Aspects of Pleistocene Glaciomarine sequences in the North Sea.

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Ph.D.

University of Edinburgh

1986

VOLUME 1

TEXT

ABSTRACT

This study is concerned with a thick sequence of glacial, glaciomarine and marine sediments which accumulated in the North Sea basin during the Pleistocene epoch. The data for this project was provided from an extensive collection of shallow seismic profiles and core material held by the British Geological Survey. The investigation concentrates especially on glaciomarine sediments and processes of deposition.

Seismic analysis of the succession allowed for the establishment of a seismic or para stratigraphy and the identification of eight seismic sequences. Most significantly, the succession can be divided into two distinctly different packages. The lower unit was deposited predominantly in a subsiding shelf marine environment and comprises the thicker part of the succession. The upper unit consists predominantly of glaciomarine and arctic marine sediments deposited in a shallow basin and contains a complex series of erosion surfaces and reflector configurations. The erosion surfaces are attributed predominantly to fluvial erosion during low sea level stands.

Dating of the seismic stratigraphy, based primarily on the criteria established by Stoker et al. (1985), suggests that the lower unit is predominantly of Lower Pleistocene age whilst the upper unit spans the Middle to Upper Pleistocene. The extensive erosion surface separating the two units is attributed to the Elsterian stage and at least three major cycles of glaciomarine sedimentation have occurred subsequent to its formaton, the last of these being ascribed to the late Weichselian.

Lithofacies analysis of the late Weichselian sequence suggests that the Scottish ice sheet at this time was relatively limited and it only encroached onto the margins of the North Sea Basin. The sediments associated with the late Weichselian glaciomarine environment were deposited from meltwater flows and melting icebergs originating from a tidewater ice front. The development, distribution and preservation of these sediments appears to have been strongly influenced by the basin configuration, although generally there is a west-east continuum of proximal to distal glaciomarine sediments reflecting a transition

from bottom current and iceberg rafting processes to overflow plumes and rare iceberg rafted debris; there is also evidence for a seasonal cover of sea ice.

Similar lithofacies are present throughout the Middle and early Upper Pleistocene succession suggesting that the predominant depositional environment was one of a shallow glaciomarine basin adjacent to a tidewater ice front. Underlying this, the Lower Pleistocene succession consists predominantly of temperate marine and deltaic sediments.

Finally, the thick sediment pile and stratigraphic architecture of the Pleistocene sequence could only have been achieved by significant basinal subsidence therefore allowing for the accumulation and preservation of several glacial advance-retreat related units, widespread erosion surfaces, and local interglacial sequences.

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ACKNOWLEDGEMENTS

The completion of this study was aided by the help of a number of people and in particular my thanks go to the following:

My supervisors, Dr Nigel Fannin and Dr Rodger Scrutton for their help with both this thesis and with the various activities in which I have participated that have made this such a rewarding project.

To all the workers in the marine units at BGS which has proved to be such a very friendly and stimulating environment in which to work. In particular I would like to thank Dr Martyn Stoker and David Long for their help and encouragement throughout the duration of my stay at BGS, and also Graham Tulloch and Eileen Gillespie for patiently carting metres of core to and from Newbattle on numerous occasions. My thanks also to Elaine Bates for carefully typing the text over such a short duration.

My parents, Reta and Alan, who have always helped and encouraged me in all my work.

Last but not least, I would like to thank Elaine for her constant patience and tolerance and especially for her help with some of the diagrams and various other bits and pieces.

DEDICATION

I would like to dedicate this work to my parents, Reta and Alan.

CHAPTER ONE

INTRODUCTION

1.1 Objectives and Location of the Project

During the past 15 years the British Geological Survey (BGS) has collected a large amount of seismic information and core material as part of their routine mapping programme of the U.K. Continental Shelf. This large data set has provided the opportunity to study an extensive sequence of Pleistocene sediments, deposited in a broad sedimentary basin in the central North Sea between 56 and 59 degrees North (Fig. 1.1).

Initial investigations by BGS staff have identified glacial, glaciomarine and marine sediments associated with a complex sequence of transgressive and regressive phases. The age of these sediments range from early Pleistocene to late Weichselian and a detailed seismic stratigraphy has now been established by Stoker et al. (1985) for the area between 56°N and 58°N. It is the aim of this project to study in greater detail these Pleistocene sediments and in particular the glaciomarine sequences with the following specific objectives:-

i. To construct a detailed seismic stratigraphy which will provide the framework for the underlying objectives. This is especially critical for the Bosies Bank and Fladen areas (Fig. 1.1) for which even a basic framework does not yet exist.

ii. To establish a definitive sequence of lithofacies using borehole and vibrocore material and to map facies geometry, lateral variations, morphological features and seismic textures using a network of seismic lines. From this it should be possible to identify specific facies and relate them to environments of deposition.

iii. To interpret processes and environments of glaciomarine deposition and to construct a model embracing these features.

iv. To compare this model or models with both contemporary analogues, and ancient sedimentary sequences believed to be of glaciomarine derivation.

v. To provide an overall understanding of the physiographic conditions in the North Sea throughout the Pleistocene especially with regard to ice limits, the location of the polar front and palaeo-sea levels.

1.2 Definitions and Nomenclature

A variety of complex and often confusing definitions of the term glaciomarine environment exists in the literature. Boulton and Deynoux (1981) define glaciomarine "as a sea area in which the water structure is dominated by glacial meltwater", whilst Powell (1984) stated that "the glaciomarine environment includes all sediment deposited in the sea after being released from grounded or floating, glacial or sea ice." Both descriptions fail to acknowledge the combination of glacial and marine processes acting to produce a continuum of glacial, glaciomarine and marine sediments. As such, a more suitable definition is to describe the glaciomarine environment as one, influenced to varying degrees by a grounded or floating ice front, in which both glacial detritus and marine sediments may be deposited penecontemporaneously by a variety of processes.

Classifying and establishing criteria for distinguishing the products of glaciomarine sedimentation has proved even more difficult and controversial than defining the term itself. Unfortunately a variety of classification systems and terms now exist to describe glacial and glaciomarine sediments each with its own group of subscribers. However two important areas of concensus have emerged. First, that the term diamict (ite) can be used to describe any "poorly sorted gravel-sand-mud admixture" (Frakes, 1978) and as such it can be applied to a variety of glacial and glaciomarine sediments regardless of genesis. Secondly, that the name till should be reserved solely for those diamictons deposited directly at the base of a glacier without subsequent reworking (Eyles et al., 1985). For the purpose of this project the term diamict will therefore be used where appropriate and a lithofacies code (Fig 1.2) has been adapted from the code of Eyles (1983) as an aid to summarising down core sediment properties. Various terms and adaptions to this code will be discussed more fully in chapter 4.

1.3 Glaciomarine Models

1.3.1 Distinguishing Criteria

The most recent appraisal of glaciomarine environments and sediments is to be found in Eyles et al (1985). In this the authors correctly emphasise the primary importance of vertical and lateral lithofacies relationships and sediment body geometry as an aid to distinguishing glaciomarine sediments and processes but they fail to recognise the importance of criteria from other sources and state that "in general, the use of criteria such as texture, fabric, compositional immaturity of sand fraction, clast shape, clast frequency, microfauna and geochemistry to establish diamict genesis is circumscribed given that these data predominantly reflect source characteristics rather than mode of sediment deposition." In fact, such criteria are invaluable where core recovery is limited, the sedimentary sequence of a monotonous nature or the outcrop of a poor quality. For example Miller (1953) in a study of glaciomarine sediments from Middleton Island, Gulf of Alaska, states that "the principal and most characteristic lithological type in the bedrock sequence is the massive sandy muds containing scattered angular to rounded fragments of gravel size." Similar sequences from the North Sea have been described as glaciomarine primarily as a result of micropalaeontological analysis, sedimentology and the presence of dropstones.

As such, many of the criteria mentioned by Eyles et al (1985) have been used with some success in specific cases, especially with regard to palaeontological, geotechnical and fabric studies. Further work on palaeomagnetic properties and micro-fabric structures may also prove useful.

1.3.2 Theoretical Models

A number of recent studies on glaciomarine sedimentation have concentrated on predicting a variety of environments, processes and lithofacies from known geological phenomena (Andrews and Matsch, 1983; Powell, 1984 and Eyles et al., 1985). The advantages of such models are that they provide a focus for comparison whilst stimulating new ideas and research, the danger is that they may induce complacency and a tendency to manipulate data to fit the model.

Most workers have now discarded the old two-fold division of Carey and Ahmed (1961) into Arctic and Antarctic glacial marine processes and have realised the complexity of glaciomarine processes and the associated lithofacies. However, there is still confusion regarding the primary factors controlling glaciomarine sedimentation. Powell (1984) in a comprehensive study of glaciomarine lithofacies identified three major variables:- the type of ice front, ice-base conditions and finally the type of ice source. In a slightly different approach Eyles et al. (1985) identified a variety of diagnostic lithofacies and lithofacies associations and suggested that the main controls on these related to the conditions of glacial sediment input and the depositional environment. Although useful such models often belie certain factors when applied to the geological record especially, for example, the nature of the palaeoice base or the type of ice sheet mass. Also they fail to include certain features more relevant to the North Sea, especially the important role of water depth, eustatic and isostatic variations, strong seasonal variations, access to open water and the position of the Polar front. In a more simplistic model Andrews and Matsch (1983) included water depth, proximity to the ice front and seasonal process fluctuations as the primary factors affecting lithofacies types and distributions. The model also allowed for the presence of sea-ice, an important factor in the North Sea.

1.3.3 Geological Models

Over the past twenty years there has been a rapid development from the qualitative models of Carey & Ahmed (1961) to quantitative models based on geological and oceanographic observations. This has come about primarily as a result of the wealth of information released by hydrocarbon exploration on today's continental shelves and the present variety of glaciomarine geological models described in the literature suggests that no single model could embrace all the data. As such, the majority of glaciomarine geological models can be grouped into four environments:- areas adjacent to and beneath large ice shelves; areas adjacent to large grounded ice sheets ending in a tidewater front; glaciated fjords and finally ocean basins which are receiving ice-rafted material.

Most of the models relating to ice-shelves are based on observations of the contemporary processes and sedimentary products associated with the large Antarctic ice-shelves (Anderson et al., 1980; Drewry & Cooper 1981; Orheim & Elverhoi, 1981). One of the main conclusions of this work is that the ice-rafting of material away from the shelf front is not a significant process and that much of the sediment is deposited at or very close to the grounding line. In contrast, glaciomarine sediments in fjord environments are often derived from fast flowing temperate outlet glaciers where the ice-rafting of debris is an important process (Powell, 1983; Elverhoi et al., 1983).

