**SEDIMENTOLOGY AND SEQUENCE STRATIGRAPHY OF SHALLOW-
MARINE SEDIMENTS IN A SYN-RIFT SETTING: HUGIN FORMATION,
SOUTH VIKING GRABEN**

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ABSTRACT

Fault growth and linkage within rift basins generates localised uplift and subsidence, which may exert significant control over syn-rift sedimentation. The shallow-marine Hugin Formation within the South Viking Graben, a salt-influenced rift basin, is studied as an example of such sedimentation. The principal aims of this thesis are to: (1) characterise the sedimentology and sequence stratigraphy of the succession, (2) investigate structural development within the graben by analysis of the hangingwall dip slope, and (3) establish the impact of syn-depositional structural development on shallow marine sedimentation.

Sedimentological analysis of the Hugin Formation has recognised fifteen facies that are grouped into six facies associations: bay-fill, shoreface, fluvio-tidal channel fill, mouth bar, coastal plain and offshore open marine. Combining the results of this analysis with biostratigraphy, the studied succession is subdivided into a series of units delineated by regionally correlatable maximum flooding surfaces. Within the stratigraphic framework delineated by these surfaces each of the main facies associations are developed coevally in distinct and different geographical locations. Higher order flooding surfaces, transgressive surfaces and candidate sequence boundaries can also be interpreted within this framework which are geographically localised.

Structural analysis of the hangingwall dip slope identified a series of distinct structural features that were evolving contemporaneous with Hugin Formation deposition. To the north, the gravity-driven Gudrun-Brynhild fault array initiated as a series of distinct segments which subsequently propagated and linked. Further south, the salt-cored highs at Dagny and Alpha Central were also developing. Palaeogeographical localisation of facies associations is interpreted to reflect deposition within these different structural locations along with proximity to the axial fluvial feeder system to the south of the basin. Sediment supply via wave-driven longshore drift from the north of the basin and local re-working of structurally controlled palaeohighs was also important.

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CHAPTER 1

Introduction

1.1 Thesis rationale

The sedimentology and facies distribution of shallow marine sediments can be complicated due to the interaction during deposition of various physical processes such as waves, storms, tides and river-mouth processes. In syn-rift settings, the depositional system is further complicated by spatial and temporal variations in localised uplift and subsidence, which make facies prediction problematic (e.g. Gawthorpe and Leeder, 2000). Analysis of facies character, facies distributions and stratigraphic architecture within these structural settings provides important insights into the topography and water depth in the area during deposition. Consequently in such settings a robust sedimentological and stratigraphic model may provide important insights into the timing and evolution of structures during rifting.

This study has focussed on the transgressive, shallow-marine Hugin Formation, deposited during the Middle Jurassic in the South Viking Graben. The Hugin Formation consists of shallow marine sediments deposited during an overall transgression that marked the southerly migration of the Brent Delta system up the axis of the embryonic South Viking Graben (Rathey and Hayward, 1993; Husmo *et al.*, 2003). This transgression coincided with the onset of rift-related faulting and halokinesis in the area (Cockings *et al.*, 1993; Sneider, *et al.*, 1995; Milner and Olsen, 1998). The Hugin Formation diachronously overlies the coastal plain deposits of the

Sleipner Formation and is diachronously overlain by the offshore marine mudstones of the Heather Formation (Vollset and Doré, 1984).

Initial sedimentological analysis of the Hugin Formation highlights the variability of the deposits within a small area of the South Viking Graben. It is hypothesised that the facies distribution in the area is controlled by early syn-rift faulting and halokinesis. A better understanding of this variability is necessary to test this initial hypothesis, and to improve facies prediction in areas away from well control. This study has important implications for exploration and production within the region, as well as providing more generic insights into the relationship between sedimentology, sequence stratigraphy and structural development that can be applied to other rift basins and areas of halokinesis.

1.2 Thesis Aim

The aim of this thesis is to determine the influence of rift-related faulting and halokinesis on the sedimentology and sequence stratigraphy of the syn-tectonic Hugin Formation and, as a corollary, to utilise the sedimentology and sequence stratigraphy of the Hugin Formation to provide insights into the Middle Jurassic structural development of the South Viking Graben. Three-dimensional seismic, wireline-log, core and biostratigraphical data are integrated with outcrop observations to examine the impact of structural development on sedimentation.

1.3 Thesis Objectives

The objectives of this investigation can be divided into three main categories, outlined below.

1.3.1 Sedimentology and sequence stratigraphy

- Establish a regional depositional model for the transgressive Hugin Formation through the recognition and interpretation of facies and facies associations
- Identify key sequence stratigraphic surfaces within the succession to establish a sequence stratigraphic framework for the studied area.
- Investigate the facies and facies architecture observed within an outcrop analogue not influenced by syn-depositional structural development (Cretaceous Almond Formation, SW Wyoming, USA), and apply the lessons of this study to the subsurface Hugin Formation.

1.3.2 Structural development of the hangingwall dipslope

- Examine the spatial and temporal evolution of rift-related faults on the hangingwall dipslope of the South Viking Graben during the Jurassic.
- Examine the influence of halokinesis of Zechstein Group evaporites on Jurassic structural development of the South Viking Graben.

1.3.3 Impact of structural development on shallow marine sedimentation

- Investigate the influence of syn-depositional structural development on shallow marine facies and facies architecture.
- Reconstruct the palaeogeographical evolution of the South Viking Graben utilising new insights developed during this study.

1.4 Approach and Methodologies

In order to address the specific aims of this thesis, several approaches have been used; (i) sedimentological and sequence stratigraphical analysis of the Hugin Formation using core and well-log data, combined with insights from the Almond Formation outcrop analogue; (ii) analysis of 3D seismic data to establish the timing of structural development within Quadrant 15 of the Norwegian South Viking Graben; (iii) sedimentological and sequence stratigraphic analysis of the upper part of the Almond Formation using outcrop data. A full discussion of the approach and methodologies is given in chapters 4 and 6.

1.5 Thesis Outline

This thesis comprises ten chapters, each one dealing with a specific aspect of the research project. Chapters 1 to 4 form the introduction and background of this study whilst chapters 5 to 9 discuss the analysis undertaken as part of this research project. Each of these latter chapters is intended to form a paper, consequently the PhD supervisors (Chris Jackson and Gary Hampson) have been invited as co-authors to review and comment on the work contained within them.

Chapter 2 comprises a review of literature pertinent to this study, focussing on three key themes: (i) rift basin development, including fault-array evolution and syn-rift sedimentation; (ii) mechanisms of salt movement and their impact on syn-tectonic sedimentation; (iii) transgressive shallow marine depositional systems. Key examples are discussed within each of these themes that are referred to within later

chapter discussions. **Chapter 3** provides a detailed introduction to the geological setting of the subsurface South Viking Graben study area, the main focus of this research project. This geological setting is summarised in the individual introductions to the main, subsurface study chapters. **Chapter 4** summarises the database available for the subsurface component of the study (3D i.e. seismic, well and biostratigraphical data), and briefly outlines the methodologies used.

Chapter 5 documents the sedimentological character of the Hugin Formation and discusses the sequence stratigraphy of the unit throughout the entire subsurface study area. The results discussed in this chapter form a key component of the study and consequently the relevant results are summarised within later chapters. These results were presented at the Petroleum Geology Conference in April 2009, and a manuscript based on this chapter is currently in review for inclusion within the special publication associated with this conference.

Chapter 6 is focussed on the Almond Formation, proposed as an outcrop analogue to the Hugin Formation. Previous work undertaken on the formation is synthesised before the results of this study are described. The sedimentology and sequence stratigraphy of the Almond Formation are described and interpreted. Observations from this outcrop analogue study were then applied to help constrain the detailed subsurface analysis presented in chapters 7 and 8. **Chapter 7** uses seismic data to investigate the temporal and spatial evolution of rift-related faults within the Sigrun-Gudrun area, before integrating well data to review the impact of syn-depositional normal faulting on depositional patterns within the Hugin Formation in this area. **Chapter 8** uses seismic data to investigate the temporal and spatial evolution of salt-

cored structures and halokinesis within the Dagny-Alpha Central area, before integrating well data to review the impact of this structural development on depositional patterns within the Hugin Formation in this area.

Chapter 9 serves as a regional synthesis for the study, integrating the results of chapters 5, 7 and 8 with previously published regional work and additional data from the UK sector to provide an updated model for the palaeogeographical reconstruction of the South Viking graben during the Middle Jurassic. **Chapter 10** comprises a series of concluding remarks that relate directly to the aims and objectives set out in this chapter.

CHAPTER 2

Study Background

The recognition of tectonic signatures in the rock record provides important information about the evolution of basins through time (e.g. Leeder and Gawthorpe, 1987; Frostick and Steel, 1993). Tectonic activity will affect topography within a basin and this, together with the sediment supply and eustasy, will affect the development and distribution of facies. Furthermore, the structural development of a basin may be linked to fluctuations in relative sea-level, promoting deposition of regressive, aggradational or transgressive sediment packages depending on the balance between local sedimentation rate and accommodation generation. The Hugin Formation was deposited within the South Viking Graben of the North Sea in the mid-Jurassic during a time of both extensional normal faulting and halokinesis (e.g. Cockings *et al.*, 1993; Husmo *et al.*, 2003). The development of these structures, in addition to eustasy, may be expected to have controlled the sedimentological and sequence-stratigraphic development of the Hugin Formation (Cockings *et al.*, 1993; Varadi *et al.*, 1998).

In this chapter models of rift evolution and halokinesis are discussed along with the resultant impact of these processes on syn-deformational sedimentation. Several key examples are highlighted from shallow-marine depositional settings analogous to that of the Hugin Formation. Transgressive deposits with similar sedimentological and sequence stratigraphic character are also discussed.

2.1 Rift Basin Development

2.1.1 Mechanisms of rifting and regional structural style

At the crustal scale, rift basins form at sites of extension when the upper crust is extended. This extension is accommodated by the development of normal faults, whilst the lower crust and lithospheric mantle stretch by plastic deformation (e.g. Allen and Allen, 1990; Kuszniir and Ziegler, 1992; Ravnås and Steel, 1998). Rift basins, therefore, overlie thinned crust and are associated with negative Bouguer gravity anomalies, high heat flow and in some cases, extensive volcanic activity (Allen and Allen, 1990).

Regionally, rifts are typically characterised by a series of tilted, elongate, grabens and half-grabens (Fig. 2.1) (Allen and Allen, 1990; Ravnås and Steel 1998). Within these half-grabens, the polarity of the basin-bounding faults and the half-grabens they control can be seen to alternate along the rift axis (Fig. 2.1) (Allen and Allen, 1990; McClay *et al.*, 2002; Schlische and Withjack, 2008). Two models have been proposed to account for the changes in fault and half-graben polarity along-strike: the hard-linked, strike-slip, transfer fault model (e.g. Lister *et al.*, 1986) and the soft-linked, accommodation zone model (e.g. Morley, 1994). The amount of stretching may increase towards the rift-basin axis, with the major basin-bounding faults showing an increase in throw (Ravnås and Steel, 1998). In addition, fault spacing also commonly shows a systematic variation and decreases towards the rift axis (Ravnås and Steel, 1998).

Figure 2.1 Conceptual model of displacement transfer in rift systems. This highlights the characteristic geometry of rift basins. Within these half-grabens, the polarity of downthrown sections reverses over transform faults which segment the rift system. (Redrawn from McClay et al., 2004)

2.1.2 Normal fault structure and growth

Extensional fault systems are recognised to comprise a series of constituent segments that grow through a process of nucleation, interaction and linkage; consequently, faults exhibit a complex, step-like growth path (e.g. Peacock and Sanderson, 1991; Anders and Schlische, 1994; Cartwright *et al.*, 1995; Gawthorpe and Leeder, 2000). During fault nucleation, isolated segments grow towards each other through radial propagation (Fig. 2.2). Displacement-to-length relationships across these segments may be asymmetrically skewed towards the interacting fault tip (e.g. Cartwright *et al.*, 1995; Gawthorpe and Leeder, 2000). As individual segments begin to overlap, relay ramps develop in the area of separation and then breach during continued displacement, allowing individual segments to link. A fault zone composed of two recently linked segments will be under-displaced with respect to its new length. During the final growth stage a through-going fault zone forms through relay ramp breaching and hard linkage. When plotting displacement against length for the through-going fault, former segment boundaries and breached relay ramps will be highlighted by displacement lows.

Displacement, uplift and subsidence rates vary along the length of individual fault segments. Typically, the greatest displacement is seen at the segment centre, although interaction between segments can cause asymmetrical slip distribution along the segment (Willemse *et al.*, 1996; Cowie *et al.*, 2000; Peacock, 2002). Displacement then decreases outwards towards the fault tips (Gawthorpe *et al.* 1997; Cowie *et al.*, 2000). During fault growth, and as fault-segment tips propagate upwards, a fault-parallel hanging-wall fold may develop (Fig. 2.3) (Gawthorpe and

Stage 1: Fault Initiation

Nucleation phase with a series of isolated small fault segments

Stage 2: Interaction and linkage

Phase of enhanced growth via propagation of fault-segment tips

Stage 3: Through going fault zone

Figure 2.2 Model of fault growth by segment linkage. Three fault segments link over time to form one through going fault zone. This defines rift-initiation through to rift-climax (redrawn from Gawthorpe and Leeder, 2000).

1. Basement fault propagating

2. Fault tip fold develops

3. Amplification of fold

4. Fault breaks surface

Figure 2.3: Extensional Forced Fold development above a propagating normal fault. As the fault tip propagates a fault-parallel hanging-wall fold may develop. These folds are comparatively short-lived monoclines before the fault breaches the surface. Within the cover, high-angle to locally reverse secondary faults and fractures develop. Of key importance is the impact of this structural development on syn-rift depocentres. Syn-rift I package thickens away from the fault due to the overlying fold (2) and becomes progressively rotated as the fold continues to grow (3) whilst the syn-rift II package thickens towards the fault (Redrawn from Gawthorpe and Leeder, 2000).

Leeder, 2000; Withjack and Callaway, 2000; Richardson *et al.*, 2005). These “extensional forced folds” are relatively short-lived monoclines which exist before the fault breaches the surface (Gawthorpe and Leeder, 2000; Withjack and Callaway, 2000).

This variation in displacement rates has implications for stratal stacking patterns and the expression of key stratal surfaces along a fault segment (Gawthorpe *et al.* 1997; Ravnås and Steel, 1998). For example, low displacement or subsidence rates at fault tips coupled with the development of growth folds will enhance marine and subaerial surfaces associated with relative sea-level fall (possible sequence boundary development). In contrast, at the centre of the fault segment, the comparatively greater subsidence rates may outpace the relative sea-level fall resulting in a correlative conformity or flooding surface within the coeval section (Gawthorpe *et al.* 1997; Ravnås and Steel, 1998).

2.1.3 Fault array evolution

Prosser (1993) recognised four stages of rift evolution: rift initiation, rift climax, immediate post-rift and late post-rift. During rift initiation the rate of fault displacement is relatively low, but this increases during the rift climax phase. The model also implies that an increase in the rate of extension is required to generate the increase in displacement rate seen during the rift climax phase. However, other workers (Gupta *et al.*, 1998; Cowie *et al.*, 2000; Dawers and Underhill, 2000; McLeod *et al.*, 2002; Peacock, 2002; Gawthorpe *et al.*, 2003) explain this relationship through fault array evolution. In this model the rate of extension stays constant and it is the linkage of fault segments, and the focusing of displacement on

these larger structures that leads to higher fault displacements (Fig. 2.2) (Gawthorpe *et al.*, 1997; Cowie *et al.*, 2000; Peacock, 2002).

The initial stage of fault array evolution (rift initiation) is thought to involve a nucleation phase consisting of a series of isolated small fault segments and associated hanging-wall sub-basins (Cowie 1998; Cowie *et al.*, 2000; Gawthorpe and Leeder, 2000; Morley, 2002; Peacock, 2002; Young *et al.*, 2002). These are separated by transfer zones or “fault domains” (Schlische and Withjack, 2008). This is then followed by a phase of accelerated fault development, characterised by the propagation of the fault-segment tips, which is itself superseded by a final phase whereby deformation is localised onto a few major fault zones whilst other faults become inactive (Fig. 2.2) (Cowie, 1998; Cowie *et al.*, 2000; Gawthorpe and Leeder 2000).

2.2 Faulting, Folding and Syn-rift Sedimentation

Syn-rift sedimentation is affected by the rate of fault-controlled subsidence, tectonically-controlled topography within the basin, the volume of sediment supply and eustasy within the basin (e.g. Leeder and Gawthorpe, 1987; Prosser, 1993; Davis *et al.*, 2000; McLeod *et al.*, 2002; Young *et al.*, 2002; Bruhn and Vagle, 2005). Stratigraphy within extensional basins commonly records an early phase of basin filling by fluvial or shallow-marine systems followed by a period of sediment starvation and corresponding deposition of lacustrine or deep-marine facies (Prosser, 1993; Schlische and Anders, 1996; Ravnås and Steel, 1998; McLeod *et al.*, 2002). This review focuses on early syn-rift, shallow-marine sedimentation patterns in rift

basins. Figure 2.4 highlights the tectono-sedimentary evolution of a normal fault array in a coastal environment for the rift-initiation and rift-climax stages.

Extensional basins comprise tilt blocks and half-grabens bounded by normal faults. Asymmetrical subsidence during extension will lead to variable slope topography, typically consisting of a gentle hanging wall dip slope and the steeper footwall scarp slope. These slope geometries impact on sediment distribution (Leeder and Gawthorpe, 1987; Prosser, 1993; Gawthorpe and Leeder, 2000). Within this overall basin topography, smaller synthetic and antithetic faults can develop that create topography within the overall dip of the slope and have a more localised effect on sedimentation patterns (Leeder and Gawthorpe, 1987; Gawthorpe and Leeder, 2000). However, this is a rather static view of rift basin structure.

It has been recognised that during the nucleation phase of fault-array development, localised depocentres develop adjacent to fault displacement maxima (Davis *et al.*, 2000; Dawers and Underhill, 2000). Where the volume of sediment supplied into the basin keeps pace with this increasing accommodation space, the surface expression of the tectonic activity may be subdued and therefore will have less of an impact on facies distribution and architecture (Frostick and Steel, 1993). Conversely, where sediment supply is low and unable to match the generation of accommodation space, positive basin topography will develop and facies distributions may be markedly affected. 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. For example, when sediment is sourced from a poorly-consolidated terrain rather than a basement terrain, a faster rate of sediment supply may be achieved that

1. Initiation stage

2. Through-going fault stage

Figure 2.4 Tectono-sedimentary evolution of a normal fault array in a coastal/marine environment. During initiation (1), numerous isolated depocentres develop and form elongate marine gulfs and lakes with marked variation in sedimentary fill. During the through-going fault stage (2) continued fault growth and linkage leads to localisation of deformation onto a limited number of major fault zones with high rates of displacement that outpace sediment supply (redrawn from Gawthorpe and Leeder, 2000).

is capable of infilling topography created even during the rift-climax stage (Ravnås and Steel, 1998). In the Gulf of Corinth, sediment supply during the Holocene was sufficient to deposit fan lobes on the actively uplifting footwall (Leeder *et al.*, 2002).

During this early phase of rift development, the development of extensional forced folds during faulting will also impact the position of depocentres in the fault hanging-wall. During the early stages of fault propagation, the extensional forced fold will cause strata to thin towards the fault zone and thicken into an adjacent hanging-wall syncline (Fig. 2.3). However, once the fault has breached the surface, the greatest accommodation will be created along the fault zone and strata will thicken towards the fault zone (Gawthorpe *et al.* 1997) (Fig. 2.3).

Antecedent drainage patterns can persist within the developing basins, despite topography created during rift basin development, and may be either axial or transverse. Axial systems commonly consist of relatively large fluvio-deltaic systems whilst transverse systems consist of relatively small antecedent systems, or of consequent hangingwall and footwall-derived alluvial fans, deltas and shorelines (Ravnås and Steel 1998; Leeder and Gawthorpe, 1987; Gawthorpe and Leeder, 2000). Several major sedimentary entry points may be recognised within rift basins: at the outlet of antecedent drainage systems, via transfer zones between fault segments, at cross-cutting faults or basinward of easily erodible substrates (Ravnås and Steel, 1998; Gawthorpe and Leeder 2000).

Another impact of topography within a developing rift basin is tidal amplification, which may occur where shallow marine systems form during the early syn-rift,

within narrow, elongate depocentres (Ravnås and Bondevik, 1997; Ravnås and Steel, 1998; Carr *et al.*, 2003; e.g. the Middle Jurassic of the Hebrides; Ravnås and Bondevik, 1997; Mellere and Steel, 1996). Larger wave fetch may occur during periods of tectonic quiescence, especially when footwall islands become completely submerged and a larger, more open-marine embayment occupies the rift basin (Ravnås and Bondevik, 1997).

2.2.1 Syn-rift sedimentation

The impact of syn-depositional structural development on sedimentation within rift basins has been studied using both subsurface and outcrop datasets (e.g. Davies *et al.*, 2000; Jackson *et al.*, 2002). Here, two relatively well-studied examples are discussed; the Tarbert Formation, deposited within the North Viking Graben, and the Nukhul Formation, deposited within the Gulf of Suez Rift. Both formations show significant similarities both to each other and to the Hugin Formation.

Tarbert Formation

The shallow-marine Tarbert Formation was deposited within the North Sea during transgression that was driven in part by the onset of Middle Jurassic extension within the region (e.g. Davies *et al.*, 2000; Dawers and Underhill, 2000). Consequently the thickness, facies distributions, sediment dispersal patterns and feeder systems within the Tarbert Formation have been influenced by syn-depositional rift-related structural development (Davies *et al.*, 2000; McLeod *et al.*, 2002; Hampson *et al.*, 2004). Two examples of this are discussed; one from the East Shetland Basin, on the western side of the Viking Graben, and the other from the Oseberg-Brage Area

located on the Horda Platform which is situated on the eastern margin of the North Viking Graben (Fig. 2.5).

Within the East Shetland Basin, three major structural features impacted Tarbert Formation deposition; the Tern-Eider Horst, the Ninian-Hutton-Dunlin fault system and the North Alwyn-Brent-Statfjord fault (Hampson *et al.*, 2004). The Tern-Eider Horst is a Caledonian basement feature that formed a structural high during Tarbert deposition whilst the Ninian-Hutton-Dunlin fault system was already a hard-linked fault at this time. The structural relief of these features served to focus sediment dispersal and feeder systems within the basin so that they were localised along the northern margin of the horst and along the hanging-wall of the fault system (Hampson *et al.*, 2004).

Conversely, the North Alwyn-Brent-Statfjord Fault system was initiated and developed through segment growth and linkage during Tarbert deposition (Dawers and Underhill, 2000; Davies *et al.*, 2000; McLeod *et al.*, 2002). As a result, Tarbert Formation thickness is variable in this area, thinning towards some footwall crests and thickening into the hangingwall of the North Alwyn-Brent-Statfjord fault system. There is evidence of growth monocline development above the extensional fault system which led to enhanced erosion/non-deposition before the fault breached the surface and an expanded succession was deposited (Hampson *et al.*, 2004). The formation is typically preserved over both the hanging-wall and foot-wall of fault segments, implying that the rate of deposition was broadly equal to the rate of subsidence. In rare cases the formation is missing from degraded footwall crests and is instead preserved within displaced sediments further down the footwall scarp

(a)

Age			Stratigraphy		Tectonic Episode	
Cretaceous		Valanginian	Cromer Knoll Group		<i>Post-rift subsidence</i>	
		Ryazanian				
Jurassic	Late	Volgian	Viking/Humber Gp	Kimmeridge Clay Fm	<i>Rifting</i>	
		Kimmeridgian				
		Oxfordian				
	Middle	Callovian	Brent Gp	Heather Fm	<i>Thermal doming</i>	
		Bathonian				
		Bajocian	Tarbert Fm			
				Ness Fm		
		Aalenian	Rannoch Fm	Etive Fm		
		Broom Fm				

(b)

Figure 2.5 Geological Setting of the Tarbert Formation. (a) Middle to Late Jurassic stratigraphy of the Northern North Sea, the tarbert Formation is highlighted (modified from Davies et al., 2000). (b) Structural setting of the Tarbert Formation; the Alwyn-Brent-Statfjord, Ninian-Hutton-Dunlin and Oseberg-Brage fault systems are highlighted and their impact on the Tarbert Formation discussed further in the text (redrawn from Færseth and Ravnås, 1998).

(McLeod *et al.*, 2002). Superimposed on regional subsidence trends, short-lived, low displacement intra-basin faults generated differential accommodation during the early part of the rift event (Davies *et al.*, 2000). Whilst activity along the North Alwyn-Brent-Statfjord fault system during Tarbert deposition resulted in thickness variations within the formation, it did not result in a major facies reorganisation due to the balance between sediment supply and subsidence (McLeod *et al.*, 2002). Fluvial systems were constrained by fault-growth monoclines and flowed parallel to the main faults, however these faults developed parallel to the antecedent drainage within the basin and therefore did not deflect it. Structurally-controlled localised depocentres were also a focus for stacked channel systems and aggradational shoreface successions (Fig. 2.6; Davies *et al.*, 2000; McLeod *et al.*, 2002). Structural development within the basin is, however, interpreted to have led to the dominance of tidal rather than wave processes during deposition of the formation (Hampson *et al.*, 2004). The strong surface expression of the fault systems and horst block would have led to development of confined basins with a small wave fetch and potentially amplified tidal currents.

In a similar fashion, the Ninian-Hutton-Dunlin fault system and the Oseberg Fault within the Oseberg-Brage area were active before and during Tarbert deposition (Færseth and Ravnås, 1998). In this area, the Tarbert Formation comprises shoreline sandstones and lower delta plain heteroliths (Ravnås *et al.*, 1997) deposited in an overall wedge-shaped package that thickens towards active fault segments. Both transgressive and regressive packages are interpreted within the formation, with transgressive segments associated with periods of increased fault rotation and subsidence (Ravnås *et al.*, 1997; Færseth and Ravnås, 1998). Strong tidal influence

Figure 2.6 Diagrammatic cross-section illustrating the distribution of Tarbert Formation facies related to active faults in the Statfjord Fault area. Fluvial systems were constrained by fault-growth monoclines and flowed parallel to the main faults, however these faults developed parallel to the antecedent drainage within the basin and therefore did not deflect it. Structurally-controlled localised depocentres were also a focus for stacked channel systems and aggradational shoreface successions (redrawn from Davies *et al.*, 2000)

is observed within the transgressive packages as a result of fault-block rotation which created smaller sub-basins that formed embayments or estuaries with limited wave fetch. Stacked shoreline bodies aggraded against exposed fault scarps.

Within the Oseberg Field, the Tarbert Formation comprises a wave-dominated deltaic depositional system that prograded from south to north (Løseth *et al.*, 2001). A transgressive tidal unit is identified locally within hanging-wall areas at the base of the formation. An unconformity caps the Tarbert Formation where the Oseberg fault block was subaerially exposed in up-dip areas and eroded material from the fault-block crest then sourced intra-Heather Formation sandstones.

Nukhul Formation

Outcrops within the Gulf of Suez rift provide several examples of rift-related faulting and folding affecting syn-rift sedimentation (Gawthorpe *et al.*, 1997; Sharp *et al.*, 2000; Carr *et al.*, 2003; Jackson *et al.*, 2005). Rifting within the region occurred during the Oligo-Miocene and was concurrent with deposition of a classic, upward-deepening rift-basin succession. Continental deposits (Abu Zenima Formation) are overlain by shallow-marine deposits (Nukhul Formation) which are themselves overlain by deeper marine sediments (Rudeis Formation) (Fig. 2.7a). Like the Tarbert Formation of the North Sea, the Nukhul Formation comprises transgressive, tidally-influenced units, within which stratal geometries and depositional patterns have been affected by syn-depositional structural development (Jackson *et al.*, 2002; Carr *et al.* 2003). Examples from the Hammam Faraun and El Qaa fault blocks are discussed here (Fig. 2.7b).

Figure 2.7 Geological Setting of the Nukhul Formation. (a) Simplified stratigraphy of the study area, the shallow-marine Nukhul Formation is highlighted and discussed further in the text (modified from Jackson *et al.*, 2002). (b) Regional plate tectonic setting of the Gulf of Suex region and a simplified fault map of the Suez rift showing major faults (>1 km throw). Examples of syn-rift sedimentation are discussed from the highlighted Hamam Faraun (East Tanka Fault Zone) and El Qaa (Baba Sidri fault zone) fault blocks (modified from Jackson, 2002).

Initial fault activity across the Hammam Faraun fault block occurred on individual fault segments (Gawthorpe *et al.*, 2003). Some of these segments grew with increased displacement and linked to form longer fault zones. Displacement then localised onto elongate border fault zones and migrated towards the western boundary of the fault block. The Nukhul Formation is thickest in hanging-wall depocentres and shows depositional thinning and onlap towards fault tips (Sharp *et al.*, 2000). Major fault-parallel variations in thickness are observed within the formation due to variations in displacement and subsidence along adjacent fault segments (Jackson *et al.*, 2002; Carr *et al.* 2003; Gawthorpe *et al.*, 2003).

Nukhul Formation deposition in the fault block occurred within tidally-influenced embayments. Depocentres were typically elongate and controlled by a number of isolated normal fault segments and growth folds (Carr *et al.* 2003; Jackson *et al.*, 2005). These narrow, elongate depocentres led to confinement of bay-head deltas and enhanced the tidal signature within the succession due to tidal amplification (Carr *et al.* 2003; Jackson *et al.*, 2005). Along-strike variation in fault displacement controlled facies development, with increased marine influence corresponding to regions of increased fault displacement (Fig. 2.8; Sharp *et al.*, 2000; Jackson *et al.*, 2005).

Only transgressive systems tracts (TST) and highstand systems tracts (HST) are observed within the Nukhul Formation (Carr *et al.*, 2003; Jackson *et al.*, 2005); lowstand and forced regression systems tracts are not recognised. The repeated aggradational to retrogradational stacking pattern is due to a balance between sediment supply and accommodation generation. TST units are interpreted to have

Figure 2.8 Nukhul Formation facies variability: hangingwall East Tanka fault zone. Block diagram highlighting facies variability along strike of the East Tanka Fault Zone, Suez Rift. Along-strike variation in fault displacement controlled facies development with increased marine influence observed corresponding to regions of increased fault displacement (from Jackson *et al.*, 2005)

been deposited when accommodation outpaced sediment supply (i.e. during periods of fault activity) whereas the HST unit were deposited when fault activity decreased and sediment supply outpaced accommodation.

Nukhul Formation deposits within the El Qaa fault block to the south comprise tidal flat and tidal channel deposits overlain by shallow-marine shoreface sandstones (Gawthorpe *et al.*, 1997). The Baba-Sidri fault zone in the northern part of this block was developing through fault propagation and linkage during Nukhul Formation deposition. The growth fold stage of fault development is defined by a shallow marine stratal wedge which thins, onlaps and is truncated towards the fault (Fig. 2.9). During this period, maximum accommodation was offset some distance into the hanging-wall of the fault zone (Gawthorpe *et al.* 1997). Once the fault had propagated and broken surface, a shallow marine unit was deposited which thickens towards the fault in a location where maximum accommodation developed.

2.3 Halokinesis

2.3.1 Mechanisms of salt movement

Historically, buoyancy was thought to be the main mechanism driving halokinesis (Jackson and Talbot, 1986; Jackson, 1995; Hudec and Jackson, 2007). However, more recent studies have begun to recognise the importance of differential loading in initiating salt movement (Jackson and Talbot, 1986; Hudec and Jackson, 2007). Three types of loading may be responsible for initiating salt flow; gravitational, displacement and thermal (Fig. 2.10; Jackson and Talbot, 1986; Waltham, 1997; Hudec and Jackson, 2007). Gravitational loading results from the weight of the

Figure 2.9 Tectono-stratigraphic development above and adjacent to an upwardly-propagating blind fault tip; the Baba-Sidri fault zone, El-Qaa fault block, Suez rift. Note basinward expansion of sequences, and the progressive rotation and enhancement of marine regressive erosion surfaces due to fold amplification and up-dip amalgamation of key stratal surfaces. (From Jackson, 2002).

Figure 2.10 Three main types of loading that may be responsible for salt flow. Salt flow is highlighted by red arrows. (a) Gravitational loading results from the weight of the overburden and gravitational forces within the salt layer. In this example, the salt flows from the area of high hydraulic head (where sand has been deposited) to areas of low hydraulic head. (b) Thermal loading is driven by temperature changes and associated volume changes within the salt layer. In this example, a localised heat source within the basement generates convection currents within the salt layer and leads to localised uplift of the salt surface. (c) Displacement loading occurs when the flanks of a salt body move closer or further apart as a result of basin extension or compression (modified from Jackson and Talbot, 1986).

overburden and the gravitational forces within the salt layer. The salt will flow like a fluid from areas of high to low hydraulic head. Displacement loading occurs when the flanks of a salt body move closer or further apart as a result of basin extension or compression. Thermal loading is driven by temperature changes and associated volume changes within the salt layer (i.e. hot salt is less dense than cold salt).

Many types of salt structure may be formed as a result of halokinesis. These are summarised in Figure 2.11. It should be noted that the lithofacies within a salt unit will impact the ability of the unit to undergo halokinesis. “Proximal” facies (e.g. carbonates and anhydrites) are less likely to undergo halokinesis, whilst more basinal facies (e.g. halite) are more likely to form salt growth structures (Jackson and Talbot, 1986; Clark *et al.*, 1998). Salt diapirs are one common type of halokinesis-related structure formed as a result of the upward movement of salt. Diapirs will continue to grow until they penetrate a lower density layer, encounter a resistant overburden, detach from the root, or the source layer is exhausted (Jackson and Talbot, 1986). The areas flanking salt diapirs overlying areas of salt depletion are commonly termed “salt withdrawal basins” (Shelley and Lawton, 2005). When the diapir detaches from the root or the source salt body is exhausted, diapirs will start to subside and former diapir crests will then become sites of high subsidence and new depocentres may form (Vendeville and Jackson, 1992). Diapir growth may occur as a result of active, reactive, or passive diapirism and salt diapirs may show stages of all three processes throughout their evolution (Fig. 2.12) (Jackson *et al.*, 1994).

Figure 2.11 Schematic summary of salt structure geometries that may be generated by salt movement. It can be difficult to image the overburden geometries within seismic data where they are overlain by salt features due to the density contrast between the salt body and surrounding strata (modified from Hudec and Jackson, 2007).

Figure 2.12 Models of diapir piercement. (a) Diapir growth may occur as a result of active, reactive or passive diapirism. Active diapirism occurs when a diapir is forcefully intrudes the overburden, driven by diapir pressure sufficient to overcome overburden strength. Reactive diapirism occurs when the salt 'reacts' to overburden thinning due to extension. Passive diapirism occurs once the diapir has pierced the overburden and continues to grow due to downbuilding. (b) During the evolution of a diapir, diapir growth may occur as a result of all three processes. In this example regional extension leads to diapir piercement. The maturity of a structure will depend on availability of salt, total amount of extension and relative rates of extension and sedimentation (modified from Hudec and Jackson, 2007).

Active diapirism

An active diapir is one which forcefully intrudes the overburden, driven by diapir pressure that is sufficient to overcome overburden strength (Schultz-Ela *et al.*, 1993). Diapir pressure can be generated by differential loading of the salt body and the density contrast with the overburden (Jackson and Talbot, 1986; Jackson *et al.*, 1994; Jackson and Vendeville, 1994; Shelley and Lawton, 2005; Hudec and Jackson, 2007). As a result this process only occurs at shallow depths of burial, e.g. < 500 to 1000 m.

Reactive diapirism

Thin-skinned extension may thin and fracture the post-salt overburden, generating a point of weakness within the overburden and lateral variation in the load above the salt (Penge *et al.*, 1993; Vendeville and Jackson, 1992; Jackson and Vendeville, 1994; Vendeville and Nilsen, 1995). Salt may then be able to migrate upwards through the fault and fractures within the thinned overburden. This type of diapirism is known as a “reactive diapirism” because the salt ‘reacts’ to overburden thinning related to regional extensional forces (Jackson *et al.*, 1994; Stewart and Coward, 1995; Hudec and Jackson, 2007). Reactive diapirism has been proposed as the main initiator of salt diapirism in many areas because extension occurs in many regions of salt diapirism. In addition, the mechanism is not significantly influenced by the thickness, lithology or density of the overburden (Schultz-Ela *et al.*, 1993). Sub-salt or thick-skinned extension will also impact diapir development as highlighted by 3D modelling undertaken by Dooley *et al.* (2005). With increasing extension, reactive diapirs form in the footwalls of the basement faults and gradually rise as the overburden is thinned. Corner points in the basement fault system (i.e. where fault

systems interact), generate complex flap structures within the overburden and localise diapiric activity on the rift margins. Growth of the diapirs is facilitated by salt flow from the footwalls of the basement system before salt flow is cut off by grounding along the basement crest (Dooley *et al.*, 2005). This type of diapirism is observed in the North Sea Basin where east-west extension, initiated during the Triassic, coupled with gravitational instability initiated salt movement and the Upper Permian salt accommodated extension by shear and gravity flow (Hodgson *et al.*, 1992). A series of N-S trending salt ridges and basins formed above the reactivated Permian faults. As predicted by Dooley (2005), salt walls were commonly located over the crests of the Permian fault blocks. Triassic sedimentation was focussed within these basins and progressively displaced salt from under this sediment “pod” into adjacent salt walls or ridges (interpod area). In most cases, subsidence of these “pod” continued until all the salt was displaced from beneath it and the pod became “grounded”. The site where the pod grounded is termed a “salt weld”. Associated growth of adjacent salt walls stopped as the salt supply was diminished and these walls began to collapse, creating localised basins. Figure 2.13 summarises the pod-interpod model.

Passive diapirism

Once a diapir has pierced the overburden and is near to, or exposed at, the sediment surface it may continue to grow by “passive diapirism” or “down-building”, (i.e. the diapir remains exposed whilst sediments accumulate around it and onlap its flanks) (Davison *et al.* 1993; Jackson *et al.* 1994; Clark *et al.*, 1998; Rowan *et al.*, 2003; Aschoff and Giles, 2005; Hudec and Jackson, 2007). In this situation, the diapir increases in relief because its base sinks relative to the sediment surface as the basin

Figure 2.13 “Pod-interpod” model reflecting salt body geometry during the Triassic to Jurassic in the Central Graben of the North Sea. (a) Initiation of pod subsidence due to regional extension, salt thickens above basement faults and Triassic sediments focussed in depocentres between these highs. (b) Triassic pod subsidence and salt wall growth. (c) Triassic pods ground when underlying salt sheet has migrated completely into adjacent salt-cored highs. (d) Salt wall collapse generates accommodation space infilled by Jurassic-Cretaceous sediments. Jurassic extension occurs where pod is weakest i.e. over collapsing salt wall resulting in half-grabens along salt-ridge crests (modified from Hodgson *et al.*, 1992).

fills (Schultz-Ela *et al.*, 1993). Salt may also grow passively in the absence of mini-basin subsidence if the salt is inflated, e.g. by shortening, (Rowan, *et al.*, 2003). Passive diapirism dominates the history of most diapiric salt bodies and ceases only when salt flow cannot keep pace with sedimentation (Rowan *et al.*, 2003).

2.3.2 Salt-influenced rift basins

Depending on the relative timing of salt deposition and rifting, extensional tectonics may be thin- or thick-skinned (Rowan and Giles, 2007). When the salt deposition is pre-rift extension affects both the supra- and sub-salt sections. However the salt layer often acts to decouple the deformation. This leads to significantly different structural styles in the sub- and supra-salt sections. Thick, mobile salt units within rift basins may strongly impact the structural development of such basins. Thicker layers of salt may form a range of structures and associated basins during extension. Thinner layers of salt may not be able to form diapirs but may act as detachment surfaces onto which supra-salt listric growth faults may develop and detach (Jackson *et al.*, 1994; Bishop, 1996; Hudec and Jackson, 2007; Jackson and Larsen, 2009). Fault-propagation folds that develop during the early stages of rift development may be enhanced in salt-influenced rift basins due to the flow of salt into their cores. In fact, the upward propagation of basement faults may be completely retarded by the presence of ductile salt within the basin fill.

2.4 Impact of Halokinesis on Syn-tectonic Sedimentation

The interaction between salt and sediment within a basin controls the structural style and geomorphological processes at the depositional surface and therefore the thickness, stratal architecture and facies distribution of syn-tectonic sedimentary units.

Syn-depositional halokinesis typically results in the formation of small basins (salt-withdrawal basins) which flank the diapir. These basins may be identified as a rim syncline when they form a local depression that develops as a border around a salt dome. Sediment accumulates within these depocentres and the facies and sequence-stratigraphic development of these units are strongly controlled by topography created during ongoing salt movement (Hodgson *et al.*, 1992; Rowan and Weimer, 1998; Shelley and Lawton, 2005). When the rate of salt removal or dissolution is greater than or equal to the sedimentation rate, sediment dispersal may be impacted and the distribution of sedimentary facies affected (Heward, 1990; Clark *et al.*, 1999; Cartwright *et al.*, 2001).

In shallow marine settings, depositional systems may develop around salt-cored highs, whilst pelagic sedimentation prevails in the salt-withdrawal basins. Conversely, in deep-marine settings, sedimentation from gravity flows may be focussed within salt-withdrawal basins flanking salt bodies. This effect will contribute to differential sediment loading that will continue to drive salt movement from under the basins and into the diapirs, thereby creating a positive deposition-subsidence feedback loop (Hodgson *et al.*, 1992; Rowan and Weimer, 1998; Rowan

et al., 2003; Gemmer *et al.*, 2004; Aschoff and Giles, 2005; Shelley and Lawton, 2005; Matthews *et al.*, 2007). When the rate of deposition is relatively rapid, salt movement will be suppressed and early-stage salt pillows may be ‘frozen’. In contrast, when the depositional rate is relatively low, salt movement is more efficient and large salt bodies flanked by salt welds may form (Gemmer *et al.*, 2004).

Within the Mesozoic succession of the North Sea, the “pod-interpod” model (Fig. 2.13) has been proposed to explain the geometry of salt bodies and their relationship to overlying stratal units (Johnson *et al.*, 1986; Hodgson *et al.*, 1992; Erratt, 1993; Clark *et al.*, 1998). In this model, Triassic sedimentation was focussed within localised basins formed through salt-withdrawal (Johnson *et al.*, 1986). Ongoing sediment loading caused the mini-basins to subside through the halite layer until they “grounded-out” on the top of the basement (i.e. they became welded) whilst underlying salt flowed into adjacent, pillow-like highs (Clark *et al.*, 1998). Continued salt-withdrawal meant that the rim-synclines eventually became grounded and started to invert, creating secondary rim synclines. Late Jurassic sedimentation was then focussed into these secondary rim synclines (Hodgson *et al.*, 1992). Conversely, other workers (e.g. Penge *et al.*, 1993) suggest that the salt diapirs developed reactively during Jurassic extension, upwelling into space generated between “rafted” Triassic fault blocks. In this model, diapirs are focussed at the sites of supra- rather than sub-salt extensional faulting.

2.4.1 Examples of salt-controlled sedimentation

Understanding the interplay between salt diapirism and sedimentation is predominantly based on subsurface datasets (e.g. Hodgson *et al.*, 1992; Stewart *et*

al., 1999), although salt-controlled sedimentary sequences exposed at outcrop have allowed for higher resolution field-based investigations. In this sub-section, examples of salt-controlled sequences are discussed from the La Popa Basin, Mexico, and the Central Graben of the North Sea. In each setting, the focus of this review is shallow marine units, comparable in terms of sedimentology to the Hugin Formation.

La Popa Basin, Mexico

In the La Popa Basin, Mexico, salt diapirs and their encasing strata are well exposed, permitting detailed study of salt-sediment relationships and interactions. The structural setting and lithostratigraphy of the La Popa basin is summarised in Figure 2.14. The basin forms a segment of a foreland basin that roughly parallels the deformational front of the Sierra Madre Oriental fold and thrust belt (Aschoff and Giles, 2005). Diapirism within the basin was triggered and sustained by compressional décollement tectonics (Millan-Garrido, 2004). This compression started no later than Campanian-Maestrichtian times and continued until at least the Tertiary (Millan-Garrido, 2004). In this example, diapir uplift strongly modified basin-floor topography and consequently modified sediment dispersal patterns (Aschoff and Giles, 2005; Shelley and Lawton, 2005).

Syn-depositional topography generated by halokinesis influenced the shallow marine sediments of the Cretaceous Delgado Sandstone Member (Potreillos Formation). Stratal units thin and become more shallow-water in facies character towards the diapir (Aschoff and Giles, 2005). Sandstone was deposited in up-dip shorelines, centred in depocentres around the flanks of the salt diapirs, whilst sediment “shadow

Figure 2.14 Geological Setting of the La Popa Basin. (a) La Popa Basin location map, major geological elements within the Sierra Madre foldbelt and foreland region, northeastern Mexico, are highlighted (Simplified from Shelley and Lawton, 2005). (b) Generalised stratigraphy of the La Popa Basin, highlighting the position of the Delgado Sandstone and Upper Sandstone members of the Potrerillos Formation (after Aschoff & Giles, 2005; Shelley and Lawton, 2005).

zones” developed down-dip (Fig. 2.15). A counter-intuitive relationship is observed at the outcrop whereby syn-halokinetic strata thin towards the axis of salt-withdrawal basins. This may be explained by the presence of a salt-cored structural high when the sediments were deposited that the units thinned onto. Subsequently, salt migrated out of this high leaving the present day salt-withdrawal basin (Aschoff and Giles, 2005). This example highlights the importance of understanding the history of salt movement within a basin, as the structural configuration of salt bodies during deposition may be markedly different to their present day configuration.

The upper sandstone member of the Palaeocene Upper Potrerillos Formation was also influenced by syn-depositional diapirism and shortening within the basin (Shelley and Lawton, 2005). The unit comprises a regressive succession of marginal marine, siliciclastic deposits. Sedimentary structures indicate asymmetrical, bi-directional palaeocurrents that trend parallel to regional structural trends. The Carroza syncline confined an ebb-dominated deltaic system and the resulting stratal accumulation within the syncline hinge caused migration of underlying salt into flanking structures. During regional regression, the syncline continued to impact deposition by forming a palaeogeographical embayment and controlling shoreline evolution.

Central Graben, North Sea

The Central Graben of the North Sea Basin lies within the North Permian evaporite basin. During the Permian up to 2 km of halite was deposited in the basin centre, passing laterally into thin carbonate and sulphate sequences at the basin margins (Penge *et al.*, 1993; Clark *et al.*, 1998; Stewart and Coward, 1999). The mechanisms

Figure 2.15 Sandstone distribution within the Delgado Sandstone Member (Potreillos Formation). Syn-depositional halokinesis modified sediment dispersal routes through the basin. Stratal units thin and lithofacies become more shallow-water in origin towards the diapir. Sandstone depocentres sourced from up-dip shorelines centred around the flanks of the salt diapirs, whilst sediment shadow zones developed down-dip of them (redrawn from Aschoff and Giles, 2005).

responsible for salt tectonics within the basin are debated and have significant implications for the timing of salt movement. The “raft tectonic” model of Penge *et al.* (1993; 1999) interprets reactive diapirism as a result of extension during the Jurassic. In contrast the ‘pod-interpod’ model (Fig. 2.13) of Hodgson *et al.* (1992) interprets that minibasin subsidence started in the Triassic as a result of reactivation of sub-salt, Permian-age faults. Regardless of timing, salt-withdrawal and dissolution within the basin generated a series of salt ridges, diapirs and flanking depocentres. This led to modification of topography which influenced sediment dispersal patterns (Hodgson *et al.*, 1992; Stewart *et al.*, 1999).

The shallow-marine Fulmar Formation was deposited during a period of Upper Jurassic rifting and halokinesis within the Central Graben (Fig. 2.16). Significant lateral variations in thickness and distribution of the formation are attributed to complex fault and salt-related topography present within the basin during deposition (e.g. Stewart *et al.*, 1999; Fraser *et al.*, 2003; Stewart, 2007). There is debate over whether the topography was created prior to (e.g. Stewart *et al.*, 1999) or during (e.g. Wakefield *et al.*, 1993) Fulmar Formation deposition, although it is agreed that both halokinesis and rift-related faulting were responsible for its creation. On a regional scale, Fulmar-age topography is dominated by salt-cored valley networks, surrounded by low relief, Triassic-cored hills. Within this system, sand deposition and preservation relied on the geometry and connectivity of the drainage network, the availability of sand and the amount of syn-depositional accommodation (Stewart *et al.*, 1999; Fraser *et al.*, 2003). Fulmar Formation distribution is interpreted to be focussed on the flanks of syn-depositional topographic highs (Fig. 2.17a). Several

Figure 2.16 Geological Setting of the Fulmar Formation. (a) Regional basin framework of the North Sea Rift with the region of interest highlighted. (b) Structural framework of the West Central Shelf and Central Graben region. Three hydrocarbon fields: Kittiwake, Puffin and Fulmar are highlighted and discussed further in the text (simplified from Fraser *et al.*, 2003)

Figure 2.17 Structural controls on Fulmar Formation distribution. (a) Idealised palaeogeography of the West Central Shelf showing possible Fulmar facies distribution with respect to local geological controls during an Upper Jurassic highstand. Triassic mini-basin fill is exposed as topography forming a series of low lying hills/islands. Sandstone deposition is focussed fringing these highs. Wave action, sand re-working and therefore sand deposition is suppressed within the valley network. Earlier shoreface deposits drowned during transgression are preserved in topographic lows e.g. fault hangingwalls (modified from Stewart *et al.*, 1999). (b) Schematic cross-section highlighting Upper Jurassic facies distribution across the Fulmar Field. The pod-shaped geometry of the Fulmar Formation is controlled by salt withdrawal. Initial sandstone deposition was sheet-like across the area before becoming restricted to the western margin of the pod and pinching out to the east into deeper water facies over comparatively short distances. This is overlain by deep water Kimmeridge Clay and mass flow deposits. The main sediment source is the Auk Ridge, a Permo-Triassic high to the west (redrawn from Johnson *et al.*, 1986).

specific examples of the complex relationship between structure and sediment distribution are discussed below.

On the West Central Shelf, located to the west of the Central Graben, salt-dissolution is interpreted to have influenced Fulmar Formation deposition. In this area, Fulmar-age shorelines developed along narrow belts that mark the position of earlier Jurassic valley systems (Wakefield *et al.*, 1993; Fraser *et al.* 2003). These valley systems developed as a result of preferential erosion of the exposed salt diapir instead of the adjacent mini-basin fill (Wakefield *et al.*, 1993; Stewart and Clark, 1999; Stewart, 2007). This erosion created topographic relief of tens of metres within which deposition was focussed.

