STRATIGRAPHIC DEVELOPMENT AND CONTROLS ON THE ARCHITECTURE OF EOCENE DEPOSITIONAL SYSTEMS IN THE FAROE-SHETLAND BASIN

ANDREW MARK ROBINSON

Submitted in partial fulfilment of the requirements for the degree of Ph.D.

Cardiff University

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SUMMARY

A detailed investigation of the Eocene stratigraphy in the Faroe-Shetland Basin was undertaken with a view to developing a basin-wide seismic-stratigraphic framework in order to describe the palaeogeographic evolution and depositional architecture of the basin. This study has tested the applicability of sequence stratigraphy on a regional and local scale and outlines the major depositional controls on the succession.

Integration of seismic, well and biostratigraphic data has allowed for the identification of four regionally correlatable seismic-stratigraphic units which document the basin fill history. Depositional systems are controlled by the interaction of fluctuating relative sea-level and local tectonic controls. Early Eocene uplift on compressional fold structures, inherited Mesozoic palaeobathymetry, dynamic uplift from mantle plume activity and prograding lava deltas all control the distribution, thickness and facies of the Eocene succession.

Case studies from a basin margin and intra-basinal setting have provided detailed evidence of the localised sedimentary response to changing basin conditions. The use of high resolution 3-D seismic data has enabled depositional sequences and the complex seismic geomorphology of palaeo-drainage systems to be recognised. A cyclicity of sedimentary response is observed in a deltaic environment which documents the evolution of relative sea-level on the southern margin of the basin.

Classical sequence-stratigraphic techniques have, in places, provided a useful guide to stratigraphic interpretation and analysis. However, attempts to test the widely used sequence stratigraphic model of sequences as regionally correlatable stratigraphic surfaces have failed largely because of the impracticability of correlating seismic reflections on a regional scale. The conclusion from this is that sequence stratigraphy is a powerful analytical tool that can be applied locally, but is likely to encounter significant difficulties on a basinal length scale. This larger scale correlatability is a corollary of the link between sequence development and eustatic sea level changes. In this thesis, this link cannot be substantiated, and local factors predominate.

ACKNOWLEDGEMENTS

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"You are not alone in this world"

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ENCLOSURES

The key figures which are referred to throughout the text are included as enclosures at the back of the thesis. These enclosures also appear as figures in the main text when they are first referred to.

Plus two papers as enclosures:

1. Robinson, A. M., Cartwright' J. A., Burgess' P. M. & Davies' R.J. 2004. Interactions between topography and channel development from 3-D seismic analysis: An example from the Tertiary of the Flett Ridge, Faroe-Shetland Basin, (UK). In: R. J. Davies, J. A. Cartwright, S. A. Stewart, J. R. Underhill & M. Lappin. (eds). 3D seismic data: Advances in the understanding of stratigraphic and structural architecture. Geological Society Memoir 29, London. 73-82.

2. Davies, R. J., Cloke, I., Cartwright, J. A., Robinson, A. M. & Ferrerro, C. 2004. Post Break-up compression of a passive margin and its impact upon Hydrocarbon Prospectivity: an example from the Tertiary of the Faeroe-Shetland Basin. AAPG Bulletin, Vol 88, No. 1, 1-20.

1. Chapter One: Introduction

1.1 Location and Regional Structural Elements

1.1.1 Geographical Setting

The Faroe-Shetland Basin is located on the northeastern margin of the Atlantic Ocean, situated some 100 km west of the coast of the Shetland Islands (**Figure 1.1a**). Lying directly between the Shetland and Faroe Islands, the basin extends for over 400 km in a northeast - southwest direction and forms part of the Northwest European continental margin.

1.1.2 Tectonic Setting of Faroe-Shetland Basin

The Northwest European continental margin comprises a number of northeast – southwest trending sedimentary basins, first documented by Bott and Watts (1971). These sedimentary basins overlie thinned and extended continental crust and form a passive continental margin (Bott 1975, Bott 1984, Bott and Smith 1984). The Faroe-Shetland Basin is one of these sedimentary basins and comprises a predominantly Cenozoic fill (e.g. Mudge and Rashid 1987, Lamers and Carmichael 1999). It is bordered to the northeast by the Møre Basin and to the southwest by the Rockall Trough (**Figure 1.1a**).

These sedimentary basins are bounded by major northeast - southwest trending structural lineaments which are seen in the entire West of Shetland area (Bott and Watts 1971, Bott 1975), The Shetland Spine Fault forms one of these lineaments and is the major basin bounding fault system that separates the stable West Shetland Platform (to the east) from the West Shetland Basin (**Figure 1.1b**). Further west, the Clair and Rona Ridges separate the West Shetland Basin from the deeper Faroe-Shetland Basin (Duindam and Van Hoorn 1987, Mudge and Rashid 1987, Hitchen and Ritchie 1987). These ridges manifest themselves as large Mesozoic tilted fault-blocks that formed during an episodic yet prolonged period of crustal extension of the Northwest European margin (e.g. Dean *et al.* 1999). The crustal structure of the study

a)

b)

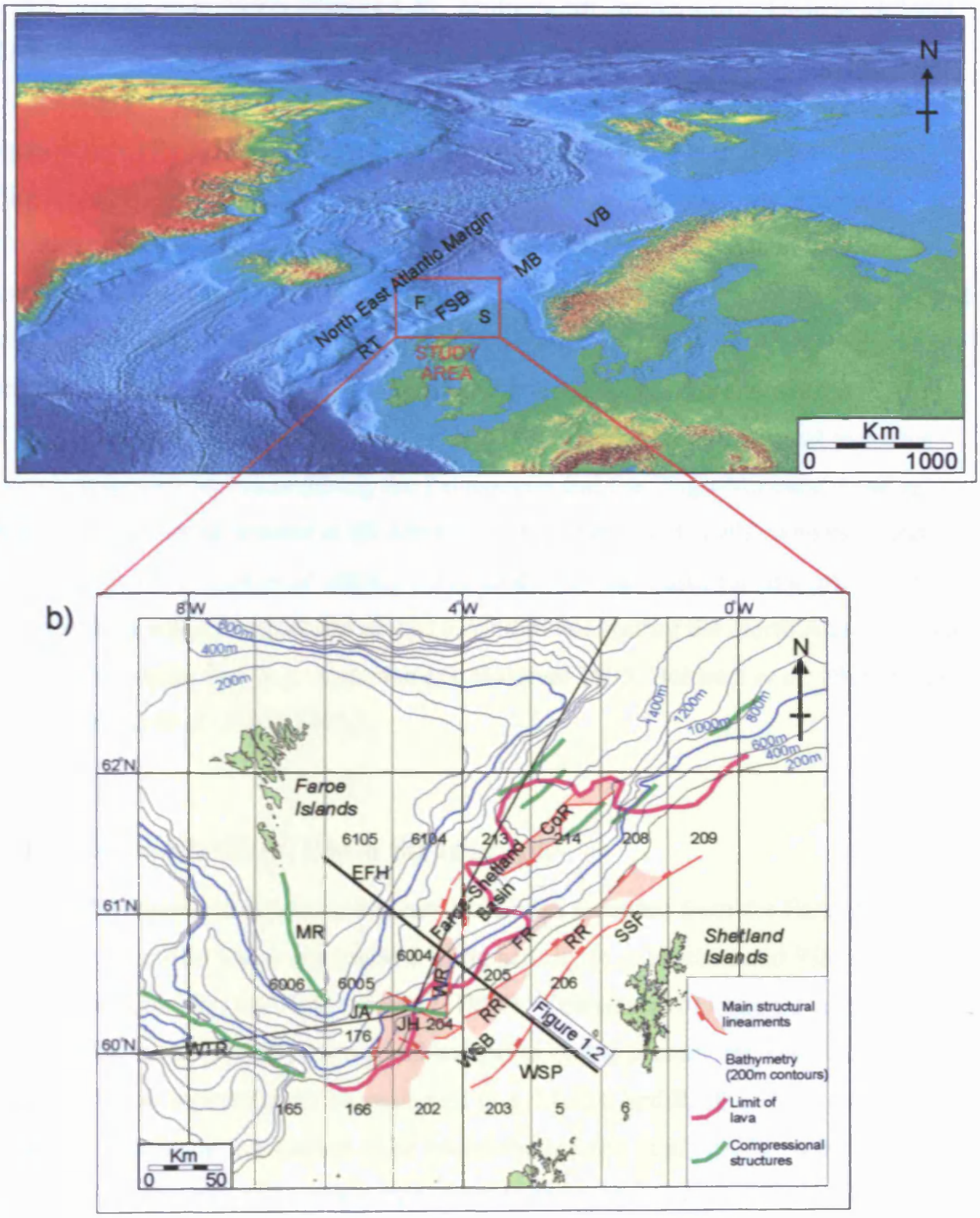


Figure 1.1:Location map of the study area. a) Bathymetry map highlighting the position of the Faroe-Shetland Basin (FSB) on the Northwest European continental margin. Abbreviations are VB=Vøring Basin, MB= Møre Basin, FSB= Faroe-Shetland Basin, RT= Rockall Trough, S=Shetland Islands and F= Faroe Islands. b) Major structural elements of the FSB. The major northeast - southwest Mesozoic tilted fault-blocks and extent of lava front are shown. Abbreviations are SSF= Shetland Spine Fault, FR= Flett Ridge, RR= Rona Ridge, CoR= Corona Ridge, WR= Westray Ridge, JH= Judd High area and Judd Fault, JA= Judd Anticline, WSB= West Shetland Basin, WSP= West Shetland Platform, EFH = East Faroe High, MR= Munkagrunnur Ridge and WTR= Wyville-Thompson Ridge. The location of the cross-section in Figure 1.2 is shown.

area can be simplified into a series of westerly dipping Mesozoic fault-blocks that are buried by a thick post-rift sedimentary and igneous cover succession (e.g. Anderton 1993, Naylor *et al.* 1999) (**Figure 1.2**). Multiple rift episodes (e.g. Dean *et al.* 1999, Doré *et al.* 1999; see **Section 2.2**) preceded the final Palaeogene break-up event, culminating in sea-floor spreading of the North Atlantic and resulted in voluminous igneous activity in the entire West of Shetland area (e.g. Knox and Morton 1988, Naylor *et al.*1999). The Corona Ridge forms the most northwesterly structural lineament seen in the Faroe-Shetland Basin, the western side of which is, in parts, covered and interbedded with Tertiary volcanics and intrusives (**Figures 1.1b** and **1.2** and **Section 2.3**). Post-rift thermal subsidence affected the whole of the margin after initiation of sea-floor spreading (Clift and Turner 1995, Turner and Scrutton 1993). This subsidence was episodically interrupted by pulses of compressional tectonics which deformed the basin during the Palaeogene and the Oligo-Miocene creating numerous inversion structures (Boldreel and Andersen 1993; 1998, Andersen and Boldreel 1995, Davies *et al.* 2004). The Faroe-Shetland Basin forms a present day narrow deep water channel and acts as a conduit connecting the North Atlantic Ocean to the Norwegian Sea (e.g. Andersen and Boldreel 1995, Boldreel *et al.* 1998, Stoker 1998, Davies *et al.* 2001; 2002,).

1.1.3 Faroe-Shetland Basin Stratigraphy

Palaeozoic to Recent sediments have been recorded from the Faroe-Shetland Basin by scientific and commercial drilling activity (e.g. Hitchen and Ritchie 1987, Mudge and Rashid 1987, Spencer *et al.* 1999). Correlation of regional seismic interpretations calibrated to wells and boreholes on the present day shelf has enabled the deep basin succession to be examined (e.g. Mudge and Rashid 1987) and a lithostratigraphic nomenclature to be established (Knox *et al.* 1997). A brief and simplified summary of the stratigraphy is outlined here, but a more detailed discussion can be found in **Chapter 2, Section 2.2**.

The oldest known rocks found in the area are crystalline gneisses and are predominantly Lewisian in age (Ridd 1981). These metamorphic basement rocks are overlain by a thick succession of Palaeozoic and Mesozoic sediments (e.g. Mudge and Rashid 1987, Hazeldine *et al.* 1987, Earle *et al.* 1989). The Mesozoic occurs as large syn-rift packages (e.g. Dean *et al.* 1999). Cenozoic sediments occur as a thick post-rift

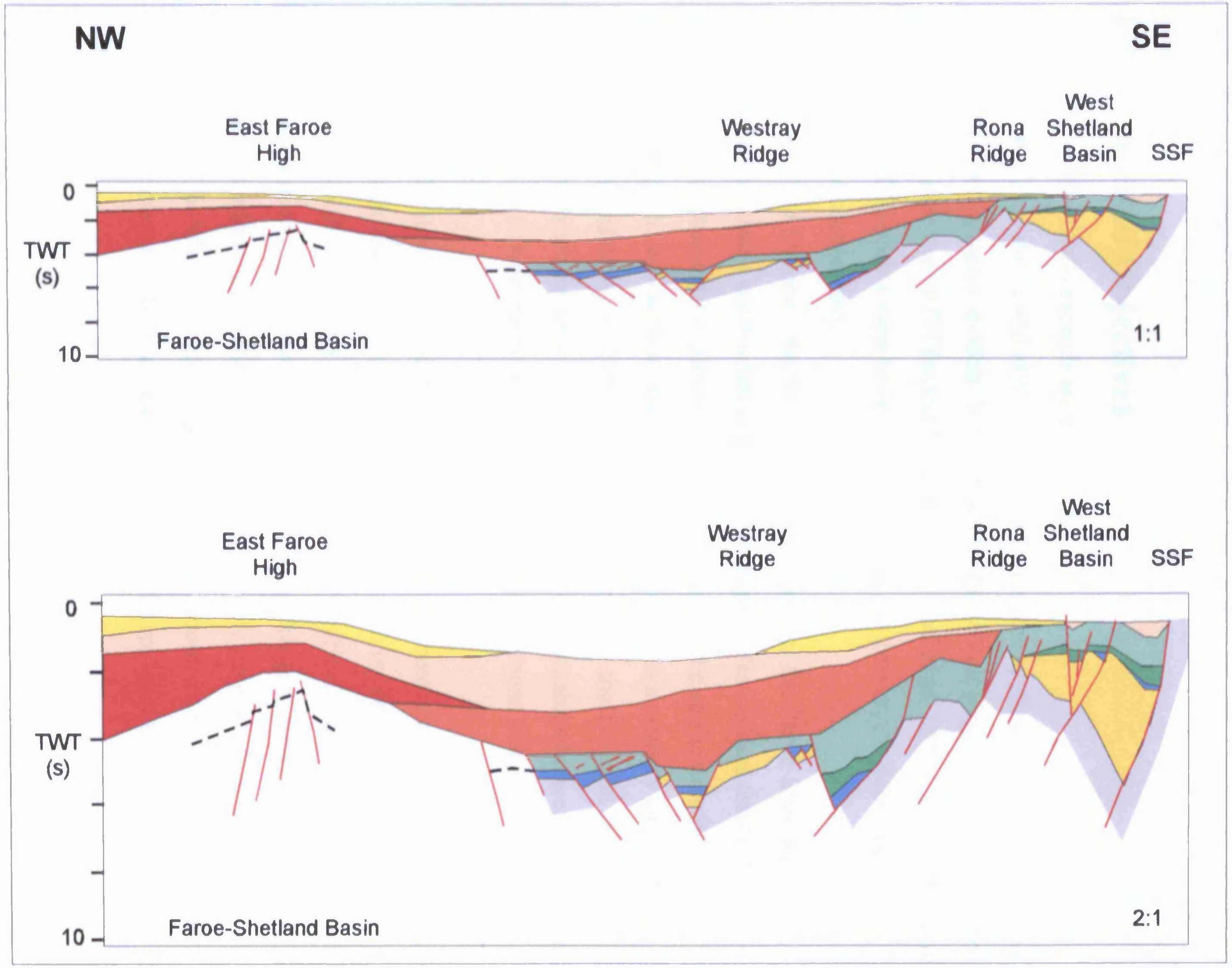
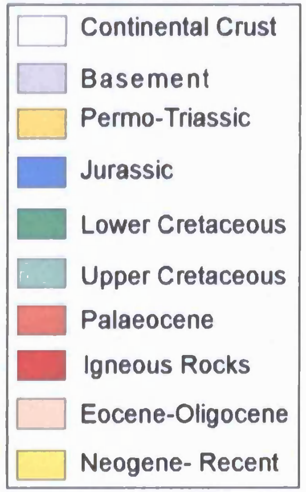


Figure 1.2. Southeast - northwest geo-seismic cross section of the Faroe-Shetland Basin (see Figure 1.1 for location) showing the major structural elements. The top geo-seismic is at a scale of 1:1, the bottom at a scale of 2:1. Major Mesozoic tilted fault-blocks formed during a multi-phased extension from the Permo-Triassic to the Palaeocene. The older West Shetland Basin is highlighted to the southeast. Palaeogene magmatism was extensive and voluminous lavas and sill complexes obscure the deep structure of the northwestern part of the basin. SSF= Shetland Spine Fault. Redrawn from Dore *et al.* (1999).



wedge that was primarily sourced from the Scottish mainland to the southeast (e.g. Ebdon *et al.* 1995, Lamers and Carmichael 1999). Voluminous lava erupted and flowed from the northwest during sea-floor spreading which began at the start of the Eocene (e.g. Knox and Morton 1988, Naylor *et al.* 1999). Contourites which form extensive Oligocene – Recent age drift deposits indicate the initiation of the North Atlantic deep water current and connection between the North Atlantic and the Norwegian Sea during this period (e.g. Stoker 1997; 1998, Davies *et al.* 2001, Knutz and Cartwright 2004).

1.2 Aims and Objectives

The aims of this research are as follows:

- This thesis will attempt to critically test the applicability of seismic and sequence-stratigraphic models first outlined by Peter Vail and his co-workers in the late 1970's (see the 1977 "Payton Volume" - e.g. Vail *et al.* 1977a, b & c, Mitchum and Vail 1977) with a view to defining the major controls on the deposition of stratigraphic successions.
- The Eocene – Recent stratigraphic record of the Faroe-Shetland Basin, located on a passive continental volcanic margin provides an ideal test-bed to evaluate some basic principles of the models. Testing the application of the methodology for sub-surface stratigraphic sub-division is a primary aim of this research. Specifically, long range correlations from the shelf margin to the basin centre can enhance the understanding of the inter-relationships between deep water fan systems and their marginal equivalents during variations in eustasy, tectonics and climate.
- To document and describe the Eocene - Recent succession of the Faroe-Shetland Basin using a seismic-stratigraphic approach in order to sub-divide the stratigraphy and understand the basin-fill history. Particular emphasis on the gross basin architecture and the sedimentary response to changing basin dynamics (tectonics versus eustasy) throughout the stratigraphic period will be examined.
- To define some of the principle tectonic controls on deposition of the Eocene succession on both a local (1 - 10 km) and a regional (10's – 100's km) scale.

- To examine specific marginal areas of the basin in order to document the depositional controls on the Early and Middle Eocene deltaic systems and to understand the response to basin subsidence.
- To define the controls on major Middle Eocene deep water fan systems and determine their spatial and temporal relationship to more marginal deltaic facies.
- To build a seismic stratigraphic framework of the Eocene succession that allows for a better understanding of the evolution of the basin and understand the major controlling factors relating to the uplift and subsidence which occurred during and immediately after the break-up of the North Atlantic Ocean.

1.3 Database

The database consists of two dimensional (2-D) and three dimensional (3-D) seismic reflection data with varying degrees of quality and resolution (**Figure 1.3**). Lithological and age information was used in this study and was acquired from composite logs, velocity logs and biostratigraphic reports from commercial wells and shallow boreholes. No potential field data were acquired in this study, although published gravity and magnetic maps were used in the interpretation of regional tectonics.

A vast seismic database of over six hundred and sixty 2-D seismic profiles from twenty seven surveys has been acquired, digitally loaded and interpreted over the whole of the Faroe-Shetland Basin in order to produce a stratigraphic evolution of the basin (see **Appendix 1**). Additionally, seven 3-D seismic surveys were acquired and interpreted in areas of interest both along the basin margin and in the basin axis (see **Appendix 2**). This seismic data collection has been an ongoing process during the research, as individual surveys became available from both oil and contractor companies.

Coupled with the seismic data, a vast well and British Geological Survey (BGS) borehole database has been constructed (see **Appendix 3**) which has been fully integrated with the geophysical data. Composite and velocity logs have been acquired both digitally and in paper format and interpretation of the gamma and sonic velocity logs have been used to aid the lithological and seismic stratigraphic interpretation. Over one hundred and sixty exploration wells have been drilled in the study area. The

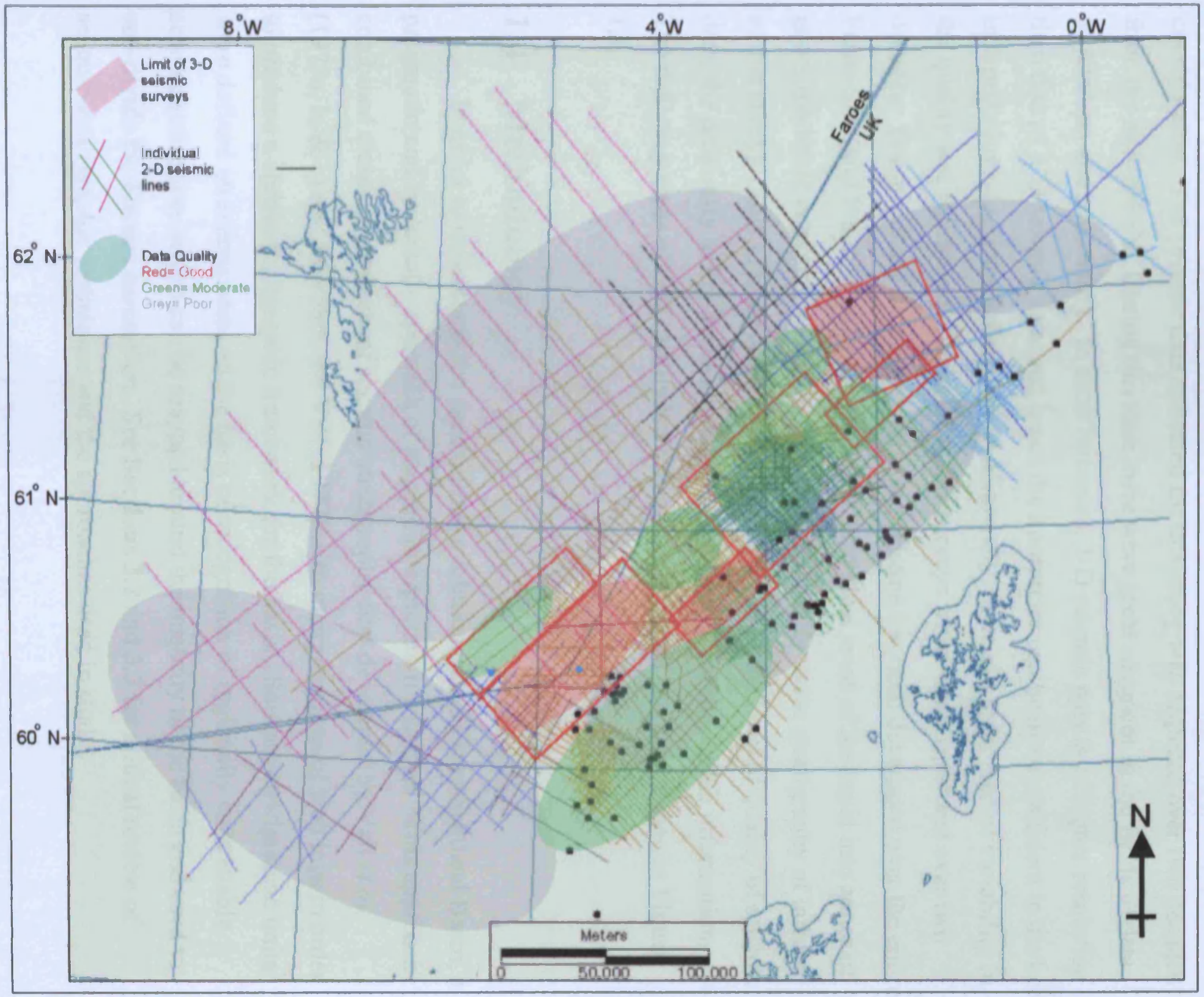


Figure 1.3: Basemap of the Faroe-Shetland Basin showing the full extent of the 2-D and 3-D seismic surveys, the position of wells (black dots) and boreholes (blue dots) used in this study. A crude data quality scheme has been created showing areas of greater quality where there is good quality 3-D seismic data and the availability of good well and biostratigraphic data from the Eocene succession. Note the Shetland margin and the Faroe Platform is generally classified as poor quality due to the lack of 3-D seismic data and well ties. However, the south of the basin has high quality seismic and good well data as too does the very northern area in the region of the recent 214/4-1 well which was made available for this study.

large majority of these have targeted hydrocarbon reservoirs below the Eocene - Recent interval that this study focuses on. However, a small number of wells provide information on lithology and stratigraphic age, and these have been classified in this study as type wells (**see Appendix 4, 5** and **6** and **section 3.5**). Integrated with these wells are biostratigraphic reports that have been used to date (where possible) the Eocene stratigraphy and aid in the interpretation of regional correlations of important marker horizons.

The quality and value of the individual database components varies considerably. The seismic data collected for this study was acquired over two decades from the mid 1980's. During this time there were great advances in the fields of data acquisition and processing. In most instances, 3-D seismic data is of higher resolution than that of 2-D seismic data, and hence the interpreter can be more confident in his/her interpretation of seismic reflections. Additionally, there is a great deal of variability in the quality and resolution of individual 2-D surveys which were collected over two decades. For this study a simple generalised scheme has been developed over the entire Faroe-Shetland Basin in order to sub-divide individual areas of the basin into zones of good, moderate or poor data quality (with reference only to the stratigraphy of interest in this study). This scheme is based on a combination of the level of quality of seismic data, the proximity to key wells and the availability of biostratigraphic information. The differing areas of data quality have been colour coded and can be seen in **Figure 1.3.**

1.4 Methodology

In order to understand the depositional evolution of the Faroe-Shetland Basin, a pragmatic and systematic approach of seismic interpretation was used. This approach combined classical methods of seismic-stratigraphy first developed by Vail *et al.* (1977a, b, & c) and Mitchum and Vail (1977) with interpretations of well logs in order to produce a chronostratigraphic framework for the basin. Seismic-stratigraphic units were defined on seismic data on the basis of recognition of regionally correlatable seismic reflections which can be mapped around the majority of the basin and used to sub-divide the Eocene succession. See **Sections 3.2** and **3.3** for critical review of seismic-stratigraphic techniques and the approaches used in study.

1.4.1 Seismic-Stratigraphic Approaches Taken

Seismic reflections are assumed to represent time-lines (Mitchum and Vail 1977, Vail *et al.* 1977a), and thus have chronostratigraphic significance. Mapping of seismic reflections therefore is a valid way of sub-dividing the succession into a number of different packages. Mitchum and Vail (1977) proposed using reflection terminations such as erosional truncation, toplap and onlap to interpret sequence boundaries (defined as an unconformity surface and its correlative conformity - see **Section 3.2** and Mitchum 1977) which form the bounding surfaces of the Exxon sequence. Sequence boundaries are interpreted to be caused by a lowering of relative sea-level (Vail *et al.* 1977a, Mitchum and Vail 1977, see **Section 3.2**).

A more pragmatic approach has been taken to regionally sub-divide the Eocene - Recent stratigraphy into a number of seismic-stratigraphic units. This approach is taken because sequence boundaries are not readily mappable on a regional basis throughout the Faroe-Shetland Basin. However this is not to say that sequence boundaries do not exist. Indeed, a number of intra-Eocene seismic markers which are the most continuous and have the greatest degree of correlatability have been mapped across the entire basin. These markers form chronostratigraphic time-lines allowing the strata to be sub-divided into time intervals. By integrating well and biostratigraphic information a more detailed, age-constrained stratigraphic breakdown can be constructed.

On a more local scale high resolution 3-D seismic data positioned in areas on the basin margins has enabled candidate sequence boundaries (which show evidence of localised incision) to be mapped across a small part of the basin. Therefore, in scenarios where data quality and location within the basin are favourable, the classical sequence-stratigraphic approach of Vail *et al.* (1977a & b) is used to sub-divide part of the Eocene succession.

The seismic-stratigraphic approaches used in both the regional basin analysis chapter (**Chapter 4**) and in the localised case studies (**Chapters 5** and **6**) are discussed in more detail in **Chapter 3**.

1.5 Organisation of the Thesis

Chapter 1 of this thesis introduces the main scientific aims of the research, the scope of this study and succinctly outlines the key problems addressed herein. The geological setting of the study area is briefly discussed and the database documented.

Chapter 2 summarises the regional geology and plate tectonic setting of the Northwest European margin, and reviews the published literature on the rift history and sedimentary architecture of the West of Shetland area, focussing specifically on the Faroe-Shetland Basin. A chronological summary using published work follows, documenting major changes in basin architecture from the Palaeozoic to Recent.

A critical review of sequence-stratigraphic methodology is detailed in ***Chapter 3***. A discussion of the seismic-stratigraphic approaches used in this study follows. This chapter discusses some of the techniques used in building the stratigraphic framework of the basin, and goes onto to highlight the accuracy and limitations of the techniques.

Chapter 4 provides a regional chronostratigraphic breakdown of the Eocene succession of the Faroe-Shetland Basin. The chapter firstly describes the earliest Eocene T50 unit of Ebdon *et al.* (1995) and goes on to fully detail four new seismic units which are bounded by regionally correlatable seismic reflections. The chapter describes in detail the gross architecture, distribution, lithology and internal architecture of each individual seismic unit. An interpretation of the environments of deposition for each seismic unit follows the description, using the evidence gathered to elucidate the basin dynamics in terms of tectonics (subsidence and uplift), sediment supply and accommodation space. A discussion on regional and local controls on the deposition of the entire Eocene succession concludes this chapter.

Chapter 5 details a case study of a small area of the basin and documents the evolution of an Early Eocene delta system. High resolution mapping on 3-D seismic data of candidate sequence boundaries allows for the classical sequence-stratigraphic approach to be tested with a view to critically examining tectonic or eustatic controls on deposition of an area susceptible to changing relative sea-level.

A second case study using high resolution 3-D seismic data is discussed in ***Chapter 6*** which looks at the effects of evolving topography on the sediment dispersal patterns and channel geometries in the intra-basinal setting of the Flett Ridge.

Chapter *7* draws together the key scientific results of the research which are detailed in the discussion points in ***Chapters 3, 4, 5*** and ***6*** and provides insights into the methodology and applicability of modern sequence stratigraphic techniques. A brief overview of the limitations and uncertainties associated with the study follows and the chapter concludes with proposals for future work.

The main conclusions of the research are listed in ***Chapter 8***.

2. Chapter Two: The Tectonic and Stratigraphic Evolution of the Northeast Atlantic Margin

2.1 Introduction and Overview

This chapter chronologically documents the geological history of the study area by putting the Faroe-Shetland Basin into a regional, plate tectonic context. In the previous thirty years of exploration over one hundred and sixty wells have been drilled in the basin and combined with a great deal of published work on the development of the Northwest European margin (e.g. Ziegler 1989), a detailed structural and stratigraphic evolution can be documented.

The following regional synthesis discusses important geological events that took place in the formation of the Northwest European margin. These major events include:

- Continental collision and accretion during three main orogenies in the Palaeozoic era.
- Mesozoic extensional tectonics on inherited Palaeozoic lineaments.
- Oceanic spreading, continental separation and compression in the Late Mesozoic and Cenozoic.

2.2 Structural and Stratigraphic Evolution of the Faroe-Shetland Basin

2.2.1 Introduction

The structural setting of the Northwest European margin and in particular the Faroe-Shetland Basin has been the focus for many studies since the late 1970's and early 1980's. Bott and Watts (1971) and Bott (1975) first described deep sedimentary basins offshore Scotland by using gravity, magnetic and sparker data to outline the West Shetland Basin. They additionally noted the occurrence of en echelon northeast trending fault arrays and postulated a northwesterly thinning of the continental crust or a northwesterly transition from continental to oceanic crust (Bott 1975, Roberts 1975). It was not until hydrocarbon exploration was well underway and boreholes had been drilled, that the nature of the basement of the West Shetland area was determined. The deeper crustal structure of the West Shetland area with particular emphasis on the Faroe-Shetland Basin is discussed in great detail in **Section 2.4**. Ridd (1981; 1983) produced initial results on the southeast margin of the continental shelf and identified a Lewisian basement structure for the West Shetland Basin. This basin is bounded to the northwest by the Rona Ridge; a northeast trending uplifted horst block first recognised by Kent in 1975. The Rona Ridge separates the West Shetland Basin to the southeast from the deeper Faroe-Shetland Basin to the northwest (see **Figure 1.1b**). By the end of the 1970's, thirty one boreholes had been drilled in the province, the vast majority of which were confined to the more marginal (up-dip) West Shetland Basin and on the crest of the Rona Ridge. The crustal structure beneath the deeper Faroe-Shetland Basin was still unknown. The Faroe-Shetland Basin has long been known to be the northerly continuation of the Rockall Trough and is contiguous with the Møre Basin to the northeast (e.g. Hitchen and Ritchie 1987, Earle *et al.* 1989). However these three basins remain bathymetrically separate. The crustal structure of the Faroe-Shetland Basin will be discussed in more detail in **Section 2.4**.

2.2.2 Pre-Cambrian to the Carboniferous.

In order to understand the structural and stratigraphic setting of the Faroe-Shetland Basin, we first need to evaluate the tectonic events the Northeast Atlantic margin experienced. The evolution of this margin has been directly affected by major periods of crustal tectonics, which can be crudely sub-divided into two periods: continent accretion in the Pre-Cambrian and Palaeozoic, and rifting, spreading and break-up from the Mesozoic. Firstly, from the Pre-Cambrian to the Palaeozoic, the whole of Northwest Europe experienced major tectonic activity in the form of sequential terrane accretion onto the North American (Laurentian) craton (Coward 1990, Ziegler 1990). In the British Isles, the terrane accretion formed northeast trending zones which developed during three major periods of orogenic compression; the Laxfordian (1800 - 1700 Ma), the Caledonian (500 - 400 Ma) and the Variscan (400 - 300 Ma) (Coward 1990). It is the Caledonian Orogeny in the Ordovician - Devonian period that brought together the Northwest Europe Pre-Cambrian basement rocks from the three tectonic domains of Laurentia (North America block), Baltica (Scandinavian block) and Avalonia (part of Gondwanaland until Early Palaeozoic),(Coward *et al.* 2003), (see **Figure 2.1**).

Northeast - southwest trending terranes dominate the structural grain of the Northeast Atlantic margin (Roberts *et al.* 1999), (**Figure 2.2**), and are bounded by high and low angle major faults or sutures (usually thrusts and strike-slip faults). These boundaries separate terranes of differing geological histories which have been juxtaposed during tectonic accretion throughout the Pre-Cambrian and much of the Palaeozoic (Roberts *et al.* 1999). The most northerly terrane (bounded to the southeast by the Moine Thrust or the Caledonian Front) consists of gniesses and granitic gniesses (from original sedimentary and granitic protoliths) of the Northwest Caledonian foreland. These rocks were originally deposited on the southeast margin of the ancient Laurasian continent. Outcrops of these Archean rocks can be found in Greenland, Newfoundland and in northwest Scotland (Lewisian) associated with their Proterozoic cover (e.g. Torridonian). However, since exploration offshore began, Lewisian gneiss has been found to form the basement rock in the Faroe-Shetland Basin. Deep seismic profiling (e.g. the West of Shetlands BIRPS profile (British Institutions Reflection Profiling Syndicate) and the FAST lines (Faroe-Shetland Traverse)) has enhanced our understanding of the deep crustal structure of Northwest Europe, when combined with conventional structural and stratigraphic field data (Coward 1990). It is now commonly agreed that the Laurentian shield has ages for the crystalline basement dating back to approximately 2900 Ma (Coward *et al.* 2003). However, the Lewisian complex was extensively reworked throughout the Proterozoic and has ages from approximately 2500 - 1600 Ma giving it a Scourian - Laxfordian age (e.g. Hitchen and Ritchie 1987). The Lewisian complex underwent several episodes of Late Archean to Early Proterozoic deformation and metamorphism giving rise to a considerable range in age of the basement.

To the southeast of the Moine Thrust, a younger basement province is seen (**Figures 2.1** and **2.2**). Average istotopic ages from Scotland and offshore are in the region of 400 Ma, signifying a Caledonian origin. These crystalline rocks formed on the western margin of the Caledonian orogenic front and consist of biotite gniesses and metabasites (Hitchen and Ritchie 1987). The Moine Thrust (both onshore and offshore delineates the northwesterly limit of thin-skinned deformation (Coward *et al.* 2003). Recently Roberts *et al.* (1999) suggested that the limit of the Caledonian compression

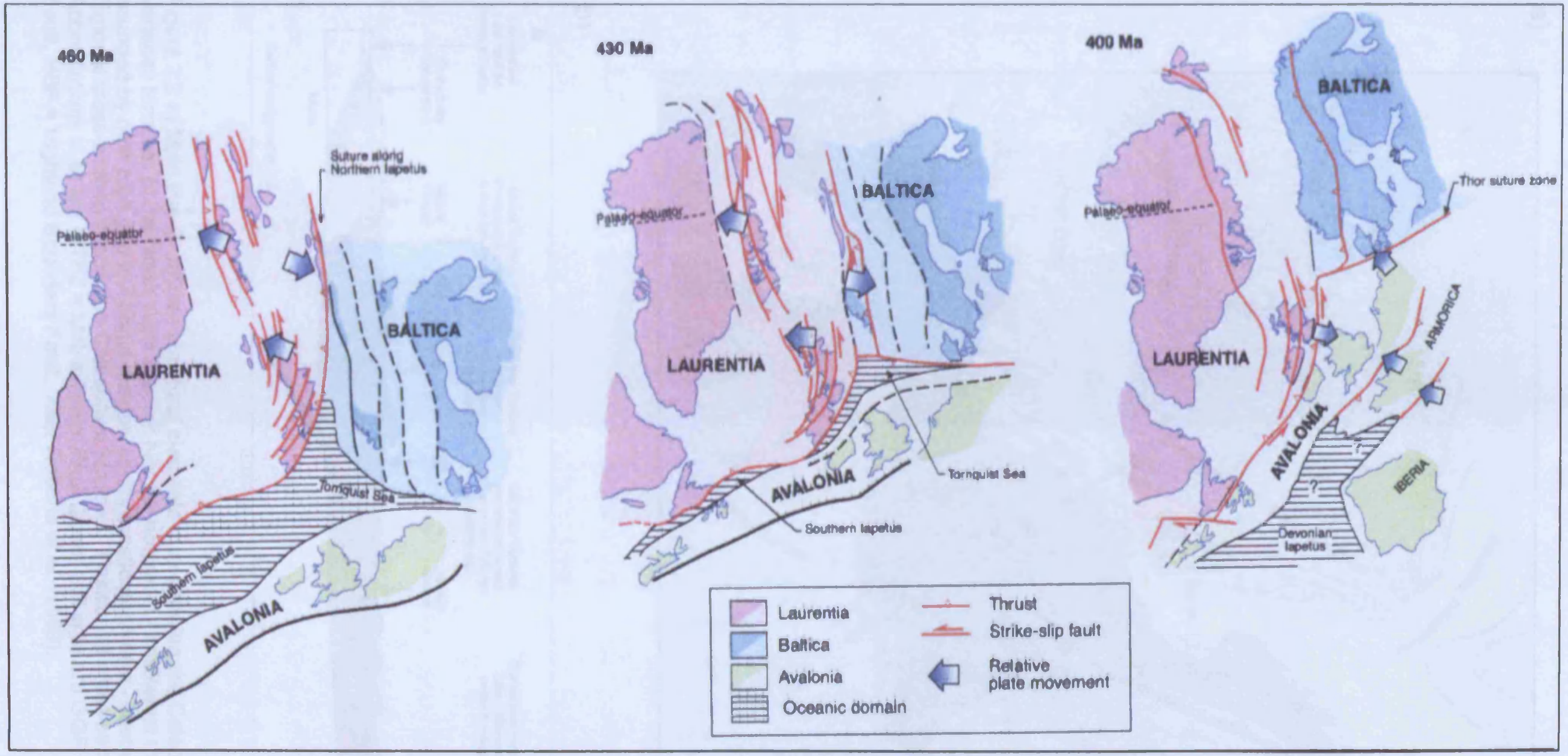


Figure 2.1. Schematic plate reconstructions of Soper *et al.* (1992a) showing the three-way collision between Baltica, Laurentia and Avalonia during the Caledonian Orogeny. The Iapetus Ocean was split into two, the southern part of which separated Avalonia and Laurentia, with the northern Iapetus closing by 460 Ma between Baltica and Avalonia. The third arm of the triple convergence zone is the Tornquist Sea that trends east - west between Avalonia and Baltica. The now, northwest trending Thor (or Tornquist) suture zone, was fundamental in the later tectonic development of Northwest Europe. This is because the amalgamated continent of Baltica became thickened lithospheric Pre-Cambrian crust on the southwest margin of Northwest Europe. Modified from Klemperer and Hobbs (1991). After Coward *et al.* (2003).

a)

b)

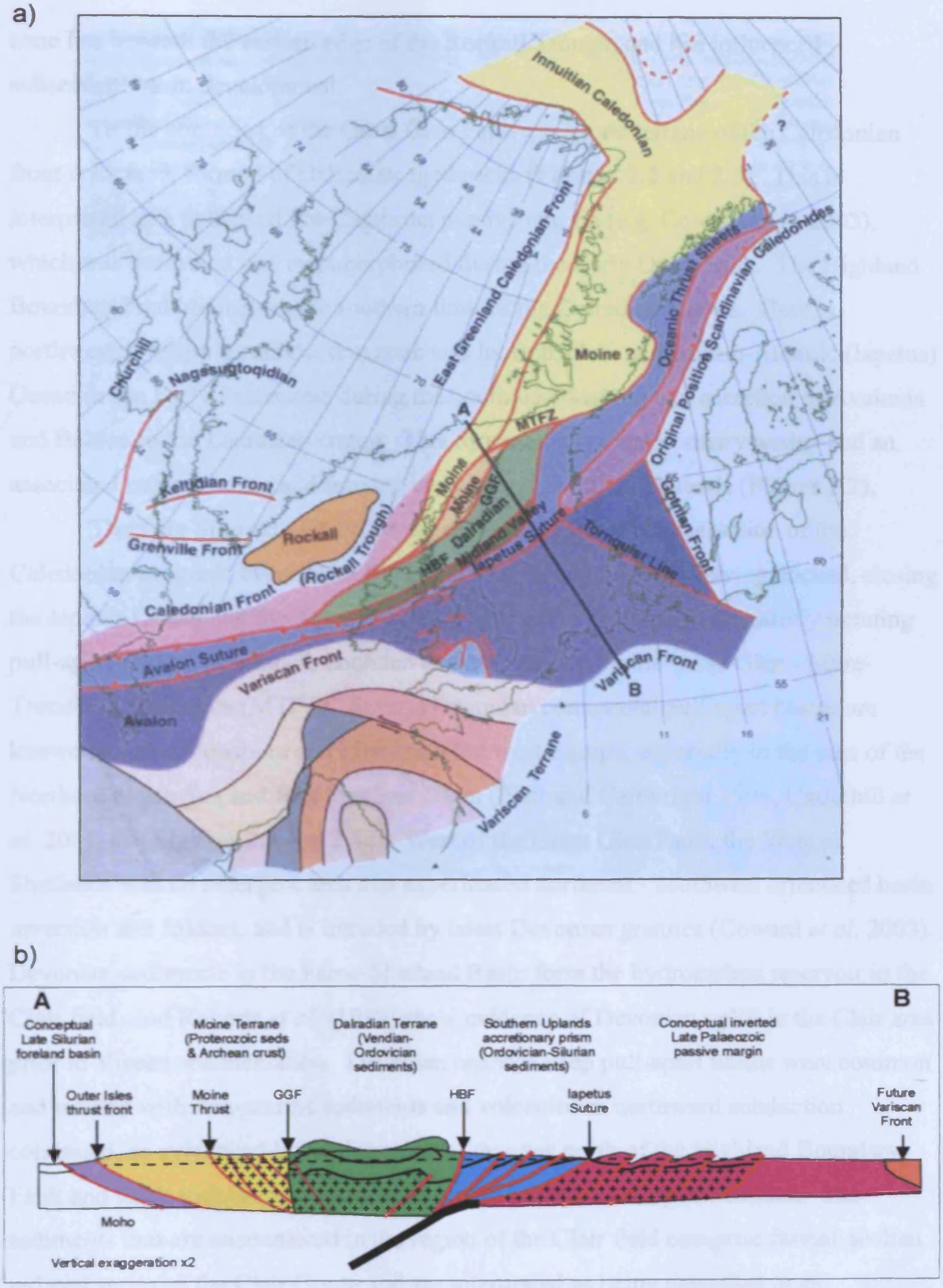


Figure 2.2 a) Main Pre-Cambrian structural elements map showing the Caledonian and Variscan terranes of the North West Europe. Nb: the northeast - southwest trending terranes bounded by both high and low angle sutures. b) Schematic northwest - southeast trending regional cross-section across the Caledonian terranes. Position of cross-section shown in a). Abbreviations in a) are MTFZ = Moine Thrust Fault Zone, and in b) are GGF = Great Glen Fault, HBF = Highland Boundary Fault. After Roberts *et al.* (1999).

zone lies beneath the eastern edge of the Rockall Trough, and has influenced subsequent basin development.

To the southeast of the Great Glen Fault a separate terrane of the Caledonian front is located, formed of Dalradian aged rocks (**Figures 2.2** and **2.3**). This is interpreted as a collapsed Pre-Cambrian passive margin (e.g. Coward *et al.* 2003), which was deformed and metamorphosed during the Early Ordovician. The Highland Boundary Fault delineates the southern limit of the Dalradian terrane. Here, a northwesterly dipping subduction zone was located, closing the proto-Atlantic (Iapetus) Ocean in the Early Palaeozoic during the continued docking and accretion of Avalonia and Baltica to the Laurasian craton. This formed a large accretionary wedge and an associated magmatic arc and fore-arc basin of the Southern Uplands (**Figure 2.2**).

The Late Silurian - Early Devonian period signified the cessation of the Caledonian orogenic event, with Baltica, Avalonia and Laurasia having docked, closing the Iapetus Ocean and the Tornquist Sea (**Figure 2.1**). A large northeasterly trending pull-apart system developed, bounded to the northwest by the Great Glen - Møre-Trøndelag fault zone (MTFZ). Several Devonian continental pull-apart basins are known from both onshore and offshore Northwest Europe, especially in the area of the Northern North Sea and East Shetland Basin (Platt and Cartwright 1998, Underhill *et al.* 2001, see **Figures 2.3** and **2.4a**). West of the Great Glen Fault, the West of Shetlands was an emergent area that experienced northeast - southwest orientated basin inversion and folding, and is intruded by latest Devonian granites (Coward *et al.* 2003). Devonian sediments in the Faroe-Shetland Basin form the hydrocarbon reservoir in the Clair field, and Roberts *et al.* (1999) show evidence of Devonian uplift in the Clair area prior to Visean sedimentation. Devonian onshore deep pull-apart basins were common and infilled with non-marine sediments and volcanics as northward subduction continued, as evidenced by the intrusion of granites north of the Highland Boundary Fault and thick successions of volcanics in the Midland Valley of Scotland. The sediments that are encountered in the region of the Clair field comprise fluvial-aeolian red-bed series of the Clair Group and are interpreted as being deposited in an extensional intermontane basin (Coward and Enfield 1987). A reactivation of Caledonian thrust structures is seen in the southeasterly dipping Devono - Carboniferous extensional faults of the West Orkney Basin (**Figure 2.3**) and the East Shetland Platform where reactivation caused half graben development (Platt and

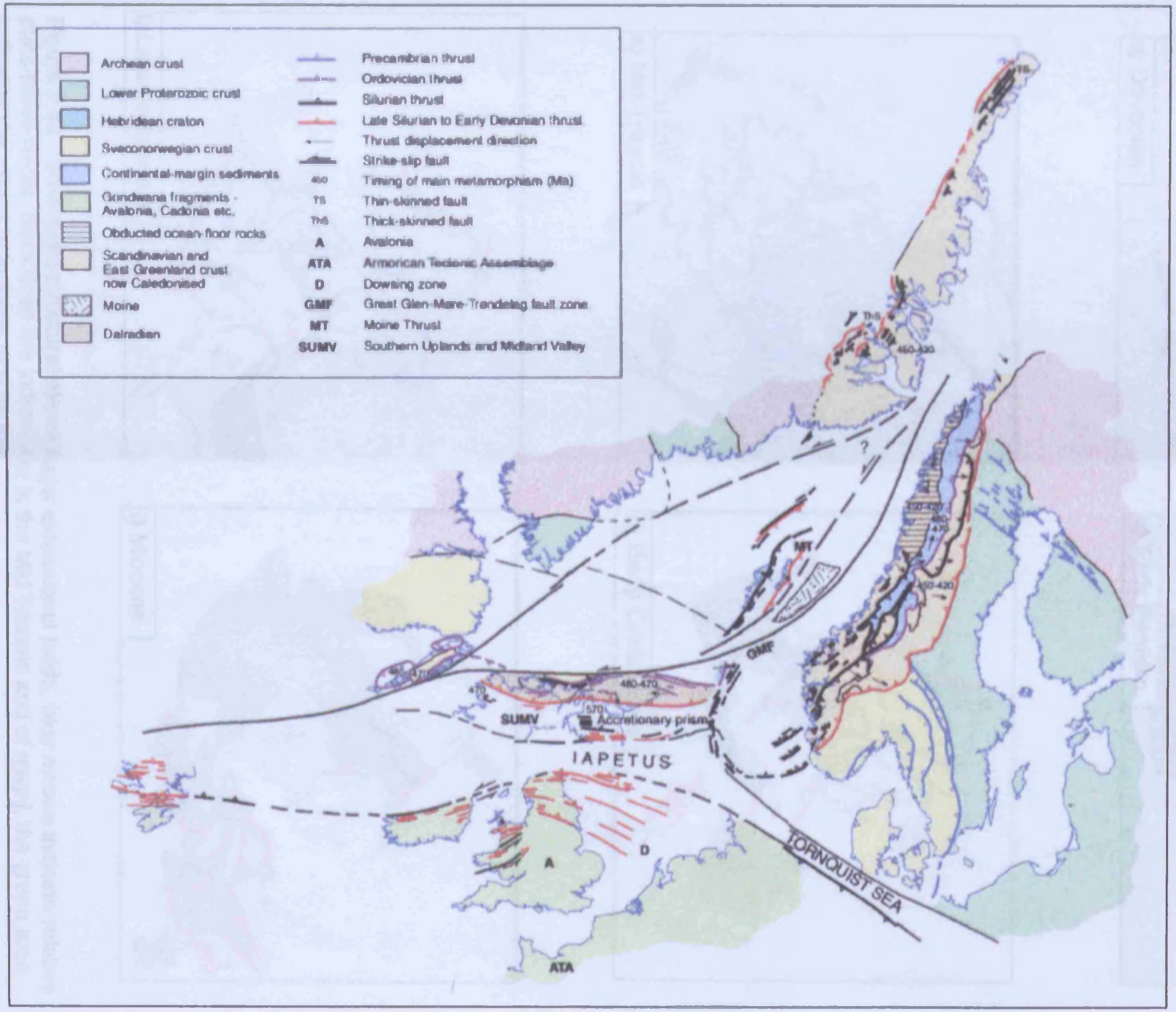


Figure 2.3. A palinspastic reconstruction of the Caledonides of the North Atlantic at approximately 430 Ma. Accretion of continental fragments and newer magmatic arc onto the Laurentian craton dominated the Caledonian and Variscan tectonic events. The accretion was southeast - northwest verging, with some important sinistral strike-slip movements along northeast trending faults (e.g. Great Glen - Møre - Trøndelag fault zone - GMF). The closure of the northern Iapetus between Baltica and Laurentia was caused due to the development of low angle thrust faults and left lateral strike-slip faults in the Scottish and Norwegian Caledonides. Nb: faults in Scandinavia are based on Hossack and Cooper (1986). Faults and terrane boundaries in Scotland and Ireland are based on Coward (1990), Dewey and Shackleton (1984), and Dewey and Mange (1999). Escher and Watt (1976) and Escher and Pulvertaft (1995) describe the position of the main thrusts in East Greenland. After Coward *et al.* (2003).

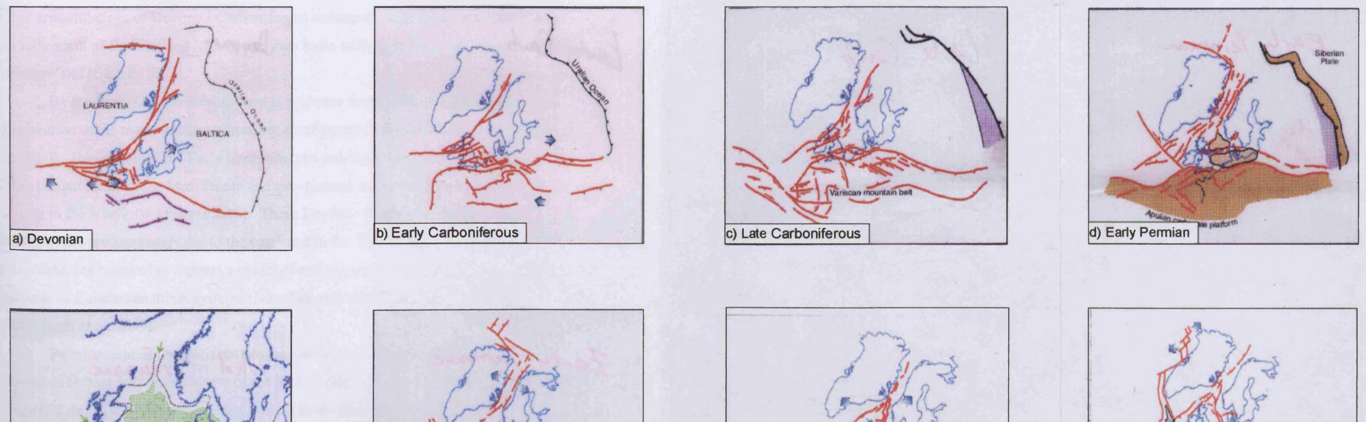


Figure 2.4a. Schematic structural elements maps of Northwest Europe showing major crustal lineaments active at various times throughout the Phanerozoic. Red lines indicate major extensional faults, blue arrows indicate relative plate movements, black lines are indicative of thrusts, purple colours shows areas of oceanic crust and brown colours show areas of uplift. Note, e) shows depositional environment in the Mid Triassic and of map j, the green area indicates the extent of Igneous activity associated with the Iceland plume. See text (section 2.2 for detailed discussion) After Coward *et al.* 2003.

Cartwright 1998). These extensional basins consist of hanging-wall depocentres for thick accumulations of Devono - Carboniferous sediments. The Sandwick Basin lying directly north of the Shetland Isles is one such basin with up to 2 km of sedimentary fill (Hitchen and Ritchie 1987).

By the Early Carboniferous, there is evidence from onshore Scotland of a reactivation in the major northeast trending shear-zones (Hartz *et al.* 1997, Ritchie *et al.* 2000). Deposition of the Early Carboniferous red-beds was slowly replaced by more deltaic facies in the Mid-Visean and experienced marine conditions by the Late-Visean to the Namurian (**Figure 2.4b**). These Devono - Carboniferous extensional faults, which predominately dip to the southeast in the West Orkney Basin and in the Clair field, are believed to support a model of extensional reactivation or gravitational collapse of Caledonian thrust systems (Duindam and Van Hoorn 1987, Earle *et al.* 1989, Dean *et al.* 1999).

Development of the Variscan orogeny in the Late Carboniferous caused intense folding in Britain with final closure of the proto-Tethys Ocean (Doré *et al.* 1999) (**Figure 2.4c**). This orogenic event is evident in the East Shetland Basin where compression and strike-slip movements of the earlier extensional faults are seen on seismic profiles (Platt and Cartwright 1998). The Northeast Atlantic margin was largely unaffected by this later tectonic activity, which will not be discussed in any more detail here.

2.2.3 The Late Palaeozoic to Recent

A second period of major crustal tectonics can be defined as Late Palaeozoic and Mesozoic - Cenozoic basin development. These latter basins formed during major periods of extension along structures that were strongly influenced by the pre-existing tectonic grain of Northwest Europe (Earle *et al.* 1989, Coward 1990, Doré *et al.* 1997; 1999). Significant thicknesses of sediment accumulated in these extensional basins offshore West Shetland, with over 6 km of succession of Mesozoic and Cenozoic recorded in the hanging-walls to major faults (Hitchen and Ritchie 1987). This Mesozoic - Recent period of tectonic events and associated sedimentation is sub-divided into individual geological periods which are discussed in greater detail in the following **Sections 2.2.3.1** to **2.2.3.8**.

2.2.3.1 Permo –Triassic

Formation of the supercontinent of Pangaea occurred at this time, with the closure of the Caledonian (Iapetus) and Variscan (proto-Tethys) Oceans. Non-marine deposition prevailed across the whole of Northwest Europe through much of the Permo - Triassic (**Figures 2.4d** and **2.5**). Thick accumulations of red-beds with occasional evaporites are seen to develop in rifted intermontane basins which trended northeast - southwest. At the same time, major salt basins occurred in the North Sea with an east - west trend (**Figure 2.4e**). These small rifted basins form half grabens which can be seen on seismic data in the Solan, West of Shetland and West Orkney Basins, (Mudge and Rashid 1987, Hitchen and Ritchie 1987, Duindam and Van Hoorn 1987), (see **Figure 1.1**). Furthermore, reservoir quality Triassic sandstones are found in the Strathmore discovery (in Block 205/26). The Shetland Spine Fault seems to be the southeastern limit of the Permo-Triassic deposits West of Shetland, although volcanics and continental deposits are found in the Midland Valley of Scotland and in Northern England (e.g. Francis 1983). The northeasterly trending rifting evidenced West of Shetland and in the Northern North Sea opened up an early seaway to the Arctic Ocean in the north (Ziegler 1990, Doré *et al.* 1999). This initial connection led to the development of the Zechstein Sea.

2.2.3.2 Jurassic

Basins of the West Shetland area show erosion of the early syn-rift Triassic rocks and this is thought to be due to post-Triassic inversion (Booth *et al.* 1993, Dean *et al.* 1999). This inversion may have continued into the Early Jurassic. Booth *et al.* (1993), for example suggested that there was removal of approximately 1.5 km of sediment from the West Shetland area. There is a suggestion of this inversion event being part of a more regional Mid-Jurassic unconformity seen out with the West Shetland area into the Hebrides Sea (Hazeldine and Russell 1987, Morton 1989) and could be related or restricted by the uplift caused by the North Sea Dome (Underhill and Partington 1993). Locally in the West of Shetlands a thin veneer of Middle to Upper Jurassic coastal and marine sediments overlie the areas of Pre-Cambrian basement and the Permo - Triassic red-beds (Duindam and Van Hoorn 1987, Earle *et al.* 1989). There is uncertainty about the tectonic origins of this regional unconformity

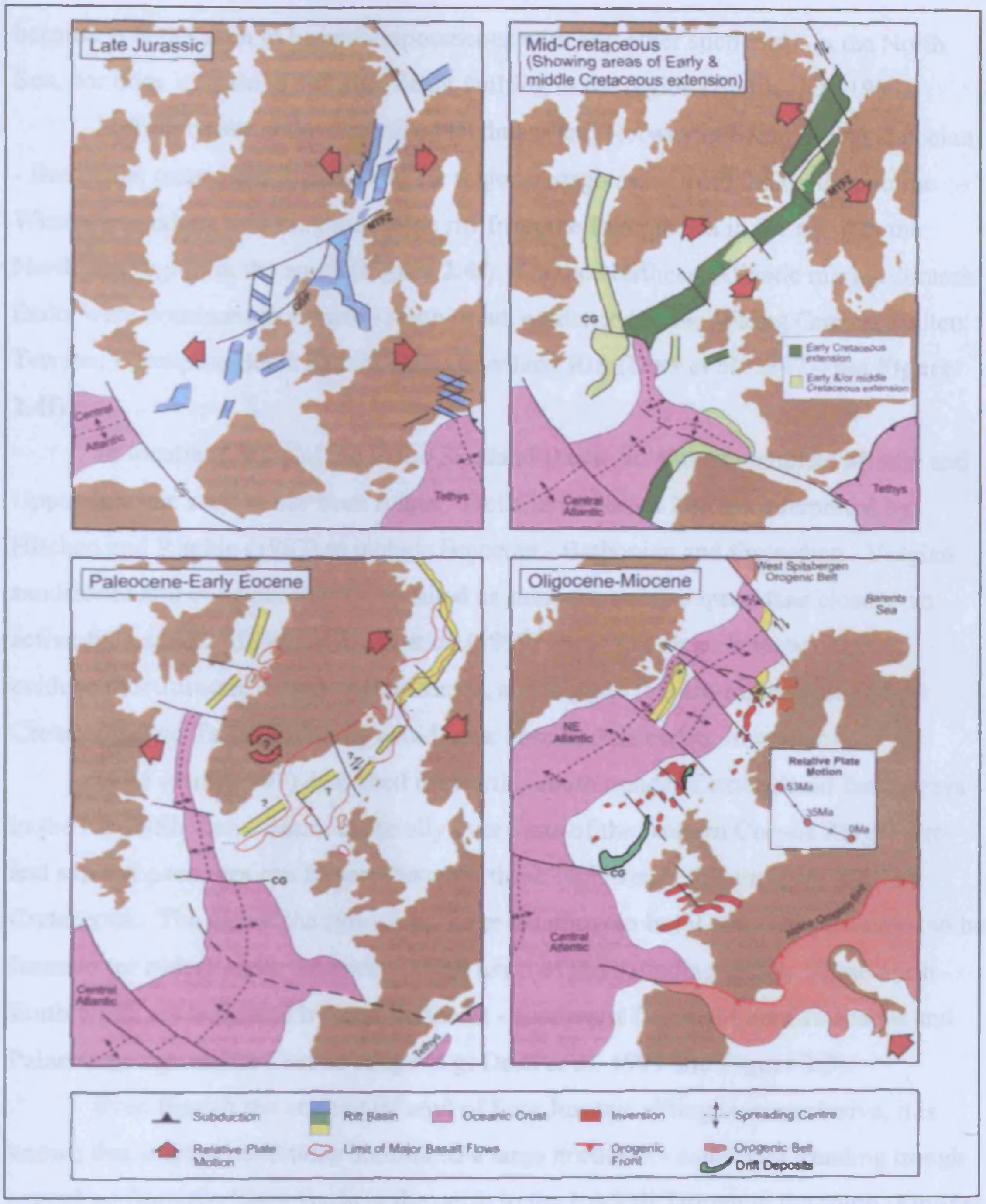


Figure 2.5. Plate reconstructions at four intervals from the Late Jurassic to the Miocene. a) Late Jurassic rifting was focussed on the North Sea and between Norway and Greenland. The rifts trended predominately north - south, though there was significant strike-slip components in the Faroe-Shetland Basin. b) In the Mid Cretaceous there was significant rifting taking place along the length of the Northeast Atlantic Margin in a northeast - southwest direction. Additional north - south rifting between Newfoundland and west Greenland occurred and oceanic spreading of the Central Atlantic had progressively moved northwards as far as the Charles Gibb Fracture Zone(CG). c) The reconstruction during the Palaeocene-Early Eocene shows oceanic spreading to the north of the Charles Gibb Fracture Zone for the first time between Newfoundland and West Greenland and some extension in the Faroe-Shetland Basin. Greenland began to rotate anti-clockwise and rifting was accompanied by voluminous magmatism. (The thin red line highlights the limit of major basalt flows). d) Oligocene-Miocene shows major oceanic spreading between East Greenland and Northwest Europe along the previous line of Cretaceous and Palaeocene rifting. A regional compressive regime occurred throughout Britain and offshore due to a squeezing effect between Alpine thrusting in the south and ridge push forces to the north. (After Dore *et al.* 1999). Nb: other abbreviations as in Figure 1.1 and text.

because it is not seen to be contemporaneous with any other such event in the North Sea, nor does it relate to any significant faulting in the region (Booth *et al.* 1993).

Rifting between the Greenland landmass and Norway initiated during Bajocian - Bathonian times, and this allowed for major transgression from the north, into the West Shetland area, linking the Arctic rift from the Barents Sea in the north to the North Sea region in the south (**Figure 2.4f**). On the Northeast Atlantic margin Jurassic faults were dominant in a north - south trend, evidenced in the Viking Graben, Halten Terrace, Porcupine Basin and the East Greenland Rift (Doré *et al.* 1997 – see **Figure 2.4f**).

In localised areas of the Faroe-Shetland Basin, an almost complete Middle and Upper Jurassic section has been found. Wells in Quadrant 206 are interpreted by Hitchen and Ritchie (1987) to include Bajocian - Bathonian and Oxfordian - Volgian sandstones and conglomerates, deposited as submarine slope apron fans close to an active fault scarp. However, Dean *et al.* (1999) argue that there is inconclusive evidence for Jurassic extensional tectonics, and suggest that later overprinting by Cretaceous and Tertiary activity could have obscured an earlier Jurassic rift.

Doré *et al.* (1997) described the north - south trends of extensional fault arrays in the Faroe-Shetland Basin, especially over parts of the western Corona Ridge. Pre- and syn-rift packages can be seen between these fault trends that pre-date the Late Cretaceous. The age of the syn-rift package is unproven but it could be postulated to be Jurassic (or older) given the north - south trend of the bounding faults. These north - south faults are truncated by later northeast - southwest faults of Late Cretaceous and Palaeocene age on the Corona Ridge (e.g. Dean *et al.* 1999 and **Figure 2.5**).

Even though the amount (if any) of Late Jurassic rifting is inconclusive, it is known that marine conditions dominated a large northeast - southwest trending trough extending from the Møre Basin in the north to the Rockall Trough in the south (**Figures 2.4 f & g** and **2.5**). Marine shales were deposited in this trough axis and the anoxic Kimmeridge Clay Formation has been found along its axis. These facies are the main source rock for hydrocarbons in the region, as is also seen in the North Sea. On the rift flanks, evidence suggests that there was deep erosion down to Permo - Triassic rocks, which locally were reworked into the basin centre, for example off the crest of the Rona Ridge.

2.2.3.3 Cretaceous

2.2.3.3.1 Early Cretaceous

At the onset of the Early Cretaceous, orthogonal extension was prevalent along the length of the Northeast Atlantic margin (e.g. Knott *et al.* 1993). This northeast - southwest trending rifting occurred in two main phases, creating what is now known as the Rockall Trough, the Faroe-Shetland Basin, the Magnus Basin and the most northerly Møre Basin (**Figures 2.5** and **2.4h**). Displacement was dominant during the Valanginian - Hauterivian which was then enhanced during the Aptian - Albian. To the south, and offshore Iberia, the Central Atlantic Ocean began to open, with sea-floor spreading beginning at 132 Ma (Coward *et al.* 2003). This spreading zone propagated north and east and bifurcated into the Porcupine Basin and Rockall Trough. Up to 70 km of crustal extension from the Barents Sea to Rockall has been estimated by Skogseid *et al.* (2000), with thinning of the crust to only a few kilometres in the Møre Basin identified by magnetic anomalies, interpreted as incipient seamounts (Lundin and Doré 1997, Olafson *et al.* 1992).

In the West Shetland area, rifting was seen on northeast trending structures, and vast thickness variations are observed within the Lower Cretaceous sediments (**Figure 2.6**). The evidence of discrete fault systems, which were active at different periods, created sub-basins that experienced both uplift and subsidence over a short period of time. In particular there is significant movement on the Westray Ridge, Shetland Spine Fault and Solan and Rona Faults (e.g. Dean *et al.* 1999). Seismic data indicate over 4500 m of Upper Cretaceous sediments are present on the downthrown (eastern) side of the Westray Ridge and in the Foula sub-basin (Lamers and Carmichael 1999).

Sedimentation during the Early Cretaceous was markedly different from the Late Jurassic, as the subdued topography of the West Shetland and Solan Basins in the Jurassic was replaced by a major structural enhanced topography which was created over a period of 70 Ma (Dean *et al.* 1999).

As rifting continued, isolated areas of uplift occurred on the rift flanks gave rise to thick accumulations of sand-rich facies in the hanging-walls of active faults scarps (**Figures 2.6** and **2.7**). The best example of this is seen in the hanging-wall of the Shetland Spine Fault. Thickness variations between 500 m and 1000 m of sand-grade material are seen in wells 205/30-1 and 205/25-1 respectively. These sandstones are interpreted as coalescing submarine fans and talus slope deposits, interbedded with

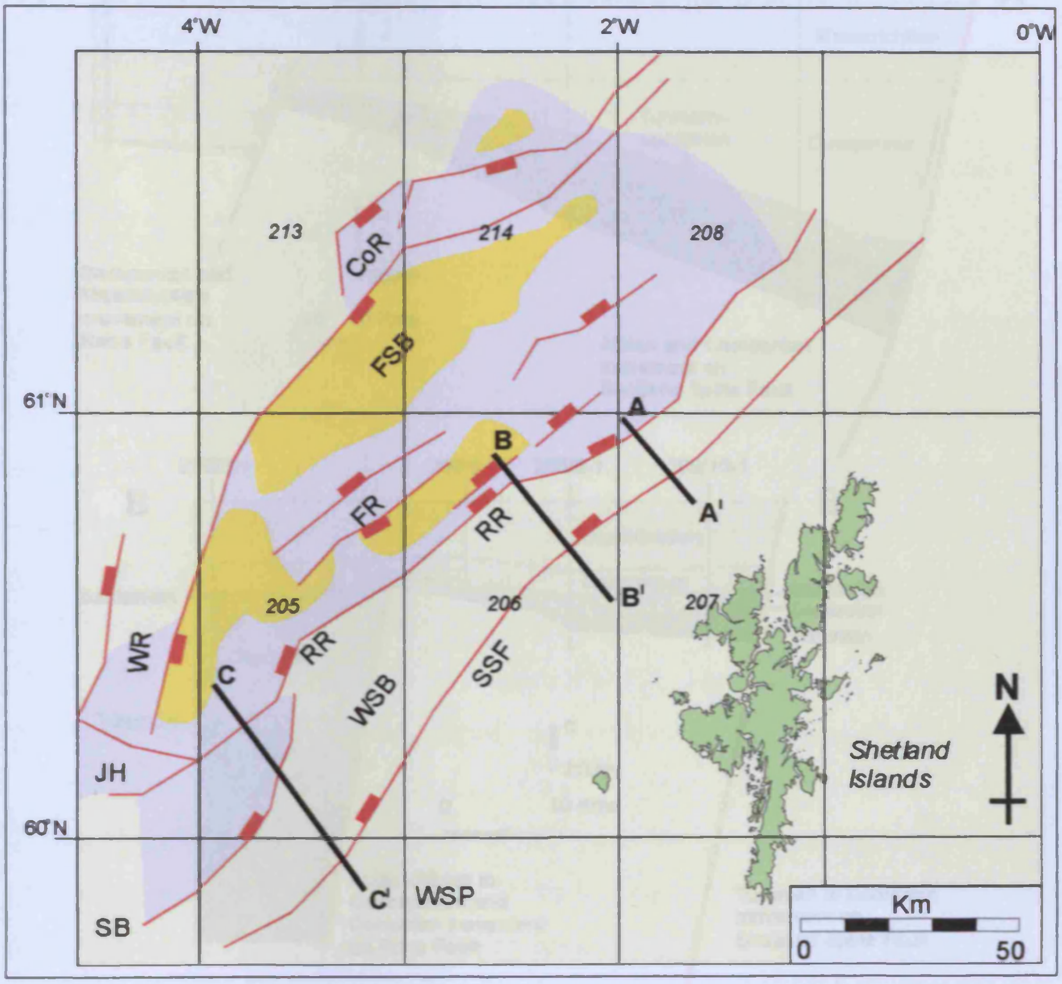


Figure 2.6. Structural elements map of the Faroe-Shetland Basin showing the distribution and orientation of the main Cretaceous rifts and depocentres. The Shetland-Spine Fault and Rona Ridge in particular experienced major periods of extension, and to the west and northwest, the Westray and Corona Ridges remained emergent. Grey shading= area of Cretaceous rifts, Yellow shading= area of main Cretaceous depocentres. Location of cross sections A', B' and C' of Figure 2.7. are shown. (redrawn from Dean *et al.* 1999).

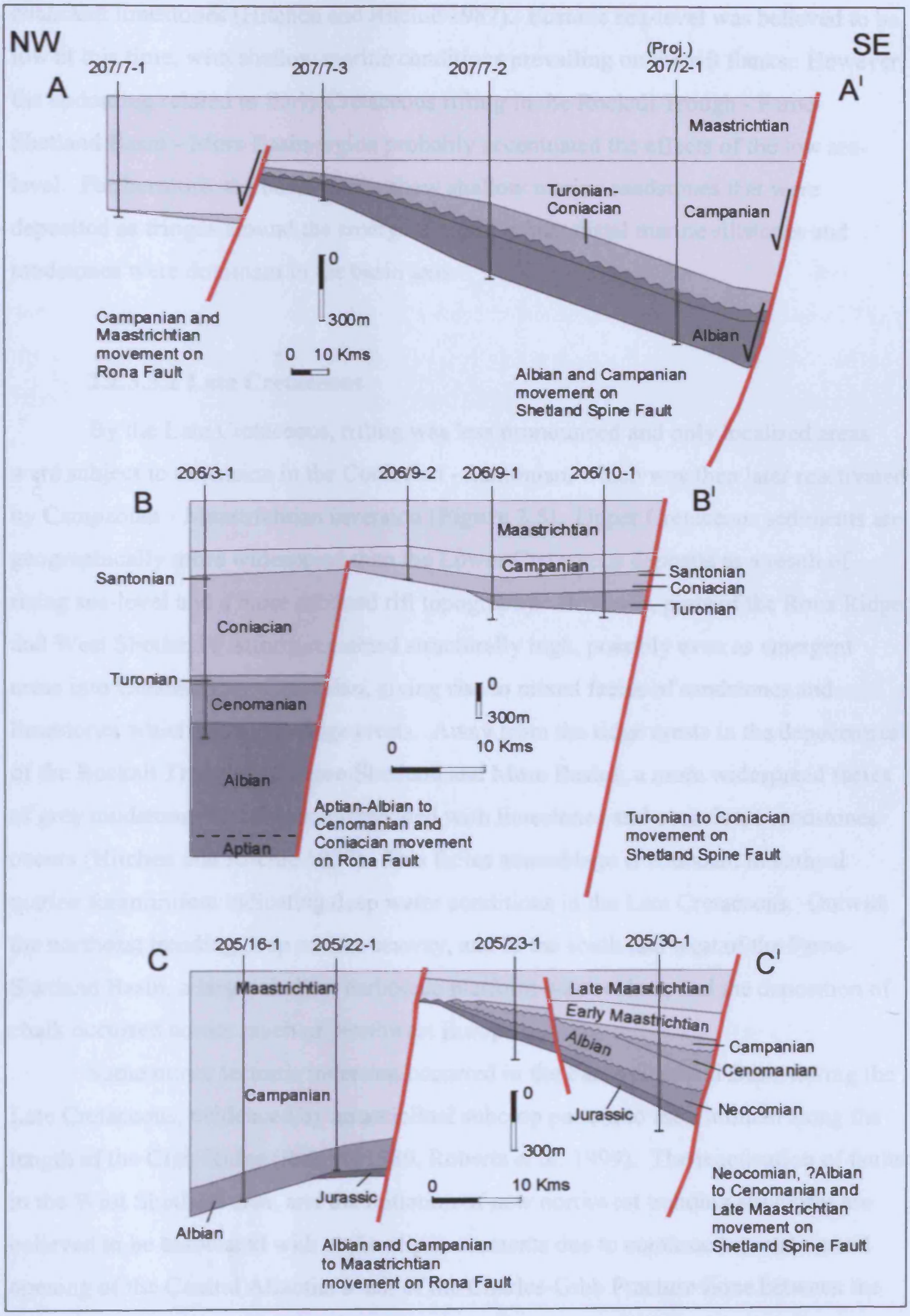


Figure 2.7. Schematic northwest - southeast cross sections from the West Shetland Basin showing the thickness variations of the Cretaceous strata on the Shetland Spine and Rona Faults. Note the difference in timing of relative movements on the faults from north to south (A' to C'). The sections have been hung on a 'datum' of base Tertiary. Locations of A', B' and C' are shown in Figure 2.6. (Redrawn from Dean *et al.* 1999).

coals and limestones (Hitchen and Ritchie 1987). Eustatic sea-level was believed to be low at this time, with shallow marine conditions prevailing on the rift flanks. However, the updoming related to Early Cretaceous rifting in the Rockall Trough - Faroe-Shetland Basin - Møre Basin region probably accentuated the effects of the low sea-level. Furthermore, the basin flanks show shallow marine sandstones that were deposited as fringes around the emergent highs, whilst distal marine siltstones and mudstones were dominant in the basin axis.

2.2.3.3.2 Late Cretaceous

By the Late Cretaceous, rifting was less pronounced and only localised areas were subject to extension in the Coniacian - Santonian, which was then later reactivated by Campanian - Maastrichtian inversion (**Figure 2.5**). Upper Cretaceous sediments are geographically more widespread than the Lower Cretaceous deposits as a result of rising sea-level and a more subdued rift topography. However, parts of the Rona Ridge and West Shetland Platform remained structurally high, possibly even as emergent areas into Cenomanian -Coniacian, giving rise to mixed facies of sandstones and limestones which onlap the ridge crests. Away from the ridge crests in the depocentres of the Rockall Trough and Faroe-Shetland and Møre Basins, a more widespread facies of grey mudstones and shales interbedded with limestones and occasional sandstones occurs (Hitchen and Ritchie 1987). This facies assemblage is abundant in bathyal marine foraminifera indicating deep water conditions in the Late Cretaceous. Outwith the northeast trending deep marine seaway, and to the south and west of the Faroe-Shetland Basin, a large, shallow carbonate platform was evident, and the deposition of chalk occurred across much of Northwest Europe.

Some minor tectonic inversion occurred in the Faroe-Shetland Basin during the Late Cretaceous, evidenced by an anticlinal subcrop pattern to the Turonian along the length of the Clair Ridge (Roberts 1989, Roberts *et al.* 1999). The reactivation of faults in the West Shetland area, and the initiation of new northwest trending structures are believed to be associated with strike-slip movements due to continued spreading and opening of the Central Atlantic, south of the Charles-Gibb Fracture Zone between the Labrador Sea and Iberia (Roberts *et al.* 1999, Coward *et al.* 2003).

This Late Cretaceous extension, inversion and associated subsidence in the Faroe-Shetland Basin with concomitant accumulations of thick sedimentary packages, and along with the Palaeocene fill helped bury the Upper Jurassic source rocks (notably the Kimmeridge Clay Formation). This led to early maturation of the kitchen area, which coincided with early migration into Cretaceous carrier beds (Lamers and Carmichael 1999, Holmes *et al.* 1999).

There is evidence to suggest that there was a minor amount of volcanic activity during the Late Cretaceous. Early work by Hitchen and Ritchie (1987) summarise the findings, noting tentative Campanian - Maastrichtian ages in Faroe-Shetland Basin wells. Similar ages are found in Helen's Reef on the Rockall Plateau (Pankhurst 1982), the Anton Dohrn seamount (Roberts *et al.* 1981) Rosemary Bank (Dietrich and Jones 1980) and basic dykes from East Greenland (Mitchell 1978). Earlier, igneous activity is possibly present on the Northeast Atlantic margin, with the Barra Ridge possibly being formed during early rifting of the Rockall Trough. A more detailed study of the associated magmatic and volcanic activity is summarised in **Section 2.3**.

2.2.3.4 Palaeocene

At the start of the Palaeocene there was a continuation of marine deposition in the Faroe-Shetland Basin from the Maastrichtian into Danian times. However, during the Palaeocene, between 65 Ma and 60 Ma there was believed to be initiation of a mantle plume below the lower crust, located somewhere beneath Greenland. The concept of mantle plumes being true geological phenomena has long been disputed and indeed at the present day the debate continues. Compelling arguments both favouring and rejecting the plume hypothesis abound and are nicely summarised in the recent debate in the Geoscientist Magazine 2003, (Volumes 5, 7, 8 and 9).

However, this proposed mantle plume is interpreted to have had a profound effect on the Northeast Atlantic margin, from the Late Palaeocene to the present day (e.g. White 1988, White 1992, Nadin *et al.* 1997, Jones *et al.* 2002). This mantle plume; known as the proto-Iceland plume, affected a vast area covering much of the Northeast Atlantic margin (**Figure 2.8**). Palaeocene uplift occurred over this large area (e.g. Nadin *et al.* 1997), in particular in mainland Scotland, the North Sea Basin and on the flanks of the Faroe-Shetland Basin and Rockall Trough. A voluminous amount of magmatic and volcanic activity was associated with the Iceland plume (e.g. Ritchie *et al.* 1999). The area of igneous activity is seen both onshore and offshore, and represents one of the largest igneous provinces (LIP's) in the world. Offshore, a vast

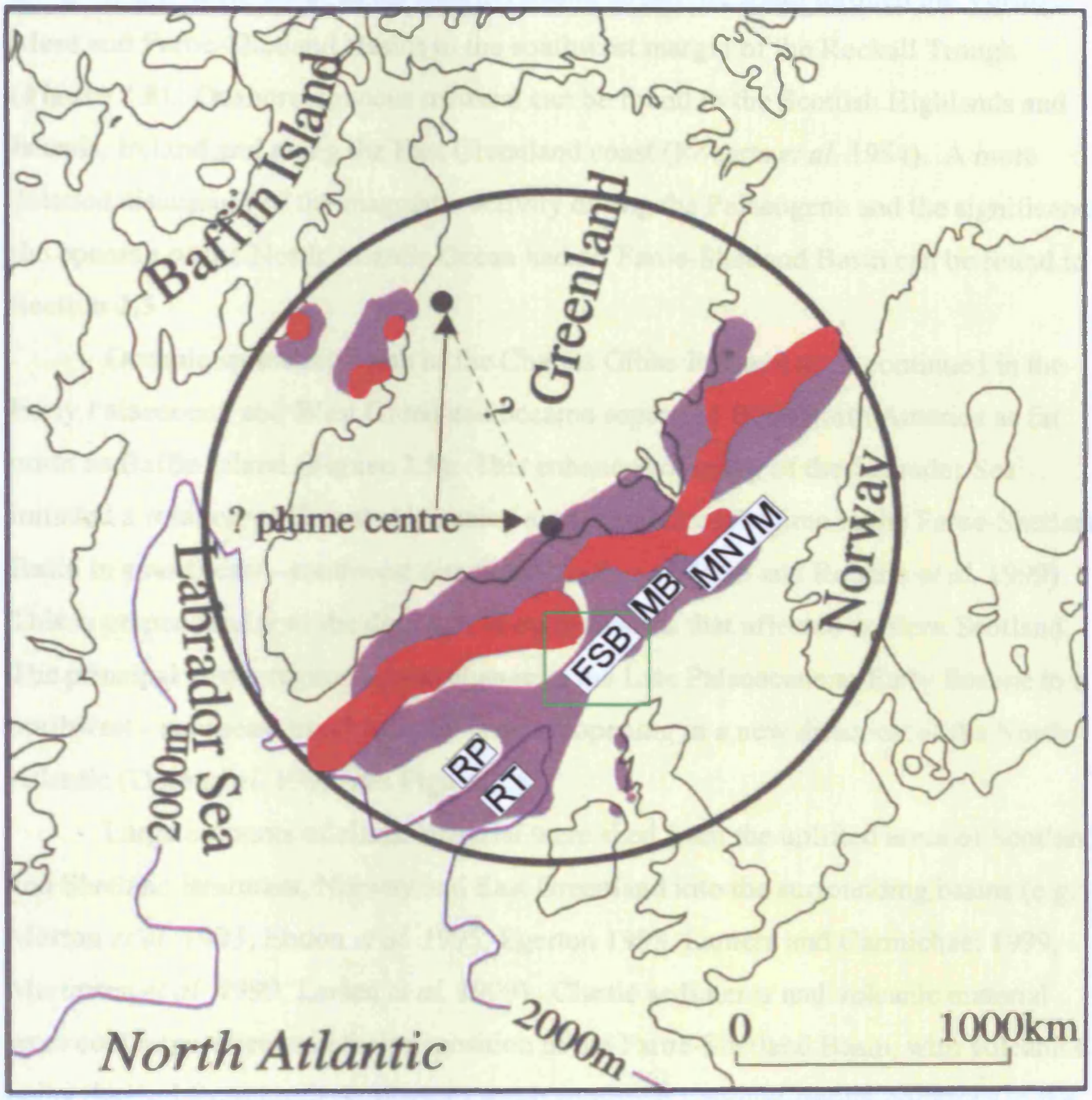


Figure 2.8. Plate reconstruction of the North Atlantic at the start of the Early Eocene (chron.24). The large circle shows the area affected by the Iceland mantle plume, with the possible positions of the plume centre at onset highlighted. The red shading represents the area of seaward dipping reflector sequences, and the light purple shows the area covered by flood basalts, sill complexes and central igneous complexes. RP= Rockall Plateau, RT= Rockall Trough, FSB= Faroe-Shetland Basin, MB= Møre Basin and MNVM= Mid-Norwegian Volcanic Margin. Green rectangle outlines Faroe-Shetland Basin which can be seen in Figure 2.18. (After Ritchie *et al.* 1999).

area affected by igneous material is seen, encompassing the East and West Greenland margins, and as far north as the Lofoten Basin, stretching south through the Vøring, Møre and Faroe-Shetland Basins to the southwest margin of the Rockall Trough (**Figure 2.8**). Onshore, igneous material can be found in the Scottish Highlands and Islands, Ireland and along the East Greenland coast (Roberts *et al.* 1984). A more detailed discussion of the magmatic activity during the Palaeogene and the significance the opening of the North Atlantic Ocean had on Faroe-Shetland Basin can be found in **Section 2.3**.

Oceanic spreading north of the Charles Gibbs Fracture zone continued in the Early Palaeocene and West Greenland became separated from North America as far north as Baffin Island (**Figure 2.5**). This enhanced opening of the Labrador Sea initiated a rotation of Greenland causing a principal stress regime in the Faroe-Shetland Basin in a northeast - southwest direction (Doré *et al.* 1999 and Roberts *et al.* 1999). This is perpendicular to the direction of dyke swarms that affected western Scotland. The principal stress regime was to change in the Late Palaeocene to Early Eocene to a northwest - southeast trend with the onset of opening in a new direction of the North Atlantic (Doré *et al.* 1999, see **Figure 2.5**).

Large amounts of clastic material were shed from the uplifted areas of Scotland and Shetland landmass, Norway and East Greenland into the surrounding basins (e.g. Morton *et al.* 1993, Ebdon *et al.* 1995, Egerton 1998, Lamers and Carmichael 1999, Martinsen *et al.* 1999, Larsen *et al.* 1999). Clastic sediments and volcanic material were contemporaneous in their deposition in the Faroe-Shetland Basin, with volcanics being derived from north northwest - south southeast trending fissure eruptions in the area of the Faroe Islands to the northwest of the basin (Naylor *et al.* 1999, Kiørboe 1999). Conversely, the clastic sediments were derived from the Scottish landmass to the southeast. Uplift over Scotland has been estimated to be in the region of 1 - 2 km by many authors (e.g. Rowley and White 1998, Thomson *et al.* 1999), increasing to the northwest nearing the plume centre, which was believed to be located in the region of central/eastern Greenland (Nadin *et al.* 1997 and **Figure 2.8**), though a dynamic support by a mantle plume is believed to be approximately 500 m on the Northwest European shelf (Jones *et al.* 2002).

Stratigraphic frameworks for the Palaeogene succession were first built in the late 1980's by integrating well, seismic and log data constrained by regional biostratigraphic zonation. A number of sequence-stratigraphic breakdowns of the

Palaeogene succession are known from the North Sea Basin. However, a regional basin-wide study of the Faroe-Shetland Basin remains speculative. Many schemes co-exist in the North Sea Basin, in particular Mudge and Bujak (1994) identified and calibrated (by dinocyst zonations) five sequences based on log correlations which used thin condensed claystones which correspond to high gamma log peaks to sub-divide the sequences. These sequences correspond closely to the stratigraphic sequence of Galloway (1989). Jones and Milton (1994) and later Ebdon *et al.* (1995) produced a sequence-stratigraphic interpretation (The so-called T-scheme) to position the Palaeogene deep water sandstones seen in both West of Shetland and in the North Sea. Neal (1996) developed a sequence-stratigraphic scheme in the North Sea Basin inferring a cyclical pattern to sedimentation by recognising five major regressions and intervening transgressions of relative sea-level in the Palaeogene.

Furthermore, lithostratigraphic schemes were developed most notably by the British Geological Survey (Knox and Holloway 1992, Knox *et al.* 1997). Biostratigraphic zonation schemes are numerous from both the Faroe-Shetland Basin and the North Sea Basin and show subtle variability depending on the individual company. A summary of the main lithostratigraphic, sequence-stratigraphic and biostratigraphic schemes used both in the North Sea Basin and in the Faroe-Shetland Basin is shown in **Figure 2.9** which is calibrated to the timescale of Berggren *et al.* (1995).

The so called 'T' - scheme, pioneered by BP and Shell, and first published by Jones and Milton (1994) and Ebdon *et al.* (1995) (see **Figure 2.9**), became the acknowledged way to sub-divide the different sand-rich packages found in the depocentres. In the Faroe-Shetland Basin a semi-regional Palaeocene palaeogeographic map can be seen in **Figure 2.10**, which shows small sub-basins which were developed in the Early Palaeocene and filled the inherited Cretaceous rift topography. Two main basins are defined by the palaeogeographic map; the Flett sub-basin and the Foinaven sub-basin which trend northeast - southwest and lie basinward of the structurally high Rona Ridge, where the Palaeocene section thins to less than 200 m on its crest. Both the Flett and Foinaven sub-basins are believed to represent post-rift thermal sag basins, with possibly some fault control active in the Palaeocene (Dean *et al.* 1999, Lamers and Carmichael 1999). Turner and Scrutton (1993) propose an Early Palaeocene rifting event followed by a Late Palaeocene thermal subsidence phase to explain the development of local basins. These sub-basins lie between the Rona and Flett Ridges

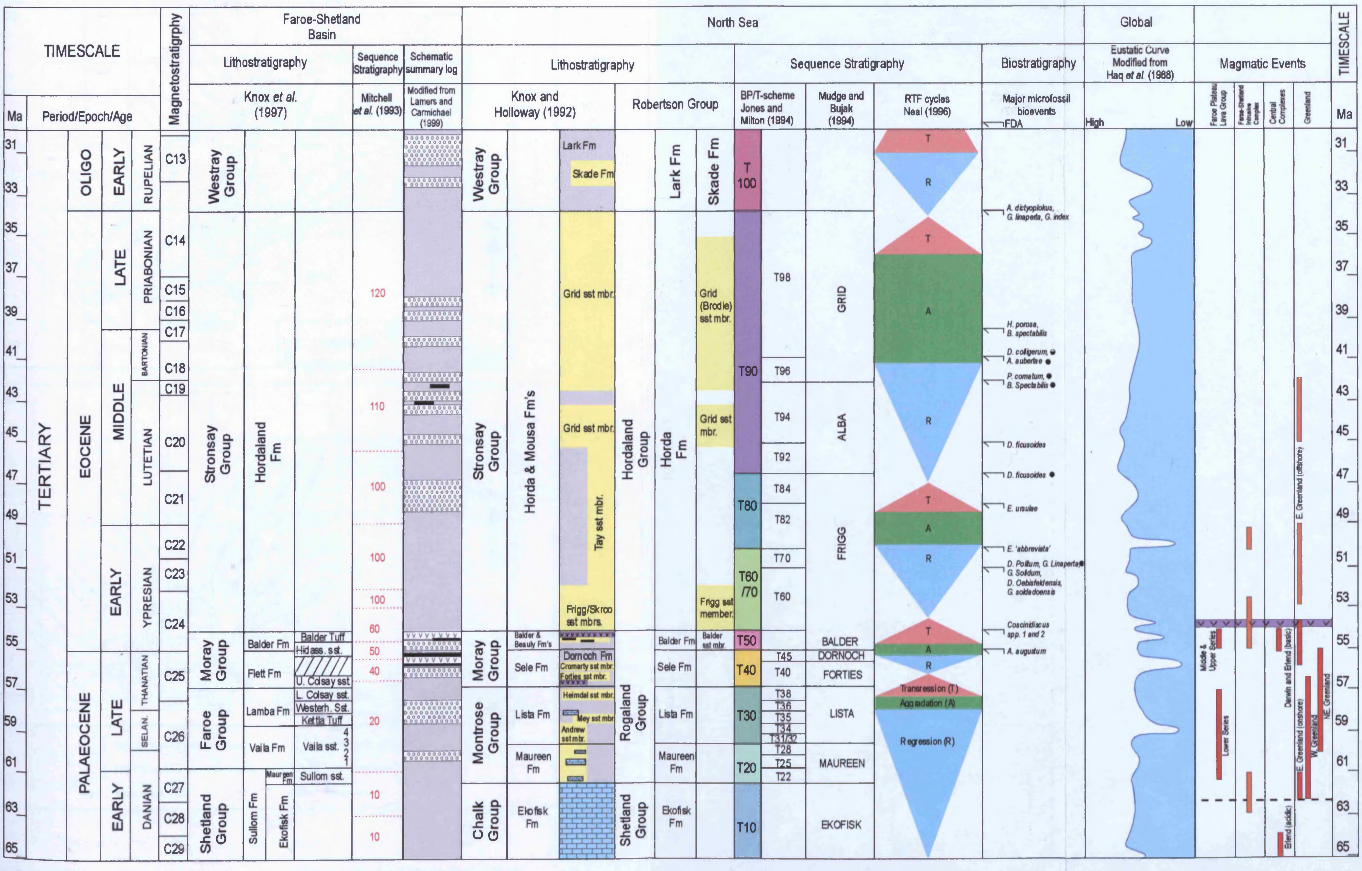


Figure 2.9. Table summarising the lithostratigraphy, sequence stratigraphy and biostratigraphy of the Palaeogene succession in the Faroe-Shetland Basin and the North Sea. Highlighted are the main schemes used to subdivide the stratigraphy. Age range taken from time-scale of Berggren *et al.* (1995), the magnetostratigraphy is also shown highlighting the age of the individual chrono-zones (from Cande and Kent 1995). The sequence stratigraphic part of the chart is based on the North Sea using, Mudge and Bujak (1994), Jones and Milton (1995), and Neal (1996). In the Faroe-Shetland Basin the sequence stratigraphic study shown is that of Mitchell *et al.* (1993). The lithostratigraphy is from Knox *et al.* (1997) for the Faroe-Shetland Basin, along with a schematic lithology log modified form Lamers and Carmichael (1999). In the North Sea, the lithostratigraphic schemes are taken from Knox and Holloway (1992) and the Robertson Group. Key publications used for the main microfossil bioevents are: Bujak *et al.* (1980), Knox and Morton (1983), Powell (1988), Gradstein *et al.* (1992), Bujak and Mudge (1994), Mudge and Bujak (1994; 1996b), Jolley (1996). The coastal onlap curve of Haq *et al.* (1988) is also shown. Finally a summary of the magmatic events from the Northeast Atlantic Margin (modified from Ritchie *et al.* (1999). Orange bars = intrusives, red bars = extrusives.

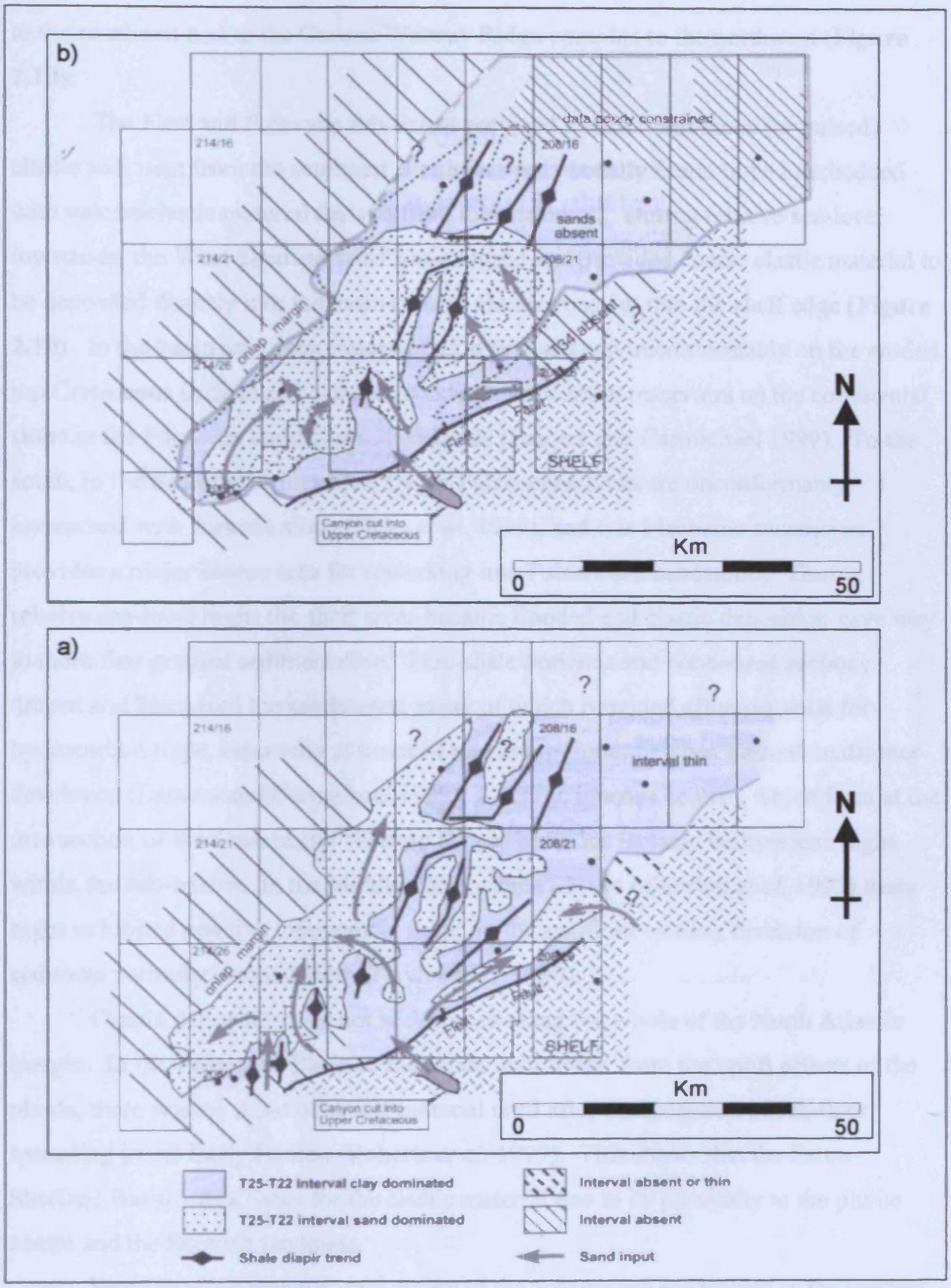


Figure 2.10. Palaeogeographic maps of the Flett sub-basin. a) (bottom) Early Palaeocene (Danian - T22-T25 times) and b) Late Palaeocene (early Thanatian times - T31-T34). Note the distribution of sands in the basin axis and the absence of sediments on the Westray Ridge to the west. The shelf break remained the same position and cut canyons into the Upper Cretaceous strata on the Flett Ridge. Differential compaction caused diapirism in the regional shale seals. (After Lamers and Carmichael 1999).

to the southeast and to the Corona/Westray Ridge complex to the northwest (**Figure 2.10**).

The Flett and Foinaven sub-basins received almost continuous (or pulsed) clastic sediment from the southeast which was occasionally found to be interbedded with volcaniclastic material derived from the northwest. During relative sea-level lowstands, the West Shetland shelf was exposed and provided coarse clastic material to be deposited directly into the basin centres via canyons cut into the shelf edge (**Figure 2.10**). In the basin axis deep water turbidite deposits rest unconformably on the eroded top Cretaceous surface and occur as ponded hydrocarbon reservoirs on the continental slope in the Foinaven and Schiehallion fields (Lamers and Carmichael 1999). To the south, in the Sea of Hebrides area, Upper Palaeocene lavas are unconformably juxtaposed with Jurassic strata (Dean *et al.* 1999), and this Mesozoic succession provides a major source area for reworking into Palaeocene sandstones. During relative sea-level highs the shelf areas became flooded and clastic deposition gave way to more fine grained sedimentation. Thin shale horizons and condensed sections draped and blanketed the sandstones, many of which provided effective seals for hydrocarbon traps, especially at times of maximum flooding when basinal mudstones developed (Lamers and Carmichael 1999). Locally, igneous centres, which form at the intersection of the Corona and Westray Ridges provided isolated bathymetric highs within the sub-basins. In the Middle Palaeocene (T31-34 of Ebdon *et al.* 1995) these highs exhibited positive topographic relief on the sea-floor causing diversion of sediment pathways around them (Naylor *et al.* 1999).

Clastic deposition was not widespread along the whole of the North Atlantic margin. In the Porcupine Basin to the south, and further from the uplift effects of the plume, there was no input of clastic material until after the inception of sea-floor spreading in the Early Eocene (Roberts *et al.* 1999). This shows that the Faroe-Shetland Basin was a focus for the clastic material due to its proximity to the plume centre and the Scottish landmass.

From detailed mapping and dating of the Palaeocene sandstones in the various sub-basins, it is possible to correlate periods of major clastic input with peaks in the amount of magmatic activity. Three periods of clastic input have been identified at approximately 62 Ma, 58 Ma and 56-55 Ma, reflecting the pulsed nature of the Iceland plume (White and Lovell 1997, Jones *et al.* 2002, Clarke 2002 unpublished PhD. thesis). Therefore for the first time a possible link between magmatic processes, dynamic uplift and sedimentary response could be postulated (White and Lovell 1997) **(Figure 2.11).**

Throughout the Palaeocene and most significantly in the latest Palaeocene the available accommodation space in the basin became progressively filled. By the time of deposition of the T35-36 units (Early Thanatian), at the southern end of the sub-basins, the shelf-slope break, which was previously controlled by the position of the Rona Ridge, began to move towards the northwest. Shelf sandstones, rich in bryozoans, were found basinward of the Rona Ridge towards the north of Quadrant 205. Sediment was still able to reach the basin centres through well established entry points in the shelf edge (e.g. 204/23-1 described first by Ebdon *et al.* 1995), **(Figure 2.10)**. This northwest movement of the facies belts signified an important part of the Palaeocene history of the Faroe-Shetland Basin. By the Middle Thanatian (BP's T34-T36 times – see Ebdon *et al.* 1995) there was a relative sea-level rise which flooded back over the basin margins, leading to the abandonment of basin floor fans (Jones and Milton 1994). Regional seals developed and transgression of the basin margin followed. The basin fill architecture changed from predominantly aggradational to largely progradational as the basin subsided and this can be related to a period of inactivity of the mantle plume (Naylor *et al.* 1999).

A renewal in mantle plume activity and therefore volcanism occurred in the Late Thanatian (BP's T40-T50) and resulted in major uplift of the basin margin and peneplanation and complete removal of the previous sequences (T36) on the shelf. Deposition at this time was confined to the north of the basin and basin floor fans were deposited west and north of the Westray Ridge and into the Faroe-Shetland Basin for the first time. Deltaic deposits prograded out across the basin and marine conditions only prevailed north of the Quadrant 213 and 214. In this northern area of marine deposition the well defined lava edge of the Faroe-Shetland Escarpment defined the north and west margins of the marine depocentre (**Figure 2.12**). This basaltic escarpment is believed to correlate with the Middle or Upper Series on the Faroe Islands, and shows a distinctive thick sequence of prograding seismic reflectors (Naylor *et al.* 1999). This striking and unique feature has been interpreted as a large lava delta which was formed by the eruption of sub-aerial Faroe lavas into a marine basin and hence was a palaeo-shoreline at the start of the Eocene. The magnitude of the foresets,

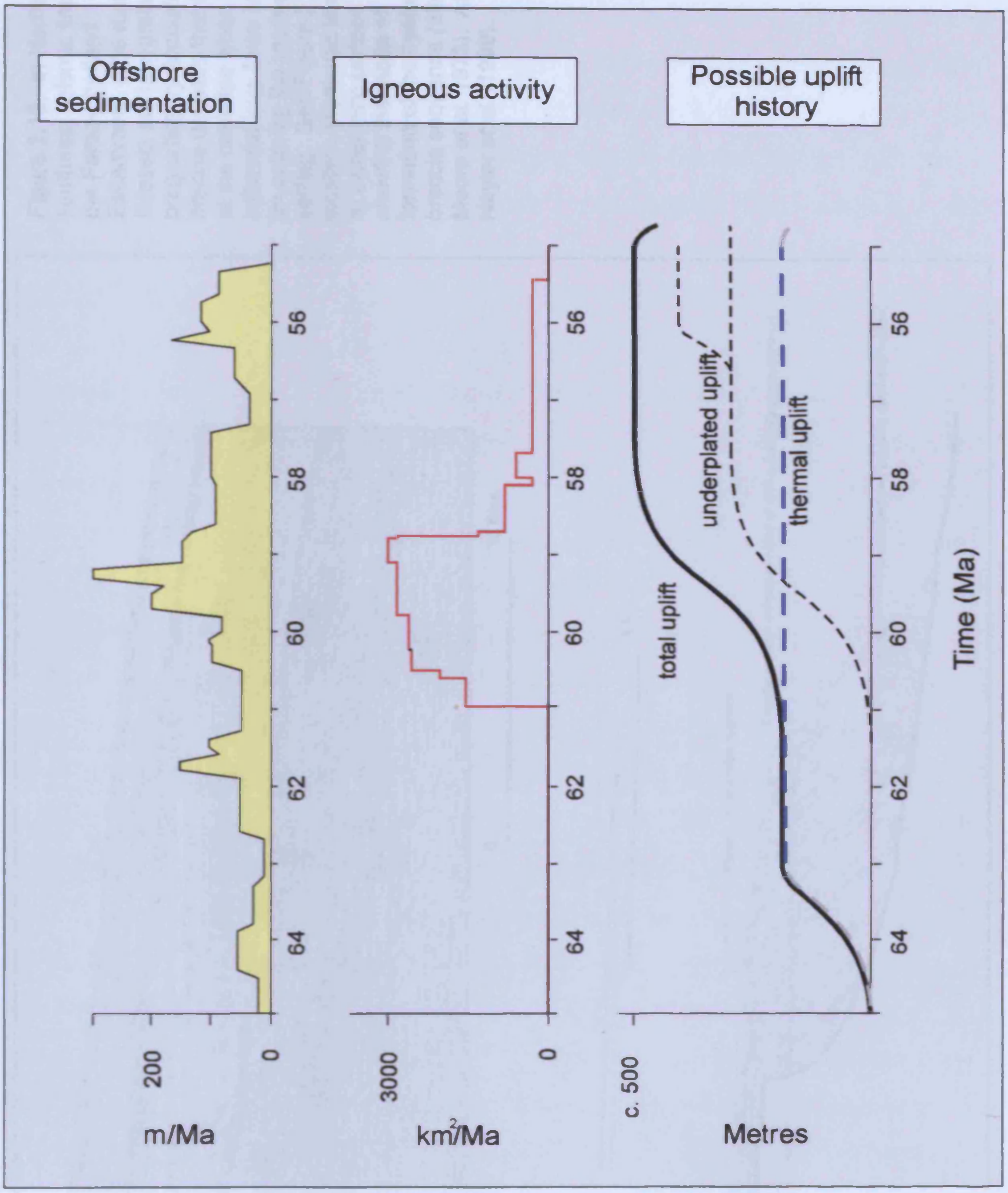


Figure 2.11. Summary graph showing comparison with the offshore sedimentation rate and the timing of peak magmatism through time. The timing of uplift, constrained by dated surface igneous activity is also shown for three case scenarios. This uplift coincides with peak sedimentation rates observed offshore Scotland. Thus a causal link between igneous related uplift by underplating and periods or pulses of clastic deposition was invoked. Modified from Clarke 2002 (unpublished PhD. thesis, Cambridge).

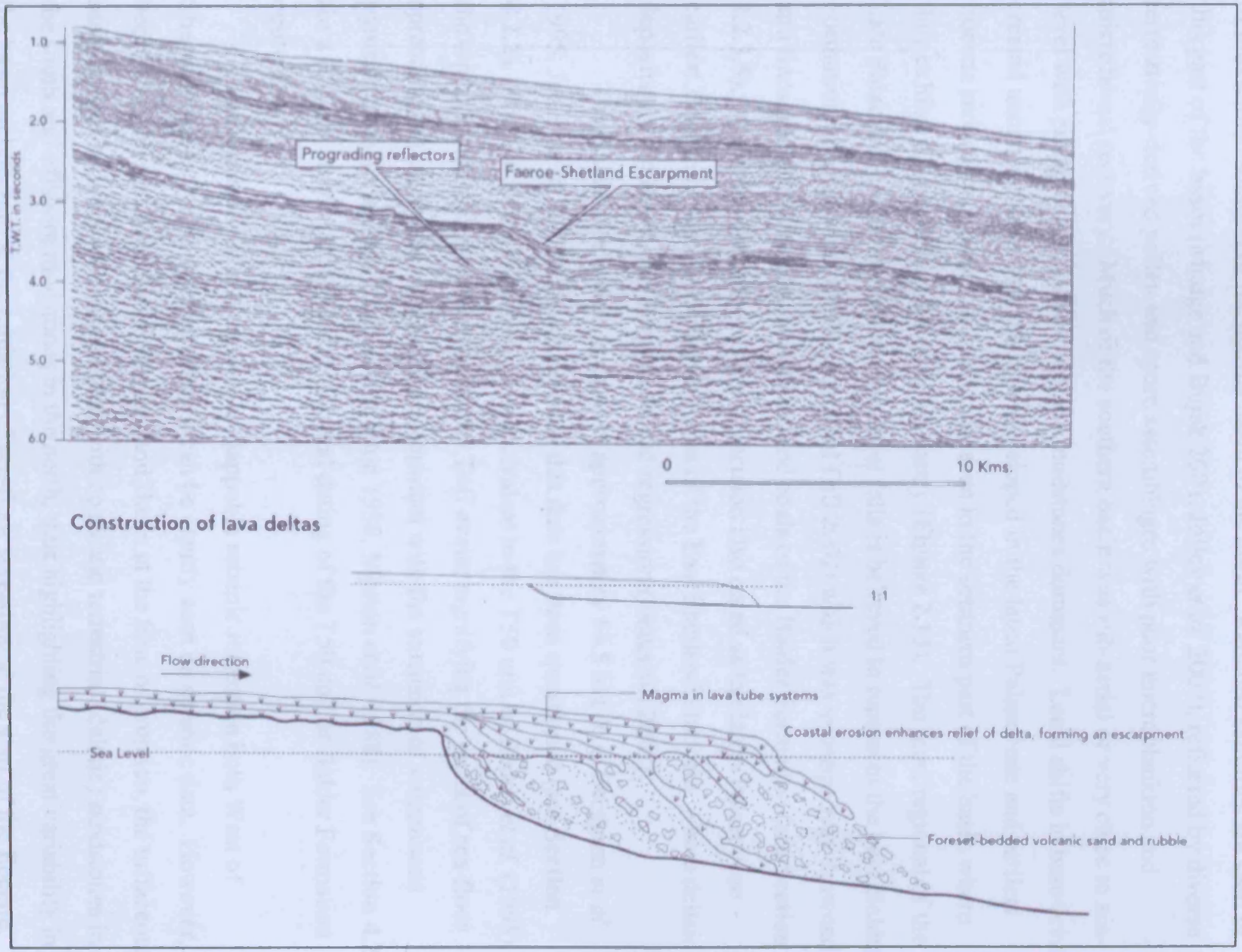


Figure 2.12. a) Northwest-southeast seismic line across the Faroe-Shetland Escarpment. The dipping foresets are interpreted as prograding hyaloclastite breccia deposits that formed at the coastline when subaerial lava flows cooled on entering the marine setting. See Figure 2.14 for location of seismic line.
b) Schematic cartoon showing the mode of formation of the hyaloclastite breccia sequence (after Moore *et al.* 1973). After Naylor *et al.* (1999).

and hence the approximate height of the escarpment gives a good indication of the variable water depths into which the lava prograded (Naylor *et al.* 1999). Thick basinal mudstones were deposited as lateral equivalents at the same time as prograding Faroe basalt lavas of the escarpment (Smythe 1983, Smythe *et al.* 1983).

By the end of the Palaeocene and into the earliest Eocene deltaic conditions prevailed over much of the southern part of the Faroe-Shetland Basin, marking a period of maximum isolation of the marine basin to the north. Reduced salinities are seen in this part of the basin (Mudge and Bujak 2001, Jolley *et al.* 2002), reflected by diverse terrestrially-derived pollen and spore assemblages with poor microplankton and microfaunal recovery. Much of the southern basin was sub-aerial or very close to sea-level with coals, deltaic sandstones and mudstones dominant. Local shifts in base-level created unconformities and channels developed in the latest Palaeocene and earliest Eocene sediments. These can be clearly seen in the southern part of the basin where they exhibit a pronounced dendritic geometry (**Figure 2.13**). The most regional of the Late Palaeocene – Early Eocene base-level falls is believed to represent the base Balder Formation unconformity (Smallwood and Gill 2002) which was subsequently drowned and transgressed with tuffs, sandstones and coals of the Balder Formation (see **Section 2.2.3.5**). A similar dendritic pattern of incision also dated as the latest Palaeocene – earliest Eocene is seen in the Bressay area of the East Shetland Platform, where deltaic deposits are cut down into during a forced regression (Underhill 2001).

The Balder Formation is dated at approximately 54.5 Ma (e.g. Berggren *et al.* 1995, Jolley *et al.* 2002) though recently this date has been questioned (see **Section 4.2.2**). The Balder Formation is time equivalent to the T50 unit of Ebdon *et al.* (1995) the top of which is defined by the Balder Tuff event; signifying the onset of sea-floor spreading in the region, which was concomitant with the extrusion of voluminous pyroclastic material (e.g. Knox and Morton 1988, Morton *et al.* 1988). See **Section 4.2** for a full discussion on the description and dating of the T50 unit or Balder Formation equivalent.

The Balder Tuff is a regionally mappable seismic reflection both West of Shetland and in the North Sea, where it can be clearly seen on seismic data. However, because of the nature of the Faroe-Shetland Basin at the time of eruption, the tuffaceous material can be seen to be interbedded with coals and terrestrial (deltaic) sandstones in the south and offshore mudstones in the north, thus highlighting the great variability in the basin architecture at the initiation of sea-floor spreading at the start of the Eocene.

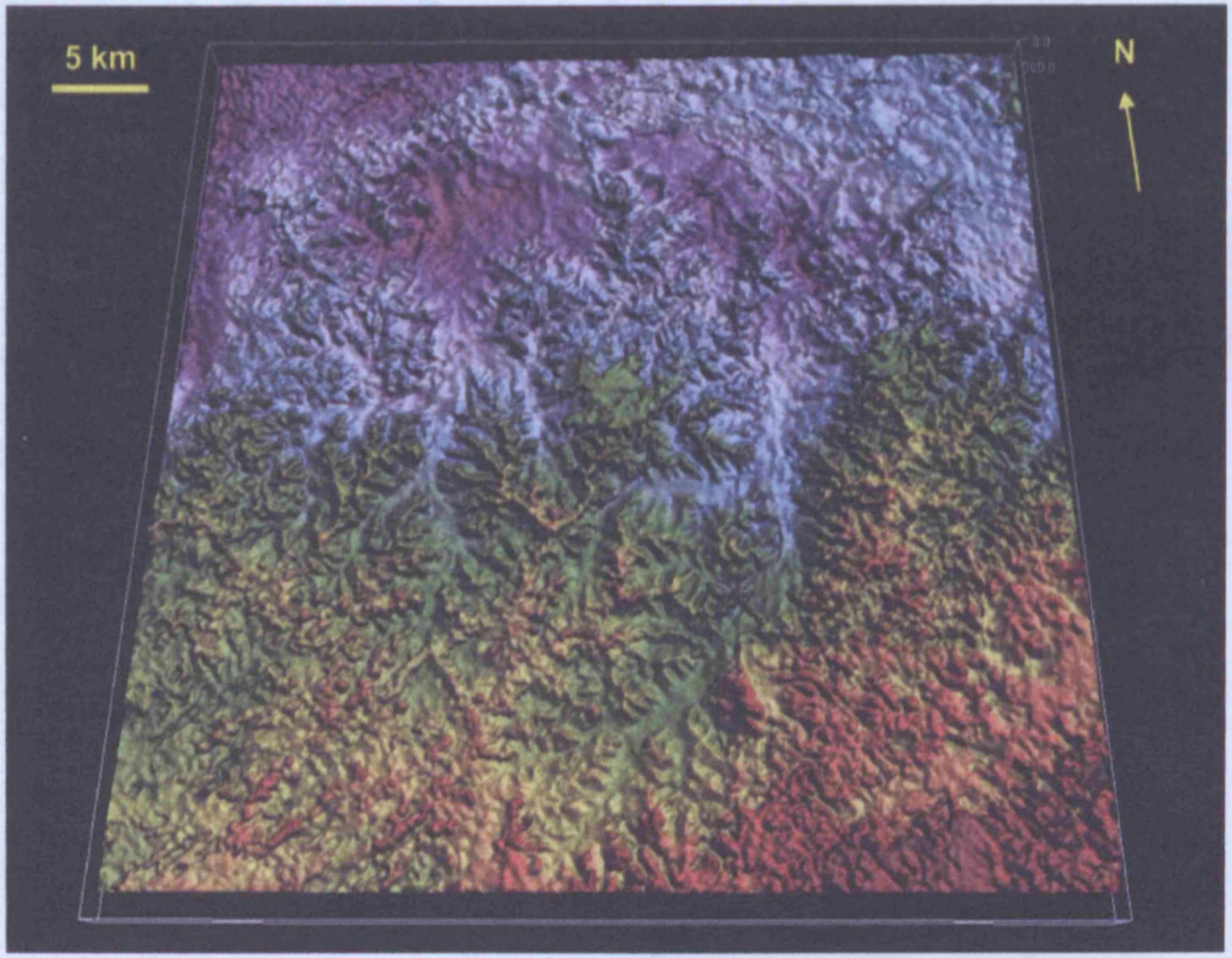


Figure 2.13. Three - Dimensional (3-D) image of depicting the base T50 (Balder) unconformity in the Judd Area in the south of the Faroe-Shetland Basin. Red to blue colours indicates thins to thicks of the T50 unit of Ebdon *et al.* (1995) (Balder Formation equivalent) isochore with a range of 20 to 320 m. The image nicely illustrates a dendritic valley drainage pattern that drained northwards into the basin. (After Smallwood & Gill 2002). For location of the map see Figure 2.14.

2.2.3.5 Eocene

The southward transgression which started during the deposition of the Balder Formation which corresponds to the T50 unit of Ebdon *et al.* (1995), continued with pace throughout the early part of the Eocene. The basin architecture was such that during deposition of the T50 unit, deep water conditions (500 m plus) were prevalent in the north of the basin where the Faroe-Shetland Escarpment bounded the depocentre to the north and west (Kiørboe 1999), whereas shallow and marginal marine conditions were dominant in the south.

In the southern part of the basin, Smallwood and Gill (2002) recognised clinoforms of Ypresian (Early Eocene) age, which downlapped towards the north. These clinoforms indicate that water depths increased enough during the Early Eocene transgression to allow for progradation of deltaic systems across a newly developed shelf. Whilst shallow marine conditions prevailed in the southern part of the basin, in the north deep water mudstones and siltstones continued to accumulate in a starved depocentre. These sediments are interbedded with the T50 unit (Balder Formation) tuffs, and show no sign of clastic input from the margins. One explanation of this vast difference in basin geometry and depth at the start of the Eocene is that hot athenosphere was placed at the base of the crust in the form of a transient mantle plume, which quickly moved away from the southern Faroe-Shetland Basin towards the northwest forming the present day Iceland-Faroe Ridge (e.g. White 1989). The evidence of a mantle plume beneath the lithosphere and the effects that it brings on a subsiding basin is discussed in more detail in **Section 2.4**.

Some attempts to sub-divide the Eocene succession in the Faroe-Shetland Basin have taken place in the past decade (Mitchell *et al.* 1993, Nielsen and Van Weering 1998). Three phases of cyclic sedimentation have been postulated by identifying unconformities that separate genetically related packages in the basin (Mitchell *et al.* 1993). Alternate slow and rapid subsidence rates in the Early - Middle and Late Eocene have been supported by apparent active rifting in adjacent basins along the Atlantic margin and the development of sand-prone basin floor deposits respectively.

In the area of the Faroe Islands, Lower Eocene tuffaceous limestones have been dredged from the sea-floor of the Faroe Platform (Waagstein and Heilmann-Clausen 1995). Bartonian and Priabonian dinoflagellates have been found in Lower Oligocene sediments around the northern and eastern Faroes, suggesting a relatively shallow marine setting over the Faroe Islands in the Middle - Late Eocene which was subsequently uplifted by the earliest Oligocene that allowed for some reworking and deposition (Waagstein 1988, Waagstein and Heilmann-Clausen 1995).

2.2.3.6 Palaeogene – Neogene Compression

After the final phase of extensional tectonism in the Late Cretaceous - Early Palaeocene had ceased with the onset of sea-floor spreading in the North Atlantic, the Faroe-Shetland Basin began to undergo thermal relaxation in the form of post-rift subsidence. This subsidence is punctuated by Late Palaeocene - Early Eocene compression across the whole of the Northeast Atlantic margin and particularly the Faroe-Shetland Basin (Boldreel and Anderson 1993; 1994). In the Faroe-Shetland Basin the areas of compression are coincident with the axis of transfer zones e.g. the Clair and Judd transfer zones (Rumph *et al.* 1993). This compressional activity is believed to be associated with ridge push forces due to the opening of the North Atlantic and production of the first oceanic crust during the Eocene at 55 Ma (Doré *et al.* 1997; 1999). The main compressional features seen are broad domes along the entire North Atlantic margin (e.g. Deidre's Arch and Helland Hansen Arch) and smaller localised pop-up features and rollover anticlines (Boldreel and Anderson 1993; 1994, Cloke *et al.* 1999, Davies *et al.* 2004 and **Figures 2.5** and **2.14**). The interaction between the northwest - southeast transfer zones and the predominant northeast - southwest Cretaceous and Palaeocene rift trend created sediment entry points for Palaeocene sediments to travel to the centre of the basin, most notably in the Foinaven and Flett sub-basins (Lamers and Carmichael 1999 – and **Figure 2.10**).

In the area of the Faroe-Shetland Basin, the main compressional features seen are the Wyville-Thompson Ridge, the Ymir Ridge and the Munkagrunnur Ridge (**Figure 2.14**). These features trend between west northwest - east southeast, with the Wyville-Thompson Ridge complex dividing the Faroe-Shetland Basin from the North Rockall Trough to the south. Additionally, there are many smaller compression anticlines that trend in a northeast direction (Cloke *et al.* 1999, Davies *et al.* 2004).

The origin of the Wyville-Thompson Ridge created much debate. It was originally believed to represent a volcanic pile erupted onto earlier Palaeocene sediments and ocean crust, resulting from fissure eruptions that aligned with the ridge (Roberts *et al.* 1983). Bott (1984) proposed an origin of volcanic loading on a belt of

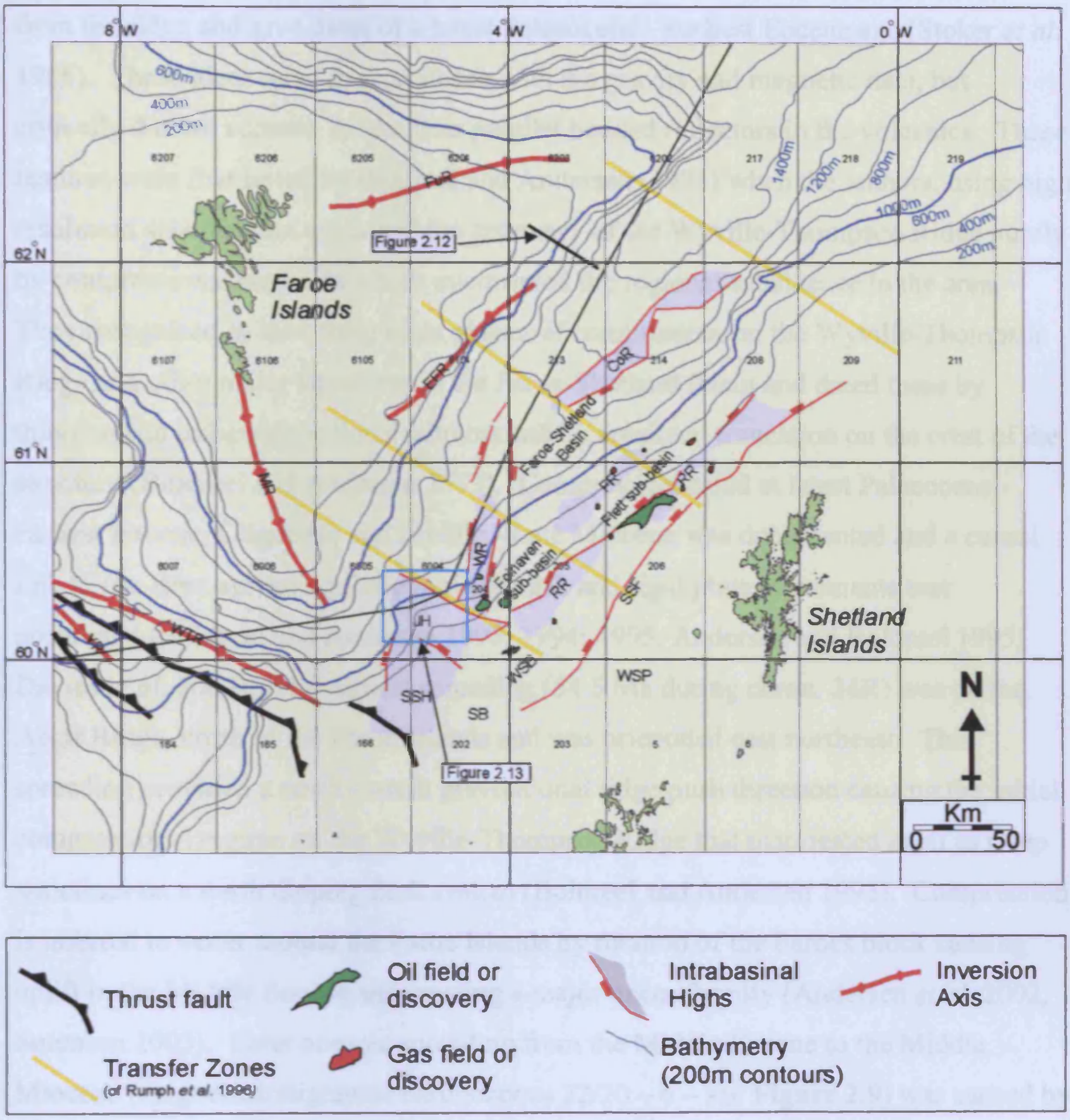


Figure 2.14. Map of the main structural elements of the Faroe-Shetland Basin, showing the main extensional, compressional and Transfer Zones seen in the basin. Note the predominantly northeast - southwest trending structures of the Rona and Corona Ridges and the north - south Westray Ridge, which depict the Cretaceous and Jurassic extensional events respectively. WTR= Wyville-Thompson Ridge, MR= Munkagrunnur Ridge, EFR= East Faroe Ridge, YR= Ymir Ridge, SSH= Sula Sgier High, SB= Solan Basin, JH= Judd High, RR= Rona Ridge, WSB= West Shetland Basin, WR= Westray Ridge, WSP=West Shetland Platform, SSF= Shetland-Spine Fault, FR= Flett Ridge, CoR= Corona Ridge and FuR= Fugloy Ridge. The location of the seismic line shown in Figure 2.12 and the position of Figure 2.13 is also shown.

otherwise uniform oceanic crust. Lavas and tuffaceous sandstones were recovered from the ridge and give dates of a latest Palaeocene – earliest Eocene age (Stoker *et al.* 1988). These ideas were in accordance with the gravity and magnetic data, but critically did not account for the near parallel bedded reflectors in the volcanics. These features were first noted by Boldreel and Andersen (1993) when the authors, using high resolution seismic data explained the geometry of the Wyville-Thompson Ridge purely by compressional tectonics which overprinted the regional subsidence in the area. They recognised at least three main phases of compression on the Wyville-Thompson Ridge and other major structures in the Faroe-Shetland Basin and dated these by thinning and onlap depositional patterns and by erosional truncation on the crest of the structure (Boldreel and Andersen 1993). Compression dated at latest Palaeocene - earliest Eocene, Oligocene and Middle - Late Miocene was documented and a causal link to sea-floor spreading through ridge push and rigid plate movements was postulated (Boldreel and Andersen 1993; 1994; 1995, Andersen and Boldreel 1995, Davies *et al.* 2004). The earliest spreading (54.5 Ma during chron. 24R) was on the Aegir Ridge, north of the Faroe Islands and was orientated east northeast. This spreading provided a north - south gravitational ridge push direction causing the initial compressional regime on the Wyville-Thompson Ridge that manifested itself as ramp anticlines on a north dipping fault system (Boldreel and Andersen 1993). Compression is inferred to occur around the Faroe Islands by rotation of the Faroes block causing uplift in the Middle Eocene and creating a major unconformity (Andersen *et al.* 2002, Sørensen 2003). Later oceanic spreading from the Middle Eocene to the Middle Miocene (magnetostratigraphic chronozones 22/20 – 6 – see **Figure 2.9**) was caused by the linking of the Reykjanes Ridge with the Aegir Ridge through a pseudo-transform fault in the vicinity of the Jan Mayen block (Larsen 1988). There was a gradual cessation in the spreading on the Aegir Ridge which caused the Jan Mayen block to rotate anticlockwise causing dextral wrench movements on the Faroe-Rockall Plateau (**Figure 2.15**). By the Mid-Miocene spreading was more orthogonal to the Faroe-Shetland Basin and northeast - southwest trending domes and inversion structures developed at this time (Davies *et al.* 2004).

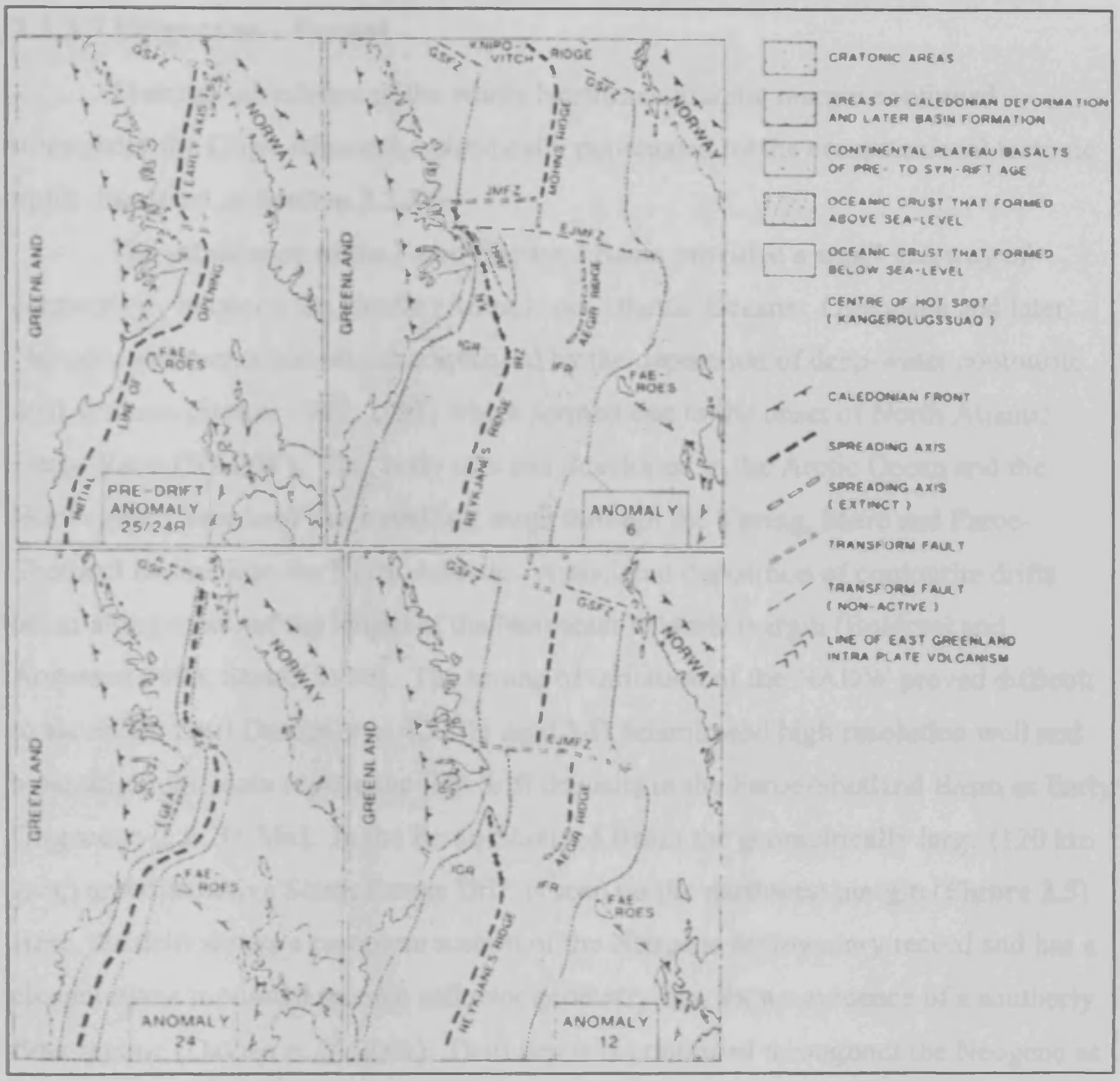


Figure 2.15. Model of the development of the North East Atlantic Rift. At Pre-Anomaly (magneto-chron.) 25/24 there was a sinuous trend along the initial line of opening, with onshore intra-plate volcanism crossing East Greenland. Upon opening this early Greenland Axis became extinct (EGEA) and oceanic spreading resulted along a continuous trend of the Reykjanes, Aegir and Mohns Ridges. By Anomaly 12, the Aegir Ridge became separated from the Reykjanes Ridge to the south by the Iceland Faroe Ridge and to the north by the Early Jan Mayen Fracture Zone (EJMFZ). The Aegir Ridge was becoming progressively extinct to the north creating a fan shape of the anomalous oceanic crust. The Reykjanes and Aegir Ridges formed a pair of propagating/retreating rifts until about Anomaly 6, when the Reykjanes Ridge tore a continental sliver off the Greenland continent which became the Jan Mayen Ridge. At this time there was an overstepping of the EJMTZ to a more northerly Jan Mayen Fracture Zone (JMFZ). This multiple and propagating rift model can account for the relative movements of the plates which results in changes in the compressional regime seen by ridge push forces throughout the Tertiary of the Northeast Atlantic Margin. After Larsen (1988).

2.2.3.7 Oligocene - Recent

Thermal subsidence of the whole Northeast Atlantic margin continued throughout the Oligo-Miocene, episodically punctuated by the compressional tectonic uplift discussed in **Section 2.2.3.6.**

The subsidence of the Faroe-Shetland Basin provided a small gateway of connectivity between the Nordic (Arctic) and Atlantic Oceans. Oligocene and later Neogene sedimentation was characterised by the deposition of deep-water contourite drift systems (Stoker 1997; 1998) which formed due to the onset of North Atlantic Deep Water (NADW). This body of water developed in the Arctic Ocean and the Norwegian-Greenland Sea travelling south through the Vøring, Møre and Faroe-Shetland Basins into the North Atlantic. Associated deposition of contourite drifts occur along much of the length of the Northeast Atlantic margin (Boldreel and Andersen 1993, Stoker 1998). The timing of initiation of the NADW proved difficult to ascertain, until Davies *et al.* (2001) used 3-D seismic and high resolution well and biostratigraphic data to date the first drift deposits in the Faroe-Shetland Basin as Early Oligocene (c.a. 35 Ma). In the Faroe-Shetland Basin the geometrically large (120 km long) and distinctive South Faroes Drift is seen on the northwest margin (**Figure 2.5**). Here, the drift shows a complete section of the Neogene sedimentary record and has a characteristic mounded seismic reflector geometry that shows evidence of a southerly flow regime (Davies *et al.* 2001). Drift deposits continued throughout the Neogene as the basin continued to experience deep water conditions as thermal subsidence continued to wane (Stoker 1998, Nielsen *et al.* 1998). Sea-floor spreading persisted along the Reykjanes Ridge far to the northwest as the effects of the Iceland plume became less apparent in the Faroe-Shetland Basin. Isolated periods of uplift in the Mid-Miocene and the Pliocene are believe to have occurred because of plate reorganisations northwest of the Faroe Islands (Larsen 1988, Boldreel and Andersen 1993; 1994). This local uplift was focussed on the Faroe Platform and caused the removal of Eocene sediments and Palaeocene basalts (and possibly later) sediments from the platform surrounding the Faroe Islands (Waagstein and Heilmann-Clausen 1995, Andersen *et al.* 2002). This provided a western source area for sediments to be redeposited on the northern and eastern margins of the Faroe Platform from Oligocene times onwards (Nielsen and Van Weering 1998). An eastern to southeastern source had dominated deposition throughout the Palaeocene and Eocene from the emergent

Scottish and Shetland landmasses (Andersen *et al.* 2000; 2002). A large eastward prograding wedge of Pliocene age has been recognised on the northern and eastern margin of the Faroe Platform and is believed to represent the final compressional uplift event of the region (Nielsen and Van Weering 1998, Andersen *et al.* 2000; 2002).

A striking cross-cutting seismic reflection exists over much of the Faroe-Shetland Basin close to the base of the drift deposits. This is a diagenetic Opal A to CT boundary and appears as a very high amplitude seismic reflection which cross cuts the stratigraphy (see **Chapter 3** and the regional correlations). It is often a good proxy for the top of the Eocene succession especially in the northern part of the basin (Davies and Cartwright 2002).

2.2.4 Summary of Structural and Stratigraphic Evolution of the Faroe-Shetland Basin

The Northeast Atlantic margin has had a complex geological evolution encompassing periods of continental collision and accretion followed by episodic but prolonged periods of extension and crustal thinning (e.g. Ziegler 1989; 1990). Furthermore, the margin experienced an oceanic realm with voluminous magmatic and volcanic material being emplaced and extruded into the previous sedimentary basins along the margin. The origin of the magmatic episode coincided with considerable crustal thinning (due to extension throughout the Mesozoic and Cenozoic) and the impingement of a mantle plume at the base of the lithosphere (White 1988; 1989). The combined effects of these to processes culminated in the initiation of sea-floor spreading at the start of the Eocene.

Sedimentary basins on volcanic margins from around the world allow us to look at the interaction between a post-rift subsiding basin and the effects a retreating mantle plume head has on the sedimentary architecture developed within basins along a the margin. By examining the post-rift fill, and in particular the Eocene succession in the Faroe-Shetland Basin it allows us to look at a unique and crucial period of time; when the basin was in its subsidence phase but was directly adjacent to a newly created sea-floor spreading zone, supported below by the proto-Iceland mantle plume.

2.3 Igneous Activity on the Northeast Atlantic Margin

2.3.1 Overview

The magmatic and volcanic activity created along the Northeast Atlantic margin covers an area of over 250,000 km^2, consisting primarily of flood basalts between East Greenland and the Faroe Islands (Waagstein 1988), 40,000 km^2 of which can be found within the Faroe-Shetland Basin alone (Naylor *et al.* 1999). The magmatism was prevalent during parts of the Palaeogene, and had a significant effect on Palaeocene and Eocene basin development. Volcanism was centred to the northwest of the Faroe-Shetland Basin at the eventual zone of break-up of the North Atlantic (e.g. Andersen 1988, White 1988, Ritchie and Hitchen 1996), **(Figure 2.8)**. The Faroe-Shetland Basin experienced basaltic volcanism and associated intrusive activity in the form of sill complexes above an active mantle plume. Additionally, large igneous centres intruded into the upper crust (see **Figures 2.16**) and were fed by even larger intrusions that caused significant lithospheric thickening by underplating (White 1988, White and McKenzie 1989). It became apparent from high resolution dating techniques ($^{40}AR/^{39}AR$) and palynological age data that the volcanic activity and sedimentation events occurred in a number of discrete pulses (Knox 1996, Jolley 1997, Bell and Jolley 1997, Jolley and Bell 2002, Jolley *et al.* 2002). Pulsed volcanism within the Faroe-Shetland Basin has be documented in detail from analysing the distribution and age of siliciclastics and volcanic rocks (lavas and tuffs) in the basins surrounding the Scotland and Shetland landmass (White and Lovell 1997, Clarke 2002 unpublished PhD. thesis), **(Figure 2.11)**.

Significant uplift on the continental margins surrounding the Faroe-Shetland Basin accompanied the main period of magmatism with an estimated uplift of 900 m in the basin compared to 375 – 525 m in the Northern North Sea Basin (Nadin *et al.* 1997). Clastic deposition was prevalent in the Palaeogene, sourced from the uplifted margins and being deposited in actively subsiding basins west of Britain (Lamers and Carmichael 1999) and eastwards into the North Sea (Jones and Milton 1994). Greater detail on the sedimentary fill of the Faroe-Shetland Basin throughout the Palaeocene can be seen in **Section 2.2.3.4**.

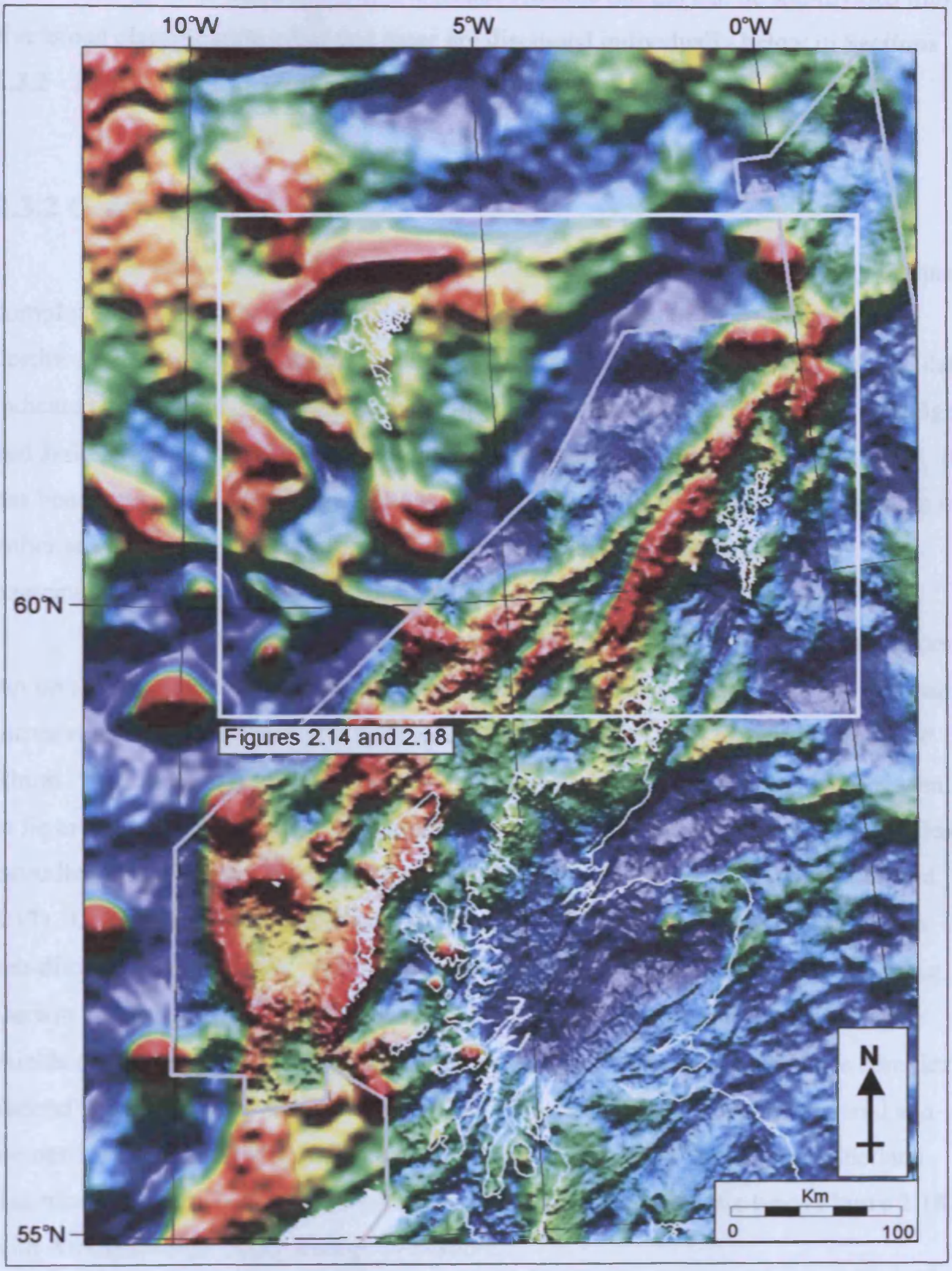


Figure 2.16. Free-Air Bouguer gravity map of the North East Atlantic Margin. Strong northeast/southwest trends can be seen along the margin with isolated circular features depicting the igneous centres along the entire margin. The white rectangle indicates area of Faroe-Shetland Basin. More detailed maps of the structural elements and Igneous features of the Faroe-Shetland Basin can be seen in Figures 2.14. and 2.18 respectively (area highlighted in white box). (Reproduced by permission of the British Geological Survey. © NERC. All rights reserved. IPR/40-51C).

The igneous material on the Northeast Atlantic margin can be sub-divided into five broad classification types and these are discussed individually below in **Sections 2.3.2 - 2.3.6.**

2.3.2 Central Complexes

In the Faroe-Shetland Basin there is little well data to indicate lithology of these complexes and much is inferred from the onshore analogues of East Greenland, northwest Scotland and northeast Ireland. However, seismic, gravity and magnetic data indicate that these plutons were intruded into uplifted horst blocks (e.g. Westray Ridge and Judd High) of the rifted and stretched continental crust (**Figures 2.16** and **2.17**). It has been noted that some of the igneous centres in the Faroe-Shetland Basin develop as either sea-floor or sub-aerially emergent highs and thus have an affect on contemporaneous sediment dispersal patterns in the Palaeocene and Eocene.

The geometry and size of these igneous centres varies in size along the margin but on average are in the region of 30 km wide and have a sheet-like centrally inclined intrusive centres (Naylor *et al.* 1999) similar to the edifices seen onshore at Mull and Rhum. With the exception of Brendans Dome, most of the large igneous centres seem to lie to the south and west of the Munkagrunnur Ridge with their associated lava piles extruding radially from the multiple centres in the Rockall Trough (**Figures 2.16** and **2.17**). Central igneous complexes of the Faroe-Shetland Basin (**Figure 2.18**) can be sub-divided into two types. Firstly, there are centres with tholeiitic compositions (e.g. Darwin and East Erlend) which are associated with the growth of substantial lava shields (Gatliff *et al.* 1984). These centres are analogous to the onshore Skye complex. Secondly, some centres show an association with substantial pyroclastic material and are not associated with the development of large lava shields. In the Faroe-Shetland Basin centres such as Judd and Westray can be categorized into this type (**Figure 2.18**), with Ardnamurchan being analogous onshore.

The distribution of age ranges of these centres is relatively focussed, with the majority of the centres having radiometric dates of an Early Tertiary age (between 60 Ma and 54 Ma). A few large centres in the Rockall Trough, such as Rosemary Bank and Anton Dohrn are believed to be of Late Cretaceous age. This is an interesting observation, suggesting that much of the magmatism on the margin occurred at a late

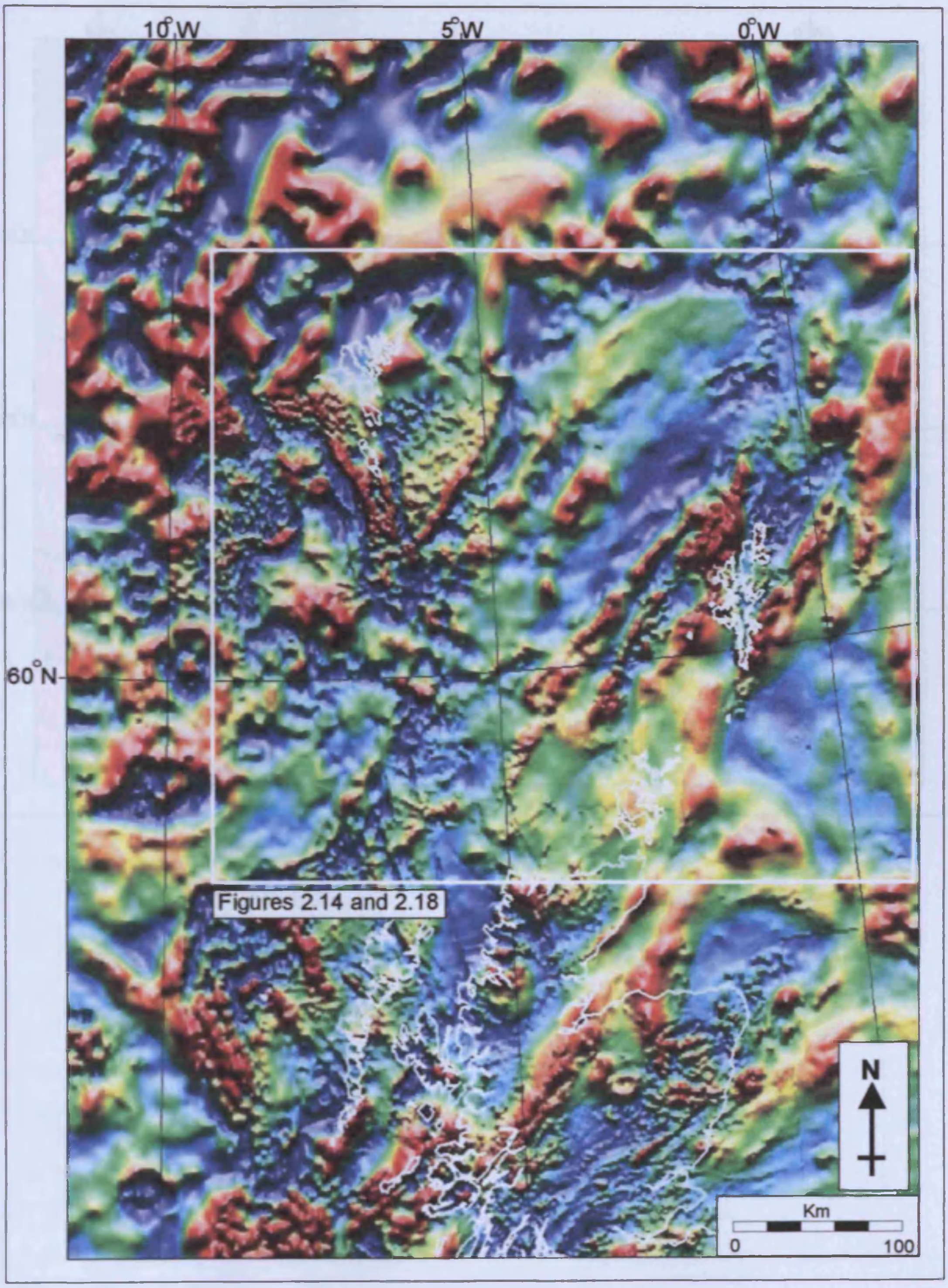


Figure 2.17. Magnetic map of the Northeast Atlantic Margin. The white rectangle indicates the Faroe-Shetland Basin shown in Figures 2.14 and 2.18. Major northeast - southwest trending features are seen just west of the Shetland Islands, and these are interpreted as major Mesozoic fault-blocks formed during extension of the continental margin. (Reproduced by permission of the British Geological Survey. © NERC. All rights reserved. IPR/40-51C).

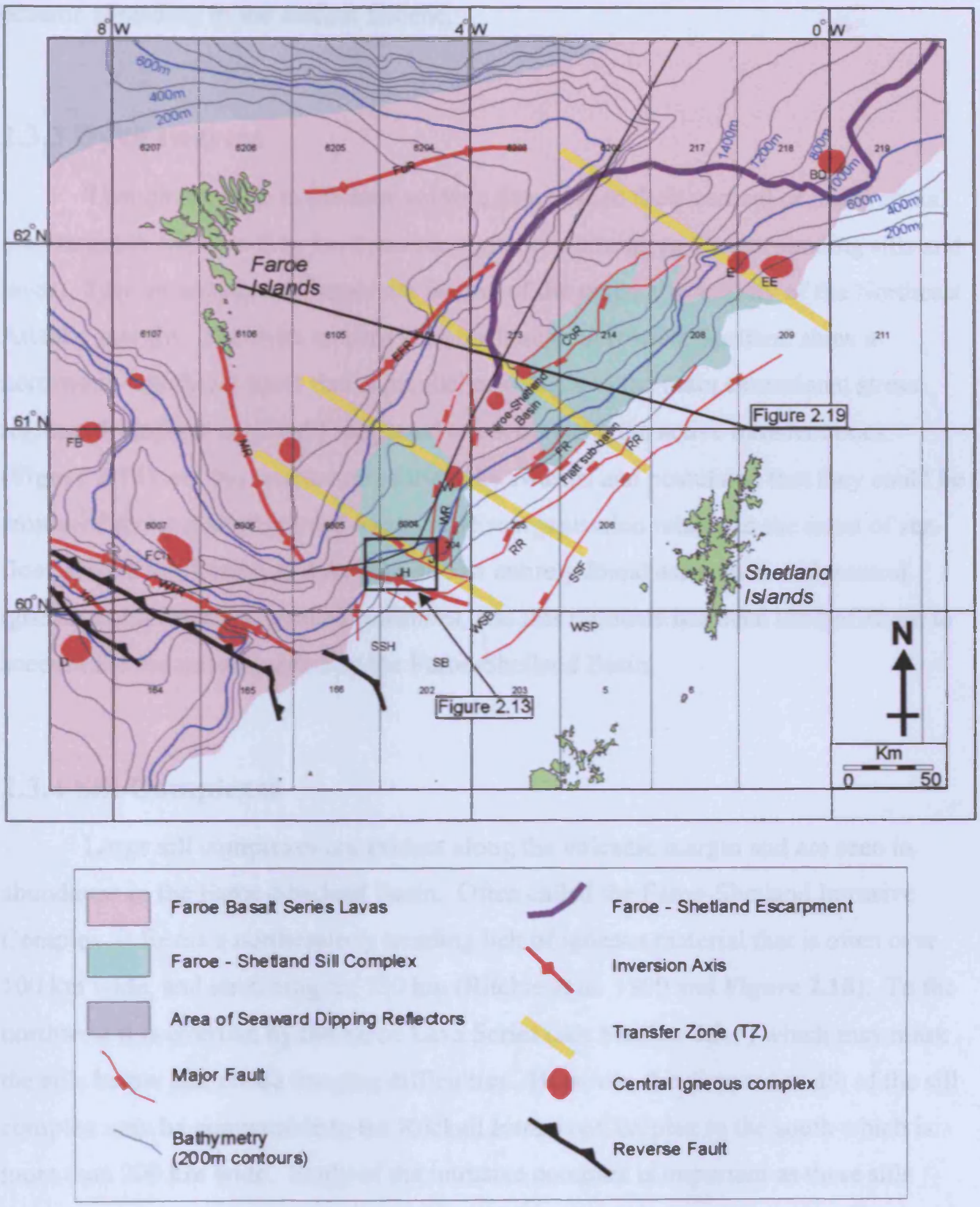


Figure 2.18. Map showing Igneous volcanics and intrusives in the Faroe-Shetland Basin. Main features shown are the central igneous complexes, seaward dipping reflector sequences, the limit of the Faroe sill and lava complexes and the position of the Faroe-Shetland Escarpment. Abbreviations of the igneous complexes shown are as follows: D= Darwin, FCK= Faroe Channel Knoll, FB= Faroe Bank, J= Judd, W= Westray, E= Erlend, EE= East Erlend and BD= Brendans Dome. For all other abbreviations see Figure 2.14. Positions of Figures 2.13 and 2.19 are shown.

stage in the rift development and continued to the time of eventual break-up and oceanic spreading in the earliest Eocene.

2.3.3 Dyke swarms

Though not seen in offshore seismic data (due to their vertical or near vertical orientation or because they have seismic signatures similar to the surrounding sills and lavas), dyke swarms are an important feature of the magmatic activity of the Northeast Atlantic margin. The dyke swarms seen onshore in northwest Scotland show a northwest - southeast trend that is parallel or oblique to the main extensional stress regime. Rumph *et al.* (1993) suggested a link between the active transfer zones **(Figure 2.14)** and the orientation of the dyke swarms and postulated that they could be emplaced along a similar trend due to plate reorganisation related to the onset of sea-floor spreading. Onshore, dykes are almost entirely found associated with central igneous complexes and central volcanoes, and this rationale has been used offshore to account for the missing dykes in the Faroe-Shetland Basin.

2.3.4 Sill Complexes

Large sill complexes are evident along the volcanic margin and are seen in abundance in the Faroe-Shetland Basin. Often called the Faroe-Shetland Intrusive Complex, it forms a northeasterly trending belt of igneous material that is often over 100 km wide, and stretching for 750 km (Ritchie *et al.* 1999 and **Figure 2.18**). To the northwest it is overlain by the Faroe Lava Series (see **Section 2.3.5**) which may mask the sills below and create imaging difficulties. However, the disputed width of the sill complex may be comparable to the Rockall Intrusive Complex to the south which is more than 200 km wide. Study of the intrusive complex is important as these sills represent the feeder systems to the vast quantities of lavas seen above. Geochemical analysis of well data from many sills in the Faroe-Shetland Basin indicates a tholeiitic olivine dolerite composition suggesting a Mid-Ocean Ridge Basalt (MORB) composition (e.g. Gibb *et al.* 1986). Detailed trace element and rare earth data has been collected from various sills and they have similar affinities with the Middle and Upper Series of the Faroe Lava Series (Gibb and Kanaris-Sotiriou 1988). This observation nicely links into the age of the intrusive complex. Numerous K/Ar and Ar/Ar age dates from many sills have been obtained over the years which suggest that the timing of intrusion for the sill complex was between 55 Ma and 53 Ma (Hitchen and Ritchie 1987, Ritchie and Hitchen 1996). This age is synchronous with the onset of sea-floor spreading in the North Atlantic suggesting that sills were intruding as ocean floor was being created to the northwest.

Some areas of the sill complex contain multiple sills with thicknesses of over 300 m (Hitchen and Ritchie 1987). It is argued that sill emplacement can provide local uplift especially when they are intruded close to the sea-bed, and this in turn can affect later sediment distribution patterns (Lamers and Carmichael 1999, Smallwood and Maresh 2002).

2.3.5 Lavas

The regional basaltic deposits of the Faroe Lava Series can be found along the entirety of the margin and onshore in places such as Antrim, Mull and Skye. They cover an area of over 40,000 km^2 in the Faroe-Shetland Basin and can be divided into two types; regional lavas from fissure eruptions found around the Faroe Islands and localised lavas from the central volcanic edifices in the basin (Naylor *et al.* 1999). Both lavas tend to have a MORB composition and are closely related to the sills below. The age range of the lavas remains uncertain with dates between 61 Ma and 54 Ma (Ritchie *et al.* 1999). The regional lavas are seen to be the thickest in the vicinity of the Faroe Islands where they reach a thickness greater than 3 km (at well Lopra 1 in the south). Here, there is a complete section of the Lower, Middle and Upper Series of lavas which thin to the south and east into the Faroe-Shetland Basin. These lavas were erupted from fissures and prograded as sub-aerial flows and deltas into the marine basin (**Figure 2.18**). Sinuous reflector geometries within the lava pile are seen from seismic data and represent the movement of the individual flows towards the southeast. This process of lava delta progradation culminated when the lava flows entered the marine basin and quenched into hyaloclastite deposits and formed the Faroe-Shetland Escarpment (e.g. Kiørboe 1999, see **Section 2.2.3.4** and **Figure 2.12**), which curves around the northern part of the basin and formed a Late Palaeocene shoreline.

2.3.6 Tuffaceous Ash Deposits

Regional tuffaceous deposits cover a vast area of the Northeast Atlantic margin and also in the North Sea. These deposits are readily identifiable on seismic data and have characteristic wireline log signatures, therefore aiding the correlation of reservoirs and seals in the basins. The tuffs are associated with the eruption of volcaniclastic material from the igneous centres and from the spreading ridges. The most widespread deposit is the Balder Tuff which is believed to represent the opening of the North Atlantic at the start of the Eocene (e.g. Knox and Morton 1983; 1988, Morton *et al.* 1988, Stoker 1988).

2.4 Crustal Structure and Subsidence History of the Faroe-Shetland Basin

Throughout the 1970's and 1980's there was a significant debate about the nature of the crust under the Faroe-Shetland Basin. In the absence of any direct evidence in the form of deep seismic reflection profiles the origin of the crust beneath the Faroe-Shetland Basin remained uncertain. Is the crust thinned and extended continental lithosphere or is it floored by a remnant, failed spreading centre of oceanic crust?

In 1984, Bott and Bott and Smith, interpreted velocity distributions within the crust from the early 1970's North Atlantic Seismic Project (NASP) refraction transects. They concluded an oceanic crust at depth in the centre of the (then-called) Faroe-Shetland Channel, and a block of continental crust underlying thick lavas on the Faroes. Similarly thinned continental crust was interpreted under the southern Rockall Trough by gravity, magnetic and single channel seismic data (Megson 1987, Earle *et al.* 1989). The oceanic crust in the centre of the Faroe-Shetland Channel was believed to separate the continental blocks of the West Shetland Platform and the Faroe Block, and was a northern continuation of the Rockall Trough.

It was not until drilling activity west of the Rona Ridge proved crystalline basement under a thick Mesozoic and Cenozoic sedimentary succession, that a continental lithosphere across the basin centre was postulated. Richardson *et al.* (1998) used wide angle seismic reflection profiles to image the crustal structure. This work highlighted that the continental crust in the centre of the Faroe-Shetland Basin has been significantly thinned when compared to the more stable crust in the Shetland Platform to the southeast. On this platform the depth to the Moho yields figures of approximately 25 - 26 km (e.g. Smallwood and Maresh 2002). On this margin, there is a sedimentary cover of approximately 4 - 5 km. Compare this to what is seen in the centre of the Faroe-Shetland Basin, where we see the Moho at about 16 - 17 km, and considerably thicker sedimentary successions reaching 10 - 11 km (**Figure 2.19**).

Hence, thinning of the continental crust from 21 km on the Shetland Platform to 7 km in the basin centre is clearly evident. Moreover, average crustal stretching factors (β factors) can be calculated from this data and taking in all known rifting events from the Triassic to Recent an average β factor of three is postulated (Dean et al. 1999). This is in agreement with Smallwood *et al.* 2001, who used the deep Faroes and Shetlands Transect (FAST) seismic line and potential field modelling to reach similar figures for the crustal stretching factors and depth to the Moho (**Figure 2.19**).

