normal regressive, sand-rich deltas and modern subaqueous deltas.
CHAPTER 6

Conclusions

The Krossfjord and Fensfjord formations, Troll Field, Horda Platform, offshore western Norway form a c. 195 m thick clastic wedge that encompasses a 13 Myr period from the Bathonian to the Callovian. They were deposited during the regional Middle-to-Late Jurassic rift event in the North Sea Basin, and are composed of sandstones that were sourced from the Norwegian mainland to the east and which pinch out basinwards to the west into offshore shales of the Heather Formation. Together the formations form part of the Viking Group, and are a secondary oil and gas reservoir in the super-giant Troll Field. They also form part of the reservoir in the Brage Field and a prospective reservoir in the area around the Gjøa Field. Despite their potential economic significance, the character, distribution, and stratigraphic architecture of these sandbodies are poorly understood, as they have not been the focus of detailed previous work.

As stated in Chapter 1, the aim of this thesis is to characterise the sedimentology and stratigraphic architecture of the Krossfjord and Fensfjord formations in the Troll Field, in order to provide insights into the distribution and character of the reservoir sandstones within the Troll Field and elsewhere on the Horda Platform. Two objectives are addressed to fulfil the aim of the thesis, as discussed below.

6.1 Sedimentology and sequence stratigraphy of the Krossfjord and Fensfjord formations, Troll Field

Core and biostratigraphic analysis presented in Chapter 3 identified eleven facies, which were grouped into six facies associations representing wave- and tide-dominated deltaic, shoreline and shelf depositional environments. No coastal-plain deposits were identified. Quantitative
Wireline-log analysis indicated limited disparity between sandstone-dominated facies associations, which has prevented detailed facies identification in uncored wells.

Wireline-log interpretations, sequence stratigraphic correlations and biostratigraphic analysis enabled the identification of regionally correlatable maximum flooding surfaces and locally developed transgressive surfaces. Maximum flooding surfaces tied to the stratigraphical “J-surface” scheme of Partington et al. (1993) were interpreted as the regional J32 (latest Bajocian, 165 Ma), J42 (early Callovian, 155.5 Ma), J44 (middle Callovian, 154 Ma) and J46 (late Callovian, 152 Ma) surfaces. These surfaces defined three ‘series’ that each contain multiple parasequences. ‘Series 1’, from J32 to J42 maximum flooding surfaces, contains the Heather “A” unit, the Krossfjord Formation and the lowermost Fensfjord Formation; ‘Series 2’, from J42 to J44 maximum flooding surfaces, contains the lower Fensfjord Formation; and ‘Series 3’, from J44 to J46 maximum flooding surfaces, contains the upper Fensfjord Formation and Heather “B” unit. Based on well correlations and seismic data, the three ‘series’ exhibit laterally uniform thicknesses, implying no major tectonic influence on sedimentation.

Palaeogeographic maps, constrained by limited core data, identified a complex distribution of depositional environments across the Troll Field. A wide (10–20 km), north-south-trending belt of wave-dominated shoreface deposits is present in the western part of the field, but its eastern part contains more irregular (0-20 km wide), north-south-trending belts of mixed wave- and tide-influenced shoreface and tide-dominated embayment deposits. Vertically through the Krossfjord and Fensfjord formations, an increase in the proportion of tide-influenced deposits identified in the east of the Troll Field was interpreted to be the result of progradational to retrogradational stacking of regressive-transgressive tongues (‘series’ or ‘genetic sequences’ sensu Galloway, 1989).
The complex temporal and spatial variations in depositional process regime identified through core and wireline-log analysis highlighted the inherent limitations of using the delta tripartite process model (e.g. Galloway, 1975). Furthermore, studies have described the dangers of using a limited subsurface well database and extrapolating from potentially atypical data points (e.g. Bhattacharya and Giosan, 2003). It was therefore intended that detailed seismic interpretation, away from the sparse well-data control, would enable a reassessment of local and regional controls on stratigraphic architecture, as described below.