Triassic-age salt withdrawal in the Kittiwake Field area led to deposition of thick, Smith Bank mudstones in 'pods'. Subsequent compaction, and possibly salt withdrawal, during the Lower Jurassic then generated accommodation space between these pods within which Fulmar Formation sandstones were deposited (Veldkamp *et al.*, 1996). A similar depositional model is envisioned for the neighbouring Puffin Field (Dickinson, 1996). During sea-level lowstand, sedimentation was restricted to the deepest parts of the interpod area whilst later transgressive deposits were focussed towards the margins of the interpod basins and over the adjacent pod crests (Veldkamp *et al.*, 1996).

Within the Fulmar Field, the Fulmar Formation sands are interpreted to have a 'pod-shaped' geometry (Johnson *et al.*, 1986). This has variably been interpreted to reflect the formation of local, syn-depositional hollows or basins resulting from late stage

withdrawal of the salt (Johnson *et al.*, 1986; Stockbridge and Gray, 1991; Fraser *et al.*, 2003). However, more recent work suggests this relationship may have been controlled by normal fault development (Kuhn *et al.*, 2003). The thickest and most complete sequences are developed in the axial part of these 'pods', whilst towards the margins the units amalgamate and unconformities may be developed. Initially sheet-like sandstones were deposited across the basin, before they became restricted to the western margin of the pod and pinched out to the east into deeper water facies over comparatively short distances (ca. 2.5 km; Fig. 2.17b; Johnson *et al.*, 1986; Stockbridge and Gray, 1991; Kuhn *et al.*, 2003).

2.5 Stratigraphic Models

Stratigraphic architecture is controlled by the interaction of accommodation space, sediment supply and sediment transport (e.g. Sloss, 1962). Accommodation space represents the space available for sediment to accumulate within, and is controlled by eustasy coupled with subsidence rate (Jervey, 1988; Posamentier *et al.*, 1988). Sediment supply in shallow-marine settings is delivered from the land via rivers, from sites along the coastline by longshore drift and onshore by various marine reworking processes (e.g. Reynolds, 1996). In these shallow marine settings, the balance between the sediment supply and relative sea-level rise controls whether facies belts prograde basinward (regression) or retrograde landwards (transgression) (Mitchum, 1977).

Regressive depositional sequences record a basinward migration of the coastline (Bates and Jackson, 1987). This may be driven by an increase in sediment supply

relative to accommodation space (normal regression; Posamentier *et al.*, 1992) or as a result of a relative sea-level fall (forced regression; Plint, 1988; Posamentier *et al.*, 1992). During normal regression, sediments will typically prograde and aggrade into the basin whilst during a forced regression sediments will prograde and downstep into the basin (Posamentier and Allen, 1999). Transgressive depositional sequences record a landward migration of the coastline during a relative rise of sea level (Posamentier and Vail, 1988; Emery and Meyers, 1996; Helland-Hansen and Martinsen, 1996; Cattaneo and Steel, 2003). The balance between sediment supply and accommodation is key to determining whether a unit will be regressive or transgressive (e.g. Catuneanu, 2002) and frequent cyclicity may be observed within shallow-marine deposits reflecting alternating periods of regression and transgression (e.g. Einsele and Seilacher, 1982; Helland-Hansen and Gjelberg, 1994). This cyclicity may be short term, e.g. seasonal, reflecting flood versus non-flood conditions within a fluvial system, or longer term, e.g. millions of years, reflecting tectonic or eustatic variations (Helland-Hansen and Gjelberg, 1994).

Accommodation space varies geographically during cycles of base-level change, for example during transgression, the shelf area enlarges and there is a tendency for more sediment to be partitioned into facies tracts in the continental or shallow marine part of the depositional system (Cross *et al.*, 1993). This results in an overall lower sediment influx into distal parts of the basin (Cross, 1988; Myers and Milton, 1996; Cross and Lessenger, 1998; Cattaneo and Steel, 2003). In contrast, during base-level fall, or forced regression, accommodation space decreases in more landward positions and the volume of sediment partitioned into facies tracts in a more seaward position increases (Cross *et al.*, 1993). Spatial and temporal variation

in accommodation relative to sediment supply during base-level cycles will impact sediment accumulation rates within the basin (Cross *et al.*, 1993; Helland-Hansen and Gjelberg, 1994). Consequently, more sediment may be incorporated into the stratigraphic record during one portion of the base-level cycle (i.e. during transgression) than during another (i.e. during regression) at any constant geographical position (Cross *et al.*, 1993).

Stratigraphic stacking patterns can be described using shoreline trajectory, defined as the cross-sectional shoreline migration path along depositional dip (Helland-Hansen and Gjelberg, 1994). This trajectory is a function of relative sea-level change, sediment supply and basin physiography, and ranges from 90-180° in transgressive deposits (Fig. 2.18). The shoreline trajectory is linked to the preservation potential of associated deposits (discussed more below; Helland-Hansen and Martinsen, 1996; Muto and Steel, 2002b; Cattaneo and Steel, 2003).

Transgressions may be continuous, when a steady rate of relative sea level rise outpaces steady sediment supply, or they may be punctuated or stepped, when the overall transgressive trend is interrupted by episodes of regression (Helland-Hansen and Gjelberg, 1994; Cattaneo and Steel, 2003). These two processes generate different stratigraphic responses and shoreline trajectories, impacting the preservation of associated strata (Helland-Hansen and Gjelberg, 1994). Where accommodation generation outstrips sediment supply, the shoreline migration path will merge with or erode the alluvial depositional surface in a landward direction to generate a composite, diachronous but physically continuous transgressive erosion (ravinement) surface (Helland-Hansen and Gjelberg, 1994; Cattaneo and Steel,

Figure 2.18 Transgressive Shoreline Trajectories (from Helland-Hansen and Gjelberg, 1994). The shoreline trajectory is the cross sectional shoreline migration path along the depositional dip in response to changes in relative sea-level, sediment supply and basin physiography. (a) Transgressive shoreline trajectories will range between 90° upwards and 180° landwards. (b) Three classes of transgressive shoreline trajectory.

2003). Increasing sediment supply will lead to the shoreline trajectory diverging relative to the alluvial depositional surface and may lead to deposition of one or more, expanded backstepping parasequences, bounded by successive ravinement surfaces that step upwards and landwards, and that typically erode the landward termination of the underlying transgressive deposits (Helland-Hansen and Gjelberg, 1994; Cattaneo and Steel, 2003). When sediment supply outpaces accommodation generation during transgression, a “step-up” or stepped transgressive surface may be formed by a series of landward- and upward- offset ravinement surfaces, each separated by small progradational sediment packages. This stacking of ravinement surfaces can lead to significant aggradation of the transgressive systems tract (Olsen *et al.*, 1999; Cattaneo and Steel, 2003). Stepped transgressions may result in several stratigraphic responses: “in-place drowning”, “overstep” and “step-up” (overstep and step-up models illustrated, Fig. 2.19). In-place drowning is typical of low-gradient settings and is characterised by vertical accretion of barrier-lagoon systems that are eventually submerged by an increased rate of relative sea level rise or a reduction in sediment supply (Sanders and Kumar, 1975; Cattaneo and Steel, 2003). Overstep will generate either collapsed backstepping parasequences or expanded backstepping parasequences, depending on the rate of sediment supply versus accommodation generation (Swift *et al.*, 1991b).

Shoreface erosion will affect the preservation of shallow marine deposits (e.g. Swift 1968; Nummedal and Swift, 1987; Van Wagoner *et al.*, 1990; Cattaneo and Steel, 2003). During forced regressions the wave base moves down with the falling relative sea-level leading to erosion of seafloor sediments that previously lay beneath wave base and forming the ‘regressive surface of marine erosion’ (Plint, 1988). During

Overstep

Collapsed backstepping parasequences $A/S \gg 1$

Expanded backstepping parasequences $A/S > 1$

Step-up

Stepped transgression surface $A/S > 1$

Figure 2.19 Stratigraphic response to discontinuous or punctuated transgression, highlighting the impact of an increasing role of sediment supply (modified from Cattaneo and Steel, 2003). Abbreviations used; wrs = wave ravinement surface, fs = flooding surface, MFS = maximum flooding surface, A/S = accommodation/supply ratio.

transgression, shoreline retreat will erode previously deposited regressive and transgressive deposits (e.g. Van Wagoner *et al.*, 1990; Helland-Hansen and Gjelberg, 1994; Reynolds, 1996; Cattaneo and Steel, 2003). The surface of ravinement may be either: (1) a wave ravinement surface (*sensu* Swift, 1968), formed by erosion by wave action during shoreface retreat and thus developed in shallow marine areas above fair-weather wave base; or (2) a tidal ravinement surface (*sensu* Swift, 1968) formed by erosion by tidal currents and therefore restricted in development to tidally-influenced and tidally-dominated coastal areas. Wave ravinement surfaces can erode large sectors of the shelf as the shoreface migrates landwards and pass basinwards into a deepening surface without associated erosion (Cattaneo and Steel, 2003). Tidal ravinement surfaces may extend over long distances up-dip within a tidal inlet or channel but are usually of limited strike extent (Cattaneo and Steel, 2003). Ravinement surfaces are omitted from “standard” sequence stratigraphic models (e.g. Van Wagoner *et al.*, 1990).

Preservation of shallow marine deposits is also controlled by the underlying topography, with the thickest and most complete transgressive sequences being preserved in topographic lows (e.g. valleys, tidal inlets, the downthrown side of faults; Belknap and Kraft, 1985; Demarest and Leatherman, 1985; Nummedal and Swift, 1987; Donselaar, 1989; Hart and Plint, 1993; Helland-Hansen and Martinsen, 1996; Cattaneo and Steel, 2003). Topography can also influence the facies characteristics, distributions and architectures developed during transgression. For example, as stated above, transgression of narrow, elongate depressions such as incised valleys can lead to tidal amplification and therefore a dominance of tidally influenced deposits (Cattaneo and Steel, 2003; Folkestad and Satur, 2008). An

understanding of the topography at the time of transgressive deposition can, therefore, improve prediction of reservoir-quality trends. Conversely, facies variability identified within transgressive deposits can help to constrain the timing and influence of structurally developed topography within a basin.

From an applied perspective, transgressive shallow marine reservoirs host a significant volume of hydrocarbons in the North Sea Basin (Husmo *et al.*, 2003) and other basins worldwide. They can be highly variable in terms of sandbody geometry and distribution, thereby making exploration for, and production from, such units somewhat challenging (Nummedal and Swift, 1987; Maxwell *et al.*, 1999; Cattaneo and Steel, 2003; Sixsmith *et al.*, 2008).

There has been a focus of research into shallow marine systems deposited during an overall basinward shift in coastline, due in part to better preservation of associated deposits (see below). Existing models thus emphasise the development of wave-dominated shorelines during regression. Tidally-influenced deposits are considered to be more significant during transgression, with deposition in the erosional topography formed during the preceding base-level fall (e.g. within an incised valley) limiting wave action and amplifying tides (e.g. Shanley and McCabe, 1994; Cattaneo and Steel, 2003; Dalrymple *et al.*, 2007; Yoshida *et al.*, 2007). Several different types of transgressive shoreline are recognised, controlled by variation in wave and tidal power (Fig. 2.20) (Dalrymple *et al.*, 1992). In areas with a large tidal range, tidal flats typically rim the margins of back-barrier lagoons, estuaries and the open coast (Dalrymple, 1992). Estuaries are a common feature of transgressive shorelines and are defined as the seaward portion of a drowned valley system

which receives sediment from both fluvial and marine sources and which contains facies influenced by tide, wave and fluvial processes (Dalrymple *et al.*, 1992). Estuaries may be either tide- or wave-dominated with implications for sand body geometry and distribution (Fig. 2.20).

Barrier islands may be another common feature of transgressive shorelines, with many of the world's (approximately two-thirds; Stutz and Pilkey, 2003) modern-day barrier islands formed by wave processes that are migrating landwards as part of a slow transgression (Reinson, 1992; Clifton, 2006). Barrier islands separate offshore muddy sediments from lagoonal muddy sediments and a key factor in their development is insufficient sand or inadequate routing of sediment within the system to fill the lagoon developed landward of the barrier (Donselaar, 1989; Devine, 1991; Reinson, 1992; Clifton, 2006). This model of barrier-lagoon development has been applied to deposits throughout the Cretaceous Interior Seaway (e.g. Devine, 1991; Sixsmith *et al.*, 2008). Approximately one-third of modern-day barrier islands are developed along regressive shorelines, fed by longshore drift from deltaic systems (Stutz and Pilkey, 2002). Regressive barrier islands have also been interpreted from the rock record e.g. Van Wagoner *et al.* (1990) identified coastal plain mudstones developing contemporaneously with shoreline progradation due to sufficient accommodation space developing during deposition. This regressive model of barrier-bay deposition has been applied to the subsurface Rannoch and Etive formations of the North Sea (e.g. in the Cormorant area, Scott, 1992).

Figure 2.20 Shoreline Classification Scheme. Regressive and transgressive shorelines can be classified on the basis of tidal versus wave power. Volumes of sediment supplied by marine and fluvial processes will also have an impact (after Dalrymple *et al.*, 1992).

2.5.1 Transgressive Deposits of the North Sea

During Jurassic rifting an overall transgression occurred in the North Sea. This transgression was associated with transgressive deposits ranging from Bajocian age in the North Viking Graben to the north, to Kimmeridgian age in the Central Graben to the south (Fig. 2.21). Several examples of these transgressive deposits are discussed here: the Tarbert, Hugin and Fulmar formations which young systematically from north to south.

The Brent Group forms a regressive-transgressive wedge of diachronous coastal-plain and shallow marine sediments in the northern part of the Viking Graben and on the flanking terraces (Yielding *et al.*, 1992). The Brent Group has been subdivided into five formations: Broom (Oseberg Formation in the Norwegian sector), Rannoch and Etive, Ness and Tarbert. These record the northward outbuilding (Broom, Rannoch, Etive and Lower Ness formations) and subsequent retreat (upper Ness and Tarbert formations) of a major wave-dominated delta fed from the south (e.g. Husmo *et al.* 2003).

The Bajocian- to Bathonian-age Ness and Tarbert formations form an overall transgressive package of sediments. Deposition of these units was, at least locally, strongly influenced by rift-related normal faults (Cannon *et al.*, 1992; Mitchener *et al.*, 1992; Glennie, 1998; Mjøs *et al.*, 1998; discussed above). These formations record early deposition in a series of lagoons to the south (Ness Formation) which were protected to the north by a retreating, wave-dominated shoreface and barrier system (Tarbert Formation; Fig. 2.22) (Hampson *et al.*, 2004, 2009). The low-energy, bioturbated, tidally-influenced shoreface deposits of the Tarbert Formation

Figure 2.21 Chronostratigraphic diagram illustrating the retreat of the Brent Delta System. During Jurassic rifting and flooding of the North Sea, several transgressive units were deposited. These young from north to south, from Bajocian-age in the North Viking Graben to Kimmeridgian-age in the Central Graben. The Tarbert, Hugin and Fulmar formations are discussed further in the text (after Partington, *et al.*, 1993).

Figure 2.22 Bajocian Palaeogeography of the Brent Delta. The Bajocian to Bathonian-age Ness and Tarbert formations of the Brent Group form an overall transgressive package of sediments. These formations record early deposition in a series of lagoons to the south (Ness Fm) which were protected to the north by a retreating, high-energy, wave-dominated shoreface and barrier system (Tarbert Fm; modified from Husmo *et al.*, 2003).

are developed above the Ness Formation and were deposited in a series of retrogradationally-stacked, regressive-transgressive tongues (Graue *et al.*, 1987; Hampson *et al.*, 2004; Hampson *et al.*, in press).

In the Hild area of the northern North Sea, five marine transgressive units interfinger with the non-marine Ness Formation (Fig. 2.23; Ravnås and Steel, 1998). Ravinement surfaces within the succession exhibit a rising landward trajectory, enabling the preservation of a thick wedge of transgressive back-barrier deposits beneath this surface. These transgressive systems tracts, capped by ravinement surfaces, are overlain by a regressive strandplain or deltaic package deposited within the highstand systems tract. Despite the overall transgressive nature of the Tarbert Formation, some sequences have an aggradational stacking pattern (Ravnås and Steel, 1998; Cattaneo and Steel, 2003).

To the south, the Brent Group is represented by the younger (Bajocian to Volgian) Vestland Group. The Hugin Formation, which is the focus of part of this PhD study, forms part of this unit. It overlies and interfingers with the continental Sleipner Formation and is itself overlain by and interfingers with the offshore marine Heather Formation (Vollset and Doré, 1984; Partington *et al.*, 1993b; Glennie and Underhill, 1998; Milner and Olsen, 1998; Folkestad and Satur, 2008). The Hugin Formation sits in a stratigraphically-similar position to the Tarbert Formation of the Brent Group, although the Hugin Formation is slightly younger and therefore not directly correlatable. When compared to the Tarbert Formation and Brent Group in general, relatively little has been published about the Hugin Formation (see Milner and Olsen, 1998, Varadi *et al.*, 1998 and Folkestad and Satur, 2008 for exceptions). It

Figure 2.23 Tarbert Formation in the Hild-South Alwyn area of the Viking Graben in the northern North Sea (from Ravnås and Steel, 1998). Each wave ravinement surface rises landwards (towards the right) and preserves a thickening wedge of transgressive back-barrier deposits below.

should be noted that several other names have been used to describe the same lithostratigraphic interval, including the Beryl Formation and the Bruce Group sands (Mitchener *et al.*, 1992; Richards *et al.*, 1993; Husmo *et al.*, 2003).

The Fulmar Formation is part of the overall backstepping “Brent Delta” system and records the first marine sedimentation within the Central Graben following subsidence of the Central North Sea dome (Underhill, 1998). Deposition within the Fulmar Formation records several discrete progradational to aggradational pulses during the overall transgression (Gowland, 1996). Within this unit, upward-coarsening successions are observed (Johnson *et al.*, 1985) but the dominant architecture is of an overall transgressive trend, recording an overall relative sea-level rise (Martin, 1993). Locally, the Fulmar Formation is capped by a sharp boundary that represents an unconformity generated by localised uplift due to halokinesis and/or footwall uplift (Martin, 1993). Regionally, however, it is still overlain by the deep marine Kimmeridge Clay Formation, suggesting a rapid deepening event and continued transgression during the rift climax phase of the Jurassic rift event. On the Western Platform, salt-influenced topography strongly affected the local deposition of the Fulmar Formation (see Section 2.4.1).

2.5.2 Transgressive deposits of the Cretaceous Interior Seaway

The Cretaceous Interior Seaway was an epicontinental sea that developed across the North American continent during the Cretaceous. The seaway extended from arctic Canada to Texas, separating North America into an eastern Appalachian mega-island of low relief and a western Cordilleran mountain chain, the Sevier orogenic belt. Flexural loading by the Sevier fold-and-thrust belt to the west combined with long-

wavelength thermal subsidence due to subduction of the Pacific plate combined causing subsidence of a large basin that was flooded rapidly by marine waters (e.g. Armstrong, 1968; Hendricks, 1983; Cross, 1986; Roehler, 1989; Liu and Nummedal, 2004). Eustatic sea-level variations during the Cretaceous, combined with variable sediment supply and thrust sheet loading along the western margin, led to deposition of a series of large-scale, transgressive-regressive depositional cycles within the basin (Molenaar, 1983; Donselaar, 1989; Krystinik and DeJarnett, 1995; Olsen *et al.*, 1999). Many of these cycles and their constituent units from the western margin of the seaway are extensively exposed in the southwestern US, including the three examples of net-transgressive strata described briefly below: the Cliff House Sandstone, Hosta Tongue and upper Almond Formation.

The Campanian Cliff House Sandstone is one of the most intensively studied transgressive deposits associated with the seaway, based on outcrops from southwestern Colorado and New Mexico. In a palaeo-landward (SW) direction, the Cliff House Sandstone pinches out into the coastal plain and lagoonal deposits of the Upper Menefee Formation (Fig. 2.24). In a palaeo-seaward direction (NE) the unit passes into the Lewis Shale (Donselaar, 1989; Mjøs *et al.*, 1998; Olsen *et al.*, 1999). In Mancos Canyon, SW Colorado, the Cliff House Sandstone comprises a stacked series of progradational shoreface sandstone tongues that were deposited within a net-transgressive system. Each tongue is up to 55 m thick and is separated by coastal plain sediments up to 15 m thick. (Olsen *et al.*, 1999). The base of each regressive sandstone tongue is marked by a wave ravinement surface (Mjøs *et al.*, 1998; Olsen *et al.*, 1999). The sandstone tongues are thinnest over the maximum sandstone development of the underlying tongue and expand landwards of the pinchout of the

Figure 2.24 Schematic chronostratigraphic cross section showing Upper Cretaceous rocks in the San Juan Basin, New Mexico. The inset map shows a schematic palaeogeographic reconstruction of the Western Interior Seaway during middle Cretaceous. Approximate position of the cross section is highlighted. The Cliff House and Hosta sandstones are highlighted and discussed further in the text (from Sixsmith et al., 2008).

underlying tongue (Olsen *et al.*, 1999). This relationship is controlled by syn-sedimentary compaction and highlights the importance of topography on sediment distribution and preserved stratal architecture within net-transgressive depositional systems. In more basinal positions, coastal plain sediments are missing and shoreface sandstone tongues directly overlie each other (Olsen *et al.*, 1999). During the overall transgression, underlying progradation tongues were partly eroded by wave ravinement at the base of the overlying back-stepping shoreface. Sediments deposited in topographic lows had the best preservation potential (Donselaar, 1989). The sediments best preserved during transgression are limited to lagoonal deposits, which include tidally-influenced channels that eroded into the underlying shoreface tongues in their most landward position (Olsen *et al.*, 1999).

In the Chaco-Tsaya Canyon area NW New Mexico, the Cliff House Sandstone comprises a 90 m thick, linear (shoreline-parallel) sandstone complex (Donselaar, 1989). This complex is composed of four sandstone bodies that range in thickness from 15-25 m along depositional strike (NW-SE), the lithofacies within the sandstone bodies remain continuous. In a dip-direction, however, the sandstones display a lenticular geometry and pass rapidly (over 9 km) from lagoonal to upper shoreface to lower shoreface (Donselaar, 1989). These geometrical and spatial relationships suggest that each sandstone body was deposited as a barrier-island complex during a regression, which entailed the lagoon behind the barrier being gradually infilled. During transgression, the barrier coastline retreated landward and significant volumes of lagoonal deposits were eroded by wave ravinement (Donselaar, 1989).

The net-transgressive Hosta Tongue (Santonian) of the Point Lookout Sandstone in New Mexico comprises a wave-dominated shoreface and a tide-dominated, back-barrier lagoon that co-existed in a mesotidal barrier island system (Sixsmith *et al.*, 2008). The back-barrier lagoon is reconstructed to have comprised a broad, sandy platform dissected by “blind” channel networks that extended landward of the tidal inlets that punctuated a barrier island chain. These channel networks were highly mobile along depositional strike and thus form a shoreline-parallel belt of channel-fill deposits, while many individual channel segments contain palaeocurrents oriented sub-parallel to the barrier-island shoreline trend. Tide-dominated deposits (channel-fill sandstones, sheet-form sandstones and tidal flat heteroliths) that are the preserved remnants of these lagoons pass abruptly down-dip into wave-dominated shoreface sandstones (Fig. 2.25). The tide-dominated and wave-dominated packages are separated by wave ravinement surfaces, whose distribution can be used to reconstruct shoreline trajectory (Fig. 2.25) (Sixsmith *et al.*, 2008).

The upper part of the Maestrichtian Almond Formation is another example of a transgressive unit deposited on the western shoreline of the Cretaceous Interior Seaway. The unit is exposed on the flanks of the Rock Springs Uplift in southern Wyoming. Part of the Upper Cretaceous Mesaverde Group, it overlies the fluvial Ericsson Formation and is overlain by and interfingers with the offshore marine Lewis Shale (e.g. Jacka, 1965; Hendricks, 1994; Garcia-Gonzales *et al.*, 1997). The Almond Formation has been subdivided into a lower alluvial unit and an upper shallow marine unit (Jacka, 1965). The upper Almond Formation records an overall transgression, and was deposited during a series of high frequency sea level rises (Jacka, 1965; Hendricks, 1994; Horn *et al.*, 2002). Several sub-environments have

Figure 2.25 Correlation panel through a depositional dip profile, the Hosta Sandstone. In a palaeo-landward (SW) direction the Hosta Sandstone pinches out into coastal plain and lagoonal deposits of the Upper Menefee Formation. In a palaeo-seaward direction (NE) the unit passes into the Lewis Shale (from Sixsmith et al., 2008).

been documented within the shallow marine Almond Formation, including tidal channel, lagoonal, estuarine, tidal delta and marsh (Hendricks, 1994; Caldaro-Baird *et al.*, 2001). Due to similarity in facies and depositional setting, the Almond Formation has been studied as an outcrop analogue for the Hugin Formation and is discussed in more detail in Chapter 6.

CHAPTER 3

Geological Setting

This chapter describes the structural and stratigraphic setting of the study area. Firstly the structural development of the North Sea Rift basin is summarised, followed by a more detailed discussion of the structural evolution of the South Viking Graben. The generalised lithostratigraphy of the Viking Graben is then discussed, before a more detailed introduction to the Jurassic stratigraphy within the study area, comprising the Sleipner and Hugin formations (Vestland Group) and the overlying Viking Group.

3.1 Tectonic Evolution of the North Sea Rift

Deposition of the Hugin Formation was coeval with the earliest phase of Mesozoic rifting, within the study area, and halokinesis of the underlying Zechstein Group salt (e.g. Cockings *et al.*, 1993; Husmo *et al.*, 2003). The structural evolution of the study area, in particular the timing and distribution of rift-related faulting and halokinesis, therefore provided a template that controlled deposition of the Hugin Formation.

The North Sea rift system lies in the offshore region between Scotland and Norway and is a failed arm of the Arctic-North Atlantic rift system that commenced rifting in the Norwegian-Greenland Sea area during the Late Carboniferous (Ziegler, 1992). It is a three-armed (trilete) failed-rift system comprising the Viking Graben, Moray

Firth Basin and the Central Graben (Fig. 3.1; e.g. Erratt *et al.* 1999; Zanella and Coward, 2003). The Viking Graben and Moray Firth Basin are asymmetrical in cross-section, with major faults dipping to the E/ESE on the western margin or SE/SSE on the northern margin of the respective grabens, whilst the Central Graben is more symmetrical in cross-section.

The North Sea rift overlies the plates of three ancient continents Avalonia, Laurentia and Baltica that collided in the Palaeozoic during the Caledonian Orogeny (Fig. 3.2a; Lyngsie *et al.*, 2006). The Iapetus Suture marks the eastern boundary of Laurentia whilst the Tornquist Suture Zone or lineament marks the southern boundary of Baltica. Since the Cambrian four major tectonic events have influenced the region: (1) Late Ordovician-Early Silurian collision of the Caledonian Orogeny, (2) Carboniferous to Permian rifting and basin formation, coeval with the Variscan Orogeny, (3) Mesozoic rifting and formation of the trilete North Sea graben system, and (4) Late Cretaceous to Early Tertiary inversion. The location of the Mesozoic North Sea rift system was partly controlled by basement lineaments, including the suture zones between these three plates (e.g. Bartholomew *et al.*, 1993; Zanella and Coward, 2003; Lyngsie *et al.*, 2006).

During the late Carboniferous to Early Permian, strike-slip deformation dominated the future position of the North Sea Basin, with extensional wrench tectonics thinning the crust (e.g. Coward, 1993). This crustal thinning led to formation and subsidence of the North and South Permian Basins, which were separated by the Mid North Sea High (Fig. 3.2b). The South Viking Graben formed an embayment at the northern margin of the North Permian Basin at this time (discussed below) and

Figure 3.1 North Sea Trilete Rift System (After Husmo *et al.*, 2003). Base Cretaceous (top syn-rift) surface map of the Central and Northern North Sea. The trilete rift system comprises the Viking Graben, Moray Firth and Central Graben. The rift system developed as a result of Mid to Late Jurassic extension during active rifting within the area.

Figure 3.2 Tectonic Evolution of the North Sea Rift. (a) Sketch map of the North Sea basement and tectonic features (After Zanella *et al.*, 2003). The North Sea rift system overlies the plates of three palaeo-continent: Avalonia, Laurentia and Baltica. These collided during the Paleozoic Caledonian Orogeny. The location of the Mesozoic rift system was partly controlled by basement lineaments, including the suture zones between these three plates. (b) Sketch map of Carboniferous to Permian structural elements (after Coward *et al.*, 2003). The position of the North and South permian Basins are shown relative to the Mesozoic rift system.

deposition of continental and evaporitic stratigraphic units was widespread across the area (e.g. Williams, 1993; Clark *et al.*, 1998; Lyngsie *et al.*, 2006).

The Early Jurassic (Toarcian-Aalenian) saw regional thermal uplift of the Mid North Sea Dome due to impingement of a mantle plume at the base of the lithosphere; this uplift preceded active rifting within the region (e.g. Ziegler, 1990; Underhill and Partington, 1993; Davies *et al.*, 1999; Coward *et al.*, 2003). The subsequent major phase of rifting during the Middle to Late Jurassic led to linkage of the Arctic and Central Atlantic rift systems and development of a triple junction between the Viking Graben, Moray Firth Basin and Central North Sea Basin. Models for the development of the triple junction fall into two broad categories: those that propose a single-phase model of opening (e.g. Roberts *et al.*, 1990; Bartholomew *et al.*, 1993) and those that infer a multi-phase model (e.g. Erratt *et al.*, 1999; Davies *et al.*, 2001). For example, Roberts *et al.* (1990) proposed that the major basins within the rift are dip-slip basins, which opened simultaneously due to orthogonal movement on the main intra-basinal and basin-margin faults. As an alternative, Davies *et al.* (2001) argued that a rotation of 135° in the stress field (from E-W to NE-SW) over ten million years would be required to generate the fault trends observed in the region. In this latter interpretation, extension was accommodated during the Bathonian and Callovian on N-S and NNE-SSW trending faults, then switched to NE-SW and E-W trending structures during the Oxfordian, and NW-SE trending structures during the main phase of Kimmeridgian and Volgian rifting (Davies *et al.*, 2001).

During the Early to Middle Cretaceous rifting migrated to the site of the proto-North Atlantic Rift. Rifting had therefore effectively ceased in the North Sea by the Mid-Cretaceous (e.g. Williams, 1993; Erratt *et al.*, 1999; Coward *et al.*, 2003). Late Campanian inversion is recorded regionally and linked to compressional stress related to the Alpine Orogeny (Erratt *et al.*, 1999; Lyngsie *et al.*, 2006). The Atlantic rift zone achieved crustal separation by the Early Tertiary, by which time the North Sea rift system had become a thermally-driven ‘sag’ basin, in which regional thermal subsidence was disturbed by localised reactivation of salt-cored structures (Ziegler, 1975; Erratt, 1999; Lyngsie *et al.*, 2006).

3.2 Structural Development of the South Viking Graben

The Viking Graben consists of three segments: the North, Central and South Viking Grabens. These segments are separated by a series of transfer zones that trend approximately WNW-ESE, interpreted to relate to older deep-seated structures (Thomas and Coward, 1996). This study focuses on the southern part of the South Viking Graben.

The South Viking Graben is a NNE-SSW-trending half-graben, bounded to the west by a series of ESE-dipping faults (Figs. 3.3, 3.4; Thomas and Coward, 1996). To the east the graben shallows onto the Utsira High, a long-lived, relatively unfaulted basement-cored structural high (Thomas and Coward, 1996). Within the South Viking Graben itself, normal faulting follows a broadly NNE-trending structural grain, although NW-SE-trending faults laterally compartmentalize the graben (Pegrum and Ljones, 1984; Thomas and Coward, 1996).

Figure 3.3 Structure map of the South Viking Graben (after Thomas and Coward, 1996). The graben is a NNE-SSW trending half graben, bound to the west by a series of ESE dipping faults. To the east, the graben shallows onto the Utsira High. The hangingwall is dissected by a series of normal faults, the majority of which developed antithetic to the main graben bounding fault.

Figure 3.4 Structural Cross section through the study area within the South Viking Graben. (a) Line from the north of the study area (modified Sneider, *et al.*, 1995). The Jurassic and Cretaceous units thicken towards the main graben bounding fault to the west. Small scale thickness variations are also observed across antithetic faults developed on the hangingwall dipslope. (b) Close up of the hanging-wall dipslope highlights the thin skinned extension and gravity-driven faulting that detaches on the underlying Zechstein layer. An inversion structure can be observed in the hangingwall of the western fault. Onlap events are observed within the lowermost Cretaceous onto the crest of the structure highlighting it was growing at this time related to the Late Jurassic-Early Cretaceous event. (c) Line from the south of the study area where thinner Zechstein units have enabled complex basement involved fault systems to develop in the supra-salt section (from Kane and Jackson, 2008).

Formation of the South Viking Graben dates from E-W extension during the Permo-Triassic rift event and thermal subsidence of the North Permian Salt Basin (Ziegler, 1975, 1992; Steel, 1993; Bartholomew *et al.*, 1993; Glennie, 1998; Coward *et al.*, 2003). The South Viking Graben formed a NNE-SSW-trending, fault-bounded embayment along its northern margin (Ziegler, 1990; Hodgson *et al.*, 1992; Glennie, 1998; Coward *et al.*, 2003). Thickening of Upper Triassic units towards the Graben Bounding Fault suggests extension continued to this time, although the magnitude of thickening and thus differential subsidence was minor compared to that of the Jurassic rift event (Roberts *et al.*, 1993; Thomas and Coward, 1996; Knott, 2001; Lyngsie, 2006; Jackson and Larsen, 2008). Mobilisation of Zechstein Group salt occurred at this time generating salt pillows, diapirs and adjacent depocentres (e.g. Pegrum and Ljones, 1984; Frostrick, 1992; Thomas and Coward, 1996; Kane *et al.*, in review). The Early Jurassic (Toarcian-Aalenian) coincided with regional thermal doming and uplift of the Mid North Sea Dome (a precursor to Jurassic rifting), erosion and/or non-deposition of Early Jurassic units and the local extrusion of volcanics (Milton, 1993; Williams, 1993; Underhill and Partington, 1993; Davies *et al.*, 1999; Knott, 2001). Late Middle Jurassic (Bajocian-Bathonian) collapse of this thermal dome and reactivation of the Graben Bounding Fault Zone coincided with the commencement of Jurassic rifting and initial flooding of the South Viking Graben (Cockings *et al.*, 1992; Thomas and Coward, 1996; Coward *et al.*, 2003).

Extension and associated activity on the Graben Bounding Fault Zone continued throughout the Middle and Upper Jurassic. Extension has variably been constrained to the following time intervals: Late Bathonian to Oxfordian (Williams, 1993); Late

Bajocian to Oxfordian (Færseth *et al.*, 1997); Late Callovian to Late Kimmeridgian (Cockings *et al.*, 1992); and Callovian and Oxfordian to Volgian (Knott, 2001). However, all authors agree that the major phase of rifting within the South Viking Graben occurred during the Upper Jurassic (Oxfordian to Volgian), after deposition of the Hugin Formation (Cockings *et al.*, 1992; Bartholomew *et al.*, 1993; Rattey and Hayward, 1993). At the same time structural development within the graben was influenced by the underlying units of mobile Zechstein Group salt, which were reactivated by the extension (Pegrum and Ljones, 1984; Kane *et al.*, in review). Where thickest the salt-rich units also acted as a décollement surface, leading to development of thin-skinned structures above the salt and the detachment of supra- from sub-salt structures (Fig. 3.4a; Pegrum and Ljones, 1984; Thomas and Coward, 1996; Jackson and Larsen, 2009). Regional tilt as a result of hangingwall rotation caused local gravity sliding, with extensional forced folds developed up-slope above basement-involved faults whilst down slope salt-cored folds developed as a result of buckling (Thomas and Coward, 1996). Where salt units are thinner or missing, complex basement-involved fault systems have developed (Fig. 3.4c; Pegrum and Ljones, 1984).

By the latest Jurassic, rift-related faults had become inactive and a period of compression with associated contractional structures has been identified within the graben (Pegrum and Ljones, 1984; Cherry, 1993; Knott, 2001; Branther, 2003; Jackson and Larsen, 2008). This period of compression dates from the Late Jurassic to Early Cretaceous (anticline developed at top of Upper Jurassic, Fig. 3.4a; Cherry, 1993; Thomas and Coward, 1996; Branther, 2003; Jackson and Larsen, 2008). Within the study area, recent work by Jackson and Larsen (2008) constrained this inversion

event further, identifying that inversion commenced in the Late Jurassic and continued for 42 Ma, from the late Early Volgian to Late Albian. The origin of compression is unknown, although the following models have been proposed: extension in neighbouring basins (Bartholomew, *et al.* 1993); a change in extension direction which led to the rotation of earlier fault blocks, causing local compression on the basin margins (Thomas and Coward, 1995; Zanella and Coward, 2003); hangingwall rollover into the graben bounding fault (Branter, 2003); and the Alpine Orogeny compressional event (Pegrum and Ljones, 1984; Lyngsie *et al.*, 2006).

3.3 Generalised lithostratigraphy of the Viking Graben

During the Permian, continental deposition prevailed over the northern North Sea in the North Permian Basin (Fig. 3.2b), with continental clastic strata of the Rotliegend Formation being overlain by the evaporitic Zechstein Group (Smith and Taylor, 1992; Thomas and Coward, 1996; Glennie, 1998; Glennie *et al.*, 2003). Within the South Viking Graben, the Zechstein Group is composed of dolomites and anhydrites to the north, near the margin of the North Permian Basin, whilst halite dominates further to the south east, towards the basin centre (Fig. 3.5; Thomas and Coward, 1996).

Continental conditions continued into the Triassic, with lacustrine and alluvial deposition being represented by the Smith Bank and Skagerrak formations (Warrington and Ivimey-Cook, 1992; Glennie, 1998; Goldsmith *et al.*, 2003). Significant thickness variations in Triassic strata occur within the graben and are

Figure 3.5 Schematic Zechstein facies map (after Thomas and Coward, 1996). During the Permian, the South Viking Graben formed a fault-bounded embayment at the northern margin of the North Permian Salt Basin. To the north the Zechstein Group is dominated by dolomites and anhydrites whilst halite dominates further to the south east. Facies and thickness of this unit impacts structural development within the graben.

interpreted to be the result of normal faulting and the movement of the underlying Zechstein Group salt (e.g. Pegrum and Ljones, 1984; Thomas and Coward, 1996).

The Jurassic saw deposition in the central and north Viking Graben of the Brent Group, which records the development of a large delta system (the Brent Delta) fed mainly from the south. The main source of sediment was from erosion of the Mid North Sea Dome to the south, and laterally from the Shetland Platform and southern Norway (Mearns, 1992; Morton, 1992a). The Brent Delta prograded northwards from the Mid North Sea Dome to the northern part of the North Viking Graben (Fig. 3.6a; Bradshaw *et al.*, 1992; Helland-Hansen *et al.*, 1992; Mitchener *et al.*, 1992; Morton *et al.*, 1992b; Husmo *et al.*, 2003). At this time, the South Viking Graben corresponded to the uplifted region of the Mid North Sea Dome and an unconformity is observed at this time within the graben. Middle Jurassic collapse of the dome and reactivation of earlier fault systems led to a relative rise in sea-level (Underhill and Partington, 1993) and deposition of coastal plain and then shallow marine deposits (Vestland Group) within the graben (Fig. 3.6b; Harris and Fowler, 1987; Cockings *et al.*, 1992). Relative sea level continued to rise during the Late Jurassic due to fault-related subsidence and eustatic rise (Cockings *et al.*, 1992), which led to continued deepening of the basin and deposition of shelfal and deep-marine deposits of the Viking Group (Bradshaw *et al.*, 1992; Fraser *et al.*, 2003).

During the Early Cretaceous, siliciclastic deposition of the Cromer Knoll Group prevailed over the Northern North Sea, with mud-prone shelfal deposits and marls being deposited in the South Viking Graben (Hancock and Rawson, 1992; Glennie, 1998). There was widespread chalk deposition in northwest Europe during the Upper

Figure 3.6 Palaeoenvironmental maps of the Northern North Sea (Husmo et al., 2003). (a) Bajocian Times - widespread regression as a result of doming in the Mid North Sea saw Brent Delta deposition to the north of the Viking Graben. (b) Mid- to Late Callovian - collapse of the North Sea Dome and reactivation of earlier fault systems led to transgression and deposition of the Vestland and Viking groups within the South Viking Graben.

Cretaceous, although Shetland Group mudstones were deposited in the deeper waters of the South Viking Graben (Hancock and Rawson, 1992; Glennie, 1998). A low-energy, offshore, relatively deep water depositional setting continued into the Cenozoic, with muds and deepwater turbidite sands being deposited within the graben (Murray, 1992; Glennie, 1998).

3.4 Stratigraphy of the Vestland Group

The Vestland Group comprises the continental Sleipner and shallow-marine Hugin formations (Fig. 3.7; Vollset and Doré, 1984). The group is overlain by and interfingers with the open marine shales of the Heather and Draupne formations that together form the Viking Group (Fig. 3.7). The lithostratigraphic units within the Vestland Group are strongly diachronous and range in age from Early Bathonian to Early Oxfordian, typically defining an overall younging from north to south along the axis of the basin (Fig. 3.7; Mitchener *et al.*, 1992; Rattey and Hayward, 1993; Partington *et al.*, 1993; Richards *et al.*, 1993).

3.4.1 Sleipner Formation

The Sleipner Formation unconformably overlies the Triassic Skagerrak Formation and is itself overlain by, and interfingers with, the Hugin Formation (Husmo *et al.*, 2003). The Sleipner Formation comprises interbedded sandstones, siltstones and mudstones with extensive coal layers (Milner and Olsen, 1998; Husmo *et al.*, 2003). It is interpreted to have been deposited in a coastal plain setting (Milner and Olsen, 1998; Husmo *et al.*, 2003), although in the Beryl embayment to the North of the study area a more estuarine depositional setting is interpreted (Richards, 1991).

Figure 3.7 Jurassic Lithostratigraphy of the Study Area. The Vestland Group comprises the continental deposits of the Sleipner Formation and the shallow marine deposits of the Hugin Formation. These lithostratigraphic units show strong diachroneity to their development and broadly young from north to south. They are overlain by the open marine shales of the Viking Group.

Several other names have been used to describe the same lithostratigraphic interval within the UK North Sea including the Pentland Formation, Middle Beryl Sandstone, Linhe Formation and Bryne Formation (Knox and Cordey, 1993).

3.4.2 Hugin Formation

The Hugin Formation was first described from the Sleipner Vest Field to the south of the study area (well 15/9-2; Vollset and Doré, 1984). In this location it was interpreted to comprise a stacked succession of nearshore-to-shallow marine sandstones and coals with some fluvio-deltaic influence. More regional data, however, suggest that the formation represents a wave-dominated barrier shoreline in some locations (Harris and Fowler, 1987). Detailed studies, focussed on the Sleipner Vest and Øst fields, have further refined the Hugin Formation sedimentological and sequence stratigraphic model. Several facies associations have been interpreted within the succession, including; alluvial plain, coastal swamp, barrier-shoreline complex (including shoreface, mouth bar, tidal channel, tidal dune, and tidal flat deposits), lagoon and /or bay, and offshore deposits, with the last corresponding to the shelfal Heather Formation (Milner and Olsen, 1998; Folkestad and Satur, 2008). Tidal signatures have been recognised throughout the Hugin Formation (e.g. Folkestad and Satur, 2008) and are interpreted to be the result of wave suppression within the narrow (*ca.* 30km), elongate (*ca.* 100-120 km) depocentre of the South Viking Graben (Richards, 1991; Folkestad and Satur, 2008). Several other names have been used to describe the same lithostratigraphic interval, including the Beryl Formation and “Bruce Group sands” in the Bruce-Beryl embayment of the UK sector (Mitchener *et al.*, 1992; Richards *et al.*, 1993; Husmo *et al.*, 2003).

3.4.3 Viking Group

The Bathonian- to Volgian-age Viking Group comprises open marine shales and siltstones of the Heather Formation and overlying Draupne Formation. The Viking Group is equivalent to the Humber Group of the UK sector. The name Heather Formation has been applied throughout the North Sea to units of grey, silty mudstone lying between the coarse clastics of the Brent Group (and its equivalents) and the organic rich shales of the Kimmeridge Clay Formation (Deegan and Scull, 1977). The lithostratigraphic term Kimmeridge Clay Formation is applied within the UK sector of the graben and is synonymous with the Draupne Formation (Underhill, 1998).

Earliest deposition of this group is interpreted to have been restricted to the graben axis, with time-equivalent coastal plain and shallow marine deposits (Sleipner and Hugin formations) persisting at the graben margins (Husmo *et al.*, 2003). The Heather Formation typically overlies and interfingers with the Hugin Formation, although in some areas it is observed to conformably overlie coastal plain deposits of the Sleipner Formation. Deep-marine turbidites deposited in the basin axis have been referred to as either the Ling Sandstone Member of the Heather Formation (Cockings *et al.*, 1992; Husmo *et al.*, 2003) or as part of the Bruce or Heather Sandstones (Richards, *et al.*, 1993; Spence and Kreutz, 2003). The source of these sandstones has been debated, with both the footwall of the main Graben Bounding Fault to the west and the coeval Hugin coastline to the south and east representing potential source areas (Cockings *et al.*, 1992; Spence and Kreutz, 2003).

CHAPTER 4

Subsurface Dataset and Methodology

Core, wireline-log and biostratigraphical data from wells have been integrated with 3D seismic data to develop an understanding of the structural development within the study area and its impact on facies distributions and stratigraphic architecture within the Hugin Formation. The dataset is summarised in this section (highlighted on Fig. 4.1) along with a brief discussion of the methodology utilised.

4.1 Well Data

Well data comprising cores, wireline logs and biostratigraphical analyses were available to the study and are summarised below. Core logs are included in Appendix 1.

4.1.1 Core data

Almost 750 m of core through the Middle Jurassic interval (Sleipner, Hugin and Heather formations) has been logged from 15 wells within the Norwegian sector of the South Viking Graben, the focus of this study, and an additional 6 wells from the UK sector of the graben (Fig 4.1). Undertaken at a scale of 1:50, the logged core data have been used for sedimentological and stratigraphical analysis of the shallow marine Hugin Formation and coeval deposits. Cored intervals have been depth shifted to calibrate with wireline data based on distinctive lithology responses. Logged wells with core, including the cored intervals and applied depth shift, are summarised in Table 4.1.

Figure 4.1 Database Map. Well data comprising core, raw wireline-log and biostratigraphical analysis were available to the study and are summarised above. The extent of the 3D seismic dataset is also highlighted.

	Well Name	Core No.	Top Depth (m)	Base Depth (m)	Length of core	Depth shift (m)
Norwegian sector wells	15/3-2R	2	4406.00	4407.63	1.63	-1.50
		3	4565.00	4573.65	8.65	-1.50
		4	4662.00	4656.00	6.00	-1.00
	15/3-3	4	4562.00	4547.00	15.00	-3.50
	15/3-4	2	3792.00	3795.60	3.60	0
		3	3796.10	3807.00	10.90	-3.70
		4	3813.00	3821.00	8.00	-2.35
	15/3-5	5	3821.00	3839.27	18.27	-3.37
		1	3971.00	3988.80	17.80	2.75
		2	3989.00	4001.00	12.00	2.75
	15/3-7	3	4020.00	4029.00	9.00	4.00
		1	4609.00	4615.67	6.67	0
		2	4616.60	4617.50	0.90	0
	15/5-1	3	4620.50	4633.60	13.10	-1.70
		4	4643.00	4668.76	25.76	-1.90
		1	3561.00	3577.89	16.89	-1.80
	15/5-2	2	3578.50	3590.20	11.70	-1.80
		3	3591.10	3600.43	9.33	-1.80
		4	3609.00	3617.40	8.40	-1.75
	15/6-2R	5	3623.00	3624.50	1.50	0
		1	4013.60	4018.75	5.15	0
		2	4032.50	4042.48	9.98	0
	15/6-4	1	3582.40	3591.46	9.06	2.80
		2	3591.50	3603.60	12.10	2.80
		3	3603.55	3618.00	14.45	3.25
	15/6-5	4	3618.00	3626.50	8.50	3.25
		5	3626.70	3641.14	14.44	3.50
		1	3207.60	3213.30	5.70	0
	15/6-6	2	3214.20	3227.60	13.40	0
		1	3629.00	3647.50	18.50	2.00
		2	3647.50	3655.67	8.17	2.00
	15/6-9S	3	3656.43	3674.03	17.60	2.00
		1	3591.00	3609.50	18.50	1.50
		2	3609.50	3625.50	16.00	1.50
	15/8-1	1	3761.30	3810.70	49.40	X
	15/9-1*	1	3658.00	3667.50	9.50	-0.25
		2	3667.50	3678.60	11.10	0.00
		3	3678.80	3680.60	1.80	?
	15/9-12	4	3681.86	3699.00	17.14	-1.00
		5	3699.00	3706.20	7.20	0.00
		1	3521.00	3539.30	18.30	X
UK sector wells	16/8A-4	1	3504.50	3522.20	17.70	X
		2	3522.25	3539.95	17.70	X
		3	3540.20	3553.33	13.13	X
	16/18-1	4	3554.00	3572.00	18.00	X
		5	3572.20	3582.20	10.00	X
		1	3685.75	3704.27	18.52	3.15
	16/23-5	1	3516.55	3533.53	16.98	X
	16/7A-33	12	4757.50	4791.80	34.30	X
		13	4791.80	4800.03	8.23	X
		14	4800.61	4823.78	23.17	X
	16/18-2	15	4823.78	4836.46	12.68	X
		16	4837.8	4847.56	9.76	X
		5	4791.00	4801.84	10.84	X
	16/07a-30Z	1	4824.98	4828.95	3.97	X
	16/7A-33	1	5595.43	5600.46	5.03	X

Table 4.1 Summary of core data logged from the 21 wells available to the study. Almost all of the wells were logged at a scale of 1:50; the exception is highlighted *. Core shifts could not be determined for wells for which wireline log data were not available; these have also been highlighted X.

The core description involved recording grain size, sorting and sphericity of matrix and clasts, colour variations, sedimentary structures, bed thickness, the nature of bounding contacts, macrofauna type and preservation state. The degree of bioturbation was recorded using the bioturbation index (BI) scheme of Taylor and Goldring (1993). This scheme involves a visual assessment of the amount of “bioturbated” versus “unbioturbated” sediment and grades this parameter on a scale of 0 to 6 (Table 4.2). Recording variability in the degree of bioturbation can provide important information on the rate of sedimentation or the occurrence of erosion, which in turn helps to define depositional facies and hiatal surfaces (Taylor and Gawthorpe, 1993).

