Late quaternary palaeovalley systems of the eastern English channel

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Thesis submitted for the degree of Doctor of Philosophy.

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September 2004

DECLARATION

This thesis is submitted in accordance with the requirements of the University of Durham for the Degree of Doctor of Philosophy.

This thesis is the result of the authors own work. References to other authors contained herein are acknowledged at the appropriate point in the text.

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Acknowledgments

This PhD research project was funded under the Natural Environment Research Council (NERC) Industrial CASE Partnership award scheme. Reference number NER/S/C/2000/03294. The research was undertaken in partnership with the Department of Geography at the University of Durham, and the Industrial CASE partners Hanson Aggregates Marine Limited and the British Geological Survey.

I gratefully acknowledge the unstinting help, advice and encouragement of my departmental supervisors Dr. Antony J. Long and Dr. David R. Bridgland. I also wish to express my gratitude to my external supervisors Dr. Ian Selby of Hanson Aggregates Marine Limited and Dr. Peter S. Balson of the British Geological Survey, and my examiners Dr .Yonqiang Zong and Dr. Brian D'Olier.

I also wish to express my thanks the following people for their friendship, help, encouragement and support throughout my time at Durham. Dr.'s Katherine Arrell, James Smith, Ben Horton and Dave Roberts.

I would also like to thank the staff and postgraduates of the Department of Geography, University of Durham, too numerous to mention, for their support and friendship. In particular I would like to thank Frank Davies and Stella Henderson for their assistance in my hours of need.

Oh and thanks again Antony ! I couldn't have pulled it off without you.

This Doctoral Thesis is dedicated to my mum who's support in whatever I do is unwavering. I did it mum !

Matthew R. Wright

Durham, September 2004.

ABSTRACT

LATE QUATERNARY PALAEOVALLEYS OF THE EASTERN ENGLISH CHANNEL

Matthew R. Wright, University of Durham, September 2004

This thesis examines the Late Quaternary palaeoenvironmental history of the palaeovalley system located on the sea bed of the eastern English Channel. The work is based on an analysis of 500 km of high resolution seismic data, 189 boreholes, and an extensive grain size, microfossil and geochemical dataset. Within the study area are two major palaeovalleys incised into Tertiary bedrock. They are on average 2 to 8 km wide and contain between 20 and 30 m of Late Quaternary sediments. Four Seismic Sequences are identified, separated by two sequence boundaries and one ravinement surface. Using an uplift-corrected eustatic sea level as a relative sea-level history for the study area, it is suggested that the palaeovalleys were cut at the transition from MOIS6 to 5e, and were infilled with transgressional shoreface sediments during the interval between MOIS5e and MOIS4. An upper seismic unit, also comprising marine sediment, developed following the Last Glacial Maximum as an extensive lag deposit. Comparison of this work with other records from the English Channel and other continental shelves demonstrate the pervasive importance of eustatic and tectonic controls in determining palaeovalley development. The major palaeovalleys of the English Channel probably originate from MOIS12, created by the drainage of a large proglacial lake in the North Sea via the Strait of Dover. The palaeovalleys in the PhD study area post-date this initial phase of valley development, and were formed by a major river flowing through the region (presumably one or more the Thames, Rhine or Meuse). The results of this PhD demonstrate the high potential of continental shelf records as archives of Late Quaternary land-ocean interaction, as well as the benefits of close liaison between industry and academic stakeholders involved in offshore aggregate and palaeoenvironmental exploration.

TABLE OF CONTENTS

LIST OF FIGURES

LIST OF TABLES

Chapter 1: Introduction

This chapter introduces the research undertaken in this PhD thesis by outlining the rationale for undertaking the study and detailing the main research questions addressed. It also provides a brief description of the location of the PhD study area and explains why this area was selected for investigation. The chapter concludes with a brief outline of the structure of the thesis.

1.1 Background

The investigation of deep sea marine sediments has revolutionised our understanding of Quaternary timescales and processes (Emiliani, 1955). Sediments from these environments have the advantage that they lie beyond the limits of the continental ice sheets, and therefore provide a relatively continuous record of sedimentation that spans many millions of years. They are, by their geographical location, ideal for determining changes in water mass characteristics through time, and thus provide a powerful means to reconstruct changes in ocean volume and glacio-eustatic sea level (e.g. Shackleton, 1987). In contrast, our understanding of the shallower continental shelves has provided a different type of information regarding Quaternary environmental change. These shelves, alternately exposed to subaerial processes during eustatic lowstands and flooded by the sea during highstands, contain valuable information regarding the link between terrestrial and marine environments. In particular, they contain a sediment archive that documents the varying importance of marine, nearshore and continental / fluvial processes during the long sequence of glacial-interglacial cycles that characterises the last 2.6 Ma. As such, shelf sea sediment archives provide an important link between the deep sea and continental records of Quaternary environmental change.

The English Channel is an important continental shelf sea that separates Britain from mainland Europe, and the North Atlantic from the southern North Sea Basin (Figure 1-1). From a Quaternary perspective, the Channel is important for several reasons. In the Western Approaches, the Channel contains important geomorphological evidence for former ice limits associated with the maximum extent of the British Late Devensian ice sheet (Scourse, 1991; Evans and Ó Cafaigh, 2003). But, perhaps more significantly, the eastern English Channel and Strait of Dover is a critical area in controlling the isolation and connection of mainland Britain to the continent (Preece, 1995). Indeed, if present sea-level dropped by between 40 to 50 m, Britain would be connected to continental Europe by a land-bridge in this region. It is now well known that this situation has occurred several times during the Late Quaternary period, and that the presence of this bridge has had a profound influence on the migration and subsequent development of the fauna and flora of not just Britain but the the British Isles more widely (Meijer and Preece, 1995). Indeed, the Strait of Dover land bridge has also played a major role as a barrier impeding the spread of marine organisms and the transfer of energy and material from the English Channel into the southern North Sea (Scourse and Austin, 1995).

Despite the importance of the eastern English Channel to the Quaternary history of the region, several factors mean that the sedimentary record from the area is neither continuous nor, in many instances, well preserved. In particular, the exposed aspect of the region combined with the shallow waters of the area mean that wave and tidal influences have a tendency to rework seabed sediments, especially during periods when relative sea-level fell or rose across the Strait of Dover threshold. Because of this, the eastern English Channel is not generally acknowledged as supporting great quantities of sea bed sediments. However, the seabed of the eastern English Channel does, in fact, contain a dense and extensive network of palaeovalleys that incise into bedrock and which have, during the Late Quaternary, acted as depocentres for the accumulation of fluvial and marine sediments during lowstands and highstands respectively. These palaeovalleys indicate that the region has repeatedly supported one or more very large river systems during the Quaternary. As a result, the area is a potential key to improving our understanding of the palaeohydrology and palaeogeographical evolution of northwest Europe during the Quaternary.

The palaeovalley systems in the eastern English Channel are also important because their overlying and infilling deposits are of significant economic value. Much of the sediments that infill and overly these palaeovalleys are sands and gravels that are of ideal quality for extraction and use as aggregate in the construction industry. Indeed, around 21% of the sand and gravel used in England and Wales is now supplied by the marine aggregates. In the United Kingdom alone, aggregate requirements equate to more than five tonnes per head of the population each year, whilst since 1955 a total of ~500 million tonnes of aggregates have been dredged from the sea and used for building and construction. Marine aggregates also play a front-line role in replenishing the beaches of Britain and mainland Europe, thus protecting the coastline from erosion and flooding by the sea industry (British Marine Aggregate Producers Association, 2000).

This PhD programme has developed from a set of complimentary interests in the Quaternary history of the English Channel and the economic potential of the marine aggregates that exist in this region. The work has been completed under a NERC Industrial CASE studentship, and has involved the close collaboration between three institutions – The University of Durham, The British Geological Survey (BGS), and Hanson Aggregates Marine Limited (HAML).

1.2 PhD study area

The study area for this PhD research was selected and defined by resource geologists at Hanson Aggregates Marine Limited (HAML), who consider that the seabed in the area may potentially yield significant gravel and sand aggregate resource. The PhD study area comprises two areas of proposed aggregate production (EEC474 and EEC475; Figure 1-1) for which HAML have submitted applications to Crown Estates United Kingdom for licences to extract marine aggregates. Eastern English Channel (EEC) License Application Areas 474 and 475 are located in the central English Channel, offshore of the English coast approximately 40 km south of Eastbourne and 75 km west of the nearest French coast (National Grid Reference 550000E, 050000N). The two license application areas lie within the English Channel Traffic Separation Scheme, which is composed of a south west bound shipping lane, a north east bound shipping lane and separation zone between the two (Figure 1-1). EEC474 is located entirely within the northeast bound shipping land and covers an area of approximately 121 km^2. EEC475 covers an area of approximately 110 km^2 of which most is also within the north-east bound shipping lane. The PhD study area lies between 2 –10 km north of the UK / France Median Line.

1.3 Research Rationale

1.3.1 Research aim

The overall aim of the thesis is *to develop a model for the Late Quaternary evolution of the eastern English Channel based on an investigation of the palaeovalleys of the region.*

1.3.2 Research approach

To achieve the above aim, the thesis undertakes the following:

1. The collection and interpretation of new high resolution seismic data from the eastern English Channel using the concepts of seismic and sequence stratigraphy.
2. The retrieval and analysis of sediment cores from the study area to determine their physical characteristics and depositional origins.
3. Comparison of these data with other records collected from analogous continental shelves elsewhere in the world.
4. To develop of a hypothetical age model for the depositional sequence in the PhD study area.

5. To test and develop further this model through its comparison with depositional sequences from elsewhere in the English Channel and the southern North Sea Basin.

1.4 Thesis Structure

This thesis is produced in two parts:

PART 1 - Text and References

Chapter 2 provides a review of literature relevant to the evolution of continental shelves in the Late Quaternary period (~last 500 ka) with particular reference to the formation of incised valley features and their associated infilling sediments.

Chapter 3 reviews the Late Quaternary history of the English Channel. The objectives of the chapter are to describe geology of the English Channel, to outline the climate and sea-level change history of the English Channel during the Quaternary period, and to critically review the physiography, deposits and morphological features of the region. This chapter forms a necessary context for the interpretation of data collected during this PhD study.

Chapter 4 describes the field programme and laboratory techniques that were employed to investigate the nature and origins of the palaeovalleys and associated infilling sediments studied in this thesis.

Chapter 5 presents the results of the field surveys (shallow seismic reflection profiling and vibrocoring) undertaken in the study area and the subsequent analysis of the data retrieved.

Chapter 6 proposes a model for the age, nature and origin of the incised palaeovalley system located in the PhD study area.

Chapter 7 uses the model proposed in *Chapter 6* as a basis to explore a range of issues relating to the wider geographical evolution of the English Channel during the Late Quaternary.

Chapter 8 presents the conclusions of this thesis, evaluates the success of the research techniques employed and makes recommendations for future work.

PART 2 – Figures, Tables and Appendices

Tables in the thesis are listed by chapter.

Figures in the thesis are grouped according to chapter.

Appendix 1 Outlines the methodology of the laboratory techniques utilised during this PhD study.

Appendix 2, 3 and 4 are presented as computer files on a CD in the back cover of the volume in both Microsoft Excel and comma delimited text formats.

Appendix 2 presents a table of the raw data obtained during grain-size analysis of sediment samples during this PhD study. File Name: GSData.xls and GSData.txt

Appendix 3 presents a table of the raw data obtained during micropalaeontological analyses of sediment samples during this PhD study. File Name: Micropal.xls and Micropal.txt

Appendix 4 presents a table of the data obtained from the analysis of multi-element chemistry of sediment samples during this PhD study. Geochem.xls and Geochem.txt

Chapter 2: Quaternary Continental Shelves

2.1 Introduction

This chapter reviews the literature relevant to the evolution of continental shelves in the Late Quaternary period (~last 500 ka) with particular reference to the formation of incised valley features and their associated infilling sediments. It begins by considering a general account of Late Quaternary erosion and sedimentation on continental shelves from across the globe. This is followed by a brief review of incised-valley research completed on continental shelves, and an account of the nature of incised valley fill sediment architecture. Finally the sequence stratigraphic approach and methodology applied to study these features is described.

2.2 Continental Shelves in the Late Quaternary

Erosion and sedimentation on continental shelves during the Late Quaternary (~last 500 ka) has been affected by five major glacial-interglacial cycles and their associated glacio-eustatic variations (Figure 2-1). These cycles, which were characterised by cold glacial stages interspersed with warmer interglacial episodes, had significant effects on the global volume of ice and therefore the related eustatic sea level. During glacial stages large volumes of water were stored on land as continental ice sheets, consequently lowering eustatic sea-levels globally by as much as 120 metres. During interglacials, warmer temperatures led to the ice sheets melting, thus increasing the volume of water in the oceans and raising eustatic sea level. Fisk (1939) was one of the first researchers to notice this interdependence between eustatic sea level and palaeotemperature change and their associated impact on continental shelf systems while studying the coastal plain and shelf of Louisiana and Texas. He saw that the origins of features such as eroded valleys, fluvial terraces and other deposits in the region could be explained as a product of processes associated with the Quaternary glacial – interglacial history of the region.

2.2.1 North America

Curray (1965) discussed the physiographic and depositional products of Late Quaternary eustatic fluctuations on the continental shelves of the United States. He demonstrated that during glacial maxima, the majority of the continental shelf around the US was subaerially exposed and subject to alluvial deposition and erosion. He suggested that during recession of continental glaciers, and subsequent transgression of sea level across the shelf, trangressional shoreline migration would have occurred at such speed as to have out-paced deposition, leaving relict sub-aerial and shallow-water landforms and deposits exposed on the shelf surface.

Gulf of Mexico

Many researchers have studied the late Quaternary morphology and stratigraphy of the Texas and Louisiana continental shelves of the Gulf of Mexico (e.g. Fisk, 1939; Bloom, 1983; Suter, 1986; Berryhill, 1986a, 1986b; Thomas and Anderson, 1994; Anderson *et al.*, 1996; Abdullah *et al.*, 1999). The dominant feature of the shelf here is the Mississippi delta plain. During the Holocene it has prograded across the entire width of the shelf and currently discharges sediment off the continental slope. Bloom (1983) suggested that in full glacial times the Mississippi river flowed though a deeply incised valley (the 'Mississippi Trench') which, during retreat of the southern quadrant of the Laurentide ice sheet margin was initially infilled by glacial outwash gravels and sands deposited in a braided river system. This was followed by aggradations of silt and clay units, deposited by a meandering Mississippi river system which flowed out from the extensive proglacial Great Lakes that developed along the southern ice margin.

Offshore of the Mississippi delta, Coleman and Roberts (1988) investigated the shelf stratigraphy in the Gulf of Mexico and related the Late Quaternary eustatic sea-level cycle inferred from the marine oxygen isotope stage (MOIS) $\delta^{18}O$ record, to sedimentary units on the shelf. Thomas and Anderson (1994) further developed this theme in the Gulf of Mexico by using the SPECMAP $\delta^{18}O$ curve of Imbrie *et al.* (1984) as a chronological and conceptual framework for the interpretation of seismic reflection profile sequences for the Late Quaternary incised-valley system of the offshore extension of the Trinity / Sabine rivers. They noted that the 100 ka and 20 ka cycles controlled the incision of the palaeovalley system with initiation of the system approximately 110 ka BP (MOIS5d) as sea level fell from the MOIS5e highstand. They also deduced that the lowstand sub-stages (MOIS5d and MOIS5b) were represented in the seismic stratigraphy as sequence boundaries, and highstand sea-level sub-stages (MOIS5e, 5c, and 5a) by deposits that infill the palaeovalleys.

Pacific Coast

On the Pacific coast of the USA, Burger *et al.* (2001) noted that the frequency of Late Quaternary channel incision by the southern Eel River on the shelf of northern California exceeded that of known glacio-eustatic lowstands over the last 500 ka BP. This suggests that tectonic and eustatic changes have combined to control sediment patterns on the shelf in the region. The significance of Quaternary tectonic uplift of the California shelf was supported by Bloom (1983). He described conspicuous sets of emerged terraces on California's mountainous coast and offshore islands, most of which were erosional. The cause of this uplift is attributed to the San Andreas Fault system, and in areas of strongest uplift Holocene terraces have emerged by as much as 40 m (Bloom, 1983; Anderson and Menking, 1994; Perg *et al.*, 2001).

Atlantic Coast

Knebel *et al.* (1979), Swift *et al.* (1980), Nordfjord *et al.* (2002) and Fulthorpe *et al.* (2002) investigated the Late Quaternary incision and related subsurface shallow stratigraphy on the New Jersey Shelf. Complex, dendritic networks of probable fluvial channels shallowly buried beneath the seabed exist in this region. Mapping of deep-towed Chirp seismic reflection profiles by Nordfjord *et al.* (2002) and Fulthorpe *et al.* (2002) revealed channel systems of a range of scales from hundreds of metres wide and metres deep up to the very substantial buried palaeo-Hudson channel (~40 m deep and 2 km wide). The complex nature of the systems was indicated by multiple cut-and-fill morphologies identified from the seismic data.

Research by Sager and Riggs (1998) produced a model for the Holocene history of Albermarle Sound, North Carolina, where integrated seismic-, litho- and chronostratigraphy revealed complex multi-sequence infilling of the incised palaeo-Roanaoke River valley. Similar studies have been undertaken to the north of this in the major coastal embayments of Chesapeake Bay and Delaware Bay (Belknap and Kraft, 1985; Foyle and Oertel, 1992, 1997; Belknap *et al.* (1994) where stratigraphic sequences representing longer Quaternary timescales have been preserved. These are discussed more fully elsewhere in this thesis.

2.2.2 South and East China Seas

Much of the work carried out on the shelf of the East China Sea has been concentrated on the delta and further offshore of the of the Yangtze (Changjiang) river (Evans *et al.*, 1995; Liu *et al.*, 2000; Li *et al.*, 2002; Yoo *et al.*, 2002). Evans *et al.* (1995) identified a series of stacked infilled channels on the shelf in the region, which were interpreted as Late Quaternary estuarine and fluvial sequences deposited in the palaeovalley of the Yangtze during the last glacial lowstand.

Liu *et al.* (2000) also collected geophysical data from the East China Sea shelf. They correlated these to a borehole retrieved from the outer shelf and inferred that six sedimentary units had been deposited since MOIS6, and that these recorded alternating continental and marine strata corresponding to glacial interglacial cycles. Li *et al.* (2002) established a Late Quaternary stratigraphic framework for the evolution of the incised valley fill of the Yangtze delta based on the analyses and correlation of some 600 boreholes, offshore on the mid- and outer-shelf of the East China Sea. Yoo *et al.* (2002) conducted analyses of seismic profile and sediment data to identify sedimentary sequences above the acoustic basement. Highstand and transgressive systems tracts were identified and interpreted as originating after the last glacial maximum.

The Quaternary sedimentary sequence offshore of Hong Kong in the South China Sea has been studied in detail (Yim *et al.*, 1988; Yim and Yu, 1993; Evans *et al.*, 1995; Fyfe *et al.*, 1997 and 1999). The Hong Kong Geological Survey (HKGS) produced 1:20000 scale mapping and associated memoir series of the offshore unconsolidated geology (Strange and Shaw, 1986; Strange *et al.*, 1990; James, 1993). They divided the Quaternary into a four-fold sequence (Figures 2-2 and 2-3):

(i) The lowermost Chek Lap Kok Formation is considered to be Mid to Late Quaternary in age (Fyfe *et al.*, 1997). It overlies a weathered bedrock surface over most of the Hong Kong offshore area. HKGS interpret the Chek Lap Kok Formation as representing predominantly fluvial sedimentation from the palaeo Pearl River Delta.

(ii) The Sham Wat Formation lies unconformably over the Chek Lap Kok. The age of the formation is proposed as being of MOIS5e age and of probable marine origin (Evans *et al.*, 1995). Locally the formation is topped with up to 4 m of palaeosol development (Fyfe *et al.*, 1999).

(iii) The Waglan Formation is recognised where a more complete Late Quaternary succession is developed in the southeastern part of Hong Kong's waters. It is considered to be MOIS5c age, with marine sediment deposited during a potential rise in global sea level (Fyfe *et al.*, 1997).

(iv) The uppermost Hang Hau Formation is the most widely distributed superficial unit in the region and underlies most of the seabed. This mud sheet is interpreted as a regressive estuarine to marine unit deposited in association with the Holocene sea level rise and subsequent highstand (Evans *et al.*, 1995).

2.2.3 Europe

In the Gulf de Lions (Western Mediterranean Sea, France), Tesson *et al.* (1990) and Rabineau *et al.* (1998) surveyed the outer continental shelf in order to reconstruct the three-dimensional architecture of well-preserved large Quaternary sand bodies. The incised features and deposits they identified were considered to be relicts of lowstand and transgressive erosion and deposition in the western part of the palaeo-Rhône river delta (Rabineau *et al.*, 1998).

To the west, in the Bay of Biscay, recent studies have been carried out into the incised valley and infilling internal stratigraphy of the Gironde River (Allen and Posamentier, 1993; 1994; Lericolais *et al.*, 2001). Following sequence stratigraphic interpretation of very high resolution seismic data obtained from the offshore incised-valley fill, Lericolais *et al.* (2001) deemed that it contained a single sequence corresponding to a fifth-order eustatic cycle (~20 ka). These findings correlated the offshore stratigraphy with earlier work onshore in the inner Gironde Estuary, by identifying continuations of the three systems tracts first described by Allen and Posamentier (1993, 1994) on to the shelf.

Such is the severity of erosion and denudation of glacial action in northern Europe that Sejrup *et al.* (1996) stated up to 90% of Late Quaternary sedimentation in the basin of the Norwegian Sea and adjacent continental margin occurred during cold stages. This sedimentation has, in some areas of the shelf and continental slope, resulted in the formation of sediment sequences in excess of 200 m thick. Similarly the continental shelf of the North Sea basin is dominated by extensive (up to 800 m) thicknesses of Quaternary deposits of glacial, glacio-marine and deltaic origin (Cameron *et al.*, 1992; Gatliff *et al.*, 1994). This thick and laterally extensive offshore depositional system is in direct contrast to the sediment starved floor shelf of the English Channel investigated in the present study.

2.3 Incised-Valley Systems on Continental Shelves

A characteristic element of the topography of continental shelves is the presence of incised valley systems. These systems are significant because they record the imprint of palaeodrainage systems comprising former fluvial networks that extended beyond their current estuarine limits onto the shelf during periods of lowered sea-level. Zaitlin *et al.* (1994) defined the most common type of incised-valley system as:

> "a fluvially-eroded, elongate topographic low that is typically larger than a single channel form, and is characterised by an abrupt seaward shift of depositional facies across a regionally mappable sequence boundary at its base. The fill typically begins to accumulate during the next base-level rise, and may contain deposits of the following highstand and subsequent sea-level cycles."

They also stated that these systems on shallow-gradient shelf settings typically extend landward of a lowstand mouth / delta of the incised valley, to a location beyond which relative sea-level changes no longer influence fluvial erosion and deposition.

Schumm and Etheridge (1994) and Thorne (1994) considered that several factors promote fluvial incision. These include: (i) relative base level fall (and commonly an increase in stream gradient), which can be caused by eustatic sea-level fall, tectonic uplift, or a combination of both; (ii) climate change resulting in increased discharge from the catchment, and; (iii) stream capture which results in an increased discharge from the combined river systems. Primarily the main control on this incision on continental shelves is relative base level fall associated with a major fall in eustatic sea level. However, in general, when sea-level falls, rivers incise to form incised valleys only if the fall results in subaerial exposure of surfaces that are steeper than the longitudinal gradient of those river systems (Posamentier, 2000; Posamentier *et al.*, 1992; Shanley and McCabe, 1994; Leeder and Stewart, 1996).

Zaitlin *et al.* (1994) identified a number of criteria for the recognition of incised valley systems as an initial step to defining generalised facies models for their identification in the geological record:

(i) The valley is an erosional palaeotopographic feature, the base of which truncates underlying strata;

(ii) The base and walls of the incised-valley system represent a sequence boundary (i.e. hiatus) which is correlative to a subaerial-exposure surface marked by soils or root horizons on the interfluves bounding the valley (Van Wagoner *et al.*, 1990);

(iii) The base of the incised-valley fill exhibits an erosional juxtaposition of more proximal (landward) deposits over more distal (seaward) facies;

(iv) Sedimentary features within the deposits of the incised-valley fill will onlap the valley sides.

They also stated that it is critical when identifying the extent of these features to examine the geometry of the sequence boundary within and outside of the incised-valleys. This is because the palaeotopography of the network of the features may allow determination of the former drainage direction which is a significant aid in palaeogeographic reconstruction.

According to Walker (1992), up to 5 - 15 m (typical depth to wave base) of sediment can be removed by marine erosion during sea-level rise, and thus only the deepest portions of incised channels are likely to be preserved in the sedimentary record. In North Carolina, Hine and Snyder (1985) discovered that on the shoreface and inner continental shelf, erosion associated with recent eustatic sea-level changes had stripped all of the Holocene, and much of the Pleistocene record, although an extensive network of incised channels retains a partial lithological record of former events.

During marine transgression over the continental shelf and coastal plain, topographically low features, such as incised valleys are often backfilled by sediments derived from and deposited in a range of coastal environments. The depositional environments that produce valley fills range from shore-parallel, wave dominated lagoonal and barrier island complexes to shore-normal fluvial and tidal channel environments flowing out from the coastline (Ashley and Sheridan, 1994). The associated coastal deposits typically include intertidal peats and muds, both of which can potentially contain valuable floral and faunal records for palaeoenvironmental interpretation. They may also contain interbedded sands and muds, and cross-bedded sands that provide an indication of palaeocurrent directions.

Ashley and Sheridan (1994) showed that the nature of these records is strongly influenced by the size of the incised valley and that the records from a relatively narrow (500-1000 m), shallow (5-10 m) drainage basin or tidal channel would be expected to be appreciably different from the fills originating in wide (several km), deep (~30 m) former river and estuarine valleys. Although the majority of incision of palaeovalley features occurs under fluvial action during lowered sea level, much of the infilling is associated with the time that sea level is rising, and thus most of the valley fills form part of the transgressive systems tract not the lowstand systems tract.

According to Hine and Snyder (1985) the lowermost incised-valley fill successions offshore become sheltered from erosion from subsequent regressions and transgressions. As a result, these portions of the valley fills may hold the fullest (if only in part) record of former eustatic change events on continental shelves, which would otherwise have normally been completely stripped from areas of interfluve.

2.4. Incised-Valley Fill Architecture

Incised-valley fill architecture has been widely analysed both on modern and ancient settings throughout the last decade, largely promoted by a growth of interest in the application and development of sequence stratigraphy (Lericolais *et al.*, 2001). An incised valley system is composed of an incised valley and its depositional fill sequence. These systems provide the most complete evidence of erosion by fluvial action during relative sea-level fall and lowstand, including what is sometimes the only record of lowstand and transgressive deposition in shelf slope and /or shallow ramp, marine depositional settings (Allen and Posamentier, 1994; Dalrymple *et al.*, 1994; Thomas and Anderson, 1994; Zaitlin *et al.*, 1994). Incised-valleys are known to occur in strata dating from the Precambrian (e.g. Dyson and von der Borch, 1994: Levy and Christie-Blick, 1994) through to the Quaternary and, indeed, right up to modern sedimentary environments. The deposits infilling incised-valleys are a volumetrically minor but scientifically and economically important component of the sedimentary stratigraphic record (Suter, 1986). The recognition of these deposits is an important criterion for the identification of sequence boundaries and, as a consequence, they are frequently recorded in older strata in association with hydrocarbon reserves (Dalrymple *et al.*, 1994; Kindinger *et al.*, 1994).

Research on more recent (Quaternary) incised-valley deposits has been undertaken for the purpose of locating offshore sand and gravel resources for aggregate production (Bellamy, 1994). At the same time the potential of future sea-level rise as a result of global warming has promoted interest in the need to understand the trangressive histories of modern (drowned former valley) estuaries, which presently serve as harbours, fisheries and recreation areas, not to mention the proximal fringing land on which a high proportion of the worlds' population lives (Zaitlin *et al.*, 1994).

The deposits infilling incised-valleys range from terrestrial through estuarine to open marine in their origins (Kindinger *et al.*, 1994). Many of the studies on estuaries and embayments of the Gulf of Mexico (Texas and Louisiana coastal plain and continental shelf), and on the shelf of the US eastern seaboard, typically originated as incised fluvial valleys that formed during the eustatic sea-level lowstand of the last (Weichselian) glaciation and were inundated by the ensuing postglacial eustatic sea-level rise. These features are infilled with sediments derived from both fluvial and marine sources (Suter *et al.* 1987; Morton and Armstrong Price, 1987; Foyle and Oertel, 1992, 1997; Anderson *et al.*, 1996). Likewise Allen and Posamentier (1993) noted that in the stratigraphy of the incised valley offshore of the Gironde Estuary, France, distinct phases of filling could be identified and attributed to fluvially, estuarine and marine derived sediments.

2.5 Sequence Stratigraphic Approach to the Study of Incised-Valleys

The origins of the now standard model of sequence stratigraphy were originally evolved by Sloss (1963) from regional mapping of sequences in North America and later developed by Payton (1977) working on seismic reflection profiling of passive continental margins. Sequence stratigraphy can be defined as the study of the relationships of rocks within a chronostratigraphic framework of repetitive, genetically related strata bounded by surfaces of erosion, non deposition or their correlative conformities (Van Wagoner *et al.*, 1988). The *depositional sequence* comprises an unconformity-bounded package. It is the fundamental unit of sequence stratigraphy. The internal geometries of these unconformable *depositional sequences* are influenced largely by fluctuating sea level (Hesselbo and Parkinson, 1996).

In essence, sequence stratigraphy is practised through the recognition of key surfaces in the geological record that define the boundaries of packages of genetically related strata (*systems tracts*), and by the recognition of facies deposited during trends of transgression and regression (Hesselbo and Parkinson, 1996).

2.5.1 Systems Tracts

Systems tracts are genetically associated stratigraphic units that accumulate under specific phases of the relative sea level cycle (Posamentier *et al.*, 1994, 1999). Each systems tract represents the collective sedimentary systems deposited during the different phases of sea level change. Such units are recognised in the geological record as three-dimensional facies assemblages, defined on the basis of bounding surfaces, their positions within a sequence and patterns of stacking of parasequence sets (USC Sequence Stratigraphy Web, http://strata.geol.sc.edu/index.html, 2002). In vertical succession, ideally all *depositional sequences* are composed of the following elements in order: sequence boundary, lowstand systems tract (LST), transgressive surface, transgressive systems tract (TST), maximum flooding surface, highstand systems tract (HST), and the following sequence boundary (Figure 2.4).

Lowstand Systems Tract (LST)

These lowermost systems tracts in the depositional sequence are equated to periods of fall in relative sea-level. They are divisible into three separate depositional units: basin-floor fan, slope fan, and lowstand wedge. Lowstand System Tract sediments often fill incised valleys that have cut down into the Highstand System Tract below creating the *sequence boundary* (USC Sequence Stratigraphy Web, 2002). Posamentier and Allen (1999) subdivided lowstand systems tracts into:

Early Phase Lowstand Systems Tract: This forms during the falling stage of relative sea level and includes all regressional deposits accumulating following the onset of relative sea-level fall. The lower bounding lies directly on the Highstand Systems Tract surface and is normally a Type 1 Sequence Boundary (described below). They are associated with:

- fluvial incision
- focussing of sedimentation at the shoreline
- forced regression caused by lowering sea level
- slope instability caused by the rapid deposition of fluvial sediments
- basin floor fan deposition

Late Phase Lowstand Systems Tract: This forms when a slow relative sea level rise occurs. This type of lowstand systems tract lie directly on the upper surface of the prograding clinoforms of the Early Lowstand and is capped by the transgressive surface formed when the sediments onlap onto the shelf margin. They are associated with:

- sedimentation outpacing the loss of accommodation space
- stabilisation of river profiles
- backfilling of valleys

Transgressive Systems Tract (TST)

Transgressive Systems Tracts generally follow on from the Lowstand Systems Tract in stratigraphy. They form when rapid relative sea level rise occurs above the shelf margin. Eustasy begins rapidly, exceeding the effects of any tectonic or isostatic uplift. The transgressing sea reworks sediments deposited by previous systems tracts to form a *ravinement surface*. The sedimentary sequence of these tracts comprises deposits that accumulated from the onset of coastal transgression until the time of maximum transgression of the coast, just prior to renewed regression. The lower bounding surface of a TST is the *transgressive surface*, initiated by the first significant flooding surface across the shelf. The upper bounding surface is the *maximum flooding surface* (see Figure 2-4).

Highstand Systems Tract (HST)

The highstand systems tract is the uppermost of a depositional sequence. Deposits form this tract when sea-level rise is out-paced by the rate of coast transgressional sediment accumulation. They are associated with:

- Slow rise of relative sea level followed by a slow fall; essentially a still stand of sea level when the slower rate eustatic change balances that of tectonic motion
- Sedimentation outpaces loss of accommodation space
- River profiles stabilize
- River valleys are dispersed laterally in a position landward of the shelf margin.

In siliciclastic shelf systems, estuaries have either been filled with sediment by the beginning of the highstand systems tract or are finally filled in the earliest phases of the highstand systems tract. Once sedimentation is no longer trapped infilling estuaries, rivers are free to build seaward and form deltas. In the portions of coastlines that lie between deltas, sandy wave-dominated shoreline deposits may form.

The basal identifier of an HST sequence is the *maximum flooding surface* (MFS) of the sea-level cycle, over which the high stand systems tract sediments prograde and aggrade. The upper unconformable bounding surface of an HST is created by subaerial exposure as sea level falls following the highstand (Type 1 or Type 2 Sequence Boundary)

Type 1 Sequence Boundary

Type 1 Sequence Boundaries are represented in the stratigraphy by an unconformity. Typically they are characterized by subaerial exposure of the shelf associated with stream rejuvenation and fluvial incision, sedimentary bypass of the shelf, and abrupt basinward shift of facies and coastal onlap. It is interpreted to form when the rate of eustatic fall exceeds the rate of basin subsidence at the depositional shoreline break, producing a relative fall in sea level at that position.

Type 2 Sequence Boundary

Type 2 Sequence Boundaries are also unconformable surfaces. They are marked by subaerial exposure and a downward shift in coastal onlap landward of the depositional-shoreline break; however, it lacks subaerial erosion associated with stream rejuvenation and a basinward shift in facies. These boundaries form when the rate of eustatic fall is less than the rate of basin subsidence at the depositional-shoreline break, so that no relative fall in sea level occurs at this shoreline position

2.6 Late Quaternary Incised-Valley Sequence Stratigraphy

Sequence-stratigraphic analysis helps to develop a chronostratigraphic and architectural framework of incised-valley fill (Thomas and Anderson, 1994). It has been argued that the lessons learned from the analyses of Quaternary sequence stratigraphy are not applicable to older strata because of the increased frequency of eustatic fluctuation during the past 2 Ma (Anderson *et al.*, 1996). However, according to Posamentier and Weimer (1993), sequence stratigraphic concepts seem to be scale-independent, both temporally and spatially. Initially depositional sequences were identified from seismic reflection data at the third order scale

(0.5-3.0 Ma). With the development of higher resolution seismic data sets, sequences have since been identified at fourth (0.08-0.5 Ma), fifth (0.03-0.08 Ma) and sixth (0.01-0.03 Ma) order scales (Posamentier and Weimer, 1993). Kindinger *et al.* (1994) applied the principles of sequence stratigraphy to delineate high frequency, sixth order (Posamentier and Weimer, 1993) depositional sequences while studying the Late Quaternary stratigraphy of the Mobile River incised-valley system on the Mississippi-Alabama shelf. Similarly Fyfe *et al.* (1997) applied sequence stratigraphical concepts to the Quaternary succession offshore of Hong Kong. They concluded that their observations of seismic and sediment architecture in the area indicate a protracted sequence of progressive marine inundation following the lowstand of MOIS6 (Figures 2-2 and 2-3).

Chapter 3: The English Channel

3.1 Introduction

This Chapter presents an introduction to the English Channel, in particular the eastern English Channel were the PhD study area is located. It begins with an introduction to the geological structure of the English Channel, including the Cretaceous and Tertiary strata that occur in the eastern Channel and underlie the PhD study area. This is followed by an outline of the climate and sea-level change history of the English Channel during the Quaternary period. Finally, the Chapter reviews previous research relating to the morphology and depositional sequences recorded in the palaeovalley system of the English Channel.

3.2 The Pre-Quaternary Geology of the Channel

The pre-Quaternary geology of the English Channel is illustrated in Figure 3.1, based on the British Geological Survey (BGS) offshore regional mapping programme (Hamblin *et al.*, 1992). This, together with an event chronology provided in Table 3.1, provide the template for the remainder of this Chapter.

3.2.1 The Geological and Structural Evolution of the English Channel

Smith and Curry (1975) and Smith (1984) briefly summarised the geological and structural evolution of the English Channel. They divided it into three distinct provinces (Western, Central and Eastern) and suggested that each province could be characterised by its own geological style and evolution (Figure 3-2).

The *Western Province* is characterised by WSW aligned near-surface structures. It consists of a series of basins, which widen and deepen in a westerly direction. It is bordered to the north by the southern coast of England and to the south by the northern coast of Brittany. The natural western extent of the province is the shelf break where it opens to the Atlantic Ocean. Here it is bordered by pre-Palaeozoic and Palaeozoic units of the Amorican and Cornubian Massifs (Smith, 1984). The eastern limit of the province occurs about the position of a line drawn from Start Bay, Devon, to the Contentin Peninsula, Normandy. The Start - Contentin line is a gentle ridge along a marked narrowing of the Channel. The Western Province has Precambrian to Pliocene lithologies exposed on the sea floor with numerous unconformities and a major fault zone extending east-northeast from Ouessant to north of Alderney.

The *Central Province* is underlain by rocks that range in age from Permian to Eocene that outcrop at the sea floor. Within this province are three major structural features, two possibly linked to the Ouessant – Alderney fault zone and the third forming the Isle of Wight-Purbeck monocline. The southeasterly alignment adopted by the monocline forms the Bembridge-St Valery-en-Caux line, dividing the Central and Eastern Provinces.

The *Eastern Province* is dominated by a relatively undisturbed open syncline (Hampshire-Dieppe Basin) of Tertiary strata flanked to the north-east by the continuation of the Weald–Artois anticlinorum of south-eastern England across into the Boulonnais area of northern France. The natural morphological limit of the eastern province is the marked shallowing and narrowing of the Strait of Dover (Pas-de-Calais).

The Hampshire-Dieppe Basin dominates much of the seabed geology of the eastern English Channel (Figure 3-5). It is a broad, gently dipping structure trending in a WNW-ESE direction and composed of mainly Lower Cretaceous (Lower Greensand, Gault and Upper

Greensand), Upper Cretaceous (Chalk) and Tertiary strata. The boundary between the youngest Tertiary sediments of the Oligocene and those of Quaternary age represents a major hiatus in the geological record in the English Channel, with Miocene and Pliocene deposits largely unknown (Hamblin *et al.*, 1992). Quaternary deposits have only been preserved in the form of palaeovalley infills and seabed lag sediments (discussed later).

3.2.2 Cretaceous Strata

The earliest strata of Cretaceous age in the eastern Channel are the uppermost part of the Purbeck beds (Figure 3-3). The characteristic defining these strata as Cretaceous is that they are of marine origin, laid down following the transgression that marks the Jurassic-Cretaceous boundary. Deposited during the Ryazanian stage they are termed the Durlston Beds and most commonly consist of shelly limestone interspersed with layers of shaley mudstone and clay containing bivalves, gastropods and ostracods (Hamblin *et al.*, 1992). After the brief transgression responsible for the formation of the Purbeck beds it is considered that a return to non-marine conditions prevailed and that the overlying Wealden sediments were deposited in predominantly freshwater conditions in a large lake that may have occupied the majority of the present Hampshire Basin and Weald areas (Allen, 1976, 1981; Lake and Shepard-Thorn, 1987) during the Valanginian, Hauterivian and Barremian stages. The Wealden Beds consist of two major units, the lower is made up of mostly sandy deposits (the Hastings Beds) and the upper more fine-grained, muddy formation (the Weald Clay). Above the Wealden Beds are the (Aptian and early Albian) Lower Greensand Strata. They are lithologically variable muddy and sandy sediments that were laid down in a variety of tidally influenced, shallow-marine and shoreline environments (Hamblin *et al.*, 1992). Despite their name, they are neither green nor are they particularly sandy, obtaining their name from the Upper Greensand with which they have been mistaken. In the Weald, the Lower Greensand is locally divided into four major lithological formations: Atherfield Clay, Hythe Beds, Sandgate Beds and Folkestone Beds. The divisions have been extended into the eastern Channel where it is considered that the differing lithology between beds has been responsible for variations in the morphology of the seafloor that arises from differential erosion of the soft and hard units, a crucial factor in the understanding of the widths, regular or irregular nature of thalwegs, and the position of palaeochannels in the area.

Following the major transgression of the Early Cretaceous, in the early Albian deeper-water conditions resulted in the basinal areas and some 40 metres of fossiliferous Gault Clay (mostly dark, bluish grey to pale grey, soft mudstone and silty mudstones) was deposited in the eastern Channel (Hamblin *et al.*, 1992). The Gault is considered to be a single continuous unit across the Channel with direct correlations possible between units at Folkestone and those at Wissant on the northern French coast (Owen, 1971). In the eastern Channel the Upper Greensand varies in both thickness and lithology. It comprises siltstones, silts, hard siliceous sandstones, and calcareous sandstones that occasionally grade into siliceous limestones. Some beds are highly fossiliferous, and contain the remains of ammonites, bivalves and sponges. The variations between the lithological nature of the upper greensand beds can have significant bearing on the seabed morphology and position of palaeochannels (D'Olier *pers. comm.*).

At the start of the Late Cretaceous, the Cenomanian transgression (caused by eustatic sea-level rise) combined with regional tectonic subsidence to cause the initiation and spread of the "Chalk Sea" which covered much of Europe. Following the initial deposition of the Glauconite Marl (basal chalk), terrigenous sedimentation was limited and Chalk deposition resulted. The Chalk is a friable biomicritic limestone consisting of a matrix of debris from planktonic algae (coccoliths and rabdoliths), with coarser components from foraminifera, calcispheres, and fragments of larger faunal remains (e.g. bivalves and echinoderms). Whittaker (1985) suggested that Late Cretaceous sedimentation rates were relatively high and that in basinal areas over 500 m of chalk may have been deposited in less than 30 million years.

Flint occurs commonly in the Upper Cretaceous chalk strata, as nodules or beds, following as well as cutting across bedding planes. The siliceous nodules of flint commonly occur infilling former crustacean burrows, the biogenically derived silica having been remineralised into them. Flint occurs mainly in the upper part of the chalk succession with only rare occurrences in the Lower Chalk.

The Lower Chalk is of Cenomanian age and has a relatively high terrigenous content. It is readily identified and mapped where it outcrops on the seabed as it forms an escarpment where the underlying softer sequences of the Lower Greensand and the Gault have been eroded away. Boreholes extracted for research prior to the beginning of the Channel Tunnel construction sampled extensive thicknesses of Lower Chalk. At the base of them is the Chloritic Marl, made up of 1 to 5 m of dark greenish grey sandy marl with glauconite. It is succeeded by Chalk Marl, 22 to 27 m of dark grey chalky marl. The 22 to 26 m thick Grey Chalk above comprises less marly pale to medium-grey chalk. Over this is the White Chalk (or White Bed) described by Destombes and Shepard-Thorn (1971) as 18 to 20 m of homogeneous, very pale, yellowish grey, marly chalk. Capping the Lower Chalk is the Plenus Marl, a thin (0.05 to 8.5 m) but widespread greenish grey marl band thought to represent a brief interval of clay deposition that interrupted generally continuous clear-water chalk deposition in the Chalk Sea. The Plenus Marl represents the end of the Cenomanian (Hamblin *et al.*, 1992).

The Middle Chalk is characterised by up to 90 m of massively bedded white chalk deposited in the early to mid-Turionian. Apart from the uppermost few metres it is generally lacking in flint. Hamblin *et al.* (1992) considered that the Middle Chalk is always characterised by a basal bed named the Melbourn Rock, a complex of hard, white nodular chalk and incipient hard-grounds.

The strata identified as the Upper Chalk ranges in age of deposition from late Turonian to Maastrichtian and is the thickest of the three Upper Cretaceous chalk units (260 m – 404 m). Flint occurs extensively throughout the unit occurring as concentrated tabular seams, with a belt of maximum development occurring in the top of the Turonian Chalk (Hamblin *et al.*, 1992). The chalk itself is white and massively bedded. Where it has been found offshore it has not been distinguished from the other chalk units.

3.2.3 Tertiary Strata

In the eastern English Channel, Dangeard (1923) first identified Tertiary sediments by an abundance of the middle Eocene foraminiferal species *Nummulites laevigatus* in samples dredged from the seabed. Estimates of the thickness of Tertiary strata in the Hampshire-Dieppe Basin in the eastern Channel suggest a sequence of up to and over 500 m. Tertiary sediments occurring in the area of interest relating to this study include Late Paleocene (Thanetian stage) and Eocene (Ypresian, Lutetian and Bartonian stages) strata (Figure 3-4).

The earliest Tertiary sediments recognised in the eastern Channel region are Late Thanetian (Paleocene) crystalline limestones containing *Microcodium elegans*. These sediments rest unconformably on the Turonian and Maastrichtian stage chalks of the Upper Cretaceous. The Woolwich Beds (marine) and the Reading (non-marine) Beds (that contain very distinctive black pebbles) complete the Late Thanetian sequence in the region. Sediments in the sequence range in their lithologies from shell beds, fossiliferous, marly clays and lignites, through to large-scale cross-bedded sands.

A major marine transgression initiated sedimentation in the Early Eocene (Ypresian). The trangression, which, it is hypothesised entered the region from the area of the present North Sea, was responsible for the deposition of the London Clay formation. The Wittering Formation overlies the London Clay to complete the Ypresian sediment sequence. Little is known about the Wittering Formation offshore, but where its onshore correlative sequence has been sampled at Bracklesham Bay (Curry *et al.*, 1977) and Sandhills (Edwards and Freshny, 1987) it ranges from 63 to 83 m thick. Plint (1988) interpreted the silty sands and clayey silts of the formation at Bracklesham Bay as denoting two transgressive sequences each with a basal pebble bed. At the top of the formation is a hiatus characterised by a complex series of channelled, fine-grained sands and lignitic silts with pebble beds, indicative of regression believed to correspond to a brief eustatic sea-level lowstand at the end of the Ypresian.

Curry and Smith (1975) noted that the Lutetian sediments offshore showed a series of strong seismic reflectors that appeared to correlate with hard calcareous beds in the middle of the sequence. The Lutetian sequence, the Earnley and Selsey Formations (part of the Bracklesham Group, Curry *et al.*, 1978) have been sampled extensively from offshore in the Channel, with recovered borehole sediments dominated in the most part by glauconitic sandy clay and sandy limestone lithologies, whereas in the middle horizons the sequence includes calcareous levels rich in nummulites especially the taxa *Nummulites laevigatus.* The upper sediments of the Selsey Formation in southern England contain abundant marine microfossils consistent with the suggestion that during Lutetian times a narrow marine embayment with a fluctuating coastline existed in the region (Hamblin *et al.*, 1992). Pollen and other floral proxies indicate the persistence of a warm climate for the region during this stage also.

The Bartonian (or Huntingbridge) Formation in the eastern Channel overlies the Selsey Formation conformably (Hamblin *et al.*, 1992). Short core sediment samples of the sequence recovered from offshore include sandy clays containing *Nummulites rectus*, indicating a correlation with the lower Barton Beds of southern England. The BGS have also retrieved boreholes in the region, one of which (75/38, Figure 3-5) penetrated some 56 m of sandy silts and silty sands representing the Bartonian. This was considered to be the equivalent to the onshore representation of the Barton Clay Formation (Edwards and Freshney, 1987), which includes sediments of Lutetian as well as Bartonian age.

Priabonian (Upper Eocene), Oligocene and Neogene strata have to date not been recorded in the eastern English Channel. Suggestions were made by Robert (1971), however, as to their possible preservation in the Hampshire-Dieppe Basin, based on the preserved thickness of the Tertiary sequence rather than information from field sampling.

3.3 Late Quaternary Climate and Sea-Level Changes in the English Channel

Glacio-eustatic variations in global sea-level during the Late Quaternary (Figure 2-1) have strongly influenced the timing and development of insularity of the British Isles. During this time sea levels have risen and fallen according to the ice sheet growth and recession associated with the major glaciations, which are forced by Earth orbital variations driving major changes in global climate. The associated variations in sea-level were such that during glacial maxima the entire English Channel became dry land, effectively extending the coastline of continental Europe to the present shelf break west of Cornwall (120 – 250 km from the nearest present coastline).

The major cold glacial maximum periods of the Late Quaternary were MOIS12, 10, 8, 6 and 2 (Figure 2-1). During these lengthy cold stages, global eustatic sea-levels dropped and the shoreline retreated back across the shelf to lowstand shorelines that occurred up to 120 metres below present. During these periods much of the current sea floor of the Channel was exposed and subjected to the processes of periglaciation associated with sub-arctic climates. During the more short-lived temperate interglacial periods of the Late Quaternary (MOIS11, 9, 7, 5e and the Holocene; Figure 2-1), global eustatic sea levels rose to highstand levels similar to present and transgressed the continental shelf of the Channel and isolating Britain from the rest of the European mainland. Because the Late Quaternary highstands are short lived compared to the lowstand conditions, a peninsular Britain connected to continental Europe has been the norm rather than the exception (Bridgland and D'Olier, 1995). Table 3-2 provides a brief summary of the Mid to Late Quaternary history of the English Channel.

3.4 Morphology and Depositional Features of the English Channel

The overall morphology and depositional features of the English Channel can be divided into six elements: (i) an extensive, gently dipping marine planation surface; (ii) submerged cliff lines; (iii) marine terraces / raised beaches; (iv) tidal sand banks and ridges; (v) sea bed sediments, and; (vi) incised channels, fosses and deeps (including the Strait of Dover).

3.4.1 Marine Planation Surface

The marine planation surface makes up the majority of the sea-floor of the English Channel, which is in general incised to 50 m lower than the surrounding English and French coasts (Curray, 1989; Reynaud *et al.*, 2003). Lericolais *et al.* (2003) stated that the sea-floor of the channel could be considered as a peneplain which begins at a depositional basin at the Western Approaches and continues in an upstream direction into the drainage basins of the rivers Somme and Seine, through the Strait of Dover and well into the southern North Sea.

The sea-floor slopes away from the coastline, and dips gently to the southwest at an average gradient of between 1:1500 (0.66 m / km, Curry, 1989; Hamblin *et al.*, 1992) and 1:5000 (0.2 m / km, Lericolais *et al.*, 2003; Antoine *et al.*, 2003). Hamblin *et al.* (1992) suggested that the geology of the planation surface is almost entirely of Neogene age, because for the majority of its area it cuts across older Palaeogene and Cretaceous sedimentary strata. Curry (1989) stated that for the most part the seabed of the English Channel is an area of erosion and not permanent deposition. According to his research the rock floor has undergone continual erosion, with the materials derived from this erosion being removed from the region by tidal scour. Stride (1990) critically examined this work and concluded that potentially half of the rock floor of the Channel is subject to erosion. However, across the remaining half, sheets of sands and gravels have been (and continue to be) deposited. These, Stride (1990) suggests, will thicken with time and preserve at least a partial record of the Channel's Holocene evolution. It is therefore probable that sedimentary records of earlier highstands will also have been preserved subaqueously in the region.

3.4.2 Submerged Cliff Lines

Three submerged cliff lines, which have been interpreted as former coastlines have been identified in the English Channel off the coasts of Devon and Cornwall (Cooper, 1948;

Wood, 1974; Kelland, 1975; Donovan and Stride, 1975). In this locality the three submarine features are located at depths of 38-49 m, 49-58 m and 58-69 m below ordnance datum (OD) (Hamblin *et al.*, 1992). Kelland (1975) identified the shallowest of the three as a relict buried coastline at 42.5 m below chart datum (CD) in Start Bay, Devon. He related his findings to those of Cooper (1948) and Clarke (1972) who identified similar features at corresponding depths near Plymouth and Torbay respectively. Kelland (1975) initially interpreted the feature as pertaining to an ancient coastline exposed during one or more periods of lowered sea level in the Pleistocene, however Wood (1974), and Donovan and Stride (1975) considered that Quaternary lowered eustatic sea-level still stands would have been too short lived to allow significant cliff lines to develop. Hamblin *et al.* (1992) suggested a Neogene (Miocene – early Pliocene) age for the features.

The British Geological Survey proved the existence of a submerged cliff line running from Shingle Bank to Beachy Head in the eastern Channel at depths of over 20 m (Figure 3-7). The feature becomes undetectable between 0° and 1°W, probably as a result of erosion by the Northern Palaeovalley. Hamblin *et al.* (1992) intimated that it reappears to the west as the 38-49 m cliff, the disparity in depth being accounted for by differential crustal motions in the region, with the western Channel subsiding relative to the eastern (Figure 3-1).

3.4.3 Marine Terraces / Raised Beaches

Pleistocene deposits on the English coast of the Channel, in particular the coastal plain of West Sussex, have been the subjects of interest by Quaternary scientists since the mid nineteenth century (Prestwich, 1859; 1892; Reid, 1892; 1903; White, 1913; 1924). Numerous publications on the subject and geological mapping by the British Geological Survey have revealed a complex depositional history that has only begun to be unravelled in recent years. A significant part of the history is the revelation that a flight of marine terraces has been recognised and the suggestion made that they are representative of successive sea-level high stands (Bates *et al.*, 1997 and 1998). Bates *et al.* (1997) suggested that there is evidence of five discreet high stands preserved in the west Sussex upper and lower coastal plains, identified in the form of raised marine terraces (or beaches) in an altitudinal range of between present sea-level and +40 m OD. In altitudinal order (highest to lowest) these raised beaches are called the Goodwood-Slindon (32 - 43 m OD) (Shepard-Thorn *et al.*, 1982), Aldingbourne (17.5 – 27.5 m OD) (Fowler, 1932; Calkin, 1934), Cams Down (15 – 18 m OD) (Palmer and Cooke, 1923; ApSimon *et al.*, 1977), Brighton – Norton (5 – 12 m OD) and Pagham (2 – 3 m OD). It is generally accepted the highest of the terraces is the oldest and that the deposits become progressively younger with decreasing altitude (Keen, 1995) (See Table 3.3). The interpretation that the raised beach deposits of the west Sussex Coastal Plain represent sea-level highstands is not conclusive. The inferred environments of deposition (Table 3.3) of the sediments associated with these features are mostly sub-intertidal estuarine/marine. Therefore, they represent limiting sea-level data i.e. at the time of deposition relative sea-level must have been at or above the altitude of the deposit.

In association with the raised beach deposits a number of large erratic boulders (Sarsen Stones) up to eight metres across, and various exotic lithologies (granite, syenite, "greenstone", diorite, quartz porphyry, biotite gneiss and mica schist) have been recognised (Goodwin-Austen, 1857; Lyell, 1871; Reid, 1892; White, 1915; Hodgson, 1964; Young and Lake, 1988). Principally derived from areas of northern France (Brittany, the Cotentin Peninsula) and the Channel Islands, these erratics were suggested to have been transported and deposited by an ice sheet (Kellaway *et al.*, 1975) a theory that would require the glaciation of northern France and the English Channel. An alternative, and more likely, explanation for their occurrence is that they originated by transportation by floating sea ice (Reid, 1892; Kidson and Bowen, 1976) and, although the formation of sea ice in the Channel is rare at present it is commonplace in marginal seas at similar latitudes in other parts of the world.

It is unlikely that during the Middle and Late Pleistocene eustatic sea-levels have attained altitudes much higher (*circa* 10 m) than at present (Siddall *et al.*, 2003; Shackleton, 1987). Thus the relative altitudinal positions of the raised beach deposits and erratics described above require an alternative explanation for their existence. The most widely cited hypothesis that southern England has experienced long-term crustal uplift. In their study of the Pleistocene sea-level and neotectonic history of the eastern Solent, Preece *et al.* (1990) attempted to quantify the rate of uplift of the area based on the inferred age of the deposits and their relative altitudes. Their best estimate of apparent uplift was some 40 m during the last 400,000 years (0.1 m per 1000 years). Antoine *et al.* (2000) describe a comparable situation in northwestern France suggesting an uplift rate of 55 – 60 m per Ma since the end of the Lower Pleistocene deduced from the analysis of the Seine and Somme rivers terrace systems.

Many suggestions have been proposed for the nature and origins of the apparent uplift. Preece *et al.* (1990) stated that it was entirely consistent with the broad uplift implied by Smith (1985a, 1985b, 1989) to have occurred during the past few million years affecting southern England and the neighbouring continent as a possible remnant of the Variscan orogeny. Likewise Long and Tooley (1995) hypothesised that a source for differential crustal movements on the southern coast of England could be the continued movements of the Variscan Front Thrust. The front can be traced through southeastern England into northwestern France and is a considered to be a remnant of the Variscan Orogeny during the Late Palaeozoic. The front is thought however, to be still active, evidence for which has been presented in the form of reconstructed historical data of the spatial extent of the earthquake in southern England and northern France on April 6th 1580 (measuring 6.2-6.8 on the Richter scale). It has been suggested that the earthquake epicentre was below the Strait of Dover, with movement probably associated with a fault related to the Variscan Front Thrust.

Preece *et al.* (1990) proposed another potential driving mechanism, noting that the well documented rapid subsidence of the southern North Sea (through loading by thick deposits of Quaternary sediment) may be compensated by associated uplift in fringing areas such as the central English Channel to the south. Bridgland (1994) promoted a means for uplift through 'erosional isostasy' in the case of the River Thames. The model sees net erosional unloading by fluvial incision on land, and net loading, by sedimentation, in sedimentary basins offshore. This is then followed by subsequent crustal realignment. Maddy (1997) considered however, that erosional isostasy was no more than a positive feedback response to uplift initiated in some other way.

Maddy *et al.* (2000) used evidence from river terraces records in southern England to demonstrate that river valley incision during the Quaternary has averaged 0.07-0.10 m per 1000 years. These authors argue that this incision is primarily caused by regional uplift. Antoine *et al.* (2000) provided supporting evidence for this from northwest France. According to their data gathered from the valleys of the Rivers Seine and Somme, a slow rate of tectonic uplift (0.05-0.06 mm per year) is required to create the incision and subsequent terrace formations that they observe in this area. Maddy *et al.* (2000) summarised the mechanisms that have been suggested as plausible explanations for uplift:

- Erosional isostasy: differential crustal movement occurring between areas of net erosion and sediment accumulation.
- Glacio-isostasy: isostatic response to loading and unloading of ice-sheets during the Quaternary.
- Hydro-isostasy: loading and unloading of the continental shelves by changing sea-level in a similar way to glacio-isostasy.
- Intraplate tectonic stress: as a result of major reorganisations of the spreading direction and rate that occurred during the Pliocene along the Atlantic spreading system.

Maddy *et al.* (2000) suggested that it is as yet impossible to attribute precisely the causes of uplift, but it is more than likely that several of the above mechanisms are involved and operate over different timescales.

3.4.4 Sea Bed Sediments, Tidal Sand Banks and Ridges

Reynaud *et al.* (2003) conducted a review of the present-day sea-floor surficial sediments of the Channel and its Western Approaches, focusing in particular on the sandy accumulations (tidal sand bank systems). Synthesising the data available of sedimentary cover of the Channel they divided it into three major sectors based on sediment nature and distribution on the sea-floor and large-scale depositional architecture of the sediments. The three sectors are the Eastern Channel, the Western Channel, and the Western Approaches. The Eastern and Western Channel sectors were separated on the basis of the dominant current direction and the zone of bed-load parting. The Eastern Channel is characterised by dominant flood tidal currents, whereas the Western Channel is ebb-dominated. Between the two sectors in the central is a bed-load parting zone north of the Cotentin Peninsula (Johnson *et al.*, 1982). The Western Channel sector is characterised by a paucity of sea-bed surface sediment (Reynaud *et al.*, 2003) which, when present, comprises a thin veneer less than a few decimetres thick. In contrast, sediment cover in the Western Approaches sector reaches several tens of metres in places.