The cumulative effect of many periods of rifting from the Triassic to the Palaeocene was a considerably thinned crust. The final rifting event at the end of the Cretaceous (Maastrichtian) (or perhaps into the Palaeocene, see **Section 2.2.3.4**) lasted for about 18 - 20 Ma, and pre-dated the eventual opening of the North Atlantic (Skogseid *et al.* 1992). Igneous activity in the region (see **Section 2.3**) occurred during the latter stages of this extension at 63 - 62 Ma (Larsen *et al.* 1992, Hitchen and Ritchie 1993). Therefore we see a late introduction of igneous activity with respect to the initiation of rifting. This is resolved by the impingement of a mantle plume at the base of the lithosphere at about 63 Ma, which gave rise to the massive volcanism seen on the margin (e.g. White 1989). Hot athenosphere in the centre of the mantle plume is believed to be approximately 100° C hotter than the surrounding mantle (Griffiths and Campbell 1991) and hence more buoyant. It can therefore preferentially upwell into areas of thinned lithosphere which not only causes volcanism, but significant uplift and erosion on the rift flanks. The area affected by the plume (after spreading) is believed to be in the region of 3000 km in diameter and encircles the North Atlantic Tertiary Igneous Province (e.g. Nadin *et al.* 1997, Ritchie *et al.* 1999, see **Figure 2.8**). Skogseid *et al.* (2000) evaluated the theoretical isostatic lithosphere uplift effects for a 100° C temperature anomaly in a plume head and this can be seen in **Figure 2.20.**

The impingement of the Iceland plume under the thinned continental crust allowed for a degree of underpinning to occur and hold up the lithosphere at anomalously high levels. Once the plume moved off the axis of the Faroe-Shetland

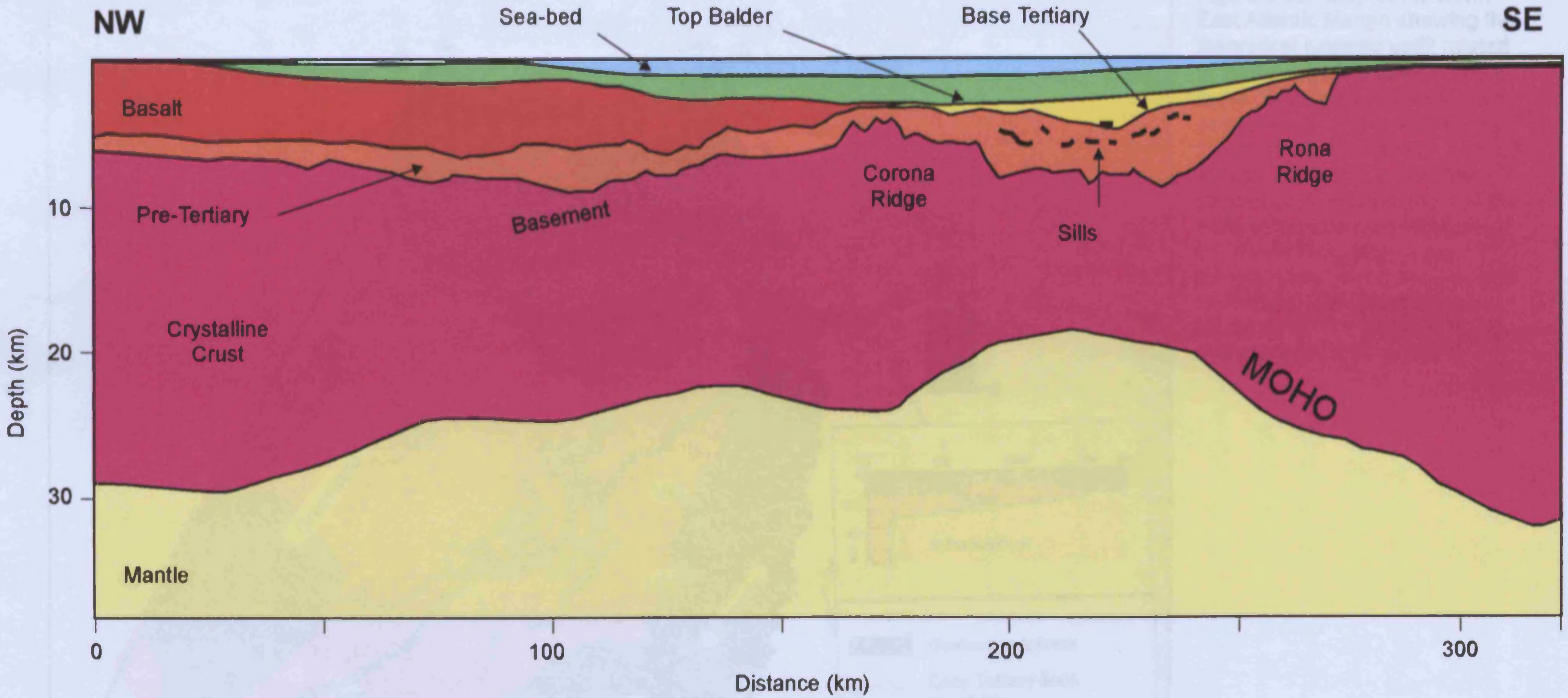


Figure 2.19. Southeast - northwest trending line drawing of deep seismic transect across the Faroe-Shetland Basin. The section is drawn from a profile from the British Institutes Reflection Profiling Syndicate (BIRPS) deep seismic reflection line "FAST" (Faroes and Shetlands Traverse) and potential field modelling (Smallwood et al. 2001). Note the thinning of the continental crust under the central part of the Faroe-Shetland Basin where the moho reaches a depth of less than 20 km, and the crust is in the region of 10 km thick compared to almost 30 km to the southeast of the Rona Ridge. Note the abundance of sills located predominantly in the Cretaceous section in the area where the crust is highly attenuated. The Eocene to recent succession is shown in green. For location of seismic traverse see Figure 2.18. From Smallwood and Maresh 2002.

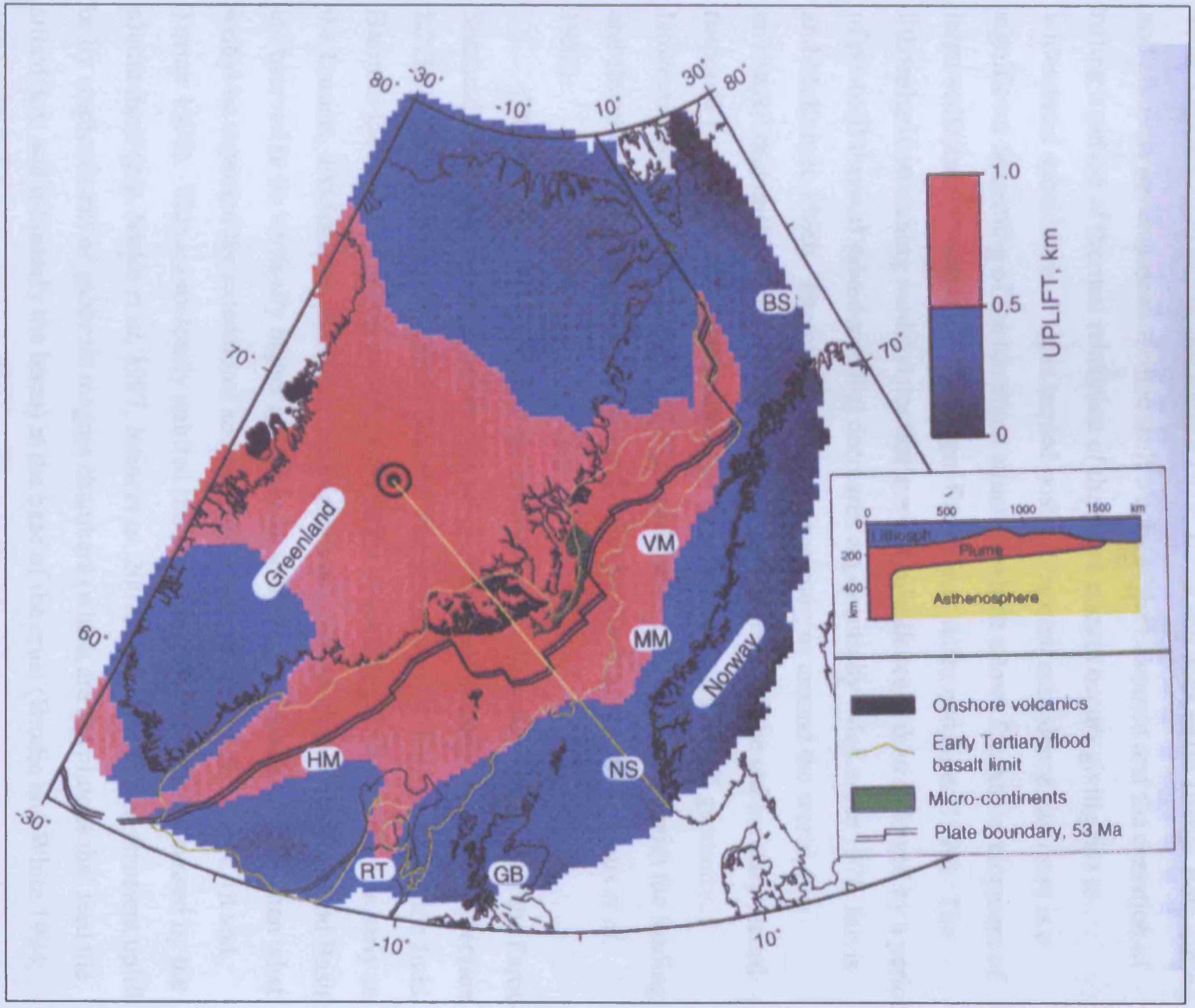


Figure 2.20. Map of the North East Atlantic Margin showing the theoretical isostatic uplift related to an average increase of 100°C in the temperature anomaly in a plume head. The resulting thickness distribution is a function of both the distance from the plume centre and the relief at the base of the overlying lithosphere. NS=North Sea, MM= Møre Margin, VM= Vøring Margin, BS= Baltic Sea, HM= Hatton-Rockall Margin and RT= Rockall Trough. (After Skogseid *et al.* 2000).

Basin at the start of the Eocene, subsidence of the basin was greatly enhanced and lead to a dramatic deepening in the basin at this time (e.g. Turner and Scrutton 1993, Clift and Turner 1998).

Thus, it is widely believed that the Eocene is marked by a period of thermal subsidence and relaxation. Subsidence constitutes a major component of any passive continental margin, be it volcanic or not at a late stage in the basins development (McKenzie 1978). Accompanying episodes of rifting and extension of the continental crust is a thinning of the lithosphere allowing for the upwelling of buoyant asthenosphere from below (McKenzie 1978, Jarvis and McKenzie 1980). Both during and after the development of tilted fault blocks of the basement and the cessation of rifting a period of thermal relaxation of the rifted margin occurs giving rise to widespread subsidence. This is termed post-rift thermal subsidence and there is a significant deepening of the basement structure which allows for the development of large sedimentary wedges on the margin flanks (e.g. Allen and Allen 1990). The lithospheric stretching model states that syn-rift subsidence is then followed by a period of post-rift thermal subsidence that decreases exponentially (McKenzie 1978, Jarvis and McKenzie 1980). Therefore, on all passive margins around the world, it is envisaged that during this post-rift phase of basin subsidence the sedimentary record responds by infilling the basin margins by the classic "Steers Head" geometry. However, this geometry occurs when the subsidence effect is coupled with the loading and flexure effects of the sedimentary infill (Watts and Steckler 1979, Watts *et al.* 1982).

The post-rift subsidence pattern observed in the Eocene succession of the Faroe-Shetland Basin is believed to be punctuated by periods of episodic local uplift (**Section 2.2.3.6**). Indeed, Sørensen (2003) believes that tectonic uplift is evidenced in the Judd Basin in the Middle Eocene, causing a cessation in subsidence which began as early as the Danian. Evidence from subsidence analysis suggests that the Faroe-Shetland Basin is observed to be vertically higher (possibly up to 900 m in the basin centre) than what would be expected for extensional sedimentary basins (Nadin *et al.* 1997, Clift and Turner 1998). This anomalously uplifted basin margin is believed to be caused by the plume itself (e.g. Nadin *et al.* 1997, Jones *et al.* 2002) creating dynamic transient uplift or by emplacement of gabbroic magma chambers (which are the plutons that feed the sills/dykes and ultimately the lavas) at the base of the crust (Brodie and White 1994; 1995). This lower crustal emplacement; termed underplating, underpins the base of the crust causing permanent uplift and hence allowing the crust to sit at a higher level than otherwise expected in the absence of any underplate. Accelerated anomalous subsidence rates are also recognised in the North Sea Basin (e.g. Joy 1992; 1993; 1996).

Subsidence patterns are thus significantly influenced by the primary effect of a plume on the base of the crust as well as the presence of underplate creating pronounced permanent uplift. However, the origin of anomalous Tertiary subsidence in basins surrounding the British Isles remains uncertain. Evidence suggests that some basins on the North Atlantic volcanic margin experienced no Tertiary subsidence and have remained close to sea-level until the present day. Other basins meanwhile have experienced significant rates of subsidence; up to 3 km of sediment-loaded subsidence in 15 - 20 Ma during the same time interval (Hall and White 1994).

No specific work has been published on the Eocene subsidence history of the Faroe-Shetland Basin with many authors only documenting subsidence histories until the end of the Palaeocene. Turner and Scrutton (1993) looked at the interval between the Late Cretaceous and the top T50 unit (Balder Formation equivalent) and documented an anomalous accelerated period of Palaeocene tectonic subsidence in the centre of the Faroe-Shetland Basin, which was not seen on the basin margins. This was achieved by producing subsidence curves from many of the released wells from both the margin and the basin axis. This high accelerated subsidence in the axis of the Faroe-Shetland Basin lasted for about 5 Ma, though was not seen in the wells from the margin. Turner and Scrutton (1993) argue that this anomalous Palaeocene subsidence cannot be explained by relative sea-level change, fluctuations in sediment supply or eustasy. They prefer to offer the explanation that the timing of high subsidence rates was coincident with the oldest recorded magmatism related to the proto-Iceland plume, hence suggesting a causal link between regional tectonics and Palaeocene subsidence patterns (Turner and Scrutton 1993, Clift and Turner 1995; 1998, Clift 1997; 1999, Clift *et al.* 1998). Furthermore, a re-alignment of the lateral outflow of the plume may have occurred during Eocene sea-floor spreading with this along axis trend reducing both dynamic support and uplift to the southeast (Nadin *et al.* 1997, Jones *et al.* 2002). After the period of high accelerated subsidence, the Faroe-Shetland Basin experienced more normal conditions indicative of thermal subsidence.

3. Chapter Three: Stratigraphic Context and Seismic-Stratigraphic Methodology and Approach

3.1 Introduction

This chapter will initially introduce and review the concepts of sequence-stratigraphy which are relevant to this regional basin study. In particular it will focus on the ongoing debate surrounding the major controlling mechanisms on depositional architecture, be it tectonics, eustasy or sediment supply. Following this brief summary, the chapter will outline the methodology of sub-dividing the Eocene succession of the Faroe-Shetland Basin taken in this study and highlight the reasons for choosing the approach. Seismic –stratigraphic units will be herein defined and then correlated on long range regional correlations of 2-D seismic data. Criterion for selecting the highest resolution well data is then discussed and the key wells, with the best age constraints are tied to the regional correlations. These correlations together with the well data will form the seismic-stratigraphic framework which is the discussed in **Chapters 4, 5** and **6.**

3.2 A Critical Review of Seismic and Sequence-Stratigraphic Concepts Related to this Study

Until the advent of geophysical seismic data, stratigraphy had fallen from favour as a core subject within the earth science community. Large scale seismic transects through sedimentary basins allowed for unprecedented geophysical insight into a previously outcrop based area of study. This evolution in the discipline began in the 1970's with the introduction of seismic-stratigraphy (Vail *et al.* 1977a, b). This new subject has its roots in the much older subject of sequence-stratigraphy that is based from preceding decades of discussion on the origins of cyclic sedimentation. This discussion proved to be an instrumental component in the continuing debate between tectonic and eustatic and climatic controls on relative sea-level change and its cyclical effect on depositional architecture (e.g. Ramsbottom 1977, George 1978, Miall 1986) (**Figure 3.1**). Sedimentary cycles have been recognised in geological strata,

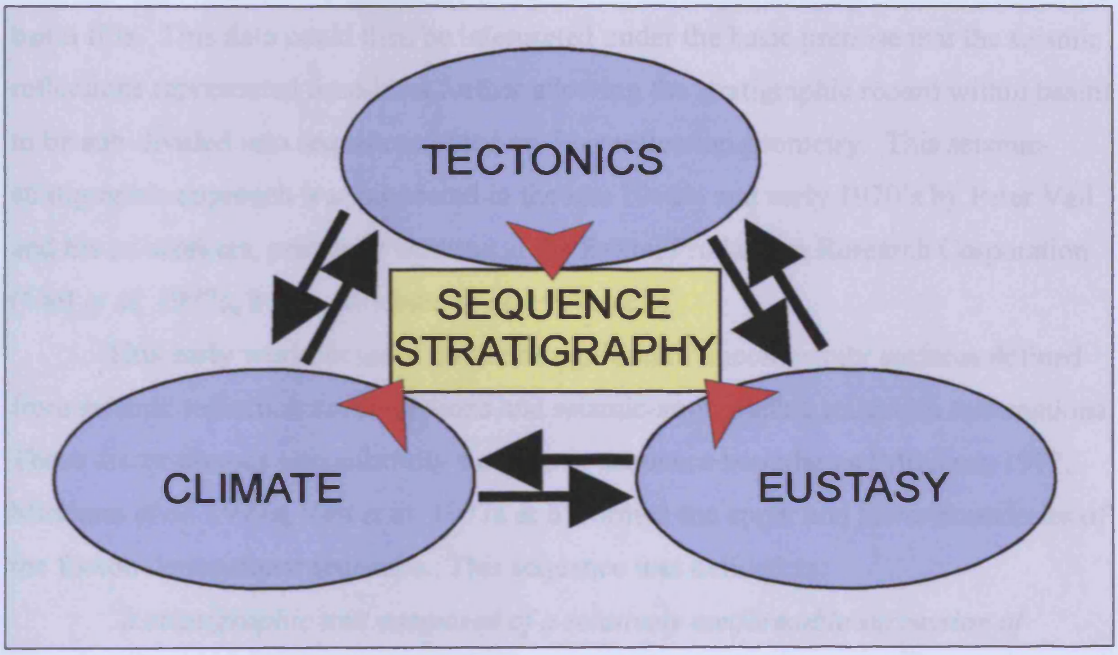


Figure 3.1. Schematic cartoon showing the relationships between eustasy, tectonics and climate. The interaction between these three combining factors controls the stratigraphic development of sequences deposited in sedimentary basins.

primarily in the field, for centuries (e.g. Steno 1669). This cyclicity has long been associated with changes in the sea-level, and more recently involving a component of tectonics.

Throughout the 1970's and until the present day, the oil industry has led the way in the recording and processing of significant amounts of seismic reflection data from sedimentary basins around the world. Initially, large scale (100's km) of 2-D seismic data, and more recently smaller, higher quality 3-D seismic surveys were acquired allowing for unprecedented imaging of both onshore and offshore sedimentary basin fills. This data could then be interpreted under the basic premise that the seismic reflections represented time-lines further allowing the stratigraphic record within basins to be sub-divided into sequences based on their reflection geometry. This seismic-stratigraphic approach was pioneered in the late 1960's and early 1970's by Peter Vail and his co-workers, primarily working at the Exxon Production Research Corporation (Vail *et al.* 1977a, b & c, Mitchum *et al.* 1977a & b).

This early work focussed on the recognition of unconformity surfaces defined from seismic reflection configurations and seismic-stratigraphic reflection terminations. These discontinuous unconformity surfaces or sequence boundaries (Mitchum 1977, Mitchum *et al.* 1977a, Vail *et al.* 1977a & b) formed the upper and lower boundaries of the Exxon depositional sequence. This sequence was defined as:

> *"a stratigraphic unit composed of a relatively conformable succession of genetically related strata bounded at its top and base by unconformities or their correlative conformities"* (Mitchum *et al.* 1977a - see **Figure 3.2**).

It is these depositional sequences that Vail *et al.* (1977a) believed were controlled by changes in global sea-level with a tectonic control only considered to be a local controlling factor that enhances or suppresses the overall eustatic effect on sequence development. In this model sequence boundaries were believed to form as a direct result of a fall in eustatic sea-level. With this in mind, sequence boundaries were identified and mapped on seismic data as surfaces defined by onlap, truncation and downlap terminations (Mitchum 1977, Mitchum *et al.* 1977a - see **Figure 3.2**). From these studies a whole new terminology began to emerge into the literature and by the late 1970's a definitive book; "The Payton Volume" summarised the early work.

It is important here to stress that seismic-stratigraphy is based resolutely on the principle assumption that seismic reflections provide a time-stratigraphic record of the depositional and structural patterns seen on seismic data. This is because primary

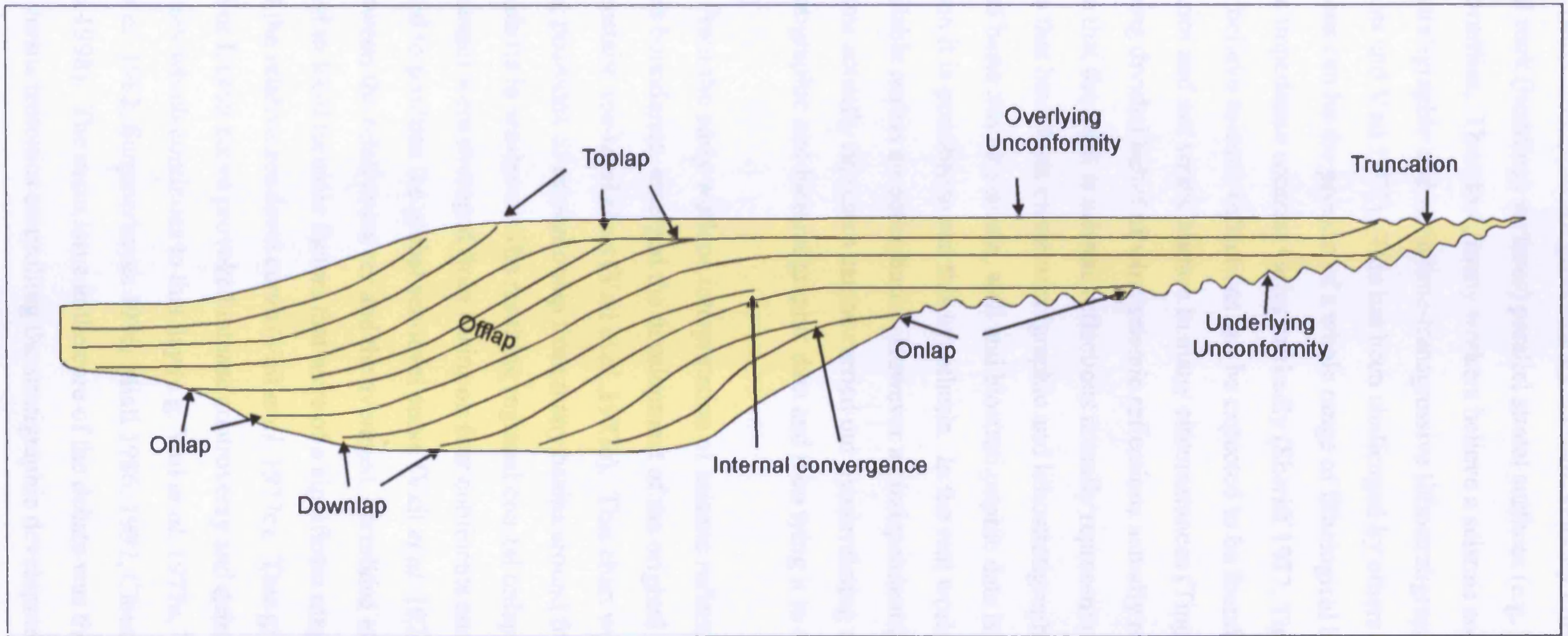


Figure 3.2. Diagram showing an idealised seismic sequence as defined by Mitchum *et al.* (1977b). This depositional sequence is defined by upper and lower unconformities (sequence boundaries) which are recognised and mapped on seismic data by a variety reflection terminations.

seismic reflections (that are generated by acoustic velocity - density contrasts at physical rock (bedding) surfaces) parallel stratal surfaces (e.g. bedding planes) and unconformities. Therefore, many workers believe a seismic section is a chronostratigraphic and not a time-transgressive lithostratigraphic record of the strata (Mitchum and Vail 1977). This has been challenged by others who believe that seismic reflections can be the product of a whole range of lithological boundaries where the acoustic impedance contrast varies markedly (Sheriff 1977, Tipper 1989). These authors believe seismic reflections can be expected to be found at lithostratigraphic boundaries and not stratal surface in many circumstances (Tipper 1993). There is a continuing divided belief of what seismic reflections actually represent and this author believes that the truth is seismic reflections actually represent many real and artefactual surfaces that have both chronostratigraphic and lithostratigraphic significance. In an idealised basin where seismic, well and biostratigraphic data is of extremely high resolution it is possible to test this hypothesis. In the real world this quality of data is not available across an entire basin. However an independent test of what seismic reflections actually represent can be carried out by correlating wells with good lithostratigraphic and biostratigraphic data and then tying it to seismic data (see **Section 3.3.2**).

From the early work on interpretation of seismic reflection packages and sequence boundaries, emerged the development of the original coastal onlap curve and global eustatic sea-level chart (Vail *et al.* 1977c). This chart was constructed by mapping positions of coastal onlap from many basins around the world and inferring relative shifts in sea-level. The resulting regional coastal onlap curves (of relative sea-level change) were averaged from basins on four continents and chronostratigraphically correlated to produce the global sea-level curve (Vail *et al.* 1977c). Any divergence seen between the relative curves and the averaged, correlated eustatic curve was attributed to local tectonic factors that were of a significant magnitude such that they affected the relative sea-level curve (Vail *et al.* 1977c). This global curve, known as the Vail or Exxon curve provoked intense controversy and debate in the geological community which continues to this day (e.g. Vail *et al.* 1977a, Vail and Todd 1981, Watts *et al.* 1982, Summerhayes 1986, Miall 1986; 1992, Cloetingh and Kooi 1990, Nystuen 1998). The main issue at the core of the debate was the argument between eustasy versus tectonics controlling the stratigraphic development of sedimentary basins. Furthermore, the accuracy of the correlation of supposed global events from basin to basin remains rigorously challenged (e.g. Miall 1991; 1992).

The original Exxon model continued to be developed and refined and it soon included well and outcrop data to give a higher degree of resolution on age constraints (e.g. Vail 1987, Van Wagoner *et al.* 1988). Additionally, tighter definitions of the terminology were required as the models developed. In particular when defining the term sequence, the definition of an unconformity became restricted to:

> *"a surface separating younger from older strata along which there is evidence of sub-aerial erosion and truncation (and in some areas correlative sub-marine erosion) and sub-aerial exposure along which a significant hiatus is indicated"* (Van Wagoner *et al.* 1987; 1988).

This change in the definition of an unconformity changed the original concept of the unconformity-bounded depositional sequence of Mitchum *et al.* (1977a) and Vail and Todd (1981) which incorporated all other unconformities other than ones with a sub-aerial origin. Therefore, the boundaries of the classical Exxon sequences could only be defined in basins where relative sea-level falls are significant enough to create sub-aerial erosion surfaces, namely basin margin zones and proximal parts of relatively shallow epicontinental basins (Nystuen 1998).

The change in definition of a sequence boundary has critical consequences. Any unconformity that develops away from the basin margin zone due to tectonic stress (Pickering *et al.* 1989, Cloetingh 1991), chemical dissolution or erosive bottom currents (Winterer 1991) is no longer classified as a sequence boundary even though they represent a change in stratigraphic development. Therefore, in an entirely marine realm where no sub-aerial erosion occurs, depositional sequences *sensu* Exxon do not occur. This concept needs to be addressed to evaluate the stratigraphic significance of deposits that occur in these scenarios.

Despite these questionable revisions, the Exxon model continued to be challenged with the main contentious issue being the application of seismic-stratigraphy as a predictive tool for eustatic sea-level change. Indeed, this culminated in 1987 when Haq *et al.* (1987) published the first global sea-level cycle chart. This eustatic sea-level chart is composed of various relative sea-level curves from different basins from around the world. The chart was interpreted to show numerous cycles of eustatic sea-level rise and fall that occur at a number of magnitudes (Haq *et al.* 1987; 1988). Four cycles were interpreted to eustatic curve and are summarised below:

a). 1st Order cycle, with a duration greater than 50 Ma.

b). 2nd Order cycle, with a duration of between 3 and 50 Ma.

c). 3rd Order cycle, with a duration of 0.5 to 3 Ma.

d). 4th Order cycle, with a duration of between 0.01 and 0.5 Ma.

These cycles were interpreted by the Exxon co-workers to advocate an observed cyclicity in eustasy. This statement has caused the most controversy and to this day, this chart remains questioned by many (e.g. Miall 1986; 1991, Hubbard 1988; Underhill 1991). This is largely due to the lack of supporting evidence which remains confidential as well as a question mark concerning the methodology used in the construction of the chart. There is still some ambiguity as to whether or not corrections have been applied for local variations in tectonic uplift and subsidence, and there have been serious questions asked about the precision of the dating of sequence boundaries that the chart implies (Miall 1991). Furthermore, other models for the sub-division of strata were developed in direct opposition to the Exxon model. The most notable of these alternatives was the model of Galloway (1989) where genetic stratigraphic units were bounded by maximum flooding surfaces, and Embry who proposed transgressive – regressive (T – R) sequence stratigraphy (e.g. Embry 1995).

A more detailed discussion of sequence-stratigraphy is required to highlight the major controls on depositional architecture and to introduce some terminology which will appear in the following sections of this chapter (**Section 3.4**).

Firstly, the concept of eustasy needs to be defined and understood. Eustasy is a change in elevation of the sea-level on a global basis relative to a stationary datum such as the centre of the earth (Kendall and Lerche 1988). Therefore in theory eustatic sea-level is a measurable amount and is independent of localised uplift or subsidence (**Figure 3.3**).

Relative sea-level is also a measurable quantity at any given point in time. It can be defined as the distance between the sea-surface and a local moving datum (such as basement or a surface within the sediment pile, (Posamentier *et al.* 1988)) (**Figure 3.3**). Relative sea-level constantly changes through time due to a combination of eustatic sea-level variations, vertical tectonic movements of a pre-existing datum and sediment compaction (**Figure 3.3**). Any change in relative sea-level gives rise to changes in the amount of space available for sediments to accumulate. This is the concept of accommodation space which also needs to be introduced when discussing sedimentary basin fill. Changes in accommodation space develop as a result of the

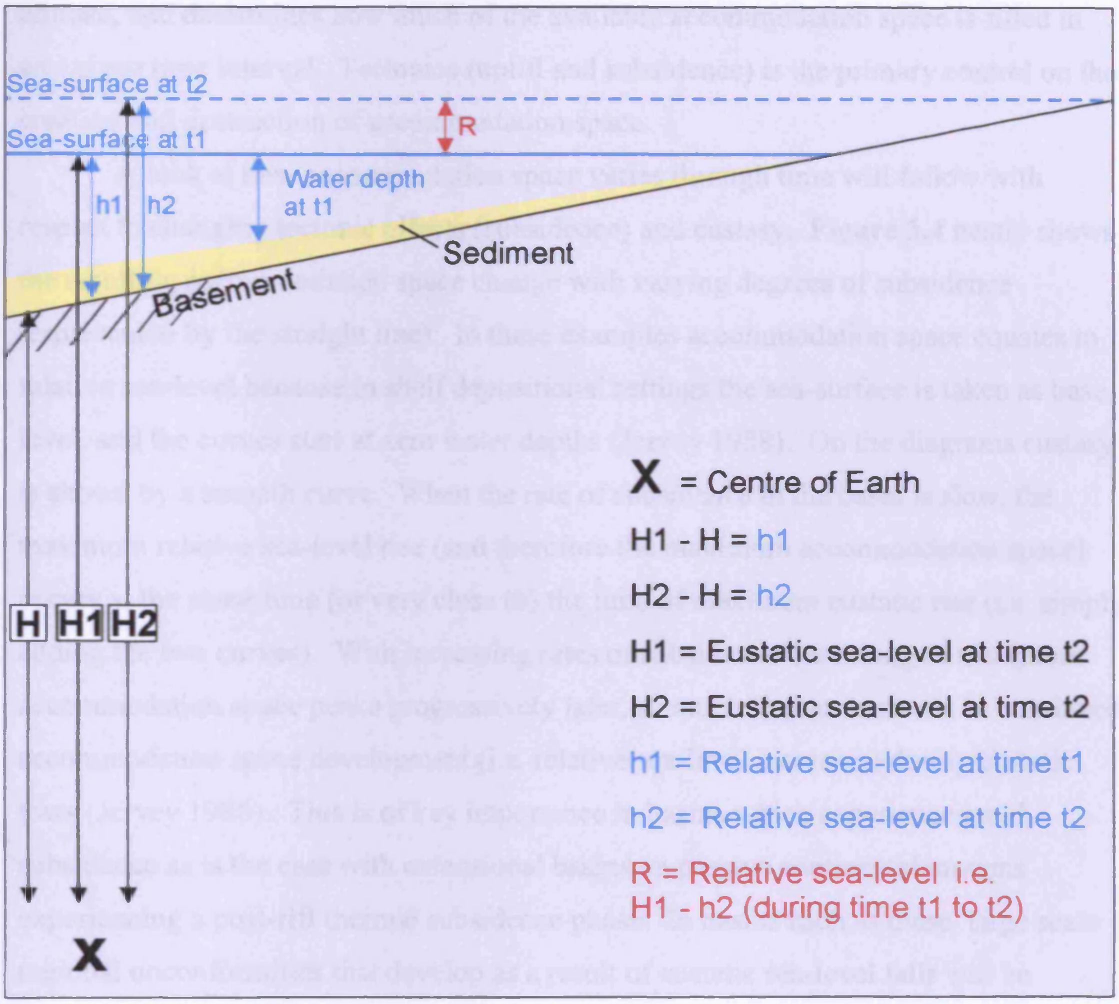


Figure 3.3. Conceptual diagram illustrating the differences between eustatic sea-level, relative sea-level and water depth. Eustatic sea-level is measured form the centre of the earth to the sea-surface at any given time e.g. H1. Relative sea-level is measured from a local datum (e.g. Basement) and is calculated by taking the value for eustatic sea-level at any given time (e.g. At t1 eustatic sea-level = H1) and subtract it from H (a constant) to give h1. Changes in relative sea-level between t1 and t2 give rise to an increase in relative sea-level of value R. Water depth is the distance between the sea-bed and sea-surface at any given time.

changes in tectonics and eustasy and compaction. Not all the accommodation space needs to be filled by sediment: it is simply the amount of space available for potential infill. Sediment supply is also controlled by tectonics and eustasy and, in addition by climate, and determines how much of the available accommodation space is filled in any given time interval. Tectonics (uplift and subsidence) is the primary control on the creation and destruction of accommodation space.

A look at how accommodation space varies through time will follow with respect to changing tectonic effects (subsidence) and eustasy. **Figure 3.4** neatly shows the resulting accommodation space change with varying degrees of subsidence (represented by the straight line). In these examples accommodation space equates to relative sea-level because in shelf depositional settings the sea-surface is taken as base level, and the curves start at zero water depths (Jervey 1988). On the diagrams eustasy is shown by a smooth curve. When the rate of subsidence in the basin is slow, the maximum relative sea-level rise (and therefore the maximum accommodation space) occurs at the same time (or very close to) the time of maximum eustatic rise (i.e. simply adding the two curves). With increasing rates of subsidence the timing of maximum accommodation space peaks progressively later, so much so that there can be continued accommodation space development (i.e. relative sea-level increases) during eustatic lows (Jervey 1988). This is of key importance in basins which experience rapid subsidence as is the case with extensional basins on passive continental margins experiencing a post-rift thermal subsidence phase. In basins such as these, large scale regional unconformities that develop as a result of eustatic sea-level falls will be obscured or even suppressed by the tectonic (subsidence) overprint. A good example of this is the Mid-Oligocene unconformity which is seen globally and well imaged in the Faroe-Shetland Basin (Stoker 1998). This is a classic example which perfectly outlines the difficulty in observing a eustatic signal within the strata in basins where local controlling factors (tectonics) suppress or enhance the global effect.

A brief discussion on the stratal patterns that develop as a result of interactions between eustasy, tectonics and sediment supply (climate) will now follow, focussing on the stratigraphic basin architecture of sequences on continental margins. Sediment supply, entering a basin is controlled by local and regional tectonism and climate. In turn, the rate at which sediment enters the basin controls both how much and where accommodation space develops in the case of a fluvially transported sediment flux for example (Emery and Myers 1996). The main process which allows sediment to enter a

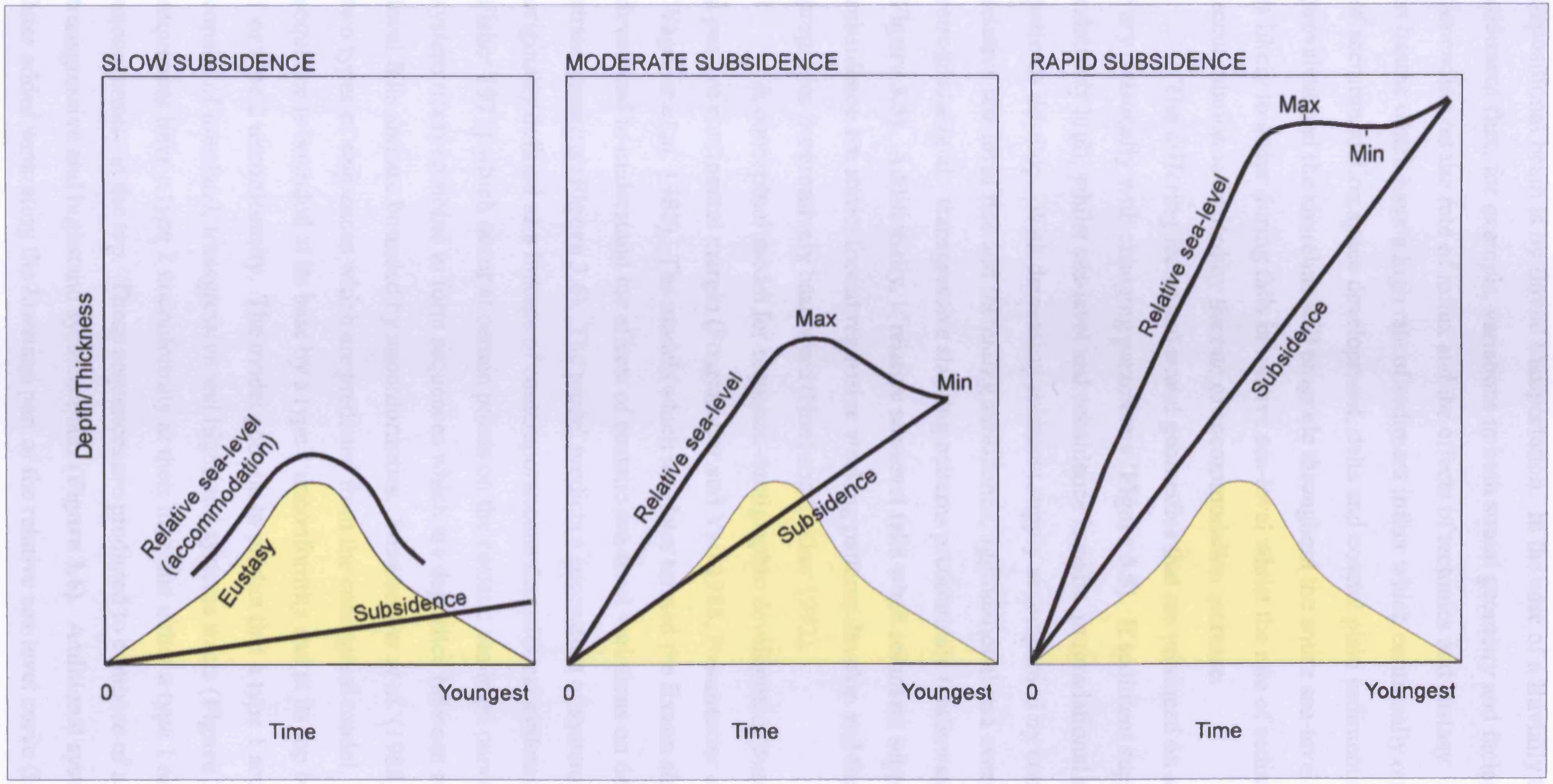


Figure 3.4. Series of diagrams showing how accommodation space development can vary through time with increasing rates of tectonics (subsidence). As subsidence in the basin increases the maximum relative sea-level rise (and hence the time of maximum accommodation space) becomes progressively later. (After Jervey 1988).

depositional basin is by fluvial transportation. In the case of a fluvially transported sediment flux, for example, variations in both stratal geometry and facies can occur depending on the rate of influx and the effects of tectonics and eustasy. For example, in basins which have a high rate of sediment influx which continually outpaces the rate of accommodation space development, delta and coastal plain sedimentation will be prevalent and the shoreline will prograde throughout the entire sea-level cycle. Erosion is likely to occur during falls in relative sea-level whilst the rate of sediment accumulation is limited by the rate of accommodation increase.

The differing facies and stratal geometries that are produced on a shelf margin vary drastically with changing parameters (**Figure 3.5**). If sediment supply remains relatively high, whilst sea-level and subsidence is static, progradational stacking patterns develop. With decreasing sediment supply accompanied by contemporaneous relative sea-level rise and increasing subsidence, aggradational and eventually retrogradational - transgressive stacking patterns predominate (Galloway 1989 and **Figure 3.5**). Additionally, if relative sea-level falls when sediment supply and subsidence are static, forced regressive stacking patterns develop and the shoreline progrades progressively basinward (Hunt and Tucker 1992).

A conceptual model for sequence-stratigraphic development was introduced for a passive continental margin (Posamentier and Vail 1988, Posamentier *et al.* 1988, Van Wagoner *et al.* 1988). The model (which was later termed the Exxon slug model) was developed to understand the effects of eustatic sea-level variations on depositional stratal patterns (**Figure 3.6**). The model predicts a succession of systems tracts originally defined as a linkage of contemporaneous depositional systems (Brown and Fisher 1977) which occur at certain points on the eustatic sea-level curve. These system tracts combine to form sequences which are deposited between relative sea-level falls and are bounded by unconformities. Posamentier *et al.* (1988) recognised two types of sequences which are predicted from the conceptual model. A type 1 sequence is bounded at its base by a type 1 unconformity and at its top by either a type 1 or type 2 unconformity. The model goes on to predict that a type 1 sequence will consist of lowstand, transgressive and highstand systems tracts (**Figure 3.6**). Type 2 sequences have a type 2 unconformity at their base and either a type 1 or type 2 unconformity at the top. These sequences are predicted to compose of a shelf-margin, transgressive and highstand systems tracts (**Figure 3.6**). Additional systems tracts were later added separating the lowstand part of the relative sea-level curve (Hunt and

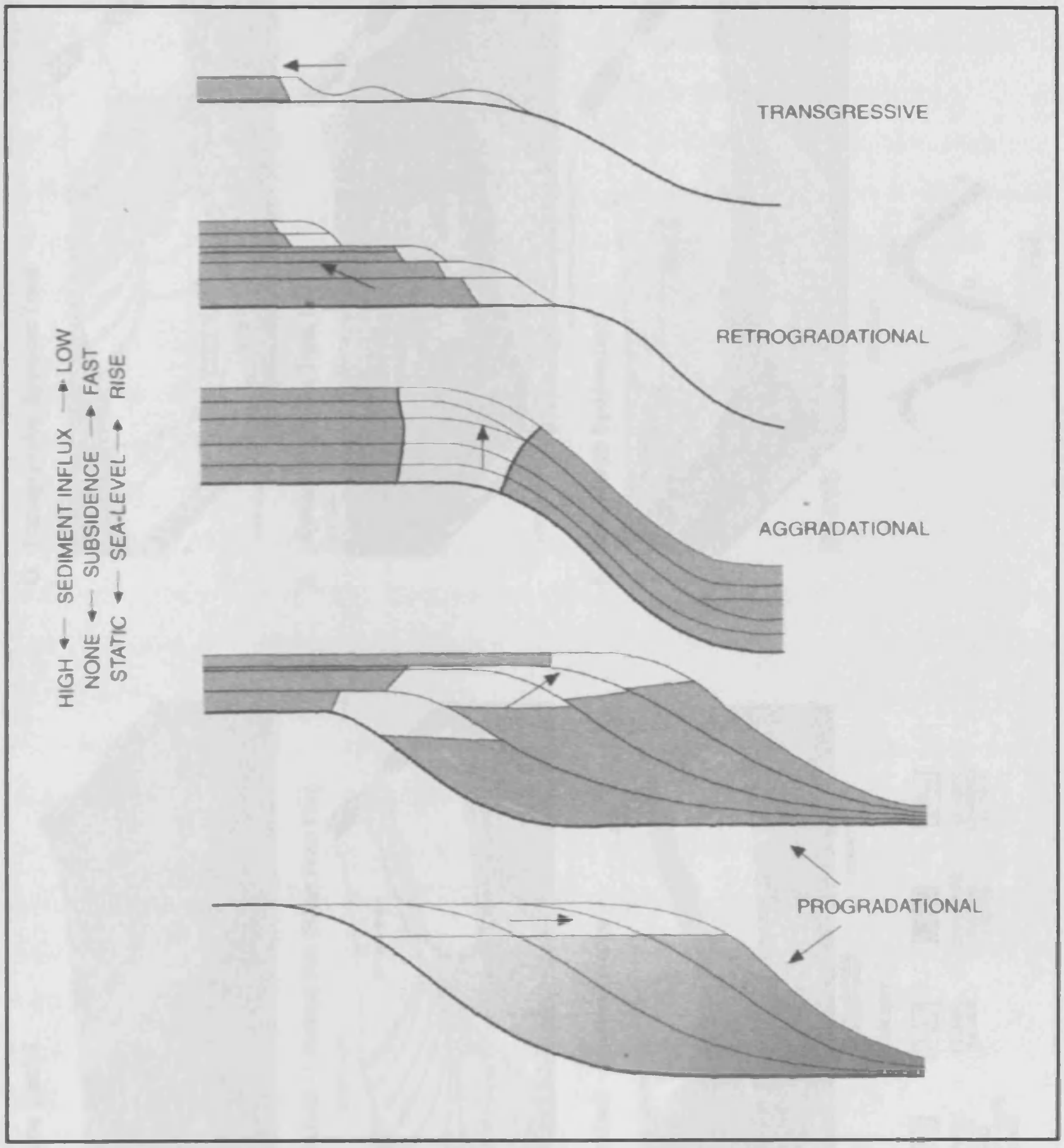


Figure 3.5. Diagram showing the changing depositional architecture on shelf margins as a result of changing sediment flux, subsidence and relative sea-level change. (After Galloway 1989).

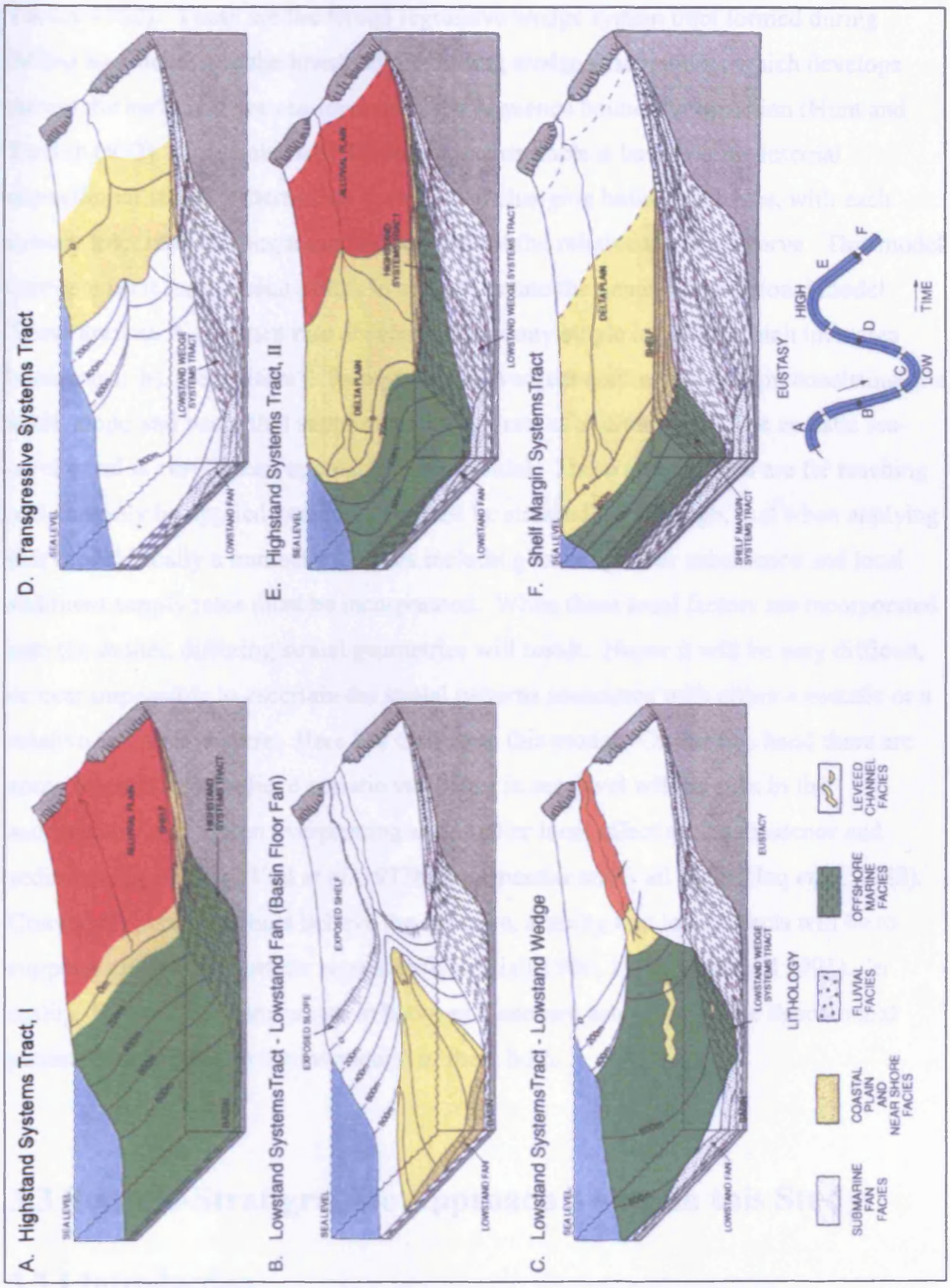


Figure 3.6: Diagram showing the conceptual Exxon model which outlines the sequence stratigraphic development on a passive continental margin. The diagram shows how different facies and systems tracts develop during varying stages of a eustatic cycle. An early highstand succession (A) of prograding clinoforms develops during a eustatic high, which then gets incised during interval B when the eustatic fall exposes the shelf to produce a lowstand fan (and a type 1 sequence boundary). A lowstand wedge develops at the lowest point in eustatic sea level (C), followed by transgression and retrogradation of facies belts during interval D. A second highstand develops with renewed progradation during interval E, and can be followed by a shelf margin systems tract (F) if a fall in eustatic sea level does not expose the shelf. (After Posamentier and Vail 1988).

Tucker 1992). These are the forced regressive wedge system tract formed during falling base-level and the lowstand prograding wedge systems tract which develops during the early relative sea-level rise after sequence boundary formation (Hunt and Tucker 1992). Recognition of different systems tracts is based on the internal depositional stratal patterns that occur due to changing basin conditions, with each system tract representing a specific segment of the relative sea-level curve. This model carries with it many assumptions in order to create the generic depositional model. These are: a). A constant rate of subsidence at any single location, which increases basinward. b). Deposition is focussed on a divergent continental margin consisting of a shelf, slope and basin that supplies a constant rate of sediment. c). The eustatic sea-level trend is curvilinear, approaching sinusoidal. These assumptions are far reaching and can only be applied generally. It must be stressed here though, that when applying this model locally a number of factors including local uplift or subsidence and local sediment supply rates must be incorporated. When these local factors are incorporated into the model, differing stratal geometries will result. Hence it will be very difficult, or near impossible to ascertain the stratal patterns associated with either a eustatic or a relative (local) signature. Here lies the key to this model. On the one hand there are some workers who believe eustatic variations in sea-level will be seen in the sedimentary succession overprinting any smaller local effect such subsidence and sediment supply (e.g. Vail *et al.* 1977b; Posamentier and Vail 1988, Haq *et al.* 1988). Conversely, many authors believe the opposite, arguing that local effects will be to suppress any global eustatic signature (e.g. Miall 1986; 1991, Underhill 1991). In reality the truth lies somewhere in between these two schools with the depositional pattern forming due to the interaction of them both.

3.3 Seismic-Stratigraphic Approach Taken in this Study

3.3.1 Introduction

After outlining the concepts of sequence-stratigraphy, this part of the chapter discusses the methodology taken in this study to sub-divide the Eocene strata of the Faroe-Shetland Basin. The Eocene succession constitutes part of the post-rift phase of basin fill (see **Section 2.4**). This phase of basin evolution represents a time of subsidence after episodic Mesozoic extensional tectonics (see **Section 2.2.3**). Two different seismic-stratigraphic approaches have been employed in this study depending on the nature of the problem which needs to be solved. The methodology of these two approaches will be discussed in the sections below (**Sections 3.3.2** and **3.3.3**). **Chapter 4** focuses on the large-scale regional controls on the basin architecture and thus utilises 2-D seismic data to grossly sub-divide the Eocene succession (**Section 3.3.2**). Conversely, **Chapter 5** details a small part of the southern margin of the basin (the South Judd Basin) and further sub-divides the stratigraphy using high quality 3-D seismic data (**Section 3.3.3**). **Chapter 6** employs both 2-D and 3-D seismic databases.

3.3.2 Regional Two-Dimensional (2-D) Seismic-Stratigraphic Approach

As has been highlighted above in **Section 3.2**, basins that undergo post-rift subsidence after a period of extension, (including the Faroe-Shetland Basin) tend to show little evidence of regional unconformities that are developed due to falls in relative sea-level (e.g. type 1 sequence boundaries). This is because during major phases of subsidence relative sea-level is inferred to rise and thus the development of sequence boundaries (which occur during major relative sea-level falls) is restricted. Therefore any eustatic fall or rise seen with a subsiding basin is suppressed or enhanced respectively. It has therefore been suggested that in many extensional basins around the world, during the post-rift subsidence phase the application of sequence-stratigraphy by conventional methods such as the Exxon model is not entirely appropriate (e.g. Miall 1991, Underhill 1991, Embry 1995). In basins where there is a paucity of regional sequence boundaries an alternative methodology for correlation of post-rift sediments is therefore required.

A simple, seismic-stratigraphic approach has therefore been used in **Chapter 4** of this study to sub-divide the Eocene succession in the Faroe-Shetland Basin. This approach stems from the need to map regionally correlative marker horizons over an area of ca. 100, 000 km^2 in order to establish a chronostratigraphic framework for the basin as a whole, an objective that has not been attempted prior to this study.

This seismic-stratigraphic framework which gives a loosely defined chronostratigraphic breakdown of the Eocene succession has been defined using varying qualities and vintages of 2-D seismic data that was collected and assembled for this study. The seismic data was acquired between 1980 and 1998 hence there is a significant variation in the acquisition and processing methodology, producing data with different resolutions and time/depth ratios. A number of regional correlations which pass through important type wells form the database template for the 2-D basin-wide chronostratigraphic evolution (see **Sections 3.4.3** and **3.5.2** respectively). Wherever possible, these regional seismic correlation panels have used the highest quality 2-D and in many cases this data has been reprocessed to improve its resolution. **Appendices 1** and **2** outline all the details of the seismic data used in this study.

Some of the regional correlations do not directly pass through one or more of the type wells (see **Section 3.5)** and in these circumstances the well is projected onto the seismic correlation using local strike lines. The density of the seismic grid is such that the wells do not need to be projected a considerable distance onto the regional correlations and in most cases the distance is approximately 2 - 3 km.

By mapping the most regionally correlatable seismic reflections and calibrating and tying them using available age data from the wells, the Eocene section can be sub-divided into a number of seismic-stratigraphic units (see **Sections 3.4** and **4.3**). These seismic-stratigraphic units form discrete packages of strata and reflections that to a good approximation represent a unique chronostratigraphic interval. The seismic units are defined and bounded by seismic reflections which are generally strong and continuous and have, over much of their lateral extent a sufficiently distinctive acoustic character, such that they can be accurately correlated. Where continuity is disrupted, the bounding reflections are correlated using an approach known as "phantoming" i.e. interpreting the reflection based on an appraisal of the stratigraphic context with underlying and overlying reflection packages. The bounding surfaces of the seismic units are thus chosen due to the strength and degree of continuity of the seismic reflection and also its geographical distribution. Specific bounding surfaces will be discussed in more detail in **Section 3.4.2.3**.

This technique of phantoming is used over much of the northern Faroe-Shetland Basin which is characterised by moderate to low amplitude, chaotic- discontinuous reflection configurations (**Figure 3.7**). This character is resultant from the development of a major polygonal fault system (Cartwright and Lonergan 1996, Cartwright and Dewhurst 1998). Therefore it is very difficult to trace reflections into this northern area. A correlation can be approximated by using any available well data

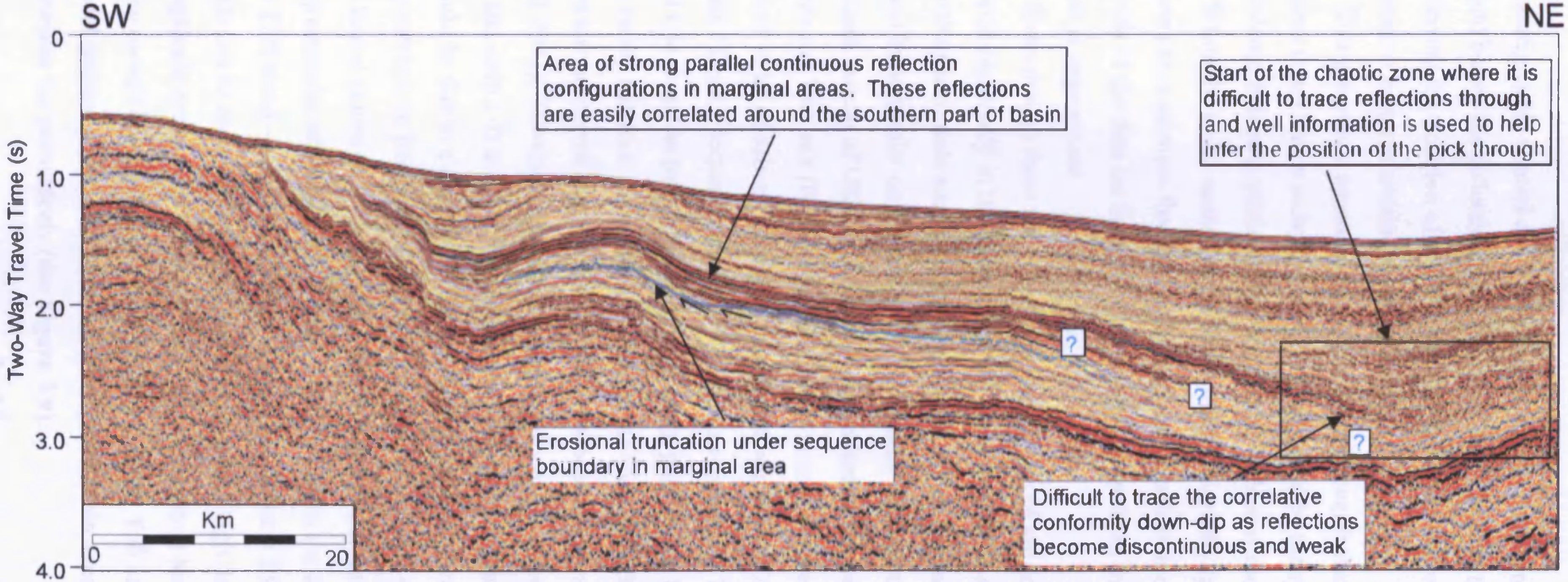


Figure 3.7. Southwest northeast trending 2-D seismic line showing the difficulty in correlating high amplitude reflections including sequence boundaries into its distal (basin-ward) position. On the margins of the basin, the seismic character is one of high amplitude, continuous parallel reflection configurations and localised sequence boundaries can be seen (e.g. blue horizon). When these reflections are traced down-dip into the distal area of the basin, the reflections become low amplitude, chaotic and semi-continuous to discontinuous. In these circumstances it is only possible to infer the position of the reflection by using good well control and by using the general appearance and thickness of the package of reflections and inferring an estimate of the reflection position. This technique is termed "phantoming", (see text section 3.3.2)

for biostratigraphic control and by looking at the general seismic character pattern both above and below the reflection interval as well as the overall thickness of the unit and then inferring the position of the interpretation based on the weak reflection configuration patterns within this chaotic zone.

This pragmatic approach of picking seismic events that are visible, strong and continuous is not a new technique, and is widely used by many seismic interpreters in the petroleum industry, particularly when tasked with constructing regional horizon maps. When the usual methodology of seismic-stratigraphy (picking sequence boundaries or maximum flooding surfaces) is not possible due to the constraints or limitations of the data (as is the case here), picking surfaces that can be tied to available well data is appropriate.

Even though there seems to be no recognition of sequence boundaries that can be mapped regionally in the Faroe-Shetland Basin it is not to say that do not exist. On small 3-D seismic data surveys sequence boundaries can be mapped locally in some parts (and in particular on the margins) of the basin. Most notably, in the southern part of the basin, (south of UKCS Quadrant 4) a localised unconformity can be mapped below the top T50 unit (Balder Formation equivalent) (**Figure 3.8**). This unconformity is believed to be of sub-aerial origin (Smallwood and Gill 2002) and therefore a candidate "Type 1 sequence boundary" of the Exxon model. When the unconformity is mapped it is found to exhibit a dendritic drainage pattern that has been interpreted as an incised valley network. This has been termed the base T50 (Balder) unconformity and signifies a short period of local uplift at the Palaeocene-Eocene transition (Smallwood and Gill 2002). However, when this unconformity is traced away from the 3-D seismic survey and onto 2-D seismic lines, the unconformity is not readily visible (**Figure 3.8**). This could be due to a decrease in the seismic resolution between the 3-D and 2-D seismic surveys, or because the unconformity is only of local significance and has a limited lateral extent (e.g. Cartwright *et al.* 1993). The nearest proxy to the base T50 (Balder) unconformity which has regional significance is the top of the T50 unit (Balder Tuff event - see **Sections 2.2.3.4** and **4.2**). The top T50 (Balder Tuff) reflection can be mapped across the entire basin with a high degree of confidence as a high amplitude continuous seismic reflection. A consistent biostratigraphic signature gives rise to well constrained markers for the top of the T50 unit (Balder Formation). A regional seismic correlation from basin margin to basin axes can be tied to wells and this illustrates the point nicely (see **Figure 3.9**).

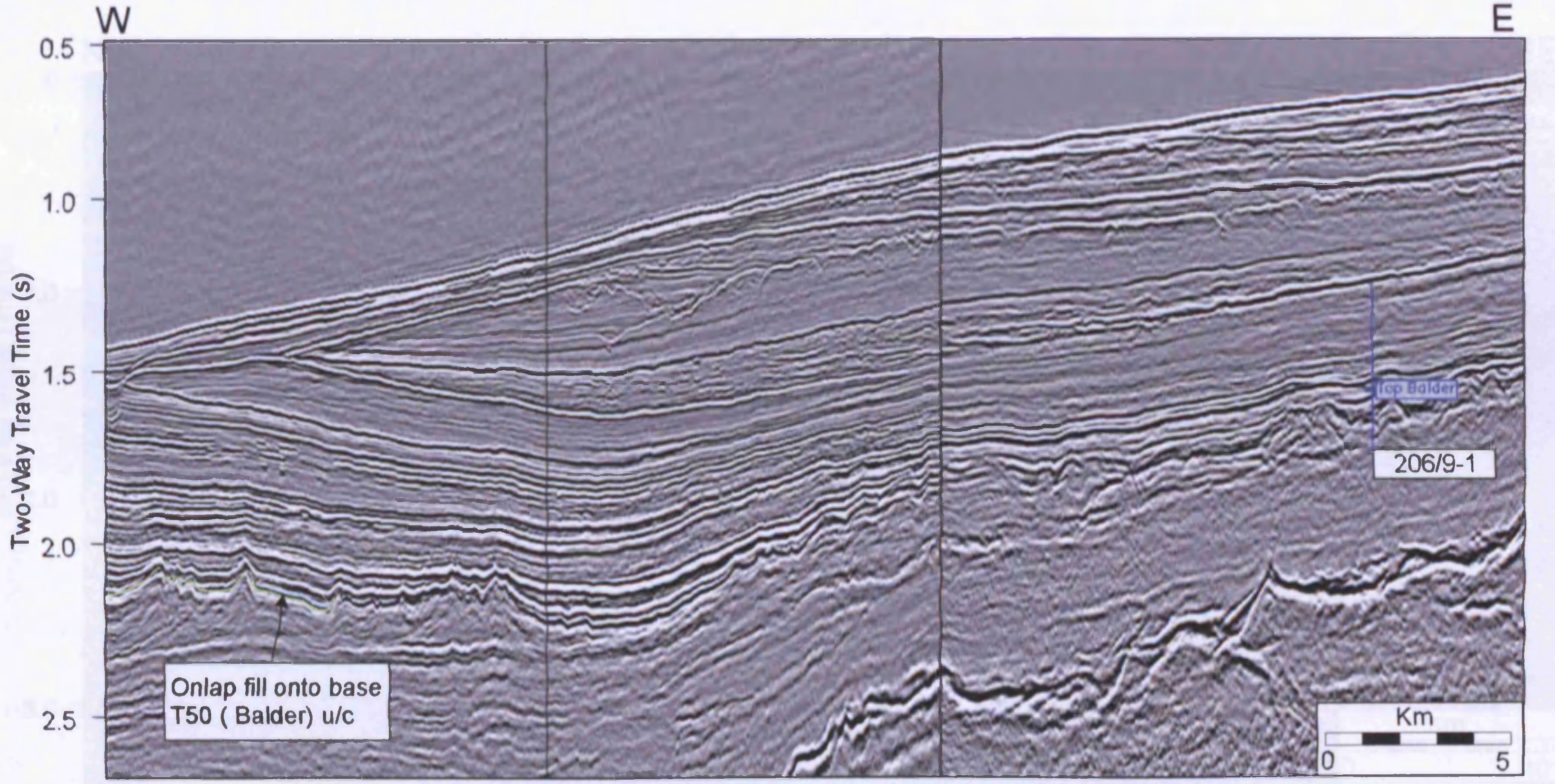


Figure 3.8. Composite 2-D/3-D seismic line (trending roughly east-west) showing the base T50 (Balder) unconformity in the southern part of Quadrant 204. The base T50 (Balder) unconformity maps out to show a dendritic drainage pattern as seen in Figure 2.11. The quality of the 3-D data allows for high vertical and horizontal resolution enabling greater imaging of seismic reflectors. Because of this, sequence boundaries can be mapped on 3-D datasets but there is difficulty when correlating surfaces onto lower resolution 2-D surveys.

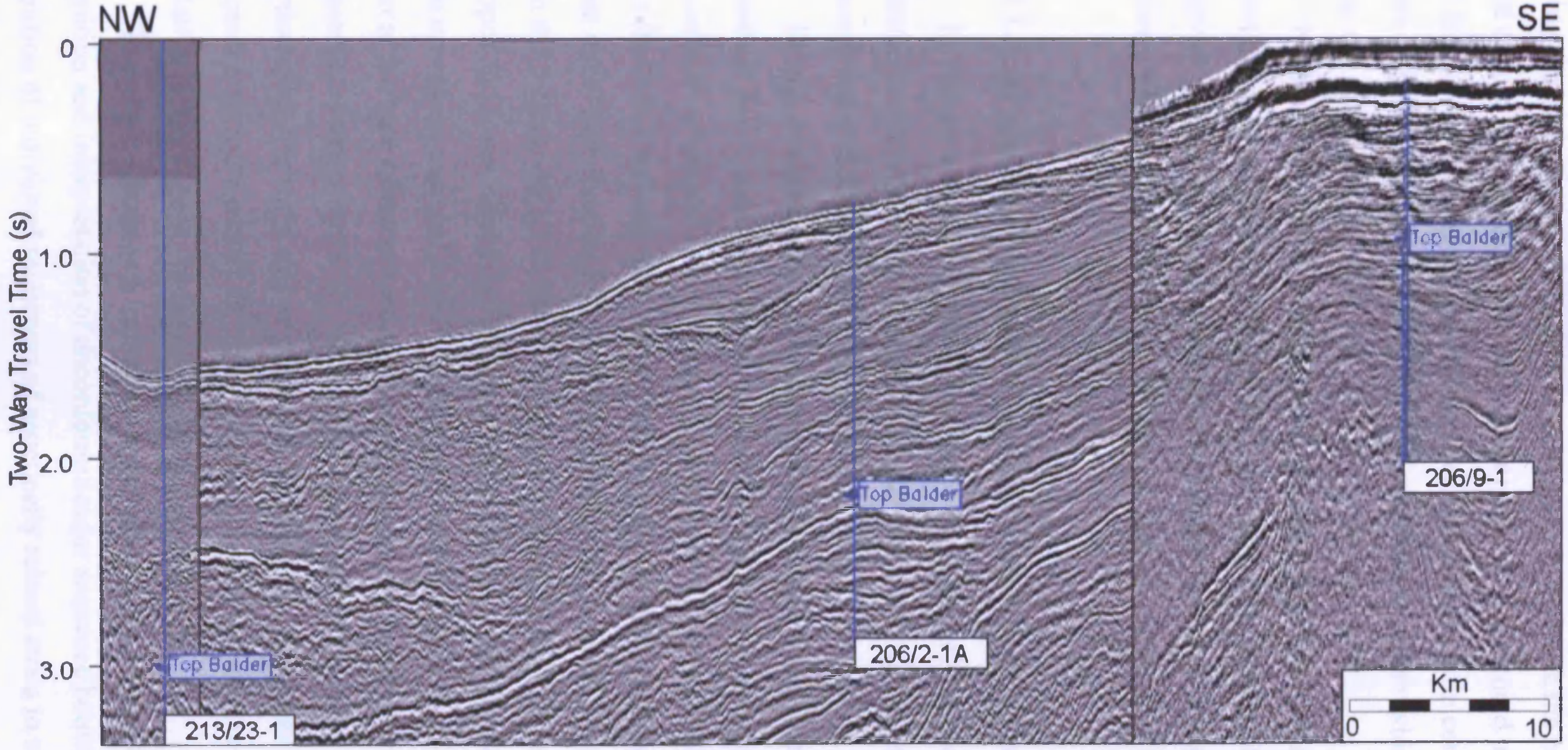


Figure 3.9. Composite 2-D seismic line trending roughly southeast-northwest through three wells from the basin margin to the basin axis. These wells have a good biostratigraphic picks of the Top T50 (Balder) marker and are seen to correlate very well with the seismic data.

A major problem when attempting to correlate seismic reflections over long distances is the change in seismic character of the reflection. Indeed in the case of mapping sequence boundaries, any truncation seen underneath the reflection will be of limited extent and correlation into its correlative conformity could prove problematic. Candidate sequence boundaries that are traced towards the basin centre and often become less discontinuous or exhibit a low amplitude character which is sometimes chaotic and featureless.

Herein lies a critical problem when mapping sequence boundaries. There is an inherent difficulty of correctly interpreting and mapping the correlative conformity throughout the zone of un-correlatable weak seismic character found within large parts of basins away from the margins.

3.3.3 Local Three-Dimensional (3-D) Seismic-Stratigraphic Approach

In **Chapter 5**, high quality 3-D seismic data located in the south of the Faroe-Shetland Basin has been used to sub-divide one particular regional seismic unit discussed above (**Section 3.3.2**).

Detailed seismic interpretation in a localised (small-scale) area is available on 3-D seismic surveys and this allows for a greater resolution of stratigraphic sub-division to be employed. On the margins of the basin the use of 3-D seismic data allows for high amplitude continuous seismic reflections to be recognised and mapped in great detail (see **Section 5.4**). The lateral extent of these reflections can be seen within the 3-D survey and often cover an area of over 15 km^2. These reflections form the upper and lower bounding surfaces of seismic units which are wholly contained within one of the regional seismic units described in **Chapter 4**. Erosional truncation is seen at the base of the reflections in many places and thus they are interpreted to represent local unconformities. Away from these areas of incision, the seismic reflection often becomes weaker in seismic amplitude and is not as easily correlated. The lower amplitude part of the reflection shows no sign of erosion and is thus the correlative conformity to the lateral erosional unconformity.

Therefore, **Chapter 5** builds a seismic–stratigraphic framework based on the recognition and interpretation of unconformities (or sequence boundaries) and uses the recognition of individual sequences of genetically related strata to sub-divide the stratigraphy in a small part of the basin. This in turn allows for the construction of a stratigraphy in a small part of the basin. This in turn allows for the construction of a chronostratigraphic evolution of the basin that follows the classical approach of stratigraphic sub-division pioneered by Exxon.

3.3.4 Seismic Interpretation Methodology

The conventional approach of seismic interpretation has been used in this study by interpreting seismic reflection configurations that possess good continuity and correlatability throughout the greater part of the basin. Seismic markers are defined and picked on the highest amplitude part of the seismic reflection wavelet (both positive and negative). However, because of the vast number of 2-D and 3-D surveys used, the polarity of the data often changes between surveys. Therefore when tying regionally correlatable seismic markers throughout the entire basin this polarity shift needed be taken into account.

All the seismic data (both 2-D and 3D) has been interpreted in the time domain. No depth conversion of the seismic data has been completed in this study. Throughout **Chapters 4, 5** and **6**, seismic lines are shown with marker horizons colour coded for ease of recognition and continuity (e.g. see **Section 4.3.1** and regional correlations).

For the 2-D regional seismic mapping, once a particular reflection was defined (based on its extensive correlatability or because it showed evidence of reflection terminations such as onlap or erosional truncation, the seismic marker was then traced laterally throughout the basin using the majority of the 2-D lines available. On the 3-D seismic data (depending on the particular feature being mapped) an initial grid of typically every tenth in-line and cross-line was interpreted (giving a spacing of 125 m). Once the manual interpretation was completed, various auto-correlation tools were utilised with varying parameters to infill the correlation. This technique is good for finalising time structure maps once the geological feature being mapped is understood and the auto-correlation tool using the manual pick of the interpreter as a constraining parameter.

The degree to which auto-correlation is necessary depends on numerous factors. If a seismic reflection is easy to pick with confidence, because of the strong, continuous (possibly high amplitude) character of the reflection (e.g. the sea-bed reflection or top T50 (Balder Tuff) event), then to get a very detailed map of the horizon the interpreter only needs to pick in-lines and cross-lines on a much coarser grid of every twentieth –

fiftieth line. However, in both **Chapters 5** and **6**, many subtle semi-continuous to discontinuous reflections have been mapped and these have been picked on a much closer spacing of one - two lines.

Sequence boundaries, by their nature, cut out stratigraphy and therefore are rarely represented as discrete, continuous seismic reflections, but are more commonly identified from stratal reflection terminations (e.g. Vail *et al.* 1977, Mitchum and Vail 1977). It can be very difficult therefore, to pick sequence boundaries using the automatic picking tool and they are interpreted to a better quality when mapped entirely manually by the interpreter, or are picked either just above or below the actual seismic event. In summary, when a seismic reflection is defined as a horizon to be mapped, the interpreter will decide, using many of the criteria above if the reflection needs to be mapped in great detail on every line or whether it is possible for the automatic picking tool to produce an equally good quality map.

Horizons which have been interpreted on 2-D data have been picked and tied on many lines in the 2-D surveys. Mis-ties in the interpretation were very common and this is primarily due to the different parameters of 2-D seismic migration and acquisition employed. The mis-ties could also be due to correlation error as well as external factors such as sea state and water temperature. The mis-ties were eliminated by an iterative process of incremental shifts of the seismic data. The methodology of picking dip-orientated lines and then strike-orientated lines was used to provide a detailed grid of seismic interpretation which could be checked by an iterative methodology which aids in the recognition of mis-ties.

From the horizons interpreted within the entire Eocene succession a suite of maps have been produced to show either the gross geometry of the basin (**Chapter 4**) or individual features on a more local scale (**Chapters 4, 5** and **6**). Time structure maps are the first maps to be created from the seismic interpretation. These maps are created by gridding and contouring the depth (in time- quantified in seconds and milliseconds) to create a contoured surface of the horizon. This surface can then be displayed using a spectrum of chosen colours to show the observed relief (in time) on the horizon.

Secondly, isochron maps can be produced to show the preserved thickness in time between two chosen seismic horizons. The production of isochron maps is a good technique to show where the distribution of a specific interval is located. The map additionally shows both the thickest and thinnest areas of preserved section (in time).

However, this cannot be interpreted to represent the location of significant depocentres or areas of little or no deposition as the effects of erosion are not taken into account. When calibrated with seismic data however (to interpret internal seismic reflection configurations and depositional environments), the maps do allow for a more valid interpretation of the specific feature of interest be that on a basin scale or in a more local setting. Basin-wide time structure and isochron maps are presented in **Chapter 4** and along with interpretation of 2-D seismic lines and well data form the basis of the palaeogeographic maps shown in the chapter.

In **Chapters 5** and **6** high quality 3-D seismic data is used to look in detail at small areas of the basin. Because of the availability of 3-D data in these two case studies, additional interpretative techniques can be utilised allowing further interrogation of the data in order to further understand the stratigraphic significance of local features too small to be seen on the 2-D seismic dataset. These techniques allow different seismic attributes to be displayed in map form and interpreted.

The first of these 3-D based techniques is seismic amplitude extraction. This technique is useful as it allows the display of the seismic amplitude value which can be used for recognition of subtle changes in the amplitude not necessarily seen on vertical seismic lines. There are many types of seismic amplitude which can be used, indeed the amplitude of the reflection itself can be maximised, minimised, averaged or root mean squared (e.g. Brown 1996). Furthermore, a time window above or below the reflection (or both) can be selected and the required amplitude extraction value displayed. In **Chapters 5** and **6** the simplest case scenario of amplitude extraction is used which takes the actual value of seismic amplitude from the reflection itself.

Maps which display the magnitude of dip of the surface are another technique which can be used in order to recognise subtle features within the 3-D surveys. In particular, these dip maps can show the sharp edges of features especially where little or no change in time and amplitude is seen. These maps therefore depict faults and lineaments very well and subtle mounded features (e.g. see **Chapter 5** and **Figure 5.1** and Brown 1996, Hart 1999).

A third technique which can be used is seismic flattening. This allows any folding of the strata to be removed at a certain chosen stratigraphic level. This has been particularly invaluable in determining the depositional history and environmental setting in the South Judd Basin (**Chapter 5**) where a large east – west trending anticline (the Judd Anticline) folds the Early Eocene succession. By removing this fold (which can be dated by its onlap) the original true depositional dip and breaks in slope prior to any compression can be restored (see **Section 5.3.3**).

Two final seismic attribute techniques have been used in this study to enhance the understanding of the depositional systems seen within the basin. The first of these is horizontal time-slicing of the 3-D seismic data which is particularly useful for visualising features in planform. This technique can be combined with flattening to slice horizontal through particular seismic reflections at specific time intervals (e.g. see **Section 6.3**). Secondly, the generation of coherency (or variance) volumes has been successful in identifying very small scale subtle features on 3-D surveys. This technique takes the original (migrated) survey and crates another volume which looks at the continuity of the traces and compares each individual trace with its neighbours. Again, this technique brings out small faults and subtle depositional features such as channels. The principles used in the 3-D mapping are summarised in Brown 1996.

3.4 Eocene Seismic-Stratigraphic Units

3.4.1 Introduction

The following section will establish the rationale underlying the method of stratigraphic sub-division of the Eocene strata in the Faroe-Shetland Basin. **Chapter 4** goes on to describe and interpret the evolution of the basin as a whole. The term seismic-stratigraphic unit is used here to define basin-wide packages of strata that are bounded by the most prominent and regionally correlatable chronostratigraphic surfaces (i.e. seismic reflections). These seismic-stratigraphic units form the building blocks which are described in detail in **Chapter 4**. Each individual seismic-stratigraphic unit is bounded above and below by seismic reflections which have the greatest regional correlatability.

The methodology of choosing these candidate seismic markers was established using the criteria discussed above (see **Sections 3.3** and **3.3.2**). This section will go on to look how the bounding seismic markers were defined and introduce the regional correlations (which are shown in **Chapter 4** and included as enclosures at the back of the thesis) which display the chronostratigraphic seismic-stratigraphic units of the Eocene succession on a basin-wide scale.

3.4.2 Definitions of Bounding Seismic Markers

3.4.2.1 Seismic Character

The seismic markers which bound the seismic units were chosen primarily on their acoustic character and lateral extent. Seismic reflections need to be continuous or semi-continuous and have a characteristic seismic character in order to allow for regional correlation. Examples of candidate regional seismic markers are shown in **Figure 3.10**. However, because of the long-range correlations (greater than 400 km) which are undertaken when sub-dividing the Eocene succession of the entire basin it is implausible to expect individual seismic reflections to show such correlatability. Indeed in particular parts of the basin, the data quality is very poor (see **Figure 1.3**) and correlation of reflections is very difficult. This is particularly the case in the northern part of the basin which is dominated by discontinuous, low amplitude seismic reflection configurations which are polygonally faulted (e.g. Cartwright and Lonergan 1996). Because of the poor data quality in this northern area the bounding seismic markers cannot be defined here.

In conjunction with the acoustic character, the stratal terminations are considered when determining the bounding seismic markers to the seismic units (**Figure 3.10**). The high amplitude, continuous, regionally correlatable reflections are often show diagnostic stratal terminations against them. Erosional truncation at the base of high amplitude reflections is seen, as well as onlap and downlap (**Figure 3.10**). Thus this suggests that these regionally correlatable seismic reflections have stratigraphic significance and in some cases represent more than just time-lines. Furthermore, in certain areas of the basin these seismic markers often separate intervals of varying seismic character, and thus are often readily visible (**Figure 3.10**).

These two techniques for qualitatively recognising regional seismic markers has enabled a chronostratigraphic sub-division of the succession to be built from detailed interpretation and correlation of regional to semi-regional 2-D seismic lines.

3.4.2.2 Geographical Location

Most candidate seismic markers which sub-divide the seismic units were defined in the south of the basin in Quadrant 204 where the Eocene almost subcrops the sea-bed and where resolution and data quality are excellent. This shallow position of

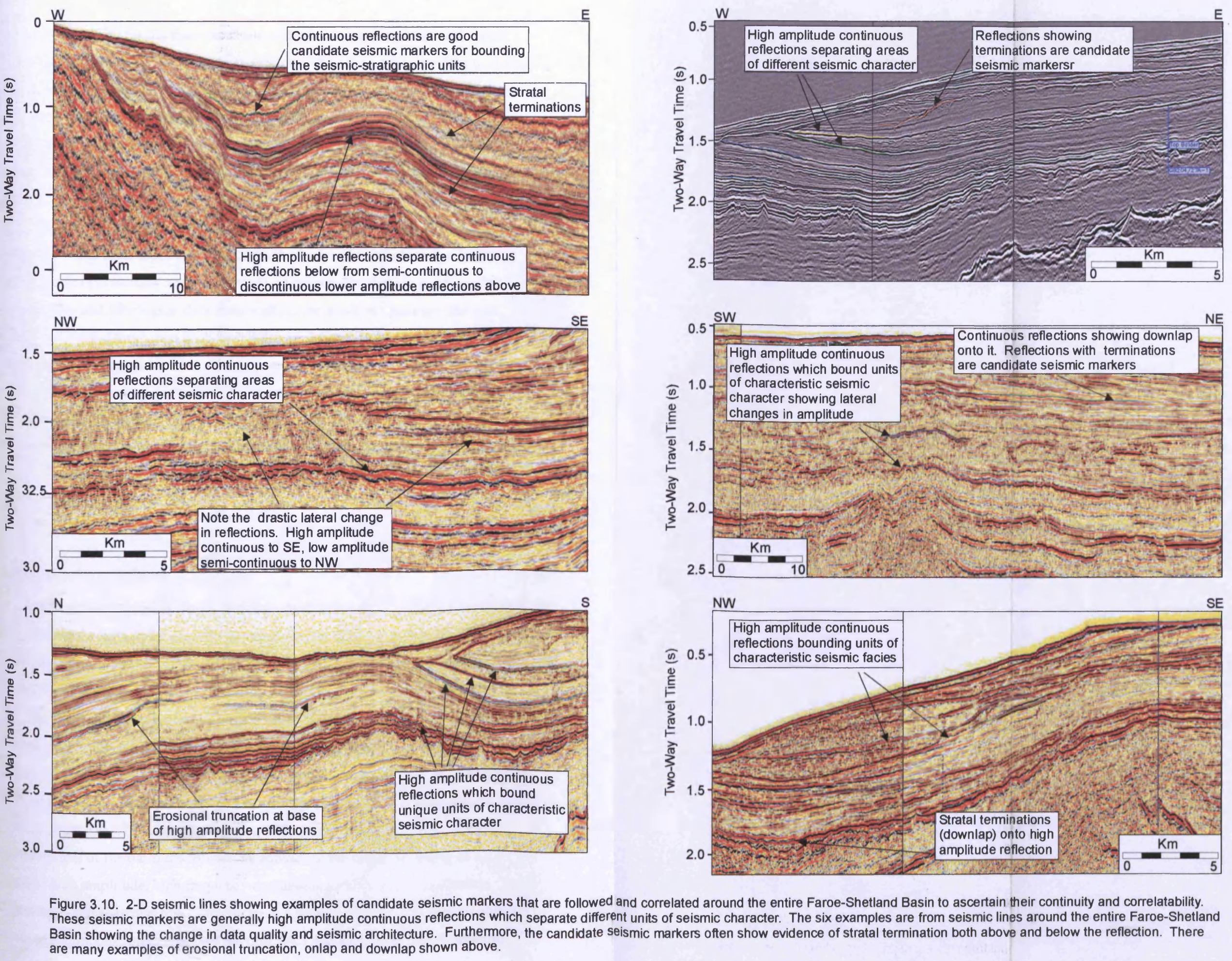


Figure 3.10. 2-D seismic lines showing examples of candidate seismic markers that are followed and correlated around the entire Faroe-Shetland Basin to ascertain their continuity and correlatability. These seismic markers are generally high amplitude continuous reflections which separate different units of seismic character. The six examples are from seismic lines around the entire Faroe-Shetland Basin showing the change in data quality and seismic architecture. Furthermore, the candidate seismic markers often show evidence of stratal termination both above and below the reflection. There are many examples of erosional truncation, onlap and downlap shown above.

the Eocene succession results from uplift and related exhumation on the east - west trending Judd Anticline due to compressive tectonic episodes in the Late Cenozoic (e.g. Davies *et al.* 2004, Boldreel and Andersen 1998). Secondly, its location is where the present day Faroe-Shetland Channel narrows before turning westwards into the Faroe-Bank Channel. At this location, highly erosive bottom currents occur and a series of dramatic cliffs known as the Judd Deeps have developed on the sea-floor (Smallwood 2004).

The Eocene seismic reflection configurations are very well imaged on the seismic data in this area, especially on three 3-D surveys (one of which has been reprocessed with the shallow section in mind). Additionally, there is good quality biostratigraphic and lithological data from wells in the south of Quadrant 204 (see **Section 3.5**). Furthermore, two BGS Boreholes are located in this area giving further age and lithology calibration. This southern area of the basin was therefore chosen as the best area to pick strong, continuous reflections which have good correlatability based on high quality seismic and biostratigraphic data. From this southern region, the bounding seismic markers were then extended by correlation to the north and east (into poorer quality data regions), primarily using long regional strike lines along the basin axis as these give the highest continuity response.

3.4.2.3 Bounding Surfaces

Five bounding surfaces which form the upper and lower boundaries to regional (basin-wide) scale units will be briefly introduced here. The amplitude, continuity and lateral variability of the reflections will be discussed here as well as the internal reflection terminations.

The basal bounding surface is the top T50 unit (Balder Formation) equivalent. (see **Sections 4.2** and **4.3** for definition of T50 unit). Throughout the entire basin the seismic reflection is seen as a high amplitude continuous negative seismic trace (shown on seismic panels as a black reflection). On all seismic panels shown throughout this thesis, this reflection will be highlighted as a white horizon and labelled Top Balder. In the southern half of the basin the reflections appears as the upper bounding of a package of high amplitude, high frequency continuous parallel seismic reflection configurations and separates this character from a weaker (moderate to low amplitude package of continuous reflections (see **Section 3.4.3** and **Figure 6.11 and regional**

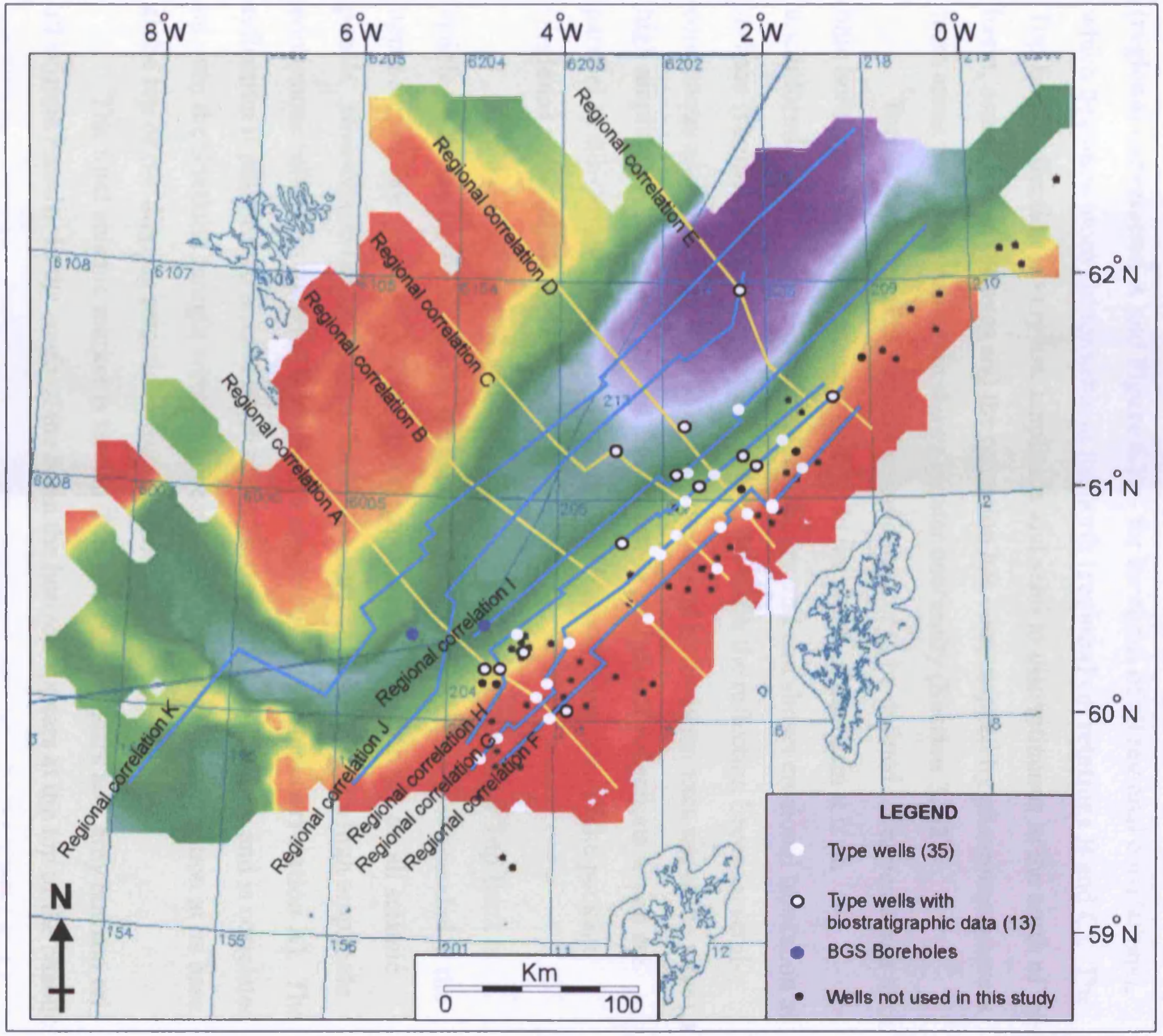


Figure 3.11. Location map and Time structure map (in TWTT) of the seabed in the Faroe-Shetland Basin, showing the positions of the eleven regional correlations. These regional correlations incorporate good quality 2-D and 3-D seismic data that have been used to subdivide the gross geometry of the Eocene succession into seismic units. The eleven correlations transect the entire basin and tie together thirty five important type wells (white dots) and two BGS boreholes (dark blue dots). Five dip orientated correlations trend SE-NW (highlighted in yellow (A-E)), and six strike orientated correlations trend southwest - northeast (highlighted in blue (F-K)). Note the skewed nature of the well locations to the UK sector (see text for explanation-section 3.4).

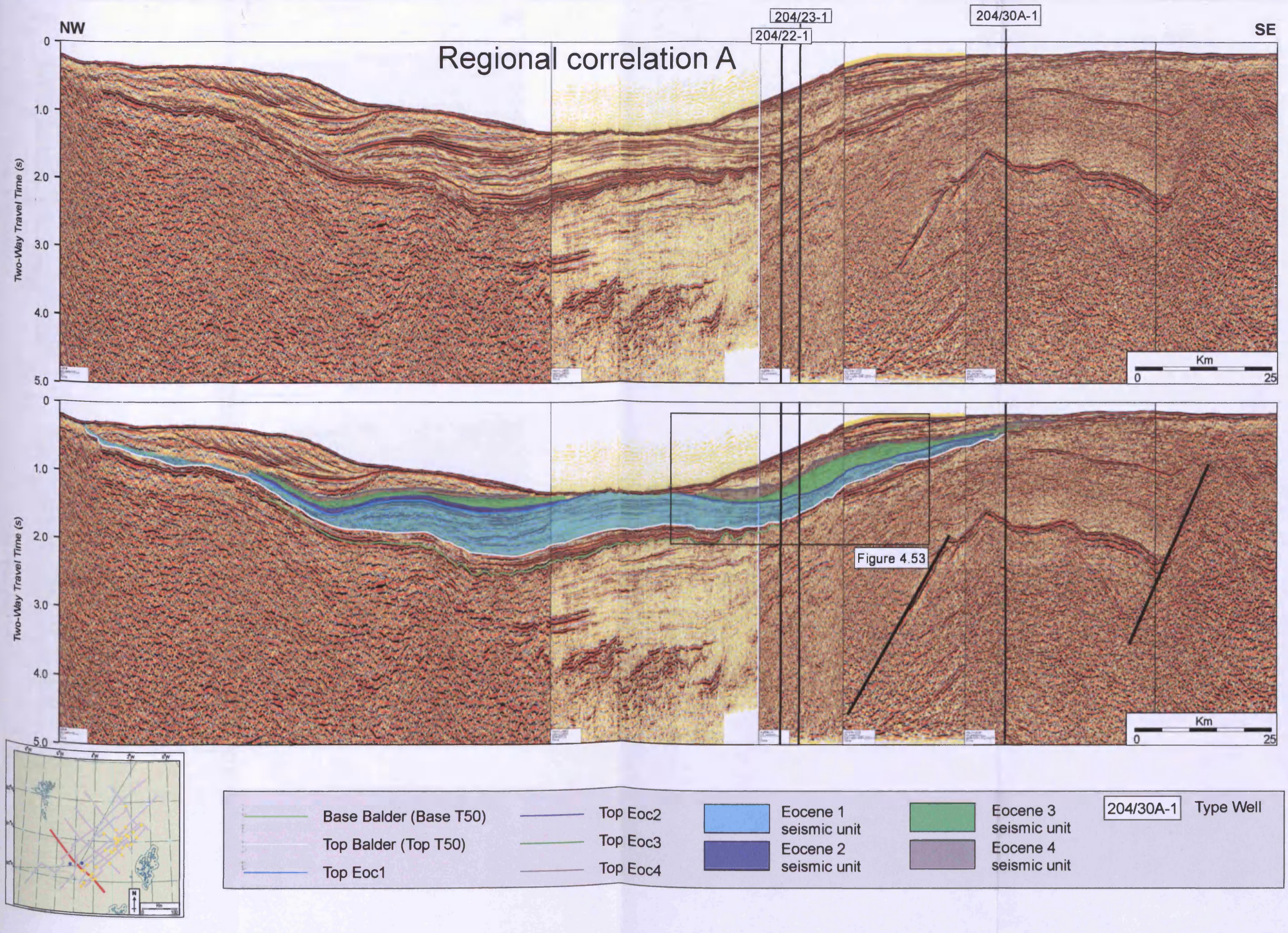
correlation A). Towards the central and northern parts of the basin the reflection is found at the top of a high amplitude discontinuous and incoherent seismic facies above which are more continuous higher frequency, more coherent reflections (see **regional correlations B** and **C** and **Figure 6.11**).

The second bounding surface is termed Top Eoc1 on all seismic panels and occurs as a pale blue horizon. It is defined by a high amplitude continuous reflection in the south of the basin and sits at the top of a high frequency, continuous moderate to high amplitude package of reflection configurations. Towards the Shetland margin this reflection is seen to thin and onlap onto the lower "Top Balder" reflection. Conversely, towards the Faroe-margin, downlap onto this reflection is seen in the south of the basin (**regional correlation A** and **Figure 6.11** - for locations of all regional correlations), which becomes more conformable to the north (**regional correlations B** and **C**). The Top Eoc1 reflection is very low amplitude and semi to discontinuous in the north of the basin, especially in the axes and the reflection has been mapped by phantoming across from areas where the reflection shows greater continuity (**Section 3.3.2**).

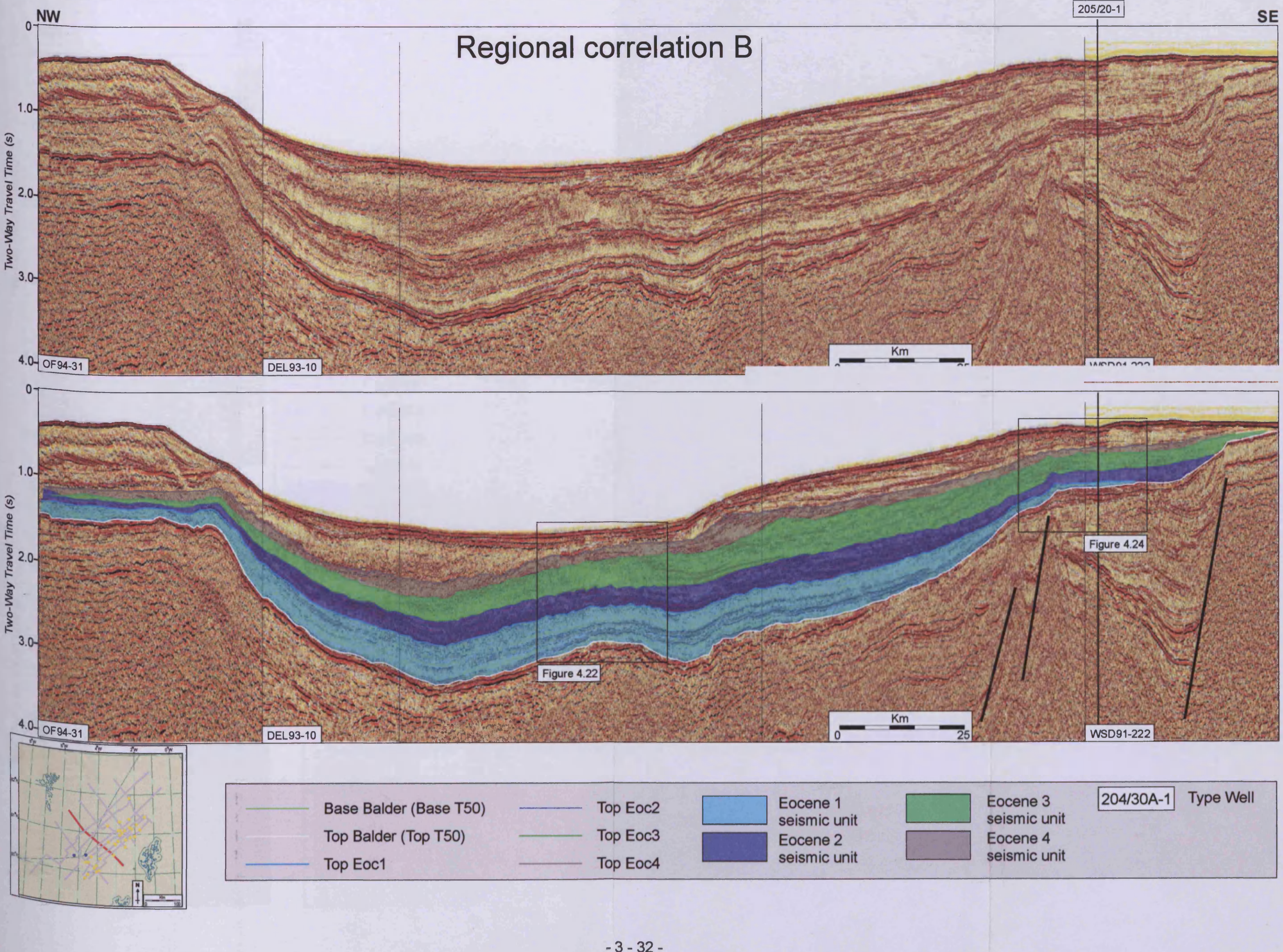
Top Eoc2 is the next regional seismic marker to be defined. It appears as a dark blue horizon on all seismic panels herein. In the south of the basin it is an unconformity which down-cuts into the younger unit and shows erosional truncation at its base (**regional correlation A**). Towards the north the reflection becomes semi-continuous and onlaps onto the Shetland margin and in the basin axes sits at the top of a high amplitude semi-continuous to discontinuous package of reflections which are parallel to sub-parallel and show more continuity towards the base of the package (**regional correlation B**).

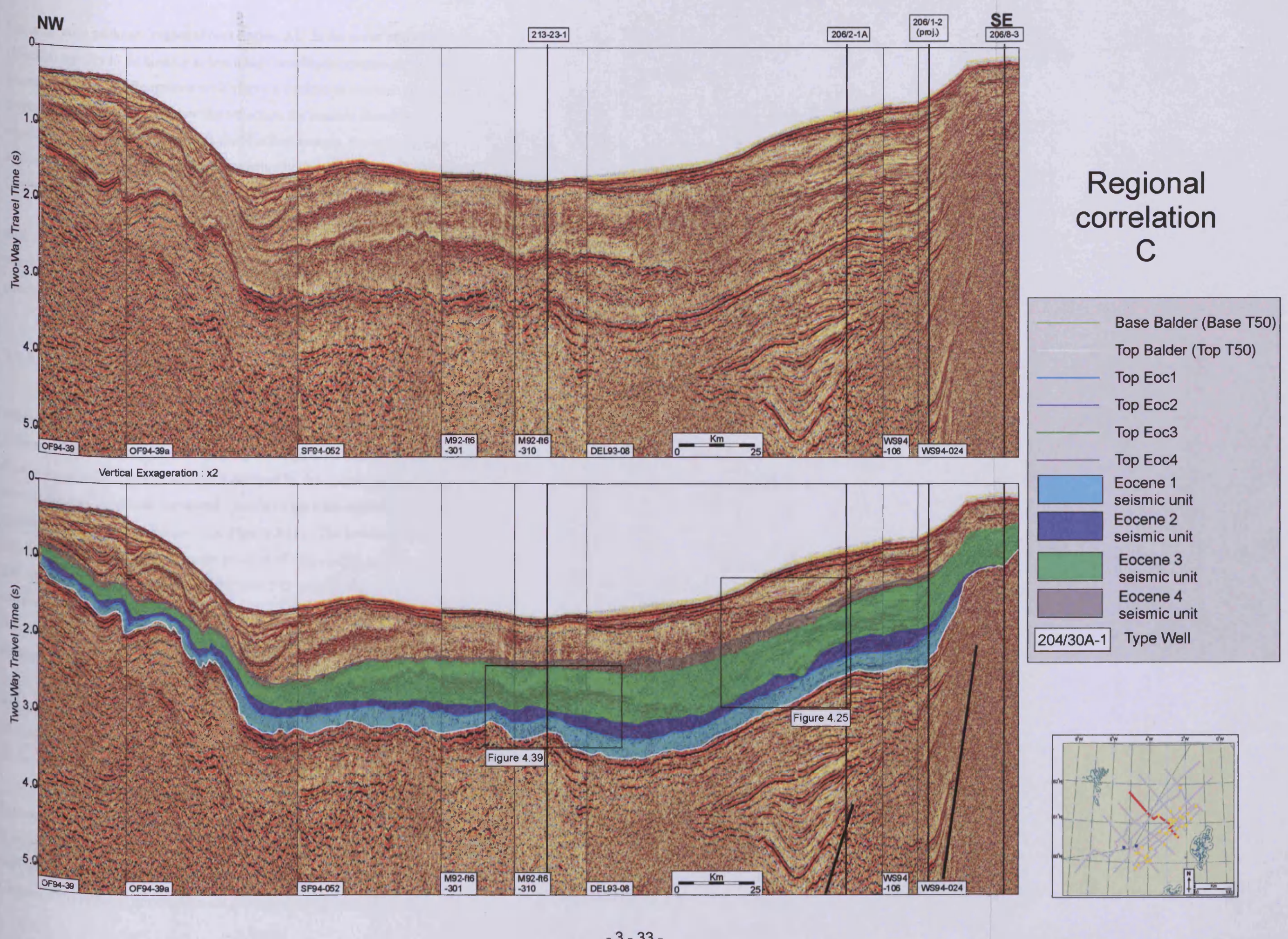
A package of dipping seismic reflection configurations above Top Eoc2 is visible on the entire southeastern margin of the basin. This package is bounded at the upper surface by Top Eoc3. This horizon appears as a green horizon on all seismic panels. However in the south of the basin, the horizon is defined as a high amplitude continuous reflection found at the base of a chaotic unit (**regional correlation A**). The reflection is phantomed around much of the northern half of the basin, and is correlated up onto the Shetland margin where it appears to exhibit erosional truncation at its base at the top of the dipping seismic reflection packages.

The final seismic marker is termed Top Eoc4 and appears as a grey horizon on all seismic panels. In the south of the basin the horizon appears at the top of the chaotic



NW
SE
Regional correlation A
204/23-1
204/22-1
204/30A-1
Two-Way Travel Time (s)
0
1.0
2.0
3.0
4.0
5.0
Km
0
25
Figure 4.53
Base Balder (Base T50)
Top Balder (Top T50)
Top Eoc1
Top Eoc2
Top Eoc3
Top Eoc4
Eocene 1 seismic unit
Eocene 2 seismic unit
Eocene 3 seismic unit
Eocene 4 seismic unit
204/30A-1 Type Well


NW
SE
205/20-1
Regional correlation B
Two-Way Travel Time (s)
0
1.0
2.0
3.0
4.0
OF94-31
DEL93-10
WSD91-222
Km
0
25
Figure 4.22
Figure 4.24
Base Balder (Base T50)
Top Balder (Top T50)
Top Eoc1
Top Eoc2
Top Eoc3
Top Eoc4
Eocene 1 seismic unit
Eocene 2 seismic unit
Eocene 3 seismic unit
Eocene 4 seismic unit
204/30A-1 Type Well


NW
SE
213-23-1
206/2-1A
206/1-2 (proj.)
206/8-3
Two-Way Travel Time (s)
OF94-39
OF94-39a
SF94-052
M92-ft6 -301
M92-ft6 -310
DEL93-08
Km
0
25
WS94 -106
WS94-024
Vertical Exxageration : x2
Figure 4.39
Figure 4.25
Regional correlation C
Base Balder (Base T50)
Top Balder (Top T50)
Top Eoc1
Top Eoc2
Top Eoc3
Top Eoc4
Eocene 1 seismic unit
Eocene 2 seismic unit
Eocene 3 seismic unit
Eocene 4 seismic unit
204/30A-1
Type Well

high amplitude package (**regional correlation** A). In the north of the basin the reflection appears to lie directly below a high amplitude continuous, parallel unit of seismic reflection configurations which have a distinctive wavy or crenulated form (**regional correlation E**). Below the reflection the seismic character is chaotic, low amplitude and featureless. Towards the Shetland margin, the horizon can be traced up a series of dipping reflections where it eventually thins by onlap onto the Top Eoc3 (**regional correlations D** and **E**).

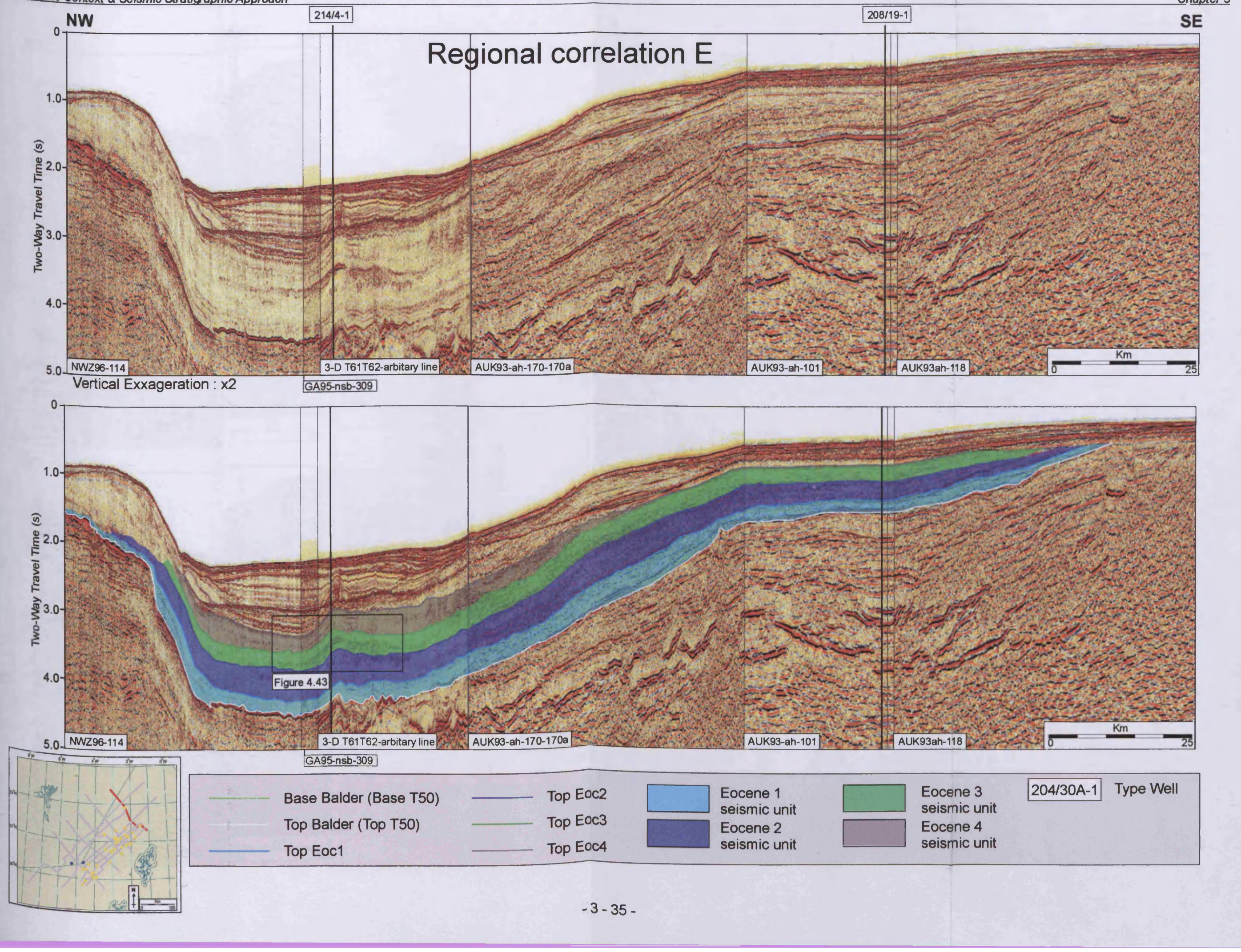
A significantly more detailed analysis of the five regional seismic markers and the intervening seismic units follows in **Chapter 4 (Sections 4.2** and **4.3.2** – **4.3.5)** where they are interpreted and described with regional correlations, local seismic panels and calibrated with lithologic and age data.

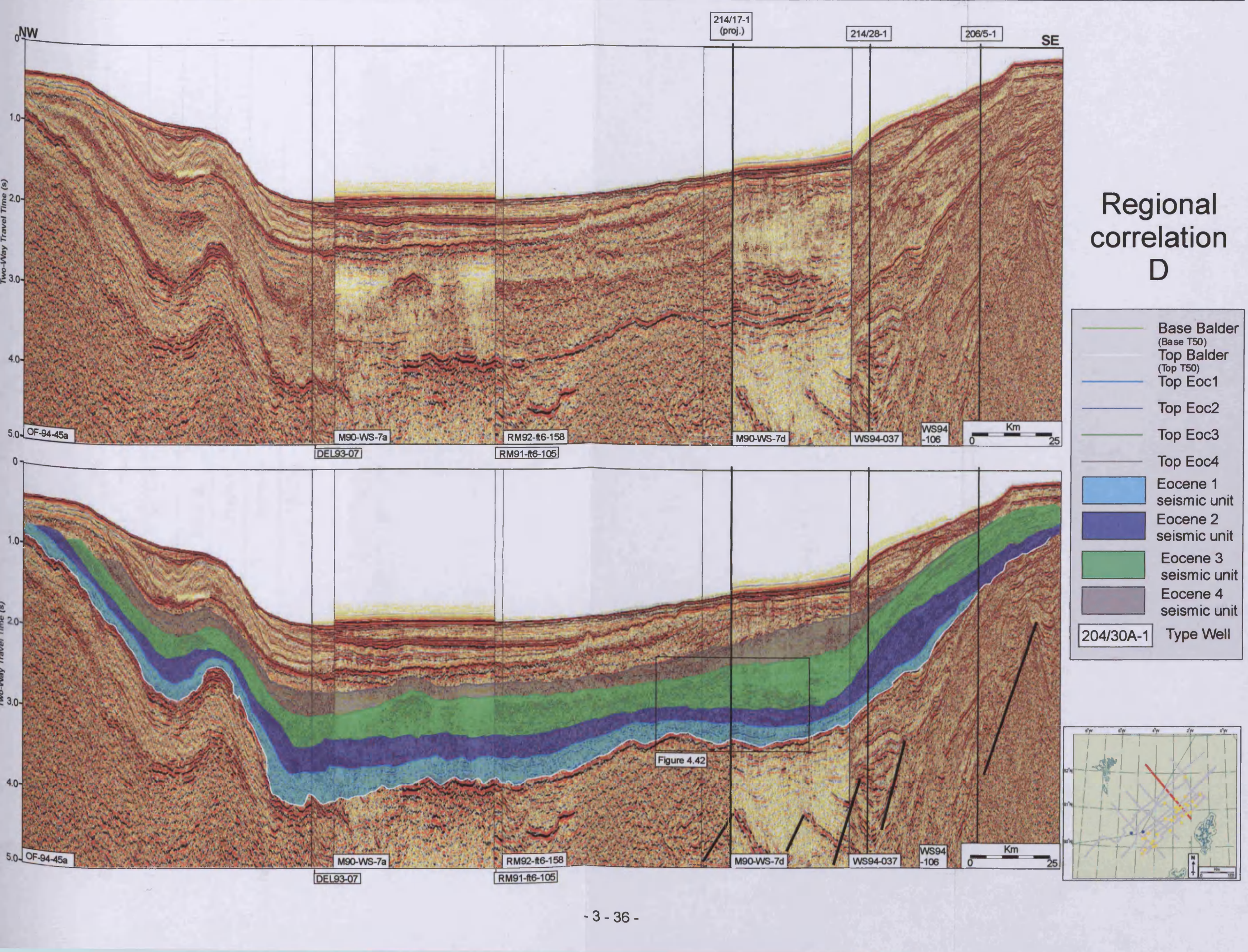
3.4.3 Regional Correlations

This section of the chapter introduces the regional seismic correlations which transect the Faroe-Shetland Basin. These correlations are tied to type wells as outlined in **Section 3.5** below and are used to show the large scale seismic architecture of the Eocene succession. The entire basin is covered by the coarse grid of the regional correlations that trend both northwest - southeast (in a dip orientation) and northeast - southwest (in a strike orientation) (see **Figure 3.11**). The locations of the regional correlations have been defined by the position of many of the type wells (see **Section 3.5**) and the location of good quality 3-D and 2-D seismic data. As such, the regional correlations were defined in the UK Continental Shelf area, in particular on the Shetland margin and in the southern part of the basin, where type wells and good quality 2-D and 3-D seismic data is available (see **Figure 3.11**).

In total there are eleven regional cross sections that traverse the basin, five trending northwest – southeast and six trending northeast – southwest. Thirty five type wells lie on or very close to the regional correlations and a summary of these wells are shown in **Tables 3.1** and **3.2** (see **Section 3.5**).

The aim of the regional correlations is to provide a gross stratigraphic framework of the Eocene succession across the entire basin and to allow a description to be made of the major architectural features. The seismic units defined below are highlighted on the regional correlations and tied to the most appropriate type wells. In **Chapter 4** the regional correlations are accompanied by isochron maps for the


Regional correlation E
NW
SE
214/4-1
208/19-1
Two-Way Travel Time (s)
0
1.0
2.0
3.0
4.0
5.0
NWZ96-114
3-D T61T62-arbitary line
AUK93-ah-170-170a
AUK93-ah-101
AUK93ah-118
GA95-nsb-309
Km
0
25
Vertical Exxageration : x2
Figure 4.43
Base Balder (Base T50)
Top Balder (Top T50)
Top Eoc1
Top Eoc2
Top Eoc3
Top Eoc4
Eocene 1 seismic unit
Eocene 2 seismic unit
Eocene 3 seismic unit
Eocene 4 seismic unit
204/30A-1
Type Well



Regional correlation D

SE UK → ← Faroes NW

SW

Regional Correlation A	204/30A-1	204/22-1	204/23-1		
Regional Correlation B	205/20-1				
Regional Correlation C	206/8-3	206/1-2 (proj.)	206/2-1A	213-23-1	
Regional Correlation D	206/5-1	214/28-1	214/17-1 (proj.)		
Regional Correlation E		208/19-1		214/4-1	

NE

Table 3.1. Table summarising the wells which lie on or close to the dip orientated regional correlations (Trending NW-SE). The positions of the wells are schematic along the length of each regional correlation. Well 214/4-1 lies in over a mile of water in the central axis of the present day Faroe-Shetland Channel (see figure 3.9 for location)

SE (top) – NW (bottom); SW (left) – NE (right)

	SW → NE
Regional Correlation F	204/26A-2 204/30A-2 206/3A-1 208/27-2 207/1-1
Regional Correlation G	202/2-1 202/3-2 204/25A-1 204/30A-1 205/16-1 206/1-1A 205/10-2B 214/30-1 208/21-1 208/19-1
Regional Correlation H	204/19-2 204/23-1 205/9-1 214/27A-1 205/2-1A 214/28-1 214/29-1
Regional Correlation I	99/3 214/27-2 214/26-1 214/19-1
Regional Correlation J	214/4-1
Regional Correlation J	UK Faroes

Table 3.2. Table summarising the wells which lie on or close to the strike orientated regional correlations (Trending NE-SW). The positions of the wells are schematic along the length of each regional correlation. Note the number of wells found in the south (in Quadrants 204 and 205 where there is shallower shelfal areas), and also the lack of any released well data from the Faroese sector.

individual seismic units showing the present day depositional limit of the unit and its overall geometry. A more detailed description of each individual seismic unit will also follow in **Chapter 4** and is accompanied by a many seismic examples highlighting specific features within each unit and the internal seismic character (see **Sections 4.3.2 – 4.3.5**).

3.5 Well, Borehole and Biostratigraphic Data

3.5.1 Introduction

At present, one hundred and eighty six exploration and appraisal wells have been drilled in the Faroe-Shetland Basin since exploration began on the UK Atlantic margin in the 1972. If development wells were included in this statistic then the figure would be well over two hundred and fifty. Of these two hundred and fifty plus wells which have been drilled, various forms of data (e.g. paper or digital wireline logs, biostratigraphic reports, sedimentological reports) from one hundred and forty one have been documented in this study and have been summarised in **Appendix 3**. In addition to this comprehensive database, lithological and biostratigraphic data from two BGS boreholes was available for this study. It has been difficult to gain any information regarding wells which were drilled since the late 1990's as these most recent wells remain confidential for a period of five years.

3.5.2 Defining "Type Wells" For This Study

A major difficulty arising from the available well data is that the majority of wells were drilled to target prospects in the Palaeocene and the deeper Mesozoic succession. Therefore the Eocene - recent succession is not well (if at all) recorded in many of the composite and velocity logs of the wells. A critical analysis of all the available well data was carried out in order to distinguish which wells would be useful in this study. A set of criteria was established for choosing type wells (see **Figure 3.12**). The criteria are all interrelated and combine to give "type wells" which will be tie together with the seismic data on the regional seismic correlations. When defining the type wells, a key prerequisite is for the well to have both composite and velocity

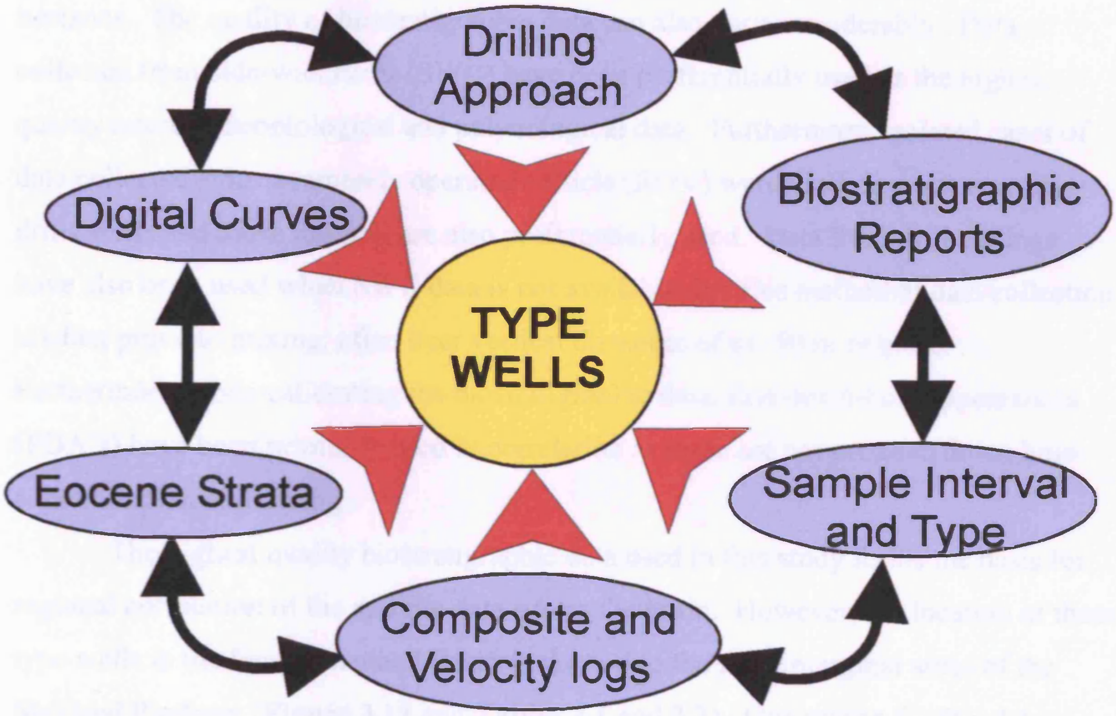


Figure 3.12. Schematic diagram showing the set of criteria that has been established for choosing the type wells in the study. Many wells penetrate the Eocene section, though due to many factors the data is not very reliable. The best wells are ones which specifically target the Eocene section with wireline log tools, or when the borehole is relatively narrow. Good biostratigraphic data and digital curves are also very important when choosing the type wells.

logs as a minimum. This is to allow for correlation from well-logs to the seismic data. In addition, type wells preferably need to have a biostratigraphic report to enable calibration of an age range for the seismic unit.

The quality of the biostratigraphic data is the main criterion for the choice of the type wells. Wells with good biostratigraphy can be tied to the velocity log and hence to the seismic data, allowing for a greater degree of age control when correlating seismic horizons. The quality of biostratigraphic data can also vary considerably. Data collected from side-wall cores (SWC) have been preferentially used as the highest quality micropalaeontological and palynological data. Furthermore, isolated cases of data collected from a remotely operated vehicle (ROV) were used on some recently drills wells and these samples are also preferentially used. Data from drill cuttings have also been used when SWC data is not available, but this method of data collection is often prone to mixing, often over vertical distances of ca. 50 m or more. Furthermore, when calibrating the biostratigraphic data, first down-hole appearances (FDA's) have been primarily used in correlation as these are not prone to down-hole contamination and caving.

The highest quality biostratigraphic data used in this study forms the basis for regional correlation of the seismic data across the basin. However, the location of these type wells in the Faroe-Shetland Basin is skewed to the more marginal areas of the Shetland Platform (**Figure 3.11** and **Tables 3.1** and **3.2**). One reason for this data skewing is the relatively recent resolution of the international boundary in the basin between the UK and the Faroes which was only settled in the last few years. To date only a few wells have been drilled in Faroese waters and all these wells remain confidential. The deep water found in the basin axes and the presence of igneous lavas and sills are other reasons for the paucity of wells in Faroese waters. The occurrence of igneous material obscures the deeper structure below resulting in a loss of seismic resolution due to seismic attenuation (e.g. Smallwood *et al.* 2001). Early wells focussed on targets on the northeast trending Rona Ridge and in the shallow West Shetland Basin as well as in the southern area of the Faroe-Shetland Basin in UKCS Quadrants 204 and 205 (see **Figure 3.11, Tables 3.1, 3.2** and **Section 2.2**). As technology advanced wells were drilled in the deeper Faroe-Shetland Basin where water depths can be as deep as 1.5 km.

Another key criterion in defining the type wells is the drilling approach which was taken on the rig. Almost all of the wells were drilled with a seventeen inch hole diameter through the Eocene succession (in an attempt to drill quickly through an interval that was not the target) and hence the quality of the data recorded onto the composite log is poor. Occasionally, an even larger hole casing is used for shallow sections often reaching thirty six inches. An ideal hole diameter of twelve and a quarter inch provides good quality lithological and wireline log information (**Figure 3.13**). Any well which used this twelve and a quarter hole diameter when drilling the Eocene (or higher) section has been classed as a type well because of the high quality of data it provides. Furthermore, wireline log tools which are inserted down the wide diameter holes after drilling tend to produce poor quality data as the tools are not in contact with the sidewall of the hole (e.g. Rider 1996). This leads to an attenuation of the log curves which can lead to misinterpretation (Rider 1996).

A final bank of data from two recent boreholes drilled by the BGS was also made available for this study. These shallow boreholes are located in the southern part of the Faroe-Shetland Basin, one in block 204/17 and the other in 6005/19 within Faroese waters (see **Figure 3.11**). They provide useful lithological data for the Eocene succession in this part of the basin as well as crude age dates, which are tied to 3-D and 2-D seismic data in the area. One of these boreholes has been included on the regional correlations discussed in **Chapter 4**.

In summary, this qualitative approach to defining type wells for use in this study and in particular with the regional correlations has eradicated many of the wells from the original database. This is because these wells didn't penetrate an Eocene succession, or had poor lithological and biostratigraphical data and thus was not required for use in this study. Furthermore, using this approach of selecting good quality well data, thirty five wells were chosen as type wells for use within this study. These wells have been chosen based on satisfying much of the criteria discussed above. They have then been integrated with the good quality seismic data (much of which is found in the south of the basin) in order to produce the regional correlations. Many of the type wells are located on or close to one or more of the correlations. Thus the intersection of type wells containing good lithostratigraphic and biostratigraphic data together with the regional seismic correlations provides the best possible technique for correlating chronostratigraphic seismic reflections from one part of the basin to the other.

However, not all of these thirty five wells have high quality lithostratigraphic and biostratigraphic data. Indeed some wells only provide a reference point and may

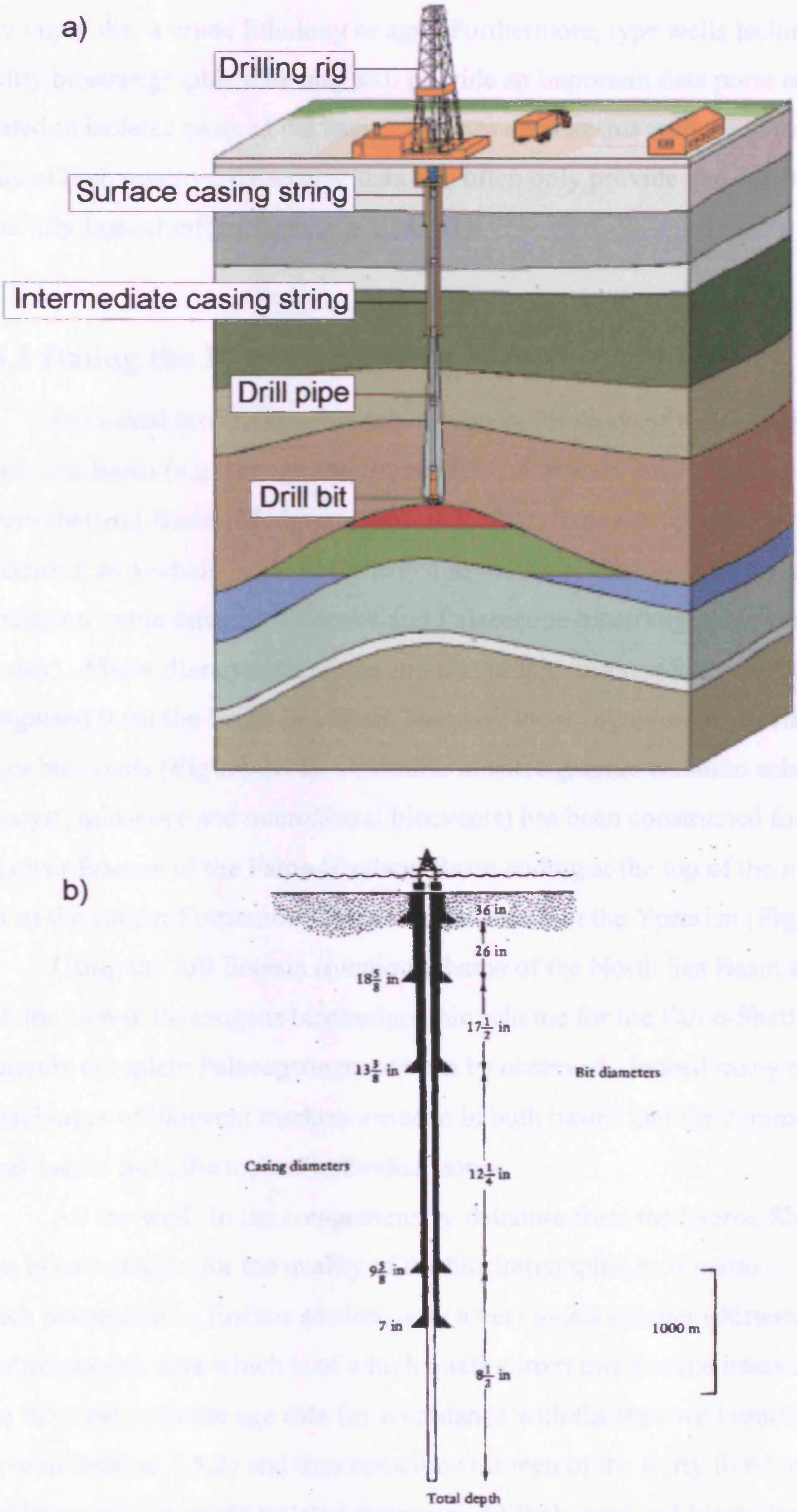


Figure 3.13. Schematic cartoons showing drilling and well operations. a). Shows location of drill rig and position of casing strings. Drill pipe and bit sit inside casing (Modified from image from Schlumberger website). b). Typical well showing decrease in casing diameter and drill bit diameter with depth. A drill bit diameter of $12^{1/4}$ inches is needed for good quality wireline logs (After Rider 1996) .

only imply that a crude lithology or age. Furthermore, type wells lacking in high quality biostratigraphic data may still provide an important data point especially when located in isolated parts of the basin. In cases such as this wells tend to be located in areas of high quality 3-D seismic data and often only provide one age date or some good lithological information (e.g. 214/4-1).

3.5.3 Dating the Eocene succession

Published biostratigraphic schemes exist for much of the Palaeogene of the North Sea Basin (e.g. Mudge and Bujak 1994; 1996a; 1996b) and more recently for the Faroe-Shetland Basin (Mudge and Bujak 2001). However, specifically the Eocene succession as a whole is not documented in the Faroe-Shetland Basin and thus correlation to the integrated Eocene and Palaeocene biostratigraphic zonation scheme is required. Major dinocyst zonations and planktonic foraminifera biozones have been recognised from the North Sea Basin based on the recognition of consistent tops of major bioevents (**Figure 3.14**). A similar biostratigraphic zonation scheme (using dinocyst, miospore and microfaunal bioevents) has been constructed for the Palaeocene to Lower Eocene of the Faroe-Shetland Basin ending at the top of the lithostratigraphic unit of the Balder Formation (T50 unit equivalent) in the Ypresian (**Figure 3.15**).

Using the full Eocene zonation scheme of the North Sea Basin and integrating it with the Lower Palaeogene biostratigraphic scheme for the Faroe-Shetland Basin, a relatively complete Palaeogene record can be observed. Indeed many of the assemblages of bioevent markers are seen in both basins and the common and abundant zonal fossils form the tops of individual zones.

All the wells in the comprehensive database from the Faeroe-Shetland Basin have been analysed for the quality of the biostratigraphic information. Of the wells which penetrated an Eocene section, only a very small number (thirteen) have biostratigraphic data which is of a high quality from this Eocene interval. These wells give the most accurate age data (in accordance with the type well criteria discussed above in **Section 3.5.2**) and thus constitute thirteen of the thirty five type wells. The remaining wells provide isolated instances of lithological and biostratigraphic information which can be used in this study though this data is not interpreted to be as accurate as the other thirteen wells (for reasons outlined above in **Section 3.5.2**).

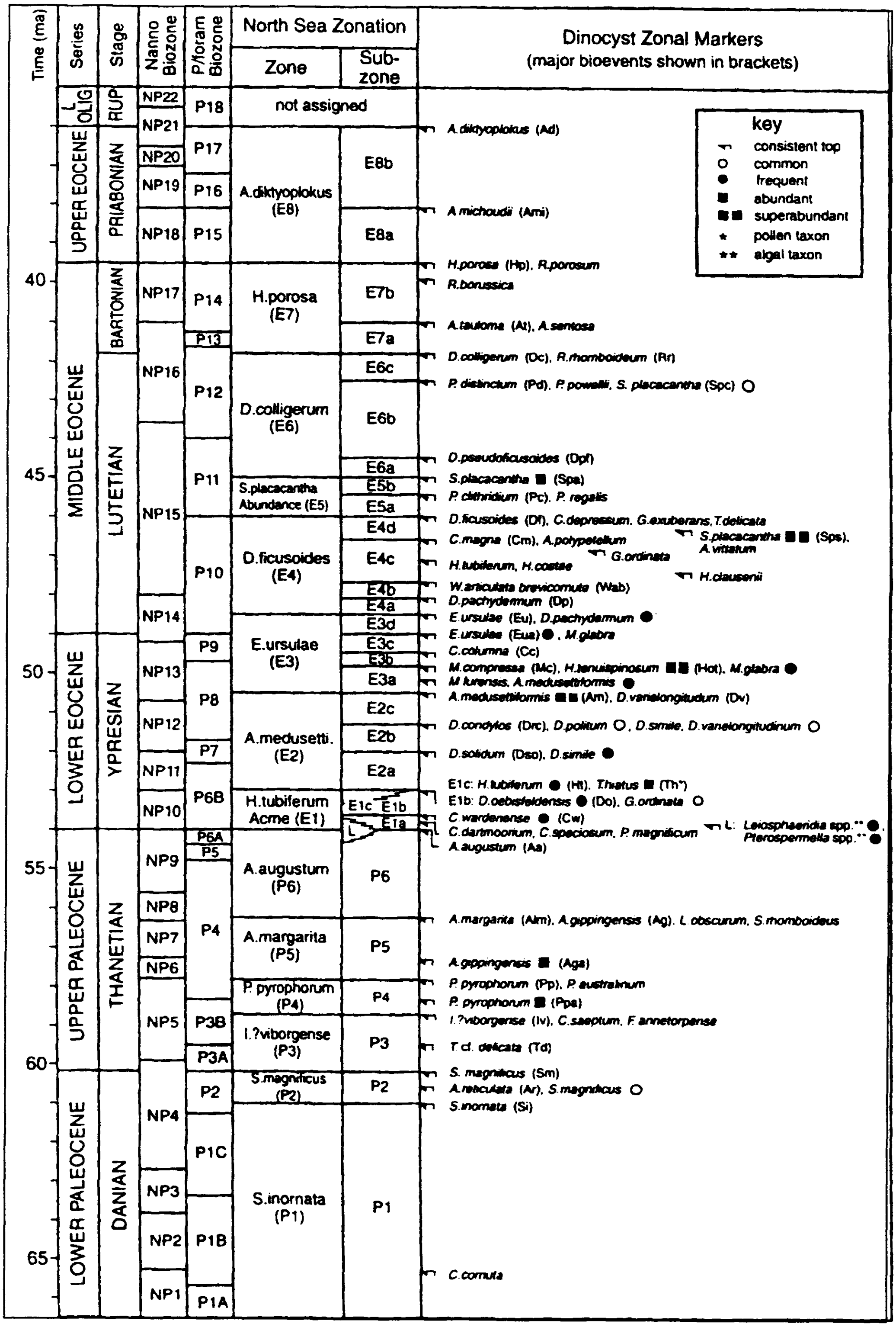


Figure 3.14. Integrated North Sea Palaeocene - Eocene dinocyst zonation scheme showing major zonal markers. Consistent tops (FDA's) of major bioevents mark the tops of individual zones and sub-zones. Absolute time-scale after Haq *et al.* (1987). After Mudge and Bujak (1996).

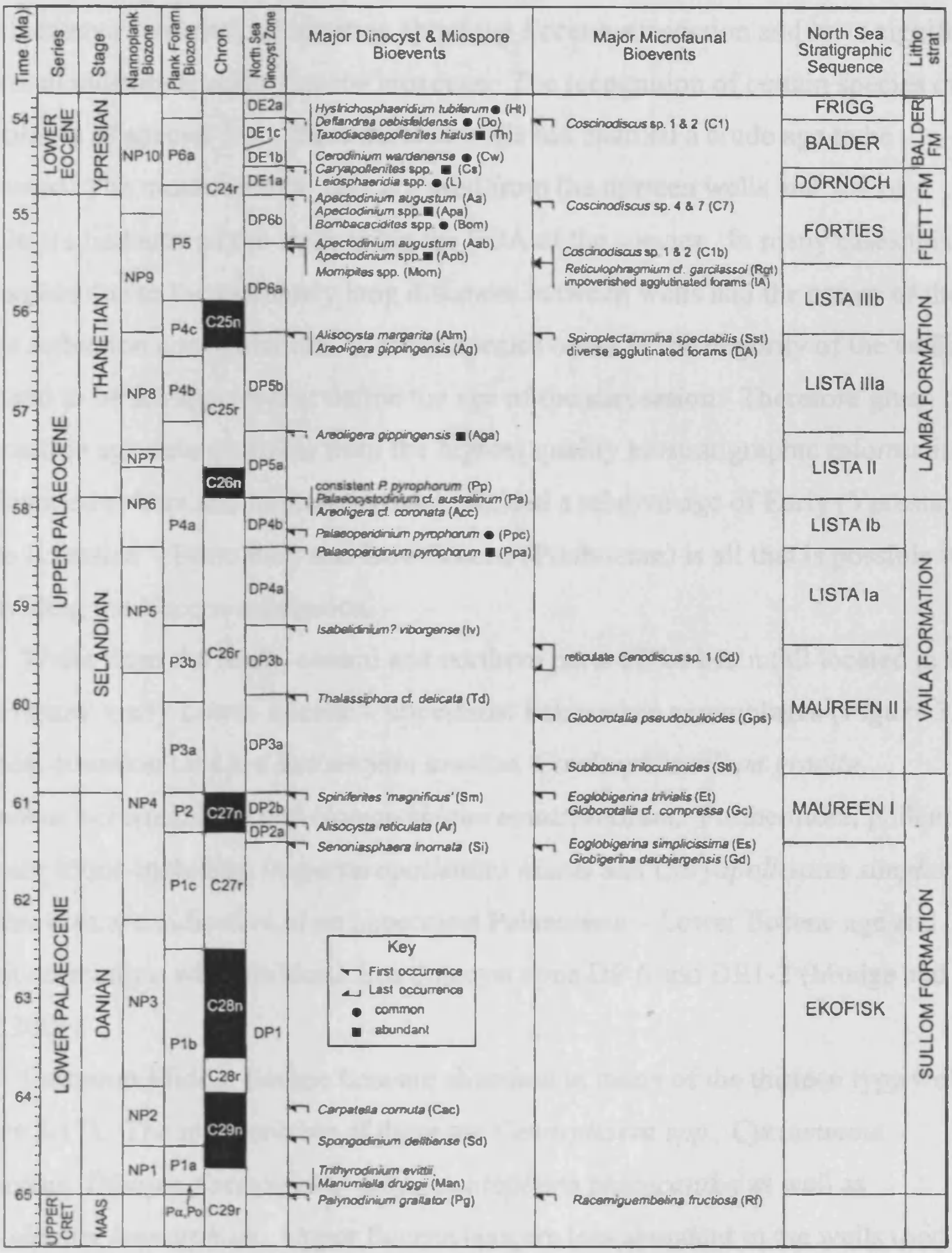


Figure 3.15. Stratigraphic chart showing the Palaeocene - Lower Eocene stratigraphy in the Faroe-Shetland Basin. The scheme is based on major dinocyst and miospore bioevents and major microfaunal bioevents. The FDA's and LDA's (first and last down-hole appearances) are indicated for the major bioevents. The stratigraphy is correlated to stratigraphic sequences of the North Sea (after Mudge and Bujak 1996b) and the Lithostratigraphy of the UK North West margin (after Knox *et al.* 1997). Absolute time-scale after Berggren *et al.* (1995). After Mudge and Bujak (2001).

These thirteen type wells form data-points on the regional correlations (discussed above in **Section 3.4.3**) and form the basis for the regional Eocene chronostratigraphic framework. These thirteen wells provide detailed biostratigraphic and palaeoenvironmental information about the Eocene succession and have significant taxa which indicate specific Eocene biozones. The recognition of certain species or assemblages of species from these thirteen wells has enabled a crude age to be ascertained. The most common taxa are used from the thirteen wells and where possible are tied around the wells using the FDA of the species. In many cases this is not possible due to the extremely long distances between wells and the nature of the seismic reflection configurations. Certain species occur in the majority of the wells and these tend to be the species that define the age of the succession. Therefore given the best possible age data available from the highest quality biostratigraphic information significant error bars and limitations exist. Indeed a relative age of Early (Ypresian), Middle (Lutetian – Bartonian) and Late Eocene (Priabonian) is all that is possible when sub-dividing the Eocene succession.

Wells from the south, central and northern parts of the basin (all located in UK waters) show many Lower Eocene – uppermost Palaeocene assemblages (**Figure 3.16**). The most common taxa are *Eatonicysta ursulae, Cordosphaeridium gracile, Deflandrea oebisteldensis* and *Homotryblium tenuispinosum*. Furthermore, pollen forms are found including *Inaperturopollenites hiatus* and *Caryapollenites simplex*. All these taxa are indicative of an uppermost Palaeocene – Lower Eocene age and suggest correlation with the North Sea dinocyst zone DP 6 and DE1-2 (Mudge and Bujak 2001).

Common Middle Eocene taxa are abundant in many of the thirteen type wells (**Figure 3.17**). The most notable of these are *Cenosphaera spp., Cyclammina amplectens, Diphyes fiscisoides and Systematophora placacantha* as well as *Apectodinium homorphum*. Upper Eocene taxa are less abundant in the wells used for this study though certain species are found to be consistent in many of the wells (**Figure 3.18**). These include *Cibicides westi, Heteraulacysta porosa* and *Phthanoperidinium spp*. Furthermore, various forms of the species *Adnatosphaeridium* and *Areosphaeridium* are seen in some of the northern wells.

A summary of the individual taxa seen in each of the thirteen wells and the depth of its occurrence (and associated two-way travel time (TWTT) in milliseconds) is

LOWER EOCENE (+ UPPERMOST PALAEOCENE)	205/9-1	205/26a-2	204/22-1	204/23-1	204/24-1	214/26-1	208/19-1	214/27-1	214/29-1	214/30-1	214/17-1	213/23-1	214/4-1
Coscinodiscus sp1. and sp2.	Y		Y	Y		Y	Y	Y		Y		Y	Y
Inaperturopollenites hiatus	Y		Y		Y		Y		Y	Y		Y	Y
Caryapollenites simplex	Y		Y		Y		Y		Y			Y	Y
Tillaepollenites microreticulatus	Y		Y	Y									
Eatonicysta ursulae	Y		Y	Y	Y	Y	Y		Y			Y	Y
Cibicoides gr. eocaenus	Y				Y							Y	
Apectodinium spp.			Y	Y								Y	
Deflandrea oebisteldensis			Y				Y	Y	Y	Y		Y	Y
Homotryblium pallidium			Y		Y								
Homotryblium tenuispinosum			Y		Y		Y		Y			Y	Y
Cordosphaeridium gracile	Y		Y		Y	Y	Y	Y	Y	Y		Y	Y
Spiniferites ramosus	Y				Y		Y	Y				Y	Y
Hystrichosphaeridium tubiferum			Y				Y	Y	Y	Y		Y	Y
Areoligera spp.	Y				Y		Y	Y	Y	Y		Y	Y
Subbotina linaperta		Y				Y		Y		Y	Y	Y	Y
Charlesdowniea columna					Y				Y	Y			Y
Dracododium spp.			Y	Y				Y	Y	Y		Y	Y
Dracododium politum			Y						Y				
A. coronata 'complex'	Y		Y		Y								
Spinozonocolpites echinatus			Y										
Globigerina sp.	Y												
Globorotalia sp.		Y											
Spiroplectammina aff. Spectabilis						Y				Y	Y		
Stellarim microtrias													Y
Fenestrella antiqua												Y	Y
Evolutinella spp.												Y	Y

Figure 3.16. Summary of the common Lower Eocene - uppermost Palaeocene taxa (both micropalaeontology and palynology) present in the thirteen type wells. Note the common occurrence of *Coscinodiscus sp1.* and *sp2.*, *Eatonicysta ursulae, Deflandrea oebisteldensis*, as well as the three forms of pollen (*Inaperturopollenites hiatus, Caryapollenites simplex* and *Tillaepollenites microreticulus*). These taxa are very common in both the North Sea and Faroe-Shetland Basin and form the tops of many North Sea dinocyst zones.

MIDDLE EOCENE	205/9-1	205/26a-2	204/22-1	204/23-1	204/24-1	214/26-1	208/19-1	214/27-1	214/29-1	214/30-1	214/17-1	213/23-1	214/4-1
Cenosphaera spp.	Y				Y	Y	Y	Y	Y		Y		
Cyclammina amplectens	Y		Y			Y		Y		Y	Y	Y	Y
Apectodinium homorphum	Y		Y	Y	Y								
Wetzelielia spp.			Y	Y	Y		Y	Y				Y	
Neoponides karsteni	Y				Y								
Turrilina brevispira			Y										
Membranophordium cf. aspinatum			Y		Y								
Diphyes fiscisoides					Y	Y					Y	Y	Y
Systematophora placacantha					Y	Y					Y	Y	Y
Stellarima microtrias											Y		
Subbotina eocaena		Y		Y		Y					Y		
Cibicides spp.	Y										Y		Y
Rhabdammina robusta	Y										Y		
Loxoconcha curryi											Y		
Deflandrea phosphoritica					Y							Y	
Pseudohastigerina micra	Y												
Truncororotaloides cf. Rohri	Y												
Diphyes colligerum													Y
Phthanoperidinium spp.												Y	Y
Areosphaeridum arcuatum												Y	Y

Figure 3.17. Summary of common Middle Eocene taxa (both micropalaeontology and palynology) present in the thirteen type wells. There is a certain degree of diversity in the assemblages seen in the samples however we see more occurrences in the Lower Eocene than in this Middle Eocene succession. Certain taxa including *Cyclammina amplectens, Cenosphaera spp.* and *Diphyes fiscisoides* are most common in the Faroe-Shetland Basin.

	205/9-1	205/26a-2	204/22-1	204/23-1	204/24-1	214/26-1	208/19-1	214/27-1	214/29-1	214/30-1	214/17-1	213/23-1	214/4-1
LATE EOCENE													
Cibicides westi	Y			Y									
Bolivina cookei	Y												
Uvigerina eocaena	Y												
Corrudinium inconpositum						Y							
Trilites spp.						Y							
Heteraulacysta porosa												Y	Y
Phthanoperidinium spp.													Y
Adnatosphaeridium spp.													Y
Areosphaeridium spp.				Y	Y							Y	Y

Figure 3.18. Summary of common Upper Eocene taxa (both micropalaeontology and palynology) present in the thirteen type wells. Note that there is a paucity of common Upper Eocene taxa found in the samples with data from only five wells found.

summarised in individual appendices. The uppermost Palaeocene – Lower Eocene, Middle Eocene and Upper Eocene is found in **Appendices 4, 5** and **6** respectively.

3.5.4 Biostratigraphic Mis-calibration

Great care has to be taken when dealing with this biostratigraphic information as it can often turn out to be spurious and misleading. A simple example of this problem can be seen on one 3-D seismic traverse which ties to three wells (**Figure 3.19**). The depth of biostratigraphic markers have been converted into two-way travel time (TWTT) using Time-Depth curves derived from check-shot data and positioned on the wells. The top T50 (Balder Tuff) event is correlatable between the three wells and ties to the seismic data, whereas the top Middle Eocene marker does not correlate between all three wells and can in no way tie to the seismic data. There is a contradiction in the biostratigraphic analysis of the top Middle Eocene marker because the ranges of the micropalaeontological and microfaunal age data are too long. Thus the top of the Middle Eocene represents a difficult event to pick accurately in this basin and is not a good event marker like the top T50 (Balder Tuff) event. Interestingly, the distance between these three wells is small (a few km) with the wells being located entirely within one 3-D seismic survey. Greater errors occur when trying to correlate intra-Eocene biostratigraphic markers around the entirety of the basin. In this particular instance it is difficult therefore to be sure where the top Middle Eocene actually is, and hence which, if any, of the biostratigraphic markers is the correct one. In circumstances such as these, the biostratigraphic data is not used unless one such data point is of much higher quality.

Therefore in this instance, the most pragmatic approach to take to determine the most accurate top Middle Eocene event is to assess which well has the best biostratigraphic information. This is based on the accuracy of the key bio-events and the type of sample taken (see **Section 3.5.2** above on sidewall cores and FDA's). The wells are then ranked according to this qualitative assessment. Higher ranked wells are given greater weight in the calibration of specific seismic marker reflections. This is the most accurate way of selecting the best biostratigraphic information though it is still carries a degree of error. The seismic interpretation that utilises the biostratigraphic information, and therefore the stratigraphic significance of that interpretation can be

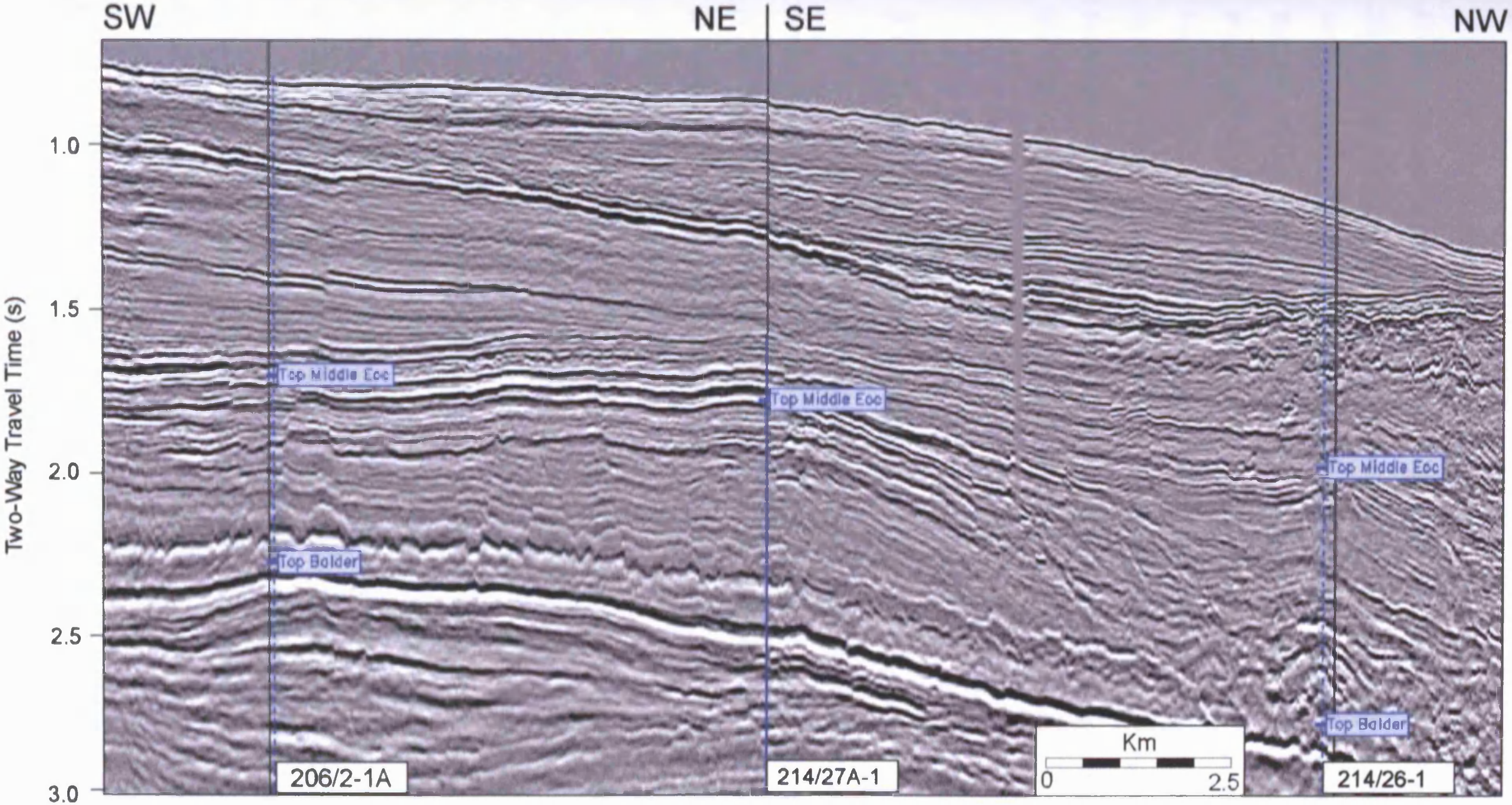


Figure 3.19. SW - NE 3-D seismic line through three wells in the south of Quadrant 214 running south into Quadrant 206. This seismic line highlights the problems associated with biostratigraphic information in the Eocene succession of the Faroe-Shetland Basin. The three wells each have individual picks taken from the composite log and biostratigraphic reports to indicate the top of the Middle Eocene pick. However when compared to the actual seismic data it is clear that the seismic does not match with the well picks. Wells 206/2-1A and 214/27A-1 have quite similar tops which may be crudely correlated, though the well to the north (214/26-1) shows a significant time shift. Therefore in these circumstances the well data must be ignored for this particular pick and the correlation must be based on the seismic alone. It is possible to see correlation of the top Balder pick across the wells, though this pick is generally consistent across much of NW Europe.

vastly different depending on which well provides the highest quality data, due to the lateral variation in the stratal reflection configurations (**Figure 3.19**).

When the biostratigraphic data is assessed and ranked to be of poor quality, a different approach needs to be taken. This is very common in the Eocene of the Faroe-Shetland Basin for reasons outlined above and there is a paucity of good biostratigraphic control in the form of FDA's and sidewall cores. When this is the case, the methodology of picking the most regionally correlatable reflections is the most pragmatic and standard approach which can be used. By doing this, and picking the highest amplitude, laterally continuous reflection which has the largest distribution and correlatability, a near top Middle Eocene reflection can be produced. This reflection is purely picked because of its seismic character and represents a close approximation to true biostratigraphic top of the Middle Eocene.

3.6 Conclusions and Summary

This chapter began with a critical review of modern seismic-stratigraphic methodology and outlined the key issues fundamental to regional basin analysis. The problems associated with picking depositional sequences *sensu* Exxon are outlined with the inherent misuse of the term unconformity; a fundamental building block of modern and ancient stratigraphy.

Recognising unconformities (or sequence boundaries) on a basin-wide scale is inferred to be implausible and it is suggested that they are relatively local in extent and that the associated correlative conformity is extremely difficult (if not impossible) to correlate around an entire basin. Indeed, in basins undergoing a post-rift thermal subsidence phase (such as the Faroe-Shetland Basin), thus with an inferred relative sea-level rise during this episode of subsidence, the widespread development of unconformities on the basin margins is expected to be highly unlikely.

The seismic-stratigraphic approaches used in this study are then introduced. Two techniques are used depending on the aim of the particular study. A regional basin-wide study of the entire Eocene succession is approached in **Chapter 4**, the aim of which is to chronostratigraphically sub-divide the strata into seismic units. A pragmatic seismic based approach is taken by mapping the most regionally correlatable seismic reflection configurations around the whole of the basin, and using the available biostratigraphic data to constrain a crude age. Five regional seismic markers are defined based on their seismic character, continuity and regional correlatability and these are the bounding surfaces of regional seismic units. These surfaces are defined in the southern half of the basin because the Eocene succession is closer to the sea-bed and thus better imaged. Furthermore, the seismic character is good here, and the surfaces show a high amplitude and continuous character. Thus all surfaces are defined here and traced to the north of the basin where the data is generally poorer in quality. A technique known as phantoming is used in the northern area as the seismic reflections are incoherent and low amplitude.

The biostratigraphic data is selected based on a qualitative approach and is assessed as to its quality by a predetermined set of criteria. Only thirteen wells from a database of over two hundred have good biostratigraphic data for the Eocene succession and these constitute some of the type wells used in this study. With these wells, a crude age range can be assigned to the seismic units described in **Chapter 4**. Furthermore, data from two shallow boreholes provides good age and lithologic information for the southern part of the basin which can be calibrated to high quality 3-D seismic data.

A more classical approach of picking sequence boundaries is used in **Chapter 5** with high quality 3-D seismic data located on the margin in the South Judd Basin allowing for the recognition of localised unconformities in the Lower Eocene.

4. Chapter Four: Regional Stratigraphic Sub-division of the Eocene Succession and Implications for Basin Evolution

4.1 Introduction

The following chapter proposes a new stratigraphic sub-division for the Faroe-Shetland Basin from the latest Palaeocene to the end of the Eocene. This chapter also discusses the controls upon basin-wide stratigraphic architecture, details the regional distribution of the latest Palaeocene and Eocene stratigraphy and documents the palaeogeographic evolution of the basin during and after a period voluminous magmatic activity.

Prior to the analysis and discussion of the new stratigraphic scheme it is required to highlight the nature of the receiving basin prior to the onset of Early Eocene deposition. Therefore the latest Palaeocene – earliest Eocene succession will be introduced. This unit has been termed the T50 unit and is taken from the T-scheme initially described by Jones and Milton (1994) in the North Sea Basin and by Ebdon *et al.* (1995) in the Faroe-Shetland Basin. This T50 unit is broadly equivalent to the lithostratigraphic unit of the Balder Formation and was deposited at the very end of the Palaeocene and into the earliest Eocene.

The chapter goes onto describe the new stratigraphic sub-division which defines four seismic stratigraphic units which have been documented within the Eocene succession (named herein Eocene 1 – 4 seismic units). These seismic-stratigraphic units are sequentially discussed in detail within this chapter. The distribution, lithology (where possible), seismic character and internal geometry of these units is discussed on a basin-wide scale and palaeogeographic reconstructions from seismic and well data have enabled the geological evolution of the Eocene to be analysed. The gross basin architecture and the basin-wide seismic character of the seismic units are shown in eleven regional correlations. Continued reference to these regional correlations is made throughout **Sections 4.2** and **4.3** and for ease, it is recommended that then reader uses the enclosures for use with the sections. Five northwest – southeast correlations show the Eocene succession from the Shetland margin to the Faroe-Platform (**regional correlations A – E** and have previously been displayed in **Chapter 3** and will not be repeated here). Six northeast – southwest correlations run along the strike of the basin from the Wyville-Thompson Ridge in the southwest, to Brendan's Dome in the northeast (**regional correlations F – K**). These six strike-orientated regional correlations are shown in the order they are referenced in this chapter and the full series of eleven regional correlations occur as a set of enclosures at the back of the thesis (**Enclosures A – K**). The basemap showing the location of the regional correlations is shown in **Figure 3.11** and **Enclosure L**.

4.2 Stratigraphy of the Upper Palaeocene – Lower Eocene

4.2.1 Introduction: the T50 unit

There was great variability in the depositional environment within the Faroe-Shetland Basin at the end of the Palaeocene to the beginning of the Eocene. A short summary of the geological setting and depositional style during this time interval will be discussed here. Particular emphasis will be focussed on the southern part of the basin in Quadrants 204 and 205 where the T50 unit is at its thickest and also at a shallow depth and therefore well imaged. This section will draw on various parts of the literature which have studied the T50 unit in detail and bring together some conclusions on this well studied basal section.

4.2.2 What is the T50 unit?

The British Petroleum Oil Company pioneered the T-scheme (abbreviation for Tertiary) which is a stratigraphic framework for sub-dividing the Tertiary interval into a series of distinguishable sand-rich units separated by more shaley intervals which in places acted as seals (see **Section 2.2.3.4, Figure 2.9**, Jones and Milton 1994 and Ebdon *et al.* 1995). The T-scheme developed through the 1990's from the broadly lithostratigraphic scheme of T10, (Ekofisk), T20 (Maureen), T30 (Lista), T40 (Forties), T45 (Sele) and T50 (Balder). The T-scheme is a hybrid scheme, which has similarities with transgressive – regressive (T - R) cycles (Embry 1993). Both maximum flooding surfaces and sequence boundaries (e.g. seen in the T50 unit) bound some of the T units.

The top of the T50 unit is defined as the top Balder Tuff event (Ebdon *et al.* 1995). The use of the term Balder is not widely used in sequence stratigraphy, indeed it is wrong to do so as it is a lithostratigraphic term. However, the top Balder Tuff event is defined and characterised by a regional ash layer (the Balder Tuff) over much of Northwest Europe and is therefore widely accepted to represent an geologically instantaneous event which is isochronous (see **Section 2.2.3.4**). This is therefore a special case, and a rare occasion when a lithostratigraphic boundary is isochronous over large distances. Ebdon *et al.* (1995) defined the base of the T50 unit as a regional unconformity that developed during the previous T40 unit. Therefore it can be seen that the T50 unit has an unconformity (sequence boundary) at its base and a regionally correlatable tuff horizon at its top, and is thus a hybrid scheme.

The T50 unit was interpreted and described using system tract terms and is interpreted as a transgressive systems tract (Ebdon *et al.* 1995) after a period of lowstand during T40 times (of the Late Palaeocene) when the basal unconformity developed. This unconformity down-cuts into the underlying Flett and Lamba Formations of the T40 and T30 units (see **Figure 2.9**) and becomes increasingly unconformable to the south and east (Smallwood and Gill 2002).

The approach of linking sequence stratigraphy and lithostratigraphy to stratigraphic sub-division is not common and should not be followed. It can cause confusion and lead to significant misleading interpretations of the stratigraphic development of a basin. The T50 unit described below simply represents a period of time at the very end of the Palaeocene which continued into the earliest Eocene and is broadly equivalent to the deposition of the Balder Formation.

4.2.2 Dating the T50 unit

The T50 unit is believed to be entirely Eocene in age according to many authors (e.g. Knox and Holloway 1992, Berggren *et al.* 1995). The basal unconformity of the T50 unit down-cuts into the underlying Flett and Lamba Formations of the T40 and T30 units (see **Sections 2.2.3.4** and **4.4** and **Figures 2.9, 2.13** and the later **Figure 4.58**) and becomes increasingly unconformable to the south and east (Smallwood and Gill 2002). As stated above the upper boundary of the T50 unit is the Balder Tuff; which is seen as a prominent regionally correlatable marker over the entire North Sea and West of Shetland area (see **Section 2.2.3.4**). The Balder Tuff is an ash deposit believed to have formed at the onset of sea-floor spreading at the incipient rift zone between

Eurasia and Greenland when phreatic eruptions were prevalent (e.g. Knox and Morton 1988).

There has been some recent debate about the age of the latest Palaeocene and earliest Eocene succession based around new biostratigraphic dates. A date of 54.98 Ma is believed as the top of the Palaeocene (Berggren *et al.* 1995) though this has recently been questioned by new U/Pb and Ar/Ar isotopic dates of Jolley *et al.* (2002) who suggest dates between 57.5 and 60.54 from lavas immediately overlying the Late Palaeocene thermal maximum. Hence an age disparity of up to 5 Ma is seen for the latest Palaeocene succession.

These recent findings have thrown doubt over both the palynological and petrological dating of the Palaeocene – Eocene interval. The new dates of Jolley *et al.* (2002) essentially question the length of time between the units which underlie the T50 unit (T40 - T45 units) and the T50 unit itself. If the dates are to be believed, the Palaeocene interval from its base (at 65 ± 0.2 Ma) to the T40 unit is possibly much shorter than originally thought. This has implications for increased sedimentation rates offshore and oceanic spreading rates throughout this interval (Jolley *et al.* 2002, Clarke 2002 unpublished PhD. thesis).

