THE RISE AND FALL OF DIAPIRS DURING REGIONAL EXTENSION AND ITS INFLUENCE ON THE DEPOSITION OF A NET-TRANSGRESSIVE COASTAL PLAIN-TO-SHALLOW MARINE SUCCESSION: MIDDLE-TO-UPPER JURASSIC, NORWEGIAN CENTRAL GRABEN

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This thesis is submitted for the degree of Doctor of Philosophy

December 2013
ABSTRACT

Regional extension in salt basins can initiate and drive salt tectonics, which may influence sediment routing systems and reservoir distribution. Structural controls on the deposition of Middle-to-Upper Jurassic, net-transgressive, shallow-marine strata preserved on the eastern flank of the North Sea Central Graben have been studied using an integrated subsurface dataset (3D and 2D seismic reflection, wireline-log, core and biostratigraphic data) in order to: (i) characterise the main structural styles in a salt-influenced rift basin; (ii) describe the sedimentology and stratigraphic architecture of the Middle-to-Upper Jurassic succession; and (iii) investigate the role that rift-related salt tectonics had on the thickness and facies distribution of the Middle-to-Upper Jurassic, syn-rift succession.

Early Triassic rifting initiated reactive diapir rise of Late Permian salt, which influenced the geometry of Triassic rafts between salt walls. Middle-to-Late Triassic differential loading and density-driven subsidence resulted in further passive diapir growth, until a depletion of salt supply led to the formation of salt welds below Triassic rafts and burial of salt diapirs. A second phase of extension in the Middle-to-Late Jurassic resulted in either diapir collapse, providing accommodation for diachronous deposition of shallow-marine reservoirs, or reactive diapir rise, which influenced the depositional thickness of these reservoirs along salt wall flanks.

Sedimentological core analysis of the Middle-to-Upper Jurassic in combination with biostratigraphic and wireline-log data has identified offshore, offshore transition, lower shoreface, upper shoreface and coastal-plain deposits. These deposits are arranged into upward-shallowing parasequences bounded by flooding surfaces. The timing of transgression is diachronous, with flooding and shoreline retreat controlled by the underlying rift topography. The resulting facies architecture reflects the balance between
fault- and halokinesis-driven accommodation creation, and intra- and extra-basinal sediment supply. This thesis highlights the key role that salt has in modifying the tectono-stratigraphic evolution of rift basins.
DECLARATION OF ORIGINALITY

The author of this thesis is responsible for all interpretations and conclusions submitted therein. No portion of this thesis has been submitted in support of any another degree or qualification. References and all other contributions to this research are fully acknowledged where applicable.

Aruna Mannie
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ACKNOWLEDGEMENTS

This research would not have been possible without your support, encouragement, guidance and advice. I express my deepest gratitude and appreciation to the following persons and institutions.

Alastair Fraser
Adrian Hartley
Christopher Jackson
Christopher Soufleris
Christopher Ward
Gary Hampson
Howard Johnson

Centrica Energi
Commonwealth Commission
London Petrophysical Society
Norwegian Petroleum Directorate
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CHAPTER 1
INTRODUCTION

1.1. Rationale

The structural evolution of rift basins developing in brittle crust is well-documented and highlights that normal faulting is a key control on syn-rift sediment transfer mechanisms and the temporal and spatial evolution of depositional environments (e.g. Leeder and Gawthorpe, 1987; Gabrielsen et al., 1990; Prosser, 1993; Nøttvedt et al., 1995; Ravnås and Steel, 1998; Cowie et al., 2000; Gawthorpe and Leeder, 2000; McClay et al., 2004). These studies show that, during rift initiation, early normal faults have low displacement, tend not to breach the surface and are not connected to each other. This creates low-relief topography providing localised accommodation in isolated depocentres. Depending on sea-level, these depocentres form lacustrine, alluvial or shallow-marine elongated embayments (Prosser, 1993; Ravnås and Steel, 1998; Gawthorpe and Leeder, 2000). During the subsequent rift climax, larger, linked fault systems form and increasing rates of fault-driven subsidence leads to the establishment of fully marine conditions (Leeder and Gawthorpe, 1987; Cowie et al., 2000; Gawthorpe and Leeder, 2000; Gawthorpe et al., 2003). Many rift basins are documented to contain ductile evaporites (salt) in the pre-rift (e.g. North Sea), syn-rift (e.g. Red Sea) and post-rift (South Atlantic basins, e.g. Kwanza, Campos and Gabon) stratigraphic succession; these ductile units can act as intra-stratal detachments, decoupling the sub- from the supra-salt cover structures (e.g. Vendeville and Jackson 1992; Nalpas and Brun 1993; Jackson et al., 1994; Stewart and Clark 1999). However, relatively few studies have adequately addressed how the presence of salt impacts the evolving syn-rift structural style, accommodation development, facies
distribution and stratigraphic architecture (e.g. Jackson et al., 1994; Stewart 2007; Jackson and Larsen 2009).

This study focusses on the eastern flank of the North Sea Central Graben (Cod Terrace, Sørvestlandet High, Norwegian-Danish and Egersund basins), where both rifting and salt tectonics influenced the deposition of the Middle-to-Upper Jurassic, net transgressive, coastal-plain to shallow-marine syn-rift succession. The North Sea is one of very few rift basins that contain relatively abundant seismic and borehole data that allows detailed study of the influence that salt can have on the development of supra- and sub-salt structures, and the influence these can have on the deposition of a syn-rift coastal-plain to shallow-marine succession. Although the North Sea is relatively mature in terms of hydrocarbon potential from the associated reservoirs, it can act as a useful analogue to characterise the structural styles, reservoir types and trapping configurations that may be encountered in other, unexplored or underexplored salt-influenced rift basins (e.g. Eastern Mediterranean).

An integrated dataset comprising 2D and 3D seismic reflection and borehole data, the latter containing wireline-logs, core, biostratigraphy, composite logs and mineralogical data, were used for this study. The biostratigraphic data was calibrated to the Partington et al. (1993) North Sea stratigraphic scheme, which allowed comparison and correlation to other parts of the North Sea.

1.2. Aim

The overall aim of this thesis is to determine the influence of rift-related normal faulting and salt tectonics on the sedimentology and sequence stratigraphy of the syn-tectonic, Middle-to-Upper Jurassic succession, and to provide further insights into the Permian-
Jurassic tectono-stratigraphy on the eastern flank of the North Sea Central Graben. The specific aims of this thesis are:

(i) to characterise the sub-, supra- and salt structural styles on the eastern flank of the North Sea Central Graben (Cod Terrace, Sørvestlandet High, Norwegian-Danish and Egersund basins)

(ii) to document the sedimentology and stratigraphic architecture of the Middle-to-Upper Jurassic, net transgressive, coastal-plain to shallow-marine succession (Bryne, Sandnes, Ula and Gyda formations)

(iii) to investigate the influence of coeval rifting and salt tectonics on accommodation development and reservoir distribution in a salt-influenced rift basin

(iv) to describe the tectono-stratigraphic evolution of a salt-influenced rift basin, with particular focus on how diachronicty in the timing of tectonic subsidence controls temporal and spatial variations in stacking patterns in a syn-rift, coastal-plain to shallow-marine succession

1.3. Dataset and Methodology

An integrated dataset comprising of 2D and 3D seismic reflection data covering a combined area of >15,000 km², and well data from 53 wells were utilised. All wells contain at least two conventional wireline logs (e.g. gamma ray, resistivity, density and sonic), and checkshot data. Twenty-six wells contain a total of 1180 m of core in the Middle to-Upper-Jurassic succession. Biostratigraphic data were available for twenty-one wells and these allowed regional correlations between sub-basins, and to tie the
stratigraphic framework to the basinwide sequence stratigraphic framework of Partington et al. (1993). See Appendix 1 for further details of the well data used in this study.

In order to achieve the aims of this thesis, the following methodology was adopted:

(i) Structural styles in a salt-influenced rift basin

- 2D and 3D seismic mapping of surfaces and faults to generate structure and isochron maps to constrain the structural framework and the tectono-stratigraphic evolution of the Cod Terrace, Sørvestlandet High, Norwegian-Danish and Egersund basins.

- use of seismic attribute analysis, in particular the cosine of instantaneous phase, dip and RMS amplitude, to constrain the stratigraphic architecture of the Middle-to-Upper Jurassic succession. The cosine of instantaneous phase seismic attribute volume, which is amplitude independent and enhances the continuity of seismic events, was derived from the 3D seismic reflectivity survey and used for detailed mapping within the Upper Jurassic reservoir interval.

- analysis of wireline-log data to investigate the initiating and driving mechanisms for salt movement in a rift-basin.

(ii) Sedimentology and sequence stratigraphy

- sedimentological core logging and trace fossil analysis in the Middle-to-Upper Jurassic Bryne, Sandnes, Ula and Gyda formations to understand the lithological and facies distribution within formations.
- integration of core analysis with wireline-log and biostratigraphic data to construct a sequence stratigraphic framework for the Middle-to-Upper Jurassic succession so that the various stratigraphic packages can be interpreted within a chronostratigraphic framework.

(iii) Impact of rift-related salt tectonics on shallow-marine sedimentation

- integration of the structural framework with the sedimentological and sequence stratigraphic framework to understand the influence of rift-related salt structures on the thickness and distribution of the Middle-to-Upper Jurassic succession.

- palaeogeographic reconstruction of the Middle-to-Upper Jurassic utilising the structural and stratigraphic data generated during this study.

A detailed discussion of the methodologies adopted throughout this research is given in each of the respective chapters (chapters 3-6).

1.4. Thesis outline

This thesis comprises of seven chapters. Chapters 1 and 2 form the introduction and background to this study whilst Chapters 3-6 deals with a specific aspect of this research project. Chapters 3-6 have been written in the form of journal papers, which have been or are in the process of being submitted to journals. Chapter 7 provides a summary of the conclusions based on the results of this research.

Chapter 2 comprises of a literature review pertinent to this study focussing on: (i) evolution of rift basins; (ii) the presence of pre-rift salt in extensional basins; (iii) impact
of faulting, folding and salt tectonics on coastal-plain/shallow-marine sedimentation in salt-influenced rifts.

**Chapter 3** describes the sub-, supra- and salt structures of the Egersund Basin. It documents depositional thickness and facies distribution of the syn-tectonic Middle-to-Upper Jurassic coastal-plain to shallow-marine succession of the Bryne and Sandnes formations.

**Chapter 4** documents the sub-, supra- and salt structures on the Cod Terrace, Sørvestlandet High and Norwegian-Danish Basin. It describes the influence of rifting on diapir collapse and the formation of minibasins that are the site of deposition of the Upper Jurassic, shallow-marine reservoirs in the Ula and Gyda fields.

**Chapter 5** explores and contributes to the debate on the origin of salt-wall confined Triassic succession in the Central North Sea.

**Chapter 6** provides syntheses results from chapters 3, 4 and 5 and integrates these with previously published regional work to reconstruct the palaeogeography of the Norwegian sector of the North Sea during the Middle-to-Upper Jurassic. Furthermore, it provides a tectono-stratigraphic model to be considered in basins where pre-rift evaporites are present.

**Chapter 7** is a summary of the conclusions that relates to the aims of this study outlined in section 1.2.
1.5. Publication status and contribution from co-authors

Chapters 3, 4, 5 and 6 are presented in this thesis as four papers. The author of this thesis is the first author on all four papers and is responsible for the interpretations and material submitted herein. Important contributions to the research are fully acknowledged at the end of each paper. An overview of the main contributions from the various co-authors is outlined below.

**Paper I: Structural controls on the stratigraphic architecture of net-transgressive shallow-marine strata in a salt-influenced rift basin: Middle-to-Upper Jurassic Egersund Basin, Norwegian North Sea.**

Aruna S. Mannie, Christopher A. – L. Jackson and Gary J. Hampson

*Published online 20th February 2014 (Basin Research)*

The first author was responsible for analysing and interpreting the seismic and borehole data and the sedimentological core descriptions. The co-authors critically reviewed and discussed interpretations, and contributed to the outline and revision of the manuscript prior to submission.

**Paper II: Shallow-marine reservoir development in extensional diapir collapse minibasins: an integrated subsurface case study from the Upper Jurassic of the Cod Terrace, Norwegian North Sea.**

Aruna S. Mannie, Christopher A. – L. Jackson and Gary J. Hampson

*Accepted (in press) American Association of Petroleum Geologists Bulletin*
The first author was responsible for interpreting the seismic and borehole data, sedimentological core descriptions and proposing the tectono-stratigraphic model. The co-authors critically reviewed and discussed interpretations, and reviewed the manuscript prior to submission.

Paper III: Salt wall-confined Triassic depocentres in the Central North Sea: depositional minibasins or raft-related remnants?

Aruna S. Mannie, Christopher A. – L. Jackson, Michael R. Hudec and Gary J. Hampson

_In the process of submitting to the Journal of the Geological Society of London_

The first author was responsible for providing the seismic interpretations and well data analysis. The co-authors discussed interpretations and reviewed the manuscript prior to submission.

Paper IV:

Tectonic controls on the spatial distribution and stratigraphic architecture of a net-transgressive succession in a salt-influenced rift basin: Middle-to-Upper Jurassic, Norwegian Central North Sea

Aruna S. Mannie, Christopher A. – L. Jackson and Gary J. Hampson

_In the process of submitting to the Journal of the Geological Society of London_

The first author was responsible for data integration and interpretation. This paper benefitted from previous interpretations in Papers I, II and III and provided a regional
understanding of the study area within the North Sea stratigraphic framework. The co-authors have discussed interpretations and reviewed the manuscript prior to submission.

Note that all four papers have been submitted or are about to be submitted to different journals and as a consequence, language style, abbreviations and references differ for each paper but are kept consistent within individual papers.
CHAPTER 2
LITERATURE REVIEW

This chapter reviews published literature in order to highlight the key aspects of the structural and stratigraphic evolution of rift basins and the influence of pre-rift salt on such. Several key examples are presented from salt-influenced rift basins analogous to the Middle-to-Upper Jurassic, coastal-plain to shallow-marine, net-transgressive succession in the Norwegian North Sea. This highlighted the gaps in understanding of the influence of pre-rift salt on shallow-marine sedimentation in which key questions were identified and formed the aims of this thesis.

2.1 Evolution of rift basins

Rift basins are elongate depressions formed due to lithospheric extension (e.g. McKenzie, 1978; Withjack et al., 2002; Allen and Allen, 2005). They can be thousands of kilometres long, tens of kilometres wide and several kilometres deep, and are composed of en echelon, parallel and intersecting sub-basins (e.g. East Africa, North Sea and the South Atlantic margin). Observations from present day and ancient rifts show their formation in areas of lithospheric extension characterised by: (i) low crustal thickness; (ii) elevated geothermal gradients that may be associated with volcanism; (iii) brittle failure of the crust dominated by normal faulting; (iv) seismicity; (v) elevated rift shoulder topography; and (vi) negative Bouguer gravity anomalies (Allen and Allen, 2005). Lithospheric extension can be described using two end-member models, active or passive rifting (McKenzie, 1978; Rosendhal 1987; Allen and Allen 2005). Active rifting is driven by convection and thinning of the lithosphere by emplacement of a thermal mantle plume (e.g. Jurassic North
Sea rift and East African rift) (e.g. Rosendhal 1987; Underhill and Partington, 1993; Ziegler and Cloetingh, 2004). A period of regional uplift is recognised as a precursor to extension (e.g. Mid North Sea Dome in the Central North Sea) (Underhill and Partington 1993) and melting of the asthenosphere may lead to rift-related volcanism (e.g. East Africa) (Rosendhal 1987). On the other end of the spectrum, passive rifting occurs as a consequence of far-field effective tensional stress in the lithosphere, resulting in the upwelling of the asthenosphere. Volcanism may be associated as a secondary process (e.g. West Africa rifts) (e.g. Rosendhal 1987; White and McKenzie, 1991). It is common for rift systems to form by a combination of processes and exhibit a range of characteristics of both active and passive models (Rosendhal 1987; Ziegler and Cloetingh 2004).

During the formation of rift basins the extension of the upper crust is accommodated by normal faults, whilst the lower crust and mantle lithosphere is stretched by plastic deformation (Allen and Allen 1990; Ravnås and Steel 1998). The evolution of rift basins is commonly divided into three main stages; pre-rift, syn-rift and post-rift. The pre-rift stage is defined as a period of tectonic quiescence, as reflected by non-deposition or deposition of uniform thickness of strata across the basin. It is followed by the syn-rift phase, which is characterised by extension, normal faulting and localised accommodation creation. During the initiation of the syn-rift phase, nucleation of normal fault arrays occurs, forming localised depocentres adjacent to the points of maximum displacement on basin-bounding faults (Prosser 1993; Gupta et al. 1998; Davis et al., 2000; Fig. 2.1a). During this initial phase, early normal faults are unconnected and have low displacements and tend not to breach the surface (Gawthorpe and Leeder 2000). The stratigraphy during syn-rift deposition is controlled by the rate of fault-controlled subsidence, tectonically-controlled topography, the volume of sediment supply and eustasy within a basin (e.g. Leeder and
Gawthorpe 1987; Prosser 1993; Young et al., 2003). Where the volume of sediment supplied into the basin keeps pace with increasing accommodation, the surface expression of the tectonic activity may be subdued and therefore have little impact on facies distribution and architecture (Frostick and Steel, 1993). However, where sediment supply is low and unable to keep pace with increasing rates of accommodation, significant relief will develop and this may affect facies distributions. Despite the relief created during rift basin development, drainage patterns can be either axial (i.e. parallel to the basin-bounding faults and basin axis) or transverse (i.e. normal to the basin-bounding faults and basin axis). Axial systems commonly consist of relatively large, fluvo-deltaic systems that build out and retreat along the basin axis, whereas transverse systems are sourced from the crests of normal fault blocks and are deposited on the hangingwall dipslope or immediate hangingwall of the basin-bounding faults. In a submarine environment turbidites, slumps and slides may be deposited along the basin axis (Leeder and Gawthorpe, 1987; Ravnås and Steel 1998; Gawthorpe and Leeder, 2000). Sediment entry points within rift basins may occur along transfer zones between fault segments, at cross-cutting faults or basinward of easily erodible substrates (Ravnås and Steel, 1998; Gawthorpe and Leeder 2000). Tidal amplification commonly occurs where shallow-marine systems form within narrow, shallow, elongate depocentres during the early syn-rift period (Mellere and Steel, 1996; Ravnås and Steel, 1998). Predominantly wave processes tend to occur during periods of tectonic quiescence when footwall islands become completely submerged and larger, more open-marine embayments occupy the rift basin (Ravnås and Bondevik, 1997).

The rift climax phase is typically characterised by through-going normal fault zones, linking fault segments into large fault systems (Gawthorpe and Leeder, 2000; Fig. 2.1b). This result in the establishment of significant relief, formation of a small number of large
rapidly subsiding half-graben basins, and the basinwide creation of accommodation (Ravnås and Steel, 1998; Gawthorpe and Leeder, 2000). The rate of sediment supply becomes critical at this stage because it dictates whether accommodation is overfilled or bypassed, starving more distal parts of the rift (Prosser, 1993). In sediment starved, distal settings that are isolated from externally supplied sediment, such as within intra-fault block basins, deposits are almost exclusively locally derived (Gawthorpe and Leeder, 2000; Fig. 2.1b). The upper limit of the syn-rift interval is commonly marked by an unconformity, which separates it from the overlying post-rift infill. There is little to no fault activity during the post-rift period and thermal subsidence dominates due to cooling of lithosphere after active stretching has ceased (Fig. 2.2). Post-rift sediments passively infill and onlap the remnant rift topography (e.g. Badley et al., 1984; Prosser, 1993; Gabrielsen et al. 2001; Fig. 2.2). In many instances, the post-rift inherits a pronounced topography of fault-bounded structural highs and deep basins from the preceding syn-rift interval. The type of depositional systems that develop and the overall pattern of sedimentation during the post-rift are strongly controlled by the configuration of the relict rift topography, in addition to changes in relative sea-level, which in turn are commonly linked back to effects of post-rift thermal subsidence (Prosser, 1993; Gabrielsen et al., 2001; Zachariah et al., 2009).

In a continental rift setting (e.g. East Africa) lakes are common in structurally isolated half-grabens. In arid climates playa lakes develop, whereas in less evaporative climates (e.g. Lake Malawi) they are fringed by fan and axial deltas (Soreghan et al., 1999). Aeolian sands are not uncommon and can develop by wind reworking of alluvial and lake shoreline sands (e.g. San Luis Basin, Colorado; Gawthorpe and Leeder 2000). During sea level highstands coarse clastic sedimentation by turbidity currents, gravity flows and
slumps occur (Ravnås and Steel, 1998; Gawthorpe and Leeder 2000). Comparison to a coastal-to-marine rift setting (e.g. Oligocene-to-Miocene, Gulf of Suez), the early syn-rift phase of basin fill is characterised by fluvial or shallow-marine systems, which is influenced by the magnitude of tidal-and wave-dominated processes (Gawthorpe et al., 1997; Sharp et al., 2000). During sea level lowstands, rifts segments can become isolated and develop into freshwater or saline lakes. Alluvial fans and fan deltas can also develop under such conditions and in areas away from significant clastic input, reefs and talus development occurs (e.g. Miocene, Suez rift; Gupta et al., 1999). In deeper water conditions, a period of sediment starvation, bypassed slopes, slump scars, debris flows and turbidity currents dominate (Prosser, 1993; Ravnås and Steel, 1998; Gawthorpe and Leeder 2000; McLeod et al., 2002). The Central North Sea is considered as a marine rift basin and therefore this literature review focuses on syn-rift sedimentation in a coastal-plain to shallow-marine rift basin (Fig. 2.1).

2.2 Influence of salt on structural style and evolution of rift basins

2.2.1 Mechanisms for salt movement

The word “salt” is typically used to include all rock bodies composed primarily of halite but that contain varying amounts of other evaporites such as anhydrite and gypsum. The ability of salt to flow like a fluid makes it unique compared to the brittle behaviour of siliciclastic and carbonate rocks, under similar geological strain rates. Its ability to flow depends on the strength of the overburden and the boundary friction within the salt layer (Hudec and Jackson 2007). Differential loading has been identified as the dominant force driving salt flow and can be subdivided into gravitational, displacement and thermal
loading (Jackson and Talbot 1986; Hudec and Jackson 2007). Gravitational loading depends on the weight of the overburden rocks overlying the salt and gravitational forces within the salt layer for salt movement (Figs. 2.3 and 2.4; Hudec and Jackson 2007). Displacement loading occurs in response to extension or compressive forces displacing the boundary of a rock relative to another rock and thermal loading is driven by temperature changes associated with volume changes within the salt layer (Fig. 2.4; Jackson and Talbot 1986).

A critical property of salt flow is that it can transform a tabular body of salt into a wide variety of salt-cored structures (Fig. 2.5; Jackson and Talbot 1986). The lithology within a salt unit will impact its ability to flow. For example, carbonates and anhydrites are less likely to deform than halite (Clark et al., 1998). Salt structures normally evolve from concordant, low amplitude structures (e.g. salt anticlines, salt rollers and salt pillows) to discordant, high-amplitude intrusions (e.g. salt walls and salt diapirs) and extrusions (e.g. salt sheets and salt canopies) (Fig. 2.5).

Salt diapirs are a common type of salt structure and they form as a result of the upward movement of salt. They continue to grow until they encounter a resistive overburden or their source layer is exhausted (Jackson and Talbot 1986). Diapir growth can occur as a result of reactive, active or passive diapirism (Fig. 2.6). Reactive diapirism occurs when regional extension allows salt to migrate upwards through faults and fractures within the tectonically thinned overburden (Vendeville and Jackson 1992a). Sub-salt extensional normal faults can also influence the location of diapir development (e.g. Dooley et al., 2005). With increasing displacement, reactive diapirs are typically formed in the footwall of sub-salt normal faults, and gradually rise as the overburden is thinned. The rate of reactive diapirism is independent of viscosity, lithology or density of the overburden, and
is also controlled by the rate of regional extension (Jackson et al., 1994). In all cases, reactive diapirs stop rising when extension ceases (Vendeville and Jackson 1992a). A transition from reactive diapir rise to passive diapirism occurs in response to forcible intrusion of the overburden driven by density difference between dense cover and less dense salt. During active diapirism, strata next to the growing diapir may be extremely deformed (Fig. 2.6). Unlike reactive diapirism, whose growth is externally controlled by the rate of regional extension, the rate of active diapir rise is limited by local stresses around the salt diapir (Vendeville and Jackson 1992a; Jackson and Vendeville 1994; Hudec and Jackson 2007). Once the diapir has pierced the overburden and is near to the exposed sediment surface it may continue to grow by passive diapirism (Fig. 2.6; Vendeville and Jackson 1992a; Hudec and Jackson 2007). A diapir can grow passively whilst sediments accumulate around it and onlap its flanks. Passive diapir growth ceases only when salt flow cannot keep up with sedimentation or salt supply is exhausted (Rowan et al., 2003).

2.2.2 Structural styles in salt-influenced rift basins

Many rift basins are associated with pre-rift salt, such as the Red Sea, Mediterranean, Persian Gulf and the North Sea. During rifting, a salt layer provides a detachment between the basement and cover structures. The degree of linkage between the basement and cover structures is dependent on the ratio of salt thickness to fault displacement (D_r) (Koyi et al., 1993; Lewis et al., 2013; Fig. 2.7). Where the salt detachment is thin, the ratio of salt thickness to basement fault displacement is low (D_r < 1), through-going normal faults connects the sub- and supra-salt stratigraphy and the earliest cover faults may be offset from the basement structure (e.g. Fig. 2.7b; Erratt 1993; Koyi et al., 1993; Jackson and
Salt rollers typically develop where the salt is relatively thin. Conversely, salt diapirs develop where the salt is relatively thick at specific locations relative to basement faults (Fig. 2.7c; Jackson and Cramez 1989; Stewart and Clark 1999). After being emplaced in this way, salt diapirs can grow until the source layer is exhausted or is pinched off by compression (Fig. 2.7d; Rowan et al., 2003).

Experimental models by Vendeville and Jackson (1992b) illustrate that regional extension (rifting) triggers reactive diapirism, initiating diapir rise and normal faulting. Once the diapir has pierced the overburden and is exposed at the surface it may continue to grow passively until sedimentation ceases or the salt supply is exhausted (Fig. 2.8; Vendeville and Jackson 1992b). Salt supply may wane during continued extension, forming salt welds, thin-skinned extensional faults, diapir-collapse minibasins and adjacent turtle-structure anticlines (Fig. 2.8; Vendeville and Jackson 1992b; Hudec et al., 2009). During extreme extension, salt structures themselves may become cannibalised, resulting in the flexure of minibasins downwards to form mock turtle anticlines (Fig. 2.8; Vendeville and Jackson 1992b).

During rifting in salt-influenced basins, extensional normal faulting affects the sub- and supra-salt stratigraphy (Vendeville and Jackson 1992; Koyi et al., 1993; Vendeville and Jackson 1994; Lewis et al., 2013). However, extensional tectonics can occur in the absence of rifting where it is confined to the supra-salt stratigraphy (thin-skinned) in response to gravity tectonics. Here, the supra-salt overburden becomes stretched two or three times its original length by normal faulting but the sub-salt stratigraphy remains the same length. Such gravity-driven extension of the supra-salt stratigraphy above a salt detachment is termed ‘rafts’ (Duval et al., 1992). It is commonly documented at the edge of continental
2.3 Impact of faulting, folding and salt tectonics on shallow-marine sedimentation in salt-influenced rifts

During syn-rift, sedimentation is affected by the rate of fault-controlled subsidence, salt-controlled topography, the volume of sediment supply and eustasy within the basin (e.g. Leeder and Gawthorpe, 1987; Prosser, 1993; Davis et al., 2000). During the early phase of rift development, extensional forced folds will impact the position of depocentres in the hangingwall of normal faults with strata thickening into the adjacent hangingwall syncline (Fig. 2.1). Syn-depositional topographic expression of salt structures influences both stratal thickness and facies distributions, with strata commonly thinning and shallowing towards salt-cored structural highs (Davison et al., 1996; Alves et al., 2003; Kieft et al., 2010). This early phase of basin fill is dominated by incipient hangingwall drainage of alluvial, fluvial or shallow-marine systems followed by a period of sediment starvation and corresponding deposition of lacustrine or deep-marine facies (Prosser, 1993; Ravnås and Steel, 1998; Gawthorpe and Leeder, 2000; McLeod et al., 2002; Fig. 2.1). It has been recognised that during the nucleation phase of fault-array development, localised depocentres develop adjacent to fault displacement maxima forming isolated basins for lacustrine deposits (Davis et al., 2000; Gawthorpe and Leeder, 2000; Fig. 2.1a). Once the fault has breached the surface, the greatest accommodation will be created along the fault zone in which strata will thicken towards the fault zone (Gawthorpe et al. 1997; Fig. 2.1b). Where the volume of sediment supplied into the basin keeps pace with increasing accommodation, the surface expression of the tectonic activity may be subdued and
therefore have less of an impact on facies distribution and architecture (Frostick and Steel, 1993). Conversely, where sediment supply is low and unable to keep up with accommodation creation, positive basin topography will develop and facies distributions will be affected (Alves et al., 2003). Therefore it is the rate of sediment supply that will control whether deposition can keep pace with developing topography, regardless of the stage in rifting.

Topography within a developing rift basin will influence the distribution of wave, tidal and fluvial processes (Rowan and Weimer, 1998; Cartwright et al., 2001; Shelley and Lawton, 2005). This is well illustrated in outcrop in the La Popa Basin, Mexico, where salt diapirs and their encasing strata are well exposed. Syn-depositional topography generated by halokinesis influenced the wave-dominated shallow-marine shoreface sediments of the Cretaceous Delgado Sandstone Member where stratal units thin and become more shallow-water in facies character towards the diapir (Aschoff and Giles, 2005). Tidal amplification may occur where shallow-marine systems form during the early syn-rift, within narrow, elongate depocentres (e.g. the Middle Jurassic Hebrides, Ravnås and Bondevik, 1997; the Middle Jurassic Hugin Formation, Keift et al., 2010). Larger wave fetch may occur during periods of tectonic quiescence, especially when footwall islands become completely submerged and a more open-marine embayment occupies the rift basin (e.g. Upper Jurassic Fife/Angus Formation, Spathopoulos et al., 2000).

In this section, two examples are discussed which describes the presence of pre-rift salt and normal faulting on syn-rift sedimentation; the Central North Sea and the Parentis Basin in the Aquitaine Pyrenean Region of southwest France.
2.3.1 Central Graben, UK sector North Sea

The Central Graben is a broadly symmetric graben located along the southern arm of the North Sea trilete rift system within the North Permian evaporite basin (Fig. 2.9a, b; Ziegler 1990; Hodgson et al., 1992; Glennie et al., 2003). Deposition of the shallow-marine Fulmar Formation was influenced by Late Jurassic rifting and salt tectonics (Fig. 2.9c; e.g. Johnson 1986; Wakefield 1993; Stewart and Clark 1999). The significant lateral variability in the thickness of the Fulmar Formation has been attributed to complex fault and salt-related topography present during the time of deposition (Wakefield et al., 1993; Fraser et al., 2002; Stewart, 2007). On a regional scale, deposition of the Fulmar Formation has been described to be dominated by salt-cored palaeo-valley networks surrounded by low relief hills cored by Triassic pre-rift strata (Fig. 2.10a). Within this framework, sandstone deposition and preservation reflects the geometry and connectivity of the palaeo-valley network and the availability of sand and accommodation creation above collapsing salt walls (Fig. 2.10a; Stewart & Clark 1999; Fraser et al., 2002). Recent work also suggests that salt-withdrawal may have been controlled by sub-salt normal fault extension (Kuhn et al., 2003).

The Fulmar Formation sandstones are interpreted to have lenticular ‘pod’ shaped cross-sectional geometries in which the thickest and most complete sequences are towards the axial part of the ‘pod’ while towards the margins strata may thin and become amalgamated (Fig. 2.10b; Johnson et al., 1986). Deposition of the Fulmar Formation records several progradational to aggradational parasequences during an overall net-transgression (Gowland, 1996). The dominant architecture records an overall relative sea level rise, and the Fulmar Formation is overlain by the deep marine Kimmeridge Clay Formation, suggesting a rapid deepening (Fig. 2.10b). Sedimentologically, the Upper Jurassic
sandstones in the Fulmar Formation, penetrated in the Fulmar, Fife and Angus fields, are predominantly very fine-to-medium grained, shallow-marine, heavily bioturbated sandstones with poorly preserved primary sedimentary structures and occurrences of belemnites and bivalve shell fragments. Depositional models to explain the distribution of bioturbated shallow-marine sandstones in localised depocentres (e.g. in the Fife and Angus fields) suggest that erosion and sediment routing via input points along the Mid North Sea High introduced sediments into a wave/storm-dominated shallow-marine environment, in which the sands were extensively reworked (Spathopoulos et al., 2000). The sands are heavily bioturbated and their primary sedimentary structures destroyed where the rate of accumulation was slow relative to the rate of biogenic reworking. If the margins of the tectonically formed structural highs flanking the basin were steep, narrow shorefaces may have formed (Fig. 2.10c; Howell et al., 1996).

Three principal models have been proposed to explain accommodation creation above salt walls (see Chapter 4). These models are based on salt dissolution (Clark et al., 1999), salt-withdrawal (Hodgson et al., 1992) and reactive diapirism (Penge et al., 1993, 1999). In the Western Shelf area salt-dissolution is interpreted to have influenced Fulmar Formation deposition. Palaeo-valley systems are interpreted as a result of preferential erosion of the exposed salt instead of the adjacent Triassic minibasin fill (Fig. 2.10a). The erosion created topographic relief of tens of metres within which deposition was focussed (Wakefield et al., 1993; Stewart and Clark, 1999; Stewart 2007). Towards the Western Graben, in the Kittiwake, Puffin and Fulmar fields, Jurassic salt withdrawal is interpreted to have created accommodation for deposition of the Fulmar Formation (Fig. 2.9b). Subsequent inversion in the Late Cretaceous led to diapir rejuvenation altering the previous rift salt geometries (e.g. Stewart and Clark 1999; Stewart 2007).
2.3.2 Parentis Basin, Aquitaine Pyrenean Region, southwest France

The Parentis Basin is the western offshore continuation of the onshore Aquitaine Pyrenean Basin (Fig. 2.11a). Its formation and evolution was controlled by the relative motion of the Iberian and Eurasian plates as the North Atlantic Ocean opened and was strongly influenced by salt tectonics (Canéro et al., 2005; Biteau et al., 2006; Ferrer et al., 2012). Two episodes of rifting (Permian-Triassic) and (Late Jurassic-to-Early Cretaceous), followed by a period of Late Cretaceous-to-Eocene convergence resulted in a wide array of salt-related structures (e.g. salt anticlines, salt diapirs, salt walls, diapir-collapse, salt glaciers), originated from the Triassic Keuper evaporites (Figure 2.11b, c; Ferrer et al., 2012). These salt-structures were interpreted to initiate during the Late Jurassic-to-Early Cretaceous in response to extension and reactive diapirism (Ferrer et al., 2012). Differential thickness of growth strata on the flanks of salt structures suggest that sediment loading also played a role during the growth of these structures (Fig. 2.11c). This is evident in the depocentre of the Parentis Basin where subsidence of a 5 km thick Barremian-Albian sequence expelled the Keuper evaporites towards the edges of the basin where salt cored anticlines formed (Ferrer et al., 2012). Aggradation to the west during the Jurassic-to-Cretaceous was faster and salt diapirs there did not extrude to the surface. Deep diapirs grew in the footwall of extensional faults. Reefs grew above the salt-cored anticlines (e.g. Antares, off the northern Parentis edge; Fig. 2.11c). The distribution of these carbonate build-up was probably controlled by eustacy, salt tectonics and palaeogeography similar to the Mesozoic in the La Popa Basin, Mexico and the modern Persian Gulf (Giles and Lawton 2002). Salt diapirs evolved to a passive mode during the Albian and many stopped growing by the Late Cretaceous. As Iberia and Eurasia collided and drove the Pyrenean orogeny in the Late Cretaceous the Parentis Basin was mildly
shortened and most of the salt structures responded readily to compression and absorbed most of the Pyrenean shortening (Biteau et al., 2006; Ferrer et al., 2012). The alternation of extension and shortening controlled the location and style of salt structures. Diapirs which were dormant and buried were rejuvenated by compression as evident by their pinched off stems (Fig. 2.11c). Locally, pre-shortening of salt walls expelled salt which advanced as glaciers over the seafloor and was subsequently buried in the Oligocene. No new diapirs formed as the Parentis Basin was inverted because the autochthonous Keuper source layer was largely depleted (Ferrer et al., 2012).

The Parentis Basin is similar to the Central North Sea Graben in that it documents the structural styles and stratigraphic record in a pre-rift basin which experienced subsequent diapir rejuvenation.

2.4 Questions thesis aim to address

Even though it is recognised that the presence of pre-rift mobile evaporites results in variable structural styles and sedimentation patterns as compared to basins where evaporites are absent, the extent of its influence remains somewhat ambiguous. While it is recorded that salt acts as a detachment between the sub- and supra- salt stratigraphy, and topographic expressions of faults and salt structures influences deposition where strata commonly thins and shallows towards the crest of salt structures and thickens towards the hangingwall of normal faults, quantitative analysis of such is limited. This leaves a number of unanswered questions in which this thesis aims to address. For example; what are the various structural styles in salt-influenced rift basins? How does location within the rift (i.e. rift margin vs. rift centre), and associated variations in sediment supply and accumulation rate, impact the net-transgressive stratigraphic architecture of shallow marine syn-rift strata in salt-influenced rifts? Does syn-deposition in salt-influenced rifts
always result in rapid lateral changes in the thickness and facies of shallow-marine syn-rift strata? Can we apply existing tectono-stratigraphic models for rift basins such as those published by Gawthorpe and Leeder (2000) to basins where salt is present?
Isolated lacustrine basins in low sediment supply basins

Alluvial basin fill due to capture of antecedent river

Major axial sediment input to growth syncline with sediment reworked by tide and wave currents

Growth monocline above blind fault

Overall deepening trend in stratigraphy due to increasing displacement rates

Antecedent drainage diverted around propagating fault tips

Consequent drainage developing at segment boundaries

Pre-rift regional palaeoslope

Fluvial/coastal plain

Alluvial fans/fan deltas

Lake

Marine gulf

Tidal sandwaves

Fault scarp degradation causes major slides generating basinal megabreccias

High subsidence rates outpace sediment supply leading to deepening of basin and slope by-pass

Reversed drainage due to uplifting footwalls

Large catchments and fans mark breached segment boundaries

Sediment by-pass on steep, sediment starved footwall scarps

Axial turbidites sourced from intra basin slides and axial/hangingwall deltas

Tilting of basin floor generates vertical stacking of axial turbidites adjacent to footwall scarp

Fig. 2.1: Tectono-stratigraphic evolution during rifting in a coastal plain-to-marine depositional environment. (a) Rift initiation. (b) Through-going fault stage (taken from Gawthorpe and Leeder, 2000).
RIFT INITIATION:
Perfect wedge shapes to reflector packages, minor onlap onto the hangingwall, discontinuous hummocky internally. Possible progradation (real or apparent), no evidence of important footwall derived sediments.