Grade	Percentage Bioturbated	Classification
0	0	No bioturbation
1	1 to 5	Sparse bioturbation: few discrete traces and/or escape structures
2	6 to 30	Low bioturbation: bedding distinct, low trace density, escape structures often common
3	31 to 60	Moderate bioturbation: bedding boundaries sharp, traces discrete
4	61 to 90	High bioturbation: bedding boundaries indistinct, high trace density with overlap of traces common
5	91 to 99	Intense bioturbation: bedding completely disturbed, limited reworking due to repeated overprinting
6	100	Complete bioturbation: sediment reworking due to repeated overprinting

Table 4.2 Summary of Bioturbation Index (after Taylor and Goldring, 1993).

4.1.2 Biostratigraphical data

The available biostratigraphical data were derived from a proprietary evaluation of Late and Middle Jurassic sediments in the South Viking Graben area by StatoilHydro ASA (Tjemsland, 2007). Previously collected data from 18 wells in the vicinity of the Gudrun and Sigrun fields and directly adjacent areas to the east and south (blocks 16/1, 15/6 and 15/5) were included in the evaluation, (Tjemsland, 2007). The wells included in this review are indicated on the database map (Fig. 4.1).

Fossil groups used in the study included pollen, spores, nannofossils and palynomorphs. Data quality typically allowed for biostratigraphical resolution and palynological zonation with a resolution close to 500,000 years within the best sampled sections of marine and marginal marine facies (Fig. 4.2; Draupne, Heather and Hugin formations; Tjemsland 2007). Lower resolution occurred, especially within the Late Jurassic, where the database was insufficient due to age and scattered descriptions of assemblages. The identified assemblages in samples from the continental Sleipner Formation consist of long ranging pollen and spores, which precludes detailed zonation of this interval (Tjemsland, 2007).

Four main stratigraphic breaks (unconformities) have been identified within the succession; Early Jurassic, Middle Jurassic, Late Jurassic and Base Cretaceous. The Early Jurassic unconformity is observed regionally at the base of the Sleipner Formation, where it directly overlies the Triassic Skagerrak Formation. This unconformity is interpreted to be the result of thermal uplift of the central North Sea as a precursor to active rifting within the region. The Base Cretaceous Unconformity

(BCU) is identified regionally and represents a missing 3-13 Ma stratigraphy (Tjemsland 2007). Development of this unconformity is interpreted to reflect a regional cessation of deposition combined with localised uplift and erosion of structures. The Middle and Late Jurassic unconformities are more locally developed and are interpreted to be the result of the development of specific structures and localised periods of marine erosion or sediment bypass.

Flooding surfaces were proposed by the biostratigrapher where marine faunal assemblages of either high diversity and/or abundance were identified. As part of this study, these observations were integrated with core and wireline-log characteristics to iteratively refine the position of the surface (i.e. where they corresponded to thick offshore mudstones overlying shallow marine deposits). Where there was uncertainty in the position of the surface based on core and wireline-log data, discussions were held with the biostratigrapher to establish the reliability of the biostratigraphy data over that interval. Specific surfaces have been identified using the nomenclature of Partington *et al.* (1992; Fig. 4.2). Within the Hugin Formation in the Sigrun-Gudrun study area (Chapter 7), higher resolution flooding surfaces of more local extent have been identified. These surfaces are characterised by large occurrences of low diversity dinocyst assemblages that indicate subtle marine incursions into a marginal marine environment. Flooding surfaces (and the units they bound) identified from core and wireline-log data have been correlated using the StatoilHydro ASA in-house scheme of zonation defined by palynological events (Fig. 4.2; Tjemsland, 2007).

4.1.3 Wireline log data

Complete suites of well logs (i.e. gamma ray, neutron porosity, density and sonic logs) were available for 17 wells within the study area, including 12 wells with available core data (Fig. 4.1). Raw wireline-log data were not available for the remaining wells logged within the UK sector of the graben, however microfiche composite logs were reviewed during core logging at the British Geological Survey core store whilst published wireline-logs were also available (e.g. Cockings *et al.*, 1992; Husmo *et al.*, 2003; Spence and Kreutz, 2003). In addition, wireline log data from a further eight wells within the study area have been utilised for regional correlation panels and to constrain palaeogeographical reconstructions. Wireline-log character and cross-plot analyses have been used to identify distinctive wireline-log responses for each of the facies identified from core data. Cross-plot analysis was undertaken using gamma ray versus sonic (GR vs. DT) and neutron porosity versus density (NPHI vs. RHOB) cross plots (Chapter 5). The characteristic wireline-log responses were then used to interpret the depositional setting of uncored intervals within the study area. The correlation panels in Chapters 5, 7 and 8 typically display gamma ray curves at a scale of 0-150 API units on the left and sonic curves at a scale of 140-40 microseconds/foot on the right.

4.2 Seismic Data

The extent of 3D seismic data available to the project is shown on the database map (Fig. 4.1). Line spacing is 12.5 m in both the inline and crossline directions, and the record length is 5.5 seconds two-way travel time (TWTT). The volume has been time- but not depth-migrated. The data have been interpreted using Landmark

SeisWorks™ and GeoProbe™ software. The seismic data are displayed in figures with a reverse polarity (SEG European convention; Brown, 1999) with a downward decrease in acoustic impedance represented by a trough (black) and a downward increase in acoustic impedance represented by a peak (red). The studied interval typically falls between 2.5 and 4 seconds TWTT, and within this interval the vertical resolution of the data is approximately 40 m.

Key lithostratigraphic or chronostratigraphic horizons have been identified within wells and then tied to the seismic data. A combination of qualitative and quantitative (synthetic seismograms) ties were made between seismic and well data. Good correlation between key horizons in wells and easily recognisable seismic markers (e.g. Base Cretaceous Unconformity) allowed for a high degree of confidence. Four horizons have been interpreted regionally across the dataset; (1) Top Zechstein (Top Permian), (2) approximate Top Callovian, (3) Intra Oxfordian and (4) Base Cretaceous Unconformity. At a regional scale, lithostratigraphic nomenclature may not be applied to the seismic events due to the diachroneity of formations throughout the study area (e.g. the Top Hugin reflection event to the north of the study area is equivalent to the Top Sleipner reflection event to the south). Several additional horizons have been interpreted locally and these are discussed in more detail in Chapters 7 and 8. Typically, horizons were picked manually in grids of every 20-50 inlines and crosslines, although in particularly complex areas the grid spacing was decreased to 5. In addition, arbitrary lines were generated perpendicular to key structural features to enable improved interpretation across them. These manual picks formed the seed grid that was then extrapolated using the mapping functions within the software (“3-D Zap” and “Interpolate”).

The “3-D Zap” function searches the seismic volume adjacent to seed points for additional points with identical or very similar seismic attributes (based on given parameters such as amplitude strength) and then creates a new seed point at those locations. Consequently, this method could only be utilised where the reflection event was of sufficient quality and continuity that the horizon could be consistently identified by the software on the basis of the given seismic parameters. For example, the sharp contrast between the underlying Draupne Formation mudstones and the overlying carbonate-rich Cretaceous succession enabled the Base Cretaceous Unconformity (BCU) to be picked using this tool. The “Interpolate” function does not employ seismic attributes and simply interpolates a straight line between interpreted seed points on lines or traces. This function was used where the quality and continuity of a seismic reflection event was not sufficient to enable the “3-D Zap” function to work effectively. For example, the facies variation and associated variation in seismic character at the approximate Top Callovian reflection event was too inconsistent to enable use of the “3-D Zap” function, and therefore “Interpolate” was utilised instead.

4.3 Seismic Analysis of Structure

Isochron maps were generated from time structure maps to demonstrate stratigraphic thickness variations which may be related to syn-depositional tectonics. Smaller-scale, more subtle thickness variations were investigated on individual seismic sections through interpretation of reflector geometry and termination configuration (e.g. onlap, downlap and truncation) after flattening on the syn-rift reflection events

(e.g. Intra Callovian). It should be noted that the effects of compaction have not been incorporated into the investigation of structural evolution and therefore stratigraphic unit thickness may have been underestimated, particularly where thick coal seams are developed. Although decompaction would change the total sediment thickness, gross thickness trends and variations are not expected to change. Integration of well data with isochron mapping allows increased confidence in thickness trends identified from seismic data.

In the Sigrun-Gudrun area (Chapter 7), fault throw analysis was undertaken to establish the fault geometry and growth history during and immediately preceding deposition of the Hugin Formation. Fault throw was measured manually on strike-perpendicular seismic sections every 200 m along the fault. Offset has been calculated at the level of the Top Sleipner reflection event, which broadly coincides with the top of the pre-rift succession. As a result, this surface records the cumulative rift-related deformation and enables accurate reconstruction of the growth of individual fault segments. Throw-distance profiles are presented in the time domain (ms TWTT) and have not been depth converted.

4.4 Application of Sequence Stratigraphic Concepts and Models

Sequence stratigraphy enables analysis of basin-fill strata based on the identification of a hierarchy of stratal units separated by chronostratigraphically significant surfaces (Vail *et al.*, 1977; Galloway, 1989; Van Wagoner *et al.*, 1990). Sequence stratigraphic concepts and models can be applied across a range of spatial and

temporal scales, but have been applied here using well data at a resolution largely below that of the seismic dataset.

Conventional sequence stratigraphic models (e.g. Van Wagoner *et al.*, 1990) emphasise the identification of key stratal surfaces which bound intervals of strata that can be related to part of a relative sea-level curve (systems tracts *sensu* Van Wagoner *et al.*, 1990). These models were developed within comparatively simple tectonic settings (e.g. passive continental margin) and implicitly assume regional or global sea-level change as the main control on stratal stacking patterns. Other factors however may influence stratal stacking patterns, including sediment supply and local tectonics. For example, syn-depositional structural development causes spatial variation in accommodation space, which consequently will have implications for stratal stacking patterns and the expression of key stratal surfaces (e.g. Gawthorpe *et al.*, 1994; Gawthorpe *et al.* 1997; Ravnås and Steel, 1998). The impact of the rift-basin setting of the Hugin Formation and associated strata is discussed in more detail in Chapter 2 (sections 2.1.2, 2.2, and 2.4); here it is sufficient to note that there are significant differences between conventional sequence stratigraphic models and the net-transgressive stratigraphic architecture of the Hugin Formation in the South Viking Graben rift. In the context of conventional sequence stratigraphic models, the Hugin Formation corresponds most closely to the transgressive systems tract of a low-order sequence.

Types of key stratal surface identified within the study dataset are summarised below. Regionally correlatable, low-order maximum flooding surfaces have been proposed on the basis of biostratigraphy (discussed above), integrated with facies

interpretations from core or wireline log data. Maximum flooding surfaces record peak transgression in the basin (e.g. Van Wagoner *et al.*, 1990), and may also be termed the surface of maximum transgression (Helland-Hansen and Gjelberg, 1994). Higher order sequence stratigraphic surfaces have been identified on the basis of facies dislocations that show a departure from Walther's Law (Walther, 1894; Van Wagoner *et al.*, 1990). Candidate high-order sequence boundaries, which record fluvial and/or shallow-marine erosion in response to relative sea-level falls, have only been recognised occasionally within the studied interval. These candidate sequence boundary surfaces cannot be correlated regionally; their limited extent is typical in both net-transgressive sequences (e.g. Cattaneo and Steel, 2003) and sequences deposited within rift basins (e.g. Gawthorpe *et al.*, 1997; Ravnås and Steel, 1998). Transgressive surfaces (or surfaces of maximum regression, *sensu* Helland-Hansen and Gjelberg, 1994) mark the onset of sea-level rise or transgression whilst high-order flooding surfaces record an abrupt increase in water depth. In some sequence stratigraphic models, these surfaces are portrayed with no or minor associated erosion (Van Wagoner *et al.*, 1988). Flooding surfaces associated with erosion are coincident with, or overlie, a transgressive erosion or ravinement surface (e.g. Swift, 1968; Nummedal and Swift, 1987; Reynolds, 1996; Cattaneo and Steel, 2003). Wave ravinement surfaces (*sensu* Swift, 1968) are formed by erosion at wave base during shoreface retreat. Similarly, tidal ravinement surfaces (*sensu* Swift, 1968) are formed by tidal-current erosion and are therefore restricted to tide-influenced or tide-dominated coastal areas. Both wave and tidal ravinement surfaces can form at the base or within a thin transgressive interval that underlies a high-order flooding surface (e.g. Cattaneo and Steel, 2003), similar to the occurrence of the transgressive systems tract beneath a maximum flooding surface in

a low-order sequence. The high-order flooding surfaces bound upward-shallowing successions of genetically related strata termed parasequences (Van Wagoner *et al.*, 1990), which are the smallest stratal units considered in this study.

CHAPTER 5

Sedimentology and sequence stratigraphy of the Hugin Formation, Quadrant 15, Norwegian Sector, South Viking Graben

The Middle Jurassic Hugin Formation has been the target of exploration within Quadrant 15 of the Norwegian South Viking Graben since the 1960s. The formation consists of shallow-marine and marginal-marine sediments deposited during the overall transgression and southward retreat of the “Brent Delta” systems. Sedimentological analysis of cores across the quadrant has identified six facies associations; bay-fill, shoreface, mouth bar, fluvio-tidal channel-fill, coastal plain and offshore open-marine. These facies associations are arranged into a series of parasequences bounded by flooding surfaces, several of which can be correlated regionally using biostratigraphical data. Within this stratigraphic framework, facies association distributions and architectures are complicated, reflecting the spatial and temporal interaction of various physical processes (e.g. waves and tides) with an evolving structural template produced by rift initiation and salt movement. The overall transgression was highly diachronous, becoming younger from north to south. The northern part of the study area (Sigrun-Gudrun field area) is characterised by a series of backstepping, linear, north-south-trending barrier shorelines and sheltered bays. The central part of the study area (Dagny field area) contains stacked, backstepping strandplain shorelines that fringed syn-depositional topographic highs. Local angular unconformities are developed around these highs, implying that they formed above fault-block crests and salt-cored structures. The southern part of the study area (Sleipner area) contains stacked deltaic shorelines

that were modified by both waves and tides. Sandbody distribution and geometry is closely related to both the depositional regime (e.g. proximity to the Sleipner Delta sediment source) and syn-depositional tectonics within the basin, therefore a robust understanding of both is critical to successful exploration of Hugin Formation reservoirs.

5.1 Introduction

The Middle Jurassic Hugin Formation has been the subject of exploration efforts in the South Viking Graben, northern North Sea, since the 1960s. The formation forms part of the Vestland Group, and comprises shallow-marine and marginal-marine deposits that interfinger with the continental deposits of the underlying Sleipner Formation and the offshore marine shales of the overlying Heather Formation (Vollset and Doré, 1984; Partington *et al.*, 1993). Despite the discovery of the Sleipner Vest and Øst Fields in 1974 and 1981, respectively (Milner and Olsen, 1998; Isaksen *et al.*, 2002), exploration has been hampered by the absence of a regional sequence stratigraphic framework supported by depositional models that allow prediction of reservoir distribution and quality within specific stratigraphic intervals. The development of an integrated sedimentological and stratigraphic understanding of the formation is complicated by two factors. Firstly, shallow-marine and marginal-marine deposits in the formation exhibit a mixed regime of physical processes (i.e. waves, tides and riverine) that varies laterally within the South Viking Graben (e.g. Richards, 1991; Folkestad and Satur, 2008), suggesting that a single depositional model is inappropriate for the area as a whole. Secondly, the construction of a stratigraphic framework is complicated by the overall

transgressive nature of the deposits along with syn-depositional rift-related faulting and halokinesis (Kessler *et al.*, 1995; Sneider *et al.*, 1995; Milner and Olsen, 1998). The resulting syn-depositional structural template was complex and dynamic. In similar rift- or salt-basin settings, an evolving structural template controls the local development of accommodation, topography and sediment routing, which in turn influence facies distributions, architectures and character (e.g. Gawthorpe and Leeder, 2000; Shelley and Lawton, 2005). The aims of this paper are to present an updated regional depositional model and sequence stratigraphic framework for the Hugin Formation within Quadrant 15. Additional wells and new biostratigraphical analyses have been used to update previous studies (e.g. Varadi *et al.*, 1998). This study is aiding current exploration within the region, as well as providing generic insights into the relationship between sedimentology, sequence stratigraphy and structural development that may be applicable to other rift basins and areas of halokinesis.

5.2 Previous studies

The Hugin Formation is an overall transgressive unit that was deposited during rift-related subsidence and subsequent flooding of the South Viking Graben during north-to-south retreat of the “Brent Delta” megasequence (e.g. Rattey and Hayward, 1993; Husmo *et al.*, 2003). The Hugin Formation is Late Bajocian to Oxfordian in age, but transgression of the graben was diachronous and the formation becomes younger from north to south (Fig. 5.1) (Mitchener *et al.*, 1992; Rattey and Hayward, 1993; Partington *et al.*, 1993; Richards *et al.*, 1993). Several other names have been used to describe the same lithostratigraphic interval, including the Beryl Formation

Figure 5.1 Jurassic lithostratigraphy of the study area. N.B. Equivalent lithostratigraphy for the UK sector of the graben is included in Figure 9.1.

and “Bruce Group sands” in the Bruce-Beryl embayment of the UK sector (Mitchener *et al.*, 1992; Richards *et al.*, 1993; Husmo *et al.*, 2003). Time-equivalent deep-marine turbidites in the basin axis have been referred to as the Ling Sandstone Member of the Heather Formation (Cockings *et al.*, 1992; Husmo *et al.*, 2003) or as part of the Bruce or Heather Sandstones (Richards, *et al.*, 1993; Spence and Kreutz, 2003). Rift-related faulting and associated localised tectonic subsidence within the South Viking Graben have influenced sediment routing, gross thickness, facies distributions and facies belt orientations within the Hugin Formation (Sneider *et al.*, 1995; Milner and Olsen, 1998; Varadi *et al.*, 1998).

The Hugin Formation was first described from the Sleipner Vest Field to the south of the study area (well 15/9-2; Vollset and Doré, 1984) to comprise a stacked succession of nearshore-to-shallow-marine sandstones and coals with some fluvio-deltaic influence. A wave-dominated barrier shoreline was interpreted from well data distributed more regionally in the South Viking Graben (Harris and Fowler, 1987). Sequence stratigraphic analysis (Varadi *et al.*, 1998) resulted in recognition of regional flooding surfaces and sequence boundaries that divide the succession into a series of cycles, each recording shoreline progradation and subsequent retrogradation. More detailed studies on the Sleipner Vest and Øst fields have defined several facies associations within such cycles; alluvial plain, coastal swamp, barrier-shoreline complex (including shoreface, mouth bar, tidal channel, tidal dune, and tidal flat deposits), lagoon and /or bay, and offshore deposits, with the last corresponding to the shelfal Heather Formation (Milner and Olsen, 1998; Folkestad and Satur, 2008). The best-quality reservoir units occur within shoreface deposits whilst the worst occur within lagoon and /or bay deposits (Harris and Fowler, 1987;

Milner and Olsen, 1998). The strong tidal signatures recognised throughout the Hugin Formation are interpreted to be the result of wave suppression within the narrow (ca. 30km), elongate (ca. 100-120 km) graben during deposition (Richards, 1991; Folkestad and Satur, 2008).

5.3 Tectonic setting

Since the Cambrian, four major tectonic events have influenced the area now defined as the South Viking Graben: (1) Late Ordovician-Early Silurian collision of the Caledonian Orogeny, (2) Carboniferous to Permian rifting and basin formation, (3) Mesozoic rifting and graben formation, and (4) Late Cretaceous to Early Tertiary inversion (Ziegler, 1990; Lyngsle *et al.*, 2006). The South Viking Graben is a NNE-SSW trending half-graben, bounded to the west by a series of ESE-dipping faults and the Palaeozoic high of the Fladen Ground Spur (Thomas and Coward, 1996; Husmo *et al.*, 2003). To the east, the graben shallows onto the Utsira High, a structurally complex area of localised faulting, tilting and erosion (Thomas and Coward, 1996; Husmo *et al.*, 2003). This study focuses on Norwegian Quadrant 15, which is located on the hanging-wall dip slope of the South Viking Graben (Fig. 5.2).

The study area is dissected by a series of intra-graben, NNE-SSW-trending normal faults that predominantly detach on the Zechstein Group salt layer (Thomas and Coward, 1996; Jackson and Larsen, 2008). These faults form a series of antithetic faults to the main graben-bounding fault array that lies to the west of the study area. In the north of the study area, these faults developed in response to gravity-driven sliding of units above the mobile Zechstein Group salt (Jackson and Larsen, 2008).

Figure 5.2 Present-day structural setting of the study area. Abbreviations used: SVG = South Viking Graben, CVG = Central Viking Graben, NVG = North Viking Graben. Modified from Zanella and Coward (2003).

Towards the south, salt-related structures become more prominent, with the occurrence of salt-cored anticlines, diapirs and salt-withdrawal basins. Locally, compressional anticlines and growth folds are associated with reverse reactivation of older extensional faults (e.g. Thomas and Coward, 1996; Knott, 2001; Jackson and Larsen, 2008). These structures are interpreted to have developed during the Late Jurassic (Thomas and Coward, 1996; Jackson and Larsen, 2008) and they therefore postdate Hugin Formation deposition.

5.4 Dataset and methodology

Almost 600 m of Hugin Formation core from 16 regional wells (Fig. 5.3) has been logged for this study. Wireline-logs were available for all of the cored wells and new, high resolution biostratigraphical analyses were available for 13. Sedimentological analysis involved recording grain size, sorting and sphericity of matrix and clasts, colour variations, sedimentary structures, bed thickness, the nature of bounding contacts, and macro-fauna type and preservation state. The degree of bioturbation is recorded using the bioturbation index (BI) scheme of Taylor and Goldring (1993), which is based on a visual assessment of the amount of 'bioturbated' versus 'unbioturbated' texture and ranges from 0 (no bioturbation) to 6 (complete bioturbation). Wireline-log character and cross-plot analyses (Sonic vs. Gamma Ray; Density vs. Neutron porosity) were employed in an attempt to identify distinctive wireline-log responses for each facies identified in core. These were used to extrapolate to uncored intervals within the studied wells. Not all of the recognised facies displayed distinctive wireline-log characteristics preventing their identification away from core control, in these cases any additional available data

Figure 5.3 Map showing the distribution of cored wells, correlation panels and seismic line used in this study (Figs. 5.4, 5.7-5.9).

was consulted (e.g. sedimentological description of side wall cuttings) and context of surrounding, more readily identified facies was considered.

5.5 Sedimentology of the Hugin Formation

Fifteen facies have been recognised within the Hugin Formation (Table 5.1). These facies have been grouped into six facies associations, which are discussed below and summarised in Figures 5.4-5.6.

5.5.1 Bay-fill facies association

Complete, well-developed successions of the bay-fill facies association are developed in the northern part of the study area. Incomplete successions of the association, comprising individual facies, are also recognised interbedded with fluvio-tidal channel and shoreface facies associations (described below) in the south of the study area.

Description. Units of this facies association are typically composed of a coarsening-upwards succession that comprises, from base to top, facies BF1, BF2, BF3 and BF4 (Table 5.1, Figs. 5.4a, 5.5a). Parallel-laminated mudstones and siltstones at the base of the succession (facies BF1) locally contain beds of densely packed, disarticulated bivalve shells. This facies passes gradationally upward into interbedded heteroliths (facies BF2) that exhibit quasi-rhythmical bedding. Very fine-grained sandstones in these heteroliths occur as non-parallel or wavy laminae and lenticular, trough and hummocky cross-stratified beds. Sandstone beds thicken and coarsen upwards in the succession, until mudstones occur only as thin laminae and drapes (facies BF3).

Facies	Description	Bioturbation	Interpretation
Laminated shale (BF1)	Parallel-laminated mudstone with rare siltstone laminae, locally sideritic. Monospecific assemblage of disarticulated bivalves oriented parallel to bedding and concentrated into thick (60-205 cm) bioclastic-rich beds. Rare gastropods. Varies 30 to 425 cm thick.	BI = 0-2 (Pl)	Restricted marine shale
Interbedded heteroliths (BF2)	Interbedded claystone and very fine-grained sandstone. Non-parallel and wavy lamination, trough cross-stratification and hummocky cross-stratification in sandstone beds. Rare disarticulated bivalve shells oriented parallel to bedding. Varies 10 to 725 cm thick.	BI = 2-3 (Te, Sk, Pl, Ar, Pa, Op)	Muddy shallow bay-fill
Mud-draped sandstone (BF3)	Fine- to medium-grained sandstone. Moderately to well sorted. Wavy bedding, high-angle cross-bedding and occasional wave ripples. Calcite cement and soft sediment deformation locally observed. Disarticulated bivalves throughout, sometimes concentrated in thin beds (<10 cm). Varies 60 to 370 cm thick.	BI = 0-4 (Pa, Dip, Ro/Cy?, Pl, Op)	Sandy shallow bay-fill
Sharp-based, bioclastic sandstone (BF4)	Coarse- to fine-grained sandstone with fining- and coarsening-upward motifs. Moderately to well sorted with sub-rounded grains. Sparse sedimentary structures include ripples and mudstone drapes. Disarticulated bivalves and shell fragments common, concentrated into thin beds (<40 cm). Locally contains calcite-cemented nodules. Sharp, possibly erosive contacts at individual bed bases. Varies 250 to 500 cm thick.	BI = 1-3 (Op, ET)	Marine channel-fill sand
Bioturbated silt- to fine sandstone (SF1)	Upward-coarsening, siltstone to very fine-grained, moderately sorted sandstone. Shell debris, quartz granules and angular carbonaceous fragments throughout. Primary sedimentary structures are poorly preserved, but contains rare thin (30-75 cm), sharp-based, very fine- to fine-grained cross-laminated sandstone beds. Varies 30 to 480 cm thick.	BI = 4-5 (Pa, Ch, Op, Pl, Ter, As?)	Distal lower shoreface
<i>Ophiomorpha</i> -rich sandstone (SF2)	Very fine- to fine-grained, moderately to well sorted sandstone. Fine-grained, carbonaceous laminae throughout. Dominantly low-angle cross bedding, with subordinate parallel lamination. Carbonate-cemented nodules. Varies 130 to 440 cm thick.	BI = 3-4 (Op)	Proximal lower shoreface
Upward coarsening, cross-laminated sandstone (SF3)	Fine- to medium-grained, well sorted, coarsening-upwards, cross-laminated sandstone. Discrete laminae of coarser sediment, carbonaceous fragments and shell fragments occur throughout. Varies 425 to 800 cm thick.	BI = 0* (crypto/Mac?)	Upper shoreface
Bioturbated sandstone with granular beds (MB1)	Highly bioturbated siltstone to very fine-grained sandstone with stringers and thin (5-20 cm) beds of rounded granules. Varies 175 to 1010 cm thick.	BI = 4-5 (Pa, Ch, Op, P)	Distal mouth bar
Sandstone with granular beds (MB2)	Very fine- to fine-grained sandstone with low-angle cross-bedding. Stringers and thin (5-15 cm) beds of rounded granules occur throughout. Monospecific trace fossil assemblage (<i>Ophiomorpha</i>). Varies 290 to 775 cm thick.	BI = 0-2 (Op)	Proximal mouth bar
Upward-fining sandstone with granular beds (FTC1)	Fining-upward very coarse- to fine-grained sandstone with possible bidirectional cross-sets and swaley cross-stratification. Stacked fining-upward beds (15 to 170 cm thick) with sparse coarse-grained stringers. Sharp, possibly erosive contacts at individual bed bases. Varies 740 to 820 cm thick.	BI = 0-1 (Sk, ET)	Tidally-influenced fluvial channel-fill sandstone
Cross-stratified sandstone (FTC2)	Fining-upward coarse- to medium-grained sandstone with unidirectional cross-bedding, carbonaceous laminae and fragments throughout. Sharp, possibly erosive contacts at individual bed bases. 580 cm thick.	BI = 0-1 (ET)	Fluvial channel-fill sandstone
Carbonaceous shales and coal (CP1)	Coal and carbonaceous shale. Basal contact typically parallel although may be undulose. Diagenetic pyrite nodules are recognised. Varies 30 to 520 cm thick.	BI = 0	Coal
Rooted interbedded heteroliths (CP2)	Interbedded dark grey mudstone and siltstone to very fine-grained sandstone. Wavy to non-parallel lamination penetrated and disrupted by roots. 60 cm thick.	BI = 3-4 (roots)	Coastal plain fines
Bioturbated shale (OM)	Mudstone with rare siltstone beds. Parallel lamination, soft sediment deformation developed locally at bed tops. Rare thin (<100 cm), finely laminated, unbioturbated beds of very fine-grained sandstone.	BI = 3-5, occ. 0 (Pl, Op, Ch, Zo, Tae, As, Te)	Open marine shale

Table 5.1 Summary of facies in the Hugin Formation. Bioturbation fabric is described using the Bioturbation Index (BI) scheme of Taylor & Goldring (1993). Individual trace fossils, listed in brackets, are abbreviated as follows: Te = *Teichichnus*, Sk = *Skolithos*, Ar = *Arenicolites*, Pa = *Paleophycus*, Op = *Ophiomorpha*, Dip = *Diplocraterion*, Ros = *Rosselia*, Cy = *Cylindrichnus*, Pl = *Planolites*, Ch = *Chondrites*, Ter = *Terebellina* (Schaub. *Fria*), As = *Asterosoma*, Mac = *Macaronichnus*, ET = Escape traces, crypto = cryptobioturbation.

Figure 5.4 Core descriptions and Gamma Ray-log characteristics of successions illustrating the principal facies associations in the Hugin Formation: (a) bay-fill, (b) shoreface, (c) mouth bar, (d) fluvio-tidal channel-fill, and (e) coastal plain.

Figure 5.5 Core photographs illustrating the selected facies in the principal facies associations of the Hugin Formation: (a) bay-fill, (b) shoreface, (c) mouth bar, (d) fluvio-tidal channel-fill, and (e) coastal plain. Photographs are located in the successions illustrated in Figure 4 where possible. Photos from: (a) BF1 - well K, 3838.70cm; BF2 – well K, 3836.80 m; BF3 – well K, 3836.00 m; BF4 – well K, 3826.35 m; (b) SF1 – well D, 3599.70 m; SF2 – well D, 3584.25 m; SF3 – well D, 3581.10 m; (c) MB1 – well K, 3615.30 m; MB2 – 3608.30 m; (d) FTC1 – well A, 3698.6 m; FTC2 – well A, 3661.25 m; (e) CP2 – well A, 3687 m. Depths given are core depths and have not been shifted to wireline-log depths.

Figure 5.6 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 principal facies associations of the Hugin Formation: (a) bay-fill (well H), (b) shoreface (well E), (c) mouth bar (well D), and (d) fluvio-tidal channel-fill and coastal plain (well A). The cross-plots are compiled with data from logged core intervals within the named wells.

Bioturbation is variable in intensity, and wavy bedding, high-angle cross-bedding and occasional wave ripples are preserved where bioturbation intensity is low. Locally this facies is truncated by sharp-based successions of coarse-grained sandstones with rare wispy lamination, ripples and sparse bioturbation (facies BF4). Although shells and shell fragments are common throughout these sharp-based successions, they are typically concentrated towards the base. Successions of this facies association are commonly capped by thin deposits of the coastal plain facies association (described below).

Interpretation. The coarsening-upwards successions that typify this facies association imply a shallowing-upward or progradational trend, with increasing proximity to the source of sand towards the top of the succession. Parallel-laminated mudstones record deposition from suspension, while sandstone laminae and beds contain structures that record episodic deposition from suspension or tractional currents. Quasi-rhythmical interbedding of mudstone and sandstone (facies BF2 and BF3) suggest regular and repeated fluctuations in energy regime. This may reflect fluctuations in fluvial discharge as a consequence of climatic or seasonal variations or tidal action, as a consequence of the diurnal cyclicity of currents (e.g. Nio and Yang, 1991). Analysis of these mud drapes to recognise distinct bundles that can be attributed to tidal cyclicity has not been undertaken as part of this study. Wave and storm influence is also present, indicated by rare wave ripples and hummocky cross stratification (e.g. Dott and Bougeois, 1982; Duke *et al.*, 1991). Sharp-based sandstones that occur locally in the upper part of the successions are interpreted as channel fills (facies BF4). The occurrence of shell fragments, similar to those preserved within facies BF1, in lags and dispersed through these sandstones,

together with rare mudstone drapes, have been recorded as common features in tidal inlet channel-fill deposits (e.g. Moslow and Tye, 1985; Israel *et al.*, 1987; Simms *et al.*, 2006). The occurrence of a monospecific shell assemblage and low-to-moderate intensity bioturbation by a restricted trace fossil assemblage indicates deposition in a restricted marine environment (Pemberton *et al.*, 1992; MacEachern and Bann, 2008). This interpretation is supported by microfossil assemblages (Tjemsland, 2007). The coarsening-upwards trend of this facies association (i.e. facies BF1 to BF3), combined with the restricted marine setting may represent deposition within a variety of settings: (1) a bay-head delta prograding into the depositional setting from the landward margin; (2) a flood tidal delta developed at the seaward margin of a barrier-fronted bay cut by tidal-inlet channels (e.g. Boothroyd, 1985); (3) a washover fan deposited during storm events when water surged over the top of a barrier (Schwartz, 1982). It is difficult to distinguish which interpretation should be applied without further data, for example palaeocurrent data which might enable distinction of seaward-directed versus landward-directed palaeocurrents. The common occurrence of coastal plain coals capping successions of this facies association may reflect abandonment of a mature flood tidal delta (e.g. Reinson, 1992) or continued progradation of the shoreline and coastal plain. The significant thickness of this coal (>5 m in wells J and H) however may favour the latter interpretation. Successions of this facies association are therefore interpreted to record periodic infilling of restricted marine embayments and/or lagoons that were locally contiguous with tidal inlet channels. The restricted marine character of these successions may be explained by deposition in sheltered lagoons and/or embayments that either lay behind coeval barriers and spits fronted by wave-dominated shorefaces (Harris and Fowler, 1987; Hampson *et al.*, 2009) or were protected by rift-generated islands and

shoals. The pervasive tidal signature in this facies association may simply reflect the absence of wave reworking in the sheltered lagoons and/or embayments, although the narrow, elongate shape of the South Viking Graben may have locally amplified tidal currents and further suppressed wave activity (Husmo *et al.*, 2003; Folkestad and Satur, 2008).

Sequence stratigraphy. At the base of successions of this facies association, lagoonal and/or bay-fill mudstones (facies BF1) representing relatively deep water (4-45 m, based on parasequence thickness) are typically juxtaposed directly above coastal plain coals. The bases of these successions may therefore represent flooding of the basin and may be interpreted as transgressive surfaces (*sensu* Cattaneo and Steel, 2003). Alternatively the observed deepening may reflect autogenic migration of the sediment feeder system and redirection of the sediment being transported into the basin to another location. An increase in bioturbation intensity is typically observed at these surfaces, combined with the occurrence of *Diplocraterion*, an indicator of more marine conditions (Taylor and Goldring, 1993). Some of these surfaces are also associated with a lag deposit of fine- to medium-grained sandstone. Where transgressive surfaces are underlain by tidal inlet channel-fill deposits (facies BF4), the basal erosion surfaces of the channels may record tidal erosion during transgression (i.e. tidal ravinement surfaces *sensu* Swift, 1968). This interpretation is supported by the occurrence of winnowed, coarse-grained and/or bioclastic lags at the bases of some channels. In one instance, a lagoonal and/or bay-fill succession is erosively overlain by upper shoreface sandstones (facies SF3, described below); the base of this sandstone is interpreted to record wave erosion during transgression (i.e. a wave ravinement surface *sensu* Swift, 1968). Each of the bay-fill successions

themselves is an upward-shallowing unit bound by transgressive surfaces (*sensu* Cattaneo and Steel) and is interpreted to represent the transgressive portion of a regressive-transgressive cycle which closely corresponds to a parasequence (*sensu* Van Wagoner *et al.*, 1990; see Chapter 7 for detailed discussion). Individual parasequence thickness varies from 4 m to 45 m, and the duration of each parasequence is approximately 500-600 ka, based on high-resolution biostratigraphy (Hampson *et al.*, 2009). The thickest parasequences occur in the western part of the study area, where greater rotation of the hanging-wall dip slope created increased accommodation space towards the graben-bounding fault.

Wireline-log character. Gamma Ray (GR) logs through successions of this facies association show an upward-decreasing trend as they become more sandstone-rich towards their tops (Fig. 5.4a). Facies BF1 and BF2 exhibit high GR values and a serrate pattern, reflecting their heterolithic character, whereas sandstone-dominated facies BF3 has a smoother, funnel-shaped GR profile. Facies BF4 has the lowest GR values and a sharp-based blocky log character, consistent with its interpretation as channel-fill deposits. The four constituent facies also have locally distinctive signatures when wireline-log cross plots are generated (Fig. 5.6a). Facies BF1 exhibits the highest GR and sonic (DT) values (41 - 308 API; 225 - 355 μ s/ft) whilst facies BF4 exhibits the lowest (7 - 35 API; 175 - 230 μ s/ft). Facies BF2 and BF3 occupy similar positions on the graph, whilst their heterolithic character means that they plot between facies BF1 and facies BF4. Sandstone-rich facies BF3 exhibits overall lower GR and higher DT values (18 - 46 API; 199 - 314 μ s/ft) whilst mudstone-rich facies BF2 exhibits higher GR and lower DT values (17 - 65 API; 189 - 268 μ s/ft). Cross plotting neutron porosity (NPHI) and density (RHOB) values

indicates that facies BF1, BF2 and BF4 have similar RHOB values (2280 - 2820 g/cm³) but show variation in NPHI values with facies BF4 exhibiting the lowest (0.038 – 0.146 v/v) and facies BF1 the highest (0.167 - 0.464 v/v). Facies BF3 shows a wide range of NPHI values (0.077 – 0.518 v/v) and the broadest range of RHOB values (1690 – 2745 g/cm³).

5.5.2 Shoreface facies association

Complete, well-developed successions of the shoreface facies association are developed in the central (Dagny field) and southern (Sleipner Vest field) parts of the study area whilst incomplete succession are seen to the north (Sigrun-Gudrun fields). Several stacked shoreface successions are recognised within the central portion of the study area, whilst to the south they interfinger in vertical successions with facies of the bay-fill, coastal plain, mouth bar and fluvio-tidal channel associations.

Description. Units of this facies association are typically composed of upward-coarsening and cleaning successions. Complete successions comprise, from base to top, facies SF1, SF2 and SF3; incomplete successions of SF1 to SF2 facies and SF2 to SF3 facies are more typical of the central part of the study area (Figs. 5.4b, 5.5b). Highly bioturbated siltstone to very fine-grained sandstone at the base of a succession (facies SF1) passes gradationally upwards into moderately to well sorted, very fine- to fine-grained sandstones (facies SF2) and then well-sorted, fine- to medium-grained sandstones (facies SF3). Sharp-based beds of very fine- to fine-grained, unbioturbated, low-angle cross-laminated sandstone (30-75cm thick) occur within facies SF1 and SF2, and cross-lamination is pervasive in facies SF3. Bioturbation decreases in intensity and diversity from the base to top of the

successions, allowing increasing preservation of cross-lamination. A monospecific assemblage of *Ophiomorpha* characterises facies SF2, whereas the ‘fuzzy’ appearance of some laminae in facies SF3 may indicate cryptobioturbation or *Macaronichnus* (Pemberton *et al.*, 2008).

Interpretation. The coarsening-upward trend in grain size and upward decrease in bioturbation intensity within the successions implies an upward increase in energy due to shallowing. The high intensity and diversity of bioturbation within fine-grained sediments is typical of the Cruziana Ichnofacies indicating a low-energy and low-stress marine environment of deposition (MacEachern and Pemberton, 1992; MacEachern and Bann, 2008) whilst monospecific bioturbation by resilient *Ophiomorpha* burrows is typical of the Skolithos ichnofacies and is indicative of a loose substrate or higher energy setting such as middle or upper shoreface (MacEachern and Pemberton, 1992; Martin, 1993). Sharp-based, sparsely bioturbated sandstone beds within facies SF1 and SF2 are interpreted to record storm events (e.g. Pemberton *et al.*, 1992). Well-sorted cross-laminated sandstones with low levels of bioturbation or cryptobioturbation (SF3) are typical of upper shoreface-to-foreshore settings (Pemberton *et al.*, 2008). This facies association is interpreted to have been deposited within a prograding shoreface with variable storm influence, as part of a strandplain, barrier island or spit shoreline.

Sequence stratigraphy. At the base of successions of this facies association, distal or proximal lower shoreface deposits (facies SF1 and SF2) representing relatively deep marine deposition below fair-weather wave base are juxtaposed directly above coastal plain coals or, less typically, fluvio-tidal channel sandstones. The bases of

these successions are therefore interpreted as flooding surfaces (*sensu* Van Wagoner *et al.*, 1988). Some such flooding surfaces occur coincident with coarse-grained, locally carbonaceous or bioclastic lag deposits 5 to 15 cm thick. The stratigraphic context of these lags records transgression, and they are thus interpreted as wave ravinement surfaces (*sensu* Swift, 1968), which suggests erosion of the uppermost part of the underlying succession during shoreface retreat. A few of these flooding surfaces are also associated with a cemented interval that may represent calcite re-precipitation of bioclastic material (e.g. within a transgressive lag). In the central part of the study area, shoreface parasequences are rarely complete, comprising successions of facies SF1 to SF2 only. Upper shoreface deposits (facies SF3) may have been removed by transgressive erosion during shoreline retreat. Alternatively, the deep *Ophiomorpha* that are characteristic of SF2 may have intensely bioturbated remnant upper shoreface sandstones after flooding, giving the appearance of facies SF2. Parasequence thickness varies from 10 m to 23 m, and the duration of each parasequence is approximately 500 ka to over 1 Ma based on high resolution biostratigraphy (Tjemsland, 2007). To the south of the study area (Alpha Central), shoreface parasequences comprise complete facies successions (SF1 to SF3) that pass upwards into coal and coastal plain deposits, recording continued progradation of the shoreline without a stratigraphic discontinuity.

Wireline-log character. The GR log character of successions of this facies association shows an upward-decreasing trend as they become more sandstone-rich towards their tops (Fig. 5.4b). The gradational contacts between facies can make recognition of distinct facies boundaries in wireline-log data difficult, although typically facies SF1 is represented by a serrate, sometimes funnel-shaped GR curve

whilst facies SF2 and SF3 have lower GR values and are more blocky in character. Cross-plotting GR against DT and NPHI against RHOB (Fig. 5.6b) shows extensive overlap between facies SF1 and SF2 which makes distinguishing these facies difficult away from core control. Both exhibit low GR with moderate DT values (25 - 46 API; 175 - 275 $\mu\text{s}/\text{ft}$) and low NPHI with moderate-to-high RHOB values (0.037 - 0.190 v/v; 2690 - 2314 g/cm^3). Facies SF3 plots as a distinctly separate group on both graphs, with moderate DT values combined with the lowest observed GR values (250 - 275 $\mu\text{s}/\text{ft}$; 15 - 25 API) and low NPHI values combined with the lowest observed RHOB values (0.094 - 0.113 v/v; 2271 - 2342 g/cm^3).

5.5.3 Mouth bar facies association

This facies association is developed in the central part (Dagny field) of the study area, where it interfingers with shoreface deposits, and to the south of the study area. Deposits of this association are similar to the mouth bar facies association described in the Sleipner Vest Field by Folkestad and Satur (2008).

Description. Units of this facies association comprise either a complete upward-coarsening succession of MB1 and overlying MB2 facies (Fig. 5.4c), or individual beds of MB1 and/or MB2 facies (Fig. 5.5c) that interfinger with shoreface deposits. The lower part of complete successions consists of moderately to highly bioturbated siltstones and very fine-grained sandstones that coarsen upward to medium- through very coarse-grained sandstones (facies MB1). Cross bedding, including coarse-grained stringers, and mudstone laminae are preserved at bed tops and as drapes along cross-bed foresets where bioturbation intensity is lowest. Complete successions are more sand-rich in their upper parts, which corresponds with a

decrease in bioturbation intensity and the loss of very coarse-grained sandstone stringers (facies MB2). Successions of this facies association are commonly capped by a coal, although some are truncated by overlying shoreface deposits.

Interpretation. The upward-coarsening grain-size trend of successions of this facies association implies a progradational trend, with increasing proximity to the source of coarse sand and granules towards the top. The intensity and diversity of bioturbation suggests a marine depositional setting, whilst poorly sorted, coarse-grained sandstones and stringers are indicative of fluvial input (Fielding *et al.*, 2005) and mudstone drapes and laminae may imply tidal action (Nio and Yang, 1991). Preservation of indicators of fluvial sediment influx suggests deposition in mouth bars and signifies low wave reworking of fluvially supplied sediment, which may support wave suppression within the narrow graben (Folkestad and Satur, 2008). This facies association is developed geographically on the fringes of the Sleipner Delta system to the south (Folkestad and Satur, 2008) and is interpreted to be fed from it.

Sequence stratigraphy. Distal mouth bar deposits (facies MB1) at the base of the succession are seen juxtaposed directly above both coastal plain coals and upper shoreface sandstones, implying a flooding surface (*sensu* Van Wagoner *et al.*, 1988) at their base. A sharp increase in bioturbation, locally coupled with coarse-grained sandstone, carbonaceous and bioclastic beds marks this surface. Where this facies association occurs interfingered with units of the shoreface facies association, autogenic lateral migration of the fluvial system is inferred rather than a relative sea-level change.

Wireline-log character. The GR log character of this facies association is variable with upward-decreasing profiles and, less commonly, upward-increasing, blocky and funnel-shaped profiles recognised. Wireline-log cross-plots (Fig. 5.6c) show significant overlap between facies MB1 and MB2 although facies MB2 exhibits a broader range of values. Facies MB2 is characterised by highly variable DT (167 - 397 $\mu\text{s}/\text{ft}$) with low to moderate GR (12 - 58 API) and a broad range of both NPHI (0.051 - 0.595 v/v) and RHOB values (1612 - 2705 g/cm^3). In contrast, facies MB1 shows little variation in DT (230 - 260 $\mu\text{s}/\text{ft}$) although may exhibit higher GR values (20 - 80 API) along with significantly lower variation in NPHI and RHOB values (0.084 - 0.169 v/v; 2327 - 2521 g/cm^3). The wireline-log character of the mouth bar facies association (particularly facies MB1) is very similar to that of the shoreface facies association, making it difficult to differentiate between the two facies associations away from core control. Within the depositional setting, mouth bar and shoreface deposits adopt similar positions within a prograding sequence between offshore marine units and coastal plain units, with similar implications for sequence stratigraphic development. Therefore, uncored intervals that exhibit this wireline signature have been identified as shoreface deposits but it is recognised that they cannot be unequivocally distinguished from mouth bars.

5.5.4 Fluvio-tidal channel-fill facies association

Identified in the southern part of the study area, this facies association occurs closely associated with mouth bar and coastal plain deposits. Facies FTC1 is broadly equivalent to the tidal channel facies association described from the Sleipner Vest Field (Folkestad and Satur, 2008).

Description. Facies FTC1 and FTC2 both comprise sharp-based, upward-fining successions of coarse- to medium-grained, cross-bedded sandstones (Figs. 5.4d, 5.5d). In facies FTC1, the sandstones contain bi-directional cross-bed sets and/or swaley cross-stratification, and asymmetrical ripple forms preserved by mudstone drapes. Bioturbation intensity is low to zero, with rare escape traces present within finer-grained intervals. The sandstones in facies FTC2 contain unidirectional cross-bed sets which lack bioturbation and mud drapes, although carbonaceous fragments and laminae are common. Some successions of FTC2 exhibit a weak upward-coarsening trend in grain size in their lower parts. Units of facies FTC1 pass vertically and laterally into units of facies FTC2, but are also overlain by bay-fill, shoreface, mouth bar and offshore deposits. Units of facies FTC2 are commonly overlain by coals and coastal plain deposits.

Interpretation. The sharp, possibly erosive, bases and upward-fining character of successions of this facies association indicate deposition within channels under conditions of gradually decreasing energy, which may record channel migration or abandonment. The possibly bi-directional cross-bed sets within facies FTC1 may be indicative of tidal influence although they may simply represent oblique sections through swaley cross stratified strata. The mud drapes within this facies may also be indicative of tides (e.g. Nio and Yang, 1991), alternatively they may reflect variations in the fluvial flow regime. The absence of such features in facies FTC2 may indicate less tidal and marine influence. This coupled with the overall coarser grain size within facies FTC2 may be interpreted to imply that facies FTC2 is the up-dip equivalent of facies FTC1. Alternatively, the lack of tidal influence within facies

FTC2 could represent deposition in rivers that lacked tidal amplification near their mouths and were associated with linear-to-lobate, deltaic shorelines; facies FTC1 would then represent deposition within funnel-shaped estuaries that locally amplified tidal currents. The overall depositional setting of this facies association is interpreted to be variably marine-influenced fluvial channel fills.

Sequence stratigraphy. The basal surfaces of successions of this facies association may not necessarily have sequence stratigraphic importance, but instead record autogenic channel migration and avulsion. However, where a unit of marine-influenced facies FTC1 directly overlies coastal plain deposits, a tidal ravinement surface is interpreted at its base, as the precursor or up-dip correlative to a flooding surface. Where such units overlie bay-fill deposits, they are difficult to distinguish from tidal inlets that formed as part of a barrier island and associated shoreface (facies BF4, Table 5.1); inlet channel-fills pinch out into bay-fill deposits, whereas fluvio-tidal channel-fills extend up depositional dip into fluvial channels in the coastal plain. Sequence stratigraphic interpretation and correlation of channel-fill sandstones that overlie bay-fill deposits is therefore uncertain, particularly in the absence of core.

Wireline-log character. This facies association is generally represented by low GR values and a blocky GR profile (Fig. 5.4d), although rare upward-decreasing and upward-increasing GR profiles are also noted. Wireline cross-plots (Fig. 5.6d) show facies FTC1 to be characterised by moderate DT values (235 - 335 $\mu\text{s}/\text{ft}$) and highly variable GR values (8 - 134 API) whilst facies FTC2 is characterised by similar DT values (241 - 297 $\mu\text{s}/\text{ft}$) combined with consistently low GR values (9 - 34 API).

Facies FTC1 also exhibits moderate RHOB values (2284 - 2570 g/cm³) and highly variable, moderate to high NPHI values (0.140 - 0.469 v/v). Facies FTC2 may be characterised by slightly lower RHOB values (2168 - 2447 g/cm³) combined with low to moderate NPHI values (0.140 - 0.289 v/v).

5.5.5 Coastal plain facies association

Thin accumulations of this facies association occur throughout the study area, although the thickest and best developed examples are closely associated with fluvio-tidal channel-fill deposits to the south.