Because of the presently favourable climatic conditions the modelling of glaciomarine environments adjacent to tidewater ice-sheets has mostly been based on the ancient record rather than modern processes and sediments. In these cases the late Pleistocene record has provided the greatest source of information (Molnia and Carlson, 1978; Vorren et al. 1982) although there are a number of examples based on much older sequences such as Lindsey's (1971) study of the Pre-Cambrian Gowganda sediments. A major stumbling block in many such models has been the inability to provide definite criteria for distinguishing between the products of ice-shelf sedimentation and a tidewater front environment. For example, Eyles et al. (1985) point to the controversy concerning glaciomarine sequences from the Puget lowlands of Washington and the difficulty of distinguishing between sedimentation below an ice-shelf and on an open marine shelf.

Models relating to deep sea environments have been primarily concerned with reconstructing palaeoclimate variations during the late Cainozoic. In these localities glaciomarine influence is almost entirely restricted to the ice-rafting of sediment and the affect that colder waters have on the biomass (Conolly & Ewing, 1965; Kirk & Clark, 1979). Recognising and explaining the products of such models tends to be straightforward due to their present setting and the limited variety of processes acting in such environments.

Clearly studies of both contemporary and ancient glaciomarine environments have concentrated on the continental margins, examples of intra-cratonic and enclosed sea locations such as the North Sea are much scarcer and apparently non-existent from modern settings. As such, any successful explanation regarding the Pleistocene glaciomarine sequence in the North Sea would have to embrace features from a variety of the models described previously, as well as including those features that are apparently unique to the study area.

1.4 Structural and Bedrock Setting

A knowledge of the pre-Pleistocene geology is important here primarily because the evolution and structural setting of the North Sea has greatly influenced the geometrical form of Pleistocene sequences. In addition to this, spatial variations in bedrock influenced the gross lithological and mineralogical characteristics of overlying Pleistocene sediments.

The structural components of the study area depicted in Fig. 1.3 can be categorised into two major provinces. First, along the western edge of the study area a stable pre-Permian platform, extending some 140km offshore, is overlain by an easterly-younging succession of Mesozoic sediments. Secondly, towards the east, a NNW to SSE trending Central graben with thick Mesozoic and Tertiary sediments. Tertiary sediments are absent over the western platform.

A number of workers have commented on the correlation between the thickness and distribution of the Pleistocene sediments and the underlying Mesozoic/Tertiary tectonic features (Caston, 1977; Holmes, 1977 and Eden et al., 1978). For example, in the Central graben combined Tertiary and Pleistocene deposits reach a thickness of over 3500m. However, sedimentation appears to have continued, unchanged, across the Pliocene - Pleistocene boundary and subdivision of the two is not straightforward (Caston, 1977 and Stoker et al., 1985). Although, along the western platform the Pleistocene is generally less than a 100m thick and oversteps the Tertiary onto harder pre-Tertiary strata where its base is easily identifiable. Also, despite the ambiguous position of the base of the Pleistocene further east, it is obvious that there is a marked increase in the thickness of the Pleistocene sequence along the western edge of the Central graben. Pleistocene sediments have therefore accumulated in areas of previously established maximum subsidence which have acted as depositional centres since Permian times (Glennie, 1984). This is perhaps not surprising given that subsidence rates were higher during the Pleistocene than at any other time during the Cainozoic (Clarke, 1973 and Zervos pers comm. 1986). The mechanism behind such rapid Pleistocene subsidence is disputed although various models have been put forward including thermal subsidence (Sc later and Christie, 1980), the lateral transfer of crustal material (Bott, 1971) and tectonic-isostatic effects (Eden et al., 1977). Clearly predictions of Pleistocene isostatic and eustatic variations are going to depend on which model is favoured.

1.5 Present Day Oceanography

1.5.1 Bathymetry

The complexity of the present day North Sea bathymetry (Fig. 1.4) can be attributed to a variety of Pleistocene processes that acted on the geological framework described previously.

To the west of 1°00'W, along the pre-Tertiary platform, the sea floor is characterised by numerous broad submarine banks and circular to elongated depressions and troughs with waters depths averaging 60m to 70m but reaching a maximum of 200m in the troughs. East of this area the water depth gradually increases into the Central graben although in the northern part the sea floor shelves more rapidly forming a large smooth bottomed basin; the Witch Ground Basin, with water depths of up to 150m. The northern margin of this basin is characterised by a series of north-south trending valley's and banks; the Fladen Deeps and Hills areas respectively.

To the south of the Fladen ground and the main 100m contour (Fig. 1.4) the water depths average between 80m and 90m forming a relatively flat plateau which stretches south to 56°00'N. Within this area a belt of valleys in the Devil's Hole locality (56°30'N, 0°45'E) form distinctive deeps reaching maximum depths of 250m.

1.5.2 Sedimentary Processes and Patterns

The North Sea is presently classified as a partially enclosed, tide dominated area where a progressive tidal wave is propagated from the Atlantic Ocean parallel to the shoreline (Reading, 1978). This produces an amphidromic process in which tidal waves move in a rotary, anticlockwise path (Fig. 1.5), although within the study area current directions are generally in a NNE-SSW direction. The resultant System has a relatively large tidal range with maximum values of 3-4m and current velocities ranging from 0.38m sec $^{-1}$ to 1.0m sec $^{-1}$ (Owens, 1981).

The conditions described above have probably existed since 5,000-6,000 years BP when sea level stood at 5m below present Ordnance datum (O.D.) (Owens, 1977), prior to which a marine transgression had reworked much of the surficial sediments leaving a lag of palimpsest deposits in the deeper water. However, at shallower depths the spatial distribution of the surficial sediments suggests that the dominant dispersal and bedform patterns (Fig. 1.6) are a product of present day processes. Namely, tidal currents and, to a lesser extent, meteorological currents (Owens, 1981). A brief description of these present day sedimentary processes and patterns will help provide an insight into the effects that Pleistocene glacial periods had on the central North Sea. This is especially important with regards to water depths which are generally agreed to have been lower during much of the Pleistocene due to the growth of large ice sheets. Late Weichselian sea levels, for example, are thought to have been some 100m below present O.D. (Jansen, 1979), and as such the North Sea would have been effectively cut off from the North Atlantic circulation system.

Fig. 1.7 depicts the present day transport paths in the North Sea, some of which are up to 400 km long, and within these paths the following depositional zones can generally be recognised:-

i. Sand ribbons are up to 15 km long and 200m wide and are deposited typically in water depths between 20m and 100m with maximum near surface current velocities in excess of 100 cm/sec (Kenyon, 1970).

ii. Sand waves have been extensively studied in the North Sea (Stride, 1963; McCave, 1971) where they form straight crested transverse bedforms with wavelengths in excess of 30m. Such bedforms occur where spring tide current velocities are in excess of 60 cm/sec. and wave activity is restricted.

iii. Sand patches form extensive areas of the North Sea and are partly relict and partly a reflection of contemporary bedload rippling, deposition from suspension and reworking by benthic organisms.

iv. Mud zone locations appear to have been affected by both tidal and wind-driven currents although generally such zones do appear to coincide with the ends of tidal current transport paths which are frequently associated with a low frequency of wave action.

Obviously tidal currents are the predominant sedimentary process in the North Sea, often completely obscuring the effects of wave driven processes and precluding their recognition. As such storm deposits appear to be presently restricted to certain areas including the German Bight (Reineck and Singh, 1972) and the south-western North Sea (Caston, 1976). In both areas typical storm couplets have been recognised at water depths of between 14 and 24m and extending up to 50 km offshore. Similarly, Jago (1981) describes the presence of a wave dominated and a tide dominated zone off the Yorkshire coast. The wave dominated zone is characterised by a seaward fining sequence passing laterally into a seaward coarsening tide dominated unit at a depth of about 30m.

It is important to realise that although such wave induced sediments are presently restricted the lower sea levels during a glacial period may well have precluded the importance of tidal currents resulting in a more isolated environment where wind and wave induced currents were predominant.

Approximate sediment sources, concentrations and destinations in the North Sea are depicted in Figs 1.7 and 1.8 (Eisma, 1981). It seems likely that these patterns would also have been radically changed during any glacial period and this is exemplified by areas of relict surficial muddy sediments in the study area, especially the Devils Hole and Fladen Ground. The absence of contemporary mud zones in the study area can be accounted for by the low suspended matter concentrations in the Scottish rivers and in the central North Sea (Eisma, 1981; Fig. 1.8).

1.6 **Previous Work**

This section will outline what are considered the most relevant studies relating to the Pleistocene sequences in the North Sea. Those relating principally to seismostratigraphic methods and palaeontological work will be discussed in more detail in chapters 2 and 3 respectively. Speculative work based only on extrapolations from land studies around Scotland, Orkney and Shetland (Mykura, 1976; and Flinn, 1967, 1978) will also not be considered here. The remaining background literature can essentially be attributed to two areas of research. First, studies based primarily on detailed seismic mapping corroborated by borehole material and secondly those relying more on core material and detailed palaeontological and lithological descriptions using only a limited seismic framework. The former are essentially restricted to BGS data and until recently many of the results were available only as internal reports. More detailed assessments of smaller areas but with limited seismic coverage have been published principally by Norwegian and Dutch workers. A third body of researchers have produced critical re-appraisals of work pertaining to the UK offshore Pleistocene successions including the North Sea (Sissons, 1981; Sutherland, 1984).

With regards to previous work on BGS data from the North Sea a number of internal publications were produced in 1977 and 1978 (Caston, 1977; Thomson and Eden, 1977; Holmes, 1977; Eden et al. 1978). Their prime objectives was to describe the Pleistocene succession, identify internal variations and to produce a seismostratigraphic framework which could be correlated with known land events. Work by Holmes (1977) and Thomson and Eden (1977) first established a Pleistocene seismic stratigraphy for the North Sea. Within this the authors identified a thick argillaceous marine unit, the Aberdeen Ground Beds, overlain by a complex series of cut and fill sequences, glacial and glacialmarine sediments and a variety of glacio-structural forms. However, on the basis of a series of radiocarbon dates (Harkness and Wilson, 1979) the whole of the Pleistocene sequence was identified as being of Devensian (Weichselian) in age. This assumption was maintained despite a thickness of Pleistocene sediments of up to 500m in the Central graben and the fact that at least three erosional phases can be identified. The ensuing hypothesis therefore envisaged very high sedimentation rates, up to 360mm per 100 years, concomitant with a complex sequence of ice advances and retreats within one stadial. Re-examination of these dates has cast doubt on their validity and it is likely that the older dates at least, were subjected to contamination by Tertiary and older material (Jansen 1979; Jansen et al., 1979; Sissons, 1981).

