

The Impact of Syn-Rift Halokinesis To Sediment Deposition And Distribution: A Case Study In The Central North Sea Basin, UK

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IMPERIAL COLLEGE LONDON

2016

A thesis submitted in fulfilment of the requirements for the degree of Doctor
of Philosophy at Imperial College London

DECLARATION:

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ABSTRACT

This study looks at the effect of rifting and halokinesis to sediment distribution in the Central Graben, which is one of a few basins that have been subjected to both rifting and halokinesis at the same time. The main aim of this study is to integrate the plethora of well and 3D seismic data in the Central Graben in order to re-evaluate the tectono-stratigraphic evolution of the rifting- and halokinesis-influenced Upper Jurassic rocks in the basin. In general, the basin was controlled by three major structural trends: (1) NW- and NE-striking Caledonian structures; (2) E-striking Variscan structures; and (3) NNW- to NW-striking Permo-Triassic structure. The initial opening of the basin during the Late Jurassic period rejuvenated and exploited the NW-striking Permo-Triassic structures before being taken over by a NNW-striking fault pattern. The complex fault pattern resulted in the CG being subdivided into several structural domains, which define areas of different salt wall patterns and distribution of the major Jurassic depocentres. The Upper Jurassic stratigraphy was divided into six major sequences bounded by major flooding surfaces. Nine sedimentary facies were also identified, associated with non-marine to deepwater depositional environments. Non-marine facies associations are found at the edge of the basin whereas shallow marine facies association could be found throughout the basin. Deep marine facies association, mainly made of low density turbidites, is rare but might be found in the axial part of each structural domains. No major submarine fan was identified, perhaps due to the lack of incised valley network that typically feed large fan system. In general, Oxfordian sequences were more influenced by halokinesis, indicated by thickening of sediments towards salt-withdrawal mini-basins and sporadic distribution of turbidites related to over-steepened Triassic's mini-basin flanks, than the younger Kimmeridgian to Tithonian sequences. However, the younger sediments on the flanks were also affected by salt diapirs resulting in a compartmentalized architecture.

ACKNOWLEDGEMENT

Firstly, I would like to express my sincerest gratitude to my supervisors, Prof Howard Johnson and Prof Chris Jackson, who have provided guidance and constantly supported me throughout this study. Both are excellent motivators and have provided invaluable feedbacks in order to improve this thesis. Their dedication towards improving my research skills and contribution in adding and propagating geological knowledge will forever be the inspiration for me in my future work.

My gratitude also goes to my external supervisors, Tom Mckie and Caroline Gill from Shell UK, who have spent their valuable time in helping me to get familiar with the data used for this study and provided regular reviews of my work, especially the initial core descriptions and seismic interpretation works. I also would like to thank Catherine Burgess and Peter Osterloff from Shell who have assisted me in the biostratigraphic picking.

This study would not have materialized without my sponsor and I would like to thank PETRONAS for giving me the opportunity to pursue this study. In particular, I owe very much to Mr Idris Ibrahim, who have encouraged me to undertake this study and have supported me throughout this endeavor. I also would like to thank Dr Mazlan Madon who had interviewed me prior to getting the approval for this study and personnels in the Staff Development Programme who have assisted me in administration matters throughout the studies.

This study also could not be conducted without the generous contribution of seismic data by PGS and well data by Shell UK. In particular, I would like to thank Richard Lamb from PGS as well as Tom and Caroline for facilitating the data collection.

I would also like to thank Dr Gary Hampson and Prof Alastair Fraser for their valuable feedback during the internal reviews. I am also thankful to my fellow PhD candidates and staffs at the Basin Research Groups in Imperial College who have provided numerous feedbacks in order to improve my work.

My love and gratitude for my beautiful wife, Ris Melati Shamsuddin, who have stood by me throughout the ups and downs of this study, as also in our lives, and provided the regular nudge when I was slacking behind. I am happy that my children have been around during this phase of my life and hope that they have enjoyed it very much as I have enjoyed their company. Love and gratitude also for my close relatives for their continuous moral supports from home particularly for my parents whom my longing to be with them was one of the motivation for me to finish this thesis as fast as I can.

Finally, all praises to the Almighty Lord of all the Worlds that is the source of all knowledge. The believe in You keeps me strong in trying times.

Contents

ABSTRACT.....	ii
ACKNOWLEDGEMENT.....	iii
Chapter 1: Introduction	1
Chapter 2: Background of study and main aims	5
Motivation.....	5
Rifting and Rift Basins	6
Salt tectonics.....	8
Net-transgressive Sediments	13
Main aims and objectives of this study.....	15
Chapter 3: Structural Evolution of a Salt-Influenced Rift Basin: A Regional Synthesis of the Central Graben, UK Central North Sea	16
Introduction	16
Current understanding of the geological setting of the Central Graben	22
Pre-Mesozoic Orogenesis Period	22
Mesozoic Rifting Period	26
Post-rift Thermal Subsidence Period	28
Data and Methods	29
Data	29
Methodology.....	33
Sub-Salt Structures.....	38
Observations	38
Interpretation.....	42
Salt Structures.....	45
Observations	45
Interpretation.....	48
Triassic Structures	49
Observation.....	49
Interpretation.....	52
Jurassic Structures	55
Observations	55
Interpretation.....	59
Lower Cretaceous Structures.....	61

Observations	61
Interpretation.....	64
Upper Cretaceous and Younger Structures	65
Observations	65
Interpretation.....	69
Structural Reconstruction	69
Discussions.....	76
Deep structural controls on the location of salt diapirs and subdivision of halokinetic provinces	76
Triassic rifting.....	78
Jurassic rifting	79
Conclusions	81
Chapter 4: The impact of halokinesis and rifting on the development and distribution of depositional systems in the Upper Jurassic of the Central Graben Basin, UK Continental Shelf	83
Introduction	83
Stratigraphic Framework	89
Jurassic rifting in the Central Graben.....	95
Main Depositional Facies	103
Coastal plain environment.....	107
Shallow marine environment.....	112
Deep marine environment.....	115
Sequence stratigraphic analysis.....	116
Lower to Middle Jurassic interval	119
Oxfordian interval (Sequences I & II)	126
Kimmeridgian-Tithonian interval (Sequences III to VI)	129
Facies distribution and depositional evolution of the Upper Jurassic.....	132
Oxfordian palaeogeography	135
Kimmeridgian-Tithonian palaeogeography	138
Controls on reservoir distribution.....	139
Tectono-sedimentary settings	141
Interpod depositional models.....	144
Synthesis	151
Conclusions	153
Chapter 5: Gravity flow deposits in salt-influenced, syn-rift setting and implications for the hydrocarbon prospectivity: Examples from the Central Graben, United Kingdom Continental Shelf.....	155

Introduction	155
Tectono-stratigraphic evolution of the Central Graben	161
Pre-Mesozoic period	162
Mesozoic Period.....	163
Post-Mesozoic period	166
Summary of Upper Jurassic gravity flow deposits in the North Sea Basin	166
Data and Methodology	174
Sedimentology of Upper Jurassic gravity flow deposits in the Central Graben	177
Massive mudstone facies (T_{E-2} and T_{E-3})	180
Laminated mudstone facies (T_{E-1})	180
Laminated siltstone facies (T_D).....	181
Ripple-laminated sandstone facies (T_C)	187
Laminated sandstone facies (T_{B-1} & T_{B-2})	188
Massive sandstone facies (T_A).....	188
Clast-rich muddy sandstone facies (D_{M-2})	189
Distribution of gravity flow deposits in the Central Graben.....	198
Oxfordian	201
Kimmeridgian-Tithonian	204
Discussions.....	207
Classification of gravity flow deposits in Central Graben	207
Structural and depositional setting of gravity flow deposits in Central Graben	208
Significance for petroleum exploration	210
Conclusions	212
Chapter 6: Conclusions and Recommendations	213
Structural and stratigraphic framework	213
Halokinesis and tectonic influence on the Upper Jurassic deposits.....	217
Recommendations for future research.....	219
References	221

APPENDIX 1: Well Logs from Jurassic Section

APPENDIX 2: Core Descriptions

APPENDIX 3: Maps

Chapter 1: Introduction

Rift basins are one of the most important hydrocarbon habitats in the world with 31% of world's giant hydrocarbon fields found in this setting (Mann et al., 2003). Many of these basins are also influenced by halokinesis. Good examples of these basins include the Central Graben in the United Kingdom and Norway (e.g. Hodgson et al., 1992), Gulf of Mexico in the United States (e.g. Seni, 1992), Whale-Horseshoe in Canada (e.g. Tankard et al., 1989), Gulf of Suez in Egypt (e.g. Carr et al., 2003), and Lower Congo in Angola (e.g. Valle et al., 2001). Those basins that are simultaneously affected by rifting and halokinesis are well-known for being tectonically and stratigraphically complex. For example, the Central Graben (CG), Whale-Horseshoe and Gulf of Suez basins all demonstrate how these complexities influence petroleum systems that develop in such basins. Specifically, syn-depositional tectonics can exert a first order impact on reservoir, source and seal distribution, and is often fundamental in terms of trapping mechanism, the genesis and timing of source rock maturity and the migration, entrapment and preservation of hydrocarbons. The superimposition of salt and associated halokinesis adds further complexity, especially with regards to reservoir distribution and trap types. However, the precise nature and timing of extensional faulting and salt movement leads to a plethora of complexity, which requires a combination of high quality data and careful structural and stratigraphic analysis. The latter is often at a level that goes beyond what is often required in other, less complex, sedimentary basins. Consequently, understanding the tectono-stratigraphic evolution of salt-influenced rift basins is the key to finding and exploiting the petroleum resources contained within these basins.

To unravel the tectono-stratigraphic evolution of a salt-influenced rift basin we need to understand the basins regional tectonic history, and how relative sea level change, sediment supply and climate have changed through time (Reading and Levell, 1996). To do this, we require high-resolution

regional datasets, comprising 2D and 3D seismic reflection surveys, well logs, biostratigraphic and core data. In this study, an extensive array of subsurface data from the CG of the North Sea is used to evaluate the complexities of a salt-influenced rift basin. The choice of basin was driven by the abundant, high quality data that are available. In addition, the basin is one of the most important and mature hydrocarbon provinces of Northwest Europe, but with the potential for significant remaining exploration potential (e.g. Erratt et al., 2010).

This study uses 22,360 km² of 3D seismic reflection data (Megamerge; courtesy of PGS) and 200 wells containing wireline logs (including gamma-ray, sonic, density and neutron logs), lithology logs, biostratigraphically-calibrated picks and time-depth surveys (courtesy of Shell UK). Wells were used to constrain the seismic interpretation by tying the seismic response to well log character.

Furthermore, the wells have an aggregate length of around 5.8 km of conventional cores, in which a total length of 1.5 km has been independently described in this study. All these data are used to generate structure, isochore and palaeogeographic maps, as well as stratigraphic correlation panels. The analysis of these data has been undertaken at both regional- and sub-basin-scales, with each conforming the other. In so doing, the aim has been to provide a rigorous and fully integrated evaluation of the basin's tectonic and stratigraphic evolution.

The aims, objectives and outcomes of this study are documented in this thesis, which is organised into six chapters. Chapter 1 outlines the contents of the five main chapters of the thesis and provides the context for the project. Chapter 2 provides a background to the study, which includes (1) study rationale, (2) geological concepts that will guide the research, and (3) the study aims. Chapters 3 to 5 contain the main results of this study, which are written in the format of journal papers. Consequently, there is an unavoidable level of repetition, most notably with regards to the geological background, datasets and methodology. However, this has been minimised by focusing

only on those aspects that are relevant to the context of each specific paper. Chapter 6 integrates the key findings of each paper into a final synthesis.

The tectonic evolution of the North Sea and how the past events have shaped the Central North Sea form the main discussion of Chapter 3. The chapter presents a synthesis of the tectono-stratigraphic evolution of the basin based on (1) an independent analysis of the available data, and (2) integration with an extensive range of previous publications. The chapter outlines a new regional structural interpretation of the basin, which extends from the Palaeozoic to the Tertiary and highlights the importance of inherited structures and their repeated reactivation, which spatial and temporal variability. This interpretation was possible because of the large subsurface dataset, which included basin-wide 3D seismic and well data, which are described and interpreted in detail in the chapter. Large basin-scale structures have been highlighted and linked to the main tectonic events. The chapter also considers the distribution and halokinetic evolution of Upper Permian salt (Zechstein Supergroup), including the structures that result from the flow of salt, and the impact this had on the geometry and evolution of Triassic and Jurassic depocentres.

The effect of Upper Jurassic rifting and halokinesis on (1) the development of accommodation space, and (2) the nature, type and distribution of depositional environments, provides the basis analyzing the stratigraphic evolution of the CG, which is outlined in Chapter 4. This chapter is illustrated with a wide range of representative illustrations, notably regional well correlations, seismic profiles and isochore maps, of the main Jurassic intervals in order to demonstrate how rifting and halokinesis impacted the stratigraphic architecture and evolution of the Upper Jurassic of the Central Graben. Palaeogeographic maps of the main Upper Jurassic intervals are also presented in order to show how accommodation creation and infill are varied through space and time. Although this chapter discusses the overall stratigraphy, it focuses more on the distribution of the main shallow-marine

environments because of the dominance of this depositional system with regards to its reservoir potential.

Chapter 5 continues to discuss on the stratigraphic development of the Central Graben, but focuses on deep-water depositional systems, specifically the type and distribution of Upper Jurassic gravity-flow deposits. The importance of this hydrocarbon play in other parts of the North Sea (e.g. in the Moray Firth, Viking Graben and East Shetland Basin) has so far not been repeated to the same degree in the Central Graben, although all the key elements of reservoir, source and seal are established. This chapter integrates core data with well logs and seismic data in order to establish the vertical and lateral relationship of the gravity-flow sandstones with other deep- and shallow-marine facies. These observations are also compared with the occurrence of gravity-flow sandstone in the Viking Graben, where the Upper Jurassic gravity-flow reservoirs are thicker, more extensive and commercially important. In addition, these reservoirs have been more extensively sampled and better understood, and are considered from an analogue perspective in relation to the origin and distribution of the gravity flow sandstones in the Central Graben.

Finally, Chapter 6 summarises the main findings of this study and recommends areas of further research. This includes topics that did not fit into the timing and scope of the current research work, but which would provide valuable, additional insight.

Chapter 2: Background of study and main aims

Motivation

The sedimentary successions in rift settings are of various environmental origins, ranging from continental to deep marine, depending mainly on the gross depositional setting of the rift basin (e.g. marine versus continental), the stage of rifting (e.g. syn-rift or post rift), sea level fluctuations and climate (Gawthorpe and Leeder, 2000; Martins-Neto and Catuneanu, 2010). In rift settings, the stacking and distribution of the sedimentary facies are also controlled by the interplay between the rate of subsidence between different fault blocks and the rate of sediment supply that entered into those fault blocks. This is in contrast with the passive margin settings where the sediment stacking is more closely controlled by the rate of relative sea level change and rate of sediment supply into the accommodation space (Lovell, 2010; Vail et al., 1977a). The relative sea level change is defined as the local sea level change relative to the sea-bed and the main driver of this change in passive margin settings is the eustatic sea level change. In contrast, the main cause in sea level fluctuations in marine rift basins is often driven by tectonics.

This relationship between tectonic subsidence and sediment supply is further complicated with the presence of halokinesis. Salt bodies may withdraw or grow (vertically and/or laterally) during rifting and this can have a profound effect on the overlying sediment, be it at the pre-, syn- or post-kinematic stages (e.g. Erratt, 1993; Hudec and Jackson, 2007). Salt also provides a decollement zone between the pre- and post-salt intervals, resulting in the decoupling of faults above and below salt stocks (e.g. Lewis et al., 2013; Stewart et al., 1996). The interaction between extension and halokinetic processes, especially where a younger rift basin develops above an underlying salt horizon, as in the case of the CG, sets up a particularly complex system, both with respect to petroleum exploration and research opportunities focused on understanding rift-salt interactions.

Rifting and Rift Basins

Rifting is a crustal thinning process in response to lithospheric extension and can be characterised by shallow depth to Moho, high surface heat flow, high volcanism and high seismic activity with predominantly normal faulting and elevated rift margins (Allen and Allen, 2005; McKenzie, 1978).

Active rifting involves the stretching of the lithosphere due to activities in the asthenosphere, such as upwelling of a hot mantle plume (Pirajno and Santosh, 2015; Turcotte and Emerman, 1983).

Passive rifting on the other hand involves mechanical stretching of the lithosphere by distant extensional forces with passive upwelling of the asthenosphere (Godfrey Fitton, 1983). Regional doming might be observed in areas affected by active rifting but not usually for passive rifting areas (e.g. Underhill and Partington, 1993).

Both types of rifting form rotated fault blocks that can create accommodation space in the hangingwall areas. In terrestrial settings, these spaces are filled with continental sediments, such as alluvial fan, fluvial deposits and/or aeolian deposits, depending on climatic conditions. A combination of climate and varying rates of extension and subsidence can create a wide range of continental rift basin facies characteristics and depositional environments. These can extend from shallow and periodically emergent lakes through to deep and permanently water-filled lakes (e.g. East African Rift; Macgregor, 2015), which results in a wide range of lacustrine deposits. Climate may vary along the length of larger rift basins where these rifts span across different climatic belts. For example, in the East African Rift System (EARS) semi-arid conditions prevail in the northern half and the lacustrine basin-fill comprises relatively shallow-water lacustrine deposits. In contrast, the climate in the southern part of the EARS is wetter and lakes are filled with deep-water lacustrine deposits (Scholz et al., 1998). In other cases, shallow-marine deposition may develop if the rift

becomes connected to adjacent oceans and where sea level is sufficiently high to place the rift in a submarine setting (e.g. North Sea Rift basin; Ziegler and von Hoorn, 1989). Eventually, deeper marine conditions may develop where increases in relative sea level cause the rift basin to experience >100-200 m water depth (e.g. Erratt et al., 1999; Mendoza-Rosales et al., 2010; Ruiz-Ortiz et al., 2006). In addition, volcanics may also accompany these syn-rift sediments as elevated Moho can heat and melt the crust (Corti, 2009). An example of this situation is locally developed where the three arms of the North Sea rift basin connect, in the northern part of the CG, in the form of the Rattray Volcanics.

Sediment delivery directions into rift basins also vary throughout the rifting period due to the nature of bounding faults on the rift edges (Gawthorpe and Leeder, 2000). Initially, a series of short, en-echelon faults with accompanying relay ramps commonly form on the proto-rift shoulder (e.g. Fraser et al., 2003). These faults typically join up to form more continuous boundary faults and, as the rift basin deepens, sediment starvation often follows (Fig. 2.1).

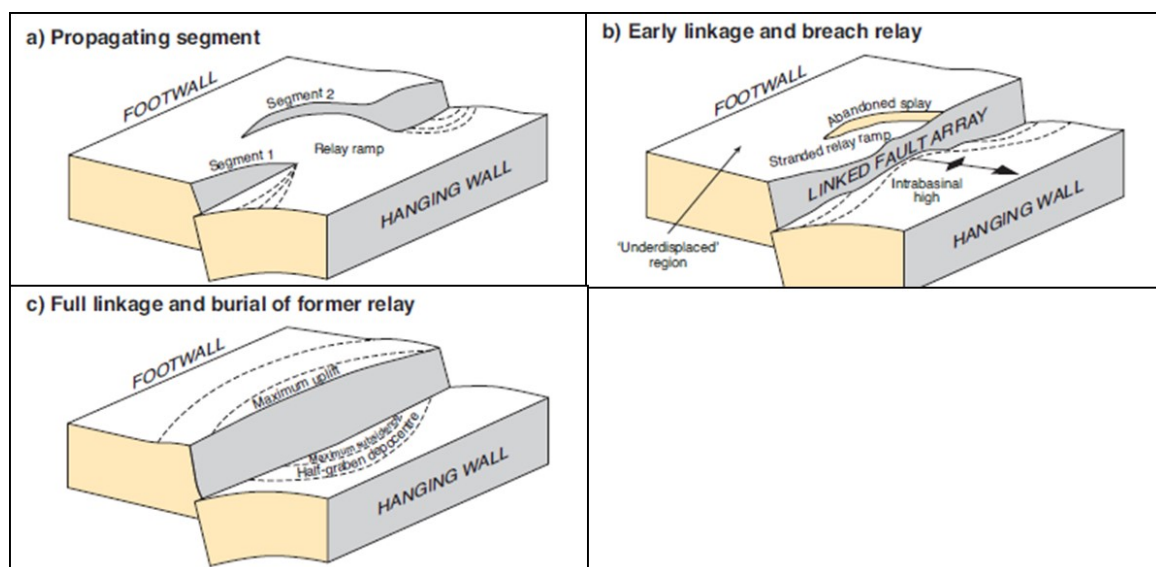


Fig. 2.1: Development and evolution of North Sea Faults (Fraser et al., 2003).

Continuous rifting eventually leads to the formation of new oceanic crust as the now separated continental crust drifts apart (Liao and Gerya, 2015; Vine, 1966; Wilson, 1966). However, a failed rift (aulacogen) may also occur if the source of extension force is halted before the separation phase (Robert and Christian, 2008). In both cases, a post-rift stage usually occurs where the area subsides with thermal cooling of the lithosphere. In most failed rift basins, the post-rift sediment fill could be of continental deposit, such as in the continental Muglad Basin in Sudan (McHargue et al., 1992) or marginal to open marine in the North Sea Basin (Ziegler, 1990b).

In general, syn-rift stratigraphy often reflects an overall deepening character, which is similar to TST in passive margins but at a longer time scale. At a shorter time scale, syn-rift stratigraphy may show periodic regressive-transgressive cycles.

Salt tectonics

Salt is a type of evaporitic rock that is composed mostly of halite but may contain a variety of other evaporitic minerals, such as gypsum, anhydrite, evaporitic carbonates, potash and sulphate, as well as non-evaporitic rocks picked up during salt movement (Hudec and Jackson, 2007). It is usually deposited when surface and near-surface brine becomes saturated as a result of solar evaporation (Warren, 2010).

About 120 of the world's evaporite basins have been affected by salt tectonics, or halokinesis (Hudec and Jackson, 2007), including the Central North Sea Basin. In generally, salt may be present in four main settings: (1) cratonic basins; (2) syn-rift basins; (3) post-rift passive margins, and (4) continental collision zones and foreland basins (Hudec and Jackson, 2007). Some of these settings exhibit more

salt-related deformation than others. Recent studies have shown that the main driving force of salt tectonics is differential loading (Hudec & Jackson, 2006). Bouyancy, although important, is not the main mechanism as shown in many undeformed salt basins, such as the East Siberian, Mid-Continental US and the West Canadian basins, where buried salts were left relatively undeformed for millions of years.

The physical attributes of salt are key to understanding how it deforms and how this causes the deformation in the surrounding rocks. Salt is incompressible and has an approximate density of 2200 kg m^{-3} , which is less than most compacted sedimentary rocks (Warren, 2010). Therefore, it is usually more buoyant than the surrounding rocks after a typical burial depth of 3000 m, and as shallow as 1600 m, when the overburden rock begins to have a density larger than rock salt (Hudec & Jackson, 2007). This makes salt inherently unstable, which results in salt basins deforming more easily than other basins.

Salt also exhibits viscoelastic deformation under geologic conditions, but due to the relatively high speed of relaxation it is generally treated as purely viscous and, therefore, flows like fluid (Hudec & Jackson, 2007). Therefore, under geological conditions, salt is much weaker than the surrounding rocks and is more susceptible to deformation. Consequently, completely dry salt deforms by creep but salt with as little as 0.05 wt.% water can deform by diffusion creep, especially when it is fine grained or if the strain rate and differential stresses are low (Hudec & Jackson, 2007). In addition, like fluid, salt is difficult to fracture, hence faults use salt bodies as decollement surface, rather than shearing through the feature.

Salt diapirs form in four different ways: 1) reactive piercement in between fault blocks due to the extension of the overburden, 2) active piercement of salt by lifting, rotating and shouldering aside the overburden rocks, 3) erosional piercement of salt after the removal of the overburden by erosion, and 4) emplacement of salt in the hangingwall of thrust faults (Fig. 2.2). All these processes can occur at various times throughout the growth of a single salt structure (Hudec and Jackson, 2007). In addition, salt diapirs might also form, albeit rarely, by ductile thinning of the roof overburden and where salt is exposed at the surface, it could rise continually with respect to the surrounding strata in a process called passive diapirism or downbuilding. Most tall salt domes and walls develop as passive diapirs (Hudec and Jackson, 2007).

Most salt diapir provinces were initiated during phases of regional extension as shown by salt bodies having the deeper part of their flanks commonly being adjoined by swarms of normal faults (Hudec & Jackson, 2007). Regional extension could result in reactive, active and passive diapirism, although salt diapirs do not necessarily progress through all of these stages. Active diapirism is controlled by gravitational forces and might continue even if regional extension stops. This stage is usually followed by passive diapirism as the salt continues to rise to the surface.

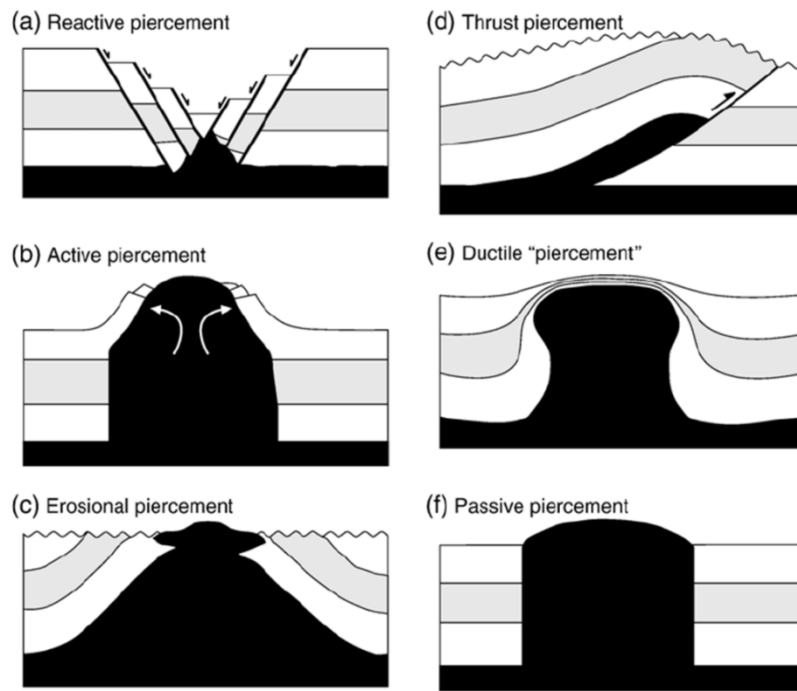


Fig. 2.2: Modes of salt diapir piercements (Hudec & Jackson, 2007)

During compression or regional shortening, the weaker salt layers usually deform more easily than the surrounding rocks, causing the overburden to buckle. Salt may then flow into the buckled overburden and amplify any existing anticlinal structures. When a pre-existing salt structure is shortened, a teardrop diapir might be formed (Fig. 2.3), creating a steep salt weld in the former waist of an hour-glass-shaped diapir. Where feeder welds are gently inclined, continued shortening of the diapir may reactivate the weld as a thrust fault.

Allochthonous salt might form where mobilised salt overlies stratigraphically younger rocks. Salt sheets comprising allochthonous salts can originate from a single feeder and advance in four ways: 1) extrusive, 2) open-toed, 3) thrust, and 4) salt wing intrusion (Hudec and Jackson, 2007).

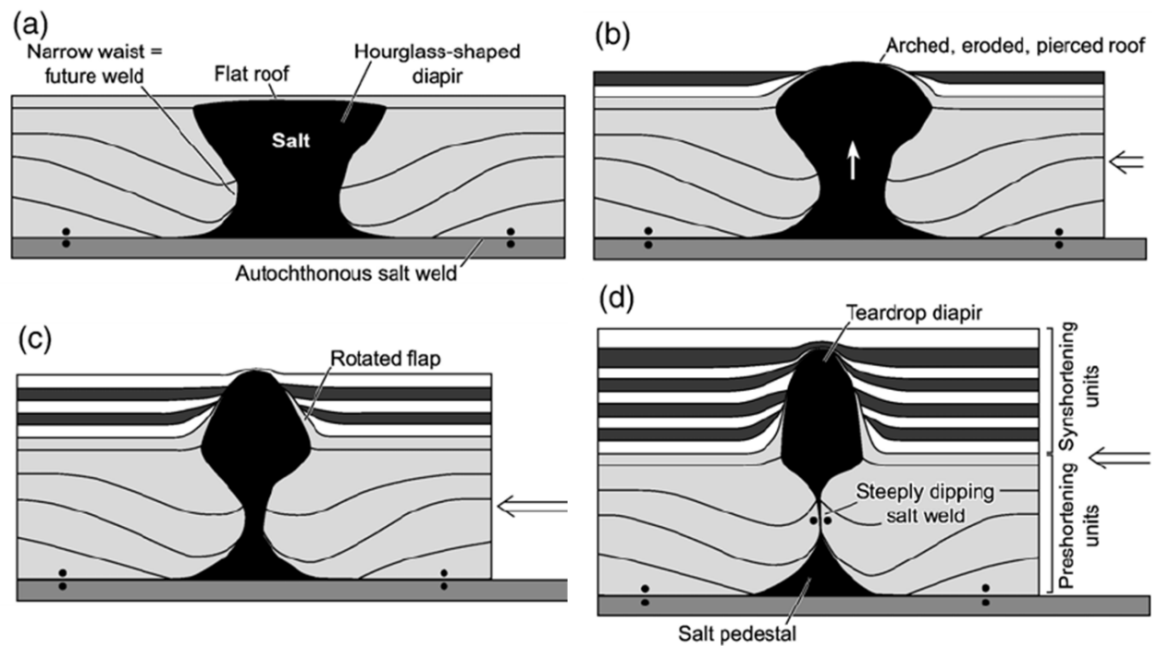


Fig. 2.3: Formation of tear-drop diapir during regional shortening (Hudec and Jackson, 2007)

Salt welds are not entirely free of salt (Wagner III and Jackson, 2011). Up to 50 m of salt can remain in an incomplete weld if salt flow is restricted, for example beneath a broad, prograding sediment wedge. Where salt flow is unrestricted, for example beneath a subsiding minibasin, about 1 m of salt could be left. In both cases, any remaining salt must be dissolved to leave a weld without any remnant salt.

Another common aspect to salt is that it dissolves easily when exposed to meteoric water (e.g. Johnson, 1981). Salt dissolution can also create accommodation space and influence sediment routing systems, which can be independent of tectonic and sea level affects (e.g. Gutiérrez, 2004). All of these salt features can be found in salt basins that have undergone tectonic extension and compression, including the Central North Sea Basin.

Net-transgressive Sediments

Transgression describes the landward marine encroachment of a shoreline and/or the drowning of the adjacent lower coastal plain. This could be achieved by a rapid increase in the eustatic sea level, especially in passive margin areas where tectonic activity is negligible. In tectonically-active areas, this often represents a balance between eustatic sea level change and rate of subsidence or uplift of the coastal area. This balance determines the rate at which sediment accommodation space is being created or destroyed, causing the fall or rise in the relative (local) sea level. Other factors that also affect the sedimentary character include sediment supply and climate. Together, all of these factors control the rate of sedimentation, stacking pattern, lateral distribution, environments of deposition and type of lithology (Reading and Levell, 1996).

The interaction between the rate of sedimentation and rate of change in the accommodation space results in the formation of different stratigraphic architectural styles, particularly vertical stacking patterns (Allen and Allen, 2005; Van Wagoner et al., 1990). If sedimentation rate exceeds that of the change in the accommodation space creation, sediments will prograde basinwards. When the opposite happens, sediments retrograde, or backstep landwards. Finally, aggradation stacking patterns form when the rates of sedimentation and accommodation space creation are balanced.

During depositional transgression, coastal facies belts may retrograde, aggrade or show a combination of both stacking patterns as the accommodation space increases due to the rise in the relative sea level. In contrast, erosional transgression occurs where sediment supply is very low and/or where the shoreline migrates rapidly landward, which results in negligible preservation of depositional environments. These conditions are often marked by a wave-cut erosion surface, or wave ravinement surface, which are often overlain by a strongly bioturbated and/or winnowed lag deposit (<1 m thick). In successions formed during depositional transgression a series of genetically

related depositional systems, or systems tract, are preserved as landward backstepping of sedimentary environments (Reading and Levell, 1996). However, at shorter time scales (higher order), sediments may also show prograding character (Van Wagoner et al., 1990).

In idealised sequence stratigraphic models, the transgressive systems tract (TST) is an example of depositional transgression. In the ideal case this follows a period of low relative sea level, or lowstand systems tract (LST), where deposition occurs in more basinward areas as a result of a decrease in accommodation space in the coastal area due to regression (Posamentier and Vail, 1988; Vail et al., 1977b; Van Wagoner et al., 1988). The model predicts that previous coastal plain or shelf areas would have been exposed, with eroded material deposited in deeper water, basinal areas. In classical sequence stratigraphic models (Fig. 2.4), lowstand sedimentation comprises three distinct episodes of deposition: 1) basin floor fan (BFF), 2) leveed channel complex (LCC) or slope fan (SF), and 3) lowstand prograding wedge (LPW). The last episode marks the time of maximum progradation of coastal environments. Subsequently, when relative sea level rises rapidly during the TST, deeper water facies accumulate above shallower water facies. The boundary between these two intervals is the transgressive surface (TS), which defines the base of the TST. Marine flooding continues until it reaches its maximum landward position, establishing a maximum flooding surface (MFS), which defines the top of TST. Subsequently, there is a period of slower rise in relative sea level where progradation dominates due to relatively stable accommodation space. This highstand systems tract (HST) period continues until the next rapid sea level fall initiates a new LST cycle.

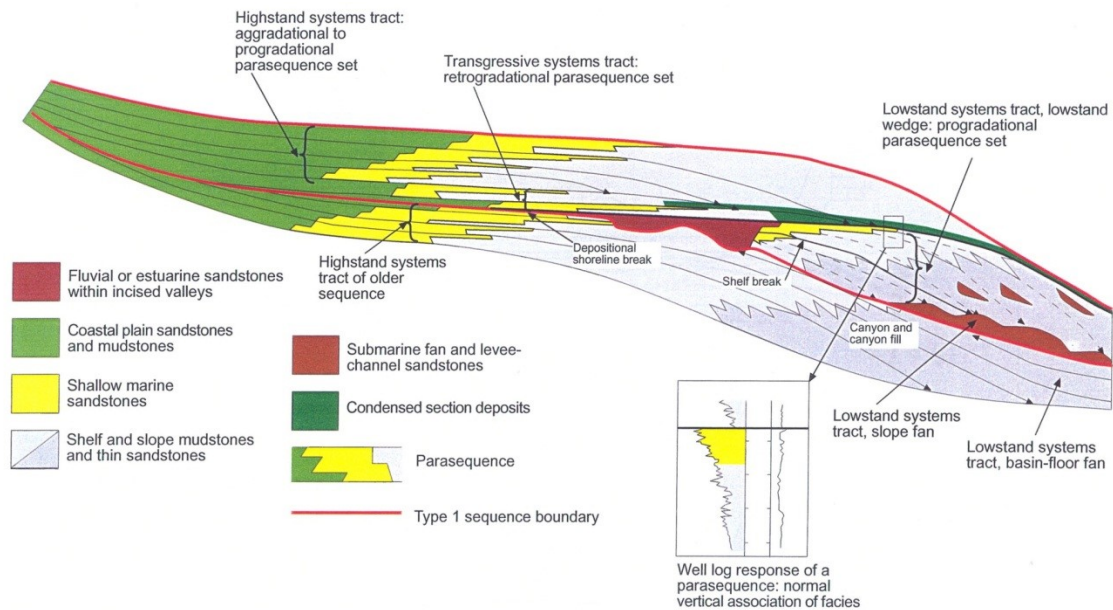


Fig. 2.4: Third Order Type I Sequence Stratigraphic model (Van Wagoner et al., 1988)

Main aims and objectives of this study

The main aim of this study is to integrate the plethora of well data and extensive 3D seismic data (PGS Mega-survey of the Central North Sea) in order to re-evaluate the tectono-stratigraphic evolution of the Upper Jurassic of the Central Graben. The study area will cover most of the Western Platform and adjoining Central Graben. In order to achieve this aim, several objectives have been identified: (1) To establish the major stratigraphic boundaries in the study area and thereby create a structural and stratigraphic framework for the Central Graben; (2) To establish an integrated sedimentary facies model and sequence stratigraphic framework for the Upper Jurassic formations; (3) To evaluate the range of Upper Jurassic sandstone preservation styles by unraveling the regional, sub-regional and local geological controls; and (4) To attempt to deconvolve the stratigraphic response to rifting, halokinesis and relative sea level rise.

Chapter 3: Structural Evolution of a Salt-Influenced Rift Basin: A Regional Synthesis of the Central Graben, UK Central North Sea

Introduction

Halokinesis has affected at least 120 of the Earth's known sedimentary basins, which occur in a wide range of tectonic settings, including cratonic, syn-rift, post-rift passive margin, continental collision zones and foreland basins (e.g. Brun and Fort, 2011; Hudec and Jackson, 2007; Jackson and Vendeville, 1994; Tankard et al., 1989). In basement-involved settings, such as in rift and continental collision systems, halokinesis of pre- or syn-kinematically deposited salt may also be synchronous with either thick-skinned deformation connected to deeper basement movement, or thin-skinned deformation triggered by gravitational forces. Some prominent examples of syn-rift halokinesis occur in the North Sea Basin (UK North Sea; e.g. Hodgson et al., 1992), the Whale-Horseshoe Basin (offshore Canada; e.g. Tankard et al., 1989) and the Gulf of Suez (offshore Egypt; e.g. Jackson and Vendeville, 1994). In these settings, the basin geometry is complex due to the interaction between rifting and halokinesis. In turn, these complexities impact the relationship between sediment dispersal, facies distribution and accommodation, and subsequent stratigraphic preservation. In petroleum-bearing basins, these processes also determine reservoir/seal relationships, trap style, size and development as well as petroleum generation, migration, entrapment and preservation. Hence, unravelling the relative impact of rifting and halokinesis on the architecture of extensional basins is critical to understand the polyphase tectono-stratigraphic history of such basins, including the prediction of ultimate hydrocarbon prospectivity and prospect evaluation.

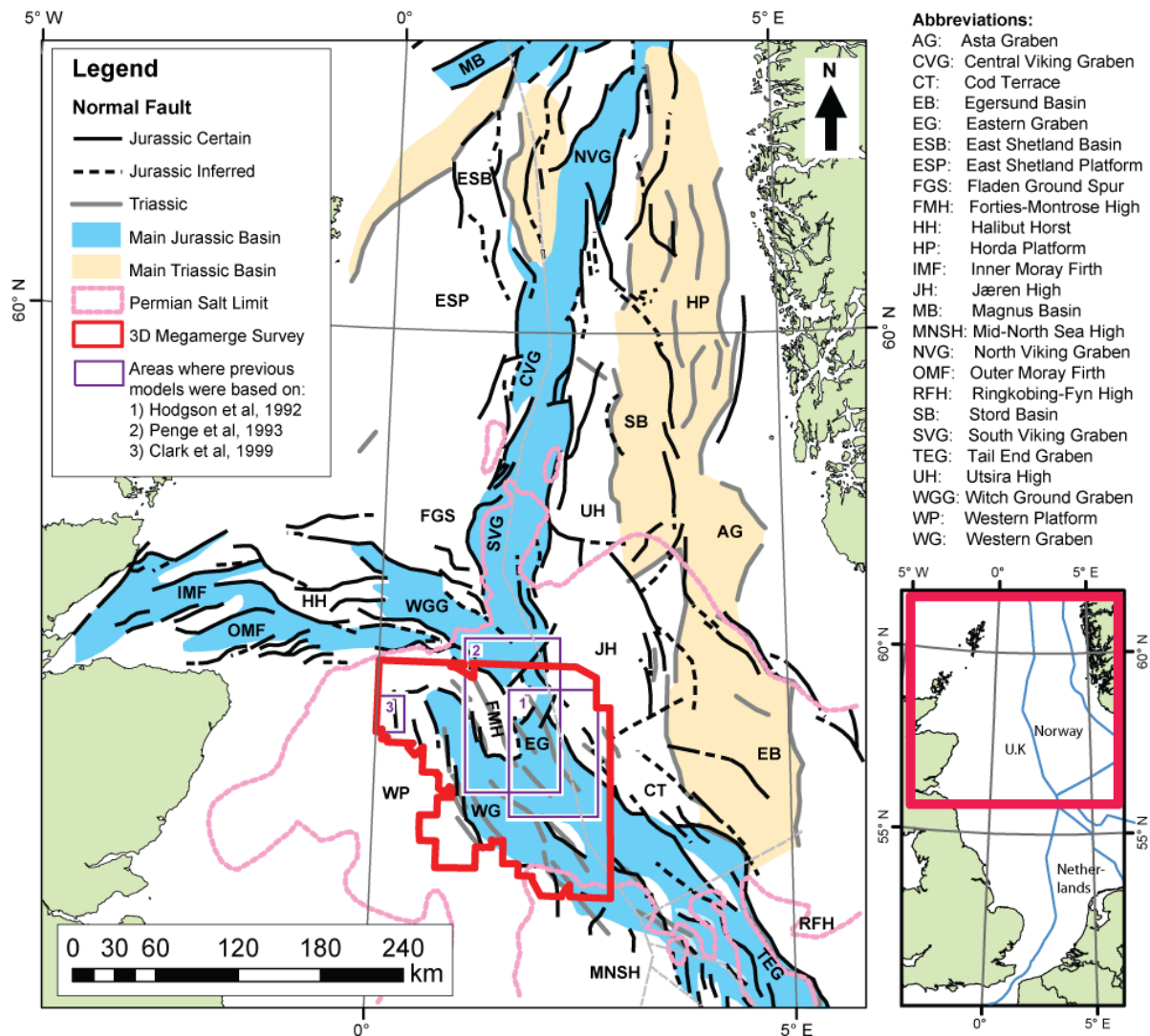


Fig. 3.1: Location of the study area with major Mesozoic structural elements in the North Sea (Fraser et al., 2003; Zanella and Coward, 2003) and limit of Permian salt. Previous halokinesis models are based on areas highlighted in purple.

The Central Graben (CG) of the North Sea Basin (Fig. 3.1) is one of the best known examples of a sedimentary basin whose long history (c. 300 Ma) has involved multiple phases of rifting and halokinesis superimposed on a diverse range of depositional systems (e.g. continental, shallow marine and deep marine settings) (Davies et al., 2001; Eggink et al., 1996; Erratt, 1993; Erratt et al., 2010; Erratt et al., 1999; Glennie, 1984; Gowers et al., 1993; Hodgson et al., 1992; Jones et al., 1999; Karlo et al., 2014; Penge et al., 1999; Penge et al., 1993; Rank-Friend and Elders, 2004; Roberts et al.,

1993; Rossland et al., 2013; Sears et al., 1993; Smith et al., 1993; Stewart and Clark, 1999; Stewart et al., 1999; Wakefield et al., 1993; Young et al., 2012; Ziegler, 1990b; Ziegler, 1992; Ziegler and von Hoorn, 1989). Consequently, the petroleum system in this basin, including petroleum generation, accumulation and preservation, is complex. More specifically, the CG is characterised by a wider range of structural, stratigraphic and combination traps than encountered elsewhere in the North Sea Basin (e.g. Rattey and Hayward, 1993; Sears et al., 1993; Stewart et al., 1997; Wonham et al., 2002). For this reason, this basin also has the largest volume of yet-to-find hydrocarbons in the North Sea, far exceeding the remaining potential from within more established areas such as the East Shetland Basin and the Viking Graben (Eriksen et al., 2003; Munns et al., 2005; Vining et al., 2005). This is elegantly highlighted by comparing the ‘creamed’ discovery profile of the Middle Jurassic play in the Northern North Sea with the ‘uncreamed’ profile of the Upper Jurassic of the Central North Sea (Eriksen et al., 2003).

Despite several extensive studies on the tectono-stratigraphic evolution of the CG there remains disagreement as to whether halokinesis is controlled by rifting, including basement-linked structures, or whether it is largely controlled by more thin-skinned-related tectonic processes (e.g. Bartholomew et al., 1993; Bishop et al., 1995; Stewart et al., 1997; Williams, 1993), although basement-involved wrench interpretation is no more preferred based on recent clearer seismic data. Part of this uncertainty stems from the fact that many previous regional studies on the structural evolution of the area have mainly utilised 2D seismic reflection data, and often with emphasis on specific parts of the basin (e.g. Hodgson et al., 1992; Penge et al., 1993). Consequently, there is a range of tectono-stratigraphic models, based on different assumptions concerning rift-salt interactions, that inform and guide tectonic and stratigraphic interpretations, most notably the following (Fig. 3.2**Error! Reference source not found.**): (1) the ‘pod-interpod’ model (Hodgson et al., 1992; Smith et al., 1993); (2) the ‘rift-raft’ model (Penge et al., 1999; Penge et al., 1993); and (3) the

'salt dissolution' model (Clark et al., 1999; Wakefield et al., 1993). Some of these models were incorporated to varying degrees in the last major regional tectonic syntheses, which were published >10 years ago (Coward et al., 2003; Fraser et al., 2003; Goldsmith et al., 2003; Zanella and Coward, 2003). Other models that have been proposed are hybrid between the above models, such as the model by Høiland et al. (1993) which invokes extension followed by dissolution and the model by Jackson et al. (2010) that incorporates dissolution and differential erosion at different stages of the basin evolution. However, both of these models were also constructed based on observations on specific area of the basin and may not be applicable in other areas.

Additionally, rift-related tectonic or structural models of the CG are also variable, based mainly on a range of inferred stress vectors during Mesozoic extension (Fig. 3.3) (Erratt et al., 1999; Roberts et al., 1990b; Zanella and Coward, 2003). Since these studies were completed, many new and larger 3D seismic reflection surveys have become available, offering new opportunities to evaluate the tectono-stratigraphic evolution of the whole basin at a wider range of scales, and at a higher imaging resolution, than was previously possible. In addition to new seismic data, results from new wells drilled in the basin during the past 10-15 years have also shed further light into the geological evolution of this complex area. Indeed, exploration drilling during the past 10-15 years also indicates significant lateral variability in the tectono-stratigraphic composition of the CG, which existing geological models fail to address.

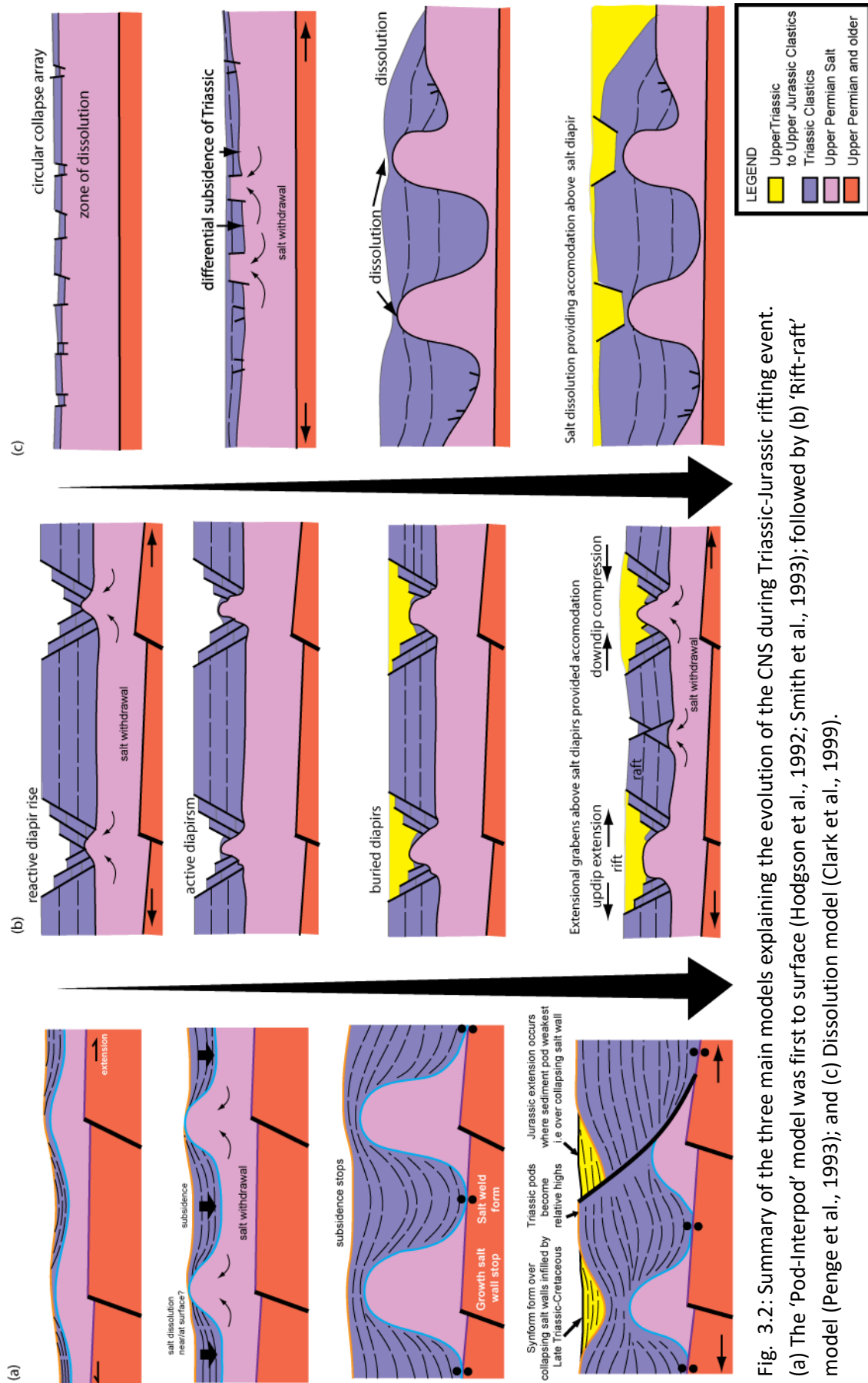


Fig. 3.2: Summary of the three main models explaining the evolution of the CNS during Triassic-Jurassic rifting event.
 (a) The 'Pod-Interpod' model was first to surface (Hodgson et al., 1992; Smith et al., 1993); followed by (b) 'Rift-raft' model (Penge et al., 1993); and (c) Dissolution model (Clark et al., 1999).

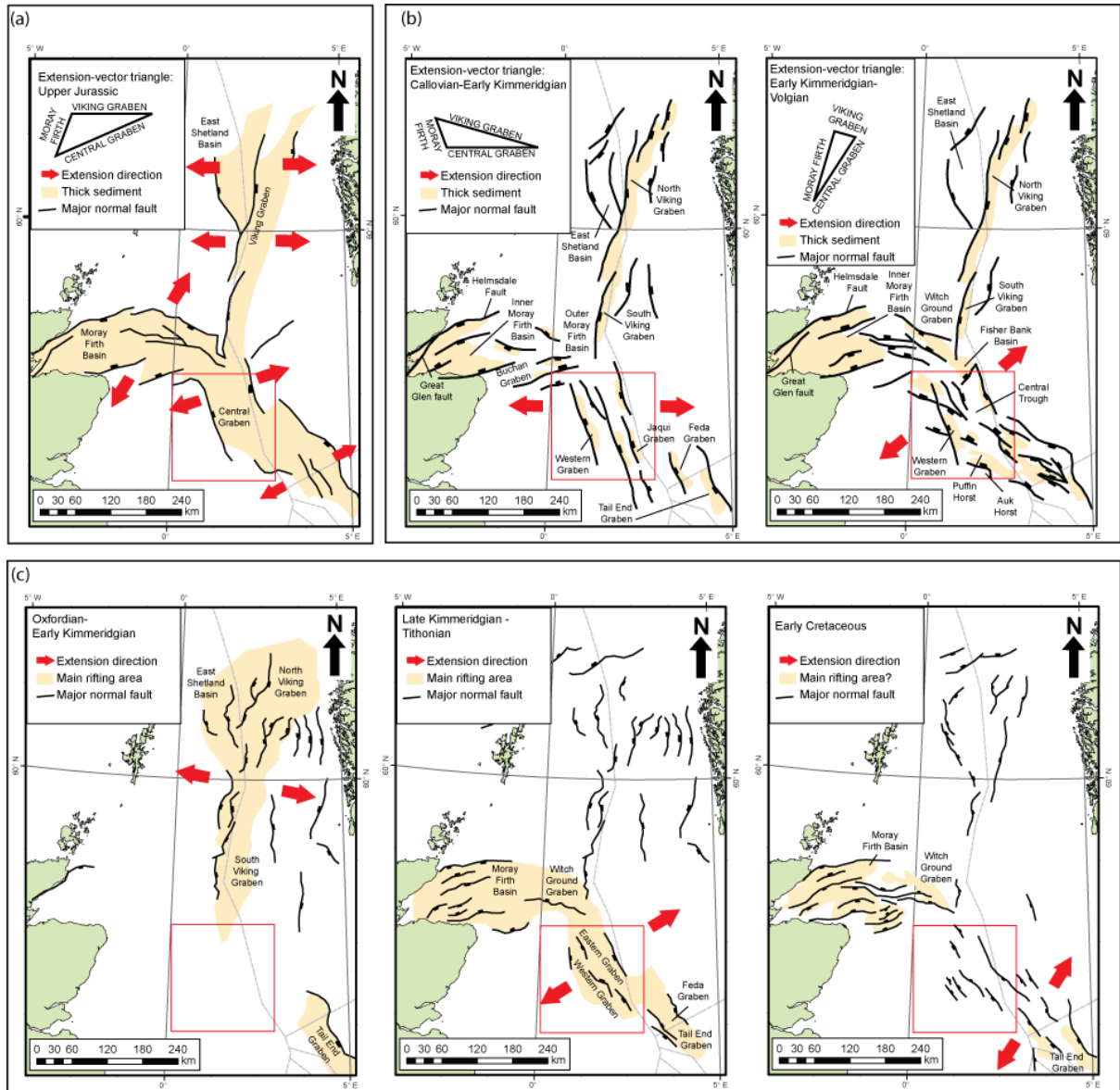


Fig. 3.3: Jurassic rift models. (a) Single-phase model (Roberts et al., 1990b) where three rift systems were spreading simultaneously. (b) Two-phase model (Erratt et al., 1999) where two rift systems occurred at different time. (c) Three-phase model (Zanella and Coward, 2003) with different extension regime occurring in different areas, at different time. Early Cretaceous faults and possible rifting areas are based on Copestake et al., (2003). Study area is in the red box.

In this paper, we focus on the nature of polyphase deformation in salt-influenced rift basin, with emphasis on the interaction between synchronous syn-rift extension and halokinesis during basin evolution, using the data-rich CG as a case study (Fig. 3.1). The three main objectives of this study are to: (1) document the basin-scale structural styles at a range of scales, including both sub- and

supra-salt levels; (2) assess the geometry and evolution of Mesozoic structures, unravelling the relative interaction of rifting and halokinesis on these structures and associated stratigraphy; and (3) critically evaluate regional tectono-stratigraphic evolution of the CG basin.

Current understanding of the geological setting of the Central Graben

The geological evolution of the CG comprises three main phases, relative to the major Mesozoic rifting events: (1) Pre-Mesozoic orogenesis; (2) Mesozoic rifting (i.e Triassic & Jurassic); and (3) Post-rift thermal subsidence periods (Cretaceous and younger).

Pre-Mesozoic Orogenesis Period

The basement and pre-Mesozoic structural framework of the North Sea Basin is largely controlled by the Caledonian Orogeny (Ordovician to Devonian), when the northern part of the Laurentia continent collided, first, with the Baltica continent and, second, with the Avalonia micro-continent (Cocks and Torsvik, 2005; Coward et al., 2003; McKerrow et al., 2000; Soper et al., 1992). These collisional events resulted in the formation of a series of NE-striking structural lineaments, where Laurentia collided with both Baltica and Avalonia (Iapetus Suture), and NW-striking lineaments, where Baltica collided with Avalonia along the Tornquist Line (Fig. 4; Lyngsie et al., 2006; Zanella et al., 2003; Ziegler and Dèzes, 2006). Subsequent post-orogenic collapse during the Devonian formed a series of non-marine basins that were filled with continental clastic deposits of the Old Red Sandstone (Marshall and Hewett, 2003; Richards, 1985; Ziegler, 1992).

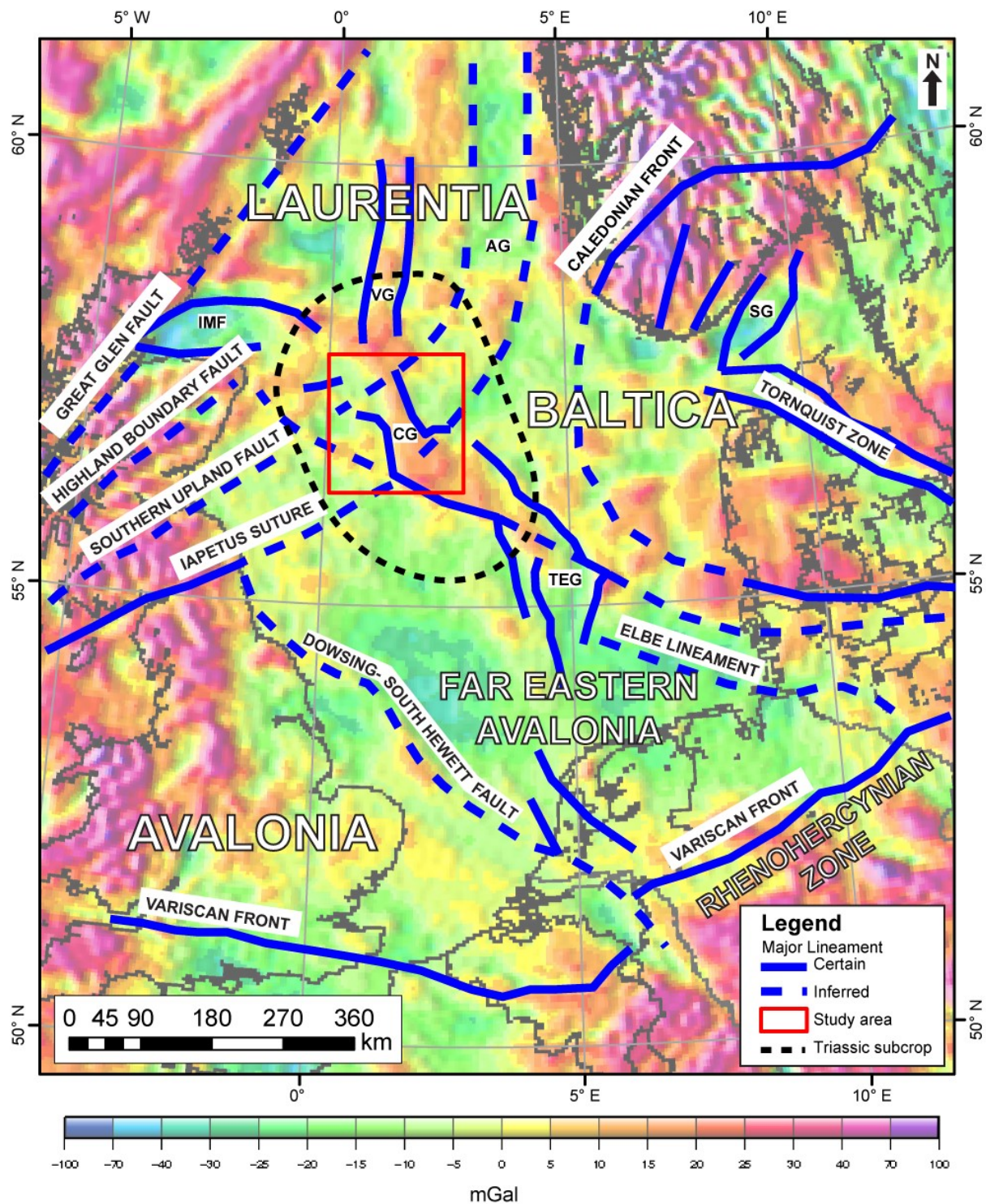


Fig. 3.4: Main structural lineaments in the CNS modified after Lyngsle et al. (2006) over Free Air gravity map (Bonvalot et al., 2012). The main Caledonian Orogeny's micro-continents are labeled in embossed letters. The outline of the Triassic subcrop represents the approximate influence of the Early to Middle Jurassic Thermal Dome, modified after Coward et al. (2003). The main basins are abbreviated as follows: AG=Asta Graben, CG=Central Graben, MF=Murray Firth, VG=Viking Graben, SG=Skagerrak Graben, TEG=Tail End Graben.

During the Late Devonian to Early Carboniferous, marine flooding transformed the North Sea area into either a broad carbonate platform or a clastic shelf, flanked to the south by the Variscan Foreland Basin (Bruce and Stemmerik, 2003; Coward, 1993; Pin, 1990; Ziegler, 1990a; Ziegler, 1992). This shelf was uplifted and eroded during the Late Carboniferous by the Variscan Orogeny, which resulted in the formation of E- to NE-striking faults (Fig. 3.4), which, in the CG, were oriented almost perpendicular to the earlier Caledonian faults (Kroner and Romer, 2013; Peacock, 2004; Stampfli et al., 2013).

Then, the change in the direction of convergence between Gondwana and Laurasia in the latest Carboniferous to early Permian meant that the consequent Variscan Orogeny was associated with transpression and, during this period, old wrench faults associated with the Caledonian Orogeny were reactivated (e.g. the Great Glen Fault; Fig. 3.4; Bartholomew et al., 1993; Ziegler, 1990b; Ziegler, 1992). This additional wrench faulting, together with intrusive and extrusive volcanisms, caused uplift of the entire North Sea area, with erosion down to the Devonian level in some areas (Heeremans and Faleide, 2004; Martin et al., 2002; Ziegler, 1990b; Ziegler, 1992). Transpression and wrench faulting waned during the latter part of the Early Permian and the CG became thermally relaxed and subsided.

Two E-W-striking Permian basins were subsequently formed, exploiting the underlying Variscan trend; these basins were separated by the Mid North Sea-Ringkøbing-Fyn High (Coward et al., 2003; Glennie, 1995; Glennie et al., 2003). Despite adjacent to the Norwegian-Greenland rift to the north (Surlyk et al., 1984) and the Wessex Channel Basin to the west (Chadwick and Evans, 1995), authors disagree on the occurrence of rifting in the CG during the Permian period (e.g. Glennie, 1995; Ziegler, 1992). However, evidence from seismic and well data indicated that rifting had started in Early Permian but became quiescent during the Late Permian (Geluk, 1999; Heeremans and Faleide,

2004; Martin et al., 2002; Neumann et al., 2004; Stemmerik et al., 2000). The Northern Permian Basin, which underlies the superimposed CG, was initially filled with volcanoclastics and then with widespread semi-arid continental clastic deposits (Rotliegend Group; Fig. 3.5; Glennie et al., 2003; Stemmerik et al., 2000). A southwards-directed marine incursion from the Boreal Ocean to the north drowned first the Northern, and then the Southern, Permian basins, resulting in carbonate and evaporate deposition under the semi-arid climatic conditions prevalent at the time (Zechstein Group; Fig. 3.5; Glennie and Buller, 1983). This marine incursion entered the CG from the Norwegian-Greenland rift via the North Irish Platform, possibly exploiting the proto-Viking Graben (Ziegler, 1992).

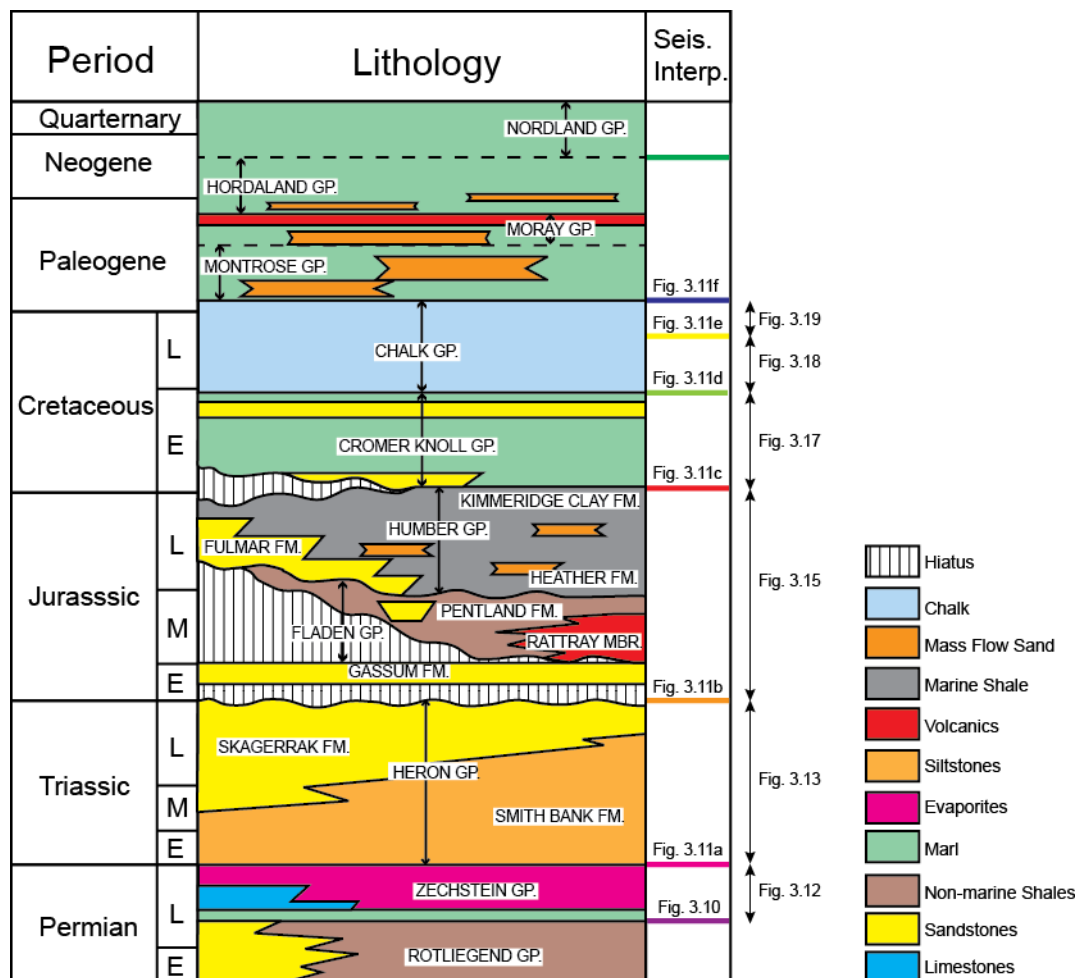


Fig. 3.5: Simplified stratigraphic column of the CNS (after Vining et al., 2005). GP: Group, FM: Formation, MBR: Member

Mesozoic Rifting Period

The earliest Triassic was defined by the onset of rifting, which was initiated when the Norwegian-Greenland Sea rift propagated into the Viking Graben, which generated the pervasive northerly structural trend, including the Horda Platform, Asta Graben and Horn Graben (Færseth, 1996; Glennie, 1995; Goldsmith et al., 2003). There was also some influence from the opening of Neo-Tethys in the south of the CG that propagated through the Polish Trough with NW-SE structural trends (Coward, 1995). The CG, together with the Egersund Basin and the Moray Firth-Witch Ground Graben, were also formed during this Early Triassic rifting period (Ziegler, 1992). Sedimentation during most of the Triassic consisted mainly of semi-arid continental red beds (Smith Bank and Skagerrak formations; Fig. 3.5; Goldsmith et al., 2003). By the Late Triassic to Early Jurassic, shallow marine conditions were gradually established following marine incursions that were again sourced from the Boreal Ocean to the north. At the end of the Triassic, rifting ceased and the rate of rift-related subsidence declined (Goldsmith et al., 2003; Ziegler, 1992).

The geometry of the North Sea during the Early to Middle Jurassic was largely controlled by the formation of a broad dome (the North Sea Thermal Dome) that developed over the Central North Sea area (Underhill and Partington, 1993). Associated with this, a large volcanic centre developed at the triple junction between the future positions of the Central, Viking and Moray Firth-Witch Ground rift arms (Ritchie et al., 1988; Sellwood and Hallam, 1974; Smith and Ritchie, 1993; Underhill and Partington, 1993; Ziegler, 1992). At this stage, the CG area was uplifted and pre-Middle Jurassic strata were deeply denuded, with the erosional products deposited in adjacent subsiding basins in the form of major prograding, sand-rich deltaic complexes such as the Brent Group (Husmo et al., 2003; Rattey and Hayward, 1993; Richards, 1992). The surface associated with the uplift that resulted in the formation of the North Sea Dome is referred to as the Mid-Cimmerian Unconformity or the Intra-Aalenian Unconformity (Davies et al., 1999; Rattey and Hayward, 1993).

Deposition resumed in the CG following the collapse of the North Sea Dome and the onset of renewed rifting during the latter part of the Middle Jurassic (Middle Bajocian and Bathonian), which accelerated during the Late Jurassic (Coward et al., 2003; Rattey and Hayward, 1993). This started with the accumulation of coal-bearing coastal-plain deposits (e.g. Pentland Formation; Fig. 3.5), with deposition interrupted in most places by differential erosion during renewed rifting and formation of the Intra-Callovian Unconformity (Davies et al., 1999). Continued denudation, relative sea-level rise and coastal onlap during the Late Jurassic resulted in the accumulation of widespread sand-dominated, coastal/shallow marine deposits (e.g. Fulmar Formation; Fig. 3.5) and laterally equivalent mud-dominated offshore/shelf deposits (e.g. Kimmeridge Clay Formation; Fig. 3.5; Fraser et al., 2003; Partington et al., 1993b; Wonham et al., 2014). This regionally-extensive period of Late Jurassic (Callovian and Oxfordian) coastal retreat and marine flooding was caused by: (1) reduced rates of sediment supply from the progressively more denuded North Sea Dome; (2) increased rates of eustatic sea-level rise in the adjacent Arctic and Tethys seas; and (3) southward-propagating and accelerating rifting (Fraser et al., 2003; Wonham et al., 2014). The latter is manifested in the Central North Sea by a series of NNW-striking, left-stepping depocenters, most notably the Central and the Tail End grabens (Erratt et al., 1999).

Rifting accelerated and reached its peak during the Kimmeridgian and continued until the Valanginian (Coward et al., 2003; Erratt, 1993; Erratt et al., 1999; Rattey and Hayward, 1993; Zanella et al., 2003; Ziegler, 1992). During this period, deeper water marine conditions were widespread across the CG, with deposition dominated by the accumulation of organic-rich, deep-marine mudstones (e.g. Kimmeridge Clay Formation, Fig. 3.5) and subordinate sandstones (e.g. Ribble Member; Johnson et al., 1986). By the Late Kimmeridgian, the Central North Sea area had become a through-going seaway, which connected the Arctic and Tethys seas (Ziegler, 1992).

Rifting gradually declined in the CG during the Early Cretaceous, although it continued in the Moray Firth (Beach, 1984; Erratt, 1993; Roberts et al., 1990a), and only the master faults delineating the Viking and Central Grabens remained active at this time (Coward et al., 2003). Coarse clastic sediments were mainly deposited as localised wedges against active fault scarps, such as the Wick Sandstones along the margins of the Inner Moray Firth Basin and the End of the World Sandstones beyond the northern edge of the Viking Graben (Copestake et al., 2003).

Post-rift Thermal Subsidence Period

The Early Cretaceous marks the cessation of rifting and the onset of post-rift thermal subsidence in the CG and over most of the North Sea Basin (Coward et al., 2003). The start of this period is marked by the regionally-extensive Base Cretaceous Unconformity (BCU) which is the remnant of post-rift erosion (Kyrkjebø et al., 2004; Rawson and Riley, 1982). Deposition was dominated by regionally-extensive transgressive marine mudstones (Cromer Knoll Group; Fig. 3.5), which accumulated in a large, epicontinental seaway where thin sands were occasionally deposited on an oxic shelf, although a relatively large, sandstone-rich fan system was developed near the Auk High (Copestake et al., 2003). During the latter part of Early Cretaceous, local movement of the main faults occurred, perhaps triggered by compaction of earlier sediments. However, most faults were inactive and gradually overlapped during steady thermal subsidence (Coward et al., 2003).

By Late Cretaceous, a combination of rising eustatic sea-level, and a reduction in siliciclastic sediment supply from the now severely denuded hinterland areas, resulted in the deposition of thick (100's metres) pelagic chalk (Chalk Group; Fig. 3.5; Surlyk et al., 2003). Inversion, associated with early Alpine collision, also occurred at this time (Coward and Dietrich, 1989). In the North Sea Basin,

inversion triggered reversal of fault movement along some of the major basin-defining faults (Coward et al., 2003). Consequently, deeper basin conditions characterised the sites of earlier Mesozoic rift axes, while the former rift shoulders were dominated by shelfal conditions. In the latter, chalk deposition was largely *in situ* (autochthonous chalk facies), but, in the former, a range of gravity flow deposits (redeposited or allochthonous chalk facies) were emplaced (Megson and Tygesen, 2005; Scholle et al., 1998).

