THE CONTROL OF FAULT ARRAY EVOLUTION ON SEDIMENT DISPERSAL AND FOOTWALL DEGRADATION IN RIFT SYSTEMS

Thesis submitted as partial fulfillment of the requirements for the degree of
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The sedimentology and stratal architecture of syn-rift deposits in extensional basins are a record of the complex interaction between the rate of accommodation generation and the rate of sediment supply. Although tectonic subsidence and the nature and location of the source terrain are recognised as the primary controls on the rates of accommodation generation and sediment supply respectively, temporal and spatial variations in these parameters remain poorly understood and largely unquantified. Through detailed investigation of an unique subsurface dataset from the northern North Sea basin, this study addresses these issues and qualifies the rôle of fault array evolution on syn-rift sediment source, dispersal and deposition. The dataset employed comprises 3D seismic, electrical log, core and biostratigraphic data; the seismic coverage encompassing >60 km of the northern end of the half-graben bounding Late Jurassic Strathspey-Brent-Statfjord fault system. Seismic interpretation of fault segments and their abandoned palaeo-tips, and investigation of displacement-length scaling relationships were employed to reconstruct the temporal and spatial evolution of the fault population in the study half-graben. The timing of activity on individual segments was determined from stratigraphic evidence. An initial population comprising a large number of short, low displacement fault segments is identified; these faults defined a graben-like geometry. As the rift event progressed, strain was concentrated on fewer structures and, after ~10 Ma, on the Strathspey-Brent-Statfjord fault alone; hence the basin subsequently evolved an half-graben geometry. As the number of active faults in the basin decreased with strain localisation so the rates of displacement increased. The evolution of the fault population strongly influenced the distribution and stratal architecture of syn-rift sediments. During the initial stages of rifting (deposition of the Tarbert and Heather formations), the rate of sediment supply largely kept pace with the rate of tectonic subsidence. Sediment was sourced externally from the southward retreating Brent delta complex. In the rift climax phase the basin was sediment starved (deposition of the Kimmeridge Clay Formation), rates of tectonic subsidence greatly outpaced rates of sediment supply. The dominant process of sedimentation was hemipelagic settling; a secondary external source of sediment supplied an axial turbidite system during the Kimmeridgian. The impact of sediment starvation was the establishment of a submarine fault bounded range front that had an elevation locally in excess of 750 m. This bathymetric high was denuded by gravity driven sliding, controlled by the rheology of the pre-rift geologies exposed in the scarp; the products are preserved both mounted on the footwall scarp (the fault scarp degradation complex) and in the proximal hangingwall. The rates of denudation were low. The total volume of sediment liberated was minor in terms of the volume of the uplifted area and the volume of accommodation space generated and also the length of time that the high was exposed (up to 60 Ma). Observations on the magnitude and rates of footwall erosion in the northern North Sea are supported by reconnaissance field studies of four latest Jurassic – Early Cretaceous faults in central east Greenland.

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and my mind observed to me,
or I to it, how ordinary
extraordinary things are or

how extraordinary ordinary
things are, like the nature of the mind
and the process of observing.

An ordinary day

Norman MacCaig, 1966

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CHAPTER 1 Introduction

1.1 Rationale

Traditional models of syn-rift sedimentation in a half-graben basin are tectonically static; the distribution of normal faults remains fixed in time and space. The accommodation of strain by a fault population through the accumulation of displacement results in hangingwall subsidence, uplift of the footwalls and the rotation of slopes in the basin – consistent with the predictions of the 2-dimensional 'rotating domino' model of normal faulting (e.g. McKenzie and Jackson 1986, Jackson *et al.* 1988). Despite their structural simplicity, traditional half-graben sedimentation models have been successful in elucidating the influence that the rates of accommodation generation and sediment supply have on the dispersal and deposition of syn-rift sediments and, hence, the facies mosaic of extensional basins (e.g. Leeder and Gawthorpe 1987, Leeder 1993, Prosser 1993).

Tectonic subsidence is recognised to be the primary control on the generation of accommodation in tectonically active basins. Recent field based studies examining the controls on the spatial distribution of facies within rifts have incorporated an understanding of the character of normal faults, consequently modifying the classic model of a 'wedge-shaped' syn-rift package (see, for example, Fig. 4 of Gawthorpe *et al.* 1997). Normal faults are discrete planes of dislocation within the crustal lithosphere; maximum displacement is located near the centre of the structure and decreases towards the margins (e.g. Rippon 1985, Barnett *et al.* 1987, Huggins *et al.* 1995). The displacement gradient along the strike of a normal fault is mirrored by the magnitude of tectonic subsidence in the proximal hangingwall – maximum accommodation is generated at the centre of, and proximal to, an isolated fault strand, accommodation space decreases towards the tips and away from the fault. The spatial variability in the magnitude of tectonic subsidence results in 'half saucer-shaped' hangingwall depocentres located adjacent to each isolated fault strand (e.g. Anders and Schlische 1994, Schlische 1995, Gawthorpe *et al.* 1997). In cross sections oriented perpendicular to the strike of the bounding fault, the geometry of these depocentres is wedge-shaped; however in cross sections paralleling the fault strike the depocentres are synclinal. Correlatively, the uplifted footwall describes a broad anticline with the hingeline oriented perpendicular to the strike of the fault.

In addition to strongly controlling the loci, magnitude and rate of accommodation generation, and hence the architecture of syn-rift deposits, the distribution and growth of the fault population also determines the topography of the rift basin. It has long been recognised that uplift of the footwall and downthrow of the hangingwall results in relative topographic highs and lows, forming sediment source areas and sediment depocentres. The consequence of the fault controlled topography is, for example, to focus axial transport systems into the proximal hangingwall (e.g. Alexander and Leeder 1987, Alexander *et al.* 1994, Peakall 1998). Integrating observations of along strike variations in fault displacement permits spatial qualification of tectonic slopes in the hangingwall and, significantly, locates intra-basin highs between adjacent fault segments (Anders and Schlische 1994). Furthermore, fault segmentation, specifically overlapping en echelon strands, results in a potentially important sediment routeway from footwall to hangingwall via a shallowly dipping transfer zone (or relay ramp) (Gawthorpe and Hurst 1993, Leeder and Jackson 1993, Jackson and Leeder 1994, Gupta *et al.* 1999).

Although expanded into 3-dimensions and illustrating the importance of along strike variations in fault displacement and fault segmentation for accommodation generation and sediment dispersal, the modified model of half-graben sedimentation remains tectonically static, representing only a snapshot of the basin with an unique configuration of the fault population. Extensional fault systems, however, are dynamic. Faults grow by the processes of radial tip propagation and linkage of segments (e.g. Peacock and Sanderson 1991, 1994, Cartwright *et al.* 1995). Numerical and analogue models of fault growth have demonstrated that normal fault populations develop as the amount of strain increases (e.g. Fossen and Gabrielsen 1996, Cowie 1998); an initial population of a large number of short, low displacement faults will evolve into a late population comprising a few large faults. The temporal and spatial evolution of the active fault population will be reflected by changes in the basin topography and changes in the location, magnitude and rate of accommodation generation. For example, the radial propagation of a fault segment will result in the coeval widening and deepening of the proximal hangingwall basin, while the linkage of two fault segments will result in the amalgamation of two isolated depocentres, the locus of deposition migrating to the centre of the longer strand (Schlische 1995). Further, fault linkage will breach a relay ramp, typically redirecting transfer zone drainage (Gupta *et al.* 1999, although see Gilpin 1998).

The relationship between fault growth and syn-rift facies architecture has long been recognised. In addition, it is clear that the distribution and stratal architecture of syn-rift sediments are a record of coeval tectonic subsidence, providing an opportunity to constrain the tectonic evolution of a rift basin: The distribution of active faults, rates of fault growth and linkage histories. However, a 4-dimensional, temporal and spatial, understanding of local variations in fault controlled subsidence rates, and the consequent impact on basin-wide sediment dispersal and deposition, has yet to be achieved. This is largely due to the limitations inherent in 2-dimensional field exposures of syn-rift sediments in exhumed continental rift basins; reconstructions of tectono-stratigraphic evolution are generally only possible for local areas and a discrete part of the rift event (e.g. Armstrong 1997, Gupta *et al.* 1999). Subsurface studies, in which the entire syn-rift stratal architecture can be resolved, have been restricted by wide spacing of seismic lines, poor age constraints and the limited area of 3D surveys (e.g. Morley 1999, Dawers and Underhill 2000).

This study integrates sedimentary and tectonic observations with the goal of qualifying the growth of extensional fault populations and determining the impact on syn-rift sediment source, dispersal and deposition through the rift event. This is achieved through detailed investigation of a unique subsurface dataset, combining ~1000 km^3 3D seismic coverage (with line spacing of $\leq$25 m) with well data providing biostratigraphically defined ages of syn-rift sediments at a resolution of 2-3 Ma. The seismic data encompasses >60 km of the length of the Strathspey-Brent-Statfjord fault system in the northern North Sea basin. In order to qualify the balance between the rate of sediment supply and the rate of tectonic subsidence this study examines the sources of syn-rift sediments, in particular quantifying the volume of sediments derived from within the basin. Observations on the rôle of footwall denudation as source of clastics in the northern North Sea basin are supported by the results of field studies of four latest Jurassic – Early Cretaceous faults in the central east Greenland rift.

1.2 The thesis

The thesis of this study is to constrain the supply of sediment to marine rift basins, and qualify the controls on syn- and post-rift sediment dispersal and deposition. This goal is addressed by examination of two key issues,

1. To gain a predictive understanding of the accommodation generated by tectonic subsidence through reconstructing the temporal and spatial evolution of the fault population. The growth of normal fault arrays also exerts a primary control on the topography of the basin floor, hence influencing sediment dispersal routeways (particularly in underfilled basins);

2. To qualify the controls on temporal and spatial variability in the rates of sediment supply and appreciate the relative importance of internal and external sources of sediment. In particular, the rôle of footwall denudation as a source of clastics in marine rift basins starved of externally derived sediments is examined.

By integrating these goals the balance between the rate of accommodation generation and the rate of sediment supply can be qualified and the controls on sedimentation determined.

1.3 Structure of report

This report documents in nine, largely self-contained, chapters the rationale and background to the thesis, the results and conclusions of the investigations and the implications of this work for other rift basins.

The following chapter, Chapter 2, considers the precedents for, and the context of, this study through examination of three topics – the growth of normal fault populations, controls on the distribution and stratal architecture of syn-rift sediments and sources of clastic sediment in continental rift basins. This review of the literature serves to demonstrate the rôle of this thesis study in marrying structural and sedimentological theory with the goal of achieving a 4-dimensional understanding of the evolution of extensional basins. This work investigates the development of two failed rift basins: The Late Jurassic northern North Sea basin and the latest Jurassic – Early Cretaceous rift system of central east Greenland. An overview of the geological history and the results of previous tectono-sedimentary studies in these areas are presented in the succeeding chapters.

Chapters 3-8 present the results and conclusions of investigations in the two study areas. A subsurface dataset comprising 3D seismic and well (electrical logs, core and biostratigraphy) data was used to examine the evolution of the half-graben basin bounded by the Late Jurassic Strathspey-Brent-Statfjord fault system in the northern North Sea. This work comprises the bulk of the thesis and can be subdivided into three parts: Chapter 3 reconstructs the growth of the normal fault population, the succeeding two chapters determine the controls on syn- and post-rift sedimentation respectively, and Chapters 6 and 7 examine the magnitude of footwall denudation and the process(es) of downslope transport. The observations from this area are contrasted with field studies of four latest Jurassic – Early Cretaceous faults in central east Greenland (Chapter 8). As syn-rift deposits in the east Greenland rift basin typically remain buried, the study focuses on the magnitude of denudation of the well exposed footwalls. Chapter 8 concludes with an assessment of the implications these results (from both study areas) for quantifying footwall degradation as a source of clastic sediment in marine rift basins.

The final chapter of this report reappraises the thesis and summarises the key conclusions. The broader implications of these results for understanding the growth of large fault systems, quantifying the controls on syn-rift sedimentation in continental rift basins and the rôle of footwall denudation as a source of clastic sediment are discussed. The report concludes with an examination of the contribution of this work to hydrocarbon exploration and exploitation.

CHAPTER 2 Background to research

2.1 Introduction

Rift basins are one of the fundamental building blocks of the earth sciences and, consequently, a large body of research exists documenting the geology and interpreting the evolution of these settings. In this chapter the structure and sedimentology of rift basins in general, and those located in a marine environment in particular, are assessed. This review is not intended to be comprehensive but rather to explain the context of this thesis study; the specific results of previous studies are integrated into the interpretation and discussions in this report. Only three fields of study are examined in detail in the following sections: The growth of normal fault systems, controls on the distribution and stratal architecture of syn-rift sediments and the sources of clastic sediment in rift basins. From this examination of the literature the two theses of this study are defined: Understanding the tectono-stratigraphic evolution of rift basins and the rôle of footwall degradation as a source of clastic sediments in rifts.

2.2 The structural evolution of rift basins

An overview of the structure of rift basins

Rifts are defined as either passive or active: In passive rift systems, regional tensile stresses cause failure of the crust resulting in lithospheric thinning and upwelling of the asthenosphere, active rift basins develop in response to the impingement of a thermal plume at the base of the lithosphere causing uplift, thinning and extension (e.g. Sengör and Burke 1978, Morgan and Baker 1983). In both types of basin, the response of the crustal lithosphere to extension is brittle failure while the subcrustal lithosphere will deform plastically. A number of mechanical models have been invoked over the last 20 years to describe the structural evolution of rift basins, these largely depend upon the distribution of strain in the lithosphere (e.g. the uniform stretching model, McKenzie 1978; the depth-dependant stretching model, Royden and Keen 1980, Scalter *et al.* 1980).

Tensile stresses in the crustal lithosphere cause brittle failure; strain is accommodated on extensional fault arrays. At the surface active normal fault planes are typically observed to be steep (45-60°) and evidence from earthquake ruptures suggest that fault planes remain steep to depths of 8-10 km (Jackson and McKenzie 1983). The gross, 2-dimensional architecture of extensional fault populations has been explained by the 'rotating domino' model (e.g. McKenzie and Jackson 1986, Jackson *et al.* 1988). In this model, high angle, planar normal faults bound rigid blocks; as displacement is accumulated on the faults, the blocks rotate about a horizontal axis. As the amount of strain increases so the fault planes are rotated to shallow angles; at dips of less than ~30° continued slip is not mechanically possible and a second generation of high angles faults are initiated, cross-cutting the early population (see Fig. 7 of Jackson and White 1989). The growth of a second generation of faults has been invoked to explain observation of low angle faults (e.g. Wernicke and Burchfiel 1982). However, alternative models of listric faults with detachments at depth and planar low angle faults that dissect the crust have also been proposed (Wernicke 1981, 1995, Westaway 1999).

The surface expression of continental rift basins is a linear, typically asymmetric, depression bounded by large displacement normal fault arrays. Scholz and Contreras (1998) explain the preferred development of half-graben type basins by a simple mechanical model. The rift initiates with the formation of a conjugate set of normal faults, but as the faults grow they intersect at depth locking one of the faults. The other fault continues to grow and evolves into the basin bounding fault. The rift bounding fault system commonly demonstrates along axis segmentation on +100 km scale defined by en echelon steps or a change (transfer) of polarity (e.g. in the East African Rift, Rosendahl *et al.* 1986, Upcott *et al.* 1996; in the Suez Rift, Colletta *et al.* 1988, Patton *et al.* 1994). The scale of segmentation has been related to the thickness of the seismogenic crust; it is considered that the thicker the crust, the longer the faults and the wider the half-graben (Hayward and Ebinger 1996, Jackson and Blenkinsop 1997, Ebinger *et al.* 1999).

The growth of normal fault arrays

Normal fault planes are discrete surfaces of slip developed in response to extensional stresses. Strain is accommodated by the accumulation of displacement across the fault plane; both uplift of the footwall and downthrow of the hangingwall accommodate coseismic slip (e.g. Stein and Barrientos 1985) (Fig. 2.1). In 3-dimensions, faults are observed to be elliptical planes of discontinuity in the crust (Fig. 2.2a) – contours of displacement illustrate that maximum slip occurs near the centre of the plane and the magnitude of slip decreases towards the margins (Rippon 1985, Barnett *et al.* 1987, Huggins *et al.* 1995). In cross section, parallel to the strike of the fault, a profile of displacement-length is shown have maximum displacement at the centre of the fault and to decline to zero at the tips (Fig. 2.2b). Field observations demonstrate a scale invariant relationship between the trace length and maximum displacement of isolated normal faults of the form $D = c.L^n$, where D is the displacement, *c* is a constant related to the rock properties, L is the length of the fault and n is some exponent (e.g. Dawers *et al.* 1993, Carter and Winter 1995). Schlische *et al.* (1996) have compiled a dataset comprising displacement and length measurements from 547 faults located in both ancient and modern compressional and extensional settings (Fig. 2.3). They demonstrate a best-fit correlation between maximum displacement and trace length for these data (covering in excess of eight orders of magnitude) of $D = 0.03.L^{1.06}$, i.e. a near linear relationship.

Two end member models have been proposed to explain the growth of normal faults by radial tip propagation. In one model (Watterson 1986, Walsh and Watterson 1988, Marrett and Allmendinger 1991, Gillespie *et al.* 1992), stresses at the fault tip are induced and entirely dissipated during each unique seismic event, i.e. slip and fault propagation are geologically instantaneous and between seismic events the stresses are relaxed. The results of this model predict a non-linear relationship between maximum displacement and trace length, of the form $D = c.L^{1.5\text{-}2}$. In contrast, in the second model, stresses at the fault tip are always just equal to the shear strength of the rock mass and are accumulated in a 'process zone' (Cowie and Scholz 1992a, 1992b, Cowie *et al.* 1993). The results of the second model predict 'bell-shaped' displacement-length fault profiles (*cf.* Fig. 2.2b) and a linear relationship between maximum displacement and trace length, consistent with the conclusion of Schlische *et al.* (1996).

The data compiled by Schlische *et al.* (1996) have a large scatter, of approximately an order of magnitude in both displacement and length (Fig. 2.3); while this scatter is in part a reflection of variable rock properties and the techniques of data collection (e.g. Cowie and Scholz 1992b, Gillespie *et al.* 1992), it is also related to the effects of fault growth by segment linkage (e.g. Peacock and Sanderson 1991, Scholz *et al.* 1993). Cartwright *et al.* (1995) examine the evolution of displacement-length scaling for fault growth by both radial tip propagation and segment linkage employing field data from the Canyonlands Graben, Utah (Fig. 2.4a). They suggest that a fault growing by radial propagation in an homogeneous rock body will follow the growth path prescribed by the displacement-length scaling relationship $D = c.L^n$ (Fig. 2.4ai). A fault growing by segment interaction and linkage will, however, have a step-like growth path. Steps represent the linkage of two fault segments; linkage increases the length of the fault but not the maximum displacement with the consequence that the ratio of maximum displacement to trace length deviates below that predicted by the scaling relationship (Fig. 2.4aii). Cartwright *et al.* propose a final stage in fault growth by segment linkage – following linkage a fault will accumulate displacement with little increase in length and thus rejoin the growth path conforming with $D = c.L^n$. Bends in the trace of the fault and fault tips abandoned in the footwall or hangingwall after breaching of the relay ramp preserve evidence of segment linkage and the geometry of palaeo-segments (Fig. 2.4b).

In the schematic linkage model presented by Cartwright *et al.* (1995) interacting fault segments are shown to have skewed displacement-length profiles; the maximum displacement is skewed from the centre towards the neighbouring structure and, hence, displacement gradients in the relay zone are steep (see also Fig. 2.2b). In addition, the central of the three segments is shown to have accumulated displacement disproportionately high for the trace length (segment b is 'over-displaced', Fig. 2.4aii). Field observations show that the displacement-length ratios of the central faults in an array can be significantly larger than those of marginal segments (e.g. Willemse *et al.* 1996). The magnitude of increased displacements (in comparison with the predictions of the scaling relationship) prior to linkage is demonstrated by the fault array described by Dawers and Anders (1995). The cumulative displacement of all the segments composing this unlinked fault array is similar to that predicted by displacement-length scaling relationships for an isolated fault growing by radial propagation (Fig. 2.2c); the implication of this observation is that when the faults link, the throughgoing fault strand will not be significantly 'under-displaced'.

Cowie (1998) explains the interaction between normal fault segments that leads to enhanced growth of overlapping segments in terms of an healing-reloading stress feedback mechanism (Fig. 2.5). In this model individual seismic rupture events are considered to determine the local stress field (King *et al.* 1994) with the consequence that the growth of some structures is enhanced and the growth of others inhibited. For example, a seismic event on fault A (Fig. 2.5) will result in a positive stress change around the tips of the fault and 'loading' of faults B and C. In contrast, fault D is located in an area where stress is relaxed following rupture of fault A ('stress shadow zone') and the continued growth of fault D is inhibited.

Cowie attributes the preferred growth of en echelon fault arrays and temporal and spatial variations in rates of fault growth (described by Nicol *et al.* 1997) to this mechanism. This model is supported by field observations presented by Roberts (1996); seismic rupture events are described that are shorter than the host fault segment and vary spatially in recurrence interval and the magnitude of slip.

The healing-reloading mechanism described by Cowie (1998) has also been invoked to explain the transition from the 'rift initiation' to the 'rift climax' phases of basin evolution. The rift initiation and rift climax phases are defined on the basis of observations of a characteristic stratigraphy in extensional basins – an early coarse grained succession of fluvial or shallow marine deposits overlain by finer grained sediments of a lacustrine or deep marine environment (Prosser 1993, Lambiase and Bosworth 1995, Schlische and Anders 1996) (Fig. 2.6). The transition between initiation and climax is considered to represent an increase in the rates of subsidence and a decrease in the rates of sediment supply. Employing the stress feedback mechanism, Gupta *et al.* (1998) demonstrate that this transition can be attributed to a localisation of strain on preferentially oriented fault segments as the rift event progresses. Slow displacement rates prevail early in the rift event as strain is widely distributed, but a sharp increase in rates of displacement (tectonic subsidence) occurs as strain is localised on a smaller number of active fault arrays. The prediction of this model is that an early fault population comprising a large number of short, low displacement faults will evolve into a late population comprising a few long, large displacement linked fault arrays.

2.3 Syn-rift sedimentation in extensional basins

Controls on accommodation space

Accommodation is the space available for deposition of sediment; the development of accommodation reflects variations in relative sea level (e.g. Posamentier *et al.* 1988). A rise in relative sea level can be driven eustatically or by subsidence in response to tectonics, isostacy or compaction. Eustatic sea level fluctuations occur at varying timescales and the changes are of variable magnitude (e.g. Haq *et al.* 1988). However, the rate and magnitude of change tend to be relatively slow and, consequently, numerous studies have concluded that tectonic subsidence is the primary control on the generation of accommodation space in extensional basins (e.g. Leeder and Gawthorpe 1987, Schlische 1991, Anders and Schlische 1994, Gawthorpe *et al.* 1994, Contreras *et al.* 1997). As normal fault populations are not static, the rate and location of tectonically generated accommodation will evolve temporally and spatially during a rift event mirroring the growth of the fault population.

Schlische (1995) has examined the implications of fault growth for the distribution of syn-rift sediments. Schlische describes the development of accommodation space in fault related extensional folds during the propagation and linkage of fault segments, and illustrates the stratal architecture of coeval syn-rift deposits (Fig. 2.7). Extensional folds are developed in response to variations in displacement along the strike of a normal fault. The length and maximum displacement of the bounding fault segments control the wavelength and amplitude of extensional folds. In the hangingwall, the hinges of folds are oriented perpendicular to, and plunging towards, the fault; the folds define fault bounded 'scoop-shaped' synclinal depocentres, separated by anticlinal intra-basin highs (IBH). Correlative opposing structures deform pre-rift geologies in the footwall. The distribution of coseismic slip (Fig. 2.1) results in higher amplitude folds in the hangingwall than in the footwall.

For an isolated fault segment growing by bilateral tip propagation, Schlische (1995) demonstrates the widening and deepening of the proximal hangingwall basin as the fault trace lengthens and the maximum displacement on the segment increases (Fig. 2.7a). Assuming that the rate of sediment supply is equal to, or exceeds, the rate of tectonic subsidence, sediments deposited during the growth of the extensional fold are shown to onlap the flanks of the fold and thicken towards the point of maximum accommodation generation, i.e. the centre of the fault. Syn-rift deposits thus demonstrate both the traditional wedge-shaped thickening towards the fault and thickness variations along the strike of the structure reflecting the differential generation of accommodation space. In the case of the linkage of two fault segments to form a single strand (Fig. 2.7b), previously isolated hangingwall depocentres will amalgamate to form a single depocentre. Consequently, the location of maximum accommodation generation will migrate to the centre of the linked fault, i.e. the centre of the amalgamated depocentre, previously an intra-basin high and location of minimum accommodation space. The interpretations of Schlische (1995) are supported by geological observations of sedimentation patterns in extensional basins (e.g. Schlische and Anders 1996, Morley 1999, Dawers and Underhill in press).

Fault growth can be shown to exert a control on the generation of accommodation, and hence sedimentation patterns, even before the fault plane has breached the surface.

Gawthorpe *et al.* (1997) present a model, derived from field studies in the Suez Rift, of the evolution in stratal geometry of sediments deposited in the hangingwall to a fault controlled monocline (Fig. 2.8). As a monocline grows in response to the vertical propagation of a blind normal fault, coeval sedimentation thins towards and onlaps the steep limb of the fold; the locus of deposition is distal of the buried fault. After the fault has breached the surface, the depocentre migrates into the proximal hangingwall and syn-rift sediments thicken into the fault. The stratigraphic model of Gawthorpe *et al.* is applicable at the growing tips of a fault, where the fault is blind, while the model of Schlische is contemporaneously applicable along strike – demonstrating the control of fault growth on sediment distribution and stratal architecture in rift basins. The models of both Schlische (1995) and Gawthorpe *et al.* (1997) have significant implications for reconstructing the temporal and spatial evolution of fault populations in ancient rift basins, using the evidence of coeval sedimentation to demonstrate the locations of active fault segments and to determine the rates of tectonic subsidence.

In addition to defining the distribution and stratal architecture of syn-rift deposits, the evolution of the fault population also determines temporal and spatial changes in the topography (bathymetry) of an extensional basin; hence influencing sediment dispersal. The control of the fault population on the topography within a half-graben type basin is illustrated by the conceptual model of Gawthorpe *et al.* (1994). In this schematic block diagram a multi-segmented normal fault array is shown to bound a subaerial range front (Fig. 2.9a). Potential drainage patterns within the basin respond to the fault controlled topography; sediment dispersal routeways are oriented either laterally or axially to the strike of the basin bounding fault.

Lateral drainage systems can be subdivided into those sourced internally or those sourced externally of the half-graben. Internally sourced drainage systems derive from degradation of the footwall scarp (short, steep basins) or of the hangingwall dip slope (broad, palmate basins). Drainage basins in both settings have restricted catchment areas; hence low rates of sediment supply to the proximal half-graben (e.g. Gawthorpe and Colella 1990). However, periodic seismic activity will elevate the crest of the scarp and rotate the hangingwall dip slope; hence rejuvenating the basins and ensuring their longevity throughout the rift phase. Footwall uplift is a barrier to lateral drainage; in order to maintain the routeway of pre-existing, externally sourced antecedent drainage systems, rates of erosional downcutting must exceed rates of tectonic uplift (Eliet and Gawthorpe 1995). As the rates of uplift increase during the rift event, drainage is typically reversed or rerouted to exploit transfer zones between fault segments (Fig. 2.10). The low angle slopes between overlapping fault strands, which connect footwall and hangingwall, are important point sources of externally derived sediment in half-graben basins (e.g. Gawthorpe and Hurst 1993, Leeder and Jackson 1993, Jackson and Leeder 1994, Gupta *et al.* 1999). However, these areas only act as a conduit for sediment for a short time prior to segment linkage; after the relay ramp is breached and the area uplifted, the sedimentary dynamics are reversed.

Similarly to antecedent and transfer zone drainage systems, the sediments supplied to a basin via axial systems are generally sourced externally. Axial drainage is typified by longitudinal fluvial systems located in the proximal hangingwall to the bounding fault system. Tectonic subsidence and hangingwall rotation focus fluvial channels into the deepest part of the basin, with the consequence of stacking fluvial deposits (e.g. Alexander and Leeder 1987, Alexander *et al.* 1994, Peakall 1999).

Controls on sediment supply

Although tectonic subsidence is recognised to be the primary control on accommodation space in extensional basins the coeval filling of the space requires the rate of sediment supply to keep pace with, or exceed, the rate of accommodation generation. Similarly to the rate of tectonic subsidence, the rate of sediment supply to a basin demonstrates temporal and spatial variability during a rift event. The flux of sediment is controlled by a multitude of diverse factors including the climate, the local bedrock geology and the size of the catchment area (e.g. Anhert 1970, Summerfield and Hulton 1994, Leeder *et al.* 1998). In general, sediment supply can be subdivided into intra- and extra-basinal sources.

The conceptual model presented by Gawthorpe *et al.* (1994) describes sediment dispersal and deposition in a subaerial basin and emphasises the rôle of externally derived sediments (Fig. 2.9a). The volume and rate of supply of sediments from intra-basin sources is low, although these sedimentary systems tend to be long-lived. The redeposition of sediments derived from the degradation of the footwall scarp and hangingwall dip slope is restricted to narrow talus slopes of amalgamated alluvial fans and broad, palmate hangingwall fans, respectively. The distribution of coseismic slip means that the volume of geologies uplifted in the footwall will be less than the volume of accommodation generated in the hangingwall (Fig. 2.1); simple mass balance demonstrates that a basin cannot be filled with internally derived sediments alone (Stein *et al.* 1988, Contreras *et al.* 1997). As a consequence, if accommodation generated during rifting is to be coevally infilled, an external source of sediment is required.

The supply of sediments derived from an extra-basinal source is typically via an axial fluvial system (or, in lacustrine settings, an axial turbidite). As fluvial channels attain their load from erosion of interfluves, the sediment flux is controlled by the rate of denudation in a distal hinterland. Conditions affecting the erosion of a terrain include rock uplift, lithology and structure, precipitation, temperature, vegetation, mean elevation, local relief and hydraulic conditions in the fluvial channel (Hovius and Leeder 1998); all these parameters are temporally and spatially variable. The processes of denudation are numerous and variable in rate, from the slow disintegration of bedrock by weathering to the catastrophic failure of a hillslope by translational sliding. Recent studies have demonstrated that the dominant mechanism in the erosion of mountainous terrain is landsliding, triggered by fluvial incision (Burbank *et al.* 1996, Hovius *et al.* 1997, Densmore *et al.* 1998, Hovius *et al.* 1998).

Extra-basinal sources of sediment, although significant in volume, are typically only important during the earliest phases of an extensional event; rift basin stratigraphy commonly recording a marine incursion, or the establishment of a lake, early in the rift history (e.g. Lambiase 1990, Prosser 1993). As a consequence of flooding the interior of a rift, the supply of sediments from extra-basinal sources declines in volume. Ravnås and Steel (1998) attribute this decrease to a partitioning of hinterland derived clastic sediments at the rift margins as the rate of tectonic subsidence in the basin rapidly outpaces the rate of sediment supply. As the basin becomes increasingly sediment starved and displacement accumulates on half-graben bounding fault arrays, a fault controlled topography will be established influencing sediment routeways and isolating the basin further. Syn-rift deposits during the climax of a rift event thus tend to be derived predominantly from low volume intra-basinal sources.

The potential intra-basinal sources of sediment in marine half-graben are suggested in Fig. 2.9a: From the water column and from degradation of the footwall and hangingwall scarps. The potential processes of sedimentation in the hangingwall are 3-fold (excluding *in situ* authigenic processes), and are comparable to the processes active in a deep marine environment identified by Stow *et al.* (1996),

1. Pelagic settling

   Pelagic sediment is generated in the open sea and is dominantly of biogenic origin. Near continental margins, where the terrigenous component of suspended sediment is >25 %, the deposits are classified as hemipelagic.

2. Semi-permanent bottom currents

   These currents are the deep water expression of oceanic thermohaline circulation. The currents are variable in velocity and location but are capable of erosion of marine sediments, transport and redeposition (as contourites).

3. Episodic resedimentation

   Encompasses all gravity driven processes; a continuum from dilute turbidity currents to large translational and rotational slide complexes. These processes are active on all slopes from steep footwall scarps to the shallowly dipping limbs of hangingwall synclines (Fig. 2.7).

Although the sediment sources and depositional processes in marine rift basins are understood, the controls on, and rates of, sedimentation in these basins have received relatively little attention, certainly in comparison to subaerial equivalents. Existing models of sedimentation in submarine half-graben are static and 2-dimensional (e.g. Surlyk 1978, 1987, 1989, Ferentinos *et al.* 1983). Indeed, in reviewing clastic sediment supply to basins, Hovius and Leeder (1998) assigned marine rift basins an *hors categorie*, as the supply of sediments sourced from a submarine topography has not been, and cannot be, examined by any of the existing landscape evolution models.

2.4 Aims of research

The tectono-stratigraphic evolution of extensional basins

The subdivision of structural geology and sedimentology in the review of rift basins presented above indicates the lack of integration of research in these two fields; yet the review also illustrates that a complete understanding of the evolution of syn-rift sedimentation cannot be achieved without constraining the dynamics of the structural development.

Although tectonic subsidence is acknowledged as the primary control on accommodation generation, models of syn-rift sedimentation developed from field data tend to be structurally static, i.e. the distribution of normal faults is fixed in time and space. While valuable in presenting a snapshot of the sedimentation patterns for a unique tectonic configuration, such models cannot examine the implications of temporal and spatial variability in accommodation generation for sediment dispersal and deposition throughout a rift event. Correspondingly, the distribution of syn-rift sediments is recognised to reflect the coeval generation of accommodation space; hence the loci of deposition will migrate as the rift event progresses, controlled by the development of the fault population. Despite this conceptual understanding, stratigraphy remains an underutilised tool in addressing problems such as temporal and spatial variability in the rate of fault growth and the location of the active fault population during a rift event.

The aim of this study is to achieve a 4-dimensional, dynamic understanding of half-graben sedimentation, building upon existing models of syn-rift sedimentation (e.g. Gawthorpe *et al.* 1994) and testing conceptual models of fault growth on a geological timescale. Sedimentological and structural observations are integrated with the goal of qualifying the growth of extensional fault populations and realising the impact on source, dispersal and deposition of sediments as total strain increases. Previous studies addressing this thesis have been largely field based and, as a consequence, inhibited by the 2-dimensional exposures in exhumed continental rift basins (e.g. Schlische and Anders 1996); however, such studies have proven successful in reconstructing the tectono-stratigraphic evolution of a local area during a discrete phase of the rift history (e.g. Armstrong 1997, Gawthorpe *et al.* 1997, Gupta *et al.* 1999). Subsurface data have also been employed to examine this thesis with the advantage that the stratal architecture of the entire syn-rift can be resolved and fault planes can be mapped. These studies have, however, been restricted by the wide spacing of 2D seismic lines, poor age constraints and the distribution and size of 3D surveys (e.g. Morley 1999, Dawers and Underhill in press).

This study has overcome the limitations identified by previous workers by employing a unique subsurface dataset combining ~1000 km^3 of 3D seismic coverage (line spacing ≤25 m) with well data which provides, through biostratigraphic dating, constraint on the age of syn-rift sediments. The subsurface study area is located in the northern North Sea basin, one arm of the trilete failed North Sea rift system (Fig. 2.11). West-east oriented extensional stresses in the northern North Sea region during the Late Jurassic resulted in ~15 % strain (e.g. Yielding 1990), largely accommodated on equally spaced, sub-parallel normal fault arrays with maximum displacements typically in excess of 1 km. The province subsequently subsided and syn-rift sediments are now buried beneath 2-5 km of passive fill. The areal extent of the seismic data available to this study encompasses >60 km of the half-graben bounding Strathspey-Brent-Statfjord linked fault array, including the northern tip of the easterly dipping fault system. Seismic interpretation permits mapping of the distribution and geometries of Late Jurassic faults in the Strathspey-Brent-Statfjord area, examination of fault scaling relationships and, from the stratal architecture of the syn-rift, definition of the timing of activity and rates of displacement on discrete fault segments.

At the initiation of rift tectonics the Strathspey-Brent-Statfjord area was located at the seaward margin of the Brent delta complex. As the cumulative strain increased, so the rate of tectonic subsidence increased and the basin was flooded (e.g. Johannessen *et al.* 1995, Ravnås *et al.* 1997). The retreat of the shoreline resulted in a decline in the supply of sediment; the character of sediments deposited during the climax of rifting is indicative of sediment starvation (e.g. Rattey and Hayward 1993). Through investigation of well data, particularly facies interpretation of cored intervals, the palaeo-environment and basin morphology can be reconstructed, and temporal and spatial variability in the rate of syn-rift sedimentation calculated. Integrating both structural and sedimentological interpretations permits quantification of the balance between the rate of sediment supply and the rate of tectonic subsidence as rifting progressed and the half-graben evolved.

The rôle of footwall degradation as a source of clastic sediment in rifts

The stratigraphy of an extensional basin is generally characterised by early flooding and, subsequently, sediment starvation – indicating that the rate of sediment supply is rapidly outpaced by the rate of tectonic subsidence (e.g. Lambiase 1990, Prosser 1993). With the flooding of a basin, the volume of externally derived sediment declines (Ravnås and Steel 1998); sediments deposited during the climax of rifting tend to be locally sourced. Although simple mass balance demonstrates that the volume of sediment liberated by the erosion of uplifted footwalls is insufficient to fill the accommodation generated by tectonic subsidence in the proximal hangingwall (Fig. 2.1), footwall denudation can be the only source of clastic sediment in marine rift basins and, consequently, the processes of denudation and the dispersal of the sedimentary products will dictate the stratigraphic fill (e.g. Surlyk 1989). Despite the potential significance of scarp erosion in marine rift basins, the controls on downslope processes for slopes located in a submarine environment remain poorly understood. This is partly a consequence of the difficulties inherent in studying modern submarine slopes, but also a more general lack of recognition of the geological remnants of slope failure processes (of both a submarine and subaerial setting). The low preservation potential of topographic highs (i.e. range fronts) and the rapid dispersal of denuded sediments both contribute to this problem.

This study examines the rôle of footwall degradation as a source of clastic sediment in rifts: Investigating the processes of downslope transport, quantifying the magnitude and rate of denudation (hence, rate of sediment supply) and determining the primary controls on sediment redistribution and deposition. This work complements the investigations into the tectono-stratigraphic evolution of rift basins described above.

Previous workers in the northern North Sea basin have recognised that the footwalls bounded by Late Jurassic fault arrays are significantly degraded (e.g. Dahl and Solli 1992, Coutts *et al.* 1996, Færseth *et al.* 1997, Underhill *et al.* 1997, Hesthammer and Fossen 1999). In addition, these authors demonstrate that the products of footwall erosion are preserved both interbedded with hemipelagic deposits in the proximal hangingwall and mounted on the denuded fault scarp. The complex of sedimentologically reworked and structurally disturbed pre-rift geologies mounted on the footwall – termed the fault scarp degradation complex by Underhill *et al.* (1997) – has been interpreted as the geological remnant of translational and rotational sliding processes (e.g. Livera and Gdula 1990). Such a complex is preserved on the degraded footwall of the Strathspey-Brent-Statfjord fault array but the displaced sediments are thin, typically <100 m in vertical thickness. Hence, while the boundaries of the complex can be defined and mapped (see McLeod and Underhill 1999), the internal structure is below the resolution of the seismic dataset. Seismic interpretation, reconstructing the eroded footwall crest, is employed to qualify the magnitude of denudation and the volume of sediments liberated. Access to >1.5 km of core samples from wells intersecting the fault scarp degradation complex and syn-rift sediments in the adjacent hangingwall permits reconstruction of the processes of scarp erosion and examination of the distribution and resedimentation of the degradation products.

Studies in the northern North Sea are supported by a field based investigation of four faults of the latest Jurassic – Early Cretaceous rift system in central east Greenland. The east Greenland continental margin has had a complex post-Caledonide evolution, experiencing a series of extensional events (see review of Surlyk 1990a) culminating in sea floor spreading and the opening of the North Atlantic in the Late Palaeocene (Talwani and Eldholm 1977, Bott 1987). A consequence of Cenozoic extension, and the impingement of the Iceland plume, has been to uplift the continental margin; the ensuing deep denudation of >2 km (Hansen 1992, Clift *et al.* 1998, Thomson *et al.* 1999) exhuming the rift geologies of the earlier extensional events. Well exposed syn-rift deposits illustrate a period of latest Jurassic – Early Cretaceous rifting (Surlyk 1978), with ~20 % strain accommodated on large 'domino-style' normal fault arrays (Price *et al.* 1997). This study undertook reconnaissance mapping of the well exposed degraded footwalls of these normal faults, and the proximal hangingwall sediments, identified from descriptions in the literature (e.g. Maync 1949, Surlyk 1978) and the work of Cambridge Arctic Shelf Program (CASP) geologists. These observations were employed to estimate the magnitude and rates of footwall erosion and constrain the controls on the denudation processes.

CHAPTER 3 Evolution of the Late Jurassic fault population in the Strathspey-Brent-Statfjord area, northern North Sea

3.1 Introduction

It is generally accepted that normal faults grow by processes of radial propagation and segment linkage (e.g. Cartwright *et al.* 1995). An early population comprising a large number of small faults will evolve into a late population of a small number of large displacement faults (Cowie 1998, Gupta *et al.* 1998). What remains poorly understood, however, is the temporal and spatial distribution of the active fault population throughout a rift event and the timing and rates of the propagation and linkage processes. Through investigation of a subsurface dataset, this study aims to address these problems and constrain the history of a >60 km long linked normal fault array: The Late Jurassic Strathspey-Brent-Statfjord fault system, located in the northern North Sea basin (Fig. 3.1).

Numerous studies (e.g. Leeder and Gawthorpe 1987, Prosser 1993, Anders and Schlische 1994, Gawthorpe *et al.* 1994, Contreras *et al.* 1997) have shown that the primary control on accommodation generation within an extensional basin is tectonic subsidence in the hangingwalls to normal faults, rather than fluctuations in eustatic sea level. Assuming the rate of sediment supply keeps pace (or exceeds) the rate of accommodation generation, the pattern of syn-rift sedimentation will document the coeval growth of the fault population. This premise is utilized in this study and, through detailed mapping of the seismic stratigraphic architecture of discrete, biostratigraphically defined packages in the syn-rift succession and examination of syn-rift sedimentology, the distribution and magnitude of accommodation space generated in the hangingwall to the Strathspey-Brent-Statfjord fault system through the rift event are reconstructed. A detailed examination of the distribution and sedimentology of the syn-rift sediments is presented in Chapter 4; addressing the balance between rate of sediment supply and rate of tectonic subsidence, hence qualifying the influence of fault activity on sedimentation. The principal conclusions of this work are that tectonic subsidence is indeed the primary control on accommodation generation and that the basin is near sediment filled during the initial 10 Ma of extension. The sedimentary evidence is given only cursory explanation in this chapter; the stratigraphic framework employed is that described in Fig. 4.2 (also Enclosure 2).

In addition to mapping syn-rift sediments, seismic interpretation is employed to describe the distribution and geometries of Late Jurassic faults in the study half-graben. Through examination of variations in displacement along the strike of the faults and the displacement-length scaling relationships of individual palaeo-segments, the mapping of abandoned palaeo- fault tips and the stratigraphic evidence of the timing of fault activity, the history of segment growth and linkage of the fault populations is constrained and quantified. A schematic restoration of the growth of the Strathspey-Brent-Statfjord fault system is proposed, with significant implications for understanding the distribution and rate of slip during the growth of this large normal fault system.

3.2 Regional setting

The fault system investigated in this study, the Strathspey-Brent-Statfjord fault, is located on the western flank of the northern North Sea rift basin. The northern North Sea basin, one arm of the trilete failed North Sea rift system, experienced ~15% extension during the Bathonian – Volgian (Yielding 1990, Roberts *et al.* 1993). The majority of the strain is accommodated on large, linked normal fault arrays with displacements typically in excess of 1 km. These trends strike N-S in the southern part of the basin and NNE-SSW in the north. Large faults occur at 10-20 km spacing. Regionally, the faults describe a graben-like geometry around a central low, the North Viking Graben.

The Strathspey-Brent-Statfjord fault is observed within the study area to be a single strand, throughgoing fault composed of a hierarchy of linked palaeo-segments (Fig. 3.2). Outwith the study area, the southward continuation of the fault bounds the eastern margin of the North Alwyn structure (Fig. 2.11b). To the north the tip overlaps with the Inner Snorre fault, a major >90 km long fault array with a throw locally in excess of 2 km (Underhill *pers.comm.* 1999). Dawers *et al.* (1999) and Dawers and Underhill (2000) have examined the Statfjord East fault, the northmost segment of the Strathspey-Brent-Statfjord fault system (Fig. 3.2). They show that the fault grew from the linkage of a number of shorter fault segments and that the growth of the fault controlled syn-rift sedimentation.