4.2.3 Distribution of the T50 unit

The geographical extent of the T50 unit is still uncertain in many parts of the Faroe-Shetland Basin. This is primarily due seismic attenuation below the large area of igneous lavas and sills (e.g. see **Figure 2.18**) that lie above and interbedded with latest Palaeocene - earliest Eocene succession. Quadrant 204 is an area of good data quality that is not significantly obscured by lavas at the same stratigraphic level. Even with the presence of the igneous rocks being prevalent throughout the basin, the top Balder Tuff (the top of the T50 unit) is recognisable over much of Northwest Europe. However, the intra and basal T50 surfaces are much more difficult to image because of the problems outlined above. Nonetheless in the southern area of the basin (in the area of Quadrant 6004, 204 and 205, where thin or spatially restricted lava is present) the top and base as well as the internal characteristics of the T50 unit can be imaged clearly on high quality 3-D seismic data. Smallwood and Gill (2002) produced an isochore map (thickness in metres) of the T50 (Balder) unit (see **Figure 2.13**) which highlights a northward draining valley network in the basal unconformity surface (base T50 (Balder) unconformity) in the very south of Quadrant 204. The map shows significant lateral changes in thickness from 20 m on the flanks of the valleys to 350 m in the axis of the valleys (**Figure 2.13**). Additionally, Smallwood and Gill (2002) produced a regional thickness map (isochore) which shows a broadly concentric depocentre centred on Quadrants 6005 and 204 (**Figure 4.1**). The thickened area is confined to the southern part of the basin only and thins to zero towards the northwest and northeast. The regional isochore map (**Figure 4.1**) shows the areas of the thickest preserved T50 succession which is approximately 300 m thick (blues and greens) and can be seen to thin (to less than 50 m) in all directions to oranges and reds. The thickest area of the isochore (in Quadrants 6005 and 204) has a curved northern edge depicted as the change from green colours (thick areas greater than 225 m) to yellow colours (thin areas less than 225 m). This arcuate northern edge trends in a northwest - southeast direction with the arc closed to the southwest (convex to the northeast) and may represent the limit of major clastic sediment input to the T50 unit delta (for a more detail discussion see **Section 4.2.6**). The thinner distal zone (shown as yellow colours) could represent an area in front of the delta on the outer shelf or possibly slope receiving a thinner mud dominated condensed section (see **Sections 4.2.4** and **4.2.6**).

The occurrence of the T50 unit therefore seems isolated to this southern area. The apparent absence (or condensed and thin equivalent) of the T50 unit in the central and northern parts of the basin may be due to sediment supply, a deeper water setting creating a natural depositional thinning towards the northwest and northeast into a condensed section or also due to erosion. Additionally, it could be a seismic imaging problem due to the presence of significantly thick lavas close to the of the top T50 unit. When tracing the T50 unit into the northern area there is great difficulty interpreting any basal unconformity as well as any intra-T50 reflections with any great confidence due to the poor quality of the sub-basaltic seismic data. On some regional 2-D seismic lines however, it is possible to see slight downlapping reflection terminations of the intra-T50 reflections that occur at a shallow angle towards the northeast onto the basal unconformity. This occurs in the location where the T50 unit thins distally in the area of northern Quadrant 204 and the southern part of Quadrant 213 (**Figure 4.2**- and **regional correlation J**). This would favour a depositional thinning of the T50 unit towards the northeast into possibly deeper water located in the more central and northern part of the basin. However, slightly further northeast of these downlapping reflections the intra-T50 unit (top T50) reflections appear to be truncated at a shallow

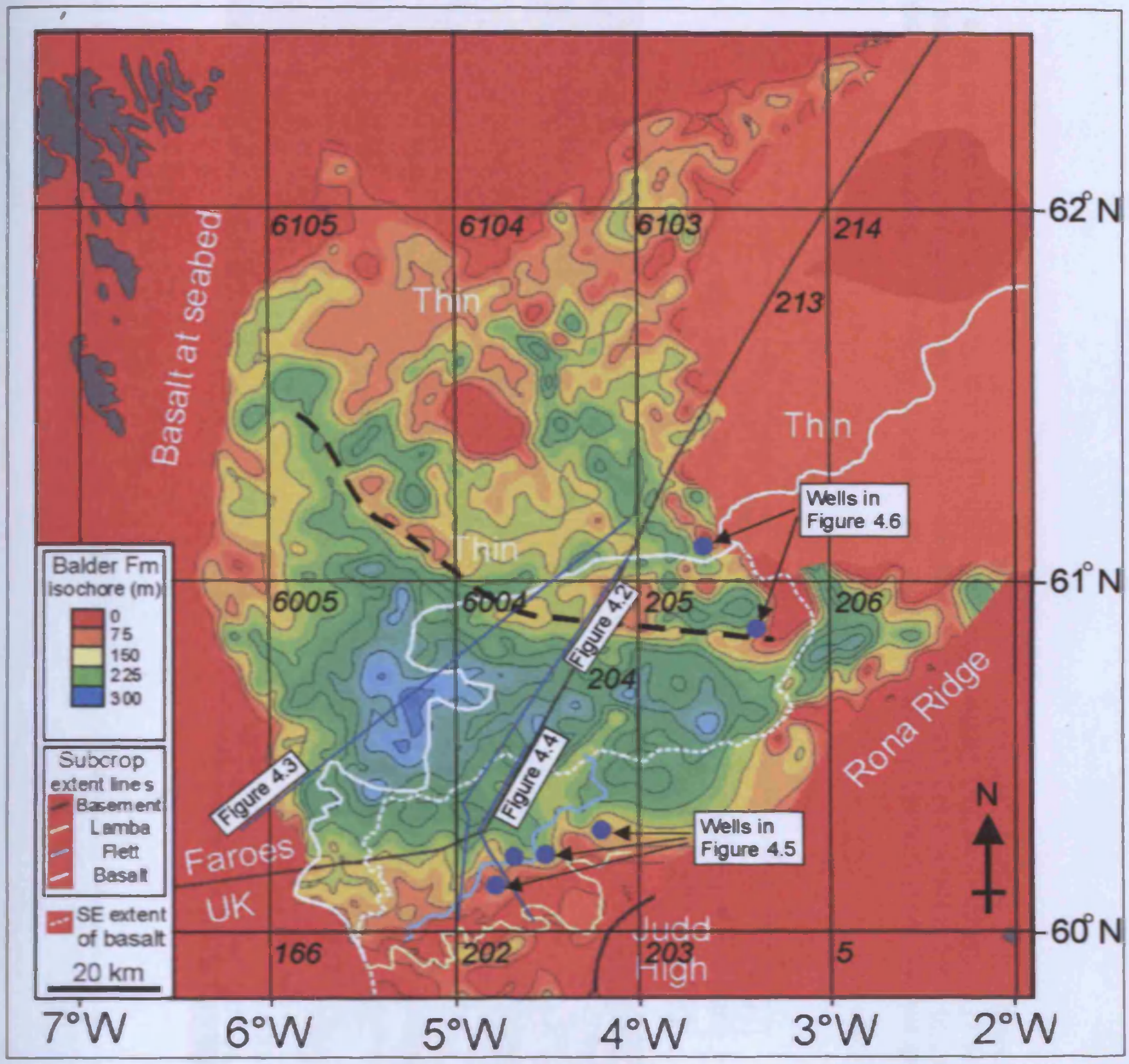


Figure 4.1. Isochore (thickness) map of the T50 unit (Balder Formation equivalent) showing gross distribution (area of preserved thickness) centred in the south of the basin over Quadrants 6005, 6004 and 204. The map shows the shape of the basin during the earliest Eocene. This southern depocentre thins to zero on its western fringe where the T50 unit subcrops the seabed. In the northeastern part of the basin the T50 unit thins significantly and may represent a condensed section in a low energy marine realm. The southeastern margin of the basin remained emergent at this time with the Rona Ridge and Judd High forming an amalgamated structural high. Progressively older Palaeocene formations subcrop the base of the T50 unit to the southeast. This basal surface thus represents a semi-regional unconformity (the base Balder unconformity of Smallwood and Gill 2002). The black dashed line delineates the thick area of the T50 unit (greater than 225 m) in the south from the thinner (less than 225 m) northern area and may represent the change from the shelf to the slope and basin axis. The positions of three seismic lines in Figures, 4.2, 4.3 and 4.4, and wells in Figures 4.5 and 4.6 are shown. (Modified after Smallwood and Gill 2002).

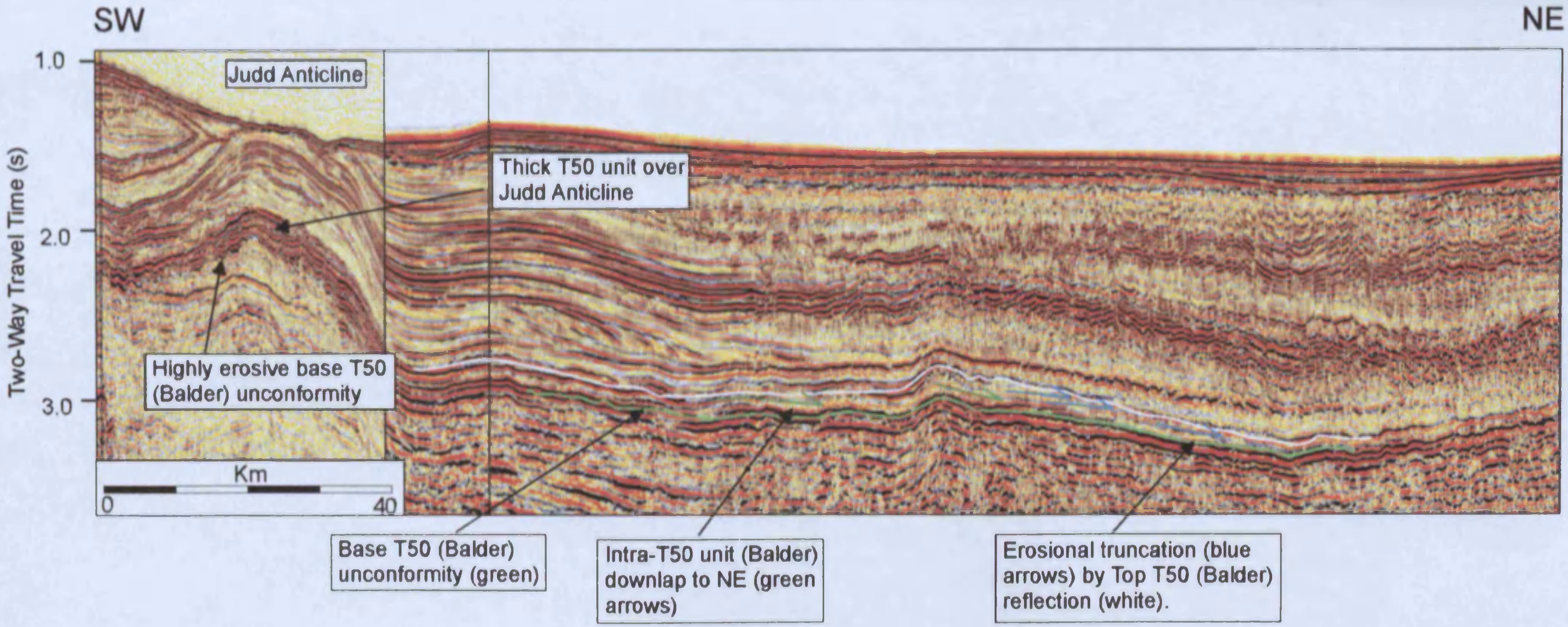
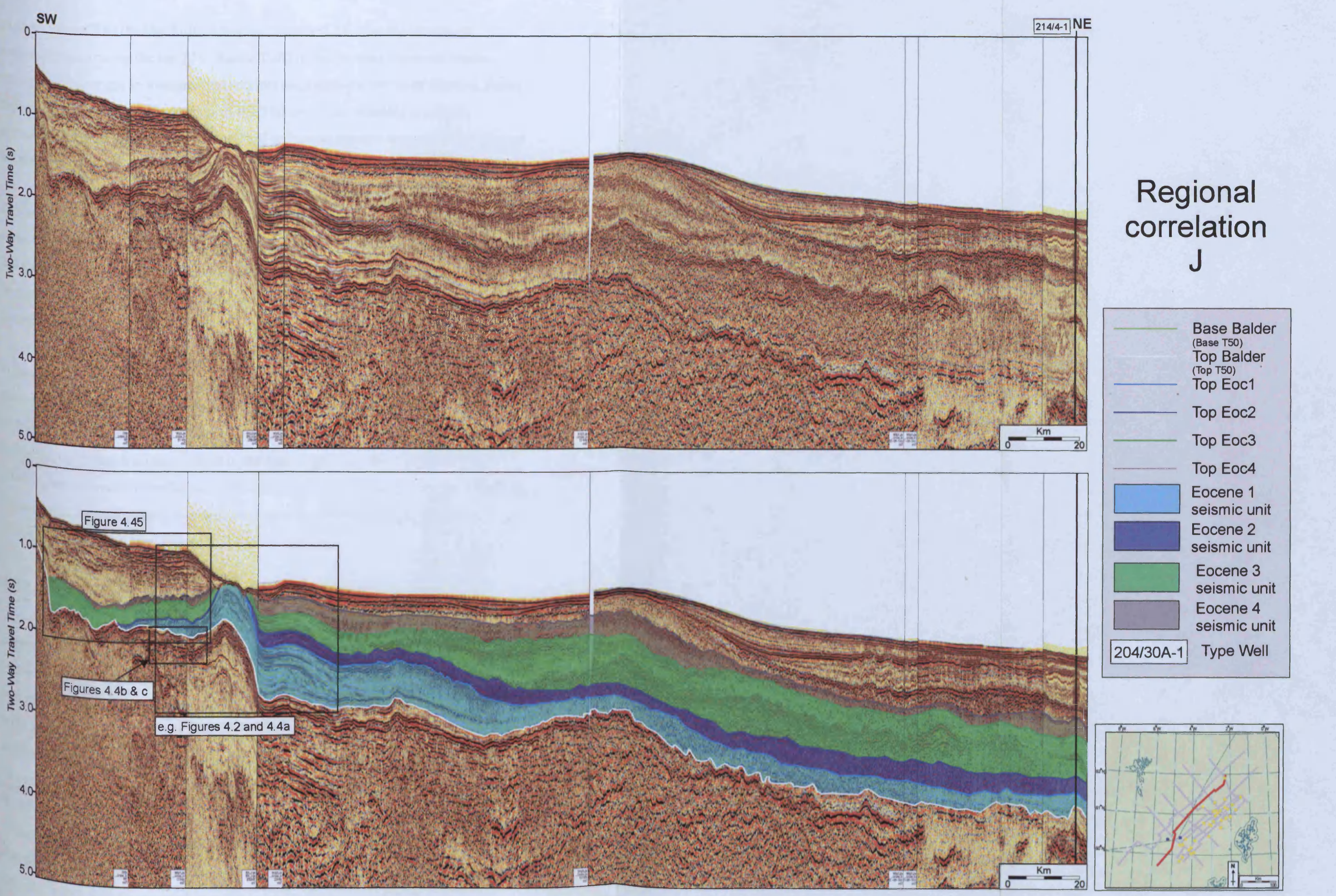


Figure 4.2. Southwest northeast trending 2-D seismic panel showing the thinning of the T50 unit to the northeast. Towards the southwest, in the area of the Judd Anticline, a thickened T50 unit is seen and the base of the unit (green horizon) shows a significant erosion surface (the base T50 (Balder) unconformity). The top (T50) Balder reflection (white horizon) thins and becomes more concordant with the base T50 (Balder) unconformity towards the northwest. For position of seismic line see Figure 4.1.


SW
NE
214/4-1
Two-Way Travel Time (s)
Figure 4.45
Figures 4.4b & c
e.g. Figures 4.2 and 4.4a
Km
0
20
Regional correlation J
Base Balder (Base T50)
Top Balder (Top T50)
Top Eoc1
Top Eoc2
Top Eoc3
Top Eoc4
Eocene 1 seismic unit
Eocene 2 seismic unit
Eocene 3 seismic unit
Eocene 4 seismic unit
204/30A-1 Type Well

angle by the top T50 (Balder Tuff) reflection (**Figure 4.2**). The appearance of erosional truncation by the top T50 (Balder Tuff) reflection may represent marine erosion or erosion due to transgression (ravinement) during a period of flooding during deposition of the youngest sediments of the T50 unit. This possible erosion is anomalous because the top T50 (Balder Tuff) reflection appears very conformable over the majority of the basin.

Towards the west and northwest of the circular depositional thick however, the T50 unit and basalt lavas below are eroded and are seen to subcrop at the sea-bed close to the Faroe Platform. This northwestern fringe of the T50 unit thins from greater than 200 ms to zero over a short distance (less than 20 km) towards the northwest and could be partly due to later uplift and erosion on the Munkagrunnur Ridge and the Faroe platform (**Figures 4.1** and **4.3**).

Progressively older stratigraphy can be seen to subcrop the basal unconformity to the southeast towards the Judd High (Smallwood and Gill 2002). The subcrop positions of the underlying units are highlighted in **Figure 4.1** and are also shown on seismic data in **Figure 4.4**. The unconformity at the base T50 unit indicates that significant uplift and erosion occurred in the Judd High area. Smallwood and Gill (2002) interpret eroded clinoforms of the T45 unit (or Flett Formation equivalent) as an indicator that a deltaic setting was prevalent prior to the development of their base T50 (Balder) unconformity (see **Figure 4.4**).

4.2.4 Lithological Description of T50 unit

No released wells are presently located in Faroese waters and hence no information on the lithology of the T50 unit is known from this particular area. However, on the eastern side of the UK/Faroe boundary there is limited well data enabling the lithology of the T50 unit succession to be determined. Additionally, some wells provide biostratigraphic data giving an indication as to the depositional environment of the T50 unit. In the southern part of Quadrant 204, the unit comprises of a series of interbedded lithologies. Sandstones, tuffs, mudstones and coals are all found in the T50 unit to a varying degree in wells located in the south of Quadrant 204 (**Figure 4.5**), thus a marginal marine to deltaic depositional environment is interpreted. The well data confirms that there is great thickness variation and lithological change laterally, which is to be expected in coastal plain and delta top settings. The

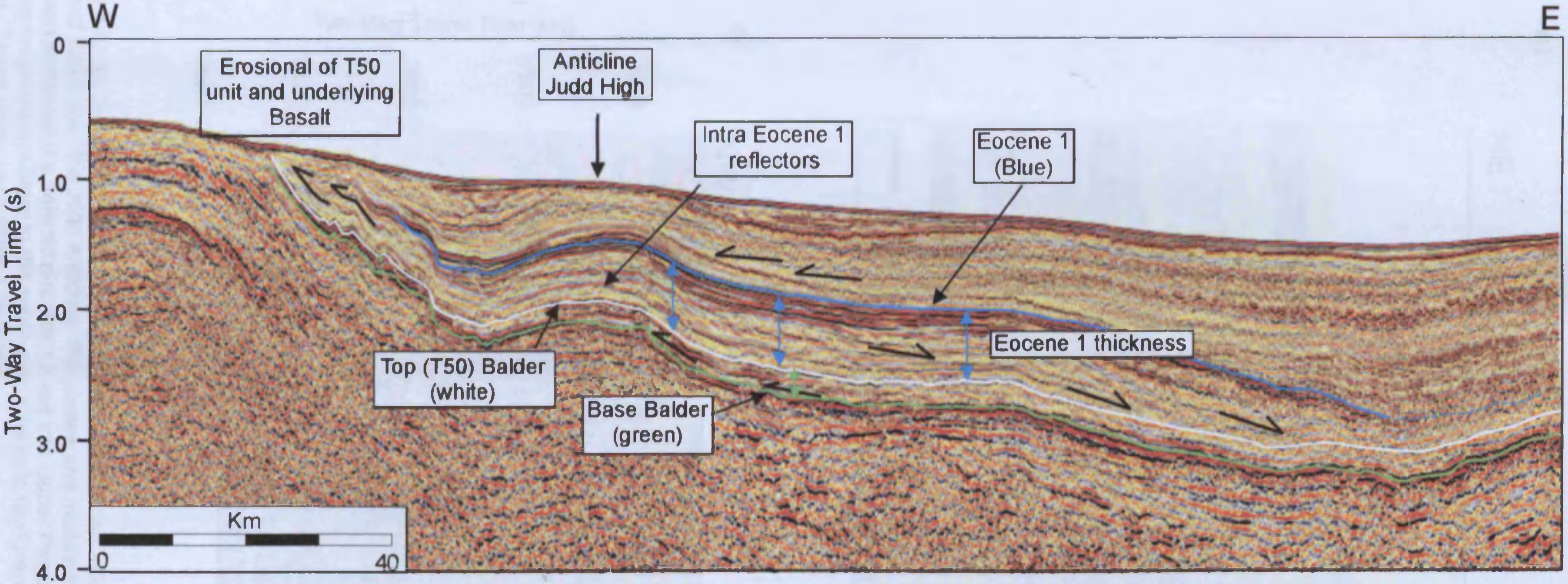


Figure 4.3. Broadly west - east 2-D seismic panel showing the high to moderate amplitude, parallel and continuous reflection configurations of the Eocene 1 unit that overlies the T50 unit. The T50 unit and the overlying Eocene 1 unit thins to zero towards the west and northwest onto the Munkagrunnur Ridge, west of the Judd Anticline. On the crest of the ridge, the entire Eocene and much of the Palaeocene section is eroded. The top Eoc1 reflection is seen close to the top of the highly reflective package which can be seen to thin and downlap onto the top T50 (Balder Tuff) reflection to the west and east. For position of seismic line see Figure 4.1.

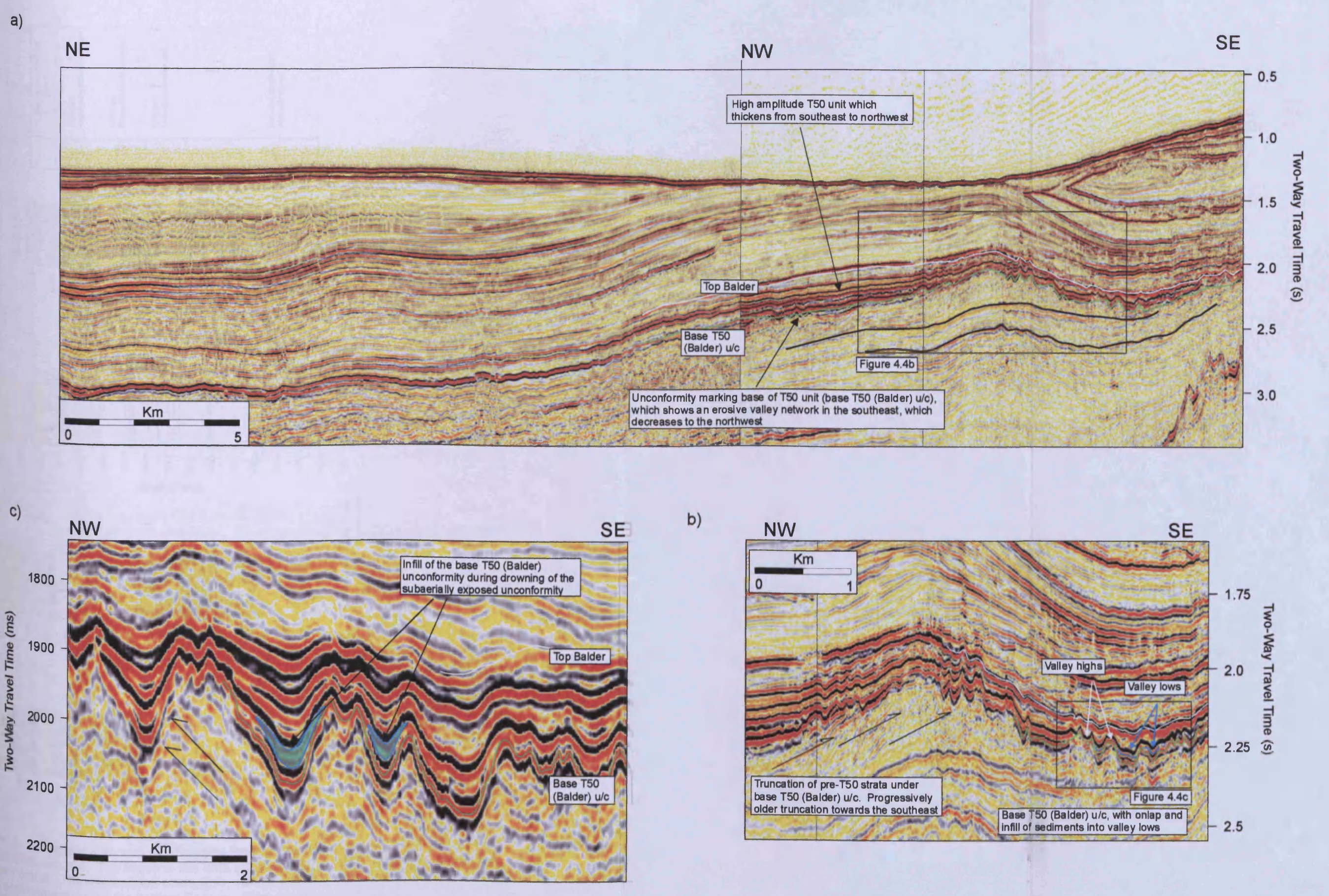


Figure 4.4. a). Composite seismic line from multiple 3-D surveys showing high amplitude seismic character of the T50 unit (Balder Formation equivalent) compared to adjacent seismic units. In the southern area of the Faroe-Shetland Basin over the Judd Anticline, the T50 unit is dominated by continuous high amplitude parallel reflections, reflecting the interbedded nature of sediments (coals, tuffs, sandstones and mudstones - see Figure 4.5). To the north and northeast the seismic character of the T50 unit becomes lower in amplitude and reflections are semi-continuous. b). The base T50 (Balder) unconformity is seen to down-cut into Upper Palaeocene sediments which become progressively older to the south and east (see also Figure 4.1 for subcrop pattern). c). Subsequent infill during deposition of the T50 unit is seen into the eroded valley lows. The relic topography is then subsequently draped. These valley lows show a maximum vertical relief in the region of 200 m. For position of seismic line see Figure 4.1.

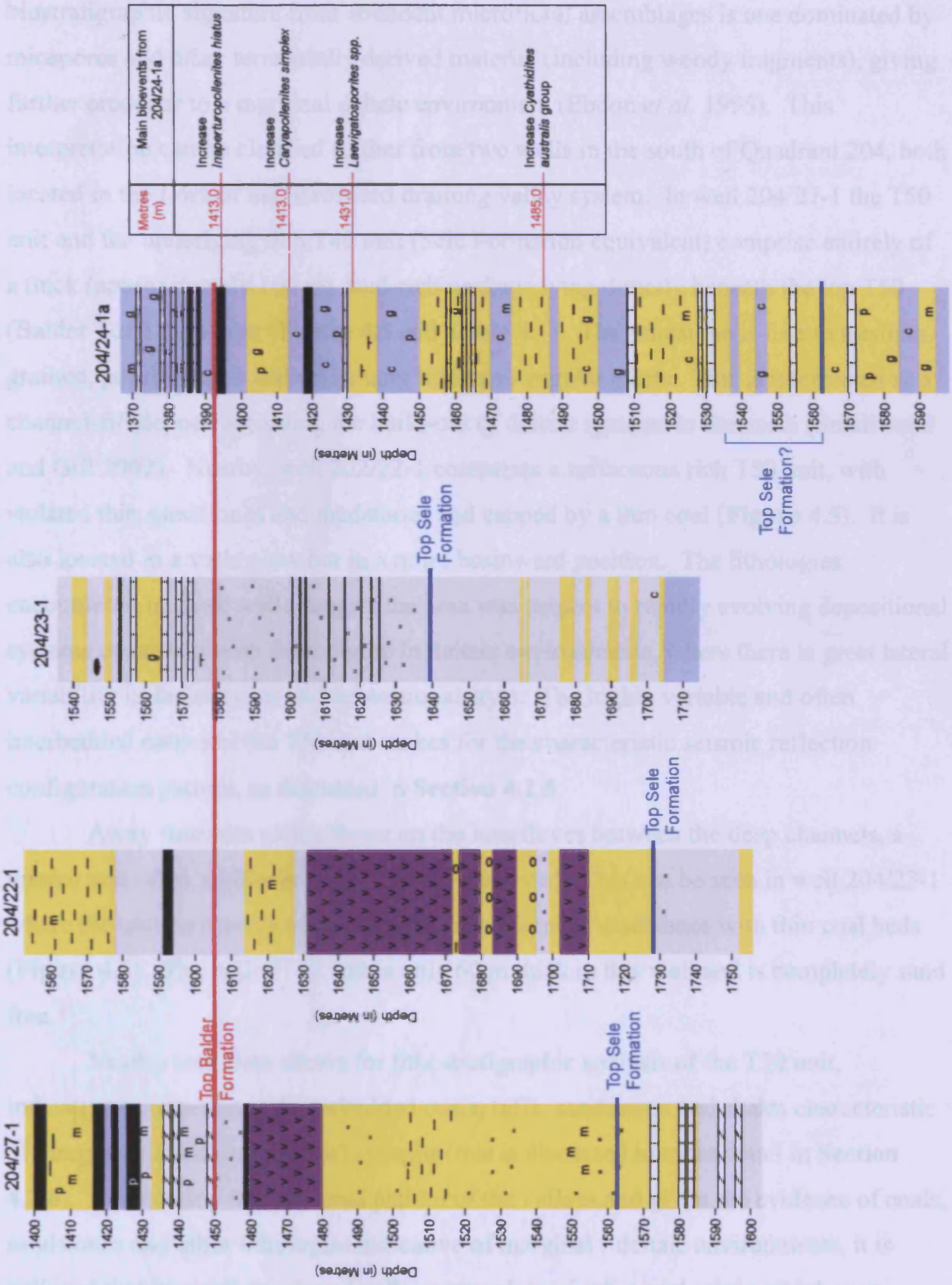


Figure 4.5. Schematic summary of lithologies found in the T50 unit from wells located in the south of Quadrant 204. The facies in this area are varied and change laterally over short distances (km's). The interbedded coals, tuffs, sandstones and mudstones indicate a marginal deltaic environment during the earliest Eocene. Biostratigraphic data from these wells (e.g. from 204/24-1a) corroborates both the age of the unit as earliest Eocene, and the palaeoenvironmental setting. A sub-aerial, marginal environment is interpreted for the southern part of Quadrant 204. All lithological logs are hung from the top T50 (Balder Tuff) event (which marks the top of the T50 unit and is taken from composite logs). This figure also shows the variable thickness of the T50 unit (bounded at the base by the top of the Sele Fm) over a small area in the south of the basin. For the full legend and abbreviations of the lithological information refer to Table 4.1.

biostratigraphic signature from abundant microfloral assemblages is one dominated by miospores and other terrestrially derived material (including woody fragments), giving further credence to a marginal deltaic environment (Ebdon *et al.* 1995). This interpretation can be clarified further from two wells in the south of Quadrant 204, both located in the lows of the northward draining valley system. In well 204/27-1 the T50 unit and the underlying thin T40 unit (Sele Formation equivalent) comprise entirely of a thick (approximately 100 m) sand-rich package lying directly beneath the top T50 (Balder Tuff) reflection (**Figure 4.5** and **Table 4.1**). The sandstone is fine to medium grained, poorly sorted and with many lithic sub-angular clasts. This is interpreted as a channel-fill deposit recording the build-out of deltaic systems to the north (Smallwood and Gill 2002). Nearby, well 202/22-1 comprises a tuffaceous rich T50 unit, with isolated thin sandstones and mudstones and capped by a thin coal (**Figure 4.5**). It is also located in a valley low but in a more basinward position. The lithologies encountered in these wells suggest the area was subject to rapidly evolving depositional systems consistent with those found in deltaic environments, where there is great lateral variability in facies types and depositional style. The highly variable and often interbedded nature of the T50 unit makes for the characteristic seismic reflection configuration pattern, as discussed in **Section 4.2.5**.

Away from the valley floors on the interfluves between the deep channels, a thinner and often argillaceous T50 unit is preserved. This can be seen in well 204/23-1 where the unit comprises of mudstones and tuffaceous mudstones with thin coal beds (**Figure 4.5**). The entire T50 unit is only 60 m thick in this well and is completely sand free.

Nearby well data allows for lithostratigraphic analysis of the T50 unit, indicating a succession of interbedded coals, tuffs, sandstones and shales characteristic of marginal deltaic depositional systems (this is discussed in more detail in **Section 4.2.6**). Considering the erosional pattern of the valleys and given the evidence of coals, sandstones and other lithologies indicative of marginal - deltaic environments, it is believed that the well developed valley network is of sub-aerial origin which was transgressed during T50 times (Smallwood and Gill 2002). This will be discussed in more detail in **Section 4.2.6**.

There is a significant difference in the depositional setting towards the centre and the north of the basin. Northwards of the southern deltaic area (discussed above) the T50 unit is thin (0 - 75 m) and difficult to image on seismic data (see **Sections 4.2.3**

a)

	Conglomerate		Tuffaceous
	Sandstone		Phosphotic/Ferrigenous Manganifereous
	Mudstone		Pebbly
	Siltstone		Calcite
	Coal/Lignite		Silty
	Tuff	p	Pyritic
	Limestone		Argillaceous
	Dolomite		Calcareous
g	Glauconitic	c	Carbonaceous
m	Micaeous		

b)

	Coastal Plain		Basin
	Inner Shelf		Deep water sands
	Outer Shelf		Lava fields
	Slope		Faroe-Shetland Escarpment

Table 4.1. a). Table of contents listing the lithology types and abbreviations used in all lithological well log summaries used in figures throughout this chapter. b). Table summarising the colours of the different depositional environments used in the palaeogeographic reconstructions throughout this chapter.

and **4.2.5**). However, the distinctive top T50 (Balder Tuff) reflection is readily visible and many wells penetrate this interval.

In the centre of the basin (north part of Quadrants 204, 205 and 206, and southern part of Quadrants 213, 214 and 208) wells encountered a T50 unit with predominately tuffs and tuffaceous mudstones (e.g. 205/9-1 and 214/26-1 – see **Figure 4.6** and **Table 4.1**). A consistent upper bathyal biostratigraphic signature was recorded in the wells in the southern part of Quadrants 214 and 208 (e.g. wells 214/30-1, 214/26-1 and 208/19-1). Slightly south of Quadrant 214, certain wells showed signs of terrestrial input in the form of occasional sandstone and lignite fragments (206/1-1a and 206/2-1a). Additionally, well 205/9-1 shows influence of locally marginal and freshwater influences in a restricted marine setting. These strands of evidence together may suggest a proximity to deltaic conditions in the south where significantly different assemblages occur with an unequivocal terrestrial domain (e.g. in southern Quadrant 204 and northern Quadrant 202). It is therefore postulated that the edge of the delta front was located just south of Block 205/9 which was situated in a proximal zone of the outer shelf (see **Section 4.2.6**). More distal (northerly) wells in Quadrants 214 and 208 were situated further out on the outer shelf or slope.

Locally in some wells in the northern part of Quadrant 206, an abundance of limestone and claystones is seen and (wells 206/3-1, 206/4-1 and 206/1-2). In this location, a lagoonal environment has been postulated by Naylor *et al.* (1999) for the older T45 unit (Flett Formation equivalent) (see **Figure 4.7**). The limestones and claystones seen in the upper T50 unit could indicate that a shallow, low energy environment such as a lagoon or outer shelf was still present at this time.

The top T50 (Balder Tuff) reflection can be mapped throughout the entire basin. The structure map of this reflection shows the cumulative effects of uplift, erosion and folding since the deposition of the ash **(Figure 4.8)**. A broad northeast – southwest trending channel is visible which broadly mirrors the present day bathymetry (e.g. compare with **Figure 1.1** in **Section 1.1.1**). The top T50 (Balder Tuff) reflection is deepest in the northeastern corner of the basin in the area of Quadrant 216 and shallows both in a southeastern and northwestern direction onto the Shetland and Faroes platforms respectively. Furthermore, a shallowing to the southeast is also seen where the basin narrows in the southern area of Quadrant 204. This top T50 (Balder Tuff) reflection has been uplifted and eroded on the margins of the basin and in particular on the Munkagrunnur Ridge. This is a major structural feature related to inversion

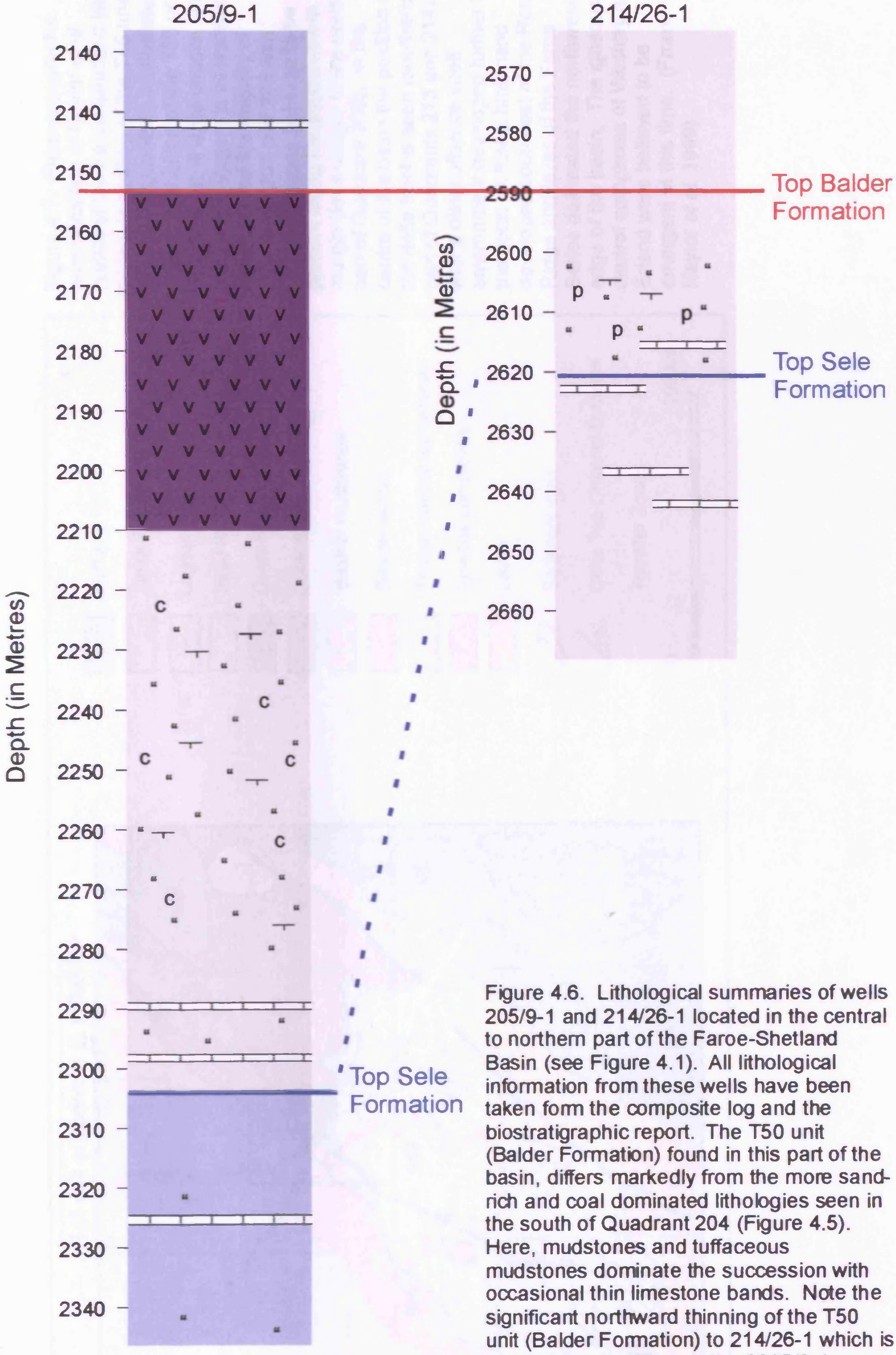


Figure 4.6. Lithological summaries of wells 205/9-1 and 214/26-1 located in the central to northern part of the Faroe-Shetland Basin (see Figure 4.1). All lithological information from these wells have been taken form the composite log and the biostratigraphic report. The T50 unit (Balder Formation) found in this part of the basin, differs markedly from the more sand-rich and coal dominated lithologies seen in the south of Quadrant 204 (Figure 4.5). Here, mudstones and tuffaceous mudstones dominate the succession with occasional thin limestone bands. Note the significant northward thinning of the T50 unit (Balder Formation) to 214/26-1 which is approximately 40km north of 205/9-1. For the full legend and abbreviations to the lithological information refer to Table 4.1

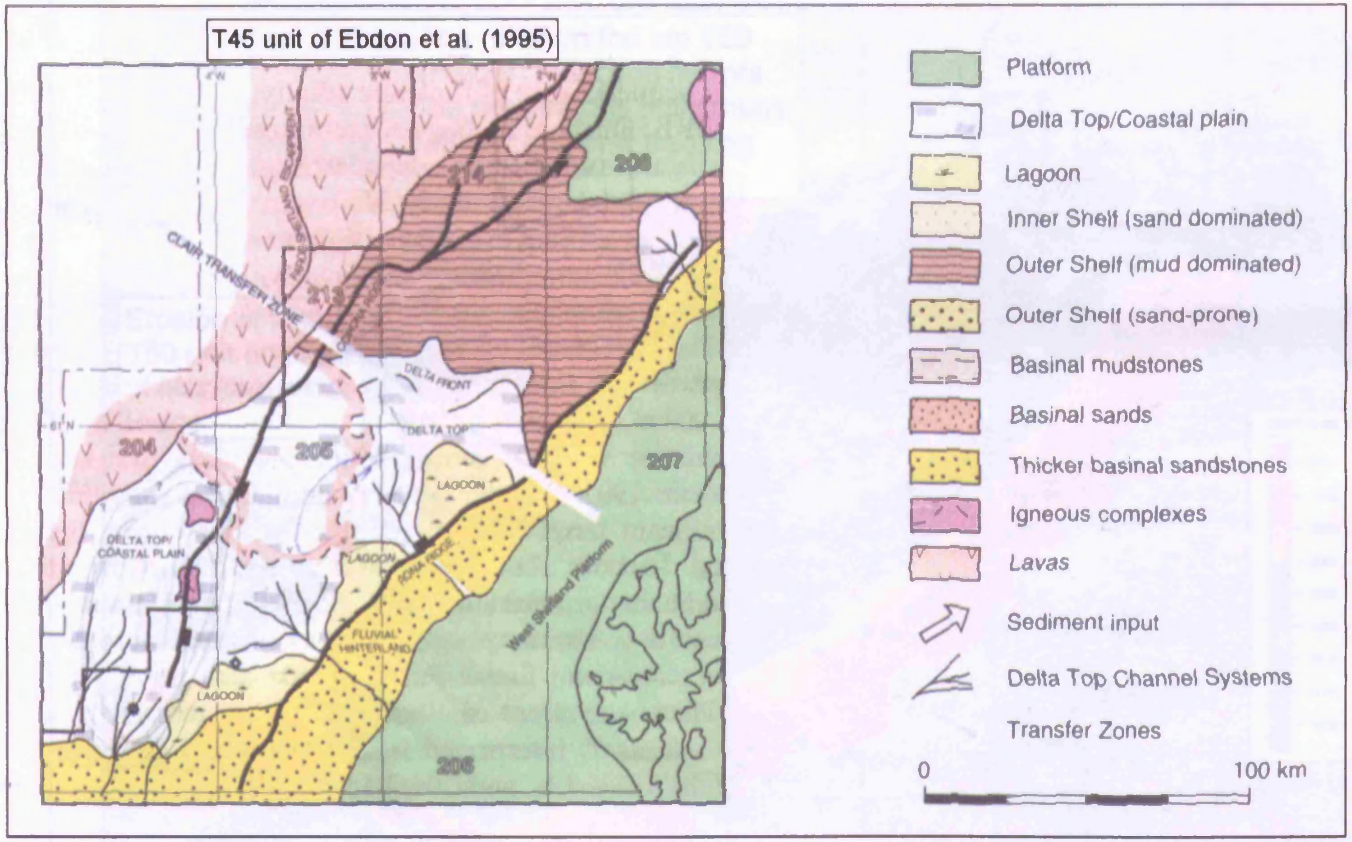


Figure 4.7. Palaeogeographic reconstruction of Naylor *et al.* (1999) of the Faroe-Shetland Basin during deposition of the T45 unit (Flett Fm equivalent) immediately prior to deposition of the T50 unit (Balder Fm). A large coastal plain and deltaic system is interpreted to have covered the majority of Quadrants 204 and 205 with isolated lagoons believed to be present along the southeastern margin (for example in the northern part of Quadrant 206). In the centre of the basin the position of the delta front is seen (southern part of Quadrants 213 and 214) with a distal offshore shelf environment developed further to the north. A fluvial hinterland developed southeast of the Rona Ridge and lavas of the Faroe Series delineated the northwestern edge of the basin. The igneous central complexes of Westray and Erlend were believed to be emergent at this time. (From Naylor *et al.* 1999).

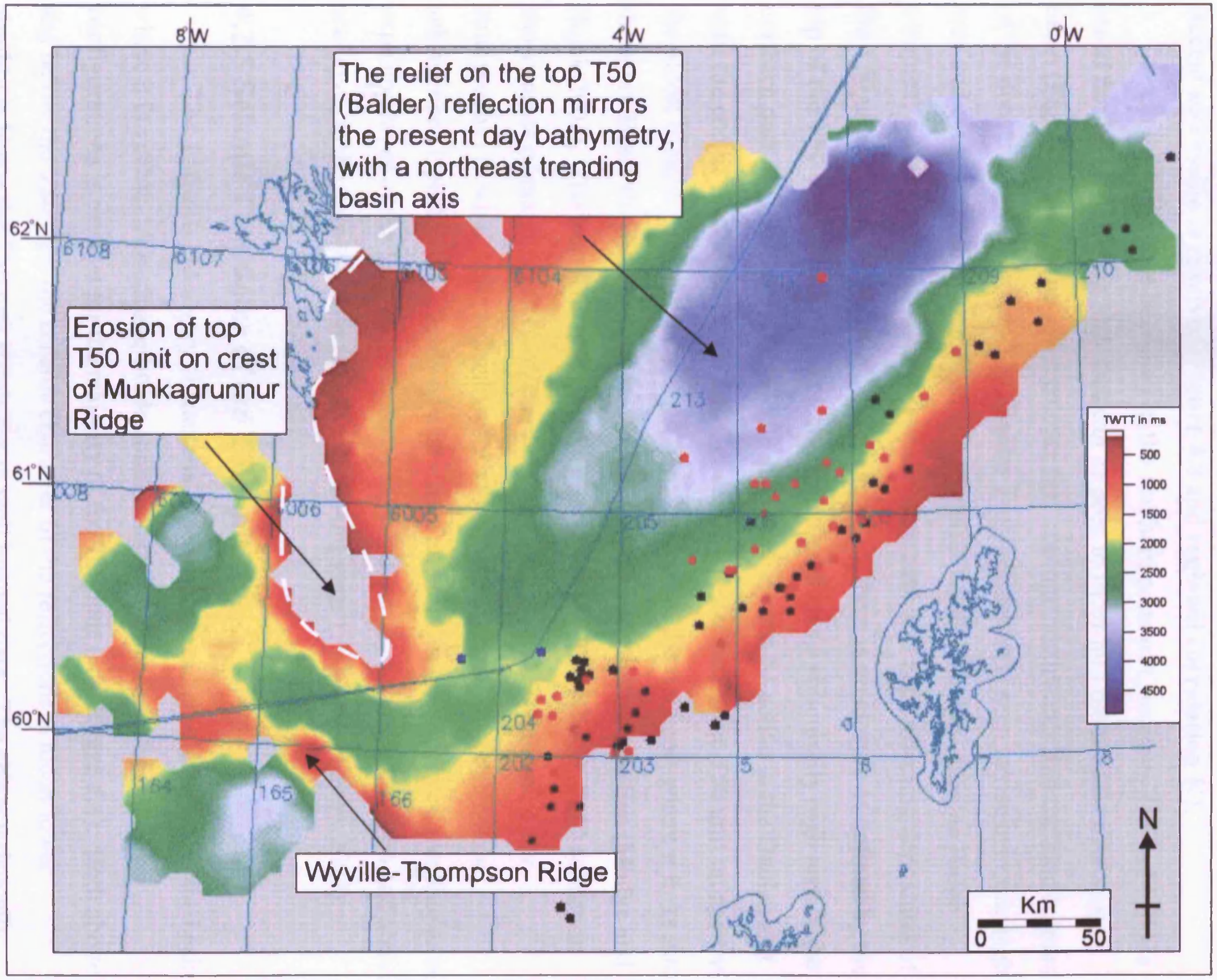


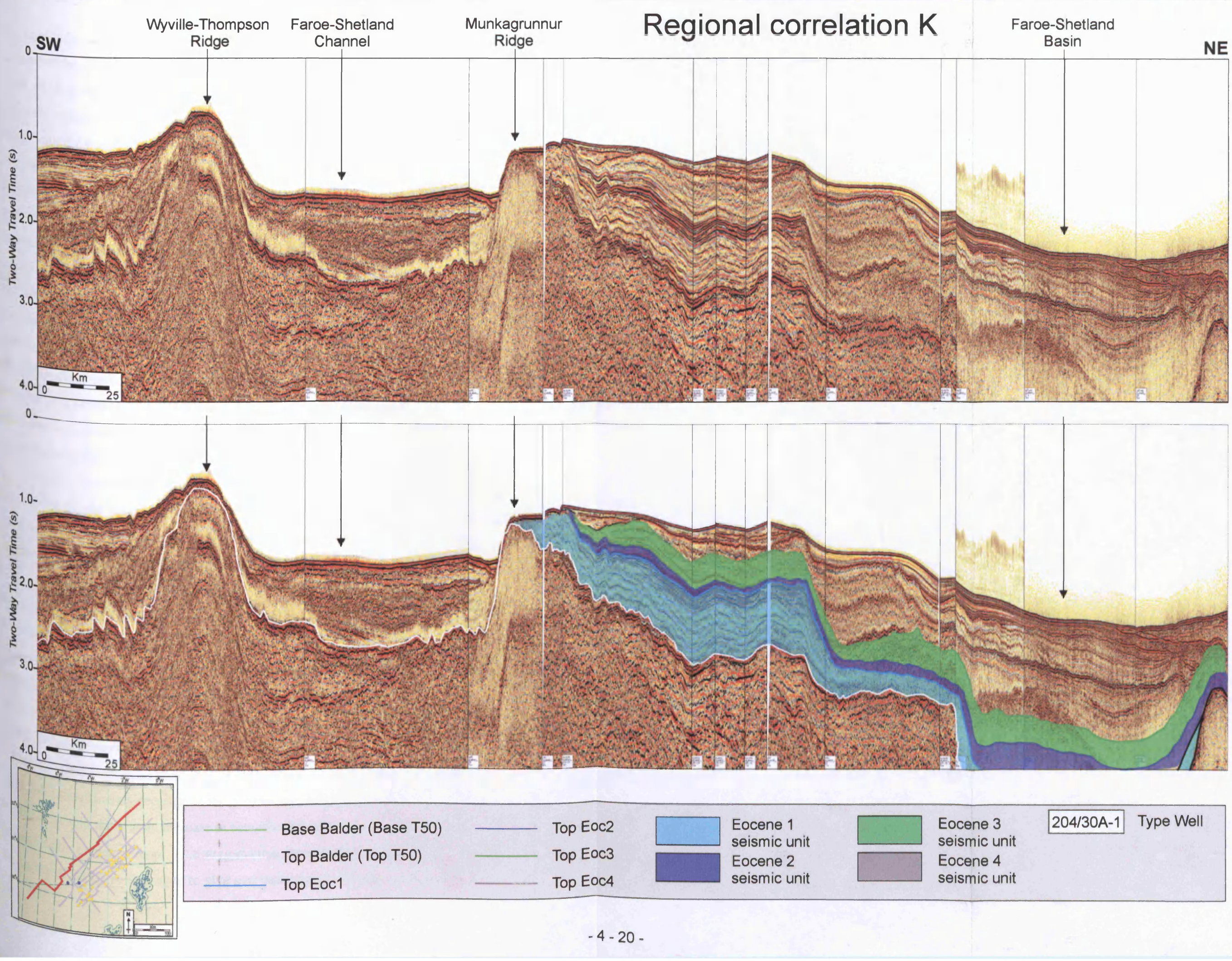
Figure 4.8. Top T50 (Balder Tuff) reflection structure map (in TWTT). The map shows a strongly northeast - southwest trending depocentre that mirrors the present day bathymetry of the basin. The area of the basin is shown in the blue colours and the shallowest by the red colours. The top of the T50 unit is eroded on the Faroe Platform in areas which experienced post depositional inversion e.g. on the Munkagrunnur Ridge and Wyville-Thompson Ridge.

tectonics which forms the southerly extension of the Faroe Islands. In Quadrant 6006, the T50 unit (and much of the Eocene succession) has been removed during a post T50 inversion event (discussed in more detail in **Section 4.2.6**). The more southeast - northwest trending inversion structure of the Wyville-Thompson Ridge is located further south and west of the Munkagrunnur Ridge. This structure folds the top T50 (Balder Tuff) reflection into a narrow anticline where the entire succession shows high amplitude, tight folding and in the majority of places along the ridge axis the Eocene to Recent succession is removed (**Figure 4.8** and **regional correlation K**).

On the UK/Faroe boundary in the southern parts of Quadrants 213 and 6103, a broad northward trending high area can be seen to incur into the broadly oval shaped basin (**Figure 4.8** - see the green colours that impinge northwards into the blue colours of the deeper basin). The top T50 (Balder Tuff) reflection is seen to rise over this high area and could represent the topographically high Westray Ridge - Corona Ridge complex; a Mesozoic structural high which may have still been affecting deposition of the T50 unit. At the northeastern end of the basin, there is no closure as defined by the top of the T50 unit. The Erlend volcanic complex forms a structurally high area in the northern part of Quadrants 209 and 208 and is believed to be active in the Danian and until the end of the Palaeocene (Gatliff *et al.* 1984) after which the T50 unit onlaps onto the oldest basalts. However, the basin remains open to the northeast where it links into the Møre Basin in Mid-Norway and possibly into the Northern North Sea (Mudge and Bujak 2001). The post-T50 unit modification of the top T50 (Balder Tuff) reflection shows areas of major uplift (with places of removal of the T50 unit and older stratigraphy) in what was predominantly a basin experiencing post extensional subsidence. Uplift was focussed on east - west and northeast - southwest trending axes exposing lavas of the Late Palaeocene on the sea-floor on the Faroe Platform and above sea-level on the Faroe Islands (Boldreel and Andersen 1993, Davies *et al.* 2004).

4.2.5 Seismic Character of the T50 unit

The T50 unit has a very characteristic seismic character in the parts of the basin where it is visible. A high amplitude continuous parallel package of reflection configurations is evident for the majority of the T50 unit (see **Figure 4.4**). Both above and below the T50 unit, the seismic character of the reflections is much lower amplitude hence the T50 unit stands out in well imaged data. The characteristically


Regional correlation K
SW
NE
Wyville-Thompson Ridge
Faroe-Shetland Channel
Munkagrunnur Ridge
Faroe-Shetland Basin
Two-Way Travel Time (s)
0
1.0
2.0
3.0
4.0
Km
0
25
Base Balder (Base T50)
Top Balder (Top T50)
Top Eoc1
Top Eoc2
Top Eoc3
Top Eoc4
Eocene 1 seismic unit
Eocene 2 seismic unit
Eocene 3 seismic unit
Eocene 4 seismic unit
204/30A-1
Type Well

high amplitude, parallel and continuous reflection configurations of the T50 unit reflect the interbedded nature of the lithology which consists of coals, tuffs, sandstones and shales which dominate the internal stratigraphy of the unit. The highest seismic amplitude values within the T50 unit occur in the south of Quadrant 204 and generally become slightly weaker northwards as the interval thins (see **Figure 4.4**). However the amplitude value remains generally strong over the whole of the unit when compared to the intervals above and below it. To the northeast and northwest in a more basinal area north of the erosive valley network, the reflection configurations remain parallel and continuous – semi-continuous to the basal T50 unconformity (base Balder unconformity of Smallwood and Gill 2002). In areas of deepest incision of the unconformity onlap into the valley lows is often seen (**Figures 4.4b & c).** In more proximal areas of the valley network one or occasionally two seismic loops over a time interval of 40 ms seem to exhibit this onlap relationship, equating to approximately 30 - 40 m of relief between valley high and valley low. The irregular erosional surface then becomes progressively draped during the deposition of the T50 unit and this has been interpreted as a gradual drowning of the basal unconformity surface during transgression in the earliest Eocene (Smallwood and Gill 2002).

Locally in the south of Quadrant 204 there seems to be narrow zones or chimneys of chaotic seismic reflections which are seen to lie consistently in regions of small valleys highs. Occasionally, these chaotic chimneys pass up above the T50 unit and into the lower amplitude package above.

Although the top T50 (Balder Tuff) reflection is generally mappable over the majority of Northwest Europe, difficulties in picking a single event have occurred in areas away from high quality well data. On the UK side of the basin, numerous wells (in the region of 100) have penetrated the top T50 (Balder Tuff) event. Tying the tuffaceous horizon to the seismic data reveals that there is a characteristic high amplitude response at the top of the T50 unit (but with both negative and positive seismic loops). However, on many composite logs in the Faroe-Shetland Basin (especially in the deep basin axis) the T50 unit shows an increase in the sonic velocity value and a simultaneous decrease in the gamma ray value (**Figure 4.9**). The increase in velocity should show as an increase in acoustic impedance on the seismic data represented by a positive loop. This suggests that there is either a significant change in the density of the tuffaceous layer to give a negative response in the acoustic

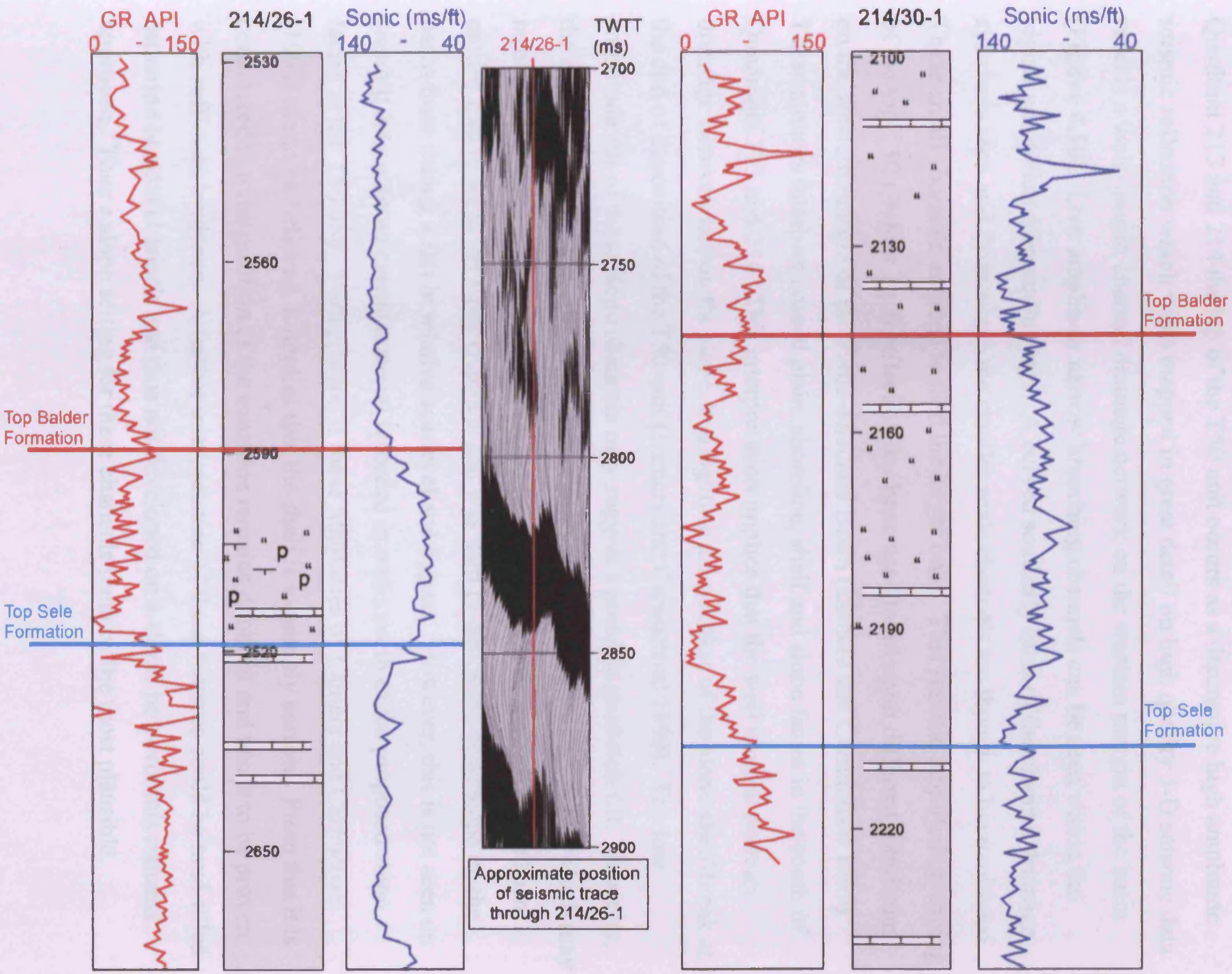


Figure 4.9. Well log summary of 214/26-1 and 214/30-1 showing the characteristic gamma ray and sonic velocity curves seen in the mud-dominated T50 unit (Balder Formation) in the central and northern parts of the Faroe-Shetland Basin. The gamma ray curve is highlighted in red and the sonic velocity curve in blue. Note the increase in the sonic velocity values and the decrease in the gamma ray values across the entire T50 unit (Balder Formation). Seismic trace shown through location of well 214/26-1 showing the observed increase in acoustic impedance for the T50 unit. For the full legend and abbreviations to the lithological information refer to Table 4.1

impedance value or that there is problems with the resolution of the seismic data and the T50 unit may indeed represent an increase in acoustic impedance.

The top T50 (Balder Tuff) reflection in the central and northern sectors of the basin is usually characterised in the wells by a tuffaceous bed (e.g. see well 205/9-1 in **Figure 4.6**) which when interbedded with mudstones and tuffaceous mudstones is interpreted to represent a marine upper bathyal outer shelf or slope setting. The microbiotic signatures of nearby wells also suggest this. In the southern part of Quadrant 213 and 214 the top of the T50 unit occurs as a distinctive high amplitude seismic reflection which when mapped in great detail on high quality 3-D seismic data reveals a well imaged channel drainage network on the western margin of the basin (**Figure 4.10**). Low amplitude narrow branching channels can be seen within the higher amplitude surrounding area. A curved southerly limit of the channel drainage system is seen and from where the smaller scale channels are though to have initiated. These small channels are seen to feed the larger ones. This seismic amplitude mapping of the top T50 (Balder Tuff) reflection has been used to interpret different lithofacies on the western margin of the Faroe-Shetland Basin (Lamers and Carmichael 1999). These authors interpret coastal plain, shoreline, shelf and slope facies in the south of Quadrants 213 and 214. This interpretation implies that the well imaged channel drainage network lies on the slope, highlighting the position of the slope-shelf-break at the end of deposition of the T50 unit (Lamers and Carmichael 1999). The low amplitude fill of these slope channels may suggest a possible mud-rich fill. However, the rest of the slope area reveals an area of high amplitude seismic character which may indicate presence of more sand-dominated facies. This sandstone could have been eroded into to form the slope channels and was then re-deposited downslope on the basin floor during a fall in relative sea-level and bypass. However, this is not seen on seismic data with no clastic material recorded in wells north of the proposed slope facies in the T50 unit. Furthermore, if these lithofacies of Lamers and Carmichael (1999) are to be believed, it implies that the shelf is extremely narrow. From this it is considered that the position of the coastline remains doubtful and needs to be proven with sufficient evidence. A further interpretation of these features could be mud-filled estuarine or tidal channels and thus not developed on a slope, however this remains unproven. Thus a slope setting for these channels remains the most plausible.

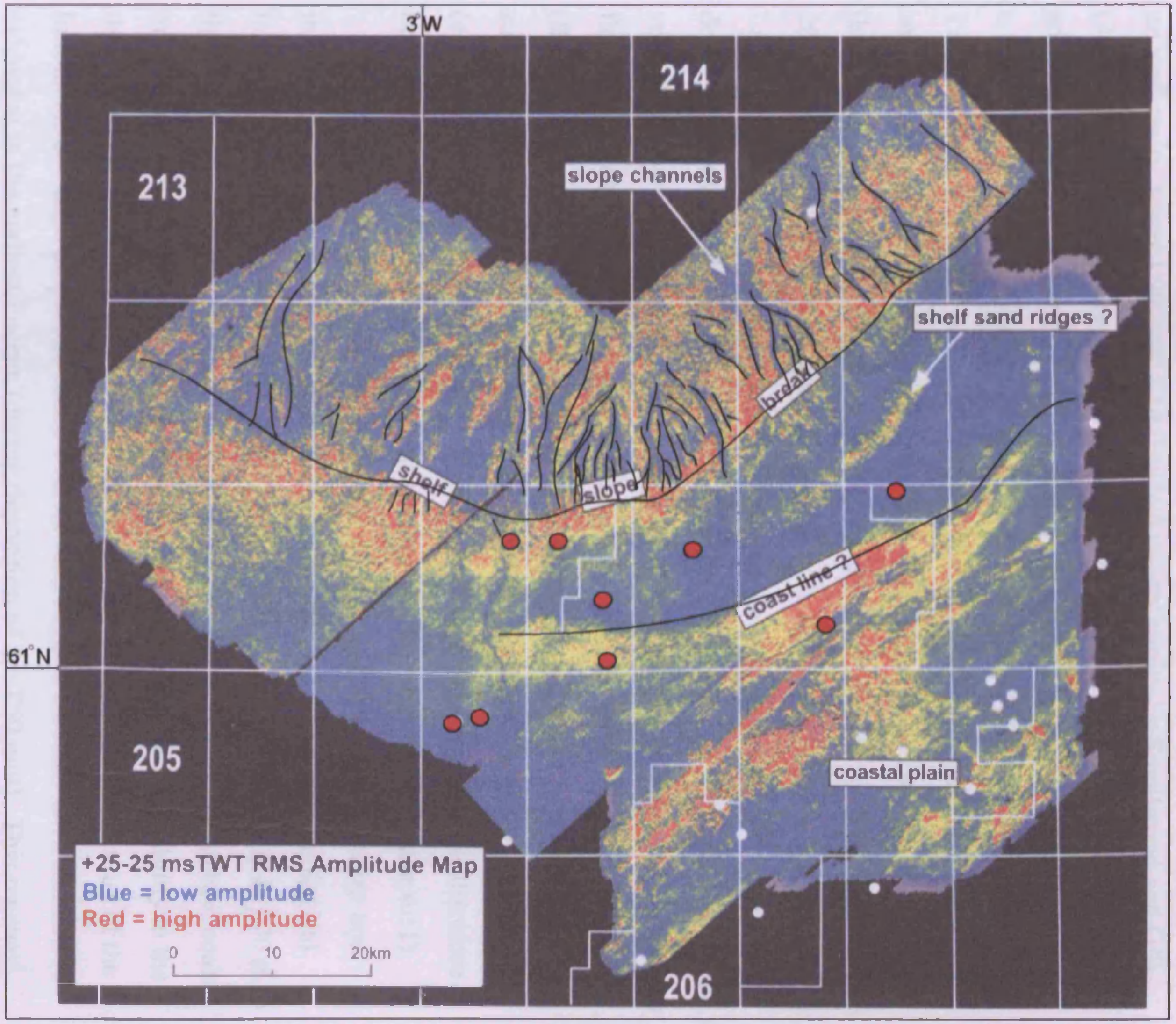


Figure 4.10. RMS (root mean squared) amplitude extraction map of the top T50 (Balder Tuff) reflection in the south of Quadrants 213 and 214, and north of Quadrants 205 and 206. The amplitude map picks out the depositional environment during the earliest Eocene during the last deposition of the T50 unit. Channel systems or gullies are interpreted to have developed on the shelf slope break. To the south there is very little evidence of channel systems and high amplitudes show the possible presence of a more sand-dominated facies in a proximal position on the coastal plain. The slope channels or gullies are picked out in the map by their characteristic low seismic amplitude values (blue colours) which may be indicative of a mudstone or siltstone fill. From Lamers and Carmichael (1999).

4.2.6 Discussion

This discussion will summarise the previous sections (**Sections 4.2.1** to **4.2.5**) into a regional interpretation of the depositional environment during T50 times. It is clear from the seismic interpretation that the T50 unit was predominantly deposited in the southern half of the basin. A broadly concentric succession is found in Quadrants 6005 (Faroes) and 204 (UK) where the thickest preserved section seen is over 300 m. The base of the unit is seen in the southern part of Quadrant 204 to be a highly erosive unconformity which created a northward draining incisional network of valleys. This unconformity, termed the base T50 (Balder) unconformity by Smallwood and Gill (2002), down-cuts into progressively older stratigraphy to the southeast (**Figures 4.1** and **4.4**) indicating a significant fall in relative sea-level to strip off areas previously covered by the Lamba and Flett Formations of units T38, T40 and T45 (Ebdon *et al.* 1995, and **Figure 2.9**). The progressive denudation of these Upper Palaeocene sediments to the south and east indicates that, towards the Judd and Sula Sgeir High, the basin was previously of greater extent and extended towards the south of Quadrant 202. Upper Palaeocene deltaic sediments of the earlier T30 and T40 (equivalent to the Lamba and Flett Formations) prograded towards the north and west and were then down-cut by the semi-regional base T50 (Balder) unconformity. This erosion created a very restricted basin at the time of deposition of the T50 unit, with the shoreline being far more basinward than the previous Late Palaeocene shoreline. The base T50 (Balder) unconformity is seen to become conformable to the northwest and northeast suggesting a more basinal setting than the more marginal areas to the south and east (e.g. **Figures 4.3** and **4.4**). Erosion of the Upper Palaeocene sediments is therefore believed to be restricted to the southern area near to the Judd High (**Figure 4.11**).

During the subsequent deposition of the T50 unit a progressive onlap and infilling of the topography is interpreted (**Figures 4.2, 4.3** and **4.4**). Lithological information from type wells show that in areas of the deepest incision (up to 200 m in the valley lows) a thick T50 succession is found that comprises of sandstones, coals, mudstones and tuffs. There is a significant vertical and horizontal variability in the thickness and lithology respectively of the T50 unit. The interbedded nature of the lithofacies and high amplitude seismic reflections suggest a deltaic environment was evident at in the earliest Eocene (during deposition of the T50 unit). This renewed sedimentation and re-establishment of deltaic condition indicates a rapid transgression

a) Earliest Eocene (early T50) Palaeogeography (time of base T50 (Balder) unconformity)

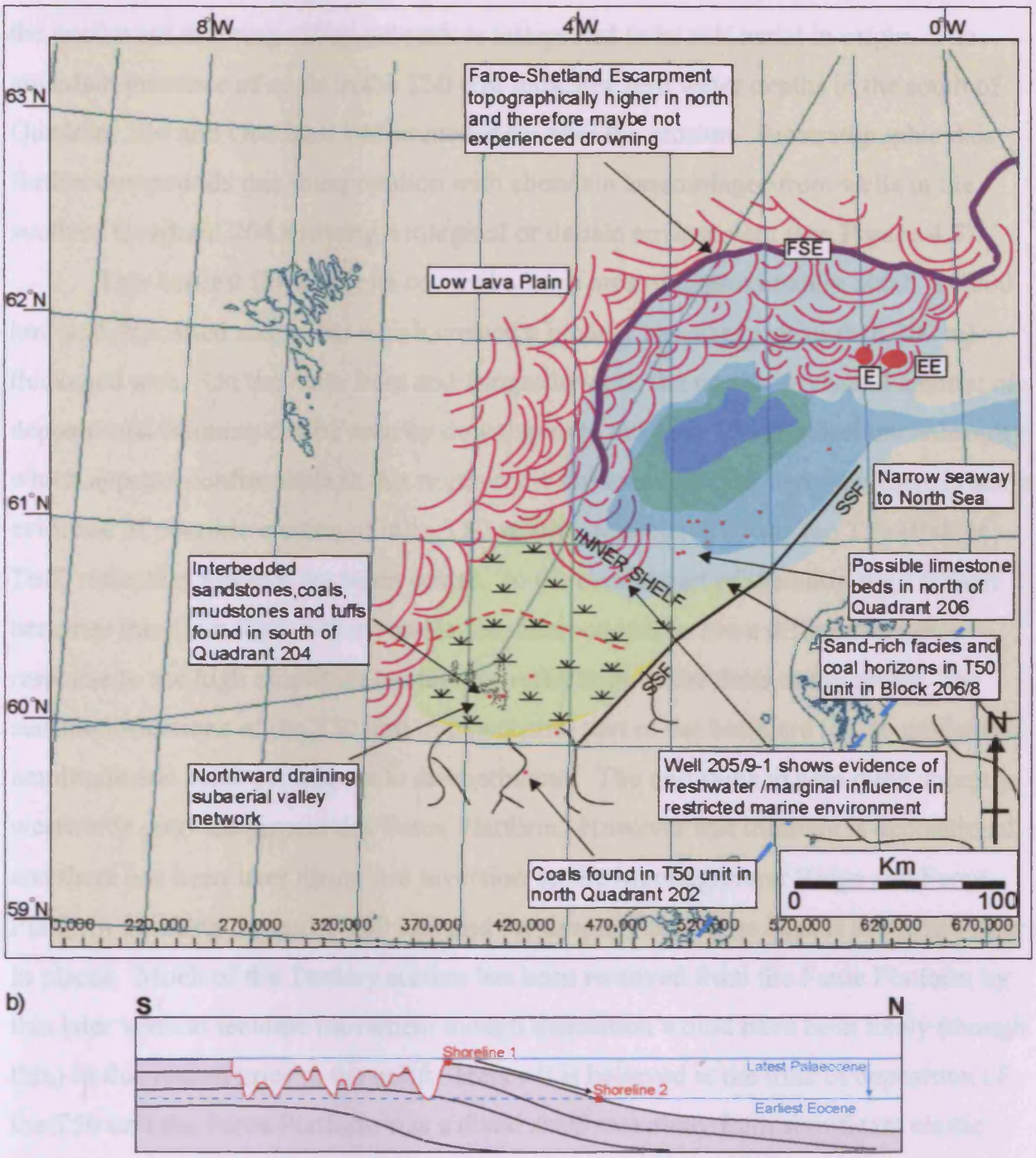


Figure 4.11. a). Schematic palaeogeographic reconstruction in the earliest Eocene (early T50 unit). For legend of the colours used for the depositional environments see Table 4.1 or Enclosure M. This palaeogeographic map represents the time of maximum retreat of the shoreline during the development of the base T50 (Balder) unconformity of Smallwood and Gill (2002). At this time, a major dendritic drainage network developed in the south of the basin that drained northward (see Figure 2.13). This drainage network cut down into the underlying Upper Palaeocene sediments (limit of T40 deposition of Ebdon *et al.* (1995) shown by red dashed line) and is interpreted to be sub-aerial in origin due to its planform geometry. To the north the basin setting was dominantly marine and restricted, with wells in the north of Quadrant 205 and the south of Quadrant 213 and 214 indicating a bathyal signature with influences of marginal marine (e.g. 205/9-1). The Erlend and east Erlend igneous complexes (E and EE) in the northeast remained topographically high, as did the western parts of the large lava plain which dominated the area of the Faroe Platform. Additionally, the northern part of the Faroe-Shetland Escarpment (FSE) remained high and controlled the western limit of the deep marine basin. The eastern margin of the basin was controlled by the position of the Shetland Spine Fault (SSF). b). Relative sea-level cartoon showing the fall in relative sea-level during the development of the base T50 (Balder) unconformity in the latest Palaeocene - earliest Eocene. Note the migration of the shoreline and downstepping to the north (from shoreline 1 - shoreline 2).

of the shoreline indicating a relative sea-level rise after the valleys were cut. Therefore the northward draining valley network is interpreted to be sub-aerial in origin. The abundant presence of coals in the T50 unit indicates zero water depths in the south of Quadrant 204 and Quadrant 202 immediately after the erosion. Biostratigraphic data further compounds this interpretation with abundant assemblages from wells in the south of Quadrant 204 showing a marginal or deltaic environment (see **Figure 4.5**).

This earliest Eocene delta covered a large area of approximately 8000 - 10,000 km^2 and deposited sediments which created a broadly circular (greater than 200 m) thickened area. On the delta front and fringes towards the northeast a small amount of depositional thinning can be seen by downlap onto the base T50 (Balder) unconformity which appears conformable in this northerly distal location. Furthermore, there is some evidence of possible erosion of intra-T50 seismic reflections by the top T50 (Balder Tuff) reflection towards the basin centre. In the central part of the basin the T50 unit becomes thin (less than 200 m) and is less interbedded, so has a different seismic response to the high amplitude continuous reflections of the delta in the south. The seismic reflections of the T50 unit in this central part of the basin are low to moderate amplitude and semi-continuous to discontinuous. The unit thins to zero quite abruptly westwards onto the present day Faroe Platform. However this thinning is depositional and there has been later tilting and inversion on the Munkagrunnur Ridge and Faroe Platform uplifting the entire T50 unit and the deeper Palaeocene basalts to the sea-floor in places. Much of the Tertiary section has been removed from the Faroe Platform by this later vertical tectonic movement though deposition would have been likely (though thin) in this region prior to the uplift. Hence it is believed at the time of deposition of the T50 unit the Faroe Platform was a distal shelf area away from significant clastic terrigenous input and received only minor amounts of sediment that was deposited conformable to the preceding Upper Palaeocene sediments (see **Figure 4.11**).

At the time of development of the base T50 (Balder) unconformity the shoreline that developed during the earliest Eocene was forced northwards into the basin during a relative sea-level fall (**Figure 4.11b**). At this point of maximum regression of the shoreline much of the southern part of the basin (Quadrants 202, 204 and southern - central 205) were sub-aerially exposed in a marginal and coastal plain environment, by a minimum of 200 m (Smallwood and Gill 2002). Therefore a minimum relative sea-level fall of this magnitude is interpreted to have occurred in the earliest Eocene that exposed a localised area in the southern part of the Faroe-Shetland Basin.

Marine conditions prevailed towards the north in the central and northern parts of the basin. A thinned T50 unit (less than 75 m) is found in Quadrants 214 and 208 in the north which can be seen to thicken southwards into Quadrants 205 and 206 (**Figure 4.1**). It is believed that this area lay basinward of the marginal -coastal plain environment in an area that was receiving little or no terrestrial input. This is further corroborated by the wells located in this thinner distal zone where mudstones and tuffaceous mudstones dominate the thinned T50 unit (see wells 214/26-1 and 214/30-1 in **Figures 4.6** and **4.9**).

A palaeogeographic summary shows depositional environments during the earliest Eocene with a broad coastal plain area covering the southern half of the basin in Quadrants 6005, 204 and 205 (**Figure 4.11**). The central and northern parts of the basin remained in a marine setting in an area on the outer shelf or slope that was experiencing upper bathyal conditions. The Flett Ridge in northern Quadrant 205 may have been a topographically high area as the T50 unit thins over the ridge (**Figure 4.11** and **Chapter 6**). A 150 m thick T50 unit comprising of mudstone and tuffaceous mudstone is seen in well 205/9-1 on the northwestern flank of the Flett Ridge. Near the crest of the Flett Ridge the unit thins to 43 m and is dominated by silty claystones which occasionally grades to siltstone and is interbedded with occasional limestone and coal beds. This suggests a slightly shallower area surrounding the Flett Ridge which is interpreted here as being a local submarine high during T50 deposition (**Figure 4.11**).

After this lowering of the relative sea-level (> 200 m) that caused major erosion and incision into the earlier Upper Palaeocene sediments, there was an immediate response (relative sea-level rise) with onlap and infilling of the relic eroded topography by earliest Eocene sediments of the T50 unit. The early initial infill was located in the valleys lows which were cut during the relative sea-level fall (**Figure 4.4**). Later infill became more widespread and indicates a phase of retrogradation of the shoreline during a transgression to deposit sediments of the T50 unit over a more widespread area outwith of the valley networks. The top T50 (Balder Tuff) reflection oversteps the intra-T50 reflections and appears to occur in more landward positions on the basin margins. This may indicate that the transgression was diachronous throughout the deposition of theT50 unit and that there was a gradual flooding back over the previously irregular topography (**Figure 4.12**). Furthermore, the top T50 (Balder Tuff) reflection appears to drape over the irregular topography formed in the earliest Eocene

Earliest Eocene (Late T50 unit) Palaeogeography (top T50 (Balder Tuff) deposition)

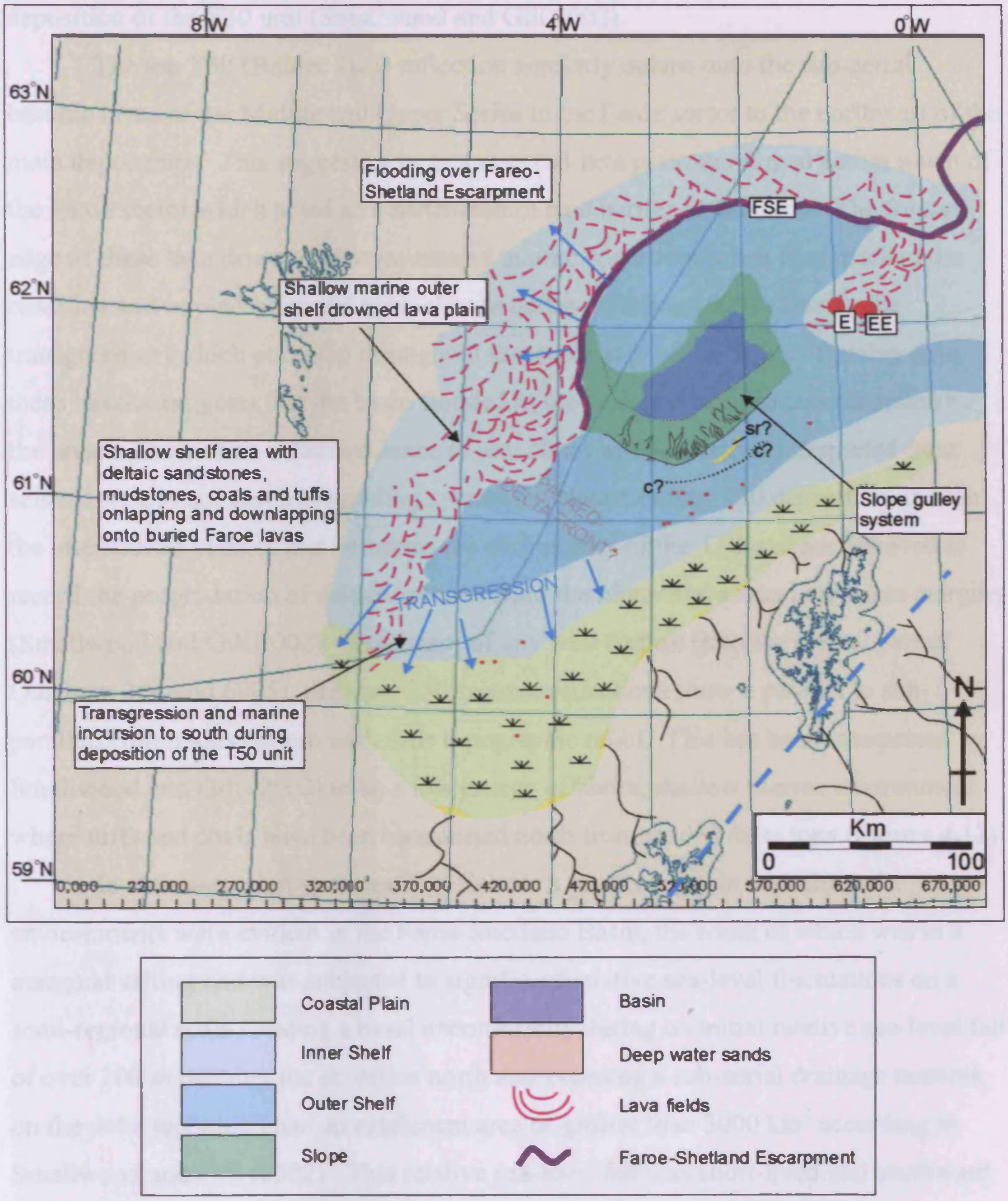


Figure 4.12. Palaeogeographic reconstruction in the earliest Eocene (during deposition of the Balder Tuff at the end of the T50 unit) showing the flooding to the south over the coastal plain and infilling of the previous sub-aerial dendritic drainage network. Back-stepping of facies belts to the south into the north of Quadrant 202 is interpreted during this relative sea-level rise (transgression) in the earliest EoceneT50 times. Flooding over the western low level lava plain also occurred. In the north of the basin, the slope-shelf break remained in a similar position to that found during early T50deposition and gullies or channels are seen. These features are interpreted as slope gullies that transport sediment to the basin centre (see text). The Faroe-Shetland Escarpment (FSE) remained as the northwestern limit to the main marine basin centre in the north. South of the slope gullies, are the interpreted positions of sand ridges (sr) and the coastline (c) of Lamers and Carmichael (1999). This study does not interpret the position of the coastline to be located here, but much further (ca. 50 km) to the southeast (see text). E= Erlend, EE= East Erlend. See Table 4.1 for key to the depositional environments.

and it is believed that this topography had been largely obliterated by the end of deposition of the T50 unit (Smallwood and Gill 2002).

The top T50 (Balder Tuff) reflection similarly onlaps onto the sub-aerial basaltic lavas of the Middle and Upper Series in the Faroe sector to the northwest of the main depocentre. This suggests a large sub-aerial lava plain developed across much of the Faroe sector which acted as a northwestern land barrier to the basin. The frontal edge of these lava flows often encountered marine conditions when they reached the coastline and in places formed hyaloclastite deltas (Kiørboe 1999). During the transgression (which occurred throughout the deposition of the T50 unit) onlap onto these basalts suggests that the basin flooded to the west and became less restricted by the lavas. There is no actual evidence of any clinoform geometries interpreted from seismic data to suggest a prograding system developed during T50 deposition, though the interbedded clastics that comprise the earliest part of the T50 unit are believed to record the progradation of deltaic systems from the hinterland around the basin margins (Smallwood and Gill 2002). Basinward of any well control (into the central part of Quadrant 204 and 6005), the intra-T50 seismic reflections show a parallel to sub-parallel, continuous pattern with little topographic relief. This has been interpreted by Smallwood and Gill (2002) to be a low energy offshore, shallow marine environment where tuffs and coals have been transported north from eroded delta tops (**Figure 4.12**).

In summary, during the earliest Eocene a great variety in depositional environments were evident in the Faroe-Shetland Basin, the south of which was in a marginal setting and was subjected to significant relative sea-level fluctuations on a semi-regional scale creating a basal unconformity during an initial relative sea-level fall of over 200 m pushing the shoreline north and exposing a sub-aerial drainage network on the delta top which had an catchment area of greater than 3000 km^2 according to Smallwood and Gill (2002). This relative sea-level fall was short-lived and southward marine flooding infilling the relic topography resumed during deposition of the T50 unit. In the central and northern Faroe-Shetland Basin the basin experienced marine, outer shelf conditions. Thus a local transient uplift of the southern Faroe-Shetland Basin is a reasonable interpretation of the stratigraphic response. The emplacement and removal of mantle plume material has recently been cited as a causal mechanism for the observed relative sea-level fluctuations (Smallwood and Gill 2002).

4.3 Eocene Seismic–Stratigraphic Units

4.3.1 Introduction

The following section will outline in considerable detail the four seismic – stratigraphic units defined in this study (**Section 3.4**). For each unit, the upper and lower boundaries (defined in **Section 3.4.2**) will be considered and colour coded for display in all corresponding seismic lines. This will be followed by a review of the areal distribution of the unit in the entire basin. The internal seismic geometry and lithological data will then be examined to allow for a concluding stratigraphic discussion for each unit.

4.3.2 Eocene 1 Seismic Unit

4.3.2.1 Introduction

The Lower Eocene succession that was deposited immediately after latest Palaeocene – earliest Eocene T50 unit of Ebdon *et al.* (1995) has a markedly different seismic character than the highly reflective T50 unit discussed above (**Section 4.2**). This section will describe the age, distribution and seismic character of the unit overlying the T50 unit, and go on to discuss the depositional history in the basin immediately after the deposition of the top T50 (Balder Tuff) reflector. This package of sediment forms the first major seismic-stratigraphic unit of this study in the Eocene succession (defined in **Section 3.4**) and is termed the Eocene 1 seismic unit.

4.3.2.2 Upper and Lower Boundaries of Eocene 1

The Eocene 1 seismic unit is the first of four seismic units defined in this study. As described in **Chapter 3**, these seismic units form the basis of the regional 2-D seismic interpretation of the entire Faroe-Shetland Basin discussed herein. The upper and lower boundaries of these seismic units are regionally correlatable seismic reflections which can be seen across a significant part of the basin. In places these reflections have stratigraphic significance (i.e. they show reflection terminations (including onlap, downlap, toplap and erosional truncation) and are often visible on high resolution data), however they have primarily been chosen as the best possible, and most regionally correlatable seismic reflections. By rigorously evaluating the seismic picks and tying these to the best possible well data for the Eocene succession it is believed that these major bounding surfaces to the regional seismic units represent best fit chronostratigraphic surfaces which can be used to build a stratigraphic framework across the entire basin. For a more detailed discussion of this pragmatic approach to sub-dividing the stratigraphy refer to **Sections 3.4** and **3.5**.

The lower boundary of the Eocene 1 seismic unit is the top T50 (Balder Tuff) reflection which represents an almost isochronous volcanic event which occurred at the time of near instantaneous eruption from the initial formation of the mid-ocean spreading ridge centre to the northwest (e.g. Knox and Morton 1988). As has been mentioned in **Section 4.2,** this reflection is readily mappable across much of Northwest Europe and has a characteristic seismic reflection at the top of the T50 unit. On all seismic displays and other figures this lower boundary is displayed as the Top Balder and appears as a white horizon.

The upper boundary of the Eocene 1 seismic unit is defined in the southern part of the basin (in the centre of Quadrant 204). The reflection is named here as Top Eoc1, and appears as a blue horizon (and named as Top Eoc1) on all seismic lines and figures. This upper boundary exists as a moderate to high amplitude continuous seismic reflection which appears close to the top of a parallel and continuous package of similar reflections (**Figure 4.3**). When the reflection is traced northwards into the basin centre, moderate to high amplitude reflections both above and below the reflection are no longer present. This upper boundary is therefore prominent towards the northern part of the basin where it appears as a high amplitude reflection in a low amplitude seismic zone (**Figure 4.3**). The Top Eoc1 reflection has good correlatability in the southern part of the basin and thus there is a high degree of confidence for the seismic pick in this area. Towards the centre and northern parts of the basin the reflection exhibits a moderate to high seismic response which becomes semi-continuous to discontinuous in places and often chaotic further north (e.g. **Figure 4.2**). In these areas the level of confidence for the reflection is moderate to low, which reflects the difficulty in correlating the pick and the paucity of good well control in this part of the basin (see **Figure 1.3**).

4.3.2.3 Distribution of Eocene 1.

The Top Eoc1 reflection can be mapped throughout the entire basin, and it is only difficult to correlate in northern areas. It is truncated in the south by younger reflections (see later **Sections 4.3.2.5** and **4.3.2.6**). The Top Eoc1 reflection is seen to dip towards the northeast and northwest of the basin and can be seen to downlap onto a parallel reflection which lies above and parallel to the top T50 (Balder) reflection (**Figure 4.2**). A time structure map of the Top Eoc1 reflection (**Figure 4.13**) shows a northeast – southwest trending depocentre which is deepest at its northeastern end and has a similar geometry to the modern day bathymetry (e.g. **Figure 1.1**). On the southeastern Shetland margin the depositional limit of the Top Eoc1 reflection lies further basinward than that of the underlying top T50 (Balder Tuff) reflection. This can be seen in the central part of Quadrant 206 and the northern part of Quadrant 207 where the southeastern edge of the top T50 (Balder Tuff) reflection comes to within approximately 25 km of the present day coastline of the Shetland Islands (**Figure 4.13**). In this region the Top Eoc1 depositional limit is (on average) a further 25 - 30 km to the northwest.

A time thickness (isochron) map of the Eocene 1 seismic unit is shown in **Figure 4.14**. This map shows a broadly circular depositional body which corresponds to an 800 ms thick succession of Lower Eocene sediments. This depositional body is located over the central part of Quadrants 6005 and 204 and has a diameter of approximately 100 km (**Figure 4.14**). A significant thinning of the circular body to less than 200 ms to the northeast and northwest and seismic data reveals a prominent northerly downlapping of reflections onto the intra Eocene 1 reflections that lie parallel to and above the top T50 (Balder Tuff) reflection (see **Section 4.3.2.5** and **Figures 4.2, 4.3** and **4.4,** plus **regional correlations A, B, J** and **K**). The downlapping terminations of reflections to the northwest (onto what is now the present day Faroe Platform) have been rotated from their original depositional position. Therefore, these terminations now appear to show a climbing onlap pattern onto the Faroe Platform (**Figure 4.15**). However, these reflections have a convex downwards geometry and bend towards the surface onto which they terminate, therefore indicating that the reflections originally downlapped onto a near horizontal surface which has since been tilted and rotated due to later vertical tectonic movements. The downlapping reflections to the northeast do not appear to look like climbing onlap as no tilting or rotation of the original near

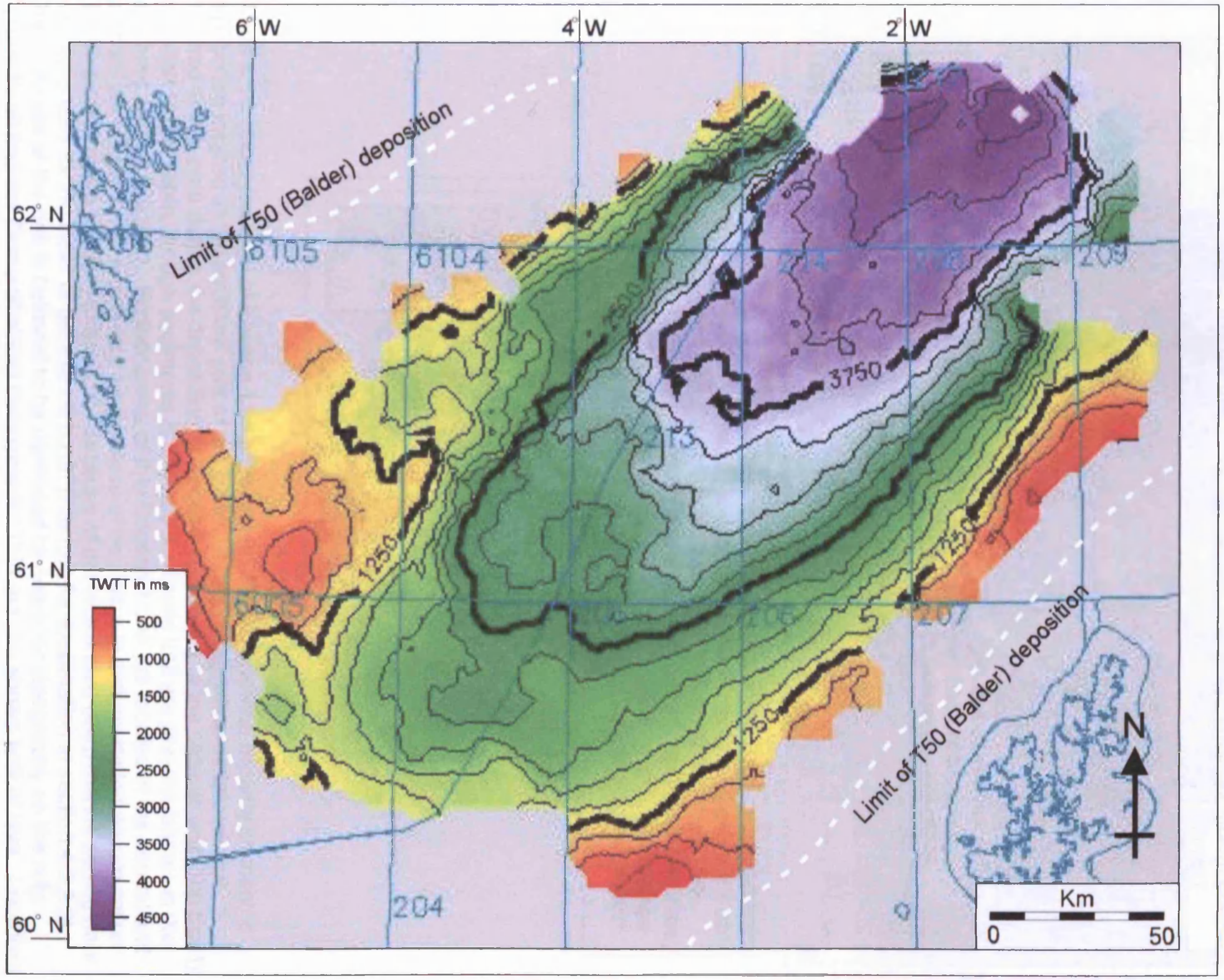


Figure 4.13. Structure map (in TWTT) showing the Top Eoc1 reflection (top of seismic unit Eocene 1). Note the limit of the top T50 unit (Balder Fm equiv) reflection (see Figure 4.8) occurs approximate 20 km more landward of the Top Eoc1 reflection on both the Shetland and Faroe margins. The deepest area of the Top Eoc1 reflection is seen in the northeast of the basin. The surface shallows to the southwest and towards both margins. Contour intervals are in milliseconds (ms).

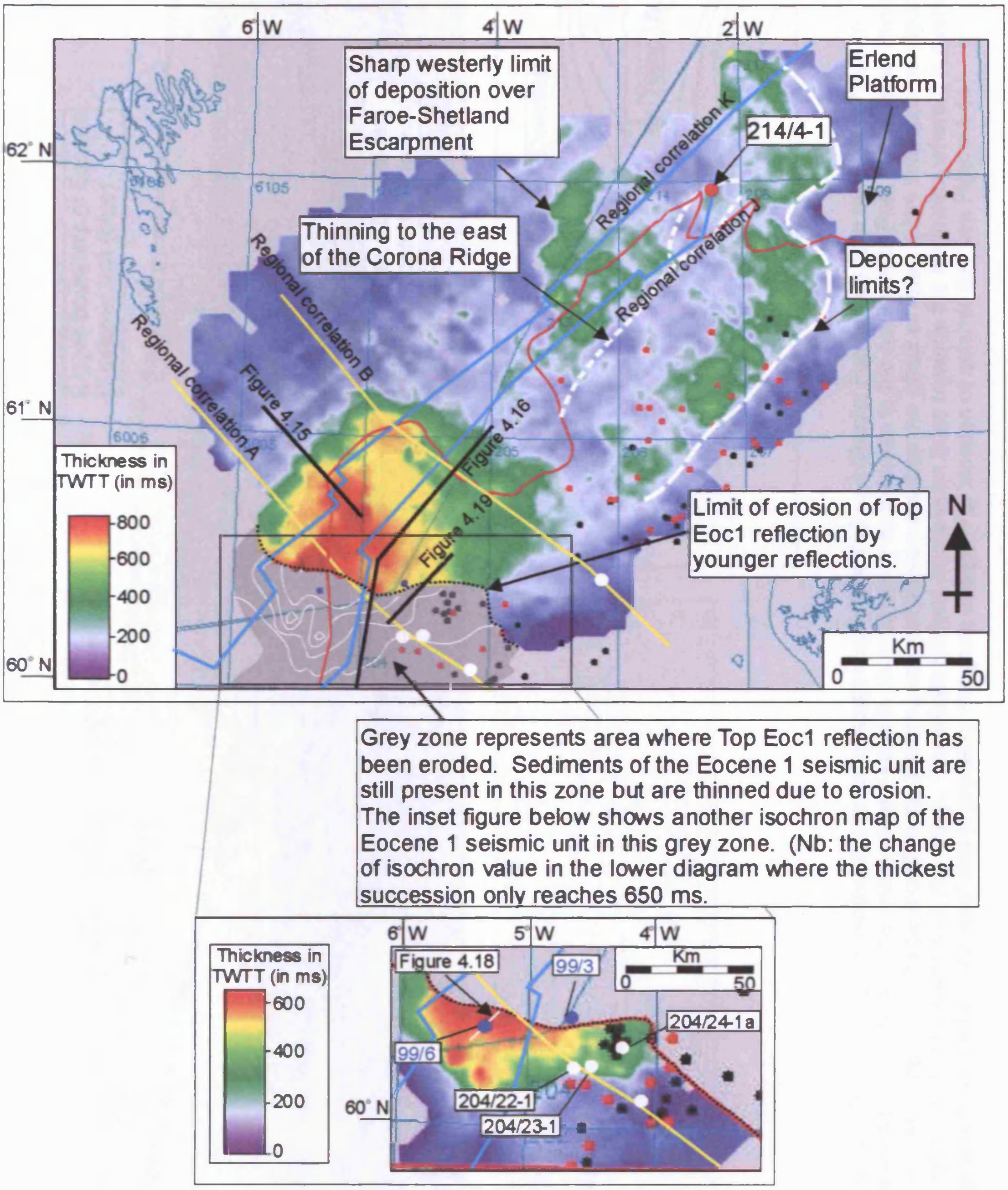


Figure 4.14. Isochron map of Eocene 1 seismic unit (in TWTT) showing broadly concentric depocentre located in the southern part of the basin. A large prograding deltaic wedge occurred in this area during the deposition of Eocene 1. The top of the seismic unit (Top Eoc1) is eroded to the south, though sediments of Eocene 1 seismic unit are present (shown in the grey area and inset figure). Further areas of thin Eocene 1 unit are located in the central and northern part of the basin which may be because of decrease in sedimentation accumulation (due to a condensed section) or erosion. Because of the interpreted deep marine setting here it is interpreted as the limits of the marine basin (see text for explanation in section 4.3.2.3). The thickness of the unit is believed to be controlled by sea-floor topography on the relic Corona Fault and the Faroe-Shetland Escarpment. Thin red line shows limit of lava. Locations of seismic lines and regional correlations (tied to type wells) are also highlighted on both figures.

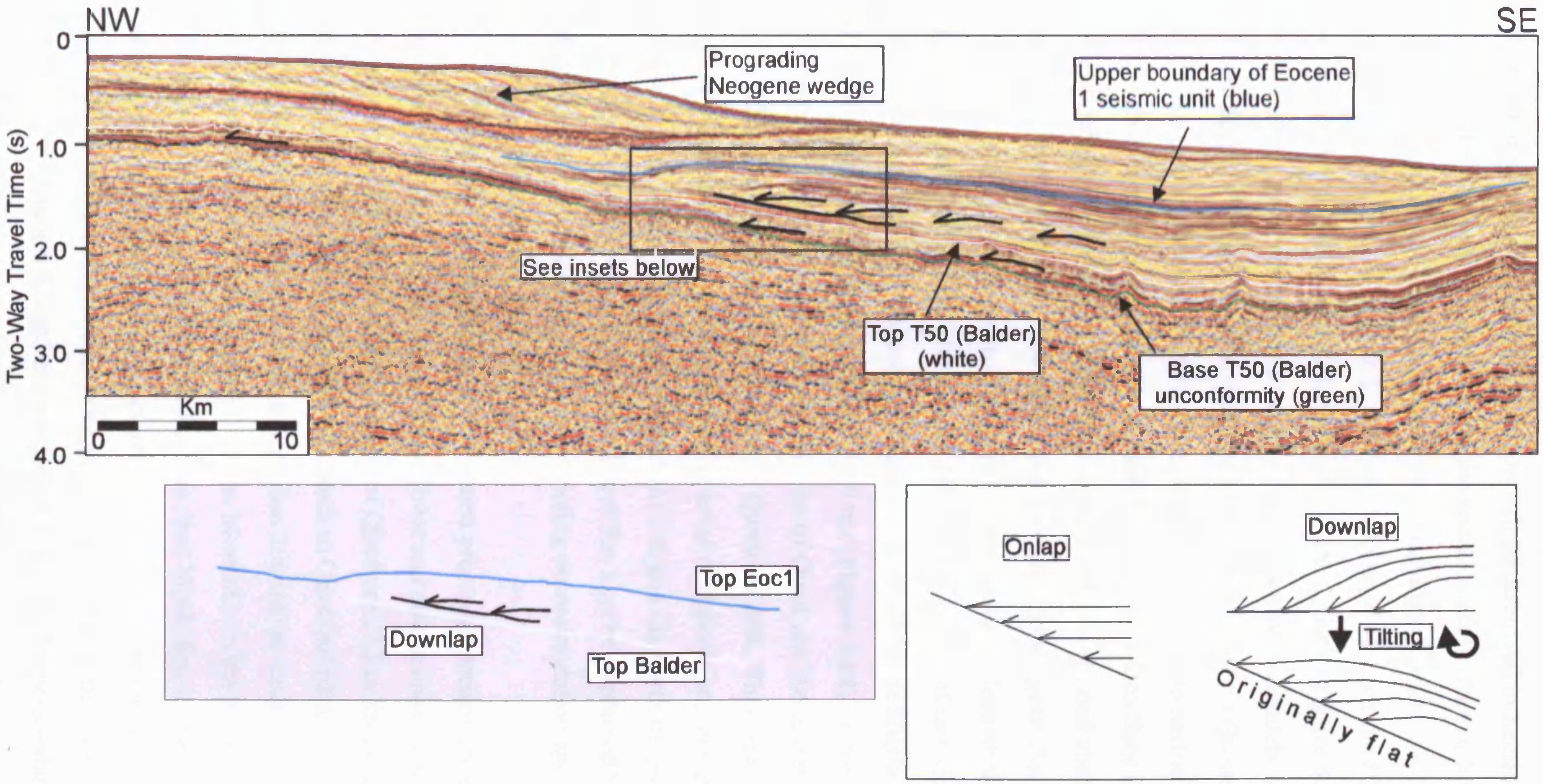


Figure 4.15. NW -SE 2-D seismic line showing the westwards thinning and onlap of the T50 unit (Balder Formation) onto the Faroe Platform. The top T50 Balder Tuff) reflection is marked in white and the unit is clearly seen to show thinning by downlap to the northwest. Post depositional inversion or tilting of the Faroe Platform and Munkagrunnur Ridge modifies the Palaeogene succession and has the affect of rotating the reflection configurations to a position which makes the reflection terminations look like onlap. The whole Eocene succession thins towards the northwest where it is overlain by a much younger deltaic wedge which progrades southeast and is probably Neogene in age. For location of seismic line see Figure 4.14.

horizontal depositional surface has occurred (**Figure 4.15**). The significance of this thickened area of the Eocene 1 succession will be discussed in **Section 4.3.2.6**.

In the northern part of the basin the seismic data is often of low resolution and also the reflections are semi-continuous to discontinuous, often chaotic (and faulted). The thinned areas of the Eocene 1 succession occur away from the 800 ms thick concentric area in the south and are between 0 and 250 ms (**Figure 4.14**). There are however, two areas where the succession thickens to approximately 400 ms. The first of these areas is located close to the political border centred on Quadrant 213 and 6103 (shown as green colours in **Figure 4.14**). Here, a northerly to northeasterly trending area (with a thickness of 300 - 400 ms) can be seen. This area thins sharply at its westerly edge to less than 200 ms. This westerly edge is located above the Faroe-Shetland Escarpment (see **Sections 2.2.3.4** and **2.3**) and suggests that the escarpment had a topographic or bathymetric influence on the locus of deposition of sediments within the basin at this time (for a more detailed discussion of this see **Section 4.3.2.6**). The second anomalously thickened northern area parallels the Shetland margin and also has a preserved time thickness of 300 – 400 ms (**Figure 4.14**). It forms a curved belt which trends northeast from the northern part of Quadrant 206 and then bends to the north and northwest in the northern part of Quadrant 208. This area may represent the shape of the basin at this time highlighting areas of deposition, though this is speculative. In the far northeast of Quadrant 208 and the north of Quadrant 209, there is no deposition of Eocene 1 seismic unit and this may be explained by the continued topographically high Erlend Platform restricting accommodation space development and marine deposition in this area.

These two northerly areas of thickened preserved section are separated by a northeast trending area of thinned section (blue and purple colours on **Figure 4.14**). This area stretches from the southern part of Quadrant 213 to the northern part of Quadrant 214 and similarly bends to the north in Quadrant 6202 (north of 214). This thin area (with a preserved section of less than 200 ms) is located above the Corona Ridge fault block which was active from the Mesozoic to the Palaeocene according to Dean *et al.* (1999) (see **Section 2.2.3**). This fault block dips to the southeast and forms the western most (known) fault block of the continental margin in the Faroe-Shetland Basin. The isochron map indicates a thickening of the Eocene 1 unit (up to 400 ms) to the west of the Corona Ridge between the ridge and the Faroe-Shetland Escarpment, whilst east of the ridge there is a thinned (approximately 200 ms) section (**Figure 4.14**).

The change in thickness of the Eocene 1 seismic unit over the Corona Ridge may suggest that there was continued relic sea-floor topography which to a degree, controlled the location of the basin centre and thus influenced sediment deposition in this area.

Locally within the thin area east of the Corona Ridge there are isolated regions of slightly thicker Eocene 1 section which are approximately 10 km^2 in area. For example, a small localised thickened area around the recently drilled type well 214/4-1 (and the unreleased 214/9-1 well) can be seen in **Figure 4.14.** This area may simply represent a subtle change in the sea-floor topography and be located in a small low, allowing for greater accumulation of sediment. The reason for this low may be explained by the development of a small elongate northerly trending mini-basin which developed in the earliest Eocene, during the deposition of T50 unit (Trude *et al.* 2003) and may have continued later affecting deposition of the Eocene 1 seismic unit. The mechanism behind the development of this mini-basin is believed to be explained by the "jack-up" of the surrounding sea-floor by shallow level sill complexes which had a significant effect on sediment dispersal patterns on the basin floor (Trude *et al.* 2003). Additionally, mapping of the Faroes Lava Group in the north of Quadrant 214 shows two southerly protruding tongues of lava which are located at the northern tip of the Corona Ridge (see **Figure 4.11**). A depositional low is interpreted in between these two tongues of lava on the sea-floor thus allowing for sediment accumulation.

The top of the Eocene 1 seismic unit (Top Eoc1) is difficult to map accurately in the central and northern part of the basin. However, by looking at the general seismic character of the interval as a whole and by looking at the general dip of the seismic package both above and below the Top Eoc1 reflection a crude, best fit seismic marker can be determined. Because, there are no continuous mappable reflections across the entire northern and central sectors of the basin, this best fit approach allows reflections in the southern part of the basin to be correlated into areas where there is greater difficulty in the seismic interpretation due to the discontinuous or chaotic reflections. This technique (known as phantoming – see **Section 3.3.2**) has its limitations and is not an accurate means of mapping seismic reflections. However, it has its merits when looking at regional basin-wide seismic analysis allowing large scale gross seismic units to be chronostratigraphically tied together over distance of a few hundred kilometres. Furthermore, this technique puts a great amount of emphasis on any available well data which can constrain the age of the succession. This is further exacerbated in the central and northern parts of the basin where there are less than ten good well ties (see **Section 3.5**) none of which are located in Faroese waters.

What this technique of correlation across seismically irresolvable data does allow, when tied to well data, is basin-scale structure and thickness maps to be produced. The majority of wells in the northern and central sector of the basin have been drilled relatively recently and therefore often have high quality biostratigraphic data collected with them due to the drilling techniques used (see **Section 3.5** and **Figures 3.12** and **3.13**). Furthermore, at least two of these recent wells were drilled specifically with Eocene targets in mind and thus wireline logs record the succession accurately.

The Top Eoc1 reflection is seen to be erosionally truncated in the south of the basin (by the younger Top Eoc2 reflection). This limit of erosion of the Top Eoc1 reflection maps out linearly in an east - west direction, on the northern flank of an east - west anticlinal fold hinge called the Judd Anticline (**Figures 4.14, 4.16** and **regional correlations A, J** and **K**). Detailed seismic mapping has shown that the Top Eoc1 reflection becomes eroded approximately 10 km north of the Judd Anticline fold axis just south of the centre of Quadrant 204 (**Figure 4.16**). Because the Top Eoc1 reflection is eroded and no longer preserved to the south of this limit, the Eocene 1 isochron map shows a zero thickness for the Eocene 1 seismic unit in the southern. However, this does not mean that no sediment is present here only that the upper boundary of the unit is no longer preserved. The isochron map in this area is shown in the inset diagram in **Figure 4.14** and is seen to show a similar pattern of deposition to the main thickness map, completing the southerly edge of the broadly concentric depositional thick. By mapping parallel intra-Eocene 1 seismic reflection configurations which are slightly older than that of the Top Eoc1 reflection it can be seen that Lower Eocene sediments are preserved over the Judd Anticline and indeed further south into the southern part of Quadrant 204 and northern part of Quadrant 202. Detailed mapping of one of the intra Eocene 1 reflections results in a similar isochron map of the depocentre (**Figure 4.17**). The thickest preserved section (approximately 400 ms) is located in the centre of Quadrant 6005 in the same position as the T50 unit and entire Eocene1 unit. However, there is still some post-depositional erosion and removal of this interval in the region of the Judd Deeps seen by erosional truncation of intra-Eocene 1 reflections close to the present day sea-floor. The areal extent of this erosion forms a narrow east - west trending region where present day erosion causes

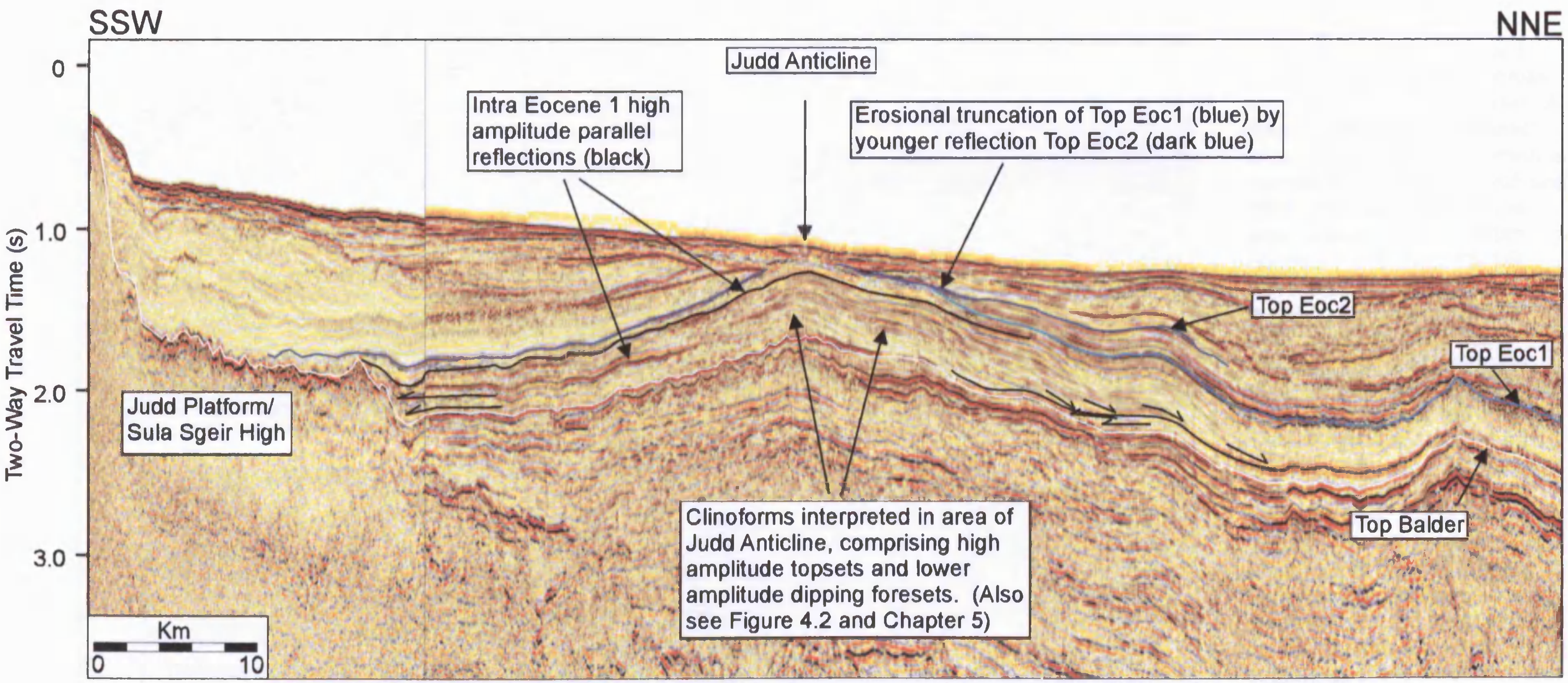


Figure 4.16. Composite 2-D seismic transect trending south southwest - north northeast across the Judd Platform and Judd Anticline. In the south the Judd Platform and Sula Sgeir High are visible onto which the Palaeocene and Eocene succession thins. The T50 unit (Balder Formation equivalent) thickens to the north northeast in this section and is characterised by has a high amplitude, high frequency parallel and continuous seismic reflection configuration. In the area of the Judd Anticline the Eocene 1 unit shows local areas of high amplitude reflections interpreted as coals. Progradation of clinoform systems is also seen over the anticline indicating a sediment source to the south and southeast. The Top Eoc1 reflection is eroded by the Top Eoc2 reflection on the northern flanks of the Judd Anticline, though intra Eocene 1 sediments are preserved to the south of the structure and over the crest. For location of line see Figure 4.14.

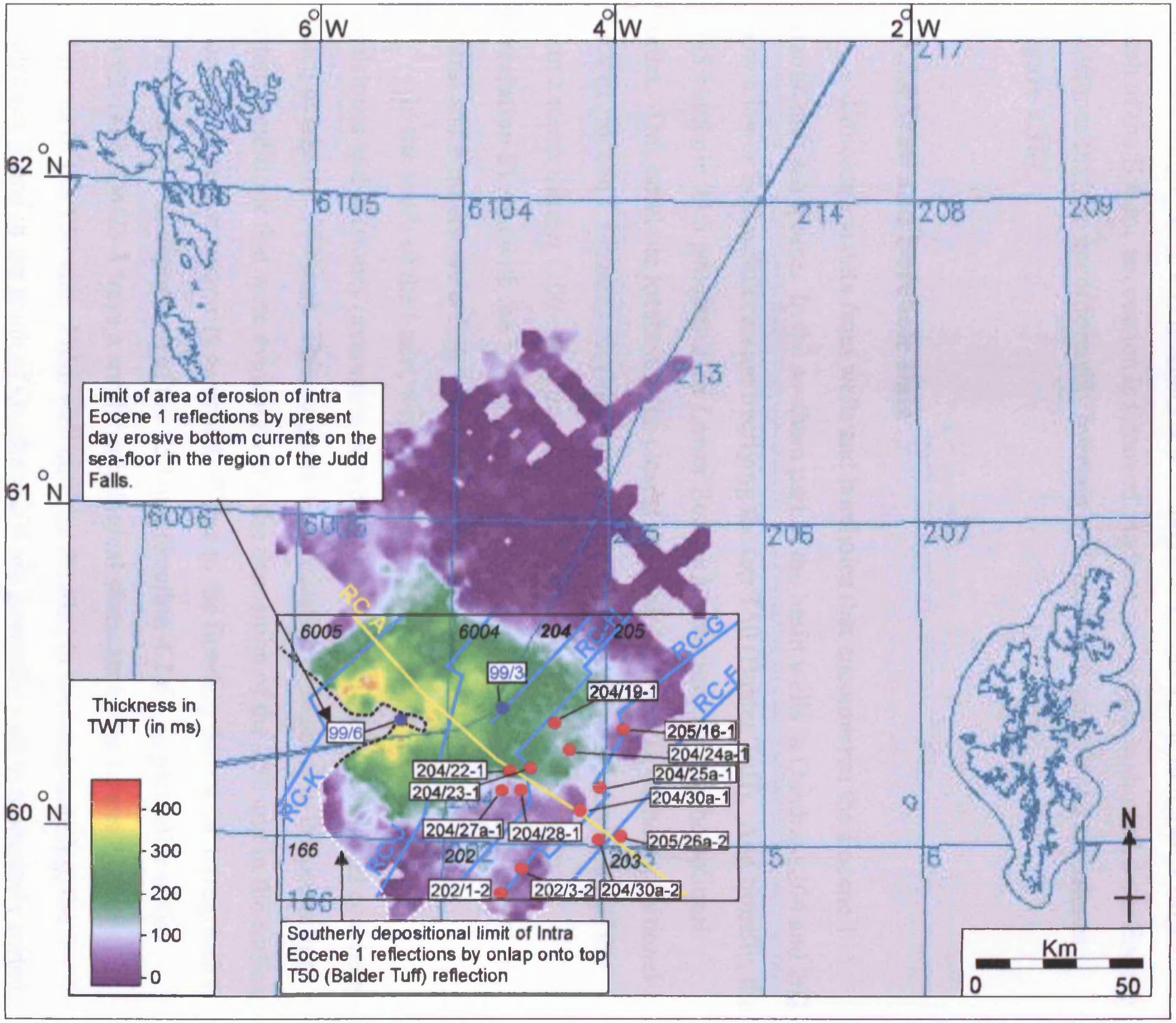


Figure 4.17. Isochron map showing the thickness (in TWTT) of an intra Eocene 1 unit which is mappable across the entire south of the basin. A roughly concentric thickened area is seen (in TWTT) which is centred on Quadrants 6005 and 6004 and 204 and similar to what is seen from the entire Eocene 1 unit (Figure 4.14). The intra Eocene 1 unit is seen to thin to the south where it onlaps onto the Top T50 (Balder Tuff) reflection in the north of Quadrant 202 and 166. In the area around the Judd Falls, there is a localised east - west trending area of erosion which is penetrated by BGS borehole 99/6. Here the Eocene unit is eroded almost down to the Top T50 (Balder Tuff) stratigraphic level in a small area on the apex of the Judd Anticline. However sediments of the Eocene 1 unit occur south of the Judd Anticline. The location of type wells, boreholes and regional correlations (RC) are highlighted.

major erosional scarps on the sea-floor. This occurs where the Faroe-Shetland Basin narrows and there is shallowing of water depths to cause significant amounts of erosion from fast flowing bottom currents that travel south and enter the Atlantic Ocean (Smallwood 2004). Further mapping of intra-Eocene 1 seismic reflections reveal that there is a thinning by onlap onto the top T50 (Balder Tuff) reflection towards the south and east, where the reflections finally pinch-out in the vicinity of the north part of Quadrants 202 and 166 (**Figure 4.16**). It can therefore be interpreted that the erosion of intra-Eocene strata is localised to the area around the Judd Deeps (Smallwood 2004). The BGS borehole 99/6 is located close to the Judd Deeps and is in the area where much of the Eocene succession is removed. Indeed, the borehole shows Lower Eocene mudstones that are unconformably overlain by Quaternary gravels and sandstones (**Figure 4.18**).

4.3.2.4 Well and Borehole Data

Lithological data from wells and boreholes that encountered the Eocene 1 seismic unit are sparse. In the southern part of the basin wells in Quadrant 204 and 202 show a lower Eocene succession overlying the top T50 (Balder Tuff). Additionally, the BGS borehole 99/3 penetrates the Lower Eocene succession close to the political border. This borehole location is the closest data point to the area of the depositional thick of Eocene 1 seismic unit (**Figure 4.14**). In the central and northern part of the basin a much thinner (200 – 400 ms) Eocene 1 seismic unit is found (**regional correlation J**). As with the T50 unit a thinned Eocene 1 succession is seen in the central and northern areas (e.g. wells 214/26-1 and 214/27-2).

In the south of the basin, wells in Quadrant 204 show thick interbedded sandstones and siltstones (averaging between 30 and 120 m) with occasional coal beds which overly the T50 unit. This suggests a continuation or renewal in the prevailing deltaic conditions that were evident during the deposition of the T50 unit in the earliest Eocene and were temporarily switched off due to the flooding during the transgression towards the end of deposition of the T50 unit (**Section 4.2.6**). In particular wells 204/22-1and 204/23-1 have a similar lithological character of the Lower Eocene sediments of the T50 unit. From well cuttings detailed in the composite log, the sandstones found in the south of Quadrant 204 are generally well to moderately sorted,

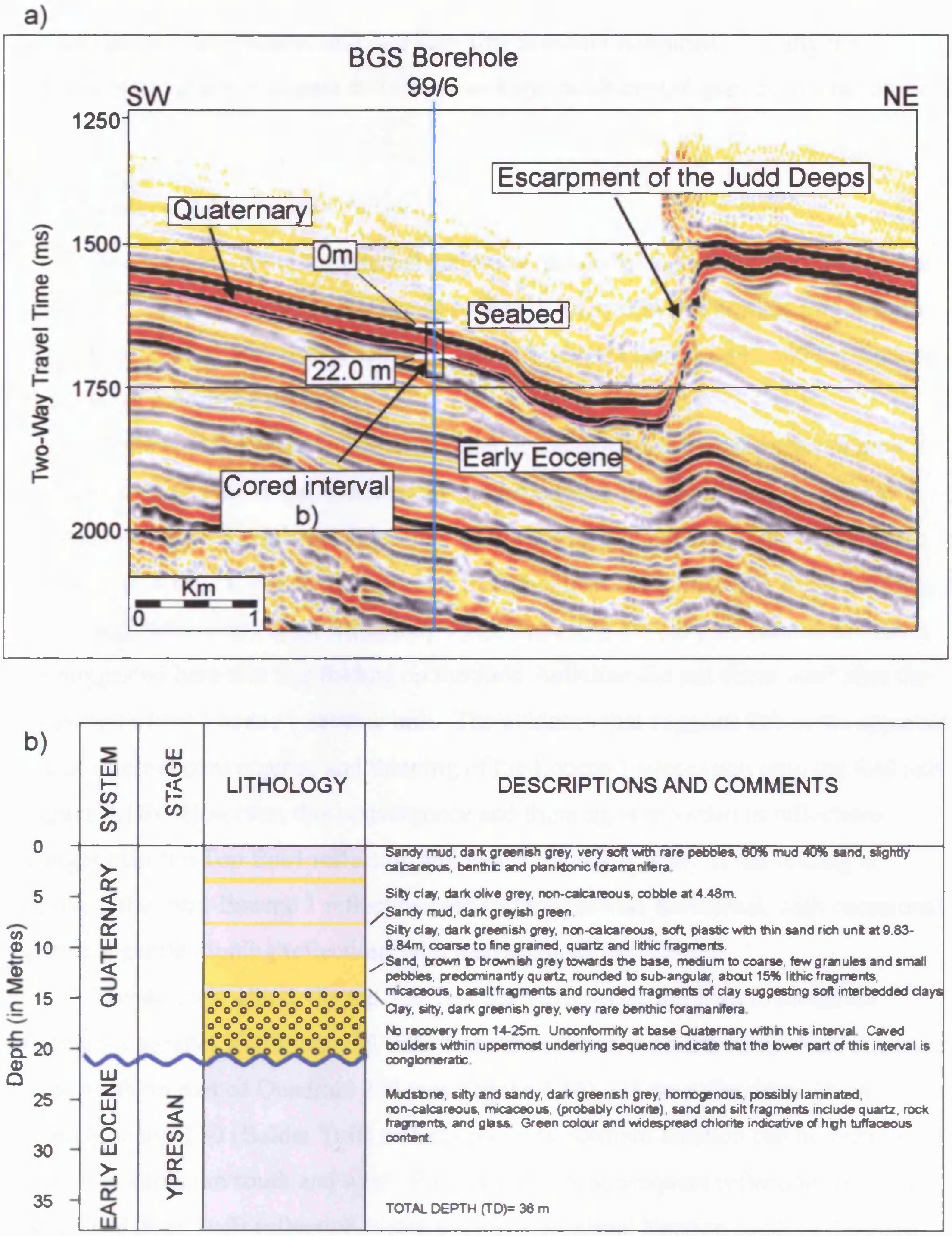


Figure 4.18. a) southwest - northeast trending 3-D seismic line across the location of BGS Borehole 99/6. This borehole is located in an area of significant erosion very close to the escarpments of the Judd Deeps (Smallwood 2004). The Eocene 1 unit is highly eroded and is overlain by Quaternary gravels and sandstones which have been cored. In this locality, there is no Middle or Upper Eocene succession preserved. b) lithological description of the cored interval of the Quaternary and Lower Eocene (Ypresian) sediments showing a major unconformity overlain by gravels and sandstones. The Ypresian sediments comprises of a mud-rich lithology which is occasionally sand and silt-rich. For location of seismic line in a) see inset in Figure 4.14. Nb: For the legend of the lithology shown in b) see Table 4.1

sub-angular to well-rounded and are very fine to medium grained. Locally the sandstones are glauconitic and micaceous and are translucent to grey brown in colour.

4.3.2.5 Internal Geometry

The section details the internal geometry and architecture of reflections within the Eocene 1 seismic unit. The majority of data presented in this section comes from the southern part of the basin where the Eocene 1 seismic unit is up to 800 ms thick on the southern margin of the basin.

In the area the Eocene 1 seismic reflection configurations are predominantly moderate to high amplitude, continuous and are parallel to sub-parallel. A general observation shows the seismic amplitude decreases further north into the basin centre (e.g. **Figure 4.16**). Close to the centre of the concentric thickened area there is a large anticlinal structure (the Judd Anticline) where the entire Tertiary succession is folded. It is suggested here that this folding on the Judd Anticline did not occur until after the deposition of the Eocene 1 seismic unit. The evidence that suggests this is the apparent lack of internal convergence and thinning of the Eocene 1 succession onto the fold axis (**Figure 4.16**). However, this convergence and thinning is recorded in reflections younger than the Top Eoc1 reflection and hence folding is later. If the folding is removed, the intra-Eocene 1 reflections are restored to near horizontal, with occasional relatively gentle dipping reflections which downlap to the north.

Immediately above the top T50 (Balder Tuff) reflection high to moderate continuous parallel to sub-parallel reflections are seen to the north of the Judd Anticline in the northern part of Quadrant 204 (see **Figure 4.16**). These reflections which parallel the top T50 (Balder Tuff) reflection in this northern location can be seen to onlap it towards the south and west. Further onlap of subsequent reflections onto the top T50 (Balder Tuff) reflection is also seen in a proximal location south of the Judd Anticline in the southern sector of Quadrant 204 and the northern sector of 202. There is a significant distance of 40 - 50 km between these two areas of onlap. This suggests a very thin package of sediment onlaps the top T50 (Balder Tuff) reflection and back-steps over an gently dipping shelf profile with a gradient of possibly less than 0.5°.

Furthermore, reflections above this onlap pattern show downlap terminations which have a low relief in the southern area (south of the Judd Anticline). This downlapping geometry continues to downlap onto the parallel to sub-parallel reflections overlying the top T50 (Balder Tuff) reflection progressively steps out northwards in a progradational stacking pattern. When the downlapping reflections are traced landward they become near horizontal and eventually thin and onlap onto the top T50 (Balder Tuff) reflector. Additionally, the reflections show an increase in seismic amplitude on the horizontal portion of the reflection. This progradational stacking pattern of reflections has been interpreted as a series of clinoforms which prograde north. The high amplitude horizontal part of the reflections is inferred to be the topset, with the dipping part interpreted to be the foreset part of the clinoform basinward of the clinoform break point (where a subtle change in angle is seen - **Figure 4.19**). The high seismic amplitude values seen in the topsets are interpreted to represent coal deposits found on the delta top.

The clinoform system is best imaged close to the fold hinge of the Judd Anticline (**Figure 4.16** and **Chapter 5,** e.g. **Figure 5.2**). Here, the clinoform break point is approximately 250 ms above the downlapping bottomsets (which in this shallow section equates to broadly the same in height in meters). Thus water depths of approximately 250 m are interpreted in the Early Eocene. The predominately progradational stacking pattern of the clinoform system is evidenced by the progressive basinward movement of high amplitude topset coals during deposition of the Eocene 1 unit (**Figure 4.16** and **Figure 5.2**). As discussed above (**Section 4.2**) the top of the T50 unit is capped by tuffs, coals, sandstones and mudstones. In this southern area, coals are seen at the very top of the T50 unit in wells 204/24-1a (see **Figure 4.5**). High amplitude seismic reflections can be seen close to the top T50 (Balder Tuff) reflection in **Figure 4.19**, that then back-step to the south. Younger stratigraphy shows a re-establishment of high amplitude topsets further north in the basin in a position close to the original coals at the top of the T50 unit (**Figure 4.19**).

It is interpreted that immediately after deposition of the T50 unit, flooding continued and pushed facies belts back landward towards the south and southeast. The result is the development of a retrogradational stacking pattern representing a transgression during relative sea-level rise. Following this relative sea-level rise deltas prograded out across the low angle shelf and facies belt began to move northwards out into the basin with coals found on the delta tops reaching positions that were previously seen at the top of the T50 unit. The delta prograded into water depths of approximately 250 m.

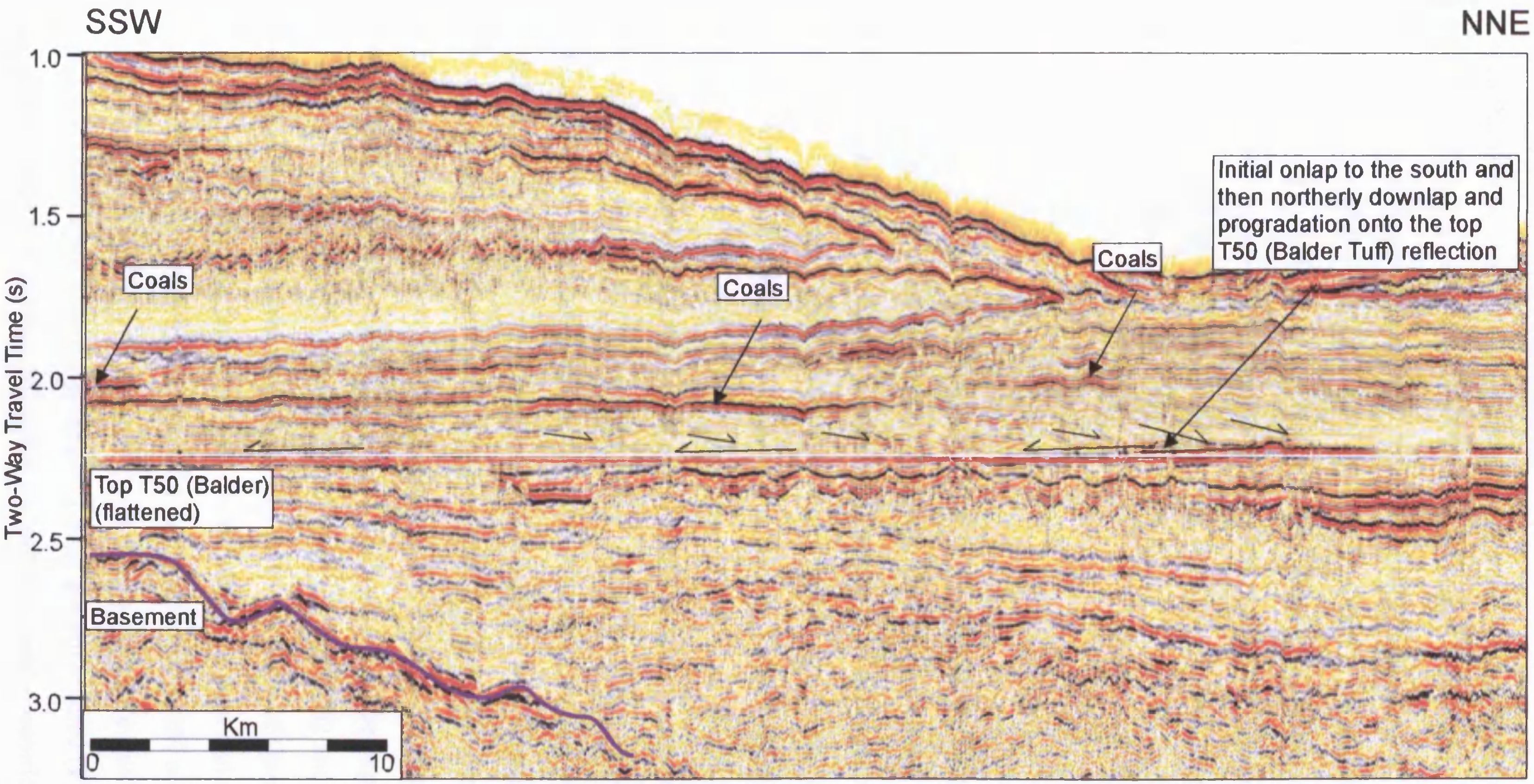
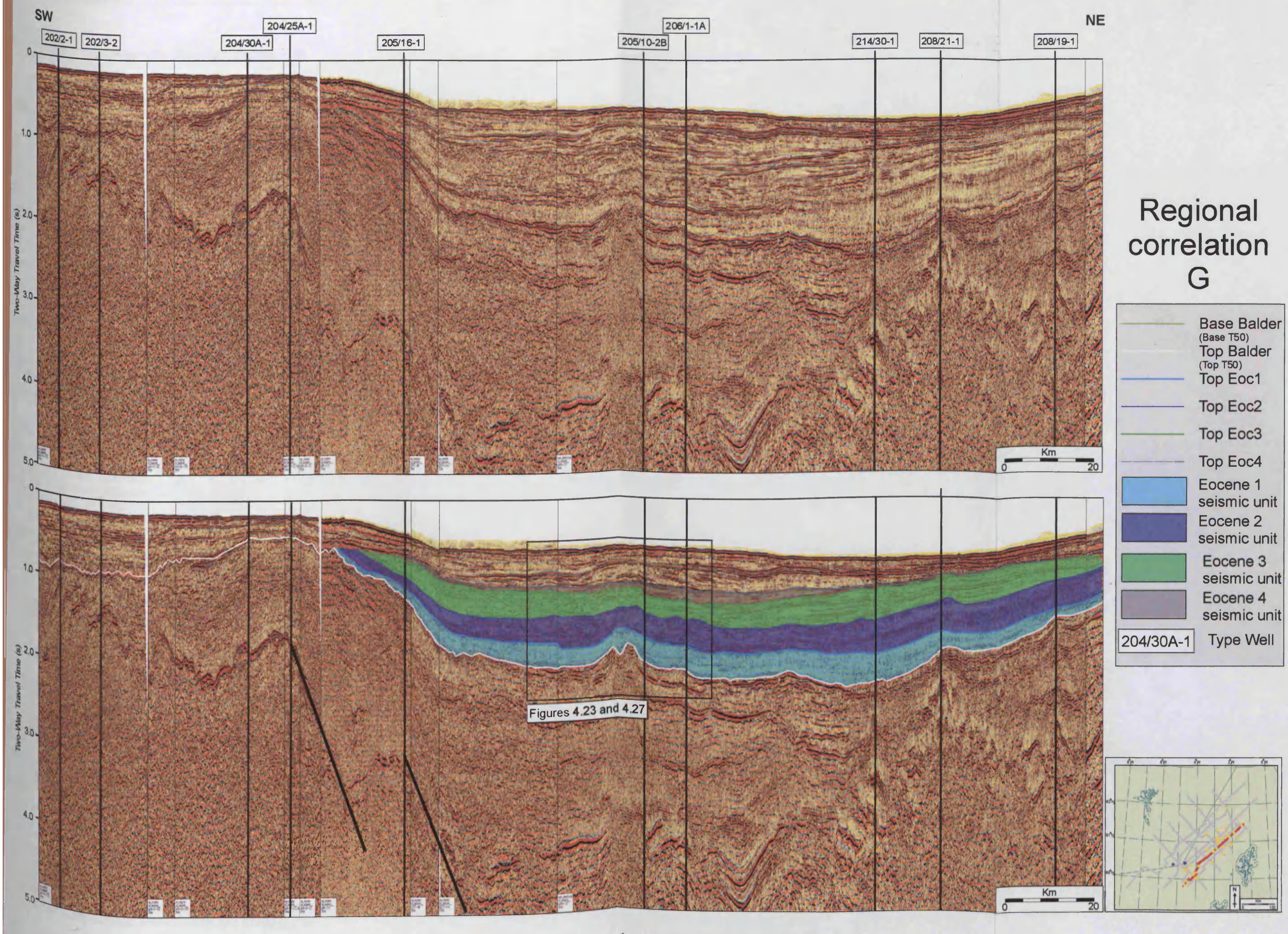


Figure 4.19. South southwest - north northeast trending 2-D seismic line flattened on the top T50 (Balder Tuff) reflection showing the northerly migration of the interpreted high amplitude topsets. These are interpreted as coals and can be seen in the T50 unit (Balder Fm) level in a basin-ward position, and then during subsequent flooding the coals are found to the south and southwest, where they appear as near horizontal high amplitude reflections. The flooding back is believed to have taken place on a very low angle gradient where facies belts can be pushed back landward significant horizontal distances with just a small amount of relative sea-level rise. This accounts for a very thin onlapping (transgressive) package which directly overlies the top T50 (Balder Tuff) reflection. The sedimentary response to this flooding back is north and northeasterly progradation of clinoform systems into the basin which attain heights of up to 250 ms. For location of the seismic line see Figure 4.14.

In the central and northern parts of the basin the Eocene 1 seismic unit appears as a low amplitude, semi-continuous to discontinuous unit of reflection configurations. Indeed the Top Eoc1 reflection is often "phantomed" across significant distances within the basin axis. However, on many 2-D seismic lines the Top Eoc1 reflection shows very distinct local lateral variations. On the Shetland margin locally (for example in the north of Quadrants 205 and 206), the Top Eoc1 reflection manifests itself as a very high amplitude reflection laterally separated by areas of moderate to low seismic amplitude (see **regional correlation G**). On this margin the Eocene 1 unit appears relatively featureless with chaotic reflections, with the top and base the only reflections with any continuity.

4.3.2.6 Discussion

It has been shown that the main depocentre at the time of Eocene 1 deposition was located in the south of the basin over Quadrants 204 and 6004, straddling the political boundary between the UK and the Faroes. This depocentre is broadly concentric in geometry and accumulated sediment during a major phase of delta progradation into the basin. Wells located in the north of Quadrant 202 and the south and central parts of Quadrant 204 show a clear marginal deltaic facies of interbedded sandstones, siltstones mudstones and coals. Furthermore, there is supporting evidence showing a high frequency of near horizontal, high to moderate amplitude, continuous parallel reflection configurations and these have been interpreted to represent topsets to the delta system. These near horizontal topsets can be traced laterally into gently dipping clinoforms. These topsets are interpreted to represent the proximal part of a delta top which experiences subtle variations in depositional environment (**Figures 4.19** and **4.20**). Locally, coals and delta top channels would be expected in this environment at this proximal area in the south of Quadrant 202 and north of Quadrant 204. This delta top feeds the delta front located to the north of Quadrant 204 and 6004. Progradational clinoform packages are seen locally in the area and reach maximum heights of 250 ms suggesting water depths of approximately 250 metres. These water depths suggest a shallow shelf area was present over the entire southern part of the basin, into which a northwardly prograding delta system entered (**Figure 4.20**). The Orkney landmass and Scottish Massif to the south probably provided the main source


SW
NE
202/2-1
202/3-2
204/30A-1
204/25A-1
205/16-1
205/10-2B
206/1-1A
214/30-1
208/21-1
208/19-1
Two-Way Travel Time (s)
Km
0
20
Figures 4.23 and 4.27
Regional correlation G
Base Balder
(Base T50)
Top Balder
(Top T50)
Top Eoc1
Top Eoc2
Top Eoc3
Top Eoc4
Eocene 1 seismic unit
Eocene 2 seismic unit
Eocene 3 seismic unit
Eocene 4 seismic unit
204/30A-1
Type Well

Early Eocene Palaeogeography (Eocene 1 seismic unit)

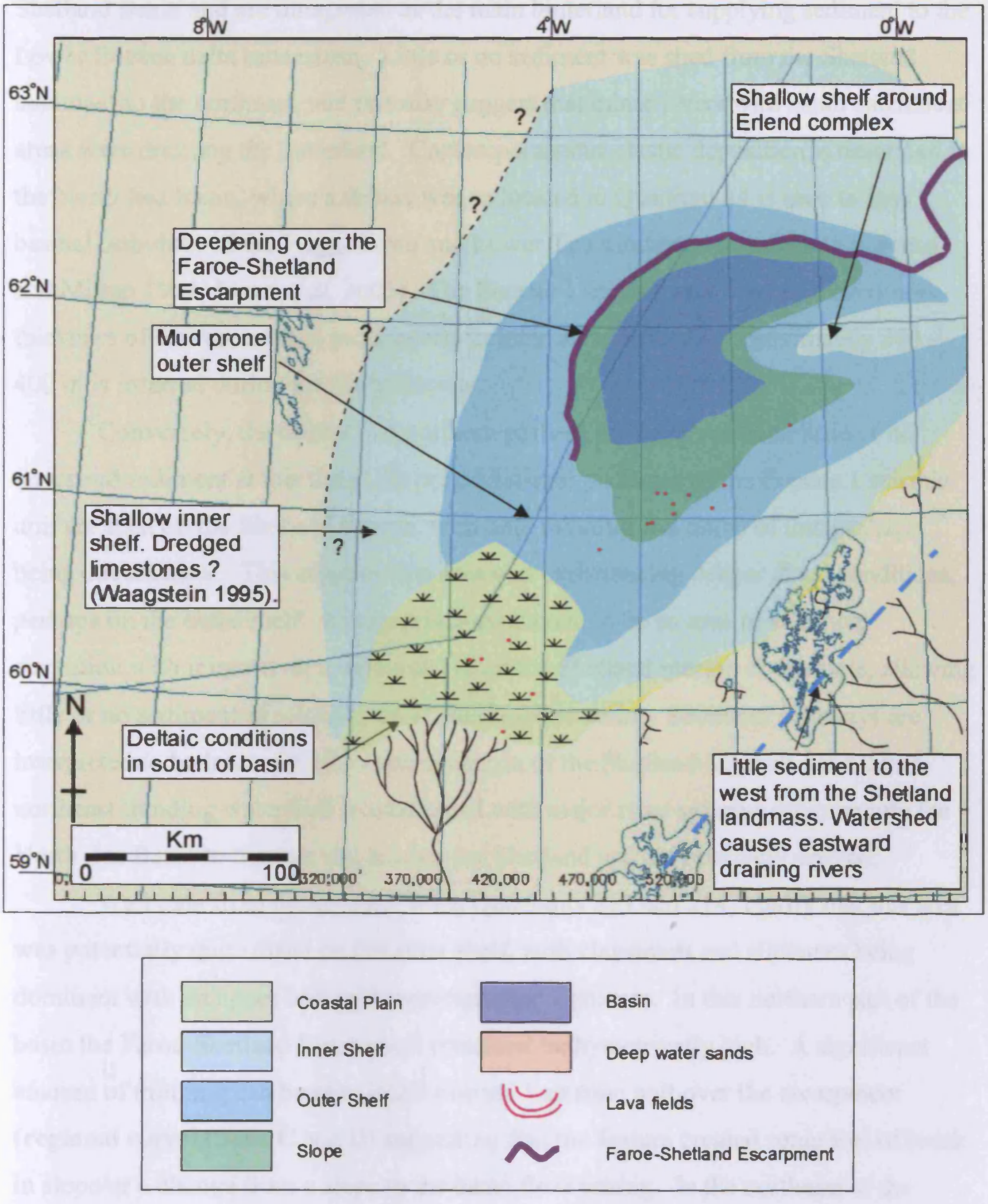


Figure 4.20. Ypresian palaeogeographic map depicting gross depositional environments during deposition of the Eocene 1 unit. The environment in the south of the basin is a dominated by deltaic system where a northerly prograding wedge is seen which is sourced form the Orkney landmass and Scottish Massif. To the north of the basin, a deep marine realm was maintained and clastic sedimentation was starved from this area. At this time, the North Sea Basin was receiving sediment (e.g. Frigg, Skroo and Tay sands) and thus a watershed is interpreted with a predominantly eastward drainage network. A mud-rich outer shelf is interpreted north of the large southern delta though a shallow shelf may have occurred towards the Faroe-Platform. The Erlend volcanic complex remained topographically high and the Faroe-Shetland Escarpment and Corona Ridge had a bathymetric expression and thus influenced sediment dispersal patterns. For full key to depositional environments see Enclosure M.

of sediment during this time. These two areas are proximal to the south of the Faroe-Shetland Basin and are interpreted as the main hinterland for supplying sediment to the Lower Eocene delta succession. Little or no sediment was shed from the Shetland landmass to the northeast, and this may suggest that minor rivers with small catchment areas were draining the hinterland. Contemporaneous clastic deposition is described in the North Sea Basin, where a deltaic wedge located in Quadrant 14 is seen to feed basinal turbidites of the Frigg, Skroo and Lower Tay sandstone members (e.g. Jones and Milton 1994, Jones *et al.* 2003). The Eocene 1 seismic unit reaches a maximum thickness of 800 m and thus pronounced tectonic subsidence of approximately 300 – 400 m is inferred during the Early Eocene.

Conversely, the central and northern parts of the basin received little or no terrestrial sediment at this time. No progradational packages of the Eocene 1 seismic unit are seen on the Shetland margin, with only thinning and onlap of this package being documented. This suggests this area was experiencing deeper distal conditions, perhaps on the outer shelf. It is therefore concluded to be an area of sediment starvation with minor river systems active on the Shetland margin at this time, allowing little or no sediment to reach the outer shelf marine realm. Sediment pathways are interpreted to be located on the eastern margin of the Shetland landmass and thus a northeast trending watershed is anticipated with major river systems draining into the North Sea Basin to the east and leaving the Shetland margin relatively starved.

Well data from north of 61° N (in Quadrants 213 and 214) clarify that this area was potentially quite distal on the outer shelf, with claystones and siltstones being dominant with an upper bathyal biostratigraphic signature. In this northern part of the basin the Faroe-Shetland Escarpment remained bathymetrically high. A significant amount of thinning can be seen in the Eocene 1 seismic unit over the escarpment (**regional correlations** C and **D**) suggesting that the feature created some sort of break in slope or a change from a slope to the basin floor setting. In the northeast of the basin, the Erlend complex remained topographically high with thinning of the Eocene 1 seismic unit occurring over the complex (**Figure 4.20**).

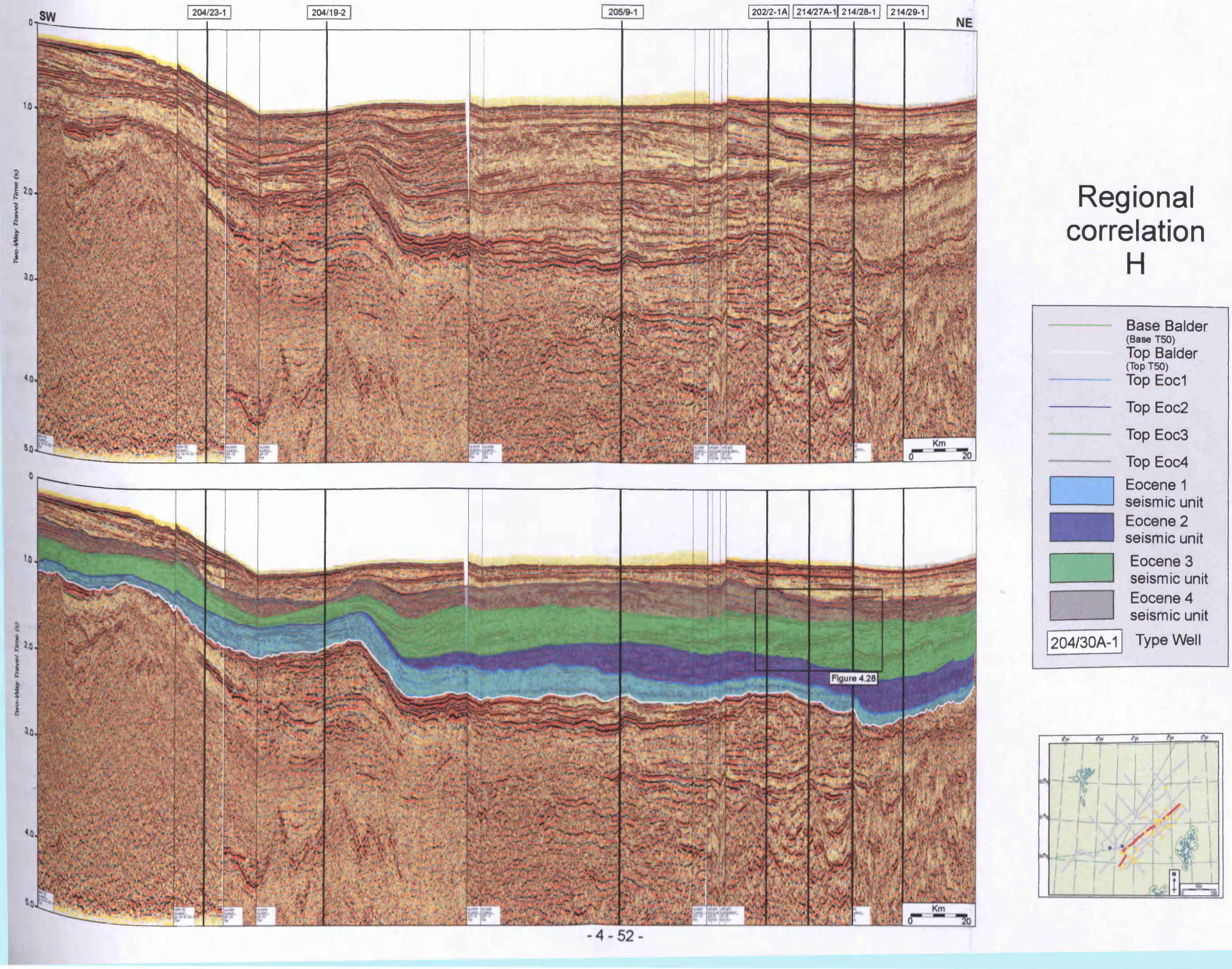
4.3.3 Eocene 2 Seismic Unit

4.3.3.1 Introduction

This section describes the distribution, the seismic character and lithology of the second seismic unit of the Eocene succession. It goes on to discuss the depositional setting of the basin during the latest Early Eocene and the earliest Middle Eocene.

4.3.3.2 Upper and Lower Boundaries

The lower boundary of this seismic unit is the Top Eoc1 reflection. As has been discussed in the previous section (**Section 4.3.2**), this is a high amplitude continuous reflection in the south of the basin, but becomes discontinuous and often chaotic towards the north. The Eocene 2 seismic unit is bounded at its upper surface by the reflection named here as Top Eoc2 (dark blue horizon on all seismic lines). This reflection is similarly defined in the south of the basin, (in the area of Quadrant 204) where it appears as a continuous high to moderate amplitude parallel reflection. This upper boundary is confidently picked in this southern area on the northern flank of the Judd Anticline where it is seen to truncate and cut out the Top Eoc1 reflection. Hence, the Top Eoc2 reflection becomes the top of both the Eocene 1 and 2 seismic units south of this truncation (see **Figure 4.14**). The confidence in the pick of Top Eoc2 reflection becomes significantly lower away from the southern area. As with the Top Eoc1 reflection, the Top Eoc2 reflection becomes semi-continuous and chaotic towards the centre and north of the basin (see **regional correlation J**). Towards the north the reflection and unit as a whole is pervasively cut by small faults, which indeed pervasively cut the entire Eocene succession. From the observations of erosion into the underlying succession this upper boundary is interpreted as a sequence boundary on the southern margin of the basin and is interpreted as a correlative conformity towards the north in the basin axis. When traced onto the Shetland margin the boundary appears to be relatively conformable with the older Eocene 1 seismic unit. However, on parts of the Shetland margin this surface can be traced up into a clinoform system that is sourced from the southeast (**regional correlation D**). Reflection configurations can be clearly seen to downlap onto the Top Eoc2 reflection. The downlap direction is to the southwest and this is best shown in **regional correlations G** and **H**.


SW
NE
204/23-1
204/19-2
205/9-1
202/2-1A
214/27A-1
214/28-1
214/29-1
Two-Way Travel Time (s)
0
1.0
2.0
3.0
4.0
5.0
Km
0
20
Figure 4.28
Regional correlation H
Base Balder
(Base T50)
Top Balder
(Top T50)
Top Eoc1
Top Eoc2
Top Eoc3
Top Eoc4
Eocene 1 seismic unit
Eocene 2 seismic unit
Eocene 3 seismic unit
Eocene 4 seismic unit
204/30A-1
Type Well

Towards the present day Faroe Platform, the Top Eoc2 unit is relatively conformable with the T50 and Eocene 1 units. Furthermore it shows a progressive thickening to the northeast as shown by **regional correlation K**. Close to the Faroe Islands, the Top Eoc2 reflection is eroded by a later (post-Eocene) unconformity, and overlain by a younger prograding system. Similarly, the entire Eocene succession is eroded and truncated on the present day Faroe Platform and this is best illustrated by **regional correlation C**.

4.3.3.3 Distribution of Eocene 2

Lower and Middle Eocene sediments of the Eocene 2 seismic unit cover a significant part of the basin. However, this seismic unit is not found in the southern part of Quadrants 205 and 204 and further west in the south of Quadrant 6006. In this southern area, the Top Eoc2 reflection becomes unconformable and down-cuts into older strata of Eocene 1 age. The preserved time thickness of Eocene 2 seismic unit varies considerably in the basin from zero to approximately 750 ms, equating to about 700 m (**Figure 4.21**). The thickest part of the succession is located in the far northeastern end of the basin in the area of northern Quadrant 208 and central Quadrant 217. This area exhibits preserved thickness of greater than approximately 650 m which dramatically thins to the east towards the Erlend complex where little or no Eocene 2 succession is deposited, suggesting that this area was still topographically high, possibly emergent at this time. A broadly north - south trending oval shaped area is represented by the thickest area over Quadrants 217 and 208 and may extend towards the northeast into the Møre Basin and eastwards into the Northern North Sea (**Figure 4.21**).

The main areas of preserved sediment thickness at the time of deposition of the Eocene 2 seismic unit can be broadly split into two parts. In the south and west of the basin, the succession is predominantly thin (between 0 and 200 ms – highlighted in purple colours on **Figure 4.21**). In this area of the basin the Eocene 2 unit thins onto the Top Eoc1 reflection in a northwesterly direction as can be seen in **regional correlation B** towards the Faroe Platform. On the platform the Eocene 2 unit is thin (less than 200 ms) though isolated thicker areas (up to 300 ms) are visible. One small circular area is located in the central part of Quadrant 6006 where the unit is

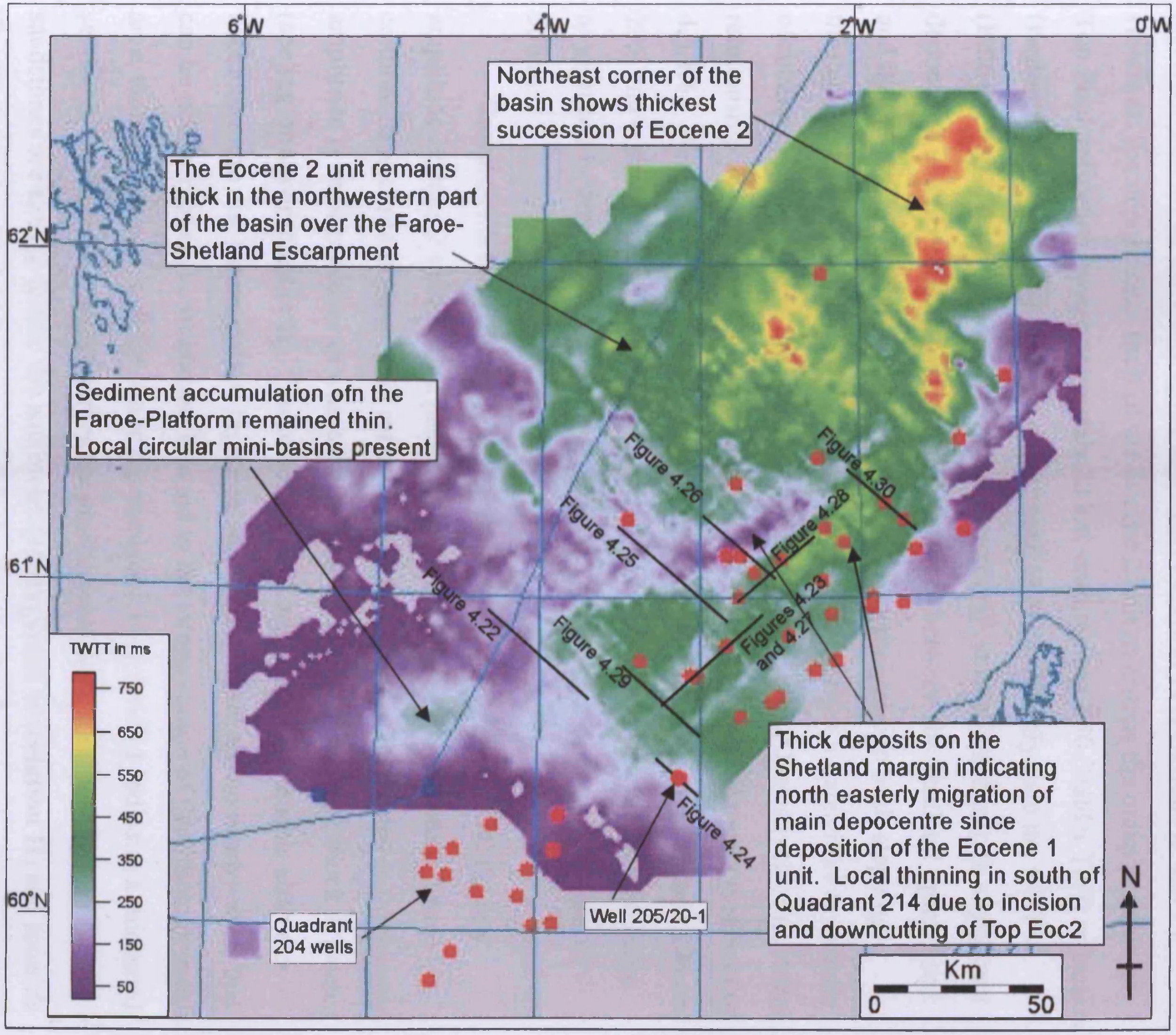


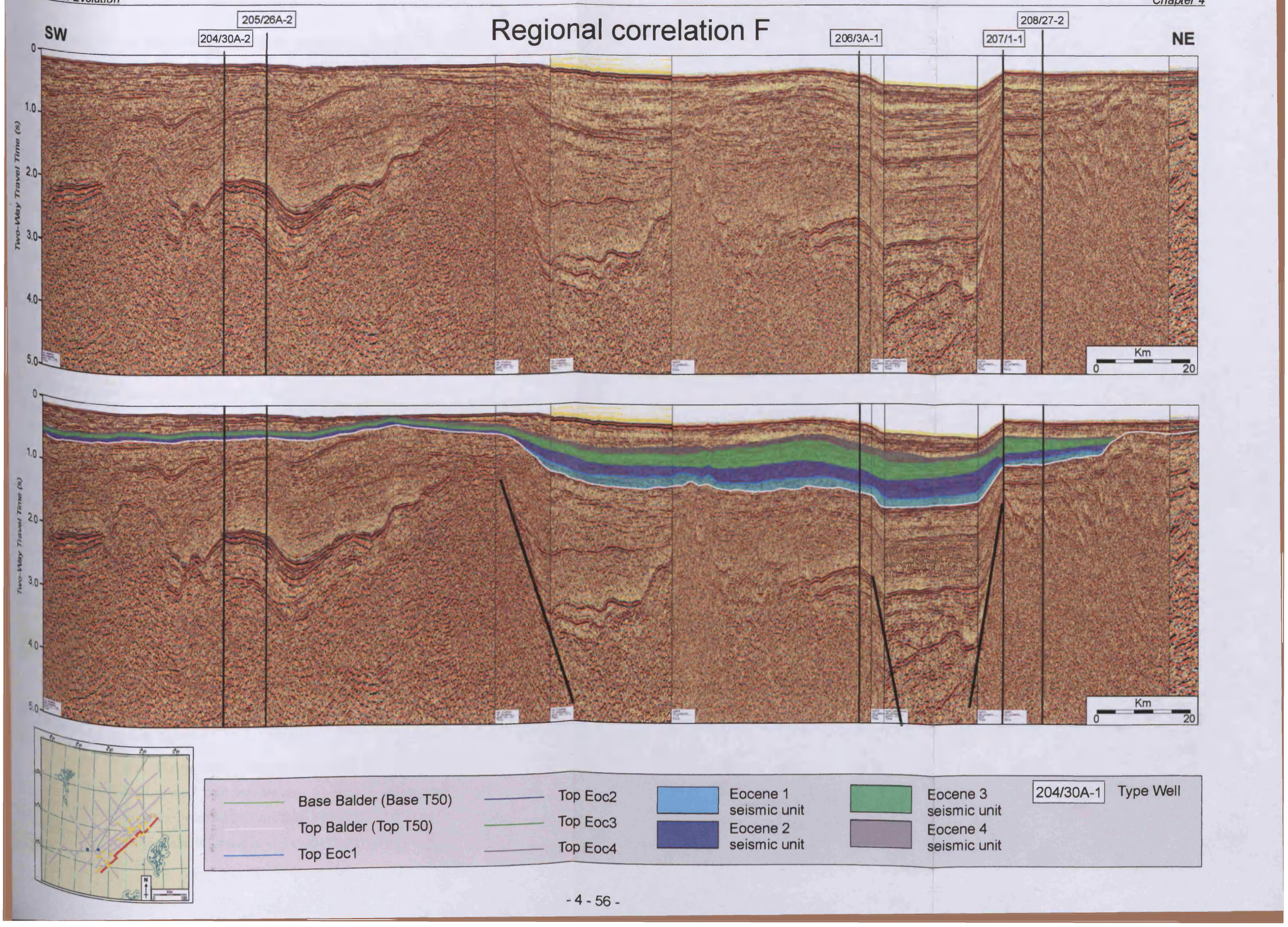
Figure 4.21. Isochron map showing the thickness (in time) of the Eocene 2 seismic unit (between the Top Eoc1 and Top Eoc2 reflections. The thickest section is seen again in the northeast of the basin. The previously thick concentric area of deposition that was located in the south of the basin in Eocene 1 times has been replaced by a thin area in this part of the basin. However, the northern part of the basin shows a preserved thickness of up to 800 ms and implies that this area either received more sediment, or is less eroded than the southern area. The deposition pattern is believed to represent a shift in the sediment source areas from the south to the northeast towards the Shetland margin). There is a northeasterly protruding spur in the basin located in the south of Quadrant 214 and this feature is explained in the text. This feature is localised and a thickening is seen both landward (to the southeast) and basin-ward (to the northwest) of this spur. Locations of all seismic lines referring to the Eocene 2 unit and type wells discussed in this chapter are shown on the figure.

approximately 250 - 300 ms, is located on the northern flank of the Judd Anticline and is approximately 30 km in diameter.