Chalk deposition was abruptly interrupted at the start of Palaeocene by an influx of coarse clastic sediments (Montrose Group; Fig. 3.5; Ahmadi et al., 2003); this was caused by major basin margin uplift related to mantle underplating associated with the opening of the North Atlantic and manifested by the Tertiary igneous province of NW Scotland and its surrounding (Lovell, 2010; MacLennan and Lovell, 2002; MacLennan et al., 2003; Mudge and Jones, 2004; White and Lovell, 1997; White, 1989). In the CG, this major uplift event caused an abrupt return to coarse siliciclastic deposition, reflected mainly in the form of Palaeogene sandy basin floor fans and mud-dominated slope and basin plain environments (Fyfe et al., 2003; Reynolds, 1994). A gradual reduction in basin margin uplift resulted in a decrease in fan body dimensions and a change to mud-dominated deposition throughout the CG area (Den Hartog Jager et al., 1993; Jones et al., 2003).

Data and Methods

Data

22,360 km² of 3D seismic reflection data were used for this project, comprising a part of 3D seismic dataset (MegaSurvey) provided by PGS. This dataset covers the area from the Forties-Montrose High in the north to the Auk Ridge to the south (Fig. 3.6) and are suitable for regional structural and stratigraphic interpretation, albeit with variable data quality, polarity and record length due to the

in this study). However, for two areas in the study area, the polarity of the seismic data seems to be the opposite as shown by the 'colour flip' in the instantaneous amplitude map of the seabed reflection (red to blue; Fig. 3.6). Seismic data peak frequency is estimated to be c. 20 Hz at the level of investigation. The horizontal resolution of the seismic data is 50 m x 50 m where the vertical resolution is around 50 m at main levels of investigation (assuming an interval velocity of 4300 m/s and peak frequency of c. 20 Hz).

The study area also contains a large number (a few 1000's) of exploration, appraisal and development wells. For the purpose of this study we have carefully selected 200 wells (Fig. 3.6), representing over 850 km of aggregate length, based mainly on: (1) geographic coverage of the various tectono-stratigraphic domains within the greater CG area; (2) representative stratigraphic sections, especially through the Mesozoic succession; and (3) a rich and consistent database suitable for the study's aims and objectives. Hence, all wells contain a full wireline log suite, including gamma ray, caliper, resistivity, density, neutron and primary (compressional) sonic, which will be useful in interpreting lithology and depositional facies in this study. In addition, direct lithology data are also provided by cuttings and conventional cores (c. 5.8 km in the Jurassic section). In this study, we have completed an independent sedimentological synthesis of 1,570 m of core data. The wells also contain reports dealing with the biostratigraphy, petrography and sedimentology of the key parts of the Mesozoic interval. For well to seismic tie, time-depth curves are provided in the form of checkshots and Vertical Seismic Profile (VSP) surveys for most of the wells.

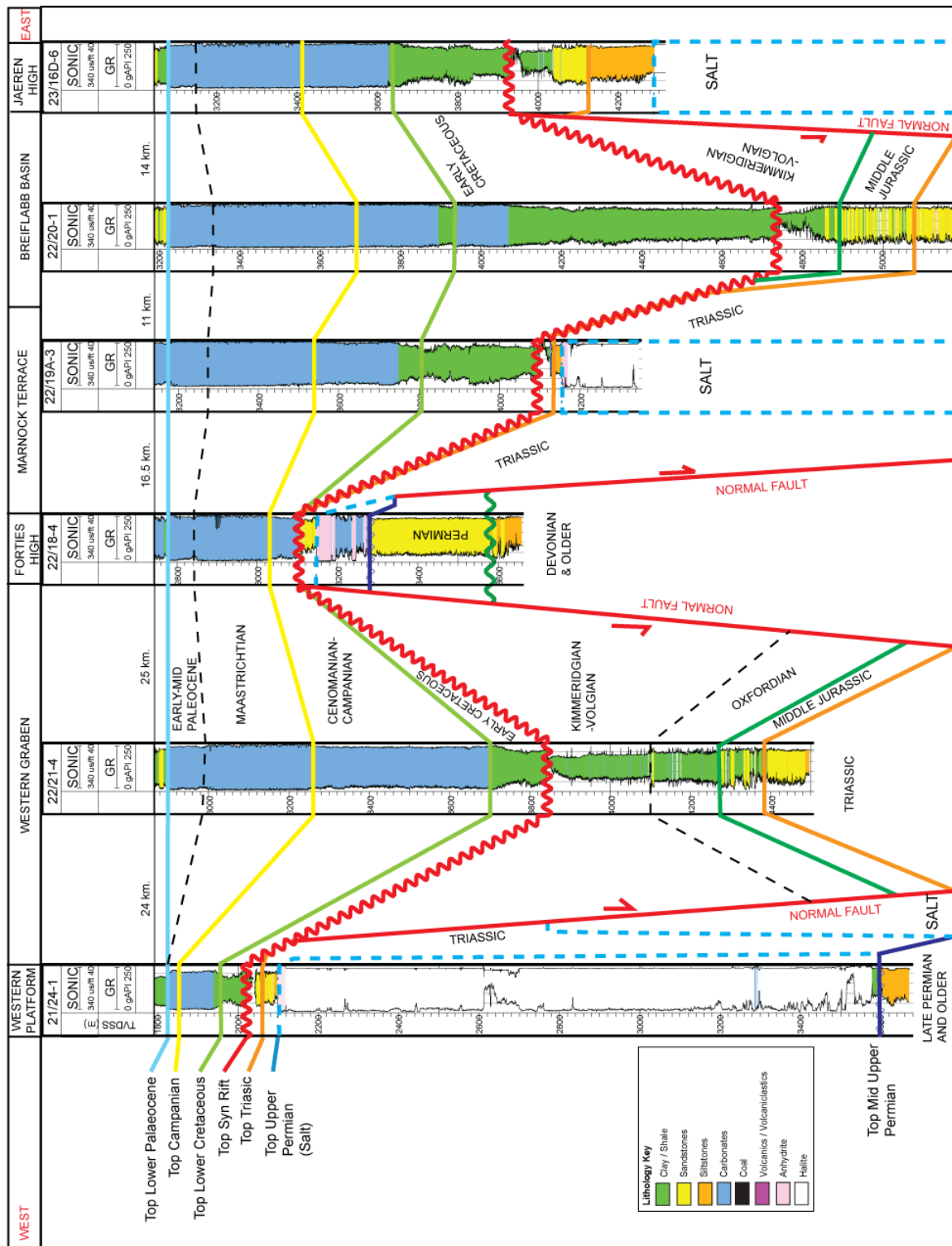


Fig. 3.7: Well section across the northern CG basin datumed on Top Lower Paleocene (Top Chalk). See Fig. 8a for the corresponding seismic line section. Gamma ray logs are on the left and sonic logs are on the right side of the wells. A number of salt diapirs are depicted here in dashed blue line, including the unpenetrated diapir adjacent to the fault to the east of 21/24-1 well.

Methodology

The stratigraphic framework was constructed from well data that were calibrated with biostratigraphic markers and then tied with wireline log signals (e.g. Fig. 3.7) and seismic events using synthetic seismograms (e.g. Fig. 3.9). The synthetic seismograms were constructed using sonic and density logs from the wells, time-depth data and a seismic wavelet extracted from the seismic data. The construction of synthetic seismograms enabled stratigraphic data (e.g. lithology, age) to be confidently tied to seismic response. However due to the regional nature of this study, this exercise was only applied to key wells, selected based on the following criteria: (1) penetration through the most complete stratigraphic successions, which often coincided with the deepest well penetrations; (2) good quality sonic, density and checkshot data; and (3) representative of the various geological domains within the study area. All other wells were tied to seismic data by using only time-depth curves.

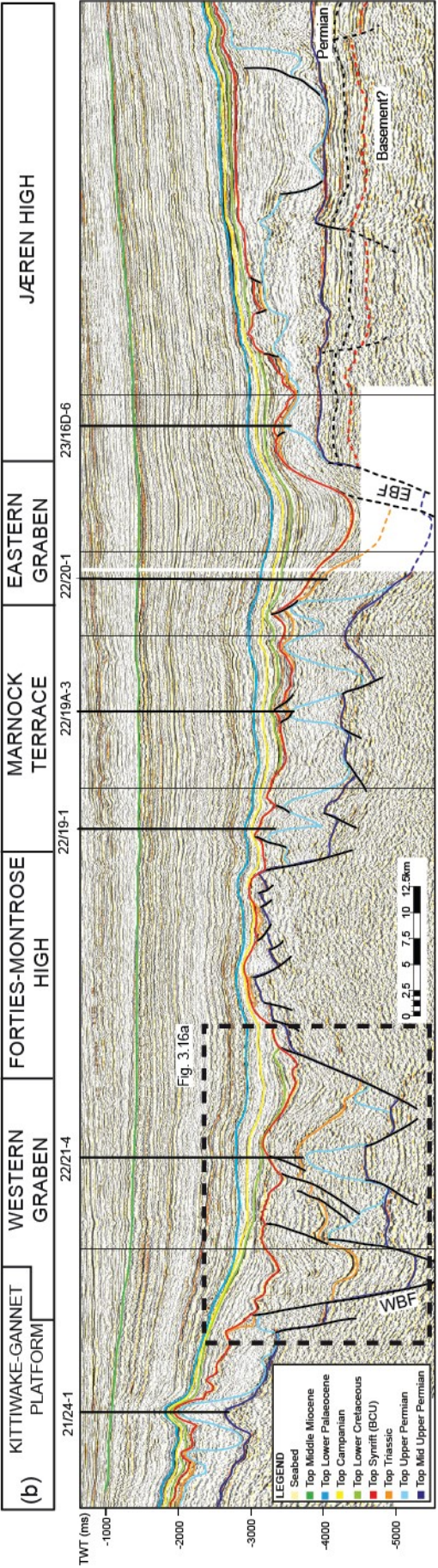
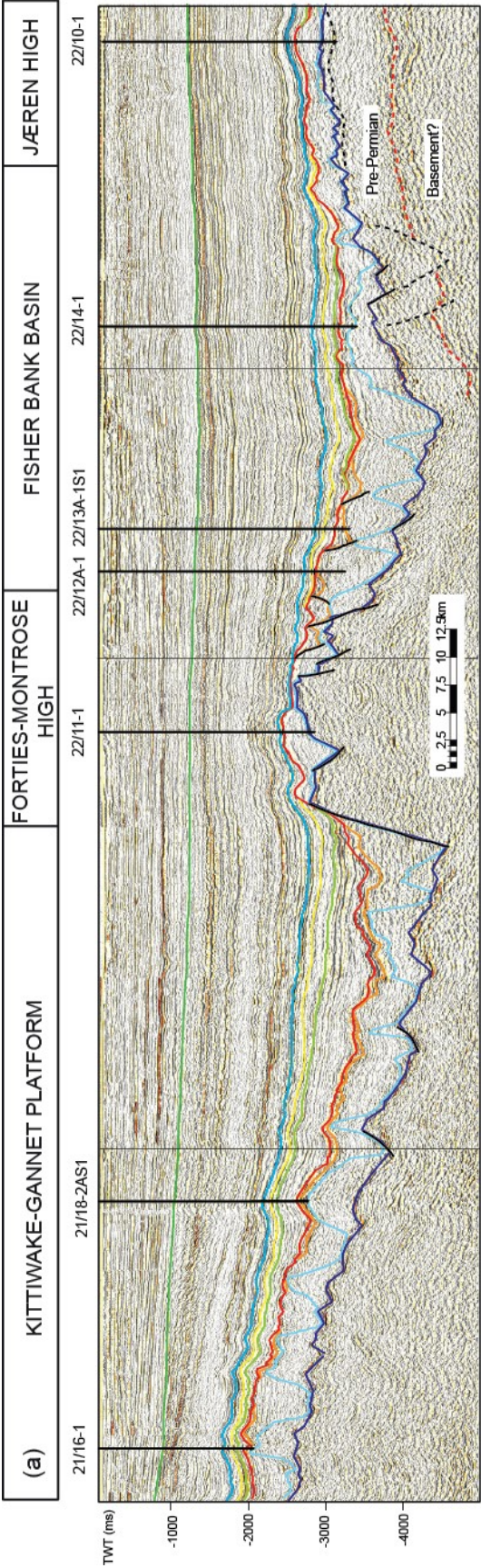
In order to define the regional tectono-stratigraphic framework of the CG the following nine stratigraphic horizons were interpreted using the 3D seismic dataset (Figs. 3.5, 3.8 & 3.9): (1) Top Mid Upper Permian (=Top Rotliegend); (2) Top Permian (=Top Zechstein); (3) Top Triassic; (4) Top Jurassic Syn-rift (=Base Cretaceous Unconformity, or BCU); (5) Top Lower Cretaceous (=Top Cromer Knoll); (6) Top Campanian (=Top Mackerel); (7) Top Lower Palaeocene (=Top Chalk); (8) Top Middle Miocene (=Top Westray); and (9) Seabed. Then, TWT structure maps were constructed for each of the interpreted levels using a 500 m X 500 m grid. The maps were then converted to depth using a velocity model constructed from time-depth data obtained from the wells (Figs. 3.10a & 3.11). For each interval, depth is determined using the following equation:

$$V=kZ+V_0$$

Where V_0 is the initial velocity at the top of the interval and k is the rate of change of velocity with depth (Z). However, the exceptions to this are for the Zechstein salt and Triassic layers. Within the

salt, which is largely incompressible, a constant interval velocity of 4500 m/s was used based on readings from the sonic logs and velocity surveys; explicitly accounting for salt thickness and velocity structure and helped us to reduce the velocity pull-up effect beneath salt diapirs thus giving us a more accurate depth representation of the sub-salt structure. An average velocity of 3890 m/s was used for the Triassic interval, again based on the velocity surveys and sonic logs. This gives sufficient accuracy (within 100 m up and down) for a basin-scale seismic interpretation project.

Finally, palinspastic reconstructions were undertaken by flattening each interpreted stratigraphic interval in order to understand the vertical relationship between successive units. However, this approach is limited in so far as it only shows vertical changes through time; horizontal changes are ignored even though they were probably significant, especially in relation to the Zechstein salt layer, which likely experienced complex, three-dimensional flow (e.g. Buchanan et al., 1996; Rowan, 1993; Rowan and Ratliff, 2012).



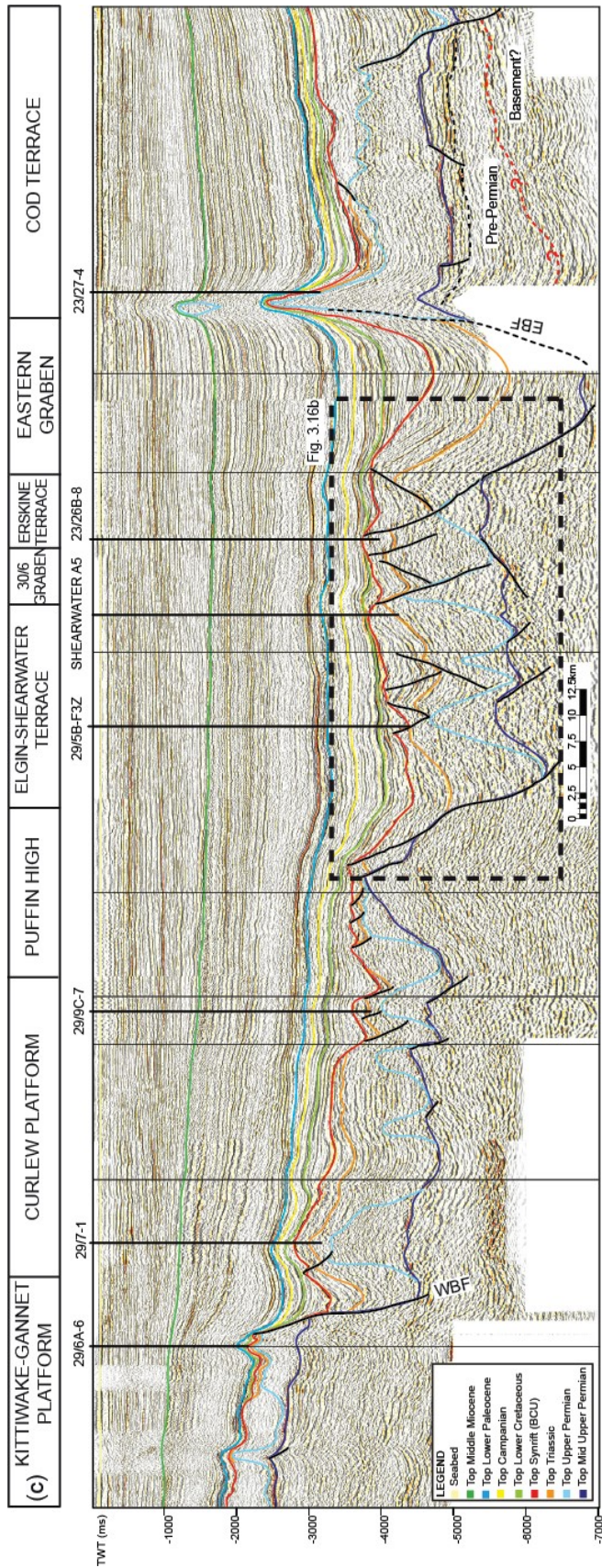
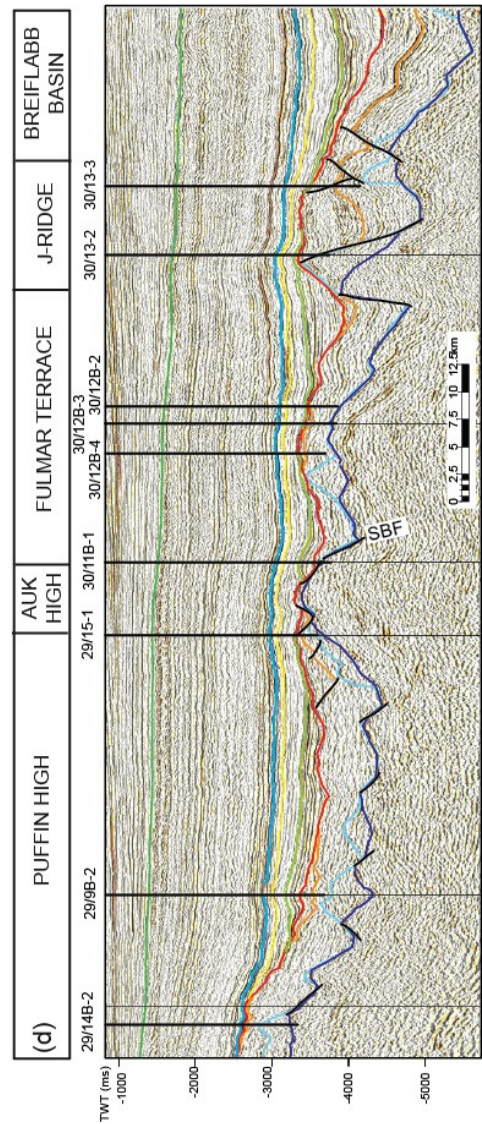


Fig. 3.8: Previous & this page - Seismic sections in the study area. (a) Two simple grabens are present in the north. (b) The grabens have developed into more complex sub-basins. (c) The Western Graben ends and the Eastern Graben becomes deeper. (d) The southern part contains less salt and thinner sediments. The index map shows the line of sections over the Top Rotliegend structure map. WBF=Western Boundary Fault, EBF=Eastern Boundary Fault, SBF=Southern Boundary Fault. For location, see Fig. 3.6.



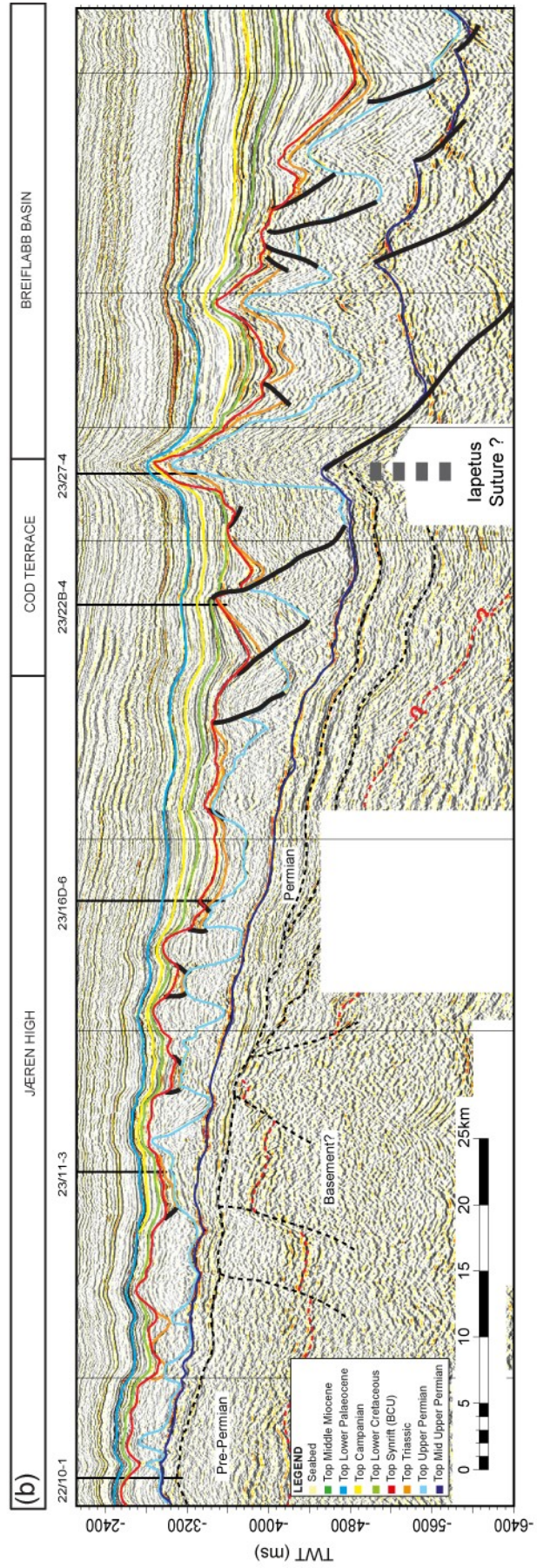
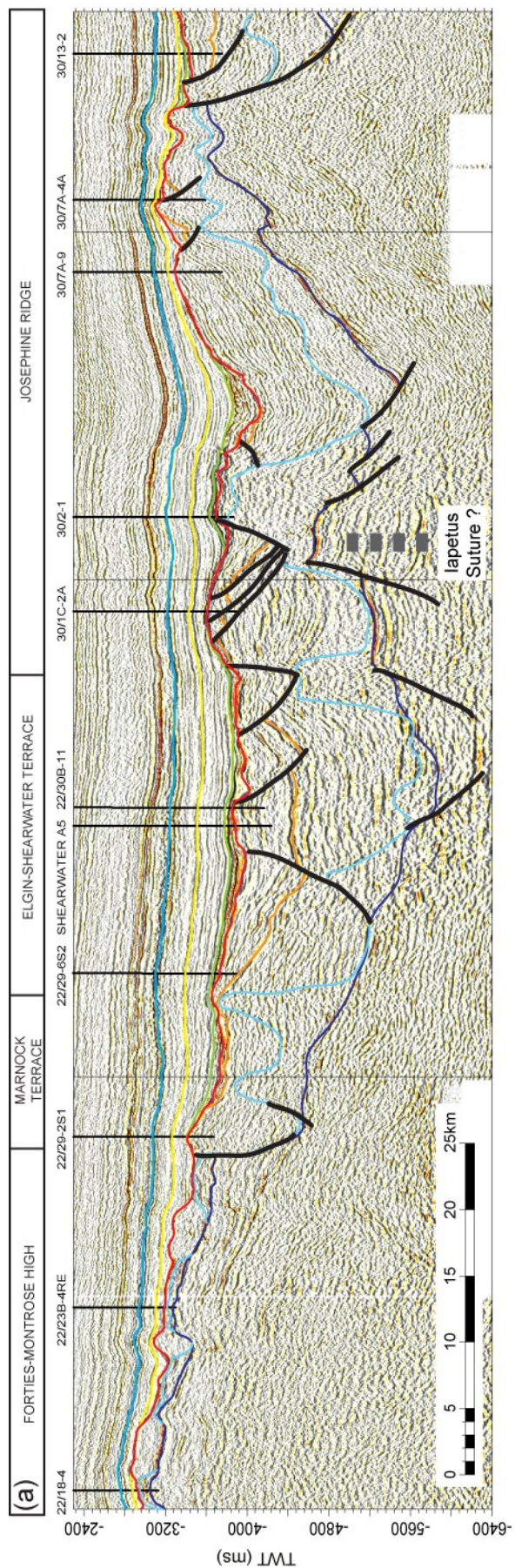


Fig. 3.9: Previous page - (a) NNW-SSE seismic section across the Forties-Montrose High to J-Ridge showing the structures in the middle of the study area. (b) NNW-SSE seismic section across the Jæren High to Breiðflabb Basin showing the structures in the eastern part of the study area. For location, see Fig. 3.6.

Sub-Salt Structures

Observations

Structures at the sub-salt level are best-defined by the Top Mid Upper Permian surface (approximate to Top Rotliegend Group, Figs. 3.5 & 3.7), which is clearly identified in the seismic data by a bright, continuous to semi-continuous hard event, typically underlain by parallel, continuous reflections and overlain by chaotic to transparent reflections (Figs. 3.8a, b, c & 3.9a, b).

There are four main structural trends based on the distribution of fault orientations on the Top Lower Permian surface (Fig. 3.10a, b): (1) a NNW-SSE trend defining the edges of the main platforms and graben; (2) a NW-SE trend occurring mainly in the southern part of the basin; (3) a diffuse, NE-SW trend, cross-cutting both graben and platform areas; and (4) an E-W trend, mainly characterising the north-western part of the Western Platform and the Puffin High.

Consequently, the CG is further divided into seventeen structural domains based on the main fault orientation and their relative elevations with respect to adjacent domains (Fig. 3.10b). For example, the Kittiwake-Gannet Platform (KGP) domain is defined by mostly E-W striking structures whereas the relatively low-lying Western Platform (WP) domain has a NE-SW striking structure (Fig. 3.10b). Wherever possible, we use the domain names identified by previous workers (e.g. Bartholomew et al., 1993; Clark et al., 1993; Erratt, 1993; Erratt et al., 1999; Lasocki et al., 1999; Stewart and Clark,

1999; Zanella and Coward, 2003), introducing new names and redefining structural boundaries as appropriate.

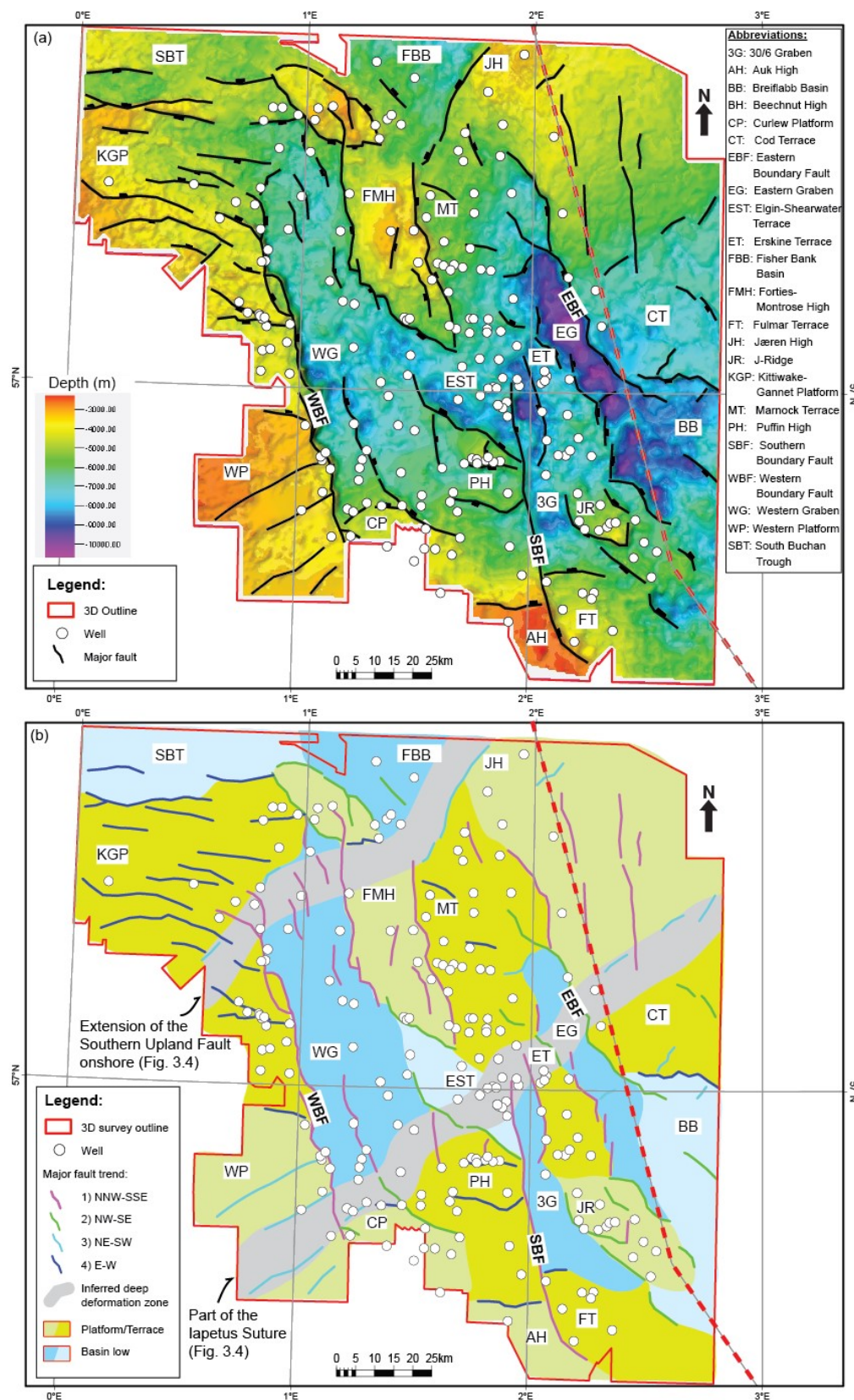


Fig. 3.10: Previous page - (a) Depth structure map of Top Mid Upper Permian (Rotliegend Group) (b) Map of major fault showing multi-azimuth fault trends and possible deep deformation zones in a NE-SW direction.

The main structural trends, NNW-SSE trend defines the dominant basin-bounding faults, most notably the Western Boundary Fault (WBF), the Eastern Boundary Fault (EBF) and the Southern Boundary Fault (SBF) systems (Figs. 3.10a, b; Eggink et al., 1996). Further description of the main faults is outlined below in order to understand their regional geological impact.

The first major bounding fault, the WBF, dips eastward and comprises of four main segments, each about 20-30 km long and up to 5 km of displacement (Figs. 3.8 & 3.10). The northern limit of the WBF terminates abruptly against a set of prominent E-W striking faults but the southern end of the fault system is not clearly imaged within the study area and therefore not fully understood. On the other side of the basin, the EBF defines the eastern edge of the CG, striking parallel to the WBF but with an opposing dip (i.e. to the west; Fig. 3.10). The EBF is divided into two main segments and has a total length of about 75 km with the southern segment accounting for c. 50 km of this length and has the greatest throw (3-4 km). Finally, the third major bounding fault, the SBF, forms the southernmost major fault (c. 70 km long with a throw of c. 3 km) in the study area, extending from the southern tip of the Forties-Montrose High and continuing outside of the southern edge of the study area (Fig. 3.10). In addition to the major bounding faults, smaller and less continuous faults with NNW-SSE trend is also dominant within the Western Graben, the Forties-Montrose High and the Jæren High.

The NW-SE structural trend comprises of somewhat less extensive fault systems, which bound the intra-basinal highs such as the Forties High (northern part of the Forties-Montrose High) and J-Ridge

(Fig. 3.10a, b). The faults are 10-30 km long with throws ranging from several hundred metres to up to 3 km. More rarely, in cases where they have relatively large throws (2-3 km), these faults also form the boundaries to sub-basins: (1) Elgin-Shearwater Terrace, where NW-striking faults define its northern and southern boundaries; and (2) Breiflabb Basin, where a NW-striking fault separates it from the J-Ridge (Fig. 3.10a).

The NE-SW structural trend is even less extensive than the previous two and most clearly defined in four areas: (1) along the southern boundary of the Fisher Bank Basin; (2) the central part of the Kittiwake-Gannet Platform (e.g. Lasocki et al., 1999); (3) the northern boundary of Cod Terrace; and (4) the central part of the Western Platform (Fig. 3.10a, b). Most of the faults defining this trend have throws of only a few hundred metres and lengths of c. 5-10 km and focussed on the two areas described above. However, the faults seem to be aligned and confined in two narrow zones (Fig. 3.10b).

Finally, the fourth structural trend has an E-W orientation and occurs mainly in the Kittiwake-Gannet Platform with a minor expression in the Puffin high (Fig. 3.10a). The faults typically have throws of several hundred metres and are c. 10-20 km long, but may merge to form fault zones >40 km in length (e.g. in the Kittiwake-Gannet Platform). Faults with this trend also form the southern boundary of the South Buchan Trough, the Western Graben and the Puffin High (Fig. 3.10a).

The majority of faults underlying the Zechstein Group do not breach the salt, with exception being the main larger, basin-bounding, NNW-SSE- (WBF, EBF & SBF) and NW-SE-striking fault systems (e.g. northern boundary of Puffin High) (Figs. 3.8 & 3.9). These faults extend upward from Permian levels,

through the salt and Triassic-Jurassic overburden, terminating near the Base Cretaceous Unconformity (Figs. 3.8a & 3.9a).

The gross thickness of Permian beneath this surface cannot be measured due to poor seismic image quality. However, based on well data the Permian thins towards the Montrose and Jæren Highs, where the Permian interval is 200-300 m thick (eg. well 22/18-4 in Fig. 3.7 and well 22/10-1 in Figs. 3.8a & 3.9b). In contrast, well data confirms that the Permian is more than 1 km within the graben axes (Glennie et al., 2003). The internal seismic character is uncertain, but in the Jæren High the Rotliegend Formation comprises continuous, parallel and high amplitude reflections overlying chaotic, low amplitude, diverging reflections (Fig. 3.9b).

Interpretation

The Top Lower Permian depth-structure map represents an amalgamation of features formed during the tectonic events that have affected the CG (Fig. 3.10a). A comparison of the structures presented here with those from other regional studies suggests that there is similarity between the fault orientations at the Mid Upper Permian level with those ascribed to pre-salt tectonic events and will be elaborated below (Coward et al., 2003; Glennie et al., 2003; Lyngsie et al., 2006; Zanella et al., 2003; Ziegler, 1992).

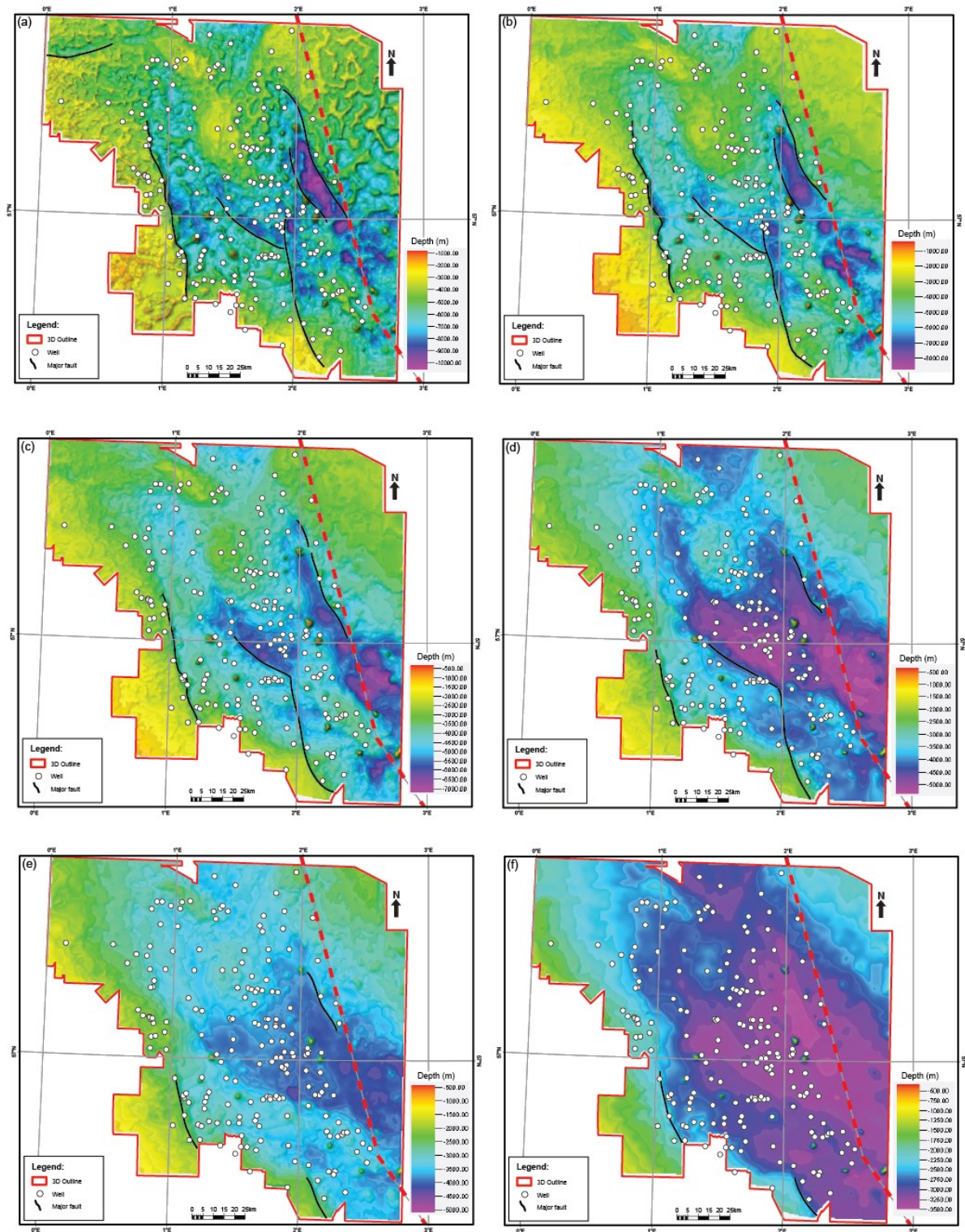


Fig. 3.11: Depth structure maps: (a) Top Permian (Zechstein); (b) Top Triassic; (c) Top Jurassic Syn-rift/Base Cretaceous Unconformity (BCU); (d) Top Lower Cretaceous (Cromer Knoll); (e) Top Campanian (Mackerel); (f) Top Lower Palaeocene (Chalk).

Firstly, the NW-SE trend may reflect the NW-SE-striking Avalonia-Baltica Suture, including the Tornquist and Elbe lineaments, which is a major gravity anomaly across the CG that extends southwards into mainland Europe (Fig. 3.4). Therefore, it is not surprising that this zone of weakness was exploited and gave the basin its overall present day shape (Fig. 3.1).

The NE-SW trend reflects the structures related to the Iapetus Suture (Figs. 3.4 & 3.10b). The NE-SW striking faults in the study area are aligned with the offshore extension of the Southern Upland Faults and the Iapetus Suture identified from gravity data (Figs. 3.4 & 3.10b). Both of these deep, basement-involved trends divide the CG into its larger-scale structural features: (1) defining the northern and southern ends of the Western Graben; (2) resulting in segmentation of the Forties-Montrose High; and (3) defining the separation between the relatively simpler horst and graben in the north (Fig. 3.8a, b) and the more complex structures in the south (Fig. 3.8c & 3.9a, b).

The E-W trend is similar to the Variscan trend (Lyngsle et al., 2006), and the CG is situated within the palaeo-realm of Variscan deformation (Glennie et al., 2003). However, the present E-W structural grain could also be younger and somehow influenced by the adjacent Outer Moray Firth Basin to the north, represented here by the South Buchan Trough, which contain predominantly Late Jurassic faults of similar strikes (Fig. 3.10a).

In addition, the NW-SE, NE-SW and E-W fault patterns also define the main structural highs and adjoining basins (e.g. Western Platform, Forties-Montrose High, Jæren High and Cod Terrace) and do not usually breach the Zechstein salt layer (Fig. 3.8a). We interpret that these faults were largely inactive during the later Mesozoic rifting periods although it is plausible that subsalt faulting was expressed and accommodated by folding of the overburden. In contrast, the major NNW-SSE striking

faults (WBF, EBF & SBF) that bound the basin's major graben have breached the salt (Figs. 3.8a, b, c). This may be attributed to extension events during the latest Triassic and/or Jurassic rifting; this will be discussed more below.

The internal seismic character below the Mid Upper Permian surface represents sandy Rotliegend Formation (e.g. well 22/18-4; Fig. 3.7), which has parallel seismic reflector overlying more chaotic and divergence reflectors of the Pre-Rotliegend intervals, containing Early Permian and older rock units (Fig. 3.9b). The observation for Rotliegend Formation is consistent with post-rift depositional setting whereas both Devonian and Carboniferous structures have a NE-SW trend (e.g. Bruce and Stemmerik, 2003; Geluk, 1999; Stemmerik et al., 2000).

Salt Structures

Observations

The present day salt structure is interpreted from the Top Permian surface (Top Zechstein Group; Figs. 3.5 & 3.11a), which is usually expressed as a bright hard event; the salt itself is characterised by largely chaotic, transparent seismic reflections (Fig. 3.8c). The Top Permian seismic reflection can be difficult to pick due to: (1) the presence of multiples; (2) the sub-vertical nature of the flanks of salt diapirs; and/or (3) the low acoustic impedance contrast between the mudstone-dominated Triassic Smith Bank Formation and the salt (Figs. 3.8b, c & 3.9a, b). Hence, the interpretation of this interval was guided by a combination of criteria: (1) the presence of evaporites in well data; (2) the presence of anticline in the overlying layers indicating the present of a salt wall underneath; and (3) the strength and continuity of the underlying Top Mid Upper Permian seismic event. Whereas the first and second criteria are obvious, in the third criterion, the Top Mid Upper Permian seismic event is weaker and less continuous where salt is thick (Figs. 3.8b, c). In contrast, the Top Mid Upper Permian

seismic event is bright and more continuous where salt is thin or beyond the tuning thickness of the seismic data (Fig. 3.8c).

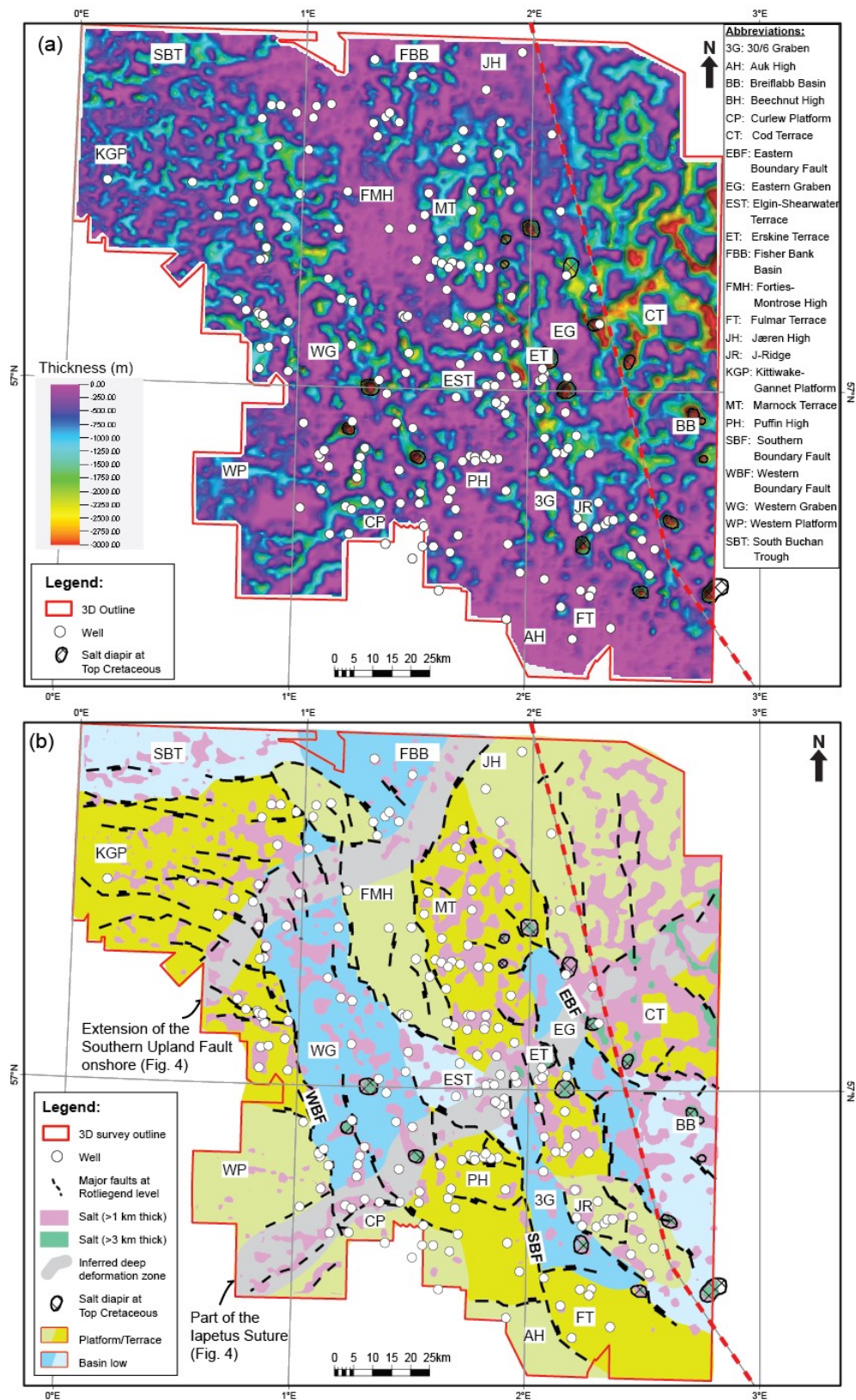


Fig. 3.12: Previous page - (a) Upper Permian (Zechstein Group) isochore map (b) Map of major salt bodies overlain over faults from the Mid Upper Permian map (Fig. 3.10). There are some similarity between the trends of salt and the underlying major faults.

An isochore map between the Top Permian and Top Mid Upper Permian surfaces defines the present-day geometry and distribution of salt bodies (Fig. 3.12a). Throughout the CG, salt has deformed mostly into a series of elongated salt walls (c. 2-3 km wide, a few km to more than 10 km long). The height of these salt walls is extremely variable (100s m to 1000s m) but the largest are concentrated within the eastern part of the CG, especially in the Cod Terrace, and the smallest are mostly found in the western area, most notably in the Kittiwake-Gannet Platform (Fig. 3.12a & b). In addition, the salt walls appear to have been welded in most areas (Figs. 3.8 & 3.9).

The trends of the salt walls broadly reflect the trends of the underlying major sub-salt faults and can be grouped by the sub-salt structural domains (Fig. 3.12b). For example, salt walls in the Kittiwake-Gannet Platform trend parallel to the E-W striking faults in the sub-salt level, whereas in the Western Platform, they parallel the NE-SW striking faults (Fig. 3.12b). In the Western Graben, salt walls trend NNW-SSE whereas in the adjacent Elgin-Shearwater Terrace, the trend is more NW-SE, parallel to the underlying fault-controlled structural grain (Fig. 3.12b). Furthermore, tall salt walls in the eastern part of the study area are also aligned with deeper structures, such as in the Cod Terrace, where the edge of the thickest salt walls coincides with the inferred NE--striking Iapetus Suture (Fig. 3.12b). In addition, within the Cod Terrace, the NNW-SSE and NE-SW oriented salt walls interact to form broadly polygonal pattern.

There are several salt stocks rising from the edges of present depocentres, especially in the Western and the Eastern grabens (green polygons; Fig. 3.12b). These are more common in the central part of

the basin, where they are up to 6 km tall and 2-5 km wide (Fig. 3.12a) and some of these salt diapirs have pierced Tertiary strata and locally have 'tear-drop' shaped cross-sectional geometries (Fig. 3.8c).

Finally, no significant variations in salt occur on the Forties-Montrose High, northern Jæren High or Auk High, in addition to parts of the Western Platform, Puffin High and Fulmar Terrace (Fig. 3.12b). Well data also show that the salt is generally absent and the Zechstein Formation composed of anhydrite and carbonate in these areas; halite is generally absent (Fig. 3.7).

Interpretation

The size, geometry and distribution of salt walls partly reflect the original pre-halokinesis salt volume. Assuming that the salt walls have undergone little dissolution, the largest salt walls would be expected to have developed closest to the areas containing the greatest original salt volume. This implies that the middle of the CG area, where the Zechstein salt is presently thickest (e.g. in the south of Western Graben and Cod Terrace; Fig. 3.12b), was also close to the centre of the Northern Permian Basin, which was oriented NE-SW and deepened towards the NE (Fig. 3.1; Coward et al., 2003; Ziegler, 1992). However, the decrease in overall salt thickness in the northern part of the study area may also reflect Early to Middle Jurassic uplift, when thermal doming resulted in regionally-extensive erosion (Underhill and Partington, 1993). This may be supported by the vertically shorter salt walls in areas such as Jæren High retaining similar lateral width to the taller salt walls in the Cod Terrace (Figs. 3.9b & 3.12a).

In addition to the doming effect, the absence of salt on prominent highs (e.g. the Forties-Montrose High and Auk High; Fig. 3.12b) could also be caused by either non-deposition on Late Permian

palaeo-highs, or removal of previously deposited salt by post-depositional salt withdrawal or dissolution, especially in patches where there are virtually no salt walls. Non-deposition of salt is our preferred interpretation, partly supported by the presence of anhydrite and carbonates as shown by the well data that indicate deposition in shallow water environment (Fig. 3.7). On the other hand, removal of salt might also be plausible since salt welds seen in seismic data on the highs may have developed in response to Triassic salt withdrawal (Fig. 3.8a, b). Salt dissolution may also have occurred, presumably at the crests of mature diapirs during Triassic, and locally Jurassic, exposure of the basin (Clark et al., 1999; Wakefield et al., 1993). However, its impact on the regional thickness of salt is thought to be minor compared to salt withdrawal and the growth of diapirs.

In the central area of study, some of the salt stocks have risen above the Base Cretaceous Unconformity (green polygons in Fig. 3.12b). These tall salt stocks are located along the boundaries of major structural domains, where the underlying structures in the Upper Permian are more complex, such as along the western boundary of Eastern Graben (Marnock Terrace to J-Ridge area) (Figs. 3.8c, 3.9b & 3.12b). At these locations, salt bodies are confined to narrow belts of thick sediments that could have compressed the salt (see Fig. 17 in Stewart et al., 1996). This could cause the salt to rise in a narrow, isolated fashion and later lead to the formation of detached salt bodies such as tear-drop shaped diapirs during the basin inversion in the Cretaceous (Figs. 2.3 & 3.8c; Dooley et al., 2009; Hudec and Jackson, 2007; Vendeville and Nilsen, 1995).

Triassic Structures

Observation

The Top Triassic surface is often overlain by bright, continuous seismic reflections (Figs. 3.8b, c, d & 3.9a, b), which are tied to the Middle Jurassic coals (Fig. 3.7). The relatively continuous nature of this

event allows it to be used as a proxy for the Top Triassic throughout most of the 3D seismic survey area. However, this reflection is less clear on the platform highs (e.g. Western Platform and Jæren High; Figs. 3.8a, d.) due to a combination of the overall thinning of the Jurassic interval and truncation by the Base Cretaceous Unconformity (BCU). In these areas, the Top Triassic has been interpreted as the base of a relatively continuous seismic package (representing Jurassic sediments) or where truncated, as the BCU (e.g. on the Jæren High; Fig. 3.8a, b).

The thickness of Triassic strata is highly variable (0-3.5 km) but, generally, thicker intervals occur within the main Mesozoic depocentres, such as the Western and Eastern grabens (Fig. 3.13a). However, the thickest parts of Triassic (>3.5 km) are located on the Cod Terrace (Figs. 3.8c, 3.9b & 3.13a). At a smaller scale, the thicker Triassic intervals are also organised into a 5-10 km-diameter circular 'pods' throughout the main depocentres (e.g. in Cod Terrace; Fig. 3.13a). Some of these are relatively elongated (up to 20 km in length) where they occur between, and have their long axes orientated parallel to, flanking salt walls. The Triassic interval is thinnest (<300 m) on top of salt walls and other structural highs (e.g. Forties-Montrose High, northern Jæren High, northern Western Platform and Auk High areas; Figs. 3.9a, b & 3.13a, b).

In relation to the major faults, there is a large increase in Triassic thickness (c. 1-2 km) from the footwall to the hangingwall of the major NNW-SSE striking faults, especially the WBF, EBF and SBF (Fig. 3.12a). In addition, smaller but significant footwall to hangingwall expansion (c. 0.5-1 km) also occurs across the NW-SE and NE-SW and E-W striking faults. The most prominent are the NW-SE striking faults, which are cut by the NNW-SSE major bounding faults, such as between the Puffin High and J-Ridge (Fig. 3.13a). In contrast, the NE-SW striking faults are restricted to the Western Platform, Cod Terrace and the Fisher Bank Basin (Fig. 3.13a). E-W striking faults are more prominent in the South Buchan Trough (Fig. 3.10a).

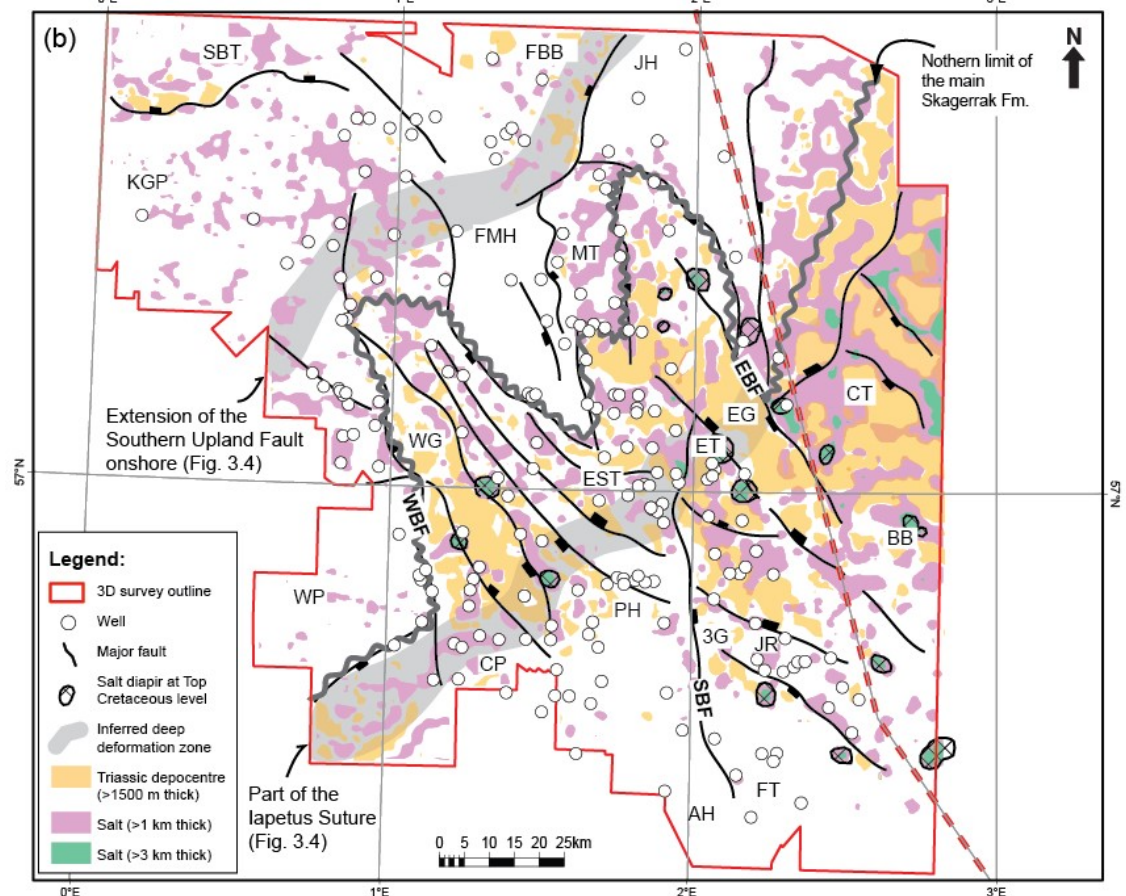
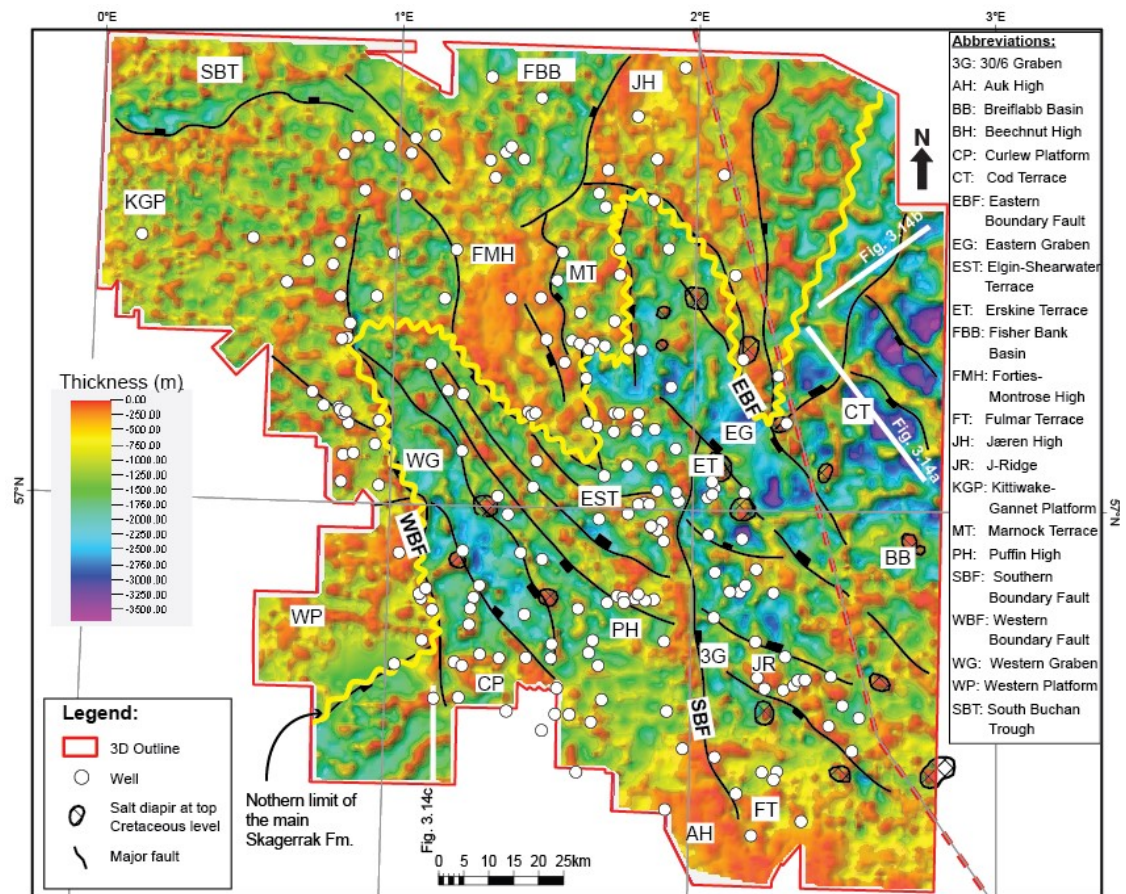


Fig. 3.13: Previous page - (a) Triassic isochore map. (b) Composite map showing the spatial relationship of the Triassic depocentres with salt bodies and older deep deformation zones. The Triassic seems to be thickening northeastwards and eroded northwards (e.g Mckie, 2014). See discussion in text.

Prominent divergent and erosional features in the Triassic on the Western Platform, Cod Terrace and Jæren High define five seismic units (Fig. 3.14): (1) Unit A contains low amplitude, parallel reflections that often follow the topography of the salt and are truncated by overlying units; (2) Unit B also contains low amplitude, parallel reflections, which are, in places, chaotic and divergent; (3) Unit C contains chaotic and low amplitude reflections that may be either divergent or truncated by the overlying unit; (4) Unit D contains high amplitude, diverging reflections, which are often truncated by the underlying unit; and (5) unit E comprises high amplitude, parallel but less continuous reflections, which onlap onto Unit D.

In the Cod Terrace and Jæren High areas, the base of Unit D approximates to the base of the sand-dominated Skagerrak Formation and is truncated by the BCU (Fig. 3.14a). The Skagerrak Formation extends over much of the study area (Fig. 3.13a) and its northern limit coincides with that of the thick Triassic depocentres (Fig. 3.13b). Unit D diverges towards the southwest in the Cod Terrace and Jæren High areas (Fig. 3.14b). This direction changes to the northeast in the Western Platform (Fig. 3.14c), where the unit is also deeply truncated by the BCU. However, further mapping of these intra-Triassic units is compromised by multiples below the BCU and by the relatively poor signal in the deeper graben areas (Fig. 3.8).

Interpretation

Triassic thickness changes were influenced by coeval halokinesis and rifting, as shown by the close association between salt walls and large extensional faults (Figs. 3.12b). The overall shape of the

Triassic succession also follows closely that of the Zechstein salt basin, as indicated by the NE-SW trend of the basinal deep, suggesting continued influence of the preceding basin-fill (Figs. 3.12b & 3.13b). However, as in the Zechstein interval, there could also be an erosional effect caused by the Early to Middle Jurassic North Sea uplift (Underhill and Partington, 1993). This is supported by the erosion of most of Skagerrak sands in the northern part of the study area (Erratt et al., 2010; McKie, 2014).

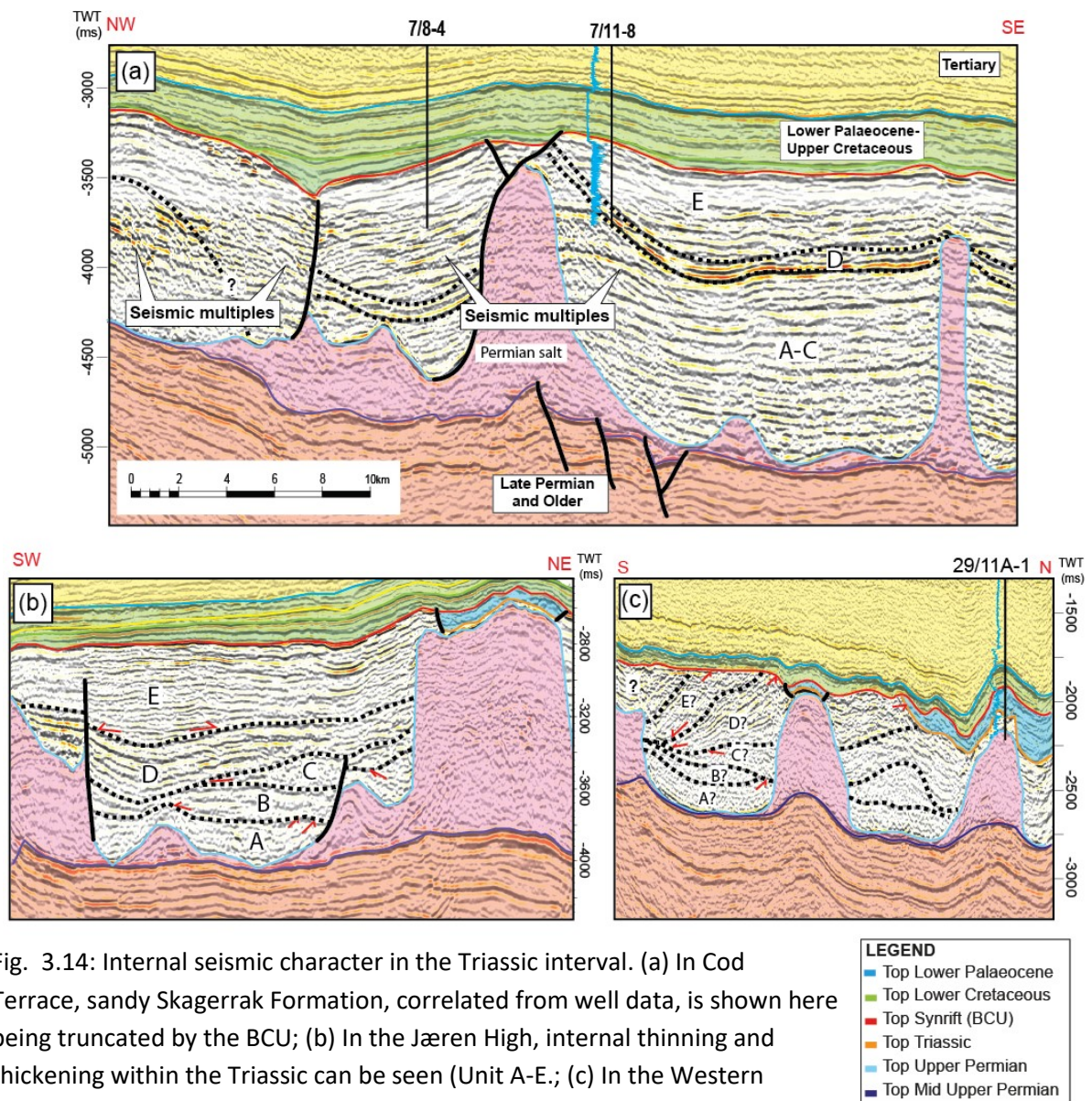


Fig. 3.14: Internal seismic character in the Triassic interval. (a) In Cod Terrace, sandy Skagerrak Formation, correlated from well data, is shown here being truncated by the BCU; (b) In the Jæren High, internal thinning and thickening within the Triassic can be seen (Unit A-E; (c) In the Western Platform, the upper units are truncated by the BCU. However, the age of the Triassic here is unclear, although well 29/11a-1 has penetrated what has been interpreted as the Smith Bank Formation before going into the Zechstein Salt. Gamma-ray log is shown here as blue line to the left of well. For location, see Fig. 3.13a.

The pronounced NNW-SSE striking boundary faults (Fig. 3.13b) conform to both the regional Triassic rift trend in the Norwegian sector and the later Jurassic rift trend (Figs. 3.1 & 3.15a, b; Erratt et al., 1999; Færseth, 1996; Zanella and Coward, 2003). However, this trend is interpreted to mainly reflect the overprinting of Jurassic rift onto the Triassic structural grain. This supported by the presence of structures seen within deeper part of the basin, which has not been completely overprinted by Late Jurassic footwall erosional effects as well as SW-divergence character of the Triassic interval (Figs. 3.13a & 3.14b; further elaborated in the discussion section).

The effect of halokinesis on the Triassic sediment accumulation was strong, as shown by the influence of pods (minibasin) and salt walls (Figs. 3.8, 3.9 & 3.13). The subsidence evolution of the pods is suggested by the changing seismic character of the sediment fill inside the pods. The parallel character of Unit A, which lies directly above salt welds in grounded minibasins, could reflect a Triassic sediment cover above pre-kinematic salt. Hence, during Triassic rifting, extension of the pre-existing Triassic sediments (Unit A) capping the salt layer could have caused salt to rise reactively creating depressions where the surrounding salt stock has withdrawn. These depressions provided the accommodation spaces for further Triassic depositions and caused more salt withdrawals until the depression became grounded. The moderate divergence patterns in Units B & C might represent early syn-rift deposition, whereas the strongly divergent Unit D may have been deposited during the main syn-rift period (Figs. 3.14b, c). Finally, Unit E could represent a post-rift succession, which is supported by the parallel reflectors and onlap base. Although these syn-rift Triassic units (B-D) appear to be punctuated, they were not restricted by the salt walls (Figs. 3.14a, b, c). Hence, halokinesis might have had a relatively localised impact on sediment accommodation. However, the main stratigraphic boundaries were controlled by the regional extension. This is consistent with the high, basin-wide continuity of sand and shale packages within the Triassic based on well log

correlations (e.g. Goldsmith et al., 2003; Goldsmith et al., 1995; McKie, 2014; McKie and Audretsch, 2005; McKie and Williams, 2009).

Jurassic Structures

Observations

The Top Jurassic horizon is identified by strong, erosional truncation and by onlapping stratal geometries on seismic data (Figs. 3.8 & 3.9). This corresponds to an abrupt, upward lithology change from black shale or marine sandstones (Upper Jurassic) into marly shales or chalk (Cretaceous) in well data and defines the BCU (Fig. 3.7). In this study, the Jurassic was not sub-divided further because of poor intra-Jurassic definition based on seismic reflections, which is due to low seismic frequency and the widespread of multiples. The exception to this is where the Jurassic interval is 100's m thick (e.g. in Western, Eastern and 30/6 grabens; Fig. 3.16).

The main Jurassic basin is compartmentalised into three generally NNW-SSE striking sub-basins (Fig. 3.15a, b): (1) the Western Graben; (2) the 30/6 Graben; and (3) the Eastern Graben (see also (Erratt et al., 1999). These sub-basins are separated from each other by highs that have existed since the Late Permian (e.g. Forties Montrose High) and areas that become high during the Jurassic (e.g. Marnock Terrace and J-ridge).

The largest of the sub-basins, Western Graben comprises two distinct segments (Fig. 3.15a): (1) a NNW-SSE striking, triangular-shaped low with Jurassic sediment diverging towards the WBF to a maximum thickness of >2.5 km in the west; and (2) a NW-SE striking low in the east adjacent to the Forties-Montrose High that contains a thinner (<1.5 km), eastwards diverging Jurassic succession.

The Western Graben is about 50 km long and 25 km wide and its length appearing to be defined by

two NE-SW striking deep deformation zones in the north and south which are thought to be the extension of the Southern Upland Fault and the Iapetus Suture, respectively (Fig. 3.15b).

The 30/6 Graben lies to the south of the inferred Iapetus Suture and is bounded by the SBF to the west and the J-Ridge to the east (Fig. 3.15b). This graben is c. 30 km long and 10 km wide, and contains a thick (c. 2.5 km) Jurassic succession in three smaller depocentres.

Unlike the Western and 30/6 grabens that are bounded by the inferred Iapetus Suture, the Eastern Graben dissects this inferred suture (Fig. 3.15b). This c. 30-km-long, 15-km-wide graben is bounded by the EBF to the east and flanked by the Marnock and Erskine terraces to the west. The northern boundary of this graben is well defined by the sharp northwards narrowing of the graben but the southern boundary is poorly defined by a series of NE-SW striking ridges that provide a transition into the Breiðflabb Basin. The centre of the graben contains c. 1.8 km of Jurassic sediments.

In addition to the main three sub-basins, the Breiðflabb Basin to the SE of the study area contains smaller, more subtle depocentres round 5-10 km long and 2-5 km wide (Fig. 3.15a, b). The depocentres are being bounded by a combination of mainly small NW-SE striking faults and salt wall highs.

The Jurassic structures above the Kittiwake-Gannet Platform, the Western Platform, Jæren High and Cod Terrace are not clearly mapped due to the coarse seismic grid. However, where the structures are large (>1 km), they resemble small, randomly oriented depocentres (Fig. 3.15a; see also Clark et al., 1999). These depocentres are mostly filled with Late Jurassic sediments directly overlying either the Top Triassic or the Top Late Permian Salt (Fig. 3.7; see also Stewart et al., 1999). Similar

depocentres are also found on the Marnock Terrace, Puffin High and J-Ridge. These depocentres contain relatively thin (<200 m) Jurassic successions and are often associated with small underlying salt diapirs (Figs. 3.8a, b; see also Clark et al., 1999).