In the eastern sector, offshore of the northwestern French coast, a series of sand banks and ridges, up to 30-70 km in length, 3-5 km wide and 10-30 m high (locally 50 m), extend (in a ENE-WNW and NNE-SSW direction) along the French border of the median line of the Channel. They include the Varne and Colbart banks, Ridens, Vergoyer, Bassurelle, Bassure de Baas, Battur, Querner and the Bassurelle de la Somme. These features are largely composed of moderately - very well sorted medium grained sands, which are mostly mature (well rounded) quartz grains (Hamblin, 1989). Johnson *et al.* (1982) suggested that tidal sand banks and ridges are approximately aligned with the strongest tidal flow. The features identified in the eastern English Channel are aligned with the dominantly northeasterly tidal flow path running from the Central Channel bed-load parting to the bedload convergence at the Strait of Dover.

Hamblin (1989) noted that in the areas of the eastern Channel where sand banks and ridges are absent, the sea floor erosion surface has a discontinuous veneer cover of lag deposits interspersed with exposures of bedrock. Bedrock is generally exposed in areas where the lithology of the solid substratum is of high or variable hardness (Hamblin, 1989). Investigations into the nature of the lag deposits show they are typically gravels, sandy gravels and gravely sands. Their sand fraction contains high (40-100%) shell-derived carbonate content whereas the gravel fraction has a conversely low content of biogenic carbonate (0-40%). The lithological content of the gravels is dominated by flints of two types: fresh, blackhearted, that was derived locally and brown, worn flint with a long history of rederivation via Tertiary deposits (Hamblin, 1989). Significant other contributions to the content of the gravel fraction are made up of chalk, limestone, sandstone, ironstone, mudstone and chert but notably only locally, and in areas where they make up the underlying solid geology. A very small proportion of the gravel is exotic to the area and, where present, is restricted to the granitic and quartz-tourmaline rocks of Cornubia and potentially Cotentin to the southwest (Hamblin, 1989) Some gravels also contain Devonian/Carboniferous Boulonnais material (D'Olier, *pers. comm.*). Examination of the larger gravel fraction has shown that pebbles and cobbles have been bored and encrusted with serpulids, bryazoa and barnacles, signifying that the lag deposits are not mobile under the transportational influence of the present current regime and have probably been *in situ* since the last transgression. This suggestion is supported by Stride (1990) who stated that these lag sediments appear to be in dynamic equilibrium with the existing peak tidal current strength and are most probably a relic of the Flandrian transgression.

Reynaud *et al.* (2003) considered that the general distribution of sediment in the English Channel demonstrated a system of two sediment prisms (a lowstand and a highstand prism) related to the glacio-eustatic cycle. These authors suggested that the larger concentrations of sediment in the Western Approaches and the eastern Channel compared with the western and central Channel can be explained by the longer persistence of coastlines in these areas, owing to their position relative to the maximum highstand and lowstand positions of the glacio-eustatic cycle (Lambeck, 1995).

3.4.5 Incised Channels, the Strait of Dover, Fosses and Deeps

The dominant feature of the gently inclined floor of the Channel is the widespread occurrence of a complex network of incised channels, fosses and deeps (Figures 3-6 and 3-7). The bathymetric map of the Channel (Figure 3-6) shows that there are three major palaeovalleys cut into the sea floor (Hamblin *et al.*, 1992): 1) the Axial 'Lobourg' Channel running through the Strait of Dover; 2) continuous from the Lobourg Channel, the

Northern Palaeovalley which runs in a generally northeast – southwesterly direction between the Strait of Dover and the Hurd Deep; 3) the Hurd Deep, lying 30 km north-west of the Cherbourg Peninsula, which may be a down-stream portion of the Northern Palaeovalley (Smith, 1985).

The Hurd Deep is an elongate depression, some 150 km long and between 2 and 5 km wide. Lericolais *et al.* (1996 and 2003) hypothesised that Neogene tectonics were responsible for the origin of this, the deepest (170 m) seabed feature in the English Channel. They note that the feature is located along the major fault zone of the Western Channel (in an ENE-WSW direction) and incises into deformed strata of the Jurassic and Upper Cretaceous (Hamilton and Smith, 1972; Evans, 1990). This hypothesis gained support from high-resolution geophysical data, which demonstrated a tectonic pre-control and origin of the feature. Further modification and infilling of the Hurd Deep was inferred to have been a product of subaerial and submarine processes during the numerous Quaternary regressive and transgressive phases (Lericolais *et al.*, 1996).

Dingwall (1975) suggested that the form and pattern of the 'incised channel and deep system' (Figures 3-6 and 3-7) is at least in part controlled by the stratigraphical and structural nature (i.e. folding and faulting) of the underlying strata of the Channel floor. Two models have been proposed to explain the remarkable incised channels and deeps that occur on the Channel floor, each of which are now reviewed.

The Glacial Model

Kellaway *et al.* (1975) argued that the system of channels is related to one or more of the Quaternary glaciations. They considered that during the Saalian (MOIS6), sea-levels were sufficiently low for ice to have been grounded near to the western margin of the continental shelf. In accordance with this hypothesis the morphological features of the Channel were described as remnants of a glaciated landscape. For example the relatively flat, low relief, floor of the channel was likened to a flat glaciated terrain. Central to the Kellaway *et al.* (1975) model was a reappraisal of the Slindon '100 ft' raised beach deposit on the Sussex coast. Rejecting its origins from a former sea level highstand, they reclassified this deposit as a remnant of a glacial lake (Lake Solent) dammed by English Channel ice. Similarly they also considered the Clay-with-flints found commonly in southern England to be the remnants of a decalcified till associated with the glaciation (Kellaway *et al.*, 1975).

In support of the theory, Destombes *et al.* (1975) also suggested that the Fosse Dangeard and associated landforms were sub-glacially excavated. They accepted Kellaway *et al.*'s (1975) proposed existence of major ice sheet glaciations in the Channel, and compared the formation of the deeps with glacial sub-drift features of East Anglia, Lancashire and Cheshire. The features they examined in the Channel were likened to 'tunnel valleys' formed at the abrupt termination of an ice sheet, eroded by debris-charged sub-glacial streams subject to intense hydrostatic pressure. Such features are characterised by overdeepened and irregular long profiles and narrow gorge-like cross profiles. Destombes *et al.* (1975) considered the likely mechanism for the sculpture of the NNE-SSW orientated deep buried valleys in the mid Channel to have been glacial scouring by ice moving in this direction across the Channel floor.

Since the theories of Kellaway *et al.* (1975) and Destombes *et al.* (1975) were invoked, the Clay-with-flints in southern England have been generally accepted as *in situ* formations that formed as a product of regolith weathering (Catt, 1986). Moreover, the landforms previously attributed to a glacial origin were considered by Kidson and Bowen (1976) to have been eroded in a marine environment. Surveys by the British Geological Survey and other workers (Oele and Schüttenhelm, 1979) in the English Channel and southern North Sea have failed to identify evidence for glaciations during the Quaternary extending further south than the traditionally accepted limits in the North Sea (Bowen *et al.*, 1986). Therefore little credence is now given to theories involving glaciation of the English Channel during the Quaternary.

Catastrophic Flooding

Smith (1985a, 1989) invoked an alternative theory to explain the seabed geomorphology of the English Channel, centred on the role of catastrophic meltwater discharge. Features pivotal to this theory were the Fosse Dangeard, and two of the other significant deeps, cut into the underlying geology in the Strait of Dover to a depth of up to 100m below the sea floor (Hamblin *et al*., 1992). Smith (1985a, 1989) argued that these features originated as plunge pools, a product of the outflow of an enormous proglacial lake that formed in the southern North Sea due to ice damming to the north by confluent Fennoscandian and British ice sheets. This lake purportedly overtopped the continuous chalk escarpment of the Wealden Anticline (between SE England and NW France) cutting the initial Strait of Dover. Beyond the Strait of Dover, the waters then spread out supposedly incising the palaoevalley network in the eastern and central channel (Figure 3-7) in a manner comparable to that of Bretz's (1923; 1959) proposed catastrophic outwash of Glacial Lake Missoula, which formed the channelled scablands of the Spokane area, Washington State, USA. Pleistocene fluviatile and marine sediments that infill an overflow river channel in the Calais region of France near Wissant were discussed by Roep *et al.* (1975), who suggested that this channel was contemporaneous to the catastrophic flooding, proposed by Smith (1985a, 1989).

Gibbard (1988, 1995) suggested that the excavation of the Strait of Dover occurred during the Anglian / Elsterian glaciation (MOIS12). He highlighted that there was indirect biostratigraphical evidence for some form of barrier between the southern North Sea and the English Channel (probably the Weald – Artois anticlinal chalk ridge) up until that time in the Pleistocene. According to Zagwijn (1979), during the early and mid Pleistocene the rivers of Belgium and Holland had drained northward through the North Sea, with little or no suggestion of southward flow through the Strait of Dover. Gibbard (1988, 1995), in agreement with Smith (1985a, 1989), proposed that at some time during the Anglian / Elsterian (MOIS12) stadial, coalescent Scandinavian and British ice sheets prevented the major northwest European rivers flowing north thus damming them in the southern North Sea. However, he suggested that the overspill of the lake at an outlet col in the Strait of

Dover might initially have been forceful if not catastrophic, but the later over spilling probably persisted for longer and was more moderate. Gibbard (1988, 1995) also hypothesised that the Rhine – Thames drainage system flowed via the Strait of Dover into the eastern Channel in all stadials since the initial breach. Bridgland and D'Olier (1995), while not disputing the timing of the initial breach, cited palaeontological evidence from Meijer and Preece (1995) to argue that Britain possibly remained joined to continental Europe in the late Anglian / Elsterian (MOIS12) and also during the early subsequent Holsteinian / Hoxnian (MOIS11) interglacial. Their theory disputed the chronological model proposed by Gibbard (1988, 1995) and argued that, for at least part of the time since the initial breach, a watershed persisted between the Weald and Artois ridges. White and Schreve (2000) also presented evidence for both peninsularity and insularity of Britain during the Holsteinian / Hoxnian (MOIS11). Their evidence, largely from Clacton-on-Sea in Essex, indicated that Britain was a peninsular to Europe for the most part of the succeeding interglacial (MOIS11) until it was severed in the later part of the Hoxnian (MOIS11) by marine transgression.

Gibbard (1995) proposed a polycyclic fluvio-marine origin for the form of the system of multiple incised channels, fosses and deeps (Figure 3-7), typical of the English Channel seabed morphology. Rather than being solely the consequence of the single catastrophic event, Gibbard (1995) suggested that their form had arisen through the persistent reoccupation of river channels, and thus down-cutting by fluvial action during low stand glacial phases, and by marine erosion (e.g. tidal scour) during periods of raised eustatic sea-level. This theory is supported by studies that have shown that many of the offshore valleys are offshore continuations or relict features of present day rivers flowing from mainland Europe and the UK (e.g. the Seine, Somme, Solent and Arun) (Dingwall, 1975; Gibbard, 1988, 1995; Bellamy, 1995; Velegrakis *et al.*, 1999).

It is now commonly accepted that the network of incised-valleys are relics of large river systems flowing across the seafloor during periods of lowered eustatic sea level. Thus the Channel was an enormous alluvial plain drained by the 'Channel River' (Fleuve Manche). Gibbard (1988, 1995) suggested that the majority of the fosses/deeps probably originated by erosion, causing channels to be over-deepened, particularly at confluence points. This method of formation was typified by Hamblin *et al.'s* (1992) discussion of St Catherine's Deep (Figure 3-7), to the south of the Isle of White, which is described as a bathymetric depression extending some 60m below the surrounding sea bed, with little or no sediment infill. The most widely accepted theory behind St Catherine's Deeps origin is tidal scour action during successive marine advances. It lies in an area of strong tidal currents, which would be expected to have been stronger when sea level was lower. Nevertheless, it may represent overdeepening of a pre-existing river valley, for its south-western end is aligned with a branch of an infilled fluvial palaeovalley. It is reasonable to expect that a river would have developed along the line of the deep, since it follows an outcrop of soft Lower Cretaceous strata immediately north of the chalk.

Testing the models described above has proved difficult due to a paucity of sedimentary evidence collected from offshore. Indeed, analyses of the stratigraphy of the infills of the palaeovalleys in the eastern English Channel have only recently been undertaken. Bellamy (1994, 1995) studied an infilled valley system to the east of Owers Bank in the eastern English Channel, thought to be the former course of the River Arun. Following detailed analyses of high-resolution shallow seismic and vibrocore sample data a series of multiple cut-and-fill deposits were identified. These deposits were interpreted as recording a history of erosion and sedimentation during the Quaternary associated with one or more cycles of relative sea-level rise and fall (Dingwall, 1975; Bellamy, 1995). Bellamy (1994, 1995) proposed a model of infilled valley evolution for the sedimentary sequences. He concluded that the sequence of events that created the cut-and-fill deposits have included the action of gravel-bed rivers in a periglacial landscape, fine-grained sedimentation in estuarine environments, and peat accumulation in temperate climatic conditions.

During the last glacial maximum (Late Devensian, MOIS2) the present form of the incised-valley network in the eastern English Channel developed. The rivers Rhine, Thames and Meuse all discharged through the Strait of Dover and Channel (Oele and Schüttenhelm, 1979; Gibbard, 1988, 1995). Similarly the rivers of southern England and north-western France flowed onto the subaerially exposed shelf cutting valleys that extend much further inland than the present coastline (Hamblin *et al.*, 1992). Offshore of south-eastern England between the Strait of Dover and Selsey Bill, a number of incised channels have been identified. Offshore extensions of the courses of the rivers Arun, Adur, Ouse and Owers (Bellamy, 1994, 1995) have been demonstrated, as well as an offshore palaeo-Solent River further west (Dyer, 1975). Offshore of northwestern France submarine extensions of the rivers Canche, Authie, Somme and Seine have also been postulated (Figure 3-7).

Chapter 4: Investigational Techniques

4.1 Introduction

This chapter describes the field programme and laboratory techniques that were employed to investigate the nature and origins of the seabed sediments studied in this thesis. The main techniques discussed in detail are continuous shallow seismic reflection profiling, seismostratigraphy, and lithostratigraphy. General descriptions of the other techniques used in the thesis (diatom, sediment grain-size and geochemical analysis) are provided, with the main methodologies for these techniques described in the Appendices. This chapter describes why each main technique was used, and reviews some of the possible sources of error and the limitations that can arise from their usage.

4.2 Data Acquisition

A large-scale field based reconnaissance survey was completed to investigate the nature and origin of the features of the seabed and sediments in the study area. This was undertaken by collecting continuous high-resolution shallow seismic reflection profile data and retrieving 189 vibrocore borehole samples with a maximum recovery of 6 metres (Le Tirant, 1979 provides a detailed introduction to the combination of the techniques). Following laboratory analyses of the retrieved material (see below), the acquired data were combined for interpretation (Chapter 6) to establish an improved depositional history of the study area during the Quaternary.

4.3 Continuous High-Resolution Shallow Seismic Reflection Profile Data and Seismic Stratigraphy

Shallow seismic reflection profiling is a technique commonly employed in the investigation of submarine geology and is used extensively in coastal and offshore waters (Evans *et al.*, 1995; Stoker *et al.*, 1997). The technique produces continuous profiles (sections) depicting the sediment and rock layers that are underlying the seabed to a depth ranging from less than five metres to over one thousand metres. Continuous, analogue shallow seismic reflection profiles are collected from a ship using a surface-towed acoustic source, hydrophone receiver and graphic recording system that translates the signal onto a continuous paper record (or profile) of the sub-sea bed strata as the vessel moves through the water (Figure 4-1). The equipment operates on the principle whereby transmitted seismic energy (from the acoustic source) incident on an acoustic interface is partly reflected by it (Sieck and Self, 1977). An acoustic interface is any interface within the strata across which there is a significant contrast in the acoustic impedance (change in strata density and subsequently sound velocity) properties of the sediment. Acoustic interfaces displayed on such profiles generally represent interfaces in the physical properties of the strata such as bedding planes, unconformities, faults and boundaries of gas zones (within the sediment) (Sieck and Self, 1977).

4.3.1 Aims of Shallow Seismic Survey Analysis and Interpretation

During this study the aims of seismic profile analyses were four fold:

1. To produce a set of stratigraphical cross sections from the study area of the Late Quaternary deposits of the inner continental shelf of the Eastern English Channel.
2. Once delineated, to map seismic sequences laterally between adjacent profiles based on the correlation of their external bounding reflectors and internal seismic signatures.

3. Using the grid of survey lines, to then build up a quasi 3-dimensional picture (in a GIS) of the study area, showing the shape, size and thickness of the seismostratigraphic units identified.
4. To build up and create a model of seismic sequence stratigraphy of the sea bed sediments in the study area.

For the purpose of this research, continuous analogue shallow seismic reflection profiles were collected in the study area using a surface-towed boomer acoustic source. Trackpoints at which the acoustic source was fired were recorded using a differential global positioning system to obtain as accurate a fix as possible for the data. At the same time the water depth was also recorded to obtain a high-resolution bathymetric survey for the study area. Two field surveys were undertaken to obtain the data for this study the first was in the summer of 1999 and the second in the summer of 2000. The seismic equipment was adjusted to obtain data on strata between the seabed and ~30 m beneath it (50 or 60 millisecond (ms) two way time (TWT)) sweep. In all some 500 kilometres of seismic line were collected. All depths in this thesis are expressed in relation to chart datum (CD) and all spatial data recorded in the Ordnance Survey of Great Britain (OSGB) coordinates.

Individual seismic profiles were described and interpreted following seismic sequence analysis and seismic facies analysis methods (see section 4.3.2 for a detailed description). These methods were used to identify depositional sequences, bounding unconformities, internal unconformities and diastems, system tracts and seismic facies, according to their amplitude, continuity and configuration geometry. The interpretational procedure adopted was a synthesis of the methods adopted by Berryhill (1986), Evans *et al.* (1995) and Foyle and Oertel (1997) while studying Quaternary continental shelf sediments. Berryhill (1986), Evans *et al.* (1995) and Foyle and Oertel (1997) further adapted the techniques described by Mitchum *et al.* (1977a and b), Mitchum and Vail (1977), Vail *et al.* (1977a and b) and Vail (1987). Seismic facies identified in this study were interpreted to represent depositional systems rather than their component lithofacies. Depositional systems are generally three-dimensional assemblages of process related lithofacies (Foyle and Oertel, 1997).

During interpretation, the seismic profiles were treated as geological cross-sections (e.g. Berryhill, 1986). The seismic profiles were examined with regard to identification of seismic sequences and then the delineation of individual seismic facies units based firstly on their external form or geometry, defined by bounding reflectors; and secondly their internal reflection configuration (seismic signature). Major reflector groupings (seismic facies) were cross-correlated between seismic lines across the study area.

4.3.2 Seismic Stratigraphy

The seismic profile (or section) often bears a striking resemblance to stratigraphic cross sections, which Sheriff (1977) suggested, invites direct interpretation. Seismic stratigraphy according to Vail and Mitchum (1977) is the geological approach to the stratigraphic interpretation of seismic reflection profile data. The unique properties of the data obtained using this technique allows the direct application of geological concepts in particular based on physical stratigraphy. It is a basic assumption of seismic stratigraphy that seismic sections are records of chronostratigraphic (time-stratigraphic) depositional and structural phenomena (e.g. faults or unconformities). Because of this Vail and Mitchum (1977) stated that it is not only possible to interpret postdepositional structural deformation from seismic profiles, but it is also possible to infer stratigraphic interpretations from the seismic reflection correlation pattern geometry. These include:

1. Geological time correlations.
2. Definition of genetic depositional units.
3. Thickness and depositional environment of genetic units.
4. Palaeobathymetry.
5. Burial history.
6. Relief and topography of unconformities.
7. Palaeogeography and geological history when combined with other geological data.

However a caveat placed upon the above interpretations is that with seismic data alone, the lithofacies and rock type can not be determined directly. To achieve the above objectives, the methods of *seismic sequence analysis* and *seismic facies analysis* developed by Mitchum and Vail (1977), Mitchum *et al.* (1977a and b), Vail et al. (1977a and b) and Vail (1987) are employed.

4.3.3 Seismic Sequence Analysis

Seismic sequence analysis consists of the identification of stratigraphic units composed of a relatively conformable succession of genetically related strata known as *depositional sequences*. Depositional sequences are three-dimensional assemblages of process related lithofacies (Foyle and Oertel, 1997), genetically linked by active (modern) or inferred (ancient) processes and environment (e.g., fluvial, deltaic, barrier-island) (Brown and Fisher, 1984). Thus each depositional sequence is made up of a succession of conformable strata bounded at the base and top by surfaces of discontinuity (unconformities or their correlative conformities) termed sequence boundaries.

Depositional sequence boundaries are, according to Mitchum *et al.* (1977), identified on seismic data by the recognition of reflectors, which suggest lateral terminations of strata and are delineated by identifying the termination of seismic reflectors at the discontinuity surface. If a termination occurs below the discontinuity it is termed the upper sequence boundary and may take the form of a discordant boundary such as *truncation* (erosional or structural) or *toplap*. Above a discontinuity (the lower sequence boundary) discordant boundaries are recognised as *onlap* or *downlap*. However terminations identifying both the upper and lower sequence boundaries may also show *concordance* (no termination). Figures 4-2 and 4-3 show simplified representations of the previously described boundaries.

4.3.4 Seismic Facies Analysis

Seismic facies analysis is undertaken within the framework of depositional sequences. It is the description and geological interpretation of seismic reflection parameters including configuration, continuity, amplitude, frequency and internal velocity, and also the external form and areal association of individual seismic facies units within the sequence (Mitchum *et al.*, 1977b). Each of the parameters can provide important information on the nature of the subsurface geology being investigated (Table 4.1).

Seismic facies units are mappable, three-dimensional seismic units composed of groups of reflections whose parameters differ from those of adjacent facies units (Mitchum *et al.*, 1977b). If the internal reflection parameters, external form and the three-dimensional form of seismic facies units can be recognised and delineated, it is possible to make a tentative interpretation of the environmental and depositional processes under which the facies unit was lain down as well as estimating the lithology of the unit.

The overall geometry of a seismic facies unit is made up of a combination of the internal configuration and external form of the unit. Descriptions of both are required to understand the geometric interrelation and depositional setting of the facies unit. Initial analysis should examine the two-dimensional features of a single seismic section and then later be combined with adjacent sections for corroboration in a quasi three-dimensional grid of sections (Mitchum *et al.*, 1977b).

4.3.5 External Form of Seismic Facies Units

Important in the analysis of seismic facies units is the understanding of the three-dimensional external form and areal association of the units. Mitchum *et al.* (1977b) divided the external geometry of seismic facies units associated with continental slope and shelf sedimentation as imaged on analogue seismic sections. They divided them into several types (Table 4.2 and Figure 4-4), some of which could in turn be subdivided into sub types.

4.3.6 Internal Reflection Configurations of Seismic Facies Units

According to Mitchum and Vail (1977), reflection configuration can commonly be used to interpret environmental setting, depositional processes, and estimates of lithology. Table 4.3 lists, and Figure 4-5 diagrammatically represents the principle internal reflection configurations outlined by Mitchum et al. (1977b). Within the external form of a single seismic facies unit, one or more reflection configurations may occur.

Parallel and *subparallel* reflection configurations most commonly form in sheet, sheet drape, and infilling units (Figure 4-5). The uniform configuration pattern is interpreted to suggest deposition under tranquil conditions (Evans *et al.*, 1995).

Divergent/Convergent reflection is characterised by wedge shaped units (Figure 4-5). Their configurations are usually associated with lateral variations in the rate of sediment deposition, or progressive tilting of the depositional surface.

Prograding reflectors occur when strata is interpreted as having been deposited due to lateral outbuilding or prograding from source. Gently sloping depositional surfaces develop laterally within the prograding units, called *clinoforms*, one of the most common depositional features (Mitchum *et al.*, 1977b). *Prograding clinoforms* occur in strata in many different external shapes (Table 4.3 and Figure 4-5). *Sigmoid* progradational configurations have S-shaped reflections, with thin gently dipping upper and lower segments, known as the *topset* and *bottomset* respectively. The middle more steeply dipping thicker segment of a clinoform is known as the *foreset*.

Oblique progradational configurations can be tangential or parallel. The topset portion of the clinoform is absent with successive foresets building out (older – younger) in a depositionally downdip direction. Ideally they consist of a number of steeply dipping clinoforms terminating at the upper boundary by toplap and downdip by downlap at the lower sequence boundary.

Combinations of alternating sigmoid and oblique progradational configurations within seismic facies units are described as *complex sigmoid-oblique*. The upper part of the facies units are characterised by complex alternations of sigmoid topsets and oblique configurations terminating with toplap.

Shingled progradation occurs as a thin prograding seismic pattern, most frequently with parallel upper and lower boundaries (Mitchum *et al.*, 1977b). Internal reflectors in shingled progradation are parallel oblique, terminating in apparent toplap and downlap. Shingled prograding clinoforms are most common in seismic facies that have been interpreted as units which prograded into shallow water (Mitchum *et al.*, 1977b).

Hummocky clinoform configurations are made up of irregular discontinuous subparallel reflections forming random hummocky patterns. These patterns are generally interpreted as signifying deposits, which have formed as interfingering clinoform lobes which built up in shallow water in a prodelta or inter-deltaic position (Mitchum *et al.*, 1977b).

A *chaotic* reflection configuration on a seismic section is characterised by discontinuous, discordant patterns of reflectors, which suggest a disordered arrangement of reflection surfaces. Chaotic configurations are interpreted as strata that was either deposited in variable, high-energy settings, or strata that was initially continuous but has been deformed post-depositionally so as to disrupt the continuity (Mitchum *et al.*, 1977b).

Areas on seismic section that are homogeneous, unstratified or highly contorted are said to be *reflection-free*. Examples of strata that might appear reflection free on a seismic section would be thick seismically homogeneous shales or sandstones (Mitchum *et al.*, 1977b).

Reflector configurations may also be described further using a series of common modifying terms outlined by Mitchum *et al.*(1977b) (Table 4.3). Examples are *even, wavy, regular, irregular, uniform, variable, hummocky, lenticular, disrupted* and *contorted.* Such terms are utilised when minor variations occur in the above outlined basic patterns.

Mitchum *et al.* (1977b) stated however, that there is no direct relationship between internal reflection configuration types and specific lithologies and that where possible, all available other sources of data should be integrated with the seismic interpretation to afford any sort of valid prediction of depositional environment and lithology.

During this study most bounding reflectors were considered to represent erosional surfaces that generally truncate the underlying units. However, bounding reflectors may also be represented by a hiatus of non-deposition that is not erosional.

4.3.7 Limitations of Shallow Seismic Analysis

The major limitations and weaknesses of shallow seismic profile interpretation arise from uncertainties in the nature and origin of seismic reflectors (Bellamy, 1995) and also in their conversion from travel time on the analogue profile to depth in metres. Stoker *et al.* (1992; 1997) working on the continental margin off the formerly glaciated northwest of Britain, showed that spatial changes in seismic signature do not always correspond to changes in lithology. Moreover acoustically homogeneous seismic responses in the profile may not be representative of homogeneous sedimentary successions, highlighting that in glaciated terrains care must be exercised in inferring geological characteristics from seismic data alone

During this study, in order to represent the interpreted seismic facies in relation to their depth a sound velocity figure of 1700 metres per second (m/s) has been adopted to convert the acoustic velocity to thickness (time-depth conversion) of Quaternary sediments in metres. Generally sound velocity in unconsolidated sediments ranges between 1500 m/s and 1800 m/s (Bennell, 1981). The figure of 1700 m/s is commonly used as an intermediary value (e.g. Needel *et al.*, 1987; Peterson and Phipps, 1992; Evans *et al.*, 1995; Stoker, 1997; Twichell and Cross, 2002; Dr. Ian Selby, *pers. comm.*). It is considered that given the shallow depths of sediment below the seabed, and the relatively short two-way acoustic travel times, the assumption will not greatly affect the accuracy of time-depth conversions from the profiles for example.

4.4 Lithostratigraphy

On the basis of preliminary examination of the seismic data obtained in the study area, sampling sites were chosen for retrieval of sediment samples to: 1) help characterise the identified seismic facies units and to aid in the determination of their thickness and extent, and 2) aid in the interpretation of the environment of deposition of the identified seismic facies units. At the chosen sites sediment samples were retrieved from the seabed using a 6 m long vibrocorer deployed from a ship (Figure 4-6). The maximum possible recovery of sediments was 6 metres due to the length of the apparatus. During sampling the recovery was commonly less than this as the vibrocorer encountered impenetrable substrates. As the data collected were for aggregate surveys, sampling sites were mainly located where initial seismic data interpretation inferred the sediments to be indicative of high-grade resource (coarse sands and gravels). There was therefore bias in the sampling towards units, which had been interpreted as coarser grained and areas with potentially fine-grained sediments were avoided.

4.4.1 Vibrocoring

Vibrocoring is a technique for collecting core samples of sub seabed sediments. The vibrating mechanism of a vibrocorer, sometimes called the "vibrahead", operates on hydraulic, pneumatic, mechanical or electrical power from an external source. The attached core tube is driven into sediment by the force of gravity, enhanced by vibration energy. When the insertion is completed, the vibrocorer is turned off, and the tube is extracted from the substrate with the aid of ship-born hoist equipment.

Vibrocoring provides relatively undisturbed samples through the sub-seabed lithology, which reveals detail on both the composition of discrete sedimentary units and the nature of transitions between units. This in turn assists seismic interpretation by assigning lithologies to seismic facies units essentially ground-truthing the data. Stoker *et al.* (1992) considered that without sample control to ground-truth seismic data, inferences drawn from its interpretation are equivocal. Therefore they suggested that if lithological samples of at least 1.5 m length can be recovered from beneath the seabed in areas where seismic data has been collected, such data could provide the necessary information to characterise most of the seismic responses encountered.

4.4.2 Limitations of Vibrocoring

For the purposes of this study the main limitation of the vibrocoring program was lack of penetration and recovery. The maximum penetration depth was 6 m (and frequently less than 6 m) below the seabed. This samples only the upper deposits and is of little value for providing lithostratigraphic control for seismic data below this depth in thicker sequences, such as the infilled palaeovalleys encountered in this study (see Chapter 5). Therefore the lack of ground-truthing sample control over certain seismic facies units means that interpretations of sedimentary history are necessarily more tentative.

As mentioned earlier another limitation to the vibrocoring program was that sampling sites were selected on the basis that core samples would potentially reveal sites of high grade aggregate resource potential, therefore areas where the seismic data was interpreted to show facies units of low grade resource potential were avoided during the program and little or no sample data was obtained for such units. The implications for this study are that very few samples of more fine grained materials such as clay, silt or peat deposits i.e. lain down in low energy conditions, have been sampled, providing they were present in the study area at all. Sampling was therefore biased to deposits of coarser grained materials (sands and gravels) of probable high-energy depositional conditions.

As a vibrocore tube penetrates the sediment the material captured inside the descending tube generally moves upward at the same rate. Occasionally, however, the increasing wall friction inside the tube can exceed the bearing strength of the sediment. So, even though the tube continues to penetrate, sediments within it fail to move up. Then the tube behaves like a solid rod. No more sediment is collected unless the tube penetrates to a harder layer and the friction inside is exceeded once more. As a result some intermediate layers of sediment may be bypassed, unbeknownst to the sampler. This process is known as a rodding or plugging effect (http://www.vibrocoring.com/VCconcepts.html, 2004). Following examination of the retrieved samples rodding does not appear to have been a problem during this study.

4.4.3 Vibrocore Samples

Some 189 sediment core samples were extracted from the study area. 140 were retrieved in 2001 using the vibrocoring system of Andrews Survey Ltd and further 49 cores were extracted in 2002. At sites where vibrocore samples were taken (Figure 4-7), the position of the cores was noted by an onboard hydrographer utilising a differential global positioning system (DGPS) and echosounder. This was undertaken to obtain as accurate a fix (easting, northing and water depth) as possible for each sampling site. The depth of the each core was calculated by the hydrographer to chart datum, following processing of the echosounding data through a swell filter and compensating for tidal variation.

4.5 Laboratory analyses

All the recovered sediment core samples were returned to the laboratory, split in half, photographed and logged using conventional lithological description techniques (outlined in Boggs, 1995). On the basis of the lithological description, sediment samples were taken from distinct lithological units for laboratory analyses (grain-size, biostratigraphical (diatom) and geochemical). These analyses (described below) were carried out to further assist seismic data interpretation and to aid in the interpretation and understanding of the nature and origin of the sediments.

4.6 Grain-Size Analysis

Where possible at least one sample was taken from each lithostratigraphic unit identified within the core samples for grain-size analysis. The laboratory techniques used for grain-size analysis during this study are outlined in Appendix 1.

4.6.1. Aims of Grain-Size Analysis

The aim of grain-size analysis was to quantitatively describe the composition of the sediment and to aid in the characterisation of the environment of deposition of lithostratigraphic units. To this end the grain-size data have been examined using the standard statistical measures of mean, median and modal grain-size, standard deviation (sorting), skewness and kurtosis. These and other measures have been claimed to provide important clues to the sediment provenance, transport history and depositional conditions (Folk and Ward, 1957; Folk, 1966, 1974; Friedman, 1967, 1979; Sly, 1977, 1978; Friedman and Sanders, 1978; McManus, 1988; Bui *et al.*, 1990; Syvitski, 1991; Boggs, 1995). For example, Friedman (1967) demonstrated that two-component variation diagrams, in which pairs of grain-size statistical parameters are plotted against one another (e.g. skewness vs. standard deviation, or mean grain-size vs. standard deviation) was one such case. Using this method he suggested that sediments could be separated into major fields relating to their environment of deposition such as beach and fluvial environments.

4.6.2 Grain-Size Statistics

The logarithmic phi (ϕ) transformation scale introduced by Krumbein (1939) has been utilised for the analysis of grain-size data produced by laboratory analysis of sediment samples in this study, whereby.

$$\text{Phi } (\phi) = -\log_2 d$$

Where d = particle diameter in millimetres

The scale was proposed by Krumbein (1939) to simplify statistical analysis of grain-size data due in the most part to the comparison of grain-size populations necessitating the consideration of a wide range of particle sizes. Therefore it is a geometric scale as opposed to an arithmetic one.

The resulting data were statistically analysed to produce the parameters outlined below employing the following commonly used method of moments formulae (Table 4.4) (McBride, 1971).

The *mean* (x_{ϕ}) is the best measure of average in the grain-size distribution of a sediment sample (McManus, 1988). It was considered by Folk (1974) to provide an indication of the minimum energy required for the sediment to have been transported. The *median* is represented by the fiftieth percentile in a distribution and is most useful in sediments with a unimodal but skewed distribution. The *modal* size is the one represented by the most common grain-size class in a distribution.

The standard deviation (σ_{ϕ}) of the grain-size distribution provides information relating to the *sorting* of the sample, i.e. the degree of uniformity produced by sediment processing (see Table 4.5). For example, the continued winnowing of wave action tends to improve the degree of sorting (reduction of standard deviation) of sediments.

The tendency of a grain-size distribution to deviate from normality leads to differences between the mean and median values. These differences can be used to characterise the asymmetry or *skewness* (Sk_{ϕ}) of the curve of the distribution (McManus, 1988). The distribution can be said to be positively or negatively skewed dependent on when more coarse or fine materials are present than in a normal distribution. Skewness values lie within the range +1 to −1 (Table 4.6). A samples skewness value is considered to be extremely sensitive to the depositional environment in which it was laid down, as it is indicative of the preferential addition or removal of particles at the coarse and fine tails of the distribution.

The *kurtosis* (K) of a grain-size sample is related both to the normality and dispersion of the distribution it has also commonly, and according to McBride (1971) inappropriately, referred to as the concentration or peakedness of the distribution. The measure of kurtosis (equation above) is a ratio of the spreads of the tails and the centre of the distribution curve. Therefore flat frequency curves of poorly sorted sediments or those with more than one modal value are platykurtic, whereas strongly peaked frequency curves in which the central part of the distribution is well sorted are leptokurtic (Baker, 1968). The different kurtosis categories are listed in Table 4.7.

Grain-size data was entered into MS Excel and processed using the GRADISTAT package of Blott and Pye (2001) to calculate the above described statistics and for subsequent intercomparison of the data.

4.7 Diatom and Geochemical Analyses

Two transects of cores (Transects A-A and B-B; Figure 4-7) were selected for diatom and geochemical laboratory analyses. The transects were selected on the basis that they could potentially yield the broadest range of information relating to the two widest and thickest infilled features recognised in the study area.

4.7.1 Diatom Analysis

Diatom analysis was undertaken on samples from boreholes in both transects 1 and 2. Samples were prepared according to well-established techniques following Palmer and Abbot (1986) (for preparational methodology see Appendix). Microscope analysis was carried out using a light transmission microscope under oil immersion with a magnification of x1000. Identification of individual diatom taxa were made with reference to Van der Werff and Huls (1958-1974), Hendey (1968) and Hartley (1986, 1996). Taxonomic nomenclature was also made with reference to the atlas of British diatoms arranged by Hartley (1996).

Alderton (1994) identified that analysis of fossil diatom assemblages in sediments can be useful and effective tool in palaeoenvironmental research providing precise information with regard to the nature of the environment in which the sediment was deposited. In particular, in the following three ways:-

1. Diatoms naturally occur in a wide range of sub-aqueous environments e.g. freshwater ponds to saline springs. Therefore demonstrating a very wide range of tolerance to salinity
2. The robust nature of the silicon frustule enables diatoms to be readily and well preserved in a range of sedimentary deposits and over long periods of time.
3. Due to their minute size and abundance only a relatively small sample of sediment is required to complete the necessary analysis.

4.7.1.1 Aims of Diatom Analysis

Within this study, diatom analysis was undertaken with the following main objectives: (i) to determine the nature of the palaeoenvironmental conditions in which the sampled unit was laid down; and (ii) to support lithostratigraphic correlation.

4.7.1.2 Diatom Classification

Diatoms have been classified for this study according to a simplified halobian classification scheme based on those of Hustedt (1953, 1957). The individual taxa are divided into the following five groups based on the salinity range of the waters in which they live:-

(a) Polyhalobian- >30 ‰ salinity, (marine)

(b) Mesohalobian- 0.2-30 ‰ salinity range, (marine-brackish)

(c) Oligohalobian-halophilous- (brackish) optimum in slightly brackish water

(d) Oligohalobian-indifferent- (fresh-brackish), optimum in freshwater, but tolerant of slightly brackish conditions.

(e) Halophobous- (fresh) intolerant of saline conditions, exclusively freshwater

Within this study many diatom taxa were not recorded intact or they were preserved in a heavily degraded or broken state hampering identification. This was due in the main to the nature of the sediments analysed. Preservation potential of diatom frustules is significantly reduced when deposited coarse grained sediments usually associated with high energy conditions. In samples where this occurred if a broken or degraded valves were encountered and could be identified they were assumed to constitute a single valve. On some occasions the degradation only enabled identification to the generic level.

Due to the low concentrations of diatom valves in the sediment samples examined it was impossible to count a significant number to obtain a statistically robust environmental interpretation. Therefore, the data presented only provides an inference of the envioronment of deposition from the diatom species found. A full listing of all counts can be found in Appendix 3

4.7.2 Multi-element Geochemical Analysis

The analysis of multi element geochemistry of a suite of sediments is a technique that has proven to be a useful tool to assist in the development of a stratigraphy and in determining their provenance (Plater *et al.*, 2000; Ridgway *et al.*, 2000). The methodology involves sampling individual lithostratigraphic units to determine the distribution of a suite of chemical elements contained within them and to then examine the degree of their variation between the units.

4.7.2.1 Aims of Geochemical Analysis

It was hoped that elemental analysis of sediments in this study would help to: 1) further establish the stratigraphy; and 2) potentially aid in their provenancing.

4.7.2.2. Previous Application of Technique

Ridgway *et al.* (2000) stated that the most important factors determining the distribution of elements in sedimentary rocks and sediments are: (1) the mineralogical and chemical composition of their source materials; and (2) the partitioning of the elements between sediment, surface and groundwater during their deposition and possible post-depositional diagenesis. Therefore the geochemistry of sediments can reflect provenance, conditions of deposition, diagenesis or combinations of the three (Haslam and Plant, 1990).

During the land ocean interaction study (LOIS) initiative (1990-1997) funded by the Natural Environment Research Council (NERC) multi-element geochemistry of sediment core samples was seen as a potentially powerful aid to the interpretation of the stratigraphy of sediments (Plater *et al.*, 1998). Studies conducted in the estuaries of the Rivers Humber (Rees *et al.*, 2000; Ridgway *et al.*, 1998, 2000) and Tees (Plater *et al.*. 1998, 2000) as well as offshore in the North Sea (Brew *et al.*, 1999) showed that recognisable stratigraphic units in cores had distinctive multi-element geochemical signatures that could be correlated between boreholes.

Ridgway *et al.* (2000) suggested that the geochemistry of the majority of clastic Holocene sediments in the UK was controlled in the most part by source and depositional processes. The sediments examined in the four main study areas of the LOIS initiative (Teesside, Humber, Fenland and north Norfolk) (Rees *et al.*, 2000; Ridgway *et al.*, 1998, 2000; Plater *et al.*, 1998, 2000) assisted significantly in developing a stratigraphy for deposits in the areas as well as in provenance determination.

4.7.2.3 Analytical Procedure

Samples for geochemical analysis were taken over the same sampling intervals as used for diatom analysis. Initially the samples were sieved through a 150 μm nylon mesh, a new piece used for each sample (to prevent cross contamination), to obtain the <150 μm fraction of the sediment so as to consistently analyse the same size fraction in each sample. Following digestion (for laboratory procedure see Appendix 1) the samples were analysed using an Elan® inductively coupled plasma mass spectrometer (ICPMS) (see Appendix for labroratory method) to determine the elements Ce, La, Nd, Pr, Sm, Gd, Dy, Er, Yb, Eu, Ho, Tb, Lu and Tm. The acquired data were then normalised to a suitable datum (the upper continental crustal average of Wedepohl, 1995; see Table 4.8) to facilitate intercomparison. The normalised data were statistically analysed using detrended correspondence analysis (DCA) to see how samples clustered according to their similarities in elemental concentrations. This allowed grouping of the samples into suites of sediments.

Chapter 5: Results

5.1 Introduction

This chapter presents the results of the field surveys (seismic profiling and vibrocoring) undertaken in the PhD study area and the subsequent laboratory analysis of these data. Firstly the results of the seismic reflection profiling surveys (1999 and 2000) are presented to establish a sequence stratigraphic framework for the seismic facies units identified in the study area. This is followed by the results of the vibrocoring surveys (2000, 2001 and 2002), which adopt the same stratigraphic framework as the seismic data. Included in this section are the results of the grain-size analysis undertaken on samples from each sample vibrocore. Finally the results of a detailed analysis of two key cross-sections in the study area are presented. This includes the presentation of more detailed lithostratigraphic, grain-size, biostratigraphic and geochemical data.

5.2 Seismic Stratigraphy

5.2.1 Introduction

This section describes the results obtained from analysis of the shallow seismic reflection profiling survey data obtained in 1999 and 2000. The combined area of Licence Application Areas EEC474 and 475, from which the data were collected, is approximately 462 km^2 of the seabed. It forms part of the gently inclined marine planation surface (discussed in section 3.4.1) of the inner continental shelf. Figure 5-1 shows the trackplot for the seismic surveys from which the seismic stratigraphic data were collected. Track lines were on average 1 km apart, with a total length circa 500 km.

The analysis of seismic sequences is based on the separation of time depositional units defined by unconformities or changes in seismic patterns (Sheriff, 1980; Sheriff and Geldart, 1994). This entails the partitioning of the seismic trace into units based on similarities in their seismic velocity, with the resulting sequences genetically, rather than lithologically, related.

Individual survey lines analysed during this study occasionally revealed more complex relationships between seismic facies. On the scale of this survey, however, it was not possible to correlate the vast majority of these features between lines or, for that matter, place them effectively in the context of a large-scale model of landscape evolution, which the study aims to accomplish.

The seismic data record a maximum sub-surface penetration of 60 ms, which equates to approximately 51 m of sediment penetration (this is based on an average velocity of 1700 m s^{-1}). The descriptive terminology used for these data is taken from Mitchum *et al.* (1977a, 1977b), Berryhill (1986) and Evans *et al.* (1995) (see Chapter 4). A summary table of the major characteristics of the seismic facies units identified is presented in Table 5.1.

5.2.2 Bathymetry of Study Area

The bathymetry in and around the study area is shown in Figure 5-2 at two metre isobath intervals. The map is based on the interpolation of bathymetric data collected during the seismic reflection profiling surveys. The seabed slopes in an east-west direction with water depths increasing from -30 m below chart datum (CD) in the east to -56 m CD in the west, at an approximate gradient of 1:2000 (0.03°). Within, and to the northwest of the study area there are a number of linear, elongated deep (10 m) hollows, trending northeast to southwest, that slope in the same direction as the seabed (Figure 5-2). They occur around the 46 metre isobath and reach depths of over -54 m CD.

5.2.3 Seismic data

The seismic profiles reveal three prominent, areally widespread, reflector surfaces. They are labelled, from the base to the top of the section, the LQ_B, the R_I, and the H_b surfaces. Between these major reflector surfaces are four seismic sequences that comprise eight different seismic facies units.

I begin by describing the gross geometry of the unconsolidated sediment sequence by separating the identified facies units into a seismic sequence stratigraphy separated by the major reflector surfaces (LQ_b, R_I and H_b). Each sequence and its associated facies units are described in stratigraphic sequence. Each of the seismic facies units is defined according to the nature of the external bounding reflectors delimiting them, the three-dimensional geometry of the bounding reflectors, and the internal seismic signature of the units (following Mitchum *et al.*, 1977b). The location of the key seismic sections discussed in this section are shown in Figure 5-3.

In some instances the lower bounding reflectors of individual facies units are identifiable on adjacent seismic sections. Where this occurs the data have been combined to create interpolated isopleth maps that reconstruct the lower bounding reflector surface of the individual facies units in relation to present sea-level.

5.2.4 Seismic Sequence I

Reflector Surface LQ_b and Seismic Facies Unit A (SFU-A)

Seismic facies unit A (SFU-A) represents the lowermost seismo-stratigraphical facies unit identified in this study. It is present over the entire study area beneath all other units. In the most part, other seismic units overlie it, but in some areas it outcrops at or very near the surface and forms the seabed.

Figure 5-4 shows an isopachyte map of the gross geometry and thickness (in metres) of the sedimentary succession overlying and infilling the upper bounding reflector of SFU-A. The overlying sediment ranges in thickness from 0 to 34 metres. Figure 5-5 shows a contour map of the depth to the upper bounding reflector surface LQ_b (i.e. bathymetry + sediment thickness). The laterally extensive surface LQ_b represents the sequence boundary between Seismic Sequences I and II. The maximum and minimum depths of the upper bounding reflector of surface LQ_b, is -73 m -33 m CD respectively. There are two areas that contain the thickest infilling sediments and achieve the greatest depths below CD of the upper bounding reflector surface of SFU-A (LQ_b):

1) Between 567000E and 575000E there is an elongated feature running from the north to the south of the study area in a NNW to SSE direction ranging in width from approximately 3000 m in the south to over 8000 m in the north of the area and depths of up to –65 m CD (Figure 5-5). The feature is infilled with sediments consistently over 20 m thick along the western edge of its length (Figure 5-4).

2) In the northeast of the study area from 583000E to the easterly limit of the area, the upper bounding reflector of SFU-A is recorded at depths ranging from -35 m to > –73 m CD (Figure 5-5). The sedimentary sequence overlying surface LQ_b in this area is locally >30 m thick (Figure 5-4).

In addition to the above are two further smaller features worthy of note. The first is in the southwest of the study area between 557000E and 559000E, where a depression in the upper bounding surface of SFU-A is infilled with up to 20 m of sediments. The second feature is also a depression in the upper bounding surface of SFU-A. It occurs at the northern limit of the study area between the two previously described major areas of infill between 578000E and 580000E. It is infilled with a sequence of deposits ranging in thickness up to 20 m.

SFU-A does not display a basal bounding reflector on the seismic profiles and is therefore considered to be the only facies unit of Seismic Sequence I in the study area. The upper bounding surface of SFU-A (LQ_b) is formed by the truncation of internal reflectors, that terminate at either the seismic facies units above (Figures 5-6 and 5-7) or the seabed (Figure 5-8). When beneath other seismic facies units in the sequence (Figure 5-6) these truncated reflectors terminate at an undulating surface apart from at the smoother almost horizontal seabed surface (Figure 5-9).

The internal reflectors within SFU-A are generally oblique to, or terminate at the base of, the overlying units (Sequence boundary surface LQ_b). The internal seismic signature of SFU-A consists of a series of moderate-high amplitude, continuous, parallel, dipping and occasionally concave reflectors (Figures 5-6). The reflectors are unevenly vertically spaced and range in their apparent angle of dip from 0.1° E in the southwest of the study area (Figure 5-9), to nearly horizontal in the centre of the study area (Figure 5-7), through to 0.35°W in the eastern part of the study area (Figure 5-10). Between reflectors the internal seismic signature of SFU-A displays either medium-high density homogeneous backscatter or is reflection free.

5.2.5 Seismic Sequence II

Seismic Sequence II is represented by the seismic facies units between the major bounding reflector surfaces of LQ_b below and surface R_1 above.

Seismic Facies Unit B (SFU-B)

Seismic facies unit B (SFU-B) is the lowermost facies of Seismic Sequence II, and overlies reflector surface LQ_b. It is limited to Licence Application Area EEC474 occurring in the northeastern portion of the study area. Figure 5-11 shows an isopachyte map of the thickness of sediments that constitute SFU-B. Figure 5-12 shows a map of isobaths indicating the depth below CD to the basal bounding reflector surface (LQ_b) of SFU-B. Also shown on both figures is the limit of the areal extent of the unit within the study area.

SFU-B is the deepest of the seismic facies units in the sequence, with the lower bounding reflector surface occurring between the depths of –36 and –73 m CD (Figure 5-13). The lower bounding reflector of SFU-B and the upper bounding reflector of SFU-A are contiguous with each other across the whole extent of SFU-B, the boundary of which displays an unconformable discordant relationship. The unit displays very undulating, low amplitude, uneven and discontinuous lower bounding reflectors (Figure 5-10, fixes 7046-7075) ranging from almost horizontal (Figure 5-13) to an apparent dip of 2.5° (Figure 5-10, fixes 7047-7049). The upper bounding reflector surface of the unit (R_1) is distinguished by an undulating (Figure 5-13, fixes 10251-10264) to relatively flat (Figure 5-13, fixes 10274-10294), moderate- to high amplitude reflection signal. The unit is commonly more than 10 km wide extending from 585000E at its western limit to 595800E in the east. The northern and southern most aerial limits of the unit are not detected, and they appear to extend beyond the boundaries of the study area.

The internal seismic signature of SFU-B appears in the main to lack any structure. It is dominated by hummocky and chaotic reflectors and high-density backscatter (Figures 5-12 and 5-12). The unit also contains occasional moderate- to low-amplitude subparallel, prograding east to west (at an approximate apparent dip of 1°) reflector patterns that downlap onto the lower bounding reflector surface (Figures 5.10). SFU-B ranges in sediment thickness up to 24 m with an average of 9 m (Figure 5-11).

Seismic Facies Unit C (SFU-C)

Contour maps of interpolated thickness of sediment (isopachyte) and the depth below CD (isobaths) of the lower bounding reflector (LQ_b) of Seimic facies unit C (SFU-C) are shown in Figures 5-11 and 5-12 respectively. SFU-C occurs on 19 of the 37 seismic profile sections analysed. SFU-C appears in both areas EEC474 and EEC475. The unit forms an elongate feature that infills an incision into surface LQ_b that trends in a NNW-SSE direction, and attains depths of between -40 m and -72 m CD. The deepest incision occurs on the western flank of the unit (Figure 5-12).

The northerly and southerly limits of SFU-C extend beyond the boundaries of the study area. The western edge of the unit trends in the same direction as the major incised feature, between 566000E in the north and 572000E in the south, and the eastern limit trending approximately north-south along the line of 575000E.

The principal differences between SFU-C and SFU-B are the spatial distribution of the units and differing internal signatures. As with SFU-B, the interface between the lower bounding reflector of SFU-C and the upper bounding reflector SFU-A is discordant and contiguous on all seismic sections examined. The nature of the lower bounding reflector (LQ_b) of SFU-C and SFU-B is also alike in so much as it is undulating, of low amplitude and discontinuous (Figures 5-6, 5-7 and 5-14), ranging from almost horizontal (Figure 5-7, fixes 8663-8671) to an apparent dip of 5°W (Figure 5-7, fixes 8654-8656). The upper bounding reflector surface (R_1) of SFU-C displays a range of features including variable-amplitude, undulating (almost jagged) to even, discontinuous to continuous reflectors. Internally no coherent reflectors are visible, the unit is slightly transparent with a predominantly chaotic seismic signature with low-medium density backscatter. Occasional low amplitude subparallel reflectors are also present (Figures 5-6, 5-7 and 5-14) but the prograding nature of the reflectors shown in in SFU-B are not apparent in this unit. The unit contains sediment thicknesses of up to 25 m (Figure 5-11).

Seismic Facies Unit D (SFU-D)

SFU-D occurs on only two of the seismic reflection profiles analysed (Lines Y5 and X6) perpendicular to one another. SFU-D occurs in the north of the study area in between seismic facies units B and C. The depth of the lower bounding reflector surface (LQ_b) ranges between –45 m and –65 m CD. This is somewhat shallower than SFU-B and –C. Figure 5-16 shows a typical example of the full extent of SFU-D.

The interface between the lower bounding reflector (LQ_b) of the unit with the upper bounding surface of SFU-A is again shown to be discordant and contiguous. The undulating boundary is identified on the seismic section by a mainly continuous moderate-amplitude reflector that cuts into SFU-A below. The upper bounding reflector surface (R_l) of SFU-D forms a mildly undulating almost wavy surface that is overlain by the unit above (SFU-G).

On Line Y5, SFU-D is some 2500 m wide and 14 m thick at its maximum (Figure 5-15). The internal seismic signature of the unit has a number of characteristics; at the base of the unit between fixes 1479-1481 the signature is almost reflection free. However, the majority of the signature is chaotic with backscatter ranging from high to low density. The unit occasionally displays dispersed hummocky type reflectors (Figure 5-15, fixes 1477 and 1490-1491).

Seismic Facies Unit E (SFU-E)

SFU-E occurs on lines 52 and Y2 only. It forms as a basin-like depression below the sea floor in the region of GR55000N, 565000E. SFU-E is approximately 500 m wide east – west and up to 20 m thick (Figure 5-16). The lower bounding surface (LQ_b), contiguous with the upper bounding surface of SFU A, is defined by a moderate- to high-amplitude reflector, undulating and discontinuous that ranges in depth from –46 m to –63 m CD (Figure 5-16). The upper bounding surface (R_l) of SFU-E is well defined with a moderate relief. It has a strong, continuous high-amplitude reflection boundary with SFU-H and SFU-J above. SFU-E is characterised by chaotic internal reflectors with a few strong (high-amplitude) reflectors apparent and is dominated by high-density backscatter.

Seismic Facies Unit F (SFU-F)

Seismic facies unit F (SFU-F) is located in the southwest of the study area around GR50000N and between GR555500E and GR560000E away from all the other major subsurface depressions. The unit only occurs on seismic lines 69, Y4 and 56, but potentially extends south beyond the limits of the study area. SFU-F is over 1500 m wide in an east-west cross section and up to 13 m thick.

Again, the lower bounding reflector of SFU-F (LQ_b) is contiguous and discordant with the upper bounding reflector of the basal sequence SFU-A (Figure 5-9). It is characterised by continuous even to mildly undulating, high-amplitude reflectors that dip from both the east and west (at approximately 0.5 - 1°) to form a trough/basin-like seismic feature. The upper bounding reflector (R_l) is contiguous with the lower bounding reflector of SFU-G (see below). Internally SFU-F has a simple chaotic reflection pattern.

5.2.5 Seismic Sequence III

Seismic Sequence III consists of only a single but very extensive seismic facies unit (SFU-G). It is bounded at its base by the reflector surface R_l and at the top by reflector surface H_b.

Seismic Facies Unit G (SFU-G)

SFU-G is the largest of the seismic facies units identified in the study area. It is present over large proportions of the area in thicknesses of up to 20 m (Figure 5-17) and caps all of the previously described facies units. Its lower bounding surface (R_l) occurs between –33m and –62 m CD (Figure 5-18), although it is mostly in the range of –44 m to –50 m CD. SFU-G unconformably overlies seismic facies units SFU-A (Figures 5-10 and 5-17), SFU-B (Figures 5-10 and 5-13), SFU-C (Figures 5-6, 5-7 and 5-14), SFU-D (Figure 5-15), SFU-E (Figure 5-16) and SFU-F (Figures 5-9). The lower boundary of SFU-G is well defined at the interface with the upper surface of SFU-A, and is characterised by high-amplitude reflectors (Figure 5-20). The boundary between SFU-G and SFU-B, -C, -D, -E and –F is less clear and can be described mostly as arbitrary (Ravenne, 1978) i.e. the seismic parameters change only gradually across the sequence (Figures 5-6, 5-7). Occasionally the boundary is more defined (SFU-B, Figure 5-10, fixes 7052-7061; SFU-C, Figure 5-14, fixes 0533-0577; SFU-D, Figure 5-15, fix 1480; SFU-E, Figure 5-16, fix 9196; SFU-F Figure 5-9, fix 7359). Across all underlying units the nature of the boundary varies from even to wavy, undulating and occasionally almost jagged (Figure 5-15). The high amplitude continuous upper bounding surface of SFU-G (H_b) is commonly horizontal when the unit represents the sea floor, and even to undulating when unconformably overlain by seismic facies units –H and –I.

Internally, SFU-G displays a broad range of seismic signatures. Towards the base of the unit are occasional parallel horizontal reflectors that onlap the lower bounding surface (Figure 5-6). Mostly, however, the unit is characterised by moderate to high amplitude, irregular, discontinuous and commonly steeply inclined, if not chaotic reflectors (Figures 5-6, 5-7, 5-15 and 5-17). Intermittently the unit also shows reflectors that dip steeply, and truncate or disrupt each other (Figure 5-14). SFU-G is therefore described as demonstrating a very varied seismic signature.

5.2.5 Seismic Sequence IV

Seismic Sequence IV is composed of two distinct seismic facies units (SFU-H and SFU-I) overlying the major reflector surface H_b. The upper bounding surface of this sequence is the seabed.

Seismic Facies Unit H (SFU-H)

Seismic facies unit H (SFU-H) is limited to an area approximately 2.8 km^2. It appears on lines 32, 31, 30, 29 and 5pt4, and is defined by a basin/trough-like feature (2000 m long N-S, 1700 m wide E-W, Figure 5-23). The lower bounding surface of SFU-H (H_b) is represented on the seismic sections by a single clear high-amplitude reflector at apparent dips of over 5°E and 0.75°W (Figure 5-6 and 5-7), truncating and incising into SFU-G. SFU-H occurs immediately beneath, and is truncated by the lower bounding reflector of SFU-I (see below). It ranges in thickness from 1 to 15 m and occurs at depths between –38 m and –53m CD.

On the western portion of SFU-H, horizontal to dipping, parallel, evenly spaced reflectors onlap the unconformable lower bounding reflector (Figure 5-6 and 5-7). Overlying reflectors in SFU-H dip in an easterly direction. These internal reflectors terminate (by toplap) against the unconformable lower boundary of SFU-I, they also downlap onto the lower bounding surface of the basin/trough (Figures 5-6 and 5-7).

Seismic Facies Unit I (SFU-I)

The uppermost unit of the seismic stratigraphy is formed by seismic facies unit I (SFU-I). This facies unit dominates sea bed exposures throughout much of the study area. In all areas this unit rests unconformably on the upper reflector (H_b) of SFU-A, SFU-G or SFU-H (Figures 5-6 and 5-7). Its lower bounding surface is most commonly is defined by a single high amplitude continuous reflector. The unit varies in thickness from a few decimetres to a maximum of around 10 m (average 0.5 m to 1 m; Figure 5-19). Internally the seismic responses Where SFU-I attains sufficient thickness on seismic profiles to display an internal signature, the unit is characterised by even to wavy, parallel reflectors, that are horizontal (Figure 5-14), or dipping at very shallow angles in an easterly direction (Figures 5-6 and 5-7) and terminating at the high amplitude reflection of the sea bed that forms the upper bounding surface of the sequence in the PhD study area.

5.3 Lithostratigraphy and Sedimentology

In this section I present the results of the lithostratigraphic and sedimentological data that have been obtained from the study area. This includes data from the 2001 and 2002 vibrocoring surveys. Figure 4-7 shows the locations of the 189 vibrocores from the surveys. Maximum penetration of the vibrocoring equipment used during the surveys was six metres below the sea bed. Therefore this was also maximum sediment recovery of the vibrocore samples.