Description. This facies association predominantly comprises coal (facies CP1) with associated heterolithic deposits of interbedded mudstone, siltstone and very fine-grained sandstone (facies CP2) (Figs. 5.4e, 5.5e). Preserved root traces are common throughout the heteroliths, disrupting lamination below coal beds. Wood fragments occur within coal beds, and angular carbonaceous fragments from the coal beds are reworked into overlying sediments. Units of this facies association are locally eroded by fluvio-tidal channel-fill deposits and pass laterally into bay-fill and shoreface deposits. The units are abruptly overlain by offshore, shoreface, mouth bar and bay-fill deposits. Coal beds and thin successions of facies CP2 cap upward-coarsening shoreface, mouth bar and bay-fill successions.

Interpretation. Coal deposition can occur in a wide range of depositional environments, however the context of coals (facies CP1) in the Hugin Formation suggests a coastal plain environment. Roots beneath the coal beds indicate that they developed *in situ*. The interbedding of sandstones and siltstones with mudstones in

facies CP2 suggests a regularly fluctuating energy regime, most likely related to variations in discharge and sediment load in adjacent fluvio-tidal channels. The association of facies CP2 with coals (facies CP1), bay-fill, shoreface, mouth-bar and fluvio-tidal channel-fill deposits supports a coastal plain environment of deposition.

Sequence stratigraphy. Thin intervals of the coastal plain facies association that cap upward-coarsening shoreface, mouth bar and bay-fill successions are interpreted to record subaerial deposition in response to shoreline progradation, and they thus form the upper part of shallow-marine and marginal-marine parasequences. Where these coastal plain intervals are abruptly overlain by offshore, shoreface, mouth bar and bay-fill deposits, their upper surfaces are interpreted as flooding surfaces that record increases in water depth. Coal fragments reworked from the coastal plain intervals into overlying deposits imply erosion during transgression.

Wireline-log character. This facies association is typically represented by a low GR, blocky to serrate log curve (Fig. 5.4e) although funnel shaped may also be observed. This facies association is characterised by high DT values combined with low GR values (Fig. 5.6d). Facies CP1 records the broader range of DT values (313 - 427 $\mu\text{s}/\text{ft}$) and lower GR values (8 - 56 API) compared to facies CP2 (DT of 358 - 415 $\mu\text{s}/\text{ft}$; GR of 32 - 60 API). Facies CP1 is also characterised by consistently low RHOB values compared to the other studied facies (1930 - 2225 g/cm^3) and moderate to high NPHI values (0.2405 - 0.529 v/v). Similarly, facies CP2 has consistently lower RHOB values (1994 - 2148 g/cm^3) and high NPHI values (0.4225 - 0.4993 v/v) than other facies.

5.5.6 Offshore open-marine facies association

This facies association is equivalent to the shales of the Heather Formation, and interfingers with shoreface, mouth bar and bay-fill deposits in the southern part of the study area.

Description. This facies association comprises highly bioturbated mudstones and siltstones that constitutes a single facies (OM). Bioturbation ranges from moderate to high in intensity, and is highly diverse. Unbioturbated, coarse-grained siltstone to very fine-grained sandstone beds occur locally, and contain parallel lamination and soft sediment deformation. Units of facies OM are gradationally overlain by upward-coarsening shoreface and mouth bar deposits, and they pass laterally into bay-fill, shoreface and mouth bar deposits. These units abruptly overlie shoreface, bay-fill, fluvio-tidal channel-fill and coastal plain deposits.

Interpretation. The fine-grained character of these deposits, combined with their parallel laminated character where not bioturbated, suggests deposition by fall-out from suspension. The high degree of bioturbation by a fully marine trace fossil suite implies deposition in an open-marine setting, below fair weather wave base. Sharp-based, coarser grained beds that lack bioturbation are interpreted to have been deposited during major storm events. Soft-sediment deformation within and below these beds is attributed to loading and water escape.

Sequence stratigraphy. This facies association represents the deepest fully marine setting observed within the studied interval. Where units of the facies association are juxtaposed directly above shoreface, bay-fill, fluvio-tidal channel-fill or coastal plain

deposits, their bases represent abrupt deepening across a flooding surface (*sensu* Van Wagoner *et al.*, 1990). The gradational upper contacts of the units represent gradual shallowing into shoreface and mouth bar deposits.

Wireline-log character. Units of this facies association have consistently high GR values, with bed-scale variations in grain size represented by a slightly serrate GR-log character.

5.6 Sequence stratigraphic framework

The facies associations described above and their bounding sequence stratigraphic surfaces have been correlated to construct a regional sequence stratigraphic framework for the Hugin Formation within the study area. Correlation between the wells has been constrained by a newly revised, high-resolution biostratigraphical and chronostratigraphical scheme that is based on the StatoilHydro ASA biostratigraphical framework for Jurassic strata in the North Sea (Tjemsland, 2007). Flooding surfaces were proposed based on the biostratigraphic data where marine faunal assemblages of either high diversity and/or abundance were identified. As part of this study, these observations were integrated with core and wireline-log characteristics to iteratively refine the position of the surface (i.e. where they corresponded to thick offshore mudstones overlying shallow marine deposits). Although fine biostratigraphical resolution (ca. 500 ka) was achieved within the marine Hugin and Heather formations, the long-ranging character of pollen and spores within the Sleipner Formation preclude a similarly detailed zonation of this unit (Tjemsland, 2007). As a result, correlation of sequence stratigraphic surfaces

between the Hugin and Sleipner formations is difficult. Regionally extensive maximum flooding surfaces have been identified where marine microfossil assemblages with high diversity and abundance occur associated with distinctive GR profiles. Where possible, these regional maximum flooding surfaces have been tied into the widely used nomenclature of Partington *et al.* (1993). One additional regional maximum flooding surface has been identified, referred to below as MFS A (Tjemsland, 2007); this surface is slightly older than MFS J42 (Partington *et al.* 1993). Higher order flooding and transgressive surfaces of sub-regional and local extent have been identified by facies interpretation of core and wireline-log data, and their correlation constrained by distinct palynological events and zones. The sequence stratigraphic framework and associated facies distributions are illustrated below using three correlations through the study area (Figs. 5.3, 5.7-9) and two gross palaeogeographical reconstructions that represent early Hugin deposition (MFS A) and mid to late Hugin deposition (MFS J44) (Fig. 5.10).

5.6.1 Regional correlation along basin axis

The studied strata are subdivided into five intervals bounded by regional maximum flooding surfaces that can be correlated along the axis of the South Viking Graben (Figs. 5.3, 5.7). Within the first, lowermost interval (below MFS J36), shallow marine deposition (Hugin Formation) was confined to the northern part of the study area within restricted marine bays and/or lagoons which pass to the south into coastal plain deposits (Sleipner Formation). In the second interval (between MFS J36 and MFS A), bay-fill deposition persisted in the north and passed towards the central part of the study area into coastal plain deposits (Sleipner Formation) and shoreface deposits (Hugin Formation) whilst being overlain locally by shoreface

Figure 5.7 Regional correlation. (a) Regional N-S-oriented correlation panel and (b) corresponding seismic section along the axis of the South Viking Graben (see Fig. 5.3 for location). In general, the Hugin Formation becomes younger towards the south.

deposits (Hugin Formation) and offshore mudstones (Heather Formation). These offshore mudstones are penetrated within the most basinward well (well F, Fig. 5.7) and imply that locally irregular shoreline trends were developed as a result of early flooding of the graben. Further south, shoreface deposits pass up-dip into coastal plain (Sleipner Formation) deposits. In the third interval (between an MFS A and MFS J44), offshore mudstone (Heather Formation) deposition dominated the study area with shoreface deposition (Hugin Formation) locally restricted to the central and northernmost parts of the study area. Interpretations of the age and depositional setting of sandstones in the northernmost well (well I, Fig. 5.7) is, however, poorly constrained by biostratigraphical and core data. Similarly, the fourth interval (between MFS J44 and MFS J46) is dominated by offshore mudstone (Heather Formation) deposition across the study area, with two exceptions. Firstly, shoreface sandstones persist in the central part of the study area (well E, Fig. 5.7). Secondly, a series of parasequences comprising fluvio-tidal channel-fill, bay-fill, mouth bar and coastal plain deposits built out from the south (well A, Fig. 5.7), as part of a major delta system in the area of the Sleipner Vest Field (Vollset and Doré, 1984; Milner and Olsen, 1998; Folkestad and Satur, 2008). This depositional pattern persists into the fifth, uppermost interval (between MFS J46 and MFS J52), with both shoreface and fluvio-deltaic deposits flooded by offshore mudstones (Heather Formation) in the upper part of this interval.

The intervals defined by regional maximum flooding surfaces can be further subdivided by flooding or transgressive surfaces of sub-regional and local extent. The limited extent of these surfaces most likely reflects variations in sediment supply and accommodation creation that were caused by rift-fault initiation and

linkage and/or by the growth of salt-cored highs (e.g. Figs. 5.7b, 5.8b). Although they may also reflect autogenic lateral migration of delta lobes. Sequence boundaries of local extent are identified in the southernmost part of the study area (well A, Fig. 5.7). These sequence boundaries are traced through the Sleipner Vest Field (Folkestad and Satur, 2008), but can not be correlated more widely. Their limited extent is typical of net-transgressive strata, within which sequence boundary expression is suppressed (e.g. Nummedal and Swift, 1987; Sneider *et al.*, 1995), and also of rift basins, where sequence boundary development is enhanced by uplift in marginal areas and suppressed by increased subsidence in axial depocentres (e.g. Mitchener *et al.*, 1992; Gawthorpe *et al.*, 1994).

5.6.2 Axis-perpendicular sub-regional correlation, Sigrun-Gudrun area

In the northern part of the study area, near the Sigrun and Gudrun fields, thick (up to 200m) successions of up to 12 stacked bay-fill and shoreface parasequences are defined by localised transgressive flooding surfaces (Figs. 5.3, 5.8). Most of this succession occurs within the second regionally mapped interval (between MFS J36 and MFS A). The underlying regionally mapped interval (below MFS J36) lies below the interval penetrated in the westward-lying wells (well J, Fig. 5.8) and is represented by coastal plain (Sleipner Formation) deposits or an unconformity in the eastern part of the Sigrun-Gudrun area. The bay-fill parasequences thicken markedly towards the west, whilst thinning to the east where they are interpreted to onlap Triassic strata. Westward thickening also occurs in the hangingwall of a normal fault in the western part of the Sigrun-Gudrun area, but this thickening is not accompanied by a noticeable facies shift. Several of the bay-fill parasequences are truncated by thick (up to 45 m) channel-fill sandstones that are not cored. These

Figure 5.8 Local correlation at Sigrun-Gudrun. (a) Local W-E-oriented correlation panel and (b) corresponding seismic section perpendicular to the axis of the South Viking Graben in the Sigrun-Gudrun area (see Fig. 5.3 for location). The Hugin Formation becomes younger towards the east. There is some ambiguity interpreting the thick sandstones as tidal inlet channels since they are uncored. They could be interpreted as fluvio-tidal channels instead. The key is the same as for Figure 5.7.

sandstones are interpreted as tidal inlet channel-fill deposits (facies BF4), based on their clean, low GR character, which is consistent with marine reworking, and the occurrence of shoreface sandstones in overlying strata, but they may alternatively represent fluvio-tidal channel-fill deposits (facies FTC1, FTC2). Shoreface sandstones occur in the uppermost Hugin Formation, above the stacked bay-fill parasequences, in the western part of the Sigrun-Gudrun area. These shoreface deposits pass eastward into bay-fill deposits. Prior to deposition of regional flooding surface MFS A, most of the Sigrun-Gudrun area had been occupied by offshore mudstones (Heather Formation). Above this surface, in the third and fourth regionally mapped intervals (between MFS A and MFS J46), coastal plain deposits (Sleipner Formation) unconformably overly Triassic strata and pass westward into bay-fill deposits and offshore mudstones (Heather Formation). The entire Sigrun-Gudrun area was flooded during the fifth regionally mapped interval (between MFS J46 and MFS J52).

Flooding of the Sigrun-Gudrun area was diachronous, becoming younger from west to east. The relatively low-energy, restricted-marine character of the bay-fill parasequences implies the presence further west of coeval barriers and spits fronted by wave-dominated shorefaces (Harris and Fowler, 1987; Hampson *et al.*, 2009) or rift-generated islands and shoals. The local occurrence of wave-dominated shoreface deposits above the bay-fill parasequences supports the former interpretation, although the latter cannot be discounted. In any case, aggradational stacking of the bay-fill parasequences indicates that sediment supply and accommodation generated by tectonic subsidence were roughly in balance during deposition of the second regionally mapped interval (between MFS J36 and MFS A).

5.6.3 Axis-perpendicular sub-regional correlation, Dagny area

In the centre of the study area, around the Dagny Field, localised flooding surfaces subdivide the five regionally mapped intervals into 11 parasequences (Figs. 5.3, 5.9), which are not correlatable to those in the Sigrun-Gudrun area (Fig. 5.8). Coastal plain (Sleipner Formation) deposition occurred across the Dagny area in the lowermost regionally mapped interval (below MFS J36). Marine flooding of the Dagny area commenced during deposition of the second regionally mapped interval (between MFS J36 and MFS A). Within this interval, offshore mudstones (Heather Formation) in the NW pass to the south through mouth bar and shoreface deposits (Hugin Formation) into coastal plain deposits (Sleipner Formation). Within the overlying regionally mapped interval (between MFS A and MFS J44), marine flooding occurred over the whole of the Dagny area, with shoreface sandstones and offshore mudstones being deposited. The base and top of this interval are marked by local angular unconformities. The lower unconformity is associated with onlap or downlap of overlying shoreface parasequences (in well B, Fig. 5.9), while the upper unconformity is marked by deep erosional truncation of underlying shoreface parasequences and correlative offshore mudstones in a different location (in well E, Fig. 5.9). The fourth regionally mapped interval (between MFS J44 and MFS J46) is present across the Dagny area and contains laterally discontinuous shoreface sandstones (in wells C and E, Fig. 5.9) that pinch out to the NW and SE into offshore shales (Heather Formation). The entire Dagny area was flooded prior to deposition of regional maximum flooding surface MFS J46, with the exception of well E (Fig. 5.9), in which shoreface sandstones were deposited. Biostratigraphical data indicate that these sandstones are capped by another locally developed

Figure 5.9 Local correlation at Dagny-Alpha North. (a) Local NW-SE-oriented correlation panel and (b) corresponding seismic section oblique to the axis of the South Viking Graben in the Dagny area (see Fig. 5.3 for location). The Hugin Formation is highly diachronous and exhibits large local thickness variations. The key is the same as for Figure 5.7.

unconformity, which lies within the fifth regionally mapped interval (between MFS J46 and MFS J52) and is overlain by offshore mudstones (Heather Formation).

The mouth bar deposits interpreted within the second and third regionally mapped intervals (between MFS J36 and MFS J44) in the NW part of the Dagny area pass up-dip into coastal plain deposits (Sleipner Formation) (Fig. 5.9), suggesting that they are linked to either the delta system in the Sleipner Vest area to the south or to small rivers draining the eastern graben margin. This subtle fluvial signature is absent in younger units, corresponding to the fourth and fifth regionally mapped intervals (between MFS J44 and MFS J52). In these younger intervals, the complex and highly localised distribution of shoreface deposits suggests that they were geographically detached from the Sleipner Vest delta system to the south and potentially from the shoreline along the eastern graben margin. The resulting shoreface sandstone distribution is interpreted to have been controlled by syn-depositional development of salt-cored highs and rotated fault-block crests that formed islands and shoals whilst deeper water depositional conditions prevailed elsewhere in the Dagny area. These “isolated” shoreface sandstones may have been deposited by in-place reworking of older sandstones that were eroded at angular unconformities, or they may have been fed by southward-directed longshore currents that nourished spit systems connected to the Utsira High, on the eastern graben margin (Fig. 5.1). The rapid facies transitions from syn-depositional topographic highs, marked by angular unconformities, into adjacent areas of expanded stratigraphy indicate that sediment supply was not sufficient to keep pace with accommodation generated by rift-fault development and salt movement.

5.7 Palaeogeographical Reconstruction

5.7.1 Palaeogeographical reconstruction, MFS A

The well correlations described above highlight the diachroneity and complex spatial distribution of flooding within the South Viking Graben during the Middle Jurassic. These patterns become clearer in palaeogeographic reconstructions.

During deposition of regional maximum flooding surface MFS A (Figs. 5.7-9), coastal plain deposition (Sleipner Formation) was prevalent throughout much of the study area, with shallow-marine sedimentation limited to the NW, in the Sigrun-Gudrun area (Fig. 5.10a). Barrier islands fronted by wave-dominated shorefaces are interpreted to have been the predominant shoreline type in the northern part of the study area. These barrier islands sheltered a restricted marine embayment to the east, on the western flank of the Utsira High. Minor river systems are interpreted to have drained from the Utsira High down the hanging-wall dipslope towards the main graben-bounding fault and into this embayment. Time-equivalent deep marine sedimentation is interpreted to the west of the shoreline, proximal to the basin axis. The barrier island system may have been supplied with sediment by southward-directed longshore drift generated by oblique wave approach from the north or NW, or by transgressive reworking of underlying bay-fill deposits during barrier retreat. The former mechanism appears more likely, because the shoreface sandstones are significantly coarser grained than the bay-fill deposits. The ultimate source of such coarse sand was most likely the Utsira High (Fig. 5.1), where Cretaceous and Tertiary sediments directly overlie eroded granitic basement (e.g. Zanella and

Figure 5.10 Palaeogeographical reconstructions of deposition in the study area during (a) Early Callovian (MFS A) and (b) Middle Callovian (MFS J44) times.

Coward, 2003), although potential drainage systems that supplied this sediment to the shoreline are not resolved in seismic and well data.

A salt-cored high is mapped in seismic data to the south of the restricted marine embayment, in an area of no well control in between the Sigrun-Gudrun and Dagny areas; Middle Jurassic strata are absent over this structure. This absence implies either non-deposition or subsequent erosion of the Middle Jurassic section; the former interpretation would indicate that the structure formed a syn-depositional topographic high that may have defined the southern limit of the restricted marine embayment.

To the south of this salt-cored high, in the Dagny area, shoreface sandstones abut directly against coastal plain deposits, implying a wave-dominated strandplain shoreline. Further south, in the area around the Sleipner Vest Field, fluvial-deltaic deposition occurred. The deposits of the “Sleipner Delta” were influenced by both waves and tides (Folkestad and Satur, 2008). Although the limited wave energy in these deposits appears to contradict the interpretation of strong wave-generated longshore currents in the Sigrun-Gudrun area, the discrepancy can be explained by the more enclosed and sheltered location of the Sleipner Vest area at the southern limit of the embryonic South Viking Graben. Folkestad and Satur (2008) argue that the narrow geometry of the graben in this area would have suppressed wave action and amplified tidal currents. Dominant fluvial drainage into the South Viking Graben is interpreted to have been from the south, along the basin axis, and the “Sleipner Delta” deposits form a small component of the regional back-stepping

Middle-to-Upper Jurassic “Brent Delta” systems (Rathey and Hayward, 1993; Milner and Olsen, 1998; Husmo *et al.*, 2003).

5.7.2 Palaeogeographical reconstruction, MFS J44

During deposition of regional maximum flooding surface MFS J44 (Figs. 5.7-9), the majority of the study area was flooded and offshore mudstone deposition (Heather Formation) was prevalent (Fig. 5.10b), with deep marine sandstones (Ling Member, Heather Formation) deposited proximal to the basin axis within the UK sector (e.g. in well 16/8A-4; Cockings *et al.*, 1992). Coastal plain deposits (Sleipner Formation) lay in the SE of the study area, fringed by shoreline deposits that are not penetrated by wells. These deposits are inferred to be shoreface sandstones rather than bay-fill heteroliths due to the wider, more open geometry of the basin at this time of deposition. “Isolated” shoreface sandstones are locally developed within the Dagny area, most likely fringing syn-depositional topographic highs formed by fault-block crests and salt-cored structures. Fluvial-deltaic deposits are not present in the study area at this time but occur further south in the area of the Sleipner Vest and Øst fields (Folkestad and Satur, 2008). The “Sleipner Delta” system later prograded to the north and west, to deposit younger fluvio-deltaic units within the study area (between MFS J44 and MFS J52 in well A, Fig. 5.7).

5.8 Summary and implications for exploration

The Hugin Fm was deposited during a tectonically-driven transgression which was associated with flooding of the South Viking Graben. Biostratigraphical data permit robust correlations of regional maximum flooding surfaces over considerable

distances (i.e. several tens of kilometres) and indicate overall younging of the unit up the hangingwall to the east, and along the main basin axis to the south. Relatively early flooding (Middle Callovian, MFS A) of the southernmost wells, however, implies that open marine conditions may have existed within the basin axis at this time. Biostratigraphically-driven correlations also indicate that each of the main facies associations was developed within a different structural location and, therefore, had local stratigraphic development controlled by markedly different tectonic styles. In addition, distance from the main fluvial sediment supply system into the basin axis, located to the south of the study area, was also a key control on facies development and distribution. Parasequence-bounding flooding surfaces and candidate sequence boundaries are also recognised and correlated locally; they cannot, however, be confidently correlated throughout the area. This is interpreted to be a function of the overall transgressive nature of the succession and, perhaps most importantly, significant spatial variations in fault- and halokinesis-controlled subsidence across the basin. Differential subsidence restricted sequence boundary development to the basin flanks and these surfaces are interpreted to pass into correlative conformities or flooding surfaces in more basinal settings.

The complex spatial variations in facies within the Hugin Formation have major implications for regional exploration within the South Viking Graben. The best reservoir intervals are likely to be in the upper part of the unit where shallow-marine rather than restricted-marine deposits are best developed. These units occur; (i) in northern parts of study area, where N-S-trending belts of shoreface sandstone occur in addition to more laterally-restricted, tidal inlet channel-fill sandstones which erode into bay-fill deposits of marginal reservoir quality. These potential reservoir

sandstones, although deposited during active growth faulting, do not appear to have had their location or areal extent determined by syn-depositional tectonic features. Therefore, predicting their occurrence based on structural location is problematic;

(ii) in central parts of study area, where shoreface sandstones with complex plan-view geometries and rapid lateral pinchouts occur. These sandstones fringe or overlie growing halokinesis-related structures and other palaeo-topographic highs;

(iii) in southern parts of study area, where N-S-trending, channelised and lobate mouth-bar sandstones of the “Sleipner Delta” remained relatively fixed in overall location. This sandstone distribution is caused by high fluvial sediment supply from the south, focussed along the main axial drainage pathway;

(iv) in the basin centre, where axially-fed deepwater fan deposits of the Ling Sandstone Member (Heather Formation) occur. These deepwater sandstones may be genetically-linked to the shallow-marine and coastal-plain deposits of the “Sleipner Delta”. These observations clearly indicate that sandstone-belt geometry and distribution is closely related to both depositional regime (i.e. shoreface, river-dominated delta, barred lagoons) and structural location. Therefore, a robust understanding of temporal and spatial variation in depositional system type and tectonically driven, syn-depositional basin geometry is critical to successful exploration of reservoirs in the Hugin Formation and coeval strata. This is discussed in more detail in chapters 7-9.

CHAPTER 6

Sedimentology and sequence stratigraphy of the upper part of the Almond Formation, Rock Springs, Wyoming, USA: an outcrop analogue for the Hugin Formation

The upper part of the Almond Formation is considered to be an outcrop analogue for the shallow-marine Hugin Formation, from a setting which lacks strong syn-depositional structural control. The upper Almond Formation was deposited during the late Cretaceous (Campanian-Maastrichtian) on the western margin of the Cretaceous Interior Seaway of North America, and is comparable to the Hugin Formation in terms of: (i) overall depositional setting (i.e. coastal system with tidally influenced lagoon/embayment and wave-influenced shoreface components); (ii) detailed facies character; and (iii) back-stepping, net-transgressive stratigraphic architecture. Studying the upper part of the Almond Formation provides insights into the facies distributions and stratigraphic architecture of shallow-marine strata deposited during transgression in a comparatively simple structural setting. Application of these insights has been used to help constrain well-based correlations in the Hugin Formation, and thus to develop more robust sedimentological models and reservoir quality predictions within the more structurally complex setting of the Jurassic South Viking Graben.

6.1 Introduction

Transgressive depositional sequences record landward migration of the coastline during a relative rise of sea level (Posamentier and Vail, 1988; Emery and Meyers,

1996; Helland-Hansen and Martinsen, 1996; Van Wagoner *et al.*, 1996; Cattaneo and Steel, 2003). From an applied perspective, net-transgressive shallow marine reservoirs host a significant volume of hydrocarbons in the North Sea Basin (Husmo *et al.*, 2003) and other basins worldwide. However, research into shallow marine systems has frequently focussed on net-regressive units that record an overall basinward migration of the coastline, due in part to better preservation of their associated deposits. Transgressive deposits can be highly variable in terms of sandbody geometry and distribution, thereby making exploration for, and production from, such units challenging (Nummedal and Swift, 1987; Maxwell *et al.*, 1999; Cattaneo and Steel, 2003; Sixsmith *et al.*, 2008).

The variability of transgressive deposits can be further enhanced by syn-depositional structural development, a common problem within transgressive syn-rift reservoirs of the North Sea such as the Tarbert Formation (Davies *et al.*, 2000), Fulmar Formation (Hodgson *et al.*, 1992) and Hugin Formation (Chapter 5; Kessler *et al.*, 1995; Sneider *et al.*, 1995; Milner and Olsen, 1998). Reconstructing facies architectures and distributions within these formations using subsurface data is difficult because most stratigraphic relationships are below seismic resolution and because core and well-log data are relatively sparse. Studying outcrop examples of transgressive units can therefore provide important qualitative and quantitative information to constrain correlations within the subsurface. This study focuses on the transgressive, upper part of the Almond Formation seen at outcrop in SW Wyoming, USA. This outcropping depositional system is proposed as an analogue for the Hugin Formation due to its similar detailed facies character and interpreted depositional setting (barrier-bay shoreline, discussed more below). Although syn-

depositional structures are documented to influence the deposition of the Almond Formation in the subsurface Washakie Sub-basin and Wamsutter Arch, which lie to the east of the outcrop belt (e.g. Jackson and Rawn-Schatzinger, 1993; Clawson and Fawret, 1997; Krystinik and Mead, 1998), the structural setting during deposition in the region of the Rock Springs Anticline was comparatively simple.

6.2 Geological setting and previous work

The Cretaceous Western Interior Seaway of North America lay within a wide (ca. 1500 km), N-S trending, asymmetrical basin produced by short-wavelength, thrust-sheet loading in the Sevier Orogenic Belt and long-wavelength, subduction-related subsidence at its western margin (Fig. 6.1a; e.g. Kauffman and Caldwell, 1993; Liu and Nummedal, 2004). The Sevier Orogenic Belt formed as a result of several phases of thrusting from late Jurassic through to Early Eocene times, following convergence between the North American and Farallon plates (e.g. Armstrong, 1968; Jordan, 1981; Cross, 1986; Kovach *et al.*, 2001; Liu *et al.*, 2005). The broad Cretaceous basin that extended eastwards from the thrust front during the Sevier Orogeny was then segmented by basement-involved tectonism during the Tertiary Laramide Orogeny (Kovach *et al.*, 2001). Tectonic subsidence, coupled with high global eustatic sea level during the Cretaceous, led to rapid flooding of the basin and formation of the Cretaceous Interior Seaway which extended from the Arctic Ocean to the Gulf of Mexico (Fig. 6.1a; e.g. Armstrong, 1968; Hendricks, 1983; Cross, 1986; Roehler, 1989; Finn and Johnson, 2005). By the close of the Cretaceous, the basin was largely infilled and marine conditions ended (Finn and Johnson, 2005).

Figure 6.1. Depositional Setting of the Almond Formation. (a) Location map showing the position of the Cretaceous Interior Seaway, an epicontinental sea that developed across the present day North America continent. The location of figure (c) is highlighted (modified from Kauffman and Caldwell, 1993) (b) Late Cretaceous Stratigraphy within the study area. The shallow marine, upper part of the Almond Formation overlies continental deposits of the lower part of the Almond and Ericson formations whilst itself is overlain by and interfingers with the open marine shales of the Lewis Shale. (c) Palaeogeographical reconstruction showing the depositional setting of the Almond Formation within the Rock Springs study area. The Rock Springs embayment lay between the Red Desert Delta to the north and an unnamed delta to the south (redrawn from Roehler, 1988).

Eustatic sea-level variations combined with laterally variable tectonic subsidence and sediment supply led to deposition of a series of large-scale, regressive-transgressive depositional cycles within the Cretaceous Interior Seaway (Molenaar, 1983; Donselaar, 1989; Keighin *et al.*, 1989; Krystinik and DeJarnett, 1995; Olsen *et al.*, 1999; Kovach *et al.*, 2001; Siedler and Steel, 2001). These cycles are expressed as siliciclastic-sediment wedges that comprise coastal plain and shallow-marine strata deposited along the western margin of the seaway (Fig. 6.1). Sediment within these wedges was predominantly eroded and transported from the Sevier orogenic fold-and-thrust belt to the west of the seaway. The Almond Formation is a net-transgressive unit deposited on the western margin of the Cretaceous Interior Seaway, and forms the upper part of a regressive-transgressive wedge, together with the underlying regressive Rock Springs and Ericson formations (Fig. 6.1b; Jacka, 1965; Hendricks, 1994; Horn *et al.*, 2002). The Almond Formation is late Campanian to early Maastrichtian in age and forms part of the Upper Cretaceous Mesaverde Group. The unit overlies the fluvial Ericson Formation and is itself overlain by and interfingers with the offshore marine Lewis Shale (Fig. 6.1b) (e.g. Jacka, 1965; Hendricks, 1994; Garcia and Gonzales, 1997; Finn and Johnson, 2005).

The Almond Formation has previously been studied where it outcrops on the flanks of the Rock Springs Uplift (e.g. Roehler, 1988) and also from the subsurface of the Laramide (Tertiary) Washakie Sub-basin and Wamsutter Arch (Fig. 6.2), where it forms an important reservoir unit (e.g. Keighin *et al.*, 1989; Krystinik and Mead, 1998). The Almond Formation has been subdivided into a lower, alluvial unit and an upper, shallow marine unit which forms the focus of this study (Jacka, 1965; Van Horn, 1979; Sturm *et al.*, 2001). Several facies and facies associations have been

Figure 6.2 Structural Setting of the Rock Springs Uplift. The Rock Springs Uplift is a basement involved feature that formed during the Cretaceous to Early Tertiary within the Greater Green River Basin of the Rocky Mountain Foreland Belt (modified from Hendricks, 1994).

interpreted previously within the shallow marine Almond Formation including; tidal channel, lagoonal, estuarine, tidal delta, marsh, shoreface and foreshore (e.g. Jacka, 1965; Roehler, 1979; Jackson and Rawn-Schatzinger, 1993; Hendricks, 1994; Schatzinger and Tomutsa, 1999; Caldaro-Baird *et al.*, 2001; Kovach *et al.*, 2001; Martinsen, 2003). The depositional setting for the Almond Formation within the Rock Springs region has typically been interpreted as a N-S trending barrier-bay shoreline with barrier island sandstones interpreted within several stratigraphic intervals (Fig. 6.1c) (e.g. Roehler, 1979; 1988) or less commonly as a component part of a large, asymmetrical delta system (Bhattacharya and Giosan, 2003).

During deposition of the Almond Formation, the area of the Rock Springs uplift is interpreted to have formed an embayment (the Rock Springs Embayment) between the Red Desert Delta to the north and an unnamed delta in NW Colorado to the south (Fig. 6.1c; e.g. Roehler, 1988; Martinsen and Christensen, 1992; Jackson and Rawn-Schatzinger, 1993; Schatzinger and Tomutsa, 1999). This embayment may have been partly tectonic in origin, due to inferred reactivation of a basement fault that underlies the Rock Springs Uplift, which produced a topographic low in the area (discussed below; Jackson and Rawn-Schatzinger, 1993; Hendricks, 1994; Martinsen, 2003).

The Rock Springs Uplift forms an elongate, N-S oriented, doubly plunging anticline situated approximately 200 km eastwards of the Sevier Thrust Belt within the Rocky Mountain Foreland (Fig. 6.2; Mederos *et al.*, 2005). The uplift lies within the Greater Green River Basin, separating it into the Green River sub-basin to the west and the Washakie and Sand Wash sub-basins to the east (Fig. 6.2). The uplift is a basement-

cored uplift that formed through a series of stages during the latest Cretaceous to early Tertiary (e.g. Garing and Tainter, 1985; Montgomery, 1996; Mederos *et al.*, 2005). The initial phase of uplift is interpreted to have occurred during the Campanian and Maastrichtian, creating a broad arch and influencing deposition of the Ericson Formation (Martinsen *et al.*, 1999; Mederos *et al.*, 2005). A series of E-W to ENE-WSW trending normal faults are identified within the Cretaceous section across the uplift and are also interpreted to have been active during deposition of the Ericson Formation (Fig. 6.3a; Martinsen *et al.*, 1999; Mederos *et al.*, 2005). Reverse movement on the basement fault underlying the Rock Springs Uplift is interpreted during Almond Formation deposition, leading to development of the Rock Springs Embayment (Hendricks, 1994; Martinsen, 2003). The uplift was reactivated during the Eocene to become the tighter structure observed present day (Mederos *et al.*, 2005).

6.3 Dataset and Methodology

Outcrop of the Almond Formation occurs almost exclusively on the eastern flank of the Rock Springs Uplift (Fig. 6.3a), such that stratigraphical relationships are exposed largely along depositional strike with only relatively limited control along depositional dip. The outcrop dataset collected from the upper part of the Almond Formation consists of 48 measured sections comprising more than 2000 m of section (Fig. 6.3). The outcrop data were collected from two study areas that lie 26 km apart on the eastern flank of the Rock Springs Uplift (Fig. 6.3b, c). A lack of good quality outcrop between these two areas, due to modification of the landscape for coal extraction, and a lack of biostratigraphical data mean it is not possible to correlate

Figure 6.3 Database Map. (a) Map of the Rock Springs Uplift. The extent of the outcrop is shown, along with the position of measured sections and correlation panels discussed later in the text (redrawn from Love and Christiansen, 1985).

(c)

(b)

Figure 6.3 Database Map (cont.) (b) Close up of the Big Flat Draw study area. (c) Close up of the Point of Rocks study area. The type log presented in Fig. 6.4 is highlighted, along with measured sections used in the presented correlation panels (Figs. 6.8 and 6.11). Note the approximate orientation of the shoreline (after Roehler, 1988).

strata between the two study areas with confidence; therefore two correlation panels are presented, one through each study area (Fig. 6.3b,c). Not all measured sections have been included on these correlation panels for two reasons; (i) where poor or limited outcrop prevented a confident correlation to other measured sections (e.g. to the north of Big Flat Draw, Fig. 6.3b); or (ii) when several sections had been measured in close proximity and including them over-complicated the correlation whilst not affecting the interpretation presented (e.g. Dead Man's Wash, Fig. 6.3c). Sedimentological analysis during logging of the measured sections involved recording grain size, sorting and sphericity of matrix and clasts, colour variations, sedimentary structures, bed thickness, the nature of bounding contacts, and macro-fauna type and preservation state. The degree of bioturbation was recorded using the bioturbation index (BI) scheme of Taylor and Goldring (1993), which is based on a visual assessment of the amount of 'bioturbated' versus 'unbioturbated' texture and ranges from 0 (no bioturbation) to 6 (complete bioturbation). Sedimentological logs have been combined with photomontages of the low ridges that form the exposures, low-level aerial photographs and more than 1350 palaeocurrent readings.

6.4 Facies and Facies Associations

Thirteen facies have been recognised within the upper part of the Almond Formation (Table 6.1). These facies have been grouped into three facies associations, which are discussed below and summarised in Figs. 6.4-6.7.

Facies	Description	Bioturbation	Interpretation
Laminated mudstone (BF1)	Parallel-laminated mudstone with rare siltstone laminae. Carbonaceous material throughout. Very rare, small bioclastic fragments, dominated by disarticulated bivalves oriented along lamination. Typically poorly exposed.	BI = 0-1 (Pl)	Restricted marine shale
Interbedded heteroliths (BF2)	Parallel-laminated mudstone with mm-cm scale siltstone and very fine-grained sandstone laminae and beds. Carbonaceous material (including leaves) throughout, sometimes concentrated into laminae. Rare wave ripples within sandstone beds.	BI = 0-4 (Pl)	Shallow bay-fill
Upward coarsening sandstone (BF3)	Upward-coarsening, siltstone to fine- and medium-grained sandstone. Sandstone beds thicken and amalgamate upwards. Planar to low-angle cross-lamination, ripple cross-lamination, symmetrical ripple forms, and carbonaceous and mudstone drapes common in sandstone beds. Sandstones also contain rare rounded mudclasts, soft-sediment deformation and hummocky cross stratification.	BI = 0-3 (Op)	Sandy bay-fill
Bioclastic-rich sandstone (BF4)	Fine- to upper medium-grained sandstone exhibiting upward-coarsening, upward-fining and uniform grain-size trends. Sharp basal contact, in places exhibits cm-scale erosional relief. Bioclastic material (predominantly disarticulated bivalves) and rounded mudclasts are common, lining set boundaries and foresets within cross-bed sets and also concentrated into distinct beds (upto 75 cm thick). Wavy lamination and bioturbation observed where bioclastic material is less.	BI = 0-2 (Op)	Bioclastic-rich bay-fill
Upward fining trough cross bedded sandstone (BF5)	Upward-fining to blocky, medium- to fine-grained, moderately to poorly sorted sandstone occurring within erosively based, laterally discontinuous units. Trough cross-sets containing bioclastic laminae and mudstone drapes, low-angle cross-lamination and ripple cross-lamination are common throughout. Mudclasts and wood fragments are preserved lining cross sets and are typically more common towards the base of the facies. Rare lateral accretion surfaces. Bi-directional palaeocurrents.	BI = 0-5 (Sk, Op)	Tidal inlet channel-fill
Hummocky cross-stratified sandstone (SF1)	Upward-coarsening successions of fine-grained, moderately sorted sandstone beds and interbedded carbonaceous siltstones. Sandstone beds thicken and amalgamate upwards. Hummocky cross-stratification is common; rare swaley and trough cross-stratification occur in the upper part of successions. Sandstone bed bases are lined by mudclasts and wood fragments. Where preserved, bed tops contain ripple cross-lamination.	BI = 2-5 (Op, Ch, Pl, Sk, Ter, Tha, ET)	Lower shoreface
Trough to swaley cross-stratified sandstone (SF2)	Blocky or upward-coarsening, fine- to medium-grained, moderately to well sorted sandstone. Trough and swaley cross-stratification is common, with subordinate low-angle planar cross-sets, planar-parallel and wavy lamination. Symmetrical ripples, carbonaceous drapes and mudstone laminae are common at bed tops. Mudclasts are rare. Distinctive silver-grey colour with rust-coloured staining of burrows in weathered exposures.	BI = 2-4 (Op, Pl, As, ET, Roots)	Upper shoreface
Parallel bedded sandstone (SF3)	Planar-parallel bedded, fine- to medium-grained sandstone. Pervasively root penetrated. Typically overlain by coal.	BI = 0-2 (Roots)	Foreshore
Coal and carbonaceous shale (CP1)	Laminated, fissile coal and carbonaceous shale with flat basal contact above root-penetarted interval. Poorly exposed.	BI = 0	Coal
Rooted interbedded heteroliths (CP2)	Interbedded dark grey, carbonaceous mudstone and siltstone to very fine-grained sandstone. Wavy to parallel lamination penetrated and disrupted by roots.	BI = 3 (roots)	Coastal plain fines
Upward-fining sandstone with unidirectional palaeocurrents (CP3)	Sharp-based, fining-upward successions of medium- to fine-grained sandstone. Trough to planar cross-beds containing carbonaceous drapes and mudclasts. Unidirectional palaeocurrents.	BI = 0-1 (Pl)	Fluvial and tidally-influenced fluvial channel-fill sandstone

Table 6.1. Summary of facies in the upper part of the Almond Formation. Bioturbation fabric is described using the Bioturbation Index (BI) scheme of Taylor and Goldring (1993). Individual trace fossils, listed in brackets, are abbreviated as follows: Pl = *Planolites*, Op = *Ophiomorpha*, Sk = *Skolithos*, Ch = *Chondrites*, Ter = *Terebellina* (Schaub. *Fria*), Tha = *Thalassinoides*, As = *Asterosoma*, ET = Escape traces.

Figure 6.4 Type sedimentary section through the upper part of the Almond Formation. The main facies, facies associations, sequence stratigraphic surfaces and vertical facies stacking patterns are highlighted. Presented example from Big Flat Draw MS6, for location see Fig. 6.3b. The section is included in the correlation panel presented in Fig. 6.8.

6.4.1 Bay-fill facies association

The bay-fill facies association commonly overlies upward-shallowing successions of the shoreface facies association, or overlies or interfingers successions of the coastal plain facies association. The thickest and most complete bay-fill successions are observed within the westward, palaeo-landward part of the studied interval (e.g. Salt Wells Creek, Fig. 6.3b; Road to Superior, Fig. 6.3c). Fine-grained components of the bay-fill facies association (e.g. facies BF1) are poorly exposed, and thus difficult to distinguish from fine-grained components of the coastal plain facies association (facies CP1, CP2) and from the open-marine Lewis Shale.

Description. Deposits of this facies association typically comprise gradational, upward coarsening successions (ranging from *ca.* 2 to 7 m thick) of facies BF1, BF2 and BF3, which are capped by thin coastal plain successions (e.g. at 10-15.5 m, Fig. 6.4). Parallel laminated mudstones and siltstones (facies BF1; Fig. 6.5a) typically form the base of these successions and pass gradationally upwards into interbedded heteroliths (facies BF2; Fig. 6.5b). In some locations, sandstone beds thicken and coarsen upwards, whilst finer grained material is limited to thin laminae or drapes (facies BF3; Fig. 6.5c). Fine- to medium-grained, bioclastic sandstones (facies BF4; Fig. 6.5d) occur in three locations: (1) in groups of beds at the base of bay-fill facies successions (0.8 to 5 m thick, up to 4.5 km wide), (2) in units that sharply, in places erosionally, overlie progradational sequences of BF1, BF2 and BF3 (<1 to 7 m thick; up to 8 km wide; e.g. at 17.5-21 m, Fig. 6.4), and (3) as thin (20-35cm) poorly exposed, laterally discontinuous (<1 km wide) groups of beds within facies BF1 and BF2. This facies contains a significantly higher proportion of bioclastic material than other facies within the association. Bioclastic material predominantly comprises a

Figure 6.5 Photograph examples of the Bay Fill Facies Association. (a) Facies BF1 - Bay-fill fines (SWC MS4 at 15m; Figs. 6.3b, 6.8). Rare bioclastic material preserved within poorly exposed unit. (b) Facies BF2 - Interbedded heteroliths (BFD MS1 at 43m; Figs. 6.3b, 6.8). Low to moderate bioturbation disrupts lamination. (c) Facies BF3 - Sandy bay fill (BFD MS11 at 10.5 m; Figs 6.3b, 6.8). Upward coarsening silt to sandstone, ripple-scale cross lamination common throughout. (d) Facies BF4 - Bioclastic sandstone (BBC MS1 at 9m; Fig. 6.3b). Sharp based, sandstone matrix support dense collection of bioclastic material. (e) Facies BF5 - Tidal-inlet channel fill (BFD MS 6 at 6 m; Figs. 6.3b, 6.8). Erosively based sandstone with prolific mudstone drapes and bidirectional palaeocurrent indicators. (f) Facies BF5 - Erosionally based tidal-inlet channel (BBC MS 1 at 5m; Fig. 6.3b).

low diversity assemblage of disarticulated bivalves, identified as *Crassostrea* (sp.) and interpreted to reflect a brackish water environment (Roehler, 1988). Cross-bed palaeocurrents indicate sediment transport to the east (offshore-directed) and northeast (shoreline-parallel). Elsewhere, the finer-grained sediments of facies BF1 and BF2 are sharply overlain by laterally discontinuous, upward-fining, cross-bedded sandstones within which bioclastic material and mudstone rip-up clasts are observed to line cross-sets, typically near the erosional base of these sandstone units (facies BF5; Fig. 6.5e,f). Bioturbation throughout this facies association is low to moderate and comprises a low diversity trace fossil assemblage.

Interpretation. Upward-coarsening successions of facies BF1, BF2 and BF3 are interpreted to exhibit a shallowing-upward trend, with increasing proximity to the source of sand towards the top of the succession. Parallel-laminated mudstones record deposition from suspension, while sandstone laminae and cross-laminated beds record episodic deposition from suspension and tractional currents, respectively. Quasi-rhythmical interbedding of mudstone and sandstone (facies BF2 and BF3) suggest regular and repeated fluctuations in energy regime, which can be indicative of tidal action (e.g. Nio and Yang, 1991). Wave and storm influence is also present, indicated by rare wave ripples and hummocky cross stratification (e.g. Dott and Bougeois, 1982; Duke *et al.*, 1991). Fine to medium-grained, cross-bedded, sharp to erosionally based bioclastic sandstones (facies BF4) were deposited from high energy, tractional currents whilst preservation of whole valves implies short transport distances. These observations, coupled with the occurrence of eastward-directed palaeocurrents (facies BF4) may indicate deposition by ebb-tidal currents within tidal channels that eroded the back-barrier succession behind a wave-

dominated barrier shoreline. Locally developed, erosively based, upward fining-sandstones with rare lateral accretion surfaces are also interpreted as channel fills (facies BF5). Mudstone drapes in these sandstones indicate fluctuations in flow velocity, consistent with tidal influence, and the occurrence of bioclastic and mudclast lags, together with mudstone drapes, is common in tidal inlet channel-fill deposits in barrier island shorelines (e.g. Moslow and Tye, 1985; Israel *et al.*, 1987; Simms *et al.*, 2006).

The occurrence of low diversity bioclastic material and trace fossils throughout this facies association indicates deposition in a restricted marine environment (Pemberton *et al.*, 1992; MacEachern and Bann, 2008). Successions of this facies association are therefore interpreted to record periodic infilling of restricted marine embayments and/or lagoons. The restricted marine character of these successions and close association with the shoreface facies association may be explained by deposition in sheltered lagoons and/or embayments that lay behind coeval barriers and spits fronted by wave-dominated shorefaces (Oertel, 1985). The tidal signatures observed throughout this facies association may simply reflect the absence of wave reworking in the sheltered lagoons and/or embayments. Alternatively, the postulated syn-depositional tectonic setting of the Rock Springs Uplift, combined with the occurrence of deltas to the north and south of the study area, may have created an embayed coastline that focussed tidal currents and limited wave action (e.g. Roehler, 1988, Jackson and Rawn-Schatzinger, 1993; Schatzinger and Tomutsa, 1999).

Sequence stratigraphy. At the base of successions of this facies association, lagoonal and/or bay-fill mudstones (facies BF1) representing moderately shallow water (2-14

m) are typically juxtaposed directly above coastal plain coals or upper shoreface sandstones (Fig. 6.4). The bases of these successions are therefore interpreted as transgressive surfaces (Embry, 1995) which separate regressive strata from subsequent transgressive strata. Alternatively, this succession may simply reflect continued progradation of the shoreline. Bay-fill successions are commonly overlain by lower shoreface and upper shoreface sandstones, across a marine flooding surface (*sensu* Van Wagoner *et al.*, 1988; e.g. at 21 m, Fig. 6.4). In places, this flooding surface is lined by a coarse-grained or bioclastic lag deposit which is interpreted to record wave erosion during transgression (wave ravinement *sensu* Swift, 1968). Thus, the bay-fill successions are interpreted to record transgressive deposition at the same scale as the upward-shallowing shoreface successions (parasequences *sensu* Van Wagoner *et al.*, 1988, 1990) that they overlie and underlie. Where bay-fill successions are stacked vertically and interfinger with coastal plain coals (e.g. 9-21 m in Fig. 6.4), it is inferred that transgression occurred via a series of “steps”, during each of which sediment supply was sufficient to infill available accommodation and allow development of a coal bed (Swift *et al.*, 1991b). Where transgressive or flooding surfaces are underlain by tidal inlet channel-fill deposits (facies BF5) or sharply-based, bioclastic rich deposits (facies BF4), the basal erosion surfaces of these units are interpreted to record tidal erosion during transgression (tidal ravinement, *sensu* Swift, 1968).

6.4.2 Shoreface facies association

Successions of the shoreface facies association are well developed throughout the study area. Several shoreface units, each with a similar internal facies succession, are recognised in sections through the upper Almond Formation, typically

intercalated with sediments of the bay-fill and coastal plain facies associations. Deposits of the shoreface facies association weather to form distinctive, low-relief, silver-grey cliffs that can clearly be identified from aerial photographs.

Description. Units of this facies association are typically composed of upward-coarsening successions that become sandier upwards. Complete successions comprise, from base to top, facies SF1, SF2 and SF3 (Fig. 6.4), although recognition of SF3 can be difficult due to significant bioturbation by roots below overlying coastal plain deposits. Sandstone beds thicken and amalgamate upwards within a succession, passing from moderately to highly bioturbated, hummocky cross-stratified, fine-grained sandstones with siltstone interbeds (facies SF1; Fig. 6.6a) gradationally upwards into moderately to well sorted, trough and swaley cross-stratified, fine- to medium-grained sandstones (facies SF2; Fig. 6.6b) and then planar-parallel bedded, fine- to medium-grained sandstones (facies SF3). Carbonaceous laminae and mudstone-drapes are common throughout facies SF1 and SF2 (Fig. 6.6b). Bioturbation decreases in intensity and diversity from the base to top of the successions, allowing increasing preservation of cross-stratification. Bioturbation in facies SF1 is characterised by moderate-to-high intensity and a high diversity of trace fossils. A monospecific assemblage of *Ophiomorpha* characterises beds within facies SF2 (Fig. 6.6c), whilst rooting commonly disrupts bedding within facies SF3 (Fig. 6.6d). Facies SF2 and SF3 are locally eroded and truncated by tidal-inlet channels (facies BF5). Cross-bed palaeocurrents indicate sediment transport predominantly towards the northeast and southeast (i.e. shoreline-parallel), although sediment transport towards the west (i.e. onshore) is also observed.

Figure 6.6 Photograph examples of the Shoreface Facies Association.
 (a) Facies SF1 - Lower Shoreface (BFD MS2 at 15m; Figs. 6.3b, 6.8). Hummocky cross- stratification is common throughout. (b) Facies SF2 - Upper Shoreface (BFD MS1 at 22m; Figs. 6.3b, 6.8). Trough cross stratification is often highlighted by mud drapes. (c) Facies SF2 - Upper Shoreface (DMW MS1 at 18 m (Figs. 6.3c, 6.11). Levels of bioturbation can be high and typically monospecific (*Ophiomorpha*), highlighted by staining. (d) Facies SF3 - Foreshore (BFD MS6 at 37m; Figs. 6.3b, 6.8). Planar parallel laminated sandstones are disrupted by rooting.