6.2 Seismic stratigraphic and clinoform architecture of the Krossfjord and Fensfjord formations, Troll Field area

Analysis of 3D seismic reflection data indicated that both formations contain seismic-scale clinoforms, the geometry, distribution, and trajectory of which could be used to help predict the lithological character and distribution of the shallow-marine sandstones. However, the limited availability and distribution of core and wireline-log data restricted the detail and confidence with which seismically imaged architectures can be interpreted. Forward seismic modelling, presented in Chapter 4, was therefore used to characterise the seismic expression of clinoforms in different deltaic depositional environments. Three outcrop analogues from the Western Cretaceous Interior Seaway, US, were studied to help constrain interpretations within the Krossfjord and Fensfjord formations. The results of the chapter indicate that clinoforms were imaged in the forward seismic models when they were: (1) spaced wider than the tuning thickness; (2) marked by pronounced interfingering of facies associations with different acoustic properties; and/or (3) lined by relatively thick (>50 cm) carbonate-cemented layers. However, where clinothems were thinner than the vertical resolution limit of seismic data, destructive interference occurred creating misleading geometrical relationships. Consequently, closely spaced clinoforms, which typify fluvial-dominated deltas, were unlikely to be fully resolved. The ability to image clinoforms was also found to be dependent on: (1) the frequency
of the seismic wavelet; (2) the overburden velocity; and/or (3) the acoustic impedance contrast at the boundary between the overburden and the clinoform-bearing reservoir target.

The methodology presented in Chapter 4 is a novel approach to characterising delta-scale clinoforms imaged in seismic reflection data. The results illustrated how delta-scale clinoforms could be used to characterise depositional process regime (e.g. Helland-Hansen and Hampson, 2009) and therefore potentially aid the prediction and distribution of genetically related sedimentary successions (Catuneanu et al., 2009). Furthermore, forward seismic modelling aided identification of small-scale heterogeneities within clinothems which could affect production strategies within deltaic reservoir sandstones (e.g. Jackson et al., 2009). Whilst Chapter 4 is specifically focussed on the characterisation of delta-scale clinoforms, the methodology established is widely applicable to other depositional systems. In particular, the methodology helped identify the effect small-scale lithological heterogeneities and variations in geometry have on seismic reflection data, which may aid the characterisation and production of deep-water (e.g. Armitage et al., 2009) and carbonate (e.g. Shekhar et al., 2014) sedimentary systems.

Seismic-stratigraphic analysis of the Krossfjord and Fensfjord formations using the detail afforded by 3D seismic datasets in the Troll Field, presented in Chapter 5, helped evolve interpretations from initial sequence stratigraphic correlations based on well data. Nine progradational clinoform sets (parasequences) were identified and were stacked into three previously interpreted parasequence sets, ('series'). ‘Series 1’ contained five parasequences referred to as ‘Series 1a’ to ‘Series 1e’, ‘Series 2’ contained one parasequence and ‘Series 3’ contained three parasequences referred to as ‘Series 3a’ to ‘Series 3c’. Parasequence stacking patterns indicated punctuated progradation of shallow-marine deposits, from east to west across the Troll Field, from ‘Series 1a’ to ‘Series 1c’ and aggradational parasequence stacking from ‘Series 1d’ to ‘Series 3c’. Detailed seismic thickness mapping of each ‘series’ indicated that
a tectonic control, driven by slip on the Svartalv and Tusse fault systems, central and western part of the Troll Field, is limited to ‘Series 3c’. Thickening of the ‘series’ across these fault systems implied normal fault growth occurred during the latest Callovian and the earliest ‘rift initiation’ phase (sensu Prosser, 1993). Therefore, both well and seismic data indicate a very limited tectonic control on sedimentation during the deposition of the Krossfjord and Fensfjord formations in the Troll Field, despite the formations having previously been described as "syn-rift" (Ravnås and Bondevik, 1997).

Forensic mapping of clinoforms, a novel approach to record the geometrical characteristics of clinoforms in each ‘series’ within the Krossfjord and Fensfjord formations, was conducted to determine the detailed stratigraphic architecture and refine previously established depositional models. The mapping, presented in Chapter 5, illustrated that laterally, within all ‘series’ in the Troll Field, average clinoform height increased from east (21 m) to west (26 m), and their lengths shortened (from 171 m to 147 m) and dips increased (from 7° to 11°) in the same direction. Clinoform strike appeared to be oriented NE-SW in the east of the Troll Field compared to NNE-SSW in the west of the Troll Field, with sigmoidal clinoform geometries and well-developed topsets exhibited in both areas. Clinoform strike in the northeast of the Troll Field appeared highly variable from NE-SW to NW-SE with oblique clinoformal geometries, consistent with the proximity to a point source of sediment input. Clinoform trajectory analysis in dip-oriented cross-sections of limited extent indicated a near-horizontal, progradational trajectory in each clinoform set. Vertically, from ‘Series 1’ to ‘Series 3’, average clinoform length (from 178 m to 132 m) and height (from 29 m to 24 m) appeared to decrease, resulting in an increase in clinoform dip (from 9° to 11°).