As is outlined in Chapter 4, the lithostratigraphic survey was undertaken to aid with the interpretation of the seismo-stratigraphic data. Therefore the results of the lithostratigraphic data are presented utilising the sequence stratigraphic model of seismic facies units presented in the previous section (i.e. seismic facies units A to I). Unfortunately not all seismic facies units identified were sampled during the vibrocore surveys, due to the limited depth below the seabed that samples could be retrieved from and the lack of penetration and recovery of the vibrocorer. There is, therefore, no lithostratigraphic or sedimentological data available for seismic facies units SFU-B, SFU-D, SFU-E and SFU-F.

To obtain quantitative data on the composition of the seismic facies units sampled during the vibrocoring surveys, some 747 subsamples were taken from the cores for grain-size analysis. Table 5.2 displays a summary of the main lithostratigraphical characteristics and sedimentological data for the seismic facies units. Textural classifications of sediment type are based on Folk (1954). The data from all 747 samples analysed are presented in Appendix 2.

SFU-A

SFU-A was sampled in 12 of the vibrocores (VC) in both Licence Application Areas 474 and 475 (VC474-2 , -10, -12, -21, -30, -45, -46 and -59; VC475-45, -47, -54 and -60: Figure 4-7). Figure 5-20 displays four photographs of example vibrocores containing sampled SFU-A. The unit is distinct in its characteristics from the units overlying it. It is generally grey to dark grey in colour occasionally becoming lighter grey/green towards the upper boundary (e.g. VC474-2: Figure 5-20). All cores sampling SFU-A show the upper boundary of the unit to have an abrupt and sharp contact with the seismic facies units that overly it (Figure 5-20).

The suites of sediments that this unit is comprised of are predominantly muddy sand and sand in their textural classification (Folk, 1954), with mean grain-sizes from 2.5 – 4.5□. The grain-size statistics of the sediments in SFU-A are poorly to very poorly sorted, negatively to very negatively skewed, with leptokurtic to very leptokurtic values (Table 5.2)

SFU-C

SFU-C was sampled in four vibrocores (VC475-15, -37, -51 and –59; Figure 5-21). This is because the upper boundary of the unit is beyond the penetration depth of the vibrocoring equipment. Figure 5- 22 shows vibrocore 475-59 in relation to a section of seismic line 69 (the closest line to the vibrocore sample site). This shows that the vibrocore sampled seismic facies unit SFU-C. The upper contact of SFU-C with the overlying SFU-G is sharp in all sample cores (Figure 5-21), although the nature of the basal contact of SFU-C is unknown as it was not sampled. The unit is light grey in colour and is predominantly composed of slightly gravely sands with occasional bands of muddy sandy gravel

Grain-size statistics for the suite of sediments represented in SFU-C have mean sizes from –1.8 to 3.6□. They range from very well to very poorly sorted, symmetrical to very fine skewness, and mesokurtic to very leptokurtic (Table 5.2).

SFU-G

SFU-G is mainly comprised of interbedded sequences of sandy gravels and gravely sands, with occasional muddy sandy gravel and gravel. The unit displays a range of colours. Many samples are dark orange-brown or orange-brown at the top of the unit and grade into brown, grey brown and light grey brown further down. The range of colours displayed by the unit displays is shown in the Figure 5-23. Also shown is the typically interbedded nature of the sandy gravels, gravely sands and sands, which make up this coarse grained facies unit. The gravel-sized fraction of samples examined from this unit is almost entirely made up of flint, which typically ranges in shape from sub angular to sub rounded. On the whole, the unit is barren of shell material and what shell material that is found is made up of fragments of nummulites probably derived from the local Tertiary bedrock.

418 samples of SFU-G were analysed for their grain-size characteristics. This large sample population shows a range of mean size values from -3.8 to 2.9 □, an almost full range of sorting from well sorted to extremely poorly sorted values, coarse to very fine skewness values and a full range of kurtosis values from very platykurtic to very leptokutic.

SFU-H

Vibrocore VC474-3 contains the only sample of SFU-H retrieved during the vibrocoring surveys. Figure 5-24 is a photograph of VC474-3 that shows the sediments recovered from SFU-H and the overlying SFU-I. Some 1.45 m of the unit is sampled it is dark grey in colour and shows no structure. Based on the single sample the unit can be described as very fine sand. The suite of sediments sampled from SFU-H has mean grain-size values in the range of -1.8 to 2.2 □. The grain-size statistics calculated for the samples show that they vary from moderately well to well sorted, fine to very finely skewed, and possess typically very leptokurtic values.

SFU-I

The uppermost seismic facies unit in the sequence examined, SFU-I, is composed of a mixture of muddy sandy gravels, sandy gravels and gravely sands. It is present in nearly all vibrocore samples in the boreholes the unit ranges in thickness from a few centimetres to more that 3.5 metres (VC474-58; Figure 5-25). The unit displays a light orange colouration in its upper portion although, where thicknesses of the unit are greater (e.g. VC 474-58), the unit becomes grey with depth. Figure 5-26 shows vibrocore 474-3 in relation to a section of seismic line 32 (the closest line to the vibrocore sample site). Examination of the clasts in the unit show that the gravel fraction is made up almost exclusively of flint which ranges from angular to subrounded in shape. SFU-I is the only unit in the study area to contain any significant amount of shell material which reaches a maximum of 44% of the total grain size measurements. Mostly, the shell material is heavily comminuted and very few complete *in situ* shells were found. Figure 5-27 shows the distribution of shell material as a percentage of the total sediment in the tops of the vibrocore samples in the PhD study area.

5.4 Transects A-A and B-B

It was impractical to sample all the vibrocores collected during this study, so two transects of vibrocores were selected for more detailed study using biostratigraphical and geochemical techniques. The transects were selected following initial analysis of the seismic data because they crossed the major palaeovalleys identified in the study area, and because they provided the best opportunity to assist in the analysis of the seismic data.

5.4.1 Transect A - A

Transect A-A (shown on Figure 4-7) is an east-west transect located in the western part of License Application Area EEC474. It crosses the elongate N-S feature that is incised into sequence boundary surface LQ_b (see discussion above). Figure 5-28 is a cross-section of the transect and demonstrates the depth to which vibrocore samples penetrate beneath the surface, their lithological characteristics and the seismic facies unit into which they penetrate.

The vibrocore samples in Transect A-A (Figure 5-28) demonstrates that seismic facies units SFU-G is composed of a mixture of interbedded coarse-grained deposits (gravelly sands, gS, and sands and gravels, SG) that do not appear to display any pattern either laterally or with depth. VC474-3 is the only unit to have sampled SFU-H and shows the unit to be composed of fine muddy sands (mS). The uppermost unit on Transect A-A, SFU-I, is also shown to be made up of coarse-grained sand and gravel laterally extensive deposits that cap the underlying units across the transect.

5.4.2 Transect B - B

Transect B-B (Figure 4-7) crosses the major incision in surface LQ_b to the east of Licence Application Area EEC474. Figure 5-29 is a cross section of the transect showing the depth to the upper bounding reflector of SFU-A (LQ_b) along the transect and the lithological characteristics of the vibrocore samples extracted from the overlying facies units. The vibrocore samples recovered along Transect B-B show very poor sediment recovery of only *circa* 2 metres, apart from VC's 474-50A and 49A that have recoveries of 4 and 6 metres respectively.

The sediments sampled from seismic facies units SFU-G show the unit to consist of interbedded fine (muddy sand, VC474-26A) to coarse (sandy gravel, VC474-54) deposits that, in a similar manner to the unit in Transect A-A, show a great degree of lateral and vertical variation between cores. The deposits of SFU-I are again shown to be coarse-grained (gravely sand, sG – gravely sand, gS) and capping the underlying units across the transect.

5.5 Biostratigraphical Data

Sediment samples from all facies units (138 in total) in selected vibrocores from transects A-A and B-B were analysed for their diatom content. Unfortunately the majority of samples were barren of diatoms, whilst those that did contain them were only found in low frequencies. Frequently however, samples also contained other two other biostratigraphical indicators, (fragments of) sponge spicules and the chitinous test linings of juvenile calcareous foraminifera (Dr. Jeremy Lloyd *pers. comm.*) that are potentially of use in indicating deposition under marine conditions. Table 5-3 displays a list of all the samples that contained diatoms and other micropalaeontological proxies (all samples examined are listed in Appendix 3). Also displayed are columns indicating whether sponge spicules and foraminifera were present in the sample and in what type of abundances. Indicators of abundance in the Table 5-3 follow the following scheme:

- R = Rare < 5 individuals recognised
- C = Common 5-20 individuals recognised
- A = Abundant >20 individuals recognised

SFU-A

Samples of SFU-A analysed under the microscope were revealed to be barren of microfossils.

SFU-G

Microfossils were occasionally identified in samples from SFU-G. These comprised rare counts of degraded siliceous sponge spicules. No evidence of diatoms or foraminifera was present. The occurrence of sponge spicules might be indicative of a marine origin for the unit. Alternatively the robust nature of sponge spicules may infer that they are allochthonous and that they were derived from older marine sediments.

SFU-H

Vibrocore VC474-3 was the only vibrocore to retrieve sediments from seismic facies SFU-H. Microscope analysis of samples from the unit revealed common counts of foraminiferal test linings and rare counts of sponge spicules. These data, although limited, do assist in the palaeoenvironmental reconstruction of the unit's depositional history by indicating that it probably accumulated in a marine setting. Because only foraminiferal test linings were recorded, it is not possible to provide a more definitive palaeoenvironmental interpretation at this stage.

SFU-I

In both transects, diatoms and other biostratigraphical data were only present in any quantity in samples taken from the uppermost section of seismic facies unit SFU-I. Frequently occuring species of diatoms were *Melosira westii. Paralia sulcata* and *Hyalodiscus scoticus*. Also found more rarely were *Diploneis crabro, D. didyma,* and *Actinoptychus senarius*. All the species identified are found in polyhalobous environments according to the classification of Hustedt (1957) and are planktonic forms common in the marine offshore waters of the United Kingdom in the present day (Hendey, 1964).

5.6 Multi-Element Geochemistry

The analyses of multi-element geochemistry was undertaken on 170 samples from the seismic facies units sampled in vibrocore Transects 1 and 2. The concentrations of the 14 rare earth elements analysed from the samples are combined to give each sample its own multi-element geochemical signature which can then be compared with other samples to see if there are any major similarities or differences between them.

To examine the similarities or differences between the multi-element geochemical signatures of the samples taken from the four seismic facies units (SFU-A, -G, -H and –I), the data were analysed using the statistical technique of Detrended Correspondance Analysis (DCA; Hill, 1979) to see whether samples grouped according seismic facies unit. Figure 5-30 displays the DCA bi-plot of the multi-element geochemistry of all the samples analysed, with samples from each of the four seismic facies units displayed as a different colour. The full table of the concentrations of the 14 elements analysed in each sample is shown in Appendix 4).

SFU-A

The multi-element chemical signature of samples from seismic facies unit SFU-A (green circles on Figure 5-30) display a geochemically distinct grouping in the upper right of the DCA biplot, with very little overlapping with the other seismic facies units analysed. This grouping indicates that the deposits of SFU-A show significantly different mutli-element chemical concentrations to those of the seismic facies units overlying it. This suggests suggests a different provenance for this unit compared with the rest of the data, and that the unit was either deposited under different conditions to those overlying it, or that the sediments of SFU-A were derived from a different source material to those of the units above. It is likely that the sediments samples of the other seismic facies units that map close to, or overlap with, the clustering of SFU-A on the biplot (Figure 5-30), may well have been derived from SFU-A during their formation.

SFU-G

The geochemical signature of samples from seismic facies unit SFU-G (black circles on Figure 5-30) show a broad scatter on the DCA biplot and indicate a variety of source material with a range of geochemical signatures.. This variability is similar to the grain-size data that were analysed for this seismic facies unit, which record a similar degree of variability.

SFU-H

The five samples analysed from this unit show a very close grouping on the DCA biplot (red circles on Figure 5-30). This suggests that the multi-element geochemical signature in each of the samples is very similar. However, this situation is not surprising, and is readily explained because the five samples of the unit analysed were taken from the same vibrocore, VC474-3, which is the only vibrocore to sample SFU-H. The samples of the unit also plot within the overlapping boundaries of the sample groupings of both SFU-G and SHU-I. Therefore it is not possible to suggest on multi-element chemistry alone, that the deposits of SFU-H are of a different provenance to these units. They may in fact have been derived from the same source material.

SFU-I

Samples of SFU-I (light blue circles on Figure 5-30) also show a widely scattered distribution on the DCA biplot. Again this indicates that this seismic facies unit lacks a distinct geochemical signature. Both SFU-I and SFU-G are predominantly coarse grained deposits (see above). Therefore, it seems likely that analysis of the <150 μm fraction of the sediment may not be reflecting the real multi-elemental concentrations of the deposit.

Chapter 6: Interpretation

6.1. Introduction

In this chapter the results presented in Chapter 5 are examined with the aim of providing an interpretation as to the nature and origin of the sequence boundary surfaces and seismic facies units identified in the PhD study area. Primarily the interpretation of the seismic facies units is based on the assumption that the bounding reflectors that define the external geometry of the seismic facies units represent unconformities in the fill stratigraphy. As a result, it is considered that the internal seismic signatures of each facies unit can be used to infer discrete and distinct lithological characteristics and depositional environments (Mitchum *et al.*, 1977b; Berryhill, 1986; Evans *et al.*, 1995). Interpretations of these data are supported by the results of the lithostratigraphical and biostratigraphical analyses (where available) of samples from vibrocores that penetrate the seismic facies units. To aid with the interpretation, the sequences are also compared to analogous studies on other continental shelves. This provides important insights to the processes that have combined to produce the features identified in the PhD study area.

Chapter 5 demonstrated the presence of three sequence boundary surfaces and nine discrete seismic facies units within the study area (Figure 6-1). The three bounding surfaces are interpreted to represent unconformities in the stratigraphy that are defined by either erosion or non-deposition.

This chapter comprises the following:

- An interpretation of the current sea bed topography based on bathymetric data
- Discussion of the Tertiary rockhead surface (LQ_b) which forms the lowest seismic facies in the study area
- Identification within the sequence of two other chronostratigraphically significant bounding surfaces (R_I and H_b)
- Interpretation of the intervening seismic facies units and examination of their relationship to changes in base level during the Quaternary
- A comparison of the features and deposits examined in the present study with analogous sequences studied on other continental shelf settings
- A hypothetical chronological model for the incision and infilling of the major features and facies units identified.

6.2 Sea Bed Topography

The sea bed topography of the study area is dominated by a number of deep, linear hollows that trend approximately south west to north east. They are incised into the broad, planated, sea bed surface lying at 42 to 45m below present sea-level (bpsl). The area lies to the south of the Northern Palaeovalley of the former Channel River, a major incised palaeoriver valley (Hamblin *et al.*, 1992). Sea bed planation, largely marine in origin (Curry, 1989), heavily disguises the old morphology of this river system and its tributaries, which is cut into a lower, earlier surface that lies some 51 to 53 m bpsl (Evans, 2002). This plain is analogous to those identified on the Atlantic continental shelf of North America and demonstrates that there is frequently no consistent relationship between palaeoriver systems and the modern bathymetry of the shelf due to postdepositional reworking during marine transgression (Suter, 1986; Knebel and Circé, 1988; Colman *et al.*, 1988, 1990; Foyle and Oertel, 1992, 1997).

The elongate depression in the west of the study area (Figure 5-2) is closely aligned to the present dominant tidal flow directions (Pingree and Griffiths, 1979; Johnson *et al.*, 1982; Figure 6-2). It appears to also be related to another depression around 556000mE/ 63000mN that reaches depths of up to -51 m bspl. The southwest to northeast alignment of the depression that trends along the northwest edge of the study area from 52000N, 550000E to 62000N, 568000E (Figure 5-2), and which may continue in an easterly direction as the sea floor shallows, is also aligned with the dominant tidal flow direction in this area. Therefore, it is suggested that the origin of these sea bed features reflects the combination of the (southwest to northeast) Holocene transgression of the shelf and differential sea bed erosion caused by tidal scour (Pingree and Griffiths, 1979; Johnson *et al.*, 1982). This interpretation is similar to that proposed by EMU Environmental Ltd. (2002a, 2002b) who have discussed these sea bed depressions and others that occur immediately adjacent to the PhD study area.

6.3 Sequence Stratigraphy

The following section combines the seismo-stratigraphic framework established in Chapter 5 with sedimentological, biostratigraphical and geochemical data to interpret the origins of the seismic facies units and sequence boundaries identified. It concludes by presenting these data as a sequence stratigraphic model for the evolution of the seabed geology in the study area.

The stratigraphic record preserved in the PhD study area can be divided, at various temporal and spatial scales, into sequences that are the fundamental units for palaeoenvironmental interpretation (Mitchum *et al.,* 1977a, 1977b; Mitchum and Vail, 1977; Vail, 1987; Thorne and Swift, 1991; Foyle and Oertel, 1992). Surfaces of geochronological significance, which include the *sequence boundary* and the *ravinement surface* (outlined in Chapter 4), form in response to changes in base level related to variation in glacio-eustacy and/or basin tectonics. In the present study the *sequence boundary* and *ravinement surface* have been identified and are therefore considered to be the important geochronological surfaces for the development of a chronostratigraphic framework for the sequence investigated in the PhD study area (section 6.4).

6.3.1 Seismic Sequence I

Depositional Sequence I

SFU-A is present over the entire study area and either occurs beneath the overlying seismic facies units or outcrops at the seabed. On the seismic records the internal reflectors of the unit demonstrate a poorly defined synclinal form, dipping at very low angles eastwards in the west of the study area (e.g. Figure 5-9), westwards in the east (e.g. Figure 5-10), and horizontal in the central area (e.g. Figure 5-7). The continuous and parallel nature of the internal seismic signature of the unit is interpreted as representing almost planar zones of acoustic impedance, similar to, if not the same as, the basal seismic facies unit identified by Bellamy (1994) offshore of the River Owers, southern England. Bellamy (1994) suggested that this internal seismic signature represented bedding disconformities (possibly the Lutetian calcareous bands; Smith and Curry, 1975) in a largely homogeneous, stratified, dipping and slightly contorted sedimentary sequence. He considered that the sequence represented the Tertiary (Palaeogene sand and clay) bedrock of the Hampshire-Dieppe Basin, sampled by Lawson and Hamblin (1989), and shown to underlie the area of his study.

The outcropping Hampshire-Dieppe Basin and its predominantly Bartonian Tertiary Geology (Curry *et al.*, 1978; Hamblin *et al.*, 1992; see Figure 3-5) continues offshore in a south-easterly direction (Hamblin *et al.*, 1992) to the present study area. Therefore, SFU-A is interpreted as an equivalent to Bellamy's (1994) lowermost seismic facies unit i.e. Tertiary bedrock. This interpretation is given further credence by the analysis of sediment samples of the sequence that were obtained during the vibrocoring surveys. These reveal similar characteristics to typical exposures of Bartonian sediments from the Hampshire-Dieppe Basin that include dark grey/green sandy clays, sandy silts and silty sands (Edwards and Freshney, 1987). However that lack of biostratigraphical evidence (such as nummulites, in particular *Nummulites rectus*, Hamblin *et al.*, 1992) in the vibrocore samples of the unit, which could have added further weight to the interpretation of the sequence, is noted. Samples of SFU-A analysed for their multi-element geochemistry were shown to cluster together in Figure 5-30. This clustering demonstrates that SFU-A has a distinctive multi-element geochemical signature in relation to other units sampled.

6.3.2 Seismic Sequence II

Sequence Boundary Surface LQ_b (Rockhead)

From the above discussion, it is clear that the upper sequence boundary of SFU-A, surface LQ_b, represents a major discontinuity interpreted to be the rockhead surface in the study area. It is identified as a major unconformity and can be used to determine the gross geometry of the sedimentary succession belonging to the overlying seismic facies units. Several authors have suggested that the submerged palaeovalleys of the eastern English Channel were eroded during the Late Quaternary (Smith, 1985, 1989; Hamblin and Harrison, 1989; Bellamy, 1994) and therefore that the rockhead surface separates Tertiary strata from Late Quaternary sediments. Gibbard (1988, 1995) and Lericolais *et al.* (1996, 2003), however, hypothesised that a fluvial drainage pattern probably evolved in the English Channel much earlier, at the end of Oligocene and the beginning of Miocene times. This hypothesis is given further support by the fact that Miocene and Pliocene strata do not occur in southern England or offshore in the Channel (Hamblin *et al.*, 1992). This implies that these strata were not deposited because the region was potentially subaerially exposed and was subject to fluvial erosion during this time. Nevertheless, whilst mindful of the observations of Gibbard (1988, 1995) and Lericolais *et al.* (1996, 2003), it seems probable that the rockhead surface LQ_b is of Quaternary, as oppose to Oligocene / Miocene age.

The rockhead relief in the PhD study area is indicative of significant erosion and incision. Similarly eroded and incised rockhead relief has been detected and studied on the inner shelf the Atlantic eastern seaboard and the south-western Louisiana shelf, northern Gulf of Mexico, USA (e.g. Suter, 1986; Knebel and Circé, 1988; Colman *et al.*, 1988, 1990; Foyle and Oertel, 1992, 1997; Swift *et al.*, 2003; Anderson *et al.*, 2004). Indeed, close similarities exist between the Quaternary geological systems of Delaware Bay, the Delmarva Peninsula, Chesapeake Bay and the Louisiana Shelf to the English Channel records. For example, these studies report seaward trending, linear depressions of up to 8 km wide, up to 35 m of relief (below sea bed), and irregular long profiles and depths of over 75 m below present sea-level (bpsl) cut into rockhead during periods of high-frequency Quaternary glacio-eustatic sea-level fluctuations (Knebel and Circé, 1987; Foyle and Oertel, 1997; Swift *et al.*, 2003). These features are interpreted as fluvially-incised valleys, formed during lowstands, that have been significantly modified and infilled during subsequent marine transgression as the drainage basins evolved into estuarine embayments and were finally submerged by transgressive marine ravinement.

In the present study, the depressions detected in the rockhead surface are of similar magnitude and depth (particularly the lower bounding surface of SFU-C) to those identified at similar latitudes on the Atlantic Shelf, USA. Therefore it is possible, if not probable, that the features share a similar origin and are related to incision associated with lowered base levels during Late Quaternary. Furthermore, the depressions in the rockhead are locally infilled and overlain with sediments (seismic facies units B-I, discussed below) that display a range of seismic characteristics indicative of a variety of sediment types. This situation also occurs on the US Atlantic shelf, most notably beneath Delaware Bay, where a system of palaeovalleys is buried by a transgressive sequence of fluvial, estuarine and shallow marine sediments (Knebel and Circé, 1987; Knebel *et al.*, 1988).

The gross geometry of the sedimentary succession infilling and overlying the rockhead surface of the study area is shown in Figure 5-4. This shows the depth below chart datum to rockhead of a diachronous surface onto which the sediments of seismic facies units B-I have accumulated. Two main morphological features in the rockhead surface are an elongate feature running north-south through the centre of the study area, and a large area of incision in the northeast of the study area. These features are interpreted to be a complex of palaeovalleys similar to those found elsewhere on the eastern English Channel sea floor (Dingwall, 1975; Bellamy, 1994, 1995; Velegrakis *et al.*. 1999). The depressions associated with the lower bounding surfaces of SFU-D, -E and –F are also inferred to represent these type of features. Between the palaeovalleys, the adjacent rockhead surface displays a relatively flat low relief at or just beneath the sea bed surface. Foyle and Oertel (1992), interpret a similar relief beneath the Delmarva Peninsula, USA, as representing former interfluves between the main palaeoriver channels, and a similar interpretation is proposed here.

Nummedal and Swift (1987) and Van Wagoner *et al.*, (1992) identify a Type 1 Sequence Boundary (see Chapter 2) that is, they suggest, potentially of inter-regional to global importance, resulting from the exposure and subsequent reworking of the continental shelf during sea level lowstands (Thorne and Swift, 1991; Foyle and Oertel, 1992). On seismic records, a Type 1 Sequence Boundary typically shows major local and regional relief and is overlain by a sequence of onlapping fluvial, estuarine and marine deposits deposited by subsequent marine transgressions. This onlapping relationship is useful in the recognition of the boundary on seismic profiles, such as those reported from the thesis study area. The updip, inner shelf, portion of the boundary is represented by incised fluvial valleys and their associated interfluves, with the former potentially exceeding depths of c. 100 m and widths of >10 km (Van Wagoner *et al.*, 1990). The relief of the boundary surface is dependent on the scale and number of the incised valleys that are, in turn, controlled by a number of factors including the time available for fluvial incision to occur, drainage basin size, slope of the subaerial landscape, and the magnitude of sea-level oscillation (Van Wagoner *et al.*, 1990; Foyle and Oertel, 1992). From a chronological perspective, a Type 1 Sequence Boundary is of particular significance since on most occasions it represents an extended period of cold stage subaerial exposure and fluvial incision of the shelf prior to a major marine transgressive event (Foyle and Oertel, 1992). Based on comparisons with this work, Surface LQ_b is also considered to be a Type 1 Sequence Boundary.

Depositional Sequence II

Depositional Sequence II comprises the seismic facies units that are bounded by the surfaces LQ_b and R_I respectively (Figure 6-1). These facies units are SFU-B, -C, -D, -E and –F. Based on their external geometries, described below, these facies units are interpreted as a complex of five palaeovalleys that are now infilled with unconsolidated sediments.

Seismic facies units SFU-B and SFU-C are the most extensive (widest and most deeply incised) of the infilling units in the PhD study area. Internally, their seismic signature is chaotic and mounded with occasional sub-parallel moderate-low amplitude aggrading reflectors. Chaotic internal configuration is usually associated with lithofacies of coarse grained sediments (Velegrakis *et al.*, 1999) and the occasional aggrading reflectors observed might suggest deposition associated with marine transgression under rising relative sea-level and deepening water depths. SFU-D, -E, and –F also display similar geometries and internal characteristics, although not all have exactly the same internal signatures (see Table 5-1). For the purposes of this interpretation, all units in Depositional Sequence II are ascribed the same mode of origin, namely coarse grained (sand and gravel) sediments deposited under marine transgression.

The seismic facies units of depositional sequence II show distinct similarities both in their internal signatures and external geometries to those studied by Anderson *et al.* (1996). These authors investigated the infilled palaeovalley systems of the Colorado and Brazos Rivers on the inner shelf of the Gulf of Mexico. The system of palaeovalleys which they considered experienced high rates of sediment supply during sea level lowstands. They observed depths of incision of up to 40 m into the underlying substratum, with individual channel widths of up to 5 km. Anderson *et al.* (1996) considered the incisions to be of MOIS 2 (last glacial maximum) age, and believed that the valleys had been back-filled primarily with sands and gravels, and then abandoned, as sea level rose during the post-glacial transgression. Seismic profiles examined from the seismic facies within the fluvial valleys showed chaotic reflection patterns. Lithological correlation provided by platform borings extracted from the infilled palaeovalleys penetrated sands and gravels on the majority of occasions and provided independent support for this interpretation. A third analogous study was that undertaken by Knebel *et al.* (1988) of a sequence that infills a system of incised-valleys in Delaware Bay, USA, and that accumulated under the rise in relative sea-level across the Late Wisconsinan / Holocene transition. They suggested that thin, coarse-grained, fluvial deposits accumulated initially within the main channels of the former drainage system as base level was elevated by rising sea level. As sea level rose further, estuarine conditions developed, and sands, silts and clays were deposited beneath the non-tidal estuarine depocenter as it migrated upwards through the embayment (Knebel *et al.*, 1988).

Studies on the Atlantic and Louisiana shelves of the USA have found only a small proportion of incised valley fills to contain any deposits of alluvial origin and that where they do occur, they comprise a thin spread of coarse sediments in the thalweg of the channels (Belknap and Kraft, 1985). Indeed, in these areas, most sediment cores that have penetrated shelf valleys pass through thick sequences of estuarine deposits overlain by lagoonal silts and clays (Oertel *et al.*, 1989). The infilling deposits are also largely composed of sediments of synchronous age to the early portion of the subsequent interglacial period. For example, as with the Delaware Bay example referred to above, Suter (1986) noted that the Quaternary incised valleys on the Louisiana Shelf were Wisconsinan in age and were infilled by sediments from estuarine systems predominantly during the Late Wisconsinan / Early Holocene transgression.

Within the present study area, seismic facies units SFU-B, -C, -D, -E and -F are considered to represent similar type infilling sedimentary sequences to those described by Anderson *et al.* (1995). However, the thin lag of coarse grained fluvial deposits that are associated with the base of many of the infilling deposits is not recognised in the PhD study area. This may be due to the removal of these deposits prior to the infilling of the channel or simply that they are not detected by the seismic survey because the signal has been attenuated by the overlying coarse grained deposits (SFU-G).

In general, the sequences of this unit (SGU-B, -C, -D, -E and -F) display a lack of marked internal reflectors and impedance contrasts and show no overlapping channels that might be expected in a complex channel fill associated with palaeovalley infilling (Berryhill, 1986). These seismic facies have, in general, an acoustically muted typical, according to Lericolais *et al.* (2003), of aggradational facies and are very similar to seismic facies units identified in a partly infilled valley of the Channel River system at the confluence of the Northern Palaeovalley and the submerged palaeovalley of the Somme (Lericolais *et al.*, 2003).

In summary, the interpretation of seismic facies units SFU-B, -C, -D, -E and –F, based on their seismic characteristics and comparisons with analogues from other similar continental shelf seismic records, is that they were derived in similar if not the same manner. They comprise a sequence of medium-coarse grained (sand and gravely sand) deposits laid down as an aggrading transgressive facies that onlapped the lowstand eroded bedrock. The facies accumulated as the palaeovalleys were transformed from a fluvial system into an open shoreface type environment with energy levels sufficiently high to transport and deposit the coarse grained deposits inferred to be the main infilling sedimentary sequence.

Unfortunately very few vibrocore or grab samples of the sediments infilling the palaeovalleys studied in this PhD exist, with material only obtained from the uppermost part of SFU-C. Though limited in number, these cores do, however, corroborate the interpretation of the seismic data by showing that the upper section of the unit is composed of medium-coarse grained deposits (slightly gravely sand and muddy sandy gravel).

6.3.3 Seismic Sequence III

Ravinement surface R_1

Above the sequence boundary of surface LQ_b, and separated from it by the deposits in the incised valley features described above, is a ravinement (transgressive/regressive) surface R_1. Ravinement surfaces are cut by erosional shoreface retreat or advancement through storm current action that extends from the surf zone down to the lower shoreface (Stamp, 1921; Swift, 1968; Swift and Thorne, 1991; Swift *et al.*, 1991) and the ravinement process may result in the loss or reworking of 10 m or more of the stratigraphic section (Swift, 1968; Belknap and Kraft, 1985; Thorne and Swift, 1991). These surfaces are unconformities that form as a result of either the landward retreating shoreface during transgression, the seaward advancement of the shoreface during regression (driven by excess sediment supply), or possibly a combination of both. The vertical thickness of deposits between ravinement surfaces and the underlying sequence boundary is very important in the preservation potential of the valley fill deposits that occupy the space between the two surfaces (Fischer, 1961; Belknap and Kraft, 1985; Foyle and Oertel, 1992). Where thicknesses between the two are large, such as over the infilled channel thalwegs that have incised in the sequence boundary, it is likely that more complete stratigraphic records of the transgressive coastal depositional environments will be preserved. However, on the interfluves between channels, Foyle and Oertel (1992) suggest that the depositional record would tend to be incomplete if not absent. This is certainly the case for the Quaternary deposits of the North Carolina Inner shelf studied by Hine and Snyder (1985).

Seismic profiles show that surface R_I frequently cuts across the lower LQ_b surface, so that between the major infilled valleys, R_I is contiguous with the rockhead (LQ_b). This indicates the removal of interfluve deposits and probable downcutting of the rockhead surface during the creation of R_I. This ravinement surface is overlain by SFU-G (see below), a coarse-grained unit that developed under high energy (regressive) shoreface conditions. From this evidence it is hypothesised that an initial ravinement surface formed during the marine transgression associated with a period of climate amelioration and rising global sea level. However, once formed, this surface was significantly modified by shoreface processes during the subsequent regression and lowstand associated with the next cold stage and associated fall in global sea level. A regression of this type, caused by a fall in eustatic sea level, is referred to as forced regression. One of the main criteria for the recognition of forced regressions in stratigraphy is the abrupt occurrence of coarser sediments (such as those associated with facies unit SFU-G, see below) that constitute a regressive (or lowstand) prograding wedge and are typically detached from the previous highstand shoreline (Posamentier *et al.*, 1992).

A similar scenario is reported by Swift *et al.* (2003) who described an occluded boundary that formed as a result of the collapse of a regressive ravinement onto an older transgressive ravinement. They noted that this type of event has the potential to destroy the entire intervening transgressive shelf system. Swift *et al.* (2003) argue that this type of sequence occurs some distance from the previous highstand shoreline, such that the regressive shoreface-shelf system has turned into a forced-regressive system capable of eroding its substrate. This was potentially the case in the PhD study area, given its location in the centre of the eastern English Channel, some distance away from the present highstand coastline.

Depositional Sequence III

Seismic facies unit SFU-G is the only depositional unit in seismic sequence III. On the summary schematic diagram of the seismo-stratigraphic sequence (Figure 6-1), this widespread unit is seen to cap SFU-B, -C, -D, -E and -F, and attains its greatest thicknesses above the palaeovalley fills. This indicates that surface R_I eroded more deeply into the unconsolidated sediments of the incised valley fills than the surrounding, more consolidated, bedrock surfaces, with the result that a greater thickness of SFU-G was deposited in the depressions created.

As stated above, R_I was immediately succeeded by the deposition of SFU-G. Thus, the sediments of this unit accumulated in high energy marine shoreface and beach environments under regressive conditions. This interpretation is supported by the many samples taken from the vibrocores extracted from this facies unit, which are predominantly (flint-rich) sandy gravel or gravely sand suggesting deposition in high energy environments. Their interbedded nature demonstrates significant variations in grain size both up-core and between cores, and this may be the product of deposition in multiple storm events. Micropalaeontological analyses of samples from the deposits of Depositional Sequence III show that diatoms are absent and only rare counts of degraded sponge spicules are present. This suggests that the sediments in the unit may have been subjected to secondary post-depositional dissolution of biogenic silica. However, many of the vibrocore samples also revealed a shelly component (albeit heavily reworked) below the deep orange stained upper portions of the unit.

In summary, Depositional Sequence III is interpreted as the product of the seaward prograding shoreface that was produced during the sea level fall and associated regression that produced the ravinement surface R_1. The variation of textural classifications present in the unit can be explained by the range in energy levels of the events (e.g. storms) that would be expected at the shoreface during the formation of the unit.

The interpretation of Depositional Sequence III assumes an origin under regressive conditions. However, it is important to note that many authors consider that the preservation of sediments under regressive regimes may not be high. For example, Field and Trincardi (1991) and Trincardi and Field (1991) suggest that regressive deposits are least likely to be preserved on broad low gradient (coastal plain) shelves, such as the eastern English Channel, particularly in a mid shelf location. This is because regressive sediments in these locations can be obliterated by subsequent transgression (Field and Trincardi, 1991). However, these authors also note that in certain cases, such as that proposed above for Depositional Sequence III, regressive coastal deposits can be preserved where their location corresponds to a deepening (relative to adjacent areas) of the underlying shelf surface such as infilled palaeovalleys. It is suggested that this is the case in the PhD study area for the deposits of depositional sequence III.

The coarse nature of the sediments in the sequence is analogous to those described by Swift *et al.* (2003) in their analysis of the dynamics of mesoscale stratigraphy in the Quaternary sediments of the Virginia Coast, USA. These authors suggest that over a sucession of transport events (such as storms of variable intensity), it is likely that the coarsest particles entrained will be retained in the base of the bed as the intensity of the event decreases. If the next event was less intense, only the upper part of the bed would be re-entrained. As a result, over several events of varying intensity, coarser sediment will be deposited preferentially at the site, with finer sediment bypassing the site and transported downstream. In the current study, this may have resulted in the preferential evacuation of fine-grained sediment to the shelf edge at the Western Approaches (see discussion in Chapter 7).

Depositional Sequence III sediments accumulated during a global eustatic sea level fall that occurred during a transitional between interglacial and glacial conditions. Following their deposition, the continental shelf in the eastern English Channel would have been subaerially exposed, and the upper surface of the sequence would have been reworked by subaerial processes. There is widespread evidence that during such intervals, significant quantities of wind borne loess were deposited on the terrestrial areas of southern Britain and northern France bordering the English Channel (Catt, 1986; Paepe and Sommé, 1970; Gallet *et al.*, 1998; Antoine *et al.*, 2000; Regnauld *et al.*, 2003). Mapped thicknesses of the deposit range from a few centimetres to several metres in these areas (Gallet *et al.*. 1998, Antoine *et al.*, 2003). Given their widespread deposition, it is probable that a significant thickness of loess was deposited on the upper surface of depositional sequence III in the PhD study area. However, examination of the fine sediment fraction of the sediment samples taken from the uppermost deposits of the sequence by laser scanning grain-size analysis failed to reveal the characteristic loessic grain-size fraction modal peaks that would have been expected. This suggests that any loessic sediments that accumulated on this surface were reworked during the production of overlying seismic facies units. Interestingly, the upper surface of depositional sequence III displays a distinctive deep orange colouration. This colouration most probably reflects a subaerially weathered surface (or palaeosol), and indeed similar palaeosols are widespread in terrestrial sequences on the coast around the eastern English Channel in southern England and northern France.

6.3.4 Seismic Sequence IV

Sequence Boundary Surface H_b

The Sequence Boundary Surface H_b is a prominent reflector lying above the surface R_1. It lies between 0 and 15 m above the R_1 surface and between 0 and 10 m below sea bed. The surface H_b is interpreted to have initially formed on the upper surface of depositional sequence III during the lowstand cold period associated with the palaeosol formation referred to above. Erosive processes modified this surface during the subsequent transgression as the shoreface retreated once more to a highstand position. Apart from the incision of SFU-H, the sequence boundary surface H_b shows a marked absence of any fluvial incisions in the PhD study area. This suggests that during the previous cold lowstand period, the area was essentially a subaerially exposed topographic high. If correct, this interpretation suggests that during this interval the Channel River did not reoccupy its previous course through the study area, but must have created a new incision or reoccupied a previous course elsewhere.

Sequence Boundary Surface H_b is also interpreted to represent a Type 1 Sequence Boundary (Nummedal and Swift, 1987, Van Wagoner *et al.*, 1992; see above). Despite the contrasting representative morphologies of Sequence Boundary Surfaces LQ_b and H_b, it is suggested that the same type of cold stage subaerial exposure followed by marine transgression has occurred during the creation of the surfaces. Foyle and Oertel (1992) investigated a similar situation in the Quaternary sedimentary sequence beneath the southern Delmarva Peninsula, Virginia, USA, where a heavily incised undulating lower Type 1 Sequence Boundary recognised in the sequence is overlain by a later Type 1 Sequence Boundary displaying a relatively low relief. The sequence was interpreted by these authors to suggest that undulating morphology of the lower Type 1 Sequence Boundary was formed by fluvial incision whereas during the formation of the upper Type 1 Sequence Boundary the fluvial system had not reoccupied their study area and therefore no significant incision had occurred during the boundary formation.

Depositional Sequence IV

SFU-H

The lower bounding reflector of SFU-H represents an incision into the underlying depositional sequence. The incision is not on the same scale of the incisions related to the palaeovalleys of sequence bounding surface LQ_b, but none the less it appears to be a substantial incision eroding up to 10 m into the underlying sequence and up to 600-700 m in cross-section.

The sub-horizontal to dipping, parallel, evenly spaced reflectors that onlap the discordant boundary with Depositional Sequence III suggest that this unit is fine grained and was deposited in a relatively low energy environment. Similar reflector patterns were found by Velegrakis *et al.* (1999) in the palaeovalleys of the Solent, and by Foyle and Oertel (1997) on the inner shelf of the Atlantic Coast Virginia, USA. However neither of their studies retrieved sediments from these units to be able to corroborate this interpretation. In the present study, vibrocore 474-3 sampled the uppermost 2 m of SFU-H and samples from the unit show that it is composed of a mixture of fine sands and muddy fine sands. This confirms the interpretation that deposition of this facies was in a low energy environment. Micropalaeontological analysis of the sediments showed that the unit contains common foraminiferal test linings indicative of deposition in marine or estuarine conditions (Dr J. M. Lloyd, *pers. comm.* 2004)

Seismic facies unit SFU-H is therefore interpreted as having been deposited in a low energy transgressive estuarine tidal channel. Due to its limited spatial extent within the study area, further inferences as to the nature and origin of the incision and infilling deposit are not possible.

SFU-I

Seismic Facies Unit I occurs at the sea bed throughout nearly all of the study area. It forms the uppermost and almost certainly the youngest seismic unit present. The high amplitude basal reflector of the unit (where detectable on profiles) suggests an erosive (high energy) contact between the unit and underlying units. In general the deposit is relatively thin (0.5 to 1 m), but is locally thicker forming substantial lenses up to 5 m thick where it rests upon areas of deeper infilling (palaeovalleys). The unit is interpreted as forming a thin veneer over the majority a sea bed in the PhD study area.

SFU-I has the greatest number of core samples of any facies in the study area. Samples indicate that the unit consists of shelly, predominantly coarse grained deposits with a (predominantly flint) gravel-rich texture. Figures 6-3 and 5-27 show the mean grain-size and calcium carbonate fraction (shell content) of the deposit at the sea bed. Figure 6-3 demonstrates that the gravels generally occur in a matrix of sands and muddy sands with isolated pockets of gravely sands that are in general poorly sorted. Where it unconformably overlies Depositional Sequence I (SFU-A), the gravel supporting matrix of SFU-I contains predominantly muddy sands. Elsewhere, sand size material is dominant. It seems likely that the gravel, and much of the sand sized fraction, are derived from reworking of the underlying units, in particular SFU-G. The high mud content is possibly derived from the underlying fine-grained Tertiary rockhead. Mixing of gravels, sands and muds to form a veneer over the seabed implies high energy conditions, as does the comminuted nature of the molluscan fauna. The high percentage of comminuted shell fragments combined with the presence of foraminiferal test linings, sponge spicules and marine planktonic diatoms such as *Paralia sulcata*, *Hyalodiscus scoticus*, *Actinopychus senarius* and *Melosira westii* indicate that this unit was deposited in high energy marine conditions.

Depositional Sequence IV began formation as a product of the transgression that created sequence boundary surface H_b, with most sedimentation occurring as the shoreface moved across the study area in response to shoreline retreat. Many workers have suggested that the uppermost facies on shelf systems formed in this way as a coarse-grained discontinuous lag (Belderson and Stride, 1966; Emery, 1968; Swift *et al.*, 1991). This implies that the formerly subaerially exposed sediment surfaces of SFU-G and SFU-H were susceptible to and in disequilibrium with wave action and tidal currents. Moreover, any features that formed higher relief surfaces would have been subject to truncation by shoreface erosion, thereby producing a marine planation surface (e.g. Swift, 1968).

Seismic facies unit SFU-I is interpreted to be the same as the lag deposit identified in the eastern English Channel by Hamblin *et al.* (1992). These authors maintain that the deposit is present over the entire area (apart from small zones of outcropping bedrock) of the offshore eastern English Channel and that some local variations in grain-size, particularly the ratio of sand to gravel, are the result of post-depositional winnowing by high velocity bottom currents. Dickson and Lee (1973) and Hamblin *et al.* (1992) considered that the lag deposits in the English Channel are not presently mobile. This is most probably the case in the PhD study area, given the coarse-grained nature of the deposits and the absence of any obvious bedforms during analysis of the seismic and sample data.

Where thicknesses are great enough (>0.5 m), SFU-I appears to show a distinctive two tone colouration between the upper and lower section of the unit (e.g. VC474-47 and VC474-49A; Figure 5-25). The lowermost section tends to be grey or orange, depending on the colour of the underlying facies unit, whereas the upper section is predominantly light brown. Moreover, the uppermost section contains common and abundant intact diatom species and abundant sponge spicules and foraminiferal test linings, whereas the lowermost is almost devoid of diatoms and contains only low frequencies of sponge spicules and foraminiferal test linings. These two sections display the same type of grain-size statistics and, therefore, the difference between the segments is most readily explained as a function of post-depositional modification. In particular, the upper section of the unit lies at the sediment water interface and is therefore biologically active under the present sea bed current regime, whereas the lower section is anoxic and devoid of biological activity explaining the lack of microfossils.

6.4. Chronostratigraphy

In this section a hypothetical chronostratigraphic framework is proposed based on the sequence stratigraphic interpretation of the data obtained from the from the PhD study area. Given the lack of absolute dating control, a relative chronology is proposed based on the principle of stratigraphic superposition. Figure 6-4 shows the proposed chronostratigraphy of the sequence boundaries and infilling depositional sequences in the PhD study area, and the following sections provide the justification for why the sequence boundary surfaces and depositional sequences are attributed to particular ages.

6.4.1 Seismic Sequence I

Depositional Sequence I

Depositional Sequence I comprises the bedrock of probable Tertiary age. Assigning a specific age to this sequence is difficult because the sequence boundary surface LQ_b incises through many layers of the Tertiary strata. Moreover, it was noted above (and in Chapter 3) that no deposits of Oligocene age are known in the eastern English Channel. Therefore, the rockhead must be of at least Middle or late Eocene, Bartonian or Priabonian age (37-40 Ma).

6.4.2 Seismic Sequence II

Sequence Boundary Surface LQ_b

Identified within the sedimentary sequence in the PhD study area are two distinct Type 1 Sequence Boundaries (Nummedal and Swift, 1987) - LQ_b and H_b. that are separated by a ravinement surface, R_I. For the sequence boundaries to have formed (in particular LQ_b), it is suggested that glacio-eustatic sea-level must have fallen well below the altitude of the sequence (-70 m bpsl) on at least two separate occasions. Moreover, there must have been at least one highstand period, sufficient to have submerged the sequence, between these lowstands. Examination of the global eustatic sea- level curve (e.g. Siddal *et al.*, 2003; Figure 6-5) indicates that on a count back basis the minimum possible age for this to have occurred was during MOIS6. It is therefore considered reasonable to suggest that the lowermost of the sequence boundary surfaces LQ_b, should be ascribed a minimum age of MOIS6. According to the principle of superposition, the rest of the units overlying this surface must be of a younger age.

Depositional Sequence II

Depositional Sequence II accumulated during the transgression from the lowstand associated with the formation of LQ_b (i.e. after MOIS6). Surface LQ_b was potentially continually incised right up until the time of the initial accumulation of Depositional Sequence II. Therefore, it is proposed that Depositional sequence II was deposited during the early transgression from MOIS6 to MOIS5e, as glacio-eustatic sea-level rose and the palaeovalleys were filled with shoreface type deposits. Deposition of the units in Depositional Sequence II probably ceased as the shoreface moved landward (towards its highstand position) across the study area, creating a ravinement surface that was reworked or destroyed during the creation of surface R_I and implying that units B, C, D, E and F may be diachronous facies of advancing sea level.

6.4.3 Seismic Sequence III

Ravinement Surface R_I

Ravinement surface R_I formed during the overall regression between MOIS5e and MOIS4, a period of over 55,000 years (Figure 6-5). During 40,000 years of this interval, eustatic sea level (and consequently the shoreface), although fluctuating, was in the range of -40 m to -60 m below present (Figure 6-5) i.e. the altitude of the PhD study area (note that the possible effects of crustal uplift in the region are discussed further in Chapter 7). This created the conditions for the development of an extensive erosion surface by nearshore marine processes under a cooling climatic trend.

Depositional Sequence III

Initial formation of the majority of Depositional Sequence III is interpreted to have been penecontemporaneous to the formation of ravinement surface R_I (MOIS5e-MOIS4). However, the sequence was subaerially exposed during the long interval of cold conditions that followed (MOIS4 to MOIS2, *circa.* 70,000 years). There is no evidence in the PhD study area to indicate that the large fluvial system that incised LQ_b returned, although such a system may have adopted a course elsewhere. The substantial length of time under which the study area was exposed was more than adequate for the palaeosol deposit recognised on the upper surface of the unit to develop.

6.4.4 Seismic Sequence IV

Sequence Boundary Surface H_b

Surface H_b was created during transgressive shoreface erosion across the study area associated with climatic amelioration and consequent glacio-eustatic sea-level rise from last glacial maximum (MOIS2) to the highstand of the Holocene (MOIS1)

Depositional Sequence IV

Units SFU-H and SFU-I were laid down contemporaneously with sequence boundary surface H_b. SFU-H is considered to represent a channel that was infilled in relatively low energy conditions during the transgression from MOIS2 to MOIS1. Since deposition, SFU-I has not undergone significant change and is not mobile under present sea bed current regime.

6.4.4 Summary

Figure 6-4 presents a summary diagram of the chronostratigraphic model of the sedimentary sequence identified in the PhD study area. It is recognised that this chronology is not absolute and maybe open to further interpretation. In particular, the starting point for the above model is the formation of LQ_b during the lowstand associated with MOIS6. It is conceivable that LQ_b is, in fact, older than this age and that there is a unrecognised hiatus between Depositional Sequence II and R_1. However, for the sequence to be assigned an alternate chronostratigraphy other sequence bounding unconformities must have originally been present in the sequence and subsequently been completely destroyed. Whilst it is not impossible for this to have occurred it is considered unlikely because at least some evidence of these surfaces would probably have been preserved. In the following Chapter, this hypothetical age model is developed further by comparing the results from the PhD study area with other data from the English Channel and North Sea Basin.

Chapter 7: Discussion

7.1 Introduction

In Chapter 6 a model was proposed for the nature and origin of an incised valley system and its associated infilling deposits on the seafloor of the eastern English Channel. In this Chapter, this model will be used as a basis to explore a range of issues relating to the wider geographical evolution of the English Channel during the Late Quaternary. It begins by comparing the record from the PhD study area and other sites in the eastern English Channel, and then scales up to consider the implications of this work for the evolution of the wider English Channel basin. From here the chapter develops further by exploring more fully the role of eustatic and tectonic processes in controlling the physical evolution of the wider study area.

7.2 The Late Quaternary Evolution of the Eastern English Channel

7.2.1 Introduction

This section opens the discussion of this thesis by considering the implications of the present study for our understanding of the Late Quaternary evolution of the eastern English Channel region. It deliberately starts at a local scale, focussing on links between the model developed in Chapter 6 and other sites located in the immediate area surrounding the study site. Its main aim is to determine the extent to which the findings of this PhD research are supported (or not) by previous work in the region. It is recognised at the outset that a major hindrance to this work is the limited nature of the absolute age control that exists both in the PhD study area, and within the wider eastern English Channel. As a result, much of what follows is, by necessity, based on an interpretation of morphological and sedimentological data, combined with seismic records where they exist. The main points for consideration are:

- The depth, dimensions and types of bedrock channels
- The chronology of incision and infilling
- The thickness and nature of the sediments that infill the bedrock channels

7.2.2 The depth, dimensions and types of bedrock channels

At a macro-scale, the multi-channel rockhead incised-valley system of the eastern English Channel is most readily explained as a result of the opening of the Strait of Dover and the formation of an extensive incised bedrock channel system by drainage of a large proglacial lake in the southern North Sea. This primary network of valleys has since been modified by subaerial fluvial action and marine transgression / regression during multiple glacial – interglacial cycles. Therefore, the channels and their associated infills are typically nested within each other, sometimes with multiple generations of channels crossing the same part of the eastern Channel floor.

The depths of the bedrock channels in the eastern English Channel are important since they provide a potential means to assess the relative age of the channels (based on their maximum depth of incision). For example, in a region that has experienced long-term uplift it is reasonable to hypothesise that, all other factors being held constant, the older channels will occur at a higher elevation than the younger channels. The nature of the channels (infilled or open) is important since they provide information regarding their depositional origin. Their analysis may make it possible to assess whether the channels originated from one or more of the main rivers that drained through the Strait of Dover at different times during the Late Quaternary, whether they originated as a result of catastrophic breaching of the Strait, or if they formed as a result of more time-transgressive processes.

Rockhead depth

Within the present study the channel features identified are incised into the Tertiary rockhead surface to depths ranging from 60 to 65 m (locally up to 70 m) below present sea-level (bpsl). Unfortunately, very little detailed work comparable in resolution to that reported here from the PhD study area has been published from other areas of the eastern English Channel, and what data do exist have mainly been centred on the offshore extensions of individual river valleys such as the palaeo-Owers (Bellamy, 1994; 1995), palaeo-Arun (Palmer-Felgate *et al.*, 2003) and palaeo-Solent (Dyer, 1975; Velegrakis, 1999), systems that are considered to be somewhat smaller than the one identified in the present study.

The British Geological Survey Dungeness-Boulogne sea bed sediments and Quaternary map sheet (Hamblin, 1990) provides a map of the rockhead in the eastern English Channel. To the northeast of the present study, the Lobourg Channel runs through the Strait of Dover with basal channel depths of 40 to 50 m bpsl, locally attaining depths of over 70 m bpsl (Figure 7-1). To the north of the PhD study area, within the Northern Palaeovalley, to which the Lobourg Channel appears connected, is a narrow deeply incised and infilled thalweg (Palaeovalley A on Figure7-2) at predominantly 70 m bpsl and locally reaching depths of 80 to 90 m bpsl (Figure 7-1). To the southwest of the study area, an infilled incision within the Southern / Median Palaeovalley (Palaeovalley B in Figure 7-2) also locally attains depths to rockhead of 70 to 80 m bpsl, although the thalweg of the system is generally between 50 and 60 m bpsl.

On the basis of the Hamblin (1990) rockhead contour map (Figure 7-1), and examination of the palaeographical relationship with the Palaeovalley B (Figure 7-2), it seems reasonable to suggest that the incised valley system investigated in this PhD study is an upslope continuation, or tributary, of this palaeovalley system (Figure 7-2). This suggestion is based on the broadly similar depth of the bedrock channel, as well as the location and orientation of the channels identified by this work. It is unlikely that the features identified by this study had any connection with the Palaeovalley A. This is because the rockhead of the latter is significantly lower (by *circa* 10 – 15 m), than that of the channel identified in this research. Given that the Palaeovalley A appears to cross-cut the study area perpendicularly, upslope of the study area, this strongly suggests that Palaeovalley A in the eastern English Channel post-dates the incised-valley system in the present study. This is a significant observation in that it confirms that the eastern English Channel supports channels of multiple ages that are, in some locations, superimposed on one another.

Channel types

According to Hamblin (1989) there is a distinct difference between the infilled and open channel features on the floor of the eastern English Channel. He suggests that the infilled valleys generally have very uneven floors with locally incised thalwegs, whereas the open valleys have relatively flat floors with no local overdeepening. The distribution of these two types of channel features is illustrated in Figure 7-3. As has been discussed in Chapter 6, the current study area is typical of the infilled valleys mapped by Hamblin (1990), with 10 to 30 m of sediment infilling a valley that has an uneven floor with rockhead undulations of 10 to 15 m. This contrasts with the open channels which contain no sediment, and are simply open bedrock incisions. Figure 7-3 shows that the open channels are geographically restricted to the Lobourg Channel in the north of the study area, and the Northern and (parts of) the Southern / Median palaeovalleys. Although Hamlin (1989) proposes this as a simple division, the reality is that in many instances the two channel types overprint one another or merge (Figure 7-3). This is especially true in the Northern Palaeovalley.

Two hypothesises that may explain the differences between the open and infilled channels of the eastern English Channel are discussed below:

(i) That the open and infilled channels differ in age;

(ii) That differing processes acted to form the two types of channel.

(i) This hypothesis suggests that the open channels, shallower than their infilled counterparts, were incised first (i.e. MOIS 12; Gibbard, 1988) and pre-date the infilled channels. The differences between the open and closed channels can thus be explained as a function of either the reworking and removal of previously deposited valley infill under multiple glacio-eustatic cycles, or simply that they were never infilled with sediment following their initial formation. Under both scenarios, the open channels would have progressively widened through time. In contrast, the younger channels have not, as yet, been subjected to multiple transgressions and regressions and therefore their depositional sequences and external geometries have remained intact. This hypothesis is, however, difficult to test, not least because of the almost complete lack of absolute age control on the nature of the channel infill sequences. Moreover, the older open channels lack any sediment and therefore any inference as to their age is particularly speculative.

Despite these problems, assuming that channel infill reworking has not lowered the rockhead of the channels, in theory it should be possible to test this hypothesis by comparing the bedrock elevation of infilled and open channels; the latter should be at a higher elevation than the former since they are of greater antiquity (a more detailed discussion of the uplift that underpins this assumption is provided later). This does appear to be the case. Thus, an inspection of Figure 7-1 shows that on average the rockhead elevations of the open channels are typically between the range of 30 to 40 m bpsl, whereas those of the infilled channels (Palaeovalleys A and B) are 50 to 70 m bpsl. This strongly suggests that the open and closed channels are of different age.

(ii) The second hypothesis for the differences in channel nature is related to their mode of formation. As discussed in Chapter 3, Smith (1985) invoked a catastrophic origin for the

Strait of Dover and the Lobourg Channel by the overflow of an ice-dammed lake from the southern North Sea during MOIS12. This hypothesis suggests that the numerous wider open channels are the downstream remnants of the ice-dammed lake overflow. In contrast, the narrower more deeply incised infilled channels are the product of "normal" processes associated with channel incision and infilling under multiple glacial-interglacial cycles of fluvial incision, infilling and marine transgression / regression.

The irregular nature and over deepening of the narrower infilled valley floors may in part be related to enhanced erosion at palaeovalley confluence points, a feature that is found in terrestrial river systems (Berry, 1979; Bridgland, 1983), or further incision that occurred as a product of tidal scour (Dingwall, 1975; Gibbard, 1988) during transgression, prior to sediment infilling, as fluvial conditions gave way to tidal currents. Alternatively, the open valley floors could also results from a combination of both processes through time.

Given the lack of chronological data, it is difficult to test these competing hypotheses. However, several implications would arise from this work were either to be correct. First, if we assume that the open channels originated as a result of the breaching of the Strait of Dover, then it is reasonable to take the base level of these channels as a datum against which other younger or older river channels can be compared. As noted above, the open channels belong almost exclusively to the Loburg, Northern and (part of) the Southern / Medial Channels. Their base levels are typically higher than those of the channel identified in the PhD study area. If the open channels are associated with the initial breach of the Strait of Dover (assumed age MOIS12, Gibbard, 1988), then the channel in the present study, because of its greater depth, must post-date this event. This in turn implies that the sea floor morphology of the Strait of Dover owes itself to a combination of catastrophic and more gradualist processes (see discussion in Chapter 3).

Alternatively, if the tidal scour hypothesis is accepted, then one must identify a logical explanation for the geographic distribution of this process. Why, in particular, is tidal scour restricted to the areas of the open channels, and is there a physical process that can explain this distribution (or lack thereof) of sediment ? In the northern part of the eastern

English Channel, the concentration of tidal flow may provide a logical explanation for the lack of channel infill and associated scour (Bridgland and D'Olier, 1995). However, the tidal scour explanation is less powerful in areas further south, where open and infilled channels (Figure 7-3) show little pattern with respect to spatial variations in current sea bed velocities (Figure 7-4). It seems reasonable to assume that current tidal flows have not remained constant during either this interglacial, nor are they necessarily typical of previous interglacial highstands. Nevertheless, the spatial distribution of the two channel types cannot readily be explained as a result of differential tidal scour under almost any sensible palaeogeographic reconstruction.

Given the above discussion, one can conclude that the most reasonable explanation for the formation of the open and infilled channels recorded on the floor of the eastern English Channel is that the channels are of different age.

7.2.3 Chronology of incision and infilling

It is widely believed (Gibbard 1988, 1995; Bridgland and D'Olier, 1995) that the Strait of Dover was cut by breaching of the Wealden-Artois ridge during the Elsterian / Anglian (MOIS12) glaciation. It was this breaching that Smith (1985) proposed created the geomorphological blueprint onto which much (but not all) of the incision and infilling of the present eastern English Channel sea floor is based. However, this is not a model that all authors support, especially in areas where drowned valleys can be related clearly to their terrestrial counterparts (Ashton and Lewis, 2002).

Bellamy (1994, 1995) suggested that the incision and infilling of the palaeo-Owers valley is likely to have occurred over several climatic cycles during the Quaternary with episodic cut-and-fill events. He indicated that the most deeply incised portion of the palaeo-Owers valley formed during the last glacial maximum (MOIS2) when relative sea-level fell to -50 m bpsl, and that much of the infilling of the valley occurred during the transgression of the Late Devensian and Early Holocene. Moreover, Velegrakis *et al.* (1999) painted a very similar picture to Bellamy (1994, 1995), proposing a comparable chronology for the incision and infilling of the incised valleys of the palaeo-Solent that date from MOIS2 onwards.

