Influence of Tectonic Inversion and Salt Mobility on Structural Styles and Reservoir Quality in the Norwegian Central Trough

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Volume I (Text)

Thesis submitted for the degree of Doctor of Philosophy

October 2010

Declaration

The results of this thesis have not been submitted for the examination of any other degree or qualification. The observations and conclusions presented in this volume are the result of my own studies except where cited or acknowledged otherwise.

Steven Sawyer

Houston,
October 2010

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Tectonic inversion and halokinesis are well documented as mechanisms for generating structural traps for hydrocarbons. Many sedimentary basins that contain kinematically active halite deposits have also experienced deformation related to positive tectonic inversion (contractional reactivation of pre-existing structures). In such cases, patterns of uplift are often complex and the relative role of competing deformation processes and their influence on structural style is poorly understood.

The focus for this study is the Norwegian sector of the Central North Sea. A major petroleum play comprises Chalk Group reservoirs, where trap development has previously been attributed to halokinesis of the Permian Zechstein Supergroup, to Mesozoic and Cenozoic tectonic inversion events, and to a combination of both, but has never been well understood. Interpretation of high resolution 3D seismic data from a 5000 km^2 area has revealed new insights into the relationship between tectonic inversion and halokinesis.

Halite of the Zechstein Supergroup became mobile during the Triassic, with the creation of minibasins and adjacent salt highs. Salt movements continued until Miocene times. Tectonic inversion was driven by far-field plate margin forces and occurred during comparatively discrete intervals; the principal events are dated Maastrichtian - Danian (contemporaneous with Chalk deposition) and Eocene – Middle Miocene. The timing and extent of salt movement prior to inversion is a major control on structural style associated with that inversion; there are consistent and predictable differences between salt-free areas as opposed to salt-prone areas. Where there is no salt (or salt has been expelled) structural styles are deep-seated, more asymmetric and localized over the site of a pre-existing structural trend.

Tectonic inversion and halokinesis have affected the porosity and permeability characteristics, and therefore the reservoir quality, of the Upper Cretaceous Chalk Group. Syn-depositional uplift exerted a strong influence over Chalk Group thickness distribution and depositional facies type. Sedimentological studies suggest initial (facies-related) matrix porosity variations were preserved or even enhanced during subsequent diagenesis.

The physical characteristics of internal fracturing are a major control over Chalk Group reservoir quality. Historically, it has been difficult to characterize sub-seismic scale 3D heterogeneities within the Chalk Group. This study has addressed the problem of fracture development in response to fold growth through integration of theoretical considerations, subsurface data and outcrop observations, using suitably chosen structural analogues. It is probable that both inversion and halokinesis directly affected and enhanced the fracture characteristics of the reservoir.

Acknowledgements

I extend sincerest thanks to my principal supervisor John Underhill. Thank you for your abundant enthusiasm and timely advice.

This study was made possible through funding from a Natural Environment Research Council (NERC) CASE studentship, NER/S/A/2004/12733, with industrial support from ConocoPhillips Norway. ConocoPhillips are gratefully acknowledged for their financial contribution and for donating the subsurface data used in this study.

I thank Hugh Beeley for some excellent guidance regarding trickier aspects of seismic interpretation and for facilitating my month in Stavanger. I thank those geologists at ConocoPhillips who assisted me during that time, especially Nina Hagen and Alex Jones.

The seismic workstation facilities at the University of Edinburgh are financially supported by Shell/Esso and Schlumberger Geco-Prakla are acknowledged for allowing use of their GeoFrame seismic interpretation software. I thank Midland Valley for generously granting me access to their 2D/3DMove software, in-house training and support. A large fraction of my time in Edinburgh was spent in the seismic laboratory. I thank the Geoscience technical support staff and Chris Place in particular, for assisting me whenever problems arose.

I have benefited from discussions with many individuals, including Chris Jackson, Al Chaney, Ian Bartholomew, Helen Lewis and Roger Scrutton. I was fortunate to arrive in Edinburgh while Caroline Gill and Nick Banbury were finishing their own PhD theses. I am indebted to you both for your advice and example.

I thank Bob Krantz for introducing me to the Uinta Mountains as a possible analogue for Norwegian Central Trough inversion. I thank the staff at Dinosaur National Monument (Diana Dilsaver and Scott Madsen in particular) for taking an interest in my survival, including assistance with access and permits.

In Edinburgh I have lived with and worked alongside a great bunch of postgraduate geologists. In particular, I thank Stefano for joining me on fun nights out in a wonderful city, Natasha for four years of company at Viewforth, and Margaret for tolerating my thesis writing madness. Extra special thanks goes to Alli. You have put up with more than your fair share of grumbles; I would never have finished without you! Final thanks are extended to my parents, Mick and Sally, who have done so much to encourage me over the years. I dedicate this thesis to you.

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1 Introduction

1.1 Rationale and Objectives

Intraplate compressional structures have been documented in a variety of settings within the continental lithosphere (e.g. Ziegler et al., 1995 (Eurasian Plate); Guiraud and Bosworth, 1997 (African and Arabian Plates); Marshak et al. 2000 (North American Plate)). Compression in such settings is attributed to the far-field stress conditions and these are governed by plate boundary interactions (e.g. Cloetingh, 1988). Intraplate compression is manifested through basin inversion; the uplift and associated deformation of sedimentary basins that occurs in response to shortening events. Inversion typically involves reactivation of the pre-existing structural fabric with an opposite sense of throw, and development of forced folds in the cover sequence.

Within petroliferous sedimentary basins inversion structures are of economic interest as potential subsurface traps for hydrocarbons. Examples from the Eurasian Plate include the Southern and Central North Sea Basins (e.g. Underhill, 2003) and the Wessex Basin of Southern England (Underhill and Stoneley, 1998). Examples from further afield include: basins in the Rocky Mountain foreland of North America (e.g. Hefner and Barrow, 1992); and relatively underexplored areas such as the West Natuna Basin of northern Indonesia (e.g. McClay, 1995).

Previous studies have identified the pre-inversion structural fabric and the magnitude and orientation of the inversion-inducing stress field as controls on the geometry of inversion (e.g. Lowell, 1995; Sandiford, 1999). However, with the exception of recent analogue experiments (Dooley et al., 2005; Del Ventisette et al., 2006) little attention has been paid to the importance of the basin infill in modifying the response to inversion. Of particular interest is the presence of easy slip horizons such as overpressured mudstones or halite within the sedimentary sequence. Such layers are common to many sedimentary basins. Salt is especially important; halite has strength and density characteristics that are markedly atypical relative to other basin-filling sediments (Davison et al., 1996). In consequence, salt layers are prone to flow when exposed to differential stress conditions, as occurs during tectonic activity (e.g. Vendeville and Jackson, 1992).

This study examines how the presence of halite within a sedimentary basin can influence the regional-to-reservoir scale characteristics of tectonic inversion. The focus for investigation is the Norwegian sector of the Central North Sea. The Norwegian Central Trough is an intracratonic basin with a complex structural configuration derived from a multiphase tectonic evolution and the presence of kinematically active Permian salt. The Norwegian Central Trough is an area with a long history of hydrocarbon exploration and this project was able to access a wealth of subsurface information including high-resolution 3D seismic reflection data. Findings from the Norwegian Central Trough have been supplemented with data from a range of outcrop localities, in order to investigate the generic implications of this research.

1.3 Thesis Structure

This thesis comprises eight largely self-contained chapters. These chapters detail the essential theory, methods, results and implications of the study. Chapter 2 presents a background to the research; principal themes and previously unanswered questions are introduced. Chapter 3 summarises the methodologies by which these previously unanswered questions were addressed. Results are presented in four consecutive chapters. Chapter 4 presents a regional scale analysis of 3D seismic data from the Norwegian Central Trough. This analysis is the basis for a more detailed investigation of inversion, halokinesis and the structural relationships between the two (Chapter 5). Chapter 6 examines the influence of inversion and halokinesis on Chalk Group reservoir quality in the Norwegian Central Trough. Chapter 7 considers the suitability of several outcrop localities as structural analogues for the Norwegian Central Trough. It also explores the generic outcomes of this research. The various strands are brought together for discussion, along with the key implications and conclusions, in Chapter 8.

2 Background

2.1 Introduction

This chapter explores the central themes of the thesis in five parts. Section 2.2 reviews the structural and stratigraphic characteristics of tectonic inversion; these characteristics form the criteria for identifying, dating and quantifying inversion. Section 2.2 also highlights the relationship between the mechanics of inversion and the resulting structural geometries. Section 2.3 reviews the causes and mechanics of halokinesis, as well as the structural and stratigraphic characteristics that define it. Section 2.4 highlights the principal knowns and unknowns in relation to the occurrence of tectonic inversion in areas also affected by salt tectonics. Section 2.5 introduces the Norwegian Central Trough study area, focusing on the existing consensus regarding tectonic inversion and salt tectonism in this area. Detailed aims of this study are presented in Section 2.6.

2.2 Inversion Tectonics

The term 'tectonic inversion' has been defined as the contractional reactivation of a pre-existing normal fault (Cooper et al., 1989). In this thesis, tectonic inversion is defined as uplift that occurs in response to structural shortening events within a pre-existing sedimentary (extensional) basin (Stewart and Clark, 1999). This broader definition is preferred because the focus for this study is not just the fault, but the adjacent sediments that may be uplifted and/or deformed in response to movements on that fault. Only through assessment of stratigraphic relationships within sediments adjacent to the reactivated fault is it possible to define the physical characteristics, timing and magnitude of tectonic reversal. For the purpose of this study, the term 'tectonic inversion' is preferred over 'basin inversion' (despite their almost synonymous nature according to the aforementioned definitions), because 'tectonic inversion' is applicable for any extensional lineament that demonstrates shortening, without the need to define the relation of that lineament in relation to other lineaments that form a basin. This becomes important in the structurally complex Norwegian Central Trough, where the demarcation of basins and sub-basins is far from unequivocal.

Examples of tectonic inversion and the associated geometries are well documented in subsurface seismic studies and in outcrop studies (e.g. Cooper and Williams, 1989;

Buchanan and Buchanan, 1995). These observations are supplemented by physical analogue and numerical models (e.g. McClay, 1989; Brun and Nalpas, 1996; Del Ventisette et al., 2006) and theoretical considerations (e.g. Sibson, 1985; Sandiford, 1999; Ranalli, 2000). Section 2.2.1 synthesises the findings from these the previous studies into criteria for the identification, dating and quantification of inversion.

2.2.1 Criteria for Identifying, Dating and Quantifying Inversion

Identifying

There are many processes that cause uplift within sedimentary basins: tectonic shortening (inversion); salt diapirism (e.g. Jackson, 1995); plume uplift (e.g. White and Lovell, 1997); basaltic underplating (e.g. Brodie and White, 1994); mantle delamination (e.g. Meissner and Mooney, 1998); and post-glacial isostasy (e.g. Ekman, 1991). Recognition of tectonic inversion requires a demonstrable link to compression in order to exclude these other possible mechanisms of uplift. The diagnostic criterion for tectonic inversion is the presence of an uplifted syn-rift sediment wedge (Figure 2.1). These features can be identified in the subsurface using seismic reflection data. Assessment of tectonic inversion by the seismic-stratigraphic interpretation of Cretaceous to Recent post-rift sediments within the Norwegian Central Trough was a major component of this study.

Dating

The timing of inversion can be constrained by identifying erosional unconformities within the post-rift sedimentary sequence. Dating them requires well calibration of the seismic reflection data. Dating precision depends on the extent to which the post-rift is preserved, its composition, and the method used to analyse it (which is in some cases limited by its composition). Detailed seismic-stratigraphic analysis coupled with absolute age dating of constituent fossils would provide the optimum resolution.

Quantifying

The intensity of inversion (amount of uplift) can be considered using the concept of a null point (Williams et al., 1989). The null point is the position on the fault where there is an apparent zero offset of beds across that fault. Everything above the null point demonstrates net reverse displacement (shortening), while everything below shows net normal displacement (extension). As the intensity of inversion increases, the null point migrates progressively down the fault (away from its tip) and progressively older syn-rift markers are put into net contraction. Typically, the shortening strain is much less than the initial extensional strain, so the uplifted cover sequence passes into net extension at depth (e.g. Figure 2.2, a seismic example from the Southern North Sea). This can give rise to unusual 3D structural geometries, such as faults that apparently change from normal to reverse along strike. This is one factor that accounts for added 3D structural complexity in sedimentary basins that have experienced compression subsequent to their initial formation under an extensional stress regime.

The geometric form of an inversion structure depends on the pre-inversion setting, the inversion-inducing stress field, and the nature of the sedimentary fill. These factors are discussed in Sections 2.2.2, 2.2.3, and 2.2.4, respectively.

2.2.2 The Pre-inversion Setting

The style and magnitude of inversion depends on the pre-inversion setting. This includes the strength of the lithosphere, its prior stress history, and brittle properties of the shallow crust, such as the orientation of pre-existing faults.

Lithospheric Strength

The strength of the lithosphere controls its susceptibility to deformation. Lithospheric strength is defined by temperature and composition (Ranalli, 1995). It depends on geothermal gradient and the relative thicknesses of the ductile and overlying brittle layers. Since temperature and composition are modified by tectonic processes (e.g. rifting), the strength of the lithosphere also depends on prior stress history (Cloetingh and Burov, 1996).

The development of rifts has been much studied: a detailed review of sedimentary basin formation models is presented in Allen and Allen (2005). Different stretching models have been proposed, including uniform (e.g. McKenzie, 1978) and depth-dependent (Wernicke, 1985) varieties. Different models often lead to different interpretations as to the evolving strength characteristics of the lithosphere. The lithosphere may undergo either strain hardening (strengthening) or strain softening (weakening) during extension. There is a trade-off between strengthening due to replacement of weak crust by strong mantle lithosphere and weakening due to increased thermal gradient. Kusznir and Park (1987) identified the rate of stretching as a critical factor governing this trade-off; strain rates $>5*10^{-15}$ s^{-1} (equivalent to 1.6% extension per 10^5 years) are predicted to produce a weakening of the lithosphere. Since this value is lower than for most commonly observed rift events (10^{-14} s^{-1} is more typical e.g. Pfiffner and Ramsay, 1982), it is reasonable to expect that lithospheric strength will be reduced by rifting events. Weakening due to a stretching event may not be permanent. However, numerical models demonstrate that the long-term thermal structure beneath rift basins can remain warmer and hence weaker than surrounding crustal blocks for 10^7 to 10^8 years after the initial stretching (Hansen and Nielsen, 2003). Thus, uplift will tend to be localised at sites of prior extension, because long-term weakness (and therefore susceptibility to deform) persists there.

<u>Material Properties of the Brittle Layer</u>

The pre-inversion structural fabric within the brittle (upper) part of the lithosphere is the dominant control over the locus and the mechanism of subsequent deformation events. The structural fabric is defined by the intensity and orientation of geological faults (fractures). The potential for fault reactivation is described by the cohesive strength and coefficient of internal friction of the fault zone relative to the surrounding (intact) rock mass, as expressed by the Coulomb criterion for failure:

$$\tau = \tau_0 + \mu(\sigma_N - P) \qquad \text{(Hubbert and Rubey, 1959)}$$

Where:

τ is the critical shear stress required for faulting

τ_0 is the cohesive strength

μ is the coefficient of internal friction

σ_N is the normal stress acting on the fault plane

P is the pore fluid pressure

Unlike intact rock, the cohesive strength of a previously fractured rock will be very small, or negligible, relative to the critical shear stress required for fracturing. Mechanical behaviour will thus tend towards reactivation rather than formation of new faults, provided the pre-existing faults are suitably oriented. Reduced friction (Krantz, 1991) and increased pore pressure (Sibson, 1985) decrease the shear strength of a fracture and increase the likelihood of its reactivation in response to subsequent compression. These concepts are summarised on a schematic Mohr diagram (Figure 2.3).

Trajectory of the Pre-existing Fault

In a homogeneous rock volume, the preferred orientation for fracture formation is that orientation in which the minimum critical stress is required for formation of a new fracture. For an Andersonian stress field (three orthogonal principal stresses, with one vertical and two horizontal) faults develop at ~30° to σ_1, in accordance with Coulomb failure theory (Anderson, 1905; see also Figure 2.3). Where the maximum principal stress is vertical, normal faults will dip at ~60° to the horizontal. Where the maximum principal stress is horizontal, reverse faults will develop at ~30° to the horizontal. The actual range of fault dips within a sedimentary basin will be more variable; for example, faults rotate in response to movement on neighbouring faults ('dominoing'). Since the preferred orientation for compressional fracture formation is ~30° to the horizontal, pre-existing normal faults with low angles of dip require less stress to reactivate in compression relative to high angle fractures (Sibson, 1995). It follows that inversion should initiate in areas of greatest former extension (shallowest dipping faults), before spreading to surrounding areas as the earliest reactivated faults steepen their neighbours by reverse dominoing.

Listric normal faults (e.g. Jackson, 1987) have a depth-variable dip and therefore a depth-variable susceptibility to inversion. Lower dips at deeper structural levels will allow compressional reactivation. At higher levels, the fault may not be reactivated and strain may instead be accommodated through the development of 'short-cut' thrusts in the footwall (e.g. Huyghe and Mugnier, 1992). This concept is illustrated in Figure 2.4.

Length of the Pre-existing Fault

Commonly, normal faults are arranged in en echelon fashion, linked by relay ramps (e.g. Peacock and Sanderson, 1994). In such instances, fault relays tend not to invert whereas the normal fault segments do (from their centres outwards). An example is the Portland-Wight Basin (within the Wessex Basin of Southern England). The northern margin of the Portland-Wight Basin is defined by several discontinuous extensional fault segments rather than one single through-going fault. The fault segments are linked by relay ramps including Chaldon

Down (south Dorset) and Lillecombe Down (Isle of Wight). Figure 2.5 shows the Chaldon Down example, where surface mapping identified a shallow-dipping region in between the uplifted and steeply dipping monoclinal fold limbs of the Weymouth and Lulworth Banks anticlines; the monoclinal folds are linked to underlying reactivated normal faults that have been identified on seismic data (Underhill, 2002). The site of the original relay ramp between two normal fault segments experienced only a limited amount of contractional deformation. Thus, the character of the extensional faults that exist prior to the onset of compression will influence the physical characteristics of that inversion.

2.2.3 The Inversion-Inducing Stress Field

Orientation Relative to the Pre-existing Fault

Typically, the direction of maximum horizontal stress during inversion will not coincide with the earlier minimum horizontal stress direction. In consequence, oblique-slip and/or strike-slip kinematics will be important during inversion (e.g. Lowell, 1995). Analogue models have investigated the significance of fault plane orientation relative to the orientation of the subsequent compressional stress regime (Brun and Nalpas, 1996; Dubois et al., 2002; Del Ventisette et al., 2006). Results demonstrate that obliquity between the fault plane and principal compressive stress direction influences both likelihood and character of inversion, although the detail of this influence is disputed. Brun and Nalpas (1996) concluded that tectonic inversion structures develop even when the compression direction is strongly oblique (including perpendicular) to the original extension direction. That a low degree of obliquity is not a prerequisite for inversion was supported further by Del Ventisette et al. (2006), but this experiment showed inversion structures are more favourably developed where obliquities are low.

Magnitude of Compression

The magnitude of compression (amount of shortening) is a major control over the resulting style of inversion. For example in Figure 2.1, the difference between partial and total inversion is due to differences in the amount of shortening. The magnitude of compression depends on the mechanism causing it and is usually attributed to the transmission of plate boundary stresses into the interior.

2.2.4 The Sedimentary Fill

The effect of the sediment fill on the character of basin inversion has been modelled using sand box (laboratory analogue) techniques (e.g. McClay, 1995; Eisenstadt and Withjack, 1995). In Dubois et al.'s (2002) analogue experiments of oblique compression, inversion was lessened by the presence of a thick post-rift sediment infill at the onset of compression. This is to be expected, because a thicker post-rift equates to a smaller difference between the maximum (horizontal) stress and the minimum (vertical) stress.

Less well understood is the influence of the syn-rift on the intensity and style of inversion. Modelling studies such as Panien et al. (2005) have demonstrated sensitivity of the structural response to the mechanical strength of the graben infill. Certain layers in the syn-rift may act as zones of weakness, influencing the style of deformation. One example from outcrop is the Lulworth Crumple (Wessex Basin), a complex deformation structure within an inverted hangingwall sequence (Figure 2.6). Folding is attributed to disharmonic movement of incompetent Wealden (clay) strata at the contact with competent limestones of the Purbeck Formation (Underhill and Paterson, 1998). The influence of a weak layer within the graben will depend on that layer's composition, thickness and distribution. Mechanically weak materials include certain clays (as above) but especially, halite.

2.3 Halokinesis

Halokinesis refers to the mobilisation and flow of salt deposits within a sedimentary basin. Halokinesis is recognised in numerous basins worldwide and a wide range of salt-structure styles have been identified (Figure 2.7). There are examples from intracratonic basins such as the North Sea (Stewart and Coward, 1995; Stewart and Clark, 1999) and from passive continental margins such as the Gulf of Mexico (Rowan et al., 1999; Trudgill et al., 1999) and West Africa (Spathopoulos, 1996; Hudec and Jackson, 2004). The causes and mechanics of halokinesis are examined in this section.

2.3.1 Physical properties of Halite

Originally horizontal salt can become mobilised into salt structures as a result of the anomalous physical properties of halite (NaCl) relative to the other typical lithologies within a sedimentary basin. Of particular importance are the strength and density characteristics of salt and their implications for flow.

Halite Strength

Byerlee's law (Byerlee, 1978) states that strength increases linearly with depth in the brittle zone of the crust. This experimentally-derived rule applies for most rocks typically found in a sedimentary basin but halite behaves differently. As Figure 2.8 demonstrates, the strength of salt is relatively low at the surface and actually decreases with depth. Factors influencing salt strength include temperature, grain size and water content (Van Keken et al., 1993); hot, fine grained, wet salt tends to be weakest.

Strength is a strain-rate dependent parameter. At high strain rates, salt can deform in a brittle manner (Davidson et al., 1996). But at typical geological strain rates salt deforms as a near Newtonian viscous fluid; deviatoric stresses are linearly proportional to strain rate and viscosity (Childs et al., 1993). This is in contrast to the typically much stronger overburden rocks, which deform in a time-independent, brittle manner.

Halite Density

On deposition, halite is typically more dense than other sedimentary lithologies (Casas and Lowenstein, 1989). Unlike siliclastic sediments, which progressively compact upon burial, salt deposits have very limited potential for compaction (Figure 2.9). Salt can in fact show a small decrease in density with increasing burial; this is due to the volume increase caused by higher temperatures at depth. In consequence, the presence of a subsurface salt layer equates to a density inversion and this introduces geodynamic instability, which can lead to flow under certain conditions.

Halite Viscosity

Most sediments can only deform by brittle mechanisms at crustal levels shallower than 8km (Talbot and Jackson, 1997). At these depths (temperatures below ~800°C) salt does not melt but it can deform by two creep mechanisms; by dislocation creep or by fluid-assisted diffusional creep (Van Keken et al., 1993). Consequently, salt deforms in a manner very different to its overburden. Both creep mechanisms are temperature sensitive and more likely to occur at higher temperatures. Dry salt will only deform by dislocation creep (Urai et al., 1986), but most natural salts contain trace amounts (≥0.05%) of water and this significantly influences the deformation mechanism. These trace amounts of water facilitate pressure solution of material at grain boundaries, enabling diffusional creep to occur (Spiers et al., 1990). As water is present only at grain boundaries, this process is strongly controlled by grain size. Van Keken et al. (1993) describe natural salt sequences as having grain sizes between 5 and 30mm; smaller grain sizes will facilitate diffusional creep. The effective viscosity of salt depends on the deformation mechanisms involved. Viscosity is lower where strain rates and temperatures are high and grain sizes are small; these sensitivities are summarised in Figure 2.10.

Halite Solubility

Halite has an extremely high solubility. Several workers have highlighted the possible importance of salt dissolution in determining the overall style of subsurface deformation (e.g. Ge and Jackson, 1998; Cartwright et al., 2001). Salt volume reduction through subsurface dissolution has been postulated to exceed 40% in the Forth Approaches Basin of the UK North Sea (Cartwright et al., 2001) or even 50% in other basins where burial has remained shallow (Hossack, 1995). Volume reduction calculations are based on estimates of the original depositional halite volume and this is very difficult to constrain. For the example of Cartwright et al. (2001), it is possible to propose a much lower value based on a more conservative estimate of the original halite thickness, by employing an only slightly different facies model. Nonetheless it remains necessary to acknowledge dissolution plays some role in deformation events within salt basins.

2.3.2 Initiation of Halokinesis

Buoyancy

The presence of subsurface salt may give rise to a density inversion as the halite's overburden compacts. In fluid systems, density inversions are inherently unstable and an overlying high density layer would sink into an underlying low density layer in response to gravity. The displaced layer would then rise up in a typically bulbous form until a gravitationally stable morphology is attained. The concept of buoyancy-driven diapirism has been applied to salt and is widely accepted (e.g. Jackson, 1995), but the role of density inversion as a lone trigger is less widely accepted. This is due to two key observations. Firstly, many salt structures formed at depths much too shallow for a density inversion to occur (<<~1km), or else they formed in settings where overburden sequences are less dense than the salt (e.g. present-day Red Sea), ruling out density inversions as the main controlling factor (Jackson and Venderville, 1994). Secondly, experimental data show that sediments in the overburden typically deform in a brittle manner at depths where salt tectonics occurs, and this holds true regardless of strain rate (Weijimars et al., 1993); the simple buoyancy model is therefore inappropriate. A more cautious hypothesis suggests that once diapirs are established by other means, buoyancy forces may cause significant uplift over the salt high, sufficient to pierce thin, weak overburdens (Davison et al., 1996). This seems the most plausible marriage between theory and observation, and is preferred here.

Gravitational Flow

Surface deposits of halite have been observed to flow down shallow inclines in response to gravity; movements as great as two metres per year have been recorded in the salt glaciers of Iran (Talbot and Rogers, 1980). Some workers have alluded to the importance of gravitational movements in subsurface deposits (e.g. Roberts et al., 1990), but the presence of an overburden (confining pressure) and the absence of significant amounts of atmospherically-derived water will dramatically impede this process, if not entirely.

As triggering mechanisms for diapirism, buoyancy and gravitational flow can be effectively ruled out for the Central North Sea. More appropriate triggers are those mechanisms that create thickness variations in the overburden; this can be achieved through differential loading (e.g. Ge et al., 1997) or extensional tectonics (e.g. Venderville and Jackson, 1992).

Differential Loading

In differential loading, a salt layer with an overburden of uneven thickness or density will experience different pressures in different areas and the resultant lateral pressure gradient will drive salt away from the loaded area. This might occur through differential sediment loading in the context of a prograding delta or alluvial fan (Jackson and Talbot, 1986) or perhaps in a compressional foreland setting (Millán-Garrido 2004). Differential erosion (i.e. differential sediment unloading) may be an important factor in the latter scenario (e.g. Huntoon, 1982). Unlike buoyancy, a differential loading mechanism can explain salt movements occurring at shallow depths.

A problem with differential sediment loading as a trigger for halokinesis is that the locus of loading will likely migrate through time. In the case of a prograding delta, locus migration may be a response to fluctuations in sediment supply, subsidence rate or sea level change. This migration, whether progradational or retrogradational, will lead to the superimposition (and partial cancellation) of differential load regimes. For diapirism to be initiated by this mechanism would require sedimentation occurring at a constant, steady rate. This is considered somewhat unlikely, whether in intraplate basins or at passive continental margins.

Extensional Tectonics

Extensional tectonics is perhaps the most widely accepted trigger mechanism for halokinesis, on the basis of considerable evidence from analogue modelling (Venderville and Jackson, 1992; Naplas and Brun, 1993; Jackson and Venderville, 1994). Extension thins and weakens the overburden and this localised thinning exerts a strong differential load on the salt layer; salt will tend to rise beneath the thinned overburden. The term 'reactive diapirism' has been used to describe salt mobility in response to extension rather than 'active diapirism', which suggests forceful intrusion of a salt body into its overburden (e.g. Schultz-Ela et al., 1993). Figure 2.11 shows the analogue model of Jackson and Venderville (1994); a reactive diapir has become established by the time the cover sequence has extended laterally by 20%.

2.3.3 Propagation and Termination of Halokinesis

After halokinesis has initiated, the continued growth of salt structures is enhanced by positive feedback mechanisms. Sediment deposited over a salt high will tend to subside less than sediment deposited in adjacent areas, where compaction effects are greater. This differential subsidence may influence the subsequent pattern of sedimentation, with greatest sedimentation (and loading) away from the salt high. The effect is to encourage salt migration into the existing high (Hodgson et al., 1993). The term 'minibasin' is used to describe sedimentary depocentres, bounded at their margins by salt walls, that are created in response to withdrawal of salt from beneath the basin and into the adjacent salt structures. The evacuated salt provides accommodation space and the process is driven by ongoing sedimentation. Minibasin development often continues until the entire salt sequence has been withdrawn from below the depocentre and the supra-salt sediment package grounds out against the sub-salt section. Since the effective viscosity of a salt layer increases as it thins, total mechanical salt withdrawal is unlikely and a salt 'weld' may be preserved (e.g. Price and Cosgrove, 1990). Figure 2.12 illustrates this concept, and is based on seismic observations from the Central North Sea basin.

A second instance of positive feedback pertains to the concept of 'passive diapirism' (Barton, 1933), applicable in scenarios where salt has penetrated its overburden or otherwise (i.e. through buoyancy) risen to the surface. In such an instance, the salt structure may continue to grow by 'downbuilding', driven by sedimentation; the diapir will maintain its salt crest near the surface throughout its development, growing relative to peripheral sediment packages that sink in response to ongoing sedimentation, thermal subsidence and compaction (Jackson, 1995). Figure 2.13 illustrates the concepts of active, reactive and passive diapirism. There is little agreement as to the applicability of any single model, but each of these three behaviours has been documented in the North Sea basin.

2.3.4 Criteria for Identifying, Dating and Quantifying Halokinesis as Opposed to Tectonic Inversion

Section 2.3 has highlighted the possibility of halokinesis causing uplift of sediments within a basin. In Section 2.2 tectonic inversion was defined in terms of sediment uplift. Since the focus of this research is to document the relative roles of inversion and salt tectonics, it is necessary to clarify how the two processes can be discriminated.

Identifying

To identify uplift caused by halokinesis it is necessary to first identify the presence of mobile salt. Presence of halite within a basin can be verified from borehole data. Significant variations in recorded downhole thicknesses of halite would be an indicator of subsurface salt mobility having occurred within the basin. Evidence of salt mobility can usually be discerned from seismic reflection data. The top and/or base of a halite-dominated sequence may be well imaged seismically, due to a sharp impedance contrast with other layers in the basin. This would facilitate three-dimensional mapping of the salt layer, and enable the presence of variable thicknesses of halite (presumed mobile) to be verified. Where salt has been mobilised, the internal seismic character of the salt layer is often chaotic, in contrast to the relatively concordant seismic facies character of non-mobile, non-salt layers.

Having identified a mobile salt layer, it is then possible to identify locations where sediments in the overburden appear uplifted (resting on salt highs). Overburden highs may have formed through diapiric activity (active diapirism), or due to differential compaction effects whereby sediments over the crest of an essentially incompressible salt diapir compact and therefore subside less than adjacent sediments, creating the appearance of uplift.