The lack of evidence for subaerial exposure in core data, twinned with well-developed clinoform topsets in seismic data, suggest the Troll Field was fully subaqueous throughout the deposition of the Krossfjord and Fensfjord formations. Whilst an alternative mechanism of transgressive
erosion to explain the absence of coastal plain deposits could not be unequivocally discounted, the majority of features identified in the Krossfjord and Fensfjord formations concur with deposition in a constantly subaqueous environment. A supporting mechanism of falling sea-level during each parasequence-scale regression, inhibiting aggradation of the coastal plain, was discredited due to the absence of other depositional characteristics resulting from forced regression (e.g. Posamentier and Morris, 2000). In the absence of forced regression, the lateral extent of the Krossfjord and Fensfjord formations noted by Steel (1993) may be attributed to high volumes of sediment supplied with efficient redistribution by hydrodynamic forces. Concurrent fault-related rifting beyond the western limits of the study area, which can significantly alter lithological distributions through variations in sediment catchments, sediment dispersal and sediment sources, may also have been influential (e.g. Gawthorpe and Leeder, 2000).

The characteristics of the seismically imaged clinoforms identified in Chapter 5 combined with the arrangement of depositional environments identified in Chapter 3 suggest the Krossfjord and Fensfjord formations in the Troll Field were deposited as part of a wave- and tide-influenced, sand-rich subaqueous delta complex. High volumes of coarse-grained sediment was supplied from the uplifted rift shoulder, via a fluvo-deltaic point source in the northeast of the Troll Field, and redistributed via southward-directed longshore currents. The relatively tectonic quiescent Horda Platform during this time promoted a continuously subaqueous delta. The novel depositional model reiterates the importance of the complex interplay between sediment supply, sediment grade, basin bathymetry, hydrodynamic forces near the shoreline (e.g. wave-vs. tide- vs. fluvial-dominated) and wave- and tide-driven currents, as well as relative sea-level changes, in controlling the sediment distribution and stratigraphic architecture in deltaic depositional settings. Furthermore, the interpretation implies sand-rich, delta-scale clinoforms formed below the fairweather wave base and at significant distances away from the shoreline, contradicting the assumption that the clinoform break in delta-scale clinoforms approximates to
the position of the shoreline (e.g. Van Wagoner et al., 1988). Whilst caution over this assumption has been previously expressed in reference to settings with a fine-grained sediment composition (e.g. Helland-Hansen and Hampson, 2009), the sand-prone, subaqueous deltaic depositional model suggests analysis of delta-scale clinoforms, without a thorough understanding of the facies composition and depositional setting, should not be governed by the assumption that the clinoforms approximate to the shoreline position.

To the southeast of the Troll Field, sigmoidal-to-shingled clinoforms of greater length (5-14 km), greater thickness (c. 50-85 m) and shallower dip (<1°), compared to clinoforms in the Troll Field, defined a finger-like sediment body that prograded to the south with westward- and eastward-dipping flanks in ‘Series 2’. The clinoform-bearing body was interpreted as a spit developed landward and alongshore of the subaqueous “Troll Delta”. The coeval development of a spit and sand-prone subaqueous delta required two distinct sediment-input points and/or alongshore redistribution of large volumes of sand by waves and tides. Whilst the limitations of the dataset prohibit the ability to select a preferred mechanism, the coeval creation of a spit and delta again highlights the complex interactions between relative sea level, basin bathymetry, sediment input and the hydrodynamic processes that occur at and near the shoreline (e.g. wave-vs. tide- vs. fluvial-dominated). Furthermore, the identification of the spit existing during the middle Callovian only may suggest temporal variations in mechanisms outside the study area such as rift-related changes in sediment routing and/or nearshore bathymetry.