IMMEDIATE POST-RIFT:
Discontinuous parallel reflectors, with possible progradational and aggradational reflectors close to the footwall. Compaction syncline over the basement footwall cut-off point.

LATE POST-RIFT:
Continuous parallel reflectors, less compaction induced deformation. Strong onlap and burial.

RIFT CLIMAX:
Chatoic zone close to the footwall scarp; aggradation and downlap if resolution is good enough. Divergence of basinal equivalents. Lozenge shapes or low angle downlaps on the hangingwall dip-slope if preserved. Minor onlap at top of hangingwall slope.

Fig. 2.2: Seismic expression of rift basin evolution (taken from Prosser 1993).
Fig. 2.3: Examples of hydraulic head gradients in salt tectonics (z = elevation head, t = thickness of the overburden, p = density).
(a) Lateral varying overburden thickness above a horizontal tabular salt layer produces a pressure gradient from Point 1 to Point 2 but no elevation head gradient. Salt will flow from left to right along the pressure head gradient. (b) A uniform overburden thickness above an inclined, tabular salt layer will produce an elevation head gradient from Point 1 to Point 2 but no pressure head gradient. Salt will flow from left to right down the elevation head gradient. (c) A uniform overburden thickness above a flat-lying salt layer produces neither elevation nor pressure head gradients even though the salt thickness varies. Salt remains at rest because there is no hydraulic head gradient. (taken from Hudec and Jackson 2007).
Fig. 2.4: Mechanisms for salt flow. (a) Gravitational loading from the weight of the overburden and gravitation forces within the salt layer. (b) Displacement loading as a result of basin extension. (c) Thermal loading driven by a localised heat source within the basement generating convection currents within the salt layer and resulting in temperature and associated volume changes within the salt layer (modified from Jackson and Talbot 1986).
Fig. 2.5: Salt structural geometries (modified from Hudec and Jackson 2007).
Fig. 2.6: Models of diapir growth as a result of reactive, active and passive diapirism (modified from Vendeville and Jackson 1992a).
(a) No salt  (b) Thin salt  (c) Thick salt  (d) Structured salt

Fig. 2.7: Schematic rise and fall of diapirs during extension (modified from Vendeville and Jackson 1992b).
Fig. 2.8: 2D schematic of structural styles associated with basement rifting with variation in pre-rift salt thickness. (a) Where salt is absent, classic half-graben geometries evolve. (b) Where salt is thin, a cover fault array develops above the basement scarp. (c) Where thick salt is present, early cover rafts separated by salt diapirs are draped above the evolving basement topography. (d) Where salt is present in locally isolated structures, reactivation of salt diapirs may occur, such that they actively rise or fall. (modified from Stewart and Clark 1999).
Fig. 2.9: Geological setting for the Fulmar Formation, UK sector of the North Sea. (a) Regional basin framework of the Central North Sea Rift. (b) Structural framework of the Central North Sea Graben highlighting various hydrocarbon fields (Kittiwake, Puffin, Fulmar, Fife and Angus) with Upper Jurassic Fulmar Formation reservoirs. (c) Generalised lithostratigraphy of the Central Graben (modified from Fraser et al., 2003).
Fig. 2.10: (a) Palaeogeography of the Western Central Shelf illustrating the structural controls during shallow-marine deposition of the Upper Jurassic Fulmar reservoirs (modified from Stewart et al., 1999). (b) Schematic cross-section of the ‘pod-shaped’ geometry of the Upper Jurassic facies distribution across the Fulmar Field (taken from Johnson et al., 1986). (c) Schematic cross-section of the ‘pod-shaped’ geometry of the Upper Jurassic facies distribution across the Fife and Angus fields (taken from Spathopoulos et al., 2000).
(a)

(b)

Fig. 2.11 c
Fig. 2.11: Geological setting for the Parentis Basin, Aquitaine Pyrenean region, southwest France. Modified from Ferrer et al., 2012.
(a) Regional basin framework of the Parentis Basin. (b) Generalised tectonostratigraphic framework of the Parentis Basin.
(c) S-N geoseismic cross-section of the Parentis Basin. See Fig. 2.11a for line of section.
CHAPTER 3

Structural controls on the stratigraphic architecture of
net-transgressive shallow-marine strata in a salt-influenced rift basin:

Middle-to-Upper Jurassic Egersund Basin, Norwegian North Sea

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3.1 ABSTRACT

In this study 3D seismic reflection, wireline log, biostratigraphic and core data from the Egersund Basin, Norwegian North Sea were integrated to determine the impact of syn-depositional salt movement and associated growth faulting on the sedimentology and stratigraphic architecture of the Middle-to-Upper Jurassic, net-transgressive, syn-rift succession. Borehole data indicate that Middle-to-Upper Jurassic strata consist of low-energy, wave-dominated offshore and shoreface deposits, and coal-bearing coastal plain deposits. These deposits are arranged in four parasequences that are aggradationally to retrogradationally stacked to form a net-transgressive succession that is up to 150 m thick, at least 20 km in depositional strike (SW-NE) extent, and >70 km in depositional dip (NW-SE) extent. In this rift-margin location, changes in thickness but not facies are noted across active salt structures. Abrupt facies changes, from shoreface sandstones to offshore mudstones, only occur across large displacement, basement-involved normal faults. Comparisons to other tectonically active salt-influenced basins suggest that facies changes across syn-depositional salt structures are observed only where expansion indices are >2. Subsidence between salt walls resulted in local preservation of coastal-plain deposits that cap shoreface parasequences, which were locally removed by transgressive erosion in adjacent areas of lower subsidence. The depositional dip that characterises the Egersund Basin is unusual and likely resulted from its marginal location within the evolving North Sea rift and an extra-basinal sediment supply from the Norwegian mainland.

3.2 INTRODUCTION

Depositional models for salt-influenced rift basins indicate that the growth of salt-cored structures is a first-order control on basin physiography, accommodation development and the resultant stratigraphic architecture of shallow-marine syn-rift deposits. For example,
syn-rift deposits in these settings may become thinner and be composed of progressively shallower water facies onto the flanks of syn-depositional, salt-cored structural highs (Davison et al., 1996; Stewart & Clark 1999; Alves et al., 2003; Aschoff & Giles, 2005; Kieft et al., 2010). In some cases, segmentation of a basin by salt structures and salt-related normal faults can result in the formation of at-seabed relief, local amplification of tidal currents, and deposition of syn-rift successions that contain a strong tidal signature (e.g. Mellere & Steel, 1996; Carr et al., 2003; Jackson et al., 2005; Kieft et al., 2010). Outcrop examples that highlight these complex tectono-stratigraphic relationships between salt-related structures and coeval shallow-marine depositional systems are rare (Alves et al., 2003; Aschoff & Giles, 2005; Giles & Rowan, 2012; Banham & Mountney 2013), and subsurface examples have tended to focus exclusively on either the structural or stratigraphic aspects of salt-influenced rift basin development and have not, therefore, provided a comprehensive account of how the two are linked (Stewart & Clark 1999; Davies et al., 2000). We therefore lack detailed case studies focused on the influence of salt tectonics on the sedimentology and stratigraphic architecture of shallow marine syn-rift deposits in salt-influenced rift basins, and our understanding of how these systems evolve is therefore relatively limited when compared to their deep-marine counterparts (e.g. Jackson et al., 2010). As a result of this paucity of studies, there are a number of unanswered questions related to this topic; for example, does syn-depositional deformation in salt-influenced rifts always result in abrupt lateral changes in the thickness and facies of shallow-marine syn-rift strata? What are the relative roles of regional tectonics, eustasy and relatively short length-scale salt-related deformation on the overall stacking pattern of transgressive strata? How does location within the rift (i.e. rift margin vs. rift centre), and associated variations in sediment supply and accumulation rate, impact the stratigraphic architecture of shallow marine syn-rift strata in salt-influenced rifts?
In this study seismic reflection and well data from the Egersund Basin, Norwegian North Sea to was used: (1) characterise the main structural styles within a salt-influenced rift basin; (2) describe the sedimentology and stratigraphic architecture of a Middle-to-Upper Jurassic, net-transgressive, coastal plain-to-shallow marine syn-rift succession (Bryne, Sandnes and Egersund formations); and (3) investigate the combined role that rift-related normal faulting and salt tectonics had on the thickness of and facies distribution within, a shallow marine syn-rift succession. Structure and thickness (isochron) maps generated from the interpretation of 2D and 3D seismic reflection data provide an understanding of the present-day and syn-depositional structural framework, and its control on the deposition of the Middle-to-Upper Jurassic succession. Detailed core descriptions allow a sedimentary facies framework to be erected for the studied succession; this framework is then integrated with wireline-log and 3D seismic data to provide an understanding of the stratigraphic architecture of the succession away from areas of core and well control. Low depositional gradients at this time resulted in the formation and preservation of a relatively wide belt of wave-dominated shoreface deposits. In the centre of the basin, thin-skinned extension resulted in the formation of normal faults and graben-like depocentres above laterally continuous (>45 km in length) salt walls that developed during an earlier period of salt movement. Furthermore, roof arching above active salt diapirs caused syn-depositional deformation of the seabed and the formation of intra-basin structural highs. Normal faulting and supra-diapir roof arching together influenced the preservation and possible original depositional thickness of the net-transgressive, shallow marine syn-rift succession; however, a lack of facies changes across these structures suggests that there was a delicate balance between sediment accumulation and accommodation, and that pronounced changes in water depth did not occur across these structures.
3.3 LOCATION AND TECTONO-STRATIGRAPHIC EVOLUTION OF THE EGERSUND BASIN

The study area is located in the Egersund Basin, which lies in the Central North Sea, to the east of the UK Central Graben and ca. 115 km offshore SW Norway (Fig. 3.1A). The Egersund Basin is bound to the west by the Sele High, the Stavanger Platform to the NE, the Flekkefjord High to the south and the Lista Nose to the SE (Fig. 3.1B) (Sørensen et al., 1992). The tectono-stratigraphic evolution of the Egersund Basin is described below in order to provide a regional context for the structures that influenced deposition during the Middle-to-Upper Jurassic rift period.

Permian

Late Carboniferous-to-Permian transtension led to the formation of the North Permian Basin, and resulted in the initial opening of the Egersund Basin and Åsta Graben (Sørensen et al., 1992). The oldest sediments encountered in the Egersund Basin are composed of non-marine clastics that belong to the Early Permian Rotliegend Group (e.g. wells 17/12-2 and 10/5-1; Fig. 3.1B). A marine transgression during the Late Permian caused flooding of the basin and resulted in deposition of the evaporite-dominated Zechstein Supergroup (Glennie et al., 2003). On structural highs, such as the footwalls of thick-skinned normal faults, the Zechstein Supergroup is relatively thin or absent and is dominated by carbonates and clastics that is not associated with large salt structures (e.g. wells 17/12-2 and 10/5-1; Fig. 3.1B) (Jackson & Lewis, in press). In contrast, in basinal areas, the Zechstein Supergroup is relatively thick and dominated by halite and anhydrite, and is associated with large salt walls and stocks (e.g. wells 9/2-4s, 9/2-8s and 9/2-2; Fig. 3.1B). Post-depositional mobilisation of the Zechstein Supergroup controlled the overall
structural style and physiography of the Egersund Basin during the Middle Jurassic-to-
Early Cretaceous rift event and, therefore, deposition of the Middle-to-Upper Jurassic syn-
rift succession considered herein.

**Triassic**

Non-marine clastics of the Smith Bank and Skagerrak formations, which are together up to
2000 m thick in the Egersund Basin (well 17/12-1R), were deposited throughout the North
Permian Basin during the Triassic (Fig. 3.2A) (Vollset & Doré, 1984; Goldsmith *et al.*, 2003; McKie & Williams, 2009). Rifting during the Triassic reactivated some of the
Permo-Triassic thick-skinned faults, and this resulted in flow of the Zechstein Supergroup
evaporites and the growth of salt structures (Fig. 3.1C). In the centre of the basin where the
Zechstein Supergroup was halite-rich and relatively mobile, a series of salt walls,
anticlines and stocks formed, which were flanked by Triassic rafts (Fig. 3.1C).

**Jurassic**

A marine transgression during the Early Jurassic led to the deposition of shallow marine
sandstones of the Gassum Formation and shallow marine mudstones of the Fjerritslev
Formation, both of which are only locally preserved as a result of Middle Jurassic uplift of
the Mid-North Sea Dome (e.g. Rattey & Hayward, 1993; Underhill & Partington, 1993). In
the Egersund Basin, much of the Early Jurassic is represented by the ‘intra-Aalenian’ or
‘mid-Cimmerian’ unconformity (Fig. 3.2) (Ziegler 1990; Underhill & Partington 1993; Fraser *et al.*, 2003). Subsidence of the Mid-North Sea Dome during the early part of the
Middle Jurassic resulted in flooding of the Egersund Basin and deposition of a net-
transgressive, Middle-to-Upper Jurassic succession that is 200-700 m thick. The lower part
of this succession comprises fluvial-to-coastal-plain deposits of the Bryne Formation,
which unconformably overlie Triassic strata across the intra-Aalenian unconformity (Fig. 3.2B). Regional biostratigraphic data suggest that the poorly dated Bryne Formation (Aalenian-to-Bathonian?) is unconformably overlain by shallow-marine deposits of the Sandnes Formation (Callovian-to-Oxfordian) (Fig. 3.2A) (Vollset & Doré, 1984); the corresponding Middle Callovian unconformity is however difficult to identify in our dataset. The close proximity of the Egersund Basin to the Norwegian mainland during the Middle Jurassic meant that the main sediment supply to fluvial and shallow marine, syn-rift depositional systems was likely from the basin margin.

Continued relative sea-level rise occurred during the latest Jurassic (Kimmeridgian and Volgian), and deposition of the open marine, mudstone-dominated Egersund, Tau and Sauda formations (Fig. 3.2A) (e.g. Vollset & Doré, 1984; Sørensen et al., 1992). This study focuses on the impact that Middle Jurassic-to-Early Cretaceous, rift-related extension and salt tectonics had on the sedimentology and stratigraphic architecture of the Bryne, Sandnes and Egersund formations, which equate to the J20- J50 units of Partington et al. (1993).

**Cretaceous**

Rifting ceased during the Early Cretaceous, although an ongoing rise in eustatic sea-level resulted in continued deepening of the Egersund Basin and deposition of a very fine-grained, open marine succession (Cromer Knoll Group; Vollset & Doré, 1984). During the Late Cretaceous, a global sea-level highstand resulted in widespread deposition of very fine-grained carbonates of the Shetland Group (e.g. Vollset & Doré, 1984; Sørensen et al., 1992). Basin inversion in the Late Cretaceous, triggered by the far-field effects of the Alpine Orogeny, resulted in minor reverse reactivation of basement-involved normal faults, and squeezing and amplification of salt diapirs (Fig. 3.1C) (Sørensen et al., 1992;
Jackson *et al.*, 2013). Thin-skinned normal faults in the centre of the Egersund Basin, which are localised above salt walls, do not appear to have been reverse reactivated during this period of inversion.

### 3.4 DATASET

This study utilises both 2D and 3D seismic reflection surveys (Fig. 3.1B). The 3D seismic reflection survey, which was acquired in 2005, comprises pre-stack, time-migrated data that covers 3600 km² and has an inline and crossline spacing of 25 m. Inlines are oriented NW-SE and crosslines are oriented NE-SW. The vertical axis is measured in milliseconds two-way time (ms TWT) and the data has a record length of 6600 ms TWT. The vertical resolution within the Middle-to-Upper Jurassic interval of interest is *ca.* 25 m (based on a dominant seismic frequency of 40 Hz and an interval velocity of 4000 m s⁻¹). The seismic data are displayed with SEG reverse polarity; a downward increase in acoustic impedance is represented by a negative or blue reflection, and a downward decrease in acoustic impedance is represented by a positive or red reflection (Fig. 3.2B) (Brown, 1999). The regional, time-migrated, 2D seismic lines are from multiple surveys that were acquired from 1997 to 2006. 2D seismic data was used to tie wells that lie outside the 3D seismic survey area. Seismic lines displayed in this paper are from the 3D seismic survey.

The following five regional, age-constrained, stratigraphic surfaces were identified and mapped on seismic in the Permian-to-Jurassic interval (Fig. 3.2B):

(i) top Rotliegend Group (258 Ma)

(ii) top Zechstein Supergroup (251 Ma)

(iii) top Triassic (*ca.* 210 Ma)

(iv) top J20 (165 Ma; i.e. near top Bathonian)

(v) top J50 (144 Ma; i.e. near top Oxfordian)
These surfaces define major changes in lithology, and they are therefore characterised by marked changes in density and velocity that manifest on seismic data as regionally mappable seismic reflections (Fig. 3.3). In the central part of the Egersund Basin, the J40 unit, which consists of the Sandnes Formation, is well developed and above the tuning thickness of the seismic data (Fig. 3.2B & 3.3). In this area, two additional J40 reflection events, which help illustrate some of the subtle stratal geometries present in the unit, were mapped (i.e. top J40 and intra-J40A; Fig. 3.2B-C).

Twenty-five wells penetrate the Middle-to-Upper Jurassic succession and they all have a standard suite of wireline log data (Table 3.1). Twenty of the wells have time-depth curves; these allowed chronostratigraphic, sedimentological and stratigraphic observations from boreholes to be tied to the seismic data and two-way travel time measurements derived from seismic data to be converted to thickness and/or depth. Fourteen wells are cored and these provide a total of 524 m of core in the Middle-to-Upper Jurassic succession; 435 m in the Sandnes Formation and 89 m in the Bryne Formation. Nine wells contain detailed biostratigraphic picks that can be correlated to the regional North Sea biostratigraphic scheme of Partington et al. (1993) (Fig. 3.1B & Table 3.1).

3.5 METHODOLOGY

This integrated study utilises a combination of geological and geophysical methods that include: (1) sedimentological logging at 1:100 scale to characterise sedimentary facies in core; (2) petrofacies characterisation and interpretation of core facies; (3) stratigraphic correlation of age-constrained surfaces and stratigraphic packages between wells; (4) tying of well data to seismic through construction of synthetic seismograms, and interpretation of seismic data to define the seismic-stratigraphic framework; (5) construction of isochrons of key stratigraphic intervals to determine the location and magnitude of structural controls.
on thickness and facies variations; and (6) generation and interpretation of seismic attribute volumes, such as coherency, to identify faults and salt structures.

3.6 STRUCTURE OF THE EGERSUND BASIN

Time-structure maps were generated for mapped seismic horizons, and isochron maps were constructed for the intervening seismic stratigraphic units. Together these maps allowed identification of the key structural features within the basin and the control they had on the thickness of the Middle-to-Upper Jurassic succession. The following three key types of structures were identified in the Egersund Basin: (1) thick-skinned, basement-involved normal faults; where ‘basement-involved’ is defined as faults that offset sub-salt strata (i.e. Rotliegend Group and older); (2) salt structures related to flow of the Zechstein Supergroup; and (3) thin-skinned, cover-restricted normal faults, which detach downward unto the salt and do not therefore deform basement (Fig. 3.4).

3.6.1 Thick-skinned, basement-involved normal faults

The top Rotliegend Group time-structure map indicates that the Egersund Basin trends NW-SE and is bounded by the Stavanger Platform and Lista Nose to the N-NE and by the Sele High to the NW (Fig. 3.4A). The Stavanger Platform is separated from the Egersund Basin by the Stavanger Fault Zone; this is a NW-SE-striking, SW-dipping fault system that has up to 1.5 km of throw and which only locally breaches the salt to the north where the salt is thin (Lewis et al., 2013). The Lista Nose is a N-S-striking, ca. 13 km wide horst, which is bound by N-S-striking faults that decrease in displacement northwards and which offset the salt. The eastern margin of the Sele High is defined by a NE-SW-striking, SE-dipping, Sele High Fault System, which has up to 1.5 km of displacement, but which does not offset the salt. In the centre of the Egersund Basin, NW-SE- and WNW-ESE-striking
normal faults are developed; these faults have relatively low displacements (<500 m) and are typically overlain by salt diapirs (outset map in Fig. 3.4A & D) (see below). Thick-skinned faults in the Egersund Basin likely formed in response to crustal extension during the Permian, Triassic and Jurassic-Cretaceous rift events (Sorensen et al., 1993; Jackson et al., 2013; Lewis et al., 2013).

3.6.2 Salt structures

Seismic mapping of the top Zechstein Supergroup allowed the following five main types of salt structure to be identified: (1) salt walls; (2) salt stocks; (3) salt rollers; (4) salt pillows; and (5) salt anticlines (\textit{sensu} Hudec & Jackson, 2007) (Fig. 3.4B). In the centre of the Egersund Basin, salt walls are located above the NW-SE- and WNW-ESE-striking, sub salt-restricted normal faults (Fig. 3.4B). Salt stocks are locally developed along these walls, and they are sub-circular in plan view, up to 5 km in diameter, and have a relief of up to 4.5 km (i.e. the Delta, Omega, Epsilon and Chi diapirs; Fig. 3.4B). The main salt walls are up to 4 km wide, up to 50 km long and have considerably lower relief than the stocks (i.e. up to 1 km) (i.e. the Beta and Alpha salt walls; Fig. 3.4B). Where sub salt-restricted normal faults change strike from WNW-ESE to NW-SE, a \textit{ca}. 8 km wide zone of salt rollers is developed, which occur in the footwalls of thin-skinned normal faults (i.e. the Lambda Roller Zone; Fig. 3.4B). The Stavanger Salt Pillow is a sub-circular, 2.5 km diameter, 700 m relief salt structure that is located in the footwall of the Stavanger Fault Zone (Fig. 3.4B). The Xi Salt Anticline is a broadly N-trending, 5 km wide, 500 m amplitude, salt-cored anticline located \textit{ca}. 10 km to the east of the Sele High Fault System (Fig. 3.4B).
3.6.3 Thin-skinned, basement-detached normal faults

Two main types of thin-skinned, basement-detached normal faults are identified in the Egersund Basin. The first type are developed above the Alpha and Beta salt walls, have up to 300 m of throw, and typically occur in oppositely-dipping pairs that define 1-2 km wide graben (Fig. 3.4C and D). These relatively planar faults offset Triassic-to-Jurassic strata, and tip-out downwards into the Zechstein Supergroup. The second type of faults is located away from the main salt walls in areas of relatively thin salt. They form 1-2 km wide symmetrical graben and tip-out downwards into the Triassic succession (Fig. 3.4C) (Tvedt et al., in press).

Supra-salt faults can form by one or a combination of three principal mechanism, at distinct times in the evolution of a salt structure: (i) by *subterranean salt dissolution*, which is accommodated by the formation of inward-dipping normal faults (Ge & Jackson 1998) – faults of this origin typically form after the main period of growth of the structure; (ii) by *active diapirism*, which results in folding (arching) and extension of the overburden – these faults typically form relatively late in the growth history of the structure (Schultz-Ela et al., 1993; Jackson et al., 1994; Hudec and Jackson, 2007; Yin & Groshong 2007); and (iii) thin-skinned regional extension, which triggers, rather than is triggered by, salt diapirism – these faults typically form before the related salt structure (Vendeville & Jackson 1992a; Vendeville & Jackson 1992b; Jackson et al., 1994; Stewart & Clark 1999; Davison et al., 2000; Stewart 2007).

Considering these mechanisms in terms of the tectono-stratigraphic setting and evolution of the Egersund Basin, salt dissolution may have triggered supra-wall extensional faulting, although the presence of a Triassic roof above many of the salt structures, even where supra-salt faults are present, argues that this was not the main mechanism. The observation that a relatively thin, Triassic-to-Jurassic interval is locally
arched above some high-relief parts of the salt walls (Fig. 3.5D) suggests that active
diapirism may locally have been important for the formation of supra-salt normal faults.
Along-strike from these locations, where the walls have considerably less relief, many of
the supra-salt faults occur in areas that do not display folding of Triassic and Jurassic strata
across the underlying structures. In these locations, the supra-salt faults formed due to
post-Triassic regional extension, which caused the salt walls to widen and collapse
(Vendeville & Jackson 1992a; Jackson et al., 1994) (Fig. 3.4D).

3.7 SEISMIC-SCALE GEOMETRY AND DISTRIBUTION OF THE MIDDLE
TO-UPPER JURASSIC SYN-RIFT SUCCESSION

3D seismic reflection data indicate that the Middle-to-Upper Jurassic syn-rift succession,
which comprises the Bryne and Sandnes formations, display prominent changes in
thickness adjacent to some of the structure described above. For example, an isochron map
of the J20-J50 interval (Fig. 3.5A) and borehole data from well 9/3-1 and 9/2-1 (Fig. 3.5B)
indicate that this unit thickens from the footwall to the hangingwall of the thick-skinned,
basement-involved, Stavanger Fault Zone (Table 3.2; Fig. 3.5A). A similar change in
thickness across the Sele High Fault System and the fault bounding the western margin of
the Lista Nose is observed on seismic data (Fig. 3.5A). In the footwall of the Stavanger
Fault System, the J20-J50 interval is folded above but does not display seismic-scale
thickness variations across the Stavanger Salt Pillow, implying that this structure
developed after the main period of Late Jurassic-to-Early Cretaceous rifting (Fig. 3.5B)
(see also Jackson et al., 2013).

In the axis of the Egersund Basin, thickness variations are commonly observed along
the major salt walls (Alpha, Beta and Gamma) and the salt anticline (Xi) (Fig. 3.5A). For
example, borehole and seismic data indicate that J20-J50 succession thins by ca. 45% from
the flank (well 17/12-3) to the crest (well 17/12-1R) of the Xi Salt Anticline (Table 3.2; Fig. 3.5C). Similar thickness variations are observed between wells 9/2-4s and 9/2-6s, which are located on the crest and flank, respectively, of the Alpha salt wall (Table 3.2; Fig. 3.5D). In this example, stratigraphic thinning indicated by borehole data is associated with stratal convergence of the J50 unit onto the underlying J20 unit, which is folded across the underlying salt wall. Thickening of the J20-J50 unit into the supra-salt wall graben are also observed on seismic data, although borehole data was used on both sides of the bounding faults to directly constrain the exact amount of stratal thickening in the specific interval of interest (Fig. 3.4D-E). These observations clearly indicate that some of the main salt-related folds and faults were growing during the Middle-to-Late Jurassic, and that they were therefore controlling the physiography and accommodation development within the Egersund Basin at this time. Having established that many of the thick- and thin-skinned normal faults, and salt-related structures were active during the Middle-to-Upper Jurassic, the stratigraphic response to the evolving rift structural style was further explored. This was achieved by undertaking a detailed sedimentological analysis of the syn-rift succession, which then allows us to erect a sequence-stratigraphic framework. Within this framework the sub-seismic variability in thickness and facies were analysed, and the detailed stratigraphic response to the structural development of this salt-influenced rift basin.

3.8 FACIES ANALYSIS OF MIDDLE-TO-UPPER JURASSIC SYN-RIFT DEPOSITS

Twelve facies were identified in the Bryne and Sandnes formations (J20-J40 units), based on lithology, grain size, degree of sorting, nature of bedding contacts, primary sedimentary structures, bioturbation intensity (using ‘bioturbation index’ of Taylor and Goldring 1993)
and trace fossil assemblages (Table 3.3). These facies are grouped into the following five facies associations: (FA1) offshore; (FA2) offshore transition; (FA3) lower shoreface; (FA4) upper shoreface; and (FA5) coastal plain (Table 3.3). The facies association nomenclature is adopted from the bioturbated shoreface model of Gowland (1996), which places mean fair weather wave base (FWWB) at the base of the lower shoreface facies association, and mean storm wave base (SWB) at the base of the proximal offshore transition facies association. The Bryne Formation of the J20 unit contains only the coastal plain facies association (FA5) whereas the J40 unit of the Sandnes Formation contains intervals of all five facies associations (FA1-FA5).

3.8.1 Facies association 1 (FA1): Offshore facies association

Description: This facies association is present in the upper part of the Sandnes Formation (e.g. Fig. 3.6A), and passes gradationally upward into open marine mudstones of the Egersund Formation. FA1 consists of structureless, unbioturbated, calcite-cemented mudstone (facies 1.1) (Table 3.3). It is only cored in well 9/2-6s, where it abruptly overlies lower shoreface deposits (FA3) (Fig. 3.6A). Biostratigraphic data from well 9/2-5 (Fig. 3.1B) suggest FA1 was deposited in a fully marine, offshore environment.

Interpretation: The very-fine grained character of FA1 suggests deposition by suspension fall-out. Lack of stratification and bioturbation is attributed to overprinting by calcite cementation. Because of its vertical and lateral association with shallow-marine deposits (FA2-FA4), in conjunction with biostratigraphic data from well 9/2-5 (Fig. 3.1B) and its lateral continuity in wells in the distal part of the basin (e.g. 18/10-1), FA1 was interpreted to document offshore deposition on a marine shelf or ramp.
3.8.2 Facies association 2 (FA2): Offshore transition facies association

*Description:* This facies association is present in the Sandnes Formation and occurs in association with the lower and upper shoreface facies associations (FA3 and FA4). FA2 comprises bioturbated silty mudstone (facies 2.1) and bioturbated siltstone (facies 2.2) (Table 3.3). Facies 2.1 is friable and poorly preserved, making it difficult to identify sedimentary structures and quantify bioturbation intensity. Facies 2.2 is highly bioturbated (BI = 5) by a diverse trace fossil assemblage (*Planolites*, *Anconichnus*, *Terebellina*, *Rosselia*, *Teichichnus*) (Fig. 6B). Rare, thin (<10 cm) intervals of hummocky cross-stratified siltstone are observed. FA2 underlies and overlies lower shoreface deposits (FA3) (e.g. Fig. 3.6B).

*Interpretation:* The fine-grain size and abundant bioturbation that characterises facies 2.1 and 2.2 suggests overall low rates of deposition, predominantly by suspension settling in a low energy environment. Rare hummocky cross-stratified beds indicate infrequent, episodic deposition by storm-generated oscillatory or combined flows below FWWB (Dott & Bourgeois 1982). The diverse trace fossil assemblage is typical of the *Cruziana* ichnofacies, which indicates a low-energy offshore-to-lower shoreface environment of deposition (MacEachern & Barn 2008). In shallow marine deposits of the Late Jurassic Fulmar Formation in the UK Central Graben, *Terebellina* and *Anconichus* characterise offshore transition and distal lower shoreface deposits (Gowland 1996; Martin & Pollard 1996). The finer grain size of facies 2.1 is interpreted to reflect a more distal depositional environment than facies 2.2.
3.8.3 Facies association 3 (FA3): lower shoreface facies association

*Description:* This is the most common facies association encountered in the cored wells, and thick (9-34 m) successions occur in the Sandnes Formation. FA3 is vertically and laterally associated with the offshore transition, upper shoreface and coastal plain facies associations (FA2, FA4 and FA5, respectively) and is restricted to the centre and eastern part of the Egersund Basin. FA3 is typically characterised by highly bioturbated, hummocky cross-stratified sandstones, which occur in successions with overall coarsening-upward or uniform grain size trends (facies 3.1) (Table 3.3). Successions can be subdivided into bedsets (Campbell 1967) that are defined by: (1) highly-bioturbated (BI = 5-6) intervals, which are separated by packages of hummocky cross-stratified beds, in which there is little or no bioturbation (BI = 0-1); or (2) by moderately-bioturbated (BI < 3), hummocky cross-stratified intervals. Bedsets are 3–34 m in thickness, but are typically 10–20 m thick (Fig. 3.6B). Trace fossil assemblages are relatively diverse (*Planolites, Chondrites, Thalassinoides, Aulichnites, Anconichnus, Rosselia, Teichichnus, Ophiomorpha, Cylindrichnus, Palaeophycus, Skolithos*). Facies 3.2 comprises predominantly massive, well-sorted sandstones containing rare planar lamination, low-angle cross-lamination and high-angle cross-bedding, which may be defined by carbonaceous drapes. Heavy oil staining in some cores hampers recognition of sedimentary and biogenic structures in FA3 (Fig. 3.6A, B & C).

*Interpretation:* The presence of hummocky cross-stratified beds (facies 3.1) is indicative of deposition during storm events (Dott & Bourgeois 1982). The tops of these beds were then bioturbated during quiescent conditions below FWWB before the next storm event. The trace fossil assemblage represents a mixture of the *Skolithos* and *Cruziana* ichnofacies, which is typical of lower shoreface deposits (MacEachern & Barn, 2008).
Thus, facies 3.1 is interpreted to document deposition above mean SWB in a lower shoreface environment (sensu Gowland 1996). The massive appearance of sandstones in facies 3.2 most likely reflects a relatively uniform grain size. The different styles of lamination likely reflect varying hydrodynamic conditions; for example, high-angle cross-bedding indicates dune migration under lower flow regime conditions, whereas planar lamination may indicate upper flow regime conditions (Harms et al., 1975). The very-fine grained, well-sorted character of the sandstones implies extensive reworking and winnowing, consistent with deposition above fair-weather wave base.

3.8.4 Facies association 4 (FA4): upper shoreface facies association

Description: FA4 is present in the Sandnes Formation and, in a similar manner to FA3, is restricted to the centre and eastern part of the Egersund Basin. In the centre of the Egersund Basin it is commonly overlain by coastal plain or lower shoreface deposits (FA5 and FA3, respectively) (Fig. 3.6C). On the Flekkefjord and Lista highs, FA4 is overlain by offshore marine mudstones of the Egersund Formation, and on the Stavanger Platform it is by offshore transition deposits (FA2). Units of FA4 consist of either well-to-moderately sorted, medium-grained sandstones that lack a vertical grain size trend (facies 4.1), or an overall upward-fining, poorly sorted sandstones that is underlain by an erosional, coarse quartz sand grain-mantled surface (facies 4.2) (Fig. 3.6B and Table 3.3). Primary sedimentary structures include planar lamination, symmetrical ripples and trough cross-bedding. In some instances, laminations are defined by carbonaceous drapes or stringers of coarse quartz sand grains (Fig. 3.8C). Bioturbation is typically absent to moderate (BI = 0-3), and the trace fossil assemblage is characterised by Aulichinites, Planolites, Skolithos and Ophimorpha, with rare Thalassinoides, Rosselia and Cylindrichnus.
**Interpretation:** Variability in sorting likely reflects the degree of sediment reworking, with the well-sorted, subrounded sand grains (facies 4.1) being intensively reworked by fair weather waves, and moderately-to-poorly sorted, subangular grains (facies 4.2) being sourced from nearby fluvial input points. The lack of fine-grained sediments in FA4 suggests deposition under the continuous action of relatively strong currents. Intercalation of symmetrical ripples, trough cross-bedded, planar and low-angle parallel stratified beds indicated deposition by oscillatory waves and unidirectional currents under fluctuating hydrodynamic energy conditions (Harms et al., 1975). The sharp, erosional base and upward-fining character of facies 4.2 indicate deposition under conditions of gradually decreasing energy, which may record channel migration and eventual abandonment. Carbonaceous drapes within this facies may reflect periods of reduced velocity related to the action of tides (e.g. Nio & Yang, 1991) or low fluvial discharge. The relatively low intensity and diversity of bioturbation is indicative of a mobile substrate or high-energy setting, and is typical of the Skolithos ichnofacies that characterises middle shoreface-to-foreshore deposits (*sensu* MacEachern & Barn 2008). In summary, the sedimentological and ichnological characteristics described above are similar to those reported from upper shoreface deposits of several Jurassic units in the central and northern North Sea (e.g. the Fulmar Formation, UK Central Graben; the Etive Formation, Horda Platform; the Ula Formation, Norwegian Central Graben; see Gowland 1996; Fraser *et al.*, 2003; Bullimore & Helland-Hansen, 2009; Jackson *et al.*, 2010).

### 3.8.5 Facies association 5 (FA5): Coastal plain facies association

**Description:** This facies association is present throughout the study area in the Bryne Formation, but is only locally preserved in the centre of the Egersund Basin in the Sandnes Formation (wells 9/2-4s, 9/2-6s, 9/2-3, 9/2-5, 9/2-9s; Fig. 3.1B). In the Sandnes Formation,
FA5 is commonly overlain by lower shoreface deposits (FA3) across a sharp or erosional surface, and underlain by upper shoreface deposits (FA4) (e.g. Fig. 3.6C). FA5 is composed of five facies that are stacked vertically into heterolithic successions of variable thickness (Table 3.3). Thin (<2 m), carbonaceous and root-penetrated sandstones (facies 5.1) underlie coals and carbonaceous siltstone (facies 5.2). Erosionally based, well- to poorly-sorted, coarse- to fine-grained, carbonaceous sandstones occur in thick (up to 18 m) successions that lack bioturbation and contain asymmetrical ripples, planar-low angle lamination, tabular and tough cross-bedding (facies 5.3). Bioturbated, organic-rich siltstones contain ripples, synaeresis cracks and fine-grained sandstone laminae (facies 5.4). Bioturbation is sparse to moderate (BI = 1-3) and is represented by a low-diversity assemblage of *Planolites* with rare *Rosselia*, *Cylindrichnus*, *Palaeophycus* and *Phycosiphon*. Non-bioturbated, organic-rich, silty mudstones occur in a thick (17 m) succession in well 9/2-2 (facies 5.5).

**Interpretation:** Coals (facies 5.2) can occur in a wide range of depositional environments, but their close association with shallow marine deposits in this context suggests a coastal plain setting characterised by a high water table and low clastic sediment input (e.g. Bohacs & Suter 1997). Rooted sandstones beneath coals (facies 5.1) indicate that peat, which eventually formed coal, accumulated *in situ*. These sandstones are interpreted as root-penetrated foreshore or coastal plain deposits within the Sandnes and Bryne formations, respectively. Asymmetrical ripples, planar-low angle lamination, tabular and trough cross-bedding in erosionally based beds sandstones (facies 5.3) suggest deposition by unidirectional currents of fluctuating velocity, most likely related to variations in discharge and sediment load in fluvial or deltaic distributary channels. Organic-rich siltstones that contain a restricted trace fossil assemblage and syneresis cracks (facies 5.4)
were deposited in an ecologically stressed marginal marine environment of fluctuating salinity, such as a lagoon. Non-bioturbated, organic-rich silty mudstones (facies 5.5) are interpreted to reflect deposition of suspended material under anoxic, low-energy conditions such as an anoxic lagoon or lake. The presence of these five facies within a single association is consistent with a coastal plain environment of deposition.