Interpretation. The coarsening-upward trend in grain size and upward decrease in bioturbation intensity within the successions implies an upward increase in energy due to shallowing. The high intensity and diversity of bioturbation within the siltstones of SF1 is typical of the *Cruziana* Ichnofacies, indicating a low-energy and low-stress marine environment of deposition (MacEachern and Pemberton, 1992; MacEachern and Bann, 2008), whilst hummocky cross-stratified sandstone beds in this facies record deposition by intermittent storm events (e.g. Dott and Bougeois, 1982; Duke *et al.*, 1991). Facies SF1 is therefore interpreted to record deposition between storm and fair-weather wave bases, in the lower shoreface (e.g. Walker and Plint, 1992). Trough cross-bedding in facies SF2 records migration of dunes in response to unidirectional currents, predominantly due to longshore drift attributed to oblique wave approach (e.g. Clifton *et al.*, 1971; Greenwood and Mittler, 1985). Swaley cross-stratification is interpreted to record storm-wave deposition, and results from poor preservation of hummocky bedforms and from unstable dune deposition (e.g. Dott and Bougeois, 1982; Dumas *et al.*, 2005). Monospecific bioturbation by resilient *Ophiomorpha* burrows is typical of the *Skolithos* ichnofacies, and is indicative of a loose substrate or higher energy setting such as middle or upper shoreface (MacEachern and Pemberton, 1992; Martin, 1993). Preserved mudstone drapes throughout facies SF2 may imply tidal influence within the succession (e.g. Nio and Yang, 1991). Tidal influence has been attributed to wave suppression due to the embayed nature of the study area during Almond Formation deposition (Jackson and Rawn-Schatzinger, 1993; Schatzinger and Tomutsa, 1999; Martinsen, 2003). The root-penetrated, planar-parallel bedded sandstones of SF3 are typical of a foreshore setting (Clifton, 1969; Walker and Plint, 1992). This facies association is interpreted to have been deposited within a

prograding, wave- and storm-dominated shoreface with variable tidal influence, as part of a strandplain, barrier island or spit shoreline.

Sequence stratigraphy. At the base of successions of this facies association, lower shoreface deposits (facies SF1) representing marine deposition below fair-weather wave base are typically juxtaposed directly above bay-fill or coastal plain successions. The bases of these successions are therefore interpreted as marine flooding surfaces (*sensu* Van Wagoner *et al.*, 1988; e.g. at 21m in Fig. 6.4). Occasionally these flooding surfaces occur coincident with coarse-grained, locally carbonaceous sandstones interpreted as transgressive lag deposits. In these cases, this surface may be interpreted as a wave ravinement surface (*sensu* Swift, 1968), which implies wave erosion of the uppermost part of the underlying succession during shoreface retreat.

6.4.3 Coastal Plain facies association

Thin (30-120 cm) intervals of this facies association occur throughout the study area, although the thickest (up to *ca.* 15m) and best developed examples are observed within the lower part of the studied interval (e.g. within Salt Wells Creek, Fig. 6.3b). Coastal plain intervals typically cap progradational successions of the shoreface and bay-fill facies associations, or interfinger with successions of the bay-fill facies association (Fig. 6.4, discussed above). Successions of the coastal plain facies association are largely fine-grained, and thus poorly exposed.

Description. This facies association predominantly comprises coal (facies CP1; Fig. 6.7a) with associated heterolithic deposits of interbedded mudstone, siltstone and

Figure 6.7 Photograph examples of the Coastal Plain Facies Association. (a) Facies CP1 - Coal (BFD MS1 at 3m; Figs. 6.3b, 6.8). Sand has been downpiped into the coal from the overlying lower shoreface deposits. The top of this unit is a flooding surface. (b) Facies CP2 - Coastal plain fines (BFD MS11 at 15m; Figs. 6.3b, 6.8). Black mottling throughout indicative of rooting. (c) Facies CP3 - Fluvial channel fill (SWC MS3 at 0m; Figs. 6.3b, 6.8). Sharply based sandstone incising poorly exposed fines which may be bay-fill or coastal plain units. Palaeocurrent data indicates unidirectional currents (n = 35).

very fine-grained sandstone (facies CP2; Fig. 6.7b). Poor exposure prevents determination of the geometry of these sandstone beds, and their spatial relationship with erosively based, channelized sandstones (facies CP3). Preserved root traces are common throughout both facies, disrupting lamination below coal beds. Wood fragments occur within coal beds, and angular carbonaceous fragments from the coal beds are reworked into overlying sediments. Erosively based, fining-upward units of medium- to fine-grained sandstone (1.3-7 m) with trough to planar cross-beds are locally developed which sharply overlie and pinch out laterally (over 30-65m) into poorly exposed finer-grained facies (facies CP3; Fig. 6.7c). Carbonaceous drapes and mudstone clasts are rare and line cross-sets whilst bioturbation levels are low and trace fossil assemblages are monospecific (*Planolites*). Cross-bed palaeocurrent measurements indicate sediment transport predominantly towards the north and east (offshore-directed), although rare westerly (onshore) directed indicators are observed. Poor exposure of the finer grained units within which these sandstones are developed make it difficult to determine whether they are coastal plain (i.e. CP2) or bay-fill (i.e. BF1 and 2) deposits.

Interpretation. The context of coals (facies CP1) in the Almond Formation suggests a coastal plain environment of deposition. Roots beneath the coal beds indicate that they were developed *in situ*. The interbedding of sandstones and siltstones with mudstones in facies CP2 suggests regularly fluctuating flow velocities in an environment that was at least partly subaqueous, most likely related to variations in discharge and sediment load in adjacent fluvial channel-fill sandstones. However, due to poor exposure, the relationship of these beds with channel-fill sandstones can not be directly observed. Cross-bedding within the erosively based, channelised sandstones (facies CP3) indicates deposition by tractional currents whilst rare carbonaceous drapes indicate fluctuations in flow velocity, which can occur due to

water-stage fluctuations within fluvial systems (e.g. Pickard and High, 1973, Jackson, 1981; Jordan and Pryor, 1992) or can be evidence for tidal influence (e.g. Moslow and Tye, 1985; Israel *et al.*, 1987). The lack of clearer marginal marine and tidal indicators within these sandstones, such as diverse trace and body fossil assemblages or abundant mudstone-drapes displaying rhythmicity, coupled with a predominantly basinward palaeocurrent direction favours the interpretation of these units as fluvial channel fills. However, the rarely observed mud drapes and landward directed (flood tidal) palaeocurrents could be evidence of limited tidal influence in some of these units.

Sequence stratigraphy. Thin intervals of the coastal plain facies association that cap upward-coarsening shoreface successions are interpreted to record subaerial deposition in response to shoreline progradation, and thus form the upper part of shallow-marine parasequences (*sensu* Van Wagoner *et al.*, 1988, 1990). Where these coastal plain intervals interfinger with bay-fill successions (e.g. at 9 and 15 m in Fig. 6.4), they are interpreted to record infilling of accommodation within restricted marine embayments. These deposits are interpreted to pass seawards into coincident marine sediments. Where the coastal plain deposits are abruptly overlain by shoreface deposits, their upper surfaces are interpreted as marine flooding surfaces (*sensu* Van Wagoner *et al.*, 1988) that record increases in water depth. Alternatively, when the coastal plain successions occur within bay-fill successions, their upper surfaces record deepening of a restricted marine embayment; such surfaces are interpreted as transgressive surfaces developed at relatively landward locations during an overall “stepped” transgression that may culminate in marine flooding (*sensu* Swift *et al.*, 1991). Coal fragments reworked from the coastal plain intervals

into overlying deposits imply an element of erosion during transgression (i.e. ravinement).

6.5 Facies architecture and sequence stratigraphic framework

The facies associations described above and their bounding sequence stratigraphic surfaces have been mapped to define aspects of facies architecture and correlated to construct a sequence stratigraphic framework for the upper part of the Almond Formation in two parts of the study area; Big Flat Draw (Fig. 6.3b) and Point of Rocks (Fig. 6.3c). Correlation between the measured sections in each of these areas has been constrained using photomontages, aerial photographs and observations in the field. The two areas are separated by an area of poor exposure preventing reliable correlation between them. However, previous workers have suggested that the succession exposed within the Point of Rocks study area lies stratigraphically above that exposed at Big Flat Draw (Roehler, 1988), implying that correlation between the two areas is not possible. The sequence stratigraphic framework and associated facies architectures are illustrated below using a correlation panel through each of the study areas.

6.5.1 Big Flat Draw correlation

This correlation traverses the study area from Dry Canyon in the south to Black Butte Creek in the north and extends for more than 27 km along the outcrop belt (Figs. 6.3b, 6.8). During deposition of the upper part of the Almond Formation, the shoreline is interpreted to have trended broadly north-south (Roehler, 1988), such that the outcrop belt is oriented approximately along depositional strike (Figs. 6.3).

Interpretation of four marine flooding surfaces and corresponding suites of related surfaces formed during transgression (tidal ravinement surfaces, wave ravinement surfaces and transgressive surfaces) has enabled the logged section to be subdivided into five parasequences and their marginal marine equivalents. The lowermost parasequence (parasequence 1) has only been logged locally within Salt Wells Creek (MS 1-4) and the central part of Big Flat Draw (MS 5, 12-14; Fig. 6.8). At Salt Wells Creek, this parasequence comprises a bay-fill succession that overlies an undifferentiated interval of coastal-plain strata (deposited below parasequence 1) containing fluvial and tidally influenced fluvial channel-fill sandstones. Towards the north, this parasequence comprises a progradational shoreface succession, capped by coastal plain deposits and a thin (7 m) succession of bay-fill deposits (e.g. BFD, MS 12). The transition from stacked bay-fill successions in the south into a shoreface and bay-fill succession to the north may be explained by either the obliquity of the correlation panel to the interpreted shoreline trend, or by a locally non-linear shoreline (Fig. 6.9a). The Salt Wells Creek exposure is located where the outcrop belt lies furthest to the west, and may be interpreted to represent a depositional setting that is proximal (palaeo-landward) of that observed within Big Flat Draw.

Parasequence 2 is exposed throughout the correlation panel (Fig. 6.8). Towards the south (SWC MS3, 4) and north (BFD MS 10-15) of the panel, a progradational shoreface succession capped by coastal plain deposits is observed. This succession pinches out into a bay-fill succession in the central part of the panel (SWC MS 1, 2). These relationships indicate that the wave-dominated shoreface was not linear (Fig. 6.9b), and that the updip (landward) pinchout of the shoreface sandstones occurs

Figure 6.9 Facies belt distribution and extent. Drawn using facies observed overlying key stratigraphic surfaces along the Dry Canyon to Big Flat Draw correlation (Fig. 6.8). (a) Base of parasequence 1; (b) Base of parasequence 2; (c) Base of parasequence 3; (d) Base of parasequence 4; (e) Base of parasequence 5.

over a short (<400 m) distance. The upper part of the shoreface succession is locally eroded by tidal-inlet channels that exhibit up to 5 m of basal erosional relief and extend for over 200 m along the outcrop (e.g. BFD MS 6; Figs. 6.8, 6.10a). A coal caps the shoreface succession and tidal-inlet deposits, suggesting that it developed after tidal ravinement associated with the tidal inlets. The coal is capped by a transgressive surface and overlain by bay-fill deposits that occur as a single succession to the north (e.g. BFD MS 3, 5) and two successions separated by a coal bed and interpreted transgressive surface to the south (e.g. BFD MS 8,6). Within the lowermost of these two latter successions, a laterally discontinuous bay-fill sandstone (facies BF3) can be traced from SWC MS1 in the south to BFD MS15 in the north over a distance of 5.8 km (Fig. 6.8). Within the upper of these two latter successions, a shell-rich bay-fill sandstone (facies BF 4) that pinches out laterally into bay-fill fines (facies BF1, BF2) can be traced from SWC MS4 in the south to BFD MS9 in the north over a distance in excess of 7 km. Within the bay-fill succession to the north of Big Flat Draw, locally developed bay-fill sandstones can not be traced over distances in excess of 2 km (e.g. BFD MS 5). Bay-fill deposits are absent from the upper part of parasequence 2 in the north-central part of the panel (BFD MS 13, 14), implying that this location was the most distal.

Within parasequence 3, shoreface sandstones are developed across the correlation panel (Fig. 6.8) however palaeocurrent data within these sandstones indicates both a NE and SW direction of sediment transport implying a non-linear or embayed coastline (6.9c). These sandstones are capped by a coal bed, which is overlain by a laterally discontinuous bay-fill succession that contains a bioclastic rich unit which

(a)

(b)

(c)

Figure 6.10 Examples of stratigraphic relationships. (a) Tidal inlet channel (facies BF5) incising upper shoreface sandstones and overlain by bay-fill facies (BFD MS6; person for scale). (b) Bioclastic rich sandstone (facies BF4) erosionally truncates bay-fill fines and pinches out laterally into them (DC MS1; person for scale). (c) Lower shoreface sandstones sharply overlie coastal plain coals (DC MS3; camera case for scale, ca. 10 cm).

can be correlated laterally for more than 6 km to the south of the correlation panel (Figs. 6.8, 6.10b; DC MS2-SWC MS2; facies BF4).

Parasequence 4 is thickest and best developed in the south of the study area (DC MS1-SWC MS2; Fig. 6.8). In Dry Canyon, a prograding shoreface succession overlies the coastal plain capping parasequence 3 (Fig. 6.9c) and is itself capped by coastal plain coal and a transgressive surface that is overlain by bay-fill fines truncated by laterally discontinuous (<200 m) tidal-inlet channels (facies BF5). To the north this shoreface pinches out into a bay-fill succession (Fig. 6.9d) within which discrete sandbodies (facies BF3) are locally developed (SWC MS2-BFD MS15). This relationship is interpreted to reflect a distal (DC MS1-3) to proximal (SWC MS2-BFD MS15) trend and may be explained by a simple, linear shoreline trending NE-SW (Fig. 6.9d). In the north-central part of the panel, this parasequence pinches out and lower shoreface sandstones of parasequence 5 are juxtaposed directly above coastal plain facies capping the upper shoreface sandstones of parasequence 3 (BFD MS 13, 14). This parasequence may either be represented within the coastal plain succession that caps parasequence 3, or it may have been completely removed by localised wave ravinement during transgression associated with the base of parasequence 5. The bay-fill succession developed at this level in the northern part of the panel (BFD MS1-12) most likely represents parasequence 4, but it may also be interpreted to form the upper part of parasequence 3.

Parasequence 5 records deposition within a progradational shoreface setting that can be correlated across the panel (Fig. 6.8; Fig. 6.9e). Locally a bay-fill succession overlies the shoreface succession, but poor exposure prevents interpretation of the

depositional extent of this unit. Tidal inlet channel-fill sandstones are locally interpreted to erode into bay-fill fines (e.g. BFD MS12) within this succession, but these sandstones can not be traced for significant distances along strike. Locally, a bioclastic rich sandstone that can be traced for 7.5 km (facies BF 4) caps the bay fill succession (SWC MS 2-BFD MS9), marking the top of the Almond Formation deposition here. This sandstone is overlain by the Lewis Shale.

The large-scale stratigraphic architecture in the Big Flat Draw area, as recorded by parasequence stacking patterns, is strongly aggradational with only a small retrogradational component. The shoreline trajectory recorded by this stacking pattern is accretionary transgressive. The development of relatively thick (up to 15 m) and, in some instances, stacked bay-fill successions in the upper part of individual parasequences indicates that accretionary transgressive shoreline trajectories were also developed at this smaller scale.

6.5.2 Point of Rocks correlation

Situated to the north of the anticline and stratigraphically above the Big Flat Draw correlation, this panel extends for approximately 17 km along the outcrop belt (Fig. 6.3c). At its southern end, the correlation is oriented approximately W-E, parallel to depositional dip, for 2.5 km, although the main part of the panel is oriented NW-SE and trends oblique to depositional dip (approximately NNE-SSW; Roehler, 1988). Outcrop exposure in this study area is predominantly limited to shallow creeks separated by 1-2 km wide regions of scrub, which means that stratigraphic units can not be mapped but are instead correlated over relatively short distances. The

relationships interpreted in this panel are therefore of lower confidence than those mapped and correlated in the Big Flat Draw area.

Flooding surfaces divide the succession into nine parasequences (Fig. 6.11). Due to limited exposure, parasequences 1-5 can only be described from the northwestern and southeastern limits of the area, but parasequences 3, 4 and 5 can be correlated across the 17 km extent of the correlation panel (Fig. 6.11). Parasequences 6-9 are more widely exposed and also extend across the entire panel. Within each parasequence, a progradational shoreface succession typically passes upwards into coastal plain and/or bay-fill deposits. In parasequence 3, bay-fill deposits occur at both the northern and southern ends of the panel, implying the development of a bay or lagoon of large extent along regional depositional dip (>13 km; Fig. 6.11). In parasequences 4 to 7, there are pronounced proximal-to-distal changes in the facies composition of the parasequences; bay-fill and upper shoreface deposits at the Road to Superior section pass down dip into lower shoreface deposits in the Point of Rocks area. However, hummocky cross-stratified sandstone beds in the lower shoreface deposits in the Point of Rocks area amalgamate and thicken upwards to define a progradational trend within each interpreted parasequence. Correlation of units within parasequence 5 implies a stepped transgressive model (*sensu*. Cattaneo and Steel, 2003), with the transgressive surfaces at the base of the bay-fill sediments to the north (RtS MS1) interpreted to be truncated by a wave ravinement surface (Fig. 6.11). Parasequence 8 appears to show the opposite proximal-to-distal trend, with bay-fill deposits being thicker towards the SE. This trend implies either a local reorganisation of the shoreline palaeogeography, or variable amounts of wave ravinement and thus preservation of bay-fill deposits at the top of the parasequence.

Figure 6.11 Point of Rocks to Road to Superior correlation. Trending SE-NW, this correlation is oblique to regional depositional strike and becomes more proximal towards the northwest. Accessible outcrop is limited to narrow creeks separated by areas of non-exposure which makes correlation difficult. Less confident interpretations are indicated by dashed lines and an opaque mask. The location of this correlation is marked on figure 6.3c.

Parasequence stacking patterns in the Point of Rocks area are aggradational, and the shoreline trajectory recorded by this stacking pattern is essentially vertical. The development of relatively thick (up to 20 m) and stacked bay-fill successions in the upper part of individual parasequences indicates that accretionary transgressive shoreline trajectories were developed during parasequence-scale transgressions.

6.6 Discussion

6.6.1 Sandbody geometry and distribution

Three facies associations are recognised within the study area: (i) coastal plain fines containing channel-fill sandstones; (ii) bay-fill heteroliths that contain both channel-fill sandstones and gradationally based, laterally discontinuous sandstones; and (iii) shoreface sandstones. Shoreface sandstones are highly continuous along depositional strike (at least 19 km), but pinch out up-dip into bay-fill units over short distances (400 m). These sandstones are also eroded into by tidal inlet channel-fill sandstones up to 200 m wide and 2-7 m thick. Sandstones within the bay-fill and coastal plain facies associations are more discontinuous along strike, with channelised sandbodies typically pinching out over less than 200 m and non-channelised sandbodies up to 2.5 m thick extending for up to 7.5 km. Many of these sandbodies pinch out into mudstones and siltstones, where they appear to be unconnected from each other in 2D cliff-face panels (Figs. 6.8, 6.11). The bay-fill successions pinch out along both cliff-face correlation panels, which reflects a combination of (1) thinning down depositional dip, (2) subtle changes in shoreline orientation between parasequences, (3) locally variable depth of wave ravinement at

the base of overlying shoreface sandstones, and (4) lateral correlation into a thick coal bed that underwent a large volume decrease during compaction. In summary, the net-transgressive strata of the upper Almond Formation contain a range of sandbody types of variable geometries in both depositional strike and dip directions, even in the absence of a syn-depositional structural control and this makes correlation extremely difficult, even in relatively well exposed areas.

6.6.2 Model of stratigraphic architecture in the upper Almond Formation

The shoreface sandstones are interpreted to have been deposited within prograding strandplains and are commonly capped by coastal plain deposits. These coastal plain deposits are commonly overlain by bay-fill deposits, interpreted to have been deposited within a back-barrier bay environment. These shoreface-to-bay-fill successions are interpreted to represent deposition during one sea-level cycle or parasequence, with shoreface sandstones representing predominantly the regressive part of the cycle whilst bay-fill sediments were deposited during the subsequent transgression that saw the previous strandplain inundated with marine water (e.g. Hendricks, 1994; Zavala *et al.*, 2001; Horn *et al.*, 2002). This facies repetition can also reflect autogenic shifting. Where several bay-fill successions are stacked across coal beds in the upper part of a parasequence, cycles of infilling and flooding of the bay are interpreted during a single episode of transgression, thus implying “stepped shoreline retreat” (cf. the conceptual models of Swift *et al.*, 1991b).

Previous workers have documented that the regressive portion of the cycle is not always present within parts of the upper Almond Formation in the subsurface

Washakie Sub-basin and Wamsutter Arch, depending on the rates of transgression, sediment supply and position of palaeohighs during deposition (Hendricks, 1994; Horn *et al.*, 2002). Palaeohighs formed focal points for transgressive erosion (Hendricks, 1994), whilst transgressive deposition was restricted to periods of high accommodation relative to sediment supply (Horn *et al.*, 2002). Alternatively, the regressive portion of the cycle may not be expected in the more proximal or landward part of the depositional setting with shoreface sandstones pinching out up-dip into bay-fill deposits. The contact between regressive and transgressive packages is typically abrupt (i.e. below well spacing in the subsurface) and can be erosive, particularly across syn-depositional topographic highs (Hendricks, 1994; Zavala *et al.*, 2001; Horn *et al.*, 2002).

6.6.3 Comparison to other net-transgressive deposits within the Western Interior Seaway

The “coupling” of regressive shoreface deposits and tidally influenced transgressive deposits is commonly recorded within the deposits of the Cretaceous Western Interior Seaway (e.g. Devine, 1991; Olsen *et al.*, 1999; Sixsmith *et al.*, 2008). Preservation of both the transgressive and regressive elements of the depositional couplet is however variable.

The facies observed within the transgressive deposits also show variability throughout the seaway. In the upper part of the Almond Formation, the transgressive bay-fill deposits are mudstone-rich successions with subordinate, laterally discontinuous sandstones. This is in marked contrast to the transgressive sediments of the Point Lookout Sandstone and Hosta Sandstone in which shoreline-parallel

belts of tidal channel-fill deposits dominate the bay-fill succession (Devine, 1991; Sixsmith *et al.*, 2008). Two factors are interpreted to control the type of backbarrier deposits; tidal range and substrate (Devine, 1991). In micro-tidal settings, backbarrier lagoons are associated with a relatively small number of tidal inlet channels separated by widely spaced barrier-island segments. Higher tidal ranges are associated with deeper and more closely spaced tidal channels, which allow transport of larger water volumes over a tidal cycle; these channels are likely to more strongly erode lagoonal sediments. Lateral migration of channels is also retarded by cohesive fine-grained bay-fill deposits (Devine 1991). Therefore the variation between the Almond Formation transgressive deposits and those within the Point Lookout Sandstone and Hosta Sandstone may be related to tidal range, with a lower tidal range within the Rock Springs embayment, and/or differences in the cohesive strength of back-barrier deposits.

Within the Cliff House Sandstone, transgressive sediments deposited within topographic lows had the highest preservation potential, including tidally-influenced channels which had eroded into underlying shoreface tongues (Donselaar, 1989; Olsen *et al.*, 1999). The aggradational to transgressive deposits within the upper Almond Formation exhibit a steeper overall shoreline trajectory compared to those in the Cliff House Sandstone, where wave ravinement removed large volumes of transgressive deposits. The shallower shoreline trajectory of the Cliff House Sandstone may account for this more pronounced ravinement.

6.6.4 Comparison to the Hugin Formation

The sedimentology of the upper part of the Almond Formation in the Rock Springs Uplift shows significant similarities to the Hugin Formation within the South Viking Graben. Both formations contain wave-dominated shoreface sandstones, tidally-influenced bay-fill successions and coal-bearing coastal plain intervals. The Hugin Formation contains more direct fluvial influence, particularly in the area of the “Sleipner Delta”, although fluvial channels are present in the lower part of the study interval in the Almond Formation. Published palaeogeographical reconstructions for the upper Almond Formation show significant fluvial-deltaic systems to the north and south of the study area, but limited fluvial input into the Rock Springs Embayment itself (Fig. 6.1c; Roehler, 1988). Parasequence thickness within the Almond Formation is markedly lower (< 30 m) than in parts of the Hugin Formation (up to 57 m). This difference may reflect in part by the different structural settings controlling the development of accommodation for the two formations; the Hugin Formation was deposited within a rift basin, whilst the Almond Formation was deposited in a foreland basin subject to more regionally extensive changes in tectonic subsidence.

Despite the similarities in facies character, the stratigraphic architecture of the Almond Formation differs from those within the Hugin Formation. Within the Almond Formation, bay-fill and shoreface facies associations form depositional couplets within the same vertical succession, similar to that which is interpreted within the uppermost part of the Hugin Formation in the Sigrun-Gudrun area. A similar stacking of regressive-transgressive packages is interpreted at Sleipner Vest (Folkestad and Satur, 2008) although, unlike the Almond Formation, fluvial

influence is observed throughout the shallow-marine succession. At Dagny-Alpha North however, transgressive packages are not interpreted within the succession and stacked successions of regressive shoreface or mouth bar units are developed. The different stacking patterns observed within the Hugin Formation are interpreted to relate to the balance between sediment supply and accommodation space within the graben, as well as the relative interplay between waves and tides. Deposition or preservation of transgressive sediments at Sigrun-Gudrun and Sleipner Vest may indicate that sediment supply was more in balance with slowly developing, rift-related accommodation whilst the lack of transgressive units at Dagny-Alpha North may reflect sediment supply was being outpaced by accommodation space created during halokinesis of the underlying Zechstein layer. Alternatively, the observed shoreface succession may simply reflect a more distal location within the basin with transgressive units preserved, but not penetrated, up-dip. There is little evidence for a similar syn-sedimentary tectonic control in the studied outcrops of the Almond Formation, although such a control has been inferred within reservoir trends in the subsurface Washakie Sub-basin and Wamsutter Arch (Van Horn, 1979; Jackson and Rawn-Schatzinger, 1993; Clawson and Fawret, 1997; Mead and Krystinik, 1997; Krystinik and Mead, 1998; Martinsen, 2003).

CHAPTER 7

Impact of syn-depositional faulting on shallow-marine sedimentation and reservoir distribution: Hugin Formation, Block 15/3, Norwegian South Viking Graben

7.1 Introduction

Syn-rift sedimentation patterns have been shown to be affected by the rate of fault-controlled subsidence, sediment supply and routing, and eustasy (e.g. Leeder and Gawthorpe, 1987; Prosser, 1993; Davies *et al.*, 2000; Bruhn and Vagle, 2005; Khalil and McClay, 2009). Investigating the temporal and spatial evolution of rift-related structures can, therefore, provide an important constraint on predicting sediment distribution and facies variability within syn-rift sedimentary successions.

This study focuses on the shallow-marine Hugin Formation deposited on the hangingwall dipslope of the South Viking Graben, North Sea Basin. This formation was deposited during the Latest Bathonian-Callovian during which the area was undergoing the initial phases of Jurassic rifting. The aim of this study is to investigate the temporal and spatial evolution of rift-related faults on the hangingwall dipslope, and to determine how the development of these structures influenced facies distribution within the earliest syn-rift, Hugin Formation. The Sigrun-Gudrun region of the South Viking Graben has been utilised for this study due to the availability of 3D seismic data, high-resolution biostratigraphy and wells,

the latter being distributed in both the hangingwalls and footwalls of syn-depositional normal fault arrays.

7.2 Geological Setting

The study area is located on the hangingwall dipslope of the South Viking Graben encompassing Norwegian blocks 15/3, 16/1 and the northern part of 15/2, extending across the Gudrun Terrace and into the Vilje Sub-basin (Fig. 7.1). The South Viking Graben is a NNE-SSW trending half-graben, bounded to the west by a series of ESE-dipping faults and the Palaeozoic high of the Fladen Ground Spur (Thomas and Coward, 1996; Husmo *et al.*, 2003). To the east, the graben shallows onto the Utsira High, a long-lived, relatively unfaulted, basement-cored structural high (Thomas and Coward, 1996; Husmo *et al.*, 2003).

Formation of the South Viking Graben dates from Late Permian to Early Triassic rifting (e.g. Coward, 1995; Roberts *et al.*, 1995; Glennie, 1998; Coward *et al.*, 2003). Thermal subsidence led to the formation of the North Permian Salt Basin, along the northern margin of which the South Viking Graben formed a NE-SW-trending, fault-bounded embayment (Ziegler, 1990; Hodgson *et al.*, 1992; Glennie, 1998; Coward *et al.*, 2003). During the Early Jurassic (Toarcian-Aalenian), at the future site of the North Sea Basin 'triple junction', regional thermal uplift occurred due to impingement of a mantle plume on the base of the lithosphere as a precursor to active rifting within the region (e.g. Ziegler, 1990; Underhill and Partington, 1993; Davies *et al.*, 1999; Coward *et al.*, 2003).

Figure 7.1 Geological Setting of the Study Area. (a) Regional section across the South Viking Graben. The study area (highlighted) sits on the hangingwall dipslope of the salt-influenced rift basin (modified Sneider *et al.*, 1995). (b) Higher resolution cross section through study area. This section shows the Gudrun and Brynhild Central fault arrays that developed as antithetics to the main graben bounding fault and dissect the hangingwall dipslope. These faults detach downwards onto the mobile Zechstein salt layer. (c) Location map highlighting the position of these cross sections with respect to the study area.

Activity on the graben-bounding fault zone is interpreted to have initiated in the Early or Late Callovian, with the main phase of extension and basin subsidence occurring during the Oxfordian to Middle Volgian (e.g. Cockings *et al.*, 1992; Milton, 1993; Williams, 1993; Knott, 2001). Movement on the N-S-trending, graben-bounding fault zone and consequent westward rotation of the hanging wall dip-slope led to gravity sliding and faulting within the Triassic to Jurassic succession (Thomas and Coward, 1996; Knott, 2001; Jackson and Larsen, 2009). Within the study area the structural style is dominantly thin skinned, being separated from basement-involved faulting by mobile evaporites of the Upper Permian Zechstein Group.

7.3 Database and methodology

The seismic dataset for this study comprises a 3-D survey covering an area of 770 km², and with an inline (N-S) and crossline (E-W) spacing of 12.5m (Fig. 7.2). The record length is 5.5 seconds two-way travel time (TWTT) and the data has not been depth converted. The studied interval lies between 2.5 and 4.5 seconds TWTT. The Zechstein Group evaporites serve to reduce the seismic quality in the sub-salt (pre-Upper Permian) section, therefore detailed interpretation is restricted to the supra-salt succession where data quality is moderate to good. The seismic data is displayed with a reverse polarity (SEG European convention; Brown, 1999) with a downward decrease in acoustic impedance represented by a trough (black) and a downward increase in acoustic impedance represented by a peak (red).

Five age-constrained stratigraphic surfaces have been identified within the pre- and syn-rift interval of interest and have been interpreted across the study area. These

Figure 7.2 Study Area Database and Fault Map. The study area lies within the Norwegian sector of the South Viking Graben. 3D seismic data mapping has been undertaken over the area highlighted. This was integrated with well data from nine wells within the study area. Core data was logged from seven of these wells, highlighted *. The presented fault map is taken from the top pre-rift (top Sleipner) time structure map.

horizons typically tie to major lithological boundaries or specific time events (i.e. flooding surfaces) within the studied succession. These horizons are relatively continuous, and exhibit moderate to high- seismic reflection amplitudes. These surfaces are used to divide the studied interval into four main units (Fig. 7.3). Unit 1 broadly represents the pre-Jurassic rift interval (Triassic- Bathonian), Units 2 and 3 represent the rift-initiation interval (Latest Bathonian-Mid Callovian Hugin Formation, which forms the focus of this study) and Unit 4 encompasses the rift climax interval (Mid Callovian-Upper Jurassic).

Seven exploration wells within the study area provide a variety of data for this study (summarised in Table 7.1; Fig. 7.2). Over 150 m of core has been logged from five of these wells and core descriptions have been integrated with wireline-log data. These well data have been utilised to develop a facies scheme and sequence stratigraphical framework for the Hugin Formation. Additional data available from four wells outside of the seismic dataset have also been used to constrain well correlations and palaeogeographical reconstructions (Chapter 5; Kieft *et al.*, *in review*). High resolution biostratigraphical analyses have enabled correlation of flooding surfaces which are of local or regional extent.

7.4 Structural Style

A fault array, which consists of four main segmented normal faults related to the Jurassic rift event, deforms the hangingwall dip slope. These normal faults are the Gudrun, Gudrun North, Brynhild and Sigrun Faults (Fig. 7.2). The structural style of these faults and their associated structures is described below. Fault throw analysis

Figure 7.3 Summary diagram of seismic reflection events picked within the study area. Five seismic horizons have been mapped through the study area and used to generate four isochrons encompassing the pre rift, earliest syn-rift and rift climax units. Quality and clarity of the mapped reflection events vary across the study area.

Well	Available data		
	Core	Wireline	Biostratigraphical analyses
15/2-1	X	X	
15/3-1S	X	X	X
15/3-2R	X	X	
15/3-3	X	X	X
15/3-4	X	X	X
15/3-5	X	X	X
15/3-7	X	X	X
16/1-2		X	X
16/1-3		X	X
16/1-5		X	X
16/1-7		X	X

Table 7.1 Summary of available well data in the Sigrun-Gudrun study area

has been undertaken using throw measurements at the Top Sleipner (i.e. pre-rift) level as this documents the cumulative syn-rift deformation and avoids displacement artefacts related to syn-tectonic thickness variations. Throw analysis has been integrated with isochron mapping to analyse the spatial and temporal development of these faults (section 7.5).

7.4.1 Gudrun Fault

The orientation and displacement of the Gudrun Fault has been previously described by Jackson and Larsen (2009). The description given here is slightly different to these authors' published work, based on observations made from seismic mapping during this study. The Gudrun Fault trends NE-SW and is more than 14 km in length. It is planar in cross-section and dips steeply (60° - 70°) to the NW (Figs. 7.2, 7.4a, b). Unlike the other faults within the study area, the hangingwall of the Gudrun Fault is rarely deformed by rift related structures. Two synthetic faults are developed at the northern tip in the footwall to the main fault (segment 4). These faults appear to splay off the main fault, exhibit minor displacement (typically less than 60 ms) and tip out to the NE.

Fault throw analysis allows subdivision of the main fault into at least four major segments based on displacement minima. Segment boundaries may be identified along the fault as kinks or bends; the clearest example occurs at the boundary between segments 3 and 4, where a marked kink is observed and the fault trace bends towards the east (Fig. 7.4c). Maximum displacement of 282 ms (TWT) is observed on segment 3 and displacement decreases towards the tips of the fault (Fig. 7.4c). Each segment displays an asymmetrical throw profile with greatest

Figure 7.4 Structural Style of the Gudrun and Brynhild Fault Zones. (a) Cross section through the southern part of the fault zones. The hangingwall of the Gudrun Fault is comparatively undeformed whilst several anti- and synthetic faults are developed in the hangingwall of the Brynhild Fault which dissect the hangingwall anticline and syncline developed to the west of this fault. (b) Cross section through the northern part of the fault zones. Throw across both the Gudrun and Brynhild faults has declined at this point whilst the anticline and syncline within the hangingwall of the Brynhild Fault exhibit significantly less relief (modified from Jackson and Larsen, 2009). See Figure 7.2 for location of lines.

Figure 7.5 (cont.) Structural style of the Gudrun and Brynhild Fault Arrays; Along-Strike Fault Throw profiles for the Top Sleipner Horizon (Top Pre-rift). (c) Gudrun fault array. The main fault can be subdivided into four segments on the basis of throw minima and kinks in the fault trace. (d) Brynhild Central, South and East fault arrays. Brynhild Central can be subdivided into five segments on the basis of throw minima and kinks in the fault trace. Throw data suggests Brynhild East (Purple) and Brynhild South (Green) can also be subdivided, each into at least two segments although these displacement minima do not correspond as clearly to kinks in the fault trace.

displacement occurring to the SW of individual segments. The fault tips out downwards into the upper part of the Zechstein Group and upwards within the Upper Jurassic Unit 4 (Heather Formation).

A fault-parallel anticline is observed in the hangingwall of the fault with magnitude diminishing northwards and southwards along the fault trace (Thomas and Coward, 1996; Jackson and Larsen, 2008, 2009). Latest Jurassic and earliest Cretaceous onlap patterns indicate growth of the structure at this time and therefore the anticline is interpreted to have formed as a result of multi-phase compression during the latest Jurassic-Early Cretaceous inversion event.

7.4.2 Brynhild Fault

Like the Gudrun Fault, the structural style of the Brynhild Fault has previously been described by Jackson and Larsen (2009). Likewise, the description given here is slightly different to these authors' published work, based on observations made from seismic mapping during this study. Importantly, due to additional data being available for this study, the previously published subdivision of the Brynhild Fault into two segments, Brynhild North and South, has been modified. An additional fault has been mapped in the north of the study area and this has been called the Brynhild North fault segment; therefore, the Brynhild North fault segment of Jackson and Larsen (2009) is herein termed the Brynhild Central fault segment. The segmented Brynhild Fault lies 5 km to the east of the Gudrun Fault, further up the hanging-wall dip slope. The fault is divided into four main segments; Brynhild North, Brynhild Central, Brynhild South and Brynhild East (Fig. 7.2). All four segments of the Brynhild Fault are listric in cross-section, shallowing and detaching downwards into

the upper part of the Zechstein Group and steepening upwards into and tipping-out within Unit 4 (Upper Jurassic) (Figs. 7.4a,b; 7.5a,b).

The Brynhild Central segment is 15 km long and trends NNE-SSW, dipping towards the WNW. To the north it is separated from the Brynhild North segment by an N-S trending relay ramp approximately 500 m wide (Fig. 7.2). To the south it is separated from the Brynhild South segment by an ENE-WSW-trending, NNW-dipping, 2 km wide relay ramp (Fig. 7.2). Brynhild Central can be further subdivided into five segments based on the recognition of displacement minima and/or bends in the along-strike trace of the fault (Fig. 7.4d). Maximum displacement of 484 ms (TWT) is observed on segment 4 towards the northern end of the fault (Fig. 7.4d). Displacement decreases to the north and south towards the fault tips but does not exhibit the marked asymmetry associated with the Gudrun Fault (Fig. 7.4c, d). An anticline and syncline are developed within the hangingwall of the fault, both of which trend parallel to the local strike of the adjacent fault segments. These features are symmetrical with gently dipping limbs and die-out along strike, corresponding to an along strike decrease in displacement on the component segments of the fault (Fig. 7.4b).

The Brynhild South fault segment is almost 6 km long and trends broadly N-S at its southern end before swinging round to trend NE-SW at its northern end (Fig. 7.2). On the basis of displacement minima and the along-strike, map-view geometry of the fault trace, this segment has been further subdivided into 2 segments (Fig 7.4d). Maximum displacement (220 ms TWT) is recorded at the northern end of the fault on segment 2 (Fig. 7.4d).

Figure 7.5 Structural Style of the Gudrun North and Brynhild North Fault Zones. (a) Cross section through the southern part of the fault zones. The main Gudrun North Fault has 3 antithetic faults developed in the hangingwall, whilst the main Brynhild North Fault has 5 antithetic faults developed. (b) Cross section through the northern part of the fault zones. Several large synthetic faults are developed in the hangingwall of the Gudrun North Fault. Fewer antithetic faults are seen developed in the hangingwall of the Brynhild North Fault however the hangingwall anticline has broadened and deepened. For key to abbreviations used see Figure 7.4.

Figure 7.5 (cont.) Structural style of the Brynhild North and Gudrun North Fault Zones; Along-Strike Fault Throw profiles for the Top Sleipner Horizon (Top Pre-rift). (c) Brynhild North fault zone. The main fault can be subdivided into five segments on the basis of throw minima and kinks in the fault trace. (d) Gudrun North fault zone. The main fault can be subdivided into six segments on the basis of throw minima and kinks in the fault trace.

Brynhild East is also approximately 6 km long and lies to the east of the Brynhild South segment, further up the hangingwall dip slope (Fig. 7.2). The fault dips to the west and has been further subdivided into two segments (Fig. 7.4d). The southernmost part of the fault trends N-S, before swinging to a NNE-SSW orientation. Maximum displacement along the two segments is comparable, with a maximum of 170 ms (TWT) recorded on both (Fig. 7.4d). The Upper Jurassic interval appears to be incomplete in the immediate footwall of the fault and there appears to be truncation of reflection events within the Upper Jurassic interval by the BCU (Fig. 7.4a). This apparent stratigraphic incompleteness may, however, simply be a function of poor resolution due to the overall thinning of the Upper Jurassic section beneath the Base Cretaceous Unconformity in this area.

The Brynhild North segment measures 22.5 km in length and extends beyond the northern limit of the seismic dataset. Within the studied area the fault comprises five segments and maximum displacement (550 ms TWT) is observed at the northern end of the fault on segment 4 (Fig. 7.5c). The fault trends NNW-SSE in the north (segment 5), is oriented N-S in the south (segments 2 through 4) and the southernmost tip (segment 1) trends ENE-WSW (Fig. 7.2). An anticline and syncline are developed within the hangingwall of this fault, both of which trend parallel to the local strike of the adjacent fault segments (Fig. 7.5a, b). These folds are asymmetrical, both exhibiting a shorter and steeper western limb. They are more pronounced at the northern end of the fault and flatten off towards the south, where the fault changes orientation and intersects with the Brynhild Central fault. The crest of the anticline is dissected by a series of antithetic faults that developed parallel to

segments 2 through 5 of the main fault (Fig. 7.5a, b). These faults are typically planar in cross section, tip out downwards within the Zechstein layer and upwards within Unit 4 (Upper Jurassic; Fig. 7.5a, b).

Unlike the Gudrun Fault, the hangingwall of the Brynhild Fault is deformed by a series of antithetic (SE-dipping) and synthetic (NW-dipping) faults that trend broadly parallel to the main fault trace (Fig. 7.2). Displacement on these faults is low, typically less than 100 ms (TWTT) (Figs. 7.4d; 7.5d). A displacement variation on these structures allows different segments to be identified. The majority of these faults are concentrated towards the northern end of Brynhild Central, adjacent to segments 2, 3, 4 and 5 (Fig. 7.2; 7.4d). These smaller faults are typically more planar in cross section than the main faults, although the former are also observed to tip out downwards within the Zechstein Group or to link with the main fault trace (Figs. 7.4a; 7.5a,b). Upwards, these smaller faults typically tip out within the Upper Jurassic Unit 4 (Figs. 7.4a; 7.5a, b).

7.4.3 Gudrun North Fault

The Gudrun North Fault is located to the NW of the Gudrun Fault, 9 km along-strike towards the main graben-bounding fault array (Fig. 7.2). The fault is almost 7 km long, trends NE-SW and dips towards the NW (Fig. 7.2). The entire structure plunges towards the SW/WSW as a result of rotation on the hangingwall dip slope. The fault displays a listric to planar geometry in cross section, tipping out within the underlying Zechstein Group, which is locally uplifted beneath this fault array (Fig. 7.5a, b). To the NE the fault tips out upwards within Unit 4 (Upper Jurassic) whilst to the SW, the top of Unit 3 (Top Hugin reflection event) is flexed rather than offset

across the fault (Fig. 7.5a, b). Fault analysis highlights that the fault comprises 6 segments and that maximum fault displacement (108 ms TWT) occurs on segment 6 at the NE end of the fault (Fig. 7.5c). Maximum recorded displacement within each of the fault segments decreases markedly along the fault system from NE to SW (Fig. 7.5c).

Several antithetic and synthetic faults which trend NE-SW are interpreted in the immediate hanging-wall of the Gudrun North fault (Fig. 7.2). These are typically more planar in cross section than the main fault and tip out downwards within Unit 1 or link with the main fault (Fig. 7.5a, b). The majority of these faults tip out upwards within the lower part of Unit 4 (Upper Jurassic Heather Formation), although some tip out within Unit 3 (Upper Hugin Formation). To the north of the main fault two additional WNW-ESE-trending faults are developed synthetic to the main fault (Fig. 7.2). These faults are listric in geometry, detaching on the underlying Zechstein Group, tip out upwards within Unit 4 (Upper Jurassic) (Fig. 7.5b) and extend northwards beyond the limits of the dataset. Unlike the Gudrun North fault, these two faults offset the intra-unit 4 reflection event (Top Heather Formation) implying that they continued to be active after the main Gudrun North Fault had ceased to develop (Fig. 7.5b). Present-day, a fault-parallel anticline is observed in the hangingwall of the Gudrun North fault (Fig. 7.5a). Relief on this structure is observed within the Upper Jurassic and lowermost Cretaceous successions. Unlike the fold associated with the Gudrun Fault, magnitude of the feature diminishes towards the northeast and the tips of the NW-SE trending synthetic faults discussed above with increasing displacement on the main fault. This anticline may have formed as a result of multi-phase compression during the latest Jurassic-Early

Cretaceous inversion event or, more likely, due to continued growth of the underlying salt-cored high.

7.4.4 Sigrun Fault

This fault is approximately 17 km long and trends NE-SW, broadly parallel to the southern segments of the Brynhild Central and South faults (Fig. 7.2). Unlike the other major faults studied within the area, it downthrows to the SE. In cross section it also exhibits listric geometry, detaching on the underlying Zechstein Group whilst upwards tips out at BCU level. Top Sleipner, Intra and Upper Hugin and Top Heather reflection events are all observed to be truncated beneath the BCU within the footwall. As a result, displacement cannot be calculated accurately across this fault because of this erosion on the footwall crest, and fault throw analysis, therefore, has not been undertaken. A thickened Zechstein layer is observed beneath the eroded footwall crest. Towards the north there is little structuring within the hangingwall; further south, however, a synthetic fault is developed to the east of the main fault, further up the hangingwall dipslope. Hangingwall anticlines and synclines, as observed adjacent to the Brynhild Central and North fault segments, are not associated with this fault.

7.5 Temporal and Spatial Evolution of Syn-rift Structures

To interpret syn-rift sedimentation patterns it is important to understand both the structural style and also the temporal and spatial evolution of the rift-related fault systems. Five regionally mapped surfaces (Fig. 7.3) have been used to generate four isochrons across the study area (Fig. 7.6). These isochrons record spatial variation in

sediment thickness (in time) which is used as a proxy for spatial variations in subsidence generated by fault movement (e.g. Young *et al.*, 2001; Childs *et al.*, 2003; Jackson and Larsen, 2009). Although sediment thickness will be controlled by a range of factors, including water depth, sedimentation rate and compaction, the studied strata comprise shallow marine packages commonly capped by coal highlighting that accommodation space was more-or-less completely infilled during deposition. Biostratigraphical analysis also highlights that time intervals are represented on both the hanging- and footwall of active faults. Thickness variations indicated by the isochron mapping have been interrogated on 2D seismic lines to enable more confidence in their interpretation.

7.5.1 Unit 1 – Top Permian to Bathonian

Description. This unit comprises the Triassic to Late Bathonian interval (i.e. Skagerrak, Smith Bank and Sleipner formations) and is therefore interpreted to pre-date the Mesozoic phase of rifting within the study area. Despite this, the unit ranges in thickness across the study area, with the greatest thicknesses (1070 ms TWT) observed to the northwest within a NE-SW trending oval depocentre (labelled A; Fig. 7.6a). To the north of this depocentre, the unit thins across and reflection events onlap an ovate high cored by Zechstein Group evaporites (labelled B; Fig. 7.6a); the northern margin of this high lies outside the study area. The unit also thins overall eastwards and south-eastwards from the main central depocentre, although this overall trend is interrupted by a series of irregular and elongate thicks and thins (Fig. 7.6a). The elongate thins lie parallel to, and in the immediate footwalls of, the Gudrun, Brynhild and Brynhild North faults, across low-relief salt pillows (labelled C-E; Fig. 7.6a). Significant thinning of the unit is also observed towards the

Figure 7.6 Structural development of the Sigrun-Gudrun area. (a) Unit 1 - Pre-rift. Isochron map generated between Top Zechstein and Top Sleipner reflection events. (b) Unit 2 - Early syn-rift. Isochron map generated between Top Sleipner and Intra Hugin reflection events. Overall thickening towards the GBFZ to the west is observed along with thinning towards the Utsira High to the east. Localised thickness variations are attributed to differential subsidence caused by halokinesis of underlying Zechstein Group and rift-related fault development.

Figure 7.6 Structural development of the Sigrun-Gudrun area (cont.). (c) Unit 3 - Early syn-rift. Isochron map generated between Intra Hugin and Top Hugin reflection events. (d) Unit 4 - Rift climax. Isochron map generated between Top Hugin and BCU reflection events. The hangingwall fault array propagates (Unit 3) to form through going fault zones (Unit 4).

southwest corner of the study area across another ovate salt-cored high (labelled F; Fig. 7.6a) and reflection events within this unit onlap the high.

Interpretation. Overall thickening of the unit towards the west indicates post-rift thermal subsidence associated with the Permian rift event or pre-Mesozoic movement on the Graben Bounding Fault Zone (GBFZ; Thomas and Coward, 1996). Although rifting within the South Viking Graben is not interpreted to have started until the Jurassic (e.g. Ziegler, 1990; Underhill and Partington, 1993; Davies *et al.*, 1999; Coward *et al.*, 2003), the graben was present as an earlier fault-bounded embayment on the northern margin of the North Permian Salt Basin (Ziegler, 1990; Hodgson *et al.*, 1992; Glennie, 1998). Thinning of the unit in the NW and SW across salt-cored highs (B and F; Fig. 7.6a) suggests a period of Zechstein Group halokinesis during the Triassic to Middle Jurassic, a period when salt movement was common within the North Sea Basin (i.e. the Central Graben; Hodgson *et al.*, 1992; Stewart *et al.*, 1999). Smaller, irregular thickness variations in the unit to the east and southeast are also interpreted to result from salt movement. Marked thinning of this unit across low-relief salt pillows developed in the immediate hangingwalls of the main faults is due to the extension-driven, gravity-induced sliding of the supra-salt units during later fault development (Figs. 7.4, 7.5).

7.5.2 Unit 2 – Latest Bathonian-Early Callovian

Description. This unit is Early Callovian in age and broadly encompasses the lower part of the Hugin Formation. The unit thickens towards the west with the thickest interval (180 ms TWT) observed within an elongate depocentre that broadly overlies the Unit 1 depocentre A (labelled A; Fig. 7.6b). The unit thins overall north-

westwards across an elongate, NE-SW oriented area (labelled G; Fig. 7.6b) although this thinning is overprinted by a series of small, localised thickened sections up to 2 km in length. The unit also thins to the SW over the previously described salt-cored high (labelled F; Figs. 7.6a, b). Overall thinning of the unit towards the east is complicated by a series of small irregularly shaped thicks and thins (Fig. 7.6b). Thickening is observed across a short (*ca.* 2 km) portion of the southern part of segment 4 of the Gudrun Fault. Poor seismic resolution in the footwall of the Brynhild North fault means it is not possible to map this unit across the fault. However, two elongate thicks are observed in the immediate hangingwall of this fault (labelled H and I; Fig. 7.6b), and the unit thins to the west of these.