In Chapter 6, it has been argued that the age of the incised valley and associated infill of the PhD study area dates from MOIS6 / 5e. As such, this hypothesis suggests an origin that post-dates the formation of the Strait of Dover (assuming a MOIS12 age) and predates the chronology developed by Bellamy (1994, 1995) and Velegrakis *et al.* (1999) from the Solent region. Moreover, because the Palaovalley A cross-cuts the upstream portion of the incised valley of the PhD study area, this suggests that the Palaoevalley A dates from either later on in the same cold period (MOIS6) or, as seems more likely, the subsequent cold lowstand periods of MOIS4 or MOIS2.

This discussion suggests that there is evidence from the eastern English Channel for at least three distinct phases of valley incision and infilling. The earliest phase, responsible for establishing the macroscale geography of the Channel floor dates from MOIS12 and the breaching of the Strait of Dover. The second phase is that recorded in the PhD study area which, it is proposed, dates from MOIS6/5e. The final and most recent phase of valley incision and infilling is that associated with the MOIS2/1 transition, as is illustrated in the Solent region (Dyer, 1775; Velegrakis *et al.*, 1999) and, to a lesser degree, in the PhD study area.

7.2.4 Channel dimensions

Channel dimensions are interesting since they provide an indication of the size (and potential source(s)) of the channels that have drained across the floor of the eastern English Channel, although it is recognised that rivers may, in the past, have occupied several channels simultaneously.

Cross-sectional profiles of the main incised channel feature identified during this PhD study indicate a width of between 3 and 8 km and depths of between 18 and 20 m, with an average width : depth ratio of c. 1:250. It is difficult to obtain accurate width : depth ratios for the deep narrow incisions of Palaeovalleys A nd B, because of the broad scale of the mapping completed to date and the relatively coarse height control on the bedrock contours. However, reasonable estimates based on the transect locations shown in Figure 7-3 are provided in Table 7.1. These demonstrate that the infilled channels have average width : depth ratios of ~1:200. These values are very similar to those proposed from the PhD study area, but are significantly larger than those from the palaeo-Owers River (~1:100). This implies that the channel identified in the PhD study area is of comparable dimensions to those of the infilled palaeovalleys A and B, and that it therefore potentially originated from a similar-scale fluvial system. Thus they probably represent the former course of the Thames-Rhine-Meuse "Channel River" system and not a smaller tributary valley.

7.2.5 Types of Deposit

This section compares the infill deposits of the PhD study area with those reported from elsewhere in the eastern English Channel. The aim is to determine the extent to which the valley infill deposits are similar (or not) to other sequences. Because of the limited data available, the main comparisons are made with the work of Bellamy (1994, 1995) and Velegrakis *et al.* (1999) based in the Solent region, where detailed seismic and borehole data exist.

The lack of gravel terraces on the interfluves identified in the study area contrasts the widespread occurrence of such landforms around the fringes of the English Channel (Allen and Gibbard, 1993; Maddy *et al.*, 2000; Antoine *et al.*, 2003), including the Thames River valley and offshore in the southern North Sea (Bridgland, 1994). Other studies on the shelf of the eastern English Channel investigating the offshore extensions of the River Owers (Bellamy, 1994, 1995) and River Arun (Dickinson *et al.*, 2003) have demonstrated well-developed gravel terraces in submarine settings within cut and fill sequences that are genetically related to palaeovalley incision and infilling. The lack of comparable terraces in the PhD study area most probably reflects their erosion during transgressive and regressive episodes (these are discussed further below).

Velegrakis *et al.* (1999) divided the unconsolidated sediments that infill the palaeovalleys of the former Solent River to the west of the PhD study area into modern sediments and palaeovalley infill deposits. These authors suggested that the former comprise a thin veneer of sediment whose spatial distribution is controlled by the prevailing hydrodynamic regime. They, like Bellamy (1994, 1995), attribute these sediments to modern processes. They are very similar in their widespread distribution and relative thinness to the uppermost stratigraphic unit identified in the present study. However, in contrast to the findings of Velegrakis *et al.* (1999), the uppermost sediments in the PhD study area comprise a coarse grained Holocene lag deposit, as previously described by Hamblin and Harrison (1989) and Hamblin *et al.* (1992). They are not considered to be mobile under the present hydrodynamic regime.

Velegrakis *et al.* (1999) identify a ravinement surface in the stratigraphy of the Solent embayment that separates the modern from the palaeovalley deposits. This boundary is defined by an inner shelf sedimentary sheet that overlies fluvial and paralic deposits that infill former sub-aerial valleys. Velegrakis *et al.* (1999) suggest that the sedimentary units infilling the Palaeo-Solent were deposited immediately before and during the Holocene transgression of the Palaeo-Solent. The thalweg of the deepest infilled channel extends to a depth of -45 m bpsl and can be related to a staircase of raised river terraces that have been mapped above present sea level across the Solent region by Allen and Gibbard (1994). The infill sequence of the palaeovalleys typically reach thicknesses of 5 to 10 m and comprise a lowermost Late-glacial fluvial gravel overlain by early Holocene transgressive paralic deposits that include intercalated units of peat, clay, silt, and aggrading sands and gravels. These later sediments mostly formed during the transgressive phase of the current interglacial.

Bellamy (1994, 1995) suggested the majority of the sediment infilling the palaeo-Owers incised valley to be fine grained sands and muds which contain biostratigraphical evidence for deposition in estuarine, intertidal, shallow sublittoral and open-shelf type environments. Bellamy (1994, 1995) lacked any absolute age control for the interpretation of the sequence he observed, but suggested that the sedimentary succession accumulated in a low energy estuarine and near-shore setting during the transgression of MOIS2 / 1.

On seismo-stratigraphic grounds, the albeit much smaller depositional sequences of the incised valley systems of the palaeo-Solent (Velegrakis *et al.*, 1999) are considered to be similar to Depositional Sequence II observed within the incised valleys in the PhD study area. In particular, they are characterised by the relatively coarse-grained nature of the sediments and an aggrading nature. This supports the suggestion that the deposits infilling the main valleys in the PhD study area also originated during a transgressional phase of an interglacial cycle, albeit an older one associated with the transition from MOIS6 to 5e.

7.3 Late Quaternary Evolution of the English Channel

7.3.1 Introduction

This section broadens the discussion further to consider the evolution of the Channel system as a whole by relating the sequences observed in the eastern English Channel with those downstream of the PhD study area in the Western Approaches (particularly the Celtic Sea Margin and the Amorican Margin) as well as the upstream areas of the southern North Sea Basin and the Strait of Dover. The objective is to explore wider patterns of palaeogeographic development and causal processes responsible for the patterns observed.

7.3.2 Links between the PhD study area and the Southern North Sea Basin and Strait of Dover

The palaeovalley system identified in this PhD forms part of a larger channel network that extends northwards via the Strait of Dover into the Southern North Sea Basin. In ascribing a potential MOIS6/5e date to the formation and initial infilling of this channel, it is appropriate to examine whether other records from these areas to the north support or refute this hypothesis. As has been discussed previously, there remains considerable uncertainty regarding the timing of the breaching of the Strait of Dover, and the formation of the main palaeovalley network in the eastern English Channel. Records from the confluence area of the Thames and the Rhine-Meuse in the southern North Sea Basin might be expected to record this diversion, perhaps by a reduction in the delivery and deposition of sediment as a result of altered flow patterns.

The most prominent and extensive of the Late Quaternary deposits in the southern North Sea Basin is the Yarmouth Roads Formation, which formed under shallow water fluvial and intertidal conditions and is derived from the offshore extension of a large (Rhine-Meuse) delta complex originating from the European Low Countries (Balson and Cameron, 1985; Zagwijn and Doppert, 1978; Zagwijn, 1979, 1989). The age of the formation is not known exactly, but Cameron *et al.* (1992) proposed an Early to Middle Quaternary age and suggest that formation of this vast delta plain took more than 1.6 Ma, growing largely independent of climate variation (Jeffrey and Long, 1989). Interestingly, according to Jeffrey and Long (1989), the Yarmouth Roads Formation ceased deposition during Cromerian complex or Elsterian time (MOIS15-12). Gatliff *et al.* (1994) hypothesised that, during this time, marine transgression converted the delta plain surface of the Yarmouth Roads Formation into a shallow sea. With the subsequent sea level fall of MOIS12, the upper surface of the formation was a ready made, low relief surface that facilitated the spread and coalescence of British and Scandinavian ice sheets and, as a direct result, ponding of a proglacial lake formed to initiate the southwards drainage of the Rivers Thames, Rhine and Meuse via the newly formed Strait of Dover into the English Channel. However, this argument is based on no absolute age controls; indeed, there is a clear element of supposition in ascribing this age, not least because this merely reflects the hypothesised Elsterian age for the breach of the Strait of Dover proposed by Gibbard (1988), again in the absence of independent absolute dating control.

The North Sea Basin record therefore provides only circumstantial evidence for a diversion of the Thames-Rhine-Meuse system southwards during MOIS12. Little can also be concluded regarding the nature of River Thames and its direction of flow during MOIS6. In the Thames, MOIS6 is represented by the Taplow / Mucking Terrace which, at the mouth of the present Thames (Dartford), has an elevation of *circa* 10 m OD (Bridgland and D'Olier, 1995). Downvalley extrapolation of this profile is complicated by possible uplift of the Strait of Dover and subsidence of the southern North Sea basin during the Late Quaternary. However, given that the depth of the thalweg of the hypothesised MOIS6 channel identified in the PhD study area lies between -60 and -70 m bpsl, it is feasible to suggest that this feature it is a downstream continuation of the MOIS6 terrace system of the Thames.

In summary, this section has demonstrated a possible link between the Late Quaternary infilling of the North Sea Basin and the breaching of the Strait of Dover. However, the age for the reduction in sediment supply associated with the upper surface of the Yarmouth Roads Formation is not known with any confidence – it could pre- or post-date MOIS12. Indeed, it would be equally plausible, based on the seismostratigraphical and lithostratigraphical evidence from this PhD study alone, to argue that the Strait of Dover was not breached until MOIS6, and that the cessation of deposition of the Yarmouth Roads Formation also dates from this time (cf. Ashton and Lewis, 2002).

Having discussed upstream links with the southern North Sea Basin, this section now turns to consider in further detail the geomorphology of the Strait of Dover. Unfortunately, within the Strait of Dover, tidal scour has removed any low-stand fluvial signature from the sea bed (Bridgland and D'Olier, 1995). This further precludes our ability to determine the age of the breach and establish a potential minimum age for the formation of the channel system mapped in this PhD study.

With regard to the palaeovalley morphology in the Eastern Channel, Smith (1985) has already demonstrated that certain palaeovalleys on the English Channel shelf could be linked with rivers onshore such as the Seine, Somme and Béthune in Northern France and the Solent and Arun of southern England. However, determining a direct terrestrial source of the channel identified in this research is not easy, due to the central Channel location of the PhD study area and the fact that most of the other major rivers have reasonably well defined offshore counterparts. For example, the location of the PhD palaeovalley precludes the possibility that it was ever part of the Seine, Somme, Solent or Arun rivers. There is, however, a possibility that the Rivers Authie and the Canche, in northeast France, drained via the channel identified in this study. This seems unlikely, for although the Rivers Authie and Canche show significant thickness of offshore deposits (Figure 7-3) they do not demonstrate rockhead incision of sufficient magnitude and depth (Figure 7-1) to have provided the erosive force to create the infilled palaeovalleys mapped in this research.

The lack of well-dated palaeovalley deposits that can be traced either upstream into the Southern North Sea Basin, or laterally to the coasts of England or France, means that correlating the palaeovalley deposits identified in this research with other records is not immediately possible.

7.3.3 Downstream links between the PhD study area and the Central and Western English Channel and the Western Approaches

Central and Western Channel

The downstream portion of the English Channel contains the downslope continuation of the Channel River system and comprises two distinct areas: (i) The central and western English Channel, westward continuations of the continental shelf planation and its associated features, and (ii) The outer shelf, the slope and the continental rise of the

Western Approaches margin, including the large depocentres on the shelf edge such as the Celtic Fan and the Americal Fans (Figure 7-5). The first area is of interest since it provides an opportunity to compare the findings of this research with other palaeovalley deposits on the continental shelf of the western English Channel. However, because of problems with erosion and sediment reworking discussed above, the second area is also considered so as to explore whether it can provide a less interrupted record of Late Quaternary environmental change in the English Channel.

Southwest of the PhD study area, the Northern and Southern Palaeovalleys and the palaeovalley of the offshore extension of the Somme coalesce to form a single "Channel River" palaeovalley that runs in a generally NNE-SSW direction. This coalescence occurs just before the palaeovalley enters the Hurd Deep and is a region examined in detail recently by Lericolais *et al.* (2003), who identify in seismic profiles what they consider to be the confluence of the Northern Palaeovalley with the palaeovalley of the Somme (Southern Palaeovalley).

Lericolais *et al.* (2003) mapped the offshore continuation of the River Seine, which upstream of Barfleur, has a relatively smoothly dipping long-profile of 0.4‰. Downstream of this point the valley floor becomes more irregular with overdeepening of 30 m or more. This represents a nick-point, downstream of which the valley floor dips more steeply with a slope of c. 0.1%. This down-valley transition coincides with water depths of c. 70 m, and Lericolais *et al.* (2003) suggest that this records a change in equilibrium profile associated with fluctuations in base level (see next section). The palaeovalley infill above this nick-point is characterised by thick accumulations of sediments that record multiple phases of incision and deposition. The incisions are between 100 and 200 m wide and are thought to have formed under conditions associated with meandering or anastomosing channels. Correlations between channel fill sequences suggest strong spatial variability, with local factors influencing the preservation and deposition of any particular sediment sequence.

This study lacks any sedimentological control – nor does it have any independent dating information. However, Lericolais *et al.* (2003) hypothesised that the initial incision of the Seine palaeovalley system was associated with the onset of "true cold climates" in the Praetiglian (MOIS100, 98 and 96). This set of three major lowstands may, they suggest, have caused the incision of the Seine above the nickpoint (identified above) when sea level fell repeatedly below about 70 m. They go on to suggest that the very large regressions of the Mid and Late Quaternary (MOIS22, 16, 10, 6 and 2) that saw sea level dropping to below 100 m, caused the entire western English Channel to emerge. Under these conditions, multiple cut and fill episodes associated with cold stage fluvial processes caused the infill of the palaeovalleys identified in the confluence area. A comparison with nearby onshore areas of southern England shows that any raised marine sediments of a Praetiglian age are only preserved capping very high ground for example the Netley Heath deposits in Surrey (John and Fisher, 1984)

In comparison to the PhD study area, there are both similarities and differences to the sequence described from the western Channel by Lericolas *et al.* (2003). First, these authors suggest that most of the sediments that they describe formed under lowstand fluvial conditions when the Fleuve Manche (Channel River) operated as a major river flowing across this relatively deep part of the Channel. This differs from the interpretation proposed for the PhD study area, where, it is proposed, trangressive marine sediments form the palaeovalley infill. It is not immediately clear why Lericolias *et al.* (2003) ascribe a fluvial origin to all of the sediments under study, although it may reflect their reliance on a three phase model for the Channel that envisages in an upstream drainage basin, a central transfer zone, and a downstream depositional basin. This model assumes that the deposits accumulate only under fluvial conditions in the central transfer zone. Secondly, the Lericolais *et al.* (2003) interpretation suggests multiple cut and fill episodes have occured, with evidence for widespread erosion in some instances. This is more similar to the sequence observed in the PhD study area, and provides further evidence that the palaoevalley systems in the Channel contain sequences that are discontinuous in time and space. Indeed, sediment packages within any palaeovalley may differ in age from those lying above or below by several glacial / interglacial cycles and as such cannot be viewed as continuous records of palaeoenvironmental change.

From a chronological perspective, there is limited scope for correlation between the sequences. Lericolas *et al.* (2003) suggest that the initial incision of the Seine system occurred during the Praetiglian – MOIS99, whereas this study proposes that the incision of the PhD study area palaeovalleys postdate this and only developed after MOIS12. If this hypothesis is correct, it implies that the sediment and water throughput that passed downstream *via* the Somme and the Seine systems is likely to have changed dramatically through time. For example, pre-MOIS12 these downstream palaeovalleys were only draining a small amount of the NW European continent in comparison to their discharge after this time, when the combined waters of the southern North Sea Basin and Channel coasts drained through them. In this respect it is interesting to note the large hiatus inferred in their model, between the initial incision and early infilling, dating from MOIS99, and the much younger phases of infilling proposed from MOIS22, 16, 10, 6 and 2. This hiatus is attributed to the magnitude of eustatic sea level change during the intervening period, which they suggest was too small to trigger fluvial deposition within these channels. An alternative hypothesis is that the flooding of the Strait of Dover was associated with widespread erosion and reworking of palaeovalley sediments throughout the English Channel and not, as is generally believed, in the eastern sector alone. This hypothesis would suggest that there might be a widespread discontinuity between older sediments that may be preserved in the base of these palaeovalleys and post MOIS12 deposits that formed after the Strait of Dover flooding.

Westward of the Hurd Deep there is little or no evidence of major incisions by the Channel River preserved on the shelf. However, this situation could arise as a result of a combination of (i) the shelf having only been sufficiently exposed for incision to occur on a handful of occasions since the Praetiglian during major lowstand periods (Lericolais *et al.*, 2003), and / or; (ii) evidence for incision having been erased by shoreface erosion and ravinement processes associated with the rapid sea level rise (for example *circa* 1.5 cm per year during MOIS2-1). A combination of these processes would heavily modify the Western Channel Shelf to form the planated surface observed today.

Western Approaches: The Continental Rise, the Slope and the Outer Shelf

Some 200 km west of the Hurd Deep, close to the shelf break of the Western Approaches, is an extensive network of sandbanks and palaeovalleys. Some of these features are connected to canyon systems on the continental slope that feed the two turbidite sytems of the Celtic Sea and Armorican deep-sea fans (Bourillet *et al.*, 2003). These represent the downstream portion of the Channel system and provide a potential depositional sink for sediments transported through the Channel "sediment conveyor".

The Celtic Sea sand banks cover an approximate area of 100 000 km^2 and extend from the northern Bay of Biscay to the Irish Channel entrance (Reynaud *et al.*, 1999) (Figure 7-5). Gibbard (1988) and Lericolais (1997) interpret these banks as the reworked remnants of mouth deposits of the lowstand Channel River, with the incised channel systems in the banks attributed to the estuary mouth or delta-system palimpsests of the Channel River (Reynaud, 1996; Reynaud *et al.*, 1999, 2003; Berné *et al.*, 1998).

Reynaud *et al.* (1999) completed a very high-resolution seismic survey of Kaiser Bank (60 km long and 4 - 6 km wide), one of the shelf tidal sand banks (100 - 150 m bpsl) belonging to the Celtic Sea shelf sandbank system. Little is know about the timing of the formation of the Celtic Sea banks, although Reynaud *et al.* (1999, 2003) speculate that their initiation could have begun in the Early Pleistocene. However these authors also propose that the main body of the Kaiser Bank was formed more recently than the Last Glacial Maximum MOIS2, suggesting that during lowered sea-levels Celtic Sea Banks would have been subjected to severe wave erosion that might have planed tens of metres of sediment from their upper surfaces. The Raynaud *et al.* (1999) survey also reveals a network of slightly sinuous to anastomosed erosional channel forms that has locally incised into the Kaiser Bank surface and subsequently been infilled. Reynaud *et al.* (1999) suggest that the Celtic Sea Shelf sand banks were deposited in an offshore setting during a transition from a tide- to a wave-dominated system and that the banks architecture reflects evolution of the area throughout the last post glacial sea-level rise (MOIS2/1).

The investigations of Reynaud *et al.* (1999, 2003) indicated cyclicity in the destruction and formation of the Celtic Sea Shelf sand banks on glacial – interglacial timescales. This suggests that evidence of deposition in this region related to the timing of the incision and infilling of the palaevalleys in the PhD study area is highly unlikely. While it is reasonable to infer that a similar type shelf sand bank system developed in the region during the transgression of MOIS6/5e, any such system would almost certainly have been destroyed during the lowstand of the Last Glacial Maximum (MOIS2).

Zaragosi *et al.* (2000, 2001a, 2001b) and Bourillet *et al.* (2003) studied the nature and origin of the shelf slope and deep sea fan sediments further offshore of the Western Approaches. In contrast to the purely seismic record reported by Reynaud *et al.* (1999), Zaragosi *et al.* (2000) obtained core samples from the fan surfaces (*circa* ~4000 m water depth) and showed that over the last 15,000 years up to 5 m of sediment had accumulated. Bourillet *et al.* (2003) undertook analyses of core data from the Western Approaches shelf slope canyon interfluves (2174 m deep), and their data provides valuable information on the nature of sedimentation rates at the shelf edge dating back to MOIS3 (Table 7.2).

Table 7.2 shows that peak sedimentation rates during MOIS2 reached 30-55 cm kyr^{-1} for the Meriadzek Terrace, compared with 40-400 cm kyr^{-1} for the Trevelyan Escarpment. Importantly, these rates vary through time, and are closely linked to the prevailing rate of sea-level rise and the timing of deglaciation. Thus, Bourillet *et al.* (2003) suggest that the highest rates of sedimentation occurred between circa 20 – 13 ka BP during the rapid deglaciation of the British and northwest European ice sheets. This implies that maximum deposition on this part of the Channel shelf edge is associated with the postglacial sea level rise of MOIS2 to MOIS1, and that high sedimentation rates were associated with remobilisation of sediment previously deposited on the exposed Western Channel floor by transgression during the initial period of the postglacial rise in sea level.

The analysis of core material from the Western Approaches demonstrates two further points of importance. First, rates of sedimentation are generally higher during cold stages compared with more temperate interglacials. This suggests that the alternating glacial / interglacial cycles of the Quaternary might be preserved in these sediment sequences as an alternating sequence of relatively fast and slow accumulated sediments, assuming no erosion and / or reworking. Second, the sediment sequences are dominated by fines – silts and sands – with no coarse grained gravels. This suggests that material of this grade, arguably one of the most significant bodies of material transported by the "Channel River" fluvial system, has either been deposited in the palaeovalleys of the English Channel, or has been reworked during successive interglacials by marine processes to form the extensive coastal accumulations of gravel that occur along much of the English and (to a lesser degree) the French coasts.

Third, the succession of glacial-interglacial sediments from MOIS3 to present on the shelf slope and deep-sea fans of the Western Approaches suggests that, at further depth, there should be a continuous succession of "Channel River" sediments for the previous glacial-interglacial climate cycles extending backward throughout the Quaternary for as long as the river system has existed. This is interesting because the depositional history of these basins should be connected to that of the North Sea Basin. In particular, one may hypothesise that after MOIS12, just as sediment supply to the North Sea Basin from the Thames-Rhine-Meuse system is thought to have been cut off (and the Yarmouth Roads Formation ceased deposition), so in the Celtic Sea and Armorican Fans there should be a compensatory increase in sedimentation rates as the waters from these rivers were diverted through the Strait of Dover and across the Channel floor.

The importance of the Strait of Dover as a control on sediment supply to these downstream locations is indicated by Bourillet *et al.*'s (2003) comparison of deposition on the Armorican and Celtic Sea Fans during the Holocene. These authors note that terrigeneous sedimentation flux ceased on the Armorican Fan *circa* 10 ka BP, but persisted on the Celtic Sea Fan until c. 7 ka BP. This difference in timing is attributed to the flooding of the Strait of Dover during the early Holocene, which effectively cut off the flow of the combined Thames-Rhine-Meuse river system and its associated supply of terrigeneous sediment (which would then have been redirected to the North Sea). Bourillet *et al.* (2003) suggest that this would have effectively cut off the main supply of sediment to the Armorican Fan, whereas the Celtic Sea Fan continued to receive a steady supply of terrigeneous sediment from the North Celtic Sea which was supplied by sediment eroded from the glacio-isostatically uplifting Irish Sea Basin and adjacent land masses.

7.4 Key controls on the Late Quaternary Evolution of the English Channel

The above discussion has highlighted the importance of tectonic and eustatic processes in controlling the evolution of the English Channel during the Late Quaternary. In the following subsections each of these are explored in further detail.

7.4.1 Tectonic Controls

The importance of long-term crustal uplift of the Eastern English Channel, allied with on-going subsidence of the North Sea Basin, are widely thought to exert a strong control on the physical evolution of the English Channel and its associated fluvial and relative sea level history (Preece *et al.*, 1990; Jones, 1999a, 1999b; Lagarde *et al.*, 2003). This section explores the potential significance of these tectonic factors for the interpretation of the data collected in the PhD study area, and for the evolution of the English Channel more widely.

Differential Uplift

Lagarde *et al.* (2003) argue that the Quaternary evolution of the Channel basin has been governed by the position of relative sea-level controlled by tectonic and glacio-eustatic variations. However they also note that, in contrast to eustatic variations, tectonic movements have varied across the region. Examples of differential tectonic movements in the region are discernable from geomorphological evidence at a range of scales.

On the larger scale, Brittany and southwest England display a southwards tilting, which is illustrated by the contrast between the well-developed cliffs on the northern Channel coasts and by contrasting flooded former river valleys on the southern coasts e.g. the estuaries of the Rivers Fal, Cornwall and Vialaine, Brittany (Antoine *et al.*, 2003a). Tilting in this area appears to occur along a NE-SW trend corresponding to the Cadomian and Variscan thrusts. Similar controls are attributed to the gentle northwards downwarping of Normandy that is defined by structural discontinuities running parallel to lines of Cadomian and Variscan faulting (see Antoine *et al.*, 2003a).

At the smaller scale, rivers in Normandy are offset along fault scarps that generate knickpoints close to fault planes (Lagarde *et al.*. 2000). This suggests that uplift is occuring faster than erosion and results in a landscape that demonstrates an "immaturity" in response to uplift (Lagarde *et al.*, 2003). Equally, the entrenchment of modern rivers in southern England and northern France has led to step-like terracing (Bridgland, 1994, 2000; Gibbard, 1985; Antoine *et al.*, 2003) and the general acceptance that at least the margins of the Channel are uplifting (Maddy *et al.*, 2000).

There is widespread evidence for differential uplift of the English Channel during the Quaternary and earlier periods. The magnitude of these crustal motions are large – for example, Lagarde *et al.* (2003) suggest that Normandy may have experienced differential tectonic movement in excess of 300 m, and southern England anywhere in between 250 m and 400m of uplift. Such differences in uplift will have had a profound impact on the Late Quaternary evolution of the eastern English Channel, and evidence for them are revealed by variations in the age and elevation of fluvial and marine deposits in the wider areas of southern and eastern Britain and northern France (Maddy *et al.*, 2000; Bridgland *et al.*, 1993, 2004; Bridgland, 1994; Allen and Gibbard, 1994; Antoine, 1994; Antoine *et al.*, 2000, 2003; Preece *et al.*, 1990; Allen *et al.*, 1996; Bates *et al.*, 1997, 1998, 2003; Westaway *et al.*, 2002).

The longest record of uplift in the eastern English Channel based on marine sediments is that of Preece *et al.* (1990) and Bates *et al.* (1999), based largely on the raised beach deposits of the Hampshire – Isle of Wight basin. The highest marine sediments here occur at Boxgrove, where they attain elevations of 40 m with an age of MOIS13, 11 or 9. Preece *et al.* (1990) correlated the ages and elevation of these deposits with the marine oxygen isosotpe record of Shackleton (1987) and, by assuming a constant uplift rate, inferred a long-term uplift rate of *c.* 10 m per 100,000 years during the Late Quaternary. A similar examination of the raised Quaternary beaches in Normandy by Lautridou (1985) and Lagarde *et al.* (2000) also reconstructed an uplift rate of 10 m per 100,000 years. However, a number of assumptions were made during these studies that potentially limit their validity. These include the extrapolation of the uplift rate beyond the time period for which the data were representative. For example, Lagarde *et al.* (2000) extrapolated Lautridou's (1985) data from intra-Saalian (MOIS7) age raised beach deposits to obtain an uplift rate for the entire Quaternary. A second limitation is in the use of the interglacial eustatic sea-level height values inferred from the oxygen isotope record of Shackleton (1987), which is itself dependent on a variety of other factors. In addition, it is commonly assumed that the raised marine deposits of the English Channel accumulated during the highest sea-level stand in each relevant interglacial, whilst tidal ranges are assumed not to have changed either spatially or temporally. Lastly, it is assumed that uplift has been linear and directionally constant, despite the evidence for slight crustal subsidence (1-2 mm yr^{-1}) observed during the late Holocene (Shennan, 1989; Shennan and Horton, 2002).

Changes in the courses of the Quaternary rivers of the English Channel may have occurred in response to variations in uplift across the area. Indeed, the Seulles River in central Normandy records an uplifted palaeovalley with an eastward deviation that has been linked to Pleistocene uplift of the Arromanches coastal plateau (Lagarde *et al.*, 2003). At a larger scale, the incised valleys on the floor of the eastern English Channel are deeper to the east of the southern limit of the Lobourg Channel (Figure 7-1) compared with areas closer to the English Channel coast. This pattern might be associated with a west to east migration of the English Channel river network through time that could have arisen as a result of differential east – west crustal motions across the region. If valid, this hypothesis would require uplift of the Dungeness area relative to the Boulogne / Calais areas. A review of the tectonic features of the region indicates that the main Variscan front thrust runs through the area, and examination of the long term uplift of the Chalklands of southern England by Jones (1999a; 1999b) suggests that during the Quaternary the Weald has experienced an eastward tilting that may have continued in to the Strait of Dover.

Uplift rates of 100 m per Ma would result in significantly different levels of channel incision. For example, if one assumes no long-term uplift then, for any given glacial – interglacial cycle (assuming they are similar magnitude), channel base level incision would be broadly similar through time. However, it was hypothesised above that the main channel identified in the PhD study area formed during MOIS6 to MOIS5e transition, and that the infilled (relatively) narrow channel (Palaeovalley A) in the Northern Palaeovalley was probably cut later during MOIS2. Using the uplift values cited by Preece *et al.* (1990) and Lagarde *et al.*, (2000, 2003), the time interval between these periods (over 100,000 years) is sufficiently long for up to 10 m greater thalweg incision of the rockhead surface in MOIS2 compared with MOIS6. This is in broad agreement with rates of uplift now recognised globally from area of normal (i.e. post-Archaen) crust, away from major subsiding Quaternary depocentres such as the North Sea (Westaway *et al.*, 2003). This is, interestingly, very similar to the observed channel thalweg differences recorded between the PhD study area and the infilled incision in the Palaeovalley A to the immediate north of the study area, where bedrock elevations reach -70 m bpsl.

7.4.2 Eustatic Controls

Late Quaternary eustatic fluctuations have certainly exercised the greatest control on the timing and magnitude of incision and infilling of the palaeovalley features and associated deposits of the eastern English Channel (Gibbard, 1988; Lericolais *et al.*, 2003). In this section Late Quaternary eustatic sea-level fluctuations are discussed in relation to the tectonic controls of the eastern English Channel. The aim is to examine the significance of relative sea-level fluctuations to the evolution of the study area, and to form a basis for the development of an evolutionary model for the eastern English Channel palaeovalley system.

7.4.2.1 Late Quaternary sea-level fluctuations and palaeovalley incision

Shackleton's (1987) eustatic sea level curve is perhaps the most widely used benchmark for reconstruucting Quaternary ice sheet volume, but in this section I choose to use a more recently developed model proposed by Siddal *et al.* (2003), based on their analysis of the $\delta^{18}O$ values in fossil foraminifera from sediment cores extracted from the Red Sea. This choice is made because the Red Sea sequence contains a high resolution record that is particularly suited to eustatic sea level reconstruction, due to the fact that the Red Sea Basin amplifies, as a result of its peculiar topography, the eustatic fluctuations in such a way to enable a very high resolution reconstruction possible over the last 470,000 years (MOIS12 to present; Figure 2-1).

As noted above, Lericolais *et al.* (2003) suggest that the deep fluvial incision in the Channel only occurs when the Channel shelf is exposed due to a fall in eustatic sea level below -70 m bpsl. If it does not fall below this level, these authors speculate that fluvial channels reoccupy the course of previous cold stage channels (incised when sea level was below -70 m bpsl) enlarging them and eroding their depositional sequences. A characteristic of this, and other analyses concerned with the palaeogeographic evolution of the English Channel, is the adoption of simple eustatic values, without taking into account the long-term pattern of crustal motions that have affected the region. In particular, as noted above in Section 7.4, the centre of the eastern English Channel has experienced uplift at a rate of c. 10 m per 100,000 years. There is not sufficient data to determine spatial variations in uplift rate across the Channel region, so in this analysis it is assumed that this value has remained spatially constant during the last 0.5 MA. In Figure 7-6, the eustatic curve of Siddal *et al.* (2003) has been modified to include this uplift rate – this therefore provides a best estimate of the relative sea-level history of the Eastern English Channel over this period of time.

Lericolais *et al.* (2003) argue that the English Channel was only fully exposed subaerially on four occasions during the Late Quaternary, during the severe glacial periods of MOIS 16, 12, 6 and 2. If the hypothesised MOIS12 age for the flooding of the Strait of Dover is correct, then any incision in MOIS16 might have been less than that of MOIS12, 6 and 2, on the grounds that the combined Rhine-Thames River did not flow through the Strait of Dover at this time. Moreover, calculation of the relative sea-level for the English Channel (Table 7.3) indicates that the lowstand of MOIS16 reached only *circa* -40 m. This estimate is some 60 m less than that based on the simple use of the eustatic curve of Shackleton (1987) and implies that, in contrast to the suggestion of Lericolais *et al.* (2003), rockhead incision during MOIS16 would most likely have been wiped out by the subsequent (MOIS16-15) transgression of the shelf and, in particular, by the major incision related to the breaching of the Strait of Dover during MOIS12.

The flooding of the Strait of Dover in MOIS12 was associated with massive incision (Smith, 1985). The reconstructed relative sea level in the region at this time (Table 7.3) suggests a value of *c.* -65 m. It seems unlikely that the lowstand incision, associated with readjustment of the fluvial network to lowered base levels, would have significantly increased the initial incision associated with the breaching. Indeed, it is likely that following the initial breach, the major river systems in the eastern English Channel adopted the course of the breach-related valleys and did not incise much further (if at all) during the rest of the lowstand of MOIS12.

Lericolais *et al.* (2003) argue that during MOIS10 and MOIS8 little if any fluvial incision occurred on the sub aerially exposed sea floor of the Channel, since sea-level did not fall to a sufficient level (-70 m). However the data presented in Figure 7-6 and Table 7.2 indicate that relative sea-level was likely to have fallen below the -70 m isobath, if only for a short period of time. As yet, no incision or deposits related to these periods have been recognised in any offshore records (seismic or stratigraphic). This may reflect their non-deposition or subsequent erosion, or that the lowstand was insufficient to cause widespread fluvial deposition. Alternatively, it may simply reflect the fact that at present very little is known about the offshore sedimentary record in the region and that deposits of this age await discovery.

During the successive highstands and lowstands of the Late Quaternary, the shifting coastline of the Channel would have been associated with migrations of the Channel River delta. As has been discussed above, this reworking is the process by which the palaeovalleys in the PhD study area were infilled with marine transgressive deposits as the Channel River delta retreated upstream towards the Strait of Dover during the MOIS6-5e transgression. However, with the exception of the palaeovalley infills, there is very little evidence of deltaic type sedimentation on the Channel floor. One theory for this situation is that deltaic deposits by their nature are generally fine grained and are therefore more susceptible to reworking and winnowing by transgression and / or by tidal currents under highstand conditions. Furthermore, the predominant lack of contemporary sedimentation, in particular fine grained sediments, on the present floor of the Channel (apart from the sand banks to the east) can be explained by the strong sea bed currents and net transport directions away from the Channel shelf towards the Western Approaches in the west or through the Strait of Dover to the southern North Sea (Johnson *et al.*, 1982). If the Holocene is at all typical of previous interglacials, then highstand sedimentation on the Channel shelf during the Late Quaternary is likely to have been very limited. Indeed, most Channel floor sediments are coarse grained and are of either of lowstand fluvial origin or (more likely), have been reworked by marine processes during the rising limb of the postglacial sea level history. This model supports the arguments of Curry (1989) that the transgression and unconformity of the English Channel sea floor reflects the fact that the ultimate depositional area for the Channel has moved well offshore to the Celtic Sea Bank and Armorican Fan (see above).

The relative sea-level reconstruction for the study area (Figure 7-6) suggests that during MOIS6, sea-level was approximately -100 m bpsl. Therefore, according to the hypothesis of Lericolais *et al.* (2003), one would expect major fluvial incision of the palaeovalleys of the PhD study area associated with readjustment of the Rhine-Thames river system to this lowered base-level. It was argued earlier that during the subsequent transgression (MOIS6 / 5), the incised palaeovalleys of the PhD study area were infilled with paralic deposits as the shoreline retreated back across the shelf. Following the MOIS5e highstand there was a *circa* 40,000 yr (MOIS5e to MOIS5a) period of overall regression as relative sea-level fell from -20 m to -60 m bpsl (Figure 7-6). This interval would have been associated with heavy erosion of the exposed rockhead surface and reworking of the uppermost portions of the palaeovalley infills. Evidence of this erosion and reworking is, I suggest, preserved in SFU-G deposited over much of the PhD study area. This erosion and reworking could have radically changed the morphology of the sea bed in the eastern English Channel and the Strait of Dover, destroying any relic features of the former fluvial network (from MOIS6). This hypothesis suggests that during the subsequent major lowstand of MOIS2, a reinvigorated combined Thames-Rhine-Meuse River system in the eastern English Channel is more likely to have adopted a new course, i.e. the major channel that trends in an east to west direction to the north of the PhD study area (Palaeovalley A, Figure 7-2).

Uplift of 10 m since MOIS6 is proposed for the area and relative sea-level at this time is predicted to have been -117 m bpsl. Therefore incision of the channel to an average of -70 m below the sea bed (locally 80-90 m) is easily reconciled. The transgression from MOIS2-1 appears very similar (in time and magnitude) to that of MOIS6-5e (Figure 7-6) when, it is proposed, the palaeovalleys of the PhD study area were infilled with shoreface type deposits. It is considered likely, therefore, that under these conditions one would expect a similar set of deposits to be infilling the feature. However without data to substantiate this claim it is recognised that this is purely speculation.

To summarise, this section has shown that it is probable that the major glacio-eustatic fluctuations of the Late Quaternary have had a profound effect on the morphology of the eastern English Channel. However, the simple application of a eustatic curve, without accounting for crustal uplift of the region, can lead to erroneous interpretations. Lericolais *et al.* (2003) suggested that it was probably only during major lowstand periods (MOIS16, 12, 6 and 2) that significant fluvial incision is likely to have occurred on the shelf. Moreover, the magnitude of relative sea-level fall needs to be considered in the wider palaeogeographic evolution of the study area. In particular, pre-MOIS12 the rivers systems were limited in their size and probably did not heavily incise the shelf. In contrast, after MOIS12 they were much larger fluvial systems capable of incising more extensive palaeovalley systems. Finally, this discussion has demonstrated that erosion and deposition associated with transgression and, in particular, regression, may have had a greater degree of control on future patterns of fluvial incision than relic patterns of former incised-valleys on the shelf.

7.5 A Model of English Channel Shelf Evolution in the Late Quaternary

This Chapter concludes by proposing a model for the Late Quaternary evolution of the English Channel basin, based on the discussion presented above. The model describes the processes that operated on the Channel during two modes of activity that typified conditions during the pre-breaching (MOIS12) and post-breaching (post MOIS12) of the Strait of Dover. There are three main periods of activity during each glacial-interglacial cycle, characterised by lowstand, transgressive and highstand conditions. Where possible, suggestive MOIS numbers are referred to for different aspects of the model, but it is worth repeating that the chronology proposed here is severely constrained by the lack of absolute age determinations from this PhD study, and other previous research in the region.

7.5.1 Pre-Breach (e.g. MOIS16-13)

Lowstand

Glacio-eustatic sea-level fall promoted the gradual extension across the Channel floor of the Rivers Somme, Seine and Solent. As relative sea level fell further, so these rivers eventually joined to form the "Channel River" that flowed to the shelf edge where it fed an outer shelf fan. The supply of sediment came from a closed 'sediment conveyor' system with terrigenous sediment delivered from the catchments of southern England and northern France. In the southern North Sea the Thames and rivers of the European Low Countries flowed out onto a large deltaic system (the Yarmouth Roads Formation). Coarse grained gravels accumulated in the active channels and as marginal terraces associated with a range of rivers of different size. Most gravel remained stored on the shelf, with limited material reaching the outer shelf fan, where sand – silt grade material was deposited.

Transgression

Climate amelioration at the end of MOIS13 was associated with a rise in eustatic sea-level rise that caused extensive planation of the Channel floor. Lowstand fluvial gravels were eroded from their terraces and channel fills, only locally being preserved in palaeovalley settings where sufficient accommodation space existed. Most gravel was transported landwards by the transgressing sea to form coastal barriers of gravel along the English and French coasts. In the Strait of Dover area, shoreface erosion of the periglacially weathered rockhead surface occurred in the embayment that then existed. Fine-grained sediments were transported to the shelf edge to supply the continental slope and deep-sea fan depositional systems.

Highstand

The highstand associated with MOIS13 is recorded in nearby onshore areas at Boxgrove (Roberts. 1999) and in the Steyne Wood Clay of the Isle of Wight (Preece *et al.* (1990), although the exact age of these deposits is debated), which show that highstand sea-level reached *c.* +40 m relative to present. The rivers of southern England and northern France retreated to their minimum shoreward extents as the English Channel becomes a shelf sea inlet (up to 60 m) deeper than present. Strong tidal sea-bed currents reworked sea bed sediments by winnowing fine sediments away and leaving a ubiquitous veneer of coarse lag deposit. Sediment delivery to the continental slope and deep sea fans was negligible, and slow accumulation of pelagic sediments dominated. In the central and southern North Sea deltaic and shallow marine conditions prevailed.

Regression

Little is known regarding the nature of the regression from MOIS13. Depending on the magnitude and rate of the regression, a combination of nearshore marine and subaerial processes would have reworked the MOIS13 coastal and seabed highstand deposits resulting in a tendency towards planation of the Channel floor. The lowstand was associated with a transition from temperate to cold stage conditions, and one would expect a full range of processes to operate that today are recorded in sub-arctic environments (e.g. seasonal sea ice, possible ice rafting, periglacial processes, loessic deposition).

7.5.2 Breaching of the Strait of Dover (MOIS12)

Lowstand

The severe cold of MOIS12 combined with the low relief surface of the central and southern North Sea delta plain to enable the spread of the British and Scandinavian ice sheets and their eventual coalescence. With their northerly route blocked, the rivers of the European Low Countries and the Thames of southern England fed into an extensive proglacial lake that extended across much of the southern North Sea Basin. Lake levels rose to eventually overtop an outlet col on the Wealden chalk ridge that joined Britain to France. This breaching may have been catastrophic (Smith, 1985) or more gradual (Gibbard, 1988), with large volumes of water discharging into the English Channel. Major incision of the floor of the eastern English Channel occurred, incising the Lobourg Channel and Southern and Northern Palaoevalleys. Previous sediment cover across the floor of the English Channel was eroded. In some instances, earlier sediment infills in palaeovalleys was completely removed, but in others partial erosion of the palaeovalley fills resulted in the creation of large hiatuses. The discharge of the lowstand rivers draining France, notably the Seinne, greatly increased. This event was associated with a massive input of sediment to the Channel shelf edge, and the deep sea fans recorded a significant coarsening –upwards sequence. With the breach of the Strait of Dover, the Yarmouth Roads Formation ceases deposition with sediment flushed through into the Channel River system and on to the depositional fan system of the Western Approaches.

7.5.3 Post-breach (MOIS11-present)

Transgression

Sea-level rise and associated shoreface erosion planated the former sea floor and areas of significant vertical relief (terrace deposits), and eroded the open channels created by the breaching of the Strait of Dover. Fine grained sediments were reworked and transported away to the shelf edge depocentre which received a massive pulse of sediment during this time, up to 400% more than during lowstand periods (Bourillet *et al.*, 2003), largely as a result of reworking of lowstand alluvial deposits from the shelf.

During the transgression from MOIS6-5E and MOIS2-1, the shoreline retreated and fluvial conditions that were responsible for incising the valleys of the Channel River were replaced by shallow marine conditions that promoted infilling of the palaeovalleys with sediments. Later on in the transgression, the upper surface of these newly formed deposits was truncated by erosion as the shoreface migrated over these areas.

Highstand

The rivers combining to form the Channel River retreated to roughly their present shoreward extent and a shelf sea developed. However, in contrast to conditions pre MOIS12, this sea was connected to the southern North Sea *via* the Strait of Dover. Continued winnowing of sea bed sediments by (probably stronger) tidal currents resulted in an extensive coverage of coarse lag deposits on the eastern English Channel sea floor. The winnowed sediments were transported to the southern North Sea and into the Channel, and nourished highstand deposition at the shelf edge throughout these areas. Long term uplift of the eastern English Channel preserved these beach and tidal channel deposits on the coastal plains of southern England and Northern France (Bates *et al.,* 1999).

Regression

The major regression phases following MOIS12 were MOIS10, 8, 6 and 2. Each of these was associated with varying degrees of channel floor exposure and subaerial reworking (see above). Within the PhD study area, the transition from MOIS5e to MOIS4 is the only regressive phase recognised. This interval was associated with an overall fall in sea level of *circa* 110 m over a period of *circa* 55,000 years, during which the forced regressive ravinement surface R_1 developed, and the accumulation of Depositional Sequence II began.

Lowstand

The drainage basin of the Channel River more than doubled in size as the large rivers of northwest Europe (e.g. Rhine, Meuse and Thames) extended across the floor of the subaerially exposed North Sea and flowed southwards (possibly forced by recurrent coalescence of the British and Scandinavian ice sheets, Dr David Evans, *pers. comm.*) through the Strait of Dover and onto the Channel shelf during the subsequent lowstand cold periods (MOIS10, 8, 6, and 2). The sediment supply to the shelf edge and deep-sea fans received an order of magnitude more sediment than during highstands (Bourillet *et al.*, 2003). Assuming relative sea-level dropped to -70 m bpsl (MOIS2, 6 and possibly 8 and 10; Siddal *et al.*, 2003; Figure 7-6), then the Channel River incised into the bedrock shelf and created an extensive system of (relatively) narrow incised palaeovalleys found on the floor of present eastern English Channel.

Chapter 8: Conclusions

8.1 Introduction

In Chapter one, the overall aim of the thesis was defined: "*to develop a model for the Late Quaternary evolution of the eastern English Channel based on an investigation of the palaeovalleys of the region*".

To address the above aim, this thesis has sought to achieve the following:

1. *The collection and interpretation of new high resolution seismic data from the eastern English Channel using the concepts of seismic and sequence stratigraphy.*
2. *The retrieval and analysis of sediment cores from the study area to determine their physical characteristics and depositional origins.*
3. *Comparison of these data with other records collected from analogous continental shelves elsewhere in the world.*
4. *To develop of a hypothetical age model for the depositional sequence in the PhD study area.*
5. *To test and develop further this model through its comparison with depositional sequences from elsewhere in the English Channel and the southern North Sea Basin.*

This chapter concludes this thesis by reviewing the main findings of the work in the context of the original aim stated above. It also reflects on the strengths and weaknesses of the work and makes recommendations for future research in the English Channel.

8.2 Main findings of the PhD research

1. *The collection and interpretation of new high resolution seismic data from the eastern English Channel using the concepts of seismic and sequence stratigraphy.*

This PhD is based on an extensive set of seismic data that were aquired prior to the start of this research by Hanson Aggregates Marine Limited. The two areas, from which the seismic data were collected are located in the eastern English Channel and were initially selected to assess the aggregate resource potential of the region. The sea bed geology of the study area includes several major palaeovalleys, and the seismic data therefore provided an excellent opportunity to examine the Late Quaternary history of these features.

Within the PhD study area, over 500 km of high resolution shallow seismic reflection had been collected, and a major component of this thesis has been the analysis and interpretation of these data. The spacing between seismic lines was typically < 1 km and the data quality is of a very high standard. Analogue seismic traces were digitised from all 500 km of data, and interpreted using the concepts of seismic sequence stratigraphy.

Interpretation of the seismic data enabled the identification of a network of palaeovalleys, separated by three major sequence boundaries. The seismic profiles reveal three prominent, areally widespread, reflector surfaces. They are, from the base to the top of the section, the LQ_B, the R_l, and the H_b surfaces. Between these major reflector surfaces are four seismic sequences that comprise eight different seismic facies units (SFUA-H).

Seismic Sequence I comprises Tertiary bedrock (SFU-A) and is separated from the overlying Quaternary sediments (Seismic Sequences II and III) by a lower sequence boundary surface (LQ_b). Where Quaternary sediments are absent, the Tertiary bedrock outcrops at sea bed. The surface relief of LQ_b varies between -73 m -33 m CD and defines two major depressions interpretated as palaeovalleys that are infilled with between 20 and 30 m of Quaternary sediments.

Seismic Sequence II comprises the seismic facies units (SFU-B, C, D, E and F) between the major bounding reflector surfaces of LQ_b below and surface R_1 above. The facies units within this sequence are interpreted as aggradational transgressive marine deposits. These facies units are recorded in distinct depressions across the study area and are interpreted as forming at the same time and under the same general conditions. These units largely fill the palaeovalleys and account for the majority of the Quaternry deposits recorded in the study area. It is feasible that these facies units could represent different periods of sedimentation during one or more transgressive cycles, but their seismic geometries and signatures are sufficiently similar that, without independent chronological control, the simplest interpretation is the most valid.

Seismic Sequence III is defined at its base by a ravinement surface R_1, that formed under conditions of a forced regression and prolonged erosion of the study area by nearshore marine processes. The sequence includes one facies unit (SFU-G) that developed penecontemperaneously to R_1 under regressional conditions. SFU-G is interpreted as a coarse-grained marine deposit.

Seismic Sequence IV is the uppermost sequence recorded in the study area and comprises two distinct seismic facies units (SFU-H and SFU-I) that overly the major reflector surface H_b. The upper bounding surface of this sequence is the seabed. SFU-H is a localised palaeovalley infill covering an area of approximately 2.8 km^2. In contrast, SFU-I is found across all areas and varies in thickness from a few decimetres to a maximum of around 10 m (average 0.5 m to 1 m). It is interpreted as a Holocene lag deposit.

2. *The retrieval and analysis of sediment cores from the study area to determine their physical characteristics and depositional origins*

One hundred and eighty-nine boreholes were collected from the study area using ship deployed vibrocoring equipment. These data provided information on the near surface deposits and were collected specifically to assess aggregate resource potential focusing therefore on the coarser grained deposits in the study area. The shallow penetration of most cores (typically <6 m, often only 1-2 m) meant that lithological control for the seismic data is provided only for seismic facies units SFU-A, -C, -G, -H and -I. Grain size analysis on 747 samples provided an extensive dataset regarding the lithological characteristics of these sediments, which ranged between muddy sand and shelly gravel. One hundred and thirty eight samples from four of these facies (SFU-A, -G, -H and -I) were also analysed for their microfossil content. The lowest facies, SFU-A, contained no microfossils but the remainder all contained variable quantities of marine diatoms, foraminiferal test linings and sponge spicules indicative of deposition under marine conditions. No freshwater microfossils were recorded in any of the samples analysed. Lastly, multi element geochemical analysis of 14 rare earth elements was completed on 174 samples, in an attempt to determine possible defining geochemical signatures for facies units SFU-A, - G, -H and –I. This demonstrated that one facies, SFU-A, contained a unique geochemical signature associated with the Tertiary bedrock.

The combined seismic, lithostratigraphic, microfossil and geochemical data enable a reconstruction of the depositional origin of the four Seismic Sequences identified in the study area. Seismic Sequence I comprises Tertiary bedrock. Seismic Sequence II is made of unconsolidated sediments of Late Quaternary age that were deposited under shallow marine conditions and infill the palaeovalleys that are incised into the bedrock of the study area. Seismic Sequence III is a regressive unit deposited in shoreface conditions. Lastly, Seismic Sequence IV blankets much of the study area with a relatively thin lag of shelly gravel that also formed under marine conditions. Nowhere through these sequences is there any evidence for deposition under fluvial conditions. However, it must be noted that the absence of microfossils cannot be used as positive evidence, and not all seismic facies units were available for analysis from the vibrocoring programme.

3. *Comparison of these data with other records collected from analogous continental shelves elsewhere in the world.*

Previous work in the English Channel has identified the presence of an extensive system of palaoevalleys and infills on the sea floor to which the two large palaeovalleys identified in this PhD study belong. Most of these previous studies lack any form of absolute age control data and are limited in their palaeoenvironmental reconstructions through their reliance on seismic data alone or, at best, restricted lithological data. Nearly all of the chronostratigraphical and palaeoenvironmental interpretations made in these studies are therefore open to alternative interpretations.

Comparisons with other continental shelf records (in particular those from the US Atlantic continental shelf and the Gulf of Mexico) demonstrate that the Channel palaeovalleys bear some close similarities to these, but also some major differences. The similarities include lowstand incision and infilling, driven by global eustatic sea-level change, as well as the operation of fluvial and subaerial processes on parts of the exposed shelves during lowstand conditions. However, a unique characteristic of the Channel palaeovalley network is its hypothesised formation as a result of the (potentially catastrophic) drainage of a large proglacial lake from the southern North Sea Basin.

The newly acquired data from the PhD study area demonstrate several other close similarities to records from other continental shelves including the following:

- The morphology of the lower sequence boundary surface LQ_b, between the Tertiary rockhead and infilling Quaternary sediments, is closely mirrored in sequences recorded on the shelf of the US Atlantic seaboard, most notably around the large inlets of Chesapeake Bay (Foyle and Oertel, 1992) and Delaware Bay (Knebel and Circe, 1987).

- The geometry and lithology of the infilling deposits in Seismic Sequence II shows a close resemblance to the sequences in the infilled palaeovalleys of offshore extensions of the Colorado and Brazos Rivers on the inner shelf of the Gulf of Mexico (Anderson *et al.*, 1996). In particular, the size and geometries of units and the nature of the infilling deposits bear a close resemblance to the Channel records.

- The sequence boundary surface H_b of Seismic Sequence IV are also comparable to the uppermost sequence on the southern Delamarva Peninsula in Chesapeake Bay examined by Foyle and Oertel (1992). For example, these authors identify a Late Devensian to Holocene Type 1 Sequence Boundary that displays similar low-relief to H_b.

4. *To develop of a hypothetical age model for the depositional sequence in the PhD study area.*

A major hindrance in this and previous work in the English Channel is the lack of absolute dating control. Therefore, a hypothetical age model for the depositional sequence recorded in the PhD study area was developed based on an interpretation of the seismic and lithostratigraphic data in comparison with a global sea level curve, and the concepts of sequence stratigraphy.

As discussed above, the deepest surface recognised in the PhD study area is the sequence bounding surface LQ_b which is a major (Type 1) Sequence Boundary that separates the Tertiary bedrock from the overlying paleovalley deposits of the Late Quaternary.

Most authors consider that the main palaeovalley network of the English Channel developed during the drainage of the proglacial lake in the North Sea Basin during MOIS12, overprinted on the pattern of the former fluvial systems of the pre-breach Channel floor. This drainage resulted in the formation of the major palaeovalley systems of the area – the Lobourg Channel, the Open Northern Palaeovalley and the Medial / Southern Channels. The palaeovalleys identified in this study are not directly connected to the Open Northern Palaeovalley, and have a higher channel base compared to this system. On these grounds, it is suggested that the surface LQ_b in the PhD study area was cut in the later cold lowstand associated with MOIS6, when relative sea-level in the English Channel fell at least 70 m below present. It is possible, that the surface LQb formed earlier, but relative sea-level fall was greater in MOIS6 compared to MOIS8 and 10, and therefore the incision is most likely to date from this period.

Seismic Sequence II infills the palaovalley incisions of LQ_b and is interpreted as a transgressive sequence deposited during the MOIS6 to MOIS5e climate amelioration and associated sea-level rise. It is hypothesised that ravinement surface R_l formed during the major regression from MOIS5e to MOIS4 and that the erosive forces responsible for the formation of this surface also destroyed any remnants of a transgressive ravinement surface and any associated sediments deposited during the MOIS5e highstand. Seismic Sequence III was laid down penecontemporaneously to surface R_l under a high energy depositional regime. The surface of Seismic Sequence III was subaerially exposed throughout to period of MOIS4 to MOIS2 (the Last Glacial Maximum). During this long interval of cold conditions the study area would have been subject to extensive periglaciation and palaeosol development, with the likely accumulation of a thick deposit of loess. The erosive forces associated with the postglacial transgression from the end of the LGM into the Holocene removed these cold stage deposits and created the sequence boundary surface H_b. Since this time, the uppermost sequence found in the study area accumulated as a Holocene sediment lag. The surface of this sequence represents the sea bed in the majority of the PhD study area and is currently still undergoing formation, and will potentially be interpreted as a submarine highstand deposit of Holocene age in geological records of the future.

5. *To test and develop further this model through its comparison with depositional sequences from elsewhere in the English Channel and the southern North Sea Basin.*

Comparisons with records from the immediate eastern English Channel show that features of similar dimensions (depths and widths) and fill types are apparent elsewhere in the region.

The palaeovalleys of the English Channel are either infilled, as is the case in the PhD study area, or open and lacking in sediment (Hamblin *et al.* 1992). It seems likely that the difference between these channel types is a function of age, with the older open channels probably associated with the initial drainage of the North Sea proglacial lake, and the smaller and narrower infilled valleys developing later. This hypothesis is compatible with the age model developed above for the PhD study area, which proposes a MOIS6 / 5e age for the valley incision and onset of infilling. Such a suggestion agrees with evidence from elsewhere in the Chanel for nested palaeovalley systems that differ in age (e.g. Lericolais *et al.*, 2003).

In terms of dimensions, the palaeovalleys in the PhD study area are typically between 2 and 8 km in width. They must, therefore, have been capable of supporting a large river system during lowstand conditions, and been major estuaries during periods of both the transgressive and regressive limbs of glacial-interglacial sea-level oscillations in the region. Assuming their original incision was driven by fluvial processes, it seems unlikely (given their location) that they originated as offshore extensions of any of the rivers that drain the English or French Channel coasts. A more likely origin for their formation is that they were created by one or more of the Thames, Rhine or Meuse rivers during regressive and lowstand conditions.

Comparisons with records from the wider region demonstrate the importance of interpreting the palaeovalleys of the eastern English Channel in the context of the combined Channel and North Sea Basin systems. For example, in Chapter 7 is was noted that the end of the Yarmouth Roads Formation, a major Late Quaternary deltaic deposit in the North Sea Basin, may date to MOIS12 and the re-routing of the Thames / Rhine / Meuse systems into the Strait of Dover and English Channel. Equally, the fine-grained depocentres represented by the Celtic Sea Fan and the Armorican Fan, should preserve evidence for a major increase in sedimentation at this same time. Testing these hypotheses through the absolute dating for the Yarmouth Roads Formation and collecting long cores from the shelf edge fans of the Western Approaches is clearly required.

Comparisons with other work in the English Channel and beyond demonstrate the importance of two main processes that have controlled the evolution of the English Channel during the Late Quaternary - tectonic movements and eustatic fluctuations.

Tectonic movements, in particular long-term crustal uplift of the Eastern English Channel, allied with on-going subsidence of the North Sea Basin, are widely thought to exert a strong control on the physical evolution of the English Channel and its associated fluvial and relative sea level history (Preece *et al.*, 1990; Jones, 1999a, 1999b; Lagarde *et al.*, 2003). Preece *et al.* (1990) and Lautridou (1985) and Lagarde *et al.* (2000) suggest typical long-term uplift rates of 10 m per 100,000 years, although this rate is likely to have varied spatially across the region. For example, differential uplift is one of the possible causes of the Thames Rhine-Meuse adopting a new course across the eastern English Channel shelf (Palaeovalley A) during MOIS2 as opposed to reoccupying its hypothesised MOIS6 course through the PHD study area.

Previous workers have generally taken global eustatic curves and applied them directly to the English Channel. In this work, one such curve, developed by Siddall *et al.* (2002) from the Red Sea, is corrected for long-term uplift in order to produce a relative sea-level curve for the region that underpins the age model developed above. This demonstrates that during (at least) the lowstands of MOIS6 and MOIS2 major fluvial incision, by a combined Thames-Rhine-Meuse River flowing through the Strait of Dover, occurred on the shelf in response to base level fall. During the subsequent transgression these incisions were backfilled. It is also hypothesised that the erosive forces associated with transgression and regression, driven by the fluctuating sea-levels of the Late Quaternary, have played an equally important role to tectonic uplift in controlling patterns of palaovalley incision during this time

8.3 Strengths and Weaknesses of the Research

During the course of this PhD a variety of factors have provided opportunities and created barriers to the research described in this thesis. Some of these are related to the logistics of collecting and interpreting a large seismic and lithostratigraphic database, others stem from decisions taken during the finite period of time provided by a programme of PhD research. It is appropriate, at this juncture of the thesis, to reflect on some of these factors in order to identify the key strengths and weaknesses of the work, not least since they provide important constraints on the main findings of this work, and also useful pointers as to possible directions of future research.