### 3.8.6 Wireline-log expression of facies and facies associations

In order to extend our facies association interpretation into uncored intervals, petrofacies interpretation of wireline-log signatures and cross-plots of absolute wireline-log values were used to identify responses for each facies association where cored (Fig. 3.7).

Only sparse core data are available in FA1 (offshore shelf) to calibrate the wireline-log character of this facies association. High gamma-ray (> 93 API) and low neutron porosity (16-23 v/v) values distinguish FA1 from underlying and overlying facies associations (e.g. FA3; Fig. 3.6A). FA2 (offshore transition) is typically represented by a serrated gamma-ray log signature with overall moderate values (> 40 API), and is characterised by relatively restricted neutron (9-13 v/v) and density values (2.47-2.65 g cm⁻³) when compared to the overlying and underlying shoreface facies associations (FA3 and FA4; Fig. 3.6B). The wireline-log expression of FA3 (lower shoreface) is defined by an overall blocky gamma-ray (< 100 API), density (2.3-2.6 g cm⁻³) and neutron (3-23 v/v) log signature, onto which small scale (<5 m) coarsening-upward trends are superimposed; it is not possible to distinguish facies 3.1 and 3.2 in wireline logs (Fig. 3.6B). FA4 (upper shoreface) has a similar petrofacies character to the lower shoreface facies association (FA3) and they are difficult to distinguish from each other using wireline logs alone (Fig. 3.6B & C). They both display an overall blocky, gamma-ray (< 100 API), density (2.3-2.65 g cm⁻³) and neutron (5-35 v/v) log signature, onto which small scale (<5 m)
coarsening-upward trends are superimposed (Fig. 3.6B & C). FA5 (coastal plain) is represented by a combination of wireline-log signatures. Coals (facies 5.2) are characterised by very low density (1.5-2.0 g cm\(^{-3}\)) log values (Fig. 3.6C & D). Sandstones of facies 5.1 and 5.3 are typically represented by funnel-shaped gamma-ray log curves of low values (< 70 API), and a distinct separation between the density and neutron log curves (Fig. 3.6D). Organic-rich siltstones (facies 5.4) in the Bryne Formation generally display higher gamma-ray log values (< 80 API) than associated sandstone and coal facies (Fig. 3.6D). Organic rich mudstones (facies 5.5) in the Bryne Formation typically display higher gamma-ray (> 100 API) and density (2.35-2.78 g cm\(^{-3}\)) log values than facies 5.4.

Quantitative wireline-log analysis indicates that facies in the offshore and offshore transition facies associations have distinct wireline-log characteristics in the sparse data available, and can be easily discriminated from shallow marine facies associations. However, facies within the lower and upper shoreface facies associations have a similar expression on wireline-log data, such that they cannot be consistently distinguished from one another; hence, for the purposes of wireline log-based interpretation and regional stratigraphic correlation, they were grouped together (facies 3.1, 3.2, 4.1, 4.2; Table 3.3; Fig. 3.7B). Facies in the coastal plain facies association (facies 5.1, 5.2, 5.3, 5.4, 5.5) display a variable petrofacies character and there is considerable overlap in terms of the sandstone facies (facies 5.1 and 5.3); low-density coals (facies 5.2) and organic-rich silty mudstones (facies 5.5) can however be easily distinguished (Fig. 3.7C). In summary, many of the facies that can be confidently recognised in core do not display distinctive wireline-log characteristics, preventing their identification away from core control. Consequently, qualitative and quantitative wireline-log-based petrofacies interpretations are less detailed than their core-based counterparts. The following four groups of facies associations were therefore interpreted for the purpose of wireline-log interpretation and correlation: offshore
(FA1); offshore transition (FA2); shoreface FA3 and FA4); and coastal plain (FA5) (Fig. 3.8).

3.9 SEQUENCE STRATIGRAPHIC FRAMEWORK OF THE MIDDLE-TO-UPPER JURASSIC SYN-RIFT SUCCESSION

Stratigraphic analysis of the Middle-to-Upper Jurassic succession, constrained by the detailed facies association framework established above, allows identification of three, relatively low-frequency (ca. 5 Myr) sequences that correspond to the North Sea basin-wide, J20?, J40 and J50 chronostratigraphic intervals of Partington et al. (1993) (Fig. 3.2B). The poorly dated J20? sequence (Bryne Formation) consists of coastal plain deposits (FA5), and the overlying J40-to-J50 interval (Sandnes and Egersund formations) records an overall landward migration of the shoreline over a period of ca. 21 Myr (Fig 3.2B). Based on facies and facies association stacking patterns, five upward-coarsening units are recognised within the J40-to-J50 succession, and they are bound by relatively high-frequency (ca. 1 Myr ) flooding surfaces (surfaces FS1-FS5 bounding stratal units A-E; Fig. 3.8). At the resolution and distribution of available data, units A-E do not appear to contain erosional unconformities and basinward facies shifts, such that there is no evidence for base-level fall or sequence boundary formation in the J40-to-J50 succession (sensu Van Wagoner et al., 1988). Instead, each unit corresponds to a single regressive succession or parasequence (sensu Van Wagoner et al., 1988). The J20-to-J40 interval therefore represents a retrogradational parasequence set (sensu Van Wagoner et al., 1988).
3.10 DISTRIBUTION AND THICKNESS OF SYN-RIFT SEQUENCES; THE ROLE OF SYN-DEPOSITIONAL TECTONICS

In this section the sub-seismic sequence stratigraphic framework and gross facies distributions in the J20-to-J50 succession, and their relationship to the rift- and salt-related structures described above, are illustrated using two regional stratigraphic correlations. The first transect is oriented NW-SE, broadly parallel to regional depositional dip, and crosses the Sele High through the Egersund Basin in the NW (i.e. palaeoseaward), and the Lista Nose in the SE (i.e. palaeolandward) (Fig. 3.1B & 3.8A). The second transect is oriented NE-SW, broadly parallel to regional depositional strike, from the Stavanger Platform in the NE, through the axis of the Egersund Basin in the SW (Figs. 3.1B & 3.8B).

To synthesize our key observations a series of palaeogeographic maps were constructed for the four main stratigraphic units (B-E) in the Sandnes Formation, based on the facies association distribution at the point of maximum regression immediately below the capping flooding surface (Fig. 3.9). To highlight the role that pre-, syn- or post-depositional subsidence had on the original depositional or subsequent preserved thickness of the studied stratigraphic unit, the thickness of the J40 unit are displayed on the palaeogeographic maps (Fig. 3.9). Seismic data resolution is insufficient and well data are spaced such that it is not possible to determine if the constituent units of J40 (to lowermost J50) change in thickness adjacent to the salt structures and thin-skinned normal faults described above; however, they all lie within a package that either: (1) thickens towards faults bounding supra-salt graben; or (2) thins towards salt-cored structural highs, implying that these structures were active at the time they were deposited.
Unit A

Description: The base of unit A, which corresponds to the Bryne Formation, is defined by the mid-Cimmerian or Intra-Aalenian unconformity. Core data indicate that unit A is composed exclusively of coastal plain deposits (Figs. 3.2B & 3.8B). Unit A is overlain by a major flooding surface that marks a regional transgression and the onset of shallow marine deposition in the Egersund Basin (J20/FS1; Fig. 3.8). Seismic and borehole data indicate that Unit A is broadly tabular and at least 150 m thick across much of the Egersund Basin, and thins abruptly (to 50 m) onto the Lista Nose across the W-dipping fault that defines its western margin (Fig. 3.8A). Locally, however, in the axis of the Egersund Basin, seismic data indicate that Unit A is wedge-shaped and clearly thickens towards faults that bound the supra salt wall graben (Figs. 3.4D & 3.5B), and that the same unit thins across the salt anticline in the NW of the basin (Fig. 3.5C).

Interpretation: Coastal-plain sediments (FA5) in Unit A were derived either from the NW or from erosion of the Mid-North Sea Dome during the initial stages of deflation and subsidence of this structure, or from the NE and from erosion of mainland Norway; provenance data from Unit A would be required to test these two hypotheses. Accommodation during the Aalenian-to-Bajocian was likely generated by a combination of regional subsidence of the Mid-North Sea Dome and relatively local subsidence caused by the initiation of thick-skinned rift-related normal faulting along the margins of the Egersund Basin. In support of the latter interpretation is the observation that Unit A thins onto topographic highs such as the Lista Nose, which are bound by thick-skinned normal faults, and thinning of the unit across the Xi salt anticline, which is genetically linked to and was triggered by the initiation of slip on, the basin-bounding, Sele High Fault System.
Supra-salt wall graben was also active at this time as indicated by the wedge-shaped geometry of Unit A adjacent to the bounding thin-skinned normal faults.

**Unit B**

*Description:* Unit B is bound below and above by FS1 and FS2 respectively (Fig. 3.8). In the south-east and central part of the Egersund Basin Unit B consists of a 0-20 m thick package of shoreface sandstones, which pass downdip to the NW and pinchout into a significantly thinner (<5 m) package of offshore shelf mudstones and siltstone (e.g. wells 18/10-1 and 17/12-1R; Fig. 3.8A). A fine-grained succession is also developed on the Sele High, despite the fact that, present-day, the Sele High Fault System has ca. 1.5 km of displacement (Fig. 3.4A). In the SE part of the basin, Unit B displays a relatively constant thickness across strike (ca. 20 m; Fig. 3.8B), and it is generally thinner (e.g. 10 m, well 10/7-1) but still composed entirely of shoreface sandstones on adjacent topographic highs (e.g. Stavanger Platform, Lista Nose, Flekkefjord High; Fig. 3.8).

*Interpretation:* Unit B heralds the onset of shallow marine sedimentation in the Egersund Basin, triggered by a regional rise in relative sea-level and flooding of the basin, which was likely eustatically driven (Hardenbol et al., 1998), but locally augmented by minor rift-related normal faulting and subsidence. A very wide (up to 70 km along depositional strike), broadly SW-NE trending shoreface facies belt was interpreted to be established in the Egersund Basin at this time, and that this passed downdip to the NW into offshore shelf deposits (Fig. 3.9A). An important observation is that this shoreface belt extended outside of the main depocentre and onto the flanking fault-bound highs implying that, at this time, these structures were associated with only minimal if any structural relief. However, subtle thinning of Unit B suggests that these structures were active, but that the
sediment accumulation rate was enough to outpace the rate of fault-driven uplift. The absence of significant relief around and within the Egersund Basin meant that deposition occurred on a low-gradient shoreline, the position of which would have been sensitive to changes in relative sea-level and sediment supply.

**Unit C**

*Description:* Unit C is bound at its base by FS2 and at its top by FS3, and comprises of coastal plain, shoreface and offshore deposits (Figs. 3.8 & 3.9B). In the centre and SE of the Egersund Basin, Unit C comprises a single shallowing-upward, regressive succession that is >100 m thick, and which passes from offshore transition (FA2), through lower shoreface (FA3), to upper shoreface deposits (FA4) (e.g. well 9/2-1; Fig. 3.8B). In wells 9/2-4s, 9/2-3 and 9/2-5 this shallowing-upward succession is capped by thin (<2 m), coal-bearing, coastal-plain deposits (FA5) (Figs. 3.8B & 3.8C). Unit C thins north-westwards to only <5 m and the shoreface sandstones pinch-out into offshore deposits (e.g. wells 18/10-1 and 17/12-1R; Fig. 3.8A). Localized shoreface sandstones are encountered on the Sele High (e.g. well 17/12-2; Fig. 3.9B).

*Interpretation:* A relatively narrow (<15 km), NE-trending shoreface facies belt was interpreted to develop in the centre of the Egersund Basin at this time (e.g. 9/2-2, 9/2-1) (Fig. 3.9B). The proximal-to-distal facies trend, with coastal plain deposits passing north-westward into shoreface deposits, implies that the shoreline built out from SE to NW. In the centre of the Egersund Basin, coastal plain deposits are present in a number of wells at the top of the sequence (9/2-6s, 9/2-3, 9/2-4s, 9/2-5 and 9/2-9s; Fig. 3.9B). Where these deposits occur, the J40 unit is thick, suggesting that these coastal plain deposits did not accumulate on salt-related topographic high, but were instead deposited or at least only
preserved in areas of relatively high subsidence. The coastal plain deposits may have been more widely developed during maximum regression, but were then removed by erosion during subsequent transgression, except in high-subsidence depocentres (Fig. 3.9B). Shoreface deposits are also penetrated locally on the Sele High (well 17/12-2), and the abrupt transition into offshore deposits south-eastward across the Sele High Fault System suggest that these sediments were deposited as part of a shoreline system that fringed the SE margin of the Sele High, which at this time possibly formed an island (Fig. 3.9B). In contrast to the Sele High Fault System, the Stavanger Fault System and the west-dipping normal fault system bounding the western margin of the Lista Nose both appear to have been inactive based on the lack of a pronounced facies shift across them. However, thinning of Unit C onto the related structural highs indicates that uplift was occurring but that sediment accumulation rate outpaced the rate of fault-driven uplift.

Unit D

Description: Unit D overlies FS3 and is capped by FS4, which defines an upward transition from lower shoreface (FA3) to offshore deposits (FA1) (e.g. well 9/2-4s; Fig. 3.8A). In the centre of Egersund Basin, Unit D is \textit{ca.} 30 m thick and consists of lower shoreface (FA3) deposits (e.g. wells 9/2-2, 9/2-1, 9/2-4s, 9/2-3 and 9/2-5; Fig. 3.8B), which pass laterally in the NW to offshore deposits (FA1) where Unit D is relatively thin (<5 m; e.g. wells 17/12-1R and 18/10-1; Fig. 3.8A). On the basin-bounding structures highs (i.e. Lista Nose, Sele High, Stavanger Platform), Unit D is also composed of shoreface-dominated (FA4) succession (well 10/7-1; Fig. 3.8A) that is thinner than that developed in the centre of the basin (e.g. well 9/3-1; Fig. 3.8B).
Interpretation:

Facies distributions in Unit D comprise of a wide shoreface facies belt in excess of 70 km along depositional dip, which passes laterally to the NW into offshore shelf deposits (wells 18/10-1, 17/12-3 and 17/12-1R; Fig. 3.9C). The proximal-to-distal facies trend, with shoreface deposits passing north-westward into offshore deposits, implies that the shoreline built out from SE to NW. Shoreface deposits are penetrated in well 17/12-2 on the Sele High in this sequence which was an area of low subsidence (Fig. 3.9C). This is similar to deposition of Unit C, where shoreface deposits are also penetrated locally on the Sele High (well 17/12-2), and the abrupt transition into offshore deposits south-eastward across the Sele High Fault System suggest that these sediments were deposited as part of a shoreline system that fringed the SE margin of the Sele High, which at this time possibly formed an island (Fig. 3.9B). In contrast to the Sele High Fault System, the Stavanger Fault System and the west-dipping normal fault system bounding the western margin of the Lista Nose both appear to have been inactive based on the lack of a pronounced facies shift across them. However, thinning of Unit D onto the related structural highs indicates that uplift was occurring but that sediment accumulation rate outpaced the rate of fault-driven uplift.

Unit E

Description: Unit E is bounded at its base by FS4 and at its top by FS5, which also defines the top of the J40 unit (Fig. 3.8). In the Egersund Basin, this sequence is up to 10 m thick and consists of upward-fining offshore deposits (FA1), whereas on the Stavanger Platform it comprises offshore transition (FA2) deposits (well 9/3-1; Fig. 3.8B). Biostratigraphic data indicate that Unit E is absent on the Flekkefjord High (wells 9/4-1 and 8/12-1; Fig. 3.8B).
Interpretation: Unit E marks the onset of relatively distal, fully marine conditions across the Egersund Basin and flooding of the adjacent structural highs, which is likely linked to the major eustatically driven, tectonically augmented transgression that characterised the latest Jurassic-to-earliest Cretaceous across much of the northern North Sea (Figs. 3.8 & 3.9D).

3.11 DISCUSSION

3.11.1 Regional controls on syn-rift stratigraphic architecture in salt-influenced rifts

Thermal deflation and subsidence of the Mid-North Sea Dome was responsible for creating regional accommodation during the Middle-to-Late Jurassic and for controlling the overall net-transgressive stratigraphic architecture of the studied succession (Underhill & Partington 1993; Hampson et al., 2009). Overall thinning and pinchout of shoreface deposits towards the WNW-NW, and the overall NNE-SSW trend of the shoreline in successive depositional sequences, strongly suggest that sediment supply was either: (i) delivered directly to the Egersund Basin from the Stavanger Platform to the E; or (ii) fed along-strike into the basin by longshore currents. Despite the Middle-to-Late Jurassic, rift-margin setting of the Egersund Basin, sediment supply to the basin was relatively low and was eventually outpaced by a rise in relative sea-level, thereby resulting in the net-transgressive stratigraphic architecture. There were times, however, when sediment supply and regional subsidence were broadly in balance, such that a thick (＞150 m), aggradational package of shoreface deposits were deposited in the basin axis. Furthermore, the wide shoreface facies belts (Fig. 3.9A, C) suggest a relatively low shoreline gradient, which
would have resulted in the shoreline being extremely sensitive to small eustatic changes (Fig. 3.9B). However, the strongly aggradational motif of the studied succession suggests that such eustatic changes did not occur or, if they did, they were minor.

3.11.2 Local controls on syn-rift stratigraphic architecture in salt-influenced rifts

The growth of thick-skinned, basin-bounding fault systems, salt structures and thin-skinned, salt-related normal faults and folds locally overprinted the more regional controls on subsidence, accommodation and stratigraphic architecture described above. In the axis of the Egersund Basin, the large (>70 km) dip extent of shallow marine shoreface deposits is suggestive of relatively low depositional gradients; in this respect, the Egersund Basin acted a relatively stable, gently-subsiding structural block, despite the overall rift basin setting. However, seismic and borehole data indicate marked changes in thickness do occur on relatively short length scales (<5 km), especially towards the crests of salt structures and across thin-skinned, salt-related normal faults. These changes in thickness are not however associated with a change in facies, which suggests that sediment aggradation rates outpaced structurally induced changes in relief and accommodation (cf. Doglioni et al., 1998). As a result, syn-depositional intra-basinal normal faults and salt structures were not expressed at the depositional surface and they did not therefore influence the relative roles of physical processes (e.g. waves and tides) occurring within the basin. These structures were however critical in terms of creating additional accommodation and thus preserving locally thickened successions of syn-rift strata.

The only example of structurally induced facies partitioning occurs in the eastern part of the Egersund Basin in association with the basin-bounding Sele High Fault System. During the onset of rifting in the Middle Jurassic, offshore deposits extended across the
Sele High, implying that the Sele High Fault System was either not active, or was active but not associated with significant relief (Fig. 3.9A); a similar interpretation can be made for the Stavanger Fault System and the structure bounding the western margin of the Lista Nose. An abrupt change in water depth, which can be attributed to an increase in slip rate of and the relief associated with the Sele High Fault System, resulted in the development of a fringing shoreline in the immediate footwall of the fault, whereas deeper water conditions and mudstone deposition persisted in its hangingwall (Fig. 3.9B, C). This abrupt facies transition reflects steeper, fault-related bathymetric gradients during the rift-climax, at a time when fault throw rates were high, differential accommodation generation was greatest, and tectonic subsidence locally outpaced sediment supply (cf. Prosser, 1993; Gawthorpe & Leeder, 2000).

### 3.11.3 Controls on the stratigraphic development of net-transgressive shallow marine systems

In order to place the Middle-to-Upper Jurassic strata of the Egersund Basin in a wider context, a comparison was made with net-transgressive, tectonically-influenced shallow-marine strata in a number of other salt-influenced and salt-free sedimentary basins (Table 3.4). Although it is focused on salt-influenced rifts, the comparison was extended to rifts that lack salt and passive margin settings. In this way the key role that intra-basin salt tectonics can have on basin physiography, accommodation development and syn-rift stratigraphic architecture can further be highlighted. To structure this comparison several key parameters were focussed on: (1) sediment provenance; (2) structural styles active during deposition; (3, 4) thickness and net-to-gross ratio of the net-transgressive strata from the crest to the flank of salt structures, and from the footwall to hanging wall of normal faults; (5) expansion index across active structures (Thorsen, 1963; Cartwright et
al., 1998, Bouroullec et al., 2004; Jackson & Rotevatn 2013); (6) sediment accumulation rate; (7) depositional process regime; (8) dip extent of the shallow-marine sandstones in areas of active structures; and (9) orientation of shorelines relative to the structural grain of the basin (Table 4). For consistency, a net-transgressive shallow-marine succession is defined at its base by an abrupt deepening of facies above coastal plain deposits and at its top by a flooding surface into fine-grained, offshore deposits that document deposition below normal storm-wave base.

Our analysis indicates that Middle-to-Upper Jurassic strata in the Egersund Basin share many similarities to the other basin types. In salt-influenced basins an increase in thickness is noted from the crest to the flank of salt structures (Table 3.4) (e.g. Alves et al., 2003; Aschoff & Giles 2005; Kieft et al., 2010) and, in salt-free rift basins, syn-rift strata thicken from the footwall to hangingwall of active normal faults (Table 3.4) (e.g. Maxwell et al., 1999; Davies et al., 2000; Kuhn et al., 2003; Jackson et al., 2005; Zecchin et al., 2006). Such increase in thickness is typically accompanied by a decrease in sandstone content (i.e. net-to-gross) (Table 3.4). For example, where stratigraphic units thin towards the crest of salt structures, the net-to-gross increases (Table 3.4) (e.g. Alves et al., 2003; Aschoff & Giles 2005; Kieft et al., 2010). Furthermore, where expansion index values are > 2.0 facies changes are noted. A good example of this relationship occurs in the Bombarral-Alcobaca sub-basin western Portugal, where shallow-marine carbonates restricted to the crests of salt pillows pass downdip into mudstone-dominated, turbidite-bearing successions in the flanking minibasins (Alves et al., 2003). Similarly, in the Hugin Formation, South Viking Graben, northern North Sea, wave- and tide-influenced, shallow-marine deposits located at the crests of salt-cored structural highs pass downdip into offshore mudstones in adjacent minibasins (Kieft et al., 2010). In contrast where expansion index values are < 2.0, sandstone content remains relatively unaffected by
thickness variations across active structures, facies variations are absent, and aggradational
successions are developed, implying that relatively low rates of subsidence did not result
in across-fault changes in water depth and therefore depositional environment (Table 3.4)
(e.g. Maxwell et al., 1999; Kuhn et al., 2003). In some cases, syn-rift strata in the
hangingwall display a strong tidal influence, implying that the depocentre-bounding
structure was associated with a subaerial or shallow marine scarp, which locally
constrained and augmented tidal energy (Mellere & Steel, 1996; Maxwell et al., 1999;
Jackson et al., 2005; Zecchin et al., 2006).

The sediment provenance in most of the basins illustrated in Table 3.4 is extra-basinal,
with the exception of the Upper Jurassic, shallow marine sandstones of the Fulmar
Formation in the UK Central Graben (Kuhn et al., 2003). The dip extent of the Fulmar
Formation, which are genetically-related to but slightly younger than the Sandnes
Formation, is limited (up to 1.5 km) due to local sourcing from a relatively small, intra-
basinal fault scarp (Table 3.4) (Khun et al., 2003). In salt-influenced basins, the
depositional dip extent of the shallow-marine sandstones is < 40 km (Alves et al., 2003;
Aschoff & Giles 2005; Kieft et al., 2010), whereas in active rifts it is < 10 km (Maxwell et
al., 1999; Davies et al., 2000; Zecchin et al., 2006) and in passive foreland basins it is < 20
km (Olsen et al., 1999) (Table 3.4). It can be suggested that the anomalously large dip
extent (up to 70 km) of net-transgressive shallow marine sandstones in the Egersund Basin
is related to the low depositional gradient in the axis of the basin and its close proximity to
a sediment source (i.e. the Stavanger Platform and mainland Norway). In general, the
orientation of the shoreline relative to the trend of syn-depositional structures does not
seem to influence the depositional dip extent of the shallow-marine sandstones.

Net-transgressive sandstones developed along the western margin of the Western
Cretaceous Interior Seaway, US, which lay within a thermally subsiding foreland basin
and was thus tectonically passive for much of its evolution (e.g. Liu et al., 2011), were wave-dominated and display little evidence for the influence of tides. Shallow-marine sandstones were sourced from the basin margin cordillera and have a depositional dip extent of 9-20 km (Table 3.4) (Olsen et al., 1999). These values are similar to those for net-transgressive shallow-marine sandstones in salt-influenced basins and salt-free rift basins, implying that the depositional dip extent of such deposits is not simply related to basin type, but rather reflects local or sub-regional substrate gradient, and the location of the basin with respect to a large, typically extra-basinal sediment source (Table 4).

3.12 CONCLUSIONS

An integrated dataset of seismic, wireline-log, biostratigraphic and core data has been analysed in order to understand the influence of structural controls on depositional thickness and gross facies distributions of a net-transgressive succession in a salt-influenced, rift basin. Mapping of key seismic horizons provides the framework to characterise the main structural styles of the basin. When integrated with facies analysis of cores and wireline logs, biostratigraphically constrained correlation and seismic attribute analysis, seismic mapping enables the role of structures on subsidence, stratigraphic thickness and depositional patterns to be elucidated. The key findings of this study are:

- Thick-skinned, basement-involved normal faults, salt structures and thin-skinned, basement-detached normal faults are identified as the main structural controls influencing deposition of the Middle-to-Upper Jurassic Bryne, Sandnes and Egersund formations of the J20-J50 unit in the Egersund Basin. Similar to that observed in other salt-influenced basins, these structures influenced depositional thickness by creating accommodation space on the flanks of salt diapirs, walls and
pillows, in the hangingwall of normal faults, and in crestal graben above salt structures.

- Sedimentological descriptions of core and wireline-log data have identified five facies associations in the J20-J40 unit of the Bryne and Sandnes formations: (1) offshore; (2) offshore transition; (3) lower shoreface; (4) upper shoreface; and (5) coastal plain. Vertical facies successions define upward-shallowing, low-energy, wave-dominated offshore and shoreface deposits, capped by coal-bearing coastal plain successions bounded by five key flooding surfaces (FS1-FS5) that sub-divide the J20-to-J40 unit into five stratigraphic units (A-E). Wave-dominated, shallow-marine depositional processes are common in net-transgressive salt-influenced basins.

- Though these five stratigraphic units (A-E) are marked by stratal thickness changes across salt structures and most normal faults, facies variability is limited where calculated expansion index values are < 2.0. This suggests that sediment supply outpaced accommodation generation across actively growing structures throughout much of the study area during deposition of these stratigraphic sequences. In other salt-influenced basins, facies changes due to bathymetric or topographic expression of salt structures are only observed where expansion index values are > 2.0 and record a shallowing of facies from the flank to crest of salt structures.

- Abrupt lateral facies change linked to differential tectonic subsidence only occurs across a basement-involved normal fault in stratigraphic Units C and D, where shoreface deposits on the slowly subsiding Sele High pass into offshore deposits in the more rapidly subsiding Egersund Basin. This lateral change indicates that accommodation generation locally outpaced sediment supply in the fault hangingwall.
• Thin, coal-bearing coastal plain deposits that cap Unit C (just below FS3) are preserved only in depocentres associated with high rates of normal-fault-hangingwall and/or salt-withdrawal subsidence. Similar coastal plain deposits may have been present in adjacent areas of low subsidence, but were removed by subsequent transgressive erosion.

• The close proximity of the Egersund Basin to the heavily-eroded structural high of the Stavanger Platform and a low shoreline gradient can account for the anomalously large depositional dip extent (up to 70 km) of the Middle-to-Upper Jurassic net-transgressive shallow marine sandstones. In other salt-influenced basins, extra-basinally sourced, net-transgressive shallow-marine sandstones have depositional dip extents of <40 km. In contrast, corresponding dip extent in rift-basins where salt is absent are < 10 km.

• In this basin reservoir distribution is more widespread than originally predicted as compared to other tectonically-active salt-influenced basins. No unexpected lateral facies changes are observed during the growth of salt structures and internal facies heterogeneity are absent, reducing the risk of reservoir quality in the basin.

The results of this study highlight the complexity of proposing a distinct depositional model for salt-influenced rift basins. Basin physiography, rate of accommodation creation and sediment flux are the main factors controlling depositional thickness and processes, net:gross variability, depositional dip extent and the resulting stratigraphic architecture of the basin.
3.13 ACKNOWLEDGEMENTS

The authors would like to thank PGS for providing access to the seismic data and for granting permission to publish the results of this study. We also thank Mike Charnock for biostratigraphic data and related discussions, and the Norwegian Petroleum Directorate for providing access to the core data. Centrica Energi Norge and the Commonwealth Commission are gratefully acknowledged for funding and support of this work. Stuart Archer, Helen Lever and David Hodgson are thanked for their critical reviews which improved the manuscript. Journal editor Jeffrey Nunn is also thanked for his insightful suggestions.

3.14 REFERENCES


Fig. 3.1: (A) Simplified map illustrating the structural setting of the central North Sea. Only structural domains related to Mesozoic rifting are shown for clarity. Modified from Glennie (1998). The location of the geoseismic cross-section shown in (C) is indicated. (B) Structure map depicting main structural elements of the Egersund Basin, distribution of wells and location of geoseismic sections (Figs. 3.2 & 3.8). Structure taken at top-Rotliegend level from Sørenson et al. (1992). (C) Simplified geoseismic section across the Central Graben illustrating the main geographic basins, structural features and their spatial relationships (after Zanella and Coward 2003).
### Table: Main tectonic events in the Egersund Basin

<table>
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<tr>
<th>System</th>
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### Diagram: Genetic Units Formations and Mapped Seismic Horizons

- **Top Triassic** (IAU)
- **Top Zechstein**
- **Top Rotliegend**
- **Top J30**
- **Top J40**
- **Top J20?**
- **Top J40A**

### Sea Level Change

- **100 m**
- **1 km**
Fig. 3.2: (A) Regional stratigraphic column modified from Vollset & Doré (1984), illustrating Permian to Triassic stratigraphy of the Egersund Basin and global sea level changes relative to present day, Hardenbol et al. (1998). (B) Details of the Triassic to early Upper Jurassic stages. The “J” and “T” nomenclature for the genetic units corresponds to that established by Partington et al. (1993). (C) Seismic reflection events mapped in this study and their interpreted stratigraphic positions are indicated. The ages of stratal surfaces such as the top of the Tr10, J20, J40 and J50 units are constrained by biostratigraphic data.
**Fig. 3.3:** Synthetic seismograms illustrating the tie between well and seismic data for the Middle Jurassic to Triassic stratigraphy in wells 17/12-1R and 9/2-2, respectively from the marginal and central parts of the Egersund Basin (Fig. 3.1B). Both wells are displayed at the same scale and with seismic horizons interpreted. In well 17/12-1R the Sandnes Formation is below tuning thickness whereas in well 9/2-2 it is 107 m thick and can be sub-divided by the top J40 and intra-J40A reflections, which are mapped locally. GR = Gamma Ray (API units); RHOB = Density (g cm$^{-3}$); DT = Sonic (μs ft$^{-1}$); RC = Reflection Coefficient; AI = Acoustic Impedance.
-1730
-4150

-250
-3900

-1790
-4150

Stavanger
Fault Zone

Sele
High

Lista Nose

Delta
diapir

Xi salt
anticline

Omega
diapir

Stavanger
salt pillow

Beta
diapir
salt wall

Epsilon
diapir

Gamma
salt wall

Lambda
roller zone

Alpha salt
smaller salt walls

Chi
diapir

NW-SE normal faults

WNW-ESE normal faults

N 3 km

Top Zechstein Group (TWT) Structure

10 km

Top Rotliegend Group Dip Map

10 km

(B)
Fig. 3.4: (A) Top Rotliegend Group structure map illustrating main basin bounding topographic highs and thick-skinned normal fault orientations. Outset dip map highlights the WNW-ESE and NW-SE normal fault orientations in the Egersund Basin. (B) Top Zechstein structure illustrating the geometries and distribution of: salt walls, salt diapirs, salt anticlines, salt pillows and salt roller zones. (C) Top J20 coherency map highlighting main structural features such as the Stavanger Fault Zone, Lista Nose, Sele High, salt walls and thin-skinned normal faults. (D) Asymmetrical crestal graben above salt walls. See Figure 3.4C for location. (E) Intra-J40A to top J20 (Middle Jurassic; Fig. 3.2B) isochron map showing thickness variations within graben and adjacent footwalls. See Figure 3.4C for location.
Fig. 3.5: (A) Middle-to-Upper Jurassic (J20-J50) isochron map. (B) Well data and seismic cross-section illustrating thickness changes from the footwall of the Stavanger Fault Zone to the Egersund Basin. (C) Well data and seismic cross-section illustrating thickness changes from the flank to the crest of the Xi Anticline. (D) Well data and seismic cross-section illustrating thickness changes from the crest to the flank of the Alpha Salt Wall. See Figure 3.5A for line locations.
Fig. 3.6: Representative core and wireline logs through facies associations (FA) and vertical facies successions: (A) FA3 to FA1 in the uppermost J40 unit, well 9/2-6s; (B) FA2, FA3 and FA4 in the J40 unit, well 9/2-1; (C) FA3, FA4 and FA5 in the J40 unit, well 9/2-3; and (D) FA4 and FA5 in the J20 unit and lowermost J40 unit, well 8/12-1. Note different vertical scales are used in parts (A) to (D). B.I. = Bioturbation Index of Taylor & Goldring (1993).
(A) Facies Association 1 & 2: (offshore and offshore transition)
Facies 1.1: Structureless calcareous mudstone; Facies 2.1: Bioturbated silty mudstone; Facies 2.2: Highly bioturbated siltstone;
Depths 1800-3200 m SS

(B) Facies Association 3 & 4: (lower and upper shoreface)
Facies 3.1: Hummocky cross-stratified and bioturbated sandstone; Facies 3.2: Very fine to fine grained massive sandstone;
Facies 4.1: Medium grained, cross-stratified, bioturbated sandstone; Facies 4.2: Fining upward bioturbated sandstone;
Depths 3150-3250 m SS
(C) Facies Association 5: (coastal plain)

Facies 5.1: Rooted sandstone; Facies 5.2: Coal; Facies 5.3: Interbedded cross-stratified sandstone; Facies 5.4: Bioturbated organic-rich siltstones; Facies 5.5: Barren, organic-rich silty shale

Depths 2400-3130 m SS

Fig. 3.7: Wireline-log cross-plots of, in left-hand column, Gamma Ray (GR) vs. Sonic (DT) data and, in right-hand column, Density (RHOB) vs. Neutron Porosity (NPHI) data. These cross-plots illustrate the quantitative log character of the facies associations of the Bryne and Sandnes formations: (A) offshore and offshore transition facies associations (FA1, FA2); (B) lower and upper shoreface facies associations (FA3, FA4); and (C) coastal plain facies association (FA5).
Flooding surfaces (FS)/Stratigraphic Sequences

- Egersund Formation
- Sandnes
- Bryne
- Triassic

- Unit A
- Unit B
- Unit C
- Unit D
- Unit E

- 17/12-2
- 17/12-1R
- 18/10-1
- 9/2-2
- 9/2-1
- 9/2-4s
- 10/7-1

- 16.2 km
- 27.8 km
- 9.0 km
- 1.8 km
- 50.2 km

- NW SE
- Sele High
- Egersund Basin
- Lista Nose

- 14.8 km
- 120 m
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**Diagram: Egersund Basin**
- **17/12-1R**
- **18/10-1**
- **9/2-2**
- **9/2-1**
- **9/2-4s**
- **10/7-1**

**NW**

**Lista Nose SE**

- **Egersund Basin**
- **Lista Nose SE**
- **Epsilon salt diapir**
- **Xi salt anticline**

**Gaps:**
- **FS1**
- **FS2**

**Legend:**
- **200 m:**
- **5 km:**
**Fig. 3.8:** Well-log correlation panels and seismic cross-sections illustrating: (A) transect oriented NW (palaeoseaward) to SE (palaeolandward), from the Sele High through the Egersund Basin to the Lista Nose, and (B) transect oriented along the basin axis from the Stavanger Platform (NE) through the Egersund Basin (SW). See Figure 3.1B for line locations. Well log depths are in TVDSS (m); GR = Gamma Ray (API); NPHI = neutron (v/v); RHOB = density (g cm⁻³) Well-log correlation panels use the top-J20 flooding surface as a horizontal datum.
maximum extent of shoreface deposits based on core and well data

maximum extent of coastal-plain deposits based on core and well data

maximum extent of shoreface deposits based on core and well data

localised shoreline

(1) (2)
Fig. 3.9: Palaeogeographic maps with basin-bounding, thick-skinned normal faults, salt structures and J40 unit well thickness illustrating the distribution of facies associations at maximum regression during the four stratigraphic sequences (B-E) shown in Figure 3.8: (A) facies distribution below FS2; (B) facies distribution below FS3; (C) facies distribution below FS4; (D) facies distribution below FS5.
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<tr>
<td>17/12-3</td>
<td>✓</td>
<td>x</td>
<td>x</td>
<td>✓</td>
<td>✓</td>
<td>✓</td>
<td>✓</td>
<td>✓</td>
</tr>
<tr>
<td>18/10-1</td>
<td>✓</td>
<td>✓ (B)</td>
<td>x</td>
<td>✓</td>
<td>✓</td>
<td>✓</td>
<td>✓</td>
<td>✓</td>
</tr>
<tr>
<td>18/11-1</td>
<td>✓</td>
<td>x</td>
<td>✓</td>
<td>✓</td>
<td>✓</td>
<td>✓</td>
<td>x</td>
<td>✓</td>
</tr>
</tbody>
</table>

✓ = data available  x = data not available  (p) = poor data quality  B = Bryne  S = Sandnes

**Table 3.1:** Listing of well data availability and quality. Wireline log curves include: GR = Gamma Ray; RHOB = Density; DT = Sonic; NPHI = Neutron; RES = Resistivity; T - D = Time-Depth.
<table>
<thead>
<tr>
<th>Well name</th>
<th>Location</th>
<th>Unit thickness (m)</th>
<th>J20</th>
<th>J40</th>
<th>J50</th>
<th>J20-J50 thickness variation (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>9/3-1</td>
<td>Footwall Stavanger fault</td>
<td>112 55 69</td>
<td></td>
<td></td>
<td></td>
<td>60</td>
</tr>
<tr>
<td>9/2-1</td>
<td>Egersund Basin</td>
<td>376 147 72</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>17/12-1R</td>
<td>crest Xi salt anticline</td>
<td>104 16 75</td>
<td></td>
<td></td>
<td></td>
<td>45</td>
</tr>
<tr>
<td>17/12-3</td>
<td>flank Xi salt anticline</td>
<td>239 31 81</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>9/2-4s</td>
<td>crest Alpha salt wall</td>
<td>limited penetration 113 53</td>
<td></td>
<td></td>
<td></td>
<td>25</td>
</tr>
<tr>
<td>9/2-6s</td>
<td>flank Alpha salt wall</td>
<td>154 72</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