During Late Jurassic rifting, the northern North Sea basin can be demonstrated to have largely had a fully submarine setting. Tectonic subsidence following thermal doming (Underhill and Partington 1993, 1994) forced the retreat of the pre-rift Middle Jurassic Brent delta complex (e.g. Helland-Hansen *et al.* 1992); the final, retrogradational package of the delta complex, the Tarbert Formation, being the only non-marginal marine sediments of the syn-rift (e.g. Ravnås *et al.* 1997, Davies *et al.* in review; Fig. 4.2). Subsequent syn-rift sediments in the study area, the Heather and Kimmeridge Clay (Draupne) formations, are generally described as the deposits of a deep marine environment. Consequently, the basin has been interpreted to have been sediment starved, tectonic subsidence rapidly outpacing sediment supply during much of the extensional event (e.g. Ravnås and Steel 1998). In contrast, sand-rich sediments of the latest syn-rift (upper Draupne Formation) have been shown in the Statfjord East area to be shallow marine in origin (Nøttvedt *et al.* 2000). This is interpreted to be a consequence of the uplift of the Snorre footwall above sea level (e.g. Dahl and Solli 1993), the crest then being eroded and shedding coarse grained sediment southwards (Dawers *et al.* 1999, Nøttvedt *et al.* 2000). Yielding *et al.* (1992) have suggested that only the largest and widest of the fault blocks (e.g. Snorre, Gullfaks) were uplifted to, or above, wavebase and, hence, substantially eroded. The Strathspey-Brent-Statfjord footwall is considered to be a medium sized fault block, which remained submarine throughout the rift event.

3.3 Dataset and methods

The area of investigation is defined by the areal extent of five 3D migrated seismic surveys (Fig. 3.1). These surveys cover the Strathspey (*strathspey_TOMS*), Brent (*brent_d*, *lm96083*), Statfjord (*st9101*) and Statfjord East (*e86*) hydrocarbon fields, which all have oil accumulations in pre-rift sediments in the footwall to the Late Jurassic normal fault system. The data were interpreted using GeoQuest IESX software at the University of Edinburgh. The vertical and lateral resolution of the seismic data is discussed below. Further details on the dataset and the density of seismic interpretation are included in Appendix 1.

The total area of coverage of the five surveys is 965 km^2, but of this only ~350 km^2 is on the downthrown side of the bounding fault system. Data quality in the hangingwall area is typically poor, certainly with respect to the footwall area, largely due to the burial depth (up to 4000 ms TWT) of the syn-rift sediments of interest (see discussion below). Hence, while the main fault system and overall syn-rift sedimentation patterns were examined over the entire area, detailed seismic investigation of the faulting in the hangingwall (Section 3.4.2) was only undertaken in the north. In this area (illustrated in Fig. 3.1), the depth to the top pre-rift reflector is <3250 ms TWT and the seismic data are generally of higher vertical resolution.

Fig. 3.3 presents a typical W-E oriented seismic cross section across the Strathspey-Brent-Statfjord fault block, illustrating the interpretation of the key seismic reflectors and the form of faults mapped in the footwall and hangingwall to the main Strathspey-Brent-Statfjord fault array. Sub-parallel westward dipping reflectors characterise the pre-rift sediments in the footwall to the fault; truncated in the palaeo- footwall scarp. Four reflectors were mapped in the footwall, including a marked onlap surface interpreted as the top pre-rift reflector and corresponding, from well ties, with near the top of the Brent Group. In the hangingwall, the deepest reflector mapped is also the top of the pre-rift. This onlap surface separates the strongly defined, gently dipping reflectors of the pre-rift from the seismically transparent syn-rift geologies. Within the syn-rift three key reflectors were mapped: Top Tarbert Formation, top Heather Formation and top Kimmeridgian. These reflectors, and the strong positive (red) waveform of the top syn-rift (base Cretaceous) reflector, define chronostratigraphically significant packages with markedly different seismic stratigraphic character. These geologies, and those of the earliest post-rift, are described further in Chapter 4.

The crest of the footwall of the Strathspey-Brent-Statfjord fault array is substantially denuded. Hence, in order to measure the throw on the fault, the eroded geologies had to be reconstructed by extrapolating both the trace of the fault plane and the interpretation of the top pre-rift reflector mapped in the footwall (see Fig. 3.3). The error in defining the depth of the pre-erosion, palaeo- footwall crest is estimated to be ~25 ms TWT. The vertical distance (in ms TWT) between the truncations of the reconstructed top pre-rift reflector in the footwall and top pre-rift reflector in the hangingwall was measured. With the depth axis of the seismic data in time, the dip of the fault is uncertain; hence, throw is measured and consistently used as a proxy for displacement in this work. It is noted, however, that a fault dip of 60° would produce an underestimate of displacement of ~15 % (for throws of 1.5 km). The length over which displacement on the Strathspey-Brent-Statfjord fault system was measured is shown in Fig. 3.2, the spacing of the throw measurements varied between 62.5-250 m. Data pertaining to the construction of displacement-length profiles are described in Appendix 2. The vertical resolution of fault throw and the lateral resolution of fault trace are discussed below.

Seismic interpretation was based on seismic-well ties and the extrapolation of key surfaces. The hangingwall of the Strathspey-Brent-Statfjord fault system is penetrated by four vertical exploration wells, 211/29-8, 211/29-9, 211/30-1 and 33/9-18, and one deviated production well, 33/9-C27. Furthermore, some 7 km beyond the southern extent of the seismic coverage of this study, well 3/10b-1 intersects the hangingwall of the North Alwyn structure (Fig. 3.1). All these wells have cored intervals, and core samples, biostratigraphy reports and electrical log data from the six wells were used in this study. Core data were available for wells 33/9-18 and 33/9-C27. Additionally, in the north of the study area, seven wells have been drilled into the hangingwall of the Statfjord East structural block. The log responses and cored intervals from the syn-rift sediments in these seven wells have been previously described by Dawers *et al.* (1999), Nøttvedt *et al.* (2000) and Davies *et al.* (in review). Composite logs and original core reports from these wells were available to this study, and reviewed in the context of the results of previous workers. Syntheses of the well data are presented in Chapter 4 and further details in Appendix 3.

Vertical and lateral resolution of seismic reflectors and faults

The vertical resolution of seismic data, that is the minimum distance between two interfaces needed to give rise to a single reflection in seismic section, is primarily controlled by the wavelength of the source signal. The shorter the wavelength (the higher the frequency) the higher the vertical resolution. Resolution will typically decline with depth due to an attenuation of the higher frequency components of a source signal, and an increase in acoustic velocity resulting in increased wavelengths. In addition, filters applied during seismic processing commonly remove the highest frequency components of the waveform. The vertical resolution of the data employed in this study is approximately 10-30 ms two-way-travel time (TWT) in the hangingwall to the Strathspey-Brent-Statfjord fault array.

Seismic energy travels as a wavefront and the region on the reflector where energy is reflected constructively is known as the Fresnel Zone. The radius of the Fresnel Zone determines the lateral resolution of the data. For migrated (3D) data, the resolution is a function of the trace spacing, the length of the migration operator, the time/depth of the reflector and the bandwidth of the data. The line spacing of these five surveys is 12.5-25 m for both W-E/WNW-ESE oriented inlines and N-S/NNE-SSW oriented crosslines.

In general, normal faults with a throw of less than one quarter of the seismic wavelet are difficult to resolve on seismic data. From this, the minimum resolution of faults defined in this study is a maximum throw of ~10 ms TWT. Consequently, the interpretation of seismic data fails to resolve both a population of faults with sub-seismic throws, and the tips of faults where displacements are sub-seismic. Previous studies have addressed these limitations by introducing predictions based on the scaling relationships (size-number and displacement-length) of normal fault populations (e.g. Badley *et al.* 1990, Pickering *et al.* 1997). Such 'corrections' are not employed in this study. Faults with a maximum throw less than the resolution of the data are ignored. Obviously, without being able to accurately map fault tips and with seismic line spacings of 12.5-25 m, the interpreted fault traces are too short. A realistic reconstruction of the tips of each fault is outwith the scope of this study.

3.4 Fault populations

Two populations of normal faults are described from the study area, distinguished by the size of the faults and the period of time, during the rift phase, when they were active. Both populations developed in response to the same regional stress regime in the Late Jurassic. The two populations are described separately below.

3.4.1 The character of the Strathspey-Brent-Statfjord fault

Location and form of fault segments

The population of faults described in this section is related to the lateral propagation and growth of the largest single fault in the study area – the Strathspey-Brent-Statfjord fault. This population can be divided into two sub-populations: (1) the main fault itself, and (2) a group of ten splays from this fault, which are located at between 5-12 km spacing along the length of the Strathspey-Brent-Statfjord fault (Fig. 3.2). These splays are found in the immediate hangingwall to the main fault.

The Strathspey-Brent-Statfjord fault is the most significant fault in the study area, in terms of length and displacement (Fig. 3.2). This 62 km long structure is a continuous, thoroughgoing normal fault strand with a maximum displacement in excess of 1.5 km. The southern tip of the fault is not covered by the seismic data available to this study. Hence, the total length of the fault is >>62 km. In the north, the fault splits into two segments – the Statfjord (west) fault and the Statfjord East fault (Fig. 3.2). The Statfjord (west) fault tips out within the area of the seismic coverage, the Statfjord East fault ~2 km to the north of the seismic coverage (Dawers and Underhill 2000).

The observed cross-sectional geometry of the fault plane is variable but, typically, near planar (Fig. 3.3). Curvature observed at depth in some areas is largely attributed to an increasing bulk rock velocity and/or to the cross section being orientated obliquely to the strike of the fault plane, rather than the fault being listric in form. The two northern segments of the Strathspey-Brent-Statfjord fault are also considered to be planar (Fig. 3.4). The two cross sections in Figs. 3.3 and 3.4 demonstrate the footwall of the Strathspey-Brent-Statfjord fault to be substantially denuded along its entire length, and the crest to be backstepped by 2-3 km from the surface trace of the fault (see Chapter 6).

The fault can be mapped along the entire N-S length of the study area, striking N-S in the south (Strathspey and Brent areas), rotating to a NNE-SSW trend to the north (Statfjord area). The change in the strike of the fault corresponds with an ~4 km eastward step of the fault trace when mapped at top pre-rift level (annotated on Fig. 3.2). The fault trace is not linear in map view, but rather has a multiple cuspate form. Low amplitude arcs come to apexes spaced at 4-7 km (see description on Fig. 3.2). It is at the apexes of some of these cusps that the ten mapped fault splays intersect with the Strathspey-Brent-Statfjord fault (Fig. 3.2). The splays arc out into the hangingwall basin from the main fault with, in general, a NE-SW strike. They are up to 3 km in length, with a maximum throw in excess of 250 ms TWT.

Displacement-length profiles, based on the throw of the top pre-rift reflector by the splays, describe a steep slope (Fig. 3.5). Maximum displacement occurs at the southwestern end of the faults, where they merge with the Strathspey-Brent-Statfjord fault. Displacement decreases along the length of the splays, typically to zero within 3 km. Such displacement-length profiles are incompatible with the 'bell-shaped' profiles described from single, isolated normal faults (e.g. Peacock and Sanderson 1991, 1994, Dawers *et al.* 1993) and predicted by fault models (e.g. Cowie and Scholz 1992). The faults rather have the form of just one half (or less) of a fault, i.e. a fault tip.

In seismic cross section perpendicular to strike, the vertical form of the splay faults is near planar (Fig. 3.6). Although demonstrated in map view to intersect with the main fault, splay faults are significantly downthrown in the hangingwall to the main fault. The vertical trace of the splays cannot be mapped through the entire syn-rift (as demonstrated in Fig. 3.6); indeed the splays rarely offset the top Heather Formation although the formation is substantially thickened in the hangingwalls (Fig. 3.2). Hence, these small faults are concluded to have only been active early in the rift event, between Early Bathonian – Late Oxfordian (see Figs. 3.2 and 3.6 and later discussion).

The observations of fault location and form presented above suggest that the splay faults represent the palaeo-tips of normal faults, which, when two overlapping segments 'hard-linked', were abandoned in the hangingwall to the new longer fault (i.e. 'footwall breaching' model of Cartwright *et al.* 1996 reproduced in Fig. 2.3b). This is compatible with the interpretation of these structures in the Statfjord East area by Dawers *et al.* (1999) and Dawers and Underhill (2000). The implication of this conclusion is that the Strathspey-Brent-Statfjord fault grew from the linkage of at least eleven shorter segments of 7-15 km length, whose locations are defined by the 'cusp-shaped' form of the fault trace and the abandoned tips now in the hangingwall. Hence, the main fault and the sub-population of splays can be considered as part of an integrated fault system associated with the growth of the Strathspey-Brent-Statfjord fault.

Observations from displacement-length profiles

The conclusions presented above are supported by observations from the displacement-length profiles of faults forming the Strathspey-Brent-Statfjord fault system. Fig. 3.7 shows the individual profiles for the main fault and each of the ten described splay faults, plotted along a 65 km long horizontal axis parallel to strike (i.e. N-S oriented in the south, NNE-SSW in the north, see Fig. 3.2). These profiles measure the throw of the top pre-rift reflector, as illustrated in Fig. 3.3.

The displacement-length profile of the Strathspey-Brent-Statfjord fault (including the summed displacements of the Statfjord (west) and Statfjord East fault segments) is shown to have a truncated 'bell-shaped' form (Fig. 3.7). Maximum displacement of 1552 ms TWT is located 12.5 km from the south of the study area (point 'A' on Fig. 3.2). D_{max} is the apex of the central part of the fault where displacement is >750 ms TWT for ~48 km. Two kilometres north of the fault splitting into two strands, the displacement begins a steep decline to zero within 10 km. A similar tip profile is predicted in the south. This profile shape and the distribution of the displacement over multiple strands at the tips are compatible with previous descriptions of large, isolated normal faults (e.g. Ebinger *et al.* 1999, Morley 1999).

Although bell-shaped, the overall displacement-length profile of the Strathspey-Brent-Statfjord fault is not smooth; it is extremely serrated with vertical variations of up to 422 ms TWT. These displacement variations are not random, but rather define a pattern, with the component units repeating every 3-6 km. This pattern is broken down in Fig. 3.8a for the central (10-30 km length on Fig. 3.7) part of the fault. Local displacement minima bound each of the units described (Fig. 3.8a). The minimum values are between 102-422 ms TWT less than the unit maxima. Maxima are skewed 1-3 km from the centre. A further 2 or 3 local minima and maxima at 0.5-3 km spacing are observed within each unit, but typically only represent displacement variations of <150 ms TWT. It is observed from Fig. 3.7 that some of the major bounding minima correlate with the locations of abandoned palaeo-tips. Hence, in addition to the 7-15 km long segments previously recognised from the locations of abandoned tips (and supported by these observations), a number of shorter segments of 0.5-3 km length, from earlier still in the rift history, are recognised.

Significantly, and in contrast to the predictions of the conceptual model of Cartwright *et al.* (1995), the along strike displacement variations in these shorter segments have not been completely equilibrated during continued fault growth after linkage. Rather, these displacement variations appear to have been preserved, as an artifact of the geometry of the earlier fault population, in the displacement-length profile of the linked system. The serrations in the displacement-length profile of the Strathspey-Brent-Statfjord fault, between 3.5-49 km along strike length, were used to estimate the minimum length of earlier segments (see Fig. 3.8a). These are minimum estimates as it is accepted that segments do not link tip to tip but instead first grow into an overlapping geometry (e.g. Willemse 1997). Palaeo-tips abandoned in the hangingwall suggest that some of the >3 km long segments may have been up to 4 km longer pre-linkage. The displacement and length data interpreted for these palaeo-segments is shown in Fig. 3.8b as a plot, on log-log scale, of the minimum lengths against the artifact of maximum displacement (as preserved on the displacement profile of the linked fault). Also shown for comparison on Fig. 3.8b is the range of maximum displacement and length data for the global dataset, as discussed by Schlische *et al.* (1996) (Fig. 2.3).

The data are presented as two groups in Fig. 3.8b. The first represents the major palaeo-segments, i.e. the 'units' described in Fig. 3.8a, of minimum length >3 km. These are commonly associated with the abandoned tips and, hence, confidently interpreted as unique structures. The second group comprises the sub-units described in Fig. 3.8a, identified by local displacement minima within the larger units. The minimum length of these faults is 0.5-4 km, and, without other supporting evidence, there is less confidence in their origin as unique faults. The whole dataset does, however, show a distinct trend, sub-parallel to, and clustering around, the trend of the global dataset. Error bars demonstrate that underestimation of segment length is skewing the population to the left; displacement is also being underestimated as the actual measure is throw. The ratios of displacement to length for this population are comparable to those of similar length faults found in the hangingwall discussed later.

These observations are surprising. They indicate that displacement variations along the length of a normal fault system developed while the fault is in a pre-linkage, multi-segment stage of its growth are *not* lost, and indeed show little equilibration at all, as the fault links and grows. The reason for the lack of equilibration of displacement after linkage to support the longer fault length is unclear. A preliminary conclusion maybe that it reflects the immaturity of the fault system, that it was 'switched off' while in the process of linkage and never had the opportunity to accumulate vertical displacement to mirror its along strike length. This conclusion is, in part, supported by the observation that the maximum displacement of the whole fault system (~1.7 km) is lower than would be predicted for its length (for faults in the Basin and Range $D_{max} \sim 0.03L$, and the maximum displacement on a fault of 90 km length is usually in excess of 2.5 km, Dawers *pers. comm.* 1999). The implications of these observations and conclusions for understanding the distribution and rate of displacement on this fault during growth are considered in Section 3.6.

On the timing of segment linkage

The observations presented in the preceding sections demonstrate that the Strathspey-Brent-Statfjord fault grew by a mappable process of segment linkage. However in order to appreciate the implications of constraining fault growth patterns for understanding both the topographic evolution of the basin and sediment dynamics, a temporal framework is required. The timing of activity on the main fault and the abandoned splays can only be constrained through examination of the seismic stratigraphic architecture of syn-rift sedimentation. Specifically, recognising thickening of sediments into active faults and sedimentation across the trace of inactive structures. Thickness variations within the syn-rift are described in some detail in Chapter 4. However, it is observed that the study area was sediment starved during the latter stages of the rift phase; hence, hangingwall accommodation space was unlikely to have been completely filled before faults became inactive.

All of the lithostratigraphic units described within the syn-rift (Tarbert, Heather and Kimmeridge Clay formations) are observed to thicken into the Strathspey-Brent-Statfjord fault (Fig. 3.3). Hence, the fault was active throughout the period of extension (Early Bathonian – Late Volgian) – approximately 20.5 Ma. In order to achieve 1.5 km of displacement, this would equate to an average displacement rate of 73 mm/ka. However, the magnitude of syn-rift thickening into the fault is extremely variable along the length of the structure and for different intervals within the syn-rift. Assuming that sediment supply was constant along the fault at any given time, this suggests that the rate of displacement and, hence, the volume of accommodation space created in the hangingwall to this fault were temporally and spatially variable throughout the rift event; as would be predicted by a model of fault growth by segment linkage.

Fig. 3.9 compares the thickness of the syn-rift succession on two seismic inlines oriented perpendicular to the fault strike. Fig. 3.9a is a seismic line across the location of maximum displacement on the fault, where accommodation space would be predicted to be at a maximum. The syn-rift is >650 ms TWT thick in this area and thickens into the fault throughout the succession. In contrast, Fig. 3.9b is a line across the eastward step in the fault trace between the Statfjord and Brent area where a major splay fault is located in the hangingwall. The syn-rift in this area is substantially thinner; in the hangingwall to the splay fault it is ~225 ms TWT and in the hangingwall to the Strathspey-Brent-Statfjord fault it is has a maximum thickness of 425 ms TWT. In addition, it is observed that the older part of the syn-rift thickens into both the splay fault and the main fault, and that the younger part only thickens into the main fault (see isopach maps in Chapter 4). In Fig. 3.9b, the splay fault and the main fault were both active early in the syn-rift, but following fault linkage the main fault was the only active fault and sediments thickened into it. Across the study area, areas where the syn-rift is observed to be thin and have less pronounced thickening into the Strathspey-Brent-Statfjord fault; these areas correspond with the locations of abandoned splays and displacement minima. Using these observations and the age of the syn-rift, it is possible to conclude when splay faults became inactive, i.e. linkage occurred (annotated on Fig. 3.2), and to map the growth of the Strathspey-Brent-Statfjord fault.

Recognising that the main period of segment linkage of the Strathspey-Brent-Statfjord fault was during the early-mid Oxfordian, it is possible to crudely assess changes in the magnitude of slip rates on the fault. Prior to linkage of the major 5-7 km long segments (associated with the abandoned tips), this study demonstrates that the segments had at least 450 m displacement (from displacement artifacts, Fig. 3.8). This corresponds with a minimum rate of displacement of 45 mm/ka at the point of maximum displacement on the faults. However, after linkage the faults accumulated up to another 1.1 km of displacement before becoming inactive in the latest Volgian. To achieve this, average displacement rates would have to have been ~105 mm/ka. An increase in the rate of displacement of this order would correlate with the transition from the 'rift initiation' to 'rift climax' phase (after Prosser 1993). This transition has previously been ascribed by Gupta *et al.* (1998) to segment linkage.

3.4.2 Faults in the hangingwall of the Strathspey-Brent-Statfjord fault

A second population of normal faults was identified by seismic interpretation in the northern part of the study area (see Fig. 3.1). Faults were mapped in the hangingwall of the Strathspey-Brent-Statfjord fault where they offset the top pre-rift reflector (Fig. 3.10). The vertical resolution of the throw of these normal faults was ~10 ms TWT and they were mapped at 125 m (N-S) spacing. Lower resolution in the area of the *e86* survey is not due to the seismic data, but the poor seismic character of the reflectors of the latest pre-rift, which made it difficult to recognise offsets. A map of the fault population described is shown in Fig. 3.11.

Of a total of 53 seismically resolvable normal faults identified, 22 are antithetic to the main fault system and 31 are synthetic. Synthetic faults are distributed within 1-3 km of the main fault system and tend to define two sub-parallel trends, approximately 0.75 km apart. Antithetic faults, in contrast, form a distinct, multi-segment single trend some 3.5 km east of the main fault. It is noted that few faults are mapped within 1.5 km of the main fault system. The overall geometry of the faults mapped in the hangingwall is graben-like (Fig. 3.10), with the graben axis ~3 km to the east of, and parallel to, the Strathspey-Brent-Statfjord fault.

Faults of the hangingwall population range in (mappable) length from 125 m to in excess of 5 km and all strike near parallel to the Strathspey-Brent-Statfjord fault. The maximum displacement on the 53 faults is between 12-210 ms TWT. As in the description of the Strathspey-Brent-Statfjord fault, throw is used as a proxy for displacement. Displacement-length profiles of 39 faults from the population (numbered in Fig. 3.11), demonstrate that the majority have a maximum displacement of <50 ms TWT and are <2 km in length (Fig. 3.12). Of these 39 faults, only 5 have a maximum displacement in excess of 75 ms TWT. These 5 faults (antithetic faults 7, 16, 22, 24 and 30) are the longest segments of the antithetic fault trend (Figs. 3.10 and 3.11).

The resolution of the seismic data, the size of the faults and the sampling interval (every 125 m) does not permit detailed analysis of the displacement-length profiles of these faults. However, general observations of, for example, skewing of the profiles of overlapping segments (faults 4 and 8; Fig. 3.13a) and a distribution of displacement between faults with two overlapping strands (faults 22 and 24; Fig. 3.13b) can be made. In addition, the ratio of maximum displacement to length of the 39 faults is demonstrated to show reasonable correlation over two orders of magnitude (Fig. 3.14), with a best-fit linear relationship for the population of $D_{max} = 0.0357L$ (correlation coefficient, $R^2 = 0.663$).

By summing the displacement of the 39 faults, the total displacement accommodated by antithetic, synthetic and all the faults of this population, in an area 16.5 km long by ~5 km wide, can be determined (Fig. 3.12). It is noted that, due to the exclusion of sub-10 m throw faults and fault tips (below data resolution) and segments of the main fault system, these total displacements are underestimated. However, with these caveats in mind, the results show that displacement accommodated by all antithetic structures averages 100.9 ms TWT, exceeding that of synthetic structures (average 38.7 ms TWT) by 2.6:1. The total displacement accommodated by these faults across the area averages 137.95 ms TWT.

Similar to the work presented in the preceding section, the time of activity on these faults can only be determined indirectly, i.e. by seismic stratigraphic techniques. Fig. 3.15 shows a W-E oriented seismic inline from the Statfjord (*st9101*) survey that crosses the path of vertical exploration well 33/9-18. The line drawing interpretation in Fig. 3.15b shows lithostratigraphic picks derived from well 33/9-18 and also the interpretation of antithetic faults 24 and 30 and synthetic fault 28. Biostratigraphic interpretation of the sediments encountered in 33/9-18 is documented in Appendix 3. An antithetic fault and a synthetic fault are shown to offset the top Tarbert Formation reflector and, indeed, the Tarbert thickens across the faults (Fig. 3.15). However, the vertical trace of the faults die in the very earliest Heather Formation. Biostratigraphic dating of these formations indicates that the faults were active only during the Early – Middle Bathonian (~2-3 Ma), suggesting an approximate maximum displacement rate of 15.5 mm/ka for fault 28. All the synthetic faults in the population have an earliest syn-rift history (see also Fig. 3.10), and became inactive in the Middle – Late Bathonian.

In contrast, the antithetic fault 24 shown in Fig. 3.15 appears to have had a longer period of activity. Not only does the Tarbert Formation thicken into this fault, but also the Heather Formation thickens. The fault does not offset the top of the Heather Formation, and thickening in the overlying Kimmeridge Clay Formation is attributed to the effects of differential compaction. Allowing for sediment starvation in the basin (Chapter 4), the duration of activity on this fault is interpreted as earliest Bathonian – ?earliest Oxfordian (~10 Ma). This equates to an approximate maximum displacement rate of 21 mm/ka for fault 24. All the main antithetic fault segments (i.e. faults 7, 16, 22, 24, 30, 32, 36) are interpreted to have been active for this longer time period (see Fig. 3.10), as would be predicted by their greater length and displacement compared to the synthetic structures.

Not all the antithetic faults were active for this prolonged period of time, as demonstrated by fault 30; some (e.g. faults 3, 20, 23) can be shown to have become inactive with the synthetic population during the Middle – Late Bathonian. All the antithetic faults that stopped significant growth at this time are <3 km in length and do not form part of the main antithetic trend. Therefore, it is suggested that only those seismically resolvable faults forming the trend of the Strathspey-Brent-Statfjord fault and the trend of the main antithetic fault continued to accumulate significant displacement after the Late Bathonian.

The observations presented above demonstrate that a large population of normal faults nucleated and grew in the hangingwall of the Strathspey-Brent-Statfjord fault, early in the rift event. Of this population, the majority became inactive after 3-4 Ma. This majority were <2 km in length, mainly synthetic to the main trend and within 3 km of that structure. The faults that continued to grow were part of a multi-segmented antithetic fault array, located ~3.5 km east of the main fault. Faults of the antithetic trend also became inactive, after ~10 Ma, when segment lengths were up to 5 km.

Fig. 3.16 shows a 5.6 km wide, W-E oriented seismic inline and interpretation of the hangingwall basin. This view emphasises the graben-like form of the fault pattern and the dominance of the antithetic faults in the basin. This line also shows that the antithetic fault and the Strathspey-Brent-Statfjord fault meet at depth. For the study area, the depth at which the two faults meet is consistently in the range 900-1000 ms TWT (Fig. 3.17a, b). The geometry of this basin is comparable with the geometry of most other documented half-graben type continental rift basins (Fig. 3.17c). That is, an asymmetric basin bounded by a major normal fault system with several kilometres of displacement, segmented along strike. The hangingwall basin is commonly observed to be dissected by a major antithetic structure (e.g. Roberts and Jackson 1991, Hayward and Ebinger 1996, Morley 1999). The growth of the antithetic structure is restricted by the inability of active faults to cross (e.g. Jackson and McKenzie 1983). This study supports the simple model of rifting (summarised by Scholz and Contreras 1998, Fig. 3.17b) in which the rift is initially symmetric, with the formation of conjugate fault sets, but faults are locked by intersection at depth. Which of two faults reaches the intersection point first will control which continues to grow. In this fault system, the Strathspey-Brent-Statfjord fault is, thus, considered to have grown most rapidly and 'switched off' the antithetic faults by locking at depth.

3.5 The tectonic evolution of the half-graben bounded by the Strathspey-Brent-Statfjord fault

An interpretation of the tectonic evolution of the half-graben bounded by the Strathspey-Brent-Statfjord fault is proposed based on the observations from the two populations of faults identified in the basin. By integrating structural observations with the stratigraphic architecture of syn-rift sediments (described in detail in Chapter 4) this study reconstructs the temporal and spatial distribution of the active fault population throughout a rift event and the timing and rates of the propagation and linkage processes. An early population comprising a large number of small faults developing into a late population of a small number of large displacement faults is demonstrated, supporting the interpretations of Cowie (1998) and Gupta *et al.* (1998). The transition from the 'rift initiation' to 'rift climax' phases is shown to have occurred ~10 Ma into the rift event, coincident with the main period of segment linkage of the basin bounding fault system and the 'switching off' of other faults.

A large number of faults are recognised to have been active in the Strathspey-Brent-Statfjord area during the earliest stages of rifting, including segments of the proto- Strathspey-Brent-Statfjord fault. Accommodation generated by tectonic subsidence in the hangingwalls to these faults controlled facies distribution and the loci of deposition of the coarse grained sediments of the Tarbert Formation (Chapter 4). Strain was accommodated on many fault strands of sub-parallel strike but no preferred dip and rates of displacement were low. A large number of the faults comprising this initial population became inactive very early in the rift history, approximately 2-3 Ma after their nucleation – coincident with the drowning of facies of the Tarbert Formation in the study area.

Faults that remained active during deposition of the Heather Formation defined two parallel trends. Each fault trend was composed of multiple segments, but the two arrays had an opposing dip and defined a graben-like structure. Palaeo-segments of the Strathspey-Brent-Statfjord fault comprised the western boundary of an ~3 km wide graben. Seismic mapping of the Heather Formation and examination of fault scaling relationships suggest that fault segments at this time were up to 5 km in length and maximum displacements exceeded 200 ms TWT. Strain localisation on fewer faults resulted in an increase in the rates of displacement.

Intersection of the planes of the graben bounding faults at depth in the crust resulted in the 'switching off' of faults of the antithetic trend and the palaeo-segments of the Strathspey-Brent-Statfjord fault became the only active faults in the basin. Subsequently, the geometry of the basin evolved from graben-like to a half-graben. Stratigraphic evidence indicates that the antithetic faults became inactive ~10 Ma into the rift event. It was also around this time that linkage of the 7-12 km long segments of the Strathspey-Brent-Statfjord fault occurred to form the single throughgoing strand. The increase in rates of displacement (hence tectonic subsidence) at this time is interpreted as the transition from the 'rift initiation' to 'rift climax' phases in this basin.

From the along strike variations in the displacement-length profile of the Strathspey-Brent-Statfjord fault the maximum displacements on the palaeo-segments prior to linkage are interpreted to be up to 450 ms TWT. The implication of this is that the majority of the displacement on the structure accumulated following linkage, when the fault accommodated all the strain in the basin. The sedimentology and distribution of the Kimmeridge Clay Formation demonstrates that the basin was sediment starved during this rift climax phase.

3.6 Discussion

A model to explain the accumulation of displacement on the Strathspey-Brent-Statfjord fault system

This study has demonstrated that the throughgoing Strathspey-Brent-Statfjord fault originally consisted of an array of shorter palaeo-segments, which propagated and linked to form a single strand. Seismic mapping of the fault, identification of abandoned palaeo-tips and an examination of fault scaling relationships enable the locations of a hierarchy of palaeo-segments to be accurately defined and the timing of linkage established. The linked fault system is shown to be 'immature' in terms of displacement-length scaling. Although the entire fault is not covered by the available data, the fault is clearly much greater than 62 km in length but the observed maximum displacement is only ~1.7 km. Additionally, large variations in displacement are preserved along the strike of the fault; interpreted as remnants of displacement variations developed when the fault was a number of palaeo-segments prior to linkage.

Interpretation of these observations suggests that the Strathspey-Brent-Statfjord fault did not grow by simple radial propagation of segments, but rather by the combined processes of tip propagation and segment linkage. In order to explain the location and rate of displacement accumulation on the Strathspey-Brent-Statfjord fault, the modified earthquake rupture model proposed by Roberts (1996) from studies of active normal fault systems in Greece is employed. This model assumes that during multiple seismic events the centre of an isolated fault will be ruptured more often than the tips and accumulate greater slip (Fig. 3.18a); the fault develops a traditional 'bell-shaped' profile. For faults with interacting local stress fields, a stress feedback mechanism results in a concentration of seismic events in the tip areas (King *et al.* 1994, Cowie 1998). In this second model (Fig. 3.18b), the tips have an increased rate of displacement (with respect to the centre of the fault) and skewed displacement-length profiles result. This model is simplistic; assumes 'healing' of the fault plane between rupture events and does not incorporate a reloading of the centre of the fault following rupture at the tip (as proposed by Cowie 1998). In addition, rupture lengths are considered to be shorter than the total length of the fault (*cf.* >100 km long single slip event proposed by Jackson and Blenkinsop 1997). It is recognised that the mechanics of slip are complex and a function of a number of parameters, significantly rheology, the geometry of the fault population and rupture repeat times.

Using this model the growth of the Strathspey-Brent-Statfjord fault system is reconstructed in Fig. 3.19, based on the observations from between 10-30 km along the strike of the fault (the displacement-length profile shown in Fig. 3.8a). The propagation and linkage of five segments of the fault system are shown (a), with the implications for displacement-length profiles and a log-log plot of maximum displacement against length for the growing fault 1 illustrated for each of the 3 stages, in (b) and (c) respectively. The locations of the palaeo-segments on the linked fault system are shown for stage 3.

In stage 1 (Fig. 3.19), five fault segments of between 3-5 km in length were observed. Each of these faults was a linked array of shorter segments – as demonstrated by along strike variations in the displacement profile. However, the faults can be considered, for the purposes of this model, to have been isolated, and, hence, conformed to the maximum displacement-length scaling relationship, $D_{max} = c.L^n$. The cumulative *rates* of displacement (illustrated by the grey regions above the fault profiles in (b) – not on the same vertical scale for all the faults) were greatest at the centre of the faults and the profiles were 'bell-shaped' (*cf.* Fig. 3.18a).

Lateral propagation of the five faults caused the modified stress regimes around the faults to overlap. Interaction between the fault strands (Fig. 3.18b) was the primary control on rates of displacement accumulation in stages 2 and 3. Prior to the linkage of the segments, skewed displacement-length profiles were established and the ratio of maximum displacement to length of fault 1 increased (compared to an isolated fault). Following linkage, these features were preserved as artifacts of the growth process on the displacement-length profile of the fault array and record the history of fault interaction.

In general, hard linkage of two faults, breaching a relay ramp and abandoning tips to form a single strand, will result in an increased segment length. As the maximum displacement does not increase instantaneously D_{max}:L will sharply decline at the time of linkage. The position of linkage is described by a displacement minimum, typically near the centre of the linked fault. Observations from palaeo-segments of the Strathspey-Brent-Statfjord fault system show that these minima, defining palaeo- segment boundaries, have been preserved and did not equilibrate with continued displacement accumulation on the fault. If a linked fault array grew in isolation following linkage then displacement would accumulate as described in Fig. 3.18a and, if the segment boundary minimum was at the centre of the structure, the size of the minimum would slowly decrease (as a proportion of the total displacement). The model illustrated in Fig. 3.18 also demonstrates how the magnitude of the linkage minimum could be increased post-linkage; if the linked fault was growing under the control of stress feedback mechanisms (i.e. interacting with another fault), and displacement preferentially accumulated at the fault tip rather than at the centre of the fault. A concentration of rupture events in the underdisplaced site of the palaeo- relay ramp is not supported by the observations from the Strathspey-Brent-Statfjord fault array or the predictions of the model, and indeed in this study area enhancement of the displacement minima appeared to have been the norm. Observations in the study area suggest that stresses due to the high displacement gradients of the overlapping tips were accommodated by the initiation of fault perpendicular structures, rather than enhanced slip on the main fault strand (the footwall faults described in Chapter 6).

3.7 Conclusions

1. The Strathspey-Brent-Statfjord fault is a >>62 km long linked array composed of a hierarchy of palaeo-segments which propagated and linked to form a single throughgoing strand. The locations of palaeo-segments are identified by abandoned palaeo-tips, displacement variations along the strike of the fault, and the stratal architecture of syn-rift sediments in the half-graben. Stratigraphic evidence also permits the timing of linkage to be constrained.

2. The displacement-length profile of the Strathspey-Brent-Statfjord fault has a truncated 'bell-shaped' form but is extremely serrated along strike. Displacement maxima and minima are interpreted to reflect displacement variations established when the fault was a number of shorter segments, and it is shown that these variations have not been equilibrated following linkage.

3. A dense population of faults is described in the hangingwall to the Strathspey-Brent-Statfjord fault array. These faults have a graben-like geometry, with the axis ~3 km east of the main fault system. The majority of the faults in the hangingwall are small (of maximum displacement <50 ms TWT and length <2 km) and were only active early in the rift history (first 2-3 Ma). The largest faults (up to 7 km length and >200 ms TWT maximum displacement) are antithetic to the Strathspey-Brent-Statfjord fault, and bound the eastern margin of the graben. These larger faults became inactive after ~10 Ma.

4. The geometry of the basin evolved with the growing fault system. The initial population of faults was dense, sub-parallel in strike, but with no preferred dip direction. Within 3-4 Ma of the initiation of rifting, a graben-like form composed of parallel fault systems was well established. A graben geometry was maintained until ~10 Ma into the rift event when the majority of the faults in the hangingwall became inactive and strain was accommodated fully on the Strathspey-Brent-Statfjord fault system. This transition corresponds with the linkage of the fault to become a throughgoing single strand. As the number of active faults in the basin decreased through the rift phase, so the strain was accommodated on fewer faults and displacement rates increased.

5. A modified rupture model (after Roberts 1996) is used to interpret the accumulation of displacement on the Strathspey-Brent-Statfjord fault array. The model suggests that interactions between the local stress fields of individual fault strands were the primary control on the location of displacement accumulated during linkage.

CHAPTER 4 Controls on Late Jurassic syn-rift sedimentation in the Strathspey-Brent-Statfjord area, northern North Sea

4.1 Introduction

Tectonic subsidence has been shown by numerous studies to be the primary control on the generation of accommodation space within extensional basins (e.g. Leeder and Gawthorpe 1987, Schlische 1991, Gawthorpe *et al.* 1994, Contreras *et al.* 1997). In particular, the transition in a basin from slow rates of subsidence, during the 'rift initiation' phase (as defined by Prosser 1993), to high rates of subsidence, during the 'rift climax', has been attributed to the growth of fault arrays by processes of fault interaction and linkage (Gupta *et al.* 1998). The propagation of normal fault segments is also responsible for the growth of extensional fault related folds, responding to along strike variations in fault displacement. These folds define isolated depocentres adjacent to fault segments which, with segment linkage, subsequently amalgamate (Schlische 1995). Consequently, the temporal and spatial evolution of the fault population has a major impact on the stratigraphic architecture of extensional basins (Schlische and Anders 1996, Morley 1999, Dawers and Underhill 2000).

The coeval generation and filling of accommodation space requires the rate of sediment supply to the basin to equal, or exceed, the rate of subsidence. Similarly to the rate of growth of the fault population, the input of sediment to a basin also demonstrates temporal and spatial variability, reflecting changes in, for example, the climate and the size, geomorphology and geology of catchment areas (e.g. Ahnert 1970, Leeder *et al.* 1998). Hence, the stratigraphic record of syn-rift sedimentation in a rift basin typically documents the interaction between the rate at which accommodation is generated and the rate of sediment supply (see Schlager 1993).

In this chapter the results of an investigation into the distribution and character of sediments in a marine half-graben type basin are presented. This study is unique in that the evolution of the sedimentary dynamics of the basin is integrated with an understanding of the temporal and spatial growth of the normal fault population (described in detail in Chapter 3). Hence, the relative importance of the rates of accommodation generation (from both tectonic subsidence and sea level fluctuations) and sediment supply in controlling the distribution and sedimentology of syn-rift sediments can be qualified and quantified.

The basin studied is the Late Jurassic half-graben bounded by the >60 km long Strathspey-Brent-Statfjord fault system in the northern North Sea (Fig. 4.1). The work has employed a subsurface dataset, using both seismic stratigraphic techniques (to interpret five 3D seismic surveys) and sedimentological and petrophysical investigations of well data. Examination of the tectonic evolution of the basin (Chapter 3) has demonstrated that the fault population grew from a large number of small (both in length and displacement) faults during the rift initiation phase to a small number of large faults during the climax of the rift event. In addition, the basin geometry changed after ~10 Ma from a graben-like form, developed during the earliest stages of rifting, to a half-graben geometry. The implications of these observations for both the topography generated during rifting, the tectonically generated accommodation in the basin and the locations and volumes of potential sources of sediment are integrated into the interpretation of the dynamics of the sedimentary systems in the study area. Through examination of key, chronostratigraphically significant, packages of syn-rift sediments the history of deposition in the basin is deduced. From this, changes in the source, supply and transport routeways of both internally and externally derived sediments can be demonstrated throughout the rift phase; changes directly related to the accommodation in the basin. The implications of these results for hydrocarbon exploration in rift basins are examined in Chapter 9.

4.2 Geological Setting

The Strathspey-Brent-Statfjord area is located in the East Shetland Basin of the northern North Sea, on the western flank of the Late Jurassic Viking Graben (Fig. 4.1). The northern North Sea is a failed rift basin, one arm of the trilete North Sea rift. W-E oriented extensional stresses during the Late Bathonian – latest Volgian resulted in approximately 15 % strain across the basin (e.g. Roberts *et al.* 1993), mainly accommodated on large, sub-parallel normal fault systems at 10-30 km spacing (see Enclosure 1). In the East Shetland Basin, the footwalls of easterly dipping fault arrays form simple structural hydrocarbon traps, including, in the study area, the Strathspey, Brent, Statfjord and Statfjord East fields –

all of which are bounded by a single strand, throughgoing normal fault, the Strathspey-Brent-Statfjord fault.

The Mesozoic stratigraphy of the study area is illustrated in Fig. 4.2 (also Enclosure 2), following the nomenclature for the Jurassic proposed by Richards *et al.* (1993). Pre-rift geologies comprise the coarse grained fluvial geologies of the Statfjord Formation of the Banks Group (e.g. Fig. 13 of Steel and Ryseth 1990), the mud-rich, shallow marine deposits of the Dunlin Group (Marjanac 1995, Parkinson and Hines 1995) and the sand-rich, marginal marine geologies of the Brent Group (Budding and Inglin 1981, Graue *et al.* 1987). In the study area, these geologies are of near constant total thickness (~430 m for the Brent and Dunlin groups). These sediments represent a period of near continuous deposition during the rapid post-rift subsidence following Permo-Trias extension. Superimposed on the pattern of thermal subsidence was the influence of the North Sea dome in the Central North Sea during the Early – Middle Jurassic (Underhill and Partington 1993, 1994). Uplift of the dome provided a source of sediment in the hinterland of the Brent fluvial system and was, in part, responsible for the rapid progradation of the delta complex.

The geologies of the Brent Group are traditionally interpreted as a regressive-transgressive wedge with the Broom, Rannoch, Etive and Ness (BREN) formations being the progradational package (e.g. Mitchener *et al.* 1992). The initiation of faulting is described, stratigraphically, by the drowning of the Brent delta (e.g. Johannessen *et al.* 1995, Råvnas *et al.* 1997, Underhill *et al.* 1997). Hence, the youngest depositional unit of the Brent Group – the transgressive Tarbert Formation – defines the base of the syn-rift in the study area (see Davies *et al.* in review). The remainder of the syn-rift package, the Humber (Viking) Group (sub-divided into the Heather and Kimmeridge Clay (Draupne) formations) is dominated by mud and silt grade sediments of marine origin (e.g. Rattey and Hayward 1993); considered indicative of basin starvation. However, in the north of the study area, Dawers *et al.* (1999), Dawers and Underhill (2000) and Nøttvedt *et al.* (2000) record sand-rich sediments, the deposits of a shoreface environment, in the hangingwall of the Statfjord East fault. These are attributed to the exposure of 'footwall islands' at the crest of the large displacement Snorre fault block (Dahl and Solli 1993).

4.3 Dataset and methods

The area of investigation for this study is described by the coverage of five 3D seismic surveys (Fig. 4.1). The surveys describe an area with approximate dimensions of 55 km in a north-south direction and 20 km from west-east. The line spacing of the surveys is between 12.5-25 m for both inlines and crosslines, and the data have a vertical resolution of approximately 10-50 ms TWT. Reflectors within the syn-rift sediments in the hangingwall of the Strathspey-Brent-Statfjord fault system are of variable resolution. In general, the data are of poorest quality in the south of the study area, where the syn-rift is deep (up to 4000 ms TWT), and quality improves northwards. As a consequence of this, and of the distribution of the hangingwall wells, interpretation of reflectors within the syn-rift was only undertaken for the areas of the Statfjord (*st9101*) and Statfjord East (*e86*) seismic surveys (see Fig. 4.1). Reflectors corresponding with the base and top of the syn-rift were mapped across the entire study area. The data were interpreted using GeoQuest IESX seismic interpretation software at the University of Edinburgh. Further details on the seismic dataset and information on the density of seismic interpretation are recorded in Appendix 1.

Five key seismic reflectors were mapped in the study area, all associated with syn-rift sediments in the proximal hangingwall to the Strathspey-Brent-Statfjord fault array (Figs. 4.2 and 4.3; Enclosure 2). The top pre-rift reflector is a strong positive waveform and a marked seismic onlap surface, separating the sub-parallel reflectors of pre-rift geologies from the chaotic reflectors of the syn-rift. Reflectors corresponding with the top of the Tarbert Formation, top of the Heather Formation and top Kimmeridgian were all derived from picks in well logs, tied to the seismic data using the velocity log (Fig. 4.4). The top syn-rift reflector represents an unconformity in the north of the study area, but is conformable in the south. It was defined by both its seismic stratigraphic character (truncation of syn-rift geologies below and as an onlap and downlap surface) and its chronostratigraphic significance, being of base Cretaceous age. Interpretation of these reflectors, and faults of the Strathspey-Brent-Statfjord fault system, was undertaken at 12.5-125 m spacing on both inlines and crosslines.