In studies of the Pleistocene sequence by Holmes (1977) and Thompson and Eden (1977) their greatest enigma was the presence of three successions of incised channels, which show no regular downslope gradient, the majority of which are unfilled, although some remain open to the present day (Fig 1.4). Flinn (1967) attributed the origin of the unfilled channels to a regional alignment of tunnel valleys formed during a stand of stagnant ice, whilst other workers have attributed them to direct ice erosion (Valentin, 1955) or even tidal scour. In assessing all the information pertinent to the formation of both the filled and unfilled channels Holmes (1973) concluded, quite reasonably, that they were generally of complex and varied origins including subglacial, glaciofluvial and glaciomarine processes.

However, none of the above workers paid any great attention to the nature of the glaciomarine processes active during the Pleistocene era. Although reference is made to ice rafted material, the presence of sea ice, assumed glacier ice-push structures, and the Wee Bankie moraine (Fig. 1.9) no attempt was made to model the depositional environment. Interpretations of palaeowater depths as little as 20m, low salinities and arctic marine conditions were based on micropalaeontological data (Gregory and Harland, 1978).

Eden et al (1978) in a broad study of Pleistocene depositional environments in the North Sea focussed more on the stratigraphy and possible age correlations, again based on radio carbon dates, together with a possible model of isostatic events during the Pleistocene. Although the timing of the model proposed by Eden et al. (1978, Fig. 1.10) is wrong, the concept behind it may help to explain the high Pleistocene subsidence rates discussed earlier.

New BGS data collected since 1978 allowed a number of workers to present a radically revised interpretation of the Pleistocene events and chronostratigraphy in the North Sea. Most significantly Stoker et al (1983, 1985) and Skinner and Gregory (1983) were able to prove that Pleistocene deposits in the North Sea range from early Lower Pleistocene to late Upper Pleistocene. This time span of over 1 million years compared with previous estimates of a maximum range of 130,000 years (Eden et al 1978) was based primarily on palaeomagnetic data, especially the identification of the Bruhnes-Matuyama boundary supported by detailed foraminiferal and dinoflagellate evidence. More recent work by Bergen university on BGS borehole (BH) 81/26 supports the presence of early Lower Pleistocene sediments in the North Sea (Sejrup et al., in press, Appendix 9).

A ramification of the new chronostratigraphy was that there was no longer a need to envisage confluent Scottish and Norwegian ice sheets covering the whole North Sea during the late Weichselian, therefore contradicting certain earlier interpretations (Holmes, 1977; Boulton et al., 1977 and Eden et al., 1978). This contention is supported by the absence of readily identifiable subglacial sediments east of the Wee Bankie Moraine (Fig. 1.9) and the probable lack of sufficient precipitation during the late Weichselian to support a Scottish ice sheet over the North Sea (McIntyre et al., 1976; Sissons, 1981). Overconsolidated sediments in the central North Sea, originally attributed to ice loading, may equally be the product of dessication or segregated ground ice (Boulton and Paul, 1976; Derbyshire et al. 1985). Similarly channels east of the Wee Bankie moraine have been re-interpreted as the product of fluvial processes (Long and Stoker, 1986) rather than subglacial meltwaters.

A more process orientated study of the Pleistocene sediments was undertaken by Cameron et al. (1986) in which they described a sequence of identifiable facies within the present stratigraphy. Inherent to this model was the recognition of three major episodes of glaciation separated by interglacial marine sediments at certain critical localities. However, both this and earlier BGS work concentrated on the North Sea area south of 58° north. It was therefore the aim of this project to concentrate firstly on data from the area between 58° and 59° north.

Interestingly certain aspects of earlier interpretations by Dutch reseachers (Jansen, 1976; Jansen et al., 1979; Oele and Schuttenhelm, 1979) regarding the North Sea Pleistocene succession, including the area north of 58°, compare well with the present framework derived from BGS data.

In their study of the Pleistocene sequence in the Fladen Ground area Jansen et al. (1976, 1979) identified both glaciomarine and overconsolidated subglacial sediments together with a possible end moraine, 'The Hills', (Fig. 1.11). These sediments were interpreted as the product of a partly buoyed off ice shelf bordering on an inland sea or proglacial lake with a series of tunnel valleys marking the maximum extent of the late Weichselian ice sheet. The implications of this model (Fig. 1.12) are therefore that the Scottish and Scandanavian ice sheets were not confluent in the central and southern North Sea and as such much of the area would have been sub aerially exposed. This interpretation is similar to that presented by Cameron et al. (Fig. 1.13, 1986), apart from some disagreement over the extent of ice cover in the northern North Sea. The idea that overconsolidation in many of the sediments is due to subaerial processes therefore appears a distinct possibility. The occurrence of Scandanavian erratics and anomalous ice flow directions (Mykura, 1976) both on land and in the UK sector of the North Sea could then be attributed to earlier, more extensive glaciations, when Scottish and Norwegian ice sheets were confluent over a much larger area. For example Oele & Schuttenhelm (1979) describe a Saalian glaciation much more extensive than the Weichselian one.

The above section reviews the advances in Pleistocene research in the North Sea over the past ten years. It also demonstrates the concentration of these studies on chronostratigraphy and event geology rather than any attempt to model the respective sedimentary environments. In addition to establishing a chronostratigraphic framework is is therefore essential to examine facies variations both within and across particular chronostratigraphic units.

1.7 Thesis Layout

The main characteristics of the various seismic units are described in chapter 2 including discussion of the possible causes of both bounding and internal reflectors together with any other features observed on the records. In chapter 3 a description of the principal biostratigraphic units and their palaeoenvironmental significance is followed by a summary of the stratigraphy, supported principally by palaeomagnetic data.

Chapters 4 and 5 concentrate on the identification, description and interpretation of the principal Late Weichselian and pre-late Weichselian sedimentary facies respectively. Their associations and implications are then assessed in chapter 6 in the form of a glaciomarine model, followed by a comparison of the model with contemporary and ancient glaciomarine sequences.

A comprehensive summary of the main conclusions of this work is given in chapter 7.

1.8 Data Base and Methods

1.8.1 Work by the British Geological Survey

Offshore studies by BGS discussed in detail by McQuillin and Ardus (1977), provide the major source of data for this project. To date, in excess of 200,000 km of seismic track have been run, complemented with the collection of cores and samples from over 25,000 stations and over 500 boreholes to a maximum depth of 235m below the seabed. A large proportion of this material is from the North Sea and the equipment and retrieval logistics are discussed briefly below.

i. Acoustic Survey:- this involves the use of continuous seismic reflection profiling techniques typically running airgun, sparker, pinger, deep tow boomer and side scan sonar operated simultaneously from a master control system. Fig. 1.14 summarises the resolution, penetration and operational frequencies of the various types of equipment. Seismic interpretation techniques are outlined in chapter 2.

ii. Drilling:- Subsequent to the seismic survey borehole points are chosen at points providing the optimum stratigraphic information. Both dynamically positioned and anchored drill ships have then been used to drill cored holes retrieving core material 77mm, 64mm or 51mm in diameter.

iii. Sampling:- In addition to the borehole sites a large number of stations are chosen for sampling using a variety of equipment depending on the sea bed lithology and physiographic conditions. At most stations samples of the sea bed sediments are obtained by using a shipek grab operated from a small winch. Recovery of sediment up to 6m below the sea bed is achieved using either gravity core or vibrocore equipment. Generally speaking the gravity corer is most effective in very soft muds and totally ineffective in clean sands although it does have a greater operational depth than the vibrocorer.

1.8.2 Field Cruises

During my three year research period I have spent a total of 3 months working aboard BGS seismic, drilling and sampling ships. Most of these cruises involved work in my particular study area.

1.8.3 Laboratory Methods

Each core collected by the sampling ship was split longitudinally, photographed, and then described visually prior to being subjected to routine engineering tests whilst on board ship. On certain cruises more detailed and thorough engineering measurements were undertaken. The cores were then kept in a cooled store and a large number were x-rayed (Appendix 5) and sub-sampled for more detailed analysis using a variety of methods briefly outlined below:-

i. Grain size analysis was performed by standard dry sieve and pipette methods (Galehouse, 1971).

ii. Clay minerals from the < 2 um fraction were analysed by x-ray diffraction using orientated aggregates on a glass slide and run on a Phillips' diffractometer (Appendix 6).

iii. The bulk geochemistry of the sediments was determined using X-ray fluorescence spectrometry. The samples were first ignited at 1100°C to remove any volatile material and then fused glass discs (45mm diameter) were made for major element analyses (Norrish & Hutton, 1969). The analyses were corrected for mass absorption effects (Theisen & Vollach, 1967) and interference effects.

iv. Micro-fabric studies of the sediments were made by mounting orientated samples some 10mm in diameter on aluminium stubs. These were then coated with gold using a splutter device and studied using a Cambridge scanning electron microscope. However, the results from this work proved inconclusive and are not included in the thesis.

v. Samples were also taken for micropalaeontological studies especially on the benthic foraminifera and dinoflagellate cyst assemblages. The method of foraminiferal analysis was similar to that described by Skinner and Gregory (1983) whilst samples for dinoflagellate analysis were processed using the sintered glass funnel procedure of Neves and Dale (1963).

CHAPTER TWO

SEISMIC ANALYSIS

2.1 Introduction

The following chapter deals firstly with the theory behind seismic analysis and the recognition and regional geometry of Pleistocene seismic sequences within the study area. This is followed by an attempt to describe and explain various seismic reflection parameters, their relationships to known environments of deposition and to the physical properties of the sediment recovered from boreholes.

In chapter three the correlations between seismic units, microfaunal and floral zones and palaeomagnetic boundaries are considered, followed by a brief discussion of the chronostratigraphic framework.

2.2 The Principles of Seismic Stratigraphy

Seismostratigraphic analysis is essentially based on the interpretation of reflectors on seismograms. These reflectors represent the energy received over a given time interval from acoustic interfaces within the geological column. Each seismogram is a graphical record of the variation in amplitude of the energy pulse, acoustic pressure, as recorded against time (vertical axis) for different points along the earths surface (horizontal axis). The occurrence of acoustic interfaces within the geological column is dependent on the presence of contrasting acoustic impedances either side of the interface. Such contrasts are essentially the product of differences in the elastic properties and density of the materials concerned.

Similarly, the velocity of the energy pulse is controlled by the physical properties of the rocks and sediments namely porosity, grain size and rigidity (Faas, 1969; Biart, 1985). All these properties are related and despite arguments to the contrary, there is not usually one specific property that can be identified as the prime factor in controlling the velocity of the energy pulse.