**Table 3.2:** Examples of thickness variations in the J20, J40 and J50 units of the Middle-to-Upper Jurassic Bryne and Sandnes formations.
<table>
<thead>
<tr>
<th>Facies Association</th>
<th>Lithofacies nomenclature</th>
<th>Lithological Descriptions</th>
<th>Bioturbation</th>
<th>Depositional Environment</th>
</tr>
</thead>
<tbody>
<tr>
<td>Offshore shelf</td>
<td>Structureless calcareous mudstone</td>
<td>Calcareous, structureless, calcite-cemented mudstone</td>
<td>BI = 0</td>
<td>Shelf</td>
</tr>
<tr>
<td>Offshore transition</td>
<td>Bioturbated silty mudstone</td>
<td>Very friable, silty, bioturbated mudstone</td>
<td>BI &gt; 3?</td>
<td>Offshore transition zone</td>
</tr>
<tr>
<td></td>
<td>Bioturbated siltstone</td>
<td>Bioturbated, carbonaceous siltstone, with thin (&lt; 10 cm scale) hummocky cross-stratification</td>
<td>BI = 5 (Pl, An, Tr, Ros, Tri, Te)</td>
<td>Offshore transition zone</td>
</tr>
<tr>
<td>Lower shoreface</td>
<td>Hummocky cross-stratified and bioturbated, sandstone</td>
<td>Very fine- to fine-grained, well sorted, carbonaceous sandstone. Sandstones are highly bioturbated which are separated by hummocky cross-stratified bedsets (3-34 m thick)</td>
<td>BI = 1-6 (Pl, Ch, Th, Au, An, Te, Ros, Op, Cy, Pa, Sk)</td>
<td>Lower shoreface</td>
</tr>
<tr>
<td></td>
<td>Medium-grained, cross-stratified, bioturbated sandstone</td>
<td>Medium-grained, well-to-moderately sorted sandstone with planar lamination, symmetrical ripples and trough cross-bedding. Lamination may be defined by carbonaceous drapes or coarse quartz sand grains</td>
<td>BI=0-3 (Op, Au, Sk, An, rare Th, Ros, Cy)</td>
<td>Upper shoreface</td>
</tr>
<tr>
<td></td>
<td>Fining-upward, bioturbated sandstone</td>
<td>Upward-fining, erosionally based, poorly sorted sandstone with a basal surface lined by coarse quartz sand or carbonaceous grains</td>
<td>BI = 0-3 (Au, Pl, Sk, Op, Th)</td>
<td>Upper shoreface (tidal channel?)</td>
</tr>
<tr>
<td>Coastal plain</td>
<td>Rooted sandstone</td>
<td>Fine-to medium-grained, thin (&lt; 2m), moderately sorted, carbonaceous sandstone with root penetrations</td>
<td>BI = 1</td>
<td>Root penetrated foreshore and coastal plain</td>
</tr>
<tr>
<td></td>
<td>Coal</td>
<td>Coal which grades in parts to a carbonaceous siltstone</td>
<td>BI = 0</td>
<td>Mire</td>
</tr>
<tr>
<td></td>
<td>Interbedded, cross-stratified sandstone</td>
<td>Coarse-to fine-grained, well- to poorly-sorted, erosionally based, carbonaceous sandstone. Lack bioturbation and contain asymmetrical ripples, planar-low angle lamination, tabular and trough cross-bedding</td>
<td>BI = 0</td>
<td>Distributary channel</td>
</tr>
<tr>
<td></td>
<td>Bioturbated, organic-rich siltstones</td>
<td>Organic-rich, carbonaceous siltstone with fine-grained sandstone laminations, ripples and synaeresis cracks</td>
<td>BI = 0-3 (Pl, rare Ros, Pa, Ph, Cy)</td>
<td>Lagoon</td>
</tr>
<tr>
<td></td>
<td>Barren organic-rich siltstone</td>
<td>Black, fissile, organic rich silty mudstone</td>
<td>BI = 0</td>
<td>Anoxic lagoon or lake</td>
</tr>
</tbody>
</table>

Table 3.3: Summary of facies and facies association of the J20-J40 stratigraphic interval in the Bryne and Sandnes formations. Bioturbation fabric is described using the Bioturbation Index (BI) scheme of Taylor and Goldring (1993). Te, Teichichnus; Sk, Skolithos; An, Anconichnus; Pa, Palaeophycus; Op, Ophimorpha; Ros, Rosellia; Pl, Planolites; Tr, Terebellina; Tri, Trichinus; Th, Thalassinoides; Au, Aulichnites; Cy Cylindrichnus; Ch, Chondrites; Ph, Phycosiphon.
<table>
<thead>
<tr>
<th>Basin name</th>
<th>Basin type</th>
<th>Stratigraphic unit</th>
<th>Age (Duration/Myr)</th>
<th>Sediment provenance</th>
<th>Structural style at deposition</th>
<th><em>1</em> Thickness (m)</th>
<th><em>2</em> Net: Gross (%)</th>
<th><em>3</em> Expansion index</th>
<th><em>4</em> Accumulation rate (m/Myr)</th>
<th>Depositional process</th>
<th><em>5</em> Depositional dip extent of shallow marine sandstone (km)</th>
<th>Shoreline orientation relative to structural grain</th>
<th>References</th>
</tr>
</thead>
<tbody>
<tr>
<td>Egersund Basin, North Sea, Norway</td>
<td>salt-influenced active rift</td>
<td>Sandnes Formation</td>
<td>Callovian - Oxfordian (5.0)</td>
<td>extra-basinal</td>
<td>Xi-salt anticline; Alpha salt wall</td>
<td>16</td>
<td>31</td>
<td>90</td>
<td>80</td>
<td>1.9</td>
<td>6</td>
<td>wave-dominated shallow marine with minor tidal influence</td>
<td>70</td>
</tr>
<tr>
<td>Bombaral - Alcobaca sub-basin, West Iberia</td>
<td>salt-influenced active rift</td>
<td>Alcobaca Formation (A2 - A3 units)</td>
<td>Oxfordian - Kimmeridgian (2.6)</td>
<td>extra-basinal</td>
<td>Bolhos-Veneiro salt pillow; Montijo-Antonio salt pillow</td>
<td>275</td>
<td>650</td>
<td>65</td>
<td>5</td>
<td>2.4</td>
<td>250</td>
<td>current and wave-dominated shallow marine; reeval carbonates; turbidites</td>
<td>40</td>
</tr>
<tr>
<td>South Viking Graben, North Sea, Norway</td>
<td>salt-influenced active rift</td>
<td>Hugin Formation</td>
<td>Callovian (3.5)</td>
<td>extra-basinal</td>
<td>salt-cored high</td>
<td>50</td>
<td>75</td>
<td>100</td>
<td>25</td>
<td>3.0</td>
<td>21</td>
<td>wave and tide-dominated shallow marine</td>
<td>&gt; 15</td>
</tr>
<tr>
<td>La Popa Basin, Mexico</td>
<td>salt-influenced foreland</td>
<td>Delgado Sandstone Member</td>
<td>Maestrichtian (3.9)</td>
<td>extra-basinal</td>
<td>El Gordo diapir; El Papalote diapir</td>
<td>10</td>
<td>30</td>
<td>100</td>
<td>25</td>
<td>3.0</td>
<td>8</td>
<td>wave-dominated shallow marine</td>
<td>20</td>
</tr>
<tr>
<td>Central Graben, North Sea, UK</td>
<td>salt-influenced active rift</td>
<td>Fulmar Formation (Unit 3)</td>
<td>Kimmeridgian (10)</td>
<td>intra-basinal</td>
<td>Fulmar Field normal fault</td>
<td>150</td>
<td>250</td>
<td>100</td>
<td>100</td>
<td>1.7</td>
<td>250</td>
<td>wave-dominated shallow marine</td>
<td>15</td>
</tr>
<tr>
<td>East Shetland Basin, North Sea, Norway</td>
<td>active rift</td>
<td>Tarbert Formation</td>
<td>Bapacoan - Bathonian (42)</td>
<td>extra-basinal</td>
<td>Statford East normal fault</td>
<td>0</td>
<td>70</td>
<td>0</td>
<td>80</td>
<td>-</td>
<td>17</td>
<td>wave-dominated shallow marine; fluval</td>
<td>7</td>
</tr>
<tr>
<td>Beryl Embayment, North Sea, UK</td>
<td>active rift</td>
<td>Beryl Formation (Unit 3)</td>
<td>Bathonian (3.0)</td>
<td>extra-basinal</td>
<td>B2 normal fault</td>
<td>75</td>
<td>120</td>
<td>100</td>
<td>95</td>
<td>1.6</td>
<td>40</td>
<td>tide-dominated shallow marine</td>
<td>10</td>
</tr>
<tr>
<td>Crotone Basin, southern Italy</td>
<td>active rift</td>
<td>Zinga (Unit 2 and 3)</td>
<td>Pliocene (1.5)</td>
<td>extra-basinal</td>
<td>Zinga normal fault; Montagna - Piasa normal fault</td>
<td>55</td>
<td>185</td>
<td>55</td>
<td>333</td>
<td>3.4</td>
<td>123</td>
<td>wave-dominated and tidal influenced shallow marine; lagoonal</td>
<td>3 - 5</td>
</tr>
<tr>
<td>Gulf of Suez Rift, Egypt</td>
<td>active rift</td>
<td>Nuukhul Formation</td>
<td>Miocene (14)</td>
<td>extra-basinal</td>
<td>East Tanka normal fault</td>
<td>0</td>
<td>45</td>
<td>0</td>
<td>80</td>
<td>-</td>
<td>32</td>
<td>tidally influenced shallow marine</td>
<td>3</td>
</tr>
<tr>
<td>Cretaceous Interior Seaway, New Mexico</td>
<td>foreland</td>
<td>Huapa Tongue Sandstone</td>
<td>Santonian (0.5)</td>
<td>extra-basinal</td>
<td>passive</td>
<td>&lt; 50</td>
<td>-</td>
<td>-</td>
<td>100</td>
<td>wave-dominated shallow marine with tidal influence</td>
<td>9 - 11</td>
<td>-</td>
<td></td>
</tr>
<tr>
<td>Cretaceous Interior Seaway, SW Colorado</td>
<td>foreland</td>
<td>Cliff House Formation</td>
<td>Campanian (2.0)</td>
<td>extra-basinal</td>
<td>passive</td>
<td>200</td>
<td>(4 individual sandstone tongues: 50 - 55 m)</td>
<td>-</td>
<td>-</td>
<td>100</td>
<td>wave-dominated shallow marine</td>
<td>10 - 20</td>
<td>-</td>
</tr>
</tbody>
</table>
Table 3.4: Parameters describing the influence of salt structures and normal faulting on net-transgressive, shallow-marine stratigraphy in salt-influenced and active rift basins.
CHAPTER 4

SHALLOW MARINE RESERVOIR DEVELOPMENT IN EXTENSIONAL DIAPIR COLLAPSE MINIBASINS: AN INTEGRATED SUBSURFACE CASE STUDY FROM THE UPPER JURASSIC OF THE COD TERRACE, NORWEGIAN NORTH SEA

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4.1. ABSTRACT

Understanding the development of accommodation above collapsing salt diapirs and its influence on reservoir distribution is critical in constraining exploration risk in salt provinces. We present a subsurface case study that uses integrated 3D and 2D seismic reflection, wireline-log, core and biostratigraphic data from the Upper Jurassic of the Cod Terrace, Norwegian North Sea, where we interpret rifting to have resulted in the rise and initial-stage collapse of salt diapirs, forming minibasins above former diapir crests. These became sites for diachronous deposition and preservation of up to 500 m thick, net-transgressive shallow-marine sandstone reservoirs. Maximum thickness is recorded in the axis of minibasins with a reduction in thickness of up to 65% noted on their flanks. The stratigraphic architecture of individual minibasins is variable. Flooding surfaces define depositional units with lateral onlap geometries. Proximal-to-distal facies variations from shoreface to offshore shelf occur over scales larger than individual minibasins, contain large sand volumes, and are not confined to areas of localized sandstone subcrop, which in combination suggest that the minibasins formed a linked network supplied by regional sediment-routing systems. The results of this study provide a new tectono-stratigraphic model for prediction of reservoir presence, thickness and continuity in diapir-collapse minibasins along salt walls in the Central North Sea, and in less mature, data-poor basins where reservoirs have been identified in depocenters above salt walls.

4.2 INTRODUCTION

Numerous studies have shown that understanding the interplay between salt movement, accommodation development and sediment dispersal is key to reducing exploration risk in salt-influenced sedimentary basins (e.g. Hodgson et al., 1992; Prather et al., 1998; Booth et al., 2000; Smith, 2004; Jackson et al., 2011; Kane et al., 2012). The majority of these
studies are concerned with the way in which salt flow controls depocenter development, accommodation creation and reservoir deposition between salt structures, principally in deep-water sedimentary environments (e.g. Prather et al., 1998; Booth et al., 2000; Smith, 2004; Kane et al., 2012), although some studies have considered the impact of salt-related tectonics on shallow-marine reservoir development on the flanks of salt structures (e.g. Giles and Lawton 2002; Aschoff and Giles 2005). Even fewer studies have focused explicitly on reservoir development in depocenters created above salt structures, despite the fact that these reservoirs and depocenters may form important components of successive petroleum systems on salt-influenced passive margins (e.g. Bay Marchand Field, Gulf of Mexico; Abriel et al., 1991).

Salt dissolution can result in the formation of supra-diapir depocenters by subterranean undermining of the diapir roof (e.g. Ge and Jackson 1998; Clark et al., 1999; Stewart and Clark 1999; Jackson et al., 2010). Alternatively, experimental models and restorations of natural examples indicate that regional extension can result in widening and collapse of pre-existing diapirs, and the formation of depocenters termed minibasins (Figure 4.1A) (sensu Vendeville and Jackson 1992b; Hudec et al., 2009). During diapir-collapse, the thickest syn-kinematic stratigraphic section is developed in the axis of the minibasin above the former diapir crest, and these strata thin onto, and is represented by a condensed section on, the flanks of the minibasin (Figure 4.1A). Even though seismic-scale thickness variations within diapir collapse-related minibasins have been documented (Abriel et al., 1991; Stewart and Argent, 2000), there is a paucity of case studies that document the detailed relationships between diapir collapse, depositional style, and reservoir geometry and continuity. Hence our understanding of the risks associated with reservoir presence and thickness in such minibasins is critical for exploration.
In the Central North Sea, one of the most prolific salt-influenced sedimentary basins in the world, three main tectono-stratigraphic models, described later, have been proposed to account for the development and distribution of Jurassic reservoir in supra-diapir depocenters: (i) the ‘pod–interpod’ model (Hodgson et al., 1992); (ii) the ‘rift-raft’ model (Penge et al., 1993, 1999); and (iii) the ‘salt-dissolution’ model (Clark et al., 1999) (Figure 4.1B-D). These models, which are described in detail in the following section, were based either on poor-quality, commonly two-dimensional seismic reflection data that offered limited constraint on the relatively subtle structural and stratigraphic relationships that underpin the related models, or on borehole datasets that provide limited three-dimensional control on the stratigraphic relationships in the supra-diapir depocenters.

This study focuses on the Central North Sea, where Upper Jurassic, net-transgressive, shallow-marine reservoirs were deposited in supra-diapir depocenters (e.g. Wakefield et al., 1993; Stewart and Clark, 1999; Spathopoulos et al., 2000; Stewart 2007) (Figure 4.2A). Here, the Upper Jurassic play is one of the most prolific in the basin and is host to over 20 hydrocarbon fields (e.g. Fulmar, Kittiwake, Ula, Angus, Fife) with estimated recoverable oil reserves in excess of 2 billion bbls (Eriksen et al., 2003) (Figure 4.2B). The study area is located on the Cod Terrace in the Norwegian Central Graben, where an integrated dataset of regional seismic reflection volumes, combined with wireline-log, core, provenance and biostratigraphic data, permit mapping of the structural framework, and detailed sedimentologic and stratigraphic analysis of the Upper Jurassic reservoir interval.

The aims of this study are threefold: (1) to understand the controls on supra-diapir depocenters; (2) to document the stratigraphic architecture, thickness, geometry and continuity of reservoirs within these minibasins; and (3) to propose a tectono-stratigraphic model that explains reservoir architecture on the Cod Terrace, and in comparable fields in
the Central North Sea and elsewhere. We hope that this study will help exploration geoscientists evaluate overlooked exploration opportunities in mature provinces such as the North Sea and that the generic aspects of this work can act as a guide for exploration in basins where supra-diapir depocenters have been interpreted but data are sparse (e.g. eastern Mediterranean; Loncke et al., 2006).

4.3 TECTONO-STRATIGRAPHIC MODELS FOR UPPER JURASSIC RESERVOIR DEVELOPMENT IN THE CENTRAL NORTH SEA

Since the discovery of the Fulmar Field in 1975, three principal models have been proposed to explain the Late Jurassic tectono-stratigraphic evolution of the Central North Sea in general, and the deposition of shallow-marine, Upper Jurassic sandstone reservoirs in particular (Figures 4.1 and 4.2B). These models attempt to explain how highly bioturbated, shallow-marine reservoirs are deposited in morphologically complex depocenters above salt diapirs. First, the “pod–interpod” model proposes that, during the Early Triassic, extension of sub-salt basement and differential loading of Triassic sediments resulted in passive diapirism and salt dissolution (Figure 1B) (Hodgson et al., 1992). Continued differential loading of the Zechstein salt by Triassic sediments led to deepening of the Triassic depocenters or ‘pods’ and growth of the flanking diapirs, until mobile salt was completely evacuated from below the ‘pods’ and salt supply to the diapirs was exhausted. Subsequent Jurassic extension resulted in diapir widening and salt withdrawal, which provided accommodation for Jurassic shallow marine reservoirs in depocenters or ‘interpods’. Second, the “rift-raft” model envisages that regional extension of cover (i.e. Triassic) and basement (i.e. sub-salt) rocks, and/or local gravity-gliding of the cover alone during the Jurassic resulted in the formation of intact Triassic ‘rafts’ separated by supra-salt ‘rifts’, in which shallow marine deposition occurred (Figure 4.1C)
(Penge et al., 1993, 1999). Most salt movement is interpreted to have been reactive and to have occurred during the Jurassic in response to regional extension, which was accommodated in downdip locations towards the basin center by squeezing of pre-existing salt structures and folding of intervening minibasins. Finally, the “salt-dissolution” model proposes that Early Triassic karstification of the Zechstein evaporites resulted in the formation of an array of sub-circular collapse features (Figure 4.1D) (Clark et al., 1999). Differential subsidence of Triassic sediments into the Zechstein salt, which may have been augmented by thin-skinned extension, resulted in the formation of Triassic depocenters. Salt structures were then preferentially eroded during basin exhumation in the Late Triassic-to-Late Jurassic, mainly during Early Jurassic formation of the Mid-North Sea Dome.

Elements of all three models have been included in previous tectono-stratigraphic models of the Cod Terrace area (Bjørnseth and Gluyas 1995). Furthermore, several studies have suggested that Late Cretaceous-to-Miocene transpression occurred along the Cod Terrace, and that this resulted in the formation of a series of complex, inversion-related structural styles, the magnitude and areal distribution of which were controlled by the degree of coupling between sub- and supra-salt deformation (Brown et al., 1992; Sears et al., 1993; Stewart 1993).

4.4 REGIONAL GEOLOGIC SETTING

The Central North Sea is a broadly symmetric graben located along the southern arm of the North Sea trilette rift system (Figure 4.2). Permian-to-Triassic and Jurassic-to-Early Cretaceous extension, followed by thermal cooling and subsidence, are responsible for the present-day structure of the basin (Zanella and Coward 2003). The study area is located on the Cod Terrace, in the Norwegian sector of the Central North Sea (Figure 4.2B), in
present-day water depths of c. 70 m. It is host to the Ula, Gyda and Tambar fields, which together contain c. 940 mm bbls of recoverable reserves (Figure 4.2B; Norwegian Petroleum Directorate, 2013). The Cod Terrace is located to the west of the Sørvestlandet High, and separates the Central Graben from the Norwegian-Danish Basin. The Cod Terrace merges southwards with the Ringkøbing-Fyn High (Figure 4.2B).

Early Permian, non-marine, syn-rift clastics of the Rotliegend Group were deposited in a laterally extensive, long-lived, intermontane trough located to the north of the Caledonian-age, Mid-North Sea High (Figures 4.2B and 4.3) (Ziegler, 1990). Post-rift thermal subsidence and a relatively restricted seaway resulted in the formation of a giant saline lake and deposition of a thick succession of evaporites (Zechstein Supergroup) (Figure 4.3). Subsequent deposition of Triassic continental clastics of the Smith Bank and Skagerrak formations was strongly influenced by halokinesis (Figure 4.2C) (McKie and Williams, 2009). Uplift of the Mid-North Sea Dome occurred during the Early Jurassic and, as a result of this uplift, Early-to-Middle Jurassic strata are absent across much of the Central Graben below the Mid-Cimmerian unconformity (Figure 4.3) (Underhill and Partington, 1993). Extension and halokinesis in the Late Jurassic resulted in relatively rapid subsidence and the development of fault- and salt-bounded depocenters that were progressively flooded as a result, at least in part, of rift-related subsidence. The net-transgressive, shallow-marine systems that were deposited at this time now form the main reservoirs of the Ula and Tambar-Gyda fields (Figure 4.2C) (e.g. Wakefield et al., 1993; Stewart, 2007). Late Cretaceous-to-Tertiary shortening caused by the Alpine orogeny and the opening of the North Atlantic, resulted in inversion and reverse reactivation of basin-bounding normal faults (e.g. Ziegler, 1990; Gowers et al., 1993), and squeezing and rejuvenation of pre-existing diapirs (e.g. Davison et al., 2000; Stewart, 2007).
4.5 DATASET

In this study we used four 3D seismic reflection surveys that cover a total area of 3,500 km² and which encompasses the Cod Terrace, the northern fringe of the Steinbit Terrace and the Hidra High (Figures 4.2 and 4.4). These post-stack, time-migrated surveys were acquired during 1992-1999 and reprocessed post-2000. The seismic cross-sections shown in this paper are displayed with SEG normal polarity (i.e. a downward increase in acoustic impedance is represented by a positive or red reflection, and a downward decrease in acoustic impedance is represented by a negative or black reflection). The surveys have record lengths of 6000 milliseconds two-way travel time (ms TWT), and E-W-trending inlines are spaced at 25.0 m and N-S-trending crosslines at 12.5 m. Based on a dominant seismic frequency of 25 Hz and an interval velocity of 2642 m s⁻¹ (obtained from checkshot data in boreholes), we calculate that the vertical resolution within the Upper Jurassic interval of interest is c. 26 m. A time-migrated 2D seismic survey that covers the Sørvestlandet High was also used to provide a regional framework for the detailed analysis presented herein (Figure 4.4). This survey was acquired in 1997, and has a vertical sample interval of 4 ms and seismic resolution of c. 50 m, based on a dominant seismic frequency of 18 Hz and an interval velocity of 3585 m s⁻¹.

We integrated our seismic reflection data with well data from 35 wells that contain original well reports and lithological cutting descriptions (Figure 4.4). Thirty-four of these wells have conventional wireline-log suites (e.g. gamma ray, resistivity, neutron, density, and sonic logs), twenty eight have time-depth curves (thereby allowing stratigraphic data to be tied to seismic reflection data), and twelve contain biostratigraphic data that can be directly tied to the regional Jurassic North Sea stratigraphic scheme of Partington et al. (1993). Biostratigraphic data were directly obtained from the published studies of Partington et al. (1993), Bjørsneth and Gluyas (1995) and Nicholson et al. (2004) (Figure...
4.3). Eleven of the 35 wells contain a total of 660 m of core in the Upper Jurassic interval of interest (Figure 4). Sandstone mineralogical composition data are also available for two wells (1/03-3 and 7/08-3; Figure 4.4).

### 4.6 METHODOLOGY

Five regional seismic horizons have been mapped to define the structural framework of the study area: (i) Top Rotliegend Group (258 Ma), which defines the sub-salt, mainly rift-related structure of the study area; (ii) Top Zechstein Supergroup (251 Ma), which defines the present-day thickness and distribution of the salt, and related structures; (iii) Top Mandal Formation (129 Ma), which marks the top of the Jurassic reservoir interval and highlights the pattern of deformation in the supra-salt interval; (iv) Top Cromer Knoll Group (99.6 Ma), which is an intra-post-rift seismic horizon that defines the cessation of supra-diapir depocenter subsidence; and (v) near-Top Shetland Group (c. 65 Ma), which marks the top of the unit that records the Late Cretaceous inversion event (Figure 4.3).

Rapid changes in stratal thickness and reflection amplitude, and poor reflection continuity preclude mapping of parasequence-scale stratigraphic units in Late Jurassic strata in the seismic reflectivity data (Figure 4.5A). The cosine of instantaneous phase seismic attribute volume, also called normalized amplitude, displays the cosine of instantaneous phase angle between -1.0 to 1.0. This attribute does not contain any amplitude information, but it can help enhances the visual appearance of seismic events and assist the detailed mapping of reservoir-bearing intervals (Duff and Mason, 1989). In this study, cosine of instantaneous phase seismic attribute was used for detailed mapping within the Upper Jurassic reservoir interval. Time-depth data available for wells allowed biostratigraphic markers to be tied to the seismic data through the construction of synthetic seismograms (e.g. Figure 4.3).
Sedimentologic facies analysis of the Upper Jurassic succession was carried out through core logging at 1:250 scale. Wireline-log responses were calibrated to core descriptions (petrofacies characterisation) in order to qualitatively interpret lithology and facies in uncored intervals of the wells. Core- and wireline-log-based facies interpretations were combined with seismic mapping and biostratigraphic data (calibrated to the scheme of Partington et al., 1993) and then used to construct a sequence stratigraphic framework for the Upper Jurassic succession (e.g. Figure 4.3).

4.7 STRUCTURAL FRAMEWORK

We use time-structure maps of five key seismic horizons (Top Rotliegend, Top Zechstein, Top Mandal, Top Cromer Knoll and near Top Shetland) to describe the present-day regional structural framework of the study area. This provided a sub-regional, tectono-stratigraphic framework for our detailed analysis of the Jurassic depocenters. Three key structural levels and associated structures are described below: (i) sub-salt involved normal faults; (ii) top salt and associated salt structures; and (iii) supra-salt restricted normal faults (Figure 4.5A-B).

4.7.1 Sub-salt involved normal faults

The Top Rotliegend time-structure map defines a series of NNW-SSE- and NE-SW-striking fault systems that dip 40-70°, predominantly westwards towards the axis of the Central Graben, and have throws of 100–600 ms (c. 250–1500 m, assuming an interval assuming an average interval velocity of 5000 m s⁻¹) (Figures 4.5 and 4.6). Many of these faults offset the base-salt surface but do not fully penetrate the salt (i.e. they are sub-salt restricted) (Figure 4.5). Normal faults of this type define the margins of the Sørvestlandet and Hidra highs, the Steinbit and Cod terraces and the Breiflabb Basin (Figure 4.6A). The
main structural element of interest to this study, the Cod Terrace, is a NW-SE-oriented, normal fault-bounded terrace that is located on the westerly, downthrown margin of the Sørvestlandet High and the easterly, upthrown side of the Brieflabb Basin. The Cod Terrace reaches a maximum width of c. 12 km and it narrows southwards towards the Hidra High and Steinbit Terrace (Figures 4.2B and 4.6A). The age of these sub-salt involved faults is difficult to determine; it is likely that they experienced movement during the main Late Jurassic rift event, but they could conceivably have accrued some displacement during the earlier Permo-Triassic rift event (Hodgson et al., 1992; Erratt 1993).

4.7.2 Salt
The Top Zechstein Supergroup time-structure map illustrates the morphology and distribution of salt structures (Figure 4.6B). Salt walls are the most common structures, and they are 2–10 km wide and display up to 3.5 km of relief (assuming an average interval velocity of 3500 m s\(^{-1}\)). On the Cod Terrace, salt walls are orientated NW-SE and they are typically located above sub-salt faults (Figure 4.6A, B). On the Sørvestlandet High, salt walls do not display a preferential strike but form a polygonal pattern (Figure 4.6B). Diapirs are locally developed along salt walls, and they are up to 3.5 km wide and display up to 4.5 km of relief (Figure 4.6B). The Zechstein Supergroup is thin between these salt structures and apparent salt welds (\textit{sensu} Wagner and Jackson, 2011) are present below subcircular-to-elongated, Triassic depocenters that are up to 3.5 km thick and 10 km wide (assuming a Triassic interval velocity of 3500 m s\(^{-1}\)) (Figure 4.6B). Although seismic imaging of Triassic strata in the salt-bounded depocenters is typically quite poor, areas of good imaging indicate that intra-Triassic stratal packages thicken towards the edges of salt walls (Figure 4.5).
4.7.3 Supra-salt restricted normal faults

The Top Mandal time-structure map illustrates the supra-salt structural framework of the study area, which is characterised by supra-salt restricted (i.e. thin-skinned) normal faults (Figure 4.5B). Two main types of supra-salt restricted faults are observed: (i) listric faults that define the edges of salt walls and decrease in dip downwards onto the top of the salt; and (ii) oppositely dipping pairs of planar faults that are developed immediately above and offset the upper part of the Triassic depocenters and the edges of Jurassic depocenters (Figures 4.5 and 4.6C). Both types of fault have throws of 10-400 ms (c. 15-530 m, assuming an interval assuming an average interval velocity of 2642 m s\(^{-1}\)), dip 45-75°, and define graben or half-graben style depocenters. The listric faults that strike parallel to the underlying salt walls and bound the Triassic depocenters are up to 10 km long.

At shallow (i.e. Cretaceous-to-Tertiary) structural levels a series of anticlines with an areal extent between 10-22 km\(^2\) are developed above the salt diapirs in which stratigraphic thinning is noted (Figure 4.5C). One such anticline defines the structural trap of the Ula field, and they are interpreted to have formed in response to post-early Cretaceous inversion and diapir rejuvenation (e.g. Davison et al., 2000).

4.8 SEDIMENTOLOGY OF JURASSIC MINIBASINS

The sedimentology of Upper Jurassic shallow marine strata in the Central North Sea has been extensively studied and documented by various authors (e.g. Fulmar Formation, Fife/Angus sandstones, Ula Formation, Heno Formation; see Johnson et al., 1986; Taylor and Gawthorpe, 1993; Wakefield et al., 1993; Gowland 1996; Martin and Pollard 1996; Spathopoulos et al., 2000; Johannessen et al., 2010). The facies and trace-fossil assemblage schemes of Taylor and Gawthorpe (1993) and Gowland (1996) in the Fulmar and Ula formations were used as a guide for interpretation of Upper Jurassic strata in the
study area. We identify six facies associations characterised by variations in lithology, grain size, degree of sorting, nature of bedding contacts, primary sedimentary structures, bioturbation intensity (described using the bioturbation index scheme of Taylor and Goldring, 1993) and trace fossil assemblages: (FA 1) bioturbated offshore shale; (FA 2) laminated offshore shale; (FA 3) offshore transition silty sandstone; (FA 4) bioturbated lower shoreface sandstone; (FA 5) cross-laminated lower shoreface sandstone; and (FA 6) cross-stratified upper shoreface sandstone (Figure 4.7).

### 4.8.1 Facies association 1 (FA1): bioturbated offshore shale

**Description**

This facies association is cored in wells on the Cod Terrace and the Sørvestlandet High. FA1 comprises silty claystone and argillaceous siltstone (Figure 4.7A). Silty claystones are calcareous, bioturbated (BI = 3-5), and contain relict patches of planar lamination. Belemnite, bivalve and shell fragments represent the main bioclastic content in FA1, and the trace fossil assemblage is composed of *Anconichnus horizontalis* and *Chondrites*, rare *Planolites*, *Terebellina* and *Teichichnus rectus*. Argillaceous siltstones are highly bioturbated (BI = 5-6) by a low-diversity trace fossil assemblage of *Anconichnus horizontalis* and *Chondrites*, with rare *Planolites* and belemnite fragments. FA1 commonly sharply overlies offshore transition silty sandstone (FA3) or bioturbated lower shoreface sandstone (FA4). It is overlain either gradationally by offshore transition silty sandstone (FA3), or abruptly by laminated offshore shale (FA2) and cross-laminated lower shoreface sandstone (FA5) (Figure 4.8A, B).
Interpretation

The very fine grained and intensely bioturbated character of FA1 suggests low rates of sediment accumulation, predominantly by suspension settling in a low-energy environment. In the Upper Jurassic Fulmar Formation of the UK Central Graben, similar bioturbated shales with Anconichnus and Chondrites trace fossil assemblages are interpreted as offshore to offshore transition sediments deposited below mean storm wave base (Gowland 1996; Martin & Pollard 1996); we apply a similar interpretation here to FA1.

4.8.2 Facies association 2 (FA2): laminated offshore shale

Description

Facies association 2 (FA2) is cored in well 2/1-8 in the Gyda Field and it passes abruptly upwards into the overlying offshore transition silty sandstone (FA3) (Figure 4.7B). FA2 consists of inter-laminated claystone and siltstone, which contain lenses of very fine-grained sandstone (Figure 4.7B). Siltstone laminations preserve low-amplitude ripple forms. Bioturbation is common (BI = 4) in FA2, and the trace fossil assemblage is dominated by Anconichnus horizontalis and Chondrites, with rare Teichichnus rectus and Terebellina.

Interpretation

The generally fine-grained, interbedded character of FA2 suggests suspension settling of mudstone with episodic deposition of siltstone and very fine-grained sandstone by unidirectional or oscillatory currents. The presence of slightly coarser laminae and lower levels of bioturbation suggest that FA2 was deposited in a higher-energy depositional environment than FA1. Similar deposits in the Upper Jurassic Fulmar Formation in the UK
Central Graben have been attributed to an offshore shelf setting and we prefer this interpretation for FA2 (Gowland 1996).

### 4.8.3 Facies association 3 (FA3): offshore transition silty sandstone

**Description**

This facies association is cored in wells on the Cod Terrace and the Sørvestlandet High. It is commonly found either overlying or underlying bioturbated and laminated offshore shales (FA1, FA2), and bioturbated and cross-stratified lower shoreface sandstones (FA3, FA4) (Figure 4.8A, B, D). FA3 comprises silty, well-sorted, subrounded-to-rounded, very-fine grained sandstone (Figure 4.7C). Sandstones are highly bioturbated (BI = 5-6) by a diverse trace fossil assemblage dominated by *Teichichnus zigzag*, *Teichichnus rectus*, *Anconichnus horizontalis*, with occasional *Planolites*, *Palaeophycus*, *Terebellina*, *Rosselia* and *Chondrites*. Preserved primary structures are rare, but low-angle and hummocky cross-stratification and high angle laminations are observed. Carbonaceous debris is common in FA3, and shell fragments and belemnites are present.

**Interpretation**

The fine grain size and abundant bioturbation that characterizes FA3 suggests a low-energy environment of deposition. Hummocky cross-stratified beds indicate episodic deposition by storm-generated oscillatory or combined flows below fairweather wave base (Dott and Bourgeois, 1982). The trace fossil assemblage is characteristic of the *Teichichnus-Anconichnus* and *Teichichnus* ichnofabric described by Taylor and Gawthorpe (1993), which is the product of repeated burrowing episodes by an infaunal deposit-feeding community during fairweather periods in an offshore-to-offshore transition
setting. FA3 is interpreted as deposits of the offshore-transition zone, in water depths between mean storm and fairweather wave base (cf. Gowland, 1996).

### 4.8.4 Facies association 4 (FA4): bioturbated lower shoreface sandstone

**Description**

This is the most common facies association encountered in the cored wells in the study area, and is commonly associated with bioturbated offshore shale, offshore transition silty sandstone, cross-laminated lower shoreface sandstone and cross-stratified upper shoreface sandstone deposits (FA1, FA3, FA5 and FA6, respectively). FA4 consists of bioturbated (BI = 3-6), well-to-moderately sorted, very fine-grained sandstone that contains intervals of cross-laminated, medium-grained sandstone and medium-to-coarse grained, moderate-to-poorly sorted sandstone. The trace fossil assemblage is generally diverse, consisting of *Thalassinoides, Ophimorpha nodosa, Teichichnus zigzag, Teichichnus rectus, Planolites*, and rare *Palaeophycus and Skolithos*, (e.g. Figure 4.7D). Intervals of medium-to-coarse grained, poorly sorted sandstone contain *Siphonichnus* burrows that overprint the background trace fossil assemblage. Carbonaceous debris, coarse quartz sand grains, shell fragments, crinoids, mudclasts and rare belemnites also occur in FA4.

**Interpretation**

The high sandstone content and trace fossil assemblages in FA4 have been interpreted as characteristic of lower shoreface deposits in the Ula Formation (Taylor and Gawthorpe, 1993). Variability in sandstone sorting reflects the degree of reworking, with the well-to-moderately sorted sandstones interpreted to have been reworked extensively by fairweather waves prior to offshore transport to the lower shoreface, and moderate-to-poorly sorted sandstones to have been transported short distances from nearby fluvial
systems (Gowland 1996). The trace fossil assemblage is characteristic of the *Thalassinoides* and *Siphonichnus* ichnofabrics described by Taylor and Gawthorpe (1993), for a lower shoreface setting.

4.8.5 **Facies association 5 (FA5): cross-laminated lower shoreface sandstone**

*Description*

This facies association is present only in wells south of the Cod Terrace, on the Hidra High and Steinbit Terrace (wells 1/03-8, 2/02-3 and 2/02-1), and abruptly overlies and underlies bioturbated offshore shales and bioturbated lower shoreface sandstones (FA1, FA4). FA5 consists of well-sorted, fine-grained sandstone beds (<1 m thick) that contain low-angle and hummocky cross-stratification, symmetrical ripples and slump folding (Figures 4.7E, 4.8C). Bed tops are weakly-to-moderately bioturbated (BI ≤ 4) by *Planolites* and *Palaeophycus*, with rare occurrences of *Anconichnus horizontalis* and *Chondrites*. Locally, sandstones appear structureless within three feet cored intervals.

*Interpretation*

Hummocky cross-stratified beds were deposited by large oscillatory or combined flows during storm events (Dott and Bourgeois, 1982). Bioturbation at bed tops records colonisation during quiescent conditions between storm events, below fairweather wave base. Where sandstones appear structureless it may reflect biogenic homogenisation of previously stratified sandstones rather than high-density gravity flows in which sharp or erosional bases are expected. The facies association records deposition between mean storm wave base and fairweather wave base, in a lower shoreface environment that had higher ambient energy than the intensely bioturbated sandstones of FA4.
4.8.6 Facies association 6 (FA6): cross-stratified upper shoreface sandstone

Description

FA6 is cored in well 7/12-6 in the Ula Field, where it abruptly overlies and underlies bioturbated lower shoreface sandstone deposits (FA4) (Figure 4.8D). FA6 consists of sparsely bioturbated (BI = 2-3), fine-grained, well-sorted sandstone that contains coarse quartz and carbonaceous grains, and shell fragments. Bioturbation is dominated by *Ophiomorpha nodosa*, with rare *Planolites* and *Palaeophycus* (Figure 4.7F). Primary sedimentary structures include tabular cross-bedding and low-angle cross-stratification.