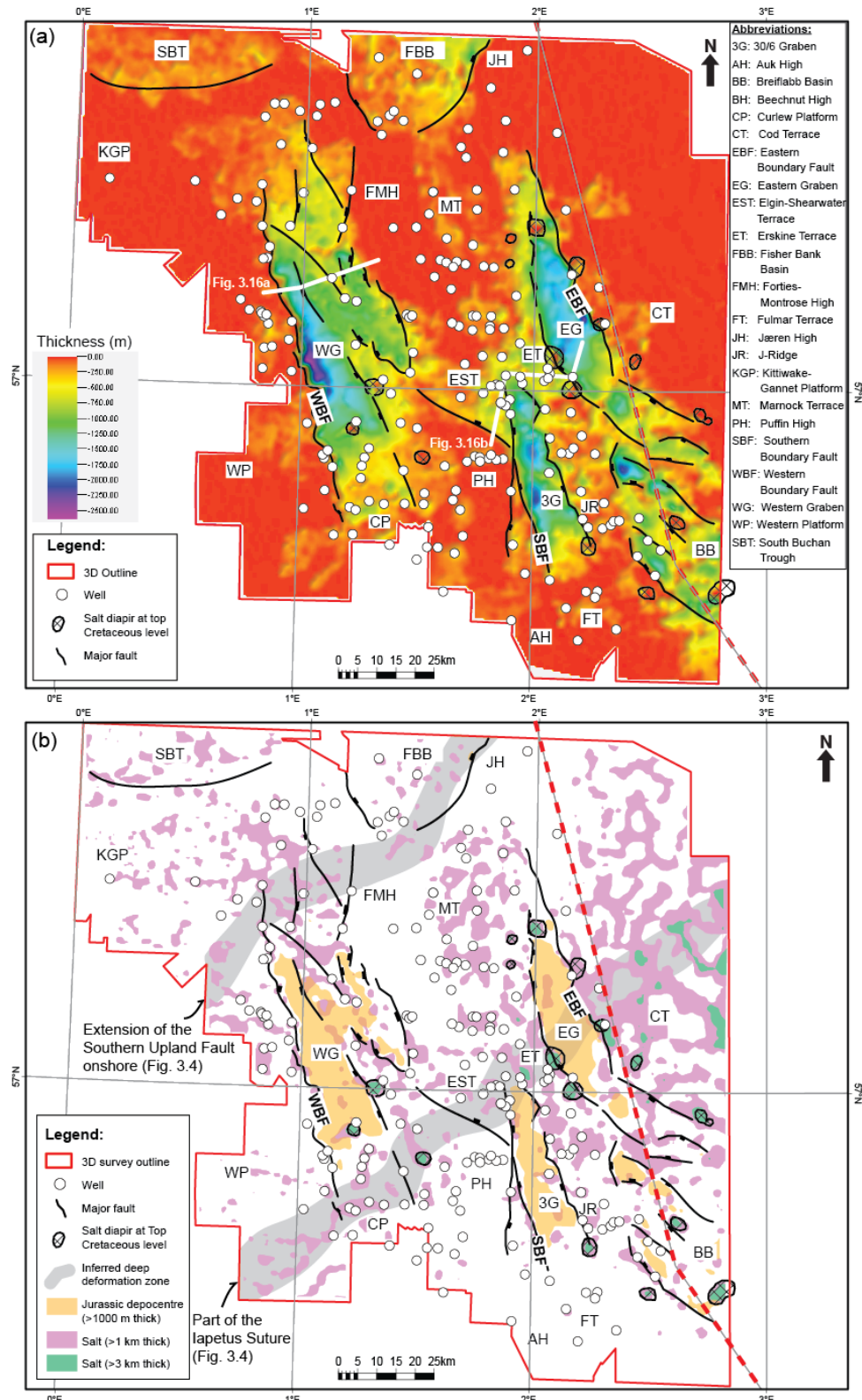


Fig. 3.15: Previous page - (a) Jurassic isochore map. (b) Composite map showing the spatial relationship between the Jurassic depocentres with salt walls. The Western and 30/6 grabens seem to abruptly change thickness at the inferred deep deformation zone while the Breiflabbb Basin seems to be more structurally independent of deep structures.

To the north, part of the study area merges with the Fisher Bank Basin, which contains up to 1 km of Jurassic strata (Fig. 3.15a; see also Clark et al., 1993). Here the basin thickens eastwards towards the Jæren High and its passes into several smaller (c. 1-3 km long by c. 1 km wide) depressions in the west containing <500 m thick sediments.

In the grabens, where the Jurassic is much thicker (e.g. 100s m; Figs. 3.16a, b) it can be clearly subdivided into three intervals: (1) Middle Jurassic, (2) early Late Jurassic (Oxfordian), and (3) the Latest Jurassic (Kimmeridgian-Tithonian). The Middle Jurassic interval comprises of parallel, continuous and high amplitude reflections, which overlie the Top Triassic surface (Fig. 3.16b). The Middle Jurassic is thinner (10s of m) in the Western Graben to the west (Fig. 3.16a) and thicker (100s of m) in the Elgin-Shearwater Terrace and the Eastern Graben to the east (Fig. 3.16b). The Oxfordian section has a more or less tabular thickness, albeit with a more chaotic character in the Western Graben (Fig. 3.16a), and a parallel, continuous character in the Eastern Graben (Fig. 3.16b). In contrast, the Oxfordian has a more SW-divergent seismic character in the Elgin-Shearwater Terrace (Fig. 3.16b). The uppermost Late Kimmeridgian-Tithonian displays a highly divergent pattern towards the major NNW-striking faults (Figs. 3.16a, b). This interval also forms the thickest section within the three major Jurassic grabens (Fig. 3.15a).

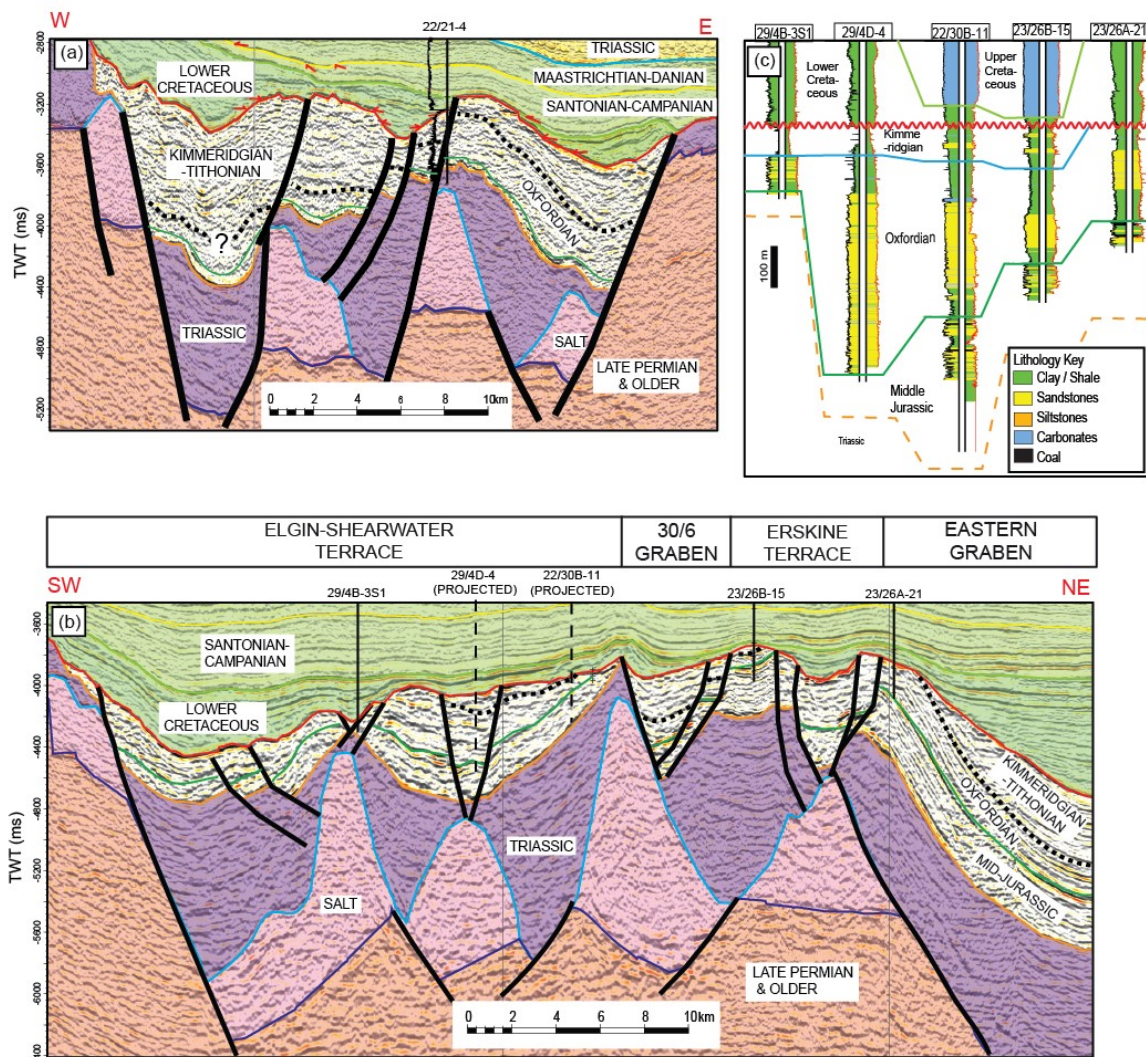


Fig. 3.16: Seismic sections showing the internal Jurassic stratigraphy, where the interval is thick and seismic data are good. (a) The Western Graben has thin (10's m) Middle Jurassic section and thick Upper Jurassic; (b) The Elgin-Shearwater Terrace has thicker (100s m) Middle Jurassic but the Upper Jurassic is thinner, perhaps due to erosion or non-deposition; (c) Well correlation in the Elgin-Shearwater area also demonstrated that the Oxfordian is thicker in than the Kimmeridgian section. See Fig. 3.15a for locations.

LEGEND					
—	Top Lower Palaeocene	—	Top Lower Cretaceous	—	Top Synrift (BCU)
—	Top Triassic	—	Top Upper Permian	—	Top Mid Upper Permian

Interpretation

Thickening of Jurassic strata towards large, salt-breaching, basement-involved fault systems indicate they were deposited during a time of rifting (Figs. 3.8b, 3.9b, c). These intervals are mainly bounded by NNW-SSE-striking faults, which imply extension perpendicular to these faults (ENE-WSW).

However, the NW-SE-striking structural trend is more common than the NNW-SSE trend in the Elgin-Shearwater Terrace and Breiflabbb Basin. This trend seems to mimic both the underlying structure at Triassic level (Fig. 3.13a) and the overlying Cretaceous structure (Fig. 3.17a). The same structural trend continues southeastwards into the Feda Graben (Fig. 3.1; see also Andsbjerg et al., 2001; Erratt et al., 1999; Tanveer and Korstgård, 2009), which is the dominant trend in this part of the basin. Following on that the extension vector is perpendicular to the fault strike argument, this could imply that different extension vector (NE-SW) has affected this area. However, lack of evidence at this stage means that the extension could also occur oblique to the strike of the faults which implies that no change in the extension vector is necessary.

The influence of salt is more subtle on the Jurassic interval than on the underlying Triassic interval since most salt stocks have been welded by the end of Triassic (Figs. 3.12b & 3.13b). This is also demonstrated by the lack of minibasins throughout the Jurassic.

Another aspect of halokinesis is the presence of dissolution features. The elongated depressions in the Kittiwake-Gannet Platform have been interpreted as salt-dissolution-induced valleys (Clark et al., 1999; Stewart et al., 1999; Wakefield et al., 1993) with modern analogues are observed in the Permian Basin of the southwestern United States (e.g. Gutiérrez, 2004; Johnson, 1981). However, alternative interpretation may also be plausible. Based on the circular and elongated Jurassic fault patterns that can be matched to the edge of salt diapirs, these depressions may be remnants of collapsed blocks on top of salt diapirs, induced by late salt withdrawal during the Late Jurassic rifting (Clark et al., 1999; Mannie et al., 2014). Diapir and overburden collapse would have led to preferential preservation of Jurassic as well as Upper Triassic sediments above salt diapirs, with this

succession being completely or strongly eroded above flanking Triassic minibasins (see also Chapter 4).

Internal seismic characteristics, especially divergence towards major extensional faults that occur in the Tithonian section, suggest that the main rifting event occurred between the Oxfordian and Tithonian periods (Figs. 3.16a, b). However, the variability in thickness between Oxfordian and Kimmeridgian-Tithonian sections in NW- and NNW-striking depressions also suggests that there could be differences in the orientation of extension in the two time period (see Discussion section).

Lower Cretaceous Structures

Observations

The Top Lower Cretaceous horizon is represented by a strong continuous seismic reflection, which is underlain by an interval of high-amplitude continuous reflectors and overlain by a largely transparent seismic facies (Figs. 3.8 & 3.9). Well data indicates an upwards change from Lower Cretaceous layered marl and carbonate beds to Upper Cretaceous relatively homogeneous chalk (Fig. 3.7).

Lower Cretaceous strata define a broadly NW-SE orientated depocentre (Fig. 3.17a); this is oblique to the main basin-bounding faults at the Jurassic level (WBF, EBF & SBF) that have a NNW-SSE strike (Fig. 3.15a) although part of the main faults are still mappable, especially in the Western and Eastern grabens. Bounded by the main faults, both grabens contain Lower Cretaceous sediments that thicken (up to c. 1.8 km) towards the main faults, mimicking that which is observed in the underlying Jurassic succession (Figs. 3.8b, c, 3.15a, & 3.17a). However, the Lower Cretaceous sediment is considerably thinner (<300 m) in the Western Graben compared to the Jurassic succession that is 1.5

km thick. In contrast, the Eastern Graben depocentre is narrower (c. 10 km wide) but contains a thicker (1.8 km) Lower Cretaceous succession compared to its Jurassic equivalent (1.3 km).

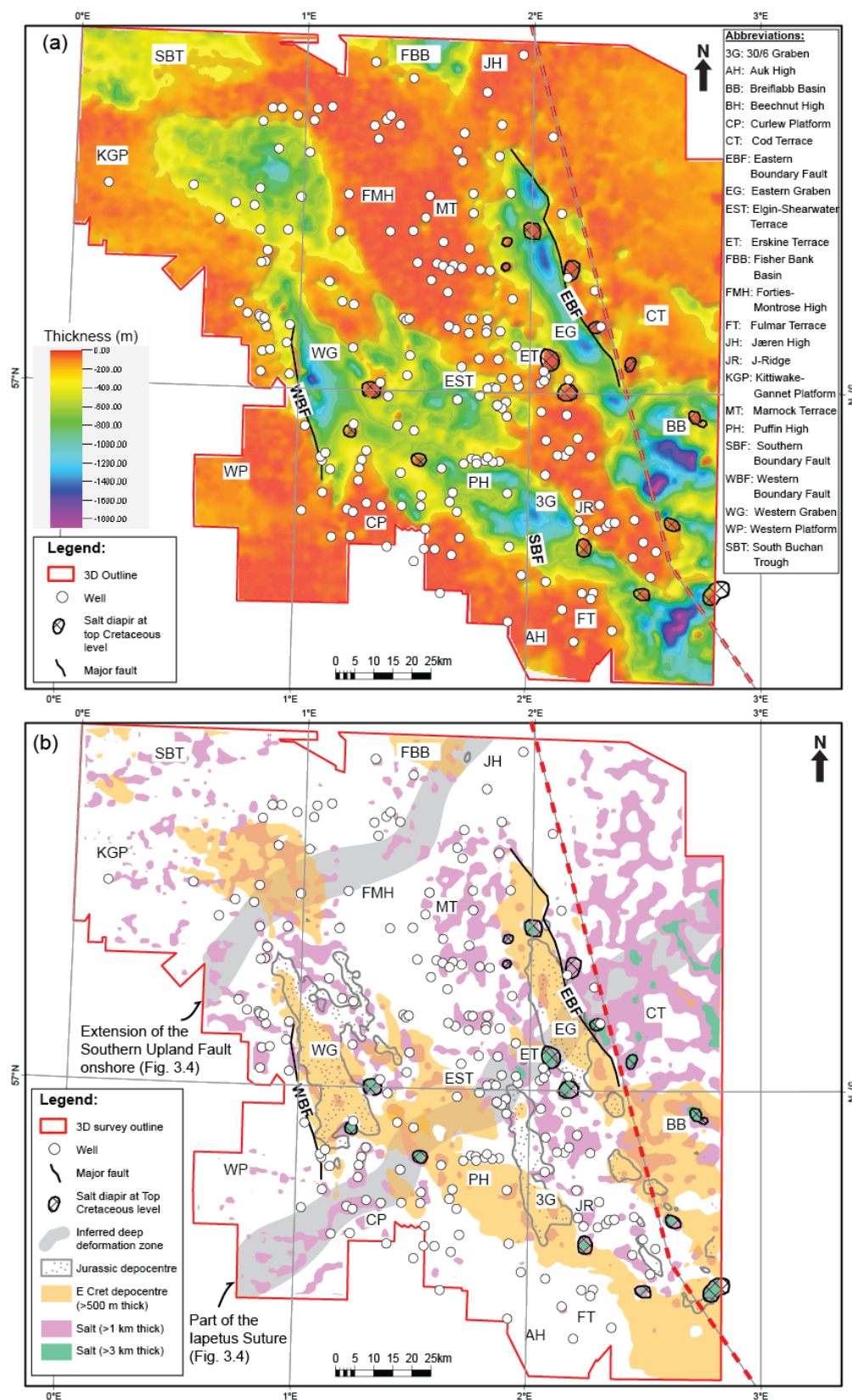


Fig. 3.17: Previous page - (a) Lower Cretaceous isochore map. (b) Composite map showing the relationship between the Lower Cretaceous depocentres with the salt walls. The interval seems to be thickest where salt is thin within the older Mesozoic basin deep.

The contrast is also prominent in the adjacent Elgin-Shearwater Terrace that have somewhat become connected with the Western Graben at the Lower Cretaceous level. Here, the NW-SE-oriented western (Elgin) part is thicker than its Jurassic equivalent (c. 1 km vs. 0.5 km) whereas the eastern (Shearwater) part is thinner (c. 0.2 km vs. 1 km; Figs. 3.15a & 3.17a).

The thickness pattern in the southern part of the study area changes dramatically compared to the Jurassic period (Figs. 3.15a & 3.17a). The 30/6 Graben has been overrun by a larger NW-SE-striking depocentre that also consumed the Puffin High and become connected to the Western Graben (Fig. 3.17b). The depressed Puffin High, which exist as a structural high during Jurassic times (Fig. 3.15a), is also being compartmentalised by E- and NW-striking ridges.

Additionally, a triangular-shaped depression (c. 50 km by 40 km) has formed to the north of Western Graben (Fig. 3.17a). This depression, which has not been seen in any of the older stratigraphic intervals, contains up to 1.2 km of Lower Cretaceous sediments and also covers part of the Kittiwake-Gannet Platform.

In the southeast of the study area, the Breiflabb Basin becomes more prominent where it contains up to 2 km thick Lower Cretaceous sediment. The depocentres, c. 10-15 km long by up to 5 km wide, are also wider than their Jurassic equivalents and slightly elongated in a NW-SE or NE-SW trend (Figs. 3.15a & 3.17a). In addition, several small (2-5 km in diameter), circular peaks are present (bright red

in the map; Fig 3.17a). These are usually located near the edge of the depocentres and coincide with the presence of tall salt diapirs (in green; Fig. 3.17b).

The internal seismic character of the Lower Cretaceous typically shows parallel, semi-continuous to continuous reflections, which onlap onto the BCU (Figs. 3.8a, b, & 3.9a, b). In addition, inverted thicker sections can be found in several areas, usually close to the main faults, most notably in the northern part of Western Graben (Fig. 3.8b) and Curlew Platform (Fig. 3.8c).

Interpretation

The general lack of abrupt, fault-controlled changes in thickness and the onlap of Lower Cretaceous strata onto the BCU indicate that the Lower Cretaceous sediments were deposited in a post-rift setting. However, abrupt thickness changes that occur across the WBF and EBF may indicate that these structures were active and that extension continued into the Early Cretaceous. An alternative interpretation is that crustal extension had ceased and that faulting and upward fault tip propagation was driven by Early Cretaceous differential compaction of the underlying, mud-dominated, Upper Jurassic succession. The absence of fault bounded, stratal divergence within the Lower Cretaceous section above the BCU supports a post-rift setting.

The concentration of maximum subsidence along the older Jurassic depocentres, especially in the Western and Eastern grabens, indicates that the locus of Early Cretaceous, early post-rift subsidence was focused above the Late Jurassic syn-rift depocentres. This is intuitive because the Upper Jurassic depocentres formed in the areas of maximum crustal stretching and hence were the site of maximum thermal cooling and subsidence. In contrast, the depocentre that combined 30/6 Graben and the Puffin High (Late Jurassic structural high), do not simply reflect post-rift thermal cooling and

subsidence. This could reflect localised late rifting without any extension orientation change in the earliest Cretaceous (Coward et al., 2003). Alternatively, it is also possible that the thermal subsidence has differentially exploited older Triassic sub-basins in this area (Fig. 3.13a).

The Lower Cretaceous strata have also undergone significant inversion along the southwestern flank of the basin especially along the Mid-North Sea High. This is shown by the arching depocentres and onlapping of overlying Upper Cretaceous strata in the northwestern depression and Curlew Platform (Figs. 3.8b, c). This interpretation is supported by other inversion features seen elsewhere along the Mid-North Sea High (e.g. Gregersen and Rasmussen, 2000).

Upper Cretaceous and Younger Structures

Observations

The Top Lower Palaeocene (Top Chalk Group; Fig. 3.5) surface corresponds to a bright, continuous, hard reflection event which is generated by an abrupt change from Cenomanian-Danian chalks to Tertiary deepwater clastics (Fig. 3.7, 3.8 & 3.9). The Chalk Group is divided into two units based on the seismic character (Figs. 3.8 & 3.9): (1) a high-frequency, semi-continuous seismic facies that onlaps onto the Top Lower Cretaceous reflection, which can be correlated to argillaceous chalk in well logs; and (2) an overlying bright, low frequency, continuous seismic facies that correlates to cleaner, thicker chalk layers (Fig. 3.7). The high amplitude, continuous Top Campanian (Top Mackerel Formation) reflection divides this two seismic facies (Fig. 3.8 & 3.9) and, in some cases, also truncates the lower Chalk Group (e.g. on Puffin High; Fig. 3.8c).

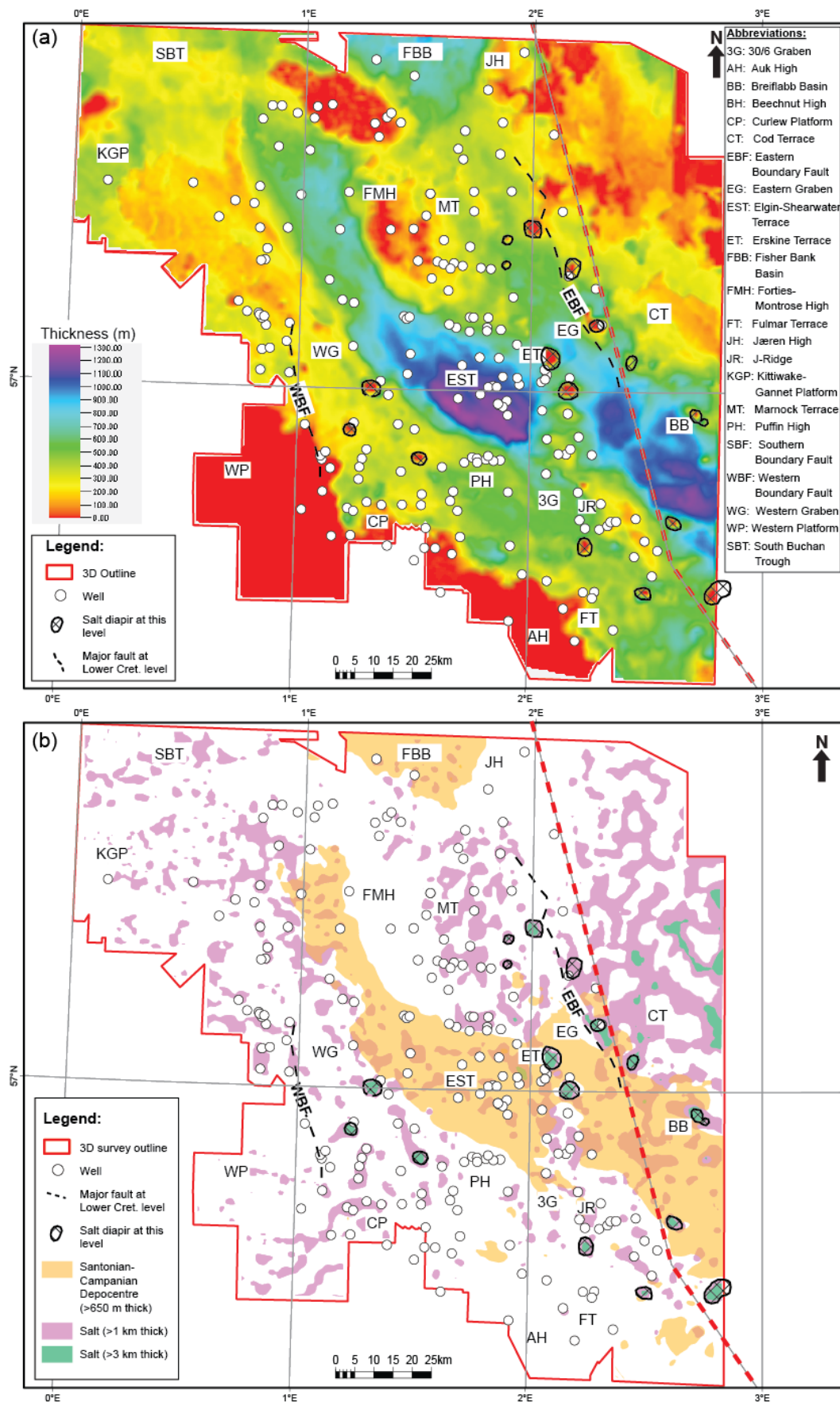


Fig. 3.18: Previous page - (a) Santonian-Campanian Chalk Group isochore map. (b) Composite map showing the Santonian-Campanian Chalk interval is largely independent in relation with the salt walls and the boundary faults.

The lower unit (Santonian-Campanian) defines two prominent thick (c. 600-1300 m), NW-oriented depressions, located north of the Auk High (Fig. 3.18a). These depressions are separated by a N-S striking ridge that extends from the J-Ridge in the south, through to Erskine Terrace and to the Forties-Montrose High in the north. The western depression (c. 90 km long by 25 km wide) extends from the Kittiwake-Gannet Platform to the Elgin-Shearwater Terrace where it thickens. The eastern depression (c. 75 km long by 25 km wide) extends from the centre of the Eastern Graben to the Breiðflabb Basin. There is a conspicuous lack of faults in this interval (Figs. 3.8, 3.9 & 3.18a), with the abrupt thickness changes corresponding to depositional onlap onto fault blocks (e.g. between Puffin High and Elgin-Shearwater Terrace; Fig. 3.8b & 3.18a). In the CG, this unit caps all structural elements, apart from Auk high and the Western Platform (Fig. 3.18a).

In contrast to the lower unit, the upper (Maastrichtian-Danian) unit's thickness patterns define a single, large NW-SE-striking basin which deepens southeastwards, up to 1.1 km thick (Fig. 3.19a). Similar to the lower unit, there is also lack of faulting in this layer and tend to pinch-out into thin (<100 m) interval on the main highs (e.g. Western platform & Jæren High; Figs. 3.8a, b, c). This relatively parallel layered interval is punched in places by tall salt diapirs (Fig. 3.19a).

Both Chalk units thin above the southern depression of the Lower Cretaceous, especially in the Curlew Platform and Puffin High areas (Figs. 3.17b, 3.18b & 3.19b). The thinning of the Upper Cretaceous to Lower Palaeocene strata also corresponds to the inverted structure of the Lower Cretaceous.

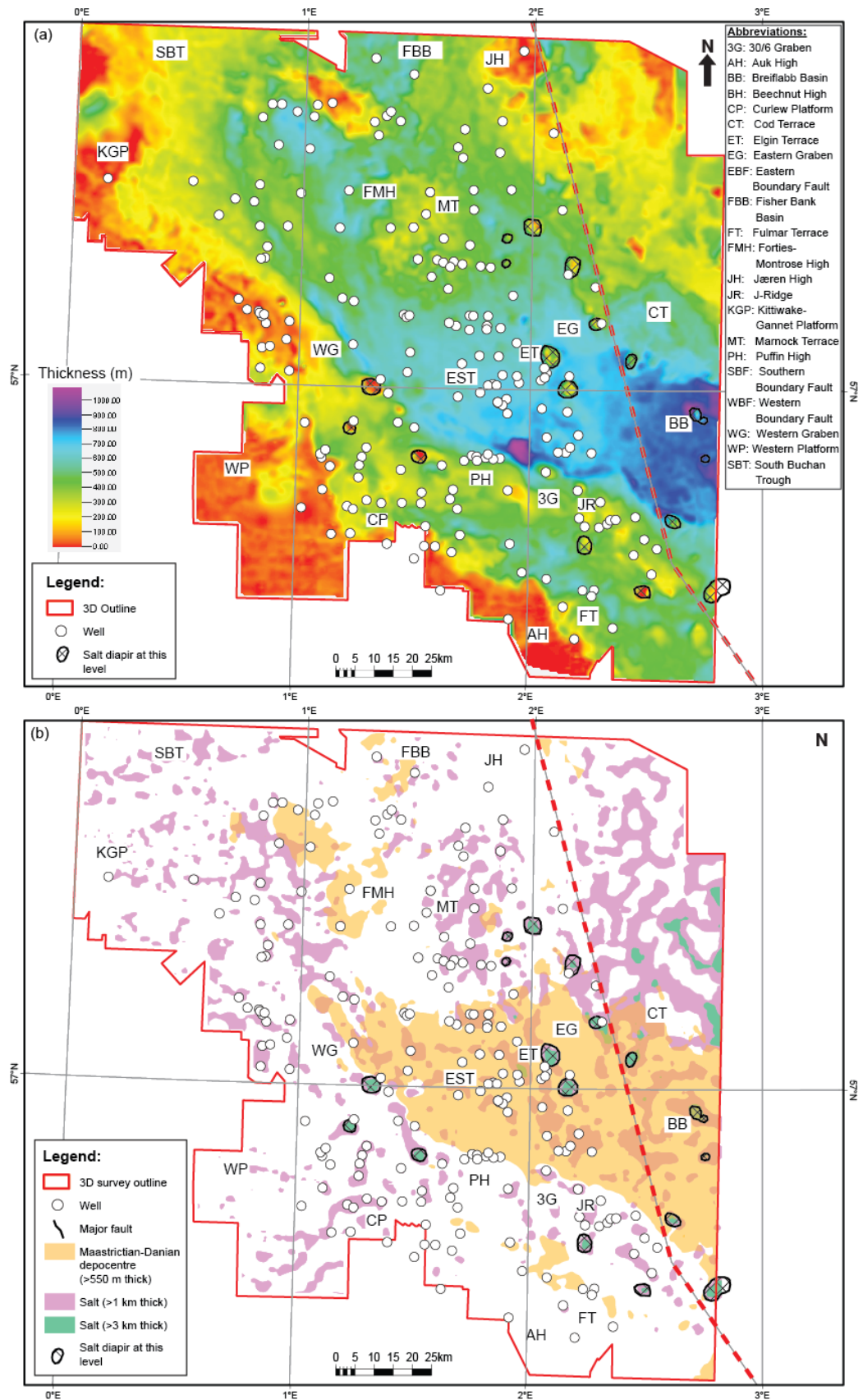


Fig. 3.19: (a) Maastrichtian-Danian Chalk Group isochore map. (b) Composite map showing that the Upper Chalk Group is largely not influenced by salt walls.

Interpretation

The Chalk Group represents almost uninterrupted deposition of chalk during the Late Cretaceous to Early Paleocene post-rift sag period, when faults were largely inactive (Surlyk et al., 2003). The Lower Chalk Group initially onlapped onto relict rift-related relief during the Santonian and Campanian, before draping all the structural elements except for the Western Platform and Auk High (Figs. 3.8c & 3.18a).

At the end of the Campanian, inversion caused uplift and folding of Santonian-Campanian strata as shown by the truncation of seismic reflections below Top Campanian in the Western Graben (Fig. 3.8b). This inversion event is also supported by observations in other areas (e.g. Erratt et al., 1999). Inversion-related uplift was greatest where the Jurassic to Cretaceous succession in Western Graben were thickest, in close proximity to the Western Platform and Auk High (Fig. 3.18a), the latter being a remnant of the Mid-North Sea High (Fig. 3.1).

The Upper Chalk Group interval represents a more tectonically quiet period where the basin slowly subsided and a general increase in sea-level during the Maastrichtian-Danian age. This eventually overwhelmed the Forties Montrose High (Fig. 3.19a), paving the way for the formation of a deep Tertiary basin covering all of the CG.

Structural Reconstruction

The structural evolution of the CG during the pre-, syn- and post-Mesozoic rifting is complex. In order to understand it, a series of palinspastic sections and block diagrams have been constructed (Fig. 3.20 & 3.21). The palinspastic sections are based on one seismic section (Fig. 3.8b), and

reconstructed for five stratigraphic levels: (a) Top Permian; (b) Top Triassic; (c) Top Jurassic; (d) Top Lower Cretaceous; and (e) Top Early Palaeocene. Each section contains vertically reconstructed surfaces interpreted from the seismic data. However, the only exception is for the End Permian, where the base of the flattened Permian salt has been derived from an estimation of the original thickness of salt shortly after deposition based on the current thickness of salt and Triassic sediments (Figs. 3.12a, 3.13a & 3.20a). The block diagrams represent the simplified areal and profile structural configuration near the end of Triassic and Late Jurassic riftings as well as the post-rift period of Early Cretaceous (Fig. 3.21).

Based on offshore seismic and well data, rifting started in the Early Permian (Jackson and Lewis, 2013; Neumann et al., 2004; Stemmerik et al., 2000). Tectonic subsidence also have been proven in the adjacent Southern Permian basin (van Wees et al., 2000). Prominent NE-SW and NW-SE faults, which are remnant of the Caledonian and Variscan orogenies, controlled the location of the structural highs during this period (Glennie et al., 2003). At the end of Permian, anhydrite and carbonates accumulated on remnants of Early Permian rift horsts (Southern Western Platform, Forties Montrose High and Jæren High; Fig. 3.20a) whereas the adjacent basins contained extensive salt, which thickened in the deeper part of the basin, most notably in the Cod Terrace (Fig. 3.12a). Evidence from seismic data in the Netherland's North Sea has shown that these evaporates are thickening towards the major extensional faults, which suggested that it have been deposited during a syn-rift period (e.g. Geluk, 1999).

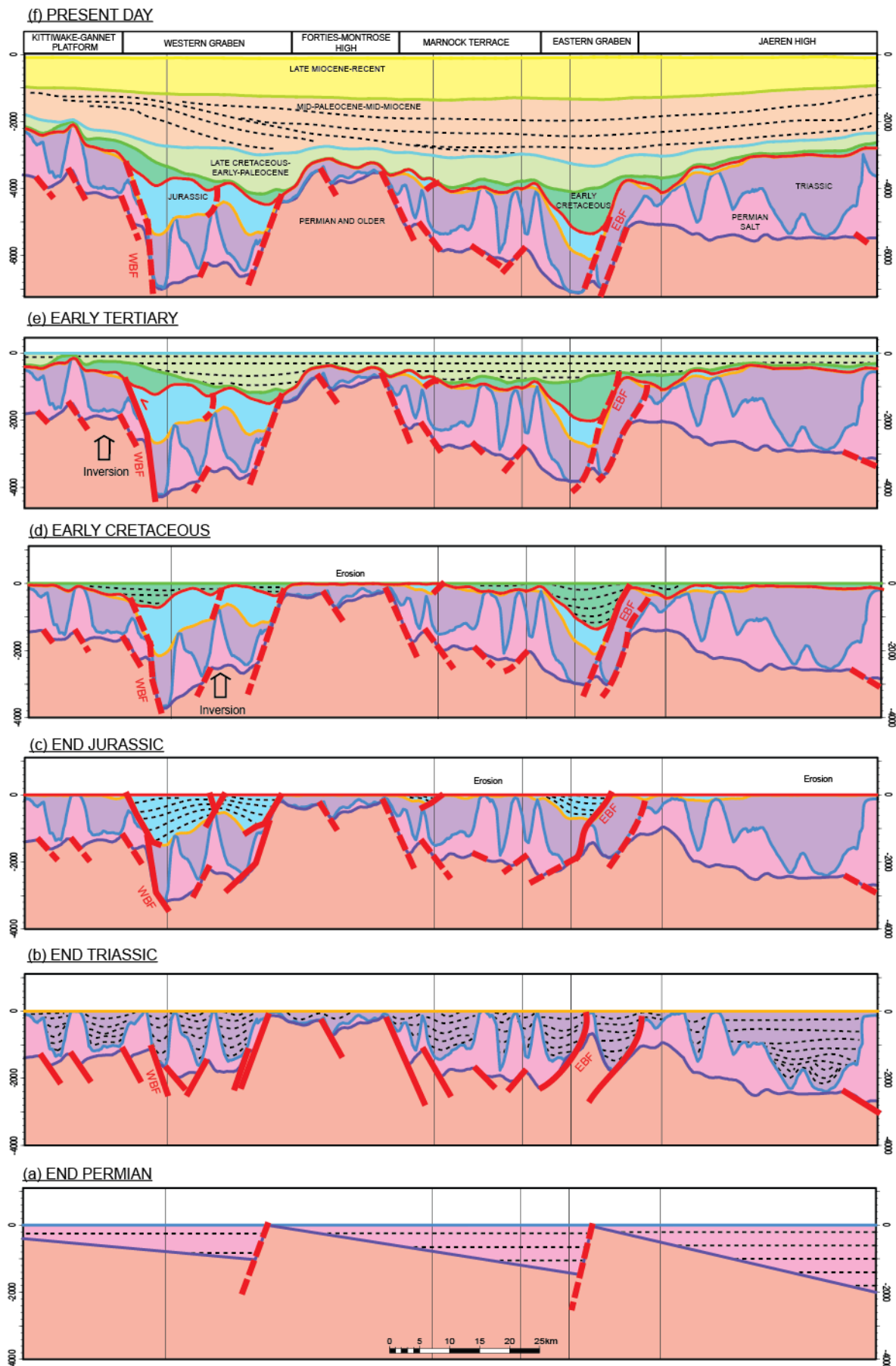


Fig. 3.20: Structural reconstruction of the Central Graben. Thick red lines represent major active faults in the time period while dashed red lines are faults that no longer active. The dashed black lines represent the internal stratal pattern observed from seismic line. For line of section and location, see Figs. 3.6 & 3.8b.

Extension in the Triassic exploited NW-striking faults that had existed since the Permian (Doré et al., 1999; Müller et al., 2005; Steel and Ryseth, 1990). Halokinesis was also most active during this period, forming salt diapirs, while the Triassic sediments infill the adjacent minibasins (Fig. 3.21a). Sub-salt faults formed during this time did not generally breach the salt, except where salt was initially thin, such as near the flanks of Forties-Montrose High and the western flank of the Eastern Graben (Fig. 3.20b). By the end of the Triassic, salt welds begun to develop, restricting further growth of flanking diapirs (Figs. 3.20b & 3.21a). Smith et al. (1993) also agreed with the formation of salt weld during the Triassic but speculated that the salt welds in deeper part of the basin (e.g. in Western and Eastern grabens) only commenced in Late Cretaceous. However, as shown by the seismic data and palinspastic reconstructions (Figs. 3.8a & 3.20b-f), salt welds are likely to have formed even in the current basin lows during the Triassic. Rifting terminated at the end of the Triassic and was followed by post-rift deposition of Early Jurassic, which is very poorly preserved, before the commencement of regional doming of the Mid-North Sea High, which resulted in partial erosion of parts of the Upper Triassic (Goldsmith et al., 2003; Husmo et al., 2003; Ziegler, 1992).

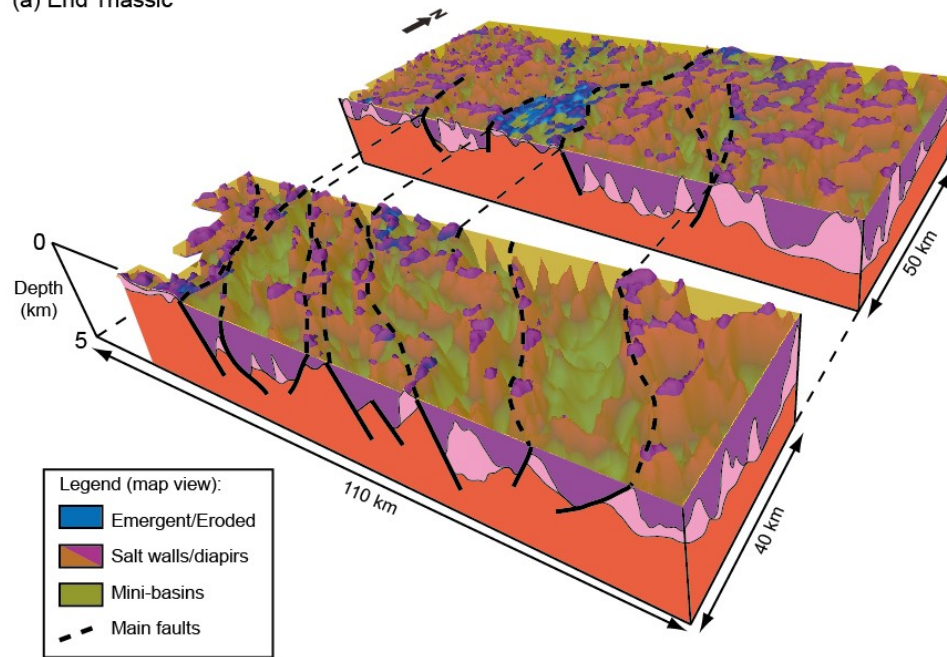
Jurassic rifting rejuvenated the NW-SE Triassic fault system and formed new NNW-SSE faults (Fig. 3.21b). The former is most common in the middle and SE corner of the CG. New faults include the WBF, EBF and SBF, where Jurassic syn-rift sediments were deposited in their hangingwalls; in association with Western, Eastern and 30/6 grabens, respectively (Figs. 3.20c & 3.21b). The formation of these narrow grabens and confinement of active faultings within them show that the Late Jurassic rifting was more focussed and limited in extent as compared to the Triassic rifting (Figs.

3.20b, c & 3.21a, b). Halokinesis was also reduced compared to during the Triassic, although some tall salt diapirs extended to the surface (Figs. 3.20c & 3.21b). The faults were usually smaller, salt-detached normal faults formed above salt diapirs, in response to the extension above the salt (Fig. 3.20c). Rifting had ceased by the end of Jurassic when the basin underwent thermal subsidence (Erratt, 1993). Widespread transgression during the Early Cretaceous eroded the Late Jurassic highs and formed the laterally extensive BCU (Kyrkjebø et al., 2004; Rawson and Riley, 1982).

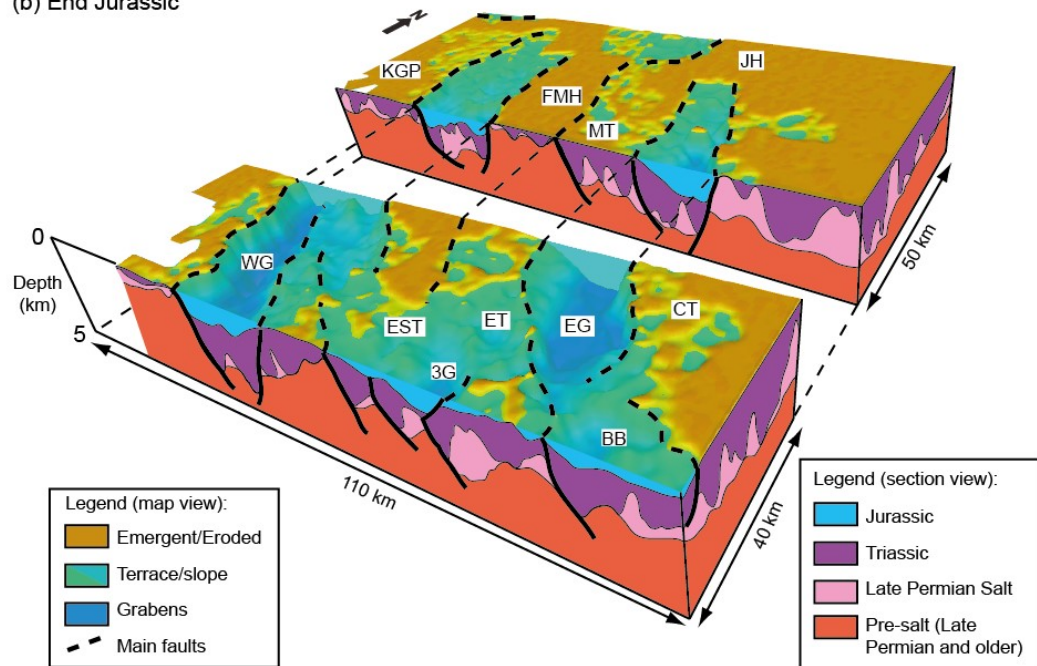
Movement on the EBF continued during the Early Cretaceous resulting with thicker Lower Cretaceous succession in the Eastern Graben compared with the Western Graben (Figs. 3.20d & 3.21c), suggesting enhanced fault slip and basin subsidence in the former. Instead, localised inversion probably occurred in the Western Graben, based on arching of the Top Jurassic (BCU) surface (Fig. 3.20d). Early Cretaceous inversion also has been proposed by Thomas & Coward (1995) in some localised sub-basins in the East Shetland Basin and attributed it to strike-slip reactivation of normal fault, so it is possible that the same mechanism has affected the CG.

Localised inversion was becoming more intense at the beginning of the Late Cretaceous and may have reversed movement on the WBF, resulting in the uplift of the basin centre in the Western Graben (Fig. 3.20e). This inversion was followed by the deposition of the Upper Cretaceous-Lower Tertiary sediments, which initially onlapped, and then completely covered reliefs associated with Late Cretaceous inversion (Fig. 3.20e). Finally, Late Tertiary sediments, derived from the western basin margin, prograded over the entire basin, which has continued to passively subside until the present-day (Fig. 3.20f).

(a) End Triassic



(b) End Jurassic



(c) End of Early Cretaceous

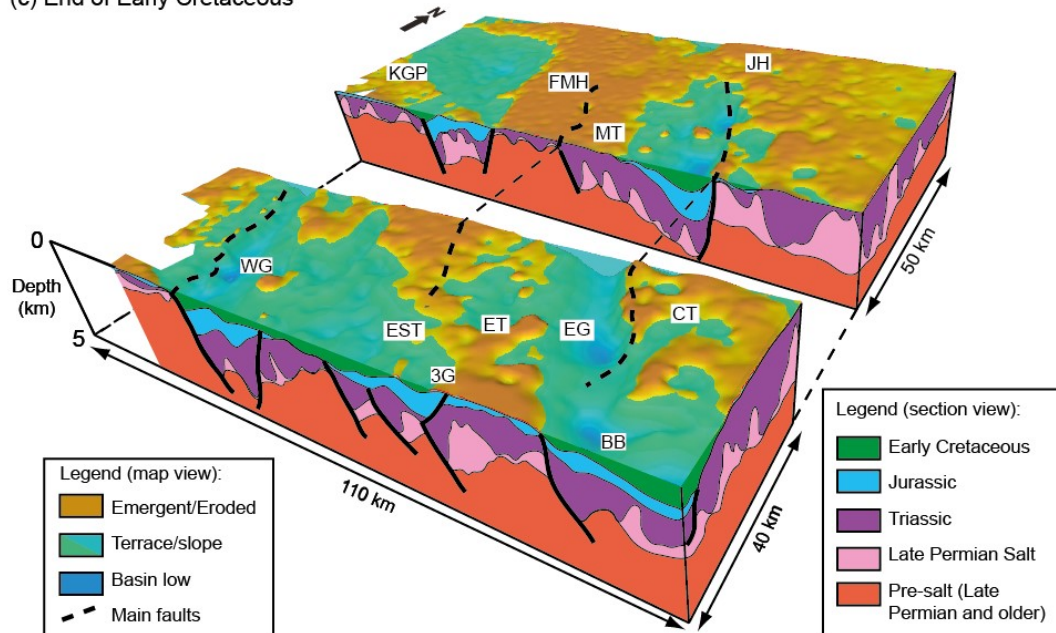


Fig. 3.21: 3D block diagrams showing the simplified structural configuration of the CNS Basin at three critical time periods: (a) At the end of Triassic, when halokinesis had been most active, the CG comprised large salt walls and mini-basins. The Triassic cover has been made transparent to show the gross internal structure of the NW-SE-trending salt walls. The associated supra-salt Triassic faults are also interpreted to have the same orientation. Some salt diapirs are to have been close to the surface, forming ridges and local topographic variations. Surface salt bodies would have been consumed by dissolution and erosion, especially during later periods of uplift (e.g during Early Jurassic thermal doming event). The Forties-Montrose High and the northwestern part of Jæren High were almost emergent features and any original salt layers appear to have been all but dissolved. (b) At the end of Jurassic the CG was dominated by the main graben system. The Late Jurassic rift phase was more focused, forming a series of narrow NNW-trending graben systems along the axial part of the basin. The main highs are also more pronounced and interpreted to have been almost emergent, except in a few embayments where dissolution of the underlying salt walls was taking place. Generally, halokinesis was minimal except near graben edges where salt diapirs continued to grow. (c) At the end of Early Cretaceous a new basin system had developed over the previous Late Jurassic grabens. This was a consequence of thermal subsidence, which exploited both NW- and NNW-trending zones of weaknesses. Salt growth continued along the edge of Late Jurassic graben edges in the form of small, isolated salt diapirs, which appear here as circular 'islands'. Note that only the major active faults are highlighted for each time-slice. See Fig. 3.10a for abbreviations.

Discussions

Deep structural controls on the location of salt diapirs and subdivision of halokinetic provinces

This study demonstrates that the sub-salt underlying structures influenced the location of large salt bodies. This is shown by the frequent confinement of salt bodies between NNE-SSW-striking faults that have also breached the salt, and the alignment of salt diapirs parallel to the other deeper faults (Fig. 3.11b). In addition, salt diapirs are preferentially located above the crest or edge of sub-salt fault blocks (Figs. 3.8 & 3.9), which is similar to that observed in the Norwegian-Danish basin (Koyi and Petersen, 1993).

The close relationship between the orientation of the salt walls and the underlying major faults (Fig. 3.12b) suggests a close genetic relationship (see also Dooley et al., 2005). Reactivation of sub-salt faults by basement-involved extension, probably during the Triassic, may have also caused overburden extension, reactive diapirism and salt wall growth. This relationship is particularly clear in the platform high areas (e.g. Kittiwake-Gannet and Western platforms) where E- and NE-striking faults are similar to the overall trend of the overlying salt wall orientation (Fig. 3.12b).

However, salt walls also form polygonal patterns, particularly where the underlying structure is more subtle, such as to the south of Jæren High (Fig. 3.12a). Where sub-salt faults have smaller throw than the initial thickness of salt, the deformation of the latter is less dependent of the former (Stewart et al., 1996). In addition, evidence from parts of the Norwegian-Danish Basin also shows that the location of salt bodies decoupled from the sub-salt structural trend (Hospers et al., 1988; Lewis et al., 2013). In these areas (e.g. along the margin of Norwegian-Danish Basin), salt diapirs are also located on a platform where the sub-salt structures are relatively subtle and display minimal impact

on the location of the diapirs. Therefore, the polygonal pattern may indicate that the formation of salt walls in thick salt above a relatively stable basement could be due to thin-skin processes such as gravity-driven subsidence and gravity slides instead of basement-induced extension (e.g. Hudec and Jackson, 2009).

The CNS has been subdivided into several structural domains in relation to salt deformation styles and their subsequent effects on supra-salt sediment deformation. Stewart (2007) divided the salt deformation areas into the following three types: (1) basement-extension, mainly within the grabens (Western, 30/6 and Eastern grabens; Fig. 3.10a); (2) minibasins, including pod-interpod and rift-raft deformations (Figs. 3.2a, b), mainly on the structural highs (Kittiwake-Gannet and Western platforms, Forties-Montrose and Jæren highs); and (3) gravity sliding, in the West Central Shelf that encompasses the Kittiwake-Gannet Platform and Western Platform. Subsequently, Karlo et al. (2014) added another two styles: (1) collapse anticlines, such as in the Norwegian-Danish Basin and the Mid-North Sea High; and (2) salt welds, based on observation of thin, dyke-like features found exclusively in the Eastern Graben. These salt deformation types are consistent with the findings of this study, apart from the salt welds interpretation in the Eastern Grabens. In this salt weld case, it is interpreted as representing well imaged, often angled sides of salt diapirs. The other sides of the diapirs could be poorly imaged by seismic data due to their sub-vertical nature or become obscured due to the direction of seismic acquisition.

In this study, we have further shown that the general orientation of salt walls is also closely related to the structural styles within the underlying Rotliegend Group (Figs. 3.10b & 3.12b). The orientation of the salt structures is also related to Triassic and Late Jurassic extension, as well as compression during the Cretaceous and Tertiary.

Triassic rifting

There is a clear evidence of Triassic rifting in the Northern North Sea Basin (Færseth, 1996; Goldsmith et al., 2003; Steel and Ryseth, 1990) and the Norwegian-Danish Basin (Høiland et al., 1993; Zanella and Coward, 2003). However, evidence for Triassic rifting in the CG, such as the development of sediment intervals that thicken towards faults, was sparse (Erratt et al., 1999). This may be due to a combination of: (1) the effect of halokinesis; (2) the homogeneous nature of Triassic's seismic response; and (3) insufficient image quality below the BCU.

The Triassic rift in the Northern North Sea was linked to the opening of the Arctic between East Greenland and Fennoscandia, which was initiated in the Late Permian and continued into the Early Triassic (Doré et al., 1999; Müller et al., 2005; Steel and Ryseth, 1990). Simultaneously, the Neo-Tethys rift in the south propagated northwestwards into the Southern Permian Basin, exploiting older Caledonian and Variscan structural fabrics (Coward, 1995; Geluk, 2005). Later, two main pulses of extension occurred in the Late Early Triassic and Early Late Triassic, mainly as the result of the opening of Neo-Tethys Ocean to the south. This resulted in the formation of the Hardegsen and Early Cimmerian unconformities (Geluk, 2005; McKie, 2014). All of these events were potentially extended into the CNS but, until recently, the main proof of this was an eastwards increase in Triassic sediment thickness based on well data (Goldsmith et al., 2003).

The Triassic isochore map presented here, integrating both seismic and well data, shows abrupt thickness changes across the NW-striking faults as well as the main NNW-striking faults (Fig. 3.13a). However, the NW-striking faults are the more plausible original Triassic extensional faults. This is supported by: (1) the southwestwards-diverging Triassic stratal patterns (Fig. 3.14b), which is

perpendicular to this fault trend; and (2) the predominantly NW-striking fault and general salt wall pattern below the Jurassic grabens (e.g. Western Graben; Figs. 3.12a & 3.13a). The NNW-SSE trend probably reflects the overprinting of Jurassic faults onto the Triassic and the abrupt thickness change across the NNW-striking faults could have been caused by footwall erosion during the Jurassic rifting (e.g. in Western Graben; Fig. 3.13a). Hence, Triassic structures located below the Jurassic grabens are more representative since these are less likely to be eroded during Jurassic rifting. Furthermore, the NW-striking faults are also parallel to the Baltica-Avalonia Suture and to the Tornquist Zone in the Norwegian-Danish Basin (Fig. 3.4), which represent the older Palaeozoic structural fabric.

Although the NW-trend is oblique to the N-S fault trend of Stord, Asta Graben and Egersund basins in the Norwegian Sector, and to some extent the Tail End Graben in the Dutch sector (Fig. 3.1), the extension vector may be similar. Firstly, the Norwegian basins are located within the main Triassic rift axis while the CG is situated near the rift flank (e.g. Erratt et al., 2010; Færseth, 1996). Therefore, the influence of the main extensional complex could be somewhat dissipated. Secondly, the oblique nature of the CG in relation to the Norwegian Triassic basins could be caused by the complex nature of the underlying Palaeozoic structure (Fig. 3.4). Similarly, in the Danish Basin the NW-striking faults of the Tornquist Zone are interrupted by the N- to NE-striking Skagerrak Graben (Fig. 3.4; Mogensen, 1995; Ro et al., 1990).

Jurassic rifting

The NNW and NW graben trends identified in the Jurassic isochore map are consistent with the two-phase (Erratt et al., 1999) and three-phase structural models (Figs. 3.5b, c; Zanella and Coward, 2003). According to the two-phase model, rifting, driven by E-W extension, has been active in the CG since the Middle Jurassic (Callovian) (Fig. 3.5b). Later, the rifting direction changed to a NE-SW trend during the latest Jurassic (Kimmeridgian), which resulted in criss-cross network of N-S and NW-SE-

striking fault sets. This model is also consistent with predictions made by the sequential dip-slip model of Davies et al. (2001) in which the E-W extensional stress regime was superseded by a NE-SW extensional stress regime. The three-phase model, however, suggests that the first opening of the NNW-striking CG was caused by transtensional movement related to the WNW-ESE-directed opening of the Viking Graben during the Late Jurassic (Beach, 1987), before switching to NE-SW-directed rifting in the latest Jurassic. The NE-SW rifting episode continued into the Early Cretaceous with only a slight change in the direction of opening (Fig. 3.5c).

This study shows that sub-basins bounded by NW-striking faults have thicker Middle Jurassic and Early Late Jurassic (Oxfordian) intervals compared the latest Jurassic (Kimmeridgian to Tithonian) interval (e.g. Elgin-Shearwater Terrace; Fig. 3.16b). In contrast, the latter tend to be thicker in NNW-striking grabens (e.g. Western and Eastern grabens; Figs. 3.8b & 3.16a). These thickness differences are linked to the different generations of bounding faults, with the NW-striking faults preceding the NNW-striking faults. The orientation of the Triassic faults supports this interpretation (Fig. 3.13a). It is probable that the Middle Jurassic faults exploited the Triassic faults in the beginning of thermal relaxation, immediately after the Middle Jurassic thermal doming. Later, during the early part of the Late Jurassic rifting period, the NW-striking grabens and terraces became the focus of sedimentation. Fault orientation switches to the NNW-SSE during the peak rifting period in the latest Jurassic.

However, evidence from Feda Graben (SE of the study area) shows a consistent N-S orientation of the depocentre throughout both the earlier and later Jurassic (Tanveer and Korstgård, 2009). This suggests a similar fault trend during both the Triassic and Jurassic rift periods in the Feda graben, which lies directly southwards of the Egersund Basin which also has a N-S trend. The original extension vector might still be similar to that of the Triassic, opening obliquely to the regional

extension in the earliest Late Jurassic, but changing in the latest Jurassic due to a more concentrated extension, resulting in the formation of smaller but deeper graben systems (Færseth, 1996).

Conclusions

We have presented a holistic interpretation across the Central North Sea, which integrates the Permian-Triassic and Jurassic rifting. This study also demonstrates the importance of sub-salt, Caledonian- and Variscan-influenced structures in shaping the overlying younger halokinetic- and rift-related structures.

Based on the internal orientation of faults in the Mid Upper Permian (Rotliegend), the CG has been further subdivided into several structural domains. These domains are largely controlled by large NNW-SSE faults, which define areas of different salt wall patterns and distribution of Jurassic depocentres.

Qualitatively, the initial salt thickness and the pre-salt structure influence the thickness, length and orientation of salt walls. The salt walls are thickest and longest where initial salt deposits were thick while the underlying structure determined the orientation of these salt walls. The original basin orientation was NE-SW during the Permian as indicated by the thickness distribution of salt bodies but this was altered to NNW-SSE during the Late Jurassic rifting phase.

Isochore maps and palinspastic reconstruction demonstrate that halokinesis was most active during the Triassic rifting but less active during the Late Jurassic rifting as most salt walls has been welded during the earlier rifting phase. Most of the salt walls reached their current height at the end of

Triassic while a few situated near the confluence of major faults continued to rise during the subsequent Jurassic rifting and Tertiary compressional periods. Therefore, the influence of halokinesis on the tectono-stratigraphic evolution of the CG gradually diminished progressively from the Triassic to the Tertiary.

The analysis of the Triassic to Cretaceous isochore maps has shown that the Triassic rifting orientation preferentially follows the underlying NW-SE-striking Baltica-Avalonia suture. This structural pattern could also be traced upwards into the earliest Upper Jurassic structures (Oxfordian). In contrast, the youngest Upper Jurassic structure (Tithonian) seems to have changed into a predominantly NNW-SSE orientation. However, this change does not necessarily reflect a change in extension orientation, as oblique opening is also feasible.

Chapter 4: The impact of halokinesis and rifting on the development and distribution of depositional systems in the Upper Jurassic of the Central Graben Basin, UK Continental Shelf

Introduction

The Central Graben (CG) of the UK North Sea (Fig. 4.1), is one of the most intensively explored salt-influenced rift basins in the world and provides an exceptional example of how the effects of synchronous rifting and halokinesis have influenced the evolution and preservation of a range of mainly shallow- to deep-marine depositional systems. The CG also contains around 50% of UK's petroleum resources, which are located mainly within Mesozoic syn-rift sediments, most notably in Upper Jurassic shallow- to deep-marine sandstone reservoirs (Cordey, 1993; Erratt et al., 2010; Fraser et al., 2003; Johnson and Fisher, 1998; Ziegler, 1990b). Since the discovery of the Fulmar Field in 1975 (Johnson et al., 1986), the estimated ultimate recovery from the Jurassic of the CG is close to 3 billion barrel of oil equivalent (BBOE). In addition, the Upper Jurassic of the CG remains the most significant play in the North Sea with regards to its remaining exploration potential (Cordey, 1993; Eriksen et al., 2003; Erratt et al., 2010) and is directly linked to the CG's highly favourable, but extremely complex, petroleum geology, most notably due to the following: (1) the presence of a world class marine source rock (Kimmeridge Clay Formation) but with lateral variability in its burial history, petroleum generation and entrapment (Cayley, 1987; Erratt et al., 2005; Kubala et al., 2003), (2) a polyphase and strongly diachronous tectonic history, involving halokinesis, rifting and inversion (Chapter 3; Bartholomew et al., 1993; Coward et al., 2003; Roberts et al., 1990b; Zanella and Coward, 2003), (3) shallow- to deep-marine sandstones, with a high degree of lateral variability in terms of reservoir thickness and quality (e.g. Fraser et al., 2003; Sansom, 2010; Wonham et al.,

2014), (4) a retrogradational stratigraphic architecture with pronounced diachroneity (Partington et al., 1993a; Partington et al., 1993b; Rattey and Hayward, 1993), (5) a wide array of trap types with variable combinations of structure, stratigraphy and reservoir quality, which has resulted in a large field size distribution (e.g. Fraser et al., 2003; Rattey and Hayward, 1993), and (6) burial depths range from c. 2,500 to almost 6,000 m sub-sea, which has caused substantial depth-related diagenesis, and associated reservoir quality variability, within proven accumulations (Table 4.1).

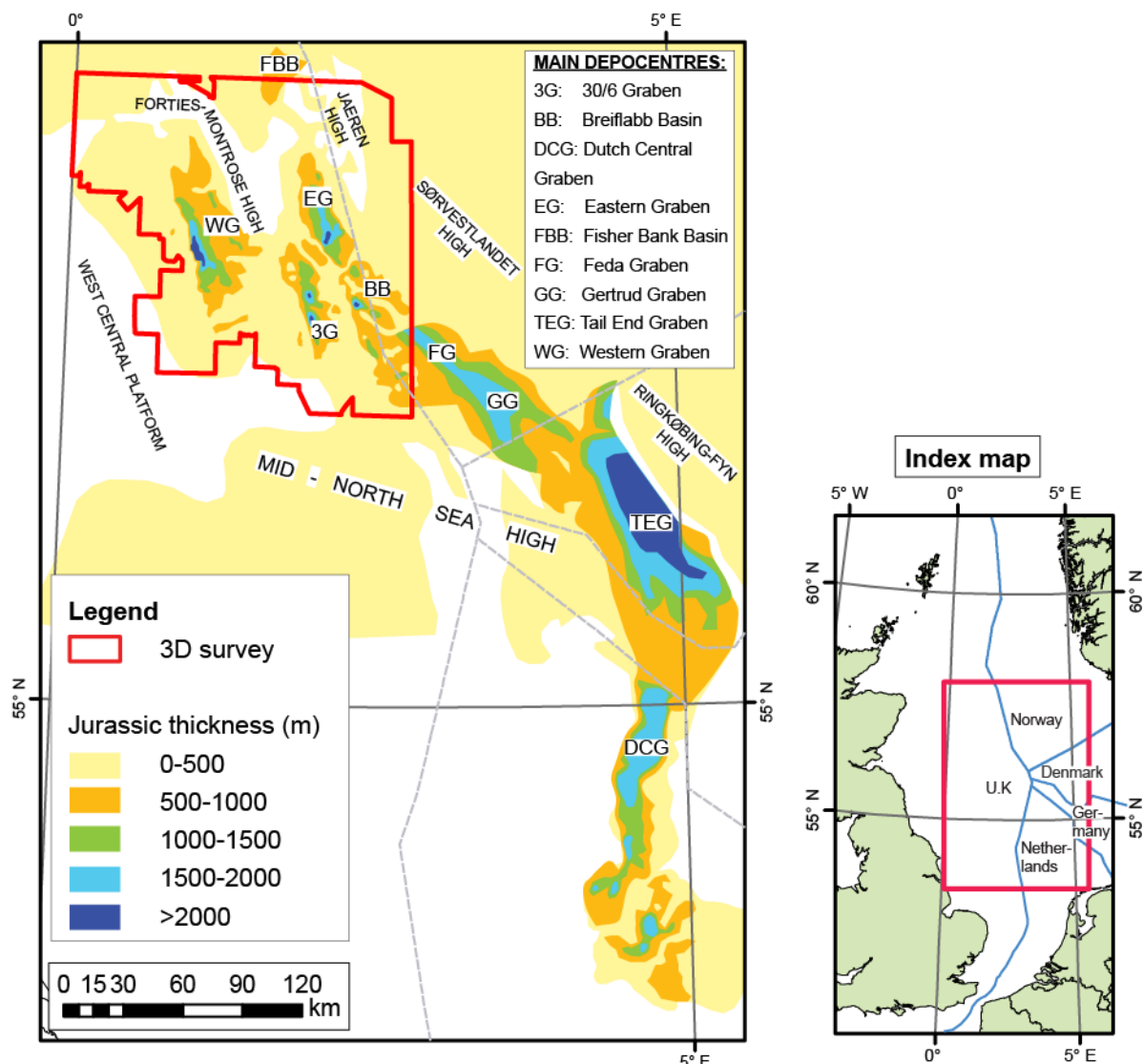


Fig. 4.1: A simplified Late Jurassic thickness map across the Central North Sea showing the highly segmented distribution of the main sub-basins and also the surrounding highs. The study area is highlighted in red, which also defines the extent of the 3D seismic survey used in this study. Thickness data from outside the study area are based on Duin et al. (2006), Fraser et al. (2003), and Sundsbø and Megson (1993)

(a)		Porosity (%) in different fields											
Field		Kittiwake	Curlew	Puffin	Shearwater	Elgin	Franklin	Erskine	Fulmar	Clyde	Fife	Fergus	
Depth (m)		2900-3000	3000-3400	4200-4400	4600-5200	5300-5700	5000-5600	3800-3900	3000-3500	3600-3800	2510-2600	2630-2680	
Facies Associations	Fluvial												
	Estuarine	} 10-27											
	Bay												
	Flood Plain	} 5-25											
	Upper Shoreface	18-32				20-25							
	Lower Shoreface	12-30	15-25	16-25	19-21	17	18	20-25	17-25	20-25	24	25	
	Mass flow					29-33				26			
(b)		Permeability (mD) in different fields											
Field		Kittiwake	Curlew	Puffin	Shearwater	Elgin	Franklin	Erskine	Fulmar	Clyde	Fife	Fergus	
Facies Associations	Fluvial												
	Estuarine	} 0.1-8000											
	Bay												
	Flood Plain	} 0.001-2000											
	Upper Shoreface	1000-3200				80-200							
	Lower Shoreface	1.0-700	30-174	0.14-25	5-25	25	10	0.1-10	2.8-784	0.1-2000	40	500	
	Mass flow					1-900				1012			
References:		Glennie & Armstrong (1991)	Eneyok et al. (2003)	Dickinson (1996)	Gilham et al. (2005); Taylor et al. (2015)	Lasocki et al. (1999)		Coward (2003)	Johnson et al. (1986); Kuhn et al. (2003)	Stevens & Wallis (1991)	Shepherd et al. (2003)		

Table 4.1: Previous page - Porosity (a) and permeability (b) ranges of different facies from various fields in the Central Graben reflecting the heterogeneity of Upper Jurassic Fulmar sandstones. The subdivision of facies has been modified to match the facies association used in this study (Table 4.2).

In addition to the above, the primary motivation for this study stems from the recognition that the interaction between tectonics and sedimentation is the most critical factor in controlling the geological variability of the Upper Jurassic of the CG, with active rifting and salt movement occurring simultaneously with source, reservoir and seal facies deposition (e.g. Erratt et al., 1999; Ziegler, 1990a; Ziegler, 1990b). For example, the variation of depositional environments is strongly influenced by active extensional tectonics and halokinesis, where the interplay between fault block rotation and salt diapir growth can influence sediment transport directions as well as the type and thickness of facies being deposited and preserved (Donovan et al., 1993; Fraser et al., 2003; Howell et al., 1996; Stewart et al., 1999; Wonham et al., 2014). Rifting may also cause salt diapirs to grow or shrink, depending on the initial location of the salt body in relation to the rotating fault blocks and the evolving rift margins (Clark et al., 1999; Erratt, 1993; Erratt et al., 1999; Hudec and Jackson, 2007; Penge et al., 1993). A major consequence of this interaction is that the wavelength of individual fault blocks and/or salt-influenced structures is much smaller (c. 2-5 km) compared with those that form in extensional basins that lack salt (e.g. 10-40 km in the Viking Graben; Rattey and Hayward, 1993). In situations where these processes were synchronous with deposition, as in the Late Jurassic of the CG, the shorter structural wavelength results in much smaller sub-basins and, therefore, more rapid lateral variability in thickness and facies distribution. This exerts a profound control on the creation and destruction of accommodation space and on the preservation potential of the broadly retrogradational Late Jurassic shallow- to deep-marine depositional systems, especially the thickness, extent and quality of sandstone reservoirs (e.g. Donovan et al., 1993; Wonham et al., 2014). The same tectono-stratigraphic interaction also controls field-scale reservoir architecture, reservoir performance and ultimate recovery, which is clearly demonstrated in the following fields: Angus (Hall, 1992; Spathopoulos et al., 2000), Durward-Dauntless (Stewart et al.,

1999), Elgin-Franklin (Lasocki et al., 1999; Lasocki et al., 1997), Fife (Currie et al., 1999; Mackertich, 1996), Fulmar (Johnson et al., 1986; Johnston et al., 1998; Kuhn et al., 2003; Spaak et al., 1999; Stockbridge and Gray, 1991), Gyda and Ula (Mannie et al., 2014; Stewart, 1993), and Shearwater (Blehaut et al., 1999; Gilham et al., 2005; Jeremiah and Nicholson, 1999).

A high-resolution understanding of the location and evolution of salt diapirs and active extensional faulting is critical in order to assess the evolution of depositional environments throughout the CG and adjacent areas. This work builds on the many previous studies that have been extensively documented, including the seminal work contained in (1) regular conference proceedings (Brooks et al., 1987; Doré and Vining, 2005; Fleet and Boldy, 1999; Illing and Hobson, 1981; Parker, 1993; Vining and Pickering, 2010; Woodland, 1975), and (2) atlas-format regional reviews (Evans et al., 2003; Ziegler, 1990a).

Building on this extensive earlier work, this contribution focuses on the analysis of an exceptionally rich and varied database, including 3D seismic, well logs, biostratigraphy, cores and a plethora of field data, and integrates them within a single, tectono-stratigraphic evaluation. There are thousands of wells and tens of thousands square kilometres of 3D seismic data blanketing the basin. However, previous studies on the depositional system of the Jurassic of the CG were mostly focussed on different areas of the basin and the results may not be applicable throughout the basin (Clark et al., 1999; Coward, 2003; Dickinson, 1996; Johnson et al., 1986; Kuhn et al., 2003; Lasocki et al., 1999; Mannie et al., 2014; Michelsen et al., 2003; Rossland et al., 2013; Spathopoulos et al., 2000; Stewart, 1993; Stewart et al., 1999; Wakefield et al., 1993). For example, there remain significant uncertainties, especially at the prospect scale, in predicting how the depositional facies distribution in the CG was affected by the synchronous halokinesis and rifting in the Jurassic (Fraser et al., 2003; Sansom, 2010; Wonham et al., 2014). In addition, this study also draws on other regional studies in

the adjacent Danish and Dutch North Sea basins and integrates them into the current study (e.g. Duin et al., 2006; Sundsbø and Megson, 1993).