The main strengths of the research described in this thesis include the following:

1. The study was undertaken as a NERC Industrial CASE studentship between the University of Durham, Hanson Aggregates Marine Limited and the British Geological Survey, which enabled an unparalleled wealth of knowledge and experience to be drawn upon from colleagues working in the industry, a Government research centre, and a University research department. Training has been received from each of these partners in different aspects of the PhD programme.

2. Hanson Aggregates Marine Limited kindly provided access to a large dataset that is, under normal conditions, considered commercially sensitive. It would not have been possible, under a normal PhD, to collect the amount or quality of the data analysed during the course of this work. The research was helped by the fact that all of the seismic data had been collected before the PhD began, which meant no time was spent in the original collection of these data.

3. The seismic dataset is of a significantly greater spatial resolution than that reported in previously published work from in and around the PhD study area in the eastern English Channel. This datset was augmented by access to a large and dense

network of vibrocore samples, the like of which are rarely described in previous published work from the region.

4. A further strength of this work has been the development of a new age model for incision and infilling of the palaeovalley system in the eastern English Channel. This model has been made possible by comparing the seismic data with other records from the region, and relating these to a modified global eustatic sea-level curve. The model has close links to, and implications for, the wider development of the southern North Sea and English Channel during the Late Quaternary.

The following weaknesses in the PhD research study are also identified:

1. The thesis lacks any absolute chronology from the PhD study area. This is because there was no material suitable for dating, beyond the shells found in the post-glacial lag (SFU-I), which was not the main interest in the work. This means that the chronostratigraphic model proposed must be viewed as provisional until further independent age determinations become available. The lack of absolute dating control means that there is a tendency for different authors, in different study areas of the English Channel, to develop their own age models based on their particular assumptions regarding uplift, eustatic variations etc. Integrating these into a wider model for the region requires an extensive dating programme (see below).

2. The study was limited by the maximum recovery of vibrocore sediment samples to only six metres beneath the seabed. Therefore the widespread sedimentary sequences seen on seismic profiles to be infilling the palaeovalleys in the PhD study were not extensively sampled to aid with further interpretation. Most of the subsamples from the vibrocores come from Seismic Sequences I, III and IV, and very little data are available for Seimic Sequence II.

3. The seismic profile data obtained were in analogue format. They required laborious digitising by hand, a process, which by its nature is subjective and open to

misinterpretation. Certainly a digital seismic dataset would enable a more robust classification of the different seismic facies units proposed here.

4. The palaeoenvironmental proxies used were not particularly helpful in the interpretation of the depositional origin of the sampled facies. In particular, dissolution or non-deposition of diatom valves meant that attributing a positive depositional environment to all facies has not been possible. The presence of foraminiferal test linings is a help, but they may be reworked from Tertiary as well as Quaternary sediments. Further palaeoecological work would be possible, such as pollen analysis of some of the finer-grained sediments, but this was beyond the scope of this research.

5. The data obtained for the study were limited in their spatial extent. However it is recognised that with the limited time available for the study there is a difficult play-off between obtaining detail at the local scale and generality at the wider scale.

6. It is clear that there is considerable complexity in the erosional and depositional record of the palaeovalley systems of the eastern English Channel. In particular, multiple incision and infilling events are the norm and therefore hiatuses and ravinements are prevalent across the system. These add considerable complexity to the palaeoenvironmental reconstruction, and make comparisons within and between sites challenging.

7. The dataset used in this thesis was driven by interests within the aggregates industry, and the selection of sampling locations was influenced by economics as oppose to any particular academic rationale. There is, therefore, a strong bias in the data towards areas and sites which are perceived as potentially valuable from an aggregate potential.

8.4 Recommendations for Future Work

This final section makes suggestions for the possible direction of future work, both in the eastern English Channel and also in the context of the wider evolution of the English Channel Basin and efforts to reconstruct Quaternary climate change.

This study has shown that one of the incised-valleys in the palaeovalley systems of the eastern English Channel contains a depositional sequence that is characterised by multiple cut-and-fill events of hypothesised Late Quaternary age. Whilst analysis of the seismic data obtained in the present study enabled features to be resolved at depths of up to 30 metres beneath the seabed the maximum vibrocore penetration was only six metres. Therefore a programme of deeper coring in the future would provide more detailed information as to the nature and composition of the deposits interpreted from the seismic data alone. In particular, deeper cores would be particularly useful if they secured samples from the greatest depth of thalweg incision, and the deposits therein. in the main N-S running palaeovalley in the study area. This would potentially yield samples for dating that, in turn, would facilitate corroboration or rejection of the proposed age model for palaovalley development. The deposits obtained would also provide useful material for further litho- and biostratigraphical analyses for palaeoenvironmental interpretation.

Further acquisition of detailed seismic and sedimentary data in areas upstream of the PhD study area would help to further elucidate the relationship with the palaeovalleys in the PhD study area and Paleaovalley A. In particular, close sampling of the area where Palaeovalley A appears to cross cut the upstream continuation of the main paleovalley in the study area would provide important information relating to the relative ages of these systems.

An urgent research priority should be the collection of long cores from the depocentres of the Celtic Sea Fan and the Armorican Fans. The potential of these areas for securing a continuous and high-resolution record of the Late Quaternary palaeohydrology of the English Channel, and by implication much of Northwest Europe, is significant. Indeed, records from these locations must have the potential to become global standards for the analysis of Quaternary land-ocean interactions.

Marine aggregates industry research surveys are a potentially major untapped source for Quaternary research in particular in an offshore setting. To undertake research offshore requires the collection of extensive data sets that to acquire are very expensive. Therefore because of this expense, offshore deposits in comparison with their onshore counterparts are poorly understood. Aggregate industry research surveys are, by their nature, biased in favour of the analysis of coarse-grained deposits, in the upper few metres of the sub seabed geology. However, during this study it has been shown that deposits of up to 30 m below the seabed are still recognised on seismic profiles and can be used for detailed mapping and interpretation of depositional sequences. It is clear that the relationship between Quaternary scientists and the marine aggregates industry should be fostered further so as to maximise the research potential of this wealth of unexplored data.

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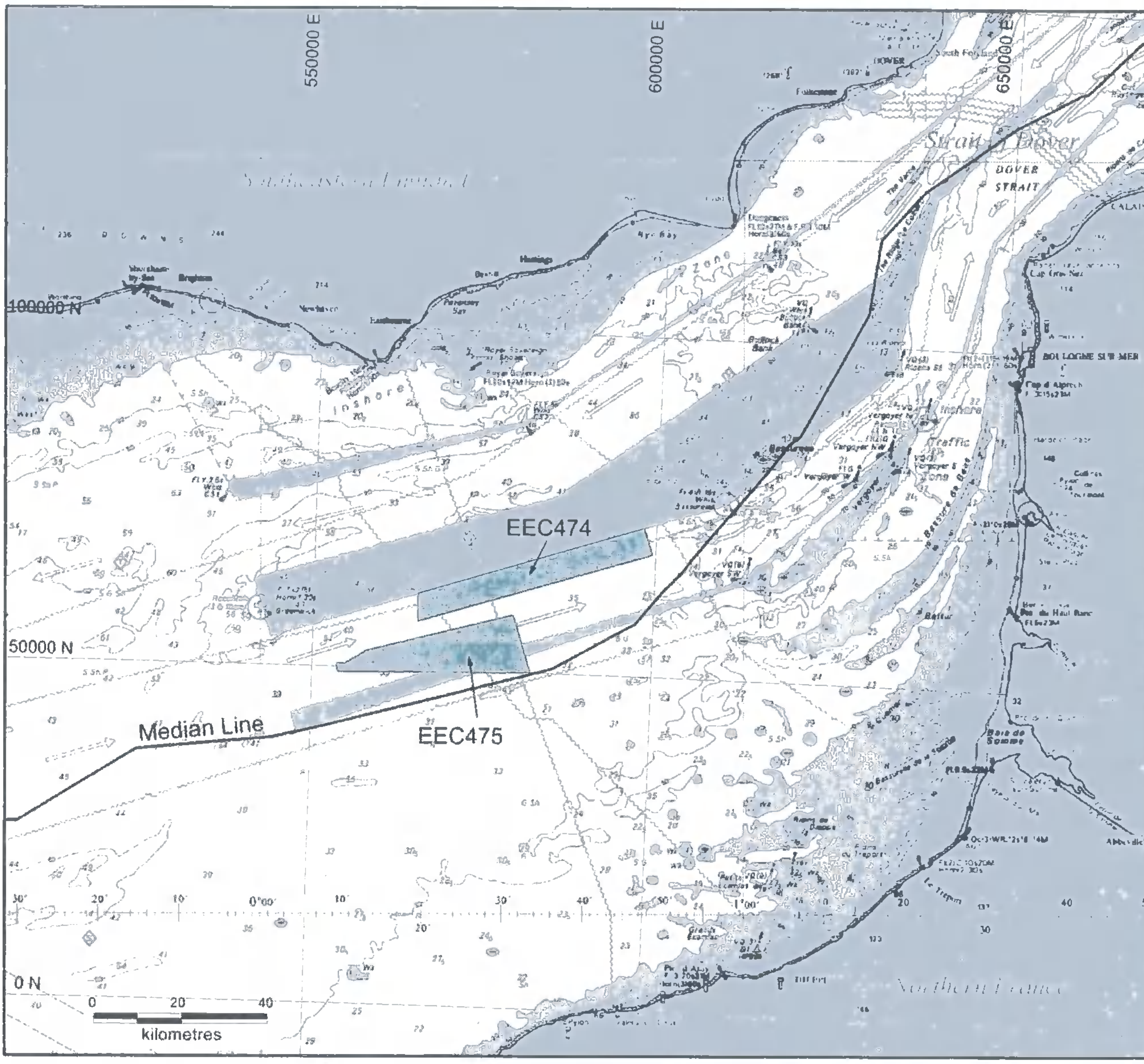


Figure 1-1: Location of the PhD study area. Licence Application Areas EEC474 and EEC 475. Coordinates are OSGB.

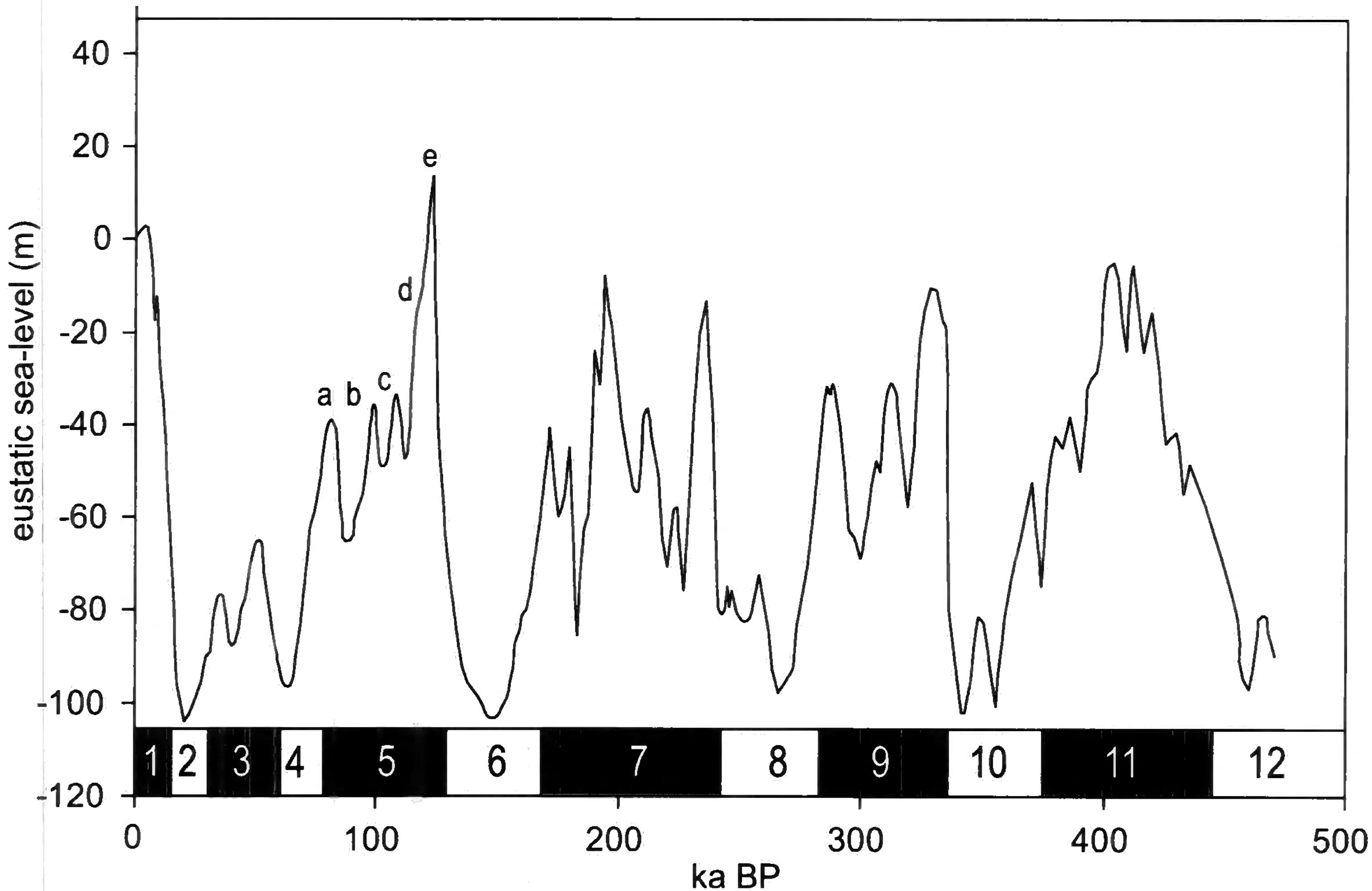


Figure 2-1: Late Quaternary eustatic fluctuations. Eustatic sea-level is derived from an amalgamation of Siddal *et al.*'s (2003) model reconstructions for Red Sea cores KL11 and MD921017. Numbers represent marine oxygen isotope stages (MOIS) (Shackleton *et al.*, 1990) even numbers represent cold glacial sea-level lowstand periods and odd numbers temperate highstands.

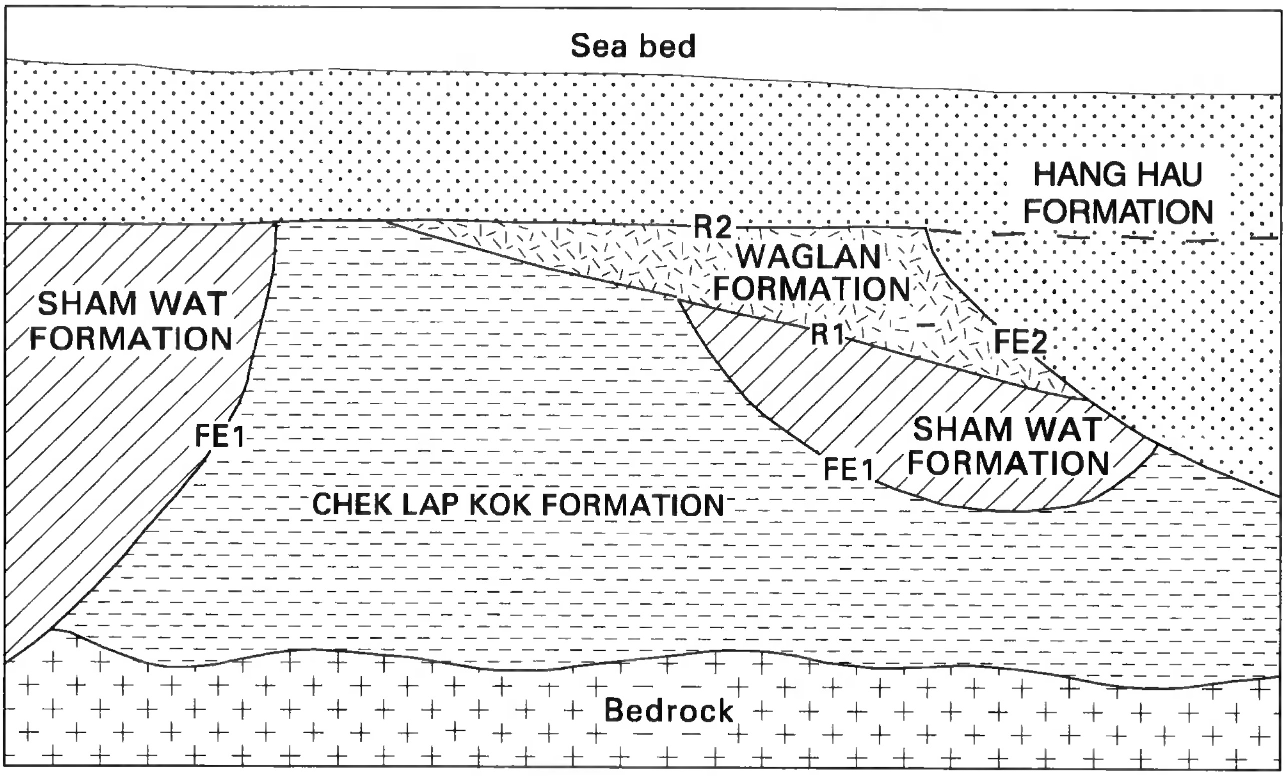


Figure 2-2: The Hong Kong Geological Survey sequence stratigraphic model for the Quaternary offshore sediments of Hong Kong (modified from Fyfe *et al.*, 1997). Age relationships of Fluvial Entrenchment surfaces (FE1 and FE2) and Ravinement Surfaces (R1 and R2) with the deposits are shown in Figure 2-3.

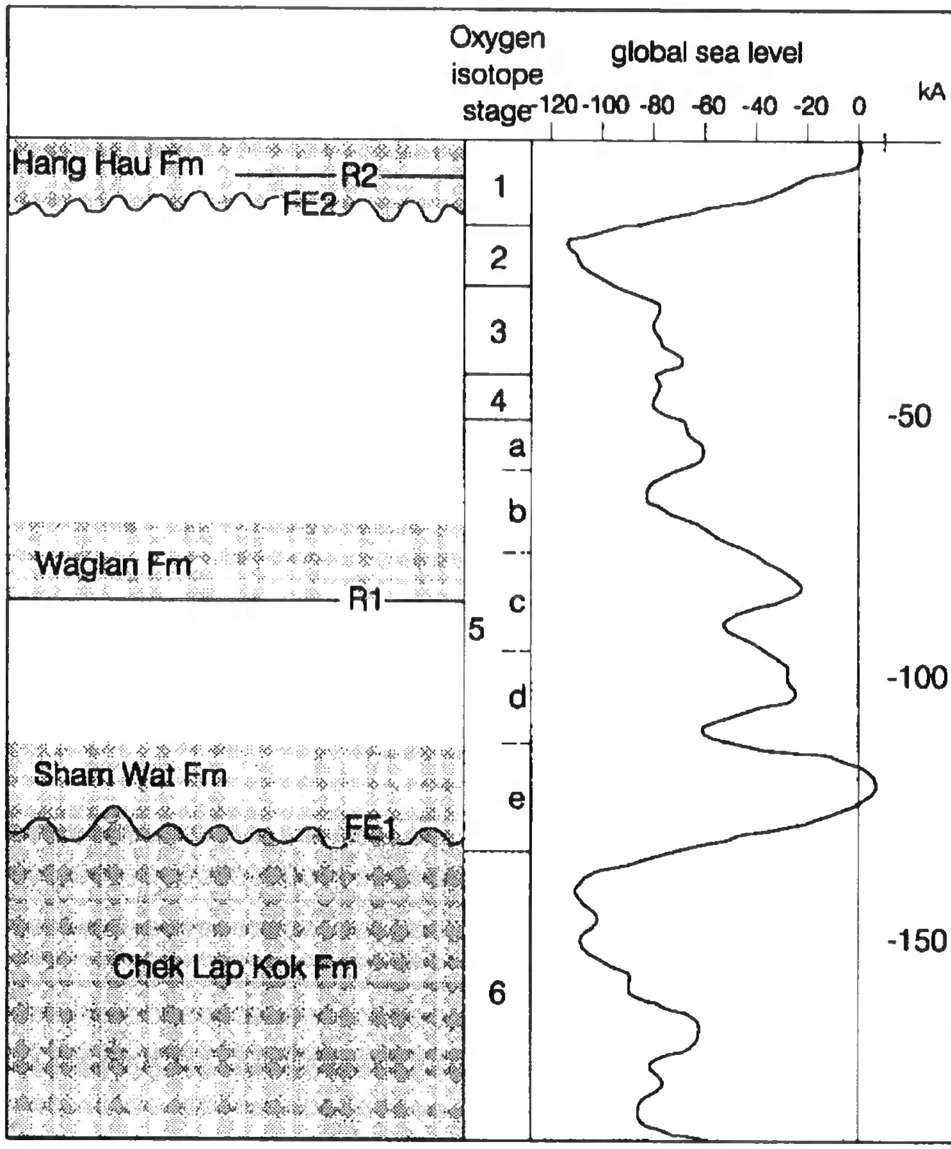


Figure 2-3: Quaternary offshore formations, Hong Kong. Ravinement (R) and fluvial entrenchment (FE) surfaces compared with oxygen isotope stage and global sea-level record (Chappell and Shackleton, 1987), after Fyfe *et al.* (1997)

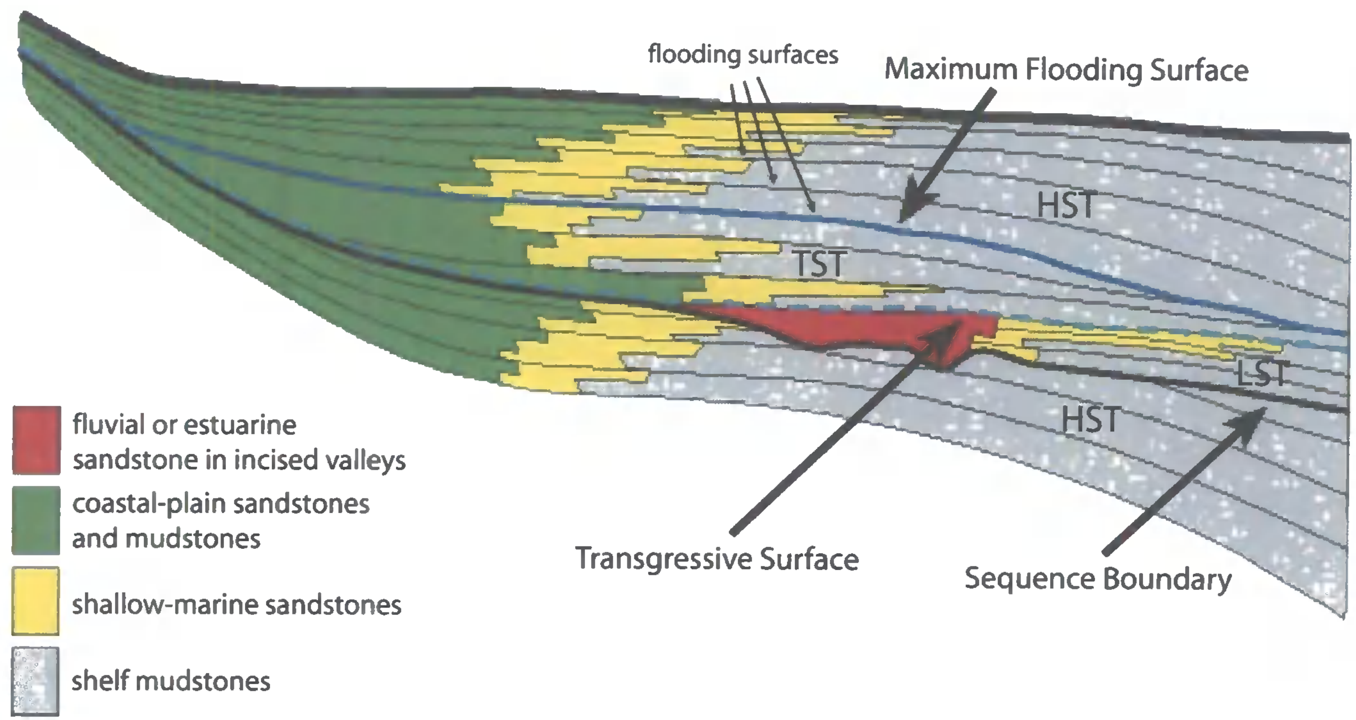


Figure 2-4: Schematic diagram of an idealised depostional sequence. Showing the highstand (HST), lowstand (LST) and transgressive (TST) systems tracts (from UGA stratigraphy Lab, http://www.uga.edu/strata/sequence/tracts.html)

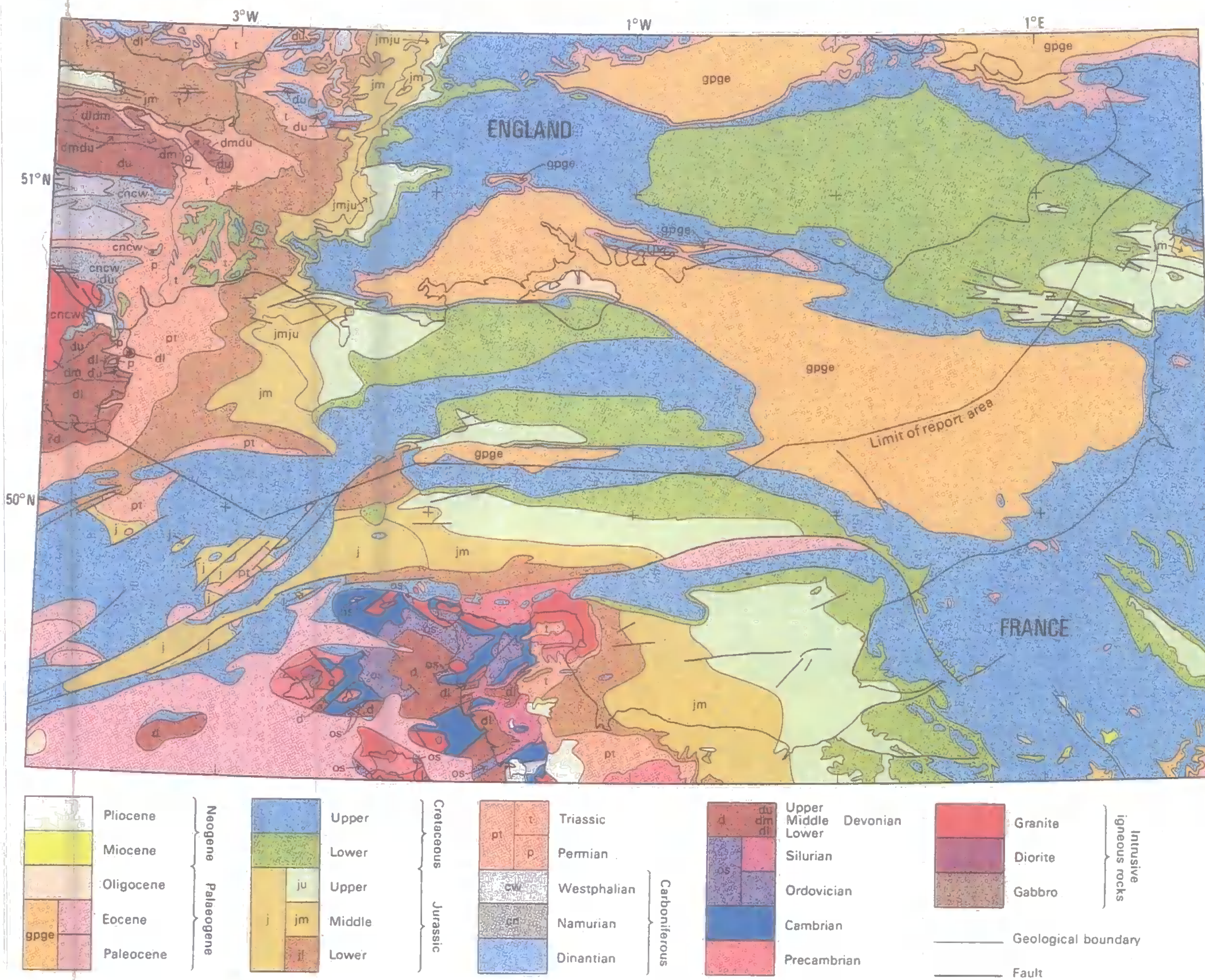


Figure 3-1: Geological map of the English Channel (from Hamblin *et al.*, 1992)

Era	Period		Epoch	Age (Ma)		Lithostratigraphic unit		Major events
CENOZOIC	QUATERNARY		HOLOCENE			(Sea-bed sediments)		Flandrian transgression
			PLEISTOCENE		0.01	(Palaeovalley infill sediments)		Palaeovalleys cut by fluvial action during glacial maxima
					1.67			
	TERTIARY	NEOGENE	PLIOCENE					Marine transgression (no deposits survive offshore)
					5.2			
			MIOCENE					Formation of cliff line now submerged at -38m to -49m OD
								Marine transgression (no deposits survive)
								Main Alpine orogeny, major structural inversion
					25.2			
		PALAEOGENE	OLIGOCENE	Chattian				
					30			Sea-level fall associated with initiation of Antarctic ice-cap
				Rupelian		Bouldnor Formation		
					36			Lacustrine and brackish-lagoonal sedimentation with brief marine incursions
			EOCENE	Priabonian		Headon Hill and Bembridge Limestone formations		
					39.4			
				Bartonian	42	Barton Group		
				Lutetian		Earnley Sand, Marsh Farm and Selsey Sand formations		Marine and brackish sedimentation
				Ypresian	49	London Clay & Wittering formations		
					54			Paris Basin and North Sea linked
			PALEOCENE	Thanetian		Woolwich & Reading formations		Marine transgression from north
					60.2			
				Danian				
					66.5			
MEZOZOIC	CRETACEOUS		LATE (Senonian)	Maastrichtian				Uplift, minor inversion of structure begins
					74			
				Campanian				
					84	Upper Chalk		
				Santonian				Most of Europe submerged
					88			
				Coniacian				
				Turonian	89	Middle Chalk		
				Cenomanian	92	Plenus Marls		
						Lower Chalk, Chalk Marl		North Atlantic continues to open
					96			
			EARLY	Albian		Upper Greensand		North Sea and Wessex-Channel Basin joined
						Gault		
					108			
				Aptian		Lower Greensand		Marine transgression and regional downwarping associated with initial opening of North Atlantic
					113			
				Barremian				
			Neocomian	Hauterivian	116.5	Wealden Beds		Active block faulting, rifting of continental margin Intrusions in Brittany, Cornubia Coastal and estuarine sedimentation
				Valanginian	121	Hastings Beds		
				Ryazanian	128	Durlston Formation } Purbeck Group		London Platform uplifted, North Sea and Wessex-Channel basins separated
	JURASSIC		LATE		131	Lulworth Formation } Purbeck Group		
				Portlandian		Portland Limestone Formation		
					138	Portland Sand Formation		Regression, leading to shallow-water sedimentation
				Kimmeridgian		Kimmeridge Clay		Renewed extension, maximum Jurassic marine transgression
					146	Corallian, including Osmington Oolite		
				Oxfordian				Uplift, carbonate-shelf sedimentation
			MIDDLE	Callovian	152	Oxford Clay		
						Cornbrash		Downwarping, renewed transgression
					157			
				Bathonian		Great Oolite, including Fuller's Earth		Uplift resulting in shallow-water carbonate-shelf sedimentation
				Bajocian	165	Inferior Oolite		
				Aalenian	171			
			EARLY	Toarcian	179	Upper Lias, including Bridport Sands		
				Pliensbachian	186	Middle Lias		Renewed extension, normal faulting coeval with initial opening of central Atlantic; shallow-water sediments of very variable thickness laid down
				Sinemurian	194			
					201	Lower Lias, including Blue Lias		
				Hettangian	205			
	TRIASSIC			Rhaetian		Penarth Group		Marine transgression
				Norian	210	Mercia Mudstone Group		Subsidence, normal faulting along basin margins, deposition of supratidal coastal playa sediments
				Carnian	220			
				Ladinian	230			
				Anisian	235	Sherwood Sandstone Group		Renewed extension, alluvial deposition in south-north axial trough
					242			
				Scythian		Aylesbeare Mudstone Formation		Deposition of floodplain alluvial complex on desert peneplain
PALAEOZOIC	PERMIAN			Tatarian	250	Teignmouth, Exe, Netherton, Oddicombe and Watcombe breccias, & Dawlish Sandstone		Lithospheric extension, block faulting, alluvial and aeolian sedimentation in fault basins
				Kazanian	255			
				Kungurian	260			
					270			
				Artinskian				
					280			
				Sakmarian		Exeter Volcanic Series		Emplacement of Cornubian granites
					290			
	CARBONIFEROUS		SILESIAN	Stephanian				
				Westphalian	300	Bude Formation	Kent: Coal Measures	Major Variscan folding (Variscan orogeny)
					310			
				Namurian		Crackington Formation		
					325			
			DINANTIAN	Viséan		(cherts)		
				Tournaisian		(limestone) (volcanics) (shales)	(limestones) (shales & limestones)	Formation of nappes
	DEVONIAN		LATE	Famennian	360	(slates)	NE France: (schists & limestones)	Emplacement of Lizard Complex
				Frasnian				
			MIDDLE	Givetian	365	Torquay and Chudleigh lsts	(sandstones)	
				Eifelian				
			EARLY	Emsian	380	Staddon Grits and Meadfoot Beds		Formation of Lizard gabbro
				Siegenian				
				Gedinnian	400	Dartmouth Slates	(schists)	
	SILURIAN			Pridoli			Normandy (Cotentin):	
				Ludlow			(siltstones, sandstones)	
				Wenlock		(black shales)	(black shales)	
				Llandovery		(mudstones, sandstones)	(sandstones)	
	ORDOVICIAN			Ashgill	425		(tillites)	
				Caradoc	450	(sandstones)		
				Llandeilo			(sandstones)	
				Llanvirn	460		(mudstones)	
				Arenig		(sandstones)		
				Tremadoc	485	(shales, sandstones)	Grès Armoricain	Emplacement of granites on Jersey
	CAMBRIAN		LATE		495		(sandstone)	
			MIDDLE			Alderney Sandstone & Rozel Conglomerate	(schists, sandstones)	
			EARLY				(grits, sandstones)	
PRECAMBRIAN					570		(volcanics)	
				Brioverian		(Volcanic Group) Jersey Shale Formation	(graywackes, schists) (schists, cherts) (arkoses, volcanics)	Cadomian orogeny
					1000		(gneiss)	
				Pentevrian				
					2000			Lihouan orogeny
					2700	Icart & Perelle gneisses	(gneiss)	Icartien orogeny

Table 3.1: Geological history and lithostratigraphy (from Hamblin *et al.*, 1992).

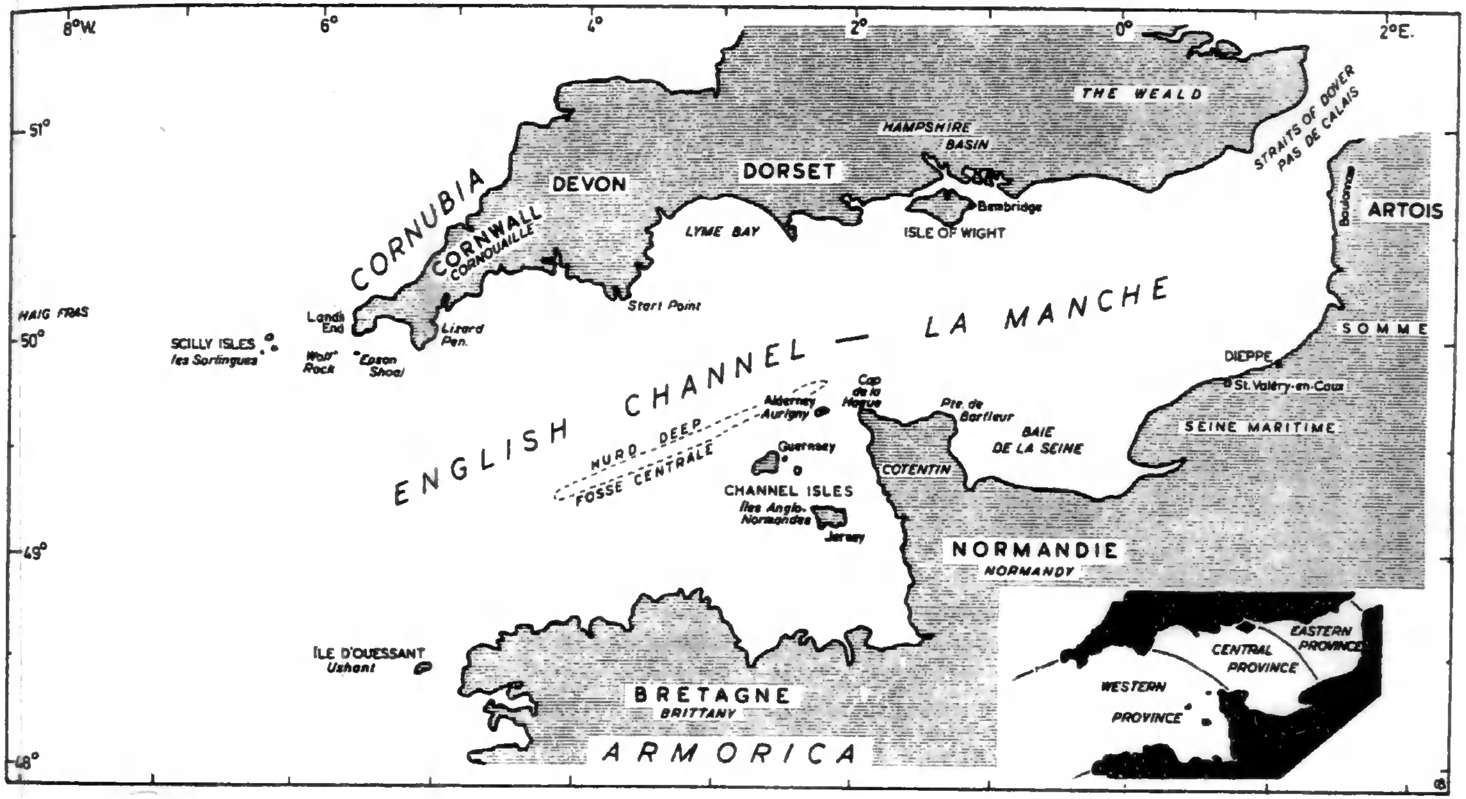


Figure 3-2: The English Channel. Inset are the three provinces from Smith and Curry (1975).

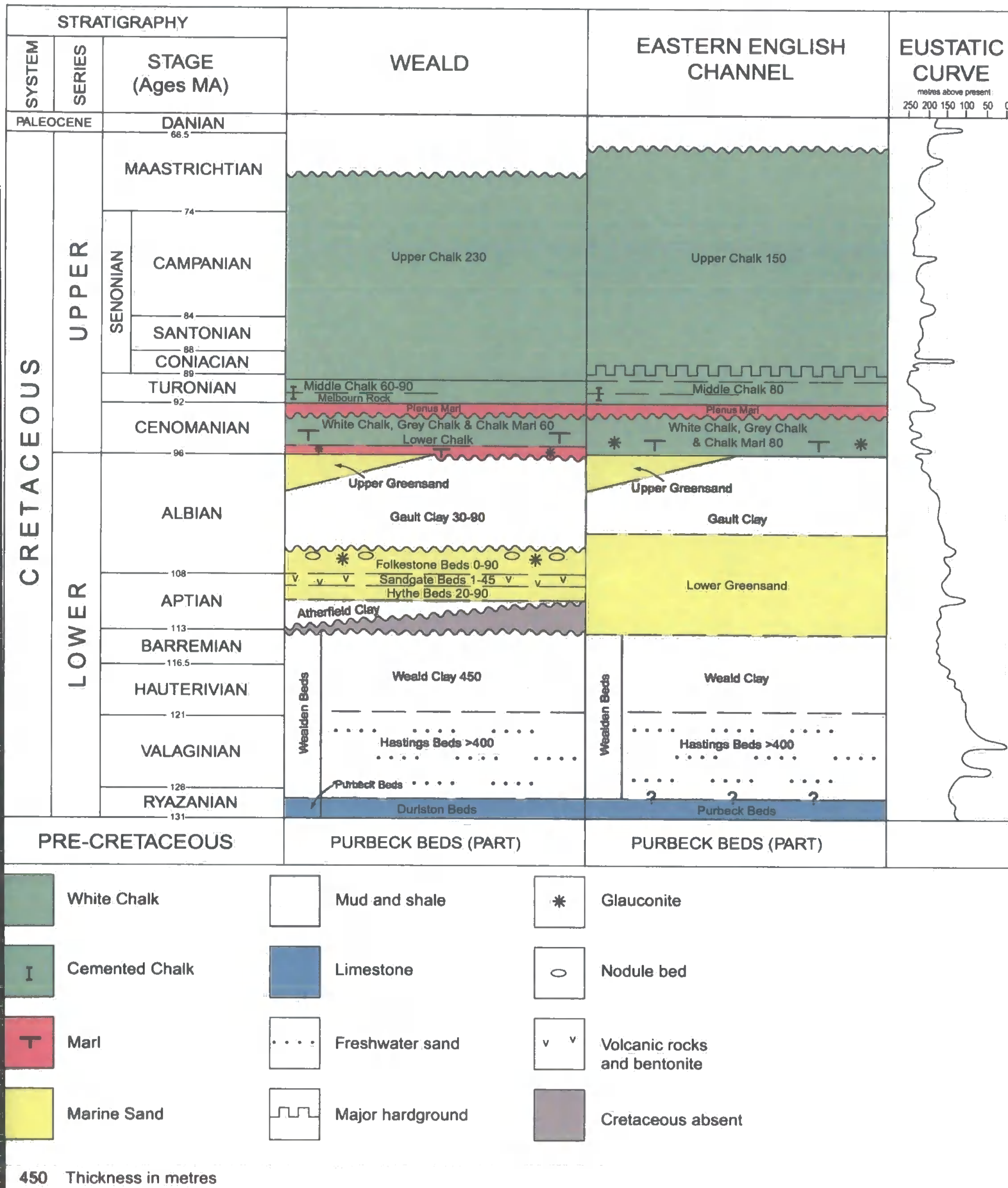


Figure 3-3: Cretaceous strata in the Weald and eastern English Channel (modified from Hamblin *et al.*, 1992).

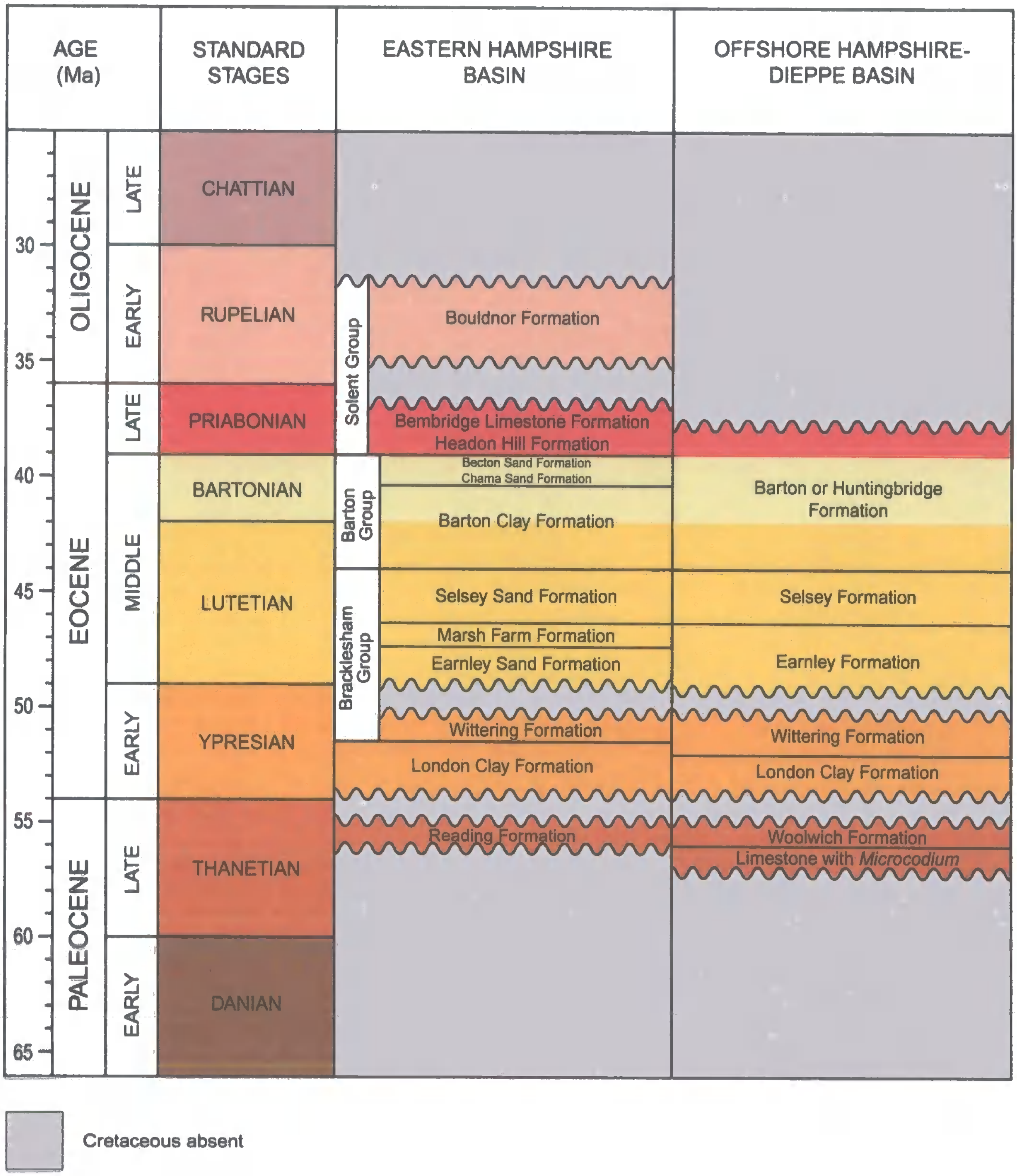


Figure 3-4: Tertiary strata in the eastern Hanpshire Basin and offshore in the Hampshire-Dieppe Basin (modified from Hamblin *et al.*, 1992).

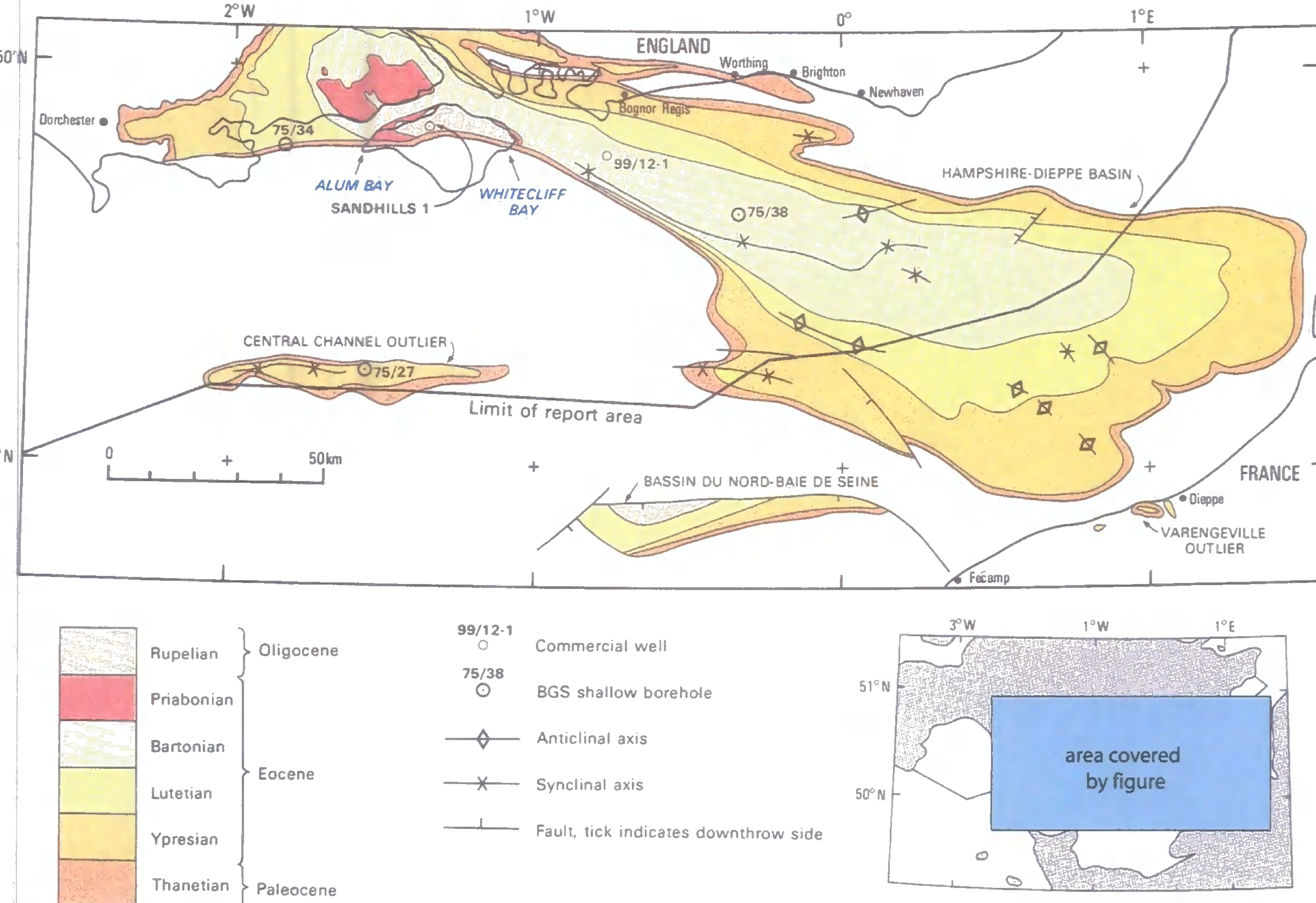
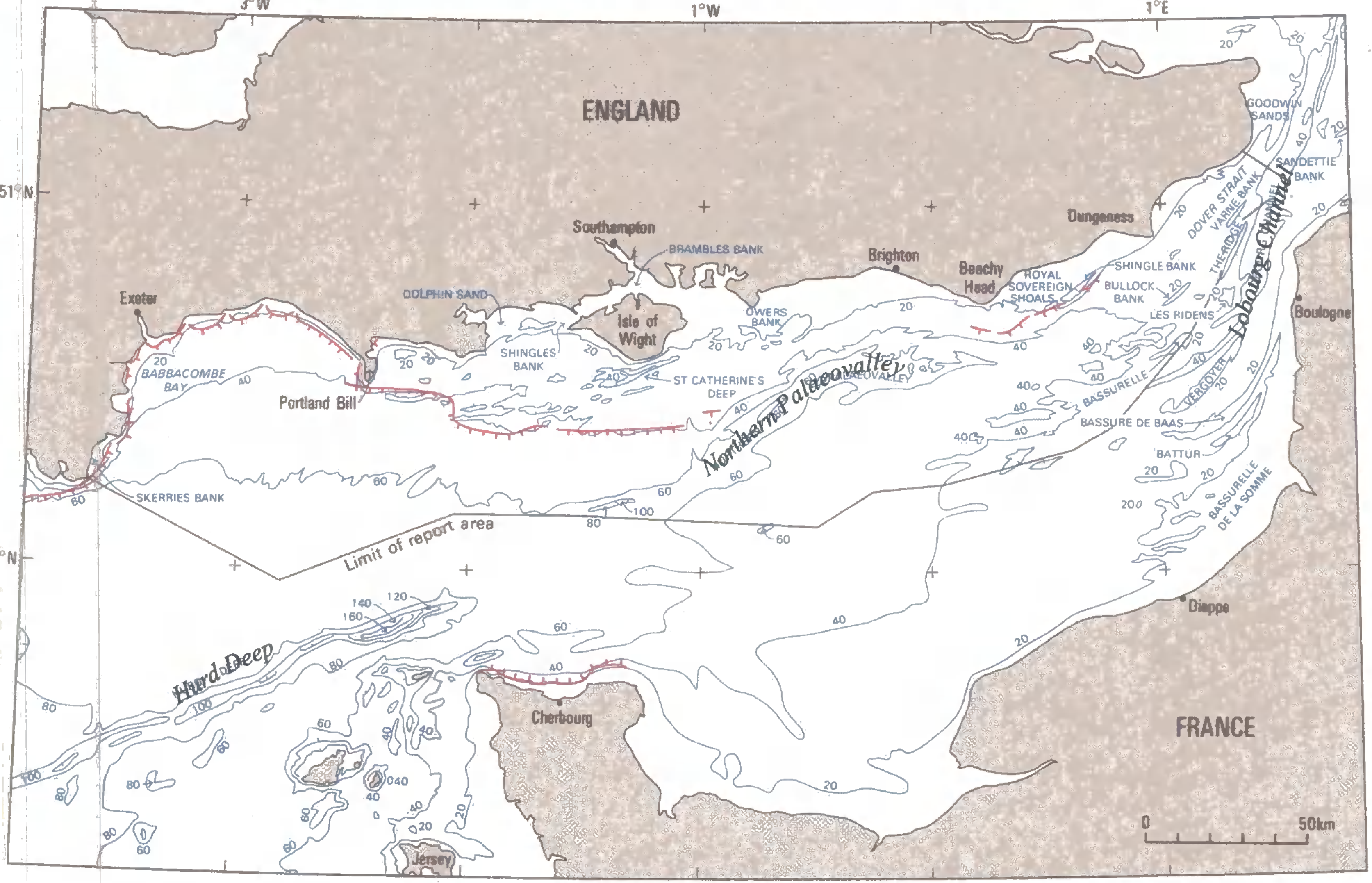


Figure 3-5: Distribution of Tertiary sediments in the English Channel. Includes location of BGS borehole 75/38 (modified from Hamblin *et al.*, 1992).



60 Bathymetric contours at 20 metre intervals below sea level. Depths or ticks on deep side of line.

Submerged cliff line; south of 50°N taken from Larsonneur et al. 1982a, west of 2°W taken from Kellaway et al. (1975), otherwise from BGS surveys

Figure 3-6: Bathymetric map of the English Channel. Shown on it are the three major palaeovalley features cut into the sea floor, the Lobourg Channel, the Northern Palaeovalley and the Hurd Deep. (modified from Hamblin *et al.*, 1992).

MARINE OXYGEN ISOTOPE STAGE	AGE (Ka)	CHRONOSTRATIGRAPHY			CLIMATIC CONDITIONS	ENGLISH CHANNEL
		SERIES		UK STAGE (NW European Stage)		
1		HOLOCENE		FLANDRIAN	TEMPERATE	Mobile sediments (sandbanks) initiated Lag sediments cover seabed as transgression completed
2-4	10 13	PLEISTOCENE	LATE	DEVENSIAN (WEICHSELIAN)	PERIGLACIAL	Palaeovalleys infilled with estuarine then marine sediments as transgression begins Channel dried out, fluvial drainage, major rivers of northern Europe flow through Dover Strait, palaeovalley system developed to present form. **Fluvial**
5	120	PLEISTOCENE	LATE	IPSWICHIAN (EEMIAN)	TEMPERATE	English Channel and North Sea joined by Dover Strait. **Marine conditions**
7	130	PLEISTOCENE	MIDDLE		PERIGLACIAL	Channel dried out, fluvial drainage, major rivers of northern Europe flow through Loburg Channel. **Fluvial**
9-11	300	PLEISTOCENE	MIDDLE	HOXNIAN (HOLSTEINIAN)	TEMPERATE	English Channel and North Sea joined by Dover Strait. **Marine conditions**
12-14	440	PLEISTOCENE	MIDDLE	ANGLIAN (ELSTERIAN)	PERIGLACIAL	Dover Strait formed as a glacial lake overflow channel in MOIS12 English Channel dried out, fluvial drainage. **Fluvial**
15	560	PLEISTOCENE	MIDDLE	'CROMERIAN COMPLEX' (CROMERIAN)	COLD/ TEMPERATE	**Marine conditions** Channel floor dried out during glacial periods ? River system developed on floor of English Channel ? **Fluvial**
16-40	770	PLEISTOCENE	EARLY	BEESTONIAN TO PRE-LUDHAMIAN (MENAPIAN TO PRAETIGLIAN)	BECOMING COLD	Shallowing marine conditions
	2550	PLIOCENE			TEMPERATE	Marine conditions over most of English Channel possible connection to southern North Sea. Dover Strait not in existence

Table3-2: Quaternary history of English Channel modified from Hamblin *et al.* (1992)

Deposit	Geographical Zone	Altitudinal Range	Inferred Depositional Environment
Pagham Raised Beach	LOWER COASTAL PLAIN	-2.0 - 3.0 m OD	inter-tidal marine marginal
Brighton-Norton Raised Beach		5.0 – 12.0 m OD	estuarine inter/sub-tidal
Cams Down Raised Beach	UPPER COASTAL PLAIN	15.0 – 18.0 m OD	inter-tidal marine marginal
Aldingbourne Raised Beach		17.5 – 27.5 m OD	inter-tidal marine marginal inter/sub-tidal
Goodwood-Slindon Raised Beach		32.0 – 43.0 m OD	estuarine inter/sub-tidal inter-tidal marine marginal

Table 3-3: Altitudinal distribution of raised marine terrace deposits of the West Sussex Coastal Plain with inferred environment of deposition (modified from Bates *et al.*, 1998).

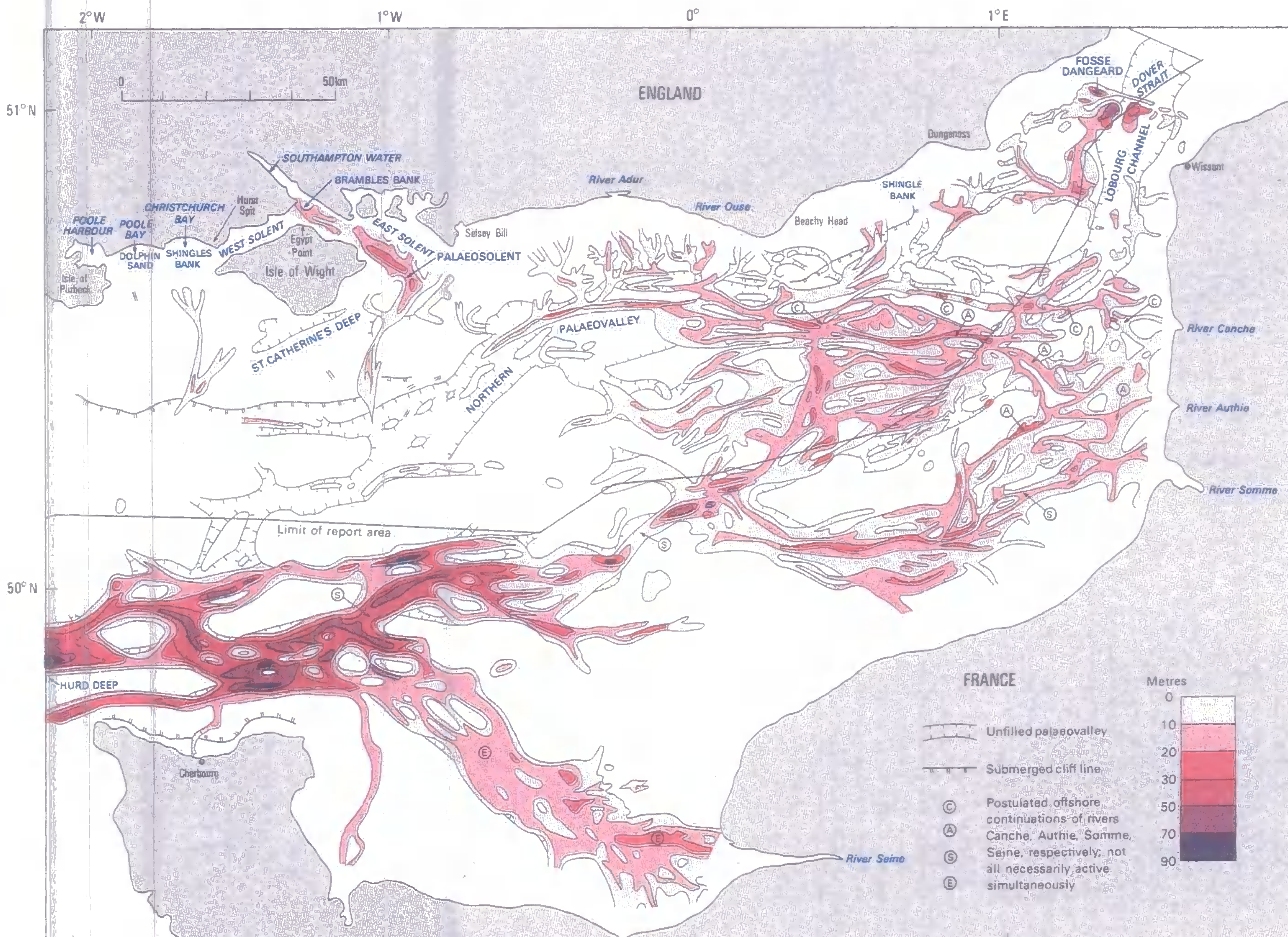


Figure 3-7: The palaoevalleys and infilling sediments of the English Channel (from Hamblin *et al.*, 1992).