Seismic Sequence	Semismic Facies Configurations	External Form	Basal Reflector Terminations	Nature of Basal Reflector	Reflection Amplitude	Reflection Continuity	Special Features	Environmental Interpretation
8	Sub-parallel, Transparent texture	Sheet	Concordant	Slightly irregular	Low	Discontinuous	Local	Shallow marine and shoaling sand processes.
	Prograding or onlapping	Channel Fill	Downlap/ onlap		Moderate	Continuous-discontinuous		Moderate-high energy, shallow-marginal marine.
	Divergent		Concordant	Irregular-highly irregular	Moderate	Continuous	Scoured channel base	Low energy, deposition from suspension.
	Sub-parallel; transparent texture	Basin fill	Concordant	Slightly irregular to even	Moderate to low	Continuous-discontinuous	Pock-marks	Low energy, primarily deposition from suspension
	Closely spaced, sub-parallel divergent at edges.			Micro irregular	High	Continuous	Gas disturbance	Low energy, deposition from suspension.
7	Chaotic or structureless	Hummocky sheet	Discordant	Sub-horizontal-irregular	Low	Discontinuous	Upstanding ridges	Variable energy, shallow marine, subglacial or ice push in parts?
	Transparent or structureless	Sheet drape	--	Irregular	Low	Discontinuous	Scoured top	Variable energy, shallow marine, high sediment concentration.
6	Structureless or chaotic, low-angle bi-directional, sub-parallel.	Wedge	Downlap concordant	Planar, locally irregular	Low low-moderate low	Discontinuous	Hummocky bedforms	Variable energy, marginal marine and shoaling sand processes.
5	Chaotic or structureless	Hummocky Sheet	--	Irregular	Low	Discontinuous		Subglacial deposition
4	Chaotic or structureless Prograding Divergent	Channel fill	-- Downlap/ onlap concordant	Highly irregular	Low moderate moderate	Discontinuous continuous continuous		Variable energy, shallow water cut and fill complexes, some slumping moderate-high energy, shallow-marginal marine or sub-aerial low energy, deposition from suspension.
	Structureless or chaotic	Sheet drape	--	Irregular	Variable	Discontinuous		Variable energy, shallow marine, high sediment concentrations.
3	Shallow, prograding reflections		Downlap		Moderate	Continuous		High energy, prograding clinoforms
	Sub-parallel-chaotic			Sub-horizontal Concordant	Moderate	Discontinuous		Variable energy, shallow marine-marginal marine.
	Parallel-sub-parallel				Moderate high	Continuous		Uniform rates of deposition on stable/gently subsiding shelf.
2	Prograding	Channel fill	Downlap/ onlap	Highly irregular	Moderate high	Continuous		Moderate-high energy, sub-aerial-marginal marine.
	Onlapping		Onlap		Moderate	Continuous		Moderate-high energy, marginal marine.
	Divergent		Concordant		Moderate	Continuous	Gas blanking	Low energy, deposition from suspension variable energy shallow marine some slumping?
	Chaotic or structureless				Low	Discontinuous.		
	Sub-horizontal or structureless	Sheet drape	Concordant	Irregular	Low	Discontinuous		Variable energy, marginal marine processes.
1	Sub-parallel-chaotic	Sheet Drape/ basin fill	--	Slightly irregular	Variable	Discontinuous		Variable energy, shallow marine-marginal marine.
	Parallel-sub parallel		Concordant locally onlapping	Irregular even	High	Continuous		Uniform rates of deposition on gently subsiding shelf.

Table. 2.1. Seismic sequences facies and interpretations.

Porosity, for example, is important in unlithified sediments as it controls the influence of pore-water on the fast wave, however a decrease in the mean grain size of the sediment will also affect velocity rates by reducing them (Horn, 1968).

The effects of the sediments physical properties on the acoustic pulse will be further considered with regards to down borehole properties at the end of this chapter.

Interpretation of the seismogram is based on the assumption that there is a one to one correspondence between a seismic arrival on the record and a geological horizon. Whilst this is broadly true there is still some dispute as to what seismic reflectors and terminations represent in a geological sense and how they can best be utilised.

If one accepts the methods of seismic stratigraphy developed by Vail et al. (1977a and b) seismic reflectors tend to parallel stratification surfaces rather than lithological units, hence the reflection alignment corresponds to a time stratigraphic horizon and not necessarily a lithological boundary. Taking this model one step further the seismic section can be divided into identifiable sequences of concordant reflectors bounded at their top and base by unconformities or correlative conformities (Mitchum et al., 1977a). Such seismic sequences are genetically related and, as they were deposited during a given interval of geological time, chronostratigraphically significant.

Although critics of the above work (Watts, 1982; Biart; 1985) have cast doubt on the validity of such models for calculating eustatic changes in sea level from coastal onlap configurations, the principles of seismic stratigraphy are generally accepted and have been employed here in a procedure outlined by Mitchum et al. (1977a) and briefly described below.

The first stage involves the identification of primary seismic reflectors and terminations and their grouping into seismic sequences, defined earlier. On the actual section respective sequence boundaries are best recognised by the disconcordance or termination of reflectors, and therefore strata, and the type of discordance is the best indicator of whether an unconformity results from erosion or nondeposition. Fig. 2.1 depicts the various termination relationships defined by Mitchum et al. (1977a). Truncation indicates an erosional hiatus, whilst onlap, downlap and toplap indicate nondepositional hiatuses.

Seismic facies analysis, the second stage of interpretation, involves the description and interpretation of various reflector parameters including configuration, continuity, amplitude, frequency and internal velocity. Not all these are applicable to the interpretation of the relatively thin and unlithified cover of Pleistocene sediments in the North Sea and in most cases identification of a particular reflection configuration is the most useful parameter. Occasionally the amplitude and continuity of the reflectors can also be used to determine the environment of deposition (Sangree & Widmier, 1977).

Following the identification of particular seismic facies some understanding of their three dimensional form is useful for subsequent analysis. This is particularly true with regards to Pleistocene sequences where the internal reflector configuration is commonly structureless or chaotic and the external shape of the sequence forms an important part of the genetic interpretation. The seven principal forms, some of which can be divided into subtypes, are shown in Fig. 2.2.

The final stage of interpretation involves combining seismic sequence and seismic facies analysis in an attempt to provide a basic stratigraphic framework and to interpret environments of deposition, palaeobathymetry and palaeogeography. With regards to this particular project the availability of borehole control, and hence chronostratigraphic and lithological information, allows for more precise and definitive stratigraphic and genetic interpretation discussed in subsequent chapers.

The following analysis is based almost totally on high resolution profiling data surveyed by a suite of equipment described in chapter 1. In this study a large number of surveys were available for interpretation, ranging in age from 1971 to 1985. However it was decided to utilise only the more recent, high quality records, and to especially concentrate on the 79/15 and 81/04 surveys from the Bosies Bank and Fladen areas respectively as these records had not yet been considered by BGS staff. Fig. 2.3 depicts the six sub-areas within the main study area and the main surveys relevant to these. It should be noted that only in the Peterhead area is there a dearth of recent surveys and as such the interpretation from this area is of a lower standard.

Resolution of the records from the chosen surveys is generally good (Fig. 1.14) and although much of the analysis is based on the identification of sequence boundaries, some areas of definitive reflection configurations have been identified and utilised. Estimates of the depths of the reflection terminations for the sequences described below are based on a two-way travel time of 1800 metres per second unless otherwise indicated.

2.3 Seismic sequences

In all eight sequences were identified bounded by eight unconformities, a schematic diagram depicting their relationships is shown in Fig. 2.4. Line interpretations from Sparker and Boomer records, along specific east-west and north-south lines, are shown in Figs 2.5 and 2.6 (back pocket). Figure 2.7 depicts, isometrically, interpretations of some of the lines from the Fladen and Bosies Bank areas not featured in Figs 2.5 and 2.6, (and Fig. 2.8 shows the exact location of the boreholes depicted in previous figures). The configuration terms used in this discussion and utilised for subsequent facies analysis are shown in Fig. 2.9. It should be stressed that in most cases the sequence boundaries were recognised not from stratal terminations but because the erosion surface itself produced a seismic reflection. The benefit of this was that the unconformity reflections were often identifiable over large areas suggesting that the reflection coefficient of the unconformity was often significantly greater than the reflection co-efficient of the underlying and overlying beds.

Further reference to Fig. 2.4 shows that the bounding surfaces generally take one of two forms. First, a planar sub-horizontal surface typical of boundaries A, C and F. Second, an irregular to highly irregular surface characteristic of boundaries B, D, E, G and H.

Over most of the study area reflector A marks the base of the Pleistocene. Along the western edge of the basin this boundary is delimited by an angular discordance; the Pleistocene reflectors clearly overstepping the underlying Tertiary, Mesozoic and Palaeozoic strata (Chesher, 1982; Stoker 1984; Skinner, in press). Line 6, project 79/15, (Fig. 2.5) clearly shows the discordant relationship, and in Fig. 2.10 the planar reflectors of sequence 1 can be seen overstepping easterly dipping Tertiary strata.

Towards the east, the basal discordance can be traced to its correlative conformity, where the Pleistocene reflectors lie conformable to the underlying Tertiary strata. The accuracy of this boundary has been verified by micropalaeontological analyses of borehole material discussed further in chapter 3. However, where the base of the sequence is beyond the penetration of BGS equipment (about 300m), tracing it has relied on the interpretation of deep seismic records and commercial borehole control such interpretations are less accurate and, in places, ambiguous.

Overlying seismic boundary A the Pleistocene succession contains eight seismic sequences of which sequence 1 is the most widespread and clearly defined (Figs. 2.5 and 2.6). Internally it is characterised by a series of strong, sub-horizontal reflectors often becoming more chaotic near the upper boundary (Figs. 2.10 and 2.11).

The nature of the upper boundary to seismic sequence 1 varies depending on which sequence overlies it (Fig 2.4). For example, Fig 2.12 shows the depth, below present sea level, of the upper surface of sequence 1. The contours to this boundary have been smoothed out but the effects of the highly irregular reflectors, B and D, can clearly be seen relative to the overall more regular nature of bounding reflectors C and F.

Seismic sequence 2 is restricted to the thicker parts of the Pleistocene succession where its irregular base cuts Sequence 1 forming a sharp truncation surface, boundary B (Figs 2.10, 2.11, 2.13 and 2.14). Only on line 54, 72/04 and line 16, 79/15 (Fig. 2.5) can this sequence be observed to cut pre-Pleistocene strata.

Internally sequence 2 displays a wide variety of reflector types with the result that its delineation relies primarily on identifying the bounding surfaces and its distinctive channel like geometry.

The upper boundary of seismic sequence 2 is most commonly represented by the relatively even basal surface of sequence 3, reflector C, which has a continuity of 10's of km's (Figs 2.9 and 2.10). The two sequences are therefore discordant, the irregular surface of reflector B, clearly being truncated by reflector C.