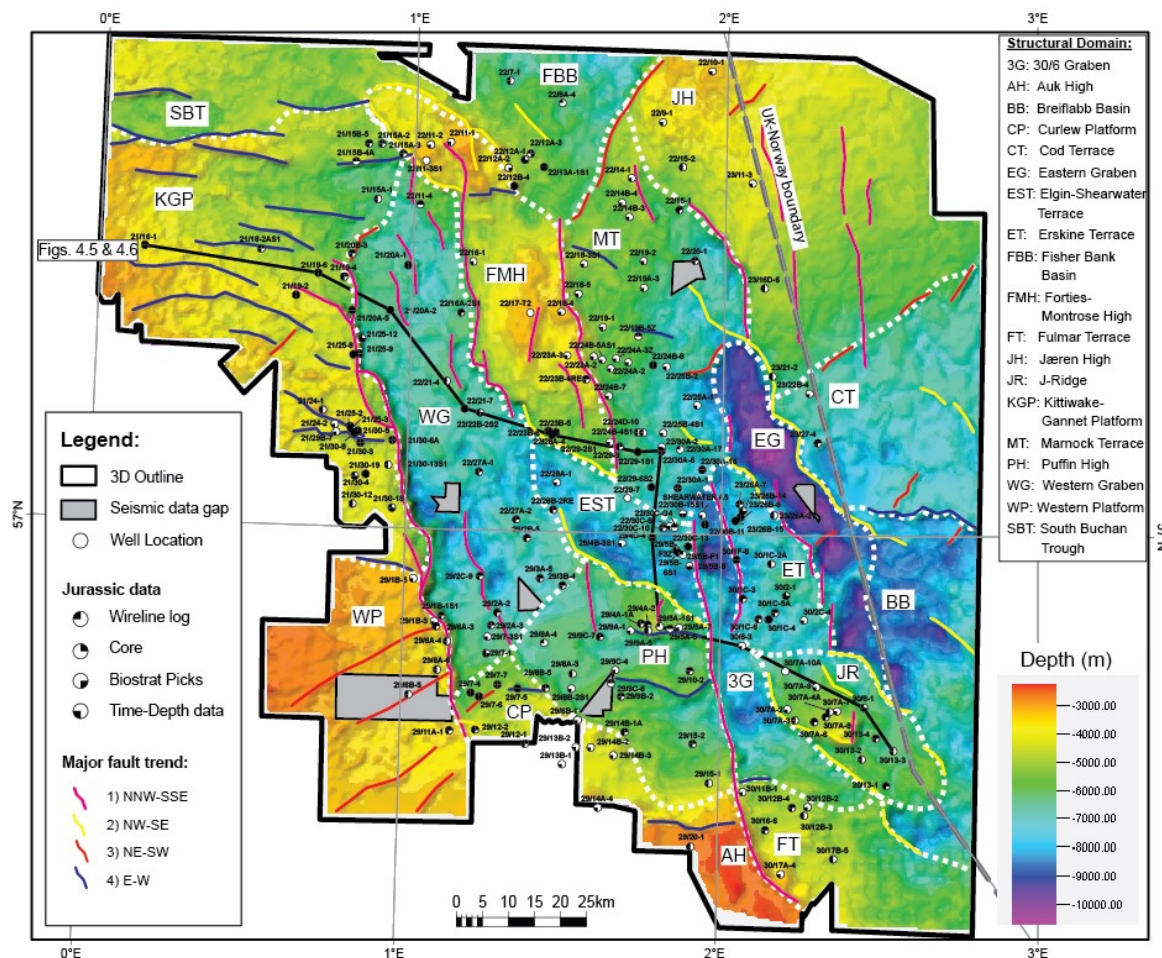


Fig. 4.2: Depth-structure map of Top Mid Upper Permian horizon (Top Rotliegend Group), which has been interpreted from the 3D seismic data (c. 22,360 km²). Superimposed on this map is the distribution of the 180 wells used in this evaluation of the Jurassic, which are coded according to the type of data available (see 'Jurassic Data' in legend). The major fault trends are also colour-coded according to their different orientations. The Late Jurassic structural domains defined in this study are outlined by the white dotted lines, which have been derived mainly from the underlying structural pattern of the Top Rotliegend horizon.

In particular, this study utilises a total of 22,360 km² of 3D seismic reflection data, comprising of a 3D seismic mega-survey provided by PGS covering the area from the Forties-Montrose High in the north to the Auk Ridge to the south (Fig. 4.2). We also have selected around 180 wells that penetrated the Jurassic section on the basis that these represent the various tectono-stratigraphic domains within

the greater CG area and also contain a full wireline log suite, together with lithology data from cuttings and cores. These wells also contain a total of length of c. 5.8 km of core, which have been reviewed, including an independent sedimentological synthesis of around 1,570 m of the core data. The latter were selected to be representative of both (1) the main depositional environments, and (2) the various tectono-stratigraphic domains across the whole of the CG (Fig. 4.2 and Table 4.2).

The main aim of this study is to provide an integrated analysis of the distribution of depositional facies within the Jurassic of the CG, including the impact of synchronous rifting and halokinesis on Late Jurassic shallow- to deep-marine depositional systems, particularly in terms of reservoir thickness development and preservation.

Stratigraphic Framework

Determining age and stratigraphic division in offshore wells of the Central North Sea has been challenging for a number of reasons: (1) initially a lack of published well and biostratigraphic data during the early stage of exploration in the basin (pre-1993); (2) subsequently, reconciling the many different stratigraphic schemes and nomenclatures used by the various companies that operate across the UK, Norway and Denmark parts of the basin (Fraser et al., 2003; Partington et al., 1993a); (3) lack of biostratigraphic samples in sandy formations (Price et al., 1993); (4) lack of preservation of palynomorphs in high temperature parts of the CG, especially in the central area of the basin (Partington et al., 1993a; Price et al., 1993); and (5) sub-seismic nature of the intra-Jurassic stratigraphic boundaries that hampers regional control of well correlation (Sansom, 2010).

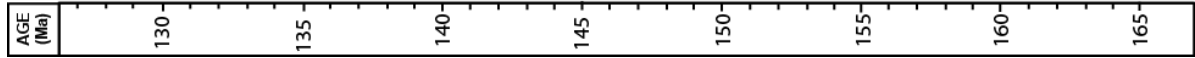
Central Graben Geological Domains	Well Name	Length of core (m)	Stratigraphy sampled
Curlew Platform	29/7-7	167.0	Sequence II & III
Elgin-Shearwater Terrace	22/30B-11	56.1	Sequence II
	22/30B-15S1	157.4	Sequence II
	22/30C-13	24.3	Sequence I & II
Erskine Terrace	23/26B-15	83.4	Sequence II
	30/01C-4	21.2	Sequence I
Fisher Bank Basin	22/12A-3	122.0	Sequence I & II
	22/13A-1	52.7	Sequence IV & V
Forties-Montrose High	22/23B-6	134.7	Sequence III, IV & V
Fulmar Terrace	30/12B-4	131.1	Sequence III & IV
	30/16-7	105.2	Sequence III & IV
J-Ridge	30/07A-8	12.2	Sequence V
	30/13-4	27.4	Sequence II
Kittiwake-Gannet Platform	21/25-2	28.6	Sequence II & III
	21/25-8	82.3	Sequence III
Marnock Terrace	22/29-1	16.2	Mid Jurassic, Sequence I & II
	22/29-3	21.0	Sequence I & II
	22/29-6S2	24.1	Sequence II
	22/30A-1	20.1	Sequence II
	22/30A-6	10.1	Sequence I & II
Puffin High	29/04A-1A	83.8	Sequence II & III
Western Graben	29/07-3	52.1	Sequence II
	22/21-7	50.0	Sequence II
	22/27A-1	32.3	Sequence II
	29/02A-2	53.9	Sequence II
Total core length		1569.2	

Table 4.2: List of cores from wells independently described in this study according to geological domain. For location of these wells, see Fig. 4.2.

Additional confusion also exists in the use of absolute time scales for the Jurassic, which have changed over the years mainly due to advancement in radiometric dating (Pálffy, 2008). The stage boundaries used by Partington et al (1993a), which is similar to the scheme used by Haq et al (1987), is c. 10 Ma younger than that used by Jeremiah & Nicholson (1999) (Fig. 4.3a, b and c). Furthermore, there are also significant differences between the definition of Upper Jurassic Stage boundaries from the Onshore UK, North Sea and other NW Europe areas due to the difficulties in correlating Boreal

and Mediterranean faunas (Cope, 2008; Pálffy, 2008). For example, the UK's Kimmeridgian Stage covers a longer period of time than its European counterpart (Cope, 2008; Price et al., 1993). However, since the sequence boundaries were tied to relative age, rather than absolute age, this discrepancy could be corrected as long as the definition of each boundary is clearly understood. Therefore, the biostratigraphic data used in this study have been tied to the International Commission of Stratigraphers (ICS) 2004 Chart (Gradstein et al., 2004) and we have used this version of stage boundaries in order to avoid further confusion (Fig. 4.3).

Partington et al (1993a) published the most complete stratigraphic framework based on a genetic sequence stratigraphic analysis, which has been widely-used ever since (Fig. 4.3b). In this scheme, the age of major flooding surfaces (MFS) was determined by offshore palynomorph fossils calibrated with onshore ammonite zones (Cope et al., 1980; Galloway, 1989; Wollam and Riding, 1983). A total of 12 MFS's were identified by Partington et al (1993a) for the Upper Jurassic section. This scheme was later refined and modified using palynomorphs from more wells (Jeremiah and Nicholson, 1999; Veldkamp et al., 1996). Another MFS was added later by Sansom (2010) to further subdivide the thick Kimmeridgian sequence (Fig. 4.3b). Consequently, most studies have divided the Upper Jurassic into seven to eight major sequences, albeit with different definitions, preferring to use either Vail-type sequence boundaries (Donovan et al., 1993; Jeremiah and Nicholson, 1999) or Galloway-type MFS boundaries (Fraser et al., 2003; Price et al., 1993).



(a) Haq et al (1987)

(b) Partington et al (1993), with modification by Samson (2010) (SA)

EPOCH	STAGE	SEQUENCE	SA
EARLY CRET.	RYAZANIAN	J76	J76
		J74	J74
		J73	J73
		J72	J71
		J71	-
		J70	J72
		J69	J66
		J68	J64
		J67	J63
		J66	J62b
		J65	J62a
		J64	J56
		J63	J54b
		J62	J54a
		J61	J52
		J60	J46
		J59	J44
		J58	J42
		J57	J36
		J56	J35
		J55	J34
		J54	J33
		J53	J32
		J52	J31
		J51	J30
		J50	J29
		J49	J28
		J48	J27
		J47	J26
		J46	J25
		J45	J24
		J44	J23
		J43	J22
		J42	J21
		J41	J20
		J40	J19
		J39	J18
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		J34	J13
		J33	J12
		J32	J11
		J31	J10
		J30	J9
		J29	J8
		J28	J7
		J27	J6
		J26	J5
		J25	J4
		J24	J3
		J23	J2
		J22	J1
		J21	J0
		J20	J-1
		J19	J-2
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		J15	J-6
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		J13	J-8
		J12	J-9
		J11	J-10
		J10	J-11
		J9	J-12
		J8	J-13
		J7	J-14
		J6	J-15
		J5	J-16
		J4	J-17
		J3	J-18
		J2	J-19
		J1	J-20
		J0	J-21
		J-1	J-22
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		J-4	J-25
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		J-9	J-30
		J-10	J-31
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Fig. 4.3: Previous page - Comparison and calibration between different North Sea Upper Jurassic stratigraphic schemes and the one used in this study (Shell, 2007). (a) The Epochs and Stages defined by Haq et al. (1987) and calibrated against the absolute time scale; (b) the stratigraphic scheme of Partington et al. (1993) and the J-sequence nomenclature defined by ammonite-defined maximum flooding surfaces (MFS) (key ammonite-defined MFS's are named), along with the further division of J62 sequence and combination of J71 & J72 by Samson (2010); (c) the Oxfordian to Volgian scheme of Jeremiah and Nicholson (1999), which was re-calibrated against the scheme of Partington et al (1993) (from Veldkamp et al., 1996); (d) the scheme from the North Sea Millennium Atlas (Fraser et al., 2003), including their sequence nomenclature; (e) the scheme used in this study utilizing the 2004 International Stratigraphic Commissions (ICS) ages (Shell, 2007 and Gradstein et al., 2004), including a revised calibration with the scheme of Partington et al (1993) shown in the last column (RP: 'Revised Partington'); and (f) the sequence nomenclature used in this study calibrated against the time scale in (c). Note that the absolute ages of the stage boundaries have changed significantly since Partington et al (1993)

However, it can be argued that the Galloway-type MFS-based approach is more appropriate compared to the unconformity-based scheme within syn-rift settings because major deepening coincides with times of major extension pulses (Martins-Neto and Catuneanu, 2010). These extension pulses generate abrupt increases of relative sea level and are more likely to be of basin-wide extent and to have a significant impact on sedimentation behaviour. This approach was also adopted by Fraser et al (2003), Gradstein et al (2004) and by Shell (2007) (Fig. 4.3d and e), and has been extended into this study (Fig. 4.3f).

This study identifies six main sequences (I-VI) within the Upper Jurassic of the CG using seven regionally-extensive, MFS-defined stratigraphic surfaces, which have been tied to both the Partington et al (1993a) and Shell (2007) biostratigraphic schemes (Fig. 4.4). These six sequences are as follows: (1) Sequence I, extends from J46 'Lamberti' MFS to J54a 'Glosense' MFS, which approximates to the early-middle Oxfordian; (2) Sequence II, is from the J54a 'Glosense' MFS to J62 'Baylei' MFS (late Oxfordian to earliest Kimmeridgian); (3) Sequence III, extends from the J62 'Baylei' MFS to J63 'Eudoxus' MFS (Kimmeridgian); (4) Sequence IV, extends from the J63 'Eudoxus' MFS to J66a 'Huddlestoni' MFS (late Kimmeridgian to early Tithonian); (5) Sequence V, extends from the

J66a 'Huddlestoni' MFS to J73 'Augiformis' MFS (Tithonian); and (6) Sequence VI, extends from the J73 'Augiformis' MFS up to J74 MFS (latest Tithonian).

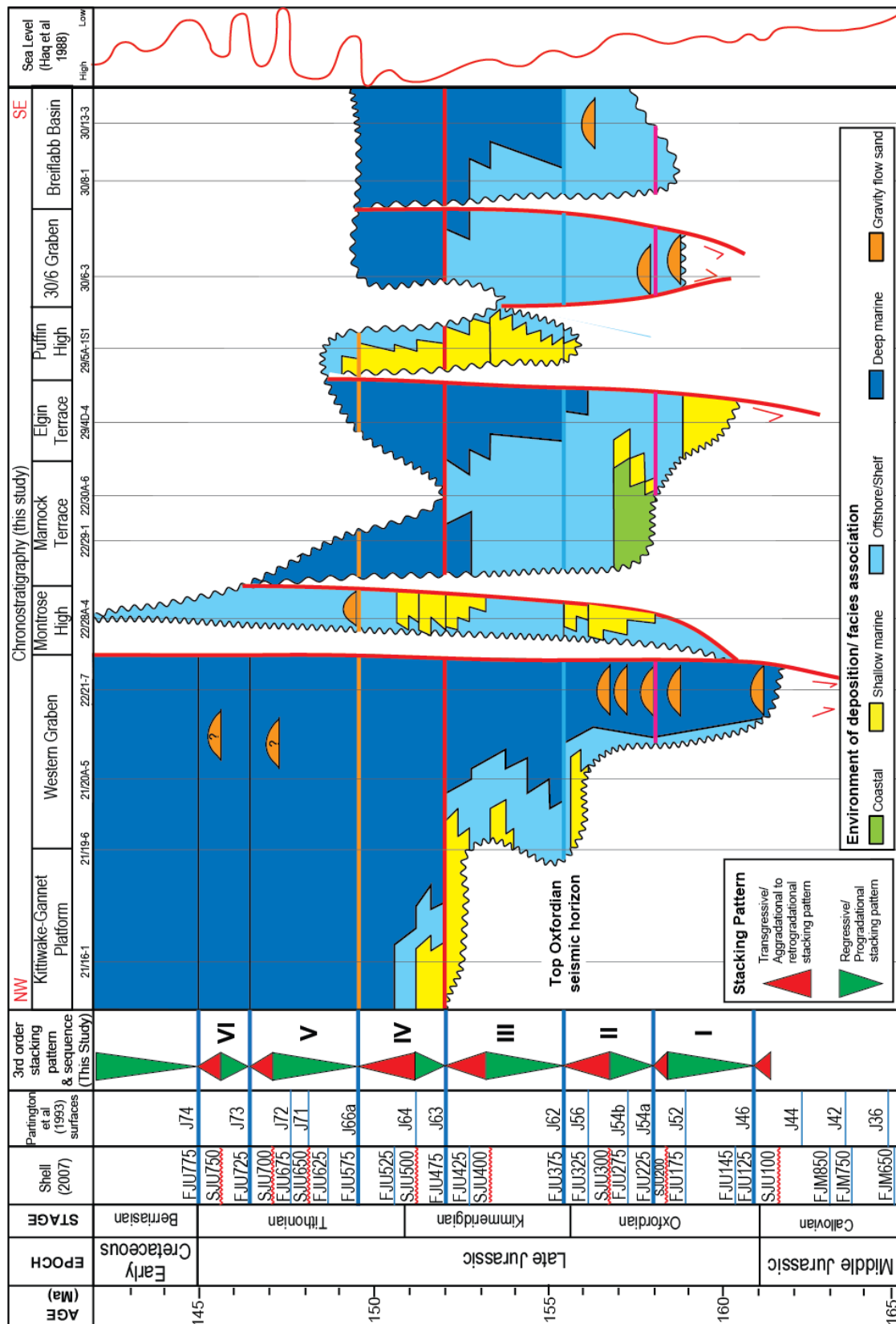


Fig. 4.4: Previous page - The chrono-, bio- and sequence-stratigraphic framework used in this study (left side) is plotted against a representative regional NW-SE oriented chronostratigraphic chart across the study area. The line of section corresponds to the well correlation panel and seismic section illustrated in Figs. 5 & 6, respectively. Seven MFS's (oldest to youngest: J46, J54a, J62, J63, J66a, J73 and J74) define six complete progradational to retrogradational successions within the Late Jurassic (Sequence I to VI). Note (1) the pronounced temporal and spatial segmentation across the various depocenters and intervening structural highs, (2) the main basin-bounding syn-rift faults, and (3) the global sea-level curve of Haq et al (1988), showing the overall rising trend during the Callovian to early Tithonian period (right column).

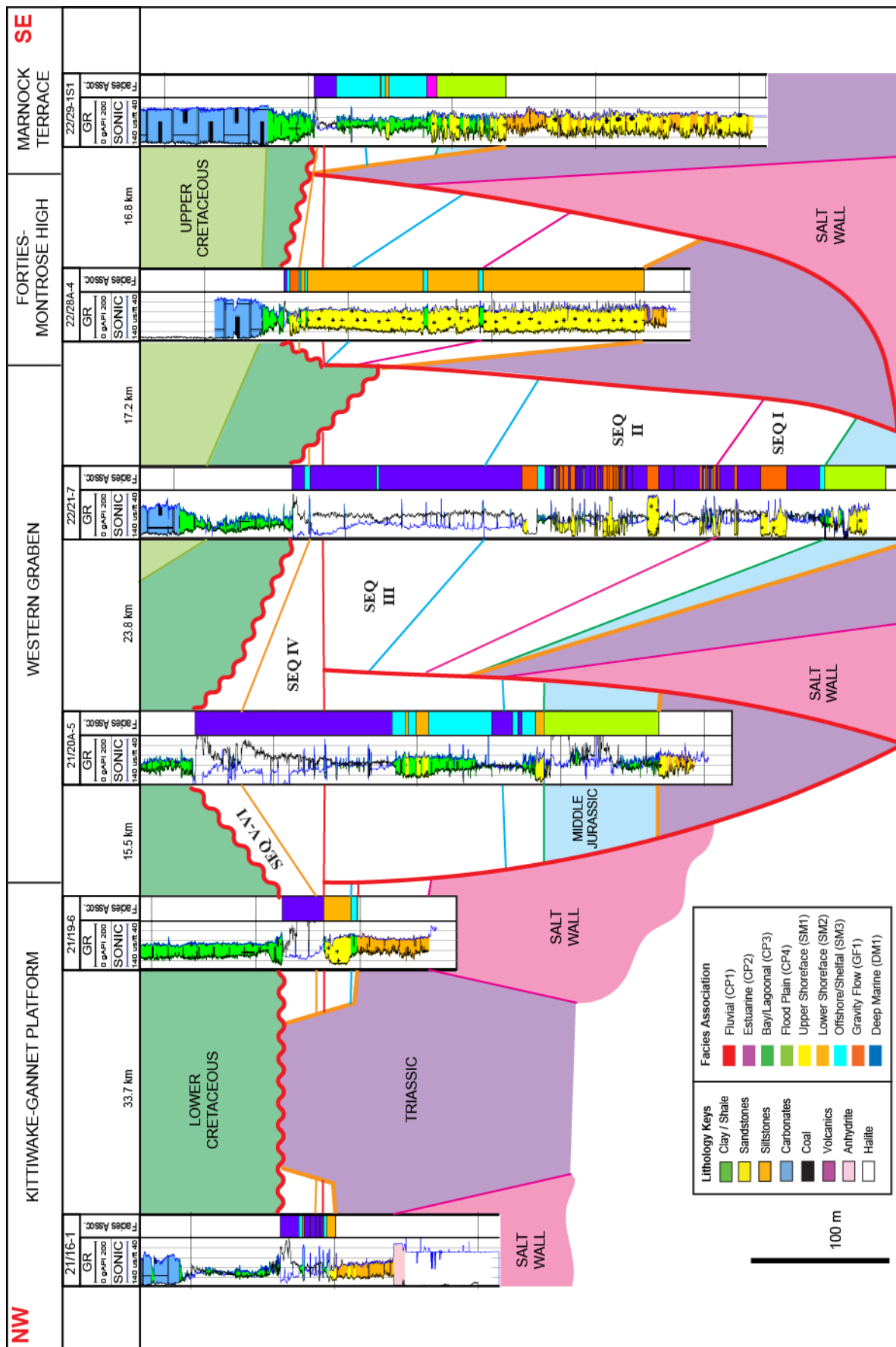
Jurassic rifting in the Central Graben

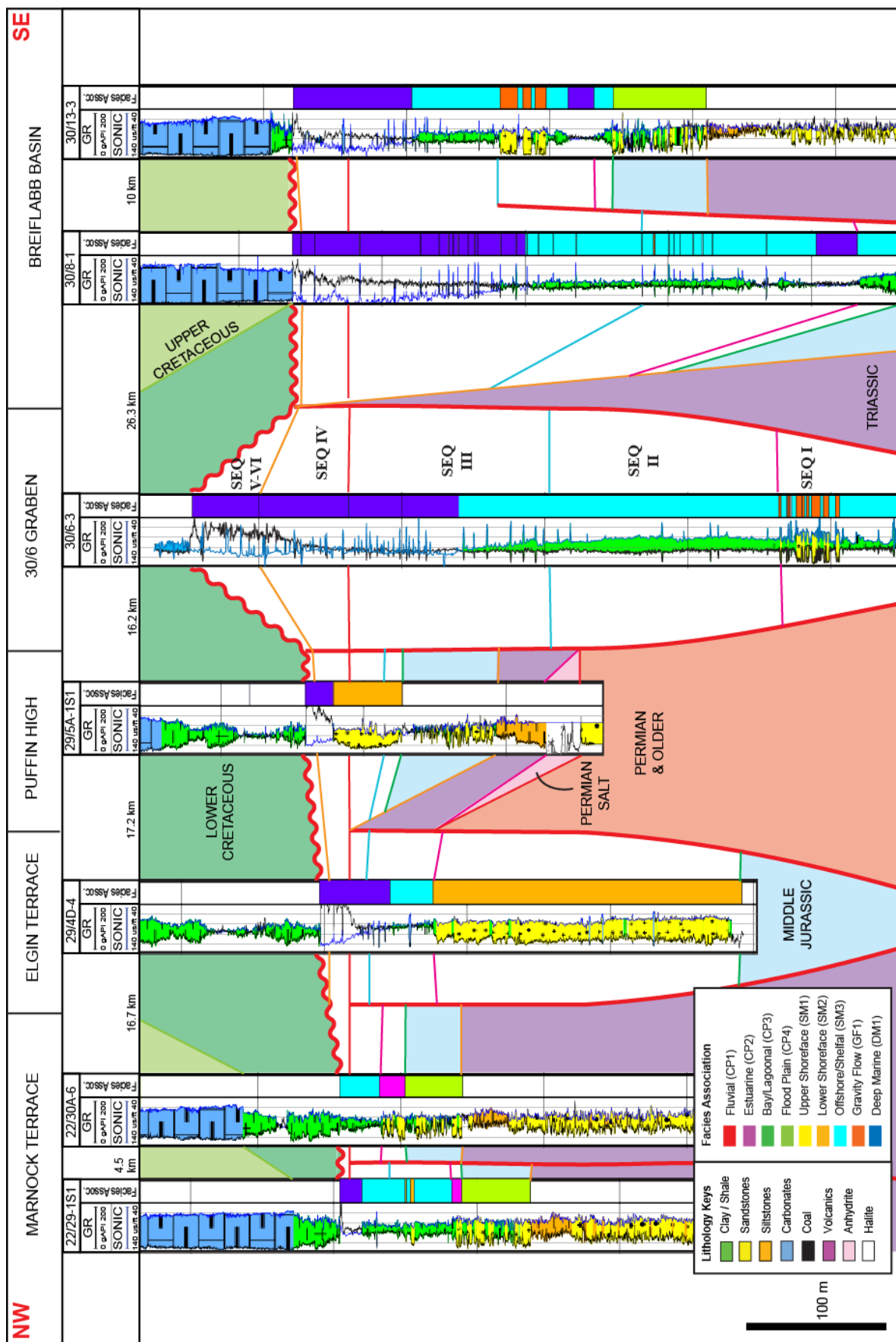
A regional NW-SE well log correlation panel (Fig. 4.5) and a corresponding seismic line extracted from the 3D seismic volume (Fig. 4.6) demonstrate the pronounced lateral variability in stratigraphy across several distinctive structural domains (see Fig. 4.2 for location and Chapter 3 for full description). It demonstrates that the distribution of Upper Jurassic depositional environments in the CG was strongly controlled by the regional structuration, most notably rifting, that has shaped both the creation and destruction of accommodation space. The present structural configuration of the CG (Fig. 4.1) is a reflection of tectonic events that occurred before, during and after the Jurassic rifting. The pre-Jurassic rift events that formed the initial structural framework of the basin comprised the following (Fig. 3.4): (1) the Caledonian Orogeny (Ordovician to Devonian), which resulted in a series of NE-SW structural lineaments where Laurentia collided with both Baltica and Avalonia (Iapetus Suture) as well as NW-SE lineaments where Baltica and Avalonia collided (Tornquist Line) (Lyngsie et al., 2006; McKerrow et al., 2000; Ziegler, 1992; Ziegler and Dèzes, 2006); (2) the Variscan Orogeny (Late Carboniferous), when a series of E-W and NE-SW were formed (Coward et al., 2003; Lyngsie et al., 2006; Ziegler, 1992); (3) the latest Carboniferous to Early Permian rearrangement of the convergence direction of Gondwana and Laurasia that formed NW-SE trending extensional faults (Clark et al., 1999; Coward et al., 2003; Ziegler, 1992); (4) Permian rifting, when two E-W trending Permian basins were subsequently formed (Glennie et al., 2003; Neumann et al., 2004; Stemmerik et al., 2000); (5) Triassic rifting that rejuvenated the older faults and

triggered the major late Permian salt withdrawals (Coward et al., 2003; Ziegler, 1992); and (6) the early to middle Jurassic North Sea Dome event (Aalenian to Bajocian) that was formed by a large volcanic centre at the triple junction between the Central, Viking and Moray Firth-Witch Ground grabens and formed the 'Mid Cimmerian' unconformity over the CG (Rathey and Hayward, 1993; Underhill and Partington, 1993; Ziegler, 1992).

The effect of this long period (>280 Ma) of tectonic activity is reflected in the array of distinctive structural trends that criss-cross the CG (Fig. 4.2): (1) NE-SW; (2) NW-SE; (3) E-W; and (4) NNW-SSE. The interaction between these trends has resulted in the subdivision of the CG into several structural domains, most of which show inheritance from earlier (pre-Jurassic) trends (Fig. 4.2). These structural domains in turn also exerted a strong influence on (1) Jurassic accommodation space history, and (2) the shape, orientation and dimensions of structures formed during the Jurassic rifting, both within the grabens, on the horst blocks and on the adjacent graben flanks (Fig. 4.2). Wherever possible, domain names identified by previous workers have been used, but new names and redefined structural boundaries have been added into this study (Bartholomew et al., 1993; Clark et al., 1993; Erratt, 1993; Erratt et al., 1999; Lasocki et al., 1999; Stewart et al., 1999; Zanella and Coward, 2003).

Fig. 4.5: Next two pages - A NW-SE oriented well correlation panel in two continuous sections across the Central Graben datumed at Top J63 surface (near Base Tithonian). The line of section is shown in Fig. 4.2. The well log display in the left track combines the Gamma Ray (GR; 0-200 API units) and Sonic (140-40 μ s/ft). The lithology interpretations in the GR/Sonic overlay (calibrated with core and cuttings data, where available) are shown in the lithology legend (lower left). The right hand track contains the Facies Association interpretations based on the analysis of about 1.6 km of core in this study (color coding shown in the legend). The calibration of core descriptions with porosity/permeability data well log response is discussed in the text and summarized in Table 4.1 and 4.3, respectively.





The Jurassic rifting itself may have already started in the Middle Jurassic (latest Bathonian), after the deflation of the North Sea Dome (Davies et al., 1999). However, lack of well penetration into the Middle and Early Jurassic, unreliable biostratigraphic data and poor seismic image have contributed to the uncertainties in determining the exact timing of the start of extension. Other workers have proposed a Callovian or Oxfordian start to Jurassic rifting (Coward et al., 2003; Erratt et al., 1999; Rattey and Hayward, 1993). Furthermore, the start of rifting also varied with geographical position, with rifting starting in the late Aalenian (Middle Jurassic) in both the Viking Graben in the north and in the Danish Central Graben in the south (Andsbjerg et al., 2001; Zanella and Coward, 2003). Hence, rifting propagated both from north (Viking Graben) and from the south (Danish Central Graben via the Gertrud and Feda grabens (Fig. 4.1), before eventually merging within the present CG area.

Initially, the early phase of rifting exploited the older Triassic structures as demonstrated by the thicker Oxfordian interval within the NW-SE-trending grabens, which also defines the main Triassic structural grain (Fig. 4.7a). However, younger phases formed almost N-S striking grabens with the most prominent being the Western, Eastern and 30/6 grabens (Fig. 4.7b). This N-S trend continues southwards into the Tail End Graben and the Dutch Central Graben (Fig. 4.1).

The effect of halokinesis on the Jurassic interval is subtle compared to its more obvious impact on the Triassic interval. Initially, the effect of halokinesis was confined to the main grabens, where the Oxfordian sediment thicknesses were varied due to the collapse of salt walls. Later, the impact of halokinesis spread on to the highs during the Kimmeridgian to Tithonian as the rifting accelerated and rising relative sea-level quickly drowned the grabens. During this latest rifting phase, basement extension had caused salt walls to collapse, creating accommodation space above the walls that was eventually infilled by sediments (Coward et al., 2003; Fraser et al., 2003; Rattey and Hayward, 1993; Zanella and Coward, 2003; Ziegler, 1992).

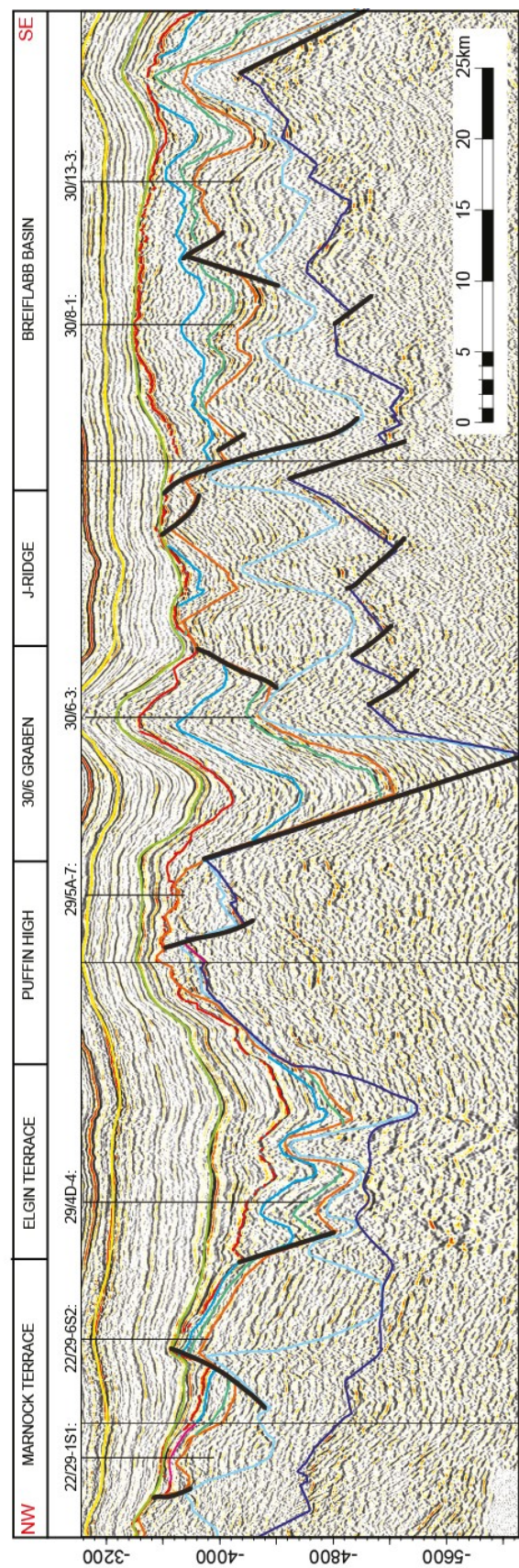
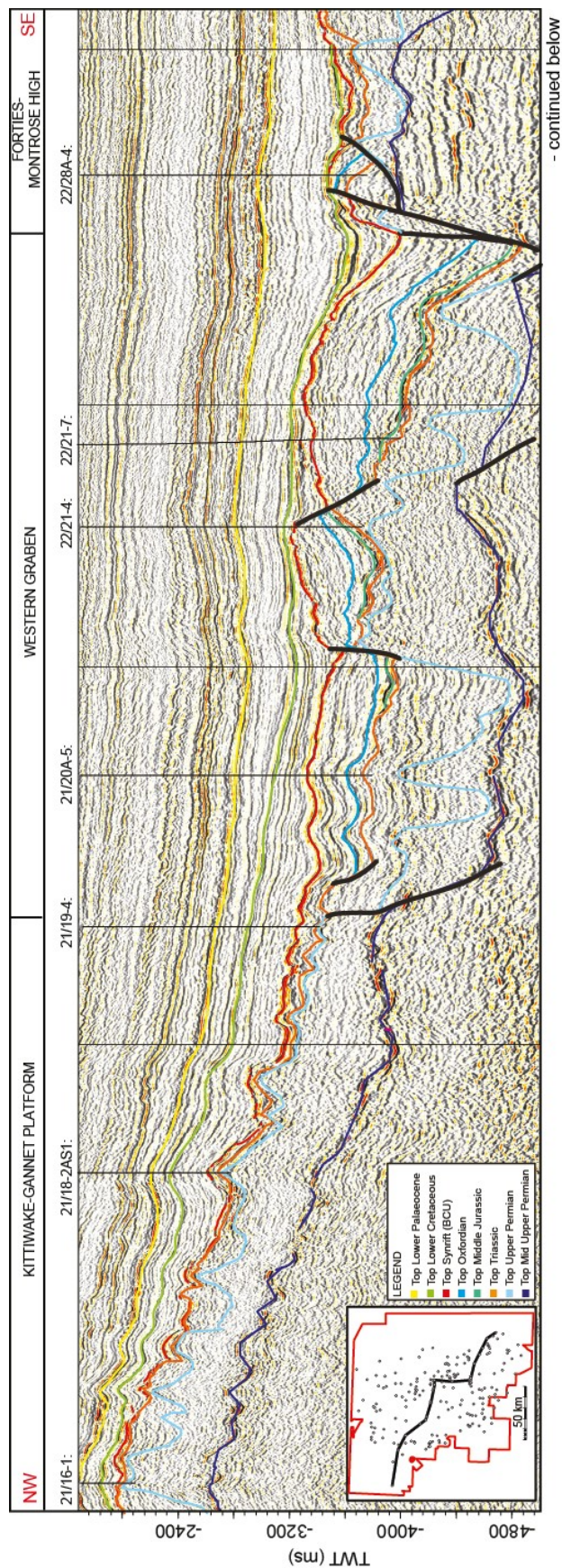


Fig. 4.6: Previous page - An interpreted, regional (c. 200 km-long), NW-SE oriented seismic section across the Central Graben (see inset map for line location). Eight seismic horizons have been interpreted throughout the 3D seismic volume and tied to the 180 wells used in the study area (see selected wells shown along the line of section and the accompanying well log correlation panel in Fig. 4.5). These data and interpretations have been used to define the regional stratigraphic architecture and structural styles throughout the region. The main structural domains are highlighted along the top of the section. See Fig. 4.5 for the detailed internal stratigraphic architecture for each well location.

Rifting gradually subsided in the Early Cretaceous, with its cessation marked by the Base Cretaceous Unconformity (BCU)(Coward et al., 2003; Ziegler, 1992). However, post-rift thermal subsidence throughout the remainder of the Mesozoic continued to exploit the Jurassic rift northerly trend (Fig. 4.7c) and Triassic north-westerly trend (Fig. 4.7d). The most notable post-Jurassic rift tectonic events that have further shaped the CG are as follows: (1) late Cretaceous Alpine Orogeny that resulted in the inversion of faults close to the Mid-North Sea High, continued growth of some Permian Zechstein salt diapirs and the formation of several unconformities within the Chalk Group (Coward and Dietrich, 1989; Surlyk et al., 2003; Ziegler and Dèzes, 2006); and (2) early Tertiary uplift associated with the opening of North Atlantic that resulted in modification of tall salt diapirs into tear-drop-shaped and regional influx of clastic basin floor fan systems above the Chalk Group (Ahmadi et al., 2003; Mudge and Jones, 2004; White, 1989).

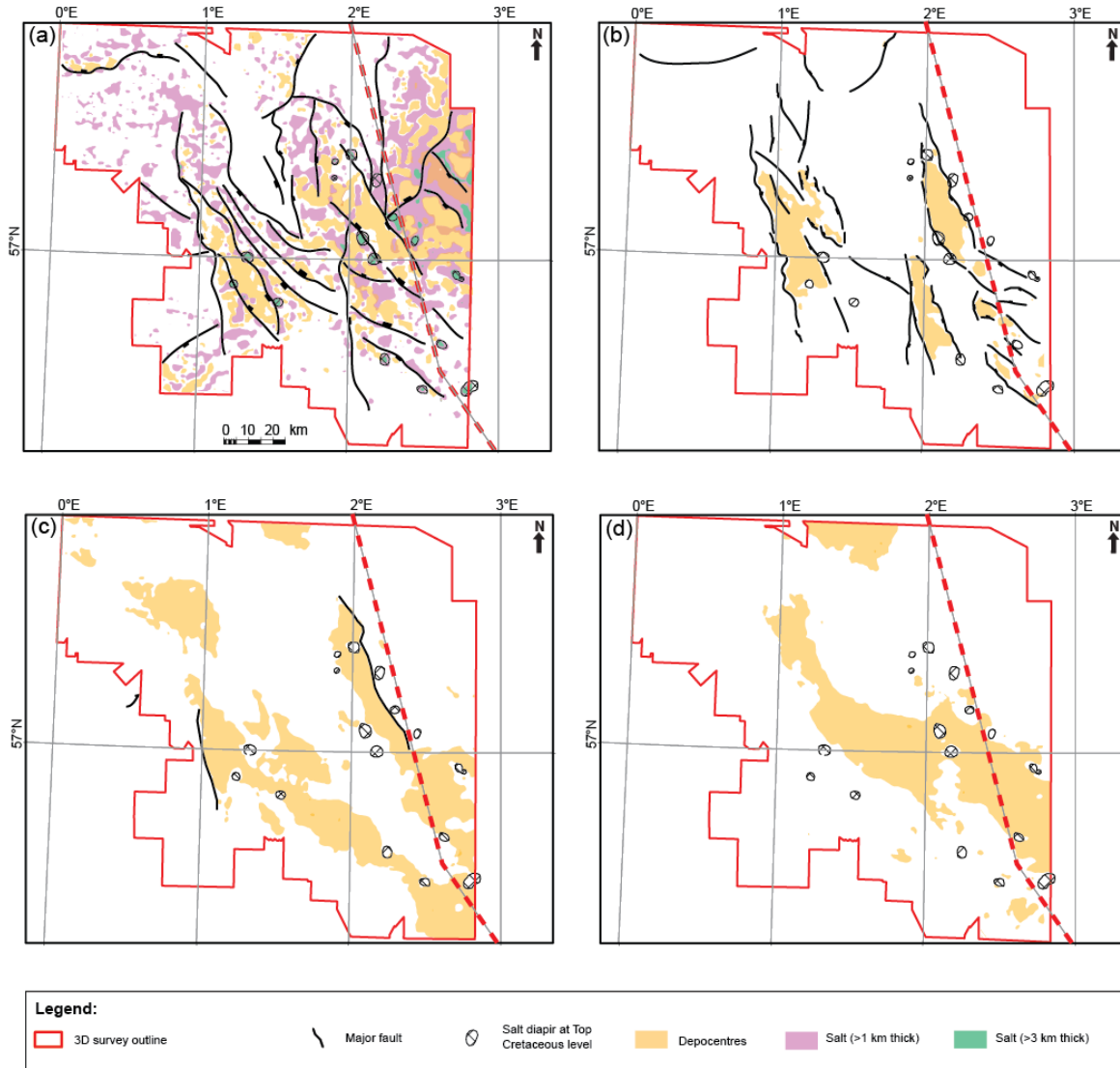


Fig. 4.7: A series of simplified, composite structure maps of the study area showing the evolution of active faults at four time periods (Triassic, Jurassic, Lower Cretaceous and Upper Cretaceous) and associated depocentres (orange areas). Prominent and active salt bodies are shown (>1 km thick = purple and >3 km thick = green), together with those that extend to the Top Cretaceous level (shaded). (a) Triassic faults have a general NW-striking pattern, which are cross-cut by NNW-striking faults. Rare NE-striking faults are overprinted by the other two fault patterns. Overall, the depocenters have a NW-SE trend, but there are numerous local variations. The salt diapirs have an irregular distribution but overall they follow the orientations of those faults that were active during this period. (b) Jurassic faults and adjacent rift-controlled depocentres are oriented mainly NNW-SSE and display stronger compartmentalisation. There are some subordinate faults that display a NW-SE orientation. (c) Lower Cretaceous depocentres have a NW-SE orientation, there are longer and less compartmentalized and the number of active faults has declined dramatically; only two NNW-trending bounding faults (bounding the Western and Eastern Grabens) remained active. (d) The Upper Cretaceous shows that the depocentres have merged into a single, connected NW-trending basin, whose axis runs along the middle of the study area.

Main Depositional Facies

The main Upper Jurassic depositional facies in the CG have been synthesised from c.1.5 km of conventional cores, which covers the full stratigraphic and geographic range of the study area (Fig. 4.2 and Table 4.2). This facies analysis was supplemented by observations from previous studies such as Gowland (1996) and Howell et al (1996), among others. The cores were examined to establish the sedimentary facies and stratigraphic boundaries within the Jurassic section, with emphasis on the documentation of lithology, grain size, sedimentary structures and trace fossils.

The facies identified in this study have been grouped into nine facies associations (FAs; Table 4.3). The main characteristics of the FAs are summarised below in relation to their distribution within three gross depositional environments (GDE; Fig. 4.8): (1) coastal plain, (2) shallow marine, and (3) deep marine environments.

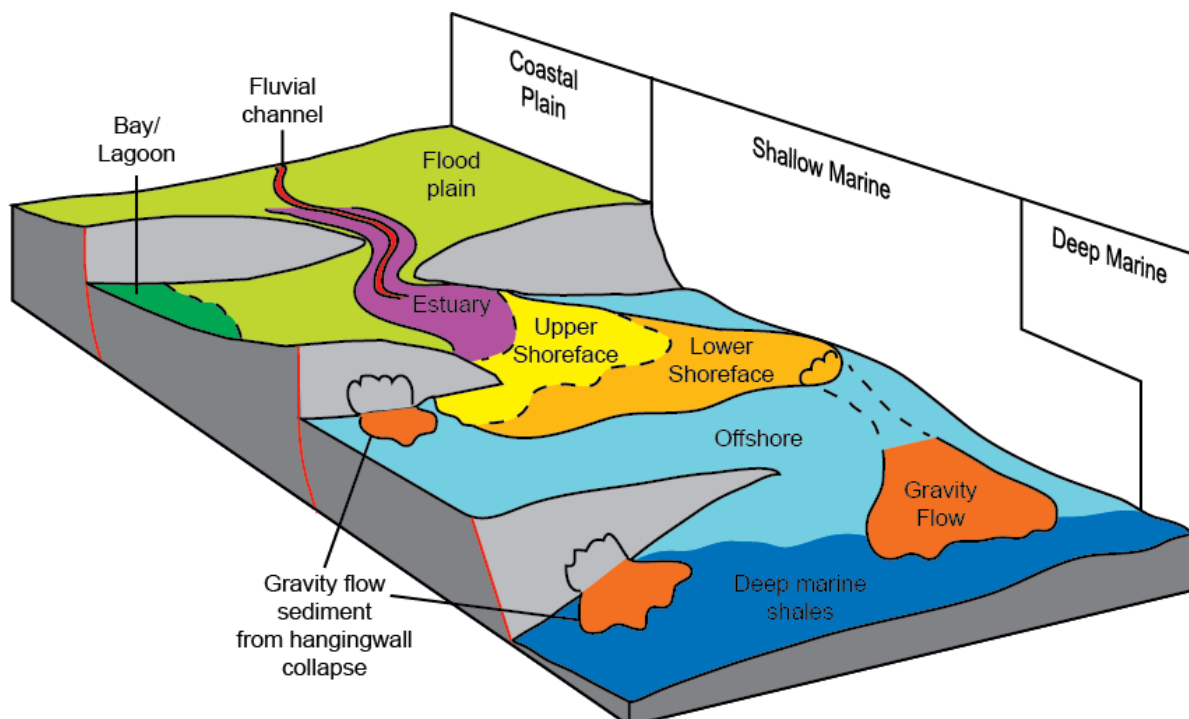


Fig. 4.8: Schematic diagram showing the general depositional environment model and facies associations used in the study (modified after Howell et al, 1996).

Facies association	Interval thickness	Sand content	Lithology and sedimentary structures	Trace fossil	Process interpretation & environment	Wireline log character	Ø and k
CP1: Fluvial (Fig. 4.9a)	5-40 m	90-100%	Medium- to coarse-grained, blocky to fining upward, unidirectional trough cross-bedded sandstones. Occasional pebbles and granules near bed bases. Usually abrupt lithological changes across beds.	Rare rootlets	High-energy, unidirectional currents depositing sinuous crested dunes in fluvial channels. Granules represent channel lags.	Blocky, relatively low GR log (<30 API) and moderate DT log (<100 µs/ft). Neutron log tracks density log but with positive separation in hydrocarbon-bearing zones. Stringers of high density (>2.5 g/cm ³) with low neutron (<9 v/v) related to channel lags (Fig.4.9a).	Ø: 10 - 27 %, k= 0.1 - 8000 mD
CP2: Estuarine (Fig. 4.9a)	1-10 m	30-50%	Very fine- to coarse-grained, blocky to fining upward, trough cross-bedded sandstones. Dispersed coal fragments. Occasionally wavy, ripple and parallel lamination, including mm-cm-scale sand-mud layers. Usually gradual lithological changes across beds.	Rare <i>Ophiomorpha</i> , <i>Thalassinoides</i> , <i>Skolithos</i> and <i>Teichichnus</i>	Low- to high-energy sand deposition with variable flow conditions. Suspension fall-out of mud. Fluvial-tidal processes in a protected, inshore/ estuarine environments.	Funnel-shaped, relatively low GR log (<50 API) and moderate DT log (<100 µs/ft). Generally, neutron and density logs are tracking each other with positive separation in hydrocarbon-bearing zones; negative separation in shaly intervals. (Fig. 4.9a)	Ø: 10 - 27 %, k= 0.1 - 8000 mD
CP3: Bay (Fig. 4.9a)	5-20 m	10-20%	Interbedded siltstones, clays and fine-grained sandstones, with disseminated coal fragments. Mainly fining upwards, but occasionally coarsening upwards, facies successions. Common cross-bedding, wavy- to flaser-bedding and parallel lamination. Occasional load structures. Usually abrupt lithological changes across beds.	Abundant, but rather limited to <i>Teichichnus</i> and <i>Thalassinoides</i> . Rare shell fragments.	Suspension settling of fine-grained sediments. Episodic introduction or coarse-grained sediments with mixed wave and tidal reworking within a restricted inshore/bay setting. A biologically-stressed environment.	Serrated, funnel- to bell-shaped, relatively moderate to high GR (50-100 API) and DT (80-120 µs/ft) logs. Generally high density (>2.5 g/cm ³) and neutron (>20 v/v) logs indicating mainly shaly intervals. Stringers of very low density (<2 g/cm ³) and high DT (>120 µs/ft) related values to coals. Streaks of high density (>2.75 g/cm ³), low DT (<80 µs/ft) with low neutron (<9 v/v) peaks associated with shell-rich intervals. (Fig. 4.9a)	Ø= 5 - 25 %, k= 0.001 - 2000 mD

Facies association	Interval thickness	Sand content	Lithology and sedimentary structures	Trace fossil	Process interpretation & environment	Wireline log character	ø and k
CP4: Flood plain (Fig. 4.9a)	1-10 m	5-10%	Interbedded siltstones, clays, fine-grained sandstones and coals in blocky to fining upwards packages. Occasionally, sandstones are coarsening upwards. Frequent cross-bedding, wavy- to flaser-bedding and parallel lamination in the fine grain strata. Biogenically-reworked, rooted and leached towards top of beds. Rare pyrite concretions and load structures. Gradual lithological changes across beds.	Rare <i>Planolites</i> . Occasional rootlets	Soil processes, churned by plant roots. Coarse sediment supplied from adjacent fluvial channels in the form of crevasse splays during river floods. Deposition of in-situ organic matter, including coal layers.	Serrated, funnel- to bell-shaped, relatively moderate to high GR (>90 API) and DT (>70 µs/ft) logs. Streaks of low density, high DT with high neutron peaks related to coals (Fig. 4.9a).	ø= 5 - 25 %, k= 0.001 - 2000 mD
SM1: Upper shoreface (Fig. 4.9b)	1-10 m	90-100%	Medium- to coarse-grained, often structureless, with occasional parallel laminations and trough cross bedding. Rare clay-sand and/or mm-cm, fine to medium grained sandy couplets. Occasional wavy and flaser bedding. Gradual lithological changes across beds with coarsening upward grain size profiles.	Occasional <i>Ophiomorpha</i> , <i>Thalassinoides</i> , <i>Skolithos</i> and <i>Planolites</i>	Migration of sand bars due to longshore and/or tidal currents above fairweather wave base. Occasional tidal-influence observed, possibly in sand bars adjacent to fluvial-tidal channel mouths.	Funnel-shaped, relatively low GR (<50 API) and DT (<100 µs/ft) logs. Neutron log tracks density log but positive separation of neutron and density logs occurs in hydrocarbon-bearing zones (Fig. 4.9b).	ø= 18 - 32 %, k= 80 - 3200 mD

Facies association	Interval thickness	Sand content	Lithology and sedimentary structures	Trace fossil	Process interpretation & environment	Wireline log character	Ø and k
SM2: Lower shoreface (Fig. 4.9b)	5-150 m	80-90%	Very-fine- to fine-grained, mainly coarsening upwards, silty sandstones. Largely destratified due to intense biogenic reworking. Rare preservation of flaser and parallel lamination where bioturbation is less intense. Gradual lithological changes across beds.	Abundant <i>Thalassinoides</i> , <i>Chondrites</i> , <i>Paleophycus</i> , <i>Planolites</i> , <i>Ophiomorpha</i> , and <i>Skolithos</i> . Occasional shell fragments.	Deposition mainly below fair-weather wave base. Occasional storm-deposition but sands largely reworked by burrowing infauna during fair-weather periods in a lower shoreface setting.	Serrated, funnel-shaped, relatively moderate to high GR (50-90 API) and DT (70-100 µs/ft) logs. Generally, neutron log tracks density log but relatively small positive separation occurs in hydrocarbon-bearing zones (Fig. 4.9b).	Ø= 12 - 30 %, k= 0.1 - 2000 mD
SM3: Offshore/shelf (Fig. 4.9b)	5-60 m	0%	Interbedded clay and siltstone (mm-cm-scale). Parallel laminated. Rare soft sediment deformation. Abrupt lithological changes across beds.	Occasional <i>Chondrites</i> , <i>Teichichnus</i> , <i>Zoophycos</i> and <i>Terebellina</i>	Suspension settling on an oxygenated shelf below fair-weather wave-base. Occasional input of silty-sand during high-energy storms.	Serrated, blocky, relatively high GR log (>50 API) and moderate DT log (>70 µs/ft). Negative separation of neutron and density logs (Fig. 4.9b).	Not applicable
GF1: Gravity flow (Fig. 4.9c)	1-30 m	95-100%	Generally medium-grained, structureless, blocky sandstones. Rare pebble layers, parallel lamination and soft sediment deformation. Abrupt lithological change across beds. Intercalated mudstone layers.	None	Coarser grained, thicker sandstone beds deposited from high-density turbidity currents. Finer grain sediments deposited by low-density turbidity currents.	Blocky, relatively low GR log (<30 API) and moderate sonic log (60-120 µs/ft). Generally, neutron log tracks density log but positive separation occurs in hydrocarbon-bearing zones (Fig. 4.9c).	Ø= 26 - 33 %, k= 150 - 1000 mD
DM1: Deep marine shale (Fig. 4.9c)	10s-100s m	0%	Organic-rich clay inter-bedded with siltstones. Pin-striped, parallel laminations. Abrupt lithological changes across beds.	None	Deposition of fine grain sediment by suspension settling. Occasional input of silts during large storms or gravity flows. Anoxic basin floor setting.	Blocky, relatively very high GR (>150 API) and DT (>100 µs/ft) logs. Negative separation of density and neutron logs (Fig. 4.9c).	Not applicable

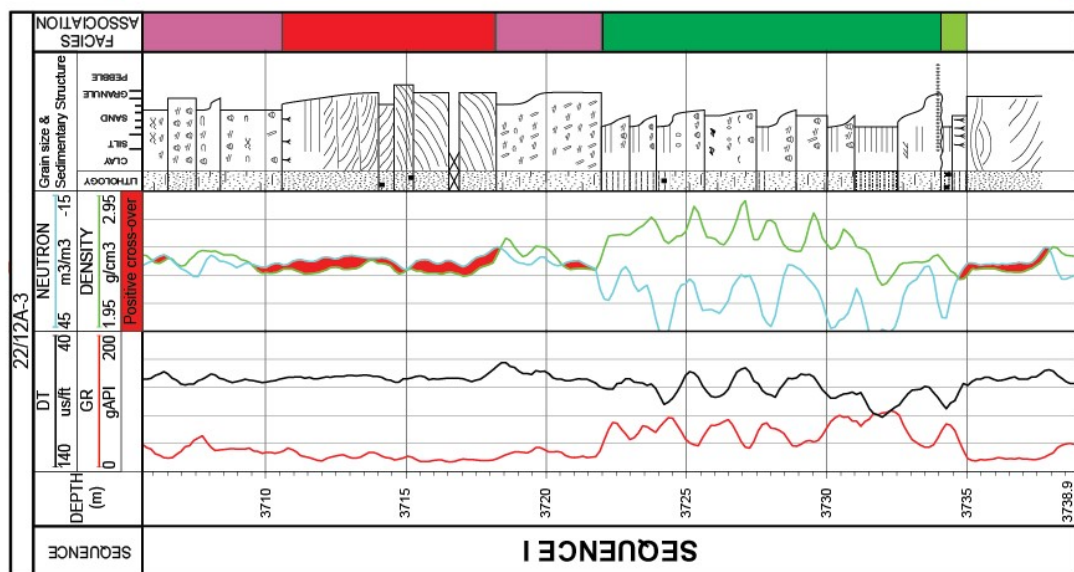
Table 4.3: Previous three pages - Description of facies associations in the Upper Jurassic interval of the Central North Sea with additional information from Howell et al. (1996) and Gowland (1996). GR: Gamma Ray, DT: Sonic, ϕ : Porosity range, k: permeability range. For source of reservoir porosity and permeability, see Table 4.1.

Coastal plain environment

The coastal plain environment comprises of four FAs (Table 4.3; Figs. 4.9a), which are listed here in order of sand content: (1) fluvial (CP1), (2) estuarine (CP2), (3) bay/lagoonal (CP3), and (4) flood plain (CP4). All the facies types within the coastal plain environment exhibit abundant plant and coaly material, which is in contrast to other environments.

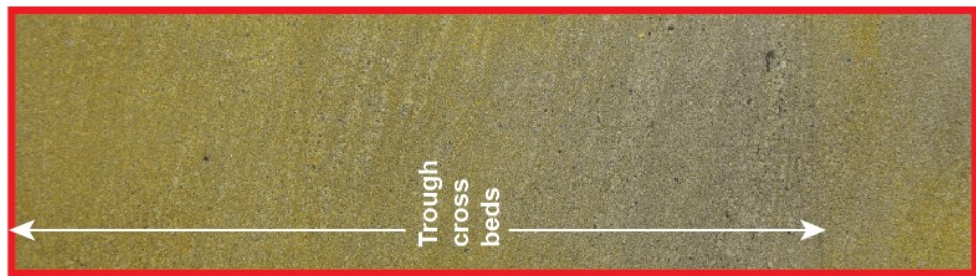
The CP1 and CP2 are generally found in vertical association with each other (Fig. 4.9a). Both comprise of blocky to fining-upwards sandstone-dominated intervals but are distinguished by their overall sand content, sedimentary structures and degree of bioturbation (Table 4.3). CP1 is usually dominated by unidirectional trough cross-bedding and rare rootlets and overlies sharp, erosional bases (Fig. 4.9a-1). CP2 also contains trough cross-bedding, but exhibits a wider array of sedimentary structures, including ripple, wavy and parallel lamination, together with a variable degree of bioturbation (Fig. 4.9a-2; Table 4.3). The basal contact of CP2 may be either erosional, sharp or gradational (Fig. 4.9a). The wireline log signature within the sandy part between these two FAs is similar. However, because CP2 is more shale-prone, its log signature is more serrated and the GR response is higher (Fig. 4.9a).

(a)



1. Fluvial (CP1)

Well: 22/12A-3
Depth: 3713 m



2. Estuarine (CP2)

Well: 22/12A-3
Depth: 3719 m



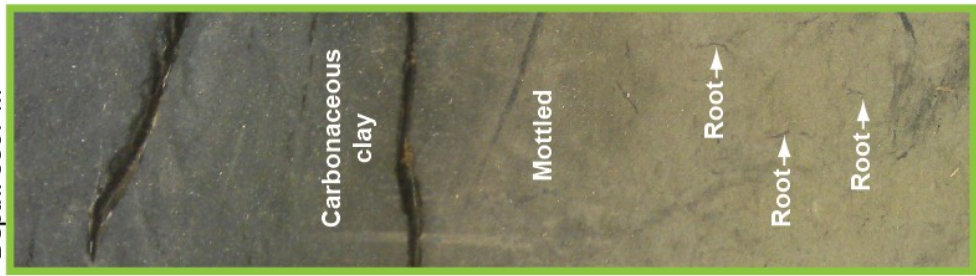
3. Lagoonal (CP3)

Well: 22/12A-3
Depth: 3728 m

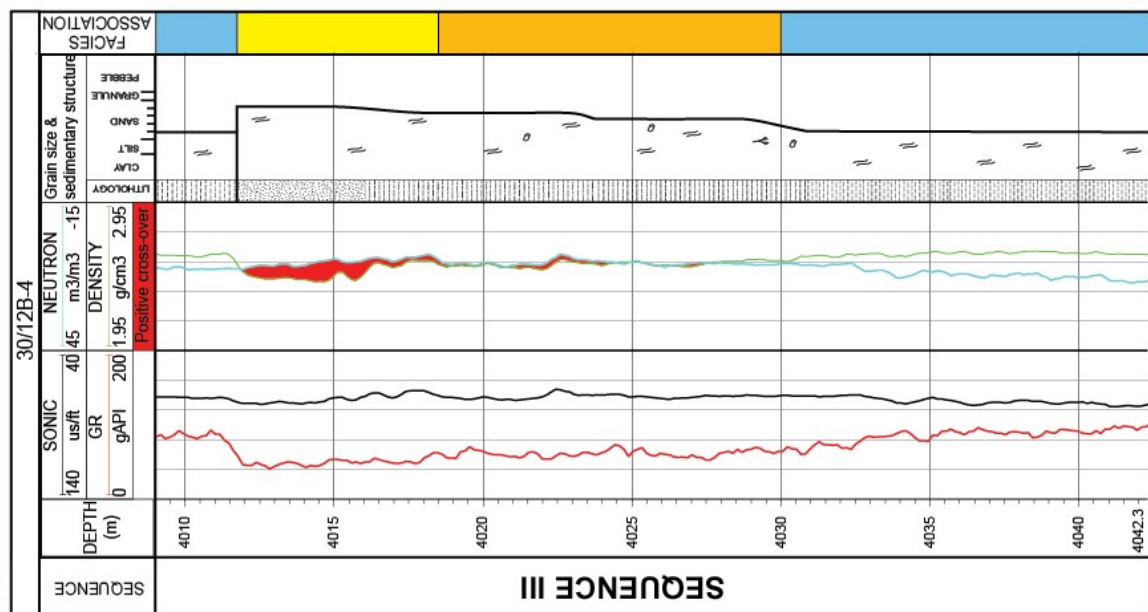


4. Flood plain (CP4)

Well: 29/7-7
Depth: 3367 m



(b)



1. Upper shoreface

(SM1)

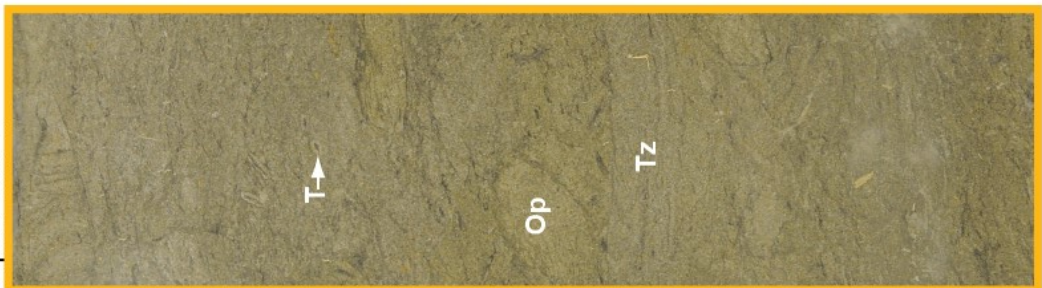
Well: 29/4A-1A
Depth: 4577 m



2. Lower shoreface

(SM2)

Well: 29/4A-1A
Depth: 4620 m



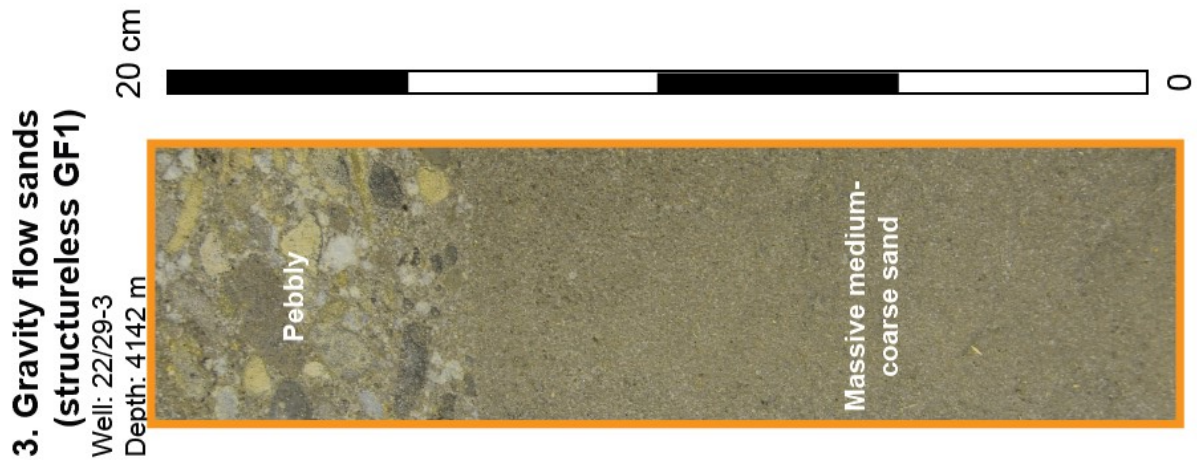
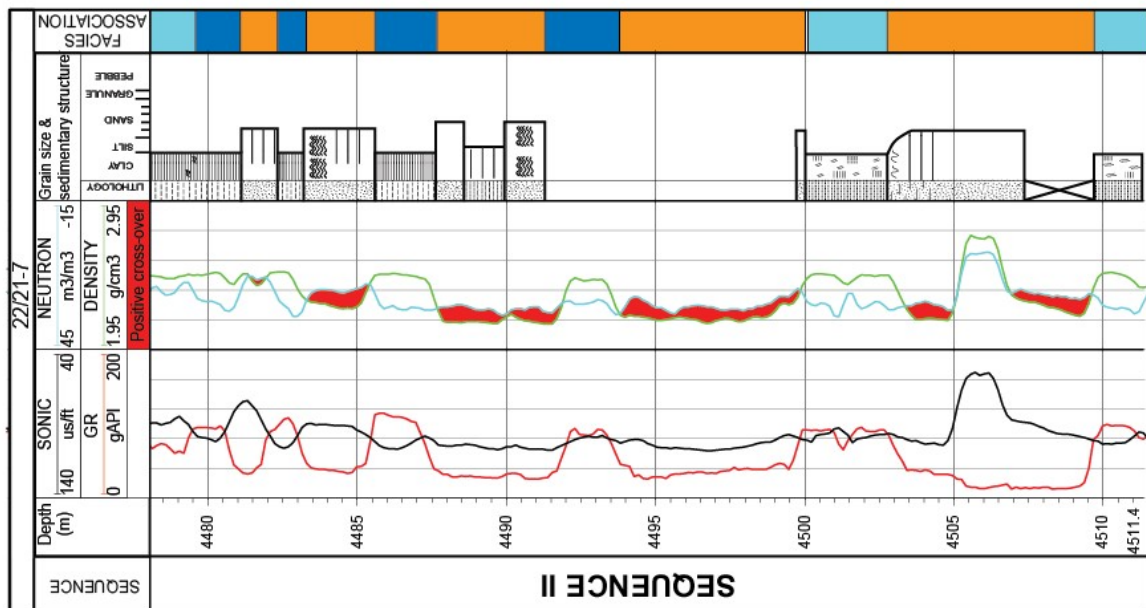
3. Offshore/shelf

(SM3)

Well: 22/23B-6
Depth: 4575 m



(c)



LEGEND FOR CORE DESCRIPTION

LITHOLOGY	SEDIMENTARY STRUCTURES		TRACE FOSSILS	
 Sandstones	 parallel lamination	 wave-ripple lamination	 Asterosoma	 Skolithos
 Silty sandstones	 cross-lamination	 HCS	 Chondrites	 Teichichnus zigzag
 Siltstones	 cross-bedding - planar	 slump structures	 Diplocraterion	 Terebellina
 Claystones	 cross-bedding - trough	 nodules	 Helminthopsis	 Thalassinoides
 Coal	 cross-bedding - herring-bone	Bed contacts:	 Ophiomorpha	 Zoophycos
 Coaly fragments	 load casts		 Phoebichnus	 Shell Debris
 Pebbles	 flaser bedding	 sharp, planar	 Planolites	 General burrow mottling
	 lenticular bedding	 sharp, irregular	 Rootlets	
		 gradational		

Fig. 4.9: Representative examples of the core-defined facies associations and their wireline log expression, shown here for SEQ I in Well 22/12A-3, which are grouped into Coastal Plain, Shallow-Marine, and Deep-Marine Gross Depositional Environments (GDE). (a) Core and well log expression of the Coastal Plain GDE (CP), which comprises four facies associations: Fluvial (CP1), Estuarine (CP2), Lagoonal (CP3) and Flood plain (CP4). Sk: Skolithos, Tz: Teichichnus Zigzag, Th: Thalassinoides. (b) Core and well log expression of the Shallow-Marine GDE (SM), shown here for which consists of three main facies associations: Upper Shoreface (SM1), Lower Shoreface (SM2) and Offshore/Shelf (SM3). Ch: Chondrites, O: Ophiomorpha, T: Terebellina. (c) Core and well log expression of the Deep-Marine GDE (DM), which comprises Deep-Marine Shale (DM1) and Gravity Flow Sands (GF1). The facies, texture and well log characteristics of the nine facies associations are summarised in Table 4.3, including a synthesis of their sedimentary processes and depositional environments.

The lack of bioturbation and the presence of roots, granules and abundant plant debris in CP1 supports a fluvial channel depositional environment. In contrast, CP2 contains an abundant but relatively low diversity of marginal marine ichnofabrics (e.g. *Ophiomorpha*, *Thalassinoides*, *Skolithos* and *Teichichnus*), which supports a marginal marine depositional environment (MacEachern and Bann, 2008; Pemberton and MacEachern, 1994; Pemberton and Wightman, 1992). The frequent vertical juxtaposition of CP2 above CP1 suggests that they are genetically-related, with CP1 representing lowstand deposits and CP2 represent the transgressive deposits. Furthermore, this succession resembles the transgressive infill of the lower reaches of an incised valley, with CP2 representing tide-influenced estuarine deposits (Dalrymple and Choi, 2007; Dalrymple et al., 1992; Zaitlin et al., 1994).

The CP3 and CP4 are more mud-rich and more heterolithic than both CP1 and CP2. There is abundant wavy- to flaser-bedding, small-scale fining or coarsening upwards beds (1-2 m thick), and occasional thin coal layers (c. 0.5 m thick; Fig. 4.9a). CP3 contains occasional shell fragments and abundant, low diversity ichnofabrics, while CP4 contains rootlets, mottled soil horizons and minor burrows (Table 4.3; Figs. 4.9a-3 & a-4). The wireline log signature for both FAs is similar, although CP3 may be recognised by the high-density spikes representing cemented shell-rich beds. The coal layers are generally too thin to be identified as low density spikes (e.g Fig. 4.9a; depth of 3734.5 m).

Both CP3 & CP4 represent deposition in a relatively low-energy, inshore to lower coastal plain setting. CP3 resembles the deposits of a restricted lagoonal/bay or estuarine environment where there was periodic reworking by fluvial and/or tidal currents. Some occurrences of CP3 could represent the lower energy parts of tidal bars and may be genetically-related to CP2, with the fining- and coarsening-upward successions representing lateral bar migration within tide-influenced channels (e.g. Dalrymple and Choi, 2007; Legler et al., 2013; Zaitlin et al., 1994). The abundant but low diversity ichnofabric supports a stressed, marginal marine environment (MacEachern and Bann, 2008; MacEachern et al., 2005). CP4 accumulated in a partly emergent floodplain environment, with minor marine influence. Tidal flats or periodically marine-inundated lower coastal plain floodplains are possible environments.

Shallow marine environment

The shallow marine environment includes three main FAs (Table 4.3; Fig. 4.9b): (1) upper shoreface (SM1), (2) lower shoreface (SM2), and (3) offshore/shelf (SM3). All three FAs contain varying degrees of marine bioturbation, ranging from intense to rare.

The vertical stacking of SM3->SM2->SM1 forms a coarsening upward facies succession, which typifies most of the shallow marine sandstone successions (Fig. 4.9b). SM1 and SM2 form the most common shallow marine sandstone reservoir in the study area (the Fulmar Formation; Johnson et al., 1986). SM1 contains medium- to coarse-grained sandstone, which is commonly structureless but occasionally trough cross-bedded. Less commonly, the cross-stratification is defined by mm-scale coarse-fine couplets, rarely with double mud layers (Fig. 4.9b-1). The paucity of mud in this system makes the detection of bioturbation difficult, while the uniform texture of the sands may obscure primary stratification. In contrast, SM2 comprises intensely bioturbated silty sandstones, where sedimentary structures were largely obliterated by burrowing infauna (Fig. 4.9b-2). Remnants of primary sedimentary structures include ripple cross-lamination and low-angle to parallel lamination, invariably detected by mm-scale muddy to silty laminae. Cemented, sharp-based, coarser grained and/or shell-rich beds may occur interbedded within both SM1 and SM2, with the highest frequency and coarsest grain size (up to granule grade) in SM1. SM3 is dominated by mudstone with intercalated siltstone and very fine sandstone laminae. Both SM2 & SM3 are largely destratified by intense bioturbation, comprising a diverse array of ichnofossils (*Chondrites*, *Teichichnus*, *Zoophycos* and *Terebellina*). The wireline log signature of a complete vertical facies succession (SM3->SM2->SM1) comprises coarsening upward patterns, which are most completely developed in high-accommodation space settings (e.g. within the main graben and basinal areas; Fig. 4.9b). Sharp or erosional contacts are observed in low accommodation space settings (e.g. platform areas; Fig. 4.10). Fining upward facies successions (SM1->SM2->SM3), are also common, especially at the top of the underlying progradational successions. The complete coarsening upward facies successions are always characterised by an increase in porosity and permeability (Table 4.3), confirming the dominant primary textural control on reservoir quality.

The high sand content, textural maturity and ichnofabrics are consistent with a shallow marine environment of deposition. The coarsening upward facies successions are widely interpreted as the result of shoreface progradation (Gowland, 1996; Howell et al., 1996; Wonham et al., 2014). The trough cross-bedding within SM1 suggests migrating dune bedforms above fairweather wave base. The mm-scale laminae defined by mud-drapes may indicate tidal currents and/or proximity to tide-influenced river mouths and estuaries (Dalrymple and Choi, 2007). The structureless nature of SM1 could be caused by any of the following processes: dewatering, rapid deposition of texturally uniform sand or by complete bioturbation. The occurrence of sharp-based sandstone beds showing waning flow characteristics and the pervasive and high intensity bioturbation are supportive of episodic sand transport, most likely by storm processes (Gowland, 1996; Howell et al., 1996), possibly enhanced by river floods in river mouth settings.