A total of 13 wells penetrate the hangingwall of the Strathspey-Brent-Statfjord fault array: Five in the Strathspey-Brent-Statfjord area and a further 7 in the Statfjord (west) – Statfjord East area (Fig. 4.1). For all these wells completion logs and biostratigraphic information (of variable quality) were available. The majority of the wells had cored intervals from the syn-rift and detailed logs were made of core from wells 211/29-8, 211/29-9, 211/30-1, 33/9-18 and 33/9-C27. A further hangingwall well, 3/10b-1, located some 7 km to the south of the seismic coverage, was investigated. The completion log and access to the cored interval were available to this study; however, no biostratigraphic information could be attained. Documentation (including core logs, facies definitions and charts of sedimentation rate) pertaining to the interpretation of each of the 6 hangingwall wells from which core samples were examined is recorded in Appendix 3. Wells in the Statfjord East area have previously been described by Dawers *et al.* (1999), Nøttvedt *et al.* (2000) and Davies *et al.* (in review). An assessment of data from the completion logs, biostratigraphy and core reports was made in light of the results of these workers.

4.4 Distribution of syn-rift sediments

An overview of syn-rift sedimentation

A map of the depth (in ms TWT) to the top syn-rift (base Cretaceous) reflector is shown in Fig. 4.5. Bar the effects of a post-tectonic 2-3° regional tilt to the south and differential sediment compaction, this surface equates to the topography of the basin at the cessation of tectonic activity. Strong highs and lows, corresponding with footwall and hangingwall respectively, are described. This emphasises the underfilled nature of the basin, which was sediment starved through much of the rift period. The controls on post-tectonic sedimentation and the infilling of the remnant accommodation in the basin are discussed in Chapter 5.

The segmented nature of the fault system is also clear in Fig. 4.5. Variations in the elevation of the footwall high reflect displacement variations along the strike of the Strathspey-Brent-Statfjord fault system. The location of the maximum displacement on the fault array (marked A on Fig. 4.5) corresponds with one of the shallowest points on the footwall. Correlatively, relatively shallow areas in the hangingwall (the intra-basin highs marked B, C and D) correspond with the positions of the largest magnitude displacement minima on the main fault system (Chapter 3, see Fig. 3.7).

The control of along strike displacement variations on the Strathspey-Brent-Statfjord fault system on basin topography is further emphasised by the form of the top pre-rift reflector. A map of the depth (in ms TWT) to this onlap surface is shown in Fig. 4.6. Again a post-tectonic southward tilt has modified the geometry of this surface. The predicted forms of footwall gently dipping away from and hangingwall dipping into the fault are observed. However, more strikingly, the reflector in the hangingwall is observed to have significant topography and describes a series of folds. A hierarchy of two scales of folding is identified (Fig. 4.6). The folds are clearly observed in N-S oriented, fault parallel seismic crosslines (Fig. 4.7). The larger structures are between 4-12 km in width and separated by intra-basin highs, up to 400 ms TWT shallower than the hinge of the syncline. Smaller folds of between 1-3 km width are described within the confines of the larger structures. These are up to 150 ms TWT in amplitude (see definition in Fig. 4.7). The hingelines of the folds are perpendicular to, and plunging towards, the main fault system (Fig. 4.6). The amplitude of the folds decreases rapidly away from the fault, such that the structures are barely visible 3-4 km to the east (Fig. 4.7d).

These folds are interpreted to be extensional fault related folds, as described by Schlische (1995) and Schlische and Anders (1996). These structures originate from along strike displacement variations on a multi-segmented fault system, and describe depocentres adjacent to each unique segment. Displacement variations are more pronounced in the hangingwall than footwall because hangingwall displacement is generally greater than footwall displacement during coseismic slip (e.g. Stein and Barrientos 1985). This interpretation is supported when the displacement-length profile of the Strathspey-Brent-Statfjord fault system is compared with the geometry of the top pre-rift reflector (Fig. 4.6b). Displacement maxima and minima on the profile correlate with the culmination of synclines and anticlines, respectively. Hence, folds can be directly related to along strike variations in displacement developed when the fault system was an array composed of numerous shorter palaeo-segments. These observations support the conclusions of Chapter 3 and demonstrate the strong tectonic control on basin floor topography.

Measuring the distance between the top pre-rift and top syn-rift (base Cretaceous) reflectors, a map of the thickness of the syn-rift interval can be generated. The thickness map shown in Fig. 4.8, as with all other thickness maps shown in this chapter, is contoured in time (ms TWT), as none of the seismic data used in this study have been depth converted. The image demonstrates an overall wedge-shaped form, thickening towards the south and into the fault. However, overprinted on that geometry, are thickening and thinning associated with the extensional folds, demonstrating the spatial variations in the volume of accommodation space. The syn-rift has a maximum thickness of ~700 ms TWT in the immediate hangingwall to the position of maximum displacment (hence, maximum accommodation space) on the fault. The first-order folds of 4-12 km width represent syn-rift thickening of up to 350 ms TWT. The small folds are described by thickening of ~150 ms TWT. The early syn-rift basin floor topography, reflecting displacement variations on a multi-segmented fault array, strongly controls the distribution of syn-rift deposits. Thickening of the syn-rift is not just associated with the main fault system but also the abandoned palaeo-tips (e.g. point A on Fig. 4.8) and with the hangingwall fault population (B on Fig. 4.8), in particular with the antithetic trend described in the Statfjord area (compare Figs. 3.11 and 4.8).

Migrating loci of deposition during the syn-rift

The sediments of the syn-rift were sub-divided in the Statfjord area by the mapping of four seismic stratigraphic packages (Fig. 4.3). The packages have very different and distinctive seismic characters and are bounded by three reflectors with good lateral continuity. Well ties demonstrate that the seismic reflectors correspond with the flooding event at the top of the Tarbert Formation, a series of carbonate horizons near the top of the Heather Formation and the top Kimmeridgian timeline (Fig. 4.4). The seismic character and the distribution of these four syn-rift packages are described below; the sedimentology of the deposits is discussed in the succeeding section.

The oldest of the four packages of the syn-rift is described on the seismic dataset by two strong positive (red) reflectors. The deeper of these is the onlap surface corresponding with the top of the pre-rift. The other reflector, which correlates with the top of the Tarbert Formation, locally onlaps the top pre-rift reflector, but, more commonly, is a strong wavelet up to 100 ms TWT above the onlap surface (Fig. 4.9). This reflector is laterally discontinuous as the package thins over the crests of the dense population of faults active during the earliest syn-rift. The thinness of the Tarbert Formation (0-100 ms TWT) and its lateral discontinuity means that the package can only be mapped on the seismic dataset where closely spaced wells permit confident definition of the upper bounding reflector as the top of the formation. For the majority of the study area this is difficult but in the Statfjord East area the well population is, relatively, dense (Fig. 4.1). The map of the thickness of the Tarbert Formation shown in Fig. 4.10 is reproduced from Davies *et al.* (in review), with the syn-depositionally active fault population superimposed.

Thickness variations observed in Fig. 4.10, and in particular local increases in the thickness of the Tarbert in the hangingwalls, demonstrate the strong tectonic control on sediment distribution. This is most marked in the south of the Statfjord East area, where the formation achieves maximum thickness of ~100 ms TWT adjacent to the Statfjord East segment of the Strathspey-Brent-Statfjord fault. Onlap and downlap geometries are observed within the formation in this area (section BB', Fig. 4.9), suggesting progradation of syn-rift sediments into growing hangingwall depocentres. To the north of the Statfjord East area (section AA', Fig. 4.9), maximum thickness is distal from the fault (M on Fig. 4.10), interpreted as a fault controlled monocline basin by Davies *et al.* (in review). In addition to the proto- Strathspey-Brent-Statfjord fault, numerous short (<3 km length), low displacement faults control the thickness of the formation (e.g. depocentres A and B on Fig. 4.10), many of these faults offsetting the deeper, but not the shallower, bounding reflectors (Fig. 4.9).

The Tarbert Formation is overlain by the Heather Formation, the contact between the two defined by a significant rise in relative sea level and regional (but not time equivalent) flooding event. The Heather Formation is seismically transparent, characterised by weak, discontinuous reflectors (Fig. 4.11). No reflectors could be mapped with confidence within the formation, which is considered to have no seismically resolvable internal structure. In contrast, the top of the formation is defined by a strong, positive reflector with good lateral continuity, emphasised by the change in the seismic character of the syn-rift across the reflector. The reflector corresponding with the top of the Heather Formation has a significant topography describing a series of extensional folds similar to those of the top pre-rift reflector (Fig. 4.6), but less pronounced in amplitude (Fig. 4.12). The depocentres defined by the top Heather reflector are wider than depocentres described by the top pre-rift reflector, this is interpreted to represent the linkage of fault segments to form longer segments with correspondingly larger hangingwall basins.

As the reflector defining the base of the Heather (top Tarbert) was only mapped in Statfjord East area (*e86* survey) but the top Heather reflector was mapped over the entire Statfjord area, for the purposes of this discussion the Tarbert and Heather formations are considered a single seismic unit. A map of the thickness of the Tarbert and Heather formations (in ms TWT) is presented in Fig. 4.12b. The seismic package describing the Heather and Tarbert formations is between <20 ms TWT (in the east, on the dip slope) and 350 ms TWT (adjacent to the fault) in thickness. It shows a general east to west thickening towards the

Strathspey-Brent-Statfjord fault system and is thicker in the south, in the Statfjord area, than the Statfjord East area. Thickness variations describe a hierachy of isolated depocentres adjacent to the main fault system – small depocentres of up to 2 km in width and larger structures of up to 7 km width. Areas where the formations are relatively thin separate the depocentres. The depocentres correspond with the location of the extensional folds described in the topography of the top pre-rift reflector and are, therefore, interpreted as infilling accommodation space produced by along strike variations in displacement on segments of the Strathspey-Brent-Statfjord fault system. This interpretation implies that, during deposition of the Tarbert and Heather formations, the fault population grew from shorter 2-4 km long segments to linked arrays up to 9 km in length.

In addition to thickening into the Strathspey-Brent-Statfjord fault array, the Tarbert and Heather formations are demonstrated to thicken substantially into some of the faults described in the hangingwall to the main structure, in particular faults comprising the antithetic array (Fig. 3.11). Amalgamated depocentres are observed adjacent to these faults, indicating their segmented nature. The hangingwall faults rarely offset the reflector corresponding with the top of the Heather Formation (Fig. 4.3) and so were inactive by the time of late Heather deposition (see discussion in Section 3.4.2).

The seismic response of the syn-rift sediments demonstrates a marked change in character between the Heather and Kimmeridge Clay formations (Figs. 4.13). In contrast to the low amplitude, discontinuous reflections describing the seismically transparent younger sediments, the Kimmeridge Clay Formation is characterised by laterally continuous, high amplitude reflectors. The strongest of the intra- Kimmeridge Clay Formation reflectors is tied in wells to the top of the Kimmeridgian and corresponds with a sharp log break indicating a change lithology. This reflector onlaps sediments of the early Kimmeridge Clay Formation on the hangingwall dip slope; a geometry indicative of increasing rates of dip slope rotation and, by implication, hangingwall subsidence.

The Kimmeridge Clay Formation is thin when compared with the underlying Heather. The formation achieves a maximum thickness of ~250 ms TWT in a fault proximal position along the length of the Strathspey-Brent-Statfjord fault system (Fig. 4.13b). It thins to below seismic resolution to the east, on the hangingwall dip slope. Thickening of the formation into individual depocentres adjacent to the fault is recognised – at both the small and large scales described previously. However, depocentres are less pronounced (of lower amplitude) than those described from the Tarbert (Fig. 4.10) and Heather (Fig. 4.12) formations. Investigation of the growth of the Strathspey-Brent-Statfjord fault array (Chapter 3) demonstrates that, during the deposition of the Kimmeridge Clay Formation, the array was a fully linked, throughgoing fault strand and rates of displacement were comparable along the length of the system. Thickening of the Kimmeridge Clay into antithetic faults, which were not active at this time, is attributed to the effects of differential compaction of underlying sediments. Hence, it is concluded that the depocentres observed adjacent to the main fault may also be an artifact preserved due to the effect of differential compaction, rather than along strike variations in displacement.

4.5 Evolution of sedimentation during the syn-rift

From observations of thickness variations in the total syn-rift alone it can be concluded that accommodation generated in the hangingwall to normal faults strongly controlled the loci of syn-rift sedimentation. Small depocentres, associated with isolated normal fault segments, grew and amalgamated during the deposition of the Tarbert and Heather formations, mirroring the growth of accommodation in the hangingwall to the faults. During deposition of the Kimmeridge Clay Formation, the Strathspey-Brent-Statfjord fault was fully linked (Chapter 3) and the entire hangingwall area subsided equally and rapidly. In this section the results of an examination of the sedimentology of the individual components of the syn-rift are presented. These data are employed to reconstruct the syn-rift palaeo-environment, specifically the processes of sediment transport and routeways within the basin. Hence, the scale of tectonic influence on sedimentation at different times of the syn-rift, compared with the rate of sediment supply and fluctuations in relative sea level, can be assessed.

4.5.1 The Tarbert Formation

Sedimentology of the Tarbert Formation

The Tarbert Formation, the oldest of the three formations of the syn-rift, is intersected by 12 of the population of hangingwall wells. Eight of these are located in the Statfjord East area; the log signatures and cored intervals of the Tarbert in these wells have been described by Davies *et al.* (in review). The other 4 wells – 3/10b-1, 211/29-8, 211/30-1 and 33/9-18 – are previously undescribed and in all, bar 211/29-8, the formation has been cored. The base of the Tarbert Formation is defined in core by a transition into more marine influenced facies (compared with the underlying Ness Formation), specifically this study defines the base by the youngest coal horizon (*cf.* Davies *et al.* in review). This boundary is biostratigraphically dated to be of latest Bajocian age. On the seismic dataset this correlates with the top pre-rift onlap surface. The top of the formation is described in well logs by the sharp lithology change, from sandstone to siltstone, marking the major flooding event at the base of the Humber Group (Heather Formation). Biostratigraphic evidence dates this flood as of early-mid Bathonian age in the study area. Permo-Trias age microfauna are commonly documented from the Tarbert Formation, possibly indicative of the source geologies.

In all the wells sampling the Tarbert Formation, the formation is observed to have a very distinctive log motif (Fig. 4.14). The base and top of the formation are described by a sharp log breaks; the base typically shows a decline in gamma count, and the top an increase in the gamma ray count and increase in the (positive) density-neutron separation. The signature of the formation is characterised by a low gamma count (typically in the range 15-100 API), a high sonic velocity (>100 μs/ft) and small, regularly positive, density-neutron separation – indicative of porous, sand-rich sediment.

From interpretation of the log signatures, a number of coarsening up, prograding depositional packages are described (Fig. 4.14). In core (Fig. 4.15), these are typically represented by marginal marine or shoreface sediments erosionally overlain by coarse channel deposits (fluvial or estuarine) – indicating an overall fall in relative sea level. These packages are capped by a marine flooding events and the cycle is commonly repeated (e.g. three times in 3/10b-1; Fig. 4.14a) with a transgressive trend. These observations are consistent with those presented by Davies *et al.* (in review); a detailed interpretation of the sedimentology of the Tarbert Formation from the Statfjord East area. Core samples also demonstrate that the formation is extremely heterolithic (including coals, bioturbated mud and fine sand laminae, unsorted coarse grained sandstones, cross-stratified sands, pebble lags; logs are contained in Appendix 3) and reflects a wide range of depositional conditions from the coastal and shallow marine environments. However, the formation is universally capped by a major flooding event and a transition into a thick succession of finer grained sediments. The low resolution of biostratigraphic information does not permit correlation of this flood throughout the study area; log signatures suggest that relative sea level rise was punctuated by progradational events (described by the serrated gamma ray profile of the upper Tarbert in 211/30-1; Fig. 4.13b). It is noted that delta plain environments tend to have a low topographic gradient, with a tidal influence deep inland (e.g. the Mississippi delta complex, Gould 1970), hence a small rise in relative sea level can flood a large area.

Sediment source, transport and deposition of the Tarbert Formation

Interpretation of core samples and log signatures from the Tarbert Formation in the study area recognises a number of small progradational units arranged with a transgressive geometry. This supports the observations of Graue *et al.* (1987), Fält *et al.* (1989), Johannessen *et al.* (1997), Råvnas *et al.* (1997) and Davies *et al.* (in review) from elsewhere in the North Viking Graben. This stratigraphic architecture is representative of high frequency fluctuations (within an overall rise) in relative sea level as the shoreline migrated south during the latest Bajocian – Middle Bathonian. Fluctuations in sea level and, hence, the complex sedimentary record of the earliest phase of rifting can be attributed to the balance between rates of sediment supply and rates of accommodation generation.

Accommodation space was generated in the basin by both fluctuations in eustatic sea level and tectonic subsidence. Seismic mapping demonstrates that tectonic subsidence in the hangingwalls to active normal faults was a significant control on the distribution of the Tarbert Formation (Fig. 4.10). However, during the initial phase of rifting, the fault population comprised a large number of short fault segments (typically <3 km) with maximum displacements of <50 ms TWT; the wide distribution of strain at this time infers low rates of growth (Chapter 3). As a consequence, the volume of accommodation generated, and the size of an area affected, during a single seismic event was limited. Seismic stratigraphy (e.g. Fig. 4.9) and well log signatures suggest that the shoreline prograded and infilled local accommodation generated by tectonic subsidence. Deposits of the Tarbert Formation are preserved throughout the basin in both the footwalls and hangingwalls of active fault segments. These observations demonstrate that the rate of sediment supply exceeded the rate of tectonic subsidence and illustrate the overprinting of local subsidence in the hangingwalls of faults by a regional generation of accommodation due to rising relative sea level (e.g. Underhill and Partington 1993, 1994). It is concluded that the basin bounded by the Strathspey-Brent-Statfjord fault system was not sediment starved during deposition of the Tarbert Formation, pronounced depocentres were not formed and all tectonically generated accommodation was infilled.

Deposits of the Tarbert Formation are the coarsest grained of the Brent Group. The rounded quartzite gravels and reworked Permo-Trias faunas sampled in cored intervals indicate that

Tarbert sediments are of different provenance to the older formations of the Brent Group. This change in provenance reflects a change in the sediment source area; most likely deeper exhumation in the North Sea Dome to the south (Underhill and Partington 1993, 1994), unroofing an older stratigraphy of coarse grained sediments (compared with previous exposure of younger finer grained lithologies, for example). The influx of pebbly sand occurred coincident with the Brent delta complex achieving its maximum northward extent in the basin. However, the Tarbert Formation was deposited as rates of subsidence were increasing due to the active tectonic regime. The implication of these observations is that, in order to maintain progradation of the delta with increasing rates of subsidence, the rate of sediment supply must have exceeded the rate of accommodation generation (by both tectonic subsidence and eustatic fluctuations), i.e. the rate of sediment supply was high.

On a regional scale, sedimentary environments during the early syn-rift are recognised to have been variable both spatially and temporally (e.g. Mitchener *et al.* 1992). A coastal plain setting to the south (corresponding with the heterolithic geologies of the Ness Formation) would have been time equivalent with the marginal marine – shoreface environment (deposition of the Tarbert Formation) in the Statfjord East area in the north. An overall diachronism can be predicted, although biostratigraphic data is not of high enough resolution to test this model in the study area. The overall signature of the Tarbert Formation is punctuated transgression (Fig. 4.14). In terms of the sequence stratigraphic model proposed by Posamentier *et al.* (1988), deposits of the transgressive system tract will thin as the coastline retreats (Fig. 4.16a). In an active extensional province thinning can be attributed to increased rates of tectonic subsidence, hence increased rate of shoreline retreat, as the rift event progresses. In the study area, the opposite is true; the Tarbert Formation thickens and coarsens to the south. In addition, the log motifs of the Tarbert Formation are comparable across the study area, as are the sedimentology and mineralogy of core samples – demonstrating that sediment source and transport mechanisms during the southward migration of the facies belt remained constant. Two factors are proposed to explain this apparent paradox. As the relative sea level rose so the shoreline was brought closer to the sediment source (Fig. 4.16b). The transport distance was shortened and sediments supplied to the coast were coarser and of increased volume (assuming the same volume of sediment was deposited in the coastal plain). The proposed source of sediments of the Tarbert Formation was the uplifted North Sea Dome, located ~350 km south of the study area (Fig. 4.17). Secondly, as rates of subsidence (accommodation generation) increased the facies belts in the shoreface narrowed (Fig. 4.16c). The impact of this was to partition coarse sediments in the coastal zone and starve the proximal shallow marine shelf (Heather Formation deposits).

4.5.2 The Heather Formation

Character of the Heather Formation

The Heather Formation is intersected by all the hangingwall wells in the study area but is generally not cored. Only wells 34/7-19 and 34/7-21 in the Statfjord East area have core samples from the Heather, short lengths at the top and base, described by Nøttvedt *et al.* (2000) and Davies *et al.* (in review). These authors document grey-black coloured, bioturbated silt and mudstones with rare sand sized grains, sediments attributed to deposition in a fully marine environment. From well 34/7-21, Nøttvedt *et al.* describe trace fauna of palaeophycus and skolithos ichnofacies, and therefore shelfal conditions can be inferred. Typical (undecompacted) sedimentation rates for the Heather Formation are >30 mm/ka.

The Heather has a very distinctive log motif, comparable between all the available wells (Fig. 4.17). In general, the formation has a relatively monotonous log signature – characterised by a gamma ray count in the range 80-120 API, a slow sonic velocity of 80-95 μs/ft and a large positive density-neutron separation. The formation can be sub-divided into three packages, defined on Fig. 4.17. The deepest sub-unit (1 on Fig. 4.17) has a declining gamma ray count, decreasing sonic velocity and the size of the positive density-neutron separation is also decreasing – in sequence stratigraphic terms, a progradational signature. In sub-unit 2 these trends are reversed and a number of high density carbonate spikes are described. The log motif of sub-units 1 and 2 is characteristic of transgression and describes a genetic stratigraphic sequence (after Galloway 1989).

The youngest sub-unit described in the Heather Formation has a very serrated (on a sub- 10 m scale) log signature, due to a series of carbonate-rich (dolomite) beds. Sediments interbedded with the carbonates have a relatively low gamma ray count in the range 40-100 API, a sonic reading of 60-95 μs/ft and large positive density-neutron separation, suggesting fine grained, quartz-rich sediments. This unit is interpreted as pulsed fine-grained clastic supply into a generally low sedimentation (carbonate dominated) environment. The overall trend of the log signatures is progradational.

The base of the Heather Formation is of mid-Bathonian age, probably older in the north than south of the study area (biostratigraphy is poor in the southern wells). The top of the formation is defined by this study as the youngest of the carbonate beds characterising the latest Heather, and is dated as near the top of the Middle Oxfordian across the study area. It is noted that deposits of Early-Middle Oxfordian age across the northwest European continental margin are characteristically carbonate-rich – the Alness Spiculite in Sutherland, the Coral Rag (Coralline Oolite) in Oxfordshire and Yorkshire and the Osmington Oolite in Dorset. These carbonate deposits are unconformably overlain by siliciclastics; the change in facies interpreted by Coe (1992, 1995) to reflect a transition from long term regression to transgression. This interpretation suggests that the youngest carbonate bed (top Heather) throughout the area is time equivalent.

Sediment source, transport and deposition of the Heather Formation

Dominated by hemipelagic deposition, the sediments of the Heather Formation were sourced from the water column. However, sedimentation rates were relatively high (>30 mm/ka (undecompacted) compared with modern hemipelagic rates of 5-15 cm/ka, Stow *et al.* 1996), although declining with time (~25 mm/ka during the Early Oxfordian). Davies *et al.* (in review) document spores from upland vegetation in wells 34/7-14 and 34/7-22; indicating that landfall was not distal and supplied material in suspension to a shelf setting. The southward retreat of the Brent delta complex, and in particular the location of the shoreline, can be constrained through mapping of the facies belts in the Beryl Embayment and South Viking Graben (Fig. 4.18). Marginal marine deposits of equivalence to the Tarbert Formation are assigned in these areas to the Bathonian – Early Callovian Beryl Formation (e.g. Dou *et al.* in review) and the Callovian – mid-Oxfordian Huggin Formation (e.g. Cockings *et al.* 1992). Hence, the shoreline can be demonstrated to have retreated ~650 km south into the Brae area by the Early Oxfordian. It is suggested that as the Brent delta complex was flooded, during a period when the volume of accommodation generated by tectonic subsidence was rapidly increasing, clastic sediments were partitioned in the nearshore zone (Fig. 4.16). The impact of this was to starve the proximal shallow marine environment, particularly of the sand-sized fraction, even although water depths were not significant. Hence, the transition from a sediment filled to an underfilled basin appears marked and dramatic.

A secondary, locally derived, source of sediments for the Heather Formation was from the degradation of footwall scarps within the basin. Maximum displacements on faults of the graben bounding arrays were <250 ms TWT during the Early Oxfordian. As the basin became sediment starved, uplifted footwalls developed relief and the scarps formed steep slopes. Semi-consolidated sediments of the Brent Group exposed in the fault scarp were transported downslope (via slides, slumps, flows) and clastic debris was shed into the basin. However, the volume of sediment liberated by this process was minor (see Chapter 6), certainly with respect to the volume of accommodation space and the volume of the sediments derived from hemipelagic fall-out. Clastics supplied by this mechanism were most likely reworked into the hemipelagic deposits by the infauna, current activity and further downslope movement of material, rather than being preserved as discrete sand-rich beds.

Although the uplifted footwalls of the active fault population were relatively unimportant in terms of sources of sediment, the location of footwalls strongly affected sediment distribution and deposition within the basin. In a sediment starved basin the basin floor is not flat, but rather has a topography controlled by the displacement on the fault population. The distribution of the Tarbert and Heather formations illustrate that the sediments are not blanketing the basin, as would be predicted for a hemipelagic deposit, but are infilling differential accommodation space (Fig. 4.12). The implication of this observation is that hemipelagic deposits are being redistributed following deposition; transported and reworked into migrating depocentres. In such a palaeoenvironmental setting, the mechanisms of sediment reworking were most likely to have been sliding, slumping and flowing on the shallow slopes on the limbs of extensional folds, possibly initiated by seismic shocks.

The rate of tectonic subsidence during much of Heather Formation deposition remained relatively low as the fault population was composed of a number of multi-segmented arrays of low displacement. Hence, low order fluctuations in relative sea level are preserved in the sedimentary record (sub-units 1 and 2). During the earliest Oxfordian, rates of subsidence in the basin increased as the Strathspey-Brent-Statfjord fault array linked to become a throughgoing fault and other faults became inactive, i.e. the basin entered the rift climax phase. The sedimentary record of this time period has an apparently progradational signature, defined by the secondary quartz-rich sediments, and is interpreted as pulsed clastic influxes into a sediment starved basin. Pulsed clastic input is incompatible with the palaeo-environmental interpretation of regional transgression, increasing water depths and a landmass becoming increasingly distant. Hence, it is concluded that these sediments are being derived within the basin, from degradation of pre-rift geologies in the uplifted footwalls. As rates of subsidence increased so the topographic highs got higher and the potential for scarp denudation also increased. Thus, sub-unit 3 is considered not to represent progradation associated with relative sea level fall, but rather an increase in fault scarp degradation during a period of rapid fault growth.

4.5.3 Kimmeridge Clay (Draupne) Formation

Sedimentology of the Kimmeridge Clay Formation

The Kimmeridge Clay Formation is sampled in all of the wells penetrating the hangingwall of the Strathspey-Brent-Statfjord fault system, and cored in wells 3/10b-1, 211/29-8, 211/29-9 and 33/9-18 (Fig. 4.1). The formation is also cored in wells 34/7-21, 34/7-21A, 34/7-23A, 34/7-23S and 34/7-24S in the Statfjord East area (described by Nøttvedt *et al.* 2000). Electrical log signatures from the Kimmeridge Clay Formation are typically extremely serrated (Fig. 4.19), with little compatibility between patterns in individual wells. The formation can, however, be sub-divided into two packages; an older unit of relatively low gamma ray count and an upper unit of exceptionally high gamma (typically in excess of 150 API). The boundary between the two packages corresponds with the top Kimmeridgian timeline (and seismic reflector, Fig. 4.13), i.e. the older sediments are of Late Oxfordian – Kimmeridgian age and the younger sub-unit is of Volgian-Ryazanian age. Average (undecompacted) sedimentation rates for the formation are <25 mm/ka – these are lower than those of the Heather Formation, indicative of basin starvation, and demonstrate the continued decline in sedimentation rates throughout the syn-rift (in well 33/9-18, the average sedimentation rate for the Kimmeridge Clay Formation is 30.3 mm/ka pre-Kimmeridgian and 17.8 mm/ka during the Kimmeridgian).

Cored sections from the older sub-unit of the Kimmeridge Clay Formation sample three facies (Fig. 4.20): (1) Fine-coarse grained sandstones beds, of cm-m thickness, (2) dark, structureless siltstones and (3) ill-sorted, chaotic beds up to 50 cm in thickness, incorporating both the sands and mudstones. The sands fine up or are massive, climbing ripple and parallel laminations are observed and, rarely, dewatering structures; the beds commonly incorporate rounded lithic gravels. These sandstones are interpreted as the mid-distal deposits of turbulent flows. Siltstones occasionally show weak lamination, are commonly micaceous and are <2 % sand of very fine grain size. These are considered to represent hemipelagic deposits. The ill-sorted third facies is typically dominated by highly deformed siltstones, but also incorporates gravels, organic material and clasts (of <5-75 cm size) of geologies comparable to the pre-rift Brent Group. This facies is relatively insignificant in the core samples, comprising less than 5 %, but is more common in wells with paths closer to the Strathspey-Brent-Statfjord fault (e.g. 211/29-8). These beds are interpreted as the deposits of debris flows reworking material on the basin floor, possibly originating as slumps and flows degrading the local footwall scarp. The biostratigraphy of this sub-unit is dominated by microfloras and microfaunas from the Early – Middle Jurassic; indicating a supply of sediments sourced from footwall erosion.

These three facies are recognised from well logs and core samples throughout the study area (although sands are less common in the Statfjord East area, Nøttevedt *et al.* 2000). A deep marine environment is interpreted; a sediment starved basin undergoing rapid subsidence in the hangingwall to the throughgoing, linked fault array. Hemipelagic deposition is the dominant process of sedimentation. The area experiences periodic influxes of sediment from degradation of the fault scarp, which locally transports clastics via debris flows, reworking the basin floor deposits. In addition, pulsed clastics are supplied by a turbidite system. The sedimentology of turbidite deposits would indicate that they are well traveled and/or have a mature source. No palaeo-current data was available to determine the direction of sediment transport. However, the similarity of these deposits throughout the basin leads this study to suggest that a longitudinal turbidite system was active, flowing parallel to, and in the proximal hangingwall of, the Strathspey-Brent-Statfjord fault.

The younger sub-unit of the Kimmeridge Clay Formation is defined by a sharp log break and marked increase in the gamma ray count (Fig. 4.19). In hangingwall wells outwith the Statfjord East area, the log signatures are less serrated than those of the deeper sub-unit, with the density-neutron separation consistently large and positive. Core samples from this sub-unit are dominated by structureless mudstones, interpreted to be of hemipelagic origin. Thin (2-150 mm) beds of fine sand sized quartz arenites are also preserved in cored intervals, generally well sorted and rarely showing cross or planar laminations (Fig. 4.20). These are interpreted as the deposits of a turbulent flows, but are of lesser frequency and size than the sand bodies in the deeper sub-unit suggesting that sediment supply to the turbidite system was significantly lower during the Volgian and largely 'switched off' at the end of the Kimmeridgian.

Volgian-Ryazanian age syn-rift sediments described from wells intersecting the hangingwall of the Statfjord East fault segment in the north of the study area are, in contrast, sand-rich (Fig. 4.21). Nøttvedt *et al.* (2000) recognise a progradational, shallowing up signature to Middle Volgian – Ryazanian sedimentation in this area, capped by early Cretaceous transgression. Core samples of the 'upper Draupne sand' are sand-rich, bioturbated and show evidence of tidal reworking; the sediments are interpreted as the deposits of a lower-upper shoreface environment. The base of the sands has been mapped in the Statfjord East area by Dawers *et al.* (1999); the red reflector in Fig. 4.13a. Dawers *et al.* illustrate that the presence of sands in the late syn-rift has a dampening effect on the amplitude of the top syn-rift reflector and, hence, constrain the distribution of these shallow marine deposits by amplitude mapping (Fig. 4.22a). Using comparable parameters, amplitude mapping in the Statfjord area demonstrates no negative anomalies in the hangingwall (Fig. 4.22b), supporting the observation that late syn-rift sands are restricted to the north of the study area.

Sediment source, transport and deposition of the Kimmeridge Clay Formation

The Kimmeridge Clay Formation was deposited during the climax of rifting; rates of displacement on the linked Strathspey-Brent-Statfjord fault were high and tectonic subsidence rapidly outpaced sedimentation. Hence, tectonic subsidence was the primary control on the generation of accommodation space and fluctuations in relative sea level had little effect on sedimentation patterns. This is documented throughout the study area with hemipelagic settling being the dominant process of sedimentation, slower rates (compared with the Heather Formation) testifying to the distance from land.

Coarse material was supplied to hangingwall from the degradation of the fault scarp, by slides, slumps and flows, although the magnitude of footwall denudation, and hence rate of supply, was low. The volume of material supplied by downslope processes was insignificant both in terms of accommodation space available and the volume of hemipelagic mudstone deposits (in core, mudstones account for ~75 % of the sample and reworked sediments <5 %). Reworked sediments were deposited proximal to the scarp from debris flows.

A second, and volumetrically more important (~20 % of core samples), sand-rich facies is also recorded in the Kimmeridge Clay Formation. Compositionally mature, well sorted sandstone beds of up to 1.5 m thickness are interpreted as the deposits of an axial turbidite system. Turbidite systems in the hangingwall to major normal faults have been documented in modern extensional basins (e.g. Lake Malaŵi, Wells *et al.* 1999; Lake Baikal, Back *et al.* 1999, Nelson *et al.* 1999) and the morphology of their deposits compared to stacked axial fluvial sedimentation (*cf.* Alexander *et al.* 1994, Peakall 1998). This system was active in the study area throughout the Late Oxfordian – Ryazanian, although these geologies are of greater abundance during the Late Oxfordian – Kimmeridgian than the Volgian-Ryazanian, indicating a decline in sediment supply at the end of the Kimmeridgian. The decreased rate of sediment supply corresponds with a shallowing in the hangingwall of the Statfjord East fault segment, and the establishment of a tidally influenced shoreface environment (the 'upper Draupne sand' of Dawers *et al.* 1999 and Nøttevedt *et al.* 2000).

The source of these mature sands is interpreted to be from the north, from the footwall of the Inner Snorre fault (Fig. 4.23a). This fault block was substantially denuded during the rift event, the subcrop of Triassic age geologies at the crest testifying to the erosion of Brent, Dunlin and Banks groups (Fig. 4.23b). The flat-topped profile of the footwall of the Inner Snorre fault has been attributed by Dahl and Solli (1993) to subaerial exposure of the crest and erosion to wavebase during the late syn-rift when the footwall defined a 'footwall island' (Fig. 4.23c). J. Underhill (*pers.comm.* 1999) presents a reconstruction of the evolution of the Zeta-Snorre-Gullfaks fault and discusses the implications for the loci and magnitude of uplift of the footwall. Changes in the rate of clastic sediment supply to the study area are interpreted to be primarily a response to the growth of the Zeta-Snorre-Gullfaks fault. In contrast to the Snorre fault system, there is no evidence to suggest that the footwall of the Strathspey-Brent-Statfjord fault system was uplifted to near or above sea level either during the rift phase or in the early post-rift (see Chapter 6). Palaeo-environmental reconstructions for deposition of the early and late Kimmeridge Clay Formation illustrating the growth of the Zeta-Snorre-Gullfaks fault system are illustrated in Fig. 4.24.

According to the model of Underhill, linkage of the en echelon Zeta and Inner Snorre fault segments during the Middle – Late Oxfordian resulted in a migration of the location of maximum accommodation generation to the hangingwall of the NW-SE oriented linking structure. Correlatively, this locality became the apex of uplift and the footwall was tilted to the southwest. As a consequence denuded footwall sediments were transported towards the southwest down the dip slope and along the Strathspey-Brent-Statfjord fault, via an axial turbidite system (Fig. 4.24a). The abundant floras and faunas of the Brent and Dunlin groups documented by biostratigraphic analysis of Late Oxfordian – Kimmeridgian age turbiditic sands support this interpretation. The 'footwall island' of the Zeta-Snorre fault, and associated shallow marine environment, would have been ~20 km to the north of the study area.

Underhill demonstrates that the Zeta-Snorre fault subsequently linked with the Gullfaks fault and the location of maximum accommodation space migrated to the hangingwall of the central Inner Snorre fault segment. The consequence of changing the location of maximum rate of displacement was to change the direction of tilt of the footwall to the west. In addition, the tip of the Strathspey-Brent-Statfjord fault system in the Statfjord East area was uplifted in the footwall to the now linked Zeta-Snorre-Gullfaks fault. Attributing the decrease in sand supply observed in the study area to the linkage of the Zeta-Snorre-Gullfaks fault system would date this linkage as late Kimmeridgian. The 'footwall island' migrated southwards and the associated shallow marine environment encompassed the uplifted Statfjord East area (Fig. 4.24b). Redistribution of shallow marine sands into deeper offshore waters, by, for example, storm currents, would have supplied a volumetrically smaller, and less active, axial turbidite system. The bulk of the pre-rift geologies exhumed and eroded from the footwall of the Zeta-Snorre-Gullfaks fault system during the Middle Volgian – Ryazanian would have been transported west down the dip slope or east into the hangingwall (although this remained significantly underfilled).

4.6 Syn-rift sedimentation in the hangingwall of the Statfjord (west) fault segment

In the north of the study area ~15 km from its tip, the Strathspey-Brent-Statfjord fault splits into two segments: The Statfjord (west) and Statfjord East fault segments. Statfjord (west) is the smaller of the two faults, both in terms of length and displacement (Fig. 3.7), and is effectively located in the footwall of the Statfjord East segment. Previous workers (e.g. Dawers and Underhill 2000) have considered the main phase of growth of the Statfjord (west) fault to post-date that of the Statfjord East fault and attributed the westward stepping of fault activity to the uplift of the Statfjord East area in the footwall to the Zeta-Snorre-Gullfaks fault system (see Fig. 4.24).

Seismic interpretation of the top pre-rift and top syn-rift reflectors in the hangingwall of the Statfjord (west) fault segment demonstrates that syn-rift deposits are up to 180 ms TWT in thickness (Fig. 4.25), substantially thinner than in the hangingwall of the Statfjord East fault. In addition, the depocentre is distal of the fault and defines the filling of a fault controlled monocline basin (see also the geometry of the top pre-rift reflector, Figs. 4.6 and 4.25). No reflectors were mapped within the syn-rift in the hangingwall of the Statfjord (west) fault segment as well penetrations did not permit confident definition of formations within the syn-rift.

A single well samples the syn-rift sediments in the hangingwall of the Statfjord (west) fault segment. Well 33/9-C27 is a deviated production well located in the proximal hangingwall to the fault. A seismic interpretation of the path of 33/9-C27 is shown in Fig. 4.26. Electrical log signatures and the biostratigraphy are reproduced in Fig. 4.27a, with plates describing representative facies from the core samples shown in Fig. 4.27b. The well samples all the formations of the syn-rift (Tarbert, Heather and Kimmeridge Clay). The log signatures and core samples from the formations are comparable to those discussed above, and, in particular, core samples from the late Kimmeridge Clay Formation record shallow marine conditions in the area (Fig. 4.27b). The Tarbert Formation is of Lower Bathonian age, but Heather Formation of only Middle Bathonian – Callovian age and Kimmeridge Clay of Middle Volgian age (equivalent to unit 2 of the formation as defined above) are sampled. Both the Heather and Kimmeridge Clay formations are thin, with respect to their thickness elsewhere in the study area. They are separated by an unique package of sediments termed the 'slump complex', which is ascribed a late Callovian – Oxfordian age from the biostratigraphy. No Kimmeridgian age sediments are described.

The slump complex has an extremely serrated log signature which can be interpreted to represent the interbedding of two facies. Low gamma count intervals, with a negative density-neutron separation, are interpreted as sand-rich. Core samples from these units (Fig. 4.27b) are of deformed geologies with a sedimentological affinity to the Brent Group. These are interpreted as reworked sediments degraded, by processes of sliding and slumping, from the footwall of the Statfjord (west) fault. The mechanics of slope denudation, the magnitude of erosion and the process of resedimentation are discussed further in the Chapters 6 and 7. Reworked sediments are interbedded with high gamma count packages – interpreted as fine grained, mud prone deposits of the Heather Formation.

The thickness of degraded sediments preserved in well 33/9-C27 is attributed to the proximity of the well path to the fault (similarly well 211/29-8, Appendix 3).

The stratigraphy of well 33/9-C27 can be interpreted in terms of the timing of growth of the Statfjord (west) and Statfjord East fault segments. Although in the footwall of the Statfjord East fault, the well samples >70 m thickness of Tarbert Formation, comparable with thicknesses in the proximal hangingwall to active segments of the proto- Statfjord East fault. This indicates that the proto- Statfjord (west) fault was active during the earliest stages of rifting. Continued activity in the Callovian and Oxfordian is suggested by the interpretation of the 'slump complex' – erosion of an active fault scarp. Geologies of the Rannoch Formation of the Brent Group are observed constraining the elevation of the scarp to, at least, 150 m. The maximum displacement on the Statfjord (west) fault segment is ~400 m at its southern extent (Fig. 3.7).

The most striking feature of the stratigraphy of well 33/9-C27 is the lack of a record for the late Oxfordian – Kimmeridgian and the extremely thin late (Middle Volgian age) Kimmeridge Clay Formation. This period of non-, or condensed, deposition is interpreted to be a consequence of the structurally high position of the basin with the potential for reworking and redistribution of sediments to a deeper basinal location. The continued growth of the Statfjord (west) fault segment is inferred, but no evidence is available which suggests enhanced growth rates and the transfer of slip from the Statfjord East fault.

4.7 Controls on syn-rift sedimentation in the Strathspey-Brent-Statfjord area

The fault population in the half-graben basin bounded by the Strathspey-Brent-Statfjord fault system evolved from a large number of short, low displacement faults to a few long, large displacement faults (Chapter 3). As strain was localised through the rift event onto a small number of structures, so the rates of displacement on the active faults increased. The implication of these observations is that the location and magnitude of accommodation generated by tectonic subsidence through the rift phase can be constrained: Initially, rates of subsidence were low and numerous isolated sub-basins developed, as the bounding fault segments grew and linked so the basins deepened, widened and linked.

The thickest deposits of the earliest syn-rift sediments (of the Tarbert Formation) occur in small, isolated fault bounded depocentres demonstrating the distribution of the earliest fault population (Fig. 4.10). However, the sand-rich deposits are preserved regionally – indicating that the rate of sediment supply exceeded the rate of tectonic subsidence and that accommodation generation was also influenced by an eustatic rise in relative sea level. The latest Bajocian – ?mid-Bathonian Tarbert Formation is extremely heterolithic; facies are characteristic of a marginal marine palaeo-environment. The sediments are coarse grained and of reworked Permo-Trias provenance. The stratal architecture is one of punctuated transgression demonstrating the complex interaction between sediment supply and accommodation generation as the Brent delta was flooded from the north. The package thickens and coarsens to the south, contradicting the model of the transgressive systems tract (Fig. 4.16).

The retreat of the coastal zone during a time of increasing rates of tectonic subsidence is interpreted to have resulted in a partitioning of coarse clastics in the shoreface. The impact of this was to starve the proximal shelf. The Heather Formation is a silty deposit of shallow marine origin. Rates of sedimentation are typically in excess of 30 mm/ka (undecompacted). The distribution of the formation is tectonically controlled (Fig. 4.12). The stratal architecture of these sediments in fault bounded basins describes the widening, deepening and amalgamation of sub-basins throughout the mid-Bathonian – Early Oxfordian. Linked fault arrays grew to 7-12 km length and faults accumulated up to 400 ms TWT displacement. The basin had a graben-like form at this time and depocentres are located adjacent to segments of both the proto- Strathspey-Brent-Statfjord fault and the main antithetic array. The Heather Formation is regionally distributed and up to 350 ms TWT thick, implying that sediment supply rates largely kept pace with rates of tectonic subsidence and that no pronounced basin floor topography was developed. Evidence of fluctuations in relative sea level recorded by the log signatures of the formation supports this interpretation (Fig. 4.11).

The sedimentary character of the Heather Formation changed in the Early – Middle Oxfordian. Rates of sedimentation declined and carbonate (dolomite) deposition was predominant. This transition is interpreted to reflect a regional change in water chemistry; this changes again into the Late Oxfordian, represented by the clastic-rich character of the Kimmeridge Clay Formation. Also during the mid-Oxfordian the basin attained a half-graben geometry; linkage of the segments of the Strathspey-Brent-Statfjord fault was completed and this was subsequently the only active fault in the basin. Average maximum rates of displacement (tectonic subsidence) increased – from ~45 mm/ka to >100 mm/ka (Chapter 3).

The distribution of the Kimmeridge Clay Formation is less focussed than that of the Tarbert and Heather formations (Fig. 4.13). The Kimmeridge Clay Formation thickens into the Strathspey-Brent-Statfjord fault alone. It is very thin, typically <250 ms TWT, indicating low and declining rates of sedimentation (undecompacted <25 mm/ka). The dominant source of sediments was from the water column; deposition by hemipelagic settling. A secondary source of sediments was from the denudation of the footwall of the Snorre fault to the north. An axial turbidite system was active during the Late Oxfordian – Kimmeridgian, but in response to the development of the Zeta-Snorre-Gullfaks fault system this bacame inactive into the Volgian-Ryazanian, although continuing to supply a shoreface environment in the Statfjord East area (Fig. 4.24). A minor input of clastics (<5 % supply) was sourced from the degradation of the footwall of the Strathspey-Brent-Statfjord fault system. Rates of sediment supply were greatly outpaced the rate of tectonic subsidence. Hence, a pronounced fault bounded range front developed with ~750 m elevation at the close of the rift phase. The infilling of this remnant accommodation space is examined in the succeeding chapter.