Where uplift is attributed to halokinesis there will always be some form of halite thick beneath the uplifted sediment. In contrast, verification of tectonic inversion always requires identification of a fault plane and of uplift in the hangingwall to that fault plane. Where there are variable thicknesses of halite and uplift within the hangingwall, it can be difficult to determine the relative contribution to uplift by tectonic inversion as opposed to salt movement. This problem is best resolved by interpreting the 'most likely' structural scenario, through considering the geometric form of the uplift in relation to structures elsewhere where salt and inversion do not occur together. This valid but somewhat subjective assessment was necessary for several of the structures examined in this thesis (e.g. Chapter 5).

Dating

Constraints on the timing of halokinesis can be derived from seismic interpretation of patterns of onlap, rotation, thinning and erosion in the overlying sedimentary sequence. These patterns, and their interpretation, are discussed in more detail later in the thesis.

Quantifying

Halokinesis can be quantified through description of the geometries that have resulted from salt movement, as interpreted from seismic reflection data.

In summary, it is possible to distinguish uplift due to salt tectonics from uplift due to the contractional reactivation of pre-existing normal faults (inversion). Where both processes occur together, it has historically been difficult to quantify the relative roles of inversion and halokinesis in the creation of structural relief. The area selected for this study (the Norwegian Central Trough) was chosen for its potential to allow such quantification, because the region incorporates areas of salt absence and salt presence in variable thicknesses, permitting opportunity to factor-out the effects of salt in an environment where the regional stress history has been approximately fixed. The manner in which seismic interpretation was undertaken is reviewed in Chapter 3, and examples of halokinesis, inversion and both processes in combination are documented in Chapters 4 and 5.

2.4 Tectonic Inversion in Salt-Prone Sedimentary Basins

This section reviews the principal knowns and unknowns regarding the characteristics of tectonic inversion in areas also affected by salt tectonics. Current understanding is based on previous subsurface studies (from the North Sea and further afield), as well as results from theoretical, physical and numerical models.

2.4.1 Normal Fault Development in Salt-Prone Settings

As a precursor to the discussion of inversion tectonics in salt-prone settings it is necessary to review the present understanding of normal fault development in such settings, remembering the character of the extensional fault system prior to inversion is a key control on the style of that inversion.

A wide variety of normal fault styles have been identified in salt-prone sedimentary basins. Styles range from kilometre-scale listric growth faults to metre-scale radial fault patterns in the overburden to salt diapirs (e.g. Stewart, 2007). Normal faults within salt-prone sedimentary basins tend to differ from those formed in salt-free settings; the presence of mobile salt affects fault linkage, displacement-length relationships, and depocentre development.

In salt-free settings, fault tip propagation and segment linkage is accepted as the principal mechanism for growth of normal fault arrays (e.g. Cowie et al. 2000; McLoed et al. 2000; Walsh et al., 2003). In salt-prone areas there have been comparatively few studies of fault growth, but nevertheless there is sufficient evidence (from those analyses of displacement-length relationships that have been undertaken) to suggest that fault segments also grow by tip propagation and linkage similar to that in salt-free settings (e.g. Richardson et al., 2005). However, salt movements during fault growth typically cause depocentres to evolve very differently from salt-free settings, with accommodation space controlled by salt movements as well as by faults. Sometimes (as in the case of the Norwegian continental shelf described by Richardson et al., 2005) salt movements rather than fault movements are the dominant control on depocentre development.

Presence of salt layers influences the propagation of faults between basement and cover. The weakness of salt on geological timescales renders it extremely inefficient at transferring shear stresses; this leads to kinematic decoupling of the sub-salt and supra-salt sequence. Thickness of the salt layer is a key control over linkage between the sub-salt and supra-salt fault populations. For example, in a study of the Dowsing Fault Zone in the Southern North Sea by Stewart et al. (1996), along-strike differences in the basement-cover relationship were identified and attributed to variations in salt layer thickness (Figure 2.14). Of particular importance is the ratio between displacement on the controlling fault and thickness of the salt (Stewart and Clark, 1999). In scenarios where salt thickness is constant, differences in the basement-cover relationship may develop due to along strike variations in fault displacement. For example, where a relay zone exists in an area of uniform salt thickness, the basement and supra-salt faults may be kinematically soft-linked within the relay but hard-linked (i.e. directly coupled) where normal offset is greater outside the relay.

In basins of the Central North Sea the sub-salt to supra-salt coupling relationships are complicated by active salt tectonics; salt thicknesses are locally highly variable and fault patterns are further complicated by this. Figure 2.15 illustrates the variety of structural styles that occur due to variations in salt thickness and distribution relative to Late Jurassic rift development in the Central North Sea area. Additional controls on the linkage between sub-salt and supra-salt fault populations have been identified from experimental modelling by Withjack and Calloway (2000). The extent of decoupling is increased where salt is less viscous, overburden is thinner and strain rate is lower.

In summary, there is a widespread consensus that normal faults within salt-prone basins develop with significantly different controls from those that develop in salt-free settings. However, there remains a lack of consensus regarding certain aspects of salt's influence on depocentre development. An important aspect of this study has been to address the lack of consensus by interpreting the influence of salt movements on the architecture and evolution of depocentres in the Norwegian Central North Sea.

2.4.2 Tectonic Inversion in Salt-Prone Settings

Where tectonic inversion occurs in a salt-prone setting, the characteristics of inversion will depend on the properties of the salt as well as the pre-existing faults. Recent analogue experiments have simulated tectonic inversion of sedimentary basins where salt occupies part of the syn-rift (e.g. Dooley et al., 2005; Del Ventisette et al., 2006). These studies aimed to replicate seismic observations of structural styles in salt-prone inverted basins and to explain these styles in terms of salt-fault interaction. Del Ventisette et al. (2006) noted the tendency of salt diapirs to develop at the basin margins and to localise fault development in the overlying sequence (Figure 2.16). In a series of sandbox experiments, Letouzey et al. (1995) examined the effect of different pre-compression salt geometries on the subsequent inversion. Figure 2.17 is a summary of their findings. Where salt is structured prior to compression there will be density inhomogeneities in the overburden; this will influence the behaviour during compression, including location of subsequent thrusts and folds (compare Figure 2.17b with Figure 2.17a).

Pre-existing salt structures (formed under an extensional stress regime) may themselves be modified by the compression that is causing contractional fault reactivation. Following detailed study of the UK Central North Sea salt diapirs, Davison et al. (2000) proposed criteria for recognition of compressionally reactivated diapirs. These criteria are summarised in Figure 2.18 and include: a thick lid of overburden strata (>300m) up-domed above the diapir; continued upward movement of the diapir even when pinched-off from source; and concentric thrust faults localised at the diapir shoulders.

With the exception of these few aforementioned studies, there remains little understanding of the role of salt tectonics in modifying the response to regional compression (and thus the characteristics of inversion). This study is designed to address the knowledge gap through the study of a structurally complex area of the Central North Sea where both salt mobility and inversion are known to have occurred to varying degrees. This study seeks to quantify these processes, to understand the relationship between them, and to identify the consequences of their (inter)relationship for the structural evolution of the area.

2.5 Introduction to the Norwegian Central Trough Study Area

The Norwegian Central Trough is that part of the Central North Sea's Permian-Cretaceous aged graben system (the Central Graben) that exists within Norwegian waters (Rønnevik et al., 1975). It is a NW-SE trending zone comprising a complex of grabens and half grabens of mainly Late Jurassic age. For the purpose of this investigation, the Norwegian Central Trough study area is defined by the extent of available seismic data; this area exceeds 5,200 km^2 and is shown in Figure 2.19. The study area includes a small region within UK waters (allows analysis across the entire graben) and omits the easternmost, southernmost and northernmost (peripheral) parts of the Norwegian sector.

2.5.1 Tectonic History

Caledonian Orogeny

Amalgamation of the continental crust that forms the present day basement to the Central North Sea occurred during the Caledonian Orogeny. The Caledonian Orogeny is defined (McKerrow et al., 2000) so as to include all the Cambrian, Ordovician, Silurian and Early Devonian tectonic events associated with the development and closure of those parts of the Iapetus Ocean that were situated between Laurentia (to the NW), Baltica and Avalonia (to the SE and East). The timings of collision are derived from palaeomagnetic data (convergence of palaeomagnetically derived latitudes) and by dating the timing of faunal mixing. These methods permit palaeocontinental reconstructions (e.g. Torsvik et al., 1996) and have demonstrated that the main closure of the Iapetus Ocean occurred in Silurian to Early Devonian times (430-400 Ma), with formation of thrust systems including the Moine Thrust, emplacement of which is dated at 425+/-5Ma (Freeman et al., 1998) . A summary reconstruction is presented in Figure 2.20.

As originally separate crustal components, Laurentia, Baltica and Avalonia had different histories and different compositions, and the boundaries (suture zones) between them may have retained physical or mechanical characteristics important to the post-Palaeozoic evolution of the region. It is therefore desirable to understand the deep structure of the Norwegian Central Trough, given its location near the centre of the triple-plate collision zone. Two dimensional deep seismic reflection profiles have been acquired in various locations in attempt to constrain the suture zones where they extend into the North Sea basin (and are buried beneath considerable thicknesses of younger sediments). The locations of surveys which have helped to constrain the Caledonian suture zones are shown in Figure 2.21. Suture zones have been identified from moderate to steeply dipping mid to deep crustal reflections, interpreted as evidence of past subduction activity at the plate-margin. For example, NNW-dipping mid-crustal reflections highlight NNW-directed subduction of Avalonia beneath Laurentia, as identified from the British Institutions Reflection Profiling Syndicate (BIRPS) NEC profile (Klemperer and Hurich, 1990). Similarly, North-dipping reflections on the MONA LISA profile 1 suggest north-directed subduction under the Baltic shield (Abramovitz et al., 1998).

In Figure 2.21, the mapped surface locations of the Laurentia-Avalonia and Baltica-Avalonia sutures are close to the study area, but both suture zones have a down-dip to the north direction, extending north from their mapped surface locations and away from the study area. The Norwegian Central Trough study area is therefore transected by neither the Iapetus nor the Tornquist sutures. The Norwegian Central Trough can be confidently located on an Avalonian basement. It seems probable that basement heterogeneities related to the Caledonian Orogeny had minimal influence on the subsequent evolution of the region. This is useful as it allows the Permian to Recent tectonics to be studied without complications of a heterogeneous basement influencing this history. It is not possible to determine the extent of internal heterogeneity within the Avalonian basement of the Norwegian Central Trough area, due to lack of evidence. On the single deep seismic line GECO NSDP 85-6 that transects the study area there is no clear evidence of a pre-Permian structural grain (Klemperer and Hobbs, 1991).

Variscan Orogeny

Avalonia was the northernmost of a group of continental fragments (microplates and terranes) originally derived from the southern continent of Gondwana. These other Gondwana-derived fragments, including Armorica and Iberia (Figure 2.20c), converged with and successively accreted onto the northern continent (Laurentia, Baltica and Avalonia now combined as 'Laurussia') through Devonian to Early Permian times, resulting in the Variscan (or Hercynian) Orogeny (e.g. Cocks et al., 1997).

The effects of the Variscan Orogeny extended throughout its foreland (e.g. Corfield et al., 1996); Variscan deformation has been identified as far north as the Midland Valley of Scotland (Underhill et al., 2008). The Variscan Orogeny was the last pervasive orogenic event to result in crustal shortening in NW Europe. Thereafter, the tectonic evolution of NW Europe is characterized by the development of Permian-Mesozoic extensional basins in an essentially intraplate setting.

Post-Variscan Tectonics

The principal halite deposit in Western Europe is the Permian-aged Zechstein Supergroup. While there are other formations comprising halite, none match the Zechstein in terms of purity and thickness accumulated. It is the Zechstein that has exerted a unique halokinetic influence on structural styles in the North Sea basin.

Zechstein salts were laid down in the North and South Permian basins; two distinct but interconnected East-West oriented salt basins that formed across a large area of Western Europe during Permian times (Figure 2.22). Development of these basins has been attributed to thermal subsidence following the cessation of Rotliegend volcanism (Taylor, 1988). The basins were largely but not completely separated by the Mid North Sea- Ringkobing-Fyn High. Cartwright (1989) identified that an Upper Carboniferous (Westphalian) to Lower Permian sequence extends along the full length of the present-day Danish Central Graben, suggesting this area formed a deep connecting channel during Zechstein times. It has been argued on sedimentological grounds that the central highs were themselves breached as early as the Late Permian Thuringian transgression (Glennie, 1986).

In the South Permian Basin, the Zechstein Group has been subdivided into at least five depositional cycles, each representing a distinct transgressive event (Glennie et al, 2003). Each Zechstein cycle began with the deposition of limestone over shallow shelf areas and organic-rich calcareous shales in the basins. This was followed by anhydrite and then halite-dominated sequences as sea level progressively fell (Figure 2.23). Observations from modern sabkha environments suggest that halite-dominated facies require some 80% of the original brine to be evaporated (Glennie, 2003), although the great thickness of accumulated halite suggests evaporation occurred in basins that were not completely cut-off from an open marine environment (in this case, the Boreal Ocean).

The lithofacies characteristics of Zechstein deposits in the North Permian Basin are more difficult to discern, and are less well understood, primarily due to the relative lack of well penetrations. Since deposition of the Zechstein Group and growth of the North Permian Basin has been linked to contemporaneous extensional tectonics (Hodgson et al., 1992), the proportion of halite probably increased up the vertical succession and the depositional thickness of halite was greater in deeper, less marginal parts of the basin. Towards the centre of the North Permian Basin, the thickness of the Zechstein Group prior to halokinesis has been estimated to exceed one kilometre (Cameron et al., 1992). The Norwegian Central Trough study area lies at the southern margin of the North Permian Basin, and the original thickness of Zechstein deposited in this area was variable and much less in places.

During the Triassic, thermal subsidence and renewed tensional stresses led to development of a network of half grabens in the North Sea area. Triassic faults striking predominantly North-South are widely documented in the Northern North Sea (Tomasso et al., 2008) but this trend is less clear in the Central and Southern North Sea due to extensive overprint by later tectonics. Most workers infer East-West Triassic extension throughout the North Sea Basin, assuming faults developed in response to the same far field stress conditions (e.g. Underhill, 2003).

Thermal subsidence appears to have continued until Early/Middle Jurassic Toarcian times, when the North Sea region began to be influenced by uplift that has been attributed to thermal doming (Underhill and Partington, 1993). Late Jurassic collapse of this thermal dome resulted in a major phase of basin formation, establishing the trilete rift arms that have dominated the structure of the North Sea Basin ever since (the Viking, Moray Firth, and Central Grabens). The kinematics of Late Jurassic faulting has been widely discussed. Some studies have attempted to explain fault trends in terms of a single regional stress regime (e.g. Roberts et al., 1990). Other workers have noted that a single stress direction requires that significant strike-slip movements occurred within individual rift arms (e.g. Bartholomew et al., 1993). However, most interpretations do not recognise major strike-slip; indeed Jackson and McKenzie (1983) present faults from the Inner Moray Firth's Beatrice Field as type examples of normal faulting. Davies et al. (2001) proposed an alternative to the strike-slip explanation; that not all Late Jurassic fault sets were active contemporaneously, and that each graben experienced its own unique stress regime. The dominant Upper Jurassic fault orientation in the Central Graben was NW-SE.

The Lower Cretaceous is characterized by thermal subsidence within the North Sea Basin (Underhill, 2003). Subsidence continued into the Late Cretaceous and Cenozoic, albeit punctuated by episodes of compression (inversion). These inversions are reviewed more fully in the next section, and a detailed tectonostratigraphic history of the Norwegian Central Trough is presented in Chapter 4, based on seismic observations from this study.

2.5.2 Tectonic Inversion

There have been few studies of tectonic inversion specifically within the Norwegian Central Trough. Early interpretations of the structural framework were based on 2D seismic data that was limited in extent and often poor in quality. This was especially problematic because of the region's structural complexity; no single structural style or trend dominates and uplifts are the product of multiple factors. Various and contradictory models (e.g. Roberts et al., 1990; Sears et al., 1993; Gowers et al., 1993) were a consequence of the low quality input data. More recent studies have done little to refine these models, focusing instead on the improved characterization of individual fields (e.g. Brasher, 1995; Farmer and Barkved, 1999). Some of the best recorded evidence for the timing and intensity of inversion comes from other parts of the European Plate. It is necessary to review this material as a precursor to the study of tectonic inversion in the Norwegian Central Trough itself.

Timing and Intensity of Inversion

The history of compression in Western Europe is complex, with more than one pulse of inversion recognised in most basins. Figure 2.24 shows the European locations where tectonic inversion has been documented. Figure 2.25 summarises the timings of these inversions, as compiled from an extensive amount of published literature. There are broad trends in the timing and intensity of tectonic inversion. In general, the intensity of basin inversion decreases with increasing distance from the Alpine thrust front and there is a systematic westward shift in both the timing of onset and the timing of greatest intensity of inversion (Ziegler, 1995).

Ziegler (1990) proposed four major pulses of compression to explain the pattern of West European inversion: Late Cretaceous (Late Turonian to Campanian), Mid Palaeocene, Late Eocene- Early Oligocene, and Late Oligocene- Early Miocene. As illustrated in Figure 2.25, the Late Cretaceous pulse is recognised in southern parts of the North Sea rift (Vejbæk and Andersen, 2002; De Lugt et al., 2003), the Broad Fourteens and West Netherlands basins (Nalpas et al., 1995; Worum and Michon, 2005), the Sole Pit basin (van Hoorn, 1987), and the Lower Saxony basin (Ziegler, 1990). Absolute ages of Late Cretaceous inversion are generally difficult to discern because of limited confidence in discriminating between uplift-induced and eustatic-induced sequence boundaries using well data, and the unsuitability of Apatite-Fission-Track-Analysis techniques or similar (e.g. Green et al., 2002), due to the Upper Cretaceous Chalk lithologies and often only mild amounts of uplift.

Late Cretaceous inversion was followed by a Palaeocene phase of intraplate compression that affected approximately the same areas, as well as basins further west (the Weald, Channel, Western Approaches, Bristol Channel, and Celtic Sea Basins). According to the recent review paper by Ziegler and Dèzes (2006), this Palaeocene phase can be dated at 65-55 Ma and was more intense than the Late Cretaceous episode. They identify it as the main phase of inversion in the Southern North Sea, where it coincides with a major disconformity at the Cretaceous-Palaeogene boundary, but inversion remains relatively mild in western basins.

The Late Eocene- Early Oligocene inversion pulse has been described in the Danish Central Graben, the Broad Fourteens and West Netherlands basins, the Lower Saxony and Sole Pit basins, and basins further west (e.g. Gale et al., 1999). A final pulse of inversion dated to Late Oligocene- Early Miocene has been identified in the Broad Fourteens and West Netherlands basins, the Sole Pit basin and in the western basins (Weald, Channel, Western Approaches, Bristol Channel, Celtic Sea, and Irish Sea Basins). This event constitutes the main inversion phase of the western basins, while its effects were relatively mild in the Southern North Sea area (Ziegler, 1990). Overall, the subdivision of observed inversions into four distinct episodes appears at best a poor fit. While there are broad similarities in timing across the European Plate, there are often major differences between adjacent basins. It is desirable to determine how inversion identified within the Norwegian Central Trough fits into this regional framework.

Causes of Inversion

Inversion is understood to have occurred in response to the transmission of plate boundary forces into the interior of the European Plate (Lowell, 1995). In addition to collision-related stresses, controlled by relative plate motions and the overall pattern of plate reorganization, ridge push has been postulated by some workers as an additional mechanism by which to transmit forces into the plate interior (e.g. Richardson, 1992; Ziegler, 1995). Ridge push is a tectonic force that arises from lateral density variations within the oceanic lithosphere. A horizontal pressure gradient develops away from the hot, low density and isostatically uplifted asthenosphere at a spreading ridge to the cooler, denser oceanic material away from it (Golke and Coblentz, 1996). The strong correlation between inferred ridge push force direction and the azimuth of absolute plate motion, as determined from present day intraplate stress direction data (e.g. Zoback, 1992; Reinecker et al., 2005), has been taken as evidence that the ridge push force accounts for a significant proportion of the intra-plate stress field.

The relative contribution of the different plate boundary processes as a cause of European inversion remains disputed. This is because the Palaeozoic to Recent evolution of NW Europe was governed by a complex, concurrent and continually changing sequence of plate boundary adjustments. The inversion history is dominated by two key events; Alpine collisional tectonics (that post-dated the closure of the Tethys Ocean) and tectonics pertaining to the opening of the North Atlantic Ocean (Ziegler, 1990; Nielsen et al., 2007). Figure 2.26 is a schematic illustration of the Cretaceous through Cenozoic history of the European Plate, as discerned primarily from palaeomagnetic reconstructions (Coward et al., 2003).

Convergence of Africa and Eurasia is considered here to be the single most importance influence on the Tertiary to Recent tectonics of NW Europe. Convergence began during the Cretaceous Normal Superchron, between 120–83 Ma (Rosenbaum et al., 2002). Africa-Arabia rotated in a counter-clockwise direction and began to collide with the southern margin of Eurasia in response to the gradual opening of the Southern (and Central) Atlantic ocean. This led to NE-directed subduction of the Eastern Tethys, with the development of back-arc basins in the Carpathians, Black Sea and Caspian regions at this time (Figure 2.26a). By 120 Ma the Atlantic Ocean had propagated northward, with seafloor spreading beginning in the Biscay area. A triple junction developed here during the Late Cretaceous, with coeval openings of the Bay of Biscay, incipient North Atlantic and Labrador Sea (Figure 2.26b). The Late Cretaceous marks the first recorded pulse of West European inversion (Ziegler, 1990), with widespread deformation in the Western Mediterranean (Alpine) area attributed to ongoing convergence between Africa and Eurasia.

Recent palaeomagnetic reconstructions identify a pronounced slowing of the rate of convergence between Africa and Eurasia towards the end of the Cretaceous (e.g. Rosenbaum et al., 2002; Smith, 2006). North- South convergence rates fell from as much as 20mm/yr to almost zero for several millennia, before rising again in the Late Eocene. Smith (2006) dates the interval of slow convergence at 75-45Ma. Noting that its onset follows the final emplacement of ophiolites in Turkey, Syria and Cyprus (figure 2.27), Smith (2006) attributes the slowing to cessation of Tethyan subduction and the accompanying change in the mode of Alpine collision. The inference here is that events occurring at a convergent plate margin can exert a primary control over rates of relative plate motion, and thus the intraplate stress regime. An alternative, but perhaps less likely, explanation for the slowing in relative motion between Eurasia and Africa is that this relative motion was increasingly taken up by displacement activity in the proto-North Atlantic. At this time, the North Atlantic was strongly affected by emergence of the Iceland Plume (White and Lovell, 1997; Doré et al., 1999).

Seafloor spreading between Greenland and Europe began during the early part of the Eocene (55.9-53.3 Ma) and continued into the Early Oligocene (Mosar et al., 2002). Spreading had ceased in the Bay of Biscay and Labrador Sea areas by this time (Figure 2.21d). Alpine tectonics during the Eocene and Oligocene are dominated by dextral translations between Europe and Africa occurring in conjunction with or as a response to differential seafloor spreading rates in the Central and North Atlantic. The Miocene (Figure 2.26e) is characterised by ongoing compression in the Alpine foreland (thin-skinned thrust tectonics), although the generation of new oceanic crust in the Western Mediterranean area suggests stretching was also locally important (Coward et al., 2003). North Atlantic spreading continued throughout the Cenozoic, but the locus of spreading migrated progressively westwards.

Figure 2.28 summarises the major Permian to present tectonic events that have affected Western Europe. Some workers have subdivided the Tethys-related orogenic episodes into separate phases termed Austrian, Laramide, Pyrenean and Late Alpine (e.g. Ziegler, 1990). For the purpose of naming and describing the West European inversion events themselves, this terminology is considered unhelpful because the inversions are in fact rather more complex products of multiple and inter-related plate boundary events, as has been demonstrated in the preceding discussion.

2.5.4 Rationale for choice of Norwegian Central Trough as study area

The Norwegian Central Trough coincides with the southern margin of the North Permian Basin. As an area where varying amounts of halite were originally deposited, it is ideally suited for studying the role of salt in modifying structural development, and the relationship between salt mobility and tectonic inversion in particular. Unlike many other inverted NW European basins, uplift occurred against a background of near-continuous Cretaceous and Cenozoic deep marine sedimentation. The sediment record has therefore preserved stratigraphic markers that can constrain the timing of inversion, unlike other locations where sediments were either never deposited, or were subsequently exhumed and eroded.

The Norwegian Central Trough is also of interest because of significant hydrocarbon accumulations within Late Cretaceous reservoirs. The tectonic inversion history of the Norwegian Central Trough remains poorly understood, in spite of the region's economic importance. Reservoir properties may have been affected by fracturing related to inversion and/or halokinesis. An advantage of the area's commercial significance is that there is a relative abundance of subsurface data with which to address the scientific objectives.

2.6 Aims of research

The overall objective of this thesis is to document the manner in which halite within a sedimentary basin can influence the characteristics of inversion. This aim is achieved through detailed study of subsurface data from the Norwegian Central Trough; a structurally complex area and an ideal setting in which to study compression in a salt-influenced basin.

Specific Aims relating to the Norwegian Central Trough

To assess the influence of mobile salt on the sedimentary and structural response to compression it is necessary to constrain:

- Timing(s) of inversion
- Timing(s) of halokinesis
- The spatial extent and magnitude of inversion
- The spatial extent and magnitude of halokinesis

These parameters were documented through detailed seismic interpretation, with particular emphasis on the accurate correlation of regional and local sequence boundaries within the Upper Cretaceous and Lower Cenozoic section. The documented evidence for inversion and halokinesis constitute a valuable output from this research.

The interpretations undertaken have sought to determine the manner in which salt-influenced inversion has affected reservoir quality of the Upper Cretaceous Chalk Group. Assessment of the sedimentologic and fracture characteristics of the Chalk Group was required to explore the way in which inversion and/or halokinesis may modify the reservoir characteristics.

General Aims

By supplementing subsurface interpretations with observations from onshore outcrop analogues, the aim is to establish a set of generic rules to explain the inversion characteristics identified in salt and inversion influenced sedimentary basins. This includes scales beneath the limit of seismic resolution.

3 Subsurface Data and Methods

3.1 Introduction

Analysis of stratigraphic relationships is the means by which to constrain the geometric and temporal characteristics of tectonic inversion and halokinesis in the Norwegian Central Trough. Three dimensional analysis of subsurface stratigraphy and structure is achieved through assessment of well-calibrated seismic reflection data. This chapter reviews the extent and quality of available data and the manner in which it was used to study structural processes occurring in the subsurface.

3.2 Seismic reflection data

Seismic interpretation was undertaken on in-house workstations located at the University of Edinburgh using Schlumberger GeoQuest IESX Version 4.2 software. The seismic reflection data used in this study consists of two partially overlapping 3D migrated volumes. The area of investigation exceeds 5,200 km^2 and is defined by the extent of this seismic coverage, delineated in Figure 3.1 (see also Figure 2.19). The studied area lies entirely within the Norwegian and UK sectors of the Central North Sea and straddles the full width of the Norwegian Central Trough.

The two seismic surveys were obtained in digital format from ConocoPhillips. They are pre-stack time migrated, with processing having conformed to standard sequences for regional surveys of this type. The larger of the two seismic surveys, *CNSmerg*e, is itself a composite of several smaller surveys acquired during the 1990s. Each of these small surveys had inline and crossline spacings (bin sizes) not exceeding 25m by 25m, and they were merged via a standard trace relocation method into 12.5m by 12.5m bins. Acquired in 1993, the *ga3D93* survey has an inline spacing of 12.5m, a crossline spacing of 25m, and partially overlaps the SE corner of the *CNSmerge* survey. Thus, the Norwegian Central Trough seismic dataset is both large and densely sampled. Appendix 1 summarises *CNSmerge* and *ga3D93* survey locations and line spacing information in more detail.

Both surveys have vertical axes in two-way travel time (TWTT). It was deemed beyond the scope of this study to undertake detailed depth conversion of all the TWTT data. Industry-standard techniques for precise time-to-depth conversion of 3D data require access to considerable computer processing power, unavailable to this study. However, specific 2D sections were converted to depth where appropriate, using a comparatively simple one-dimensional approach (Section 3.5). The reflection data extend to a minimum of 6.0 seconds TWTT. This corresponds to depths of ~8km, sufficient to allow all sediments within the Norwegian Central Trough to be studied. The seismic data used in this study are therefore unlike most data previously shot across the Norwegian Central Trough, which target only the shallower prospective parts of the basin.

A consequence of working with a composite of seismic surveys is that there are inconsistencies in seismic character between surveys. In particular, there are often differences in the amplitude of the seismic response. This observation can be explained by differences in acquisition parameters and processing methodology from one survey to the next. In general, the boundaries between merged surveys are easy to identify and do not impair the geologic analysis of the composite dataset.

The densely sampled nature of the cnsmerge and ga3D93 surveys allows for a detailed three dimensional study. It was deemed unnecessary to make interpretations at the maximum level of detail; that it is possible to preserve the integrity of interpretation when using coarser inline and crossline spacings was confirmed by empirical observation. For this study, interpretations were performed at spacings of not more than 1,000m, and typically 400m. It was then possible to interpolate these line interpretations into a complete 3D image. Criteria for the choice of line spacing were the perceived importance of the horizon for pursuit of thesis objectives, and the lateral complexity of the surface. More complexly folded or faulted surfaces required more detailed interpretation. A full discussion of the approach to seismic interpretation, including details of the horizons interpreted, is presented in Section 3.4.

3.3 Well data

The seismic data was calibrated using borehole information. Well data is important for the dating and regional correlation of seismic horizons, and it provides direct information about the nature and sedimentologic characteristics of the basin-fill.

Hydrocarbon exploration and production activity has been ongoing in the Norwegian Central Trough since the late 1960s and hundreds of wells have been drilled in the course of this activity. It was neither practical nor necessary to make use of all of this well data. Accurate regional correlation of seismic horizons across the Norwegian Central Trough was achieved using depth constraints from 43 wells (locations shown on Figure 3.1). The 43 wells were chosen such that they were widely distributed throughout the study area. Few wells drilled in the Norwegian Central Trough penetrate beneath the Upper Cretaceous Chalk Group. Thus, for the purposes of regional correlation the deepest wells were favoured because these wells provided stratigraphic constraints down to the (sub-salt) Permian Rotliegend Formation.

Well tie information was provided by ConocoPhillips (COP), as measured depths recorded in metres. The information (depth to formation tops) has been determined from lithologic and biostratigraphic interpretations of core and/or geophysical logs. The precision of the recorded depths was variable, often given to the nearest millimetre but sometimes stated to only the nearest metre. This reflects variability in the confidence with which certain lithostratigraphic tops can be constrained. Regardless, the precision with which all tops are given is beneath the limits of seismic resolution so any uncertainty is deemed insignificant for the mapping of seismic horizons. Independent verification of the values provided by COP was beyond the scope of this study.

Measured depth values were converted to two way travel times using check shot data provided by COP. It was possible to verify the gross reliability of the depth to time conversion for each well by comparing the computed top Rogaland Group and top Humber Group TWTTs with their TWTTs inferred directly from seismic: the top Rogaland Group and top Humber Group horizons are readily identified on seismic data because of their consistent and strongly defined seismic signatures. A degree of non-trivial mistie was often recognised but typically did not exceed 20msec and thus fell within the limits of seismic resolution, allowing TWTTs calculated for other events to be plotted with confidence.