SM2 is believed to have been deposited below the fair weather wave base in a lower shoreface setting, where marine fauna was able to flourish during fairweather periods in-between storms. SM3 is interpreted to have been deposited in an offshore environment within a relatively quiet but well-oxygenated shelf setting that was favourable for benthic infauna. Sharp-based shoreface successions display abrupt basal contacts with underlying offshore mudstone facies and are preferentially developed in platform areas (e.g. Western Platform; Fig. 4.10a). Fining upward facies successions record gradual deepening from shoreface to shelf environments, with relative sea-level rising faster than sediment supply (e.g. Fig. 4.10a and b). This is commonly developed at the top of progradational successions in areas of high-accommodation space creation, both in basinal and platform areas, such as on the Kittiwake-Gannet Platform (Donovan et al., 1993). Sharp tops to the progradational successions reflect a combination of rapid deepening, as developed in the Fulmar Field (Johnson et al., 1986). Elsewhere, sharp upper contacts may also reflect wave ravinement, with

the latter capable of generating thin, reworked pebbly lag deposits, as seen in the Fife and Flora fields (Currie et al., 1999; Hayward et al., 2003).

Deep marine environment

This environment comprises of two FAs (Table 3; Fig. 4.9c): (1) deep marine shale (DM1), and (2) gravity flows (GF1). Both of these FAs conspicuously lack both the bioturbation intensity of the shallow marine environment and the abundance of coaly debris of the coastal plain environment.

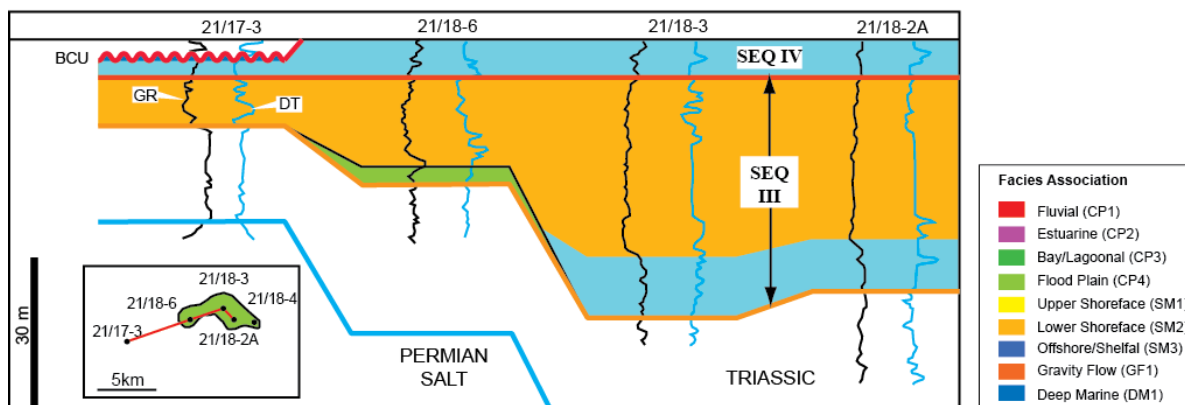
DM1 contains mostly organic-rich shales interbedded with minor siltstones layers that are usually devoid of any other internal sedimentary structures apart from thin parallel laminations (Fig. 4.9c-1). Bioturbation is non-existence. The gamma-ray logs of the DM1 often resembles that of the SM2 but due to its distinctively low velocity, the sonic log reading is often high, to the point of giving a cross-over effect when displayed together with gamma-ray log (Fig. 4.9c). DM1 is interpreted to be deposited in a deep marine, anoxic environment where there is an absence of benthic infauna and high preservation potential of organic material.

In contrast to DM1, GF1 mainly comprises of medium grained sandstones with rare pebble layers (Fig. 4.9c-3). The sandstones are generally massive except for occasional wavy and parallel lamination (Fig. 4.9c-2 & c-3). Dewatering structures are also occasionally present. GF1 sandstones are invariably found in vertical association with DM1 mudstones. However, GF1 sandstones are also occasionally found interbedded with the more distal shallow marine FAs (SM3 intervals; e.g Well 30/6-3 in Fig. 4.5). Based on lack of bioturbation, textural maturity and evidence of rapid emplacement, GF1 is interpreted to be the product of gravity flow processes within the moderately deep, mud-dominated basinal setting of DM1.

Sequence stratigraphic analysis

The facies associations have been calibrated with their well log response and integrated into the stratigraphic framework outlined earlier. The biostratigraphic data were utilised in constraining the key MFS and tying these into the well log and seismic datasets. The resulting integrated sequence stratigraphic analysis, incorporating well log correlation and seismic interpretation with the genetic stratigraphy, is presented in the form of a series of representative seismic (Fig. 4.11) and well correlation panels (Figs. 4.12, 4.13, 4.14, 4.15 & 4.16). These panels trend NE-SW, across the structural grain of the basin (see the separate inset maps on each figure), and capture the internal stratigraphic architecture of the main structural domains defined earlier. However, the relatively low vertical resolution of the 3D seismic data (e.g Figs. 4.11b, c), has restricted the seismic stratigraphic analysis to the identification of two internal MFS, which has enabled the construction of three isochore maps: (1) Lower to Middle Jurassic (pre-Sequence I; Fig. 4.17), (2) Oxfordian (Sequences I to II; Fig. 4.18a), and (3) Kimmeridgian-Tithonian (Sequences III to VI; Fig. 4.18b). The sequence stratigraphic analysis of these three intervals is discussed in the following sections.

(a) Kittiwake-Gannet Platform



(b) Curlew Platform

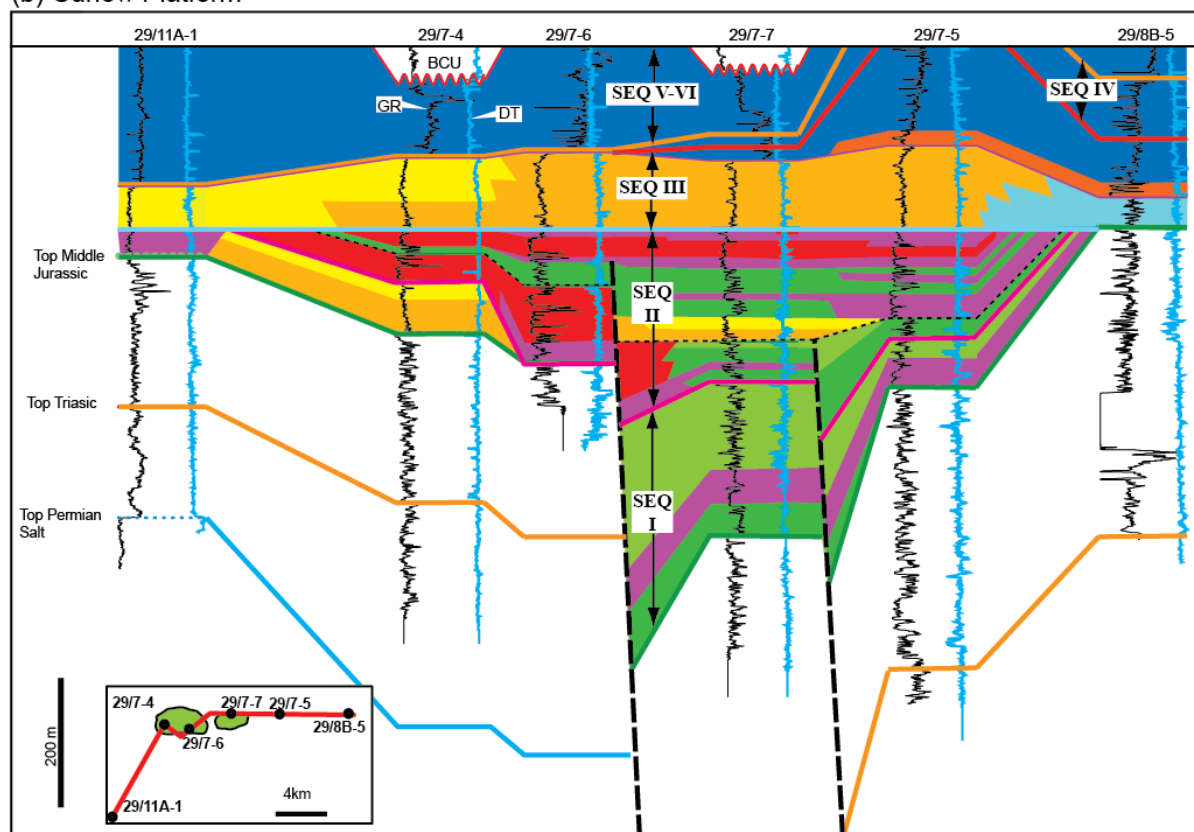


Fig. 4.10: Previous page - The vertical and lateral relationships of the facies associations illustrated in well log correlation panels from two contrasting structural domains: (a) Kittiwake-Gannet Platform (Kittiwake Field), datumed on the J63 MFS (Top Seq III) (after Stewart et al, 1999). (b) Curlew Platform complex, datumed at J62 level (Top Seq II). The GR log (black/left) and sonic log (blue/right) are shown for each well. The seismic expression of the Kittiwake-Gannet Platform section and the Curlew Platform section are shown in Fig. 4.11a and c, respectively.

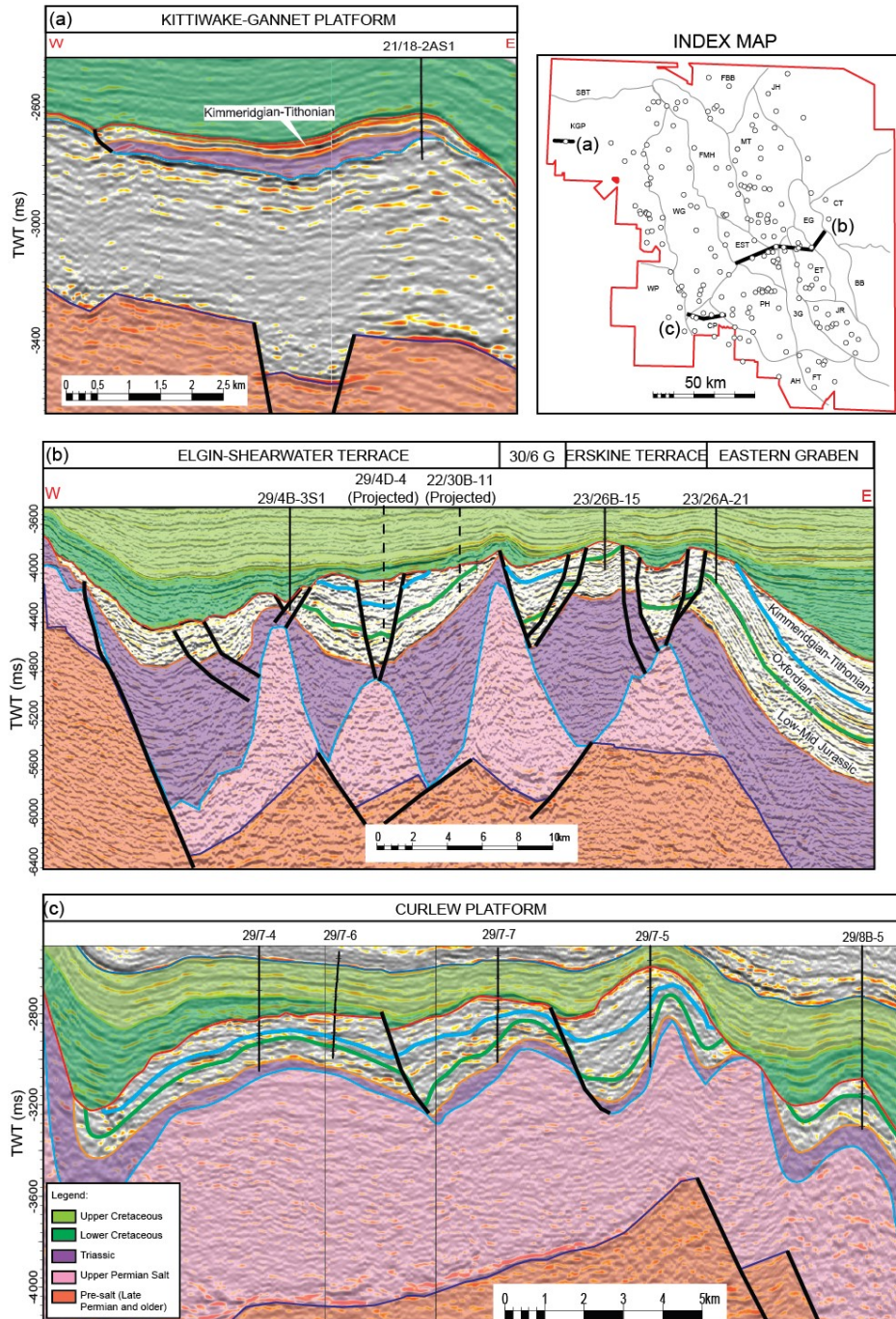


Fig. 4.11: Representative sections showing the 3D seismic expression of three contrasting structural domains: (a) E-W section through part of the Kittiwake-Gannet Platform domain (across the Kittiwake Fi-eld), where the Jurassic interval is thin (10-60 m) and the seismic expression is strongly attenuated. (b) WSW-ENE section through the Western Graben domain (Elgin-Shearwater and Erskine terraces), where the Middle Jurassic, lower Upper Jurassic and Triassic sections are thicker and where both syn-rift faulting and salt movement influenced Upper Jurassic accommodation space creation. (c) E-W section across part of the Curlew Platform domain where the Oxfordian section thickens towards the supra-salt faults (note also the thick Permian salt interval). The faulted pre-salt (Rotliegend) interval is visible in all three sections. The index map shows structural domain boundary. For key to abbreviations, see Fig. 4.2.

Lower to Middle Jurassic interval

The Lower to Middle Jurassic interval is incompletely preserved throughout the study area, where it unconformably overlies the Top Triassic surface (e.g in Western Graben; Fig. 4.6). The top of the Lower Jurassic is not recognisable within the seismic dataset but the top of the Middle Jurassic is usually coincident with a high amplitude and continuous seismic reflector, which defines the upper boundary of this interval (Figs. 4.6 & 4.11b, c). The Lower to Middle Jurassic interval contains high amplitude and continuous seismic characters, especially where the interval is thicker than 100 m but otherwise it merges with the underlying Triassic reflectors into a bright and continuous amplitude event (Figs. 4.6, & 4.11b, c).

Lithologically, the Lower to Middle Jurassic interval comprises interbedded sandstones, claystones and coals (Figs. 4.5, 4.12, 4.13, 4.14, 4.15 & 4.16). Volcanics are also locally developed, such as in the Puffin High and the Auk High (e.g Well 29/4A-1A in Fig. 4.15). The gross depositional environment is interpreted to be coastal plain based on the occurrence of *in situ* coal layers, abundance of disseminated land plant material, the absence of marine fauna and limited bioturbation (Husmo et al., 2003). The volcanics in the Puffin High have been dated as Aalenian, with their occurrence probably related to the Mid-Jurassic thermal doming (Smith and Ritchie, 1993). However, the thickest sections, especially in the Eastern Graben, have not been penetrated by wells, so the depositional environment cannot be confirmed there.

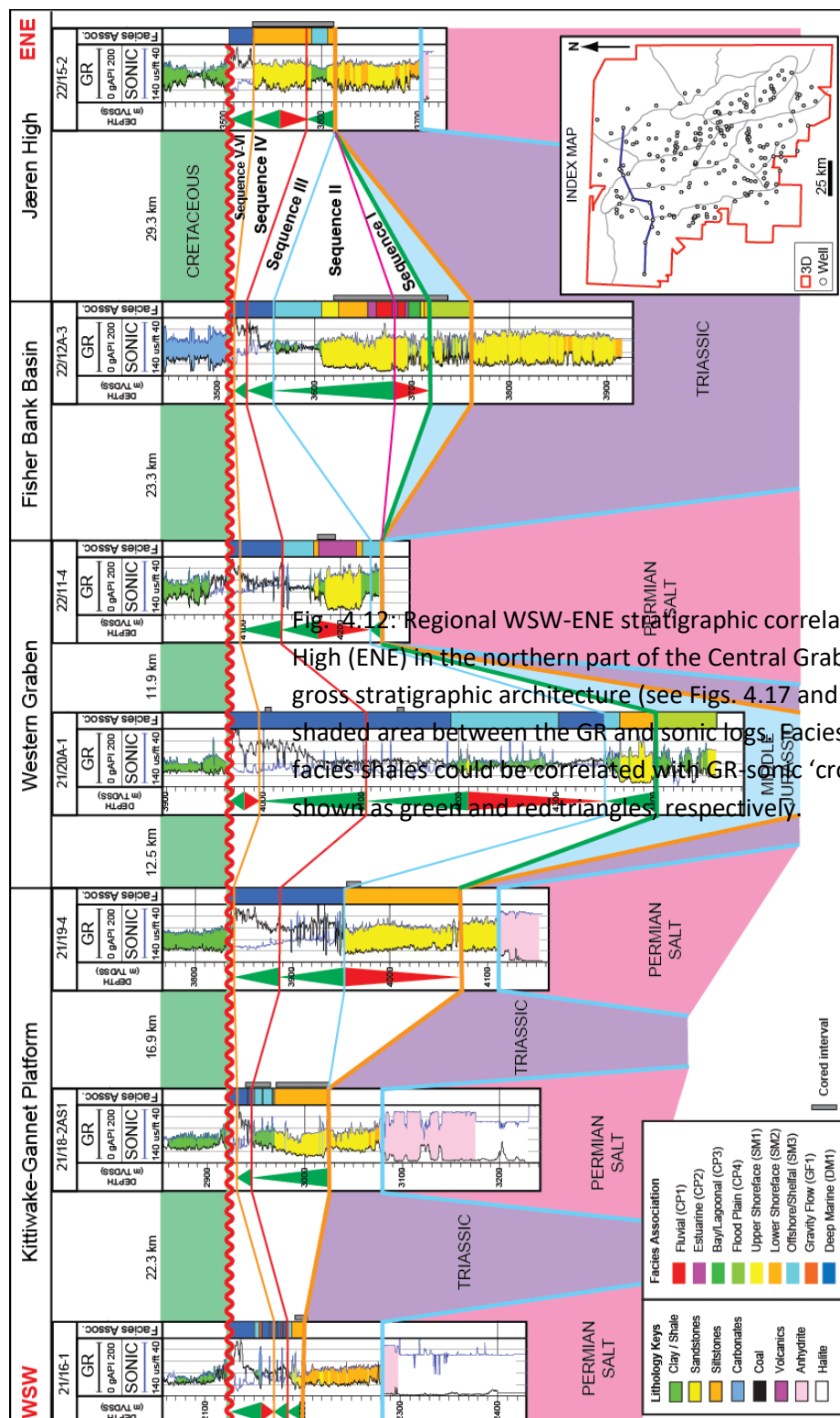


Fig. 4.12: Regional WSW-ENE stratigraphic correlation panel running from the Kittiwake-Gannet Platform (WSW) in the southern part of the Central Graben to the Jæren High (ENE) in the northern part of the Central Graben. Note the relationship between the major sequences and the gross stratigraphic architecture (see Figs. 4.17 and 4.18 for isochore expression). The lithology and facies associations are displayed to the right of the well logs. The shaded area between the GR and sonic logs indicates the facies 'cross-over'. Interpreted progradational and aggradational sequences are shown as green and red triangles, respectively.

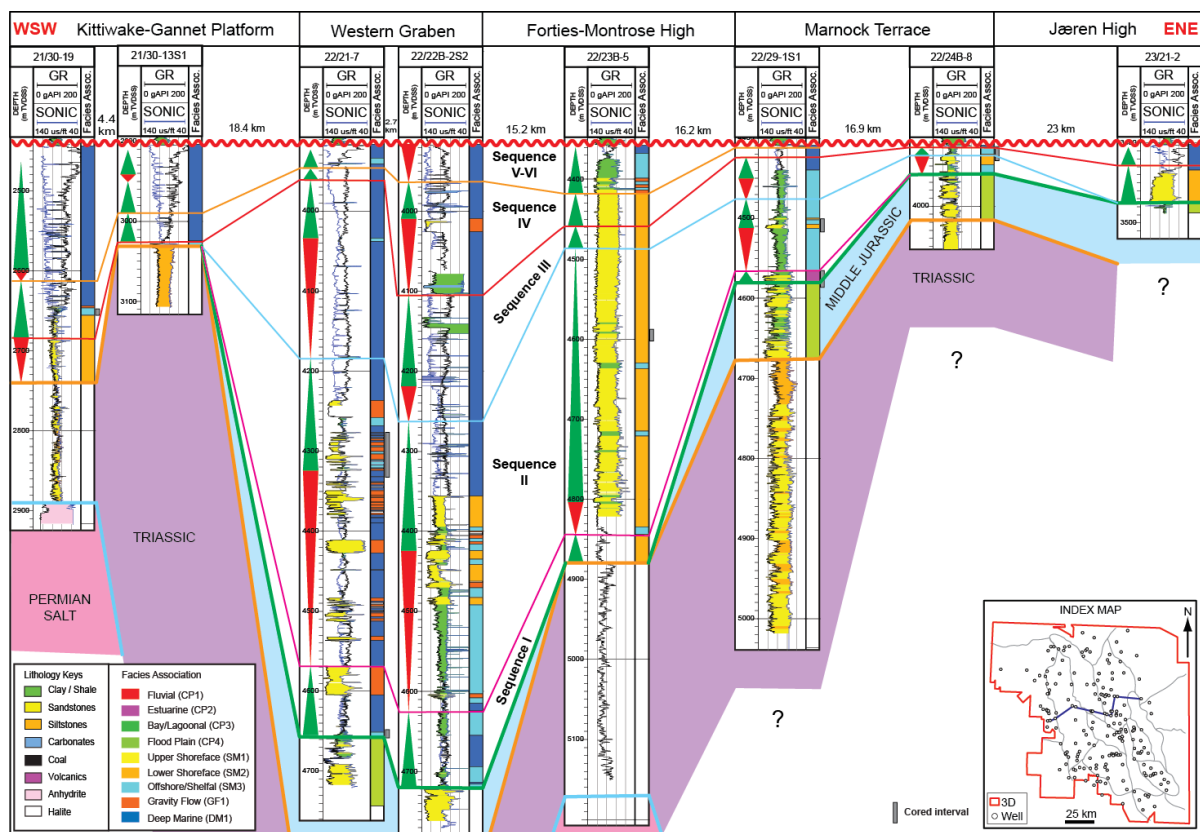


Fig. 4.13: Regional WSW-ENE stratigraphic correlation panel running from the Kittiwake-Gannet Platform (WSW) to the Jæren High (ENE) in the middle part of the Central Graben. The Upper Jurassic interval is considerably thicker within the Western Graben compared with the more northerly section in Fig. 12. The Forties-Montrose High is virtually absent at this location (see Figs. 17 and 18 for isochore expression). The lithology interpretation is shown in the shaded area between the GR and sonic logs. Facies association is displayed to the right of the logs. Note that deep marine facies shales could be correlated with GR-sonic 'cross-over'. Interpreted progradational and retrogradational successions are shown as green and red triangles, respectively.

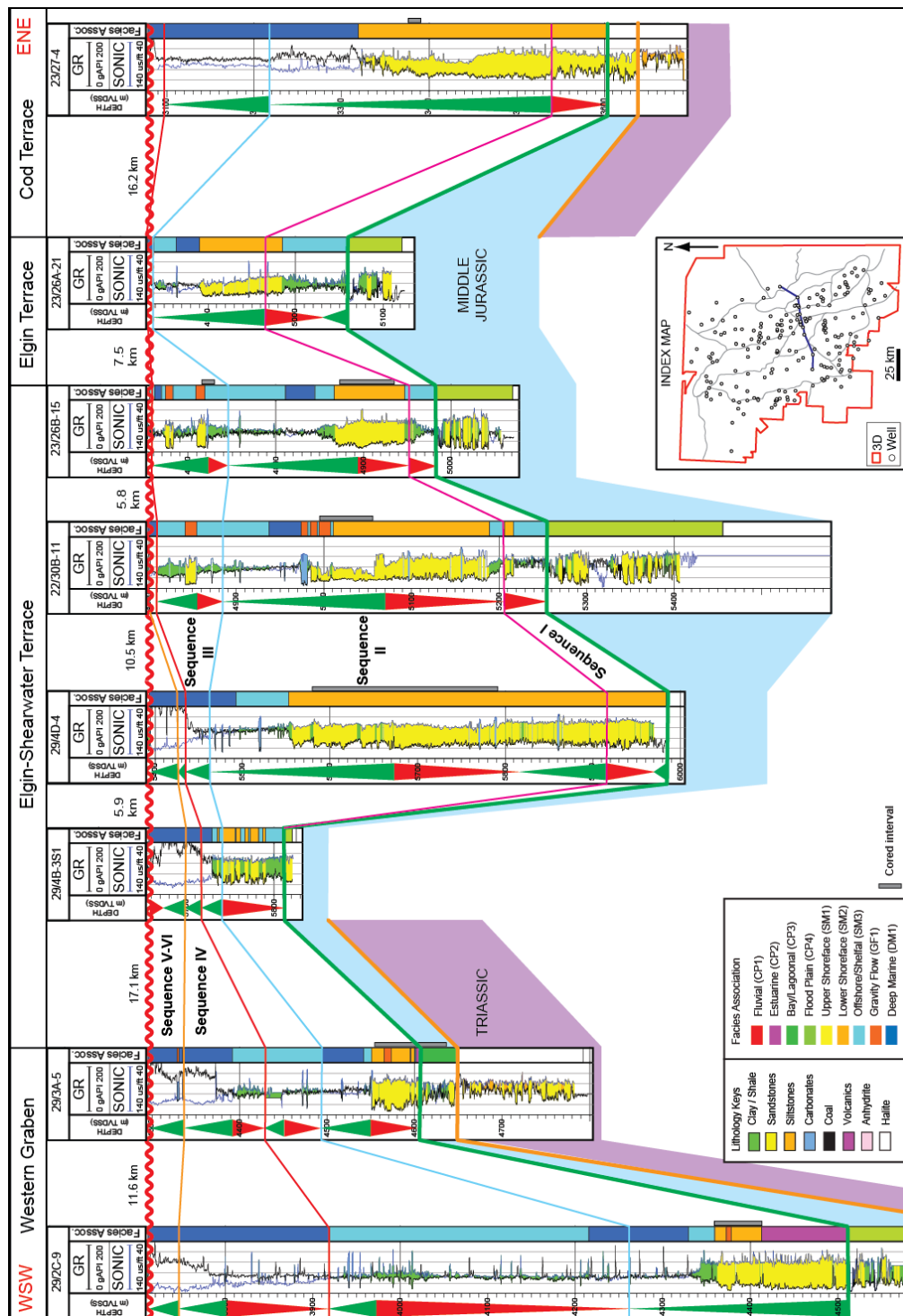


Fig. 4.14: Regional WSW-ENE stratigraphic correlation panel running from the Western Graben (WSW), through the Elgin-Shearwater Terrace and across to the Cod Terrace (ENE). The Upper Jurassic interval is considerably thinner within the Western Graben compared with the previous, more northerly section in Fig. 4.13 (see Figs. 4.17 and 4.18 for isochore). The lithology interpretation is shown in the shaded area between the GR and sonic logs. Facies association is displayed to the right of the logs. Note that deep marine facies shales could be correlated with GR-sonic 'cross-over'. Interpreted progradational and retrogradational successions are shown as green and red triangles, respectively.

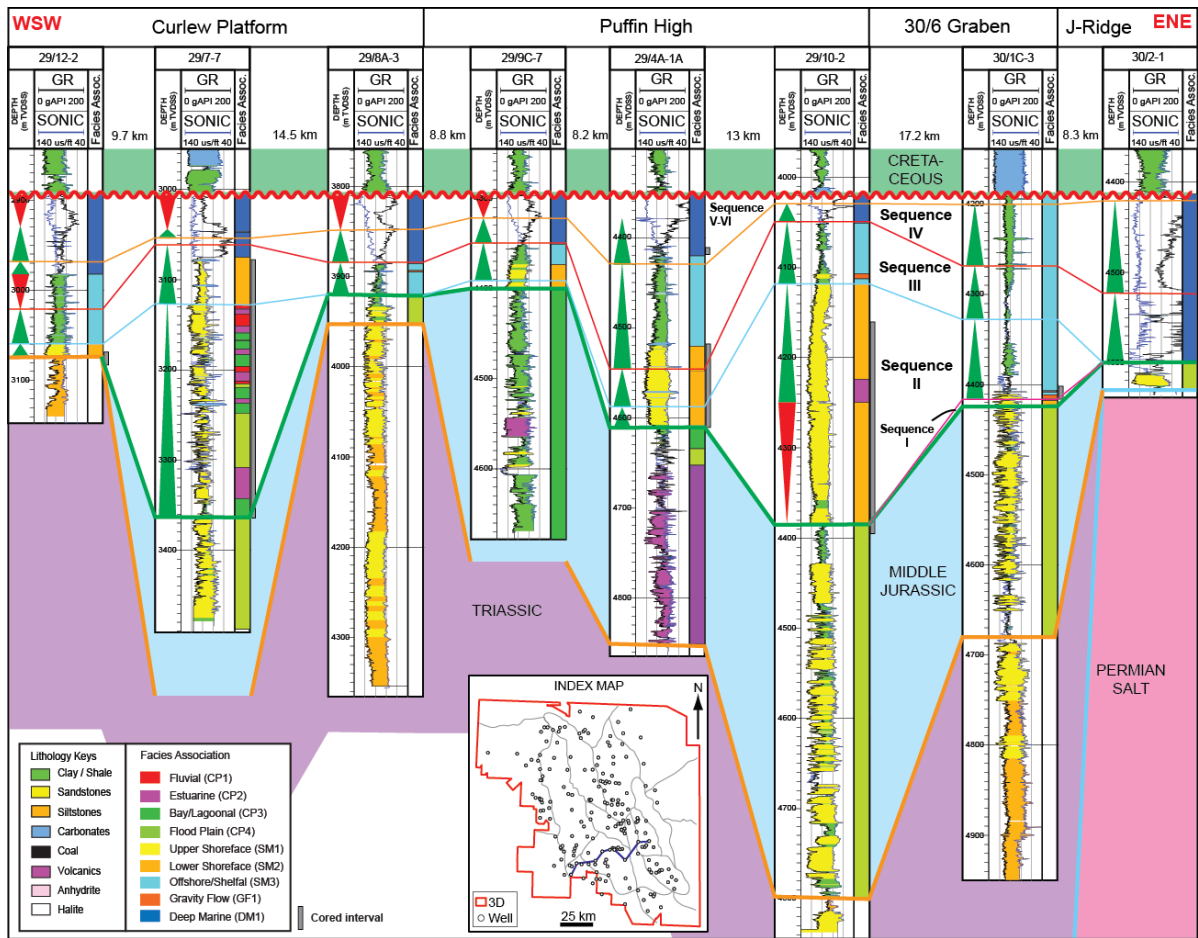


Fig. 4.15: Regional WSW-ENE stratigraphic correlation panel running from the Western Platform/Curlew Terrace (WSW), through the Puffin High and across to the Breiflabb Basin (ENE) in the southern part of the Central Graben. The Upper Jurassic interval shows considerably more thickness variability compared with the middle sections in Figs. 4.13 and 4.14 (see Figs. 4.17 and 4.18 for isochore expression). The GR and sonic logs are displayed to the left and right of the facies association column, respectively. The lithology interpretation is shown in the shaded area adjacent to both the GR and sonic logs. Interpreted progradational and retrogradational successions are shown as green and red triangles, respectively.

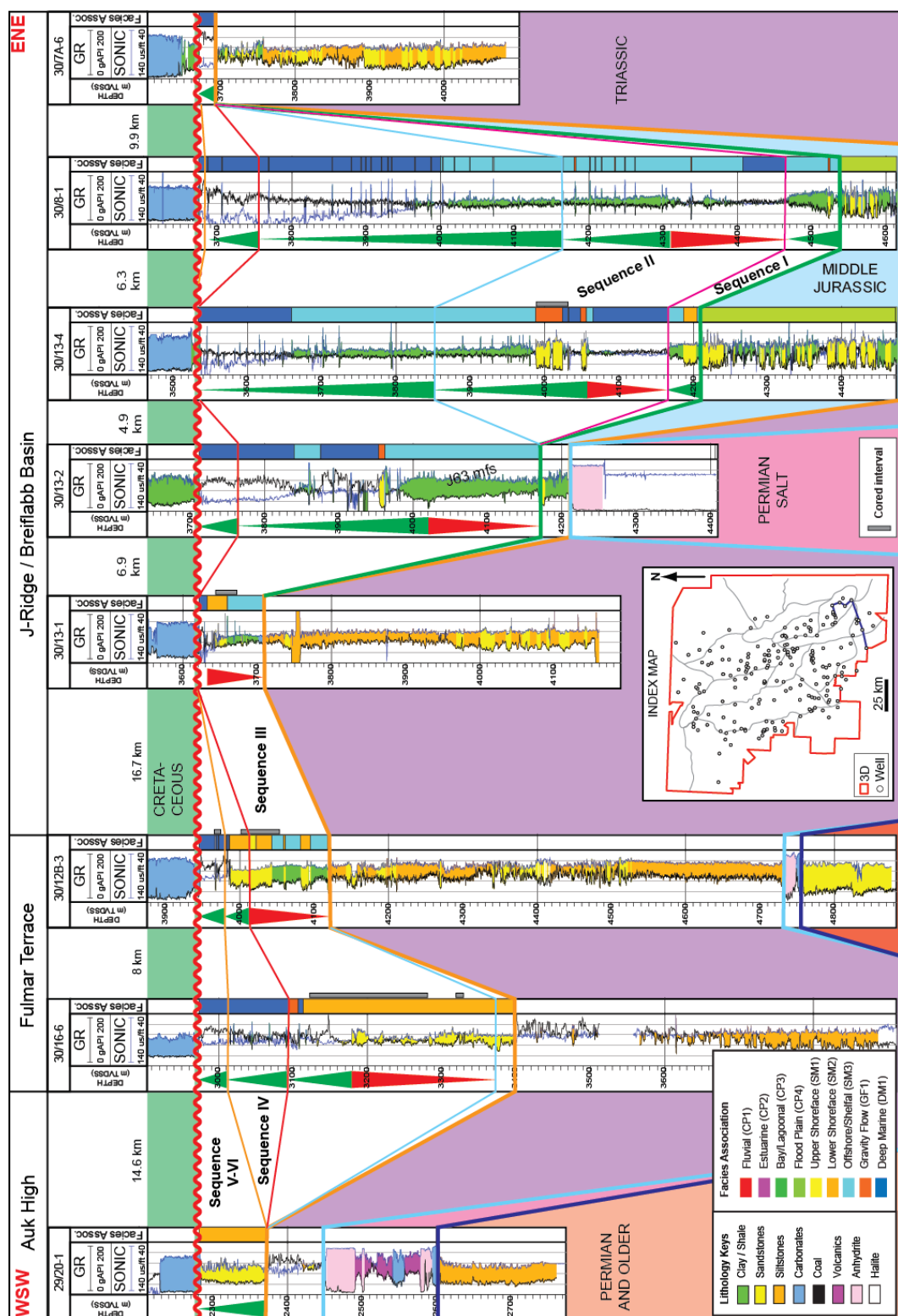


Fig. 4.16: Regional WSW-ENE stratigraphic correlation panel running from the Auk High (WSW), through the Fulmar Terrace and onto the J-Ridge (ENE) in the most southern part of the Central Graben (see Figs. 4.17 and 4.18 for isochore expression). The GR and sonic logs are displayed to the left and right of the facies association column, respectively. The lithology interpretation is shown in the shaded area adjacent to both the GR and sonic logs. Interpreted progradational and retrogradational successions are shown as green and red triangles, respectively.

The Lower to Middle Jurassic interval generally thickens eastwards towards the Eastern Graben but occurs in three main areas (Fig. 4.17): (1) western area (Western Graben), (2) eastern area (Eastern & 30/6 grabens and the Breiflabbb Basin), and (3) northern area (the Fisher Bank Basin). A northerly-striking ridge, which runs from the Forties-Montrose High in the north to the Auk High in the south, separates the western and eastern areas. The northern area is separated from the rest of the basin by the Jæren High. The Lower to Middle Jurassic interval is absent over most of the highs as a consequence of either non-deposition or erosion during the Mid-Jurassic thermal doming and/or Late Jurassic rifting (Husmo et al., 2003).

Distinctively, the western and northern areas are relatively flat (same isochore) and generally comprise of successions less than 200 m thick, whereas the eastern area is more variable with thicknesses ranging from less than 100 m to more than 1000 m (Fig. 4.17). The latter has also been compartmentalised into smaller, NW-striking sub-basins (Eastern Graben, Breiflabbb Basin, Erkinne Terrace, Elgin-Shearwater Terrace and 30/6 Graben). This structural pattern is similar to that seen at the Triassic level (Fig. 7a) but the thickness variation at least partly reflects post-depositional erosion of the Lower to Middle Jurassic interval. This is exemplified by the truncation of parallel seismic reflectors by the BCU in the Elgin-Shearwater Terrace and Eastern Graben areas (Fig. 4.11b). This erosional truncation usually coincides with salt walls, which suggests that the truncation occurs due to the re-growth of the underlying salt diapirs during the Late Jurassic rifting period (Fig. 4.19a).

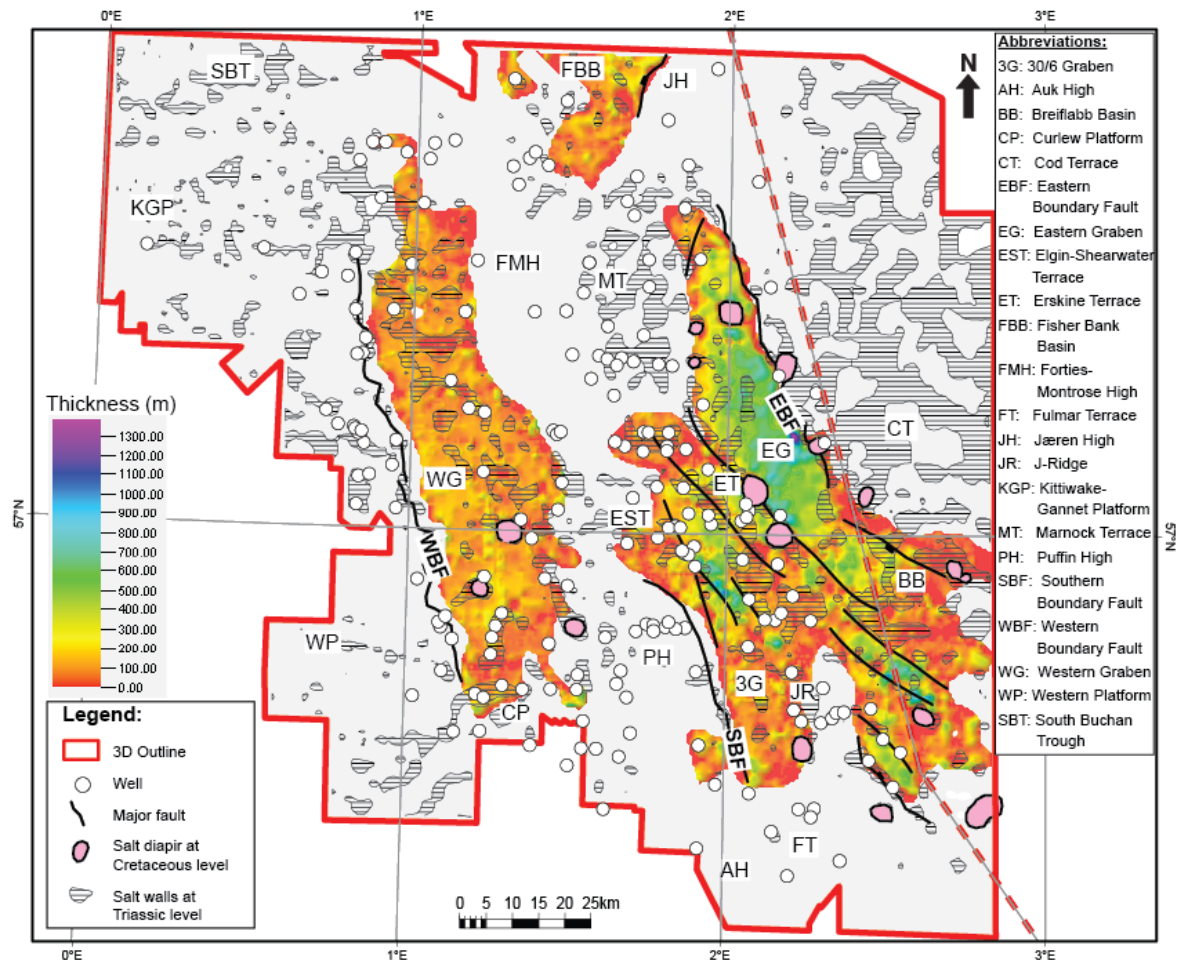


Fig. 4.17: Isochore map of the Lower to Middle Jurassic showing an overall eastward thickening from the Western to Eastern Grabens. The dominant NNE-SSW thickness pattern reflects an overprint of the Upper Jurassic graben system. However, there is a hint of NW-SE structural trend, especially in the thicker eastern compartment, which is more likely to be of Lower-Middle Jurassic age. The relationship with the distribution of salt walls distribution in the underlying Triassic is also shown.

Oxfordian interval (Sequences I & II)

The Oxfordian interval is represented by Sequences I and II (Fig. 4.4). In the seismic data, this interval conformably overlies the Lower to Middle Jurassic succession and seems to share similar high amplitude and continuous internal reflections with the older sediments (Figs. 4.6 & 4.11b, c). However, the top of the Oxfordian is poorly defined in term of amplitude intensity but can be identified by onlap onto this surface (Figs. 4.6 & 4.11b). However, this poor seismic definition is

complimented by tracing the Top Sequence II biostratigraphic pick (J63 MFS; Fig. 4.4) and is further constrained by well log data (Figs. 4.5, 4.12, 4.13, 4.14, 4.15 & 4.16).

The Oxfordian succession comprises mainly sandstones and shales, which are predominantly shallow marine in origin (e.g Fig. 4.9b), but with locally preserved coastal plain deposits, such as in well 22/12A-3 in the Fisher Bank basin (Fig. 4.9b & 4.12) and in well 29/7-7 in the Curlew Platform (Fig. 4.15). Both of these examples, however, also contain marine burrows and are interpreted as marine-influenced lower coastal plain deposits, most probably marking the onset of marine influence in the CG at the start of Oxfordian. In addition, this inferred Oxfordian interval also contains occasional gravity flow sediments, especially within the thicker part of the main depocentres, such as in wells 22/21-7 in the Western Graben (Fig. 4.13), 22/30B-11 in the Elgin-Shearwater Terrace (Fig. 4.14) and 30/13-4 in the Breiflabbb Basin (Fig. 4.16).

Generally, the Oxfordian succession thickens eastwards, similar to that of the underlying Lower to Middle Jurassic interval (Fig. 4.18a). However, unlike the older interval, the Oxfordian succession in the Western Graben also thickens slightly westwards against the Western Boundary Fault and extends into the Elgin-Shearwater Terrace, which forms a connection between the eastern and western areas (Fig. 4.18a). In addition, the Oxfordian succession is also laterally more extensive in the Breiflabbb Basin and continues eastwards into the Feda Graben in the Norwegian Central Graben (Fig. 4.1). The succession also thickens further south in the 30/6 Graben where the Oxfordian interval is up to 1 km thick (Fig. 4.18a).

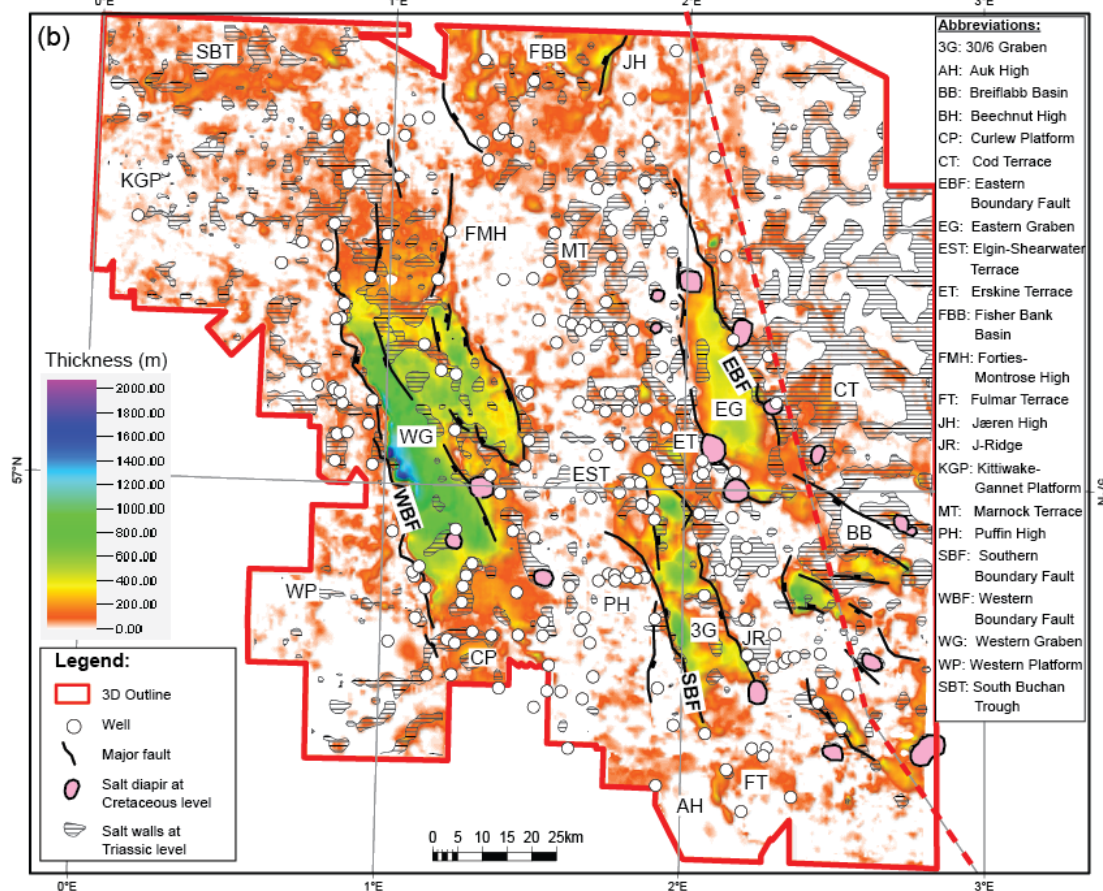
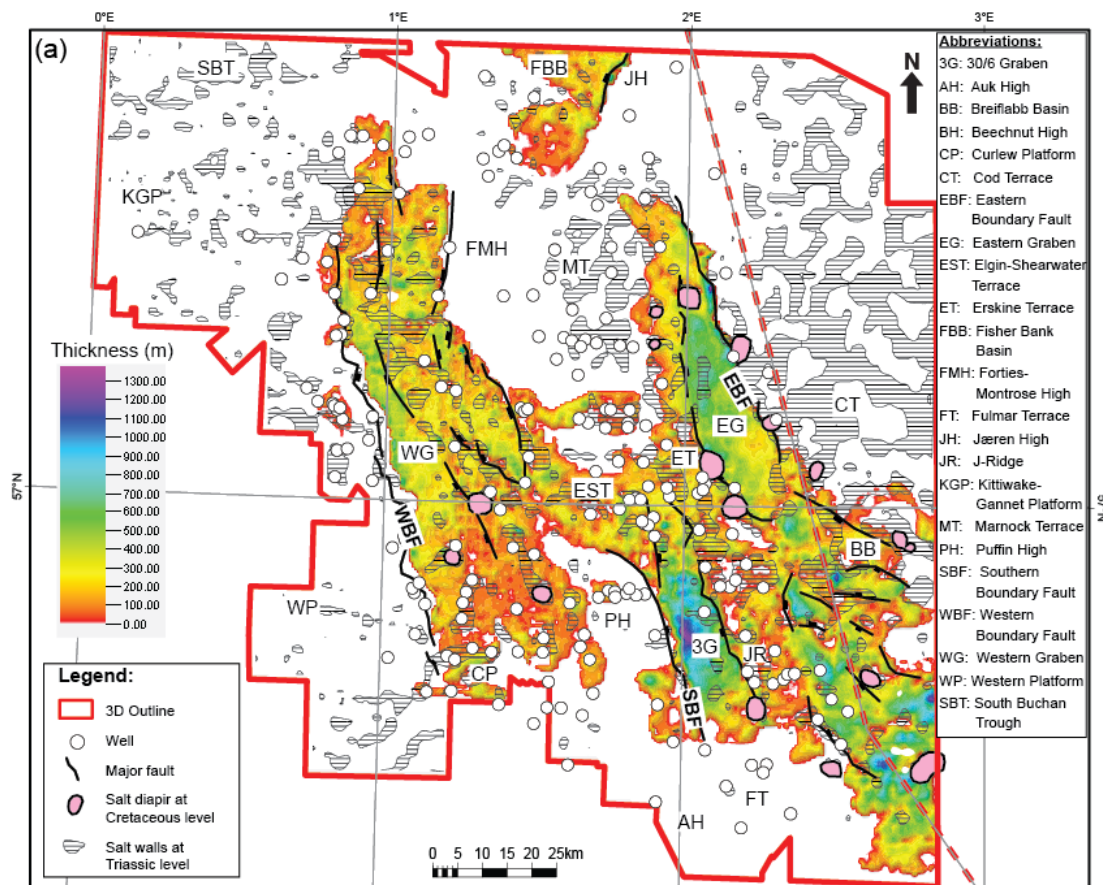


Fig. 4.18: Previous page - Isochore maps of the Upper Jurassic: (a) Oxfordian interval. (b) Kimmeridgian-Tithonian interval. Salt walls are overlain to show the overall influence to the thickness pattern.

The Oxfordian has an overall NNW-striking structural pattern that closely resembles the overall Jurassic structural trend (Figs. 4.7b & 4.18a). However, the thick Oxfordian succession relative to the Kimmeridgian-Tithonian sediment in a NW-striking depocentre, such as the Elgin-Shearwater Terrace (Figs. 4.11b & 4.14), indicates inheritance of Triassic and Lower Jurassic NW-striking depocentres (Figs. 4.7a & 4.17). This is different from the previous interpretation that the rift has formed NNW-striking grabens before changing to NW in the latest Jurassic (Erratt et al., 1999; Zanella and Coward, 2003). The NNW-striking grabens, in the current case, formed later in the Late Oxfordian or Kimmeridgian as the rifting accelerated, thus overprinting the NW-striking structures.

The successions above salt diapirs, apart from being truncated by the BCU, also appear to show convergence and thinning (e.g. Fig. 4.11c). In this situation, the salt diapir growth is interpreted to be syn-depositional with the Oxfordian sediment (Fig. 4.19b). The growth of salt in the flanks of the major grabens might be triggered by the commencement of rifting where salt in the grabens could be subjected to rift extension, as well as loading from the accumulation of syn-rift sediments, resulting in the collapse and withdrawal to the flanks.

Kimmeridgian-Tithonian interval (Sequences III to VI)

The Kimmeridgian-Tithonian interval comprises of Sequence III to VI (Fig. 4.4). The seismic expression of the base of this interval shows onlap onto the Top Oxfordian surface, especially near the flanks of horst blocks, but becomes more conformable within the adjacent grabens (Fig. 4.11b). The top is defined by the basin-wide Base Cretaceous Unconformity (BCU). The internal seismic

reflections show moderate to high amplitudes with high continuity, albeit with less continuity than the underlying Oxfordian interval (Fig. 4.11b).

In the thicker, basinal areas, the Kimmeridgian-Tithonian section is dominated by black, organic-rich shales, which contain pelagic fauna (e.g. ammonites). Thin, structureless sandstone layers also appear intermittently within the black shale (e.g. well 22/22B-2S2 in Fig. 4.13), which lack bioturbation (Fig. 4.9c-3). In contrast, bioturbated sandstones and shales occur around the edges of the main depocentres (e.g. well 21/16-1 & 21/18-2AS1 in Fig. 4.12; well 21/30-19 in Fig. 4.13).

The black shales accumulated within an anoxic, deep-marine environment (Howell et al., 1996; Stow and Atkins, 1987). However, the presence of marine bioturbated sandstones and shales indicate that contemporaneous, shallow-marine, oxygenated conditions occurred around the periphery of the main depocentres, most likely representing shelf to shoreface environments.

The Kimmeridgian-Tithonian interval contains the most widespread Upper Jurassic sediments in the Central North Sea, although thickness patterns reflect strong structural compartmentalisation (Fig. 4.18b): successions developed on platform highs are usually less than 100 m thick (e.g. Kittiwake-Garnet Platform; Fig. 4.18b), whereas the main NNW-trending graben depocentres exceed 1000 m in thickness (e.g. Western, Eastern and 30/6 grabens; Fig. 4.18b). The Western Graben is a particularly prominent half-graben, with the Kimmeridgian-Tithonian interval thickening westwards against the Western Boundary Fault (Figs. 4.18b). The Eastern and 30/6 grabens also show strong compartmentalisation between the graben and adjacent platform areas.

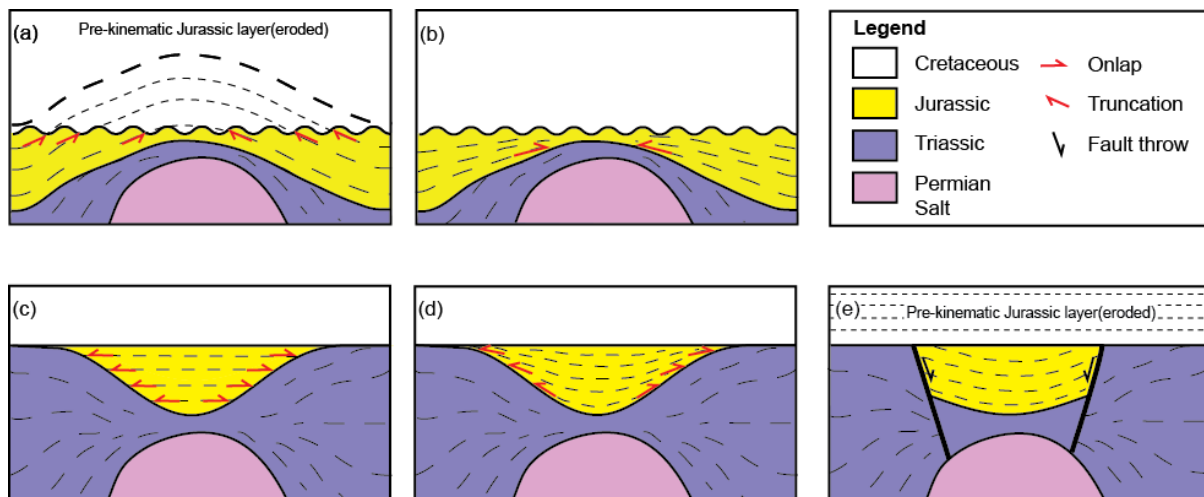


Fig. 4.19: Conceptual deformation styles of Jurassic sediments above salt walls in response to halokinesis of the underlying salt wall. (a) Post-depositional salt growth resulted in the erosion of sediments above the salt wall. (b) Syn-depositional salt growth resulted in the thinning of sediment above the salt wall. (c) Pre-depositional salt withdrawal, where the withdrawal happened during the Triassic and the space created was only filled during Late Jurassic period. (d) Syn-depositional salt withdrawal, where the withdrawal was concurrence with the deposition of Upper Jurassic. (e) Post-depositional salt withdrawal, where the Upper Jurassic sediments initially blanketed the area and then part of the sediments above the salt wall collapsed onto the withdrawing salt. The remaining sediments above the Triassic pods were then eroded during the Early Cretaceous period.

The Kimmeridgian-Tithonian successions on platform areas are more affected by the underlying salt than their basinal equivalents. Thicker successions on the platform areas coincide with underlying networks of salt walls, especially in the Kittiwake-Gannet Platform, Montrose High, Jæren High and Cod Terrace (Figs. 4.11a & 4.18b). The genesis of the Upper Jurassic accommodation space above the salt walls, often termed as interpods (Hodgson et al., 1992; Smith et al., 1993), is still subject to debate, in part due to the poor seismic imaging and ambiguous well data (Fig. 4.11a). One key question is whether the interpods were formed pre-, syn-, or post-depositional, relative to the Kimmeridgian-Tithonian successions (Fig. 4.19c, d, e). The pre-depositional collapse or shrinkage of the salt walls has been popular due to the valley-like shape of the accommodation space as well as the apparent limitation and parallel nature of the Upper Jurassic sediment within the interpods (Clark et al., 1999; Stewart et al., 1999; Wakefield et al., 1993) However, syn-depositional salt

withdrawal model has also been proposed, citing the presence of onlap onto the flanks as well as diverging character of the seismic reflection patterns towards the centre of the interpods (Hodgson et al., 1992; Mannie et al., 2014). Post-depositional collapse given the similarity in fining upwards character within adjacent interpods, implying that they could have been part of a more widespread, connected depositional system (Fig. 19e). The genesis of the interpod feature will be discussed more fully after examining the distribution of facies throughout the basin in the section below.

Facies distribution and depositional evolution of the Upper Jurassic

Four palaeogeographic maps have been constructed in order to capture the depositional evolution of the CG during Sequence I to IV (Figs. 4.20 & 4.21). These maps are based exclusively on the dataset analysed in this study, and can be compared with those of previous workers (e.g. Fraser et al., 2003). In the palaeogeographic maps, the upper and lower shoreface facies associations (SM1 & SM2) are grouped into a single shoreface environment, due to their close genetic relationship at a regional scale. In addition, the palaeogeographic reconstructions have been constrained by the following two key factors: (1) facies distributions and environmental interpretations are based on well log and core data; and (2) basin and sub-basin geometries have been guided by the isochore maps (Figs. 17 & 18). The maps also incorporate the concepts that underpin accommodation space creation and sand body preservation (Fig. 4.19), which are discussed separately (see Discussion).

Overall, the paleogeographic maps show that the CG evolved progressively from a relatively narrow NW-orientated basin in the Oxfordian, into a wider and more complex basin during the Tithonian (Figs. 4.20 & 4.21). Initially, the main depositional environment was shallow-marine, dominated by the shoreface (SM1 & SM2) and shelf (SM3) depositional environments. However, contemporaneous deep-marine (DM1) systems were also developing in the deepest parts of the sub-basins, where subordinate gravity flows (GF1) occur as relatively small and isolated sand bodies (Fig. 4.20a). The

extent of the deep-marine environment increased progressively at the expense of the shallow-marine systems, as relative sea-level rise outpaced sediment supply during the latter part of the Late Jurassic (Fig. 4.21b).

Several fluvial sediment input points are recognised based on the occurrence of estuarine and fluvial depositional environments, especially in the western, northern and southern parts of the study area. These are more evident in Sequences I-II compared to Sequences III-IV (cf. Figs. 4.20 and 4.21). Sandstone mineralogy, including heavy minerals, show that Jurassic (Fulmar Formation) and Upper Triassic (Skagerrak Formation) reservoirs have similar compositions (Stewart et al., 1999). However, Upper Jurassic sandstones are coarser grained (mainly fine- to medium-grained, but occasionally very coarse-grained to granule grade) than sandstones in the Skagerrak Formation within the CG (Stewart et al., 1999). This implies similar source areas, albeit with Late Jurassic fluvial systems capable of transporting coarser bedload material, perhaps reflecting a combination of a wetter climate, more persistent rivers and/or steeper gradients (Stewart et al., 1999; Wakefield et al., 1993). The inferred fluvial sediment entry points shown on the maps (Figs. 4.20 and 4.21) coincide with the location of the coarsest grained sediments, for example well 29/7-7 in the Curlew Platform. Extrapolating these river systems landwards, suggest that there were at least three main extra-basinal sediment source areas: (1) the Forth Approaches Basin in the west, which supplied sands through the Kittiwake-Gannet Platform (Stewart et al., 1999) and the Fisher Bank Basin (Clark et al., 1993), (2) the Mid North Sea High in the south and south-west, which supplied sands to the Curlew Platform and the Auk/Fulmar Terrace areas (Weibel et al., 2010), as well as the conglomeratic sediments found in the Fife-Angus area further the south-east (Currie et al., 1999; Mackertich, 1996; Spathopoulos et al., 2000), and (3) the north-east, which fed sands into the Jæren High and Cod Terrace.

Fig. 4.20: Previous page - General depositional environment maps at the main flooding events during the earliest part of Late Jurassic. (a) Intra-Oxfordian Seq I (b) Lower Kimmeridgian/Oxfordian Seq II.

Some authors have suggested that the Upper Jurassic sands were mainly reworked from local Triassic Highs (Hodgson et al., 1992; Wonham et al., 2014). However, the presence of granule to conglomeratic extra-basinal sediment is more consistent with direct fluvial sediment supply. Nonetheless, all workers agree that the proportion of fluvial facies within the Upper Jurassic of the CG is limited both spatially (widely spaced river mouths; Figs. 4.20 and 4.21) and temporally (upward reduction in coastal plain deposits; Fig. 4.4). This partly reflects limited preservation of fluvial facies, which would be consistent with (1) the gross retrogradational stratigraphic stacking pattern (Fig. 4.4), and (2) erosional transgression during shoreface retreat and associated wave ravinement (Fraser et al., 2003). This is also consistent with the recognition of estuarine FAs, which preserve drowned river mouths and the location of potentially significant extra-basinal sediment input points. These results confirm the early sequence stratigraphic model of Howell et al. (1996).

Oxfordian palaeogeography

Oxfordian palaeogeography is represented in maps constructed for Sequence I (Fig. 4.20a) and Sequence II (Fig. 4.20b). Sequence I shows a narrow (c. 25-km wide) NW-oriented depocentre that cuts obliquely across the CG, extending from the northern part of Western Graben, through the Elgin-Franklin Terrace and into the Breiflabb Basin (Fig. 4.20a). The basin-fill is dominated by shallow marine sediments (SM1, SM2 & SM3 FAs). Deep-marine shales (DM1) in the south-east, in the J-Ridge and Breiflabb Basin areas, may be part of a regional basin deepening trend that extended into Feda and Gertrud grabens (Fig. 4.1). Shoreface facies accumulated as early as the Callovian in the Danish Central North Sea (Andsbjerg et al., 2001), suggesting that the initial marine flooding into the UK's CG came from the south-east.

The main fluvial entry point during deposition of Sequence I was interpreted to be located in the north-west, where the basin was at its narrowest, based on the structural trend at the end of Triassic and occurrence of marginal marine facies (bay/lagoonal; CP3) in well 22/27A-1 (Figs. 4.21a). This is further supported by the non-occurrence of Sequence I in wells in the north and southern part of the Western Graben (Fig. 4.20a). This supports a regional, south-east-dipping palaeoslope along the western margin of the CG. The fluvial systems might have exploited the pre-existing zones of weakness in areas such as the Western Platform (Central Graben Shelf of Stewart et al., 1999), where the Triassic was originally thin and where salt walls were abundant (e.g. on the Kittiwake-Gannet Platform; Figs. 4.7a, 4.18a). Other fluvial sediment entry points may have also exploited pre-existing structures, most notably (1) a northeastern source (Scandinavian) that entered the CG via the Jæren High, and (2) southern source (Mid-North Sea High) that entered the CG via Auk High (Fig. 4.20a). Gravity flow facies are rare within Sequence I, but based on limited occurrence in the Western Graben (e.g. Well 22/21-7; Fig. 4.5), they appear to occur preferentially downdip of fluvial entry points, suggesting a genetic linkage (Donovan et al., 1993).

During deposition of Sequence II the basin expanded and the shoreface systems retreated landwards (Fig. 4.20b). The wider basins display a NNW orientation and there is the first evidence of basin segmentation into the Western and Eastern grabens, and other smaller sub-basins (Fig. 4.20b). The latter is manifested in a series of NNW-oriented ridges that created several deep-marine depocentres, especially in the southern part of Marnock Terrace (Fig. 4.20b). In addition, the J-Ridge during this time period was emergent, providing physical separation of the Breiflabb Basin and 30/6 Graben, which was a W-dipping half-graben. A new sediment input point appeared in the Curlew Platform, with rivers flowing from the Mid-North Sea High towards the north-east and into the Western Graben (Fig. 4.20b).

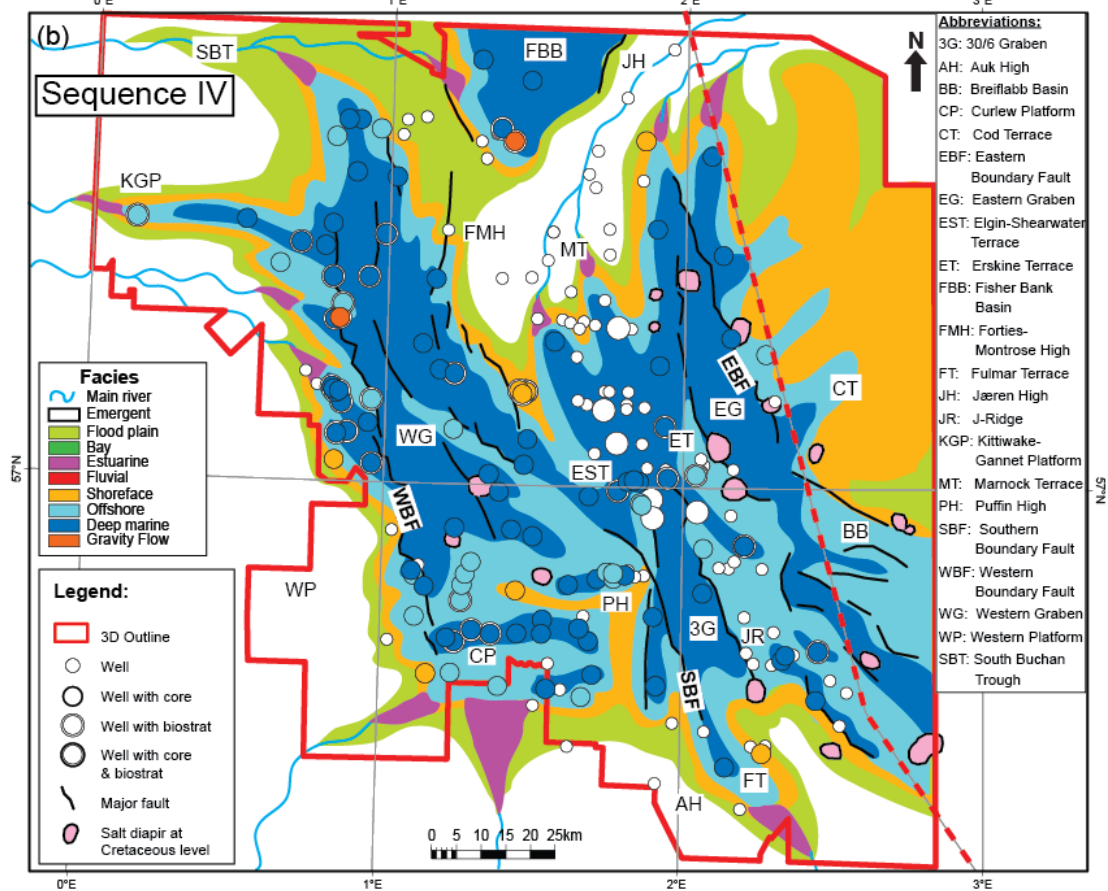
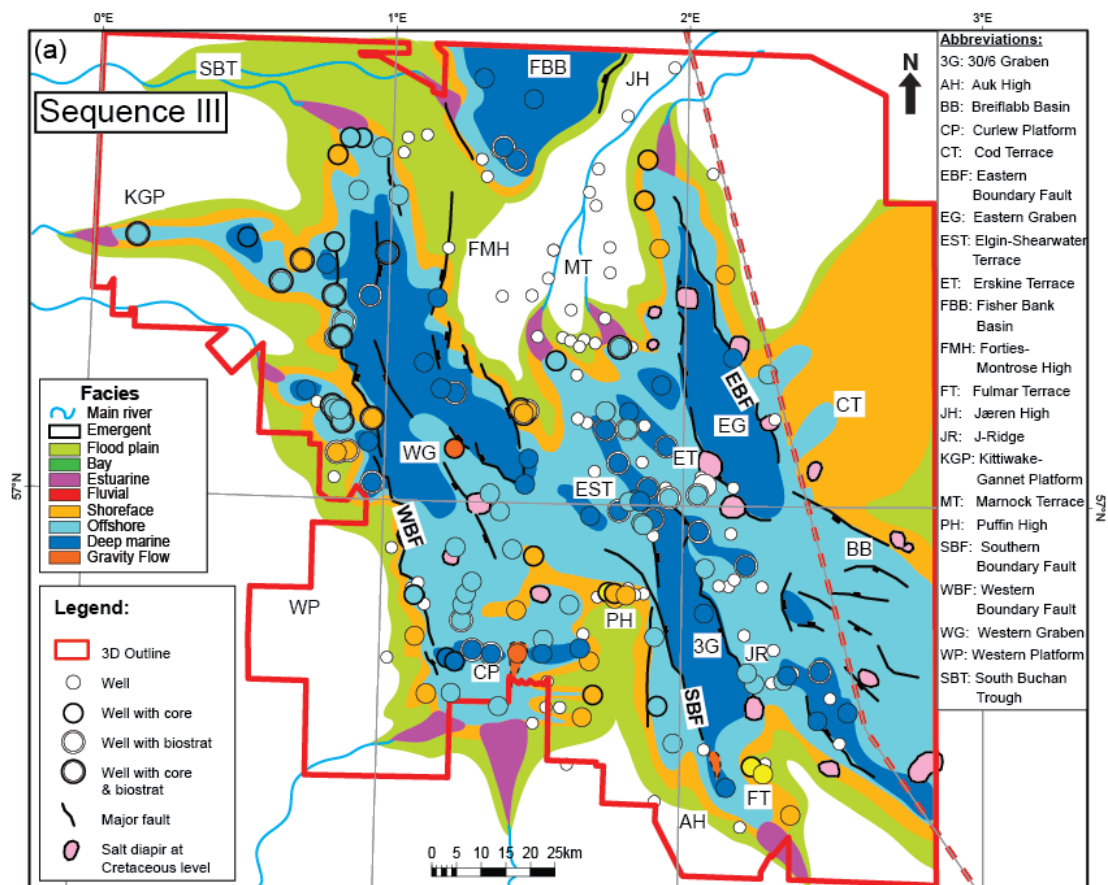


Fig. 4.21: General depositional environment maps at the main flooding events during the later part of Late Jurassic. (a) Intra-Kimmeridgian Seq III (b) Intra-Tithonian Seq IV.

Kimmeridgian-Tithonian palaeogeography

Kimmeridgian-Tithonian palaeogeography is represented in maps constructed for Sequence III (Fig. 4.21a) and Sequence IV (Fig. 4.21b). Shoreface retreat and basin deepening continued during this period, with a gradual expansion of deep-marine sedimentation within a wider and better connected series of sub-basins (Fig. 4.21a and b).

During deposition of Sequence III (Kimmeridgian), the Western, Eastern and 30/6 grabens were dominated by deep-marine FAs (Fig. 4.21a). The southern Marnock Terrace remained compartmentalised with NNW-striking ridges and multiple, smaller deep-marine depocentres. The shoreface systems have now retreated beyond the previous fault-defined basin boundaries and onto the adjacent platforms (e.g Kittiwake-Gannet Platform, Cod Terrace and Jæren High; Fig. 4.21a). Three fluvial sediment input points are recognised in the north-west exploited the proto-South Buchan Trough and the lows in the Kittiwake-Gannet Platform (Fig. 4.21a). Further south, the Mid-North High continued to supply fluvial sediment across the Western Platform, through the Curlew Platform and into the Western Graben (Fig. 4.21b). The shallow-marine incursion onto the Cod Terrace was more widespread than the Western Platform (Fig. 4.21a).

Sequence IV was characterised by widespread organic-rich mud deposition, which extended throughout the CG (Western, Eastern and 30/6 grabens; Fig 4.21b). Deep-marine conditions dominated both the basin centres and across the former platform margin areas, including the Kittiwake-Gannet Platform, the eastern edge of Western Platform and the western edge of Jæren High and Cod Terrace (Fig. 4.21b). The submerged CG was up to around 100 km wide at this time and

the coastline was pushed to the periphery of the basin. However, there is very little sign of deposition on the Forties-Montrose High, Jæren High and northern Marnock Terrace, which could have been exposed. Hence, fluvial sediment transport into the basin is still proposed for the (1) Jæren High to the north-east, (2) Buchan Trough to the north-west, and (3) Mid-North Sea High to the south and south-west (Fig. 4.21b). However, the southern part of Jæren High and Cod Terrace were both flooded, creating a large shallow-marine shelf that extended into the Norwegian Central North Sea.

The Base Cretaceous Unconformity (BCU), along with correlatable conformities within the basin centre, marked the end of Jurassic rifting and the start of post-rift thermal subsidence. In addition, this regionally-extensive, sub-marine erosion event caused truncation of the latest Jurassic stratigraphy, especially on the crest of rotated fault blocks and highs (e.g. in well 22/23A-6 in Fig. 4.5b; 23/26A-21 in Fig. 4.14; 30/13-4 in Fig. 4.16).