4.8 Conclusions

1. The Strathspey-Brent-Statfjord area is described as an underfilled half-graben basin developed in response to extensional stresses in the latest Bajocian – Early Ryazanian. The majority of the strain is accommodated on the >60 km long Strathspey-Brent-Statfjord fault system.

2. The distribution of syn-rift sediments in the half-graben is shown to be primarily controlled by tectonic subsidence and in particular illustrates the differential accommodation generated along the strike of the fault populations due to displacement variations. The generation of accommodation in the basin is shown to evolve from small depocentres, in the immediate hangingwall of isolated fault segments, which deepen and amalgamate as the faults lengthen and link. During the early stages of rifting the rates of tectonic subsidence are slow and fluctuations in eustatic sea level also impact upon

sedimentation patterns. With the linkage of the main half-graben bounding fault array, rates of subsidence increase, the basin becomes a single large depocentre and sea level fluctuations no longer have an influence upon sedimentation.

3. The initiation of rifting is documented stratigraphically by the drowning of the pre-rift geologies of the Brent delta. The extremely heterolithic marginal marine geologies of the Tarbert Formation have a complex stratigraphic architecture generally describing punctuated transgression. All the accommodation generated in the basin was filled during the earliest stages of rifting. Increased rates of subsidence, and hence rapid flooding, occurred coincident with the linkage of fault segments and the concentration of strain on fewer faults.

4. As the shoreline migrated southwards, so coarse grained sediments were partitioned at the coast, and the environment in the basin became a sediment starved, shallow marine shelf characterised by fine-grained deposition. However, rates of subsidence did not greatly outpace sediment supply as fluctuations in relative sea level are recorded in the log signatures of the early Heather Formation. Increased rates of tectonic subsidence, and the transition into the 'rift climax', occurred with the linkage of the Strathspey-Brent-Statfjord fault array during the Early – Middle Oxfordian. Subsequently, this was the main structure in the basin, accommodating the majority of the strain.

5. Sedimentation rates during the climax of the rift phase were extremely low and dominated by hemipelagic deposition. Two sources of coarse clastic deposits are recognised: (1) from degradation of the proximal fault scarp, and (2) from denudation of the Snorre footwall high to the north. Sands sourced in the Snorre area are the more voluminous of the two sediment supplies, and were transported in the study area by an axial turbidite system in the proximal hangingwall of the fault. The evolution of this system is intimately linked to the evolution of the Zeta-Snorre-Gullfaks fault system and the stratigraphy can be used to time the growth and linkage of this structure.

CHAPTER 5 Controls on early post-rift (Cretaceous) sedimentation in the Strathspey-Brent-Statfjord area, northern North Sea

5.1 Introduction

The consequence of syn-rift sediment starvation in an extensional province is the establishment of a fault controlled topography. In the preceding chapter it was demonstrated that the half-graben basin bounded by the Strathspey-Brent-Statfjord fault system was sediment starved during the latter stage of the rift event and, as a result, substantially underfilled. The map of depth (in ms TWT) to the top syn-rift (base Cretaceous) reflector, presented in Fig. 5.1a, approximates the topography of the basin floor at the end of the rift event. Although the effects of post-tectonic compaction and tilting mean that this is not an entirely true representation of the topography, the image demonstrates that the footwall of the Strathspey-Brent-Statfjord fault was a relative high at the end of rifting, up to 800 ms TWT above the adjacent hangingwall low. At the cessation of active tectonics, the study area was below sea level. The water depth can be constrained only in the Statfjord East area where depths of 10-50 m are inferred from Volgian – Early Ryazanian shoreface sediments (Nøttvedt *et al.* 2000). Hence, the base Cretaceous map approximates a bathymetry and preserves a minimum estimate of the accommodation space in the basin at the end of rifting.

This chapter documents, and interprets the controls on, the distribution and character of post-rift sediments filling the remnant accommodation space in the Strathspey-Brent-Statfjord area. Following the cessation of rift tectonics in the latest Jurassic – earliest Cretaceous, the northern North Sea began a phase of gentle post-tectonic subsidence that continues to the present day, interrupted by a transient uplift event during the Palaeocene (e.g. Nadin and Kusznir 1995). Only the earliest post-rift sediments are examined in this study, those belonging to the Cretaceous Cromer Knoll and Shetland groups (Fig. 4.2). Silty sediments assigned to the Shetland Group finally buried the footwall high during the Late Cretaceous; although it is observed that this buried topography continued to control sedimentation throughout the Tertiary. Seismic stratigraphy divides the early post-rift into 3 packages, of Ryazanian-Campanian age. Through mapping of these seismic stratigraphic packages temporal and spatial variations in the loci of deposition are illustrated. It is concluded that differential compaction and basin floor topography (initiating gravity driven sediment reworking) were the primary controls on post-rift sedimentation in the Strathspey-Brent-Statfjord area.

5.2 Dataset and methods

The work presented in this chapter is based on interpretation of the dataset described in Chapters 3 and 4. This consists of five 3D seismic surveys and fourteen wells penetrating the hangingwall to the Strathspey-Brent-Statfjord fault system (Fig. 4.1). The seismic surveys describe an ~965 km^2 area with approximate dimensions of 55 km in a north-south direction and 20 km from west to east. The line spacing of the surveys is between 12.5-25 m for both inlines and crosslines, and the data have a vertical resolution of approximately 10-30ms TWT. Further detail on the seismic dataset and information on the density of seismic interpretation are recorded in Appendix 1. For all the wells completion logs and biostratigraphic information (of variable quality) were available. Core samples of sediments of the earliest post-rift were logged from wells 211/29-8, 211/29-9 and 33/9-18. Documentation pertaining to the interpretation of well data, including core logs, interpretation of log signatures and charts of sedimentation rate, is recorded in Appendix 3.

Three seismic reflectors from within the post-rift interval were mapped during this study (Fig. 5.2a). The deepest of the reflectors correlates with the top of the Cromer Knoll Group and the two shallower picks – intra-shetland 1 and intra-shetland 2 – are within Shetland Group. Each of the three represents a positive acoustic impedance contrast and could be regionally picked with confidence.

The age and stratigraphic significance of the three reflectors was determined from correlation with electrical well logs using velocity log data. The seismic cross section shown in Fig. 5.2a maps the path of vertical exploration well 211/29-9. The gamma ray, density and sonic log signatures and the age (determined from biostratigraphy) of the early post-rift sediments sampled in this well are reproduced in Fig. 5.3, with the depth of the three post-rift reflectors superimposed. The deepest of the three picks correlates with an unconformity at the top of the Cromer Knoll Group, of Albian age in 211/29-9. The picks within the Shetland Group correspond with density spikes, interpreted as carbonate-rich horizons, and are dated as Lower and Upper Campanian in age. The top Cromer Knoll and intra-shetland 2 reflectors correlate with strong log breaks in 211/29-9, the break corresponding with the intra-shetland 1 reflector is much weaker. However, although the signatures described in this well are diagnostic regionally, all the reflectors vary in amplitude locally while demonstrating lateral continuity (e.g. Fig. 5.4). The rates of deposition and the palaeo-environment of sedimentation were also interpreted from well data.

5.3 Distribution of early post-rift (Ryazanian-Campanian) sediments

The seismic character of the early post-rift is illustrated in W-E and NNE-SSW oriented seismic cross sections, shown in Figs. 5.2a and 5.4 respectively. In the underfilled hangingwall of the Strathspey-Brent-Statfjord fault system, sediments of the Cromer Knoll Group and the lower part of the Shetland Group onlap the footwall scarp to the west and hangingwall dip slope to the east. The crest of the footwall was completely buried by the time the carbonate-rich horizon corresponding with intra-shetland 1 was deposited. The Shetland Group has a generally transparent seismic character, composed of weak, laterally discontinuous reflections (as would be predicted by the log signatures, Fig. 5.3) with poorly developed sub-parallel reflector geometry. The Cromer Knoll Group, in contrast, is bounded by two high amplitude reflectors and has a number of strong internal reflections. Reflectors within the Cromer Knoll Group onlap the top syn-rift (base Cretaceous) surface to the west and downlap onto the top syn-rift to the east.

Flattening the section on the top syn-rift reflector (Fig. 5.2b), reveals that the thickest deposits of Cromer Knoll Group, and indeed Shetland Group, are located some 1.5 km to the east of the base of the scarp; hence, the origin of onlap and downlap patterns observed in the Cromer Knoll Group. The implication of this observation is that loci of deposition for early post-rift sediments was not proximal to the fault scarp, as was the case during syn-rift sedimentation (Fig. 4.8), but rather is sited ~1.5 km basinward of the base of the slope. This can be attributed to a buttressing effect by the footwall geologies. As sediments post-depositionally compacted, the syn-rift sediments compacted more than underlying pre-rift sediments in the footwall. Hence, accommodation space created by differential compaction was at a maximum where the syn-rift sediments overlay pre-rift sediments in the hangingwall rather than those in the footwall.

Maps of the depth (in ms TWT) to the top syn-rift (base Cretaceous) reflector and post-rift reflectors are shown in Fig. 5.1. The reflector corresponding with the top of the Cromer Knoll Group was mapped in the hangingwall area, and could only be resolved seismically in the south of the study area. In all of the maps, the surfaces have significant topography. Highs occur over the footwall of the Strathspey-Brent-Statfjord fault and lows over the hangingwall. Relative highs and lows within the hangingwall area correlate with the locations of depocentres described by the syn-rift sediments (see Fig. 4.8). The magnitude of the topographic variation declines with time, i.e. the base Cretaceous reflector has more topography (~1400 ms TWT over the study area) than intra-shetland 2 (~350 ms TWT).

From these picks, maps of the thickness (in ms TWT) of the total early post-rift (Ryazanian-Campanian) sediments, the Cromer Knoll Group and the interval between the base Cretaceous – intra-shetland 1 reflectors and intra-shetland 1 – intra-shetland 2 reflectors were produced. These maps are shown in Fig. 5.5. The total thickness of the early post-rift sediments (i.e. top syn-rift – intra-shetland 2 reflectors) in the study area is 175-1075 ms TWT. This thickness map is directly comparable with the topography of the base Cretaceous reflector (Fig. 5.1a), as would be predicted if the post-rift sediments were infilling a remnant topography. The post-rift interval is thin on the footwall high, and thinnest at the crest of the Statfjord structure (point A on Fig. 5.1a), close to the point at which the top syn-rift reflector is shallowest. The post-rift sediments are thickest in the hangingwall area. The sediments thicken to the south, mirroring the deepening of the top syn-rift reflector, and are thickest in the hangingwall of the Strathspey structure.

By breaking down the post-rift section into three packages, the infilling of remnant accommodation space can be mapped.

The Cromer Knoll Group is only seismically resolvable in the south of the study area. The group is up to 350 ms TWT thick (Fig. 5.5b), thickening towards the south where the accommodation space was greatest at end syn-rift times. Fig. 5.5b suggests the possible extent of deposition of the Cromer Knoll Group – thinning and onlapping to the north, onto the footwall scarp in the west and onto the dip slope in the east. The loci of deposition were distal of the base of the scarp, as described in Fig. 5.2b. Sub-basins corresponding to along strike variations in the displacement of the fault, documented by the distribution of the syn-rift (Fig. 4.8), are observed and related to differential compaction of the underlying sediments. These observations on the distribution of the Cromer Knoll Group demonstrate that deposition during early post-rift times was restricted to the deepest parts of the basin and that elsewhere non-deposition resulted in an unconformity. Thus, rather than 'blanket-deposition' of post-rift sediments, the topography of the hangingwall basin controlled the sedimentation. The south to north onlap pattern implies that the oldest post-rift sediments should be found in the south, the location of maximum accommodation space, and should young northwards.

In Fig. 5.5c, the thickness of sediment (in ms TWT) between the top syn-rift and intra-shetland 1 reflectors is shown. This equates to the thickness of the Cromer Knoll Group and the lower Shetland Group (to Lower Campanian age). This sedimentary unit is of thickness 50-900 ms TWT, up to 650 ms TWT thicker in the hangingwall than in the proximal footwall (e.g. line B on Fig. 5.5c). The top Cromer Knoll reflector onlaps both the footwall scarp and dip slope, whereas the intra-shetland 1 reflector is shallower than the highest point on the footwall. Hence, this package represents the complete infilling of the end syn-rift topography. The major thickness variations within this package support this observation, again demonstrating a direct correlation between thickness and the amount of end syn-rift accommodation space (enhanced by differential compaction).

In contrast, the final post-rift package, intra-shetland 1 – intra-shetland 2 (Fig. 5.5d), is of relatively uniform thickness across the study area. This package varies in thickness from 75-300 ms TWT, again thinnest on the footwall high and in the Statfjord East area and thickest in the hangingwall of the Brent and Strathspey structures. The syn-rift topography was now buried; hence, during deposition of this sedimentary package the sea bed was probably near planar. Thickness variations are, thus, attributed to accommodation space produced by differential compaction. The effects of differential compaction in generating accommodation space were likely to have been significant due to the varying thicknesses and sedimentary character of underlying syn- and post-rift sediments.

5.4 Process(es) of sedimentation during the early post-rift (Ryazanian-Maastrichtian)

A progressive infilling of the basin, from deep to shallow, controlled by the remnant topography of the rift event and the effects of differential compaction is demonstrated by the observations presented above and is supported by calculations of sedimentation rates from hangingwall wells. For eight wells (located on Fig. 4.1), the total thicknesses of Cromer

Knoll and Shetland groups were measured and rates of sedimentation during the Ryazanian-Maastrichtian calculated (rates not calculated for well 3/10b-1 as no biostratigraphy was available). The results are presented in Fig. 5.6; note that in the calculation of sedimentation rates the sediments were not decompacted.

As predicted from the seismic interpretation, both the Cromer Knoll and Shetland groups show a northwards thinning in the hangingwall area, and are thickest in well 3/10b-1, located ~7 km south of the coverage of the seismic data. Where the Cromer Knoll Group is below seismic resolution, in wells 33/9-18, 34/7-21, 34/7-24S and 34/7-9, the group is <20 m thick. More significant than the thickness variations are the temporal and spatial variations in sedimentation rates (Fig. 5.6b). In general, sedimentation rates were low (average = 2 mm/ka) during deposition of the Cromer Knoll Group and increased by more than 10-fold during Shetland Group deposition (average = 33 mm/ka). Superimposed on these temporal variations are spatial variations in sedimentation rates. Obviously, a decline in the thickness of early post-rift sediments from south to north corresponds with a general decrease in the rate of sedimentation, but this does not take into account the effect of unconformities.

Across the study area, two periods of low rates of sedimentation or non-deposition are recognised. The older of the two (unconformity 1) is dated to correspond with the earliest post-rift, Ryazanian-Hauterivian (~12 Ma), and can be recognised in wells 211/29-8, 211/30-1, 33/9-18, 34/7-24S and 34/7-19. As described above, during the earliest post-rift (deposition of the Cromer Knoll Group) the basin was being filled by passive onlap of sediments from the south towards the north (i.e. from deepest to shallow). Unconformity 1 can be attributed to this onlap and, consequently, the length of time represented by the unconformity is greater in the wells in the north than in the south. The second of the periods of low sedimentation and non-deposition (unconformity 2) occurs during the latest Barremian – Coniacian (~37 Ma) and correlates in wells with the boundary between the Cromer Knoll Group and the Shetland Group (Fig. 5.6). The length of time represented by this event is variable between wells, probably partly reflecting the accuracy of the biostratigraphical information, but the unconformity is generally a more distinct event in the north of the study area. The basin remained underfilled at this time but this event is apparent even in the deepest areas indicating a basinwide control on the decline in rates of deposition, a regional unconformity. The origin of this event is unknown.

The environment of deposition of the Ryazanian-Maastrichtian post-rift sediments can only be determined indirectly from interpretation of the wireline log responses. The log signatures in well 211/29-9 (reproduced in Fig. 5.3) are discussed and interpreted below in order to assess sediment source, transport and palaeo-environment. The log signatures of the post-rift interval are comparable across the study area, so the observations from 211/29-9 are considered to have regional applicability.

The base of the Cromer Knoll Group is described by a sharp log break from the high gamma ray count shales of the latest syn-rift below. Sediments of the Cromer Knoll Group sampled in 211/29-9 can be sub-divided into two packages – units 1 and 2 on Fig. 5.3. The division between the two is marked by a strong log break and a major change in the character of the log signatures. The upper unit 2 is similar in log response to the Shetland Group and so is described and interpreted with that group below.

Unit 1 of the Cromer Knoll Group has an extremely serrated log signature. The unit is 240' (71.6 m) thick and of Ryazanian – Upper Albian age. The gamma ray count is in the range 30-125 API, sonic readings are between 80-130 μs/ft and the bulk density is 2.1-2.7 gr/cc (no other tools were run for this interval). These indicate that the unit is extremely heterolithic, although dominated by shale deposits. The basal 1-2 m of the Cromer Knoll Group has been cored in this well and also in wells 211/29-8 and 33/9-18 (see logs in Appendix 3). The core samples are of a carbonate-rich mudstone typically demonstrating intense soft sediment deformation and reworking (Fig. 5.7). The sediments also show evidence of prolonged exposure at the sea bed (chemical precipitates) and carbonate clasts. From these observations, it is concluded that unit 1 in 211/29-9 represents a condensed horizon, indicative of a period of exposure at the sea bed in a deep marine environment, with no clastic deposition. This is compatible with the biostratigraphical evidence, presented above, demonstrating that this unit was deposited during a period characterized by low rates of sedimentation and non-deposition (unconformity 1).

Well 211/29-9 is located in the centre of the study area and in a fault proximal position; indeed, the well path intersects with the fault at depth. The results of seismic stratigraphy would predict that this well samples a condensed section from the earliest post-rift. A more complete post-rift section is predicted in well 3/10b-1 to the south of the study area (no biostratigraphy was available for this well). In this well, the Cromer Knoll Group is 431 m thick and the suite of electrical well logs has a monotonous signature comparable to the overlying Shetland Group. Hence, the lower unit described from the Cromer Knoll Group is interpreted to represent condensed deposition of different lengths of time in the basin as the accommodation space was filled.

Unit 2 of the Cromer Knoll Group and the Shetland Group in 211/29-9 have a relatively monotonous log signature: Gamma ray count in the range 30-70 API, sonic velocity of 100-110 μs/ft and a bulk density reading of typically 2.45 gr/cc. On the basis of these measurements (no other tools were run in the post-rift interval), the group is interpreted as a quartz-rich siltstone. Abundant density (~2.7 gr/cc) and sonic (~90 μs/ft) spikes suggest thin carbonate-rich stringers; seismic interpretation maps these limestone beds (intra-shetland 1 and intra-shetland 2) and illustrates their lateral continuity throughout the basin. The carbonates are interpreted to represent periods of low clastic supply in an area otherwise dominated by hemipelagic sedimentation. Rates of sedimentation in modern environments dominated by hemipelagic sedimentation are in the range 5-15 cm/ka (Stow *et al.* 1996) – comparable with those calculated for the Shetland Group (when decompacted). The palaeo-environment is suggested to have been marine, below storm wavebase, with water depths probably in the range 50-750 m. As the remnant basin floor topography was filled during deposition of the Shetland Group, water depths shallowed.

Although sediment supply was dominantly fall out from the water column (hemipelagic/pelagic), the preserved strata is clearly not 'blanket-like' in distribution, indicating post-depositional transport. Two possible mechanisms for sediment movement are suggested. As described in Fig. 5.1a, the end syn-rift topography (bathymetry) of the basin floor was substantial. The initiation of gravity-driven slides and flows on slopes of <0.5° has been demonstrated in mud-rich delta front environments (Prior and Coleman 1978). With significant sea bed gradients, the redistribution of the silty sediments found in the study area by gravity slides, possibly evolving into turbulent flows (*cf.* Eyles and Lagoe 1998, their Fig. 11), appears a realistic model. The second mechanism of sediment redistribution proposed also results from the topography of the basin floor. In the early post-rift the topography represented by the footwall was probably up to 750 m, i.e. a submarine range front of >60 km length and up to 750 m elevation. Studies of modern deep water movements have suggested that the bathymetry may influence the path of bottom water currents and, in particular, currents will travel along the contours of a bathymetric slope (e.g. Pickering *et al.* 1989). If such currents were active in the study area during the Early Cretaceous then they provide a mechanism for sediment redistribution.

5.5 Controls on early post-rift (Cretaceous) sedimentation in the Strathspey-Brent-Statfjord area

The Strathspey-Brent-Statfjord area was sediment underfilled at the close of the Late Jurassic rift event. Remnant accommodation space was substantial; it can be estimated that only one third of the accommodation in the hangingwall to the Strathspey-Brent-Statfjord fault system was filled. A schematic reconstruction of the Early Cretaceous topography (bathymetry) in the basin is illustrated in Fig. 5.8. The western margin of the basin was defined by a linear, fault bounded range front, with an elevation locally in excess of 700 m above the proximal basin floor. The footwall scarp was significantly denuded during the period of active tectonics. To the east, pre-rift sediments exposed in the low angle hangingwall dip slope were onlapped by syn-rift geologies. In the deepest part of the basin, where syn-rift sediments were deposited, the basin floor had a subtle, and evolving, topography. In the study area, along strike displacement variations on the Strathspey-Brent-Statfjord fault had the consequence of producing a gentle, southward dipping slope in the hangingwall, i.e. dipping towards the position of maximum displacement on the fault. This slope was maintained, and enhanced, by differential compaction of syn- and pre-rift geologies; syn-rift sediments compacted more than pre-rift sediments, hence the volume of accommodation generated reflected the thickness of the underlying syn-rift deposits. By this mechanism, the depocentres established early in the syn-rift, and filled with syn-rift sediments, were also depocentres during the post-rift. As post-rift sediments were not deposited evenly in the basin, differential compaction continued to control accommodation generation into the Tertiary.

While differential compaction was the primary control on the distribution of accommodation during the post-rift, sedimentation processes were characterised by reworking. During the early post-rift period the Strathspey-Brent-Statfjord area was located in a mid-distal shelf setting. Rates of sediment supply were initially very low (Cromer Knoll Group). Following a basin wide period of non-deposition (unconformity 2), rates of supply increased 10-fold (Shetland Group). The dominant source of sediments was from the water column, i.e. deposition by hemipelagic settling. By this process of deposition, rates of sedimentation would be equal throughout the basin. However, deposition was not onto a flat basin floor, but rather a subtle topography. Gravity driven processes reworked the sediments into the deepest parts of the basin (with the consequence of continuing to enhance the effects of differential compaction). It is suggested that sediments of both the post-rift and the latest syn-rift (Kimmeridge Clay Formation) were reworked by these processes; hence syn-rift sediments were cannibalised into the post-rift. The basin was infilled from deep to shallow. This is schematically illustrated in Fig. 5.8. It is noted that similar mechanisms of sediments reworking were active during the syn-rift period, accounting for the lack of syn-rift deposits in the structurally high areas of the basin.

5.6 Conclusions

1. The Strathspey-Brent-Statfjord area is a Bathonian – Early Ryazanian age marine half-graben type basin bounded by the >60 km long Strathspey-Brent-Statfjord fault system. At the cessation of active extension the basin was substantially underfilled; remnant accommodation space resulting in a basin floor topography with up to 850 ms TWT elevation between footwall and hangingwall.

2. The loci of early post-rift deposition were not at the base of the linear footwall range front, but rather sited ~1.5 km basinward. This was a consequence of differential compaction in the hangingwall area, specifically a buttressing effect by the underlying footwall. Differential accommodation generation due to differential compaction of underlying sediments also had the consequence of defining post-rift depocentres mirroring the geometry of syn-rift depocentres.

3. A progressive infilling of the topography in the basin, from deep to shallow, is described. The post-rift sediments onlap the southwards dipping palaeo-slope in the hangingwall from north to south, i.e. the oldest sediments are in the south while an unconformity occurs at the base of the post-rift in the north. The effect of differential compaction in generating differential accommodation is recognised throughout the early post-rift, and the buried hangingwall of the Strathspey-Brent-Statfjord fault system remains the location of a topographic low in the basin into the Tertiary.

4. Remnant accommodation space within the basin was completely filled, i.e. the footwall was buried, ~60 Ma after the end of the rift event. Sedimentation rates were extremely low in the earliest post-rift (deposition of the Cromer Knoll Group), but increased after the Coniacian. A prolonged period of low sedimentation rates and non-deposition is recognised throughout the basin during the mid-Cretaceous.

5. Although the dominant process of sedimentation during the early post-rift was hemipelagic settling, post-rift sediments are not 'blanket-like' in distribution, indicating post depositional transportation. Two possible mechanisms are proposed: (1) Gravity-driven slumping and sliding, evolving into sediment flows, and (2) Current-driven transportation, possibly along the base of the bathymetric slope.

CHAPTER 6 The magnitude of footwall denudation of the Strathspey-Brent-Statfjord fault system, northern North Sea

6.1 Introduction

Sediment supply to rift basins can be either externally sourced, i.e. from a distal hinterland, or sourced from within the interior of the basin. The principal source of clastic sediments within an extensional basin is from the erosion of geologies in the uplifted footwalls of normal faults (e.g. the 'footwall-derived' sediments of Gawthorpe *et al.* 1994, see Fig. 2.9a). The footwalls of normal fault systems define relative topographic or bathymetric highs of asymmetric shape in cross section, bounded by a steeply dipping fault scarp and a low angle dipslope. These highs can be maintained for geologically significant periods of time as a single increment of coseismic slip will result in both uplift of the footwall and downthrow of the hangingwall, with respect to an original zero datum (Stein and Barrientos 1985). Hence, this is a source of sediment not depleted with time, but which grows, mirroring the accumulation of displacement on the fault. Although the volume of sediment liberated by footwall degradation is relatively small, certainly with respect to the volume of accommodation space generated in the adjacent hangingwalls during rifting (e.g. Stein *et al.* 1988, Contreras *et al.* 1997), this supply of sediment is important in basins without a significant external sediment source. For example, in basins located in a deep marine environment erosion of the footwall maybe the only source of clastic sediments and the processes of denudation and the dispersal of the sedimentary products will dictate the stratigraphic fill (e.g. Surlyk 1978, Ravnås and Steel 1998 see their Fig. 4C).

The processes by which fault bounded range fronts will degrade in a subaerial environment have received a great deal of attention (e.g. Harbor 1997, Allen and Hovius 1998, Densmore *et al.* 1998) and, consequently, the majority of landscape evolution models are subaerial in setting (Hovius and Leeder 1998). In recent years, numerous field and analogue studies have demonstrated that submarine downslope processes are comparable to subaerial processes, that is they are driven by gravity and constitute slides, slumps and flows (e.g.

Dingle 1980, Schwarz 1982, Stoker *et al.* 1998). However, largely because submarine slope failure processes are rarely observed, questions remain about the rates of slope denudation (and, hence, rates of sediment supply) and the primary controls on sediment redistribution and deposition.

The rôle of sediments sourced from footwall denudation in the filling of marine half-graben is examined in the following three chapters. The aims of this work address two issues,

1. To constrain the processes of denudation and consider the implications for the character of sediments sourced from these processes;

2. To gain a predictive understanding of the contribution of clastic sediments sourced from footwall denudation to the architecture of marine rift systems.

This and the succeeding chapter document the consequences of fault scarp denudation in the half-graben bounded by the >60 km Strathspey-Brent-Statfjord fault system; complementing the tectono-stratigraphic interpretation of this basin presented in Chapters 3-5. These results are based on analysis of a subsurface dataset, comprising five 3D seismic surveys and core, biostratigraphy and electrical log data from 13 wells. An overview of the denudation of the footwall of the Strathspey-Brent-Statfjord fault is documented in this chapter; defining and quantifying the area of scarp erosion, the magnitude of crestal retreat and the volume of pre-rift sediments eroded. A complex of displaced and degraded sediments, interpreted as the remnants of scarp denudation processes, are preserved mounted on the fault scarp of the Strathspey-Brent-Statfjord fault. The character of these sediments – the fault scarp degradation complex – is examined in Chapter 7, and the processes and sedimentary products of footwall erosion assessed. These results are contrasted with observations from field studies in central east Greenland, presented in Chapter 8. Excellent field exposures of the footwalls of latest Jurassic – Early Cretaceous fault systems in Greenland permit interpretation of the controls on the post-erosional scarp morphology. Chapter 8 concludes with a qualification of intra-basin sources of clastic sediment and examines the implications of these results for the stratal architecture and sedimentology of the basin fill.

6.2 Geological setting

The Strathspey-Brent-Statfjord fault system is a >60 km long, N-S to NNE-SSW striking, easterly dipping normal fault with a maximum displacement of in excess of 1.5 km. In the north, the fault splits into two strands (Fig. 6.1) – the Statfjord (west) and Statfjord East fault segments – which tip out within the study area, the fault continues beyond the south of the study area to bound east of the North Alwyn structure (Fig. 2.11a). The fault propagated and grew from an array of shorter fault segments during the Late Jurassic (Bathonian – early Ryazanian), in response to a W-E oriented regional extensional stress regime. The northern North Sea failed rift accommodates strain of ~15 % (e.g. Yielding *et al.* 1992) on a number of large normal fault arrays, sub-parallel to the Strathspey-Brent-Statfjord fault system and centred around the North Viking Graben (see Enclosure 1). The evolution of the fault population in the study area and growth of the Strathspey-Brent-Statfjord fault array are described in Chapter 3.

The half-graben bounded by the Strathspey-Brent-Statfjord fault system was sediment starved during the rift event and, as a consequence, significantly underfilled (see Chapter 4). Syn-rift geologies belong to three formations: The Tarbert Formation of the Brent Group and the Heather and Kimmeridge Clay (or Draupne) formations of the Humber (Viking) Group (Richards *et al.* 1993; Fig. 4.2 and Enclosure 2). The Tarbert Formation is a sand-rich package deposited in a marginal marine environment during the punctuated southward retreat of the Brent delta complex; retreat controlled by increasing rates of tectonic subsidence and forced regional sea level rise (e.g. Johannessen *et al.* 1995, Råvnas *et al.* 1997, Davies *et al.* in review). The Heather and Kimmeridge Clay formations are characterised by mud-rich sediments, dominated by marine geologies of hemipelagic origin. In the Strathspey-Brent-Statfjord area, these deposits record the declining rate of sedimentation and the increase in water depths during the rift event. Syn-rift sediment starvation is ubiquitous throughout the East Shetland Basin, typically demonstrated by a substantial palaeo-topography between the crest of the footwall and the top of the syn-rift deposits in the hangingwall (i.e. remnant accommodation), commonly in excess of 500 m (for example see Fig. 3 of Underhill *et al.* 1997).

Erosion of normal fault bounded footwall scarps is documented throughout the East Shetland Basin by shallowing of the fault scarp and crestal retreat (e.g. Alwyn North, Johnson and Eyssantier 1987; Brent, Coutts *et al.* 1996; Gullfaks, Færseth *et al.* 1997;

Ninian, Underhill *et al.* 1997; Snorre, Dahl and Solli 1992; Statfjord, Hesthammer and Fossen 1999). In addition, a complex of structurally deformed and sedimentologically reworked sediments is frequently described mounted on the denuded scarp. These sediments have been termed the 'fault scarp degradation complex' by Underhill *et al.* (1997) and are considered to be the remnants of mass wastage events on the slope (Livera and Gdula 1990). The fault scarp degradation complex is composed of displaced pre-rift geolgies: The sand-rich, deltaic Brent Group, the muddy, shallow marine deposits of the Dunlin Group and the fluvial sandstones of the Banks Group (Fig. 4.2). The geology of the fault scarp degradation complex in the study area is described in detail in the following chapter.

Although documented regionally, detailed studies of the fault scarp degradation complex have only been published for the Brent (Coutts *et al.* 1996, McLeod and Underhill 1999), Ninian (Sawyer and Keegan 1996, Underhill *et al.* 1997) and Statfjord (Hesthammer and Fossen 1999) areas of the northern North Sea. These authors describe the complex to be of comparable form – with a ramp-flat basal décollement surface in cross section perpendicular to the strike of the fault and an amalgamated cuspate geometry in plan view. Internally, pre-rift geologies of Brent, Dunlin and Banks groups are recognised, these sediments show little deformation and are stratigraphically the right-way-up. From these observations, geologies of the fault scarp degradation complex are interpreted to be the remnants of translational and rotational slides exploiting muddy décollement surfaces in the Dunlin and lower Banks groups. Sliding is considered the dominant mass wastage process on the scarp during the syn-rift, with secondary shallow erosion by mass flows. Redeposition of these sediments in the proximal hangingwall in a talus complex of stacked slide blocks has been interpreted by Underhill *et al.* (1997) and McLeod and Underhill (1999).

6.3 Dataset and methods

Five 3D seismic surveys (*strathspey_TOMS*, *brent_d*, *lm96083*, *st9101* and *e86*) were available to this investigation (Fig. 6.1), with a line spacing of 12.5-25 m for both inlines and crosslines and a vertical resolution of 15-30 ms TWT. The seismic data have not been depth converted and, hence, have a vertical scale in two-way-time (TWT). These data were interpreted on the basis of correlation with electrical well logs using velocity data. Significant reflectors in the pre-rift succession were extrapolated from 8 vertical exploration wells penetrating the footwall of the Strathspey-Brent-Statfjord fault system (located on Fig. 6.1). These picks were also determined in 13 wells with paths intersecting geologies of the fault scarp degradation complex. Further details on the seismic dataset and the density of seismic interpretation are contained in Appendix 1.

Three key seismic reflectors were mapped in the footwall of the Strathspey-Brent-Statfjord fault system (Fig. 6.2). From well correlations these reflectors are shown to correspond with near the top of the Brent Group (top pre-rift onlap surface), the top of the Statfjord Formation and the top of the Hegre Group (Fig. 4.2). Each reflector represents a sharp change in the petrophysical characteristics of the sediments (i.e. lithology change) and, hence, has a strong seismic response. These reflectors have excellent lateral continuity across the study area. Reflectors corresponding with pre-rift geologies are typically truncated in the eroded footwall scarp; in order to construct a subcrop map for the scarp face and to quantify the crestal retreat, seismic markers were employed to describe the truncations of reflectors (Fig. 6.2). The reflectors are truncated below the fault scarp degradation complex; the definition of this complex and the seismic character of the denuded sediments are described in the succeeding chapter. In addition to mapping reflectors in the pre-rift, the strong positive reflector describing the top of the syn-rift (base Cretaceous) was mapped in both the footwall and hangingwall to the fault array. This reflector is a marked seismic onlap surface; post-rift sediments of the Cromer Knoll and Shetland groups onlapping the footwall high. The half-graben bounding Strathspey-Brent-Statfjord fault array was mapped throughout the study area (Chapter 3). Normal faults offsetting pre-rift geologies in the footwall were mapped with a vertical resolution (minimum throw) of ~20 ms TWT.

The denuded crestal area of the footwall of the Strathspey-Brent-Statfjord fault system was reconstructed by extrapolating the trace of the fault and the pre-rift reflectors in the footwall (see Fig. 6.2). The location of the palaeo-crest (pre-erosion) was described by a seismic marker. It can be estimated that the error in the depth of the palaeo-crest, due to manual vertical extrapolations on data with a vertical axis in time, was ±50 ms TWT. The volume of sediments eroded from the footwall was determined using the hydrocarbon reserves evaluation software, REP. Surfaces describing the reconstructed crest and the top syn-rift on the scarp were contoured and then depth converted using the time-depth function for well 211/30-1 (Fig. 6.3). It is recognised that this technique introduces a number of errors, particularly undertaking depth conversion on a single point of information. Hence, the value derived is not an absolute, but an estimate. This technique is discussed further below.

6.4 The structure of the footwall of the Strathspey-Brent-Statfjord fault

The structure of the footwall of the Strathspey-Brent-Statfjord fault system is described by the maps of the depth (in ms TWT) to the reflectors within the pre-rift (top pre-rift, top Statfjord Formation and top Hegre Group) and the top syn-rift (base Cretaceous) reflector. These maps are presented in Fig. 6.4. A post-tectonic regional tilt of 2-3° towards the south and the effects of differential compaction during burial to depths in excess of 2 km have modified the morphology of all these surfaces. The topography described by the form of the top syn-rift reflector represents the topography of the basin floor at the end of the rift phase. The hangingwall of the fault system is defined by a relative low on the map of the top syn-rift reflector (Fig. 6.4a), and the footwall by a relative high – testifying to the underfilling of the basin during the syn-rift. The divide between footwall and hangingwall is described on Fig. 6.4a by a white dashed line, which approximates the lower limit of degradation on the scarp.

Maps of depth (in ms TWT) to the near top Brent Group, top Statfjord Formation and top Hegre Group reflectors in the footwall demonstrate that the pre-rift geologies have a shallow, but constant, westerly dip, veering to a northwesterly dip in the Statfjord area (Fig. 6.4b-d). Consequently, the footwall stratigraphy is described as layer-cake and structurally simple. This supports the observations of the footwall well correlation shown in Fig. 6.5. The easterly truncation of each of these surfaces is erosional, in the denuded scarp, or tectonic, by the fault (Fig. 6.2).

Overprinted on the westward dip of the pre-rift sediments is a gentle N-S oriented structuration (see definition on Fig. 6.4b). This is clearly represented by the morphology of the top syn-rift (base Cretaceous) reflector. This reflector directly overlies pre-rift geologies on the eroded footwall scarp and at the crest of the footwall, where no, or seismically unresolvable, syn-rift deposits are preserved. Distinct highs and lows at the crest of the block can be described (see Fig. 6.4a), which mirror the locations of closure of the wide, low amplitude structures described by the pre-rift reflectors. Highs occur at 7-10 km spacing. Lows are approximately 100-150 ms TWT deeper than their adjacent highs and three major crestal lows in the footwall are identified, between the Strathspey and Brent structures, the Brent and Statfjord structures and between the Statfjord and Statfjord (west) structures (annotated A, B and C respectively, on Fig. 6.4a). Hence, the crest of the block can be described as a series of low amplitude (<150 m) folds (with W-E veering to NW-SE oriented hingelines) on a 7-10 km scale. These structures lie to the west (down dip) of the crestal erosion, thus, are primary structures and not erosional.

The folds defined on Fig. 6.4b are interpreted as fault related extensional folds (after Schlische 1995). They developed in response to along strike variations in displacement on the Strathspey-Brent-Statfjord fault when it was composed of a number of shorter, unlinked segments. This is supported by the coincidence of the hinges of synclines with displacement minima on the Strathspey-Brent-Statfjord fault system (see Chapter 3, Fig. 3.7) and with the location of intra-basin highs in the hangingwall. The amplitude and resolution of these structures is substantially less than the fault related folds described by the topography of the top pre-rift reflector in the hangingwall of the fault (see Chapter 4; Fig. 4.6). This reflects the distribution of coseismic slip on a normal fault, the majority resulting in downthrow of the hangingwall and typically less than one fifth resulting in footwall uplift (e.g. Stein and Barrientos 1985) – the ratio of uplift to downthrow defining the magnitude of the fault controlled topography.

The maps of reflectors in the pre-rift also describe the locations and geometries of a number of normal fault populations in the footwall of the Strathspey-Brent-Statfjord fault system. These faults offset the pre-rift geologies and control the thickness of the syn-rift sediments. A map of the distribution of the seismically resolvable faults in the footwall is shown in Fig. 6.6. The faults are shown where they offset the reflector corresponding with the top of the Statfjord Formation and are colour coded with respect to the magnitude of maximum throw on the fault.

Three populations of faults are identified, each associated with the locations of the main crestal lows, identified as A, B and C on Fig. 6.4a. The southernmost of the populations (A on Fig. 6.6), located between the Strathspey and the Brent structures, is composed of nine normal faults with seismically resolvable displacement. The largest of these has a throw of ~80 ms TWT, although the majority have sub-25 ms TWT throw. The faults offset pre- and syn-rift sediments, have thickened syn-rift in the hangingwall, tip at different depths the syn-rift and do not affect the post-rift, i.e. are of syn-rift age (Fig. 6.7a). All of these faults, and those of the central population, are oriented perpendicular to the strike of the main fault system and have no preferred dip direction; although within each population the largest displacement faults define graben-like structures (Fig. 6.8). The central population (B), located between the Brent and the Statfjord structures, has ten normal faults of up to 200 ms TWT throw, also of syn-rift age. The largest of the faults dips to the northeast and demonstrates significant thickening of syn-rift sediments in the hangingwall (Fig. 6.7b).

The orientation of the faults in populations A and B is parallel to the axis of principle stress (σ_3) in the basin, an orientation not compatible with structures developed to accommodate the regional stress. This study interprets these faults as syn-rift age accommodation structures originating from stress concentration during the uplift and folding of brittle geologies in the footwall. A sketch interpretation of the distribution of stresses and the deformation of the footwall of the Strathspey-Brent-Statfjord fault system during W-E oriented extension is illustrated in Fig. 6.8c. In this model, displacement variations along the strike of the Strathspey-Brent-Statfjord fault system result in steep displacement (and hence topographic) gradients at the tips of fault segments. Tensile stresses, with an orientation perpendicular to σ_3, are concentrated at the segment boundaries. Brittle failure and the establishment of local fault populations in the footwall are the response. Hence, the steep limbs of the fault related extensional folds in the footwall are dissected by normal faults. The fault populations are distributed coincident with displacement minima on the main throughgoing fault system where displacement gradients would have been highest. Extensional stresses in pre-rift sediments are accommodated by brittle rather than ductile strain – a comparable pattern of normal faults (below seismic resolution) are predicted at structural highs in the hangingwall, defining faulted antiforms (horst blocks).

The timing of displacement accumulation on the footwall faults, with respect to the evolution of the Strathspey-Brent-Statfjord fault and the progress of the rift event, is uncertain, as the syn-rift geologies are thin. However, the higher displacement faults can be mapped through the majority of the syn-rift sediments; indicating that the faults were active late in the rift event, which would be compatible with this interpretation. It is also observed that the maximum displacement on these faults is at a shallow structural level (i.e. the offset of syn-rift or late pre-rift sediments), and that displacement decreases with depth (*cf.* tension cracks).

The northernmost population of footwall faults occurs in the Statfjord (west) area (C) and, in contrast to the other two groups, the faults have a conjugate pattern (oriented NNW-SSE and NNE-SSW); hence do not strike perpendicular to the main fault system. The throw on this group of eleven normal faults is typically small (<25 ms TWT), bar one higher displacement structure which occurs isolated, 2-3 km to the south of the others. All of these faults are of syn-rift age, again demonstrating syn-rift thickening in the hangingwall. The stresses responsible for the development of these faults are unclear; however, it is observed that these faults are located in the area of overlap of the Strathspey-Brent-Statfjord and Zeta-Snorre-Gullfaks fault systems. It is suggested that these structures reflect the complex stress regime resulting from the interaction between two large normal fault arrays and, in particular, the uplift of the Statfjord East area in the footwall of the Zeta-Snorre-Gullfaks fault array during the latest syn-rift.

6.5 The magnitude of crestal retreat and scarp denudation

In addition to the regional mapping of reflectors corresponding with the top of the pre-rift, the top of the Statfjord Formation and the top of the Hegre Group, the point of erosional truncation of these reflectors in the denuded footwall scarp of the Strathspey-Brent-Statfjord fault system was defined by a seismic marker (Fig. 6.2). From this interpretation, a map of the subcrop of the pre-rift geologies in the fault scarp was determined (Fig. 6.9a). The pattern of subcrop has N-S linearity, veering towards NE-SW in the Statfjord area; oriented parallel to the strike of the Strathspey-Brent-Statfjord fault system. The reflector describing the top of the Hegre Group is exposed (and the Hegre Group below degraded) in the footwall scarp along most of the length of the fault studied. Where the fault splits into two strands of lower displacement – the Statfjord (west) and Statfjord East faults – the Statfjord Formation and Hegre Group are buried in the footwall and not exposed in the scarp.

The map presented in Fig. 6.9b shows the upper and lower limits of erosion on the scarp of the Strathspey-Brent-Statfjord fault system. The upper limit corresponds with the location of the eroded crest of the footwall. Also delineated is the position of the crest as it would have been without any scarp erosion, as determined by crestal reconstruction (see definition in Fig. 6.2). The area of the eroded scarp face is shaded; the area affected by erosion totals 167.7 km^2 for the entire study area. The scarp is degraded along the entire length, including both the Statfjord (west) and Statfjord East structures and on the abandoned hangingwall splay fault (*stat_01*) between the Brent and Statfjord structural blocks.

The distance by which the crest of the footwall has retreated (distance between the upper limit of erosion and the position of the palaeo-crest) and the width of the denuded scarp (distance between the upper and lower limits of erosion) was measured at regular intervals perpendicular to the strike of the bounding fault system. For the Statfjord, Brent and Strathspey areas, these two values remain fairly constant – the backstepping of the crest is around 2-2.5 km and the width of the eroded scarp is typically in the range 2.5-3.5 km. As would be predicted by the subcrop patterns (Fig. 6.9a), the width of both these measures sharply decreases where the Strathspey-Brent-Statfjord fault splits into the Statfjord (west) and Statfjord East fault segments, and declines northwards.