Interpretation. Thickening towards the west indicates movement on the graben-bounding fault, interpreted to have been active throughout the Jurassic (Fig. 7.7a, b; e.g. Thomas and Coward, 1996) or may represent continued thermal subsidence along the graben axis post-Permo-Triassic rifting (Badley *et al.*, 1988). Thinning to the NW (G; Fig. 7.6b) broadly coincides with the previously described salt-cored feature and therefore may indicate continued growth of the structure (Fig. 7.7a). Localised thickened sections in this area are interpreted to indicate early movement on the Gudrun North fault and associated antithetic faults (Fig. 7.7a). Thinning of the unit towards the SW coincides with the previously described salt-cored high (F; Figs 7.6a,b) suggesting continued halokinesis of this structure, whilst the irregularly shaped thickened areas to the east of the study area are also interpreted to be the result of halokinesis and localised salt withdrawal. Thickening of the unit in the hangingwall of a small segment of the Gudrun Fault indicates that this fault had initiated and was developing at this time (Fig. 7.7b). Whilst the unit has not been

mapped across the footwall of the Brynhild North Fault, the fault parallel thickened succession in the immediate hangingwall of the fault (H and I; Fig. 7.6b) and the thinned succession to the west of this may represent the initial development of the rollover anticline and associated hangingwall depocentre implying activity on this fault system (Fig. 7.7a). In detail, these thickness distributions within the hangingwall suggest that segments 2 and 5 of the Brynhild North Fault were active at this time.

7.5.3 Unit 3 – Early-Mid Callovian

Description. This unit is Early-Mid Callovian in age and encompasses the upper part of the Hugin Formation. The unit is thickest (120 ms TWT) to the west of the study area (labelled A; Fig. 7.6c) whilst thinning to the north, east and south. Thickening is observed over a ca. 8 km length of the northern half of the Gudrun Fault, over segments 3 and 4 (Fig. 7.6c). Thickening of the unit is also observed between segment 4 at the northern end of the Brynhild Central fault and the parallel antithetic fault (Fig. 7.6c). To the north of the study area, an elongate thick is again observed in the hangingwall of the Brynhild North fault and the unit thins to the west of this (Fig. 7.6c). The unit also thins to the NW (labelled G; Fig. 7.6c) broadly coincident with the thinned zone previously identified (labelled G; Fig. 7.6b). This overall north-westwards thinning is again interrupted by a series of more localised thickness changes. Truncation of the Top Hugin reflection event by the BCU along the footwall crest of the Sigrun Fault, towards the eastern margin and in the SW corner of the study area prevents the thickness of the unit being determined in these regions.

Figure 7.7 Schematic reconstruction of Mid-Late fault development within the study area. Sections are presented across (i) the Gudrun North and Brynhild North fault zones and, (ii) the Gudrun and Brynhild fault zones at the following times; (a & b) Top Unit 2 - Early syn-rift. (c & d) Top Unit 3 - Early syn-rift. (e & f) Top Unit 4 - Rift climax.

Interpretation. Westwards thickening of the unit implies continued movement on the GBFZ. In addition, thickening across the Gudrun Fault indicates it had propagated to the south, with activity on segments 2 and 3 since deposition of Unit 2 (Fig. 7.7d). Activity was also continuing on the Gudrun North (i.e. segment 1), Brynhild Central (i.e. segments 3 and 4) and Brynhild North (i.e. segments 2, 3 and 4) faults, generating locally thickened sections in their hangingwalls (Fig. 7.7 c, d). The antithetic faults to the Gudrun North segment are active at this time generating the localised thicks and thins observed in the NW of the study area. Thickness variations, however, in this area are extremely subtle, making recognition of individual faults and associated depocentres difficult. The truncation of the Top Hugin event to the east of the study area beneath the BCU suggests that rotation of the hangingwall uplifted this part of the slope above sea level either during or after deposition of the unit. This part of the study area was, therefore, either a site of non-deposition at this time or the unit has been subsequently eroded as rotation of the hangingwall continued (Fig. 7.7a, c, e). In the SW of the study area, the absence of Unit 3 indicates continued halokinesis of the previously identified salt-cored high (labelled F; Figs. 7.6a,b) which meant this area was either a site of non-deposition or that any deposits were subsequently eroded.

7.5.4 Unit 4 – Late Jurassic

Description. This unit is Late Callovian to Volgian in age, encompassing the deep marine shales of the Heather and Draupne formations, and was deposited during the main period of Late Jurassic extension (e.g. Milton, 1993; Williams, 1993; Knott, 2001). In general, the unit is thickest in the west of the study area (labelled A; Fig. 7.6d) and thins eastwards. Marked thickening is observed across the entire lengths of

the Gudrun fault and the Brynhild Central, South and East faults (Fig. 7.6d). Marked thinning is observed across the hangingwall anticline of the Brynhild North Fault (labelled J; Fig. 7.6d), westwards of an elongate thick developed in the immediate hangingwall of this fault, and across the Brynhild Central hangingwall anticline (labelled K; Fig. 7.6d). Minor thickness variations are also observed across the antithetic and synthetic faults associated with the main faults. In the NW and SW of the study area, the unit thins over the previously identified salt-cored highs (labelled G and F; Fig. 7.6d). The Top Heather event is truncated by the BCU to the east of the study area and Unit 4 is absent across the footwall of the Sigrun Fault.

Interpretation. Westward thickening of the unit indicates rotation of the hangingwall dip slope as a result of continued movement on the GBFZ. Thickening across the majority of faults within the study area indicates they were all active at some point during deposition of Unit 4. However, it has been observed that some faults tip out within this unit (section 7.4) indicating that some of the structures ceased to be active at this time (Fig. 7.7e, f). Thinning across the Brynhild North fault rollover anticline highlights continued displacement along this fault leading to further development of the hangingwall anticline, syncline and associated antithetic faults (Fig. 7.e). Truncation of the unit to the SW and thinning to the NW over previously identified salt-cored highs indicates continued growth of these structures.

7.5.5 Summary of Structural Development

During deposition of pre-rift Unit 1, despite possible activity on the GBFZ, none of the fault arrays within the study area were active. The main structural control on sedimentation within the study area at this time was post-rift thermal subsidence

associated with Permo-Triassic rifting and halokinesis of the underlying Zechstein Group, which formed salt-cored highs to the NW and SW of the study area that continued to influence subsequent deposition. Halokinesis also occurred in the SE of the study area, where a series of depocentres developed that overprint the overall west-to-east shallowing and thinning of the unit.

Fault activity within the study area started during deposition of Unit 2 (lower part of the Hugin Formation). At this time, activity was focussed on short constituent segments of the Gudrun, Gudrun North and Brynhild North faults (Figs. 7.7a,b). Movement on numerous short, low displacement fault segments is typical of the earliest phases of rifting or rift initiation (e.g. Gawthorpe and Leeder, 2000). Continued growth of salt structures influenced the thickness of the unit to the NW and SW of the study area leading to a structurally controlled embayed depocentre that persisted throughout deposition of Unit 2 and most of Unit 3. This embayment deepened towards the west and the Graben Bounding Fault Zone (GBFZ), whilst towards the east was bounded initially by rotation of the hangingwall dipslope and later by activity on the Gudrun and Brynhild fault arrays.

Propagation of previously active segments and initiation of activity on new segments occurred during deposition of Unit 3 (upper part of the Hugin Formation; Fig. 7.7c,d). However, many of the faults remained inactive over most of their lengths until deposition of Unit 4 (Heather and Draupne formations) and the main phase of regional extension (e.g. Cockings *et al.*, 1992; Rattey and Hayward, 1993) during which most of the fault segments developed (Fig. 7.7e,f). Extension within the study area propagated from north to south, up the graben axis, and from west to east, up

the hangingwall dipslope (see also Jackson and Larsen, 2009). There are some exceptions to these broad trends, however, with larger structures remaining active for longer and smaller features becoming inactive sooner despite being located further to the east or south. In these cases, it is interpreted that the larger structures acted as a focus for displacement, leading to the smaller features becoming inactive sooner than might have otherwise been predicted. Activity on the Gudrun North Fault, for example, post-dates activity on the Gudrun Fault, despite the former being located to the NW.

The development of large through-going fault segments that accommodate the bulk of displacement (e.g. the main faults studied) is more typical of the rift climax phase (e.g. Gawthorpe and Leeder, 2000). In this case they formed during the Upper Jurassic when the fault and basin configuration observed present-day became established.

7.6 Sedimentology and Sequence Stratigraphy of Early Syn-rift Succession

Facies and sequence stratigraphic analysis of the Hugin Formation has been undertaken to investigate the impact of the syn-depositional structures described above on the sedimentology and stratigraphical architecture of the early syn-rift succession. The Hugin Formation is highly variable within the South Viking Graben, comprising fluvially-dominated deposits to the south (Milner and Olsen, 1998; Folkestad and Satur, 2008), proximal to the main graben feeder system, that pass northwards into shoreface and bay-fill successions (Fig. 7.8) (Harris and Fowler, 1987; Cockings *et al.*, 1992; Varadi *et al.*, 1998; Kieft *et al.*, in review.).

Figure 7.8 Early Callovian depositional Setting of the Hugin Formation. The Sigrun-Gudrun study area is highlighted. At this time, Hugin Formation deposition occurred within restricted marine embayments and/or lagoons that were at times fronted by barrier islands. The main feeder system to the basin was located to the south whilst localised drainage from the Utsira High is also inferred (from Kieft *et al.*, in review).

Facies Assoc.	Facies	Description	Bioturbation
Bay-fill	Restricted marine shale	Parallel-laminated mudstone with rare siltstone laminae, locally sideritic. Monospecific assemblage of disarticulated bivalves oriented parallel to bedding and concentrated into thick (60-205 cm) bioclastic-rich beds. Rare gastropods. Varies 30 to 425 cm thick.	BI = 0-2 (Pl)
	Muddy shallow bay-fill	Interbedded claystone and very fine-grained sandstone. Non-parallel and wavy lamination, trough cross-stratification and hummocky cross-stratification in sandstone beds. Rare disarticulated bivalve shells oriented parallel to bedding. Varies 10 to 725 cm thick.	BI = 2-3 (Te, Sk, Pl, Ar, Pa, Op)
	Sandy shallow bay-fill	Fine- to medium-grained sandstone. Moderately to well sorted. Wavy bedding, high-angle cross-bedding and occasional wave ripples. Calcite cement and soft sediment deformation locally observed. Disarticulated bivalves throughout, sometimes concentrated in thin beds (<10 cm). Varies 60 to 370 cm thick.	BI = 0-4 (Pa, Dip, Ro/Cy?, Pl, Op)
	Marine channel-fill sand	Coarse- to fine-grained sandstone with fining- and coarsening-upward motifs. Moderately to well sorted with sub-rounded grains. Sparse sedimentary structures include ripples and mudstone drapes. Disarticulated bivalves and shell fragments common, concentrated into thin beds (<40 cm). Locally contains calcite-cemented nodules. Sharp, possibly erosive contacts at individual bed bases. Varies 250 to 500 cm thick.	BI = 1-3 (Op, ET)
Shoreface	Upper Shoreface	Fine- to very coarse-grained sandstone, dominated by upward coarsening motifs. Moderately to well sorted. Cross-bedding dominates although is locally obscured by cement. Rare bioclastic and carbonaceous material observed. Varies 425 to 800 cm thick.	BI = 0-2 (Th, Te, Pl, Op)
Coastal Plain	Coal	Coal and carbonaceous shale with occasional silt partings. Basal contact typically parallel although may be undulose. Diagenetic pyrite nodules are recognised. Varies 30 to 520 cm thick.	BI=0

Table 7.2 Facies summary of the Hugin Formation within the Sigrun-Gudrun study Area (Chapter 5; Kieft *et al.*, in review). Bioturbation fabric is described using the Bioturbation Index of Taylor and Goldring (1993). Individual trace fossils, listed in brackets, are abbreviated as follows: Pl = *Planolites*, Te = *Teichichnus*, Sk = *Skolithos*, Ar = *Arenicolites*, Pa = *Paleophycus*, Op = *Ophiormorpha*, Dip = *Diplocraterion*, Ro = *Rosselia*, Cy = *Cylindrichnus*, ET = *Escape trace*, Th = *Thalassinoides*.

Figure 7.12 Facies and sequence stratigraphic summary of the Study Area. (a) Bay-fill facies association (well 15/3-4). (b) Remnant shoreface sandstone of a barrier island system (well 15/3-7). Abbreviations: ts = transgressive surface, trs = tidal ravinement surface, wrs = wave ravinement surface, fs = flooding surface.

Core and wireline-log data have been integrated from six wells within the study area, which has enabled the identification of six main facies that are grouped into three facies associations; bay-fill, shoreface and coastal plain. These facies and associations are summarised below (Table 7.2; Fig. 7.9; Kieft *et al.*, in review). The Hugin Formation comprises a series of stacked, progradational parasequences that record periodic infilling of a single, large restricted marine embayment and/or lagoon (Fig. 7.8). At times the embayment or lagoon was completely infilled, allowing the development of regionally extensive coastal plain deposits that cap the parasequences.

The embayment or lagoon is interpreted to have been developed at the same time as tidal creek or inlet channels that erode into bay-fill and coastal plain deposits. A possible tidal signature is interpreted throughout the succession; this may reflect the absence of wave reworking in the sheltered lagoon or embayment, or it may be a function of the narrow, elongate shape of the South Viking Graben which may have locally amplified tidal currents and suppressed wave activity (Husmo *et al.*, 2003; Folkestad and Satur, 2008). Seismic mapping (section 7.5) indicates that salt-cored and fault-controlled structural highs delimited the margins of a large embayment within the study area that deepened to the west and shallowed to the east during Hugin Formation deposition.

Cross-bedded sandstones of the shoreface facies association are identified in well 15/3-7 during latest Hugin Formation deposition at this location (Fig. 7.9b). These sandstones are interpreted to be the preserved remnants of a coeval barrier or spit

system that delimited the western margin of the embayment for at least the latter stages of Hugin Formation deposition. This barrier system was coincident with the approximate position of the Gudrun Fault and its development may have been focussed along this fault system, implying that its footwall formed a topographically high feature at this particular time (Fig. 7.8). However, the inferred barrier shoreline lay in a more westward (basinward) position during deposition of the underlying bay-fill units, which suggests that the Gudrun Fault did not influence its position during deposition of the lower Hugin Formation. The interpretation of a barrier shoreline during Hugin Formation deposition is in agreement with other interpretations of depositional setting within the study area (Harris and Fowler, 1987).

Barrier islands are a common feature of transgressive shorelines, forming as a result of onshore transport and accretion of sand (Fig. 7.10a; Davis, 1994). During subsequent progradation, the embayment developed landward of the barrier will be infilled and capped by coastal plain deposits, and, with enough sediment supply, the barrier island expands to become part of a prograding strandplain or wave-dominated delta (Fig. 7.10b). The stacked succession of bay-fill sediments within the Hugin Formation are interpreted to represent the landward portion of a coastal system that underwent several regressive-transgressive cycles, implying that the majority of the regressive sediments occur to the west of the Sigrun-Gudrun area (i.e. in a more basinward position). The data available from the Hugin Formation is not sufficient to constrain the development of the regressive deposits within each parasequence, although two end member scenarios may be proposed: (i) stacked series of strandplain sandstones of significant extent down depositional dip; or (ii) a

Figure 7.10. Depositional model for the Hugin Formation within a barrier-bay setting at Sigrun-Gudrun. (a) During transgression, a barrier shoreline migrates and accretes slowly landwards whilst bay-fill sediments prograde to infill the back-barrier basin. (b) During regression, with sufficient sediment supply, the back-barrier basin has been completely infilled and a strandplain shoreline system is developed. This will continue to prograde into the basin with sufficient sediment supply (modified from Clifton, 2006). The position of key sequence stratigraphic surfaces is highlighted. Abbreviations used; wrs = wave ravinement surface formed at the base of the retreating shoreline; trs = tidal ravinement surface formed by erosion by tidal currents during transgression; ts = transgressive surface recording the onset of transgression; fs = flooding surface capping the most landward retreat of the shoreline (*sensu* Cattaneo and Steel, 2003).

narrow, vertically aggrading, barrier sandstone (or stacked series of such sandstones). These two end members will be controlled, in part, by sediment supply with the former scenario being favoured by high sediment supply and the latter by low sediment supply. In the Sigrun-Gudrun area, no obvious fluvial feeder systems are recognised and longshore drift is interpreted to be the main mechanism for supply of sediment. Consequently, a narrow, vertically aggrading barrier sandstone may be more likely in this area.

Four key surfaces are recognised within the studied transgressive succession in the Sigrun-Gudrun area (Figs. 7.9, 7.10a,b). Transgressive surfaces (*sensu* Cattaneo and Steel, 2003) are interpreted at the base of the bay-fill successions recording the first evidence of deepening within the succession (Figs. 7.9a, 7.10a,b). These may be associated with tidal ravinement surfaces (*sensu* Swift, 1968; Cattaneo and Steel, 2003) which lie at the base of tidal channels (Figs. 7.9a, 7.10a,b). The base of the shoreface sandstone is marked by a wave ravinement surface (*sensu* Swift, 1968; Cattaneo and Steel, 2003; Fig. 7.9b, 7.10a,b) whilst the juxtaposition of open marine shales directly above shoreface sandstones is interpreted as a flooding surface (*sensu* Van Wagoner *et al.*, 1988; Fig. 7.9b, 7.10a,b). It is worth emphasising that the interpretation of these four stratigraphic surfaces depends on the interpretation of a barrier-island and lagoon setting. Similar bay-fill successions could be developed in interdistributary bays within a deltaic setting (e.g. in the “Sleipner Delta” system), without a barrier island fronting them. In such a context, the base of a bay-fill succession could be interpreted to record deepening across a flooding surface (*sensu* Van Wagoner *et al.*, 1988), and the bay could be infilled by regressive deltaic deposits that initiate as part of a bayhead delta and then expand outside of the bay

during the later stages of regression (e.g. in Hugin Formation in the Sleipner Field; Folkestad and Satur, 2008).

Although barrier sandstones can often be removed by ravinement during transgression, their preservation will be partly controlled by the shoreline trajectory during the transgression (Helland-Hansen and Martinsen, 1996; Muto and Steel 2002). The stacked bay-fill parasequences within the study area are interpreted to represent an accretionary transgression (*sensu* Helland-Hansen and Martinsen, 1996) during which accommodation was continuously generated, due to rotation of the hangingwall dip slope, and repeated period of bay infill behind the retreating shoreline.

7.6.1 Sequence Stratigraphic Correlation

Within the study area, a thick (>200 m) Hugin Formation succession can be subdivided into a series of 16 parasequences, separated by locally correlatable flooding surfaces. As part of this study, these observations were integrated with core and wireline-log characteristics to iteratively refine the position of the surface (i.e. where they corresponded to thick offshore mudstones overlying shallow marine deposits). These parasequences typically stack in an aggradational pattern. Three well correlation panels have been generated through the study area utilising the facies associations and sequence stratigraphic surfaces summarised above and described fully in Chapter 5. It should be noted that the presented correlations are one interpretation of the available data and alternative interpretations may be equally viable. Correlation of sequence stratigraphic surfaces between the wells has been constrained by a newly revised high-resolution biostratigraphical and

chronostratigraphic scheme based on the StatoilHydro ASA regional database. These data suggest that the parasequences are on average 500-600 ka duration. The mapped seismic horizons have also been related to stratigraphic intervals on the correlation panels.

Axis-perpendicular correlation

This correlation traverses the study area from west to east crossing the Gudrun, Brynhild and Sigrun fault arrays (Figs. 7.2, 7.11). The lower eight bay-fill parasequences of the Hugin Formation are only developed in the west of the study area and the oldest two are not penetrated in the most westerly well (15/3-1S; Fig. 7.11). These parasequences thicken towards the west towards the GBFZ, whereas towards the east they thin and onlap underlying Triassic strata. Several of these bay-fill parasequences are truncated at their tops by thick (up to 45 m) channel-fill sandstones. Although not cored, these sandstones are interpreted as tidal inlet channel-fill deposits, based on their clean, low-GR character, which is consistent with marine reworking, and the occurrence of shoreface sandstones in overlying strata. Alternatively, they may represent fluvio-tidal channel-fill deposits (see Chapter 5). Parasequence 6 is capped by a thick coal which is developed over the western (more basinward part) of the study area. The shoreface deposits described above are only developed at the western end of the correlation panel in parasequence 7, and pass up-dip into a time-equivalent bay-fill succession. This supports the interpretation of this sandstone as a barrier or spit system that was broadly coeval with an embayment or lagoon. Parasequences 6 and 7 are both observed to thicken across the Gudrun fault. This may in part relate to differential compaction of the sediments, especially the variable thickness of coal between the two wells. However

Figure 7.11 Axis Perpendicular Correlation. This correlation transects the study area from west to east, up the hangingwall dipslope. Facies have been interpreted from core data and extrapolated into uncored intervals based on wireline log response. Hugin Formation development was diachronous across the area and shows parasequence scale thickness variations, in part related to structural development.

seismic mapping (section 7.5) suggests that segments of the Gudrun fault were active at this time. Importantly, this thickening is not accompanied by a noticeable facies shift.

Within parasequence 8, open marine conditions (Heather Formation) are established in wells 15/3-1S and 15/3-7 due to flooding in the west of the study area, whereas more proximal, marginal marine condition persisted in well 15/3-4. An unconformity is identified at the top of parasequence 8 in wells 15/3-4 and 16/1-3, which dies out westwards. The restricted distribution of this surface may reflect overall rotation of the hangingwall dipslope or erosion of exposed footwall-crests, as is observed at the Sigrun Fault (Section 7.4.4). This surface, therefore, may provide evidence for movement on the Sigrun fault at this time. The top of parasequence 8 records an important flooding event within the study area, the shoreline moves ca. 10 km up the hangingwall dipslope and the broadly aggradational stacking pattern (parasequences 1-7) becomes retrogradational (parasequences 9 to 16). Parasequences 9 to 13 show open marine conditions prevailing over the western portion of the study area, whilst shallow marine Hugin deposition is restricted to well 16/1-5 in the east. Here the Hugin Formation is interpreted to comprise stacked shoreface sandstones, based on its wireline log character. These parasequences thicken towards the east, indicating that a major basement-involved fault to the east of well 16/1-5 had become active at this time (Fig 7.2). This area was flooded and open marine conditions became established during deposition of parasequence 14. Throughout parasequences 9 to 14, coastal plain deposition prevailed at well 16/1-3.

Hugin Formation development within the final two parasequences (15 and 16) is restricted to well 16/1-3, and these deposits are interpreted from wireline log signature to comprise bay-fill facies. These units pass to the west and east into open marine shales of the Heather Formation. They are also observed to thicken to the east between wells 16/1-3 and 16/1-5 and the west between 15/3-4 and 15/3-7, suggesting that the regional fault identified to the east of well 16/1-5 was still active whilst the southern segment of the Brynhild Central (segment 2) was also active at this time. Within well 16/1-3, parasequence 16 is overlain by open marine shales of the Heather Formation, which extended across the entire basin at this time.

Local axis-parallel correlation

This correlation trends SW-NE, broadly parallel to the Brynhild Central and Sigrun fault arrays, on the footwall of the Sigrun fault (Figs. 7.2, 7.12). No structural features are mapped between the two wells on this correlation which are less than 5 km apart (Fig. 7.2). Only the lower eight parasequences of the Hugin Formation are recognised in this correlation; parasequences 9 to 17 described above are represented by open marine shales of the Heather Formation. The parasequences within both wells comprise stacked bay-fill successions. However, the deposits to the NE are typically more sand rich with the bay-fill successions being eroded by tidal-inlet channels or creeks interpreted to have tidal ravinement surfaces at their bases. The succession to the SW is more mud-rich and channel sandstones are not developed. This section serves to highlight the variability that may be observed within the Hugin Formation away from areas of obvious structural control. The occurrence of more sand-rich deposits containing more abundant tidal channels in well 15/3-4 implies

Figure 7.12 Local Axis Parallel Correlation. Oriented SW to NE, this correlation highlights the variability of shallow marine depositional systems away from structural control. Some structural control is in evidence due to the parasequence scale thickness variations and the local erosion observed at the top of parasequence 8. This is interpreted to relate to movement on the Sigrun fault to the east of these wells and erosion of its footwall crest.

that this well is located closer to a long-lived tidal inlet system that supplied marine-reworked sediment into the Hugin Formation embayment or lagoon.

In well 15/3-4 the biostratigraphical analysis suggests a palynological zone is missing from the top of parasequence 8 (Figs. 7.11; 7.12). This coincides with thinning of the parasequence across a short distance (Fig. 7.12). As the final parasequence before open marine deposition (Heather Formation) across this area the missing section could have been removed by ravinement during transgression. The well also lies close to the footwall crest of the Sigrun Fault across which seismic mapping (section 7.5) indicates Callovian and Upper Jurassic units have been removed therefore the missing section may relate to this later erosional event associated with footwall uplift.

Regional axis-parallel correlation

This correlation trends N to SW and intersects four wells within the study area (Figs. 7.2, 7.13). It broadly parallels the Gudrun and Brynhild North fault arrays and, as such, does not cross any structures interpreted from seismic data to have been active during Hugin Formation deposition (Fig. 7.2). Parasequences 1 through 3 comprise bay-fill deposits at the northern end of the correlation panel but pass southwards into time equivalent coastal plain deposits of the Sleipner Formation. Within these parasequences, several thick (up to 40 m) tidal inlet channels erode into the bay-fill successions. Parasequence 4 records a major flooding event within the study area within which even the southernmost well (15/2-1) was flooded with shallow marine sediments that persisted until parasequence 8. Thick tidal inlet channels erode into the bay-fill successions in wells 15/3-2R and 15/3-3. The thick coal layer identified

Figure 7.13 Regional Axis Parallel Correlation. Traversing the study area in a broadly north south direction, this correlation panel trends parallel to the structural features within the study area. Overall the interval floods from north to south and, like figures 7.13 and 7.14 parasequence scale thickness variations are observed highlighting the influence of structural development.

on the axis perpendicular correlation (well 15/3-7; Fig. 7.11, 7.13) in parasequence 6 can be correlated southwards for over 17 km, highlighting the regional extent of this unit. No coal is observed in well 15/3-2R, which may suggest that the bay in this location was not infilled sufficiently to have accumulated thick coastal plain deposits. Alternatively the coal may have been eroded by a tidal inlet channel. Parasequence 6 thins markedly southwards (i.e. 15/3-2R to well 15/2-1; Fig. 7.13), probably due to differential compaction of the thick coal unit in the south of the correlation.

The shoreface sands previously identified within parasequence 7 (wells 15/3-1S and 15/3-7; Fig. 7.11) can also be correlated over the length of the section. To the north, these units sharply overlie restricted marine shales of the bay-fill facies association and are interpreted to have a wave ravinement surface at their base (e.g. well 15/3-2R; Fig. 7.13). To the south, the restricted marine claystone appears to pinch out and is replaced by a sandier interval juxtaposed immediately above the coals capping parasequence 6 (well 15/2-1). Here the shoreface succession is interpreted to comprise both lower and upper shoreface deposits, indicating that the barrier or spit system was more fully developed in this location. Thinning of this unit in the north corresponds to seismically mapped thinning of the upper part of the Hugin Formation across the hangingwall anticline of the Brynhild North fault (Fig. 7.6; section 7.5).

Parasequence 8 indicates that open marine conditions and shale deposition (Heather Formation) occurred across the majority of the study area. Although shales are observed at this level in well 15/3-2R, a lack of biostratigraphical and core data

through the interval makes it difficult to interpret the depositional setting of these shales. This well (15/3-2R) is also the only well within the correlation to have sandstones deposited in parasequences 9 and 10, which are represented in the rest of the correlation by open marine shales. A short (<2 m) core interval through these sandstones shows that they are cross-bedded, medium- to coarse-grained sandstones, similar to the upper shoreface sandstones of the underlying parasequences. These sandstones are thus interpreted have been deposited within progradational shoreface units which pinch out to the south into open marine shales of the Heather Formation. Alternatively, given the ambiguity of sedimentological features within the sandstones and the lack of biostratigraphical data, these sandstones may have been deposited as turbidites within the open marine shale. Intra-Heather and Draupne Formation turbidites have been identified within or proximal to the study area (e.g. Cockings *et al.*, 1992). The thickness variations observed in parasequences 9 and 10 across the correlation are interpreted to be due to differential compaction between the sandstones and open marine shales.

7.6.2 Palaeogeographical Reconstruction

The well correlations described above highlight the diachroneity of flooding within the study area during Hugin Formation deposition. Three palaeogeographical reconstructions are presented for the study area, two from Early Callovian times (seismically-defined Unit 2/parasequence 5, Fig. 7.14a; seismically-defined Unit 3/parasequence 7, Fig. 7.14b) and one from Mid Callovian times (lower part seismically-defined Unit 4/parasequence 10; Fig. 7.14c). These maps are snapshots in time, recording the final shallow-marine deposition at the top of the named parasequences prior to subsequent transgression. To provide better constraint for the

Figure 7.14 Palaeogeographical reconstruction of deposition within the Sigrun-Gudrun. The results of structural mapping have been integrated with well data to constrain depositional patterns. Maps presented show maximum regression within parasequences. (a) Early Callovian (Parasequence 5; structural analysis of Unit 2). (b) Early Callovian (Parasequence 7; structural analysis of Unit 3).

(c)

KEY

Coal	Lower Shoreface
Restricted Marine Claystone	Upper Shoreface
Marine channel fill sand	Open Marine Heather Fm.
Sandy bay fill	Continental Sleipner Fm.
Muddy bay fill	○ Constraining wells
Missing section present day	— Inactive normal fault
--- Seismically mapped thin	— Active normal fault

Figure 7.14 Palaeogeographical reconstruction of deposition within the Sigrun-Gudrun (cont.). The results of structural mapping have been integrated with well data to constrain depositional patterns. Maps presented show maximum regression within parasequences. (c) Mid Callovian (Parasequence 10 - Structural analysis of Unit 4).

palaeogeographical maps, data from two additional wells (16/1-7 and 16/1-2) located to the NE of the study area have also been utilised, in addition to those utilised in the correlation panels described earlier (i.e. Figs. 7.11-7.13). Biostratigraphical and wireline-log analysis from these two wells has enabled the timing of Sleipner Formation and Heather Formation deposition in this region. It must be stressed that the presented palaeogeographical reconstructions are one interpretation of the data and other realisations may be equally viable.

Early Callovian 1 (Seismic Unit 2-Parasequence 5)

During the Early Callovian, coastal plain deposition (Sleipner Formation) prevailed in the east of the study area (i.e. wells 16/1-7, 2 and 3; Fig. 7.14a). Biostratigraphical data suggest that this unit is not preserved within well 16/1-5, implying this was either an area of subaerial exposure and non-deposition or that the unit has been subsequently eroded. Time-equivalent deep marine sedimentation is interpreted to the west of the embayed shoreline, proximal to the basin axis. Overall thickening of the unit to the west implies movement of the GBFZ and rotation of the hangingwall dipslope. As a result, it is interpreted that regional drainage in this area was towards the west, down the hangingwall dipslope. Wells in block 15 were dominated by facies of the bay-fill succession, except for wells 15/2-1 and well 15/3-7 where coastal plain coals were developed. It is interpreted that this coal developed atop the bay-fill facies succession, which prograded and infilled the restricted-marine embayment in this area. The southern margin of the embayment is interpreted to have been controlled by the salt-cored high mapped in the SW of the study area, whilst the eastern margin is interpreted to have been controlled by rotation of the hangingwall dipslope. The prominent coal within this unit is developed in a

basinward position, fringed by bay-fill sediments implying it developed on a back-barrier bar. To the north (well 15/3-3) a tidal inlet channel was developed, close to the northern tip of the Gudrun fault segment which was active at this time (Fig. 7.6b); the channel system may have been focussed at or diverted around the propagating fault tip. To the north, interbedded heteroliths were developed (well 15/3-2R). These deposits are interpreted to have fringed a restricted marine embayment bounded to the NW by an active salt-cored high (Fig. 7.6b) and to the NE by the footwall crest of several, active segments associated with the Brynhild North fault (Fig. 7.6b).

Early Callovian 2 (Unit 3)

Coastal plain deposits of the Sleipner Formation still prevailed immediately to the east of the study area (Fig. 7.14b). At well 16/1-5 this interval is not penetrated suggesting, in a similar manner to Early Callovian 1 times, that this area was emergent and a site of non-deposition or that the unit was subsequently eroded. To the north east of well 16/1-5, however, open marine shales of the Heather Formation are penetrated (well 16/1-2). Regional data indicate that a normal fault to the east of this area became active at this time, allowing open marine flooding of a fault-controlled embayment (Kane and Jackson, 2008; H.C. Ronnevik, *pers. comm.*). There is a comparatively rapid transition (<5 km) from open marine shales of the Heather Formation (well 16/1-2) to coastal plain deposits of the Sleipner Formation (well 16/1-7). This suggests that the Hugin shoreface was relatively narrow at this time, possibly due to steep, tectonically-generated gradients. Although rotation of the hanging-wall dip slope still resulted in the bulk of drainage off the Utsira High being

directed westwards, more localised drainage to the east and northeast into this fault-bounded embayment may have occurred.

Shoreface sandstone development prevailed to the west of the study area and these deposits are penetrated in wells 15/2-1, 15/3-1S, 15/3-7, 15/3-3 and 15/3-2R (Fig. 7.14b). These shoreface sandstones are interpreted to have formed a broadly NE-SW-trending barrier shoreline that protected the restricted marine embayment developed behind it. To the south, the position of this shoreline is interpreted to have been controlled by the active salt-cored high interpreted from seismic mapping (Fig. 7.6). Passing to the north, the shoreface sandstones appear to be focussed along the Gudrun Fault array, which is interpreted to have been active along its northern segments at this time (Figs. 7.6c, 7.14b). This structure may have acted as a focus point for sedimentation, with sands accumulating above and against the footwall crest that may have formed a topographical high at the seabed. It should be noted, however, that the same facies is observed in both the hanging and footwall of these active fault segments (wells 15/3-1S and 15/3-7). This relationship implies a high sediment supply relative to the rate of structural development therefore accommodation space was infilled and sea bed topography unable to develop. To the north (well 15/3-2R) these shoreface sandstones are interpreted to be focussed along the axis of the rollover anticline developed in response to movement on the Brynhild North fault array located 4 km to the east (Fig. 7.6c). Wells 15/3-4 and 3-5 lie within the restricted marine embayment at this time and penetrated relatively proximal bay-fill facies (tidal dune and tidal inlet channel). These facies are interpreted to lie close to the eastern margin of the embayment, which may have been controlled by rotation of the hangingwall or movement on the Sigrun Fault.

Mid Callovian (Lowermost Unit 4)

By this time, the bulk of the study area had been flooded by open marine shales (Heather Formation; Fig. 7.14c). Shallow marine Hugin Formation deposition is interpreted to continue at well 15/3-2R to the north and well 16/1-5 to the SE of the study area. Well 16/1-3, also located towards the SE, is still a site for coastal plain deposition (Sleipner Formation). Although there is some ambiguity in interpreting the sandstones at 15/3-2R as shoreface deposits (see discussion above), this interpretation implies that the shoreface fringed the rollover anticline located 4 km to the east of the Brynhild North fault however it is also possible that the sands extended eastwards at this time over the footwall of the fault. The source of these sandstones is uncertain but four possible source regions can be interpreted: (1) local sourcing from the footwall crest of the Brynhild North fault, an area of known erosion (Figs. 7.6d, 7.14c); (2) reworked from underlying sediments and deposited during transgression; (3) sourced from the coastal plain penetrated in well 16/1-3 by a feeder system focussed along the hangingwall anticline ahead of the Brynhild North fault array; (4) transported from the north where thick, wave-influenced shelf sandstones are penetrated (well 24/12-1R; Kane and Jackson, 2008) by wave-driven, longshore currents. Shoreface sandstones are developed within 16/1-5 at this time. Regional data indicate that a fault to the east of this well may have been active at this time, allowing a fault-controlled embayment to develop within the hangingwall of the graben and form a focus for shoreface sandstone deposition (Kane and Jackson, 2008; H.C. Ronnevik, *pers. comm.*). The wave-dominated character of the shoreline in this inferred embayment is consistent with the wider, less confined geometry of the South Viking Graben at this time.

7.7 Discussion

7.7.1 Structural development

The study area is dissected by a rift-related fault array which comprises four main segments and which formed as a result of gravity-driven sliding above the mobile Zechstein Group salt (this study; Jackson and Larsen, 2009). This style of rift-related faulting has been described from several other basins within northwest Europe that contain Zechstein Group salt (cf. rift-raft tectonics of Penge *et al.*, 1999). Whilst the studied faults formed as a result of gravity-driven sliding, they are still recognised to have initiated as a series of individual segments which subsequently propagated and linked to form through-going fault zones. This type of fault growth has been recognised in extensional fault systems in numerous other settings (e.g. Peacock and Sanderson, 1991; Anders and Schlishe, 1994; Cartwright *et al.*, 1995; Gawthorpe and Leeder, 2000). During this process, relay ramps between fault tips are breached, thereby allowing individual segments to link. The preservation of relay ramps between the different segments of the Brynhild Fault system suggest that extension ceased before linkage was achieved. Extension may have ceased following the end of movement on the GBFZ, although movement on the GBFZ is not interpreted to have ceased until the Late Jurassic (e.g. Williams, 1993, Faereth *et al.*, 1997; Coward *et al.*, 2003) before which the Brynhild East, South and Central segments had all ceased to be active (i.e. Early to Middle Volgian; Jackson and Larsen, 2009). Alternatively, the focus of later extension may have migrated up the hangingwall dip slope, with extensional strain being accommodated on faults to the east of the study area. The migration of faulting progressively updip with time has been

recognised within the study area (see also Jackson and Larsen, 2009) as well as within physical analogue models of gravity-driven fault growth (Childs *et al.*, 1993).

Not only did the Zechstein Group salt influence structural development by acting as a detachment, but halokinesis of this unit also formed several large salt-cored highs within the studied area. Mobilisation of the salt in these areas was initiated in the Triassic, a time when halokinesis has been recognised throughout the North Permian Basin (e.g. Pegrum and Ljones, 1984; Frostrick, 1992; Thomas and Coward, 1996). Growth of these structures continued throughout the Jurassic. The position of the salt-cored highs may be influenced by basement faults which have been recognised to focus salt flow on their footwalls (e.g. Hodgson *et al.*, 1992; Dooley *et al.*, 2005), although this relationship can not be resolved in the current seismic dataset.

7.7.2 Influence of structural development on sedimentation

During deposition of the Hugin Formation, fault displacement was limited and focussed on short (2-5 km) segments; this is typical of the rift initiation phase (e.g. Gawthorpe and Leeder, 2000). The associated increase in accommodation space in the hangingwall of these low-displacement segments was not accompanied by a facies shift across these faults. This relationship, along with the predominantly aggradational stacking pattern observed within the lower part of the Hugin Formation (i.e. parasequence 1-6, Fig. 7.13), suggests that the volume of sediment supplied to the basin was sufficient to keep pace with fault-generated accommodation space, such that the faults did not have a topographic expression that could influence facies distributions (Frostrick and Steel, 1993). A similar relationship is observed within the syn-rift Tarbert Formation in the North

Viking Graben, where a balance between sediment supply and subsidence meant that facies distributions were not influenced by the developing structures (e.g. Davies *et al.*, 2000; McLeod *et al.*, 2002; Hampson *et al.*, 2004). The aggradational stacking pattern observed within the syn-rift Nukhul Formation is also attributed to a balance between sediment supply and accommodation (Carr *et al.*, 2003; Jackson *et al.*, 2005). The retrogradational stacking pattern observed within the upper Hugin Formation deposits (i.e. parasequences 7-16, Fig. 7.13) coincides with increased fault displacement which indicates that at this time, sediment supply was insufficient to keep pace with subsidence. Accordingly, at this time, it may be interpreted that facies belts were being focussed along syn-depositional structures (e.g. barrier shoreline focussed along the footwall of the Gudrun Fault; Fig. 7.14b). Salt-cored structures seem to be more important in the study area than rift-related faulting during deposition of the Hugin Formation, as they form a focal point for shallow-marine deposition (e.g. shoreface sandstone at 15/2-1) and also delimit the northern and southern margins of the embayed coastline. Focussing of shallow marine sandstones above salt-cored highs has been recorded within the Fulmar Formation to the south of the study area (e.g. Stewart *et al.*, 1999; Fraser *et al.*, 2003) and also within the La Popa Basin in Mexico (Aschoff and Giles, 2005).

Despite the balance between sediment supply and subsidence interpreted within the study area, structural development may still have influenced the sedimentology of the Hugin Formation within this area. Tidal signatures observed within the succession may record amplification of tidal currents, a common observation within rift basins (e.g. Ravnås and Bondevik, 1997; Ravnås and Steel, 1998; Carr *et al.*, 2003). Alternatively it may record wave suppression within the embayment, either

due to the presence of a coeval barrier shoreline or due to the partitioning of the basin by developing structures, particularly the salt-cored highs (labelled F and G, Fig. 7.6) to the north and south of the embayment. Strong tidal influence is observed within the Tarbert Formation where fault-block rotation has partitioned the basin into smaller sub basins with limited wave fetch (e.g. Ravnås *et al.*, 1997; Faerseth and Ravnås, 1998; Hampson *et al.*, 2004).

CHAPTER 8

Impact of syn-depositional halokinesis on sedimentology and sequence stratigraphy of the Hugin Formation

8.1 Introduction

The movement (halokinesis) of evaporite-rich units (herein referred to as ‘salt’) within sedimentary basins can be responsible for marked temporal and spatial variations in topography within sedimentary basins. This impacts accommodation development which in turn can influence sediment transport pathways, impacting sedimentology and sequence stratigraphic architecture. Halokinesis may influence the thickness, stratal architecture and facies distribution of syn-halokinetic sedimentary units (e.g. Heward, 1990; Hodgson *et al.*, 1992; Rowan and Weimer, 1998; Clark *et al.*, 1999; Stewart *et al.*, 1999; Cartwright *et al.*, 2001; Fraser *et al.*, 2003; Aschoff and Giles, 2005; Stewart, 2007).

Field and subsurface studies demonstrate that syn-halokinetic stratal units thin and lithofacies may record deposition in shallower water towards salt-cored topographic highs. Depending on the relief of the salt-related topography, depositional systems developing adjacent to these structures may pinch out basinwards over relatively short distances into deeper marine facies (e.g. Johnson *et al.*, 1986; Stockbridge and Gray, 1991; Aschoff and Giles, 2005). Shallow marine shorelines or deltaic systems may be focussed within depocentres formed by salt-withdrawal or dissolution whilst the salt-cored highs and their flanking slopes are areas of non-deposition and/or erosion (Wakefield *et al.*, 1993; Stewart and Clark, 1999; Shelley and Lawton, 2005; Stewart, 2007). Alternatively, where coarse-grained sediment input into the basin is

focused, sediment ‘shadow zones’ may develop down-dip from salt-cored structures (Aschoff and Giles, 2005). The focussed deposition of shallow marine units can contribute to differential sediment loading that will continue to drive salt movement (passive diapirism) from under the depocentres and into the salt-cored highs, thereby creating a positive deposition-subsidence feedback loop (Hodgson *et al.*, 1992; Rowan and Weimer, 1998; Rowan *et al.*, 2003; Gemmer *et al.*, 2004; Aschoff and Giles, 2005; Shelley and Lawton, 2005; Matthews *et al.*, 2007). Consequently, the thickest and most complete stratigraphic sequences may be expected in the axial part of the depocentre whilst towards the margins depositional units amalgamate and unconformities are developed. As a corollary, understanding the distribution and timing of formation of syn-depositional salt-cored highs can provide important insights into facies distribution within shallow-marine successions. However the continued growth of these salt-cored highs, which may be driven in part by ongoing spatially-variable deposition, means that the present-day structural configuration can vary markedly from the syn-depositional configuration. Therefore, it is important to reconstruct the growth of salt-cored topographic features to ascertain their geometry during deposition.

This study focuses on the shallow-marine Hugin Formation (Middle Jurassic) deposited on the hangingwall dip-slope of the South Viking Graben, northern North Sea Basin, offshore Norway. The formation was deposited during the Middle Jurassic during the initial period of a major Jurassic rift event which was also associated with halokinesis of the underlying Zechstein Group (Upper Permian). This study has two main aims; (i) to document the temporal and spatial evolution of a series of salt-cored structures associated with halokinesis of the Zechstein Group;

and (ii) to determine the influence that the development of these structures had on the stratal architecture and facies distribution within the Hugin Formation. The study area has been selected due to the availability of 3D seismic data, high-resolution biostratigraphy and spatial distribution of wells.

8.2 Geological Setting

8.2.1 Structural Setting

The study area is located across the Dagny-Alpha Central area which lies within the Vilje Sub-basin on the southern part of the hangingwall of the South Viking Graben (Fig. 8.1a,b). The area encompasses Norwegian exploration blocks 15/5, 15/6 and 15/8 and extends into part of UK exploration block 16/18. The South Viking Graben is a NNE-SSW trending half-graben, bounded to the west by the Graben Boundary Fault Zone (GBFZ) which itself comprises a series of ESE-dipping faults and delineates the eastern margin of the Fladen Ground Spur (Thomas and Coward, 1996; Husmo *et al.*, 2003). To the east, the graben shallows onto the Utsira High, a long-lived, relatively un-faulted, basement-cored structural high (Thomas and Coward, 1996; Husmo *et al.*, 2003). In the southern part of the graben, three major fault blocks are developed which are bounded by a series of N-S-trending, W-dipping faults (i.e. antithetic to the GBFZ) and which down-step towards the west (Pegrum and Ljones, 1984).

The South Viking Graben initially formed during the Late Permian to Early Triassic rift event and a period of major basin formation within the North Sea (e.g. Coward, 1995; Roberts *et al.*, 1995; Glennie, 1998; Coward *et al.*, 2003). Thermal subsidence

Figure 8.1 Geological Setting of the Study Area. (a) Location map. (b) Schematic Zechstein Group facies map (after Thomas and Coward, 1996). The South Viking Graben formed a fault-bounded embayment at the northern margin of the North Permian Salt Basin. To the north the Zechstein is dominated by dolomites and anhydrites whilst increasing salt influence is observed to the south. Facies and thickness of the formation impacts structural development within the graben. (c) Regional section across the South Viking Graben. The study area (highlighted) is located on the hangingwall of the salt-influenced rift basin.

after this rift event led to the formation of the Northern Permian Basin, with the South Viking Graben forming a NE-SW-trending, fault-bounded embayment along its northern margin (Ziegler, 1990; Hodgson *et al.*, 1992; Glennie, 1998; Coward *et al.*, 2003). Evaporite-rich units of the Zechstein Group were deposited at this time and they have impacted the subsequent structural development within the basin as a result of halokinesis (discussed below) and by forming an intra-stratal detachment horizon which decoupled sub- and supra-salt deformation (e.g. Pegrum and Ljones, Thomas and Coward, 1996; Jackson and Larsen, 2009; Kane *et al.*, *in review*).

During the Early Jurassic (Toarcian-Aalenian), regional thermal uplift occurred due to the impingement of a mantle plume at the base of the lithosphere as a precursor to active rifting within the region (e.g. Ziegler, 1990; Underhill and Partington, 1993; Davies *et al.*, 1999; Coward *et al.*, 2003). Jurassic rifting and movement on the GBFZ within the study area is interpreted to have initiated in the Early or Late Callovian with the main phase of extension and basin subsidence occurring during the Oxfordian to Middle Volgian (e.g. Cockings *et al.*, 1992; Milton, 1993; Williams, 1993; Knott, 2001). The majority of rift-related faults became inactive during the latest Jurassic and there is evidence for the initiation of compression and basin inversion during the early Middle Volgian (e.g. Cherry, 1993; Thomas and Coward, 1996; Knott, 2001; Fraser *et al.*, 2003; Jackson and Larsen, 2008).

8.2.2 Stratigraphic framework

During the Permian, continental deposition prevailed over the study area and evaporite-rich units of the Zechstein Group were deposited. These units are in excess of 1 km thick in the axis of the graben where they comprise ‘basinal’ evaporite facies of dolomite, anhydrite and halite (Thomas and Coward, 1996). These units

pass northwards and eastwards towards the graben margins into carbonate and anhydrite-dominated, 'marginal' evaporite facies (Fig. 8.1c; e.g. Hodgson *et al.*, 1992; Thomas and Coward, 1996; Glennie, 1998; Glennie *et al.*, 2003). Continental conditions continued into the Triassic, with lacustrine and alluvial deposition being represented by the Smith Bank and Skagerrak formations (Warrington and Ivimey-Cook, 1992; Glennie, 1998; Goldsmith *et al.*, 2003). Halokinesis of the Zechstein Group began in the Late Triassic and led to significant lateral thickness variations within the syn-halokinetic units (Pegrum and Ljones, 1984; Thomas and Coward, 1996). This halokinesis is interpreted to have continued into, and partly controlled sedimentation patterns during, the Jurassic within the South Viking Graben (Pegrum and Ljones, 1984; Jackson and Larsen, 2009), in a similar manner to that documented further south in the Central North Sea Basin (Wakefield *et al.*, 1993; Stewart *et al.*, 1999; Fraser *et al.*, 2003; Stewart, 2007).

Uplift of the North Sea Dome during the early Jurassic resulted in the development of a major unconformity in the South Viking Graben and time-equivalent northwards progradation of the Brent Delta in the North Viking Graben (e.g. Bradshaw *et al.*, 1992; Helland-Hansen *et al.*, 1992; Mitchener *et al.*, 1992; Morton *et al.*, 1992b; Husmo *et al.*, 2003). The Brent Delta was fed from the south (axially) by sediment eroded from the North Sea Dome, and from the east and west (laterally) by material derived from the Shetland Platform and southern Norway (Mearns, 1992; Morton, 1992a). The subsequent Jurassic rift event led to a relative rise in sea-level (Underhill and Partington, 1993) and deposition of fluvial (Sleipner Formation) and then shallow marine (Hugin Formation) deposits within the graben (Harris and Fowler, 1987; Cockings *et al.*, 1992). Relative sea level continued to rise during the

Late Jurassic due to fault-related subsidence and eustasy, resulting in deepening of the basin and deposition of shelfal and deep-marine deposits (Heather and Draupne formations; Bradshaw *et al.*, 1992; Fraser *et al.*, 2003).