Apart from its distinctly planar basal surface seismic sequence 3 is best characterised by the presence of strong, sub-horizontal reflectors. Where the base of this sequence directly truncates sequence 1 it commonly forms a strong reflecting, non-angular, unconformity due to the constructive interference of sub-horizontal refectors either side of the unconformity.

On a regional scale seismic sequence 3 can be identified over much of the area, except in the Fladen Ground and Marr Bank areas where it has been eroded out by overlying sequences to form a steeply dipping truncation surface (Fig. 2.5., lines 10 and 15, 79/15, and Fig. 2.6, line 4, 81/04). Where the sequence is not eroded out it can be seen gradually thinning onto sequence 1 (Fig 2.5, line 15, 79/15).

In the Forties and Devil's Hole areas the upper boundary of sequence 3 commonly outcrops at the sea bed where it acts as an effective acoustic basement on boomer and pinger records. Elsewhere the upper boundary is most commonly marked by the irregular to highly irregular basal surface of seismic sequence 4, reflector D.

Intraformational reflector configurations in seismic sequence 4 vary from strongly sub-parallel to chaotic or structureless (Fig 2.11) and as such it is very similar to sequence 2. It occurs over the whole area where its external form varies from a continuous blanket like deposit in the central Fladen area (Fig. 2.11) to a more irregular based sequence around the edges of the Witch Ground Basin (Fig. 2.14). However, in the Forties and Devil's Hole areas the sequence is discontinuous, consisting of a series of essentially isolated channel like features ranging from 20m to 180m deep (Fig. 2.15).

Where the upper surface of sequence 4 does not cropout at the seabed it is marked by the irregular reflectors G or H, except in the south-west where it is bounded by the strongly planar reflector, F (Fig 2.5, line 36, 80/03).

Seismic sequence 5 appears to be relatively local, occurring only along the western edge of the Peterhead and Marr Bank areas, although its real extent is possibly obscured by the resolution of the profiling equipment. The sequence itself is characterised by hummocky upper and lower surfaces, the former often occurring at the sea bed, and an opaque or chaotic reflection pattern.

In most places sequence 5 directly overlies pre-Pleistocene strata (Fig. 2.5, L30, 80/03) where the unconformity between the two is strongly discordant and clearly visible. In contrast, where it does overly sequence 1 the unconformity is weaker and discontinuous, the generally chaotic nature of both sets of intraformational reflectors making it difficult to identify any reflection termination.

To the east and north-east seismic sequence 5 passes laterally into sequence 6. A strong sub-horizontal basal surface, reflector F, delineates the base of this sequence, whilst internally it is characterised by discontinuous planar and dipping reflectors. The sequence cuts a variety of sequences, including pre-Pleistocene strata, and the disconformable relationship of the internal reflectors against the underlying sequences further highlights the presence of the unconformity, (Figs. 2.15 & 2.16).

The upper boundary of sequence 6 generally outcrops at seabed except in areas where it is cut by reflector H, as seen along lines 15 and 36, 80/03, Fig. 25.

Spatially, sequence 6 is restricted to the south-west where its base dips gently to the north-east. Towards the coast the sequence oversteps the underlying units eventually lying directly on pre-Pleistocene strata. The eastern and northern limit of the sequence is more difficult to trace as in places it appears to run laterally into sequence 7 whilst elsewhere it ends as an abrupt ridge or is cut out by reflector H (Fig. 2.6., L28, 80/03).

Seismic sequence 7 is restricted to the northern area, the sequence is bounded by an even or slightly irregular base, reflector G, whilst the upper surface either outcrops at the seabed or is delineated by reflector H. Internally the reflector pattern is most commonly chaotic or irregular and discontinuous (Fig. 2.17), with frequent bright spots often disrupting the reflector pattern.

In the Fladen area the upper boundary, reflector H, equates with that described by Stoker and Long (1984) being very irregular above a depth of 160m (below O.D), especially where the surface outcrops at sea bed (Fig. 29/31). Below 160m reflector H is more regular and the whole sequence forms a blanket like infill in the central basin. The basal boundary in this area, reflector G, is often discontinuous suggesting a low reflection coefficient between this sequence and the underlying sequence 4, and in places, the boundary between the two sequences is best delineated by differences in seismic texture not configuration (Fig. 2.11).

The uppermost seismic sequence, sequence 8, can essentially be subdivided into two correlative units. The first is restricted to the Fladen and Bosies bank areas where it forms an acoustically well layered and continuous sequence (Fig. 2.5, Line 14, 81/04) whose base, reflector H, was described in the preceding section. Towards the top of the sequence the layering becomes weaker and the upper boundary, the sea bed, is pitted by distinctive pockmark features (Fig. 2.10). The second subdivision occurs over the remaining area and is characterised by an irregular base, often forming linear erosive features varying from 20 to 150m deep (Fig. 1, Line 30, 80/03 and Fig. 2.10), connected by thinner interchannel areas, and more regular blanket type deposits.

Internally, the seismic layering is again well defined, becoming weaker towards the top, and where the sequence infills channel features the layering forms a strongly discordant relationship with the adjacent sequence (Figs. 2.15, 2.18 and 2.19).

2.4 Seismic Facies Interpretation

Within the seismic sequences a number of diagnostic reflection configurations can be recognised. Such configurations are now widely used, usually within a known framework, to establish conditions and environments of deposition and to make some estimates of possible relevant lithologies. In this case it is now widely accepted in the literature that within the North Sea Pleistocene succession we are dealing primarily with sequences of glacial, glaciomarine and shelf marine facies concomitant with extensive erosion surfaces and evidence of periods of emergence. As such, this background knowledge was utilised when interpreting the various seismic facies therefore helping to analyse the more ambiguous configurations. A brief summary of the identified seismic facies and their interpretation is given in table 2.1.

2.4.1 **Sequence 1**

Reflector patterns within this sequence define two seismic facies (Fig. 2.19). The first is characterised by high amplitude, parallel and sub-parallel reflectors which display good continuity and low amplitude variations. In the central basin these reflectors dip to the east at angles between 0.2-0.5°, mirroring irregularities in the basal surface and suggesting a concordant draping over the surface. Only along the north-west edge of the sequence, where strong reflectors onlap onto the pre-Pleistocene strata, is there evidence of non-concordant reflectors within this facies, (Fig. 2.5, Lines 6 & 16, 79/15). The second type is typified by a chaotic configuration displaying variable amplitude and poor continuity. This facies generally occurs towards the top of the sequence, except along the edges of the basin where it appears to overstep the underlying parallel reflectors to become the predominant pattern.

Geometrically sequence 1 appears to form a wedge shaped unit which reaches a thickness of over 200m in the central basin and thins rapidly towards the coast in the west, (Figs. 2.5 & 2.7), where it terminates in one of two manners. In the south it is abruptly cut out and overstepped by the overlying sequence, whilst in the north it appears to thin imperceptibly below the resolution of the profiling equipment. Despite the wedge shaped appearance of the sequence within the study area it is probable that it eventually thins to the north and east and is better described as a sheet drape unit.

Environmentally the lower seismic facies suggests uniform rates of deposition in a stable, uniformly subsiding area. In the central basin the concordance of this facies with the underlying Tertiary strata may indicate uninterrupted deposition through the Tertiary-Pleistocene boundary. However, further west a Pleistocene transgression surface separates this facies from the underlying pre-Pleistocene strata.

In contrast the overlying chaotic facies is consistent with deposition in a variable energy environment and shallower water depths (Mitchum, 1976). This interpretation is consistent with the predominance of this facies in the shallower parts of the sequence.

2.4.2 **Sequence 2**

A variety of seismic facies can be identified within this sequence reflected by a number of different configurations and external forms. In the Devils Hole a single, continuous facies is characterised by a wavey and sub-parallel, more rarely chaotic, configuration concordant with an irregular base. The reflectors are generally discontinuous and of a low amplitude, often broken by areas of acoustic masking attributed to gas pockets. Externally, the geometrical form of this facies is one of an irregular based sheet drape unit.

To the north-west at least three seismic facies can be identified infilling the highly irregular surface cut into sequence 1. The first, and most common type, displays chaotic or structureless configurations, (Fig. 2.13), and occurs predominantly in the Fladen and Forties area. The second is characterised by moderate amplitude, divergent reflectors, forming a symmetrical infill in the channel feature. Elsewhere the channels contain an onlap infill but with similar moderate amplitude reflectors. The final facies type displays inclined reflectors, usually dipping towards the central Fladen basin, as shown on line 6, 79/15, and line 24, 81/04, Fig. 2.5.

Complex fills, consisting of two or more seismic facies, take a variety of forms but most commonly a lower facies of onlapping or divergent reflectors is unconformably overlain by, low amplitude, chaotic reflectors (Fig. 2.20). More complex infills occur in some of the Fladen channels where the basal facies consists of low amplitude divergent reflectors overlain by a chaotic facies, and finally high amplitude onlapping reflectors separating areas of chaotic infill, (Fig. 2.5, line 18, 81/04).

Inherent to the interpretation of seismic facies within this sequence is a genetic explanation of the irregular basal boundary, especially in the North where it defines a series of channel like features. Originally this surface was attributed to subglacial meltwater erosion (Holmes, 1977). However it is equally possible that it relates to fluvial erosion during a low sea level stand, although it may have been locally modified by subglacial and proglacial processes. Such low sea level cycles are common throughout the Pleistocene as a result of glacio-eustatic variations, and the magnitude and timing of these events has been well documented by a variety of workers, (Evans, 1979; Jelgersma, 1979 and Dinter, 1985). Furthermore similar infilled channels have been recorded from a number of continental shelves and attributed to fluvial erosion during the Pleistocene era. Hine and Snyder (1985) describe a series of such features incised into shelf sediments off North Carolina and infilled with facies ranging from simple to complex in appearance. When sampled these facies proved to consist of estuarine and shelf fossiliferous muds, the latter being attributed to a mid-Pleistocene transgression. Similarly, Johnson et al. (1982) have interpreted infilled channels (complex and simple layering), in the Great Barrier Relief Shelf, as the product of fluvial erosion subsequently backfilled by fluvial, deltaic and marine sediments.

Recent work by Long and Stoker (1986a) favoured the predominantly fluvial nature of both the irregular base to this sequence and some of the overlying irregular sequence boundaries. Figs. 2.5 & 2.6 clearly shows that the base of these channel features generally fit a concave base level, depressed in the central basin.

Accepting that the base of this sequence was, at least in part, fluvially eroded, then both the chaotic infill type configurations, and the discontinuous and low amplitude reflectors, can be related to a variable energy, probably fluvial cut and fill complexes or a nearshore environment (Sangree & Widmier, 1977). Penecontemporaneous slump structures are also associated with chaotic configurations (Mitchum et al., 1977b). The genesis of the structureless facies is uncertain although it possibly reflects homogenous sediments deposited primarily from suspension. Gas blanking may also contribute to the structureless configuration at certain locations.