Towards the north and east of the basin there is increased accumulation of sediments (approximately 300 – 750 ms) which are highlighted on **Figure 4.21** as green and red colours. On the southeastern margin of the basin (in the northern part of Quadrant 205 and 206) the Eocene 2 unit thickens eastwards away from the Quadrants 6005, 6004 and 204. The source for this sediment remains debatable, although a thickened area towards the Shetland margin may suggest a southeasterly provenance (see **Section 4.3.3.6**). This unit looks relatively conformable with the underlying Eocene 1 unit along the strike of the basin (see **regional correlations G** and **F**). However, the depositional limit of the Eocene 2 unit oversteps the onlap position of Top Eoc1 reflection by approximately 20 km onto the top T50 (Balder Tuff) reflection (**regional correlation B**). Here, it can be seen to onlap directly onto the top T50 (Balder Tuff) reflection in the area of the remnant West Shetland Basin (a significant depocentre active in Jurassic and Cretaceous times between the Shetland Spine Fault and the Rona/Clair Ridge - see **Section 2.2.3**). This progressive back-stepping throughout the Early and Middle Eocene may be evidence for a transgressive episode of relative sea-level along the northeastern Shetland margin. Alternatively, this stratal relationship could be explained by onlap and fill by deep water sediments (see **Section 4.3.3.6**). Biostratigraphic reports from wells in the northern part of Quadrants 206 and 205, and wells further to the northeast favour an outer shelf environment with some oxygenated bottom waters, suggesting water depths interpreted to be between 200 – 500 m.

Towards the basin centre (in the north of Quadrant 205 and 204) the low amplitude Eocene 2 seismic unit passes laterally into moderate amplitude, semi-continuous to continuous parallel reflection configurations. The change in the seismic amplitude is apparent close to where the hinge at the Rona Ridge fault block is located (see **regional correlation B**). Basinward of the Rona Ridge, the seismic character becomes more discontinuous and higher in amplitude. There are no nearby wells that can be tied to **regional correlation B** to aid in the interpretation of the lithology in this area, though approximately 20 km to the northeast, well 205/9-1 indicates a thickened section of shallow marine sandstones are predominant in this interval. These sandstones are seen to thin to the southwest (see **regional correlation H**) and towards Shetland margin. The thickness of the Eocene 2 unit remains relatively constant in

Regional correlation F



Quadrant 204 and 205 and this is further highlighted from **regional correlation B**, where only a local area of thinning occurs (see **Figure 4.21**) before the package re-thickens in the basin centre and thins towards the Faroe Platform. The moderate amplitude, parallel refection configurations found basinward of the low amplitude area (on **regional correlation B**) may represent an interbedded shelf unit of sandstones and siltstones to give the package the distinct higher frequency seismic character. Further basinward, the amplitude response of the seismic unit becomes even lower in a zone which is relatively close to the position of the present day shelf edge. In this area the base of the unit shows a distinctive high amplitude, wavy reflection pattern highlighted in **Figure 4.22.**

In the southern part of Quadrant 214, a narrow (15 - 20 km) northeast trending spur can be seen which runs parallel to the shelf edge of the Shetland margin (**Figure 4.21**). This spur which has a length of approximately 50 - 60 km highlights an area where there is a distinct thinning (less than 250 ms) of the Eocene 2 unit. This area of thinning is bounded by thicker areas (250 – 550 ms) both proximally to the southeast and into the basin axis to the northwest.

Along the entire southeast Shetland margin, there is a general thickening of the unit to the northeast into the region immediately west of the Erlend complex in northern Quadrants 208 and 209. Generally, the unit has a higher degree of continuity on this part of the margin than the lower Eocene 1 unit which shows relatively low amplitude, semi-continuous to discontinuous seismic reflection configurations below the strong amplitude reflector of Top Eoc1 (**Figure 4.23**). Conversely, there is significant local lateral variation seen within the Eocene 2 seismic unit which is more continuous either side of the Flett Ridge structure (**Figure 4.23**). Additionally, there seems to be a slight thickening of the Eocene 2 unit to the northeast which is evident from the thickness map seen in **Figure 4.21**.

4.3.3.4 Well and Borehole Data

Biostratigraphic and lithological data from wells that encountered a Lower to Middle Eocene succession (Eocene 2 unit) are also rare. In addition to the type wells in the southern part of the basin (in Quadrant 204 and 202) further type wells on the Shetland margin also provide useful information and could be tied to the seismic data.

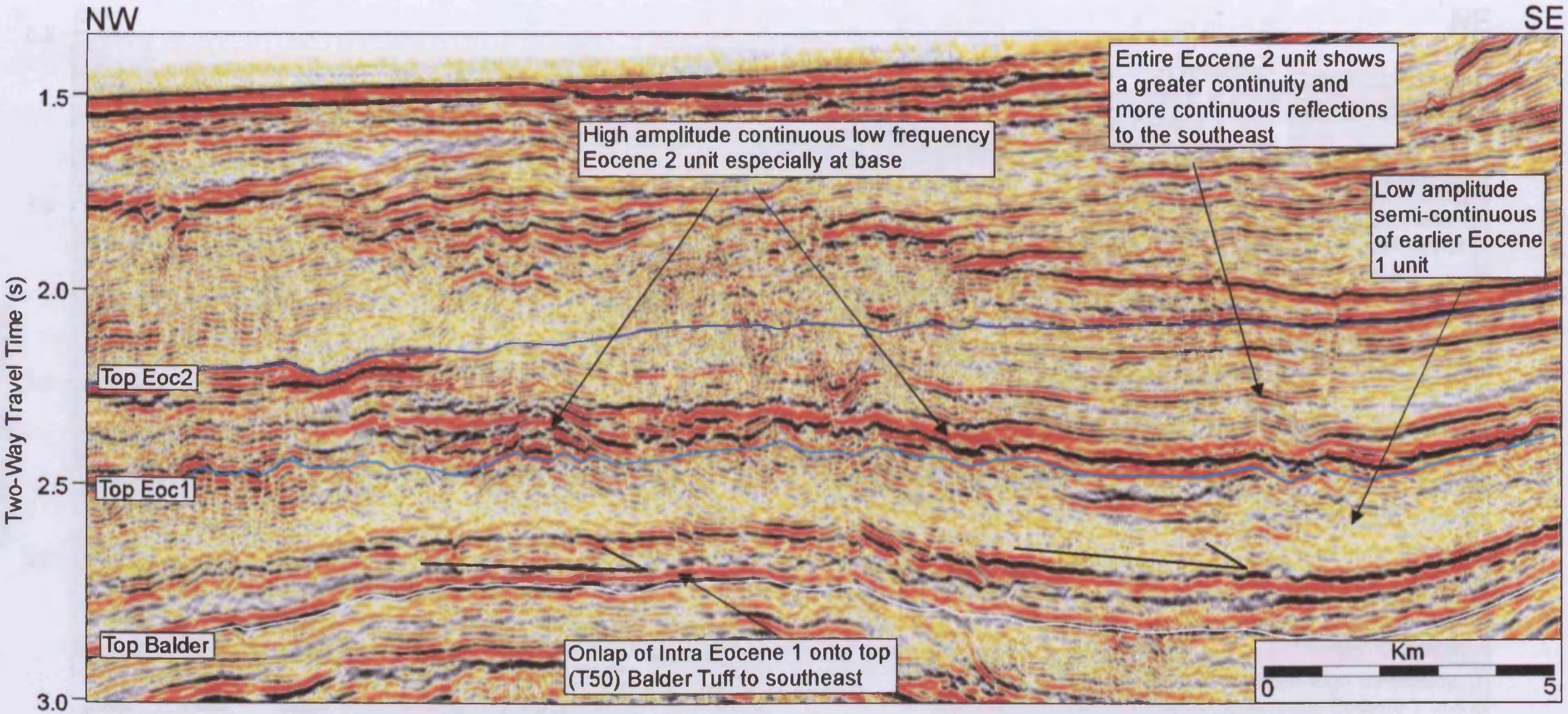


Figure 4.22. Southeast - northwest trending 2-D seismic line, showing a zoomed in portion of regional correlation B (on the southeastern slope of the basin). This figure shows the difference in seismic character between the Eocene 1 and 2 units, where the Eocene 2 unit has a higher more continuous seismic character to the low amplitude Eocene 1 unit. High amplitude mounded seismic reflections are located near the base of the Eocene 2 unit and lie directly over a significantly lower amplitude area of the Eocene 1 unit. This differs from the more proximal seismic line (Figure 4.24) which shows a low amplitude Eocene 2 unit and a higher amplitude Eocene 1 unit which has coals and limestones lying immediately above the top T50 (Balder Tuff) reflection within a massive sandstone. See Figure 4.21 for line location and also refer to regional correlation B.

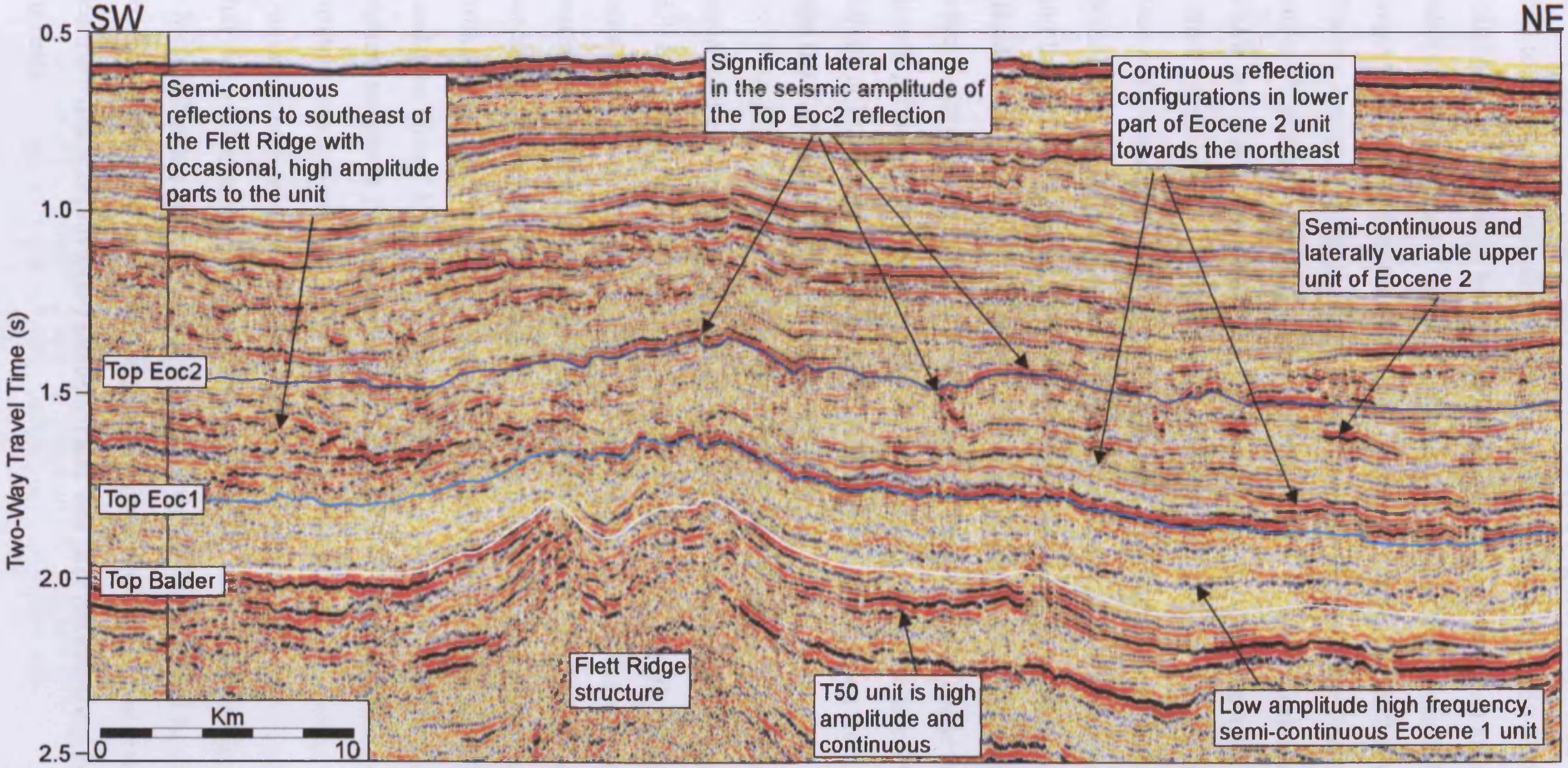


Figure 4.23. Southwest - northeast trending 2-D seismic line across the Flett Ridge structure (in the north of Quadrant 205) comparing the seismic character of the Eocene 1 and Eocene 2 units. In this part of the basin the Eocene 2 unit is shows higher amplitude seismic reflection configurations which are more continuous. There is significant lateral variation in the Eocene 2 unit seen both at its upper bounding surface (Top Eoc2) and in its internal geometry. High amplitude continuous reflections occur locally along the Top Eoc2 reflection and are 2-5 km in length and pass laterally into a moderate or low amplitude reflection. A more discontinuous area of seismic reflections is seen over the Flett Ridge. In comparison the Eocene 1 unit exhibits a low amplitude, homogenous semi-continuous seismic character. For location of line see Figure 4.21 and regional correlation G.

Located on the Shetland margin and close to the crest of the Rona Ridge, well 205/20-1 intersects **regional correlation B** (see **Figure 4.21**). This well exhibits a very fine to coarse sand-rich Eocene section that overlies the top T50 (Balder Tuff) reflection and is seen to fine upwards. This well is located in an area where there is a relatively thin Eocene 1 unit and a thicker Eocene 2 unit and was drilled for a deep target in 1972. Information from the biostratigraphic report for this well shows a crude Eocene - recent age for the sand-rich succession, with a Lower Eocene assemblage of pollens and dinoflagellates predominating. However, the occurrence of *Wetzeliella ovalis* and *homoromorpha* may suggest part of the sand-rich section to have a tentative Middle Eocene age. Localised traces of limestone and lignite beds are found near the base of the sandstone and the biostratigraphic assemblage suggests a marginal marine, deltaic setting with restricted marine conditions and terrestrial influences. These coals and limestones can be interpreted on 2-D seismic data and lie above the top T50 (Balder Tuff) reflection and below a low amplitude zone of semi-continuous reflections which is interpreted as the sand-rich interval (**Figure 4.24**). The absence of dinoflagellates in the underlying Palaeocene section indicates an exclusive continental, deltaic or littoral environment and further suggests a deepening from latest Palaeocene into the Early and Middle Eocene on this margin.

Additional well data on the margin of the basin is scarce, but close to the basin centre wells in the southern part of Quadrant 214 show a predominantly clay and mud-rich facies (e.g. in wells 214/28-1, 214/27-1 and 214/17-2 with occasional sand-rich units between 20 and 100 m thick with traces of limestone (seen in well 214/29-1). A marine, outer shelf to upper bathyal signature is recorded in the Lower Eocene succession of well 214/29-1 with the biostratigraphic report only focussing on strata from 1690 m, and the top of the Lower Eocene not seen. However, the basal thin sandstone (found between 1700 and 1720 m) is given a Lower Eocene age by the highest sample at 1690 m in the above claystone. Furthermore, abundant reworked Jurassic taxa are recovered throughout the Lower Eocene succession including *Gonyaulacysta jurassica* at 1690 m and *Nannoceratopsis pellucida* at 1810 m. The fact that this well is unique in the appearance of small isolated sandstone units in the Lower – Middle Eocene succession and is the only well with reworked Jurassic taxa may suggest that there is a link between the two lines of evidence and that this well received sediment that was previously deposited and was later reworked e.g. by turbidity currents.

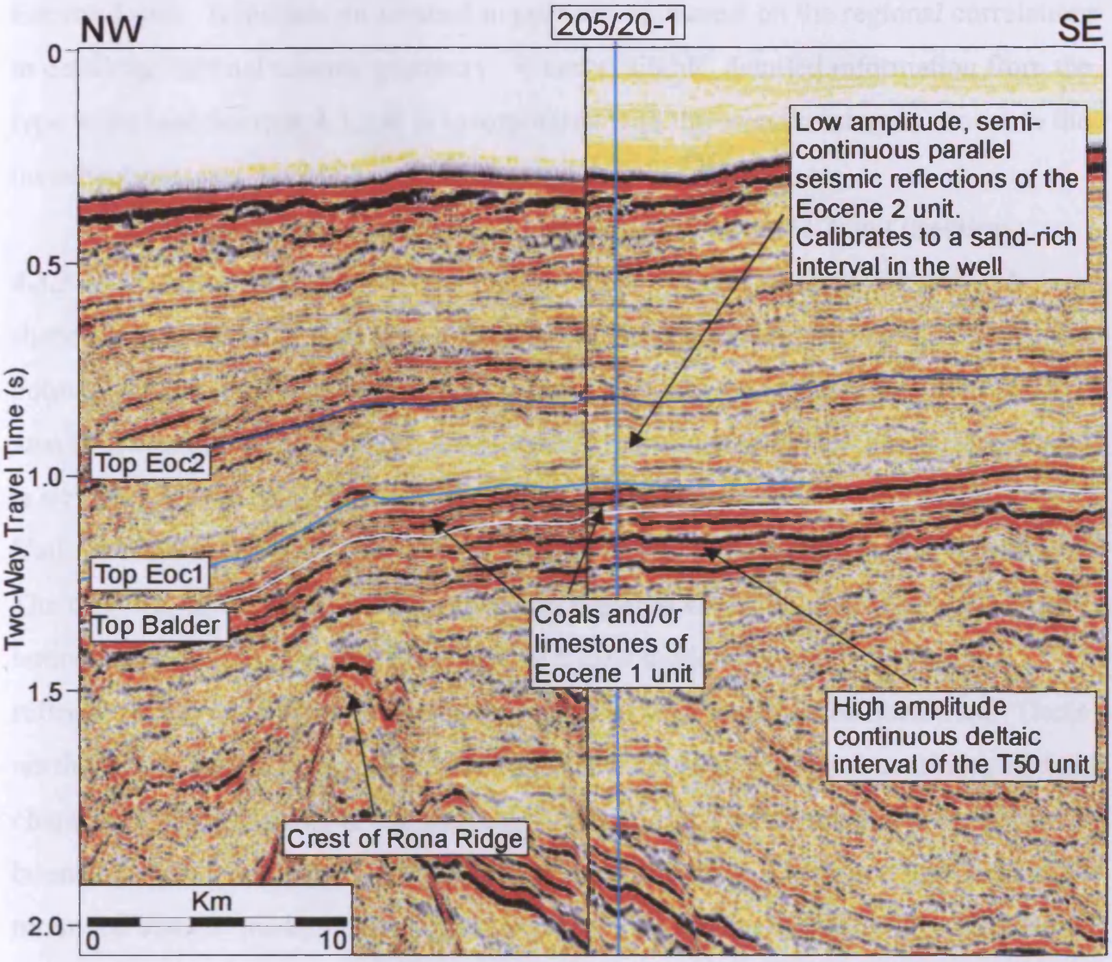


Figure 4.24. Southeast - northwest trending 2-D seismic line showing a zoomed in portion of regional correlation B highlighting the difference in seismic character between the Eocene 1 and 2 units. On the Shetland margin the Eocene 1 unit is thin and in well 205/20-1 the composite log records a sand-rich interval with coals and limestones is seen immediately overlying the top T50 (Balder Tuff) reflection. The overlying Eocene 2 unit is thicker and exhibits a low amplitude seismic character reflecting the massive sandstone interval seen in the well. The lithology encountered in the well throughout the Eocene succession is entirely sand-dominated and becomes finer upwards. The accompanying biostratigraphic report suggests a marginal marine, deltaic setting with restricted marine conditions and terrestrial influences. Early Eocene and occasionally Middle Eocene ages are given for the Eocene 1 and Eocene 2 units. The underlying T50 unit has a more pronounced terrestrial (continental) signature (from the ExxonMobil in-house biostratigraphic report) with an assemblage dominated by pollen and free of dinoflagelletes. This evidence suggests a transgressive episode occurred on this part of the Shetland margin from latest Palaeocene times and into at least the Early, possibly Middle Eocene. For location of seismic line see Figure 4.21 and regional correlation B.

4.3.3.5 Internal Geometry

This section highlights the key observations of the internal architecture of the Eocene 2 unit. It focuses on zoomed in portions of interest on the regional correlations to detail the internal seismic geometry. When available, detailed information from the type wells (see **Section 4.3.3.4**) is incorporated with the seismic interpretation into the investigation.

As discussed in the section on distribution of the Eocene 2 unit (**Section 4.3.3.3**), an area of thinning occurs in the southern part of Quadrant 213 and 214 shaped as a small spur. The area of thinning coincides with an area of where the upper bounding reflection (Top Eoc2) shows a distinctive erosive nature over a very small area (**Figure 4.25**). This distinct down-cutting represents a marked unconformity and is a candidate for a major sequence boundary *sensu* Vail *et al.* (1977), Mitchum and Vail (1977) and Vail and Todd (1981). It is however, of very limited in areal extent. The thinning of the unit can be best seen on **regional correlation C** where the upper bounding reflection (Top Eoc2) appears to follow moderate to high amplitude reflections that dip relatively steeply towards the basin centre to the northwest. These northwesterly dipping seismic reflections have a continuous to semi-continuous seismic character and are parallel to sub-parallel. The Top Eoc2 reflection forms the upper boundary of this more steeply dipping zone and cuts below a localised high amplitude mounded seismic package (**see Figure 4.25** and **regional correlation C**). This local incised feature is only seen on two 2-D seismic lines perpendicular to the southeast margin. It represents a constituent part of the Eocene 3 seismic unit, as it sits directly on top of the Top Eoc2 reflection. A 3-D seismic survey does cover the feature however and it will be discussed in more detail in the following Eocene 3 section (**Section 4.3.4**).

A second high amplitude local incised feature is seen in close proximity to the first. The Top Eoc2 reflection again cuts down locally at the base of this feature and forms a strong reflection which dips towards the northwest (**Figure 4.26**). The internal seismic reflections of the Eocene 2 seismic unit show a more continuous character to the southeast where they form parallel to sub-parallel high to moderate amplitude reflection configurations. To the northwest of the local incised feature at the top of the unit, the reflections are generally of lower amplitude (moderate to low) and are

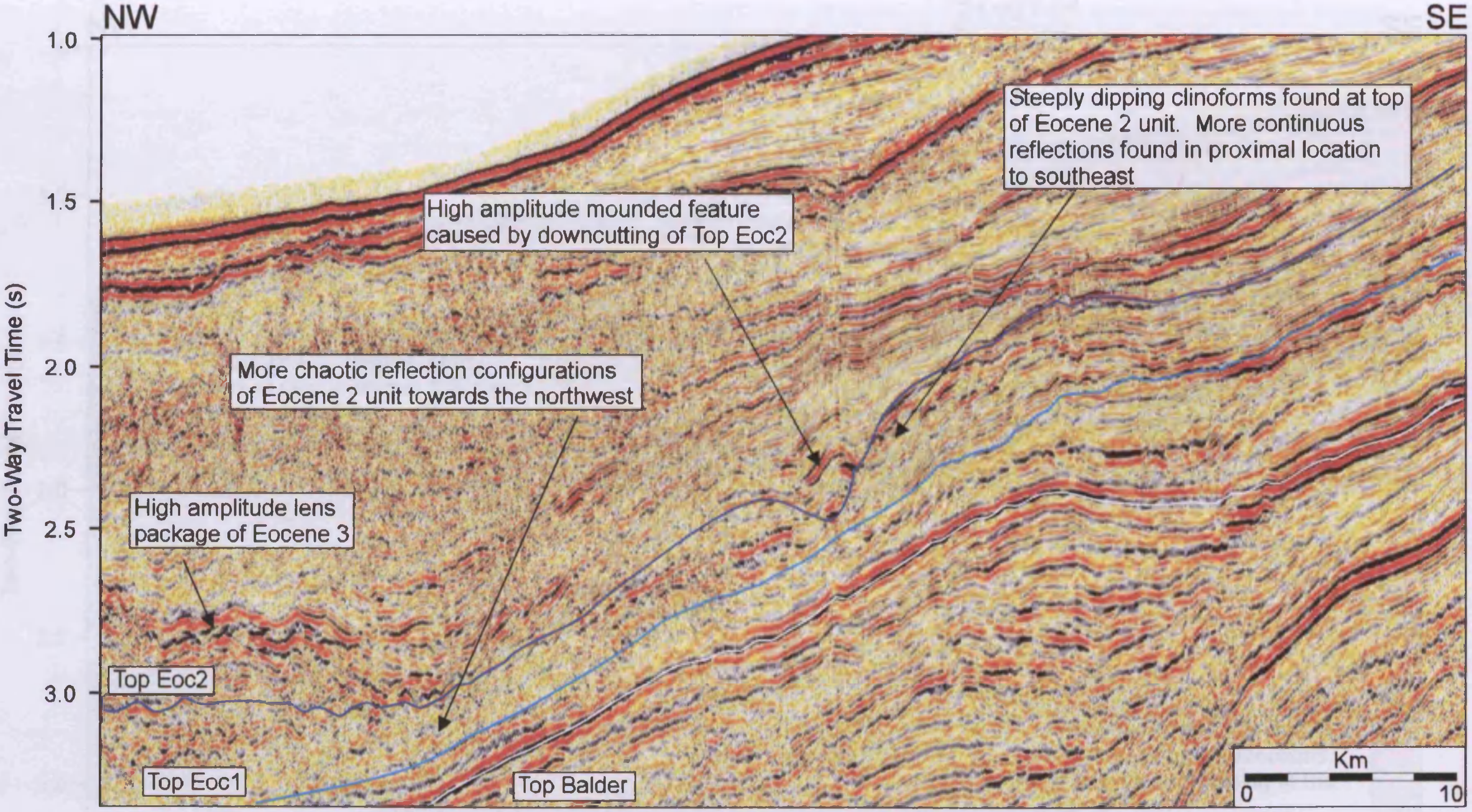


Figure 4.25. 1. Northwest southeast trending 2-D seismic line located on the Shetland margin in the southern part of Quadrant 213. This line highlights the localised thinning of the Eocene 2 unit created by the down-cutting of the Top Eoc2 reflection underneath a high amplitude incisional feature. This feature is only 2- 4 km in width but is bounded to the southeast by steeply dipping reflections which are interpreted as the foresets of clinoforms that shallow in angle up-dip to form topsets. The Eocene 2 unit consists of high amplitude reflections which diverge to the northwest towards the local incisional feature. Further northwest of this feature the Eocene 2 unit seismic reflections exhibit chaotic, low to moderate amplitude configurations. For location of the seismic line see Figure 4.21.

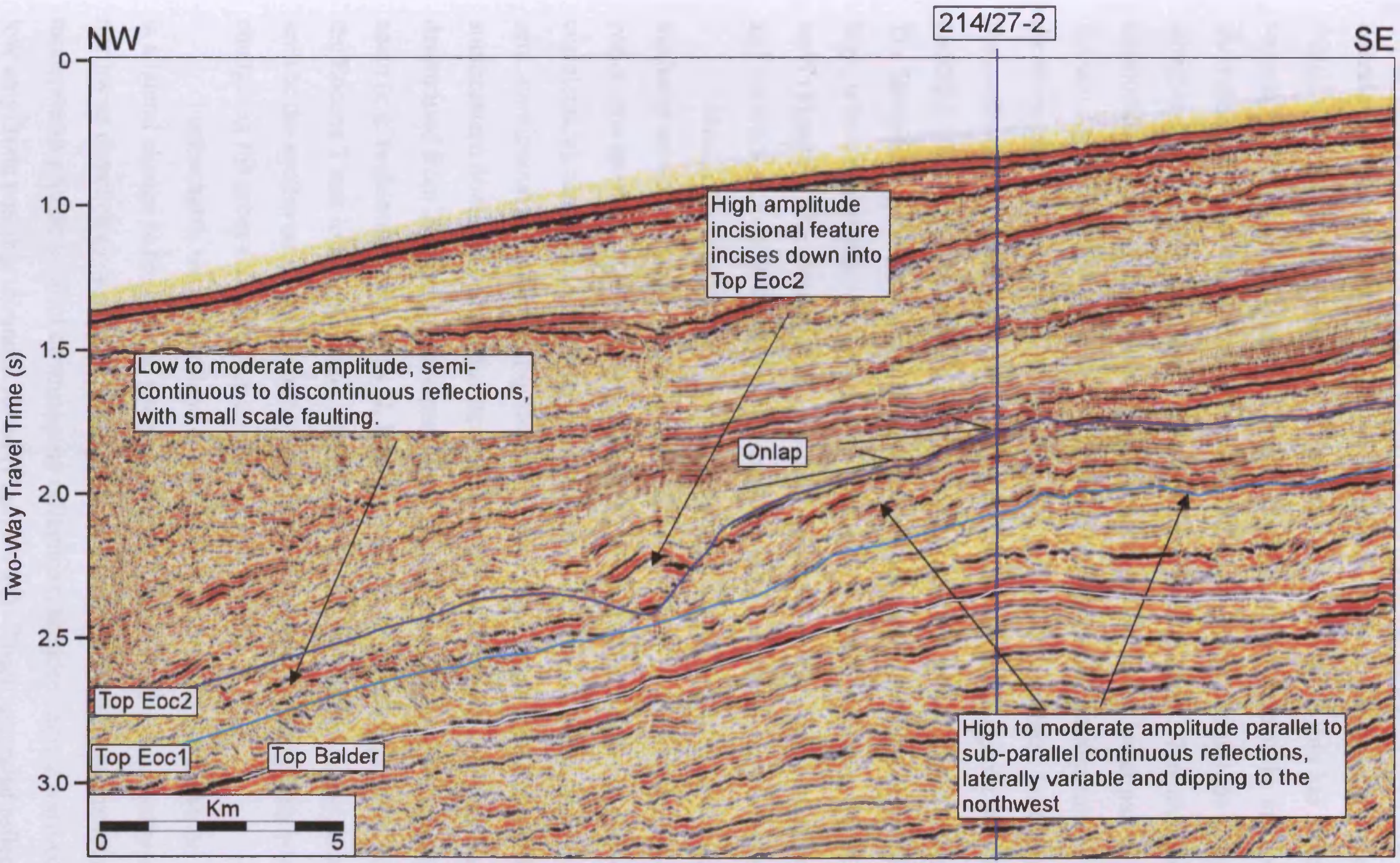


Figure 4.26. Northwest - southeast trending 2-D seismic line showing second localised high amplitude incisional feature which down-cuts into the Eocene 2 unit. This feature has very similar dimensions to previous incisional feature (shown in Figure 4.25). High amplitude continuous to semi-continuous sub-parallel reflections are seen to the southeast of the incisional feature and a more semi-continuous to discontinuous seismic reflection character is seen to the northwest. For location of the seismic line see Figure 4.21.

occasionally semi-continuous to discontinuous. A degree of internal polygonal-style faulting is also evident (**Figure 4.26**).

A more detailed look along this strike direction from a portion of **regional correlation G** shows that there is significant lateral variation in seismic character within the Eocene 2 unit. **Figure 4.27** shows that the top of the unit has a variable seismic character and that there are distinct changes in the value of the amplitude over short distances (less than 5 km). The reflection has very high amplitude components along its length which pass laterally into low or moderate zones. Towards the northeastern end of the strike line (**regional correlation G**) the lower part of the Eocene 2 unit has a strong continuity to the seismic reflection configurations and when these are traced to the southwest, they become difficult to correlate. Furthermore, in the southwest, the Eocene 2 unit is generally discontinuous throughout the whole package, whereas this is only the case to the upper part of the package to the northeast. The lateral change in seismic character occurs close to the location of a small structural high, which is manifested as an anticline structure at a pre T50 (Balder) stratigraphic level (**Figure 4.27**). This structure occurs as two folds with an intervening small saddle and the whole structure spans approximately 10 km in northeast - southwest direction.

Much of the internal architecture within the Eocene 2 unit is not well imaged in the basin axis and into the Faroese sector. Away from the southern part of the basin (where the seismic reflections are generally continuous, high amplitude and readily correlatable), the seismic character is one of generally low amplitude, discontinuous to semi-continuous reflection configurations. However, there remain areas on the southeastern Shetland margin that have some internal architecture that can be determined from 2-D seismic interpretation. Large-scale strike sections through the basin (e.g. **regional correlations G, H** and **I**) show that there is a gross thickening of the Eocene 2 unit to the northeast. Above the Top Eoc2 reflection major downlap is seen to the southwest (in an opposite sense to the thickening) highlighting possible changes in the gross sediment dispersal (see **Figure 4.28**).

Furthermore, on the Shetland margin in the northern part of Quadrant 205, there is a lateral change in the internal geometry of the Eocene 2 unit. In a southeast – northwest direction (towards the basin axis) there is a change from continuous, high to moderate amplitude parallel downlapping reflections, to more chaotic semi-continuous low amplitude possible mounded forms (**Figure 4.29**). These mounded reflections

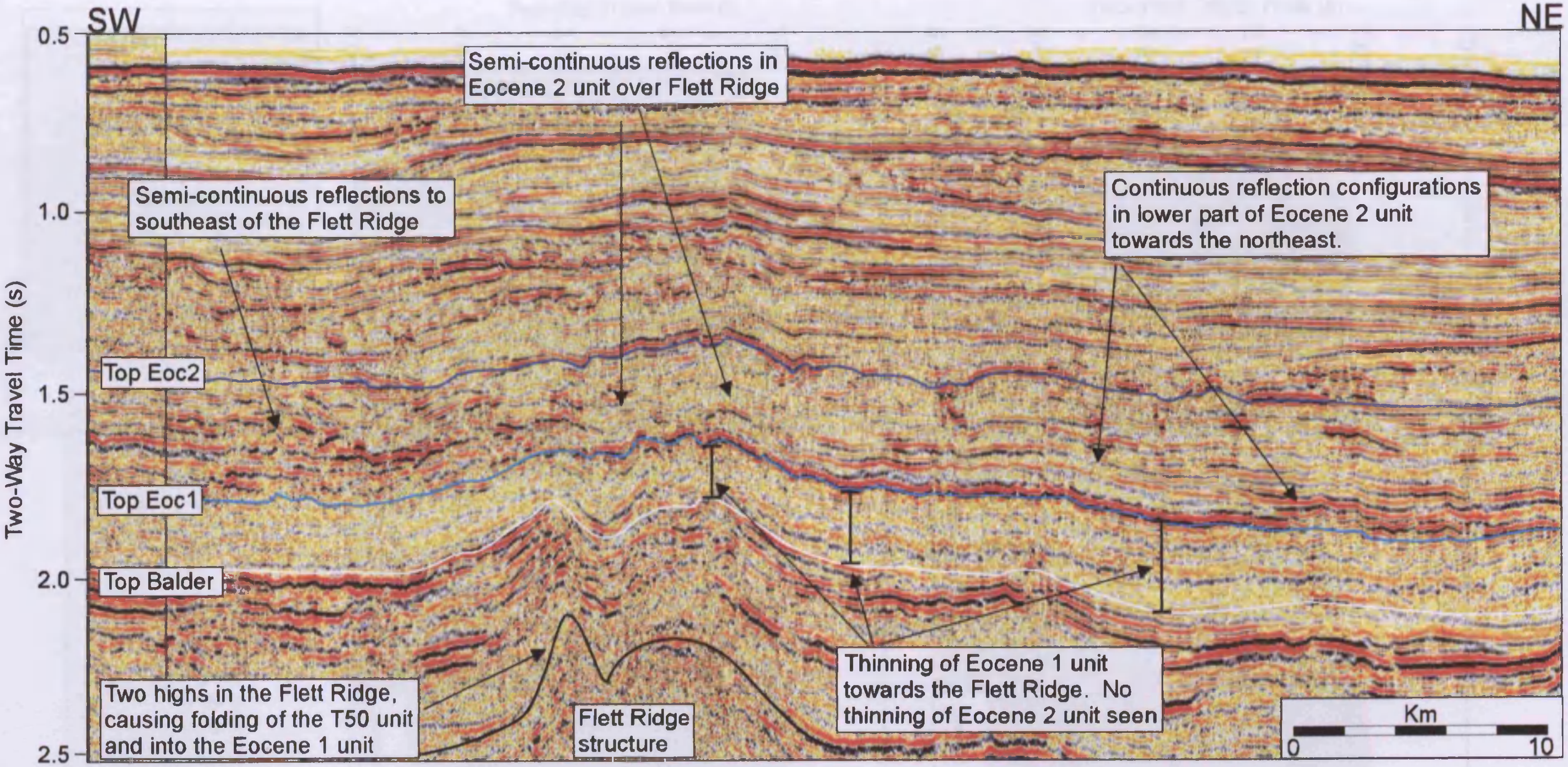
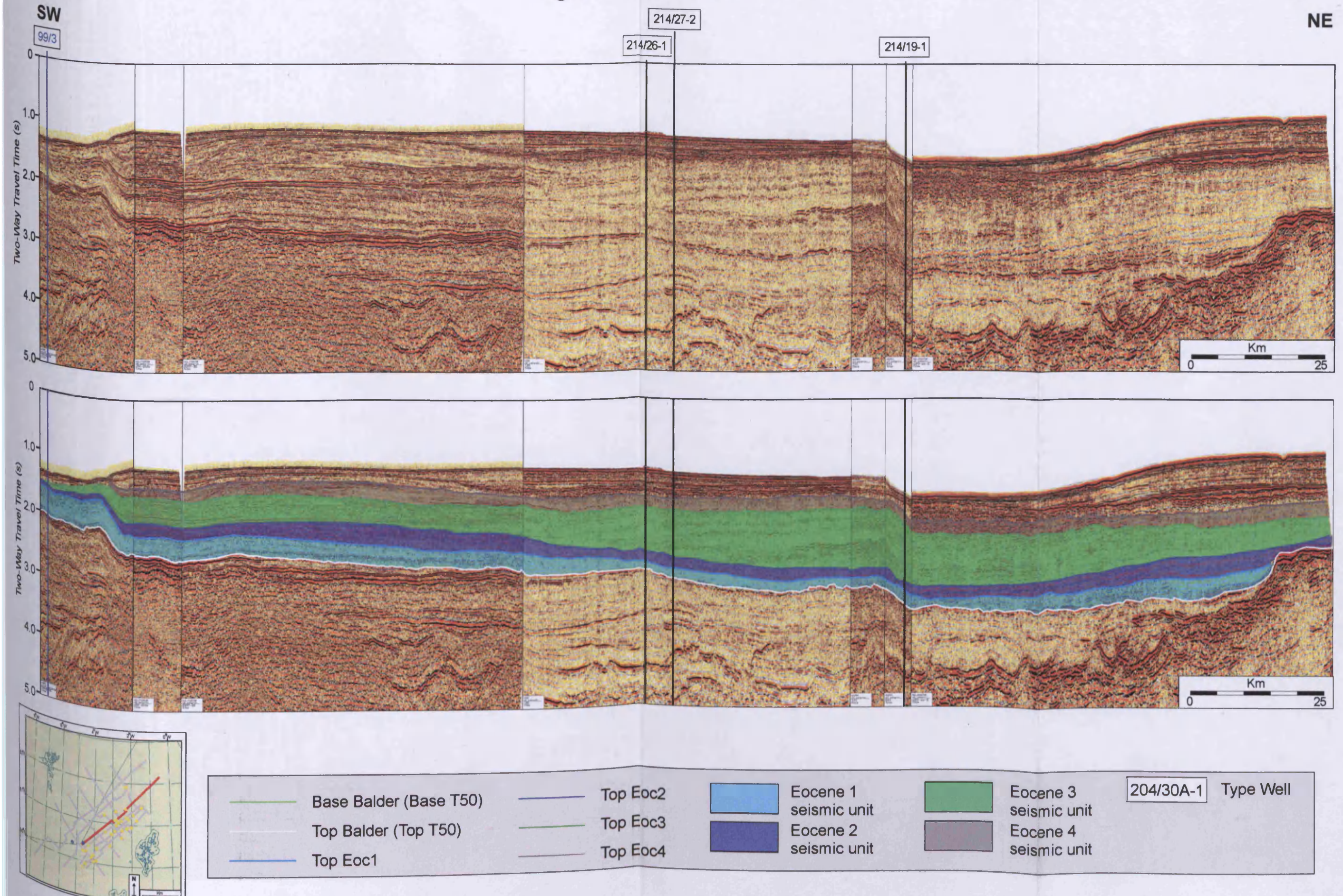


Figure 4.27. Southwest - northeast trending 2-D seismic line across Flett Ridge, showing the two highs and intervening saddle at depth (at the Palaeocene or possibly Cretaceous stratigraphic level. The Eocene 1 unit thins over the crest of the Flett Ridge where as the Eocene 2 unit is seen to be more or less isochronous across the structure. Internally, there is great variation in the Eocene 2 unit. For location of seismic line see Figure 4.21.

Regional correlation I



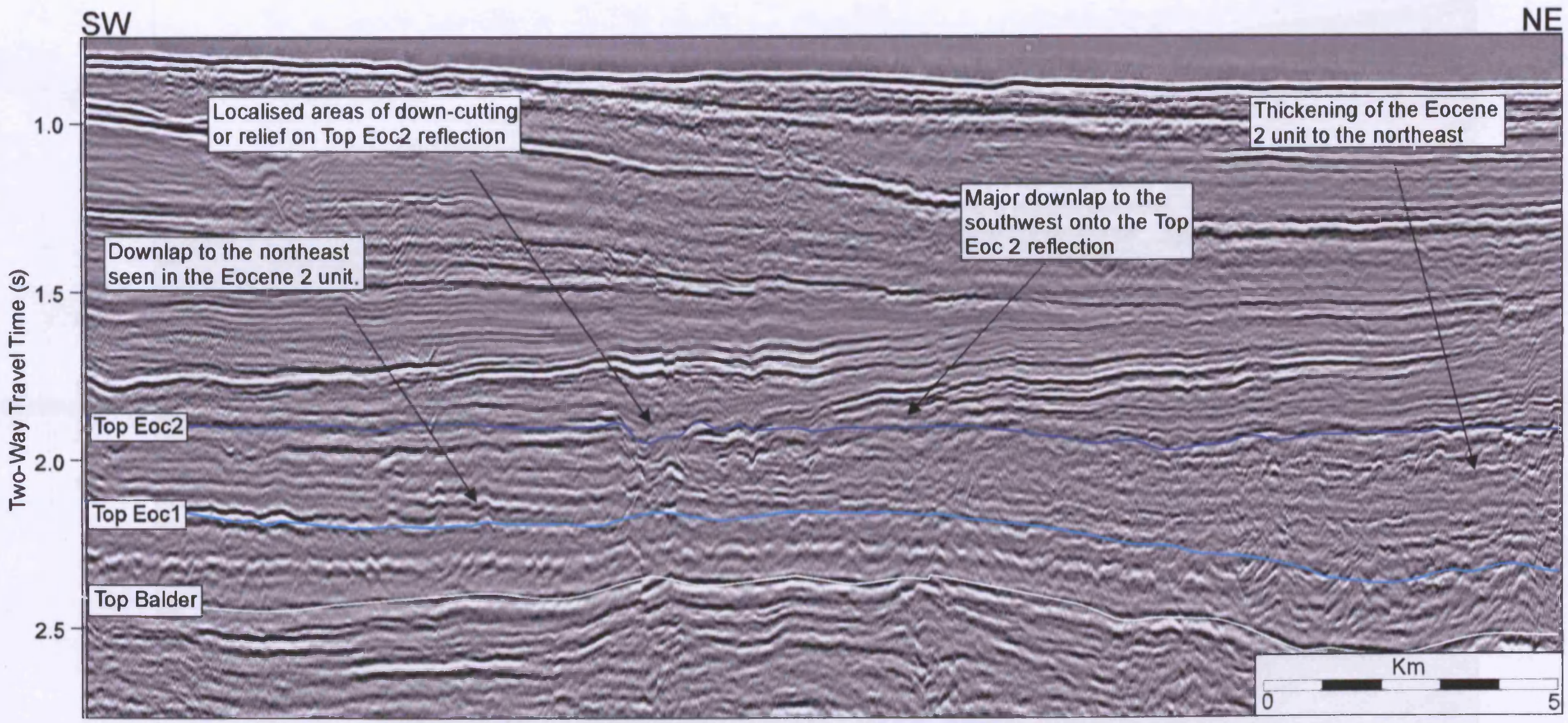


Figure 4.28. Southwest - northeast trending 2-D seismic line showing the change in downlap direction between the Eocene 2 and 3 units. The Eocene 2 unit shows downlap to the northeast which then changes direction to the southwest in the Eocene 3 unit. This observation could reflect changes in gross sediment architecture between Eocene 2 and 3 units throughout the Middle Eocene on the Shetland margin that may represent changes in sediment supply to the marginal areas. This seismic line is located on the Shetland margin (see Figure 4.21) and shows a general thickening of the Eocene 2 unit

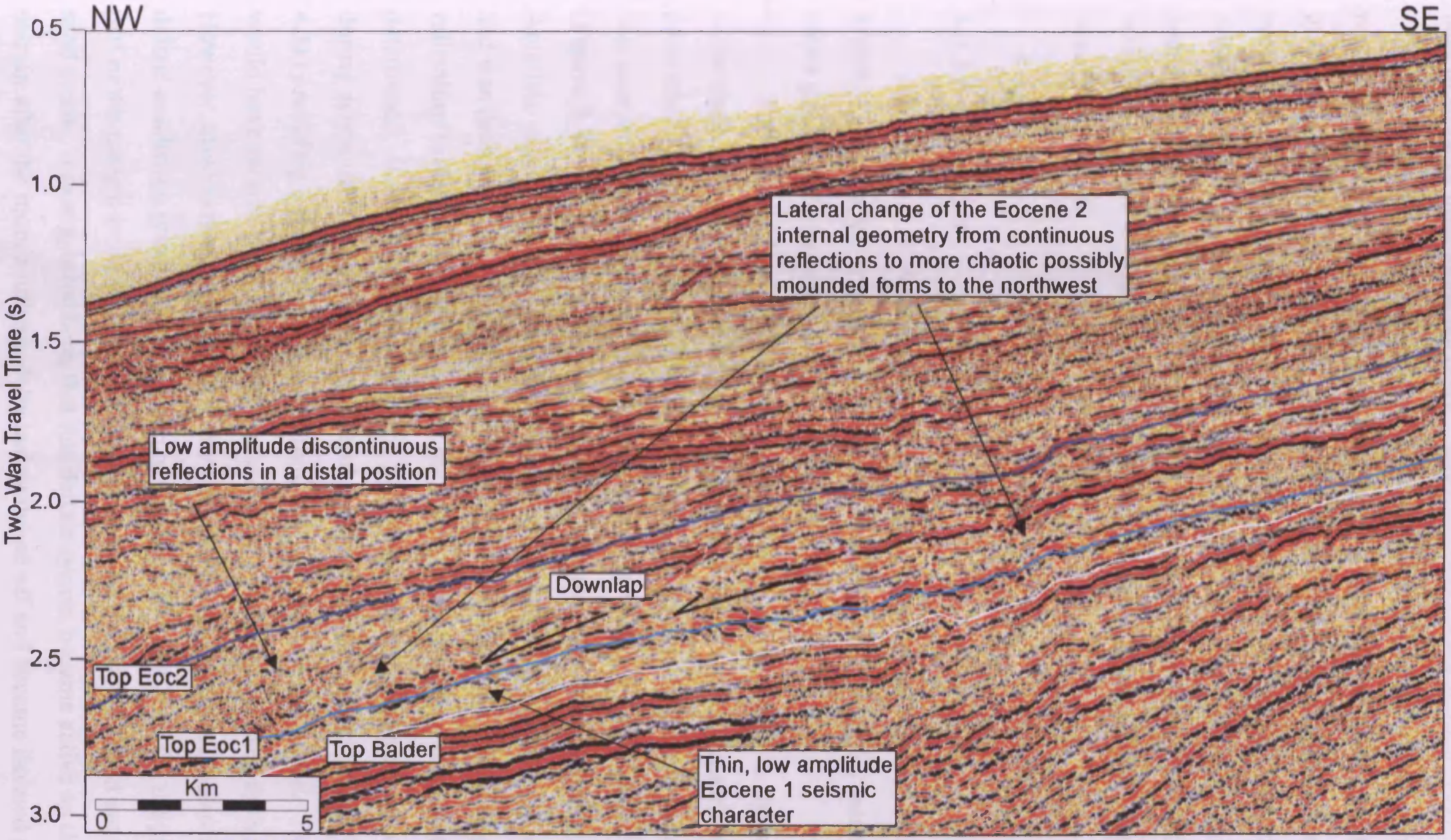


Figure 4.29. Southeast - northwest trending 2-D seismic line on the Shetland margin showing the lateral change in reflection geometry from the basin margin towards the basin centre. The Eocene 2 unit is parallel, continuous and conformable towards the southeast, and becomes lower in amplitude and more discontinuous to chaotic towards the northwest. In this northwestern position, there is evident downlap of reflections which show some degree of localised mounding. Additionally, note the slight change towards the centre of the seismic panel from parallel to divergent reflections in the post Eocene 1 units. For location of seismic line see Figure 4.21.

show possible evidence of bi-directional downlap which may represent some form of base of slope or intra-slope fan body.

Locally, on the northeastern end of the Shetland margin, in the central, western part of Quadrant 208, the Top Eoc2 reflection is locally very variable. On the margin of the basin (to the southeast) the reflection appears as a high to moderate amplitude continuous reflection which dips gently to the northwest. Further to the northwest, the reflection becomes a very high amplitude mounded reflection. These mounds are approximately 1 km apart and are focussed over a small part (5 km) of the Top Eoc2 reflection (**Figure 4.30**) overlying a small hinge at the T50 unit (Balder Formation) stratigraphic level.

4.3.3.6 Discussion

A summary of the basin evolution during the latest Early Eocene to the earliest Middle Eocene at the time of deposition of the Eocene 2 unit will now establish the major geological events that occurred during this time.

There was a drastic change in the environment of deposition in the southern part of the basin between the deposition of the Eocene 1 and 2 units. The earlier southerly delta which prograded north in the Early Eocene had ceased by this time and this area was now subject to relatively little sedimentation, possibly on a distal outer shelf (**Figure 4.31**). Significant onlap and thinning onto the east – west trending Judd Anticline suggests that this structural feature was undergoing compression and uplift and was thus growing during the deposition of the Eocene 2 unit. This convergence of reflections towards the anticline allows the dating of the growth of the structure to be determined. It is postulated here that the Judd Anticline became emergent at some time during deposition of the Eocene 2 unit and may have formed a land bridge (**Figure 4.31**) providing a possible route for mammalian migration to Greenland. This in turn would have created a baffle or barrier to sediment being sourced from the south. However, clinoforms prograded from the Shetland margin towards the northwest and deltaic conditions prevailed in the north of Quadrant 205 and 206. On the proximal part of the margin in these quadrants the sediments of the Eocene 2 unit are particularly sand prone. It is suggested here, that this deltaic system became active on the Shetland margin after the more southerly delta had switched off and became isolated (**Figure 4.31**). This relocation of the main depocentre suggests a major shift in the sediment

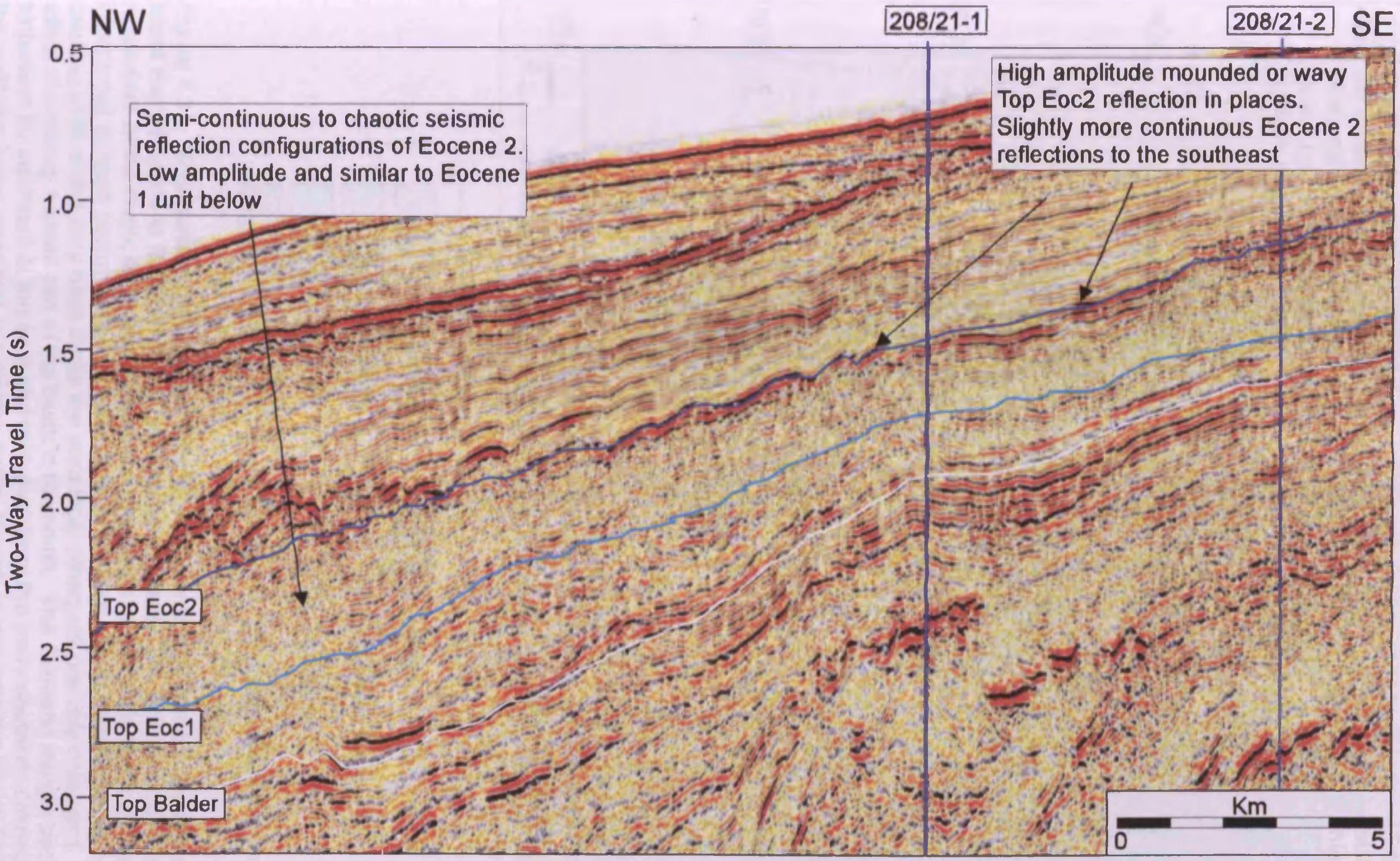


Figure 4.30. Northwest - southeast trending 2-D seismic line in the northeast corner of the basin showing the internal architecture of the Eocene 2 unit on the Shetland margin. In this northern part of the basin the Eocene 2 unit exhibits a moderate amplitude with semi-continuous reflection configurations. Its upper bounding surface however is displays a very high amplitude seismic response and has a mounded or wavy morphology. These mounds are 1-2 km in length and are localised along this margin. They may represent local areas of lithologically different strata and could potentially be composed of carbonate, organic rich shale or coals. For location of seismic line see Figure 4.21.

Latest Early Eocene - earliest Middle Eocene (Ypresian - Lutetian) Palaeogeography (during deposition of the Eocene 2 unit).

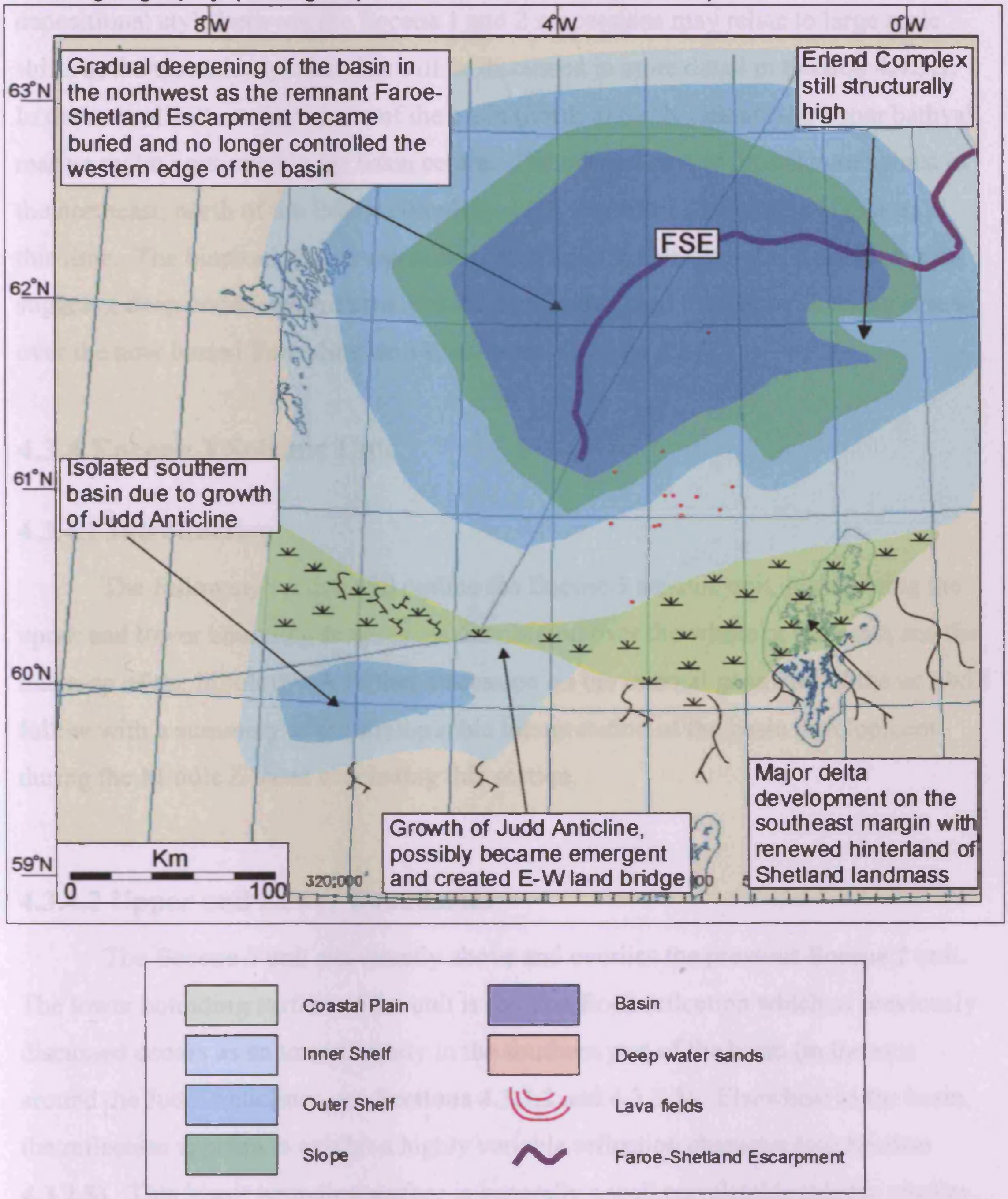


Figure 4.31. Schematic palaeogeographic map at the time of Eocene 2 deposition during the latest Early Eocene to Middle Eocene (Ypresian - Lutetian). The first major observation is the movement of the major depocentre away from the south of the basin to the northeast. This is interpreted to have occurred due compression on the east - west trending Judd Anticline and causing uplift which may have made the area local emergent possibly emergent. This had the affect of isolating a small part of the basin in the south. The Shetland margin became the major hinterland for sediment at this time and took over from the more southerly Orkney Platform. In the north the basin remained in a marine realm, though at this time the Faroe-Shetland Escarpment (FSE) was not structurally high and became breached to the west. Major sediment deposition occurred on the shelf of the Shetland margin in the area of northern Quadrant 205 and 6 and southern parts of 214 and 208. For full key of depositional systems see Enclosure M.

source with the hinterland moving from the Orkney landmass towards the northeast towards the present day Shetland Islands. The causes of this change in the gross depositional style between the Eocene 1 and 2 successions may relate to large scale shifts in the tectonic style but this will be discussed in more detail in **Section 4.4.3.1**. In the central and northern parts of the basin (north of 61° N latitude) an upper bathyal marine realm continued in the basin centre. The deepest part of the basin remained in the northeast, north of the Erlend complex which was still a structurally high area at this time. The biostratigraphic signatures from wells in this central and northern area suggest a deep water environment dominated this area, and little or no thinning is seen over the now buried Faroe-Shetland Escarpment (**Figure 4.31**).

4.3.4 Eocene 3 Seismic Unit

4.3.4.1 Introduction

The following section will outline the Eocene 3 seismic unit, highlighting the upper and lower bounding surfaces, its distribution over the whole of the basin and the make-up of the lithology. A further discussion on the internal geometry of the unit will follow with a summary of the stratigraphic interpretation of the basin development during the Middle Eocene concluding this section.

4.3.4.2 Upper and Lower Boundaries

The Eocene 3 unit sits directly above and overlies the previous Eocene 2 unit. The lower bounding surface of the unit is the Top Eoc2 reflection which as previously discussed occurs as an unconformity in the southern part of the basin (in the area around the Judd Anticline - see **Sections 4.3.3.2** and **4.3.3.3**). Elsewhere in the basin, the reflection appears to exhibit a highly variable reflection character (see **Section 4.3.2.5**). This lower bounding surface is generally a well correlatable seismic marker throughout the southern part of the basin and it can be traced onto the south eastern Shetland margin where along strike it appears as a major downlapping surface (e.g. see **regional correlation G**). On this margin, in a dip sense the Top Eoc2 reflection manifests itself as a variably dipping surface that occurs at the base of a moderate to high amplitude seismic package of continuous to semi-continuous parallel reflection configurations. Onlap onto the reflection is seen locally on this margin and this will be discussed later in **Section 4.3.4.6**. The variable dip of the reflections is best visualised in areas to the north of Quadrants 205 and 206 (and in the southern parts of Quadrants 213 and 214). Here there is a pronounced local steepening of the gradient to the northwest and dips down an interpreted clinoform system and under a localised high seismic amplitude incised feature (discussed in more detail in **Section 4.3.3.6**). Southeast of the dipping clinoform part of the reflection, the Top Eoc2 reflection becomes parallel to sub-parallel to the underlying and overlying units and is interpreted as the topset. In this position of topset development, the seismic character of Top Eoc2 reflection is very laterally varied with both low and high amplitude portions to the reflection being common. This could be indicative of varied lithofacies on the delta top (see later **Section 4.3.4.6**).

The upper boundary of the Eocene 3 unit is named here as the Top Eoc3 reflection and appears as a dark green horizon on all seismic lines. Like the previous regionally correlatable reflection, it was defined in the southern area of the basin as a high seismic amplitude continuous reflection. Additionally it was mapped and defined on the southern part of the Shetland margin where it appears as a high amplitude seismic marker which forms the top of a major progradational package, namely the Eocene 3 unit. This observation can be seen on **regional correlation A** on the southeastern margin of the basin, where the Eocene 3 unit is relatively thin but seen to prograde to the northwest.

4.3.4.3 Distribution of Eocene 3

The distribution of the Eocene 3 seismic unit is again very widespread and is found across the entirety of the basin. As with the previous regional seismic markers the Top Eoc3 reflection has been modified since deposition, through a prolonged period of subsidence with occasional episodic compressional pulses (see **Section 2.2.3.5** and **2.2.3.6**). These tectonic events together give rise to the present day structure of the Top Eoc3 reflection seen in **Figure 4.32**. The common northeast - southwest trending depocentre is again seen, with the Top Eoc3 reflection deepening to the northeastern corner of the basin. The structure map (**Figure 4.32**) shows the geometry of the basin to be relatively symmetrical about its northeast trending axis with both the Shetland and Faroe margins having a similar gradient. In the southern part of the basin, the Top

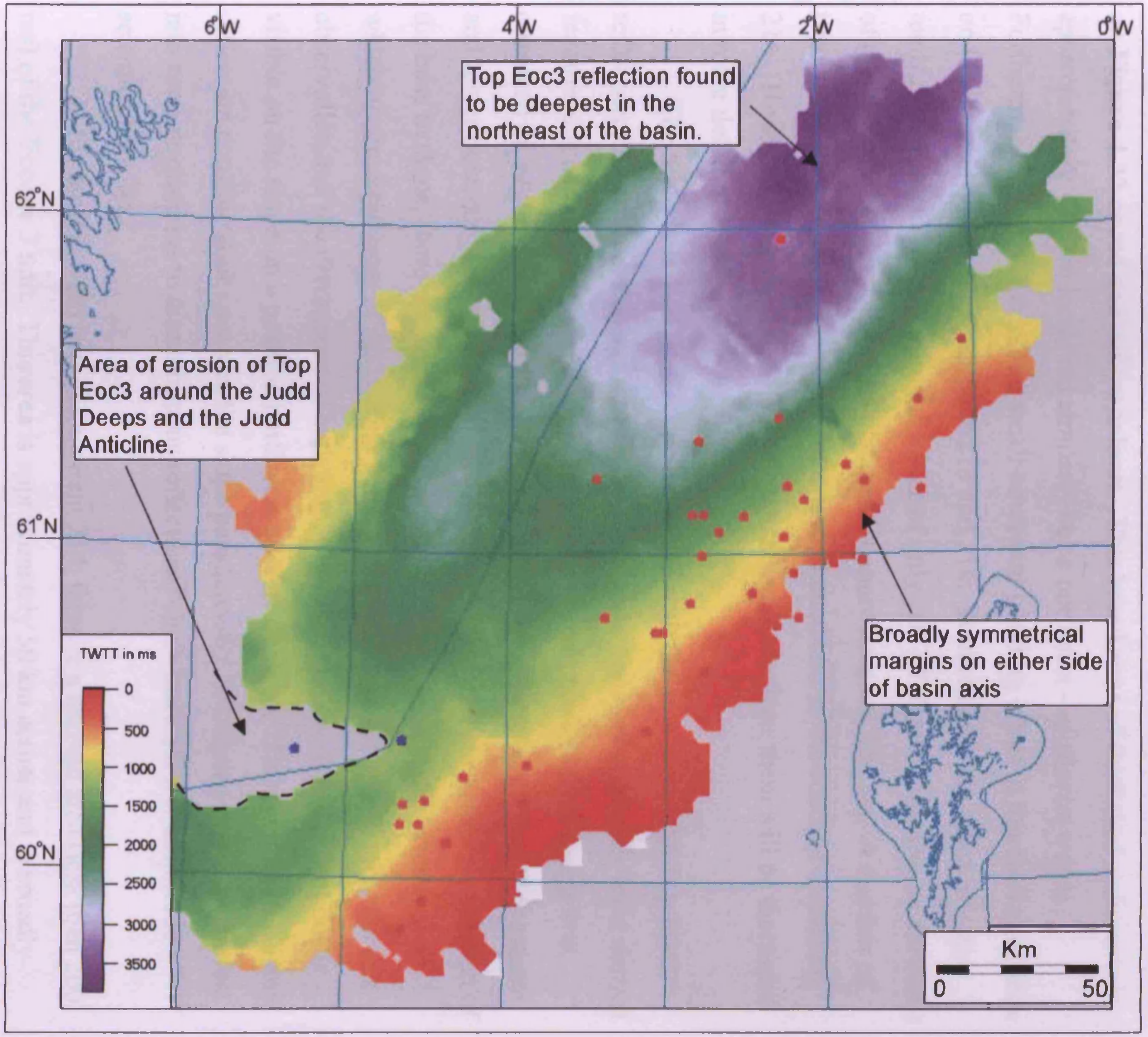


Figure 4.32. TWTT structure map of Top Eoc3 reflection. The main features to be noted are the continued northeast - southwest trending basin axes which mirrors the present day bathymetry. The deepest area of the basin is found in the northeastern corner. The margins of the basin look to show an symmetry when compared to the underlying Top Eoc1 and 2 reflections. The Top Eoc3 reflection is also seen to be eroded in the south of the basin in a broadly east - west trending area close to the Judd Anticline and Judd Deeps and the Munkagrunnur Ridge.

Eoc3 reflection is seen to be preserved to the south of the east - west Judd Anticline. However, there is an area where the Top Eoc3 reflection has been eroded by younger seismic reflections around the Judd Deeps (**Figure 4.32**).

The thickest area of the preserved Eocene 3 seismic unit is seen in the central part of the basin in the southern part of Quadrant 214 and to an extent 213 (**Figure 4.33**). Here, in a small northeast - southwest trending lens, the Eocene 3 succession reaches a thickness of over 900 ms. This lens is approximately 50 km across and has a width of about 22 km and is centred on block 214/18 and 214/19. A generally thick area covers the southern part of Quadrants 213 and 214 and this thick area is depicted on **Figure 4.33** as red and yellow colours. This broad zone of thickened sediments is approximately 100 km long (and similarly has a northeast - southwest trend). Furthermore, there seems to be a small northwest trending offset in this thickened area on the boundary between Quadrants 213 and 214. Southeast of this offset, the area tends to be thinner, with maximum values of only 700 - 750 ms, compared to northeast of the offset, where the maximum thickness is approximately 950 ms. A number of type wells are located in the vicinity of this thickened lens in the south of Quadrant 214. These type wells and the lithological data available from them will be discussed in more detail in **Section 4.3.4.5**.

The thick lens discussed above coincides with the area where the Top Eoc2 reflection (lower bounding surface) dips steeply under the high amplitude local incised features discussed in **Section 4.3.3.6**. The geometry of this thickened lens is best observed on **regional correlation D**, where significant divergence between the upper and lower bounding surfaces can be seen on the Shetland margin. Near the position of the base in slope, there appears to be an area of high amplitude seismic reflections which occur at the base of the Eocene 3 unit. **Regional correlation C** confirms this observation and the divergence of high amplitude continuous seismic reflections is visible on the slope, in a position northwest of well 206/2-1a. The continuous seismic reflection configurations seen on the slope pass laterally and diverge to the northwest into semi-continuous to discontinuous reflections which have a lower amplitude seismic character.

In the northeastern part of Quadrant 214, there is a thinned area (less than 200 ms) of the Eocene 3 unit. This area is approximately 50 km across and is broadly circular, though it has a slight north - south trend and lies to the south of well 214/4-1 (**Figure 4.33**). To the east and west of this thinned zone, the area thickens (to over 300

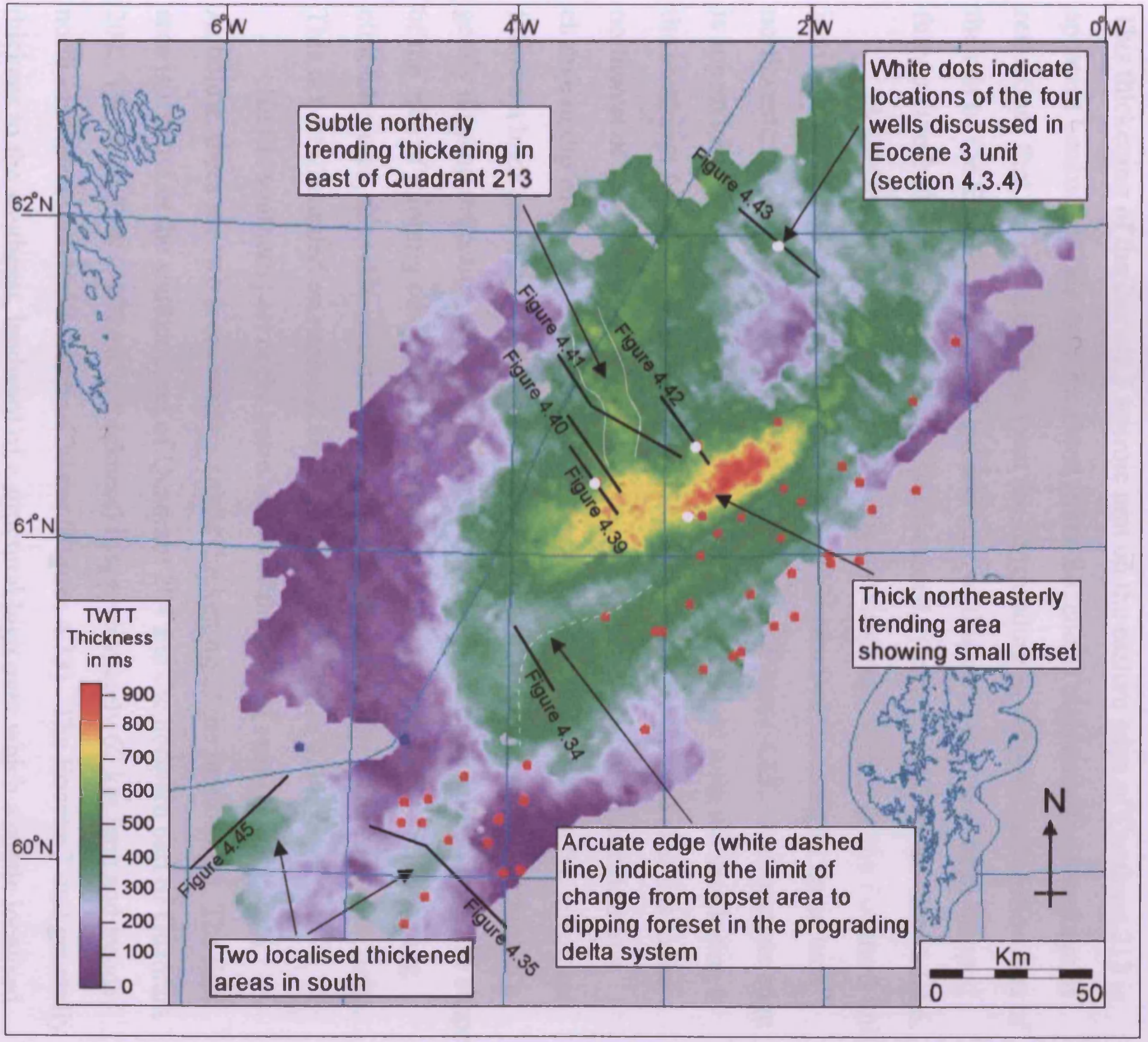


Figure 4.33. Isochron map in TWTT showing the preserved thickness of Eocene 3 unit. The thickest area is located in the centre of the basin. A northwest - southwest trending offset is seen in this thick area, which reaches a maximum of over 900 ms. A thinning to the north of this feature in the northern part of Quadrant 214. A subtle north - south trending slightly thick area is seen on the eastern edge of Quadrant 213 and this corresponds with one of the high amplitude features seen within the Eocene 3 unit. The Eocene 3 unit is eroded in the south of the basin near the Judd Anticline and Judd Deeps. Finally, a subtle arcuate feature (white dashed line) is seen in the north of Quadrant 205 and this is visible due to the thinning and downlap of a clinoform system in this area. The positions of numerous seismic lines and wells used to describe the appearance of the Eocene 3 unit are shown.

ms, shown by the green colours). Additionally, a thinning of the Eocene 3 unit occurs over the western edge of Quadrant 213 where the Faroe Shetland Escarpment is located.

On the eastern side of Quadrant 213, a subtle area of thickening (feint yellow colours in **Figure 4.33** showing values up to 650 ms) can be seen which trends northwards along the whole length of the Quadrant. This thickening coincides with an area of high amplitude seismic reflections (intra-Eocene 3 reflections) which form a thick package and occur as a large lens shape. The high amplitude reflection configurations are often continuous to semi-continuous, especially near the top of the package but can often be discontinuous to chaotic (see **regional correlations C** and **D**). This thickening of the Eocene 3 seismic unit on the eastern edge of Quadrant 213 is apparent because of this extra localised (intra-Eocene 3) high amplitude package of reflections that is very distinctive from the surrounding lower amplitude reflections of the rest of the Eocene 3 unit. The top and base of this extra unit have been mapped fully on the 2-D seismic grid and will be discussed in greater detail in **Section 4.3.4.6**.

A subtle northwesterly thinning (to approximately 250 ms) of the Eocene 3 unit is visible towards the north of Quadrant 205. This area is delineated by an arcuate northwesterly edge (shown in light purple colours on **Figure 4.33**). This arcuate edge is approximately 60 km long and trends towards the northeast with the arc closing to the southeast (convex to the northwest) **(Figure 4.33)**. The thinned area lies to the northwest of the arcuate feature and is seen on seismic data to occur because of a change in dip of the Top Eoc3 reflection. On the southeastern margin the reflection follows a high amplitude reflection that is near horizontal and laterally changes to a gently dipping reflection to the northwest **(Figure 4.34)**. An interpretation of the shape of the area of thinning could reflect the change from the topset area to the dipping clinoform area, thus this arcuate edge is interpreted as the edge of a prograding delta. This is best illustrated on **regional correlation B** and **Figure 4.34**.

In the southern part of the basin, south of the area of erosion at the Judd Anticline, there are two areas where a subtle thickening of the unit occurs. The first area is located in the southern part of Quadrant 204 and the northern part of Quadrant 202. This area (300 – 400 ms in thickness) is approximately 60 km long and trends north - northeast and has a width of 20 km **(Figure 4.33)**. The Eocene 3 unit generally thickens to the southeast, landward of a structural high over which a subtle localised thinning occurs **(Figure 4.34)**. In the north of Quadrant 202 and the south of Quadrant

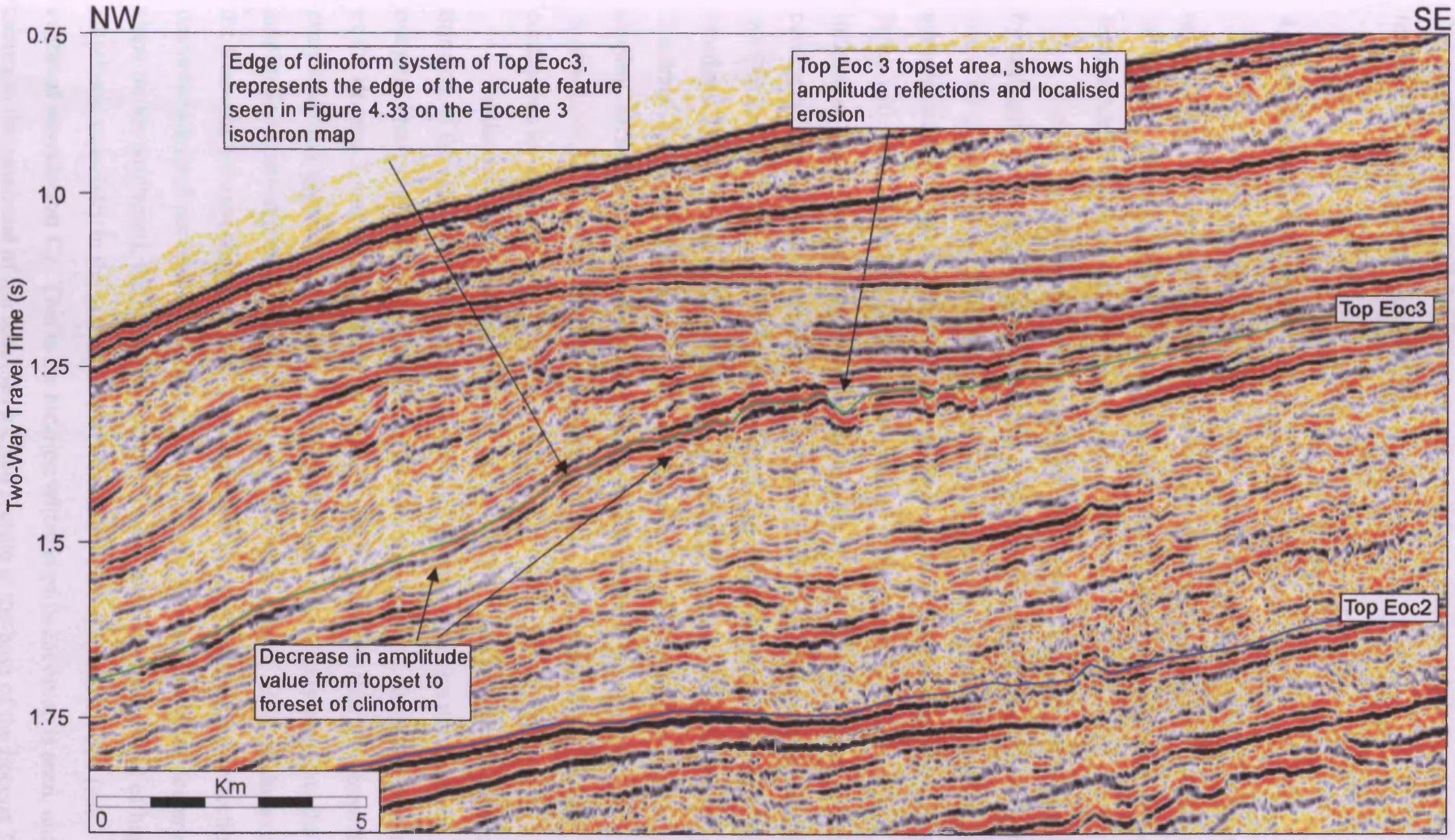


Figure 4.34. Southeast - northwest trending 2-D seismic line showing the thinning of the Eocene 3 unit to the northwest in the south of the basin. The edge of the clinoform system is seen on the Eocene 3 isochron map (Figure 4.33) where a subtle arcuate feature is observed. This is interpreted to be at the break in slope where the sigh amplitude topset dips to the northwest into the lower amplitude foreset. See Figure 4.33 for location of the seismic line.

204 parallel to sub-parallel reflection configurations are seen to downlap and thin towards the northwest (**Figure 4.35**).

4.3.4.4 Internal Geometry

The Eocene 3 seismic unit has a very varied and complex internal stratigraphic architecture. This section will summarise (using examples) all the local variations in the internal geometry of the reflection configurations and provide evidence of local features identified from the 2-D seismic data.

Firstly, on the Shetland margin, the internal geometry of the unit is dominated by high amplitude, gently dipping, parallel reflection configurations. The seismic character changes along these parallel reflections and this is best illustrated on **regional correlation H**. This variation along the seismic reflections reflects a change in seismic facies with the highest amplitude values possibly representing localised coals or limestones in the Eocene 3 unit, (see **Section 4.3.4.5**). The gently dipping reflection configurations dip and downlap to the southwest where they become conformable with the Top Eoc2 reflection in the northern part of Quadrant 202 and the southern part of Quadrant 214. As seen from **regional correlation H**, further to the southwest, (into Quadrant 205) the gently dipping seismic reflections pass laterally into shallower dipping, parallel to sub-parallel reflections, which show little sign of downlap. Furthermore, these reflections have moderate seismic amplitudes and are semi-continuous to continuous.

In the area to the south and east of Quadrant 214, there is major onlap and thinning of the Eocene 3 unit onto the underlying Eocene 2 unit on the Shetland margin. From **regional correlation B** and to a certain extent **C**, the internal geometry of the Eocene 3 unit is one that shows major dipping reflections that are interpreted as progradational systems that downlap to the northwest. Downlap is seen onto the high amplitude localised features discussed above in **Section 4.3.4.3**. Here, in the axes of the basin the seismic character is of low amplitude and the downlapping reflections are discontinuous and pervasively faulted above the high amplitude localised features. Up-slope (to the southeast), the reflections become much more continuous and of higher amplitude especially in the area close to the northern part of Quadrant 206 (see **regional correlation C**). This is the location where significant onlap is seen, and this occurs to the southeast of the localised incised feature at the base of the Eocene 3 unit.

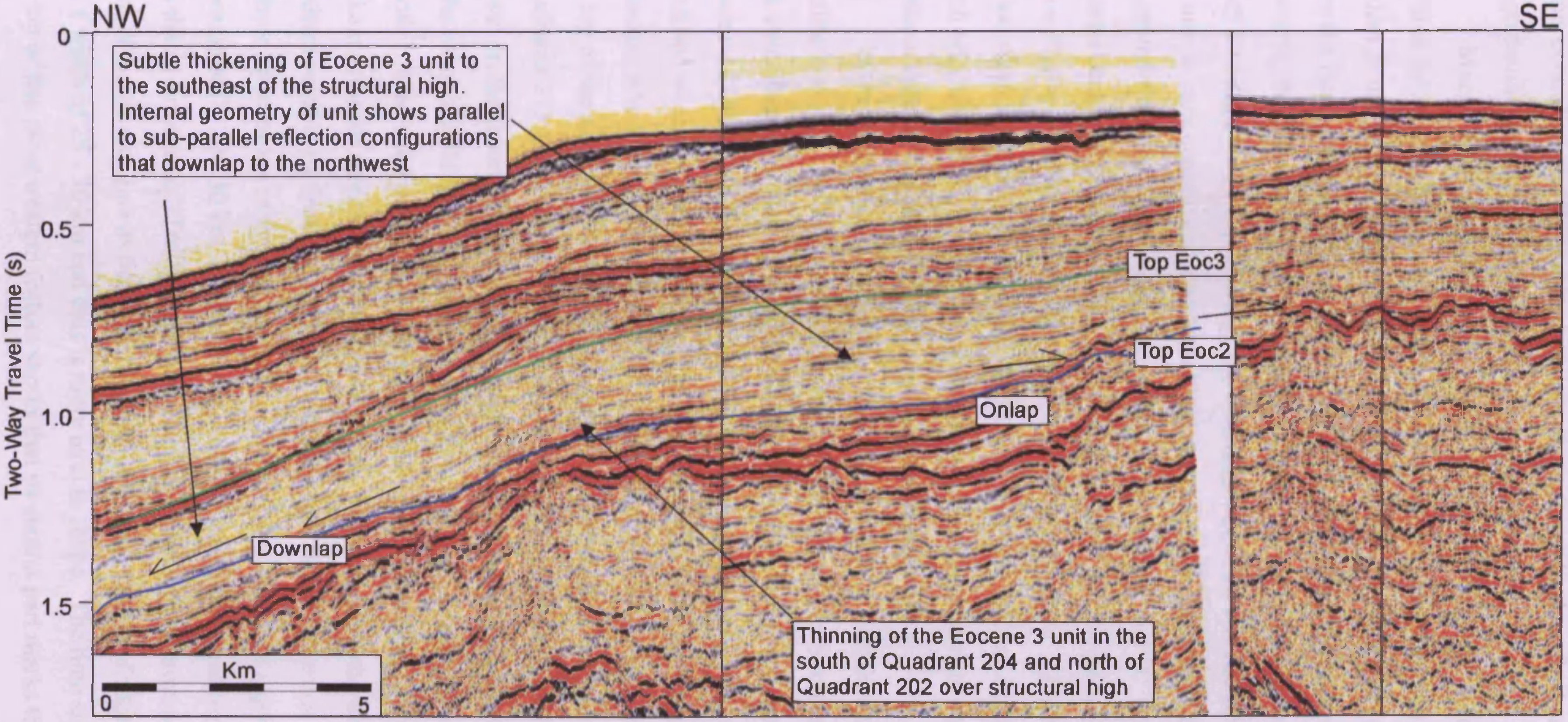


Figure 4.35. Composite 2-D line trending broadly southeast - northwest in the southern part of the basin off the Shetland margin. This seismic line shows the thinning of the Eocene 3 unit to the northwest. In the basin centre (towards the northwest) the Top Eoc3 reflection appears as a high amplitude reflection below which becomes lower in amplitude to the southeast. The Eocene 3 unit displays a package of low amplitude continuous to semi-continuous reflection configuration which remains the same from the margin to the basin centre. This unit thins by onlap to the southeast and also thins to the northwest by downlap. For location of seismic line see Figure 4.33.

In this central part of the Shetland margin, the unit can be described as having a gross progradational geometry being sourced from the southeast.

Much of the internal geometry of the Eocene 3 unit on the Faroes margin and towards the west of the basin is generally very poorly imaged. In particular this is evident in the northwestern corner of the basin in the area of Quadrant 6105 and 6104. Here the Eocene 3 unit thins onto the Faroe margin and is between 0 and 300 ms thick. However, the 2-D data has a large grid spacing in this part of the basin (approximately every 20 km), and the resolution of the data is generally poor. The internal geometry of the unit is very chaotic and very little character is seen in the seismic data. This is best illustrated from the dip orientated **regional correlations C** and **D**. However, on **regional correlation D**, there is a local high amplitude chaotic package that sits on the Faroe Platform on the western side of a major folded structure. This high amplitude package can be seen to be truncated along with the entire Eocene succession under a much later unconformity close to the northwestern edge of the data on the Faroe Platform (see northwestern edge of **regional correlation D**).

In the central and northern part of the basin, there are distinct areas of high amplitude seismic reflectors, which were briefly mentioned in **Section 4.3.4.3**. This area coincides with a subtle thickening of the Eocene 3 unit on the eastern side of Quadrant 213 seen in **Figure 4.33**. Three high amplitude packages have been identified which exhibit a high amplitude continuous seismic reflection at its upper boundary which is occasionally faulted. Because of the distinctive seismic character at the top of the package, the area of high amplitude reflections (which appear in Quadrants 213 and 214) have been extensively mapped on the 2-D seismic data. However, the lower boundaries of the high amplitude packages are difficult to correlate as the seismic reflection configurations are generally low amplitude and discontinuous - chaotic. A detailed map of the three individual upper boundaries of the high amplitude packages reveals three narrow elongate features which trend north - south from the southern reaches of Quadrant 213 and 214 (**Figure 4.36**). The central and most eastern of these features are in the region of 100 - 120 km long and have a general width of approximately 20 - 30 km. However, the width of the features narrows to the south to less than 1 km across. The third of these features, which is the most western of the three, is not as elongate as the other two, and only has a length of approximately 50 km and a width of 25 - 30 km and thus is more oval in shape. The time structure map of the top of this most western feature shows that its central part marks the crestal region

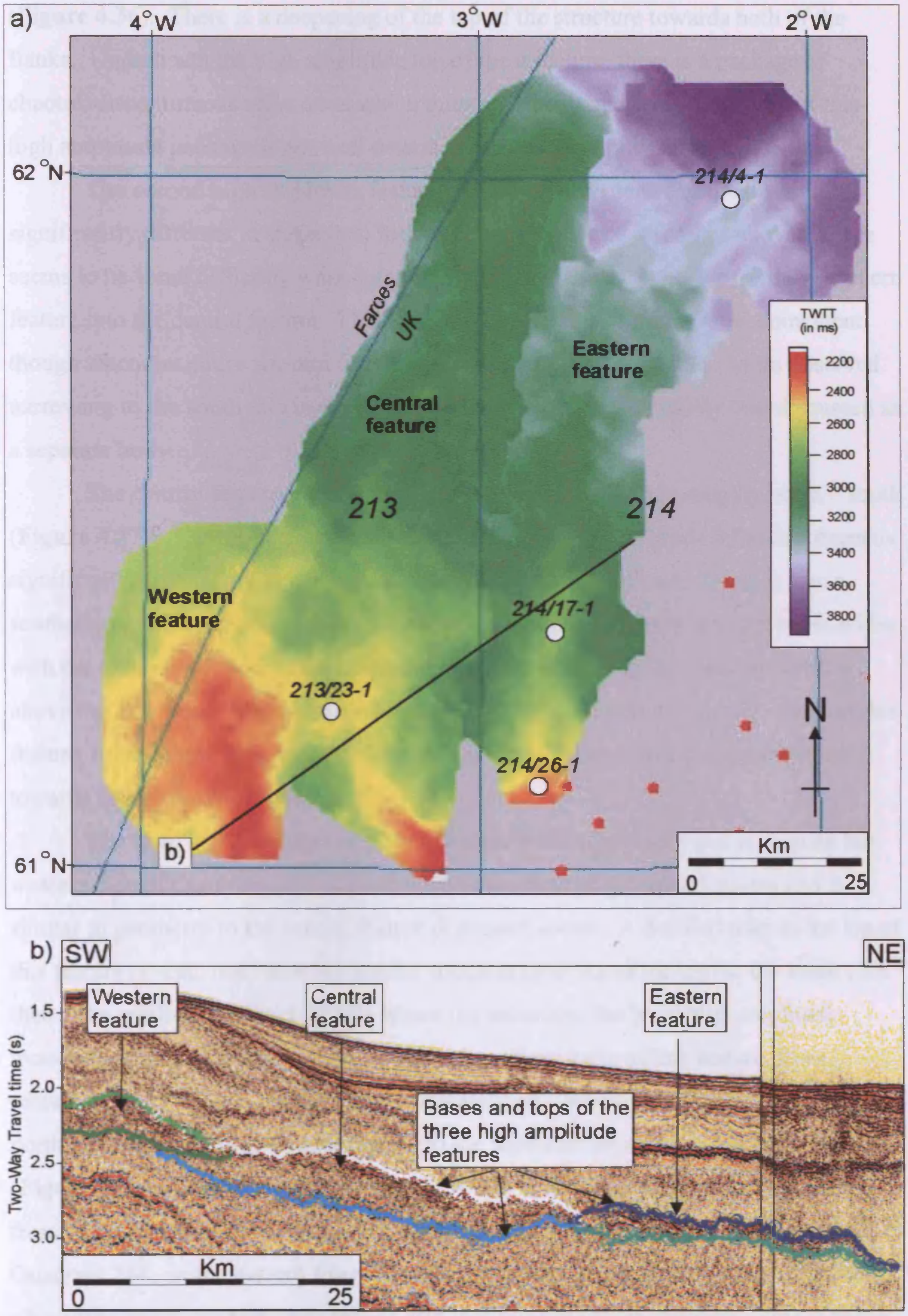


Figure 4.36. a), Time structure map of the top of the high amplitude features seen in the central and northern parts of the basin. This shows the broadly circular western feature and two more elongate narrow features to the east. All three features show a narrowing to the south where the features are shallowest and terminate up-dip The positions of the four wells which penetrate these features is highlighted. b). Shows northeast - southwest 2-D seismic line through the three high amplitude features, showing younging to the northwest.

(**Figure 4.36**). There is a deepening of the top of the structure towards both of the flanks. Underneath the high amplitude top of the structure, there is a package of chaotic, discontinuous reflections which thins to the east and west. The base of this high amplitude package is not well determined and is generally diffuse.

The second high amplitude feature is located in the centre of the three and is significantly different in shape than the western most feature discussed above. There seems to be some difficulty when interpreting the top reflection across from the western feature into the central feature. This is because they seem in places to be coincident, though when the entire western feature is mapped out there appears to be an observed narrowing to the south into its tip (**Figure 4.36**) and hence is probably best discussed as a separate body.

The central feature is elongate for over 100 km and trends roughly north – south (**Figure 4.37**). The top structure map shows that the high amplitude reflection deepens significantly from south to north from 1500 ms to nearly 4000 ms. There is also a southerly tapering (or narrowing) of the high amplitude top structure and this coincides with the location of the localised high amplitude incised features which sit directly above the Top Eoc2 reflection (see **Section 4.3.3.6**). Towards the northern limit of this feature, there seems to be a slight change in the trend where the top structure turns towards the northeast (**Figure 4.37**).

The third and final high amplitude feature within Eocene 3 unit is seen on the western side of Quadrant 214. This is the eastern most of the three features and is similar in geometry to the central feature discussed above. A detailed map of the top of this feature reveals that there is a similar southern tapering of the unit to the south (less than 2 km width) into Block 214/26 where the second of the local high amplitude incised features is located. In the central and northern parts of this feature, there remains a fairly constant width of approximately 20 - 30 km, but this thickens out in the north of Quadrant 214 and turns slightly to the northeast into a broad circular lobe (**Figure 4.38**). This lobe is approximately 50 km in diameter and differs significantly from the elongate narrow central and southern parts of the feature. In the central part of Quadrant 214, on the eastern fringe of the main north - south feature, there seems to be a localised small circular, oval shaped area that joins the main structure. This is in the region of Block 214/15 and is approximately 10 – 15 km in diameter.

In the area of the three high amplitude features there is some degree of well control for calibration with the seismic data. Four wells are seen to penetrate the high

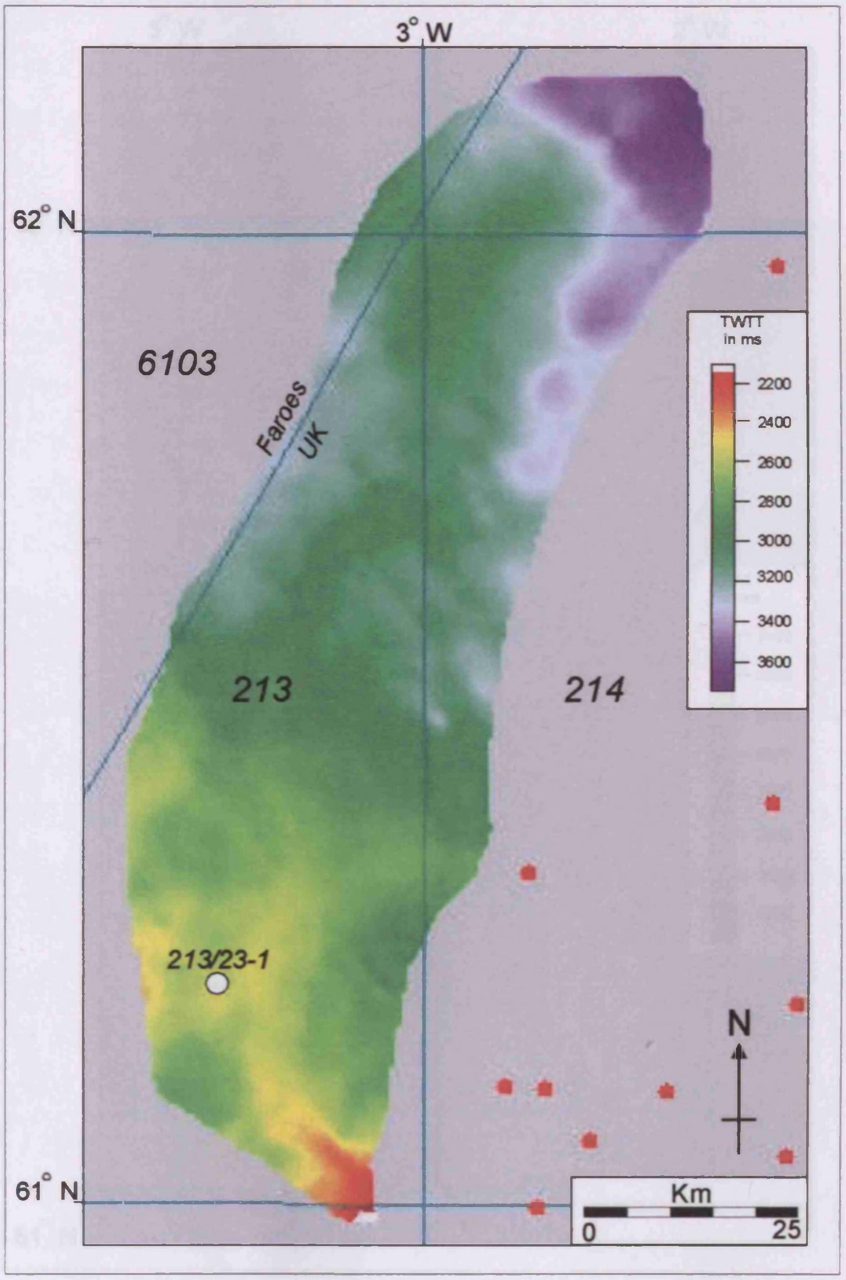


Figure 4.37. TWTT structure map of top of central high amplitude feature seen on the eastern side of Quadrant 213. The feature maps out as a narrow elongate body which has a centrally high crestal area especially towards the south where it is structurally highest. The feature is an average of 25 - 30 km wide and is over 100 km long and trends roughly north - south, though it slightly bends towards the east, especially near to the northern limit. The feature shows a significant narrowing towards the southern tip where it is less than 5 km in width. The feature has been drilled by one well (213/23-1) which is located close to the crestal structure and towards the south part of the feature.

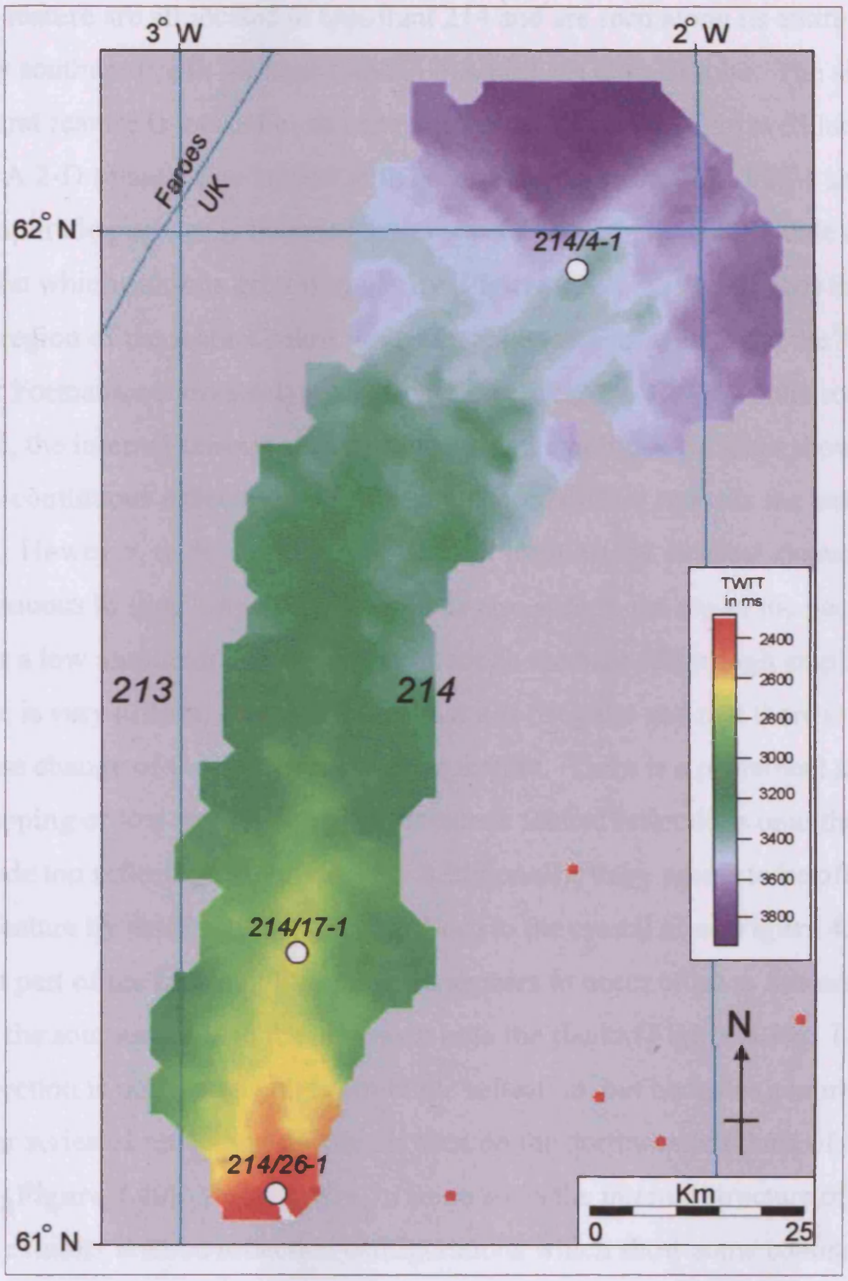


Figure 4.38. TWTT structure map of the top of the eastern most high amplitude feature located in Quadrant 214. This feature is also extremely elongate and is only 20 - 25 km in width. It stretches for over 100 km in a broadly north - south direction and has a gradual bend towards the east, creating an arcuate geometry. It has many similarities to the central feature (Figure 4.35) including the overall dimension, a central high crestal area and its orientation. However, this feature has a lobate curved outer northern limit which is approximately 50 km in width. Three wells penetrate this feature (white dots), one located at the southern narrow tip, one on the central crest in the middle and one at the northern lobate end.

amplitude features seen within the Eocene 3 unit. Three of these wells are located in the eastern high amplitude feature and one in the central feature. The three wells in the eastern feature are all located in Quadrant 214 and are seen along its entire length; at the very southern tip, in the centre and in the northern circular lobe. The single well in the central feature is located in its central part (see **Figure 4.36** for well locations).

A 2-D seismic line located in the central feature at well 213/23-1 shows that the high amplitude package is bounded at the upper limit by a high amplitude continuous reflection which exhibits great irregularity (**Figure 4.39**). The well goes through a high crestal region of the feature below which there is a folded structure at the T50 unit (Balder Formation equivalent) stratigraphic level (**Figure 4.39**). To the southeast of the well, the internal seismic character of the high amplitude package show continuous to semi-continuous reflections which become more diffuse towards the base of the feature. However, to the northwest of the well location, the internal character is more discontinuous to semi-continuous and this is also seen in the top of the package which exhibits a low amplitude response. Even though the base of the high amplitude package is very diffuse, it is clear to see that it is irregular and that there is considerable thickness change of the package across the feature. There is a prominent northwesterly downlapping of low amplitude semi-continuous faulted reflections onto the high amplitude top reflection (**Figure 4.39**). Additionally, there seems to be offset of the top of the feature by small faults, especially close to the crestal area (**Figure 4.40**). The thickest part of the high amplitude feature appears to occur close to this crest and thins both to the southeast and to the northwest onto the flanks (**Figure 4.40**). Locally, the top reflection is not a single high amplitude reflection, but becomes a more chaotic irregular series of reflections and this is seen on the northwestern flank of this central feature (**Figure 4.40**). Furthermore, in some areas the internal structure of the central feature exhibits seismic reflection configurations which show some continuity which can be traced locally and hence may allow for a further sub-division of the high amplitude package.

On a larger scale, the entire high amplitude package of the central feature can be seen to thin considerably to the northwest over the remnant high of the Faroe-Shetland Escarpment (**Figure 4.41**). This suggests that the escarpment still had a significant topographic expression at sea-floor during deposition of the Eocene 3 unit and thus influenced sediment dispersal.

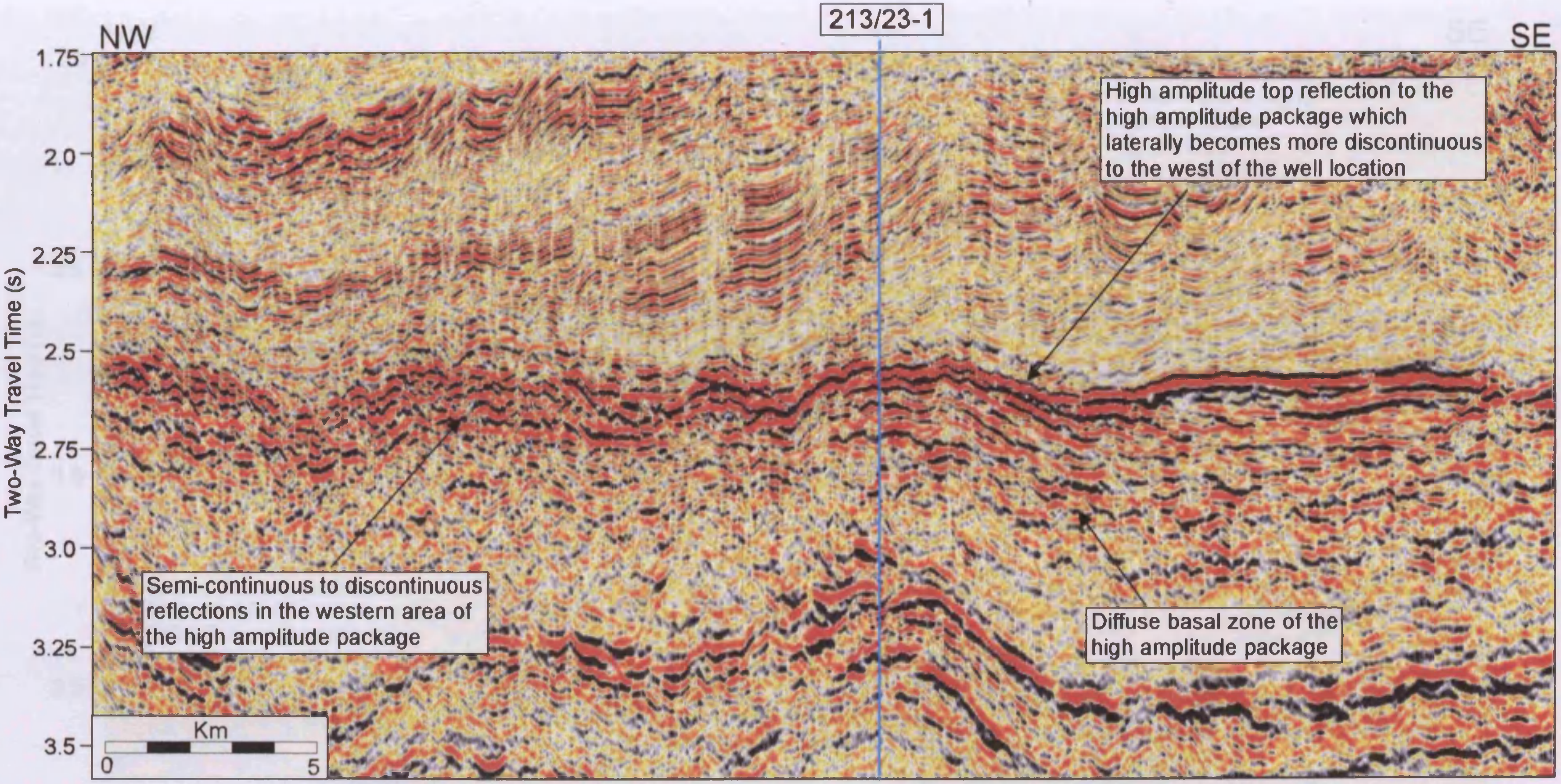


Figure 4.39. Southeast - northwest trending 2-D seismic line through the well location of 213/23-1. This seismic panel shows the high amplitude package which sits within the Eocene 3 unit. The top of the high amplitude unit is marked by a continuous and high amplitude reflection which has an irregular top. The base of the high amplitude package is not as clear to interpret with a diffuse area seen close to the Top Eoc2 reflection. Internal reflections of the high amplitude package are variable across the feature with more continuous reflections appearing towards the southeast of the well location, and more semi-continuous to discontinuous reflections appearing to the northwest of the well location. See Figure 4.33 for location of seismic line.

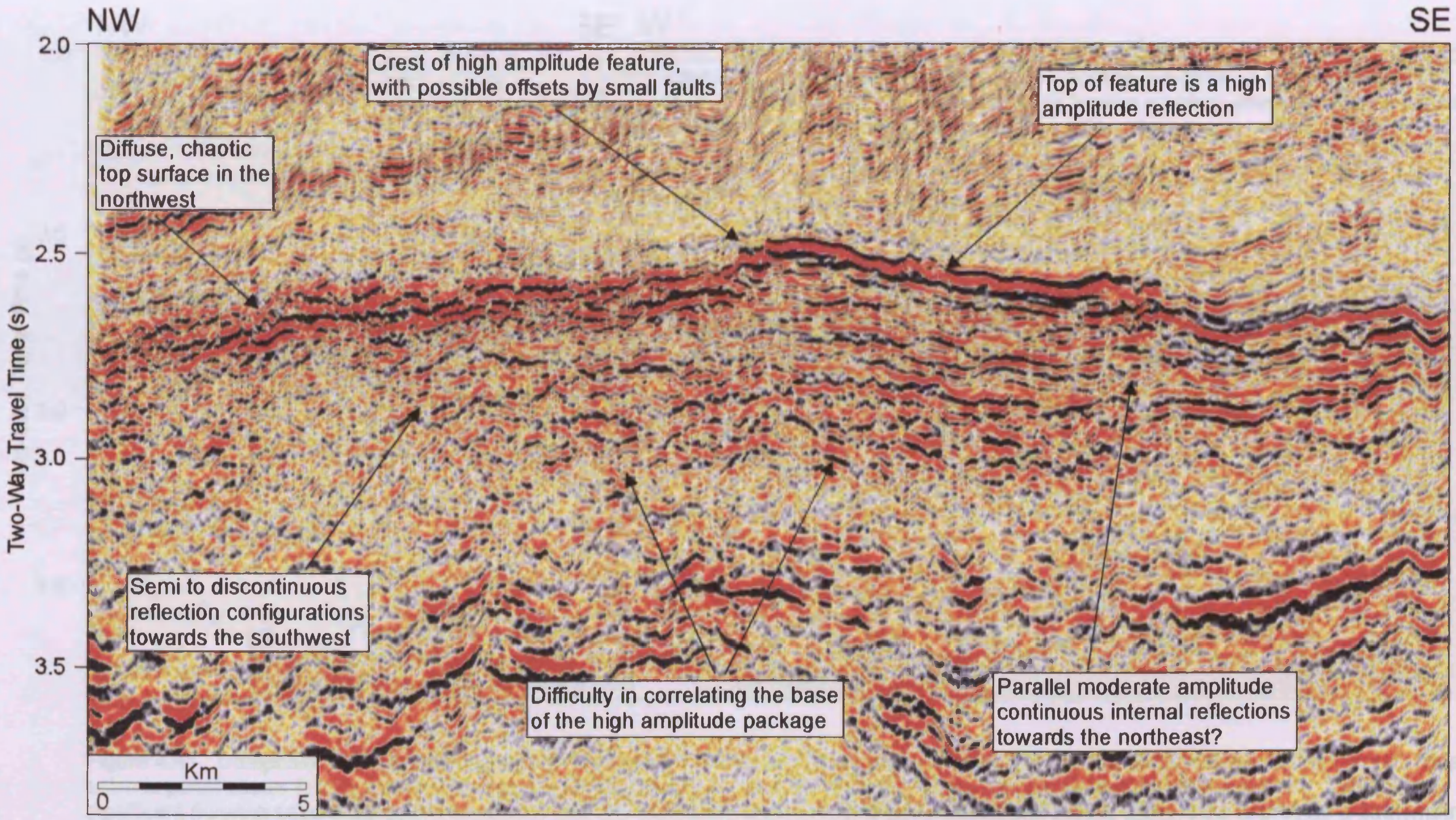


Figure 4.40. Southeast - northwest trending 2-D seismic line across high amplitude intra-Eocene 3 central feature. This seismic line draws attention to the characteristic top of the feature which has a continuous high amplitude reflection and the base which is much more diffuse and chaotic and harder to correlate. However, internal reflections are visible within the high amplitude feature, though these are semi to discontinuous. The whole package is downlapped to the northwest by low amplitude highly faulted discontinuous reflections. See Figure 4.33 for the location of the seismic line.

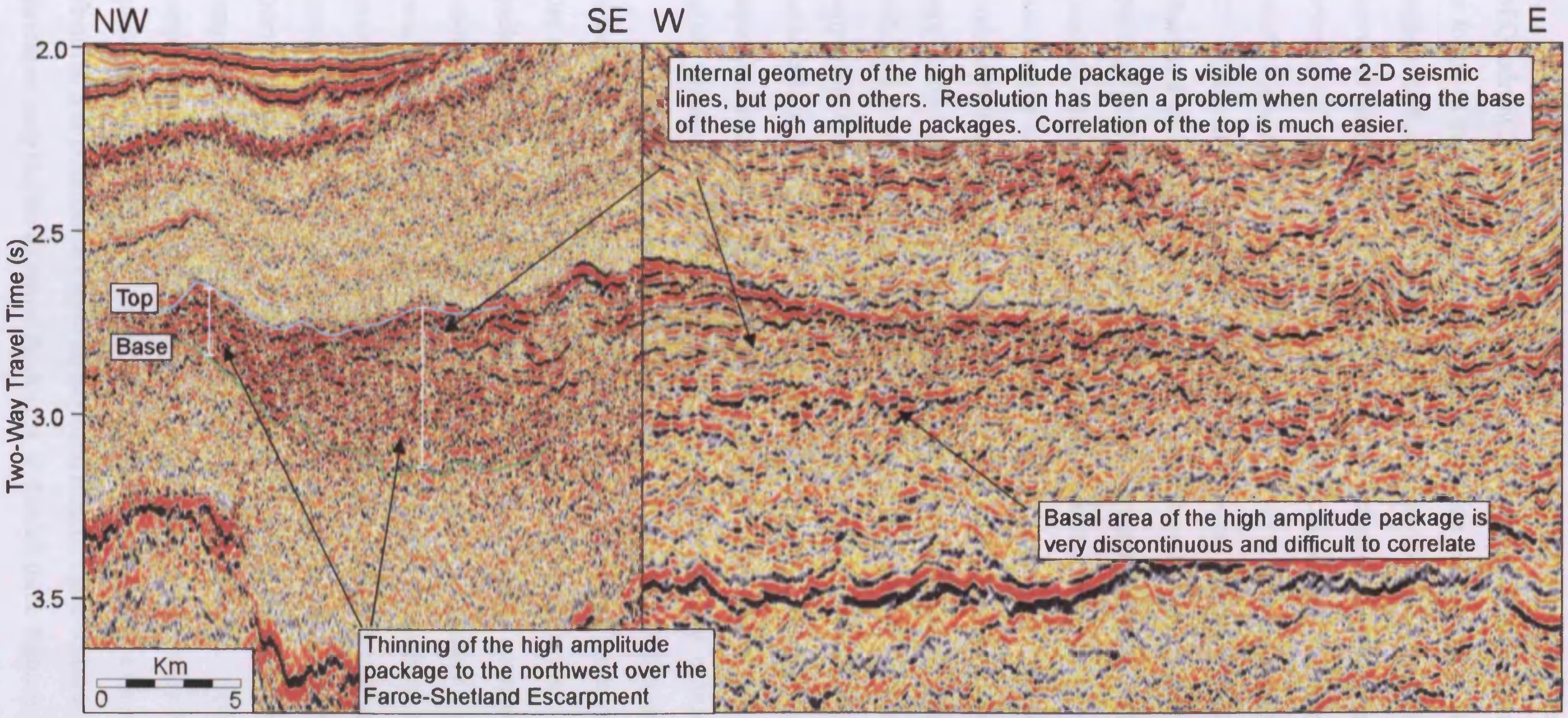


Figure 4.41. Composite 2-D line trending broadly southeast - northwest across the central high amplitude feature of the Eocene 3 unit. This seismic panel shows the thinning of the high amplitude package on the flanks, and the thick central crestal part of the feature. There is a significant thinning seen towards the northwest over the remnant Faroe-Shetland Escarpment and thus is interpreted to provide a bathymetric control on deposition of sediments. The internal architecture of the high amplitude feature is often seen on certain 2-D seismic lines, but is not visible on others and this composite line displays the change in resolution between lines. See Figure 4.33 for location of seismic line.

Finally, a detailed description of the eastern feature located in the western part of Quadrant 214 will follow. Well 214/17-1 is located in the centre of this feature and is found in the crestal area of the high amplitude package. It is observed that this eastern feature shows slightly steeper flank dips than the central feature discussed above (**Figure 4.42**). The flanks also show a degree of symmetry not seen in the central high amplitude feature. At the well location, this high amplitude feature has a relatively narrow width (in the region of 6 - 7 km), and this compares to a maximum of 25 - 30 km as seen from the structure maps shown in **Figures 4.37** and **4.38**. Again there is a similarity with the other two features where the base of the high amplitude package is difficult to map on 2-D seismic data and shows a much more diffuse character than the top. This central elongate feature turns to the west at its northern limit in the far north of Quadrant 214. At the location of well 214/4-1 there is a very high amplitude horizontal reflection which is seen close to the top of the high amplitude feature (**Figure 4.43**). This is a good example of a "flatspot" and it is believed to be a direct hydrocarbon indicator. Away from the crest of the high amplitude feature and flatspot, the internal structure of the high amplitude package can be well imaged. However, directly under the flatspot, there is very little seismic character and no continuous reflection configurations are visible. This may be a result of the presence of hydrocarbons above causing attenuation of the seismic data.

As has been discussed above, there is a degree of difficulty when trying to interpret the base of these high amplitude packages. The bases differ markedly from the tops which appear as high amplitude, continuous reflections. However, from looking at the gross seismic character of the package and comparing it with the underlying lower amplitude character it is possible to produce a good estimate of the near basal reflection of the high amplitude packages. This base is not intended to be an accurate pick on the base of each high amplitude feature but represents a near basal zone above which the high amplitude semi-continuous to discontinuous reflections occur. **Regional correlations C, D** and **E** show that the basal zone of the high amplitude features are diffuse and not accurately mappable. However, the high amplitude seismic configurations can be differentiated from the lower amplitude package below. By producing three individual bases for the three high amplitude features it is possible to create an isochron map of these features (**Figure 4.44**). This isochron map shows a central thick area within all three features, with thicknesses of the features varying form 0 - 200 ms. Generally, there is a thinning to the flanks of the

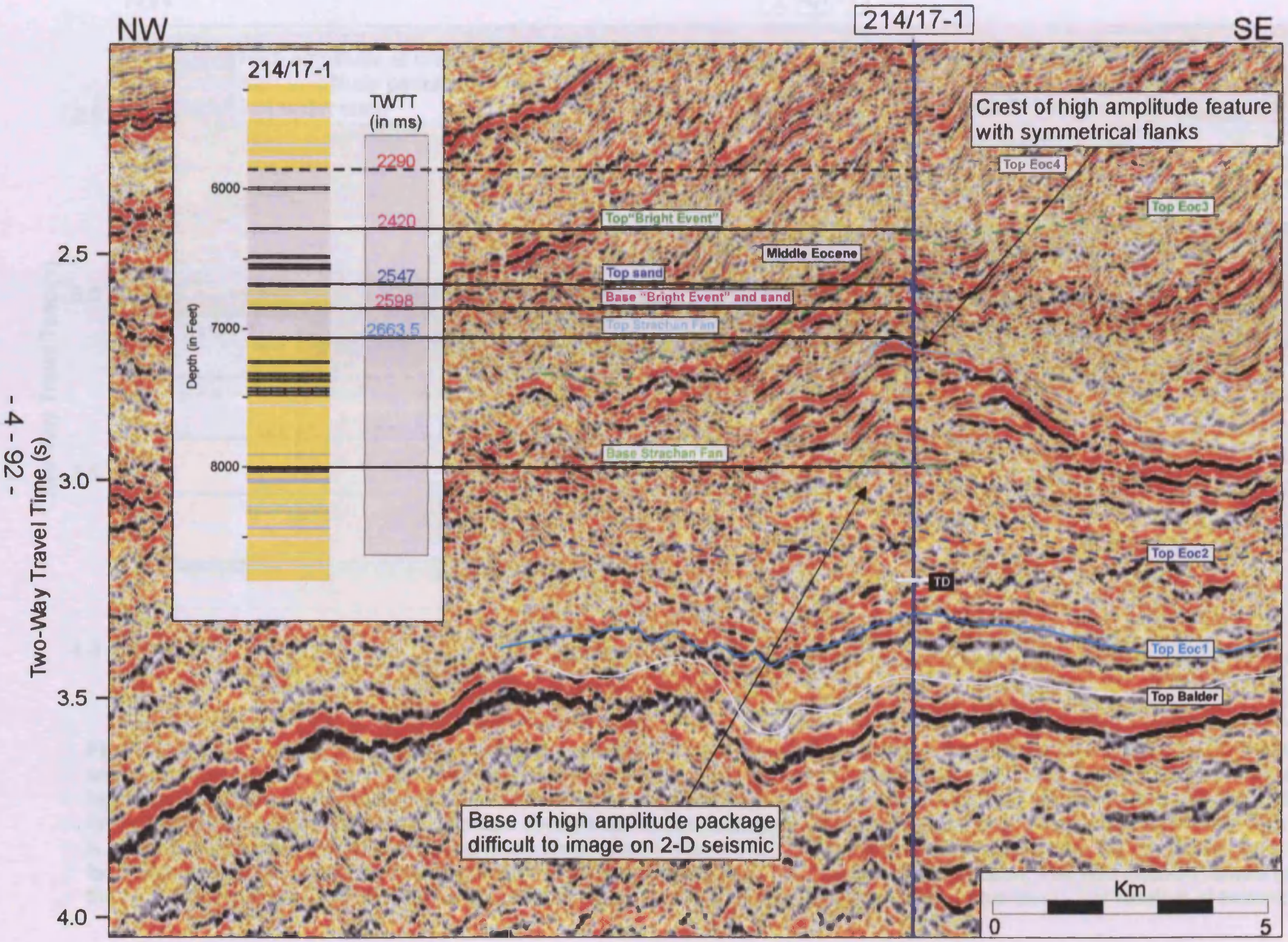


Figure 4.42. Southeast - northwest trending 2-D seismic line through the well 214/17-1 showing the central intra - Eocene 3 high amplitude feature. The well goes directly through the crest of the high amplitude feature and this seismic line shows there are symmetrical flanks to the feature. The top of the high amplitude package is characterised by a high amplitude reflection. The inset diagram shows the lithology through the Strachan Fan the top reflection can be calibrated to the top of the sandstone interval. The base of the high amplitude feature is difficult to see on 2-D seismic data. However, it is still marked on from the well, and may show no change in seismic amplitude because of the presence of additional thin sands below the fan base. Approximate position of regional seismic markers is shown. See Figure 4.33 for line location.

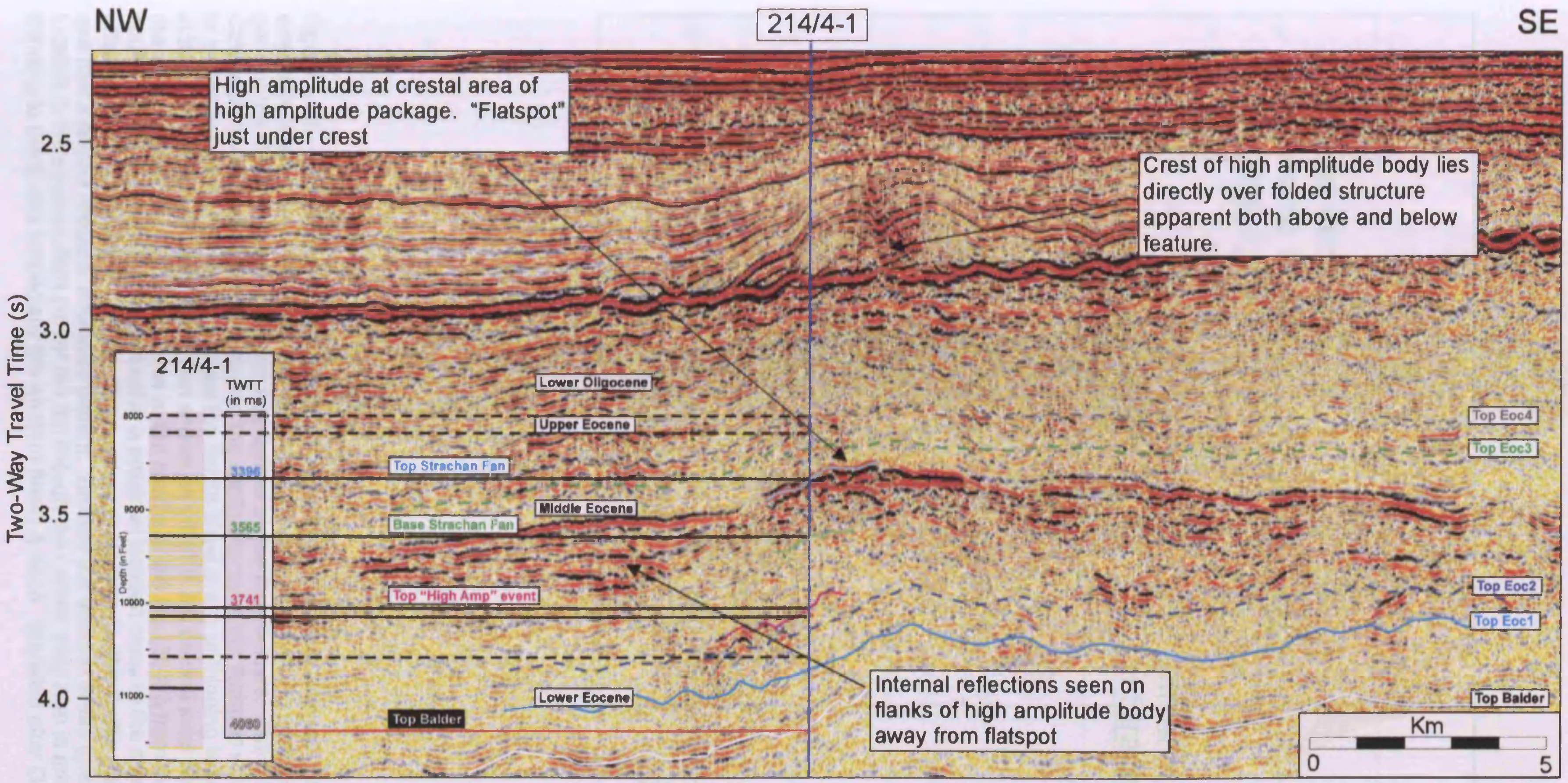


Figure 4.43. 2-D seismic line trending southeast - northwest across the northern edge of the eastern high amplitude feature through well 214/4-1. The crest of the feature has a very high amplitude at its top reflection and appears to show a characteristic "flatspot" just underneath the top reflection. The high amplitude top corresponds to the top of the Strachan Fan seen in the inset figure. Away from the crest of the structure, the flanks show continuous to semi-continuous internal seismic reflection configurations not seen under the crest, particularly towards the northwest. The base of the Strachan Fan is not well imaged on seismic data and is much more diffuse than the top. Thinner sands are visible from the composite log and these correspond to local high amplitude reflections seen between the fan base and the top T50 (Balder Tuff) reflection. The well location, drilled on the crestal structure of the high amplitude fan which lies on an anticlinal structure, seen up to the Neogene stratigraphic level. For location of seismic line see to Figure 4.33.

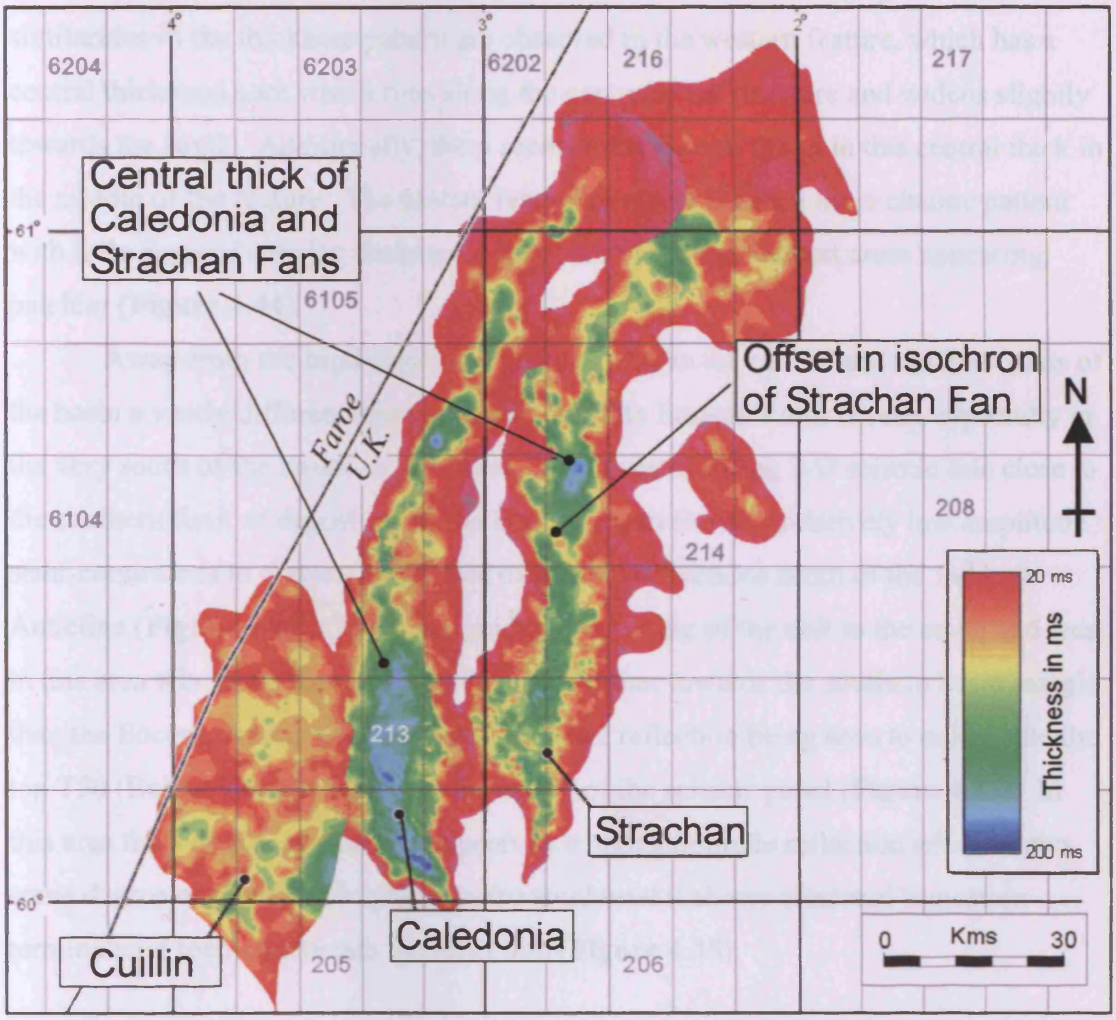


Figure 4.44. Isochron map (thickness in time) showing the three elongate Eocene 3 high amplitude fans in the northern part of the basin. This map shows the locations of the thickest parts of the individual fans which are highlighted by the blue colours. The thickest part of the Caledonia and Strachan Fans is seen in the central cores of the features where 200 ms is seen to be the maximum. In these two fans the flanks show areas of thinning to below 20 ms (shown in the red colours). The Strachan Fan shows an offset in the central axial part of the fan and it reaches its maximum thickness close to the middle of the fan (50 km from the canyon in the south). The Caledonia Fan is thickest in a proximal location close to the southerly tip and no offset is seen in the axis of the fan. The western most feature (the Cuillin Fan) is less elongate and has a patchy irregular thickness pattern. Broadly the thickest areas (green colours) are located in the southeastern part of the fan though the central area also is relatively thick and a thinning to the flanks (especially the western flank) is seen. Modified after Davies *et al.* (2004).

features and towards the north. In the central feature an oval shaped thick area located towards its southern reaches which reaches a maximum time thickness of 200 ms. This central thick area narrows into an elongate area towards the north **(Figure 4.44)**. Some similarities in the thickness pattern are observed in the western feature, which has a central thickened core which runs along the centre of the structure and widens slightly towards the south. Additionally, there seems to be a small offset in this central thick in the middle of the feature. The eastern feature however shows a more chaotic pattern with little signs of a major thickened central core with the thickest areas appearing patchier **(Figure 4.44)**.

Away from the high amplitude features seen in the central and northern parts of the basin a vastly different seismic character of the Eocene 3 unit is seen, especially in the very south of the basin. A northeast – southwest trending 2-D seismic line close to the southern limit of deposition of the Eocene 3 unit shows a relatively low amplitude semi-continuous to discontinuous area of seismic reflections south of the Judd Anticline **(Figure 4.45)**. There is a gradual thickening of the unit to the south and west in this area where the Eocene 3 unit is found further towards the southern basin margin than the Eocene 2 unit, with the basal Top Eoc2 reflection being seen to onlap onto the top T50 (Balder Tuff) reflection in the centre of the seismic panel **(Figure 4.45)**. In this area the Top Eoc3 reflection appears as a high amplitude reflection which shows some degree of relief and in places to the southwest it shows erosional truncation terminations seen underneath the reflection **(Figure 4.45)**.

4.3.4.5 Well and Borehole Data

This section discusses the wells which penetrate the Eocene 3 unit, and in particular it highlights wells in the central and northern part of the basin. As has been discussed above **(Section 4.3.4.4)** there are four wells drilled in the areas that are covered by the elongate high amplitude features. The wells that have been drilled here are recent wells and have specifically in some cases targeted the Eocene and Palaeocene section. These wells constitute some of the type wells discussed in **Section 3.5.2** as they have used a narrow drill and casing diameter of 12 ¾ inches to give accurate well data. Furthermore, these wells have high resolution biostratigraphic data which used sidewall core and remote operated vehicle (ROV) sampling techniques. A summary of the lithological information available from the four wells (three from

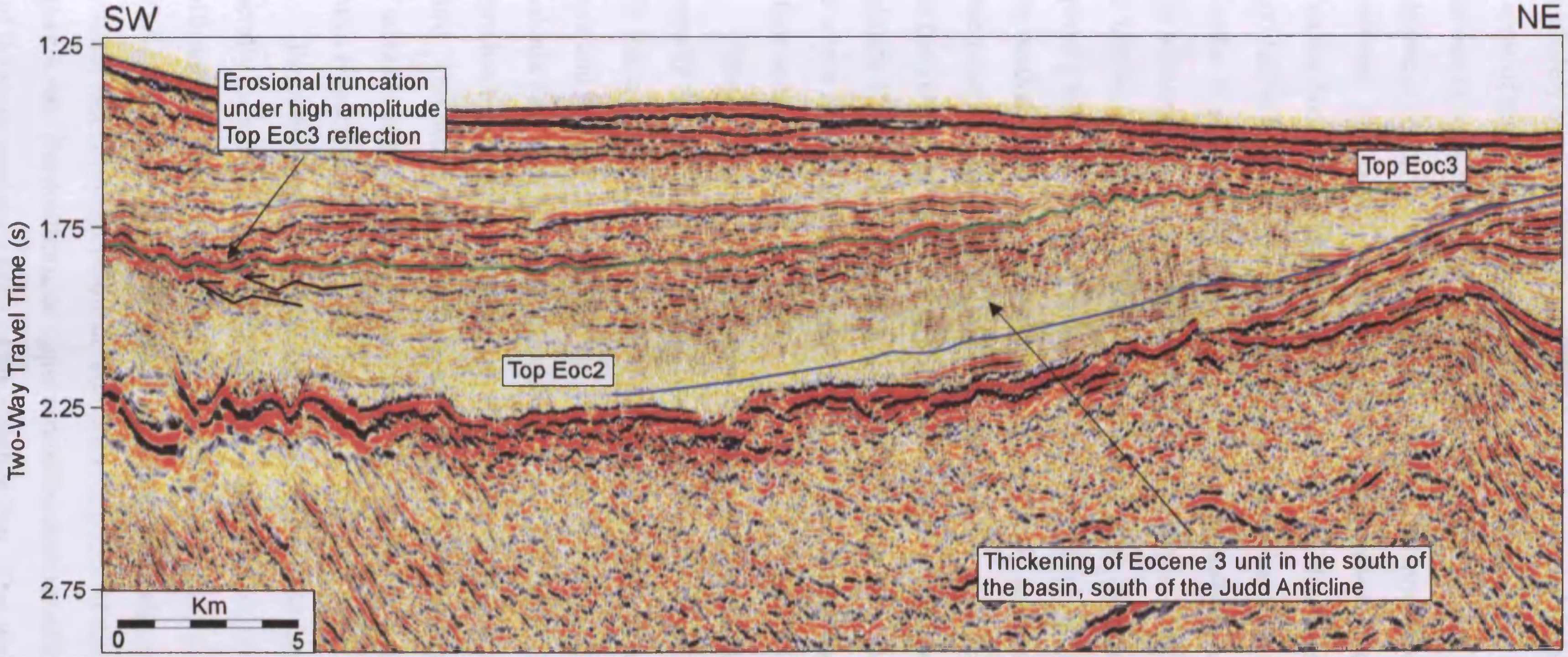


Figure 4.45. Southwest - northeast trending 2-D seismic panel illustrating the thinning and onlap of the Eocene 3 seismic unit onto the southern flank of the Judd Anticline. This stratal relationship demonstrates that there was folding on the anticline prior or during the deposition of the Eocene 3 unit. Towards the southwest the Top Eoc3 reflection becomes high amplitude, continuous and slightly folded and sits at the base of a high amplitude package. See Figure 4.33 for location of seismic line.

Quadrant 214 and one from Quadrant 213) can be seen in **Figure 4.46.** What is immediately clear is the apparent abundance of a sand-rich facies which corresponds to the areas of high amplitudes seen on the seismic data. All of the four wells penetrate sandstone in what the composite logs date as Middle Eocene, and in most cases the sandstone is usually entirely encased in Lower and Upper Eocene shales and mudstones. The sandstone intervals make up a significant proportion of the facies, and individual beds of sandstone vary from well to well. However, a maximum thickness of 200 feet is common, though the range of thicknesses varies down to a few feet (in particular in well 214/4-1). The sharp transition from shales to the top of the sand-rich facies accounts for the very high amplitude reflection seen at the upper boundary of the high amplitude features. The general elongate north – south orientation and the deepening to the north, combined with the thinning to a high amplitude incised feature at the southern tips of the features suggest a southerly source for this clastic material. Furthermore, the easternmost feature of the three has a broad circular lobe-like northern tip in the vicinity of 214/4-1 and this may suggest a distal lobe termination of the high amplitude feature. The summary of the well lithologies and ages is shown in **Figure 4.46** where the nomenclature of the sand-rich intervals packages has been taken from the four composite logs.

The sandstone intervals of the eastern feature have been grouped together and informally termed the Strachan Fan by ExxonMobil (who drilled two of the three wells in the feature). However, Conoco drilled well 214/26-1 in the southern tip of this feature and termed it the Lewis Fan. The central feature is informally termed the Caledonia Fan (courtesy of ExxonMobil who drilled 213/23-1). The western-most feature has been informally termed the Cuillin Fan, though no wells penetrate this feature. This study will use the more widely accepted Cuillin, Caledonia and Strachan Fan nomenclature from this point on when discussing the western, central and eastern features respectively.

From the well information, it is possible to see the base of these fan features. Generally the sand-rich interval is encased in shales or claystones hence giving rise to a significant change in acoustic impedance at the top of the sandstone. However, in well 214/17-1 there appears to be a significant amount of sandstones and interbedded claystones and siltstones below the composite log pick of the base of the Strachan Fan (**Figure 4.46**). Furthermore, an upper sandstone dominated interval is seen above the top of the main sandstone body of the Strachan Fan. This well differs from the other

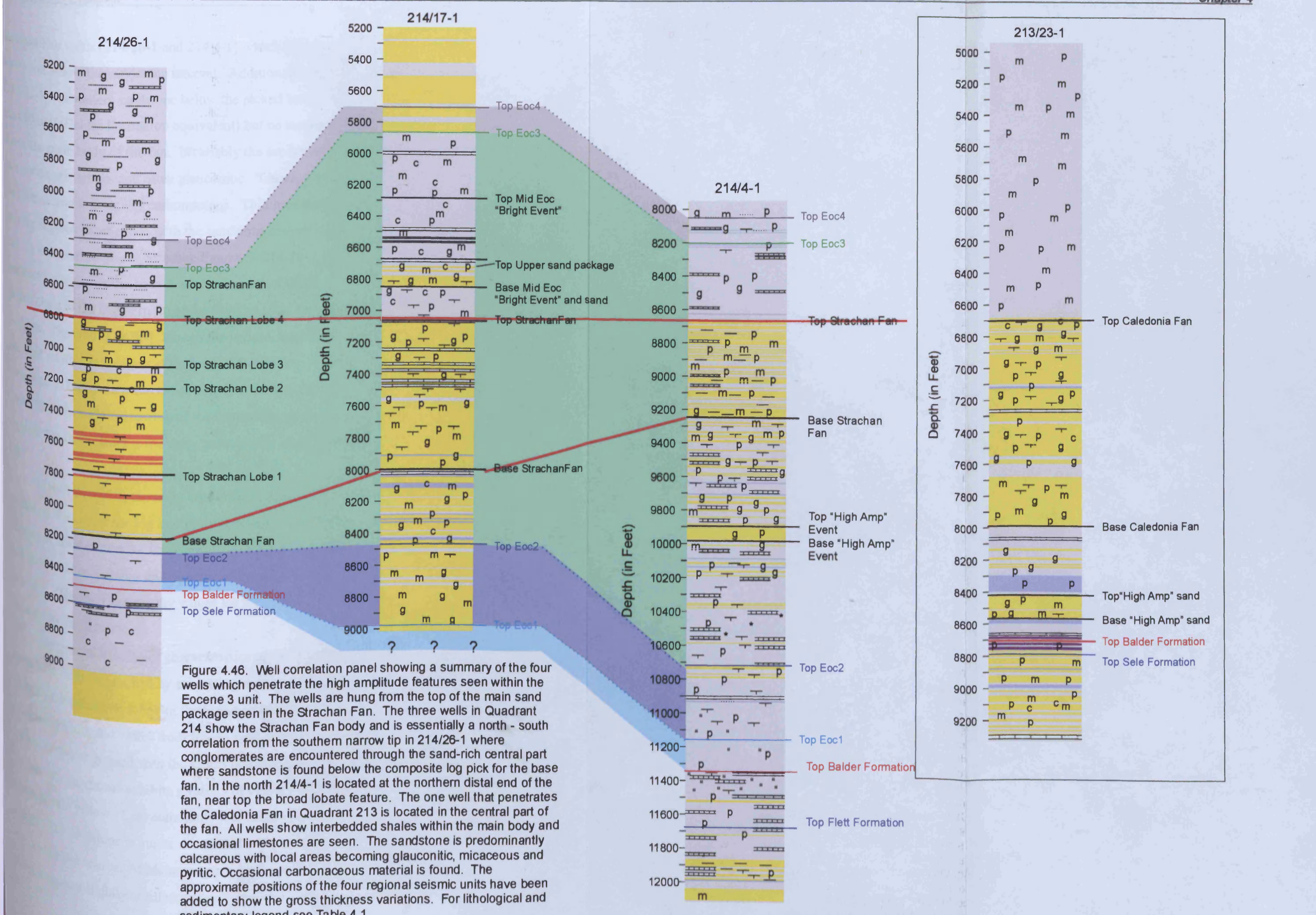


Figure 4.46. Well correlation panel showing a summary of the four wells which penetrate the high amplitude features seen within the Eocene 3 unit. The wells are hung from the top of the main sand package seen in the Strachan Fan. The three wells in Quadrant 214 show the Strachan Fan body and is essentially a north - south correlation from the southern narrow tip in 214/26-1 where conglomerates are encountered through the sand-rich central part where sandstone is found below the composite log pick for the base fan. In the north 214/4-1 is located at the northern distal end of the fan, near top the broad lobate feature. The one well that penetrates the Caledonia Fan in Quadrant 213 is located in the central part of the fan. All wells show interbedded shales within the main body and occasional limestones are seen. The sandstone is predominantly calcareous with local areas becoming glauconitic, micaceous and pyritic. Occasional carbonaceous material is found. The approximate positions of the four regional seismic units have been added to show the gross thickness variations. For lithological and sedimentary legend see Table 4.1.

Strachan Fan wells (214/26-1 and 214/4-1) which have thick claystone intervals both above and below the sandstone interval. Additionally, the sole Caledonia Fan well (213/23-1) encounters sandstone below the picked base of the main fan body overlying the T50 unit (Balder Formation equivalent) but no sandstone intervals are observed above the main body of the fan. Invariably the sandstone found in these fans is calcareous, micaceous and often glauconitic. Traces of pyrite are also seen in parts of the interval and it is locally carbonaceous. Thin limestone beds or nodules are seen in varying degrees interbedded with the sandstone and are more prevalent in the Strachan Fan. In the most southerly Strachan Fan well (214/26-1) there are thin isolated (10 – 20 feet thick) conglomeratic beds towards the base of the main fan body. This well is located close to the southern tip of the Strachan Fan where it narrows to approximately 1 – 2 km. The well is drilled through the incised feature that shows high amplitude mounded seismic reflections at the base of the Eocene 3 unit. This feature is interpreted to represent a canyon setting located close to or at the edge of the shelf break. The southerly narrowing of the fan and the presence of conglomeratic beds suggests that this feature is the entry point for the clastic material found in the Strachan Fan.

In well 214/26-1, the main sandstone interval of the Strachan Fan has been sub-divided on the composite log into four lobes separated by thin 10 – 30 feet thick siltstone or shale intervals. The gamma ray signature for this interval of the 214/26-1 well shows a classic fining up sequence of the sandstone (**Figure 4.47**) throughout the four lobes. The high gamma ray values for the intervening shales and siltstones are also seen on this figure.

The general lithofacies characteristics of the sand-rich facies within the fan bodies have been ascertained by evaluating the composite logs of the four wells. Generally the sandstone found in the Strachan and Caledonia Fans is clean, occasionally loose and varies from fine to very coarse. However, it is more commonly medium to coarse grained with the grains being dominantly sub-rounded to sub-angular. The sandstone exhibits a variety of colours though it is often translucent, pale orange or pale yellow. Less commonly it is locally reddish, greenish and occasionally pinkish. The sandstone is quartz dominated with feldspar and lithic fragments which have a bi-model source. Moderate to poor sorting of the sandstone is common and in places it appears slightly argillaceous with mudstone clasts. However there is generally

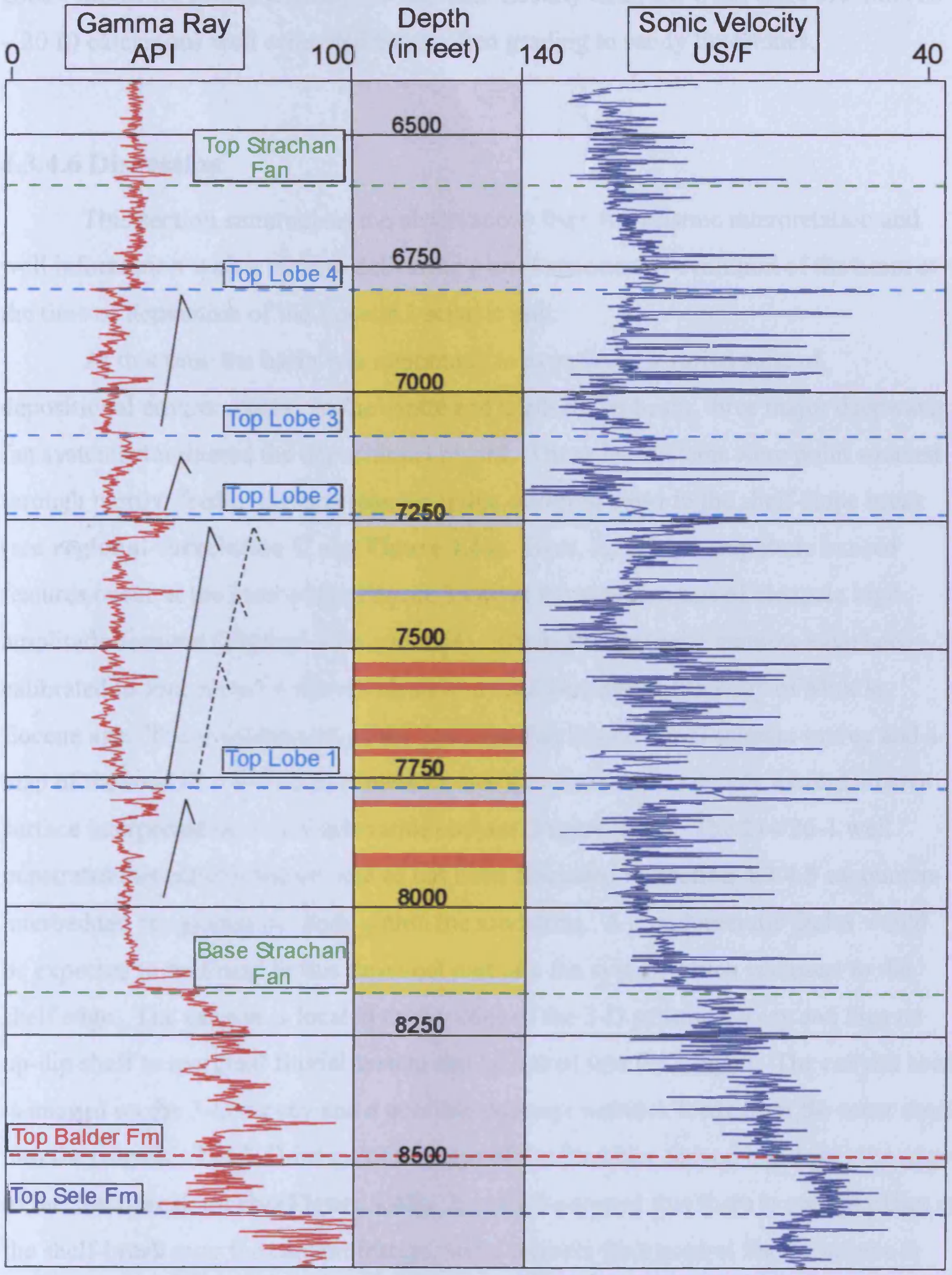


Figure 4.47. Gamma ray and sonic velocity wire-line log of the 214/26-1 well located at the southern tip in the canyon of the Strachan Fan. The gamma ray log on the left (red) nicely shows the high gamma ray values for the thin siltstones and mudstones which sit at the top of individual units (or lobes) of sandstone and conglomerate. Four lobes have been defined and identified (by Conoco Ltd.) from the composite log and have fining up packages seen in the gamma ray response. The conglomerate is confined to the lower two lobes and the upper two lobes are dominated by sandstone, suggesting that through time the deposits of the fan system became more distal. The main fan body is entirely enclosed in mudstones and siltstones and is dated at Middle Eocene in age (Eocene 3 seismic unit) by micropalaeontological data. For lithological legend see Table 4.1.

good - moderate visible porosity throughout. Locally in all the wells there are thin (10 – 20 ft) calcareous well cemented bands often grading to sandy limestones.

4.3.4.6 Discussion

This section summarises the observations from the seismic interpretation and well information with a view to delivering a brief synopsis of evolution of the basin at the time of deposition of the Eocene 3 seismic unit.

At this time the basin was continuing to experience a varied suite of depositional environments. In the centre and north of the basin, three major deepwater fan systems dominated the depositional record. These fan systems were point sourced through narrow feeder systems (canyons) that occurred close to the shelf-slope break (see **regional correlation C** and **Figure 4.44**). Here, local high amplitude incised features occur at the base of the Eocene 3 unit at the southern tips of elongate high amplitude features (**Figures 4.36** and **4.44**). The high amplitude features have been calibrated to four recent wells which show a sand dominated lithology of Middle Eocene age. The southern part of the Strachan Fan lies on a 3-D seismic survey and a map of the base of the high amplitude incised feature reveals a deeply incised erosive surface interpreted here as a submarine canyon (**Figure 4.48**). The 214/26-1 well penetrates this canyon feature and as has been discussed in **Section 4.3.4.5** encounters interbedded conglomeratic beds within the sandstone. A conglomeratic facies would be expected to be found in this proximal part of a fan system which is closest to the shelf edge. The canyon is located on the edge of the 3-D seismic survey and thus no up-dip shelf to marginal fluvial system can be linked into the canyon. The canyon head is imaged on the 3-D survey and a possible drainage network feeds from the outer shelf into the canyon. The 3-D image of the base of the Strachan canyon highlights the edge of the shelf-break nicely (**Figure 4.48**). It could be argued that there is a small offset in the shelf-break over the canyon feature, and a tectonic fault control for the canyon is one possible mode of formation and cannot be ruled out. A small crestal graben occurs in the overlying strata above the canyon (**Figure 4.48**) which could be caused by differential compaction of the sediments.

At the time of deposition of the Eocene 3 unit the Shetland margin was dominated by a shelf environment which in places would have been sand-rich allowing for the redistribution into the fan feeder systems and eventually to the basin floor.

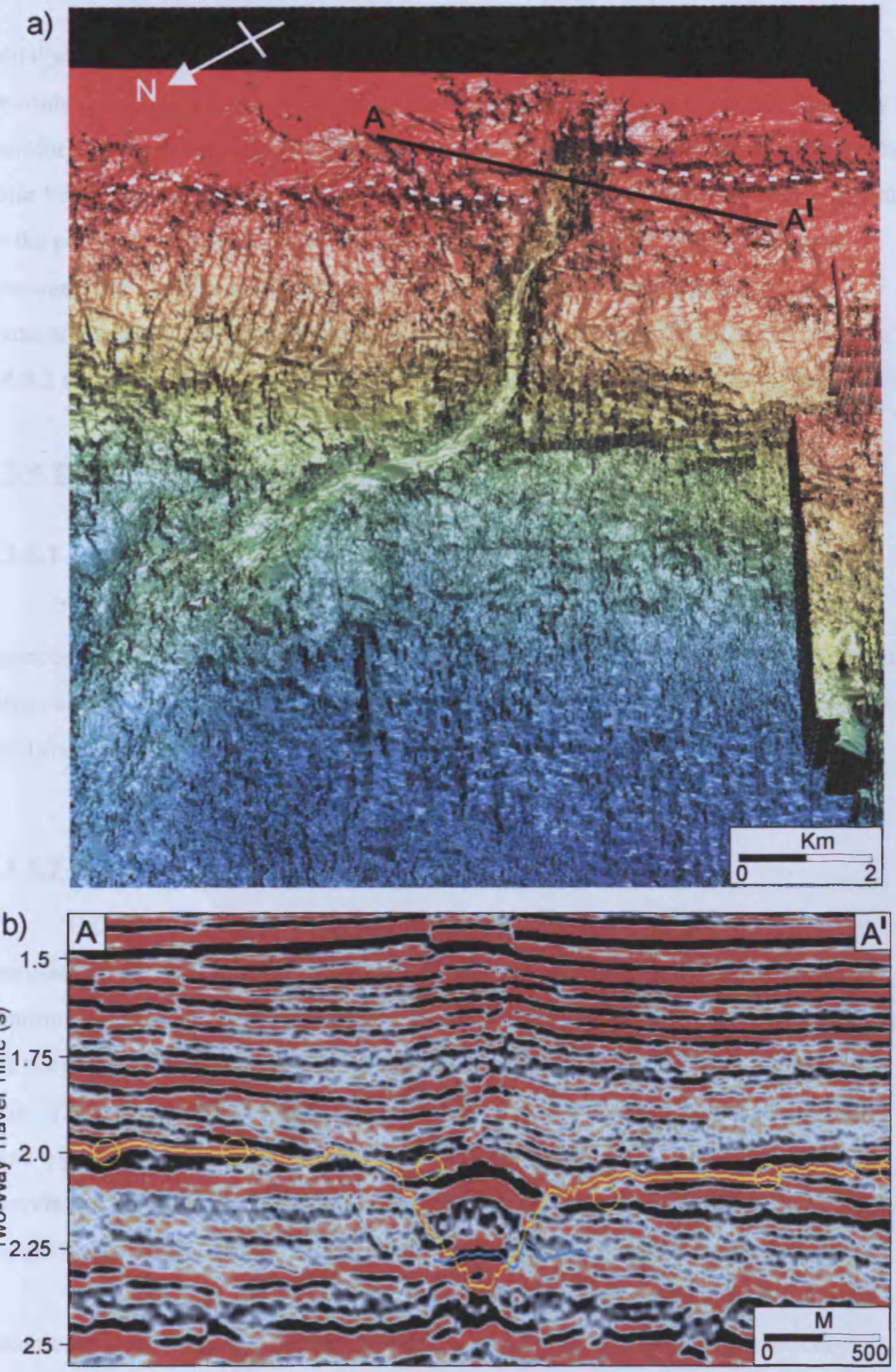


Figure 4.48. a) 3-D perspective view of the seismic reflection at the base of the Strachan canyon located at the southern tip of the Strachan Fan. The canyon trends roughly northwards for approximately 5 km from its proximal position on the shelf-slope break. Note the apparent offset in the shelf-slope break (dashed white line) either side of the canyon which may indicate a fault controlled origin to the position of the canyon. On the shelf, a drainage pattern seems to feed into the head of the canyon. b) 3-D seismic line (position shown in a), showing strike line through the upper reaches of the Strachan canyon (Yellow horizon indicates base of canyon mapped in a). High amplitude mounded seismic reflections appear in the canyon feature and a crestal graben forms in the overlying sediments above the canyon.

Initial analysis of the biostratigraphic reports of these four wells suggest possible Devonian and Jurassic reworked fauna found in the sandstone of the fans. This may therefore imply a southeasterly hinterland as the source area, and perhaps suggests the older West Shetland Basin. The reworking of sand along the margin may be confirmed by the presence of varied sandstone and siltstone lithologies in wells in the north of Quadrant 208. A palaeogeographic reconstruction of the Eocene 3 seismic unit summarising the main environments of deposition is shown in **Figure 4.49**. **Section 4.4.3.2** continues this discussion with respect to the basin evolution during the Eocene.

4.3.5 Eocene 4 Seismic Unit

4.3.5.1 Introduction

The Eocene 4 seismic unit represents the last of the four regional Eocene seismic-stratigraphic units described in this chapter. This section will detail the seismic observations, the geographical extent of the unit and evaluate the internal architecture and lithology of the unit.

4.3.5.2 Upper and Lower Boundaries

This unit is bounded at its base by the Top Eoc3 reflection which has been discussed in **Section 4.3.4.2**. This reflection appears exhibits a high amplitude continuous character in many places in the south of the basin and on Shetland margin, but becomes lower in amplitude and semi-continuous to discontinuous in the basin axis. The correlatability of the reflection is poor in much of the central and northern parts of the basin where it is seen to be is pervasively faulted (e.g. **regional correlations D** and **E**). This poor correlation of the Top Eoc3 reflection continues to the northwest on the Faroe margin, where no continuous reflections appear.