The boundaries defined by the seismic reconstruction of the eroded crest of the Strathspey-Brent-Statfjord fault system and the top syn-rift reflector on the fault scarp describe the volume of material eroded from the scarp. Using these two surfaces, the total volume of pre-rift geologies eroded from the footwall of the Strathspey-Brent-Statfjord fault system was calculated to be 35.24 km^3. As discussed in Section 6.3, this measure is only an estimate due to errors introduced by the methods employed in the calculation. In direct comparison, Nøttvedt *et al.* (2000) estimate that approximately 3.5 km^3 of Brent Group sediments have been denuded from the main Statfjord and the Statfjord East structures. This study investigates an area at least three times larger than that of Nøttvedt *et al.*, and quantifies the total volume of erosion (including the Dunlin and Banks groups). Hence, the value derived by this study is considered of a comparable order of magnitude to that determined by Nøttvedt *et al.*.

Limitations imposed by the software available to this study constrained the quantification of sediment supply from footwall erosion. Future work, using the GeoFrame software package, will separate the volumes of Brent, Dunlin and Banks groups eroded and, ultimately, qualify the component of sand grade material liberated by footwall denudation. When integrated with the model for fault array evolution (Chapter 3), this should permit reconstruction of the temporal and spatial changes in sediment supply in the basin. In addition, further seismic interpretation using this software will examine the accommodation space generated by tectonic subsidence in the hangingwall and, hence, permit the disparity between the supply from footwall erosion and the volume of accommodation to be quantified.

Although only preliminary observations, the results presented in this chapter do demonstrate a conspicuous along strike continuity in the magnitude of erosion of the fault scarp and in the post-erosional morphology of the slope. This linearity is interpreted to reflect the lack of subaerial exposure of the scarp; the consequence of wave action would have been to flatten the crest of the structure (see, for example, the Snorre footwall, Fig. 4.23). A map of the dip of the scarp face is presented in Fig. 6.10 (based on interpretation of the top syn-rift reflector, Fig. 6.2). The subcrop of the top pre-rift and the top of the Statfjord Formation in the scarp are illustrated. The dip of the slope between these two markers is extremely consistent (0.1-0.2 ms/m equates to ~6-12°). In addition, the locations of palaeo-segments of the fault system have a similar magnitude of denudation and angle of slope to the locations of maximum displacement.

The mapped morphology of the footwall scarp of the Strathspey-Brent-Statfjord fault system is the product of a prolonged period of uplift, syn-tectonic denudation and post-rift submarine exposure. Slow rates of post-rift sedimentation resulted in continued exposure of the scarp for ~60 Ma after the end of the rift event (Chapter 5). Although a seismic trigger for slope failure was absent during the post-rift period, it is likely that the slope continued to degrade. Gravity driven processes were the primary mechanisms of denudation. Recent studies have demonstrated that such gravity driven processes are dominant in the rapid denudation of elevated areas (Hovius *et al.* 1997). However, these processes can only be triggered on steep slopes and rely on modification of the slope morphology by fluvial processes (i.e. undercutting triggering landslides). In a submarine setting the angle of the slope is controlled only by gravity driven processes. Hence, the scarp angle can shallow to the point that gravity driven failure can no longer be initiated (as the forces resisting slope movement are greater than the forces driving downslope motion, e.g. Newmark 1965). By implication, the mapped morphology of the scarp reflects the stable end member of the denudation process. This proposal is examined further in the discussion in Chapter 8.

6.6 Conclusions

1. The topography of the footwall of the Strathspey-Brent-Statfjord fault system reflects along strike displacement variations on the fault. A number of low amplitude, large wavelength folds are identified and attributed to fault related folding of the footwall during differential uplift.

2. Clusters of normal faults oriented perpendicular to the strike of the main fault are described offsetting pre- and syn-rift geologies of the footwall. The fault populations are located at crestal lows, the culmination of footwall synclines. The faults are interpreted as syn-rift accommodation structures developed in response to local extensional stresses in the footwall. Brittle, rather than ductile, strain accommodated the majority of the tensile stresses in the footwall.

3. The footwall of the Strathspey-Brent-Statfjord fault system is substantially denuded. The area of the eroded scarp face is ~168 km^2 and the footwall crest has been backstepped by up to 2.5 km. The total volume of degraded sediments is estimated to be ~35.2 km^3. The magnitude of footwall denudation and the morphology of the scarp face demonstrate marked along strike continuity. This is interpreted to be a consequence of the prolonged exposure of the scarp, during both the syn- and post-rift, and the limitations of the denudation processes in a submarine environment.

CHAPTER 7 The geology of the fault scarp degradation complex, Strathspey-Brent-Statfjord fault system, northern North Sea

7.1 Introduction

Field studies and analogue models have demonstrated the similarity between processes of gravity driven mass wastage on subaerial and submarine slopes (e.g. Prior and Coleman 1979, Schwarz 1982, Nemec *et al.* 1988). The potential processes of hillslope failure are a continuum of transverse displacement (including creep, slides and flows), describing an increase in the degree of dislocation and disaggregation of the transported material (e.g. Varnes 1978, Nemec 1990). For slopes located in a subaerial setting, the controls on, and the rates of, gravity driven downslope transport processes are well understood; predictive numerical models constrain the conditions leading to slope failure (Jibson and Keefer 1993). In contrast, the controls on downslope processes for slopes located in a submarine environment remain poorly understood. This is partly a consequence of the difficulties inherent in studying modern submarine slopes (although shallow seismic technology is permitting greater investigation of this problem, e.g. Nielsen *et al.* 1998), but also a more general lack of recognition of the geological remnants of slope failure processes (of both a submarine and subaerial setting; e.g. Woodcock 1979). The low preservation potential of topographic highs (i.e. range fronts) and the rapid dispersal of denuded sediments both contribute to this problem.

Normal fault scarps are unique slopes in the respect that movement on the fault will maintain a relative positive elevation (i.e. the slope) for a geologically significant time. These slopes will be prone to gravity driven, seismically triggered denudation processes (e.g. Keefer 1988, 1999); supplying reworked sediments to the proximal hangingwall basin. In sediment starved basins, for example the Late Jurassic submarine half-graben in the northern North Sea, fault bounded range fronts of significant elevation can be established. The denudation of such scarps occurs over a long period of time and the geological products are transported into an environment with low sediment supply from other sources.

Examination of the degradation of fault scarps in these settings provides the opportunity to constrain the character of sediment supplied from footwall denudation and assess the controls on, and the rates of, submarine erosion processes.

This study aims to reconstruct the downslope processes by which the footwall of the Strathspey-Brent-Statfjord fault system was denuded through investigation of the reworked sedimentary products. Analysis of subsurface data from the northern North Sea has described underfilled half-graben type basins bounded by degraded footwall scarps and the debris of ancient slope failure events, preserved both mounted on the scarp and in the proximal hangingwall (e.g. Dahl and Solli 1992, Coutts *et al.* 1996, Færseth *et al.* 1997, Underhill *et al.* 1997, Hesthammer and Fossen 1999; see also chapters 4 and 6). In particular, this chapter examines of the character of a complex of displaced sediments preserved mounted on the footwall scarp, termed the fault scarp degradation complex (after Underhill *et al.* 1997). The sedimentologically reworked and structurally disturbed geologies of the fault scarp degradation complex have been interpreted as the geological remnants of the scarp degradation processes (e.g. Livera and Gdula 1990). This complex, although laterally continuous along the scarp, is thin, typically <100 m in vertical thickness. Hence, while the boundaries of the complex can be defined and mapped (see McLeod and Underhill 1999), the internal structure is below the resolution of the seismic dataset. However, from examination of core samples and electrical log signatures it is possible to demonstrate that translational sliding was the primary mechanism of downslope transport and identify the principle décollement surfaces.

7.2 Dataset and methodology

The geology of the fault scarp degradation complex was investigated using a subsurface dataset, comprising five 3D seismic surveys and electrical logs, core samples and biostratigraphy from well penetrations. The areal extent of the surveys is illustrated in Fig. 7.1, as are the locations of wells in the hangingwall and footwall to the Strathspey-Brent-Statfjord fault system discussed in this chapter, including the 14 wells with core samples from the fault scarp degradation complex. The seismic data are described in more detail in Chapter 6 and Appendix 1. Chapter 6 also assesses the regional geological setting of the study area and previous studies on the geology of fault scarp degradation complexes (Section 6.2).

Seismic interpretation of the fault scarp degradation complex follows the convention of McLeod and Underhill (1999, their Fig. 5). The thinness of the fault scarp degradation complex (typically <150 ms TWT) and a chaotic, seismically transparent seismic character precludes the interpretation of reflectors from within the complex; however, the upper and lower bounding reflectors could be confidently defined (Fig. 7.2). The upper bounding surface is a seismic onlap surface, with the eastward dipping reflectors of the post-rift Cromer Knoll and Shetland groups onlapping the chaotic reflectors of the fault scarp degradation complex. This horizon corresponds with the top syn-rift reflector discussed in previous chapters, and was mapped regionally with confidence. By this definition, the fault scarp degradation complex is an integral part of the syn-rift. The lower bounding surface, essentially the base of the syn-rift on the scarp, is described by the erosional truncation of the gently dipping pre-rift geologies in the scarp (see Chapter 6). The lower surface was not mapped regionally, although the horizon could be described throughout the study area, but only in the vicinity of wells with cored intervals. Both of these surfaces are of seismic stratigraphic significance, but do not describe timelines or structural discontinuities (fault planes).

The fault scarp degradation complex is intersected by in excess of 150 wells in the study area (see, for example, Coutts *et al.* 1996, Hesthammer and Fossen 1999) but, of these, only 14 have samples of core from the complex; the majority of the cored wells are located in the Statfjord area. These wells are located on Fig. 7.1 and a summary of the key observations and techniques applied in each well are presented in Fig. 7.3. Electrical well logs, biostratigraphy and core samples from the 14 cored wells were available to this study, but no image logs, dipmeter data or production (pressure) results were accessible. Conventional core analysis (CCA) data from ½" core plugs, in the form of porosity, permeability (vertical and horizontal air, Klinkenberg corrected) and grain density, were supplied for the 12 wells in the Statfjord area by Statoil.

Approximately 1350 m of core from the fault scarp degradation complex was logged for structure and sedimentology, including fault counting (per metre of core) and a visual measure of porosity and permeability; detailed facies analysis was not undertaken. In addition, a complete Brent Group succession, cored in exploration well 33/9-3 from the footwall of the Strathspey-Brent-Statfjord fault, was logged as a standard for facies and formation interpretation and to characterise the pre-existing deformation. Logging was undertaken at 1:15 – 1:500 scale on generally good quality A-cut sections at Statoil (Stavanger) and Shell (Aberdeen) corestores. Graphic logs are presented in Appendix 4.

For 4 wells, the hole describing the location of the core plug samples (used for CCA measurements) was characterised and a 'k-factor' assigned to the plug (see Fig. 7.15). This factor describes the host rock lithology, the number of faults intersecting the plug, the type of the faults and the orientation of the faults with respect to bedding. The k-factor was directly compared with the CCA data to quantify the effects of brittle deformation on the porosity and permeability of the bulk rock. The observations of k-factor and comparisons with CCA data are presented in Appendix 5 along with the results of fault counting and porosity determinations.

Further investigations of the low displacement brittle deformation were undertaken on the microscopic scale. 37 localities in 6 wells were marked by the author for ½" core plugs (see Fig. 7.3). The plugs were taken with their long axes parallel to bedding and polished thin sections, oriented perpendicular to bedding, were made by Statoil. Thin sections were examined using standard microscopic techniques and a number were carbon coated for study using the Phillips XL 30CT scanning electron microscope (SEM) at the University of Edinburgh. Of particular interest were grain size, shape and orientation and the mineralogy of fault zones (with respect to the host rock). Semi-quantitative analysis of grain composition was also undertaken on the SEM. Using back-scattered electron (BSE) images, crude image analysis was undertaken to quantify microporosity of the bulk rock and within the low displacement faults. This technique involved 'binning' a range of colours (from grades of greyscale) and pixel counting those assigned to porosity (as a proportion of the total) (see Fig. 7.17). 25 images, 14 of faults and 11 of the undeformed host rock, from 6 thin sections of the Rannoch Formation in well 33/9-C6 were analysed. This technique, including the errors involved and sensitivity of results, is discussed further in Section 7.5.3.

7.3 The external morphology of the fault scarp degradation complex

This study follows the definition of the seismic form of the fault scarp degradation complex of Underhill *et al.* (1997). The complex described by Underhill *et al.* is mounted on the footwall scarp of the fault system bounding the Ninian hydrocarbon field, located to the west of the Strathspey-Brent-Statfjord area. A line drawing of the interpretation of

Underhill *et al.* is reproduced in Fig. 7.4a (their Fig. 15) and compared with the morphology of the complex in the study area (Fig. 7.2). The two sections, both oriented perpendicular to the strike of the bounding faults, illustrate the similarity in the morphologies of the fault scarp degradation complexes. Four key features are recognised,

1. Seismic reflections within the two complexes are weak and chaotic, the complexes are thin (0-150 ms TWT) and, hence, difficult to resolve on the seismic dataset;

2. In contrast to the internal form, the boundaries of the fault scarp degradation complexes can be clearly defined. The top is a clear onlap surface, with gently dipping syn- and post-rift reflectors onlapping the palaeo-slope from the east. This surface corresponds with the top syn-rift reflector described above. The base of the fault scarp degradation complex is defined by truncation of the sub-parallel, westward dipping pre-rift reflectors in the footwall;

3. In section perpendicular to the strike of the main fault system, the top bounding surface is easterly dipping and near planar in form, but the lower boundary has a distinct ramp-flat morphology. Steep ramps correlate with the subcrop of the sand-rich Brent and Banks groups in the footwall below, and flats overlie the muddy Dunlin Group and the top of the Hegre Group. A dip map of the outer bounding reflector illustrates the near planar form of the scarp and consistency of the dip (generally <20°) across the study area (Fig. 6.10);

4. The complexes are arranged as multiple tiers on the scarp, defined by the ramp-flat lower bounding surface.

These features are described in published examples of the fault scarp degradation complex throughout the northern North Sea (e.g. Alwyn North, Johnson and Eyssantier 1987; Brent, Coutts *et al.* 1996; Gullfaks, Færseth *et al.* 1997; Snorre, Dahl and Solli 1992; Statfjord, Hesthammer and Fossen 1999).

Using these definitions of the bounding surfaces of the fault scarp degradation complex, McLeod and Underhill (1999) describe an envelope of degradation products and present a map of the thickness of the complex for the northern Brent area (from interpretation of the *brent_d* survey). This map is reproduced in Fig. 7.4b (from their Fig. 6). Similar techniques have been employed in-house at Shell (for the Brent area), Statoil (for the Statfjord area, 'base of slope failure' surface defined by Hesthammer and Fossen 1999) and Texaco (for the Strathspey area) with comparable results. The key observations on the thickness and distribution of the fault scarp degradation complex are,

1. The complex has two linear (in an along strike sense) zones of maximum thickness – a structurally shallower zone overlying the subcrop of the Brent Group and a deeper zone overlying the Banks Group. The complex thins between the two zones, where it overlies the Dunlin Group, supporting the observation of multiple tiers (e.g. Fig. 7.2).

2. Within the linear thicker zones, the thickness variations within the fault scarp degradation complex describe 'spoon-shaped' packages in map view; with maximum thickness at the centre, thinning to the margins. The cusps are of 0.25-1 km in along strike width and 0-150 ms TWT thick.

On the basis of these geometric observations previous workers on the fault scarp degradation complex have interpreted the complex as the remnants of gravity driven rotational-translational slides (see particularly Livera and Gdula 1990, Coutts *et al.* 1996, Underhill *et al.* 1997, Hesthammer and Fossen 1999, McLeod and Underhill 1999). Key décollement surfaces are described in the Ness Formation, the mid-Dunlin Group and the lower Banks Group.

7.4 The sedimentology and structure of the fault scarp degradation complex

Due to its size, complexity of internal structure and the masking effect of the strong top syn-rift (base Cretaceous) reflector, the internal structure of the fault scarp degradation complex is below the resolution of the seismic data (Fig. 7.2). Direct evidence of the sedimentology and structure of the sediments in the complex is only available through investigation of core samples and, indirectly, from electrical log signatures. The observations and interpretations presented in this, and the succeeding section, are from study of the 14 wells in the Strathspey-Brent-Statfjord area which have core samples from the fault scarp degradation complex (a total length of core of ~1350 m). These wells are located on Fig. 7.1 and a summary of the key observations from each cored interval is presented in Fig. 7.3.

Seismic interpretation has demonstrated that the fault scarp degradation complex has two, strike-parallel, zones of maximum thickness which overlie the subcrop of the Brent Group and Statfjord Formation. These zones of maximum thickness are termed in this study zones S and B, respectively (see Fig. 7.2). Zone B, the structurally higher zone of the complex, is a secondary hydrocarbon reservoir and its position on the scarp is structurally above the hydrocarbon bearing area of the Statfjord Formation. Hence, zone B has a dense suite of well penetrations (over 150 in the study area), largely targeted at the underlying Statfjord reservoir. 14 of these wells have cored sections, the majority are located in the Statfjord area and no cored wells are located in the Strathspey and southern Brent areas (Fig. 7.1). Zone S, in contrast, has a less economically significant position and, hence, far fewer well penetrations. As a consequence of this distribution of data the study of the fault scarp degradation complex has been focused on zone B and, indeed, the work described below is concerned only with that zone.

7.4.1 Overview of the sedimentology of core samples from zone B of the complex

All the core samples from zone B of the fault scarp degradation complex in the study area were logged for both sedimentology and structure. It is recognised that cored sections represent an extremely small sampling of the complex and that the cored wells have a wide distribution across the study area (Fig. 7.1). The scale of the sample is trivialised further as the complete section of the fault scarp degradation complex intersected in each well is generally not cored; hence, different wells sample a different part and a varying proportion of the complex. Consequently, the core data represents limited, widely spaced and very diverse sampling of the fault scarp degradation complex. However, similarities in structure and sedimentology between the core samples in different wells are frequently observed (see Fig. 7.3), in particular,

1. In at least 10 of the 14 wells, stratigraphically consistent (i.e. depositionally younger sediments overlying older sediments) packages composed of the formations of the Brent Group are observed. In these cored intervals, the characteristic primary sedimentological features of the formations are generally well preserved, the sediments show little evidence of post-depositional tectonic deformation and are directly comparable with *in situ* pre-rift deposits in the footwall;

2. In contrast, core samples from 3 of the wells are highly deformed, exhibiting both ductile and brittle deformation. These sediments are interpreted the deposits of

debris flows and slumping. Geologies of the Brent Group are recognised as clasts incorporated into these sequences, but the sediments are partially disaggregated and deformed, having lost their primary depositional characteristics and depositional sequence.

From each of the 14 wells with core samples from the fault scarp degradation complex (bar wells 33/9-C31 and 33/12-B18 which sampled deformed Dunlin Group geologies), observations from the cored interval were consistent with one or other of these conclusions. The wells located on Fig. 7.1 are annotated as either type 1, type 2 or, rarely, type 1/2 corresponding with the key observations above. The ubiquity of the observations from a very disparate sample population suggests that the cored intervals may be representative of the structure and sedimentology of the fault scarp degradation complex as a whole across the study area. The descriptions below expand on these two observations, with examples from specific wells. The small scale deformation, i.e. low displacement faults and fractures, observed in the cored intervals is documented below but examined in greater detail in Section 7.5.

7.4.2 Type 1: Core samples from the fault scarp degradation complex incorporating undeformed sediments assigned to the Brent Group

This category includes all the cored intervals from the fault scarp degradation complex from which sediments of affinity to the formations of the Brent Group are interpreted. These geologies show very little tectonic disturbance and generally maintain the depositional stratigraphy of the Brent Group (i.e. Tarbert Formation overlies Ness Formation etc.). These characteristics are interpreted in core from 10 of the 12 wells – wells 33/9-9, 33/12-1, 33/9-A17, 33/9-A23, 33/9-B6, 33/9-B23A, 33/9-C6, 33/9-C11, 211/29-C15 and 211/29-D38 (Appendix 4). The description below examines core samples from two wells in detail, 33/9-C6 and 33/9-C11; these core are considered representative of the group and demonstrate the main structural and sedimentological features.

Wells 33/9-C6 and 33/9-C11 are deviated production wells located on the footwall of the Statfjord (west) fault segment (Fig. 7.1). Seismic interpretation (Fig. 7.5) illustrates that the wells intersect the footwall scarp where the subcrop is the Dunlin Group, i.e. Brent Group geologies interpreted in these wells cannot be *in situ*. Both wells sample the fault scarp degradation complex and core samples are from within the complex. Summary graphic logs of the cored intervals in these wells, and the electrical log signatures from the fault scarp degradation complex, are illustrated in Fig. 7.6. These observations are compared with the log signatures and core observations from *in situ* Brent Group in well 33/9-3, situated within 2 km of both wells. Core photographs of sediments assigned to the Brent Group in wells 33/9-C6 and 33/9-C11 are presented in Fig. 7.7.

Well 33/9-C6 intersects 218.1 m measured depth (m MD), 128 m true vertical depth (m TVD), of sand-rich geologies in the fault scarp degradation complex; sediments assigned to the Brent Group by interpretation of core and log signatures (Fig. 7.6). The upper 183.9 m MD of the fault scarp degradation complex is cored (with 99.6% recovery), from which the Humber (Viking) Group (9.2 m loggers thickness), Tarbert Formation (35.9 m), Ness Formation (55.8 m), Etive Formation (41.5 m) and Rannoch Formation (>50.7 m) are interpreted. From the completion log alone, the uncored lower part is interpreted as Rannoch Formation (total thickness 85.6 m MD) and Broom Formation (1.3 m). Sediments assigned to the Brent Group are underlain in 33/9-C6 by 245.7 m MD (151.7 m TVD) of Dunlin Group and >144 m MD (>93.1 m TVD) thickness of Statfjord Formation.

Although all the formations of the Brent Group are recognised in stratigraphic order in well 33/9-C6, a comparison of the true vertical thicknesses with those sampled by proximal west flank well 33/9-3 shows that the Brent Group is thinned, with particular thinning affecting the Tarbert, Ness and Rannoch (5A) formations (Fig. 7.8). The typically funnel-shaped gamma ray profile of the Rannoch Formation has a condensed centre in 33/9-C6, but no discrete slip surfaces can be defined. Thinning is possibly associated with an unusually high resistivity 'tight' zone (3960-3985 m MD), supported by the observation of intense deformation in the core sample from this interval. The heterolithic nature of the Ness Formation means it is impossible to locate the missing stratigraphy and no distinct slip planes are observed in the core or from interpretation of the log signatures. Core observations indicate that the Tarbert Formation is stratigraphically intact, hence, it is suggested that the top of the formation may have been eroded. In addition, the Dunlin Group is thinned in well 33/9-C6 but the Statfjord Formation is of comparable thickness to the formation in 33/9-3. The thicknesses of the Dunlin Group and Statfjord Formation are consistent with the structural observations from seismic interpretation (Fig. 7.5a). From interpretation of the electrical log signatures, it is suggested that the majority of Dunlin Group thinning has been accommodated in the Drake and Cook formations, thinned (in comparison to 33/9-3) from 41.2 m TVD to 18.4 m TVD and 74.7 m TVD to 39.7 m TVD, respectively, but no distinct slip surfaces can be defined.

Well 33/9-C11 is located ~1 km to the south of C6. This well intersects 123.5 m MD (93.7 m TVD) of sediments in the fault scarp degradation complex that have been assigned to the Brent Group, of this 100 m MD is cored. From interpretation of the cored interval and the electrical logs (Fig. 7.6), the well path samples Cromer Knoll Group (~10 m MD thick), Humber Group (0.8 m MD), Tarbert Formation (42 m loggers depth), Ness Formation (34 m), Etive Formation (13.8 m), Rannoch Formation (30.9 m MD), Broom Formation (1.9 m TVD), Dunlin Group (225.2 m TVD) and Statfjord Formation (158.7 m TVD).

The thicknesses of the Dunlin Group and the Statfjord Formation in well 33/9-C11 are comparable with those in the undegraded west flank (Fig. 7.8), as would be inferred from the seismic interpretation (Fig. 7.5b). Within the Brent Group, however, the Etive, Ness and Rannoch formations are thinned significantly. This thinning suggests that major slip surfaces dissect the formations. The lack of deformation observed in the completely cored Ness and Etive formations indicates that the faults are discrete single slip planes. In contrast, the cored Rannoch Formation is pervasively deformed, and thinning of the formation is related to intra-formational faulting.

The most striking feature of the cored sections in both of these wells (and the other wells in this category) is the lack of deformation and, hence, ease of identification of Brent Group sediments (see core photographs, Fig. 7.7). Indeed, geologies of the Brent and Dunlin groups are recognised to the exclusion of all others within zone B of the fault scarp degradation complex. The excellent preservation of the primary sedimentology and depositional structure of the Brent Group observed in the core samples is mirrored by the form of the electrical well log signatures. Diagnostic log patterns associated with facies units in the footwall are recorded from the sediments in the fault scarp degradation complex. For example, the classic funnel-shaped gamma ray profile of the Rannoch Formation observed in west flank well 33/9-3 is also the profile of the Rannoch Formation interpreted in fault scarp degradation complex well 33/9-C6 (Fig. 7.6).

As the near identical log signatures would suggest, sediments within the fault scarp degradation complex have comparable porosity and permeability values to their sedimentological counterparts in the west flank. Fig. 7.9 illustrates the porosity and permeability measurements (from CCA data) for Tarbert, Ness, Etive and Rannoch formations in wells 33/9-3, 33/9-C6 and 33/9-C11. All three plots are drawn to the same vertical scale. Low displacement faults observed in sediments from the complex do affect local porosity and permeability (e.g. the Rannoch Formation in 33/9-C11); these structures are described in detail in Section 7.5.

The main features of the core samples from the fault scarp degradation complex in the 10 wells assigned to this category are, (1) a lack of pervasive brittle or ductile deformation, (2) excellent preservation of primary sedimentary structures and petrophysical characteristics, (3) stratigraphic integrity and way-up on >50 m scale, and (4) thinning of formational units, associated with large slip planes or intra-formational shearing. From these observations it is concluded that the sediments sampled from the fault scarp degradation complex in these wells form part of large translational-rotational slides of competent Brent Group geologies. The slides are bounded by major discrete slip planes but have little internal deformation. The main basal dècollement is within the Dunlin Group, rather than within the Brent. This conclusion supports the morphological observations from seismic interpretation (Section 7.3).

7.4.3 Type 2: Core samples from the fault scarp degradation complex incorporating pervasively deformed sediments

This second category includes core samples in which the sediments were observed to be pervasively deformed (by both ductile and brittle deformation). Such deformation was recorded in cored intervals from 3 wells – 33/12-B16, 33/12-B21 and 211/29-C15 (Fig. 7.1). In all these wells, characteristic features of Brent Group geologies were recognised in the sediments suggesting that the deformation maybe, in part, due to reworking. This is demonstrated in the summary graphic log of the core sample from the fault scarp degradation complex for well 33/12-B16 shown in Fig. 7.10 (no wireline log signatures were available for this well). The thicknesses of lithostratigraphic units in the 3 wells are compared with those in well 33/9-3 in Fig. 7.8. Core photographs illustrating the characteristic features of the deformation of sediments included in this category are presented in Fig. 7.11. Detailed sedimentary logs of the cored intervals can be found in Appendix 4.

The locations of the cored intervals from the 3 wells in this category are described, on seismic cross sections oriented perpendicular to the strike of the Strathspey-Brent-Statfjord fault, in Fig. 7.12. In all 3 wells the cored intervals are sampled from a shallow structural position on the scarp directly underlying the onlapping syn- and post-rift sediments. However, the wells penetrate the footwall scarp in different structural positions. Well 211/29-C15 is located nearest to the crest of the footwall; where the fault scarp degradation complex overlies the subcrop of the Dunlin Group. This well intersects a total thickness of 219 m TVD of the complex, interpretation of the core sample assigns the top 91 m TVD to this category and the lower 128 m TVD to type 1. The complex overlies a thinned Dunlin Group (with respect to the west flank) and complete Statfjord Formation. In contrast, wells and 33/12-B16 and 33/12-B21 penetrate the fault scarp degradation complex above the subcrop of the Statfjord Formation. The complete interval of fault scarp degradation complex sampled in these wells is of type 2 and the complex is thin – 53.5 m MD in 33/12-B16 and ~40 m TVD in 33/12-B21. Both wells overlie thinned successions of the Dunlin Group and Statfjord Formation (in comparison to well 33/9-3, Fig. 7.8).

Deviated production well 33/12-B16 was plugged and abandoned when the Statfjord Formation target was found to be thinner than expected. From interpretation of core samples and electrical log signatures, the well is shown to penetrate 53.5 m MD of sand-rich, displaced sediments which overlie of 191 m MD thickness of Dunlin Group mudstones and the thinned Statfjord Formation (Fig. 7.10). There are two cored intervals in the well – from the fault scarp degradation complex and the Dunlin Group – and both samples are observed to have a pervasive ductile deformation (Fig. 7.11). This observation and the conspicuous absence of the Cook Formation (from interpretation of log signatures) suggests that sediments assigned to the Dunlin Group may not be *in situ*. This interpretation is consistent with the observed thinning of the Statfjord Formation below.

The sediments sampled in the 15.5 m of core (100% recovery) from the fault scarp degradation complex in well 33/12-B16 are sedimentologically chaotic and exhibit extreme soft sediment deformation (Fig. 7.11a). Within a highly distorted muddy matrix (composing <10% of the sample), clasts of <10 cm to in excess of 1.5 m in size are interpreted. The margins of individual clasts are sharp or deformed and incorporated into the matrix (Fig. 7.11b, d). The sediments composing the clasts are typically sand-rich and partially maintain sedimentological integrity; primary depositional structures, however, commonly displaced by intense deformation, both brittle and ductile (Fig. 7.11c, d). From the primary structures, geologies of affinity to the Ness, Etive, Rannoch and Broom formations can be identified, suggesting that these are reworked, displaced sediments of the Brent Group.

This cored section is interpreted as the deposits of a debris flow. A sketch interpretation of the structure of the fault scarp degradation complex in the vicinity of well 33/12-B16 is shown in Fig. 7.13, drawn to scale from interpretation of core, electrical log and seismic data. This interpretation illustrates the main body of the complex to be constructed of large slide blocks of competent Brent Group (in zone B) and Statfjord Formation (zone S) geologies, as described above. The slide blocks are overlain by a thin veneer of disaggregated sediments locally derived from surface reworking by debris and granular flows, transporting sediments down the scarp. Hydroplaning of the flow (*cf.* Mohrig et al. 1998) could have inhibited the complete disintegration of clasts within flows. Debris flows transported sediments into the hangingwall (see Chapter 4, Fig. 4.20) and the debris also ponded in lows on the footwall scarp. The sketch interpretation of the fault scarp degradation complex is compared in Fig. 7.13 with the structure of a translational slide from the south coast of the UK, although the geologies constructing the slopes are significantly different and the sliding occurred in different environments (subaerial and submarine). The Bindon landslip is being rapidly eroded due to undercutting by waves and small scale fluvial activity on the slide; these processes are obviously not active in submarine settings.

The sediments described from core samples from wells 33/12-B21 and 211/29-C15 have comparable features to those in 33/12-B16 (see logs in Appendix 4) and are also interpreted as the deposits of debris flows and slumps on the footwall scarp. It is emphasised that these geologies are thin and locally derived from shallow reworking processes on the scarp.

7.4.4 Processes of degradation of the footwall of the Strathspey-Brent-Statfjord fault

Interpretation of core and electrical log signatures identifies two components to the fault scarp degradation complex mounted on the footwall of the Strathspey-Brent-Statfjord fault system: The remnants of translational-rotational slides overlain by a veneer of debris flow deposits. The source of sediment for both components is reworking of pre-rift geologies in the underlying footwall. Geologies incorporated into the slide deposits are largely undeformed and form a multi-tiered complex of 'cup-shaped' units up to 200 m thick.

Principal décollement in the Rannoch Formation and mid-Dunlin Gorup are described. In contrast, the debris flow deposits are chaotic in structure and generally thin (<25 m), although ponding in local lows on the scarp is recognised. Comparable debris flow deposits are recognised interbedded with hemipelagic mudstones in the syn-rift in the proximal hangingwall (e.g. well 33/9-18).

The implication of these observations, and the interpretation of syn-rift sediments (Chapter 4), is that degradation of the footwall of the Strathspey-Brent-Statfjord fault array was largely by sliding processes with superficial reworking of sediments by gravity driven sediment flows. This interpretation supports the results of previous seismic based studies. On this basis, Hesthammer and Fossen (1999) and McLeod and Underhill (1999) have constructed dynamic models for the denudation of the Statfjord and Brent structures, respectively. They demonstrate that deep-seated sliding will only occur when the décollement horizons are exposed in the scarp face and, hence, propose a multi-phase erosion history controlled by the increasing displacement on the bounding fault systems.

7.5 Brittle deformation in the fault scarp degradation complex

From the interpretation of core samples and electrical logs of wells penetrating the fault scarp degradation complex faults with displacements of decametres are interpreted to bound the margins of discrete, large (10-100's m) slides. Offsets across these faults results in complex, stratigraphically inconsistent juxtapositions of formations and apparent thinning or thickening. In addition to the large displacement structures, faults with mm-cm displacement are also observed, deforming displaced sediments within the slide complex (e.g. Fig. 7.7). Comparable structures are recorded within clasts in the veneer of debris flow deposits overlying the slides on the scarp (Fig. 7.11). This section documents the characteristics of these low displacement faults and examines the mechanics of their origin. The results of these investigations – fault counts, CCA data and porosity calculations – are recorded in Appendix 5.

7.5.1 Characteristics and distribution of small displacement faults

In this study, 'brittle' is used to describe discrete planes across which shear displacement has occurred with little associated contortion of the surrounding sediment. Using this criterion, the structure of the core samples in all the wells was logged. Three types of fault with low displacement are recognised (Fig. 7.14),

1. Faults of type 1 are restricted to medium – coarse grained (reservoir) sandstones and are typically pale in colour without hydrocarbon staining (Fig. 7.14a, b). The fault planes are 2-4 mm in width and have offsets of up to 1 cm. The faults typically occur at a high angle with respect to bedding. The faults are generally observed in low concentration as isolated single planes, but branching, multi-stranded fault systems and concentrations of faults are also recorded;

2. Type 2 structures occur in sediments dominated by very fine – fine sand grained sandstones, typically interbedded (on a mm scale) with mud-rich laminae. In contrast to type 1 faults, these faults are dark in colour and the fault plane is thin, of 1-2 mm. Displacements can exceed 3 cm, but are typically <1 cm. The faults appear to have no preferred orientation with respect to bedding, although faults sub-parallel to bedding are probably numerically dominant. Type 2 faults can occur in isolation (Fig. 7.14c) or high concentration where they have anastomosing, cross-cutting geometries (Fig. 7.14d, e);

3. Interbedded sand and mud horizons (on a mm-cm scale) are offset by faults of type 3. These faults have sharp, thin planes commonly with a mud smear and offsets in the mm-cm range (Fig. 7.14f). The faults are rare, occur in isolation and at a high angle to bedding.

Faults of all three types are observed throughout the core samples. In the highly deformed sediments interpreted as debris flow deposits the faults occur within clasts, rather than throughout the deposit. The orientation of the faults could not be determined, as the wells are deviated and core samples have not been reoriented. The development of the faults is considered to be syn-degradation. Low displacement faults (of all types) are recorded in equivalent pre-rift geologies in the west flank (e.g. in well 33/9-3), but are less pervasive and of lower density. The micromechanics of faulting is the same in similar lithologies in the west flank and the degradation complex (Section 7.5.4), but the processes of degradation enhanced the growth of normal faults.

The distribution of the small displacement faults was quantified by counting the number of faults per metre of core in 4 wells – 33/12-B21, 33/12-B23A, 33/9-C6 and 33/9-C11 (Fig. 7.1). Plots of the fault density in wells 33/9-C6 and 33/9-C11 are presented in Fig. 7.6. These plots illustrate that, although faults are recognised throughout the fault scarp degradation complex, they are generally of low concentration (<3 ft/m) and there is an overall lack of brittle deformation. High densities of faults (of up to 150 ft/m) do occur locally, in zones of several metres thickness, dissecting the sediments incorporated into slides; such zone are observed particularly in geologies assigned to the Tarbert and Rannoch formations.

The type of fault (as defined above) was also recorded during fault counts and the results support the qualitative observation of the strong lithological control on fault distribution. Type 1 faults are restricted to medium – coarse grained, porose sands and, in particular, the well sorted but massive sandstones of the Tarbert Formation. Type 1 faults are also observed to deform comparable sands of the Etive and Ness formations. These faults generally occur in isolation and are rarely multi-stranded, hence the maximum number of faults recorded in a metre is <20. In the Tarbert Formation, however, the sandstones are poorly cemented and prone to disintegrate in core sample, so fault densities are most likely underestimated.

In contrast, faults of type 2 occur in very fine – fine grained silty sandstones. Rare isolated fault planes are recorded from the Ness and Etive formations, but the densest populations of type 2 faults occur in the Rannoch Formation. A maximum of 131 ft/m was recorded in geolgies assigned to the Rannoch Formation in well 33/9-C6. The Rannoch Formation is a shoreface deposit, coarsening up (over ~70 m) from silt to fine – medium grained sand. Faulting is largely restricted to the very fine grained sandstones of the lower Rannoch Formation (5B), where the sediment is micaeous and bedding is on the sub-cm scale. Planar bedding and low angle truncations are well developed throughout the formation, defined by fining up units of mm-cm scale and capped by micaceous, mud-rich laminae. Offset of the regular bedding structures demonstrates that displacement on the faults is low, typically <1 cm. However, although the displacement on a single fault is small, the cumulative displacement of many faults is on the metre scale. High densities of faults occur in shear zones of 10-150 cm thickness, separated by areas up to 70 cm thick of very few faults. Fault strands in the high density populations are anastomosing and cross-cutting, in particular faults are observed to be sub-parallel to bedding and link via short high angle faults. The timing of movement on the individual strands with respect to each other is unclear as fault displacements are small and the population dense; but bedding parallel structures are generally older, and offset by, the linking faults oriented at a high angle to bedding.

7.5.2 The impact of fault types 1 and 2 on bulk rock properties

Core plugs in samples from the fault scarp degradation complex in wells 33/9-A17, 33/12-B21, 33/12-B23A, 33/9-C6 and 33/9-C11 (Fig. 7.1) were examined and assigned a k-factor. This measure (described in Fig. 7.15) defines the bulk rock lithology sampled by the plug and the geometry, number and type of the small displacement faults intersecting the hole. For example, a k-factor of '*mod 1 ss*' would refer to a core plug of sandstone (<5% silt grade) intersected by 4-6 faults of type 1 (as defined above), the faults not oriented parallel to bedding. These observations are employed to assess the impact of the deformation on bulk rock porosity and permeability, comparing CCA data from plugs intersected by faults with data from plugs sampling undeformed host rock in the fault scarp degradation complex and the footwall (well 33/9-3).

Fig. 7.16 presents a synthesis of the porosity and permeability data, k-factor measurements and core observations from well 33/9-C11. The primary control on porosity and permeability is observed to be lithology (Fig. 7.16a), hence the individual formations of the Brent Group can be differentiated. Lithological variations, in particular associated with bedding structures, are also the principal control on vertical, compared with horizontal, permeability. Consequently, all the permeability data presented in this section is horizontal (near bedding parallel) permeability (KAH). While lithology and sedimentary structure are the primary controls on bulk rock properties, in core plug samples intersected by faults the faults are observed to have a secondary influence on porosity and permeability. In well 33/9-C11, a high density of type 2 faults – a shear zone – is recorded in the lower Rannoch Formation. Permeability measurements decrease significantly in core plug samples from the sheared sediments compared with those in the undeformed rock (Fig. 7.16b, see also Fig. 7.9).

Measures of porosity, permeability and the k-factor of individual core plugs are presented in Figs. 7.16c and 7.16d. The ranges of porosity and permeability for undeformed medium – coarse grained sandstones of the Etive Formation are shown to be 25-33 % and 107-6200 mD, respectively (Fig. 7.16c). Two higher measures of porosity and permeability are recorded from core plugs with a k-factor of *mod 1 ss*; intersected by 4-6 faults of type 1. Similarly high porosity and permeability are measured in plugs sampling type 1 faults from the medium sandstones of the upper Rannoch Formation (Fig. 7.16d). Typical measures for the planar bedded, silty sands of the Rannoch Formation are a porosity of 22-30 % and a permeability of 362-5100 mD (decreasing with increasing silt content). Measurements in plugs sampling type 2 faults show a substantial decrease in permeability (<1-704 mD); however porosity measurements are comparable to the undeformed rock (19-29 %). The orientation of the faults with respect to bedding is significant; type 2 faults oriented bedding parallel typically have a more limited effect on bulk rock properties, but faults oriented at an angle to bedding have a significant impact.

In conclusion, it is observed that faults of type 1, when occurring in sufficient density, *enhance* the porosity and permeability of the bulk rock. In contrast, type 2 structures *inhibit* porosity and permeability, particularly when the faults are oriented at an angle to bedding, reducing measures of these properties by up to an order of magnitude.

7.5.3 Microstructure of type 1 and type 2 faults

A total of 37 core plugs were sampled in 6 wells (Fig. 7.3) and thin sections cut, oriented perpendicular to bedding (see Section 7.2). Detailed microscopic studies examined the mineralogy, grain size and shape and structure of the type 1 and type 2 faults sampled in these thin sections, complemented by simple image analysis to quantify porosity. The technique employed to calculate porosity is described in Fig. 7.17. No facilities were available to determine the micropermeability of the faults.

Type 1 faults

The microstructure of a type 1 fault from the Tarbert Formation in well 33/9-C6 is illustrated in Fig. 7.18. The host rock for this fault is a well sorted, medium grained sandstone, with a poorly developed cm scale planar bedding defined by fining up towards crude micaceous, silty laminae. The sandstone is a quartz arenite of typically sub-angular –

sub-rounded equant grains, rounded accessory ('heavy') minerals compose ~5 % of the sand. Cementation is poor, consequently porosity is high. The location of the fault is illustrated in Fig. 7.18 by the offset of the bedding plane. The mineralogy of the fault zone is comparable to that of the bulk rock, bar the loss of the silt and clay fraction. It is suggested that fluid flow along the fault zone, either syn-displacement or subsequently, has removed the fine grains; the implication is that this fault was exploited as a dewatering structure. The lack of the fine grained sediments, which bind hydrocarbon molecules, probably accounts for the pale colour of the faults in core sample. Sediments incorporated into the fault zone are dilated (of increased porosity), with respect to the surrounding undeformed sandstone.

Type 2 faults

The structure of type 2 faults is in stark contrast to that of type 1 faults. Fig. 7.19 presents a type 2 fault from the Rannoch Formation in well 33/9-C6. The host rock is a well sorted, very fine – fine grained quartz arenite of sub-rounded grains, with ~10 % feldspar and <2 % mica lathes. Mica content increases, as does the proportion of accessory phases, in the diffuse bedding planes, capping fining up sub-parallel beds of <1 cm thickness. Cementation is ill-developed. The type 2 fault dissecting the microphotograph shown in Fig. 7.19 has multiple anastomosing strands, splitting into two in the upper right. The margins of the faults are sharp.

Within the fault zone, quartz and feldspar grains are sub-rounded – sub-angular and of very fine – fine sand size; comparable to the grains in the host rock. Point count transacts, measuring grain size and grain angularity, demonstrate no change in these characteristics in the fault zone compared with the rock. The grains remain intact with only local grain cracking at grain-grain contacts. The bright birefringence colours in the fault zone (in contrast to the host rock, Fig. 7.19b) reflect the abundance of <10 μm size mica and clay (particularly illite) clasts. In the fault zone, these form a matrix to, and coat the margins of, the quartz and feldspar grains (Fig. 7.20d). The intact mica lathes of the host rock are absent in the fault zone and partially degraded clasts at the margins of faults testify to the break down of the grains into smaller component parts.

The abundance of silt and clay sized broken mica lathes, and their distribution within and between the quartz and feldspar grains, would have had a significant impact upon the diagenetic environment within the fault zone. The area of free surface, important for the nucleation of new minerals, is substantially increased (with respect to the undeformed rock). In addition, the low pH environment associated with mica dissolution would be pervasive in the fault zone, rather than localised around a single grain in the host rock. Low pH conditions would enhance the dissolution of quartz and feldspar grains, further enhanced by increased grain-grain contacts and the veneer of mica lathes. This is demonstrated by the ragged appearance of grains, particularly feldspar, incorporated into the fault zone compared with their equivalents in the host rock (Fig. 7.20d). The low permeability of the fault zone would restrict the movement of mineral rich fluids and encourage local precipitation of new mineral phases. The growth of siderite and pyrite during early diagenesis was enhanced by the plenitude of mica, and these mineral phases are more abundant in the fault zone than host rock. The growth of siderite crystals within partially deformed mica grains (Fig. 7.20c) demonstrates that fault zone formation preceded the growth of siderite, i.e. that faulting occurred at shallow depths of burial.

7.5.4 The origin of type 1 and type 2 faults

The brittle failure of porous sandstones in response to extensional stresses is described by a 3-fold hierachy of structures: (1) a single deformation band, (2) a zone of deformation bands and (3) a slip surface associated with a zone of deformation bands (Aydin 1977, Aydin and Johnson 1978). These structures are interpreted as a process continuum, accommodating greater strain (Fig. 7.21a). A deformation band is defined as 'a quasi-tabular band of crushed grains about 1 mm in width' with shear offset in the mm-cm range (Antonellini and Aydin 1995). Similar structures without offset are termed compaction bands (Mollema and Antonellini 1996). Petrophysically, deformation bands are characterised by a porosity of one order of magnitude less than the host rock and permeability of up to three orders of magnitude less than the host rock (Pittman 1981, Antonellini and Aydin 1994); these factors are controlled by the amount of cataclasis, clay content and diagenesis. In the field, deformation bands form anastomosing segments linked via ramp or eye structures (Fig. 7.21b), an individual band can have great lateral continuity.