Controls on reservoir distribution

Sandstone distribution in the Upper Jurassic of the CG is complex, variable and frequently unpredictable (e.g. Donovan et al., 1993; Erratt et al., 2010; Stewart et al., 1999; Veldkamp et al., 1996; Wakefield et al., 1993; Wonham et al., 2014). Initially, this seems unusual since (1) the main reservoir comprises wave- and storm-influenced shoreface sandstones (Fulmar Formation and its lateral equivalents), and (2) palaeogeographic maps in this study (Figs. 4.20 & 4.21) and earlier examples (e.g. Fraser et al., 2003) predict widespread distribution of the shoreface systems (c. 25-75 km). This is consistent with the distribution of similar coastal sandstones in other basin settings that typically form regionally extensive and highly correlatable reservoirs (e.g. Bowman and Johnson, 2014; Hampson, 2010; Hampson et al., 2009; Hampson and Storms, 2003; Plint, 2010). However, this contrasts markedly with the CG area where preserved shoreface thicknesses are extremely variable,

both in basinal areas (c. 10's m to c. 300 m; e.g. Fulmar and Elgin/Franklin fields) and in adjacent platform locations (c. 0-50 m; e.g. Kittiwake and Durward/Dauntless fields). This variability largely reflects the unusual geological setting of the CG, in particular the following aspects of the and tectono-stratigraphic setting and related accommodation space creation: (1) the gross stratigraphic stacking pattern reflects short-term (100,000's to 1-2 million years) coastal regressive-transgressive (R-T) cycles superimposed on longer term (5-10 million years) coastal retrogradation (Fig. 4.4; Fraser et al., 2003; Hampson et al., 2009); (2) transgression during shoreface retreat was accompanied by transgressive erosion and wave ravinement, as manifested by winnowed coarse sand to pebbly lags in areas of stratigraphic condensation (e.g. Veldkamp et al., 1996); (3) accommodation space creation varied between the various structural domains at highly variable lengthscales (c. 1-25 km), reflecting both salt withdrawal and/or dissolution and extensional faulting (e.g. Figs. 4.17 and 4.18; Wonham et al., 2014); (4) antecedent topography, developed over a period of around 40 million years of erosion and non-deposition between the Late Triassic and latest Middle Jurassic, had been established prior to Upper Jurassic deposition (e.g. Stewart, 2007; Stewart et al., 1999); (5) syn-depositional accommodation space may have been created by incised valleys (Donovan et al., 1993); (6) non-incised 'valleys' may have been created by the syn-depositional collapse of salt walls, rather than by fluvial incision (e.g. Wakefield et al., 1993), and (7) post- and syn-depositional erosion added an additional complexity, most notably that resulting from initially the Base Volgian-Ryazanian Unconformity (Sansom, 2010) and, finally, the Base Cretaceous Unconformity (Donovan et al., 1993). The regional geological processes controlling the above have been extensively documented as follows: (1) regional tectonically-driven subsidence, (2) the onset and acceleration of rifting, (3) rift-related triggering of renewed Zechstein salt remobilisation, (4) collapse of the Early-Middle Jurassic 'Mid-North Sea Dome', and (5) regional relative sea-level rise during the Late Jurassic. However, there has been considerable debate as to the relative importance of these processes, particularly in relation to how they influenced the spatial and temporal distribution of depositional environments and reservoir sandstone bodies (c.f. Johnson et al., 1986; Wakefield et

al., 1993; Donovan et al., 1993; Veldkamp et al., 1996; Howell et al., 1996; Stewart et al., 1999; Erratt, 1999; Fraser et al., 2003; Sansom, 2011; Mannie et al., 2014; Wonham et al., 2014). These studies show a relatively high level of agreement with regards to the gross depositional environments, as synthesised earlier, but considerably less agreement with regards to the controls on reservoir thickness and quality distribution. The latter is exacerbated by the internal stratigraphic architecture occurring on a vertical scale that is below seismic resolution (e.g. Veldkamp et al., 1996). Consequently, well log correlation has to be guided by models that capture the interaction between depositional environments and accommodation space behaviour: i.e. the tectono-sedimentary setting.

Tectono-sedimentary settings

The result of seismic interpretation shows that the CG could be divided into different structural domains (Fig. 4.2). These domains could be further classified into tectono-sedimentary settings. In the most recent synthesis of the geological controls on Upper Jurassic sandstone distribution in the CG, Wonham et al (2014) identified three distinct tectono-sedimentary settings (Fig. 4.22): (1) Type A (hangingwall setting); (2) Type B (reactivated pod setting), and (3) Type C (interpod setting).

Type A settings comprise of classical syn-rift hangingwall successions, which are associated with major faults that offset the deeply buried Rotliegend, such as during initial Oxfordian rifting. This type of setting results in traditional syn-rift stratal geometries, with thickening into hangingwall faults, where salt was either thin or absent (e.g. North Viking Graben, South Viking Graben, Moray Firth, etc.). In the CG, Type A settings are found in the Western, Eastern and 30/6 grabens (Fig. 4.2).

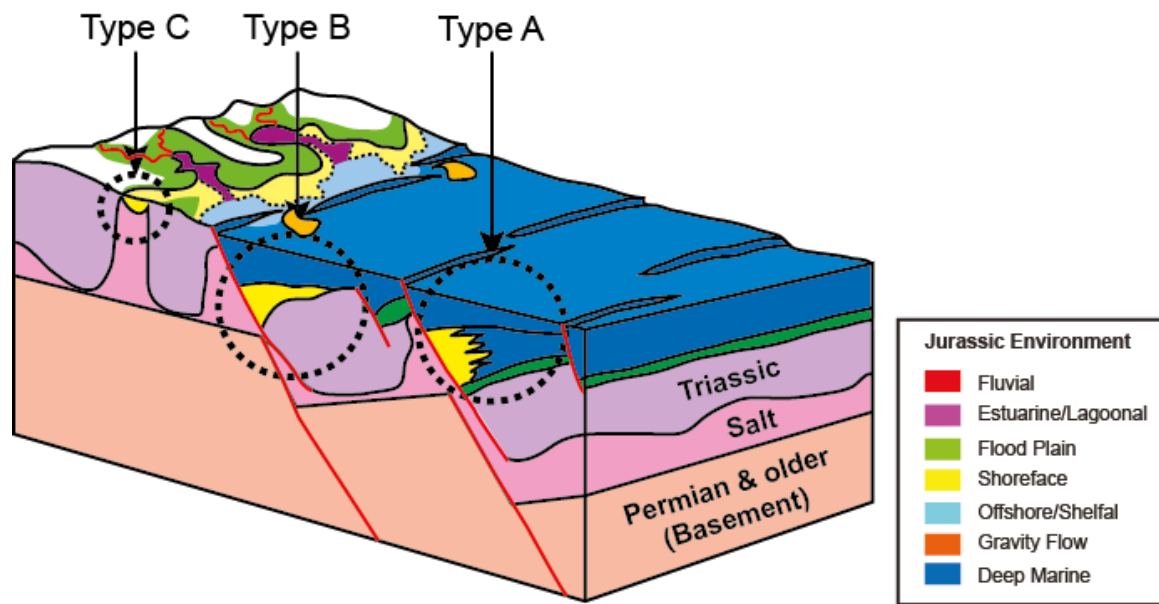


Fig. 4.22: Tectono-sedimentary models (Wonham et al., 2014). See text for explanation of the types.

Type B settings are also developed within the main CG, but involve remobilisation of Zechstein Salt and overlying Triassic pods. Salt walls that were formed during the earlier Triassic, and usually located at the crest of Permian rotated blocks, have re-grown during Late Jurassic rifting as the blocks re-rotated along with the Triassic pods (Fig. 4.19a). Subsequently, these growing salt walls have also influenced the deposition of syn-rift Oxfordian sediments, resulting in the thickening of these successions towards the centre of the Triassic pods, away from the salt walls, and thinning onto the crests of the salt walls (Fig. 4.19b). In the CG, Type B settings could be found in the Elgin-Shearwater Terrace (Fig. 4.11b) and Fulmar Terrace (wells 30/16-6 & 30/12B-3 in Fig. 4.16).

The most significance contribution of Type B setting is that it created the largest syn-depositional accommodation space in the CG during the Late Jurassic, which resulted in the classical, amplified, prograding shoreface successions, first seen in the Fulmar Field (Johnson et al., 1986). Hence, Type B settings comprise the thickest accumulations of high-quality shallow-marine sandstone reservoirs (up to c. 300 m thick) and coincide with the largest hydrocarbon accumulations (e.g. the Elgin-

Franklin and Fulmar fields). The reservoir successions are diachronous (e.g. Oxfordian sandstones in Elgin-Franklin and Early Kimmeridgian in Fulmar), strongly aggradational and, at field- and sub-regional-scales, show variations in thickness and stratigraphic architectural patterns that reflect salt withdrawal (Fig. 4.23; Kuhn et al., 2003). The temporal change in the relative interplay between salt withdrawal and rifting can result in Type A settings evolving into Type B settings, as seen in the Fulmar Field (Johnson et al., 1986; Kuhn et al., 2003), including changes in basin bathymetry and depositional environments (e.g. from shallow- to deep-marine).

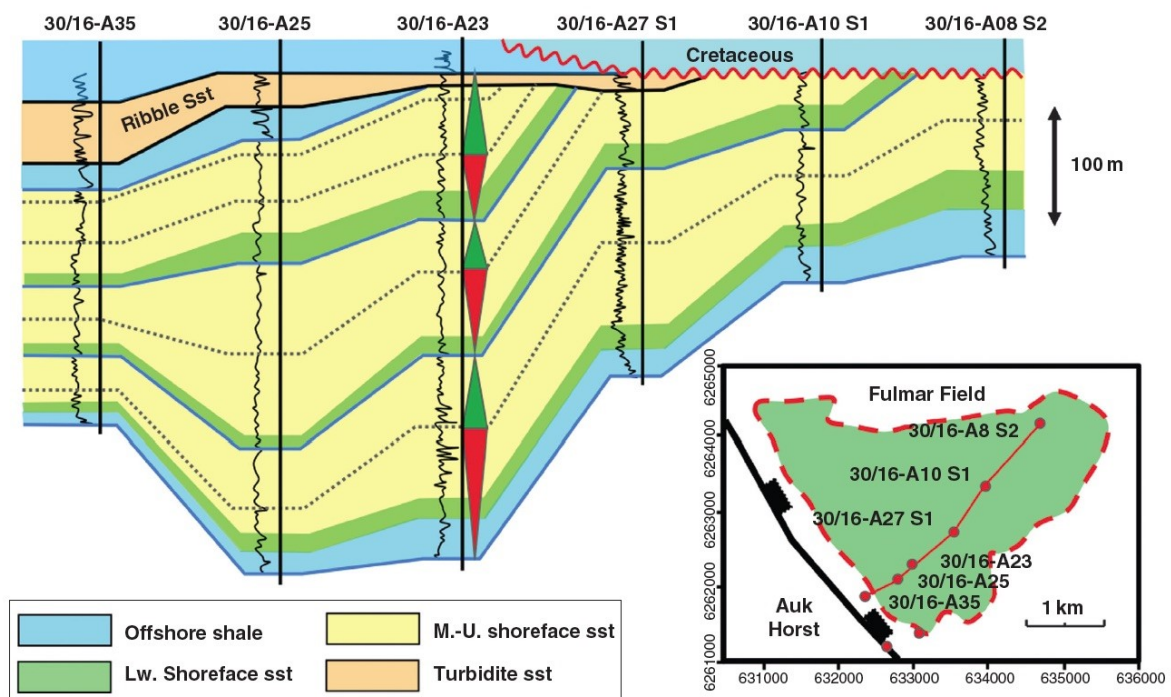


Fig. 4.23: Well correlation in across the Fulmar Field showing variation in thickness pattern due to salt withdrawal (Kuhn et al., 2003; redrawn by Wonham et al., 2014).

Whilst the two settings described above are found in the grabens or lows, Type C settings are most typical of platform areas on either side of the CG (the Western Platform in the west and the Jæren High/Cod Terrace in the east). However, Type C settings also occur on the larger intrabasinal highs (e.g. the Forties-Montrose High and the J-Ridge). Type C settings contain variable reservoir thickness, which are up to c. 50 m thick on the Western Platform (e.g Kittiwake, Durward/Dautless fields;

Stewart et al., 1999) and up to c. 500 m in the Cod Terrace (e.g Gydda and Tambar fields; Mannie et al., 2014). These settings are characterised by incomplete vertical shoreface successions, which reflect multiple coastal regressive-transgressive (R-T) cycles, usually with a strong overall retrogradational stacking pattern. They are dominated by typical shoreface facies, but with variable well log profiles reflecting different sediment supply/accommodation space scenarios. In the absence of unique seismic-based stratigraphic architectural styles the Type C/interpod setting has resulted in the greatest exploration uncertainty, with a range of depositional models, in part driven by detailed, local observations, which are reviewed below.

Interpod depositional models

Stratigraphic successions beyond the more rapidly subsiding grabens, such as the Western Platform and Cod Terrace (Figs. 4.2 & 4.11a), are characterised by syn-depositional withdrawal of salt from elongate salt walls, which created accommodation space during the Late Jurassic. Salt withdrawal probably resulted from a combination of dissolution and remobilisation of salt, in part triggered by ongoing rifting and sediment loading (Stewart, 2007). This resulted in the collapse of the roof and formation interpod features, which are located either side of larger pods infilled mainly by thick Triassic mudstones (Smith Bank Group). A few models have been proposed to explain the origin of these interpods could be classified as: (1) pre-depositional origin (Fig. 4.19c), including palaeovalley (Donovan et al., 1993; Wakefield et al., 1993) and palaeotopography models (Høiland et al., 1993; Jackson et al., 2010; Stewart, 2007; Stewart et al., 1999); (2) syn-deposition origin (Fig. 4.19d), particularly the salt minibasin collapse model (Hodgson et al., 1992; Mannie et al., 2014); and (3) post-depositional origin (Fig. 4.19e), particularly the extensional fault collapse model (suggested but not preferred by Høiland et al., 1993). There is uncertainty associated with all the models, mainly due to the relatively low resolution of both seismic data, especially over the platform areas, and biostratigraphic data (e.g. Veldkamp et al., 1996). This hampers the detailed examination of the

relationship between the infill of the Jurassic interpods and the flanks of the Triassic pods, which is the main source of the interpretational uncertainty. In addition, some of the proposed models straddle the four process-based models listed above. These uncertainties require a synthesis of the strengths and weaknesses of these models, which are discussed below.

The Palaeovalley Model (1) was developed primarily from the elongate geometry of the interpods on the Western Platform, which take on the superficial appearance of a network of roughly E-W oriented valleys (Clark et al., 1999; Høiland et al., 1993; Stewart et al., 1999; Wakefield et al., 1993; Wonham et al., 2014). This builds on two main concepts: (1) the pod/interpod concepts developed originally for Triassic continental depositional systems, which show a close linkage between fluvial systems and accommodation space created from salt migration (Hodgson et al., 1992; Smith et al., 1993), and (2) the salt dissolution concepts developed from the Southern Oman Salt Basin. Between the end of the Triassic and the onset of Late Jurassic transgression (c. 40 Myr) the Western Platform, and the greater Central North Sea area, was uplifted, eroded and gradually deflated.

Accommodation space during the Late Jurassic was created by rising relative sea-level and subsidence within the interpods (flanked by pods composed of Triassic Smith Bank Group mudstones), with the latter due to the following: (1) dissolution of salt from elongate salt walls caused by 'repeated ground-water penetration', which created Late Jurassic valleys, and (2) 'collapse of the valley fill' (Wakefield et al., 1993). During the Bajocian to Oxfordian the dendritic valley network was occupied by Jurassic rivers, which transported sands basinwards to the CG.

Subsequently, the Late Oxfordian to Kimmeridgian transgression resulted in erosional transgression, wave ravinement was widespread original fluvial deposits within the palaeovalleys were reworked. Transgressive sands were preserved within syn-transgressive subsiding 'valleys'. However, Wakefield et al (1993) stressed that the 'valleys' were not incised valleys *sensu* Van Wagoner et al., 1990); they were not 'caused by a drop in relative sea-level (and) incision by rivers'. Instead, these palaeo-valleys

were generated purely by halokinesis, with the rivers being passive features that utilised these palaeo-valleys during the Bajocian to Oxfordian. It was stressed that there was no palaeotopography, but the term 'palaeovalley' was adopted on the basis that they had once been occupied by Late Jurassic rivers. This interpretation contrasts with that of Donovan et al. (1993) who interpreted the same features as the result of lowstand fluvial incision, which created an incised topography that was subsequently infilled during transgression. The following aspects of Model 1 are problematic: (1) a Western Platform facies map (Fig. 4.21a) shows Lower Kimmeridgian shoreface sands forming a laterally extensive, c. N-S to NNW-SSE trending facies belt (c. 20-30 km wide and >100 km long) that extends far beyond the width of the interpod areas (c. <10 km), (2) the infill of the palaeovalleys comprise transgressive bioturbated shoreface sandstones, which show minimal lateral variation and an open marine ichnofauna, (3) complete absence of fluvial deposits, (4) absence of estuarine deposits, which would be expected in a marine flooded valley (e.g. tidal channels, tidal bars, tidal flats, lagoons, etc.), and (5) the vertical stratigraphic architecture comprises a transgressive, back-stepping shoreface depositional system (Veldkamp et al., 1996; Hampson et al., 2009). The lack of fluvio-estuarine facies preservation was explained by extensive wave ravinement during erosional transgression in the Kimmeridgian (Veldkamp et al., 1996) but this remains a contentious problem for this model. In addition, the Kimmeridgian-Tithonian sediments that filled the interpods display an almost identical, fining upwards vertical facies succession, implying high lateral facies continuity (e.g. Sequence III in wells 21/16-1, 21/18-2A1, & 21/19-4 in Fig. 4.12). If the individual valleys had evolved as separate, disconnected systems, one would expect that the stacking patterns would also be different. Despite the above, Wakefield et al (1993) concluded that the primary factor for the preservation of the Fulmar shoreface sands was the generation of salt-related accommodation space (see Model 3). This was supported by the isopach maps in the Kittiwake Field, which resemble those from other salt-related accumulations (e.g. Fulmar Field), albeit with much lower thicknesses.

The Palaeotopography Model (2) highlights the common truncation of intra-Triassic seismic reflectors and the presence of cap rock on top of interpod Zechstein salt walls (Stewart et al., 1999; Fraser et al., 2003). This is used to support the concept of differential erosion of salt-cored interpods and show that exposed crests of salt highs can erode faster than the crests of adjacent mudstone-filled pods (as in Model 1). In the context of the Western Platform (=West Central Shelf of Stewart et al., 1999), the basal surface of the Late Jurassic transgression was 'largely controlled by pre-existing topography, which had evolved during a significant period of (late Triassic-pre Upper Jurassic) subaerial exposure'. During this c. 40 Myr period the salt walls (interpods) became 'valleys' that were flanked by emergent, low-relief hills, which comprised cohesive mudstones (pods of Smith Bank Group mudstones). The pods were less prone to fluvial erosion, whereas dissolution and reprecipitation preferentially removed halite from the interpods. The formation of these palaeotopography is somewhat supported by the occurrence of thick (up to c. 100 m) Late Triassic Skagerrak fluvial sandstones above anhydrite deposits in the interpods of the Kittiwake-Gannet Platform (Fig. 4.12). Then, the sedimentation seems to stop until these lows were filled by Late Jurassic sediments.

Interpretation of onlapping stratal geometries from well log correlation and isopach maps of the Late Jurassic (Kimmeridgian) infill, based mainly in the Kittiwake and Durward/Dauntless field areas (Stewart et al., 1999), was used to infer the passive infill of an existing palaeotopography during the Late Jurassic. The topographic relief between salt valley floors and minibasin hill summits, immediately prior to the Late Jurassic transgression, was estimated to be some tens of metres (Stewart, 2007). The accommodation space for the Late Jurassic transgressive sedimentary system, which retrograded westwards over this irregular, interpod surface, was estimated to be around 80% palaeotopography and 20% salt withdrawal (Stewart et al., 1999). The nature of the infill was considered to be a function of the (1) connectivity of individual salt valleys to the transgressing

shallow marine shelf, and (2) the shape and depth of the valley floor relative to the lowest point on the surrounding valley wall (Stewart, 2007). Once infilled, the younger (Volgian to Ryazanian) mud-dominated shallow- to deep-marine successions display contrasting subsidence patterns, interpreted as mainly related to salt withdrawal and regional eastwards tilting. The following aspects of Model 2 are problematic: (1) the sands infilling of the palaeotopography comprise transgressive, open marine bioturbated shoreface sandstones, (2) absence of terrestrial deposits, (3) absence of estuarine and related marginal marine deposits, (4) the transgressive, retrogradational stacked shoreface succession (Hampson et al., 2009; Veldkamp et al., 1996), (5) multiple coarse grained to pebbly lag deposits representing wave ravinement surfaces within the palaeotopographic infill (Glennie and Armstrong, 1991; Veldkamp et al., 1996). It is questionable as to how such a long-lived topographic surface could survive and how it could be passively infilled during aggressive marine erosion. Evidence for the latter is widespread in the form of wave ravinement surfaces (or shoreface 'razor'), which are effective at creating peneplaning pre-existing topography (e.g. Sixsmith et al., 2008; Swift, 1968). This is consistent with the gross transgressive evolution of the Kimmeridgian succession (e.g. Hampson et al., 2009). The published stratal geometries inferred from well log correlations in the Kittiwake Field (Stewart et al., 1999) can also be interpreted as the result of a combination of thinning against an evolving surface and truncation by wave ravinement. The trajectory of wave ravinement surfaces in Kittiwake is in the range of 0.03 to 0.07 degrees. This compares favourably with outcrop-based measurements from the transgressive Upper Cretaceous Hosta Sandstone, which are c. 0.05-0.2 degrees, with steeper dips at the landward termination of around c. 0.3 degrees (Sixsmith et al., 2008). This emphasises the very low-angle dip of stratal surfaces within the sandy part of the interpods (see also Hampson et al., 2009), which appear deceptively steeper in many published sections because of the large vertical exaggeration (e.g. Stewart et al., 1999).

The Salt Minibasin Collapse Model (3) includes components of the reactivated pod tectono-sedimentary setting (Type B) but on the much smaller scale of interpods (Type C) (Wonham et al., 2014). This model stems from the fact that in places there are significant coarse Triassic sediments above the salt walls (50-100 m), which could inhibit dissolution of the underlying salt. Proponents of this model argue that the Kimmeridgian-Tithonian sediment show thickening towards the centre of interpods, such as in the Kittiwake Field (Glennie and Armstrong, 1991) and in the Ula-Gyda fields (Mannie et al., 2014) (Fig. 4.24). These thickness patterns resemble those reactivated pods, such as the Fulmar Field (Johnson et al., 1986), which may be their larger equivalents. The trigger for the collapse of salt in this case could be extension or sediment loading (Hodgson et al., 1992).

Accommodation space creation was synchronous with shallow-marine sand deposition, which removes the need for any fluvial or terrestrial influence on their formation. This could also explain the lack of fluvial facies within the interpods, since minibasin/interpod collapse is synchronous with initial marine transgression (Hodgson et al., 1992). This model has been further developed on the Cod Terrace (Mannie et al., 2014). Here the interpods are more accurately referred to as 'supra-diapir minibasin-style depocentres' (abbreviated to 'salt minibasin collapse'). Essential elements of this model are as follows (Mannie et al., 2014): (1) rifting triggered the initial rise and fall of salt diapirs, and the formation of supra-diapir minibasin-style depocentres above collapsing salt walls (i.e. interpod depocentres); (2) Late Jurassic infill comprised up to 500 m of net transgressive shallow-marine (shoreface) sandstone reservoirs, with maximum thicknesses located within the minibasin axes and a gradual reduction in thickness towards the flanks; (3) internal stratigraphic architecture varies laterally within and between separate minibasins; (4) proximal-distal facies variations (shoreface to offshore) can be mapped over scales much larger than individual minibasins, while facies belts extend across multiple minibasin margins; (5) the minibasins formed a linked network that were filled with sediment supplied from regional shoreface-shelf sediment-routing systems, with a strong alongshore sediment transportation consistent with the wave-dominated shoreline interpretation. The following observations resemble the observations of Wakefield et al.

(1993): (1) shoreface-shelf facies belts are much wider (c. 10s km) than the salt diapir collapse minibasins (c. 2-10 km); (2) facies belts cross-cut the salt diapir collapse minibasins, often at high angles; and (3) there is no genetic relationship between the shoreface-shelf depositional system and the syn-depositional, salt-related accommodation space creation. This suggests that the model is also applicable to the Western Platform on the opposite side of the CG. Optimum reservoir thickness and quality occurs where proximal shoreface trends intersect collapsing minibasins. This model is a smaller-scale version of the reactivated pod model (Type B) of Wonham et al., (2014).

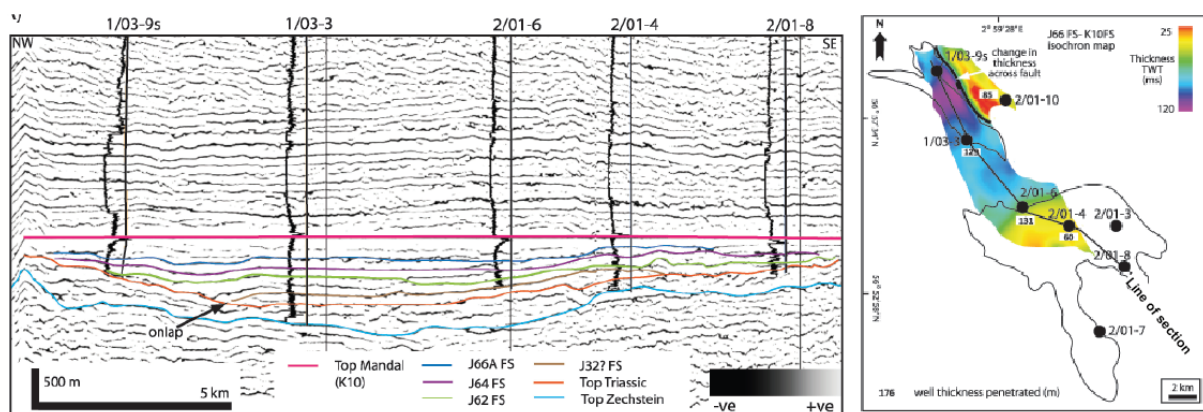


Fig. 4.24: Seismic section across Tambar and Gydda fields in the Norwegian North Sea showing the divergent seismic-stratigraphic character inside the minibasin/interpod (Mannie et al., 2014).

The Fault Collapse Model (4) does not require the formation of valleys but involves salt wall roof collapse due to differential loading in the latest Jurassic period. It invokes a greater degree of extensional faulting, with salt withdrawal and/or dissolution playing a secondary role (Høiland et al., 1993). The process initially involved widespread shallow marine deposition above the variable Triassic substrate during the Kimmeridgian R-T cycles. This was followed by the collapse of the incompetent salt due to differential loading or extension during the latest Kimmeridgian to Tithonian period (Fig. 4.19e). The Kimmeridgian sediments on the Triassic pods, which would have formed ‘turtle back highs’, would now be subject to erosion, while the collapsed interpods would be filled by

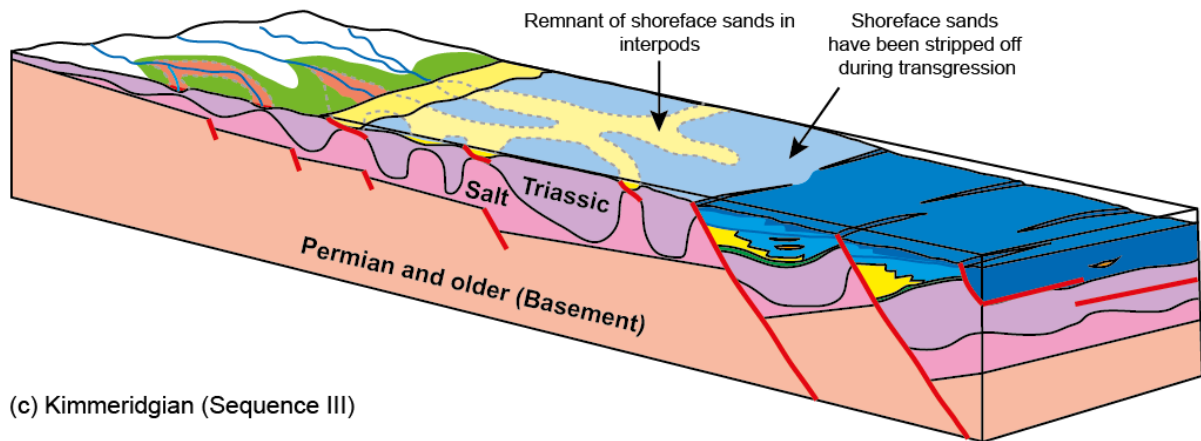
deeper water sediments. When rifting stopped, the turtle-back-highs would be eroded, with remnants of Upper Jurassic deposition preserved within the down-faulted interpods. This could explain the similarity in Kimmeridgian stacking pattern throughout the Kittiwake-Gannet Platform. However, this model may have generalised the Late Jurassic fillings of the interpods. Variations in stacking pattern do occur between the different interpods of the Jæren High, for example (e.g wells 22/15-2 in Fig. 12 and 23/21-2 in Fig. 4.13).

Synthesis

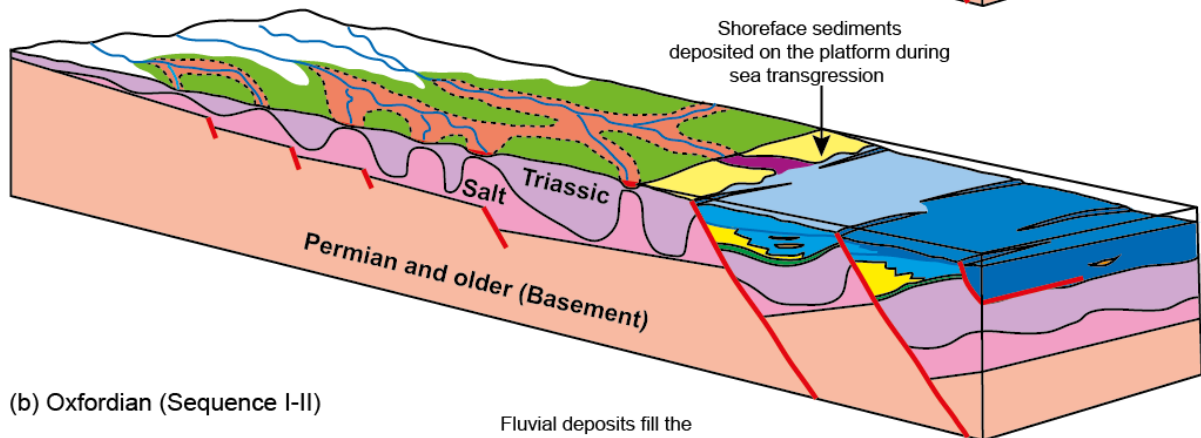
The current study could not pin-point the exact model for the genesis of the interpods since (1) the seismic data resolution has not improved in order to determine the nature of the Upper Jurassic-Triassic boundary on the flank of the interpods and (2) the well data are not sampling the edge of these interpods. However, analogues based on the seismic and well data analysis of Mannie et al. (2014) could be used. This analogue shows a clear diverging pattern towards the centre of the interpod/minibasin in Gydda Field, indicating a syn-sedimentary salt collapse. The space in the interpods was probably created in the Late Triassic as salt walls were reduced by dissolution and at the same time filled by Upper Triassic Skagerrak fluvial deposits (Fig. 4.25a). This process could leave the Triassic pods as topographic highs during the Late Triassic and explains why Skagerrak fluvial deposits have not been found above the Triassic pods. The entire topography was later uplifted by the early to middle Jurassic Thermal Dome Event and resulted in the erosion of the entire area.

Fig. 4.25: Next page – Summary of the main depositional settings during the evolution of the Western Platform. (a) Depositional setting during Late Triassic. (b) Depositional setting during Sequence I to II (Oxfordian). (c) Deposition setting during Sequence III (Kimmeridgian). (d) Depositional setting during Sequence IV to VI (Tithonian). See text for discussion.

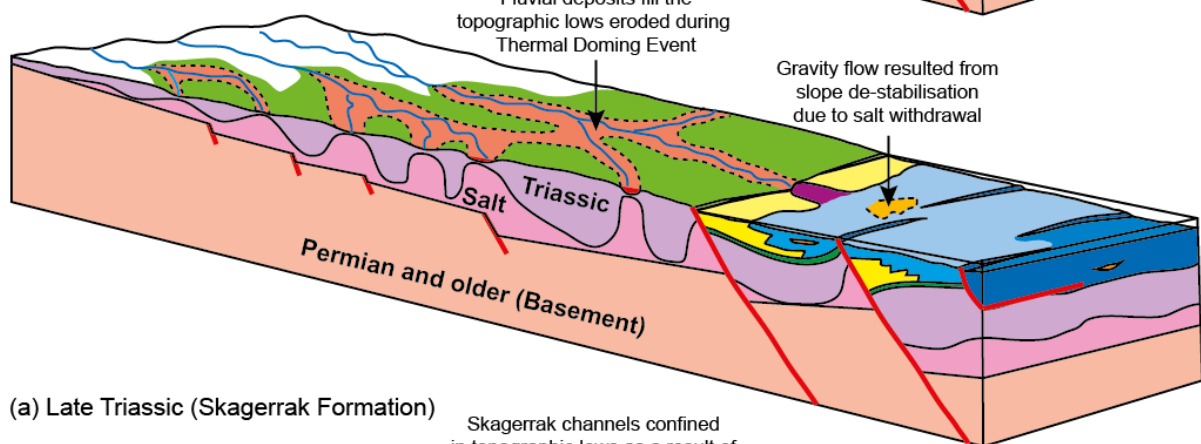
(d) Tithonian (Sequence IV and later)



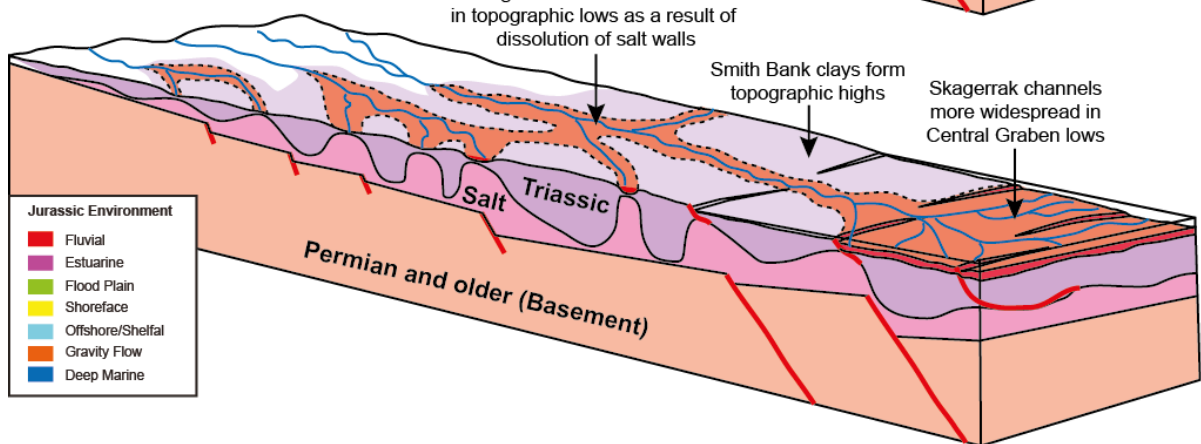
(c) Kimmeridgian (Sequence III)



(b) Oxfordian (Sequence I-II)



(a) Late Triassic (Skagerrak Formation)



Jurassic Environment	
Fluvial	Estuarine
Flood Plain	Shoreface
Offshore/Shelfal	Gravity Flow
Deep Marine	

When the Jurassic rifting occurred, these interpods were reactivated again as conduits for sediment supply into the Central Graben (Fig. 4.25b). As the sea level was rising, the shoreface sediments backstepped onto the platform edge, while estuarine sediments were deposited in the accommodation space formed by the previous fluvial incisions (Fig. 4.25c). The evidence for fluvial or non-marine facies were then wiped-out by transgressing shoreline during the Kimmerigian-Tithonian, which have flooded the Western Platform and Cod Terrace, leaving a few small pockets of non-marine facies in the Kittiwake area (e.g Well 21/18-6 in Fig. 10a; Fig. 4.25d). The major transgression also explains the occurrence of thin shallow marine sandstones over the Smith Bank-filled pods (e.g Well 21/17-3 in Fig. 10a).

Conclusions

Nine sedimentary facies associations have been identified within three main depositional environments. The main depositional environment within the Upper Jurassic interval of the Central Graben is shallow marine but deep marine and coastal plain environment can also be found. The coastal plain environment which includes fluvial, estuarine and lagoonal facies associations also exhibit some bioturbation which indicate some marine influence. Gravity flow facies associations are also rare but could be found both in the deep marine as well as shallow marine environment.

The Upper Jurassic interval has been divided into six major sequences bounded by major flooding event and their correlative unconformities, using genetic sequence stratigraphic method. This method is more suitable for rift basin due to the nature of sedimentation that responds more to extensional pulse rather than to global sea level rise.

In this study, we found that the distribution of Late Jurassic sedimentary facies follows a net-transgressive model, starting with almost a single depocentre during the Oxfordian and expanded into several depocentres during the Kimmeridgian before becoming one single basin during the Tithonian. Thickness maps have shown that at the start of rifting, the Oxfordian depocentre was oriented in a NW strike, in similarity to the older Triassic structural trend. This demonstrated that the initial extension had used the older structural fabric and also may have rejuvenated some Triassic faults. When rifting was at its peak during the Kimmeridgian-Tithonian period, the grabens were orientated in a more northerly strike pattern.

We also have shown that the distribution of Upper Jurassic sediments on the platform highs in the basin margin areas are complex due to the halokinetic process of the underlying Permian salt walls. The top of these salt walls were subject to dissolution and was filled with Upper Triassic fluvial deposits. Later, after a general uplift during the Thermal Doming Event, these topographic lows were used to carry sediments into the deeper part of the Jurassic basin. However, transgression during the Kimmeridgian-Tithonian period have wiped-out almost all of the evidence of the non-marine deposits. What is left now are remnant deposits of transgressive marine sands that have subsided into the lows.

Chapter 5: Gravity flow deposits in salt-influenced, syn-rift setting and implications for the hydrocarbon prospectivity: Examples from the Central Graben, United Kingdom Continental Shelf

Introduction

Density-driven flows, including turbidity currents, result in a distinctive suite of coarse- to fine-grained event beds and facies types (Talling et al., 2012) and are the primary mechanism for transporting large volumes of fine- and coarse-grained sediments into the deepwater. Coarse-grained, or non-cohesive, sediments display a range of external geometries and internal architectures, and may be host to numerous hydrocarbon accumulations in a wide variety of deep-water settings throughout the world (Nelson et al., 2011; Slatt and Roger, 2013). This includes >30 giant discoveries containing >500 BBOE of hydrocarbons, mostly in prolific Tertiary reservoirs in the deep-water basins of the Gulf of Mexico, Brazil and West Africa (Pettingill and Weimer, 2002; Slatt and Roger, 2013). The gravity-flow-related discoveries in these latter regions are located on passive margins, with their deposition controlled by varying degrees of syn-depositional tectonics related to halokinesis, and gravity-driven shortening, thrusting and folding (Hudec and Jackson, 2009; Morley et al., 2011; Slatt and Roger, 2013; Valle et al., 2001). Numerous studies have shown how these tectonic processes exert a major control on sand body distribution in both slope and basin floor settings, adding significant complexity to reservoir development and distribution (Anderson et al., 2000; Gee and Gawthorpe, 2006; Olafiranye et al., 2013).

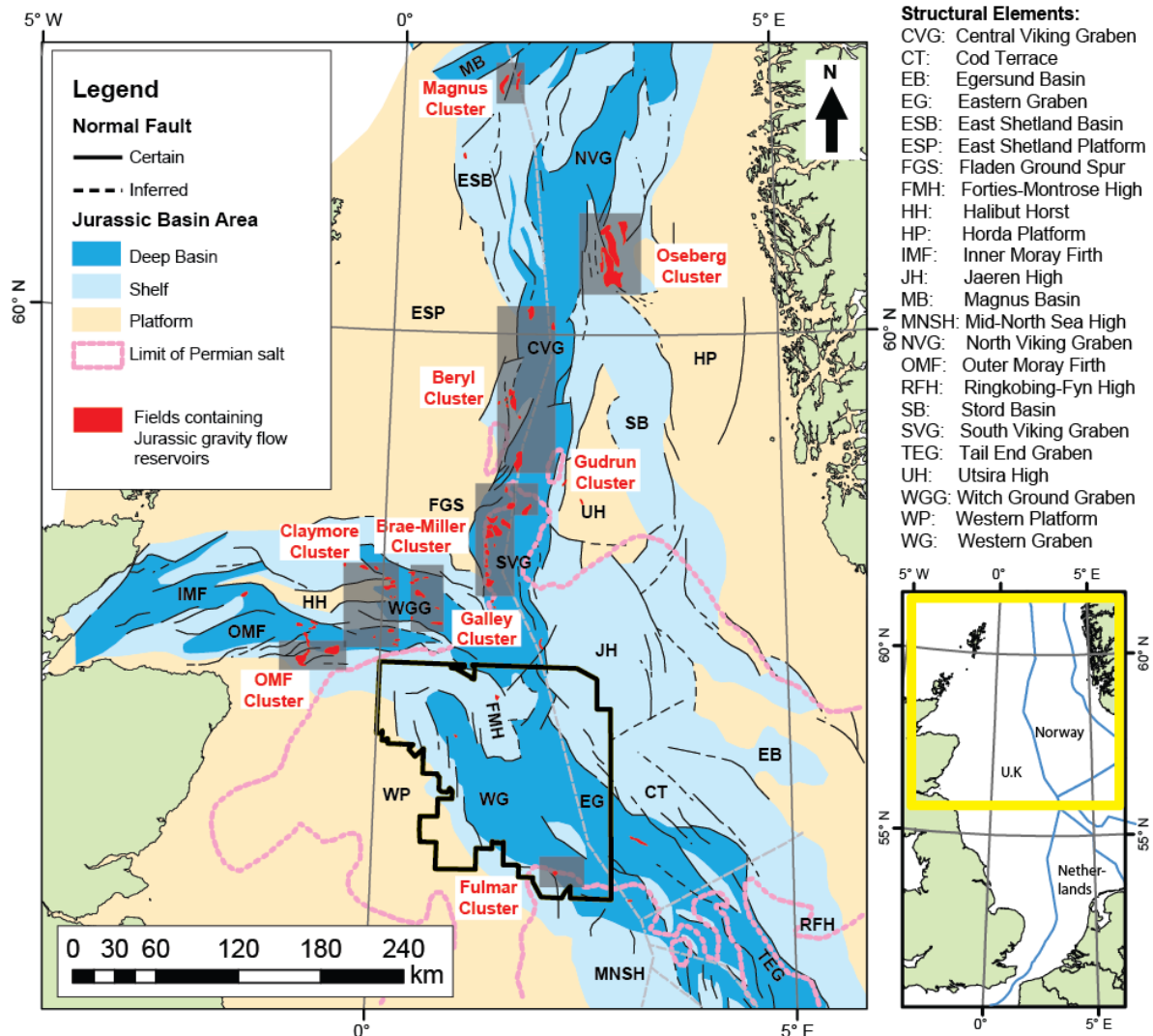


Fig. 5.1: Map showing major Upper Jurassic structural elements in the North Sea (after Fraser et al, 2002). Also drawn is the limit of Permian salt, which have affected the Mesozoic structure of the Central Graben through the underlying halokinetic structures. Selected hydrocarbon fields containing proven Upper Jurassic gravity flow reservoirs are highlighted in red while the PGS' 3D mega-survey used to ring-fence the study area is in black. The fields could be grouped into clusters containing lithologically similar reservoirs. See text for detail. Note that there is a lack of proven major gravity flow reservoirs in the Western and Eastern Graben and the cause of it is the main discussion of this study.

In contrast to passive margin settings influenced by typically thin-skinned tectonic processes, there have been far fewer studies into the distribution of gravity flow sediments in syn-rift settings, although the Mesozoic of the North Sea region is a notable exception (Jackson et al., 2011; Nielsen et al., 2015; Ravnås and Steel, 1997). In this region, Upper Jurassic to Lower Cretaceous gravity flow

deposits form the reservoirs occur in around 70 oil and gas fields (Fig. 5.1; Erratt et al., 2010; Fraser et al., 2003). Unfortunately, the seismic expression of these gravity flow deposits is compromised by limited seismic data quality and resolution, often reflecting a combination of low frequency and the presence of multiples (Erratt et al., 2010). However, this limitation is at least partly offset by an abundance of high-quality data from exploration, appraisal and development wells (e.g. cores, logs, fluids, production data, etc.). Consequently, the majority of discoveries in Mesozoic gravity flow reservoirs in the North Sea, most notably in the Viking Graben and Moray Firth Basin, have necessitated the integration of seismic, geological and petroleum engineering data analysis to aid reservoir characterisation and production (Fig. 5.1).

One of the most striking features of Upper Jurassic gravity flow reservoirs in the Mesozoic North Sea Basin is their extreme variability in terms of facies characteristics, internal stratigraphic architecture, external geometry (e.g. shape, thickness and orientation), and temporal and spatial distribution (Table 5.1). In all cases, this variability reflects a combination of largely submarine rifting superimposed on, and augmenting, an overall eustatic rise in sea-level. Extensional faults influenced both sediment supply and sea-bed bathymetry, with the overall basin physiography being characterised by variably connected sub-basins, particularly in the Viking Graben and Moray Firth (Fraser et al., 2003). In the case of the Central Graben (CG), bathymetry was further complicated by synchronous halokinesis superimposed on extensional faulting. Although oil-bearing deep-water sandstones have been known to exist in the CG since the early days of exploration (Johnson et al., 1986), there are no standalone fields producing exclusively from deep-water reservoirs. Nevertheless, many wells (>30 available in this study) in the basin have encountered deep-water sands and some have been cored (Fig. 5.2; Table 5.2). This wealth of well data and the availability of a 3D seismic mega-survey that covers the basin (>20,000 km²) has provided an opportunity to study the gravity flow deposits in the CG in order to establish the controls on their thickness, quality and

distribution. In particular, this study aims to increase our understanding of the character and distribution of deep-water sediments in relation to the control exerted by halokinesis within an overall syn-rift setting. Furthermore, with the potential hydrocarbon resource estimated at 8 billion barrel of oil equivalent (BBOE), Upper Jurassic gravity flow reservoirs stand-out as a relatively underexplored but highly promising play type within the North Sea (Munns et al., 2005).

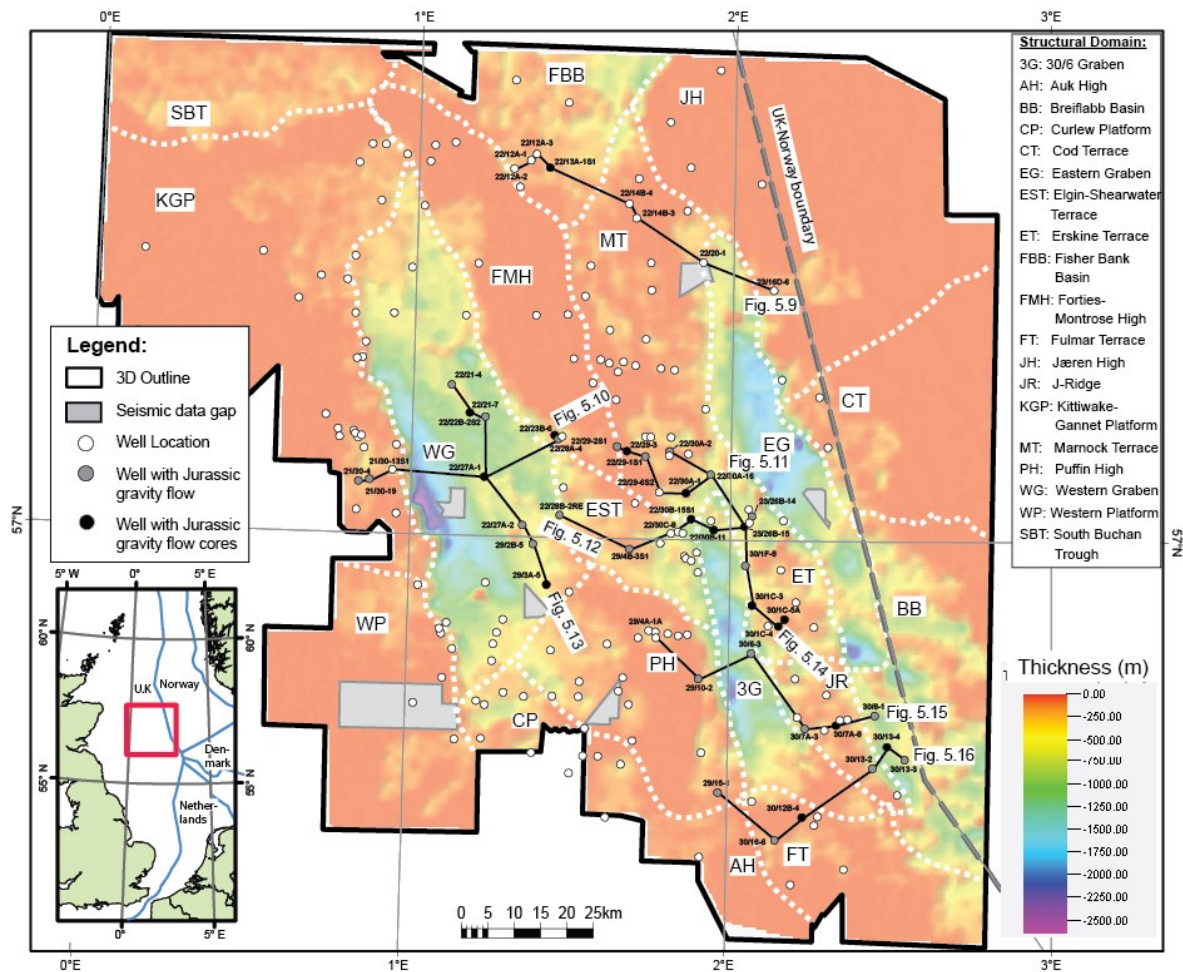


Fig. 5.2: Isochore map of the Jurassic interval, which has been interpreted from the 3D seismic data (c. 22,360 km²). Superimposed on this map is the distribution of the wells used in this evaluation highlighting wells that encountered gravity flow sediments (grey circles) and conventional core examined (black circles). The Late Jurassic structural domains defined in this study are outlined by the white dotted lines, which have been derived mainly from the underlying structural pattern of the Top Lower Permian as well as the Jurassic interval. Note that only wells mentioned in this paper are being named here.

Cluster	Basin	Field example	Type (after Reading & Richards, 1994; Fig. 5)	Age (Stage)	Local Name	Gross Sand thickness (m)	Length (km)	Width (km)	Av. Net to gross (%)	Average Porosity (%)	Average Permeability (mD)	Structural setting	Main References
Brae-Miller	South Viking Graben	Central Brae	Gravel-rich slope-apron	Kimmeridgian to Early Tithonian	Brae Fm	200-500	5	2	60	12	100	Half Graben Hangingwall	Fletcher (2003b)
		North Brae	Gravel-rich slope-apron	Kimmeridgian to Tithonian	Brae Fm	240-350	8	2	85	15	300	Half Graben Hangingwall	Brehm (2003)
		South Brae	Gravel-rich slope-apron	Kimmeridgian to Middle Tithonian	Brae Fm	0-500	4	3	75	12	131	Half Graben Hangingwall	Fletcher (2003a)
		Toni	Gravel-rich slope-apron	Late Oxfordian to Early Tithonian	Brae Fm	180-220	5	3	45	11	150	Half Graben Hangingwall	Cherry (1993), Kerlogue et al (1995)
		Tiffany	Gravel-rich slope-apron	Late Oxfordian to Early Tithonian	Brae Fm	240-260	5	3	79	10	75	Half Graben Hangingwall	Cherry (1993), Kerlogue et al (1995)
		Thelma	Gravel-rich slope-apron	Early Tithonian	Brae Fm	180-300	5	3	52	12	500	Half Graben Hangingwall	Cherry (1993), Gambaro & Donagemma (2003)
		East Brae	Sand-rich submarine fan	Kimmeridgian to Middle Tithonian	Brae Fm	300-350	n/a	n/a	85	15	500	Half Graben Hangingwall	Branter (2003)
		Miller	Sand-rich submarine fan	Oxfordian to Middle Tithonian	Brae Fm	700-800	15	10	85	17	1500	Half Graben Hangingwall	Garland (1993), Fraser et al (2003)
		Kingfisher	Sand-rich submarine fan	Callovian to Middle Tithonian	Brae Fm	300-500	n/a	n/a	70	n/a	n/a	Half Graben Hangingwall	Fraser et al (2003)
Gudrun	South Viking Graben	Gudrun	Sand-rich submarine ramp	Late Oxfordian to Middle Tithonian	Draupne Fm	100-300	10	5	70	25	n/a	Tilted fault block	Jackson et al (2011)
Beryl	Central Viking Graben	Beryl	Mixed sand-mud submarine fan	Oxfordian	Katrine Unit	90	n/a	n/a	43	15	150	Half Graben Hangingwall	Robertson (1993)

Cluster	Basin	Field example	Type (after Reading & Richards, 1994; Fig. 5)	Age (Stage)	Local Name	Gross Sand thickness (m)	Length (km)	Width (km)	Av. Net to gross (%)	Average Porosity (%)	Average Permeability (mD)	Structural setting	Main References
Magnus	Magnus Trough	Magnus	Sand-rich submarine fan	Late Oxfordian to Tithonian	Ptarmi-gan & Magnus Mbr	150-200	20	10	85	21	80	Half Graben	Shepherd (1991), Ravnås & Steel (1997), MacGregor et al (2005)
Greater Horda	North Viking Graben	Oseberg	Mixed sand-mud submarine fan	Late Oxfordian to Tithonian	Draupne Fm	80	n/a	n/a	10	n/a	n/a	Tilted fault block	Ravnås & Steel (1997), Ravnås & Bondevik (1997)
Galley	Witch Ground Graben	Alder	Sand-rich submarine fan	Tithonian	Galley Mbr	200-350	20	10	60	n/a	n/a	Graben	Fraser et al (2003)
		Saltire	Sand-rich submarine fan	Late Kimmeridgian	Galley Mbr	60	15	5	80	15	1190	Half Graben Hangingwall	Casey et al (1993)
		North Galley	Sand-rich submarine fan	Tithonian	Galley Mbr	100-150	20	10	60	n/a	n/a	Graben	Fraser et al (2003)
		South Galley	Sand-rich submarine fan	Tithonian	Galley Mbr	200	20	10	60	n/a	n/a	Graben	Fraser et al (2003)
Claymore	Witch Ground Graben	Claymore	Sand-rich submarine fan	Early Tithonian	Claymore Mbr	400-420	15	10	80	n/a	n/a	Half Graben Hangingwall	Moseley et al (1993), Fraser et al (2003)
		Tartan	Sand-rich submarine fan	Early Tithonian	Claymore Mbr	70-80	15	5	60	n/a	n/a	Half Graben Hangingwall	Moseley et al (1993), Fraser et al (2003)
Outer Moray Firth	Outer Moray Firth	Buzzard	Sand-rich submarine fan	Late Oxfordian to Tithonian	Buzzard Mbr	90	10	5	70	24	9000	Graben	Doré & Robbins (2005)
Fulmar	Central Graben	Fulmar	Mixed sand-mud submarine ramp	Late Kimmeridgian	Ribble Mbr	3-50	5	1.5	82	25	3000	Salt minibasin	Johnson et al (1986), Kuhn et al (2003)

Table 5.1: This and previous pages - Proven fields containing gravity flow deposits in the North Sea. For examples of wireline logs, see Fig. 5.5b.

Central Graben Geological Domains	Well Name	Length of core with gravity flow (m)	Stratigraphy sampled
Elgin-Shearwater Terrace	22/30B-11	12.8	Sequence II
	22/30B-15S1	6.1	Sequence II
Erskine Terrace	23/26B-15	1.8	Sequence II
	30/01C-4	0.6	Sequence I
Fisher Bank Basin	22/13A-1	36.1	
Forties-Montrose High	22/23B-6	13.1	Sequence IV
Fulmar Terrace	30/12B-4	22.6	Sequence III & IV
J-Ridge	30/07A-8	10.7	Sequence V
	30/13-4	27.1	Sequence II
Marnock Terrace	22/29-3	8.2	Sequence II
	22/29-6S2	18.3	Sequence II
	22/30A-1	11	Sequence II
Western Graben	22/21-7	10.8	Sequence II
	22/27A-1	6.1	Sequence II
Total core length		164.4	

Table 5.2: List of cores containing gravity flow sediments from wells independently described in this study according to geological domain. For location of these wells, see Fig. 5.2.

Tectono-stratigraphic evolution of the Central Graben

The North Sea has undergone multiple tectonic events since the Palaeozoic, which has resulted in a complex structural configuration (Chapter 3; Coward et al., 2003; Erratt et al., 2010; Lyngsie et al., 2006; Ziegler, 1990a; Ziegler, 1990b; Ziegler, 1992). The main Mesozoic structural framework comprises three rift-related basins that formed immediately after formation of the Early Jurassic Mid-North Sea Dome (Fig. 5.1): (1) Viking Graben; (2) Central Graben (CG); and (3) Moray Firth Graben (Rattee and Hayward, 1993; Underhill and Partington, 1993; Ziegler, 1990a). The Central Graben is the southernmost and most complex of the three basins due to its long tectonic evolution and the occurrence of Permian salt within the basin, which was periodically mobile during the Mesozoic and Cenozoic.

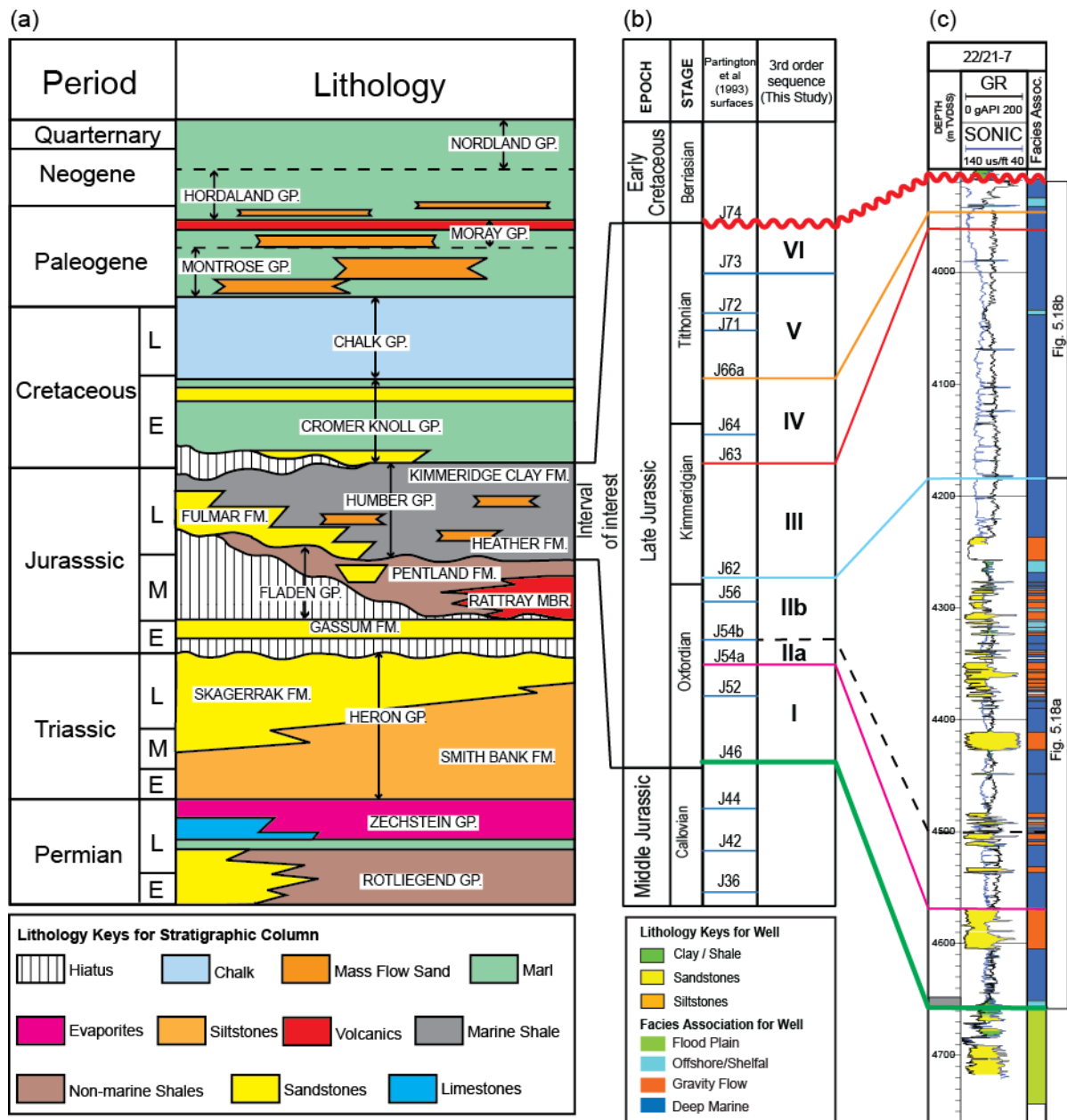


Fig. 5.3: Stratigraphy of the Central North Sea. (a) Simplified stratigraphic column (after Vining et al., 2005). (b) Upper Jurassic stratigraphic column showing the current stratigraphic sequence based on genetic stratigraphy and compared to Partington et al (1993) biostratigraphic scheme. (c) Wireline log from well 22/21-7 showing example of gravity flow deposits in the Central Graben Basin. For facies description, see text. GP: Group, FM: Formation, MBR: Member.

Pre-Mesozoic period

The oldest strata penetrated by wells in the CG are Devonian, consisting mainly of continental clastic deposit of the Old Red Sandstone (Marshall and Hewett, 2003; Ziegler, 1992). The basin was

periodically submerged during marine transgression in the early Carboniferous, which gave rise to marine sandstone and carbonate deposition (Bruce and Stemmerik, 2003; Coward, 1993). A subsequent compressional and/or transpressional tectonic regime caused uplift of these basins during the Late Carboniferous to Early Permian, resulting in denudation down to the Devonian in some parts of the CG leaving the Permian to be deposited unconformably above the Devonian (Heeremans and Faleide, 2004; Martin et al., 2002).

Deposition in the CG resumed during the Late Permian, when semi-arid, continental depositional environments (preserved as the Rotliegend Group; Fig. 5.3a) extended throughout much of NW Europe, potentially in a series of active rifts (Geluk, 1999; Stemmerik et al., 2000). The basin was partially submerged during the latest Permian, following the creation of a seaway connecting the CG with the Norwegian-Greenland rift in the north. This resulted in the extensive deposition of evaporitic rocks throughout both the Northern (CG) and Southern Permian Basins (Zechstein Supergroup; Fig. 5.3a; Glennie and Buller, 1983; Ziegler, 1990a; Ziegler, 1990b; Ziegler, 1992).

Mesozoic Period

The Mesozoic period started with a major rifting phase in the Triassic (Coward, 1995; Glennie, 1995), when the basin was filled by syn-rift, semi-arid, continental deposits (Heron Group; Fig. 5.3a) (Goldsmith et al., 2003; McKie and Audretsch, 2005). This period also coincided with the commencement of halokinesis of Zechstein salt, which resulted in the formation of minibasins, or pods, between rising salt diapirs (Hodgson et al., 1992; Penge et al., 1993; Smith et al., 1993). Marine conditions were re-established at the end of Triassic to Early Jurassic, following a major marine incursion from the Boreal Ocean in the north (Goldsmith et al., 2003).

Another period of uplift occurred in the CG during the Early to Middle Jurassic; this resulted in widespread erosion and/or non-deposition of Lower Jurassic sediment (Underhill and Partington, 1993). Lower Jurassic deposition continued beyond the limits of the area of uplift (the so-called Mid-North Sea Dome), such that marine strata of the Dunlin and Brent groups were deposited further north in the Viking Graben (Husmo et al., 2003).

Another phase of rifting in the CG followed collapse of the Mid-North Sea Dome. The initiation of this final phase of rifting may have been as early as the Late Bathonian (Davies et al., 1999) but, due to limitations in both biostratigraphic and seismic data, slightly later (Callovian or Oxfordian) dates have also been proposed (Chapter 4; Coward et al., 2003; Erratt et al., 1999; Rattey and Hayward, 1993). During this period of rifting, the location and kinematics of newly formed structures were variably controlled by older structural trends (e.g. Palaeozoic and/or earlier Mesozoic), which resulted in the formation of a structurally complex basin defined by numerous smaller, semi-isolated sub-basins (Fig. 5.2). In addition, this extension regime also caused further flow of Zechstein salts, which resulted in the initial growth and later collapse of a series of salt diapirs (e.g. Mannie et al., 2014). As a result of this complex structural setting, the seabed bathymetry was extremely variable; for example, seabed-breaching structures (e.g. normal faults, diapirs) represented barriers to bedload transportation, whereas subsidence in the hangingwalls of the same structures provided accommodation for sediment accumulation, the latter being especially active during the accelerated period of rifting in the Kimmeridgian and Tithonian (Clark et al., 1999; Erratt, 1993; Hodgson et al., 1992; Stewart et al., 1999). Rifting was coeval with a Late Jurassic phase of eustatic sea level rise, which resulted in the deposition of a net-transgressive marine succession, ranging from paralic sediments along the rift flanks to deep-water sediments in the basin centre (Humber Group; Fig. 5.3a; Fraser et al., 2003).

As the result of extensional faulting and syn-rift halokinesis, the CG becomes structurally-compartmentalised into the various sub-basins during the Late Jurassic to Early Cretaceous (Fig. 5.2; see also Chapter 3). The tectono-stratigraphic framework of the CG shows that the Late Jurassic basin-fill comprises a regionally-extensive and highly diachronous retrogradational succession, with superimposed higher-order regressive-transgressive cycles (Fraser et al., 2003). The larger-scale retrogradational succession comprises three main elements (Fig. 5.3): (1) coastal plain deposits ('Pentland Formation'-type facies) that infill a complex, partly remnant topography, initially created during Middle Jurassic North Sea doming and further influenced by synchronous faulting, (2) coastal-shelf deposits (mainly shoreface sediments of 'Fulmar Formation'-type facies and shelfal equivalent mudrocks of 'Heather Formation'-type facies), which overlie coastal plain successions either gradationally (e.g. in basin centre locations) or erosionally (e.g. in basin margin and footwall crest locations), and (3) mud-dominated deeper water basinal deposits ('Kimmeridge Clay Formation'-type facies) containing subordinate intercalations of sand-rich gravity flows (variably named as members, such as the Ribble Member in the Fulmar Field). This large-scale succession reflects regional deepening, seaway expansion and shoreface retreat (Fig. 5.4).

Rifting ceased at the end of Jurassic. Along the basin flanks, where less rift- or salt-related subsidence had occurred, a regional unconformity is identified above eroded, typically Mesozoic strata (the Base Cretaceous Unconformity, or 'BCU'). In the adjacent basins, where subsidence and sedimentation had been more-or-less continuous, this surface is more conformable and represented by a broadly continuous stratigraphic succession (Kyrkjebø et al., 2004; Rawson and Riley, 1982).

Post-rift deposition continued into the Early Cretaceous with transgressive shallow-marine mudstone accumulating throughout the basin (Cromer Knoll Group; Fig. 5.3a), followed by widespread, deeper water conditions. This resulted in the shutdown of siliciclastic sediment supply

throughout the region, which was followed by chalk deposition during the Late Cretaceous (Chalk Group; Fig. 3a) (Copestake et al., 2003; Surlyk et al., 2003). The basin was also subjected to several inversion events in the Late Cretaceous related to the earliest phases of the Alpine orogeny, which reactivated some of the major basin-bounding faults and caused rejuvenation of existing salt structures (Coward and Dietrich, 1989).

Post-Mesozoic period

The start of the Tertiary saw the re-commencement of coarse clastic sediment supply into the basin in the form of deep-water submarine fans (Montrose & Moray groups; Fig. 5.3a). Coarse clastic sediment input was caused by basin margin uplift, most notably in NW Scotland, as a result of underplating associated with the opening of the North Atlantic Ocean (Ahmadi et al., 2003; MacLennan and Lovell, 2002; MacLennan et al., 2003; Mudge, 2014; Mudge and Jones, 2004; White, 1989). The rate of uplift gradually declined and during the remainder of Tertiary, from the Late Palaeocene onwards, the basin continuing to subside in a deep-water environment characterised by deposition of mainly deep-water mudstones and minor turbidite sandstones (Hordaland & Nordland groups; Fig. 5.3a) (Fyfe et al., 2003).

Summary of Upper Jurassic gravity flow deposits in the North Sea Basin

Upper Jurassic gravity-flow deposits, including high quality, turbidite reservoir sandstones, occur in the syn-rift intervals of most North Sea sub-basins (Fig. 5.1). These sandstones were mainly deposited in relatively deep-water, basinal settings. However, reservoir thickness, lateral extent and prospectivity are variable and unpredictable, with the play remaining somewhat 'enigmatic' (Erratt et al., 2010). Despite several notable early discoveries (e.g. in the Magnus and Brae fan systems) and positive predictions concerning the future potential from subtle traps within this deep-water play, widespread exploration success has been elusive (Boldy and Fraser, 1999; Cordey, 1993; Fraser et

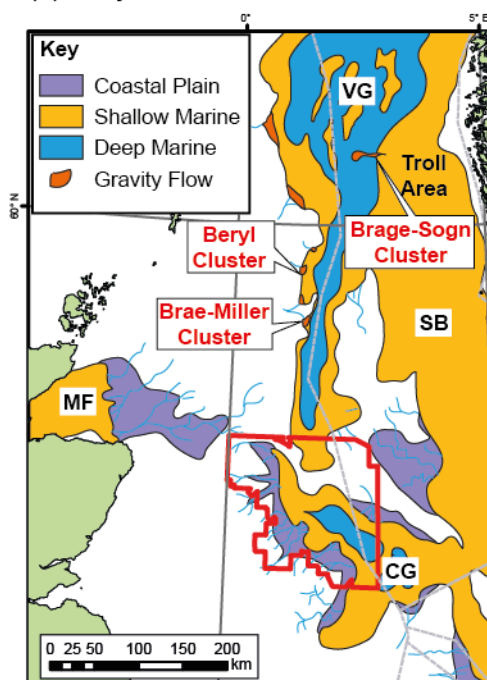
al., 2003; Munns et al., 2005; Partington et al., 1993b). This contrasts with the much greater prospectivity of the laterally extensive shallow-marine plays associated with the underlying, pre- to early syn-rift Middle (Brent) and Upper Jurassic (Fulmar and its lateral equivalents) successions (Johnson and Fisher, 1998). Deep-water fan systems deposited during Late Jurassic rifting and rapid basin subsidence accumulated within overall mud-dominated basins; as a result, these reservoirs are directly juxtaposed with or occur within organic-rich source rocks (Kimmeridge Clay and Draupne formations), which may also act as vertical and lateral seals and thus enhance the potential for stratigraphic trapping. Combination stratigraphic-structural traps are common in relatively old producing fields such as those occurring along the 'Brae Trend' in the South Viking Graben (Roberts, 1991), and in more recent discoveries such as the Buzzard Field in the Outer Moray Firth (Doré and Robbins, 2005). The unrealised promise of the Upper Jurassic play is reflected in the shape of the associated creaming curve (Eriksen et al., 2003; Munns et al., 2005; ØStvedt et al., 2005). The key to translating this potential into exploration success is linked to integrated tectonostratigraphic analysis, especially the availability of exploration-scale 3D seismic reflection data, as shown by the Upper Jurassic Buzzard Field, and the vertical extension of the play type into the Lower Cretaceous Kopervik Sandstone trends in, for example, the Outer Moray Firth (Doré and Robbins, 2005; Erratt et al., 2010; Wilson et al., 2005).