6.3 Further work

The Krossfjord and Fensfjord formations, together with the stratigraphically-younger Sognefjord Formation, form a long-lived (c. 20 Myr from Bathonian to Oxfordian) subaqueous delta deposited in the area that now encompasses the Troll Field. Whilst numerous modern subaqueous deltas have been documented (e.g. Nittrouer et al., 1986; Nittrouer et al., 1996;
Kuehl et al., 1997; Hori et al., 2002; Liu et al., 2006; Liu et al., 2013), there are comparatively few ancient systems (e.g. Hampson, 2010). The reservoir sandstones of the Troll Field therefore may be treated as a novel shallow-marine reservoir type. The limited number of ancient examples of this reservoir type suggests others may have been overlooked or misinterpreted where delta-scale subaqueous clinoforms are recognised. A re-evaluation of sandstone-rich shallow-marine reservoirs, which have similar sedimentological and stratigraphic architectures to the Viking Group in the Troll Field, may permit better prediction of the character, distribution and connectivity of the reservoir sandbodies.

The geographical limitations of the 2D and 3D seismic datasets prevented the identification of the southerly termination of the subaqueous delta and coeval spit. Whilst there are large variations in the down-current extent of subaqueous deltas documented in the published literature (from 10's to 100's of km), the southerly extent of the Troll Field subaqueous delta should be assessed. Furthermore, the southward-prograding spit to the southeast of the Troll Field is also potentially composed of reservoir-grade material and its full extent should be established. Analysis of well and seismic data outside the immediate study area may also provide evidence for the spit being restricted to the middle Callovian.

Further afield, the Krossfjord and Fensfjord formations are identified: (a) west of Troll in the Brage Field (e.g. well 31/4-11) where they form an oil reservoir; (b) immediately north of Troll in the Fram Field (e.g. 35/11-4) where the Sognefjord Formation is the principal reservoir; and (c) further north of Troll in the Gjøa Field (e.g. well 35/9-1) where they form part of the reservoir containing gas above a relatively thin oil zone, with traps defined by several tilted fault blocks (NPD, 2013a). The connectivity between these reservoir sandstones and those on the Troll Field is poorly understood and requires further work. Provenance analysis using detrital garnets (e.g. Morton et al., 2004) could be conducted to ascertain whether these sandstones are genetically related to single or multiple catchment areas, which would
potentially reveal their sediment transport pathways and reasons for their distribution. Furthermore, the establishment of a high resolution chronostratigraphic framework within the Krossfjord and Fensfjord formations would allow greater constraint on the variations in volume and timing of sediment transport across the shelf and beyond the shelf edge. The results may further refine the depositional model on the Troll Field as well as providing a better understanding of the connectivity between the reservoir sandstones. It may also indicate other areas within the basin where the Krossfjord and Fensfjord formations may occur.

Lastly, Chapter 4 presented the first application of forward seismic modelling to understanding the seismic expression of clinoforms in different deltaic depositional environments. Whilst the results presented are applicable to other shallow-marine reservoir sandstones, for which the outcrops used are considered to be sedimentological analogues, further deltaic outcrop analogues should be analysed to widen the applicability and rigorously test the methodology. For example, tide-dominated deltaic units are difficult to recognise in subsurface data (e.g. Berg, 1982) and forward seismic modelling may therefore improve the ability to identify these units in seismic data. If successful, the robust interpretation of subsurface stratigraphic architectures enabled by this method highlight the potential for it be used as an additional tool in frontier exploration, reservoir characterisation and production strategies.
REFERENCES


References


Hampson, G.J. 2010. Sediment dispersal and quantitative stratigraphic architecture across an ancient shelf. *Sedimentology, 57 (1)*, 96-141.


References


**APPENDICES**

**SECTION A1**

**Sedimentary Core Logs from Troll Field**

### Lithology
- Sandstone
- Siltstone
- Calcite Cement
- Unconsolidated Core
- No Core
- Graded Bed
- Wood Fragment
- Carbonaceous Material
- Mica
- Calcite Nodule
- Pyrite
- Siderite
- Glaucnite

### Bio
- Articulated Bivalve
- Oyster
- Shell Layer
- Shell
- Shell Fragment
- Belemnite
- Chondrites
- Terrebelina
- Planolites
- Burrow

### Structures
- Planer Lamination
- Tabular Cross Bedding
- Low Angle Cross Bedding
- Trough Cross Bedding
- Wavy Laminae
- Symmetric Ripple
- Asymmetric Ripple
- Lenticular Bedding
- Sharp Contact