8.3 Database and Methodology

The seismic dataset for this study comprises a 3-D reflection dataset covering an area of 1025 km² with an inline (N-S) and crossline (E-W) spacing of 12.5m. The record length is 5.5 seconds two-way travel time (TWTT) and the data has not been depth converted. The studied interval lies between 2.5 and 4.5 seconds TWTT. The Zechstein Group evaporites serve to reduce the seismic quality in the sub-salt (pre-Upper Permian) section, therefore detailed interpretation is restricted to the supra-salt succession where data quality is moderate to good. The seismic data is displayed with a reverse polarity (SEG European convention; Brown, 1999) with a downward decrease in acoustic impedance represented by a trough (black) and a downward increase in acoustic impedance represented by a peak (red).

Three age-constrained seismic horizons (top Zechstein Group, top Sleipner Formation and Base Cretaceous Unconformity; Fig. 8.2) which can be mapped across the study area have been identified within the pre- and syn-rift interval of interest. These horizons are moderate- to high-amplitude and relatively continuous, and tie to major lithological boundaries identified in well data. Two additional seismic horizons (intra-Hugin Formation and top Hugin Formation; Fig. 8.2) have been interpreted within the syn-rift succession in the Alpha Central case study area (see Fig. 8.4 for location), where the succession is thickest. These two horizons

Figure 8.2 Stratigraphic column illustrating the studied interval. Also shown is a summary diagram of seismic reflection events picked within the study area. Three seismic horizons can be picked regionally; Top Zechstein, Late Bathonian and Base Cretaceous Unconformity. In the southern part of the study area two additional horizons can be mapped; Early Callovian and Late Callovian.

cannot be interpreted within the Dagny-Alpha North case study area (see Fig. 8.4 for location), due to thinning of the syn-rift succession over salt-cored structural highs.

Ten wells within the study area provide a variety of sedimentological and stratigraphic data (summarised in Table 8.1). Core totalling 361 m in length has been logged from nine of these wells and integrated with wireline-log data. These data have been utilised to develop a facies scheme and sequence stratigraphic framework for the Hugin Formation. Additional published data was available from the neighbouring Sleipner Vest and Øst fields (Milner and Olsen, 1998; Folkestad and Satur, 2008; Kieft *et al.*, *in review*) and has been used to place the study area within a regional palaeogeographical framework. Biostratigraphical data were used to aid identification of age-constrained locally and regionally-developed flooding surfaces.

8.4 Structural Style

The South Viking Graben generally deepens from east to west, from the Utsira High into the immediate hangingwall of the GBFZ (Fig. 8.1c). This overall deepening trend is interrupted by a series of normal faults and salt-cored highs (Fig. 8.3). Three key structural highs (the Dagny, Alpha North and Alpha Central Highs; Fig. 8.3) are recognised within the study area and their structure is described below. Development of these highs is described in section 8.5.

8.4.1 Dagny High

The Dagny High is a NW-SE trending, triangular-shaped salt-cored high, approximately 7 km long and up to 4 km wide at the top pre-rift level (Figs. 8.3;

Well	Available data		
	Core	Wireline-log	Biostratigraphical analyses
15/5-1	X	X	X
15/5-2	X	X	X
15/6-2R	X	X	X
15/6-4	X	X	X
15/6-5	X	X	X
15/6-6	X	X	X
15/6-7		X	X
15/8-1	X	X	X
UK 16/18-1	X	X	X

Table 8.1 Summary of available well data in the Dagny-Alpha Central study area

Figure 8.3 Structural Setting of the Study Area. (a) Top Pre-rift (Top Sleipner) time structure map. The study area is interrupted by a series of salt and fault controlled highs. Three key features are described here in detail and highlighted on the map (Dagny, Alpha North and Alpha Central highs. (b) Study area fault map. All of the mapped faults offset the Jurassic succession. The well and seismic database is highlighted, along with the location of well transects.

8.4). The high is underlain (cored) by Zechstein Group evaporites. It is bound to the west by a broadly NW-SE trending, planar normal fault which downthrows to the WSW (Figs. 8.3b, 8.5b). To the north, the high deepens into a salt withdrawal-related structural low, whilst to the south a fault-bounded depocentre separates the structure from the Alpha North High (Fig. 8.5a). To the east the Dagny High merges with a NE-SW-trending, 12 km long salt wall which is bound to the north by a NW-dipping normal fault (labelled Ermintrude High in Figs. 8.3a, 8.5b). In detail, at the Upper Bathonian (top Sleipner Formation) structural level, the Dagny High is observed to consist of two structural culminations; a broader (2.4 by 3.4 km), circular one to the north and an elongate (1.9 by 1.4 km), more ovate one to the south. These are separated by a narrow (>1 km), shallow (>100 ms TWTT relief) saddle. The northern crest is highly faulted by a series of radial normal faults that offset Triassic, Jurassic and lowermost Cretaceous units (Figs. 8.3b, 8.5b).

8.4.2 Alpha North High

The N-S trending Alpha North High is 4 to 9 km in width and is more than 8.5 km in length (Fig. 8.3a). Published data show that this structure forms part of the Sleipner Vest Terrace and extends beyond the available dataset to the south (Husmo *et al.*, 2003; Folkestad and Satur, 2008). To the NW the high is separated from the Dagny High by a fault-bounded depocentre (*ca.* 2 km wide by 1.5 km long; Fig. 8.5a). The western flank is delimited by a series of broadly N-S trending normal faults that dip towards the west (Figs. 8.3, 8.5c) and which define the eastern margin of an elongate depocentre that separates the Alpha North High from the Alpha Central High to the west; these structures are discussed further below. To the east, the structure deepens off into the Sleipner Graben (Fig. 8.5c; Kane *et al.*, in review). The Sleipner Graben

Figure 8.4 Top Zechstein Time Structure Map. (a) Dagny-Alfa North study area. Present-day the highs are underlain by uplifted Zechstein Group evaporites. (b) Alpha Central study area. The high is underlain by a salt-pillow cored by evaporites of the Zechstein Group. (c) Regional index map highlighting the spatial relationship between the two study areas.

Figure 8.5 Structural Setting Dagny High region. Seismic and geoseismic sections presented. (a) NW-SE cross section through 15/5-2, Dagny and Alpha North highs. (b) E-W cross section through the Dagny and Ermintrude highs. Termination of reflection events has been highlighted and utilised to determine timing of structural development. The Middle to Upper Jurassic succession thins markedly and is truncated over the structural highs. Vertical black lines on the seismic section represent a join and change in direction of the presented line.

Figure 8.5 Structural Setting Dagny High region (cont.) (c) E-W cross section through the Alpha North High. The Middle to Upper Jurassic succession thins markedly across the high before thickening westwards into the Sleipner Graben system.

is approximately 30 km long and 6 km wide, and is bound to the east by a series of normal faults which are interpreted to have propagated up through the Zechstein Group salt layer and deformed the Jurassic succession (Figs. 8.3b, 8.5c). In detail, the crest of the Alpha North High is characterised by a series of sub-structures which are related to both halokinesis and normal faulting (Figs. 8.4; 8.5c). The faults trend N-S to NW-SE, are steep, downthrown to either the west or east, and exhibit throws ranging between 50 to 140 ms TWT T.

8.4.3 Alpha Central High

The Alpha Central High is a salt-cored anticline that plunges gently to the north (Fig. 8.4b). The anticline is *ca.* 6 km wide by at least 20 km long (it extends beyond the southern margin of the dataset; Fig. 8.3a). The anticline has a maximum amplitude of 400 ms (TWT) at Top Callovian level. The crest displays a variable orientation, trending broadly N-S in the south, switching to NE-SW along its central portion, before being represented by a NW-SE oriented spur at its northern tip. The western flank of the anticline is dissected by a series of NNW-SSE to N-S oriented, normal faults that dip to the west (Figs. 8.3b, 8.6b). These faults tip out downwards within the Zechstein Group interval and displace the Jurassic succession. The faults die out within the Upper Jurassic succession and do not offset the BCU, although this horizon is folded across the crest of the anticline (Fig. 8.6b). The anticline is separated from the Sleipner Vest High to the east by a N-S elongate, fault-bounded depocentre. This depocentre can be subdivided into northern and southern sub-basins separated by an E-W-trending normal fault which downthrows to the south (Fig. 8.3a). The southern sub basin is broadly ovate in geometry with a long axis which

Figure 8.6 Structural Setting of the Alpha Central High. Seismic and geoseismic sections are presented. (a) N-S (strike) section through anticline which plunges towards the north. (b) E-W section through anticline. The nose and flanks of the structure is dissected by a series of normal faults with low displacement that offset the Triassic and Jurassic successions. (For key see Fig. 8.5, position of sections shown in Fig. 8.3a).

trends E-W, and deepens towards the north, perpendicular to the dividing fault. The northern sub-basin is more elongate in geometry and trends NNE-SSW.

To the north, the nose of the anticline is cross-cut by a series of NE-SW trending faults that are downthrown to the north and offset Triassic-to-Upper-Jurassic strata (Figs. 8.3b; 8.6a). These faults coincide with deflection of the anticline crest to a NNE-SSW orientation. Offsets across these faults are typically low (20-50 ms TWT) with the greatest throw (*ca.* 100 ms TWT) being observed on the most northerly of the faults. To the north, the anticline is separated from another small, NE-SW trending high (the J-structure) by a low-relief saddle that also separates the major basin of the South Viking Graben to the west from a smaller depocentre to the NE.

A major N-S to NNE-SSW trending, westward dipping extensional fault which tips out upwards into the Zechstein layer has been interpreted to underlie the anticline (Fig. 8.6b; Kane and Jackson, 2008). The evaporites of the Zechstein Group effectively decouple sub-salt faulting from supra-salt deformation. Based on this spatial relationship, the Alpha Central High is interpreted to have formed as a forced fold or fault-propagation fold above the sub-salt normal fault (*cf.* Withjack *et al.*, 1990; Gawthorpe *et al.*, 1997). The low-offset faults to the north and west of the structure are interpreted to have accommodated extensional strains related to folding.

8.5 Temporal and Spatial Structural Development

It may be anticipated that halokinesis and normal faulting within the study area would have impacted the thickness, stratal architecture and facies distribution within

syn-tectonic sedimentary units such as the Hugin Formation (e.g. Stewart *et al.*, 1999; Gawthorpe and Leeder, 2000). To assess the impact of development of the structural features described above on deposition of the Hugin Formation, the study area has been subdivided into two regions; Dagny to Alpha North, and Alpha Central. Mapping of seismic surfaces (Fig. 8.2) has allowed isochrons to be generated across the structural features of interest, which help to constrain the timing of their development.

8.5.1 Structural Development of Dagny and Alpha North highs

Three key reflection events have been mapped (top Zechstein Group, Top Bathonian and BCU), which have been used to generate two isochrons. Note that due to poor internal reflectivity and thinning of the Hugin Formation (and the Jurassic succession in general) across the crest of the Dagny High, it has not been possible to accurately map reflection events within this interval.

Unit 1 - Top Zechstein to Top Bathonian

Description. This unit is Triassic to Lower Jurassic in age (i.e. Skagerrak, Smith Bank and Sleipner formations) and therefore predates the main period of Mesozoic rifting within the graben. The unit is up to 200 ms (TWTT) thick, with the greatest thickness recorded to the northwest and west of the study area within a series of depocentres that are characterised by irregular platform geometries and predominantly gradual thickness changes at their margins (labelled A-E; Fig. 8.7a). The unit thins overall eastwards and south-eastwards, with the thinnest section observed to the south of the study area to comprise two thinned regions (labelled F

Figure 8.7 Structural development of the Dagny to Alpha North Study area. (a) Unit 1 Isochron: Top Zechstein to Late Bathonian. (b) Unit 2 Isochron: Late Bathonian to Base Cretaceous Unconformity. The present day structural highs appear to have developed during this interval. For location of this study area see Fig. 8.4.

and G; Fig. 8.7a). The section thickens again to the south and east of this thinned region.

Interpretation. Overall thickening of the unit towards the west may indicate post-rift thermal subsidence associated with the Permo-Triassic rift event (Badley *et al.*, 1988) or Pre-Mesozoic movement on the GBFZ (Thomas and Coward, 1996). The significant local thickness variations are attributed to halokinesis of the Zechstein Group which was common within the North Sea Basin at this time. Halokinesis is interpreted to have been driven by regional extension during the Jurassic, superimposed on a template defined at least in part by passive diapirism caused by Triassic sediment loading (e.g. Central Graben; Hodgson *et al.*, 1992; Stewart *et al.*, 1999). An important observation is that the salt-cored Dagny and Ermintrude highs were actually areas of salt withdrawal at this time (Fig. 8.8a). Thinning to the south (F and G, Fig. 8.7a) coincides in part with the Sleipner Graben suggesting this feature had not started to subside. Thinning is also observed across the northern part of the Alpha North High indicating this structure was growing at this time (Fig. 8.8a).

Unit 2 – Late Middle and Upper Jurassic

Description. This unit comprises the Middle and Upper Jurassic successions (i.e. Hugin, Heather and Draupne formations) and was deposited during the main phase of Mesozoic rifting. The unit shows variable thickness across the study area and attains a maximum thickness of in excess of 150 ms TWTT (labelled I; Fig. 8.7b) within a fault-bounded basin fringing the western margin of the Dagny High. In addition, a locally thickened zone is developed to the east comprising four distinct

Figure 8.8 Schematic reconstruction of the structural development of the Dagny-Alpha North study area. Reconstruction based on sequentially flattened seismic section shown in Fig. 8.5 (see Fig. 8.3 for location). (a) Top Bathonian (pre-rift). Highlights thickness variation within the Triassic-Lower Jurassic succession due to halokinesis. The northern crest of the Dagny High is a depocentre at this time whilst the southern crest and Alpha North area are comparative highs. (b) Base Cretaceous Unconformity (post-rift). Thickening is observed over the faults highlighting they are active at this time. Alpha North remains a high whilst Dagny starts to grow. (c) Present day section. Highlights continued growth of salt cored highs into the Cretaceous.

regions (labelled J-M; Fig. 8.7b). The thinnest intervals are observed to the north and south of the area (labelled N-P; Fig. 8.7b).

Interpretation. Westwards thickening of the unit indicates movement on the GBFZ related to the main Mesozoic rift event (e.g. Husmo *et al.*, 2003). Thickening to the east (labelled J-M; Fig. 8.7b) coincides with the Sleipner Graben and indicates that the fault system bounding this feature was active and the area subsiding during deposition of this unit (Fig. 8.8b; Kane *et al.*, *in review*). Abrupt thickening to the SW of the Dagny High (labelled I; Fig. 8.7b) suggests that the major fault bounding its western flank was active at this time. Thinned zones (labelled N-P; Fig. 8.7b) coincide with the Dagny, Ermintrude and Alpha North highs indicating that all of these features were being uplifted during deposition of this unit (Fig. 8.8b). The geometry of these features as indicated by this isochron does not match their present day geometry indicating salt movement continued after deposition of this unit (Fig. 8.8c).

8.5.2 Structural development of the Alpha Central High

Seismic resolution and stratal unit thicknesses across the Alpha Central High are sufficient to enable five reflection events to be mapped within the Jurassic interval; Top Zechstein, Top Bathonian, Intra Callovian, Top Callovian and BCU. These have been used to produce four isochrones which help to constrain the growth of this high.

Unit 1 - Top Zechstein to Top Bathonian

Description. This unit comprises the Triassic to Late Bathonian interval (i.e. Skagerrak, Smith Bank and Sleipner formations) and ranges in thickness from 120

ms (TWTT) to the east to more than 800 ms (TWTT) in the south, north and NW (Fig. 8.9a). Four main, locally thickened sections, which define irregularly shaped depocentres with highly variable orientations, can be identified (labelled A-D; Fig. 8.9a), separated by a series of narrow (*ca.* 2 km), thinned regions of the unit (Fig. 8.9a). Within the main depocentres (A-D; Fig. 8.9a), localised thickness variations define a complex series of sub-depocentres. The unit thins markedly to the east central part of the study area (labelled E; Fig. 8.9a) and towards the southwest of the study area (labelled F; Fig. 8.9a).

Interpretation. The irregular shape, gradational boundaries and patchy distribution of thickened sections (A-D; Fig. 8.9a) are interpreted to reflect the formation of a series of salt-withdrawal basins due to movement of the underlying Zechstein Group evaporites (Fig. 8.10a). Importantly, the Alpha Central High had not begun to develop and was actually represented by a depocentre at this time. Marked thinning to the west and east (labelled E and F, Fig. 8.9a) is interpreted to reflect the growth of two salt-cored structural highs at this time which do not coincide with the Alpha Central and Sleipner Vest highs.

Unit 2 - Top Bathonian to Intra Callovian

Description. This unit is Early Callovian in age (i.e. the lower part of the Hugin Formation) and ranges in thickness from almost 0 ms (TWTT) to 140 ms TWTT (Fig. 8.9b). The Intra-Callovian reflection event cannot be mapped across the Sleipner Vest bounding fault to the east, due to a change in seismic character. Thickness variation within this unit is discrete although a series of E-W and NE-SW trending thickened zones are identified (labelled G to J; Fig. 8.9b). Within area J

Figure 8.9 Structural development of the Alpha Central High (ACH) study area. For location of this study area see Fig. 8.4. Key is the same as Fig. 8.7. (a) Unit 1: Top Zechstein to Top Bathonian reflection events (pre-rift). The Alpha Central High lies within a salt-withdrawal basin at this time whilst salt diapirs are interpreted to the east and west of the study area. (b) Unit 2: Top Bathonian to Intra Callovian reflection event. Minor thinning across crest of ACH to south of study area.

(c)

(d)

Figure 8.9 Structural development of the Alpha Central High (ACH) study area (cont.). (c) Unit 3: Intra Callovian to Late Callovian reflection events (early syn-rift). Minor thinning adjacent to the Alpha Central High. (d) Unit 4: Late Callovian to Base Cretaceous Unconformity reflection events (rift climax). Alpha Central High develops present day geometry and thinning towards Sleipner Vest High.

Figure 8.10 Schematic reconstruction of the structural development of the Alpha Cental study area. (a) Top Bathonian (pre-rift). Thickness variation within Triassic-Lower Jurassic succession due to halokinesis. ACH coincides with salt-withdrawal basin. (b) Top Callovian (early syn-rift). Thickness variation across active faults whilst anticline starting to uplift. (c) Base Cretaceous Unconformity (post-rift). Thickening across faults and thinning across crest of ACH observed therefore structure active at this time. (d) Present day section. Thinning and flexure of units highlights continued growth of salt cored anticline into the Cretaceous. Reconstruction based on Fig. 8.6b, for location see Fig. 8.3.

rapid thickening is observed across an E-W trending fault (Fig. 8.9b). An elongate, broadly N-S-trending thickened section is also observed to the east of the area (labelled K; Fig. 8.9b). The unit thins towards the southern margin of the study area and, importantly, an elongate zone of thinning trending approximately N-S is interpreted to the west of thickened regions H, I and J (Labelled L; Fig. 8.9b).

Interpretation. The elongate zone of thinning (labelled L; Fig. 8.9b) is broadly coincident with the crest of the Alpha Central High implying this feature was starting to develop at this time (Fig. 8.10b). Likewise, the elongate thickened section (labelled K; Fig. 8.9b) lies parallel to the Sleipner Vest bounding fault system and may indicate subsidence associated with this fault at this time; however, as Unit 2 cannot be interpreted across the fault, this cannot be equivocally determined. Between these two developing structures the series of poorly defined, irregularly shaped depocentres (labelled G-J; Fig. 8.9b) are interpreted to relate to salt-withdrawal and ongoing movement of the Zechstein Group evaporites (Fig. 8.10b). Rapid thickening observed within zone J coincides with an E-W trending fault suggesting it was active at this time.

Unit 3 - Intra Callovian to Top Callovian

Description. This unit is Late Callovian in age (upper part of the Hugin Formation) and varies in thickness from 10 to 180 ms (TWTT) (Fig. 8.9c). The unit is thickest in an E-W trending depocentre which runs across the centre of the study area and is subdivided in two (labelled M and N; Fig. 8.9c) by a narrow, NE-SW thinned zone. This thinned zone extends towards the SW of the study area (labelled P; Fig. 8.9c).

The unit thins markedly across a WNW-ESE-trending lineament in the north (labelled O; Fig. 8.9c) and also to the SE (labelled Q; Fig. 8.9c).

Interpretation. Thinning of the unit across an elongate feature that extends to the SW (labelled P; Fig. 8.9c) broadly coincides with the Alpha Central High implying the structure was continuing to develop at this time (Fig. 8.10c). Thinning to the SE (labelled Q) broadly coincides with the margins of the Sleipner Vest High implying this feature was also developing at this time (Fig. 8.10c). Marked thinning of the unit across a WNW-ESE lineament in the north of the study area coincides with a salt-pillow (Fig. 8.6a). This linear feature may have formed as a result of active diapirism above a sub-salt fault (e.g. Dooley, 2005) although this is not clearly imaged on seismic data. Thickening of the unit northwards (labelled M and N, Fig. 8.9c) is interpreted to reflect salt-withdrawal from this region, which may have migrated into the nearby growing Alpha Central and Sleipner Vest highs.

Unit 4 - Top Callovian to Base Cretaceous Unconformity

Description. This unit comprises the Upper Jurassic succession (Heather and Draupne formations) which was deposited during the main phase of Mesozoic rifting within the South Viking Graben (e.g. Thomas and Coward, 1996; Husmo *et al.*, 2003). The unit varies in thickness from less than 20 ms to in excess of 440 ms (TWTT). The unit is thinnest towards the SE (labelled R, Fig. 8.9d) and across an elongate, broadly N-S trending feature (labelled S; Fig. 8.9d) which exhibits a bulbous geometry, being narrower to the south and wider to the north. A series of linear features trending NE-SW and N-S cross cut this thinned region highlighting small scale thickness variations. The section thickens off into two major depocentres

to the east and west (labelled T and U; Fig. 8.9d). Zone T forms a narrow thickened area which appears to extend beyond the margins of the dataset to the west (Fig. 8.9d). Zone U is elongate and sinuous, comprising a series of linked thickened sections, the thickest of which lies to the south, trending E-W and is sharply delimited on its northern margin (Fig. 8.9d).

Interpretation. Thinning of the unit to the west (labelled S, Fig. 8.9d) coincides with the Alpha Central High indicating the anticline was developing at this time (Fig. 8.10c). Linear thickened zones that trend NE-SW and N-S across this feature are interpreted to relate to faults mapped across the nose and flanks of the anticline and support the interpretation that it was developing at this time. Thickening of the unit to the west (labelled T; Fig. 8.9d) may be indicative of continued movement on the GBFZ to the west or localised salt withdrawal (Fig. 8.10c). Thickening of the unit towards the east (labelled U; Fig. 8.9d) coincides with the depocentre that separates the Alpha Central High and Sleipner Vest High, indicating subsidence of this region, possibly due to salt-withdrawal and probable salt flow into the adjacent salt-cored highs. The sharply delimited E-W thickened zone within this feature coincides with a mapped E-W fault and is therefore interpreted to be fault-controlled. Thinning to the east suggests growth of the Sleipner Vest High during this time (Fig. 8.10c), an observation in agreement with published work (Husmo *et al.*, 2003; Satur and Folkestad, 2008).

8.5.3 Summary of structural development

Initial halokinesis within the study area pre-dated the main Jurassic rift phase and impacted the distribution of both the Triassic and Lower Jurassic intervals. This Pre-

Mesozoic salt movement may reflect passive diapirism caused by differential loading during the Triassic, a mechanism proposed for diapirism within the Central Graben at this time (e.g. Hodgson *et al.*, 1992; Stewart *et al.*, 1999). At this time, the bulk of the Dagny, Ermintrude, Alpha North and Alpha Central highs were not positive structural features, but were actually areas of salt withdrawal above which basins developed and thickened successions were deposited (Figs. 8.8; 8.10). However, discrete thinning over portions of the southern part of the Dagny High and the Alpha North High may highlight these features, were also topographic features at this time.

During the Middle Jurassic, the previously established template of salt-cored highs and withdrawal basins began to change, and faults within the area initiated or were reactivated. The salt-withdrawal basin in the vicinity of the Dagny High was locally inverted to form the geometry observed present-day, and continued to be uplifted into the earliest Cretaceous (Fig. 8.8c). Similarly, the depocentre located beneath the Alpha Central High was inverted, experiencing minor growth during the Callovian whilst the main phase of growth was during the Late Jurassic (Fig. 8.9). This reorganisation of the structural style coincides with the main period of rifting within the South Viking Graben, suggesting this regional extension was a driving force behind salt-movement within the area. Due to poor seismic imaging of the sub-salt succession it is unclear how important sub-salt structure may have been on the development or localisation of the Jurassic salt-cored highs. Importantly, the majority of the major salt-cored highs within the study area were evolving during the deposition of the Middle to Upper Jurassic successions. Therefore, it can be predicted that these features will have had an impact on stratigraphic architecture and facies distributions within the Hugin Formation.

8.6 Sedimentology and Sequence Stratigraphy of the Hugin Formation

Facies and sequence stratigraphic analysis of the Hugin Formation and coeval strata has been undertaken to investigate the impact of the syn-depositional structural development (described above) on the sedimentology and stratigraphic architecture of the Callovian succession. A detailed analysis of the sedimentology and sequence-stratigraphy of the Hugin Formation has previously been described by Kieft *et al.* (*in review*) and this, in addition to the regional palaeogeography (Fig. 8.11), is summarised here.

8.6.1 Sedimentology

Core and wireline-log data have been integrated from nine wells. Twelve facies have been identified that have been grouped into five facies associations; (i) bay-fill; (ii) shoreface; (iii) mouth bar; (iv) fluvio-tidal channel fill; and (v) coastal plain. These facies associations are summarised below (see also Table 8.2 and Fig. 8.12).

The Hugin Formation comprises a series of stacked, progradational parasequences that record deposition within a wave- and tide-influenced deltaic system and laterally-equivalent strand plain (Fig. 8.11). Restricted-marine shales and tidal dunes are interpreted to form within interdistributary bays of the delta, with a strong tidal signature attributed to tidal amplification within the narrow graben (e.g. Husmo *et al.*, 2003; Folkestad and Satur, 2008). These bay-fill units are incised by channel-fills that record variable fluvial to tidal influence (Fig. 8.12), with tidal influence interpreted to be greater in down-dip locations. Where tidally-influenced channels

Figure 8.11 Early Callovian depositional setting of the Hugin Formation. The Dagny to Alpha Central study area is highlighted. At this time, Hugin Formation deposition occurred within a progradational deltaic environment and attached strandplain. The delta system formed the main feeder system to the basin whilst localised drainage from the Utsira High is also inferred (from Kieft *et al.*, in review).

F.A.	Facies	Description	Bioturbation
Interdistributary bay fill	Restricted marine shale	Parallel-laminated mudstone with rare siltstone laminae, locally sideritic. Monospecific assemblage of disarticulated bivalves oriented parallel to bedding and concentrated into thick (60-205 cm) bioclastic-rich beds. Rare gastropods.	BI = 0-2 (Pl)
	Sandy shallow bay-fill	Fine- to medium-grained sandstone. Moderately to well sorted. Wavy bedding, high-angle cross-bedding and occasional wave ripples. Calcite cement and soft sediment deformation locally observed. Disarticulated bivalves throughout, sometimes concentrated in thin beds (<10 cm).	BI = 0-4 (Pa, Dip, Ro/Cy?, Pl, Op)
Shoreface	Distal lower shoreface	Upward-coarsening, siltstone to very fine-grained, moderately sorted sandstone. Shell debris, quartz granules and angular carbonaceous fragments throughout. Primary sedimentary structures are poorly preserved, but contains rare thin (30-75 cm), sharp-based, very fine- to fine-grained cross-laminated sandstone beds.	BI = 4-5 (Pa, Ch, Op, Pl, Ter, As?)
	Proximal lower shoreface	Very fine- to fine-grained, moderately to well sorted sandstone. Fine-grained, carbonaceous laminae throughout. Dominantly low-angle cross bedding, with subordinate parallel lamination. Carbonate-cemented nodules.	BI = 3-4 (Op)
	Upper shoreface	Fine- to medium-grained, well sorted, coarsening-upwards, cross-laminated sandstone. Discrete laminae of coarser sediment, carbonaceous fragments and shell fragments occur throughout.	BI = 0* (crypto/Mac?)
Mouth Bar	Distal mouth bar	Highly bioturbated siltstone to very fine-grained sandstone with stringers and thin (5-20 cm) beds of rounded granules.	BI = 4-5 (Pa, Ch, Op, P)
	Proximal mouth bar	Very fine- to fine-grained sandstone with low-angle cross-bedding. Stringers and thin (5-15 cm) beds of rounded granules occur throughout. Monospecific trace fossil assemblage (<i>Ophiomorpha</i>).	BI = 0-2 (Op)
Fluvial-tidal channel fill	Tidally-influenced fluvial channel-fill sandstone	Fining-upward very coarse- to fine-grained sandstone with possible bidirectional cross-sets and swaley cross-stratification. Stacked fining-upward beds with sparse coarse-grained stringers. Sharp, possibly erosive contacts at individual bed bases.	BI = 0-1 (Sk, ET)
	Fluvial channel-fill sandstone	Fining-upward coarse- to medium-grained sandstone with unidirectional cross-bedding, carbonaceous laminae and fragments throughout. Sharp, possibly erosive contacts at individual bed bases.	BI = 0-1 (ET)
Coastal Plain	Coal	Coal and carbonaceous shale. Basal contact typically parallel although may be undulose. Diagenetic pyrite nodules are recognised.	BI = 0
	Coastal plain fines	Interbedded dark grey mudclaystone and siltstone to very fine-grained sandstone. Wavy to non-parallel lamination penetrated and disrupted by rooting	BI = 3-4 (roots)
Open Marine	Open marine shale	Mudstone with rare siltstone beds. Parallel lamination, soft sediment deformation developed locally at bed tops. Rare thin (<100 cm), finely laminated, unbioturbated beds of very fine-grained sandstone.	BI = 3-5, occ. 0 (Pl, Op, Ch, Zo, Tae, As. Te)

Table 8.2 Summarised facies of the Hugin Formation within the study area. Bioturbation fabric is described using the Bioturbation Index (BI) scheme of Taylor & Goldring (1993). Note that the bioturbation scheme cannot be used to describe cryptobioturbation, where sediment is pervasively bioturbated with little to no distortion of primary structures (e.g. facies SF3). Individual trace fossils, listed in brackets, are abbreviated as follows: Te = *Teichichnus*, Sk = *Skolithos*, Ar = *Arenicolites*, Pa = *Paleophycus*, Op = *Ophiomorpha*, Dip = *Diplocraterion*, Ros = *Rosselia*, Cy = *Cylindrichnus*, Pl = *Planolites*, Ch = *Chondrites*, Ter = *Terebellina* (Schaub. *Fria*), As = *Asterosoma*, Mac = *Macaronichnus*, ET = Escape traces, crypto = cryptobioturbation.

Figure 8.12 Facies summary of the Study Area. (a) Shoreface Facies Association (15/5-1). (b) Mouthbar Facies Association (15/6-2R). (c) Coastal Plain Facies Association (well 15/8-1). (d) Fluvio-Tidal Channel-fill Facies Association (15/8-1).

incise coastal plain deposits their basal surface may be interpreted to represent a candidate flooding or transgressive surface, or autogenic migration of the channel system. To the north, these units pass laterally into shoreface deposits of the strand plain, which interfinger with mouth bar deposits interpreted to record the locations of deltaic distributary channels (Fig. 8.12). This close association of facies is interpreted to reflect autogenic migration of the fluvially-fed distributary channels across the strand plain. In places however the relationship between the facies is interpreted to have sequence stratigraphic significance and be interpreted as a flooding surface e.g. where distal mouth bar deposits overlie upper shoreface deposits. Preservation of coarse-grained stringers within the mouth bar facies implies limited wave reworking of fluvially supplied sediment, possibly due to wave suppression within the narrow graben (e.g. Richards, 1991; Folkestad and Satur, 2008). Coastal plain deposits (Fig. 8.12) cap shallow-marine parasequences and record subaerial deposition in response to shoreline progradation. Open marine shales are observed locally to interfinger with the deltaic and strand plain deposits and are interpreted to be equivalent to the Heather Formation. These shales are observed at the base of parasequences overlying the basal bounding flooding surface and represent periods of significant marine transgression and flooding.

8.6.2 Sequence stratigraphic correlations

Three well-based correlation panels have been generated through the study area utilising the facies associations and sequence stratigraphic surfaces described above and described more fully by Kieft *et al.* (in review). Correlation of sequence stratigraphic surfaces between the wells has been constrained by a newly revised, high-resolution biostratigraphical and chronostratigraphical scheme based on the

StatoilHydro ASA regional database. Four maximum flooding surfaces can be correlated regionally subdividing the studied succession into four main units and identified using the nomenclature of Partington *et al.*, 1993). Locally correlatable flooding surfaces have also been identified which permit further subdivision of the succession into nine parasequences in the Dagny-Alpha North High area and ten parasequences in the Alpha Central High area.

Dagny to Alpha North Correlation

Nine parasequences have been identified and locally correlated in the Dagny-Alpha North area (Parasequences 1-9; Fig. 8.13). Initial shallow marine deposition within this area (parasequences 1 and 2; Fig. 8.13) is restricted to the north of the correlation panel (wells 15/5-1 and 15/6-2R) and passes southwards into time equivalent coastal plain deposits (Sleipner Formation; wells 15/6-6 and 15/6-5). These shallow marine units comprise interfingered shoreface and mouth bar deposits recording fluvial influence to the shoreline at this time. The succession penetrated on the northern flank of the Dagny High (well 15/5-1), is significantly thinner and biostratigraphy suggests only parasequence 1 is represented at this location, bound by locally developed unconformities (Fig. 8.13). The section penetrated on the southern flank of the Dagny High (well 15/6-2R) is greatly expanded, no biostratigraphical zones are recognised to be missing and the unit has conformable basal and upper contacts. Where present, parasequence 2 is capped by a thick, regionally correlatable coal (wells 15/6-2R, 15/6-6, 15/6-5; Fig. 8.13). Deposition of parasequences 3 and 4 saw the initiation of shallow marine deposition on the flank of the Alpha-North High (15/6-5) with a thick (*ca.* 30m) shoreface sandstone (parasequence 3) overlain by a progradational mouth bar unit (parasequence 4; Fig.

Figure 8.13 Dagny to Alpha North Correlation Panel. Within the lower parasequences, a fluvial signature is observed throughout and facies can be correlated between wells. In the upper parasequences, shoreface sandstones pass laterally to open marine shales over short distances. This is due to structural development during Hugin Formation deposition and is both salt- (15/5-1) and fault-controlled (15/6-6). See Fig. 8.3b for location of line and Fig. 8.15 for key.

8.13). This succession correlates with a stacked lower shoreface which is only locally-developed on the southern flank of the Dagny High (parasequences 3 and 4; well 15/6-2R; Fig. 8.13). These parasequences are absent from the crest of the Alpha North High (well 15/6-6) and northern flank of the Dagny High (well 15/5-1) where they are represented by unconformities.

Shoreface sandstones developed in 15/6-6 within parasequence 5 represent the first time shallow marine deposition occurred towards the crest of the Alpha North High. These sandstones pinch out rapidly (<2 km) down dip to the southeast into a thickened succession of open marine shales (Heather Formation; well 15/6-5; Fig. 8.13), with these abundant fines probably being supplied from the Sleipner Vest delta to the south. The shoreface sandstones are interpreted to thin and possibly onlap onto the Alpha North High itself to the WNW. Parasequence 5 comprises open marine shales on the southern flanks of the Dagny High (well 15/6-2R) whilst it is absent on the northern flank of the Dagny High and represented by an unconformity (well 15/5-1; Fig. 8.13).

Parasequences 6 and 7 saw aggradation of locally developed shoreface sandstones towards the crest of the Alpha North High (well 15/6-6) and at the northern flank of the Dagny High (well 15/5-1; Fig. 8.13). These sandstones pinched out over comparatively short distances (<2 and 5 km respectively) into open marine shales deposited on the flank of the Alpha North High (well 15/6-5) and the southern flank of the Dagny High (well 15/6-2R). Open marine conditions spread towards the crest of the Alpha North High (well 15/6-6) during parasequence 8 whilst shallow marine shoreface deposition was limited to the northern crest of the Dagny High (well 15/5-

1; Fig. 8.13). Shoreface sandstone deposition persisted on the northern flank of the Dagny High (well 15/5-1) into parasequence 9 whilst the southern flank of the Dagny High and the Alpha North High were flooded and open marine shales (Heather Formation) deposited (Fig. 8.13). This parasequence is capped by a locally developed unconformity on the northern flank of the Dagny High (well 15/5-1) which is sharply overlain by much younger shales.

Locally developed unconformities at the northern flank of the Dagny High (well 15/5-1) are interpreted to represent periods of uplift and subsequent erosion across the northern flank of this high. It is not possible to correlate these surfaces to the southern flank of the high (well 15/6-2R) suggesting that uplift was localised and implying that the southern flank of the high was not uplifted to the same extent as the northern flank during Hugin Formation deposition. The locally developed unconformity near the crest of the Alpha North High (well 15/6-6) is also interpreted to represent a period of uplift and subsequent erosion related to rotation of the fault block into the Sleipner Graben to the east. These relationships indicate a more complex timing of structural development than indicated by seismic mapping (section 8.5.1). Aggradation of shoreface sandstones at the northern flank of the Dagny High (well 15/5-1) and towards the crest of the Alpha North High (well 15/6-5) implies these structures were uplifting during Hugin Formation deposition forming topographical features along which shallow marine sedimentation was focussed (discussed further section 8.6.3).

Alpha Central Correlation

In the Alpha Central area the studied succession has been subdivided into ten parasequences (parasequences 1 to 10; Fig. 8.14). Parasequence 1 comprises restricted marine claystones sharply overlain by fluvio-tidal channel fill and capped by coastal plain deposits. The parasequence thickens from *ca.* 32 m in the west (crest proximal well 16/18-1) to *ca.* 42 m in the east (flank of Alpha Central High, well 15/8-1; Fig. 8.14) but this thickness change is not associated with a facies shift. This parasequence is overlain by an isopachous, thick (*ca.* 75 m) succession of open marine shales (Heather Formation) containing a sharp-based, blocky sandstone bed (> 5m thick, parasequence 2; Fig. 8.14). These sandstones may represent deep marine deposition, possibly as part of a turbidite or mass flow complex offshore of a coeval delta front. This interpretation may imply that the sandstones have a candidate sequence boundary at their base (e.g. Posamentier and Vail, 1988). Biostratigraphical data proposes a maximum flooding surface capping this parasequence (Fig. 8.15).

During deposition of parasequence 3, open marine shale deposition continued across the Alpha Central High (wells 16/18-1 and 15/8-1; Fig. 8.14). These open marine shales are sharply overlain by a fluvio-tidal channel complex interfingered with coastal plain deposits with a sequence boundary interpreted at the base. This is overlain by the bay-fill to coal-bearing coastal plain successions of parasequences 4 to 6. Each of these parasequences is observed to thicken from the west, near the crest of the Alpha Central High (well 16/18-1) to the east, down the flank of the Alpha Central High (Well 15/8-1). Facies associations can be correlated between the two wells despite the thickness variation, however several coal beds (2-4 m thick) and

Figure 8.14 Alpha Central Correlation. The wireline log character of individual facies is not distinctive enough to enable their recognition away from core control although facies associations have been interpreted. Early flooding of the region (parasequence 2) is followed by repeated progradation and transgression of a deltaic shoreline with variable tidal influence. See Fig. 8.15 for key.

channelised sandstone units (>5 m thick) are only locally developed on the flank of the high (e.g. channelised sandstone at the base of parasequence 5, well 15/8-1; Fig. 8.14). This implies differential subsidence between the two wells, and increased accommodation space was developed on the flank of the high at this time. This observation agrees with seismic mapping (Section 8.5.2) which indicated the Alpha Central High was developing throughout Hugin Formation deposition.

Shoreface sandstones are developed across the high during the early part of parasequence 7 and are overlain by interfingered fluvio-tidal channel and coastal plain deposits whilst parasequence 8 records bay-fill deposition across the structure, capped by a coal (Fig. 8.14). Both parasequences are broadly isopachous across the structure. Parasequence 9 thickens from west (*ca.* 13m) to east (*ca.* 20m) across the structure and comprises bay-fill sediments capped by fluvial-tidal channel fill. Within the expanded bay-fill section to the east, i.e. on the flank of the structure (well 15/8-1), a channelised sandstone is identified which can not be correlated to the west. A 15m thick shoreface sandstone succession is developed within parasequence 10 but, similarly to the channelised unit within the bay-fill of parasequence 9, is only identified to the east of the structure (well 15/8-1). A lack of biostratigraphical constraint at this depth means it is not possible to determine if parasequence 10 is penetrated towards the crest of the structure (i.e. within well 16/18-1) and simply represented by a different facies association (i.e. open marine shales overlying parasequence 9, well 16/18-1; Fig. 8.14). Alternatively, the shoreface sandstones could pinch out up-dip towards the crest recording differential subsidence across the structure at this time i.e. more on the flank compared to the crest. Seismic reflection events have been identified which onlap towards the crest

of the structure between the two wells (Fig. 8.6b), providing supporting evidence to the interpretation that parasequence 10 onlaps the flank of the developing high. Alternatively the shoreface sandstones may have been deposited and subsequently uplifted and eroded prior to deposition of the overlying open marine shale.

Whilst facies can typically be correlated across the 2 km between the wells, channel-fill sandstones, coal beds and shoreface sandstones are identified which are limited to the eastern flank of the Alpha Central High (Fig. 8.14). This localised development of distinct facies may be interpreted to reflect differential subsidence across the Alpha Central High, with greater subsidence to the east on the flank of the structure (well 15/8-1) and less subsidence close to the crest of the structure (well 16/18-1). This relationship is interpreted to have led to focussing of channel systems onto the flanks of the high causing deposition of the locally developed channelised sandstones interpreted (parasequences 5, 6 and 9; Fig. 8.14). However, these eastwards thickening parasequences are interfingered with parasequences which are isopachous across the structure, indicating that uplift of the structure was an episodic event, rather than continuous throughout Hugin Formation deposition.

Regional Correlation Panel

This correlation traverses from the Alpha Central High in the south (well 15/8-1) across the Alpha North (well 15/6-5) and Dagny highs (well 15/5-1) to well 15/5-2 in the north (Figs. 8.3b; 8.15). The studied interval has been divided into four units, bounded by maximum flooding surfaces, which can be correlated along the axis of the South Viking Graben (Units 1-4; Fig. 8.15). Within these units, flooding surfaces

Figure 8.15 Regional Strike Correlation. The structural setting of each of the wells is highlighted. The studied succession can be subdivided into four units on the basis of regionally correlatable maximum flooding surfaces, these four units coorespond to those identified on Figs. 8.13 and 8.14. This correlation highlights the complex pattern of flooding within the study area.

have been identified although these can not be correlated between the Alpha Central and Dagny-Alpha North areas.

During Unit 1, coastal plain deposition was predominant across the southern part of the study area (wells 15/8-1 and 15/6-5) and passed northwards through shallow marine sandstones (well 15/5-1) into open marine shales (well 15/5-2). In addition, shallow marine Hugin Formation sediments are observed at the top of this unit in the southernmost well (15/8-1), indicating transgression of the area was not a simple, north to south younging event as proposed for the graben as a whole (e.g. Mitchener *et al.*, 1992; Rattey and Hayward, 1993). It should be noted however that wells 15/8-1 and 15/5-2 lie to the west of the study area; proximal to the graben axis therefore flooding of these areas may be linked to deepening within the axial portion of the graben. The complex flooding of the area is further evidenced during Unit 2, during which open marine shale deposition was established across the Alpha Central High in the south (well 15/8-1; Fig. 8.15) whilst also continuing to the north (well 15/5-2). Shallow marine deposition persisted for the majority of this unit on the flank of the Alpha North High (well 15/6-5) whilst the unit is not penetrated on the northern flank of the Dagny High where it is correlated to a locally developed unconformity (well 15/5-1). During Unit 3, shallow marine conditions were re-established across the Alpha Central High in the south (well 15/8-1; Fig. 8.15) indicating progradation of the shoreline following the major transgression during Unit 2. These shallow marine units pass northwards into open marine shales across the flank of the Alpha North High (well 15/6-5) and can be correlated to an aggradational succession of shoreface sandstones on the northern flank of the Dagny High (Well 15/5-1) which pass northwards into further open marine deposits (well 15/5-2). This spatial

relationship highlights the spatial complexity associated with flooding of this area and also implies the deltaic system developed across the Alpha Central High is not linked to the shoreface established at Dagny at this time. This pattern of sedimentation persisted into Unit 4, with shallow marine deposition limited to the Alpha Central High (well 15/8-1) and the northern flank of the Dagny High (well 15/5-1; Fig. 8.15). In the upper part of Unit 4, the southern part of the study area is flooded (well 15/8-1), whilst the locally persistent shoreface sandstones (well 15/5-1) are capped by an unconformity and sharply overlain by younger open marine shales.

The complex spatial distribution of flooding within this area is interpreted to reflect structural control on sedimentation patterns. The Alpha Central, Alpha North and Dagny Highs are all interpreted to have been actively growing during deposition of the Hugin Formation (see also section 8.5). Development of the Dagny High is interpreted to have focussed aggradational shoreface deposition along its northern flank (well 15/5-1) whilst subsidence of the flank of the Alpha North High into the Sleipner Graben enabled comparatively early flooding in this location (well 15/6-5). Seemingly the structural development of the Alpha Central High was not sufficient to prevent early transgression of the feature (Unit 2) however sediment supply was later sufficient for the shoreline to prograde back into the basin and re-establish shallow marine conditions in this area (units 3 and 4). Early flooding towards the south and north may be explained by the location of these wells which are both proximal to the axis of the South Viking Graben, a region interpreted to have had marine influence throughout the Callovian (coincident with deposition of the Hugin Formation) and even as far back as the Bathonian (e.g. Cockings *et al.*, 1992).

8.6.3 Palaeogeographical reconstruction

The well correlations described above highlight marked facies and thickness variability within the Hugin Formation, in addition to marked diachroneity in drowning of shoreline systems towards the top of the unit. Based on these data and the structural analysis provided previously, three palaeogeographical reconstructions are presented for the study area that record maximum regression of the shoreline beneath regionally correlatable maximum flooding surfaces. These correspond to the Early Callovian (top Unit 1, beneath MFS A), approximately Middle Callovian (top Unit 2; beneath MFS J44) and Late Callovian (top Unit 3; beneath MFS J46). The well data discussed in detail above has been integrated with published data from the Sleipner Vest and Øst fields to constrain the reconstructions in the southeast corner of the study area (Milner and Olsen, 1998; Folkestad and Satur, 2008).

Early Callovian (MFS A)

During the Early Callovian (Unit 1. MFS A), coastal plain deposition (Sleipner Formation) prevailed to the east and south of the study area (i.e. 15/6-4, 15/6-7, 15/6-6; Tjemsland, 2007) whilst open marine shale deposition (Heather Formation) prevailed to the north (i.e. 15/5-2) (Fig. 8.16a). Although not observed at this level within study area wells, a deltaic system is interpreted to lie between the Alpha Central and Dagny-Alpha North areas which contributed to the fluvial signature observed within Early Callovian shallow marine deposits preserved on the northern flank of the Dagny High (well 15/5-1, see section 8.6.2). This deltaic system is thought to form part of the Sleipner Delta system, an axial feeder system which is interpreted to be the main source of sediment to the graben at this time. MFS A can not be identified on the northern flank of the Dagny High (well 15/5-1; Fig. 8.16a)

Figure 8.16 Palaeogeographical reconstruction of deposition within the Dagny-Alpha Central study area. (a) Early Callovian. Majority of study area is site for coastal plain deposition whilst to the north open marine deposition prevails. (b) Mid Callovian. Open marine shales deposited across most of the study area.

therefore this area was either a site of non-deposition at this time or the units were deposited and subsequently eroded. Shoreface deposits are inferred to front the extensive coastal plain away from the fluvial influence (Fig. 8.16a).

Middle Callovian (MFS J44)

The Middle Callovian (Unit 2, MFS J44) saw a major regional flooding event and the deposition of open marine shales (Heather Formation) across the majority of the study area (Fig. 8.16b). Shallow marine and coastal plain deposition at this time was limited to the flank of the Utsira High to the east (e.g. well 15/6-7) and to the south east of the study area across the Sleipner Vest Terrace (Folkestad and Satur, 2008). The Sleipner Delta is interpreted to have backstepped towards the SE since the Early Callovian (*cf.* Figs. 8.16a). Isolated shoreface sandstones are seen developed near the crest of the Alpha North High (well 15/6-6) and are interpreted to be focused along a footwall crest that rotated with subsidence of the Sleipner Graben to the east at this time (Fig. 8.16b). These sandstones may be sourced by; (i) long shore drift of material from the Sleipner Delta to the south or the Utsira High to the east, however sands sourced from the Utsira High may have struggled to bypass the developing Sleipner Graben; (ii) eroded material transported down dip from the crest of the structure; or (iii) re-working of underlying units. This time interval (MFS J 44) is represented by an unconformity on the northern flank of the Dagny High (well 15/5-1) and at the crest of the Ermintrude High (well 15/6-4; Tjemsland, 2007). Deposition may have occurred in this region and been subsequently eroded. Alternatively, the salt-cored highs may have formed a topographic feature at this time which may have focussed shallow marine sedimentation along its flanks, similar to the relationship observed at Alpha North. In this case, shoreface

sandstones are interpreted to fringe the northern margin of this structure and sourced by similar mechanisms to those discussed for the isolated sandstones preserved at Alpha North.

Late Callovian (MFS J46)

During the Late Callovian (MFS J46) open marine shale deposition (Heather Formation) prevailed over the majority of the study area, including the crest of the Alpha North High (well 15/6-6; Fig. 8.16c). Isolated shoreface sandstones were deposited focussed on the northern flank of the Dagny High (well 15/5-1) whilst the southern flank of the Dagny High, which was experiencing diminished rates of uplift, was a site of deeper-water, open marine deposition. The extent and source of these sandstones is uncertain. They may extend to the NE, forming a spit system sourced by long shore drift of material from the Utsira High. Alternatively they may be more localised in extent and sourced by re-working of underlying deposits or localised erosion across the crest of the Dagny High. Open marine shale deposition across the crest of the Ermintrude High to the NE (well 15/6-4) may be evidence to support the latter interpretation, although a lack of well penetrations to the north of this means a spit system sourced from the Utsira High (similar to that proposed during the Middle Callovian) can not be discounted. In the south of the study area, the Sleipner Vest delta system has prograded eastwards (since the Middle Callovian) across the Alpha Central High (wells 16/18-1 and 15/8-1; Figs. 8.16b,c).