A key aspect of the complexity of this play is the highly variable nature of sand delivery, which ranges from footwall collapse ('fault scarp degradation complexes' comprising various slumps, slides and mass transport deposits; McLeod and Underhill, 1999; Yielding, 1990; Yielding et al., 1992), through local footwall erosion and resedimentation (e.g. slope apron and fans of the Brae Trend; Fraser et al., 2003), to more extensive systems typically linked to deltas (e.g. the Greater Horda Platform area and Troll Delta; Ravnås and Steel, 1997). Overall, Upper Jurassic fan systems were relatively small, the sub-basins were structurally-compartmentalised and the basin-fills were mud-dominated and relatively starved of sand-grade sediment. In addition, Upper Jurassic deep-water

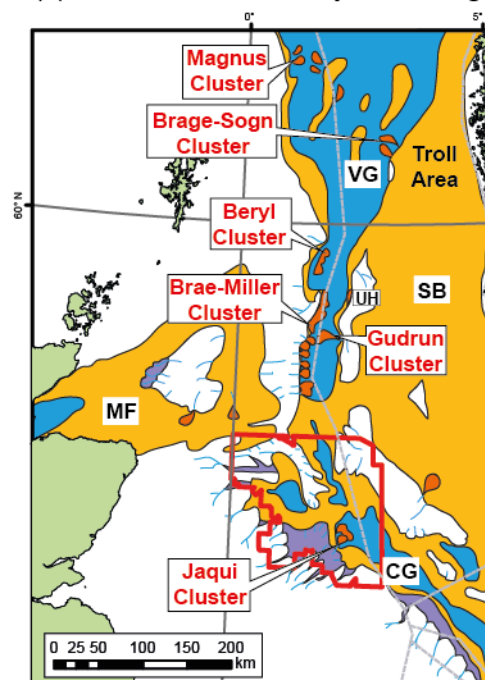
sandstones occur in a variety of geological settings, which can be summarised as follows (Figs. 5.1 & 5.4; approx. from north to south): (1) East Shetland Basin/Magnus fan system (e.g. Magnus Field; De'Ath and Schuyleman, 1981; Evans and Parkinson, 1983; Macgregor et al., 2005; Shepherd, 1991); (2) Troll area related fan systems (e.g. Fram, Brage and Oseberg fields; Fraser et al., 2003; Ravnås and Bondevik, 1997; Ravnås et al., 2000; Ravnås and Steel, 1997); (3) Beryl fan systems (Frostick et al., 1992; Richards, 1991; Robertson, 1993); (4) Gudrun fan systems (Jackson et al., 2011); (5) South Viking Graben fan systems (e.g. Brae, Tiffany/Toni/Thelma and Miller fans; Cherry, 1993; Stow, 1983; Stow et al., 1982; Turner and Allen, 1991); (6) Outer Moray Firth fan systems (e.g. Claymore, Galley and Buzzard fields; Witch Ground Graben/South Halibut Trough; Casey et al., 1993; David, 1996; Doré and Robbins, 2005; Harker and Rieuf, 1996); and (7) Central Graben (e.g. Kimmeridgian sands in the Shearwater, Erskine and Fulmar fields; Erratt et al., 2005; Jeremiah and Nicholson, 1999; Johnson et al., 1986; Taylor et al., 2015).

Upper Jurassic gravity-flow deposits were initially deposited during the onset of rifting in the latest Callovian to early Oxfordian around the Brae-Miller Cluster (Figs. 5.4a & 5.5b), which includes the T-block fields (Toni, Tiffany & Thelma) along the western margin of the South Viking Graben (Table 5.1; Figs. 5.4a; Fraser et al., 2003; Kerlogue et al., 1995). Gravity-flow deposits in the Brae-T-Block area are dominated by sand- and clay-matrix conglomerates and medium- to coarse-grained sandstones that form a slope apron (Fig. 5.5a) against the main syn-rift boundary fault of the strongly asymmetric South Viking Graben (Cherry, 1993; Fletcher, 2003a; Fletcher, 2003b; Gambaro and Donagemma, 2003; Reading and Richards, 1994). The sediment source area comprised Devonian and other pre-Jurassic rocks of the immediately adjacent and uplifted Fladen Ground Spur to the west (Fig. 5.1). The initial slope apron succession is succeeded by fine- to medium-grained sand-rich submarine fans, which extended further basinwards into the Miller-Kingfisher area (Fraser et al., 2003; Garland, 1993). Submarine fan deposition continued into the latest Jurassic (Fig. 5.4d).

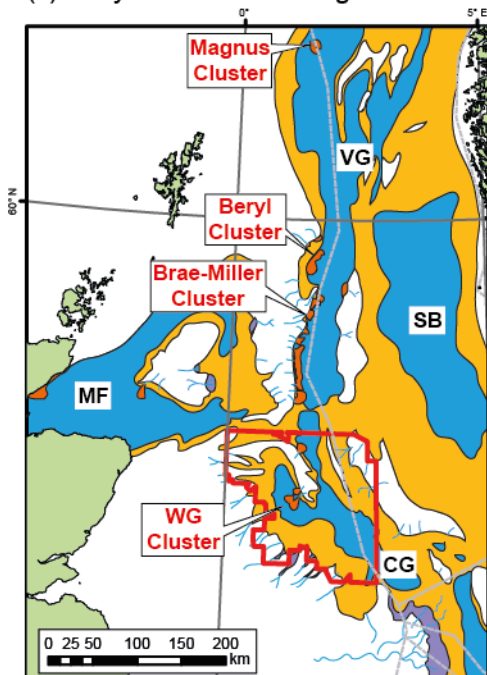
(a) Early to Late Oxfordian



(b) Late Oxfordian to Early Kimmeridgian



(c) Early to Late Kimmeridgian



(d) Late Kimmeridgian- Early Tithonian

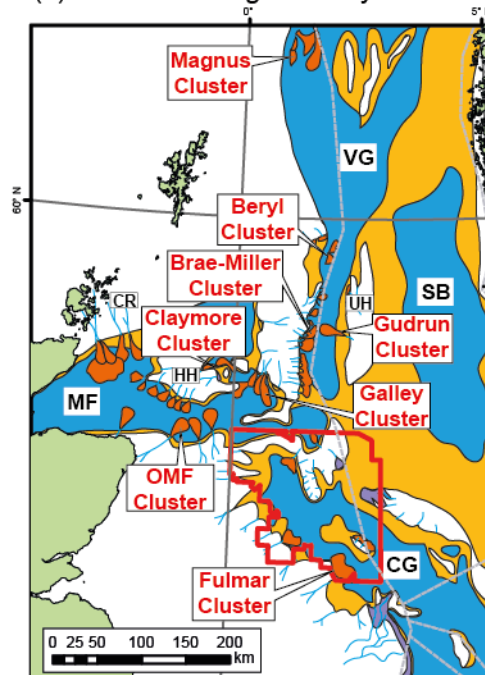


Fig. 5.4: Evolution of the depositional environment in the North Sea rift system in the Late Jurassic (modified from Fraser et al, 2003). (a) In the Early Oxfordian, three distinct basins radiating from the centre of North Sea Dome developed. The Viking Graben (VG) was connected to the main Atlantic rift in the north, hence become the earliest to contain deep basin. Gravity flow deposits started to appear in the Brage, Brae and Beryl areas. (b) By Late Oxfordian, the Central Graben (CG) also became connected to the sea from the south expanded its deep basin. More gravity flow deposits could be found along the western edge of the VG and sporadically along the eastern flank,

downdip of Troll area and the Utsira High (UH). Gravity flow deposits also started to appear in the CG. (c) In the Kimmeridgian period, deep marine environment also spread into the Moray Firth (MF) and Stord Basin (SB) while shallow marine environment transgressed onto the platform highs and connected various sub-basins in the CG. Gravity flow deposition continues in the south of VG and in Western Graben (WG) and started to appear in the MF. (d) Towards the end of the Late Jurassic, deep marine environment dominated the North Sea basins as rifting accelerated and global sea level increases. There was an increase of gravity flow deposition in the MF south of Caithness Ridge (CR) and Halibut Horst (HH).

Early Oxfordian gravity flows were deposited in a small, sand-rich submarine fan in the Beryl Cluster (Table 5.1; Fig. 5.4a); these fine-grained sandstones account for only 30% of the overall mud-dominated succession (Fig. 5.5b; Robertson, 1993). These sands were sourced from the East Shetland Platform, with deposition continuing into the latest Jurassic (Fig. 5.4d). The Brage-Sogn Cluster in the eastern corner of Viking Graben also contains early Oxfordian gravity-flows (Table 5.1; Fig. 5.4a). Here, the mixed sand-mud-prone fan system, which is located further away from the main boundary fault, is similar to that of the the Beryl Cluster (Fig. 5.5b; Ravnås and Steel, 1997). However, the provenance of this fan system was from the Greater Horda Platform area to the west (Fig. 5.1 & 5.4b), which was linked to the large Troll Delta (Holgate et al., 2013).

The Magnus fan system is located in the western part of the North Viking Graben and was initially deposited during the Late Oxfordian, with deposition continuing into the Early Tithonian (Table 5.1; Fig. 5.4b). This sand-rich submarine fan system consists of medium- to coarse-grained sandstones, which were derived from the Margareta Spur/Manet Ridge to the north (Ravnås and Steel, 1997). Similar age fans (Late Oxfordian to Middle Tithonian) occur in the Gudrun Cluster in the eastern side of the South Viking Graben (Table 5.1; Fig. 5.4a). This sand-rich submarine ramp system contains medium- to coarse-grained sand derived from the Utsira High to the east, with deposition occurring on the slope of a large, salt-influenced half-graben (Jackson et al., 2011).

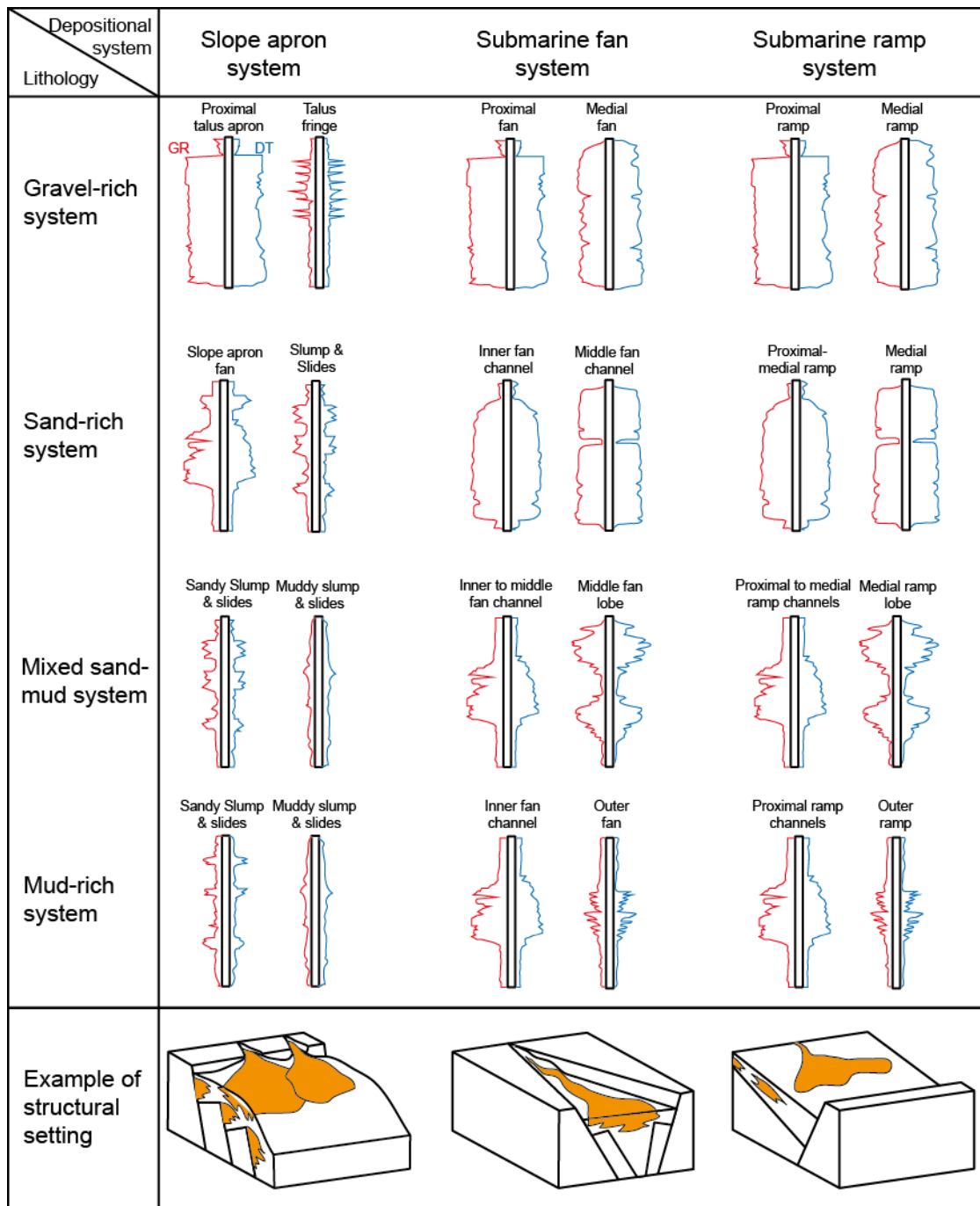
In the Outer Moray Firth, Late Oxfordian, fine to medium-grained turbidite sandstones form the reservoir in the Buzzard Field. These sandstones were largely derived from exposed Permian and Triassic rocks exposed in the Grampian Spur to the west (Doré and Robbins, 2005). However, in the Moray Firth the largest and most long-lived fan system is located in the Inner Moray Firth. Here, Kimmeridgian fans were derived from the Caithness Ridge and deposited on the Smith Bank High and in the South Halibut Basin, where they interfingered with fans originating from the Halibut Horst (Fig. 5.4d). The deposition of submarine fan system continued south of Halibut Horst until the Late Cretaceous (Aptian Kopervick sands), which saw the final phase of rifting in the Moray Firth (Harker and Rieuf, 1996).

The Witch Ground Graben also received significant quantities of gravity flow-deposited sediments, albeit somewhat later (i.e. Tithonian) than in other areas (Fig. 5.4d). In the Witch Ground Graben, two regions or 'clusters' of deep-water deposits are found; (i) the Claymore Cluster – this occurs in the western part of the Witch Ground Graben and comprises coarse grained sandstones sourced from Devonian rocks exposed on the Halibut Horst (Moseley, 1999); and (ii) the Galley Cluster - this is located in the eastern part of the Witch Ground Graben and comprises medium-grained sandstones sourced from the Fladen Ground Spur to the east (Casey et al., 1993; Fraser et al., 2003).

In the CG, the relatively simple and predictable overall retrogradational facies succession belies the much more complex distribution of gravity-flow deposits, especially with regards to the sand-rich, turbidite-deposited intervals, which are of variable age (Oxfordian to Tithonian; Fig. 5.3) and are often assigned member status (creating a complex lithostratigraphy), and whose geographical distribution is widespread: (1) Ribble Member in the Fulmar Terrace area (Johnson et al., 1986; Kuhn et al., 2003), (2) Freshney Sandstone Member in the J-Ridge area (Fraser et al., 2003; McArthur et al., 2011; Wonham et al., 2014), (3) Heather Sandstone Member in the Eastern Graben area

(Shearwater, Elgin-Franklin and Erskine fields) (Jeremiah and Nicholson, 1999; Lasocki et al., 1999; Wonham et al., 2014; Wonham et al., 2002), and (4) 'Unnamed' sandstone member in the Western Graben (this study).

(a)



(b)

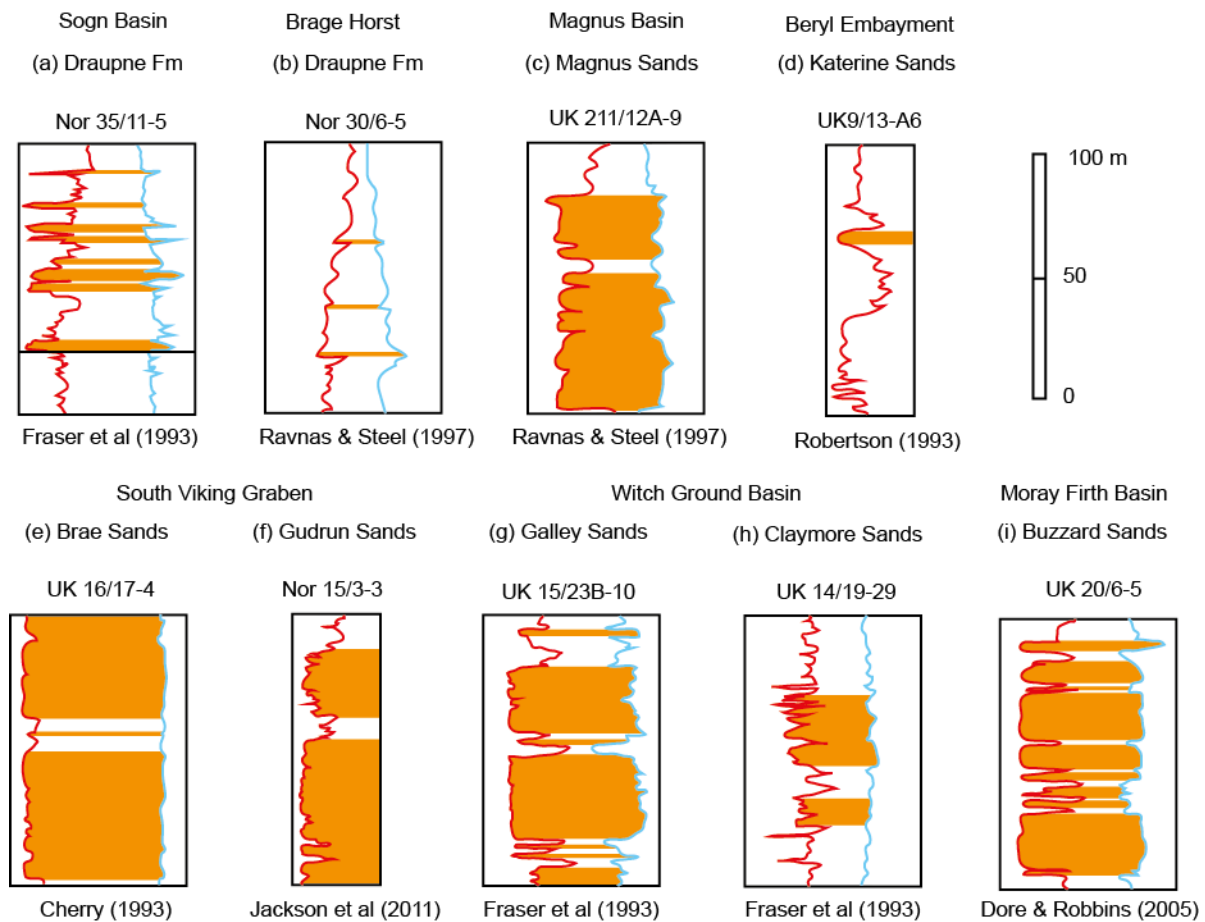


Fig. 5.5: Previous and this pages - Summary of wireline log characters for different types of gravity flow settings. (a) General log characters (after Reading and Richards, 1994). Each types have proximal and distal or medial character. (b) Log characters of selected gravity flow sands in the Viking Graben and Moray Firth Basin. GR: Gamma-ray log; DT: Sonic log.

In addition to the complex temporal and spatial distribution of these gravity-flow deposits, considerable debate surrounds the provenance of these sediments, with the following models most commonly proposed: (1) local, stratigraphically deep erosion of Triassic and older rocks from immediately adjacent footwall highs (e.g. the Ribble Sands in the Fulmar Field derived from the adjacent Auk Horst; gravity flows that rim the J-Ridge; Kuhn et al., 2003); (2) local, stratigraphically-shallow erosion related to shoreface collapse during tectonically-related 'transgressive shoreface reworking' (fautling and/or grounding of salt-related pods), which was triggered by a combination of

earthquakes, tropical storms and oversteepening of shoreface profiles (Franklin C Sandstone turbidites in Elgin-Franklin and an alternative hypothesis for the Ribble Sands in the Fulmar Field; Wonham et al., 2014); (3) local, stratigraphically-shallow erosion related to post-salt withdrawal shoreface collapse, with intra-pod turbidite accumulation (e.g. Jacqui turbidites; McArthur et al., 2011; Wonham et al., 2014); (4) regional, stratigraphically deep erosion that created lowstand fans linked to updip incised valleys with large drainage areas (Carruthers et al., 1996; Howell and Flint, 1996), including river-connected canyon systems with river mouth bypassing by hyperpycnal flows (e.g. a model proposed for the Buzzard Sands in the Ettrick Basin of the Outer Moray Firth; Doré and Forster, 2005; Wonham et al., 2014), and (5) regional drainage systems forming tectonically-active basin margin deltas, which resulted in prodelta-delta front collapse (e.g. an alternative model for the Buzzard Sands; Ray et al., 2010). Understanding the provenance of Upper Jurassic deep-water sandstones is important because this may exert a fundamental control on reservoir quality (e.g. high-quality sandstones may be derived from quartz-rich, 2nd-cycle material, whereas poorer-quality reservoirs are derived from feldspathic, 1st-cycle material) and the volume of deep-water sandstone (e.g. large volumes of sand may be derived from large, basin-margin uplands, whereas intra-basin structural highs may yield only limited volumes of sand).

Data and Methodology

A total of 164 m of core from 15 wells, which contain a range of Upper Jurassic gravity flow deposits, have been independently described in this study (Fig. 5.2, Table 5.2). In addition, a further 23 wells (out of a total of 200 wells analysed across the whole study area) have also been evaluated in terms of their wireline log and lithology (drill cuttings data) characteristics (Fig. 5.2, Table 5.3). Regional knowledge was used to select core and well log data representative of the different gravity-flow facies types and occurrences encountered in the Upper Jurassic of the CG. In addition to the Facies

Associations (FAs) identified and described in Chapter 4 (Table 4.5), seven sedimentary facies have been described in detail for the Gravity Flow FA in this study.

Based on core and wireline logs from 200 wells and an extensive biostratigraphic database, the Upper Jurassic of the CG was divided into six major sequences, mainly bounded by basin-wide major flooding surfaces (see Chapter 4; Figs. 4.4 & 5.3). These surfaces were correlated to the maximum flooding surfaces (MFS) of Partington et al. (1993a): (1) Sequence I, from the Base Upper Jurassic to J54a MFS; (2) Sequence II, from the J54a MFS to J62 MFS; (3) Sequence III, from the J62 MFS to J63 MFS; (4) Sequence IV, from the J63 MFS to J66a MFS; (5) Sequence V, from the J66a MFS to J73 MFS; and (6) Sequence VI, from the J73 MFS to the Top Jurassic (Fig. 5.3b). In addition, Sequence II has been locally divided further into Sequences IIa and IIb, mainly where Sequence II is sufficiently thick to enable the intra-sequence J54b flooding surface (Top Sequence IIa) to be correlated, at least within specific structural domains (e.g. Figs. 5.9-16).

Where seismic resolution permits, the sequences and their bounding surfaces have been identified and mapped within the 22,360 km², 3D seismic reflection 'mega-survey' (Fig. 5.2). The main limitation was the variable quality of the seismic data and its limited vertical resolution (50 m; based on a calculated interval velocity of 4300 m/s and peak frequency of c. 20 Hz), especially at significant (>2 km) burial depths. Less commonly, where the data contains multiples, the Upper Jurassic stratigraphy can be completely masked. The main consequence of these seismic limitations is that only three Upper Jurassic stratigraphic surfaces have been mapped throughout the basin (Fig. 5.3b): (1) Top Jurassic or Base Cretaceous Unconformity (BCU); (2) Top Sequence II (Top Oxfordian); and (3) Top Middle Jurassic. This enabled two isochore maps to be generated for the analysis of syn-depositional deformation on basin physiography and accommodation development, and stratigraphic thickness patterns and internal architecture.

Structural Domain	Well	Sequence	Core	Interpreted lithology from wireline log and cores (where available)
30/6 Graben	30/06-3	II	N	Blocky sandstones between clays
Breiflabb Basin	30/08-1	II	N	Blocky sandstones between clays
	30/13-3	II	N	Blocky sandstones between clays
	30/13-4	II	Y	Blocky, massive sandstones with slumps and clay fragments
Elgin-Shearwater Terrace	22/28B-2RE	II	N	Fining upwards sandstones between clays
	22/30B-11	II	Y	Massive sandstones between shales & bioturbated sands
		III	N	Blocky to fining upwards sandstones within clays
	22/30B-15S1	II	Y	Massive sandstones between shales & bioturbated sands
Erskine Terrace	23/26B-14	III	N	Thin sandstones between clays
	23/26B-15	III	Y	Massive sandstones within bioturbated clays
	29/04B-3S1	II	N	Blocky sandstones between clays
		III	N	Blocky sandstones between clays
	30/01C-3	II	Y	Massive sandstones within biturbated siltstones
	30/01C-4	I	Y	Cross-bedded sandstones within bioturbated clays and non-bioturbated shales
	30/01C-5A	II	Y	Laminated sandstones between bioturbated clays and non-bioturbated shales
Fisher Bank Basin	22/13A-1S1	V	N	Blocky sandstones between clays
		IV	Y	Laminated blocky sandstones between non-bioturbated shales
Forties-Montrose High	22/23B-6	IV	Y	Massive & laminated sandstones within bioturbated sands and siltstones
		V	Y	Massive & laminated sandstones bioturbated sands and siltstones
	22/28A-4	V	N	Blocky sandstones between clays
Fulmar Terrace	30/12B-4	III	Y	Massive sandstones with ripple and slump
		IV	Y	Massive sandstones with ripple and slump
	30/16-6	III	N	Fining upwards sandstones between clays
J-Ridge	30/07A-3	II	N	Blocky & fining upwards sandstones between clays
	30/07A-8	IV	Y	Clast-rich muddy sandstone (debrite)
	30/13-2	III	N	Blocky sandstones between clays
Kittiwake-Gannet Platform	21/30-19	IV	N	Thin, blocky sandstones between clays
	21/30-4	IV	N	Blocky sandstones within clays
Marnock	22/29-2S1	II	N	Thin sandstones between clays

Structural Domain	Well	Sequence	Core	Interpreted lithology from wireline log and cores (where available)
Terrace	22/29-3	II	Y	Massive & laminated sandstones with pebble bed
	22/30A-1	II	Y	Massive sandstones within bioturbated silty sandstones
	22/30A-16	II	N	Blocky sandstones between clays
	22/30A-2	II	N	Thin sandstones between clays
Puffin High	29/10-2	III	N	Fining upwards sandstones between clays
	29/15-1	III	N	Fining upwards silty sandstones between clays
Western Graben	22/21-4	III	N	Blocky sandstones between clays
	22/21-7	I	N	Thick & blocky sandstones between clays
		II	Y	Massive, laminated and rippled sandstones within shales
	22/22B-2S2	II	N	Blocky to fining upwards sandstones within clays
	22/27A-1	III	N	Blocky sandstones
		IV	Y	Massive sandstones with mud clasts between bioturbated clays
	22/27A-2	V	N	Blocky calcareous sandstones within clays
	29/02B-5	V	N	Thin sandstones between clays
	29/03A-5	II	Y	Massive medium-grained sandstones

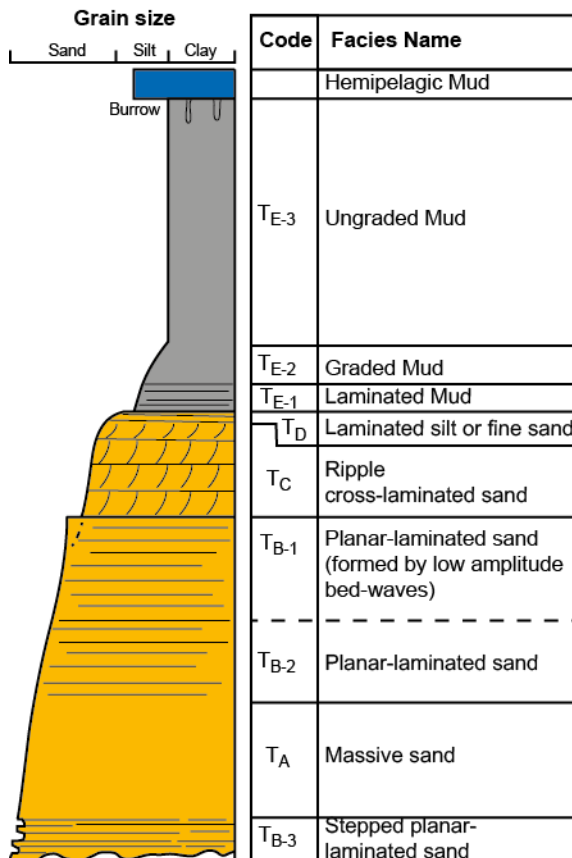
Table 5.3: Previous and this page - Wells in the Central Graben with gravity flow deposits identified from core and/or wireline logs. Where available, core interpretation would include common sedimentary structures. See Fig. 5.2 for well location.

Sedimentology of Upper Jurassic gravity flow deposits in the Central Graben

In order to provide consistency with established gravity-flow facies terminology and classification, we have used the scheme of Talling et al. (2012) (Fig. 5.6). Core descriptions resulted in the recognition of the following seven sedimentary facies in the gravity flow facies association: (1) massive (mainly ungraded) mudstone (mainly T_{E-3} with minor T_{E-2}); (2) laminated mudstone (T_{E-1}); (3) laminated siltstone (T_D); (4) current ripple cross-laminated sandstone (T_C); (5) laminated sandstone (T_{B-1} & T_{B-2}); (6) massive sandstone (T_A and/or D_{CS}); and (7) clast-rich, muddy sandstones (D_{M-2}). See

Fig. 5.6 for the abbreviations/bed code nomenclature and Fig. 5.7 for representative core photographs, and Table 5.4 for a summary of the main features and process interpretation of the facies analysed in this study.

(a) Turbidite facies



(b) Debrite facies (additional to (a))

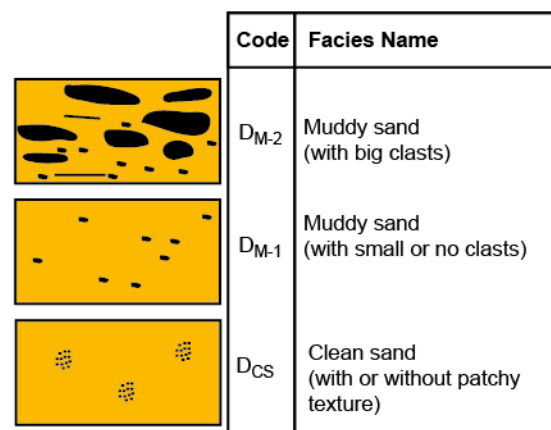


Fig. 5.6: Summary of gravity flow facies (after Talling et al., 2012). (a) Facies deposited by turbidity flow. In this classification, TD facies may or may not be easily distinguished from TE-1 due to the small change in modal grain size. (b) Facies deposited by debris flow. These are usually deposited below the TE-1 or TD facies.

Having recognised the key sedimentary facies/facies associations in core, the core-defined facies have been calibrated to their wireline log response in order to interpret uncored intervals. In this study, the gamma-ray (GR) and sonic logs (DT) were deliberately placed in the same track using a gamma-ray scale of 0 to 200 API and a sonic scale of 40 to 140 $\mu\text{s}/\text{ft}$, thus enhancing the

differentiation of the deep-marine shale facies association from the other two, more sandstone- and/or siltstone-rich facies associations (Fig. 5.8a). By using this scale, organic-rich deep-marine shales (Kimmeridge Clay Formation) correspond to intervals in which the sonic log deflects to the left of the gamma-ray log (Fig. 5.8b); this contrasts with shallow-marine, non-organic-rich shales, which corresponds to the sonic log being clearly separated from, and positioned to the right of, the gamma-ray log (e.g. Fig. 5.8g).

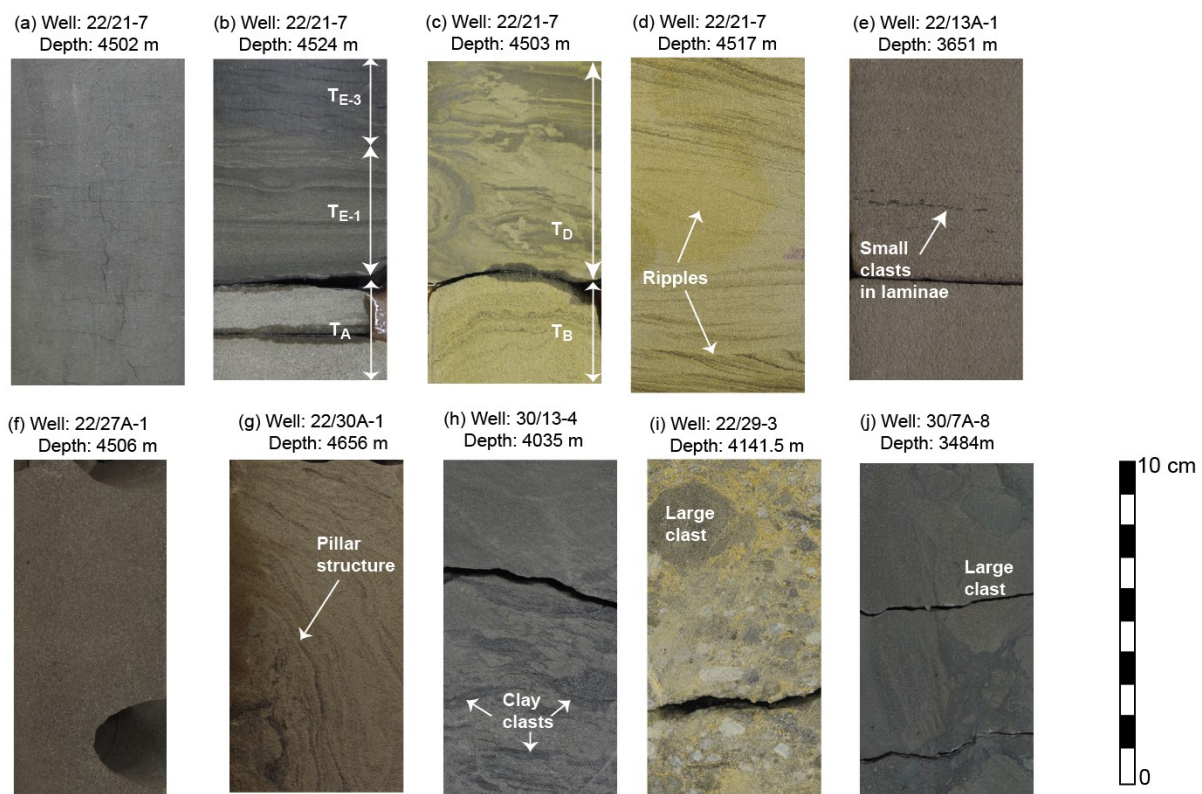


Fig. 5.7: Core photographs showing examples of sedimentary structures in gravity flow deposits in the Central Graben. (a) Massive mudstone (T_{E-2} or T_{E-3}). (b) Laminated mudstone (T_{E-1}) between T_{E-3} and T_A facies. (c) Slumping in laminated mudstone/siltstone (T_D). (d) Ripple-laminated sandstone (T_C). (e) Laminated sandstone facies that includes mm-scaled mud clasts (T_{B-1} or T_{B-2}). (f) Massive structureless sandstone facies (T_A or D_{CS}). (g) Water escape feature (pillar structure) in massive sandstone facies (T_A or D_{CS}). (h) Shale clasts in massive sandstone (T_A). (i)-(j) Mudstone and sandstone clasts in muddy sandstone matrix (D_{M-2}). See Fig. 5.8 for location in wells.

Massive mudstone facies (T_{E-2} and T_{E-3})

This facies occurs in 2 to 4-m-thick intervals lacking sedimentary structures or bioturbation (Fig. 5.7a and Table 5.4). This facies rarely contains relatively thin (<10m) sandstones although, overall, sandstone content is negligible. Talling et al (2012) differentiate ungraded massive mudstone (T_{E-3}) and graded massive mudstone (T_{E-2}) (Fig. 5.5), but these two facies types are indistinguishable in the available core data. In wireline-log data, this facies is characterised by GR > 80 API and DT > 90 μ s/ft, with the two tracks typically not crossing (Fig. 5.8a; Table 5.4). These facies are usually sandwiched between coarser (siltstone to sandstone) facies and could have sharp contact with the overlying facies type but, in addition, have a transitional contact with the underlying facies.

The massive, non-laminated and non-bioturbated nature of these mudstones suggest they were deposited rapidly from a gel or from a relatively high concentration, fluid mud layer, possibly from mud-rich, cohesive debris flows (Talling et al., 2012). In contrast, graded massive mud layers imply some segregation-settling during deposition, whereby coarser mud and silt settles out preferentially from gels that can no longer support coarser materials. In addition, observations from other study areas have indicated that both of these facies are often deposited in large flows that extended across basin plains or are confined to intra-slope sub-basins, including ponded topographic lows (Talling et al., 2012).

Laminated mudstone facies (T_{E-1})

This facies comprises mudstones with mm-scale silty laminations (Fig. 5.8b), occurring in intervals usually <1 m thick. This facies often underlies the massive mudstone facies and is usually bounded by sharp contacts. This facies typically overlies sandy gravity flow deposits (Fig. 5.6). The wireline log response of this facies is similar to that of the massive mudstone facies described above, since the mineralogical composition of the mud appears broadly similar at the hand-specimen scale (Table

5.4). Furthermore, the mm-scale lamination identified in this facies is far below the vertical resolution of most logging tools.

Laminated siltstone facies (T_D)

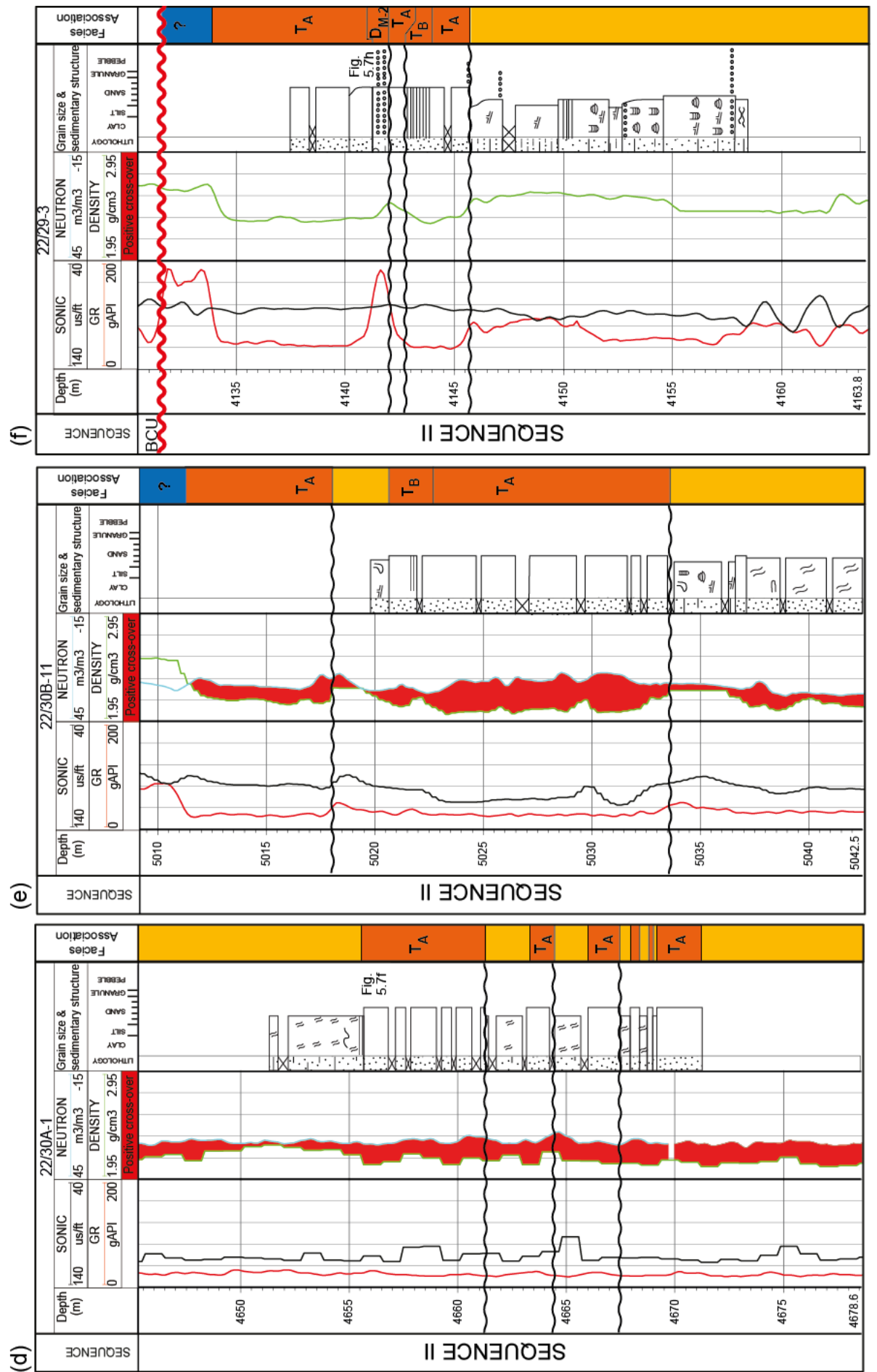
The laminated siltstone facies comprises finely laminated, mud- to silt-grade sediments occurring in <1 m thick beds. This facies commonly gradually overlies very fine-grained sandstone (Fig. 5.7c; 5.8a), but slumping may cause sharp basal contacts with underlying sandstones (Fig. 5.7c). In some cases, it may be difficult to differentiate muddy, laminated siltstone (T_D) and the overlying silty laminated mudstones (T_{E-1}), especially where the former is very thinly developed. Hence, it is not uncommon for this facies (T_D) to be classified as part of the T_{E-1} facies (Talling et al., 2012). This facies has a serrated character on wireline logs, with moderate (40-80 API) gamma ray values that do not cross-over with the sonic log. Density and neutron logs show a negative separation (Table 5.4).

The following interpretations have been proposed for this facies: (1) deposition from lower stage plane bed conditions, (Harms and Fahnestock, 1965; Walker, 1965); and (2) deposition from dilute, waning flows, when the duration of bedload reworking is too short to allow the formation of ripples or dunes (Lowe, 1988). Confidently making a depositional process-based interpretation is difficult because of a lack of unequivocal calibration between model results and experimental observations (Talling et al., 2012). However, T_D facies is retained to denote the non-cohesive part of dilute gravity flow that differs from the overlying, mainly cohesive part (T_{E-1}).

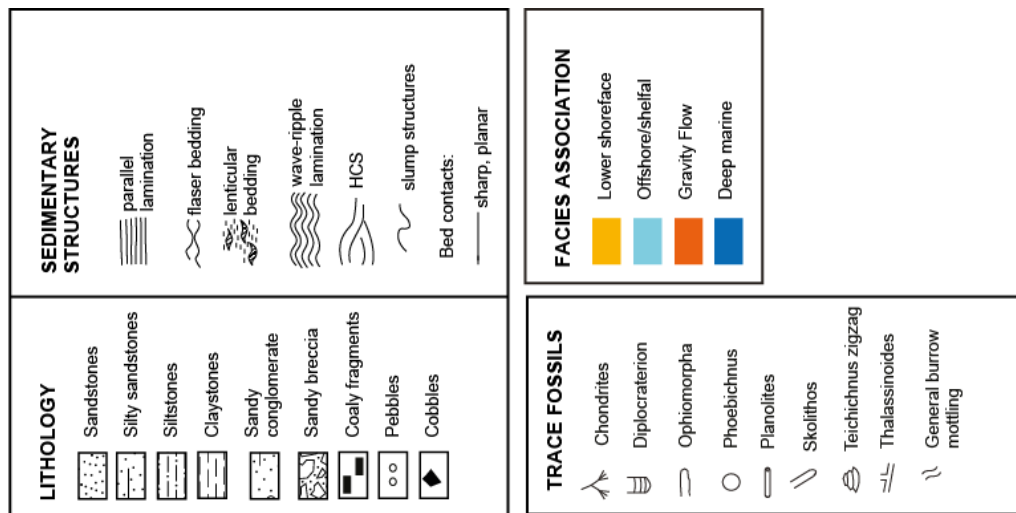
Facies	Bed thickness	Sand content	Lithology and sedimentary structures	Process interpretation & environment	Wireline log character
Massive mudstone (T _{E-3} or T _{E-2} ; Fig. 5.7a)	2-4 m	0%	Mudstone with no apparent lamination or grading. Usually massive and occasionally may be graded.	Transported by debris flow and deposited <i>en masse</i> from gels or high concentration mud layer (T _{E-3}) and segregation settling (T _{E-2}). Indistinguishable in this study.	Blocky & serrated. GR>80 API; DT>90 μs/ft; Density>2.4 g/cm ³ ; neutron>30 v/v. Slight or no overlap between GR and DT logs. Have negative separation between density and neutron logs (Fig. 5.8a).
Laminated mudstone (T _{E-1} ; Fig. 5.7b)	~1 m	0%	Mudstone with stripes of siltstones (mm-scaled). Lenticular lamination is common and bioturbation is rare.	Transported by turbulent flow current and deposited when denser flocs break up from the flow.	Same as above. Usually very difficult to distinguish in wireline logs. (Fig. 5.8a)
Laminated siltstone (T _D ; Fig. 5.7c)	<1 m	0-1%	Siltstone with stripes of mudstones (mm-scale). Contain very fine sand grains. Occasionally slumped.	Transported by turbulent flow and deposited by waning of dilute flow.	Serrated. GR=40-80 API; DT=80-90 μs/ft; Density>2.5 g/cm ³ ; Neutron>20 v/v. No overlap between GR & DT logs. Have negative separation between density and neutron logs (Figs. 5.8a, c).
Ripple-laminated sandstone (T _C ; Fig. 5.7d)	1-2 m	100%	Very fine- to fine-grained sandstone with ripple lamination (cm-scale). Very rare bioturbation.	Transported by turbulent flow and deposited when decelerating current started to stabilise.	Blocky to bell-shaped. GR<40 API; DT=90-120 μs/ft; Density<2.3 g/cm ³ ; Neutron<20 v/v. Typically, density & neutron logs are tracking but positive separation occurs in hydrocarbon layer (Fig. 5.8a).
Laminated sandstone (T _{B-1} or T _{B-2} ; Fig. 5.7e)	1-3 m	100%	Fine- to medium- and occasionally coarse-grained sandstone with fine lamination (mm to cm-scale). Occasionally contain mm-scaled mud clasts.	Transported by dilute (T _{B-1}) or high density turbulent flow (T _{B-2}). Indistinguishable in this study.	Blocky. Generally the same as above. Difficult to distinguish between T _C & T _B in wireline logs. (Figs. 5.8a, b, e, f, g)
Massive sandstone (T _A or D _{CS} ; Fig. 5.7b, f, g)	1-10 m	100%	Medium- to coarse-grained. Structureless and well sorted. Dewatering features are common. Occasionally, contains layers of mud clasts or lithic pebbles. Commonly oil stained.	Transported by turbulent flow (T _A) or liquefied density flow (D _{CS}) and deposited by rapid fallout of sediment from the flow. Indistinguishable in this study.	Generally the same as above. Difficult to distinguish in wireline logs. (Figs. 5.8a-h).

Facies	Bed thickness	Sand content	Lithology and sedimentary structures	Process interpretation & environment	Wireline log character
Clast-rich, muddy sandstone (D_{M-2}; Fig. 5.7h, i)	1-10 m	10-20%	Pebble-to cobble-sized clasts suspended chaotically in mud and sand matrix. Clasts consists of lithic fragments of clay and sand.	Transported by debris flow. High- to moderate-strength cohesive debrite.	Serrated. GR=40-100 API; DT=60-100 μ s/ft; Density>2.3 g/cm ³ ; Neutron>20 v/v (Fig. 5.8h).

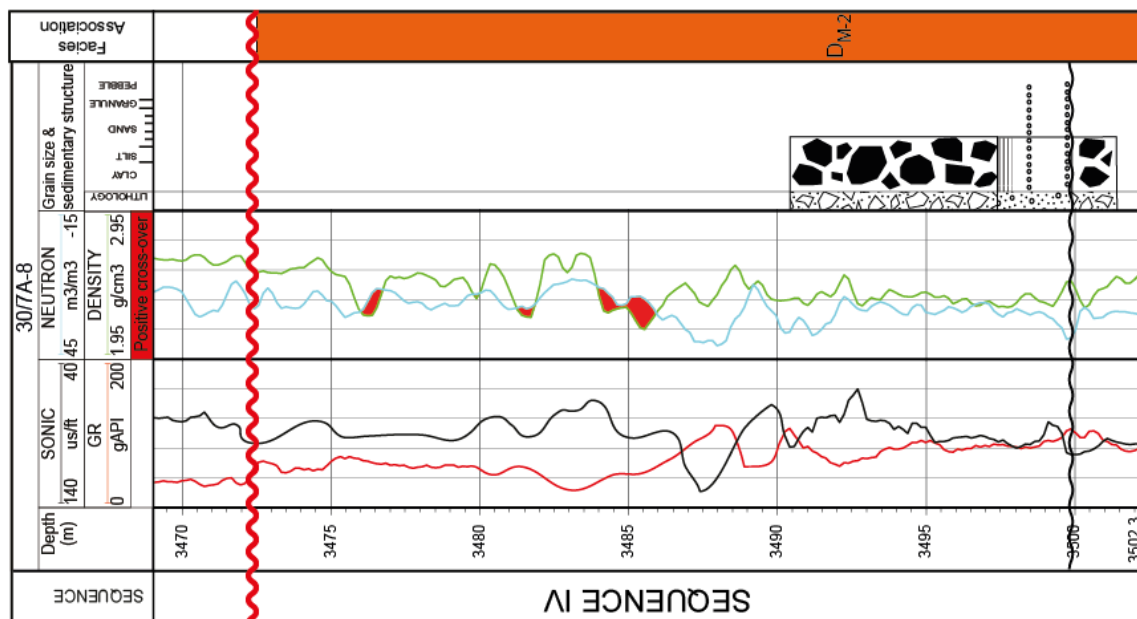
Table 5.4: Previous and this pages - Description of gravity flow facies in the Upper Jurassic interval of the Central North Sea. Processes based on Talling et al (2012). GR: Gamma Ray, DT: Sonic



LEGEND FOR CORE DESCRIPTION



(h)



(g)

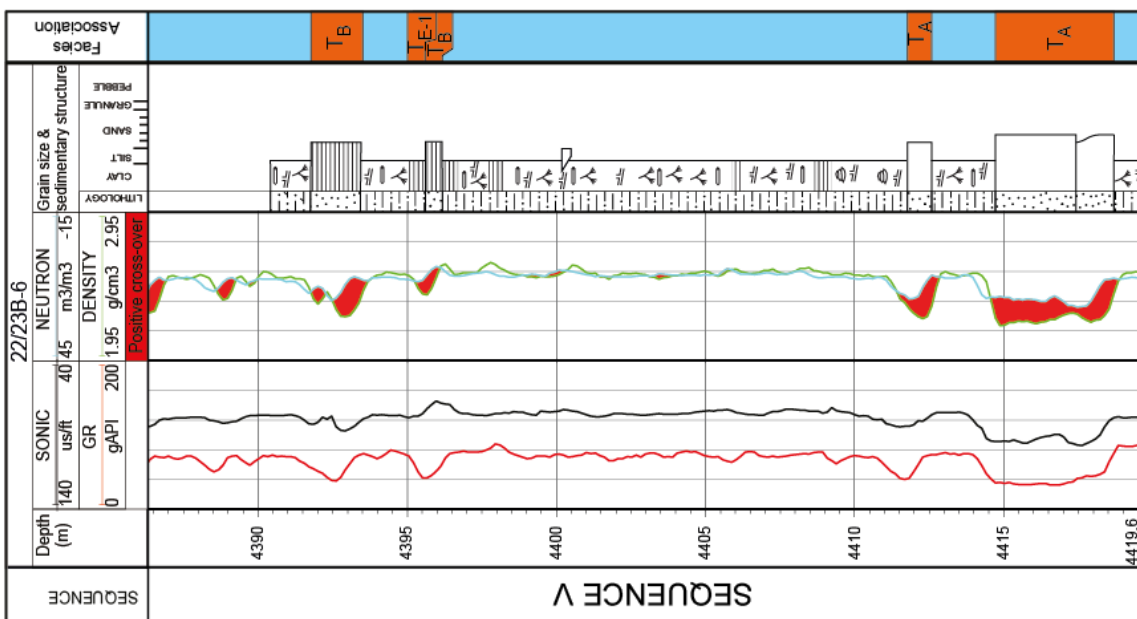


Fig. 5.8: Previous three pages - Core and log expressions of gravity flow deposit in the Central Graben. (a) Mixed of low density (bottom) and high density (top), fine-grained turbidite sandstones deposited in the Western Graben. (b) Thick, fine-grained, high density turbidite sandstone deposited in Fisher Bank Basin. (c) Thick medium-grained, high density turbidite sandstones intercalated with mud clast-rich layers in the J-Ridge area. (d) Medium-grained structureless sandstones in between bioturbated, silty, very fine-grained sandstone in southern Marnock Terrace. (e) Fine-grained, structureless sandstones resting on top of shallow marine silty sandstone in the Elgin-Shearwater Terrace. (f) Medium grain structureless sandstones with beds of pebbles resting on shallow marine sandstones in the southern Marnock Terrace. (g) Fine-grained turbidites on the edge of the Forties Montrose High. (h) Breccia and conglomerate deposits in the J-Ridge. Some were deposited within deep marine environment (a-c), expressed here by the laminated shales (often black) and overlapped gamma-ray and sonic logs, and shallow marine environment (d-h), expressed here by the presence of bioturbation in muddy to fine-grained sandstones. For location, see Fig. 5.2.

Ripple-laminated sandstone facies (T_C)

This facies comprises very fine- to fine-grained sandstones with abundant 1-2 cm-thick sets of current ripple cross-lamination, including climbing ripples with black, muddy lamina (Fig. 5.7d). Bioturbation is rare, but can include isolated burrows (Fig. 8a). This facies usually underlies the T_E or T_D facies and overlies the T_B facies (Fig. 5.8a). Wireline logs display a bell-shape to blocky character with low (<40 API) gamma-ray and moderate (80-90 μ s/ft) sonic log values without any overlap (Fig. 5.8a; Table 5.4). Density and neutron logs track each other or have positive cross-over in hydrocarbon-bearing zones.

These current rippled fine-grained sandstones are widely attributed to deposition during the stabilisation of decelerating flows, when there was enough time for the formation of lower flow regime, current ripple bedforms (Harms and Fahnestock, 1965; Talling et al., 2012). The rare bioturbation probably represents displaced, shallow-water organisms carried by the turbidity flow. The base of T_C facies usually represents a grain size break from fine- to medium-grained. This may define a period of sediment bypass defined by the abrupt transition from the underlying, high-

density turbidity current deposits (T_{A-B} interval) from the overlying, low-density turbidite (T_C) deposits (Sumner et al., 2008; Talling et al., 2012).

Laminated sandstone facies (T_{B-1} & T_{B-2})

This facies is common in the studied cores, where it typically forms 1-3 m thick intervals. This facies comprises fine- to medium-grained sandstones with subordinate thin (mm-scaled) laminations of silt- to mud-grade sediment, including clay granules (Figs. 7e, & 8a, b, f, g). This facies usually underlies the current ripple cross-laminated sandstone facies (T_C) and overlies the massive sandstone facies (T_A). The wireline log character of this facies is similar to that of the T_C facies, albeit the gamma-ray log character is blockier and perhaps, therefore, indicative of overall lower mud content (Fig. 8b).

Parallel laminated sandstones could be formed by two processes: (1) dilute turbidity flows, which form the T_{B-1} facies (John, 1982); or (2) high density turbidity flows, which form the T_{B-2} facies (Sumner et al., 2008). In our case, there is no unequivocal evidence to enable confident discrimination between these processes.

Massive sandstone facies (T_A)

Massive sandstones form the most common facies found in wells used this study (Fig. 7b, f & g; Table 2). The sandstones are fine- to medium-grained, moderate to poorly sorted, and occur in intervals that are 1-10 m thick. Individual beds have sharp upper and lower boundaries. Internally, these sandstones are almost devoid of sedimentary structure except for water escape structures, such as dish and pillar structure, which pick-out a weak stratification (Fig. 7e). Pebble-sized clasts of

mudstones are found in some intervals (Fig. 7h; 8c). The wireline log character of this facies is similar to the T_B facies; thus, the two facies cannot be distinguished from their log response alone.

Massive sandstones with dewatering structures are most likely to have been deposited by high density turbidity flows (Talling et al., 2012). The massive structure forms as a result rapid fall-out of sediment from the main highly turbulent flow as the near-bed velocity fluctuates, causing sedimentation in a layer-by-layer fashion (Kneller and Branney, 1995) Massive sandstones can also be deposited from debris flows, in which sand is dumped *en masse* from a non-cohesive laminar flow (Mulder and Alexander, 2001). The latter may have patchy grain size distribution, which give it a swirly fabric character, but may lack dewatering features (Talling et al., 2012). The origin of mud clasts is thought to come from rip-up clast from the muddy sea floor, possibly further updip as the flows cut through previously deposited cohesive layers (Johansson and Stow, 1995). It is not possible to unequivocally distinguish these two processes in a purely massive sandstone bed, but the close association with other facies observed in this study suggests that high-density turbidity currents represent the most likely process.

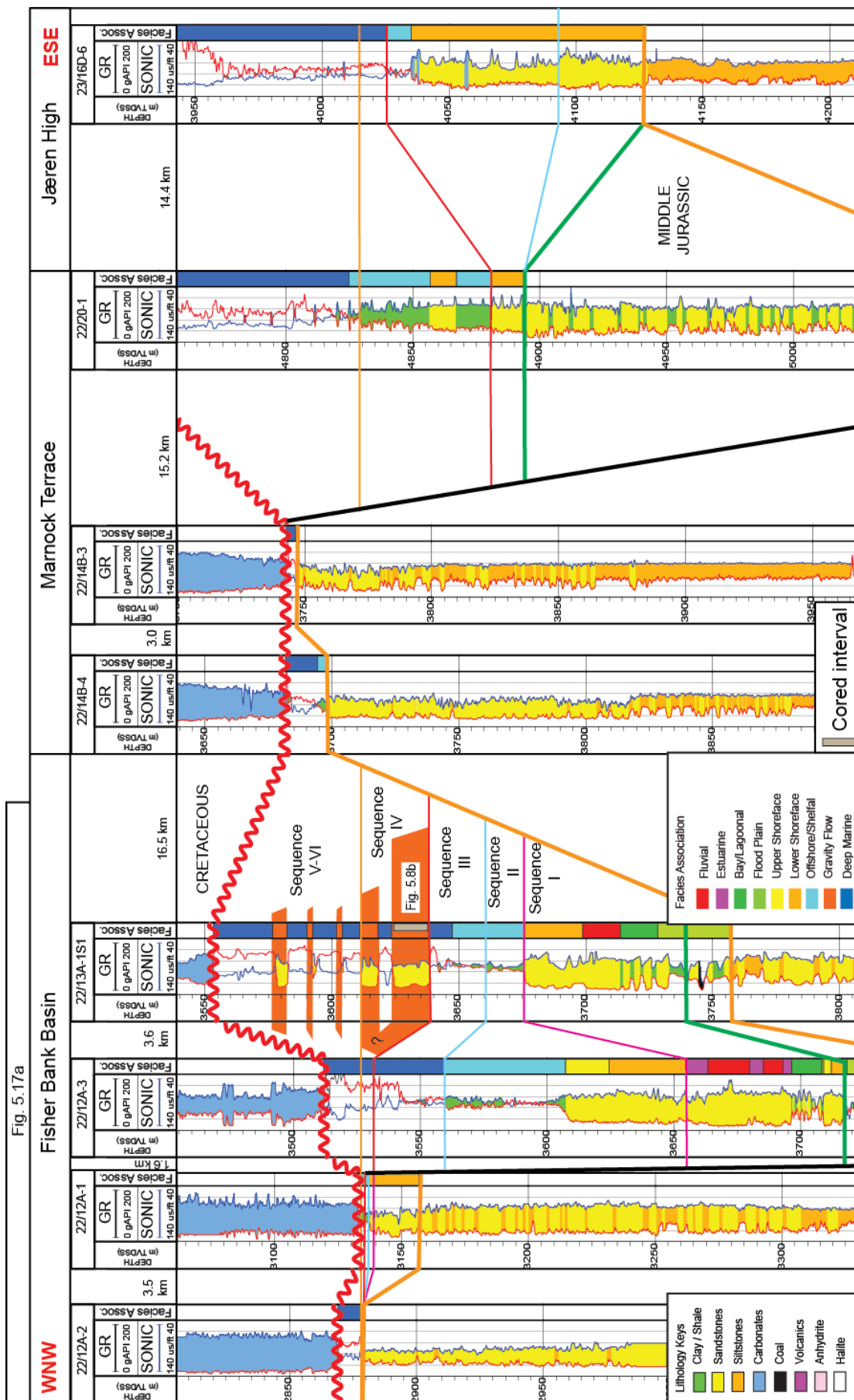
Clast-rich muddy sandstone facies (D_{M-2})

This facies comprises clasts of sandstone and mudstone within a muddy matrix; it is relatively rare, having been encountered in only two wells (22/29-3 and 30/07A-8; Table 2; Figs. 7i, j). The matrix is a muddy, poorly-sorted, fine- to medium-grained sandstone. Clasts are a mixture of sandstone and mudstone, which range in size from granules (mm-scaled) to cobbles (dm-scaled). In well 22/29-3, the clasts comprise rounded clasts of sandstones and mudstone in beds around 1 m thick (Fig. 7i), whereas in well 30/07A-8 the clasts consist of larger angular blocks of similar material and in layers up to 25-30 m thick (Fig. 7j). Wireline log response for this facies is variable depending on the proportion of mud in the matrix and the overall proportion of mudstone clasts. In the intervals

sampled in this study, GR values are very high in the thinner layers (>150 API) and low to moderate in the thicker brecciated intervals (40-80 API).

The presence of thick layers of relatively large, chaotically distributed clasts (well 30/07A-8) suggests deposition from high-strength cohesive debris flows (Talling et al., 2012). In contrast, the smaller clasts and sandier matrix in the much thinner layers (well 22/29-3) probably originated from less cohesive debris flows. The thicker intervals, with their large, angular blocks, could have been deposited close to slope setting (Laberg and Vorren, 2000). In contrast, the thinner layers can form in a diverse range of settings, as far as 100s km into the basin, including distal fan locations (Talling et al., 2013).

Fig. 5.9: Next Page - Well correlation across the Fisher Bank Basin (FBB) in the west to Jaeren High (JH) in the east, datumed at Top Sequence IV (Early Tithonian), emphasizing on the gravity flow deposits in well 22/13A-1S1 (cored; Fig. 5.8b). The deposits are thick in Sequence IV but thinner in Sequence V-VI and seems to be the product of erosion from the west (Forties Montrose High). For seismic section, see Fig 5.17a. For location, see Fig. 5.2.



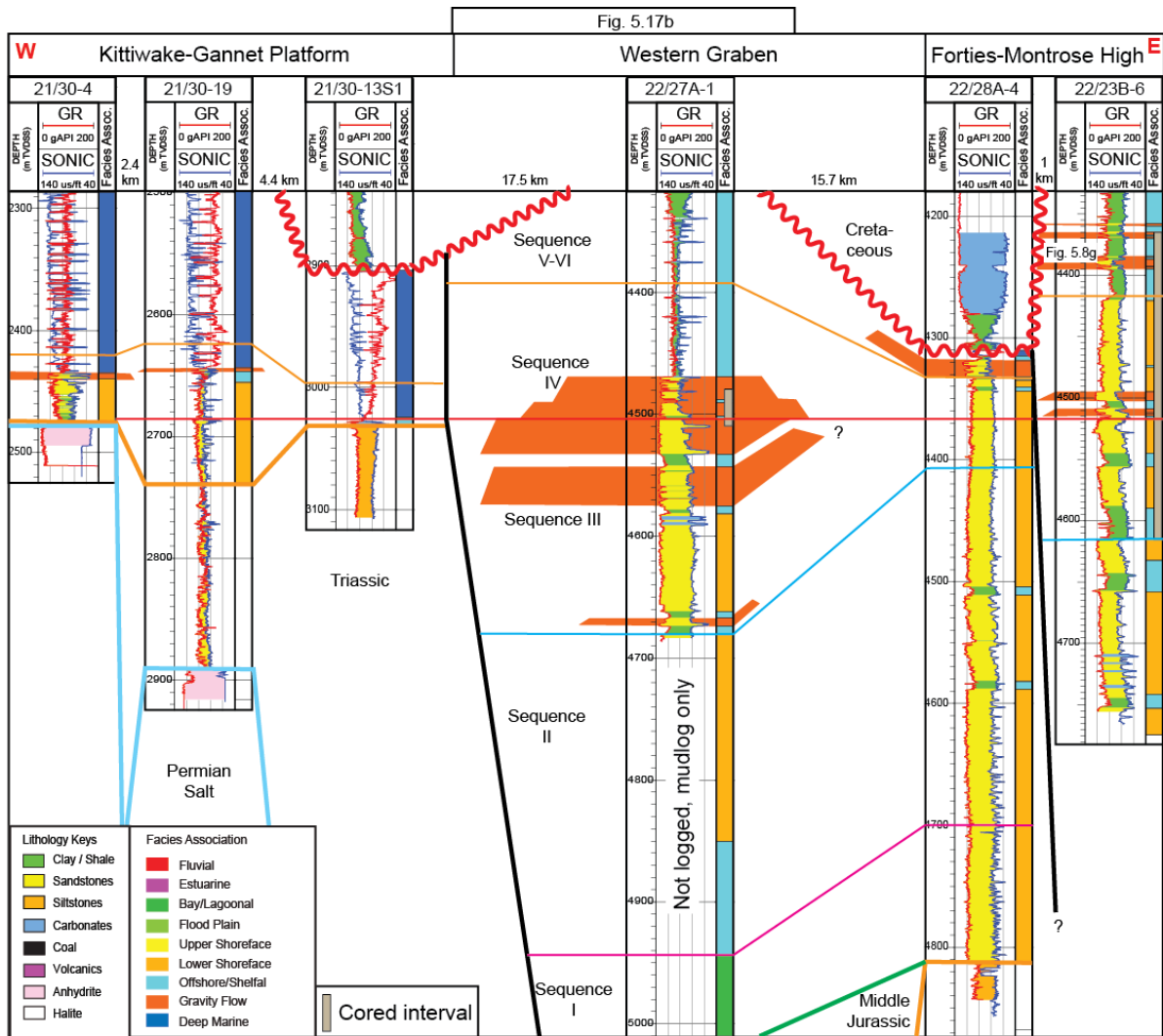


Fig. 5.10: Well correlation across the Kittiwake-Gannet Platform (KGP) in the west to the Forties-Montrose High (FMH) in the east, datumed at Top Sequence III (Near Top Kimmeridgian), showing clusters of gravity flow deposits in the Western Graben (WG) and the edge of the FMH. Thin deposits also presence in the KGP above the shallow marine sediments but unlike the WG and FMH deposits (Fig. 5.8g), the KGP's deposits are not confirmed by core data but interpreted based on the log characters. The deposits seems to spread mainly in Sequence IV-VI (Kimmeridgian-Tithonian) throughout the section and older deposits are confined in the WG (well 22/27A-1; see seismic section in Fig. 5.17b). For location, see Fig. 5.2.

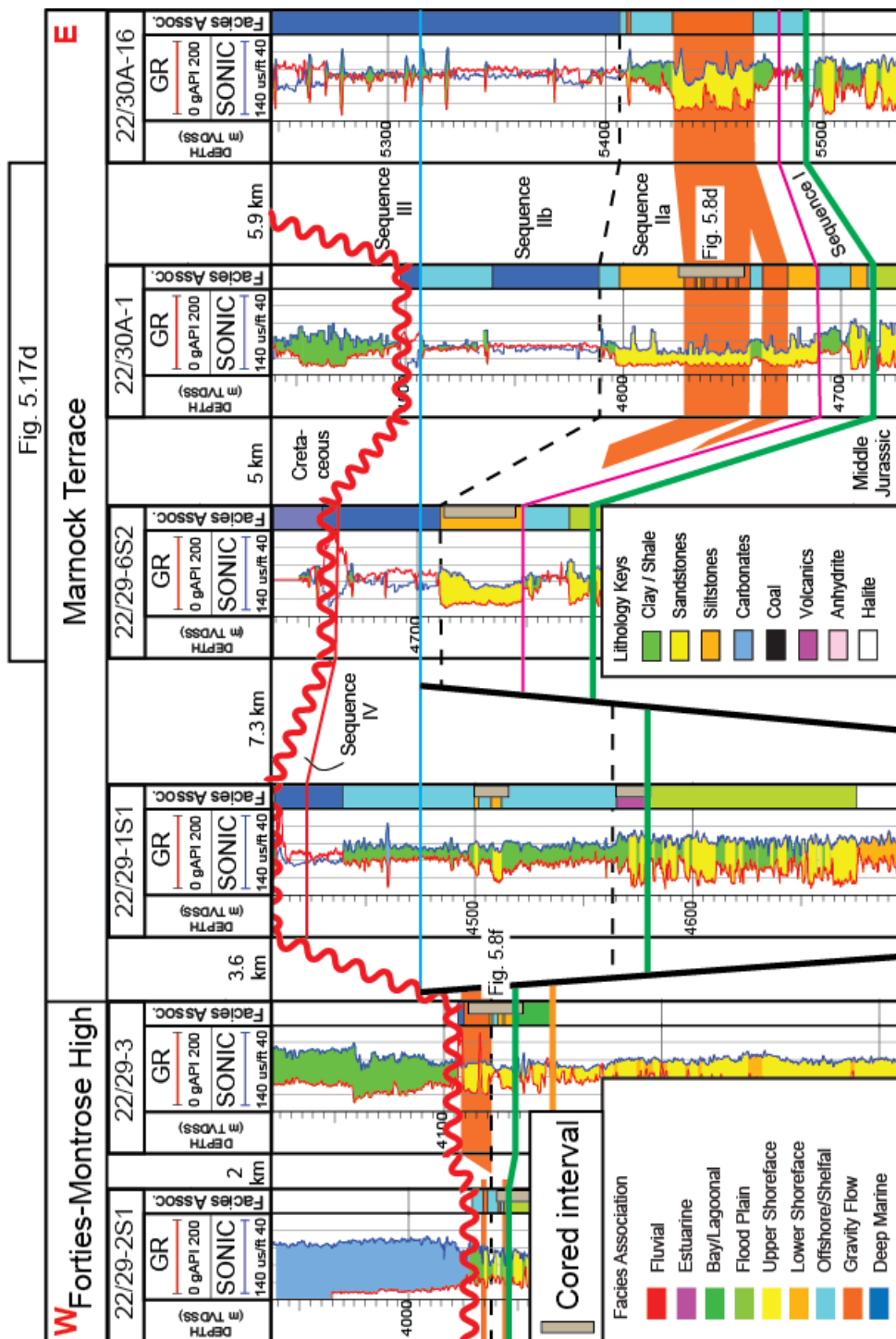


Fig. 5.11: Well correlation across Forties-Montrose High (FMH) in the west to Marnock Terrace (MT) in the east, datumed at Top Sequence II (Near Top Oxfordian). Thin gravity flow deposits in the eastern edge of the FMH and thick deposits in the southern edge of MT. The deposits seems to be confined in Sequence II (Oxfordian) in both areas and also interbedded with shallow marine deposits where the former could be distinguished by coarser grain, generally massive character and the absent of bioturbation in core samples (Fig. 5.8d, f). For seismic between wells 22/29-6S2 and 22/30A-1, see Fig. 5.17d. The source of sediments is not clear hear but could be from Triassic or older sediments from the FMH to the northwest. For location, see Fig. 5.2.

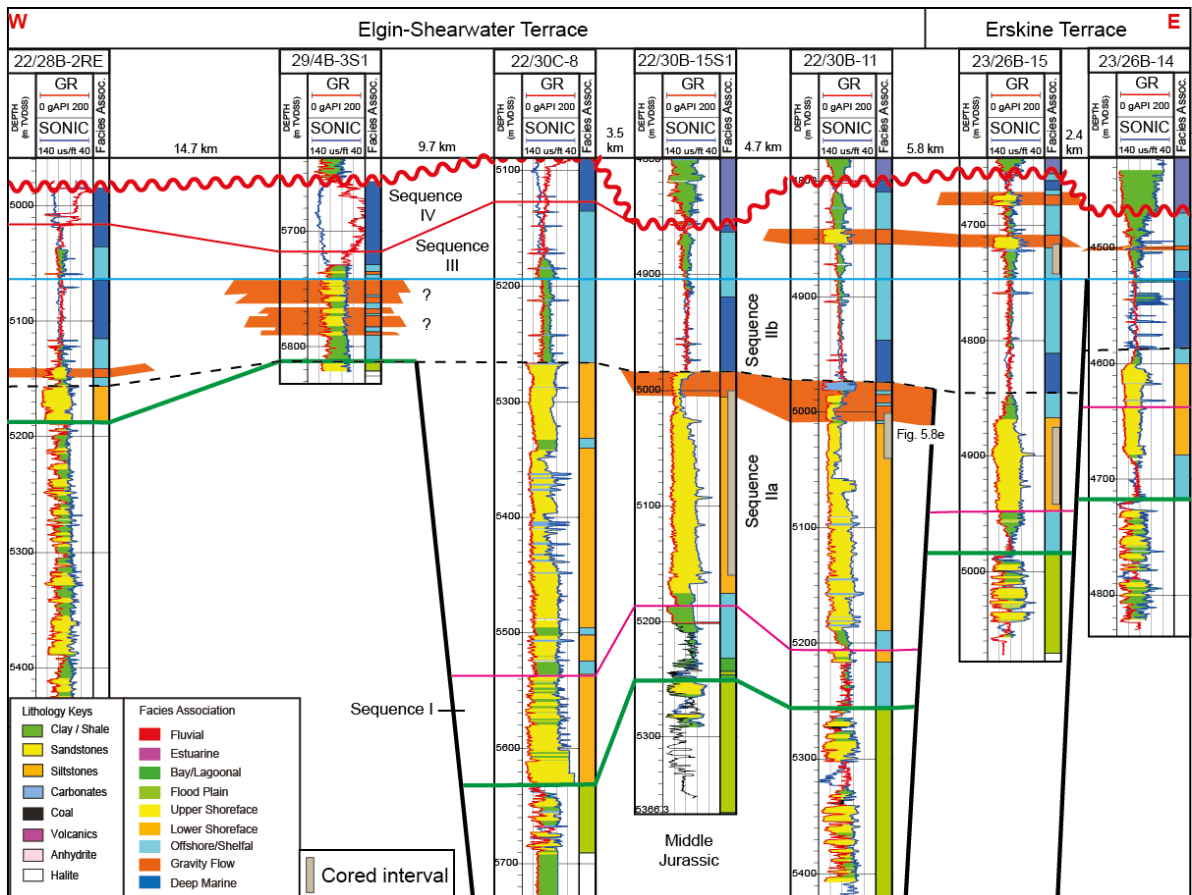


Fig. 5.12: Well correlation across Elgin-Shearwater Terrace (EST) in the west to Erskine Terrace (ET) in the east, datumed at Sequence II (Near Top Oxfordian). Gravity flow deposits mainly occurred in Sequence II (EST) with minor occurrence in Sequence III (ET). Thick deposits can be found above very thick shallow water successions in wells 22/30B-11 (Fig. 5.8e) & 22/30B-15S1 that could be the product of erosion of Middle Jurassic interval in the west (well 29/4B-3S1). The deposits in well 29/4B-3S1 and the upper Sequence III could be coming from the FMH in the north. For location, see Fig. 5.2.