The upper boundary of the unit is the Top Eoc4 reflection and is seen on all seismic lines as a dark grey horizon. In the south of the basin it appears as a high amplitude continuous reflection that lies above a high amplitude chaotic package (e.g. see **regional correlation A**). To the southeast, this reflection becomes coincident with Top Eoc3 reflection and to the northwest it is truncated by a later unconformity which is very close to the sea-bed. In the central and northern parts of the basin centre the Top Eoc4 reflection is not a well correlatable seismic reflection and appears as a low to

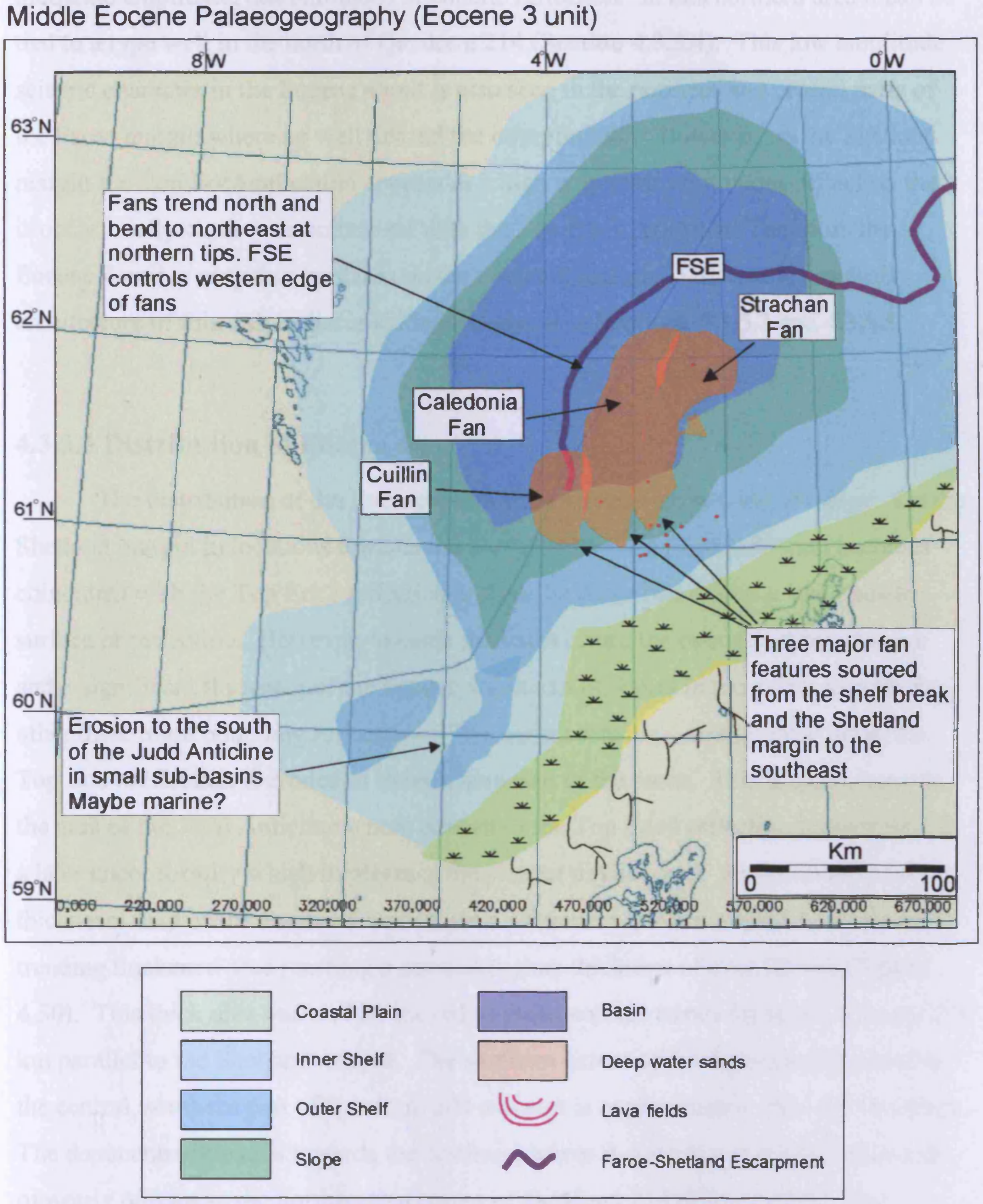


Figure 4.49. Schematic map showing the palaeogeography of the basin during the Middle Eocene at the time of deposition of the Eocene 3 unit. The basin at this time is dominated by the deposition of large clastic deepwater fan systems in the central and northern part of the basin. These fans form elongate sand-rich fairways which trend roughly north - south and stretch for over 100 km. The fans are fed from the shelf-break area where canyons systems are developed. These canyons funnel sand from the shelf area into the deeper basin floor. In the proximity of the canyons conglomerates have been found in wells with evidence of slumping and mass flow. This may represent debris flows caused by instability on the edge of the shelf which may have collapsed, causing some form of remobilisation of sediment. In the south of the basin there was little clastic sediment supply reaching the basin which resembled an outer shelf setting. However, some evidence of erosion is seen south of the Judd Anticline and this maybe marine erosion. FSE= Faroe-Shetland Escarpment. For the legend of the colours used for the depositional environments see Table 4.1 or Enclosure M.

moderate amplitude, discontinuous or chaotic reflection. In this northern area it can be tied to a type well in the north of Quadrant 214 (**Section 4.3.5.4**). This low amplitude seismic character in the Eocene 4 unit is also seen in the northern and central parts of the Faroe margin where no well ties aid the interpretation. However, on the Shetland margin the Top Eoc4 reflection appears as a high amplitude continuous reflection that is occasionally close to or coincident with the Top Eoc3 reflection. Therefore the Eocene 4 unit is very thin in places on the Shetland margin. The detailed internal architecture of this unit is discussed in more detail in **Sections 4.3.5.3** and **4.3.5.5**.

4.3.5.3 Distribution of Eocene 4

The distribution of the Eocene 4 unit shows a near basin-wide coverage. On the Shetland margin in locations towards the northeast the Top Eoc4 reflection becomes coincident with the Top Eoc3 reflection and can be described as being a composite surface or reflection. However, towards the basin centre the two reflections diverge and a significant thickness of the Eocene 4 unit can be found in the axes. As with the other three main bounding reflections of the earlier seismic – stratigraphic units, the Top Eoc4 reflection is eroded in the southern part of the basin. This is again found in the area of the Judd Anticline where erosion of the Top Eoc4 reflection is expressed in a later unconformity which in places is the present day sea-bed. An isochron (time thickness) map of the Eocene 4 unit shows a prominent linear northeast to southwest trending thickened area reaching a maximum time thickness of over 500 ms (**Figure 4.50**). This thick area outlined by the red to green colours trends for approximately 250 km parallel to the Shetland margin. The southern extent of the depocentre is found in the central, southern part of Quadrant 204 where it is approximately 200 - 250 ms thick. The depocentre thickens towards the northeast where it reaches a maximum thickness of nearly 600 ms in the northeastern corner of Quadrant 214 (**Figure 4.50**). This northeast trending thickened area is very linear and is narrow in the south and central part of the basin, where it averages a width of 25 - 30 km. Towards the northeast and in the centre of Quadrant 214, the depocentre is seen to widen out and reaches a maximum width of over 100 km in the very north of the basin (**Figure 4.50**).

There is a visible offset in the depocentre which is seen in the north of Quadrant 204 and 205 (**Figure 4.50**). The offset has a left-stepping direction and is highlighted in the green colours. The size of the offset appears to be in the region of 15 - 20 km.

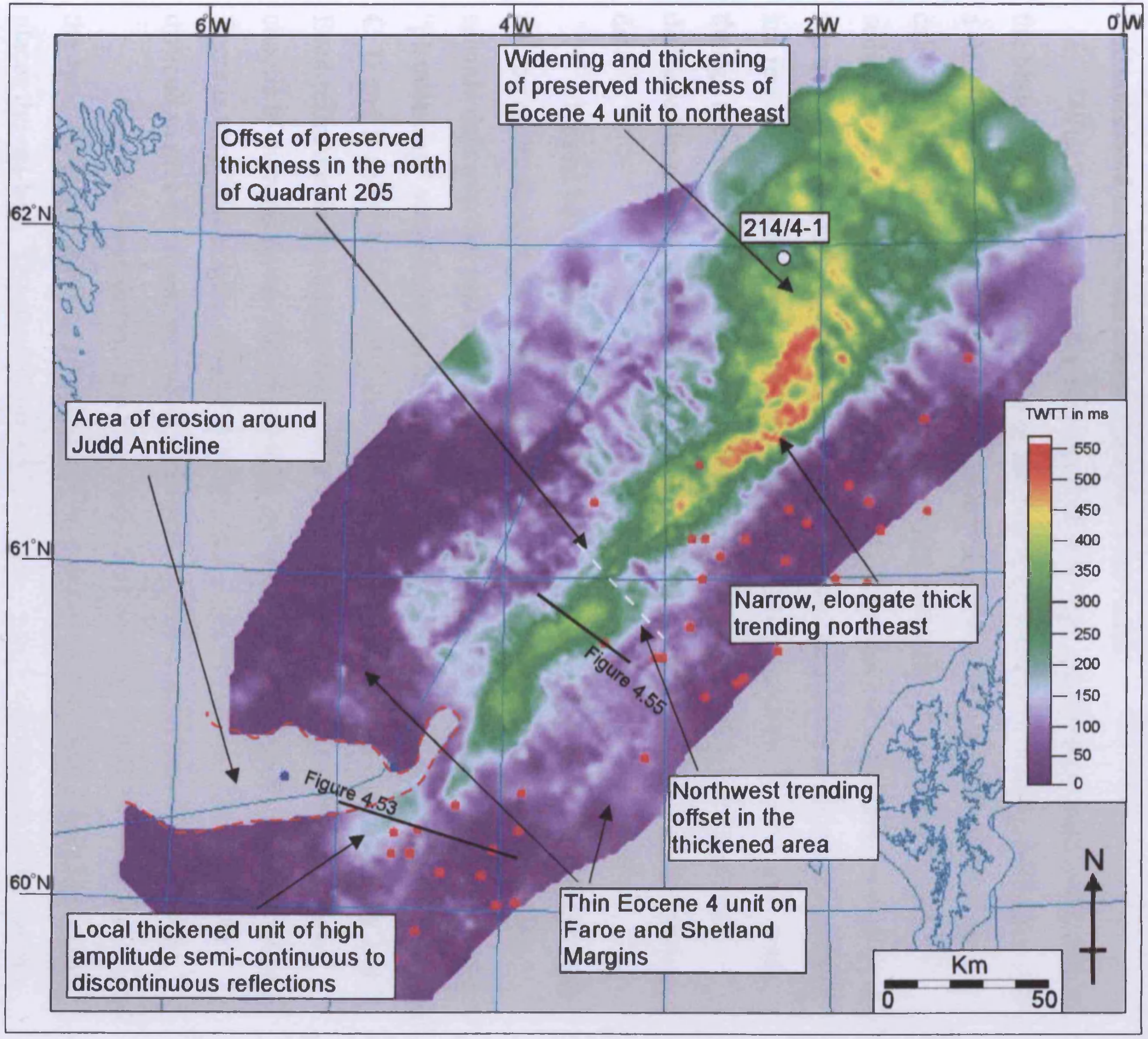


Figure 4.50. Isochron map (in TWTT ms) of the Eocene 4 seismic unit showing a very narrow northwest trending elongate area of preserved thickness in the axes of the basin. This possible depocentre is approximately 20 - 25 km wide in the south and it widens to over 100 km in the northeast. The thickest accumulation of the Eocene 4 unit is in Quadrant 214 where it reaches a thickness of over 500 ms. This narrow area is seen to be offset by 10 -15 km in the centre of the basin towards the north of Quadrant 205 (white dashed line). Both the Faroe and the Shetland margins of the basin have a very thin succession of the Eocene 4 unit and on the regional correlation panels A - E the Top Eoc4 and Top Eoc3 reflections are seen to be coincident in many places. The Top Eoc4 reflection is eroded by younger reflections and the sea-bed in the south of the basin (see red dashed line). The locations of seismic lines which discuss the Eocene 4 unit are highlighted. The location of the type well 214/4-1 is also highlighted which is a key data point for the Eocene 4 unit in the centre of the basin axes.

This offset coincides with an increase in the width of the depocentre, which can be seen to widen to the northeast of the offset zone (**Figure 4.50**).

Both the Shetland and Faroese margins show a thin accumulation of sediments of the Eocene 4 unit and generally have little more than 100 ms of section. As has been mentioned above (in **Section 4.3.5.2**) on the more proximal part of the Shetland margin, the Top Eoc3 and Top Eoc4 reflections are either coincident or close to coincident and only locally do they diverge. This is best observed from the successive strike orientated **regional correlations F, G** and **H**, where a progressive increase in thickness between the two reflections occurs. In particular on **regional correlation H**, the Top Eoc4 reflection appears to be a consistent high amplitude reflection that can be well correlated on the southeastern margin of the basin.

Well 214/4-1 provides the most northerly high quality data point in the axis of the basin to help in the correlation of the Top Eoc4 reflection. During drilling in 1999, a ROV sample grab allowed for the detailed Neogene to Palaeogene record to be calibrated in this well and gave a confident pick to the top of the Upper Eocene sediments in the basin centre for the first time. This data point has been used to tie the top of the Eocene succession (top of the Eocene 4 seismic unit) in an area were there are few wells. Indeed, this well lies over 100 km away from the nearest other wells, though these did not use an ROV to collect samples. The 214/4-1 well will be discussed in more detail in **Section 4.3.5.4** where it will be correlated to the seismic data.

It must be stated here that there are few available data points to help correlate the top of the Eocene succession and it has been purely correlated throughout the seismic data over the vast majority of the northern part of the basin by the technique of "phantoming" away from the 214/4-1 well (see **Section 3.3.2**). **Regional correlations C, D** and **E** show the difficulty that is encountered when trying to correlate the Top Eoc4 reflection in the basin centre, where the reflection occurs in a broad zone of chaotic to discontinuous reflection configuration patterns that are pervasively faulted. There is very little seismic character within the Eocene 4 unit and this has made it difficult to pick the upper bounding surface accurately in many parts of the basin.

Towards the southwestern limit of the basin, in the area close to the location of the Judd Anticline, the Top Eoc4 reflection is eroded by later unconformities and in places the sea-bed. The erosion occurs in an east - west trend, though at the eastern edge of the area of erosion there is a thin north - northeasterly spur where the Top Eoc4 reflection has also been removed close to the Judd Deeps (**Figure 4.50**).

4.3.5.4 Well and Borehole Data

As has been highlighted in the **Section 4.3.5.3** above, the most northerly well in the basin has provided a good data point for the top of the Eocene succession. This well is 214/4-1 which is drilled in the central axis of the basin and, at the time, was the deepest water well drilled on the United Kingdom Continental Shelf (UKCS) in 1999. Because of its unique location in the centre of the basin axis it encountered a near complete Neogene to Palaeogene stratigraphic record that was extensively sampled using an ROV. The findings from the biostratigraphic samples enabled Davies *et al.* (2001) to suggest a new approach for dating the initiation of the North Atlantic Deep Water (NADW) as Oligocene in age. This was based on the recognition of contourite drift bodies in the Upper Palaeogene succession which was calibrated with the good biostratigraphic data obtained from the 214/4-1 well. A summary of the latest Eocene to recent stratigraphy found in this well is summarised in **Figure 4.51**. Numerous regional studies along the Northwest European Margin exist outlining the Late Cenozoic and in particular the Neogene stratigraphy and the associated sedimentary architecture. These studies have provided a useful datum for the sediments that overlie the interval of interest in this study which sees a change from major progradational systems in the Eocene to more axially fed contourite drift systems that dominate the post-Eocene succession (Stoker 1997; 1998).

4.3.5.5 Internal Geometry

The Eocene 4 unit shows a very varied internal seismic geometry from one part of the basin to another. It has already been mentioned that over much of the northern part of the basin there is very little seismic character and very few internal seismic reflections can be correlated over any distance. In the far north of the basin the 214/4-1 well was drilled in the very centre of the basin axis and shows that above and below the high amplitude fan body of the Eocene 3 seismic unit (see **Section 4.3.4.4**) there is very low amplitude chaotic and discontinuous reflection configurations (**regional**

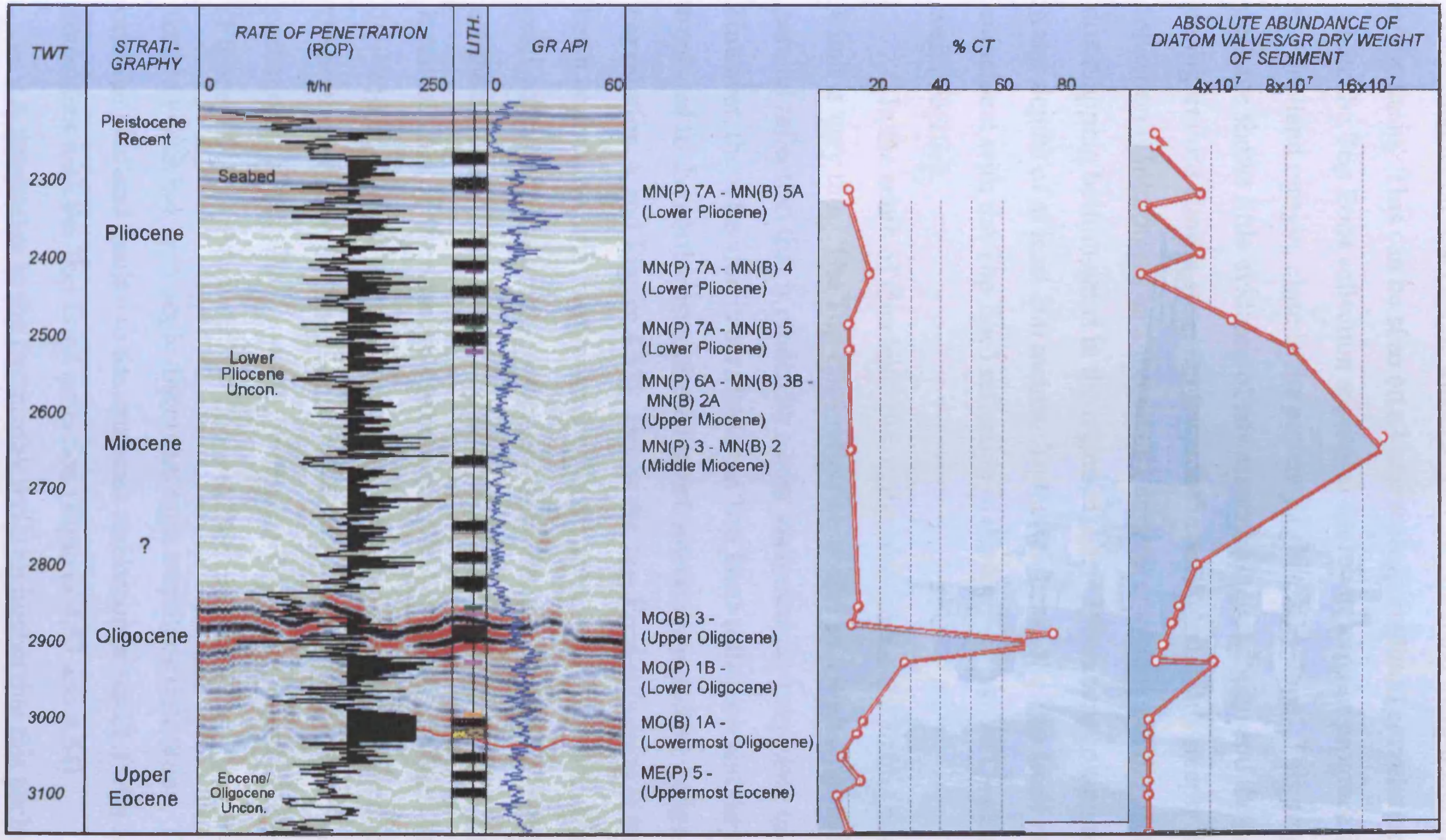


Figure 4.51. Stratigraphic summary of the results from the Neogene and Upper Eocene succession from well 214/4-1. This well drilled an almost complete succession in the basin centre and penetrated Upper Eocene strata just below 3100 ms. Also highlighted is the corresponding well to seismic tie which shows the change in seismic character between the Neogene and the Palaeogene either side of the diagenetic Opal A/CT transition boundary. Note the increase in the rate of penetration (ROP) throughout the hard Opal A/CT layer and the dramatic increase in the % of CT. This well has been used in this study to give a good correlation of the top of the Eocene succession in the northern part of the basin in an area with little well control. For well location see Figure 4.37. From Davies *et al.* 2001)

correlation E). This corresponds to a predominant claystone succession in the well which encases the sand-rich fan package (**Figure 4.52**).

The internal seismic reflection configuration of the Eocene 4 unit becomes more visible and much less chaotic towards both the Shetland margin and the southern part of the basin. This can be seen on a basin scale in **regional correlations B** and **A,** where the Top Eoc4 reflection appears at the top of a major progradational package on the Shetland margin, close to the present day break in slope. This progradational package shows little evidence of any associated aggradation and the reflection configurations show a steep dip towards the northwest. This steeply dipping part of the reflection is interpreted as a foreset of a clinoform system and the depth to the downlapping bottom-set is in the region of 200 – 250 ms suggesting progradation into water depths of at least 200 metres. To the southeast, the Top Eoc4 reflection becomes coincident with the Top Eoc3 reflection in the northern part of Quadrant 205 close to well 205/20-1.

In the south of Quadrant 204, in the area around the Judd Anticline, the Eocene 4 unit is very thin. The Top Eoc4 reflection occurs as a high amplitude continuous parallel reflection that is eroded by a later unconformity very close to the sea-bed. However, the reflection is parallel to the Top Eoc3 reflection and they both become truncated to the northwest under a distinct younger prograding wedge (see **regional correlation A** and **Figure 4.53**). Below the Top Eoc4 reflection in the south of the basin on the Shetland margin is a localised area of high amplitude discontinuous to semi-continuous reflection configurations that thins towards the margin (towards the southeast) and thickens towards the northeast and is truncated (**Figure 4.53**). This localised area of discontinuous reflections also has a northeast trend and appears as a 20 – 25 km elongate area, with a width of approximately 10 km and is located just to the northwest of wells 204/22-1 and 204/23-1 (**Figure 4.54**). This local high amplitude package thins and onlaps against the dipping Top Eoc3 reflection to the southeast (**Figure 4.54**) close to well 204/23-1 and then re-thickens into a small lens-shaped feature which has high angle discordant high amplitude reflections. This feature thins onto the Shetland margin to the southeast and becomes parallel with and eventually coincident with the Top Eoc3 reflection (**Figures 4.53** and **4.54**).

A correlation to the two nearby wells highlights that this package consists of a moderately sorted medium grained sandstone which is colourless to white (occasionally translucent) and has rounded sub-spherical grains. The sandstone is slightly glauconitic

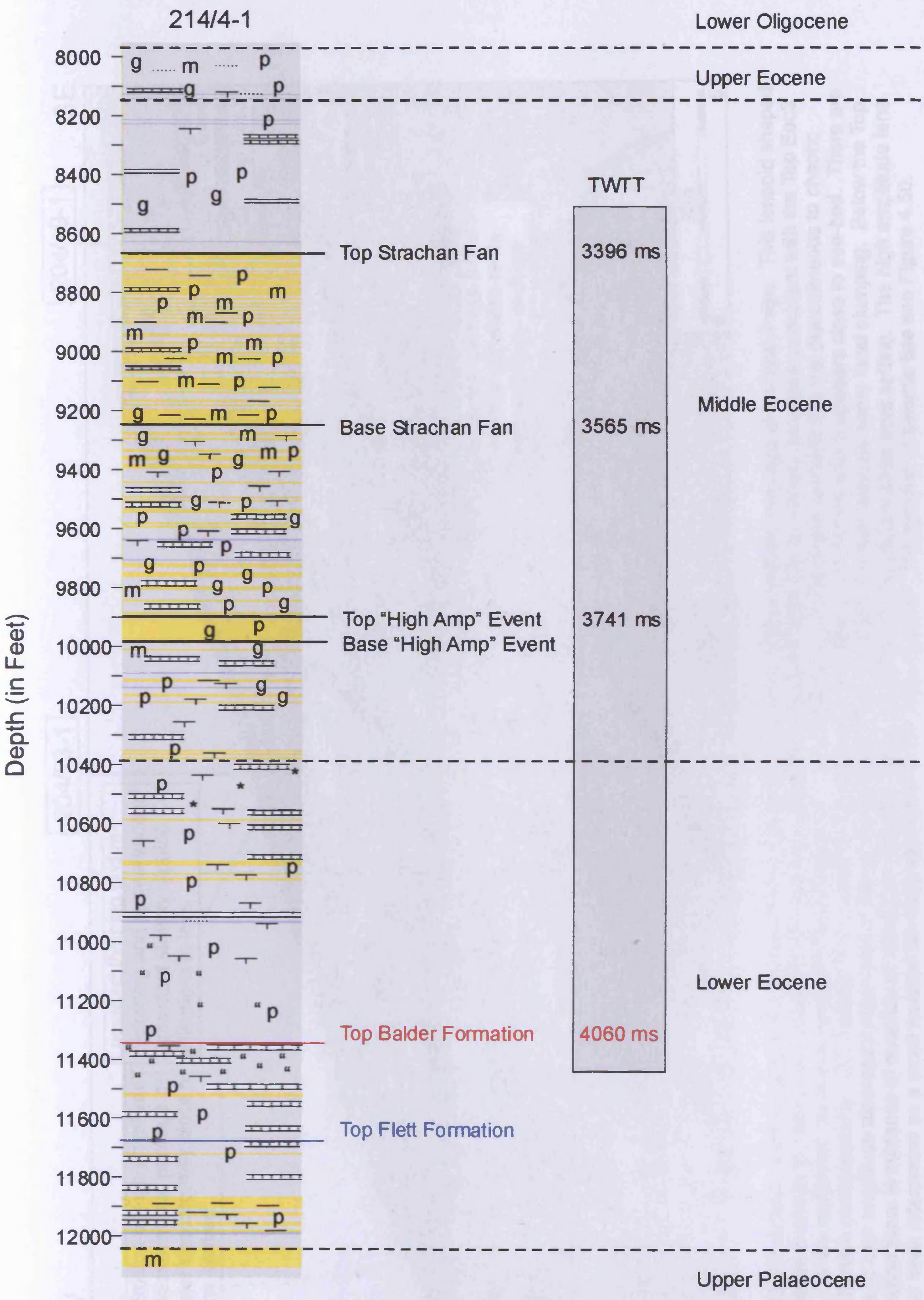


Figure 4.52. Lithological summary of well 214/4-1 (taken from the composite log and the ExxonMobil in-house biostratigraphic report - see Figure 4.50 for location) showing the sandstone to claystone ration in the main body of the Strachan Fan. This well is located at the northern distal end of the Strachan Fan and consists of thin interbedded sandstones and claystones with minor beds of limestone throughout the interval. There is also a thick sandstone interval under the base of the Strachan Fan which shows a distinctive high amplitude character on seismic data. The top of the Eocene 4 unit is provided by the biostratigraphic information from this well. For lithological and sedimentary legend see Table 4.1.

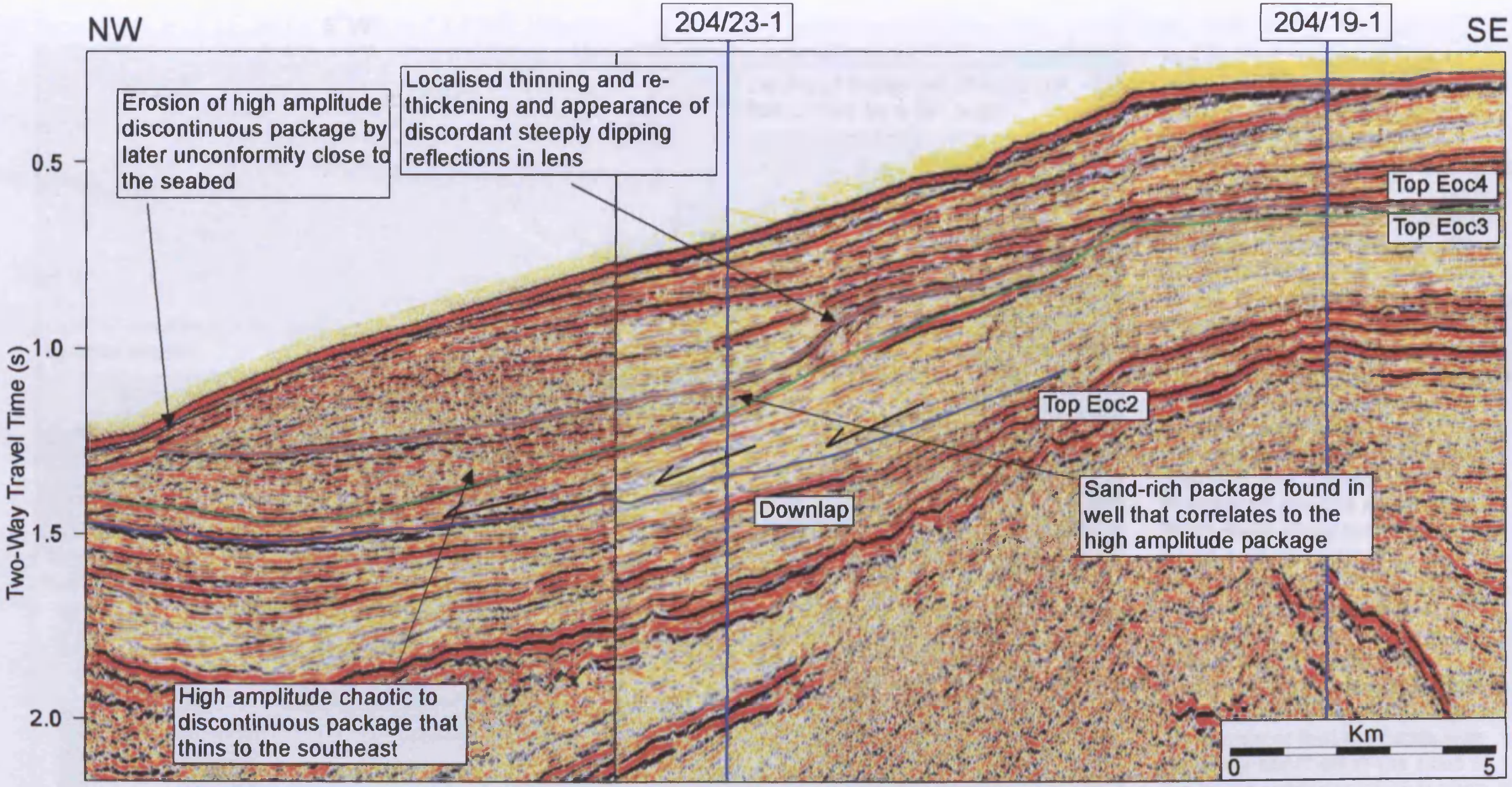


Figure 4.53. Southeast - northwest trending 2-D seismic line showing a localised high amplitude package of Eocene 4 age. This lensoid shaped high amplitude package is seen to thin up-dip to the southeast where the Top Eoc4 reflection eventually becomes coincident with the Top Eoc3 reflection. To the northwest, there is a significant thickening of the high amplitude package which contains internal discontinuous to chaotic seismic reflection configurations. The package is truncated to the northwest by a later unconformity which appears close to sea-bed. There are local signs of high amplitude discordant reflections in the up-dip position of the lens and this may indicate some local slumping. Below the Top Eoc3 reflection there is evidence of downlap of clinoforms onto the Top Eoc2 reflection indicating a delta front setting. The high amplitude lens feature has been interpreted as a small isolated fan body deposited at the base of slope. For location of seismic line see Figure 4.50.

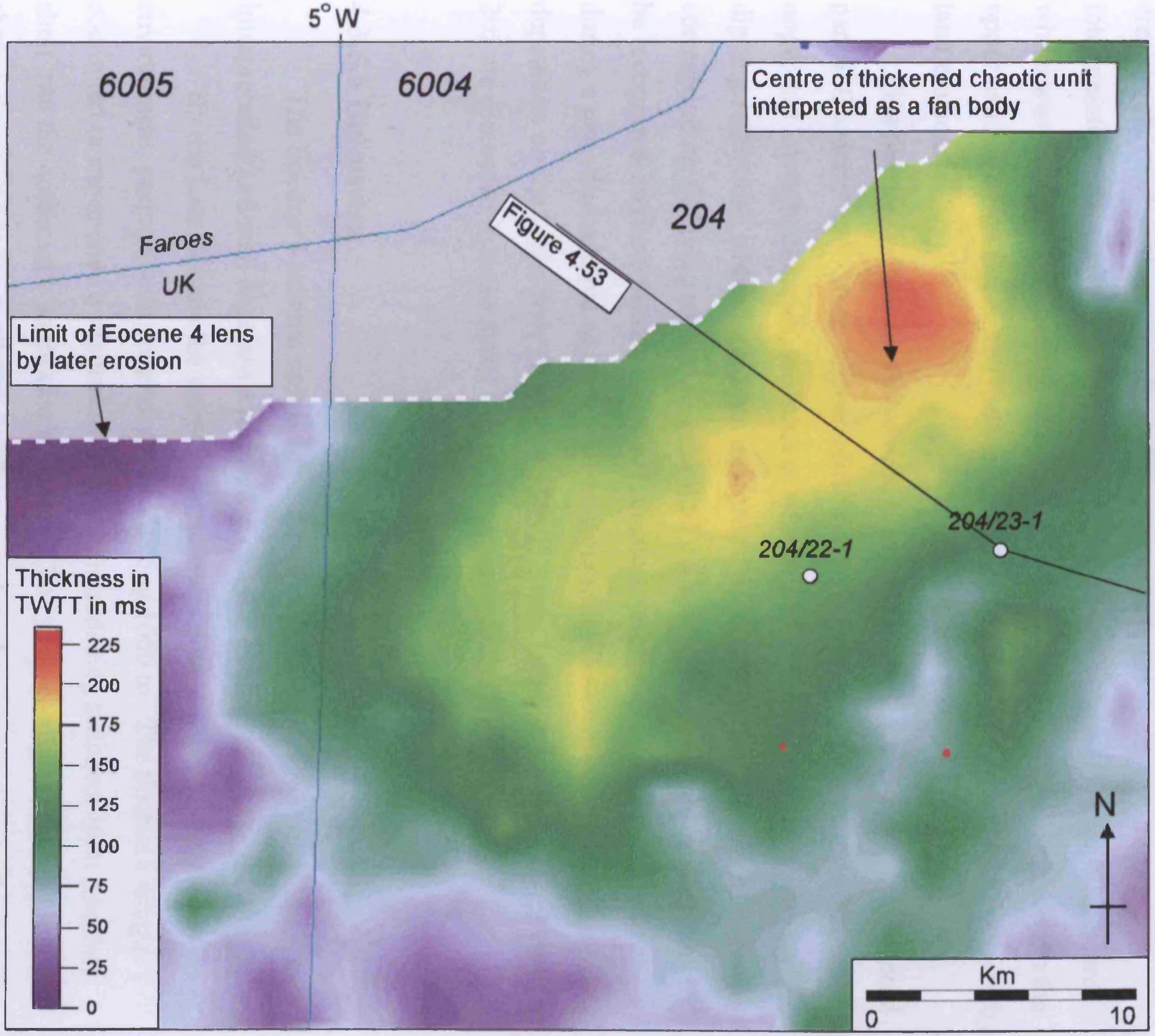


Figure 4.54. Isochron map (thickness in TWTT) of the thickened area of the Eocene 4 unit located in the south of Quadrant 204. The map highlights the overall geometry of the body which has a maximum thickness of nearly 250 ms. The broadly circular or lens shape body thins up dip to the southeast and also to the southwest and has a general oval shape that trends northeast. The northwestern edge of the lens has been eroded by a later unconformity that is close to the sea-bed (see Figure 4.53). Wells 204/22-1 and 204/23-1 are highlighted and the lithology of these wells show that the Eocene 4 unit here consists of a sandstone dominated succession. It is therefore interpreted to represent a base of slope fan body sourced up-dip to the southeast. Its narrow (elongate in a northeast - southwest) shape may suggest that the basin was relatively confined in the axes of the basin similar to what is seen at present day. The location of the 2-D seismic line in Figure 4.53 is also shown.

and micaceous and has occasional igneous rock fragments and rare shells. From the composite log, the sandstone has an interpreted age as Late Eocene to Early Oligocene (Priabonian – Rupelian). From these observations this local high amplitude package has been interpreted as a small base of slope fan system that was fed from the Shetland margin and ponded in a relatively narrow basin axes. This prominent northeast – southwest geometry of the fan may suggest that the basin axes or near base of slope zone was relatively narrow (similar to modern day) and the fan system may have just filled the available bathymetry. Additionally, it may suggest the fan system experienced significant along slope currents which affected the sediment dispersal creating the oval northeast trending sedimentary thick that is observed. An interpretation of a gravity driven deep water fan system is postulated for this feature which was fed from the Shetland margin and possibly shows signs of instability in the upper reaches of the slope. The distal edge of the fan system has since been removed by later erosion.

Further basin floor fans have been identified in the Eocene 4 unit in the central part of Quadrant 205. Here, high amplitude chaotic packages have strong continuous upper and lower boundaries, the bases of which can be traced landward up steeply dipping reflections which have been interpreted as foresets of clinoforms into the corresponding flat lying topsets (**Figure 4.55**). On single 2-D lines local incision can be recognised on these topsets and may indicate an erosional surface that occurred during a period of relative sea-level fall creating sedimentary bypass which allowed for deposition on the basin floor (**Figure 4.55**). The fans found in the north of Quadrant 205 are discussed in more detail in Flett Ridge case study (**Chapter 6**).

4.3.5.6 Discussion

The Eocene 4 seismic unit is summarised here, incorporating both the seismic interpretation and well data into a regional geological evolution for the entire basin.

By the Late Eocene the entire basin axis was now in a deep marine environment, perhaps with water depths exceeding 1000 m. The Shetland margin continued to experience a shallow shelf setting introducing sediment from a narrow shelf into the northeast – southwest basin axis. The basin configuration continued to show a deepening to the northeast and there may have been connection to the more northeasterly Møre Basin into Quadrants 218 and 219. The Erlend complex now

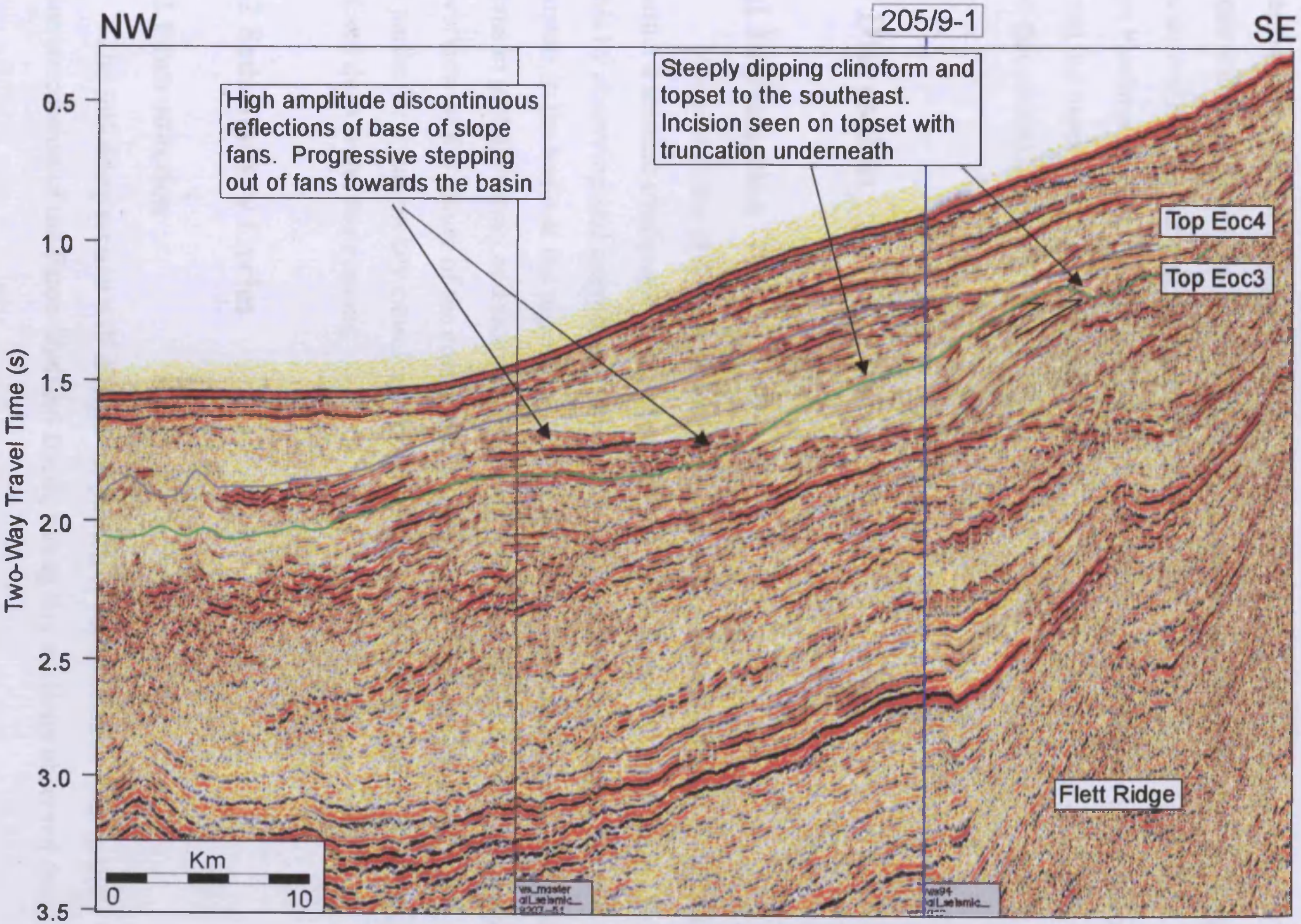


Figure 4.55. Southeast - northwest trending 2-D seismic line showing high amplitude discontinuous reflections found at the base of slope of the Shetland margin (in Quadrant 205). Steeply dipping foresets of clinoforms can be traced up dip into high amplitude tilted topsets where local incision can be seen. These base of floor fans occur in the Eocene 4 unit and are seen to progressively step out towards the basin axis. The incision seen in the up dip topset area represents a candidate sequence boundary (of Vail *et al.* 1977) which developed due to a fall in relative sea-level and possibly created a period of bypass to allow clastic deposition to the basin floor setting. At least two possibly three base of floor fans can be seen in Quadrant 205 and they step and young to the northwest. The location of the seismic line is shown in Figure 4.50.

provided little or no structural relief. This narrow elongate geometry to the depocentre is interpreted to be a result of continued subsidence in the central part of the basin axis (**Figure 4.56**). Little clastic sedimentation is evident at the time of deposition of the Eocene 4 unit compared to the more sand-rich Eocene 3 unit, and only local deep marine fan systems are seen in the south of the basin (south of Quadrant 204). The basin narrowed and became shallower to the southwest where a small circular fan system is found in the basin axes (**Figure 4.56**). This fan differs dramatically from the elongate fan systems of the older Eocene 3 unit which are significantly greater in dimensions and have an elongate, narrow geometry. This Late Eocene aged fan (of the Eocene 4 unit) is more lens-shaped and may suggest that the bathymetry of the basin floor controlled the shape of the fan in this restricted narrow part of the basin. The Faroe Platform remained a stable outer shelf area or basinal area though little sediment entered the basin from the northwest (Andersen *et al.* 2002 - see **Section 4.4.3.2** for more discussion).

4.4 Discussion

4.4.1 Introduction

This part of the chapter will firstly bring together all of **Sections 4.2** and **4.3** and construct a seismic stratigraphic framework for the entire Faroe-Shetland Basin. It will do this by observing and interpreting key stratigraphic patterns seen within the sediments in the basin at the regional scale and relate these observations to changing patterns in global eustasy and tectonics. Secondly, this part of the chapter will provide an overview of the nature of the controls on deposition in the Faroe-Shetland Basin with particular focus on key causal mechanisms that affect both gross depositional style and local deep water fan systems.

4.4.2 Sedimentary Cycles

4.4.2.1 Introduction

The following section will discuss the pattern of sedimentation seen in the Eocene succession of the Faroe-Shetland Basin, noting any cyclicity observed from the

Late Eocene Palaeogeography during deposition of the Eocene 4 unit.

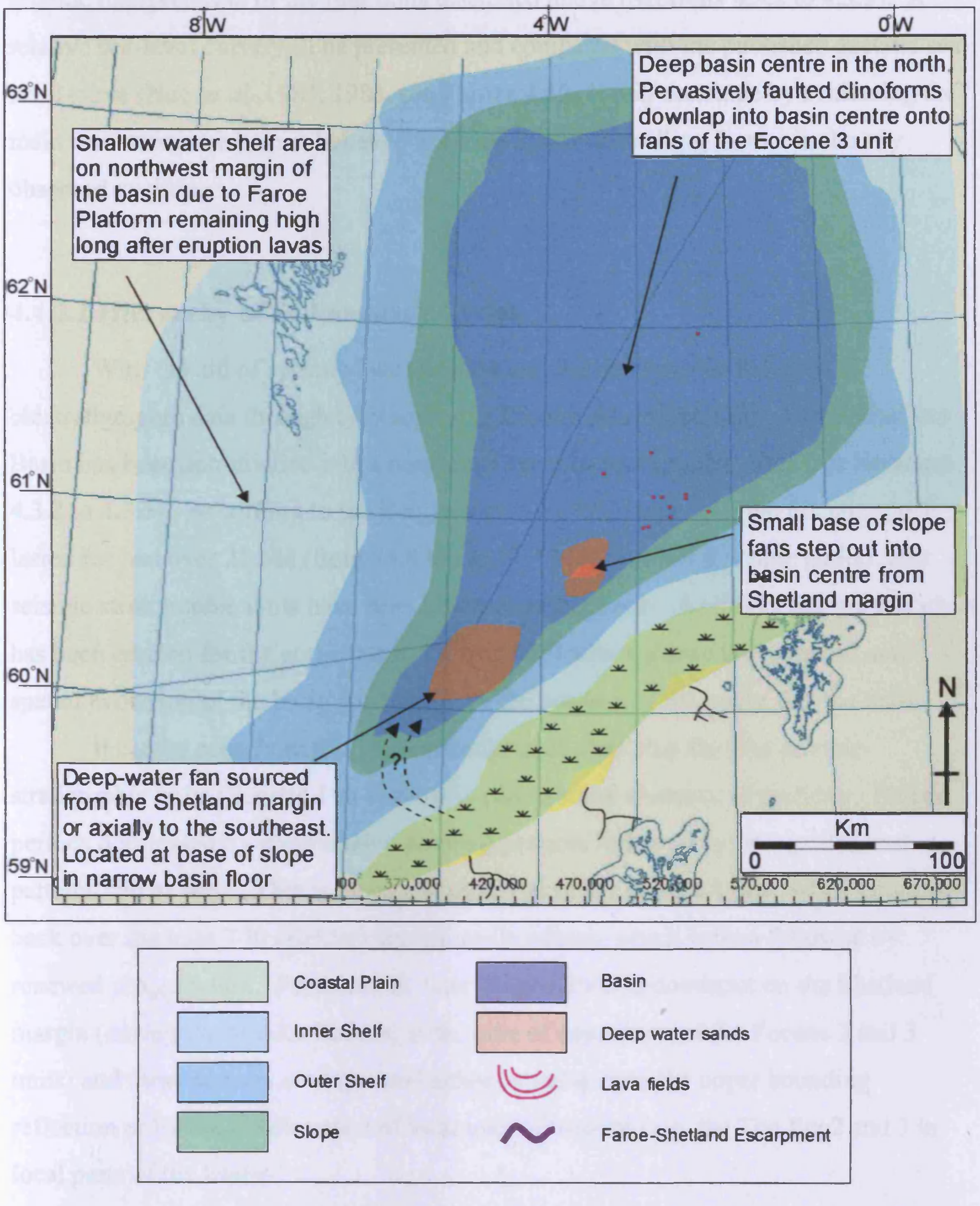


Figure 4.56. Schematic palaeogeographic map showing the generalised basinal setting during the deposition of the Eocene 4 unit. This unit broadly represents deposition during the Late Eocene to Early Oligocene. The basin was experiencing very little clastic deposition at this time and the majority of the basin (especially in the north) was dominated by hemi-pelagic fallout of clays and muds which was the background sedimentation, though clinoforms can be interpreted to downlap onto the Middle Eocene fans in the centre of the basin and indicate sediment entry from the southeast. The major fan systems had ceased by this time in the northern and central parts of the basin. In the south of the basin, small isolated fan bodies were deposited and sourced from the southern part of the Shetland margin. These fans differ from the elongate fans of the Eocene 3 unit and may reflect different basinal conditions. These fans are circular and lobate and have a northeast - southwest long axes orientation. This may reflect the narrow basin floor towards the south or the possibility of axial currents modifying the deposits. These fans are seen to step out and young into the basin centre (as seen in the central part of Quadrant 205). Here the bases of the fans can be traced up-dip into clinoform systems and localised eroded topsets. For legend of the colours used for the depositional environments see Table 4.1 or enclosure M.

seismic interpretation of the four units discussed above (**Sections 4.3.2** to **4.3.5**). A relative sea-level curve will be presented and compared with the published eustatic sea-level curve (Haq *et al.* 1987; 1988, see **Figure 4.57**) It will conclude by evaluating the main mechanisms what are believed to be the major controlling factors in the any observed cyclicity.

4.4.2.2 Hierarchy of Sedimentary Cycles

With the aid of seismic interpretation and the ability to tie the strata to biostratigraphic data through type wells, the Eocene succession of the Faroe-Shetland Basin has been sub-divided into a number of seismic-stratigraphic units (see **Sections 4.3.2** to **4.3.5**). According to the Berggren *et al.* (1995) timescale, the Eocene epoch lasted for just over 21 Ma (from 54.8 Ma to 33.7 Ma). Within this time period, four seismic stratigraphic units have been observed in the basin. A chronostratigraphic chart has been erected for the entire basin (**Figure 4.58**) which shows the temporal and spatial evolution of the basin and highlights the position of the major Eocene units.

It can be seen from this chronostratigraphic chart that the four seismic-stratigraphic units (Eocene 1 to 4) show in places some evidence of cyclicity. Indeed periods dominated by progradation stacking patterns followed by retrogradational patterns can be seen. This is especially observed in the Eocene 1 unit, when flooding back over the base T50 (Balder) unconformity occurs, which is then followed by renewed progradation. Furthermore, later progradation is dominant on the Shetland margin (during the Middle Eocene, at the time of deposition of the Eocene 2 and 3 units) and these periods are separated either by onlap onto the upper bounding reflection or by the development of local unconformities (e.g. the Top Eoc2 and 3 in local parts of the basin).

As has been mentioned throughout this study, the biostratigraphic data used herein does not allow for an accurate age date for the units to be ascertained. However, a crude age can be put on the units and this shows that they have durations of between 7.5 and 2.3 Ma and therefore share a close resemblance to the duration of 2nd or 3rd order cycles. It must be stated that the error bars on these age data are vast and it is by no means suggested here that these four units are indeed 2nd or 3^{rd-}order cycles.

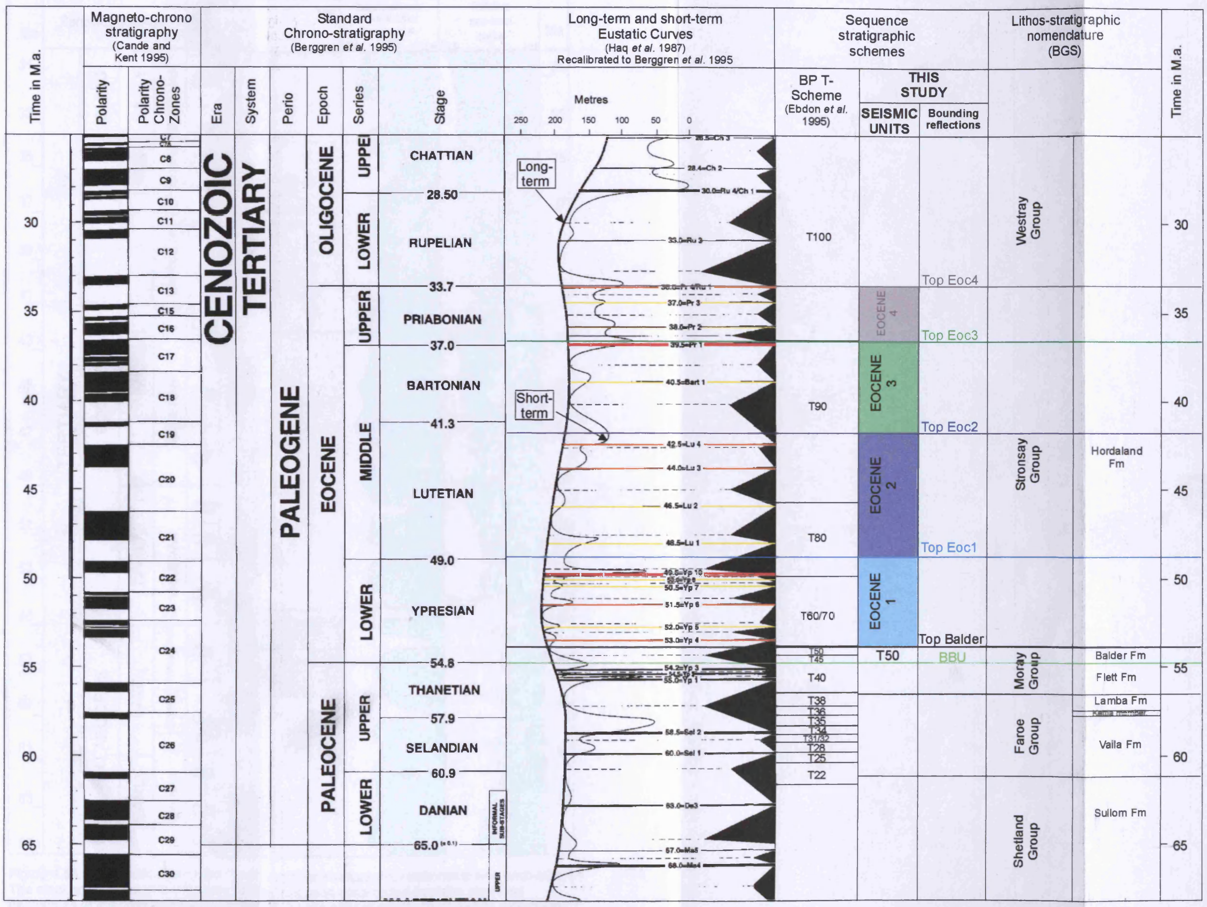


Figure 4.57. Stratigraphic chart summarising the chrono, magneto, litho and sequence stratigraphy of the Faroe-Shetland Basin. The stratigraphic position of the four Eocene seismic units defined in this study and there relationship to the eustatic sea-level curve of Haq *et al.* (1988) is shown. However, the biostratigraphic control on the four seismic units is poor and as such has large error bars. The four bounding surface s of the Eocene units are colour coded with respect to the seismic reflections they represent. The eustatic sea-level curve of Haq *et al.* (1988) recognises fifteen cycles of sea-level in the Eocene epoch and these have been interpreted to represent 3rd order cycles of global sea-level change. These cycles are bounded by sequence boundaries which have been given major (red), medium (orange) and minor (yellow) significance. Two major sequence boundaries are noted in the Eocene, one at 49.5 Ma and one at 39.5 Ma. The top of the Eocene 3 unit seems to correlate with the later major sequence boundary of Haq *et al.* (1988), though it is not possible to date the Top Eoc3 reflection accurately and it is schematically shown at the top of the Bartonian. The base T50 (Balder) unconformity, which is possibly the most significant unconformity seen in the Palaeogene of the basin does not correlate with the major sequence boundaries, though may correlate with a sequence boundary of medium significance. The non-correlation of the Eocene units defined here with the eustatic curve is expected due to the lack of any biostratigraphic control on the surfaces.

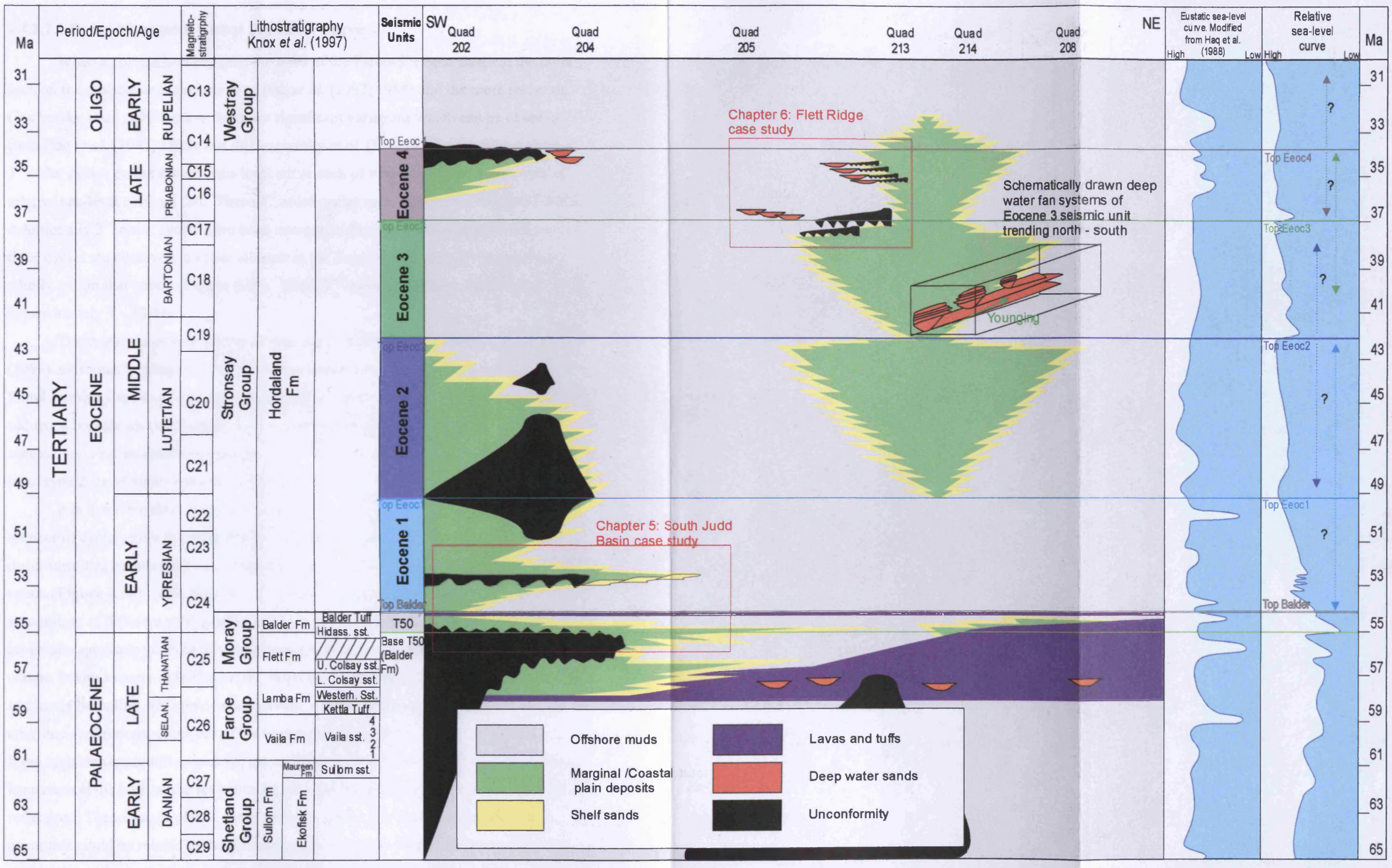


Figure 4.58. Schematic chronostratigraphic chart for the Eocene of the Faroe-Shetland Basin. This figure shows the temporal and spatial position of the four Eocene seismic units and their stratigraphic relationship with each other. The stacking patterns of the Eocene succession show periods of progradation and retrogradation of the facies belts during fluctuations in relative sea-level. A relative sea-level curve has been constructed from the observed geometries of the units and this is compared to the eustatic curve of Haq *et al.* (1988). The biostratigraphic resolution of the data used to date the four Eocene units is poor and thus the units are schematically shown in their crude position. The error bars on the absolute ages of the upper and lower bounding reflections are large. Nonetheless, the position of the deltaic and marginal facies and their relationship to the deep water fans of the Eocene 3 unit is shown. The unconformities can be seen to be of local extent (e.g. over the Judd Anticline) and this suggests a local tectonic control on the basin caused the observed fluctuations in the relative sea-level.

4.4.2.2.1 Comparison with Global Sea-Level Curve

When a comparison between the units of the Faroe-Shetland Basin to the units seen on the eustatic sea-level curve of Haq *et al.* (1987; 1988) and the more recent de Graciansky *et al.* (1999) curve there are significant variations which can be observed. Both Haq *et al.* (1987; 1988) and de Graciansky *et al.* (1998) recognise fifteen similar 3rd order cycles on the eustatic sea-level curve each of which represent a full cycle of relative sea-level rise and fall. These 3rd order cycles vary between 0.5 Ma and 2.5 Ma. Additionally 2nd order cycles have been interpreted from the Palaeogene period and three cycles are observed to occur at times in the Eocene, with one (TA3) occurring wholly within that epoch **(Figure 4.57)**. These 2nd order cycles have durations of approximately 9 – 11 Ma.

The eustatic sea-level curves of Haq *et al.* (1987; 1988) and de Graciansky *et al.* (1998) additionally recognises that sequence boundaries of both type 1 and type 2 are found forming the bases to transgressive parts of many of the cycles. Two major type 1 sequence boundaries (see **Section 3.2**) are seen on this global curve in the Eocene succession. Five sequence boundaries are ranked as medium and eight are believed to be of type 2, or of minor relevance **(Figure 4.57)**.

It is therefore clear to say that there is a significant discrepancy between the number of cycles, their duration and the number of major and minor type 1 or type 2 unconformities between this study and the expected global signature seen in the Eocene epoch **(Figure 4.58)**. This discrepancy could be a manifestation of the individual recognition of different stratigraphic units. It is clear from the seismic mapping and the pragmatic approach taken in this study that it was not possible to map and correlate the classic Exxon sequences in this basin. What has been mapped and defined as the upper and lower bounding surfaces of each seismic – stratigraphic unit are continuous high amplitude reflections which are of the best regional extent. In places within the basin these high amplitude reflections appear as either sequence boundaries showing erosion truncation at the base of the reflection or of major onlap or downlap surfaces onto the reflection. Therefore, in places, these surfaces represent a significant stratigraphic event that could be related to the necessary interaction of eustasy, tectonics and sediment supply required to produce such a depositional architecture.

However, this does not alter the view of many authors who are proponents of the eustatic sea-level curve who believe that global sea-level variations are manifested in basins throughout the world as sedimentary cycles (e.g. Vail *et al.* 1977a, Mitchum and Vail 1977, Mitchum *et al.* 1977a, - see **Section 3.2**). The fifteen 3rd order cycles recognised during the Eocene epoch from the eustatic sea-level curve are not recognised in the Faroe-Shetland Basin and the question remains as to why. The lack of recognition of these 3rd order cycles in the basin may be explained by many lines of argument.

Firstly, the eustatic cycle may be in some way incorrect or inaccurate and may be explained by something other than global sea-level change. The origins of the global eustatic sea-level curve are in the recognition of the relative changes in the position of coastal onlap packages. Herein lies an assumption that the onlap interpreted from seismic data or from outcrops is entirely of coastal origin and represents the relative movement of the shoreline in response to fluctuations in relative sea-level. This may not always be the case and in areas of poor well and lithological control marine and coastal onlap may sometimes be misinterpreted.

Conversely, the data in the Faroe-Shetland Basin is of varying quality and there is only local evidence of major sequence boundary development in the basin. The base T50 (Balder) unconformity (of Smallwood and Gill 2002) is one such example of a type 1 sequence boundary that can be extensively mapped and correlated around the southern part of the basin and is known to be of sub-aerial origin. However, this unconformity is believed to be developed due to local tectonic factors and not as a result of global sea-level fall. Indeed the area of uplift is believed to be a minimum of 3000 km^2 (Smallwood and Gill 2002) covering an area restricted to the south of the basin in the south of Quadrants 6004, 204 and 205 and the north of Quadrant 202 (see **Section 4.2**). The favoured model put forward for this localised unconformity is the transient impingement of hot upwelling asthenosphere in the form of the proto-Iceland mantle plume (see **Section 4.4.3.1.3**).

Other localised unconformities have been identified in the Faroe-Shetland Basin throughout this study. One such example of these are the canyons (e.g. the Strachan Canyon) that feed the elongate deep water fans of the Eocene 3 seismic unit (**Section 4.3.4.4**). However, these canyons are of very limited areal extent (less than a few km) and may form by a variety of processes. One of these processes is a fall in global sea-level to expose the shelf and hence create the classic lowstand fan model (Posamentier *et al.* 1988). However, additionally canyons can be initiated by oversteepening of the slope (Ross *et al.* 1994) causing rotational slides and slumps. Furthermore, a purely tectonic control on canyon formation can exist with local faulting creating the focus to sediment bypass (see **Section 4.4.3.2.2** for further discussion).

The global sea-level curve identifies many sequence boundaries within the Eocene, of which, two are interpreted by the authors to be major or type 1 sequence boundaries (Haq *et al.* 1987; 1988). In the Faroe-Shetland Basin, the base T50 (Balder) unconformity is a good candidate to be classified as a type 1 sequence boundary though as discussed earlier is not believed to be caused as a result of global sea-level variations. Therefore, if the global sea-level curve is to be believed, additional type 1 sequence boundaries should be recognisable throughout the Eocene succession in the Faroe-Shetland Basin. These are not apparent from the 2-D seismic data used in this regional study, especially from the basin margins where they would be expected to be developed. Candidate sequence boundaries have nevertheless been recognised from the southern part of the basin and on the Shetland margin purely from reflection termination mapping (e.g. the Top Eoc2 reflection, cutting out Top Eoc1 reflection in the south and the Top Eoc4 reflection becoming coincident with Top Eoc3 reflection in the northeast part of the Shetland margin). Unfortunately, these candidate sequence boundaries cannot be calibrated with accurate age and lithological data to infer their stratigraphic position, thus their chronostratigraphic and lithostratigraphic significance is therefore unknown.

On the Shetland margin, a candidate sequence boundary occurs at the top of a major progradational package of the Eocene 3 unit which is seen to source base of slope fans. This is located in the central part of Quadrant 205, where the base of the high amplitude fan packages can be traced up-dip and up the dipping clinoform onto a flat topset. Incision is seen on these topsets (into the Top Eoc3 reflection) but this is only seen locally and a broad unconformity over much of the shelf is not seen. However this local incision may represent a fall in relative sea-level and cause sediment bypass to the basin floor (see **Section 4.3.5.5**). This suggests that there are local sequence boundaries that developed on the margins of the basin, but are not always recognisable in basin axes highlighting the limited lateral extent of these surfaces (Cartwright *et al.* 1993). This overall lack of recognition may be due to the data quality or simply to the location of the 2-D lines. High quality 3-D seismic surveys positioned on basin margins may provide the key to the identification and recognition of local sequence boundaries in the Faroe-Shetland Basin. Furthermore, 3-D surveys were available for this study and **Chapters 5** and **6** will look at two local case studies, one from the southern basin margin and one located in an intrabasinal setting. Unfortunately no 3-D surveys are located on the southeastern Shetland margin in a proximal setting which has been shown to be a major entry point into the basin throughout the Palaeogene.

Finally, there is a difficulty in dating the sequence boundaries found in the Faroe-Shetland Basin. This is because the biostratigraphic resolution of the Eocene succession in the basin is generally very poor. As previously highlighted in **Section 3.5**, very little accurate data is available for the Eocene succession due to drilling techniques and deeper target reservoirs. Hence, biostratigraphic information is sparse and when available, does not constrain the age with any real precision. This has prevented the accurate dating of the four seismic stratigraphic units detailed in this study and thus are better regarded in a relative age scheme than an absolute one (see **Section 3.5**). The thirteen wells which have the highest resolution of biostratigraphic data have allowed for a crude age range to be assigned to the units. However, the paucity of FDA's has meant that the general diverse micropalaeontological and microfaunal assemblages have been used to identify an Early (Ypresian), Middle (Lutetian – Bartonian) or Late (Priabonian) Eocene age. Occasionally, the age can be constrained to a greater degree in some wells, e.g. a Late Ypresian age though there is a lack of consistency between individual contractor schemes.

4.4.2.3 Palaeogeographic Evolution

Sections 4.3.2 to **4.3.5** have summarised in detail each individual seismic-stratigraphic unit of the Eocene succession. The gross pattern of deposition has been described on a basin-wide scale using the combined tools of seismic interpretation and well data analysis, with a view to producing a series of palaeogeographic maps representing time intervals throughout the Eocene. These maps show the change in depositional setting of the basin throughout the Eocene. The earliest Eocene T50 unit (Balder Formation equivalent) was dominated by the presence of a sub-aerial swampy area that covered much of the southern part of the basin. Coals and paralic, marginal facies were common and are interpreted to be sourced from the Orkney Platform to the south. In the north of the basin, marine conditions prevailed with the position of the shoreline being controlled by the remnant Faroe-Shetland Escarpment which was still structurally high at this time (**Figure 4.59**). Marine siltstones and shales dominated this

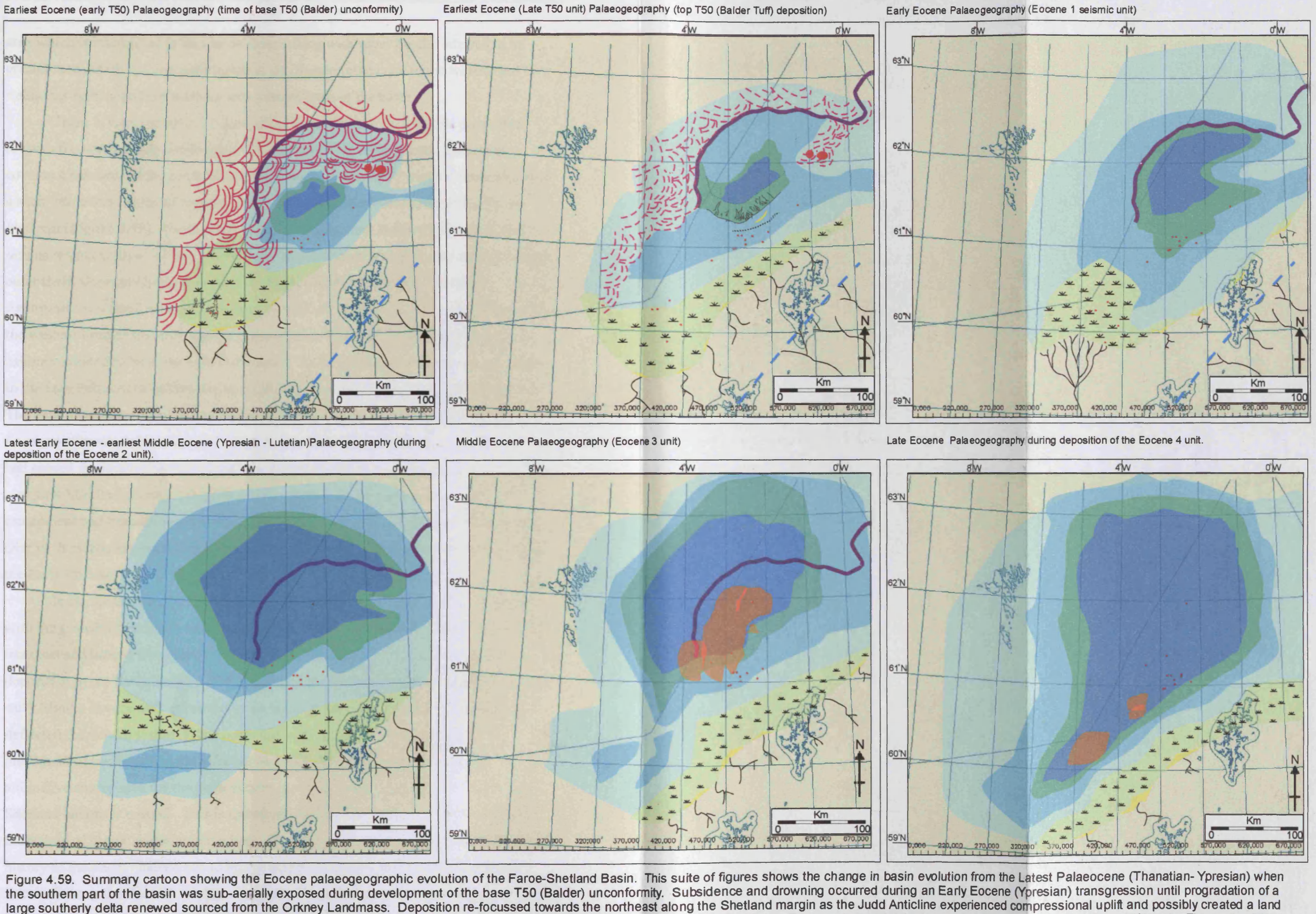


Figure 4.59. Summary cartoon showing the Eocene palaeogeographic evolution of the Faroe-Shetland Basin. This suite of figures shows the change in basin evolution from the Latest Palaeocene (Thanatian- Ypresian) when the southern part of the basin was sub-aerially exposed during development of the base T50 (Balder) unconformity. Subsidence and drowning occurred during an Early Eocene (Ypresian) transgression until progradation of a large southerly delta renewed sourced from the Orkney Landmass. Deposition re-focussed towards the northeast along the Shetland margin as the Judd Anticline experienced compressional uplift and possibly created a land bridge restricting the basin. This compression was localised and short lived as the basin returned to a marine setting and by the Middle Eocene, deepwater fans were fed through canyons at the shelf break. The north of the basin remained in a marine realm throughout the entirety of the Eocene. Smaller, possibly axially fed fans were deposited towards the south of the basin in the Late Eocene.

area which is interpreted to be a basin floor setting with water depths inferred to be greater than 500 m. A low sedimentation flux is interpreted due to the widespread lava fields that surrounded the northern and western parts of the basin.

There is clear evidence to show the presence of a large delta that prograded towards the north during the deposition of the Eocene 1 seismic unit. This delta covered a vast area in the south of the basin (up to 10,000 km^2) where it prograded onto a shelf into water depths of approximately 250 m and northwest towards the Faroe Platform (**Figure 4.59**). North of this deltaic environment, there remained a deeper marine realm (north of 61° N) where upper bathyal conditions prevailed in a mud prone outer shelf, slope and basin floor setting (**Figure 4.59**). The Faroe-Shetland Escarpment remained an important feature and continued to control sedimentation and the western limit of the deep marine environment. The escarpment can be used to calibrate water depths in the Late Palaeocene. Indeed, the escarpment was a shoreline in the Late Palaeocene as lava deltas entered the sea and cooled (e.g. Kiørboe 1999). The escarpment continued to provide a westerly control on the accumulation of Lower Eocene sediments (of the Eocene 1 unit) in the north of the basin, though subsidence had caused the initial flooding of the escarpment by this time. By the late Early Eocene – earliest Middle Eocene, during the onset of deposition of the Eocene 2 unit, the escarpment had been drowned and no thinning of the Eocene succession is observed over it. It is thus interpreted to no longer represent a bathymetric feature by the end of the Early Eocene.

In the south of the basin, considerable basin reorganisation was taking place, with the growth of the east - west trending Judd Anticline forming a baffle to sediment transport and having the effect of switching off the delta system that had dominated deposition in the Early Eocene. This continued growth of the Judd Anticline during the early Middle Eocene influenced the gross sedimentary architecture of the basin and deflected the deltaic system to the north and east, up the Shetland margin. This movement of the delta had the effect of isolating the Orkney hinterland and an alternative source area for the delta system (of the Eocene 2 seismic unit) from the Shetland landmass ensued. This is therefore a classic example of tectonics affecting the supply of clastic material into a basin, thus exerting a dominant control on the infill history of the basin. Indeed a land bridge may have formed at the end of the Early Eocene linking the Scottish landmass with the Faroe Islands to the northwest. Furthermore, in the south of Quadrant 204 and the north of Quadrant 202, a small isolated basin, with water depths of less than 500 m developed south of the growing Judd Anticline. There was continued deepening of the basin in the north which was subsiding and little or no sediment was being deposited from the starved and lava rich Faroe margin **(Figure 4.59)**.

The Middle Eocene (Eocene 3 unit) is dominated by the deposition of deep water sandstones which were fed through canyons on the shelf-slope break. These sandstone deposits were formed by gravity-driven processes and developed as elongated deep water fan systems that trend approximately north - south. Conglomerates and interbedded sandstones are found in well 214/26-1 which is located in the canyon that feeds the westernmost Strachan Fan. These conglomeratic beds suggest that this well is located in the very proximal part of the fan and that towards the basin centre, the conglomerates become progressively interbedded with sandstones. The controls on the development of these deep water fan bodies will be discussed in more detail in **Section 4.4.3.2**.

These deep water fans were sourced from a shelf area on the Shetland margin, where progradational clinoform systems are seen up-dip from the canyons. Water depths greater than 250 m are interpreted for the shelf area and thus the fans are interpreted to be deposited in significantly deeper waters in a base-of-slope to basin setting. During the Middle Eocene, a sand-rich inner shelf may have fed sediment into the outer shelf and ultimately down into canyons (see **Section 4.4.3.2.2**). In the south of the basin, the Eocene 3 unit remained relatively thin and deltaic conditions gave way to an outer shelf environment with little clastic sediment. Subsidence in the central part of the Faroe-Shetland Basin continued, and the Faroe Platform was relatively uplifted and continued to experience outer shelf conditions. This subsidence had the effect of creating the dominant northeast – southwest trend that the bathymetry displays today, and thus an open synclinal form to the basin is believed to have developed as early as the Eocene. The Judd Anticline remained bathymetrically high and two localised depocentres trending northeast - southwest are present in the south of Quadrant 204.

Finally, in the latest Middle Eocene to earliest Late Eocene (at the time of deposition of the Eocene 4 unit) the northeast – southwest basin depocentre was experiencing deep marine conditions, with accommodation space and water depths believed to be a at a maximum during the Eocene (possibly greater than 1000 m in places). However, the basin narrowed significantly towards the southwest where small circular, lens shaped fans occurred which can be traced up-dip into their shelfal equivalents. It is suggested that this may possibly be the result of tectonic focussing of the slope into a narrow axis.

The north of the basin was prone to the deposition of mudstone at this time and the only clastic systems were to be found in Quadrants 205 and 204. The palaeogeographic evolution throughout the Eocene is therefore regarded here as one of deepening of the basin by subsidence. The northern part of the basin remained in marine setting throughout the entire Eocene, though water depths are interpreted to have increased from 500 m to greater than 1000 m during this time, whilst the southern part of the basin was affected by episodic and localised compressive tectonic events which created local unconformities. Subsidence is interpreted in this thesis to have occurred at some point in the Early Eocene prior to the development of the Eocene 3 fans that dominate the northern part of the basin in the Middle Eocene. It is apparent from the mapping that shifting sediment source areas on a semi-regional scale occurred from the Early to Middle Eocene and deltas that originated on the southern margin of the basin then migrated to the northeast allowing for major progradation and feeding of the deep water fans in a northeast - southwest trending depocentre.

Extensional tectonics is considered by recent studies to have ceased in the Palaeocene (Dean *et al.* 1999). Based on this, it is thought most likely that this Eocene deepening is due to the post-rift subsidence which manifests itself along basin margins after rifting had ended (see **Section 4.4.3.1.1** for further discussion). Alternatively, progressive loading of the passive margin by sedimentary wedges may cause significant flexure resulting in continued subsidence of the margin (e.g. Watts and Steckler 1979).

Additionally, the subsidence may also be viewed as a result of the collapse of the plume from under the lithosphere in which thermal contraction of the plume head would have been associated with significant subsidence. Furthermore, the transient movement of the plume head to the northwest (towards the present day site under Iceland) away from the Faroe-Shetland Basin may have provided relative uplift in the area of the Faroe Islands and thus the neighbouring basin may have experienced relative subsidence.

To further complicate the pattern of semi-regional uplift and subsidence, this regional basin-wide subsidence was punctuated by local tectonic controls including compressive episodes on the Judd Anticline and changing sediment supply which affected the stratigraphic architecture of the basin. The individual controlling mechanisms of the regional basin fill including eustasy, subsidence, the Iceland mantle plume and compression will be discussed in greater detail in the following section (**Section 4.4.3.1**). Local controls on the deposition of the Middle Eocene deep water fans will be discussed in **Section 4.4.3.2**.

4.4.2.4 Summary of the Relative sea-level history in the Faroe-Shetland Basin

A brief explanation of the relative sea-level as shown from **Figures 4.58** and **4.59** will now finalise this section. An interpreted relative sea-level curve for the basin has been constructed and compared to the eustatic curve (e.g. Haq et al. 1987; 1988) and is shown in **Figure 4.48**. The general trend of the curve shows a gradual increase in water depths from a period of low relative sea-level during deposition of the T50 unit to a relative high relative sea-level by the end of the Eocene and start of the Oligocene. This general relative sea-level rise is punctuated by small falls in relative sea-level throughout the Eocene, however, as has been discussed in **Section 4.4.2.2.1** the absolute ages of these falls cannot be constrained well by the available biostratigraphic data. The trend of the relative sea-level curve (**Figure 4.58**) can be explained by a progressive deepening of the basin which is highlighted by the six palaeogeographic reconstructions (**Figure 4.60**). This increase in the relative sea-level caused a flooded the basin margin initially to the south, and then progressively on to the Shetland and Faroe margins. Water depths in the northern end of the basin axes remained relatively deep throughout the Eocene (between 250 – 100m), and was probably continually in an upper bathyal setting. Conversely, the basin margins experienced significant falls in relative sea-level and thus changes in the basin setting were common; indeed localised unconformities were created by these relative sea-level falls. The possible individual controls on the relative sea-level of the Faroe-Shetland Basin are discussed independently in the flowing section (**Section 4.4.3**).

4.4.3 Driving Mechanisms

This section will discuss possible key causal mechanisms of the processes that controlled the development of the basin and affected the sediment dispersal patterns both on a regional and local scale. Each mechanism will be discussed individually focussing on the evidence seen from the seismic and well data. The merits of each mechanism will then be assessed.

4.4.3.1 Controls on Regional Basin-wide stratigraphic Architecture

Large scale (10 - 100's km) controls on Eocene deposition will be discussed here with particular emphasis on regional tectonic events which had the effect of configuring the entire basin architecture into its final form.

4.4.3.1.1 Post-rift Subsidence History

After the cessation of an extensional tectonic episode, passive margins undergo a period of thermal subsidence which is documented by a negative exponentially decreasing curve (McKenzie 1978). A brief review and discussion of the subsidence observed in the Faroe-Shetland Basin and adjacent basins will now follow, highlighting published work and integrating the stratigraphic observations of this study.

The post-rift sedimentary fill of the North Sea Basin was the first to be extensively studied and patterns of subsidence were derived from both well and seismic data (e.g. Bertram and Milton 1988, Joy 1992, White and Latin 1993). From these studies it was noted that there is a discrete period of accelerated subsidence seen in the Palaeogene.

Separate studies focussing on the Faroe-Shetland Basin were conducted which showed the same pattern of accelerated Palaeocene subsidence from the basin centre of approximately 100 – 150 m/Ma (Turner and Scrutton 1993). However, coeval with the basin subsidence in the centre of the Faroe-Shetland Basin, there was widespread flank uplift on the basin margins. This uplift allowed for the development of a Mid-Palaeocene unconformity on the basin flanks, which was noted in many other North Atlantic basins implying a regional control on this uplift. This anomalous pattern of subsidence was initially attributed to igneous underplating which was synchronous with re-extension in the basins (Hall and White 1994). Furthermore, Brodie and White (1994; 1995) linked the permanent uplift of the UKCS and the British Isles to the lower crustal intrusion of gabbroic material which was responsible for the regional uplift and the missing post-rift subsidence in the adjacent Mesozoic basins.

Therefore in the Faroe-Shetland Basin, there is coeval flank uplift and basin subsidence and the origin of these conflicting vertical tectonic movements needs an explanation. There is a suggestion of Palaeocene extension in the Faroe-Shetland Basins that may account for the observed increase in tectonic subsidence rates (Hitchen and Ritchie 1987, Mudge and Rashid 1987). More recently Palaeocene extension has been documented on the most northwestern ridge complex in the Faroe-Shetland Basin; the Corona Ridge (Dean *et al.* 1999). Similarly in the North Sea and Porcupine basins, extension in the Palaeocene is documented (Ziegler 1990) though the syn-sedimentary faults observed here may be caused by differential compaction over buried pre-existing fault scarps. However, just a simple small pulse of Palaeocene extension which only accounts for a very small amount of the total extension cannot explain the observed simultaneous flank uplift and accelerated basin subsidence (e.g. Joy 1992).

Alternative hypotheses were examined to account for the anomalous subsidence patterns seen in the basins of the North Atlantic. In the Faroe-Shetland Basin a simple eustatic or sedimentary explanation can be dismissed on the grounds that there appears to be a mismatch between eustatic sea-level variations and the observed changes in water depths and tectonic subsidence rates (Turner and Scrutton 1993). This mismatch occurs by examining known water depths from wells in the basin (e.g. by the recognition of coal beds or topsets) and correlating periods of shallowing (decreasing water depths) to periods when rates of tectonic subsidence remained high.

Intra-plate stresses have been suggested as a possible cause of rapid pulses of tectonic subsidence in the Quaternary of the North Sea (Cloetingh *et al.* 1990). This model suggests bending of the lithosphere due to compressive intra-plate stresses from the Alpine deformation to account for the flank uplift and accelerated basin subsidence. Models predicted that 10 - 100 m of uplift and greater than 150 m of basin centre subsidence is expected when feasible values of intra-plate stresses of 100 - 500 MPa are used (Karner 1986, Cloetingh *et al.* 1990). These estimates of vertical movements are approximately similar to the observations seen in the Faroe-Shetland Basin and may provide an alternative model for the origin of the accelerated subsidence and uplift seen.

Another key observation in this complex pattern of Palaeocene tectonics is the initiation of magmatism (related to the Iceland plume) at the same time as the inception of rapid basinal subsidence. This coincidence of magmatism and subsidence may imply a causal link between igneous activity and subsidence (Joy 1992). The period of accelerated subsidence is approximately 4 Ma and it is believed that longer timescales are required to generate lithospheric heating effects. Therefore a mechanical origin is envisaged to be more likely rather than a thermal one for the response seen in the subsidence rates (Turner and Scrutton 1993). A more detailed discussion of the effects of the Iceland plume and lithospheric heat flow will follow in **Section 4.4.3.1.3**.

Rates of subsidence are seen to return to normal (post-rift thermal) levels in earliest Eocene times, when magmatism had ceased. At this time, to the northwest of the Faroe-Shetland Basin, the inception of sea-floor spreading and continental break-up occurred. Any effect the Iceland plume had on both uplift and subsidence was short-lived as the transient plume head migrated to the northwest and into its present position beneath Iceland. Hence it is probable that the support under the lithosphere had moved away and allowed the entire Faroe-Shetland Basin to subside and sag.

A gross isochron (time thickness) map shows the preserved thickness of the entire Eocene succession **(Figure 4.60)** and highlights a central depocentre trending northeast - southwest, deepening to the northeast and showing semi-regional areas of uplift and erosion which was caused by intra and post Eocene compressive episodes (see **Section 4.4.3.1.2**). There is an asymmetrical thinning of the Eocene succession to the Shetland and Faroe margins, and from the northwest – southeast **regional correlations A – E** it can be seen that the classic "Steers Head" geometry does not exist. On the Shetland margin, there is good evidence of landward convergence of reflections, whilst on the Faroe margin, parallel reflection configurations are seen to be eroded by a much later (post-Eocene) unconformity. This suggests that the Faroe margin may have experienced uplift whilst the Shetland margin was undergoing subsidence during and after Eocene deposition. An important factor to take into consideration when analysing this subsidence pattern is the effect of loading the crust with large sedimentary wedges. By creating a rift basin and then infilling it with sediment, the extra loading caused by the syn and post-rift succession causes flexure and bending of the flanks to produce the "Steers Head" geometry (Watts and Steckler 1979, Watts and Thorne 1984). Couple this with periods of compression and this may account for a complex basin fill as seen in the gross Eocene succession.

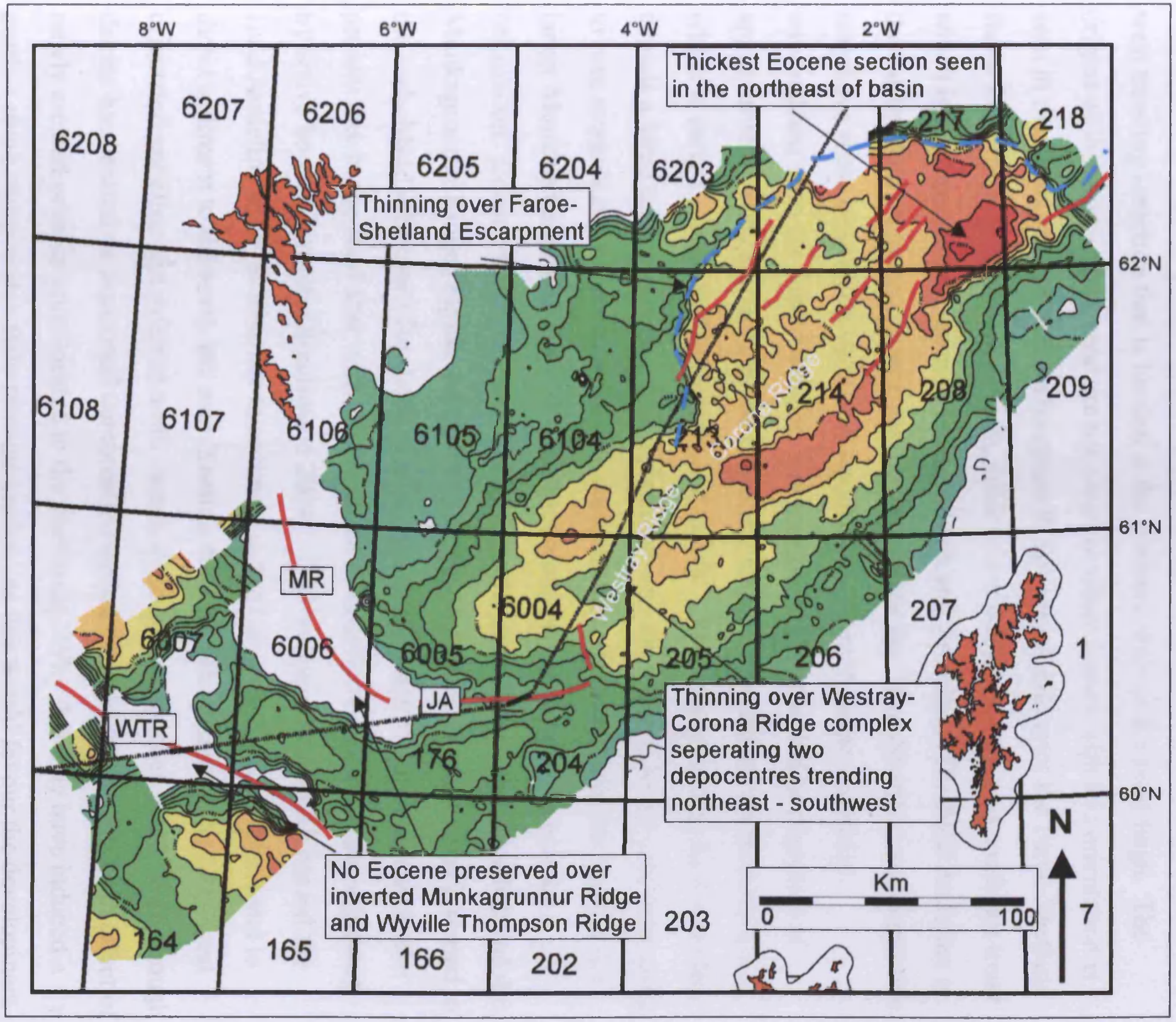


Figure 4.60. Isochron map (thickness in TWTT) showing the gross Eocene succession (Top T50 (Balder) - Top Eoc4). A northeast - southwest preserved thickness is dominant with the thickest succession seen in the northeast corner of the basin. There are specific controls on the gross basin architecture which are apparent. Firstly, the Faroe-Shetland Escarpment (blue dashed line) in the west of Quadrant 213 shows a dramatic change in preserved thickness of the Eocene strata over the escarpment. This feature remained topographically high and affected the controlled the western limit of the three Middle Eocene deep-water fans. The north - south to northeast - southwest trending Westray - Corona Ridge complex also shows thinning over it and has the effect of separating the basin into two distinct depocentres. In the south, intra and post Eocene compression caused inversion anti-clines (red lines) which and erosion of the Eocene succession in the south of the basin on the Munkagrunnur Ridge, Wyville-Thompson Ridge and Judd Anticline, (MR, WTR and JA).

4.4.3.1.2 Compressional Episodes

The effects of Palaeogene and Neogene compression have long been documented in the Faroe-Shetland Basin (Boldreel and Andersen 1993; 1998, Andersen *et al.* 2000, Davies *et al.* 2004). Major northeast - southwest trending inversion anticlines have been identified and have been dated as predominately post-Eocene in age (commonly Oligo-Miocene and older - Boldreel and Andersen 1993, Davies *et al.* 2004). However, in this study, intra-Eocene compression has been identified from the south of the basin in the Judd area (south of Quadrant 204).

This early compression is unique in the sense that the Judd Anticline is an east – west trending structure that is located at the northern limit of the Judd High. The origins of this east – west trend are not clear: no other feature with this orientation is seen in the basin. The Wyville-Thompson Ridge, which separates the Faroe-Shetland Basin from the North Rockall Trough, exhibits a west northwest – east southeast trend which is the closest orientation to the Judd High and Anticline. The Judd Anticline is the earliest known inversion structure present within the Faroe-Shetland Basin and this may have some bearing on its unique orientation. From thinning and onlap terminations onto the Judd Anticline, it is possible to date the timing of growth at approximately early Middle Eocene (Eocene 2 seismic unit) times (see **Section 4.3.3**), which is early Lutetian in age (Smallwood 2004). The geometry of the Judd Anticline depicts a large east – west spur which protrudes into the basin, with the fold decreasing to zero towards its eastern edge. At the western limit, the Judd Anticline links to the larger Munkagrunnur Ridge which has a predominantly northwest – southeast orientation. Eocene sediments are missing from the crest of the Judd Anticline and the Munkagrunnur Ridge (**Figure 4.60**) due to localised uplift and erosion which started in the early Middle Eocene (Top Eoc2 erosion, see **Section 4.3.3**) and continued to the present day in the Judd area with the formation of deepwater erosional scarps created by active bottom currents (Smallwood 2004). This compressional uplift caused the Judd Anticline to act as a barrier to sediment in the Lower to Middle Eocene and to direct sediment to the north and east (**Section 4.4.2.3**). The origin of the east - west orientated anticline and inferred north - south directed compression is unknown, though debate has centred on horizontal compression stresses related to the ridge push effect of newly created oceanic crust located to the northwest. Why this may have induced a north – south compressive state remains unclear, as this would favour the development of the more commonly observed northeast – southwest post-Eocene compressional anticlines which are common in the basin (Davies *et al.* 2004). However, Boldreel and Andersen (1993) describe a Late Palaeocene north – south compressive episode to form the Wyville-Thompson Ridge and Ymir Ridge and propose this early compression to be related to the oblique nature of the Aegir Ridge spreading centre to the northwest (Boldreel and Andersen 1993). This does not, however, explain the east – west Judd Anticline which formed some 5 -10 Ma later. In summary, whilst the evidence for significant compressional tectonics in the Early Eocene and later is clear, the origin of this stress configuration is still unresolved.

4.4.3.1.3 Mantle Plumes

The notion of the presence of a mantle plume beneath the Faroe-Shetland Basin has been debated for decades (e.g. White 1988; 1989, Nadin *et al.* 1997). From observations in the Faroe-Shetland Basin, evidence supporting a mantle plume underneath the lithosphere is primarily to be found in the associated magmatic activity and the short-lived uplift event that was evident in the earliest Eocene (**Sections 4.2** and **4.3.2**). The development of the base T50 (Balder) unconformity is seen to be confined to the southern part of the basin and is thus a localised feature. Mantle plume activity has been cited as the causal mechanism for tectonic uplift and generation of the base T50 (Balder) unconformity (Smallwood and Gill 2002). Tectonic uplift greater than 200 m is advocated from evidence of the amount of incision within incised valley lows (Smallwood and Gill 2002). This uplift event appears to be short-lived and may have only lasted for a period of approximately 0.5 Ma (Smallwood and Gill 2002, Jolley *et al.* 2002). This brevity and transient nature of this uplift period may favour a plume origin, in which the plume head was moving away and provided only a momentary pulse of uplift at the base of the crust. However, the localised nature of the uplift (only affecting the southern part of the basin) could equally be taken to imply that a more localised mechanism of uplift was responsible. The location of the plume centre remains uncertain; however it is believed to have been in the region of East Greenland or under the Faroe Islands (White 1989; 1992, Nadin *et al.* 1997, Jones *et al.* 2002) which would then account for the formation of the Greenland-Iceland-Faroe Ridge, which consists of thick oceanic crust formed during interaction between upwelling material from the plume and sea-floor spreading (Nadin *et al.* 1997).

The Faroe-Shetland Basin lies in close proximity to the proposed location of the plume and it is therefore speculated here that the whole of the basin would be expected to have experienced uplift related to plume activity. Transient dynamic uplift in the Palaeocene is recorded in the Faroe-Shetland Basin with a magnitude of up to 900 m on the Flett Ridge (Nadin *et al.* 1997). However, some of this uplift may be explained by permanent underplating (Brodie and White 1994; 1995). Uplift would be expected to be greater in the Palaeocene due to the proximity of the plume to the basin prior to the onset of sea-floor spreading and to the increased associated heat flow. By the Earliest Eocene the plume centre was further to the northwest and the reduction in observed uplift may be explained by the shift to an along-axis trend of the plume outflow rather than an axi-symmetric trend prior to break up (Nadin *et al.* 1997). However, the fact remains that no evidence for uplift is seen in the northern part of the Faroe-Shetland Basin during the earliest Eocene. It has been shown that the base T50 (Balder) unconformity is of local extent and developed during a short period of uplift that was soon replaced by the resumption in Eocene subsidence. If a plume driven uplift is to be believed, it is speculated here that evidence of uplift in the central and northern parts of the basin should be observed. This is most definitely not the case, but this could simply be due to the fact that the marine basin to the north was deep enough to suppress the uplift event, thus concealing any expression of an uplift event. Taking the minimum uplift of Smallwood and Gill (2002) in the south of the basin as 200 m, water depths of approximately 750 - 1000 m in the northern part of the basin would seem to have to be in place at the time of uplift for no evidence to be recorded. This figure seems feasible, though no water depth value is defined from this area of the basin, and it can only be loosely described as upper bathyal setting based on biostratigraphic information.

More controversially, the observed localised nature of the uplift may imply that the dynamic support generated by the plume was not the cause of the uplift associated with the development of the base T50 (Balder) unconformity. It may suggest that the causal mechanism was another local control or possibly a combination of interrelated factors. Indeed, a local plume with a narrow head of a few km cannot be totally excluded as a possible cause for such local transient uplift, but seems highly contrived nonetheless.

The presence of uplift on a local scale may reflect the size of the plume head or indeed the position of the plume in relation to the basin. The lack of any supporting evidence of uplift in the north and northeast of the basin, may rule out a mantle plume hypothesis for the uplift as it would be expected to be have a more regionally widespread affect. The effects of a mantle plume on the surrounding area are numerous. As well as the effect of significant uplift on a regional scale, there is additional heat flow into the lower crust and the associated intrusive and extrusive magmatic material. An ephemeral focussed thermal mantle anomaly will significantly alter the relative sea-level over a short period of time and have the effect of changing the sedimentary architecture. Furthermore, when combined with the more regional post-rift subsidence of a margin, an isolated uplift event such as the impingement of a plume can create semi-regional short-lived change in the basin architecture like that seen during the deposition of the T50 unit in the south of the basin. The effects of a mantle plume are not widely seen in the north of the basin as this remained a marine environment throughout the Palaeogene dominated by marine mudstones and silts. The recognition of short-lived unconformities or their correlative conformities in the northern part of the basin is difficult, both because of the basinal setting and the data quality.

4.4.3.1.4 Eustasy

A consideration of the global sea-level history throughout the Palaeogene is discussed here, as eustasy has a primary effect on sedimentary successions of passive margins. Originally, Haq *et al.* (1987; 1988) and more recently de Graciansky *et al.* (1998) produced global sea-level curves for the Mesozoic and Cenozoic eras. These curves were constructed using the positions of relative coastal onlap onto basin margins (from reflection configurations on seismic data and outcrop data) from many basins around the world. (For full explanation on the construction of the original Haq curve see **Section 3.2**). As has been discussed in **Section 4.4.2** above, the global sea-level curves of both Haq *et al.* (1987; 1988) and de Graciansky *et al.* (1998) highlight 3rd order cycles which have been tied to the eustatic curve (**Figure 4.57**). This pattern of 3rd order cycles is vastly different to the gross sedimentary cycles seen in this study as has been discussed in **Section 4.4.2.2.1**. Analysis of the long term eustatic curve (**Figure 4.57**) shows that during the Early Eocene (Ypresian) there was a slight initial rise in sea-level before a gradual sea-level fall is seen throughout the Middle Eocene (Lutetian - Bartonian). Towards the very end of the Eocene a final very small rise is seen in the Priabonian which continues into the Oligocene (**Figure 4.57**). Overall, a decrease in the eustatic sea-level from the start to the end of the Eocene epoch is observed despite the initial rise in the Ypresian. The amount of this decrease is approximately 50 m (from about 220 – 170 m).

However, analysis of the short term eustatic cycle shows major variations on a roughly million year frequency. These eustatic variations exhibit larger changes in the sea-level of up to 120 m (at the end of the Ypresian). The changes tend to average between 25 and 75 m, and these smaller variations are seen in the Early Eocene (Ypresian) and throughout the Middle and Late Eocene (**Figure 4.57**). These sharp falls in eustatic sea-level (highlighted on the short-term curve eustatic curve in **Figure 4.57**) occur at or immediately after major sequence boundaries interpreted by Haq *et al.* (1987) and dated by correlation with the Berggren *et al.* (1995) timescale. As has been mentioned earlier in **Section 4.4.2.2.1**, these sequence boundaries are not identified in the Faroe-Shetland Basin for a number of reasons.

In summary, the cycles observed from the published eustatic sea-level curves do not correspond well with any cycles seen in this study. This is to be expected due to the difficulty in categorically stating that falls or rises in sea-level can be unequivocally linked to eustasy. These rises and falls are observed from seismic data and in places can be correlated to periods of compression and plausibly with mantle plume activity. Therefore a relative sea-level reconstruction for parts of the Eocene is viable though a eustatic one remains impossible. Thus the history of sea-level fluctuations seen in this study does not match the global sea-level curve. This is because: (a) the published eustatic curve is considered to be suspect and (b) this study has recorded fluctuations in relative sea-level which have been shown to be heavily affected by local tectonic controls causing uplift and subsidence that varied both temporally and spatially throughout the basin.

4.4.3.2 Controls on Middle Eocene Deep Water Canyon – Deep Water Distal Lobe Systems

This section will look at more local or semi-regional factors that control the deep water fan systems and the canyon feeder systems described in the Eocene 3 seismic unit (see **Section 4.3.4**).

4.4.3.2.1 Controls on Fan Distribution

A first-order control on the location of the Eocene fans was the geometry of the basin at that time. Seismic interpretation has led to the recognition of the shelf-slope break to have been located in the southern part of Quadrant 213 and 214 with the base of slope to have been approximately located 60 km to the north (e.g. **Figure 4.49**). The position of the shelf-slope break and base of slope were similar during the earliest Eocene (during deposition of the T50 unit) (Lamers and Carmichael 1999). Slope canyons confined gravity currents that transported sediment captured from the shelf (the Shetland Platform) downslope (see **Section 4.4.3.2.2**). Basinward of the base of slope, deposition was less confined and the fans broadened out and became distal lobes (Davies *et al.* 2004).

Landward of the fans and feeder canyons, clinoforms are identified on the shelf. These clinoform geometries on this south eastern margin of the basin vary in height throughout deposition of the Middle Eocene aged fan system. However, from these clinoforms, water depths on the shelf at this time are estimated to have been approximately 250 m and thus the fans down-dip from the feeders are interpreted to be deposited in greater depths possibly more than 1000 m.

The examination of the Bouguer gravity map of Davies *et al.* (2004) (**Figure 4.61**) in the region where the distal lobe deposition occurred shows that there is a very close spatial correlation between the relatively low Bouguer gravity values (-0 mGal to –11 mGal) and the location of the northern parts of the Strachan and Caledonia fans that trend in a more northeast – southwest orientation (**Figure 4.61**). The relatively high Bouguer gravity values (red and yellow colours) correspond to basement highs formed by footwall uplift of extensional faults that have been mapped in this region by Dean *et al.* (1999). Whereas the relatively low Bouguer gravity values correspond to the hanging-walls of extensional faults where sedimentary rock thickness is greater. Eocene sediments appear to have been focused into these hanging-wall depocentres. However, the extensional faults are Cretaceous in age. Even faults active during the Palaeocene tip-out below the Eocene (Dean *et al.* 1999) and no rift-related faults are interpreted to have been active during the Eocene. Instead much of the Eocene interval is characterized by polygonal fault networks (e.g. Davies *et al.* 1999) especially seen in the northern half of the basin that are considered to be related to dewatering of smectite-rich claystones (e.g. Cartwright and Dewhurst 1998). Therefore

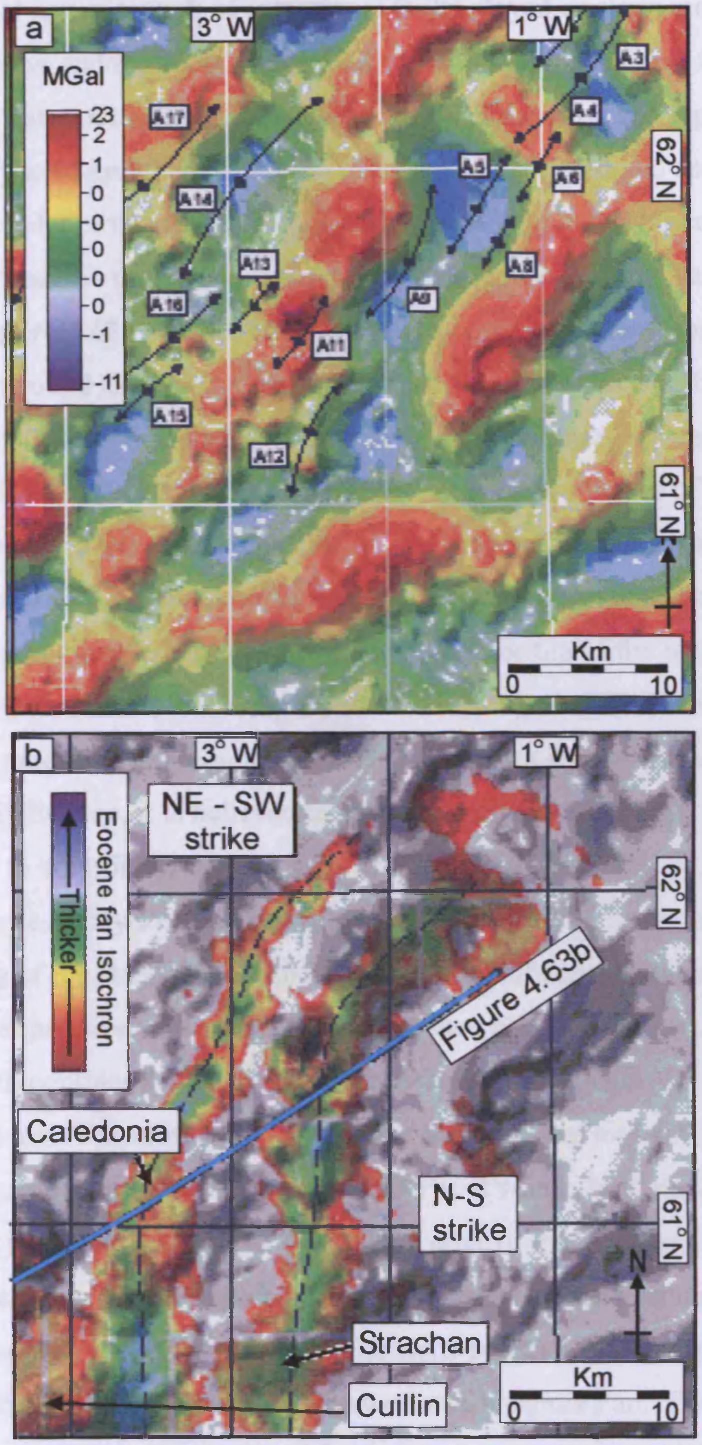


Figure 4.61. a) Residual Bouger gravity map (with 40-km high pass filter) from northern part of the Faroe-Shetland Basin showing the northwest - southeast trending gravity highs (red colours e.g. the Corona Ridge)and the compressional axes located. b) This map is the same Bouger gravity map as in a) but with grey shading. Superimposed on this is the isochron map of the Strachan and Caledonia fan systems. From this map there is a close spatial correspondence between the location of gravity lows (marked in blue colours in map a) and the thickest fan development. Red and blue colours on the isochron map are equivalent to thicknesses of approximately 50 m and 200 m respectively. Note the more southerly north - south trend present in the southern and central parts of the fan which then switches to a more northeast - southwest trend towards the distal end of the fan. The position of the seismic line in Figure 4.63b is shown. Modified from Davies *et al.* (2004).

it is unlikely that active growth of extensional faults played a role in modifying sediment pathways and sediment thicknesses. The lack of extensional faulting in the Eocene section rules out a direct fault control. Instead the close correspondence between gravity lows and the thickest development of the Caledonia and Strachan fans is compelling evidence to suggest that post-rift compaction above basement highs generated by Mesozoic rifting and enhanced by Palaeocene faulting was the primary control (see **Figure 4.62**). Davies *et al.* (2004) propose that compaction generated bathymetry controlled the orientation of the terminal northern ends of Eocene fans which are likely to be composed of weakly confined distributary channel complexes.

Towards the southern limits of the fans (e.g. southern halves of quadrants 213 and 214), there is positive relief on the top of the Strachan, Caledonia and Cuillin fans (**Figures 4.36** and **4.61**). This relief is interpreted as depositional and can also be seen in overlying draping claystones as well as onlap onto the top of the fans.

On a regional scale, the fan systems are never observed to the west of the Faroe-Shetland Escarpment. Even though this palaeo-lava delta is onlapped and buried during the Early Eocene, it is believed to have been structurally higher than the basin centre in order to control the western limit of the fan fairways. The three fan systems are seen to progressively step and young to the west. The process controlling this lateral stepping of the stacking pattern may be due to differential compaction and bathymetric compression of the sand bodies and their clay-rich host sediments through time, along with continued post-rift subsidence which is greater in more western (basinal) positions. Therefore, once the earliest Strachan Fan was deposited, continued subsidence focused subsequent deposition to the west where the Caledonia Fan was then deposited in the depressions on the western edge of the Strachan Fan. These marginal fan sediments of the Strachan Fan are significantly less sand prone than those in the axial sand-rich belt (**Figure 4.63**). The same sequence of events also occurred with the Cuillin Fan being deposited to the west of Caledonia Fan. The outcome of this differential compaction is a stacking of the fans to the northwest, which mirrors the progressive movement of the entire depocentres seen on the Northeast Atlantic margin (from the oldest Jurassic and Cretaceous basins of the West Shetland Basin through the Palaeocene Flett sub-basin to the Eocene to recent Faroe-Shetland Basin). This localised stacking of the three fan systems is termed compensational stacking (e.g. Mutti and Sonnini 1981). The trend of this compensational stacking towards the

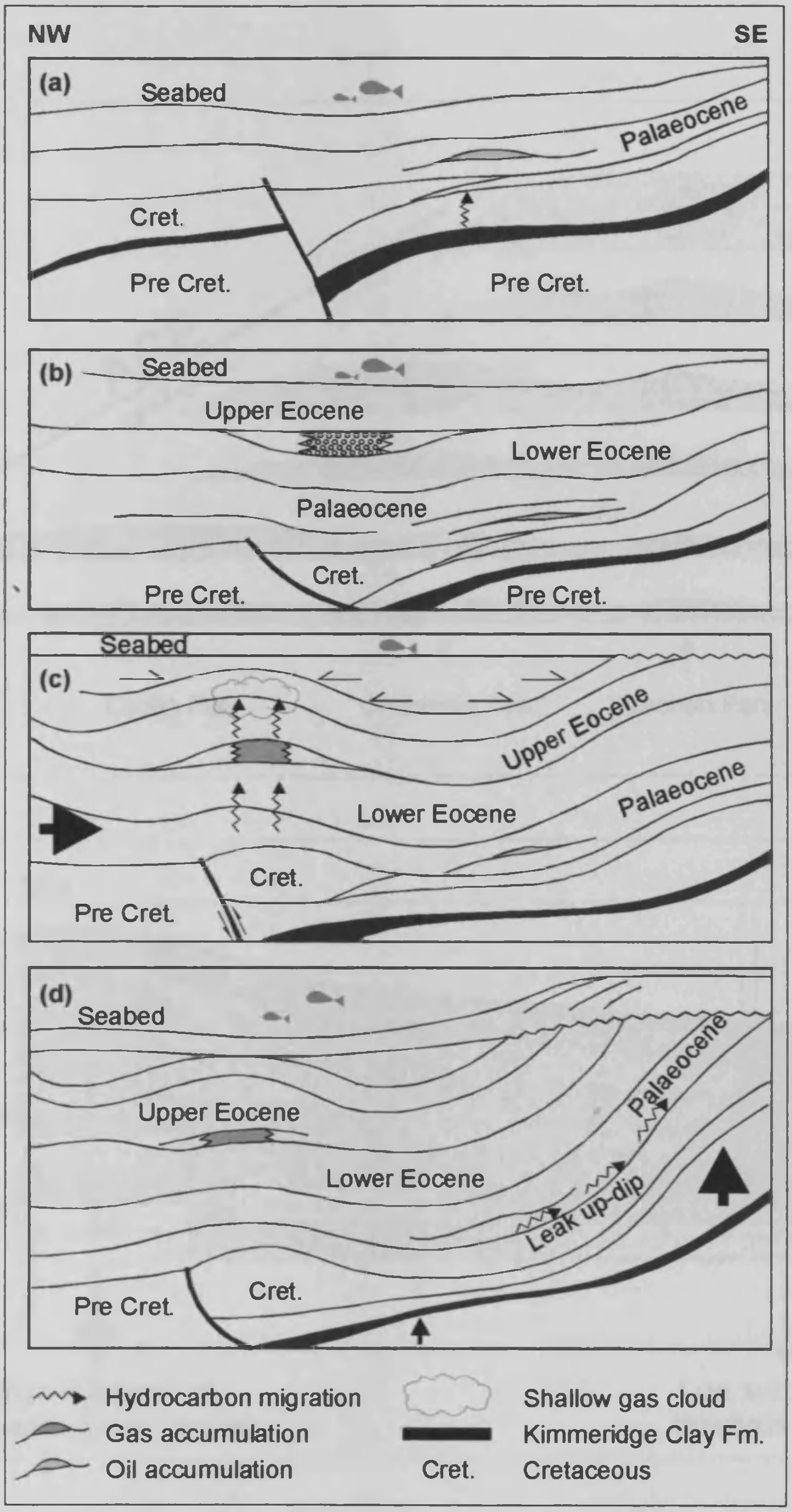


Figure 4.62. Schematic cartoon (southeast - northwest) illustrating the impact of compression and differential uplift on the development of the fan systems. Extension occurred during the Cretaceous. a). Compaction of Palaeocene and Eocene sediments created bathymetric depressions above the hanging-walls of the extensional faults. b) Eocene submarine fans fed from the south were confined by northeast - southwest striking hanging-wall bathymetric lows. c). Compression during the middle Miocene caused contractional reactivation of Mesozoic faults and hanging-wall anticlines developed, folding the Eocene fans at the crests. d). Differential uplift of the Shetland Platform during the Early Pliocene caused tilting of Palaeocene traps and up-dip leakage of hydrocarbons. Large black arrows in c) and d) indicate compression and uplift/subsidence respectively. From Davies *et al.* 2004.

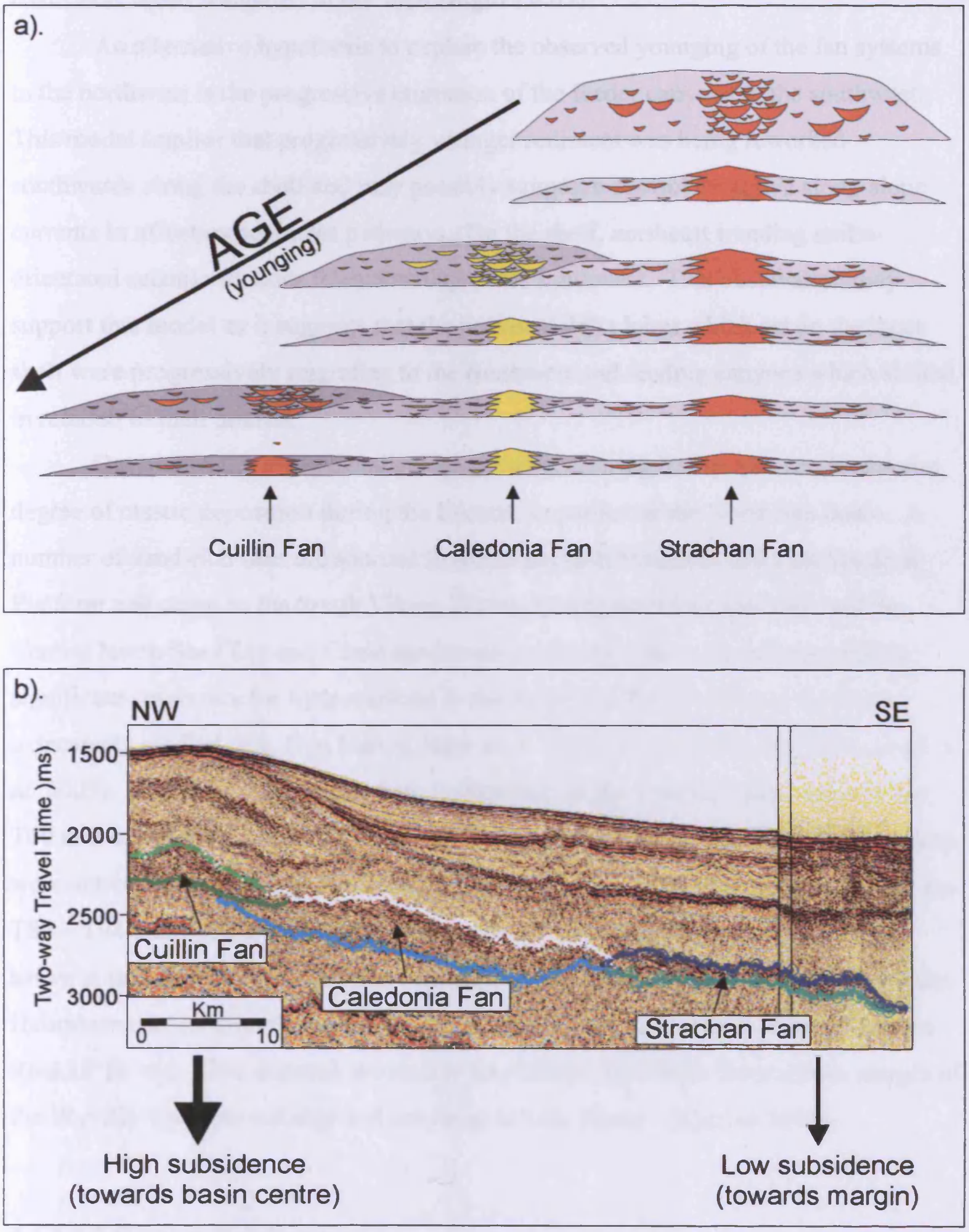


Figure 4.63. a). Schematic cartoon illustrating the development of the three major fan systems that were deposited during the Middle Eocene (Eocene 3 unit). The earliest fan (Strachan) develops with a high sand content in the central part of the channel system. On the flanks of the fan, there is significantly less sand-grad material and more mudstone and siltstone. Differential compaction of the Strachan Fan causes the flanks to form side lows into which subsequent deposition of the next channel/fan system (Caledonia) develops. This is known as compensational stacking and in this instance the fans young and stack to the northwest as shown in the seismic panel b). The channels may stack and young to the northwest because of greater subsidence in the central part of the basin (towards the northwest) than towards the southeast on the Shetland margin. The location of the seismic line is shown in Figure 4.61b.

northwest in this case may be due to the increased relative rates of subsidence in the basin axes when compared to the shelf (**Figure 4.63**).

An alternative hypothesis to explain the observed younging of the fan systems to the northwest is the progressive migration of the feeder canyons to the southwest. This model implies that progressively younger sediment was being reworked southwards along the shelf and may possibly suggest a control by active along-slope currents in affecting sediment pathways. On the shelf, northeast trending strike-orientated seismic sections lines downlap to the southwest. This observation may support this model as it suggests that the proximal delta lobes which sat on the inner shelf were progressively migrating to the southwest and feeding canyons which shifted in relation to their source.

Outwith of the Faroe-Shetland Basin, neighbouring basins also received some degree of clastic deposition during the Eocene, in particular the North Sea Basin. A number of sand-rich fans are sourced from the Scottish Mainland and East Shetland Platform and occur in the South Viking Graben (Frigg sandstone member) and the Central North Sea (Tay and Caran sandstone members). These sandstones provide significant reservoirs for hydrocarbons in the North Sea Basin and have thus been extensively studied (e.g. Den Hartog Jager *et al.* 1993, Jones and Milton 1994, Jones *et al.* 2003). The large Frigg fan system is deposited in the Ypresian (equivalent to the T60 unit of Jones and Milton (1994) and Ebdon *et al.* (1995)). The Tay and Caran fans were active over a longer duration though the main clastic material occurred during the T84 – T94 units of the Middle Eocene (Lutetian) and are more likely to have been active at the same time as the three fans observed in the Faroe-Shetland Basin. On the Hebridean margin significant Palaeogene clastic deposits are seen in the northeastern Rockall Trough. The deposits are seen to be point sourced from the southern margin of the Wyville-Thompson Ridge and are dated as Late Eocene (Egerton 1998).

4.4.3.2.2 Canyon Development

The fan systems exhibit well developed canyons which have been mapped on both the 3-D and 2-D seismic data. Both the Strachan and Caledonia fans have well developed canyons on the 3-D data and are found in blocks 213/29, 213/30 and 214/26 respectively. The Cuillin Fan to the west has a less well developed canyon seen on 2-D seismic data. All of the canyons have an overall broadly north - south orientation. At the base of the slope the canyons broadly bend to give a northeast - southwest orientation of the more distal lobes (which are sand dominated compared to the conglomeratic canyons).

As has been discussed above, these three canyons young to the southwest, and two hypotheses have been put forward to explain this observation. An additional observation is that the Middle Eocene-aged canyons appear in the same location as the much older Palaeocene canyons documented by Naylor *et al.* (1999) and Lamers and Carmichael (1999) (see **Figure 2.10**). This is an interesting observation bearing in mind that between the Late Palaeocene and the Middle Eocene there was a major transgression of the coastline during deposition of the earliest Eocene T50 unit (Balder Formation equivalent) which drowned the Late Palaeocene aged deltas. Middle Eocene aged deltas are seen to prograde out across the shelf and the shelf-break is located in a similar position.

The observation of the similar positions of the Palaeocene and Eocene canyons may be a coincidence or it may suggest that there is a fundamental control over the development of canyons. This control may be tectonic in that the canyon positions may be controlled by a pre-existing lineament or possible proposed transfer zones in the basin (e.g. Rumph *et al.* 1993). Conversely a stratigraphic control may exist including the amount of progradation and the angle of clinoform development prior to possible failure at a delta front. The outer shelf of a progradational margin can be starved of sediment during a transgressive episode if the rate of relative sea-level rise is greater than the rate of sediment supply (Ross *et al.* 1994). During subsequent progradation an oversteepening of the slope ensues. During slope re-adjustment slope erosion (canyons), sediment bypass and submarine fan-aprons may develop (Ross *et al.* 1994). There is some evidence to support the latter hypothesis of slope failure from well 214/26-1. This well was located in the Strachan canyon and encountered conglomerates interbedded with sandstones. Within one of the conglomeratic beds, a single large boulder has been identified which is clearly inverted and shows evidence of bioturbation. The boulder consists of interbedded sand and mud and probably represents an outer shelf deposit which has been incorporated into the conglomerate during reworking. Smaller clasts within the conglomerates are similarly bioturbated and several show signs of slumping and resedimentation prior to incorporation into the conglomerate. Oversteepening of the delta front and proximal sourcing of large clasts and boulders is therefore a valid mode of formation for the deposition of these conglomerates. The interbedded sandstones may reflect more distal equivalents of the conglomerates.

A larger scale control on the amount of delta progradation may be apparent through the crustal loading of large sedimentary wedges on the margin. A threshold of sediment thickness may be reached creating flexural loading and bending causing instability on the outer shelf and slope and thus causing the gravity flow deposits. Additionally, the hinge position on the margin may affect the gross scale of the sedimentary wedge and have a deep seated control on the development of prograding systems on the margin. A similar situation has been reported from the Namibian volcanic margin where a pronounced hinge-line is controlled by deep crustal structure and influences slope stability throughout the Late Mesozoic and Cenozoic (Clemson *et al.* 1997). The position of the deep basin bounding faults on the Shetland margin may have a fundamental control on why sediment progrades to the same position on the shelf before canyons developed at the shelf-slope break.

Simply relying on a purely stratigraphic control in determining the position of the shelf-break seems implausible. A reason for this could be the vast lithological difference between the Palaeocene and Eocene successions. The Palaeocene sediments have a much higher sand-grade content that was principally dominated by the deposition of turbidites and other clastic deep water gravity flows (e.g. Lamers and Carmichael 1999). The Eocene succession in comparison is not sand dominated, the three fan systems constituting the main deep water sand fairway throughout the whole of the epoch.

4.4.4 Two-Dimensional (2-D) Seismic Data: Limitations and Difficulties.

This final section will briefly discuss the nature of the pragmatic 2-D regional seismic interpretation approach that has been used in this study highlighting uncertainties and limitations of the dataset. Limitations with this kind of interpretation are inevitable because of the large area the dataset covers and in relatively small detail. Because of the large distances between 2-D lines (which varies throughout the basin from a minimum of 1 - 2 km to, in places, over 20 km) a lot of the detail in the seismic data is missed. The detail which is missed is often very important and, even though large amounts of information can be obtained from looking at the regional 2-D surveys, when analysing small-scale features a different approach needs to be taken.

2-D seismic data is usually the most practical tool to use when undergoing a basin analysis study, unless the basin is completely covered by 3-D data. It is ideal for the correlation of regional seismic reflections across large areas and allows for broad scale stratigraphic frameworks to be constructed. Occasionally, 2-D seismic data can give the interpreter clues as to the stratigraphic significance of seismic reflections and sequences or units can be defined. However, what is seen is invariably only a small percentage of the true amount of information which can be gleaned from high resolution 3-D seismic data.

Couple the problems of 2-D seismic interpretation with the poor biostratigraphic control in the Faroe-Shetland Basin and it becomes apparent that this regional study represents a broad generalised attempt to chronostratigraphically sub-divide the Eocene of the Faroe-Shetland Basin.

4.5 Conclusions

- A new seismic-stratigraphic framework for the Eocene succession has been constructed for the entire Faroe-Shetland Basin using 2-D seismic data and where possible tied to high quality biostratigraphic data from wells and BGS boreholes.
- The stratigraphic succession of the Eocene has been sub-divided into four seismic-stratigraphic units. The surfaces bounding these units are seismic reflections that are the most correlatable around the basin. The seismic reflections chosen have the most widespread geographical extent and often appear as high amplitude continuous to semi-continuous configurations.
- An unconventional approach of mapping seismic reflections is used in this study as the conventional approach of mapping sequence boundaries or maximum flooding surfaces is not possibly in this basin. This is because the long-range correlations in this basin are so prone to significant errors that it is unfeasible to attempt a study based on huge data problems. Therefore chronostratigraphic timelines based on seismic reflections that are given the highest degree of confidence (tied to wells) are used.

- Early Eocene deposition (of the Eocene 1 seismic unit) is focussed in the southern half of the basin with a large deltaic environment developed in Quadrants 204 and 6004 and 6005. This deltaic succession was deposited during a relative sea-level rise which developed in response to subsidence related to the continued movement away of the Iceland mantle plume. Sediments were fed from the Orkney landmass to the south, whilst the north of the basin remained starved and experiencing deep water conditions.

- Subsequent deposition in the early Middle Eocene (Eocene 2 seismic unit) was focussed on the Shetland margin and the southern part of the basin which was an active depocentre during the Early Eocene became an area of erosion (by Top Eoc2). This large scale shift in focus of the depositional systems is believed to because of early compression on the Munkagrunnur Ridge, Wyville-Thompson Ridge and the Judd High/Anticline complex, which caused erosion over the anticline and possible separation of the southern part of the basin.

- Post Early Eocene sedimentation was dominated by clinoform progradation from the Shetland margin. Little or no sediment is interpreted to have come from the slowly subsiding Faroe Platform during the Eocene.

- Major deep water fan systems are seen to dominate the Middle Eocene succession (during deposition of the Eocene 3 seismic unit). Three major fans are documented are controlled by a number of factors including the position of canyons located on the shelf-edge and the basin physiography and bathymetry. Indeed, the Faroe-Shetland Escarpment and the Corona Ridge have an impact on the location and thickness of the fan systems. Some degree of slumping or collapse at the shelf-edge may contribute to the gravity processes depositing the fan systems. Three fans are documented and they are seen to young to the northwest.

- Deep water fans are seen to develop in the Late Eocene (during deposition of the Eocene 4 seismic unit) however their geometry is vastly different. The Middle Eocene elongate fans have been replaced by oval or circular bodies which are much smaller in dimension. This may be a consequence of the basin physiography or a decrease in sedimentary supply.

5. Chapter Five: Seismic-Stratigraphy of the Lower Eocene in the South Judd Basin

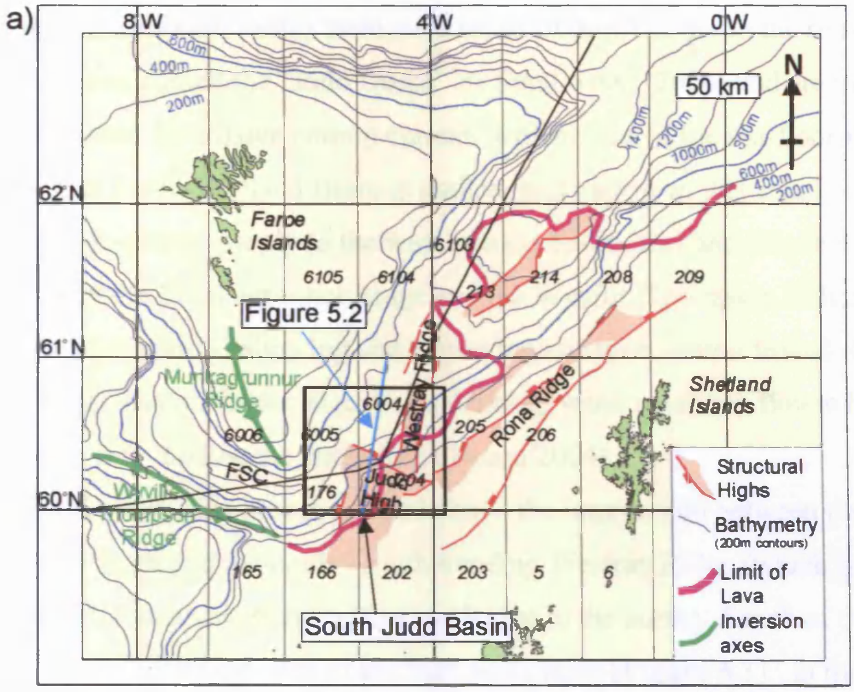
5.1 Introduction

This is the first of two chapters that will use multiple 3-D seismic data volumes to describe in detail a small part of the Faroe-Shetland Basin in a specific part of the Palaeogene succession.

A detailed study of depositional systems from the southern margin of the Faroe-Shetland Basin is examined here. The Early Eocene delta system (Eocene 1) which was discussed in **Chapter 4 (Section 4.3)** is sub-divided into a number of sub-seismic units. This area is crucial for the understanding of the early development of the basin as this was the depocentre for much of the early sediments and was influenced by regional compressional tectonics creating folding and uplift on many local features including the Judd Anticline (see **Section 5.2**). This basin which is located in the southern part of the Faroe-Shetland Basin is herein termed the South Judd Basin (**Figure 5.1**). The focus of this chapter is to detail the sub-division of the Eocene 1 seismic unit to highlight the response of the depositional systems in a marginal area which was susceptible to changing relative sea-level.

5.2 Regional Geological Setting

The study area is located close to the political boundary between the UK and the Faroese owned waters. The area concerned covers approximately 150 – 200 km^2 over Quadrants 204 and some of 205 (in the UK sector) and Quadrants 6004 and 6005 on the Faroes side (**Figure 5.1**). This part of the basin is found in relatively shallow waters averaging between 200 and 400 m but deepening to a maximum of about 800 m to the north and west of the study area (**Figure 5.1**). The south eastern part of the study area is located in the area of the present day shelf-break which trends northeast (see regional correlation A). The edge of the shelf-break is visible on the dip map shown in **Figure 5.1b** and recent debris flows and slumps are seen to be present on the slope. Furthermore, in the central part of the study area (located in all three of the 3-D seismic



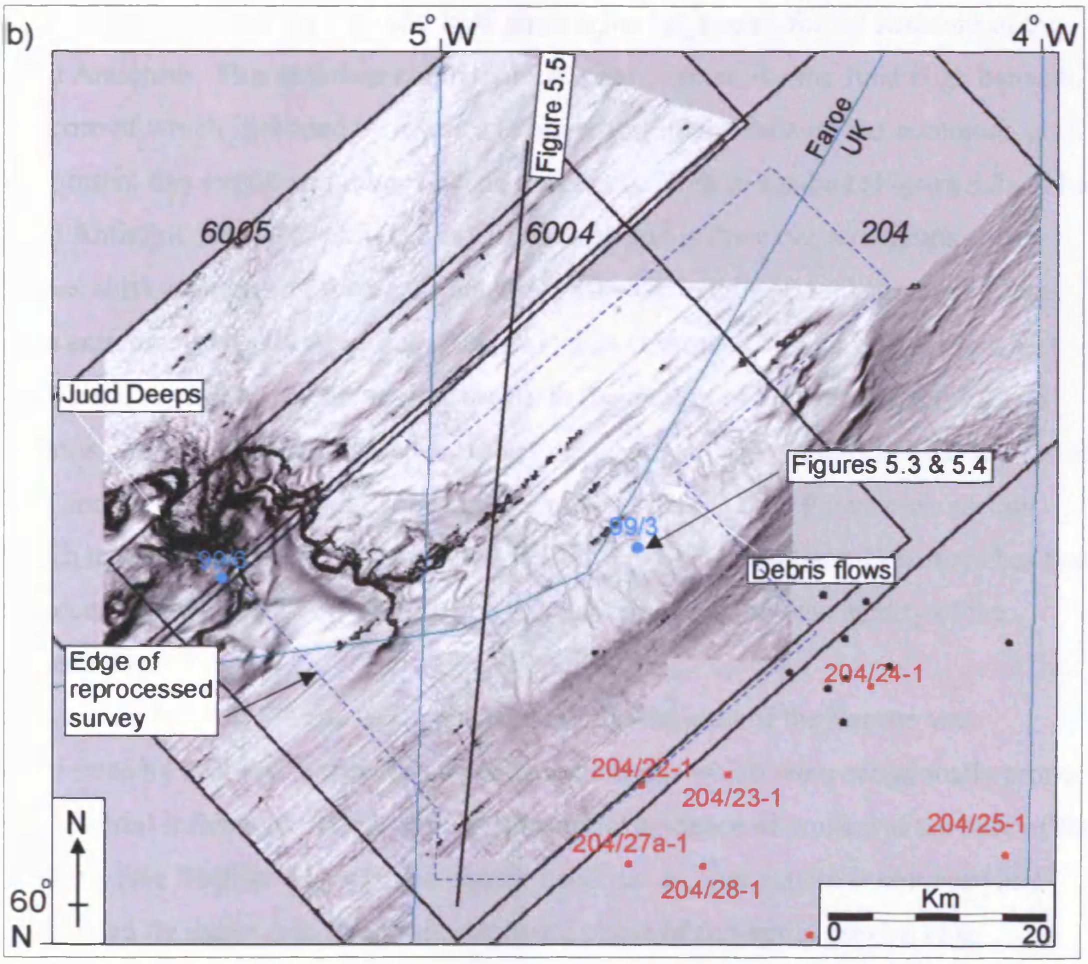


Figure 5.1. Location maps showing position of the case study in relation to the Faroe-Shetland Basin. a) Study area located in south of basin in area of relatively shallow water depths (see contours). Location of the 2-D seismic line in Figure 5.2 is highlighted. b) Zoomed in location map showing dip image of the seabed from the three 3-D seismic surveys used in this case study. Note the cuspate erosional scarps of the Judd Deeps and the present day debris flows from the shelf-break. Key wells (red) and boreholes (blue) are highlighted as well as locations of Figures 5.3, 5.4 and 5.5.

surveys used in this case study) there are a series of deep scarps on the sea-floor. These scarps have been named the "Judd Deeps" by Smallwood (2004) and are interpreted to have been formed by erosive bottom currents which scoured the sea-floor in the Neogene. The Faroe-Shetland Basin is shallow and narrow in this area before the basin axis changes direction sharply to the west into the Faroe-Shetland Channel or Munken Basin between the Munkagrunnur Ridge and the Wyville-Thompson Ridge (**Figure 5.1a**). This southward shallowing and narrowing creates a natural funnel-shaped geometry to the basin which concentrates the deep water southerly flow which causes erosion down into the Eocene strata (Smallwood 2004).

Additionally, this area is situated above the intersection between the east - west trending Judd High and the north - south trending Westray Ridge (which links up to the more northeast-southwest trending Corona Ridge in the north). South of the Judd High is the more stable platform area of the Sula Sgeir High (**Figure 5.1**). In the Palaeogene strata both above and north of the Judd High is the large open folded structure of the Judd Anticline. This anticline similarly trends east – west like the Judd High beneath, the core of which is eroded by a series of unconformities. Erosion has continued until the present day exposing Lower Eocene rocks at or close to sea-bed (**Figure 5.2**). The Judd Anticline is seen to plunge towards the east and is therefore stratigraphically higher at its western end where it joins the northwest trending Munkagrunnur Ridge. This anticline is well imaged in 3-D seismic data (**Figure 5.2**) where its distinct east - west trend is visible. In this area of the basin the quality of the data is good for two reasons. Firstly, the shallow depth of the Eocene stratigraphy results in high resolution data and there are no igneous rocks (sills and lavas) in the Late Palaeocene section which attenuates the seismic data below. Additionally, one of the 3-D surveys has been reprocessed for the shallow section and this has enhanced the data quality of the Eocene interval.

The basin at the end of the Palaeocene and the start of the Eocene was dominated by shallow, marginal marine environments which were occasionally prone to terrestrial influences. Here, there is substantial evidence of erosion at the base of the T50 unit (see **Section 4.2**) where a classic dendritic erosion pattern is observed and interpreted by many authors as representing a phase of sub-aerial erosion (e.g. Smallwood and Gill 2002). Furthermore, there are observations of high amplitude seismic reflections within the T50 unit (Balder Formation equivalent) which may reflect occurrences of coals and prevalent deltaic sediments including interbedded

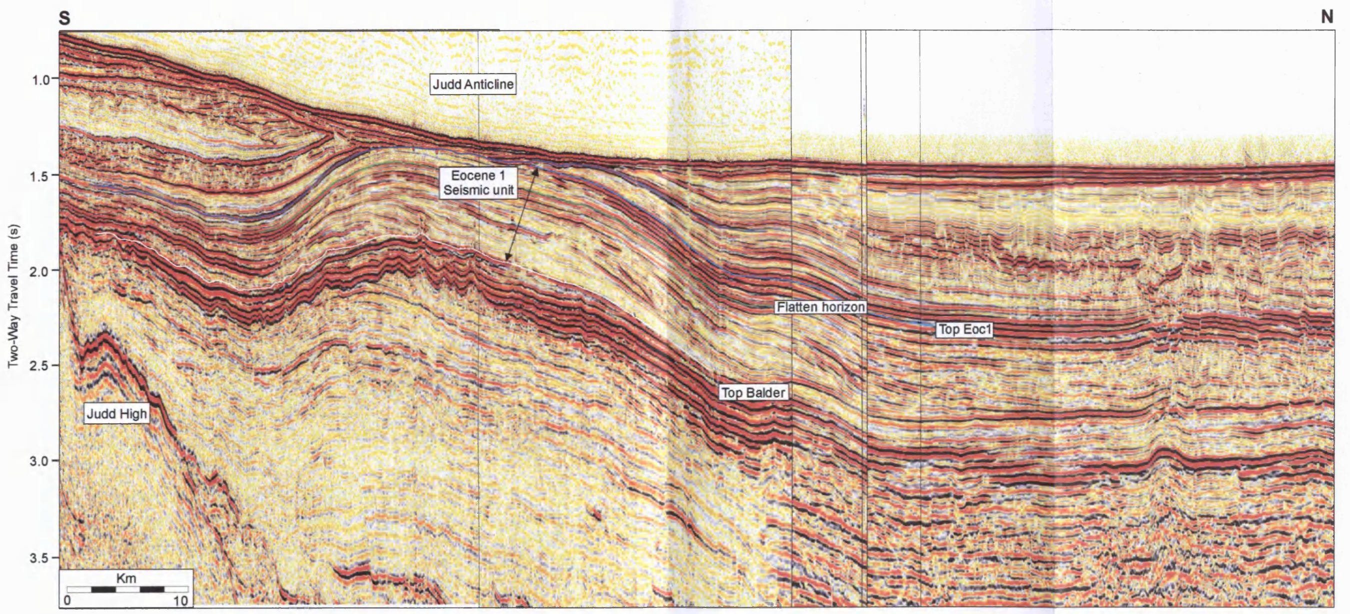


Figure 5.2. Composite seismic traverse broadly trending north - south across the South Judd Basin. Main features highlighted are the Judd High in the south and the east-west trending Judd Anticline that sits to the north of the high. The Eocene 1 seismic unit is seen to be relatively thick over the anticline which was folded at a later date. Within the Eocene 1 unit, other high amplitude continuous reflections are seen which have been mapped in detail throughout the three 3-D surveys. Additionally, the technique of seismic flattening has been used in this chapter to take out the post-depositional effects of the Judd Anticline. This flattening as been used on a strong continuous reflection that is not eroded by the Top Eoc2 reflection. For location of seismic line see Figure 5.1a.

sandstones, siltstones and mudstones. These lithofacies are present in many wells in the area seen in the south of Quadrant 204 and the north of Quadrant 202 (see **Section 4.2.4** and **Figure 4.5**).

5.3 Eocene 1 Seismic Unit: Regional Character

5.3.1 Introduction

This section will focus entirely on the Eocene 1 seismic unit (defined and broadly discussed in **Section 4.3.2.3**) and detail the internal architecture within this unit. In **Chapter 4** this unit was not sub-divided by internal reflections but the availability of high quality 3-D seismic surveys has enabled a higher resolution of mapping that allows for further sub-division of the unit.

5.3.2 Biostratigraphic Calibration

Within the South Judd Basin, there is substantial well and borehole data to calibrate a crude age to the Eocene 1 succession. No absolute ages for the succession are available as the biostratigraphic data is relatively poor. However, it is possible to use the two boreholes 99/3 and 99/6 and a number of wells (**Figure 5.1**) in the north of Quadrant 202 and the south of Quadrant 204 to allow for calibration of the Eocene 1 seismic unit (**Section 4.3**). The 99/3 borehole is most useful in constraining the relative age of the sediments. It is located in block number 204/17 and drilled to a depth of 166.5 m and consists of Quaternary sediments immediately underlain by an Eocene succession (**Figure 5.3**). This Eocene succession comprises of a very thin (7 – 10 m) Upper Eocene aged siltstone and claystone which is underlain by a Middle Eocene succession of mainly mudstone (which is calibrated as early – mid Lutetian by the appearance of the dinoflagellate cysts *Diphyes ficusoides* and *Heteraulacacysta leptalae* (Riding *et al.* 1999a, 1999b) and is approximately 70 m thick. Lower Eocene (Ypresian) sandstone and shale underlies the Lutetian mudstone (**Figure 5.3**) and calibrated by the occurrence of *Dracodinium vareilongitudum* (BGS 99/3 internal report). A shallow sparker seismic line through the location of the borehole reveals

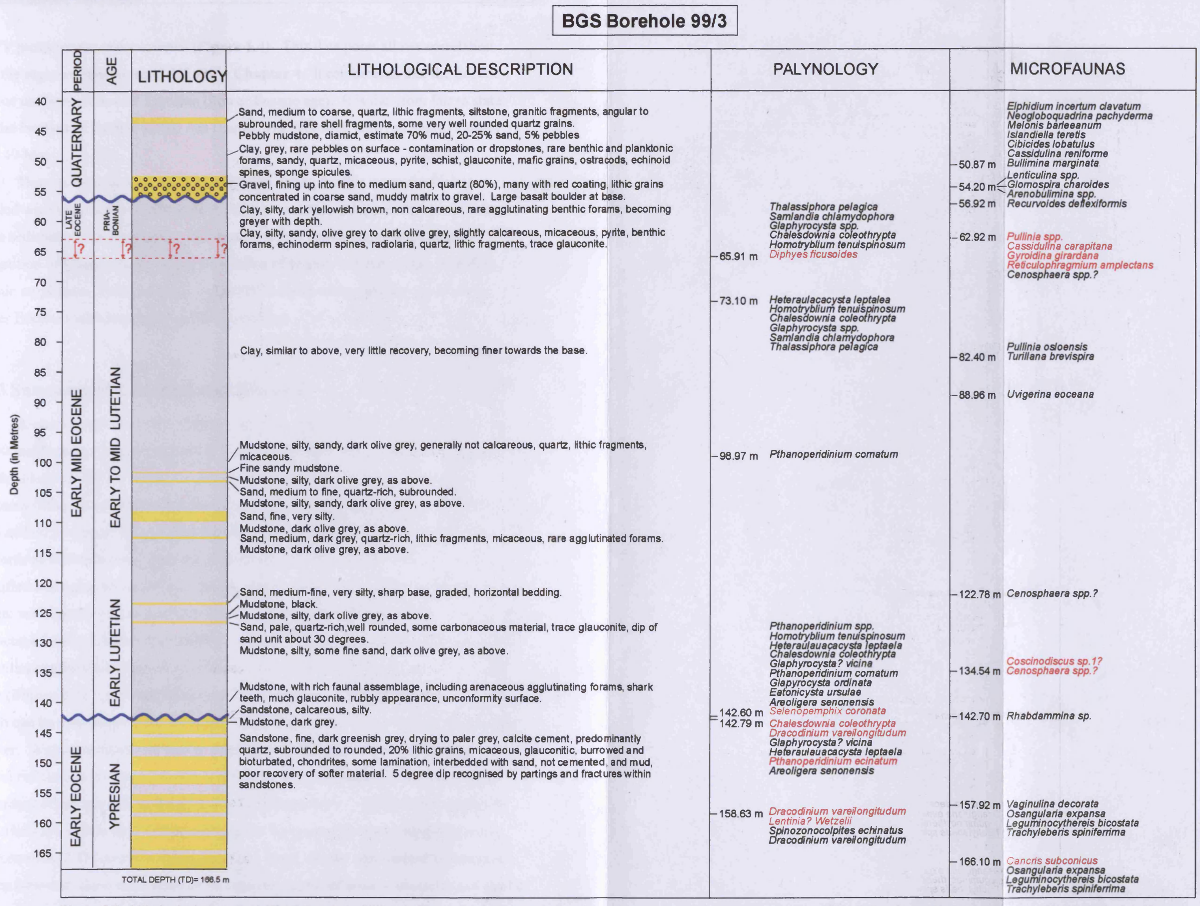


Figure 5.3. Lithological summary of BGS Borehole 99/3 (see Figure 5.1 for location) showing a thin Quaternary cover succession underlain by a very thin Upper Eocene (Priabonian) claystone. An approximately 70 m mudstone dominated interval of early-mid Lutetian (Middle Eocene) age is recorded underlain by a much more sand-grade Ypresian (Lower Eocene) succession. This borehole has been used to date the Eocene 1 seismic unit (of chapter 4) as Ypresian and is the best data point for the Lower Eocene sediments discussed in this chapter. Red bio-events are indicative to specific age, black bio-events are common occurrences given general epoch or age. The position of the corresponding seismic reflections of the age boundaries are shown in Figure 5.4.

TWTT picks for the entire interval (**Figure 5.4**). This data point allows correlation with the regional seismic units defined in **Chapter 4**. It can be seen that the Eocene 1 seismic unit is therefore of Ypresian (Lower Eocene age). It is therefore fair to state that the interval of interest within this chapter was deposited between approximately 55 Ma - 49 Ma.

There is therefore generally poor quality dating for this interval which is coupled with the paucity of core data in order to ascertain a depositional environment of the sediments. Because of this data quality it puts a great importance on the recognition of seismic facies and interpretation of seismic attribute maps, including seismic amplitude. There remains a great difficulty in calibrating the age of these Lower Eocene sediments to any greater resolution.

5.3.3 Summary of Distribution of Eocene 1

A quick overview of the Eocene 1 seismic unit shows it is found as a thick (up to 800 ms) broadly circular package in the southern half of the Faroe-Shetland Basin (south of latitude 61°N). Here, the Eocene 1 unit is found to be at a maximum thickness (of up to 800 ms) in the central part of the circular body located in Quadrants 6004 and 6005 west of the political boundary (see **Figure 4.14** and **Section 4.3.2.3**). In the north of the basin away from the thick circular body, a thin Eocene 1 unit has been identified averaging about 200 ms, though locally increasing to 400 ms on the Shetland margin and directly east of the Faroe-Shetland Escarpment (**Figure 4.14**). The seismic character of the unit is one of relatively moderate to high amplitude continuous seismic reflection configurations which parallel one another but are also sub-parallel to each other (**Figure 5.2**). However, of interest in this unit are the internal reflections, some of which can be mapped over the entire area of the three 3-D seismic surveys used in this chapter. Within the moderate to high amplitude continuous and parallel reflections, certain reflections are locally visible which have a higher amplitude and are slightly discordant or sub-parallel to the overall reflection geometry. These higher amplitude reflections are visible on 2-D data, although in this part of the basin the grid spacing between these 2-D lines is in the region of 1 - 3 km. On the reprocessed 3-D seismic survey however, these reflections show a greater degree of seismic character and can be mapped extensively.

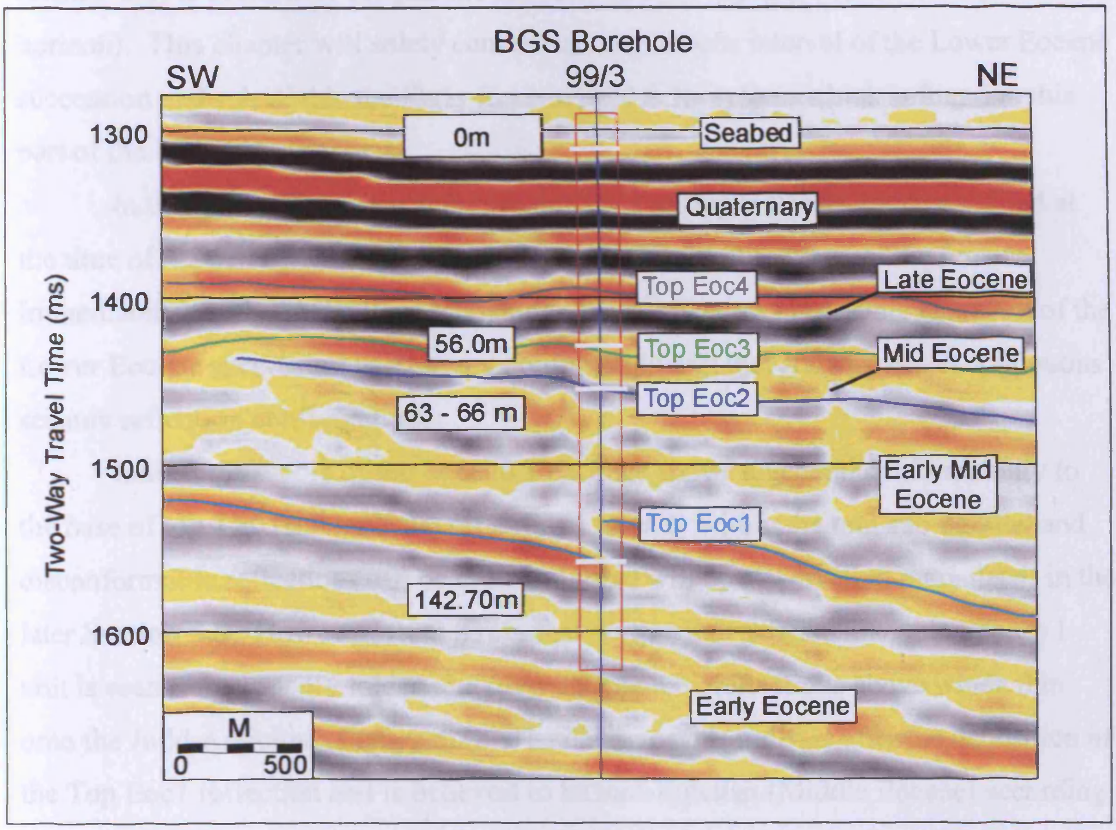


Figure 5.4. Southwest - northeast seismic line through the location of borehole 99/3 showing the positions in Two-Way Travel Time (TWTT) of the relevant seismic markers of the Quaternary to Lower Eocene succession. The Lower Eocene succession comprises of sandstones and shales that are Ypresian in age (see Figure 5.3 for detailed lithological descriptions) and are wholly found within the Eocene1 seismic unit described in chapter 4. It is therefore fair to state that the interval of interest within this chapter is entirely are of Lower Eocene in age and represent deposition of sediments between approximately 55 Ma - 49 Ma (Ypresian). However it remains difficult to calibrate the age of these Lower Eocene sediments to any greater resolution. See Figure 5.1 for location of borehole.

The deposition of the Balder Tuff was a geologically isochronous event and marking the start of sea-floor spreading in the North Atlantic (Knox and Morton 1988 and **Section 2.3**). The top Balder Tuff reflection defines the top of the T50 unit of Ebdon *et al.* (1995) and base of the Eocene 1 seismic unit. The top of the Eocene 1 seismic unit is defined by the seismic reflection called Top Eoc1 (shown as the blue horizon). This chapter will solely concentrate on this one interval of the Lower Eocene succession and sub-divide the Early Eocene aged delta system which is found in this part of the basin.

In this part of the basin a marginal marine setting is known to have existed at the time of deposition of the Balder Tuff (top T50 unit) at the start of the Eocene. Immediately overlying the top T50 (Balder Tuff) reflection the seismic character of the Lower Eocene succession is seen as a conformable unit that shows parallel continuous seismic reflection configurations.

Indeed the whole of the Eocene 1 package shows a general conformability to the base of top T50 (Balder Tuff) reflection. Locally within the unit sub-parallel and disconformable reflections can be seen and these will be discussed in more detail in the later **Section 5.5**. However, from a regional north-south seismic line, this Eocene 1 unit is seen to be broadly folded and then onlapped by reflections above which thin onto the Judd Anticline (**Figure 5.2**). The dating of this uplift is after the deposition of the Top Eoc1 reflection and is believed to be mid-Lutetian (Middle Eocene) according to Smallwood (2004). This is constrained by major thinning and onlap of Middle Eocene seismic reflections onto the east - west trending Judd High.

Prior to this tectonic activity during the Early Eocene there are parallel high amplitude reflections representing the Lower Eocene sediments which would have been near horizontal at the time of deposition. It is possible to flatten the seismic data on any mapped reflector and this has the effect of taking out the folding on the Judd Anticline and therefore restores the seismic data to its approximate (original) position at a specific time. Flattening of the seismic data has been carried out using a strong continuous parallel reflection which is readily mapped around the whole of the seismic survey. Furthermore it is essential that when flattening the data, a reflection which is part of the folded stratigraphy is chosen so the folding is removed (**Figure 5.2**). In this example a continuous reflection just under the Top Eoc1 reflection is used as this is entirely within the folded Eocene 1 seismic unit and is not eroded by later unconformities (green horizon in **Figure 5.5**). The Top Eoc1 cannot be used to flatten

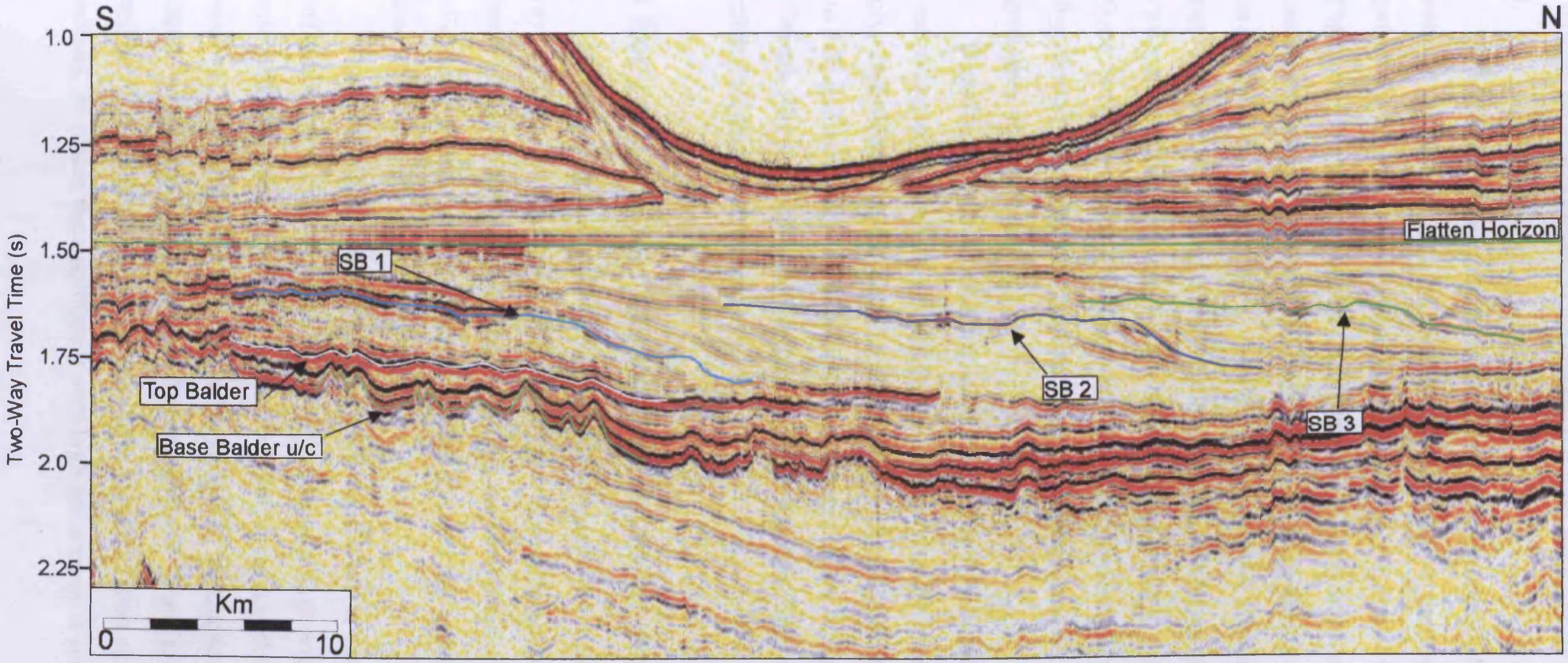


Figure 5.5. South - north 3-D seismic line that has been flattened on the green horizon shown (also see Figure 5.2). When flattening of the seismic is implemented, the original depositional slope can be seen. In this example, the high amplitude near horizontal reflections of SB 1 are seen to change dip towards the north and downlap onto the Top (T50) Balder reflection. Two other high amplitude reflections termed here SB 2 and SB 3 are seen to step out towards the north into the basin. Each of these three features show some degree of incision at their bases, and the significance of this will be discussed in the text. For location of the line see Figure 5.1.

the seismic on as it is eroded at the crest of the Judd Anticline by Top Eoc2 (see **Figure 5.2**).

When this flattening of the seismic data is implemented shallow dipping seismic reflections are visible which become close to horizontal in the south of the survey (**Figure 5.5**). These reflection geometries have been interpreted as a progradational package consisting of topsets and clinoforms. This stacking pattern is indicative of a deltaic system which included high amplitude topsets considered to be coals (seen in the in the south of survey). As a whole the clinoform package seems to have an average height in the region of 100 – 150 ms giving an approximate estimate of the water depth of 100 – 150 m at this time (assuming an average velocity of sediment of 2000 ms $^{-1}$) (**Figure 5.5**). There is a variation in the geometry of the reflection configuration packages throughout the Eocene 1 succession with different degrees of progradation and aggradation.

Detailed analysis of the stacking patterns of this generally progradational unit can aid in the understanding of the interactions between relative sea-level and sediment supply. Individual stacking patterns of each unit will be discussed later (see **Sections 5.5** to **5.8**) and as a whole in **Section 5.10**. By mapping the position of the depositional shelf-break (the point between the topset and the clinoform) the movement of the shoreline can also be examined.

5.4 Eocene 1 Seismic Unit Sub-Division Methodology

Numerous seismic reflections have been mapped throughout the three 3-D surveys and the geological evolution of the basin is constructed using structure, isochron and seismic amplitude maps to outline the local basin architecture at the time. The Eocene 1 seismic unit of **Chapter 4** is sub-divided into a number of higher order seismic units which is discussed herein.

The methodology of sub-dividing the Eocene 1 unit is based on the recognition of surfaces of stratal discontinuity which often appear as high amplitude continuous seismic reflections and show evidence for localised erosion of the underlying reflections. Throughout the three 3-D surveys used in the chapter, four such high amplitude reflections have been identified and mapped. These reflections have a limited areal extent and exhibit significant amplitude variation along the length of the reflections. In places the amplitude is very high and erosion is seen under small local incisions, however in other places (in particular towards the basin centre in the north), the seismic amplitude decreases and the reflection is often seen to thin and downlap distally to the north.

These four intra-Eocene 1 seismic reflections have been mapped across one or more of the surveys and are classified as candidate sequence boundaries (CSB's) (*sensu* Vail *et al.* (1977), Mitchum and Vail (1977) and Vail and Todd (1981)). These CSB's bound the sub-units of the Eocene 1 seismic unit which are termed the 1A, 1B, 1C and 1D seismic units respectively. These CSB's show evidence of erosion and are interpreted to represent local unconformities and their correlative conformities (*sensu* Vail *et al.* 1977, Mitchum and Vail 1977) which cuts out and erodes areas of the underlying unit. In the context of the localised nature of the study area these candidate sequence boundaries are thus of regional extent and are thus termed sequence boundaries (SB's) in context of the South Judd Basin. This does not imply that they have a basin-wide regional extent. Detailed maps of these erosional surfaces (SB's) have been produced in order to understand the planform geometries of these surfaces in order to interpret their palaeogeographic setting. The four SB's are called SB 1, 2, 3 and 4 with SB 1 being the oldest and 4 the youngest. **Table 5.1** summarises the upper and lower boundaries and the names of the individual sub-units discussed here.

The following sections (**Sections 5.5 – 5.8**) will discuss the age and examine the internal reflection architecture of the four sub-units (1A - 1D) and detail the bounding surfaces (SB 1 - SB 4).

5.5 Seismic Unit 1A

5.5.1 Introduction and Age

The upper bounding reflection of seismic unit 1A is termed SB 1. The basal bounding surface to this unit is the top T50 (Balder Tuff) reflection. The top T50 (Balder Tuff) event has been dated at approximately 54.9 Ma. As has been previously discussed in detail above (**Section 5.3.2**) there is difficulty accurately dating the Eocene 1 succession to anything more accurate than Ypresian (Early Eocene). It is fair to assume that because seismic unit 1A is conformable to the top T50 (Balder Tuff)

Bounding Reflection	Name of Eocene 1 sub-unit
SB 4	
	Seismic Unit Eocene 1D
SB 3	
	Seismic Unit Eocene 1C
SB 2	
	Seismic Unit Eocene 1B
SB 1	
	Seismic Unit Eocene 1A
Top T50 unit (Balder Fm equiv)	T50 unit of Ebdon *et al.* (1995)

Table 5.1. Table summarising the upper and lower bounding seismic reflections of the sub-divisions in the Eocene 1 seismic unit. In the south of the Faroe-Shetland Basin it has been possible to sub-divide this unit into four sub-units (1A - 1D) based on the recognition of high amplitude continuous seismic reflections that are seen on three 3-D surveys. The upper and lower bounding surfaces are termed sequence boundaries (SB's) sensu Vail *et al.* 1977 and show aerially limited areas of incision into the underlying seismic reflections.

reflection, the age of the unit is Early Eocene and comprises of Lower Ypresian interbedded sandstones and mudstones.

5.5.2 Seismic Reflection Interpretations and Stacking Patterns

In the very southern part of the reprocessed survey in an area close to the border between the southern parts of Quadrant 204 and 6005 (**Figure 5.1**), unit 1A consists of very high amplitude seismic reflections that appear conformable and parallel to the top (T50) Balder reflection. However, these high amplitude reflections become convergent with the top T50 (Balder) reflection and eventually downlap onto it towards the north (**Figure 5.6**) thus defining a low relief progradational system. The reflection has a low seismic amplitude in this distal position. Internal seismic reflections of the unit 1A are continuous to semi-continuous and are sub-parallel to the top T50 (Balder Tuff) reflection. Some of the dipping reflections show a high amplitude continuous seismic character and the most distal of these dipping reflections and its coincident horizontal part has been mapped extensively across the 3-D survey and forms the top of the unit.

The top of the unit is the SB 1 reflection which in places is very high amplitude, in particular on the horizontal, parallel to sub-parallel part of the reflection. At the base of the high amplitude sub-parallel part there may be some evidence of erosional truncation. It is here where the reflection shows some relief on its surface and rises structurally to form a small local mounded feature (500 – 100 m in width) close to the position where it downlaps onto the top T50 (Balder Tuff) reflection (**Figures 5.6** and **5.7**). North of this small mounded feature the SB 1 reflection thins by true or apparent downlap onto the Top T50 (Balder Tuff) reflection (**Figure 5.5**). To the south of the sub-parallel reflections the seismic character becomes lower frequency, more continuous and higher amplitude (**Figure 5.6**).