Interpretation

The fine-grained, well-sorted character of the sandstones implies extensive reworking, consistent with deposition above fairweather wave base in a high-energy setting. Tabular cross-bedding records the migration of dunes due to unidirectional currents, probably due to longshore currents. The relatively low intensity and diversity of bioturbation reflects high energy conditions that are typical of upper shoreface environments in the Ula Formation (Taylor and Gawthorpe, 1993).

4.8.7 Wireline-log character

Due to their distinct sedimentologic characteristics, in particular the sedimentary and biogenic structures they contain, it is relatively simple to discriminate between facies associations in core. However, where core is lacking, it is difficult to distinguish facies associations within, for example, sandstone-rich, shallow marine sandstones, using wireline log data alone. We therefore interpret three groups of facies associations in wireline logs: offshore (FA1, FA2); offshore transition (FA3); and shoreface (FA4-FA6).
Offshore facies associations (FA1 and FA2) are represented by a serrate gamma-ray log signature and are characterised by overall absolute high gamma-ray values (> 80 API), which allows them to be distinguished from offshore transition (FA3) and shoreface (FA4-FA6) deposits (Figure 4.8A, B). FA1 and FA2 are characterized by a relatively wide range of neutron (10-44 v/v) and density values (2.37-2.67 g cm\(^{-3}\)), and occur in successions 10-130 m thick.

FA3 is characterized by an overall blocky or funnel-shaped, ‘upward-cleaning’ gamma-ray log signature, which form successions 5-40 m thick with absolute values > 55 API (Figure 4.8A, B, D). FA3 is thus characterised by absolute gamma-ray values that are generally higher than FA4-FA6 and lower than FA1-A2. FA3 is represented by restricted neutron values (10-27 v/v) and a wide spread of density values (2.25-2.67 g cm\(^{-3}\)).

FA4, FA5 and FA6 are characterised by an overall blocky gamma-ray signature and relatively low absolute gamma-ray ( < 85 API), density (2.22-2.66 g cm\(^{-3}\)) and neutron (4-25 v/v) values. These facies associations form successions 5-200 m thick, onto which thin (<5 m), coarsening-upward trends are superimposed in some wells (Figure 4.8A, C, D).

### 4.8.8 Depositional models

Previous sedimentologic studies of Upper Jurassic strata in the Central North Sea have proposed three depositional models for the shallow-marine sandstones: (1) high-energy, storm-influenced shoreface; (2) low-energy, bioturbated shoreface; and (3) low-energy, bioturbated shelf (Johnson et al., 1986; Taylor and Gawthorpe, 1993; Gowland 1996; Martin and Pollard 1996; Johannessen et al., 2010). Four of the facies associations described herein (FA1, FA3, FA4 and FA6) can be placed within the low-energy, bioturbated shoreface model of Gowland (1996) (Figure 4.9A), and two (FA2, FA5) within the high-energy, storm-influenced shoreface model of Gowland (1996) (Figure
4.9B). High-energy, storm-influenced shorefaces may have been fronted by deeper water, which reduced frictional damping of storm waves, or along stretches of the shoreline that were more open to direct wave approach than low-energy, bioturbated shorefaces. As we describe later, deposits of the two shoreface types occur in different geographic locations and at different stratigraphic intervals.

4.9 SEQUENCE-STRATIGRAPHIC FRAMEWORK AND ARCHITECTURE OF JURASSIC MINIBASINS

A sequence stratigraphic framework for the net-transgressive, Upper Jurassic shallow-marine succession has been constructed using the procedure outlined below. First, seismic mapping of the top-Zechstein and top-Mandal horizons in the context of the regional structural framework, as described earlier, allowed the overall distribution and the geometry of Upper Jurassic minibasins to be constrained. Second, further reflection events within the Triassic and Jurassic intervals were mapped on every inline and crossline using the cosine of instantaneous phase seismic attribute volume. Third, several of these locally mapped seismic horizons were tied to wells using synthetic seismograms (e.g. Figure 4.3). Published biostratigraphic interpretations of these wells enabled identification of five basin-wide, Late Jurassic maximum flooding surfaces (J32?, J62, J64, J66 and J71 surfaces of Partington et al., 1993), which correspond to the locally mapped seismic horizons. The J50-J70 interval, which is the focus of this study, represents a period of c. 21 Myr (Partington et al., 1993). Biostratigraphically calibrated seismic mapping demonstrates that Jurassic minibasins occur only above salt walls, and that the Ula Field reservoir was deposited in a separate minibasin from the reservoir in the Tambar and Gyda fields (Figure 4.10). Sedimentologic detail is added to this chronostratigraphic framework using the core-and wireline log-based facies analysis described above. Facies associations are arranged
into upward shallowing successions that are bounded by surfaces of abrupt deepening (i.e. parasequences bounded by flooding surfaces; *sensu* Van Wagoner et al. 1990). Intra- and inter-field correlation of parasequences below seismic resolution is constrained by additional published biostratigraphic interpretations, calibrated to the scheme of Partington et al. (1993).

The sequence stratigraphic architecture and gross facies-association distributions in the study area are illustrated and described using three regional stratigraphic-correlation transects. The first transect is oriented NW-SE and crosses the Ula Field parallel to depositional dip; the second transect is oriented NW-SE and crosses the Tambar and Gyda fields parallel to depositional dip; and the third transect is oriented SW-NE perpendicular to depositional dip and crosses the Hidra and Sørvestlandet highs (Figure 4.11).

### 4.9.1 Transect 1: Ula Field

A flattened NW-SE-oriented, cosine-of-instantaneous-phase seismic cross-section across the Ula Field illustrates the pod-shaped geometry of the Upper Jurassic minibasin, its structural setting on top of a salt wall and the relatively thin (100 ms or 200 m thick) Triassic succession that underlies it (Figure 4.12A). The J71 and J64 maximum flooding surfaces were seismically mapped across the field, defining overall thinning of the entire Jurassic interval towards the margins of the Ula minibasin (Figure 4.12A). This observation is consistent with well data, which indicate that the J50-to-J70 succession is up to 280 m thick in the center of the Ula minibasin (well 7/12-2) and thins by up to 65% towards its flanks to 126 m (well 7/12-9) over a distance of c. 5 km (Figure 12B, C). Biostratigraphic data allow recognition and correlation of a two further, sub-seismic, Late Jurassic maximum flooding surfaces (J62, J72) within the minibasin (Figure 4.12B), which enables the J50-to-J70 succession to be divided into five stratigraphic units (J56, J63-J62,
J66-J64, J71, J76-J72), each of which corresponds to a parasequence (Figures 4.3 and 4.12B).

J56 (Early Kimmeridgian) is the oldest stratigraphic unit encountered within the Ula minibasin and it is bounded below by the top-Triassic horizon or Mid-Cimmerian Unconformity and above by the K10 (top/intra Ryazanian) maximum flooding surface. The J56 unit is broadly tabular and is 5-10 m thick in the center and SE of the minibasin, and it is absent due to inferred depositional onlap on the NW margin of the minibasin. J56 is composed entirely of offshore transition deposits (well 7/12-6) (Figures 4.8D, 4.12B).

The J63-J62 unit (Early Kimmeridgian) is bounded below and above by the J62 and J64 maximum flooding surfaces, respectively. In the center of the minibasin the unit is up to 125 m thick (well 7/12-2), where it represents the majority of the sandstone-dominated fill of the lower part of the Ula minibasin (Figure 4.12B). Core data from well 7/12-6 indicate that the lower part of the J63-J62 unit is composed of offshore transition (FA3) and sharp-based upper shoreface sandstones (FA6), which in turn are overlain by lower shoreface deposits (Figure 12B), implying that an additional, higher-order flooding surface is developed within the unit. The J63-J62 unit thins markedly onto the SE flank of the minibasin (30 m in well 7/12-9), although this thinning is not associated with a change in facies association. Likewise, thinning of this unit onto the NW flank of the minibasin (16 m in well 7/12-10; Figure 4.12B) is not associated with any discernible changes in facies association. However, we infer that the intra-J63-62 flooding surface onlaps the NW flank of the minibasin, and that the lowermost part of the J63-J62 unit is absent in this northernmost part of the minibasin (Figure 4.12B).

The J66-J64 unit (Middle-to-Late Kimmeridgian) is bounded below and above by the J64 and J71 maximum flooding surfaces, respectively. Its thickness varies from 15-30 m on the NW and SE flanks of the Ula minibasin, to 40 m just to the SE of the basin axis.
The J66-J64 unit is composed of offshore transition deposits across the width of the minibasin (Figure 4.12B).

The J71 unit (Early Portlandian) is bounded at its base by J71 maximum flooding surface, and at its top by J72 maximum flooding surface, and thins from 75 m in the axis of the Ula minibasin (well 7/12-2) to approximately 45 m on its SE and NW flanks (Figure 4.12B). Across most of the Ula minibasin, the J71 unit is composed of offshore shelf deposits, but its upper part passes laterally into a 20 m thick package of offshore transition deposits towards the SE flank of the minibasin (well 7/12-9) (Figure 4.12B).

The J76-J72 unit (Early Portlandian-to-Ryazanian) is dominated by offshore shelf deposits and documents final drowning of the Ula minibasin. It is bounded below and above by the J72 and K10 maximum flooding surfaces, respectively (Figure 4.12B). Thickness changes in the J76-J72 unit are relatively subdued; it is slightly thicker in the axis of the minibasin (40 m in well 7/12-2) than on its NW and SE margins (25 m in wells 7/12-10, 7/12-9) (Figure 4.12B).

4.9.2 Transect 2: Tambar and Gyda fields

A flattened NW-SE-oriented, cosine-of-instantaneous-phase oriented seismic cross-section along the axis of the Tambar-Gyda minibasin illustrates its pod-shaped geometry and the overall seismic-stratigraphic architecture of its fill (Figure 4.13A). At the base of the minibasin, Triassic strata are up to 150 ms (300 m) thick and separate Jurassic strata above from Zechstein Supergroup evaporites below. J32?, J62, J64 and J66A maximum flooding surfaces are recognised on seismic data and are mapped across the Tambar and Gyda fields. They are sub-parallel and bound apparently tabular bodies in the center of the minibasin, and they progressively onlap Triassic strata onto the NW and SE flanks of the minibasin. Three further maximum flooding surfaces (J56, J71, J72) are identified in wells
using previous biostratigraphic interpretations and these allow the J50-to-J70 net-
transgressive succession to be sub-divided into seven stratigraphic units (J54, J56, J63-J62,
J64, J66, J71, J76-J72), which each corresponds to a single parasequence (Figure 4.13B).
Well data indicate that the J50-to-J70 succession is up to 500 m thick in the axis of the
Tambar-Gyda minibasin (well 1/03-3) with a reduction in thickness of up to 64% on its
flanks (180 m, well 2/01-8), over a distance of c. 12 km (Figure 4.13B, C, D). Isochron
maps generated from seismic mapping also show abrupt thickening of the unit across a
NW-SE-striking, SW dipping normal fault developed at the NW end of the minibasin
(Figure 4.13D).

Cuttings descriptions from composite logs suggest that the top-Triassic horizon (Mid-
Cimmerian Unconformity) is overlain by a c.25- 50 m thick, Middle Jurassic (Early
Bathonian?) coastal-plain succession (wells 1/03-3, 2/01-6), which is capped by the J32
(Early Bathonian?) maximum flooding surface (Figure 4.13B). It is important to note that
this coastal-plain succession was absent from the Ula minibasin 10 km to the north, where
Upper Jurassic shallow marine deposits directly overlie Triassic deposits across the Mid-
Cimmerian Unconformity.

The J54 unit (Late Oxfordian) is bounded at its base by the J32 maximum flooding
surface and its top by the J56 maximum flooding surface (Figure 4.13B). It is the oldest
Upper Jurassic unit encountered in the Tambar-Gyda minibasin, and an age-equivalent unit
is not developed in the Ula minibasin. The J54 unit consists exclusively of shoreface
deposits, which thicken from the flanks of the Tambar-Gyda minibasin (e.g. 12 m in well
2/01-4) into its axis (e.g. 120 m in well 1/03-3) (Figure 4.13B).

The J56 maximum flooding surface marks an upward transition from shoreface (J54
unit) to offshore shelf deposits (J56 and J63-J62; Early Kimmeridgian), which occur
across the minibasin (Figure 4.13B). The J56 unit varies slightly in thickness from 8 to 20
m, whereas the J63-J62 unit is notably of relatively uniform thickness across the basin (c. 60 m; Figure 4.13B).

The J64 unit (Middle Kimmeridgian) is bounded at its base and top by J64 and J66 maximum flooding surfaces, respectively. It comprises a single shallowing-upwards succession of offshore shelf through offshore transition to lower shoreface deposits, which is thickest (>175 m) in the minibasin axis (well 1/03-3) and <50 m towards the SE basin margin (well 2/01-8) (Figure 4.13B).

The J66 unit (Late Kimmeridgian) is bounded by the J66 and J71 maximum flooding surfaces. It consists of a 67 m thick, shallowing-upwards succession of offshore transition to lower shoreface deposits in the NW of the minibasin (wells 1/03-9s, 1/03-3), which thins laterally into lower shoreface deposits (< 28 m in wells 2/01-6, 2/01-4) and then onlaps underlying deposits toward the SE (Figure 4.13A, B).

The J71 unit (Early Portlandian) is bounded at its base by the J71 maximum flooding surface and is capped by the J72 maximum flooding surface. A shallowing of facies associations from distal-to-proximal is recorded from the NW-to-SE, as offshore shelf deposits (wells 1/03-9s, 1/03-3, 2/01-6) pass laterally into offshore transition deposits (wells 2/01-4, 2/01-8) (Figure 4.13B). The unit is 5-20 m thick (Figure 4.13B).

The uppermost J76-J72 unit consists of offshore shelf deposits throughout and documents ultimate flooding of the minibasin. It is thickest in the axis of the minibasin (50 m, well 1/03-3) and thins to 20 m on its SE flank (well 2/01-8) (Figure 4.13B).

4.9.3 Transect 3: Hidra High to Sørvestlandet High

A flattened, SW-NE-oriented seismic reflectivity line illustrates thickness variations in Upper Jurassic strata between the Hidra and Sørvestlandet highs (Figure 4.14A). To the west of the Cod Terrace, on the Hidra High (well 1/03-8 in Figure 4.14B), five of the
maximum flooding surfaces identified within the Tambar-Gyda minibasin (J56, J62, J64, J71, J72) can be identified using previous biostratigraphic interpretations of well data. The J66 maximum flooding surface is also identified in the Tambar-Gyda minibasin (well 2/01-6 in Figure 4.14B). These maximum flooding surfaces are used to subdivide the J50-to-J70 net-transgressive succession into seven stratigraphic units (J54, J56, J63-J62, J64-J66, J71, J76-J72). Only the J71 and J72 maximum flooding surfaces are correlated to the Sørvestlandet High (wells 2/01-2, 8/11-1 in Figure 4.14B). The J50-to-J70 succession, which is up to 500 m thick in the axis of the Tambar-Gyda minibasin (well 1/03-3 in Figure 4.14B), thins to c. 340 m on the Hidra High (well 1/03-8 in Figure 4.14B) and to <90 m on the Sørvestlandet High (wells 2/01-2, 8/11-1 in Figure 4.14B).

Pre-J32 coastal-plain deposits (Early Bathonian?), which are penetrated in the Tambar-Gyda minibasin on the Cod Terrace, are not developed on the Sørvestlandet and Hidra highs (well 2/01-6 in Figure 4.14B). The overlying J54 unit (Late Oxfordian), which is present in the Tambar-Gyda minibasin, is also developed on the Hidra High, where it directly overlies Triassic strata and is composed of a 90 m thick succession of lower shoreface sandstones (well 1/03-8 in Figure 4.14B).

The J56 and J63-J62 units (Early Kimmeridgian) are penetrated on the Hidra High, where it is composed of an offshore shelf succession that is up to 160 m thick (well 1/03-8 in Figure 4.14B) and therefore substantially thicker than on the Cod Terrace (80 m in well 2/01-6 in Figure 4.14B).

The J64 and J66 units (Middle-to-Late Kimmeridgian), which are together represented by a succession of offshore-to-shoreface deposits up to 240 m thick in the Tambar-Gyda minibasin (Figure 4.13; well 2/01-6 in Figure 4.14B), cannot be subdivided on the Hidra High, where they are only represented by a thin interval (<20 m) of offshore deposits (well 1/03-8 in Figure 4.14B). Thin (<20 m) shoreface deposits of the J66 unit are locally
present on the Sørvestlandet High (well 2/01-2 in Figure 14B), where they directly overlie
and onlap Triassic strata. It is important to note that the Upper Jurassic succession is
highly condensed on the Sørvestlandet High, and that J32, J54, J56, J63-J62 and J64 units
are not penetrated (Figure 4.14B).

The J71 unit (Early Portlandian) is present on the Hidra and Sørvestlandet High, where
it is bounded at its base and top by the J71 and J72 maximum flooding surfaces,
respectively. The unit comprises offshore transition deposits that are c. 25 m thick on the
Sørvestlandet High (wells 2/01-2, 8/11-1 in Figure 4.14B), and which pass laterally into a
c. 15 m thick, offshore shelf succession in the Tambar-Gyda depocenter (well 2/01-6 in
Figure 14B) and on the Hidra High (well 1/03-8 in Figure 4.14B). Offshore shelf deposits
of the J76-J72 unit (Early Portlandian-to-Ryazanian) are as laterally extensive as J71, and
gradually thicken from 20 m on the Sørvestlandet High (well 8/11-1 in Figure 4.14B) to 60
m on the Hidra High (well 1/03-8 in Figure 4.14B).

4.9.4 Chronostratigraphic summary

A chronostratigraphic cross-section across the Gyda, Tambar and Ula fields in the south to
the Sørvestlandet High in the north illustrates diachronous deposition, characterized by
northwards onlap and younging of stratigraphic units (Figure 4.15) as the area of
deposition progressively increased. Stratigraphic units J54 to J76 in the Late Jurassic
(Oxfordian-Ryazanian) succession consist predominantly of low-energy, bioturbated
shoreface, offshore transition and offshore shelf deposits, with the exception of high-
energy, wave-influenced shoreface deposits in the J54 unit to the south of the Gyda and
Tambar fields (Figure 4.15). Coastal-plain and marginal-marine deposits are absent from
these units.
Younging of stratigraphic units to the north is interpreted to reflect a regional influence on accommodation, and can be attributed to the collapse of the Mid-North Sea Dome (Figure 4.2A). Previous studies have documented onlap onto the dome across the Mid-Cimmerian Unconformity, associated with southward retreat and younging of shallow-marine deposits (Rattey and Hayward 1993; Underhill and Partington 1993).

4.10 SANDSTONE PETROGRAPHY, PROVENANCE AND DISPERSAL

Upper Jurassic sandstones in the Central North Sea exhibit a large spread of bulk petrographic compositions (Figure 4.16). Sandstones from well 7/08-3, which is located c. 20 km north of the Ula Field (Figure 4.4), are dominantly subarkosic (green points in Figure 4.16) (Harris, 2006). Further south, in the Tambar Field, sandstones in well 1/03-3 are arkosic (brown points in Figure 4.16). Sandstones from the Tambar Field are thus compositionally similar to those encountered in the Fulmar Field (blue points in Figure 4.16) in the UK Central Graben (Figure 4.2B) (Johnson et al., 1986). In the UK and Danish sectors of the North Sea, petrographic data from the Fife, Angus and Freja fields (Figure 4.2B) indicate that sandstones in these fields display a wide range of compositions and contain a high proportion of lithic fragments (purple and red points in Figure 4.16) (Spathopoulos et al., 2000; Weibel et al., 2010). These data suggest that the sandstones do not have a shared provenance, but instead were supplied from multiple sources or the provenance contained varied lithologies.

Compositional variability within the Upper Jurassic reservoirs of the Fife and Angus fields (Figure 4.2B, purple points in Figure 4.16) led Spathopoulos et al. (2000) to suggest that a mixture of lithologies contributed to sandstone composition, with Devonian and Carboniferous sandstones of the Mid-North Sea High (Figure 4.2B) suggested as the most likely extra-basinal source. The northwards and eastwards retreat of Upper Jurassic
shallow-marine sandstones (Figure 4.15), and proximal-to-distal (SE-NW) facies variations within depositional units in the study area (Figures 4.12-4.14) imply that sediment was supplied from the north and east. Triassic and Permian rocks were likely exposed on the Sørvestlandet, Jaeren and Ringkøbing-Fyn highs during the Late Jurassic (Figure 4.2B), and these structures were probably the source of extra-basinal sediment for the reservoirs in the Ula and Tambar-Gyda fields than the Mid-North Sea High (Spathopoulos et al., 2000).

Several authors have suggested that Triassic structural highs were the local, intra-basinal source areas for sediment that was delivered into Jurassic minibasins in parts of the Central North Sea (Wakefield et al., 1993; Bjørnseth and Gluyas, 1995). However, based on the relatively limited size of these intra-basinal Triassic highs compared to the volume of sediment contained in the Jurassic minibasins, other authors have argued that (extra-basinal) sediment was instead supplied from the basin margins (Weibel et al., 2010). A subcrop map of the Mid-Cimmerian Unconformity in the study area (Figure 17A) shows that the Ula and Tambar-Gyda minibasins are mainly underlain by Triassic sandstones and siltstones of the Skaggerak Formation, whereas Jurassic minibasins on the Sørvestlandet High are mainly underlain by Triassic mudstones of the older Smith Bank Formation. This suggests that, where sandstones lithologies subcrop the Mid-Cimmerian Unconformity, at least some of the Upper Jurassic sandstones in the supra-salt minibasins could have also been sourced locally from intra-basinal Triassic highs.

**4.11 TECTONO-STRATIGRAPHIC CONTEXT AND EVOLUTION OF JURASSIC SUPRA-DIAPIR MINIBASINS**

Based on our regional and sub-regional structural interpretation, and analysis of the sedimentology and sequence stratigraphy of the Jurassic minibasins, we propose a tectono-
stratigraphic model for the Cod Terrace area (Figure 4.18). We illustrate our model with a broadly NW-SE-oriented, schematic cross-section that qualitatively reconstructs the key structural and seismic-stratigraphic geometries observed on the seismic profile shown in Figure 4.10. Seven time intervals are represented in our model; additional detail is provided for the key Middle-to-Upper Jurassic interval of interest, drawing on observation from the facies preservation maps illustrated in Figure 4.17. These maps show facies-association distributions at the point of maximum regression in each of the stratigraphic units and are interpreted in the broader context of the genetically related, Jurassic supra-diapir minibasin array present on the Cod Terrace and surrounding structural terraces.

4.11.1 Early to Middle Triassic
Seismic cross-sections indicate that Early to Middle Triassic strata are preserved in basins between salt walls, in which salt walls are locally developed above the upper tips of basement-involved, sub-salt-restricted normal faults (Figures 4.5, 4.6B and 4.10A-B). The most widely held interpretation for these Triassic depocenters is that they represent minibasins formed as the result of density-driven subsidence of Triassic deposits into the salt (e.g. Hodgson et al., 1992). However, density-driven subsidence of minibasins would require that the average density of Triassic sediments was greater than the density of salt (c. 2200 kg/m³; Hudec et al., 2009), which is unlikely to have occurred until the Middle-to-Late Triassic where these sediments would have been sufficiently buried at this time to have reached a density of 2200 kg/m³ (based on Zechstein well penetrations 7/12-5, 1/03-9s, 8/10-2, 1/03-3, 2/01-2; 2/01-4). An alternative interpretation is that these Triassic depocenters actually represent extensional rafts formed during Late Jurassic extension (Penge et al., 1993, 1999). We therefore suggest that flow of the Zechstein salt, and growth of walls and diapirs, resulted from extension and faulting of a supra-salt roof of Early to
Middle Triassic strata (cf. ‘turtle structure horst’ of Vendeville and Jackson 1992b; ‘pod’ of Hodgson et al. 1992; ‘raft’ of Penge et al. 1993; and ‘minibasin’ of Stewart 2007) (Figure 4.18A). Extension during the Early Triassic has been proposed by Penge et al. (1993, 1999) and Hodgson et al. (1992) (Figure 4.1B, C). Based on the density-inversion required for density-driven subsidence, we suggest that differential loading and density-driven subsidence of the minibasins and growth of the flanking salt walls would not have occurred until the Middle to Late Triassic (Figure 4.18B), as suggested by Hodgson et al. (1992) and Clark et al. (1999) (Figure 4.1B, D).

4.11.2 Late Triassic to Early Jurassic

Well data indicate that the majority of salt walls are capped by locally thick (>300 m) Triassic successions (e.g. wells 2/01-4, 2/01-6, 7/12-6, 7/12-9 in Figures 4.12B and 4.13B), suggesting that the sediment accumulation rate eventually outpaced salt wall growth, most likely during the Late Triassic (Figure 4.18C). This change may reflect either an increase in sediment supply to the basin and/or declining rates of salt flow into the walls due to local exhaustion of the primary salt layer and formation of welds below the adjacent Triassic depocenters (Figures 4.10 and 4.18C). Early Jurassic uplift of the Mid-North Sea Dome led to formation of the Mid-Cimmerian Unconformity and variable erosion of Triassic strata (Figure 4.18D), although the preservation of Triassic strata above salt diapirs indicate that salt was not exposed at the surface and major faults for ground-water infiltration and dissolution of large quantities of salt is not observed on seismic data. A key interpretation arising from this observation is that the majority of the overlying Jurassic minibasins were not formed principally by salt dissolution, which is in direct contrast to the model proposed for the origin of Jurassic depocenters on the Western Platform (cf. Clark et al., 1999). To help understand the potential source areas for sediments filling the
minibasins, a map showing the age of units that subcrop the Mid-Cimmerian Unconformity has also been constructed and is shown in Figure 4.17A; this map is based on well penetrations because subdivision of Triassic strata into the sandstone-prone Skagerrak Formation and mudstone-prone Smith Bank Formation is not possible using seismic data. This map shows that Jurassic minibasins are variably underlain and flanked by Permian evaporites (Zechstein Group), Triassic mudstones (Smith Bank Formation) and Triassic sandstones (Skagerrak Formation).

4.11.3 Middle Jurassic
Determining the Middle Jurassic (Aalenian-Callovian) tectono-stratigraphic evolution of the Cod Terrace and surrounding areas is problematic, due to the relatively limited preservation of related strata. Sparse well data indicate that relatively thin (<50 m), coastal-plain strata are locally preserved below the Early Bathonian (J32?) flooding surface (e.g. wells 1/03-3, 2/01-8 and 2/01-10; Figures 4.13B and 4.17B; see also Figure 4.15); seismic data indicate that these strata and the capping flooding surface onlap the flanks of some of the supra-diapir minibasins (e.g. Tambar-Gyda minibasin; Figure 4.10). Local preservation of this unit may reflect local deposition in minibasins that had begun to develop above collapsing salt walls or local preservation within these minibasins, which would imply that the deposits were originally more widespread but were later eroded; however, based on the observation that stratigraphic onlap of the unit occurs onto the flanks of the supra-diapir minibasins, we prefer the former interpretation.

4.11.4 Late Jurassic (Oxfordian- to- Portlandian)
The patchy distribution of Upper Jurassic strata on the western margin of the Central Graben has been proposed to reflect post-depositional erosion of the unit across structural
highs, with salt collapse-related minibasins preserving only remnants of Upper Jurassic strata that were originally more extensive (Clark et al., 1999). However, our observation that Upper Jurassic strata onlap the margins of supra-diapir minibasins (Figures 4.10E, 4.13A) suggests that the distribution of these strata is instead the result of syn-depositional accommodation creation within these minibasins. The areal extent of increasing fine-grained depositional belts progressively expanded during the Late Jurassic filling of the minibasins, as sandstone-rich shorelines retreated towards the basin margin (Figure 4.17C-G).

Now that we have established that the stratigraphic architecture of the Upper Jurassic succession indicates that supra-diapir minibasins had started to develop during the Late Jurassic, we now explore what mechanism or mechanisms were responsible for their formation. We earlier argued that salt dissolution was not responsible for the formation of supra-diapir Jurassic minibasins, thus we must appeal to alternative mechanisms. Previous studies have indicated that regional extension of the North Sea Basin occurred in the Late Jurassic (Oxfordian-to-Volgian; Figure 4.6C) (Hodgson et al., 1992; Penge et al., 1993, 1999). Regional extension involved both sub-salt and supra-salt strata, and likely resulted in reactivation of sub-salt, basement-involved faults and supra-salt faults present above the salt walls. Stretching of the supra-salt cover would have resulted in widening of salt walls and diapirs, and subsidence above these structures would have resulted in the formation of diapir-collapse minibasins that may locally have been bounded by normal faults that dipped in towards the diapirs (Figure 4.18E) (cf. ‘interpods’ of Hodgson et al. 1992, and ‘rifs’ of Penge et al. 1993; see also modelling results of Vendeville and Jackson, 1992b). The intervening Triassic depocenters would have formed structural highs at this time (cf. ‘turtle structure anticlines’ Vendeville and Jackson 1992b) and may have represented localized sediment source areas (Figure 4.18E).
Now that we have established the likely origin of the supra-diapir minibasins, we document their detailed sedimentary infill history. High-energy, storm-influenced shoreface sandstones (J54 unit) overlie the Middle Jurassic (Early Bathonian; Figure 4.17B) coastal plain deposits and are developed only locally in Jurassic minibasins in the southernmost part of the study area (Figure 4.17C). Further north, low-energy, bioturbated shoreface sandstones are developed in the Tambar-Gyda minibasin (Figure 4.17C). The J54 shallow marine sandstones are capped by offshore shelf deposits, which fill a number of Jurassic depocenters located in the southern part of the study area, including the Tambar-Gyda minibasin (Figure 4.17D). In contrast, the Ula minibasin contains offshore transition deposits, implying that this minibasin was probably connected to the Tambar-Gyda minibasin during this time (Figure 4.17D), to give a consistent proximal-to-distal trend in facies associations from north to south. Two distinct groups of linked Jurassic minibasins are identified during deposition of J64 (Figure 4.17E). On the north of the Cod Terrace, low-energy, bioturbated shoreface sandstones are present in the Ula minibasin, and potentially in other salt-bounded basins developed further west (Figure 4.17E). Further south, in a series of linked Jurassic minibasins, which extend from the westernmost margin of the Sørvestlandet High to the Cod Terrace, low-energy, bioturbated shoreface sandstones pass eastward into offshore transition and then offshore shelf shales, implying increasing water depth towards the basin center (Figure 4.17E). We interpret that these two groups of minibasins were separated from one another because of the relatively abrupt facies association transition (i.e. shoreface to offshore) that occurs just to the south of the Ula minibasin across a salt-cored structural high. Facies-association distributions in J66 imply that Jurassic minibasins on the Sørvestlandet High were connected to those on the Cod Terrace and Hidra High, because low-energy, bioturbated shoreface sandstones pass westwards, via a network of linked minibasins, into offshore transition and offshore shelf
shales (Figure 4.17F). Offshore transition and offshore shelf deposits occur in the Ula minibasin, and potentially further west towards the axis of the Central Graben (Figure 4.17F). All of the Jurassic minibasins appear to be connected during deposition of J71 unit (Figure 4.17G). Offshore transition deposits dominate Jurassic minibasins on the Sørvestlandet High, and these pass southwards into low-energy, bioturbated shoreface sandstones (Figure 4.17G), and westward, onto the Cod Terrace and Hidra High, into offshore shelf shales (Figure 4.17G).

4.11.5 Latest Jurassic to Early Cretaceous (Portlandian-to-Ryazanian)

Rift-related extension and subsidence rates declined during the latest Jurassic (Portlandian-to-Early Cretaceous, as the Central North Sea entered a period of post-rift thermal subsidence (e.g. Ziegler 1990; Zanella and Coward 2003). Regional thermal subsidence, potentially coupled with a rise in eustatic sea-level, caused flooding of the Triassic highs that bound the Jurassic minibasin, migration (backstepping) of sand-rich shorelines towards the basin margins and widespread deposition of offshore shales (Figure 4.18H). The J72-J76 unit, which lies below the K10 maximum flooding surface, contains offshore shelf shales throughout the study area (Figure 4.15H). Unlike the underlying units, deposition at this time was not confined to Jurassic minibasins, but also occurred above the flanking Triassic-cored structural highs (Figure 4.17H). The J72-76 unit represents the culmination of a progressive increase in depositional area, water depth, and the associated net-transgressive retreat of facies-association belts that characterise Middle-to-Late Jurassic deposition on the Cod Terrace and surrounding areas (Figure 4.17B-H).
4.11.6 Late Cretaceous

Regional shortening during the Late Cretaceous-to-Miocene associated with the far-field effects of the Alpine orogeny has been documented in the Central North Sea, where salt structures are interpreted to absorb almost all of the shortening and were rejuvenated to form squeezed diapirs with pinched stems (Sørensen et al., 1992; Ziegler 1992; Stewart, 2007). Alternatively, differential sedimentary loading in the Tertiary can drive passive diapirism and also result in similar salt-related structural styles which represent the main anticlinal trap-forming mechanisms for many fields in the Central North Sea (e.g. Banff, Kyle, Pierce, Machar, Monan; Davison et al., 2000). Compression and inversion structures such as force folds and reverse slip on normal faults have been documented in parts of the Central North Sea (Vendeville and Nilsen 1995; Lewis et al., 2013). However, such structures are not recorded in the study area. Several authors used relatively low-quality, regional, 2D seismic reflection data to propose that this phase of regional shortening led to the development of complex, thick- and thin-skinned transpressional structures on the Cod Terrace (e.g. thrusts, buttress folds and strike-slip faults; Brown et al., 1992; Sears et al., 1993; Stewart 1993). We see no evidence for such structures in our modern, high-quality, 3D seismic reflection dataset, although thinning of Upper Cretaceous strata across the arched roofs of salt walls on the Cod Terrace suggest that diapir squeezing and rejuvenation also occurred in the study area during the Late Cretaceous, and that this phase of deformation may have been responsible for the formation of the Ula Field trap (Figure 4.5C).
4.12 DISCUSSION

4.12.1 A revised tectono-stratigraphic framework for Jurassic supra-diapir minibasins

The tectono-stratigraphic model described above suggests that Late Jurassic extensional diapir-collapse was the main mechanism providing accommodation for the deposition of the shallow-marine reservoirs on the Cod Terrace and, potentially, surrounding areas. This model differs from the ‘pod-interpod’ model which suggests salt-withdrawal, the ‘rift-raft’ model which interprets reactive diapir rise, and the ‘salt-dissolution’ model that have been previously proposed to explain the Late Jurassic tectono-stratigraphic evolution of the Central North Sea (Figure 4.1B-D; Hodgson et al., 1992; Penge et al., 1993, 1999; Clark et al., 1999). Detailed mapping within minibasins identified pod-shaped geometries above diapir crests which differs from the geometries of the depocenters described by Hodgson et al. (1992) (Figure 4.1B) and the crestal graben geometries in the Penge et al. (1993, 1999) models (Figure 4.1C). As we have argued earlier, the presence of Upper Jurassic syn-depositional normal faults (Figure 4.13D) and thick Triassic successions capping many of the salt walls make it unlikely that groundwater infiltration and salt dissolution was the main mechanism controlling the formation of Jurassic minibasins (cf. Clark et al., 1999) model since normal fault formation and dissolution would have to be synonymous in the Upper Jurassic (Figure 4.1D).

Our revised tectono-stratigraphic model can be applied when risking reservoir presence for exploration prospects in the Upper Jurassic play of the Central North Sea. More specifically, less risk can be assigned to prospects located directly above salt diapirs than those located above Triassic turtle-structures (Figures 4.15 and 4.18). Furthermore, relatively late-stage shortening related to Cretaceous inversion can alter the original
geometry of the pod-shaped, supra-diapir minibasins and result in the formation of anticlines that form structural traps (e.g. Ula Field). Onlap of Upper Jurassic strata on to the margins of the diapir-crest minibasins may also provide a component of stratigraphic trapping, particularly where thick Triassic mudstones (Smith Bank Formation) provide basal and lateral seals. These features are not considered in existing published exploration models.

The sedimentology and sequence stratigraphic architecture of the Upper Jurassic in the Central North Sea have been studied by various authors (Johnson et al., 1986; Taylor and Gawthorpe, 1993; Gowland 1996; Martin and Pollard 1996; Johannessen et al., 2010; Sansom 2010). However, these studies have utilised only core, wireline-log and/or biostratigraphic data, and have not attempted to integrate relatively local observations from these data types with the regional tectono-stratigraphic context that is provided by seismic reflection data. As a result, the influence that syn-depositional salt flow and faulting has on accommodation creation and the stratigraphic architecture, thickness, geometry and continuity of reservoirs within minibasins, is poorly understood. Our study suggests that the thickest reservoirs are preserved in the axes of the supra-diapir minibasins and that reservoirs are thin or absent on the minibasin flanks; in our study we find no evidence for regional, Late Jurassic, post-depositional erosion of shoreface sandstone reservoirs as described by Sansom (2010). We note however that reservoir deposition is strongly diachronous and that proximal-to-distal changes, from shoreface-to-shelf facies associations, occur within almost all of the main reservoir-bearing stratigraphic intervals. A key implication of this observation is that reservoir quality can be highly variable within any one stratigraphic unit, and that this can result in highly variable reservoir quality (e.g. J66-J64 unit, Figure 4.12B). Very fine-grained depositional units (e.g. shelf facies associations) may act as potential barriers to flow within otherwise high-quality, shoreface
reservoirs (e.g. J63-J62 units in the Tambar-Gyda fields, Figure 4.13B). Therefore, we advise caution when using existing fields as analogues for ‘new’ discoveries. We suggest that, in existing fields, seismic mapping and detailed sedimentologic studies need to be carefully integrated to aid placement of future production and injection wells.