Should this mismatch have proved a more significant problem, an improvement in the well-to-seismic tie could have been sought through the construction of synthetic seismograms from geophysical log data. A summary of the well data used for regional correlation is included in Appendix 1.

3.4 Resolution and Uncertainty

Resolution is defined as the minimum separation of two features before their individual identities are lost (Sheriff, 1991). The resolution of seismic data limits the degree of detail with which subsurface structures can be described.

Vertical resolution is the minimum distance between two interfaces needed to allow two distinct reflections on a seismic profile. It is controlled by the wavelength of the source signal: the minimum vertical resolution is generally one quarter of a seismic wavelet (e.g. Badley, 1985). Thus, shorter wavelengths (higher frequencies, lower velocities) correspond to improved vertical resolution. Since deeper-travelling seismic waves tend to have a lower dominant frequency due to progressive loss of higher frequencies by absorption, and higher velocities due to the effects of sediment compaction, vertical resolution decreases with increasing depth. For depths of interest in this study and given that typical acquisition frequencies were involved, vertical resolution is estimated to vary within the range of 25 to 50m. This is very small relative to the dimensions of the basin under investigation.

The horizontal resolution of the seismic data also depends on the dominant wavelength of the source signal. For unmigrated seismic data, horizontal resolution is defined by the Fresnel Zone radius, which is the radius of the area on the reflector surface from which energy is returned within a quarter wavelength after the onset of reflection (and is thus returned as a coherent, in-phase arrival). The Fresnel Zone radius is usually ~250m to 750m depending on the frequencies, velocities and TWTTs involved (Sheriff, 1991). Migration improves horizontal resolution. For the densely gridded 3D-migrated seismic data in this study, lateral resolution is similar to vertical resolution, and is not considered a limitation for structural and stratigraphic interpretations.

An additional factor to consider when discussing seismic data quality is the presence of Seismically Obscured Areas (SOAs) that affect small areas of the dataset. SOAs exist exclusively within and above certain Chalk Group antiforms, and are attributed to the presence of leaked gas in the reservoir overburden (e.g. Farmer et al., 2006). An SOA over the Ekofisk field is the most prominent example, and measures some 2.4 km (W to E) by 2.7km (S to N). Although SOAs are a major problem for Chalk field production activities and reservoir monitoring, their presence does not hamper this regional interpretative study.

3.5 Seismic Interpretation technique

The purpose of seismic interpretation was to map the structural and stratigraphic characteristics of the Norwegian Central Trough in sufficient detail to enable analysis of the role of tectonic inversion and halokinesis in the evolution of the basin. In order to examine the influence of Permian-aged evaporites upon structural style it was necessary to map sub-salt as well as supra-salt horizons. Hydrocarbon exploration activity in the Norwegian Central Trough has tended to focus on relatively shallow strata of the Upper Cretaceous Chalk Group, rather than the more deeply buried (including sub-salt) strata. These deeply buried sediments are mostly post-mature with respect to hydrocarbons and consequently there is no commercial incentive for their study. Mapping the deeper structure of the Norwegian Central Trough in this study is therefore an important result in its own right.

Eight horizons were interpreted across the entire Norwegian Central Trough survey area, covering approximately 5,200 square kilometres: top Rotliegend Formation (base salt), top Zechstein Group (top salt), top Humber Group, top Cromer Knoll Group, top Hod Formation, top Chalk Group, top Rogaland Group and top Hordaland Group. These horizons are presented in their correct stratigraphic context in Figure 3.2. They are among the most prominent seismic markers in the basin, able to constrain the tectonostratigraphic history of the Norwegian Central Trough in some detail. Seismic characteristics of the individual surfaces are discussed in Chapter 4. Further constraints on the geometry and chronology of basin evolution were derived from consideration of additional reflections within the basin-fill sequence. In some instances it was not possible to map these reflections as fully regional 3D surfaces (often because lateral correlation was hindered by weak acoustic amplitudes and inadequacies in well control); in other instances it was not considered necessary to do so.

Mapping was undertaken in accordance with the principles of seismic stratigraphy (e.g. Mitchum et al., 1977; Berg and Woolverton, 1985), whereby age relationships and thus structural histories are revealed through identification of seismic reflection geometries (e.g. downlap, onlap, erosional truncation) and associated unconformities and correlatable conformities. From the initial eight seismic horizon maps, a set of isochron (time thickness) maps were generated using GeoQuest IESX software. Isochrons record the vertical travel time difference between two horizons. They are a useful tool to assess syn-sedimentary growth and hence, the structural history of the basin.

Constraining the magnitude and timing of fault activity is essential for the understanding of tectonic inversion in the Norwegian Central Trough. Recognition of fault displacement is limited by the vertical seismic resolution, such that throws of less than approximately one quarter of the seismic wavelength (c.25 msec) are imperceptible. By contrast, the largest faults have throws in excess of 1,000 msec and were the dominant controls on regional-scale basin development. Due to the vastness of the seismic volume, it was not possible to map each and every observable fault in three dimensions. Interpretation focused on faults with large offsets (>100 msec), faults with uplifted hanging walls and faults connected to the Zechstein Group, since these features are of most relevance to the compressional and halokinetic history of the basin. With this in mind, footwall and hangingwall cutoffs corresponding to the mapped horizons were defined during line interpretation. This allowed manual creation of accurate fault polygons on the plan view horizon maps.

For any interpreted horizon, the IESX software is able to extract various seismic attributes; the resultant attribute maps may facilitate the geological interpretation of the subsurface volume. In this study, the attributes of dip-magnitude and dip-azimuth proved the most useful. These attributes facilitate the identification of discontinuities in the seismic data, and were used for fault and fracture characterization. Dip-azimuth maps were created from several of the horizon interpretations (top Rotliegend, top Humber and top Chalk surfaces).

3.6 Depth conversion

Ideally, seismic time sections should be converted into depth sections so as to provide a velocity independent picture of the subsurface. Given the sizeable area under investigation, depth conversion using a 3D velocity model (e.g. the layer-cake method) was impractical. A one-dimensional approach was considered instead. Time-depth data (checkshot surveys and velocity logs) from the 43 project wells were plotted onto a single graph of two-way-travel-time against true vertical depth (Figure 3.3). The graph shows that velocity at any given depth is relatively uniform and remarkably so for depths less than 3000m. This suggests lateral velocity variations are minor and confirms the suitability of a one-dimensional approach to depth conversion.

At depths greater than 3000m there is some divergence in velocity values. This is attributed to the presence of heterogeneous thicknesses of halite at these greater depths. There is no well-derived velocity information for depths greater than 5,200m (~4,300 msec), but the time-depth curve must be extended to 6,000 msec to ensure the deepest parts of the basin are included in the depth conversion. A best fit time-depth function was calculated using a curve-fitting algorithm within Microsoft Excel:

$$TVD(m)= 9*10^{-5}*[TWT(msec)]^2+0.7918*[TWT(msec)]$$

The best fit time-depth function yields a plausible estimate of the deep velocity structure. The curve has an R^2 value (coefficient of determination) of 0.9943; this is considered an agreeable fit. It is possible to improve the R^2 value by including higher order polynomial terms in the best-fit equation, but these higher order terms were not included because they generate unjustifiably high TVD values when extrapolated beyond 4,300 msec.

The depth conversion was undertaken with Midland Valley's 2D Move software. In terms of overall geometries, the depth converted sections show little difference from the time sections, and this supports the decision to make geological interpretations directly from time sections rather than necessitating depth conversions in every case.

3.7 Section Restoration

Palinspastic restoration of seismic cross-sections is a means by which to visualise the evolution of basin geometry and is achieved through backstripping and fault-block restoration. Backstripping is a technique that models the progressive removal of sediment loads in a manner that incorporates the decompaction and isostatic responses to unloading (Watts, 2001). Restorations were undertaken using 2DMove software in light of these considerations.

Decompaction

Sediment burial is associated with rock volume decrease through loss of porosity. Porosity data for typical basin-filling lithologies (sandstone, shale, chalk) have been investigated by several workers (e.g. Sclater and Christie, 1980; Halley and Schmoker, 1983). Under conditions where pore pressures are hydrostatic (as opposed to over- or under-pressured) the relationship between porosity-at-surface (Ø_0) and porosity at depth (Ø) is observed to conform to a simple exponential relationship of the form:

$$\text{Ø} = \text{Ø}_0 \, e^{-CZ}$$

Where C is the porosity-depth coefficient (in cm^{-1}) and Z is the depth (in m) to which the sediment is buried. The values Ø_0 and C are lithology specific; they were defined by Sclater and Christie (1980) for a typical sandstone, for shale, for shale-sand intermediate and for chalk. The Ø_0 and C values presented in Sclater and Christie (1980) are the guide values for this study; Ø_0 and C values for each formation are shown on the next page. For the halite-dominated Zechstein Group, Ø_0 and C values of 0.01 were chosen in order to reflect the essentially incompressible character of the sequence due to the presence of salt within.

Within the 2DMove software, the decompaction algorithm works by rebuilding the cross-section along vertical ray traces after applying the prescribed correction.

Strata	$Ø_0$	$C^* 10^{-5} cm^{-1}$
Nordaland Group (shale)	0.63	0.51
Hordaland Group (shale)	0.63	0.51
Rogaland Group (sand-shale)	0.56	0.39
Chalk Group (chalk)	0.70	0.71
Cromer Knoll Group (shale)	0.63	0.51
Humber & Hegre Groups (shale)	0.63	0.51
Zechstein Group (halite)	0.01	0.01
Rotliegend Group (sand)	0.49	0.27

Isostatic Adjustment

In addition to modelling compaction due to burial-related porosity loss, it is necessary to consider isostatic adjustments associated with sediment loading. There are two competing approaches to modelling isostasy. One method involves 1D Airy isostatic modelling (e.g. Steckler and Watts, 1978), and the other involves 2D flexural modelling (Watts et al., 1982). The principal difference between the two methods is that in the Airy approach, loads are compensated locally (immediately beneath the load) while with the flexural approach loads are compensated regionally. In the flexural case, crust has a finite (non-zero) lateral strength.

After comparison of both Airy-type and flexural restorations, the flexural backstripping approach was preferred for this study. When applied along a cross-section, the 1D Airy approach tended to impart unrealistic distortions to fault block geometries whereas no such distortions were created by the flexural correction. This preference for flexural backstripping is supported by the observations of other workers; Roberts et al. (1998) state 'predictions of β (stretching factor) from flexurally backstripped models are more in accordance with predictions of β from forward modelling than are the predictions of 1D Airy backstripped models'.

Isostatic adjustments were made using 2DMove software. The following parameters were specified: load density (2680 kgm^{-3}), mantle density (3300 kgm^{-3}), elastic thickness (Te= 5 km) and Young's modulus ($E= 5.0*10^{10}$ Pa). The load density refers to the density of the layer being backstripped from the model, and a constant value of 2680 kgm^{-3} was considered appropriate for a basin comprising mixed siliciclastics (sand density 2650 kgm^{-3}, shale density 2720 kgm^{-3}). The Te value was taken from Barton and Wood (1984) and is considered appropriate for the Central North Sea in general. The E value is correct for a typical shale (Sheriff, 1991), shale being chosen because it is the dominant basin filling lithology in the Norwegian Central Trough.

The validity of section restorations is sensitive to another factor; the presence of halite. Palinspastic restoration of sections with mobile evaporites is somewhat unreliable due to the potential for evaporites to flow out of the plane of a section. This invalidates the plane strain criterion on which 2D restoration depends (e.g. Hossack, 1995). Consequently, section restorations presented in this study are driven (in part) by assumptions based on seismic-stratigraphic interpretation of timing and magnitude of salt movements, rather than themselves being an independent means to infer these observations. Section restorations are nevertheless a helpful tool for capturing insights into structural evolution, advancing understanding of section geometries in terms of tectonic subsidence, uplift and erosion, and global sea level changes.

4 *Tectonostratigraphic Framework for the Norwegian Central Trough*

4.1 Introduction

This chapter presents the results of regional seismic interpretation. Structural styles and stratigraphic relationships are demonstrated from key representative 2D regional seismic lines. The 3D tectonic and stratigraphic history of the Norwegian Central Trough is examined using seismic sections, TWTT horizon maps and isochron maps. The results provide a necessary context for detailed study of Late Cretaceous and Cenozoic inversion tectonics.

4.2 Selection and Interpretation of Regional Seismic Lines

Figure 4.1 shows the location of the nine seismic cross sections presented in Figure 4.2. The lines comprise five parallel and approximately equidistant sections oriented at 065°, and four parallel sections oriented orthogonal to the first five. With lengths in excess of 50km, these nine sections provide a regional overview of the study area. The section orientations were carefully chosen. Lines are oriented parallel and perpendicular to the overall trend of the Central Graben as it passes through Norwegian and Danish territory (see Figure 2.19b). Similarly orientated sections were presented in previously published work from the Danish Central Graben (e.g. Cartwright 1989; Vejbæk and Andersen, 2002).

Figure 4.2 shows TWT interpretations that summarise the major characteristics of structure and stratigraphy within the Norwegian Central Trough. The vertical to horizontal scale is fixed for all displayed sections. Using the time-depth relationship plotted in Figure 3.3, this fixed scale can be interpreted as an approximately two times vertical exaggeration. This vertical exaggeration causes fault dips to appear slightly steeper than in reality, and salt bodies more vertically elongated. This exaggeration was acknowledged during interpretation. Figure 4.2 interpretations are discussed in conjunction with the TWTT horizon maps and isochron maps that are presented chronologically in the following text.

4.3 Palaeozoic

4.3.1 Pre Permian

It is not possible to resolve pre-Permian structures on the seismic data; the top Rotliegend is the deepest mappable reflection. Chaotic reflection signatures in the pre-Permian sequence are indicative of crystalline basement rocks (e.g. at ~5.0s TWT beneath well 1/3-5 on figure 4.2a). On the basis of plate tectonic reconstructions (e.g. Cocks et al., 1987; Torsvik et al., 1996) and deep seismic profiles (e.g. Klemperer and Hurich 1990; Abramovitz et al., 1998), the basement is considered to comprise the Gondwana-derived Archaean and Proterozoic crust of the Avalonia plate (refer to Section 2.5.1 for discussion; also Figures 2.20 and 2.21). The material properties of the crystalline basement (especially internal heterogeneities) will exert some degree of control over the response to plate-scale stresses. An exploration of this influence falls outside the remit of this study.

4.3.2 Lower Permian (Rotliegend Group)

The top of the Rotliegend Group is constrained by seven wells in the Norwegian Central Trough. The horizon is important because it is the deepest seismically mappable surface within the basin and effectively defines the sub-salt and basement structure. The top Rotliegend is mostly recognised as a high amplitude negative reflection; it records a major impedance contrast between high velocity, low density halite (above) and low velocity, moderate density Rotliegend sandstones (below). In some areas the horizon is poorly imaged. This is often true of the deepest parts of the basin, where reflection multiples tend to obscure the primary reflectivity, which is weak on account of the depths involved. Typically, however, the top Rotliegend horizon is oriented differently from horizons within the supra-salt sequence (e.g. Figure 4.2a, NE) such that it is seldom confused with multiples. Thus, the top Rotliegend reflection has been mapped with confidence across the Norwegian Central Trough despite a lack of well constraints relative to the other, shallower horizons. Figure 4.3 shows the top Rotliegend TWTT surface as mapped across the study area. The principal tectonic provinces in the Norwegian Central Trough have been labelled in accordance with the nomenclature of Rønnevik et al. (1975) and Gowers and Sæbøe (1985). The division of the Feda Graben into North-East, South-West and Western segments is new for this study.

The top Rotliegend surface is extensively faulted, as demonstrated in the figure 4.2 sections. This faulting corresponds to net extension, with the amount of extension varying up to ~10% depending on line orientation and location (having been quantified by 2DMove structural restoration, as illustrated in Chapter 5). Extension occurred during Permian to Early Cretaceous times; the syn-rift character of the post-Permian, pre-Cretaceous sequence is clearly shown on several lines (e.g. Figure 4.2a, NE; Figure 4.2d, NE; and Figure 4.2g, NW). The supra-salt section is less affected by these very large offset faults that disrupt the sub-salt sequence. This is largely attributed to the role of salt movements in accommodating strain in the supra-salt sequence. The large majority of the top Rotliegend faults do not extend into the supra-salt sequence. This is an important observation: the Zechstein Group acts as a detachment layer where it overlies the Rotliegend Group. Where there is no overlying salt (or it is very thin) basement faults may extend up into the Lower Cretaceous (e.g. Figure 4.2d, SW; Figure 4.2f, SE).

The reliability of the time surface as a true structural interpretation has been considered. In general, the validity of time section interpretations depends on the velocity structure of the basin; where the velocity-depth profile is smooth, and lateral velocity variations are minimal, the approximation is a good one. In the Norwegian Central Trough, the presence of high velocity halite within the Zechstein Group and the irregular geometries of their distribution cause problems for the time section appearance of the top Rotliegend surface. Specifically, it is possible that a 'pull-up' effect exists where the top Rotliegend underlies a thick body of salt (i.e. beneath a salt diapir). This is important because it could alter the fault interpretations that form the basis for later consideration of structural inheritance and inversion within the Norwegian Central Trough. In such instances it is necessary to differentiate between the true and apparent basement geometries. The expected influence of salt can be quantified using simple calculations based on assumptions of the seismic velocity of the Zechstein Group relative to the seismic velocity of other basin-filling lithologies. A further, qualitative means to discriminate pull-up from basement uplift beneath salt highs is to compare sections passing through a salt high with adjacent parallel sections (passing just beyond the salt high). If the pattern of faulting outside the influence of the salt high was consistent with the pattern identified beneath the salt high, then basement faulting was considered an acceptable interpretation. If not, then pull-up was the preferred explanation, and a stylized interpretation was made through the problem area. For typical depths of 4 to 5 seconds TWTT to the base salt in this study, pull-up was identified as a significant problem only where salt thicknesses exceed 1,000 msec, and the effects are therefore confined to the base of only the most prominent salt structures.

While the throw of faults at top Rotliegend level is often clear, the orientation of the fault planes is less easy to discern from individual sections. Fault orientations were determined by 3D interpretation, with adjacent lines providing insight into along-strike variation in magnitude and azimuth of offset. The larger faults (those with offsets greater than approximately 200 msec) have been mapped and displayed on Figure 4.3. On sections in Figure 4.2, smaller-offset faults are also interpreted where identified. It was neither practical nor necessary to map these smaller faults in three dimensions; the exception being wherever salt-fault relationships appeared interesting or unusual. Trajectories of the smaller-scale faults were interpreted stylistically, such that they conform to the larger pattern of rifting. The trajectories of all 2D-mapped faults were reviewed on depth converted sections and their dips were confirmed as geologically plausible.

Within the Norwegian Central Trough, individual faults are commonly oriented NW/NNW-SE/SSE, while the bounding envelope of the basin is oriented approximately WNW/NW-ESE/SE (see Figure 2.19b). This bounding envelope is not defined by continuous fault segments, and is therefore unlike the Danish sector of the Central Graben to the immediate south of the studied area. Greatest throws at top Rotliegend level are observed on the NW/NNW-SE/SSE trending faults (e.g. the Skrubbe, Piggvar, and Mandal faults) whereas the more WNW/NW-ESE/SE trending faults typically have smaller throws. The WNW/NW-ESE/SE trending faults may be interpreted as transfer faults to a dominant NNW-SSE dip-slip trend. These observations make sense in the context of the Central Graben as a whole (Figure 2.19b).

4.3.3 Upper Permian (Zechstein Group)

The top of the Zechstein Group is constrained by nine of the 43 project wells. Figure 4.4 is a time horizon map for the top of the Zechstein Group. Relief on the top Zechstein horizon is considerable, varying from c.6500 msec to c.1500 msec. Most well penetrations are at salt highs, whilst borehole constraints on the deeper geometry of the Zechstein Group are poor. The strength of top Zechstein reflection is highly variable. Reasons for this variation are threefold: different lithologies in contact at the top Zechstein boundary (hence lateral variations in impedance contrast); different dips (steeply dipping surfaces cannot be accurately imaged using conventional seismic reflection techniques); and different depths (primary signal strength diminishes as depth increases). The top Zechstein reflection is typically poor to moderate away from salt highs and interpretation is challenging in these areas. However, three further considerations have facilitated the mapping of the top Zechstein reflection. Firstly, the Zechstein Group has a distinct seismic character due to the absence of coherent, concordant internal surfaces common to other basin-filling lithologies. The transparent or chaotic seismic facies character of the Zechstein Group is sufficient to constrain its irregular geometry in some places (e.g. Figure 4.2b, beneath well 1/6-3; Figure 4.2f, beneath well 1/9-2). Secondly, the Zechstein Group can be constrained from recognition of 'rim synclines' or similar seismic-stratigraphic patterns within sediments adjacent to a salt body. Rim synclines are local depressions that develop adjacent to salt diapirs, forming in response to salt withdrawal at depth. Identifying such features is a useful aid when mapping the steeply-dipping flanks of a salt diapir (e.g. the diapirs in Figures 4.2b, 4.2h and 4.2i). Thirdly, the geometry of the overburden to the Zechstein Group can often be directly related to the shape of the underlying salt; this can facilitate interpretation in areas of poor data quality.

At the present day, the Zechstein Group extends across most but not all of the Norwegian Central Trough study area (Figure 4.4). In addition to the nine wells that constrain the depth to the top Zechstein, there are four wells (2/7-19, 2/7-9, 2/9-2 and 2/9-3) that record an absence of Zechstein deposits despite penetrating into the stratigraphically lower Rotliegend Group. West of the Feda Graben (wells 2/7-19 and 2/7-9) absence of the Zechstein Group corresponds to high basement of the Grensen Nose. The Figure 4.5 Zechstein Group isochron map shows that salt thins southward and westward onto the basement high; this is also seen on Figures 4.2c and 4.2d. These 2D sections show a relatively thin post-Permian to pre-Cretaceous interval. The preferred interpretation here is one of non-deposition rather than salt withdrawal, dissolution or erosion.

Formation of the North and South Permian Basins in which Zechstein Group facies were deposited (Figure 2.22) is most often attributed to thermal subsidence (e.g. Erratt et al., 1999), but some workers favour interpretations involving contemporaneous extensional tectonics. For example, Glennie and Underhill (1998) used age constraints from Lower Rotliegend volcanics of the South Permian Basin to estimate basin subsidence rates of

~250m/Myr, a value they considered too high to be the product of thermal subsidence alone. Likewise, the non-deposition of halite in the Grensen Nose area despite significant deposition to the immediate east (on the other side of the Jurassic fault, in the present-day Feda Graben) suggests that Permian extensional faulting may have been important in the area, with Jurassic faulting having a mechanical connection to the Permian basin-forming process. Thus, it is postulated here that the present-day Skrubbe fault may have originated in the Permian as a North-South trending normal fault formed in response to East-West extension. It is possible that other, similarly N-S oriented faults existed. Their presence would allow for considerable variations in the depositional thickness of Zechstein deposits, with significant implications for later deformation.

East of the Feda Graben on the Mandal High there is evidence for the original presence of Zechstein facies; well 2/6-3 penetrates a 62m (TVD) interval of Zechstein deposits yet it is not possible to trace these deposits laterally on seismic data, suggesting loss through salt withdrawal, dissolution or erosion. On the adjacent Piggvar Terrace, wells 2/9-2 and 2/9-3 record a total absence of Zechstein facies. The seismic character of the post-Permian, pre-Cretaceous section in the vicinity of these wells is discordant and apparently deformed (Figure 4.2e). Consequently, halite removal rather than non-deposition is the preferred explanation for the absence of the Zechstein Group in this area.

The most pronounced top Zechstein highs on Figure 4.4 correspond to the crests of salt diapirs. Figure 4.5 is a Zechstein Group isochron map, which identifies the salt thicks (salt walls and diapirs) more clearly. Most of the Zechstein thicks are not radially symmetric (simple diapirs), but are elongate or otherwise irregular in shape. Their geometries likely reflect their complex and varied origins, having formed at the margins of, and in association with, developing mini-basins of variable shapes and sizes (see again Figure 2.12). The most prominent salt diapirs are those of Feda West and Delta in the West Feda Graben, and Mode and Trud in the Feda Graben (SE). However, salt structures are quite evenly distributed throughout the Norwegian Central Trough and there is no clear trend between the height of a salt structure and the depth to its base.

The top Rotliegend fault interpretation has been superimposed onto the Zechstein isochron map (Figure 4.5). Salt thicks exist at a variety of locations relative to the sub-salt fault network: deep within the basin (e.g. the Affleck and SE Tor diapirs in the West and NE Feda Graben, respectively); in the immediate hangingwall to basin-bounding faults (e.g. the Valhall and Eldfisk salt thicks at the Skrubbe fault on the west margin of the SE Feda Graben); or on elevated footwall blocks in shallower parts of the basin (e.g. the Ekofisk and Tor salt thicks). There is no straightforward, consistent link between sub-salt fault location and the siting of salt structures in the Norwegian Central Trough. However, the sections presented in Figure 4.2 suggest the latter scenario is most common; salt walls typically root to the footwall side of basin-bounding faults. This observation is discussed further in chapter five.

4.4 Mesozoic

4.4.1 Triassic (Hegre Group)

Several studies of the Central North Sea have focussed on the complex Triassic and Jurassic evolution of the basin (e.g. Goldsmith et al., 2003; Husmo et al., 2003; Rattey and Hayward, 1993; Fraser et al., 2003). Due to time constraints and in light of the specific objectives of this thesis it was decided not to undertake detailed three-dimensional mapping of sequences within the post-Permian, pre-Cretaceous section. However, it was desirable to confirm a Triassic age for the onset of halokinesis, as has been recorded in other parts of the Central North Sea (e.g. Hodgson et al., 1993). This was achieved through targeted local mapping of the top Triassic reflection.

Relative to the Jurassic, there is a high degree of uncertainty in constraining chronostratigraphy within the Triassic (limited biostratigraphic data due to the environment of deposition). However, a top Triassic pick has been recorded in 15 of the 43 project wells (Appendix 1). The pick is the top Tr50 stratigraphic marker of Goldsmith et al. (2003), and corresponds to the last appearance of a suite of Triassic palynological markers. In most of the wells where it is constrained, the Triassic is a thin sequence preserved over a basement high (e.g. well 1/3-5, figure 4.2a; well 2/9-3, figure 4.2e; and well 2/7-19R). An exception is the Triassic beneath well 2/5-7; a top Tr50 pick at 3760 msec highlights the presence of a thick Triassic package (Figure 4.2i). This Triassic thick is best explained as a minibasin that formed in response to the syn-depositional withdrawal of halite into a salt wall to the immediate NW. The salt wall subsequently collapsed (the appearance of the post-Triassic, pre-Late Cretaceous thick above the Zechstein suggests collapse occurred during the Late Jurassic and Early Cretaceous). Importantly, the presence of a Triassic minibasin confirms that major movements of the Zechstein salt had already occurred before the end of the Triassic; a phenomenon that has been previously assumed for the Norwegian Central Trough (Gowers et al., 1993) and observed more clearly in other parts of the Central North Sea (e.g. Hodgson et al., 1993).

In deeper parts of the basin the top Triassic reflection can be difficult to trace. This is presumed due to a lack of lithological contrast at the boundary between the Hegre and Humber Groups. Evidence for a pod-interpod style of Triassic deposition is less clear and in some areas the Triassic layers appear relatively uniform in thickness (e.g. within the Feda Graben SE, Figure 4.2i SE; and within the Feda Graben NE, Figure 4.2c). This suggests Triassic accommodation space in these areas was generated predominantly through extensional tectonics rather than salt movements, but the evidence to support this assessment remains unclear. Figure 4.6 highlights the extent to which Top Triassic (Tr50) TWTT interpretations were possible across the study area.

4.4.2 Jurassic (Humber Group)

The top Humber surface is constrained by 27 of the 43 project wells. It is characterised by a high amplitude reflection that is especially prominent in basin areas. The surface is often strongly onlapped by overlying reflections. In some areas it truncates underlying reflections and therefore forms an erosive (base Cretaceous) unconformity. The top Humber Group TWTT horizon map is displayed in Figure 4.7; the surface was mapped with ease on account of the good well control and strength of the reflection.

From the Figure 4.2 sections, it is evident that the Jurassic corresponds to a major rifting episode. The dominant characteristic of the upper part of the Triassic-Jurassic sequence is a thick asymmetric infill, and rotational fault block geometries can be clearly seen (e.g. Figure 4.2a NE; Figure 4.2g NW). The observation of major rift activity conforms to the model of Late Jurassic rift development in response to the collapse of a Mid Jurassic thermal dome (Underhill and Partington, 1993), whereupon the Norwegian Central Trough formed part of the southern arm of this trilete, failed rift (the Central Graben). Because the top Humber is in places an erosive unconformity, it was desirable to observe intra-Jurassic reflections in addition to the top Humber reflection. Regional correlation of Upper Jurassic strata is complicated due to the marked facies variability and pronounced diachroneities associated with major tectonic movements occurring at this time; also due to the presence of reflection multiples. A top J50 well pick (corresponding to a major Early Kimmeridgian flooding surface in the stratigraphic scheme of Rattey and Hayward, 1993) was picked as a gamma ray spike on well logs from 20 of the 43 project wells. These picks were tied to seismic, allowing a loose grid of interpretations to be made, and the general style of Late Jurassic sedimentation to be documented (Figure 4.2 sections).

Figure 4.8 shows an isochron map for the top Zechstein to top Humber (Triassic and Jurassic) section. This interval encompasses the major phase of extension in the Norwegian Central Trough. The map shows that the thickest sequences were deposited in the Feda Graben SE, where net extension was most localized, and rotational movement was most pronounced on the Piggvar (e.g. Figure 4.2d) and Skrubbe (e.g. Figure 4.2e) faults. Patterns in Triassic-Jurassic thickness are complicated by the presence of salt structures of varying geometry within this layer, including some that pierce it (pink blobs on Figure 4.8). Evidence that diapir growth during the Jurassic occurred through downbuilding comes from recognition of the considerable topography of some salt structures (the Feda West and Delta diapirs in particular) that apparently cannot be volumetrically justified by halite withdrawal from intervening areas.

4.4.3 Lower Cretaceous (Cromer Knoll Group)

The top Cromer Knoll reflection was constrained by 31 well penetrations; it is a moderate to strong positive amplitude reflection. At structural crests where the Lower Cretaceous thins beneath the limits of seismic resolution, the base Upper Cretaceous is seismically indistinguishable from the top Humber horizon (e.g. Figure 4.2a at well 1/3-5), but can nonetheless be accurately inferred (because the top Humber reflection has also been mapped). The top Cromer Knoll TWTT horizon is shown in Figure 4.9.

Seismic reflections within the Cromer Knoll Group are typically parallel, onlapping against the top Humber surface (e.g. Figure 4.2g; Figure 4.2i). The parallelism of onlaps suggests basement extension had effectively ceased by the Early Cretaceous; had the Lower Cretaceous accommodation space been formed by active basement faulting, the onlapping units would diverge towards the faulted margins. Further evidence that rifting had abated is that many of the faults that control base Cretaceous structural relief die out in Lower Cretaceous strata (e.g. Figure 4.2f, NW). The observation that Early Cretaceous sedimentation occurred in a post-tectonic setting is in agreement with the consensus view among previously published studies from the Central Graben (e.g. Sundsbø and Megson, 1993; Copestake et al., 2003).