Antonellini *et al.* (1994) expanded the definition of deformation bands through the definition of three discrete types: (1) bands with little or no cataclasis, (2) bands with grain cataclasis and (3) bands with clay smearing. Deformation bands with little or no grain cataclasis are further subdivided into dilational, compactional and those with little volume change, with respect to the host rock. The type 1 and type 2 faults described by this study occur in high porosity sand-rich sediments, have shear offset but no grain cataclasis, hence, can be assigned to Antonellini *et al.*'s type 1 deformation bands – bands with little or no grain cataclasis. Type 1 faults are dilational with respect to the host rock, type 2 faults are compactional. Antonellini *et al.* propose that deformation bands with and without cataclasis are genetically linked and form a continuum. Faults with no grain cataclasis form initially and accommodate small amounts of strain, but as individual deformation bands grows and accumulate displacement, so grain fracturing and cataclasis occurs.

The micromechanics of high porosity sandstone deformation in non-hydrostatic stress regimes have been elucidated by triaxial laboratory experiments. Such experiments have identified, and in some instances constrained, the rheological parameters influencing rock failure. The results of this experimental work contribute significantly to this study, particularly to understanding the processes of deformation within slide blocks. Hence, a brief resume of the key conclusions of laboratory studies is presented below. For more detail, including on the experimental procedure and the properties of the sandstones tested, the reader is referred to Zhang *et al.* (1990), Lockner *et al.* (1992), Menéndez *et al.* (1996), Wong *et al.* (1997) and Zhu and Wong (1997).

Two modes of failure have been recognised in high porosity sandstones (Fig. 7.21c),

1. Shear localisation. Low stresses result in elastic strain until the critical stress state (C') is achieved, marking the onset of shear induced dilation. The local shear stresses are sufficient to rupture the grain-grain contacts, permitting the grains to move relative to each other. Grain movements result in relative dilation within the fault zone, but a decline in permeability as the tortuosity increases. As stress increases and the peak stress is attained, so Hertzian fractures begin to develop. Coalescing of the intra-grain fractures results in shear localisation. This is reflected by an enhancement of grain contact tensile stress and pervasive comminution within the shear band;

2. Cataclastic flow. In this deformation regime, the mean effective stress is too high to permit the movement of grains relative to each other, as described above. The high stresses at grain contacts result in the initiation of Hertzian fractures which coalesce leading to grain crushing and pore collapse. This process is described by a decline in

porosity and permeability, as the comminuted particles compact, and strain hardening.

The primary control on the mode of failure is certainly the stress regime (overburden pressure). The experimental work, however, has also recognised a number of other factors influencing deformation. These include, the initial grain radius, porosity, cementation (cements 'cushion' grain boundaries inhiting cracking), primary sedimentary structures and mineralogical composition.

The results of experimental work indicate that both type 1 and type 2 faults were formed by the process of shear localisation. Significantly, the process appears to have 'stalled' in the early stages – indicative of extremely low effective mean stresses, certainly insufficient to cause grain contact cracking. The implication of this conclusion is that the faults were formed at shallow depths within the earths crust, suggesting an early genesis and supporting the association with scarp degradation processes.

7.6 Conclusions

1. Interpretation of seismic data demonstrates that a complex of displaced sediments is preserved mounted on the degraded footwall scarp of the Strathspey-Brent-Statfjord fault system. The fault scarp degradation complex is thin (<150 ms TWT), and its internal structure is below the resolution of the seismic data. Mapping of the boundaries of the complex illustrates that it forms two, strike parallel zones of maximum thickness on the scarp: Zone B and zone S. Within these zones, the complex demonstrates cuspate thickness variations on a sub-km scale.

2. Logging of ~1350 m of core samples from zone B of the fault scarp degradation complex separates the complex into 2 categories: (1) sediments of affinity to the Brent Group, and (2) pervasively deformed sediments.

3. In core samples from 10 wells from the complex, geologies of the Brent Group are interpreted which are not *in situ*. These sediments have little post-depositional deformation, retain their sedimentological integrity, petrophysical characteristics and stratigraphic continuity. They are interpreted as the remnants of translational and rotational slides of competent Brent Group sediments.

4. In contrast, in 3 core samples from the fault scarp degradation complex, pervasive deformation (both ductile and brittle) of the sediments is observed. These samples are all from a shallow structural position in the complex, commonly overlying the undeformed Brent Group geologies incorporated into slides. Within a mudstone matrix, clasts of deformed sediments of affinity to Brent Group geologies are interpreted. These sediments are interpreted as the deposits of debris flows, locally derived from reworking of the shallow sediments on the scarp.

5. Small displacement faults are recognised to deform sediments throughout the fault scarp degradation complex, within the slides and in clasts in the debris flows. 3 types of fault are recognised, different in character and each restricted to a specific lithology. Of the 3 types, types 1 and 2 are observed to occur in high density populations (at up to >150 faults/m), particularly within sediments assigned to the Tarbert (type 1 faults) and Rannoch (type 2 faults) formations.

6. Faults of type 1 are interpreted as dilatational deformation bands that enhance the porosity and permeability of the host rock. The origin of these structures is ascribed to dewatering, with fluid movement associated with small amounts of shear displacement. Dewatering of the pre-rift sediments probably occurred during catastrophic slope failure.

7. Faults of type 2 are interpreted as compactional deformation bands, with associated reductions in rock porosity and permeability. No evidence of quartz or feldspar grain cataclasis was observed, but mica lathes are fragmented and the flakes mantle the unfractured grains. A unique diagenetic environment is interpreted within the fault zones as a consequence of the mica disaggregation. These structures commonly describe shear zones of several metres thickness, and define a major décollement in the Rannoch Formation.

CHAPTER 8 A reconnaissance study of the degraded footwalls of latest Jurassic – Early Cretaceous faults, central east Greenland

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8.1 Introduction

Syn-rift footwall denudation has been invoked as a potentially significant source of clastic sediment in marine rift basins without an externally derived sediment supply (e.g. Gawthorpe *et al.* 1994, Ravnås and Steel 1998). This thesis has been examined by this study through an assessment of the magnitude, processes and temporal evolution of the degradation of the footwall of the Strathspey-Brent-Statfjord fault system in the northern North Sea (presented in Chapters 6-7). In this area, the volume of sediment liberated by footwall erosion is demonstrated to be relatively small in comparison with both the volume of accommodation generated and the volume of the uplifted area, and also considering the period of time over which denudation occurred (~60 Ma). The products of footwall degradation are preserved in the proximal hangingwall to the Strathspey-Brent-Statfjord fault, interbedded with hemipelagic and turbidite deposits, but comprise <5 % of the syn-rift sections examined (Chapter 4). The implication of these observations is that footwall denudation is relatively insignificant as a source of syn-rift sediments in the Strathspey-Brent-Statfjord area. Indeed, the similar morphologies of other denuded footwalls in the northern North Sea basin suggests that a low magnitude of footwall erosion is characteristic of the province (e.g. Yielding *et al.* 1992; Alwyn North, Johnson and Eyssautier 1987; Gullfaks, Fossen and Hesthammer 1998; Ninian, Underhill *et. al.* 1997). The rôle of footwall degradation as a source of syn-rift clastic sediments is examined further in this chapter; the observations from the footwall of the Strathspey-Brent-Statfjord fault are compared with observations from a number of large displacement faults in the central east Greenland rift.

The results of field studies of four, latest Jurassic – Early Cretaceous faults in central east Greenland are presented in this chapter. For each fault system, the morphology of the footwall, evidence for the process(es) of denudation and the character of sediments deposited in the proximal hangingwall are examined. Two-dimensional reconstructions of the denuded footwalls of these faults illustrate that the magnitude of denudation is low and, significantly, is comparable in scale to the magnitude of denudation of the northern North Sea example. Additionally, although syn-rift deposits remain largely buried in east Greenland, degradation products are observed interbedded with hemipelagic post-rift sediments but comprise only a small fraction of these deposits. The implications of the results of these studies in central east Greenland and the northern North Sea for qualifying and quantifying sediment supply to marine rift basins are discussed in conclusion to this chapter. The ramifications of this work for hydrocarbon exploration and exploitation are considered in Chapter 9.

8.2 Geological Setting

The east Greenland continental margin has experienced a series of extensional events during its complex post-Caledonide evolution (see review of Surlyk 1990a). The most recent rift phases were during the latest Jurassic – Early Cretaceous and, subsequently, during the Early Tertiary; the latter event culminating in sea floor spreading and the opening of the North Atlantic in the Late Palaeocene (Talwani and Eldholm 1977, Bott 1987) (Fig. 8.1a). During the Tertiary, large normal fault systems were developed, some reactivating ancient structures (Price *et al.* 1997), and concurrent extrusive and intrusive magmatics were voluminous (e.g. Larsen *et al.* 1989). A consequence of the extension was to uplift the continental margin and the ensuing deep denudation of >2 km (Hansen 1992, Clift *et al.* 1998, Thomson *et al.* 1999) has exhumed the rift geologies of the earlier mid-Mesozoic event.

W-E oriented extensional stresses affected the central east Greenland region during the Middle Volgian – latest Valanginian (Vischer 1943, Maync 1947, 1949, Surlyk 1977, 1978), as evidenced by syn-rift deposits in the Wollaston Foreland area. The continuation of active extension into the mid-Albian has been proposed by Whitham *et al.* (1999) based on the interpretation of outcrops in the Hold With Hope area (discussed further in Section 8.3.2). The limitations imposed by 2-dimensional field exposures, buried syn-rift strata and a Tertiary structural overprint mean that the evolution of the late Jurassic – Early Cretaceous fault population in central east Greenland rift is poorly constrained, both in terms of the distribution of active faults and the timing of activity. However, it is clear that the majority of the strain was accommodated on N-S trending, 'domino-style' fault systems composed of large displacement (in excess of 1 km), easterly-dipping, sub-parallel normal fault arrays located at 5-30 km spacing. These faults are currently exposed between Milne Land and Store Koldeway, in a thin coastal strip east of the Post-Devonian Main Fault (PDMF, as defined by Vischer 1943). The PDMF defines the eastern boundary of a hinterland of Caledonian metasediments (Fig. 8.1b). The magnitude of crustal stretching during the latest Jurassic – Early Cretaceous rift event is estimated to have been ~20 % (Price *et al.* 1997).

During the Middle Jurassic, prior to the onset of rifting, the central east Greenland area was located in a tidally influenced, shallow marine setting with a high supply of coarse clastic sediments derived from the west (sands of the Vardekløft Formation, Fig. 8.2). This environment was gradually flooded from the north during a major, regionally significant transgression, in part driven by a rise in eustatic sea level (Surlyk 1990b); hemipelagic sedimentation became the dominant depositional process (mudstones of the Bernbjerg Formation, Fig. 8.2) (Surlyk *et al.* 1973, Callomon and Birkelund 1980, Surlyk 1991). By the latest Oxfordian, fully, and deepening, marine conditions were established from Wollaston Foreland in the north (Surlyk and Clemmensen 1983) to Milne Land in the south (Surlyk 1987). Hence, as the rift system was initiated, the palaeo-environment in the study area was that of a distal shelf sea.

The W-E oriented cross section, perpendicular to the strike of the Dombjerg fault on Wollaston Foreland, reproduced in Fig. 8.2b from the work of Surlyk (1978, his Fig. 1), demonstrates the nature of syn-rift sedimentation in the east Greenland area. In the west of the section, proximal to the scarp, are a 2-3 km thickness of coarse clastic sediments, interpreted as the deposits of footwall sourced submarine rockfall and debris flow processes. Laterally, these sediments become dominated by gravity flow deposits, fining eastwards onto the dip slope. Syn-rift sediments are <100 m thick on the distal dip slope. Surlyk interprets the depositional environment of these sediments as closely spaced, interfingering submarine fan systems sourced from the growing and degrading Dombjerg fault (his Fig. 41). Surlyk also records a subtle evolution in the character of the syn-rift sedimentation. Proximal facies of Volgian-Ryazanian age (Rigi Member) are extremely coarse (clasts >20 m), however, Valanginian age equivalents of the Young Sund Member are finer grained (clasts <1 m), and the facies belts are narrower. Surlyk considers this change to be a consequence of the drowning of 'crestal islands' during the Valanginian, resulting in a decline in the rate of erosion of the footwall and the evolution into an open marine basin.

Elsewhere in the basin, syn-tectonic deposits remain largely buried below a thick post-rift succession. Post-rift sediments are assigned to the 'mid-Cretaceous sandy-shale sequence' (Barremian-Campanian) and interpreted as the deposits of a sediment starved distal shelf sea (e.g. Whitham *et al.* 1999). These mud-prone sediments onlap fault bounded footwall highs. Where the onlap surface is exposed, the footwall scarps are observed to be significantly denuded (e.g. Maync 1949, Whitham and Price 1997), as predicted by Surlyk's depositional model. In addition, reworked pre-rift sediments are preserved in both the proximal post-rift sediments and mounted on the footwall scarp (Whitham and Price 1997, Kelly *et al.* 1998). For example in the north of Clavering Ø, Maync (1949) reports large clasts of basement derived geologies embedded as part of debris flow deposits within Middle Albian mudstones.

8.3 Field Observations

A preliminary 8-week field season to central east Greenland was undertaken during July-August 1997. The aim of this season was to do reconnaissance mapping (of between 3-10 days in each area) of a number of well exposed degraded footwalls of normal faults and proximal hangingwall sediments, identified from descriptions in the literature (e.g. Maync 1949, Surlyk 1978) and the work of Cambridge Arctic Shelf Program (CASP) geologists. The location of the faults studied is illustrated in Fig. 8.1b, and the observations made during that time are reported in this section. The results were used to plan a second field season for the summer of 1998.

In the months following the 1997 field season, work on the subsurface dataset from the northern North Sea (presented in Chapters 3-7) developed and, consequently, the thesis of this study became more focussed. In light of this, a decision had to be made on whether to expand on the ideas generated from the North Sea studies or to continue work in east Greenland. Taking into account other factors (such as constraints on the length of the season, problems of bad weather and logistical difficulties), the decision was made against further study in Greenland. However, even with only one short season, conclusions from the field season in Greenland do have a bearing on the thesis and, hence, the preliminary results are presented in this chapter. The observations are the outcome of a single field season, so are incomplete and by no means conclusive. Of the four fault scarps studied, located on Fig. 8.1b, the volume and quality of the information gathered in the field varied substantially.

Consequently, the descriptions presented below are more extensive where observations were sufficient to propose a (preliminary) interpretation with some confidence (Månedal fault, Geographical Society Ø and Fosedalen fault, Hold With Hope) but shorter where observations were insufficient to draw firm conclusions and/or little could be added to previous interpretations (Clavering fault, Clavering Ø and Månedal fault, Traill Ø).

Although the faults studied grew under the influence of the same regional stress regime, the areas of exposure are widely distributed and, as a result, temporally equivalent deposits are of different facies in different areas. A correlation of facies and lithostratigraphic nomenclature are shown in Fig. 8.2a. The brevity of the field time meant that the majority of the observations made were spatial and macro-structural and that little detailed sedimentology (facies definition) or structural analysis could be undertaken. For more detailed sedimentological descriptions and palaeo-environmental interpretation the reader is referred in particular to the work of Maync (1949), Surlyk (1977, 1978), Surlyk and Clemmensen (1983), Price and Whitham (1997) and references therein.

8.3.1 Månedal fault, Geographical Society Ø

Structural Setting

The Månedal fault is a N-S striking, easterly dipping normal fault which can be traced for >50 km across Geographical Society Ø and Traill Ø in central east Greenland (Fig. 8.1b). At its southern extent in the Svinhufvuds Bjerge area of Traill Ø, the fault splits into two strands – fault A and fault B (see Section 8.3.4 and Fig. 8.22) – and, to the north, in central Geographical Society Ø the fault tips out. In the south of Geographical Society Ø, the Månedal fault is oriented sub-parallel to, and delineates the eastern flank of, the NNE-SSW trending, U-shaped valley of Tvaerdal (Fig. 8.3). The western margin of this valley is also defined by the trace of an easterly dipping normal fault – the Tvaerdal fault (the northward extension of the Bordbjerget fault on Traill Ø). The Månedal and Tvaerdal faults are near parallel in strike and spaced ~5 km apart. Hence, exposure in Tvaerdal is of geologies in the footwall of the Månedal fault (see cross section in Fig. 8.3). A lithostratigraphy for the Tvaerdal area has previously been defined by Maync (1947), Donovan (1955), Surlyk (1977) and Price and Whitham (1994) (see Fig. 8.2a).

Field observations demonstrate that the Tvaerdal fault was active during the Tertiary. Pre-rift sediments currently exposed in the footwall of the fault are of Carboniferous age (Fig. 8.4). Exposures in the proximal hangingwall are of continental red bed sediments of Triassic age and assigned to the Wordie Creek Formation; these sediments are unconformably overlain by the coarse, shallow marine sands of the Bajocian-Bathonian Vardekløft Formation (Donovan 1955). Pre-rift sediments are onlapped by northerly dipping, Late Albian mud-rich sediments (see Section 30, Enclosure 1 in Nøhr-Hansen 1993), deposited during the post-rift of the latest Jurassic – Early Cretaceous rift event. No sediments of Volgian-Valanginian age (i.e. the syn-rift of that event) are exposed. Footwall and hangingwall geologies are juxtaposed across a breccia belt defining the trace of the fault. This zone incorporates angular clasts of Carboniferous age sediments, a matrix of red muds (comparable in outcrop to Triassic mudstones in the hangingwall) and carbonate precipitates. Characteristically for the region, a dolerite dyke has intruded the fault plane. This study does not find any evidence indicating that the Bordbjerget-Tvaerdal fault was active during the latest Jurassic – Early Cretaceous and subsequently reactivated (as proposed by Price and Whitham 1994). In particular, the lack of Volgian-Valanginian age (syn-rift) sediments in a proximal position to the Tvaerdal fault suggests that there was no significant topography across the structure at this time (*cf.* Dombjerg fault, Surlyk 1978). Displacement in excess of 1 km accumulated on the Tvaerdal fault during the Early Tertiary.

In contrast, the Månedal fault shows no evidence of activity during the Tertiary. Albian age mudstones of the mid-Cretaceous sandy-shale sequence are continuous in bedding from the hangingwall across the trace of the fault and, although sedimentologically monotonous, show no evidence of offset. These northeasterly dipping sediments onlap (with a dip 2-7° less) northwesterly dipping, Triassic continental red beds in the footwall of the Månedal fault (Fig. 8.5). The steeply dipping denuded scarp of the Månedal fault is exposed on the north face of Tørvestakken on the eastern flank of Tvaerdal (Figs. 8.5 and 8.6). The displacement on the Månedal fault is cannot be constrained; pre-rift geologies in the hangingwall of the fault remain buried below mudstones of the post-rift. However, it can be concluded that displacement is in excess of 750 m (see section in Fig. 8.3) and accumulated during the Late Jurassic – Early Cretaceous.

Palaeo-topography on the footwall scarp of the Månedal fault

The footwall of the Månedal fault is significantly denuded, the crest of the structure backstepped and post-rift sediments onlap across the trace of the fault onto the scarp. The palaeo-topography on the footwall scarp, immediately prior to burial, is preserved at the interface between pre- and post-rift sediments. Exposures in the north side of the mountain of Tørvestakken permit examination of this unconformity and reconstruction of the morphology of the footwall. A series of views illustrating the palaeo-topography on the scarp are shown in Fig. 8.6.

In Fig. 8.6a the view is towards the south and the trace of the fault is to the left of the field of view (to the east) (Fig. 8.3). Below the summit of Tørvestakken a thick sill is exposed, intruding Triassic age pre-rift sediments in the footwall to the fault. Two prominent, strongly indurated beds – termed the 'hard bed' and the 'sandy bed' – are defined and demonstrate the northwesterly tilt of the sediments. The 'hard bed' is a diagenetically modified evaporite and the sandy bed a poorly sorted, pebbly fluvial deposit (Fig. 8.7). Dark, mud-prone sediments of the post-rift can be observed to directly onlap the eroded footwall across the fault trace. This contact records the palaeo-topography on the scarp. At the head of the valley (in the left of the view) mudstones unconformably onlap a bench of the 'hard bed'. In outcrop, the unconformity steepens to both the north and south. To the south, the palaeo-slope is steep and can be traced for *c.*150 m above the 'hard bed' until the scarp is truncated by the sill.

The continuation of the unconformity to the north is shown in Fig. 8.6b, a view of the northern flank of the valley shown in Fig. 8.6a. Again the dip of the pre-rift geologies is shallow and towards the northwest; the 'hard bed' is the prominent horizon in the view, the 'sandy bed' is exposed at the crest of the ridge. Outcrop of the 'hard bed' demonstrates that the footwall is dissected by a number of dominantly synthetic normal faults with <5 m displacement. Proximal to the palaeo-scarp, these faults decrease in spacing. The unconformity defining the palaeo-slope, visible in the right of the view, is apparently steep.

In Fig. 8.6c the view is further still to the north, but looking towards the east at the front of the ridge shown in Fig. 8.6b. The 'hard bed' is again the prominent bedding surface in this view and demonstrates a northerly dip, in contrast to the northwesterly dip observed in the south (Figs. 8.6a and 8.6b). The dip of the 'hard bed' increases towards the north (to in excess of 60°), defining a monocline. Late Albian mudstones of the post-rift are observed in the left of the view to directly onlap the top surface of the 'hard bed' where it forms the limb of the fold. As the dip of the footwall sediments shallows towards the crest of the ridge, the palaeo-scarp remains steep and younger sediments, overlying the 'hard bed' are truncated in the scarp.

From these observations the palaeo-topography of the footwall scarp can be reconstructed for this area. A schematic 3-dimensional model of the topography of the Månedal fault scarp pre-burial is presented in Fig. 8.8a. In this reconstruction, the upper bedding plane of the 'hard bed' horizon forms a distinct topographic bench on the scarp; overlying this bench the poorly indurated sediments, and the 'sandy bed', are retreated westwards. The primary controls on the morphology of the scarp are footwall lithology (specifically, the relative 'strengths' of pre-rift geologies) and the dip of bedding. In this model the monocline affecting geologies in the northeast of the study area is attributed to fault-controlled folding; invoking a blind, W-E oriented fault. Although perpendicular to the dominant strike of faults in central east Greenland, the proposed blind fault is located at the northernmost traceable extent of the Månedal fault, i.e. the fault is decreasing in displacement and probably tips out in north Tvaerdal. Cross-faults at a high angle to the main trend have been described in similar positions by Larsen (1988), Gawthorpe and Hurst (1993), Trudgill and Cartwright (1994), Cartwright *et al.* (1996) and Gilpin (1998) and related to the fault linkage process.

Palaeo-topography on the dip slope of the footwall of the Månedal fault

Recent glaciations have eroded the much of the footwall of the Månedal fault to below the top of the pre-rift. Hence, the current outcrop pattern only permits local reconstruction of the pre-burial topography on the dip slope. The palaeo-topography can be demonstrated where the unconformity between the pre-rift and the post-rift can be mapped, i.e. in the north of the study area and in the hangingwall of the Tvaerdal fault. The form of the contact is illustrated in the views shown in Fig. 8.4, of the western flank of Tvaerdal, and in Fig. 8.9, where Cretaceous mudstones overlie Jurassic sands.

The dip of Triassic sediments in the footwall to the Månedal fault is generally shallow (10-15°) towards *c.*285°. The top surface of these geologies is not concordant with a bedding plane, but rather has an erosional relief. This is demonstrated by the map of the pre-rift – post-rift contact in the northwest of Tvaerdal ('A' on Fig. 8.3). This contact can be traced across the valley the post-rift overlying progressively younger stratigraphy from east to west, Triassic mudstones then Jurassic sandstones. In the south of the study area ('B' on Fig. 8.3), the strike of the Triassic sediments changes to a WSW-ENE trend and the strata overlying the pre-rift sediments progressively ages to the south. The change in strike of the pre-rift geologies is mirrored by the Cretaceous sediments and defines a wide, low amplitude fold – interpreted as a fault controlled extensional structure (Schlische 1995) developed during the growth of the Tvaerdal fault in the Tertiary.

The erosional relief of this contact is best demonstrated around the exposure of the Vardekløft Formation (Fig. 8.9). An *c.*200 m high exposure of Bathonian-Bajocian sediments, the deposits of a fluvio-deltaic environment, form a ridge on the west side of Tvaerdal (Figs. 8.3 and 8.4). An angular unconformity with the underlying red beds is observed near the base of the exposure and, at the top, the Jurassic sands are unconformably onlapped by Late Albian mud-rich sediments. To the north and south of the ridge, the sands are erosionally truncated, describing steep palaeo-slopes, the unconformity on the surface representing ~70 Ma. The view in Fig. 8.9 of the southerly of the two palaeo-slopes bounding the Jurassic outcrop illustrates that the relief on the slope is not insignificant and that the dip of bedding in the marginal marine sediments controls the angle of the palaeo-slope.

A schematic 3-dimensional representation of the eroded footwall of the Månedal fault constructed from these observations is shown in Fig. 8.8b. Similarly to the reconstruction of the scarp, strongly indurated horizons are shown to form benches in the palaeo-topography (e.g. the 'hard bed') and form erosion resistant caps to ridges (e.g. the Jurassic sands). Ridges and gullies, with up to 250 m of topography, are suggested on the dip slope. Again, lithology and the dip of the stratigraphy are interpreted to be the primary controls on the palaeo-terrain.

Process(es) of, and products from, the degradation of the footwall of the Månedal fault

Limited evidence is available to indicate the process(es) by which the footwall of the Månedal fault was degraded. Sediments shed to the east during the rift phase remain deeply buried in the hangingwall of the fault. On the dip slope, the contact between pre- and post-rift geologies is 'clean' – no syn-rift deposits are preserved. Subsequent activity on the Tvaerdal fault has exhumed and eroded sediments deposited further to the west. However, the cessation of tectonic activity in central east Greenland is interpreted as during the latest Valanginian (Surlyk 1978), and the mudstones burying the footwall of the Månedal fault are of Albian age. This indicates that the footwall was a topographic high for ~20 Ma following the end of the rift event. As a high, the footwall would have continued to be eroded post-tectonically and the products of this erosion should be preserved in proximal onlapping sediments.

Exposed sediments of mid-Cretaceous age in the study area are mud-rich. The dominant facies is a dark, micaceous mudstone with a sparse macrofauna, deposited in a fully marine basin below storm wavebase. The muds are interbedded with thinly bedded, fine grained quartz-rich sandstones, interpreted as the deposits of turbulent flows. It is suggested that some of the sand-sized sediment may have been derived from degradation of footwall geologies. The insignificant volume of footwall derived debris in the mid-Cretaceous age deposits suggests that either sediment being supplied from footwall degradation was transported and deposited elsewhere in the basin or that the rates of footwall erosion were extremely slow at this time. Considering the total magnitude of footwall denudation was relatively small (reconstruction in Fig. 8.3) and the palaeo-environment during the Cretaceous, this study prefers the latter of these two options.

In addition to the sands observed in the post-rift stratigraphy, displaced sediments attributed to scarp erosion are preserved mounted on the steep palaeo-scarp below the summit of Tørvestakken (Figs. 8.3 and 8.6a). In this area, the palaeo-slope has an apparent dip of 60-70°; heterolithic Triassic age sediments in the footwall are onlapped by mudstones of the post-rift. The slope can be traced for a vertical height of 250-300 m between a bench on the 'hard bed' below and truncation in the thick sill below the summit of the mountain. The displaced sediments form only a thin veneer, 0.1-3 m thick, on the palaeo-scarp, and are illustrated in Fig. 8.10. Typically, the contact between pre-rift and post-rift is not a storage area for degradation products, and this is the only locality in the study area where the remnants of degradation are preserved.

In outcrop, the displaced sediments take the form of elongate 'teardrops' (Fig. 8.10a). Below Tørvestakken three lobes are observed, of up to 15 m length and thickening downslope from 0.1 m to >3 m. Internally, the structures are composed of sediments of similar character to the geologies of the immediately underlying footwall. Primary bedding and depositional structure within the lobes is well preserved, although bedding is substantially rotated with respect to the footwall – in the lobes bedding is oriented near parallel to the palaeo-slope. In contrast to the main body of these structures, sediments in the basal ~70 cm show evidence of intense shearing (Fig. 8.10b). A strong, base parallel fabric is developed in the sediments, primary sedimentary structure is lost and mud clasts are stretched parallel to the fabric. However, the contact with the underlying footwall is sharp and sediments immediately beneath the lobes have an undisrupted primary depositional structure.

From observations of location, morphology and internal structure, these structures are interpreted as the deposits of a process of sediment sliding, i.e. the lobes are rotational slide blocks (using the classification scheme of Varnes 1978) that were transported on a basal shear complex of mud-rich sediment. The similarity of the displaced sediments to underlying footwall geologies and the lack of deformation suggest that the downslope transport distance was short. It is suggested that small displacement faults (e.g. Fig. 8.6b) may have been exploited as décollement horizons for slide blocks. Whether these deposits are the basal remnants of much larger structures, the body of which has been removed, is unclear. It is noted, however, that these structures were probably formed during the long exposure of the footwall high prior to burial, when gravity was the only driving force for downslope sediment movement and there was no seismic trigger. Additionally, there is no preservation of large (+100 m scale) slide blocks of rotated, competent footwall geologies in the footwall underlying the top pre-rift unconformity.

8.3.2 Fosedalen fault, Hold With Hope

Structural Setting

The Fosedalen fault is the most significant, in terms of length and displacement, of number of easterly-dipping normal faults in the Home Foreland area (Gulelv-Kap James) of the Hold With Hope peninsula, central east Greenland (Fig. 8.11). Although the trace of the fault is largely buried by alluvium, the Fosedalen fault can be mapped parallel to, and dissecting, the ~10 km length of the N-S oriented, U-shaped valley of Fosedalen. The fault offsets contemporaneously extruded Ypresian (Upton *et al.* 1995) flood basalts, downthrowing the lavas several hundred metres in the east (see Upton *et al.* 1980, their Fig. 3). This clearly indicates that the fault was active during the Early Tertiary. Pre-Tertiary sediments below the basalts in Home Foreland have previously been mapped and described by Vischer (1943), Maync (1949), Nøhr-Hansen (1993) and Kelly *et al.* (1998). This study follows the lithostratigraphic framework described by Kelly *et al.* (1998) (Fig. 8.2).

To the east of Fosedalen, on the downthrown side of the fault, outcrop below the basalts is dominated by Albian-Santonian mudstones interbedded with thin sandstones (Kelly *et al.* 1998). The muds are separated from the flood basalts by a 2-3 m thick conglomerate, which has been ascribed a Tertiary age by correlation with similar exposures in Clavering Ø (Vischer 1943). To the west of Fosedalen, on the footwall of the fault, equivalent Middle Albian – Cenomanian age mudstones are exposed. These onlap, to the west, a steep palaeo-slope with at least 850 m of relief (see cross section in Fig. 8.11b). The westerly dipping sediments forming the slope are of Triassic – Early Albian age, with angular unconformities between the Wordie Creek and Vardekløft formations and the Vardekløft and Steensby Bjerge formations (Fig. 8.12). A further small angular unconformity separates the Home Foreland Formation and the flood basalts. The palaeo-slope is interpreted as the denuded scarp of a proto- Fosedalen fault. The age of the onlapping sediments and evidence of rotation of geologies in the footwall constrains activity on this structure to the late Jurassic – Early Cretaceous.

The sedimentary succession exposed in the palaeo-slope is heavily faulted (Fig. 8.11). Easterly dipping, N-S striking, synthetic normal faults at sub-km spacing offset the geologies by up to 300 m. The distribution and thickness of Early Albian – Triassic sediments, including the easterly thinning Steensby Bjerge Formation, are unaffected by the faults. Early Tertiary age intrusions are generally not offset by the structures. The sedimentological monotony of the Home Foreland Formation makes it difficult to determine if the sediments have been displaced by the faults. Detailed sedimentological and palynological work (A. Whitham *pers. comm.* 1997) would suggest they have not, and, hence, constrain the movement on these faults to a short period in the mid-late Albian, between the deposition of the Steensby Bjerge and Home Foreland formations. In addition to these large structures, the sedimentary succession is heavily fractured and accommodates numerous, sub-m throw normal faults.

Palaeo-topography on the footwall scarp of the Fosedalen fault

Similarly to the Månedal fault on Geographical Society Ø, palaeo-topography on the footwall scarp of the Fosedalen fault is preserved at the interface between the pre- and post-rift sediments of the mid-Mesozoic rift event.

The easternmost exposures of pre-rift geologies in the footwall of the Fosedalen fault are located at the mouth of River 23, ~5 km west of the current outcrop of the fault. In fjord-side exposures, Middle – Late Albian (after locality P2069 of Kelly *et al.* 1998) mudstones are observed to onlap, to the west, a steep slope of interbedded red mudstones and yellow sandstones of the Wordie Creek Formation (Fig. 8.13). The contact between pre- and post-rift can be mapped along the steep sided flanks of, and is dissected at a low angle to strike by, the SW-NE oriented valley of River 23. On the northwestern margin, heavily fractured muds and sands of the Wordie Creek Formation are exposed. On the opposite flank, the lower *c.*100 m of outcrop is also heterolithic Wordie Creek Formation, unconformably overlain by mudstones of the Home Foreland Formation (Fig. 8.14). A 45 m thick dolerite sill caps the southeastern side of the valley. For much of the length of the southeastern flank the palaeo-slope corresponds with the top of a 1 m thick, intensely bioturbated, fine grained quartz arenite. At the head of the River 23 valley, the Wordie Creek Formation is truncated on both flanks defining *c.*60 m high palaeo-slopes. Muds of the Home Foreland Formation onlap the slopes.

To the west of River 23, the unconformity beneath the Home Foreland Formation can be traced around the flanks of another river valley, Rødelv. Exposure in Rødelv is dominated by poorly indurated, fine grained sediments of Triassic age. Overlying these are easterly thinning (*c.*300-50 m) sand-rich successions, which form an amphitheatre of steep cliffs around the wide valley. Mud-rich sediments of the Rødelv Member of the Steensby Bjerg Formation and the Home Foreland Formation are more prone to erosion, and are exposed above a prominent topographic bench defining the top bedding surface of the Barremian – Early Albian sands. The mud-mud contact between the Rødelv Member and Home Foreland Formation is difficult to confidently map in the field; however, strata of the Rødelv Member can be locally demonstrated to be truncated and onlapped, suggesting an erosional relief as inferred in Fig. 8.11.

Figure 8.11b presents a 2-dimensional palinspastic reconstruction for the Home Foreland area to the top of the Home Foreland Formation, based on the cross section AA'. This reconstruction qualifies the degradation of the footwall and illustrates that only to the east of the River 23 area was denudation deep. Quantifying the magnitude of erosion is not possible as the location of and the throw on the bounding fault are unknown. Again, strongly indurated beds (in this instance a quartz arenite) are shown to form topographic benches on the palaeo-slope (Fig. 8.14). Significantly, the coarse sandstone successions exposed in the footwall thin to the east; hence the majority of the degraded sediments are fine grained geologies of the Rødelv Member and the Wordie Creek Formation. It is suggested that the strongly indurated sandstones may have formed a protective cap over the footwall, inhibiting further crestal retreat.

Process(es) of, and products from, the degradation of the footwall scarp of the Fosedalen fault

This study describes a >5 km wide palaeo-slope with at least 850 m of relief, now exposed in the River 23 – Rødelv area of Home Foreland. The slope is interpreted as the degraded footwall scarp of the Fosedalen fault, currently buried to the east of the area. The magnitude of erosion was strongly controlled by the relative strengths of the footwall geologies. The contact between pre- and post-rift sediments is sharp – there is no mantle of displaced, degraded sediments on the scarp. Adjacent footwall sediments, while heavily fractured, are not recognised to form part of a slide block and, indeed, shear surfaces are not observed on the slope. The sediments onlapping the slope are of Middle Albian – Cenomanian age. In a proximal location, these geologies are dominantly of mud grain-size, with only thin coarse clastic pulses, although clasts attributed to eroded Jurassic and Triassic footwall lithologies have been documented in Fosedalen, 3-4 km to the east (Kelly *et al.* 1998).

With this information from only a reconnaissance season, the process(es) of fault scarp degradation cannot be defined. Further work on this problem needs to be undertaken (see Section 8.4).

8.3.3 Clavering fault, Clavering Ø

Geology of the Dolomitdal-Djævlekløften area, Clavering Ø

The N-S striking, easterly dipping Clavering fault splits Clavering Ø into two geological regimes (Fig. 8.1b). In the east of the island in the hangingwall to the fault, sediments of Mesozoic age are exposed. In contrast, outcrop in the west is of Carboniferous geologies and, in the proximal footwall, metasediments deformed during the Caledonian Orogeny. According to the maps of Koch and Haller (1971), Clavering Ø is located in an area of overlap of two segments of the basin bounding Post-Devonian Main Fault (Fig. 8.1b); the fault system steps ~30 km to the southwest between Wollaston Foreland and Hold With Hope (*cf.* transfer zones of Rosendahl *et al.* 1986). The Clavering fault has been interpreted by Maync (1949), Koch and Haller (1971) and Surlyk (1977, 1978) as the southward extension of a segment of the PDMF: The Dombjerg fault on Wollaston Foreland. The implication of these observations is that the Clavering fault is a segment near the southernmost tip of a major basin bounding fault system.

In NE Clavering Ø, the Clavering fault is exposed in the deep, U-shaped valleys of Dolomitdal and Djævlekløften. The axes of these valleys trend approximately W-E, near perpendicular to the strike of the fault; hence footwall geologies are exposed in the head of the valleys and sediments in the hangingwall near the coast (Figs. 8.15, 8.16 and 8.17). In the hangingwall an *c.*500 m thick succession of Bathonian clastics, dominated by coarse sands deposited in a shallow marine setting, are exposed (assigned to the Vardekløft Formation by Surlyk 1977; see Fig. 8.2) (Fig. 8.16). The sands are overlain, with an erosional contact, by up to 70 m of conglomeratic sediments of Volgian-Valanginian age, which incorporate rounded – sub-angular clasts (<1-50 cm) of quartzite, gneisses and limestone (assigned to the Rigi and Young Sund members by Surlyk 1978) (Fig. 8.20). Equivalent along strike sediments, in the proximal hangingwall of the Dombjerg fault in Wollaston Foreland, are 2-3 km in thickness, conglomeratic and have been interpreted as footwall sourced debris redeposited in a fan delta environment (see Section 8.2). In the study area, Middle Jurassic and Early Cretaceous geologies, and the metasediments and carbonates exposed in the footwall, are buried by Middle Albian (Nøhr-Hansen 1993) mud-prone sediments. To the south of Djævlekløften, Maync (1949) has documented coarse debris flow deposits composed of footwall lithologies interbedded in the mid-Cretaceous mudstones ('Boulder Ridge', his Fig. 35; see Figs. 8.18 and 8.21).

The Clavering fault outcrops in two localities in the study area: Dislokationdal and Kontaktravine (Fig. 8.15). The fault plane itself is poorly exposed but, in both localities, sands of the Vardekløft Formation and basement lithologies are observed to be truncated by, and juxtaposed across, the fault (e.g. Fig. 8.19). The mid-Albian mudstones deposited across the trace of the fault have not subsequently been offset. Hence, activity on the Clavering fault can be constrained to the latest Jurassic – Early Cretaceous, and the fault was not reactivated during the Early Tertiary; Volgian-Valanginian conglomerates were deposited during rifting.

The magnitude of denudation of the footwall scarp of the Clavering fault

The sediments exposed in the Dolomitdal-Djævlekløften area have previously been mapped and described in some detail by Vischer (1943), Maync (1947, 1949) and Surlyk (1977, 1978). Both Maync and Surlyk have demonstrated that the footwall scarp of the Clavering fault was a topographic high, being eroded by rockfall processes, from the Volgian until, at least, the mid-Albian. A reconstruction of the palaeo-topography during the Albian is illustrated in Mayncs' (1949) Fig. 36, reproduced in Fig. 8.21. Observations made during this reconnaissance field season support the interpretations and palaeo-environmental reconstructions of the previous workers and emphasise the denudation of the scarp.

The W-E oriented cross section in Fig. 8.15b reconstructs the morphology of the footwall of the Clavering fault. The throw on the fault in the Dolomitdal-Djævlekløften area is unknown. However, assuming that the Vardekløft Formation was deposited directly on the basement and no Carboniferous-Triassic strata were deposited in the area, as suggested by exposures on Wollaston Foreland (Fig. 8.2b), the current exposure indicates a minimum displacement of >1 km. Significantly, the fault plane is exposed in Dislokationdal and Kontaktravine, permitting the magnitude of backstepping of the scarp to be constrained – between <0.5 km (B on Fig. 8.15) and ~2.5 km (A on Fig. 8.15). It is recognised that the retreat of the crest was significantly higher (as suggested on Fig. 8.15b), but these measures do illustrate the variable morphology of the footwall scarp across the study area. For example, at the mouth of Djævlekløften (Fig. 8.18), the exposed palaeo-slope is steep (+60°) and backstepped <0.5 km from the trace of the fault. In contrast, in Dolomitdal (Fig. 8.17), the contact between pre- and post-rift geologies is substantially shallower before steepening at the head of the valley. In Dolomitdal, the footwall is backstepped ~1.75 km (C on Fig. 8.15).

The syn-rift geologies in the Dolomitdal-Djævlekløften area are <100 m thick – dramatically thinner than in the equivalent fault proximal position adjacent to the Dombjerg fault on Wollaston Foreland. This suggests that debris derived from syn-rift degradation of the scarp of the Clavering fault was reworked and transported in the hangingwall rather than deposited locally. As the location on Clavering Ø is near the tip an ~100 km long segment of the PDMF (Fig. 8.1b), it is suggested that a northward palaeo-slope (a consequence of extensional folding) may have redirected sediment axially from the Dolomitdal-Djævlekløften area into along strike hangingwall depocentres.

8.3.4 Månedal fault, Traill Ø

Coastal exposures on southern Traill Ø dissect a W-E oriented dip section through an ~11 km wide normal fault bounded block (Figs. 8.1 and 8.22). The structure is bounded by easterly dipping, large displacement (>1 km) faults – the Bordbjerget fault in the west, and the 2 splays, A and B, of the Månedal fault in the east (*cf.* Tvaerdal, Section 8.3.1). The footwall between the faults is composed of Triassic, unconformably overlain by Middle Jurassic, pre-rift strata (assigned to the Wordie Creek and Vardekløft formations, Christiansen 1990) (Figs. 8.22 and 8.23). These sediments are intruded by numerous Tertiary magmatics. Post-magmatic movement is observed on both large and small displacement faults (see Price *et al.* 1997), including the Bordbjerget fault and fault A (Fig. 8.23).

Exposure between faults A and B in the Svinhufvuds Bjerge area indicates that fault B accumulated significant displacement during the latest Jurassic – Early Cretaceous. A denuded palaeo-slope of Traissic and Jurassic geologies, onlapped by Upper Albian mud-rich sediments (Nøhr-Hansen 1993), can be mapped over a vertical distance of ~800 m (Figs. 8.23 and 8.24). This slope is interpreted as the palaeo-scarp of fault B, established during latest Jurassic – Early Cretaceous rifting and exposed until the latest Albian. The post-rift sediments onlapping this slope, while dominated by mud-rich hemipelagic deposits, incorporate sandy horizons. Sand-rich beds vary in thickness from mm-m scale, but are typically composed of well sorted sands and, hence, interpreted as the deposits of gravity driven mass flows, possibly travelling axially in the proximal hangingwall of fault B. In addition, clasts of sandstone varying in size from <1 cm to +10 m are preserved, either in isolation embedded in the post-rift muds or as part of chaotic beds interpreted as the deposits of debris flows (Fig. 8.25a). These clasts were most likely proximally sourced, from degradation of the footwall. Reworked sand and mud laminae, the deposits of sediment slumping, are also observed (Fig. 8.25c,d). Isolated exposures of conglomeratic beds (80-100 cm thick), incorporating small (<5 cm), angular quartzite and gneiss pebbles and rounded concretions, are observed within the post-rift (Fig. 8.25b). These sediments are comparable in character to the 'Pachyteuthis Beds' described by Donovan (1955) from elsewhere on Traill Ø. Similar geologies exposed on Wollaston Foreland (Rødryggen Formation) have been interpreted by Surlyk (1978) as condensed deposits. The occurrence of these strata interbedded with post-rift mudstones suggests that the sediments maybe reworked, possibly derived from crestal deposits.

The palaeo-scarp of fault B is mantled along 200 m of its length by a block of displaced sediment (Fig. 8.24). This unit is 'lobe-shaped', increasing in thickness downslope to ~20 m; the along strike continuity of these sediments is unknown. It is composed of undeformed sands of affinity to the Vardekløft Formation, with a narrow (<20 cm) basal shear zone and local 'jigsaw breccia' (shattered, but not disaggregated, sediments; after Shreve 1965, 1968). This unit is interpreted as the remnant of a slide block (*cf.* Tvaerdal, Section 8.3.1), and has a similar morphology and characteristics to the tail of a long runout landslide (e.g. Campbell *et al.* 1995).

8.4 Conclusions from reconnaissance study and plans for future work

It is clear from the observations reported in the preceding section that the exhumed rift geologies of the latest Jurassic – Early Cretaceous extension in central east Greenland present an excellent opportunity to study partially degraded footwalls. This preliminary field investigation has demonstrated that:

1. The magnitude of denudation of the footwall, on both the scarp and dip slope, of these large displacement normal fault arrays is relatively small, certainly in terms of the uplifted volume. The process(es) of degradation and, hence, the morphology of the eroded footwall scarp are primarily controlled by the rheology and structure of the pre-rift geologies subcropping in the slope.

2. The unconformable contact between pre-rift and post-rift geologies on the fault scarp is generally not a storage location for sediments degraded from the footwall. Such displaced sediments were only observed in one locality (Tvaerdal, Section 8.3.1), where m-scale lobes of pre-rift geologies are interpreted as the remnants of slides.