### Contacts
- Undulatory Contact
- Gradational Contact
WELL 31/2-1

KEY
- Troll Field
- Cored Well
- Normal Fault

Lomre Terrace
Svartalf Fault
Troll Fault

31/2-4R
31/2-3
31/2-1
31/5-5
31/6-1
31/6-6
31/3-1
31/6-5

WELL 31/2-1

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**m**

**GR**

**gAPI**

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**40**

**70**

**1 6**
# APPENDICES

## SECTION A2

### Photo Panoramas and Sedimentary Logs from USA Fieldwork

**Sedimentary Log Key**

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**Cements & Nodules**

- Nodule
- Ironstone
- Glauconite
- Mica
- Pyrite
- Carbonaceous
Chimney Rock Tongue at the CR2 locality

Key
- Wave-dominated middle shoreface
- Wave-dominated lower shoreface
- Offshore transition
- Cement
- Outcrop outline
- Major transgressive surface
- Transgressive surface
- Subaerial unconformity
- Measured Section
- Pseudo Section

Photo Panorama

Photo Panorama Interpretation

Pseudo-log Correlation (flattened on subaerial unconformity)
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**GEOLOGIST: N.E.Holgate**

**DATE: 26/09/2011**
**UNIT: Chimney Rock Tongue**

**LOCATION: 10.1**

**GEOLOGIST: N.E. Holgate**

**DATE: 26/09/2011**

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**Subaerial Unconformity**

Top CS6

Highlighted by significant erosion

Slightly cemented (<30%)

Massive, blocky units

Sandstone commonly structureless

Discreet coarsening-upwards profile

**Top of outcrop**

**Middle Shoreface**

**Thick amalgamated HCS**
UNIT: Chimney Rock Tongue

LOCATION: 11.4

GEOLOGIST: N.E. Holgate

DATE: 27/09/2011

**MEASURED SECTION**

**SEDIMENTOLOGY DATA SHEET**

**COMMENTS**

- HCS units becoming more frequent
- Subaerial Unconformity Top CS6
- Well cemented bed (80%)
- Clinoform
- Clinoform
- Sandstone units increasing in thickness and frequency
- Major Transgressive Surface Top CS4

No exposure due to vegetation

Clinoform

Top CS6

Middle Shoreface

Middle Shoreface

Middle Shoreface

Lower Shoreface

Offshore Transition

Major Transgressive Surface Top CS4

Subaerial Unconformity
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Chimney Rock Tongue at the CR3 locality

Photo Panorama

Photo Panorama Interpretation

Pseudo-log Correlation (flattened on MFS)

Key
- Orange: Outer-estuary tidal bars
- Blue: Estuary-mouth barrier
- Purple: Fluvial Channel
- Yellow: Wave-dominated upper shoreface
- Green: Wave-dominated lower shoreface
- Yellow-green: Wave-dominated middle shoreface
- Black: Outcrop outline
- Yellow-orange: Measured Section
- Red: Subaerial unconformity
- Pink: Pseudo Section
- Purple: Subaerial unconformity
- Grey: Offshore transition
- Light blue: Major transgressive surface
- Blue: Transgressive surface
- Gray dotted: Cement
### Sedimentology Data Sheet

#### Chimney Rock Tongue

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- **No exposure due to vegetation**
- **Subaerial Unconformity Top CS6**
- **Major Transgressive Surface Top CS4**
- **O.T. (Offshore Transition)**
- **Clinoform**
- **Intense bioturbation throughout**
- **Thin, isolated beds of HCS and/or wave-ripples**
## Measured Section

### Sedimentology Data Sheet

**Unit:** Chimney Rock Tongue  
**Location:** 11.2  
**Date:** 27/09/2011

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1. Common very fine, siltstone lenses c. 5-10 cm thick
2. Decreasing occurrence of siltstone beds
3. Subtle increase in grain size
4. HCS beds are becoming increasingly amalgamated
5. No exposure due to vegetation
6. Major Transgressive Surface Top CS4
7. Clinoform
8. Clinoform
9. Clinoform
10. Clinoform
11. Clinoform
12. Clinoform
13. Clinoform
14. Clinoform

**GEOLOGIST:** N.E. Holgate
**DATE:** 27/09/2011
MEASURED SECTION
SEDIMENTOLOGY DATA SHEET

UNIT: Chimney Rock Tongue
LOCATION: 11.2
GEOLOGIST: N.E. Holgate
DATE: 27/09/2011