Post-rift regional subsidence, post-rift passive infill of pre-existing bathymetry, and locally, subsidence related to salt movement (salt collapse structures) appear to have been important in the Norwegian Central Trough. The clearest evidence for regional subsidence rather than passive infill comes from recognition of deformation of the earliest Early Cretaceous onlaps in a manner harmonious with deformation of the top Humber surface (e.g. Figure 4.2c, SW). Regional subsidence can be differentiated from salt collapse by examining the underlying seismic for evidence of pronounced disruption to the layering within the Triassic and Jurassic sequence, which would be indicative of the latter process (e.g. Figure 4.2b NE; Figure 4.2i NW). Comparison of the Lower Cretaceous isopach with the Chalk Group or later isopachs also aids this differentiation.

Figure 4.10 shows an isochron map for the Lower Cretaceous interval. There are two major Lower Cretaceous thicks in the West Feda Graben; one in the vicinity of well 2/7-16 (Figure 4.2d, SW), and a second further west in UK waters to the south of well 1/5-2 (Figure 4.2c, SW). In both these examples, subsidence appears to have been the dominant process in creating accommodation. Elsewhere, and especially within the SE Feda Graben where Late Jurassic extensional rotation was most pronounced, passive infill appears the dominant mode of Early Cretaceous sedimentation. Recognition of the dominantly passive nature of Lower Cretaceous deposition is important because it validates the technique of seismic restoration by flattening the top Cromer Knoll. This flattening provides an insight into end Jurassic basin geometries and aids the analysis of salt body geometries prior to their modification during Cretaceous and Cenozoic times (see Chapter 5 restorations).

4.4.4 Upper Cretaceous (Chalk Group)

Picks from 43 wells were used to constrain the top Chalk horizon. As a high amplitude reflection, the top Chalk TWTT horizon was interpreted with confidence (Figure 4.11).

Chalks of the Upper Cretaceous blanket the existing topography and accommodation space appears controlled by post-rift thermal subsidence for that reason (Figure 4.2 sections).

Figure 4.12 displays an isochron map for the Chalk Group. Sedimentation is greatest in the Breiflabb Basin (Figure 4.2a) to the NW of the Norwegian Central Trough, and is at a minimum at the western margin of the SE Feda Graben (Figure 4.2d beneath well 2/7-1, and Figure 4.2e). The Chalk Group thin is associated with a top Chalk structural high and the pattern of onlaps is indicative of uplift occurring during Upper Cretaceous sedimentation. The cause of uplift was inversion on the Skrubbe fault that underlies the top Chalk high. Additionally, salt is present in the immediate hangingwall to the Skrubbe fault and may have influenced the structural style and amplitude of uplift that developed in response to Late Cretaceous compression. The interplay between halite and fault-driven uplift is examined more closely in the next chapter.

Several of the Norwegian Central Trough's pre-Cretaceous normal faults show a slight uplift of the early Cretaceous sequences in the hangingwall (e.g. Figure 4.2c NE; Figure 4.2f NW; and Figure 4.2g beneath well 1/6-1 and NW). In general, the larger the local normal fault movements were in the Late Jurassic/Early Cretaceous, then the larger the subsequent inversion. Certainly, the fault with the greatest pre-Cretaceous displacement (the Skrubbe fault) also shows the most intense inversion. Constraining precise timings of inversion requires reliable biostratigraphic dating, coupled with accurate seismic correlation of syn-inversion onlap. An intra-Chalk Group surface (the top Hod Formation) was interpreted to aid this attempt to constrain the age of syn-inversion post-rift deposition. This is discussed in more detail in Chapter 6.

4.5 Cenozoic

4.5.1 Upper Palaeocene (Rogaland Group)

A top Rogaland time horizon map (based on constraints from 43 wells) is displayed in Figure 4.13, and an isochron map for the Rogaland Group is shown in Figure 4.14. There is a broad regional increase in Rogaland Group thickness towards the NW, and the isochrons do not closely mirror patterns in the underlying Chalk Group. This suggests sedimentation was not controlled by the uplift of particular tectonic features, but by regional subsidence; tectonic inversion appears not to have occurred during the Upper Palaeocene.

4.5.2 Eocene to present (Hordaland and Nordaland Groups)

The top of the Hordaland Group is constrained by all 43 of the project wells. A top Hordaland time horizon map is displayed in Figure 4.15, and an isochron map for the Hordaland Group is shown in Figure 4.16. The top Hordaland horizon is an unconformity that corresponds to a regional rotation occurring in the Middle Miocene. There is some evidence of convergent reflection patterns and onlap against structural highs within the Hordaland interval and also within the lower part of the Nordaland Group (e.g. Figure 4.2e, near well 2/7-24). This is indicative of a second phase of compression-related structural uplift.

The Cenozoic evolution of the Norwegian Central Trough is dominated by thermal subsidence, with sedimentation basin-wide and relatively rapid. Evidence for the rapidity of sedimentation is that several of the salt diapirs that grew by passive downbuilding are outpaced by sedimentation within the cooling, subsiding basin. Additional evidence for rapid sedimentation comes from recognition of polygonal faults (Figure 4.17). Polygonal faults have been identified in other parts of the Central Graben (e.g. Cartwright, 1996) and have been attributed to fluid expulsion from mudrocks in response to significant early compaction (rapid burial).

4.6 Summary

A tectonic framework for the Norwegian Central Trough has been presented. This framework, encompassing Permian to Recent strata, has established the chronology of principal tectonic events affecting the basin. Halite of the Zechstein Group was mobilised during the Triassic under an approximately East-West extensional stress regime. The major phase of extension occurred during Late Jurassic times, with an especially thick sequence of Humber Group shales deposited in the SE Feda Graben. The Early Cretaceous is characterised by passive rift infill and sedimentation due to the onset of post-rift thermal subsidence that continues into the Cenozoic. This subsidence is punctuated by pulses of compression. Tectonic inversion can be recognised from seismic-stratigraphic patterns indicative of syn-sedimentary uplift in the Late Cretaceous and Early Cenozoic strata. These patterns are examined in more detail in the next chapter.

5 Compressional Structural Styles in the Norwegian Central Trough

5.1 Introduction

A regional structural framework was presented in Chapter 4. This chapter focuses on the detailed interpretation of structural styles and addresses the relative roles of fault movement and salt mobility in accommodating compression. Five of the seismic lines displayed in Figure 4.2 were structurally restored using 2DMove, and the results of these restorations are presented here. The relationship between tectonic inversion and halokinesis is further examined from additional, suitably selected seismic sections.

5.2 Criteria for identifying compression in the salt prone extensional basin

Criteria for identifying tectonic inversion, and halokinesis as opposed to inversion, were proposed in thesis Sections 2.2.1 and 2.3.4, respectively. Figure 5.1 restates these criteria using cartoon examples based on Figure 4.2 section observations. The figure summarises observations used to discriminate between the following compressional structural styles: salt-absent tectonic inversion; compressional salt movements unrelated to faulting; and salt-influenced tectonic inversion. The observations pertain to seismic stratigraphic patterns, and include the recognition of salt structures acting as fault detachment surfaces (or otherwise demonstrably affecting fault behaviour), the identification of uplifted syn-rift sequences, and the proximity of such sequence to salt structures.

The task of identifying Late Cretaceous and Cenozoic compression in the Norwegian Central Trough is complicated by the presence of Zechstein salt in most localities (Figure 4.5). This salt was mobilised earlier in the history of the basin, which complicated structural styles by increasing lateral heterogeneity of the basin fill, and decoupling the sub-salt and supra-salt sequences. In spite of this structural complexity, criteria uniquely indicative of compression were identified in several localities; the best examples are shown later in this chapter (see Figures 5.11 and 5.17). It is because of this early history of salt mobilisation that the compressional structures in the Norwegian Central Trough are so varied in character.

5. 3 Quantifying Shortening

Figures 5.2 to 5.6 show section restorations for five of the seismic lines previously displayed in Figure 4.2 (sections AA', CC', EE', GG' and II'). The initial GeoFrame interpretations have been depth converted, then sequentially restored for compaction, isostasy and faulting. The restoration work was undertaken with the aid of 2DMove software, according to the method outlined in Section 3.6. The goal of restoration was twofold: to quantify line-length changes since the Permian (i.e. to quantify the history of extension and shortening); and to assess the influence of differential compaction effects (as opposed to fault driven uplift or salt buoyancy) on the creation of geometry over salt highs.

The results of restoration are presented in a consistent manner for each of the five figures. In addition to a depth-converted present day section, there are reconstructions for the Middle Miocene (Top Hordaland), End Danian (Top Chalk), End Albian (Top Cromer Knoll), End Ryazanian (Top Humber) and End Permian (Top Zechstein). Restoration to Middle Miocene required unloading of and isostatic compensation for the Nordaland Group sediments; restoration to End Danian required the likewise removal of Nordaland, Hordaland, and Rogaland Group sediments; restoration to End Albian required the same, with additional removal of the Chalk Group. The restorations to earlier time steps typically incorporate more assumptions and are bound by greater uncertainties than the earlier time steps.

One issue is a lack of certainty regarding palaeobathymetries. In this study, palaeobathymetries are predicted from the model. Where accurately obtained from independent data (for example using biostratigraphy), palaeobathymetries can be used to calibrate a given restoration step and thus improve the interpretation of deeper geometries. This approach was not deemed appropriate for the Norwegian Central Trough restorations, because precise palaeobathymetries cannot be stated. An advantage of not specifying palaeobathymetries as a model input is that this enabled prediction of relief at end Danian times, based solely on the present day bathymetry and the parameters used for decompaction and isostatic adjustment. That there was substantial relief during deposition of the Chalk Group has been widely acknowledged (e.g. Bramwell et al., 1999, and discussion in thesis Sections 6.2.3. and 6.4), but there have been no previous attempts to quantify this relief, despite its importance as a control on possible sedimentary (slope) movements that may have been occurring at the time. (On Figure 5.4, the Danian crest of the Lindenes Ridge at Valhall is calculated to dip towards the SW by as much as 7°; a very large slope, sufficient for allochthonous processes to have been significant).

The restoration to end Albian has been flattened on top Cromer Knoll (i.e. a flat top Cromer Knoll palaeobathymetry is assumed). This was considered acceptable because, as discussed in the previous chapter, the pattern of Early Cretaceous onlap is indicative of essentially passive infill (although there are occasional, localised, suggestions to the contrary, such as the Early Cretaceous pattern of sedimentation on section EE', Figure 5.4d). It should be noted that the Lower Cretaceous grabens did not fully fill in areas to the north of the Norwegian Central Trough (Roberts, *pers. comm.*, 2009); flattening would be inappropriate in this case. In spite of these possible limitations, it is apparent that the half graben geometries are more realistic after this flattening than before it. Flattening does not alter the assessment of line-length changes, and is of minimal significance for assessment of differential compaction, since this phenomenon is more important for shallower sediments that drape salt diapirs.

Line length changes were calculated by restoring offset on faults. Within 2DMove, the same approach was used to restore movement on all faults. Offsets were restored using an inclined shear algorithm that reconnected hangingwall to footwall along a trajectory defined at 60° to the horizontal. This value was chosen because it is the typical dip of a normal fault, and most of the interpreted basement faults were normal faults mapped with dips within a few degrees of this value.

The top Rotliegend surface was restored for each of the five sections. In every case, there has been a net line lengthening (extension) between end Permian and present day. The detailed timing of movement cannot be defined due to a lack of stratigraphic control because of the decoupling of overburden by salt. β-factors for extension between end Permian and end Ryazanian were: 1.116 (AA'), 1.026 (CC'), 1.122 (EE'), 1.129 (GG'), 1.027 (II'); that is to say they vary from as little as 2.6% to as much as 12.9%. The measured extension is no greater on the SW-NE oriented sections (AA', CC', EE') than on the SE-NW sections GG' and II'; this is surprising given the dominant regional extension direction during Triassic and Jurassic times is known to have been approximately ENE-WSW (e.g. Underhill, 2003), suggesting extension would be greater on the SW-NE oriented sections. It may be explained by the Norwegian Central Trough's situation at a complex stress transfer zone, where oblique faults offset north-south oriented grabens (the Central Graben and Danish Central Graben); neither SW-NE sections nor NW-SE sections capture the full magnitude of extension.

The magnitude of extension across the Norwegian Central Trough has been calculated by other workers, using different techniques. Simple estimates have been made from crustal thickness data (assumes thickness is directly related to stretching factor), as well as more sophisticated estimates that incorporate subsidence analysis (e.g. White and Latin, 1993). The values calculated using these methods are greater than those computed from my 2DMove approach (restoration of seismically mapped faults). For example, White and Latin (1993) calculated a β factor as high as 2.0 for the Central North Sea triple junction area. They did this by calculating the amount of lithospheric thinning that would be required to generate a subsidence curve that best explains the known basin history, under the assumption of a 'normal' thermal regime and uniform stretching model. (The assumption of a normal thermal regime is supported by isotopic evidence from volcanics related to the Jurassic rifting).

There would be less of a mismatch between the subsidence analysis result and my own fault restoration methods if stretching was depth-dependent (rather than uniform), with more extension being accommodated in the deeper, ductile zone. However, the primary reason for the mismatch is presumed to be the under-sampling of faults (and therefore extension) in the restoration model. Under-sampling may be attributed to any or all of three factors. Firstly, the lines of section restored were not long enough to capture the full length of the rift zone. Secondly, low-angle faults may not have been clearly imaged on seismic reflection data; such faults would provide a major contribution to overall extension. Thirdly, potentially large numbers of smaller faults, with sub-seismic scale offsets, were disregarded (although potentially large in number, the total magnitude of offset on sub-seismic faults would be small relative to the offset on the fewer, seismically resolvable faults).

There are relatively few faults in the supra-salt sequence, and restoration of them is often complicated by the presence of underlying salt. There are examples of supra-salt fault restorations in Figures 5.4 and 5.5. In Figure 5.4, the End Ryazanian reconstruction has incorporated restoration of offset on two salt-involved faults; one in the centre of the section, and the other two-thirds towards the NE end of the section. An associated β (stretching) factor of 1.005 was calculated. However, the restoration of offset on these faults could be interpreted in an alternate manner; by invoking movement of the salt body (adjacent to the fault) during this time step. In this instance, there is insufficient seismic-stratigraphic evidence to distinguish between the two.

In Figure 5.5, the Lower Cretaceous thick at Albuskjell is interpreted as an inverted crestal collapse graben overlying the salt structure (see also Figure 5.17). The forward restoration step from End Ryazanian (Figure 5.5e) to End Albian (Figure 5.5d) illustrates the collapse of a salt structure and the creation of a localized graben over its crest, rather than a regional line-length change. A difficulty with interpretation of such salt-involved structures is that the model cannot itself predict the timing or amount of halite movement; this must be inferred from seismic stratigraphic relationships and incorporated into the model. For many cases where there are salt-involved faults, restoration appears best modelled through a combination of both salt and fault movements, and the solution can be considered in terms of end-member scenarios, rather than specified exactly.

The 2DMove restorations reinforce the observation that most salt movement occurred during the Triassic to end Jurassic interval, with only minor movements occurring after end Ryazanian times. These minor movements include collapse of salt highs (as discussed in the previous paragraph), continued growth of salt diapirs by passive downbuilding, and the squeezing of salt diapirs during compression (e.g. Figures 5.2 and 5.6). An original Zechstein thickness (that is to say, thickness of the Zechstein sequence prior to its post-depositional mobilisation) has been calculated for each of the AA', CC', EE', GG' and II' section restorations. These thickness values are to be treated as approximate (order of magnitude) values only, because the estimates require assumption of constancy of area (i.e. that there is no salt movement perpendicular to the plane of section), and of uniform initial thickness. Sites of inferred non-deposition (see Figure 4.5) were honoured when calculating original Zechstein thickness (e.g. Figures 5.4f and 5.5f). The range of recorded values, 225 to 410m, appears feasible in light of the limitations of the approach. Values compare acceptably with the initial thickness estimates of Cameron et al., 1992, (Section 2.5.1, p23), who anticipated thicknesses up to in excess of 1km in the deepest parts of the basin on the basis of palaeoenvironmental models.

The ability to quantify shortening by restoring contractional movement on supra-salt faults (and calculating the associated β factors) is hampered by the paucity of reverse faults with measurable (seismically-resolvable) offsets in the shallow section. It is likely that shortening was accommodated by folding as well as contractional faulting. For example, there will have been shortening associated with Late Cretaceous inversion of Early Cretaceous sediments (compare Figures 5.5d and 5.5c), that cannot be quantified through fault restoration. The amount of fold-related shortening may be as much as a few percent. However, the actual amount of shortening will depend on how much of this relief is attributed to differential compaction rather than shortening. This is discussed further in Section 5.4.

5.4 Quantifying Differential Compaction

Differential compaction is a mechanism for generation and enhancement of relief over salt bodies, due to the relative incompressibility of halite as opposed to other typical basin filling lithologies (e.g. Figure 2.9). It is necessary to quantify this phenomenon, in order to be satisfied that the present day relief on key marker horizons (the top Chalk surface in particular) is attributable to mechanisms of uplift (i.e. tectonic inversion and/or salt mobility) rather than solely due to mechanisms of differential subsidence. 2DMove software was used to do this.

The effects of compaction were modelled using porosity and depth coefficient values chosen for each stratigraphic interval on the basis of assumed lithology (as discussed in Section 3.6), and in agreement with Sclater and Christie (1980). In addition, decompaction restorations were undertaken using a wide range of feasible porosity values for each interval to test the sensitivities in the model. Of particular interest for sensitivity testing were the lines in Figures 5.2, 5.3 and 5.6, because these sections include prominent salt diapirs and differential compaction effects are likely to be greatest at these locations.

The five section restorations show that differential compaction accounts for some, but certainly not all, of the present day relief over the halite thicks. The pattern of onlaps within the Chalk and Hordaland Groups cannot be entirely attributed to differential compaction;

onlaps remain rotated onto the flanks of the diapir after decompaction, suggesting rotation is due to salt movement rather than solely due to differential compaction. Differential compaction effects are most significant where the height of the diapir (amplitude of salt structuration) is greatest.

On seismic data, there is discordancy in reflections over the crest of some diapirs (e.g. Figure 4.2f). The gradual process of doming due to differential compaction would yield a concordant pattern of reflections; discordancy is better explained by compression, because uplift and therefore sedimentary patterns are time variable. In most sedimentary basins, doming by compression creates highs over basin centres, whereas doming due to differential compaction would create relative highs over the less compressible basin margins (e.g. Gomez and Verges, 2005). The relationship is not so simple in the Norwegian Central Trough, due to the presence of laterally heterogeneous, incompressible salt within the basin fill. Doming due to differential compaction occurs not just at the basin margins, but wherever sediment has accumulated over significant local thicknesses of Zechstein halite.

5.5 Compressional Styles in Salt-Free Areas

The Norwegian Central Trough is situated at the southern margin of the North Permian Basin (Figure 2.22). This marginal setting was chosen with the intent to document structural styles in areas both affected and unaffected by salt tectonics. Seismic interpretation suggests that there are few unequivocally salt-absent areas within the Norwegian Central Trough: those areas where salt is absent are outside the principal Mesozoic grabens, and structurally quiescent. In consequence, there are no compressional structures entirely uninfluenced by the presence of halite. (On any given cross-section through the Norwegian Central Trough there may be salt-absent 'harpoon-like' asymmetric inversion geometries, but these are seen in close association with salt-present symmetric inversion geometries, along strike of the same fault. For example, compare the Piggvar fault in Figures 5.11 and 5.12. These contrasting structural styles are discussed in Section 5.6.2, and are regarded as variable structural styles within salt-prone areas, rather than truly salt-absent structural styles).

The lack of compressional structures uninfluenced by the presence of halite is not evidence for absence of compression, but evidence for the predominance of salt and salt tectonic behaviour throughout the basin. The salt-absent inversion schematic in Figure 5.1 is in fact based on a more complicated structure (Figure 4.2e, NE) where Chalk Group geometries have also been affected by salt withdrawal movements in the deeper basin.

Compression can still be interpreted, albeit in areas where salt is present. It is often not straightforward to discriminate compressional structural styles due to tectonic inversion in areas affected by halokinesis from those styles due to halokinesis alone. The value of the Norwegian Central Trough dataset is that it allows the interpretation and comparison of a range of salt-influenced structures.

5.6 Compressional Styles in Salt-Prone Areas

Evidence of uplift due to compression was recognised at relatively few locations within the Norwegian Central Trough, but these locations include areas within the four principal basins; the Feda Graben (NE, SE and West) and the Breiflabb Basin. This evidence is discussed with reference to seismic interpretations, shown in Figures 5.8 to 5.15, and 5.17; the locations of these lines summarised on Figure 5.7. As in the chapter 4 examples, the seismic data presented here has a vertical scale in TWTT that approximates to a two times vertical exaggeration. This vertical exaggeration is useful because it highlights those subtle onlaps in the seismic that hold the key to deciphering the structural history of the basin.

Uninterpreted sections have typically not been included in this chapter, to allow space for enlarged display of interpreted sections. For a visual reminder as to the quality of the data that underlies the seismic interpretations the reader is encouraged to consult the uninterpreted sections in Chapter 4. Uninterpreted enlargements of key section details have been included in Figures 5.12 to 5.15., in order to illustrate the stratigraphic detail on which interpretation is based. There is some uncertainty regarding the interpretation of any given fault or salt structure, but the uncertainty range is usually constrained within an acceptably narrow window relative to the end member possibilities. The greatest uncertainty is not the present day geometry, but the 3D structural history associated with that geometry.

5.6.1 Feda Graben NE

Figure 5.8 shows four SW to NE oriented sections through a domed sequence of Chalk Group sediments in the NE Feda Graben. The Chalk Group high is the trap for the Tor oilfield (e.g. well 2/5-1 on section 2C2C'). On each of the four sections, a fault is interpreted to offset sediments in the supra-salt sequence, and a salt body is interpreted beneath the domed Chalk Group. The amount of salt identified on the sections varies on the different lines, with the amount of salt increasing westwards from 2A2A' to 2D2D'. The magnitude of reverse throw on the supra-salt fault, and its connectivity to sub-salt faulting, decreases westwards from 2A2A' to 2D2D'. The amount of doming of Chalk Group sediments increases westwards from 2A2A' to 2D2D'. On each section, onlaps onto the top Hod Formation are identified suggesting that doming was occurring during the later phase of Chalk Group deposition. Cartoon reconstructions for the End Albian have been created to illustrate the inferred contrasting appearances of sections 2A2A' and 2D2D' prior to Late Cretaceous uplift. A second phase of onlap is identified on figure 2D2D'; there are marked onlaps within the top Rogaland to top Hordaland interval, suggesting discrete uplift during that time. Its discrete nature is indicative of uplift rather than differential compaction as a mechanism for the doming. On the other sections, the pattern of Cenozoic sedimentation appears more uniform; there is no strong evidence for uplift during this time. One reason may be that the Cenozoic phase of uplift was relatively subtle and so only identified on structures that were already the most pronounced. The most significant Cenozoic event is a regional tilting dated as top Hordaland by onlaps onto this surface. From the observations described above, the timings of uplift in the North East Feda Graben are interpreted as Late Cretaceous and Eocene-Oligocene pulses of compression.

5.6.2 Feda Graben SE

Figures 5.9 to 5.15 show a series of seismic sections across the SE Feda Graben. Comparison of these figures illustrates the manner in which salt has influenced the structures produced by inversion. Of particular interest are the fault and salt relationships at the Skrubbe and Piggvar faults (labelled) that define the margins of the basin in this area.

Figure 5.9 has been annotated with some of the key observations relating to the salt mobility and inversion history of the line. There is evidence of Early Cretaceous salt withdrawal; the Zechstein Group had been mobilised into salt structures prior to Chalk Group deposition. Onlaps within the Late Cretaceous section, and latterly within the post-Palaeocene to mid-Miocene section, are evidence of two main compressive pulses in the basin.

Figures 5.10 and 5.11 offer interesting comparison of the variation in Chalk Group geometry at the Lindesnes Ridge in relation to differences in salt presence in the immediate hangingwall to the Skrubbe fault. The subsidiary caption on Figure 5.10 is a cartoon illustrating the postulated appearance of the Lindesnes Ridge at 3B3B', in the case of salt absence from the hangingwall to the Skrubbe fault (in which case, geometries more closely resemble those mapped on Figure 5.11).

Figure 5.12 is another section across the SE Feda Graben. The presence of the large 'Mode' diapir on this line explains the variability of salt presence in the hangingwall to the Skrubbe fault on the other lines: salt movements have occurred outside the plane of these serial sections, including (for some, such as Figure 5.11) withdrawal from the immediate hangingwall of the Skrubbe fault into the Mode diapir. The uninterpreted enlarged seismic included on this figure was included because of the well resolved pattern of crestal faulting that can be interpreted over the Piggvar salt dome. These faults are most prevalent within the Lower Hordaland Formation; the same sediments that were identified to be offset by polygonal faults in Figure 4.16. Rather than forming a polygonal pattern of faults over the Chalk Group high, the sediments have developed a fault network that, from its radiating pattern, is interpreted as a product of the existence of that high. The fracture networks that developed over this salt dome differ from the fracture patterns over the 'harpoon-like' asymmetric inversion structure on Figure 5.11.

Figures 5.13, 5.14 and 5.15 are further examples showing salt and fault variability across and along the SE Feda Graben. As with Figure 5.9 to 5.12, these interpretations demonstrate that salt influences the structures that are formed through tectonic inversion. The geometry of Chalk Group doming varies along the strike of the fault. Figure 5.16 summarises the along-strike variability identified at the Skrubbe fault in Figures 5.9 to 5.15. The top Rotliegend fault throw, the Zechstein salt thickness, and the top Chalk elevation have been recorded on, against and over the Skrubbe fault, respectively. The figure demonstrates how variability in fault displacement and salt-involvement at the Skrubbe fault influences top Chalk Group relief. This relationship is not simple, but in general, the elevation at top Chalk level is greatest where fault throw and/or salt thickness is greatest. The error bars presented on this figure are a qualitative attempt to capture the uncertainty in the posted values, based on confidence with which the distances can be resolved and measured from the seismic data. The thickness of Zechstein salt in the hangingwall is the least precise measurement, because the top Zechstein pick can be difficult to constrain with as much confidence as the top Rotliegend and (especially) top Chalk surfaces.

5.6.3 Breiflabb Basin

Serial sections through part of the Breiflabb Basin are shown in Figure 5.17. These sections correspond to the Albuskjell oilfield. Observations from these three sections are considered best explained in terms of an inverted crestal collapse graben, as illustrated by the accompanying cartoon sketches of End Albian geometries. The sections show that inversion geometry changes as the shape and amount of salt changes along strike.

5.6.4 West Feda Graben

The West Feda Graben is an area dominated by salt structures, many of which pierce the top Chalk Group (e.g. Figure 4.11). Salt structures here include the Feda West, Delta, Tommeliten Gamma and Tommeliten Alpha diapirs, and display some or all of the characteristics of squeezed diapirs that were summarised in Figures 5.1 and 2.18. There is no clear evidence of reverse faulting or other contractional fault movements in this region.

5.7 Regional Synthesis

Based on the observations presented in Chapters 4 and 5, it is possible to summarise the timings, spatial extent and magnitude of compression in the Norwegian Central Trough. Two main phases of compression are identified: Late Cretaceous and Cenozoic. Late Cretaceous inversion is most pronounced during deposition of the Upper Chalk interval (the top Hod to top Chalk section typically onlaps onto local highs, indicative of Maastrichtian - Danian (c.70-63 Ma) uplift; e.g. Figures 5.8 and 5.15). Cenozoic compression has been identified through recognition of onlaps within the Early Hordaland to Earliest Nordaland sequence (Eocene to Middle Miocene). Most commonly, onlaps are within the Middle Hordaland, indicative of a Late Eocene - Early Oligocene (c.45-30 Ma) peak to the Cenozoic phase of inversion (e.g. Figures 5.8d and 5.9).

Locations within the Norwegian Central Trough where inversion has been observed are summarised in Figure 5.18. Not every fault inverts. The majority of inverted faults are oriented WNW-ESE to NNW-SSE, suggesting there was a preferential orientation for inversion. On initial inspection, this appears to infer a NE-SW orientation of compression (perpendicular to the trend of the inversion axes). However, it should be recalled from Section 2.2.3 that previous studies (e.g. Bruhn and Nalpas, 1996) have demonstrated that tectonic inversion features may develop even when the compression direction is strongly oblique to the original extension direction. The approximately NW-SE fault orientation is inherited from the extensional regime and tells us little about the compression direction(s). Furthermore, the presence of halite within a basin adds to the difficulty in resolving orientations of compression relative to those localities where salt-absent inversion has occurred, because of the additional complications associated with salt movement. The Ekofisk uplift is an interesting example, where the primary axis of inversion is North-South (parallel to the salt wall), but the structure is underlain by an East-West basement fault, which imparts a second order control on the inversion geometry.

The magnitude of compression in the Norwegian Central Trough is difficult to quantify; primarily due to the role of salt in accommodating regional stresses through movement. It was demonstrated that present day relief in the post-salt sequence cannot be attributed to differential compaction alone; shortening is required (Section 5.4). The amount of shortening was shown to be variable and less than the 3 to 13% post-Permian extension calculated from 2D restoration of variably oriented (NW-SE and NE-SW) transects (Section 5.3).

An alternate measure of shortening comes from the amount of uplift. Figure 5.16 charted Chalk Group uplifts in the immediate hangingwall to the Skrubbe fault of as much as 400msec (equivalent to approximately 400m). In salt free basins, such hangingwall uplifts can provide evidence of shortening. In the Norwegian Central Trough, uplifts can only provide evidence of shortening where not associated with an underlying salt diapir. This is because uplift may have resulted independently, from non-compressional salt movements.

Salt movements occurred throughout the evolution of the basin. This study has identified that most movement occurred prior to the Late Cretaceous, with comparatively minor halokinesis thereafter. Evidence for halite movements during Late Cretaceous and Cenozoic times was nonetheless documented in several examples (most notably Figure 4.2h). These observations render it possible to falsify the hypothesis that all salt moved prior to the Late Cretaceous. It is postulated herein that salt mobilisation occurred soon after deposition because of its inherent instability under loading. As loading continues, lateral variations in load stress become less significant relative to the total load. The potential for additional salt movement is limited in places where upwelling has already occurred and salt welds have formed, restricting the potential supply for further diapir growth. Where Late Cretaceous and Cenozoic salt movements have occurred, halokinesis appears to be associated with the compression of existing salt structures, causing renewed growth and uplift of the overburden.

In the previous chapter, a map showing the Zechstein isochron with top Rotliegend faulting superimposed (Figure 4.5) was briefly discussed. The sections presented in Chapter 5 reinforce the assessment that there is no straightforward, consistent link between sub-salt fault location and the siting of salt structures in the Norwegian Central Trough. Salt thicks may exist deep within the basin (e.g. the Affleck and SE Tor diapirs in the West and NE Feda Graben, respectively); in the immediate hangingwall to basin-bounding faults (e.g. the Valhall and Eldfisk salt thicks at the Skrubbe fault on the west margin of the SE Feda Graben); or on elevated footwall blocks in shallower parts of the basin (e.g. the Ekofisk and Tor salt thicks). The extensional history post-dating salt deposition appears to have been a more critical control on halite mobilization than the basin characteristics prior to salt deposition (see also Figure 8.3). This assessment is supported by observations from other salt basins, most notably the Gulf of Mexico (Beeley, *pers. comm.*. 2009), where min-basins and salt walls have formed in the absence of any basement faulting.