These convergent dipping reflections are of significantly lower amplitude than the parallel (horizontal) reflections seen towards the south and are interpreted as low relief foresets, representing clinoforms that downlap onto the top T50 (Balder Tuff) reflection as bottomsets. The more southerly, horizontal very high amplitude reflections are interpreted as topsets and may represent coal horizons on the top of a delta plain. If they do indeed represent coals then they have limited lateral extent as shown by the high amplitude areas in the south (**Figure 5.8**). On a local (1 - 2 km) scale the clinoforms have a complex sigmoid-oblique or tangential oblique

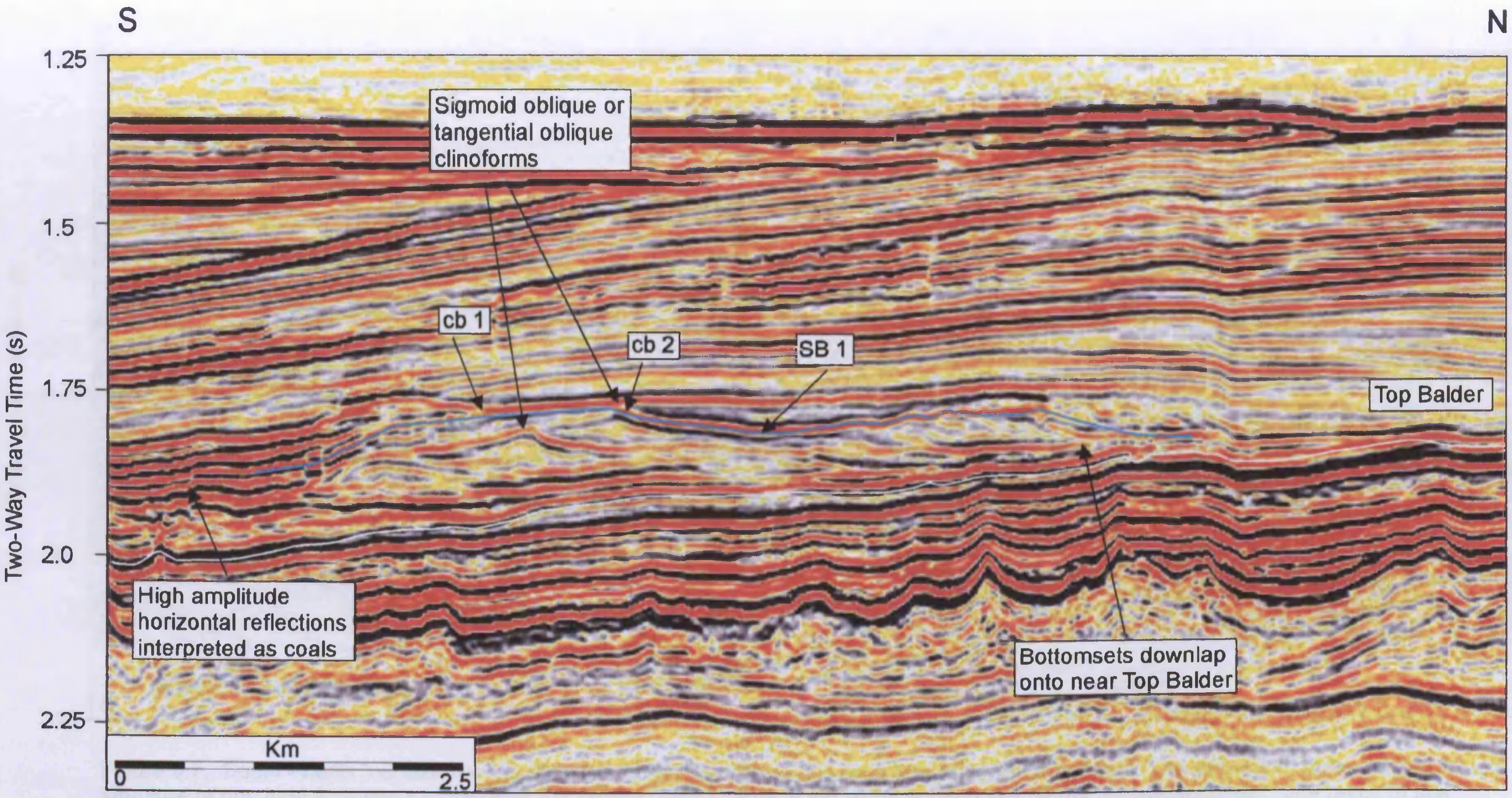


Figure 5.6. 3-D seismic line running roughly N-S showing early progradation of clinoforms of unit 1A. The patterns of the clinoforms are oblique-tangential or complex sigmoid-oblique (cb 1 and cb 2 indicate positions of two successive clinoform break points). This clinoform geometry may suggest a static sea-level, no subsidence or high sediment supply. South of the clinoforms are higher amplitude reflections which are interpreted as topsets and may represent coals. The most basin-ward clinoform is interpreted as SB 1 (blue) and shows erosional truncation at its base and downlaps towards the north. These reflections are near horizontal when the folding of the Judd Anticline is removed. The clinoform system is then seen to be covered and transgressed by the overlying unit as recognised by onlap. For line location see Figure 5.8.

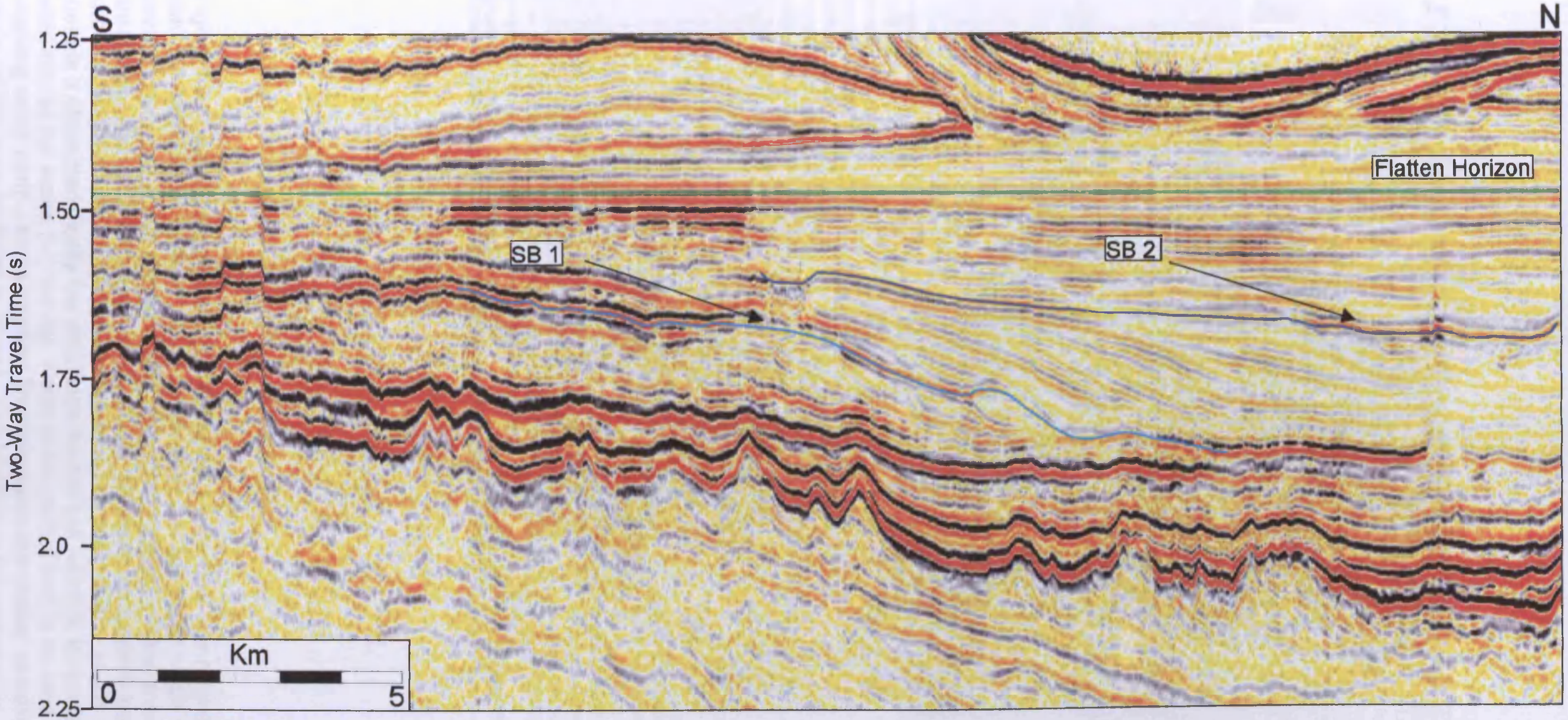


Figure 5.7. South - north 3-D seismic line showing the affect of flattening on a pre-determined horizon. This technique takes out the folded structure of the Judd Anticline and restores the seismic data to what it would have been like before folding occurred. The result of this flattening is better visibility of the dipping reflections which are interpreted here as clinoforms and the more horizontal higher amplitude reflections to the south which are interpreted as topsets. Some of the reflections show signs of incision of underlying strata and these have been termed sequence boundaries (SB's) and are mapped in great detail. The location of this seismic line is shown in Figure 5.8.

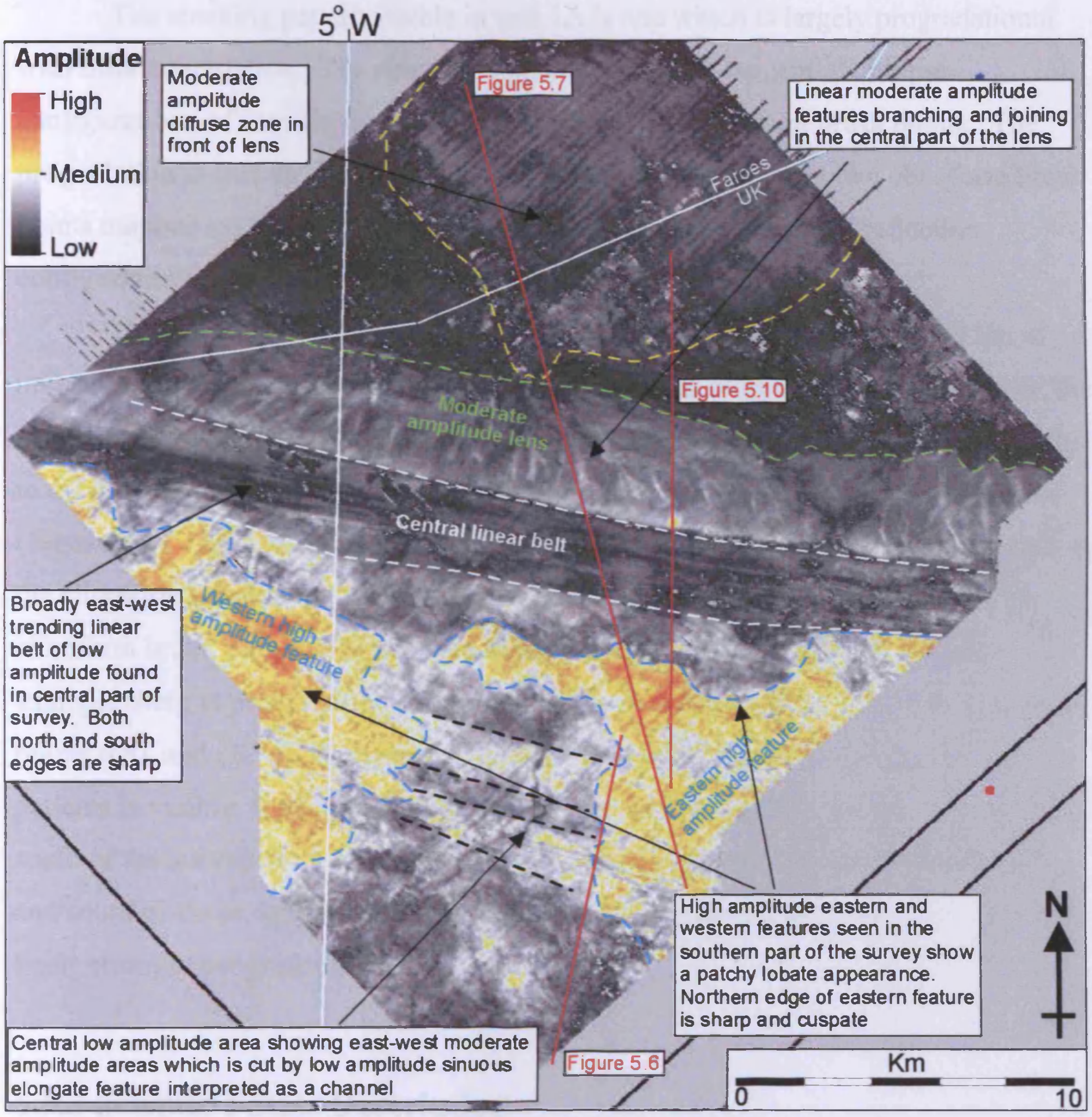


Figure 5.8. Amplitude extraction map of the SB 1 reflection. The map details a complex laterally variable depositional system. In the south, there are two areas which show a high amplitude seismic response which are separated by low to moderate central zone. The high amplitude areas have sharp cuspate northern edges (in particular the eastern feature). The central zone show a thin (200 - 500 m) channel which trends broadly north - south. North of these features, there is a characteristically low amplitude central belt which is featureless and trends east - west across the entire 3-D survey. North of the central belt, is an area of moderate amplitude showing small north - south trending linear features interpreted as being channelised and feed into a central area (outlined in green). The location of the seismic line shown in Figures 5.6, 5.7 and 5.10 are highlighted.

configuration where little or no resolvable aggradation at the topset is seen. A strongly progradational stacking pattern is visible within this unit.

The stacking pattern visible in unit 1A is one which is largely progradational with little aggradation. The complex sigmoid-oblique or tangential oblique configurations of the clinoform reflections have sharp clinoform break points. The progradation is further emphasised by the trajectory seen between two clinoform break points mapped on SB 1 which show a progressive down-stepping of reflection configurations (**Figure 5.9** and **5.6**).

The SB 1 reflection represents an oblique clinoform that occurs at the top of unit 1A and shows some incision on the foreset portion of the clinoform. However, the mapped SB 1 reflection represents the time of maximum amount of progradation prior to onlap and is thus interpreted as a composite surface which follows two clinoforms (**Figure 5.9**). SB 1 is mapped on a high amplitude horizontal reflection which bends at the clinoform break point (cb1) and merges into another horizontal reflection to its clinoform break point (cb2) and then down the foreset to the bottomset. In this example there is progradation of approximately 2 km between the two clinoform break points (cb1 and cb2 on **Figures 5.6** and **5.9**). Little or no aggradation of stacking patterns is visible, though this may be due to the limit of the 3-D survey. In the very south of the survey the high amplitude reflections are continuous and of low frequency and south of these, aggradation may be found. Unit 1A as a whole can be described as being strongly progradational.

5.5.3 Seismic Facies Description

An amplitude extraction map of the SB 1 reflection shows an extremely variable pattern of seismic amplitude response which reveals a complex drainage pattern (**Figure 5.8**). This drainage represents a short period of time near to the end of deposition of seismic unit 1A. A discussion of this seismic amplitude map and its importance when determining the depositional environments will follow in the next section (**Sections 5.5.3.1** and **5.5.3.2**). The amplitude map will be discussed geographically starting with the southern part and followed by the north.

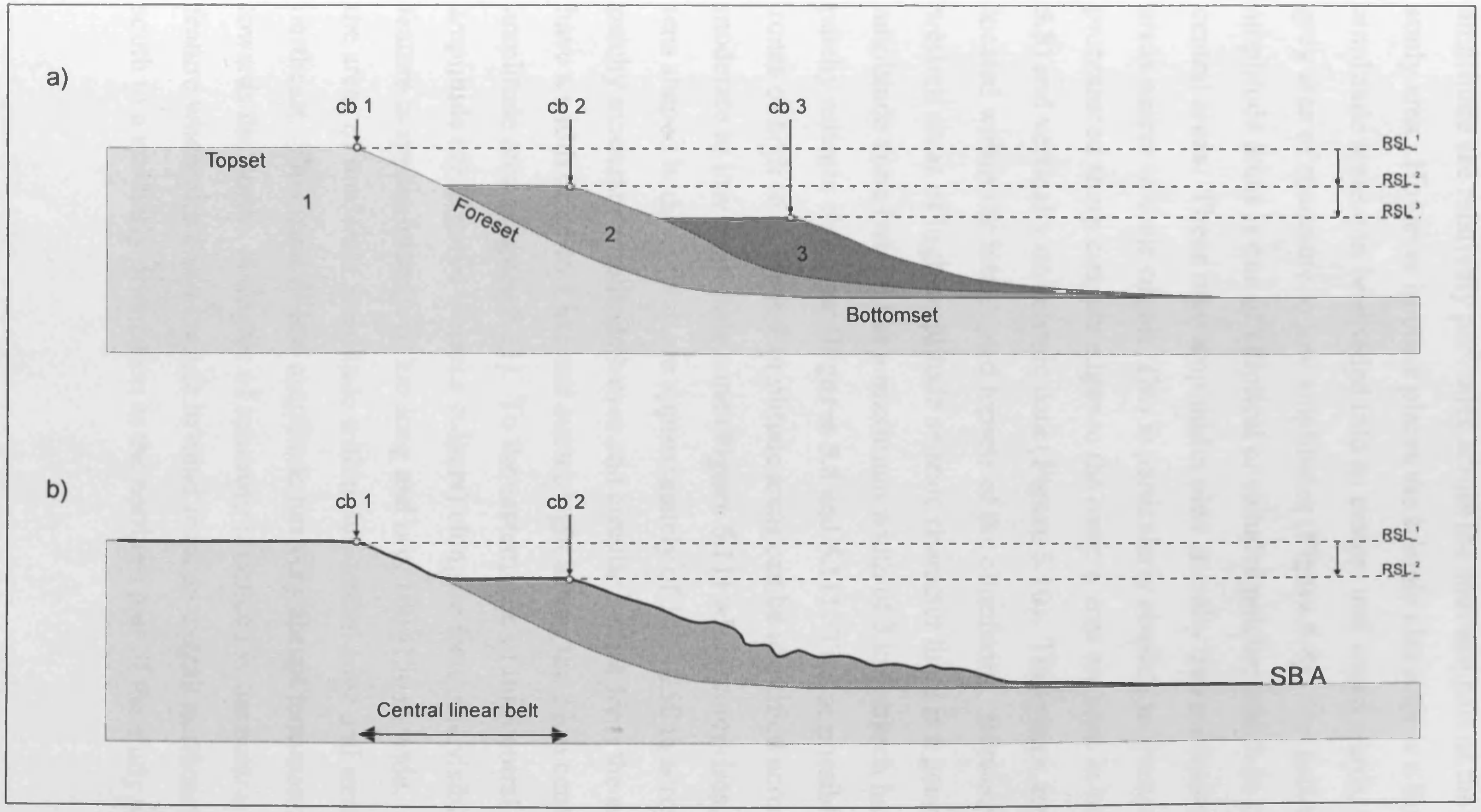


Figure 5.9. Schematic cartoon showing formation of progradational stacking pattern during a relative sea-level fall. a) shows the development of unit 1 with associated topset, foreset and bottomset of the delta system which was deposited at RSL1. Relative sea-level falls to RSL2 and exposes the clinoform break point (cb 1) of unit 1. Sedimentation of unit 2 commences further into the basin developing with the topset and clinoform break point (cb 2) at a lower stratigraphic level than the previous unit. A decrease in accommodation space occurs during this fall in relative sea-level. The same process occurs between RSL2 and RSL3. b) shows the mapped sequence boundary A (SB A) which is a merged surface representing time of maximum progradation into the basin (see Figure 5.6). The foreset of the clinoform package shows evidence of relief and erosion. The two clinoform break points (cb 1 and cb 2) are seen on the amplitude extraction map (Figure 5.8) and represent two successive shorelines which developed during relative sea-level fall.

5.5.3.1 Observations from the Southern Area

In the southern half of the reprocessed survey the seismic amplitude map shows a significant areal dominance of high amplitudes (warm colours). These areas of high amplitude are relatively pervasive across the southern part of the South Judd Basin study area. However in other places the seismic character is a little patchy and the high amplitude areas can be divided into an eastern and western area, which is divided by a grey area of moderate to low amplitudes **(Figure 5.8)**. The pattern of these high amplitude areas is one of elliptical or rounded patches which have high amplitude central areas. These high amplitudes pass laterally into moderate and low amplitude areas nearer to their edges. This is particularly obvious in the north where there are pronounced sharp cuspate edges to the eastern area are seen in both plan view **(Figure 5.8)** and vertically on seismic data **(Figure 5.10)**. These high amplitude areas are located within the interpreted topsets of the clinoforms. Between the eastern and western areas of high amplitude seismic character there is a generally moderate to low amplitude zone (which has a maximum width of 3 km) which has a very discontinuous patchy seismic character **(Figures 5.8** and **5.11)**. Two or possibly three east - west zones of high to moderate amplitude areas can be identified across this central moderate to low amplitude zone **(Figure 5.11)** which comprises of broadly circular or lens shaped bodies which are approximately of 100 - 250 m across. Together these patchy moderate amplitude lenses and circular bodies form the east - west zones which have a width of up to 1 km and stretch right across the 3 km central moderate to low amplitude area **(Figure 5.11)**. To the eastern side of this central zone an extremely low amplitude (dark greys – blacks colours) elongate feature is visible **(Figure 5.11)**. This feature is approximately 10 km long and is c. 100 -150 m wide. It is seen to cut across the areas of moderate amplitude within this central zone and trends roughly north-northeast. This zone of low amplitude has very abrupt terminations to it and widens towards the north. A degree of sinuosity is noticed in the central part of the 10 km feature where there is a change in trend from an overall northeasterly orientation in the south to a northerly orientation in the northern part of the study area **(Figure 5.11)**.

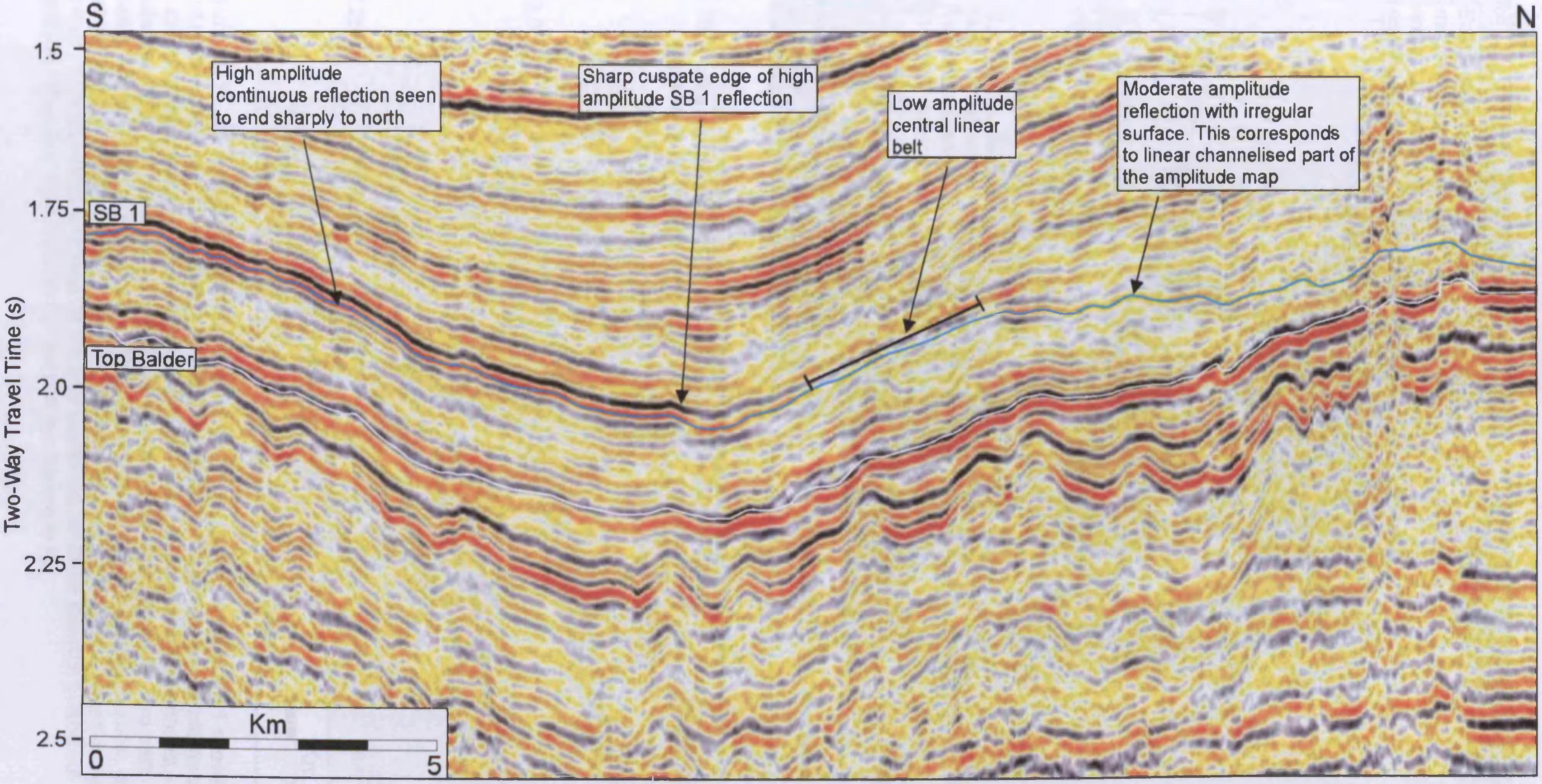
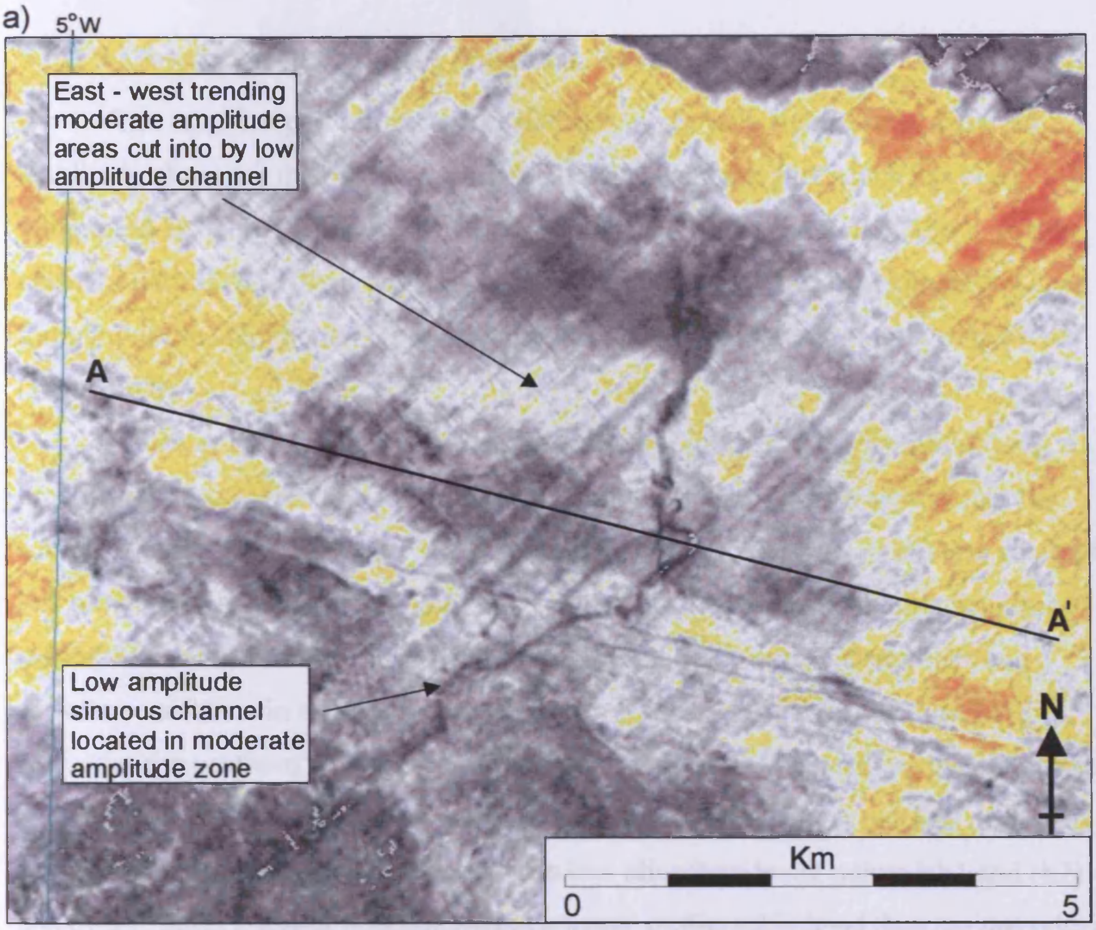


Figure 5.10. South - north seismic line showing edge of the high amplitude continuous SB 1 reflection. The change in seismic character is apparent along the reflection, from a high amplitude continuous character in the south to the more moderate and low amplitudes in the north where the reflection eventually downlaps onto the Top (T50) Balder reflection. In the centre is the low amplitude central belt which is a distinctive feature on the amplitude map (Figure 5.8). For location of seismic line see Figure 5.8.

a)


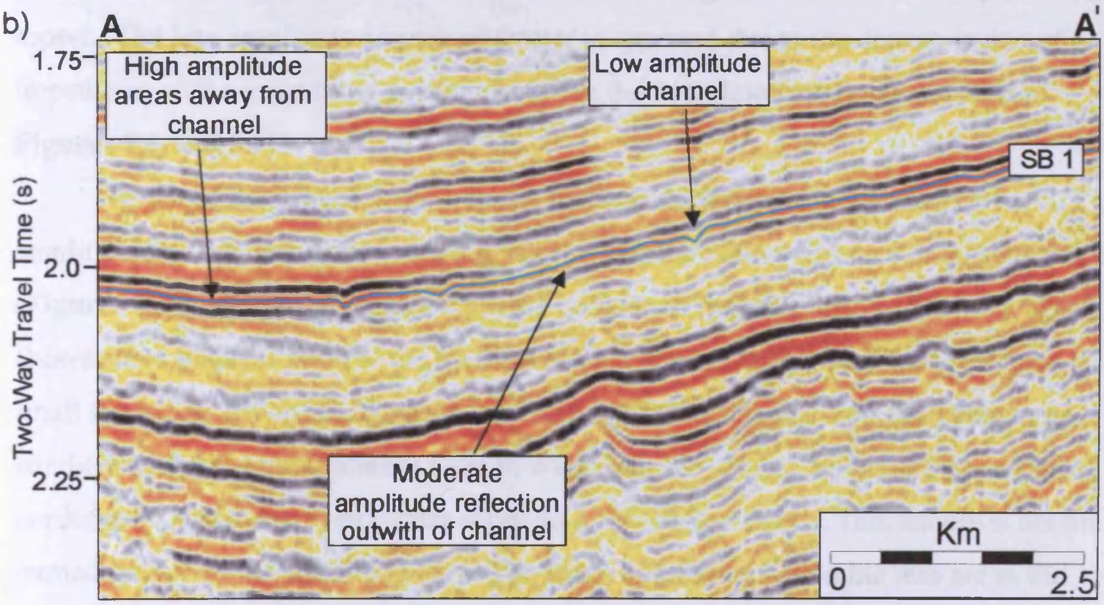
b)


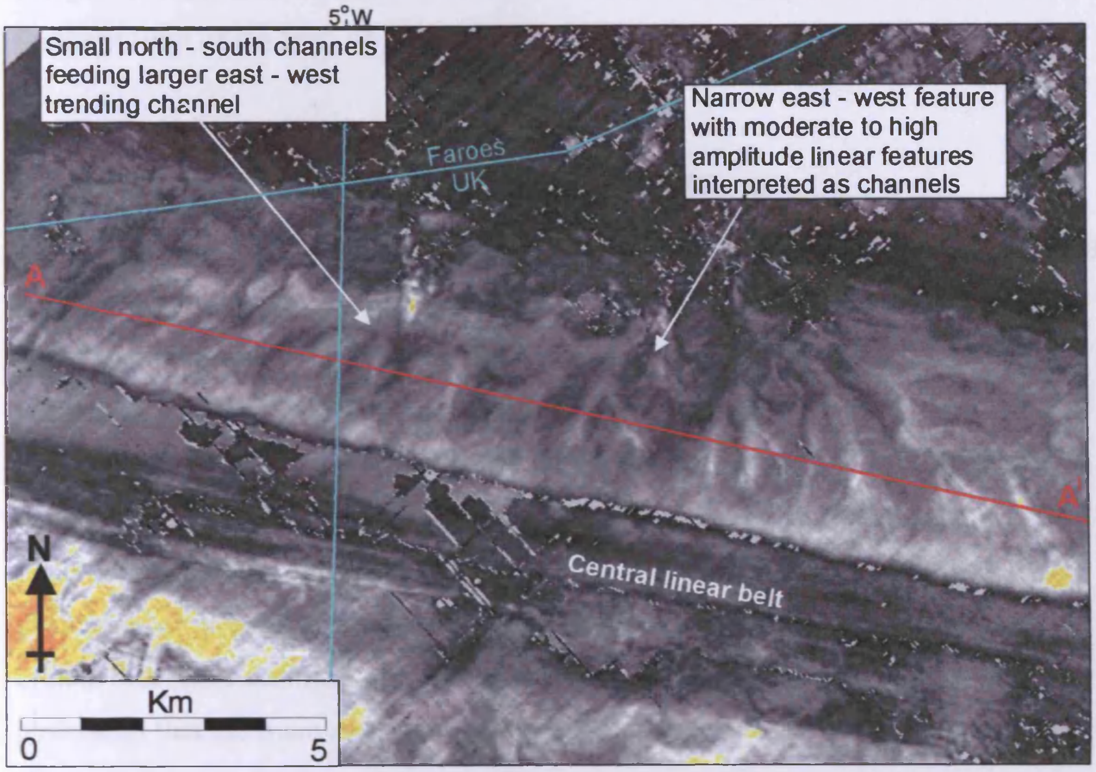
Figure 5.11.a). Amplitude extraction map (reds =high, greys =lows) of SB 1 showing a close up of the southern part of the survey showing the central moderate to low amplitude zone with the low amplitude channel. The channel is seen to change course from a northeast - southwest trend to a north - south trend, and shows some slight sinuosity. Two east - west trending moderate amplitude belts are seen across the central low amplitude zone. b). east - west seismic line across the low amplitude channel and higher amplitude zones away from the channel. This seismic line highlights the very subtle nature of the channel picked out in the amplitude extraction map. Location of seismic line shown in a).

5.5.3.2 Observations from the Northern Area

The northern part of the reprocessed survey shows a significantly different amplitude character than the previously mentioned southern part which has been interpreted as being dominated by high amplitude coals (see **Section 5.5.3.1**). In this northern part of the 3-D survey the general amplitude character is predominantly low throughout the area though it occasionally increases to moderate amplitudes in places (**Figure 5.8**).

North of the high amplitude abrupt cuspate edge of the high amplitude area, a broadly east - west trending low amplitude 1 km wide zone appears right across the survey (**Figure 5.8**). There are a few isolated patches of moderate amplitudes south of this broad parallel zone or belt, but these are not attached to the main high amplitude eastern and western areas. This 1 km wide low amplitude zone is featureless and shows no variation in seismic character; however it has abrupt terminations, in particular the northern edge which forms a very noticeable lineament on the amplitude map (**Figure 5.12**). This feature has been termed the central linear belt and represents part of the reflection which occurs between two clinoform break points (cb1 and cb2) and thus follows a topset which occurs at a lower stratigraphic level than the preceding topset. The low amplitude seismic character is apparent due to the change in acoustic impedance of the underlying package between the two clinoform break points (**see Figures 5.6** and **5.9**).

North of the central linear belt is a feature which shows moderate to low amplitude seismic character revealing an east – west trending elliptical or lensoid shape (**Figures 5.8** and **5.12**). It is located in the position of the clinoform of the SB 1 which shows some positive relief at the location of the elliptical body. This feature displays small (less than 150 m) moderate to high amplitude lineaments. This moderate to low amplitude lens is approximately 2 km in width and 18 – 20 km across and has a curved northern edge and sharp, straighter southern limit (**Figure 5.12**). This feature is herein termed the moderate amplitude lens. The small lineaments within the lens are in the region of 1 – 2 km long and trend broadly north - south, though this varies laterally along the length of the feature where they are seen to trend towards a central part of the lens which is highlighted in **Figure 5.12**. Indeed there is evidence to suggest that small north - south trending lineaments feed into larger and broader east - west orientated features which trend towards the central part of the lens. The lineaments are seen to



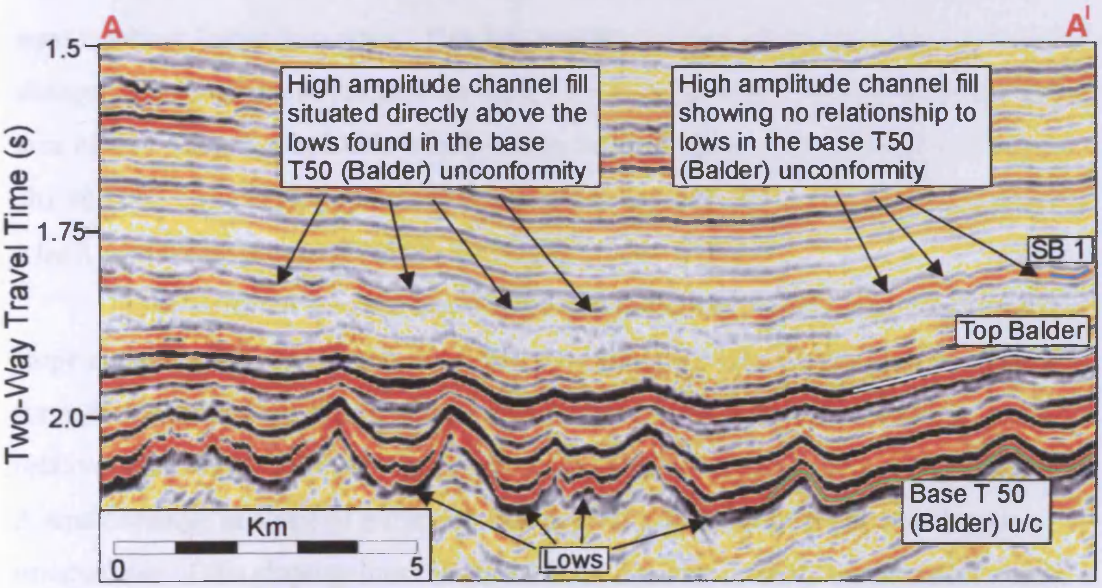


Figure 5.12. a). Close up of amplitude extraction map of SB 1 in central part of survey (see Figure 5.8) showing narrow elongate east - west area which has north - south trending moderate to high amplitude linear features. These linear features have been interpreted as channels. They are seen to bifurcate and lead into a central area by means of linking into larger channels that trend more east - west. b). Seismic line (location shown in a) through the lens shaped body showing the moderate to high amplitude channels. An important observation is that some of these channels lie directly over lows in the base T50 (Balder) unconformity and are thus being controlled by pre-existing topography, whilst others seem to have no relationship with the underlying structure.

merge into each other and become a wider more moderate amplitude area towards the northern limit of the lens.

5.5.4 Interpreted Depositional Environments

The dominantly high amplitude seismic character of the southern part of the SB 1 amplitude map (**Figure 5.8**) corresponds to the topset portion of the clinoform surface is interpreted as a delta top area. Coal deposits are interpreted to be found in the east and west areas which pass laterally into a central zone.

The low amplitude central area located between the coals is dominated by a small fluvial channel system which sits on the delta top and deposits terrigenous material towards the delta front. The high amplitude coals are interfluves to the main channel axis. It is suggested here that this small river channel is not the main river system that feeds sediment to the delta because it does not reach the delta front. It is postulated that the main river may occur outside the location of the 3-D seismic surveys.

The central linear belt is characterised by its strikingly low amplitude east – west trending featureless zone. This low amplitude seismic character is a result of a change in seismic facies underneath the SB 1 reflection which represents an isolated area of low seismic amplitude which occurs between the two clinoform break points cb1 and cb2. These two clinoform break points are thus inferred to be two shorelines which down-step progressively to the north.

In the northern area the linear moderate amplitude features are interpreted as a slope channel gullies or slope rills which feed from the delta top and acts as a conduit for sediment to the north into a basin centre. This slope part of the clinoform is relatively narrow (2 km) and may suggest very shallow water depths during this time. A small change in slope of a fraction of a degree may occur to create the changing orientations of the slope gullies. They are seen to curve towards the curved northern front of the slope.

North of the slope the seismic character returns to a featureless low amplitude area which covers a large part of the northern part of the survey. However a slightly higher amplitude area, which is very subtle, is seen in a position which is coincident with the down-dip terminations of the linear channels at the northern curved apex of the gullied slope (**Figure 5.12**).

The underlying control on these small discrete channel systems on the slope may be the pre-existing topography. The base T50 (Balder) unconformity (of Smallwood and Gill 2002) can be mapped in great detail on this reprocessed survey. A seismic line through the lens body which shows the slope gullies in many places sit directly above the lows in the base T50 (Balder) unconformity. It could be argued that the comparison between the high amplitude areas in SB 1 and the low areas of the base T50 (Balder) unconformity is striking (**Figure 5.12**), and suggests a fundamental control on the sediment dispersal patterns. Correspondingly, there are low amplitudes seen on the SB 1 reflection which are located directly above the highs in the base T50 (Balder) unconformity (**Figure 5.12**). However it can be seen that these slope gullies have their own order and are relatively evenly spaced across the slope and indeed do not sit directly over the base T50 (Balder) lows. It may therefore just be a coincidence that the two features are seen in the same position.

An isochron map between the top T50 (Balder Tuff) reflection and the SB 1 reflection shows the slope picked out by a slightly thicker area (**Figure 5.13**). To the east and south of the slope the familiar dendritic pattern cut during the development of the base T50 (Balder) unconformity (Smallwood and Gill 2002) is visible and suggests a submarine drainage pattern was prevalent during deposition of unit 1A which had superseded the sub-aerial pattern which occurred at the start of the T50 unit.

5.5.5 Relative Sea-Level Interpretation

Prior to deposition of the T50 unit the availability of accommodation space was low in early T40 times (Sele Fm and Flett Fm equivalents) during a period of progressive exhumation in the South Judd Basin to develop the semi-regional base T50 (Balder) unconformity. Seismic reflections progressively onlap onto this unconformity during renewed flooding at the time of T50 deposition. The seismic reflections immediately overlying the top of the T50 unit (of Ebdon *et al.* 1995) shows complex sigmoid-oblique or tangential oblique clinoforms which correlate southwards into very high amplitude near horizontal seismic reflections. This is interpreted to represent a continuation of marine transgression that was occurring throughout the deposition of the T50 unit (Balder Formation equivalent) (Ebdon *et al.* 1995, Smallwood and Gill 2002). Development of small clinoforms in the region of 50 – 100 m high are interpreted from the position of the clinoform break point and its height above positions

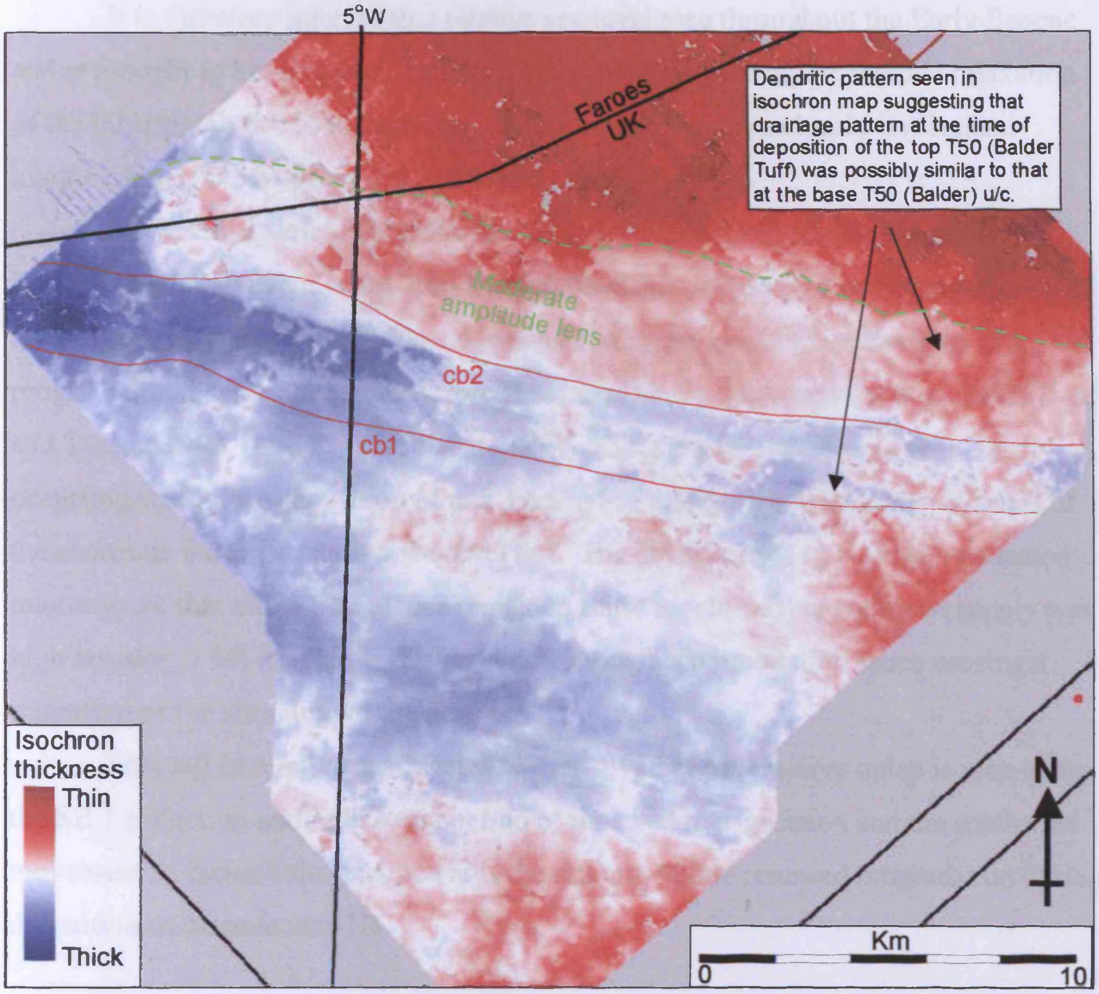


Figure 5.13. Isochron thickness map between the top T50 (Balder Tuff) reflection and the SB 1 reflection showing the familiar dendritic drainage pattern seen at the base T50 (Balder) stratigraphic level (Smallwood and Gill 2002 and Figure 2.13). The fact that there remains a dendritic drainage pattern picked out between these two reflections may suggest that at the end of the T50 unit (end of Balder Formation equivalent) similar drainage networks were present. However, it may just be an artefact of the high relief on the basal unconformity. The lows in the drainage network may have continued to influence deposition of the Early Eocene succession though this may also have been due to differential compaction over the eroded topography (see Figure 5.12). The positions of the two clinoform break points (cb1 and cb2) representing shorelines are also marked.

of bottomset downlap (on to the top T50 reflection) (see **Figure 5.6**). Hence an increase in water depths is inferred to have occurred into the period of deposition of unit 1A.

It is therefore inferred that relative sea-level rose throughout the Early Eocene and is thought to be in response to increased subsidence during the thermal relaxation of the lithosphere due to the north-westward drift of the anomalously hot Iceland mantle plume (see **Sections 2.4** and **4.4.3.1.3**).

However, it should be noted that the transgression and flooding over the top T50 (Balder Tuff) event is not long-lived as there is evidence of downward stepping of the prograding stacked units. This downward stepping of seismic reflections in a progradational system is termed a forced regression (e.g. Posamentier *et al.* 1992, Hunt and Tucker 1992, - see **Figures 5.4** and **5.9**). This evidence of a forced regression occurring within the deposition of unit 1A implies that there was a downward shift of the shoreline during a relative sea-level fall. The development of a forced regression may suggest that either subsidence remained static (or slowed) or sediment supply was high creating a fall in relative-sea level decreasing accommodation space causing a migration of the shoreline to the north.

This fall in relative sea-level was short-lived as progressive onlap is seen above the SB 1 reflection and indicates a period of renewed transgression and the southward movement of facies belts back towards the margin, before renewed progradation during deposition of seismic unit 1B.

5.5.6 Implications for South Judd Basin Evolution

From the observations and interpretations outlined above it can be seen that there was a complex interplay between the depositional style and the changes in relative sea-level during the deposition of unit 1A. An early period of continued flooding occurred over the base T50 (Balder) unconformity and the southern part of the South Judd Basin was at or close to sea-level during deposition of the top T50 (Balder) Tuff event. A small deltaic system prograded towards the north into shallow waters as subsidence waned and eventually relative sea-level fell, causing down-stepping progradational stacking patterns during a forced regression. Sediment supply may have been relatively high at this time with complex sigmoid-oblique or tangential oblique clinoforms developing (Mitchum *et al.* 1977, Mitchum and Vail 1977).

A palaeogeographic reconstruction at the end of unit 1A shows a delta top environment in the south of the study area consisting of coals and small fluvial channel systems that fed sediment to the delta front (**Figure 5.14** and **Table 5.2**). A narrow northerly dipping linear shoreline sits at the clinoform break point, two of which are interpreted on the seismic amplitude map seen in **Figure 5.8**. The two successive shorelines seen are approximately 2 km apart and represent and show the progressive northward movement of the shoreline through time during a forced regression. The narrow slope and the height of the clinoform systems indicate progradation into shallow waters close to the shoreline.

5.6 Seismic unit 1B

5.6.1 Introduction and Age

The second sub-unit consists of the strata found between SB 1 and SB 2. This unit is found to the north and thus more basinward than the earlier SB 1 unit (**Figure 5.11**). A crude age of Early – Mid Ypresian can be assigned to this unit from nearby boreholes which prove this interval is no later than Middle Ypresian in age (see **Section 5.3.2** and **Figures 5.3** and **5.4**).

5.6.2 Seismic Reflection Interpretations and Stacking Patterns

Immediately overlying the complex sigmoid-oblique to tangential oblique clinoforms of seismic unit 1A, onlap is seen onto a strong high amplitude continuous reflection that clips the top of the clinoform package and SB 1 (**Figure 5.15**). The onlap is progressive towards the south and becomes parallel with the high amplitude low frequency continuous reflection configurations of the 1A unit. The general seismic character of the 1B unit is spatially similar to the underlying unit in terms of high to moderate amplitude parallel to sub-parallel reflections in the south of the survey area which decrease in amplitude to the north and become more convergent with underlying reflections. However, where this unit differs is the higher frequency of the strongest amplitudes seen south of the 3-D surveys on 2-D seismic lines in the very south in the study area above the Judd High (**Figures 5.15** and **5.2**).

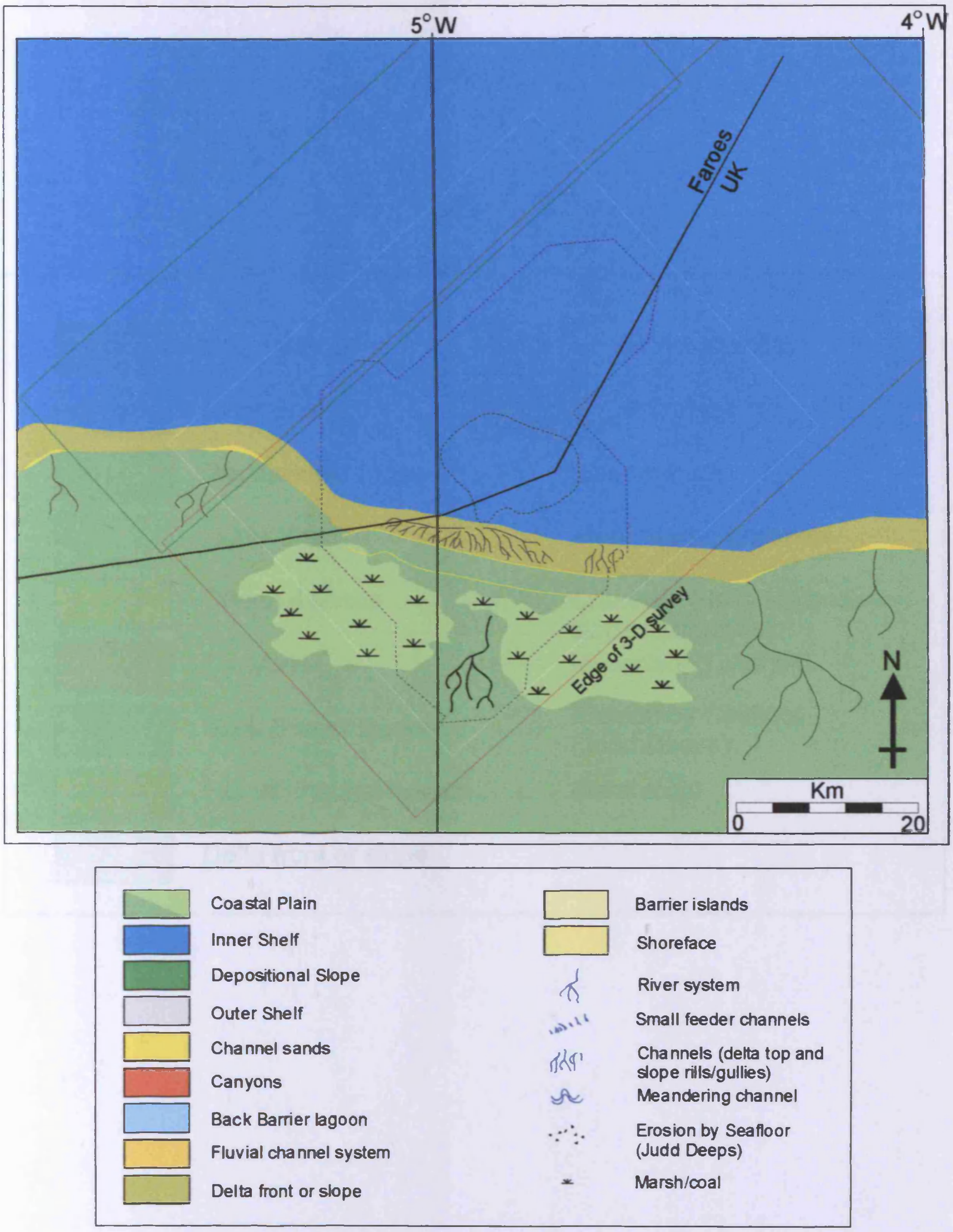


Figure 5.14. Schematic palaeogeographic map at the time of SB 1 deposition. In the southern part of the basin delta top conditions dominated and small river systems cut down into the top sets to feed north to the shoreline and slope. Coals are interpreted to have formed on the interfluves in the south due to the high seismic amplitudes and they had sharp northern fronts, possibly suggestive of collapse at the delta front close to the shoreline. Two successive shorelines are interpreted at the clinoform break points (cb1 and cb2 - yellow lines) from the amplitude extraction map (Figure 5.8) which are stranded due to a subtle fall in relative sea-level during a forced regression. A very narrow (2 - 4 km) northerly dipping slope is interpreted to sit in front of the delta top which shows small (200 - 500 m) wide slope gullies or rills down into the base of slope. These slope gullies are perpendicular to the linear clastic shorelines but focus to a central point and may suggest a subtle variation along the strike of the slope or possible slight axial control on channel systems. For legend of the colours used for the depositional environments see Table 5.2 or Enclosure P..

Coastal Plain	Barrier islands
Inner Shelf	Shoreface
Depositional Slope	River system
Outer Shelf	Small feeder channels
Channel sands	Channels (delta top and slope rills/gullies)
Canyons	Meandering channel
Back Barrier lagoon	Erosion by Seafloor (Judd Deeps)
Fluvial channel system	Marsh/coal
Delta front or slope	

Table 5.2. Table of contents showing colours used to describe the depositional environments and features seen in the four palaeogeographic maps for of the seismic units 1A - 1D (Figures 5.30 - 5.33 inclusive).

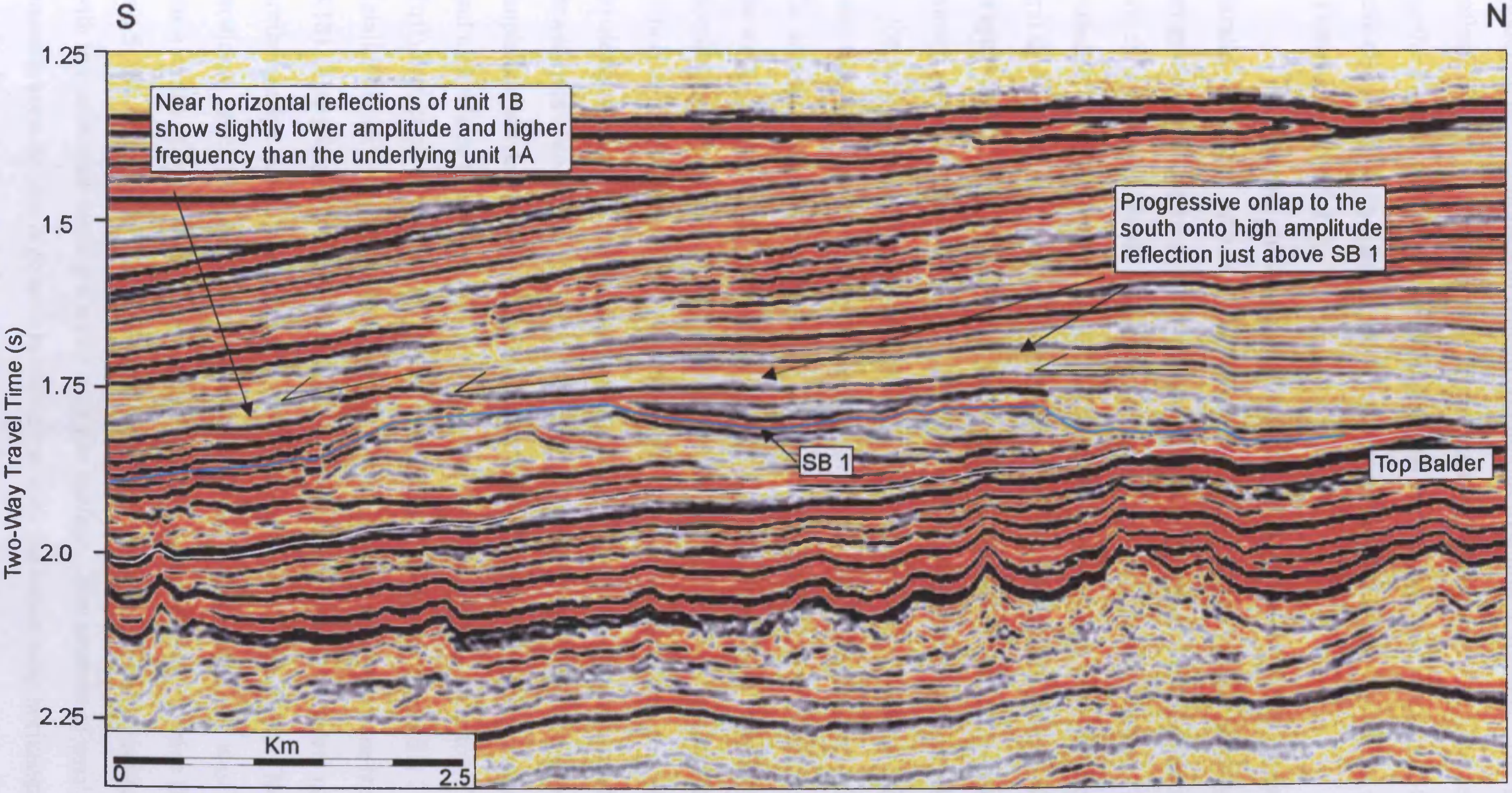


Figure 5.15. North - south trending 3-D seismic line showing the initial progressive southerly onlap onto the 1A unit. On the far southern end of the seismic line the reflections of the 1B unit are near horizontal (when flattened) and are moderate to high amplitude. They differ from the underlying higher amplitude and lower frequency of the near horizontal reflections of the 1A unit. The onlap occurred during a transgressive episode and this is discussed in more detail in the text (see section 5.6.5 and 5.6.6). For location of seismic line see Figure 5.8.

Towards the north of the 3-D survey area the reflections become low to moderate in amplitude and remain continuous to semi-continuous. The internal reflection configurations of unit 1B are seen to downlap onto the SB 1 reflection to the north. Further north they are similarly seen to downlap onto the Top T50 (Balder Tuff) reflection, indeed the entire unit 1B sits more basinward than the earlier unit 1A (**Figures 5.7** and **5.16**).

The internal architecture of unit 1B shows reflection configurations to be sub-parallel with both internal onlap and toplap apparent locally, in a broadly progradational stacking pattern (**Figure 5.16**). The top of the unit is defined by SB 2 which is distinguishable from its continuous high amplitude reflection. The SB 2 reflection can be mapped as a high amplitude (positive) reflection (warm red colours) and shows erosional truncation of underlying lower amplitude reflections at its base (**Figure 5.17**). The highest amplitudes of SB 2 are distributed in the central zone of the survey in blocks 6004/11, 12, 13, 16 and 17 in the Faroese sector and will be discussed in the following section (**Section 5.6.3**). Upon closer examination of SB 2, it can be seen that the high amplitude continuous reflection is very much a local feature. When the seismic data is flattened on the continuous high amplitude seismic reflection (green horizon in **Figure 5.17**) which is located just below the eroded Top Eoc1 horizon (see **Sections 5.3.3** and **5.5.3**), the high amplitude part of SB 2 can be seen to be at or close to horizontal (**Figure 5.17**). Under the high amplitude horizontal part of the reflection erosional truncation is seen which cuts out dipping reflection configurations. Internal downlap is seen below the horizontal part of the SB 2 reflection (**Figure 5.17**). The amplitude value of the SB 2 reflection decreases away from this central erosive part and also where the reflection converges and downlaps towards the top T50 (Balder Tuff) reflection to the north. When examining this high amplitude SB 2 reflection in detail a broad concave upwards geometry with a near horizontal base is seen (**Figure 5.18**). The edges of this structure show a higher relief than the central part and the northern edge has a height that is a maximum of 80 – 100 m higher. The positive erosional relief on the northern edge trends broadly east northeast - west southwest and has a width of approximately 4 km and is thus a sort of barrier or ridge (**Figures 5.17** and **5.18**). The SB 2 reflection on this northern edge also shows a highly erosive nature with the reflection having a wavy irregular surface. The amount of small narrow incision seen on this high area is consistent with the broad near horizontal (higher amplitude) area to south thus suggesting that this incision was coeval. This observation

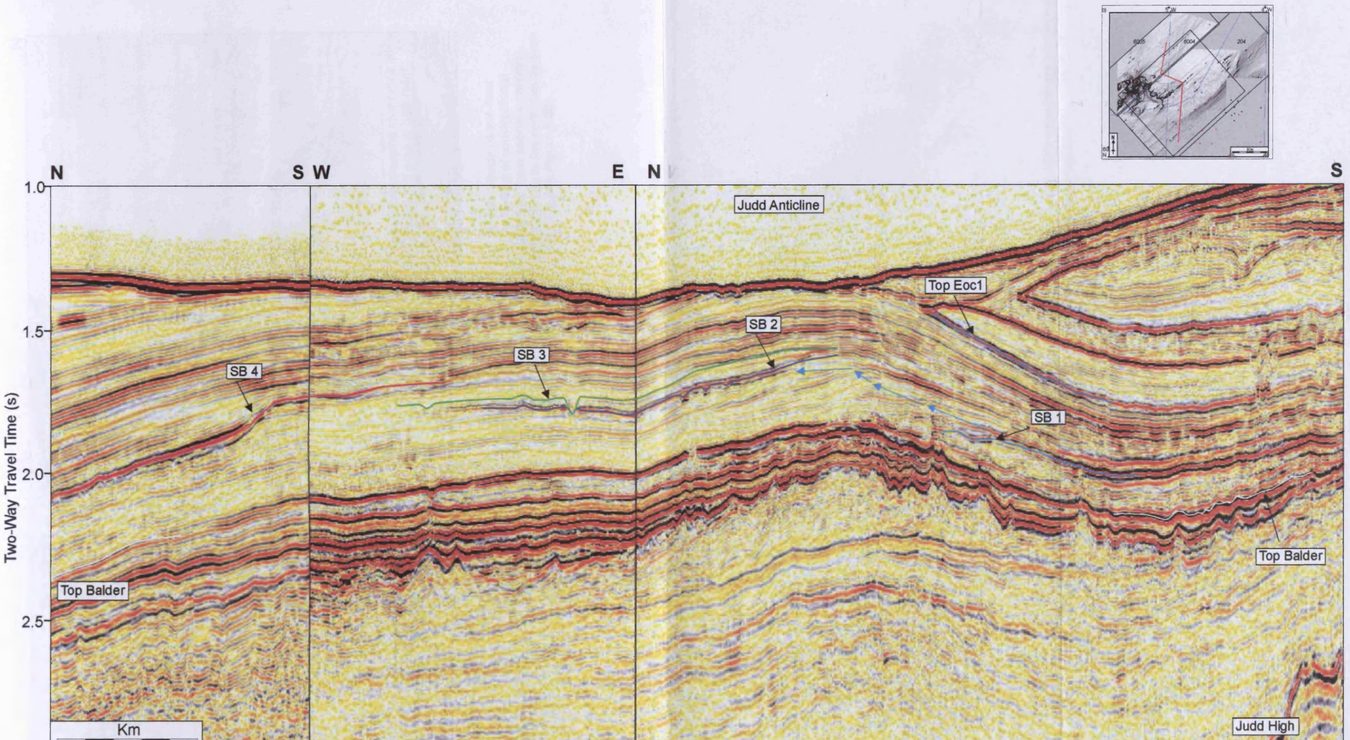


Figure 5.16. Composite seismic line from three 3-D surveys showing the generalised locations of the four seismic units (1A - 1D) discussed in the text. These seismic units are separated by incisional surfaces termed sequence boundaries (sequence boundaries) step out further to the north and west (into the basin centre) through time. The pale blue arrows indicate the progradation between unit 1A and 1B. The sequence boundaries are seen to be picked out easily by their high amplitude seismic response in an otherwise low to moderate amplitude seismic unit. It additionally shows the different morphology of each of the sequence boundaries and highlights that they are variable in form. Note that SB 2 shows numerous small incisions, whilst SB 4 has a smooth larger cut that has no evidence of small incisions.

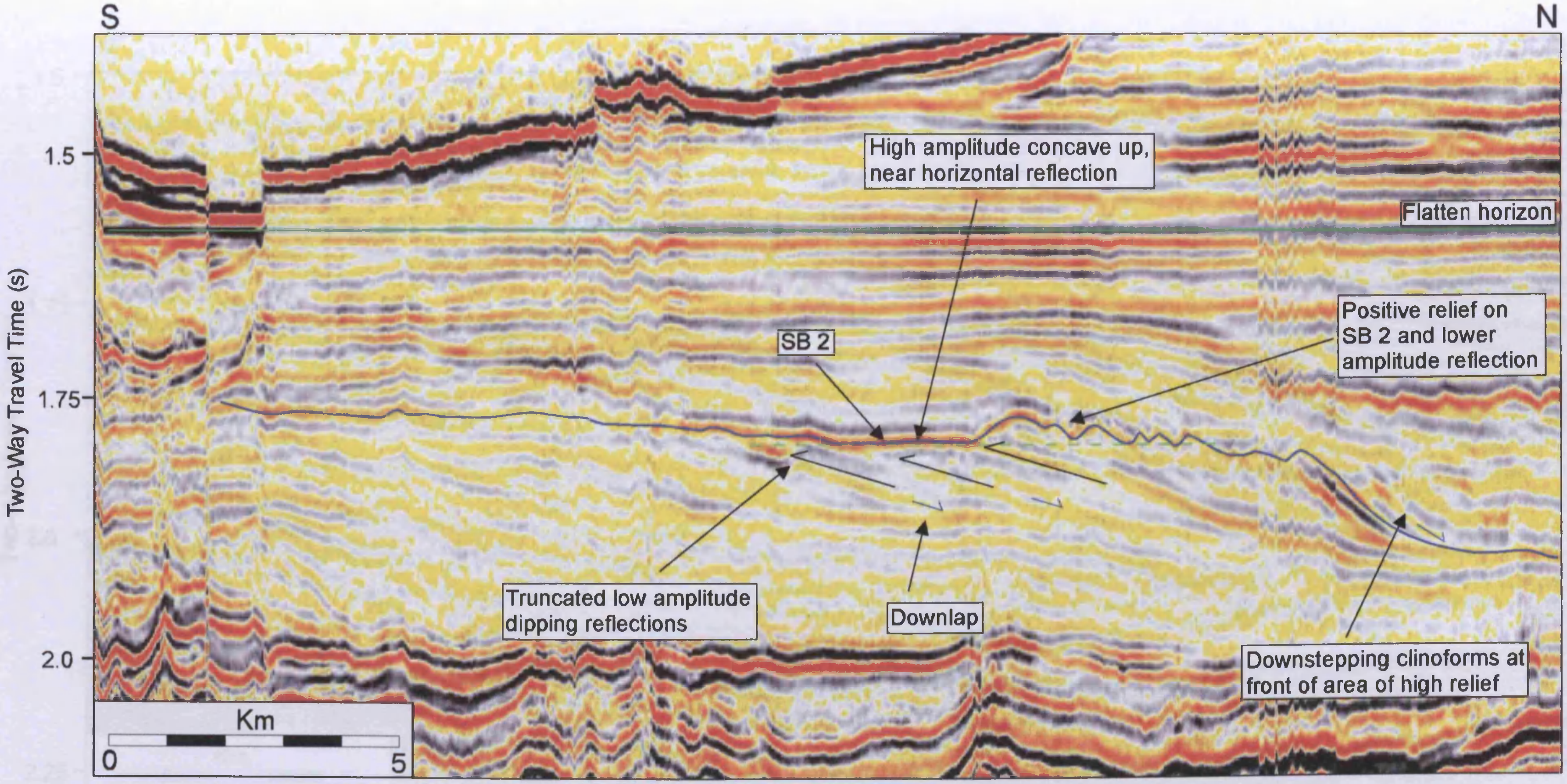


Figure 5.17. South - north trending flattened 3-D seismic line over the central part of SB 2. Towards the south, the reflections are near horizontal and they are seen to dip gently towards the north. Note the local positive relief of SB 2 which shows some form of constructional topography. Erosion of underlying reflections can be clearly seen underneath the high amplitude reflection of SB 2 and there seems a common base level (green dashed line) of erosion both of the flat part of the reflection and on the wavy higher relief part. Downlap to the north can be seen of the 1B unit and in front of the high relief area down-stepping clinoforms are interpreted. For location of seismic line see Figure 5.20.

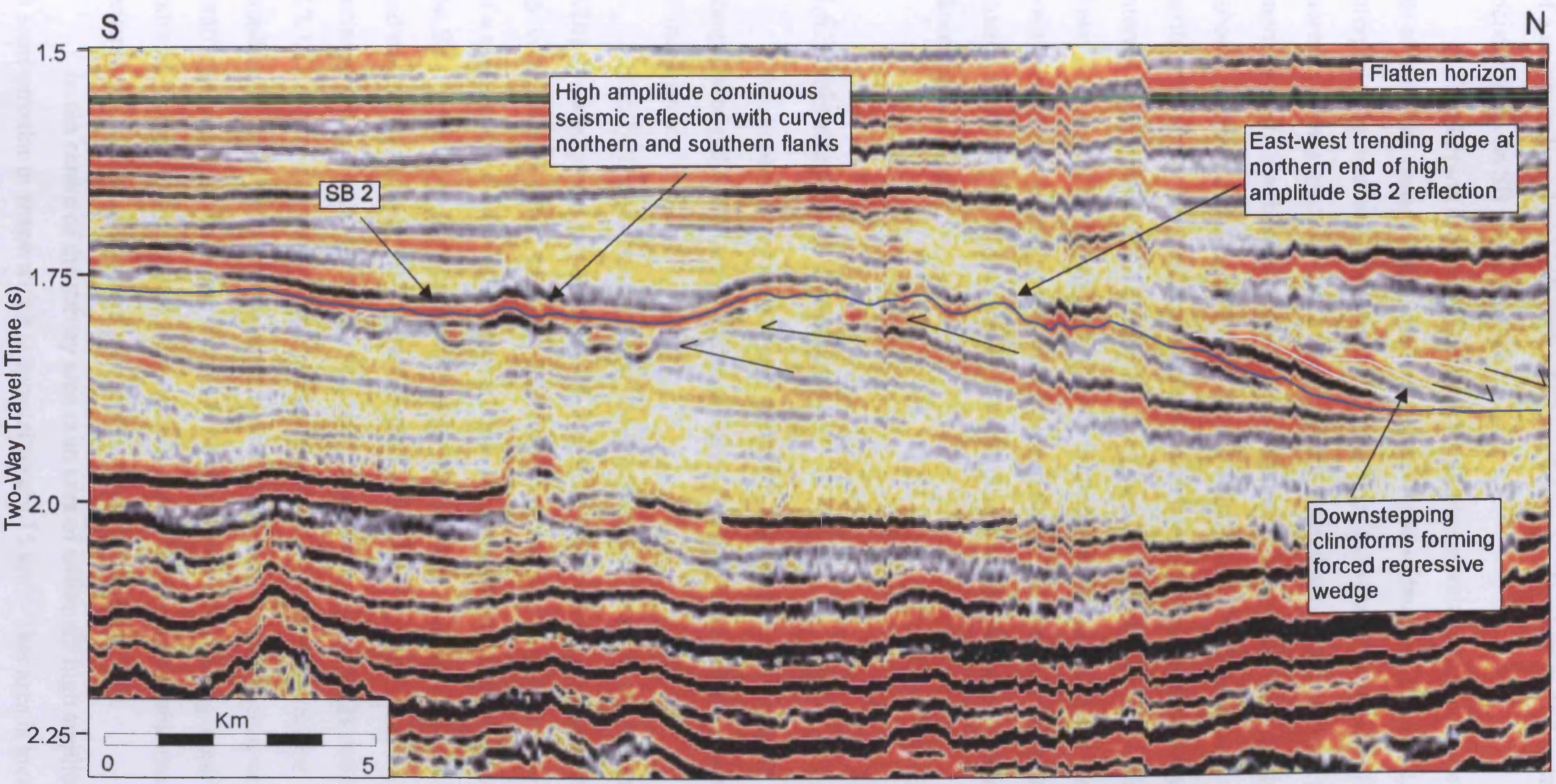


Figure 5.18. Broadly north - south trending flattened 3-D seismic line across the high amplitude upper bounding surface of unit 1B. This high amplitude reflection (SB 2) shows a relatively flat, near horizontal central part and has curved edges noticeably in this example on the northern side. This northern edge shows a high positive relief which trends in an east - west arcuate ridge approximately 8 - 10 km long. A down-stepping pattern of reflections is seen at the front of the ridge in the area of maximum progradation of unit 1C. For location of seismic line see Figure 5.20.

of a common base level indicates a fall in relative sea-level up to 100 m caused the incision seen on SB 2 (see **Section 5.6.5**).

The dipping seismic reflection configurations which are seen to be truncated in the area of this concave high amplitude erosive part of the SB 2 reflection, are interpreted to represent clinoforms that have been truncated. These clinoforms have a more classical sigmoidal shape than the more tangential or oblique forms seen in the previous unit. The bottomsets of the clinoforms are seen to downlap to the north and become parallel to sub-parallel with the top T50 (Balder Tuff) reflection. North of the northern edge the SB 2 reflection dips towards the north and itself downlaps onto internal reflections of the 1B unit. North of the area of erosion, progradation and down-stepping of clinoforms and topsets is interpreted at the northerly limit of the positive relief erosive edge (**Figures 5.18** and **5.19**). Therefore outside the area of erosion it is possible to see clinoform break points which can be interpreted as shoreline positions.

5.6.3 Seismic Facies Description

An amplitude extraction map of the SB 2 reflection (**Figure 5.20**) reveals a complicated pattern of varying seismic facies located within central and northern part of the survey.

Firstly, it must be stated here that there are features seen on this map (**Figure 5.20**) that are caused by the effects of overlying and underlying seismic reflections both above and below SB 2. In this example the observed map is therefore a merged imaged of a number of seismic reflections and does not truly represent the amplitude values of the SB 2 reflection. This is most strikingly seen where the SB 2 reflection is truncated and cut out by the later SB 3 reflection (see **Section 5.7**) giving rise to a spectacular series of north – south trending low amplitude canyons (**Figure 5.20** and **Section 5.7.3.1**). The truncation by the canyons of an east - west trending high to moderate amplitude channel planform is evidence alone to infer that the canyons sit at a higher stratigraphic level. Therefore care must be taken when interpreting amplitude extraction maps to filter out what is actually real. The distinguishable (real) features of seismic facies seen in this map will now be discussed individually.

In the centre of the survey area is an area of strikingly high amplitude area that is semi-circular in shape which is approximately 15 km^2. This area is located in the

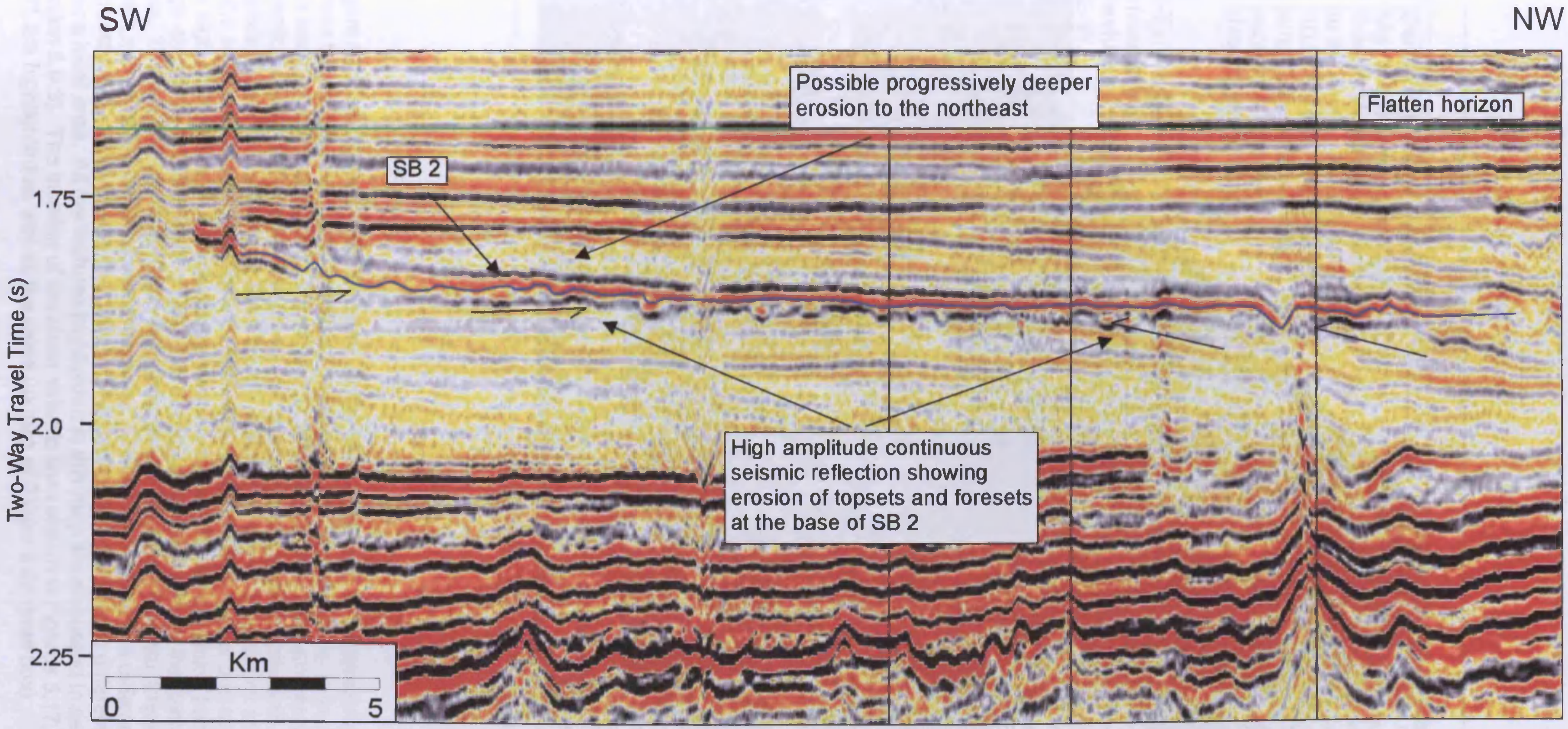


Figure 5.19. Broadly southwest - northeast flattened composite 3-D seismic line running through the centre of the high amplitude channelised area of the SB 2 reflection. This seismic line shows the flat nature of the SB 2 reflection. Erosion of dipping reflections at the base of the reflection is interpreted to represent incision into delta topsets and foresets. There is a suggestion of a progressively deepening of the erosion level to the northeast and a small local cut is seen at the northeastern end. For location of the seismic line see Figure 5.20.

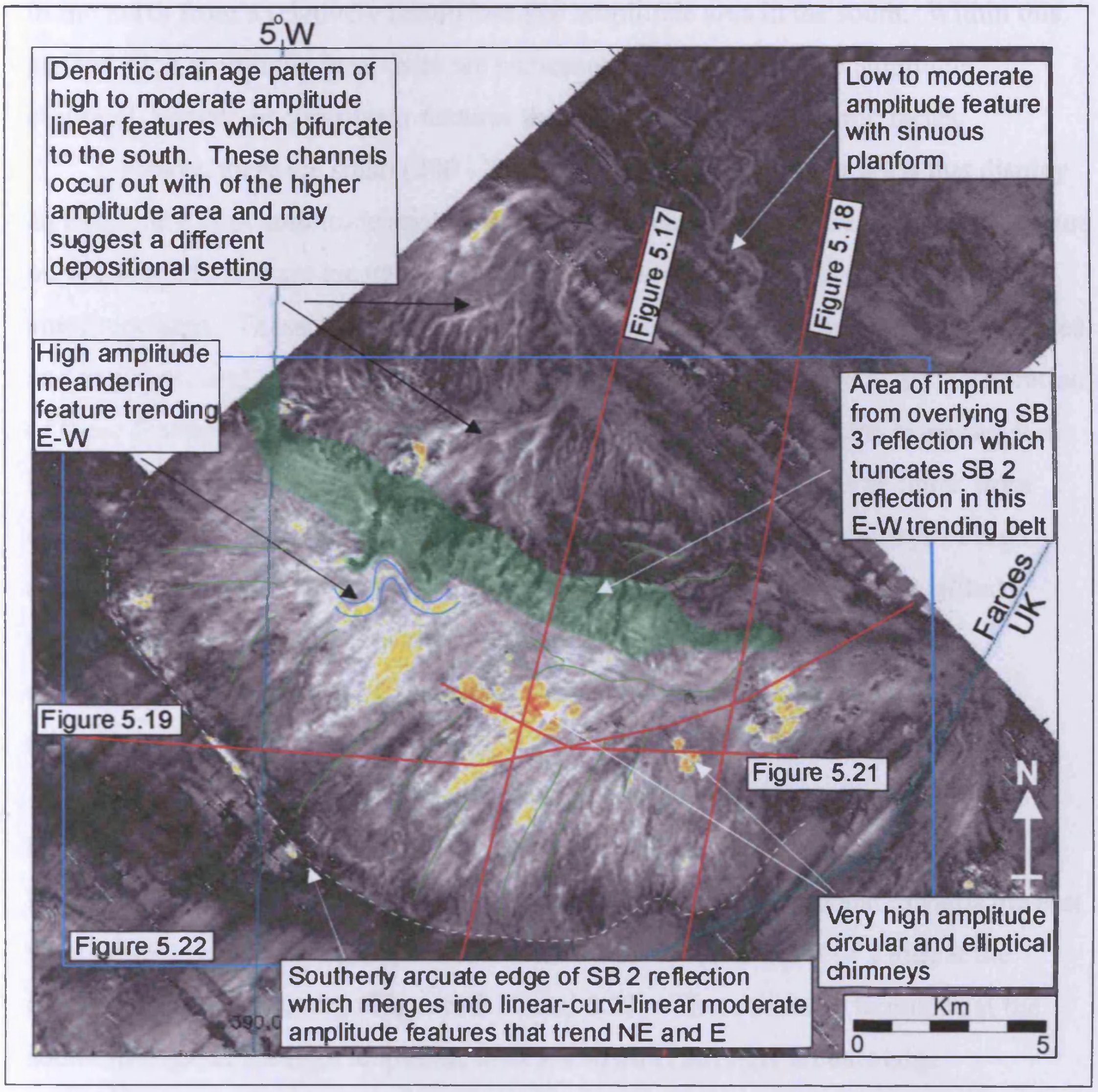


Figure 5.20. Amplitude extraction map of the SB 2 reflection. In the central part of the map (green transparent area) low amplitude incisions of the SB 3 reflection form an imprint in the map and are thus not related to SB 2 (see text for discussion). The SB 2 amplitude map reveals a complicated drainage network in a semi-circular shaped high amplitude area. High to moderate elliptical, linear and curvilinear features are visible (thin green lines) which are seen to originate at the southerly arcuate edge (white dashed line) and trend northeast and east. To the west of the high amplitude area a meandering planform is seen (blue line) which trends east - west. High amplitude elliptical bodies (250-300 m in diameter) are seen in the central high amplitude area. To the north, dendritic finger like geometries of high to moderate amplitudes are seen which merge to the north into wider features. These channels have a different planform and may suggest subtle variations on the delta top environment to give rise to different channel planforms over a local area. All the features highlighted in this map are discussed in detail in the text (see section 5.6.3). The location of the close seismic lines shown in Figures 5.17, 5.18, 5.19 and 5.21 are highlighted as well as the close up map of Figure 5.22 (blue box).

highly erosive and high amplitude part of the SB 2 reflection (**Section 5.6.2**). The high amplitude area has a strong curved southerly edge which delineates the high amplitudes in the north from a relatively featureless low amplitude area in the south. Within this general high amplitude area, there are numerous high and moderate amplitude elliptical, arcuate or curvilinear features that show a variety of seismic facies.

Firstly, there are small (200 - 300 m) elliptical often circular areas that display an extremely high amplitude seismic character. These appear as the darkest red colours on **Figure 5.20** and are located in the central part and eastern parts of the high amplitude area. These elliptical circular high amplitude features are seen to be isolated and in groups and often make up an area greater than 1 km^2 (**Figure 5.20**). Calibration of these features by a seismic line through both the central and eastern examples it is believed that these very high amplitudes are related to escape of gas or fluids from deeper down in the stratigraphic section (**Figure 5.21**). Vertical chimneys of high amplitude are visible throughout much of the stratigraphy. These high amplitude chimney features are often located over highs seen in the base T50 (Balder) unconformity and these local highs may have acted as conduits to assist in the migration of fluids or hydrocarbons (e.g. Løseth *et al.* 2001, Berndt *et al.* 2003).

Secondly, within the high amplitude area arcuate and curvilinear moderate to high amplitude features are seen that have a broadly northeast - southwest trend. Indeed there is good evidence that these features change direction and towards the east where they are seen to converge into a narrow zone (with a width of 2 km) at the eastern edge of the survey (**Figures 5.20** and **5.22**). These features terminate at the southern edge of the high amplitude area at a distinct strongly arcuate edge.

At this southerly arcuate feather edge, linear narrow dendritic features can be seen on the seismic amplitude map which bifurcates to the south at their proximal ends. These features are small (approximately 50 - 100 m wide) (**Figure 5.22**) and are located around the entire southerly edge of the high amplitude feature, a distance of over 25 km. Towards the north they merge into the larger moderate to high amplitude arcuate to curvilinear features discussed above. These small dendritic channels and their larger more arcuate counterparts are interpreted to represent a channel system that is feeding to the north and northeast. The dendritic channels on the sharp southerly edge of this feature, south of which has no seismic character, are interpreted to form from a process of mass wasting, possibly at the top of a slope (see **Section 5.6.4**). Indeed an indentation in the curved feather edge is seen east of the political boundary

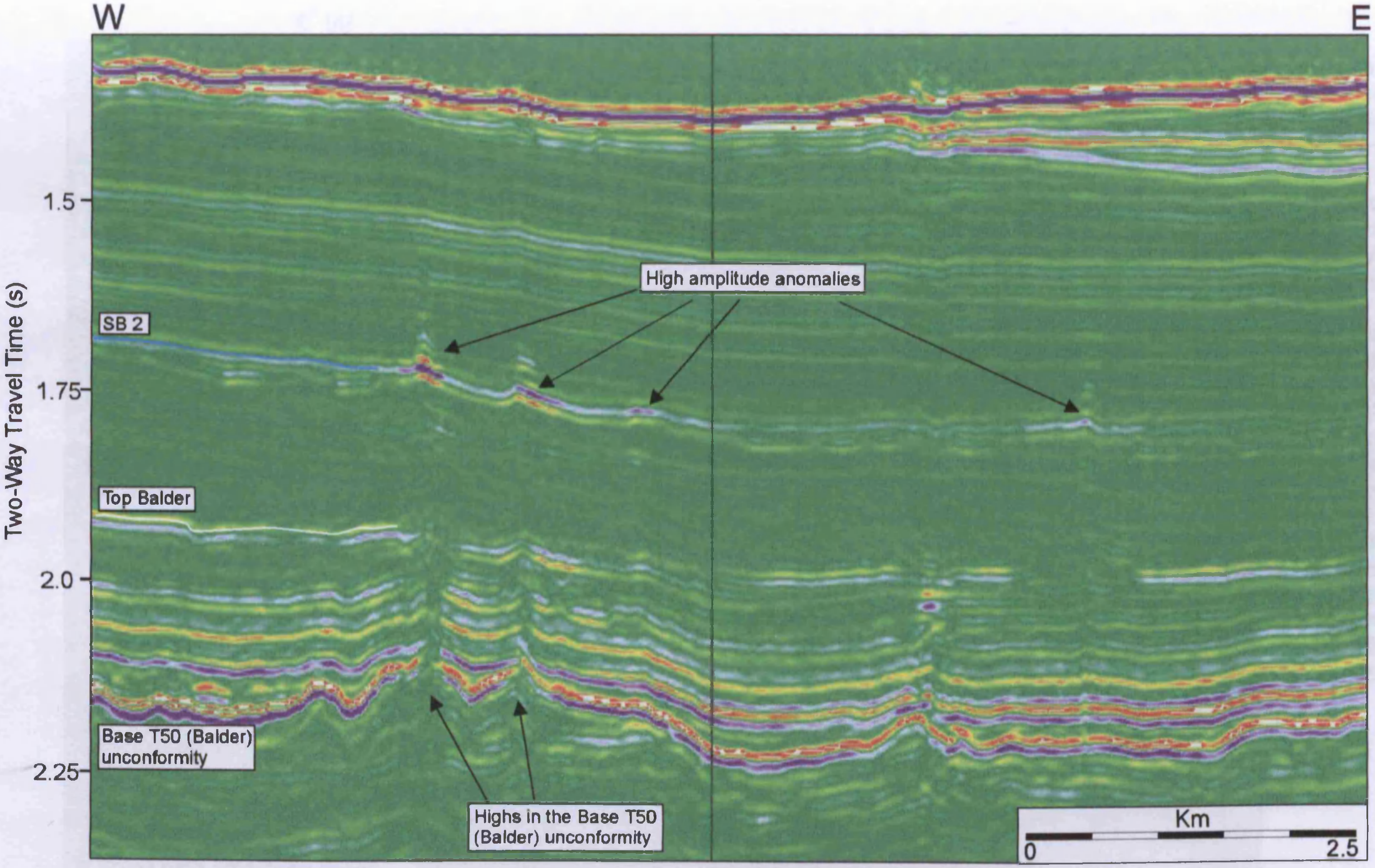


Figure 5.21. East - west trending seismic line through the high amplitude circular and lobate features seen in the SB 2 amplitude extraction map (see Figure 5.20 for location of seismic line). These circular and elliptical features in plan view are found in the central part of the high amplitude area. This seismic line through these features shows that some of the high amplitude features are directly related to highs in the base T50 (Balder) unconformity. The colour scheme of this particular figure has been altered to enhance the difference between the high and low amplitudes. The high amplitudes are shown in purples, low amplitudes in reds and the moderate amplitudes are shown in the green colour range.

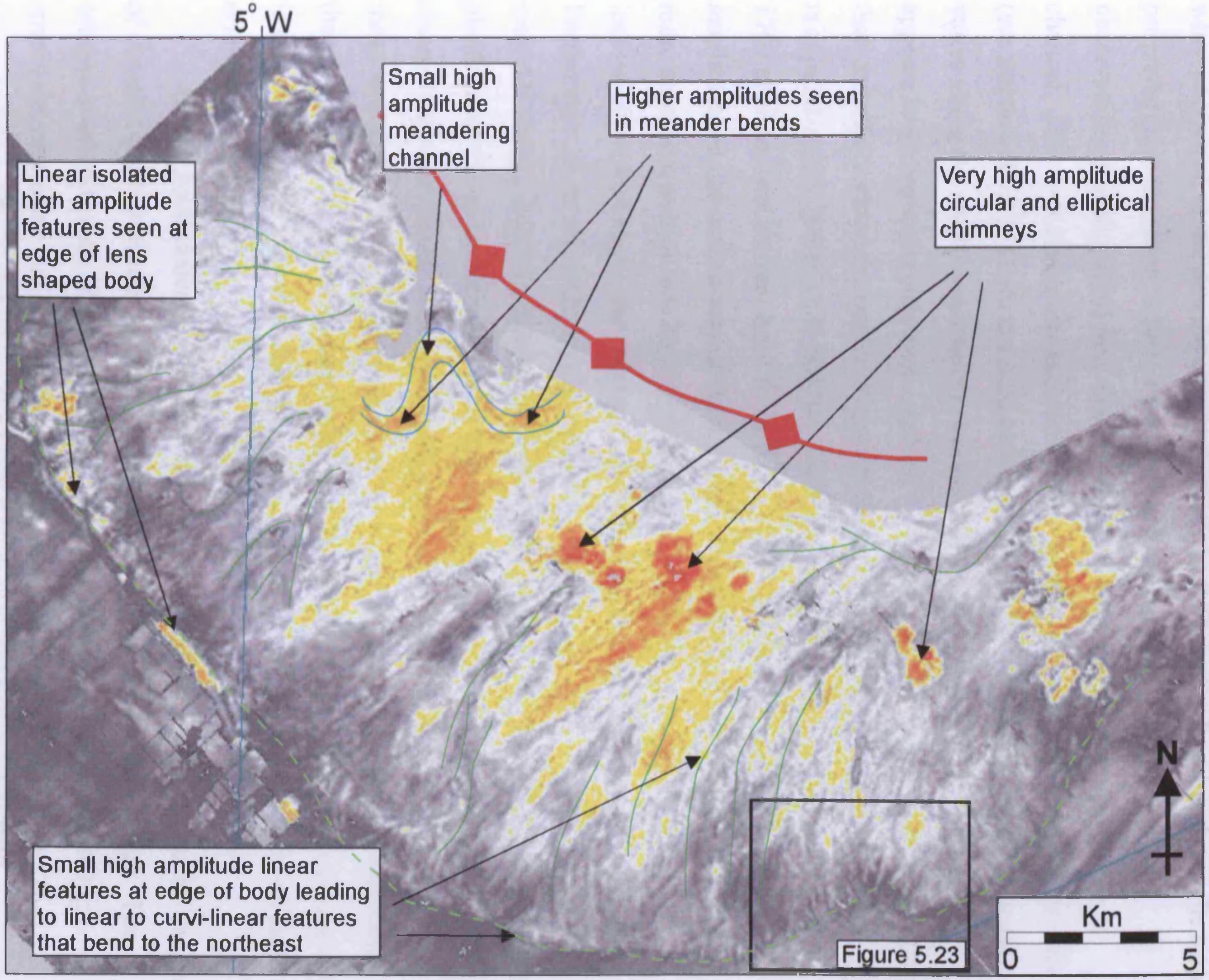


Figure 5.22. Close up image of the seismic amplitude extraction map of SB 2 showing the semi-circular high amplitude area. Towards the northwestern corner of this area a meandering channel can be identified (blue lines) with three meander loops. This is cut off to the north and east by the later erosion and truncation of SB 3 in the area of the east - west trending ridge (thick red line- see Figure 5.20). The high amplitude circular and elliptical features shown in Figure 5.21 are highlighted as well as the southerly arcuate edge pale green dashed line) which exhibits isolated linear high amplitude along its length and is where northeast - east trending linear and curvilinear channels originate (green lines). The location of Figure 5.23 is shown.

and is interpreted to represent a possible collapse zone greater than 1.5 km^2 in area (**Figure 5.23**).

Towards the northwestern edge of this higher amplitude area is a moderate amplitude highly curved seismic amplitude anomaly. This feature is approximately 2 km long and has a width (picked out by the moderate amplitudes) in the region of 200 – 300 m (**Figures 5.20** and **5.22**). The planform of this feature is highly meandering even though on the amplitude map there are only signs of three meander loops being present. The meandering feature trends broadly east – west. This seems ambiguous when compared to the downlapping reflections on the seismic data that shows the main progradational direction to be towards the north and northeast. From these observations it is suggested here that this highly curved feature is a meandering channel. However it remains unclear whether the channel is fed from the east or west (see **Section 5.6.4**). There is a suggestion that the width of this meandering channel varies along the short length seen in the amplitude map. Noticeably, the channel appears to be approximately twice as thick on the two southerly convex meander loops than the sole northerly meander (**Figure 5.22**). The widest part of the meander channel is approximately 200 m wide and the channel as a whole has a wavelength of about 1700 m (from one sinuous bend to the next). There are subtle indications of higher amplitudes in the outer bends of the meanders (**Figure 5.22).** This is indicative of a more sand-rich section that has a higher acoustic impedance contrast that may be located in the thalweg of the channel than the background low amplitude claystone. Furthermore, a subtle evidence of a meander cur-off loop is seen at the northern meander though this is interpretation is speculative (**Figure 5.22).** The orientation and planform of this meandering channel may be slightly ambiguous as the majority of the channels are curvilinear and they trend towards the north or northeast. This meandering channel has an expression on the underlying seismic reflection and thus there may be a slight imprint seen at the SB 2 stratigraphic level. However, the feature is real and remains important and may indeed suggest an evolution of planforms and orientation through time.

On the southeasterly edge of the high amplitude area there appears to be a series of elongate high amplitude linear features that lie parallel to this arcuate edge. The seismic amplitude map (**Figures 5.20** and **5.22**) shows three or four elongate high amplitude anomalies that are isolated and separated by low amplitude areas. The longest of these elongate features is 2 km long and has a width of only 100 -200 m. It

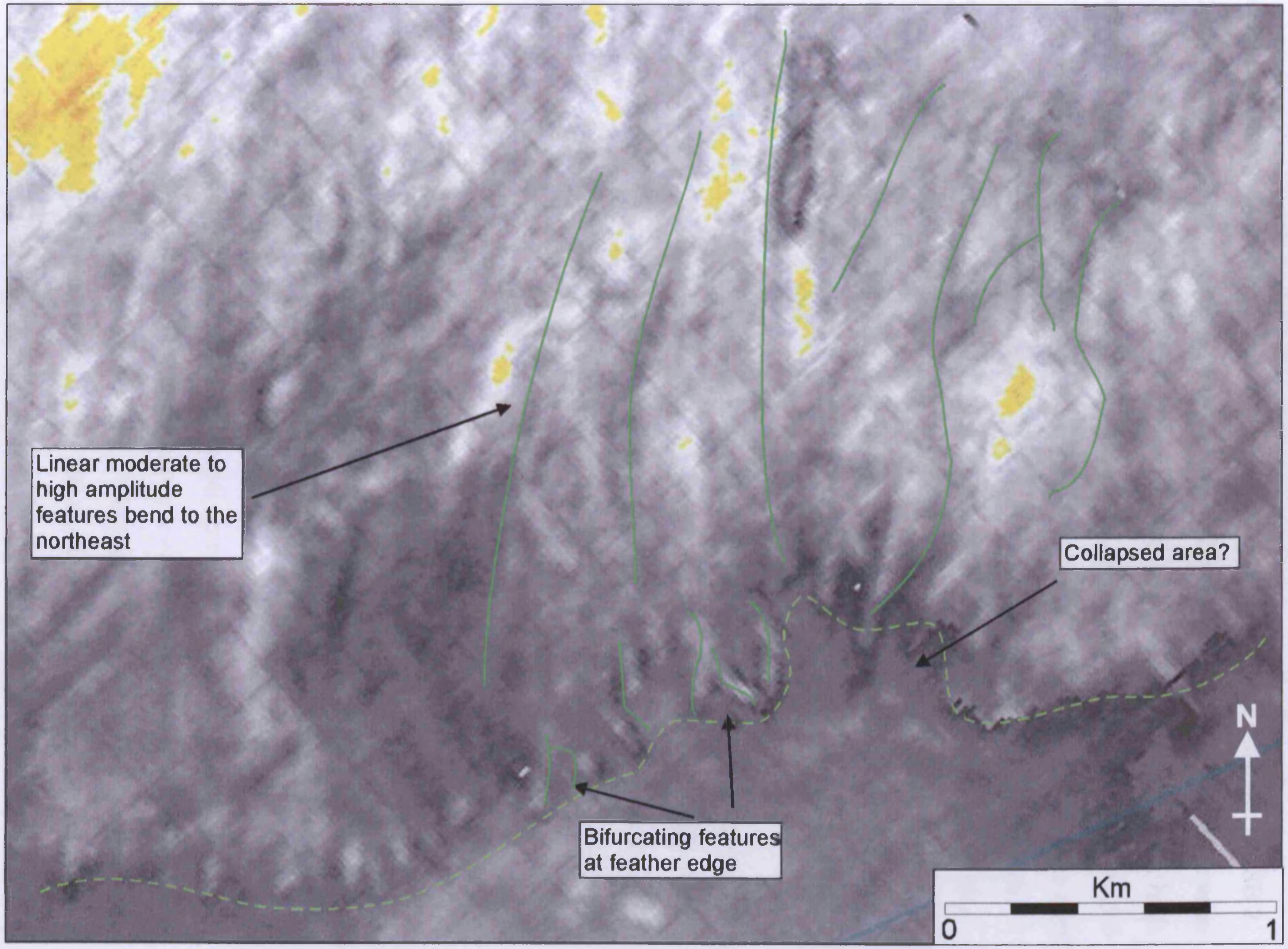


Figure 5.23. Close up of the southerly feather edge pale green dashed line) of the semi-circular high amplitude area of the SB 2 amplitude map. Small moderate amplitude linear features (green lines) are seen to originate from the southerly edge of the feature and merge into each other to the north. These features become higher in amplitude to the north and pass into linear or curvilinear to slightly sinuous amplitude highs. These are interpreted as small channels which trend north - south and bend to towards the east. The southerly edge of the feature is not entirely curved and smooth with one particular area showing a large embayment in the edge and this may represent a collapsed area of channel system.

is observed to have a slightly cuspate southerly edge to the feature and a more curved northern front. However, other high amplitude areas on this southerly fringe do not exhibit the same dimensions, and a more patchy elliptical form is seen on the fringes (where the dimensions are 1 km by 200 - 300m). Together these features form the southerly limit of the high amplitude area, south of which a low amplitude featureless seismic character is seen (**Figure 5.20**).

A final observation of **Figure 5.20** is found north of the high amplitude area and indeed just north of the later (SB 3) east –west canyon system that truncates the SB 2 reflection. Here, high amplitude linear and curved features (50 -100 m wide) are seen to form a finger-like geometry that trends north northeast. Numerous curvilinear high amplitude features join into each other towards the north and become wider (up to 200 m). A further observation is that these features become more sinuous towards the north of the survey, though here the seismic amplitude value becomes lower and they are more difficult to see (**Figure 5.20**). Additionally just north of the eastern end on the superimposed (SB 3) canyon belt there is evidence of high amplitude linear features that trend east – west and bifurcate to the west. All these features to the north of the canyons are interpreted as channel systems that feed predominantly to the north east, though occasionally eastwards.

In summary, the high amplitude area located in the northern part of the survey is interpreted to represent an area of complex drainage systems that show a variety of channel networks (**Figure 5.20**). The interpretation of the environment of deposition of all the features seen on the SB 2 amplitude extraction map will now follow in **Section 5.6.4**).

5.6.4 Interpreted Depositional Environments

From interpretation of seismic lines and the use of seismic facies analysis to interpret channel systems, a depositional environment can be constructed for unit 1B, and in particular the significance of the SB 2 reflection can be examined.

The stacking pattern of unit 1B is largely progradational with some initial aggradation in the very south of the study area above the interpreted coals of the unit 1A. Unit 1B progrades in the region of 20 -30 km and from clinoform geometries into water depths believed to be approximately 100 m. The SB 2 reflection truncates numerous topsets and indeed the upper part of the foresets in the prograding wedge.

Clinoform break points are therefore difficult to observe because they have been removed by erosion, however towards the north outwith the area of erosion clinoform break points are observed and show a down-stepping pattern. From these observations it clear to suggest that as a whole unit 1B prograded north over the earlier unit 1A with the shoreline lying some 15 – 20 km to the north of the SB 1 break point (**Figure 5.24** and **Table 5.2**). The recognition of these clinoform break points is crucial as it allows for the depositional environment to be inferred. The channel systems are all located to the south of the inferred SB 2 shoreline position and thus are interpreted to be sub-aerial in origin and occupy a position on the delta top. In marginal – shallow water environments, there may only be a slight incline on the shelf and thus only small variations in relative sea-level may move facies belts and depositional systems significant distances (**Section 5.6.5**). The observation of different orientations in the drainage network with the meandering channel trending in a broadly east - west orientation and the smaller more linear channels draining north to northeast suggests that the main direction of drainage on the delta top was variable at this time. However the main shoreline and basin slope and depocentre are interpreted to be towards the north or northeast as this is where the channels ultimately find their most favourable course.

Given that the drainage network is accepted as being sub-aerial in origin, it is now necessary to explain the variety of channel planforms and other features seen in the seismic amplitude map. The most striking of these features is the large high amplitude area which has the southerly arcuate edge and curvilinear and meandering channel systems developed. The southerly edge has a higher relief than the centre of the reflection (which is broadly concave up) and the striking arcuate geometry of this feature may suggest a failure scarp, implying a slump or slide origin. It has not been possible to establish if this southerly edge joins the northern edge and thus closes to form an elliptical geometry. The apexes of any possible ellipsoid sit outwith of the 3-D survey area. However, small dendritic channels are seen to "waste" from the fringe and a possible collapse zone is inferred towards the eastern side of it (**Figure 5.23** and **Section 5.6.3**). However, it remains clear that the position of this erosive area lies in the delta topsets (indeed clinoforms are truncated under the SB 2 reflection in this area) and thus no significant depositional slope is envisaged in this area to create instability to excavate a vast area of over 15 km^2. Furthermore, the northern edge of the concave up high amplitude SB 2 reflection (located in the area of the later canyons) forms a

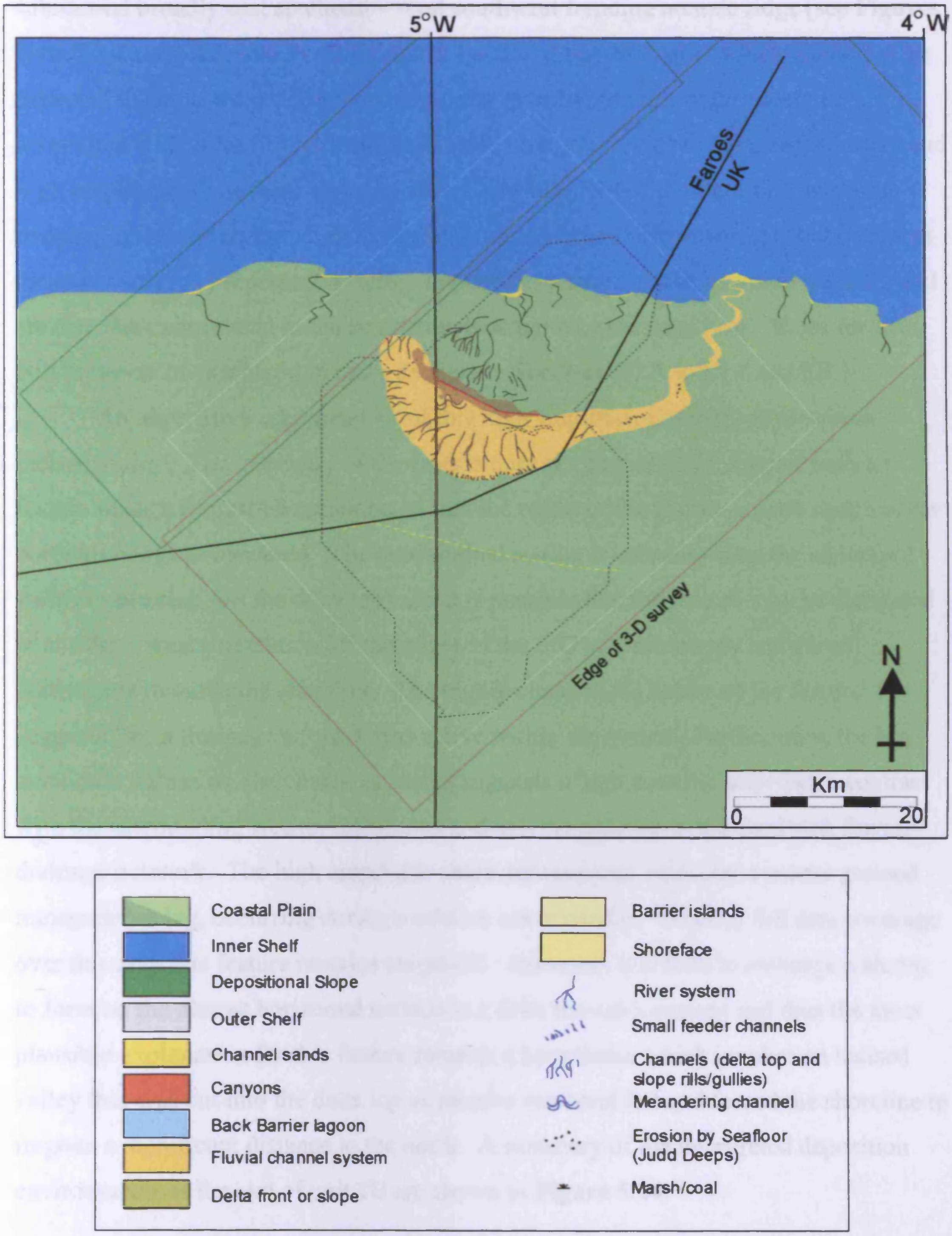


Figure 5.24. Schematic palaeogeographic map of the SB 2 incision surface. The interpreted shoreline is towards the northern end of the survey area. When compared to the shoreline which is interpreted from the SB 1 reflection (yellow dashed line and Figure 5.14) there is a migration northwards movement in the region of 30 -35 km. A large incised valley system (greater than 15 km 2) is interpreted to sit some 15-20 km inland. The northerly limit of the incised valley remained high and formed a positive barrier (shaded brown area and red line) and thus the incised valley is interpreted to swing to the east and eventually out to the shoreline to the east of the survey. There is a prominent arcuate trend to the channel systems that develop in the incised valley, and meandering and curvilinear forms all trend east or northeast. On the southerly edge of the incised valley there are small rills or runnels which feed into larger channels. For legend of the colours used for the depositional environments see Table 5.2 or Enclosure P.

substantial broadly east southeast – west southwest trending arcuate ridge (see **Figures 5.17, 5.18** and **5.22**) that forms a ridge or barrier of positive relief which would not be expected to occur from a slide or slump. The possible arcuate failure scarp is coincident with areas of high amplitude small elongate features. By their geometry and high amplitude alone these could be interpreted as sand-prone ridges or shoals or a fringing carbonate reefs though this is highly speculative. However, it is believed that these features may represent a further imprint from seismic reflections above SB 2 and are therefore not related to the sub-aerial drainage network seen here. These features will however be discussed in more detailed in **Sections 5.7.3.3, 5.7.4** and **5.9**.

An alternative interpretation of this high amplitude channelised area is an incised valley. The geometry of the feature does not immediately suggest such a feature though it must be remembered that the edges of the feature are not seen and the northern fringe is obscured. The depositional setting is certainly ideal for an incised valley to develop (on the delta top) and it is possible that the feature may be elongated in an east – west orientation off the edges of the 3-D seismic survey and curve northwards towards the shoreline. The highly channelised nature of the feature suggests that a drainage network was active within the system. Furthermore, the high amplitude values of the channel systems suggests a high acoustic impedance contrast with the surrounding weaker reflections and may be indicative of a sand-rich fluvial drainage network. The high amplitude character may also represent a coarse grained transgressive lag, occurring during a relative sea-level rise. Without full data coverage over this area this feature remains enigmatic. However, it is hard to envisage a slump to form on the almost horizontal surface in a delta top environment and thus the most plausible explanation for this feature remains a hypothesis which invokes an incised valley that was cut into the delta top as relative sea-level fell and forced the shoreline to migrate a significant distance to the north. A summary of the interpreted deposition environments at the end of unit 1B are shown in **Figure 5.24**.

5.6.5 Relative Sea-Level Interpretation

The marked progradation of the deltaic system between units 1A and 1B is dramatic with over 25 km of migration of the shoreline. However, immediately after the incision which created SB 1 relative sea-level rose and transgression of the shoreline occurred. This transgression is evident from the southward progressive onlap of seismic reflections onto SB 1. No evidence of retrogradational stacking patterns are seen in the early stages of unit 1B and this is because the gradient of the shelf is believed to have been extremely low (less than 1°) and thus any small rise in relative sea-level would have flooded back quickly and hence would have covered the near horizontal delta top environment and moved the position of the shoreline vast distances. This transgression is interpreted to be relatively short-lived as clinoform systems of the next highstand are seen to prograde across the shelf, though this is not possible to confirm due to poor age dating. However, transgressive events are often condensed and thus represent a long period of time. Therefore it is not possible to make a link between thickness of sediments and time. Thus only a very thin succession of sediments is preserved before the system progrades north again during the continued sea-level rise. The clinoform packages downlap to the north onto the flooded SB 1 reflection and further north, eventually onto the top T50 (Balder Tuff) reflection. Water depths are interpreted to be in the region of 100 m (by inferring the height between any observed clinoform break point and their corresponding downlapping bottomset position) and this gave rise to a more sigmoidal geometrical shape of clinoform system.

At the end of unit 1B there was a significant fall in relative sea-level (approximately 80 – 100 m) which created a major type 1 unconformity (SB 2). This relative sea-level fall exposed the delta top area and caused significant amounts of incision with the development of an extensive drainage network perhaps with incised valley systems. North of the area of delta top incision a small unit of down-stepping clinoform packages is seen basinward of the shoreline and this constitutes a small forced regressive wedge of sediments which bypassed the delta top area during this period of relative sea-level fall.

5.6.6 Implications for South Judd Basin Evolution

The location of the shoreline some 30 km north of the shoreline at the end of unit 1A indicates that this marginal environment was highly susceptible to small changes in relative sea-level. Indeed it is advocated here that the SB 2 reflection was created by a substantial fall in sea-level in the order of 80 – 100 m and this had the effect of exposing the delta top and creating highly incised drainage networks. Because of the evidence of exposure and incision into the delta top the SB 2 reflection can be defined as a type 1 SB (Van Wagoner *et al.* 1988). Furthermore, the unit 1B represents a sequence of the Exxon model (**Section 3.2**) which is bounded above and below by unconformities or their correlative conformities (Mitchum *et al.* 1977a). In between these two sequence boundaries one would expect to find a transgressive (or ravinement) surface though the transgression episode is represented by a condensed section. Progradation renews after the transgressive episode though the duration of the transgression is unknown and may represent along period of time. SB 1 seen at the top of unit 1A shows a similar pattern during a forced regression, though with no erosion of topsets or the upper part of the foreset evident it is difficult to estimate the fall in relative-sea level. However the development of a forced regressive wedge at both SB 1 and SB2 indicates that the shoreline position moved basinward and down-stepped thus producing type 1 SB's. Therefore it is here concluded that unit 1B is a type 1 sequence.

The lateral extent of SB 2 shows that whilst being of local to semi-regional extent (an evident erosive area of at least 20 km^2) within a basin-wide context this is very small indeed and it is only with the positioning of 3-D surveys on marginal areas that such SB's can be accurately defined. Indeed, SB 2 is not entirely contained on the 3-D survey and thus its true lateral extent is not known. An adjacent or nearby 3-D survey may not show any erosion at this stratigraphic level and a type 1 sequence will not be inferred. However, in reality the lack of erosion indicates the correlative conformity to SB 2 though it would not be interpreted to represent this.

One final point of consideration is the cause for this relative sea-level fall. The shoreline is interpreted to have advanced northwards with accommodation space decreasing throughout the relative sea-level fall. The fall in relative sea-level could be explained by the rate of eustatic fall which outpaced the rate of subsidence to produce a forced regression. However, it is more likely that local tectonic controls leading to uplift and compression on localised inversion axes caused the fall in relative sea-level fall. It is known that the Wyville-Thompson Ridge was experiencing a phase of north – south directed compression in the latest Palaeocene –earliest Eocene which is attributed to ridge push forces during the early opening of the North Atlantic (e.g. Boldreel and Andersen 1993). However additional structures may have been in their early stages of compression at this time (e.g. the Judd Anticline and the Munkagrunnur Ridge). This discussion will be continued in **Section 5.9**.

5.7 Seismic unit 1C

5.7.1 Introduction and Age

The third seismic unit bounded at the top by the SB 3 reflection and at the base by the SB 2 reflection, will be discussed here. In the central and northern area of the survey the SB 3 reflection is found very close to the top of the SB 2 reflection and in places truncates the lower surface (see **Section 5.6.2**). Unit 1C is thus very thin or even non-existent throughout much of the study area due to erosion. An Early – Middle Ypresian age is assigned to unit 1C with this relative age being as accurate as it is possible to date the succession (see **Section 5.3.2**).

5.7.2 Seismic Reflection Interpretations and Stacking Patterns

Unit 1C is very thin compared to the other three units as the SB 3 reflection is found in places very close to the SB 2 reflection. However, major incision is seen at the base of SB 3 and because of its moderate to high amplitude and continuous nature it is possible to map the reflection around much of the study area.

Unit 1C shows initial onlap south onto the SB 2 reflection and then the system progrades out to the north into the basin (**Figures 5.25a** & **b**). The unit appears slightly basinward of unit 1B (in a same way as unit 1B is found basinward of 1A). In the areas of the close proximity of the two SB's no internal architecture is visible in unit 1C. However to the south of the survey area unit 1C is dominated by a suite of dipping seismic reflection configurations which downlap and infill the high amplitude near horizontal incised valley system seen on SB 2. These dipping reflections are continuous, moderate amplitude and show some slight divergence to the north (**Figure 5.25a**). Further south these reflection configurations flatten out and become higher in amplitude. The dipping reflections are interpreted as foresets to clinoform systems that are prograding to the north. The horizontal part of the reflection is interpreted as a delta topsets. The clinoforms are prograding into water depth is interpreted to be approximately 50 - 70 m (**Figure 5.25a**).

The top of the unit (SB 3) displays moderate amplitude continuous configuration which laterally becomes higher amplitude in discrete areas of approximately 1 km across (**Figure 5.25a**). These higher amplitude areas are often

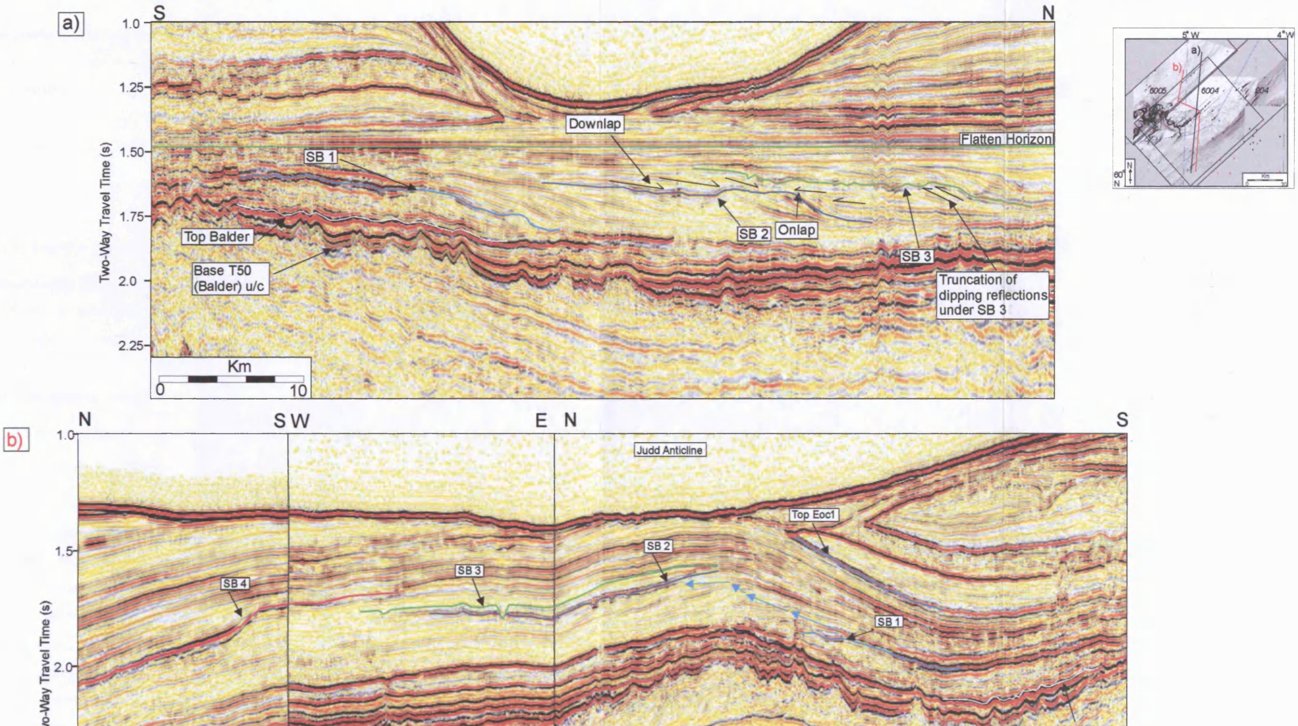


Figure 5.25. a) South - north trending 3-D seismic line that has been flattened on the green horizon shown and shows the progradational nature of the early 1C unit that downlaps and infills the high amplitude concave up (near horizontal) incised valley system that developed during fall in relative sea-level at SB 2. The 1C unit sits basinward (to the north) of the preceding units and is then a fall in relative sea-level creates the erosion truncation seen under SB 3 and the observation of truncation of dipping reflections at the maximum extent of progradation. b) Composite 3-D seismic traverse trending broadly north - south showing the very thin nature of the 1C unit with respect to the other three. SB 3 can be seen to lie very close to SB 2 in parts of the study area and does indeed truncate the lower sequence boundary. This lower seismic line also shows the northward migration of the units from 1A -1D and the characteristic high amplitude continuous seismic configurations the sequence boundaries exhibit. For location of the see map in top right hand corner

located at the position of small incisions into the underlying unit. However, this is not always seen and some local areas of incision remain moderate amplitude (e.g. **Figure 5.25b** shows a significant incision truncating the lower SB 2 reflection). One final interpretation of the seismic reflection configurations is the observed truncation of the dipping foresets underneath the SB 3 reflection close to the maximum limit of progradation.

5.7.3 Seismic Facies Description

Three strikingly different depositional features are seen in the seismic amplitude map (**Figure 5.26**). A general summary of the features seen in the map highlights both high amplitude and low amplitude features of varying geometries which take the form of a number of different shaped incisions and these will be discussed individually in the following sections (**Sections 5.7.3.1** to **5.7.3.3**).

5.7.3.1 Low Amplitude Incisional Features

Firstly, a spectacular series of incisional surfaces are observed in the northern part of the survey (**Figure 5.26**). These features are also seen in the amplitude map of the underlying SB 2 reflection (see **Figure 5.20**) creating an imprint in from above. However, as will be proved here, these low amplitude incisional features are wholly associated with erosion of the SB 3 reflection and their geometry and seismic facies will be discussed here.

There is a strong alignment to these incisions which are orientated in a broad east - west trending curvilinear belt that stretches for approximately 8 km. The incisions are picked out by their characteristic low seismic amplitude fill (seen by the light colours in **Figure 5.26**). This set of incisions are situated to the north of the earlier SB 2 channel features (**Section 5.6.3**) and consists of up to eighteen individual incisions of varying length and width. Because of their geometry and shape and most importantly because they display erosional truncation at their bases, these features will herein be termed canyons, and their morphology will be discussed in greater detail.

The largest of these canyons (**Canyon A**) is located at the western end of the canyon belt and is almost 2000 m long with an average width of 250 m (**Figure 5.26**). Canyon A has a distinct curvature at its southern limit in the head region, where it

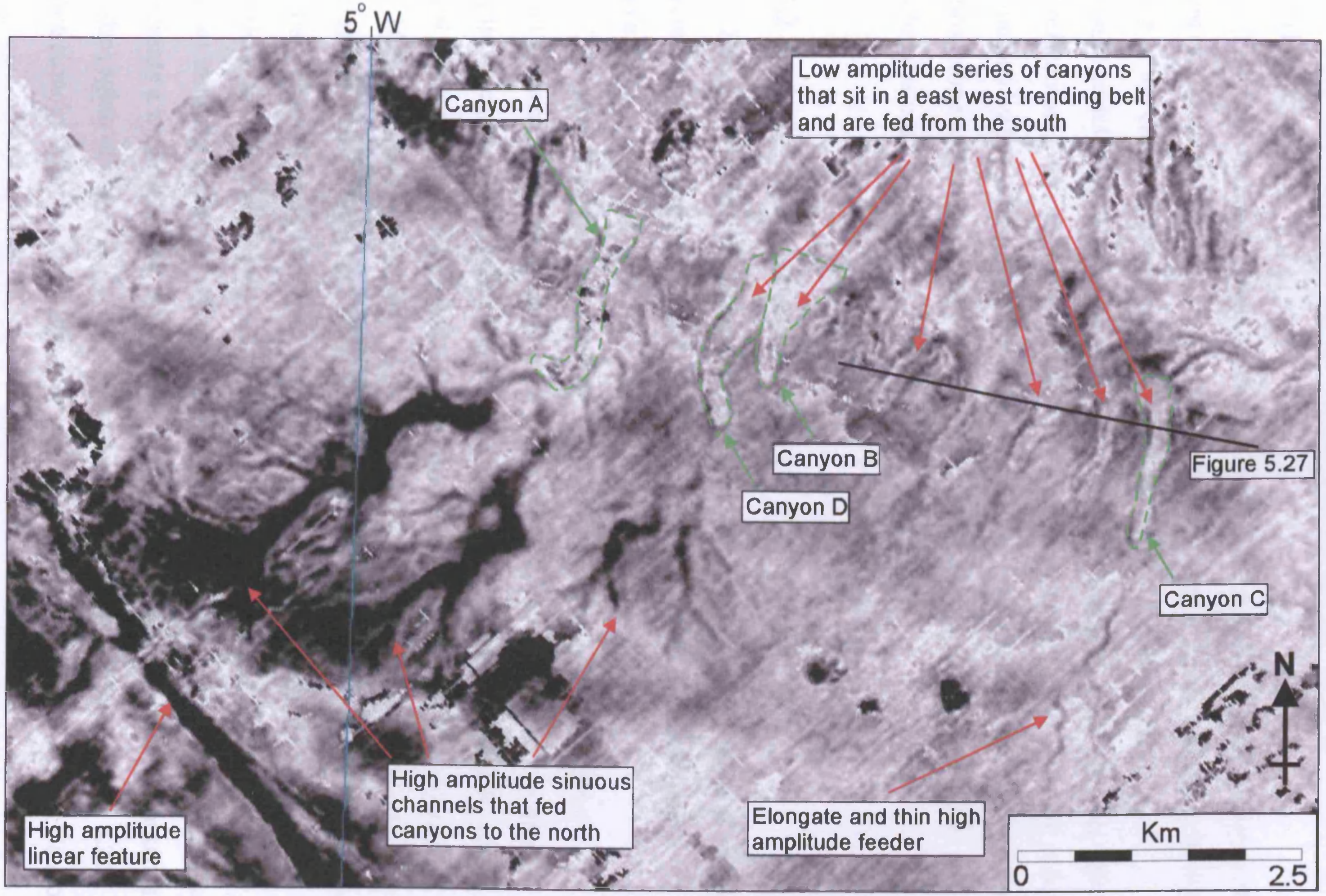


Figure 5.26. Amplitude extraction map of the SB 3 reflection in the central part of survey are in the same area as the channelised incised valley system seen in the SB 2 amplitude map (Figure 5.22). Numerous features displaying varying seismic facies and planforms are visible. On this map the high amplitude values are shown in black colours and the low amplitude values shown in light grey to white colours. An east west trending belt of low amplitude linear features are seen and termed canyons (highlighted in green). Feeding these canyons are sinuous bifurcating high amplitude channels that vary in shape and planform. Furthermore a high amplitude northwest southeast trending linear feature is visible to the south and west. (See section 5.7.3 for discussion)

bends dramatically westwards almost at right angles. This seems to be unique in the set of canyons as all the others are predominately narrow and have no sharp bends along their length. Canyon A is the longest of the canyons when compared to the others. All the canyons have varying lengths between 400 m and 2000 m long, though the average length of the canyons is approximately 800 - 1000 m, with only four of the eighteen canyons exceeding 1200 m. Most canyons are in the region of 200 - 300 m in width but one canyon (Canyon B) opens out more at its distal end into a 500 m wide erosional valley (**Figure 5.26**).

The depth of these canyon features is quite uniform with a maximum depth of erosion of approximately 60 – 70 ms. There does not seem to be a relationship between the size of the canyons (length and width) and the depth of incision. This canyon incision is seen to truncate the earlier SB 2 reflection in this area of the canyons. The SB 2 reflection is of much higher amplitude than the overlying SB 3 and can be seen to be truncated by the individual canyons (**Figure 5.27**). Using the average width of these canyons, slope angles in the region of 25° are calculated as an average for these erosional systems.

5.7.3.2 High Amplitude Sinuous Features

South of the canyons discussed above (**Section 5.7.3.1**) further features with a characteristic high amplitude response are seen from the seismic amplitude map (**Figure 5.26**).

The seismic amplitude map of the SB 3 reflection shows high amplitude channel systems which can be seen to be to link into the head-reaches of the larger canyons (discussed above in **Section 5.7.3.1**). Most strikingly are the largest of these channel systems found in the western part of the map that are up to 3.5 km long and 500 – 1000 m wide (**Figure 5.26**). These channel systems are seen to act as feeders to the largest canyon (Canyon A). The Canyon A feeder systems show two channels that have main channel axes that are fed by small branching tributaries that are seen to bifurcate to the south into a low amplitude area north of an elongate northwest – southeast trending linear belt. The impression gained from the observed pattern is that it represents a well organised, dendritic network (e.g. Schumm 1987). The tributaries are in the region of 200 m wide and have a maximum length of 500 m (**Figure 5.26**). The two sinuous channels join at the head of Canyon A to the north. Similarly Canyon

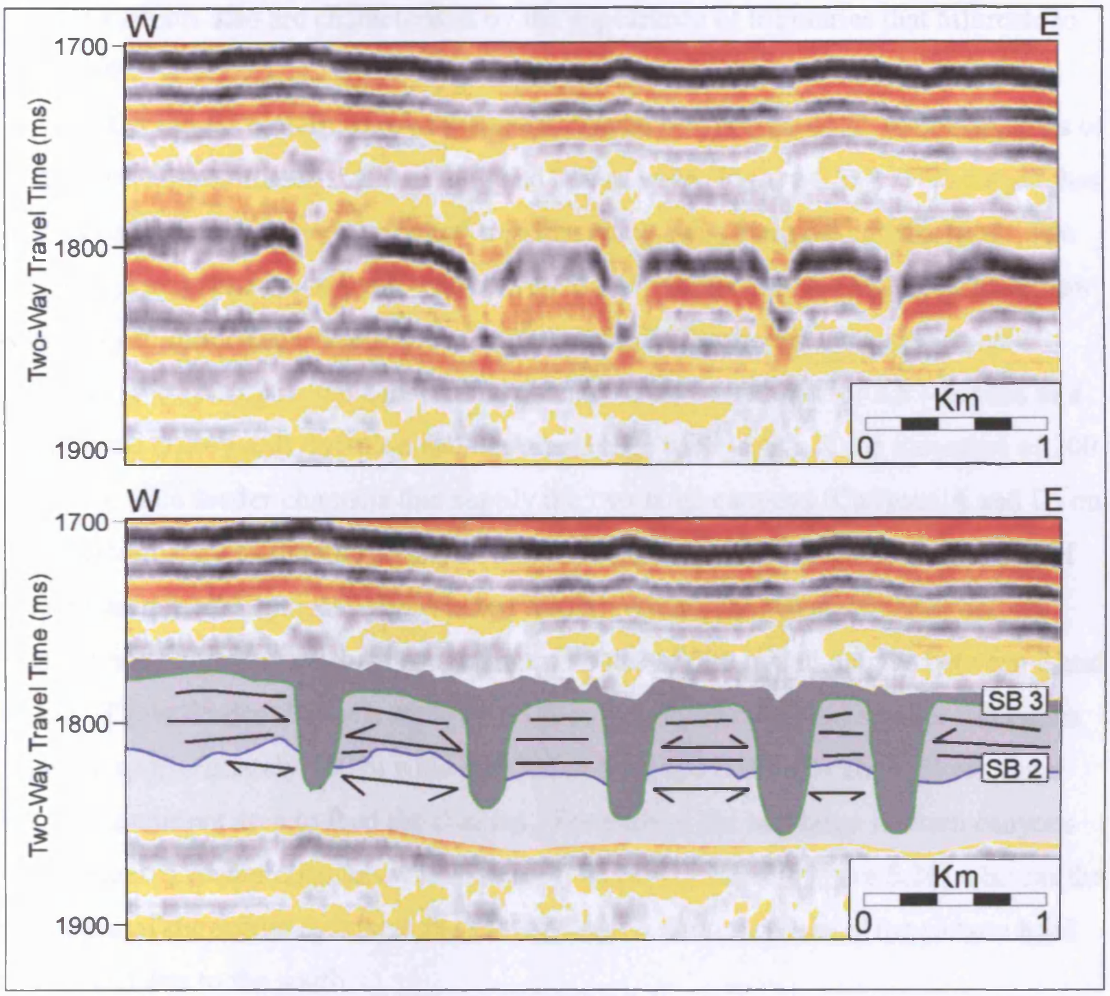


Figure 5.27. East west trending 3-D seismic line through the low amplitude incisional features (termed canyons) shown in Figures 5.26 and 5.20 which form an elongate linear east - west trending belt of 18-20 canyons. The top seismic panel shows un-interpreted section showing high amplitude SB 2 reflection which is eroded by the lower amplitude SB 3 reflection. The lower seismic panel shows the interpreted section with the deep canyons interpreted and showing significant down-cutting into the SB 2 reflection. Individual canyons are in the region of 200-400 m in width and are approximately 50-70 ms in vertical time. The length of the individual canyons varies, though an average of 800-1000m is seen. For location of seismic line see Figure 5.26.

D, seen 1 km west of Canyon A shows the same pattern of two sinuous feeder systems that join at the head of the canyon. However, the channels are smaller in size (approximately half the dimensions of the feeder systems that feed Canyon A). These feeder systems also are characterised by the appearance of tributaries that bifurcate to the south.

On the far eastern edge of the linear canyon belt is Canyon C, the dimensions of which are approximately 1400 m long and 200 m wide (**Figure 5.26**). This canyon has a very different feeder system found trending south away from the canyon head. The feeder system appears as a very narrow (up to 100 m) moderate to high amplitude, low sinuosity individual channel which is only seen to bifurcate once into similar size channel (**Figure 5.26**). This individual channel system stretches for 3.5 – 3.7 km in a southward (landward) direction and has an average wavelength of the meanders of 200 - 250 m. The feeder channels that supply the two large canyons (Canyons A and D) on the western edge of the linear canyon belt have entirely different planforms to that of this eastern feeder system. The western feeder systems show less sinuosity and are much wider in geometry, having maximum widths of 500 m and lengths between 2 and 3 km. These feeder channels show bifurcation into more landward smaller tributaries that are approximately 100 m wide and 300 m long and this bifurcation allows for a larger catchment area to feed the channel. For each of the two large western canyons there are two feeder channels which meet at the canyon head (**Figure 5.26**) whereas the channels of the feeder system to the east of Canyon C do not meet at the canyon head but over 1 km to the south.

A final observation of these feeder and canyon systems is that the dimensions of each system seem to possibly show an inter-relationship. The largest feeder systems in the west seem to feed the largest canyons (Canyons A and C). Moreover, the smallest canyons in the centre of the belt do not show any signs of feeder systems, though this may be a seismic resolution problem. Conversely, Canyon C in the east is relatively large (1400 m long) but it is only fed by a small feeder system, though as has already been discussed it has a very different planform.

A summary of the canyons and the feeder systems is shown diagrammatically in a schematic figure (**Figure 5.28**) which draws out the important observations of both the feeder systems and the canyons.

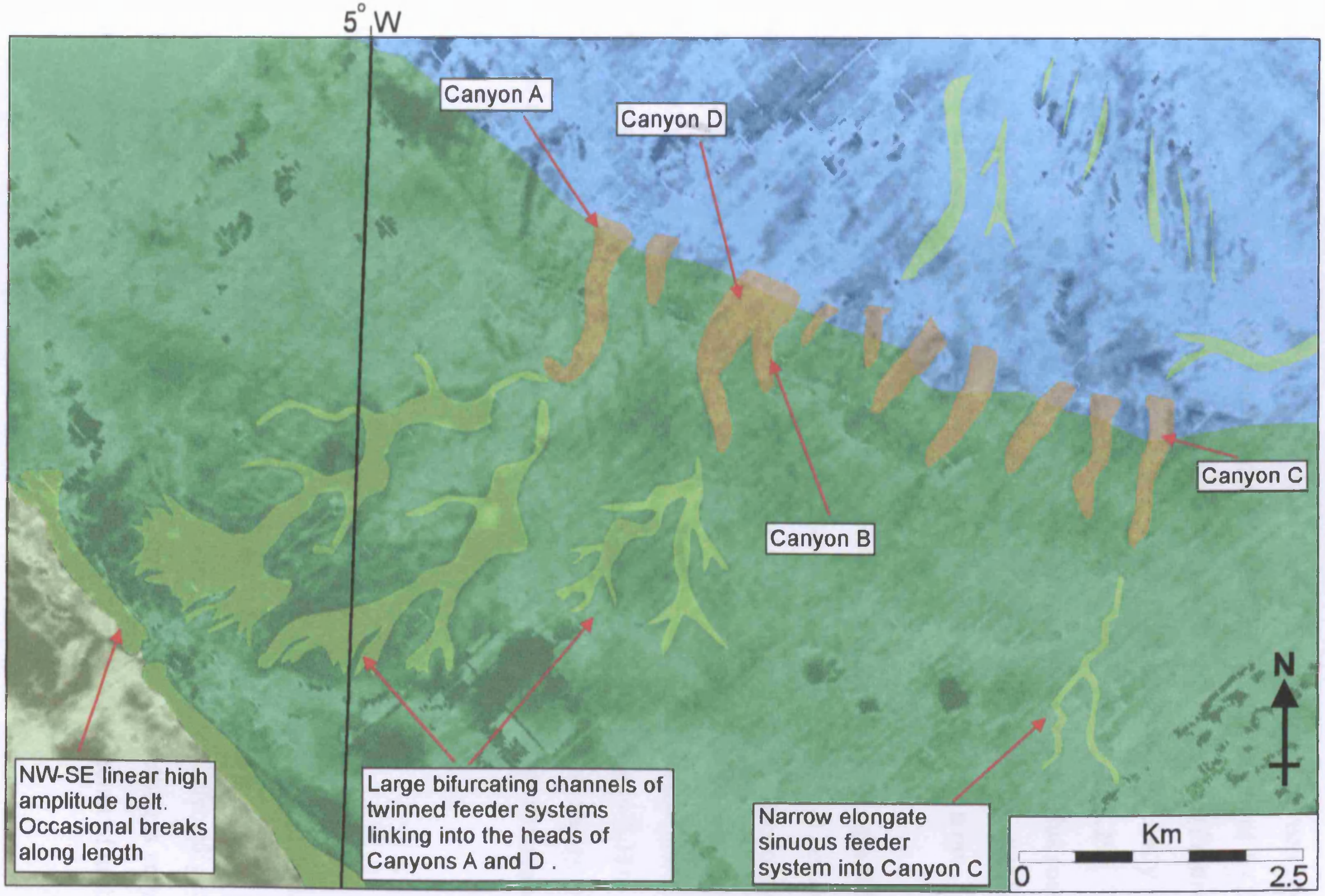


Figure 5.28. Schematic cartoon diagram overlain over the amplitude extraction map of Figure 5.26. This diagram summarises the key depositional features seen in the SB 3 reflection. The green and blue colours resemble the position of a delta top and possible delta front. The yellow coloured channel systems on the delta top indicate the feeder systems that are sinuous and link into large canyons (shown in orange). The yellow channels seen north of the canyons and may be on the delta front and show different geometries and some are orientated perpendicular to the canyons and trend east - west.

5.7.3.3 High Amplitude Linear - Arcuate Feature

The final feature identified within the amplitude map of SB 3 is the southerly linear high amplitude belt that is located just south of the feeder systems discussed above (**Section 5.7.3.2**). A detailed map of this feature is shown in **Figure 5.29** which highlights it shape and form.

This feature forms a 16 km long elongate belt and has a northwest - southeast orientation and is arcuate in planform. The thickness of the belt varies between 200 – 300 m and it consists of four individual segments which each have a length of 1 - 2 km. The breaks in between the segments are approximately 200 – 500 m across (**Figure 5.29**). The high seismic amplitudes seen in this elongate feature are significantly higher than that of feeder systems (as shown by the red colours on **Figure 5.29**). The highest amplitudes are found on the southern edge of the arcuate feature. This southern limit has a sharp edge which is slightly cuspate. However in comparison the northern edge is more diffuse, has a slightly lower amplitude character and shows a more cuspate fringe (**Figure 5.29**).

The arcuate feature is seen to be constructional along much of its length (**Figure 5.30**). The seismic line shown in **Figure 5.30** shows a north – south transect across the high amplitude linear belt. It is located at the edge of a high amplitude reflection that is concave up and has a near horizontal base. There is a therefore a slight change in slope at the position of the elongate arcuate feature. A zoomed in seismic traverse shows the feature to have in the region of c. 25 - 30 m of constructional relief (**Figure 5.31**). Additionally, there is evidence of differential compaction over the feature shown by the thinning of moderate amplitude low frequency seismic reflection configurations seen above the constructional feature (**Figure 5.31**).

5.7.4 Interpreted Depositional Environments

From observations of the SB 3 seismic amplitude map coupled with analysis of seismic reflection configurations and stacking patterns a detailed deposition environment can be interpreted. The palaeogeographic reconstruction differs slightly from the earlier SB 2 map though in general terms the environmental setting is very similar. The reason for this is because there is only a thin preserved amount of unit 1C

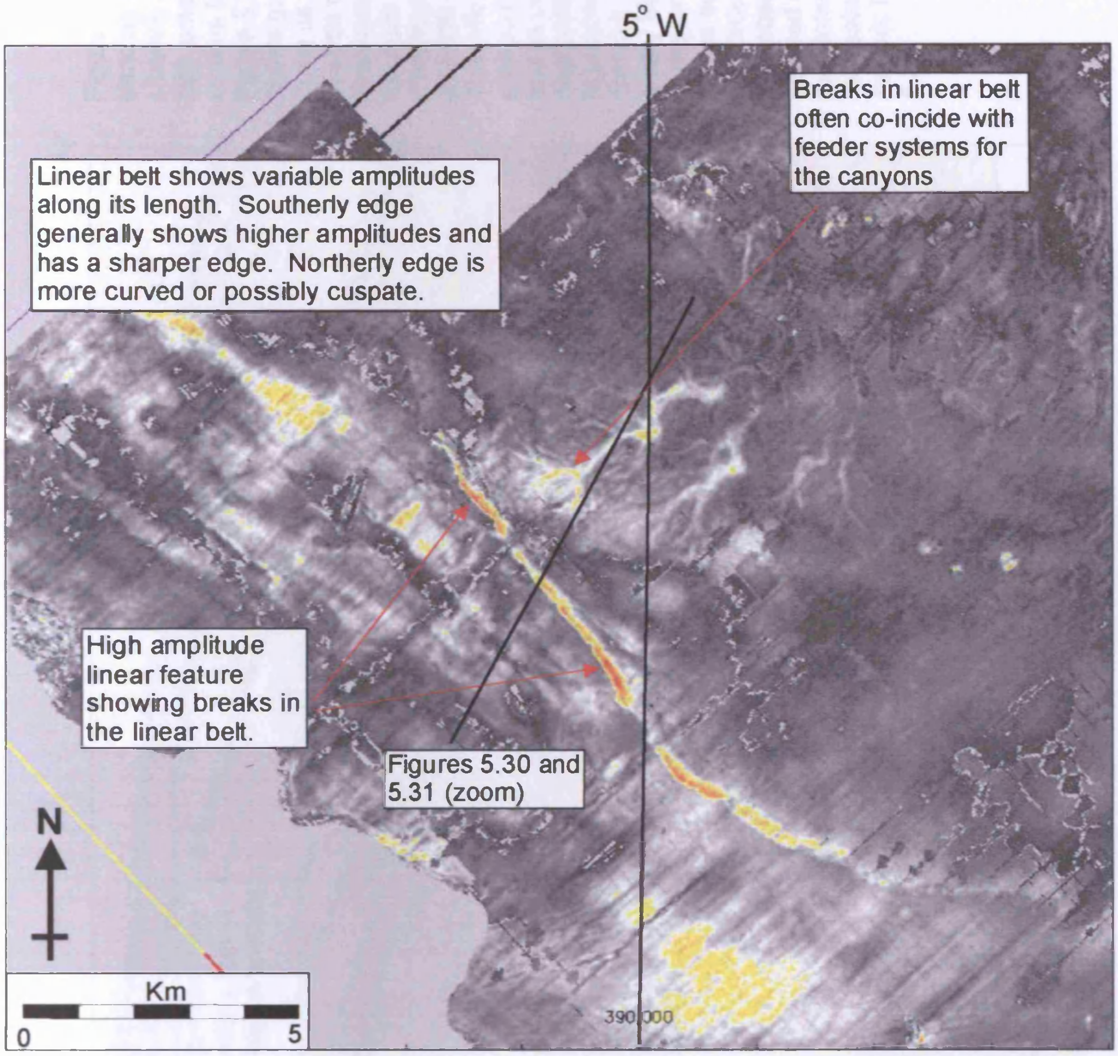


Figure 5.29. Amplitude extraction map of the SB 3 reflection. To the south of the feeder channels a long elongate very high amplitude belt is seen (red and bright yellow colours). This belt stretches for 16 km in a broadly arcuate northwest - southeast orientation. The high amplitude belt is not continuous and is seen to break in three or more places, with individual segments of the belt varying between 1.5 and 4 km long. The strength of the amplitude varies both along the length of the belt and the also axially within individual segments. Indeed, the highest amplitude values are seen to lie on the southern fringe of the feature. The northern fringe is more diffuse or cuspate in comparison to the sharper southern fringe. Towards the northwestern end of the elongate feature, there is a break in the belt which corresponds to the position of the sinuous feeder systems. The location of Figures 5.30 and 5.31 are shown.

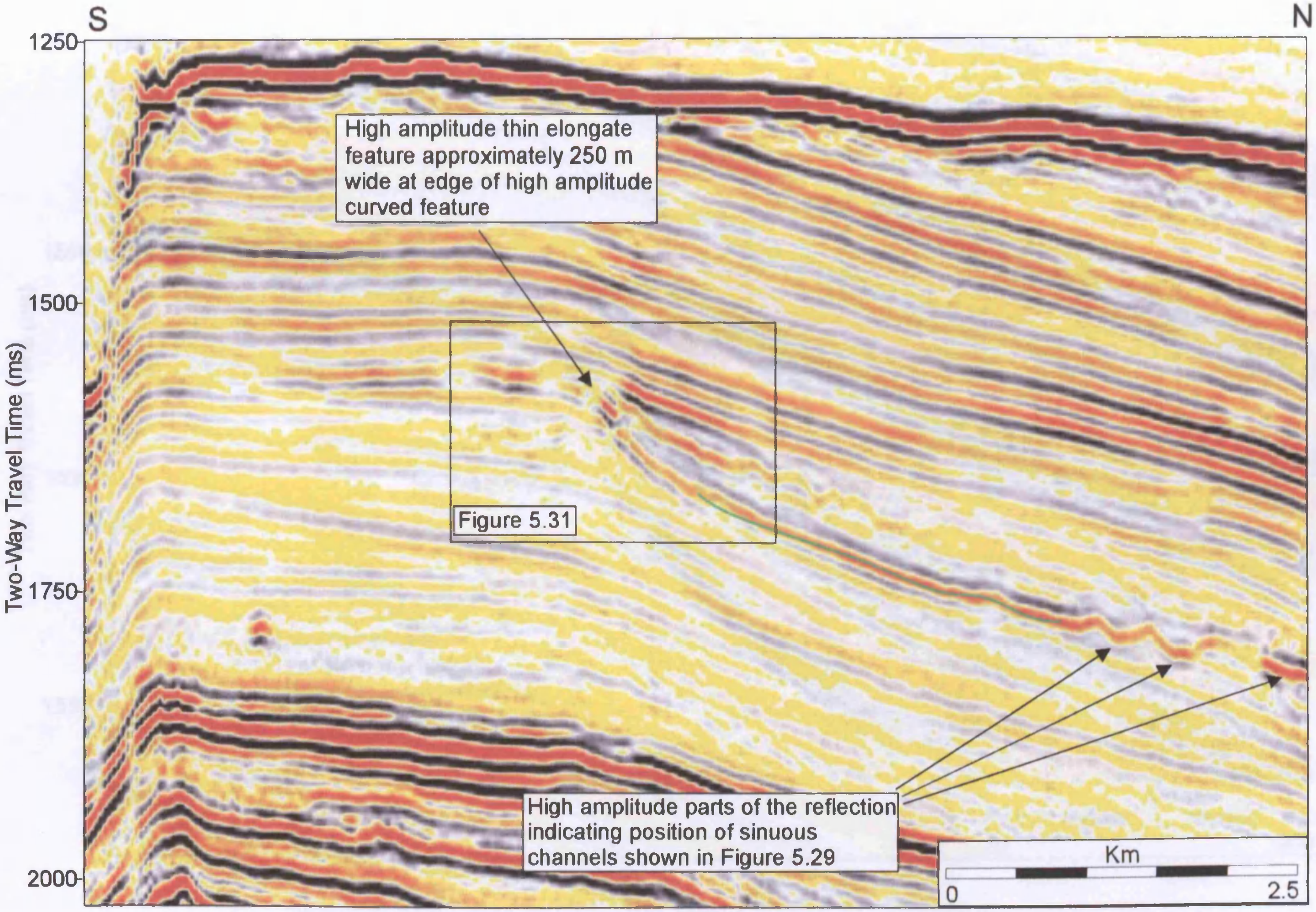


Figure 5.30. South - north trending 3-D seismic line showing location of the high amplitude elongate arcuate feature seen in Figure 5.29. The reflection shows great variation along its length where the position of sinuous channels develop. The elongate feature averages a width of approximately 250-300 m along its length. The reflection of the high amplitude linear feature and its associated sinuous channels shows a very high acoustic impedance contrast and thus produces a very high amplitude (one of the strongest amplitude values seen within much of the Eocene 1 package). This observation therefore be used as an attribute and may infer the lithology of the feature. For location of seismic line see Figure 5.29.

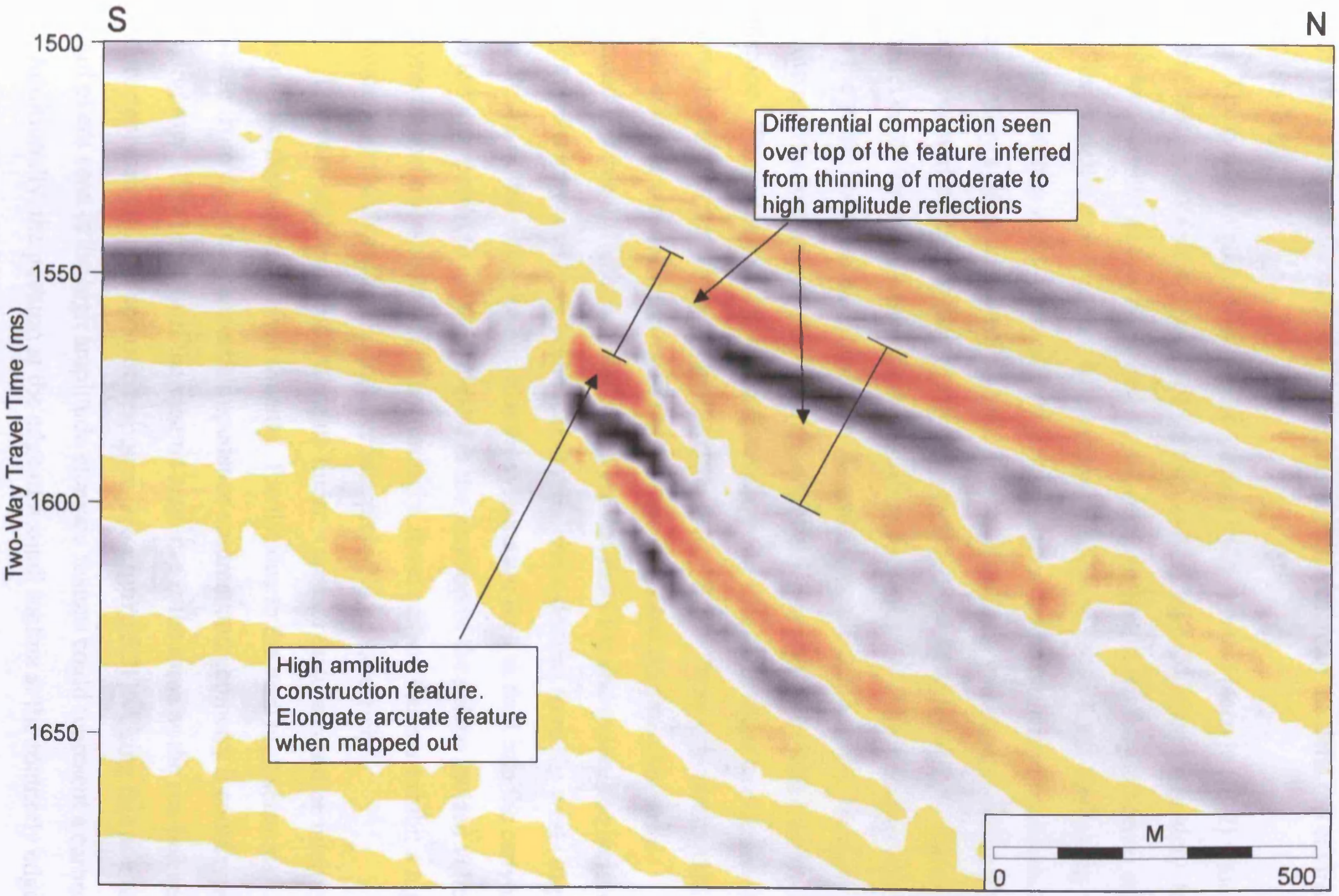


Figure 5.31. Zoomed in close up of Figure 5.31 showing south - north trending 3-D seismic line across the high amplitude arcuate feature (shown in planform in Figure 5.29). This anomalous feature appears close to a slight change in slope on the edge of a high amplitude near horizontal concave upwards reflection where sinuous channels are found. The high amplitude feature is approximately 250 - 300 m across and is seen to be constructional as moderate to high amplitude reflections are seen to thin onto it showing a degree of differential compaction. For location of seismic line see Figure 5.29.

which is believed to represent a short period of time between the development of the two sequence boundaries (see **Section 5.7.6**).

A suite of clinoforms are seen to prograde to approximately the same position as the SB 2 reflection and thus the shoreline is interpreted to have remained relatively static between these two units. Thus the delta top area in the south of the study area remained. However, the recognition of dramatic and highly variable seismic amplitude features has allowed for the local depositional architecture of the delta top environment to be detailed.

The northern edge of the broad incised valley system (seen in SB 2) which had a strong east –west arcuate trend and formed as a positive ridge approximately 80 -100 m high became cut and cannibalised during development of SB 3 by subaerial canyons or gullies which trended north – south (**Figure 5.32** and **Table 5.2**). These canyons are only found on this east – west ridge and thus suggest that the east northeast trending incised valley system may have become abandoned or changed course and the drainage network reacted by cutting their paths north and through the palaeo-high. The canyons are believed to have all formed by the same relative sea-level fall as the have they all cut to the same common base-level into the underlying SB 2 reflection. Because of their distinctive low amplitude fill the canyons are interpreted to be filled with mudstone. These canyons are thus not developed at a break in slope as is quite often the case in submarine settings (e.g. Pickering *et al.* 1989). There is evidence that channel systems trended east – west north of the ridge and thus the planforms continued to have different orientations on the flat delta top and indeed the incised valley system may well have been receiving sediment derived from the west (**Figure 5.32**). South of the ridge sinuous channel systems are seen to feed into the canyons. These are interpreted to have developed in the location of the palaeo incised valley embayment and trend northeast. Contrastingly, these canyon feeders exhibit a high amplitude seismic character and thus are interpreted to be sand-rich.

The very high amplitude elongate linear – arcuate feature could be interpreted as one of two depositional environments. Firstly, due to the very high seismic amplitude from the strong acoustic impedance contrast and geometry it could represent a fringing reef. The polarity of the seismic response is the same as the sea-bed and therefore represents a hard acoustic response (see **Figure 5.31**). Using this as a rule, the hard event seen at the high amplitude elongate feature could represent a carbonate lens. Additionally, the position at the edge of a small incline at the southerly edge of a

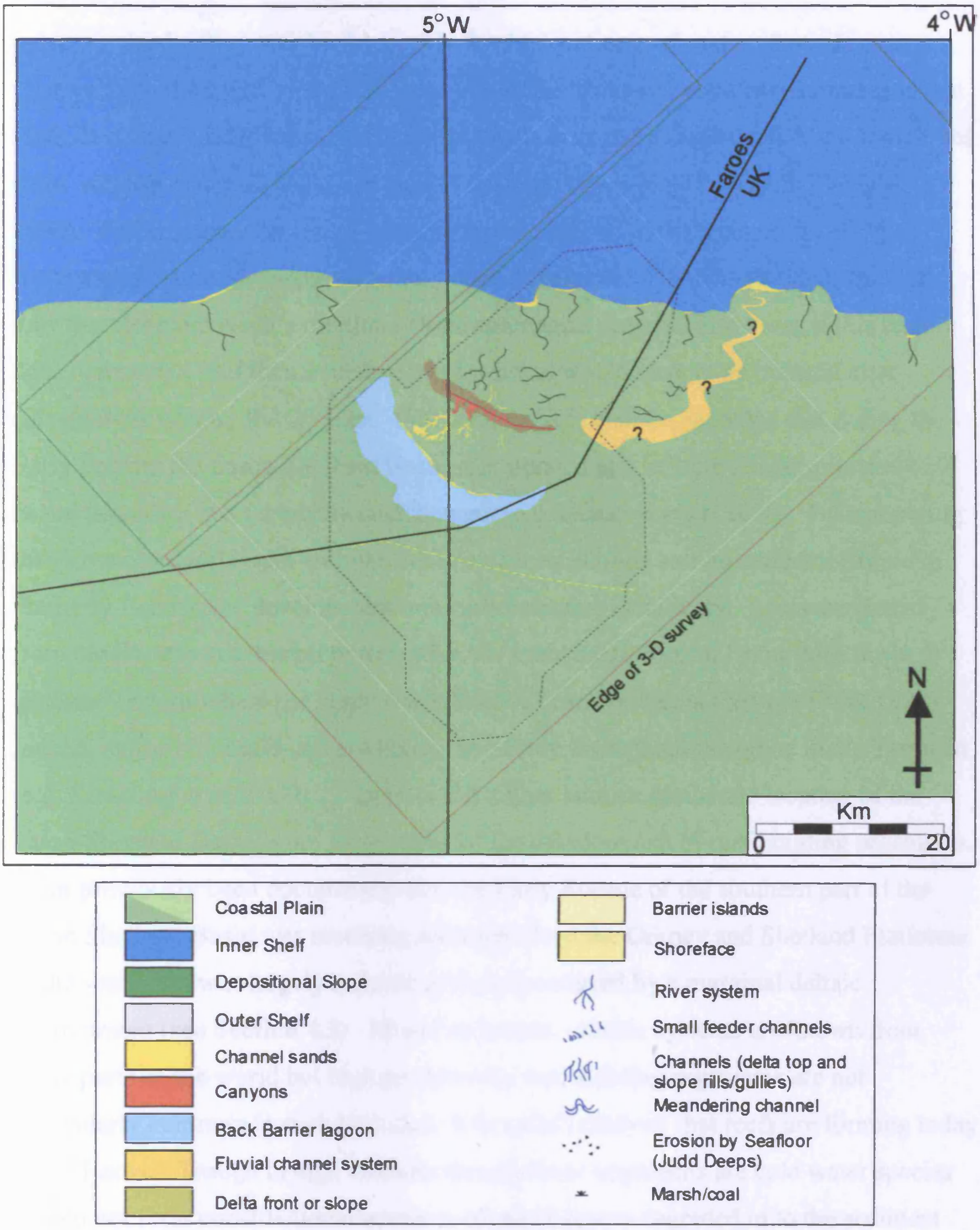


Figure 5.32. Schematic palaeogeographic map showing the environments of deposition occurring at SB 3. After incision during SB 3 a rapid transgression and subsequent fall in relative sea-level occurred and this had the effect of pushing sedimentary facies belts across the shelf with the shoreline remaining in the same place. There is very little time interval between SB 2 and 3 (indeed the SB 3 reflection sits almost directly above and truncates the SB 2 reflection in places). The northern ridge of the SB 2 incised valley is cannibalised during development of SB 3. This has the effect of producing a confined series of canyons that trend broadly east-west and are fed by sinuous feeder systems. These canyons are not therefore developed at a break in slope and suggest that drainage to the previous incised valley had ceased and channel systems cut into the previous positive relief ridge and reached the shoreline to the north. During the subsequent relative sea-level rise after incision during development of SB 3, a chain of barrier islands were formed at the previous edge of the incised valley system. This barrier island separated a back barrier lagoon environment from the coastal plain and shoreline to the north. The abandoned incised valley may have changed orientation and receive fluvial sediment from elsewhere. For legend of the colours used for the depositional environments see Table 5.2 or Enclosure P.

palaeo-incised valley may be favourable for the location reef system that sits on a relative high at a small break of slope. A common feature of such environments is that reefs lie on the border between a stable platform area and a slope profile and sometimes show varying water depths either side of the reef system (e.g. Tucker and Wright 1990). Furthermore, the feeder systems that are located at the front of the elongate feature may represent mixed carbonate-clastic channels. The cuspate northern edge may therefore represent a proximal slope apron talus deposit. However, if this feature does represent a reef then a relative sea-level rise would have to be inferred after subaerial erosion by the canyons. Furthermore, it is well documented that during the Early Eocene the Faroe-Shetland Basin was situated at a latitude of approximately 50° N and therefore in a temperate environment (e.g. Dercourt *et al.* 2000). By comparing the Faroe-Shetland Basin with the sub-tropical conditions seen in southern Britain in the Early Eocene the development of a carbonate reef is unlikely. Reefs are absent from the Eocene succession preserved in the more semi-tropical Hampshire Basin in southern Britain where the climate would have been warmer at the time (Plint 1988). Indeed, major reef build-ups are known to be rare throughout the globe in the Ypresian (e.g. Kiessling *et al.* 1999). Therefore the higher latitude temperate location of the Faroe-Shetland Basin is not favourable for the development of reef building organisms. It has previously been documented that the Early Eocene of the southern part of the Faroe-Shetland Basin was receiving sediment from the Orkney and Shetland Platforms to the south and was largely a clastic system dominated by a marginal deltaic environment (see **Section 4.3**). Mixed carbonate - clastic systems are known from many parts of the world but high productivity reef building organisms are not particularly common in such latitudes. It is noted however, that reefs are forming today in the Rockall Trough in high latitudes though these organisms are cold water species and do not form constructional barrier reefs and become degraded in to the sediment pile soon after formation (e.g. Tucker and Wright 1990). Because of the interpreted depositional environment of subaerial canyons on a delta plain and the temperate climatic setting the hypothesis of a fringing reef is rejected here.