LENGTH (m) | UNIT | LITHOLOGY | GRAIN SIZE AND SEDIMENTARY STRUCTURES | BIOTURBATION | PHOTOGRAPHS | FACIES ASSN. | DEPOSITIONAL ENVIRONMENT | COMMENTS
---|---|---|---|---|---|---|---|---
24 | Chimney Rock Tongue | Top of outcrop | | | | | | Dark brown sandstones, uni-directional trough cross beds
23 | | | | | | | | Erosively-based fining-upwards sequences
22 | | | | | | | | Subaerial Unconformity Top CS6 Minor cementation at top (<30%)
21 | | | | | | | | Amalgamated HCS beds
20 | | | | | | | | Clinoform
19 | | | | | | | | Minor cementation at top (<30%)
18 | | | | | | | | HCS beds are becoming increasingly amalgamated
17 | | | | | | | | Major Transgressive Surface Top CS5
16 | | | | | | | |
PANTHER TONGUE AT THE PT1 LOCALITY

(a) Map showing the locations of states such as Wyoming, Utah, Idaho, Nevada, and Colorado, with Salt Lake City marked.

(b) Close-up map of the Helper area, showing Quitchupah Creek, Spring Canyon, and other landmarks.

(c) Detailed map highlighting the locations of Minnes Gap, The Glades, and other geographical features.

(d) Diagram indicating the palaeoflow direction with measured sections and studied localities marked.

KEY:
- Palaeoflow Direction
- Measured Section
- Studied Locality

0 km 100 km
## Panther Tongue

### Top of Outcrop

- **FA 4**: Proximal Delta Front
- **FA 3**: Mouth Bar

### Depth (m)

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**Thickness of sandstone beds increasing**

**Clinoform**

**Minor cementation (<20%) at base of bed**

**Clinoform**

**Clinoform**

**Clinoform**

**Clinoform**

**Clinoform**

**Clinoform**

**Clinoform**

### Location: 15.6

**GEOLOGIST:** N.E. Holgate  
**DATE:** 01/10/2011

---

Appendices
Panther Tongue at the PT3 locality

Photo Panorama

Photo Panorama Interpretation

Pseudo-log Correlation (flattened on transgressive surface)

KEY
- Mouth Bar
- Proximal Delta Front
- Distal Delta Front
- -- Outcrop outline
- Transgressive surface

Vertical Exaggeration x2

14.2

14.3

20 m

100 m
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Panther Tongue

FA 4
Proximal Delta Front
Very fine-grained units heavily bioturbated
Clinoform

FA 5
Distal Delta Front
Very discreet lenses of sandstone
Clinoform

Geochemist: N.E. Holgate
Date: 30/09/2011
### Sheet 2 of 2

**Measured Section**

**Sedimentology Data Sheet**

**Unit:** Panther Tongue

**Location:** 14.2

**Geologist:** N.E. Holgate

**Date:** 30/09/2011

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**Panther Tongue**

- **Mouth Bar**
- **FA 3**

**Appendices**

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Sheet 1 of 2

MEASURED SECTION
SEDIMENTOLOGY DATA SHEET
LOCATION: 14.3

UNIT: Panther Tongue
MEASURED SECTION
LOCATION: 14.3

GEOLGIST: N.E.Holgate
DATE: 30/09/2011

Panther Tongue
FA 4
Proximal Delta Front
FA 5
Distal Delta Front
FA 3
Mouth Bar

LENGTH (m)
UNIT
LITHOLOGY
GRAIN SIZE AND SEDIMENTARY STRUCTURES
PHOTOGRAPHS
FACIES ASSN.
DEPOSITIONAL ENVIRONMENT
COMMENTS

Strong alternation between
grainsize over cm-scale
Clinoform
Clinoform
Clinoform
Clinoform
Clinoform
Clinoform
Clinoform

Sandstone beds contain
rippled tops
Clinoform

Very rare sandstone beds
cm-thick
Clinoform
Clinoform

Appendices
Page 300
KF-2 TONGUE OF THE FERRON SANDSTONE MEMBER AT THE IV1 LOCALITY

Appendices
Kf-2 tongue of the Ferron Sandstone Member at the IV1 locality

Photo Panorama Interpretation

Pseudo-log Correlation (flattened on MFS top Kf-1-lv[c])

KEY
- Wave-modified upper shoreface
- Wave-modified middle shoreface
- Wave-modified lower shoreface
- Fluvial-dominated proximal delta front
- Fluvial-dominated distal delta front
- Offshore transition
- Cement
- Transgressive surface
- Major transgressive surface
- Channel Base
- Outcrop outline