8.7 Discussion

8.7.1 Structural development within the area

Significant thickness variations are observed across the studied area within the Triassic to Lower Jurassic succession (Section 8.5). Predominantly irregular in shape, these thickness changes are attributed to withdrawal and uplift of the underlying, Permian-age, evaporite-rich Zechstein Group which commenced in the Triassic, and continued throughout the Jurassic and into the lowermost Cretaceous (section 8.5). Mobilisation of the Zechstein Group layer during the Triassic has been recognised throughout the North Sea basin (e.g. Pegrum and Ljones, 1984; Frostrick, 1992; Hodgson *et al.*, 1992). Regional extension led to remobilisation and continued halokinesis of the Zechstein Group during the Jurassic (e.g. Pegrum and Ljones, 1984; Fraser *et al.*, 2003). Within the study area, Jurassic extension led to a reorganised pattern of halokinesis and Triassic-age salt-withdrawal basins were inverted to create the structural highs observed. This pattern of structural development is different to that observed within the West Central Graben and Central Shelf of the North Sea (e.g. Fraser *et al.*, 2003). In this area, during the Jurassic Triassic depocentres continued to sink until they grounded and the underlying salt sheet had migrated completely into adjacent salt-cored highs. In this region, Jurassic-age depocentres coincide with Triassic depocentres, but are also located above collapsing salt-cored structures and within areas of salt-dissolution (e.g. Wakefield *et al.*, 1993; Fraser *et al.*, 2003). The key differences between salt-structural styles observed within the South Viking Graben when compared to those described from other regions of the North Permian Salt Basin could be attributed to several factors; (i) salt pillows within the studied area are not recognised to have

been exposed at the sediment surface at any time, as they were in the Central Graben (e.g. Wakefield *et al.*, 1993). This means creation of topography as a result of salt-dissolution and preferential erosion could not have happened; (ii) Salt-withdrawal basins are not recognised to have “grounded” consequently deflation of flanking salt-cored highs would not have occurred (cf. “pod-interpod” model, Hodgson *et al.*, 1992). However sub-salt seismic resolution within the study area makes it difficult to recognise base salt and this interpretation is speculative.

8.7.2 Structure and sedimentation

Shallow marine deposits are recognised to fringe local structural highs within the study area. These structural highs were formed as a result of halokinesis (e.g. well 15/5-1, Dagny High) and rift-related extension with associated fault-block rotation (e.g. well 15/6-6, Alpha North High). Localisation of shallow marine systems fringing salt-cored structures has been previously documented within the Central Graben of the North Sea (e.g. Stewart *et al.*, 1999; Fraser *et al.*, 2003) and within the La Popa Basin, Mexico (e.g. Aschoff and Giles, 2005). Similarly, localisation of shallow marine systems fringing fault-block highs has also been previously documented within the North Sea Basin (e.g. Fraser *et al.*, 2003). These shallow marine systems are observed to pinch-out rapidly down-dip into open marine shales, over comparable scales (over 2-5 km) to those recognised in other salt-basins (e.g. Johnson *et al.*, 1986; Aschoff and Giles, 2005). The aggradational stacking pattern of shallow-marine sediments against these structural highs (e.g. well 15/5-1) may be interpreted to reflect a steep (i.e. near-vertical) transgressive shoreline trajectory (Cattaneo and Steel, 2003), despite the retreat recorded elsewhere within the basin. The source of the shallow marine sandstones is uncertain as they may be locally re-

worked from the exposed crest of the structural high or fed via longshore drift from a regional source. These two alternatives have significant implications for the volume of sediment available, with the former model implying smaller volumes compared to the latter. It is proposed that the topographic feature formed during uplift of the Dagny High may have prevented backstepping of the shoreline during transgression and led to aggradational stacking on the flanks of the evolving structure.

Channelised units within the Hugin Formation are also interpreted to preferentially flank growing salt-cored structures (e.g. Alpha Central High, well 15/8-1). This relationship has been recognised within the Porterillos Formation in the La Popa Basin, Mexico, within which a deltaic system was confined to an evolving syncline (Shelley and Lawton, 2005). In this example, the preferential stacking of sediments within the syncline is interpreted to have caused migration of underlying salt into the flanking structures, creating a positive deposition-subsidence feedback loop and driving passive diapirism (e.g. Hodgson *et al.*, 1992; Rowan and Weimer, 1998; Shelley and Lawton, 2005). It is unclear whether the focussing of channels onto the flank of the Alpha Central Anticline would have provided sufficient sediment load to drive development of the structure but it may have contributed, along with propagation of an underlying sub-salt fault.

CHAPTER 9

Palaeogeographical Reconstruction of the Middle Jurassic Transgression within the South Viking Graben

This chapter focuses on the Bathonian-Callovian net-transgressive succession within the South Viking Graben, with the specific aim of putting the detailed results presented in Chapters 7 and 8 into a more regional context. Whilst the main focus of this study has been the Norwegian Sector of the South Viking Graben (Quad 15), so as to provide a regional context it is necessary to review and synthesise previous studies which have focused on coeval tectonic evolution and deposition within the UK Sector of the basin. In this chapter the temporal and spatial evolution of key structures and the sedimentological and stratigraphic variability associated with their development are summarised based on detailed analysis within the main study area (Norwegian Quadrant 15). Through the recognition and correlation of key biostratigraphical markers, palaeogeographical reconstructions are presented that integrate these detailed analyses with regional data.

9.1 Introduction

Middle Jurassic sedimentation within the South Viking Graben is ascribed lithostratigraphically to the Bathonian to Oxfordian-age Vestland Group which comprises the continental Sleipner Formation (Pentland Formation in the UK sector) and shallow marine Hugin Formation (Fig. 9.1; Vollset and Doré, 1984; Husmo *et al.*, 2003). The Hugin Formation is an overall transgressive unit that was deposited

Figure 9.1 Jurassic Lithostratigraphy of the South Viking Graben. The interval of interest to this study is highlighted.

during rift-related subsidence and subsequent flooding of the South Viking Graben during the north to south retreat of the Brent Delta megasequence (Harris and Fowler, 1987; Cockings *et al.*, 1992; Underhill and Partington, 1993). At this time, the South Viking Graben formed a long (100-120 km), narrow (30 km) marine embayment located to the west of the Stord Basin (Fig. 9.2; Husmo *et al.*, 2003).

Continuing relative sea-level rise during the Late Jurassic, driven by rift-related subsidence and eustasy, led to continued deepening of the basin and deposition of the shelfal and deep-marine deposits of the Viking Group (Heather and Draupne/Kimmeridge Clay formations; Bradshaw *et al.*, 1992; Fraser *et al.*, 2003). Previous studies have suggested that shallow marine sediments were spatially restricted to the south and east of the basin due to the asymmetrical geometry of the developing half graben (e.g. Harris and Fowler, 1987; Færseth and Pederstad, 1988; Fält *et al.*, 1989). However, although less well-documented, coeval shallow-marine sediments have been penetrated within several fields along the western margin of the graben, including Beinn, Kingfisher and the 'T-block' cluster of fields (e.g. Chandler and Dickinson, 2003; Gambaro and Donagemma, 2003; Spence and Kreutz, 2003).

9.2 Previous Studies

The Middle Jurassic Hugin Formation has been the target of exploration within the Norwegian sector of the South Viking Graben since the 1960s and forms an economically important reservoir unit within several fields (e.g. Sleipner Vest and Øst). Due to its economic significance, several previous studies have investigated its distribution and the Middle Jurassic tectono-stratigraphic evolution of the South

Figure 9.2 Mid-Late Callovian Palaeoenvironmental Map of the North Sea Basin (Husmo *et al.*, 2003). Collapse of the Mid North Sea Dome and reactivation of older fault systems led to widespread transgression within the North Sea Basin. Offshore marine sedimentation dominated in the axis of basins whilst coastal plain and shallow marine deposition was focussed around the flanks. The South Viking Graben formed a narrow, elongate marine embayment between the Utsira High and East Shetland Platform. Norwegian Quadrant 15 which formed the focus of this study is highlighted within the South Viking Graben.

Viking Graben in general (e.g. Cockings *et al.*, 1992; Varadi *et al.*, 1998; Husmo *et al.*, 2003).

Sedimentological and sequence stratigraphic analyses were undertaken by Cockings *et al.* (1992) and a series of palaeogeographical reconstructions of the Middle Jurassic succession were presented (Fig. 9.3). Seven main depositional environments were recognised; (i) fluvial; (ii) delta front; (iii) delta plain; (iv) shoreface; (v) shelf; (vi), open marine; and (vii) basin floor turbidites (Cockings *et al.*, 1992). The influence of the developing rift system on patterns of Middle Jurassic sedimentation was recognised, with the earliest marine influence observed during the Bathonian to be restricted to the western portion of the graben and coinciding with movement on the Graben Bounding Fault Zone (GBFZ; Thomas and Coward, 1996). The graben then progressively flooded from north to south and west to east, driven by a combination of subsidence caused by slip on the GBFZ and eustatic sea-level rise (Haq *et al.*, 1987).

Deposition of the Hugin Formation within a barrier shoreline setting was proposed by Harris and Fowler (1987) and Varadi *et al.* (1998). In particular, Harris and Fowler (1987) interpreted a regionally extensive barrier shoreline that extended from the Sleipner Vest Terrace in the south of the graben to the Gudrun Terrace in the north (Fig. 9.4). Varadi *et al.* (1988) also stressed the importance of fluvial-deltaic systems sourced from the Utsira High in their depositional model and in a series of palaeogeographical reconstructions which focussed on the northern part of the South Viking Graben (Fig. 9.5). They interpreted several drainage systems sourced from the Utsira High which supplied sediment to a shallow marine coastline fringing the

Figure 9.3 Schematic regional palaeogeographical reconstruction of Middle Jurassic depositional environments in the Bathonian to Callovian (a) Bathonian; (b) Early Callovian; (c) Mid-Callovian; (d) Late Callovian (Redrawn from Cockings *et al.*, 1992).

Figure 9.4 Schematic regional palaeogeographical reconstruction of depositional environments during the Callovian (Redrawn from Harris and Fowler, 1987). This interpretation is comparatively 'low-resolution' compared to the other regional palaeogeographical reconstructions presented.

Figure 9.5 Palaeogeographical reconstruction of depositional environments during the Middle Jurassic within Quadrant 15. (a) Late Bathonian; (b) Early Callovian; (c) Mid Callovian; (d) Late Callovian (Redrawn from Varadi *et al.*, 1998). Maps indicate palaeogeographical reconstruction at the point of maximum progradation of the system immediately prior to the subsequent flooding event.

high. In addition, they linked these shallow marine deposits to the deeper water turbidite deposits observed within the UK sector (Ling Sandstone Member, Cockings *et al.*, 1992), thereby implying that these turbidites were fed from the eastern basin margin.

In contrast to the models proposed by Cockings *et al.* (1992), Varadi *et al.* (1998) interpret that an open marine environment persisted across a consistent area of the graben throughout the Middle and Late Jurassic. Both Cockings *et al.* (1992) and Varadi *et al.* (1998) recognise the importance of local variation in subsidence and uplift in affecting sedimentation patterns and flooding of the graben. Localised variation in rift-related subsidence along the western flank of the graben formed focal points for shallow marine sedimentation, for example, the slowly subsiding Thelma Terrace formed a focal point for deltaic sedimentation (Cockings *et al.*, 1992), whilst differential salt movement was interpreted to have influenced depositional patterns, in particular the position of fluvial systems (Varadi *et al.*, 1998).

9.3 Regional Structural Setting

To the east the South Viking Graben shallows onto the Utsira High whilst to the west it deepens towards the GBFZ and the Fladen Ground Spur (Fig. 9.6). The South Viking Graben initially formed as a result of rifting during the Permo-Triassic but became reactivated during the Mesozoic rift event (e.g. Ziegler, 1975, 1992; Steel, 1993; Bartholomew *et al.*, 1993; Thomas and Coward, 1996; Glennie, 1998; Coward *et al.*, 2003; Coward *et al.*, 2003). The dominant N-S to NE-SW structural trend

Figure 9.6 Structural map of the South Viking Graben (after Thomas and Coward, 1996). The South Viking Graben is a NNE-SSW trending half graben, bound to the west by a series of ESE dipping faults. To the east, the graben shallows onto the Utsira High. The hangingwall is dissected by a series of normal faults, the majority of which developed antithetic to the main graben bounding fault, and two Permian-age transfer zones that trend northwest-southeast.

observed within the South Viking Graben relates to this later Mesozoic rift event (e.g. this study; Pegrum and Ljones, 1984; Thomas and Coward, 1996; Jackson and Larsen, 2008; 2009; Kane *et al.*, in review). A series of NW-SE trending transfer zones also dissect the graben and are interpreted to relate in some way to the Permian rift event (Fig. 9.6; e.g. Northern and Southern Transfer Zones; Thomas and Coward, 1996). These zones coincide with pronounced eastwards steps along the GBFZ and also coincide with a change in fault orientation along the western flank of the Utsira High (Thomas and Coward, 1996). The GBFZ exhibits variable structural styles along its length, from broad fault terraces at the Thelma Terrace in the south (Harris and Fowler, 1987; Cherry, 1993) to areas of more complex fault geometry such as the South Brae Sub-basin (Cockings *et al.*, 1992).

9.3.1 Structural Elements of the Study Area (Quadrant 15)

The Hugin Formation was deposited during a period of both rift- and salt-related tectonics. Therefore it is critical to understand the temporal and spatial development of these structures to understand the stratigraphic development of the unit. Seismic mapping has been undertaken to establish the detailed structural framework and, more importantly, timing of structural development within the main study area (Norwegian Quad 15; Chapter 7, section 7.5; Chapter 8, section 8.5). Due to diachroneity of rifting and flooding within the basin it is difficult to produce a regional pre-rift map to illustrate these structures; therefore the key structural elements within the study area are here defined based on a regional top syn-rift (Base Cretaceous Unconformity) map (Fig. 9.7). Six main structural elements are identified in the Norwegian sector of the South Viking Graben.

Figure 9.7 Middle Jurassic Structural Elements of the Study Area. (a) Time structure map at the Base Cretaceous Unconformity (BCU) level i.e. top syn-rift. (b) Structural features of study area. Basinal areas are identified in blue whilst relative highs are identified in green.

The Utsira High extends into the eastern part of the main study area and forms a structural high that trends broadly NE-SW (Fig. 9.7). The western margin of the high is delimited by Caledonian-age normal faults that trend NNE-SSW and WNW-ESE, and which downthrow to the west (Pegrum and Ljones, 1984). Jurassic strata are missing on the crest and proximal margins of the Utsira High, where Cretaceous and Tertiary strata directly overlie heavily eroded granitic basement (e.g. Zanella and Coward, 2003). As a result of this missing section and the large period of time it represents, timing the growth of the high is difficult. However, several authors suggest that the structure formed a major topographic feature during both the Permo-Triassic and Mesozoic rift events (Cope *et al.*, 1992; Zanella and Coward, 2003).

A series of normal fault-bounded, broadly N-S elongate rotated blocks lie immediately to the west of the Utsira High. The most northerly of these blocks is the Gudrun Terrace which is bound to the west by the Brynhild Fault system (Fig. 9.7b and described in detail in Chapter 7, section 7.4). Although the Brynhild Fault system predominantly trends NNE-SSW, the northernmost segment trends NW-SE; this is interpreted to be due to interference from the Permian-age Northern Transfer Zone (*sensu* Thomas and Coward, 1996). The Gudrun Terrace passes westwards into the Vilje Sub-basin, within which the prominent, NE-SW to NW-SE trending high of the Brynhild Rollover Anticline is developed (Chapter 7, section 7.4). This anticline, which is bound to the west by the Gudrun Fault system, lies to the west of a hangingwall syncline developed in the immediate hangingwall of the Brynhild Fault system. These fault systems developed as antithetic structures to the GBFZ in response to gravity-driven sliding of supra-salt units following Jurassic extension and hangingwall tilting (Chapter 7; Thomas and Coward, 1996; Jackson and Larsen,

2008). Seismic isochrons show that these structures initiated during the Late Bathonian-Early Callovian, with initial displacement focussed on discrete, short (ca. 2 km) fault segments (Chapter 7, section 7.5; Jackson and Larsen, 2008; 2009). Segment linkage and maximum displacement was achieved during the Late Jurassic (Late Callovian-Volgian) during the main period of rift-related extension (Chapter 7; Thomas and Coward, 1996; Jackson and Larsen, 2009).

To the south, the Utsira High is separated from the main axis of the South Viking Graben by two salt-cored structural blocks: the Sleipner Vest Terrace and the Alpha Central High (Fig. 9.7b; Chapter 8; Pegrum and Ljones, 1984). Seismic data indicate that these structures were most active during the Middle and Late Jurassic (Chapter 8, section 8.5; Pegrum and Ljones, 1984; Kane *et al.*, in review), although the Sleipner Vest Terrace may have initially developed during the Permo-Triassic. The Sleipner Vest Terrace is itself separated from the Utsira High by the N-S-trending Sleipner Graben, which is bound on its eastern margin by a west-dipping normal fault zone (Fig. 9.7b; Kane *et al.*, in review). To the west, The Sleipner Vest Terrace is separated from the Alpha Central High by a synclinal low.

The Ve Sub-basin is located to the north of the Sleipner Vest Terrace and Alpha Central Anticline and is internally partitioned by several salt-cored structural highs. The Glitne High lies to the west of the Gudrun Terrace, within the axis of the South Viking Graben and immediately south of the Brynhild Rollover Anticline (Fig. 9.7). Triassic and Jurassic units thin towards this structure and become progressively eroded, such that at the crest of the structure Cretaceous units directly overlie Upper Permian Zechstein Group evaporites (Chapter 7, section 7.5). Progressive thinning of

units towards this structure implies that the lack of deposition across its crest was due to growth and subaerial exposure of the high during both Triassic and Jurassic times. Alternatively, any sediments deposited on the crest may have been subsequently uplifted and eroded.

The Dagny-Ermintrude High lies to the SE of the Glitne High and forms an elongate, broadly NE-SW-trending structural high within the Ve Sub-basin (Fig. 9.7). Seismic isochron mapping provides important insights into the temporal development of this structure and indicates that it had a complex history (Chapter 8; section 8.5, 8.6). During the Triassic-Early Jurassic, this structural high coincided with a salt-withdrawal basin. However, during the Middle Jurassic to Early Cretaceous, ‘inversion’ of this basin occurred due to re-migration of evaporites of the Zechstein Group (Chapter 8, section 8.5). In addition, seismic-stratigraphic observations indicate that growth of the structure during the Middle and Late Jurassic was, in detail, very complex, with different portions of the high growing at different times (Chapter 8, section 8.5).

The Alpha Central High lies to the SW of the Dagny-Ermintrude High and to the west of the Sleipner Vest Terrace (Fig. 9.7). This elongate, salt-cored anticline trends broadly N-S and plunges gently to the north. The western flank of the anticline is offset by a series of relatively small-displacement (ca. 100 m) normal faults that dip towards the west; the eastern flank of the structure dips eastwards into the hangingwall of a fault-bounded depocentre. This anticline is interpreted to have developed as an extensional forced fold (cf. Withjack and Callaway, 2000; Richardson *et al.*, 2005) related to upward propagation of a N-S-trending, basement-

involved normal fault beneath the Zechstein Group evaporites during the Middle and Upper Jurassic (Chapter 8, section 8.5; Kane and Jackson, 2008).

9.4 Sedimentology and Stratigraphic Evolution of the Middle Jurassic w within Quad 15

Detailed sedimentological analysis was integrated with biostratigraphical analysis (provided by StatoilHydro ASA) to establish a depositional model and sequence-stratigraphic framework for the Middle Jurassic succession within the Norwegian sector of the South Viking graben (Quad 15; Chapter 5; Kieft *et al.*, in review). Six facies associations were recognised that record deposition in three main, temporally-related, shallow-marine environments; (i) restricted marine embayment and associated barrier shoreface (bay-fill and shoreface facies associations); (ii) wave-influenced strandplain (shoreface facies association); and (iii) tidally-influenced delta (fluvio-tidal channel fill and mouth bar facies association). These facies associations and their component facies are summarised in Table 9.1. Whilst the three main shallow marine environments are temporally related, they were geographically disparate, with their locations being strongly influenced by the prevailing structural template; (i) a restricted marine embayment environment, which was controlled by both fault- and salt-related structures, developed on the Gudrun Terrace; (ii) the strandplain environment dominated around the salt-controlled Dagny-Ermintrude High; and (iii) the tidally-influenced deltaic environment was focussed to the south across and on the margins of the Alpha Central High and Sleipner Vest Terraces (this is discussed more in section 9.5 and highlighted in figs. 9.8 and 9.9). Within each of these coupled structural-depositional systems, the local stratigraphic development was controlled by the markedly different tectonic styles

Facies Assoc.	Facies	Brief Description	Interpretation
Bay-fill	BF1	Parallel-laminated mudstone with rare siltstone laminae and monospecific assemblage of disarticulated bivalves.	Restricted marine shale
	BF2	Interbedded claystone and very fine-grained sandstone. Non-parallel and wavy lamination, trough cross-stratification and hummocky cross-stratification in sandstone beds.	Muddy bay fill
	BF3	Fine- to medium-grained sandstone with rare mudstone drapes. Wavy bedding, high-angle cross-bedding and occasional wave ripples.	Sandy bay fill
	BF4	Coarse- to fine-grained sandstone with fining- and coarsening-upward motifs. Sparse sedimentary structures include ripples and mudstone drapes. Disarticulated bivalves and shell fragments common, concentrated into thin beds (<40 cm).	Marine channel fill sand
Shoreface	SF1	Upward-coarsening, siltstone to very fine-grained, moderately sorted sandstone. Primary sedimentary structures are poorly preserved due to high levels of diverse bioturbation.	Distal lower shoreface
	SF2	Very fine- to fine-grained, moderately to well sorted sandstone. Fine-grained, carbonaceous laminae throughout. Dominantly low-angle cross bedding, with subordinate parallel lamination.	Proximal lower shoreface
	SF3	Fine- to medium-grained, well sorted, coarsening-upwards, cross-laminated sandstone.	Upper shoreface
Mouth bar	MB1	Highly bioturbated siltstone to very fine-grained sandstone with stringers and thin (5-20 cm) beds of rounded granules.	Distal mouth bar
	MB2	Very fine- to fine-grained sandstone with low-angle cross-bedding. Stringers and thin (5-15 cm) beds of rounded granules occur throughout.	Proximal mouth bar
Fluvio-tidal channel fill	FTC1	Fining-upward very coarse- to fine-grained sandstone with possible bidirectional cross-sets and swaley cross-stratification.	Tidally-influenced fluvial channel-fill sandstone
	FTC2	Fining-upward coarse- to medium-grained sandstone with unidirectional cross-bedding, carbonaceous laminae and fragments throughout.	Fluvial channel-fill sandstone
Coastal plain	CP1	Coal and carbonaceous shale.	Coal
	CP2	Interbedded dark grey mudstone and siltstone to very fine-grained sandstone. Wavy to non-parallel lamination penetrated and disrupted by roots.	Coastal plain fines
Open marine	OM	Mudstone with rare siltstone beds. Parallel lamination, soft sediment deformation developed locally at bed tops.	Open marine shale

Table 9.1 Summary of the facies and facies associations identified within the Hugin Formation, Quad 15, South Viking Graben

Figure 9.8 Regional Chronostratigraphical Diagram. The diagram highlights overall north to south and west to east flooding of the study area, plus additional structurally controlled complexities that result in local hiatuses and anomalous facies distributions. Parasequences and relationships within the gross time intervals are schematically represented and should be compared to detailed correlations in Chapters 7 and 8 (Figs. 7.11-13 & 8.13-15). Coastal plain deposits frequently cap the shallow marine parasequences but have not been represented for simplicity.

(i.e. fault vs. salt control) and distance from the main axial basin feeder system located to the south (Kieft *et al.*, in review).

Flooding of the South Viking Graben during the Jurassic phase of rifting was strongly diachronous and spatially complex (Fig. 9.8). Overall, the age of marine flooding became younger from north to south, along the graben axis, and also from west to east up the hangingwall dip slope; this is discussed further in section 9.5. A linear zone of deep water became established and persisted in the basin axis during the Middle to Late Jurassic, and deep marine turbidite sandstones were deposited within the Heather Formation in this location; importantly, deposition of these turbidites was coeval with deposition of the Hugin Formation at the basin margins (Cockings *et al.*, 1992; Spence and Kreutz, 2003). Variably termed the Ling Sandstone Member, Bruce Sandstone or Heather Sandstone (Cockings *et al.*, 1992; Richards *et al.*, 1993; Husmo *et al.*, 2003), these deepwater turbidite sandstones are interpreted to have been derived either from a point source located on the Fladen Ground Spur (Cockings *et al.*, 1992) or to form the distal part of a linked delta-shelf submarine fan system sourced directly from a coeval Hugin shoreline situated to the southeast (Varadi *et al.*, 1998; Spence and Kreutz, 2003). Critically, these deep marine sandstones pre-date (by ca. 6 Myr) deposition of the Brae Formation, a conglomerate and sand-prone fan system developed in the immediate hangingwall of the GBFZ sourced by erosion of the rapidly uplifting footwall (Fraser *et al.*, 2003).

9.5 Regional data and basin-scale palaeogeographical reconstructions for the South Viking Graben

The detailed sedimentological and stratigraphic analysis of the Hugin Formation summarised above was supplemented with core data from five wells within the UK sector, in addition to previously published data (e.g. core descriptions, wireline log data) from the UK sector. This allowed basin-scale palaeogeographical maps to be constructed for the South Viking Graben. Key to permitting the integration of the detailed analysis from Quad 15 with published data was the recognition of key stratal surfaces which could be correlated across the basin. Maximum flooding surfaces were identified using high resolution biostratigraphical and well data within the main study area (Tjemsland, 2007). These were related to the Middle Jurassic sequence stratigraphic framework established for the Central and Northern North Sea by Partington *et al.* (1993). This framework has been utilised by workers within the UK sector of the graben, thereby enabling the work undertaken in Quad 15 as part of this study to be correlated to that which has been previously published for the UK sector.

Although well data and biostratigraphical data are relatively common for the Middle Jurassic in Quad 15, wells within the UK sector typically target the Upper Jurassic Brae Formation with the Middle Jurassic succession not being penetrated. This results in a poorer understanding of the sedimentological and stratigraphic framework of the Hugin Formation in the UK sector of the South Viking Graben than in the Norwegian Sector. These variations in temporal and spatial resolution of datasets notwithstanding, four palaeogeographical reconstructions are presented that show distribution of depositional environments in response to marine flooding

highlighted within Figure 9.8. Maps have been constructed which show maximum regression beneath regionally correlatable maximum flooding surfaces as defined by StatoilHydro's in-house biostratigraphical framework correlated to the North Sea-wide sequence-stratigraphic nomenclature of Partington *et al.* (1993); J36, A, J44, J45 (Fig. 4.2).

9.5.1 Late Bathonian (MFS J36)

Where penetrated, the Late Bathonian succession typically comprises continental deposits of the Sleipner Formation (Fig. 9.9a). The exception to this is observed on the eastern flank of the Vilje Sub-basin and western part of the Gudrun Terrace where restricted marine bay-fill units were deposited within a back-stepping, linear, N-S trending barrier and associated back-barrier bay shoreline system (wells 15/3-2R, 15/3-3; Fig. 9.9a). A similar bay-fill unit, which interfingered with continental deposits of the Sleipner Formation and is identified as Early Bathonian in age, was also observed to the south (well 15/5-2; Tjemsland, 2007). Bathonian flooding and the establishment of shallow marine (i.e. Hugin-type) conditions has previously been documented within axial parts of the graben (Cockings *et al.*, 1992; Varadi *et al.*, 1998) and is interpreted to reflect relatively rapid deepening associated with the initial stages of movement on the GBFZ (e.g. Williams, 1993). Shallow marine deposits have also been identified within the Bathonian succession beneath the Kingfisher Field near the south-western margin of the Gudrun Terrace (Fig. 9.9a; Spence and Kreutz, 2003).

Figure 9.9 Palaeogeographical reconstructions for deposition within the South Viking Graben during the Middle Jurassic. (a) Bathonian MFS J36. (b) Early Callovian MFS A. (c) Middle Callovian MFS J44. (d) Late Callovian MFS J46. Overall the graben floods from north to south and west to east.

KEY

9.5.2 Early Callovian (MFS A)

During the Early Callovian, relative sea-level continued to rise in response to ongoing fault-related basin subsidence. As a result, marine conditions spread eastwards and southwards into the South Viking Graben (Fig. 9.9b). The northern part of the Gudrun Terrace was flooded and barrier islands, fronted by wave-dominated shorefaces, developed (Kieft *et al.*, in review; Chapter 7, section 7.6). These barrier islands sheltered a restricted marine embayment to the east, whilst open marine (Heather Formation) deposition occurred to the west toward the basin axis (Fig. 9.9b). The shoreface sandstones are interpreted to have been supplied by southward-directed longshore drift of sediment derived from the heavily eroded Utsira High. Longshore drift is interpreted to have been generated by oblique wave approach from the open marine basin to the north or northwest (Kieft *et al.*, in review). Localised drainage into the embayment is inferred from the Utsira High to the east, down the hangingwall dip slope. The embayment was, in part, delimited by salt-cored highs and the Brynhild Rollover Anticline, all of which were actively growing during the Bathonian to Late Callovian (Chapter 7, section 7.5). Coastal plain sedimentation persisted to the south on the western flank of the Utsira High, behind a strandplain shoreline fronted by a wave dominated shoreface (Fig. 9.9b). To the southeast, across the Sleipner Vest Terrace and the Alpha Central High, fluvial-deltaic deposition occurred within the tidally-influenced Sleipner Delta system (Chapter 8, section 8.4; Folkestad and Satur, 2008; Kieft *et al.*, in review). These two structural features are interpreted to have begun to develop during this time, although the lack of obvious facies variation across them implies that sediment supply was sufficient to keep pace with the developing structurally controlled topography (Chapter 8; section 8.5). Coeval deltaic and estuarine or paralic deposits have been

interpreted on the Thelma Terrace to the SW of the area (Cockings *et al.*, 1992, Gambaro and Donagemma, 2003). A major marine incursion occurred in the axis of the basin at this time, in response to continued activity on the GBFZ. This resulted in the establishment of relatively deep-water, open marine conditions (Heather Formation), which extended eastwards up the hangingwall towards well 15/5-2 (Cockings *et al.*, 1992; Varadi *et al.*, 1998; Kieft *et al.*, in review). Within the western part of the Ve Sub-basin (Fig. 9.7b), locally developed argillaceous sandstones within the South Brae Sub-basin are interpreted to be sourced by submarine re-working of sand from fault block crests near the western basin margin (Cockings *et al.*, 1992).

9.5.3 Middle Callovian (MFS J44)

By the Middle Callovian, the majority of the South Viking Graben was flooded and open marine conditions (Heather Formation) prevailed eastwards across the Gudrun Terrace and southwards across the uplifting Alpha Central High (Fig. 9.9c). The Sleipner Delta had backstepped to the southeast and fluvial-deltaic deposits were developed across the Sleipner Vest region at this time. Coastal plain and shallow marine environments were restricted to the eastern basin margin. Due to the open geometry of the graben at this time, this shallow marine belt is inferred to have been characterised by a wave-dominated strandplain, rather than a tidally-influenced barrier-bay environment (Fig. 9.9c).

Importantly a series of “isolated” shoreface sandstones developed within the Dagny and Alpha North areas at this time (Chapter 8; Kieft *et al.*, in review). Seismic mapping indicates the presence and growth of a number of salt-cored highs and fault

blocks within this area; these topographic highs may have been interpreted to have provided a focal point for shoreface sandstones to have been deposited or reworked (Chapter 8; section 8.7). The source of these sandstones is uncertain; they may have been sourced from re-working of the underlying units due to localised erosion across the crests of the salt and/or fault-controlled topographic features, or by longshore drift that transported sediment from the north.

In the UK sector, a series of intra-basin tilted fault blocks, which are parallel to the GBFZ and form part of the Thelma Terrace, formed relative structural highs or terraces located within the immediate hangingwall of the basin-bounding fault zone. As a result, shallow marine sedimentation persisted in this location along the western margin of the South Viking Graben (Cockings *et al.*, 1992; Gambaro and Donagemma, 2003). Deltaic sediments are also recognised to the NW, within the South Brae Sub-basin (Fig. 9.6), and were probably sourced from relay ramp to the NW (Cockings *et al.*, 1992; C. Turner, *pers. comm.*). To the west and basinward of these localised areas of shallow marine deposition, deep marine turbidites of the Ling Sandstone Member were deposited in the South Brae Sub-basin, immediately southwards of a major 'jog' in the GBFZ (e.g. Cockings *et al.*, 1992; Knox and Cordey, 1993). The source of the Ling Sandstone Member within block 16/8 is uncertain and various interpretations have been proposed, including; (i) sourcing from the SE and transport in a N-NW direction (Spence and Kreutz, 2003); (ii) sourcing from coeval Hugin Formation shorelines to the E and SE (Cockings *et al.*, 1992; Varadi *et al.*, 1998; Fraser *et al.*, 2003); or (iii) sourcing from a deltaic shoreline located on the relay ramp to the NW (Kane and Jackson, 2008).

9.5.4 Late Callovian (MFS J46)

Overall, depositional patterns during the Late Callovian were similar to those documented during the Middle Callovian. Open marine deposition (Heather Formation) dominated across the South Viking Graben whilst coastal plain and shallow marine deposition was restricted to its flanks (Fig. 9.9d). Shoreface sandstone deposition was also locally focussed on the northern flank of the Dagny High which, at this time, represented a major intra-basin topographic high (section 8.5.1). The Sleipner Delta system had prograded north- and westwards into the basin over the previously flooded Alpha Central High. This structural feature continued to grow and is interpreted to have focussed fluvial-tidal channels along its flanks (Chapter 8, section 8.6.2). In the UK sector, shallow marine deltaic or estuarine sedimentation continued on the Thelma Terrace (Cockings *et al.*, 1992; Gambaro and Donagemma, 2003). The Thelma and Sleipner Vest deltas are interpreted to be separated by an open marine seaway (Cockings *et al.*, 1992), although sediment supplied during the progradation of the Sleipner Delta system may have been sufficient to infill this area. Related to this suggestion, it is noted that fluvial-tidal sediments at this approximate stratigraphic level have been interpreted within well 16/18-2 which lies between the deltas; biostratigraphical data is not available to accurately date these sediments, however. In the South Brae Sub-basin, turbidites of the Ling Sandstone Member, which were also identified in the Middle Callovian, are interpreted to have become more widespread, with their distribution being controlled, in part, by syn-depositional faulting (Cockings *et al.*, 1992).

9.6 Discussion

9.6.1 Tectonic controls on Middle Jurassic deposition within the South Viking Graben

Previous studies have suggested that rift-related faulting strongly influenced sediment routing, gross thickness patterns, facies distributions and facies belt orientations within the Hugin Formation (Harris and Fowler, 1987; Cockings *et al.*, 1992; Sneider *et al.*, 1995; Milner and Olsen, 1998; Varadi *et al.*, 1998). The results of this study suggest that halokinesis of the underlying Zechstein Group was a more significant control on deposition than previously thought, especially on the hangingwall dipslope and within the basin axis. In particular, salt-cored topographic highs are interpreted to have; (i) partly delimited the spatial extent of a shallow marine embayment that developed on the Gudrun Terrace during the early stages of rifting; (ii) formed prominent intra-basin focal points for sandstone deposition within the basin at a time when relative sea-level was high and shallow marine deposition was otherwise limited to the flanks of the graben; and (iii) resulted in focusing of channels within the Sleipner Delta system, at the margins of developing structures.

In contrast, the Gudrun and Brynhild fault zones, located in the north of the basin where halokinesis was more subdued, are interpreted to have had less of an impact on sedimentation patterns. The Brynhild North fault zone and associated rollover are the only part of this fault array interpreted to have had any impact. The rest of the fault array was only undergoing relatively low rates of displacement during this rift-initiation phase, compared to the high sedimentation rate associated with deposition of the shallow marine Hugin Formation. Consequently, sediment supply was

sufficient to infill the developing accommodation. A similar relationship is observed within the Middle Jurassic syn-rift Tarbert Formation in the North Viking Graben (Davies *et al.*, 2000; Hampson *et al.*, 2004).

With increased rift-related subsidence during the Middle and Late Callovian and continued transgression of the graben, tectonically-generated accommodation space outpaced sediment supply and prominent intra-basin highs became important focal points for shallow marine deposition. For example, shoreface sandstones were deposited focussed on the flanks of the salt-cored Dagny-Ermintrude High and pinched out downdip into open marine shales. The change in the balance between sediment supply and accommodation within the graben at this time meant that rift-related faulting was also observed to impact facies distributions within the shallow-marine succession. For example, on the Alpha North High shoreface sandstones are focussed along the footwall crest of an active normal fault and pinch out rapidly (ca. 2 km) downdip into open marine shales. Rift-related faulting also had a marked impact on facies distributions around the GBFZ, where sediment supply from the Fladen Ground Spur was limited and locally restricted. Displacement along the GBFZ outpaced this limited sediment supply and shallow marine sedimentation was only able to persist where rates of tectonically-generated subsidence were less (e.g. deltaic sediments which are locally developed on the Thelma Terrace in the south and on the relay ramp to the northwest of the South Brae Sub-basin).

The marked control of halokinesis and normal faulting on deposition of the Hugin Formation is reminiscent of the net-transgressive Fulmar Formation deposited in the Central Graben of the North Sea (Hodgson *et al.*, 1992; Stewart *et al.*, 1999). The

Central Graben, in a similar manner to the South Viking Graben, lies at the margins of the North Permian evaporite basin (Ziegler, 1990; Thomas and Coward, 1996; Coward *et al.*, 2003). Shallow marine units within the Fulmar Formation were focussed on the flanks of salt- and fault-controlled syn-depositional topographic highs (cf. the sandstones flanking the Dagny-Ermintrude Highs in the South Viking Graben) and within valley systems related to salt withdrawal and dissolution (Wakefield *et al.*, 1993; Stewart *et al.*, 1999; Fraser *et al.*, 2003). The Hugin Formation also shares many stratigraphic similarities with the syn-rift Tarbert Formation (Bajocian-Bathonian) of the North Viking Graben, although deposition of this unit was not influenced by syn-depositional halokinesis. In a similar manner to the Hugin Formation adjacent to the Gudrun Fault, the Tarbert Formation can be correlated across active fault segments with the lack of facies shift indicating no discernable increase in water depth from footwall to hangingwall. This implies that the rate of deposition was broadly equal to the rate of subsidence (McLeod *et al.*, 2002). However, structurally-controlled depocentres did form a focus for stacked channel systems similar to that proposed for the Sleipner Delta system on the flank of the Alpha Central High (Davies *et al.*, 2000; McLeod *et al.*, 2002). The same relationship may also be predicted within the rollover syncline in the hangingwall of the Brynhild Fault system; however, this region is currently undrilled.

The overall syn-rift setting may have contributed to the tidal signatures observed within the Hugin Formation, especially to the south of the South Viking Graben within the Sleipner Delta system. Apparent tidal dominance or frequent tidal signatures may be typical of syn-rift successions where fault-controlled topography creates confined basins with a small wave fetch and potentially serves to amplify

tidal currents (e.g. Ravnås and Bondevik, 1997; Ravnås and Steel, 1998; Carr *et al.*, 2003). In apparent contradiction, the development of a barrier shoreline on the Gudrun Terrace implies that a degree of wave action was required to generate longshore drift. However, the geographical setting of the two depositional settings is markedly different; the Gudrun Terrace lies to the north of the South Viking Graben within a more open basin, whilst the Sleipner Delta system was developed in a more enclosed location at the southern limit of the embryonic South Viking Graben. In addition to the overall basin geometry, the Alpha Central High, Sleipner Vest Terrace and Sleipner Graben system are interpreted to have been actively growing during shallow-marine deposition within the southern part of the basin. These structures may have served to further partition the graben into a series of narrow, broadly N-S trending embayments which dampened wave energy and enhanced the relative importance of tides within this part of the graben through funnelling and shoaling effects.

9.6.2 Comparisons to existing models for Middle Jurassic sedimentation in the South Viking Graben

The overall patterns of deposition interpreted during this study are broadly similar to those previously published, with progressive flooding of the South Viking Graben from north to south and west to east (e.g. Cockings *et al.*, 1992). However, biostratigraphical analysis available to this study implies an earlier (Bathonian) onset of shallow marine sedimentation, especially on the Gudrun Terrace. The importance of rift- and halokinesis-related differential subsidence on depositional patterns has also been previously recognised (e.g. Cockings *et al.*, 1992; Varadi *et al.*, 1992). However, the distinct geographical partitioning of depositional environments due to

syn-depositional structural growth as interpreted in this study has not been previously recognised. In contrast, previous studies have suggested more regionally correlatable facies belts (cf. the barrier shoreline interpreted by Harris and Fowler, 1987; Fig. 9.4). The difference in spatial distribution of the interpreted shoreline types between this study and those previously presented may reflect, in part, the more numerous well and biostratigraphical data available to the present study. Alternatively, these differences may reflect the integrated, seismic and well-based procedure used here.

Whilst it is inferred in this study that the heavily eroded Utsira High was an important source of shallow marine sediments which were eventually transported southwards by longshore drift, distinct E-W-trending drainage systems, also derived from the Utsira High, have previously been interpreted by Varadi *et al.* (1998); these systems have not been convincingly recognised in this study. Thick, apparently channelised sandstones within the area of the Gudrun Field may be interpreted as fluvial channels sourced by E-W trending drainage systems from the Utsira High, however, rather than more locally developed tidal inlets; these sandstones have not been cored (well 15/3-7; Kieft *et al.*, in review). The positions of east-west trending drainage systems (Varadi *et al.* 1998) are interpreted to have been influenced by salt-cored structures. A similar observation has been made in this study where channelised sediments within the Sleipner Delta system are interpreted to have been focussed by salt-cored structures.

9.7 Conclusions

- The transgression of the South Viking Graben during the Middle Jurassic was a diachronous event, with flooding younging both along the main basin axis to the south and eastwards up the hangingwall dip slope.
- The oldest shallow-marine (Hugin-type) deposits are dated as Late Bathonian and are developed on the Gudrun Terrace to the north, whilst the youngest deposits are dated as Early Oxfordian and are developed across the Alpha Central High.
- Different gross depositional environments (bay-fill, shoreface, fluvial-deltaic) are effectively partitioned by markedly different tectonic styles, in addition to proximity to the main fluvial sediment supply system.
- Maximum transgression occurred during the Middle Callovian with shallow-marine sedimentation limited to the flanks of the Utsira High and small segments of the Fladen Ground Spur flank. Locally developed shoreface sandstones were deposited adjacent to and re-worked material from relatively small syn-depositional, intra-basin topographic highs.
- Syn-depositional structural highs are interpreted to have; (i) delimited the spatial extent of a prominent shallow marine embayment in the north of the basin; (ii) formed prominent intra-basin focal points for sandstone deposition; and (iii) resulted in focusing of channels within the Sleipner Delta System.

CHAPTER 10

Conclusions

The results of the study presented in the preceding chapters are summarised here in a series of concluding remarks that are linked back to the overall objectives of the thesis (Chapter 1, section 1.3).

10.1 Sedimentology and sequence stratigraphy

Sedimentological analysis of the Hugin Formation has recognised fifteen facies which have been grouped into six facies associations; bay-fill, shoreface, mouth bar, fluvio-tidal channel fill, coastal plain and offshore open marine. Facies analysis in combination with biostratigraphy data has enabled the identification of regionally correlatable maximum flooding surfaces and locally developed transgressive surfaces, flooding surfaces and candidate sequence boundaries. These surfaces subdivide the succession into a series of stacked parasequences which demonstrate an overall aggradational to retrogradational stacking pattern.

Within the stratigraphic framework defined by these maximum flooding surfaces, several of the main facies associations were developed coevally but in distinct and different geographical locations. The northern part of the study area (Sigrun-Gudrun) is characterised by a series of stacked, progradational bay-fill parasequences which can be correlated across the area and exhibit predominantly aggradational stacking patterns. This suggests that sediment supply was able to keep pace with

accommodation space development. The stacking pattern became more retrogradational towards the end of Hugin Formation deposition, when shoreface sandstones were developed within the area. These parasequences are interpreted to represent deposition within a sheltered bay with associated backstepping barrier shoreline.

In the central part of the study area (Dagny-Alpha North) stacked, backstepping shoreface sandstones are developed, with variable fluvial influence, reflecting deposition within a series of strandplain shorelines. Correlation of these shoreface sandstones between wells is complicated by the occurrence of locally developed unconformities and rapid pinching out down-dip into time-equivalent open marine shales. These relationships indicate the importance of structural development to Hugin Formation deposition in this area, discussed more below. To the south of the study area (Alpha Central) stacked deltaic shorelines that were modified by both waves and tides were developed within parasequences that exhibit an aggradational to progradational stacking pattern. Typically, facies units can be correlated across the structure, although some facies units are only locally developed on the flank of the structure and corresponding parasequence-scale thickness variations are also observed.

The observed palaeogeographical localisation of facies associations, along with local development of flooding surfaces, transgressive surfaces and candidate sequence boundaries is interpreted to reflect deposition within different structural locations which impacted local stratigraphic architectures. These different structural locations are discussed more below. Sediment supply also varied throughout the study area and consequently influenced stratigraphic architecture, with distance from the axial fluvial

system in the south of the basin interpreted to be a significant control on facies development. Wave-driven longshore drift from the north of the basin and local reworking of material from structurally controlled palaeo-highs was also important. The different stacking patterns observed within the Hugin Formation are interpreted to relate to the balance between sediment supply and accommodation space within the graben, as well as the relative interplay between waves and tides.

The net-transgressive strata of the upper Almond Formation outcrop analogue contain a similar range of facies and facies associations to the Hugin Formation, deposited in a series of wave-dominated barrier shorelines and back-barrier lagoons. A variety of sandbody types are documented within the unit, with variable geometries in both depositional strike and dip directions in the absence of a syn-depositional structural control. Shoreface sandstones are highly continuous along depositional strike (at least 19 km), but pinch out up-dip into bay-fill units over short distances (400 m) and are locally eroded into by tidal inlet channel-fill sandstones (up to 200 m wide and 2-7 m thick). Sandstones within the bay-fill and coastal plain facies associations are more discontinuous along strike with channelised sandbodies typically pinching out over less than 200 m whilst non-channelised sandbodies extend for up to 7.5 km. The relationships observed within the Almond Formation highlight the complexity of shallow-marine depositional systems even when a syn-depositional structural control is absent. The observed complexities, dimensions and geometries within the Almond Formation were considered when constraining sub-seismic well-log correlations in the Hugin Formation.

10.2 Structural development of the hangingwall dipslope

During the Middle Jurassic, salt-cored highs and rift-related faults were developing on the hangingwall dipslope of the South Viking Graben and may be expected to have influenced sedimentation patterns (section 10.3). To the north of the study area (Sigrun-Gudrun), the hangingwall dipslope was dissected by a rift-related fault array that formed as a result of gravity-driven sliding above the mobile, salt-bearing Zechstein Group layer. This fault array initiated as a series of individual segments during the latest Bathonian and subsequently propagated and linked to form through-going fault zones during the Upper Jurassic (coincident with Heather and Draupne formation deposition).

Seismic analysis indicates halokinesis of the underlying Zechstein Group initiated during the Triassic, possibly as a result of early movement on the Graben Bounding Fault Zone or as a consequence of differential loading by Triassic depositional systems. Salt-cored structures in the Sigrun-Gudrun area formed long-lived structural features that initiated during the Triassic and continued to develop into the Cretaceous. In the Dagny-Alpha Central area, Jurassic reactivation of the Zechstein Group layer meant that Triassic-age salt-cored highs were deflated and previous salt-withdrawal basins were uplifted.

10.3 Impact of structural development on shallow marine sedimentation

The different structural features developing across the hangingwall dipslope during the Middle Jurassic are interpreted to have had variable influence on shallow marine sedimentation within the graben. During deposition of the Hugin Formation in the Sigrun-Gudrun area, fault displacement was limited to short segments, and the increase in accommodation space in the corresponding hangingwalls was not necessarily accompanied by a pronounced facies shift. This relationship, along with the predominantly aggradational stacking pattern observed in this area, suggests that the volume of sediment supplied to the basin was sufficient to keep pace with fault-generated accommodation space, such that fault-generated topography was unable to develop and consequently facies distributions were not impacted.

Further south, in the Dagny area, shallow marine deposits are locally truncated by unconformities and pinch out rapidly into open marine shales. These units are interpreted to locally fringe structural highs that formed as a result of halokinesis and rift-related extension with associated fault-block rotation. This relationship suggests that sediment supply was outpaced by accommodation space development, leading to localised facies development and abrupt lateral facies changes down steep bathymetric gradients. In the Alpha Central area, facies can typically be correlated across the developing salt-cored high implying that sediment supply was predominantly sufficient to keep pace with accommodation space generation and differential topography did not develop. This is interpreted to reflect proximity to the main axial feeder system to the graben. However, some channelised units are interpreted to have been preferentially deposited flanking the growing salt-cored structure.

The results of this study identify the importance of salt-controlled, rather than fault-controlled, differential subsidence in influencing deposition patterns within the Hugin Formation despite deposition coeval with regional extension and rifting. Key to this relationship is the balance between sediment supply and subsidence, which determined whether structurally generated topography was able to develop at the seabed.

10.4 Further work

The balance between sediment supply and structural development within the Hugin Formation was key to determining whether a developing structure would accumulate seabed topography and consequently impact depositional patterns. In order to improve understanding of sediment source and supply within the basin, provenance work could be undertaken, with the aim of establishing the sediment source areas and transport routes in different stratigraphic intervals and palaeogeographical locations. For example, although significant section is missing from the Utsira High, little work has been undertaken to date in linking this missing section to deposits on the flanks of the high.

Where shoreface sandstones are developed locally fringing structural highs, two possible sources of sand are envisaged: (1) longshore drift, implying a regional sediment source to the north of the study area, and (2) local crestal erosion and reworking of sediment from the structural high up-dip. These different sources of sand have major implications for the volume of sand that may be predicted, as well

as the spatial extent over which it may be distributed. An assessment of the volume of material eroded from the crest of structural highs would allow the feasibility of the latter model to be established.

Provenance of the deep marine sandstones deposited within the graben coeval to Hugin Formation deposition (e.g. Ling Member) is also uncertain despite these sandstones forming an additional reservoir target. Whilst the feeder system of the Ling Member is poorly constrained, provenance work may help to determine whether sandstones were derived from the east or west flank of the graben. As a result, it may be possible to identify other areas in which this play may be developed.

Detailed structural reconstruction using 3D and long-offset 2D seismic data may help to constrain the position of syn-depositional structural highs and consequently facies belts. This is particularly important with respect to halokinesis in the south of the study area and could provide a more detailed insight into the position of syn-depositional structural highs which may have focussed shallow marine depositional systems and controlled reservoir distribution in this area. It may also have impacted the routing and position of deep marine sandstones similar to those of the Ling Member.

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