An alternative more plausible interpretation therefore of this feature is a barrier island which consists of clastic material. The geometry and form of the high amplitude feature suggests this is a convincing interpretation and the implausibility of the reef interpretation gives greater credibility for this known clastic feature. A relative sea-level rise would once again have to be inferred to accommodate this feature though they would be expected to occur during a transgression (see **Section 5.7.5**). During transgression (and associated subsidence) delta abandonment creates delta margin islands which take the form as barrier islands produced by wave reworking on a former delta shoreline. The Chandeleur Islands in front of the modern day Mississippi River is a classic example of this (Coleman and Gagliano 1964). Many more barrier islands are known today from around the world (e.g. Galveston Island) and occur on coastlines that are influenced by a tidal, micro-tidal and wave regimes (e.g. Reading 1986). An estimation of a very shallow water depth is appropriate if indeed this feature represents a barrier island close to a coastline.

Thus an interpretation of a barrier island is preferred because of the latitude of the South Judd Basin at the beginning of the Eocene (see **Section 5.7.3.3**). A back barrier lagoon is interpreted between the barrier island and the coastline which was experiencing tidal and or wave conditions (**Figure 5.32**). However the coastline which was moving south during transgression (after SB 3) is not visible on the seismic data, unless a large back lagoon exist, though it may occur further south or is not well imaged.

5.7.5 Relative Sea-Level Interpretation

From examination of unit 1C a cycle of relative sea-level rise and fall can be documented. An initial relative sea-level rise produced onlap onto the SB 2 reflection during a period of transgression and drowning at the time of the earliest deposition of unit 1C. The onlap to the south onto the flat delta top pushed the shoreline back for a short while and sediment responded by prograding out to the north into approximately 50 -70 m of water. Therefore a relative sea-level rise of this amount is inferred during deposition of unit 1C after the subaerial erosion that created SB 2. As the delta system progrades to a similar position to that which developed at the end of unit 1B, relative sea-level begins to fall and a small forced regressive wedge is produced. By the end of unit 1C deposition relative sea-level fell by more than the 50 – 70 m causing widespread erosion of delta topsets and foresets. The barrier island is thus interpreted to form after erosion during SB 3 and occurred during the subsequent transgression during the earliest deposition of unit 1D. Thus it is suggested here that unit 1C also represents a type 1 depositional sequence of the Exxon model bounded above and above and below by type 1 unconformities (sequence boundaries) (*sensu* Vail *et al.* 1977, Mitchum and Vail 1977, Vail and Todd 1981, Van Wagoner *et al.* 1988).

5.7.6 Implications for South Judd Basin Evolution

It was suggested above (in **Section 5.7.4**) that unit 1C was deposited in a relatively short period of time because it is represented by a very thin suite of seismic reflection configurations. However, it has been demonstrated that unit 1C is bounded above and below by type 1 unconformities and thus a significant amount of strata may be missing due to erosion. No age dates are available to calibrate this time period and because the relationship of the units is unconformable, any amount of time may have elapsed between the two erosional events when SB 3 ultimately down-cut into SB 2. However, if a long period of time elapsed during deposition of unit 1C, prior to erosion by SB 3, clinoforms may be expected to prograde further north and if relative sea-level rose water depths would be expected to increase. Clinoforms are interpreted in unit 1C but these are relatively gently dipping (into shallow waters) and suggest only the upper part of the foreset and the topset ifs removed by SB 3 erosion. A period of decrease in the rate of sediment input (or starvation) and relative sea-level still-stand may facilitate this thinned but relatively complete unit 1C. Therefore it is easier to advocate a short period of time between the two unconformities.

In essence the lower bounding unconformity (SB 2) may have exposed the delta top and a long-lived drainage network may have developed prior to marine transgression at the start of unit 1C. Nonetheless, without the ability to accurately date the SB 2 and SB 3 it is impossible to put an absolute timeframe on the deposition of the 1C unit.

5.8 Seismic unit 1D

5.8.1 Introduction and Age

The fourth and final seismic unit bounded above by SB 4 is discussed here. This SB has also been identified to the west of the previous sequence boundaries on a second 3-D survey in the area of Quadrant 6005 (see **Figure 5.1**). There is evidence to suggest the previous three sequence boundaries discussed above (**Sections 5.5, 5.6** and **5.7**) may be truncated by the SB 4 reflection in this western part of the study area.

5.8.2 Seismic Reflection Interpretations

Extensive mapping of the final sequence boundary reflection has shown that it covers a significantly larger area than the previous three sequence boundaries. SB 4 can be traced some 15 km to the west of the main area of incision which is located in the central part of Quadrants 6004 and 204 (see **Figures 5.6, 5.13** and **5.20** for SB's 1, 2 and 3 respectively).

The upper bounding reflection exhibits a high amplitude seismic value it is seen to have erosional truncation at its base (**Figure 5.33**). Indeed SB 2 and 3 are eroded by SB 4 in this area (e.g. **Figure 5.25b**). The general seismic character of unit 1D is seen to be moderate amplitude continuous parallel to sub-parallel seismic reflection configurations. The reflections show a gentle dip to the north where they are seen to downlap onto the SB 3 reflection (**Figure 5.33**). In the north of the study area the high amplitude SB 4 reflection is near horizontal and exhibits a slightly curved edge on the southeastern edge where erosional truncation is seen (**Figure 5.34**). Unit 1D is a relatively thin seismic unit but shows variable thickness changes depending on how much erosional truncation of SB 4 is seen. However across the study area the thickness is generally no more than 100 ms. Small incisions (less than 100 m wide) are seen to cut the SB 4 reflection and the amplitude value decreases at these features (**Figure 5.35**). These are only found in the eastern part of the survey area.

5.8.3 Seismic Facies Description

In the same survey as the three previous sequence boundaries, the seismic amplitude map of SB 4 shows a striking high amplitude area to the north (**Figure 5.36**). This map shows a diffuse curved periphery to the high amplitude northern area which exhibits small bifurcating slightly sinuous high amplitude features that become wider to the east. The high amplitude area is completely closed off to the west and these sinuous features (in the region of 100 m across) feed two larger features (up to 250 m across) that trend almost east – west. Furthermore, these high amplitude features which broadly trend towards the east and northeast are subsequently seen to be cut by later

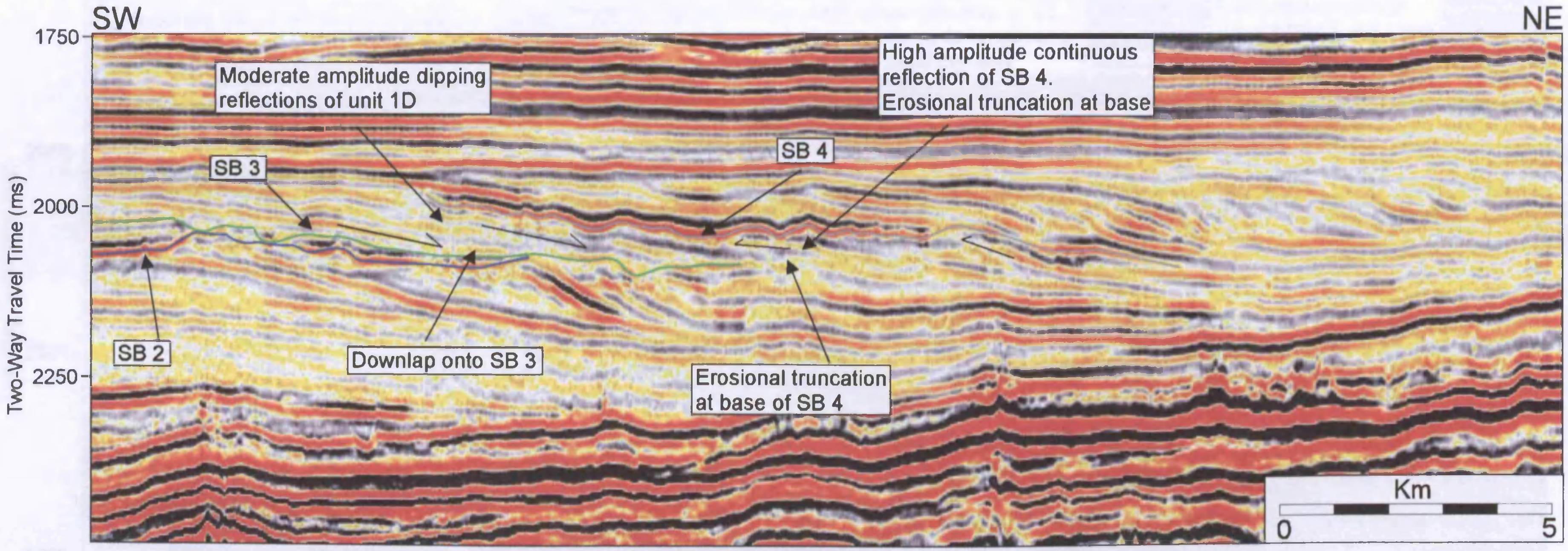


Figure 5.33. Southwest - northeast trending flattened 3-D seismic line showing internal architecture of the 1D unit. Moderate continuous reflections dip and downlap to the northwest onto the high amplitude SB 2 and 3 reflections. Erosion of these dipping reflections is seen under the high amplitude SB 4 reflection. This erosive part of the SB 4 reflection lies further north than the previous three sequence boundaries. Above the SB 4 reflection, the stacking pattern is dominated by a progradational unit. See Figure 5.36 for location of seismic line.

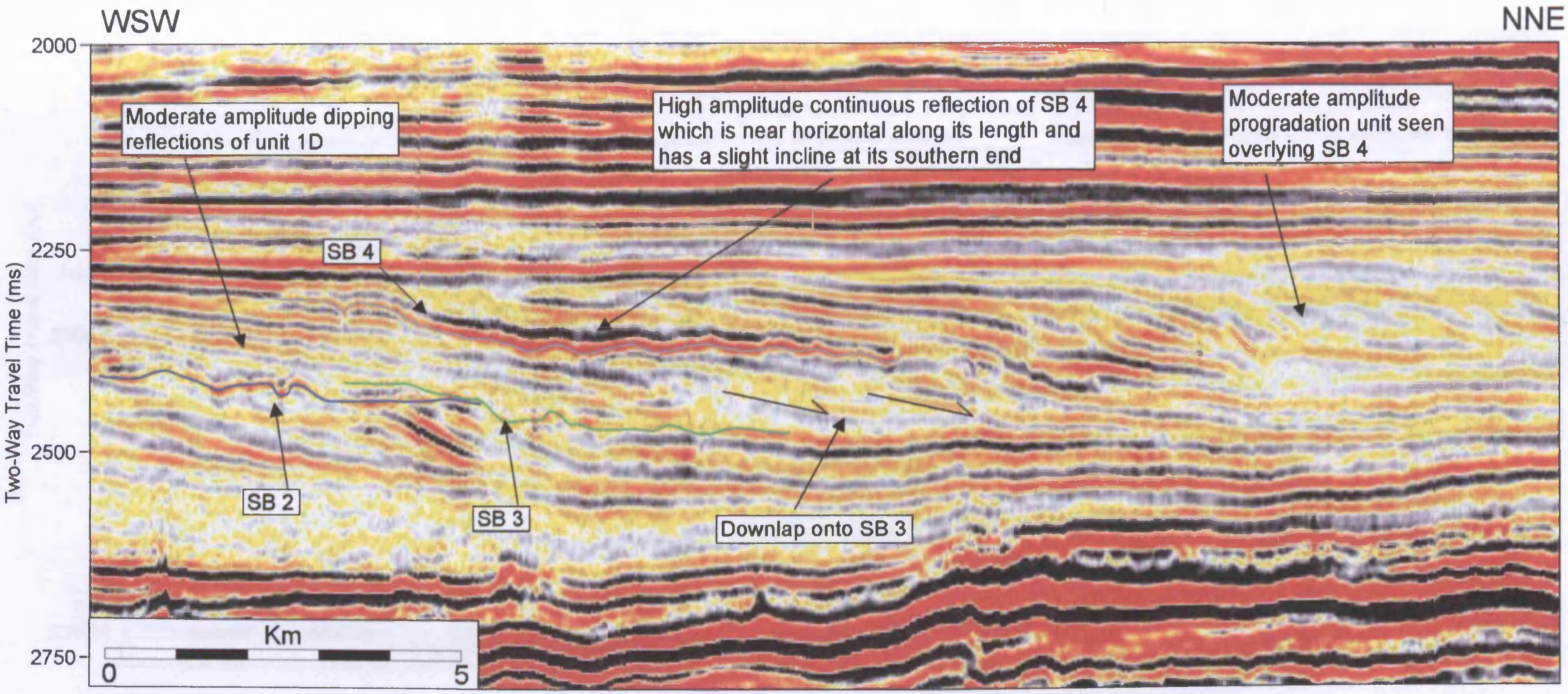


Figure 5.34. West southwest - east northeast trending flattened 3-D seismic line showing near horizontal base of the SB 4 reflection. A slight incline is seen at the southern end of the reflection and this corresponds to the edge of incision and a decrease in the seismic amplitude. The unit directly overlying the SB 4 reflection is strongly progradational, as is the eroded 1D unit. See Figure 5.36 for location of seismic line.

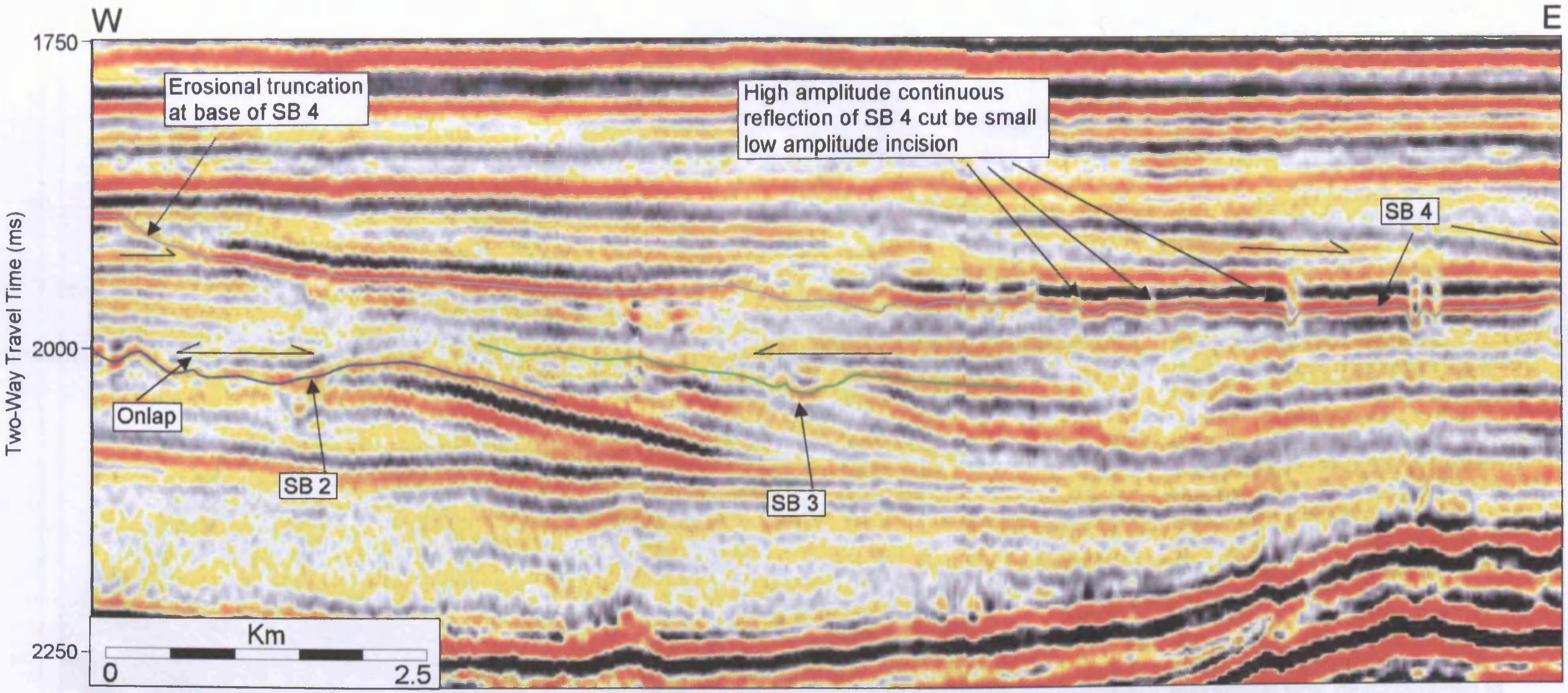


Figure 5.35. West - east trending flattened 3-D seismic line showing small low amplitude incisions in the SB 4 reflection in the east of the survey. There is evidence of progressive erosion at the base of SB 4 towards the east where approximately 80 ms of erosion is seen The internal seismic architecture of the 1D unit shows moderate amplitude continuous reflections which are parallel to sub-parallel and onlap into the lows of the previous sequence boundaries. Above the SB 4 reflection downlap is seen in overlying unit. See Figure 5.36 for location of seismic line.

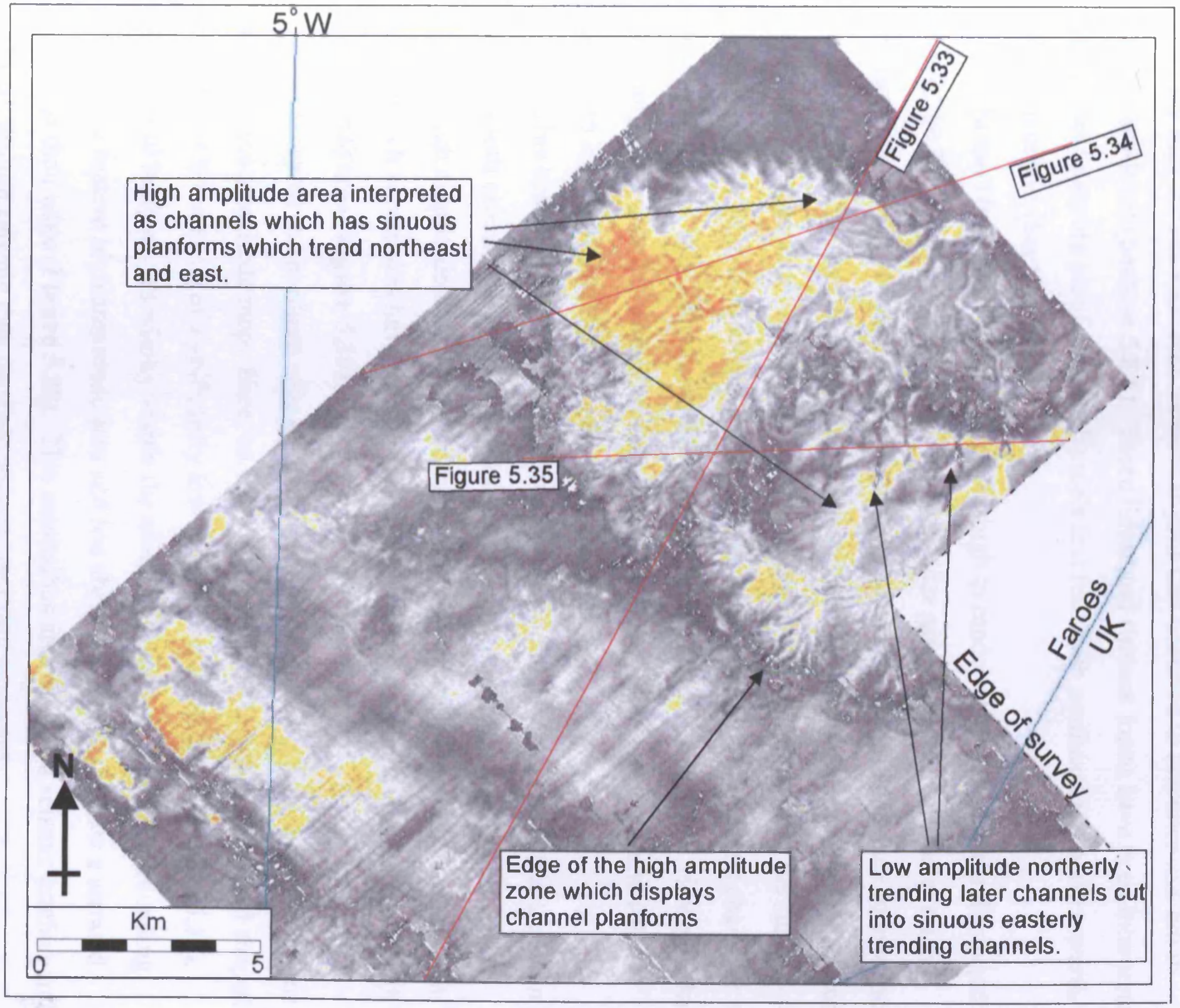


Figure 5.36. Amplitude extraction map of SB 4. The southern part of the survey area is pretty featureless apart from patches of high amplitude. However in the northern part of the survey, a high amplitude (red and yellow colours) zone can be delineated (by the green dashed line) which shows an area of sinuous channels that feed to the northeast and east. The size of these channels varies along the length., They are seen to start off small (100m wide) but become larger towards the east and northeast where they reach up to 200 -250 m and become more sinuous. A curved edge to the catchment area is seen which swings around the entirety of the channel system opening out to the east. Later low amplitude (grey colours) northerly trending channels are seen to cut down into the earlier sinuous channels. These channels are less sinuous, though they occasionally are seen to bifurcate and of much smaller dimensions and spatial limit, being confined to the eastern edge of the survey. Locations of seismic line sin Figures 5.33, 5.34 and 5.35 are shown.

low amplitude linear features (**Figure 5.36**). These low amplitude features trend obliquely to the earlier higher amplitude more sinuous forms in a north - south orientation. The dimensions of these northerly trending forms are approximately 100 m across and up to 7 or 8 km long. At their northerly limits these low amplitude features meet the distal ends of the easterly – northeasterly larger high amplitude features close to the edge of the survey (**Figure 5.36**). This may suggest that the depositionally lowest part of the drainage system was to the north and northeast at this time. A final observation of the low amplitude linear features is that they are longer in length but similar in geometry to the canyons of SB 3 (both in seismic amplitude and orientation), and they may indeed represent further canyons that occurred in response to a fall in relative sea-level (**Section 5.8.5**). These linear and sinuous forms have been interpreted to represent varying planforms of channels that feed both northeast and north towards the basin centre (**Section 5.8.4**).

Some 10 - 15 km south west of the high to moderate amplitude area discussed above, the SB 4 reflection is seen to have a similar general seismic character on the westernmost survey (**Figure 5.37**). However, in this survey, there is a lack of evidence depicting any obvious sinuous or linear channel planforms within the seismic amplitude response. A large area (over 15 km^2) in the southern and central parts of the survey shows a patchy high to moderate amplitude seismic response. The seismic character in this broad area is very diffuse and the most of the perimeter is not sharp. However the high amplitude area has abrupt arcuate southerly edges where the amplitude value is highest in a broadly circular 4 km diameter area.

This broad area is separated in two by a region of low amplitude that trends in a north – south orientation (**Figure 5.38**) which is seen to narrow to the south. The eastern high amplitude area has a random pattern of low amplitude zones (up to 2 km wide) which sometimes have a linear element which separate parts of the grossly high to moderate area (**Figure 5.38**).

Towards the northern edge of the eastern high amplitude area, is an anomalous region in the amplitude map. Here, an east – west trending lensoid or elliptical shaped body (7 – 8 km across) of significantly low seismic amplitude exists (**Figure 5.38**). The lensoid body occurs wholly within the moderate to high amplitude surrounding area of the eastern high amplitude area and has abrupt edges which show a serrated pattern at their edge (**Figure 5.38**). This anomalous area may be a seismic artefact and not real, though seismic data does not suggest anything untoward.

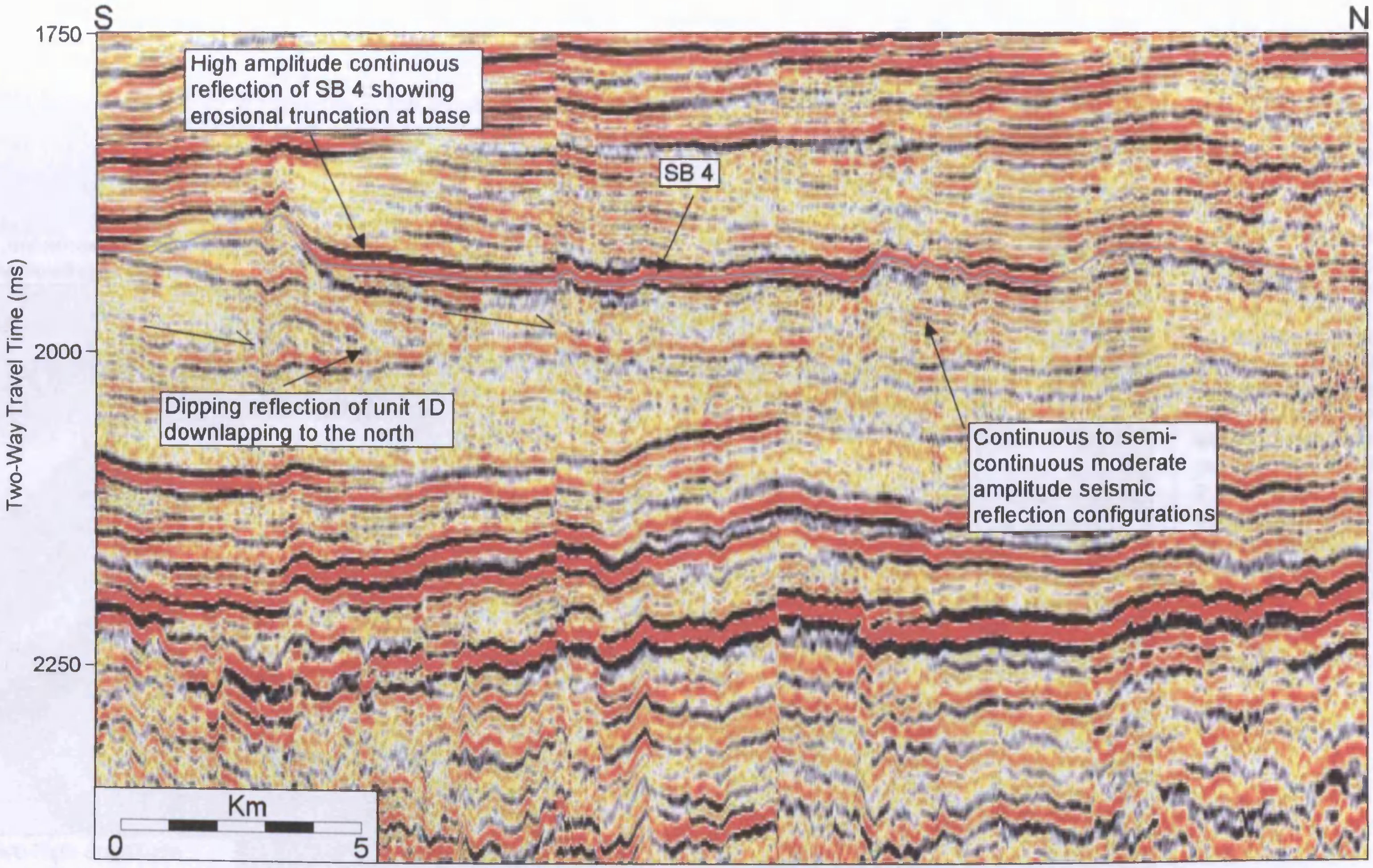


Figure 5.37. South - north trending flattened 3-D seismic line showing SB 4 in the western survey. SB 4 appears as a high amplitude continuous reflection that shows erosional truncation at the base. Downlap of internal reflections of the 1D unit towards the north is evident. See Figure 5.38 for location of seismic line.

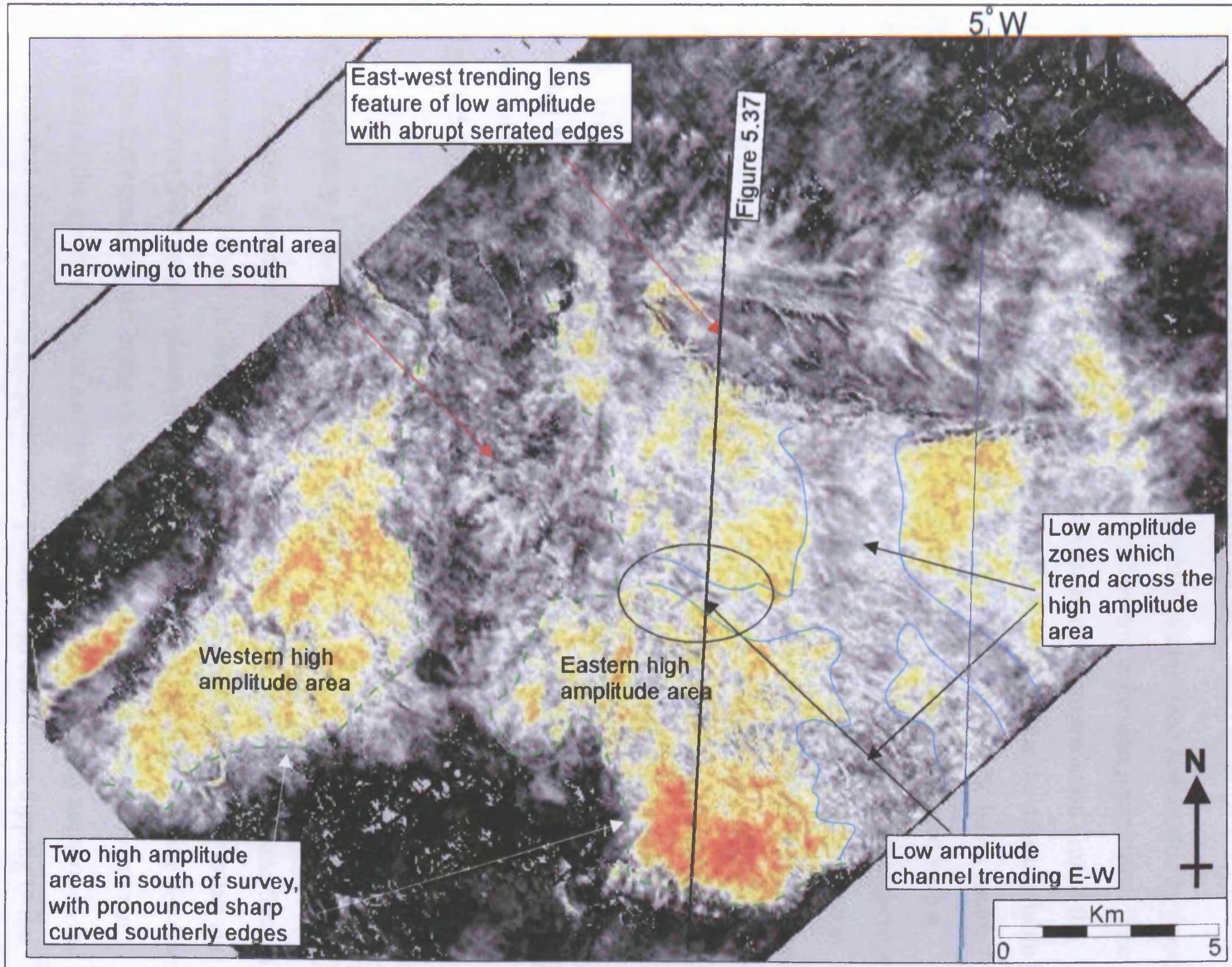


Figure 5.38. Amplitude extraction map of SB 4 in the survey to the west of the three previous sequence boundaries. The map shows two southerly regions of high amplitude (bright colours) in the south and central parts of the survey. These high amplitude areas show a curved southerly edge and an increase in amplitude to the south. Delineating these two regions is a north-south orientated low amplitude region in the centre. Towards the north of the survey there is an anomalous area which trends east-west and is in the region of 8-10 km long and 2.5 km wide. This area seems be a lens or body and has sharp serrated edges to it. The location of the seismic line in Figure 5.37 is highlighted.

Subtle features on the seismic amplitude map may infer some channel systems to have developed across the patchy high amplitude area. These are highlighted on **Figure 5.38** and have generally low amplitude values and various trends. One low amplitude candidate channel system is seen in the centre of the eastern high amplitude area. The low amplitude broad zones of low amplitude may too represent some form of channel system (see **Section 5.8.4**).

5.8.4 Interpreted Depositional Environments

An interpretation of a delta top setting which was incised and eroded by sinuous channels has been documented by the seismic facies and reflection geometries. The delta top area extended towards the north and east as delta progradation draped the previous sequence boundaries (**Figure 5.39** and **Table 5.2**). Thus the area available for the development of accommodation space migrated north towards the basin centre. The small low amplitude north trending incisions are seen to be associated with a slightly later incision which cuts down into the SB 4 reflection but infers that there was changing channel planforms active at approximately the same time on the delta top. In the western part of the study area the channel free area has been interpreted as a large interfluve area which may explain its high to moderate amplitude. The slight evidence of channels seen in this area trend east – west and is interpreted to represent small feeder channels that ultimately link to the main axes of channels in the east (**Figure 5.39**). The elliptical lens shaped feature remains enigmatic and may represent an erosional feature though this is purely speculation. The main depositional features seen across the two surveys in the SB 4 reflection are summarised in a dip attribute map (**Figure 5.40**).

5.8.5 Relative Sea-Level Interpretation

Relative sea-level has been interpreted to rise by more than 100 m and then fall by approximately 50 - 80 m during deposition of unit 1D. Initial filling by onlap and draping of the SB 3 (and in places SB 2) reflection occurred during a relative sea-level rise. The transgressive part of the unit is not seen as it is very thin and beyond the resolution of the seismic data, as no retrogradational stacking patterns are seen in the unit. The unit shows a strong progradational pattern with clinoforms dipping to the

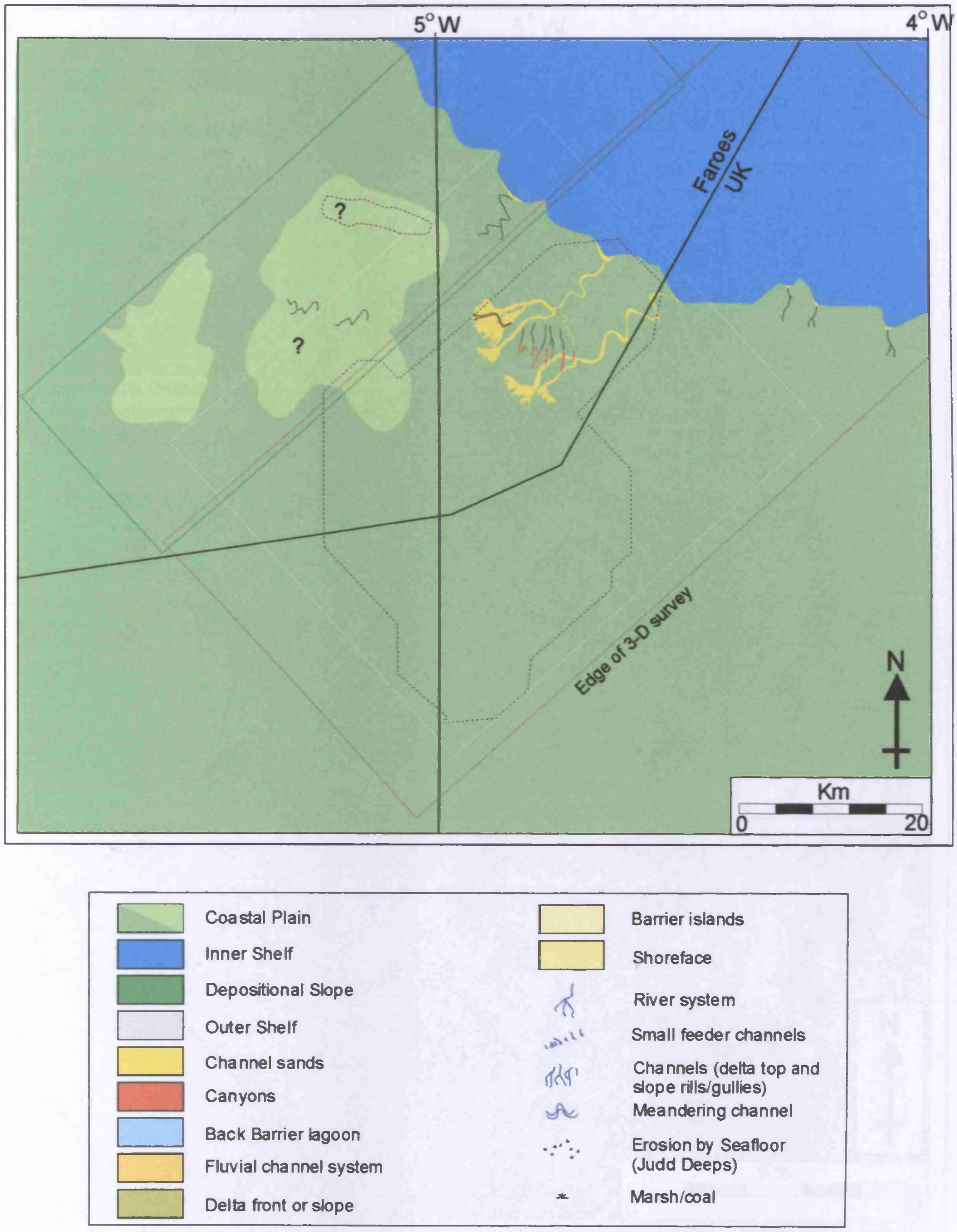


Figure 5.39. Schematic palaeogeographic map showing the depositional environments interpreted at the end of unit 1D deposition (during development of SB 4). In the eastern surveys sinuous channels of high amplitude are seen to trend towards the northeast and east. These high amplitude channels are seen to cut into the foresets and topsets of clinoforms in the underlying 1D unit. These sinuous channels are later cut by smaller canyons with little or no sinuosity which trend north and have a low amplitude fill. In the western survey, little or no evidence is seen for channel systems and a coaly delta top is interpreted. An enigmatic lens shaped feature is seen ton the north of this area and may represent an erosional feature. For legend of the colours used for the depositional environments see Table 5.2 and Enclosure P.

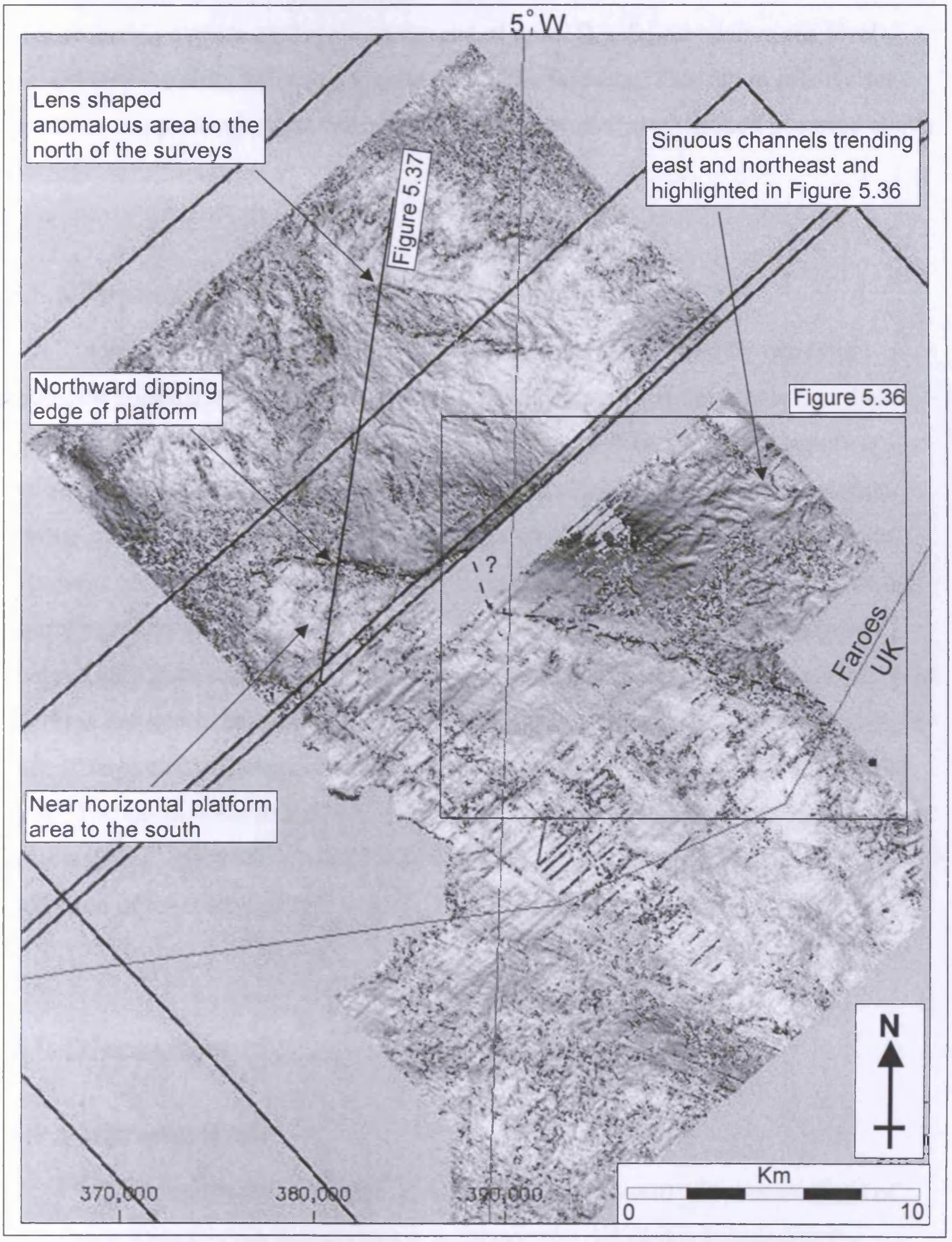


Figure 5.40. Dip attribute map summarising the main features of the SB 4 reflection. The dip attribute takes the pick of the seismic reflection and computes the change in angle between two neighbouring traces. If the angle of dip is large it assigns a dark colour (in this case, dark greys - blacks). If the angle of dip is low then it assigns a light colour to the pick. This attribute s good for picking out sharp lineaments and edges to particular bodies, such as faults, channels etc. The location of the close up of the sinuous channels in the eastern survey (Figure 3.36) is shown as well as the location of the seismic line shown in Figure 5.37.

north and northeast. This progradation occurred during the continued relative sea-level high or during a static sea-level. At the end of unit 1D, a fall in relative sea-level incised into the delta topsets and upper part of the foresets. This fall in relative sea-level was not prolonged and flooding back and extensive progradation occurred above the SB 4 reflection.

5.8.6 Implications for South Judd Basin Evolution

Unit 1D represents another depositional sequence bounded by two type 1 sequence boundaries. It is clear from the detailed mapping of this whole unit that a cyclicity of relative sea-level rise followed by fall is evident. Sediment supply is believed to have kept pace with periods of relative sea-level highs as progradation further into the basin is seen. The relative sea-level highs have been interpreted to represent periods of subsidence on the basin margin leading to a transgression of the shoreline. However this general subsidence of the basin which is believed to be thermal subsidence after cessation of the major Mesozoic rifting event (e.g. Dean *et al.* 1999) is thought to have been episodically punctuated by periods of relative sea-level fall. This cyclicity cannot be dated because of the lack of any good well data in the area. The cause of these falls are either eustatic falls in sea-level or an impact of local tectonics (e.g. uplift and compression of the margin or reactivation and tectonic inversion of local structures).

5.9 Discussion

5.9.1 Introduction

The internal seismic reflection configurations coupled with observations of planforms from seismic facies analysis has enabled a detailed palaeogeographic evolution of the South Judd Basin to be developed. This section will go on to discuss some of the possible causes of the observed patterns and cyclicity seen in the Eocene 1 seismic unit.

5.9.2 Eocene 1 Sub-Division

Four high amplitude reflections have been mapped extensively on high quality 3-D seismic data throughout the study area. These surfaces show various levels of erosion at their base and are thus termed sequence boundaries. These sequence boundaries form the upper and lower bounding surface to the sub-units 1A – 1D. The erosional extent of each sequence boundary is localised, with each individual sequence boundary displaying a basal erosion surface that cuts into the stratigraphy of the underlying unit.

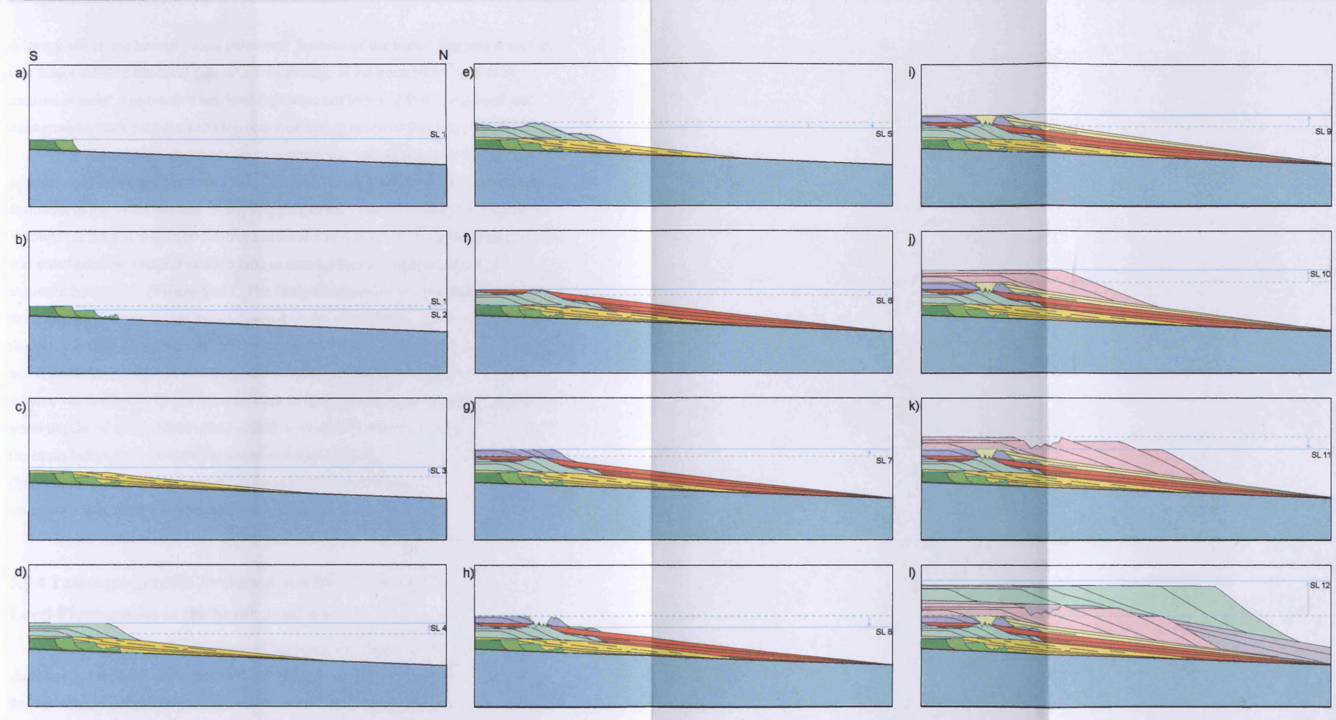
Examination of the four seismic amplitude extraction maps of the individual sequence boundaries highlights a great variety in channel planforms. The four surfaces represent a palaeogeographic "snapshot" during periods of incision during the Early Eocene. The whole Eocene 1 unit is dominated by a prograding clinoform package that exhibits subtle variations in the stacking pattern of the seismic reflection configuration. The strongly progradational succession is thought to have been punctuated by episodic periods of transgression. Erosional truncation of delta topsets and foresets is interpreted to have occurred during the periods of relative sea-level fall.

Indeed this erosion of the deltaic succession indicates that the South Judd Basin was at or close to sea-level during the Early Eocene (Ypresian). Towards the central and northern parts of the basin, the depositional environment changed to a shallow marine setting.

5.9.3 Relative Sea-Level Interpretation

The four seismic units discussed above (in **Sections 5.5 – 5.8**) record cyclicity in relative sea-level which can be documented by the recognition of progradational stacking patterns that are separated by erosional unconformities (**Figure 5.41**). The prograding systems dominantly occurred during periods of relative sea-level rise when the shoreline is seen to migrate northwards. An additional observation is the overall clockwise rotation of the shoreline that is seen through time from unit 1A - 1D (**Figure 5.41**).

In between these periods of progradation, forced regressive wedges are interpreted to have developed and the shoreline down-steps basinward during relative sea-level falls. Landward of these forced regressive wedges are areas of incision on the



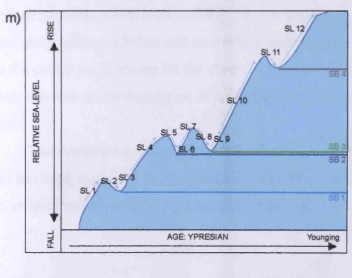


Figure 5.41. Schematic south - north cross sections through the South Judd Basin showing the evolution of the observed stratal geometries seen on 3-D seismic data. The twelve cartoons (a - l) show the interpreted fluctuations in relative sea-level (RSL) throughout the Early Eocene (Ypresian). Early clinoforms prograde towards the north (a) and SB 1 develops at a fall in RSL during a forced regression (b). A RSL rise during a transgression then follows (c) and covers the clinoforms before progradation further north continues (d). A further fall in RSL causing erosion into delta topsets occurs (again creating a forced regressive wedge (e) creating an incised valley system (SB 2) that feeds north. A further rise in RSL pushes facies belts back to the south when the delta top is re-flooded (f). Progradational stacking patterns continue above the SB 2 and downlap towards the north (g). Canyons of SB 3 develop during a further RSL fall and cut down into the SB 2 (h) and subsequent re-flooding over the delta top is seen (i). A final progradation of clinoforms occurs during renewed RSL rise (j) which pushes the shore-face further north past the SB 2 and SB 3 shorelines. A final RSL fall occurs (SB 4) north of the previous three (k), before RSL rise continues after SB 4 creation. A relative sea-level curve for this cyclicity of fluctuations is shown in cartoon m). However, because of poor biostratigraphic data the error bars on these sequence boundaries are very large.

delta top which are laterally quite extensive. Incision of the underlying unit is seen in four major relative sea-level falls where an average of between 50 and 100 m of incision is seen. The relative sea-level falls were not believed to be long-lived and transgression back over the delta top occurred during renewed flooding (**Figure 5.41**).

It is argued here, that the basin as a whole was undergoing post-rift thermal subsidence (Turner and Scrutton 1993, Clift and Turner 1998) with extension being dominant in the Mesozoic and finally stopping in the Palaeocene (Dean *et al.* 1999). Throughout the Early Eocene relative sea-level was rising, but this general relative rise was punctuated by isolated relative falls as seen by the development of the four sequence boundaries (**Figure 5.41**). The likely transgressive events expected to follow the sequence boundaries are not recognised on the seismic data, and this suggests that they are too thin to be resolved. However, progradation followed the transgressions and clinoforms are seen to downlap onto the previous sequence boundary. Therefore relative sea-level rises led to flooding back of the delta plain during subsidence and water depths of up to 100 m were reached, with clinoforms prograding further out into the basin before the next relative sea-level fall and incision. A schematic chronostratigraphic chart summarises these observations both spatially and temporal of changing sea-level and a schematic curve is shown (**Figure 5.42**).

5.9.4 Palaeogeographic Evolution and Mechanisms for Relative Sea-Level Fluctuations in the South Judd Basin

The depositional environment of the South Judd Basin and the movement of the shoreline towards the north and northeast throughout the Ypresian have been documented in the palaeogeographic reconstructions for each individual sub-unit (1A – 1D) (see **Sections 5.5.4, 5.6.4, 5.7.4 & 5.8.4** respectively). A brief summary of the palaeogeographic evolution will now follow and then continue with an investigation into the causal mechanisms for the observed fluctuations in relative sea-level and the effects this has on the translation of facies belts as well as the available accommodation space.

The development of sequence boundaries and the northwards migration of the delta has been explained by rises and falls in relative sea-level (**Section 5.9.3**). The depositional environments of the margin are variable through time depending on

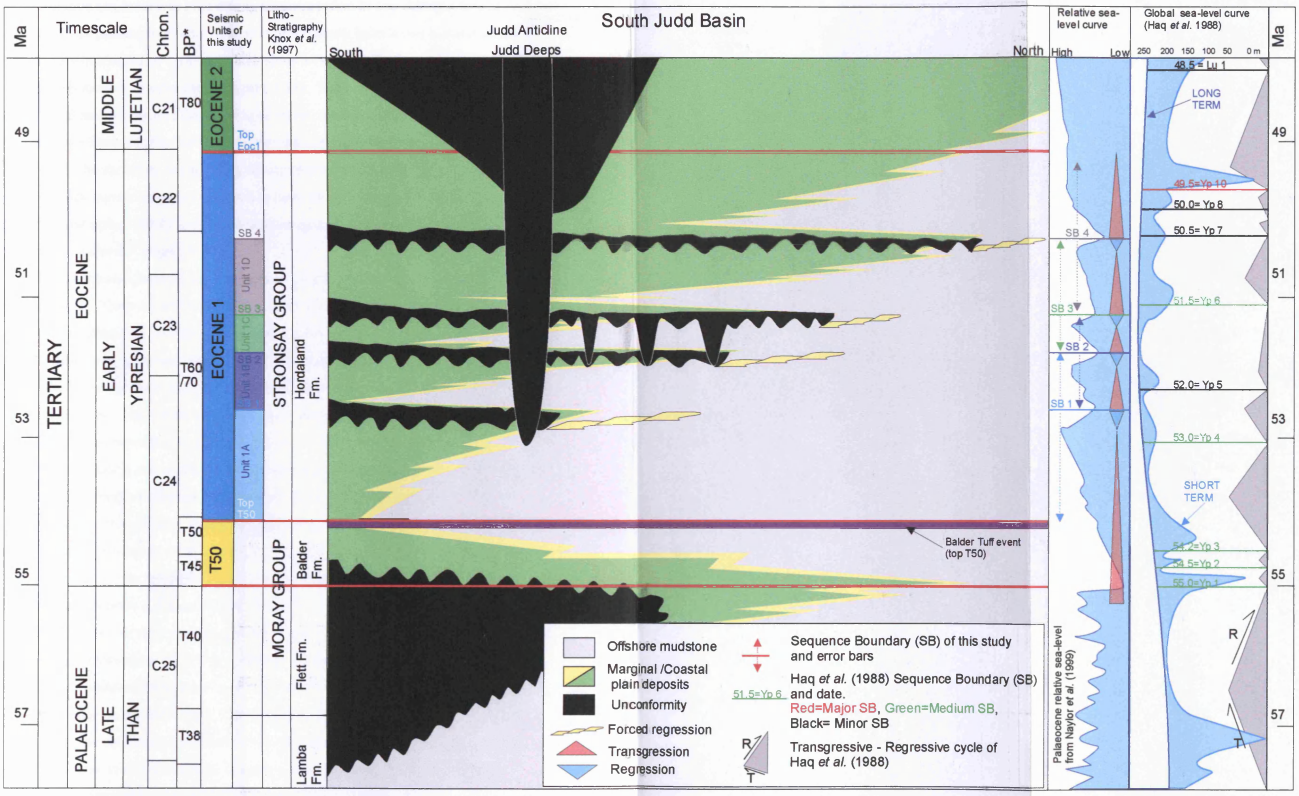


Figure 4.42. Chronostratigraphic chart showing the temporal and spatial distribution of the Eocene 1 seismic unit in the South Judd Basin. A cyclicity of transgressions and regressions occur creating forced regressive wedges in a distal position that developed at the same time as unconformities (sequence boundaries) in the delta top area. Note that SB 3 down-cuts into the earlier SB 2 and that later unconformities (post Eocene 1 deposition occur in the area of the Judd Anticline and the Judd Deeps. The sequence boundaries are correlated to the eustatic curve of Haq *et al.* (1987) who recognised ten sequence boundaries in the Ypresian. Because of the poor quality of biostratigraphic data the four sequence boundaries recognised in this study cannot be dated accurately and thus not possible to calibrate to the eustatic curve. Timescale after Berggren *et al.* (1995).

whether relative sea-level was high during a transgression or low during a forced regression. The palaeogeographic evolution of the South Judd Basin is summarised schematically in a series of cartoons which show a "snapshot" of the margin at the time of sequence boundary development (**Figure 5.43**). The northerly build out of the shoreline and facies belts and the location of rivers, incised valleys and canyons are all shown through time (**Figure 5.43**). However marine transgressions occur between each SB with the shoreline retrograding during relative sea-level high. The barrier island and back barrier lagoon is believed to have formed during one such transgression (after SB 3 and before SB 4) and a further palaeogeographic map of this depositional environment is shown (**Figure 5.44**).

Causal mechanisms for the fluctuations in relative sea-level in the South Judd Basin during the Ypresian will now be explored. Firstly, the movement of shorelines and clockwise rotation of shorelines and facies belts towards the north and northeast could be influenced by a compressive phase that was responsible for the folding and uplift of the Munkagrunnur Ridge to the west or Wyville-Thompson Ridge to the southwest. If these ridges experienced an Early Eocene phase of compression causing associated vertical movements, a local effect would be to push facies belts and coastlines to the north and northeast. Indeed, on a much larger scale, the Lower Eocene sediments of the regional Eocene 1 seismic unit discussed in **Chapter 4** move eastwards and north eastwards up the Shetland margin through time. By the time of deposition of the Eocene 2 seismic unit (Early to Middle Eocene) the main depocentre which was centred over the South Judd Basin had shifted a minimum of 50 km to the northeast (see **Sections 4.3.2** and **4.3.3**). Therefore changing patterns of sediment dispersal is interpreted both on large and small scale (from km's to tens of km's) on the south and eastern margins of the Faroe-Shetland Basin.

This postulated folding and uplift on the Munkagrunnur Ridge, Wyville-Thompson Ridge or the Judd High/Anticline complex is known from onlap and thinning onto the structures (see **Figure 5.2**). If compression was starting to occur in the Early Eocene its affect would be to cause a relative lowering of the sea-level. Interestingly, compression has been documented on the Wyville-Thompson Ridge and dated as Late Palaeocene (e.g. Boldreel and Andersen 1993). However, if subsidence was occurring elsewhere in the South Judd Basin (away from the axial hinges of the Judd Anticline and Munkagrunnur Ridge and Wyville-Thompson Ridge) and the compression was therefore an extremely local effect, then a relative sea-level rise

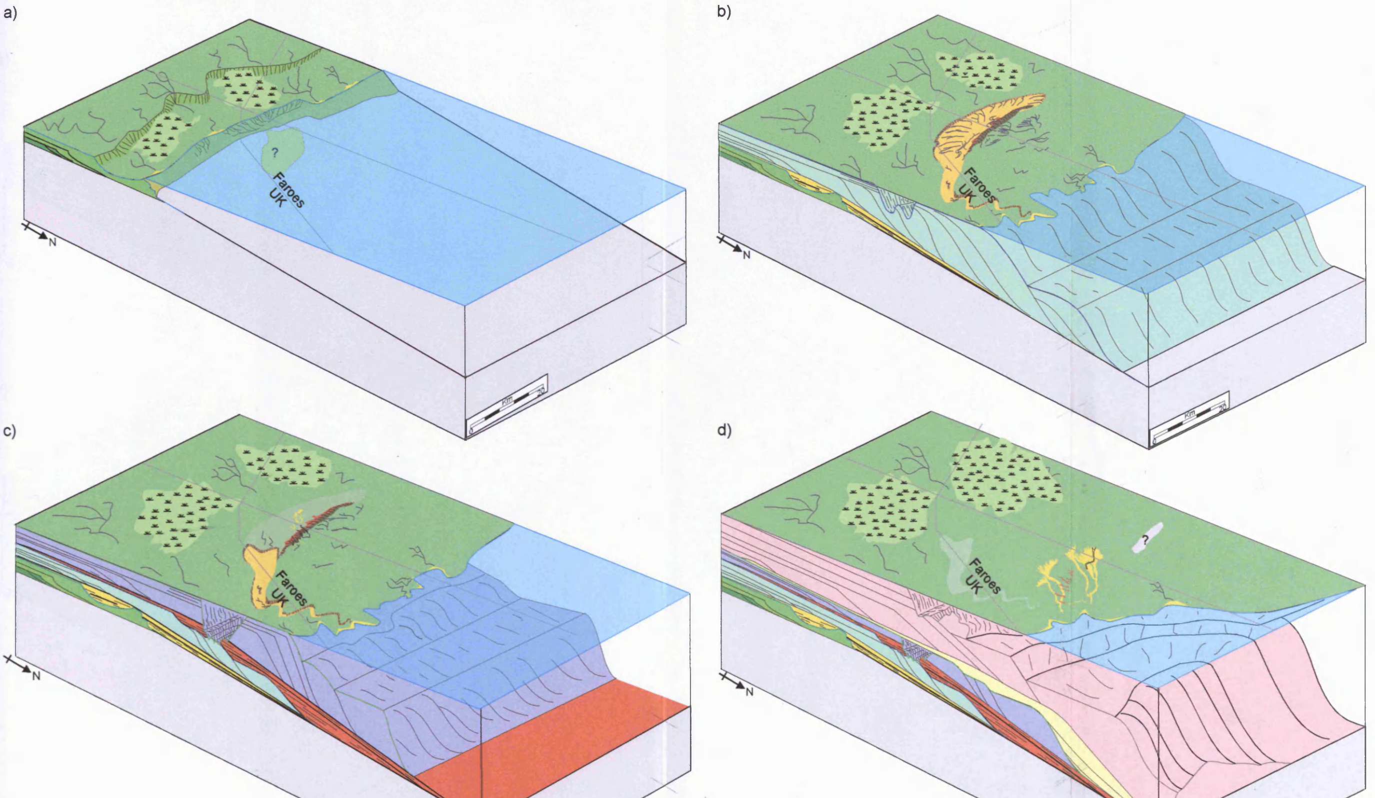
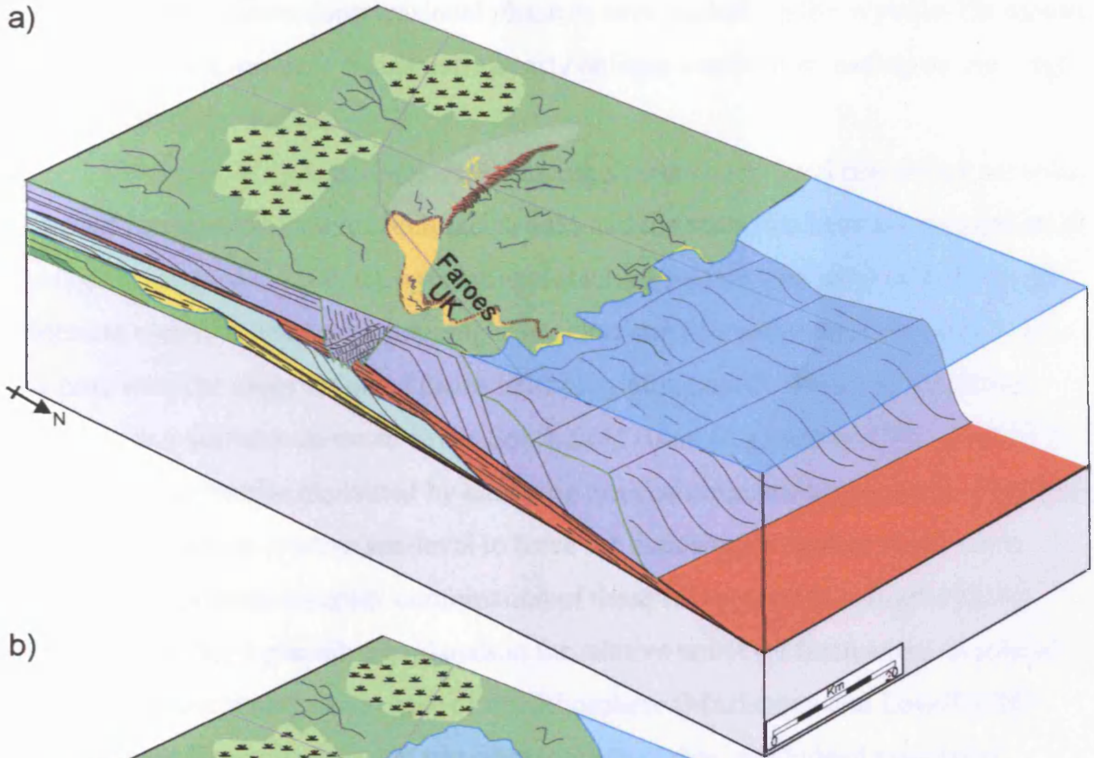


Figure 5.43. Schematic palaeogeographic maps showing the environments of deposition at the times of sequence boundary development throughout the Ypresian (SB 1-4 represented by a-d). An overall northward migration of the shoreline occurs throughout the Ypresian, with the development of forced regressive wedges towards the north at the time of sequence boundary development on the delta top. In between these relative sea-level falls the shoreline is pushed back south during re-flooding of the delta top.

a)



b)

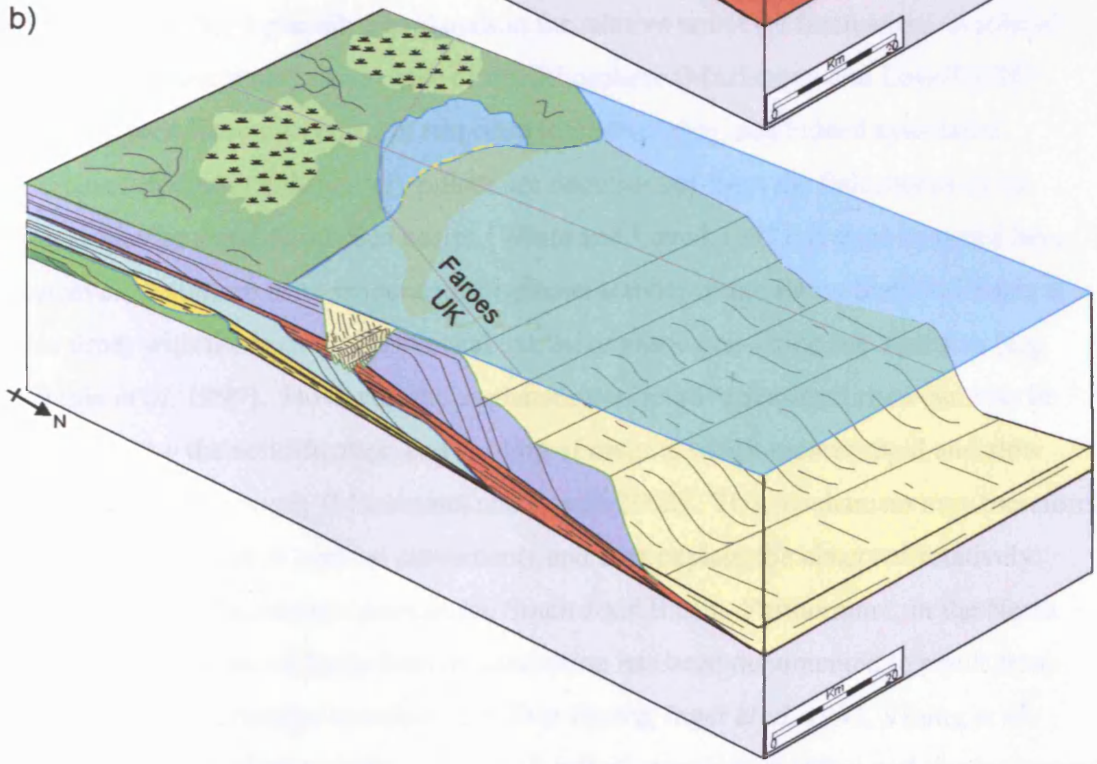


Figure 5.44. Schematic palaeogeographic cartoons at time of SB 3 incision. a) shows the environments of deposition that were present during the RSL fall (creating SB 3), the diagram is the same as the 5.43c and creates canyons in the east - west ridge that are fed by dendritic channels. However, after the RSL fall (creating SB 3) a subsequent RSL rise (b) flooded the delta-top and palaeo-incised valley. On the southern fringe of this incised valley a series of elongate barrier islands develop during this RSL rise. Behind these barrier islands a back barrier lagoon is interpreted. Further south the delta top is found.

would be occurring elsewhere at the same time. Indeed, Boldreel and Andersen (1993) suggest a north – south compressional phase to have pushed up the Wyville-Thompson Ridge and Munkagrunnur Ridge during early oblique sea-floor spreading on the Aegir Ridge north of the Faroes.

However, if the basin was experiencing a relative sea-level rise during the time, it would be expected to force both facies belts and the shoreline back towards the south during a transgression (i.e. retrogradational stacking pattern) reducing or shutting off sediment supply to the basin, trapping it in deltas and decreasing erosion. This is not the case with the observation of facies belts prograding north. However, candidate transgressive surfaces do occur in the South Judd Basin (e.g. between SB 1 and SB 2). This may therefore be explained by changing rates of sediment supply to the delta front or by subtle falls in relative sea-level to force the depositional system to the north. Furthermore, a more complex combination of these two processes is highly likely.

Secondly, a plausible explanation for relative sea-level fluctuations is episodic magmatic underplating at the base of the lithosphere (Maclennan and Lovell 2002). Uplift is well known to occur in response to underplating and indeed associated increases in clastic sedimentary pulses are documented from the Palaeocene in the Faroe-Shetland and North Sea basins (White and Lovell 1997). It must be noted here however, that there is no evidence for igneous activity in the Faroe-Shetland Basin at this time, with the major intrusive and extrusive phase pre-dating the Ypresian (e.g. Ritchie *et al.* 1999). However, the explanation of relative sea-level rises can also be explained by the solidification and cooling of magma which creates rapid and slow subsidence respectively (Maclennan and Lovell 2002). This mechanism may therefore account for the subtle vertical movements and thus explain the observed relatively rapid sea-level fluctuations seen in the South Judd Basin. Furthermore, in the North Sea, the Palaeocene to Early Eocene succession has been documented to result from rises and falls in relative sea-level (e.g. Den Hartog Jager *et al.* 1993, Vining *et al.* 1993, Jones and Milton 1994). An overall fall of greater than 800 m and rise by greater than 500 m throughout this interval is inferred by Jones and Milton (1994).

A third and final control on these sea-level fluctuations is changes in global sea-level. These possible sudden rises and falls in relative sea-level may be accounted for by small variations (less than 100 m rise or fall) in eustasy. No glacial control on eustasy is possible as no major ice sheets were present on Earth during the Palaeocene and Eocene. Furthermore, no available accurate dating is available for the Ypresian succession in the Faroe-Shetland Basin (see **Figure 5.42**). The global sea-level curve of Haq *et al.* (1988) exhibits one major sequence boundary in the Ypresian interval, five medium sequence boundaries and five minor sequence boundaries **Figure 5.42**). Thus a total of eleven sequence boundaries are documented in the Ypresian succession from the global sea-level curve. The sequence boundaries detailed in this study are believed to represent type 1 sequence boundaries (*sensu* Van Wagoner *et al.* 1988) and thus may be expected to be major sequence boundaries on the global sea-level curve. However, because no dating is available, it is impossible to correlate the sequence boundaries analysed in this study with any of the sequence boundaries seen on the global sea-level curve (**Figure 5.42**). Furthermore, if this was possible, because of the vast number of possible sequence boundary events seen on the global sea-level curve then correlation will almost always succeed (Miall 1992). This is because the interval of time between each individual sequence boundary event on the global sea-level curve is less than the potential error bar expected when dating these events with the best available chronostratigraphic and geochronologic techniques (Miall 1992). Thus stratigraphic correlation to the global sea-level curve is almost inevitable with present day dating techniques and thus the curve remains of questionable value (Miall 1992). The sequence boundaries documented in this study are therefore interpreted to be localised within the context of the South Judd Basin and do not have any global (eustatic) significance.

However, there is no doubting the fact that there appears to be some cyclicity in the sequence boundaries caused by relative sea-level variations in the South Judd Basin and this phenomenon will be discussed briefly here.

The lack of absolute age data does not allow for exact dates on the succession and in particular the sequence boundaries to be identified. Thus correlation of the sequence boundaries to surfaces in other basins around the world remains impossible. This study has dated the seismic unit as Ypresian in age from nearby wells (see **Section 5.3.3**) and boreholes in the south of Quadrant 204 (see **Section 4.3.2.4**). This date agrees with Smallwood (2004) who suggests the same age for this succession though no reference to a source is highlighted. A crude observation shows that the sequence boundaries are relatively evenly spaced in time within the Eocene 1 seismic unit and this might allude to some sort of depositional cyclicity which may be controlling their temporal position within the succession. However, this hypothesis is unproven and has not been calibrated with any high resolution biostratigraphic data and therefore remains highly speculative. If the continued progression northwards of the shorelines and the sequence boundary incision surfaces are interpreted as being cyclical events it may imply a pulsed or episodic control on the compressional activity or underplating. Authors have argued throughout the 1990's on the possibly pulsed nature of the Iceland plume and its affect on uplift and in turn on sedimentation throughout the Palaeocene (e.g. White and Lovell 1997, Clarke 2002 unpublished PhD. thesis). A possible cause of episodic compression and associated uplift may well be intrinsically linked to the start of ridge push tectonics to the northwest of the Faroe Islands where sea-floor spreading had been initiated. Furthermore, periods of ridge shifts may well have a role to play in defining episodes of changing regional compression. Smallwood and White (2002) and Smallwood *et al.* (1999) has produced initial results on small lateral ridge jumps which create characteristic "V" shaped ridges of magnetic anomalies which occur during the initial few million years of sea-floor spreading. The timing of these ridge jumps need to be accurately calibrated and compared with good dates of the sequence boundaries seen in the South Judd Basin. Until then a link between sequence boundaries and pulsed compression remains extremely tenuous.

5.9.5 Depositional Environment Interpreted from Seismic Facies Analysis

Analysis of the seismic amplitude coupled with the knowledge of the geometry of the bodies and the depositional setting can provide evidence for the lithology of certain features seen on amplitude maps. Interpreting depositional features and systems using high quality 3-D seismic data is a relatively new arm of seismic interpretation. Termed seismic geomorphology; this science has been borne out by the significant advances in 3-D seismic technology (e.g. Posamentier 2002, 2004).

The best example of the using 3-D seismic to understand geomorphic features from this study is the occurrence of the high amplitude linear feature seen in the SB 3 amplitude extraction. This feature could be interpreted as a reef or a barrier island from its planform, though the northerly latitudes of the Faroe-Shetland Basin during the Early Eocene favour a barrier island interpretation. Therefore a barrier island is the preferred interpretation exhibiting a constructional geometry and is believed to occur at the edge of a palaeo-incised valley where a slight break in slope occurred.

It is well documented from both ancient and modern day examples that barrier islands are made up of clastic material that is often relatively coarse grained or even pebbly. The presence of a barrier island may indicate a highstand in relative sea-level due to the enhanced preservation potential at this point on the relative sea-level curve. (e.g. Reading 1986). Low amplitude continuous reflections that make up many of the clinoforms are herein interpreted to be dominantly claystone. Following this logic, the high amplitudes within this seismic character are likely to be either sandstones, siltstones or carbonates. It is therefore fair to state that the barrier island is made up of a sandy or pebbly facies and that this corresponds with the high amplitude seen in the seismic data. If this is the case, the surrounding reflections, which are very continuous and low to moderate in amplitude, may be inferred to probably represent a siltstone or claystone lithology. Therefore the principal deltaic conditions during the deposition of the Eocene 1seismic unit are interpreted to be dominated by low energy mudstones and siltstones that were deposited on a broad shelf (see **Section 4.3.2**). Lithological data from the nearest BGS borehole (99/6) also suggests the Early Eocene is dominated by mudstone at this time (see **Figure 4.18** and **Sections 4.3.2.3** and **4.3.2.4**). However, sandstones are found to the south in wells 204//22-1 and 204/23-1 (see **Figure 5.1** for location of wells). If the low to moderate amplitude of the interpreted clinoforms suggests mud-dominated clays and silt-grade deltaic conditions, then it may imply that this delta was initially prograding and then experienced a relative sea-level falls creating periods of incision which are associated with the deposition of higher amplitude sandstones in the sinuous channels.

Furthermore, the canyons north of the sinuous channels are of significantly low seismic amplitude which suggests that they are also filled with claystone or siltstone. Therefore the canyons are interpreted to fill during a transgressive episode after the initial fall which created the incision. Indeed, Plint (1988) suggests a two-stage transgressive phase after a relative sea-level fall. This comprises an early stage of estuarine deposits which develop during the initial rise into the incised lows, followed by a marine mud-dominated phase which marks the classic onset of marine transgression.

A depositional history of an initial lowering of relative sea-level creating incision and associated channel and canyon development is documented. This is followed by a transgressive episode allowing for deposition of mudstones to back fill the canyons, pushing facies belts back south and creating a barrier island at a slight change in gradient during a period of high sea-level. Further to this, is the possible affect of waves and tides on the near-shore and coastal depositional style. Barrier islands are known to be common in an environment that is influenced by both waves and tides, and if, as believed, the Lower Eocene delta succession prograded out into a large basin and onto a broad shelf, there would be significant scope for the impingement of tidal processes. Ebb and tidal deltas may be expected to form close to the barrier islands in this depositional setting although none were identified on the 3-D data available. These have not been recognised in the seismic data either in front or behind the barrier island.

5.10 Conclusions

- In the South Judd Basin the Eocene 1 seismic unit can be generally seen to consist of moderate to high amplitude parallel – sub-parallel seismic reflection configurations that form a largely progradational package. The seismic unit has been sub-divided into four sub-units based on the recognition of high amplitude, continuous, often erosional reflections, termed sequence boundaries.
- The use of high quality 3-D seismic data allows for the recognition and detailed mapping of the sequence boundaries. The four sequence boundaries show differing patterns of drainage and incision which occurred during their creation as relative sea-level fell exposing the delta top. Seismic amplitude extraction maps have been used to build a palaeogeographic evolution for the South Judd Basin. The sequence boundaries are believed to represent sequence boundaries of Mitchum and Vail (1977) and Mitchum *et al.* 1977a. Seismic facies analysis coupled with planform geometries have identified sinuous and meandering channels, incised valleys, barrier islands and canyons that are all interpreted to be located on the delta top.
- The sequence boundaries are interpreted to represent type 1 unconformities (developed during significant falls in relative sea-level which expose the shelf). Thus the seismic units are believed to represent the classical depositional sequences of Exxon.
- Seismic unit 1A (bounded at the top by SB 1) shows truncated clinoforms that are seen to have a complex sigmoid-oblique or tangential oblique configuration. Water depths were interpreted to be low (in the region of 50 – 100 m)

from looking at the depth between the clinoform break points and the bottom-set positions. Clinoforms that post-date SB 1 are more sigmoidal and prograde into greater water depths through the Eocene 1 seismic unit. Subsequent sequence boundaries (SB 2 – SB 4) are found further basinward and similarly appear to show down-cutting into the underlying unit.

- Lithological information can be interpreted from seismic facies analysis and the barrier island probably consists of coarse grained sandstones that form at the height of transgression in wave and tidal dominated settings. The barrier island develops north of the coastline and a back barrier lagoon sits behind it sheltered from the main sea or ocean.
- Evidence of prograding and down-stepping clinoforms throughout the succession implies periods of relative sea-level fall during development of a forced regressive wedge. This is followed by periods of relative sea-level rise causing a transgression when the previous sequence boundary is flooded and renewed progradation during reduced rates of relative sea-level rise occurs.
- Cycles of relative sea-level have been interpreted from the succession, with observed variations in the region of 50 – 100 m. These fluctuations are interpreted to be caused by possible local uplift (relative sea-level fall) on an overall thermally subsiding basin margin (relative sea-level rise).
- Successive shorelines are seen to migrate to the north and rotate in a clockwise direction. This is interpreted to be caused by a). A response to compressional activity on the Munkagrunnur Ridge to the west, Wyville-Thompson Ridge to the southwest or early subtle folding on the nearby Judd High/Anticline complex. b). or by rapid uplift and subsidence by intrusion and cooling of gabbroic material at the base of the lithosphere (underplating).
- The Ypresian sedimentary succession shows a degree of cyclicity recorded in the development of four seismic units bounded by sequence boundaries. Accurate age dating of these sequence boundaries is not achievable and thus correlation with the global sea-level curve is not possible. However pulsed compressional activity remains a possibility, though this remains unproven.

6. Chapter Six: Structural Controls on Channel Development on the Flett Ridge

6.1 Introduction

This chapter will show how 3-D seismic data can be utilised to describe the detailed stratigraphic evolution of a small part of the Faroe-Shetland Basin throughout Palaeogene times. It will detail the second of the two case studies which have used high quality small 3-D seismic datasets to study the depositional systems in great detail (see **Section 5.1** for brief summary of resolution). This case study looks at the role of tectonics on the deposition of the sedimentary succession and documents the stratigraphic and structural evolution of an intra-basinal area which is located close to the Shetland margin on the southeastern side of the basin. A series of seismic units have been mapped and described throughout the Palaeogene (see **Section 6.4**) to give a more detailed description of the Flett Ridge area. For a concise summary of this chapter see Robinson *et al.* (2004).

6.2 Regional Geological Setting

The study area discussed in this chapter is located close to the Shetland margin in the northern part of Quadrant 205 (**Figure 6.1**). This area is on the middle of the present day continental slope, in water depths of approximately 500 m and situated in the area of a Mesozoic structural high; the Flett Ridge (e.g. Dean *et al.* 1999 and **Section 2.2.3**). This ridge trends northeast – southwest and forms a common lineament along the entire Northeast Atlantic margin. The position of the Flett Ridge is of some significance because it forms in an intra-basinal setting northwest of the major bounding fault systems of the Shetland Spine Fault and the Rona Ridge (**Figure 6.1**).

The Flett Ridge structure is an intrabasinal structural high and forms the southeasterly lineament which separates the Foula sub-basin from the more northwesterly Flett sub-basin (**Figures 6.1** and **6.2**). The Foula sub-basin shows evidence of extension during the Early Cretaceous and possibly earlier (Lamers and Carmichael 1999). However

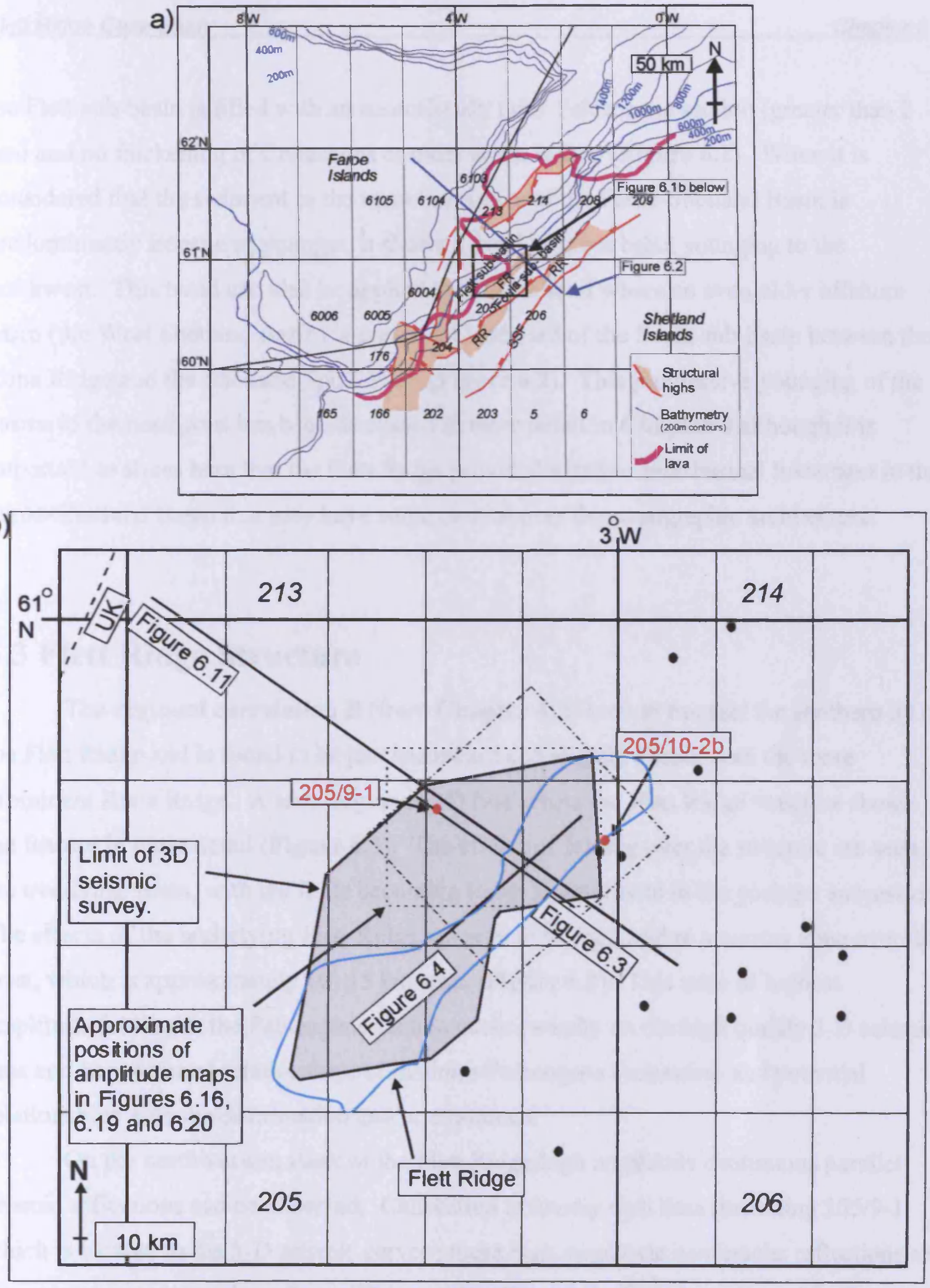


Figure 6.1. a) Major Mesozoic structural lineaments that trend northeast - southwest in the Faroe-Shetland Basin. Note the position of the Flett Ridge in an intra-basinal area, northwest of the main bounding faults. SSF = Shetland Spine Fault, RR = Rona Ridge, FR = Flett Ridge. The Flett Ridge separates the Foula sub-basin to the southeast from the Flett sub-basin to the northwest. The locations of Figures 6.1b and 6.2 are shown. b). Detailed location map of the Flett Ridge area. The study area is located in the northern part of Quadrant 205 and sits over the Mesozoic Flett Ridge structure (blue line). Highlighted on this map are the positions of selected seismic panels used in this chapter (Figures 6.3, 6.4 and 6.11) as well as the approximate positions of time structure and amplitude maps also shown (Figures, 6.16, 6.19 and 6.20). The study area is also covered in a coarse grid of 2-D surveys with spacings between the lines of approximately 1 km as well as a large 3-D survey (grey area). The red dots indicate the two type wells used for sub-dividing and calibrating the age of the Palaeogene succession in this chapter.

the Flett sub-basin is filled with an anomalously thick Palaeocene section (greater than 2 km) and no thickening of Cretaceous or older strata is seen (**Figure 6.2**). When it is considered that the sediment in the main basin axes of the Faroe-Shetland Basin is predominantly Eocene or younger, it shows a trend of gross basin younging to the northwest. This trend can also be applied to the southeast where an even older offshore basin (the West Shetland Basin) is preserved landward of the Foula sub-basin between the Rona Ridge and the Shetland Spine Fault (**Figure 6.2**). This progressive younging of the basins to the northwest has been discussed in more detail in **Chapter 4** although it is important to stress here that the Flett Ridge provided a major intra-basinal lineament in the Faroe-Shetland Basin that may have some control over the stratigraphic architecture.

6.3 Flett Ridge Structure

The **regional correlation B** (from **Chapter 4**) is seen to transect the southern tip of the Flett Ridge and is found to be just basinward and slightly deeper than the more prominent Rona Ridge. A semi-regional 3-D line across the Flett Ridge structure shows the feature in more detail (**Figure 6.3**). The effects of folding over the structure are seen in the overlying strata, with the folds becoming lower in amplitude in the younger succession. The effects of the underlying Flett Ridge are seen to be localised to a narrow zone over its crest, which is approximately 10 -15 km wide (**Figure 6.3**). This zone of highest amplitude folding in the Palaeogene section occurs wholly on the high quality 3-D seismic data and hence stratal relationships of the intra-Palaeogene succession and potential relationships with the deformation can be examined.

On the northwestern flank of the Flett Ridge high amplitude continuous parallel seismic reflections can be observed. Calibration to nearby well data (including 205/9-1 which is located in the 3-D seismic survey) these high amplitude continuous reflections are interpreted as volcanic lava flows (**Figure 6.3**). These lava flows have been dated as belonging to the Upper Series of the Faroe Plateau Lava Group and are dated at approximately 55 Ma (Ritchie *et al.* 1999).

By comparing the 2-D data (from **regional correlation B**) with the higher quality 3-D seismic survey a number of additional features can be seen. Firstly the 3-D data has

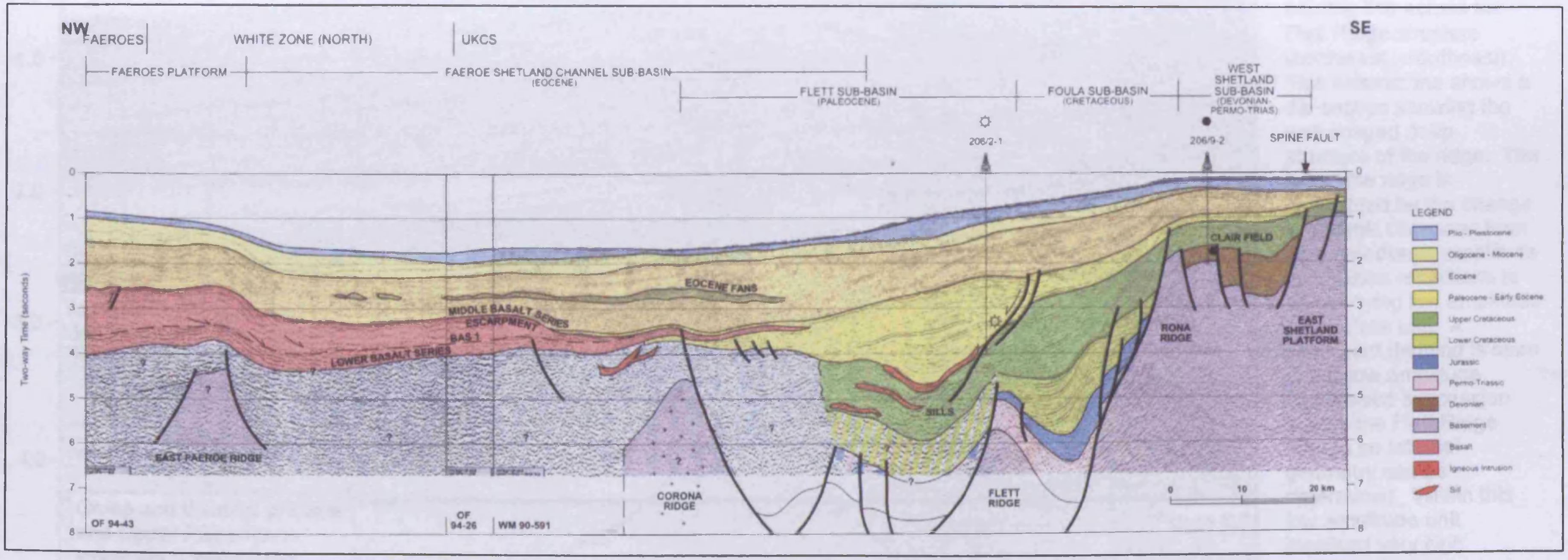


Figure 6.2. Regional 2-D geo-seismic line showing the position of the Flett Ridge within a basin-wide context. Note that the Flett Ridge occurs in an intra-basinal setting with the large Rona Ridge found landward towards the southeast. The Flett sub-basin is found to the northwest of the Flett Ridge and shows a thickened (up to 2 km) Palaeocene package compared to the Foula sub-basin which sits to the southeast of the ridge (and shows a thickened syn-rift Lower Cretaceous section of approximately 1.5 km. The Flett Ridge is believed to have a Permo-Triassic and Jurassic cover above the basement. Igneous sills intrude into the Cretaceous and Lower Palaeocene section in the Flett sub-basin. From Lamers and Carmichael 1999.

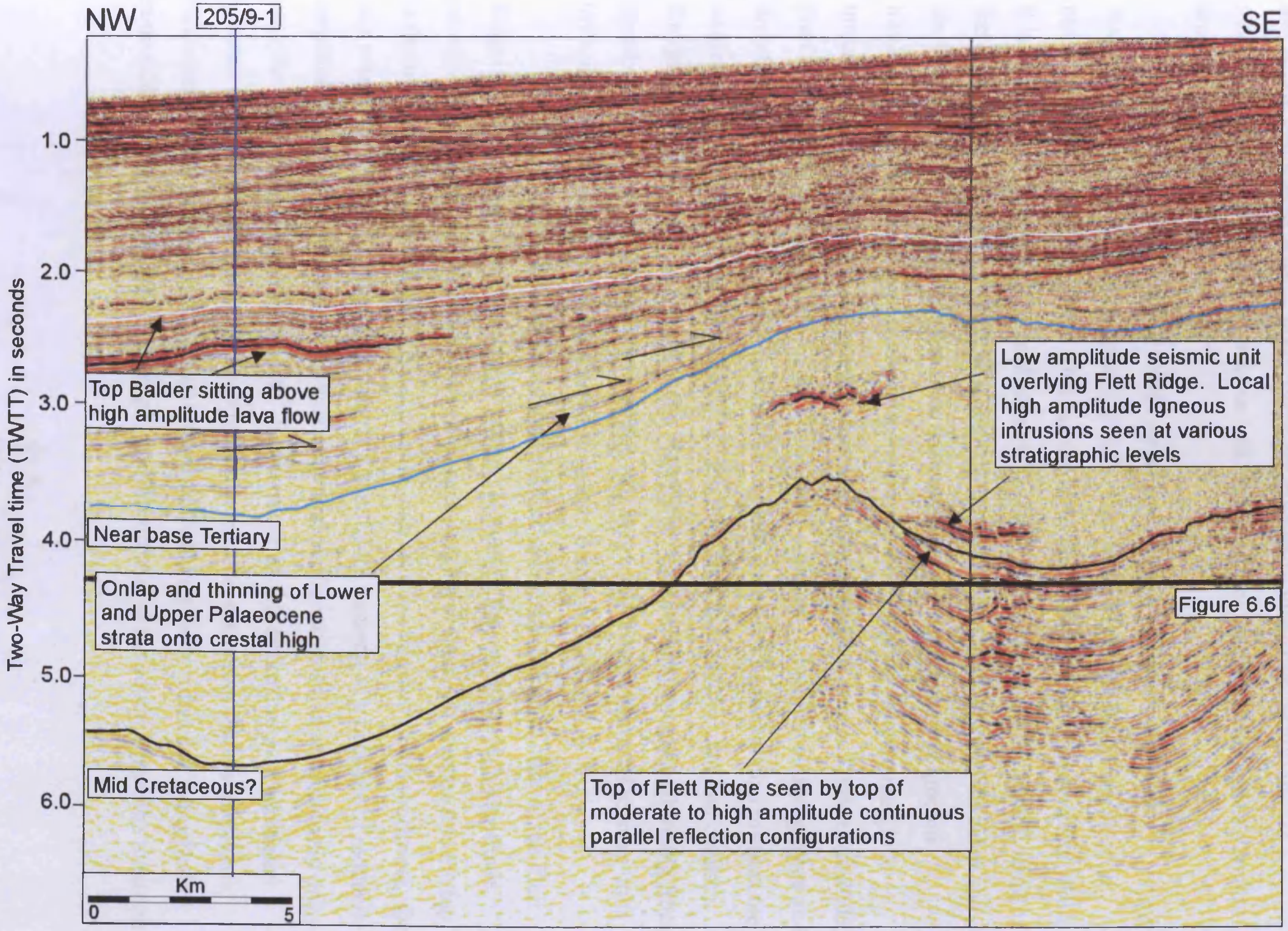


Figure 6.3. Representative 3-D seismic line across the Flett Ridge structure (northwest - southeast). This seismic line shows a dip section showing the well imaged deep structure of the ridge. The top of the ridge is highlighted by the change in seismic character from high - moderate amplitude continuous reflections to an overlying low amplitude featureless unit. A significant thinning is seen in this low amplitude Cretaceous succession across the Flett Ridge though no internal geometry can be determined. Within this low amplitude unit localised very high amplitude continuous seismic reflections that occur at different stratigraphic levels. These have been interpreted as igneous sills (see text). The post-Cretaceous succession initially onlaps and thins onto the rear base Tertiary reflection. See Figure 6.1b for line location.

much greater resolution at depth than the 2-D seismic data (e.g. under 4 seconds) where there is an increase in the amplitude of the seismic data. The top of this moderate to high amplitude seismic unit has been mapped as the top of the Flett Ridge structure, and it is overlain by a thick (greater than 1000 ms) unit which is characterised by a very low amplitude featureless seismic unit. There are little or no mappable reflection configurations visible in this seismic unit as they are either discontinuous and of very low amplitude (**Figure 6.3**).

Over the crest of the Flett Ridge structure there is a very high amplitude discontinuous reflection that is approximately 4 km long (**Figure 6.3**) and has an areal extent of approximately 10 km^2. This feature turns upwards towards its eastern end where it is seen to cross-cut very low amplitude reflections. Additionally, on the eastern flank of the Flett Ridge another very high amplitude reflection is seen to onlap onto the ridge and abuts against it (**Figure 6.3**). These two features have been interpreted as igneous intrusions and have been noticed along the whole of the Northeast Atlantic margin to intrude into the Palaeocene strata and earlier Cretaceous succession (e.g. Trude *et al.* 2003, Smallwood 2004 and see **Section 2.3**). The crestal structure of the Flett Ridge appears as a broad dome feature that is broadly symmetrical, with a western flank that dips towards the northwest and basin centre (**Figure 6.3**). In the low amplitude seismic unit that contains the igneous intrusions which overlies the Flett Ridge a significant amount of thinning from approximately 1750 ms – 1000 ms (over the ridge) is seen. This thinning of seismic units towards the Flett Ridge will be discussed in greater detail in **Section 6.4**.

The seismic data is relatively poorly imaged along the structural crest of the Flett Ridge (in a northeast – southwest orientation). However, the top of the high amplitude continuous package seen at depth can still be mapped (**Figure 6.4**). It appears that the top is faulted on the southern edge of the structural high and downthrown to the southwest. On this southwestern flank of the ridge there is a significant thinning of a package of moderate amplitude continuous to semi-continuous reflection configurations that progressively onlap onto the crest of the ridge (**Figure 6.4**). Furthermore, in this strike direction additional igneous intrusions can be seen at different stratigraphic levels within the low amplitude, weak (featureless) seismic unit. These igneous intrusions show evidence of being both concordant (to the north of the ridge high) and discordant (to the south) where the intrusion

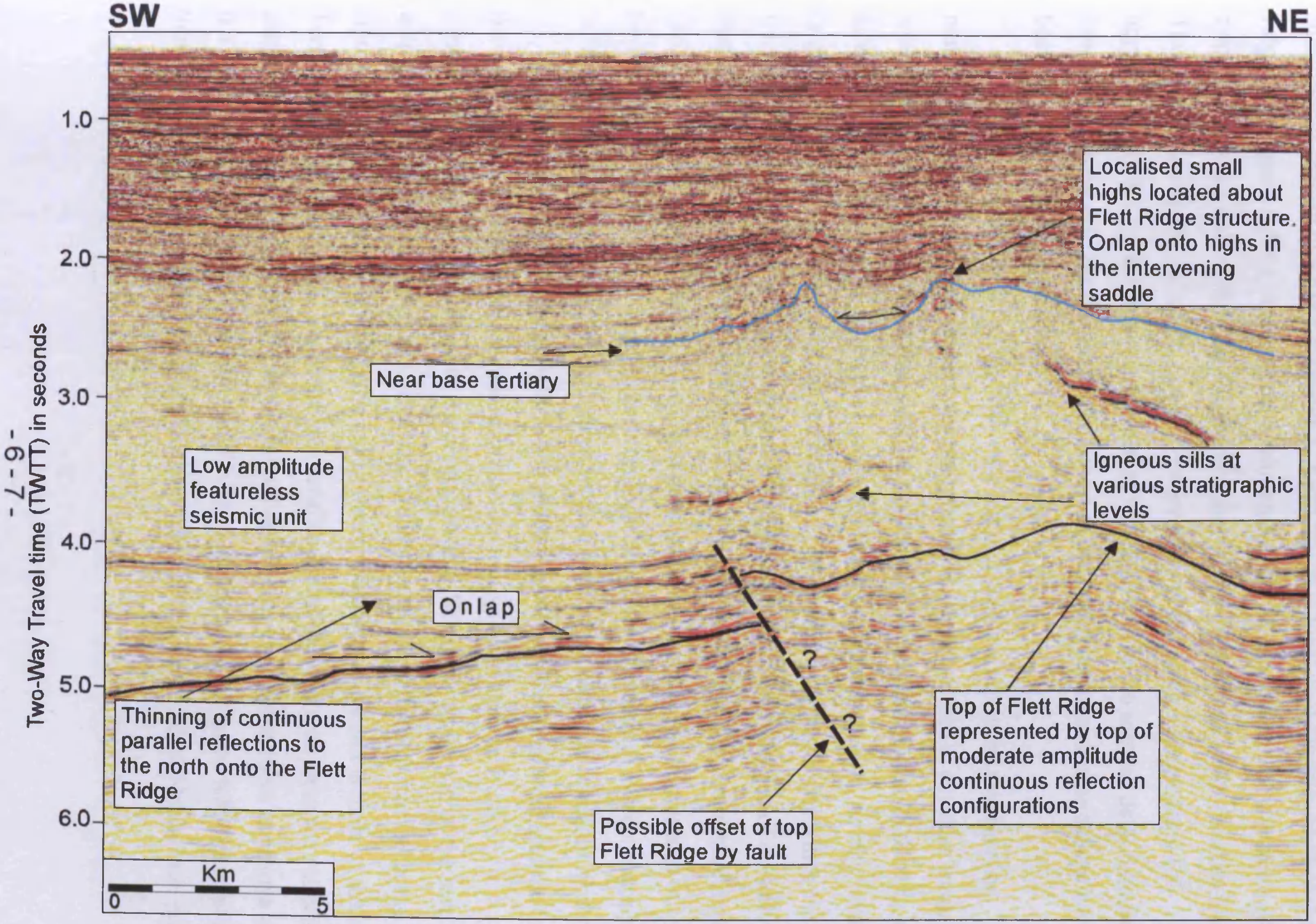


Figure 6.4. Northeast - southwest trending strike line (see Figure 6.1b for location) along the Flett Ridge structure where the top of a high amplitude continuous reflection package depicting the top of the ridge. There is possible evidence of an offset in the top of the ridge which is then onlapped from the south. Above the Flett Ridge is a low amplitude unit which contains bright high amplitude local igneous sills at different stratigraphic level. Above this low amplitude unit is a moderate amplitude package which consists of continuous parallel reflections. Near to the base of this package there are two localised structural highs which are separated by an intervening saddle feature. These two highs are located over the crest of the Flett Ridge and may be related to the growth of the structure or be associated with the intrusion of the igneous bodies. There is later onlap into the saddle feature and this suggests that the highs created some positive bathymetric relief on the sea-floor. The most northerly local high seems to have a localised core of higher amplitude discordant reflections which appear as a diamond shape and this may be some high level small igneous body. See Figure 6.1 for location.

looks to be cuspate, bowl shaped and concave upwards (**Figure 6.4**), a common feature in igneous intrusions in this margin (e.g. Trude 2004). A 3-D perspective view of the Flett Ridge structure displayed in **Figure 6.5** highlights the broad dome at its crest.

A seismic time-slice through the Flett Ridge structure reveals a near perfect circular structure at depth (**Figure 6.6**). This seismic time-slice is taken at 4250 ms and highlights this circular feature that has a diameter of 7 km which is located at the very crest of the ridge (see **Figure 6.3** for vertical profile showing depth of time-slice). To the south of the circular body, a smaller identical feature is found with a diameter of approximately 2 km. This feature is slightly detached from the larger body to the north. At this level (of approximately 4.5 – 5 km when calibrated to well 205/9-1) in the seismic data, igneous intrusions can be seen locally to the southwest of the two circular bodies and these have been interpreted as igneous sills.

It is postulated here that the two near circular features represent two igneous plutons that were the magma source for the igneous intrusions found at a higher level. No wells have ever been drilled to these depths on this part of the Flett Ridge, so there is no direct evidence pointing to an igneous origin for these features. However, nearby wells (e.g. 205/10-2b) have penetrated what have been interpreted as igneous sills which have a felsic origin and this has led some authors to believe this indicates a degree of crustal re-melting, possibly by the emplacement of a magma chamber within the continental crust (Smallwood and Maresh 2002). Thus, combining this evidence with the presence of the two perfect circular bodies at depth seen on **Figure 6.6**, the presence of a deep plutonic or volcanic edifice under or indeed emplaced in the Flett Ridge is viewed as highly likely.

Above the Flett Ridge structure, and indeed above the low amplitude seismic unit, there are continuous to semi-continuous moderate to high amplitude reflection configurations which also thin onto the crestal area of the ridge. At the top of the low amplitude unit there is a mappable high amplitude continuous reflection that has some significant relief on it. This reflection is correlated to nearby wells and is close to the base Tertiary stratigraphic level. There are two localised folds that are seen from the reflection geometry and these folds exhibit an antiformal geometry. In between these antiforms there is a small saddle (synform) which is about 1.5 to 2 km in length (**Figure 6.4**). Interestingly these two features are positioned directly above the Flett Ridge and could be influenced by

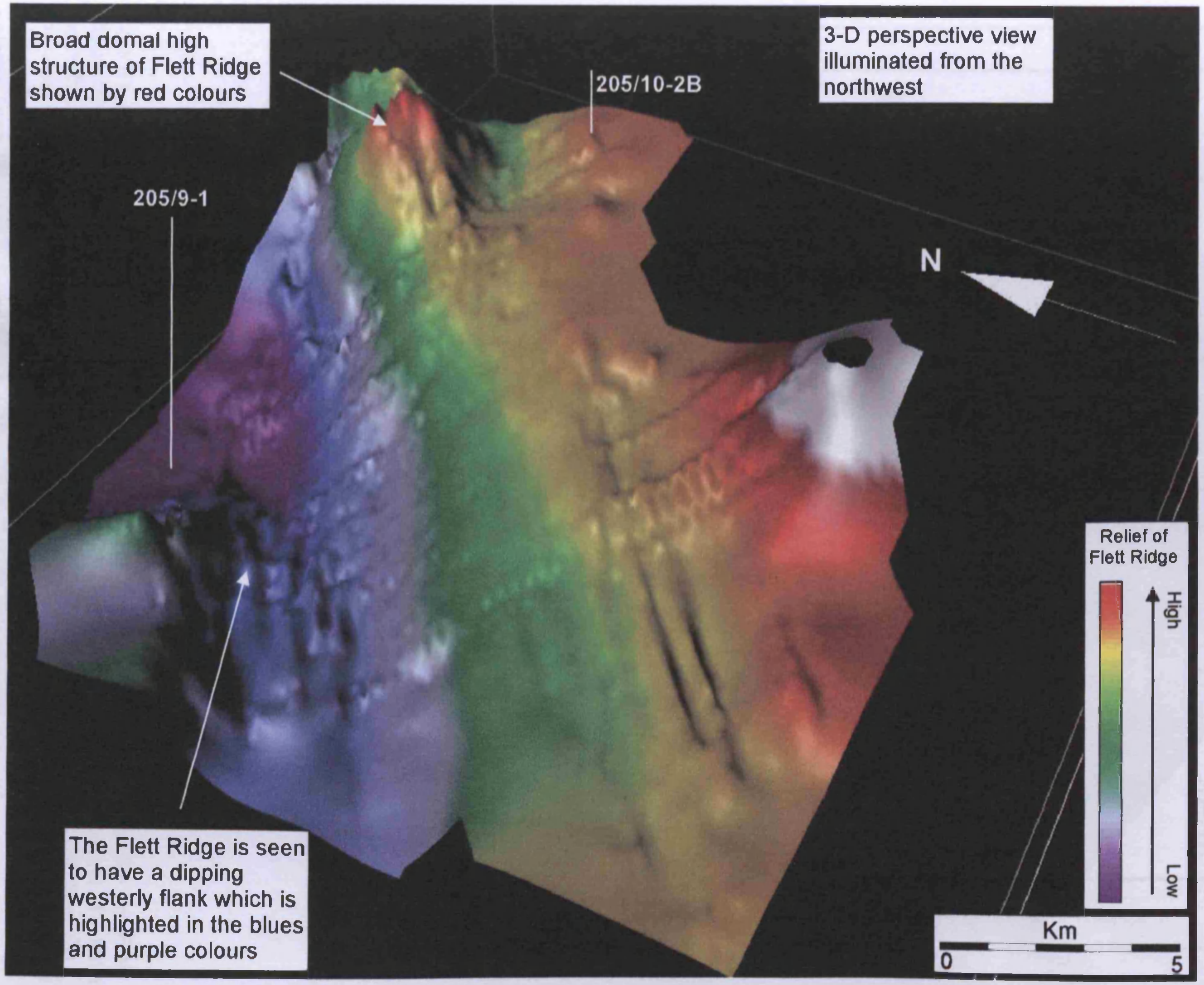


Figure 6.5. 3-D perspective view (looking northeast) of the Flett Ridge structure. This structure is broadly domal in shape and sits at the northern end of the 3-D survey. It is elongated in a northeast southwest direction and is approximately 2.5 - 3 km in diameter. Above this domal crest are two small localised highs are seen at the near base Tertiary stratigraphic level (see Figure 6.7). High amplitude continuous seismic reflections can be seen terminating close to the crest of the Flett Ridge. These high amplitude, cuspate, bowl shaped reflections are interpreted as igneous sills, one of which was drilled by well 205/10-2b. These sills occur at many stratigraphic levels in the low amplitude Cretaceous section and are shown in Figures 6.3 and 6.4. This image is vertically exaggerated by a factor of 5.

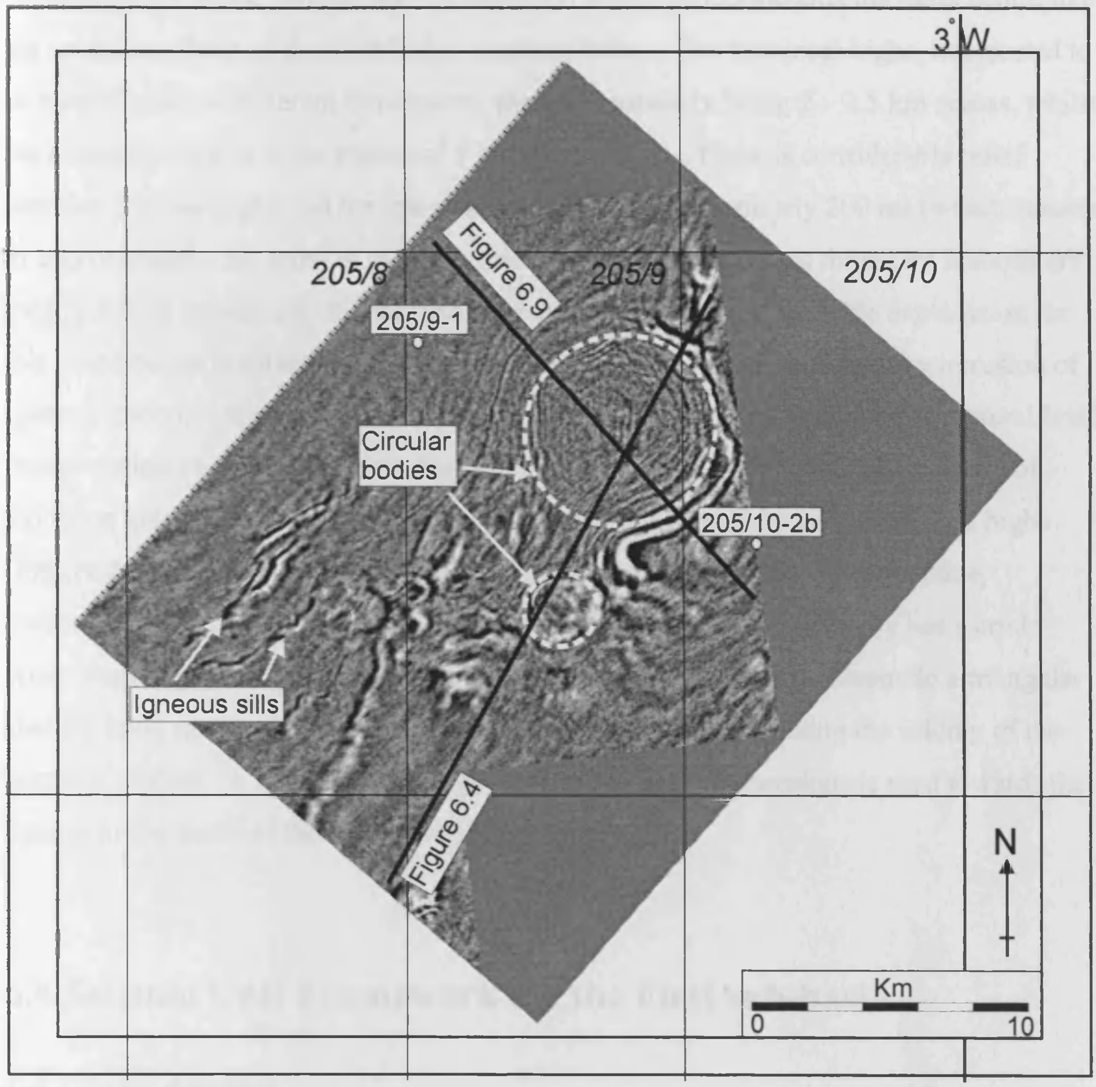


Figure 6.6. Seismic time-slice at 4260 ms showing two circular bodies at depth under the Flett Ridge structure. The bodies show different dimensions the largest being in the region of 7 km in diameter. The smaller of the two features is seen to be south southeast of the larger one. To the west of these circular features igneous sills have been interpreted at this level. The circular bodies have been interpreted to represent igneous plutons (see text). The positions of two seismic lines (Figures 6.4 and 6.9) and the two key wells are also highlighted.

the growth of the ridge or associated with the igneous intrusions. When mapped throughout the entire 3-D survey this reflection is seen to dip towards the basin centre like the northwest flank of the Flett Ridge structure below. The two local highs, interpreted to be anticlines have different dimensions, the most northerly being 2 – 2.5 km across, whilst the southerly high is in the region of 1 km (**Figure 6.7**). There is considerable relief between the two highs and the intervening saddle of approximately 200 ms (which equates to approximately the same in metres when depth converted). Thus these two features are locally 200 m structurally higher than the surrounding area. A plausible explanation for this could be for emplacement of additional material locally (for example by intrusion of igneous material) to push up a small area of the overlying strata to a higher structural level. Interpretation of the seismic data has revealed the presence of a small area or body of moderate amplitude situated just below the most northerly of the local structural highs (**Figure 6.4**). This very small discrete feature is discordant to the low amplitude, predominantly featureless seismic data that surrounds it. This small body has a crude prism shaped geometry with steep flanks or sides and may possibly resemble a triangular shallow level igneous pluton that is sitting directly under and causing the folding of the northern feature. A more regional high, with much larger dimensions is seen towards the margin in the south of the 3-D survey (**Figure 6.7**).

6.4 Seismic Unit Framework for the Flett sub-basin

6.4.1 Introduction

This section will outline the stratigraphic development of the Palaeogene succession in the area surrounding and adjacent to the Flett Ridge. A number of continuous, moderate to high amplitude seismic reflections have been mapped throughout the 3-D seismic survey. The Palaeogene succession is particularly thick (up to 2 km) in this part of the basin with the older Palaeocene sediments infilling the Flett sub-basin. A series of dipping seismic reflection configurations interpreted as an extensive progradational deltaic wedge has been mapped in the area overlying the Flett Ridge and Flett sub-basin (**Figures 6.1** and **6.3**). Biostratigraphic and lithological information is

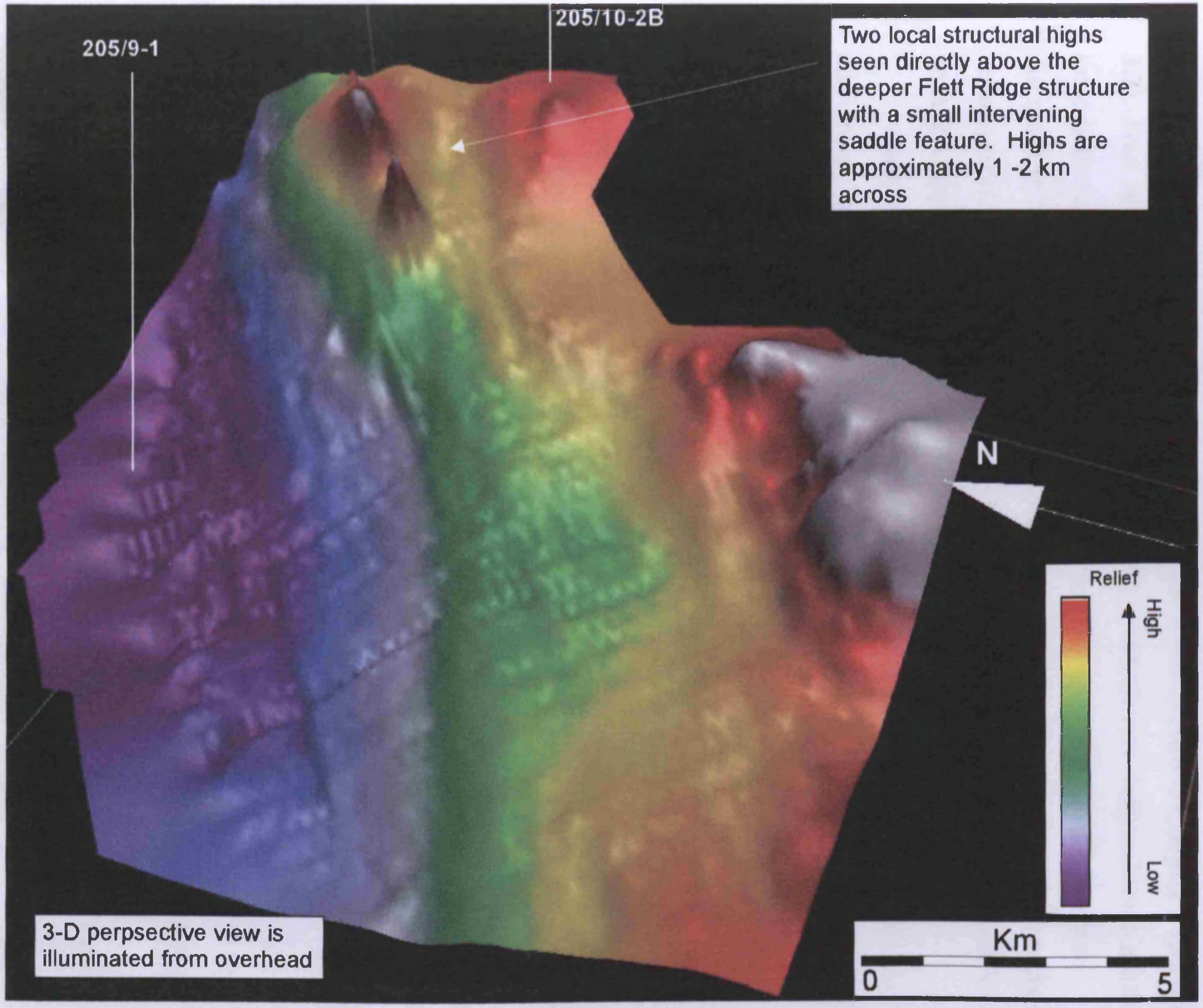


Figure 6.7. 3-D perspective view of the near Base Tertiary reflection showing the 2 localised highs that sit directly above the Flett Ridge structure (shown in the red colours). These highs are approximately 200 m high and are then onlapped by the overlying sediments. Under the most northerly of the highs is a discordant high amplitude body (see Figure 6.4) which could represent a shallow level igneous body which jacked up the high causing the deformation. The highs are between 1 and 2 km across and have along axes trending northeast - southwest. Note that the image is vertically exaggerated by a factor of 5.

sparse, with only five nearby wells. Consequently there is only limited knowledge about lithology, depositional environment and water depth. However, within the area of 3-D coverage (**Figure 6.1**) a number of mappable semi-regional seismic reflections can be seen and these have been tied to two key wells, one of which lies in the area of the 3-D survey and the other just on the edge of it (highlighted in red in **Figure 6.1b**). These reflections form the boundaries of seismic units that represent a package of seismic reflection configurations that are chronostratigraphically the same (**Figure 6.8**). This approach of seismic interpretation and stratigraphic sub-division was defined in **Chapter 3** (**Section 3.3**). Four major reflections mapped above the top T50 (Balder Tuff) reflection are the same as those that have been mapped around the whole of the basin and have been discussed in broad detail in **Chapter 4** (**Section 4.3**). These four continuous, correlatable seismic reflections, along with the top T50 (Balder Tuff) reflection bound the four seismic units (Eocene 1 - Eocene 4) discussed here (**Figure 6.8**).

However, some additional reflections have been mapped within the older Lower and Upper Palaeocene succession. These reflections aid in the stratigraphic sub-division of the earlier succession. Three seismic units are identified in the Palaeocene succession, one in the Early Palaeocene and two in the Late Palaeocene (**Figure 6.8**). These units are herein termed Pal 1, 2 and 3 respectively. An overview of the Palaeogene stratigraphy used in this study is calibrated with previously published schemes of sequence stratigraphy and lithostratigraphy (**Figure 6.8**). A schematic lithology log from the Flett sub-basin complied from age data and cuttings from the nearby two wells is also shown. Both of these wells have been assigned as type wells in this study (see **Section 3.5**) and thus represent wells which have the best available biostratigraphy and lithological information from the Eocene succession. The location of the two wells (205/9-1 and 205/10-2b) are shown on **Figure 6.1**.

6.4.2 Early Palaeocene - Danian

The Lower Palaeocene succession is penetrated in both type wells and was recognised by its biostratigraphic signature. Results from the most recent micropalaeontological data from the contractors well report supports an Early Palaeocene

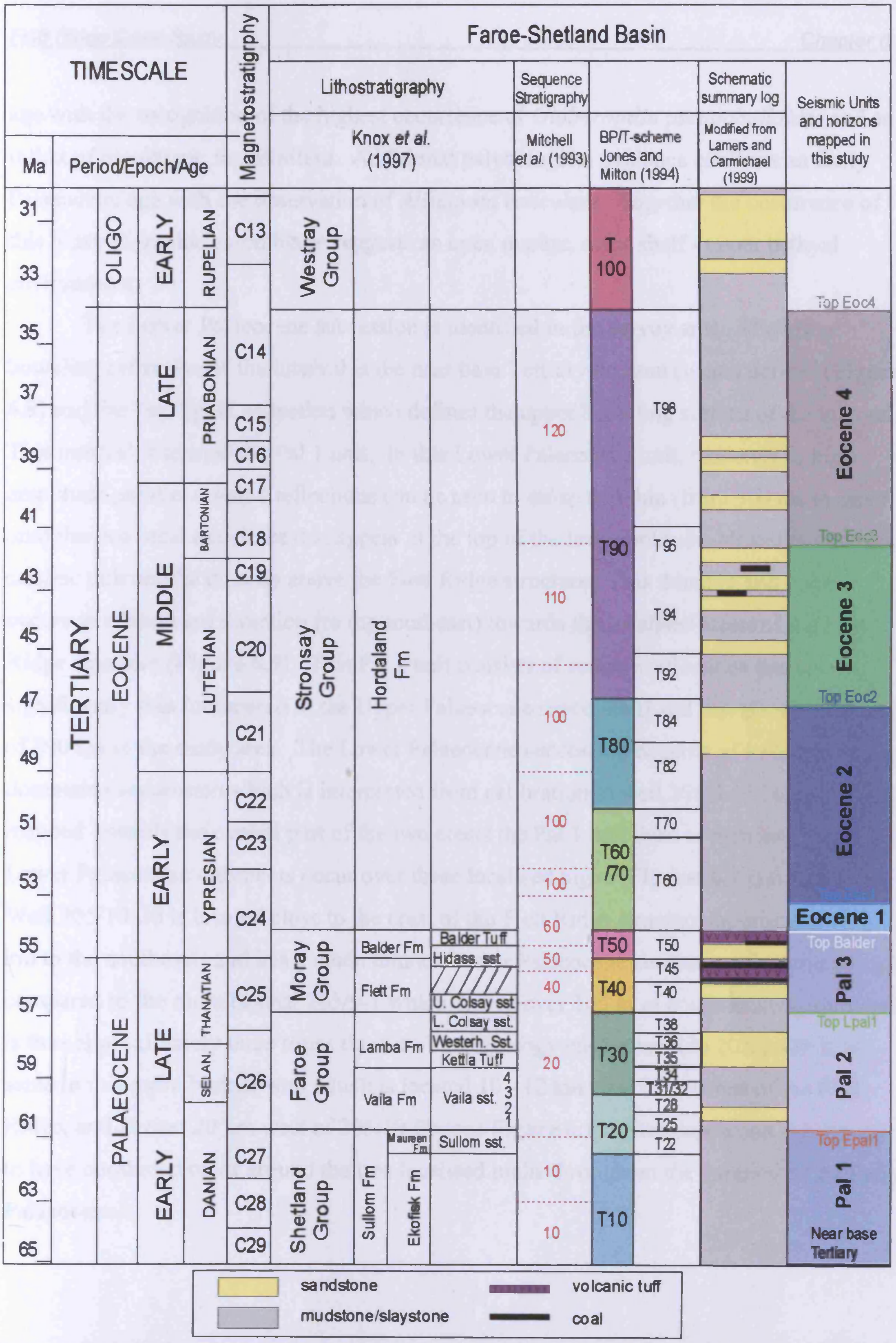


Figure 6.8. Stratigraphic chart summarising main lithostratigraphy and sequence stratigraphy of the Palaeogene strata of the Flett sub-basin. The seismic units mapped in this study are shown and calibrated with the T-scheme units of Jones and Milton (1994) and Ebdon *et al.* (1995). The seismic reflections which mark the tops of the Eocene aged seismic units are the same as the regional seismic markers described in chapter 4. However, additional Palaeocene units are mapped in this study. A summary log of the Flett sub-basin shows the position of sands in the sub-basin. Timescale from Berggren *et al.* (1995).

age with the recognition of the highest occurrence of *Globorotalia pseudobulloides* and an influx of planktonic foraminifera. Additional palynological evidence confirms an Early Palaeocene age with the observation of *Alisocysta reticulate.* Together the occurrence of this biostratigraphic assemblage suggests an open marine, outer shelf - upper bathyal environment.

The Lower Palaeocene succession is identified in the survey area. The lower bounding reflection of the interval is the near base Tertiary horizon (which defines (**Figure 6.8**) and the Top Epal1 reflection which defines the upper bounding surface of the interval. This interval is termed the Pal 1 unit. In this Lower Palaeocene unit, moderate to high amplitude parallel seismic reflections can be seen to onlap and thin (from 300 ms to zero) onto the two local anticlines that appear at the top of the low amplitude, almost featureless seismic unit and lie directly above the Flett Ridge structure. This thinning and onlap occurs in a landward direction (to the southeast) towards the localised crests of the Flett Ridge structure (**Figure 6.9**). This Pal 1 unit consists of seismic reflections that are significantly thin (compared to the Upper Palaeocene succession) and reaches a maximum of 300 ms in the study area. The Lower Palaeocene succession consists of a shale/mud dominated succession which is interpreted from calibration to well 205/9-1. Indeed, when mapped towards the central part of the two crests the Pal 1 unit thins to zero and hence no Lower Palaeocene sediments occur over these localised highs (**Figures 6.9** and **6.10**). Well 205/10-2b is located close to the crest of the Flett Ridge structure (approximately 6 km to the southeast) and has a much thinner Lower Palaeocene succession (less than 40 m) compared to the more basinal 205/9-1 which drilled over 105 m of comparable section and is thus approximately three times thicker. The lithology encountered in 205/10-2b is the same in this more basinal well which is located 10 - 12 km west of the crest of the Flett Ridge, and almost 20 km west of 205/10-2b (see **Figure 6.1**). Non-deposition is believed to have continued on or around the two localised highs throughout the duration of the Early Palaeocene.

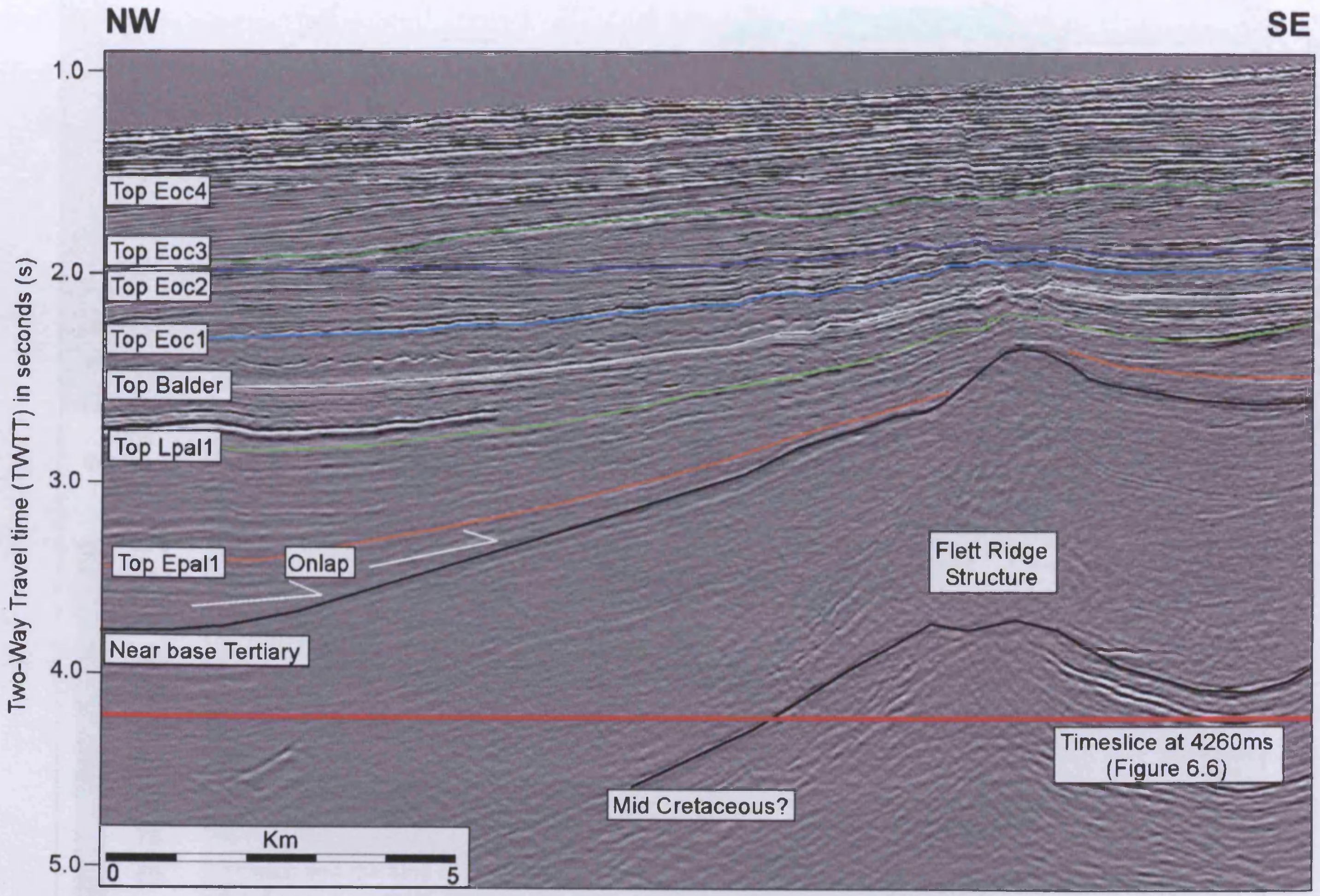


Figure 6.9. Southeast - northwest trending seismic line from a 3-D survey across the Flett Ridge. The deep structure of the Flett Ridge is visible and picks out by high amplitude parallel and continuous reflections. At the near base Tertiary level the ridge forms an anticline that effects the deposition and areal extent of sediments in the Cretaceous and Early Palaeocene. Later onlap and thinning onto this anticline is seen in the in the Late Palaeocene. Two major incision events occur within the Eocene succession close to the crest of the Flett Ridge. For the line location see the deep time-slice in Figure 6.6.

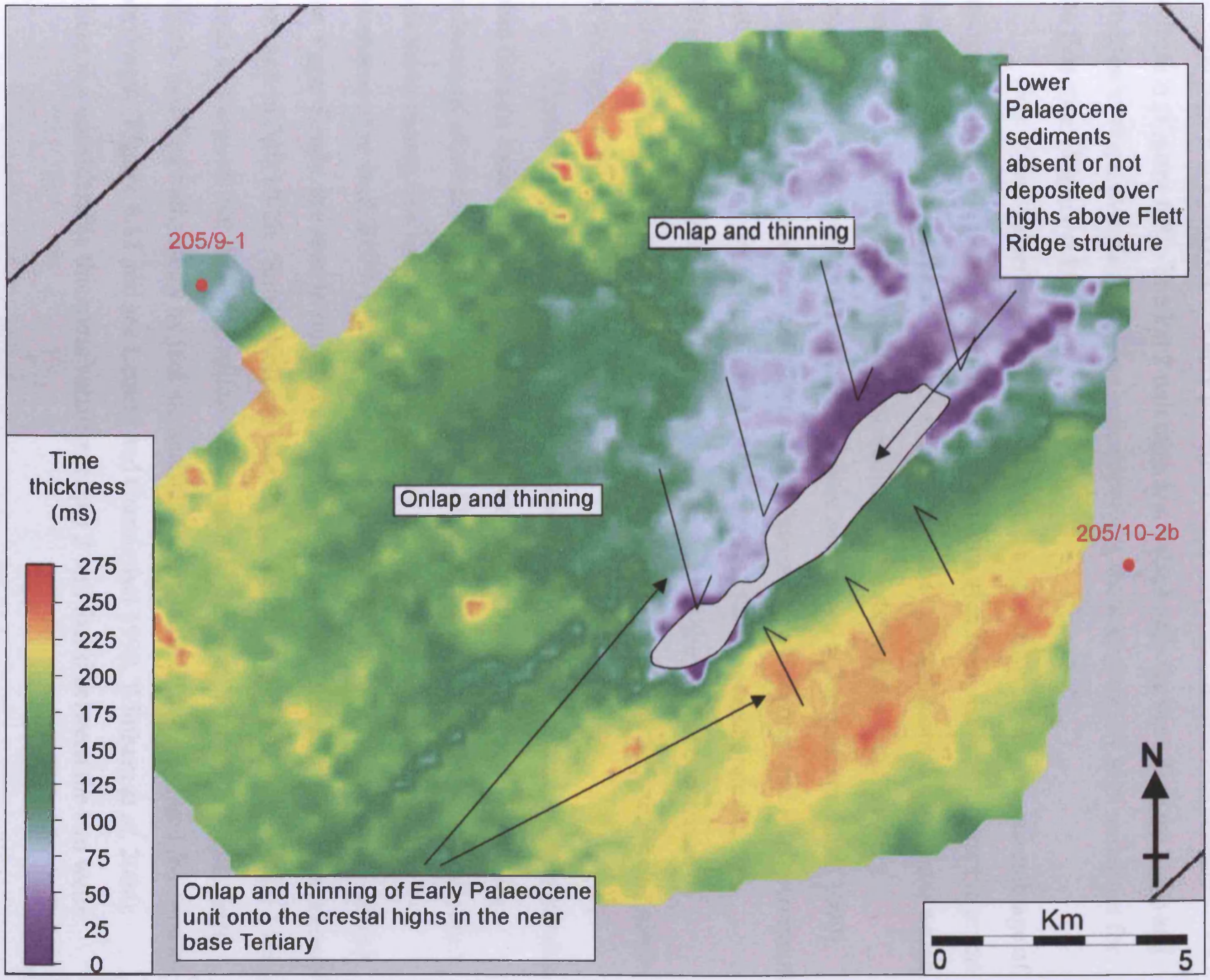


Figure 6.10. Isochron map (thickness in time) showing the onlap and thinning of the Lower Palaeocene succession onto the near base Tertiary highs located above the Flett Ridge. No deposition occurred over the highs during the Early Palaeocene (shown by the grey region in the centre) as thinning and onlap are seen onto the anticlines. Most of this thinning and onlap is seen on the basin-ward (northwest) flank of the highs.

6.4.3 Late Palaeocene - Selandian-Thanatian

The Upper Palaeocene succession in the Flett sub-basin is divided into two units. The lower unit is (Pal 2) is discussed in this section and represents the majority of Upper Palaeocene interval and conformably overlies the Lower Palaeocene unit (Epal1). The Pal 2 unit is defined at its base by the Top Epal1 horizon and at its top by the Top Lpal1 horizon which sits just below a very prominent high amplitude continuous seismic reflection (**Figure 6.9**). The Pal 2 unit thins and onlaps onto the base Tertiary highs and are seen to drape over the anticlines hence covering these structurally high anticlines for the first time (**Figure 6.11**).

This Upper Palaeocene succession is dated by an extremely diverse assemblage of micropalaeontological and micropalynological fauna seen from the biostratigraphic report from well 205/9-1. Important marker events (from the micropalaeontological data) which confirm a Late Palaeocene age for these sediments are the highest occurrences of *Spiroplectammina spectabilis* and *Reticulophragmium sp.1* (Charnock and Jones 1990). Additionally, *Rhabdammina abyssorum* with *Haplophragmoides* spp. and locally common radiolarians and diatoms of the *Coscinodiscus* species further corroborate this age. Palynological data also gives a Late Palaeocene age with the highest occurrence of *Alisocysta margarita* and the increased abundance of *Areoligera cf. coronata* being some of the main events.

Significant thicknesses of Upper Palaeocene sediments continued to onlap and thin onto the two local highs seen at the base Tertiary level (**Figures 6.9** and **6.11**). Large volumes of siliciclastic material fed from the Shetland margin were deposited in the Flett sub-basin during this interval including locally the deposition of sands that were encountered in well 205/9-1 (see **Figure 6.8**). These sands, which are informally termed the Vagar Sands, are seen to thin and pass laterally into more argillaceous, mud dominated intervals in 205/10-2b. Smallwood and Maresh (2002) highlighted the distribution of these sands and argued that there deposition was controlled by igneous sill intrusions at depth which modified bathymetry by jack-up and that these sands were sourced and fed from the northwest (**Figure 6.12** and see Lamers and Carmichael 1999, Whitham *et al.* 2004). There is a considerable thickness variation of pal 2 unit found between the two wells

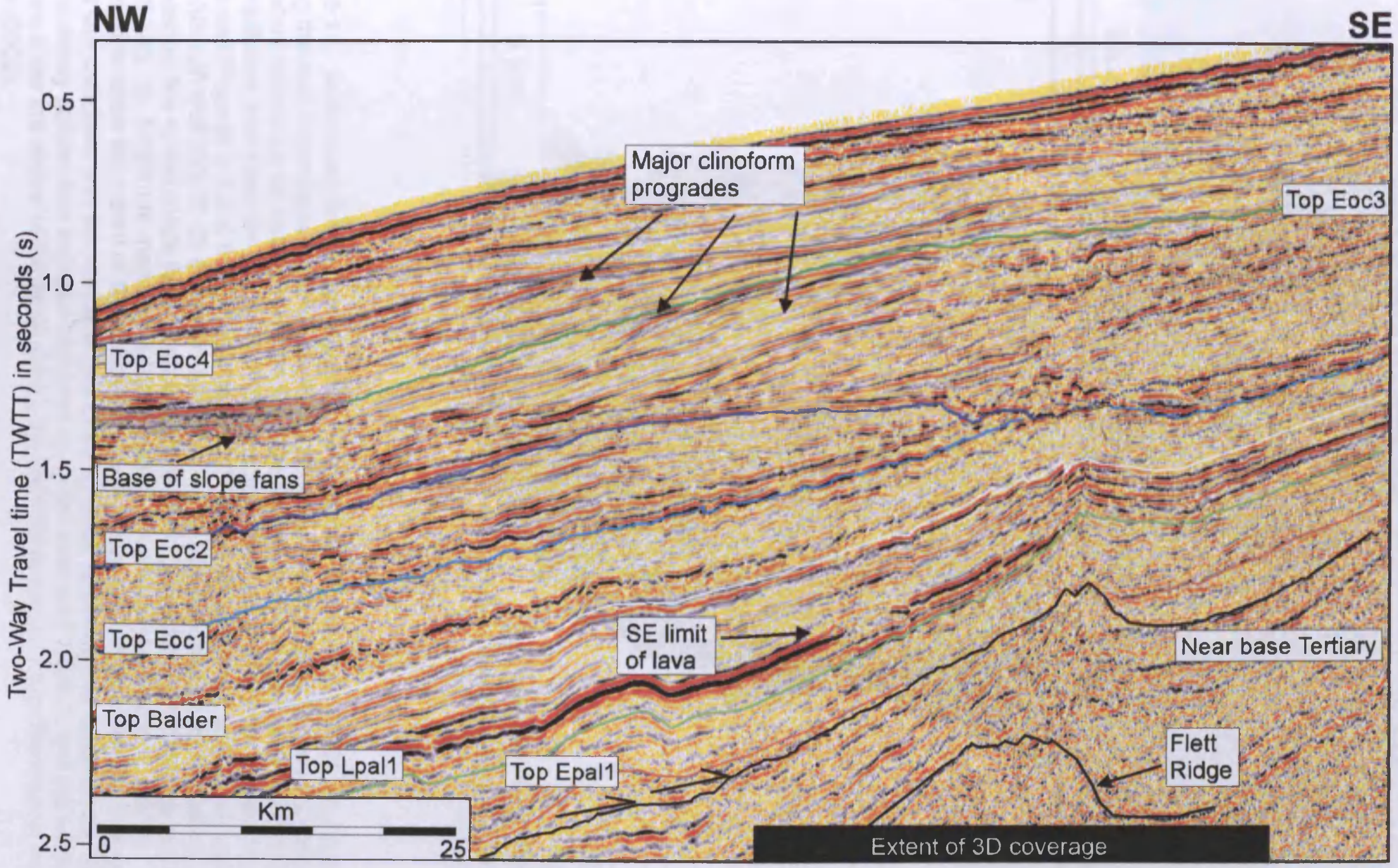


Figure 6.11. Regional southeast - northwest trending 2-D seismic line crossing the Flett Ridge study area. This seismic panel shows the general stratigraphic architecture of the Eocene succession on the Shetland margin close to the Flett Ridge structure. The south easterly limit of the Faroe Lava series is highlighted which lies conformably within the upper part of the Palaeocene section. Also highlighted are the areal extent of the 3-D survey, the Flett Ridge structure, and the major northwesterly dipping seismic reflection configurations that are interpreted to represent a progradational of Middle Eocene age (corresponding to the Eocene 3 seismic-stratigraphic unit. Down-dip from these clinoforms a high amplitude chaotic unit is seen and is interpreted to be base of slope fans. For line location see Figure 6.1.

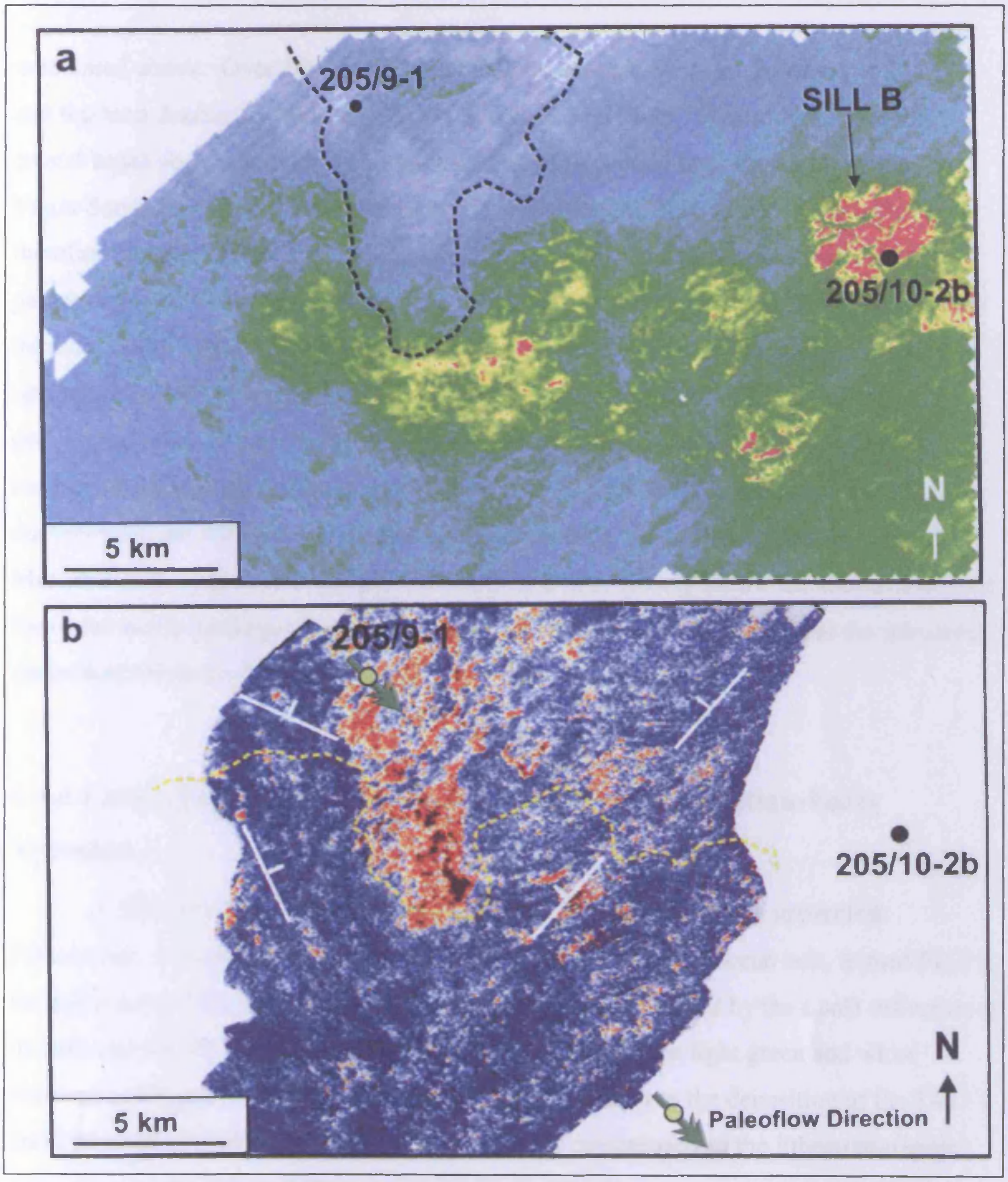


Figure 6.12. Seismic attribute maps taken from Smallwood and Maresh (2002). a). Attribute map showing the root mean square (RMS) amplitude between the near base Tertiary and Mid Cretaceous reflections as shown in Figure 6.9. The brighter colours (pinks and yellows) indicate high amplitudes and show the positions of the igneous sills seen in this generally low amplitude seismic unit (Figures 6.9 and 6.3). The locations of the two wells referred to in this chapter are shown (though well 205/10-2b does not lie within the 3-D survey available for this study). The black dashed line corresponds to the high amplitude Late Palaeocene reflections shown here in Figure 6.10b. b). Amplitude map from a Late Palaeocene reflection tied to well 205/9-1 interpreted to show the extent of the turbidite sandy package of the Vagar Sands. The yellow dashed line coincides with the northerly limit of the igneous sills shown in figure 6.10a. Note that a southeasterly palaeo-flow indicator has been recorded from well 205/9-1, and the white dip indicators show the shape of the basin during the Late Palaeocene. From Smallwood and Maresh (2002).

mentioned above. Over 1600 m of Upper Palaeocene sediments are recorded in 205/9-1 and this unit drastically thins to just over 300 m in 205/10-2b. This suggests that the crestal highs above the Flett Ridge were still topographically high during this time. The Vagar Sands have been interpreted as turbidite fans by Smallwood and Maresh (2002), therefore suggesting that the basin was relatively deep at this time though significant palaeo-bathymetry was evident over the Flett Ridge structure allowing for the significant thinning of the Upper Palaeocene succession towards the east. Detailed mapping of the igneous sills and the Vagar Sands shows that there is a strong relationship between the geometries of the two bodies (**Figure 6.12**). There is also a close spatial relationship between the lower sills with the sands which may suggest that the intrusion of the sills directly affected the sediment dispersal of the turbiditic Vagar Sands (Smallwood and Maresh 2004). The northern limit of the sills is seen to closely follow the southern limit or toe of the sandy package (**Figures 6.12a & b**) leading to the conclusion that the intrusives controlled sediment deposition.

6.4.4 Latest Palaeocene - Earliest Eocene - Late Thanatian-Early Ypresian

A distinct change in the seismic reflection geometry is seen in uppermost Palaeocene - lowermost Eocene sediments. This Palaeocene – Eocene unit, termed Pal 3 is the upper part of the Late Palaeocene aged interval and is defined by the Lpal1 reflection at its base and the top T50 (Balder Tuff) reflection at its top (see light green and white horizons in **Figure 6.11**). The Pal 3 unit is time equivalent to the deposition of the T40 and T50 units of Ebdon *et al.* (1995) which closely corresponds to the lithostratigraphic formations of the Sele and Balder respectively in this area.

A dramatic thinning of this unit capped by the top T50 (Balder Tuff) reflection can be observed over the crestal highs seen at base Tertiary level (e.g. **Figures 6.11** and **6.3**). This draping and blanketing of sediments continued from the previous earlier Late Palaeocene interval (Pal 2 unit). This stratal relationship differs greatly from the observed onlap onto the highs seen in the Early Palaeocene interval (Pal 1 unit) and is of great importance when considering the stratigraphic and depositional setting (see later discussion in **Section 6.5**).

Well calibration of the Pal 3 unit shows abundant coals, sandstones, mudstones and tuffaceous deposits which strongly suggest that deltaic conditions prevailed during the latest Palaeocene and at the Palaeocene - Eocene transition. The lithofacies seen in this well within this interval show great variation and are markedly different from the turbidite sands seen in the Lower Palaeocene succession (Pal 2 unit). A significant amount of shallowing is interpreted to have occurred throughout the Late Palaeocene to give rise to the marginal deltaic deposits encountered. An increase in sediment supply from the margin may have caused significant delta progradation to take place. This could explain why the crests of these local highs became blanketed with sediments during the Late Palaeocene. Additionally, the sub-marine conditions that the area was previously experiencing suggests that earliest Late Palaeocene marine conditions (which deposited the turbiditic Vagar Sands of the Pal 2 unit discussed above in **Section 6.4.3**) gave way to a more dominant shallow marine - marginal deltaic environment during a fall in relative sea-level, allowing the deltaic sediments to bury the local crestal highs by the end of the latest Palaeocene to earliest Eocene.

A well established biostratigraphic record for the Pal 3 unit is available from well 205/9-1 which confirms a latest Palaeocene – earliest Eocene age (from the contracted company well report). *Coscinodiscus sp. 1* and *2* and rare agglutinated foraminifera *Rhabdammina abyssorum* as well as a single *Spiroplectammina spectabilis* are evidence for this age. The very sparse *in situ* microfaunas from this interval is consistent with a restricted marine possibly hypohaline palaeo-environment. Abundant pollen is seen in the palynological record including *Inaperturopollenited hiatus, Caryapollenites simplex* and *Tiliaepollenites microticulatus.* Furthermore, the brackish marine indicator *Apectodinium homomorphum* is found in low numbers and a pteridophytic freshwater algae *Azolla* spp. possibly suggests lagoonal-type facies present. Finally, the presence of the spores *Cicatricosisporites* spp., *Laevigosporites* spp. and *Cyathidites australis* 'group', often found associated with the coaly horizons suggest mature vegetation, perhaps a humid, marshy environment which gives rise to an influx of *Stereisporites* spp. seen in this interval. Therefore it is reasonable to conclude that a great deal of information can be derived from the biostratigraphic analysis both on the age of this interval and the depositional setting of the study area.

The dramatic thickness variation of the uppermost Palaeocene and lowermost Eocene sediments of the Pal 3 unit over these crestal structures at base Tertiary level above the Flett Ridge is coeval with the deposition of the T40 and T50 units of Ebdon *et al.* (1995) (in this area equivalent to the Sele and Balder Formations) the latter of which coincides with the start of the Early Eocene sea-floor spreading event in the North Atlantic and the concomitant widespread deposition of tuffaceous and volcaniclastic sediments (Knox and Morton 1988, Ritchie *et al.* 1999). The thinning of sediments of the latest Late Palaeocene interval (equivalent to the Balder Formation) indicate that the local highs at base Tertiary level directly overlying the Flett Ridge still showed some positive topographic relief that had an affect on the sediment dispersal of the latest Palaeocene - earliest Eocene deposition (**Figure. 6.11**).

A small interval of the Pal 3 unit reveals a high amplitude continuous seismic reflection can be seen abutting against the northwestern flank of the localised highs over the Flett Ridge (e.g. **Figure 6.11**). The lithology of this feature has been defined as dolerite and interpreted as a lava flow based on information from cuttings from well 205/9-1. These volcanics appear in the well between coals of both latest Palaeocene and earliest Eocene age and may suggest that deltaic conditions prevailed throughout the deposition of the Pal 3 unit and the lava flows erupted in a marginal environment at or close to sea-level. The lava has also been dated and is believed to represent the feather edge of the Upper Series of the Faroe Plateau Lava Group (Ritchie *et al.* 1999 and **Section 2.3**) and is concordant with the surrounding strata suggesting these lavas were extruded and onlapped the northwest flank of the Flett Ridge structure. The surrounding sedimentary strata adjacent to the volcanic lava flows found in well 205/9-1 have been assigned to the Flett Formation (which is approximately equivalent to the upper part of the Sele Formation of Knox and Holloway 1992 and Knox *et al.* 1997) by biostratigraphic data and represents an age of around 55 - 56 Ma (e.g. Knox 1996, Ritchie *et al.* 1999).

6.4.5 Early Eocene - Ypresian

The Lower Eocene interval is defined by the top T50 (Balder Tuff) reflection (shown as a white horizon on figures) at its base and the Top Eoc1 reflection (light blue horizon at its upper limit, and equates to the Eocene 1 unit introduced in **Chapter 4** (**Section 4.3.2**). This unit is relatively thin over the whole of the 3-D survey (reaching a maximum of approximately 200 ms towards the northwest – **Figure 6.9**) though the unit has a distinctive and variable seismic character.

The Early Eocene reflection configurations exhibit a low amplitude response, and are seen to be discontinuous to semi-continuous parallel to sub-parallel (**Figures 6.9** and **6.11**). Biostratigraphic data from wells 205/9-1 and 205/10-2b support a marine setting showing that siltstones, mudstone and occasional limestones were deposited at this time in a shelf environment. The occurrence of *Eatonicysta ursulae* and single occurrences of *Cibicidoides gr. eocaenus* and *Globigerina* sp. as well as a caved appearance of *Spiroplectammina navoarroana* all point to an Early Eocene age. Palaeo-environment analysis further suggests outer shelf oxygenated bottom waters and very weak open marine influences at this time.

This suggests that the previous deltaic environment of the latest Palaeocene – earliest Eocene was short-lived and that fully marine conditions resumed during deposition of the Lower Eocene sediments. A relative sea-level rise is therefore inferred to have occurred throughout the Late Palaeocene into the Early Eocene as the delta was flooded and the succession became progressively more marine dominated. This relative sea-level rise can be attributed to continued subsidence throughout the Early Eocene on the southeast margin of the Faroe-Shetland Basin.

A high amplitude continuous reflection that is offset by faults is observed in the Eocene 1 unit. This reflection is continuous around the entire 3-D survey and is easily recognisable by its high amplitude appearance in an otherwise low amplitude seismic package (**Figure 6.13**). This faulted reflection occurs approximately 50 ms above the high amplitude continuous reflection of the top T50 (Balder Tuff) reflection. The faults are clearly visible on seismic data and show a range of throws. An average fault throw of between 20 – 40 ms is observed for the high amplitude reflection. The entire low

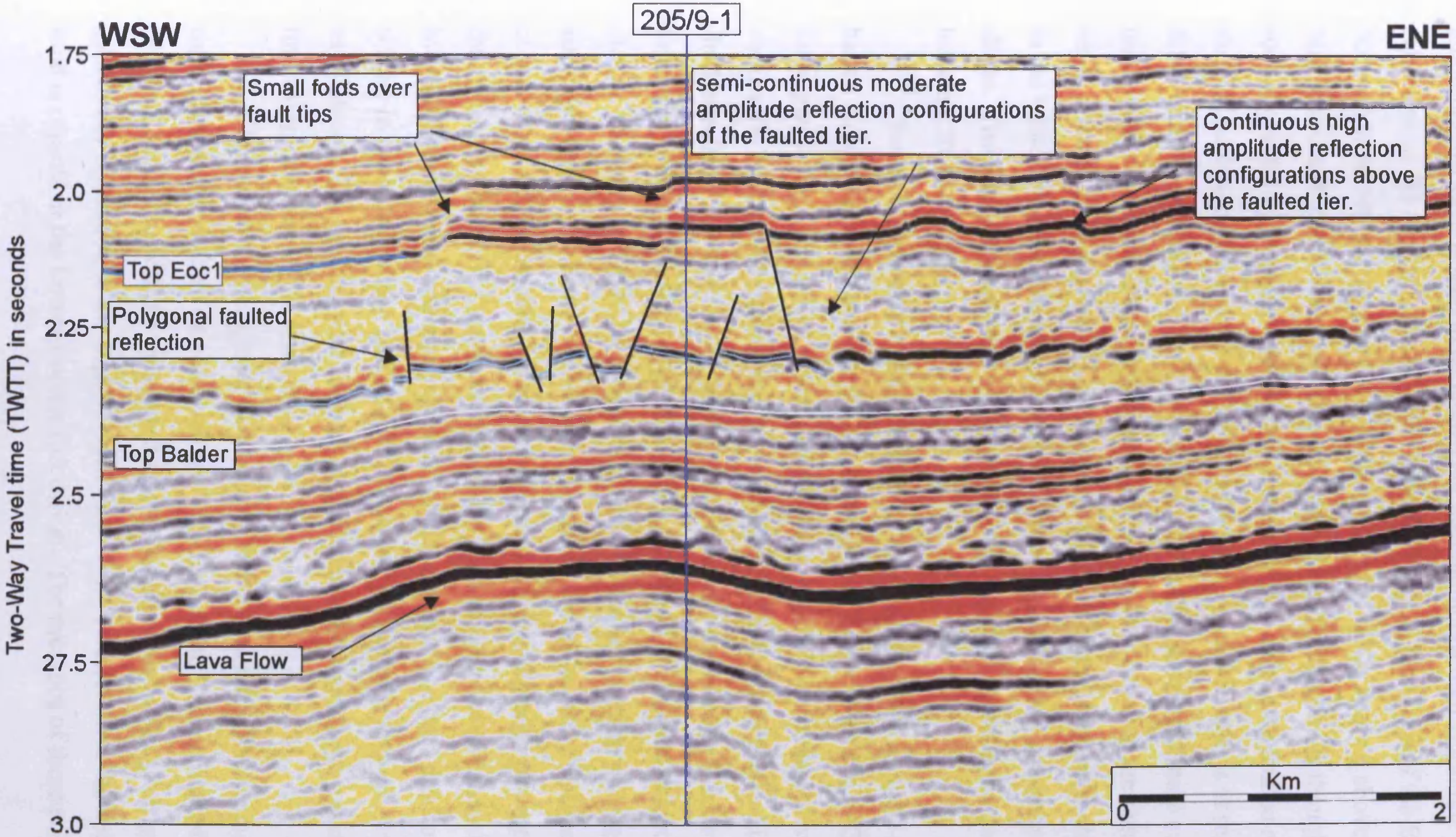


Figure 6.13. West southwest - east northeast trending 3-D seismic line through the well location of 205/9-1 showing the high amplitude faulted nature the unit between the top T50 (Balder Tuff) reflection (white horizon) and the Top Eoc1 reflection (blue horizon). This package equates to the Eocene 1 seismic unit discussed in chapter 4 (section 4.3.2). The faults have throws of varying amounts though on average they tend to be in the region of 20-40 ms. This high amplitude reflection appears in a very low amplitude discontinuous to semi-continuous reflection package and is easily recognisable. It occurs just above the top (T50) Balder interval and when calibrated to well 205/9-1 the interval is entirely composed of siltstones, mudstones and occasional limestone beds. For location of the seismic line see Figure 6.13.

amplitude package is seen to be faulted but the continuous high amplitude parallel seismic reflection of the top T50 (Balder Tuff) reflection defining the base of the Eocene 1 unit is not seen to be offset. Similarly, the seismic character of the interval above (which is also high amplitude, parallel and continuous) is not offset by these faults though small folds are seen above the upper tips of the faults. When this high amplitude faulted reflection is mapped across the whole of the seismic survey a polygonal pattern is observed (**Figure 6.14**). These polygons are best illustrated when looking at an amplitude extraction map of the seismic reflection, where the edges of the polygons (the faults) are seen as high amplitudes (red colours) and the internal part of the polygon is low amplitude and shaded in the white colours. These polygons are seen to vary in size but generally have a common shape. The dimensions of these polygons are also relatively similar, having diameters of between 500 m – 1 km.

These polygonal faults have been noticed on seismic data in many sedimentary basins around the world and were first observed in the early 1990's (e.g. Cartwright and Lonergan 1996). It is believed that this particular type of fault is caused by shrinkage and volumetric contraction of clay and mud rich material (often smectitic) which is caused by fluid expulsion and dewatering during burial and early compaction (Cartwright and Lonergan 1996). This suggests that this high amplitude continuous reflection could be clay or mud-rich and that the fault nucleated from this horizon and propagated into the siltstone above and below. As has been discussed above the entire Eocene 1 unit consists of numerous siltstones, with occasional mudstones and limestone beds appearing throughout the unit. The distinct nature of the high amplitude event may conversely indicate that the reflection may represent a limestone bed or a mud rich horizon (contained wholly in the siltstone rich interval) which the fault propagates through. However, calibration of the seismic with the well logs does not show any large variation in the interval transit time in the sonic log.

Ultimately, the polygonal faults tip out with in the Early Eocene aged tier of reflections and do not offset either the top T50 (Balder Tuff) reflection below or the Middle Eocene interval above, suggesting that the lithology of this package to be genetically dissimilar to the adjacent intervals. From correlation to the nearby wells no sand is present in the Lower Eocene succession. The presence of these polygonal faults

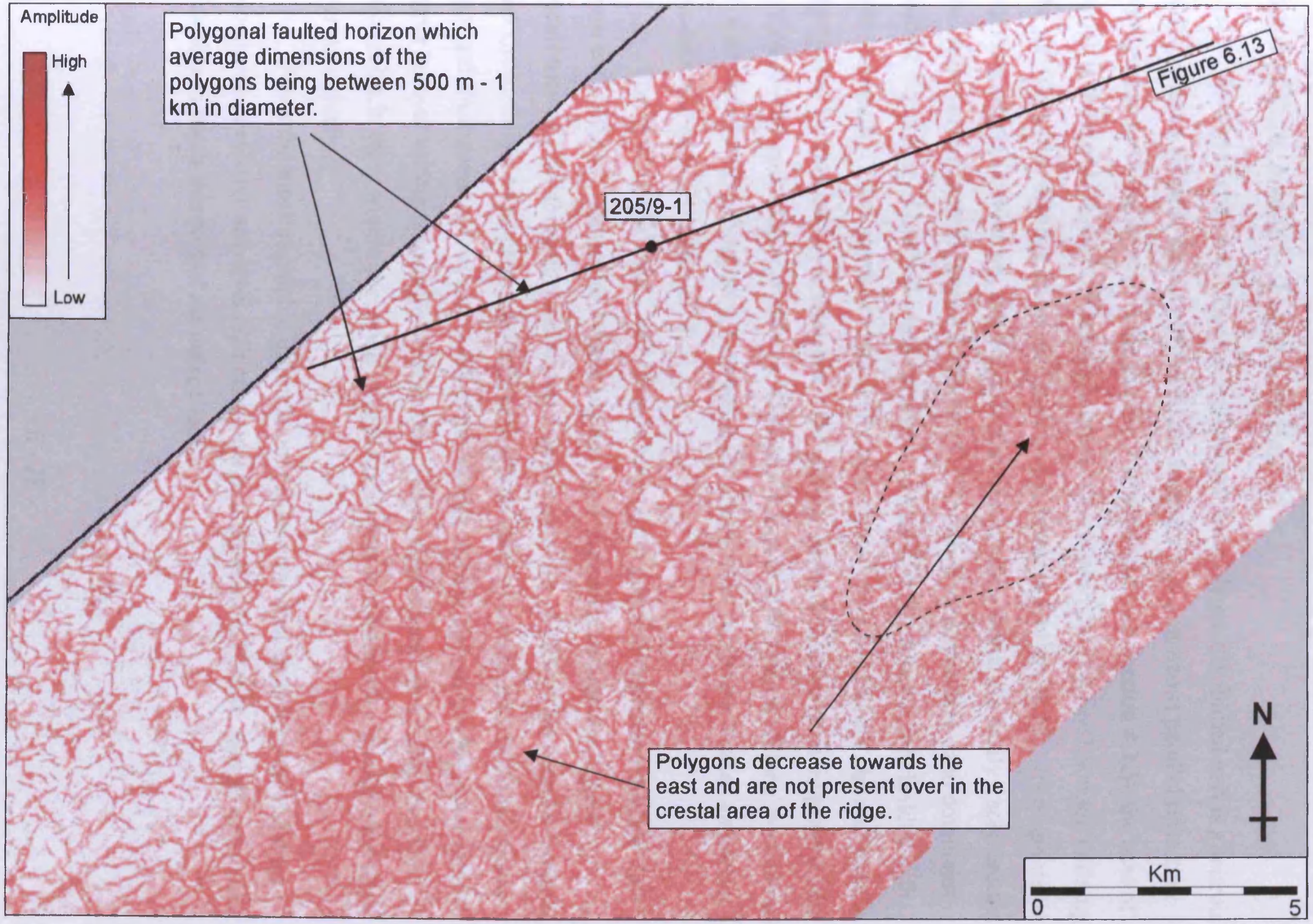


Figure 6.14. Amplitude extraction map of the intensely faulted horizon of the Eocene 1 seismic unit shown in Figure 6.13. When mapped out the faults form small polygons in the region of 500m - 1 km in diameter and are picked out by the high amplitude edges (red colours). The centres of the polygons appear to show a low amplitude as highlighted by the white shading. One observation is that the polygons are not observed in the eastern most part of the survey towards the area of the Flett Ridge crest. The location of the seismic line in Figure 6.13 is shown.

suggests a low energy mud dominated marine environment. Thus it is interpreted that a transgressive episode dominated the Early Eocene aged unit and flooding over the delta plain occurred during a rise in relative sea-level through out the Ypresian.