3. Footwalls remained topographic highs for geologically significant periods of time (locally in excess of 30 Ma) after the cessation of fault activity until buried by post-rift strata. Denudation of the footwall continued during the post-rift but the products of degradation are rarely observed interbedded with proximal post-rift (hemipelagic) sediments, suggesting that rates of erosion and, hence, sediment supply were low (assuming local redeposition of degradation products).

The implications of these conclusions, and the conclusions of subsurface investigations from the northern North Sea (presented in Chapters 6-7), for sediment supply and the nature of fault proximal deposition in marine rift basins are discussed in Section 8.5 below.

The preliminary conclusions from the reconnaissance study demonstrate the potential to address a number of key issues through further studies in the field area. Hence, a second field season to central east Greenland was planned for the summer of 1998. Plans for future work realised the limitations of the study area, particularly the 2-dimensional nature of the exposure and the typically deep burial of associated fault proximal syn-rift deposits. Three principal areas of investigation were determined and are summarised below.

The geometry of extensional folds

The response of pre-rift geologies to the stresses imposed by extensional folding is unknown. Analogue models of deformation at the tip of a vertically propagating fault have shown that brittle failure, in both the reverse and normal sense, accompanies the development of a monocline (e.g. Withjack *et al.* 1990, Mitra 1996; *cf.* field studies of Sharp *et al.* 1998). Brittle failure would also be predicted to accommodate the local stresses associated with large wavelength folding of both the hangingwall and footwall developed in response to along strike displacement variations on normal faults. This is supported by the observations of the deformation of the footwall of the Strathspey-Brent-Statfjord fault presented in Chapter 6 (Fig. 6.8). Reconnaissance observations on Geographical Society Ø document large scale folding of pre-rift geologies and indicate faulting parallel to the axis of the principal stress (σ_3); future work was proposed to map the structure of the pre-rift geologies in this area to examine the relationship between extensional folds and brittle deformation.

The morphology of degraded footwalls

The rheology of pre-rift geologies in the footwalls of normal faults is a primary control on the mechanism of slope failure, and, ultimately, the morphology of the scarp. The lithological control on slope failure is well quantified for the subaerial realm and numerical models are available which successfully predict downslope movements on a slope for specified rheological and environmental parameters (e.g. Newmark 1965, Graham 1984). However, the effects of a submarine environment on the mechanics of downslope transport are less well constrained and field studies present the opportunity to attempt to replicate the erosion of a palaeo-scarp by modification of numerical models (*cf.* Jibson and Keefer 1993). The palaeo-scarp of the Fosedalen fault on Hold With Hope is dissected by two valleys – River 23 and Rødelv – the exposure affording an element of 3-dimensions to the slope. Hence, a (local) 3-dimensional map of the palaeo-slope (generated by surveying using differential GPS or plane table/altimeter techniques) could be directly related to the rheology of the underlying footwall geologies. This morphology could then be contrasted with the results of numerically modelling the failure of the slope, using 'real' field derived rock properties.

Footwall denudation during the early post-rift

One of the main limiting factors in studying the process(es) of scarp denudation in central east Greenland is that syn-rift deposits, and hence the products of degradation, largely remain deeply buried in the hangingwall. However, as the basin was sediment starved during the rift phase, footwalls remained topographic highs for up to 30 Ma following the cessation of tectonic activity. As a consequence, scarp denudation continued into the early post-rift. The relative importance of post-rift denudation, in terms of rates, process(es) (particularly considering the lack of a seismic trigger for failure) and volume of sediment liberated, compared to syn-rift slope degradation are unknown. In addition, the rôle of reworking and transportation in the hangingwall is unconstrained. The opportunity exists to address these issues in Greenland, where the products of slope erosion are documented interbedded with the post-rift mid-Cretaceous sandy-shale sequence, e.g. on Traill Ø (Fig. 8.25) and Clavering Ø (by Maync 1949).

8.5 The rôle of footwall degradation as a source of sediment in marine rift basins

Reconnaissance field observations presented in this chapter have been employed to reconstruct, in 2-dimensions, the morphology of the degraded footwalls of latest Jurassic – Early Cretaceous faults exposed in central east Greenland (e.g. Figs. 8.3 and 8.11). Comparable observations are presented in Chapters 6-7 from study of the denuded footwall of the >60 km long Strathspey-Brent-Statfjord fault, located in the northern North Sea basin. The results from the two study areas are examined in this section with the goal of qualifying and quantifying the rôle of syn-rift footwall degradation as a source of clastic sediment in the basins. The implications of these results for hydrocarbon exploration are examined in Chapter 9.

The magnitude of crestal denudation of footwalls in the northern North Sea and central east Greenland is striking and comparable (Fig. 8.26). Considering that all the footwalls studied were topographic highs for in excess of 25 Ma and had up to 1 km of elevation with respect to the proximal hangingwalls, the magnitude of erosion is pitiful. The Månedal fault in Tvaerdal (Fig. 8.26a), for example, has lost approximately 125 m of topography at the crest of the footwall; an elevated region between the latest Volgian – Late Albian. This equates to an average erosional decline in elevation of <3 mm/ka. In extreme contrast, Burbank *et al.* (1996) have shown that some areas in the northwestern Himalayas are denuding at rates of >1 mm/yr, largely attributed to a 2-12 mm/yr vertical bedrock incision by the Indus River. The consequence of such rapid denudation is the large sediment yields (sediment load divided by the area of the drainage basin) of rivers draining mountain belts (e.g. Milliman and Syvitski 1992). Obviously with much lower rates of erosion, the volume of sediment liberated by the degradation of the footwalls considered in this study was significantly less than that of modern subaerial ranges. The total volume of sediments denuded (over 60 Ma) from the footwall of the Strathspey-Brent-Statfjord fault system (Fig. 8.26c) was approximately 35 km^3, equating to an average erosion rate (sediment supply) of <600 m^3/ka. Although the average rate of denudation was low it is noted that the rate was probably temporally variable; the bulk of the erosion occurred during the climax of the rift phase (~12 Ma) when the scarp was growing and slope failure processes were seismically triggered. In addition the processes of downslope transport mean that sediment supply from scarp erosion was probably pulsed, for example, by geologically instantaneous events such as large slides.

The development of substantial elevation and the low rates of denudation of the footwalls described in this study can be attributed to the submarine setting common to both the northern North Sea and central east Greenland basins during the climax of rifting, when the majority of the displacement was accumulated on the faults (see Chapter 3). Significant elevation change across a fault plane, from uplifted footwall to downthrown hangingwall, will develop if a basin is sediment starved and the accommodation space is underfilled, i.e. the rate of tectonic subsidence exceeds the rate of sediment supply. Low rates of sediment supply are predicted in the interior of submarine rift basins as a consequence of partitioning the extra-basinal clastic supply at the basin margins; hence within the basin sediment supply is internally derived, i.e. from the water column and erosion of local bathymetric highs

(Ravnås and Steel 1998). The dominant process of syn-rift sedimentation in the study areas was hemipelagic settling and the rates were low (estimates in the Strathspey-Brent-Statfjord area are 3-25 cm/ka, assuming 30 % compaction). During the rift climax the bulk of the syn-rift sediments were locally derived. As a consequence the basins were starved and a substantial topography developed at the crests of the footwalls. This is emphasised by the volume of remnant accommodation in the basins at the end of the rift event (Fig. 8.26).

A secondary requirement in order that topography will be established across a fault plane is that the rate of erosion of the footwall geologies must be lower than the rate of uplift. As the elevation of a footwall scarp increases, mirroring the accumulation of displacement on the bounding fault, so the potential energy of the footwall increases – resulting in slope failure events, largely triggered by seismic shocks (Keefer 1984, 1999). The balance between the driving (gravity) and resisting (coherency of the rock mass) forces controls the initiation of slope failure. The flux of sediment can be categorised as dominated by diffusive (essentially sediments degraded by weathering) or landsliding processes (e.g. Allen 1997). Recent studies have illustrated that fluvial undercutting, steepening slopes and triggering landslides, is the most important process in the rapid erosion of mountain belts (Burbank *et al.* 1996, Densmore *et al.* 1997, Hovius *et al.* 1997, Densmore *et al.* 1998, Hovius *et al.* 1998). In a submarine environment the processes of downslope transport are limited; only gravity driven processes (flows, slumps, slides, falls) can occur. Submarine slopes are generally not modified (incised and steepened) by channelised flow. Consequently it is suggested that a slope can achieve 'stability', i.e. the gravitational potential to drive downslope processes is insufficient to overcome the resisting forces.

It can be concluded that a footwall high will be established as a consequence of the low rates of sediment supply in a submarine environment, but that the submarine environment will inhibit the processes, and hence rates, of erosion of the uplifted geologies. Significantly, it is observed that the fault scarps studied in the North Sea and east Greenland remained submarine during the climax of rifting and the early post-rift (up to the time when the footwalls were finally buried).

The morphology of the footwalls studied in the northern North Sea and central east Greenland rift basins represent near end members: The rates of syn-rift sedimentation were low and major fault bounded bathymetric highs were established, but the rates of erosion were low and, consequently, the sediments supplied from degradation were of minor volume. This combination of factors is attributed to the submarine environment of the half-graben basins during the climax of the rift event and the early post-rift period. The impact of submarine conditions in limiting the processes of erosion are emphasised by comparison of Figs. 8.26b and 8.26c, fault perpendicular cross sections across the footwalls of the Strathspey-Brent-Statfjord and Zeta-Snorre-Gullfaks fault systems, respectively. The two sub-parallel fault arrays grew contemporaneously and proximally (see Chapter 4) – the southern tip of the Inner Snorre fault overlaps with the northern tip of the Strathspey-Brent-Statfjord fault (Fig. 4.23). However, the morphology of the footwall of the Inner Snorre fault is strikingly different to that of the Strathspey-Brent-Statfjord fault. The crest of the Inner Snorre structure is demonstrated to be near planar (accounting for a regional post-tectonic tilt to the south) and the volume of pre-rift sediments denuded from the footwall is three-four times greater than the volume degraded from the footwall of the Strathspey-Brent-Statfjord fault.

The flat-topped profile of the footwall of the Inner Snorre fault has been be interpreted to reflect peneplanation to wavebase, with the implication that the uplifted footwall was an isolated 'crestal island' (see Fig. 9 of Dahl and Solli 1993). Processes of footwall erosion were thus subaerial, or at least enhanced by wave action, and rates of erosion were relatively high, certainly in comparison to the degradation of the proximal Strathspey-Brent-Statfjord structure exposed for an equivalent period of time. The substantially less denuded footwall of the Strathspey-Brent-Statfjord fault system shows no evidence of uplift to at or above sea level. The magnitude of uplift of a footwall has been attributed to the width of the fault block (e.g. Yielding *et al.* 1992): The wider the fault block the greater the uplift. This is consistent with the predictions of models relating strain distribution to rates of displacement (Gupta *et al.* 1998). The footwall bounded by the Inner Snorre fault is of greater width than the width of the footwall bounded by the Strathspey-Brent-Statfjord fault: Approximately forty kilometres and ~15 km, respectively. Maximum displacement on the Inner Snorre fault is in excess of 2.5 km (J. Underhill *pers.comm.* 1999), significantly greater than the maximum displacement on the Strathspey-Brent-Statfjord fault system (~1.7 km).

This study has demonstrated that the majority of the denudation of the Inner Snorre footwall occurred during the climax of the rift event; the sediments liberated supplied an axial turbidite system active in the hangingwall to the Strathspey-Brent-Statfjord fault (Chapter 4). While local footwall erosion is relatively unimportant as a source of syn-rift sediment in the Strathspey-Brent-Statfjord area, it is shown that the erosion of the footwall of a proximal fault is important. These observations imply that footwall degradation can be a voluminous source of clastic sediment in marine rift basins, but the depth of the crest with respect to wavebase primarily controls the magnitude of degradation.

CHAPTER 9 Implications and conclusions

9.1 Introduction

This study has examined a simple thesis: To determine the controls on syn- and post-rift sediment source, dispersal and deposition in marine rift basins. This aim has largely been addressed through a detailed subsurface study of the half-graben bounded by the Strathspey-Brent-Statfjord fault system; reconstructing the history of the fault population and the evolution of syn- and post-rift sedimentary systems. In particular this work has emphasised the rôle of the evolving fault population in controlling the location and magnitude of accommodation space generated during the rift event and, consequently, determining the dispersal and deposition of sediments in the basin. Supported by additional field studies in central east Greenland, the denudation of submarine, fault-bounded footwall highs has been demonstrated to be minimal, certainly in terms of the period of exposure of footwalls prior to post-rift burial (commonly in excess of 50 Ma). Hence, it is concluded that footwall erosion generally liberates an insignificant volume of clastic supply to marine rift basins.

While the observations presented in this thesis are specific to the unique study areas, the interpretations and conclusions do still have important implications for the wider extensional provinces of the northern North Sea and central east Greenland and, indeed, for rift basins in general. In this chapter, the key conclusions of this study are presented and the broader implications of these results for understanding the evolution of large normal fault arrays, the controls on syn-rift sedimentation and the rôle of footwall denudation as a source of sediment are assessed. The chapter concludes with an examination of the ramifications of this work for hydrocarbon exploration and exploitation in rift basins.

9.2 Understanding the evolution of normal fault populations

Key conclusions from study of the Strathspey-Brent-Statfjord fault population

- The fault population in the Strathspey-Brent-Statfjord area evolved during the rift event. Initially, a large number of short, low displacement fault segments were active but, during the 'rift climax' phase, displacement became concentrated on the >60 km long linked array of the Strathspey-Brent-Statfjord fault. As the number of active faults in the basin decreased, so the strain was localised on fewer faults and displacement rates increased. These observations support the conclusions of Cowie (1998) and Gupta *et al.* (1998).

- The geometry of the basin evolved with the growing fault system. Within 3-4 Ma of the initiation of rifting, a graben-like form composed of sub-parallel fault systems was established. The graben-like geometry was maintained until ~10 Ma into the rift event when the majority of faults became inactive, strain was concentrated on the Strathspey-Brent-Statfjord fault system and the basin subsequently developed a half-graben geometry.

- The half-graben bounding Strathspey-Brent-Statfjord fault is a single strand, throughgoing fault composed of a hierarchy of linked palaeo-segments. The locations of palaeo-segments can be identified by abandoned palaeo-tips, the syn-rift stratigraphic architecture and along strike variations in displacement on the linked array. It can be demonstrated that along strike displacement variations, established when the fault comprised a number of shorter unlinked segments, were not equilibrated following linkage.

Implications for understanding the evolution of fault populations in rift basins

Study of the fault population in the half-graben bounded by the Strathspey-Brent-Statfjord fault has emphasised that rift basins are dynamic; the distribution of active faults in the basin will evolve throughout the rift event. Hence, the structure of the basin at the cessation of extension is comparable with, but not equivalent to, the structure at the beginning of, or during, the rift event. As fault growth (accommodation generation) and sediment dynamics are intimately linked, the implication of this conclusion is that syn-rift sediment dispersal and deposition in the basin cannot be understood without reconstructing the growth of the fault population. This conclusion supports the results of numerous recent studies, for example, Schlische and Anders (1996), Gupta *et al.* (1998) and Dawers and Underhill (2000).

Through mapping of the fault population and examination of the stratal architecture of syn- and post-rift sediments, this work has demonstrated that the tectonic evolution of a basin can be reconstructed. Fault scaling relationships have been employed to illustrate that displacement minima observed on the displacement-length profiles of linked fault arrays define the locations of palaeo-segment boundaries. The linkage zones are frequently located coincident with palaeo- fault tips, abandoned in the hangingwall following hard linkage of two segments (*cf.* Dawers and Underhill 2000). The timing of linkage, and rates of fault growth, can be constrained through study of the stratigraphic architecture; depositional patterns record coeval fault growth. This reconstruction of the tectono-stratigraphic evolution of the half-graben bounded by the Strathspey-Brent-Statfjord fault was limited by the resolution of biostratigraphic dating, the distribution of well penetrations and the underfilled nature of the basin during the rift climax phase. However, these results demonstrate the potential of subsurface studies of ancient rifts to constrain fault growth rates on geological timescales and to test numerical models (e.g. Cowie 1998).

A distinct three-fold hierarchy of palaeo-segment lengths is recognised from interpretation of the Strathspey-Brent-Statfjord fault: 1-3 km, 7-12 km and 18-25 km length, each segment a linked array of shorter faults. Utilising observed fault scaling relationships (e.g. Schlische *et al.* 1996), this work demonstrates that along strike displacement variations established when the Strathspey-Brent-Statfjord fault was a number of shorter segments were not equilibrated following linkage. Rather, it is observed that displacement accumulates evenly along the strike of the linked array (Fig. 3.7). The implication of this observation is that, following breaching of the boundary to form a single fault strand, palaeo- segment boundaries are not the principal control on the continued accumulation of displacement on the throughgoing fault system, i.e. that breached segment boundaries are not a barrier to subsequent ruptures. This questions the proposal that the structural complexity of palaeo-segment boundaries means that they are a preferred site of rupture nucleation or arrest (e.g. Crone and Haller 1991, Janecke 1993). Rather this work suggests that, while the deformation around palaeo- linkage zones will impact upon the local stress regime and, hence, influence rupture location, other factors, for example fault interaction, may have an equally significant rôle to play.

Jackson and Blenkinsop (1997) have concluded from studies in the East African rift system that linked fault arrays of length in excess of 25 km can behave as a single fault. That interpretation is supported by the observations presented in this study. However, Jackson and Blenkinsop, and indeed other workers in modern continental rift provinces, have related maximum fault segment length and half-graben width to the thickness of the local seismogenic layer (e.g. Jackson and White 1989, Ebinger *et al.* 1999). Hence, the 100 km long segment described by Jackson and Blenkinsop in Malawi represents the anomalously thick upper crust in the region (~35 km). The current thickness of the thinned crust in the northern North Sea basin is 21-30 km (Holliger and Klemperer 1989), thus, following the relationship proposed by Ebinger *et al.* (1999) (their Fig. 7a), the corresponding maximum fault segment length would be 60-80 km. Fault segment lengths and half-graben widths within the northern North Sea rift province are, however, highly variable (compare for example the Strathspey-Brent-Statfjord and Zeta-Snorre-Gullfaks structures, Fig. 8.26). In addition, the pre-rift geologies, strain rates and heat flux of the northern North Sea rift system are unique – questioning the validity of comparisons with other rift provinces (particularly the very different Basin and Range, Greece and East African areas). It is suggested by this study that subsurface datasets in ancient rift basins, such as the northern North Sea, provide an opportunity to systematically examine and test the relationship between fault segment length, half-graben width and the seismogenic and elastic thickness of the continental crust.

9.3 Controls on syn-rift sedimentation in rift basins

Key conclusions from study of the syn-rift in the Strathspey-Brent-Statfjord area

- The source, dispersal patterns and loci of deposition of syn-rift sediments evolved throughout the extensional event. The migration of the loci of deposition (maximum accommodation generation) can be directly related to the growth of the fault population: Initially depocentres were small, numerous and widely distributed, however, as the rift event progressed, they widened, deepened and amalgamated until, during the climax of rifting, the entire half-graben acted as a single depocentre. Accommodation space generated as a consequence of differential compaction became increasingly important during the latter stages of the rift event (and was the primary control on post-rift deposition).

- The thickest deposits of the Tarbert Formation – the oldest sediments that may be confidently ascribed to the syn-rift – occur in isolated depocentres adjacent to short (<3 km), low displacement (<50 ms TWT) faults. The sand-rich marginal marine sediments are, however, preserved outwith the fault-controlled depocentres, indicative of a basin-wide accommodation generation in response to a relative rise in eustatic sea level. The rate of sediment supply exceeded the rate of tectonic subsidence.

- The Tarbert Formation is characterised by an overall transgressive trend, punctuated by minor progradational events. Contrary to the classical model for a transgressive systems tract (e.g. Postamentier *et al.* 1988), deposits of the Tarbert Formation are observed to thicken as the coast retreated. This is interpreted to be the response to an increase in sediment supply, as the shoreline migrated closer to the sediment source, and a narrowing of the facies belts (Fig. 4.16). As a consequence, coarse sediments were partitioned at the coastline, resulting in starvation of the proximal shallow marine shelf (Heather Formation) although rates of tectonic subsidence remained low.

- Increased rates of tectonic subsidence, a decline in sedimentation rates and deepening of water depths occurred in the study area during the mid-Oxfordian; coincident with the linkage of the palaeo-segments of the Strathspey-Brent-Statfjord fault to form a single throughgoing strand.

- During the climax of rifting, sediments were largely sourced from within the basin – from the water column (>75 %) or from degradation of the proximal footwall (<5 %). An external source of sediments, the products of crestal erosion of the footwall of the Zeta-Snorre-Gullfaks fault, supplied a Late Oxfordian – Kimmeridgian axial turbidite system and a shoreface environment in the Statfjord East area during the Volgian – Ryazanian. The rate of sediment supply can be directly related to the linkage history of the Zeta-Snorre-Gullfaks fault (J. Underhill *pers.comm.* 1999).

Implications for qualifying the controls on syn-rift sedimentation in rift basins

The stratigraphy of extensional basins commonly documents an early phase of basin filling (recorded by shallow marine or fluvial facies) succeeded by a period of sediment starvation (deep marine or lacustrine facies) (e.g. Lambiase 1990, Prosser 1993, Råvnas and Steel 1998). Syn-rift deposits described in the hangingwall to the Strathspey-Brent-Statfjord fault system conform to this stratigraphic pattern: Early marginal marine sand-rich sediments of the Tarbert Formation are overlain by the shallow-deep marine, fine-grained geologies of the Humber Group. As the rift event progressed, so the rates of sedimentation declined.

The transition in syn-rift depositional environments has been attributed to increased rates of tectonic subsidence, possibly invoked by strain localisation as the number of active faults in the basin decreases (Gupta *et al.* 1998). In the Strathspey-Brent-Statfjord area, the temporal and spatial distribution of syn-rift sediments is shown to document the generation of accommodation by tectonic subsidence. The thickest deposits of the Tarbert Formation are preserved in numerous, small fault controlled depocentres. The distribution of the Heather Formation defines wider and deeper sub-basins controlled by syn-sedimentary movement on 7-12 km long unlinked segments of the proto- Strathspey-Brent-Statfjord fault and segments of the antithetic array. In contrast to the Tarbert and Heather formations, the distribution of the Kimmeridge Clay Formation is less focused, thickening regionally into the fully linked Strathspey-Brent-Statfjord fault – the only active fault during deposition of the formation. The stratigraphic transition from Heather to Kimmeridge Clay formations occurs coincident with the concentration of strain onto the Strathspey-Brent-Statfjord fault; implying support for the mechanism proposed by Gupta *et al.* (1998).

The results of this study demonstrate that tectonic subsidence is the primary control on the generation of accommodation space, supporting the observations of previous workers (e.g. Leeder and Gawthorpe 1987, Contreras *et al.* 1997, Dawers and Underhill 2000). The rate of sediment supply was outpaced by the rate of tectonic subsidence for the majority of the rift phase. Interpretation of the stratal architecture of the Tarbert Formation, however, suggests that during the earliest syn-rift the rate of sediment supply exceeded the rate of tectonic subsidence. Indeed, as rates of tectonic subsidence increased, although the shoreline was gradually transgressed, there continued to be periods of shoreline progradation. This can be interpreted largely as a consequence of the partitioning of coarse sediments in a narrowing coastal zone, starving the shelf. Even so, (undecompacted) sedimentation rates for the Heather Formation were >30 mm/ka at a time when the estimated displacement rates on faults were <21 mm/ka; the rate of accommodation generation only slightly outpaced the rate of sediment supply. The implication of these observations is that the balance between accommodation generation (due to tectonic subsidence) and sediment supply is delicate during the rift initiation phase, resulting in a complex stratigraphic architecture. In the Strathspey-Brent-Statfjord area the rift initiation phase lasted ~10 Ma.

The supply of sediments to a rift basin from an external source depends upon modification of the pre-rift drainage pattern. Previous studies have demonstrated the rôle of transfer and antecedent drainage in controlling sediment dispersal and deposition (e.g. Schlische 1992, Jackson and Leeder 1994, Eliet and Gawthorpe 1995, Gilpin 1998, Gupta *et al.* 1999). Significantly, in the Strathspey-Brent-Statfjord area, the strike orientation of topographic highs, either footwall crests or fault growth monoclines, paralleled the pre-rift sediment transport direction. Hence, transverse drainage was unimportant; axial systems migrated in response to the differential accommodation generation and changing basin floor topography (*cf.* Leeder and Jackson 1993, Jackson and Leeder 1994). As a consequence, the evolution of fluvial systems, the shoreline and longshore currents during deposition of the Tarbert Formation was likely to have been complex and dynamic (see Davies *et al.* in review). In particular, Davies *et al.* describe stacked, multistorey fluvial channels, recording frequent avulsion events, in the proximal hangingwall to active faults (*cf.* Alexander and Leeder 1987, Peakall 1998).

The retreat of the Brent delta complex during the latest Bajocian – Bathonian effectively 'switched off' a voluminous external supply of sediment to the study area. Subsequently, the basin was starved, with respect to accommodation generated, and the dominant source of sediment was from the water column, by hemipelagic settling. Although the majority of the accommodation space from tectonic subsidence was generated during the rift climax phase, this accommodation was largely unfilled as rates of sedimentation were at a minimum. Hence, thickness variations in the syn-rift emphasise differential accommodation generation during the rift initiation, when sediment supply largely kept pace with subsidence. With a large volume of unfilled accommodation at the cessation of tectonic activity, the basin remained marine, the rate of sediment supply remained low and the half-graben took ~60 Ma to be filled.

9.4 The rôle of footwall denudation as a source of sediment in rift basins

Key conclusions from study of erosion of the footwall of the Strathspey-Brent-Statfjord fault, northern North Sea and the footwalls of four latest Jurassic – Early Cretaceous faults, central east Greenland

- The footwall of the Strathspey-Brent-Statfjord fault experienced differential uplift, reflecting along strike variations in fault displacement. As a consequence, pre-rift geologies were folded into low amplitude anticlines and synclines; the culmination of anticlines corresponds with topographic highs in the footwall and the locations of displacement maxima on the fault. Local tensile stresses were established in response to the steep displacement gradients at palaeo- fault segments. These stresses were accommodated by brittle strain, the growth of low displacement fault populations oriented perpendicular to the strike of bounding fault array.

- The area of the denuded scarp of the Strathspey-Brent-Statfjord fault is ~170 km^2 and the footwall crest has been backstepped by up to 2.5 km. The total volume of degraded sediments is estimated to be ~35 .2 km^3. The magnitude of denudation and the morphology of the slope demonstrate marked along strike continuity; the angle of the scarp is typically <20°. This is interpreted to be a consequence of the prolonged period of scarp exposure (>60 Ma) and the limitations of erosional processes in a submarine environment. These observations are characteristic of the large fault arrays throughout the northern North Sea rift province.

- A complex of sedimentologically degraded and structurally displaced sediments is recognised mounted on the footwall scarp of the Strathspey-Brent-Statfjord fault array. The fault scarp degradation complex has two components: Largely undeformed, stratigraphically intact sequences of Brent Group geologies of up to 200 m thickness, overlain by a thin veneer of chaotic, reworked debris, the deposits of debris flows. These geologies are interpreted as the geological remnants of the scarp denudation process; illustrating gravity driven slope failure by translational-rotational sliding and shallow sediment flows. This interpretation is supported the observation of debris flow deposits, comprising reworked pre-rift geologies, interbedded with hemipelagic syn-rift sediments in the proximal hangingwall.

- Observations and interpretations from study in the northern North Sea are supported by field studies of four latest Jurassic – Early Cretaceous faults of the central east Greenland rift. Structural cross sections are employed to demonstrate that the magnitude of footwall denudation of these large displacement normal faults is low, even although the structural highs were exposed for >20 Ma. The low magnitude of denudation is attributed to the submarine environment of the study area during the syn- and early post-rift. In addition, the products of footwall erosion are rarely observed mounted on the scarps of faults in central east Greenland, suggesting that the scarp is generally not a storage location for reworked sediments.

Implications for internally sourced sediment supply in rift basins

Study of the denuded footwalls of large normal fault arrays in the northern North Sea basin and the central east Greenland rift has shown that the magnitude of crestal degradation is low. Even although these structures were topographic (bathymetric) highs for up to 60 Ma with as much as 1 km of elevation over proximal hangingwall basins, the footwall crests have lost only 100-250 m of stratigraphy. This equates to an average rate of erosion of <3 mm/ka (*cf.* rates of >1 mm/yr in the northwestern Himalaya, Burbank *et al.* 1996). For the footwall of the Strathspey-Brent-Statfjord fault system the total volume of eroded pre-rift sediments is estimated at ~35 km^3 – an average erosion rate of <600 m^3/ka. Although these erosion rates are averaged over significant time periods and it is more likely that denudation was concentrated in the syn-rift period, the rates are still extremely low.

The low rate of denudation of these footwall highs is attributed to the submarine environment in which they were exposed. In a submarine setting the processes of downslope transport are limited; only gravity driven processes can occur and the slopes are generally not modified by channelised flow. Consequently, a submarine slope can achieve 'stability', i.e. the gravitational potential to drive downslope processes is insufficient to overcome the resisting forces. It is concluded that the establishment of a footwall high will be a consequence of the low rates of sediment supply in a submarine environment, but also that the submarine environment limits scarp degradation processes and, hence, inhibits the supply of sediments from footwall erosion.

Previous studies have cited footwall denudation as an important source of clastic sediment within rift basins (e.g. Surlyk 1989, Gawthorpe *et al.* 1994, Ravnås and Steel 1998). In particular, this process is considered significant during the climax of rifting when the basin has been flooded and external supplies of sediment no longer reach the interior of the rift system. The results of this study, however, demonstrate that footwall erosion liberates a limited volume of sediment at very low rates of supply. In addition, evidence from the geological remnants of slope failure (the fault scarp degradation complex) suggest that catastrophic failure by sliding was the dominant process of downslope transport in the northern North Sea. This implies denudation by geologically instantaneous events, probably seismically triggered, each supplying a large volume of sediment. The distribution of syn-rift deposits in the hangingwall illustrates that sediment supplied to the basin will subsequently be reworked and transported to fault controlled depocentres located at positions of maximum displacement on the basin bounding fault array.

The contrast in rates of erosion of subaerial and submarine terrain is illustrated by comparison of the footwalls bounded by the Strathspey-Brent-Statfjord and Zeta-Snorre-Gullfaks fault arrays (Fig. 8.26). While the denudation of the Strathspey-Brent-Statfjord structure is limited, the footwall of the Zeta-Snorre-Gullfaks fault system is deeply eroded and the volume of sediment liberated is estimated to be three to four times greater. The flat-topped profile of the Zeta-Snorre-Gullfaks footwall has been interpreted to reflect peneplanation to wavebase as the crest was uplifted above sea level during the late syn-rift. The reworked sedimentary products of erosion are recognised to have supplied an axial turbidite system active in the hangingwall to the Strathspey-Brent-Statfjord fault during the Kimmeridgian. The scale of the Zeta-Snorre-Gullfaks fault, in terms of maximum displacement and half-graben width, is approximately twice that of the Strathspey-Brent-Statfjord fault array, accounting for the greater uplift of the structure. Hence, it is suggested that some large faults have the potential to uplift their footwalls above sea level, and as a consequence liberate significant volumes of sediment, but the dispersal and character of the reworked debris will be a function of the structure of the footwall and the nature of the pre-rift geologies exposed at the crest.

9.5 Implications for hydrocarbon prospectivity

Partitioning of footwall reservoirs

The magnitude of displacement along the strike of normal fault systems is variable, illustrated by serrated displacement-length profiles. This variability reflects the growth of fault systems by the linkage of a hierarchy of shorter palaeo- fault segments; along strike variations in displacement developed during the propagation of the shorter segments are preserved following fault linkage. The consequence of along strike variations in the magnitude of displacement is differential uplift of the footwall and the development of topography along the crest of the structure. For the >60 km long Strathspey-Brent-Statfjord fault array, discrete highs and lows are described at the crest of the footwall defining low amplitude (>250 ms TWT) folds of 15-20 km wavelength. Footwall lows (the hinges of synclines) correspond with the locations of displacement minima on the bounding fault system and first-order palaeo- segment boundaries.

As a consequence of differential uplift, folding brittle pre-rift geologies, local stress regimes are established in the footwall. In particular, tensile stresses are concentrated around the steep displacement gradients at segment boundaries. In the footwall of the Strathspey-Brent-Statfjord fault, these stresses are accommodated by brittle failure and the growth of shallow normal fault populations striking perpendicular to the block bounding fault.

Pre-rift geologies tilted in the uplifted footwalls of large displacement normal faults are a primary exploration target in extensional basins. Optimal production from such reservoirs requires a good understanding of the structure of the footwall and deformation of pre-rift sediments. The observations presented in this thesis suggest that footwall traps can be viewed as anticlines, the wavelength and amplitude of the fold a function of the evolution of the bounding normal fault. The structural culmination of the footwall will correspond with the position of maximum displacement on the fault. In addition, the reservoir in the Strathspey-Brent-Statfjord area is shown to be partitioned by small displacement normal faults, accommodation structures developed in response to local tensile stresses established during differential uplift. In the study area, these stresses are accommodated by brittle failure, however, in a different geological regime pre-rift geologies may response differently. Previous workers in the northern North Sea basin have interpreted footwall cross fault systems as a reflection of basement inherited Triassic structuration or even as later Late Cretaceous – Cenozoic age structures.

Exploration for secondary hydrocarbon reservoirs

Rift basins have long been explored as provinces of hydrocarbon prospectivity and, as a consequence, many are now achieving maturity in terms of production. With the depletion of reserves in first-order, 'giant' structural footwall traps, the focus in such basins is to extend the production life, and utilise existing infrastructure, through exploration for secondary hydrocarbon reservoirs. While technological improvements contribute to this goal, with previously unviable targets now potentially economic, significant continued development is only really achievable through an improved, and predictive, understanding of the tectonic evolution of the rift basin and the controls on syn- and post-rift facies architecture. Two exploration targets can be identified from this work: The fault scarp degradation complex and syn-rift geologies in the proximal hangingwall.

THE FAULT SCARP DEGRADATION COMPLEX

In the northern North Sea rift province, the sedimentologically degraded and structurally dislocated geologies of the footwall mounted fault scarp degradation complexes are hydrocarbon bearing and, hence, important secondary reservoirs. Indeed, with the decline in reserves in footwall traps that have been in production for >25 years, some studies have estimated that up to one third of the remaining recoverable hydrocarbons may be contained within the complex (Schulte *et al.* 1994, Coutts *et al.* 1996). Although it has been recognised for some time that these sediments are the remnants of scarp degradation by translational-rotational sliding processes (e.g. Livera and Gdula 1990), effective production has been hampered by the poor understanding of the internal structure of the fault scarp degradation complex. This is largely due to the thinness of the deposits (below seismic resolution) and the masking effects of the strong base Cretaceous (top syn-rift) seismic reflector (see Hesthammer and Fossen 1999, McLeod and Underhill 1999). As a consequence, production results are extremely variable (e.g. Coutts *et al.* 1997, Underhill *et al.* 1997).

Through examination of core samples from the fault scarp degradation complex in the Strathspey-Brent-Statfjord area, this study has achieved an improved understanding of the internal structure of the complex. Although the dataset is limited (samples are widely spaced and <1.5 km total length), a number of key observations are reported. In particular, an overall lack of deformation within the slide blocks is emphasised and, hence, the reservoir properties of geologies incorporated into the complex are demonstrated to be comparable with *in situ* equivalents in the footwall. The macro-structure of the complex remains poorly resolved; it is suggested that deformation increases in intensity away from the west flank (as shown by Coutts *et al.* 1996). Micro-structuration within the complex is, however, shown to be lithology controlled and, hence, predictable. Low displacement faults both inhibit and enhance the productivity of reservoir geologies. Indeed, excellent production from the Tarbert Formation in well 33/9-C6 (exceeding that from the west flank) (A. Welbon *pers. comm.* 1998) can be attributed to enhanced fluid flow along dilational fault zones.

The implications of these observations and interpretations for exploiting footwall mounted degradation complexes in other rift basins are limited. Similar structures are not recorded from elsewhere in the literature and field studies in central east Greenland suggest that the fault scarp is typically not a storage location for syn-rift sediments. It is clear that the unique geological and environmental conditions will control the denudation processes and the distribution and deposition of reworked sediments. It can be concluded that the preservation of footwall highs for a geologically significant time will only occur in a submarine setting.

SYN-RIFT SEDIMENTS IN THE HANGINGWALL

Hydrocarbon exploration in rift basins is becoming increasingly focused on the search for secondary, subtle stratigraphic traps in proximal hangingwall syn-rift sediments (see Gabrielsen *et al.* 1995). This effort relies upon achieving a thorough understanding of the tectono-stratigraphic evolution of the target basin, specifically syn-rift sediment dispersal and depositional patterns. In the North Sea area, successful plays have been discovered in the hangingwall exploiting, for example, late syn-rift sand-rich sediments of a shoreface mantling an exposed crestal high (Dahl and Solli 1993, Nøttvedt *et al.* 2000), and turbidite and submarine fan systems (Cherry 1993, Ravnås and Steel 1997). These plays are largely sealed by background mud prone hemipelagic/pelagic deposits, which also act as the hydrocarbon source rocks.

Detailed study of the stratal architecture of syn-rift sediments in the half-graben basin bounded by the Strathspey-Brent-Statfjord fault array has permitted reconstruction of the depositional environments throughout the rift phase. This work has emphasised the rôle of fault growth in determining sediment source, dispersal and deposition. The fault population in the study area has a N-S trend, paralleling the orientation of Brent fluvial channels. Models of transfer zone drainage leading to the development of base of ramp fans (e.g. Underhill 1998) and the incision of antecedent drainage are thus not applicable in this area where there was no pre-existing transverse drainage network. The early flooding of the basin, with tectonic subsidence forcing the retreat of the shoreline, resulted in a decline in clastic sediment supply. During the climax of the rift event the majority of the deposits were sourced within the basin, bar an axial turbidite system derived from erosion of the footwall of the overlapping Zeta-Snorre-Gullfaks fault. Clastics compose <25 % of late syn-rift sediments and are not prospective. In contrast, sand-rich early syn-rift sediments of the Tarbert Formation were deposited in small, isolated synclinal depocentres.

Results from study of the half-graben bounded by the Strathspey-Brent-Statfjord fault demonstrate that the classic syn-rift plays models (e.g. relay ramp fans, axial turbidite systems) are not applicable in this area. This can largely be attributed to the location of the study half-graben in the interior of the northern North Sea rift province. Hence, clastic sediment was only supplied to the area during the earliest syn-rift and, during the climax of rifting, potential sources of sediment were limited and of low volume. The general observation of early flooding of rift basins suggests that sediment supply is a ubiquitous problem in terms of syn-rift play potential. It is concluded that the exploration for subtle stratigraphic traps in syn-rift sediments in the proximal hangingwalls to large normal faults must rely on the examination the unique evolution of each individual basin. In addition, the recognition of the potential sources of clastic sediment requires an understanding of the development of the region, and particularly the migration of sediment routeways as the structure of the basin evolves.

References

Alexander, J. and Leeder, M.R. 1987. Active tectonic control on alluvial architecture. *In:* Ethridge, F.G., Flores, R.M. and Harvey, M.D. (eds.), *Recent Developments in Fluvial Sedimentology,* SEPM Special Publication, **39**, 243-252.

Alexander, J., Bridge, J.S., Leeder, M.R., Collier, R.E.Ll. and Gawthorpe, R.L. 1994. Holocene meander-belt evolution in an active extensional basin, southwestern Montana. *Journal of Sedimentary Research,* **64,** 542-559.

Allen, P.A. 1997. *Earth surface processes.* Blackwell Science, London, 404pp.

Allen, P.A. and Hovius, N. 1998. Sediment supply from landslide-dominated catchments: implications for basin-margin fans. *Basin Research,* **10,** 19-35.

Anders, M.H. and Schlische, R.W. 1994. Overlapping faults, intrabasin highs and the growth of normal faults. *Journal of Geology,* **102,** 165-180.

Anhert, F. 1970. Functional relationships between denudation, relief and uplift in large mid-latitude drainage basins. *American Journal of Science,* **268,** 243-263.

Antonellini, M. and Aydin, A. 1994. Effect of faulting on fluid flow in porous sandstones: Petrophysical properties. *AAPG Bulletin,* **78**, 355-377.

Antonellini, M. and Aydin, A. 1995. Effect of faulting on fluid flow in porous sandstones: Geometry and spatial distribution. *AAPG Bulletin,* **79**, 642-671.

Antonellini, M., Aydin, A. and Pollard, D.D. 1994. Microstructure of deformation bands in porous sandstones at Arches National Park, Utah. *Journal of Structural Geology,* **16**, 941-959.

Antonellini, M., Aydin, A., Pollard, D.D. and D'Onfro, P. 1994. Petrophysical study of faults in sandstone using petrographic image analysis and x-ray computerised tomography. *Pageophysics,* **143**, 181-201.

Arber, M.A. 1973. Landslips near Lyme Regis. *Proceedings of the Geologists Association,* **84**, 121-133.

Armstrong, B. 1997. *The temporal and spatial evolution of clastic syn-tectonic sedimentation on and adjacent to a developing relay ramp: An example from the Suez Rift.* Unpublished M.Res. Thesis, University of Edinburgh, 182pp.

Aydin, A. 1978. Small cracks formed as deformation bands in sandstones. *Pageophysics,* **116**, 913-930.

Aydin, A. and Johnson, A.M. 1978. Development of faults as zones of deformation bands and as slip surfaces in sandstone. *Pageophysics,* **116**, 931-942.

Back, S., De Batist, M., Strecker, M.R. and Vanhauwaert, P. 1999. Quaternary depositional systems in Northern Lake Baikal, Siberia. *Journal of Geology,* **107,** 1-12.

Badley, M.E., Freeman, B., Roberts, A.M., Thatcher, J.S., Walsh, J., Watterson, J. and Yielding, G. 1990. Fault interpretation during seismic interpretation and reservoir evaluation. *In:* The integration of geology, geophysics and petroleum engineering in reservoir delineation, description and management. *Proceedings of the 1st Archie Conference, Houston, Texas,* 224-241.

Barnett, J.A.M., Mortimer, J., Rippon, J.H., Walsh, J.J. and Watterson, J. 1987. Displacement geometry in the volume containing a single normal fault. *AAPG Bulletin,* **71,** 925-937.

Bott, M.H.P. 1987. The continental margin of central East Greenland in relation to North Atlantic plate tectonic evolution. *Journal of the Geological Society, London,* **144,** 561-568.

Budding, M.C. and Inglin, H.F. 1981. A reservoir geological model of the Brent sands in Southern Cormorant. *In:*Illing, L.V. and Hobson, G.D. (eds.), *Petroleum geology of the continental shelf of northwest Europe.* London, Heyden, 326-334.

Burbank, D.W., Leland, J., Fielding, E., Anderson, R.S., Brozovic, N., Reid, M.R. and Duncan, C. 1996. Bedrock incision, rock uplift and threshold hillslopes in the northwestern Himalayas. *Nature,* **379,** 505-510.

Callomon, J.H. and Birkelund, T. 1980. The Jurassic transgression and the mid-late Jurassic succession in Milne Land, central East Greenland. *Geological Magazine,* **117,** 211-226.

Campbell, C.S., Cleary, P.W. and Hopkins, M. 1995. Large-scale landslide simulations: Global deformation, velocities and basal friction. *Journal of Geophysical Research,* **100,** 8267-8283.

Carter, K.E. and Winter, C.L. 1995. Fractal nature and scaling of normal faults in the Española Basin, Rio Grande rift, New Mexico: Implications for fault growth and brittle strain. *Journal of Structural Geology,* **17,** 863-873.

Cartwright, J.A., Trudgill, B.D. and Mansfield, C.M. 1995. Fault growth by segment linkage: an explanation for scatter in maximum displacement and trace length data for the Canyonlands grabens of S.E. Utah. *Journal of Structural Geology,* **17,** 1319-1326.

Cartwright, J.A., Mansfield, C. and Trudgill, B. 1996. The growth of normal faults by segment linkage. *In:* Buchanan, P.G. and Nieuwland, D.A. (eds.), *Modern developments in structural geology.* Special publication of the Geological Society, London, **99,** 163-177.

Childs, C., Watterson, J. and Walsh, J.J. 1995. Fault overlap zones within developing normal fault systems. *Journal of the Geological Society, London,* **152,** 535-549.

Christiansen, F.G., Olsen, H., Piasecki, S. and Stemmerik, L. 1990. Organic geochemistry of Upper Paleozoic lacustrine shales in the East Greenland basin. *Organic Geochemistry*, **16**, 287-294.

Clift, P.D., Carter, A. and Hurford, A.J. 1998. The erosional and uplift history of NE Atlantic passive margins: constraints on a passing plume. *Journal of the Geological Society, London,* **155,** 787-800.

Cockings, J.H., Kessler, L.G., Mazza, T.A. and Riley, L.A. 1992. Bathonian to mid-Oxfordian sequence stratigraphy of the South Viking Graben, North Sea. *In:* Hardman, R.F.P. (ed.), *Exploration Britain: Geological insights for the next decade.* Special publication of the Geological Society, London, **67,** 65-105.

Coe, A.L. 1992. *Unconformities within the Upper Jurassic of the Wessex Basin, Southern England.* Unpublished Ph.D. Thesis, University of Oxford.

Coe, A.L. 1995. A comparison of the Oxfordian successions of Dorset, Oxfordshire and Yorkshire. *In:* Taylor, P.D. (ed.), *Field geology of the British Jurassic.* Geological Society, London, 151-172.