While there are no recent publications regarding the Norwegian Central Trough itself, observations made in this study can be compared with those from the Danish sector of the

Central North Sea, where interpretations based on recent seismic data have been published (e.g. Vejbaek and Andersen, 2002; Esmerode et al., 2008). Unequivocal salt-absent inversion has been documented at the Coffee Soil fault; a major basin-bounding fault that is linked by relay to the Mandal fault studied in this thesis. The northern part of the Danish Central Graben is relatively, and in places completely, devoid of Zechstein halite which allowed structures in this region to be mapped more confidently than has been possible in the Norwegian Central Trough (styles are similar to those shown in the Southern North Sea example in Figure 2.2).

There has been some discussion of tectonic inversion in salt-prone basins in a recent paper by Stewart (2007), which reviewed attempts to interpret salt structural styles in terms of the underlying mechanisms. This paper is based on interpretations from the UK North Sea, and therefore does not include data from the Norwegian Central Trough. As in this study, Stewart (2007) demonstrated that Cretaceous and Cenozoic shortening affected regions of the North Sea basin with pre-existing salt structures, and this compression resulted in a variety of reactivation styles that do not fit with currently used definitions of inversion tectonics (see Section 2.2).

5.8 Summary

The Norwegian Central Trough is an area of net extension. This net extension has been quantified on five regional transects. Post-Palaeozoic uplifts are locally recognised, and have been attributed to Late Cretaceous and Cenozoic shortening. Differential compaction and salt buoyancy are alternative mechanisms for generation of relief in the post-rift sequence, but unlike compression, neither process can explain all observations. The magnitude of Post-Palaeozoic shortening is considerably less than the magnitude of prior extension, but has nonetheless been an important mechanism for the generation of relief (i.e. trapping geometries) within the basin.

The presence and orientation of basement structures affects the compressional structural style. Salt movements are interpreted to have occurred in association with Late Cretaceous and Cenozoic shortening, and have also affected compressional structural styles. Of particular significance is the salt geometry prior to inversion, and the amount of salt involved in the structure during deformation. Where there is no salt in the hangingwall of an inverted fault, the structural styles are more asymmetric, with uplift localized over the fault. Where salt is present, the style of uplift becomes complex and variable. Inversion geometries can change along strike and at different stratigraphic levels in response to these changes in salt-involvement both along strike and through time.

6 Tectonic Inversion, Halokinesis and Reservoir Quality

6.1 Introduction

The Norwegian Central Trough is an important petroleum province and major hydrocarbon discoveries have been made within the Chalk Group. Although fine grained, chalk can be high in porosity (while low in original permeability) and in which case it may act as a very effective petroleum reservoir if adequately fractured. This chapter addresses the role that tectonic inversion and halokinesis may have exerted on Chalk Group reservoir quality.

6.2 The Chalk Group

The Chalk Group consists predominantly of micrite in the form of coccolith fragments. Coccoliths are the calcareous exoskeletons of pelagic unicellular algae (coccolithophorids). In addition to coccoliths, chalks have subordinate carbonate contributions from foraminifera (pelagic and benthic), and coarser skeletal material (e.g. shell fragments). Chalks typically contain a siliceous biogenic fraction, derived from the remnants of organisms such as radiolarian tests and sponge spicules. Chalks also contain small amounts of both silica and clay of terrigeneous origin (e.g. Tucker, 1991).

Chalk Group sedimentation occurred over a large area in Western Europe during Late Cretaceous and Early Palaeocene (Danian) times (Figure 6.1). Pelagic deposition occurred through the settling of coccoliths and coccoliths aggregated into faecal pellets. Sedimentation rates are estimated to have been on the order of 150-250 mm/kyr (Tucker, 1991). The rate of coccolith sedimentation was primarily controlled by coccolithophorid productivity and is typically low relative to sediment input into a basin from other (i.e. terrigeneous) sources. High eustatic sea level stands of the Late Cretaceous caused flooding of continental areas and resulted in reduced supply of terrestrially-derived clastics; these conditions were favourable for coccolith sedimentation and gave rise to the Chalk Group.

6.2.1 Chalk Group Stratigraphy

The first formal lithostratigraphic nomenclature for the Central North Sea was published by Deegan and Scull (1977). The Cretaceous and Cenozoic nomenclature for the Norwegian sector was later revised by Isaksen and Tonstad (1989). This study adheres to the terminology of the latter and thus the Chalk Group is divided into five lithostratigraphic intervals; the Hidra, Blodøks (Plenus Marl), Hod, Tor and Ekofisk Formations. Identification of the lithostratigraphic subdivisions can be difficult and usually requires a combination of techniques. Figure 6.2 summarises the seismic, gamma ray log (GR), sonic log (DT), biostratigraphic and lithostratigraphic criteria for distinguishing between formations of the Chalk Group.

On seismic data, the base of the Chalk Group usually appears as a moderate to strong amplitude positive polarity reflection, corresponding to the contact between argillaceous chalks, calcarenties and (minor) siliciclastics that are underlain by shales of the Cromer Knoll Group. The top of the Chalk Group (strong negative reflection) is also readily identifiable, but reflections from intermediate formation tops are not. For this reason, the top Hidra, top Blodøks, and top Tor Formations have not been seismically mapped as 3D surfaces. Seismic interpretation was aided by access to GR and DT data for the 43 project wells (Figure 3.1). The top Hod Formation is an identifiable pick on well logs (Appendix 1) and was mapped as a TWTT horizon on seismic sections in Chapter 5.

The Hidra Formation is seldom cored, but in most of the cases where it has been examined, it is interpreted as a pelagic deposit, with occasional evidence of turbidity currents in the form of thin calcarenties (Kennedy, 1987). The Blodøks Formation comprises black, pyritic shales with high organic carbon content; a local manifestation of the Cenomanian-Turonian black shale horizon known across Europe. The shale-rich Blodøks Formation is usually recognised as a positive spike in the gamma ray log.

The top of the Hod Formation is recognised as an increase in the gamma and sonic responses relative to the overlying Tor Formation. The top Hod is also defined by biostratigraphy, and can be most confidently picked from core. The Ekofisk and Tor Formations are the most commercially important reservoir formations and are the most drilled and highest in porosity (lowest GR and low DT). These chalks, deposited in Maastrichtian and Early Palaeocene times, are relatively pure. Recognition of the boundary between the Tor and Ekofisk Formations can be difficult due to the low contrast in physical properties between them. Transition from the Tor to the Ekofisk Formation typically involves a gradual (but well defined) increase in GR, but generally its identification requires additional interpretations. The contact is usually most easily recognized from biostratigraphy because of the distinct fauna either side of K-T boundary.

6.2.2 Chalk Group Lithofacies

A range of chalk lithofacies types can be identified from the study of chalk core sections and from chalk outcrop observations. Chalk lithofacies were examined during a reconnaissance study of several hundred metres of Norwegian Central Trough Chalk Group core at the ResLabs core facility in Stavanger, Norway, in October 2006. The purpose was to identify the range of depositional environments in which chalks of the Norwegian Central Trough were deposited. Criteria commonly used for the determination of siliciclastic lithofacies (e.g. grain size, clast composition and relative abundance) are often difficult to distinguish in chalks; a Chalk Group lithofacies classification scheme is proposed here on the basis of those sedimentary features which are readily discernible in hand specimen. It was deemed possible to recognise seven key lithofacies types; homogeneous, argillaceous, laminated, burrowed, hardground, conglomeratic and deformed chalks, and combinations of the aforementioned.

Observation of different Chalk Group lithofacies is evidence of a variety of depositional environments and early burial conditions. Figure 6.3 is a schematic illustration of the lithofacies types identified, and the environmental conditions that may have created them. However, a particular lithofacies may not be evidence of a specific depositional environment. For example, homogeneous chalks may form: through progressive accumulation of pelagic material of uniform composition and steady terrigeneous input; due to intense bioturbation within an initially heterogeneous sequence; due to dewatering processes removing the initial sedimentary structure; or as a consequence of sorting processes during resedimentation as a debris flow or turbidite. It has long been clear that not all of the chalks in the Norwegian Central Trough were deposited in quiet platform-type environments as thick but slowly accumulating pelagic sequences. The deformed chalk lithofacies is indicative of allochthonous resedimentation; mass movement through (submarine) slides, slumps, debris flows or turbidites.

6.2.3 Allochthonous Chalks

Although it is a pelagic sediment, fine-grained calcareous ooze may be mobilized prior to lithification and redeposited as allochthonous units (Bromley and Ekdale, 1987). Previous studies have identified allochthonous chalks within the Norwegian Central Trough (e.g. Watts et al., 1980; D'Heur, 1984; Brewster and Dangerfield, 1984). Using core data, allochthonous chalks can sometimes be recognized by direct observation of deformed chalk lithofacies in core. Dipmeter logs can also facilitate their recognition because they may appear as packets of bedded chalk with steeper dips than strata above and below. Redeposited chalks may be identified on wireline logs. Characteristics would include very low gamma ray response, high neutron porosity, low formation density and high sonic velocity (Rider, 2000). Allochthonous chalks may also be recognized from biostratigraphy. For example, Watts et al. (1980) studied core from the Albuskjell field (well 1-6/3) and identified repeated units within the Danian. Repeated units are most readily explained as mass-flow deposits of older material redeposited elsewhere at a later date.

Allochthonous chalks have been classified as slides, slumps, debris flows and turbidites, and their appearance and dimensions vary accordingly. Slides and slumps comprise packages of internally deformed chalk, up to ~30m in thickness; internal deformation may be plastic or semi-brittle in character (Sikora et al, 1999). Debris flows are typically less than 3m thick. Poorly sorted, they comprise small pebbles (lithified chalk fragments) supported by the micritic chalk matrix. Turbidites comprise layers of generally fine grained, fining upwards sequences.

There has been some considerable debate as to the causes of chalk resedimentation. Candidate explanations include slope instability due to sea level changes; storm wave activity; marine channel processes; or slope instability due to active tectonics. In one instance, meteorite impact has been postulated as the cause (Rider and Kroon, 2003). To isolate the role of marine as opposed to tectonic processes, it is necessary to constrain the depths below sea level at which the chalks were deposited. Benthic foraminifera are commonly used to derive palaeobathymetric information, but this is of limited use in the Norwegian Central Trough because the Late Cretaceous to Danian Chalk basin is a depositional system with no close modern analogue. Most of the benthic foraminifera within the Chalk Group have been assigned palaeobathymetric limits based on their occurrences in terrigeneous passive margin sequences. According to Sikora et al. (1999), who examined chalk core from the Valhall field, most benthic foraminifera are deep water forms, indicative of outer-shelf to upper-slope water depths of ~200-500m (e.g. *Stensioina beccariiformis*) or deeper (e.g. *Nuttallides truempyi*). Ichnofacies assemblages are little help in constraining palaeoenvironments and therefore palaeobathymetries of chalk deposition, because they are difficult to identify. The most commonly identified trace fossils, *Chondrites*, *Zoophycos* and *Planolites* (burrows of deposit-feeding worm-like organisms) support the notion of deep water deposition, since they are seldom recognised in shelf-sea chalk sequences (Ekdale and Bromley, 1984). *Thalassinoides*, the dwelling burrow of a crustacean, is widespread in shelf-sea chalks, but conspicuously absent from cored chalks of the Norwegian Central Trough (Tucker, 1991). This would appear to confirm their deeper-water nature. Whilst rare, hardgrounds have been identified within chalks of the Norwegian Central Trough. Their carbon isotope ratios are only slightly less positive than those of homogeneous chalks (Bramwell et al., 1999) and are therefore presumed to have formed in sub-aqueous marine environments, rather than as subaerial exposures, or surfaces that formed at close to wave base. To summarise, these various lines of evidence suggest that chalks of the Norwegian Central Trough were mostly, if not always, deposited at depths below the storm wave base and were therefore unaffected by shallow marine processes.

Chalk outcrops in the Etretat area of Haute Normandie, France, were examined during a reconnaissance visit in April 2005. These chalks exhibit a variety of sedimentary structures, including slumps and slides. Figure 6.4 shows examples of the main sedimentary structures identified in outcrop. Figure 6.4a shows an example of the large-scale concave-up geometries, up to 60m deep and 1km wide, that were first described by Kennedy and Juignet (1974) and explained as part of a carbonate bank complex. Subsequent workers have favoured an erosional explanation; that the mounds are in fact residual highs adjacent to erosive tidal channels (e.g. Quine and Bosence, 1991). Some of the sedimentary structures incorporate brittle elements (e.g. Figure 6.4b), while others appear more plastic in nature (e.g. Figures 6.4c and d). Plastic deformation may well have been an important process in the uppermost metre or metres of the sediment column, where freshly deposited chalk was transitioning from fluid suspension (ooze) to a solid-phase through compaction and dewatering. Late Cretaceous bathymetries at Etretat are inferred, in general, to have been shallower than those within the Norwegian Central Trough; marine processes were likely more prevalent where bathymetries were shallower. Nonetheless, the observed allochthonous sediments at Etretat, and examples 6.4b, 6.4c and 6.4d in particular, are considered representative of the allochthonous deposits that are known to exist, but cannot be sampled at such large scale, within the Norwegian Central Trough.

It is interesting and important to note that allochthonous chalks are more common in the Norwegian Central Trough than in most Chalk outcrop localities. The interpretation herein is that the role of structural feedback in the form of active faulting and/or halokinesis was critical in their genesis. There is one previously documented example of allochthonous chalks being attributed to active tectonics: Gale (1980) identified slump folds at the Down End quarry near Portsmouth, Hampshire, and attributed their origin to downslope movement of chalks that had been deposited on the slopes of the Portsdown anticline. A tectonic trigger was the preferred hypothesis because the Portsdown anticline had been identified as a tectonically active axis at the time of deposition.

From the Norwegian Central Trough Chalk Group isochron map (Figure 4.11) it is clear that there was a very marked variation in the depositional thickness of the Chalk Group. With the exception of an early effort by Watts (1980), there has been no attempt to relate the timing and relative abundance of the Norwegian Central Trough allochthonous chalks to the events that caused them. A goal of this research was to examine the spatio-temporal distribution of allochthonous chalks in relation to the interpreted pattern of tectonic inversion and salt mobility. Our understanding of allochthonous chalk in the Norwegian Central Trough utilises data derived from the production history of the basin.

6.2.4 Chalk Group Oilfields

The Chalk Group is reservoir for a major complex of hydrocarbon accumulations in the Norwegian sector of the Central North Sea, as well as smaller accumulations in the Danish and UK sectors. Figure 6.5 names and shows the approximate locations of the principal Central North Sea Chalk Group fields. For ease of reference, these fields have also been labelled on seismic sections presented previously, in Chapters 4 and 5. Many of the chalk fields have had a long production history (the very first hydrocarbon discovery in the North Sea occurred in 1966 in the Chalk Group) and remain active today. The long production history is advantageous in that it has led to a sizeable quantity of subsurface data being collected throughout the Norwegian Central Trough. Hundreds of wells have been drilled and these wells provide valuable core and wireline information, some of which was utilised in this study (from the Eldfisk, Ekofisk, Tor and Albuskjell fields).

6.3 Controls on Chalk Group Reservoir Quality

Chalk can act as both a seal and a reservoir rock. Whether it is sealing or a reservoir depends on the porosity and permeability characteristics. These are controlled by the lithofacies type (primary sedimentology), by diagenesis, and by internal fracturing. These factors are relevant to this study because of the potential for salt mobility and/or tectonic inversion to directly or indirectly influence them, and are discussed below.

6.3.1 Primary Sedimentology

Oakman and Partington (1998) observe that the initial porosity of redeposited chalks (up to ~80%) tends to exceed that of pelagic ooze (~70%), and the average initial permeability of redeposited chalks (100mD to 1D) tends to exceed that of autochthonous chalks (10-100mD). In addition, allochthonous chalks are widely considered to retain their relatively high porosities at depth (e.g. Watts et al., 1980; Kennedy, 1987; Taylor and Lapré, 1987; Brasher and Vagle, 1996). The increased porosity of allochthonous chalks has been attributed to the presence of a more rigid support framework that resists early compaction, or to the destruction of early diagenetic cements (Bramwell et al., 1999). In contrast, Maliva and Dickson (1992) studied chalk from the Eldfisk field and suggested that mode of deposition has little direct influence on porosity. They found that the non-carbonate content is a significant influence on porosity. Higher porosity chalks comprise smaller percentages of non-carbonate material (clay and silica) and only a few percent of this material can destroy reservoir potential. Allochthonous reworking is a conceivable mechanism by which to disperse the non-carbonate fraction within chalk. Without dispersal, the clay component may destroy reservoir potential due to the accumulation and preservation of seasonal varves, essentially impermeable to fluids.

6.3.2 Diagenesis

Mineralogically, coccoliths were formed and deposited as low magnesium calcite, the most stable polymorph of $CaCO_3$ at near-surface temperature, pressure and water chemistry conditions (Tucker, 1991). This gives chalk an unusually pronounced chemical stability for a carbonate rock; polymorphic transformation does not occur, and leaching and reprecipitation processes are impaired until pressure solution commences at greater depths of burial. Instead of chemically-controlled porosity loss, initial porosity reduction (from ooze values of ~70% to soft chalk values of ~40%) is attributed to material consolidation by mechanical compaction. Oakman and Partington (1998) suggest that chalk sediments must reach a depth of at least 600-900m, before overburden pressure becomes high enough for pervasive pressure solution and associated cementation to occur. This depth has been constrained using oxygen isotope data; the $\delta^{18}O$ value of a carbonate cement corresponds to the temperature at which that cement was precipitated, and this can be equated to a depth on the basis of assumed isotherms. This technique has confirmed that cementation is an important reason for porosity loss but that cementation occurred relatively late and at depth (Maliva and Dickson, 1992).

Figure 6.6 is an idealised summary diagram for the porosity-depth history of chalk, based on the comparative study of dozens of North Sea wells (Brasher and Vagle, 1996). It identifies five principal controls on porosity: burial depth, overpressuring, presence of hydrocarbons, chalk lithofacies type, and original grain size. The effect of high fluid pressures (overpressuring) is to support a proportion of the lithostatic load and thereby reduce the effective stress at grain contacts. Evidence for overpressure includes the fact that subsidence on the scale of metres has affected oil platforms sited over the crest of the Ekofisk and Valhall fields (e.g. Zoback and Zinke, 2002), because the onset of oil production resulted in pore pressure reductions. The effect of early hydrocarbon migration into chalk pores is to inhibit the dissolution and precipitation of carbonate. Mallon and Swarbrick (2008) recognise that deeply buried, non-reservoir chalks of the Central North Sea have much lower porosities and permeabilities than the reservoir chalks, and they attribute much of this difference to the role of hydrocarbon migration in preserving reservoir porosity. While overpressuring and hydrocarbon presence are important for the porosity-depth trajectory, the key observation from this figure is that porosity differences attributed to primary sedimentology tend to be preserved or even accentuated during the burial process. The mode of deposition and character of deposits can have real consequences for reservoir quality.

6.3.3 Fracturing

Fractures are important to the reservoir quality and performance of the Chalk; a high permeability fracture system is critical for the dynamic performance of a low matrix permeability reservoir. For reservoir chalks of the Norwegian Central Trough, well tests indicate effective permeabilities up to 300mD while the measured matrix permeability of the best reservoir is generally less than 10mD (Megson and Hardman, 2001). In this study, a fracture is considered to be any discontinuity that has accommodated strain by brittle failure and is at sub-seismic resolution. By this definition, fractures will include joints (opening mode fractures) and faults (shear mode fractures). There are many factors that affect the mechanism, intensity, and orientation of open fracturing. These factors relate to rock properties, the present-day stress state, and tectonics (including inversion and halokinesis).

Regarding rock properties, two key elements are the strength of the rock layer and the bed thickness. All else being equal, more competent (stronger) rocks will be more highly fractured because lithologies with a higher Young modulus will fail at lower extensional strains and thus, more often (Narr and Suppe, 1991). The strength properties of chalk depend on porosity and fluid character. The uniaxial compressive strength of chalk samples from UK outcrop localities was shown to vary in inverse relation to chalk porosity (Matthews and Clayton, 1993). Data from Risnes (2001) suggest the strength of oil-saturated rock is greater than the strength of water-saturated rock. In addition, anisotropies within the chalk (e.g. fossils and bedding plane irregularities) will act as sites of fracture initiation. Consequently, less pure, lower porosity, oil-saturated chalks are most prone to fracturing.

A relationship between bed thickness and mean fracture spacing has been documented widely in outcrop studies (e.g. Narr and Suppe, 1991; Peacock and Mann, 2005). The mean spacing of fractures is commonly proportional to bed thickness, such that fracture frequencies tend to be higher in thinner beds. The mechanical characteristics of adjacent beds may also affect fracture intensity. Fractures tend to stop at bedding planes that connect two beds of markedly different competence, whereas stacked units of similar high competence will tend to act as a single mechanical unit, and fractures will tend to be through-going and more widely spaced (Helgeson and Aydin, 1991). Relationships between bed thickness and fracture spacing that conform to these predictions were identified in the fractured chalks that wall the opencast Laegerdorf quarries of NW Germany, examined on reconnaissance visits in 2005 and 2006. Descriptions of the Chalk Group fractures have been presented previously in the scientific literature (Koestler and Erhrman, 1991).

The style and geometry of fracturing is influenced by tectonic setting, because fractures will develop in orientations related to the stress conditions at the time of their formation (e.g. Figure 2.3). Extensional fractures commonly form perpendicular to the direction of least compressive stress, whereas shear fractures typically form in conjugate pairs at an acute angle to the maximum compressive stress direction (Peacock and Mann, 2005). Tension fractures are most common when differential stresses are relatively low (i.e. the load stress is not much greater than the horizontal compressive stresses). This might occur at shallow depths, or under high fluid pressures at reservoir depths. The palaeostress history is also important because it controls the sequence of fracture development and therefore the geometry of the fracture network.

This section has emphasised the importance of sedimentology and fracturing as controls on Chalk Group reservoir quality. The next two sections of this chapter examine evidence for Late Cretaceous tectonic inversion and salt movements having influenced Chalk Group sedimentology and fracturing, and therefore reservoir quality.

6.4 Influence of Active Tectonics on Chalk Group Sedimentology

Seismic interpretation has highlighted considerable thickness variations in the Chalk Group across the Norwegian Central Trough (Figure 4.11; Chapter 4 and 5 section interpretations). The existence of allochthonous chalks is envisaged as the principal (albeit not the only) explanation for the observed dramatic variations in Chalk Group thickness. From seismic sections across the Lindesnes Ridge presented in Chapter 5, including Figure 5.4c which is depicted without vertical exaggeration and with compaction effects removed, it is evident that slopes exceed 5° in some localities. Such slopes are significantly steeper than those of approximately 1° from horizontal that are deemed sufficient to trigger slope instability in poorly consolidated siliciclastic sediments deposited in marine environments (e.g. Pratson et al., 1996).

The interpretation of core samples by workers at ConocoPhillips has suggested that the Ekofisk and Tor Formations comprise the highest proportion of allochthonous chalks (Beeley, 2005, *pers. comm.*; Figure 6.2). While it has not proved possible to map the Ekofisk and Tor Formations separately, due to the low contrast in physical properties between them (see Section 6.2.1), it was possible to map the top Hod reflection regionally and thereby generate an isochron map for the combined thickness of the Tor and Ekofisk Formations. This map is shown in Figure 6.7. The Tor and Ekofisk Formations thicken markedly away from the Lindesnes Ridge. This is interpreted as evidence the ridge was uplifting at the time of Tor and Ekofisk Formation deposition (i.e. Maastrichtian into Early Palaeocene times), with allochthonous chalks being actively shed from the ridge into the Feda Graben. From the pattern of contours it appears there may have been a northward directed transport of allochthonous material along the ridge, as well as shedding of material from the ridge.

The observations from Figure 6.7 have not been developed into a more substantial interpretation of allochthonous movements, because this approach belies the true complexity expected from the varied nature of allochthonous lithofacies recognised in the Norwegian Central Trough, indicative of varied mass movement processes. Many allochthonous flows may have been triggered by sub-seismic scale bathymetric features, rather than the kilometre-scale structural elements documented in Chapter 5. Evidence for this is that allochthonous chalks are recognised in core samples from at or near crestal locations, as well as locations further from Chalk Group highs. Crestal locations may have experienced minor horst and graben development in response to outer-arc extension. A particular problem is to understand the distance travelled by allochthonous units away from their original site of deposition. Transport distances are likely to have been variable, due to the varied nature of the processes occurring. Slumps may have transported material only tens of metres, whereas turbidity currents may have transported material tens of kilometres. The relative absence of wells drilled away from Chalk Group highs limits the opportunity to describe the importance of allochthonous material in basinal areas, and to compare with elevated (crestal) locations.

In spite of its limitations, Figure 6.7 usefully emphasises the very considerable thickness differences in the shallowest part of the Chalk group across the Norwegian Central Trough, and reinforces the importance of active tectonics at this time as an explanation for such variability in thickness of a pelagic deposit. While the complete picture of chalk dispersal cannot be reached, it is evident that palaeorelief generated by compression has influenced deposition, and this has implications for reservoir quality. It is interesting to observe that thinning of the top Hod to top Ekofisk interval is at its greatest where salt is more involved in the inversion structure, at the southernmost localities (A, B) on the Lindesnes Ridge (Figure 5.16).

6.5 Influence of Active Tectonics on Chalk Group Fracturing

Constraints on the fracture system can be derived from core, borehole image data, flow data and certain types of seismic data (e.g. azimuthal-anisotropy studies, such as those of Hall and Kendall, 2003). Core can be a definitive source of fracture information and procedures for describing it are well established (e.g. Kulander et al. 1990). Natural fractures are abundant in chalk cores from the Norwegian Central Trough and can be readily discerned. Four types of fracture were observed during reconnaissance analysis at the ResLabs core facility in October 2006; tectonic, stylolite-associated, healed and irregular. Figure 6.8 shows the two types considered herein to be important for reservoir quality; tectonic and stylolite-associated fractures.

Tectonic fractures (Figure 6.8a) tend to be planar, parallel or conjugate surfaces with steep dips; dips are usually inclined at approximately 70° to bedding (indicative of shear fractures formed where maximum compressive stress is vertical). Tectonic fracture types appear also to include sub-vertical joints (common) and bedding parallel fractures (less common). To distinguish between shear and tension fractures in chalk core is difficult, because stratigraphic offsets tend not to be resolvable. Tectonic fractures appear important in enhancing vertical and horizontal permeability, with inferred lengths on the order of metres or more.

Stylolite-associated fractures are also interpreted as important for reservoir productivity (Figure 6.8b). Stylolites are uneven surfaces of insoluble residues that form perpendicular to the maximum compressive stress direction. Stylolite-associated fractures tend to comprise anastamosing networks of subvertical fractures, up to several tens of centimetres in length. Unlike tectonic fractures, stylolite-associated fractures do not cross-cut stylolite boundaries and are presumed to enhance horizontal rather than vertical permeability; their overall contribution to permeability appears low relative to tectonic fractures. Stylolites appear to be relatively uncommon features in chalk core from the Ekofisk and Tor Formations. Stylolite-associated fractures have sometimes been referred to as 'tension gashes' in previous studies (e.g. Watts, 1983).

Healed fractures are also relatively uncommon. Their presence may reflect early fracturing in only semi-lithified sediments, with fractures being rapidly infilled with chalk matrix. This is in contrast to the tectonic and stylolite-associated fractures, that are filled with sparitic calcite or silica if not open. Some fractures cannot be obviously categorised as tectonic, stylolite-associated or healed, and these are described as irregular. They may include fractures related to drilling, core recovery, sample preparation or handling; all of which might conceivably cause modification of natural fractures or the creation of new ones.

A considerable body of evidence has been presented to support the notion that fracture intensity and orientation is related to fold development, and that the most intense fracturing tends to be concentrated in high strain zones (e.g. Hanks et al., 1997; Jamison, 1997). The highest strain zones are typically the hinge zones of a fold. The stress states within a developing fold can be predicted from simple theory (e.g. Ramsey and Huber, 1987): flexures may cause fractures associated with layer-parallel extension in the outer arc of a fold, and fractures associated with contractions in the inner arc.

Numerical modelling experiments of strain development in forced folds (e.g. Couples et al., 1998) have been used to infer strain states through time within reservoir structures. The calculated strain history can then be used to predict fracture patterns. This numerical modelling approach is advantageous because it allows sensitivity testing of parameters that affect the strain characteristics of a forced fold. Thus, the method permits a degree of control over fracture predictions that is otherwise absent. The quality of the predictions depends on the suitability of the experimental setup. One month was spent at the petroleum engineering institute of Heriot Watt University in Edinburgh, examining the feasibility of modifying an existing numerical model (Structural Analysis Via Finite Element Modelling code, as implemented previously by Couples and Lewis, 1998) for the specific case of Chalk Group folds within the Norwegian Central Trough. It was decided not to pursue this work further, given the time constraints and greater goals of this study. A recommendation for follow up is to research the best way that physical behaviour of halite can be incorporated into existing numerical models for the prediction of basin strain history.

Where it is not possible to model strain directly, curvature (i.e. rate of change of dip) is commonly used as a proxy for strain, and therefore for fracture density assessment (e.g. Masaferro et al, 2003; Hennings et al., 2000; Sanders et al., 2004). The advantage with curvature is that this is a parameter that can be evaluated directly from seismic mapping. For example, curvature can be inferred from dip maps computed in GeoFrame.

Figure 6.9a is a dip-azimuth map for the top Chalk Group; the greyscale colour varies as a function of azimuth. This figure allows the identification of prominent slopes and discriminates tilted regional lineaments from more localised structural highs. The reliability of the top Chalk dip-azimuth map is related to the density of top Chalk TWT interpretation, from which the attribute is derived. Dip-azimuths calculated within the cnsmerge survey area are more reliable than those calculated within the ga3d93 survey area. On the dip-azimuth map the ga3d93 area is partially obscured by a NW-SE oriented striping artefact. The artefact is due to a relatively low density of TWT line interpretations in this area; inline and crossline interpretation spacings of approximately 1km, rather than the less than 400m elsewhere (Figure 4.10 shows the interpreted line locations), were not sufficient to produce a quality attribute interpolation. In hindsight, the dip-azimuth map would have been improved in the ga3d93 area by a more detailed interpretation; 3D auto-tracking tools would have facilitated the timely completion of this additional interpretation.

Several trends are clear on Figure 6.9a. Firstly, the NNW-SSE oriented Lindesnes Ridge that overlies the Skrubbe fault can be clearly seen. The axis of the ridge has been labelled; it divides a westerly dipping slope on the west side of the crest from an easterly dipping slope on the east side. At first glance, the azimuth of the slope appears uniform on either side of the fault trace. In detail, dip azimuths vary by tens of degrees from the axis-perpendicular azimuth, suggesting second order complexity along the axis. This complexity, in the form of local highs along the ridge, is considered the result of the variable presence of salt in the immediate hangingwall to the Skrubbe fault. If there was no salt involvement, the pattern of dip-azimuths associated with the inverted Skrubbe fault is likely to have been more consistent. Secondly, while there is a clear NNW-SSE trend to ridge axes in the SE Feda Graben and far NW corner of the study area, ridges trend in a more WNW-ESE orientation in the northern part of the Feda Graben and in the Breiflabb Basin. This emphasises that there has been structural uplift on pre-existing faults of variable orientation. Thirdly, relief in the West Feda Graben is dominated by localized rather than linear trends, indicative of the relative importance of salt movements in developing relief in that area.