6.4.6 Middle Eocene - Early Lutetian

Overlying the Lower Eocene argillaceous and polygonally faulted unit is a seismic unit that has a variable amplitude and higher frequency of continuous parallel reflection configurations. This interval is the Eocene 2 seismic unit (see **Section 4.3.3**), the base of which is defined by the Top Eoc1 (light blue) horizon with the Top Eoc2 horizon defining the top (dark blue horizon on seismic panels, e.g. **Figures 6.11** and **6.15**). It comprises of high amplitude, continuous, parallel reflection configurations that are easily traced around the entire survey (e.g. **Figures 6.11** and **6.15**). This unit corresponds to sand dominated Middle Eocene aged interval when correlated to well 205/9-1 which is interbedded with silts and muds. These sands are informally termed the Upper Judd Sands and are found in the well to be over 400 m in thickness. The sand is generally fine to medium grained and is locally silty and often argillaceous. Composite logs describe this sandstone as moderate to well sorted with sub-rounded to sub-angular grains. The sandstone is often colourless occasionally pinkish or yellowish.

Dating of the Eocene 2 unit is again taken from good micropalaeontological data from well 205/9-1. Particular Middle Eocene aged marker events are the highest occurrence of *Psuedohastigerina micra*, *Truncorotaloides cf. rohri*, *Neoeponides karsteni* and *Cyclammina amplectens*. Additionally, observation of occurrences of *Cenosphaera* spp. and *Rhabdammina robusta* confirm the Middle Eocene age. A marine, outer shelf – upper bathyal setting is interpreted from this diverse microfaunal assemblage that had dysaerobic bottom waters becoming oxygenated higher in the succession with some open marine influence.

From the biostratigraphic information it is believed that shelf deposition during conditions of relative sea-level high continued into the Middle Eocene, though more sand was able to reach this part of the outer shelf.

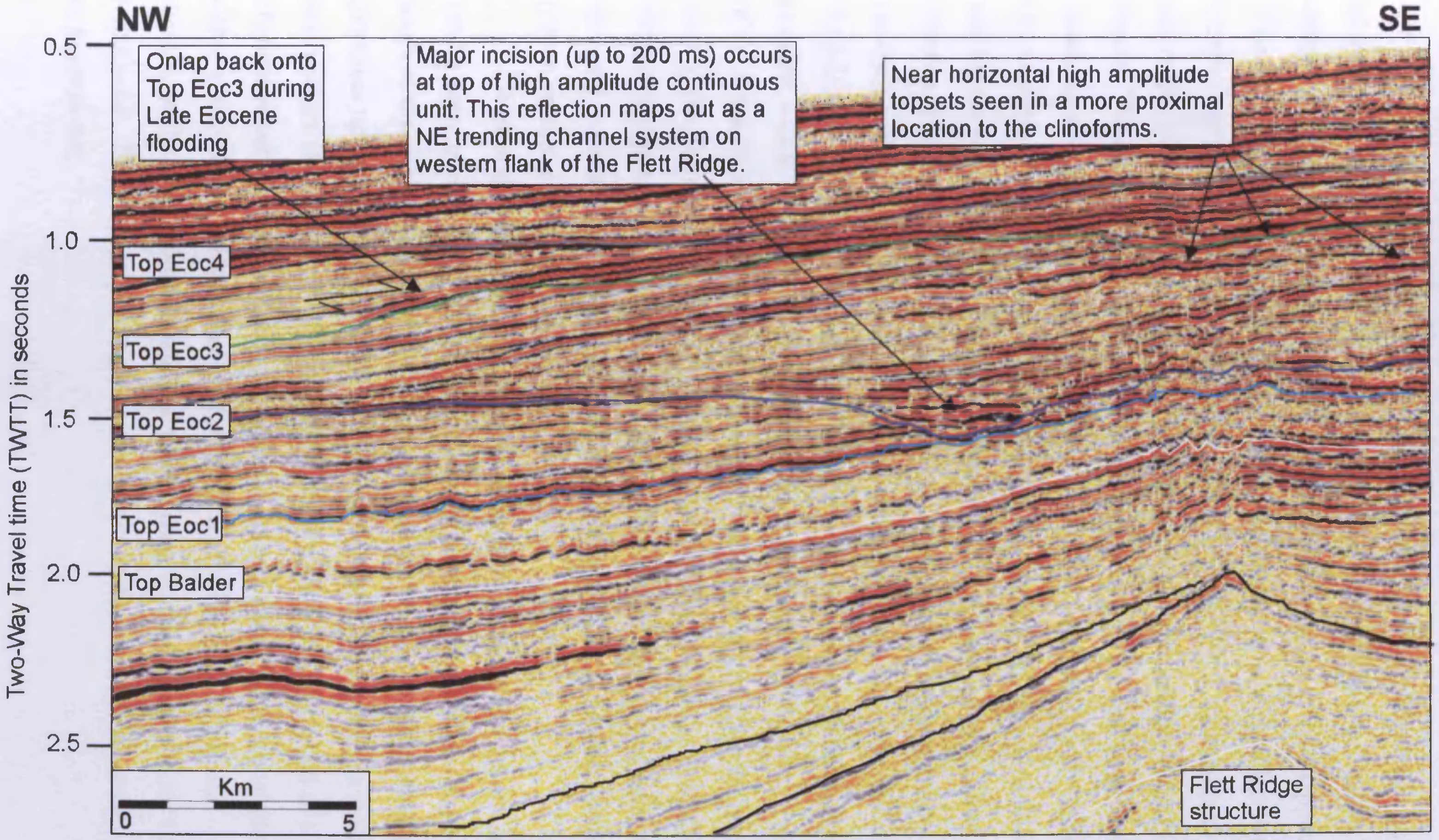


Figure 6.15. Southeast - northwest trending 3-D seismic line showing position of a deeply incised channel system in the early Middle Eocene (Eocene 2 seismic-stratigraphic unit) on the northwestern flank of the Flett Ridge. When mapped out, this incisional feature trends perpendicular to the shelf edge and bends around the palaeo-structure of the Flett Ridge (see Figure 6.16). The incisional feature is filled with higher amplitude reflection configurations which is then downlapped to the northwest by a series of dipping reflection configurations that are interpreted as clinoforms that developed during a renewed highstand in relative sea-level. See Figure 6.16 for line location.

However, a major incision surface with over 100 ms of erosional relief is recognised at the top of the continuous high amplitude parallel reflection configurations and recognised by erosional truncation of internal reflections (**Figure 6.15**). This 100 ms of incisional relief is an average for the channel and equates to approximately the same in metres at this relatively shallow depth. When this erosive pattern is mapped around the 3-D survey it appears to be present only on the northwest flank of the Flett Ridge and trends broadly northeast – southwest (**Figures 6.15** and **6.16**). When correlated to well 205/9-1, this major incision event occurs at the top of the sand-rich package of the Upper Judd Sands. The detailed mapping of the erosive feature shows a broad channel-like feature that trends parallel to the shelf edge. This channel system is approximately 2.5 km wide and can be seen on the amplitude attribute map of this horizon by the distribution of higher amplitudes seen within the axis of the channel (**Figure 6.16**). The width of the channel narrows to less than 1 km in the north. It can be seen to be trending north - northeast along the shelf and then turn abruptly downslope to the north and west near the northern limit of the 3-D seismic survey (**Figure 6.16**). Additionally, there may be some smaller channel systems, which appear to be tributaries to this main channel and these are seen to the east of the main channel feeding into its axis and are being sourced from the area above the crest of the Flett Ridge. Similarly, west of the main channel axis, and southwest of well 205/9-1 a high to moderate amplitude area trends northwest and broadens out to cover an area of 4 – 5 km. This feature has been interpreted as a channel system, though no incision is seen on the reflection.

Within the northeast – southwest trending channels system the amount of incision can be seen to vary along its length. Incision of approximately 80 m can be seen in the south of the survey (in a more proximal position with regard to the shelf edge) and increases up to a maximum of 200 m in the north where the channel turns towards the basin centre (**Figure 6.16**). This erosional feature has been interpreted as a sub-marine channel which has a deeper incised canyon at its northern end. It is postulated here that either a lowering in relative sea-level over the shelf triggered the incision of this channel system, or that it was fault-controlled, though there is a little evidence for the latter hypothesis. The stratigraphic significance of this feature will be discussed in greater detail in **Section 6.5**.

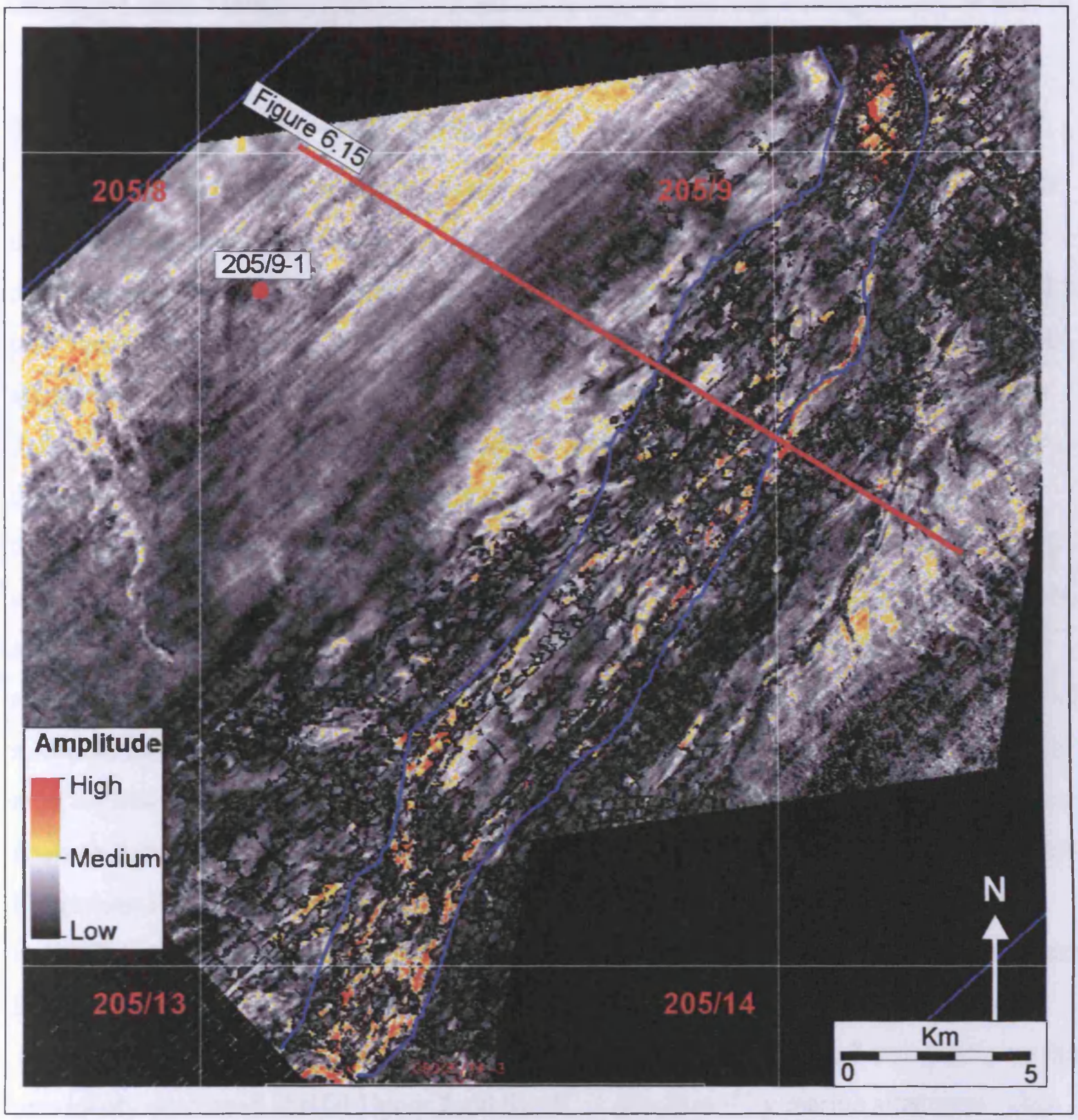


Figure 6.16. Amplitude extraction map of early Middle Eocene channels system. The channel can be mapped across the entire 3-D survey and forms an elongate 1 - 2 km wide belt that runs parallel to the shelf edge in a broadly northeast-southwest trend. Along the length of the channel system there are areas where the amplitude changes and there appears to be a rather patchy appearance to the amplitude of the channel fill. Generally, there is a higher acoustic amplitude value in the southern part of the channel system, and also to the very north of the 3-D survey where the channel narrows to approximately 1 km and the incision deepens. Outwith of the channel axis, the general amplitude is low with occasional higher amplitude areas depicting possible shelf perpendicular channel features and some tributaries to the main channel are possibly interpreted. The location of the seismic panel in Figure 6.15 is shown.

6.4.7 Middle Eocene - Mid Lutetian

After the incision and development of the channel system during the early part of the Middle Eocene (see **Section 6.4.6**) a thick interval of seismic reflections is seen to downlap over the entire sandy unit including the channel system itself (**Figure 6.15**). This package is dominated by reflections that are moderate to high amplitude, continuous and are seen to dip and downlap towards the west and northwest. The top of this package is defined by a high amplitude reflector which is termed Top Eoc3 (dark green horizon), and the base of the package is Top Eoc2. Thus it represents the Eocene 3 seismic unit first described in **Chapter 4** (see **Section 4.3.4**). The dip of these reflections in this Eocene 3 unit are seen to shallow to the east where they become near horizontal and often higher in amplitude **(Figure 6.15)**. From seismic observations these reflections which downlap towards the west and northwest are interpreted to represent clinoforms that formed part of a Middle Eocene delta that was sourced from the Shetland Platform to the southeast. The near horizontal parts of the reflection are interpreted here to represent topsets and may be coals or delta top sands. The 205/10-2b well which is located near to the crest of the Flett Ridge (see **Figure 6.1**) shows a higher proportion of sand than the more distal 205/9-1 well. This would be an expected observation in deltaic systems and the well data conforms to this idea of sourcing from the southeast.

From correlation with well 205/9-1 the lithology of he Eocene 3 unit overlying the previously discussed shelfal Upper Judd Sands, is dominated by marine siltstones. This suggests that after incision into the shelf during a relative sea-level fall, the sea-level rose again and pushed facies belts back landwards during the subsequent transgression. The shelf was flooded during this transgressive episode before progradation of clinoforms responded during a relative sea-level high. The location of well 205/9-1 is found in a basinal setting about 15 – 20 km from the crest of the Flett Ridge. The silts found here may represent the distal equivalents (e.g. bottom-sets) to a more sand-rich succession in the proximal location, on the foresets and topsets of the clinoform. The biostratigraphic data of these silts has been interpreted as a marine outer shelf to upper bathyal setting.

The direction of clinoform progradation is from the southeast and resulted from a period of relative sea-level rise which is interpreted to have been caused either by subsidence on the margin or by a eustatic sea-level rise. This rise in relative sea-level resulted in the creation of significant accommodation space which filled by both aggradation and progradation of deltaic systems. The clinoforms of this deltaic succession have relatively steep slope angles of approximately 1.5 degrees, though this is not corrected for compaction of the sediments. The clinoform relief is also significant and from measurements of position of the topset (which at present day has little or no rotation) to the position of the most basinward downlap a rough estimation of the water depths can be ascertained. Using this basin method it is believed that the delta prograded into water depths that were up to a maximum 400 m at the base of slope during the Middle Eocene.

6.4.8 End of the Middle Eocene - Bartonian

At the top of the Middle Eocene package (Eocene 3 unit), defined by the Top Eoc3 reflection (dark green horizon) there is visible incision seen in the region of the near horizontal topset reflections (see **Figures 6.15** and **6.17**). This incision is not as obvious as the larger scale early Middle Eocene incision discussed above (**Section 6.4.6**) and looks to be of much smaller scale (both horizontally and vertically). This second phase of pronounced incision occurred into delta top deposits near the end of the Middle Eocene and it is not a single event. It is seen on the 3-D seismic data as a series of closely spaced incision surfaces, cut into successive delta topsets (**Figure 6.17c**). Locally, these five separate incision surfaces can be identified over a relatively small time interval with incision of only tens of milliseconds rather than hundreds (**Figure 6.17c**). Three examples of the late Middle Eocene incision event are shown with seismic profiles both perpendicular and parallel to the shelf margin (**Figure 6.17**). Detailed mapping of these individual local incision events show complex channel planforms that trend in a northwest - southeast direction (**Figures 6.18** and **6.19**). These channel systems have a low sinuosity, and are generally 500 m - 1000 m in width with low amplitude meanders and braided planforms, though they do occur at a range of scales. The channels are seen to be possibly affected by later east northeast - west southwest trending faults (**Figure 6.18**). These channel systems are located across the whole of the topset region and often further downslope and are found laterally across the majority of the 3-D seismic survey.

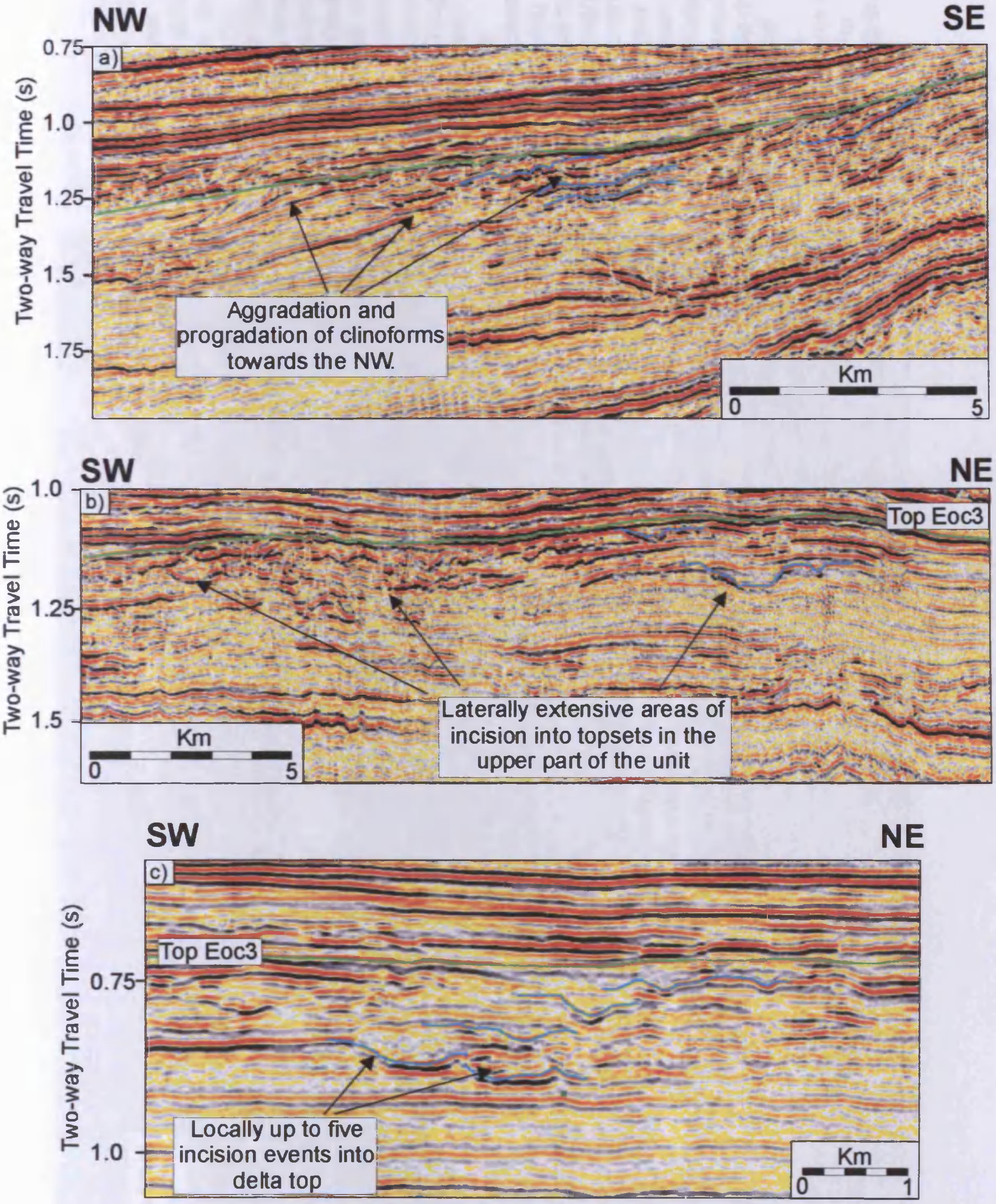


Figure 6.17. 3-D seismic lines showing examples of the localised areas of incision near the top of the Middle Eocene unit (Eocene 3) close to the Top Eoc3 reflection. a) Dip line (southeast - northwest) showing both aggradation and progradation of the delta system and the nature of the local incisions down the length. Notice the position and size of these later incisions in comparison with the earlier lower Middle Eocene channel system below. b) Strike line (southwest - northeast) perpendicular to major incisions in the upper Middle Eocene unit showing their positions laterally on the delta topsets. c) Strike line (southwest - northeast) showing a close up of 5 individual distinct incision events seen. Note the scale of these feature s vary from 50 - 400 m across. For locations of seismic line a) and b) see Figure 6.18 and for c) see Figure 6.19.

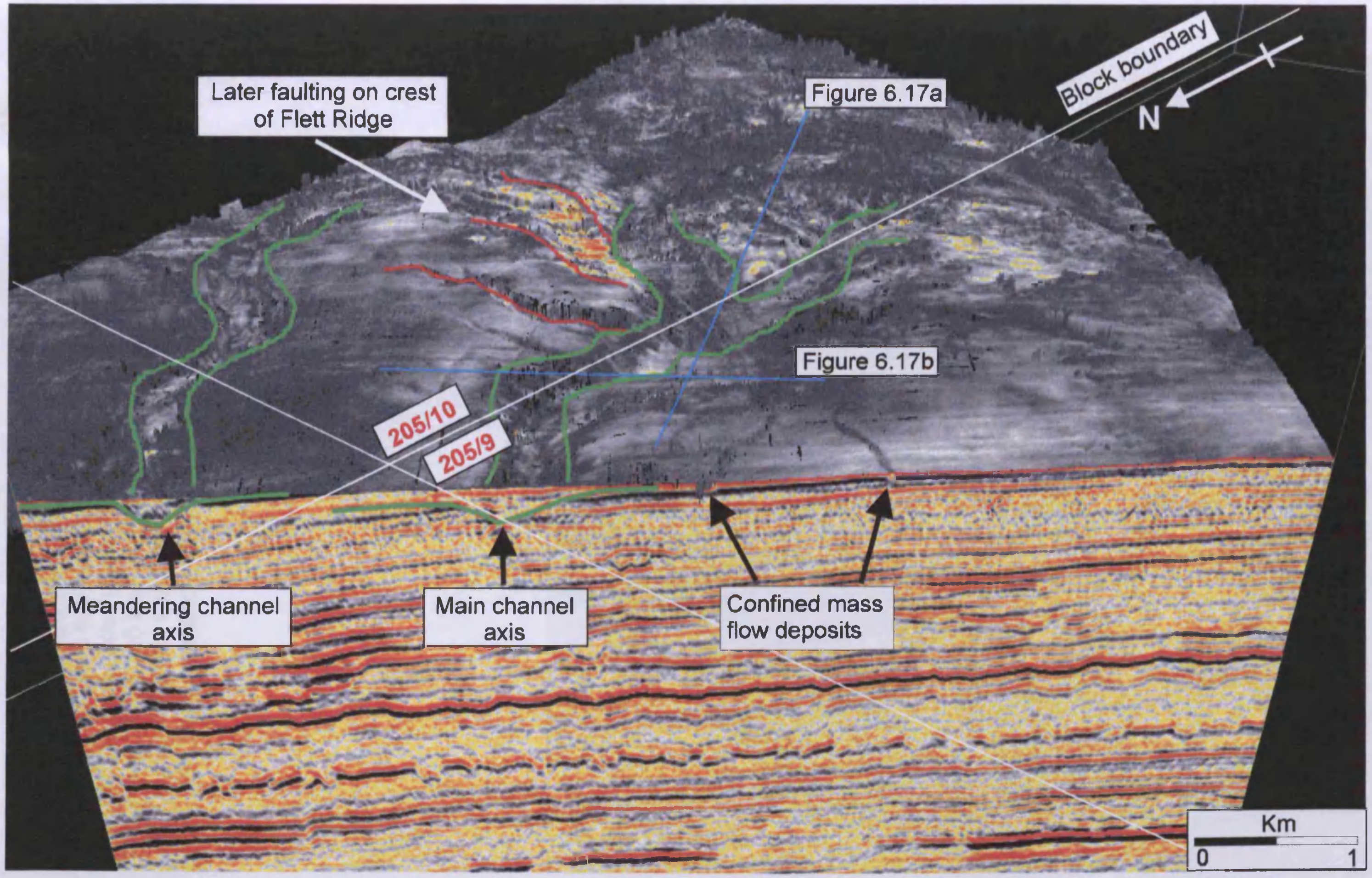


Figure 6.18. 3-D perspective image (looking southeast) of one of the Late Middle Eocene incisions into the delta top. Here the channels seen are picked out in the amplitude map (in dark green) which has been draped over the depth map (in TWTT). On the front a 3-D seismic line shows the cross sectional geometry of the channel systems. In this image the complex and varied geometries of the channels can be seen showing meandering forms and perhaps mass flow or confined channels on a much smaller scale. Broadly north - south trending faults bisect the crest of the Flett Ridge (in the area of high amplitude) though these faults do not seem to control the channel positions. Locations of Figures 6.17a & b are shown. For location of seismic line see Figure 6.20.

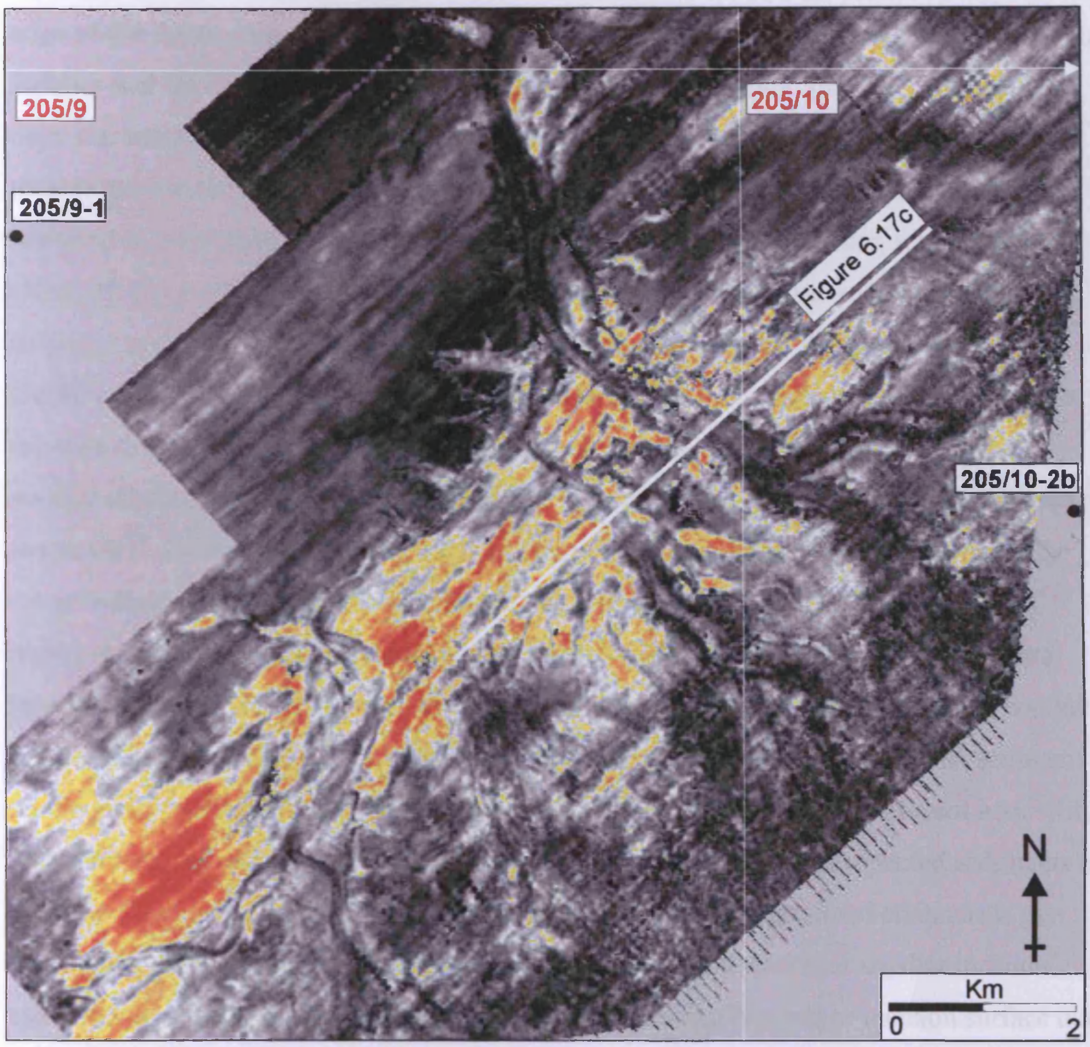


Figure 6.19. Amplitude extraction map of one of the many incision surfaces seen in the Late Middle Eocene in the Flett sub-basin. This amplitude map picks out the low amplitude fill (dark greys to blacks) of the channel systems and shows the complex geometry of the drainage network. The channels are seen to cut into a broad northeast - southwest trending high amplitude zone (red colours) which may reflect the presence of a coal horizon or a shore-face sand on the delta top. Some of the channels are seen to merge down-dip towards the basin centre. The location of seismic line 6.17c is highlighted.

Furthermore, the channels tend to be focussed near the shelf-slope break between the topset and the pro-delta region of the clinoform, suggesting that they were incised into the edge of the delta plain, close to the delta front (**Figure 6.17**). Seismic and well data (wells 205/9-1 and 205/10-2b) shows that the channels cut into a Middle Eocene succession of delta top sands interbedded with silts, coal and limestone fragments, and the incisional geometries are thus interpreted as an expression of a relative lowering of sea-level, and response to base level fall exposing the whole of the delta top. The complex network of channels observed on the delta top and pro-delta slope is believed to represent a modest sediment source entering the basin, but also provide evidence for a discrete focus for clastic input from the Shetland margin. Additionally, two small-scale, confined channels are seen to possibly break out of the main broad channel axis (**Figures 6.20**). These smaller channels are narrow (in the region of 100 m) and are 3 – 4 km long, perpendicular to the shelf. Their upper reaches are narrow and become broader (up to 250 m) distally towards their termination in lobe units. These lobe units have central part which is of higher amplitude than the channels edges (**Figure 6.20**). These particular channels are located on the transition between the topset and the clinoform and could represent conduits for mass flow deposits, which broke out of the main channel axis and deposited sediment on the pro-deltaic clinoform slope. However, they could alternatively represent instability at the delta front and be more indicative of collapse of unstable, uncompacted sediments at the delta front transition. Another plausible idea for these two confined channels is that they occur just below the main incision surface and are thus an earlier smaller incision event that has become incorporated into the amplitude map of a major incision surface due to the vertical seismic resolution of the data.

6.4.9 Latest Eocene - Early Oligocene - Priabonian-Rupelian

After the end of Middle Eocene incision progradational reflection geometries are interpreted to indicate clinoforms that downlap and prograde northwest into the basin draping the incision surface. This renewed progradation is interpreted to have been caused by a relative sea-level rise during the Late Eocene which caused flooding back across the delta top. Onlap back onto the Top Eoc3 reflection is seen in **Figure 6.15.**

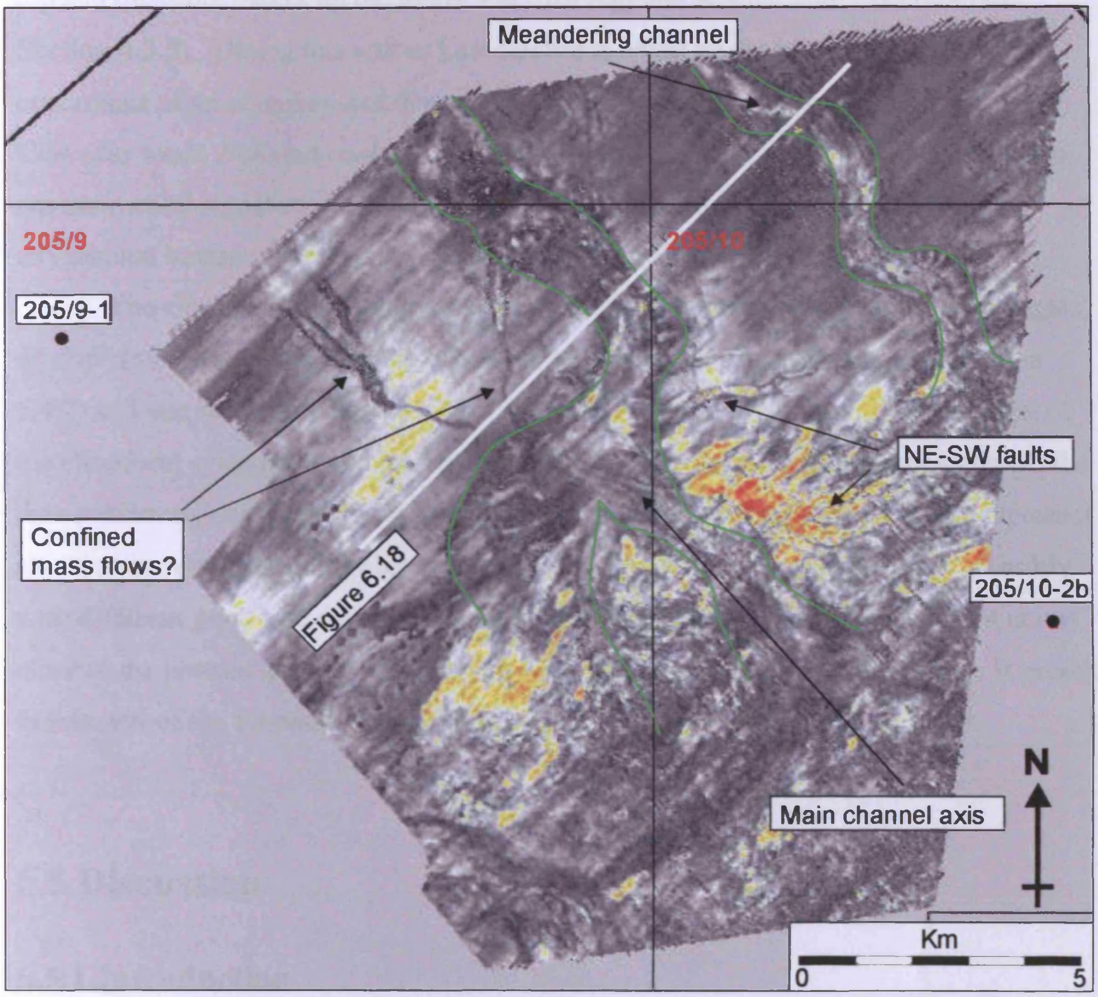


Figure 6.20. Amplitude extraction map of another reflection showing incision in late Middle Eocene (same incision as shown in draped surface in Figure 6.18). This map shows the planform of specific varieties of channels as picked out by the low amplitude colours (greys). Three types of channel type are seen here, meandering in the far north with low sinuosity, a larger main channel in the centre and small confined channels in the northern part of Quadrant 205. Northeast - southwest trending faults are seen across the area above the Flett Ridge structure which also coincides with an area of high amplitude seismic character, which may be lithological or maybe gas escaping from the crest of the Flett Ridge. See text for discussion on these channels. Position of vertical seismic panel shown in Figure 6.18 is shown..

This Upper Eocene interval is defined by the Top Eoc3 reflection at its base and by a high amplitude continuous seismic reflection named Top Eoc4 (dark grey horizon) at its top and therefore makes up the fourth and final regional seismic unit: Eocene 4 (see **Section 4.3.5**). Dating this unit as Late Eocene is based on the recognition on the occurrence of an abundant and diverse benthonic foraminifera assemblage including *Cibicides westi, Bolivina cookie* and *Uvigerina eocaena* from well 205/9-1. The palaeo-environmental signature from this assemblage remained an outer shelf setting with oxygenated bottom waters with some open marine influence.

The clinoforms that prograded during the Late Eocene seem to have lower angles of slope (<1°) than the earlier clinoforms of Middle Eocene age (discussed in **Section 6.4.7**) and can be seen in **Figure 6.15**. This shallower angle and smaller mean height of the clinoform system may represent progradation of the delta into shallower water depths than previously seen, perhaps in the region of 100 m. However, they could also represent a situation in which the delta edge was receiving a lower rate of sediment supply possibly with different grain size distributions (i.e. higher silt to sand ratio). The Eocene 4 unit is close to the present day sea-bed and only a thin post Eocene to recent succession is present in this part of the Faroe-Shetland Basin.

6.5 Discussion

6.5.1 Introduction

This section will summarise the results from **Section 6.4** and construct a Palaeogene basin evolution and outline the major controls on the two incision events in the Middle Eocene. It will continue with a discussion on the role of palaeo-bathymetry on channel development. The section will end with questioning the origin of the observed relative sea-level fluctuations and highlighting whether they were caused by local tectonic factors or by a eustatic influence.

6.5.2 Controls on Middle Eocene Channel Development.

Here, a discussion will follow on the possible controls on the channel systems seen in the Middle Eocene succession (Eocene 3 seismic unit) of the Flett Ridge area. From the observations of the two major incision events discussed above (**Sections 6.4.6** and **6.4.8**) it is clear that drainage networks developed at these times of relative sea-level fall and reveal varying characteristics giving rise to a significantly different scales of channel system and varying planforms.

Taking the early Middle Eocene channel system first, it can be observed that the course the channel system takes trends in a northeast – southwest direction. When the amplitude map of this channel system is superimposed onto a contour map of the base Tertiary horizon which highlights the localised highs above the Flett Ridge (see **Figure 6.7**) many interesting correlations can be observed (**Figure 6.21**). Firstly, this early Middle Eocene aged channel axis is seen basinward (west) of the localised highs above the ridge crest. By following the ridge contours the channel is seen to run parallel to the palaeo-shelf edge (**Figure 6.21**). There is also evidence for small channel features (or tributaries) diverging off the ridge crest which may link to the main channel axis to the west, though also may drain off to the east away from the main channel. This may indicate that the localised highs above the ridge crest could have been a topographically high area and provided sediment from a minor source area into or away from the main channel. At the northeast limit of the ridge this early Middle Eocene channel is seen to bend sharply basinward and become much more deeply incised and canyon-like. It may be a coincidence that there is a pronounced bend in the channel system, though it is an interesting observation that this directional change and deepening of the channel appears at the northern limit of the localised highs (**Figure 6.21**). The planform of the early Middle Eocene channel shows that it is diverted around the highs above the Flett Ridge structure and indicates that these highs and therefore possibly the ridge had some positive topographic or structural relief at this time. Terrestrial river systems are known to cut downwards and incise into structurally elevated features during base level fall if the rate of erosion matches or exceeds the rate of any uplift (Jackson *et al.* 1996; Burbank *et al.* 1996; Humphrey and Konrad 2000). In contrast, channels in a wholly marine environment would be more likely to trace and follow the contours of a submarine bathymetric high. It

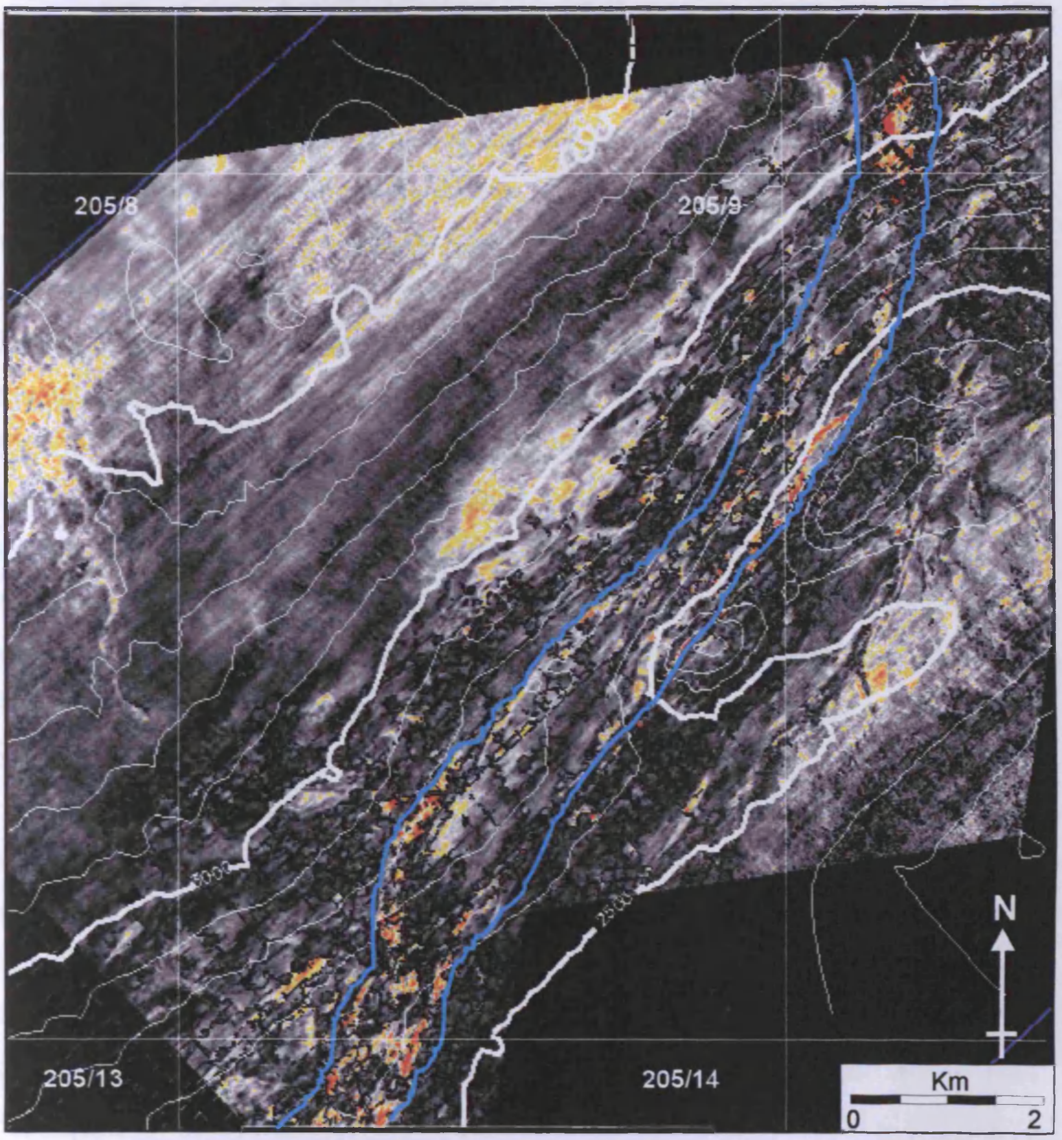


Figure 6.21. Amplitude extraction map of the early Middle Eocene channel that trends northeast - southwest overlain by contours of the near base Tertiary reflection showing the location of the two crestal highs which sit directly above the deeper Flett Ridge. Note that this channel system sits on the western flank of these highs and seems to follow the contours nicely. It is believed that this channel is parallel to the shelf break at the time and is influenced by the topography of the two crestal highs. North of the highs the channel swings westwards and becomes significantly deeper and may even become a canyon feature in the shelf break at the northern limit of the 3-D seismic survey. Possible smaller channels are seen on the eastern side of the crestal feature and this may be another drainage network sourced from this localised high.

is therefore inferred that the incision in this early Middle Eocene example and was not the result of fluvial erosion and more likely to have occurred entirely in the marine realm.

The late Middle Eocene channel morphologies show a markedly different pattern of channel networks to the early channel system discussed above. Furthermore, these channels, seen near to the top of the Eocene 3 unit (topped by the dark green Top Eoc3 horizon) have smaller dimensions than the broader early Middle Eocene channel. By superimposing these later channels on the contour map of the base Tertiary horizon the channels can be seen to traverse directly over the local highs above the ridge crest. Two examples from localised incision surfaces from the late Middle Eocene are shown in **Figures 6.22** and **6.23**. Both channel networks are seen to run perpendicular to the palaeo shelf-edge. The highs above the ridge crest seem to have had no effect or topographic control on the channel architecture and pattern (**Figures 6.22** and **6.23**). The earlier of the two channels shown in **Figure 6.22** channel bifurcates to the southeast of the ridge highs and becomes one main channel on the basinwards side of the two high crests of the ridge. An area outwith the channel axis, (possibly an interfluve or coaly floodplain based on its high amplitude seismic response) sits directly over the small saddle feature between the ridge crests (**Figure 6.22**). Additionally this may be interpreted as shallow gas which has escaped up faults which form from the crest of the Flett Ridge structure. The architecture and spatial variation of these late Middle Eocene channels and their relationship to the underlying highs over the Flett Ridge suggests that the highs were no longer a significant topographic feature, with the sediment dispersal not being influenced by any remnant structural grain. Instead, the channel pattern is interpreted to have incised as a geomorphic response to a fall in base-level on the delta top, without structural control of depositional or erosional topography. Furthermore, obscure confined channels exist and are seen to break–out from the main channel axis and head down-dip towards the basin centre (**Figure 6.22**). However these confined channels terminate after a short distance (2 - 3 km) and may represent a turbulent mass flow system that formed at the front of the delta. A summary of the change in channel morphology and the controls on their distribution with respect to the Flett Ridge can be seen in **Figure 6.24**.

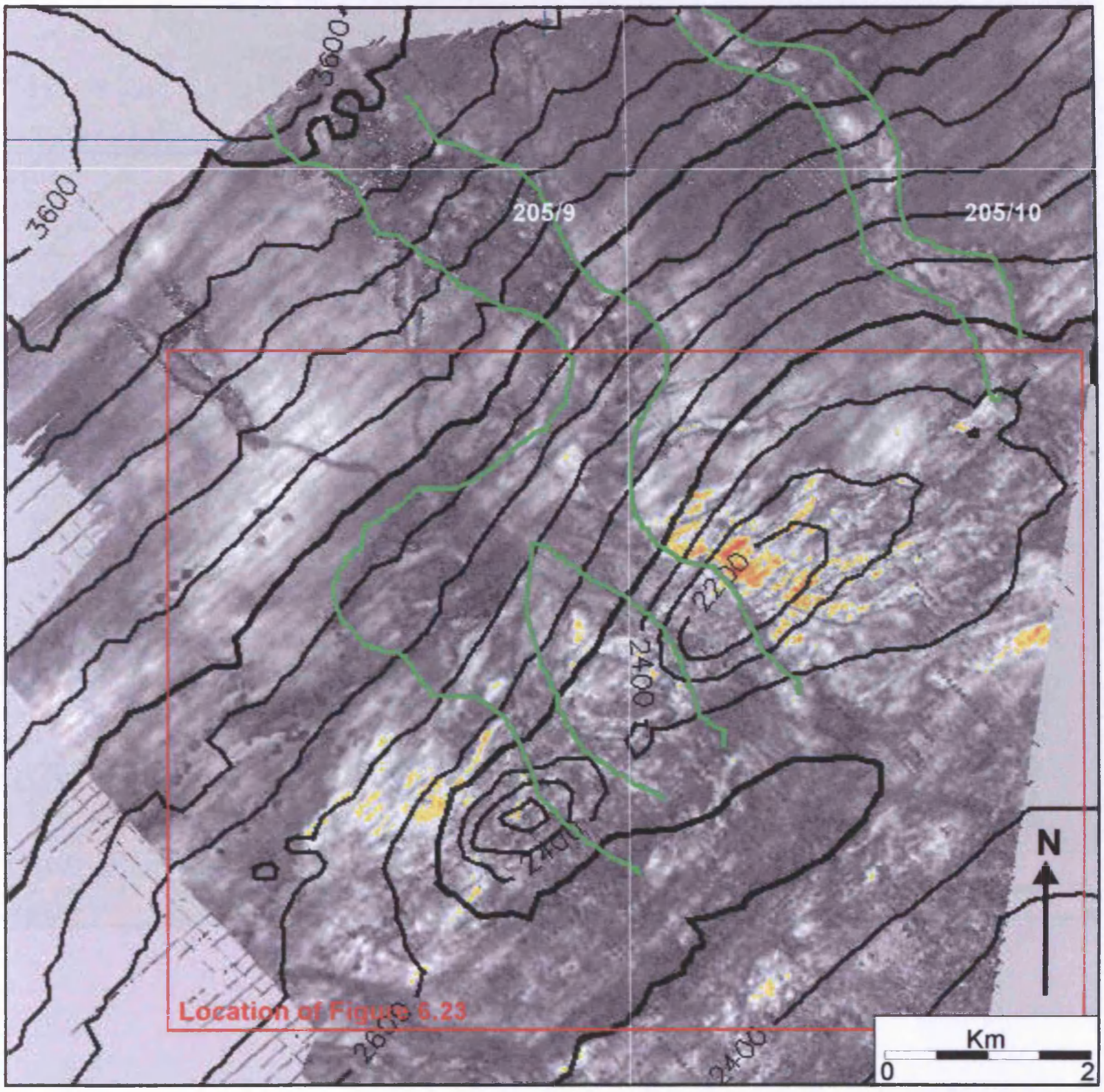


Figure 6.22. Map of amplitude extraction of a late Middle Eocene channel systems (highlighted in green) as seen cutting into delta top clinoforms, overlain on the by contours of the near base Tertiary reflection showing the location of the two crestal highs which sit directly above the deeper Flett Ridge. Note these channels are seen to directly cut over the highs above the Flett Ridge and are not controlled by the topography whatsoever. These channel systems are both smaller in scale and are perpendicular to the shelf break at the time. It is interpreted that the local crestal highs above the Flett Ridge had no topographic or bathymetric control on these channels in the late Middle Eocene. The position of Figure 6.23 is also highlighted.

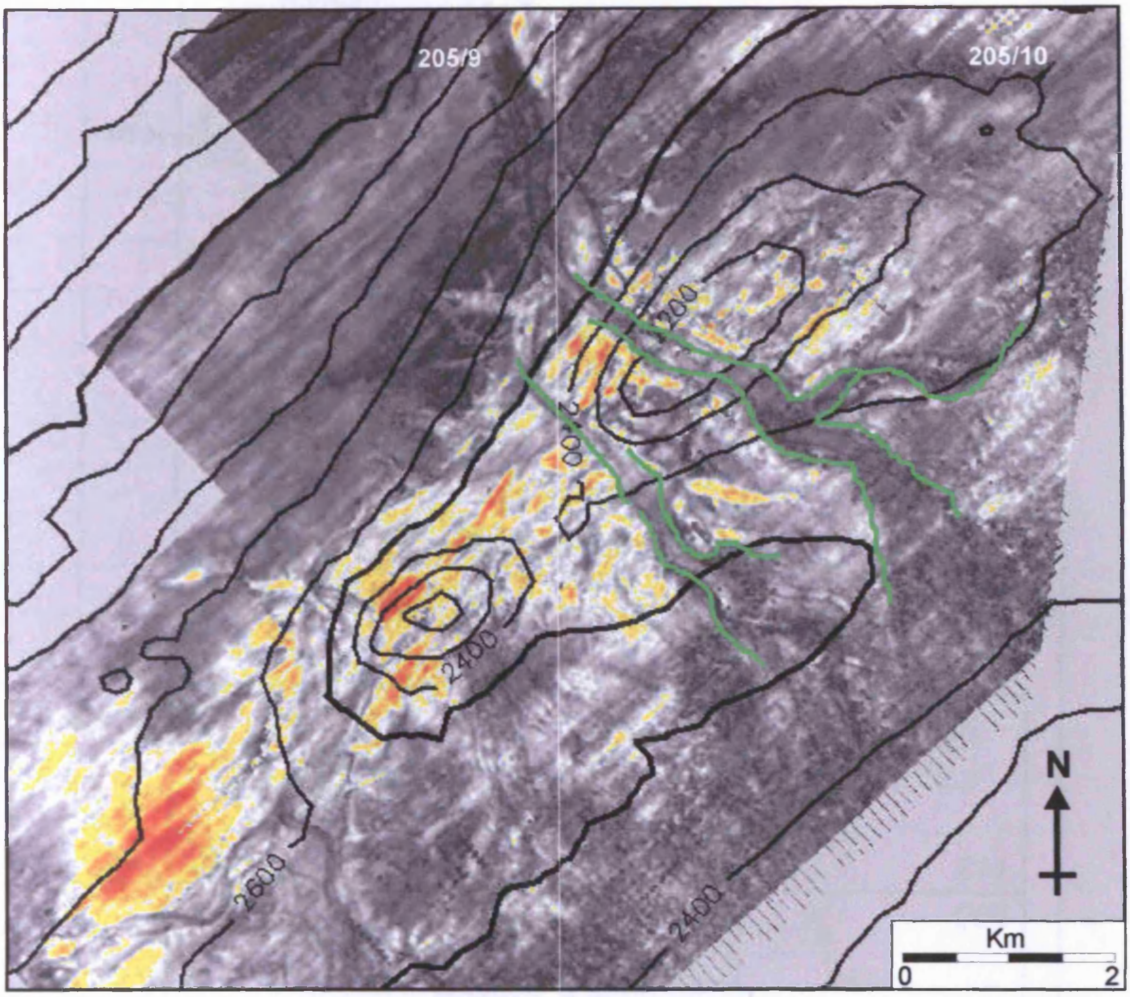


Figure 6.23. Further example of an amplitude extraction from an incisional surface of the late Middle Eocene channel systems overlain on the by contours of the near base Tertiary reflection. As with Figure 6.22 these channels are seen to also cut over the localised highs suggesting no topographic control on these channel systems. Note however the change in planform of these channels and how they differ from the channels shown in Figure 6.22. This change in planform occurs over a very small amount of time during the late Middle Eocene and may indicate variations in sediment supply or the hinterland.

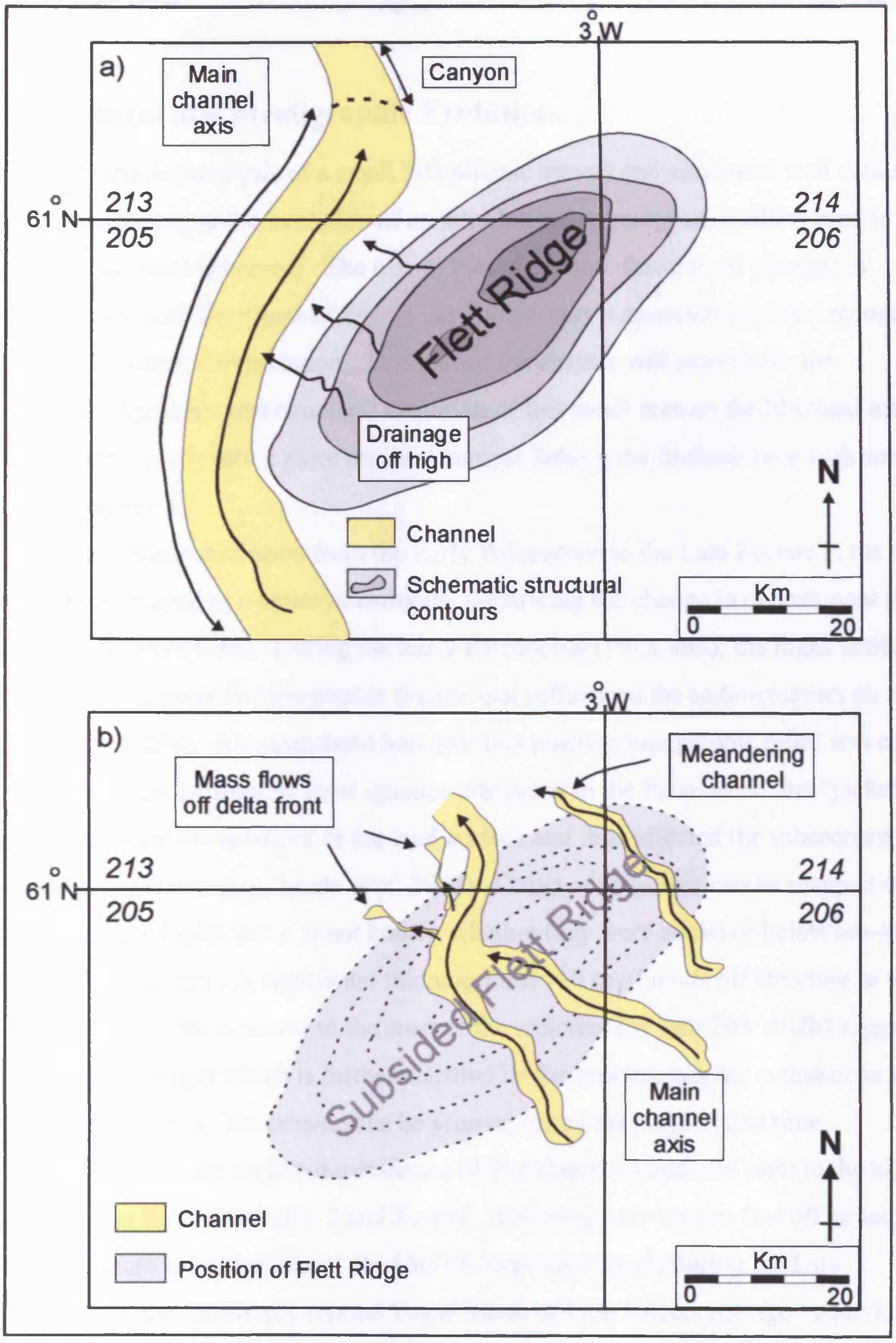


Figure 6.24. Schematic cartoon showing the geometry of the early and late Middle Eocene channels systems and there relationship to the underlying highs in the near base Tertiary and the underlying Flett Ridge. a) The early Eocene channel system flows around the feature and is controlled by the topography on the local highs. b) By the late Middle Eocene the highs show no topographic control on sediment dispersal and the channels cut directly over the palaeo Flett Ridge, and solely developing due to relative sea-level fall creating incision and bypass on a delta top (see text).

6.5.3 Structural and Stratigraphic Evolution.

From detailed analysis of a small 3-D seismic survey and additional well data a comprehensive stratigraphic evolution of an intra-basinal to marginal, shallow marine shelf environment has been achieved. The role of buried tectonic features on changes in relative sea-level and the effects of this on the sedimentary succession has been recognised form detailed seismic interpretation. This part of the chapter will summarise the Palaeogene stratigraphic and structural evolution of this small area on the Shetland margin and put this local study into a more regional context linking the findings here with some results of **Chapter 4**.

The schematic evolution from the Early Palaeocene to the Late Eocene in the case study area is displayed as a series of cartoons, illustrating the change in depositional style through time (**Figure 6.25**). During the Early Palaeocene (Pal 1 unit), the highs above the Flett Ridge were a positive topographic feature that influenced the sedimentation on its flanks (**Figure 6.25a**). It is postulated here that this positive topographic relief was caused by the emplacement of shallow level igneous intrusions in the Palaeocene that "jacked up" local areas of either the sea-floor or the land surface and thus affected the subsequent sedimentary succession (e.g. Trude *et al.* 2003). Onlap of reflectors can be mapped onto the flanks of these highs, but it is not known whether they were above or below sea-level in the Early Palaeocene. A significant thinning from 110 m of muds off structure in well 205/9-1 to less than 40 m nearer to the crest of the structure (in well 205/10-2b) suggests a more sub-marine origin which is further clarified by the biostratigraphic signature in these argillaceous sediments. No erosion can be seen on top of the ridge at this time. Additionally, there is no supporting evidence of downlap of clinoforms onto to the highs ridge in the Late Palaeocene (Pal 2 and 3 units). However, it is known that off structure in well 205/9-1 there were significant deposits of deep-water sands during the Late Palaeocene (e.g. the informally termed Vagar Sands of Late Palaeocene age - equivalent to the T35 unit of Ebdon *et al.* (1995). At this time however, sediment was deposited over the crest of the ridge highs and this indicates that there was a sufficient increase in water depths by subsidence or eustasy to allow the accumulation of significant sediment on the

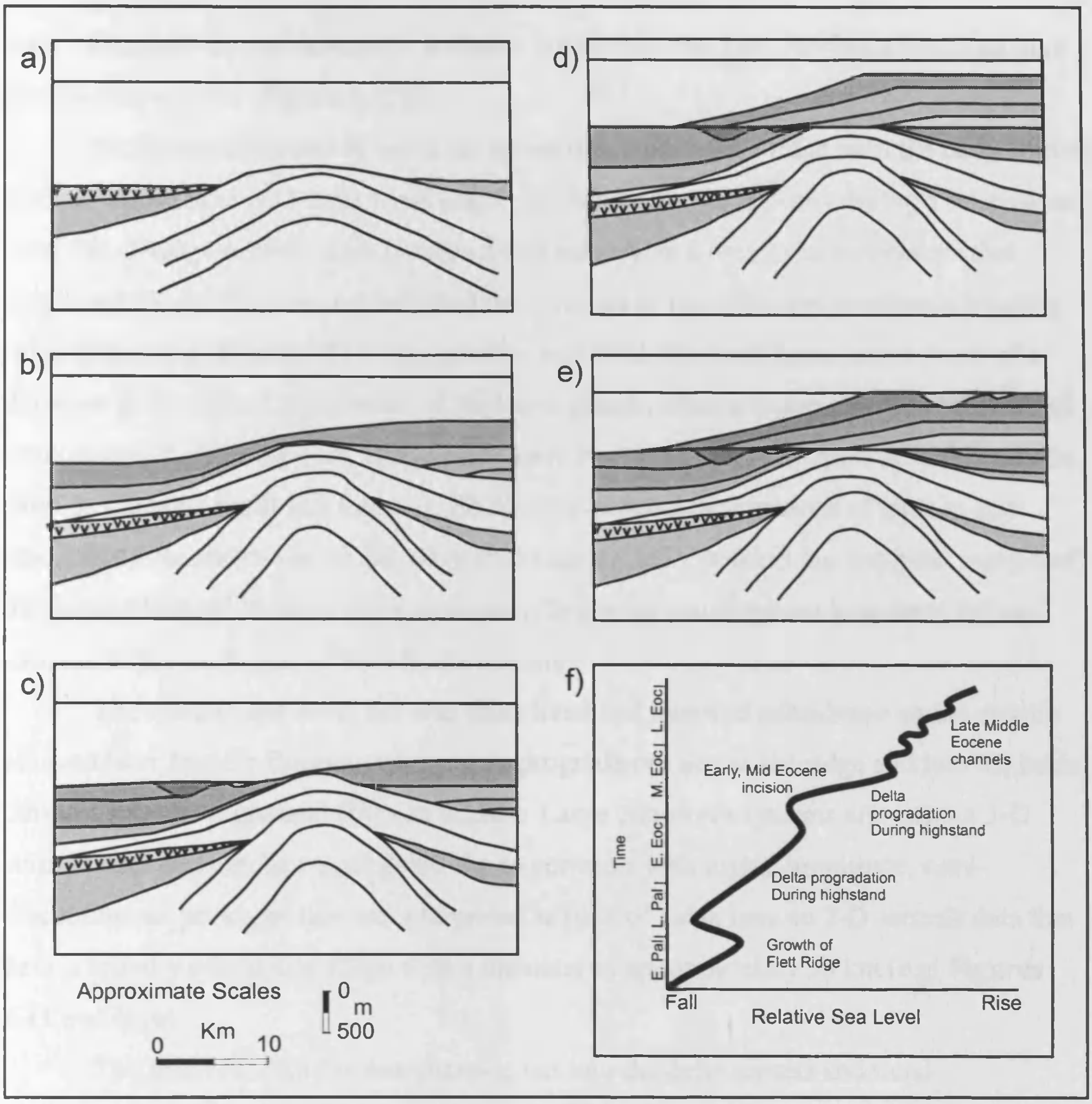


Figure 6.25. Schematic cartoons showing stratigraphic summary throughout the Palaeogene. a) Early Palaeocene - thinning and onlap onto near base Tertiary highs. b) Late Palaeocene-Early Eocene delta development and draping of highs. c) Major incision during the early Middle Eocene parallel to shelf. d) Renewed delta progradation during the Middle Eocene. e) Late Middle Eocene incision into delta tops incising channels that run perpendicular to the shelf. f) Schematic relative sea-level curve for Flett Ridge area throughout the Palaeogene. The sea-level was believed to be relatively deep in the Early Palaeocene with the deposition of the turbiditic Vagar Sands. This then gave way to deltaic deposition in the Late Palaeocene which continued into the Early Eocene. At the start of the Middle Eocene a relative sea-level fall created a channel system which was short-lived and subsequent delta deposition renewed. Near the end of the Middle Eocene small falls in relative sea-level created incisions in the area of the Flett Ridge structure that led to bypass of sediment that fed base of slope fans to the northwest of the 3-D survey. For greater detail on channel development and the stratigraphic evolution of the Flett Ridge area see section 6.4 and for a comparison of the relative sea-level curve with the eustatic sea-level and the spatial and temporal positions of sediments in the Flett sub-basin see the chronostratigraphic chart in Figure 6.27.

crest. Alternatively an increase in sediment supply into the Flett sub-basin could account for this observation (**Figure 6.25b**).

Sediment continued to cover the submarine highs of the ridge until the early Middle Eocene times (Eocene 3 unit) when major incision occurred, possibly during a relative sea-level fall (**Figure 6.25c**). This incision event resulted in a broad channel system that paralleled the shelf-break and followed the contours of the northeast -southwest trending ridge (**Figures 6.16** and **6.21**). Any relative sea-level fall could have been a result of a decrease in the rate of subsidence on the basin margin when combined with a eustatic fall or more specifically the Flett Ridge could have undergone relative uplift or faulting in the form of footwall uplift and rotation. Shallow level igneous intrusions of plutons and associated sills and dykes on the ridge could have locally uplifted the southeast margin of the Faroe-Shetland Basin relative to surrounding areas causing local base level fall and incision in the early part of the Middle Eocene.

The relative sea-level fall was short lived and renewed subsidence on the margin allowed later Middle Eocene sediments to prograde out across the ridge and into the basin during a relative highstand (**Figure 6.25d**). Large clinoform systems are seen on 3-D seismic data and can be traced down-dip to correlate with higher amplitude, semi-discontinuous packages that are interpreted as base of slope fans on 2-D seismic data that have a broadly concentric shape with a diameter of approximately 30 km (e.g. **Figures 6.11** and **6.26**).

The latest Middle Eocene channels cut into the delta topsets and trend perpendicular to the shelf-break traversing the Flett Ridge. These channels are both braided and meandering and are found across the whole of the delta top at this level (**Figure 6.25e**). There is a high frequency of incision events at this stage, with erosional truncation seen on virtually every seismic reflection (or loop) (**Figure 6.17c**). This can be interpreted as either small scale oscillations in the eustatic sea-level over a relatively short time period, or by minor changes in the rate of discharge and sediment supply coming off the hinterland.

An additional cause of these high frequency relative sea-level falls could be localised tectonic uplift of the Flett Ridge. These later channel systems are interpreted as subaerial because they cut into successive delta topsets, feed base of slope fans, and are not

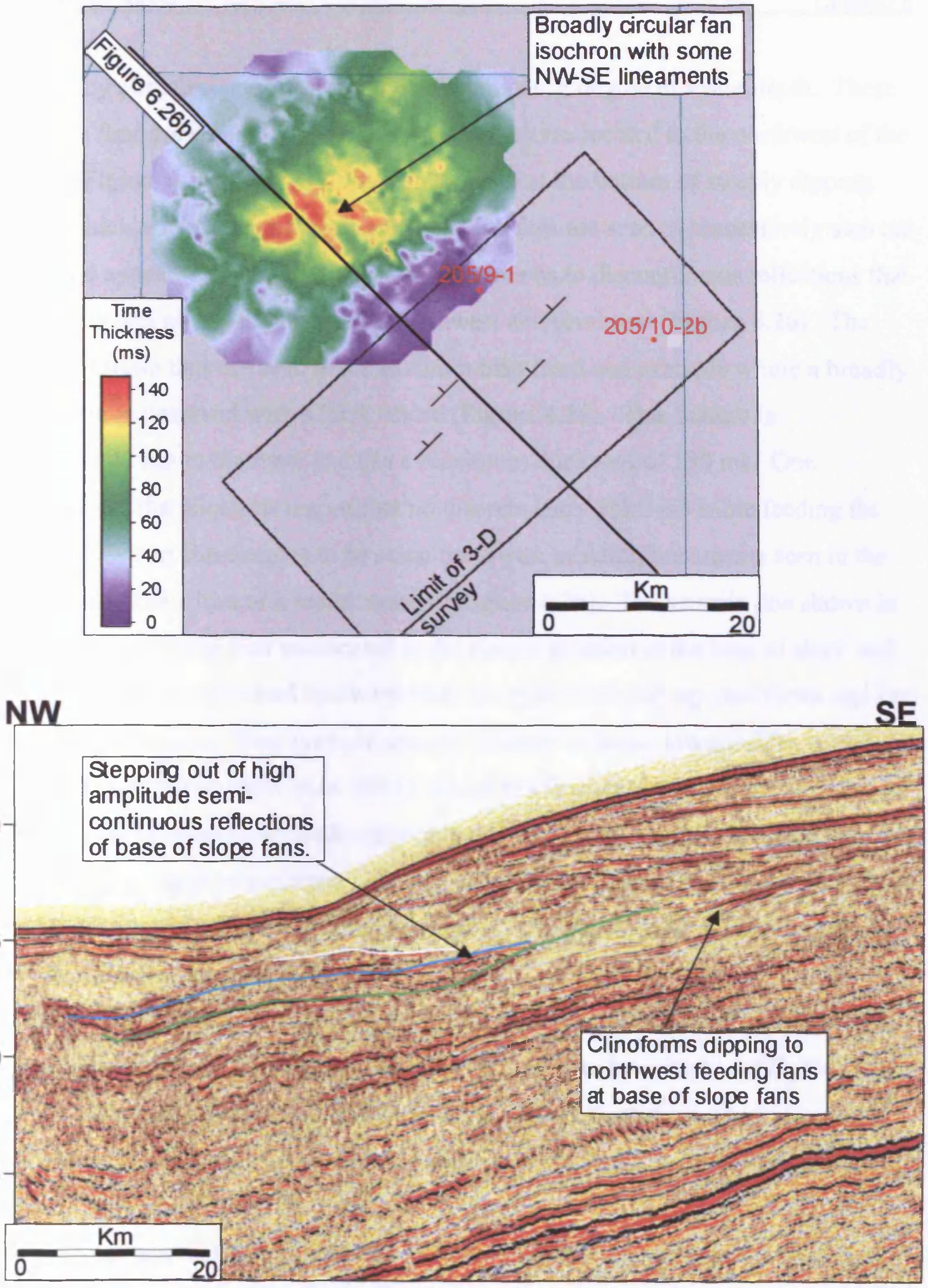


Figure 6.26. a) Isochron map of a base of slope fan found northwest of the 3-D survey. Red colours show thickest part of the fan in the centre of a broadly circular body. The dip symbols indicate the direction of clinoform dip. b) Northwest southeast trending 2-D seismic line through the centre of the fan that is shown in the isochron map above (top = blue, base = green). The fans are seen to step out into the basin centre (see subsequent fan with white horizon top) and are highlighted by discrete packages of high amplitude semi-continuous to discontinuous seismic reflections. Location of seismic line is seen in the isochron map (a) above.

downlapped by clinoform systems which would indicate a degree of water depth. These base-of-slope fans are well imaged on 2-D data as they are located to the northwest of the 3-D survey (**Figure 4.26**). The fans are seen to occur at the bottom of steeply dipping reflections which are interpreted as clinoforms. The fans are seen to successively step out into basin and appear as high amplitude semi-continuous to discontinuous reflections that are seen to thin and pinch-out both to the northwest and southeast (**Figure 4.26**). The dimensions of these fans are seen in the isochron map from one example where a broadly circular feature is observed with a thick centre (**Figure 4.26**). This feature is approximately 40 km in diameter and has a maximum thickness of 150 ms. One observation from the thickness map is that no discrete entry point is visible feeding the circular fan, however there seems to be some northwest trending lineaments seen in the isochron and may be a hint of a feeder system (**Figure 4.26**). The seismic line shown in **Figure 4.26**, shows these fans are located in the classic position at the base of slope and seismic reflections can be traced landward from the base of the fan up clinoforms and into higher amplitude topsets. This example appears to show a classic lowstand fan of the Exxon model (e.g. Posamentier *et al.* 1988), where the fans are deposited during a fall in relative sea-level creating an unconformity on the shelf. In this scenario a certain amount of sediment bypass may be expected. The incisions seen in the late Middle Eocene on the delta top maybe represent a significant unconformity caused by small relative sea-level falls and exposure of the shelf.

However, uplift and compression of the Faroe-Shetland Basin during the Middle - Late Eocene and throughout the Oligo-Miocene (Boldreel and Andersen 1993; Davies *et al.* 2004) could explain the fall in relative sea-level, though eustatic variation cannot be ruled out especially since global sea-level is believed to have fallen by 50 - 100 m during the Eocene – Miocene (Haq *et al.* 1988).

From observations and interpretations, a relative sea-level history for the Palaeogene has been erected for the southeast margin of the Faroe-Shetland Basin (**Figure 6.25** and Robinson *et al.* 2004). This assumes that both channel systems caused by the incision events were a result of sea-level fall, the first of which was probably not sufficient to expose the shelf.

The first shelf margin-parallel channel system (seen in the early Middle Eocene) is therefore interpreted here as a sub-marine channel which may have been additionally associated with faulting and along-shelf bottom currents. The late middle Eocene channels are interpreted as sub-aerially incised valleys, based on the evidence highlighted above. A schematic chronostratigraphic chart of the Flett sub-basin highlights the spatial and temporal position of the Palaeogene sediments of the study area (**Figure 6.27**). However, as with the chronostratigraphic charts shown in **Chapters 4** and **5**, the biostratigraphic resolution of the Eocene succession of the Faroe-Shetland Basin is generally not of a good quality. Thus accurate absolute age dates for the Eocene succession are unobtainable from the biostratigraphical analysis. Therefore the error bars on the crude relative dates obtained are potentially great. The significance of this is that there is no worthwhile correlation of the relative sea-level curve constructed for the Flett sub-basin with the eustatic sea-level curve of Haq *et al.* (1988). Indeed, when the two curves are compared (**Figure 6.27**) no semblance of correlation occurs and this is to be expected when considering the available age data.

This is not to say that either the relative sea-level is incorrect, or that there are errors in the eustatic curve, it is just that such a disparity would be expected between any relative sea-level curve from any basin and the standard global curve. This is because of local variations in tectonics, sediment supply and climate which all influence the relative sea-level and thus cause significant diversions from the global signature. How much of a diversion these local factors contribute to the recorded change in sea-level remains questionable, but so much so that the eustatic signal may be completely suppressed or enhanced depending on basin conditions.

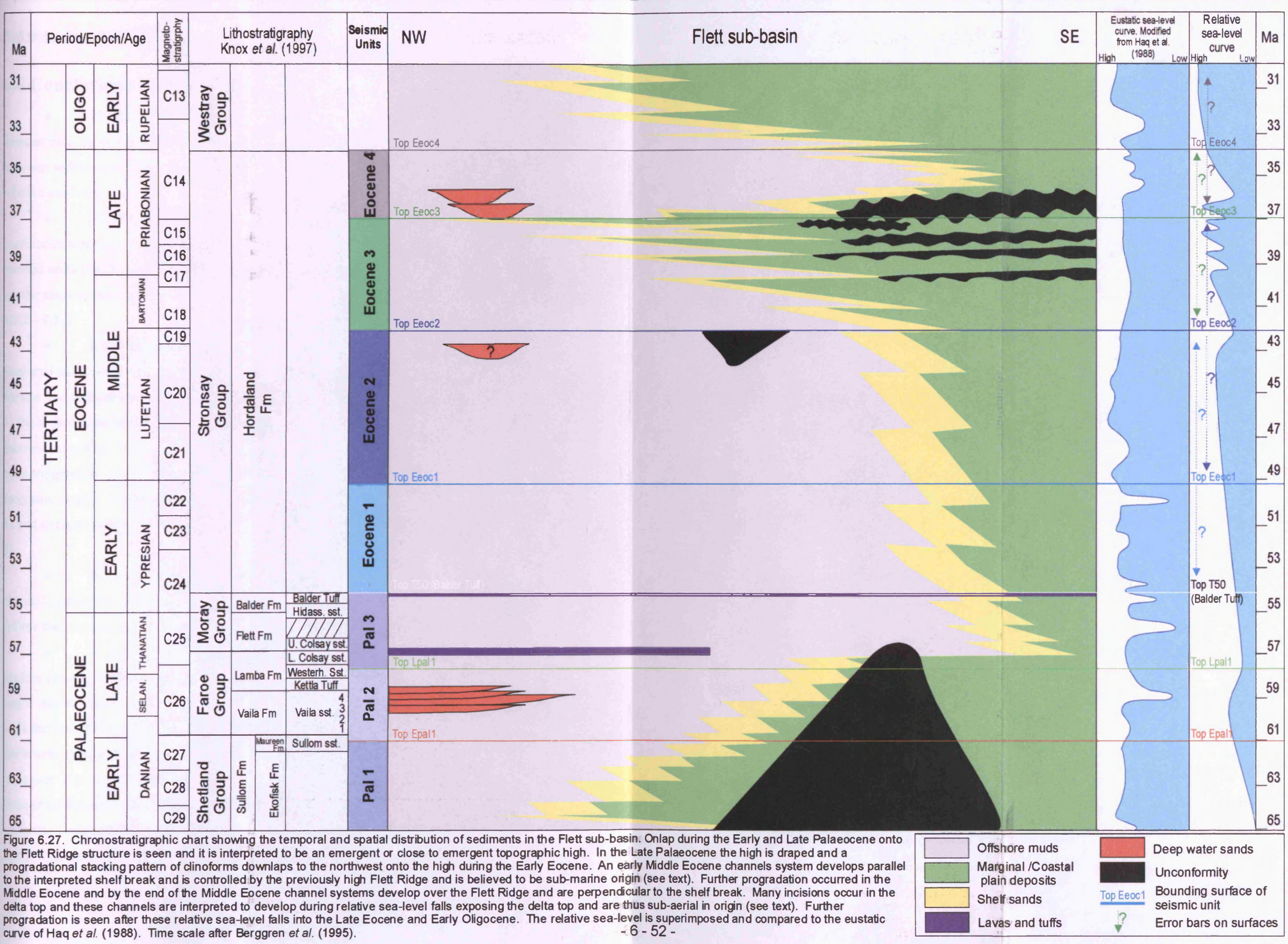
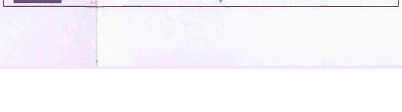


Figure 6.27. Chronostratigraphic chart showing the temporal and spatial distribution of sediments in the Flett sub-basin. Onlap during the Early and Late Palaeocene onto the Flett Ridge structure is seen and it is interpreted to be an emergent or close to emergent topographic high. In the Late Palaeocene the high is draped and a progradational stacking pattern of clinoforms downlaps to the northwest onto the high during the Early Eocene. An early Middle Eocene channels system develops parallel to the interpreted shelf break and is controlled by the previously high Flett Ridge and is believed to be sub-marine origin (see text). Further progradation occurred in the Middle Eocene and by the end of the Middle Eocene channel systems develop over the Flett Ridge and are perpendicular to the shelf break. Many incisions occur in the delta top and these channels are interpreted to develop during relative sea-level falls exposing the delta top and are thus sub-aerial in origin (see text). Further progradation is seen after these relative sea-level falls into the Late Eocene and Early Oligocene. The relative sea-level is superimposed and compared to the eustatic curve of Haq *et al.* (1988). Time scale after Berggren *et al.* (1995).



6.6 Conclusions

- The study of 3-D seismic data from the margin of the Faroe-Shetland Basin demonstrates its potential use in aiding the reconstruction of complex relationships between sedimentation and tectonics, and in interpreting relatively short-term changes in relative sea-level.
- Seven seismic units are defined in the study area by the correlation of continuous seismic reflection configurations. These reflections bound three Palaeocene seismic units (Pal 1, 2 and 3) and four Eocene seismic units (Eocene, 1, 2, 3 and 4) which are the same as the regional (basin-wide) seismic units describe in **Chapter 4 (Sections 4.3.2 - 4.3.5**).
- It has been shown that during the Early Palaeocene (Pal 1) there was a northeast -southwest trending structural high, the Flett Ridge; a positive topographic feature that influenced sediment dispersal patterns. During the Late Palaeocene (Pal 2 and 3) the high was onlapped and covered during a possible relative sea-level rise, or by the increase in sediment supply from the margin. This first blanketing of the ridge could have been triggered by increased subsidence on the basin margin or merely by continued sediment supply. Sediments continued to drape the margin into the Early Eocene (Eocene 1) and a major incision event occurred near the start of the Middle Eocene (Eocene 2 and 3).
- It is contended that the Flett Ridge had a major impact on the position of the channel systems at this time, resulting in a shelf-parallel axis of channel belts, which follow the contours of the ridge.
- A major period of progradation and delta development followed the early Middle Eocene incision event, and by the end of the Middle Eocene a second period of major incision took place (end of Eocene 3). This incision exhibits no spatial relationship with the buried ridge and the channel systems traverse this underlying structure in a northwest -southeast direction, perpendicular to the shelf edge and across the ridge structure. It is argued therefore that by the end of the Middle Eocene, the Flett Ridge had little or no impact on controlling the spatial variation of the channels caused by the earlier incision. Renewed progradation into shallower water depths occurred in the Late Eocene – Early Oligocene (Eocene 4).

- It is believed that the channel system seen at the start of the Middle Eocene incision event is sub-marine in origin, whilst the late Middle Eocene channels are interpreted as being sub-aerial.
- From all of these interpretations a relative sea-level curve and chronostratigraphic chart for this particular part of the Faroe-Shetland Basin has been constructed.

7. Chapter Seven: Discussion

7.1 Introduction

The specific aims of this thesis set out in the first chapter (**Section 1.2.2**) outlined the scope of this research. These aims have been achieved by the integration of a vast seismic, well and biostratigraphic database from the Faroe-Shetland Basin. A primary aim was to construct a chronostratigraphic based framework for the Eocene succession. By using the Faroe-Shetland Basin as a natural laboratory for the study of depositional systems, the erection of the new stratigraphic framework has been used to document the tectono-stratigraphic evolution at a unique time in the basin's history.

A fundamental issue relating to this study is the applicability and use of seismic and sequence stratigraphy in large-scale basin settings where vast differences in the quality of the database exist. Insights from areas of high resolution data has allowed for an understanding into the controls on the depositional architecture in localised areas of the basin. Thus a model for the inter-relationships of fluctuations in relative sea-level, changes in the rate of sediment supply and the role of eustasy can be sought.

This chapter will firstly summarise the most notable results and findings of the research which were described and discussed in **Chapters 4, 5** and **6**. The limitations of the study will then follow in a discussion focussing on uncertainties in the dataset with further explanation into any unresolved issues pertaining to the research. Thirdly, a summation of possible future avenues of research will conclude this thesis.

7.2 Summary of Results

The main results of the research are two-fold and stem from the detailed seismic interpretation on both a regional 2-D data (100's km) and local 3-D data (1 - 10 km's) scale.

7.2.1 Results from the Regional Eocene Study (Chapter 4)

A unique and pragmatic approach to seismic interpretation was conducted in order to sub-divide the Eocene stratigraphic succession. The varying quality of the geophysical, well and biostratigraphic database meant there was an intrinsic difficulty in using a standard approach of sequence stratigraphy. Indeed, this study has outlined the limitations and difficulties in the approach when attempting long-range correlation (see **Section 7.3**). An alternative methodology was set up to chronostratigraphically sub-divide the succession. This was based on the recognition of basin-wide (regionally) correlatable seismic reflection configurations that had a high amplitude continuous seismic character. Reflections with the highest degree of continuity define the upper and lower boundaries of seismic units. Four seismic units (Eocene 1 – 4) are identified, their age, distribution, and internal architecture being described in detail in **Chapter 4**. These seismic observations, coupled with the calibration of high quality lithological data has allowed for a systematic evolution of the sedimentary architecture to be documented.

The southern part of basin experienced marginal to shallow water conditions throughout the Late Palaeocene - Early Eocene when two large deltaic systems, separated by a major subaerial hiatus, prograded northwards into the basin. An Orkney and Scottish landmass source is inferred for the deltaic deposits. A marine environment, possibly experiencing outer shelf to basin floor conditions prevailed in the north of the basin. Tectonic movements in the form of a localised compressional activity were dynamic in the early Middle Eocene causing localised uplift which may have become emergent thus causing a separation of the southern part of the basin. This compression, possibly related to forces inherently related to the newly opening North Atlantic, had the effect of deforming the basin physiography which in turn led to the movement of sediment entry points to the northeast through time. A southeasterly Scottish and Shetland source for Middle Eocene deep water fans is invoked which are fed from point sourced canyons cut into the shelf break. The position of the canyons, the palaeo-bathymetry, the location of relic Mesozoic half grabens in the basin axis and the position of a buried lava escarpment are the major controls on the distribution and areal extent of the fan bodies. Minor submarine fans sourced form the southeast continued till the Late Eocene as the basin deepened, with progradational systems giving way to a more dominantly axial deposition occurring into the Oligocene and Neogene.

A chronostratigraphic chart was constructed for the entire basin showing the temporal and spatial evolution of the stratigraphy. This schematically shows the relationship between the marginal deltaic sediment and their deepwater basinal equivalents. Deposition of sandstones in fan lobes was contemporaneous with delta build out on the Shetland margin. However, no evidence of a lowering of relative sea-level can be calibrated with the clastic deposition. A thorough evaluation of key issues of the seismic stratigraphic approach follows this section outlining the results (**Section 7.3**).

A degree of cyclicity can be seen in the stratigraphic record where progradational units are bounded by localised unconformities. However, evidence within this study suggests that these local unconformities are developed due to relative sea-level falls associated with tectonic activity. Comparison of unconformities identified in the Faroe-Shetland Basin with suggested global sequence boundaries of the eustatic sea-level curve (Haq *et al.* 1988) show no correlation whatsoever. However, the biostratigraphic control of the Eocene seismic units defined in this study is such that a correlation would not be expected due to its poor resolution (see **Section 7.3**). The lateral extent of the unconformities and their localised distribution within the basin has been a major factor in understanding the causal mechanisms behind such falls in base-level. This study advocates compressional tectonics (inversion structures on relic Mesozoic lineaments) and dynamic uplift from a mantle plume as two plausible mechanisms for the observed fluctuations in sea-level. These events interrupted the general basin dynamics with an overall thermal subsidence phase being dominant since initial break-up occurred at the start of the Eocene. As no eustatic signal is recognised in this basin a global control cannot be ruled for or against. Biostratigraphic data used to construct the curve remains unproven and thus the validity of the eustatic curve remains questioned and hence would explain the reason for the non-correlation with the relative sea-level curve of the Faroe-Shetland Basin.

7.2.2 Results from the Localised Case Studies (Chapters 5 and 6)

High quality seismic data has shown that a higher resolution of stratigraphic architecture can be determined. Detailed analysis of the Eocene 1 unit described in chapter 4 has enabled four sub-units to be recognised. These units represent classical depositional sequences of the Exxon model and thus are bounded above and below by unconformities (sequence boundaries). Therefore in a small part of the basin a sequence stratigraphic framework of the Early Eocene has been constructed. This has enabled the palaeogeographic setting to be interpreted which shows very subtle changes in the depositional architecture which can be related to fluctuations in relative sea-level. Indeed regressive episodes pushed the shorelines forward and exposed the delta top areas causing dendritic drainage networks with a forced regressive wedge developed at the delta front. Marine inundation flooded the delta top during subsequent relative sea-level rise. A high degree of cyclicity is apparent with four transgressive – regressive cycles identified within the Ypresian. The delta progressively built out towards the north into water depths that deepened through time until subsidence caused eventual drowning.

With the detailed 3-D coverage of seismic data, an evolution of a marginal setting during fluctuations in sea-level has enabled recognition of small scale depositional and geomorphic features including barrier islands, incised valleys, deeply incised gullies and their associated feeder systems. More importantly, an understanding of the inter-relationship of these features during a full cycle of sea-level has been documented.

Detailed seismic interpretation in the area of the Flett Ridge has documented further controls on deposition of the Eocene succession. The role of local tectonics and the interaction with sediment dispersal patterns is emphasised. Indeed channel systems are primarily controlled by the position of the ridge which is interpreted to have undergone uplift due to emplacement an igneous body at depth creating a bathymetric high. Channel systems traversed the structure during renewed subsidence of the ridge.

7.2.3 Results Pertaining to Seismic-Stratigraphic Analysis

7.2.3.1 Introduction

This study has highlighted the approach of two stratigraphic methods which have been used to sub-divide the Eocene succession. The need to use different approaches of stratigraphic sub-division is borne out by the nature of the different datasets and the specific aim of the analysis.

Classical methods of sequence –stratigraphy (first described by Vail *et al.* 1977) have not been used in the regional basin analysis as the mapping of sequence boundaries have proved impractical over such vast distances – but the main concept that seismic reflections are isochronous and are thus time lines is used. Moreover, it is believed that sequence stratigraphic techniques are idealistic and impractical where the data quality is not consistent. A pragmatic approach was taken to sub-divide the stratigraphy which has enabled a seismic-stratigraphic framework to be established (see **Chapter 4** and **Section 7.2.1**) and a geological evolution of the basin has then been documented. This approach was favoured over of a more classical scheme because it was more practical to use with the data available.

The recognition of sequence boundaries from 2-D data has been difficult. Indeed this study has brought into question the lateral extent of sequence boundaries especially on a passive margin which is experiencing post-rift subsidence. The validity of sub-dividing strata based on sequence stratigraphy is thus examined here with a view to summarising the key issues involved.

Data resolution is paramount in basin analysis and underpins the entire strategy for which a stratigraphic succession is sub-divided. Integration of seismic, well and biostratigraphic data, all of high resolution, is required for an ideal sequence stratigraphic approach to be taken.

7.2.3.2 Sequence Boundaries and their Correlative Conformities

3-D seismic data used in this study has recognised sequence boundaries from the southern margin of the basin in the South Judd Basin (**Chapter 5**). However, these unconformities which bound depositional sequences are very limited aerially and cannot be correlated around the entire basin. This local case study therefore highlights an inherent problem when correlating sequence boundaries. Significant erosion is seen on the delta top during the development of the sequence boundaries in the South Judd Basin (SB1 – SB4 of **Chapter 5**) and thus subaerial exposure is interpreted during a relative sea-level fall. However the difficulty remains in the regional correlation of the correlative conformity to this sequence boundary. Indeed, away from the margin of the South Judd Basin either in the basin centre or on the Shetland margin sediments were deposited at the same time as the non-deposition within the areal extent of the sequence boundary. This begs the question as to the significance of the correlative conformity as it implies that areas of little or no deposition are time equivalent to areas of significant deposition. A geological scenario can be set up to explain this further. Erosion occurs in an isolated local area over a small high such as the Judd Anticline and basinward of this is, the corresponding correlative conformity is found. However, deltaic conditions prevail on the margin where no erosion is seen. In this scenario erosion (at the sequence boundary), non-deposition and varying amounts of deposition occur simultaneously in time though not spatially. Thus a package of sediment which occupies a 3-D area in space and time on a chronostratigraphic chart corresponds to a similar area which is characterised by an area of erosion and non-deposition. These two parts of the chart are joined by an area which resembles the correlative conformity or correlative conformities, which is characterised by an area (or areas) experiencing no erosion and continued deposition through time (with no significant hiatus).

Thus it is clear that the correlative conformity (or conformities) represent a period of time that corresponds to the development of a sequence boundary. How much time remains unclear though as an unconformity can develop in stages from initial creation until the final flooding event and indeed stages may differ between a type 1 and type 2 unconformity. Does a correlative conformity thus occur at (a) the time from initial incision (i.e. during erosion or bypass of the first delta topsets?), (b) during erosion and bypass at the offlap break (where the topset terminates basinward and becomes the foreset?) or, (c) at the time when there is maximum areal extent of the unconformity (at the inflection point, as defined by Posamentier *et al.* 1988) possibly with the development of an associated lowstand wedge? In the case of the final point, the inflection point occurs on the falling limb of the relative sea-level curve at the point of greatest sea-level fall (Posamentier *et al.* 1988). In reality, the correlative conformity is represented by all of the above scenarios on a falling relative sea-level curve. However the most rapidly falling sea-level is most often used to define the position of the true correlative conformity and this is defined by the latter of the three scenarios. The question remains though if it is correct to put the correlative conformity at this position which may correspond to the age of the first sediments deposited in a basin floor (lowstand fan) or in fact it is more correct to place the correlative conformity anywhere where a conformable succession is seen and is deposited at the time of sequence boundary development (**Figure 7.1**). This argument, that a correlative conformity does not exist in one place and is indeed unscientific is summarised by Embry (1995), who highlights the difficulty in recognising this event in the stratigraphic record.

From the observations highlighted above (and summarised in **Figure 7.1**) it can be stated that a correlative conformity could have a seismic expression which is one single seismic reflection representing a minimum value of sediment which occurs as hemi-pelagic fallout and deposits a thin veneer which may represent a long period of

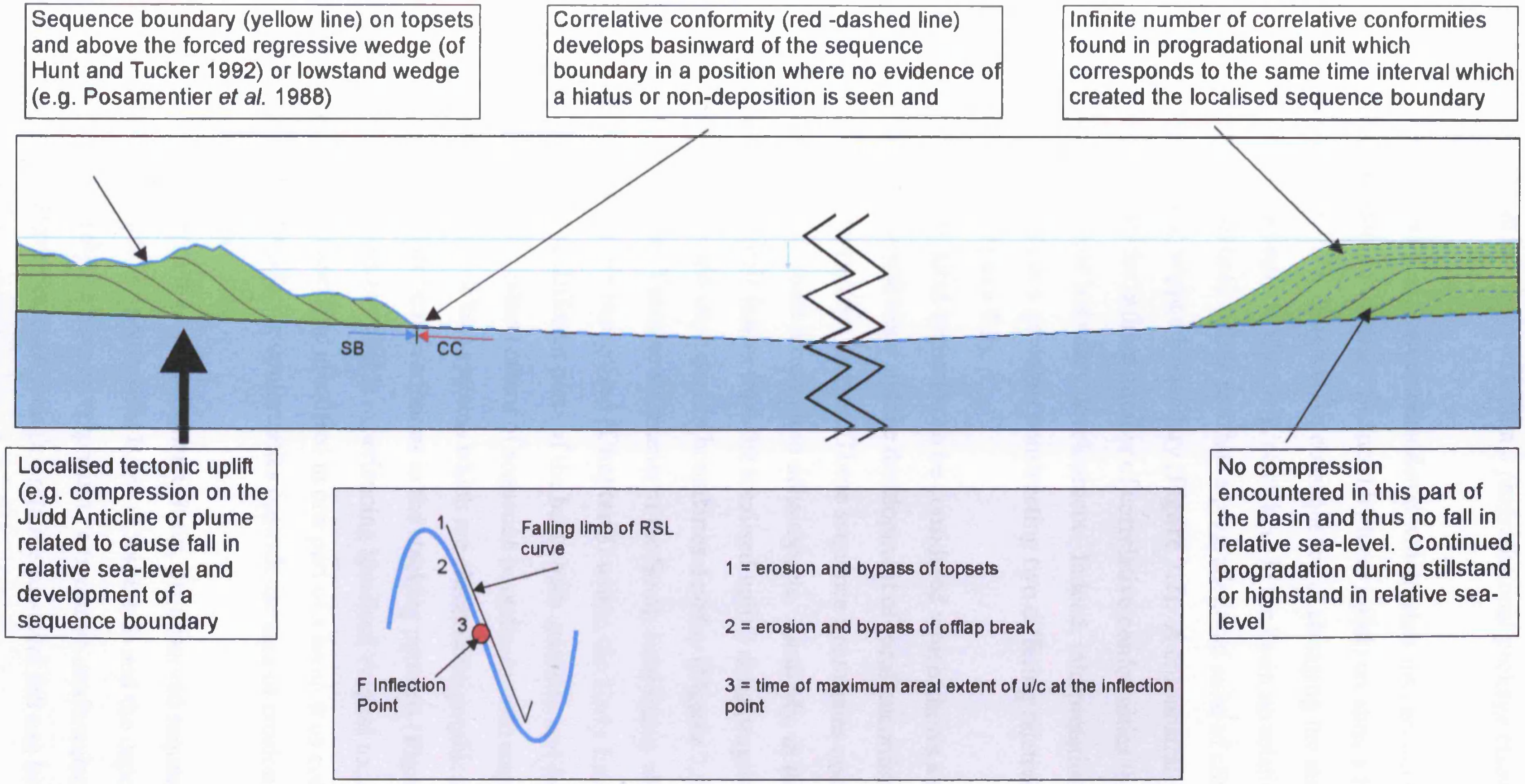


Figure 7.1. Schematic cartoon showing the effects of correlating an unconformity (sequence boundary) and its correlative conformities around a basin when the sequence boundary is created by a relative sea-level fall (e.g. due to uplift and compression on Judd Anticline or Flett Ridge, or due to mantle plume activity). The sequence boundary develops during the falling limb of the relative-sea-level curve though is at its maximum areal extent at the inflection point (Red dot). Elsewhere in the basin away from the local tectonic factors progradation continues during stillstand or highstand and thus correlative conformities occur throughout the progradational package. In between these areas a thin correlative conformity exists where the basin is sediment starved and only a thin veneer of hemi-pelagic sediment occurs. (See text for discussion)

time. Conversely, a suite of infinite correlative conformities representing the same period of time can be observed within a progradational package elsewhere in the basin (**Figure 7.1**).

A further scenario can be considered which takes into account a relative sea-level fall in one part of the basin (induced by local uplift) creating a localised sequence boundary (**Figure 7.2**). This has the consequence of changing the stacking pattern of a progradational package. Conversely, elsewhere in the basin no relative sea-level fall is seen (due to no localised uplift) and thus a progradational suite of clinoforms is seen with no intervening sequence boundary (**Figure 7.2**). A chronostratigraphic chart of this scenario shows the infinite number of correlative conformities that occur at the same time as sequence boundary development. Indeed, interpretation of the stacking patterns would result in a geologist constructing two differing relative sea-level curves for the two areas (**Figure 7.2**).

A final geological scenario can be considered which shows a cyclicity of transgressions and regressions and the development of local unconformities with their associated forced regressive wedges. These sequence boundaries occur due to relative sea-level falls which create erosion into delta topsets. Similarly, in this scenario elsewhere on the margin (away from the localised uplift) delta progradation is seen throughout this time and no sequence boundaries develop (**Figure 7.3**). Part of this scenario is seen in the Ypresian sediments of the South Judd Basin where four sequence boundaries are recognised (**Chapter 5**) within the Early Eocene delta. However, by showing different parts of the basin both spatially and temporally this scenario highlights the lateral extent of sequence boundaries with respect to localised uplift. Two individual deltaic systems which are chronostratigraphically the same are seen to experience significant variations in the stacking patterns (**Figure 7.3**) due to their position within a basin that is experiencing localised vertical tectonic movements. The four sequence boundaries identified in one part of a basin thus correlate with an infinite number of correlative conformities outwith the area of erosion (**Figure 7.3**) elsewhere in the basin.

Coming back to the Faroe-Shetland Basin, the observed sequence boundaries on the southern margin of the South Judd Basin (**Chapter 5**) and the depositional sequences they bound thus have corresponding correlative conformities throughout the rest of the basin. However, even though a relative sea-level fall can be interpreted for the development of the sequence boundaries, no eustatic signal is seen due to the poor

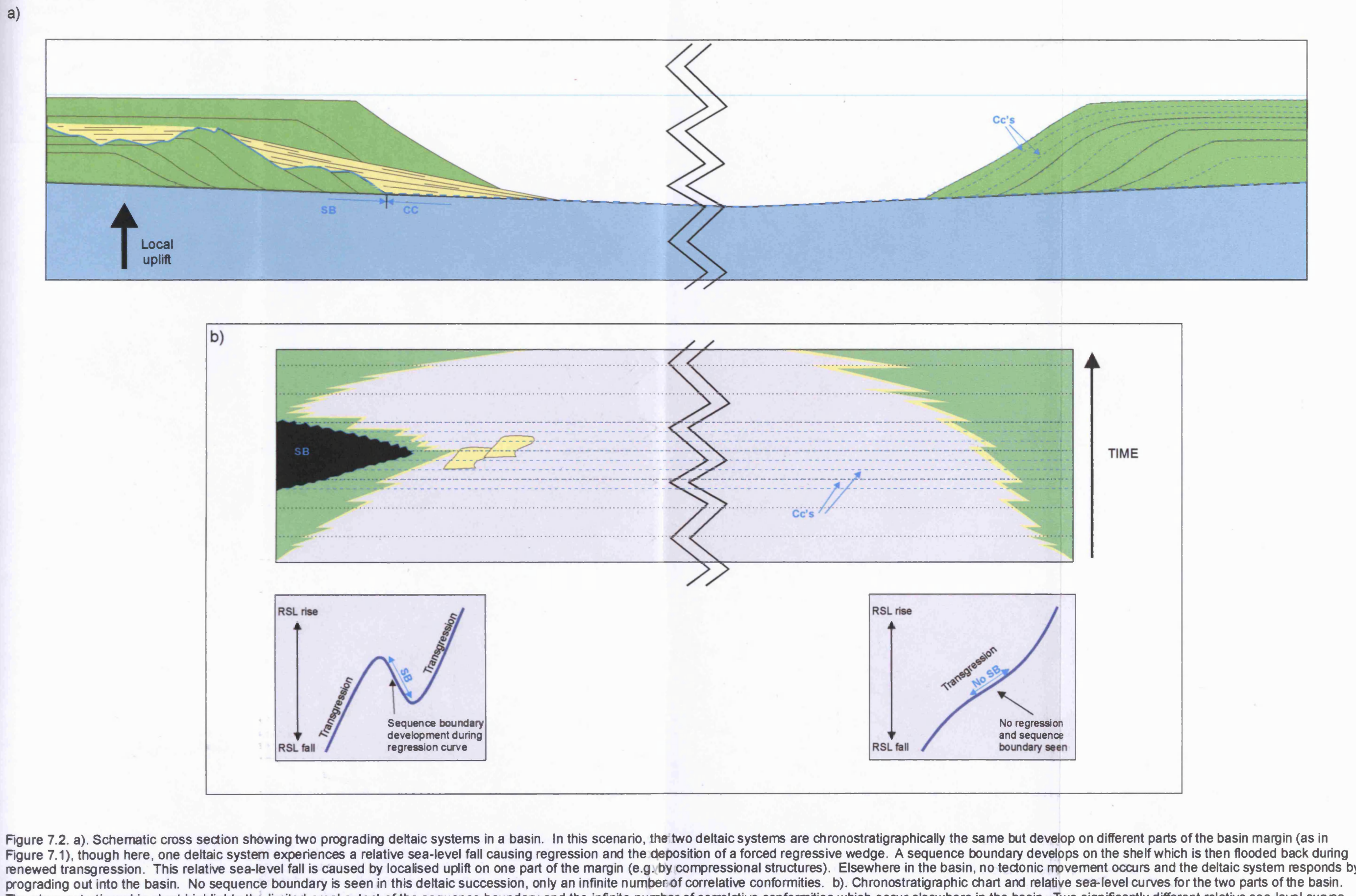


Figure 7.2. a). Schematic cross section showing two prograding deltaic systems in a basin. In this scenario, the two deltaic systems are chronostratigraphically the same but develop on different parts of the basin margin (as in Figure 7.1), though here, one deltaic system experiences a relative sea-level fall causing regression and the deposition of a forced regressive wedge. A sequence boundary develops on the shelf which is then flooded back during renewed transgression. This relative sea-level fall is caused by localised uplift on one part of the margin (e.g. by compressional structures). Elsewhere in the basin, no tectonic movement occurs and the deltaic system responds by prograding out into the basin. No sequence boundary is seen in this deltaic succession, only an infinite number of correlative conformities. b). Chronostratigraphic chart and relative sea-level curves for the two parts of the basin. The chronostratigraphic chart highlights the limited areal extent of the sequence boundary and the infinite number of correlative conformities which occur elsewhere in the basin. Two significantly different relative sea-level curves can be constructed from the two parts of the basin. This diagram highlights the methodology and difficulties when attempting to accurately pick the true correlative conformity.

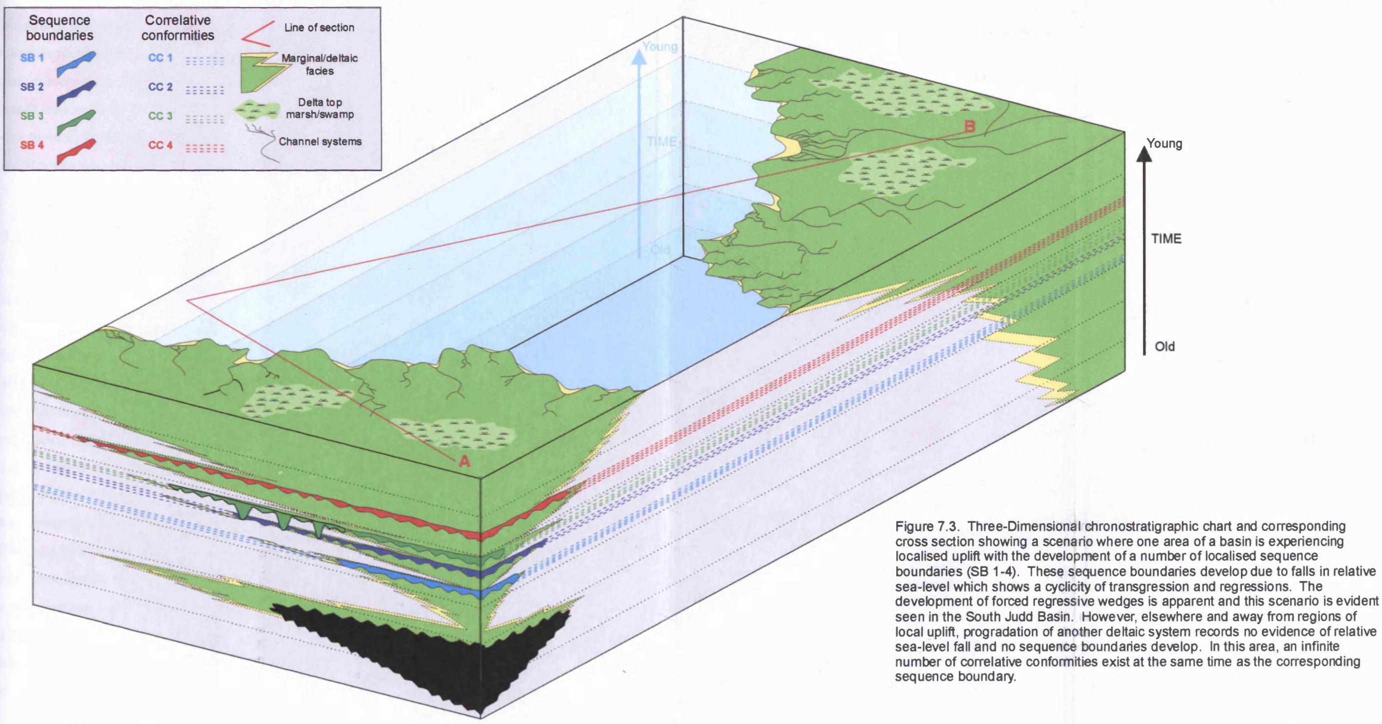
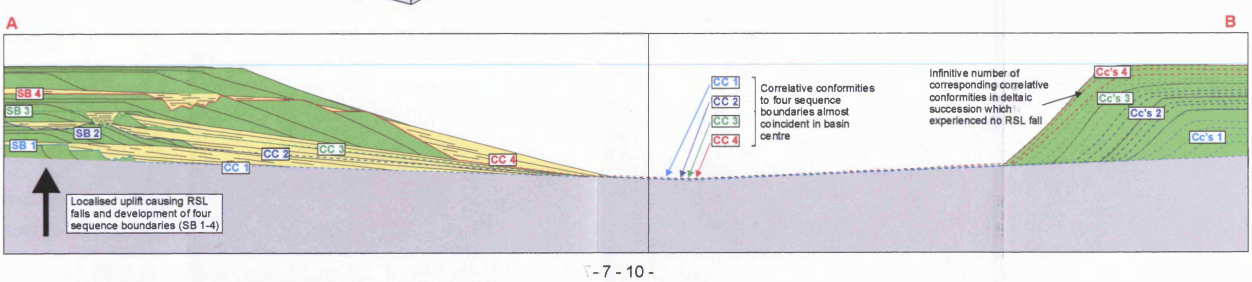


Figure 7.3. Three-Dimensional chronostratigraphic chart and corresponding cross section showing a scenario where one area of a basin is experiencing localised uplift with the development of a number of localised sequence boundaries (SB 1-4). These sequence boundaries develop due to falls in relative sea-level which shows a cyclicity of transgression and regressions. The development of forced regressive wedges is apparent and this scenario is evident seen in the South Judd Basin. However, elsewhere and away from regions of local uplift, progradation of another deltaic system records no evidence of relative sea-level fall and no sequence boundaries develop. In this area, an infinite number of correlative conformities exist at the same time as the corresponding sequence boundary.



biostratigraphic control. It has been documented throughout **Chapters 4, 5** and **6** that unconformities develop locally due to local tectonic controls. This has been particularly noticed in areas experiencing compression causing uplift over a relatively small area such as on the Judd Anticline and Flett Ridge. Unconformities that develop due to vertical tectonic movements are thus of intra-basinal significance and have no merit when contemplating eustatically driven sedimentary cycles.

7.2.3.3 What Constitutes a Sequence Boundary?

A further area of contention within the sequence stratigraphy model is what actually constitutes an unconformity in a sequence stratigraphic sense. The original definition of Mitchum *et al.* (1977) was revised by Van Wagoner *et al.* (1988) to be more restrictive to a surface where there is evidence of subaerial erosion (and in some areas, correlative submarine erosion) or subaerial exposure. This definition of an unconformity in the sequence-stratigraphic sense therefore is presumed to include submarine canyons. However, it is fair to say that canyons can only be included if they can be linked up-dip to their shelfal and ultimately fluvial counterparts which show a basin ward shift in facies (an unconformity). This is rarely possible in reality and can be shown with the case of the Middle Eocene submarine fans seen in the Faroe-Shetland Basin. The three canyons lie at the shelf-slope break and the data quality does not allow for the canyons to be traced up-dip into the eroded delta tops or fluvial systems which may be expected in the classic lowstand fan model (Posamentier *et al.* 1988). Canyons developing during either a subtle or large fall in relative sea-level may be one way for delivering clastic material to the basin floor although other ways do exist.

Firstly oversteepening of a delta at a shelf edge may cause a permanent failure on the slope with delta front collapse. This in turn could cause a canyon head to develop. Therefore no relative sea-level fall needs to be invoked for this canyon development with the principle mechanism for the canyon generation being failure due to faulting or by an earthquake. Furthermore, modern river fluxes to shelf deltas have been analysed and show that significant transport of clastic systems occurs during third-order highstands of relative sea-level (Burgess and Hovius 1998). These authors thus argue that a relative sea-level fall does not need to be invoked to deposit clastic material in the basin and in fact shelf dynamics, shelf dimensions, drainage basin characteristics and climate are all locally important and must be incorporated into the stratigraphic model (Burgess and Hovius 1998).

A final tectonic control on canyon development may occur though not be related to any oversteepening and failure. The similar position of Palaeocene and Eocene aged canyons in the south of Quadrant 214 is not believed to be a coincidence and may be related to a deep-seated tectonic control on the hinge or to a sediment loading factor (see **Section 4.4.3.2.2**).

7.2.3.4 Comparison of the Faroe-Shetland Basin with other basins along the Northeast Atlantic Margin

This section will briefly summarise the evolution of adjacent basins during the Eocene and compare any findings with what has been documented in this study, in order to evaluate the bigger picture.

Firstly, it must be stated that there is very little documented work which focuses on the Eocene succession in the Rockall Trough, Møre and Vøring Basins. As with the Faroe-Shetland Basin, detailed studies have concentrated on the prospective hydrocarbon plays, thus, the Palaeocene and deeper successions have been identified and studied. Regional reviews are thus used here to ascertain the geological history of the neighbouring basins with isolated references to key studies. To the north of the Faroe-Shetland Basin (north of the Brendan's Dome volcanic edifice) the Møre Basin exhibits a similar northwesterly thinning sedimentary wedge of Eocene and Oligocene that appears to be thinner than the comparable succession in the Faroe-Shetland Basin (see Figure 2 of Doré *et al.* 1999). Furthermore, the Neogene to Recent succession in these more northerly basins is observed to be significantly thicker than in the central and southern parts of the Faroe-Shetland Basin. Indeed a thickening of this post Eocene succession is seen towards the North Vøring Basin (Doré *et al.* 1999).

In the Rockall Trough, (south of the Wyville-Thompson Ridge) a thick Eocene to Oligocene succession is observed with no Neogene to Recent cover in the basin centre (Doré *et al.* 1999). A simplified Eocene sequence stratigraphic framework has been established in the Northeast Rockall Trough which records an overall deepening throughout the Eocene (Waddams and Cordingley 1999). The authors record a similar stratigraphic architecture to the Faroe-Shetland Basin, with clastic systems (both shelfal and basinal sands) dominating the early part of the Eocene with a local unconformity separating this earlier unit from the later one that shows a more northwesterly prograding shelf system with isolated basinal sands (Waddams and Cordingley 1999). Indeed, Egerton (1988) suggests basin floor turbidites to be point sourced form the locally high Wyville-Thompson Ridge to the northeast.

The entire Northeast Atlantic Margin was experiencing voluminous igneous activity at the time of break-up and the northwestern margins of all these basins and it is believed that an axial area of central uplift occurred in the north separating East Greenland from the Møre and Vøring Basins (Brekke *et al.* 1999). Thus, a central uplifted lineament trending northeast - southwest is interpreted at the time of break-up with the northwesterly basin margins being delineated by a continuation of the Faroe-Shetland Escarpment (in the Møre Basin) and the more northerly Vøring Escarpment which is situated more basinward due to offset on the Jan Mayen Lineament (Brekke *et al.* 1999, Roberts *et al.* 1999).

Thus it is concluded here that during the Eocene, the adjacent basins experienced similar conditions to that of the Faroe-Shetland Basin. Towards the northeastern end of the margin little clastic deposition from either margin was deposited in the basin, though locally high areas may be the reason for deep water clastic systems to be observed in the more southerly basins. Regional thermal subsidence is thus believed to be the overwhelming factor controlling the depositional architecture that is observed and only local tectonically active areas provided uplift.

7.3 Limitations and Uncertainties

7.3.1 Biostratigraphic Resolution

With all scientific research it is important to know the accuracy of the data one is working with. Knowing the limitations and error bars involved with the data is key to assessing the validity of the results. The major uncertainty which can be drawn from this study is the lack of high quality biostratigraphic data. The lack of this data has proved crucial in determining the accuracy of dating the seismic units of the Eocene succession. Indeed, the results have shown that it is only possible to give tentative relative ages to the seismic units such as Early (Ypresian), Middle (Lutetian – Bartonian) and Late (Priabonian) Eocene. No absolute age dates (in Ma) are possible from the observed biostratigraphical information for this study. This is because there is a dearth of good well data available (see **Section 3.5.2** on definition of type wells) with only thirteen wells providing any real usable data for age dating.

The fundamental problem with the biostratigraphic data is the paucity of first and last down-hole appearances (FDA's and LDA's) in the samples taken for which to assign particular bioevents. The North Sea biostratigraphic zonation uses common occurrences of assemblages of dinocysts and miospores to correlate to a more accurate age range (Mudge and Bujak 1996). This zonation has been applied in the Faroe-Shetland Basin (Mudge and Bujak 2001) but remains of little worth until high quality biostratigraphic data is available.

Due to this irresolvable issue with biostratigraphic data it has not been possible to correlate either the regional (basin-wide) Eocene units **(Chapter 4)** or the local sub-units of the South Judd Basin **(Chapter 5)** to the eustatic sea-level curve of Haq *et al.* (1987; 1988). Relative sea-level curves have been produced for the basin as a whole and for the two case study areas and it is suggested that a correlation to the eustatic curve is unlikely because of the inherent problems and pitfalls associated with the construction (see **Section 4.4.2.2.1** for discussion). However, what has been possible in this study is the construction of a basin-wide chronostratigraphic framework which documents and describes the entire Eocene succession of the basin. What remain untested are the absolute age constraints for this new chronostratigraphic framework.

7.3.2 Skewed Nature of Database

This study as benefited from a vast database which spans seismic, well and biostratigraphic forms. However this data is not ideal and is skewed towards the UK side of the basin. A consequence of this distorted nature of the data is thus conveyed into the observations and interpretations documented in this study. No well data has been acquired from Faroese waters and for this reason, lithological data from this vast area has been confined to isolated published reports which rely primarily on dredge samples. Additionally, seismic data from this area is for the most part 2-D and the detail seen from the 3-D seismic data is thus not on hand and hence part of the geological story is probably underrepresented. One particular related geological issue to this lack of data is the provenance areas for the clastic systems documented in the Palaeogene. The informally termed Late Palaeocene "Vagar Sands" are interpreted to have been sourced from the northwest opening up a new potential source area. It is believed that the lava plains of the Faroes Platform would not have provided much clastic sedimentation in the Early Eocene, though long-lived drainage systems may have connected to the East Greenland area to provide a possible source. High resolution 3-D seismic data from Faroese waters would enable this idea to be validated since it remains untested.

7.3.3 Variation in 2-D and 3-D Seismic Data Quality

This brief section will highlight the difference between the 2-D seismic data used to produce a regional chronostratigraphic framework in **Chapter 4** and compare it directly to the high quality (and more localised) 3-D seismic surveys used in **Chapters 5** and **6**. Furthermore it will highlight the advantages when using 3-D seismic data and point to the limitations of the 2-D realm. The 3-D seismic data, by nature has no data holes or windows which are seen in grids of 2-D seismic. Thus, there is a significant amount of data that is not available to the interpreter (i.e. missing) when using 2-D seismic. Using 3-D seismic therefore gives complete areal coverage and allows for data missed in the 2-D to be seen in the 3-D seismic.

The three localised 3-D seismic surveys used in **Chapters 5** and **6** have a line spacing which is usually acquired and processed to produce either a 25 x 25 m or 12.5 x 12.5 m grid. Indeed, the horizontal resolution of 3-D data can be in the region of 10 – 15 m. Couple this high lateral resolution with a vertical resolution of between 10 and 20 m and it allows for seismic interpretation to be carried out with much greater detail than with the 2-D data, especially when looking at small scale geological phenomena.

Conversely, 2-D regional seismic lines provide the geologist with an overview of the gross geometry of a stratigraphic succession with the spacing between neighbouring 2-D lines varying from approximately 1 km to tens of kilometres. Therefore the application of the two categories of seismic data is different depending on the specific aim of the seismic interpreter. In the 3-D realm, subtle, small scale (localised) features of a much smaller dimension and magnitude than seen on 2-D data can be extensively mapped. This has been proven in **Chapter 5** with the recognition of sequence boundaries (SB1 - SB4) and discrete small scale channel systems. Other features may include individual channel levee complexes, canyons, barrier islands, shelf ridges and faults which may only have dimensions of a few tens to a few hundreds of metres and are thus not seen on many (if any) 2-D lines. Identification of these subtle features and calibration with modern day analogues may enhance the understanding of depositional systems. This novel branch of seismic interpretation termed seismic geomorphology is in its infancy but may prove to be a useful additional tool in increasing knowledge on modern and ancient depositional systems (e.g. Posamentier 2002; 2004).

This thesis has outlined some of the generic and technical issues which arise when using both a coarse grid of 2-D seismic data and small localised 3-D seismic surveys. Both datasets have strengths which vary with regard to certain aspects of geological insight. Indeed they are complementary and a combination of 2-D and 3-D seismic interpretation is ideal in order to recognise features on a variety of scales. This is the recommended practise which will aid the overall stratigraphic or structural interpretation of the basin.

7.4 Future Avenues of Research

7.4.1 Biostratigraphy

A fundamental issue within the study of stratigraphy is the degree to which correlation of biostratigraphic markers can be used to objectively date individual units or sequences. There is a paucity of good biostratigraphic information for the Eocene succession of the Faroe-Shetland Basin which has been emphasised by the need to qualitatively assess the suitability of all wells and biostratigraphic data. Detailed micropalaeontological and micropalynological work would be a prerequisite in order to ascertain absolute age dates for the Eocene section for any future work. Re-sampling of the cuttings from many of the wells (in particular those drilled relatively early in the basins exploration history) would prove extremely worthwhile and be expected to provide better quantitative data which would aid with stratigraphic sub-division. Many of the type wells defined in **Section 3.5.2** have been recently drilled and are often located in the axis of the basin. Therefore by focussing on acquiring better biostratigraphic data from wells on the Shetland margin, it will enable the deep water Middle Eocene sands to be correlated to the shallow water equivalents.

7.4.2 Provenance Studies

Heavy mineral analysis of sandstones found in the basin may provide an insight into potential source areas during the Eocene. Further to this, the positions of palaeo-sediment pathways which were active at the time may be possible to predict with extensive analysis of the sedimentology and trace element chemistry. Particular terranes from the Scottish landmass (including the Shetland and Orkney Islands) and East Greenland may allow for a reconstruction of catchment areas throughout the Palaeogene. Both southeasterly and westerly source areas are known for Late Palaeocene turbidites in the basin, with the former believed to be the wholly dominant hinterland. However, with the paucity of knowledge and wells on the Faroes side important play fairways may remain unexplored. In addition to this provenance work, integration of data from many published studies of Palaeocene and Eocene clastic deposits of the North Sea Basin would provide an important regional study into the palaeogeomorphic evolution of onshore and offshore northern Britain. From this work a detailed understanding of fluvial, deltaic and ultimately deep water systems active in both a sheltered basin and on a volcanic passive margin is possible which can then be incorporated with ongoing uplift, exhumation and denudation studies.

7.4.3 Internal Architecture of Deep water Fans

An understanding into the depositional processes and controls on the Middle Eocene deep water fans may be possible by looking at high resolution 3-D seismic data. This study has highlighted the importance of the canyon setting, but has not been possible to trace the feeder systems into more marginal areas due to the lack of any 3-D seismic data on the shelf. When studying deep water systems it is always imperative to have an understanding of the shelfal and shallow water equivalents and this knowledge must be integrated with the deep water depositional system. Detailed interpretation of the individual fan bodies would allow for individual lobes, identified in wells to be correlated. This would allow for amore robust depositional model for the fans to be constructed. Individual lobes could be mapped out and their distribution, thickness and geometry could be examined. Further to this, an insight into the evolution of fan systems could be considered by comparing the stacking patterns throughout the entire fan body with the geometry of the individual lobes.

7.4.4 Subsidence Analysis

This thesis has constructed a seismic-stratigraphic framework which has a loosely defined chronostratigraphic significance. From this initial work detailed subsidence analysis from both well and seismic data and back-stripping of the stratigraphy can allow for an insight into the timing, location and amount of post-rift subsidence on both the flanks and centre of the basin. The regional thickness maps (isochores) would provide the template for this arm of study, from which the subsidence of a volcanic margin can be analysed.

7.4.5 Published and Planned Papers

The research from this thesis has produced two published paper. The results from **Chapter 6** have been summarised in a short paper describing how 3-D seismic data have been used to describe the interaction of growing bathymetry/topography on channel development in the Flett Ridge area (Robinson *et al.* 2004). A more regional study on the episodic compressional history of the basin and its impact on the prospectivity of hydrocarbons and this research has provided ideas and stimulus for this paper (Davies *et al.* 2004).

Further to these two papers, it is proposed that the author will focus on publishing the results pertaining to the work in **Chapter 5**. Several lines of enquiry are being pursued with particular emphasis on the seismic architecture of the geomorphic features encountered in the Ypresian sediments of the South Judd Basin. Suggested paper publications include:

- A description of the depositional systems observed in the South Judd Basin and the implications for fluctuations in relative sea-level.
- The depositional controls on an Early Eocene deltaic system and the role of relative sea-level on marginal channel development and abandonment.
- Seismic geomorphology of marginal systems: forced regressions and incised valleys versus transgression and barrier island development.

8. Chapter Eight: Conclusions

This thesis has investigated the stratigraphic evolution and depositional systems of the Eocene succession of the Faroe-Shetland Basin. The following statements will précis the concluding remarks from the chapters, providing a summation of the research.

8.1 General Conclusions

1. This study has shown that to a close approximation seismic reflections are isochronous and thus represent timelines.

2. A regional seismic-stratigraphic framework for the Eocene succession of the Faroe-Shetland Basin has been constructed. This has been achieved by using two techniques of seismic-stratigraphic interpretation which has led to the sub-division of the succession on both a basin-wide and local scale.

3. Sequence boundaries, though not readily identifiable on 2-D seismic data have been recognised on localised high resolution 3-D seismic surveys and thus relative sea-level falls are inferred in what is believed to be a post-rift thermally subsiding basin. Three-dimensional (3-D) seismic data is thus critical in this study for the construction of local palaeogeographic reconstructions (**Chapters 5** and **6**). Two-dimensional (2-D) seismic data however is fundamental when undertaking regional basin analysis as in **Chapter 4**.

4. Relative sea-level curves have been produced for the entire basin and for two local regions and show no correlation with the eustatic sea-level curve. This is to be expected for two reasons: 1. the basin experienced localised controls throughout deposition of the Eocene succession including a post-rift subsidence phase, local compressional tectonics, igneous underplating and mantle plume activity. 2. the biostratigraphic resolution is poor and thus only a loosely defined chronostratigraphic control can be placed on the newly developed seismic-stratigraphic framework and thus it is not feasible to test the relative sea-level curves against the eustatic curve.

5. Major sequence boundaries that developed on the margins of the basin have been attributed to relative sea-level falls (due to the local controls discussed in point 4 above) and not as a result of global changes in sea-level.

8.2 Concluding Remarks from the Basin-wide Study

1. A sequence-stratigraphic framework for the Eocene succession has been constructed for the entire Faroe-Shetland Basin using 2-D seismic data and where possible tied to biostratigraphic data from wells and BGS boreholes. This seismic-stratigraphic framework has a loosely defined chronostratigraphic component and together they have formed the template to produce a series of palaeogeographic maps which document the basin evolution.

2. By using a pragmatic approach of seismic mapping, which differs from the more conventional seismic stratigraphic approaches known, four regional seismic-stratigraphic units (Eocene 1 - 4) have been identified. The surfaces bounding these units are seismic reflections that are the most correlatable around the basin and thus the fundamental concept of seismic stratigraphy; that seismic reflections are equivalent to timelines is followed (Vail *et al.* 1977). The seismic reflections chosen have the most widespread geographical extent and often appear as high amplitude continuous to semi-continuous configurations.

3. Regional compressional structures cause changes in the depositional architecture of the Eocene succession. Indeed, Early Eocene deposition is focussed in the southern half of the basin. Subsequent deposition in the Middle Eocene is more dominant on the Shetland margin and the southern part of the basin became an area of erosion. This large scale shift in focus of the depositional systems is believed to because of early compression on the Munkagrunnur Ridge, Wyville-Thompson Ridge and Judd High and Anticline which caused erosion over the anticline and was perhaps emergent creating a possible land-bridge between Greenland and the Scottish landmass for migration of mammals.

4. The Middle Eocene succession is dominated by a series of elongate deep water fan systems which are interpreted as base of slope turbidites. The principle controlling factor on the deposition of these turbidites is the position of shelf-edge canyons which link the deep water clastic system with the shallow deltaic succession towards the Shetland margin. The canyons appear to be located in the same position as earlier Late Palaeocene canyons which fed clastic material into the Flett sub-basin and a long-lived inherent, possibly tectonic control on their position is postulated. Further controls on fan deposition include the basin physiography and bathymetry with the

Mesozoic Corona Ridge and the Faroe-Shetland Escarpment forming major topographic controls.

8.3 Concluding Remarks from the Local Case Studies

A more detailed seismic investigation from two parts of the basin allowed for a thorough evaluation of the controls on the stratigraphic development of small-scale depositional systems. Firstly, 3-D seismic data from the South Judd Basin provides a high resolution case study located in a marginal setting during Early Eocene (Ypresian) delta progradation. The intra-basinal structure of the Flett Ridge provides a second case study which focussed on the effects of topographic controls on channel development.

8.3.1 South Judd Basin Case Study

1. More classical methods of sequence stratigraphy have been used in **Chapter 5,** with the recognition of four sequence boundaries allowing the Eocene 1 unit to be sub-divided into four type 1 sequences (Posamentier *et al.* 1988). The sequence boundaries develop during falls in relative sea-level with the shoreline advancing to the north throughout the Ypresian.

2. The development of the sequence boundaries caused major periods of erosion of the delta top. Forced regressive wedges have been identified in the strongly progradational stacking pattern. Cyclicity in relative sea-level fluctuations (regression – transgression) is identified throughout the Ypresian with facies belts and depocentres migrating towards the north. Fluctuations in the order of 50 - 100 m are interpreted though no absolute time-scale can be put on the timing of sea-level change due to the poor biostratigraphic resolution.

3. Causal mechanisms that are postulated for the observed relative sea-level fluctuations are vertical tectonic movements. Specifically uplift on the Munkagrunnur Ridge, Wyville-Thompson Ridge and the local Judd Anticline caused compression associated by sea-floor spreading in the newly created North Atlantic Ocean. Additionally, underplating or plume activity may have a role to play and could explain the pulsed or cyclical nature of the relative sea-level.

4. Seismic attributes have been used to image the complex drainage network of channel systems. Recognition of sinuous and meandering channels, incised valleys, barrier islands and canyons on the sequence boundary has aided the stratigraphic interpretation and relative sea-level history. Indeed, barrier islands are believed to develop during the very late stages of transgression and occur at the edges of palaeo-incised valleys where a small change in gradient occurs. The barrier islands may indicate a wave or tidal dominated environment.

5. Incised valley systems are seen to develop on the delta top and have topographically high areas on their flanks. Flooding of the valley networks ensue during transgressive episodes and subsequent down-cutting into the high flanks by highly incised canyons or gullies is evident during following relative sea-level fall.

8.3.2 Flett Ridge Case Study

1. Detailed seismic interpretation in the Flett Ridge area has proven the ridge had an important role in controlling the deposition of sediments on the Shetland margin. This intra-basinal ridge forms a topographic high which influenced the dispersal patterns of Palaeogene sediments and caused deflection of channel systems. The complex interplay between tectonics and sedimentation is highlighted and the mechanisms driving the observed sedimentary architecture explored.

2. The Flett Ridge is interpreted to be a Mesozoic tilted fault block which was intruded into by a 7 km diameter igneous body. This in turn fed sills which intrude into the Cretaceous section and jack-up the overburden causing significant topographic relief in the Early Palaeocene. Indeed, westerly sourced turbidites of the earliest Late Palaeocene are seen to fill lows in the sea-floor topography which is locally uplifted by sills.

3. The ridge remained prominent into the Early and Middle Eocene and a relative sea-level fall is interpreted as shelf parallel channels bend round the ridge. Subsidence of the ridge is inferred by the end of the Middle Eocene as smaller delta top channels traverse the ridge structure.

4. Fluctuations in relative sea-level of less than 50 m are believed to cause the erosion seen on the delta top allowing for bypass of clastic sedimentation which is ultimately deposited in the base of slope region as deep water fans.

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Appendix 1. 2-D Seismic Database

The following 2-D surveys were used in this study:

SURVEY NAME	**NUMBER OF LINES AND KM (where available)**
AHL-94-202	29
AUK-92-AH	33
C-92	48
C-93	23
DEL-93	19 (1519 km)
DGS-95	6
E-84	24
GA-95-NSB	40
M-90-WS	10
M-91-FT-6	14
M-92-FT-6	6
NEST-90	1 (1160 km)
NSB-96	4
NWZ-96	14
OF-94	37
OF-95	23
RM-90-WS	1
RM-91-FT-6	27
RM-92-FT-6	19
SF-94	72 (2894 km)
SFE-95	21 (2843 km)
WG-RR-84	8
WS-94	64 (3283 km)
WSD-91	38 (1650 km)
WSD-92	23 (2400 km)
92-07	61

Appendix 2. 3-D Seismic Database

The following 3-D surveys were used in this study:

SURVEY NAME	**Survey area (square km)**
MC3D-WOC-96	587
RSF-96 (Re-processed)	2050
SF-96	2600
SF-97	2007
T-4	1350
T-61/T-62	2361
WSMERGE_KM	2487.26

Appendix 3. Well Database

The following wells were used in this study (nb. wells in red are the thirty five wells and two BGS boreholes that have been classified as "type wells"):
All these wells are now released (i.e. they were drilled over 5 years ago).

Quadrant 202

Well Name	X location	Y location	Composite Log	Velocity Log	Biostratigraphic Report	Year drilled	Current Owner	TD
202/02- 1	408936.7423	6635392.942	Y	Y	Y	1974	ExxonMobil	1219.81m
202/03- 1A	415210.5757	6637011.514	Y	Y	Y	1974	BP	1775.52m
202/03- 2	416165.7613	6644671.556		Y	Y	1977	BP	1524.51m
202/03a- 3	412528.8429	6652273.716	Y			1987	BP	2553.34m
202/08- 1	414017.0942	6629403.981	Y			1974	Shell	1685.54m
202/09- 1	427010.6682	6629284.448	Y	Y		1987	ExxonMobil	1665.12m
202/18- 1	418230.3438	6583399.641	Y			1991	Total	3102.86m
202/19- 1	422873.0213	6578134.499	Y			1984	Shell	3188.82m

Quadrant 204

Well Name	X location	Y location	Composite Log	Velocity Log	Biostratigraphic Report	Year drilled	Current Owner	TD
204/19- 1	425696.9469	6695460.223		Y		1986	BP	4683.25m
204/19- 2	429529.5925	6694909.028	Y	Y	Y	1991	BP	967.98m
204/22- 1	409318.4363	6678205.027	Y	Y	Y	1989	BP	2774.29m
204/22-2	404567.2121	6684724.037		Y		1994	BP	3711.0m
204/23- 1	417155.7799	6679141.945	Y	Y	Y	1987	BP	3881.99m
204/24- 1A	430067.757	6687121.622	Y	Y	Y	1990	BP	3019.8m
204/24A-2Y	425908.2901	6685841.315		Y		1992	BP	954.12m
204/25- 1	441068.1734	6688583.449	Y	Y		1991	Amerada	2904.74m
204/27a- 1	408162.3556	6671083.802	Y	Y		1990	BP	2179.69m
204/28- 1	414900.4888	6670815.408	Y	Y		1981	BP	1974.0m
204/29- 1	424741.495	6664614.168	Y	Y		1986	Talisman	2243.33m
204/30- 1	438532.8396	6663552.34	Y	Y	Y	1975	Amerada	1981.81m
204/30a- 2	443655.6327	6653401.599	Y	Y		1991	Amerada	3442.11m

Quadrant 205

Well Name	X location	Y location	Composite Log	Velocity Log	Biostratigraphic Report	Year drilled	Current Owner	TD
205/09- 1	479717.1108	6741448.283	Y	Y	Y	1989	BP	4748.78m
205/10- 1A	496841.2137	6735383.939	Y			1981	BP	2453.6m
205/10- 2B	496950.0404	6736438.887	Y	Y		1984	BP	998.52m
205/10- 3	499491.0229	6736333.923	Y	Y		1985	BP	2790.39m
205/14- 1	482755.9945	6711555.364	Y			1990	Shell	2738.33
205/16- 1	452282.7054	6690972.638	Y	Y		1986	BP	4313.0m
205/20- 1	493622.9671	6702898.024	Y	Y	Y	1974	Total	623.32m
205/21- 1A	453235.4768	6672190.859	Y	Y	Y	1974	Shell	1367.94m
205/21- 2	450962.9591	6678554.399	Y	Y		1990	Shell	3646.93m
205/22- 1A	457570.7403	6681758.735	Y			1974	BP	3229.0m
205/23- 1	476168.7222	6674154.126	Y	Y	Y	1975	ExxonMobil	2743.2m
205/25- 1	495120.0189	6670626.929	Y		Y	1977	ConocoPhillips	2600.25m
205/26- 1	449540.0614	6664324.598	Y		Y	1975	BP	2136.04m
205/26a- 2	450184.6008	6654175.332	Y	Y	Y	1982	BP	2467.97m
205/26a- 3	444385.397	6656747.813	Y			1990	Amerada	2875.36m
205/27a- 1	460631.2363	6659552.138	Y			1982	Shell	2626.77m
205/30- 1	490118.8164	6665801.184	Y	Y	Y	1974	Talisman	2194.56m

Quadrant 206

Well Name	X location	Y location	Composite Log	Velocity Log	Biostratigraphic Report	Year drilled	Current Owner	TD
206/01- 1A	508849.309	6746237.071	Y	Y		1985	BP	3032.15m
206/01- 2	506507.2186	6758919.017	Y	Y		1986	Shell	4532.68m
206/02- 1A	513026.8746	6762473.595	Y	Y	Y	1980	Shell	4493.97m
206/03- 1	530408.8905	6749528.343	Y	Y		1985	BP	4940.81m
206/04- 1			Y			1996	ConocoPhillips	4149.24m
206/05- 1	544941.8528	6756610.937	Y	Y	Y	1976	Total	4145.28m
206/08- 1A	524393.9367	6727888.273	Y	Y		1977	BP	2327.0m
206/08- 2	528298.6861	6733620.947	Y			1978	BP	1890.0m
206/08- 3	527185.299	6729190.62	Y			1978	BP	1945.0m
206/08- 3Z	527114.204	6729252.955	Y	Y		1978	BP	2577.0m
206/08- 4	527283.1307	6726592.24	Y	Y		1979	BP	3008.01m
206/08- 5	524515.3335	6727789.631	Y			1980	BP	2135.0m
206/08- 6A	522430.8349	6734233.352	Y	Y		1980	BP	2739.3m
206/08- 7	524528.6704	6727356.759	Y	Y		1985	BP	2334.4m
206/08- 8	524982.9821	6728647.311	Y	Y		1991	BP	2499.36m
206/08- 9Z	524931.2626	6728623.671	Y	Y		1992	BP	3725.88m
206/09- 1	538705.7679	6738301.258	Y	Y		1977	BP	2777.95m
206/09- 2	534455.0593	6739127.524		Y		1978	ExxonMobil	2464.92m
206/10a- 1	546274.4663	6741609.269	Y	Y		1980	ExxonMobil	2989.48m
206/11- 1	503147.26	6719272.333	Y			1977	BP	4620.01m
206/12- 1	512766.5381	6718026.045	Y			1972	Amerada	1727.0m
206/12- 2	513907.003	6723106.152	Y	Y		1977	Amerada	2563.98m

Quadrant 207

Well Name	X location	Y location	Composite Log	Velocity Log	Biostratigraphic Report	Year drilled	Current Owner	TD
207/01- 1	557948.4161	6762313.157	Y	Y		1977	ChevronTexaco	1466.7m
207/01- 2	560237.868	6757891.48	Y	Y	Y	1977	ChevronTexaco	1761.74m
207/01- 3	558666.896	6759705.753	Y	Y		1977	ChevronTexaco	1434.39m
207/01a- 4Z	555919.0653	6751440.597	Y			1990	ChevronTexaco	2682.24m
207/02- 1	568410.647	6761157.097	Y	Y		1974	Shell	2058.92m

Quadrant 208

Well Name	X location	Y location	Composite Log	Velocity Log	Biostratigraphic Report	Year drilled	Current Owner	TD
208/15- 1A	602361.2791	6836923.912	Y		Y	1979	BP	3165.17m
208/15- 2			Y			1995	BP	2549.56m
208/17- 1	564527.6831	6814401.036	Y	Y	Y	1985	BP	4846.32m
208/17- 2			Y			1995	BP	3702.84m
208/19- 1	587635.5754	6816214.017	Y	Y	Y	1983	BP	3334.51m
208/21- 1	562524.7673	6793723.684	Y	Y		1985	Total	3610.33m
208/22- 1	568717.209	6788148.501	Y	Y	Y	1986	ExxonMobil	2286.0m
208/23- 1	579867.8135	6782858.554	Y		Y	1983	ENI	2066.85m
208/24- 1A	588653.1071	6785344.585	Y	Y		1986	BP	2159.51m
208/26- 1	563558.1588	6776670.174	Y			1983	BP	3901.99m
208/27- 1	569106.2451	6773831.986	Y			1979	BP	1525.52m
208/27- 2	572857.727	6778686.23	Y	Y		1982	BP	1401.78m

Quadrant 209

Well Name	X location	Y location	Composite Log	Velocity Log	Biostratigraphic Report	Year drilled	Current Owner	TD
209/03- 1A	627532.6291	6860776.88	Y			1980	ExxonMobil	2316.48m
209/04- 1A	641860.7152	6868561.066	Y			1985	ENI	4051.71m
209/06- 1	612727.4001	6839641.388	Y		Y	1980	ChevronTexaco	3902.66m
209/09- 1	640148.0576	6850752.333	Y			1980	BP	2699.92m

Quadrant 213

Well Name	X location	Y location	Composite Log	Velocity Log	Biostratigraphic Report	Year drilled	Current Owner	TD
213/23- 1	476306	6788177	Y	Y	Y	1999	ExxonMobil	4374.79m

Quadrant 214

Well Name	X location	Y location	Composite Log	Velocity Log	Biostratigraphic Report	Year drilled	Current Owner	TD
214/04- 1	540175.13	6870659.85	Y	Y	Y	1999	ExxonMobil	4349.5m
214/17- 1	511909	6800924	Y	Y	Y	1998	ExxonMobil	2698.39m
214/19- 1			Y			1996	Shell	4867.0m
214/26- 1	509192.17	6776311.91	Y	Y	Y	1996	ConocoPhillips	2743.2m
214/27- 1	518747.2494	6770299.237	Y	Y	Y	1985	ChevronTexaco	5030.72m
214/27- 2	513722.1947	6776197.527	Y	Y	Y	1986	ChevronTexaco	4425.7m
214/28- 1	527376.3534	6775721.239	Y	Y	Y	1984	ExxonMobil	5124.3m
214/29- 1	541015.7927	6768409.936	Y	Y	Y	1985	Total	5032.71m
214/30- 1	548343.2414	6781254.688	Y	Y	Y	1984	BG	3323.84m

BGS Boreholes

Well Name	X location	Y location	Composite Log	Velocity Log	Biostratigraphic Report	Year drilled	Current Owner	TD
99/3	409047.46	6699788.17			Y	1999	BGS	166.5m
99/6	371966.87	6696922.79			Y	1999	BGS	36m

Appendix 4. Uppermost Palaeocene - Lower Eocene Biostratigraphic Database

The following biostratigraphic information from the thirteen type wells were used in this study:

	205/9-1		205/26a-2	
LOWER EOCENE	**M**	**TWTT**	**ft**	**TWTT**
(+ UPPERMOST PALAEOCENE)				
Coscinodiscus sp1. and sp2.	occur @ 2285m			
Inaperturopollenites hiatus	occur @ 2294m			
Caryapollenites simplex	occur @ 2294m			
Tillaepollenites microreticulatus	occur @ 2372m			
Eatonicysta ursulae	occur @ 2180m			
Cibicoides gr. eocaenus	occur @ 2189m			
Apectodinium spp.				
Deflandrea oebisteldensis				
Homotryblium pallidium				
Homotryblium tenuispinosum				
Cordosphaeridium gracile	occur @ 2180m			
Spiniferites ramosus	occur @ 2216m			
Hystrichosphaeridium tubiferum				
Areoligera spp.	occur @ 2180m			
Subbotina linaperta			occur @ 1416' (SWC)	
Charlesdowniea columna				
Dracododium spp.				
Dracododium politum				
A. coronata 'complex'	occur @ 2180m			
Spinozonocolpites echinatus				
Globigerina sp.	occur @ 2225m			
Globorotalia sp.			occur @ 1680'-1800'(SWC)	522
Spiroplectammina aff. spectabilis				
Stellarim microtrias				
Fenestrella antiqua				
Evolutinella spp.				

	204/22-1		204/23-1	
LOWER EOCENE	**M**	**TWTT**	**M**	**TWTT**
(+ UPPERMOST PALAEOCENE)				
Coscinodiscus sp1. and sp2.	high occur @ 1625m		influx @ 1890m	1966.58
Inaperturopollenites hiatus	increase occur @below 1500m			
Caryapollenites simplex	occur @ 1360m	1542.9		
Tillaepollenites microreticulatus	occur @ 1360m	1542.9	occur @ 1450m	
Eatonicysta ursulae	occur @ 1320m	1509.7	high occur @ 1260m	1439.1
Cibicoides gr. eocaenus				
Apectodinium spp.	occur @ 1630m-1680m		occur @ 1530m	1660
Deflandrea oebisteldensis	fda @ 1620m	1762		
Homotryblium pallidium	infux @ 1380m	1558.5		
Homotryblium tenuispinosum	infux @ 1380m	1558.5		
Cordosphaeridium gracile	increase @ 1380m	1558.5		
Spiniferites ramosus				
Hystrichosphaeridium tubiferum	increase occur @ 1620m			
Areoligera spp.				
Subbotina linaperta				
Charlesdowniea columna				
Dracododium spp.	occur @ 1320m	1509.7	occur @ 1200m	1387.7
Dracododium politum	fda @ 1420m	1592		
A. coronata 'complex'	occur @ 1420m	1592		
Spinozonocolpites echinatus	fda @ 1400m	1575.6		
Globigerina sp.				
Globorotalia sp.				
Spiroplectammina aff. spectabilis				
Stellarim microtrias				
Fenestrella antiqua				
Evolutinella spp.				

	204/24-1		214/26-1	
LOWER EOCENE	**M**	**TWTT**	**ft**	**TWTT**
(+ UPPERMOST PALAEOCENE)				
Coscinodiscus sp1. and sp2.			top @ 8595' (SWC)	2781.2
Inaperturopollenites hiatus	fda @ 1395m	1480		
Caryapollenites simplex	increase occur @ 1413m			
Tillaepollenites microreticulatus				
Eatonicysta ursulae	occur @ 1233m	1344	high occur @ 8166' (SWC)	2672.8
Cibicoides gr. eocaenus				
Apectodinium spp.				
Deflandrea oebisteldensis				
Homotryblium pallidium	high occur 1125m	1246		
Homotryblium tenuispinosum	high occur 1125m	1246		
Cordosphaeridium gracile	increase @ 1233m	1344	occur @ 8166' (SWC)	2672.8
Spiniferites ramosus	increase occur @1341m			
Hystrichosphaeridium tubiferum				
Areoligera spp.				
Subbotina linaperta			top @ 8350' (SWC)	2720.28
Charlesdowniea columna				
Dracododium spp.				
Dracododium politum				
A. coronata 'complex'	occur @ 1314m			
Spinozonocolpites echinatus				
Globigerina sp.				
Globorotalia sp.				
Spiroplectammina aff. spectabilis			top @ 5860'	2101
Stellarim microtrias				
Fenestrella antiqua				
Evolutinella spp.				

	214/27-1		214/29-1	
LOWER EOCENE	**ft**	**TWTT**	**M**	**TWTT**
(+ UPPERMOST PALAEOCENE)				
Coscinodiscus sp1. and sp2.	top @ 7840'	2520.2		
Inaperturopollenites hiatus			fda @ 1830m	1924.4
Caryapollenites simplex			occur @ 1840m	1931.7
Tillaepollenites microreticulatus				
Eatonicysta ursulae			occur @ 1690m	1802.1
Cibicoides gr. eocaenus				
Apectodinium spp.				
Deflandrea oebisteldensis	top @ 7190'	2366.1	occur @ 1750m	1860.3
Homotryblium pallidium				
Homotryblium tenuispinosum			occur @ 1690m	1802.1
Cordosphaeridium gracile	top @ 7250'	2382	occur @ 1690m	1802.1
Spiniferites ramosus				
Hystrichosphaeridium tubiferum	top @ 7370'	2413.5	occur @ 1780m	1887.1
Areoligera spp.	occur @ 7190'	2366.1	influx @ 1720m	1831.2
Subbotina linaperta	top @ 7220'	2374		
Charlesdowniea columna			occur @ 1690m	1802.1
Dracododium spp.	occur @ 7490'	2442.5	top solidum @ 1780m	1887.1
Dracododium politum				
A. coronata 'complex'				
Spinozonocolpites echinatus				
Globigerina sp.				
Globorotalia sp.				
Spiroplectammina aff. spectabilis				
Stellarim microtrias				
Fenestrella antiqua				
Evolutinella spp.				

	214/30-1		214/17-1	
LOWER EOCENE	ft	TWTT	ft	TWTT
(+ UPPERMOST PALAEOCENE)				
Coscinodiscus sp1. and sp2.	fda @ 7040'	1117.8		
Inaperturopollenites hiatus	fda @ 7040'	1117.8		
Caryapollenites simplex				
Tillaepollenites microreticulatus				
Eatonicysta ursulae				
Cibicoides gr. eocaenus				
Apectodinium spp.				
Deflandrea oebisteldensis	fda @ 7040'	1117.8		
Homotryblium pallidium				
Homotryblium tenuispinosum				
Cordosphaeridium gracile	occur @6700'	1073.7		
Spiniferites ramosus				
Hystrichosphaeridium tubiferum	fda @ 7040'	1117.8		
Areoligera spp.	fda @6580'	1057.3		
Subbotina linaperta	fda @ 6500'	1046.2	min influx 7175' (SWC)	
Charlesdowniea columna	occur @ 6500'	1046.2		
Dracododium spp.	fda @6580'	1057.3		
Dracododium politum				
A. coronata 'complex'				
Spinozonocolpites echinatus				
Globigerina sp.				
Globorotalia sp.				
Spiroplectammina aff. spectabilis	fda @ 6600'	1060	top 6135' (ROV)	
Stellarim microtrias				
Fenestrella antiqua				
Evolutinella spp.				

	208/19-1		213/23-1	
LOWER EOCENE	**ft**	**TWTT**	**ft**	**TWTT**
(+ UPPERMOST PALAEOCENE)				
Coscinodiscus sp1. and sp2.	top @ 4210'	1418.2		
Inaperturopollenites hiatus	occur @ 4180'	1409	top abundant@ 8963' (SWC)	
Caryapollenites simplex	top @ 4210'	1418.2	top @ 8963'(SWC)	
Tillaepollenites microreticulatus				
Eatonicysta ursulae	occur @ 4060'	1363.9	top @ 8272' (SWC)	
Cibicoides gr. eocaenus			occur @ 8400'(SWC)	
Apectodinium spp.			occur @ 9470'	
Deflandrea oebisteldensis	top @ 4240'	1427.5	top freq @ 8693'(SWC)	
Homotryblium pallidium				
Homotryblium tenuispinosum	occur @ 4060'	1363.9	top @ 8272' (SWC)	
Cordosphaeridium gracile	occur @4060'	1363.9	top @ 8580'(SWC)	
Spiniferites ramosus	occur @4060'	1363.9	top @ 8580'(SWC)	
Hystrichosphaeridium tubiferum	top @ 4210'	1418.2	top @ 8677'(SWC)	
Areoligera spp.	top @ 4210'	1418.2	top @ 8570'(SWC)	
Subbotina linaperta			Occur @ 8570' (SWC)	
Charlesdowniea columna				
Dracododium spp.			top @ 8272' (SWC)	
Dracododium politum				
A. coronata 'complex'				
Spinozonocolpites echinatus				
Globigerina sp.				
Globorotalia sp.				
Spiroplectammina aff. spectabilis				
Stellarim microtrias				
Fenestrella antiqua			reworked @ 8670'(SWC)	
Evolutinella spp.			Top @ 9250'(SWC)	

	214/4-1	
LOWER EOCENE	**ft**	**TWTT**
(+ UPPERMOST PALAEOCENE)		
Coscinodiscus sp1. and sp2.	occur @ 11480'	
Inaperturopollenites hiatus	occur @ 11360'	
Caryapollenites simplex	occur @ 11360'	
Tillaepollenites microreticulatus		
Eatonicysta ursulae	top @ 10700' (SWC)	
Cibicoides gr. eocaenus		
Apectodinium spp.		
Deflandrea oebisteldensis	top @ 11507'	
Homotryblium pallidium		
Homotryblium tenuispinosum	occur @ 11015'	
Cordosphaeridium gracile	occur @ 11015' (SWC)	
Spiniferites ramosus	occur @ 11469'	
Hystrichosphaeridium tubiferum	top @ 10950' (SWC)	
Areoligera spp.	top @ 11250'	
Subbotina linaperta	top @ 10773'	
Charlesdowniea columna	top @ 10837' (SWC)	
Dracododium spp.	top @ 11103' (SWC)	
Dracododium politum		
A. coronata 'complex'		
Spinozonocolpites echinatus		
Globigerina sp.		
Globorotalia sp.		
Spiroplectammina aff. spectabilis		
Stellarim microtrias	occur @ 11480'	
Fenestrella antiqua	top @ 11480'	
Evolutinella spp.	top @ 11480'	

Appendix 5. Middle Eocene Biostratigraphic Database

The following biostratigraphic information from the thirteen type wells were used in this study:

	205/9-1		205/26a-2	
MIDDLE EOCENE	**M**	**TWTT**	**ft**	**TWTT**
Cenosphaera spp.				
Cyclammina amplectens	high occur @ 2030m			
Apectodinium homorphum	occur @ 2198m			
Wetzelielia spp.				
Neoponides karsteni	high occur @ 1710m			
Turrilina brevispira				
Membranophordium cf. aspinatum				
Heteraulacacysta porosa				
Diphyes fiscisoides				
Systematophora placacantha				
Stellarima microtrias				
Subbotina eocaena			high occur @ 1636' (SWC)	509
Cibicides spp.				
Rhabdammina robusta	occur @ 2130m			
Loxoconcha curryi				
Deflandrea phosphoritica				
Pseudohastigerina micra	high occur @ 1290m			
Truncororotaloides cf. rohri	high occur @ 1290m			
Diphyes colligerum				
Phthanoperidinium spp.				
Areosphaeridum arcuatum				

	204/22-1		204/23-1	
MIDDLE EOCENE	**M**	**TWTT**	**M**	**TWTT**
Cenosphaera spp.				
Cyclammina amplectens	high occur @ 1430m	1605.3		
Apectodinium homorphum	fda @ 1380m	1558.5	occurr @ 1690m +	
Wetzelielia spp.	occur @ 1300m	1492.5	occur @ 1200m	1387.7
Neoponides karsteni				
Turrilina brevispira	occur @ 1310m			
Membranophordium cf. aspinatum	occur @ 1300m	1492.5		
Heteraulacacysta porosa				
Diphyes fiscisoides			occur @ 1230m	1414.3
Systematophora placacantha				
Stellarima microtrias				
Subbotina eocaena				
Cibicides spp.			occur @ 1210m	1397

Rhabdammina robusta				
Loxoconcha curryi				
Deflandrea phosphoritica				
Pseudohastigerina micra				
Truncororotaloides cf. rohri				
Diphyes colligerum				
Phthanoperidinium spp.				
Areosphaeridum arcuatum				

	204/24-1		214/26-1	
MIDDLE EOCENE	**M**	**TWTT**	**ft**	**TWTT**
Cenosphaera spp.	single occur @ 1242m	1351.5	top @ 5440'	1986
Cyclammina amplectens			top @ 5440'	1986
Apectodinium homorphum	increase occur @ 1368m			
Wetzelielia spp.	fda @1152m	1271		
Neoponides karsteni	high occur 1080m			
Turrilina brevispira				
Membranophordium cf. aspinatum	high occur 1152m	1271		
Heteraulacacysta porosa	high occur 1044m			
Diphyes fiscisoides	high occur 1125m	1246	top @ 6050'	2135
Systematophora placacantha			top @ 5910'	2115
Stellarima microtrias				
Subbotina eocaena			occur @6240'	
Cibicides spp.				
Rhabdammina robusta				
Loxoconcha curryi				
Deflandrea phosphoritica	top persistent 1044m			
Pseudohastigerina micra				
Truncororotaloides cf. rohri				
Diphyes colligerum				
Phthanoperidinium spp.				
Areosphaeridum arcuatum				

	214/27-1		214/29-1	
MIDDLE EOCENE	**ft**	**TWTT**	**M**	**TWTT**
Cenosphaera spp.	occur @ 7160'	2337.2	occur @ 1700m	1812.6
Cyclammina amplectens	base @ 7220'	2374		
Apectodinium homorphum				
Wetzelielia spp.	base @ 7490'	2442.5		
Neoponides karsteni				
Turrilina brevispira				
Membranophordium cf. aspinatum				
Heteraulacacysta porosa				
Diphyes fiscisoides				
Systematophora placacantha				
Stellarima microtrias				
Subbotina eocaena				

Cibicides spp.				
Rhabdammina robusta				
Loxoconcha curryi				
Deflandrea phosphoritica				
Pseudohastigerina micra				
Truncororotaloides cf. rohri				
Diphyes colligerum				
Phthanoperidinium spp.				
Areosphaeridum arcuatum				

	214/30-1		214/17-1	
MIDDLE EOCENE	**ft**	**TWTT**	**ft**	**TWTT**
Cenosphaera spp.			top 6030' (ROV)	
Cyclammina amplectens	base @ 6800'	1087.4	top 6320'	
Apectodinium homorphum				
Wetzelielia spp.				
Neoponides karsteni				
Turrilina brevispira				
Membranophordium cf. aspinatum				
Heteraulacacysta porosa				
Diphyes fiscisoides			top consistent 7500'	
Systematophora placacantha			top consistent 6030' (ROV)	
Stellarima microtrias			top 6560'	
Subbotina eocaena			min influx 6674' (SWC)	
Cibicides spp.			min influx 8300'	
Rhabdammina robusta				
Loxoconcha curryi				
Deflandrea phosphoritica				
Pseudohastigerina micra				
Truncororotaloides cf. rohri				
Diphyes colligerum				
Phthanoperidinium spp.				
Areosphaeridum arcuatum				

	208/19-1		213/23-1	
MIDDLE EOCENE	**ft**	**TWTT**	**ft**	**TWTT**
Cenosphaera spp.	occur @ 4120'	1418.2		
Cyclammina amplectens			top @ 6660' (SWC)	
Apectodinium homorphum				
Wetzelielia spp.	occur @4060'	1363.9	top @ 8100'	
Neoponides karsteni				
Turrilina brevispira				
Membranophordium cf. aspinatum				
Heteraulacacysta porosa				

Diphyes fiscisoides			top @ 7600' (SWC)	
Systematophora placacantha			top @ 5837' (SWC)	
Stellarima microtrias				
Subbotina eocaena				
Cibicides spp.				
Rhabdammina robusta				
Loxoconcha curryi				
Deflandrea phosphoritica			occur @ 4882'	
Pseudohastigerina micra				
Truncororotaloides cf. rohri				
Diphyes colligerum				
Phthanoperidinium spp.			top @ 6650'	
Areosphaeridum arcuatum			top @ 6650'	

	214/4-1	
MIDDLE EOCENE	**ft**	**TWTT**
Cenosphaera spp.	occur @ 8230'	
Cyclammina amplectens		
Apectodinium homorphum		
Wetzelielia spp.		
Neoponides karsteni		
Turrilina brevispira		
Membranophordium cf. aspinatum		
Heteraulacacysta porosa		
Diphyes fiscisoides	top @ 8931 (SWC)	
Systematophora placacantha	increase @ 8230'	
Stellarima microtrias		
Subbotina eocaena		
Cibicides spp.	occur @ 8725' (SWC)	
Rhabdammina robusta		
Loxoconcha curryi		
Deflandrea phosphoritica		
Pseudohastigerina micra		
Truncororotaloides cf. rohri		
Diphyes colligerum	top @ 8230'	
Phthanoperidinium spp.	top @ 8230'	
Areosphaeridum arcuatum	top @ 8390'	

Appendix 6. Upper Eocene Biostratigraphic Database

The following biostratigraphic information from the thirteen type wells were used in this study:

	205/9-1		205/26a-2	
LATE EOCENE	**M**	**TWTT**	**ft**	**TWTT**
Cibicides westi	occur @ 1250m			
Bolivina cookei	influx @ 1270m			
Uvigerina eocaena	occur @ 1250m			
Corrudinium inconpositum				
Trilites spp.				
Heteraulacysta porosa				
Phthanoperidinium spp				
Adnatosphaeridium spp.				
Areosphaeridium spp.				

	204/22-1		204/23-1	
LATE EOCENE	**M**	**TWTT**	**M**	**TWTT**
Cibicides westi			occur @ 1210m	
Bolivina cookei				
Uvigerina eocaena				
Corrudinium inconpositum				
Trilites spp.				
Heteraulacysta porosa				
Phthanoperidinium spp				
Adnatosphaeridium spp.				
Areosphaeridium spp.				

	204/24-1		214/26-1	
LATE EOCENE	**M**	**TWTT**	**ft**	**TWTT**
Cibicides westi				
Bolivina cookei				
Uvigerina eocaena				
Corrudinium inconpositum			occur @ 5290'	
Trilites spp.			occur @ 5290'	
Heteraulacysta porosa				
Phthanoperidinium spp				
Adnatosphaeridium spp.				
Areosphaeridium spp.				

	214/27-1		214/29-1	
LATE EOCENE	**ft**	**TWTT**	**M**	**TWTT**
Cibicides westi				
Bolivina cookei				
Uvigerina eocaena				
Corrudinium inconpositum				
Trilites spp.				
Heteraulacysta porosa				
Phthanoperidinium spp				
Adnatosphaeridium spp.				
Areosphaeridium spp.				

	214/30-1		214/17-1	
LATE EOCENE	**ft**	**TWTT**	**ft**	**TWTT**
Cibicides westi				
Bolivina cookei				
Uvigerina eocaena				
Corrudinium inconpositum				
Trilites spp.				
Heteraulacysta porosa				
Phthanoperidinium spp				
Adnatosphaeridium spp.				
Areosphaeridium spp.				

	208/19-1		213/23-1	
LATE EOCENE	**ft**	**TWTT**	**ft**	**TWTT**
Cibicides westi				
Bolivina cookei				
Uvigerina eocaena				
Corrudinium inconpositum				
Trilites spp.				
Heteraulacysta porosa			**occur @ 5349'**	
Phthanoperidinium spp				
Adnatosphaeridium spp.				
Areosphaeridium spp.			**occur @ 4600'**	

	214/4-1	
LATE EOCENE	**ft**	**TWTT**
Cibicides westi		
Bolivina cookei		
Uvigerina eocaena		
Corrudinium inconpositum		
Trilites spp.		
Heteraulacysta porosa	**top @ 7812'**	
Phthanoperidinium spp	**common @ 7958'**	

Adnatosphaeridium spp.	**top @ 8150'**	
Areosphaeridium spp.	**top @ 7812'**	