4.12.2 Wider implications for exploration and production in extensional, salt-influenced sedimentary basins

In frontier basins (e.g. offshore Angola, Hudec and Jackson 2002; eastern Mediterranean, Loncke et al., 2006) where extension is known to have occurred, salt welds are observed, and depocenters above salt walls or former diapir crest have been recognised on seismic data, characteristics of this model can be applied to the interpretation and prediction of reservoir presence for exploration prospects. Extensive studies in salt-influenced basins (e.g. North Sea, Mexico, Red Sea and Santos Basin) document accommodation creation between salt structures on the flanks of diapirs during diapir rise (Davison et al., 1996; Davison et al., 2000; Aschoff and Giles, 2005; Guerra and Underhill 2012). However, during diapir collapse, accommodation is created above former diapir crests, which become sites for deposition of younger strata in minibasins that are not time-equivalent to strata deposited between salt structures during diapir rise (e.g. turtle-structures; Figure 4.1A). Strata infilling diapir-collapse minibasins may be thin or absent in between diapirs. Diachronous deposition and proximal-to-distal facies trends in such minibasin-fill strata can be expected, and will reflect complex, localized linkages in paleobathymetry, paleogeography, and sediment routing between minibasins during deposition of their fill strata. When exploring in salt-influenced extensional basins such diapir-collapse minibasins are often overlooked in favor of depocenters between salt structures. However,
these supra-salt minibasins can be prospective targets as illustrated by the Upper Jurassic play of the Central North Sea.

4.13 CONCLUSIONS

- Upper Jurassic shallow-marine reservoirs in the Cod Terrace and surrounding areas, Norwegian North Sea are deposited in ‘pod-shaped’ minibasins occurring above salt walls. Maximum reservoir thickness is recorded in the axis of the minibasins (280-500 m). Thinning of reservoir strata on the minibasin flanks is accommodated by both thinning and lateral onlap of stratigraphic packages.

- Shallow-marine strata infilling the minibasins comprise six facies associations: (FA 1) bioturbated offshore shale; (FA 2) laminated offshore shale; (FA 3) offshore transition silty sandstone; (FA 4) bioturbated lower shoreface sandstone; (FA 5) cross-laminated lower shoreface sandstone; and (FA 6) cross-stratified upper shoreface sandstone. These facies associations occur in upward-shallowing successions that represent low-energy, bioturbated shoreface parasequences (FA1, FA3, FA4, FA6) and high-energy, wave-influenced shoreface parasequences (FA2, FA5). The former are more abundant, whereas the latter occur only in the older fill strata of minibasins to the south of the Cod Terrace, near the main axis of the North Sea rift basin.

- Minibasins penetrated by multiple wells have different ages and stratigraphic architectures, which consist of aggradationally-to-retrogradationally stacked parasequences (Ula Field) and retrogradationally-to-progradationally-to-retrogradationally stacked parasequences (Tambar and Gyda fields). Thinning of reservoir strata on minibasin flanks is accommodated by both thinning and lateral
onlap of parasequences, such that not all parasequences penetrated in the minibasin axes are encountered on their flanks.

- Proximal-to-distal trends from shoreface-to-offshore shelf facies associations are recorded within all parasequences, but extend over scales larger than individual minibasins. Shallow-marine sandstones also occur in large volumes, and are not confined to areas of localized sandstone subcrop. In combination, these features suggest that the minibasins formed a linked network supplied by regional sediment-routing systems.

- A revised tectono-stratigraphic model for reservoir development on the Cod Terrace is proposed, based on the rise and collapse of salt diapirs. Early Triassic extension resulted in reactive diapir rise and formation of Triassic depocenters. Differential loading of Triassic clastic sediments in these depocenters resulted in the onset of passive diapirism. During the Late Triassic, salt supply for continued passive diapirism had been exhausted and salt welds formed beneath the Triassic depocenters. Late Jurassic extension resulted in diapir-collapse, providing accommodation for development of Upper Jurassic shallow-marine minibasins above collapsing salt walls. This model implies a high risk for Jurassic reservoir presence away from salt walls. Jurassic minibasins may also be affected by later, active diapir rejuvenation, which alters their geometry to form anticlinal traps (e.g. Ula Field).

- The results of this study offer improved prediction of Upper Jurassic reservoir presence, thickness and continuity in the Central North Sea, and can be used as an analogue for interpretation in other salt-influenced basins affected by diapir-collapse.
4.14 ACKNOWLEDGEMENTS

We thank the Norwegian Petroleum Directorate for providing access to seismic and core data. The London Petrophysical Society and the Commonwealth Commission are gratefully acknowledged for funding and support of this research. Centrica Energi is thanked for the provision of funding and access to data and Schlumberger Limited for use of Petrel software via an academic software donation. We extend our gratitude to Professor Alistair Fraser for his inspiring conversations and constructive comments and to Craig Magee and Olly Duffy for remarks on an earlier draft of this manuscript. Reviewers Philip Ball, Christopher Elders, Paul Wilson are thanked for their critical reviews which improved the manuscript and journal editor Michael Sweet for his insightful suggestions.

4.15 REFERENCES CITED


of the continental crust: the legacy of Mike Coward: Geological Society of London, Special Publication 272, p. 361–396.


Figure 4.1: Conceptual models for the structural and stratigraphic architecture of depocenters above salt diapirs. (A) Generic model of initial-stage diapir collapse due to extension and salt withdrawal (sensu from Vendeville and Jackson 1992b). Tectono-stratigraphic model illustrating proposed mechanisms for accommodation creation above salt walls and associated development of Upper Jurassic minibasins in the Central North Sea: (B) ‘pod-interpod’ model of extension over collapsing salt walls (Hodgson et al. 1992); (C) ‘rift-raft’ model of extensional grabens (Penge et al. 1993, 1999); and (D) ‘salt-dissolution’ model (Clark et al. 1999).
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Partington et al. (1993)
Figure 4.3: Composite stratigraphic column for the Central North Sea (modified after the Norwegian Petroleum Directorate, 2013) showing mapped seismic horizons and biostratigraphic calibration of basinwide Upper Jurassic flooding surfaces (Partington et al., 1993).
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diapir stock

Triassic depocentres (salt weld)

salt wall

Triassic thickness (m)

sub-salt, normal faults

Breiflabb Basin

Cod Terrace

Norwegian-Danish Basin

Steinbit Terrace

Sørvestlandet High

Hidra High

No Data

N.D.I.T.W.T (ms)

Depth TWT (ms)

-1600

-5600

10 km
supra-salt thin-skinned normal faults defining edge of salt walls

Depth TWT (ms)

-3800
-1100
3 km

56° 57' 41" N 3° 01' 32" E 3° 05' 15" E
Figure 4.7: Core photos illustrating facies associations (FAs) in Upper Jurassic shallow-marine strata: (A) silty claystone of FA1, with belemnite (B) fragments, Anconichnus (An) and Chondrites (Ch) trace fossils (3624 m in well 7/12-10); (B) interbedded claystone and siltstone lenses of FA2 (3955 m in well 2/01-8); (C) highly bioturbated silty sandstones of FA3, with Teichichnus zig zag (Tz) trace fossils assemblage (3333 m in well 2/01-2); (D) highly bioturbated sandstones of FA4, with Thalassinoides (Th) trace fossils (4318 m in well 2/01-6); (E) low-angle (hummocky?) cross-stratified sandstones of FA5 (3893 m in well 2/02-3); (F) sparsely bioturbated, cross-stratified sandstone of FA6, with Ophimorpha (O) trace fossils (3510 m in well 7/12-6). Facies 6.1 Measured depths are quoted.
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- FA: Formation Age
- J63-J62, J66-J64, J69-J64, J71-J72: Depth intervals

**Properties:**
- RHOB g/cm³: 0.5 0
- NPHI v/v: 1.95 2.95
- GR, API: Various graphs showing changes over depth
Figure 4.8: Representative core and wireline logs through facies associations (FA) and vertical facies-association successions: (A) net-transgressive succession comprising units of FA1, FA3, and FA4 bounded by the J71FS and J64FS flooding surfaces (well 7/12-9, Ula Field); (B) succession of FA1, FA2, and FA3 containing the J64FS flooding surfaces (well 2/1-8, Gyda Field); (C) succession of FA5 (well 1/3-8); and (D) net-transgressive succession comprising units of FA3, FA4, FA6 and containing the J62FS and un-named flooding surfaces (well 7/12-6, Ula Field). See Figure 4.9 for legend. B.I. = Bioturbation Index of Taylor & Goldring (1993). GR = gamma-ray; RHOB = density; NPHI = neutron.
Figure 4.9: Shoreline perpendicular, proximal-to-distal cross-sections through depositional models of: (A) a low-energy, bioturbated shoreface; and (B) a high-energy, storm-influenced shoreface (modified from Gowland, 1996). Facies associations observed in the Ula Formation, and the ichnofabrics of Taylor and Gawthorpe (1993), are placed in the context of these models. (MFWB: mean-fair weather wave base; MSWB: mean-storm wave base)
Figure 4.10: Seismic cross-section displayed using the cosine of instantaneous phase attribute and flattened on the top-Mandal reflector, illustrating Jurassic minibasins above salt walls in the Ula, Tambar and Gyda fields. See Figure 4.4 for line of section. Gamma-Ray logs are displayed for well penetrations.
Figure 4.11: Top Zechstein time structure map showing the location of three seismic and well-log transects (Figures 4.12-4.14).
Figure 4.12: (A) NW-SE-oriented cosine-of-instantaneous-phase seismic cross-section through the Ula minibasin, showing gamma-ray logs along well penetrations; (B) NW-SE-oriented well-log correlation and facies interpretation through the Ula minibasin; and (C) top-Triassic to top-Mandal (K10 maximum flooding surface) seismic isochron map of the Ula depocenter. Panels use the top-Mandal seismic horizon (corresponding to the K10 maximum flooding surface) as a horizontal datum. GR = Gamma Ray (API); NPHI = neutron (v/v); RHOB = density (g cm-3).
Figure 4.13: (A) NW-SE-oriented cosine-of-instantaneous-phase seismic cross-section through the Tambar-Gyda minibasin, showing gamma-ray logs along well penetrations. The line location is presented in Figure 4.11; (B) NW-SE-oriented well-log correlation and facies interpretation through the Tambar-Gyda depocenter; (C) J32? maximum flooding surface to J66 maximum flooding surface seismic isochron map over the Tambar and Gyda fields; and (D) J66 maximum flooding surface to top-Mandal (K10 maximum flooding surface) seismic isochron map over the Tambar and Gyda fields. Panels use the top-Mandal seismic horizon (corresponding to the K10 maximum flooding surface) as a horizontal datum. GR = Gamma Ray (API); NPHI = neutron (v/v); RHOB = density (g cm-3).
Figure 4.14: (A) SW-NE-oriented 2D seismic cross-section from the Hidra High to the Sørvestlandet High, showing gamma-ray logs along well penetrations; (B) SW-NE-oriented well-log correlation and facies interpretation from the Hidra High to the Sørvestlandet High. Panels use the top-Mandal seismic horizon (corresponding to the K10 maximum flooding surface) as a horizontal datum. GR = Gamma Ray (API); NPHI = neutron (v/v); RHOB = density (g cm⁻³).
Figure 4.15: Chronostratigraphic summary of Upper Jurassic (Bajocian to Ryazanian) deposition on the Cod Terrace and surrounding areas.
Figure 4.16: Ternary Quartz-Feldspar-Lithic (QFL) plot of detrital composition of Upper Jurassic sandstones in the Central North Sea (compiled from End of Well Reports; Johnson et al., 1986; Spathopoulos et al., 2000; Harris, 2006; Weibel et al., 2010). The sandstone classification scheme of Folk (1980) is shown. Fields from which sandstones were sampled are located in Figure 4.2B.
(A) Mid-Cimmerian unconformity subcrop map

(B) Below J32?FS 'Parkinsoni' (Middle Jurassic)

Legend:

Well unit thickness (m)

- 0
- 51-100
- 1-50
- > 101

Facies associations groups:
- Offshore shelf
- Shoreface
- Coastal-plain
- Offshore transition

Subcrop map lithology:
- Triassic sandstones and siltstones (Skaggerak Formation)
- Triassic mudstones (Smith Bank Formation)
- Evaporites (Zechstein Group)

Topographic features & faults:
- Permian Rotliegend high
- Non-deposition
- Triassic high
- Seismically mapped extent of salt walls

Maps depict geographical features and geological formations, including subcrop maps and facies associations.
(F) Below J71FS
’Fittoni’
J66-J64 units

(G) Below J72FS
’Okusensis’
J71 unit

(H) Below K10FS
’Stenomphalus’
J76-J72 units
**Figure 4.17:** Maps illustrating the preserved extent, thickness and facies-association distributions of Upper Jurassic chronostratigraphic units (Figure 4.15): (A) subcrop beneath the Mid-Cimmerian Unconformity; (B) maximum regression below J32 maximum flooding surface; (C) maximum regression below J56 FS maximum flooding surface; (D) maximum regression below J62 maximum flooding surface; (E) maximum regression below J64 maximum flooding surface; (F) maximum regression below J71 maximum flooding surface; (G) maximum regression below J72 maximum flooding surface; and (H) maximum regression below K10 maximum flooding surface.
(A) Early Triassic
Sub-parallel deposition

(B) Early-to-Middle Triassic
Extension and reactive diapir rise

(C) Late Triassic
Differential loading, diapirs buried, salt welds form

(D) Early-to-Late Jurassic
Uplift of Mid North Sea Dome and erosion of uppermost Triassic sediments
Figure 4.18: Proposed tectono-stratigraphic model of the rise and collapse of salt diapirs, to explain the creation of accommodation accounting for the geometry, architecture and continuity of Upper Jurassic strata, Cod Terrace, Central North Sea. The line of cross-section is oriented NW-SE and qualitatively reconstructs the NW-SE trending seismic profile shown in Figure 4.10.
CHAPTER 5

Salt wall-confined Triassic stratigraphy in the Central North Sea: minibasins or rafts?

Abbreviated title:
Raft Tectonics, Central North Sea

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5.1 ABSTRACT

The minibasin or raft origin of the salt-bound Triassic (1–3.5 km thick) stratigraphic successions in the Norwegian Central North Sea was investigated using an integrated wireline-log and seismic dataset. The structural framework can be defined by sub-salt restricted normal faults, salt structures such as walls and welds, and supra-salt normal faults which deform the Triassic and Jurassic stratigraphy. Seismic stratigraphic analysis of the Triassic documents either wedge-shaped reflection packages which thickens towards salt walls or tabular, isochronous reflection packages. Wireline-log density analysis of the Triassic indicates at least 1600 m of clastic sediments are required to be deposited for density-driven subsidence and minibasin formation. The results of this study suggest that the salt-bound Triassic stratigraphic successions were initially formed as rafts by Early Triassic extension of the supra-salt cover, rather than the common misconception that all thick successions are minibasins formed by density-driven subsidence. It provides further insight on the various structural styles in a salt-influenced rift basin and contributes to the understanding of the minibasins versus raft tectonics debate for the Triassic successions in the Central North Sea.

5.2 INTRODUCTION

The origin of salt-bound Triassic stratigraphic successions in the Central North Sea is controversial and two competing, end-member hypotheses have been proposed: (i) density-driven subsidence initiated passive diapirism and the formation of minibasins, which occurred after deposition of only few hundred feet of clastic sediment (sensu ‘pods’ Hodgson et al. 1992; Clark et al. 1999; Stewart, 2007; Jackson et al. 2010); and (ii) Late Jurassic extension of the Triassic overburden and formation of rafts, which was associated with reactive diapirism of the autochthonous salt layer (Penge et al. 1993, 1999). The mechanisms
of formation for rafts and minibasins are quite distinctive. Rafts form due to thin-skinned, gravity-driven extension of the cover strata above salt and in response to reactive diapirism (Fig. 5.1a; Jackson & Cramez, 1989; Duval et al. 1992; Penge et al. 1993). In these situations, the thickness of salt influences the structural style of the thin-skinned extensional system, influencing whether large relief (i.e. salt diapirs) or small relief (i.e. salt rollers) salt structure form, and the degree of rotation and separation that occurs within and between rafts (Fig. 5.1a; Jackson & Cramez 1989). In contrast, minibasins form by a variety of initiation and driving mechanisms which include; (i) density-driven subsidence (e.g. Hudec et al. 2009); (ii) extensional diapir fall (e.g. Vendeville & Jackson 1992b); (iii) diapir shortening (e.g. Rowan 2002); (iv) dissolution (e.g. Levell et al. 1988; Clark et al. 1999); (v) sedimentary topographic loading (e.g. Hudec et al. 2009); (vi) subsalt deformation (e.g. Vendeville et al. 1995); and (vii) decay of salt topography (e.g. Jackson & Talbot 1986) (Fig. 5.1b) (see review by Hudec et al. 2009).

It is a common misconception that all thick (1–5 km) salt-bound stratigraphic successions characterised by syn-depositional strata which onlap onto its margins are ‘minibasins’ formed due to density-driven subsidence of sediment into salt; despite the fact that mechanical conditions illustrate otherwise. For example, Hudec et al. (2009) demonstrated that in the Gulf of Mexico, minibasins should contain at least 2300 m of clastic sediment before density-driven subsidence can occur, alluding to the fact that some other mechanism must trigger minibasin formation and drive subsidence. However this concept has yet to be tested outside the Gulf of Mexico. We therefore define the principal aim of this study to investigate the underlying mechanism(s) responsible for the formation of locally thick successions of Triassic strata between salt walls in the Central North Sea. The Central North Sea is covered by relatively good-quality 3D seismic reflection data and borehole data is relatively abundant, thus it provides an ideal natural laboratory to critically appraise the origin and evolution of
salt-bound stratigraphic successions on the Cod Terrace, Sørvestlandet High, Norwegian-Danish and Egersund basins (Fig. 5.2). The results of this study will provide further insight on the structural styles in a salt-influenced rift basin and will contribute to the understanding of the minibasins versus raft tectonics debate for the Triassic successions in the Central North Sea (Hodgson et al. 1992; Penge et al. 1993, 1999; Clark et al. 1999).

5.3 REGIONAL GEOLOGICAL SETTING

The Central North Sea is a broadly symmetric graben located along the southern arm of the North Sea trilete rift system (Fig. 5.2a; Zanella & Coward 2003). Three main tectonic phases have affected the basin; Permian-Triassic extension, Late Jurassic-to-Early Cretaceous extension and Late Cretaceous-to-Cenozoic inversion (Fig. 5.3; Zanella & Coward 2003).

The oldest sediments encountered in the basin are Early Permian, syn-rift, non-marine clastics of the Rotliegend Group, which were deposited in a laterally extensive, intermontane trough (North Permian Basin) located to the north of the Mid-North Sea High (Fig. 5.2a). During the Late Permian, marine flooding and formation of a relatively-restricted sea in the North Permian Basin, coupled with several cycles of evaporation, basin desiccation and marine flooding and recharge, resulted in deposition of a succession of evaporites (Zechstein Supergroup), which at least partly infilled pre-existing, Permian rift-related topography (Fig. 5.3; Clark et al. 1998; Stewart & Clark 1999; Jackson & Lewis, 2013). The Zechstein Supergroup originally consisted of up to 2 km of halite in the basin centre, which was located in the Norwegian-Danish Basin (Smith et al. 1993), and thin carbonate-, anhydrite- and siliciclastic-dominated units towards its flanks (e.g. Western Central Shelf, Forth Approaches Basin; Stewart & Clark 1999).

Early Triassic regional extension caused stretching of the sub-salt basement and supra-salt cover (Zanella & Coward 2003). This initiated salt movement and the formation of reactive
diapirs (Hodgson et al. 1992; Zanella & Coward 2003), whose growth influenced deposition of the predominantly continental clastics, and subordinate anhydrite and carbonates of the Smith Bank and Skaggerak formations. Presently, the Triassic succession is largely salt-bound, and is very thin and locally absent above flanking diapirs (Figs 5.2c and 5.3; McKie & Williams 2009). The Smith Bank Formation is composed of red, arenaceous mudstone that is interpreted to be deposited in a continental playa or in a hypersaline lake (Goldsmith et al. 1995). The overlying Skaggerak Formation is composed of mainly fluvial deposits, and consists of sheetflood sandstones that are interbedded with thin, overbank and lacustrine deposits (Goldsmith et al. 1995). On platform areas, such as the Forth Approaches Basin and West Central Shelf, these salt-bound Triassic successions are up to 1000 m thick (Stewart & Clark 1999) and are typically bound by listric faults that define the margins of salt walls (Fig. 5.2c) (Clark et al. 1999; Fraser et al. 2002; Stewart 2007). In plan-view, the distribution of the Triassic is highly irregular (i.e. linear, irregular, amorphous and dendritic); in cross-section, tabular and wedge-shaped reflection packages are present, which are typically tilted eastwards towards the axis of the Central Graben (Fig. 5.2c; Fraser et al. 2002). On the intrabasinal Jaeren and Forties-Montrose highs the Triassic successions is significantly (i.e. order of magnitude) thinner than observed on the platform areas of the Western Central Shelf. Furthermore, on these intra-basin structures, the Triassic is bound by normal faults and small salt rollers, and are separated by Jurassic stratigraphy (Fig. 5.2c). The basinal area of the Central Graben documents the thickest (>3000 m) salt-bound Triassic successions; in this location they are separated by very tall (c. 4 – 8 km) salt diapirs (Fig. 5.2c; Davison et al. 2000).

Uplift of the Mid-North Sea Dome during the Early Jurassic resulted in the absence of Early-to-Middle Jurassic strata across much of the Central Graben and the formation of the Mid-Cimmerian Unconformity (Figs 5.2a and 5.3; Underhill & Partington 1993). On
platform areas, such as the Western Central Shelf, Late Jurassic-to-Early Cretaceous rifting resulted in moderate collapse of diapirs and formation of shallow marine Jurassic depocentres above their crests (Fig. 5.2c) (Hodgson et al. 1992; Jackson et al. 2010). On the Jaeren and Forties-Montrose highs, Late Jurassic-to-Early Cretaceous rifting resulted in extreme extensional collapse of the diapirs, which is documented by welding of Jurassic depocentres onto sub-salt (Lower Permian) strata (Fig. 5.2c). Diapir rejuvenation in the Late Cretaceous-to-Tertiary resulted in the inversion of these salt structures (e.g. Davison et al. 2000; Stewart, 2007; Jackson & Lewis, 2012; Jackson et al. 2013).

5.3.1 Tectono-stratigraphic models for the formation of Triassic depocentres in the Central North Sea

Three principal models have been proposed to explain the origin of locally thick Triassic successions in the Central North Sea: (i) the “pod-interpod” model; (ii) the “rift-raft” model; and (iii) the “dissolution model” (Fig. 5.4). The “pod-interpod” model proposes that regional extension of sub-salt basement and the salt layer itself during the Early Triassic, in almost the complete absence of a pre-kinematic overburden (supra-salt) layer, initiated salt diapirism and provided accommodation for the accumulation of non-marine Triassic clastics in a series of salt-bound minibasins termed “pods” (Fig. 5.4a; Hodgson et al. 1992). Hodgson et al. (1992) suggest that deposition of only a few hundred feet of clastic sediment in these “pods” was enough to then drive density-driven subsidence of the Triassic into the salt. Subsidence ceased and formation of a salt weld occurred during the Late Jurassic, after all the salt had been expelled from beneath the “pods” into surrounding salt structures. It is likely that the Central Graben “pods” would have been thicker and grounded later than those on the graben flanks due to the greater thickness of underlying salt (Clark et al. 1999). Where the “pods” had become welded, subsequent Jurassic extension resulted in salt withdrawal along salt
walls, which resulted in local subsidence above parts of the diapirs and the formation of Jurassic ‘interpods’ (Fig. 5.4a). Because salt movement is predominantly vertical in this model, no kinematically linked shortening of the overburden is required.

The “rift-raft” model proposes that most of the salt movement was post-Triassic, and occurred in response to regional tilting and thin-skinned extension. In updip areas, the originally isopachous, regionally extensive Triassic strata was extended, whereas downdip, kinematically linked shortening, in the form of basement-detached buckle folding, occurred (Stewart & Clark 1999). Gravity sliding of the cover resulted in the formation of Triassic “rafts” separated by supra-salt “rifts”, the latter representing Jurassic shallow marine depocentres (Fig. 5.4b; Penge et al. 1993, 1999). The “rift-raft” model differs to the “pod-interpod” model in terms of: (i) the timing of salt flow and diapir formation; (ii) the dynamics of salt structure growth; (iii) the origin of locally thick Triassic succession; and (iv) the types and age of kinematically linked structures that are predicted.

The “salt-dissolution” model proposes that Early Triassic exposure and dissolution of the Zechstein salt resulted in the formation of an array of circular collapse features. Differential subsidence of Triassic sediments into the salt, augmented by thin-skinned extension, resulted in the formation of salt-bound Triassic depocentres (Fig. 5.4c). This model differs to the previous two models in terms of the initiating mechanism that formed the salt-bound Triassic stratigraphic successions (i.e. dissolution rather than extension) and the timing of formation (i.e. Middle to Late Jurassic rather than Early Triassic).

5.4 DATA SET

We have used five (S1-S5) 3D seismic reflection surveys and a time-migrated 2D seismic survey for this study (Fig. 5.5). Surveys S1-S4 were acquired during 1992-1999 and
reprocessed post-2000, are post-stack time-migrated and have a record length of 6000 milliseconds two-way time (ms TWT). Together these surveys cover 3500 km² of the Cod Terrace and the northern part of the Sørvestlandet High (Fig. 5.5). Inlines trend E-W and are spaced at 25 m, and crosslines trend N-S and are spaced at 12.5 m. The vertical resolution within the Triassic interval of interest is estimated to be c. 45 m (based on a dominant seismic frequency of 20 Hz and borehole-derived interval velocity of 3500 m s⁻¹). The S5 dataset is a pre-stack time-migrated survey that was acquired in 2005, and which covers 3600 km² of the Egersund Basin (Fig. 5.5). Inlines are oriented NW-SE and crosslines are oriented NE-SW, with a spacing of 25 m and a record length of 6600 ms TWT. The vertical resolution within the Triassic interval of interest is c. 25 m (based on a dominant seismic frequency of 35 Hz and a borehole-derived interval velocity of 3500 m s⁻¹). The 2D seismic survey was acquired in 1997, covers an area of 8,500 km² on the Sørvestlandet High and Norwegian-Danish Basin, and has a vertical seismic resolution of c. 50 m in the Triassic succession (based on a dominant seismic frequency of 18 Hz and a borehole-derived interval velocity of 3500 m s⁻¹).

Twenty-seven wells penetrate the Triassic and each one contains a suite of conventional wireline-log data (e.g. gamma ray, resistivity and sonic logs) and time-depth curves that were used to tie stratigraphic data (i.e. age and lithology of seismically mapped units) to the seismic reflection data (Fig. 5.5). Furthermore, twenty-one of these wells contain density logs, which allowed us to constrain the density of the Triassic and investigate the role that density-driven subsidence may have played in the formation of the locally thick Triassic depocentres (Fig. 5.5). Ten wells contain biostratigraphic data that delineated the Triassic-Jurassic boundary (Mid-Cimmerian Unconformity). Well penetrations are restricted to the Late Triassic where it is relatively thin above salt walls as opposed to the adjacent Triassic depocentres; the exception to this is well 7/12-11, which penetrated c. 60 m of Triassic strata (Fig. 5.5).
5.5 METHODOLOGY

In order to investigate the underlying mechanism(s) responsible for the genesis of the salt-bound Triassic stratigraphic successions in the Central North Sea the following methodology was adopted.

5.5.1 Mapping of structural geometries

Four regionally extensive, moderate-to-high amplitude seismic reflections have been mapped in order to define the structural framework of the study area: (i) Top Rotliegend Group (c. 258 Ma or top Lower Permian) horizon, which defines the base of the salt/top of the sub-salt succession; (ii) Top Zechstein Group (c. 251 Ma or top Permian), which corresponds to the top of the salt and thus defines the present distribution and geometry of salt structures, and the location of flanking, locally thick Triassic successions; (iii) the Mid-Cimmerian unconformity (c. 210 Ma or top Triassic), which corresponds to the top of the salt-bound, locally thickened Triassic successions; and (iv) Top Mandal or Top Flekkefjord Formation (c. 129 Ma or near top Jurassic), which marks the top of a regional flooding event and allows the salt-related deformation of the overburden to be characterised (Fig. 5.3). The lack of well penetrations in the locally thick Triassic successions, coupled with a poor impedance contrasts within and between the Smith Bank and Skaggerak formations, meant it was difficult to map in detail individual stratigraphic units within the Triassic. We therefore made detailed seismic-stratigraphic interpretations where data permit and, where relevant, indicate where we are uncertain of the timing of specific structural events (e.g. rafting). Despite a lack of well data, the thickness of the Triassic were estimated using the Top Triassic – Top Zechstein isochron and an average Triassic interval velocity of 3500 ms\(^{-1}\) derived from sonic logs.
5.5.2 Mechanism(s) for formation of salt-bound Triassic stratigraphic successions

In order to investigate the role of density-inversion on the formation and subsidence of the salt-bound Triassic stratigraphic successions, we adopted the methodology of Hudec et al. (2009). Density-inversion can only drive subsidence when the average density of the entire basin-fill is greater than that of salt. Analysis of Cenozoic, marine siliciclastic depocentres in the Gulf of Mexico suggests that this is mechanically unlikely until at least 2300 m of sediments have accumulated (Hudec et al. 2009). Unlike the Gulf of Mexico, the Triassic of the Central North Sea is comparatively old and buried beneath a few kilometres of younger strata, was deposited in a non-marine environment, and was subjected to various phases of tectonics and diagenesis (Fig. 5.3). As a result of these differences in overall tectono-stratigraphic setting, the Triassic succession studied in the Central North Sea is expected to have different compaction coefficients (Yang & Alpin 2004) compared to the Cenozoic succession studied in the Gulf of Mexico by Hudec et al. (2009). To determine the required thickness of the Triassic necessary for density-inversion, depth versus density plots were constructed. In order to capture the uncertainty within the data, minimum (Min), most likely (ML) and maximum (Max) curves were derived, varying the surface-density between 1400 – 1600 kg/m³ (Athy 1930). Since the buoyancy is determined not by the density of the sediments at the bottom of the basin but by the average density of the sediment column, average-densities vs depth plots and curves were computed and compared to that of salt density. The density of pure halite is 2165 kg/m³ (Alger & Crain 1966). However, salt in the subsurface is rarely pure and consists of proportions of other minerals (e.g. anhydrite, gypsum, dolomite; Warren 2010). Zechstein well penetrations in the study area (i.e. wells 8/09-1, 7/09-1, 7/12-5, 8/10-2, 1/03-3, 2/01-4; Fig. 5.5) suggest an average density of approximately 2200 kg/m³.
5.6 REGIONAL STRUCTURAL FRAMEWORK

Time-structure and isochron maps were generated for the regionally mapped seismic horizons (Fig. 5.3). These maps allowed identification of key structural features at three key levels: (i) basement-involved, sub-salt restricted normal fault system that deform the sub-salt succession but which tip out within the salt; (ii) salt structures, such as walls, stocks and rollers that formed due to flow of the Zechstein Supergroup; and (iii) supra-salt normal faults, which deform the Triassic and Jurassic stratigraphy and die-out downwards into the salt (Fig. 5.6). In the following section these structures are described so as to provide a framework for our detailed analysis of the Triassic succession.

5.6.1 Basement-involved, sub-salt restricted normal fault system

The top Rotliegend time-structure map indicates that a series of basement-involved, sub-salt restricted normal fault systems are present in and define the present-day margins of, the Cod Terrace, Sørvestlandet High, and the Norwegian-Danish and Egersund basins (Fig. 5.7). The Cod Terrace is a NW-SE-oriented, normal-fault bounded terrace that is located on the eastern flank of the Central Graben. NNW-SSE- and NE-SW-striking fault systems dip 40–70° predominantly westwards towards the axis of the Central Graben, and have throws of 100–600 ms (c. 300–1500 m). To the east of the Cod Terrace is the Sørvestlandet High, a NW-trending horst that separates the Central Graben from the Norwegian-Danish and Egersund basins (Fig. 5.7). Its main basin bounding faults dip 40–70°, have throws of 100–350 ms (c. 300–875 m) and, to the east, form the boundary between this intra-basin high and the NW-trending Norwegian-Danish Basin (Fig. 5.7). To the NE of the Norwegian-Danish Basin lies the NW-trending Egersund Basin, which is bounded to the Stavanger Platform and Lista Nose to the N-NE and by the Sele High to the NW (Fig. 5.7). The Stavanger Platform is separated from the Egersund Basin by the NW-SE-striking Stavanger Fault Zone which is a
SW-dipping fault system that has up to 1.5 km of throw. The Lista Nose is a N-S-striking, c. 13 km wide horst, which is bound by N-S-striking faults. The easterly margin of the Sele High is defined by a NE-SW-striking, SE-dipping fault system which has up to 1.5 km of displacement. In the centre of the Egersund Basin, NW-SE- and WNW-ESE-striking, moderate displacement (<500 m), sub-salt normal faults are present and these are overlain by salt diapirs (Fig. 5.7). It is likely that these sub-salt normal faults in the Egersund Basin were active during the main Late Jurassic rift event, although they could conceivably have accrued some displacement during the earlier Permo-Triassic rift event (Hodgson et al. 1992; Erratt et al. 1993). The basement-involved, sub-salt restricted normal fault system that flank the Egersund Basin offset the base-salt surface, but only locally breach the salt where it is thin (e.g. Stavanger Fault Zone; Lewis et al. 2013).

5.6.2 Salt structures

The structure map of the Zechstein Supergroup defines the morphology and distribution of salt diapirs and welds (Fig. 5.5). On the Cod Terrace, salt walls, which are 2–8 km wide and 2–3 km tall, typically overlie the upper tips of basement-involved normal fault systems (Fig. 5.5). In contrast, salt walls on the Sørvestlandet High, which are narrower (up to 5 km wide) but equally as tall (2–3.5 km) as those on the Cod Terrace, form a polygonal network that bound broadly sub-circular to elongate, NW-SE and NNE-SSW-elongate Triassic depocentres (Fig. 5.5). Sparse 2D seismic reflection data in the Norwegian-Danish Basin indicate that a mosaic of salt walls, which are <8 km wide and 2.5–3.5 km tall, trend broadly NW-SE and are sub-parallel to the basement-involved normal fault systems (Fig. 5.5). In the Egersund Basin, salt walls are up to 5 km wide and 1–2.5 km tall, and are situated above and trend parallel to underlying basement-involved normal fault systems (Fig. 5.5). A c. 8 km wide zone of low-relief (< 0.5 km) salt rollers is developed in the centre of the Egersund
Basin (Fig. 5.5). These structures are anomalous in that they occur in the footwalls of thin-skinned, listric normal faults that were only active during the Triassic, as opposed to other salt structures that occur next to Jurassic structures (see below). Salt stocks are locally developed along salt walls, and they are particularly common and anomalously large (3.5 km diameter and up to 4.5 km relief) along the salt walls that define the eastern margin of the Sørvestlandet High (Fig. 5.5). Salt stocks in the Egersund Basin display a similar relationship to salt walls and, in this location, they are up to 5 km in diameter and display relief of up to 4.5 km.

### 5.6.3 Supra-salt normal fault system

The Top Mandal time-structure map illustrates the supra-salt structural framework of the study area, which is characterised by supra-salt restricted, thin-skinned normal faults (Fig. 5.6). Three types of thin-skinned normal faults are identified in the Triassic-Jurassic succession. The first type is listric, define the edges of salt walls and decrease in dip downwards onto the top of salt (Figs 5.6b and 5.8). They locally define the margins of wedge-shaped reflection packages of Triassic strata (see below) (Fig. 5.6b) and their upper tips are located in the Late Jurassic-to-Early Cretaceous interval (Figs 5.6b and 5.8). The second type consists of oppositely dipping pairs of planar normal faults that are developed immediately above and offset the upper part of the Triassic depocenters, and define the edges of Jurassic depocenters (Figs 5.6 and 5.8). These faults have throws of 10–400 ms (c. 15–530 m), dip 45-75°, and define graben or half-graben style depocenters. The third type of fault is only locally developed in the centre of the Egersund Basin; they strike NW-SE, are listric in cross-section, have up to (250 ms) 425 m of throw, are stratigraphically restricted to the Triassic-Early Cretaceous succession, and have salt rollers in their footwalls (Fig. 5.9).
5.7 SALT-BOUND TRIASSIC STRATIGRAPHIC SUCCESSIONS

5.7.1 Triassic geometries

Adjacent to salt walls, where the Zechstein Supergroup is relatively thin and salt welds are developed, thick Triassic successions are present (Figs 5.5 and 5.6). A Triassic isochron defines the thickness, distribution and geometry of the Triassic succession (i.e. above salt welds; Fig. 5.10), and this indicates that the Triassic is 2–3 km thick on the Cod Terrace, generally increases in thickness eastwards towards the Norwegian-Danish Basin (up to 3.5 km), before thinning again into the Egersund Basin (< 2 km) (Fig. 5.10b).

Salt-bound Triassic strata on the Cod Terrace occur in NW-to-WSW-trending units that are 3–7 km wide and 2–3 km thick. Wedge-shaped packages thicken towards salt walls (Figs 5.6a and 5.6b). The majority of these packages dip and thicken westward, although a few eastward-dipping and thickening packages are locally observed (Fig. 5.10b). The relationship between the thickness of the Triassic succession and the bounding salt structures is illustrated by isochron maps between intra-Triassic seismic horizons, which can be locally mapped in a number of depocentres (X and Y; Fig. 5.10c). Figure 5.10d shows that the maximum thickness sediment within the depocentres occurs at its central axis, where the salt is welded at its base.

On the Sørvestlandet High, salt-bound Triassic strata occur in 5–10 km wide, 2–3.5 km thick bodies, and generally thicken northwards (Fig. 5.10b). The lowermost parts of these units are better imaged as compared to the Cod Terrace. They are up to 500 ms (c. 875 m) thick, broadly isochronous and composed of sub-parallel reflections (Figs 5.6c and 5.6d), which are overlain by distinct, wedge-shaped packages that thicken towards salt walls (Figs 5.6b and 5.6c). Wedge-shaped packages are common within these locally thick Triassic bodies, and in a similar manner to that observed on the Cod Terrace (Fig. 5.6a). The wedge-shaped packages generally dip to the west on the western margin of the Sørvestlandet High,
to the east on the eastern margin and to the south on the northern margin (Fig. 5.10b). In the Norwegian-Danish Basin the poor quality of the 2D seismic data precludes a detailed interpretation of reflection geometries in the Triassic. However, where they are imaged, they are sub-parallel and dip towards the west and east (Fig. 5.6a).

In the Egersund Basin, high-quality 3D seismic data indicate that the Triassic occurs in 2.5–3 km thick units that are tabular, isochronous and that are typically flat-lying or, more rarely, dip gently (< 2°) towards flanking salt walls (Figs 5.6a and 5.6d). Marked local changes in Triassic thickness are however observed in the centre of the Egersund Basin (Figs 5.9 and 5.10a). In this location, wedge-shaped Triassic reflection packages dip toward and terminate against, NE-SW-dipping, NW-SE-striking, thin-skinned, listric normal faults, which have salt rollers in their footwalls and which tip-out at the top of the Triassic succession (Figs 5.9 and 5.10b).