Channel
Bay-fill
Foreshore

20 m
Proximal Delta Front

- No exposure due to vegetation cover
- Common carbonaceous material throughout
- Load deposits common

Lower Shoreface

- Occasional planar laminations although pervasive bioturbation

Clinoform

Load deposits common

Common carbonaceous material throughout

Appendices
UNIT: Ferron Sandstone

LOCATION: 19.1

GEOLOGIST: L.V. Forman
DATE: 05/10/2011

Sedimentary Structures

- Abundent planar laminations
- 10-20 cm thick cemented hard bands, not laterally extensive
- Beds are mostly structureless with minor bioturbation
- Moderately- to well-sorted
- Common HCS in minor sandstone beds

Comments
**UNIT:** Ferron Sandstone  
**LOCATION:** 19.1  
**GEOLOGIST:** L.V. Forman  
**DATE:** 05/10/2011

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Top of outcrop  
Kf-2-Iv[c]  
Clinoform  
Top of clinoform unit weakly cemented (<20%)  
Abundant amalgamated trough cross-bedding  
Planar laminations in beds  
Kf-2-[v]  
Clinoform  
Top of clinoform unit weakly cemented (<20%)  
Moderately- to well-sorted  
Clinoform  
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Clinoform  
Sandstone contains roots and numerous burrows
KF-2 TONGUE OF THE FERRON SANDSTONE MEMBER AT THE IV2 LOCALITY
Kf-2 tongue of the Ferron Sandstone Member at the IV2 locality

Photo Panorama

Photo Panorama Interpretation

Pseudo-log Correlation (flattened on MFS top Kf-2-Iv[b])

KEY

- Bay-fill
- Foreshore
- Wave-mod upper shoreface
- Wave-mod middle shoreface
- Wave-mod lower shoreface
- Fluvial-dom proximal delta front
- Fluvial-dom distal delta front
- Offshore transition
- Cement
- Outcrop outline
- Transgressive surface
- Major transgressive surface
Ferron Sandstone

- Common oyster coquina layers in lower beds
- Grain size generally coarsening upwards
- No exposure due to vegetation cover
- Coal layer topping Kf-1-Iv[c] is assumed to be absent due to vegetation cover
- Rare cemented nodules <10 cm diameter
- Grain size generally coarsening upwards
- Common oyster coquina layers in lower beds

HCS amalgamated in beds
Weakly cemented throughout
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Thickly bedded, massive unit with minor planar laminations

Clinoform

Sparse laminations throughout

Unit is generally massive with very minor grain size alterations throughout
UNIT: Ferron Sandstone
LOCATION: 19.2

GEOLOGIST: N.E. Holgate
DATE: 05/10/2011

LENTH (m) | UNIT | SAND | CLAY | SILT | PEBBLE | V \( f \) | V \( t \) | V \( m \) | V \( c \) | V \( g \) | COMMENTS
---|---|---|---|---|---|---|---|---|---|---|
37 | Top of outcrop | | | | | | | | | |
36 | | | | | | | | | | Common roots and burrows throughout
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DEPOSITIONAL ENVIRONMENT

CLINOFORM

Trough cross-bedding appears in isolated beds

TOP OF OUTCROP

Common roots and burrows throughout

WEAKLY CEMENTED IN PATCHES (<10%)
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**Sheet 1 of 3**

**Measured Section**

**Sedimentology Data Sheet**

**Location:** 19.3

**Unit:** Ferron Sandstone

**Geologist:** N.E. Holgate

**Date:** 05/10/2011

**No exposure due to vegetation cover**

**Coal layer topping Kf-1-Iv[c] is assumed to be absent due to vegetation cover**

**HCS amalgamated in beds**

**Clinoform**
Ferron Sandstone

**Middle Shoreface**
- Sparse laminations throughout
- Unit is generally massive with very minor grain size alterations throughout

**Upper Shoreface**
- Weakly cemented in patches, not laterally extensive (<10%)

**Lower Shoreface**
- Weakly cemented in patches not laterally extensive (<20%)
KF-1 TONGUE OF THE FERRON SANDSTONE MEMBER AT THE JP1 LOCALITY

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Appendices
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**UNIT:** Ferron Sandstone

**LOCATION:** 19.7

**GEOLOGIST:** N.E. Holgate

**DATE:** 05/10/2011

**LITHOLOGY:**
- Ferron Sandstone
- Tununk Shale

**GRAIN SIZE AND SEDIMENTARY STRUCTURES:**
- Planar laminations
- Restricted to beds d-cm thick
- Very rare laminations, almost completely bioturbated