Colletta, B., Le Quellec, P., Letouzy, J. and Moretti, I. 1988. Longitudinal evolution of the Suez rift structure (Egypt). *Tectonophysics,* **153,** 221-233.

Contreras, J., Scholz, C. and King, G.C.P. 1997. A model of rift basin evolution constrained by first-order stratigraphic observations. *Journal of Geophysical Research,* **102,** 7673-7690.

Coutts, S.D., Larsson, S.Y. and Rosman, R. 1996. Development of the slumped crestal area of the Brent Reservoir, Brent Field: an integrated approach. *Petroleum Geoscience*, **2**, 219-229.

Coutts, S.D., Jurgens, M., van Kessel, O., Pronk, D. and Ward, V.C. 1997. Phase 2 development of the slumped crestal area of the Brent reservoir, Brent Field. *Society of Petroleum Engineers,* **38476,** 135-144.

Cowie, P.A. and Scholz, C.H. 1992a. Displacement length scaling relationship for faults: Data synthesis and discussion. *Journal of Structural Geology,* **14,** 1149-1156.

Cowie, P.A. and Scholz, C.H. 1992b. Physical explanation for displacement-length scaling relationship of faults using a post-yield fracture mechanics model. *Journal of Structural Geology,* **14,** 1133-1148.

Cowie, P.A., Vanneste, C. and Sornette, D. 1993. Statistical physics model for the spatio-temporal evolution of faults. *Journal of Geophysical Research,* **97**, 17911-17920.

Cowie, P.A. 1998. A healing-reloading feedback control on the growth rate of seismogenic faults. *Journal of Structural Geology,* **20,** 1075-1087.

Crone, A.J. and Haller, K.M. 1991. Segmentation and coseismic behaviour of Basin and Range normal faults: Examples from east-central Idaho and southwestern Montana, USA. *Journal of Structural Geology,* **13**, 151-164.

Dahl, N. and Solli, T. 1993. The structural evolution of the Snorre Field and surrounding areas. *In:* Parker, J.R. (ed.), *Petroleum Geology of Northwest Europe: Proceedings of the 4th Conference.* The Geological Society, London, 1159-1166.

Davies, S.J., Dawers, N.H., McLeod, A.E. and Underhill, J.R. in review. Controls on the spatial and temporal evolution of early syn-rift deposition: The Middle Jurassic Tarbert Formation, northern North Sea. Submitted *Basin Research*, April 2000.

Dawers, N.H., Anders, M.H. and Scholz, C.H. 1993. Fault length and displacement: Scaling laws. *Geology,* **21,** 1107-1110.

Dawers, N.H. and Anders, M. 1995. Displacement-length scaling and fault linkage. *Journal of Structural Geology,* **17,** 604-614.

Dawers, N.H., Berge, A.M., Häger, K.-O., Puigdefabregas, C. and Underhill, J.R. 1999. Controls on Late Jurassic, subtle sand distribution in the Tampen area, Northern North Sea. *In:* Fleet, A.J. and Boldy, S.A.R. (eds.), *Petroleum geology of northwest Europe: Proceedings of the* 5^{th} *conference.* Geological Society, London, 827-838.

Dawers, N.H. and Underhill, J.R. 2000. The role of fault interaction and linkage in controlling syn-rift stratigraphic sequences: Statfjord East area, northern North Sea. *AAPG Bulletin*, **84**, 45-64.

Densmore, A.L., Anderson, R.S., McAdoo, B.G. and Ellis, M.A. 1997. Hillslope evolution by bedrock landslides. *Science,* **275,** 369-372.

Densmore, A.L., Ellis, M.A. and Anderson, R.L. 1998. Landsliding and the evolution of normal fault-bounded mountains. *Journal of Geophysical Research,* **103**, 15203-15219.

Dingle, R.V. 1980. Large allochthonous sediment masses and their role in the construction of the continental slope and rise of southwestern Africa. *Marine Geology,* **37,** 333-354.

Donovan, D.T. 1955. The stratigraphy of the Jurassic and Cretaceous rocks of Geographical Society Ø, East Greenland. *Meddelelser om Grønland,* **111,** 150pp.

Doré, A.G. 1991. The stuctural foundation and evolution of Mesozoic seaways between Europe and the Arctic. *Palaeogeography, Palaeoclimatology, Palaeoecology,* **87**, 441-492.

Dou, L., Turner, J.D. and Underhill, J.R. in review. Evidence from subsidence analysis for the timing, location and significance of fault activity in rift systems: the Middle to Late Jurassic extensional history of the Beryl Embayment, UK North Sea. Submitted *AAPG Bulletin,* November 1999.

Ebinger, C.J., Jackson, J.A., Foster, A.N. and Hayward, N.J. 1999. Extensional basin geometry and the elastic lithosphere. *Philosophical transactions of the Royal Society of London A,* **357,** 741-765.

Eliet, P.P. and Gawthorpe, R.L. 1995. Drainage development and sediment supply within rifts, examples from the Sperchios basin, central Greece. *Journal of the Geological Society, London,* **152,** 883-893.

Elliott, D. 1976. Energy balance and deformation mechanisms of thrust sheets. *Philosophical transactions of the Royal Society of London A,* **283,** 289-312.

Eyles, C.H. and Lagoe, M.B. 1998. Slump-generated megachannels in the Pliocene-Pleistocene glaciomarine Yakataga Formation, Gulf of Alaska. *GSA Bulletin,* **110,** 395-408.

Færseth, R.B., Knudsen, B.-E., Liljedahl, T., Midboe, P.S. and Soderstorm, B. 1997. Oblique rifting and sequential faulting in the Jurassic development of the northern North Sea. *Journal of Structural Geology,* **19,** 1285-1302.

Fält, L.M., Helland, R., Jacobsen, V.W. and Renshaw, D. 1989. Correlation of transgressive-regressive depositional sequences in the Middle Jurassic Brent Vestland Group megacycle, Viking Graben, Norwegian North Sea. *In:* Collinson, J.D. (ed.), *Correlation in hydrocarbon exploration.* Norwegian Petroleum Society (NPF), Graham and Trotman, London, 191-200.

Ferentinos, G., Papatheodorou, G. and Collins, M.B. 1988. Sediment transport processes on an active submarine fault escarpment: Gulf of Corinth, Greece. *Marine Geology,* **83,** 43-61.

Fossen, H. and Gabrielsen, R.H. 1996. Experimental modelling of extensional fault systems by use of plaster. *Journal of Structural Geology,* **18,** 673-687.

Fossen, H. and Hesthammer, J. 1998. Structural geology of the Gullfaks Field, northern North Sea. *In:* Coward, M.P., Johnson, H. and Daltaban, T.S. (eds.), *Structural geology in reservoir characterisation.* Special publication of the Geological Society, London, **127,** 231-261.

Fowles, J. and Burley, S. 1994. Textural and permeability characteristics of faulted, high porosity sandstones. *Marine and Petroleum Geology,* **11**, 608-623.

Gabrielsen, R.H., Steel, R.J. and Nøttvedt, A. 1995. Subtle traps in extensional terranes: a model with reference to the North Sea. *Petroleum Geoscience,* **1**, 223-235.

Galloway, W.E. 1989. Genetic sequence stratigraphic sequences in basin analysis 1: Architecture and genesis of flooding-surface bounded depositional units. *AAPG Bulletin,* **73,** 125-142.

Gawthorpe, R.L. and Colella, A. 1990. Tectonic controls on coarse-grained delta depositional systems in rift basins. *In:* Collela, A. and Prior, D.B. (eds.), *Coarse-grained deltas.* International association of sedimentologists special publication **10,** 113-127.

Gawthorpe, R.L. and Hurst, J.M. 1993. Transfer zones in extensional basins: their structural style and influence on drainage basin and stratigraphy. *Journal of the Geological Society, London,* **150**, 1137-1152.

Gawthorpe, R.L., Fraser, A.J. and Collier, R.E.Ll. 1994. Sequence stratigraphy in active extensional basins: Implications for the interpretation of ancient basin fills. *Marine and Petroleum Geology,* **11,** 642-658.

Gawthorpe, R.L., Sharp, I., Underhill, J.R. and Gupta, S. 1997. Linked sequence stratigraphic and structural evolution of propagating normal faults. *Geology,* **25,** 795-798.

Gillespie, P.A., Walsh, J.J. and Watterson, J. 1992. Limitations of dimension and displacement data from single faults and the consequences for data analysis and interpretation. *Journal of Structural Geology,* **14,** 1157-1172.

Gilpin, R.E. 1998. *Controls on structural geometries and sediment dispersal in dissected relay ramps: Evolution of the Baba fault, eastern Gulf of Suez.* Unpublished M.Res. Thesis, University of Edinburgh, 209pp.

Gould, H.R. 1970. The Mississippi delta compex. *In:* Morgan, J.P. and Shaver, R.H. (eds.), *Deltaic sedimentation modern and ancient.* SEPM Special Publication **15,** 3-30.

Graham, J. 1984. Methods of stability analysis. *In:* Brunsden, D. and Prior, D.B. (eds.), *Slope instability.* John Wiley and Sons, 171-215.

Graue, E., Helland-Hansen, W., Johnsen, J.R., Lømo, L., Nøttvedt, A., Ronning, K., Ryseth, A. and Steel, R.J. 1987. Advance and retreat of the Brent Delta system. *In:* Brooks, J. and Glennie, K. (eds.), *Petroleum Geology of North West Europe.* Graham and Trotman, London, 915-937.

Gupta, S., Cowie, P.A., Dawers, N.H. and Underhill, J.R. 1998. A mechanism to explain rift basin subsidence and stratigraphic patterns through fault array evolution. *Geology,* **26,** 595-598.

Gupta, S., Underhill, J.R., Sharp, I.R. and Gawthorpe, R.L. 1999. Role of fault interactions in controlling syn-rift sediment dispersal patterns: Miocene, Abu Alaqa Group, Suez Rift, Sinai, Egypt. *Basin Research,* **11,** 167-189.

Hansen, K. 1992. Post-orogenic tectonic and thermal history of a rifted continental margin: the Scoresby Sund area, East Greenland. *Tectonophysics,* **216**, 309-326.

Haq, B.U., Hardenbol, J. and Vail, P.R. 1988. Mesozoic and Cenozoic chronostratigraphy and eustatic cycles. *In:* Wilgus, C.K., Hastings, B.S., Ross, C.A., Posamentier, H.W., Van Wagoner, J. and Kendall, C.G.St.C. (eds.), *Sea-level changes: An integrated approach.* SEPM Special Publication **42,** 71-108.

Harbor, D.J. 1997. Landscape evolution at the margin of the Basin and Range. *Geology,* **25,** 1111-1114.

Harland, W.B., Armstrong, R.L., Cox, A.V., Craig, L.E., Smith, A.G. and Smith, D.G. 1990. *A Geologic Timescale.* Cambridge University Press, Cambridge.

Hayward, N. and Ebinger, C. 1996. Variations in the along-axis segmentation of the Afar Rift system. *Tectonics,* **15,** 244-257.

Helland-Hansen, W., Ashton, M., Lømo, L. and Steel, R.J. 1992. Advance and retreat of the Brent delta: recent contributions to the depositional model. *In:* Morton, A.C., Haszeldine, R.S., Giles, M.R. and Brown, S. (eds.), *Geology of the Brent Group.* Special publication of the Geological Society, London, **61,** 109-127.

Hesthammer, J. and Fossen, H. 1999. Evolution and geometries of gravitational collapse structures with examples from the Statfjord Field, northern North Sea. *Marine and Petroleum Geology,* **16**, 259-281.

Hippler, S.J. 1993. Deformation microstructures and diagenesis in sandstone adjacent to an extensional fault: Implications for the flow and emplacement of hydrocarbons. *AAPG Bulletin,* **77**, 625-637.

Holliger, K. and Klemperer, S.L. 1989. A comparison of the Moho interpreted from gravity data and from deep seismic reflection data in the northern North Sea. *Geophysical Journal,* **97**, 247-258.

Hovius, N., Stark, C.P. and Allen, P.A. 1997. Sediment flux from a mountain belt derived by landslide mapping. *Geology,* **25,** 231-234.

Hovius, N. and Leeder, M. 1998. Clastic sediment supply to basins. *Basin Research,* **10,** 1-5.

Hovius, N., Stark, C.P., Tutton, M.A. and Abbott, L.D. 1998. Landslide-driven drainage network evolution in a pre-steady state mountain belt: Finisterre Mountains, Papau New Guinea. *Geology,* **26,** 1071-1074.

Huggins, P., Watterson, J., Walsh, J.J. and Childs, C. 1995. Relay zone geometry and displacement transfer between normal faults recorded in coal-mine plans. *Journal of Structural Geology,* **17,** 1741-1755.

Jackson, J. and McKenzie, D. 1983. The geometrical evolution of normal fault systems. *Journal of Structural Geology,* **5,** 471-482.

Jackson, J.A., White, N.J., Garfunkel, Z. and Anderson, H. 1988. Relations between normal fault geometry, tilting and vertical motions in extensional terrains, an example from the southern Gulf of Suez. *Journal of Structural Geology,* **10,** 155-170.

Jackson, J. and White, N.J. 1989. Normal faults in the upper continental crust: observations from regions of active extension. *Journal of Structural Geology,* **10,** 155-170.

Jackson, J. and Leeder, M. 1994. Drainage systems and the development of normal faults: an example from Pleasant Valley, Nevada. *Journal of Structural Geology,* **16**, 1041-1059.

Jackson, J. and Blenkinsop, T. 1997. The Bilila-Mtakataka fault in Malawi: An active, 100-km long, normal fault segment in thick seismogenic crust. *Tectonics,* **16,** 137-150.

Janecke, S.U. 1993. Structures in segment boundary zones of the Lost River and Lemhi Faults, east central Idaho. *Journal of Geophysical Research,* **98**, 16223-16238.

Jibson, R.W. and Keefer, D.K. 1993. Analysis of the seismic origin of landslides: Examples from the New Madrid seismic zone. *GSA Bulletin,* **105,** 521-536.

Johannessen, F.P., Mjøs, R., Renshaw, D., Dallard, A. and Jacobsen, T. 1995. Northern limit of the 'Brent delta' at the Tampen Spur: A sequence stratigraphic approach for sandstone prediction. *In:* Steel, R.J. Felt, V., Johannessen, E.P. and Mathieu, C. (eds.), *Sequence stratigraphy on the northwest European margin.* Norwegian Petroleum Society (NPF) Special Publication **5,** 213-256.

Johnson, A. and Eyssautier, M. 1987. Alwyn North Field and its regional geological context. *In*: Brooks, J. and Glennie, K. (eds.), *Petroleum Geology of North West Europe*. Graham & Trotman, London, 963-977.

Keefer, D.K. 1984. Landslides caused by earthquakes. *Geological Society of America Bulletin,* **95,** 406-421.

Keefer, D.K. 1999. Earthquake-induced landslides and their effects on alluvial fans. *Journal of sedimentary research,* **69**, 81-104.

Kelly, S.R.A., Whitham, A.G., Koraini, A.M. and Price, S.P. 1998. Lithostratigraphy of the Cretaceous (Barremian-Santonian) Hold With Hope Group, NE Greenland. *Journal of the Geological Society, London,* **155,** 993-1008.

King, G.C.P., Stein, R.S. and Lin, J. 1994. Static stress changes and the triggering of earthquakes. *Bulletin of the Seismological Society of America,* **84,** 935-953.

Koch, L. and Haller, J. 1971. Geological map of East Greenland 72°N-76°N lat (1:250,000). *Meddelelser om Grønland,* **183,** 26pp.

Krantz, R.W. 1988. Multiple fault sets and three-dimensional strain. *Journal of Structural Geology,* **10,** 225-237.

Lambaise, J.J. 1990. A model for tectonic control of lacustrine stratigraphic sequences in continental rift basins. *In:* Katz, B.J. (ed.), *Lacustrine basin exploration: case studies and modern analogs.* AAPG Memoir **50,** 265-276.

Lambaise, J.J. and Bosworth, W. 1995. Structural controls on sedimentation in continental rifts. *In:* Lambaise, J.J. (ed.), *Hydrocarbon habitats in rifts.* Special publication of the Geological Society, London, **80**, 117-159.

Larsen, L.M., Watt, W.S. and Watt, M. 1989. Geology and petrology of the Lower Teriary plateau basalts of the Scoresby Sund region, East Greenland. *Bulletin Grønlands Geologiske Undersøgelse,* **157,** 162pp.

Larsen, P.-H. 1988. Relay structures in a Lower Permian basement-involved extension system, East Greenland. *Journal of Structural Geology,* **10,** 3-8.

Lee, M.J. and Hwang, Y.J. 1993. Tectonics and structural styles of the East Shetland Basin. *In:* Parker, J.R. (ed.), *Petroleum Geology of Northwest Europe: Proceedings of the 4th Conference,* Geological Society Publishing House, 1137-1149.

Leeder, M.R. 1993. Tectonic controls upon drainage basin development, river channel migration and alluvial architecture: implications for hydrocarbon reservoir development and characterization. *In:* North, C.P. and Prosser, D.J. (eds.), *Characterization of Flvial and Aeolian Reservoirs.* Special publication of the Geological Society, London, **73**, 7-22.

Leeder, M.R. and Gawthorpe, R.L. 1987. Sedimentary models for extensional tilt-block/half-graben basins. *In:* Coward, M.P., Dewey, J.F. and Hancock, P.L. (eds.), *Continental extensional tectonics.* Special publication of the Geological Society, London, **28,** 139-152.

Leeder, M.R. and Jackson, J. 1993. The interaction between normal faulting and drainage in active extensional basins, with examples from the western United States and central Greece. *Journal of the Geological Society, London,* **148,** 331-343.

Leeder, M.R., Harris, T. and Kirkby, M.J. 1998. Sediment supply and climate change: Implications for basin stratigraphy. *Basin Research,* **10,** 7-18.

Livera, S.E. and Gdula, J.E. 1990. Brent Oil Field. *In*: Beaumont, E.A. & Foster, N.H. (eds.), *Structural Traps II, Traps associated with tectonic faulting.* Atlas of Oil and Gas Fields, American Association of Petroleum Geologists Memoir, 21-63.

Lockner, D.A., Byerlee, J.D., Kuksenko, V., Ponomarev, A. and Sidorin, A. 1992. Observations of quasistatic fault growth from acoustic emissions. *In:* Evans, B. and Wong, T.-F. (eds.), *Fault mechanics and transport properties of rocks.* Academic press, San Fransisco, 1-31.

Lowe, D.R. 1979. Sediment gravity flows: their classification and some problems of application to natural flows and deposits. *In:* Doyle, L.J. and Pilkey, O.H. (eds.), *Geology of Continental Slopes,* SEPM Special Publication, **27**, 75-82.

Marjanac, T. 1995. Architecture and sequence stratigraphic perspectives of the Dunlin Group formations and proposal for new type- and reference-wells. *In:* Steel, R.J., Felt, V., Johannessen, E.P. and Mathieu, C. (eds.), *Sequence stratigraphy on the northwest European margin.* Norwegian Petroleum Society (NPF) Special Publication **5,** 143-165.

Marrett, R. and Allmendinger, R.W. 1991. Estimates of strain due to brittle faulting: Sampling of fault populations. *Journal of Structural Geology,* **13,** 735-738.

Maync, W. 1947. Stratigraphie der Jurabildungen Ost grönland zwischen Hochstetterbugten (75°N) und dem Kejser Franz Joseph Fjord (73°N). *Meddelelser om Grønland,* **132(2),** 223pp.

Maync, W. 1949. The Cretaceous beds between Kuhn Island and Cape Franklin (Gauss Peninsula), northern East Greenland. *Meddelelser om Grønland,* **133(3),** 291pp.

McGrath, A. 1992. *Fault propagation and growth: a study of the Triassic and Jurassic from Watchet and Klive, North Somerset.* Unpublished M.Sc. Thesis, Royal Holloway, London, 165pp.

McKenzie, D.P. 1978. Some remarks on the development of sedimentary basins. *Earth and Planetary Science Letters,* **40,** 25-32.

McKenzie, D.P. and Jackson, J. 1986. A block model of distributed deformation by faulting. *Journal of the Geological Society, London*, **143**, 349-354.

McLeod, A.E. 1996. *The temporal and spatial evolution of a zone of Middle - Late Jurassic fault scarp degradation, North Brent Field, UK sector northern North Sea.* Unpublished M.Res. Thesis, University of Edinburgh, 120pp.

McLeod, A.E. and Underhill, J.R. 1999. Processes and products of footwall degradation, northern Brent Field, northern North Sea. *In:* Fleet, A.J. and Boldy, S.A.R. (eds.), *Petroleum geology of northwest Europe: Proceedings of the 5th conference.* Geological Society, London, 91-106.

McMillan, R.A. 1975. *The orientation and sense of displacement of strike-slip faults in continental crust.* Unpublished Bachelor Thesis, Carleton University, Ottawa.

Menendez, B., Zhu, Wenlu and Wong, T.-F. 1996. Micromechanics of brittle faulting and cataclastic flow in Berea sandstone. *Journal of Structural Geology,* **18**, 1-16.

Milliman, J.D. and Syvitski, J.P.M. 1992. Geomorphic/tectonic control of sediment discharge to the ocean: the importance of small mountainous rivers. *Journal of Geology,* **100,** 525-544.

Mitchener, B.C., Lawrence, D.A., Partington, M.A., Bowman, M.B.J. and Gluyas, J. 1992. Brent Group: Sequence stratigraphy and regional implications. *In:* Morton, A.C., Haszeldine, R.S., Giles, M.R. and Brown, S. (eds.), *Geology of the Brent Group.* Special publication of the Geological Society, London, **61,** 45-80.

Mitra, S. 1993. Geometry and kinematic evolution of inversion structures. *AAPG Bulletin,* **77**, 1159-1191.

Mohrig, D., Whipple, K.X., Hondzo, M., Ellis, C. and Parker, G. 1998. Hydroplaning of subaqueous debris flows. *GSA Bulletin,* **110**, 387-394.

Mollema, P.N. and Antonellini, M.A. 1996. Compaction bands: A structural analog for anti-mode I cracks in aeolian sandstone. *Tectonophysics,* **267**, 209-228.

Morgan, P. and Baker, B.H. 1983. Introduction – processes of continental rifting. *Tectonophysics,* **94,** 1-10.

Morley, C.K. 1999. Patterns of displacement along large normal faults: Implications for basin evolution and fault propagation, based on examples from East Africa. *AAPG Bulletin,* **83,** 613-634.

Muraoka, H. and Kamata, H. 1983. Displacement distribution along minor fault traces. *Journal of Structural Geology,* **5,** 483-495.

Nadin, P.A. and Kusznir, N.J. 1995. Palaeocene uplift and Eocene subsidence in the northern North Sea Basin from 2D forward and reverse stratigraphic modelling. *Journal of the Geological Society, London,* **152,** 833-848.

Neimi, T.M. and Ben-Avraham, Z. 1994. Evidence for Jericho earthquakes from slumped sediments of the Jordan River delta in the Dead Sea. *Geology,* **22,** 395-398.

Nelson, C.H., Karabanov, E.B., Colman, S.M. and Escutia, C. 1999. Tectonic and sediment supply control of deep rift lake turbidite systems: Lake Baikal, Russia. *Geology,* **27,** 163-166.

Nemec, W. 1990. Aspects of sediment movement on steep delta slopes. *In:* Colella, A. and Prior, D.B. (eds.) *Coarse Grained Deltas.* Special publication of the international association of sedimentologists, **10**, 29-73.

Nemec, W., Steel, R.J., Gjelberg, J., Collinson, J.D., Prestholm, E. and Øxnevad, I.E. 1988. Anatomy of a collapsed and re-established delta front in Lower Cretaceous of eastern Spitsbergen: gravitational sliding and sedimentation processes. *AAPG Bulletin,* **72**, 454-476.

Newmark, N.M. 1965. Effects of earthquakes on dams and embankments. *Geotechnique,* **15**, 139-160.

Nicol, A., Walsh, J.J., Watterson, J. and Underhill, J.R. 1997. Displacement rates of normal faults. *Nature,* **390,** 157-159.

Nielsen, T., Van Weering, TJ.C.E. and Andersen, M.S. 1998. Cenozoic changes in the sedimentary regime on the northeastern Faeroes margin. *In:* Stoker, M.S., Evans, D. and Cramp, A. (eds.), *Geological processes on continental margins: Sedimentation, mass-wasting and stability.* Special publication of the Geological Society, London, **129,** 167-171.

Nøhr-Hansen, H. 1993. Dinoflagellate cyst stratigraphy of the Barremian to Albian, Lower Cretaceous, North-East Greenland. *Bulletin Grønlands Geologiske Undersøgelse,* **166,** 171pp.

Nøttvedt, A., Berge, A., Dawers, N.H., Færseth, R.B., Häger, K.-O., Mangerud, G. and Puigdefabregas, C. 2000. Syn-rift evolution and resulting play models in the Snorre-H area, northern

North Sea. *In:* Nøttvedt, A., *et al.* (eds.), *Dynamics of the Norwegian Margin*. Special Publication of the Geological Society, London.

Ophiem, J.A. and Gudmundsson, A. 1989. Formation and geometry of fractures, and related volcanism, of the Krafla fissure swarm, northeast Iceland. *Bulletin of the Geological Society of America,* **101,** 1608-1622.

Parkinson, D.N. and Hines, F.M. 1995. The Lower Jurassic of the North Viking Graben in the context of Western European Lower Jurassic stratigraphy. *In:* Steel, R.J., Felt, V., Johannessen, E.P. and Mathieu, C. (eds.), *Sequence stratigraphy on the northwest European margin.* Norwegian Petroleum Society (NPF) Special Publication **5,** 97-107.

Patton, T.L., Moustafa, A.R., Nelson, R.A. and Abdine, S.A. 1994. Tectonic evolution and structural setting of the Suez Rift. *In:* Landon, S.M. (ed.), *Interior rift basins.* AAPG Memoir **59,** 9-55.

Peacock, D.C.P. 1991. Displacements and segment linkage in strike-slip fault zones. *Journal of Structural Geology,* **13,** 1025-1035.

Peacock, D.C.P. and Sanderson, D.J. 1991. Displacements, segment linkage and relay ramps in normal fault zones. *Journal of Structural Geology,* **13,** 721-733.

Peacock, D.C.P. and Sanderson, D.J. 1994. Geometry and development of relay ramps in normal fault systems. *AAPG Bulletin,* **78,** 147-165.

Peakall, J. 1998. Axial river evolution in response to half-graben faulting: Carson River, Nevada, U.S.A. *Journal of Sedimentary Research,* **68,** 788-799.

Pearce, A.J. and Watson, A.J. 1986. Effects of earthquake-induced landslides on sediment budget and transport over a 50-yr period. *Geology,* **14**, 52-55.

Pickering, G., Peacock, D.C.P., Sanderson, D.J. and Bull, J.M. 1997. Modeling tip zones to predict the throw and length characteristics of faults. *AAPG Bulletin,* **81**, 82-99.

Pickering, K.T., Hiscott, R.N. and Hein, F.J. 1989. Deep marine environments: clastic sedimentation and tectonics. Unwin Hyman, London, 416pp.

Pittman, E.D. 1981. Effect of fault-related granulation on porosity and permeability of quartz sandstones, Simpson Group (Ordovician), Oklahoma. *AAPG Bulletin,* **65**, 2381-2387.

Posamentier, H.W., Jervey, M.T. and Vail, P.R. 1988. Eustatic controls on clastic deposition 1 – conceptual framework. *In:* Wilgus, C.K., Hastings, B.S., Ross, C.A., Posamentier, H.W., Van Wagoner, J. and Kendall, C.G.St.C. (eds.), *Sea-level changes: An integrated approach.* SEPM Special Publication **42,** 109-124.

Price, S.P. and Whitham, A.G. 1994. *Mesozoic-Tertiary geology of the Traill Ø region, East Greenland: an overview.* Unpublished East Greenland Report (CASP) **31.**

Price, S.P. and Whitham, A.G. 1997. Exhumed hydrocarbon traps in East Greenland: Analogues for the Lower-Middle Jurassic play of Northwest Europe. *AAPG Bulletin,* **81,** 196-221.

Price, S.P., Brodie, J., Whitham, A. and Kent, R. 1997. Mid-Tertiary rifting and magmatism in the Traill Ø region, East Greenland. *Journal of the Geological Society, London,* **154,** 419-434.

Prior, D.B. and Coleman, J.M. 1978. Disintegrating retrogressive landslides on very low-angle subaqueous slopes, Mississippi delta. *Marine Geotechniques,* **3,** 37-60.

Prior, D.B. and Coleman, J.M. 1979. Submarine landslides – geometry and nomenclature. *Zeitschrift für Geomorphologie,* **23,** 415-426.

Prosser, S. 1993. Rift-related linked depositional systems and their seismic expression. *In:* Williams, G.D. and Dobb, A. (eds.), *Tectonics and seismic sequence stratigraphy.* Special publication of the Geological Society, London, **71,** 35-66.

Rattey, R.P. and Hayward, A.B. 1993. Sequence stratigraphy of a failed rift system: The Middle Jurassic to Early Cretaceous of the Central and Northern North Sea. *In:* Parker, J.R. (ed.), *Petroleum Geology of Northwest Europe: Proceedings of the 4th Conference.* The Geological Society, London, 215-249.

Ravnås, R. and Steel, R.J. 1997. Contrasting styles of Late Jurassic syn-rift turbidite sedimentation: a comparative study of the Magnus and Oseberg areas, northern North Sea. *Marine and Petroleum Geology,* **14,** 417-449.

Ravnås, R. and Steel, R.J. 1998. Architecture of marine rift-basin successions. *AAPG Bulletin,* **82,** 110-146.

Ravnås, R., Bondevik, K., Helland-Hansen, W., Lømo, L., Ryseth, A. and Steel, R.J. 1997. Sedimentation history as an indicator of rift initiation and development: The Late Bajocian - Bathonian evolution of the Oseberg-Brage area, northern North Sea. *Norsk Geologisk Tidsskrift,* **77,** 205-232.

Richards, P.C., Lott, G.K., Johnson, H., Knox, R.W.O'B. and Riding, J.B. 1993. Jurassic of the central and northern North Sea. *In:* Knox, R.W.O'B. and Cordey, W.G. (eds.), *Lithostratigraphic nomenclature of the UK North Sea.* British Geological Society, Nottingham, 219.

Rippon, J.H. 1985. Contoured patterns of the throw and hade of normal faults in the Coal Measures (Westphalian) of north-east Derbyshire. *Proceedings of the Yorkshire Geological Society,* **45,** 147-161.

Roberts, A.M., Yielding, G., Kusznir, N.J., Walker, I.M. and Dorn-Lopez, D. 1993. Mesozoic extension in the North Sea: Constraints from flexural backstripping, forward modelling and fault populations. *In:* Parker, J.R. (ed.), *Petroleum Geology of Northwest Europe: Proceedings of the 4th Conference.* The Geological Society, London, 1123-1136.

Roberts, G.P. 1996. Noncharacteristic normal faulting surface ruptures from the Gulf of Corinth, Greece. *Journal of Geophysical Research,* **101,** 25255-25267.

Roberts, J.D., Matthieson, A.S. and Hampson, J.M. 1987. Statfjord. *In:* Spencer, A.M. *et al.* (eds.), *Geology of the Norwegian Oil and Gas Fields.* Graham & Trotman, London, 319-340.

Roberts, S. and Jackson, J. 1991. Active normal faulting in central Greece: an overview. *In:* Roberts, A., Yielding, G. and Freeman, B. (eds.), *The geometry of normal faults.* Special publication of the Geological Society, London, **56,** 125-142.

Rosendahl, B.R., Reynolds, D.J., Lorber, P.M., Burgess, C.F., McGill, J., Scott, D., Lambiase, J.J. and Derksen, S.J. 1986. Structural expressions of rifting: Lessons from Lake Tanganyika, Africa. *In:* Frostick, L.M., Renaut, R.W., Reid, I. and Tiercelin, J.J. (eds.), *Sedimentation in the African Rifts.* Special publication of the Geological Society, London, **25,** 29-43.

Royden, L. and Keen, C.E. 1980. Rifting processes and thermal evolution of the continental margin of eastern Canada determined from subsidence curves. *Earth and Planetary Science Letters,* **51,** 343-361.

Sawyer, M.J. and Keegan, J.B. 1996. Use of Palynofacies Characterisation in Sand-dominated Sequences, Brent Group, Ninian Field, UK North Sea. *Petroleum Geoscience*, **2**, 289-297.

Schlager, W. 1993. Accommodation and supply – a dual control on stratigraphic sequences. *Sedimentary Geology,* **86,** 111-136.

Schlische, R.W. 1991. Half-graben basin filling models: New constraints on continental extensional basin development. *Basin Research,* **3,** 123-141.

Schlische, R.W. 1995. Geometry and origin of fault-related folds in extensional settings. *AAPG Bulletin,* **79,** 1661-1678.

Schlische, R.W. and Anders, M.H. 1996. Stratigraphic effects and tectonic implications of the growth of normal faults and extensional basins. *In:* Beratan, K.K. (ed.), *Reconstructing the history of Basin and Range extension using sedimentology and stratigraphy.* Geological Society of America Special Paper **303,** 183-203.

Schlische, R.W., Young, S.S. and Ackermann, R.V. 1996. Geometry and scaling relations of a population of very small rift-related normal faults. *Geology,* **24,** 683-686.

Scholz, C.H., Dawers, N.H., Yu, J.Z. and Anders, M.H. 1993. Fault growth and fault scaling laws: preliminary results. *Journal of Geophysical Research,* **98,** 21951-21961.

Scholz, C. and Contreras, J. 1998. Mechanics of continental rift architecture. *Geology,* **26,** 967-970.

Schulte, W.M., Van Rossem, P.A.H. and Van de Vijver, W. 1994. Current Challenges in the Brent Field. *Journal of Petroleum Technology*, **46**, 1073-1079.

Schwarz, H.-U. 1982. Sub-aqueous slope failures – experiments and modern occurrences. *Contributions to Sedimentology,* **11,** 116pp.

Sclater, J.G., Royden, L., Horvath, F., Burchfiel, B.C., Semken, S. and Stegena, L. 1980. Formation of the intra-Carpathian basins as determined from subsidence data. *Earth and Planetary Science Letters,* **51**, 299-302.

Sengör, A.M.C. and Burke, K. 1978. Relative timing of rifting and volcanism on Earth and its tectonic implications. *Geophysical Research Letters,* **5,** 419-421.

Sharp, I.R., Gawthorpe, R.L., Underhill, J.R. and Gupta, S. in press. Fault propagation folding in extensional settings: examples of structural style and syn-rift sedimentary response from the Suez rift, Sinai, Egypt. *GSA Bulletin.*

Shreve, R.L. 1966. The Sherman Landslide, Alaska. *Science,* **154,** 1639-1643.

Shreve, R.L. 1968. The Blackhawk landslide. *Geological Society of America Special Paper* **108,** 47pp.

Steel, R.J. and Ryseth, A. 1990. The Triassic – early Jurassic succession in the northern North Sea: Megasequence stratigraphy and intra-Triassic tectonics. *In:* Hardman, R.F.P. and Brooks, J. (eds.), *Tectonic events responsible for Britain's oil and gas reserves.* Special publication of the Geological Society, London, **55,** 139-168.

Stein, R.S. and Barrientos, S.E. 1985. Planar high-angle faulting in the Basin and Range: geodetic analysis of the 1983 Borah Peak, Idaho earthquake. *Journal of Geophysical Research,* **93**, 11355-11366.

Stein, R.S., King, G.C.P. and Rundle, J.B. 1988. The growth of geological structures by repeated earthquakes. 2. Field examples of continental dip-slip faults. *Journal of Geophysical Research*, **93**, 13319-13331.

Stoker, M.S., Evans, D. and Cramp, A. (eds.) 1998. *Geological processes on continental margins: Sedimentation, mass-wasting and stability*. Special publication of the Geological Society, London, **129,** 355pp.

Stow, D.A.V., Reading, H.G. and Collinson, J.D. 1996. Deep seas. *In:* Reading, H.G. (ed.), *Sedimentary environments: processes, facies and stratigraphy.* Blackwell Science, London, 395-453.

Struijk A.P and Green, R.T. 1991. The Brent Field Block 211/29, UK North Sea. *In:* Abbotts, I.L. (ed.), *United Kingdom Oil and Gas Fields: 25 Years Commemorative Volume*, Geological Society Memoir, **14**, 63-72.

Summerfield, M.A. and Hulton, N.J. 1994. Natural controls of fluvial denudation rates in major world drainage basins. *Journal of Geophysical Research,* **99,** 13871-13885.

Surlyk, F. 1977. Stratigraphy, tectonics and palaeogeography of the Jurassic sediments of the areas north of Kong Oscars Fjord, East Greenland. *Bulletin Grønlands Geologiske Undersøgelse,* **123,** 56pp.

Surlyk, F. 1978. Submarine fan sedimentation along fault scarps on tilted fault blocks (Jurassic - Cretaceous boundary, East Greenland). *Bulletin Grønlands Geologiske Undersøgelse,* **128,** 108pp.

Surlyk, F. 1987. Slope and deep shelf gully sandstones, Upper Jurassic, East Greenland. *AAPG Bulletin,* **71,** 464-475.

Surlyk, F. 1989. Mid-Mesozoic syn-rift turbidite systems, controls and predictions. *In:* Collinson, J.D. (ed.), *Correlation in hydrocarbon exploration.* Norwegian Petroleum Society (NPF), Graham and Trotman, London, 231-241.

Surlyk, F. 1990a. Timing, style and sedimentary evolution of Late Palaeozoic - Mesozoic extensional basins of East Greenland. *In:* Hardman, R.F.P. and Brooks, J. (eds.), *Tectonic Events Responsible for Britain's Oil and Gas Reserves.* Special publication of the Geological Society, London, **55,** 107-125.

Surlyk, F. 1990b. A Jurassic sealevel curve for East Greenland. *Palaeogeography, Palaeoecology, Palaeoclimatology,* **78,** 71-85.

Surlyk, F. 1991. Sequence stratigraphy of the Jurassic - Lowermost Cretaceous of East Greenland. *AAPG Bulletin,* **75,** 1468-1488.

Surlyk, F., Callomon, J.H., Bromely, R.G. and Birkelund, T. 1973. The stratigraphy of the Jurassic - Lower Cretaceous sediments of Jameson Land and Scoresby Land, East Greenland. *Bulletin Grønlands Geologiske Undersøgelse,* **105,** 76pp.

Surlyk, F. and Clemmensen, L.B. 1983. Rift propagation and eustacy as controlling factors during Jurassic inshore and shelf sedimentation in northern East Greenland. *Sedimentary Geology,* **34,** 119-143.

Talwani, M. and Eldholm, O. 1977. Evolution of the Norwegian-Greenland Sea. *GSA Bulletin,* **88,** 969-999.

Thomson, K., Green, P.F., Whitham, A.G., Price, S.P. and Underhill, J.R. 1999. New constraints on the thermal history of North-East Greenland from apatite fission-track analysis. *GSA Bulletin,* **111,** 1054-1068.

Trudgill, B. and Cartwright, J. 1994. Relay-ramp forms and normal fault linkages, Canyonlands National Park, Utah. *GSA Bulletin,* **106,** 1143-1157.

Underhill, J.R. 1998. Jurassic. *In:* Glennie, K.W. (ed.), *Petroleum geology of the North Sea: Basic concepts and recent advances.* Blackwell Science (4th edition), London, 245-293.

Underhill, J.R. and Partington, M.A. 1993. Jurassic thermal doming and deflation in the North Sea: Implications of the sequence stratigraphic evidence. *In:* Parker, J.R. (ed.), *Petroleum Geology of Northwest Europe: Proceedings of the 4th Conference.* The Geological Society, London, 337-345.

Underhill, J.R. and Partington, M.A. 1994. Use of genetic sequence stratigraphy in defining and determining a regional tectonic control on the 'Mid-Cimmerian Unconformity': implications for North Sea basin development and the global sea-level chart. *In:* Weimer, P.J. and Posamentier, H.W. (eds.), *Siliciclastic sequence stratigraphy.* AAPG Memoir **58,** 449-484.

Underhill, J.R. and Woodcock, N.H. 1987. Faulting mechanisms in high-porosity sandstones; New Red Sandstone, Arran, Scotland. *In:* Jones, M.E. and Preston, R.M.F. (eds.), *Deformation of sediments and sedimentary rocks.* Special publication of the Geological Society, London, **29**, 91-105.

Underhill, J.R., Sawyer, M.J., Hodgson, P., Shallcross, M.D. and Gawthorpe, R.L. 1997. Implications of fault scarp degradation for Brent Group prospectivity, Ninian Field, Northern North Sea. *AAPG Bulletin,* **81,** 999-1022.

Upcott, N.M., Mukasa, R.K., Ebinger, C.J. and Karner, G.D. 1996. Along-axis segmentation and isostasy in the Western rift, East Africa. *Journal of Geophysical Research*, **101**, 3247-3268.

Upton, B.G.J., Emeleus, C.H. and Hald, N. 1980. Tertiary volcanism in northern E Greenland: Gauss Halvø and Hold With Hope. *Journal of the Geological Society, London,* **137,** 491-508.

Upton, B.G.J., Emeleus, C.H., Rex, D.C. and Thirwall, M.F. 1995. Early Tertiary magmatism in NE Greenland. *Journal of the Geological Society, London,* **152,** 959-964.

Varnes, D.J. 1978. Slope movement types and processes. *In:* Schuster, R.L. and Krizek, R.J. (eds.), *Landslides – analysis and control.* National Academy of Sciences Transportation Research Board Special Report **176**, 12-33.

Van Wagoner, J.C., Mitchum, R.M., Campion, K.M. and Rahmanian, V.D. 1990. Siliciclastic sequence stratigraphy in well logs, cores and outcrops: Concepts for high-resolution correlation of time and facies. *AAPG methods in exploration series* **7,** 55pp.

Villemin, T., Angelier, J. and Sunwoo, C. 1995. Fractal distribution of fault length and offsets: Implications of brittle deformation evaluation – Lorraine Coal Basin. *In:* Barton, C.C. and La Pointe, P.R. (eds.), *Fractals in the earth sciences.* Plenum Press, New York, 205-226.

Vischer, A. 1943. Die Postdevonische Tektonik von Ostgrönland zwischen 74° und 75° BR. *Meddelelser om Grønland,* **133(1),** 194pp.

Walsh, J.J. and Watterson, J. 1988. Analysis of the relationship between displacements and dimensions of faults. *Journal of Structural Geology,* **10,** 239-247.

Watterson, J. 1986. Fault dimensions, displacement and growth. *Pure and Applied Geophsysics,* **124,** 365-373.

Wells, J.T., Scholz, C.A. and Soreghan, M.J. 1999. Processes of sedimentation on a lacustrine border-fault margin: Interpretation of cores from Lake Malawi, East Africa. *Journal of Sedimentary Research,* **69,** 816-831.

Wernicke, B. 1981. Low-angle normal faults in the Basin and Range province – nappe tectonics in an extending orogen. *Nature*, **291**, 645-648.

Wernicke, B. 1995. Low-angle normal faults and seismicity – a review. *Journal of Geophysical Research*, **100**, 20159-20174.

Wernicke, B. and Burchfiel, B.C. 1982. Modes of extensional tectonics. *Journal of Structural Geology*, **4**, 105-115.

Westaway, R. 1999. The mechanical feasibility of low-angle normal faulting. *Tectonophysics*, **308**, 407-443.

Whitham, A.G. and Price, S.P. 1994. *Albian-Cenomanian sedimentation in the Traill Ø region of East Greenland.* Unpublished East Greenland Project (CASP) Report **30.**

Whitham, A.G. and Price, S.P. 1997. *A photo library of the degradation products of a Jurassic footwall slope, S Svinhufvuds Bjerge, Traill Ø*. Unpublished East Greenland Project (CASP) Report.

Whitham, A.G., Price, S.P., Koraini, A.M. and Kelly, S.R.A. 1999. Cretaceous (post-Valanginian) sedimentation and rift events in NE Greenland (71-77°N). *In:* Fleet, A.J. and Boldy, S.A.R. (eds.), *Petroleum geology of northwest Europe: Proceedings of the 5th conference.* Geological Society, London, 325-336.

Willemse, E.J.M. 1997. Segmented normal faults: Correspondence between three-dimensional mechanical models and field data. *Journal of Geophysical Research,* **102,** 675-692.

Willemse, E.J.M., Pollard, D.D. and Aydin, A. 1996. Three-dimensional analysis of slip distributions on normal fault arrays with consequences for fault scaling. *Journal of Structural Geology,* **18,** 295-310.

Withjack, M.O., Olson, J. and Petersen E. 1990. Experimental models of extensional forced folds. *AAPG Bulletin,* **74**, 1038-1054.

Wong, T.-F., David, C. and Zhu, W. 1997. The transition from brittle faulting to cataclastic flow in porous sandstones: Mechanical deformation. *Journal of Geophysical Research,* **102**, 3309-3025.

Woodcock, N.H. 1979. Sizes of submarine slides and their significance. *Journal of Structural Geology,* **1**, 137-142.

Yielding, G. 1990. Footwall uplift associated with Late Jurassic normal faulting in the northern North Sea. *Journal of the Geological Society, London,* **147**, 219-222.

Yielding, G., Badley, M.E. and Roberts, A.M. 1992. The structural evolution of the Brent Province. *In:* Morton, A.C., Haszeldine, R.S., Giles, M.R. and Brown, S. (eds.) *Geology of the Brent Group.* Special publication of the Geological Society, London, **61**, 27-43.

Zhang, J., Wong, T.-F. and Davis, D.M. 1990. Micromechanics of pressure-induced grain crushing in porous rocks. *Journal of Geophysical Research,* **95**, 341-352.

Zhu, W. and Wong, T.-F. 1997. The transition from brittle faulting to cataclastic flow: Permeability evolution. *Journal of Geophysical Research,* **102**, 3027-3041.