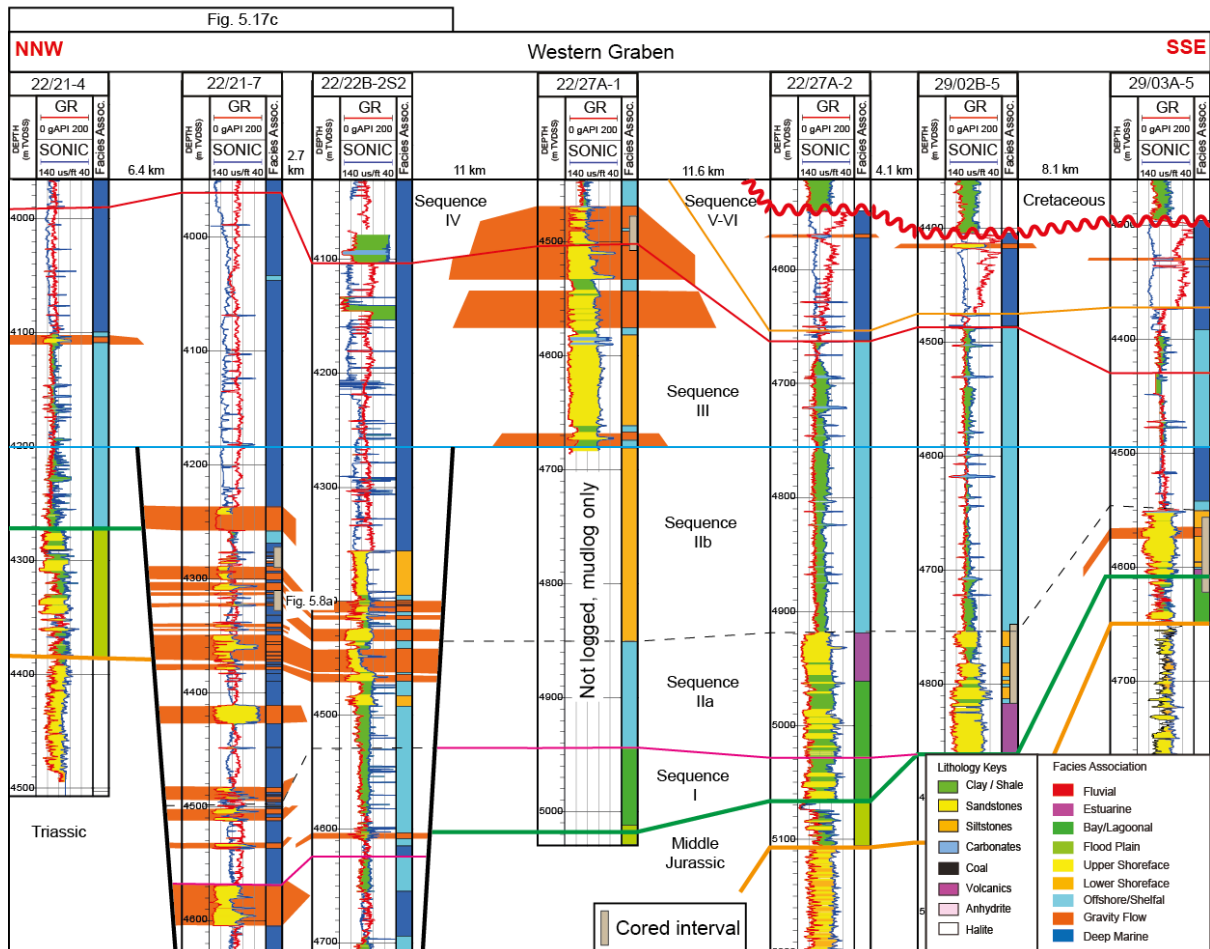


Fig. 5.13: NNW-SSE well correlation across the Western Graben (WG), datumed at Top Sequence II (Near Top Oxfordian). The main deposition of gravity flow deposits appear to be in the northern part within Sequence II (Wells 22/21-7 & 22/22B-2S2; Figs. 8a) and in the middle part within Sequence III. See Figs. 17c & 17b for seismic section. Minor deposits also can be found in the southern part in between shallow marine sands (Well 29/03A-5). For location, see Fig. 2.

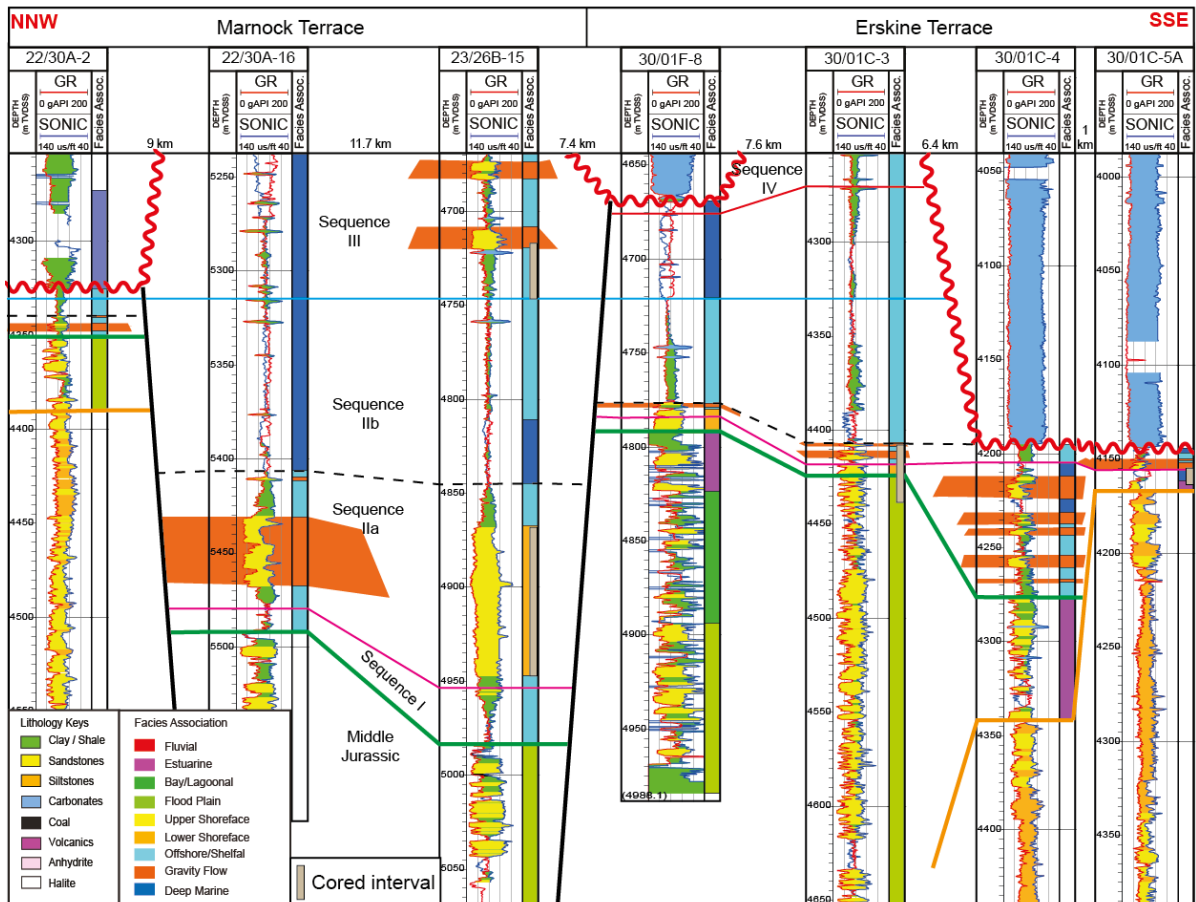


Fig. 5.14: Well correlation across the Marnock Terrace (MT) in the NNW to Erskine Terrace (ET) in the SSE, dated at Top Sequence II (Near Top Oxfordian). The gravity flow deposits seems to be more confined to the Sequence II with minor occurrence in Sequence III (well 22/26B-15) in MT but older deposits of Sequence I are presence in ET (well 30/01C-4) where the source of sediments could be the product of erosion from the Middle Jurassic sediments in the north (e.g well 30/01F-8). For location, see Fig. 5.2.

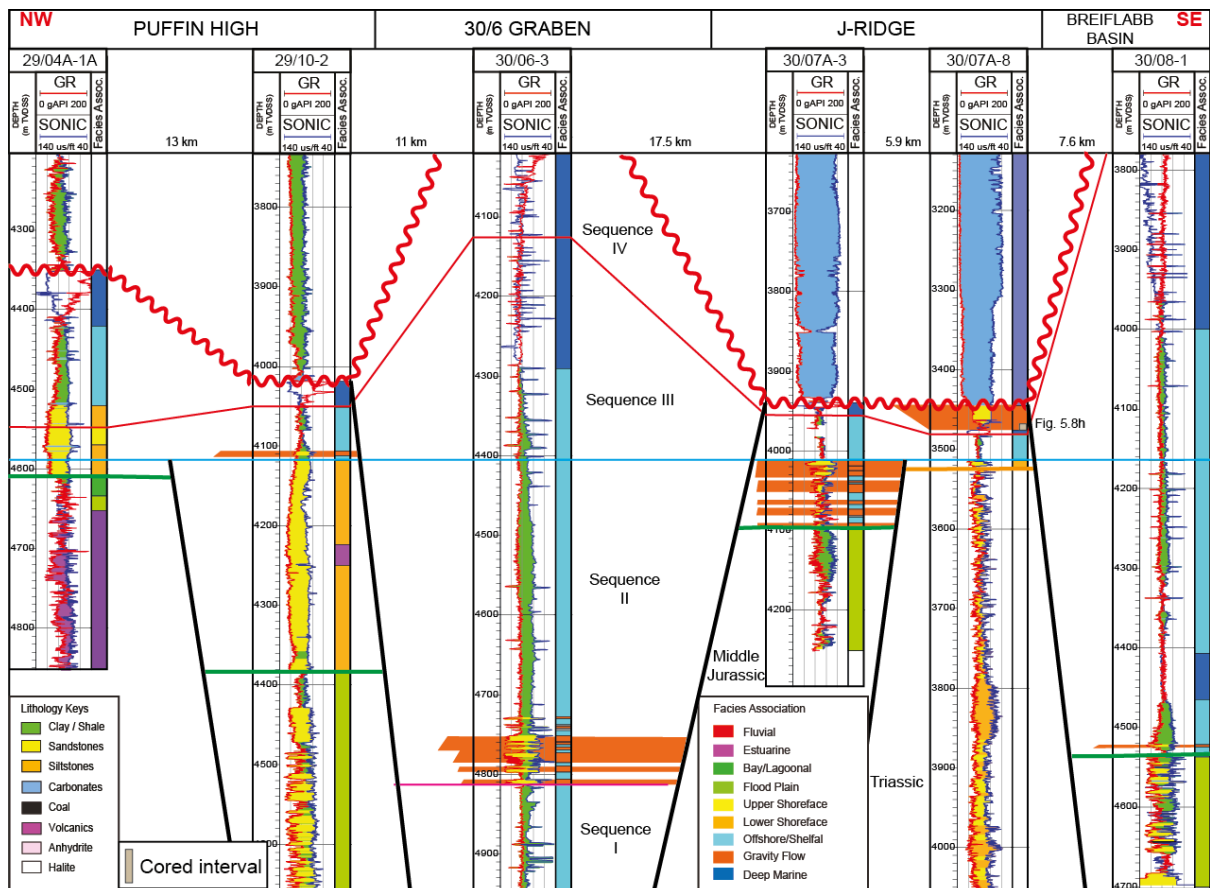


Fig. 5.15: Well correlation across Puffin High (PH) in the NW to Breiflabb Basin (BB) in the SE, datumed at Top Sequence II (Near Top Oxfordian). Thick gravity flow deposits could be found in the 30/6 Graben (3G) and J-Ridge (JR) towards the base of Sequence II. Well 30/07A-8, which lies at the flank of JR contains mostly breccia deposit (cored; Fig. 5.8h). The sediment source could come from the JR. For location, see Fig. 5.2.

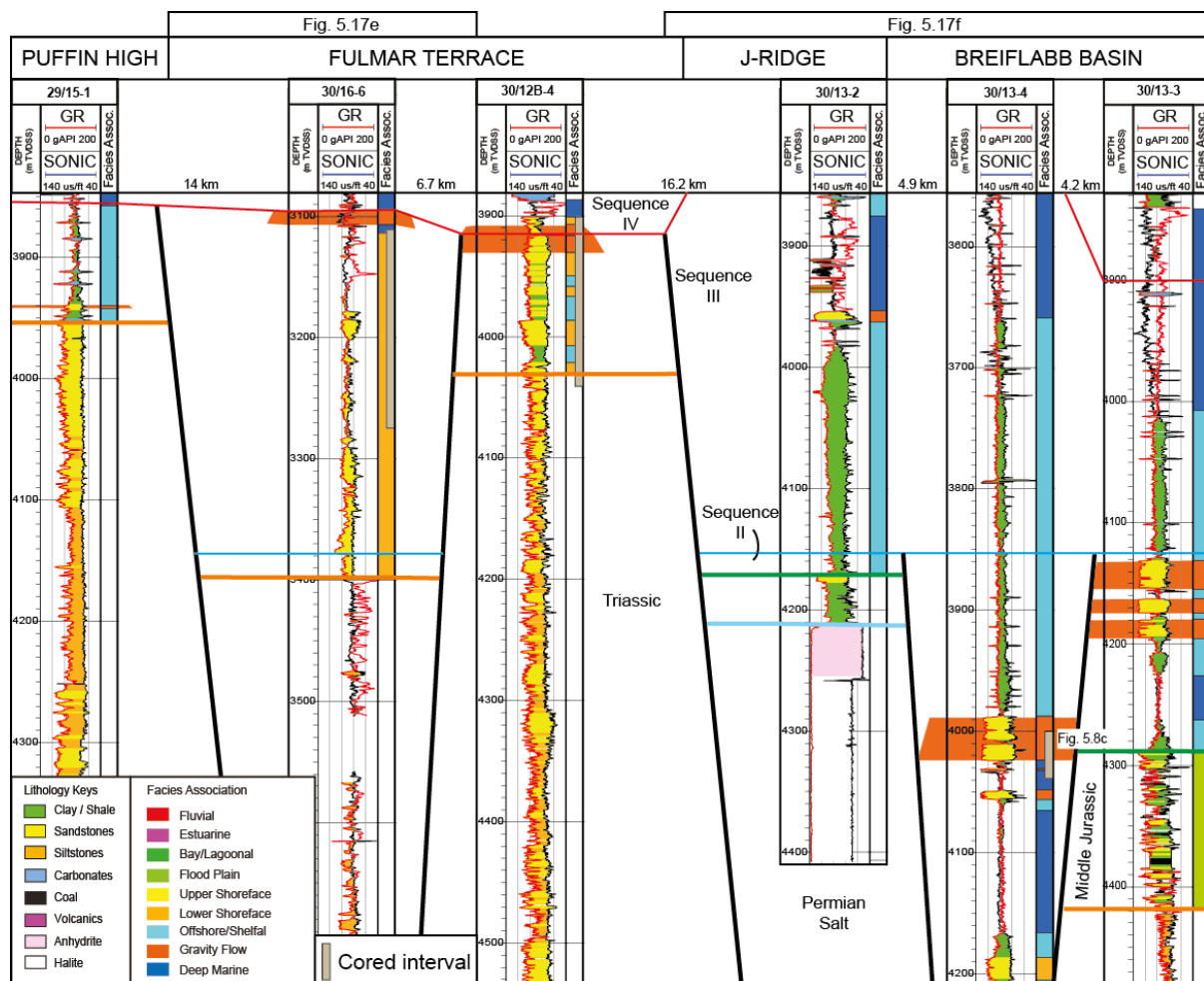
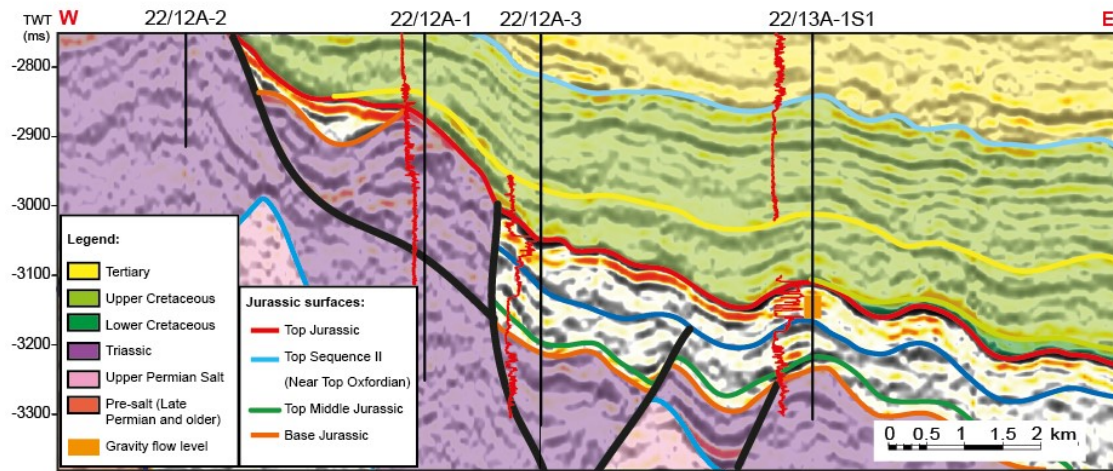


Fig. 5.16: Well correlation across the Puffin High (PH) in the east to the Breiflab Basin (BB) in the west, dated at Top Sequence II (Near Top Oxfordian). The Sequence III gravity flow deposits are thin in PH but becoming thicker in the adjacent Fulmar Terrace (FT). In the BB, thick deposits are found in Sequence II (see also Figs. 8c & 17f). However, based on seismic section, the source of sediments for the FT gravity flow may not originate from the adjacent basement highs (PH) but rather from localized high (Figs. 5.17e). For location, see Fig. 5.2.

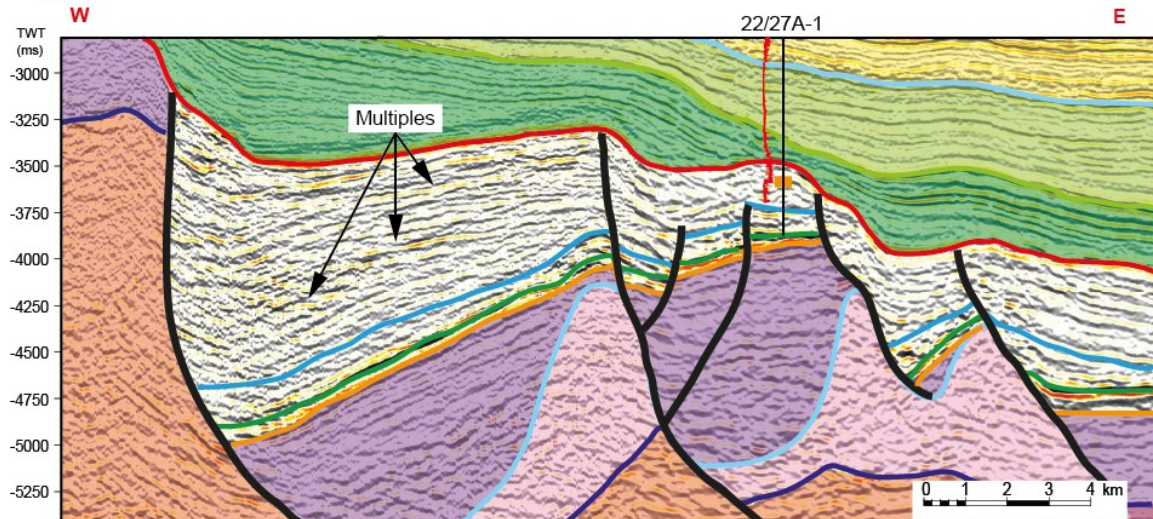
Distribution of gravity flow deposits in the Central Graben

Upper Jurassic gravity flow deposits occur at almost every stratigraphic level, within most structural domains within the CG (Figs. 5.9 to 5.19; Table 5.3), and in both deep- (e.g Figs. 5.8a, b, c) and shallow-marine settings (e.g. Figs. 5.8d, e, f, g). We here describe these characteristics of Upper Jurassic gravity flow deposits, with particular emphasis placed on their distribution relative to the seismically defined Oxfordian (Seq I and II) and Kimmeridgian-Tithonian (Seq III-VI) sequences.

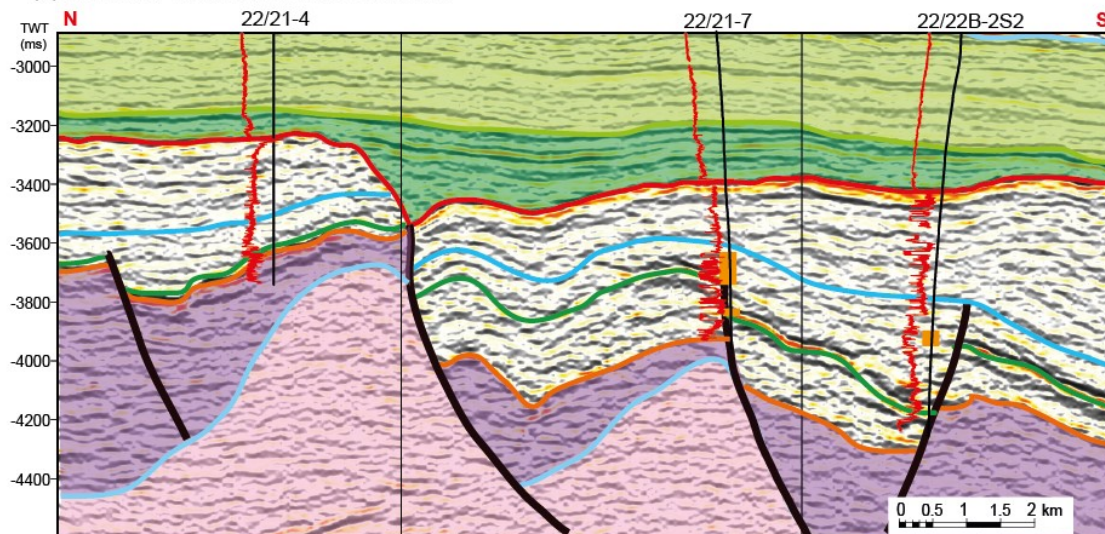
(a) Fisher Bank seismic section



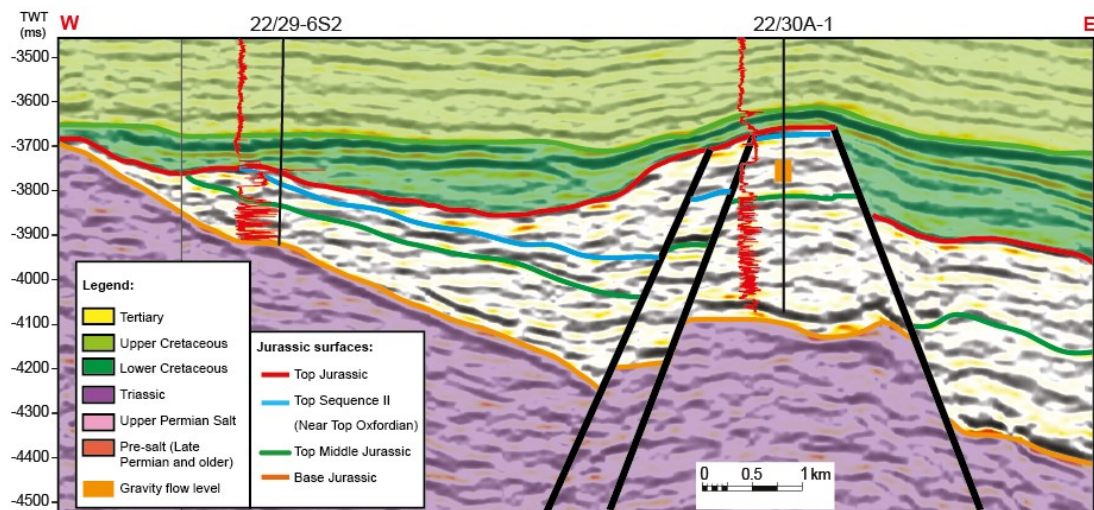
(b) Western Graben seismic section 1



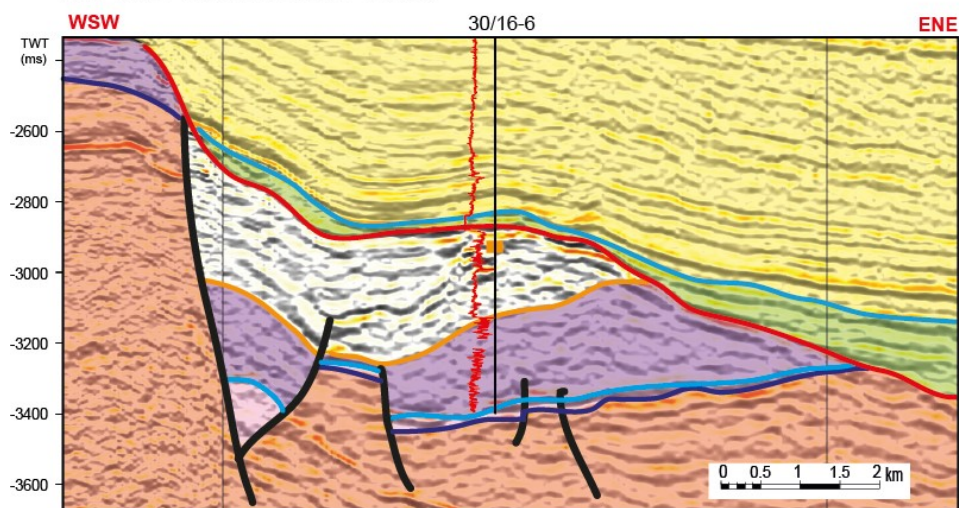
(c) Western Graben seismic section 2



(d) Marnock Terrace seismic section



(e) Fulmar Terrace seismic section



(f) J-Ridge to Breiðflabb Basin seismic section

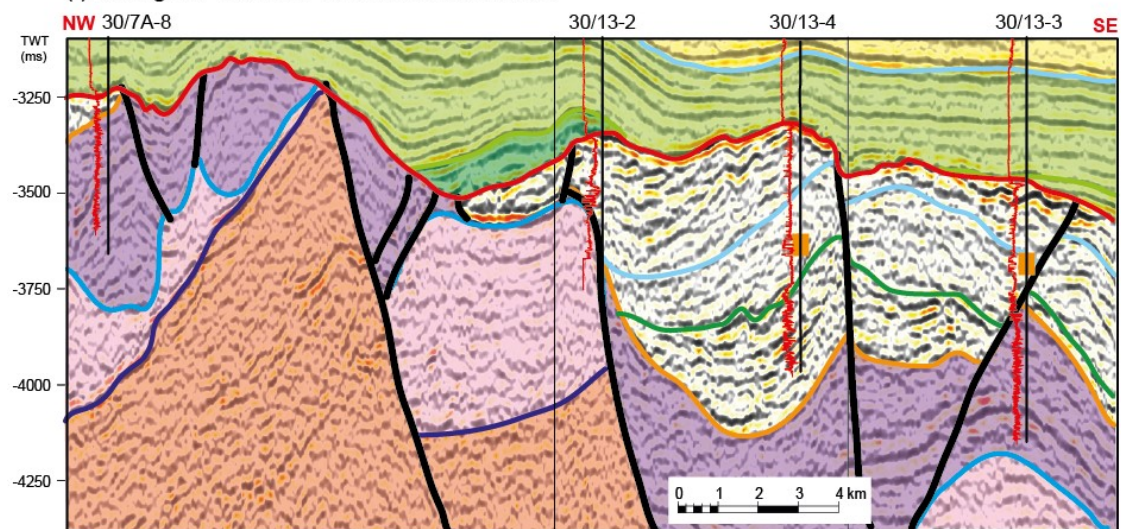


Fig. 5.17: Previous two pages - Seismic sections over the selected well correlation (Figs. 5.9-5.16) with gravity flow intervals highlighted in orange on the well path. (a) W-E section across the Fisher Bank Basin (FBB) showing the interval with gravity flow deposits just below the Top Jurassic in well 22/13A-1S1 having more transparent seismic character than the surrounding sediments in what is the best image of gravity flow in this dataset. (b) W-E section across the Western Graben (WG) showing the position of deposits in well 22/27A-1. However, the image is not clear due to the occurrence multiples that mask the reflections within the Jurassic interval. (c) N-S section across the WG showing the structural setting for gravity flow deposits in well 22/21-7 and 22/22B-2S2. Both appears to be confined to a 3-km-wide graben that may have been formed by salt withdrawal rather than rifting. (d) W-E section across the Marnock Terrace (MT) showing the structural setting for the gravity flow and shallow marine interbedded deposits in well 22/30A-1. The well is positioned almost in the middle of a rotated half-graben system that was later uplifted. (e) WSW-ENE section across the Fulmar Terrace (FT) showing the gravity flow interval dipping westwards and possibly eroded at the top, signalling that the source of deposits is not from the high in the west. Salt withdrawal may have caused the rotation and gravity flow deposits were source from the exposed older Jurassic sediments to the east (f) NW-SE section across J-Ridge (JR) to Breiflabb Basin (BB) showing the position of breccia in well 30/7A-8 and thick deposit in BB in relation to the JR. Both may have originated from the Triassic sub-crops on the JR but the deposits in BB travelled further, hence containing sand rather than breccia.

Oxfordian

The Oxfordian (Sequence I & II) gravity flow deposits are usually fine- to medium-grained and structureless based on core data (e.g. Fig. 5.8a, c, d, e & f). A key observation we make is that these Oxfordian gravity flow sandstones tend to be slightly coarser than age-equivalent shallow marine sandstones (usually very fine- to fine-grained), suggesting the former are not directly derived from the latter. Furthermore, Oxfordian gravity flow sandstones are also commonly interbedded with shallow-marine facies, which could be identified in core by the presence of bioturbation in intercalated silty sandstones and mudstones (e.g. Figs. 5.8d, e, f). Such an example found in cores in the Marnock Terrace area (wells 22/30A-1 and 22/30A-16), where gravity flow sandstones are intercalated with bioturbated lower shoreface deposits (Figs. 5.7f, 5.8d and 5.11).

The oldest gravity flow sandstones in Oxfordian Sequence I are relatively rare, with an average gross sand thickness of 30-35 m (in wells 22/21-7 & 30/01C-4; Figs. 5.13 & 5.14, respectively). Both of

these occurrences are in the Western Graben and Erskine Terrace areas, within close proximity (<10 km) to the Forties-Montrose High (Fig. 5.18a). In both locations, the sandstones are mainly intercalated with organic-rich mudstone that accumulated in a deep, basinal setting (Fig. 5.13). However, gravity-flow sandstones in the lower part of Oxfordian succession encountered on the Erskine Terrace (well 30/01C-4) are initially intercalated with shallow- rather than deep-marine mudstones (Fig. 5.14).

In contrast, the younger Sequence II contains relatively thick gravity flow sandstones; in fact, this interval contains the thickest gravity flow sandstones encountered in the Central Graben, usually in the deep grabens such as the Western Graben (110 m in well 22/21-7; Figs. 5.8a, 5.13 & 5.19b) and the 30/6 Graben (60 m in well 30/06-3; Fig. 5.15). Substantial thickness also present adjacent to the J-Ridge (60 m in well 30/07-3; Fig. 5.15), the Breiflabb Basin (40-50 m in wells 30/13-3 & 30/13-4; Figs. 5.8c, 5.16 & 5.19b), Marnock Terrace (43 m in well 22/30A-1; Figs. 5.8d, 5.11 & 5.19b), and Erskine Terrace (40 m in well 29/04B-3S1; Figs. 5.12 & 5.19b), which represent areas between the thick and thin sediments within the Upper Jurassic interval.

Oxfordian gravity flow sandstones are 2-110 m thick, with an average gross thickness of c. 30 m (Fig. 5.19a). Some of the thickest gravity flow sandstones occur where the Oxfordian is overall thick, such as in the 30/6 Graben and Breiflabb Basin areas (Figs. 5.18a & 5.19b). However, there is no correlation in the J-Ridge, Erskine Terrace, Marnock Terrace and Western Graben areas, where the gross Oxfordian interval is relatively thin (<200 m) and yet the sandstones are anomalously thick (average gross thickness between 14 to 40 m). The presence of bioturbated facies that intercalated with these sandstones also indicate that those sands were deposited in a shallow marine environment (e.g. wells 22/30A-1 & 22/30A-16 in Figs. 5.8d, 5.11 & 5.17d).

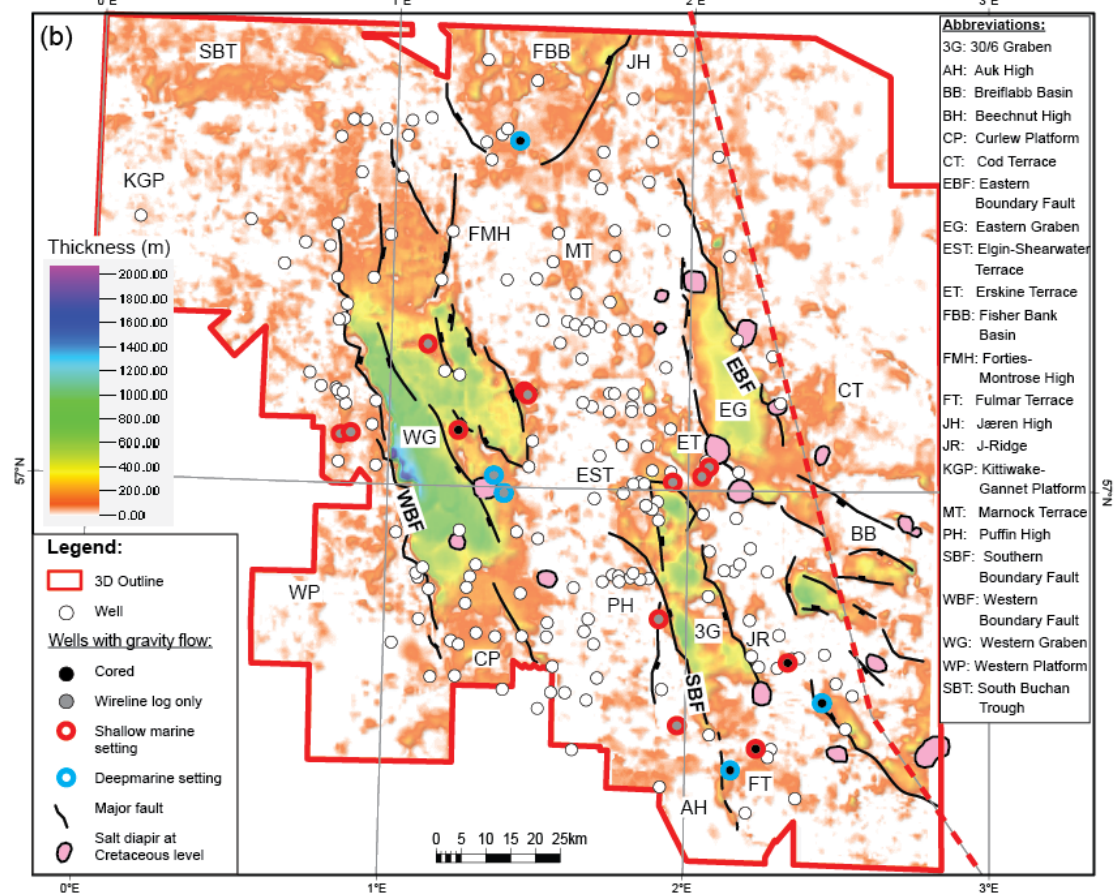
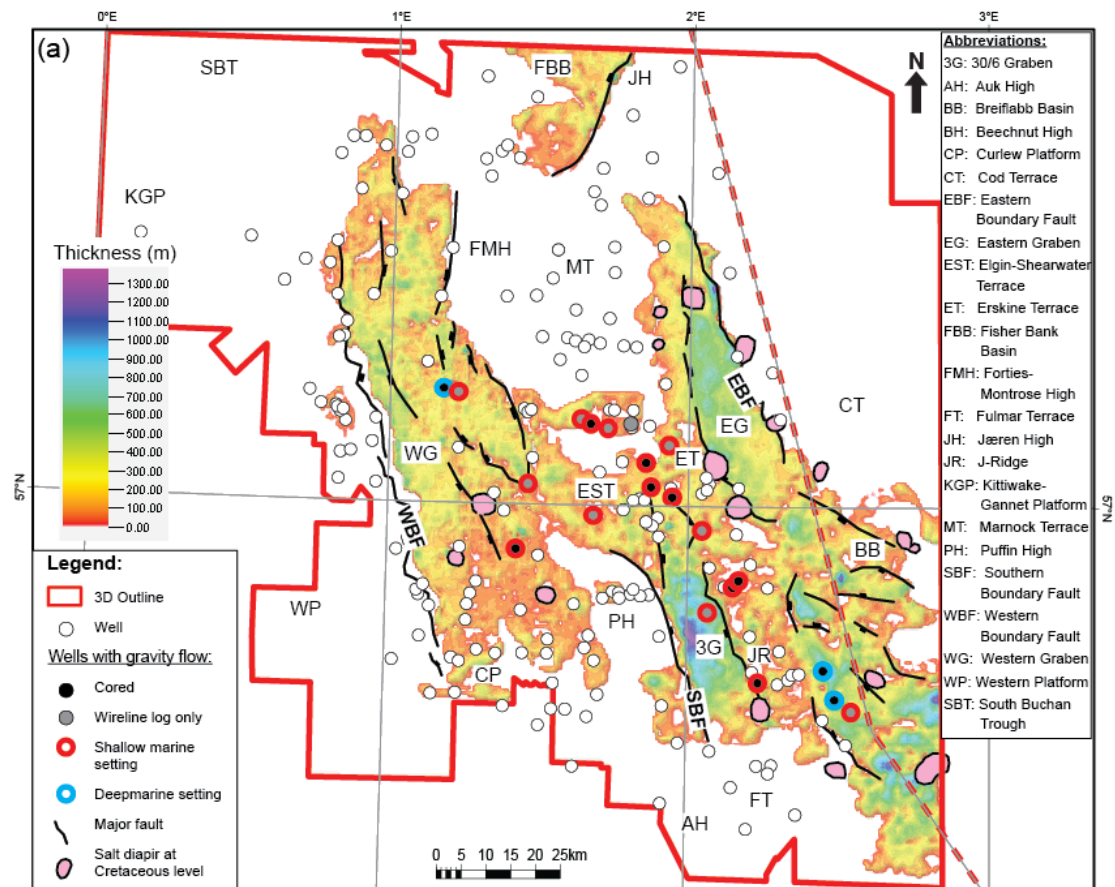


Fig. 5.18: Previous page - Isochore maps of the major sections of the Upper Jurassic interval overlain by the location of wells used in this study, highlighting the wells containing gravity flow deposits and their environmental setting based on core and/or log observations. (a) Isochore of the Oxfordian (Sequence I to II) showing that generally the basin deepens to the east. There seems to be a general positive correlation between thicknesses with environmental setting of the deposits. (b) Isochore of the Kimmeridgian-Tithonian (Sequence III-VI) showing the presence of more isolated deep sub-basins as a result of accelerated rifting. The gravity flow deposits are more sporadic and deepwater setting also have transgressed onto the Oxfordian highs (e.g. FT & PH).

Kimmeridgian-Tithonian

The Kimmeridgian-Tithonian (Sequence III to VI) consists of mainly fine- to very fine-grained sandstones with cumulative gross sand thickness of 8-17 m (Figs. 5.8b & 5.19b), although one well contains a c. 65 m thick sandstone-bearing interval (well 22/27A-1; Figs. 5.10, 5.13 & 5.19a). These sandstones are either interbedded with organic-rich, non-bioturbated deepwater shales (e.g. Fig. 5.8b) or bioturbated shallow marine mudstones (e.g. Fig. 5.8g).

Kimmeridgian Sequence III sandstones are the more widespread than in the younger intervals (Fig. 5.19b). Apart from the presence of thick sandstones in Western Graben (65 m in well 22/27A-1; Figs. 5.10 & 5.13), other significant sandstone-rich occurrences occur in Erskine Terrace (25 m in well 23/26B-15; Figs. 5.12 & 5.19b), and on the Fulmar (15 m in wells 30/12B-4 & 30/16-6; Figs. 5.16 & 5.19b) and Elgin-Shearwater terraces (well 22/30B-11, which could also be the extension from deposits in well 23/26B-15 in Erskine Terrace; Figs. 5.12 & 5.19b). These sandstones occur within the generally thinner Kimmeridgian-Tithonian interval and are often intercalated with bioturbated shallow-marine mudstones.

The overlying Kimmeridgian-to-Tithonian Sequence IV is less widespread than Sequence III, but generally contains thicker deep-water sandstone (Fig. 5.19a). The most significant sandstone-rich

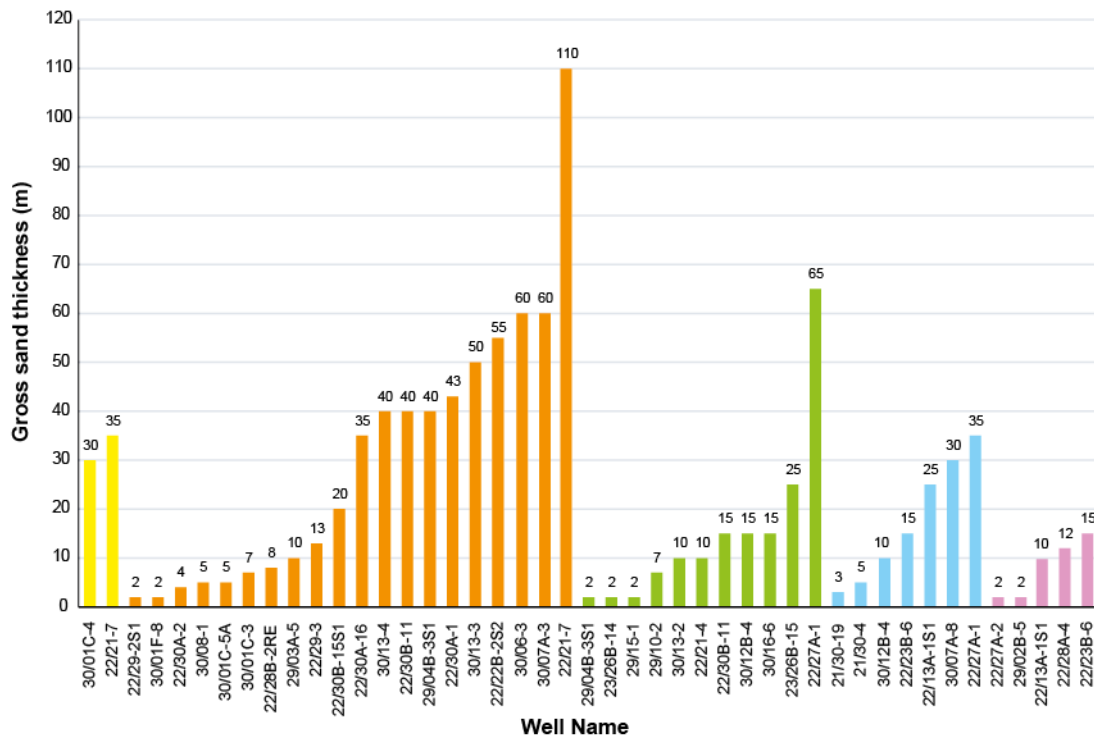
intervals occur in the Western Graben (35 m in well 22/27A-1; Figs. 5.10, 5.13 & 5.19a), J-Ridge (30 m in well 30/70A-8; Figs. 5.15 & 5.19b), and Fisher Bank Basin (25 m in well 22/13A-1S1; Fig. 5.9 & 5.19b). These gravity-flow sandstones also occasionally occur on present structural highs, such as on the Kittiwake-Gannet Platform (e.g. in wells 21/30-4 & 21/30-19; Fig. 5.13) and on the Forties-Montrose High (well 22/23B-6; Fig. 5.10). Whereas some of the sandstones in Sequence IV accumulated in a shallow marine environment, albeit at a relatively distal location (e.g. well 22/27A-1 in Fig. 5.10), thicker sandstone successions occur within truly deep-marine environments (e.g. well 22/13A-1S1 in Fig. 5.9, wells 21/30-4 & 21/30-19 in Fig. 5.10 & wells 22/27A-1 & 29/02B-5 in Fig. 5.13). These deep-marine-associated sandstones are mainly very fine- to fine-grained and laminated (Fig. 5.8b).

Cobble-rich gravity-flow sandstones also occurs in the J-Ridge area (e.g. in well 30/07A-8; Figs. 5.8h & 5.15). The sandstones are currently truncated by the Base Cretaceous Unconformity but overlie deep-marine shales (Fig. 5.15) and located in an over-steepened area of the J-Ridge (Fig. 5.17f).

Sequence V is only locally developed, typically on or near structural highs such as on the Forties-Montrose High (12-15 m in wells 22/28A-4 & 22/23B-6; Fig. 5.10). This sequence is thin overall and deep-water sandstones reach an average gross sand thickness of c. 8 m (Fig. 5.19a), typically being intercalated with offshore to deep-marine deposits (e.g. Figs. 5.8g & 5.13).

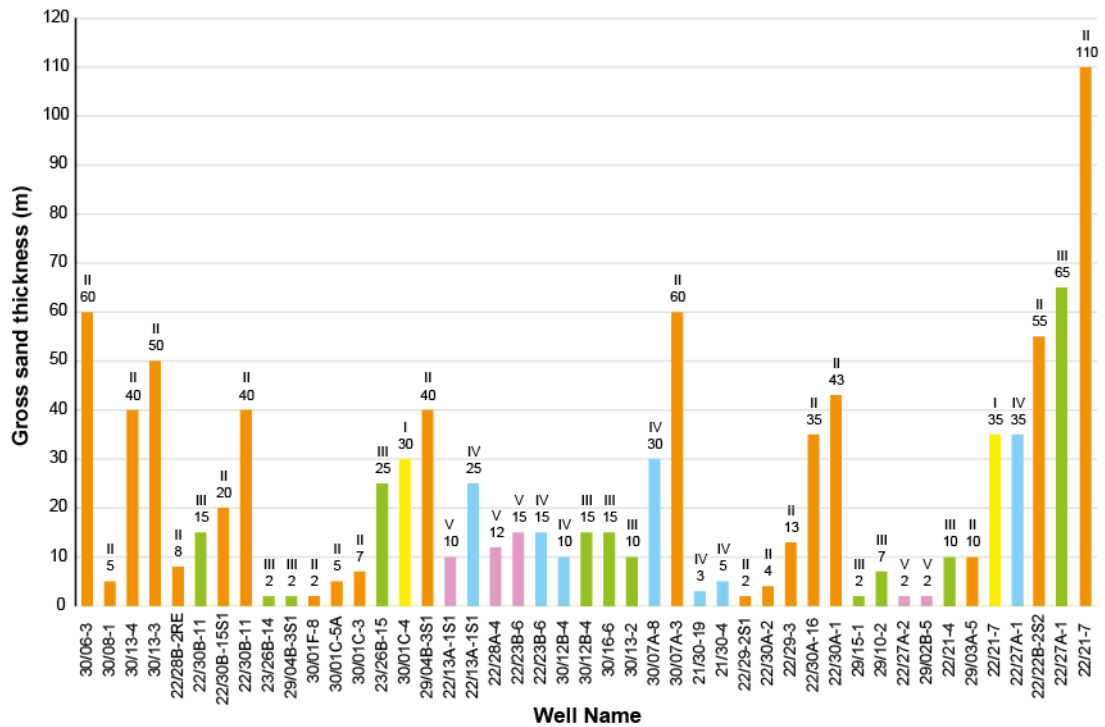
So far we have not encountered gravity flow sandstones in Sequence VI (Fig. 5.19a). This could be caused by data limitation or a general sediment starvation near the end of the Upper Jurassic rifting period.

(a) Gross thickness of gravity flow sand according to stratigraphic level



Sequence	I	II	III	IV	V
Average Thickness (m)	32.5	30.4	15.2	17.6	8.2

(b) Gross thickness of gravity flow sand according to structural domain



Structural Domain	3G	BB	EST	ET	FBB	FMH	FT	JR	KGP	MT	PH	WG
Average Thickness (m)	60	31.7	20.7	14.1	17.5	14	13.3	33.3	4	19.4	4.5	36

Fig. 5.19: Previous Page - Plots showing the distribution of gravity flow deposit thickness throughout the Central Graben. (a) Sort according to stratigraphic level, thick (>30 m) deposits are more likely to be encountered in Sequence I & II. There is a general decrease in thickness towards the younger sequences. (b) Sort according to structural domain, the deposits are very likely to be found in the Western Graben (WG) but encounters in the 30/6 Graben (3G), J-Ridge (JR), Erskine Terrace (ET) and Elgin-Shearwater Terrace (EST) are likely.

Discussions

Classification of gravity flow deposits in Central Graben

The core and wireline log observations and interpretations can be compared with the deep-water facies models of Reading and Richards (1994) (Richards et al., 1998) and gravity flow deposits elsewhere in the North Sea (Figs. 5.5a, b). The overall thickness of the gravity flow sandstones in the CG is almost an order of magnitude thinner when compared with other gravity flow deposits in the North Sea (Fig. 5.19; Table 5.1). However, there are some similarities between the deposits in the Western Graben and Breiflabb Basin with the deposits in the Outer Moray Firth and Witch Ground Graben in terms of thickness and log characters, especially with the Galley and Buzzard members (Fig. 5.5b; Table 5.1).

Another noticeable difference is the stacking character of the sandstones. The majority of the gravity flow sandstones observed in this study have blocky to slightly fining upward trends (e.g. Figs. 5.8b, c, d, e, f), which could resemble sand-rich submarine fan to submarine ramp systems (Fig. 5.5a). However, thickness-wise, especially in the latest Oxfordian (Sequence II) and all of the Kimmeridgian-Tithonian (Sequence III to V), the gravity flow deposits are relatively thin and fine-grained across much of the CG. These characteristics hold for sediments deposited above previous Oxfordian 'highs', most notably the Kittiwake-Gannet Platform (Fig. 5.10), as well as the Forties Montrose (Fig. 5.10) and Puffin highs (Figs. 5.15 & 5.16). These deposits, therefore, closely resemble mixed sand-mud submarine fan to slope apron systems (Fig. 5.5a). In addition, the thick debris flow

that was penetrated by well 30/07A-8 on the western edge of the J-Ridge (Figs. 5.7j, 5.8h, 5.15 & 5.17f) could be part of a relatively localised, gravel-rich slope apron system (Fig. 5.5).

Structural and depositional setting of gravity flow deposits in Central Graben

Although sharing broadly similar facies characteristic to Upper Jurassic gravity flow successions encountered elsewhere in the North Sea, CG deposits differ in that gravity flow sandstones were deposited in both shallow- and deep-marine environments (Table 5.1). Other than the significant thinning of sandstones found in the CG than in other part of the North Sea, the grain-size also does vary across the CG from fine-grain in the basin lows (e.g. in the Western and 30/6 grabens; Figs. 5.8a) to medium- to coarse-grain near the terraces and highs (e.g. in the Elgin-Shearwater Terrace; Fig. 5.8e; and Marnock Terrace; Fig. 5.8f). The variation of grain size seems to suggest different sediment origin. The fine grain materials are similar to the Upper Jurassic shallow water sediments (Chapter 4; Table 4.3), suggesting that the gravity flows in the basin lows could have been originated from the reworking of the contemporaneous shallow marine sediments. In contrast, the coarser grain materials, which have been deposited closer to the highs, may be the product of erosion of the Triassic or older sediments originated from the local highs such as the Forties-Montrose High or the J-Ridge. This situation is similar to that in the Witch Ground Graben, where grain size also vary according to the origin of the sediment (see Summary of Upper Jurassic gravity flow deposits in the North Sea Basin section above; Casey et al., 1993; Moseley, 1999).

Based on the structural setting of the Central Graben, where regional extension and halokinesis were synchronous, the steepening of the slope could trigger gravity flow deposition. This is supported by observation made using the seismic data in the Marnock Terrace, where gravity flow sandstones are found with shallow marine silty sandstones deposited in a half graben (Figs. 5.11 & 5.17d), and in the J- Ridge area, where cobble-rich sandstones occur in the updip position of a half

graben arm (Well 30/7A-8 in Fig. 5.17f). The latter deposits are interpreted to be a product of erosion of the adjacent J-Ridge by McArthur et al. (2011) who studied the deposits in detail.

In addition, salt withdrawal also played an important role in sediment dispersal in the Fulmar and Elgin-Shearwater terraces, where extension-induced salt movement caused rotation and erosion of Jurassic and Triassic strata, with the erosional products deposited in adjacent basins (Figs. 5.12, 5.16 & 5.17e; Johnson et al., 1986; Wonham et al., 2014; Wonham et al., 2002). The associated sandstone deposits tend to be massive, poorly sorted and ungraded, possibly reflecting deposition from high-density turbidity currents or debris flows, rather than dilute, low-density turbidity currents (Figs. 5.8d, e, f). Furthermore, based on seismic and well location, the run-out distance of these flows is 2-5 km, which is short relative to what have typically been associated with turbidity currents (e.g. 10s of km; Talling et al., 2012).

There are areas where gravity flow-deposited sandstones were deposited in deep marine shales (e.g. Fisher Bank and Breiflabbbasins, and the Western Graben; Figs. 5.8a, b, c, 5.9, 5.13 & 5.16). All three areas show evidence of dilute turbidity flow with the presence of ripple-laminated sandstone facies (T_c) and/or laminated siltstone facies (T_D). However, while the deposits in Fisher Bank Basin and Western Graben have very fine to fine grain size, the deposits in Breiflabbbasins are more mixed-grained, perhaps reflecting the differences in their provenance. The Western Graben and Fisher Bank Basin provenance are close to the Forties Montrose High, whereas the provenance of the Breiflabbbasins is thought to be either from the Auk High (which include the proto-Fulmar Terrace high) or axially further north from the uplifted part of the Cod Terrace or beyond (Fig. 5.18a; Fraser et al., 2003; McArthur et al., 2011). The structural settings of the three areas are also varied with the Fisher Bank Basin's and Western Graben's sands were deposited in a half graben whereas the

Breiflabb Basin's sands were deposited in a collapsed Triassic pod (Figs. 5.17a, c, f, 4.22; Wonham et al., 2014).

The Oxfordian section (Sequence I & II) contains the thickest and more numerous occurrences of gravity flow sandstones deposits in the Upper Jurassic in the CG (Fig. 5.19a). This is perhaps due to the active rejuvenation of salt walls at the start of rifting and the proximity to exposed shoreline, which could provide the instability and source of sediments reworked into deeper waters in flanking depocentres. Amongst the structural domains in the CG, the Western and 30/6 grabens as well as the Breiflabb Basin contains the thickest Oxfordian gravity flow deposits, with average gross sand thicknesses of 30-60 m. This is perhaps due to those areas being the deepest in the CG during the Oxfordian, although the Western Graben's Oxfordian interval have been significantly eroded later (Fig. 5.18a).

The Kimmeridgian-Tithonian represents an accelerated phase of rifting and very rapid increase in sea level, which have pushed the shoreline further onto the platform and shut down the coarse clastic input into the basin. This has resulted in thinner and sporadic distribution of the gravity flow deposits towards the end of Jurassic. The exception is in the Fisher Bank Basin, where thick gravity flow deposits are found in the Kimmeridgian-Tithonian section due to slightly younger basin formation than the main CG (Clark et al., 1993).

Significance for petroleum exploration

The gravity flow deposits found to date in the CG are very thin compared to those discovered elsewhere in the North Sea. Whereas gravity flow-deposited sandstone reservoirs in the Viking Graben are relatively thick (>100 m), similar deposits in the CG are, on average, only c. 30 m thick

(Table 5.1; Fig. 5.19). However, in term of reservoir quality, the clean sands that dominated CG's deposits are at least comparable to that of the other neighbouring basins as shown by the discoveries in Fulmar, Shearwater, Jaqui and King Lear.

The main challenge in exploring for gravity flow reservoir in the CG is seismic data resolution. As shown in this study, even 40-50 m thick sand-rich interval is beyond resolution with the current 3D seismic data, for example the hydrocarbon-bearing sandstones in the Breiflabb Basin, near the J-Ridge (Figs. 5.7c & 5.17f). However, all hope is not lost as seismic indicators of the deposits do exist, for example in the Fisher Bank Basin, which appears as transparent, chaotic seismic character (Figs. 5.8b & 5.17a). We have no information on the vintage, acquisition or processing parameters for this part of the 3D mega-survey but the observation is a significant one to take into considerations when looking for the gravity flow deposits or in setting up parameters for future seismic data acquisition and processing.

Most importantly, the proper understanding of the structural and depositional settings of these gravity flows must precede any other exploration activity. In this study, we have demonstrated that the gravity flow deposits in the CG could be found not just in the deep-water setting but also in shallow water terrains. The deposits are often found in the deep palaeo-bathymetry close to salt diapirs and growing half graben system, the combination of which could provide the instability that lead to the collapse of soft sediments or exposed rocks in the updip location. This location could only be identified by detail mapping of the interval of interest and correlating the relationship between the isochore with well data that is beyond this study's objectives and time frame.

Finally, the findings of this study indicate that the petroleum potential for the gravity flow deposits in the CG is not as attractive as in the Viking Graben or the Moray Firth Basin due to the limited thickness and areal spread of the deposits. The best gravity flow reservoirs so far are found in the Oxfordian level in Western Graben and Breiflabbb Basin, and Kimmeridgian level in the Fisher Bank Basin. However, these reservoirs lack the lateral extent to be attractive targets on their own. These deposits will remain to be secondary targets with the Upper Jurassic shallow marine deposits being the primary targets.

Conclusions

This study has shown that the main control of the deposition of the gravity flow deposits are proximity to salt structures and rift shoulders since the trigger mechanism of slope failure here are halokinesis and rifting. Therefore, the deposits are more likely to be found in the palaeo-deep close to salt highs and growth faults.

The majority of these deposits are located in the Oxfordian interval and more likely to be found in the palaeo-deep of the Western and 30/6 grabens as well as the Breiflabbb Basin. The exception is for the Fisher Bank Basin where the main deposits can be found in the Kimmeridgian-Tithonian interval.

Future exploration potential will have to rely on good understanding of the structural and depositional setting of each structural domain in the CG, although new 3D seismic acquisition and processing technology might illuminate some hydrocarbon-bearing patches in this basin. However, based on the likelihood that the sandstones are thin (<50 m) and areally confined (<3 km), the patches are not expected to be big and have to be explored together with shallow water deposit targets.

Chapter 6: Conclusions and Recommendations

A comprehensive synthesis of the Central Graben (CG) has been conducted based on the integration of a large 3D seismic dataset with 200 selected wells that include about 5 km of core data. This, combined with literature from the public domain, has provided a wealth of information that has been used to build a robust structural and stratigraphic framework for the entire CG basin. The initial challenge of putting these data together was met with an even greater challenge in integrating the interpretation results with those from the extensive literature that extends to over 800 articles. The latter includes many studies undertaken at the field- and/or sub-basin-scales, which have been integrated into the interpretations presented here. The initial, larger-scale framework was developed to analyse the effect of structuration from halokinesis and rifting on the Upper Jurassic deposition. Therefore, this final chapter starts with the summary of this geological framework, followed by its impact on the spatial and temporal evolution of sedimentation and stratigraphy. At the end of this chapter, I will outline some recommendations for future research work in this area.

Structural and stratigraphic framework

Three major structural trends controlled basin development and Upper Jurassic sedimentation in the CG: (1) NW- and NE-striking Caledonian structures; (2) E-striking Variscan structure; and (3) NNW- to NW-striking Permo-Triassic rift structure. These structural trends are mappable using the 3D seismic data and can be observed clearly on Top Rotliegend depth structure map (Fig. 3.10a). Based on these structural patterns, the CG has been further subdivided into several smaller structural domains, which determined the location, trend and accommodation space history of multiple Late Jurassic sub-basins. These are largely controlled by large NNW-striking faults, which define areas of different salt wall patterns and distribution of the major Jurassic depocentres. In addition, smaller NW- and E-striking faults provided subordinate, but still significant, structural boundaries that were also active during the Late Jurassic (Fig. 3.10b).

The significance of these structural domains can also be seen in their influence on the distribution of the Upper Jurassic depocentres (Figs. 4.20 & 4.21). The deeper parts of the depocentres are usually controlled by the NNW- and NW-striking structural patterns (e.g. in Western, Eastern and 30/6 grabens; Figs. 4.20 & 4.21). In contrast, the shallower parts are more strongly influenced by the W- and NE-striking structural patterns (e.g. Kittiwake-Gannet Platform and Cod Terrace; Fig. 4.21). The thickest and most deeply buried successions were deposited in Jurassic rift grabens, which tend to rejuvenate NW- & NNW-striking rift-related faults. The thinner and less deeply buried successions were deposited on the platform highs that were less structurally influenced by rifting and more influenced by NE-striking Caledonian and E-striking Variscan structures.

The NW-striking faults are also related to the Triassic rifting, which is proven based on the divergent character seen in the seismic data and abrupt thickness change across the NW-striking fault (Figs. 13a). The main opening direction (NE-SW) of the Triassic rift seems to be related to the opening of the Arctic to the north and the Neo-Tethys in the south. The analysis of the Triassic to Cretaceous isochore maps has shown that the Triassic rifting orientation preferentially follows the underlying NW-striking Caledonian structure (second pattern after NE-striking faults). This structural pattern can also be traced upwards into the earliest Upper Jurassic structures (Oxfordian). This demonstrates that the initial Jurassic extension has exploited this older structural fabric and may have also rejuvenated some Triassic faults (Erskine Terrace in Fig. 4.17 and Elgin-Shearwater Terrace in Fig. 4.18a). In contrast, the youngest Upper Jurassic structures (Tithonian) seems to have changed into a predominantly NNW-SSE orientation (Fig. 4.18b). However, this change does not necessarily reflect a change in extension orientation, as oblique opening is also feasible, but more of preferential reactivation of Triassic faults earlier in the Jurassic rifting period.

The study also has shown that halokinesis started during the Triassic, as shown by the isochore and palinspastic structural reconstruction (Figs. 3.12a & 3.20). The geometry of the resultant salt walls, including their length and orientation, was controlled by the underlying structure that had been rejuvenated during the Triassic rifting (Fig. 3.12b). The thickness distribution of the Permian Zechstein Salt shows that salt thins gradually in a westwards direction from the Cod Terrace to the Western Platform, which confirms that the original basin orientation was E-W during the Permian (Fig. 3.12a). However, the current NNW-SSE orientation was probably imposed during the Late Jurassic rifting phase. Most of the salt walls have been welded during the Triassic rifting phase and reached their current height at the end of the Triassic. However, a few of the salt bodies, such as those situated near the confluence of major faults, continued to rise during the subsequent Jurassic rifting and Tertiary compressional periods. Therefore, the influence of halokinesis on the tectono-stratigraphic evolution of the CG gradually diminished from the Triassic to the Tertiary.

The stratigraphic architecture of the Upper Jurassic interval has been divided into six major sequences (Sequence I to VI) bounded by major flooding events and their correlative unconformities (Fig. 4.4), using a genetic sequence stratigraphic method. This method is more suitable for marine-filled rift basins due to the nature of sedimentation, which is more responsive to localised tectonic pulses rather than to global sea level fluctuations. The stratigraphic subdivisions presented here are closely tied to both the Partington et al (1993a) biostratigraphic scheme and to the sequence stratigraphic framework of Fraser et al (2003) (Fig. 4.5). The tight calibration with previous schemes reflects the reliance on biostratigraphic data, which is driven by the sub-seismic nature of the Upper Jurassic stratigraphic sub-division. This also reflects the highly compartmentalized nature of the basin, which hampers any attempt to use seismic data to correlate between the multiple sub-basins.

Subsequently, nine (9) sedimentary facies associations (FAs) have been identified within the three main depositional environments (coastal plain, shallow-marine and deep-marine) based on core data in the Upper Jurassic interval (Table 4.3). The most widespread depositional environment is shallow-marine, which comprises upper shoreface, lower shoreface and offshore FAs with highly variable degrees of bioturbation (Fig. 4.9b). However, where original sedimentary structures are preserved, there is evidence of tidal current reworking, as interpreted from observations of double clay laminae and coarse-fine couplets within cross-bed sets (Fig. 4.9b-1).

Coastal plain environments are also identified in cores, especially in the Western Platform. The coastal plain environment includes fluvial, estuarine and lagoonal facies associations, which exhibit marginal marine influence, supported by variable biogenic fabrics (Fig. 4.9a). The bioturbation signature can be used to differentiate between Upper Jurassic (mainly lower coastal plain sediments) and Middle Jurassic (mainly upper coastal to alluvial plain sediments), with the latter typically lacking bioturbation. Interestingly, lagoonal and fluvial FAs, often found in association with each other, are found to north and south of the Western Grabens, and are particularly abundant in the Curlew Platform in the south (Fig. 4.20). The occurrence of these facies indicates the location of sediment entry points into the CG, which are inferred to have been associated with funnel-shaped palaeo-topography (see paleogeographic maps), which are generally oriented orthogonal to the basin axis. This provides the basis for interpretation of other fluvial entry points into the CG Basin (Figs. 4.20 & 4.21).

The CG also contains deep-water FAs, which comprise of thick mudstones and subordinate gravity flow sandstones (Fig. 4.9c). These are mainly found in the deeper, axial parts of the basin but gravity flow FAs can also be found intimately associated with the shallow-marine FAs (Fig. 5.8d, e, f). The gravity flows deposited two main types of beds; (1) high-density turbidites; and (2) low-density

turbidites. The high-density flows were sand-dominated and most commonly deposited sediments very close to local highs, where steep gradients were developed. In contrast, low-density flows were mud-rich and deposited classical, Bouma-type turbidites within central and axial basin locations. The origin of these gravity flows are still debatable but the mineralogical and textural similarity Triassic rocks could support derivation of high-density turbidity flows from nearby structural highs. However, low-density turbidity flows might be transported from a more distal locations, including derivation from active river mouths (e.g. McArthur et al., 2011). One conclusion from this study is that suitable conditions for major submarine fan systems (e.g. deeply incised valleys feeding canyon systems and basin floor fans) were never developed within the CG. Instead, the gravity flow deposits probably represent multiple, relatively small source areas, which were triggered by oversteepening of basin margins caused by local tectonics, including both halokinesis and rifting. Hence, the distribution of deep-water sandstones is highly variable, but volumetrically small.

Overall, the distribution of depositional environments in the CG is closely linked to the net-transgressive evolution of the Upper Jurassic, which has been widely recognized by previous workers: (1) this started with small, restricted depocentres during the Oxfordian, (2) it expanded into numerous and more widely distributed depocentres during the Kimmeridgian, and (3) this terminated with a single, large and fully connected basin during the Tithonian, which was coincident with a period of rising global sea-level (Figs. 4.20 & 4.21).

Halokinesis and tectonic influence on the Upper Jurassic deposits

The Upper Jurassic sediments were deposited during a syn-rift period and this was demonstrated by the presence of large diverging character seen in the seismic data (e.g. in Western Graben; Figs. 3.16a & 5.17b). However, this diverging character is only clearly developed for the Kimmeridgian to Tithonian levels but not for the Oxfordian level, which shows a more parallel seismic reflection

character (e.g. Fig. 3.16). In addition, the Oxfordian interval shows thickening towards depocentres that formed as a result of salt withdrawal (e.g. in the Curlew Platform & Elgin-Shearwater Terrace; Figs. 3.16b, c, 4.10b & 4.11b, c). These observations indicate that the Oxfordian sediments are more influenced by halokinesis than rifting, which had just started during that period.

In contrast, the stratigraphic pattern of the Upper Jurassic sediments on the platform areas and structural highs does not show any diverging character comparable to that in the main grabens. Here, the sediments are mostly of Kimmeridgian to Tithonian period and are dominated by transgressive shallow-marine sandstones and mudstones. The deposition was synchronous with peak rifting period and rising sea-level. However, these successions are more influenced by halokinetic processes associated with the underlying Permian salt walls than its age-equivalent in the main CG, as shown by thickening towards the centres of collapsed salt highs (interpods). This is most clearly illustrated in the two opposite- facing flanks of the basin: (1) Kittiwake-Gannet Platform and Western Shelf/Platform in the west (Fig. 4.10a), and (2) the Cod Terrace in the east (Fig. 4.24). Consequently, preserved successions within the interpods mainly represent stratigraphic remnants of sediments preserved by the collapsing salt walls during overall transgression (Fig. 4.25). Hence, predicting sandstone preservation remains difficult, reflecting the delicate balance between sediment supply, salt withdrawal and rising sea level (Fig. 4.25). This process also operated within the CG, as demonstrated in the Fulmar Terrace (Fig. 4.23), where both sediment supply and relative sea level rise were much greater.

The effect of halokinesis is also evident in the distribution of gravity flow deposits, which mostly contain coarse grain materials from Triassic and older rocks. These deposits are commonly found within the Oxfordian interval in the palaeo-deep areas of the Western Graben, 30/6 Graben and the Breiflabb Basin. The only exception is the Fisher Bank Basin where the main deposits occur within

the Kimmeridgian-Tithonian interval. The deposition within the salt-influenced Oxfordian interval, and the proximity of these deposits to salt walls, indicates that their genesis is linked to the instability caused by halokinesis. However, the role of rifting cannot be ruled out since some of the deposits can also be found on over-steepened rift shoulders (e.g in Well 30/7A-8 near the J-Ridge; Fig. 5.17f).

Jurassic rifting renewed halokinesis in the Central Graben, which had reached its initial peak during the Triassic. The structural movement related to the renewed halokinesis had profound and superimposed effects on the syn-rift successions, that include the following: (1) focusing of sediment accumulation towards local depocentres created by salt withdrawal/salt wall collapse, (2) preserving transgressive sediments within collapsing salt walls and above collapsing salt diapirs (i.e. interpods), and (3) creating locally oversteepened palaeoslopes, both around the basin margins and sometimes within sub-basins, which triggered multiple gravity flows. These processes continue to exert a profound effect in predicting sandstone reservoir distribution throughout the CG, but especially within the interpods of platform areas. In the case of shallow-marine reservoirs, establishing shoreface trends that extend beyond individual interpods provides the best means of making sandstone thickness and quality predictions.

Recommendations for future research

This study has achieved its main aim of synthesising the tectono-stratigraphic evolution of the Upper Jurassic of the Central Graben. The influence of halokinesis and rifting to sedimentary processes at regional-scale is now better understood. However, at local sub-basin scales, halokinetic processes, and resulting Jurassic reservoir preservation could be more complex. Consequently, this requires a more detailed seismic interpretation in order to decipher the interaction between halokinesis, rifting and relative sea level and the resulting impact on sedimentary processes, environments and

preservation. Several areas of interest for detailed case studies have been identified based on the complexity of the underlying salt features as well as within the Upper Jurassic sediments. These areas include (1) the Elgin-Shearwater Terrace, (2) the Curlew Platform, and (3) the eastern half of the Western Graben, which have all been affected by remobilisation of salt during the Oxfordian (Type B setting of Wonham et al. (2014)). The Curlew Platform, for example, contains the most heterogeneous sedimentary section in the study area. However, the lateral distribution of the sedimentary facies could not be mapped in detail due to a lack of well calibration to the north and south of the area. Therefore, improved seismic imaging of the salt bodies is necessary, preferably using Pre-Stack Depth Migration (PSDM) processed 3D seismic data. Additional well data would also be advantageous in order to provide additional calibration points in order to reconstruct more accurate facies maps.

The provenance of the deep-water sandstones warrants further examination in order to confirm the origin of the sediment source area(s). This study should involve a detailed mineralogical study of the Jurassic as well as Triassic and older rocks in the adjacent highs in order to re-construct the sediment supply directions. Areas of interest stemming from this study include the eastern part of J-Ridge, where relatively thick gravity flow sandstones have been found, and the Western Graben, where multiple gravity flow sandstones have been recorded in several wells.

Finally, 3D structural modelling could be used to validate and/or improve the current interpretations. This would involve estimating the amount of erosion on the rift shoulders and salt walls as well as modelling the quantifying the rate and amount of movement along the major boundary faults.

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