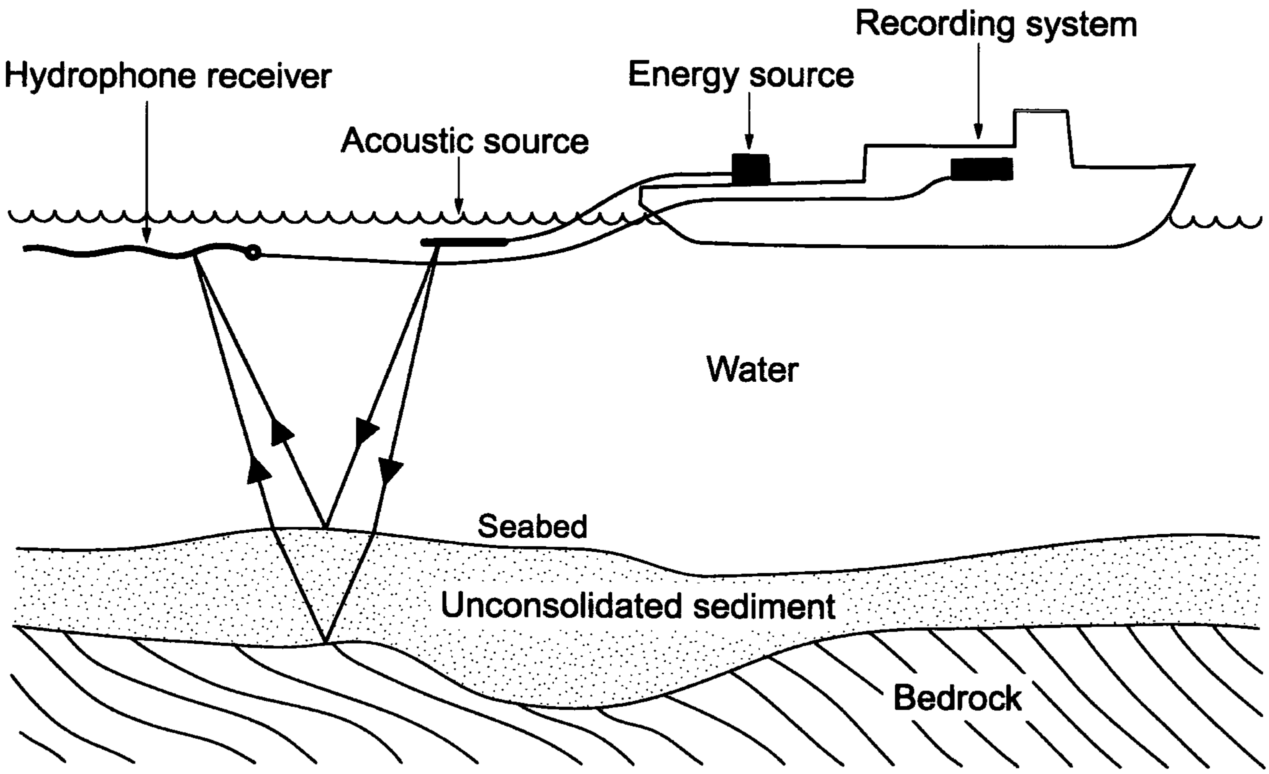


Figure 4-1: Schematic of continuous shallow seismic reflection profile data acquisition (modified from Evans *et al.*, 1995).

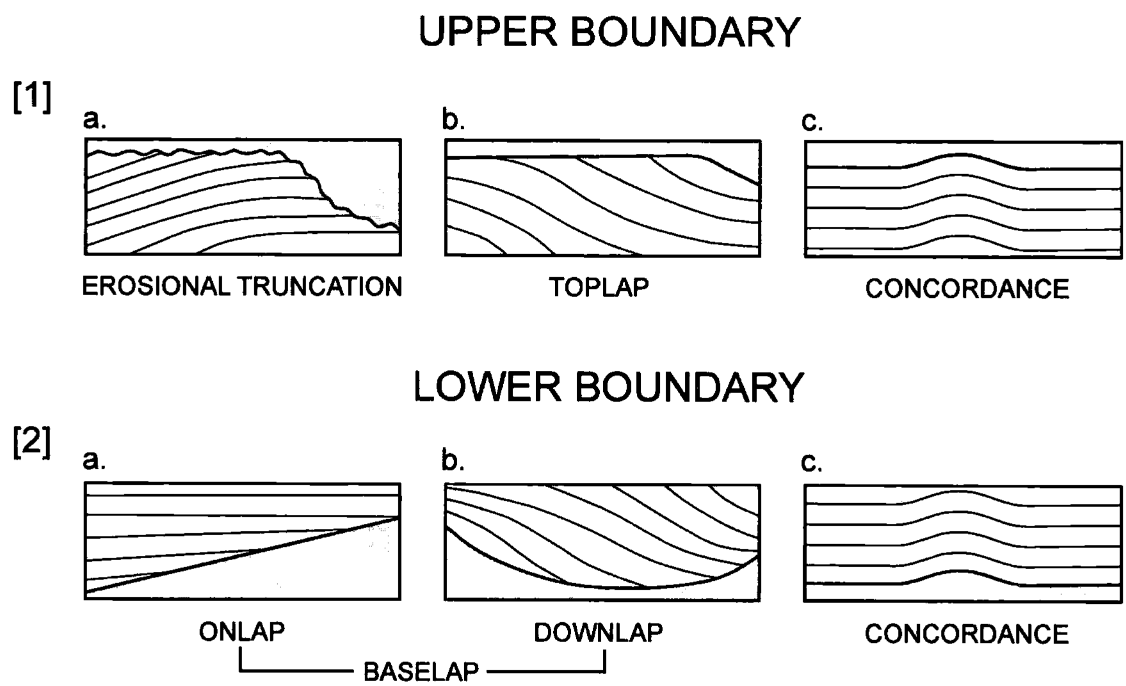


Figure 4-2: Relations of strata to boundaries of depositional sequences.

[1] Upper Boundary. **a.** *Erosional truncation*: strata at the top of a given sequence termiate against upper boundary mainly as a result of erosion (e.g. tilted strata terminating against an overlying horizontal erosion surface, or horizontal strata terminating against later channel incision surface). **b.** *Toplap*: initially inclined strata at the top of a given sequence terminate agains the upper boundary mainly as a result of nondeposition (e.g. foreset strata terminating against an overlying horizontal surface where no erosion of deposition took place). **c.** *Upper concordance*: the relation in which strata at the top of a given sequence do not teminate against the upper boundary.

[2] Lower Boundary. **a.** *Onlap:* at the base of the sequence initially horizontal strata terminate progressively against an initially inclined surface or initially inclined strata terminate in an updip manner progressively against a surface of greater inclination. **b.** *Downlap*: at the base of the sequence initially inclined strata terminate in a downdip manner progressively against an initially horizontal or inclined surface. **c.** *Basal concordance*: strata at the base of the sequence do not terminate against the lower sequence boundary.

Modified from Mitchum et al. (1977a).

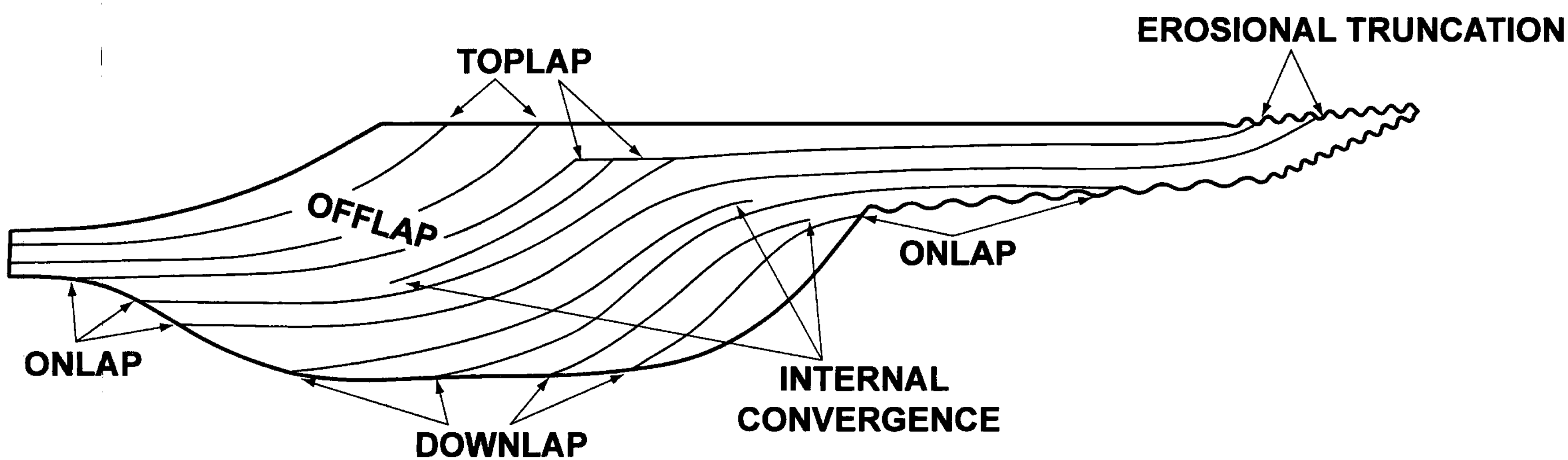


Figure 4-3: Examples of discordant seismic stratigraphic sequence boundaries within an idealised sequence (modified from Mitchum *et al.*, 1977b)

Seismic Facies Parameters	Interpretation
Reflection Configuration	- Bedding Patterns - Depositional Processes - Erosion and Palaeotopography - Fluid Contacts
Reflection Continuity	- Bedding Continuity - Depositional Processes
Reflection Amplitude	- Velocity – Density Contrast - Bed Spacing - Fluid Content
Reflection Frequency	- Bed Thickness - Fluid Content
Interval Velocity	- Estimation of Lithology - Estimation of Porosity - Fluid Content
External Form and Areal Association of Seismic Facies Units	- Gross Depositional Environment - Sediment Source - Geological setting

Table 4.1: Seismic reflection parameters used in seismic stratigraphy and their geological interpretation (taken from Mitchum et al., 1977b).

External Forms
Sheet
Sheet drape
Wedge
Bank
Lens
Mound (subdivided into generalised and fan types)
Fill (subdivided into channel, trough, basin or slope front types)

Table 4.2: External forms of seismic facies units

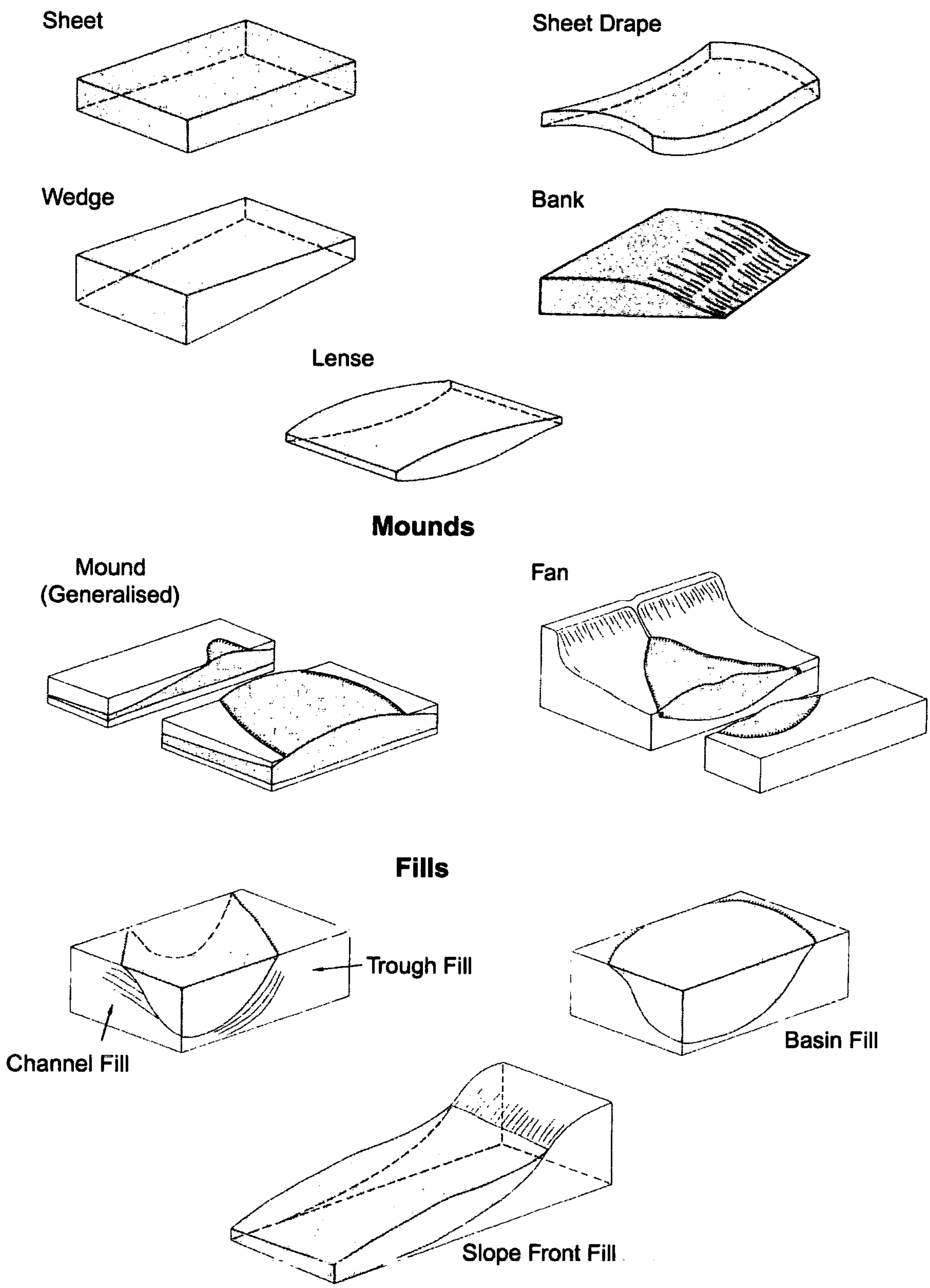


Figure 4-4: Schematic representation of some external forms of seismic facies units (modified from Mitchum et al., 1977b)

Internal Reflection Configurations	
Principal Stratal Configuration	**Modifying Terms**
Parallel	Even
Subparallel	Wavy
Divergent/Convergent	Regular
Prograding Clinoforms	Irregular
-Sigmoid	Uniform
-Oblique (tangential or parallel)	Variable
-Complex Sigmoid-Oblique	Hummocky
-Shingled	Lenticular
-Hummocky Clinoforms	Disrupted
Chaotic	Contorted
Reflection-Free	

Table 4.3: Internal reflection configurations of seismic facies units (modified from Mitchum et al., 1977b).

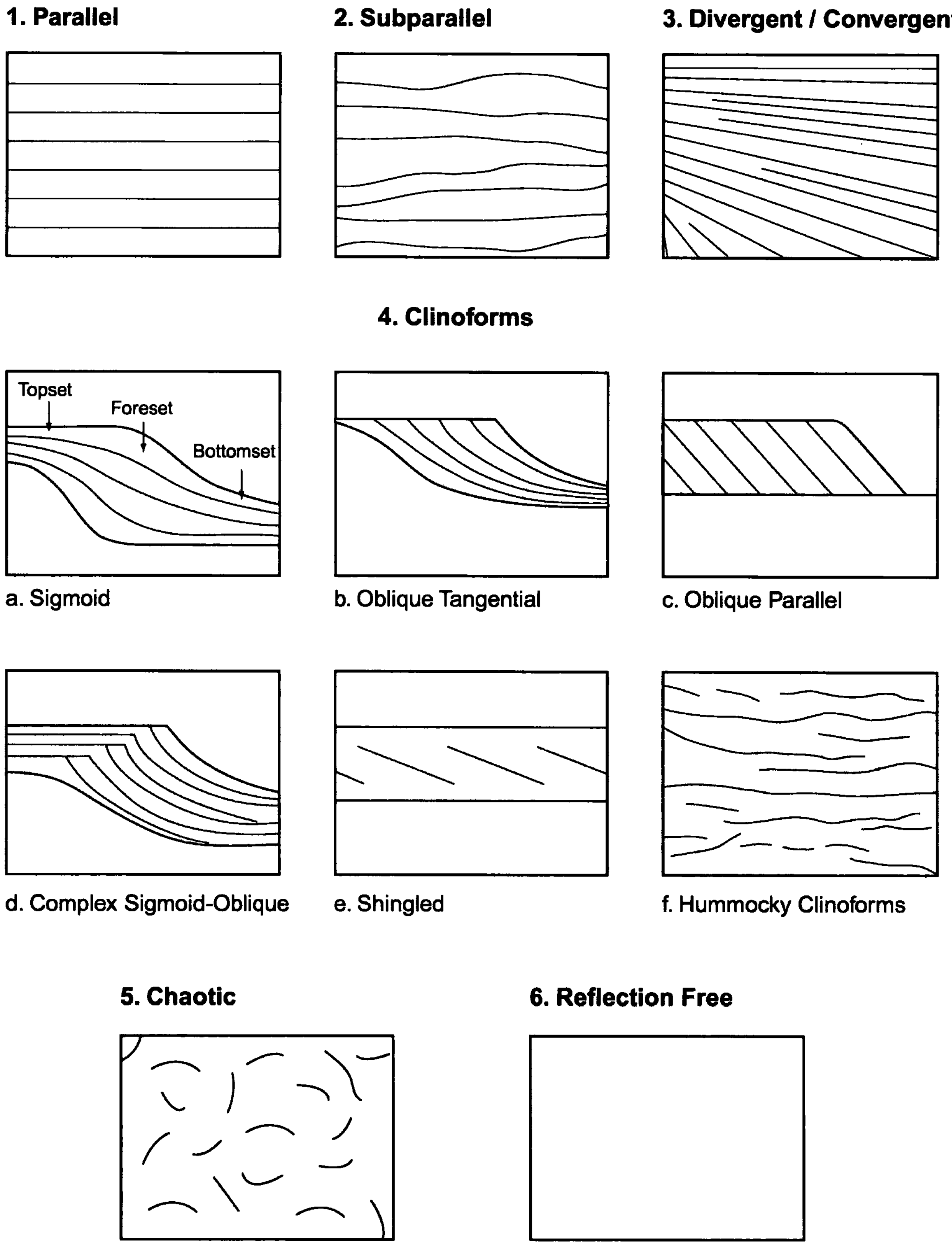


Figure 4-5: Schematic representation of commonly used seismic internal reflection configurations (after Mitchum *et al.*, 1977b)

Figure 4-6: Deployment of a vibrocorer from the rear deck of the ship (photograph, the author).

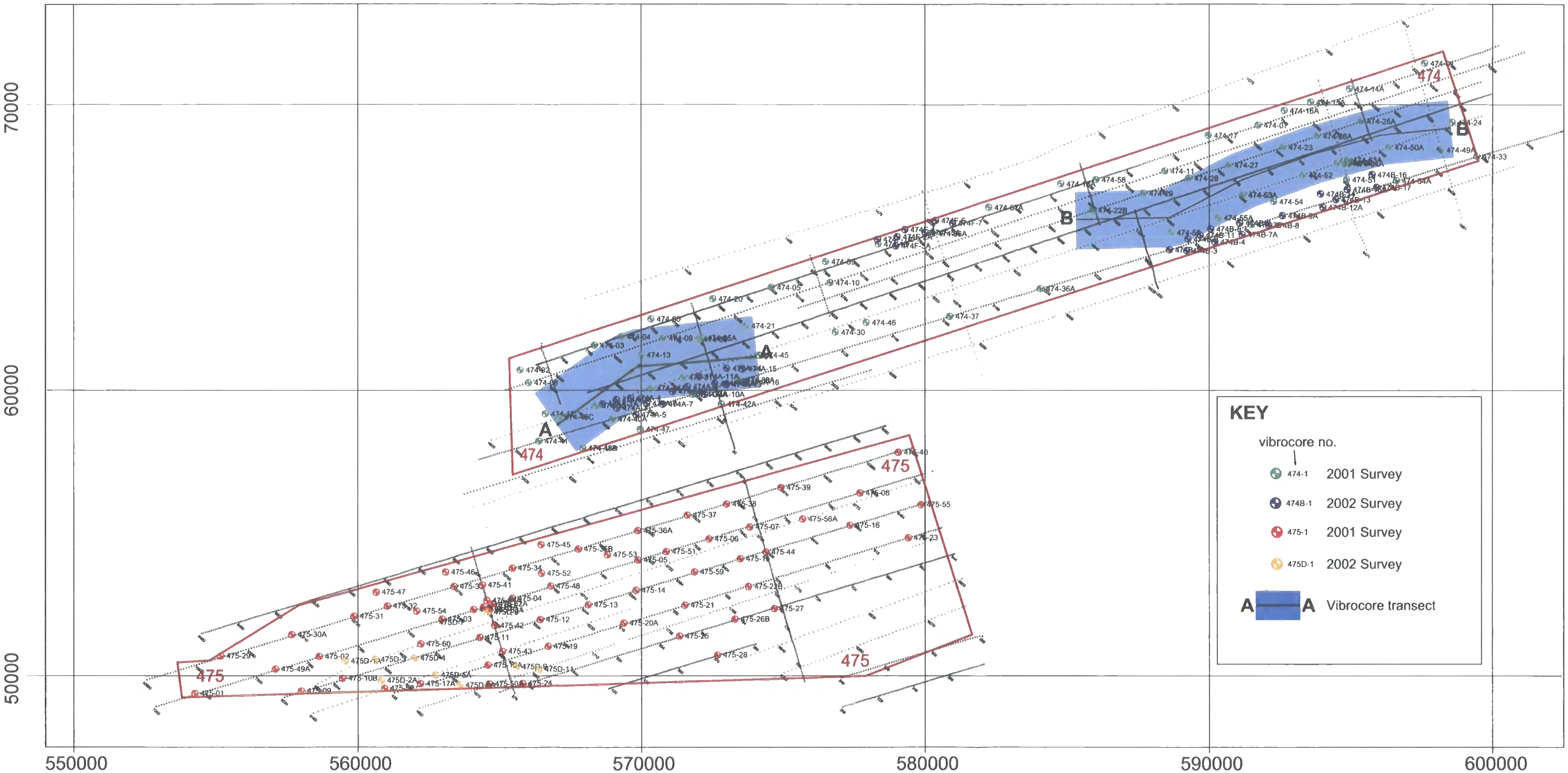


Figure 4-7: Locations of vibrocore samples from the 2001 and 2002 vibrocoring surveys in Licence Application Areas 474 and 475. Also shown are the locations of the transect lines A - A and B - B. Coordinates are OSGB.

$$\text{Mean} \quad \bar{x}_\phi = \frac{\Sigma fm}{n} \tag{1}$$

$$\text{Standard deviation} \quad \sigma_\phi = \sqrt{\frac{\Sigma f(m - \bar{x}_\phi)^2}{100}} \tag{2}$$

$$\text{Skewness} \quad Sk_\phi = \frac{\Sigma f(m - \bar{x}_\phi)^3}{100\,\sigma_\phi^3} \tag{3}$$

$$\text{Kurtosis} \quad K_\phi = \frac{\Sigma f(m - \bar{x}_\phi)^4}{100\,\sigma_\phi^4} \tag{4}$$

where:

f = weight percent (frequency) in each grain-size grade present,
m = midpoint of each grain-size grade in phi (□) values
n = total number in sample which is 100 when f is in percent.

Table 4.4: Statistical method of moments formulae used in calculating grain-size parameters. The table shows the way calculations are made for the four moment statistics. The technique of computing the parameters is commonly called the method of moments, because, computations involve multiplying a weight (frequency, in percent, by a distance; this is analogous to computing moments of inertia for the grains in each size grade in the distribution (McBride, 1971).

Sorting (σ_ϕ)	**Verbal Description of Sorting**
Under 0.35	Very well sorted
0.35 - 0.50	Well sorted
0.50 - 0.71	Moderately well sorted
0.71 - 1.0	Moderately sorted
1.0 - 2.0	Poorly sorted
2.0 - 4.0	Very poorly sorted
Over 4.0	Extremely poorly sorted

Table 4.5: Descriptive terms applied to sorting values.

Skewness (Sk_{ϕ})	**Verbal Description of Skewness**
>1.30	Very fine skewed
0.43 to 1.30	Fine skewed
-0.43 to 0.43	Symmetrical
-0.43 to -1.30	Coarse skewed
< -1.30	Very coarse skewed

Table 4.6: Descriptive terms applied to skewness values

Kurtosis (K_ϕ) Value	**Verbal Description of Kurtosis**
Under 1.70	Very platykurtic
1.70 – 2.55	Platykurtic
2.55 – 3.70	Mesokurtic
3.70 – 7.40	Leptokurtic
Over 7.40	Very leptokurtic

Table 4.7: Descriptive terms applied to kurtosis values

Element	UCC	Element	UCC
Ce	65.7	**Er**	2.1
La	32.3	**Yb**	1.5
Nd	25.9	**Eu**	0.95
Pr	6.3	**Ho**	0.62
Sm	4.7	**Tb**	0.50
Gd	2.8	**Lu**	0.27
Dy	2.9	**Tm**	0.30

Table 4.8: Element concentrations (in parts per million) in the Upper Continental Crust (UCC) taken from Wedepohl (1995).

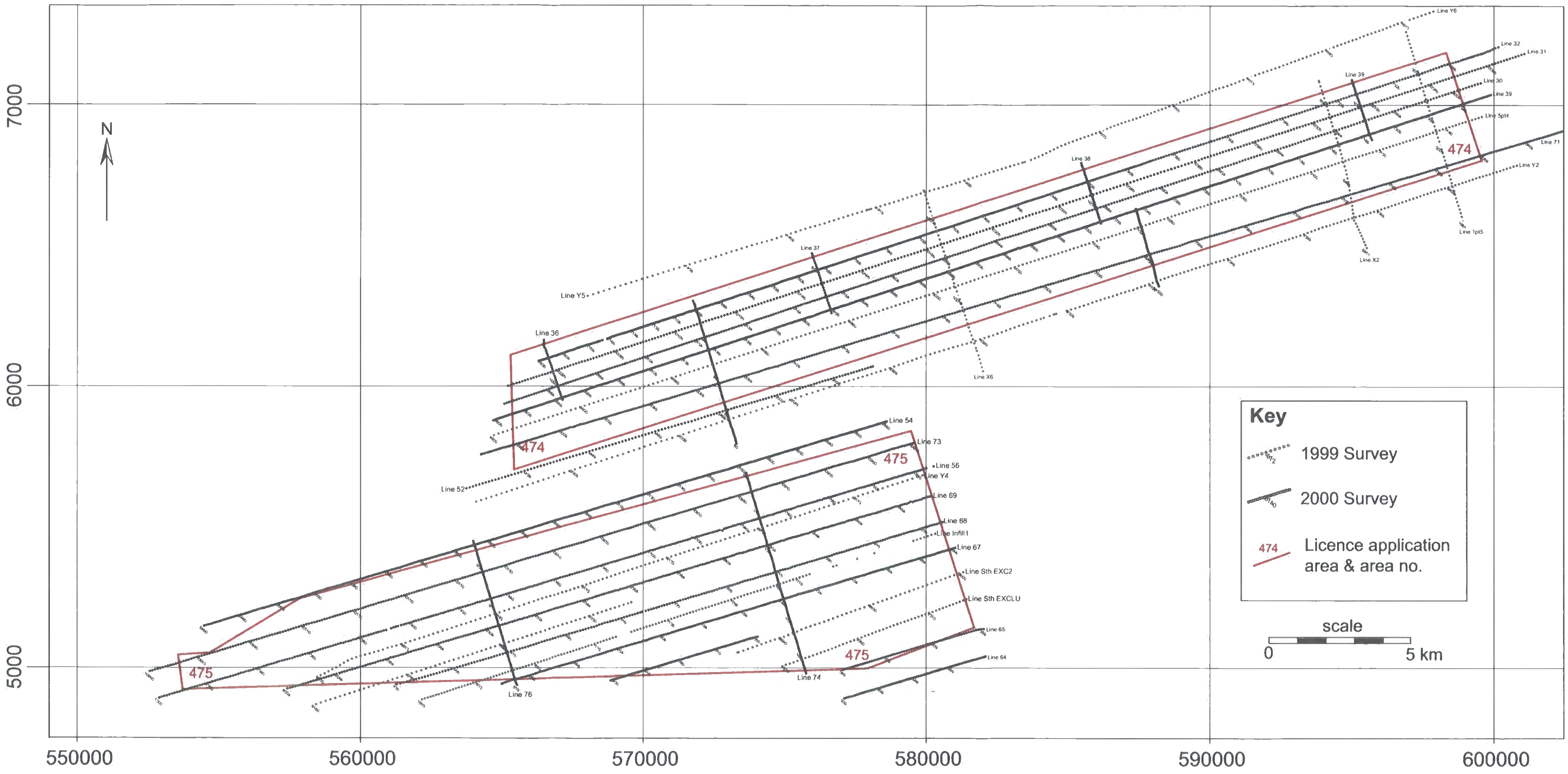


Figure 5-1: Trackplot of 1999 and 2000 seismic reflection profiling surveys in the study area. Coordinates are OSGB.

Seismic Facies Unit	Thickness	Lower Bounding Surface Depth range (CD)	Lower Bounding Surface Reflector Characteristics	Upper Bounding Surface Reflector Characteristics	Principal Internal Reflector Characteristics	Areal Extent	Appearance on Seismic Lines	Type Sections (Figure No.)
A	Unknown	Unknown	Unknown	Low amplitude undulating – even horizontal	Moderate-high amplitude; dipping – horizontal; continuous, parallel	Occurs throughout study area	All	On all sections
B	0 to 24 m	-36 to –76 m	Low-amplitude, uneven, discontinuous, horizontal-dipping	Moderate-high amplitude, undulating to flat	Moderate-low amplitude chaotic, occasional subparallel prograding.	N-S: beyond limits of study area E-W: 585000E – 595800E	Y6, 32, 31, 30, 29, 5pt4, 71, Y2, 1pt5, 39, X2, 38.	5-10, 5-13
C	0 to 25 m	-40 to –72 m	Low-amplitude, discontinuous, horizontal to dipping	Variable amplitude, undulating - even, discontinuous	Moderate amplitude chaotic / mounded, occasional subparallel	N-S: beyond limits of study area E-W: 566000E – 575000E	Y5, 32, 31, 30, 29, 5pt4, 71, 52, Y2, 54, 73, 56, Y4, 69, 68, Infill1, 67, 66, Sth EXC2, Sth EXCLU,	5-6, 5-7, 5-14
D	0 to 14 m	-45 to –65 m	moderate-amplitude, Continuous and undulating	Mildy undulating – even	Moderate amplitude chaotic and occasional dispersed hummocky type reflectors reflection free /transparent towards base	Forms a depression around 65000N, 580000E	Y5, X6	5-15
E	0 to 20 m	-46 to –63 m	Moderate- to high-amplitude	High amplitude, continuous	Chaotic occasionally mounded	Forms a depression around 55000N, 565000E	52, Y2	5-16
F	0 to 13 m	-46 to –56 m	Discordant high-amplitude, continuous, dipping from both east and west	Arbitrary	Chaotic	N-S: 50000N E-W5: 55500E-560000E	69, 56, Y4	5-9
G	0 to 20 m	-33 to –62 m	High-amplitude, continuous with SFU-A, arbitrary with other units	High amplitude, continuous, horizontal at sea bed, even to undulating when below other units	Varied: Dominated by moderate - high amplitude irregular, discontinuous steeply inclined and chaotic, occasional onlapping parallel	Present over most of the study area	All	5-6, 5-7, 5-9, 5-10, 5-11, 5-12, 5-13, 5-14, 5-15
H	1 to15 m	-38 to –53 m	High-amplitude, dipping E and W	Single high amplitude continuous	Parallel, horizontal to dipping onlap and downlap	Forms a depression around 61000N, 568000E	29, 30, 31, 31, 5pt4	5-6, 5-7
I	10 cm to10 m	Just below sea bed	High-amplitude, continuous	Seabed	Even to wavy parallel reflectors, horizontal or shallow dipping	Present over most of the study area	All	On all sections

Table 5-1: Main characteristics of seismic facies units identified in the PhD study area.

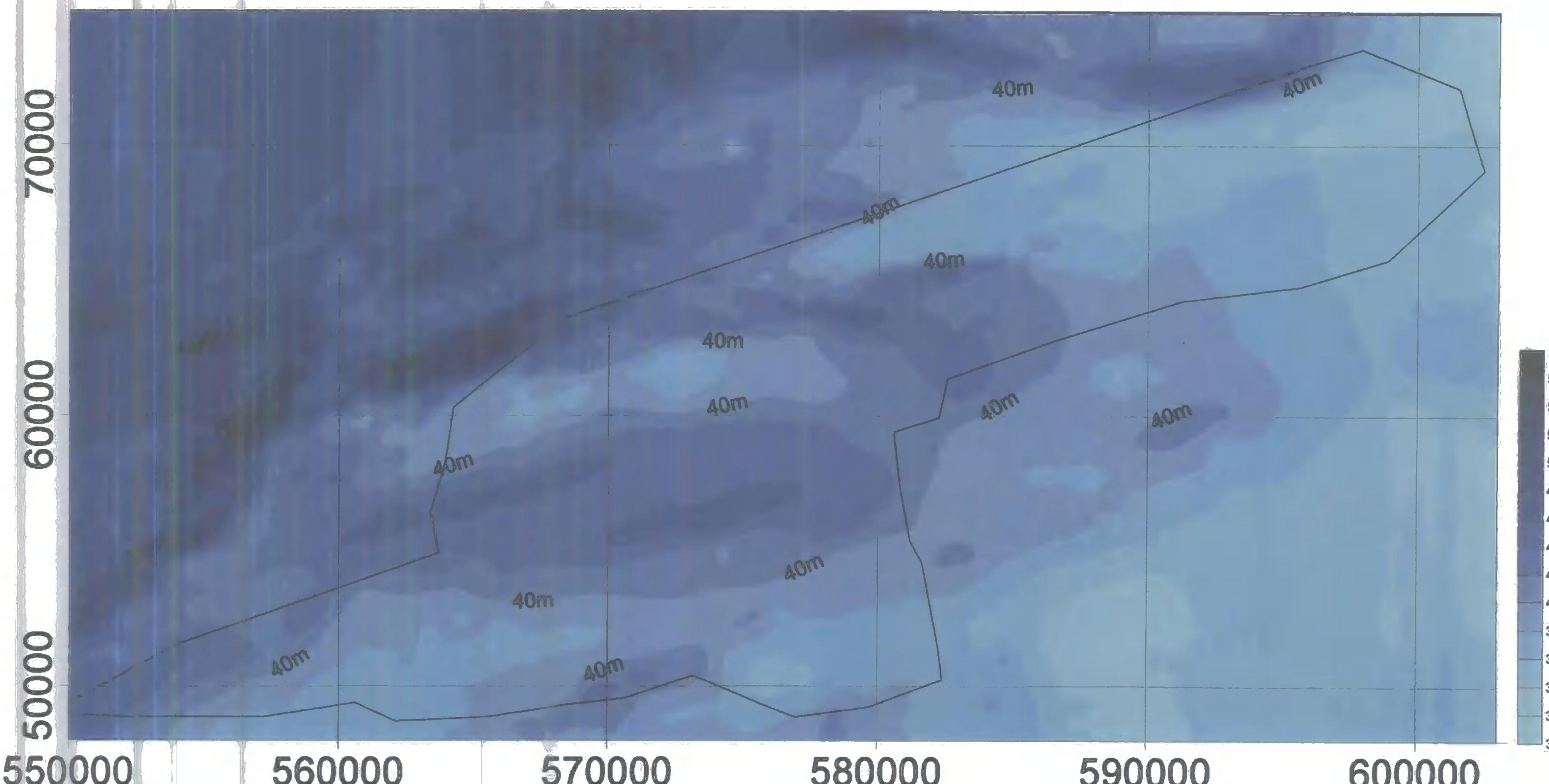


Figure 5-2: Contour map of bathymetry in metres below CD in the study area

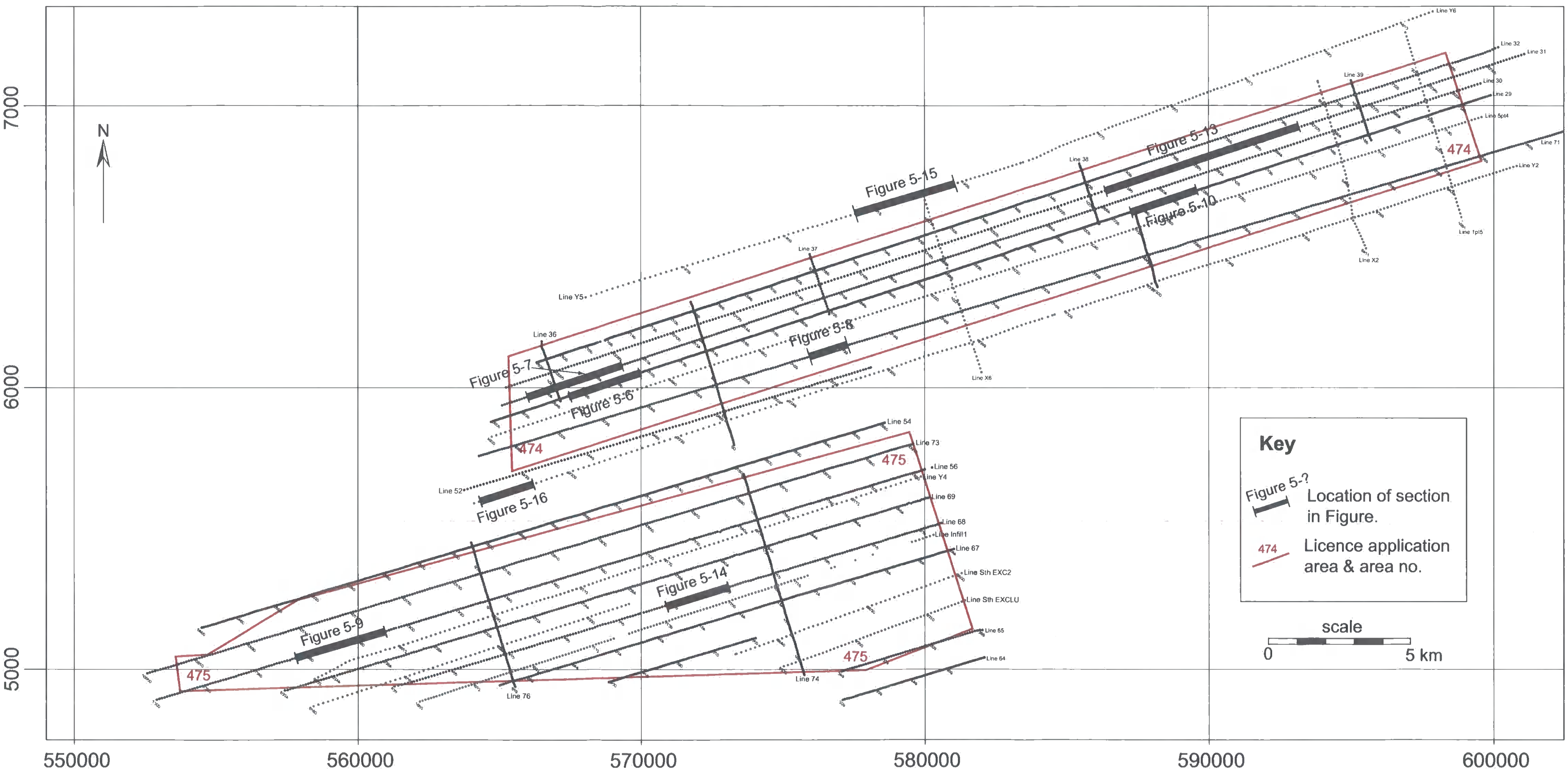


Figure 5-3: Trackplot showing the locations of reflection profile sections referred to in the text. Coordinates are OSGB.

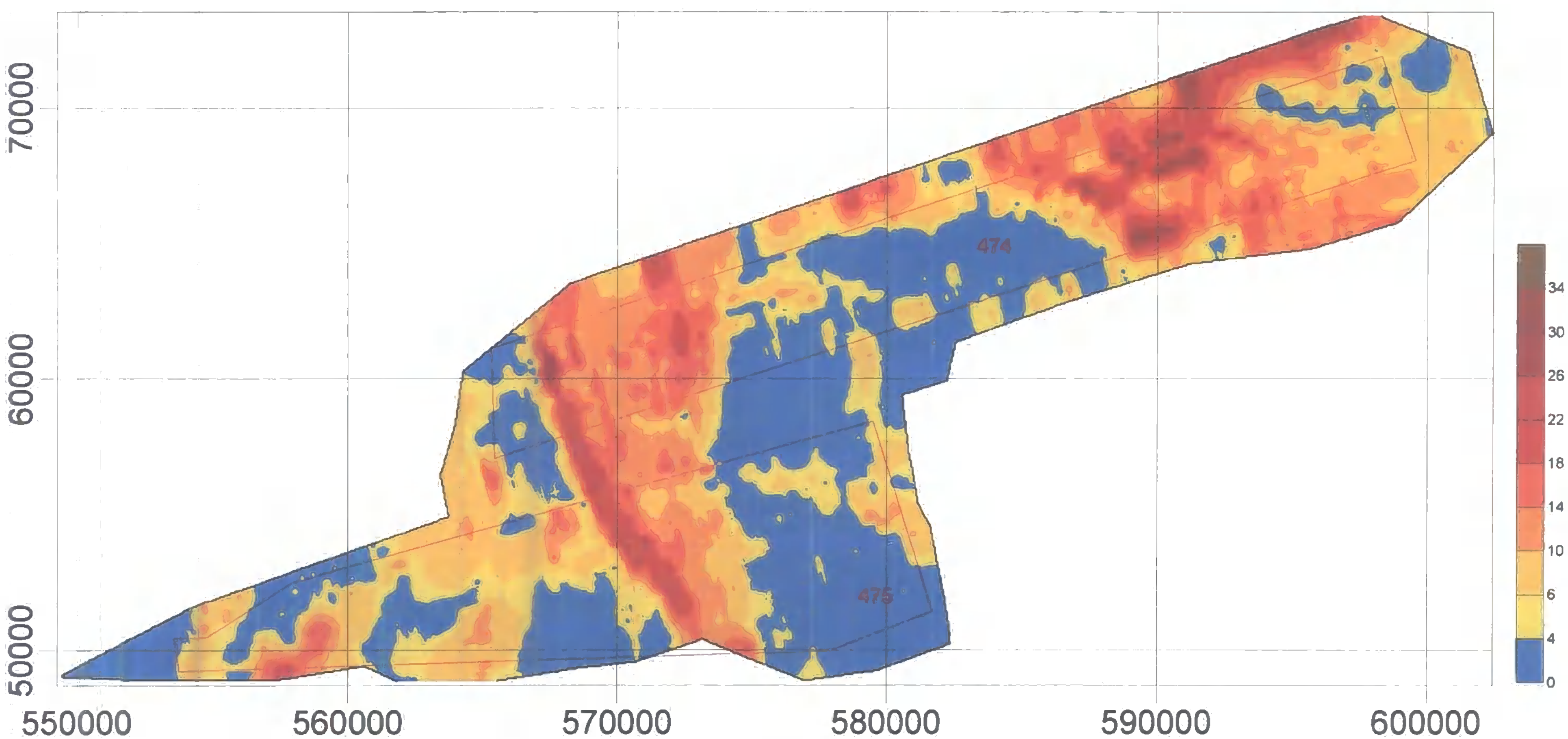


Figure 5-4: Isopachyte map of infilling sediment thicknesses in the study area

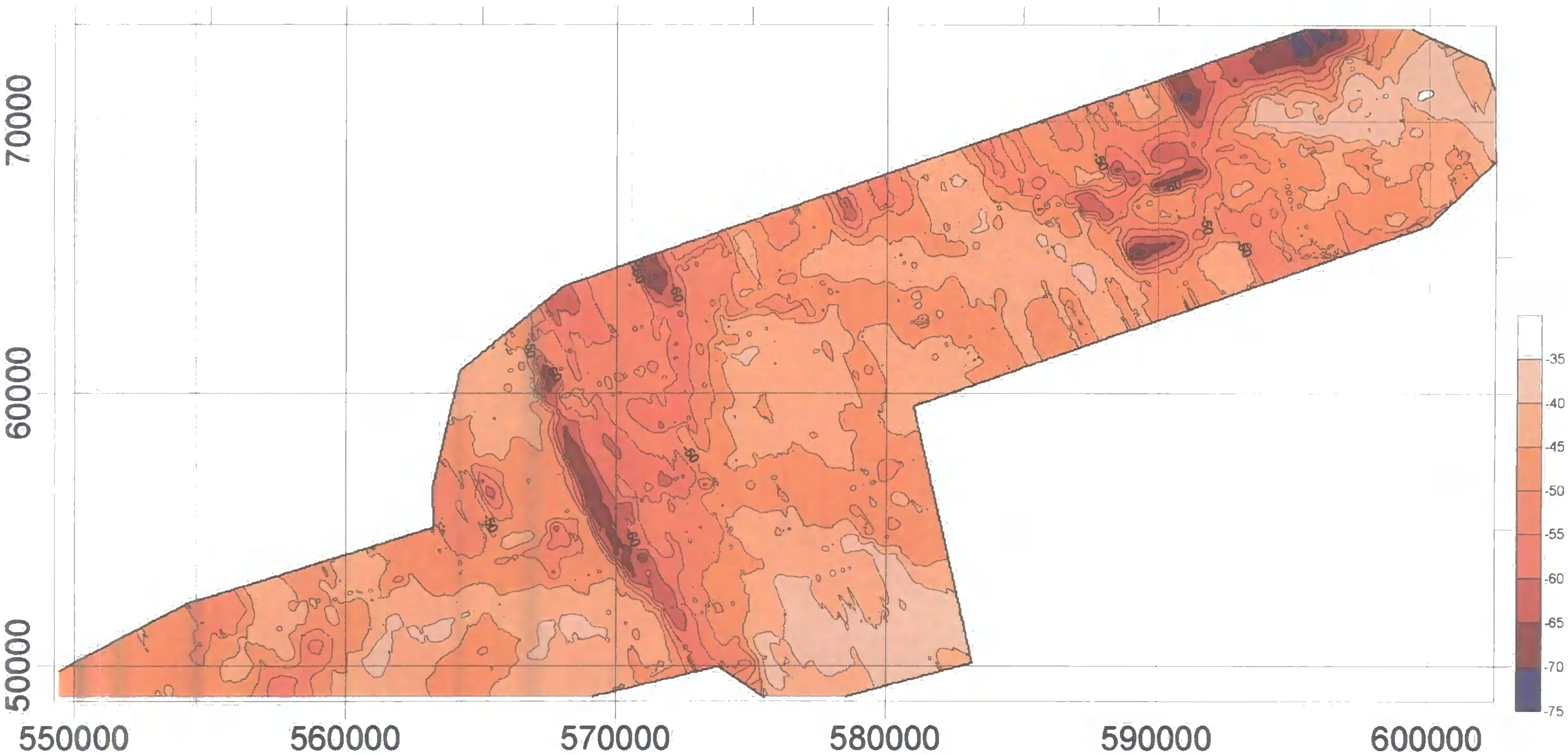


Figure 5-5: Contour map of depth in metres below CD to sequence boundary surface LQ_b (rockhead) in the study area

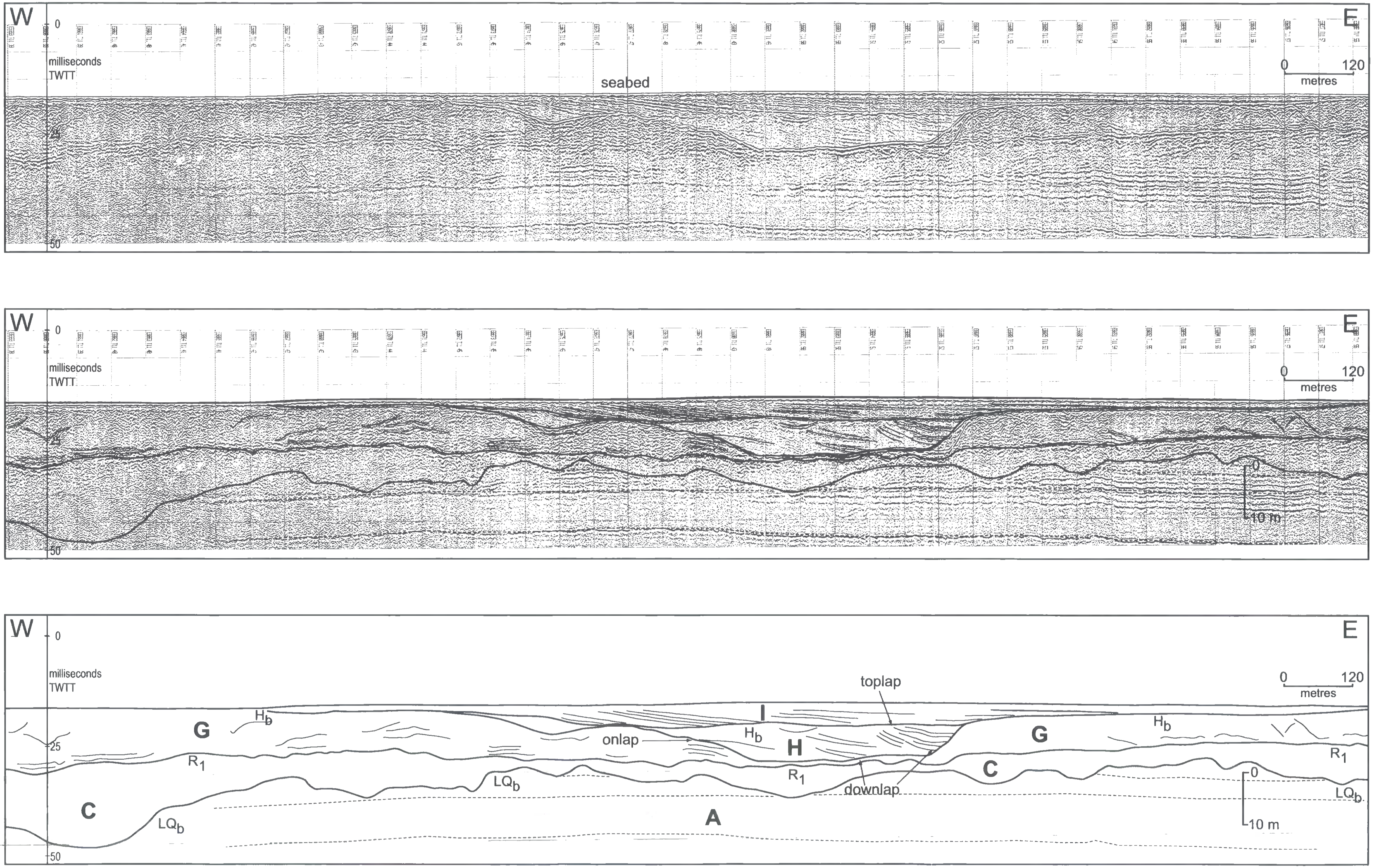


Figure 5-6: Boomer seismic profile section and Interpreted line drawing of Line 29 from Fixes 6659 - 6698, showing the sequence boundary surfaces LQ_b and H_b and Ravinement surface R_1. Also shown are seismic facies units A, C, G, H and I. Location of section shown on Figure 5-3.

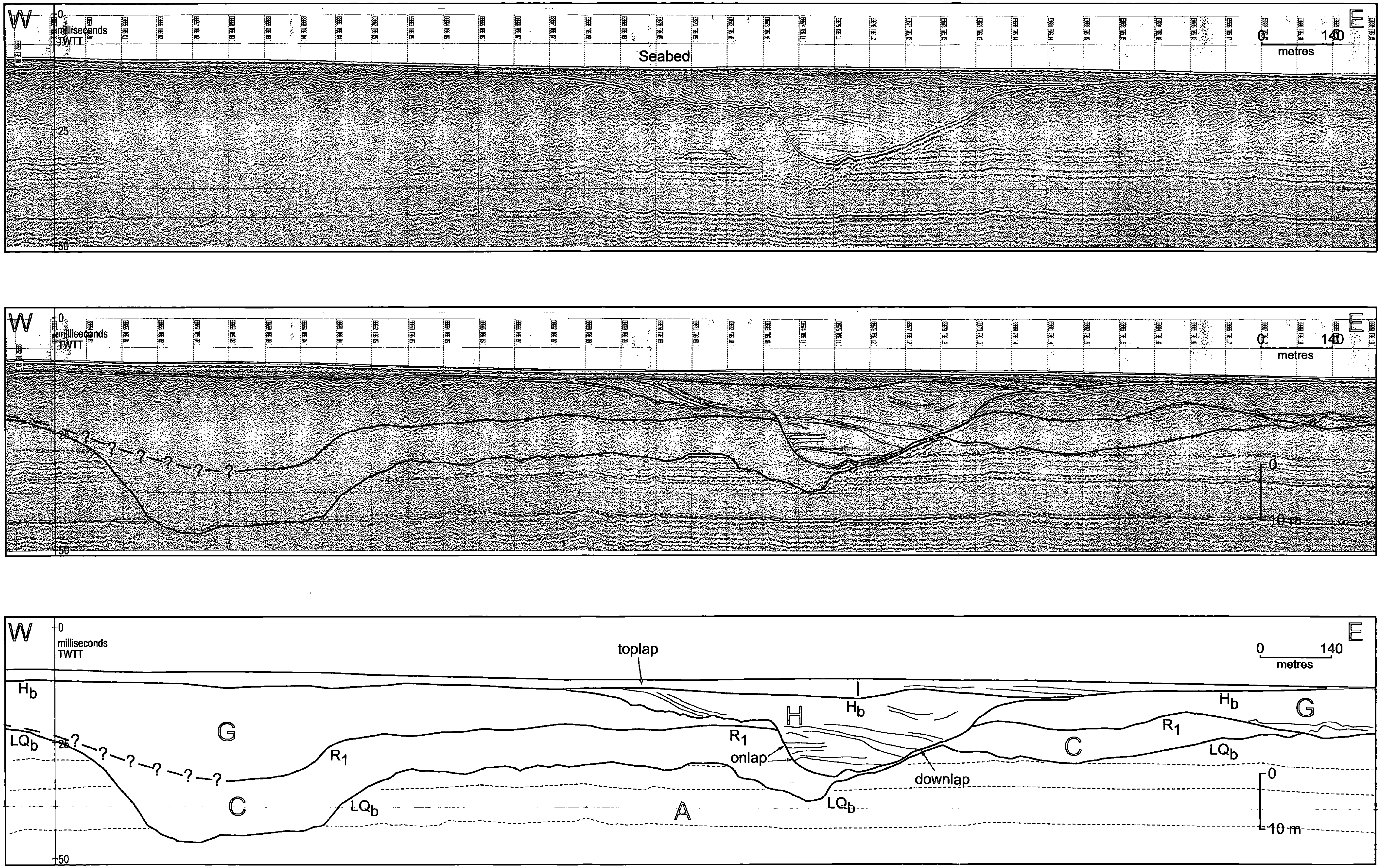


Figure 5-7: Boomer seismic profile section and interpreted line drawing of Line 30 from fixes 8653 - 8690, showing sequence boundary surfaces LQ_b and H_b and and ravinement surface R_1. Also shown are seismic facies units A, C, G, H and I. Location of section shown on Figure 5-3.

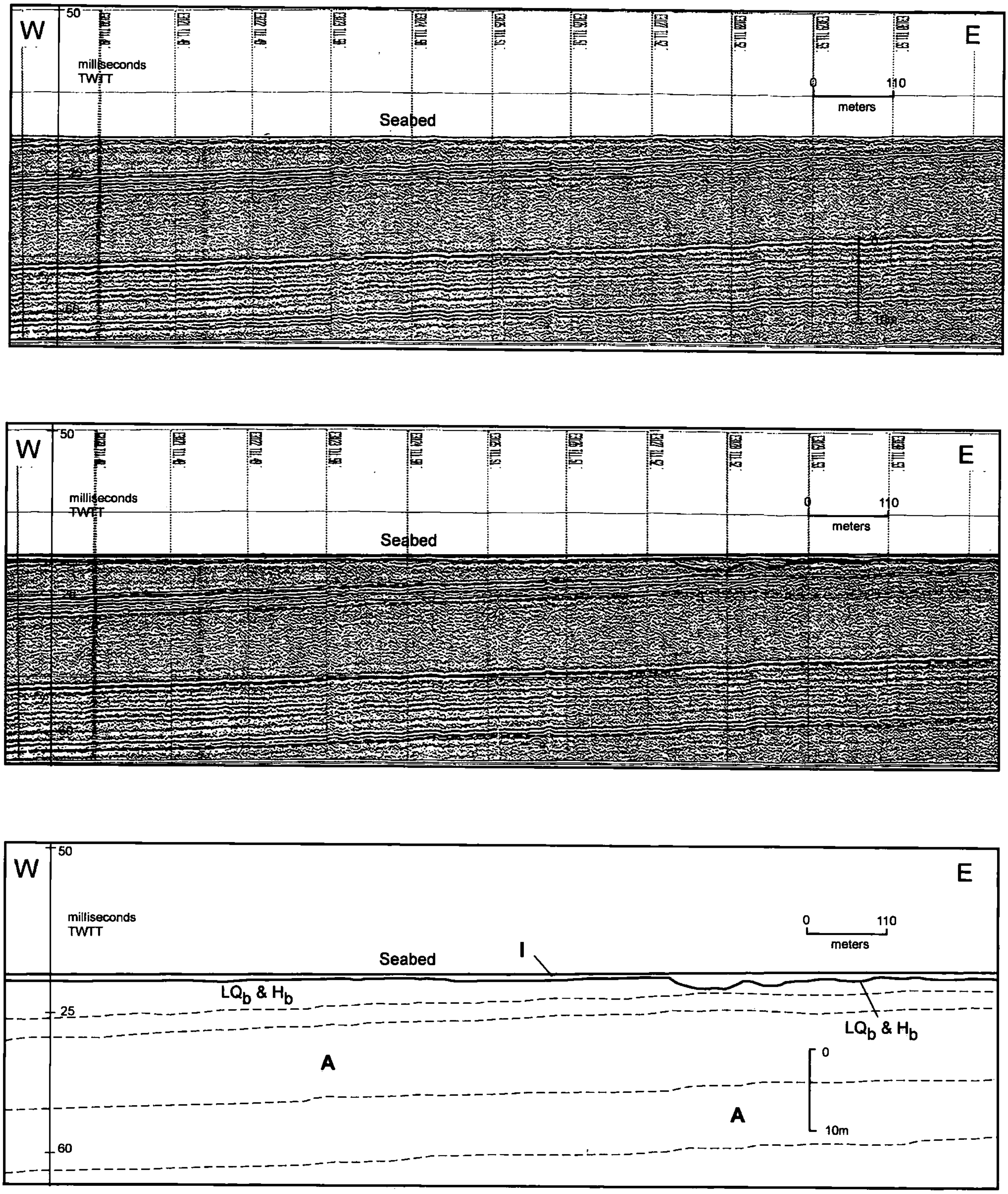


Figure 5-8: Boomer seismic profile section and interpreted line drawing of Line 71 from fixes 3119-3131, showing contiguous sequence boundary LQ_b and Hb. Location of section shown on Figure 5-3.