Symmetrical onlap configurations are characteristic of moderate or low energy channel and basin infills, deposited premominantly from bottom processes. Similarly divergent patterns are indicative of a low energy environment but with a greater component of sedimentation from suspension and possible lateral variations in the rates of deposition. In a series of flume experiments McKee (1957) found that symmetrical onlap infills were typical of subaerial streams whilst bottom concordant and divergent facies formed in submerged channels under low energy conditions.

Asymmetric layered facies within sequence 2 are open to a variety of interpretaions but are most likely the product of either progradational deposition or alternatively lateral accretion normal to the principal current direction. In further flume experiments McKee (1957) observed that the channel infill became asymmetrical if a current passed diagonally over the channel, and it is likely that in the majority of cases asymmetric infills in sequence 2 reflect progradational deposition in shallow water conditions. An exception to this is shown in Figs. 2.21 and 2.22, where a complexly layered facies is clearly defined with internal reflectors dipping in a westerly direction for up to 3 km. The dip of these reflectors varies from 3° to 13° and their configuration appears to be concordant and associated with the sides of a partially infilled, north-south trending channel. As such, the facies at this location has been interrupted as the product of westward lateral accretion in a large meandering channel. The remnants of this feature are recorded by a present day open channel.

2.4.3 **Sequence 3**

This sequence basically contains two seismic facies with a few local variations. The lower facies directly overlies a strong basal unconformity and is characterised by parallel and sub-parallel reflectors. In the Bosies Bank area these reflectors are laterally continuous and display moderate to high amplitudes (Fig. 2.14) which become slightly divergent as the facies thins to the west. Further south this facies is typified more by moderate amplitude, sub-parallel reflectors, which are discontinuous in places.

Ovelying the layered configuration, the second seismic facies is characterised by moderate to low amplitude, discontinuous, sub-parallel and irregular reflectors (Figs. 2.23). This facies, although present in the North, appears to be predominant in the Forties and Devils Hole area, where it often forms the majority of sequence 3 and occurs right up to the sea bed. Also isolated topographic highs with internally chaotic or structureless patterns often appear to be associated with this facies (Fig. 2.24).

Two local seismic facies were also identified within this sequence. The first, seen in the Devils Hole area, consists of low amplitude complex reflection configurations, which become almost transparent in places, bounded by strong sub horizontal reflectors. (Fig. 2.25). The second occurs in the Fladen area where there are two units, each about 30m thick, of shallow progradational configurations dipping towards the central basin (Fig. 2.26). In both cases the reflectors are moderate amplitude and occur over an area of about three to four kilometres.

The geometrical form of sequence 3 is difficult to establish because it generally ends abruptly against overlying irregular truncation surfaces and is absent from most of the Fladen basin. A contour map of the base of sequence 3 (Fig. 2.27) displays this abrupt truncation whilst also highlighting two possible depositional centres, each 200m below present day sea level, and a third smaller depression in the north-east corner of the area. The latter may be associated with westerly prograding clinoforms observed at this locality.

Subsequent to the formation of the planar base of sequence 3 by a marine transgression the two succeeding seismic facies reflect deposition in environments similar to those described for sequence 1. This would suggest that the transgression was relatively rapid and a stable shelf environment with uniform deposition rates was quickly established. A further drop in sea level resulted in the deposition of the upper facies in a shallower and generally less stable marginal marine environment. The predominance of this upper facies in the south reflects shallower water depths and more variable in conditions, and this is corroborated by the overall north-easterly dipping nature of sequence 3, (Fig. 2.27).

The genesis of locally complex configurations in the Devils Hole and Forties area is uncertain, Holmes (1977) suggested that such configurations could be the result of glacial tectonism.

However, it is equally possible that the configurations represent high energy sand prone deposits.

The progradational configurations are interpreted as prograding clinoforms suggesting a relatively high sediment supply and little or no subsidence concomitant with a higher energy regime. Jansen (1976) identified this facies as a deltaic unit, and foreset configurations suggest outwash directions were from both the south-west and north-east towards the central Fladen basin.

2.4.4 Sequence 4

In the Fladen and Bosies Bank areas this sequence, with rare exceptions, contains one continuous seismic facies. It is characterised by an uneven base which becomes more regular towards the central basin, and internally displays structureless or even transparent configurations, occasionally passing either laterally or vertically into areas of chaotic reflector patterns with irregular discontinuity surfaces, (Figs. 2.14 & 2.17). Reflectors, where present, are usually of a variable amplitude, irregular, and discontinuous. Where this facies outcrops at the sea bed it forms an irregular surface which becomes distinctly hummocky along the western edge of the area, (Fig. 2.5, L6, 79/15).

Further south a variety of seismic facies have been identified, broadly similar to those described for sequence 2, infilling a highly irregular basal surface and forming a series of channel like features. The most common infill pattern, seen throughout the Marr Bank area, consists of two distinctive facies. The lower one displays westerly inclined or symmetrical onlapping reflectors, the upper is characterised by divergent reflectors (Fig. 2.28), both facies display moderate amplitude reflectors with poor to moderate continuity.

In the Devils Hole and Forties area the infill consists, most commonly, of a single facies displaying structureless or chaotic configuration patterns, broken by high amplitude reflectors. More complex infills consist of chaotic, onlapping and prograding seismic facies in a variety of combinations, (Figs. 2.15 and 2.29).

The highly irregular nature to the base of sequence 4 is again primarily attributed to fluvial erosion during low sea level stands. The tendency for this basal surface to even out towards the central Fladen area suggests that marine conditions were maintained in this area and only the surrounding highs were subject to fluvial erosion.

Structureless configurations from the relatively continuous northern facies are attributed to generally homogenous sedimentary sequences, probably muds, deposited in an environment lacking bottom traction currents or downslope resedimentation or where large volumes of suspended sediment are available (McCave, 1971). Such an interpretation is consistent with the blanket like geometry of this facies, especially in the central basin, and it is important to note that such an environment would not necessitate a deep water setting but could occur in comparatively shallow water.

Similarly structureless and chaotic channel infill facies, further south, probably reflect the homogenous nature of the sediment, whilst irregular intra-facies reflectors represent re-activation surfaces and a further period of erosion.

Layered channel infill facies, described earlier, are open to the same interpretations as for sequence 2 and it is suffice to point out that such configurations are generally restricted to the south-west suggesting shallower water depths and relatively higher energy environments.

2.4.5 **Sequence 5**

This sequence can be treated as a single seismic facies whose main diagnostic features are the presence of point-source hyperbolic reflections within a seismically structureless configuration.

Geometrically sequence 5 forms a veneer, usually exposed at the sea bed, of between 5 and 25m which often displays a distinctly hummocky upper surface (Fig. 2.30). It is relatively local in extent and within the study area it occurs only along the western edge of the Marr Bank and Peterhead areas although it can be traced into the onshore area to the west (Stoker et al., 1985).

As such, sequence 5 has been widely interpreted as a subglacial lodgement till (Thomson and Eden, 1977; Holmes, 1977; and Stoker et al. 1985), and this would appear to be consistent with the internal features and landward extent of this sequence.

To the north this sequence is thought to equate to the western section of seismic sequence 7 which displays similar diagnostic features. However, the poor quality of the data from the Peterhead area generally prevents a direct correlation of the two sequences.

2.4.6 Sequence 6

Within this sequence three laterally equivalent facies can be identified. The first, and most extensive facies displays sub-parallel, low amplitude, discontinuous reflectors which are associated, in places, with hummocky bedforms. The second is characterised by low angle, bi-directional downlapping reflector configurations (Fig. 2.31) occurring over areas between 3 and 6 km^2. Structureless and chaotic configurations are diagnostic of the third facies which occurs mainly in the thinner areas of the sequence, along its south and western edge.

Figs. 2.5 and 2.6 show the wedge like geometry of sequence 6 and the distinctive planar base to the sequence (Fig. 2.16). Contours drawn to the base of this sequence (Fig. 2.32) indicate a maximum gradient, towards the north-east, of 1 in 1000 decreasing seaward to 1 in 2000. Fig. 2.32 also shows the edges of sequence 6 which appears to either terminate as a low-ridge (Fig. 2.6, Line 28, 80/03), or to pass laterally into sequence 5 (Fig. 2.32).

The extensive planar base to this sequence has been interpreted as the product of wave erosion (Thompson and Eden, 1977; and Stoker and Graham, 1985) succeeding deposition of the lower facies of sequence 4. The overlying seismic facies are all indicative of unstable, variable energy environments dominated by marginal marine processes. Bi-directional reflectors are interpreted as wave induced structures formed when the sea bed was at or slightly below the wave base.

2.4.7 Sequence 7

Accurate identification of individual facies within this sequence is precluded by the structureless and chaotic internal configurations. Seismic interpretations of this sequence has therefore relied on the nature of the bounding surfaces and the overall geometrical form.

On this basis two seismic facies can be recognised within the sequence. The first occurs in the Fladen and east Bosies Bank areas where it forms a sheet drape type unit bounded at the base by a low amplitude, discontinuous reflector. Towards the edges of the central basin this basal reflector passes laterally into the base of sequence 8, suggesting that in parts the two sequences are lateral equivalents. The upper surface of this facies displays a distinct micro-relief, of 1 to 6m, (Fig. 2.11) which appears to become more even towards the deeper parts of the central basin (Fig. 2.17).

The second seismic facies is restricted to the Bosies Bank area, especially along the western edge, where it occurs as a blanket of irregular thickness. Intraformational reflectors are low amplitude and highly chaotic, especially where the facies forms relatively upstanding features (Fig.2.5, Line 15 & 16, 79/15 and Fig. 2.33).

The generally structureless configurations displayed by both facies, similar to those seen in parts of sequence 4, are again attributed to homogenous sedimentary sequences deposited primarily from suspension. In the central Fladen basin, the discontinuous nature of the basal reflector where this sequence overlies sequence 4, suggests that in parts deposition between the two sequences may have been continuous. The irregular nature of the upper surface in the Witch Ground Basin was attributed by Stoker and Long (1984) to scour by sea ice similar to that described from the Beaufort sea (Reimnitz and Barnes, 1974). The age of this surface and its implications for palaeosea levels will be discussed further in chapter 3. Further west, in the Bosies Bank area, the upstanding features within sequence 7 are possibly the product of ice push processes which would also explain their highly chaotic configurations and restriction to the shallower parts of the study are.