**BIOTURBATION:**
- Clinoform

**DEPOSITIONAL ENVIRONMENT:**
- Offshore
- Proximal Delta Front
- Distal Delta Front
- Clinoform

**COMMENTS:**
- Planar laminations restricted to beds d-cm thick
- Very rare laminations, almost completely bioturbated
SAND
VF FM VC GR
PEBBL. SILT CLAY

GRAIN SIZE AND
SEDIMENTARY STRUCTURES

LENGTH (m)

PHOTOGRAPHS

LITHOLOGY

BIOTURBATION

UNIT

PARASEQUENCE

DEPOSITIONAL
ENVIRONMENT

COMMENTS

UNIT: Ferron Sandstone

LOCATION: 19.7

GEOLOGIST: N.E. Holgate

DATE: 05/10/2011

Top of outcrop

Full sequence absent at top of outcrop

Clinoform

HCS appears in isolated beds

Clinoform

Clinoform

Very rare planar and ripple laminations

Bay-fill deposits on top of channel

Appendices
MEASURED SECTION
SEDIMENTOLOGY DATA SHEET

UNIT: Ferron Sandstone
LOCATION: 20.13
GEOLOGIST: N.E. Holgate
DATE: 06/10/2011

LENGTH (m) | UNIT | LITHOLOGY | GRAIN SIZE AND SEDIMENTARY STRUCTURES | BIOTURBATION | PARASEQUENCE | DEPOSITIONAL ENVIRONMENT | COMMENTS
---|---|---|---|---|---|---|---
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- Clinoform
- Proximal Delta Front
- Distal Delta Front

Appendices
### Unit: Ferron Sandstone

**Measured Section**

**Sedimentology Data Sheet**

**Location:** 20.13

**Geologist:** N.E. Holgate

**Date:** 06/10/2011

#### Lithology

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#### Observations

- **Top of Outcrop**
- **Proximal Delta Front**

**Appendices**

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KF-1 TONGUE OF THE FERRON SANDSTONE MEMBER AT THE JP2 LOCALITY

(a) Map showing the locations of Wyoming, Utah, Idaho, and Nevada.

(b) Detailed map of the study area around the JP2 locality.

(c) Close-up map of the Helper area, showing Spring Canyon.

(d) Map of The Glades area, showing the location of the CR1 and CR2.

KEY
- Palaeoflow Direction
- Measured Section
- Studied Locality

0 km 1 km

Appendices
Kf-1 tongue of the Ferron Sandstone Member at the JP2 locality

Photo Panorama

Photo Panorama Interpretation

Pseudo-log Correlation (flattened on MFS top Kf-1-Iv[a])

Vertical exaggeration 2x

Major transgressive surface

Cement

Outcrop outline

Transgressive surface

Proximal Delta Front

Distal Delta Front

Mouth Bar

Correlation Key

20.12
20.11
20.10
20.9
20.6
20.5
20.4
20.3
20.2
20.1
20.0

20 m

20 m
## Sedimentology Data Sheet

**Location:** 20.06

**Unit:** Ferron Sandstone

**Date:** 06/10/2011

**Geologist:** N.E. Holgate

### Sedimentological Data

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**Depositional Environment:**
- Offshore
- Distal Delta Front
- Proximal Delta Front
- Clinoform
## MEASURED SECTION
### SEDIMENTOLOGY DATA SHEET

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**LOCATION:** 20.06  
**DATE:** 06/10/2011  
**SCALE:** 1:50  
**GEOLOGIST:** N.E. Holgate

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**Location:** 20.08  
**Geologist:** N.E. Holgate  
**Date:** 06/10/2011

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**Measured Section:**

- Top of Kf-1-Iv[a]

**Depositional Environment: Clinoform:**

- Mouth Bar
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Appendices
### MEASURED SECTION

**Sedimentology Data Sheet**

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**Location:** 20.09

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**GEOLOGIST:** N.E. Holgate  
**Date:** 06/10/2011

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GEOLOGIST: N.E. Holgate
DATE: 06/10/2011
UNIT: Ferron Sandstone
LOCATION: 20.12
GEOLOGIST: N.E. Holgate
DATE: 06/10/2011

MEASURED SECTION
SEDIMENTOLOGY DATA SHEET

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