5.7.2 Triassic densities

Having established the thickness, distribution and seismic-stratigraphic architecture of the Triassic succession, we now investigate the role of density-inversion to provide additional data/insights into the formation and subsidence of the salt-bound Triassic stratigraphic successions. The density of Triassic strata within the 2700 – 5200 m depth range is quite variable (2210– 2830 kg/m³) but, as would be predicted, increases with depth (Fig. 5.11a). Based on measured present day surface density of clays deposited in continental environments similar to that of the Triassic in the Central North Sea(1400 – 1600 kg/m³; Athy 1930), a minimum (Min), most likely (ML) and maximum (Max) line of best fit were calculated for these data (Fig. 5.11a). Average densities were calculated for each of these cases (Min, ML, Max) and then compared to an average density versus depth cross-plot for Cenozoic mudstones from the Gulf-of Mexico (Fairchild and Nelson, 1989), and derived
Based on the density calculations described in the preceding section it seems unlikely that the Triassic stratigraphic successions are minibasins that formed in response to density-driven subsidence. By applying a subsurface salt density value of 2200 kg/m³, our calculations indicate that density-inversion and spontaneous subsidence of Triassic strata into salt would realistically only occur after at least 1600 (ML) m of Triassic sediments were deposited, which is significantly greater than that predicted by Hodgson et al. (1992) and Clark et al. (1999). Furthermore, our seismic-stratigraphic observations imply that the Triassic was initially deposited as a 500 ms thick, tabular, broadly isochronous unit capping the salt; evidence for early subsidence (i.e. basal minibasin units progressively thinning towards bounding salt structure; see Hudec et al. 2009) are thus lacking (Figs 5.6c and 5.6d). According to Hudec et al. (2009), vertically stacked depocentres and raised bathymetric rims are two additional key criteria for identifying density-overturn as the main driving
mechanism for minibasin subsidence. Intra-Triassic isochron maps (Figs 5.10c and 5.10d) indicate that depocentres shifted through time and that vertically stacked depocentres are rare if present at all. However, the relatively poor resolution of the seismic data within the Triassic interval precludes the identification of subtle, raised bathymetric rims.

Although mechanical considerations and seismic-stratigraphic patterns suggest that density-driven subsidence was not responsible for the formation and initial (i.e. Early Triassic) subsidence of the salt-bound Triassic successions, it is plausible that this process may have driven subsidence in the Middle-to-Late Triassic, when the sediment thickness was greater and the average sediment density exceeded that of salt. Based on the compelling evidence that the Lower Triassic was deposited as a broadly tabular unit prior to the initiation of salt movement (Fig. 5.6c), it seems unlikely that other mechanisms such as extensional diapir fall, salt dissolution, diapir shortening, sedimentary topographic loading, decay of salt topography, and subsalt deformation triggered minibasin formation and subsidence (i.e.; Fig. 5.1b) (see section on Regional Structural Framework; Figs 5.5, 5.6, 5.7 and 5.8). Salt dissolution has been proposed for the Triassic minibasins in the Central North Sea in association with density-driven subsidence and extension (Clark et al. 1999; Cartwright et al. 2001; Fig. 5.4c). The presence of an anhydritic “cap-rock” at the top of Zechstein well penetrations (e.g. wells 8/9-1, 8/12-1, 8/10-2) above salt walls and diapirs suggests some extent of dissolution; even though its timing is uncertain. Physical experimental models carried out by Ge and Jackson (1998) to determine the role of dissolution for accommodation creation suggests that it is unlikely for ground-water infiltration to dissolve diapirs alone especially in extensional settings for minibasin formation (Ge & Jackson 1998). Thus, it is not considered as a sole mechanism for minibasin formation in the Triassic based on the supra-salt extensional setting of the study area (see section on Regional Structural Framework; Figs 5.5, 5.6, 5.7 and 5.8).
Our preferred interpretation is that ‘raft’ (Duval et al. 1992) tectonics is responsible for the geometry and distribution of the Triassic succession in the East Central Graben. Raft tectonics is defined as thin-skinned extension of the supra-salt succession, and the formation of structural blocks or ‘rafts’ that are separated from one another by salt-detached normal faults. These faults typically have salt rollers or diapirs in their footwalls (Fig. 5.1a). In situations where salt is thin or the magnitude of extension is mild, low-relief salt rollers form in the footwalls of the thin-skinned faults (e.g. Egersund Basin; Fig. 5.9); in contrast, where salt is thick and/or the magnitude of extension is moderate to extreme, large diapirs may occur (e.g. Cod Terrace; Fig. 5.6b) (Fig. 5.1a). The structural geometries observed on seismic data are geometrically different to the ‘rift-raft’ model of Penge et al. (1993, 1999) (compare Fig. 5.4b with Figs 5.6 and 5.9), which can be explained by the difference in timing of extension and the thickness of salt present. The Penge et al. (1993, 1999) model proposes that extension occurred during the Late Jurassic, whereas we argue that the wedge-shaped geometry of intra-Triassic units, such as that recorded on the Cod Terrace and Sørvestlandet High, is compelling evidence for syn-Triassic extension (Figs 5.6a and 5.6b). However, in the Norwegian-Danish and Egersund basins, where the Triassic is tabular and isopachous (Figs 5.6c and 5.6d), it is likely that rafting occurred during the Late Jurassic rather than Triassic, which is thus in accordance with the timing of that in the Penge et al. (1993; 1999) model.

5.9 TECTONO-STRATIGRAPHIC MODEL FOR THE TRIASSIC EVOLUTION

Our regional structural interpretation of the Permian-Jurassic interval in the eastern Central Graben, coupled with a detailed analysis of the Triassic seismic-stratigraphic architecture and densities of the Triassic succession has provided important insight into the tectono-stratigraphic evolution of this hitherto poorly understood part of the North Sea Basin. A critical element of this model is the mechanism(s) that led to preservation (rather than
deposition) of locally thickened successions of Triassic strata between regionally extensive salt walls. We illustrate our model with a schematic cross-section that qualitatively reconstructs the structural and stratigraphic geometries observed on a regional seismic profile that crosses the core of our study area; i.e. the Cod Terrace, the Sørvestlandet High and the Norwegian-Danish Basin (Fig. 5.12).

5.9.1 Early-to-Middle? Triassic

Seismic cross-sections indicate that the basal Lower Triassic interval comprise of up to a 500 ms thick, broadly isopachous, sub-parallel reflections. On the Cod Terrace, Sørvestlandet High and locally in the centre of the Egersund Basin this is overlain by wedge-shaped packages, which thicken towards bounding salt diapirs and dips in various three-dimensional directions (Figs 5.6c, 5.9 and 5.10b). On the Cod Terrace and Sørvestlandet High regional dips of Triassic strata are broadly towards the west (Fig. 5.10b). In contrast, the Norwegian-Danish Basin and across much of the Egersund Basin, the Lower Triassic succession is fairly tabular and the dips of Triassic strata are in variable directions (Figs 5.6d and 5.10b). This variability suggests a difference in timing of the formation of salt structures between the Cod Terrace and Sørvestlandet High, and the Norwegian-Danish and Egersund basins. Our analysis of density data suggests it is highly unlikely that the Triassic succession was sufficiently thick at this time to initiate and sustain density-driven subsidence of the Triassic into the underlying salt and, therefore, trigger halokinesis. We therefore suggest that Early-to-Middle? Triassic extension reactivated sub-salt normal faults which triggered reactive diapirism of Upper Permian salt on the Cod Terrace, Sørvestlandet High and in the centre of the Egersund Basin and supra-salt extension and gravity sliding of a relatively thin Triassic overburden. The direction of gravity sliding can be interpreted by the overall dips within the Triassic strata. On the flanks of this rift basin (i.e. Cod Terrace and Sørvestlandet High), dips
of the Triassic strata are broadly towards the west (Fig. 5.10b) which suggests gravity sliding of the cover structure towards the axis of the Central Graben in response to reactivation of sub-salt normal faults. It can be expected that such updip extension is accommodated by buckle folding and shortening towards the axis of the Central Graben, similar to that documented in the Triassic on the Western Central Shelf and the West Central Graben in the UK sector of the North Sea (see Fig. 5.2b for location; Stewart & Clark 1999). However, such downdip shortening lies outboard of the study area and is difficult to identify in the subsurface due to latter stages of regional extension and compressional tectonic events experienced in the Central North Sea (Fig. 5.3).

5.9.2 Middle-to-Late Triassic

Well (e.g. 7/12-10, 1/03-3, 9/02-1; Fig. 5.5) and seismic data indicate that the salt walls are capped by relatively thick (c. 300 m), broadly tabular, Triassic roof, implying that they were buried during the Late Triassic in response to an increase in sediment accumulation rate and emplacement of a thick roof that could not be breached by salt, and/or a decrease in salt flow into walls due to welding and exhaustion of the autochthonous salt source (see Mannie et al. 2014). Based on the observations that salt welds are developed below most of the rafts and that the Late Triassic was characterised by a relatively higher sedimentation rate of fluvial deposits of the Skagerrak Formation as compared to the Early Triassic lacustrine, Smith Bank deposits, we suggest that both processes together resulted in burial of the salt walls. Our density analysis suggests that, in areas where the Triassic attained a thickness of >1600 m (e.g. Cod Terrace, Sørvestlandet High and the Norwegian-Danish Basin; Fig. 5.10b), density-inversion may have now been able to drive subsidence of the Triassic rafts into the salt and the eventual formation of salt welds (Figs 5.11b and 5.12c).
5.9.3 Early-to-Middle Jurassic

Early Jurassic uplift of the Mid-North Sea Dome led to formation of the Mid-Cimmerian Unconformity and the absence of much of the Early-to-Middle Jurassic stratigraphy across the eastern Central Graben (Figs 5.2a and 5.12d; Underhill & Partington 1993). Due to a lack of biostratigraphic data within the Triassic, we cannot constrain exactly how much Triassic stratigraphy was eroded during the Early Jurassic. However, based on the Mid-Cimmerian subcrop maps presented by Underhill and Partington (1993) and Andsbjerg et al. (2001), it is clear that Triassic strata was tilted southwards, away from the apex of the Mid North Sea Dome (Fig. 5.2a), and that progressively older strata are preserved southwards away from the location of the dome. It has been hypothesised that a regional southward tilt caused by Early Jurassic thermal doming could have resulted in gravity sliding on the platform areas flanking the Central Graben (e.g. Western Central Shelf Fig. 5.2c; see Stewart 2007). However, uplift of the Mid-North Sea Dome during the Early Jurassic (Underhill & Partington 1993) post-dates (syn-Triassic) gravity sliding and extension documented in the Cod Terrace, Sørvestlandet High and locally in the Egersund Basin. Furthermore, this tectonic event pre-dates the second period of halokinesis that occurred during the Late Jurassic-to-Early Cretaceous (see below).

5.9.4 Late Jurassic-to-Early Cretaceous

On the Cod Terrace and Sørvestlandet High, a second phase of extension during the Late Jurassic resulted in diapir collapse, and the formation of supra-salt wall depocentres and Triassic turtle-structures (Fig. 5.12e; see Mannie et al. 2014). Extension would have initiated movement of salt along strike to feed diapir growth along salt walls and further tilting of Triassic rafts (Fig. 5.7b). In the Norwegian-Danish and Egersund basins, the relatively tabular and isopachous reflection packages in the Triassic suggests that the main phase of
supra-salt extension and reactive diapirism initiated in the Late Jurassic-to-Early Cretaceous (Fig. 5.12e; see Mannie et al. 2014).

5.10 CONCLUSIONS

The structural styles of the Permian-Jurassic strata of the Cod Terrace, Sørvestlandet High, Norwegian-Danish and Egersund basins of the eastern flank of the North Sea Central Graben can be defined by sub-salt normal faults, salt structures and supra-salt normal faults. Salt walls bound Triassic stratigraphic successions which comprise of either wedge-shaped reflection packages which thickens towards salt walls or tabular, isochronous reflection packages. Wireline-log density analysis of the Triassic suggests at least 1600 (ML) m of sediments are required to be deposited before density-driven subsidence acts as a driving mechanism for minibasin subsidence. This contradicts previous models for the Central North Sea which suggests that the salt-bound Triassic successions in the Central North Sea were formed by density-driven subsidence after only a few hundred metres of deposition of continental clastics. The results of this study suggest that the salt-bound Triassic stratigraphic successions were initiated as rafts formed by extension of the supra-salt cover rather than minibasins formed by density-driven subsidence. On the Cod Terrace, Sørvestlandet High and locally in the centre of the Egersund Basin these salt-bound Triassic stratigraphic successions are interpreted to form by Early-to-Middle Triassic extension, which initiated reactive diapirism and salt movement resulting in supra-salt listric normal faulting of the Triassic cover structure. Continued Late Triassic deposition resulted in density-driven subsidence, welding of the Zechstein Supergroup and the burial of salt walls. A Late Jurassic second phase of extension resulted in diapir fall and formation of Jurassic depocentres on the Cod Terrace and Sørvestlandet High. During this time the main phase of extension is interpreted to form the Triassic rafts in the Norwegian Danish and Egersund basins.
5.11 ACKNOWLEDGEMENTS

The authors would like to express their gratitude to the Norwegian Petroleum Directorate and PGS for providing access to the seismic data. Centrica Energi Norge and the Commonwealth Commission are gratefully acknowledged for funding and support for this post-graduate study. We thank Mike Charnock for the biostratigraphic analysis and related discussions. Alastair Fraser and Adrian Hartley are thanked for their critical reviews of this manuscript.

5.12 REFERENCES


Fig. 5.1: Mechanisms for rafts and minibasin formation. (a) Raft formation in thick and thin salt (modified from Jackson & Cramez 1989). (b) Seven initiating and driving mechanisms for minibasin formation (modified from Clark et al. 1999; see review by Hudec et al. 2009).
Fig. 5.2: (a) Regional geographic setting of the Central North Sea trilete rift system and the outer limit of the Zechstein distribution in the North Permian Basin (taken from Smith et al. 1993 and Zanella & Coward 2003).
(b) Simplified structure map indicating the major present day topographic features in the Central North Sea (modified from Glennie et al. 2004).
(c) Simplified SW-NE geoseismic section across the Central Graben illustrating the main geographic elements, structural styles and their spatial relationships (modified from Fraser et al. 2002).
Fig. 5.2c
Western Central Shelf
Forties-Montrose
High

Jaeren High

East Central Graben

Salt diapirs

Salt-bound Triassic strata

Jurassic depocentres

North Sea time in seconds

25 km

248
Fig. 5.3: Regional tectono-stratigraphic column for the eastern flank of the North Sea Central Graben (modified from Vollset & Doré 1984) showing major mapped seismic reflection events.
Fig. 5.4: Conceptual models for the evolution of Triassic depocentres in the Central North Sea. (a) ‘pod-interpod’ model of Hodgson et al. (1992). (b) ‘rift-raft’ model of Penge et al. (1993, 1999). (c) ‘salt-dissolution’ model of Clark et al. (1999).
Fig. 5.5: Distribution of seismic, biostratigraphic and wireline-log density data used for this study overlying the Zechstein (TWT) structure map. Salt structures are characterised by regional scale salt walls, salt diapirs, salt rollers and salt welds. Sub-salt, basement-involved normal faults define basin bounding topographic features.
Norwegian-Danish Basin
Sørvestlandet High
Cod Terrace
Norwegian-Danish Basin
Egersund Basin

Fig. 5.6b
Fig. 5.6d

Sub-salt
Salt wall
Jurassic depocentre
Listric faults
Wedge-shaped reflections
Tabular, isochronous reflections

500 ms 5 km
300 ms 2 km
Fig. 5.6: (a) Regional uninterpreted and interpreted SW-NE trending seismic line flattened on the top Jurassic showing the interpretation of mapped seismic horizons defining salts structures, Triassic and Jurassic depocentres. See Fig. 5.5 for line of section. (b) Seismic section illustrating the wedge shaped reflection geometries within Triassic depocentres on the Sørvestlandet High. See Fig. 5.6a for location. (c) Seismic section illustrating the tabular, sub-parallel reflection geometries with the Triassic depocentres on the Sørvestlandet High. See Fig. 5.5 for location. (d) Seismic section illustrating flat lying, sub-parallel reflection geometries within Triassic depocentres in the Egersund Basin. See Fig. 5.6a for location.
Fig. 5.7: Top Rotliegend (TWT) structure map depicting sub-salt, basement-involved normal faults and basin bounding topographic features.
Fig. 5.8: (a) Top Mandal (TWT) structure map depicting the supra-salt thin-skinned listric normal faults along salt walls and planar normal faults with graben geometries. See Fig. 5.5 for location. (b) Seismic cross-section illustrating the supra-salt normal faults shown in plan view in Fig. 5.6a. See Fig. 5.8a for location and Fig. 5.6 for key.
**Fig. 5.9:** SW-NE seismic cross-section of the salt roller zone in the Egersund Basin, illustrating the supra-salt thin-skinned listric normal fault detaching on salt rollers. See Fig. 5 for location and Fig. 5.6 for key.
Fig. 5.10: (a) Triassic (TWT) isochron map defining the geometry and orientation of Triassic depocentres and salt structures. (b) Plan view line drawing of Triassic depocentres illustrating their maximum thickness (km), characteristics of reflection packages and their overall dip directions. (c) Intra-Triassic X-to-intra-Triassic Y seismic isochron map. See Fig. 5.5 for location. (d) Intra-Triassic X-top Zechstein isochron map. See Fig. 5.5 for location.
Triassic thickness (km)
- 3 - 3.5
- 2.5 - 3
- 2 - 2.5
- < 2

wedge-shaped reflection packages
sub-parallel reflection packages
overall Triassic strata dip direction
Top Rotliegend normal faults
Fig. 5.11: (a) Wireline-log depth versus density plot for wells in the study area with a Min, ML and Max best-fit curves. See Fig. 5 for well locations. (b) Comparison of average-density vs depth plot for the study area to compaction curves for different lithologies and environments assuming a grain density of 2600 kg/m³. (Modified from Hudec et al. 2009). Average sediment density does not exceed salt density until at least 2300 m of sediment have accumulated.
Fig. 5.12: Proposed model to explain the evolution of the Triassic depocentres on the eastern flank of the North Sea Central Graben. The line of section is oriented SW-NE along the same line of section as Fig. 5.6a and qualitatively reconstructs this seismic profile. Plan view of line of section is shown on Fig. 5.5 (highlighted white). (a) Early Triassic deposition. (b) Early-to-Middle Triassic extension and diapirism. (c) Late Triassic differential loading. (d) Early-to-Middle Jurassic erosion; (e) Late Jurassic-to-Early Cretaceous extension.
CHAPTER 6

Tectonic controls on the spatial distribution and stratigraphic architecture of a net-transgressive succession in a salt-influenced rift basin: Middle-to-Upper Jurassic, Norwegian Central North Sea

Abbreviated title:
Structural and stratigraphic architecture in a salt-influenced rift basin

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6.1 ABSTRACT

The influence of structural controls on the sub-regional stratigraphic architecture of net-transgressive strata in rift basins was investigated using an integrated borehole and seismic dataset from the Norwegian Central North Sea. The rift-related structural framework is strongly modified where an Upper Permian salt layer is sufficiently thick and mobile to act as an intra-stratal detachment, decoupling rift-related basement and cover structures, giving rise to structural styles defined by sub-salt normal faults, various salt structures and supra-salt normal faults. Rift-related normal faulting and the growth of salt structures influence the depositional thickness and facies distribution within the coastal-plain to shallow-marine, Middle-to-Upper Jurassic syn-rift succession. The resulting facies architecture reflects a delicate balance between fault- and halokinesis-driven accommodation creation and intra- and extra-basinal sediment supply. Where sediment supply exceeded accommodation, little or no change in facies is observed across syn-depositional structures. In contrast, where accommodation outpaced sediment supply, localised, footwall-attached shorelines are developed, with deeper water conditions in the adjacent hangingwall. Flooding was diachronous and influenced by the underlying rift topography of intra-basinal horsts and graben. This paper highlights the key role that salt plays in modifying the tectono-stratigraphic evolution of rift basins.

6.2 INTRODUCTION

Rift basins may contain salt in the pre-rift succession (e.g. Red Sea, Barents Sea, Central North Sea) and, if sufficiently thick and mobile, it may act as an intra-stratal detachment that decouples rift-related basement and cover structures, thus giving rise to a range of complex structural styles typified by folding and gravity sliding (e.g. Stewart et al. 1996; Stewart & Clark 1999; Corfield & Sharp 2000; Stewart 2007; Marsh et al. 2010; Lewis et
al. 2013; Wilson et al. 2013). For example, salt may flow from the hangingwall to the footwall of rift-related normal faults, resulting in the development of large salt diapirs, and basement tilting may lead to thin-skinned extension of the cover and formation of reactive diapirs (e.g. Vendeville & Jackson 1992b; Stewart & Clark 1999). In contrast, where salt is thin and largely immobile, basement and cover may be fully coupled, thus resulting in structural styles that are similar to those documented in rift basins that lack salt (Leeder & Gawthorpe 1987; Prosser 1993; Ravnás & Steel 1998; Gawthorpe & Leeder 2000).

Variable structural coupling between basement and cover during rifting, in addition to the growth of salt diapirs, adds further complexity to coeval syn-rift sediment routing systems. For example, the growth of salt structures locally influences stratigraphic thickness, stratal architectures, and facies distribution, with strata commonly thinning and a shallowing of facies recorded from the axis of depocentres towards adjacent salt-cored structural highs (Davison et al. 1996; Alves et al. 2003; Aschoff & Giles 2005; Kieft et al. 2010; Mannie et al. 2014). Furthermore, relatively late extensional collapse of salt diapirs may result in the formation of minibasins that provide local accommodation for thick sedimentary successions (Wakefield et al. 1993; Stewart & Clark 1999; Mannie et al. 2014). These purely halokinesis-driven depositional styles may be superimposed on or spatially partitioned from those associated with salt-free rift basins (Erratt 1993; Gawthorpe & Leeder 2000; Stewart 2007). The structural style of rifts containing salt, which may be characterised by both extensional and halokinetic structures, may therefore be more complex than those that lack salt (e.g. Alves et al. 2003; Stewart 2007).

Our understanding of the regional, syn-rift tectono-stratigraphic evolution of salt-influenced rift basins is limited because: (i) most studies are relatively local, and focus solely on either purely salt- or fault-controlled settings in individual fields or sub-basins (e.g. Clark et al. 1998; Alves et al. 2003; Jackson et al. 2005; Kieft et al. 2010); or (ii) the
structural style, and syn-rift sedimentology and stratigraphy are insufficiently integrated (e.g. Johnson et al. 1986; Wakefield et al. 1993; Spathopoulos et al. 2000). Because of these issues there are a number of unanswered questions; for example, how does the presence of salt influence stratigraphic architecture in rift basins? Can existing tectono-stratigraphic models for rift basins, which are principally based on salt-free settings (Gawthorpe & Leeder 2000), be applied to basins where salt is present?

The aim of this study is to constrain the tectonic controls on syn-rift depositional patterns in a salt-influenced rift basin. In particular, we aim to identify the various sub-, supra- and salt structural styles and their influence on the deposition of a net-transgressive, shallow-marine succession of early syn-rift age. We focus on the eastern flank of the Norwegian Central North Sea Graben, which is an ideal location to undertake this type of study because seismic reflection and borehole-related data are relatively abundant, and a basinwide biostratigraphic framework allows us to link periods of deformation and specific depositional styles between various structural domains.

6.3 REGIONAL TECTONO-STRATIGRAPHIC FRAMEWORK

The Viking Graben, the Central Graben and the Moray Firth Basin represent the three arms of the trilete North Sea rift system (Fig. 6.1). During the Late Permian, the Central Graben was located near the margin of a marine seaway (the North Permian Basin) that was subject to high rates of evaporation. Several flooding-evaporation cycles resulted in deposition of a thick, salt-bearing succession (Zechstein Supergroup) (Clark et al. 1998; Stewart & Clark 1999). Early Triassic rifting triggered reactive diapirism of the Zechstein salt and the formation of large salt walls, which influenced the deposition of continental clastics of the Smith Bank and Skagerrak formations (McKie & Williams 2009) (Fig. 6.2).
Uplift of the Mid-North Sea Triple Junction Dome during the Early Jurassic caused erosion of the uppermost Triassic succession, and resulted in the absence of Early- to-Middle Jurassic strata across much of the Central Graben due to non-deposition or post-depositional erosion (Mid-Cimmerian unconformity) (Underhill & Partington 1993). Thermal deflation and subsidence of the Mid North Sea Dome during the Middle Jurassic was synchronous with the initiation of rifting, and together these triggered marine flooding and prolonged (c. 30 Myr) retreat of shoreline systems southward along the axis of the northern and central North Sea and deposition of a net-transgressive shallow marine succession (Fig. 6.3; Rattey & Hayward 1993; Underhill & Partington, 1993; Andsbjerg et al. 2001). Activity on rift-related faults initiated during the latest Bajocian-to-early Bathonian in the North Viking Graben; further south, in the Moray Firth, faulting did not commence until the Bathonian-to-Callovian (e.g. Fraser et al. 2002; Husmo et al. 2003). Rifting and halokinesis in the Central Graben led to complex temporal and spatial variations in the timing of flooding and diachronous deposition within various parts of the graben (Fig. 6.3). For example, in the Norwegian sector of the Central Graben, flooding is recorded by the coastal-plain deposits of the Bryne Formation (Bajocian-Bathonian?), which are conformably overlain by shallow-marine deposits of the Sandnes Formation (Callovian) in the Norwegian-Danish and Egersund basins (Fig. 6.2). In contrast, on the Cod Terrace, the Bryne Formation is unconformably overlain by shallow marine deposits of the Ula/Gyda Formation (Oxfordian-Kimmeridgian) (Fig. 6.2). Continued relative sea-level rise during the Late Jurassic-to-Early Cretaceous, driven by rift-related and salt-influenced subsidence, led to an increase in water depth and deposition of the shelfal and deep-marine deposits of the Tau, Sauda, Farsund, Mandal and Flekkefjord formations (Fig. 6.2). This study focuses on the impact that Middle-to-Upper-Jurassic, rift-related extension
and salt tectonics had on the stratigraphic architecture of the Bryne, Sandnes and Ula/Gyda formations (J20- J70 units of Partington et al. 1993).

6.4 DATASET
To achieve our study aims, we use 2D and 3D seismic reflection data that cover a combined area of >15,000 km² on the eastern margin of the Central Graben, and data from 53 wells (Fig. 6.4). Seismic surveys S1-S4, which were acquired between 1992 and 1999, and reprocessed after 2000, are post-stack time-migrated and have a record length of 6000 milliseconds two-way time (ms TWT). The S5 dataset is a pre-stack time-migrated survey that was acquired in 2005 and has a record length of 6600 milliseconds two-way time (ms TWT). The 2D seismic survey was acquired in 1997 and is post-stack time-migrated. All wells fully penetrate the Middle-to-Upper-Jurassic succession, contain at least two conventional wireline logs (e.g. gamma ray, resistivity, density and sonic), and have checkshot data that allowed us to tie stratigraphic data (i.e. age and lithology) to the seismic reflection data. Twenty-six wells contain a total of 1180 m of core in the Middle-to-Upper-Jurassic succession. Biostratigraphic data were available for twenty-one wells and these allowed regional correlations between sub-basins and enabled us to tie our stratigraphic framework to the basinwide sequence stratigraphic framework of Partington et al. (1993).

6.5 METHODOLOGY
Four seismic horizons have been mapped to define the regional structural framework of the study area (Fig. 6.2): (i) Top Rotliegend Group (258 Ma), which defines the sub-salt, predominantly rift-related structure of the study area; (ii) Top Zechstein Supergroup (251 Ma), which defines the present-day distribution and geometry of salt structures; (iii) Top
Egersund Formation (144 Ma); and (iv) Top Mandal Formation (129 Ma), which broadly defines the top of the Late Jurassic interval and highlights the pattern of rift- and salt-related deformation in the supra-salt interval.

Biostratigraphic data allowed us to identify a number of age-constrained flooding surfaces (e.g. K10FS, J52FS, J32?FS) within the J20-J70 stratigraphic units that could be correlated to the North Sea-wide, Jurassic stratigraphic framework scheme of Partington et al. (1993) (Fig. 6.2). Checkshot data allowed these biostratigraphically defined flooding surfaces to be tied to seismic data through the construction of synthetic seismograms (Mannie et al. 2014). Core data permitted detailed sedimentological facies analysis of the Middle-to-Upper-Jurassic succession. Based on the vertical facies relationships observed in core, higher-order flooding surfaces were identified, such as those within the J40 unit (FS2 – FS4; Fig. 6.2). Wireline-log responses were calibrated to core descriptions (petrofacies characterisation) in order to qualitatively interpret lithology and facies in uncored intervals. Flooding surfaces interpreted in core and wireline-log data were combined with seismic mapping and biostratigraphic data (calibrated to the scheme of Partington et al. 1993) in order to construct a sequence stratigraphic framework for the Middle-to-Upper-Jurassic succession.

6.6 REGIONAL STRUCTURAL AND STRATIGRAPHIC FRAMEWORK

In order to understand the temporal and spatial development of rift- and halokinesis-related structural styles, and their control on the Middle-to-Upper Jurassic (J20-J70) syn-rift stratigraphic development of the eastern Central Graben, sub-regional structure and isochron maps were integrated with biostratigraphic data, sedimentological core descriptions and wireline-log analysis.
The main structural elements within the study area are: (i) the platform areas of the Cod Terrace and Stavanger Platform; (ii) intra-basinal highs such as the Sørvestlandet High, Sele High and Lista Nose; and (iii) basinal areas of the Norwegian-Danish and Egersund basins (Fig. 6.4). The main structural styles include: (i) basement-involved, sub-salt normal faults, which formed in response to Permo-Triassic and Jurassic-to-Early Cretaceous rifting; (ii) salt structures that formed in response to flow of the Zechstein salt; and (iii) supra-salt normal faults, which form in response to a combination of flow of the Zechstein salt and Jurassic-to-Early Cretaceous rifting (Figs 6.5a and 6.6).

Sedimentological facies analysis of core from the Ula/Gyda Formation (J50-J70 unit) on the Cod Terrace, and the Bryne and Sandnes formations in the Egersund Basin (J20-J40) identified four main facies associations (Table 6.1). (i) offshore shelf: which consists of a silty mudstone and a pinstriped siltstone dominated by Anconichnus and Chrondrites trace fossil assemblages; (ii) offshore transition: which comprises bioturbated and hummocky cross-stratified siltstone to very-fine grained sandstone in which Teichichnus and Terebellina trace fossil assemblages are common; (iii) shoreface: which comprises either a highly bioturbated, hummocky cross-stratified sandstone or a cross-stratified sandstone with planar lamination, symmetrical ripples and trough cross-bedding. Trace fossil assemblage in the shoreface are quite diverse, and are dominated by Thalassinoides, Ophimorpha, Skolithos and Teichichnus; and (iv) coastal-plain: which is variable in character but comprises rooted sandstones, coal, organic-rich mudstones, barren shales and cross-stratified sandstones. Organic-rich mudstones contain a low-diversity Planolites trace fossil assemblage (Table 6.1).

A regional SW (Cod Terrace) -NE (Stavanger Platform) stratigraphic cross-section through Middle-to-Upper Jurassic strata suggests that the poorly dated coastal-plain deposits of the J20? unit (Aalenian-Bajocian) are present on the Cod Terrace and in the
Norwegian-Danish and Egersund basins, but are absent on the Sørvestlandet High (Fig. 6.7). On the Cod Terrace these coastal-plain J20? deposits (Aalenian-Bajocian) are overlain by the shallow-marine deposits of the J50 unit (Oxfordian-Kimmeridgian), whereas to the east of the Sørvestlandet High, in the Egersund Basin, they are overlain by the older, coastal-plain to shallow-marine J40 stratigraphic unit (Callovian – Oxfordian). On the Sørvestlandet High, the offshore transition silty sandstones and marine shales of the J70 unit (Portlandian-Ryazanian) directly overlie the Triassic across the Mid-Cimmerian unconformity (Fig. 6.7).

We now describe the structural styles and stratigraphic succession present in the main geographical provinces using seismic cross-sections and isochron maps in Figure 6.5, structure maps in Figure 6.6, wireline-log and chronostratigraphic correlations in Figures 6.7 and 6.8, and thickness and facies variations in Table 6.2.

**Cod Terrace**

The structure and evolution of the Cod Terrace has been described in detail by Mannie et al. (2014), thus only a brief overview of the key aspects pertinent to our sub-regional analysis is presented here. The Cod Terrace is a NW-trending, normal-fault bounded terrace that is located on the eastern flank of the Central Graben (Fig. 6.6a). Salt walls are common and they typically overlie sub-salt, basement-involved, normal faults (Fig. 6.6b). Supra-salt, listric normal faults define the edges of salt walls and offset the upper part of the Triassic-Jurassic stratigraphy (Fig. 6.5a). Supra-salt, planar normal faults are developed immediately above salt walls, offset Triassic and Jurassic strata, and define graben and half-graben that contained thickened Upper Jurassic successions (Figs 6.5a and 6.6c).
On the Cod Terrace coastal-plain deposits of the J20 unit (Aalenian-Bajocian) represent the oldest penetrated Jurassic stratigraphic unit. These deposits are unconformably overlain, across an unconformity that documents c.16 Myr hiatus, by the shoreface sandstones-to-offshore mudstones of the J50-J70 (Oxfordian-Ryazanian) succession (Fig. 6.7). Detailed seismic mapping of Upper Jurassic strata (J56-to-J71 FS; Oxfordian-Kimmeridgian) indicate that they were deposited in a series of minibasins formed in response to the extensional collapse of salt walls (Fig. 6.5b; Mannie et al. 2014). An isochron map between the J66 and K10 flooding surfaces (Kimmeridgian-Ryazanian) also indicates abrupt thickening of Upper Jurassic strata across syn-depositional normal faults (Fig. 6.5d).

Sørvestlandet High

The NW-SE-trending horst of the Sørvestlandet High lies to the east of the Cod Terrace (Fig. 6.6). In contrast to the Cod Terrace, this is defined by an array of sub-parallel, relatively elongate salt walls, salt walls on the Sørvestlandet High form a polygonal network. Salt stocks are locally developed along the salt walls, and are particularly common along the eastern margin of the Sørvestlandet High (Fig. 6.6b). In a similar way to the Cod Terrace, supra-salt listric and planar normal faults define the edges of salt walls and offset the upper part of the Triassic-Jurassic stratigraphy, where they define small graben and half-graben (Fig. 6.5a).

On the Sørvestlandet High, the offshore transition to offshore shelf deposits of the J70 unit (Portlandian-Ryazanian) constitute the oldest Jurassic strata that directly overlie the Triassic across the Mid-Cimmerian unconformity, which represents a c. 74 Myr hiatus here (Fig.6.7). Detailed seismic mapping of Upper Jurassic strata indicate that the J70 unit was deposited in a series of minibasins formed in response to the extensional collapse of
Norwegian-Danish and Egersund basins

The Norwegian-Danish Basin is bounded to the west by the normal fault system at the eastern margin of the Sørvestlandet High, and to the east by the Egersund Basin (Figs 6.4 and 6.6a). In the Norwegian-Danish Basin, a mosaic of salt walls, which trend broadly NW-SE and are sub-parallel to the basement-involved normal fault systems, are present (Fig. 6.6b). Supra-salt normal faults in the Norwegian-Danish Basin are planar and are developed in Triassic-Jurassic strata, where they define small graben and half-graben (Fig. 6.5a).

To the northeast of the Norwegian-Danish Basin is the NW-SE-trending Egersund Basin, which is bound to the NE, W and E by large displacement (up to 1.5 km) normal fault systems (Lewis et al. 2013; Mannie et al. 2014) (Fig. 6.6a). In the Egersund Basin, the dominant salt structures are elongate, broadly NW-SE-trending salt walls that are situated above and trend parallel to basement-involved normal fault systems (Fig. 6.6b). Salt anticlines and salt rollers are also locally developed in the Egersund Basin (Fig. 6.6b), and supra-salt normal faults are either listric and define the margins of salt rollers, or planar and bound graben or half-graben developed above salt walls (Figs 6.5a, 6.5c and 6.6d).

In the Egersund Basin, J20 coastal-plain deposits (Aalenian-Bajocian) constitute the oldest Jurassic strata that directly overlie the Triassic across the Mid-Cimmerian unconformity, which here represents a c. 39 Myr hiatus (Figs 6.7 and 6.8). These deposits in turn are overlain by the coastal-plain to shallow-marine deposits of the J40 unit (Callovian–Oxfordian) and offshore shelf mudstones of the J50-J70 units (Oxfordian-
Ryazanian) (Figs 6.7 and 6.8). Isochron maps indicate that the Bryne and Sandnes formations (J20-J40 units) thin towards the crest of several of the salt structures in the centre of the Egersund Basin (Fig. 6.5e and Table 2). In the Egersund Basin, the J20-J70 units (Bajocian-Portlandian) are laterally continuous, unlike on the Cod Terrace and Sørvestlandet High where they are restricted to depocentres above salt walls (Figs 6.7b and 6.8).

6.6.1 Timing of rifting and salt movement

On the Cod Terrace and Sørvestlandet High, salt movement initiated in response to Early Triassic rifting, resulting in reactive rise of salt, formation of diapirs and the formation of extensional normal faults and Triassic rafts in the overburden (Mannie et al. 2014). Differential loading of the salt by non-marine clastic sediment drove subsequent salt movement in the Late Triassic (Fig. 6.5a; Mannie et al. 2014). Late Jurassic rifting reactivated the Permo-Triassic sub-salt normal faults, and resulted in stretching of the overburden, widening and extensional collapse of pre-existing diapirs, and formation of supra-salt normal faults and minibasins that provided accommodation for deposition of the Middle-to-Upper Jurassic (J20-J70) succession (Fig. 6.5b; Mannie et al. 2014). Unlike the Cod Terrace and Sørvestlandet High, in which salt movement and diapir formation strongly influenced deposition during the Triassic, salt movement in the Norwegian-Danish and Egersund basins occurred to a lesser degree in the Early Triassic, occurring predominantly in the Middle-to-Late Jurassic, and was associated with the formation of reactive diapirs that influenced the thickness of the J20-J70 succession (compare Figs 6.5b and 6.5c; Mannie et al. 2014).
6.7 TECTONO-STRATIGRAPHIC RECONSTRUCTIONS FOR THE NORWEGIAN CENTRAL NORTH SEA

In order to understand the impact of salt-influenced rifting on deposition of the net-transgressive Middle-to-Upper Jurassic (J20-J70) succession in the Central North Sea, we constructed sub-regional tectono-stratigraphic maps using the seismic and borehole data described above. Facies distributions at the point of maximum regression below age-constrained flooding surfaces were combined with the Middle-to-Upper-Jurassic structural template derived from time-structure and isochron maps to construct maps for six time intervals; Late Callovian (below FS2), Late Oxfordian (below J56FS), Early Kimmeridgian (below J62FS), Late Kimmeridgian (below J66AFS), Portlandian (below J71FS) and Late Ryazanian (below K10FS) (Fig. 6.9).