2.4.8 Sequence 8

Individual facies within this sequence are best defined on high frequency boomer and pinger records. From these four seismic facies have been identified, the first two are restricted to the Witch Ground Basin and peripheral areas (Fig. 2.34), and the remainder occur at shallower water depths to the south and south-west.

In the Witch Ground Basin the lower seismic facies forms a basin fill unit, 5 to 20m thick, characterised by closely spaced, high amplitude, continuous reflectors. These define a divergent, occasionally onlapping, layered configuration which tends to be draped over basal irregularities (Fig. 2.35). Internally this configuration is commonly punctuated by bright spots, vertical disturbance zones and gas blanking (Fig. 2.36), whilst over highs the reflectors pass laterally into zones of discontinuous, low amplitude and irregular reflectors (Figs. 2.37 and 2.38). Further diagnostic features of the facies include a near transparent, discontinuous basal layer some 1 to 3m thick, and, in places, a concordant band of very closely spaced reflectors (Fig. 2.39).

The upper seismic facies in the Witch Ground Basin generally lies conformably on the lower facies although locally this relationship is unconformable. It tends to have a relatively transparent texture, within which concordantly layered reflectors, commonly discontinuous, are of a moderate to low amplitude and less closely spaced than in the underlying facies, (Figs. 2.37 and 2.38). Disturbance of the reflector configuration is less common, although the upper boundary is commonly punctuated by cone shaped notches. These are usually between 1 and 3m deep and 30m wide, but reaching a maximum size of 15m deep and 200m wide. Such features, termed pockmarks, have been identified in the area by a number of workers and generally attributed to gas escape processes (Van Weering et al., 1973; Caston, 1974; Eden, 1975; McQuillin and Fannin, 1979; Hovland, 1980 and 1982; Green et al., 1985). Similar depressions in the top boundary of the lower facies are therefore attributed to buried pockmarks. Their distribution and genesis along with other gas related features will be discussed in more detail in appendix 1.

The remaining two seismic facies occur in the second subdivision of sequence 8, described previously. An upper and lower facies can again be identified and they appear to be the lateral equivalents of the upper and lower facies in the Witch Ground Basin. The lower is characterised by continuous, low to moderate amplitude reflectors, forming a basally concordant, often divergent, configuration. It occurs along channel margins and as a channel infill feature, where the base is often highly irregular and overlain by a transparent layer, (Figs. 2.40, units 1 and 2, and Figs. 2.41 and 2.42).

In contrast the upper facies contains downlapping or onlapping configurations, often lying discordant to the underlying facies, and therefore delineating a marked unconformity between the two (Figs. 2.43 and 2.44). Where the reflectors form a prograding type infill they generally dip, at shallow angles, to the east and north. The pattern of the prograding reflectors is usually a shingled or sigmoidal configuration (Figs. 2.45, 2.46 and 2.47).

Some fill features contain only one seismic pattern, usually a divergent configuration, that often forms an area of positive, or mounded relief, at the sea bed (Figs. 2.19 and 2.47). Truncation of the reflectors, either at the sea bed or by a thin transparent layer 0.5 to 1.0m thick, is a common feature over many of the channel fill sequences.

Complex infills within sequence 8, consisting of three or more seismic facies, have been identified along the western edge of the area (Fig. 2.5, line 36, 80/03 and line 6, 79/15). Fig. 2.48 shows a typical example of a complex fill in which two lower chaotic facies (1 & 2) are overlain by prograding reflectors (3). These features are conspicuous by the fact that they are up to 200m deep (below sea level) compared with average base level depths of 80m along the western edge of the study area.

Where the second sub-division of sequence 8 does not occur as channel infill it forms a single seismic facies. This is characterised by a blanket like geometry and by low amplitude, sub-parallel, intra formational reflectors. Lateral terminations of this facies are often in the form of a small ridge as shown in Fig. 2.5, line 15, 80/03.

In the Witch Ground Basin the concordantly layered drape fill unit suggests that this facies was deposited predominantly from suspension in a relatively low energy environment or concomitant with high sediment loads. The divergent configuration of both the upper and lower facies indicates a steady rise in sea level or slow subsidence of the basin.

Disruption of the layering over topographic highs is attributed to a less stable environment with variable energies, indicating shallower water depths and wave reworking of the sediments. In the Fladen area such zones occur at or near the sea bed in water depths between 130-140m. This would suggest palaeo-wave bases at about 130m below the present day sea level indicating that the upper and lower facies were deposited in a shallow water basin. The cut off depth of sequence 8 in the Fladen area (Fig. 2.34) at about 140m further supports the interpretation that it was deposited in shallow water. In the Bosies Bank area, along the periphery of the basin, shallower cut off depths of about 120m indicate that the basin has been tilted or that sequence 8 is a diachronous unit.

Outside the Witch Ground Basin sequence 8 occurs primarily as a channel fill facies. The lower concordantly layered facies relates to suspension deposition, although the relatively transparent texture on boomer records may indicate the presence of coarser material relative to the lower facies in the Witch Ground Basin. Progradational reflectors from the upper facies suggest a primary sediment source from the west with a subordinate supply from the south. These clinoforms are interpreted as having built out into shallow water under moderate to high energy conditions. Truncation of the reflectors in the upper facies is attributed to the last transgression to occur over the area.

The majority of channel features infilled by sequence 8 appear to fit a regional base level of about 80m in the west gradually increasing to 150m in the east. A fluvial origin would therefore again be consistent with these features, assuming that the shape of the base level is the product of increasing subsidence to the east. The highly irregular base to many of these channels is thought to be the result of scour by river ice in a similar manner to that described by (Collinson, 1971).

Where the channels do not fit a regional base level alternative processes to fluvial erosion must be sought. Singular channels in the Marr bank and Peterhead areas cut deeply into the underlying rockhead, and are interpreted by Long and Stoker (1986a) as the products of subglacial erosion. Symmetrical, chaotic facies in the base of these channels are therefore thought to be subglacial tills. A similar, but much wider, feature can be seen in the south-west corner of the Bosies Bank area (Fig. 2.5, Line 6, 79/15) and is possibly of a similar subglacial origin. Other, overdeepened channels were possibly caused by catastrophic sea level lowering or meltwater discharge.

Away from the coast, in isolated channels which have remained open or become only partly infilled, occasional units of a highly chaotic nature with a hummocky upper surface and abrupt lateral terminations were interrupted as slump or debris flow deposits. Fig. 2.21 shows the occurrence of such facies in the axis of an open channel. These mass transport deposits are considered in greater detail in the next section. Disturbed units on the Scotion slope (Piper et al., 1985) although formed in much deeper water, are very similar to those recorded in Fig. 2.21 especially with regard to the abrupt lateral passage into undisturbed, acoustically layered facies. Piper et al. (1985) attributed this disturbed zone to a variety of processes including slumping and sediment creep.

The blanket like facies within the southern part of sequence 8 is thought to be the product of marginal marine process, and the ridge-like terminations possibly represent palaeo-beach scarps.

2.5 Mass Transported Deposits

Seismic evidence of these deposits is restricted to channel features, both open and infilled. A large number of the filled channels, at various stratigraphic levels, are partially or totally infilled with sediment which displays a chaotic reflector configuration (Figs. 2.13, 2.14, and 2.20). This type of reflector configuration has previously been interpreted here as reflecting sedimentation in shallow, variable energy marine environment. Equally, it is possible that on the channel margins such a configuration pattern is the product of mass transport mechanisms (Mitchum et al., 1977b).

More definitive geophysical evidence of mass transported deposits occurs in certain open channels and detailed seismic profiling over one of these facies revealed the presence of extensive mass transported deposits (Fig. 2.21) both on the channel margins and in the channel axis. Two types of mass transport or movement deposits were identified, as defined by Nardin et al (1979a). First, slide or slump deposits are characterised by a hummocky topography, back scarps and deformed or chaotic reflectors. The latter are not however a pre-requisite for identifying slumps or slides and in fact a slump deposit may still display continuous or slightly deformed internal reflectors (Nardin et al. 1979a). Other criteria, outlined above, are typical of slump or slide deposits as described by various workers (Nardin et al., 1979a, 1979b; Field and Clarke, 1979 and Piper et al., 1985).

The second type of Mass transport deposits identified are mass flow deposits characterised by a hummocky or mound shaped surface, discrete hyperbolae, and a general absence of internal reflectors. The latter is due to the high water content or deformational homogenization of the sediment mass. Boomer profile 53 (Fig. 2.22) was shot along the axis of the channel (Fig. 2.21) and clearly shows the presence of extensive mass flow deposits which probably originated on the channel flanks to the north-east and south west.

Both types of Mass transport deposits may have been instigated by any one of a variety of mechanisms. However, there is no evidence of any gas charged sediments in the immediate area and slump triggering by this mechanism is thought unlikely. More possible, is the action of freshwater flushing weakening the ionic bond of the clays during periods of sub-aerial exposure. Periglacial type processes have also been suggested as a possible mechanism for other such features in the study area (Long and Stoker, 1986b) whilst undercutting by erosion could also have induced slumping.

2.6 The Relation of Seismic Sequence Boundaries to Down Borehole Physical Properties

The positions of cored boreholes from the study area within the previously described seismic stratigraphy are shown in Fig. 2.49. The actual locations of these were depicted in Fig. 2.8.

On certain drilling programmes the core was subject to a variety of engineering tests at relatively close sampling intervals. The wealth of results described in various BGS internal reports (Lambert and Hallam, 1976; Hobbs and Long, 1978; Long and Hobbs; 1979) allow for a study into their relationship with bounding seismic reflectors.

i. BH 77/2:- Two strong breaks in the geotechnical profile (Fig. 2.50) at 19m and 40m both correspond with the calculated depths for strong reflectors, representing the base of sequence 7 and the base of sequence 4. Below the 19m break there is a strong increase in the shear strength and bulk density of the sediment, and corresponding decreases in the porosity and void ratio. However the particle size distribution, although highly variable, is generally similar for 6m either side of the break. On the seismic record this break appears to correlate with a slightly irregular, often diffuse, and probably non-erosional boundary (reflector G).

The break at 40m is most conspicuous by the sharp increase in shear strength, from less than 50KN/m^2 to greater than 200 KN/m^2, below the boundary. Acoustically this break appears to relate to basal sequence boundary of seismic sequence 4. The boundary itself is only slightly irregular as compared with the highly irregular nature of this boundary associated with channel features more common outside the Witch Ground Basin. Unfortunately the probable position of an acoustic break at about 65-70m between sequences 1 and 2 was not sampled.

ii. BH 77/3:- No sharp breaks in the geotechnical profile were detected although two minor breaks do occur at 37m and 90m (Fig. 2.51). Above 17m the profile is highly irregular and difficult to interpret.