The likely fracture patterns associated with development of top Chalk highs with the various geometries identified in this and earlier figures have been considered. Figures 6.9c and 6.9d illustrate idealized patterns of tectonic fracturing expected for an inversion monocline and for the crest of a salt diapir, respectively. The fracture characteristics of the different Norwegian Central Trough uplifts most likely comprise a range of intermediate forms between 'inversion monocline' and 'salt diapir' end members. Fracture patterns are expected to vary within any given uplift, where the presence of salt beneath the uplift, as well as the timing and magnitude of movement, has varied. In some instances, monoclines may have evolved into domes as ongoing salt movements modified a pre-existing high.

Fracturing is expected to be greatest where the rate of change of dip is greatest, such as at the flanks of a monocline rather than at its crest. The significance of concentric relative to radial fractures is uncertain. According to a review of cross-disciplinary literature by Stewart (2007), radial faults are more common relative to concentric faults where a structural high has developed through doming rather than basin subsidence.

There are a number of papers that have examined fracture development within the Chalk Group in relation to folds that developed through salt doming. Examples include folds over the Machar (Foster and Rattey, 1993) and Banff (Evans et al., 1999; Davison, 2000) salt diapirs in the UK sector of the Central North Sea. Evans et al. (1999) documented radial and concentric fracture patterns over the Banff field and Davison (2000) demonstrated that radial normal faults are a common structural feature overlying the UK Central North Sea chalk fields. According to Davison (2000), the greatest amount of fracturing occurs in crestal areas, where (in the case of the Banff field) as much as 80% of the chalk is affected by fractures with spacing less than 10cm.

The present day stress state is a key indicator of whether fractures are open to conduct reservoir fluids. Fractures perpendicular to the maximum compressive stress direction are considered more likely to be closed than those fractures perpendicular to the minimum compressive stress direction (Crampin, 1987). The orientations of present-day in situ stresses can be determined from the orientations of borehole breakouts and drilling-induced fractures (e.g. Teufel et al., 1991). In the Norwegian Central Trough, the present-day minimum horizontal compressive stress direction is oriented NW-SE, and so open fractures are inferred to be preferentially oriented in a NW-SE direction.

6.6 Summary

Allochthonous chalks tend to have higher porosities and permeabilities than autochthonous chalks from their initial deposition and throughout their burial history. Allochthonous chalks are relatively uncommon in Western Europe but are abundant in parts of the Norwegian Central Trough. The relative abundance of allochthonous chalk is best explained by Late Cretaceous active tectonics. Chalk group sedimentation began at a time of relative structural quiescence in a slowly subsiding basin. Allochthonous chalks were initially rare but became more prevalent at later times; especially during Maastrichtian and Early Palaeocene times, contemporaneous with deposition of the Ekofisk and Tor Formations. By these later stages of Chalk Group deposition, the morphology of the Norwegian Central Trough was controlled by the growth of major tectonic inversion and salt structures. The mechanisms of fault movement and salt mobility were both causing (or contributing to) syn-depositional uplift, and therefore encouraging allochthonous activity.

The patterns of tectonic fracturing within the Chalk Group are inferred to be variable across a range of scales. This is a reflection of the varied geometries of uplift, and the complex histories of their development.

7 *Supplementary Results from Onshore Field Observations*

7.1 Introduction

The previous chapter addressed reservoir-scale implications of tectonic inversion and halokinesis. In the Norwegian Central Trough, as with other subsurface localities, limited resolution and density of data impede the detailed understanding of reservoir properties. This chapter addresses the potential of outcrop analogues to provide additional insights into the relationship between reservoir-scale characteristics and trap-forming mechanism.

7.2 Justification

The principal advantage of onshore outcrop studies is that they allow a degree of observation not possible in the subsurface. They enable documentation of sub-seismic resolution (meso-scale) characteristics of salt and inversion tectonics. Outcrop measurements can be dense and three dimensional. Measurements can encompass all parts of a structure, unlike in the subsurface where typically only the (prospective) shallow crestal areas that are sampled by boreholes.

The original idea for a research project was to examine the relationship between the seismically resolvable and sub-seismic scale characteristics of tectonic inversion. Thus, outcrop studies were intended to form a major element of the research. Could a generic relationship be identified in outcrop between sub-seismic structural parameters (i.e. fracture intensity and orientation) and the larger scale seismically-observable geometric parameters of inversion (e.g. location on the fold and fold curvature)? Might such a relationship allow prediction of sub-seismic fracture properties using only seismic reflection data? The seismic data of interest had been acquired in the Norwegian Central Trough, where tectonic inversion had been described or invoked previously (see Section 2.5.2), but never thoroughly investigated nor fully understood. Quantifying this tectonic inversion would constitute the additional major aim of research. Ultimately, however, this second aim (which evolved into the interpretation of salt and fault relationships at the seismic scale) became the emphasis for research.

An outcrop locality was sought that might have the greatest potential to establish a link between macro- and meso-scale characteristics of tectonic inversion. It was felt that the ideal field study site would be a tectonically-inverted setting with excellent, three-dimensional exposures of folds developed within the inverted post-rift, occurring on scales similar to those of the Norwegian Central Trough (fold amplitudes on the order of 100m). The initial suggestion by ConocoPhillips, the CASE sponsor, was to observe tectonic inversion in the continental interior of the United States. Large areas of the North American interior have been affected by intracratonic compression. In some areas, near continuous sedimentation throughout the Phanerozoic has combined with moderate uplifts to preserve a unique record of the tectonic history. Such areas, where well exposed and easy to access, are favourable locations for the study of meso-scale inversion tectonics. As the focus of several key papers concerning the relationship between fracture patterns and fold growth (e.g. Mitra and Mount, 1998; Couples and Lewis, 1998; Hennings et al., 2003; Cooper et al., 2006), the Laramide structures of Utah, Wyoming and Colorado would be an appropriate choice. The Uinta Mountains of NE Utah were selected for field study. Tectonic inversion structures have been identified here in outcrop (e.g. Davis, 1978; Rowley et al., 1985) and from subsurface data (e.g. Hefner and Barrow, 1992) where over 400 oil wells have been drilled into anticlinal closures attributed to Laramide compression.

After an initial three month field season in the Uinta Mountains, and several months working with the Norwegian Central Trough seismic reflection data, it became evident that the applicability of field outcrop observations would be limited by the complexity and uniqueness of structures being mapped in the subsurface. Specifically, salt tectonics was recognised as a critical control over Norwegian Central Trough structural styles, with tectonic inversion geometries being affected by the presence of halite and the manner and timing of its movement.

The stratigraphic column in the Uinta Mountains does not include a thick halite-dominated formation akin to the Zechstein Group of the Norwegian Central Trough. In order to address this limitation, focus was initially shifted towards the Paradox Salt Basin of East Central Utah, where many of the Uinta lithologies have been deformed by mobile Carboniferous halites of the Paradox Formation (Matthews et al., 2004). The salt structures of East Central Utah have been previously examined as analogues for North Sea salt mobility (e.g. Stewart et al., 1999; Banbury, 2006). Later investigation of reservoir data from the Norwegian

Central Trough (Section 6.3) confirmed the importance of lithology (i.e. the Chalk Group) in controlling reservoir properties. In acknowledgement of this fact, focus shifted towards Chalk Group outcrops in inverted settings (South Dorset, UK) and halokinetically deformed settings (Laegerdorf, NW Germany).

The Paradox Basin and Laegerdorf localities were not studied in the same detail as the Uinta and Dorset localities; this work contributed little to the understanding of reservoir scales in the Norwegian Central Trough and so has been omitted from this thesis. In contrast, observations from the Uinta Mountains and South Dorset are reported in some detail, and their implications for the Norwegian Central Trough are considered.

7.3 Tectonic Inversion in the Eastern Uinta Mountains

7.3.1 Location and Geological Setting

Large areas of the United States continental interior have been affected by intracratonic compressional deformation. This is exemplified by the presence of numerous 'Laramide' basement-cored uplifts in present day Utah, Colorado and Wyoming. These uplifts have been dated as Late Cretaceous to Early Palaeogene (e.g. c.75 to 35Ma by Bird, 1998) and are attributed to NE-directed compression associated with tectonic plate convergence that has been occurring along the western margin of the North American craton since Jurassic times (Bird, 1998).

There are uncertainties concerning the exact mechanism that caused compression far within the plate interior, but the evidence for tectonic inversion related to this compression is well documented. The Uinta Mountains are a spectacular example (e.g. Hansen, 1986), and the Eastern end of the uplift was selected for study. The Uinta Mountains are situated in northeast Utah and northwesternmost Colorado, where they form an East-West oriented topographic high, some 200 km in length, 40 km in width, and up to ~3 km in elevation. Field mapping allied with borehole and seismic investigations has revealed that the Uinta Mountains are bounded to the North and South by reverse faults (e.g. Rowley et al., 1985; Stone, 1993). Figure 7.1 shows the principal tectonic elements that define the region.

The timing of uplift on the bounding faults has been constrained to Late Cretaceous- End Eocene, based on recognition and dating of thick synorogenic deposits in the Uinta and Green River Basins. This age, coupled with the distinctive basement-involved style of deformation, has caused uplift to be attributed to the Laramide Orogeny rather than the partially contemporaneous (c.119 Ma to 50 Ma according to Bird, 1998) Sevier Orogeny. The Uinta Mountains are situated in the foreland of the east-verging Sevier fold and thrust belt and partially overlap the Laramide Orogeny in location as well as time (Bradley and Bruhn, 1988). However, the structural styles are markedly different; Sevier deformation is characterized by 'thin-skinned' thrusts that detach above basement, rather than the basement-cored character of the Laramide Orogeny.

7.3.2 Stratigraphy

A detailed stratigraphic framework for the Eastern Uinta Mountains was proposed by Rowley et al. (1985), on the basis of extensive field mapping. This stratigraphy is summarized in Figure 7.2.

At the base of the section is the Precambrian Uinta Mountain Group. This thick succession of Middle Proterozoic siliciclastics, dated to ~1.1Ga (Hansen et al., 1983), forms the core of the partially-eroded anticline that constitutes the Uinta Mountains. From boreholes, the thickness of the Uinta Mountain Group has been demonstrated to exceed 7.5 km in places (Rowley et al., 1985). The deposition of such a thick sequence has been attributed to sedimentation within an active rift system (e.g. Sears et al., 1982). Subsequent deposition occurred in an essentially intraplate environment, although Late Cretaceous and Early Palaeogene (Laramide) compression was significant. The reverse faults that bound the present day Uinta uplift are interpreted as Laramide reactivations of the original rift elements (see Section 7.3.3).

Palaeozoic strata include a transgressive Cambrian sandstone (the Lodore Formation); a sequence of Cambrian through Permian shelfal carbonates and shales (the Madison, Doughnut, Humbug, Round Valley and Morgan Formations); and a major Late

Carboniferous (Pennsylvanian) to Permian aged aeolian sandstone, the Weber Formation (Krantz, 1995). The Triassic to Early Cretaceous sections includes marginal marine to terrestrial facies, such as grey shales of Morrison Formation, renowned for the presence of terrestrial dinosaur fossils within. These Mesozoic sediments reflect a transition from coastal plain to continental interior environments. This was followed in the Late Cretaceous by transgression of an epicontinental sea (the Cretaceous Western Interior Seaway), which led to the accumulation of marine shales at this time.

Cenozoic lithologies include thick Palaeocene and Eocene marginal to lacustrine facies that fill the Uinta and Green River Basins (including the Uinta Formation to the south of the Uinta Mountains, and the Green River Formation to the north). Similar sediments were for the most part not deposited in the High Uintas, because this area was undergoing uplift at this time, and in fact acted as a sediment source. The principal Cenozoic material in the High Uintas is the Oligocene-aged Bishop Conglomerate Formation that was deposited on a broad pediment surface ('the Gilbert Peak erosion surface') that has been K-Ar dated to 29 Ma (Hansen, 1986[b]). The presence of a pediment surface is indicative of a long period of quiescence having occurred after deposition of the youngest Eocene rocks. Today, this pediment surface is only partially preserved, and there are excellent exposures of the entire stratigraphic section across the range, tectonically undisturbed since the Laramide Orogeny.

7.3.3 Evidence of Tectonic Inversion

Some constraints on the manner and timing of faulting have been observed directly from structural and stratigraphic relationships in outcrop. Further constraints have been derived from seismic data and exploratory boreholes, which help to characterise faults at depth. These techniques have identified the presence of reverse faults beneath monoclinal folds that essentially drape the basement faults. Some of these reverse faults are interpreted as contractionally reactivated Precambrian basement-penetrating normal faults, because strata of the Uinta Mountain Group have been shown to thicken across the faults at depth (e.g. Stone, 1993). The structural style is demonstrated in Figure 7.3; a simplified cross section through the Eastern Uinta Mountains, along a North-South transect.

The north flank of the Uinta uplift is defined by a single major reverse fault, albeit divided into separate west (the North Flank-Henry's Fork Fault) and east (the Uinta-Sparks Fault) segments. Interpretation of the Uinta-Sparks Fault as an inversion structure is supported by hangingwall and footwall well constraints. Osmond (1986) documents three wells that cut the Uinta-Sparks fault; the Husky 7-3 Clay Basin well, the McMoran-Freeport 43-2A well, and the Champlin 31-19 Bear Springs well (Figure 7.1). The reversal of throw on the Uinta-Sparks fault occurred prior to deposition of the Middle Palaeocene Wasatch Formation within the Green River Basin (Hansen, 1986), as evidenced by the inclusion of pebbles of Cretaceous Mowry Shale as clasts within formation. Figure 7.4 shows the Uinta-Sparks Fault in outcrop.

The southern flank of the Uintas is particularly well-imaged seismically due to extensive hydrocarbon exploration in the northern part of the Uinta Basin. The pattern of faulting on the southern flank of the Uinta Mountains is relatively complicated, and there is a complex array of monoclines linked to faults at depth (e.g. Hefner and Barrow, 1992). Levels of exposure are exceptional, having been aided by the lithologies involved, by climate and by river incision processes. The area provides an ideal opportunity to study a tectonically inverted setting in three dimensions, and in particular, the manner in which moderate to steeply dipping reverse faults pass up-section into asymmetric folds in the stratigraphic cover. Figure 7.5 is an overview map of the studied area, and Figure 7.6 is an aerial photograph in which several of the monoclinal folds can be clearly seen.

7.3.4 Methodology

The goal was to study the deformation mechanisms associated with development of tectonic inversion structures, to record the fracture characteristics, and to compare these characteristics in different parts of the fold (e.g. to compare the steeply and gently dipping limbs of an asymmetric anticline). An initial reconnaissance was undertaken to ascertain accessibility and quality of outcrops. It was recognised that the most common outcrop lithology was the Weber Sandstone; the principal reservoir rock of the Rangely oilfield. The Weber Formation comprises high porosity and only partially cemented aeolian sandstone, with sub-rounded quartz grains typically 200-500 microns in diameter (Evans et al., 1995). It is, in respect of its near monominerallic purity and high porosity, somewhat akin to the

Chalk Group of the Norwegian Central Trough, and there may be similarities in its mechanical behaviour. Consequently, measurements concentrated upon this lithology alone.

It was decided to undertake fold and fracture observations at localities along four transects oriented approximately NW-SE. This orientation is perpendicular to the Island Park and Mitten Park fault trends, and perpendicular to the inferred Laramide compression direction of Bird (1998). In combination, these transects would enable trends to be examined both along-strike and perpendicular to strike across the various monoclinal folds.

Field observations of Weber sandstone outcrops were made at a total of 188 different localities during the 2005 field season. Some or all of the following parameters were measured at each locality: bedding orientation, fracture orientation, style, and intensity (fractures per unit length), fracture length, termination patterns, cross-cutting relationships and mineral fill. Where exposures were good, a line sampling technique was used to gather fracture intensity data in a systematic fashion, with as many as 30 fractures sampled at a given locality.

7.3.5 Results

Styles of deformation within the Weber Formation include folding, faulting, jointing and cataclasis (deformation band development). Folds, faults and the relationships between them are seen most clearly on the regional transects BB', CC', DD' and EE' displayed in figure 7.7. On section BB', the Split Mountain anticline has an essentially box fold geometry, with uplift attributed to movement on the Island Park fault that remains buried here. The Cliff Ridge monocline (that passes eastwards into the Blue Mountain monocline, e.g. transect CC') is more asymmetric, with a southern limb that dips by as much as 90° showing strong and penetrative deformation. Along transect DD' there is an especially tight NW-verging antiform-synform pair. The fold is located at the intersection of the Ruple Point and Mitten Park monoclines and is interpreted as an interference structure. At this locality the Weber Sandstone is heavily influenced by deformation bands and jointing. The north-verging Yampa monocline observed on transect DD' is linked to an underlying fault exposed along transect EE', where it juxtaposes the Precambrian Uinta Mountain Group against the Triassic Moenkopi Formation.

The majority of fractures appear to be dilatational (joints), due to their sub-vertical and/or bedding-perpendicular orientations and because they do not show any indication of shear movement. Joint densities are higher on the fold hinges than elsewhere, but this relationship was less pronounced than expected from theoretical considerations of strain variation along a fold hinge (e.g. Lisle, 1994; Stearns and Friedman, 1972). The observed fracture pattern may have been strongly influenced by other processes, such as the Cenozoic unroofing that post-dated tectonic inversion.

Cataclasis is demonstrated by the presence of deformation bands (granulation seams) within the Weber Sandstone. The formation of these bands is attributed to grain crushing (with associated porosity reduction) and shear (e.g. Aydin and Johnson, 1978). Deformation bands are a phenomenon specific to pure sandstones and do not occur in chalks. However, it is postulated here that deformation bands in sandstone are in many respects akin to pore collapse structures in chalk (laboratory tests of chalk demonstrate its tendency to deform by the crushing of individual coccoliths). As with deformation band development, pore collapse is a mechanism of distributed shear deformation that can occur as opposed to (or more typically in addition to) shear failure through faulting.

There were 188 field measurement localities and these have been grouped into a smaller and more manageable number of 'fracture' localities for the purpose of displaying and interpreting fracture patterns. This grouping was considered acceptable, provided grouped field localities share a common structural position (i.e. bed orientation), and are in close proximity (typically not much more than tens of metres apart). Standard stereogram software was then used to plot the cumulative fracture orientation data; a stereonet for each of the 19 fracture localities is shown in Figure 7.9. There is one bedding measurement for each stereonet, deemed representative of all bedding values for that given fracture locality. The field data from which these stereonets are sourced is listed in Appendix 2.

For a given fracture locality, joint orientations are typically consistent, comprising one or more sets, of which the dominant (most common) set is usually related to bed dip magnitude and azimuth. That joints on different fold limbs strike in different directions suggests these fractures are most likely the product of fold growth, rather than other mechanisms, such as

Cenozoic unroofing, that might give rise to uniform fracture orientations across the region. The dominant joints strike approximately parallel to bedding, with steep dips in the opposite direction to bedding (e.g. B1, B4, D1, E5). There are exceptions to this observation, however (e.g. B6, D2, E4, E7; joints oriented at high angle to bedding). At fracture localities B1 to B4 (Split Mountain, southern limb) a dominant WNW trending fracture set is clearly developed. This joint set is in approximate alignment with the axis of the Split Mountain anticline and shows the clearest relationship to bedding of all the folds. It is possible that the relationship between bedding and joint orientation is less clear for NE-SW striking folds because of the overprint of the E-W trend in the eastern part of the studied area (e.g. D2). It is probable that joint patterns are less strongly developed where beds are shallow dipping (compare C1 with B1), but there is no strong evidence that the steepest dipping fold limbs give rise to the most tightly aligned joint sets (e.g. bed dip at B4 is steeper than at B3, but the dominant joint set is more consistently aligned at B3). In contrast to the joint sets, deformation band orientations are more variable and less obviously related to bed dip.

Figure 7.8 shows the Mitten Park fault in outcrop. It juxtaposes Pennsylvanian Round Valley Limestone against Precambrian to Mississippian aged strata. Near-horizontal strata are folded to near vertical in the vicinity of the fault. Folding is interpreted not as fault drag, but as the synclinal portion of a fault-propagation fold that has subsequently been faulted through. This conclusion is drawn because of the appearance of the same structure along strike; it has the form of a monoclinal fold above a fault.

Fracture patterns in Weber Sandstone of the footwall were examined along a transect towards the Mitten Park fault. This enabled variations in the character of fracturing to be observed in relation to position on the synclinal fold. It was observed that intensity of deformation band development increases markedly as the fault is approached (and therefore as a function of the dip of the fold limb). Where horizontally bedded, there is relatively little deformation within the Weber Sandstone. Within approximately 150m from the synclinal axis, bedding-parallel deformation bands can be observed and are widely spaced (~3m apart). Closer than approximately 50m to the synclinal axis, bedding-parallel deformation bands are more intensely developed and occur in association with deformation bands that cut the bedding at high angles (linking the bedding-parallel bands). Minor (centimetre-scale) offset reverse faults are also observed here. Within a few metres of the synclinal axis, the deformation bands are at their most developed, tending to occur in thick bands (with thicknesses > ~1cm and up to ~10cm) and forming complex linkage networks. The dramatic increase in the intensity of deformation band development and shear-mode fractures is evidently related to the development of the forced fold.

From study of the Mitten Park fault, it appears that the effect of initial shortening was to cause rotation of the cover sequence, involving both bedding-parallel slip and folding. As beds within the cover continue to steepen, the angle between the bedding and the horizontal (principal stress direction) increases. This makes further bedding-parallel slip less likely. Further shortening occurs by increased displacement on the basement faults and may result in a faulted fold (such as the Mitten Park structure, Figure 7.8). In this case, after initial development of the fold the underlying fault has broken through the steep limb region, and the footwall syncline has been preserved.

7.4 Tectonic Inversion in South Dorset

7.4.1 Location and Geological Setting

There are coastal outcrops of the Chalk Group across much of Southern England. Of relevance to this study is the Chalk of South Dorset, which forms a folded cover to the inverted Purbeck normal fault (Underhill and Paterson, 1998; Figure 2.6). Figure 7.10 shows the area of interest; a 1km stretch of coastline between Durdle Door (50°37'05N, 02°16'05W) and Bat's Head (50°37'05N, 02°17'05W). At this location the structural position of the Chalk in outcrop corresponds to the steeply-dipping (north) limb of the approximately east-west striking, east-plunging, north-verging Purbeck monocline. South Dorset was therefore considered a suitable location to observe the fracture patterns associated with the steeply dipping forelimb of an inversion-related monocline where the lithology is Chalk. As such, it is deemed the most appropriate structural analogue for monoclines within the Norwegian Central Trough that were documented in Chapters 4 and 5.

Seismic studies of the area (e.g. Underhill and Paterson, 1998; Underhill and Stoneley, 1998) have cast light on the relationship of the fold to the underlying Purbeck fault. The Purbeck fault represents the northern boundary to the English Channel and Portland-Wight Basins, formed through Jurassic and Cretaceous (predominantly Late Kimmeridgian) extension. Tectonic inversion has been dated to Oligocene and Miocene times, with compressive stresses attributed to the Alpine Orogeny. The surface expression of Purbeck fault varies along strike but tends to comprise multiple low-angle fault strands, interpreted as short-cut thrusts (e.g. Figure 7.10b). Similar complexities may exist in the Norwegian Central Trough; the Lindesnes Ridge in Figures 5.14 and 5.15 are the most obvious locations that might be affected by short-cut thrusts of this type.

There are along-strike variations in fold curvature at the Purbeck monocline. On account of its easterly plunge, the more westerly parts of the monocline correspond to deepest parts of the cover, and are the most intensely folded and deformed. This same phenomenon is observed at the Lindesnes Ridge, where deformation is more pronounced in the Late Cretaceous section than in the overlying Earliest Cenozoic strata. The Purbeck monocline was studied where it outcrops between Durdle Door and Bat's Head; at this locality the monocline's northern limb dips at more than 80° and is overturned in places.

7.4.2 Fractures in the Chalk Group

The fractured Chalks of the Purbeck monocline were first studied by Arkell (1938), and subsequently by Phillips (1964), and Bevan (1985). Arkell (1938) identified slickensides on bedding planes, and attributed this observation to (bedding-parallel) flexural slip occurring during fold development. As with the Uintas study, bedding-parallel slip can be interpreted as an early response to compression and folding in the competent beds. They have also analysed fractures oriented obliquely with respect to bedding, and have presented different interpretations to explain these features.

In this study observations of mesoscale fractures were made from the Chalk Group of Swyre Head (Figure 7.11) and Bat's Head (Figure 7.12). In both cases, the most prominent features in the Chalk are meso-scale shear fractures, many of which exhibit a pronounced shear offset. They form two sets; one north-dipping and one south-dipping. These appear to be conjugate, symmetrically arranged relative to the bedding. The shear fractures are best explained as a result of a bedding-perpendicular maximum principal stress direction ($\sigma 1$).

Both north and south-dipping fractures are observed to offset bedding and are deemed to have formed after flexural slip had occurred. These concepts are illustrated in Figure 7.12c.

The Chalks outcropping in Dorset are Turonian in age; relatively flinty, hard and nodular chalks of the Upper Cretaceous Chalk Group (House, 1985). They are more indurated than the Ekofisk and Tor Formation Chalks of the Norwegian Central Trough observed in core. For this reason, flexural slip is expected to have been a less significant phenomenon in Chalks of the Norwegian Central Trough. There may, however, have been a significant amount of conjugate shear fracturing on the more steeply dipping slopes in the Norwegian Central Trough, in a manner akin to that observed in the Dorset outcrops.

7.5 Applicability of Onshore Observations to the Norwegian Central Trough

Reservoir-scale observations documented in this chapter remain somewhat limited in their applicability to the Norwegian Central Trough discussions outlined elsewhere in the thesis. Given the number of differences between datasets, it was difficult to compare and contrast non-salt involved features from outcrop with the salt-involved features of the Norwegian Central North Sea. To make comparisons at the reservoir scale more thoroughly would require significant additional data gathering and/or access to subsurface datasets and would be a major undertaking, much beyond the scope of this thesis.

It is evident that while the Uinta Mountains afforded an exceptional opportunity to study salt-absent tectonic inversion at outcrop, parallels to tectonic inversion in the Norwegian Central Trough are restricted by the number of unconstrained variables. One example is the lithological contrast between the aeolian sandstones of the Weber Formation versus the chalks of the Norwegian Central Trough; this contrast will contribute towards different mechanical behaviours. (There is however, some precedent for comparing deformation in chalks with deformation in porous sandstones (e.g. Schultz and Siddharthan, 2005; Peacock and Mann, 2005)). Given more time it would have worthwhile to develop the Uintas observations into a thorough, stand-alone analysis. A recommendation for future work would be to undertake additional studies in the southern Uintas in order to fully unravel the fracture chronology and thereby improve understanding of stress evolution in the area. It proved not possible to do so using data collected in a single field season.

The folds and faults examined in the Uintas are in some cases difficult to prove as inversion structures. It is interesting to consider the extent to which it might affect reservoir scale properties if these features were reverse faults and not inversion structures. An inverted fault would have an orientation that was related to the extensional stress regime and may not be aligned with the subsequent compressive stress regime. This would have implications for the minor fold and fracture characteristics of the structure. In general, one would expect a more complex pattern of folding and fracturing when dealing with inversion structures rather than reverse faults.

7.6 Summary

Previous studies of tectonic inversion have tended to focus on evidence from seismic data. By considering supplementary results from onshore field observations, this study has addressed the problem of seismic resolution as a limiting factor in the ability to document the fracture patterns associated with inversion. The Uinta Mountain and South Dorset uplifts described in this chapter are most directly comparable, in terms of geometry and scale, with the least salt-prone uplifts of the Norwegian Central Trough.

8 *Discussion and Conclusions*

8.1 Introduction

This final chapter summarises the results of research, building on arguments presented in the preceding chapters. Results are reviewed against the background of previous Central North Sea and other relevant studies, and relative to the initial objectives of this study. Remaining uncertainties are clarified and recommendations for future work are given.

8.2 New Insights into the Tectonic History of the Norwegian Central Trough

Interpretation of seismic reflection data constrained by subsurface information from wireline logs has revealed new insights into the tectonic history of the Norwegian Central Trough; a 5,200km^2 region within the Central North Sea that comprises a complex system of Permian-Cretaceous grabens and half-grabens. Fresh insights have come from mapping the pre-Zechstein basement, laterally heterogeneous Zechstein salt deposits and a complex post-salt basin fill that revealed evidence of active movements during sedimentation. Original products have included detailed seismic line interpretations, two-way travel time maps, section restorations and conceptual diagrams (this chapter).

A wide variety of structural behaviours and interactions have been documented. This study has utilised some of the most comprehensive, high-resolution 3D seismic reflection data yet gathered over the Central North Sea in order to undertake these interpretations, helping to address previously unanswerable questions relating to the highly complex structure seen within the post-Variscan section of the Norwegian Central Trough. The approach to interpretation differed from that undertaken by hydrocarbon explorationists who have worked with similar data. Rather than focusing on the geometric description of structural closures, this study has sought to comprehend the mechanisms that generated these closures and to explain observed structural variety in terms of these mechanisms. The large size of the dataset has allowed comparison of different structures on a whole-of-basin scale. It was a novel undertaking to compare areas within the basin in terms of salt-presence, tectonic inversion, and salt-involvement in inversion.

The Zechstein salt interval and six supra-salt sequence boundaries were mapped in three dimensional detail and independently of other workers. The resulting time surfaces are acceptable proxies for their corresponding depth surfaces (see Section 3.6): these maps and their associated isochrons were presented in Figures 4.3 to 4.16. The locations of interpreted sections are posted on the two-way travel time maps, emphasising the density and therefore robustness of results and the amount of original work required to generate detailed surfaces from an initial, uninterpreted seismic volume. The necessity of such fine line spacing was discussed in Section 4.2. In hindsight, it would have made for a more efficient and effective research exercise if the initial seismic interpretation work had been undertaken in closer collaboration with ConocoPhillips. To have done so would have facilitated access to industry-standard software support and guidance regarding best practise techniques. For example, auto-tracking and surface gridding tools could have been implemented more successfully, resulting in the timelier production and tidier appearance of final maps.

The top and base Zechstein (top Rotliegend) maps are a valuable product from this study. The top Rotliegend map (Figure 4.3) defines the structural framework and is a significant update and improvement upon early published basement interpretations. Interpretations of the structural framework that do exist in published scientific literature are based on older, 2D seismic data that was more limited in extent and typically poor in resolution; a problem exacerbated by the region's structural complexity. Previous studies (e.g. Roberts et al., 1990; Sears et al., 1993; Gowers et al., 1993) postulated varied, at times contradictory models; whilst later studies (e.g. Brasher, 1995; Farmer and Barkved, 1999) have done little to refine these models, focusing instead on the improved characterization of individual oilfields. The top Rotliegend map incorporates a greater number of structural lineaments than were proposed by Gowers et al. (1993), and the orientation and magnitude of offset on faults is captured more definitively than in Sears et al. (1993).