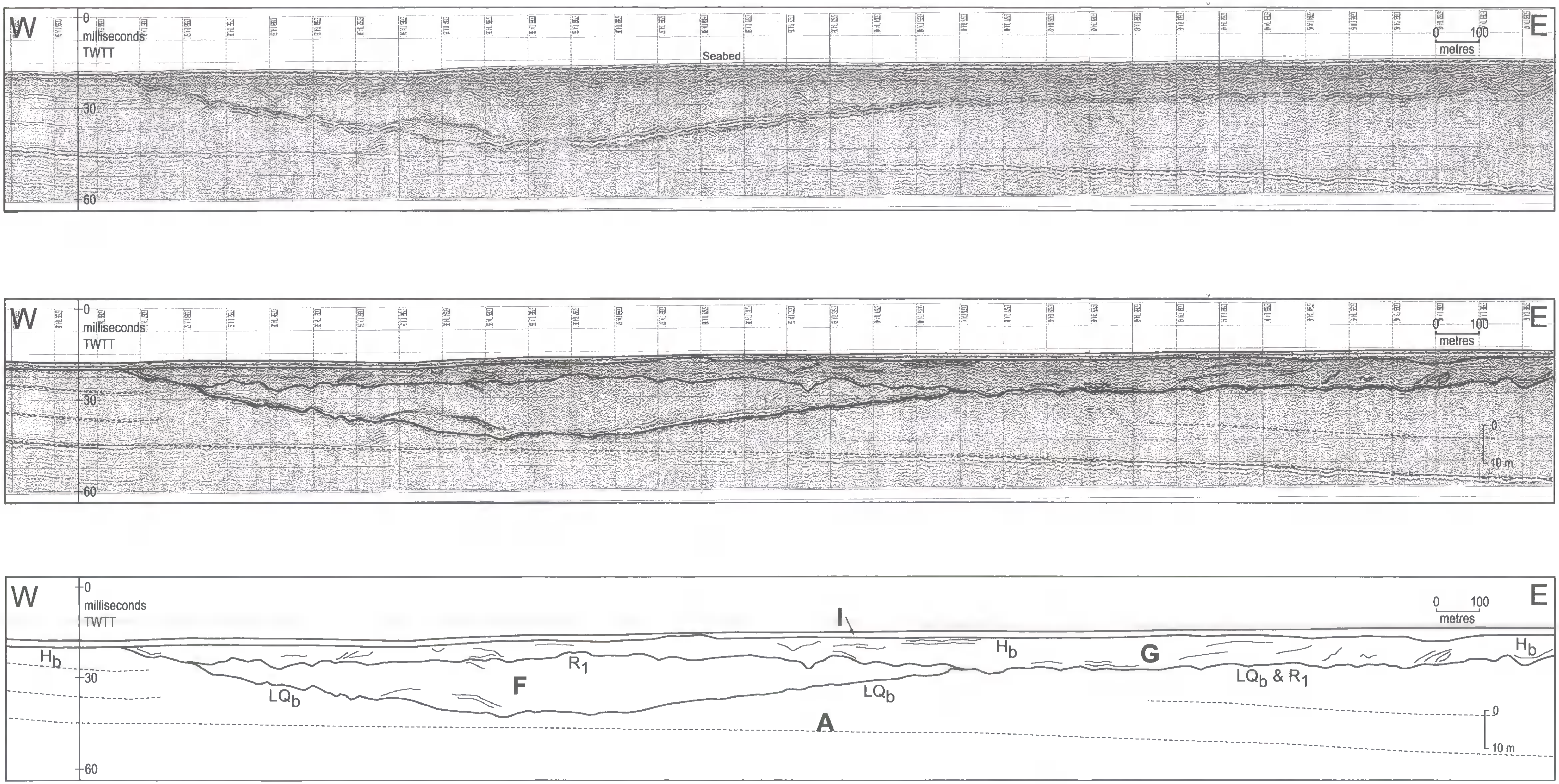


Figure 5.9: Boomer seismic profile section and interpreted line drawing of Line 56 from fixes 7354 - 7389, showing sequence boundary surfaces LQ_b and H_b, and ravinement surface R_1. Also shown are seismic facies units A, F, G and I. Location of section shown on Figure 5-3.

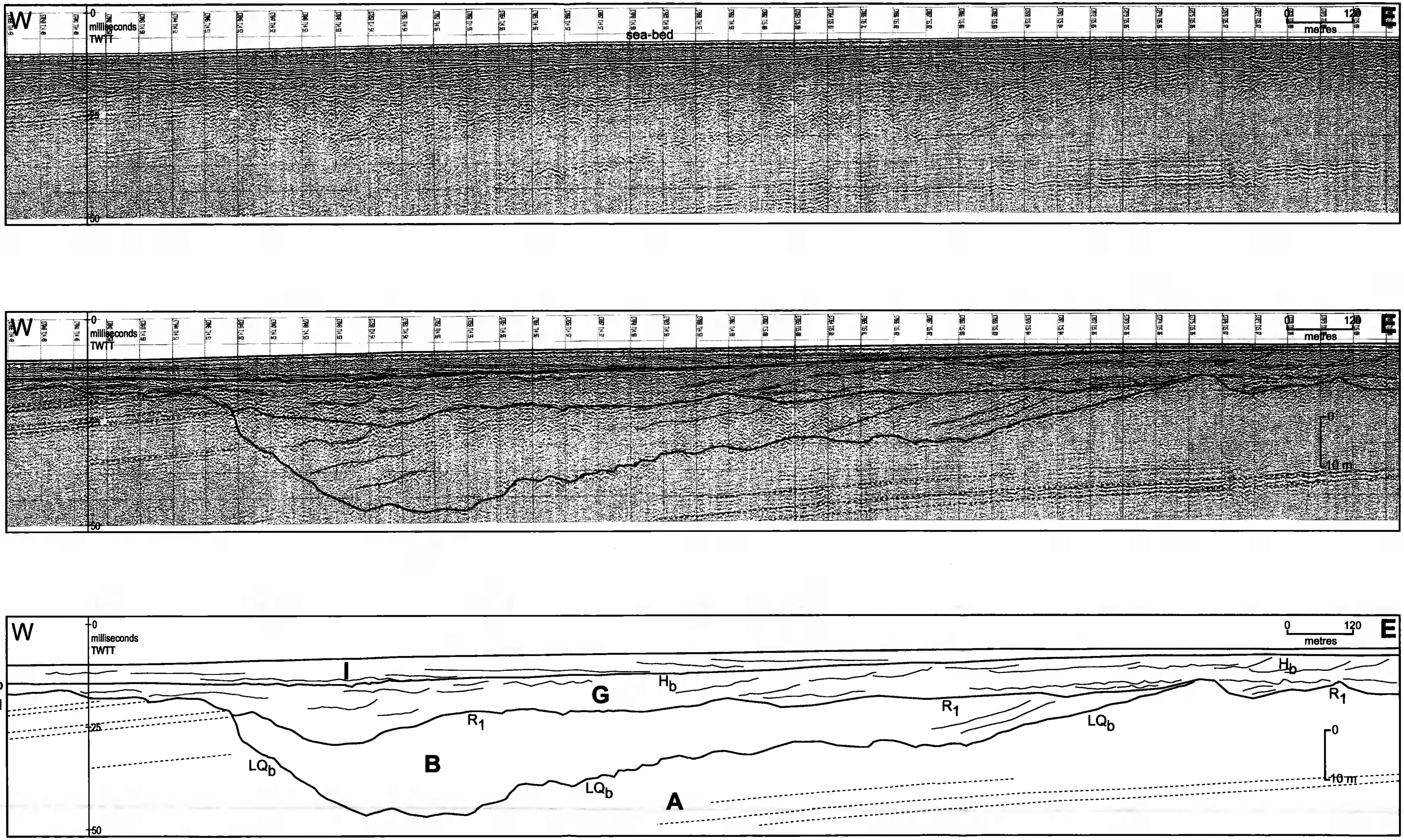


Figure 5-10: Boomer seismic profile section and interpreted line drawing of Line 29 fixes 7039 - 7081, showing sequence boundary surfaces LQ_b and H_b, and ravinement surface R_1. Also shown are seismic facies units A, B, G and I. Location of section shown on Figure 5-3. The vertical scale is two-way travel time in milliseconds.

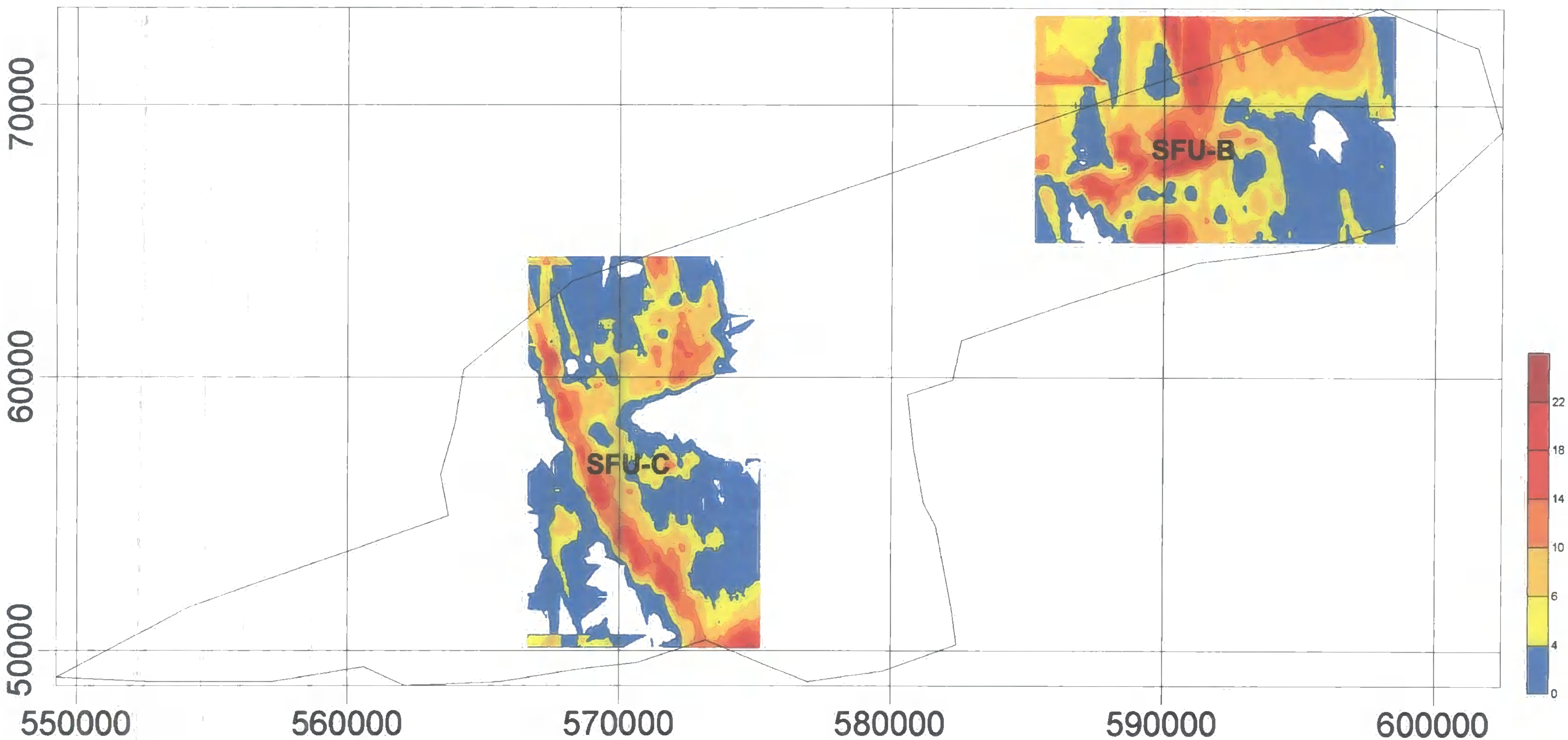


Figure 5-11: Isopachyte map of sediment thickness (in metres) of SFU-B and SFU-C

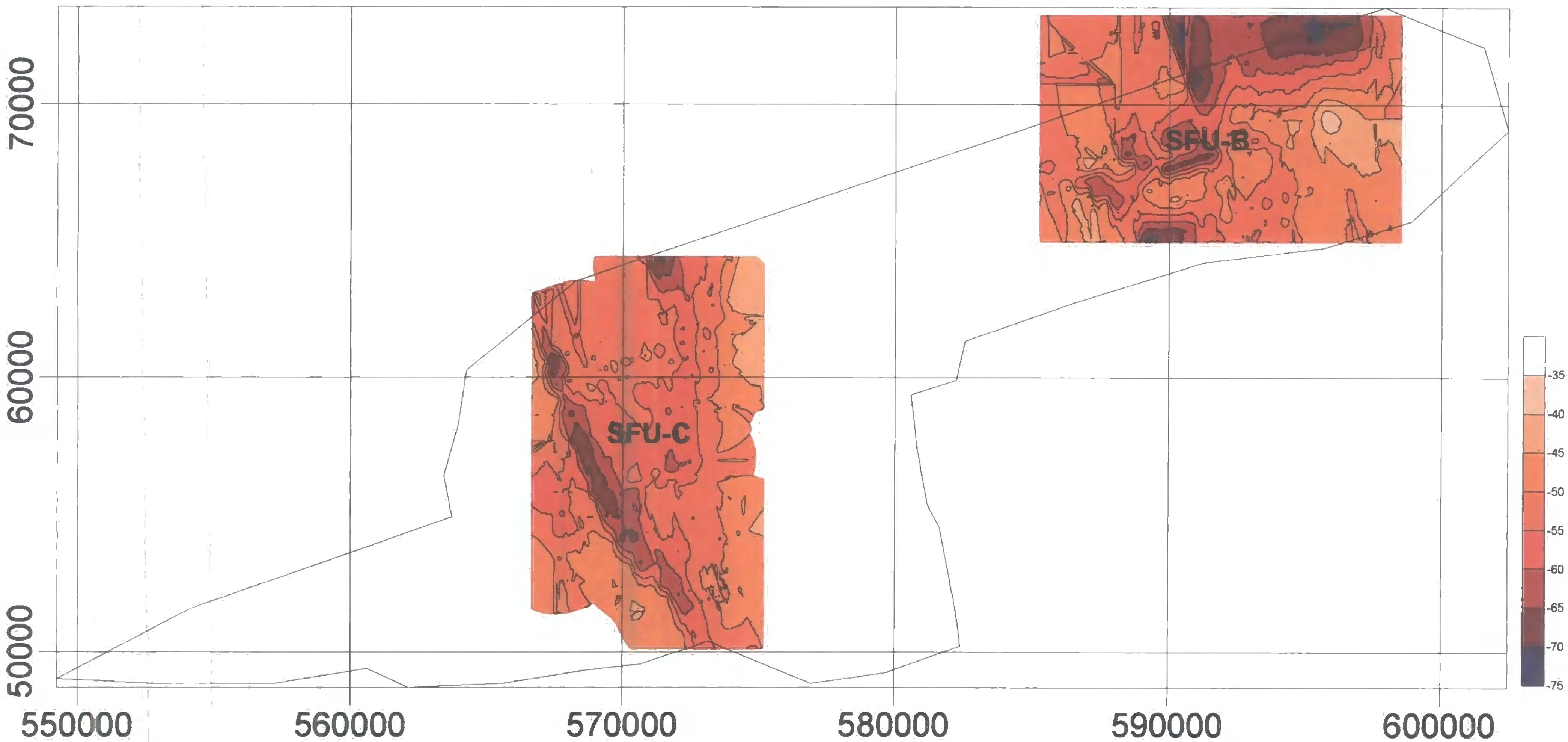


Figure 5-12: Contour map depth (in metres) to the lower bounding reflector surface (LQ_b) of SFU-B amd SFU-C.

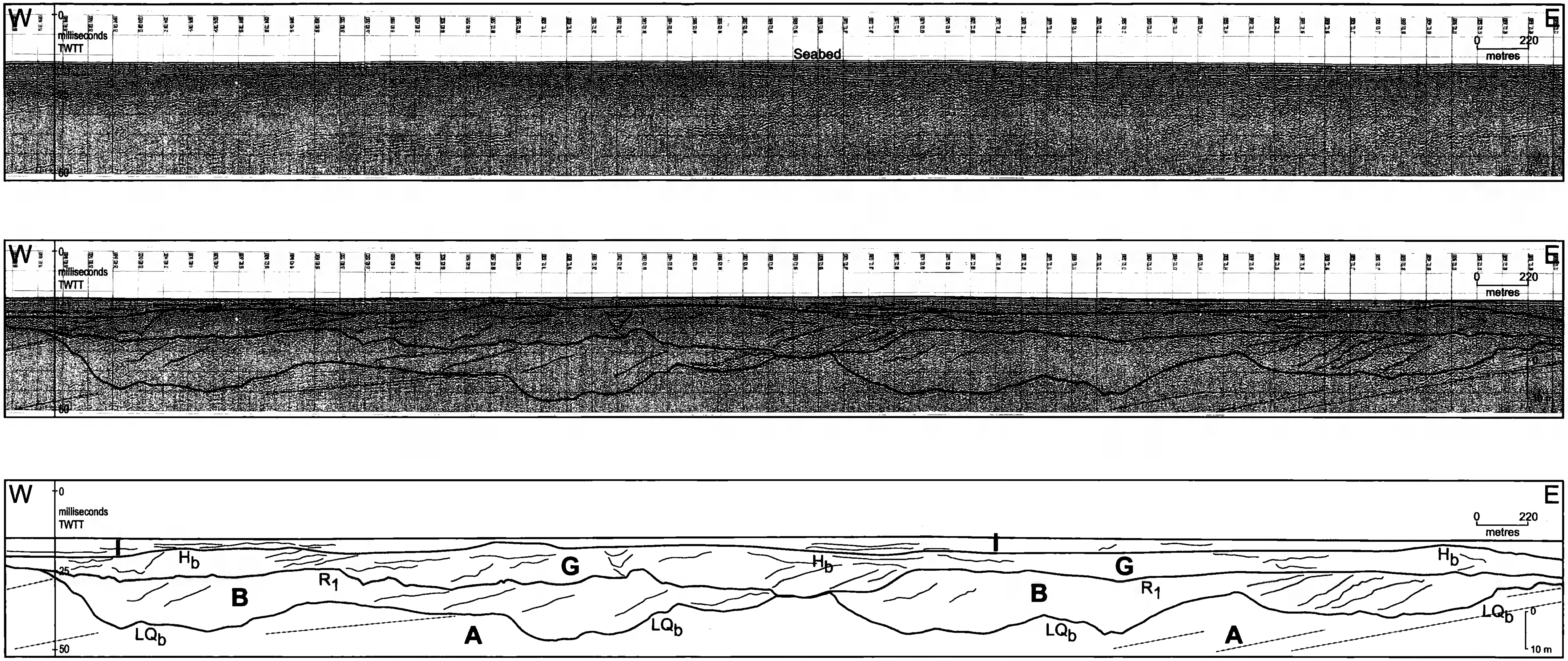


Figure 5-13: Boomer seismic profile section and Interpreted line drawing of Line 31 from fixes 10238 -10299, showing sequence boundary surfaces LQ_b and H_b and ravinement surface R_1. Also shown are seismic facies units A, B, G and I. Location of section shown on Figure 5-3.

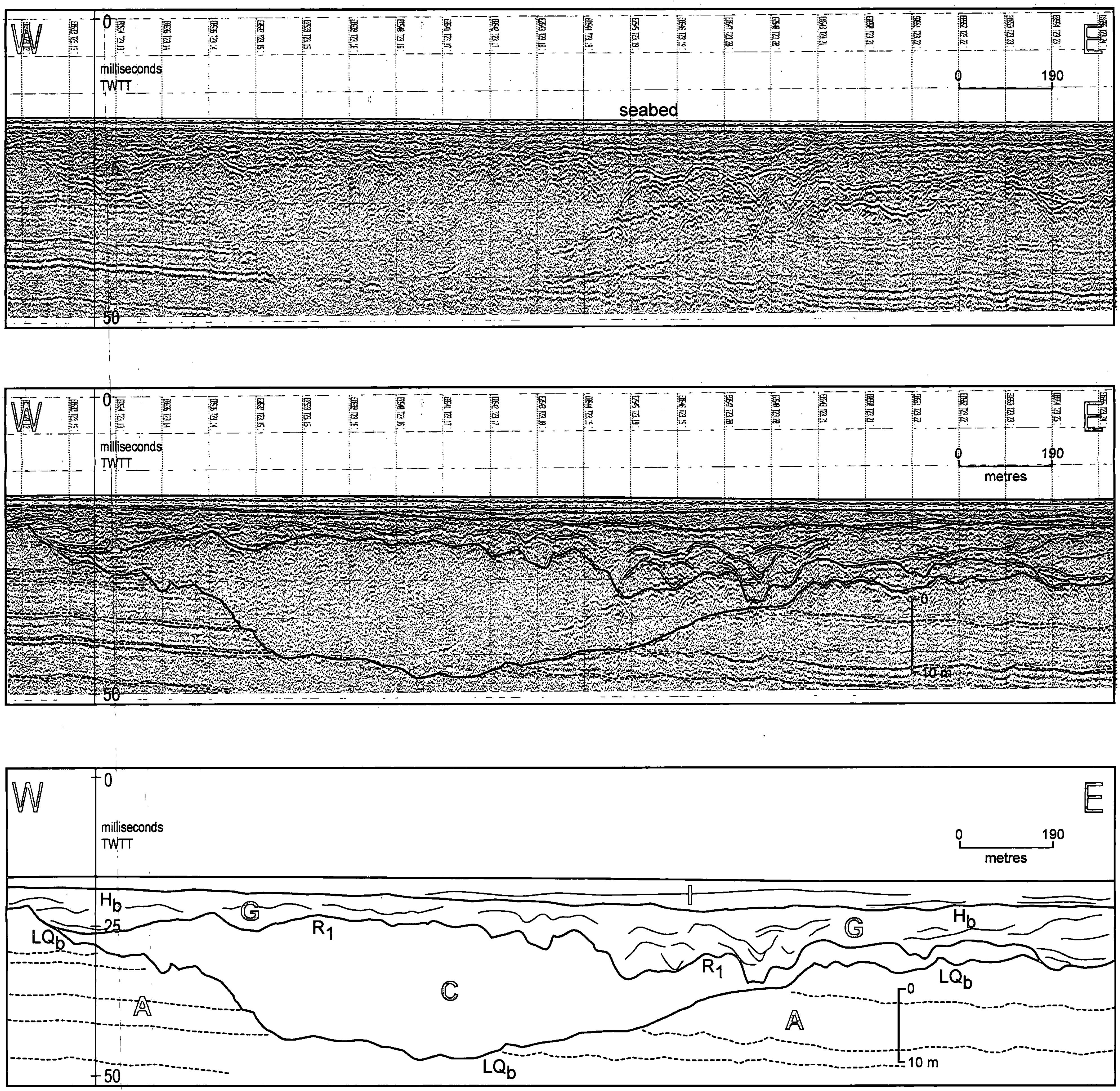


Figure 5-14: Boomer seismic profile section and interpreted line drawing of Line 68 from fixes 0532 - 0555 showing sequence boundary surfaces LQ_b and H_b and ravinement surface R_1. Also shown are seismic facies units A, C, G and I. Location of section shown on Figure 5-3.

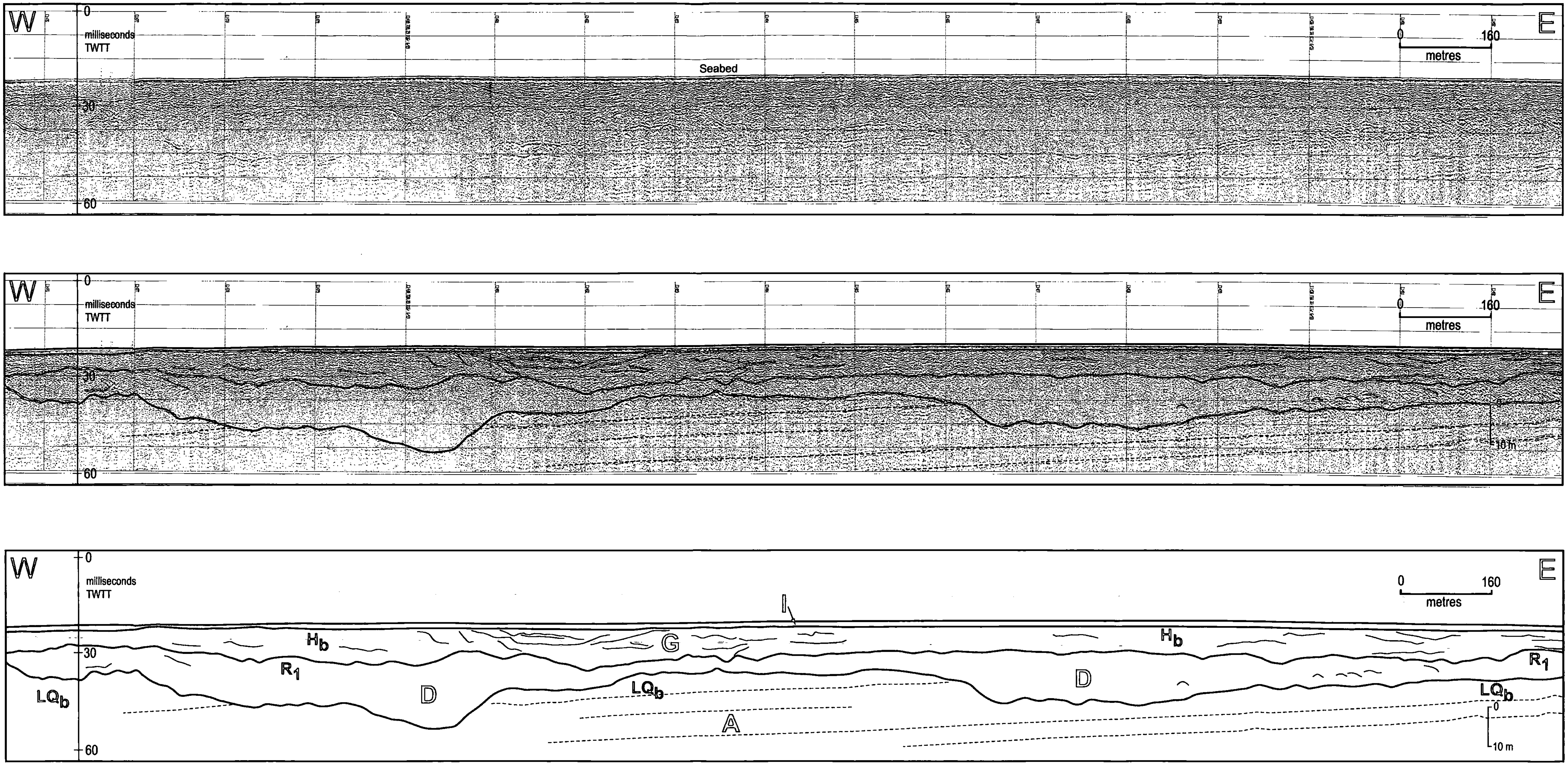


Figure 5-15: Boomer seismic profile section and interpreted line drawing of Line Y5 from fixes 1472 -1496, showing sequence boundary surfaces LQ_b and H_b, and ravinement surface R_1. Also shown are seismic facies units A, D, G and I. Location of section shown on Figure 5-3.

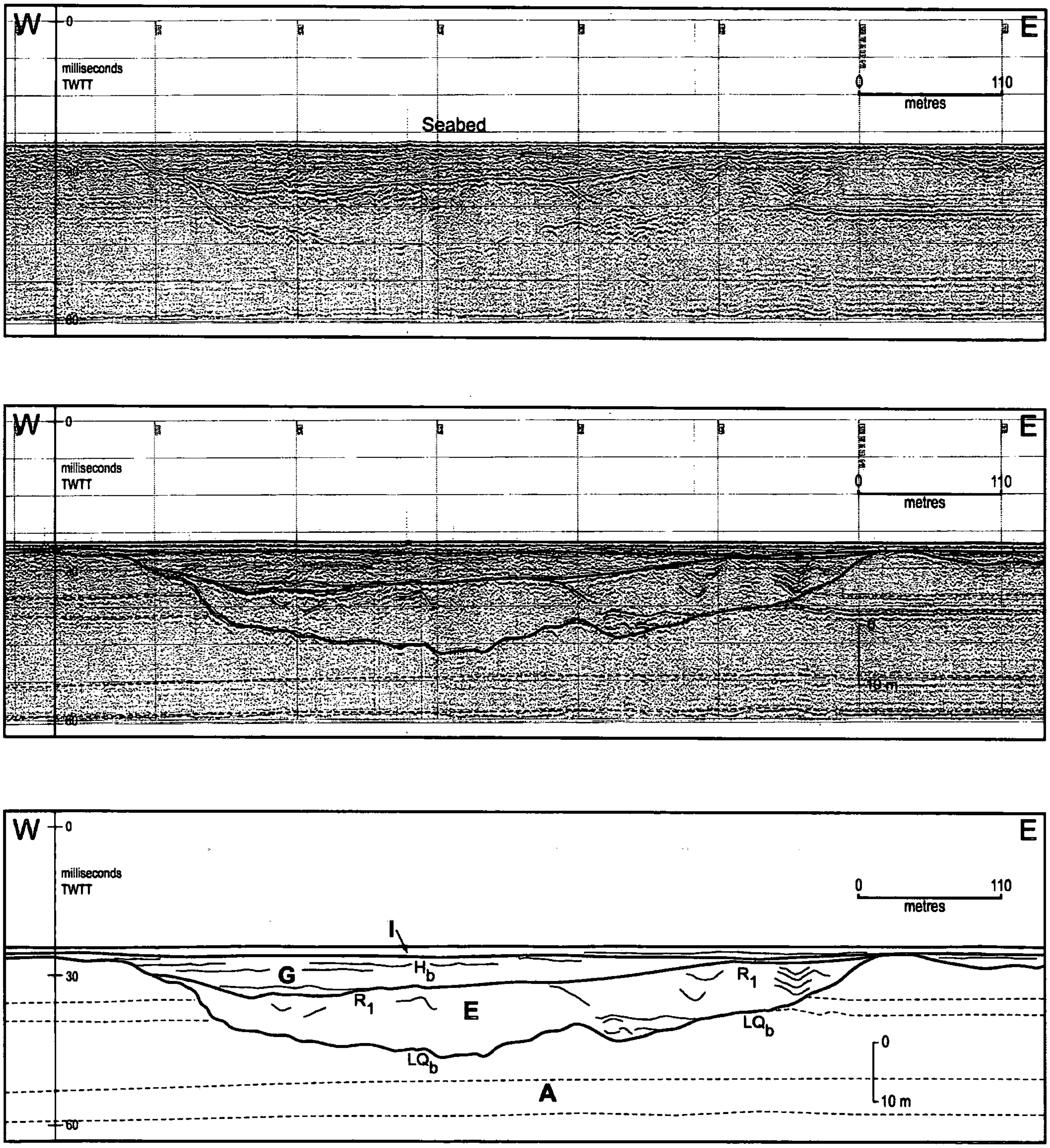


Figure 5-16: Boomer seismic profile section and interpreted line drawing of Line Y2 from fixes 9194 - 9201, showing sequence boundaries LQ_b and H_b and ravinement surface R_1. Also shown are seismic facies units A, E, G and I. Location of section shown on Figure 5-3.

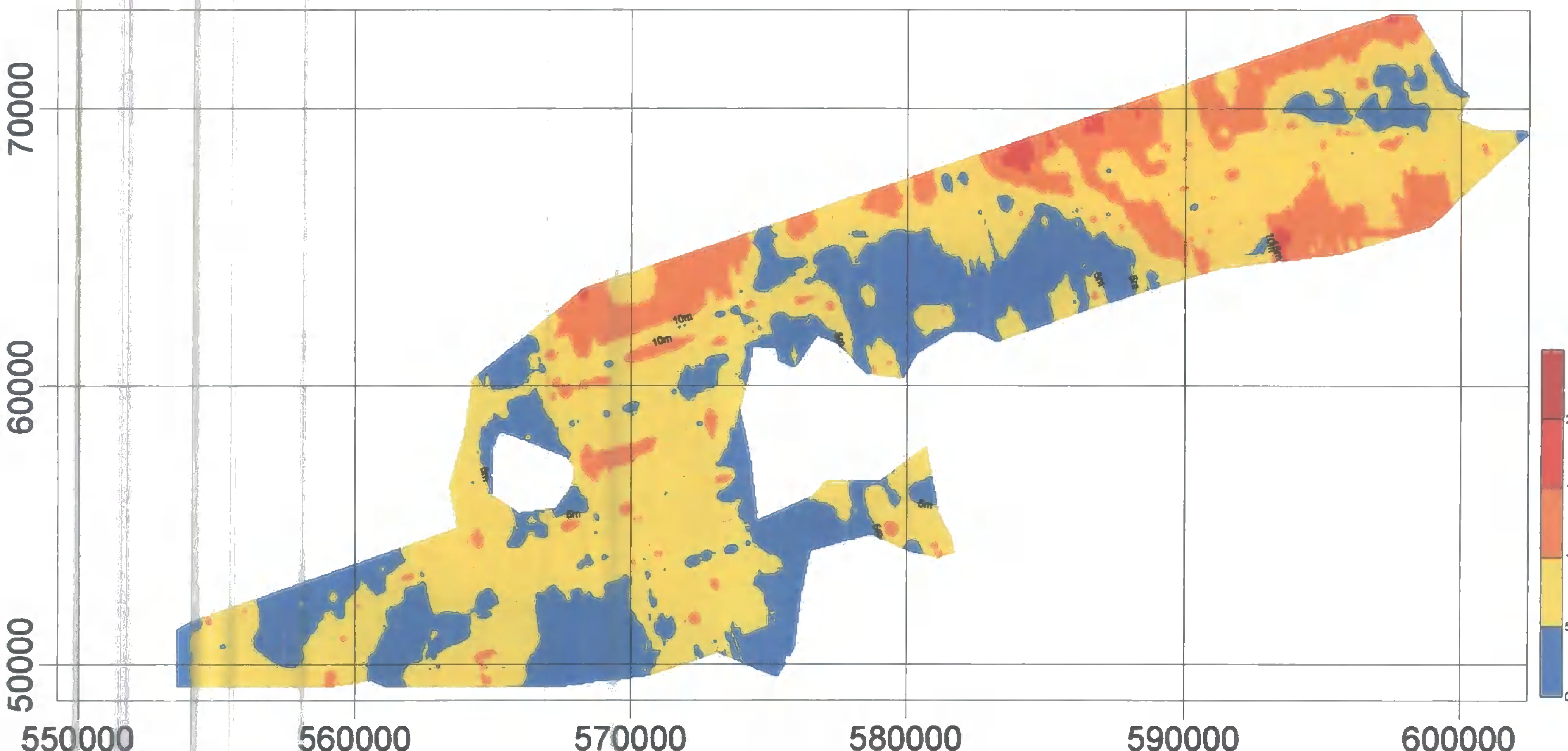


Figure 5-17: Isopachyte map of sediment thickness (in metres) of SFU-G.

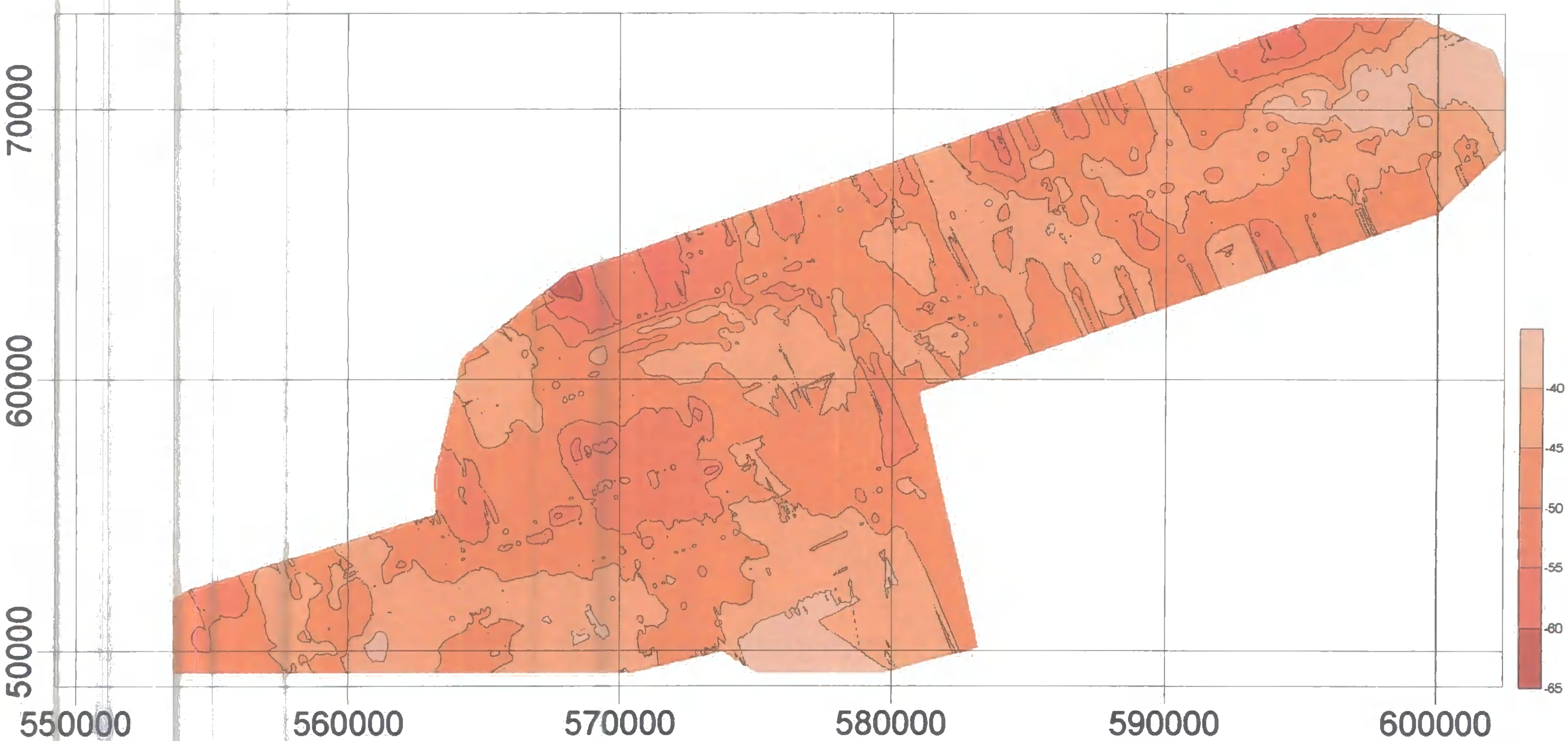


Figure 5-18: Contour map of depth (in metres) to the lower bounding reflector surface (R_1) of SFU-G

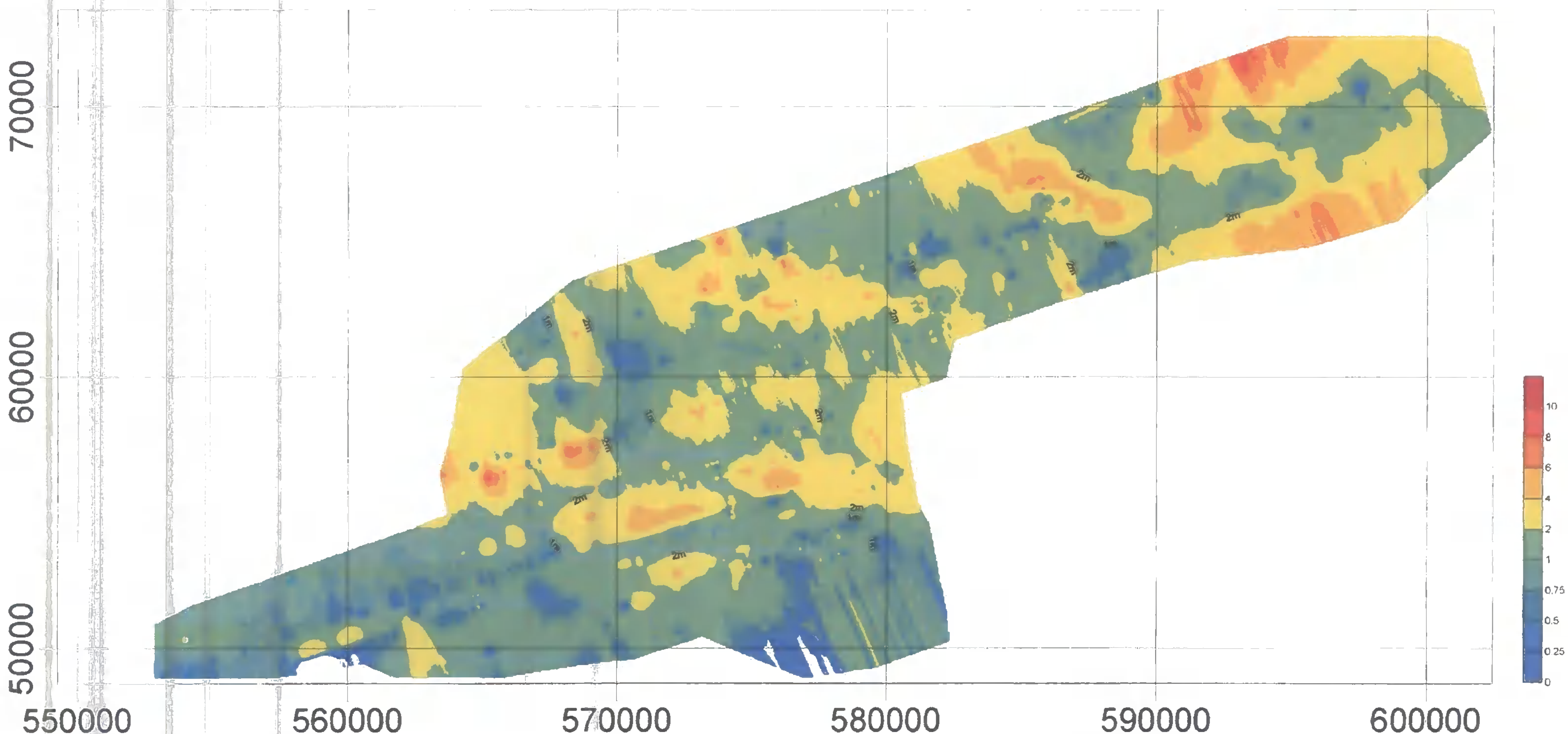


Figure 5-19: Isopachyte map of sediment thickness (in metres) of SFU-I.

Seismic Facies Unit	No. of samples analysed	Dominant Textural Classifications (Folk, 1954)	Mean grain size (ϕ)	Sorting		Skewness		Kurtosis	
				Value	Description	Value	Description	Value	Description
A	44	Muddy Sand - Sand	2.5 – 4.5	1.2 – 2.9	poorly to very poorly	0 – 3.5	Symmetrical to very fine	2 – 19.5	platykurtic to very leptokurtic
B	Not Sampled								
C	4	Slightly Gravely Sand – Gravely Sand	-1.8 – 3.5	0.9 – 3.0	moderately to very poorly	0.5 – 5.0	Fine to very fine	2 - 30	platykurtic to very leptokurtic
D	Not Sampled								
E	Not Sampled								
F	Not Sampled								
G	418	Sandy gravel – Gravely sand	-3.8 - 2.9	0.6 - 4.9	Moderately well to very poorly	-1.25 - 4.7	Very coarse to very fine	1.6 - 37	Platykurtic to very leptokurtic
H	8	Muddy Sand - Sand	3.0 – 3.6	1 – 1.35	moderately to poorly	3.9 – 5.6	Fine to very fine	20 - 42	very leptokurtic
I	204	Muddy sandy gravel – Gravely sand	-3.1 – 1.9	0.9 - 4.5	Moderately to extremely poorly	-0.8 - 2	Coarse to very fine	1.5 - 22	Very platykurtic to very leptokurtic

Table 5.2: Sediment grain-size statistics (method of moments) for the seismic facies units sampled during the vibrocore surveys in the PhD study area.

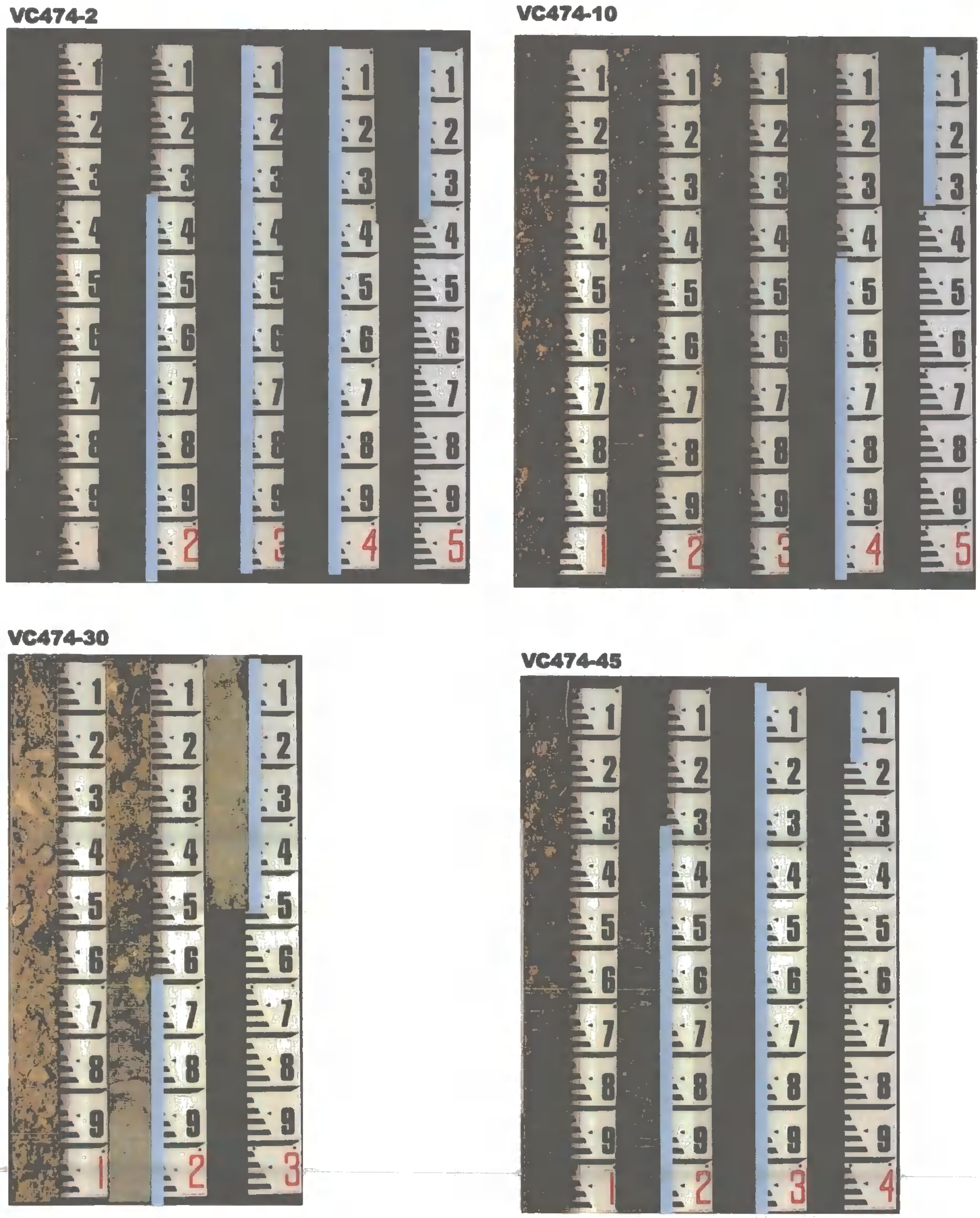


Figure 5-20: Type vibrocore sections for seimic facies unit SFU-A. Highlighted by light blue.

Figure 5-21: Type vibrocore sections of seismic facies unit SFU-C. Highlighted in light blue.

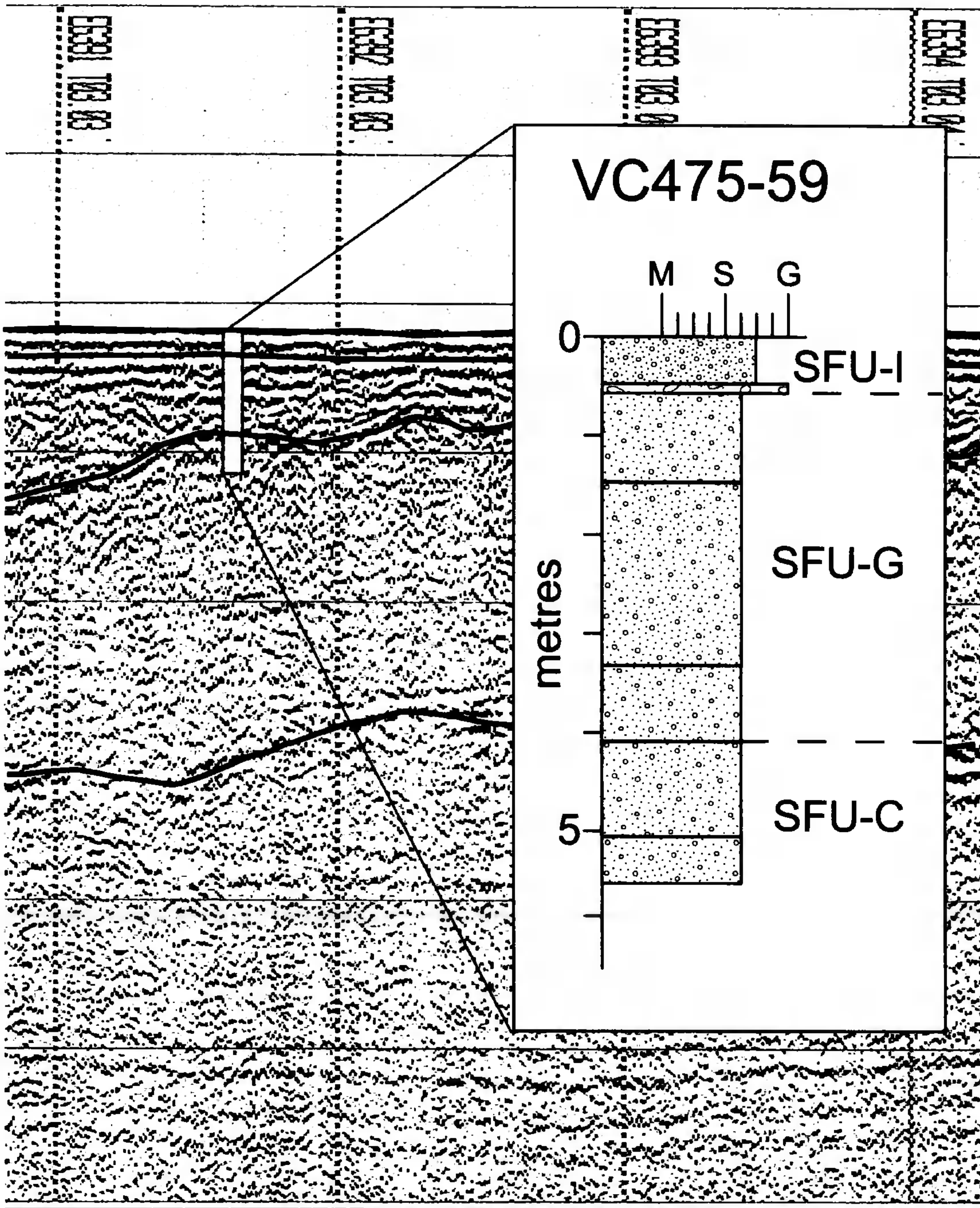


Figure 5-22: Intercomparison of vibrocore sample VC475-59 with seismic reflection profile from Line 69 closest to sample site. Depth of units in vibrocore show an excellent match with the sequence boundaries identified in the seismic interpretation.

Figure 5-23: Type vibrocore sections of seimic facies unit SFU-G. Highlighted in light blue

VC474-3

Figure 5-24: Type vibrocore section of seismic facies unit SFU-H. Highlighted in light blue.

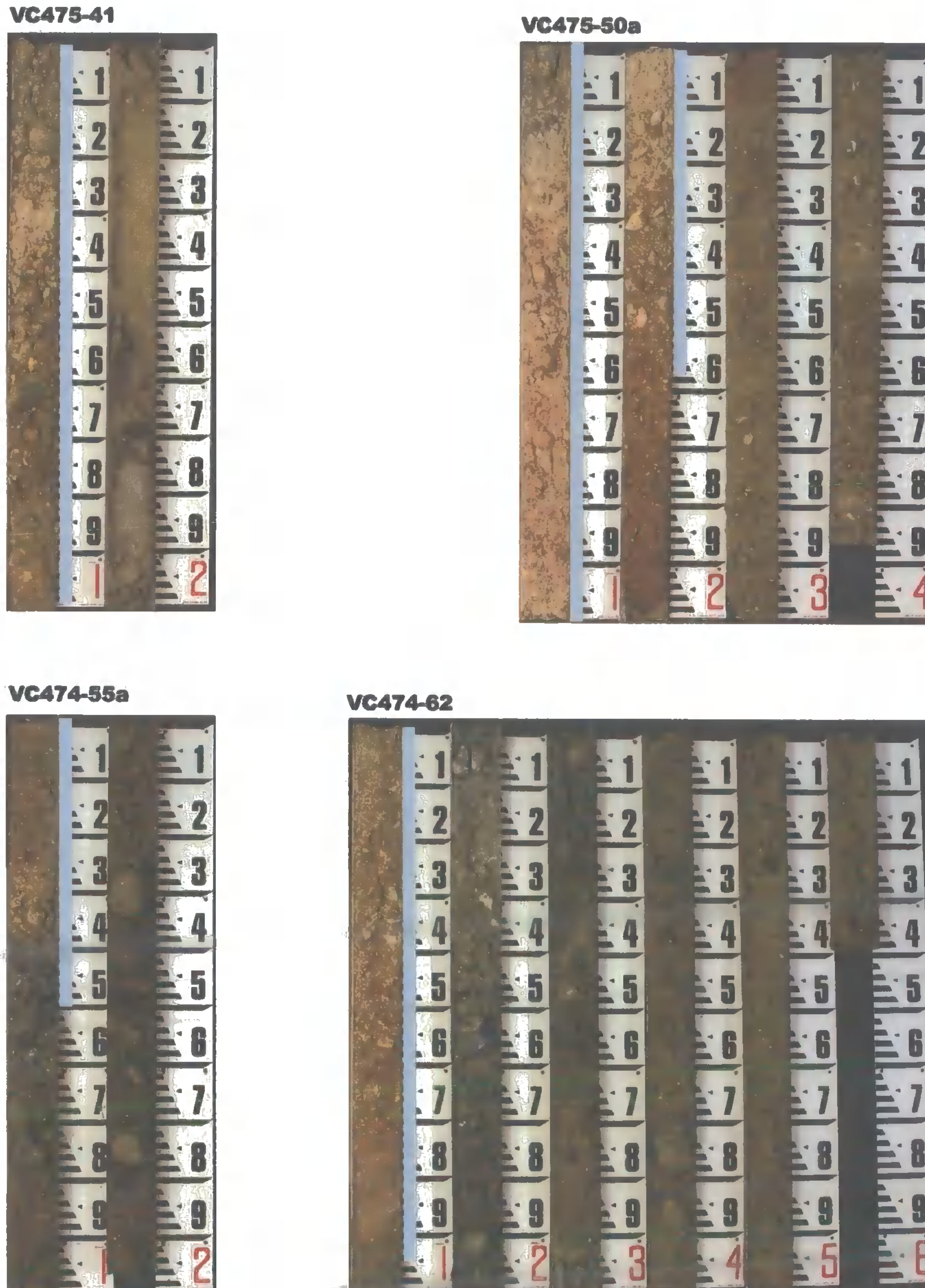


Figure 5-25: Type vibrocore sections of seismic facies unit SFU-I. Highlighted in light blue.

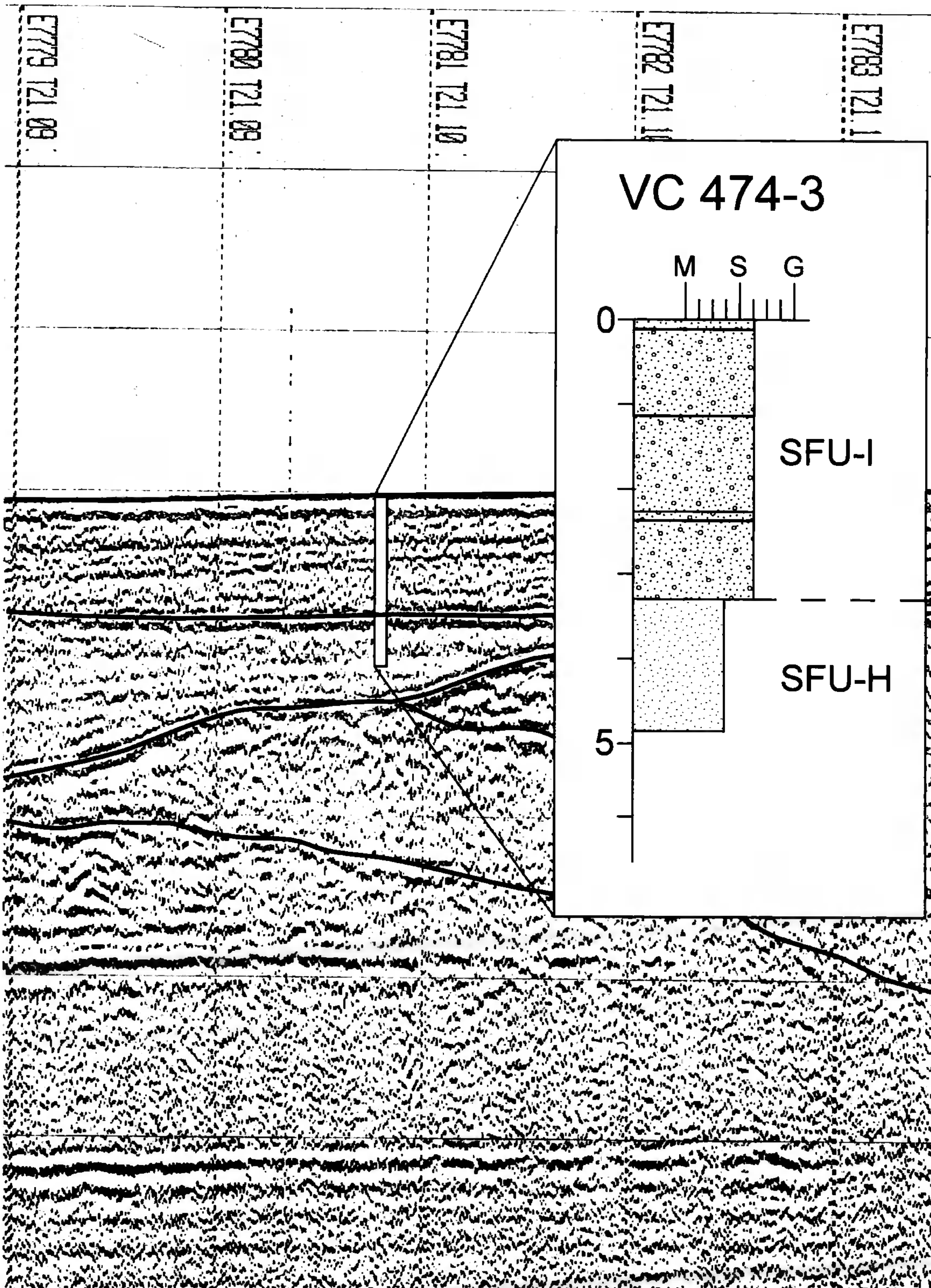


Figure 5-26: Comparison of vibrocore VC474-3 with seismic interpretation of a section of Line 32 closest to sample site showing excellent agreement between interpretation and ground truthed sediment data.