6.7.1 Late Callovian (below FS2)

Below the FS2 flooding surface, shoreface and offshore shelf deposits of the Sandnes Formation (J40 unit) are present in the Norwegian-Danish and Egersund basins, with facies becoming progressively more distal from southeast to northwest (Figs 6.8 and 6.9a). A detailed NW-SE cross-section through the Egersund Basin illustrates marked diachroneity in the timing of deposition of the Sandnes Formation (Fig. 6.8), indicating that flooding to the east of the Sørvestlandet High occurred from the northwest during the Callovian, and lasted c. 29 Ma yrs (Figs 6.8 and 6.9a). Although thickness changes that are attributable to Late Callovian, salt- and rift-driven differential subsidence are observed in the J40 strata, a general lack of facies changes suggests that sediment supply kept pace with the rate at which accommodation was generated, such that structural relief and water depth changes were relatively subdued (Figs 6.5c, 6.7 and Table 2; Mannie et al. 2014).
An exception to this trend occurs to the west of the Egersund Basin on the Sele High, where localised, footwall-attached shoreface deposits are present, indicating that faulting had a greater influence on differential subsidence across the basin-bounding fault on the eastern margin of the Sele High and resulted in development of a localised shoreline (Fig. 6.9a). The absence of the Sandnes Formation or its lateral age-equivalent on the Sør Vestlandet High and Cod Terrace suggests that perhaps they were topographic highs during the Callovian, and may have represented sediment sources that supplied sediment to the adjacent Norwegian-Danish and Egersund basins (Fig. 6.9a).

6.7.2 Late Oxfordian (below J56FS)

During the Late Oxfordian, the Norwegian-Danish and Egersund basins continued to deepen in response to ongoing extension rift-related normal faulting, and this is recorded by deposition of the marine shales of the Egersund and Tau formations above the shallow marine Sandnes Formation (Figs 6.2 and 6.9b; Vollset & Doré, 1984). Late Oxfordian deposits are absent on the Sørvestlandet High and the northern part of the Cod Terrace, and Portlandian strata directly overlie the Sørvestlandet High (Figs 6.8 and 6.9b); these stratigraphic relationships suggests that these structures were topographic highs during the Late Oxfordian (Fig. 6.9b). However, on the southern part of the Cod Terrace, Late Oxfordian shoreface deposits of the Ula/Gyda Formation were locally preserved in minibasins that formed due to extensional diapir collapse as part of a connected network of valleys, with later differential erosion and preservation of strata playing only a minor role in the present-day thickness of these strata (Figs 6.5b and 6.9b; Mannie et al. 2014).
6.7.3 Kimmeridgian (below J62FS and below J66AFS)

Rift-related subsidence and basin deepening continued into the Kimmeridgian in the Norwegian-Danish and Egersund basins, resulting in the deposition of anoxic shales of the Tau Formation (Fig. 6.2; Vollset & Doré 1984). In contrast, shallower water depths persisted on the northern part of the Cod Terrace with deeper offshore conditions to the south during the Early Kimmeridgian (below J62FS) (Fig. 6.9c). Increased accommodation above salt walls on the northern Cod Terrace during the Late Kimmeridgian (below J66AFS) resulted in shoreface deposition with offshore conditions persisting to the south (Fig. 6.9d). This increase in localised accommodation northwards on the Cod Terrace reflects the temporal and spatial variation in diapir collapse, which formed a linked network of depocentres, in contrast to the active basin-bounding faults in the Norwegian-Danish and Egersund basins which provided regional subsidence in this rift margin location (Fig. 6.9c and d). On the Sørvestlandet High and the northernmost part of the Cod Terrace, Kimmeridgian deposits are absent which suggests that perhaps they were topographic highs during the Callovian, and may have represented sediment sources that supplied sediment to the adjacent Norwegian-Danish and Egersund basins (Fig. 6.9c and d).

6.7.4 Portlandian (below J72FS)

In the Norwegian-Danish and Egersund basins, marine shales of the Sauda Formation were deposited in the Portlandian due to continued rift-climax conditions and subsidence in this rift margin location (Fig. 6.2; Vollset & Doré 1984). On the Sørvestlandet High, preservation of offshore transition deposits above salt walls reflects increased accommodation and/or increased preservation of these deposits above salt walls due to
extensional diapir fall and transgression of this long-lived topographic high (Fig. 6.9e). To the west of the Sørvestlandet High on the Cod Terrace, preservation of offshore mudstones above salt walls suggests increased accommodation in this location, probably related to increased slip on the westerly bounding fault of the Cod Terrace (Fig. 6.9e). We interpret that, at this time, the linked interconnected network of depocentres above salt walls on the Cod Terrace was connected to a similar network of supra-salt depocentres on the Sørvestlandet High (Fig. 6.9e).

6.7.5 Late Ryazanian (below K10FS)

In the Late Ryazanian, relatively deep-marine conditions had become established across the study area, resulting in the deposition of the marine shales of the Flekkefjord Formation in the Norwegian-Danish and Egersund basins and the Mandal Formation on the Cod Terrace and Sørvestlandet High (Fig. 6.9f). Regional thermal subsidence, potentially coupled with a rise in eustatic sea-level, caused flooding of the Triassic highs and Jurassic minibasins on the Cod Terrace and Sørvestlandet High (Fig. 6.9f; Mannie et al. 2014), with increased fault-related extension in the Norwegian-Danish and Egersund basins (Lewis et al. 2013). This resulted in the migration (backstepping) of sand-rich shorelines towards the basin margins and widespread deposition of offshore mudstones (Fig. 6.9f).

The Middle-to-Upper Jurassic sequence stratigraphic framework established by Partington et al. (1993) has been utilised by many workers in the central and northern North Sea, thereby enabling the work undertaken in this study to be compared to and integrated with the results of previously published regional studies, including the benchmark compilations presented in the Millennium Atlas (Fraser et al. 2002; Husmo et
al. 2003). In particular, the results of this study highlight the importance of using an integrated, high-resolution, structural-stratigraphic approach compared to other studies, which have adopted a relatively low-resolution, stratigraphy-focused approach. The higher-resolution integrated structural and stratigraphic results of our study challenge the laterally continuous, regional play fairways interpreted at relatively low stratigraphic resolution. Our palaeogeographic maps differ to those previously interpreted (see Millennium Atlas; Fraser et al. 2002; Husmo et al. 2003), in that we explicitly account for complex temporal and spatial variation of marine incursion that resulted from the influence of underlying rift- and salt-related structures.

6.8 DISCUSSION

6.8.1 Tectono-sedimentary evolution of salt-influenced rift basins

The tectono-stratigraphic evolution of rift basins that lack salt has been documented in many studies (e.g. Leeder & Gawthorpe, 1987; Gabrielsen et al. 1990; Prosser 1993; Nøttvedt et al. 1995; Ravnås & Steel, 1998; Cowie et al. 2000; Gawthorpe & Leeder 2000). During the initial stage of rifting (‘rift initiation’; sensu Prosser 1993), in a coastal plain-to-shallow-marine setting, abundant small normal faults and growth folds are developed, bounding numerous isolated depocentres (Fig. 2.1; Gawthorpe & Leeder 2000). At this time, subsidence rates are relatively low, and sediment accumulation rates may equal or exceed the rate at which accommodation is generated, thus leading to filled or overfilled basin conditions. Early syn-rift depocentres display marked variability in sedimentary fill, with sediment routing influenced by evolving fault- and fold-related patterns of differential subsidence and surface relief (Fig. 2.1b) (Leeder & Gawthorpe 1987; Ravnås & Steel 1998; Gawthorpe & Leeder 2000). Shallow platforms that develop
along the axis of the rift may become sites for shallow-marine sedimentation where they are connected to an external seaway (Mellere & Steel 1996; Ravnås & Steel 1998). As rifting continues, faults grow, interact and link, and tectonic subsidence becomes focussed within a small number of large, deep, interconnected depocentres. Fault slip rates and the rate of tectonic subsidence typically increase during the so-called ‘rift climax’ ([sensu Prosser 1993]), which accelerates marine transgression. The rift climax phase is typically characterised by a few large, through-going normal fault systems characterised by high displacement rates, such that tectonic subsidence outpaces sediment supply in hangingwall depocentres, leading to widespread transgression at depocentre margins and sediment starvation in the basin axis if major intra-basin sediment sources are absent (Fig. 2.1c; Gawthorpe & Leeder 2000).

The presence of salt in rift basins adds to the complexity of structural styles, and the temporal and spatial evolution of syn-rift basin fill ([compare Figs 2.1 and 6.10]). The growth and decay of salt diapirs can impact the stratigraphic thickness and facies distribution of syn-kinematic strata in two key ways. First, adjacent to growing diapirs and salt-cored structural highs, strata typically become thinner and contain increasingly more proximal and/or shallower-water facies (Davison et al. 1996; Alves et al. 2003; Kieft et al. 2010; Table 6.2). Second, collapse of mature diapirs may result in the formation of fault-bound crestal minibasins that provide local accommodation (Wakefield et al. 1993; Stewart & Clark 1999; Mannie et al. 2014; see also Fig. 6.5b and Table 6.2). Unlike rift basins that lack salt, within which accommodation and sediment routing is controlled by through-going, basement-involved normal faults and associated folds, the controls in salt-influenced rift basins are more complex, with sub-salt and supra-salt faults, and salt diapirs all playing key roles (Fig. 6.10). We now discuss the influence that these structures have
on the stratigraphic thickness and facies distribution during the deposition of the net-
transgressive, Middle-to-Upper Jurassic syn-rift succession:

**Salt structures**

In the Egersund Basin, extension resulted in the growth of salt structures such as salt
anticlines and salt walls, which influenced the depositional thickness of the Middle-to-
Upper Jurassic, coastal-plain to shallow-marine succession (Fig. 6.10). This is illustrated
by well data, which indicate a reduction in thickness of this unit from the flank to the crest
of the Xi salt anticline (25%) and the Alpha salt wall (45%) respectively (Fig. 6.6b and
Table 6.2; Mannie et al. 2014). However, these thickness changes were not accompanied
by a change in facies, suggesting that sediment accumulation rate outpaced the rate of
accommodation creation across growing salt structures, and that minimal surface relief
developed (Figs. 6.9 and 6.10; Mannie et al. 2014). We suggest that high sediment
accumulation rates, at least relative to the rate of salt diapir growth, was a function of the
proximity of the Egersund Basin to the Norwegian mainland during the Middle Jurassic,
which would have represented a major extra-basinal sediment source. However, in other
salt-influenced rifts, such as the Bombarral-Alcobaca sub-basin in Western Iberia, the rate
of accommodation generation outpaced sediment supply and accumulation, such that
stratigraphic thinning towards salt pillows (21-42%) is accompanied by major facies
change from shallow-marine carbonates that are restricted to the pillow crests, to
mudstone-dominated, turbidite-bearing successions on the pillow flanks and adjacent
minibasins (Alves et al. 2003).

Exhaustion of autochthonous salt coupled with ongoing extension can result in the
collapse of diapirs and formation of supra-diapir minibasins (e.g. Masson 1972; Mannie et
al. 2014; Fig. 6.10). On the Cod Terrace, formation of these minibasins during Late Jurassic regional extension resulted in the local deposition and preservation of up to 500 m of coastal-plain to shallow-marine strata. In comparison, fault-related subsidence in the Egersund Basin was responsible for development of a laterally continuous net-transgressive succession up to 150 m thick (Mannie et al. 2014). As is typical of minibasins, these successions on the Cod Terrace are thickest in the axis of the basin and thin (by up to 65%) towards the basin margins, where they may be associated with a proximal to distal facies change from the flanks to the axis of the minibasin (Fig. 6.5d and Table 6.2; Mannie et al. 2014).

Sub-salt normal faults

Through-going normal faults which breach the salt, connecting the sub- and supra-salt stratigraphy are observed on the eastern margin of the Sele High where salt is thin and, potentially, rich in non-halite lithologies (e.g. anhydrite, carbonate) that deform in a brittle rather than ductile fashion (Lewis & Jackson 2013). Formation of a large displacement, basement-involved normal fault resulted in major footwall uplift and rotation, and deposition of a hangingwall shoreline that developed downdip to the west of, and that fringed, the Sele High (Figs 6.6a and 6.9a and illustrated by the intra-basinal high in Fig. 6.10; Mannie et al. 2014). In the hanging-wall of this fault, deeper-water conditions existed (Fig. 6.10). The tectono-stratigraphic setting of the Sele High thus conforms to the predictions of tectono-stratigraphic models developed for salt-free rifts, with a strongly fault-controlled structural style resulting in abrupt, across-fault changes in thickness and facies (Gawthorpe & Leeder 2000).
Tectonically active sub-salt normal faults can also influence sub-regional subsidence and the timing of transgression as observed on the westerly bounding fault of the Sørvestlandet High (Fig. 6.9e and f). Increased accommodation in this location, probably related to increased slip on the westerly bounding fault of the Cod Terrace at this time, linked interconnected network of depocentres above salt walls on the Cod Terrace to a similar network of supra-salt depocentres on the Sørvestlandet High (Fig. 6.9e).

**Supra-salt normal faults**

Supra-salt normal faults can form relatively early in the salt-related basin history and bound the margins of rising salt walls (e.g. Cod Terrace); in other circumstances, these faults may form relatively late in response to the extensional collapse of salt structures (Figs 6.5a, 6.5b and 6.10). Faults related to rising diapirs clearly controlled bulk thickness changes in the Middle-to-Upper Jurassic succession, especially in the Egersund Basin (e.g. Figs 6.5e). The lack of well penetrations in graben that are bound by these faults means that we are unable to document whether these thickness changes are also associated with facies changes; however, based on well penetrations adjacent to salt-cored structural highs (see above), we consider that such changes are unlikely (Fig. 6.9a). This interpretation again suggests that high sediment supply and accumulation rates, which are directly linked to the basin-margin setting of the Egersund Basin, suppressed the development of significant intra-basin relief and water depth variations, and associated facies distributions, in the syn-rift succession (Fig. 6.9a).
6.8.2 Comparisons between salt- and non-salt influenced active rift basins in the North Sea

Comparisons between the salt-influenced Norwegian Central North Sea and non-salt influenced rift areas of the North Sea such as the East Shetland Basin and the Oseberg-Brage area located on the Horda Platform (Fig. 6.11a), highlight the presence of a ductile evaporite layer in decoupling the sub- and supra-salt normal fault systems. In these evaporite-free areas the structural styles are dominated by hard-linked normal fault relay systems and growth-fault monoclines, where the underlying rift topography controls accommodation, sediment distribution pathways and depositional processes.

The underlying structural topography of the Tern-Eider Horst, the Ninian-Hutton-Dunlin and the North Alwyn-Brent-Statfjord fault systems in the East Shetland Basin influenced the sediment dispersal and feeder systems along the hanging-wall of normal faults during the deposition of the syn-rift, shallow-marine, Middle Jurassic Tarbert Formation (Fig. 6.11a). As a result, the Tarbert Formation thickness is variable, thinning towards footwall crests and thickening towards the hangingwall of normal faults with enhanced erosion/non-deposition above growth fold monoclines prior to normal faults breaching the surface (Fig. 6.11b) (Davies et al. 2000; Hampson et al. 2004). While activity along the fault system resulted in thickness variations it did not result in a major facies reorganisation due to the balance between sediment supply and subsidence (McLeod et al. 2002). Structurally-controlled localised depocentres on the hangingwall of normal faults were a focus for stacked channel systems and aggradational shoreface successions, which led to the dominance of tidal rather than wave processes during deposition (Hampson et al. 2004). Similar thickness variations without major facies changes are observed in the Egersund Basin within the study area (Table 6.2). Structurally controlled localised depocentres on the hangingwall of normal faults (e.g. Egersund Basin) and
within salt-collapsed structures (e.g. Cod Terrace) documents aggradational wave-dominated shoreface successions, rather than localised embayments with strong tidal influence (see Section 6.6).

Similarly, the Ninian-Hutton-Dunlin fault system and the Oseberg Fault in the Horda Platform area were active during deposition of the wedge-shaped packages of the shoreface sandstones and the lower delta plain heteroliths of the Tarbert Formation, which thicken towards active fault segments (Fig. 6.11c; Færseth & Ravnås 1998). Strong tidal influence is observed where there is fault block rotation which created smaller sub-basins that formed embayments or estuaries with limited wave fetch. Within the Oseberg Field, the Tarbert Formation is predominantly wave-dominated with local tidal influence in transgressive units in the hangingwall of normal faults (Ravnås et al. 1997). During the rift climax phase the Oseberg fault block was subaerially exposed in up-dip areas and eroded material from the footwall crest and deposited turbidites and debris flow deposits in the hangingwalls (Færseth & Ravnås 1998). Within the study area such deeper-water deposits were not recorded. However, the large-scale trend of the Middle-to-Upper Jurassic syn-rift infill in the Northern and Central North Sea represents an overall transgression which reflects a deepening trend as the rift evolved.

Comparisons between the Northern North Sea and the Norwegian Central North Sea highlight the differences in the structural styles and stratigraphic architecture between rift basins that contain salt and those that do not (compare Figs 2.1 and 6.10). A single depositional model cannot be proposed for either case since the resulting stratigraphic architecture is dependent on the interplay between basin physiography, rate of fault- and salt-generated accommodation and sediment supply.
6.9 CONCLUSIONS

We used an integrated dataset of seismic and borehole data in order to reconstruct the Middle-to-Upper Jurassic tectono-stratigraphic evolution of the Norwegian sector of the central North Sea. Our findings illustrate that the presence of salt in rift basins influences structural style and stratigraphic architecture, which implies that established tectono-stratigraphic models for rift basins cannot be applied where such basins contain mobile salt. Key results are summarised below.

- Unlike rift basins where salt is absent and the structural style is dominated by growth folds and normal faulting, the presence of salt in a rift basin acts as a detachment between the sub- and supra-salt cover structure. The resulting structural styles can be defined by sub-salt normal faults, salt structures (e.g. salt walls, salt diapirs, salt rollers and salt anticlines) and supra-salt normal faults.

- Extensional normal faulting and the growth of salt structures during syn-rift times influence depositional thickness and facies distributions. Stratigraphic thickness reduces from the flank to the crest of salt structures. Facies distributions are a function of accommodation creation and sediment supply. Where sediment supply exceeds accommodation creation, no change in facies is observed during the growth of structures (e.g. Egersund Basin). In contrast, where creation of accommodation outpaces sediment supply localised shorelines are developed on the footwall of normal faults with deeper water conditions in the hangingwall (e.g. easterly bounding fault of the Sele High).

- Similar to rifts that lack salt, the timing of flooding is diachronous and localised topographic highs can result in the temporal and spatial variation of marine incursion.
The results presented in this paper highlight the distinct depositional evolutions of rift basins in the presence and absence of salt. During syn-rift times, depositional facies distributions vary in relation to subsidence rates, sediment supply and basin physiography.

6.10 ACKNOWLEDGEMENTS

The authors express their gratitude to the Norwegian Petroleum Directorate and PGS for providing access to seismic data. Centrica Energi Norge and the Commonwealth Commission are gratefully acknowledged for funding and support for this study, and Schlumberger Limited for use of Petrel software via an academic software donation. We extend our gratitude to Mike Charnock for biostratigraphic analysis and related discussions. Alastair Fraser and Adrian Hartley are acknowledged for their critical reviews and discussions of this manuscript.

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Fig. 6.1: Regional geographic setting of the North Sea trilete rift system, showing the limit of the Zechstein Supergroup in the North Permian Basin (taken from Smith et al. 1993). (VG: Viking Graben; MF: Moray Firth; CG: Central Graben)
## System Lithostratigraphy

<table>
<thead>
<tr>
<th>System</th>
<th>Series</th>
<th>Stage</th>
<th>Genetic Unit</th>
<th>Flooding Surfaces after Partington et al. (1993)</th>
<th>Lithostratigraphy</th>
<th>Main tectonic events</th>
<th>Interpreted surfaces</th>
</tr>
</thead>
<tbody>
<tr>
<td>Triassic</td>
<td>Early</td>
<td>Ryazanian</td>
<td>J70</td>
<td>J76 _Aestomphalus J76 _Kochi J74 _Anguiformis J73 _Angulifer J72 _Oklosoma J71</td>
<td>Shetland Gp.</td>
<td>Compression; Inversion; diapir rejuvenation</td>
<td>Top Mantal K10FS</td>
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<tr>
<td></td>
<td>Late</td>
<td>Portlandian</td>
<td>J60</td>
<td>J66 _Aestomphalus J65 _Oklosoma J64 _Angulifer J63 _Anguiformis</td>
<td>J62 _Baybi J61 _Rijk J60 _Angulifer</td>
<td>Extensional diapir collapse and formation of Jurassic minibasins on the Cod Terrace and High; reactive diapir rise in the Norwegian-Danish and Egersund basins; supra-salt normal faults</td>
<td>J62FS J66AFS</td>
</tr>
<tr>
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<td>Kimmeridgian</td>
<td>J50</td>
<td>J56 _Aestomphalus J55 _Oklosoma J54 _Angulifer J53 _Anguiformis</td>
<td>J52 _Baybi J51 _Rijk J50 _Angulifer</td>
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<td>J52FS Top Egersund J56FS</td>
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<td>J40</td>
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<td>J30</td>
<td>J36 _Baybi J35 _Rijk J34 _Angulifer J33 _Anguiformis</td>
<td>J32 _Baybi J31 _Rijk J30 _Angulifer</td>
<td></td>
<td>J32FS J36FS</td>
</tr>
<tr>
<td></td>
<td>Late</td>
<td>Aalenian</td>
<td>J10</td>
<td>J16 _Baybi J15 _Rijk J14 _Angulifer J13 _Anguiformis</td>
<td>J12 _Baybi J11 _Rijk J10 _Angulifer</td>
<td>Extension and reactive diapir rise on the Cod Terrace and Sarvesandet High; Triassic rift tectonics; listric normal faults at the edge of salt diapirs</td>
<td>Top Zechstein Top Rotliegend</td>
</tr>
<tr>
<td></td>
<td>Early</td>
<td>Toarcian</td>
<td>J00</td>
<td>J06 _Baybi J05 _Rijk J04 _Angulifer J03 _Anguiformis</td>
<td>J02 _Baybi J01 _Rijk J00 _Angulifer</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

**LEGEND**
- chalk
- evaporites
- non-marine
- coastal plain
- shallow-marine sandstones
- shallow-marine siltstones
- marine shales

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**Fig. 6.2:** Regional tectono-stratigraphic chart for areas between the Cod Terrace and the Norwegian-Danish Basin, on the eastern flank of the Central Graben (modified from Vollset & Doré 1984), showing major flooding surfaces documented in regional literature (Partington et al. 1993), main tectonic events, and surfaces interpreted in seismic, well-log and core data from the study area. See Figure 6.1 for location of cross-section.
local sediment input point (into N-S cross-section)

Coastal-plain Shallow-marine sandstones Offshore shales Non-marine Pre-Jurassic

Fig. 6.3: SW-NE chronostratigraphic cross-section illustrating diachronous deposition of net-transgressive Upper Jurassic strata in the Central North Sea (modified from Partington et al. 1993 and Hampson et al. 2009). See Figure 6.1 for location of cross-section.
Fig. 6.4: Simplified structure map illustrating major present day topographic features in the study area, and the distribution of seismic and well data used for this study. (SH: Sele High; EB: Egersund Basin; LN: Lista Nose; N-DB: Norwegian-Danish Basin; SVH: Sørvestlandet High; CT: Cod Terrace; CG: Central Graben)
Sørvestlandet
High
Stavanger Platform
Egersund Basin
Norwegian-Danish Basin

Cod Terrace Egersund Basin

change in thickness across fault

25
Thickness TWT (ms)

120

2/01-6

Fig. 6.5b

supra-salt grabens

Fig. 6.5c

LEGEND
Rotliegend  J66A FS (Late Kimmeridgian)
Zechstein  J62 FS (Early Kimmeridgian)
Triassic - Jurassic  J32? FS (Early Bathonian)
near top Jurassic  sub-salt normal faults
Top Triassic  supra-salt normal faults
Fig. 6.5: Interpreted SW-NE regional line of section from the Cod Terrace to the Stavanger Platform illustrating the presence and distribution of sub-salt normal faults, salt structures, and supra-salt normal faults. (b) Cosine of instantaneous phase section from the Cod Terrace showing detailed mapping and onlap of Jurassic reflectors on to the margins of minibasins developed above salt walls (located on Fig. 6.5d). (c) Cosine of instantaneous phase section from the Egersund Basin showing laterally continuous Upper Jurassic reflectors that record gradual lateral thinning from the flank to the crest of salt structures (located on Fig. 6.5e). (d) Isochron map between the J66FS-K10FS flooding surfaces on the Cod Terrace showing stratigraphic thickness variations across supra-salt normal faults (Figs 4 and 5b). (e) Isochron map between the near top Jurassic-Top Triassic surfaces showing thickness changes between supra-salt grabens in the Egersund Basin (Figs 6.4 and 6.5c).
Fig. 6.6: Two-way time (TWT) structure maps for: (a) top Rotliegend, illustrating sub-salt, basement structures; (b) top Zechstein, illustrating various types of salt structures, including salt walls, salt diapirs, salt rollers, salt pillows and salt welds; (c) top Mandal, illustrating supra-salt, cover structures on the Cod Terrace (located on Fig. 6.6b); and (d) top Egersund illustrating supra-salt cover structures in the Egersund Basin (located on Fig. 6.6b). (SH: Sele High; EB: Egersund Basin; LN: Lista Nose; N-DB: Norwegian-Danish Basin; SVH: Sørvestlandet High; CT: Cod Terrace; CG: Central Graben)
Fig. 6.7: (a) SW-NE wireline-log correlation panel and (b) corresponding chronostratigraphic cross-section across the Cod Terrace, Sørvestlandet High, Norwegian-Danish Basin, Egersund Basin, and Stavanger Platform illustrating the temporal and spatial variation of Middle-to-Upper Jurassic net-transgressive strata of the Norwegian Central Graben. See Figure 6.4 for location.
Fig. 6.8: (a) NW-SE wireline-log correlation panel and (b) corresponding chronostratigraphic cross-section across the Egersund Basin, illustrating the temporal and spatial variation of Middle-to-Upper Jurassic net-transgressive strata of the Norwegian Central Graben. See Figure 6.4 for location.
LEGEND

- sub-salt basin bounding normal faults
- salt walls
- exposed post-salt Triassic strata
- offshore shales
- offshore transition
- shoreface

(a) eastern flank of Sels High

(b) southern Cod Terrace

(c) east of Sørvestlandet High
Fig. 6.9: Series of palaeogeographic maps that illustrate the tectono-stratigraphic evolution of the Middle-to-Upper Jurassic succession in the Norwegian North Sea: (a) Late Callovian (below FS2); (b) Late Oxfordian (below J56FS); (c) Early Kimmeridgian (below J62FS); (d) Late Kimmeridgian (below J66AFS); (e) Portlandian (below J71FS); and (f) Late Ryazanian (below K10FS). (SH: Sele High; EB: Egersund Basin; LN: Lista Nose; N-DB: Norwegian-Danish Basin; SVH: Sørvestlandet High; CT: Cod Terrace)
Fig. 6.10: 3D schematic illustrating the structural styles and depositional model for the salt-influenced Norwegian Central North Sea rift basin.
Fig. 6.11: Examples of non-salt influenced syn-rift in the Northern North Sea: (a) Location map of the East Shetland Basin and the Horda Platform (Færsetha & Ravnås, 1998; (b) Thickness variations from the hangingwall to the footwall in fault-controlled accommodation in the Statfjord East Field (Davies et al. 2000); (c) Syn-rift faulting in the Oseberg Field controlling accommodation in the hangingwall of normal faults, resulting in thickness variations and tidal processes (Færsetha & Ravnås, 1998).
### Summary of facies associations in the Upper Jurassic Ula, Gyda and Sandnes formations (Mannie et al. in press)

<table>
<thead>
<tr>
<th>Facies Association</th>
<th>Brief Description</th>
<th>Trace fossil assemblage</th>
</tr>
</thead>
<tbody>
<tr>
<td>Offshore shelf</td>
<td>Consists of a silty mudstone with relic patches of planar lamination and a pinstriped siltstone with flattened lenses and low-amplitude ripple forms composed of silt or very fine-grained sand with common bioturbation. Belemnites are commonly observed.</td>
<td>Anconichnus, Chrondrites with rare Planolites, Terebellina, and Teichichnus</td>
</tr>
<tr>
<td>Offshore Transition</td>
<td>Comprise of a bioturbated and hummocky cross-stratified carbonaceous siltstone to very fine-grained sandstone which are subrounded-rounded and well-sorted. Presence of bivalve/shell fragments.</td>
<td>Teichichnus, Terebellina, Trichinus, Anconichnus, Planolites, Palaeophycus, Chrondrites, rare presence of Rosselia</td>
</tr>
<tr>
<td>Shoreface</td>
<td>Consists of a highly bioturbated, very fine- to fine-grained, well sorted sandstone which are separated by hummocky cross-stratified bedsets or a very-fine to medium-grained, well to moderately sorted sandstone in which planar lamination, symmetrical ripples and trough cross-bedding are noted. Sparse presence of shell fragments, crinoids and mud clast are recorded.</td>
<td>Thalassinoides, Ophiomopha, Teichichnus, Planolites, Palaeophycus, Skolithos, Siphonichnus, Anconichnus, Aulichnites, Rosselia</td>
</tr>
<tr>
<td>Coastal Plain</td>
<td>Comprise of a fine-to medium-grained rooted sandstone, coal, bioturbated organic-rich mudstones with sandstone laminations, ripples and synaeresis cracks, barren fissile shales, and an interbedded, cross-stratified coarse-to fine-grained, well- to poorly-sorted, sandstone.</td>
<td>Bioturbated organic mudstones comprise of Planolites, rare Rosselia, Palaeophycus, Phycosiphon, Cylindrichnus</td>
</tr>
</tbody>
</table>
Table 6.2: Summary of the thickness and facies variations for the various sub-, supra- and salt structural styles in the study area. See Figure 6.5 for structural styles identified.
CHAPTER 7
CONCLUSIONS

The results of the study presented in Chapters 3-6 are summarised here in a series of concluding remarks that are linked back to the overall aims of the thesis, (Chapter 1, section 1.3) and the key unanswered research questions that this thesis aimed to address (Chapter 2, section 2.4).

7.1. STRUCTURAL STYLES IN A SALT-INFLUENCED RIFT BASIN

The Zechstein Supergroup in the Central North Sea rift basin is sufficiently thick and mobile that it locally acts as an intra-stratal detachment between sub- and supra-salt cover strata (e.g. Egersund Basin and Cod Terrace), similar to that recorded in other salt-influenced rift basins (e.g. Red Sea, Persian Gulf and Mediterranean). In these locations, deformation in the sub- and supra-salt may be fully decoupled. In contrast, where the salt is thin and/or largely immobile, deformation in the sub- and supra-salt cover strata may be fully coupled by through-going normal faults (e.g. Stavanger Platform), thus resulting in structural styles that are similar to those documented in rift basins that lack salt (e.g. North Sea Viking Graben). This study aimed to characterise these sub-, supra- and salt structural styles in this salt-influenced rift basin, which can be defined by: (i) basement-involved, sub-salt involved normal faults, which formed in response to Permo-Triassic and Jurassic-to-Early Cretaceous rifting; (ii) salt structures (e.g. salt walls, salt pillows, salt anticlines) that formed in response to flow of the Zechstein salt; and (iii) supra-salt restricted normal faults, which formed in response to a combination of flow of the Zechstein salt and Jurassic-to-Early Cretaceous rifting.
Multiple stages of rifting, in addition to salt thickness and mobility, are at least partly responsible for the variability of the structural styles documented in the eastern Central Graben. For example, on the Cod Terrace and Sørvestlandet High, salt flow was triggered by Early Triassic rifting, resulting in reactive rise of salt and the formation of salt walls and stocks, and extensional normal faults and Triassic rafts in the overburden. Differential loading of the salt by non-marine clastic sediments drove subsequent salt movement in the Middle-to-Late Triassic. Late Jurassic rifting reactivated the Permo-Triassic sub-salt normal faults and resulted in stretching of the overburden, extensional collapse of pre-existing diapirs, and formation of supra-salt normal faults and minibasins that provided localised accommodation for deposition of the shallow marine, Upper Jurassic succession. In contrast, major salt movement in the Norwegian-Danish and Egersund basins did not occur until the Middle-to-Late Jurassic, and was associated with the formation of reactive diapirs that influenced the deposition of the laterally extensive, coastal plain-to-shallow marine, Middle-to-Upper Jurassic succession. The difference in timing of rifting during the Middle-to-Upper Jurassic resulted in diachronous flooding that occurred earlier in the Egersund Basin (i.e. Callovian) than on the Cod Terrace (i.e. Late Oxfordian).

7.2. SEDIMENTOLOGY AND SEQUENCE STRATIGRAPHY OF THE NET-TRANSGRESSIVE, MIDDLE-TO-UPPER JURASSIC SUCCESSION

The transition from syn- to post-rift strata in a coastal-plain to shallow-marine environment typically records a net-transgressive succession. This study aimed to document the sedimentology and sequence stratigraphic architecture of the syn- to post-rift, Middle-to-Upper Jurassic stratigraphic succession. Sedimentological core analysis of the Bryne, Sandnes and Ula/Gyda formations identified thirteen facies which have been grouped into five facies associations; offshore, offshore transition, lower shoreface, upper
shoreface and coastal plain. Facies analysis integrated with biostratigraphic data and interpretation of wireline-log data has enabled the identification of regionally and locally correlatable flooding surfaces. These surfaces divide the Middle-to-Upper Jurassic succession into a series of stacked, upward-shallowing parasequences that define an overall net-transgressive, aggradational to retrogradational stacking pattern.

In the Egersund Basin, the stratigraphic framework of the Bryne and Sandnes formations is defined by four parasequences that are aggradationally to retrogradationally stacked to form a net-transgressive succession that is up to 150 m thick, at least 20 km in depositional strike (SW-NE) extent, and ≥70 km in depositional dip (NW-SE) extent. The close proximity of the Egersund Basin to the heavily eroded structural high of the Stavanger Platform can account for high sediment influx, shoreline progradation over large distances, and the anomalously large depositional dip extent (up to 70 km) of these shallow marine sandstones. In contrast, the Ula/Gyda Formation on the Cod Terrace and Sørvestlandet High is confined to diapir collapse-related, supra-wall minibasins, within which up to 500 m thick, net-transgressive, shallow-marine sandstone reservoirs successions were deposited. Proximal-to-distal facies variations, from shoreface to offshore shelf, occur over scales larger than individual minibasins and are not confined to areas of localized sandstone subcrop, suggesting that the minibasins formed a linked network of depocentres supplied by regional sediment-routing systems.

7.3. IMPACT ON RESERVOIR DISTRIBUTION AND CHARACTER
Rift-related normal faulting and salt structures influence shallow-marine sedimentation; for example, syn-kinematic strata thin, and a decrease in water depth inferred from facies transitions are noted towards the crest of salt structures and on the footwall of normal
faults. This study aimed to quantify such and to determine the influence of syn-
depositional structures on reservoir distribution in salt-influenced rifts.

In the Egersund Basin, laterally extensive, Middle-to-Upper Jurassic stratigraphic units of the Bryne and Sandnes formations (Units A-E; see Chapter 3) change thickness across salt structures and supra-salt normal faults, but facies variability is limited to areas where across-fault expansion indices are <2.0. This suggests that sediment supply outpaced accommodation generation across actively growing structures during deposition of these stratigraphic sequences, at least within this part of the basin. The close proximity of the Egersund Basin to the Norwegian mainland can account for a high, extra-basinal sediment supply. Abrupt lateral facies change linked to differential tectonic subsidence only occurs across through-going sub-salt normal faults, in locations where the Zechstein salt is relatively thin (Units C and D; see Chapter 3). In this case, shoreface deposits on the slowly subsiding Sele High pass abruptly across a major normal fault into offshore mudstones deposited in the more rapidly subsiding Egersund Basin. This lateral change indicates that accommodation generation locally outpaced sediment supply in the fault hangingwall. In the Egersund Basin, shallow marine reservoirs are more laterally continuous than predicted in other tectonically active, salt-influenced basins. Growth of these structures did not result in abrupt, across-fault facies changes, thus reducing risk associated with reservoir distribution and quality in the basin. In contrast, on the Cod Terrace and Sørvestlandet High, the Middle-to-Upper Jurassic Ula/Gyda Formation was deposited locally in minibasins occurring above salt walls (Chapter 4). These minibasins formed a linked network supplied by regional sediment-routing systems and, unlike the Egersund Basin, this complexity places a higher risk for Upper Jurassic reservoir presence away from salt walls. The results of this study highlight the variability in reservoir distribution that can occur during the growth of fault- and salt-related structures during
early syn-rift, shallow-marine sedimentation. It is worth noting the role that salt plays in modifying the tectono-stratigraphic evolution of rift basins; in fact, this study suggests that existing tectono-stratigraphic models (e.g. Gawthorpe and Leeder 2000) for salt-free rifts should be applied with caution to basins where salt is present. The resulting stratigraphic architecture and reservoir distribution is dependent on accommodation creation, sediment supply and basin physiography.

7.4. FURTHER WORK

- Balanced structural restoration in 2D or possibly 3DMove, which utilise regional seismic profiles from large 3D seismic volumes or long line-length 2D seismic data, the latter being likely to give better imaging of sub-salt structure and stratigraphy. These restorations should focus on transects that go from the flanks to the axis of the North Permian Basin, and they should focus on constraining and validating the structural geometries of normal faults and salt structures which would provide detailed insight into the syn-depositional topographic expressions during deposition of the coastal-plain to shallow-marine Middle-to-Upper Jurassic succession. This would also allow any regional controls on accommodation to be constrained.

- In order to improve our understanding of the sediment-routing systems during the evolution of the Upper Jurassic minibasins in the vicinity of the Cod Terrace and Sørvestlandet High, bulk petrographic and heavy mineral analysis should be undertaken with the aim of establishing the provenance of sediment source areas and transport routes in the different stratigraphic intervals and palaeogeographic
locations which will provide further insight into the understanding of reservoir
distribution.

• Well penetrations are limited and seismic imaging within the Triassic succession is
of low quality. In order to improve our understanding of the tectono-stratigraphic
evolution during the Triassic, well penetrations and higher-resolution 3D seismic
surveys, from the Western Shelf and Central Graben areas of the UK Central North
Sea should be studied. These areas will provide additional information on the
Triassic in which the various conditions required to initiate density-driven
subsidence and minibasin formation can be constrained. They are also structurally
similar to the study area and will serve as useful analogues for understanding the
Triassic.
References