The focus for this study has been the analysis of salt and fault interactions identified during mapping work. To the oil industry, the Zechstein and deeper geology of the Norwegian Central Trough has previously been considered a secondary interest, given source rocks of Jurassic age and Chalk Group reservoir, such that the entire petroleum system is contained within the post-salt section. However, this study has highlighted that the manner of salt and fault interaction is fundamental to the story of trap generation. Of the 18 seismic sections presented in chapters four and five, all have contributed insights into the tectonic history of the Norwegian Central Trough and the nature of salt-fault interactions occurring within it. Five of these sections were quantitatively restored using 2DMove software (Figures 5.2 to 5.6), in an additional and innovative bid to constrain structural controls. The inclusion of a top Triassic map (Figure 4.6) was a further aid to the interpretation of salt structural history, identifying regional differences in the character of Triassic sedimentation. This work has resulted in the most comprehensive description of salt-fault behaviour in the Norwegian Central Trough to date.

8.3 Timing and Intensity of NCT Inversion Tectonics

It has been conclusively demonstrated in this thesis that tectonic inversion occurred in the Norwegian Central Trough. The main evidence derives from the Lindesnes Ridge, where hangingwall uplift cannot be explained by salt movements alone and requires compression to be invoked as an explanation. The results of decompaction modelling using 2DMove allows differential compaction to be disregarded as the sole mechanism for Chalk Group uplift along the ridge. In addition to the Lindesnes Ridge, several other inversion axes are documented; most notably at the Albuskjell and Tor structures, as well as at the western margin of the Piggvar Terrace. Consequently, tectonic inversion has been recognised further north in the Central North Sea than previous studies have indicated (see Figure 2.24). It has also been shown that the manner of inversion in the Norwegian Central Trough differs markedly from the classically interpreted concept of tectonic inversion (e.g. Cooper et al., 1989), and that this is due to the role of salt tectonics. The role of salt in influencing inversion tectonics is discussed further in Section 8.4.

Timing of inversion was constrained through recognition of seismic-stratigraphic relationships, relative to the age-calibrated regionally interpreted seismic horizons. The timings of uplift are broadly consistent across the Norwegian Central Trough, while the magnitude of uplift is more variable. Inversion occurred during two discrete, albeit pulsed, intervals. The first event occurred during the Late Cretaceous, with an approximately Turonian-Santonian onset and a Maastrichtian-Danian climax (synchronous with deposition of the Tor and Ekofisk Formations). The Upper Cretaceous uplift was followed by a second, post-Danian phase of tectonic inversion, occurring during deposition of the Hordaland and Lowermost Nordaland Groups. The peak of inversion is constrained to an approximately Late Eocene- Early Oligocene age.

Broadly, the inversion timings identified in the Norwegian Central Trough are consistent with those found elsewhere in NW Europe (these timings were summarised in Figure 2.25). That the timings of compression generally accord with those described for other basins on the European Plate suggest the causes of compression are linked and related to plate scale stress conditions, governed by plate boundary forces. It has not been possible to resolve more than two distinct compressional events, in contrast to Ziegler (1990), who proposed four distinct pulses of West European inversion dated Late Cretaceous (Late Turonian to Campanian), Middle Palaeocene, Late Eocene to Early Oligocene, and Late Oligocene to Early Miocene. To some extent, this may reflect the limited resolution of inversion in the Norwegian Central Trough, relative to other West European basins that have experienced greater uplift. However, many previous analyses of these other West European basins have also identified fewer than four discrete pulses of compression.

Along with the opening of the North Atlantic Ocean, Alpine collisional tectonics is widely accepted as the key event driving NW European inversions (see Section 2.5.2). Elements that contribute to the Late Cretaceous and Cenozoic stress field most likely include: NW directed 'Alpine' compression; North directed 'Pyrenean' compression; SE directed ridge push associated with North Atlantic opening; East directed ridge push associated with Central Atlantic opening; and NE directed ridge push from the opening of the Bay of Biscay. It is not possible to definitively link an observed episode of compression with a specific plate boundary event. In part, this is because of the imprecision with which orientations and timings of inversion are constrained. Significantly, it is also because plate boundary events are themselves highly complex, overlapping in time and variable in their magnitude. The compressional history of the Norwegian Central Trough is considered best described as the sum of a set of continually changing plate boundary conditions.

In the Norwegian Central Trough, the post-Danian phase of compression was both less intense and somewhat different in structural style relative to the Late Cretaceous phase. This observation can be attributed to broad differences in the underlying mechanisms. A candidate explanation relates to the history of relative movements between the African and European Plates. Late Eocene to Oligocene compression coincides with the resumption of Africa-Europe convergence rates to as much as 20 mm/yr after several millennia of negligible convergence during the Early Cenozoic (Figure 2.27; Smith, 2006). A distinct difference in Late Cretaceous versus Cenozoic compressional structural style has also recently been documented in the Central Danish Basin by Nielsen et al. (2007). The Cretaceous phase was characterized by narrow uplift zones, reverse reactivation of faults and crustal shortening. In contrast the Cenozoic (mainly Eocene) phase was characterized by dome-like uplift of a wider area with only minor fault movements.

Inversion in the Norwegian Central Trough differs from that in most locations in NW Europe due to the involvement of salt from the basin fill. In addition, the presence of salt impedes the ability to interpret inversion. For example, it is difficult to ascertain the absolute amount of Cretaceous and Cenozoic shortening within the Norwegian Central Trough, due to the presence of mobile Zechstein halite that invalidates the plane strain criterion on which 2D restoration depends. Determination of shortening is also hindered by uncertainty in bathymetries at the time of deposition.

8.4 Role of Salt in Influencing Inversion Tectonics

The influence of salt movements on structural styles has been documented in a variety of previous studies: key findings from these studies were reviewed in Section 2.3. The majority of observations relate to salt behaviour in extensional (basin-forming) environments. Comparatively few studies have considered the role of salt in basin inversion despite the existence of many sedimentary basins, across the globe, with mobile halite infill and a history of late compression. This study of the Norwegian Central North Sea has documented salt behaviour during the Late Cretaceous and Cenozoic compressions that affected much of Europe. Seismic sections in Chapters 4 and 5 provide constraints on the timing of salt movement, extension, and compression. These sections have illustrated how salt has influenced the structures produced by inversion; a complex relationship here shown to be fundamentally linked to the extensional (pre-inversion) phase of basin history.

The extent to which halite affects inversion tectonics is underlined by the difference between structural geometries in salt-free versus salt-prone inverted sedimentary basins. In salt-prone basins, compressional uplifts are less clearly related to the extensional framework of the basin (e.g. compare Figure 5.10 with Figure 2.2). This disconnect is due to halite's anomalous physical properties and its mobile behaviour in response to the prevailing stress conditions. The presence of salt typically results in a structural decoupling between supra-salt strata and underlying basement. For example, basement faults rarely extend into the post-Permian section, except where halite is thin or absent. Also, the presence of salt in the basin fill is indicative of a complex sediment fill history. Lateral variations in salt presence are indicative of pre-inversion structural heterogeneities. These heterogeneities affected the response to compression.

A salt movement chronology is presented in Figure 8.1. This figure also summarises the relative amounts of halokinesis during basin evolution. It illustrates these movements in relation to the inversion timings (as reviewed in Section 8.3) and to the principal plate tectonic events affecting Western Europe. The earliest halokinetic activity in the Norwegian Central Trough occurred during the Triassic and was related to the formation of minibasins in many areas (e.g. the Steinbitt Terrace). As demonstrated in Figure 4.6, there is insufficient data resolution in some areas of the Norwegian Central Trough to discern whether salt movements were occurring there during the Triassic. This is especially true for the deepest parts of the basin where seismic reflections from the Triassic and Jurassic infill are weak and interpretation is therefore ambiguous. Nonetheless, the Triassic constituted a major phase of salt movement, and halokinesis established the pattern of salt highs and intervening lows that persisted (and was typically augmented) thereafter.

Salt movements continued throughout the evolution of the basin. Times of greatest halokinetic activity correspond broadly to those times of maximum tectonic activity. There was a major pulse of salt mobility associated with the Late Jurassic extension: Jurassic strata are strongly rotated and onlapping against salt diapirs, indicative of syn-depositional movements (e.g. Figure 5.9). Development of salt structures continued into the Early Cretaceous, but their growth was more passive in character. Salt diapirs were rejuvenated by Late Cretaceous compression. Rapid Cenozoic subsidence and associated sedimentation tended to outpace the growth of salt diapirs, although a final halokinetic pulse is identified and attributed to Eocene-Miocene compression. During compressional pulses, salt movements accommodated much of the shortening that would otherwise have been taken up through the formation of classic inversion structures. The consequent geometries define a suite of 'salt-influenced inversion structures' that have not previously been characterised in the scientific literature.

Figure 8.2 schematically illustrates the difference between salt-prone and salt-absent basins when inverted. There are consistent and predictable differences between salt-absent areas as opposed to salt prone areas, because the presence of halite within a sedimentary basin influences the character of tectonic inversion that occurs. Where there is no salt, structural styles are deep-seated and localized over the site of a pre-existing structural trend, producing more asymmetric structures (Figure 8.2d). Where salt is present in the basin fill, its impact on inversion depends on the characteristics of the salt prior to inversion. Where salt is present and has been mobilised into a significant salt structure prior to the onset of compression, this salt structure will dominate the structural response to compression (Figure 8.2f): uplift tends to be broad and amplitudes in the post-rift exaggerated. Uplift is associated with the salt highs rather than being localized over the immediate hangingwall to the (inverted) extensional fault.

The manner of salt involvement depends on the salt thickness distribution relative to the pre-inversion (extensional) structural framework. Figure 8.3 illustrates this concept. Salt structures may exist in the immediate hangingwall to a normal fault (e.g. Figure 8.3d), or over the footwall (e.g. Figure 8.3b). Post-inversion geometries will differ as a result; compare Figures 8.3e and 8.3c respectively. As noted in Section 4.3.3, arguably the most common scenario involves salt walls rooted to the footwall. Where salt is present in the hangingwall to a fault, that fault typically exhibits a large throw (e.g. the Skrubbe and Piggvar faults). The relationship between salt structure and the Jurassic rift is complex because salt movements had initiated prior to the main (Late Jurassic) phase of basin formation. The stress regime and sedimentation patterns associated with Late Jurassic extension differed from the regime under which the salt structures initially developed. For example, Figure 5.9 showed evidence of shifting stress regimes during Triassic- Jurassic times in the hangingwall to the Piggvar fault. The variable location of salt thicks relative to the structural framework has resulted in varied inversion geometries relative to non-salt areas. Major salt and fault interactions in the Norwegian Central Trough have been systematically documented, in accordance with the thesis aims.

8.5 Implications of Salt-Influenced Inversion for Chalk Reservoir Quality

Seismic interpretation highlighted significant Chalk thickness variations in the Norwegian Central Trough. Such thickness variation in essentially pelagic sediment is indicative of significant palaeobathymetry, and thus active tectonics, during Chalk deposition. This preponderance of allochthonous chalks relative to other localities of Chalk deposition in Western Europe (e.g. South Dorset) is attributed to the syn-depositional uplifts occurring there. It has been argued that allochthonous chalks have generally superior reservoir properties relative to autochthonous chalks. Thus, it is inferred that Upper Cretaceous inversion was a factor enhancing reservoir quality in the Central North Sea.

Allochthonous movement pathways in the Norwegian Central Trough have been postulated (Figure 6.7). It is desirable to do so, in anticipation that these flows will yield good quality reservoir. However, there are limits to the extent that such movements can be predicted. Allochthonous flows would have a primary relationship with regional dip magnitude, but actual movements would likely be complex, due to secondary dips associated with faults that exist on scales beneath the seismic resolution of this study. Complexity of allochthonous flow is expected to have been exacerbated by the Norwegian Central Trough's multiple fault trends. Furthermore, the data to support such predictions is sparse, with a sampling bias towards the near-crestal locations that have been drilled (and the lithofacies therefore documented).

A potential follow-up to the regional predictions herein requires study of allochthonous chalk on the scale of individual oilfields, tying high quality seismic reflection data to detailed, proprietary, wireline information from the many production wells (in addition to the exploration well data used in this study). A template for such work is the seismic analysis by Esmerode et al. (2008) undertaken in the vicinity of the Adda field in the Danish Central Graben, which revealed unequivocal evidence of slope processes within the Chalk Group.

The fracture styles associated with inversion were investigated, but lack of reliable data from the Norwegian Central Trough precluded detailed interpretation and analysis. Chalk core fracture interpretations are inherently affected (indeed, compromised) by the journey from subsurface to the laboratory. Open fractures in chalk appear to have been particularly vulnerable to disturbance because they weaken the rock and when recovered it often has the appearance of uninterpretable rubble. Even where subsurface fracture interpretations are possible, to build a sufficient database upon which to establish statistically significant (meaningful) relationships would require a substantial, focused effort and was beyond the scope of this study. A possibility for future work is to synthesise existing fracture data, in the form of previously published studies of chalk core and data from reservoir engineering (flow data) provided by ConocoPhillips. This information should be used to evaluate fracture characteristics (style, intensity and orientation of fracturing) in relation to primary structural parameters (e.g. curvature and structural position). And ultimately, to develop a predictive model for sub-seismic resolution fracture characteristics based on observations from seismic reflection data.

Predictions regarding fracture development in response to doming were made in Figure 6.9. From this study it is expected that both inversion and halokinesis affect the lithofacies and fracture characteristics of the reservoir. Both concentric and radial faults are expected to develop due to the strain implications of compression and uplift. Radial faults are likely to cluster at the ends of elliptical salt bodies (i.e. where the curvature of the salt body is greatest). From Norwegian Central Trough examples, radial fracturing is expected to be more pronounced in the northern part of the Ekofisk field than the central and southern parts, and more pronounced in the eastern part of the Tor field rather than the western part. The density of fracturing will depend on the amount of doming that has occurred; the most intense fracture development is expected over the crests of salt diapirs and inversion structures with the greatest amplitude. Figure 5.12's inset showed radial faults developed over a diapir in the Hordaland sediments. The expectation is that layers regionally prone to polygonal faulting (see Figure 4.17a) will have a tendency to organize into radial faults over a diapir.

8.6 Generic Implications of this study

The overall objective of this thesis was to document the manner in which halite within a sedimentary basin can influence the characteristics of inversion. This objective has been successfully achieved, primarily through the detailed interpretation of salt-fault interaction in the Norwegian Central Trough, allowing the underlying mechanisms to be isolated in accordance with the degree permissible given the limitations of the data.

A broad range of structural styles have been documented, from salt absent inversion through salt influenced inversion to 'classic' salt diapirism. It has been shown that structural complexity is greater in inverted basins that have salt structures than those that do not. This observation has fundamental implications for hydrocarbon assessment. In addition, the timing and extent of salt movement prior to inversion was recognised as a major control over the post- inversion geometry.

It is evident that salt dominated the structural styles and response to compression in the Norwegian Central Trough. What is less evident is the degree of separateness of the salt movements relative to inversion tectonics. The interpretations outlines herein suggest that salt movement during basin evolution is generally less discrete than tectonic movement, but tends to occur most markedly when tectonic movements are also greatest. Based on observations in this thesis, I advocate use of the term 'inversion' being expanded for salt prone environments, to include all salt involved structures that demonstrate evidence of compression.

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Appendix 1: 3D seismic interpretation, Central North Sea

1.1. Survey Location

This study made use of two seismic surveys, with locations and line spacings as follows:

CNSmerge survey

Survey coordinates:

Minimum Latitude 56.233100 N

Maximum Latitude 56.812303 N

Minimum Longitude 2.410978 E

Maximum Longitude 3.734936 E

Inlines are oriented South to North and numbered from 16,952 (West) to 23,576 (East)

At 12.5m spacing

Crosslines are oriented West to East and numbered from 5,152 (South) to 10,168 (North)

At 12.5m spacing

Depth Range: 0 to not less than 6.0 seconds two-way travel time (TWTT)

Ga3D93 survey

Survey coordinates:

Minimum Latitude 56.160519 N

Maximum Latitude 56.570425 N

Minimum Longitude 3.491603 E

Maximum Longitude 4.093046 E

Inlines are oriented SW to NE and numbered from 1 (NW) to 1,245 (SE)

At 25m spacing

Crosslines are oriented SE to NW and numbered from 1 (NE) to 3,273 (SW)

At 12.5m spacing

Depth range: 0 to 7.0 seconds two-way travel time (TWTT)

1.2. Well Data

This section details the wells ties used to constrain seismic interpretation. Well data is recorded in metres (Measured Depth) and was calibrated to Two-Way-Time using check shot data. The specific curves used for time-depth conversion are listed by their names as supplied by ConocoPhillips.

The X-Y locations of the 43 well tops are expressed in UTM format. When plotting (sea bottom) well locations onto the seismic surveys, it was necessary to convert from UTM into Latitude and Longitude. Conversion between these different spatial referencing systems was performed using an online calculator:

http://www.uwgb.edu/dutchs/UsefulData/UTMConversions1.xls

The conversion from latitude and longitude to UTM is accurate to within one metre and the reverse conversion is accurate to within 10 metres. This level of accuracy is deemed satisfactory relative to other constraints on resolution affecting the data.

CNSmerge survey: 35 wells used

1/3-5

Near-Vertical exploration well

Location (sea bottom)	Location (TD)
6292054.00 N	6292072.00 N
493528.875 E	493529.00 E

Time-Depth curve: mjs_td_135

Lithostratigraphic tops:

Name	*Depth (m MD)*	*Depth (ms TWT)*
Hordaland Group	1664	1714
Horda Fm	2807	2868
Rogaland Group	3063	3094
Chalk Group	3288	3242
Tor Fm	3384	3324
Hod Fm	3853	3496
Blodoeks Fm	no pick	unknown
Cromer Knoll Group	4604	3831
Humber Group	4739	3903
Triassic (Tr50)	4739	3903
Zechstein Group	4739	3903
Rotliegend Group	4771	3962
TD (Lower Permian)	4850	3980

1/5-2

Vertical exploration well

Location
6270599.00 N
477943.00 E

Time-Depth curve: 0001000_HQGL_CAL

Lithostratigraphic tops:

Name	*Depth (m MD)*	*Depth (ms TWT)*
Hordaland Group	1636	1684
Horda Fm	2646	2685
Rogaland Group	2832	2845
Chalk Group	3069	3015
Tor Fm	3155	3057
Hod Fm	3431	3174
Blodoeks Fm	no pick	unknown
Cromer Knoll Group	3892	3342
Humber Group	4205	3573
J50	4229	3584
Triassic (Tr50)	4229	3584
Zechstein Group	4229	3584
TD (Upper Permian)	4285	3608

1/6-1

Non-Vertical (?) exploration well

Location (sea bottom)	Location (TD)
6276784.08 N	unknown
499827.93 E	unknown

Time-Depth curve: cs_cep_pa_794

Lithostratigraphic tops:

Name	*Depth (m MD)*	*Depth (ms TWT)*
Hordaland Group	1697	1738
Horda Fm	2750	2800
Rogaland Group	2947	2984
Chalk Group	3145	3139
Tor Fm	3247	3195
Hod Fm	3564	3334
Blodoeks Fm	3987	3524
Cromer Knoll Group	4112	3575
TD (Lower Cretaceous)	4822	

1/6-2

Non-Vertical (?) exploration well

Location (sea bottom)	Location (TD)
6269005.92 N	unknown
491682.49 E	unknown

Time-Depth curve: cs_cep_pa_594

Lithostratigraphic tops:

Name	*Depth (m MD)*	*Depth (ms TWT)*
Hordaland Group	1706	1715
Horda Fm	2699	2686
Rogaland Group	2898	2859
Chalk Group	3024	2961
Tor Fm	3090	2993
Hod Fm	3294	3077
TD (Upper Cretaceous)	3383	

1/6-3 T3

Vertical exploration well

Location
6277837.00 N
495728.97 E

Time-Depth curve: 000105_HQ_GL_CAL

Lithostratigraphic tops:

Name	*Depth (m MD)*	*Depth (ms TWT)*
Hordaland Group	1710	1744
Horda Fm	2694	2754
Rogaland Group	2910	2946
Chalk Group	3130	3116
Tor Fm	3225	3178
TD (Upper Cretaceous)	3324	3227

1/6-7 T2

Near-Vertical exploration well

Location (sea bottom)	Location (TD)
6268058.50 N	6268227.00 N
494043.88 E	493782.25 E

Time-Depth curve: td_167_100901

Lithostratigraphic tops:

Name	*Depth (m MD)*	*Depth (ms TWT)*
Hordaland Group	1680	1708
Horda Fm	2850	2821
Rogaland Group	3072	3052
Chalk Group	3278	3204
Tor Fm	3389	3263
Hod Fm	3744	3411
Blodoeks Fm	4152	3582
Cromer Knoll Group	4287	3650
Humber Group	4403	3738
J50	4709	4010
TD (Upper Jurassic)	4995	4166

1/9-2

Non-Vertical (?) exploration well

Location (sea bottom)	Location (TD)
6250516.36 N	unknown
495450.00 E	unknown

Time-Depth curve: cs_cep_ue_594

Lithostratigraphic tops:

Name	*Depth (m MD)*	*Depth (ms TWT)*
Hordaland Group	1508	unknown
Horda Fm	2720	2706
Rogaland Group	2933	2903
Chalk Group	3129	3058
Tor Fm	3195	3090
Hod Fm	3307	3142
TD (Upper Cretaceous)	3459	

1/9-4 T4

Non-Vertical (?) exploration well

Location (sea bottom)	Location (TD)
6260133.08 N	unknown
495898.82 E	unknown

Time-Depth curve: cs_cep_pa_794

Lithostratigraphic tops:

Name	*Depth (m MD)*	*Depth (ms TWT)*
Hordaland Group	1526	1492
Horda Fm	2792	2740
Rogaland Group	2941	2859
Chalk Group	3115	2989
Tor Fm	3210	3041
Hod Fm	3311	3090
Blodoeks Fm	3483	3167
Cromer Knoll Group	3608	3222
Humber Group	3608	3222
Zechstein Group	3689	3266
TD (Upper Permian)	3710	

1/9-5

Non-Vertical (?) appraisal well

Location (sea bottom)	Location (TD)
6260869.22 N	unknown
497859.76 E	unknown

Time-Depth curve: Erico_dataset_VEL

Lithostratigraphic tops:

Name	*Depth (m MD)*	*Depth (ms TWT)*
Hordaland Group	1484	1301
Horda Fm	2814	2627
Rogaland Group	2997	2771
Chalk Group	3208	2916
Tor Fm	3285	2945
TD (Upper Cretaceous)	3412	

2/2-1

Near-Vertical exploration well

Location (sea bottom)	Location (TD)
6294075.50 N	6294093.000 N
538959.38 E	539048.875 E

Time-Depth curve: td_221_240901

Lithostratigraphic tops:

Name	*Depth (m MD)*	*Depth (ms TWT)*
Hordaland Group	no pick	unknown
Horda Fm	2492	2393
Rogaland Group	2658	2530
Chalk Group	2815	2647
Tor Fm	2847	2665
Hod Formation	3128	2796
Blodoeks Fm	no pick	unknown
Cromer Knoll Group	3225	2835
Humber Group	3314	2893
J50	3733	3121
Triassic (Tr50)	3811	3172
Zechstein Group	3947	3243
TD (Upper Permian)	4003	3263

2/4-12

Near-Vertical exploration well

Location (sea bottom)	Location (TD)
6264556.00 N	6264601.00 N
508833.69 E	508836.25 E

Time-Depth curve: td_cep_pa_794

Lithostratigraphic tops:

Name	*Depth (m MD)*	*Depth (ms TWT)*
Hordaland Group	1713	1702
Horda Fm	2826	2794
Rogaland Group	3014	2960
Chalk Group	3195	3100
Tor Fm	3299	3155
TD (Upper Cretaceous)	3383	3194

2/4-16 R

Near-Vertical exploration well

Location (sea bottom)	Location (TD)
6281583.00 N	6281639.00 N
509221.09 E	509193.03 E

Time-Depth curve: Erico_dataset_CVL_1

Lithostratigraphic tops:

Name	*Depth (m MD)*	*Depth (ms TWT)*
Hordaland Group	1675	1698
Horda Fm	2881	2892
Rogaland Group	3146	3120
Chalk Group	3318	3243
Tor Fm	3435	3293
Hod Fm	3890	3480
Blodoeks Fm	4594	3768
Cromer Knoll Group	4638	3789
Humber Group	4738	3853
J50	4869	3938
TD (Jurassic)	4996	

2/4-17 T3

Non-Vertical exploration well

Location (sea bottom)	Location (TD)
6282380.93 N	unknown
514043.23 E	unknown

Time-Depth curve: CS_AF_edited_051102

Lithostratigraphic tops:

Name	*Depth (m MD)*	*Depth (ms TWT)*
Hordaland Group	1721	1706
Oligocene	2143	2118
Horda Fm	2826	2801
Rogaland Group	3050	3001
Chalk Group	3193	3099
Tor Fm	no pick	unknown
Hod Fm	3729	3320
Blodoeks Fm	no pick	unknown
Cromer Knoll Group	4183	3520
Humber Group	4190	3525
J50	4238	3584
Triassic (Tr50)	4446	3675
Zechstein Group	4486	3696
Rotliegend Group	4520	3755
TD (Lower Permian)	5258	4088

2/4-19 B

Non-Vertical (?) exploration well

Location (sea bottom)	Location (TD)
6269153.66 N	unknown
512526.87 E	unknown

Time-Depth curve: td_2419b_240701

Lithostratigraphic tops:

Name	*Depth (m MD)*	*Depth (ms TWT)*
Hordaland Group	1600	1679
Horda Fm	2652	2725
Rogaland Group	2833	2870
Chalk Group	2981	3002
Tor Fm	no pick	unknown
Hod Fm	no pick	unknown
Blodoeks Fm	no pick	unknown
Cromer Knoll Group	3870	3416
Humber Group	4073	3560
TD (Jurassic)	4605	3969

2/4-2

Near-Vertical exploration well

Location (sea bottom)	Location (TD)
6265865.00 N	6265920.50 N
512208.19 E	512234.38 E

Time-Depth curve: TD_SNSMerge_AF_101103

Lithostratigraphic tops:

Name	*Depth (m MD)*	*Depth (ms TWT)*
Hordaland Group	1553	1587
Horda Fm	2643	2690
Rogaland Group	2904	2930
Chalk Group	3042	3042
Tor Fm	3197	3138
TD (Upper Cretaceous)	3306	3198

2/4-3

Near-Vertical exploration well

Location (sea bottom)	Location (TD)
6266910.00 N	6267196.50 N
515295.69 E	515307.63 E

Time-Depth curve: TD_SNSMerge_AF_211103

Lithostratigraphic tops:

Name	*Depth (m MD)*	*Depth (ms TWT)*
Hordaland Group	1668	1686
Horda Fm	2702	2703
Rogaland Group	2959	2918
Chalk Group	3104	3030
Tor Fm	3247	3138
TD (Upper Cretaceous)	3431	3238

2/4-4

Near-Vertical exploration well

Location (sea bottom)	Location (TD)
6263380.50 N	6263383.00 N
513051.72 E	513069.25 E

Time-Depth curve: TD_SNSMerge_AF_101103

Lithostratigraphic tops:

Name	*Depth (m MD)*	*Depth (ms TWT)*
Hordaland Group	1693	1723
Horda Fm	2789	2806
Rogaland Group	2977	2954
Chalk Group	3122	3062
Tor Fm	3253	3137
TD (Upper Cretaceous)	3425	3221

2/4-6

Non-Vertical (?) exploration well

Location (sea bottom)	Location (TD)
6269306.42 N	unknown
505261.16 E	unknown

Time-Depth curve: Erico_dataset_VEL

Lithostratigraphic tops:

Name	*Depth (m MD)*	*Depth (ms TWT)*
Hordaland Group	1705	1725
Horda Fm	2755	2774
Rogaland Group	2957	2959
Chalk Group	3117	3085
TD (Upper Cretaceous)	3412	

2/4-7

Near-Vertical exploration well

Location (sea bottom)	Location (TD)
6276913.50 N	6276949.50 N
517123.72 E	517123.72 E

Time-Depth curve: 000059_HQGL_CAL

Lithostratigraphic tops:

Name	*Depth (m MD)*	*Depth (ms TWT)*
Hordaland Group	1697	1712
Horda Fm	2832	2829
Rogaland Group	3049	3018
Chalk Group	3175	3110
Tor Fm	3304	3173
TD (Upper Cretaceous)	3494	3265

2/4-9

Near-Vertical exploration well

Location (sea bottom)	Location (TD)
6273826.50 N	6273866.00 N
505260.72 E	505253.09 E

Time-Depth curve: TD_SNSMerge_AF_241103

Lithostratigraphic tops:

Name	*Depth (m MD)*	*Depth (ms TWT)*
Hordaland Group	1696	1722
Horda Fm	2805	2838
Rogaland Group	3022	3027
Chalk Group	3215	3173
Tor Fm	3299	3214
Hod Fm	3683	3378
TD (Upper Cretaceous)	3752	3408

2/5-1

Non-Vertical (?) exploration well

Location (sea bottom)	Location (TD)
6277377.68 N	unknown
520576.09 E	unknown

Time-Depth curve: td_cep_nj_594

Lithostratigraphic tops:

Name	*Depth (m MD)*	*Depth (ms TWT)*
Hordaland Group	1676	1701
Horda Fm	2710	2731
Rogaland Group	2918	2928
Chalk Group	3041	3028
Tor Fm	no pick	unknown
Hod Fm	no pick	unknown
Blodoeks Fm	no pick	unknown
Cromer Knoll Group	3666	3291
Humber Group	3864	3363
TD (Jurassic)	3972	3429

2/5-10 A

Non-Vertical exploration well

Location (sea bottom)	Location (TD)
6283054.46 N	6282902.00 N
530071.86 E	530005.69 E

Time-Depth curve: td_2510_190901

Lithostratigraphic tops:

Name	*Depth (m MD)*	*Depth (ms TWT)*
Hordaland Group	1665	1711
Horda Fm	no pick	unknown
Rogaland Group	3053	3034
Chalk Group	3215	3153
Tor Fm	3342	3218
Hod Fm	3735	3384
Blodoeks Fm	no pick	unknown
Cromer Knoll Group	4252	3594
Humber Group	4299	3617
J50	4482	3750
Triassic (Tr50)	4667	3852
TD (Triassic)	4715	3878

2/5-6 T2

Non-Vertical (?) exploration well

Location (sea bottom)	Location (TD)
6269872.15 N	unknown
538176.28 E	unknown

Time-Depth curve: td_cep_pa_794

Lithostratigraphic tops:

Name	*Depth (m MD)*	*Depth (ms TWT)*
Hordaland Group	1601	1610
Horda Fm	2481	2488
Rogaland Group	2908	2890
Chalk Group	3048	2966
Tor Fm	no pick	unknown
Hod Fm	no pick	unknown
Blodoeks Fm	no pick	unknown
Cromer Knoll Group	3587	3212
Humber Group	3596	3218
J50	3723	3336
Triassic (Tr50)	4087	3606
TD (Triassic)	4132	3636

2/5-7

Non-Vertical (?) exploration well

Location (sea bottom)	Location (TD)
6286563.43 N	unknown
522459.46 E	unknown

Time-Depth curve: Ed_cksts_257_210901

Lithostratigraphic tops:

Name	*Depth (m MD)*	*Depth (ms TWT)*
Hordaland Group	1723	1754
Horda Fm	no pick	unknown
Rogaland Group	3028	3011
Chalk Group	3204	3140
Tor Fm	no pick	unknown
Hod Fm	no pick	unknown
Blodoeks Fm	no pick	unknown
Cromer Knoll Group	4077	3510
Humber Group	4108	3526
J50	4292	3664
Triassic (Tr50)	4403	3760
TD (Triassic)	4531	3815

2/7-1

Non-Vertical (?) exploration well

Location (sea bottom)	Location (TD)
6253994.297407 N	unknown
512579.236713 E	unknown

Time-Depth curve: Erico_dataset_VEL

Lithostratigraphic tops:

Name	*Depth (m MD)*	*Depth (ms TWT)*
Hordaland Group	1533	1622
Horda Fm	2622	2615
Rogaland Group	2822	2799
Chalk Group	2935	2898
Tor Fm	2997	2937
Hod Fm	3031	2957
Blodoeks Fm	3293	3094
Cromer Knoll Group	3419	3159
Humber Group	3702	3372
TD (Jurassic)	4573	

2/7-10

Near-Vertical appraisal well, publicly released through the Norwegian Petroleum Directorate

Location (sea bottom)	Location (TD)
6259202.23 N	unknown
505183.86 E	unknown

Time-Depth curve: Erico_dataset_VEL

Lithostratigraphic tops:

Name	*Depth (m MD)*	*Depth (ms TWT)*
Hordaland Group	1774	1648
Horda Fm	2807	2718
Rogaland Group	3033	2920
Chalk Group	3192	3048
Tor Fm	3282	3096
Hod Fm	3370	3142
TD (Upper Cretaceous)	4701	

2/7-16

Vertical exploration well

Location
6253267.50 N
505857.94 E

Time-Depth curve: TD_SNSMerge_AF_211103

Lithostratigraphic tops:

Name	*Depth (m MD)*	*Depth (ms TWT)*
Hordaland Group	1648	1633
Horda Fm	2773	2749
Rogaland Group	2993	2958
Chalk Group	3182	3114
Tor Fm	3264	3156
Hod Fm	3367	3203
Blodoeks Fm	3968	3480
Cromer Knoll Group	4158	3583
TD (Lower Cretaceous)	4818	4076

2/7-19 R

Near-Vertical exploration well

Location (sea bottom)	Location (TD)
6243890.00 N	6243845.00 N
506410.09 E	506448.66 E

Time-Depth curve: TD_SNSMerge_AF_070305_AF

Lithostratigraphic tops:

Name	*Depth (m MD)*	*Depth (ms TWT)*
Hordaland Group	1530	1526
Horda Fm	2697	2690
Rogaland Group	2989	2944
Chalk Group	3131	3054
Tor Fm	3183	3078
Hod Fm	3397	3171
Blodoeks Fm	4066	3443
Cromer Knoll Group	4211	3508
Humber Group	4584	3747
J50	4698	3835
Triassic (Tr50)	4841	3907
Zechstein Group	4870	3928
Rotliegendes Group	4870	3928
TD (Permian)	4876	3932

2/7-2

Non-Vertical exploration well

Location (sea bottom)	Location (TD)
6235931.56 N	unknown
509440.65 E	unknown

Time-Depth curve: TD_SNSMerge_AF_241103

Lithostratigraphic tops:

Name	*Depth (m MD)*	*Depth (ms TWT)*
Hordaland Group	1521	1522
Horda Fm	2520	2500
Rogaland Group	2856	2793
Chalk Group	2965	2881
Tor Fm	no pick	unknown
Hod Fm	no pick	unknown
Blodoeks Fm	no pick	unknown
Cromer Knoll Group	3748	3242
Humber Group	3861	3285
TD (Jurassic)	3875	3295

2/7-24

Non-Vertical exploration well

Location (sea bottom)	Location (TD)
6240676.00 N	6240433.00 N
519997.41 E	519851.38 E

Time-Depth curve: TD_SNSMerge_AF_081103

Lithostratigraphic tops:

Name	*Depth (m MD)*	*Depth (ms TWT)*
Hordaland Group	1542	1540
Horda Fm	2383	2388
Rogaland Group	2674	2654
Chalk Group	2735	2706
Tor Fm	2744	2711
Hod Fm	2744	2711
Blodoeks Fm	2966	2842
Cromer Knoll Group	2994	2859
Humber Group	3195	3012
J50	3928	3649
TD (Upper Jurassic)	5023	4341

2/7-28 T3

Non-Vertical exploration well

Location (sea bottom)	Location (TD)
6248667.00 N	6248649.50 N
514740.53 E	514685.44 E

Time-Depth curve: TD_SNSMerge_090305_AF

Lithostratigraphic tops:

Name	*Depth (m MD)*	*Depth (ms TWT)*
Hordaland Group	1584	1649
Horda Fm	2514	2519
Rogaland Group	2729	2717
Chalk Group	2850	2820
Tor Fm	2898	2854
Hod Fm	2902	2855
Blodoeks Fm	3091	2961
Cromer Knoll Group	3169	3002
Humber Group	3339	3113
J50	3797	3439
Triassic (Tr50)	3814	3447
Zechstein Group	3839	3459
TD (Upper Permian)	3892	3482

2/7-3

Near-Vertical exploration well

Location (sea bottom)	Location (TD)
6248997.00 N	6248880.50 N
515195.63 E	515117.84 E

Time-Depth curve: TD_SNSMerge_AF_241103

Lithostratigraphic tops:

Name	*Depth (m MD)*	*Depth (ms TWT)*
Hordaland Group	1584	1575
Horda Fm	2506	2532
Rogaland Group	2704	2708
Chalk Group	2804	2805
Tor Fm	2878	2869
Hod Fm	2878	2869
Blodoeks Fm	3158	3034
Cromer Knoll Group	3240	3078
Humber Group	3342	3152
J50	3911	3572
Triassic (Tr50)	4177	3752
Zechstein Group	4188	3758
TD (Upper Permian)	4358	3836

2/7-9

Near-Vertical exploration well

Location (sea bottom)	Location (TD)
6243054.00 N	6243035.50 N
514931.94 E	514988.91 E

Time-Depth curve: TD_SNSMerge_AF_241103

Lithostratigraphic tops:

Name	*Depth (m MD)*	*Depth (ms TWT)*
Hordaland Group	1573	1558
Horda Fm	2607	2600
Rogaland Group	2949	2877
Chalk Group	3086	2985
Tor Fm	3163	3024
Hod Fm	3278	3073
Blodoeks Fm	3890	3338
Cromer Knoll Group	3962	3369
Humber Group	4193	3486
J50	4196	3488
Triassic (Tr50)	4203	3491
Zechstein Group	4203	3491
Rotliegend Group	4203	3491
TD (Lower Permian)	4448	3611

2/8-14 T2

Near-Vertical exploration well

Location (sea bottom)	Location (TD)
6235612.50 N	6235625.50 N
522078.63 E	522117.00 E

Time-Depth curve: TD_SNSMerge_110305_AF

Lithostratigraphic tops:

Name	*Depth (m MD)*	*Depth (ms TWT)*
Hordaland Group	no pick	unknown
Horda Fm	no pick	unknown
Rogaland Group	2570	2575
Chalk Group	2614	2614
Tor Fm	2617	2616
Hod Fm	no pick	unknown
Blodoeks Fm	no pick	unknown
Cromer Knoll Group	2874	2778
Humber Group	3196	3036
TD (Jurassic)	4392	4038

2/8-2

Vertical exploration well

Location
6261750.20 N
529305.51 E

Time-Depth curve: TD_SNSMerge_AF_211103

Lithostratigraphic tops:

Name	*Depth (m MD)*	*Depth (ms TWT)*
Hordaland Group	no pick	unknown
Horda Fm	no pick	unknown
Rogaland Group	2824	2802
Chalk Group	2947	2903
Tor Fm	3042	2954
Hod Fm	no pick	unknown
Blodoeks Fm	no pick	unknown
Cromer Knoll Group	3227	3042
TD (Lower Cretaceous)	3246	3057

Ga3D93 survey: 8 wells used

2/6-2 T2

Near-Vertical exploration well

Location (sea bottom)	Location (TD)
6263607.50 N	6263589.00 N
543757.688 E	543774.438 E

Time-Depth curve: 000238_HQGL_CAL

Lithostratigraphic tops:

Name	*Depth (m MD)*	*Depth (ms TWT)*
Hordaland Group	no pick	unknown
Horda Fm	no pick	unknown
Rogaland Group	3018	2959
Chalk Group	3160	3063
Tor Fm	3252	3108
Hod Fm	3598	3261
Blodoeks Fm	no pick	unknown
Cromer Knoll Group	3812	3350
Humber Group	4118	3527
J50	4518	3798
Triassic (Tr50)	4590	3840
Zechstein Group	4723	3904
TD (Upper Permian)	4760	3918

2/6-3

Near-Vertical exploration well

Location (sea bottom)	Location (TD)
6265551.50 N	6265604.00 N
550665.812 E	550701.312 E

Time-Depth curve: 000352_HQGL_CAL

Lithostratigraphic tops:

Name	*Depth (m MD)*	*Depth (ms TWT)*
Hordaland Group	no pick	unknown
Horda Fm	no pick	unknown
Rogaland Group	2817	2774
Chalk Group	2945	2870
Tor Fm	3021	2910
Hod Fm	3271	3034
Blodoeks Fm	no pick	unknown
Cromer Knoll Group	3391	3088
Humber Group	3391	3088
J50	3499	3164
Triassic (Tr50)	3639	3249
Zechstein Group	3729	3297
Rotliegend Group	3791	3318
TD (Lower Permian)	4060	3425

2/7-15

Vertical exploration well

Location
6250374.50 N
519450.43 E

Time-Depth curve: TD_SNSMerge_AF_241103

Lithostratigraphic tops:

Name	*Depth (m MD)*	*Depth (ms TWT)*
Hordaland Group	16231	1627
Horda Fm	2662	2655
Rogaland Group	2899	2875
Chalk Group	3005	2962
Tor Fm	no pick	unknown
Hod Fm	no pick	unknown
Blodoeks Fm	no pick	unknown
Cromer Knoll Group	3395	3167
Humber Group	3584	3298
J50	4191	3805
TD (Jurassic)	4423	4011

2/8-12 S

Non-Vertical exploration well

Location (sea bottom)	Location (TD)
6250476.50 N	6249940.00 N
527013.69 E	526443.06 E

Time-Depth curve: td_cep_295_to

Lithostratigraphic tops:

Name	*Depth (m MD)*	*Depth (ms TWT)*
Hordaland Group	no pick	unknown
Horda Fm	no pick	unknown
Rogaland Group	2874	2754
Chalk Group	3045	2862
Tor Fm	3116	2897
Hod Fm	no pick	unknown
Blodoeks Fm	no pick	unknown
Cromer Knoll Group	3257	2973
Humber Group	4000	3402
J50	4548	3826
Triassic (Tr50)	5238	4262
TD (Triassic)	5289	4286

2/8-3

Vertical exploration well

Location
6240656.50 N
527741.25 E

Time-Depth curve: 000072_HQGL_CAL

Lithostratigraphic tops:

Name	*Depth (m MD)*	*Depth (ms TWT)*
Hordaland Group	no pick	unknown
Horda Fm	no pick	unknown
Rogaland Group	2728	2734
Chalk Group	2780	2782
Tor Fm	2808	2795
Hod Fm	no pick	unknown
Blodoeks Fm	no pick	unknown
Cromer Knoll Group	3200	2984
Humber Group	3537	3220
J50	4092	3650
TD (Jurassic)	4115	3665

2/9-1

Vertical exploration well

Location
6247594.50 N
541178.00 E

Time-Depth curve: 000079_HQGL_CAL

Lithostratigraphic tops:

Name	*Depth (m MD)*	*Depth (ms TWT)*
Hordaland Group	no pick	unknown
Horda Fm	no pick	unknown
Rogaland Group	2963	2920
Chalk Group	3092	3014
Tor Fm	3205	3070
Hod Fm	3482	3195
Blodoeks Fm	no pick	unknown
Cromer Knoll Group	3551	3342
TD (Lower Cretaceous)	3552	

2/9-2

Vertical exploration well

Location
6245460.00 N
557711.94 E

Time-Depth curve: TD 2/9-2 ag 140605

Lithostratigraphic tops:

Name	*Depth (m MD)*	*Depth (ms TWT)*
Hordaland Group	no pick	unknown
Horda Fm	no pick	unknown
Rogaland Group	2930	2900
Chalk Group	3036	2984
Tor Fm	3126	3026
Hod Fm	3411	3152
Blodoeks Fm	no pick	unknown
Cromer Knoll Group	3471	3178
Humber Group	3650	3308
J50	4304	3774
Triassic (Tr50)	4316	3781
Zechstein Group	4325	3784
Rotliegend Group	4325	3784
TD (Lower Permian)	4367	3806

2/9-3

Non-Vertical exploration well

Location (sea bottom)	Location (TD)
6254838.50 N	6254748.00 N
549162.438 E	549164.50 E

Time-Depth curve: TD 2/9-3 ag 140605

Lithostratigraphic tops:

Name	*Depth (m MD)*	*Depth (ms TWT)*
Hordaland Group	no pick	unknown
Horda Fm	no pick	unknown
Rogaland Group	3030	2972
Chalk Group	3168	3076
Tor Fm	3254	3122
Hod Fm	3596	3273
Blodoeks Fm	no pick	unknown
Cromer Knoll Group	3834	3375
Humber Group	3888	3408
J50	4541	3872
Triassic (Tr50)	4600	3882
Zechstein Group	4670	3941
Rotliegend Group	4670	3941
TD (Lower Permian)	4859	4025

Appendix 2: Fieldwork Observations, Eastern Uinta Mountains

Field observation of Weber sandstone outcrops were made at a total of 188 different localities during the 2005 field season. This table summarises the principal recorded measurements; orientation data for bedding, joints, faults and deformation bands (db). The latitudes and longitudes of field localities were initially recorded using a hand-held GPS device while in the field. The accuracy of these positions was subsequently verified using an internet tool for deriving spatial coordinates from satellite map images (**http://itouchmap.com/latlong.html**).

The many field localities were grouped into a smaller and more manageable number of 'fracture' localities. This was considered acceptable means to improve the display of results, provided grouped field localities share a common structural position (i.e. bed orientation), and are in close proximity (typically not much more than tens of metres apart). A stereonet for each fracture locality is shown in figure 7.9

Fracture Locality	Field Locality	Latitude Deg.	Latitude Min.	Latitude Sec.	Longitude Deg.	Longitude Min.	Longitude Sec.	Strike	Dip	Dip Direction	Style
B1	23	40	26	50.7	-109	15	35.1	**276**	**60**	N	joint
B1	31	40	27	0.5	-109	16	5.1	**296**	**55**	N	joint
B1	31	40	27	0.5	-109	16	5.1	**272**	**46**	N	joint
B1	41	40	26	52.7	-109	15	39.9	**287**	**54**	N	joint
B1	41	40	26	52.7	-109	15	39.9	**279**	**60**	N	joint
B1	41	40	26	52.7	-109	15	39.9	**254**	**38**	N	joint
B1	41	40	26	52.7	-109	15	39.9	**270**	**57**	N	joint
B1	41	40	26	52.7	-109	15	39.9	**282**	**62**	N	joint
B1	41	40	26	52.7	-109	15	39.9	**076**	**38**	S	bedding
n=9											
B2	17	40	26	43.1	-109	15	15.5	**283**	**49**	N	joint
B2	17	40	26	43.1	-109	15	15.5	**284**	**74**	N	joint
B2	17	40	26	43.1	-109	15	15.5	**280**	**64**	N	joint
B2	18	40	26	43.5	-109	15	15.8	**190**	**56**	N	db
B2	18	40	26	43.5	-109	15	15.8	**147**	**56**	W	db
B2	18	40	26	43.5	-109	15	15.8	**271**	**44**	N	db
B2	18	40	26	43.5	-109	15	15.8	**280**	**62**	N	db
B2	19	40	26	43.9	-109	15	15.5	**284**	**68**	N	db
B2	20	40	26	44.1	-109	15	15.9	**100**	**52**	S	bedding
B2	22	40	26	44.7	-109	15	15.9	**158**	**29**	W	joint?
n=10											
B3	6	40	26	42.7	-109	15	14	**274**	**68**	N	joint
B3	6	40	26	42.7	-109	15	14	**150**	**12**	S	joint?
B3	6	40	26	42.7	-109	15	14	**270**	**50**	N	joint
B3	6	40	26	42.7	-109	15	14	**282**	**50**	N	joint

B3	6	40	26	42.7	-109	15	14	**174**	**25**	W	joint
B3	6	40	26	42.7	-109	15	14	**292**	**53**	N	joint
B3	6	40	26	42.7	-109	15	14	**270**	**55**	N	joint
B3	6	40	26	42.7	-109	15	14	**291**	**38**	N	joint
B3	6	40	26	42.7	-109	15	14	**067**	**24**	S	bedding
B3	7	40	26	42.9	-109	15	14.3	**273**	**59**	N	joint
B3	11	40	26	43.7	-109	15	13.1	**264**	**54**	N	joint
B3	11	40	26	43.7	-109	15	13.1	**280**	**55**	N	joint
B3	11	40	26	43.7	-109	15	13.1	**279**	**50**	N	fault
B3	11	40	26	43.7	-109	15	13.1	**276**	**50**	N	fault
n=14											
B4	101	40	26	41.6	-109	12	59.3	**316**	**47**	N	joint
B4	101	40	26	41.6	-109	12	59.3	**265**	**61**	N	joint
B4	101	40	26	41.6	-109	12	59.3	**290**	**34**	N	joint
B4	101	40	26	41.6	-109	12	59.3	**306**	**73**	N	joint
B4	101	40	26	41.6	-109	12	59.3	**270**	**28**	S	joint
B4	101	40	26	41.6	-109	12	59.3	**316**	**31**	N	joint
B4	101	40	26	41.6	-109	12	59.3	**114**	**50**	S	bedding
B4	102	40	26	41.6	-109	12	59.3	**265**	**51**	N	db
B4	102	40	26	41.6	-109	12	59.3	**282**	**70**	N	db
B4	102	40	26	41.6	-109	12	59.3	**277**	**77**	N	db
B4	104	40	26	40.7	-109	12	42.3	**300**	**69**	N	joint
B4	104	40	26	40.7	-109	12	42.3	**280**	**67**	N	db
B4	104	40	26	40.7	-109	12	42.3	**292**	**70**	N	db
B4	104	40	26	40.7	-109	12	42.3	**297**	**56**	N	joint
B4	104	40	26	40.7	-109	12	42.3	**302**	**46**	N	joint
B4	104	40	26	40.7	-109	12	42.3	**288**	**70**	N	db cluster (fault)*
B4	104	40	26	40.7	-109	12	42.3	**117**	**27**	S	bedding
B4	105	40	26	41.6	-109	12	40.3	**284**	**78**	N	joint?
n=18											
B5	182	40	21	0.1	-109	9	28	**046**	**66**	S	joint
B5	183	40	21	2.5	-109	9	18.9	**050**	**69**	E	fault
B5	183	40	21	2.5	-109	9	18.9	**211**	**11**	W	db
B5	183	40	21	2.5	-109	9	18.9	**037**	**61**	E	db cluster (fault)
B5	184	40	21	26.2	-109	8	33.4	**056**	**68**	S	joint
B5	184	40	21	26.2	-109	8	33.4	**058**	**63**	S	joint
B5	184	40	21	26.2	-109	8	33.4	**138**	**25**	S	bedding
n=7											
B6	114	40	19	59.5	-109	9	53.6	**254**	**58**	N	db
B6	114	40	19	59.5	-109	9	53.6	**300**	**47**	N	db
B6	114	40	19	59.5	-109	9	53.6	**306**	**36**	N	db
B6	114	40	19	59.5	-109	9	53.6	**025**	**51**	E	db
B6	116	40	20	13	-109	10	12.2	**037**	**36**	E	joint?
B6	118	40	20	12.7	-109	10	8.2	**079**	**59**	S	db
B6	118	40	20	12.7	-109	10	8.2	**082**	**53**	S	db
B6	119	40	20	14.4	-109	9	58.6	**243**	**53**	N	db cluster
B6	119	40	20	14.4	-109	9	58.6	**188**	**52**	N	db
B6	119	40	20	14.4	-109	9	58.6	**265**	**48**	N	db cluster
B6	119	40	20	14.4	-109	9	58.6	**298**	**48**	N	db
B6	119	40	20	14.4	-109	9	58.6	**226**	**42**	N	db
B6	119	40	20	14.4	-109	9	58.6	**244**	**26**	N	db
B6	119	40	20	14.4	-109	9	58.6	**332**	**35**	N	db

B6	119	40	20	14.4	-109	9	58.6	**248**	**48**	N	db
B6	119	40	20	14.4	-109	9	58.6	**064**	**62**	S	bedding
B6	120	40	20	15.3	-109	9	57.1	**312**	**18**	N	db
B6	121	40	20	18.9	-109	9	51.2	**337**	**27**	N	db
B6	121	40	20	18.9	-109	9	51.2	**012**	**25**	N	db
B6	121	40	20	18.9	-109	9	51.2	**282**	**40**	N	db
B6	121	40	20	18.9	-109	9	51.2	**308**	**22**	N	db
B6	121	40	20	18.9	-109	9	51.2	**348**	**24**	N	db
B6	121	40	20	18.9	-109	9	51.2	**328**	**32**	N	db
n=23											
C1	188	40	22	54.3	-109	1	52.8	**300**	**90**		joint
C1	44	40	23	13.3	-109	0	0.1	**268**	**58**	N	joint
C1	44	40	23	13.3	-109	0	0.1	**180**	**62**	W	joint
C1	44	40	23	13.3	-109	0	0.1	**141**	**85**	E	joint
C1	44	40	23	13.3	-109	0	0.1	**236**	**56**	N	joint
C1	44	40	23	13.3	-109	0	0.1	**024**	**30**	E	joint
C1	44	40	23	13.3	-109	0	0.1	**092**	**14**	S	bedding
n=7											
D1	107	40	30	48.5	-109	7	17.5	**320**	**61**	N	joint
D1	108	40	30	46.9	-109	6	59.7	**192**	**40**	N	joint
D1	108	40	30	46.9	-109	6	59.7	**182**	**36**	W	joint
D1	109	40	30	47.9	-109	6	58.2	**320**	**53**	N	joint
D1	110	40	30	47.9	-109	6	58.2	**185**	**35**	W	bedding
D1	112	40	30	48.5	-109	6	56.3	**319**	**55**	N	joint
D1	112	40	30	48.5	-109	6	56.3	**343**	**48**	N	joint
D1	112	40	30	48.5	-109	6	56.3	**312**	**60**	N	joint
n=8											
D2	141	40	29	59.2	-109	2	20.1	**038**	**78**	W	db
D2	141	40	29	59.2	-109	2	20.1	**024**	**70**	W	db cluster
D2	141	40	29	59.2	-109	2	20.1	**270**	**45**	N	db
D2	141	40	29	59.2	-109	2	20.1	**106**	**78**	S	db
D2	141	40	29	59.2	-109	2	20.1	**227**	**19**	N	db
D2	141	40	29	59.2	-109	2	20.1	**170**	**32**	W	db
D2	141	40	29	59.2	-109	2	20.1	**091**	**7**	S	db
D2	143	40	29	55	-109	2	24.9	**299**	**69**	N	joint
D2	143	40	29	55	-109	2	24.9	**290**	**81**	N	joint
D2	141	40	29	59.2	-109	2	20.1	**020**	**84**	E	bedding
n=10											
D3	139	40	30	26.2	-109	1	30.8	**180**	**54**	W	db
D3	139	40	30	26.2	-109	1	30.8	**186**	**72**	W	db
D3	139	40	30	26.2	-109	1	30.8	**093**	**42**	W	db
D3	139	40	30	26.2	-109	1	30.8	**210**	**85**	W	db
D3	139	40	30	26.2	-109	1	30.8	**062**	**58**	W	db
D3	139	40	30	26.2	-109	1	30.8	**135**	**47**	W	db
D3	139	40	30	26.2	-109	1	30.8	**110**	**46**	W	db
D3	139	40	30	26.2	-109	1	30.8	**087**	**48**	W	db
D3	139	40	30	26.2	-109	1	30.8	**086**	**32**	S	db
D3	139	40	30	26.2	-109	1	30.8	**082**	**46**	S	db
D3	139	40	30	26.2	-109	1	30.8	**075**	**64**	W	db
D3	139	40	30	26.2	-109	1	30.8	**284**	**41**	N	joint

D3	139	40	30	26.2	-109	1	30.8	**022**	**82**	S	bedding
n=13											
D4	151	40	29	29	-109	2	57.8	**042**	**68**	E	db
D4	148	40	29	26.8	-109	3	13.8	**291**	**59**	N	bedding
n=2											
D5	95	40	30	26.2	-109	1	30.8	**262**	**88**	N	joint
D5	95	40	30	26.2	-109	1	30.8	**312**	**90**	N	joint
D5	95	40	30	26.2	-109	1	30.8	**257**	**78**	N	joint
D5	95	40	30	26.2	-109	1	30.8	**325**	**75**	N	joint
D5	95	40	30	26.2	-109	1	30.8	**295**	**74**	N	joint
D5	95	40	30	26.2	-109	1	30.8	**130**	**23**	S	bedding
n=6											

E1	124	40	32	13.2	-108	59	47.1	**039**	**88**	E	fault
E1	129	40	32	7.3	-109	59	51.9	**261**	**76**	N	db
E1	129	40	32	7.3	-109	59	51.9	**253**	**56**	N	db
E1	129	40	32	7.3	-109	59	51.9	**293**	**70**	N	db
E1	129	40	32	7.3	-109	59	51.9	**220**	**90**		joint
E1	133	40	32	7.3	-109	59	51.9	**195**	**60**	W	db
E1	129	40	32	7.3	-109	59	51.9	**029**	**72**	E	bedding
n=7											
E2	16	40	32	3.2	-109	0	31.2	**005**	**90**		db
E2	16	40	32	3.2	-109	0	31.2	**240**	**85**	N	db
E2	16	40	32	3.2	-109	0	31.2	**325**	**90**		db
E2	16	40	32	3.2	-109	0	31.2	**085**	**10**	S	bedding
n=4											
E3	67	40	31	53.4	-108	59	41.9	**100**	**24**	W	db
E3	67	40	31	53.4	-108	59	41.9	**350**	**20**	E	db
E3	67	40	31	53.4	-108	59	41.9	**180**	**22**	W	db
E3	68	40	31	55.8	-108	59	43	**132**	**15**	S	db
E3	68	40	31	55.8	-108	59	43	**070**	**56**	W	db cluster
E3	69	40	31	54.6	-108	59	48.5	**035**	**60**	E	fault
E3	69	40	31	54.6	-108	59	48.5	**047**	**58**	E	bedding
E3	70	40	31	56.6	-108	59	51.9	**077**	**47**	S	db
E3	71	40	31	57.5	-108	59	53.2	**046**	**57**	E	db
n=9											
E4	63	40	31	33.1	-108	59	41.4	**253**	**72**	N	db cluster
E4	63	40	31	33.1	-108	59	41.4	**242**	**69**	N	db
E4	63	40	31	33.1	-108	59	41.4	**076**	**74**	S	db
E4	63	40	31	33.1	-108	59	41.4	**042**	**69**	S	db
E4	63	40	31	33.1	-108	59	41.4	**250**	**68**	N	db
E4	63	40	31	33.1	-108	59	41.4	**231**	**75**	N	db
E4	63	40	31	33.1	-108	59	41.4	**135**	**30**	W	joint
E4	64	40	31	36	-108	59	41.4	**253**	**66**	N	db
E4	64	40	31	36	-108	59	41.4	**037**	**77**	S	db
E4	65	40	31	37.8	-108	59	41.2	**008**	**89**	W	joint
E4	65	40	31	37.8	-108	59	41.2	**138**	**9**	S	bedding
n=11											
E5	162	40	26	31.6	-108	53	18	**085**	**58**	S	joint
E5	162	40	26	31.6	-108	53	18	**087**	**54**	S	joint
E5	162	40	26	31.6	-108	53	18	**087**	**46**	S	joint

E5	162	40	26	31.6	-108	53	18	**264**	**30**	N	bedding
E5	163	40	26	28.2	-108	53	17.8	**071**	**32**	S	db
E5	163	40	26	28.2	-108	53	17.8	**068**	**33**	S	db cluster (fault)
E5	164	40	26	28.2	-108	53	17.8	**046**	**39**	S	fault
E5	164	40	26	28.2	-108	53	17.8	**278**	**39**	N	bedding
E5	168	40	26	24.8	-108	53	16.5	**030**	**58**	E	joint
n=9											
E6	180	40	22	51.2	-108	45	2.2	**129**	**30**	W	db
E6	180	40	22	51.2	-108	45	2.2	**127**	**32**	S	db
E6	180	40	22	51.2	-108	45	2.2	**105**	**64**	S	db cluster
E6	180	40	22	51.2	-108	45	2.2	**138**	**35**	S	db
E6	180	40	22	51.2	-108	45	2.2	**140**	**30**	S	db
E6	180	40	22	51.2	-108	45	2.2	**91**	**44**	S	db
E6	180	40	22	51.2	-108	45	2.2	**100**	**46**	S	db
E6	180	40	22	51.2	-108	45	2.2	**95**	**6**	S	bedding
n=8											
E7	175	40	16	46.8	-108	39	37.8	**042**	**90**		joint
E7	175	40	16	46.8	-108	39	37.8	**082**	**90**		joint
E7	175	40	16	46.8	-108	39	37.8	**065**	**90**		joint
E7	175	40	16	46.8	-108	39	37.8	**045**	**68**	E	joint
E7	175	40	16	46.8	-108	39	37.8	**041**	**90**		joint
E7	175	40	16	46.8	-108	39	37.8	**036**	**90**		joint
E7	175	40	16	46.8	-108	39	37.8	**056**	**90**		joint
E7	175	40	16	46.8	-108	39	37.8	**038**	**90**		joint
E7	175	40	16	46.8	-108	39	37.8	**084**	**90**		joint
E7	175	40	16	46.8	-108	39	37.8	**060**	**90**		joint
E7	175	40	16	46.8	-108	39	37.8	**078**	**90**		joint
E7	175	40	16	46.8	-108	39	37.8	**061**	**90**		joint
E7	175	40	16	46.8	-108	39	37.8	**051**	**90**		joint
E7	175	40	16	46.8	-108	39	37.8	**073**	**90**		joint
E7	175	40	16	46.8	-108	39	37.8	**064**	**90**		joint
E7	175	40	16	46.8	-108	39	37.8	**082**	**90**		joint
E7	175	40	16	46.8	-108	39	37.8	**120**	**5**	S	bedding
E7	176	40	16	52.9	-108	39	32.7	**034**	**90**		joint
E7	176	40	16	52.9	-108	39	32.7	**052**	**90**		joint
E7	177	40	17	46.4	-108	39	42.9	**092**	**90**		joint
E7	177	40	17	46.4	-108	39	42.9	**081**	**90**		joint
E7	177	40	17	46.4	-108	39	42.9	**072**	**90**		joint
E7	177	40	17	46.4	-108	39	42.9	**076**	**90**		joint
E7	178	40	18	27.4	-108	40	20.1	**073**	**90**		joint
n=24											

*db cluster identifies a group of immediately adjacent deformation bands, rather than a single band. A cluster indicates more pronounced grain crushing and shear than is associated with a single band (e.g. Aydin and Johnson, 1978).