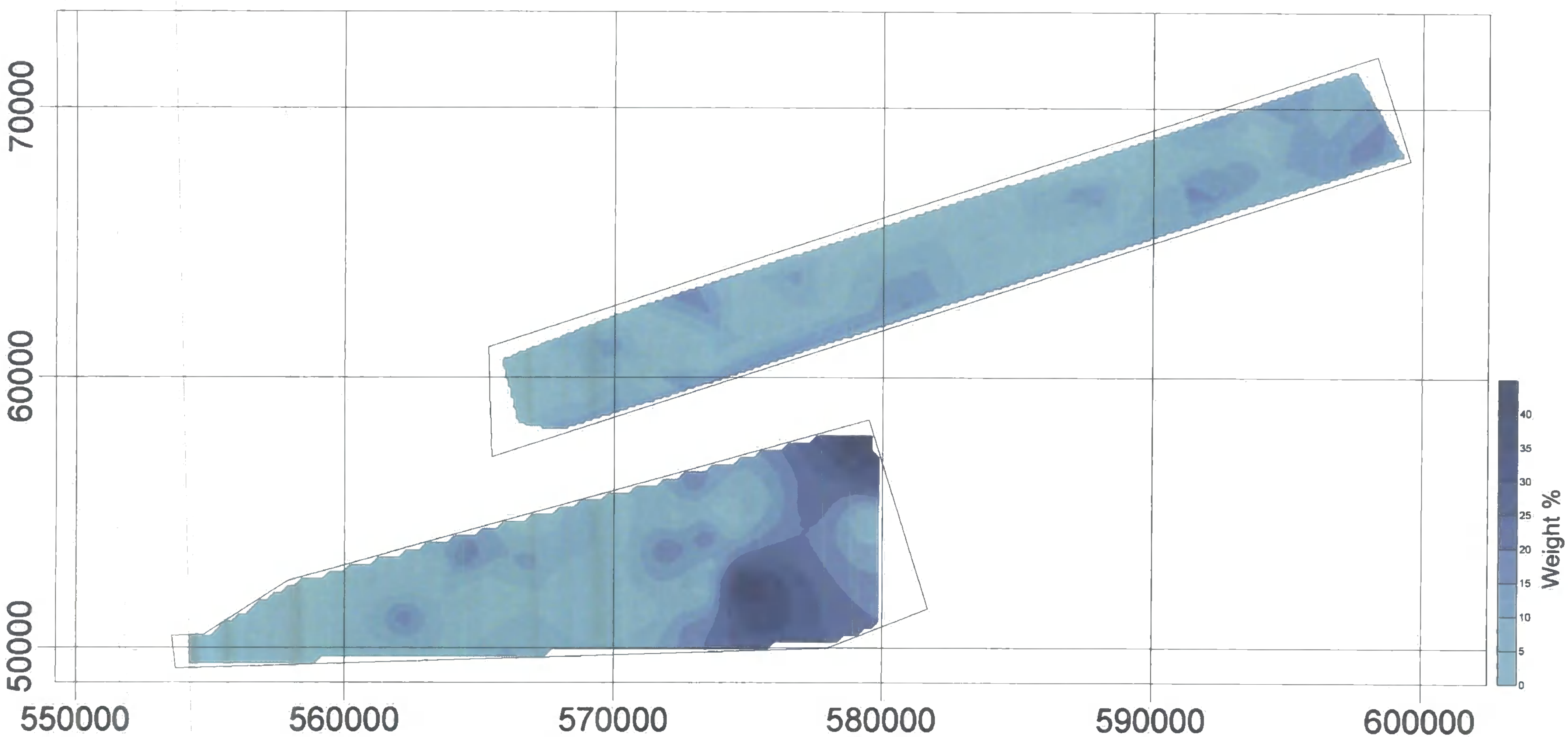


Figure 5-27: Shell content in sea bed sediments of the PhD study area as a percentage of the total sediment weight. Calculated from tops of the vibrocore samples.

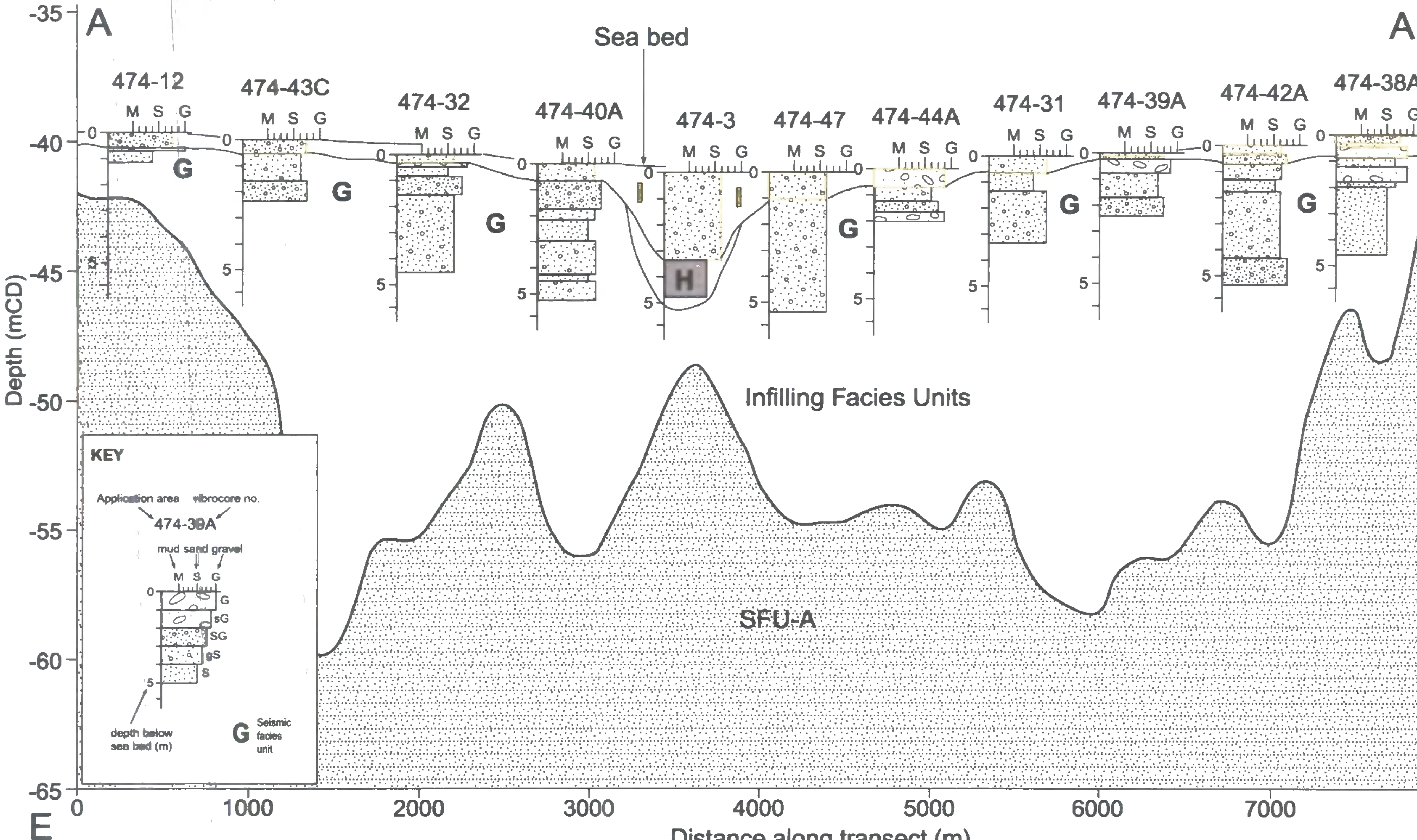


Figure 5-28: Transect A-A: Showing depth to upperbounding reflector of SFU-A along the transect and lithological logs of vibrocore samples extracted from the overlying facies units. Also shown are the facies units sampled (SFU-I in yellow)Location of transect shown in Figure 4-7.

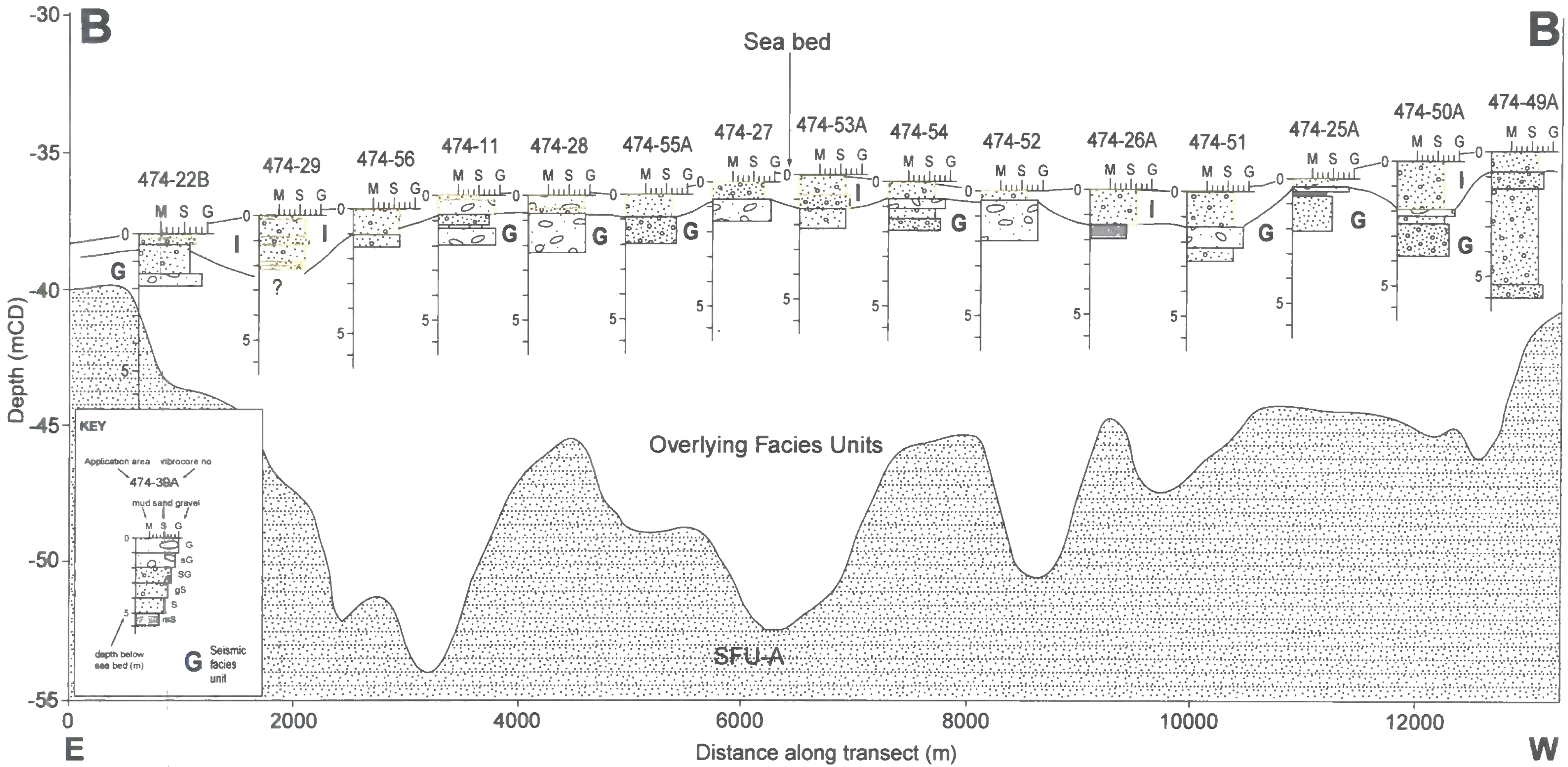


Figure 5-29: Transect **B-B**: Showing depth to upperbounding reflector of SFU-A along the transect and lithological logs of vibrocore samples extracted from the sea bed. Also shown are the facies units sampled (SFU-I in yellow). Location of transect shown in Figure 4-7.

Sample	Depth (cm)	*Actinoptychus senarius*	*Diploneis crabro*	*Diploneis didyma*	*Hyalodiscus scoticus*	*Melosira westii*	*Parala sulcata*	*Thallassiosira spp.*	Sponge	Forams
474-3	420-422									R
	470-472								R	R
474-12	10 - 14	R	R			C	R		A	C
	50-54								A	
474-32c	34-38								R	
	54-56								R	
	450-454								R	
474-40 A	10-14			R	R	C	R		A	A
	50-54								C	R
474-47	80-84								R	
474-44A	15-19					R	R		A	C
474 -31A	13-17		R	R	R	R	C		A	A
	16-20			R		C	C		A	C
474-42A	20 -24						R		C	C
	64-68								R	
	250-254								A	
	350-354								A	
	526 – 530								R	
474-60	30-32								R	
	40-42								C	R
	50-52								A	
	60-62								R	
	70-72								R	
	80-82								R	
	170-172								R	R
474-22 B	30-34					R			A	A
	84-88								C	
	170-174								R	
474-29	50-54				R		R		A	R
	150-154						R		A	R
474-56	20-24								R	
474-11	20 - 24					R	R		A	A
	90 - 94								A	R
	116-120								A	R
474-55A	20-25								C	C
	41-42								R	R
	50-54								R	C
474-53 A	10-14	R	R		R	A	A	R	A	A
	40-44								C	R
474-54	16-20			R	R	R			A	A
474-52	16-20					R			A	C
474-51	20-24				R	R			A	A
	86-90								R	C
474-25 A	10-14		R	R		R	R		A	A
	50-54								A	R
	146-180								R	R
474-3	56-60				R	R			A	A
	150-154								A	R
	230-234								R	R
	280								A	R

Table 5.3: Diatom and other biostratigraphical indicator abundance data.

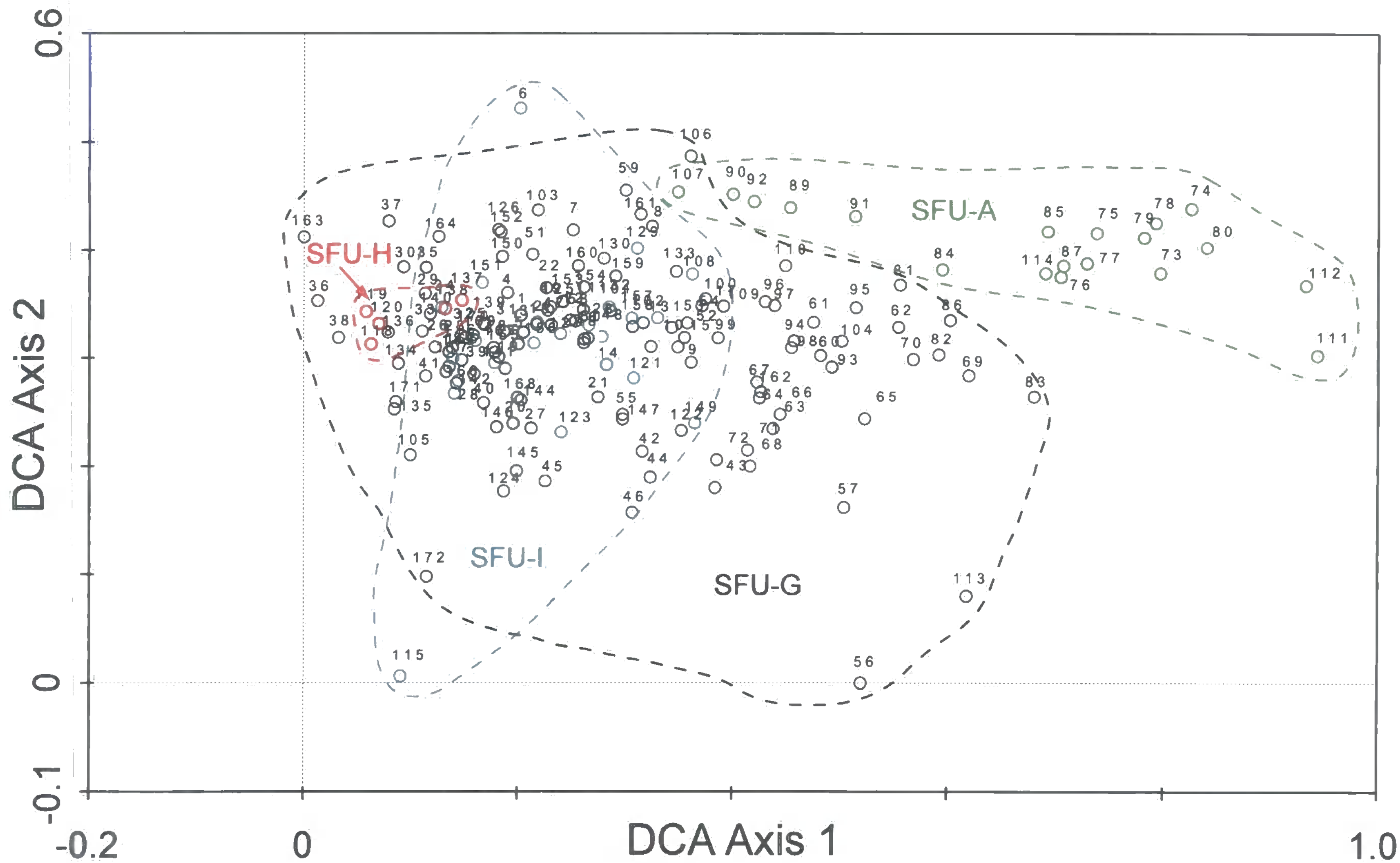
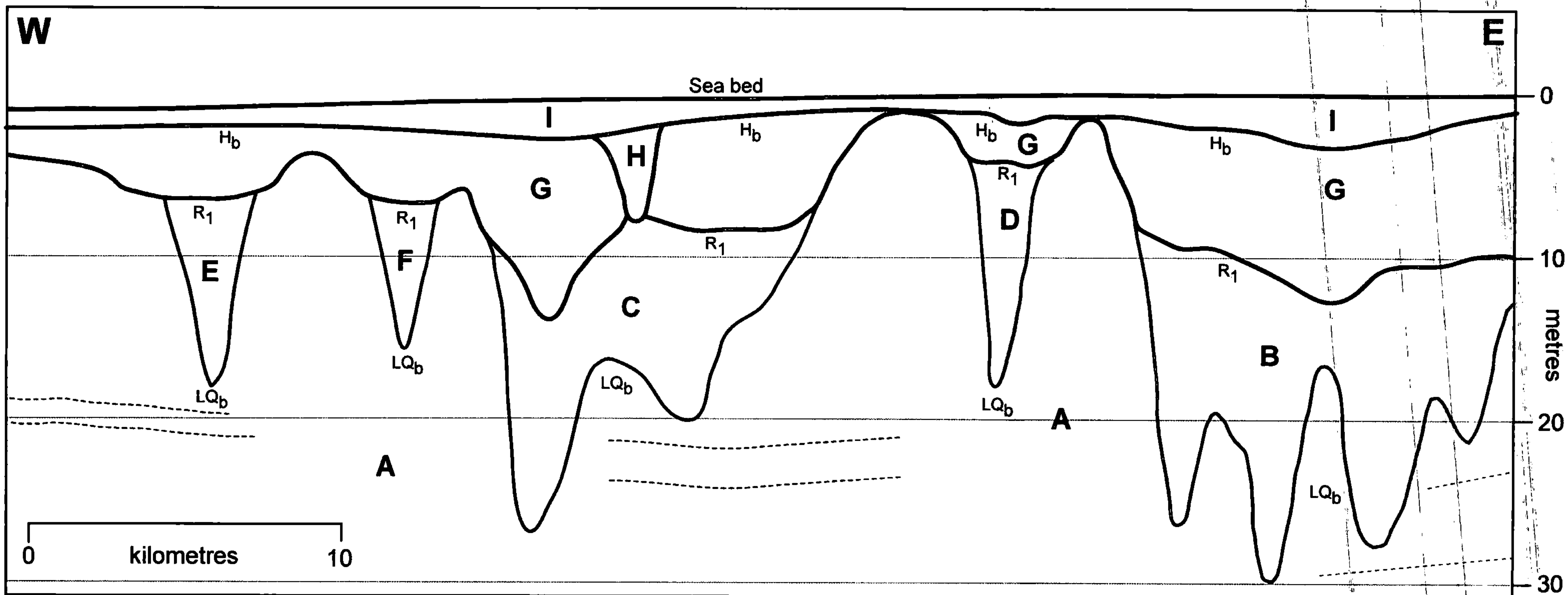
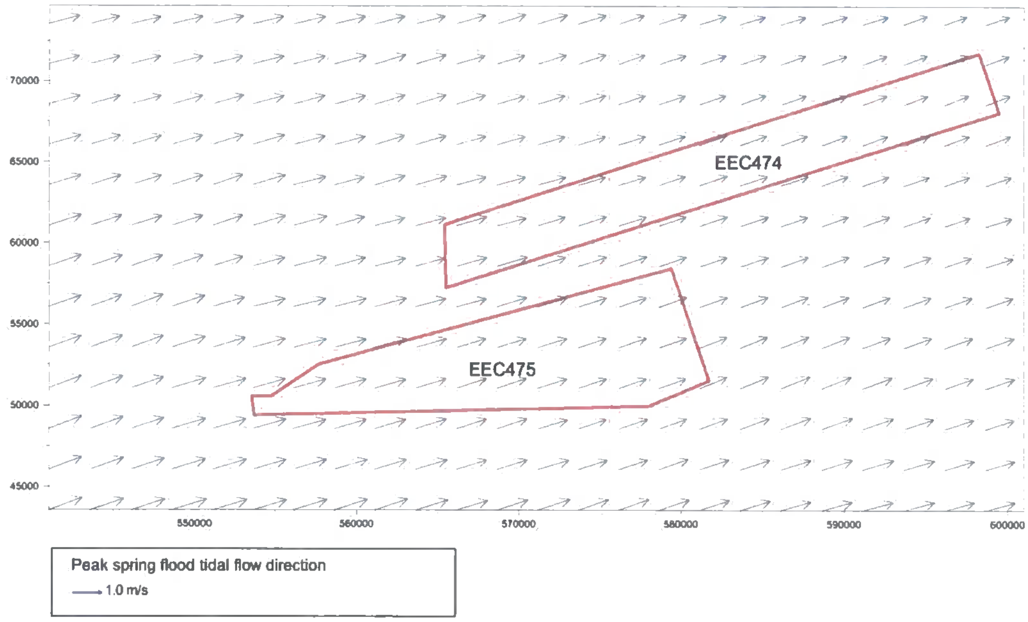


Figure 5-30: Biplot showing detrended correspondance analysis (DCA) of the 171 samples analysed for their mulit-element geochemistry . Highlighted in green are samples of deposits from SFU-A, black - SFU-G, red - SFU-H and blue - SFU-I.



vertical exaggeration x500

Figure 6-1: Schematic representation of an E-W cross-section of the study area showing the three sequence bounding surfaces LQ_b, R_1 and H_b and the eight seismic facies units A-I identified.



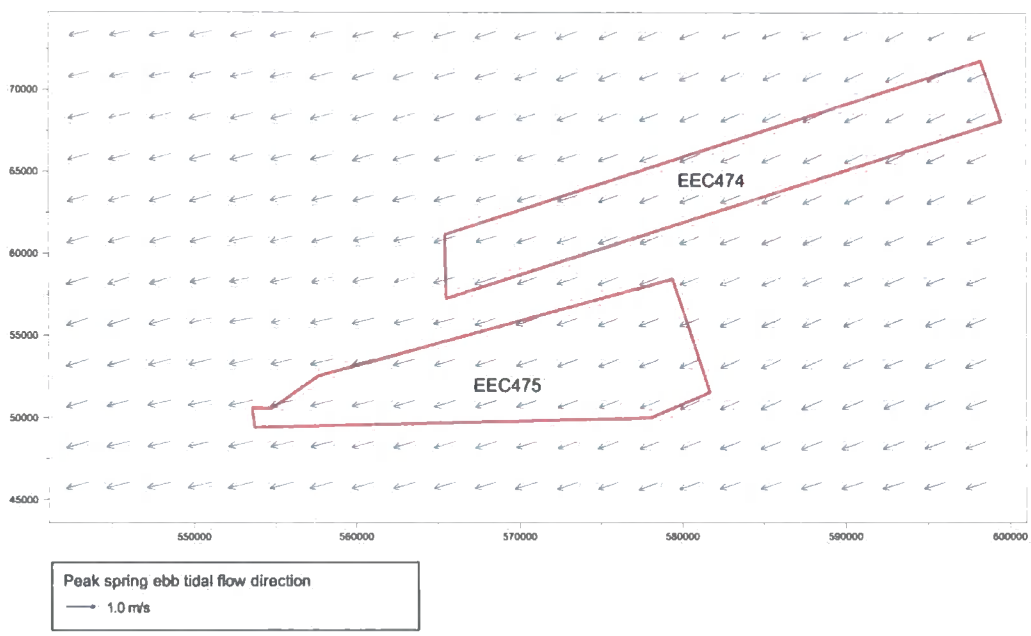


Figure 6-2: Peak tidal flow directions across the PhD study area (modified from EMU Environment Ltd., 2002a)

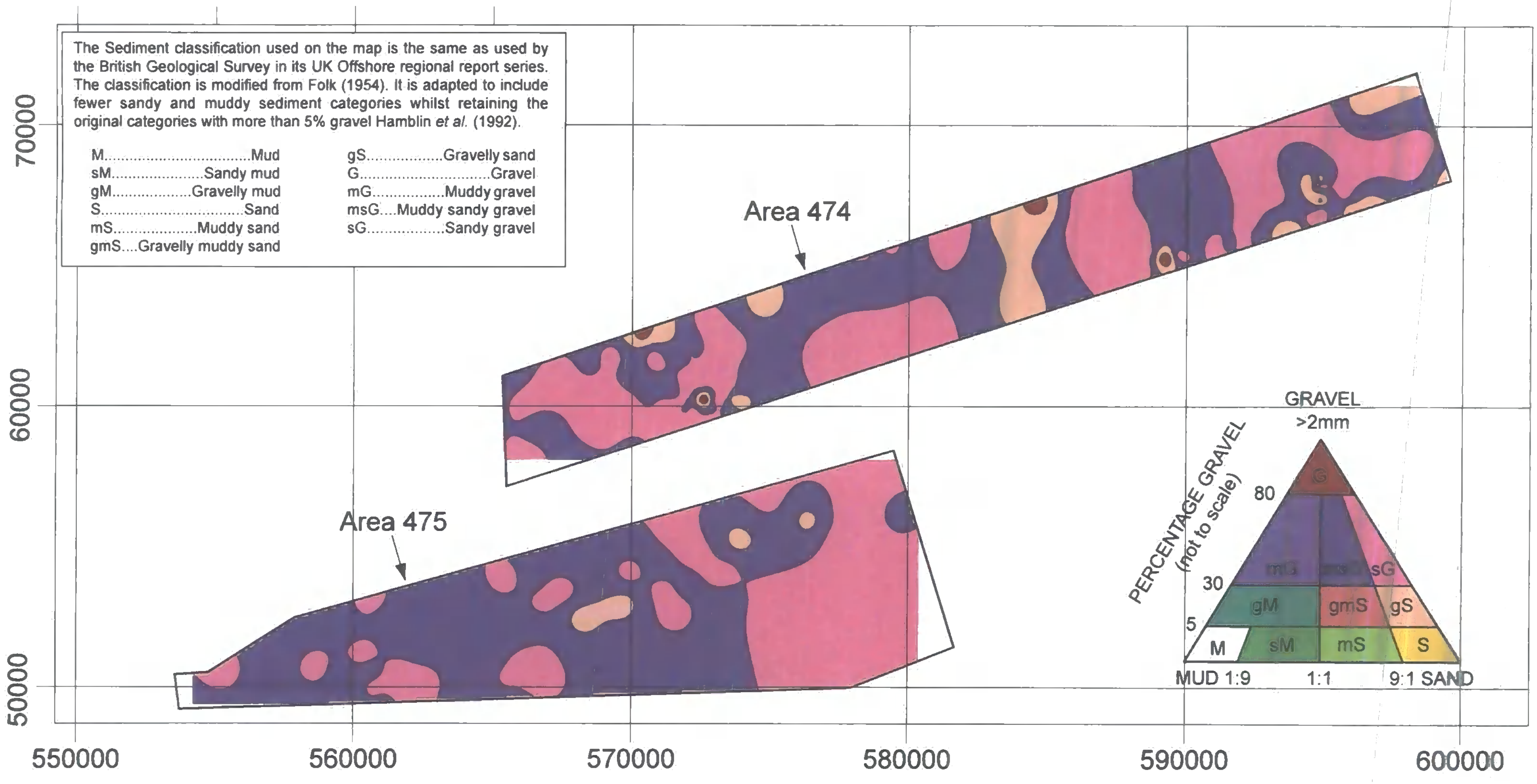


Figure 6-3: Sea bed sediments in the study area as interpreted from the vibrocoring survey. Coordinate system OSGB

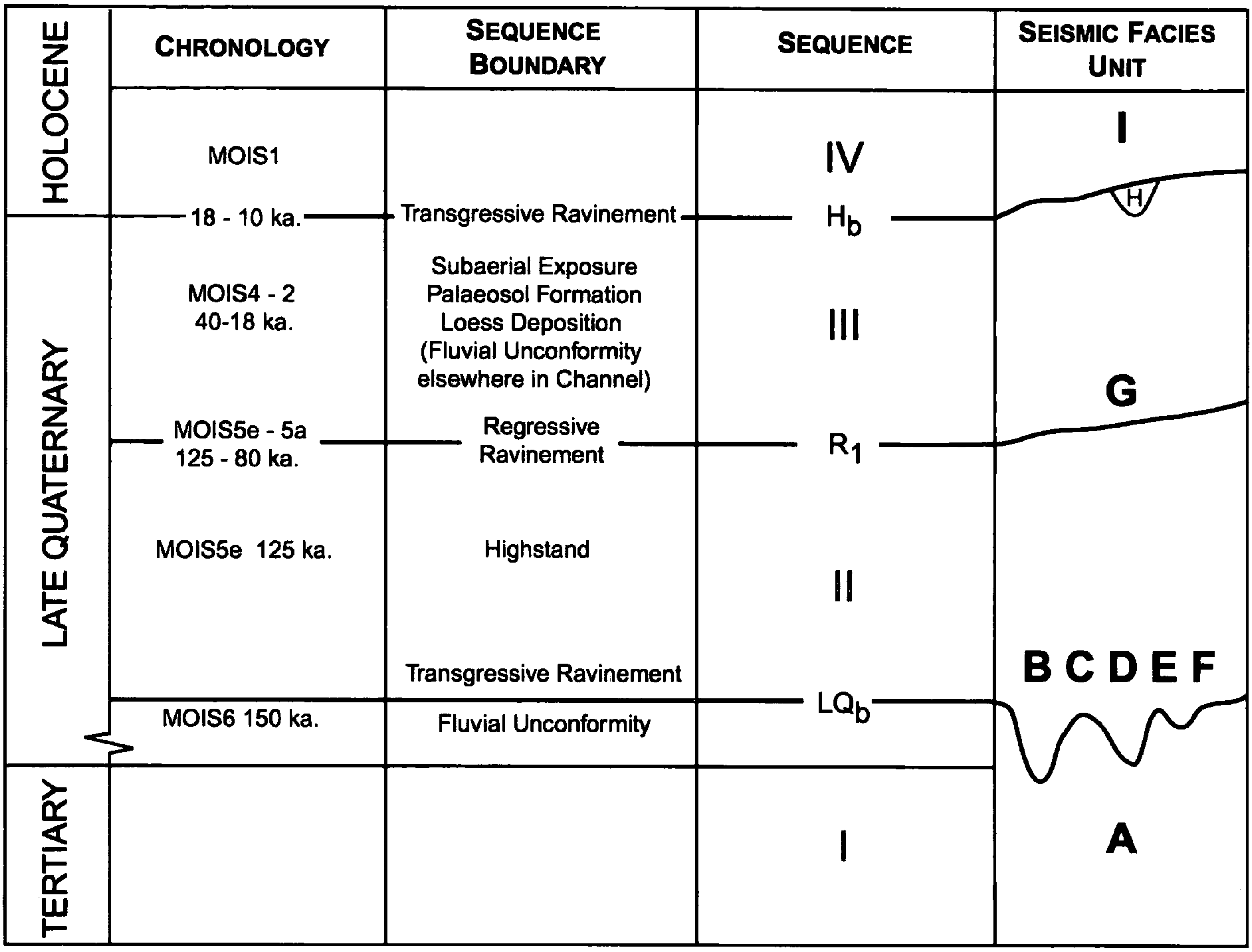


Figure 6-4: Quaternary stratigraphy of the PhD study area.

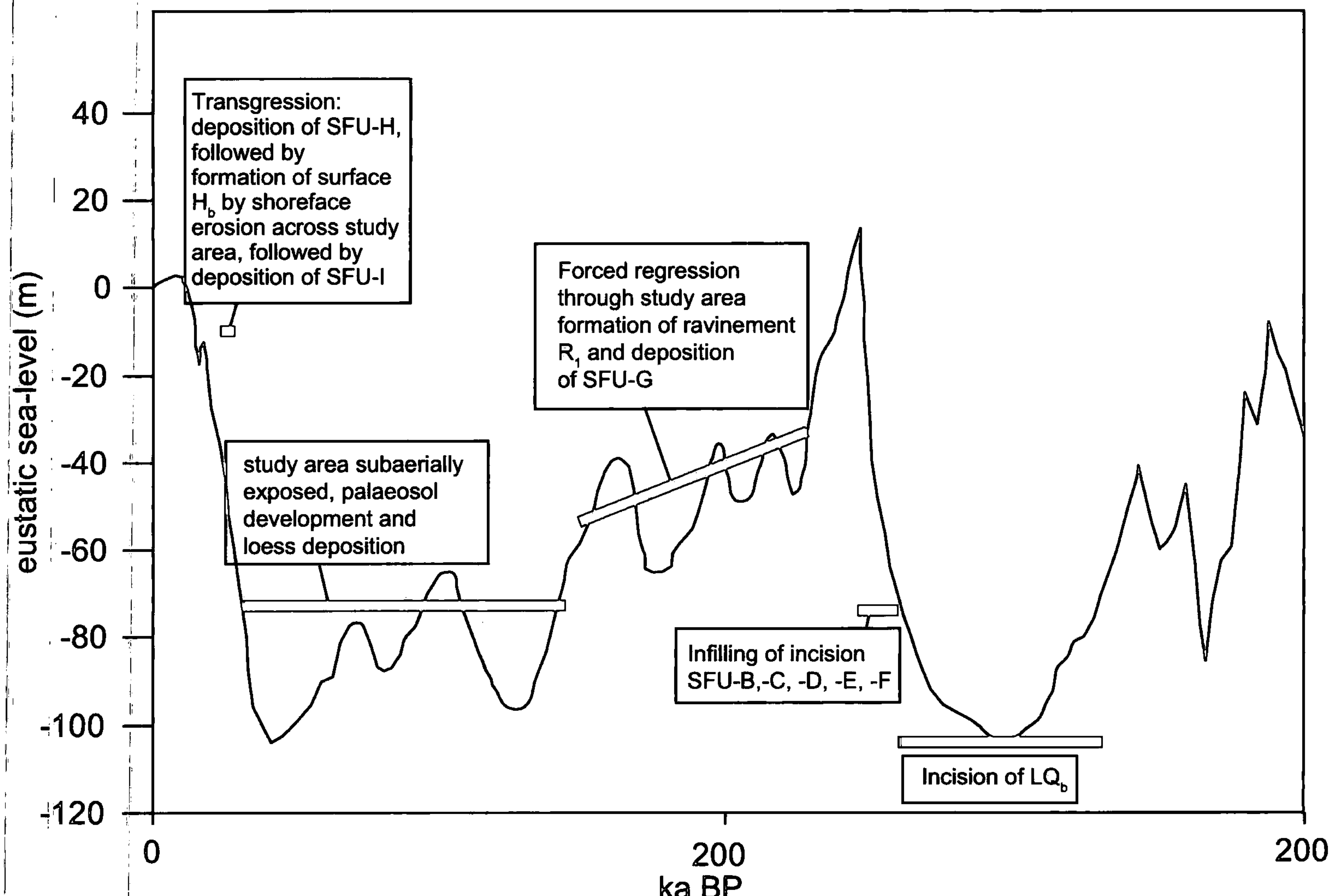


Figure 6-5: Conceptual chronological model of the development of the sequence bounding surfaces and seismic facies units identified in the PhD study area.

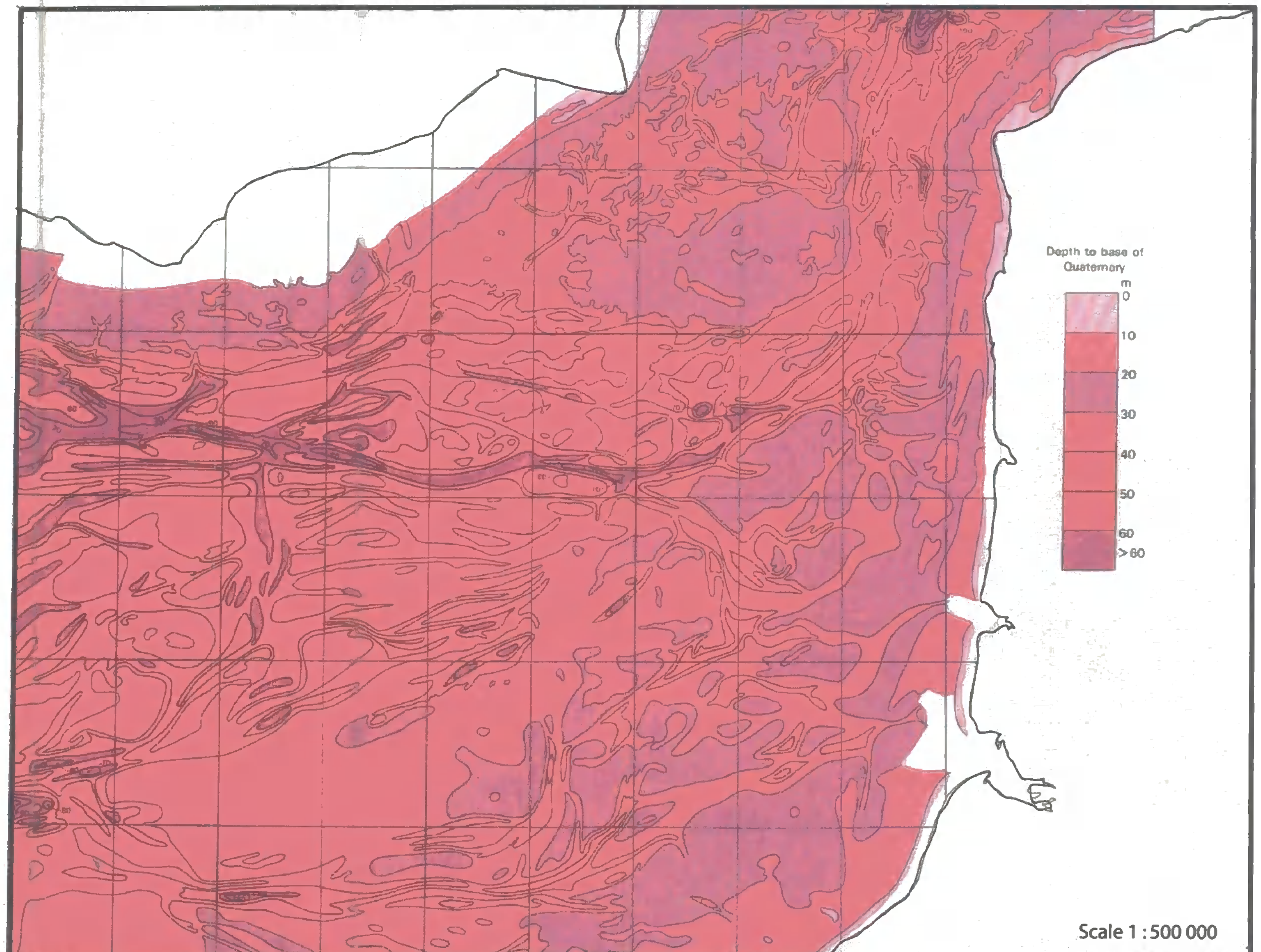


Figure 7-1: Map of rockhead in the eastern English Channel showing the palaeovalleys incisions (from Hamblin, 1989)

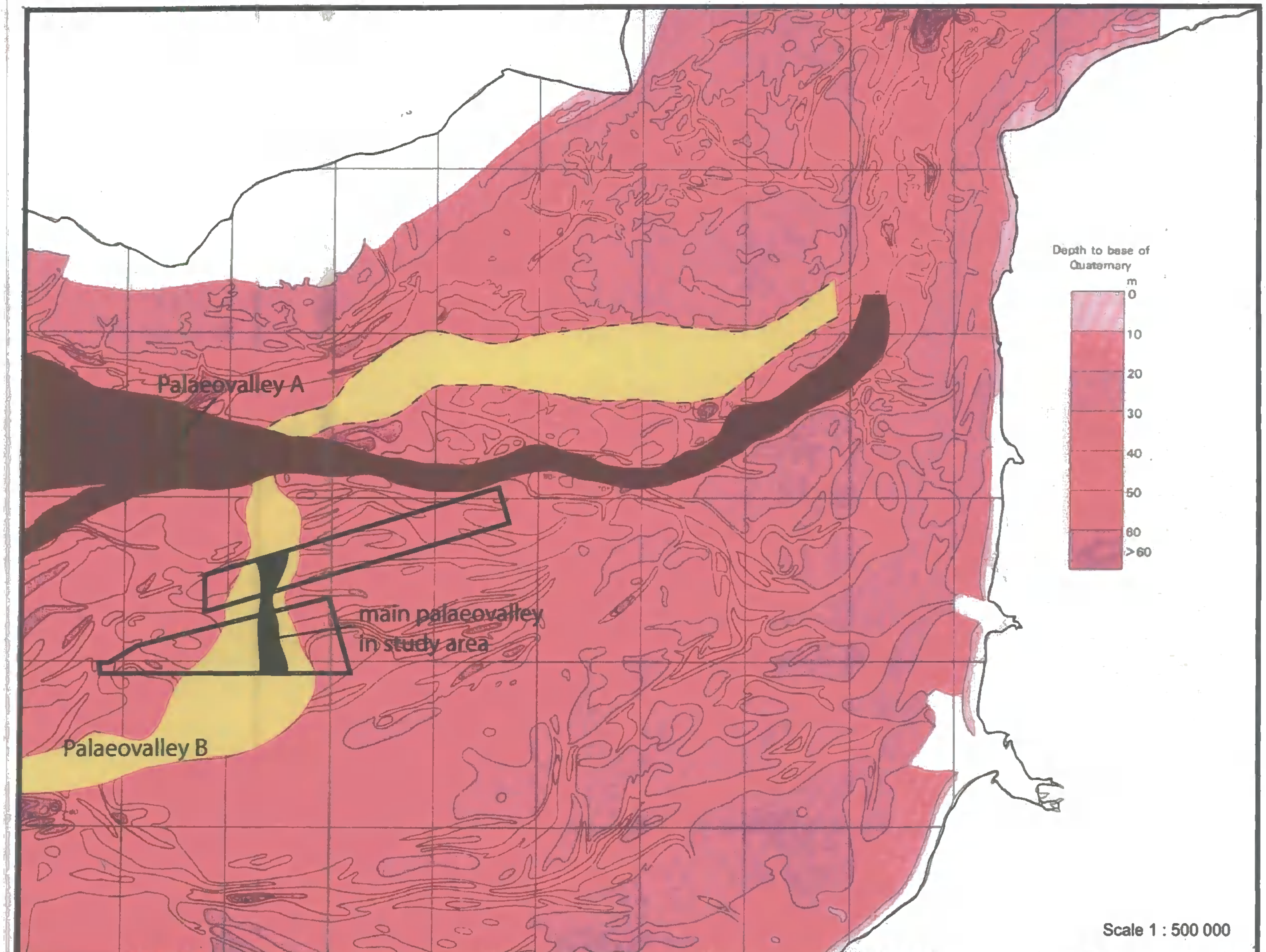


Figure 7-2: Locations of Palaoevalleys A and B and the main palaeovalley running through the study area.

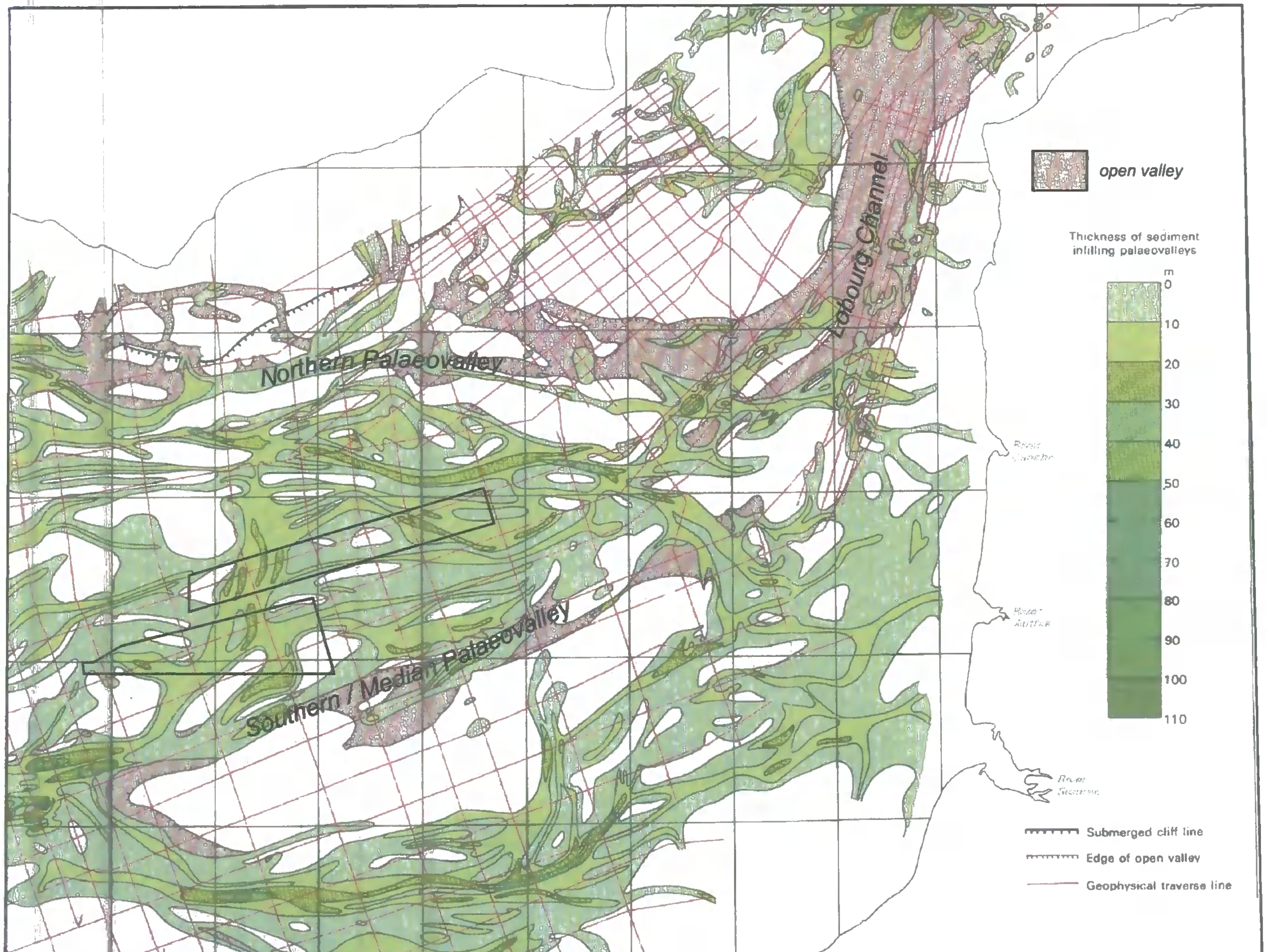


Figure 7-3: Palaeovalleys and infilling sediments of the eastern English Channel. Showing the open and infilled (closed) nature of the features (modified from Hamblin, 1989).

Infilled channel	Approximate channel width (km)	Average thickness of infill (m)	Approximate width : depth ratio
PhD Study Area	3-6	18-20	1:225
Palaeovalley A	3-5	15-25	1:200
Palaeovalley B	3-8	20-30	1:200
Owers Palaeovalley	1-3	16-20	1:100

Table 7.1: Width:depth ratios of palaeovalleys in the eastern English Channel.

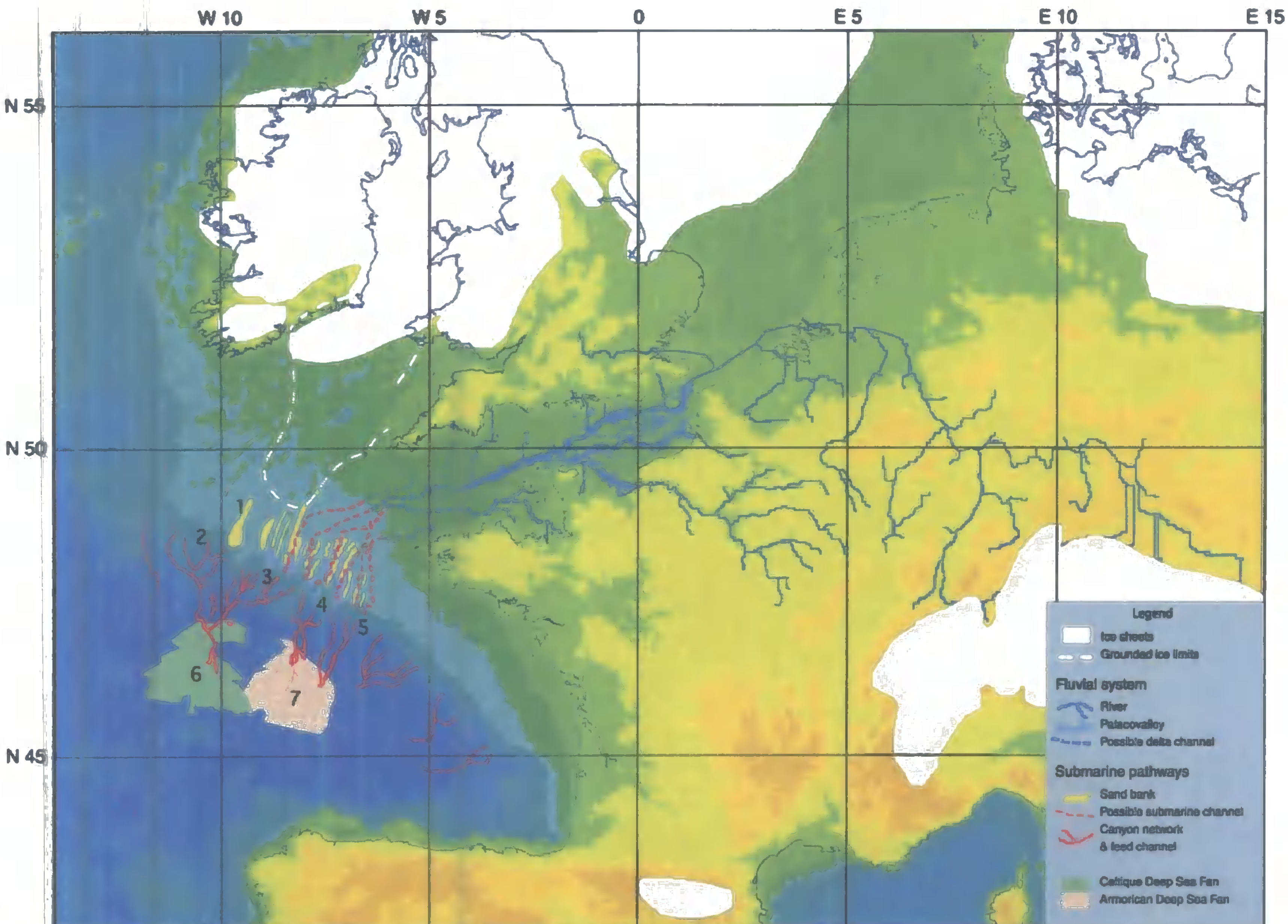


Figure 7-5: The English Channel and inferred submarine pathways at the LGM:1, sand banks of the outer shelf; 2, 'Grande Sole' slope drainage basin; 3, 'Petite Sole' slope drainage basin; 4, 'La Chapelle' slope drainage basin; 5, Ouest Bretagne' slope drainage basin; 6, Celtic Deep Sea Fan; 7, Armorican Deep Sea Fan (modified from Bourrillet *et al.*, 2003).

			Armorican Fan	Celtic Fan	Trevelyan Escarp.	Meriadzek
Local stratigraphy	**Isotopic stage**	**Age (B.P.)**	**72 104 Ma KS03**	**MK S03**	**AK S01**	**MD95-2002**
Holocene	1		12	7	3	10
Würm4-(Younger Dryas)	1	10 000				
Würm 3/4-(Bölling-Alleröd)	1	11 000	65-70	60	4-5	60-70
Würm 3	2	12 000 12 700			4-5	70
Würm 3	2	14 000			55	400
Würm 2/3	2	15 000 16.5 18.5 20 000			50	100
Würm 2/3	2				20-30	40
Würm 2	3	24 000 28 000				40

Sedimentation rates (cm/ka)

Table 7.2: Late Quaternary and Holocene sedimentation rates at the shelf edge and in the deep sea fans of the Western Approaches (from Bourillet et al., 2003).

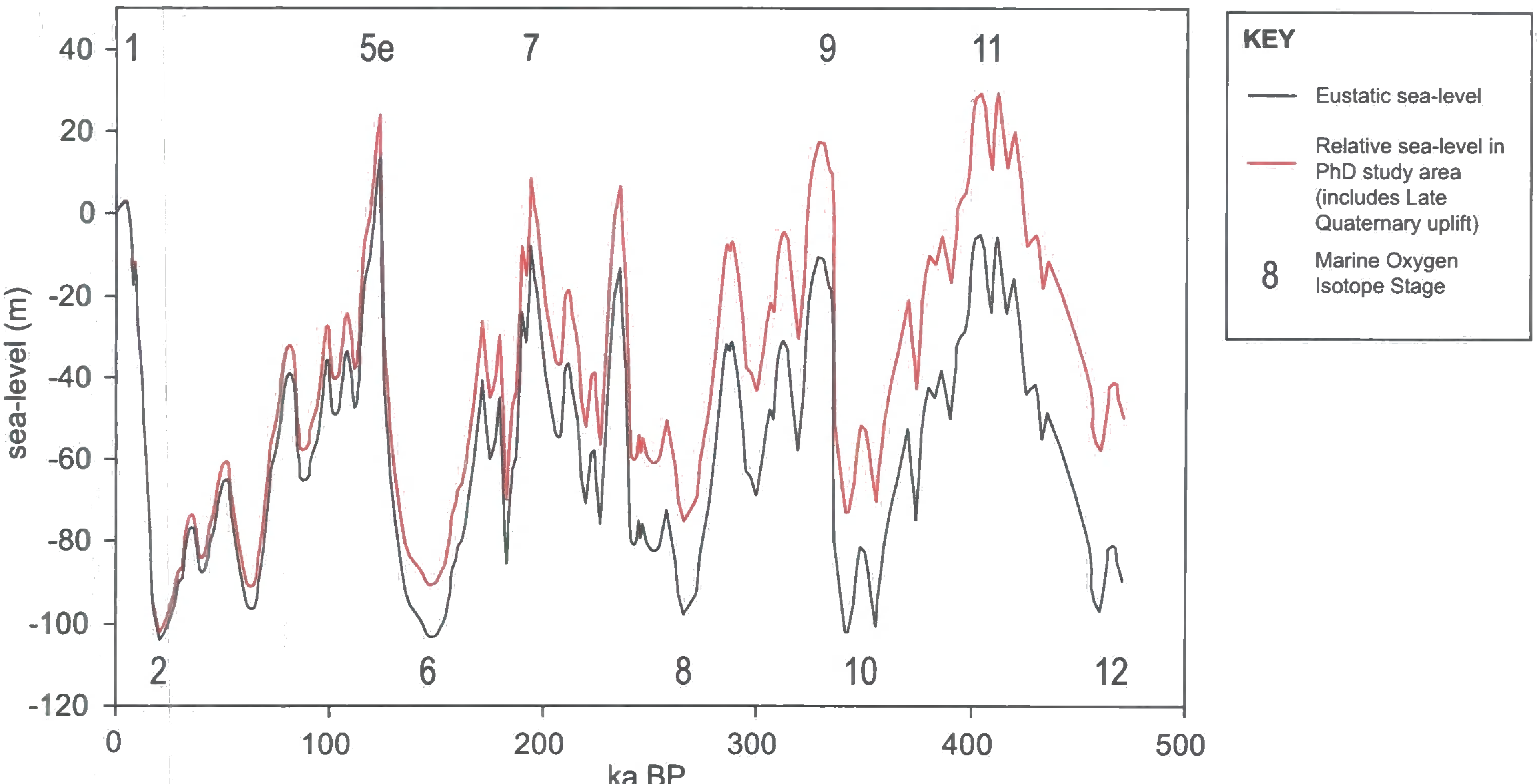


Figure 7-6: Late Quaternary eustatic and relative sea-level curves. Eustatic sea-level is derived from an amalgamation of Siddal *et al.*'s (2003) model reconstructions for Red Sea cores KL11 and MD921017. Relative sea-level curve includes the inferred uplift of the PhD study area (10 m per 100 ka)

Marine Oxygen Isotope Stage	Age (ka BP)	Eustatic Sea-level (m)	Sea level relative to present (m)	Fluvial incision of shelf
2	18	106	104	✓
6	150	106	-92	✓
8	270	98	-75	?
10	350	104	-72	?
12	470	95	-60	✓ Breach of Strait of Dover

Table 7.3: Eustatic and relative sea levels in relation to the eastern English Channel during the major lowstand periods of the Late Quaternary.

APPENDIX

Grain-Size Analysis

Sand and Gravel Fraction

Wet sieving of the sand and gravel sediment fractions were carried out by Andrews Survey (2001) in accordance with British Standard Institute, Methods of Tests for Soils for Civil Engineering Purposes; Part 2; Classification Tests (BS1377: Part 2: 1990).

<2 mm Fraction

The <2mm grain-size fraction was measured using standard techniques (Department of Geography, University of Durham). Samples were first sieved 2 mm. 0.5 g of sediment was transferred into a plastic 50 ml centrifuge tube and weighed on a top pan balance to four decimal places. A further 20 ml deionised H_2O_2 was then added and the samples placed in a hot water bath for 2-3 hours at +85°C to remove all organic material. Samples were then centrifuged at 4000 rpm for four minutes and half the supernatant decanted off. Tubes were topped up with deionised H_2O and transferred back to the centrifuge for four minutes. 20 ml of deionised H_2O and 2 ml of Sodium Hexametaphosphate ($(NaPO_3)_6$) solution was then added to reduce sediment flocculation. Samples were then analysed on a Coulter Laser Granulometer LS230 particle size analyser equipped with a fluid module and PIDS (Polarisation Intensity Differential Scatter) attachment. This machine uses a 5 mW, 750 nm laser beam and 126 detectors placed at a range of angles up 35° to the laser beam, measuring particle sizes from 0.4 μm-2 mm. Samples were mixed thoroughly before analysis and loaded into the counter vessel until obscuration values of 45-55 % and PIDS values of 8-12 % were obtained. Background readings were undertaken for each run. Samples were run continuously with offsets automatically measured every hour and detectors aligned every two hours; sample run time = 90 seconds; grain size parameters were modelled using the Fraunhofer model with sand/silt/ clay divisional boundaries defined by Wentworth (1922). The sub-2mm grain-size data was exported directly from the Coulter LS2000 software as xls-files and then the relevant grain-size channels extracted using the program GRADISTAT (Blott and Pye, 2001).

Diatom Analysis

Diatoms were prepared from core and surface samples using standard techniques. Samples were first sieved (0.5 mm) to remove coarse material. Between 0.3-0.4 g of dry sediment were then transferred to 50 ml plastic centrifuge tubes. Tubes were then arranged sequentially in test tube racks. The tubes were put in rows of eight, followed by an empty row, to avoid contamination. 30 ml of 30% H_2O_2 was added to each sample and then observed. If the samples did not react vigorously, they were transferred to a heated water bath for 2-3 + hours at (+ 85°C). After about one hour, samples were lightly shaken and returned to the water bath. Samples were removed from the heat and 1-2 drops of 50% HCl were added to dissolve any carbonates. The digested samples were then filled with distilled (d) H_2O and centrifuged for 4 minutes at 1200 rpm. The supernatant was then decanted off and the sediment pellet resuspended. This cleaning step was repeated four times. For slide preparation, appropriate suspension concentration was achieved by noting the turbidity and experience from test slides (Palmer and Abbot, 1986). 0.5 ml of suspension was then placed on a cover slip and left to settle overnight. The high refraction mountant Naphrax® was used for mounting the cover slip on the slide. Diatoms identification was carried out using Nikon Alphashot 2 microscope, fitted with a Zeiss 100/1.25 oil immersion objective.

Multi-element Geochemistry

0.5 gram of sediment was first sieved through a 150 μm nylon mesh (nylon mesh was used in place of a standard brass sieve to avoid contamination). 0.25 g of the sieved sediment was then treated to a microwave assisted acid digest using 9 mL of concentrated nitric acid and 3mL hydrofluoric acid for 15 minutes. After cooling, the digested sample was left to settle, diluted to volume and analysed on Perkin-Elmer Elan DRC+ intercoupled plasma mass spectrometer (ICP-MS) at the Department of Geography, University of Durham.