A calculated depth to sequence boundary D of about 90m appears to relate to the lower break in the geotechnical profile which is characterised by a decrease in the shear strength and bulk density, and a corresponding increase in the porosity. On the seismic record this boundary appears as a highly irregular and erosive surface cutting into the underlying sequence.

The upper break in the profile, at 40m, corresponds to the calculated depth for an intraformational, irregular and erosive surface within sequence 4.

The sequence boundary between 8 and 4, at approximately 17m, appears to correlate best with a sharp reduction in the silt: clay ratio below 17m, indicating a lithological difference between the two channel fill sequences.

iii. BH 75/33:- The first break on the geotechnical profile (Fig. 2.52) from this borehole occurs at 16m where there is an increase in the shear strength, bulk density and percentage of sand. This appears to relate to the basal boundary of sequence 8(M) and as such is of a similar nature to the break recorded in 77/2.

At 25m, a moderate increase in the bulk density of the sediment appears to correspond with the slightly irregular basal boundary of sequence 7 (G). Laterally this surface is discontinuous and diffuse and not strongly erosive.

2.7 Discussion

The seismic stratigraphy clearly shows that the Pleistocene succession thickens eastwards into the central North Sea Basin reaching a maximum thickness of over 400m (Fig. 2.5). It is also apparent that the succession thins again towards the East Shetland platform and as such the Pleistocene isopachs correspond closely to the Mesozoic graben structures (Fig. 1.3). This association has been recognised by a number of workers (Caston, 1977; Bjorlykke, 1985; and Cameron et al., 1986) and, as mentioned in chapter 1, corresponds with rapid sedimentation and subsidence rates in the thickest parts of the sequence.

As much of the Pleistocene succession here is associated with glacial periods sedimentation would have been controlled by a combination of isostatic and eustatic influences described by Morner (1980), together with tectonic movement inherent to the North Sea Basin. However, within the discussed seismic framework there is no evidence of fault movement in the Pleistocene, although further south the base of the Pleistocene is sometimes displaced by up to 100m (Balson and Cameron, 1985) and Ringdal (1983) has documented evidence of modern seismic events to the north of the study area. As Jelgersma (1979) points out, objective evaluation of the relation between tectonic, eustatic and glacial isostatic effects is difficult to assess and at this point it is suffice to stress that during the Pleistocene the majority of the study area consisted of a subsiding basin environment interrupted by isostatic and eustatic effects.

Evidence of sea level variations, whatever the cause, are best documented by the presence of basin-wide irregular erosion surfaces. Boundary B is the most significant of these surfaces in that it separates two significantly different successions. The lower succession, sequence 1, reflects deposition in an essentially non-glacial and relatively stable marine shelf environment probably characterised by uniform rates of subsidence. Above boundary B, the remaining seismic sequences reflect deposition under a variety of energy regimes, and essentially shallow water environments with periods of emergence. Such a melange of sequences above B represents an increaed influence of glacial and glaciomarine processes concomitant with isostatic and eustatic effects.

Highly irregular bounding surfaces, such as B, have been attributed to fluvial erosion and, where they have been penetrated by boreholes, they appear to correlate with a slight reduction in shear strength. Possible explanations for this include reworking of the sediment at the unconformity surface or a weakening of the clay framework by electrolyte dilation during subaerial emergence (Buchan et al., 1972). Where sequence boundaries are represented by an increase in shear strengths, as in 75/33 and 77/2, they are often less irregular, and not necessarily the product of fluvial erosion. Such boundaries do however still reflect periods of emergence, both dessication and freeze thaw processes are known to consolidate the soil (Derbyshire et al., 1985) and as such to generate high reflection coefficients. From this it would appear that the bounding surfaces or primary seismic reflectors, in agreement with Vail et al. (1977a), are generated by chronostratigraphic stratal surfaces rather than lithostratigraphic units. Although such surfaces are not necessarily time-synchronous as they are often laterally variable in duration due to erosion or nondeposition.

With regard to the variability and extent of the bounding surfaces within the framework a number of interesting features are apparent. The fact that channel facies within sequence 2 become more predominant to the north, whilst further south the lower boundary (B) is less irregular and more continuous, suggests that the shallowest water depths occurred to the north around the Witch Ground Basin. This is verified by the presence of large channel accretion complexes in the Bosies Bank area (Figs. 2.21 and 2.22) and their association with only partially infilled channels or valleys.

In contrast, the channel facies in sequences 4 and 8 are restricted predominantly to the south of the Witch Ground Basin and in fact the basal surfaces of both sequences tend to become much less irregular in the Witch Ground Basin. This would therefore suggest that, as at present, water depths were shallowest in the southern part of the study area allowing for emergence and fluvial erosion prior to the deposition of sequences 4 and 8. In agreement with this, a wave cut erosion platform (boundary F) is restricted to the southern part of the area (Fig. 2.32).

To conclude, the structural setting of the area is recognised as one with a high long term preservation potential for glacial and glaciomarine sediments (Nystuen, 1985 and Bjorlykke 1985) and as such one might expect to find a more complete vertical record of glacial and interglacial periods than is documented from terrestrial Pleistocene outcrops in Europe. The presence of eight seismic sequences and at least three basin-wide erosion surfaces would appear to support the assumption. Possible chronostratigraphic ages of these boundaries and their relevance to glacially induced eustatic and isostatic effects will be discussed in detail in the next chapter.

CHAPTER THREE

PALAEOECOLOGY, BIOSTRATIGRAPHY AND CHRONOSTRATIGRAPHY

3.1 Introduction

Pleistocene stratigraphy is unique in the fact that it is primarily controlled by climatic variations and the effects these have had on the palaeontological and lithological components of any particular sequence. In this chapter I shall be dealing principally with the micropalaeontological components of the Pleistocene sequence, namely the dinoflagellate cysts and benthic foraminifera. The former were identified from numerous borehole and vibrocore samples by Rex Harland (BGS, Keyworth) and the foraminifera were analysed by Dianne Gregory (BGS, Keyworth). Their results are contained in a series of internal reports held by BGS.

The aims of the chapter are as follows:-

i. To briefly discuss the criteria used for interpreting microfauna and flora from the North Sea Pleistocene sequence.

ii. To define a series of diagnostic units for the dinoflagellate cysts and foraminifera respectively.

iii. To combine these units into environmentally diagnostic bio-units for each particular seismic sequence described in chapter 2. These units are essentially delineated by the presence of favourable or unfavourable assemblages. However, the fact that they have been constructed within seismic divisions or boundaries precludes their use as biozones (sensu-strictu).

iv. To attempt to date respective bio-units using a variety of methods, and to extrapolate data sequences using the seismic network.

With regards to the bio-units it should be stressed that they reflect the ecological response of organisms to environmental changes rather than evolutionary changes in flora and fauna. As such, they define a climatostratigraphy or geological-climatic units, each unit representing an inferred widespread climatic episode defined from a subdivision of Quaternary rocks (American

Commission 1961). In this particular area glacial and interglacial stages constitute the principal geologic-climatic units, while stadials and interstadials form units of lesser rank. Identification, ordering and correlation of these stages and chronozones forms the basis for the Pleistocene stratigraphy.

3.2 Palaeoecology

In chapter 1 it was mentioned that the present day water currents in the North Sea follow an anticlockwise pattern (Fig. 1.7). It was in fact Laevastu (1963) who first described the hydrographic setting in the North Sea in terms of an environment dominated by an anticlockwise flowing branch of the North Atlantic Drift. This current enters the North Sea between Orkney and Shetland, flows south along the east cost of Scotland, then east and eventually northwards along the Norwegian coast. The result of this is a hydrographic duality. In the northern North Sea the water mass is thermally stratified whilst in the southern North Sea a holothermal regime exists. However, it is likely that during glacial periods, as occurred through much of the Pleistocene, this division would have been precluded as a result of the southward extension of the Polar front (McIntyre et al., 1972) and the exclusion of warm North Atlantic Drift water from the North Sea.

In the oceanic record the feasibility of detecting such shifts in ocean currents and water masses, using various micropalaeontological and lithological evidence, has been widely recognised and used as a major characteristic for climatically subdividing the Pleistocene (McIntyre et al., 1972; Ruddiman and McIntyre, 1973; Lamb, 1974; McIntyre et al., 1976; Kellog, 1976 and Ruddiman and McIntyre, 1976). Furthermore such climatic variations could also be correlated to those in terrestrial sequences where the variations had long been used as a basis for stratigraphic subdivision.

The basis for studies of climatic variation in the oceanic record lies in being able to map variations in the position of the Polar Front (Ruddiman and McIntyre, 1976); that is the interface between cold, low-salinity, Polar water and warmer, more saline Atlantic water. Variations in the position of this are reflected by changes in both the benthos and phytoplankton and also in lithology, and these can often form an uninterrupted record in the Oceanic sediment.

However, as noted by Gregory and Harland (1978), relatively little attention has been paid to documenting climatic change in the North Sea and relating it to the known oceanographic history. In fact variations in the degree of influence of the North Atlantic Drift in the North Sea, given the effects they had on the fauna and flora (Harland et al., 1978), can be used as a climatic indicator as they must have been closely related to changes in the latitude of the Polar Front, which in turn occurred in response to palaeotemperature variations.

In this study, evidence of palaeoenvironmental and climatic change in the North Sea is based primarily on the identification of various dinoflagellate cyst and benthic foraminiferal assemblages. As such the distribution of dinoflagellate cysts in the North sea (Reid, 1975 and Reid and Harland, 1975) appears to be primarily controlled by the North Atlantic Drift and modified by a series of water masses described by Laevastu (1963). The distribution of benthic foraminifera in the North Sea (Murray, 1971) relates to a series of hydrographic provinces (Dietrich and Kalle, 1957), characterised by depth, temperature and salinity variations and again affected by the North Atlantic Drift. The criteria by which the dinoflagellate cyst and foraminiferal assemblages were defined and interpreted will be briefly described below.

Dinoflagellate cysts are a major group of phytoplankton found in almost all present day aquatic environments (Dale, 1985). They are particularly useful in that most cysts are palynomorphs composed of sporopollenin-like material and as such are relatively unaffected by dissolution problems. In the following sections, interpretation of the dinoflagellate assemblage is based primarily on cyst abundance and species variation. With regards to the former, a good dinoflagellate cyst recovery is interpreted as indicating a favourable environment whilst poor cyst recoveries relate to unfavourable conditions. Typical unfavourable conditions include harsh extremes of temperature, waterdepth and hyposalinity; all characteristic of glacial periods. This assumption is supported by the occurrence of only low numbers of cysts in both recent arctic sediments (Harland and Reid, 1977) and in Pleistocene glaciomarine sequences (Gregory and Harland, 1978).

Favourable conditions, which allowed for a greater dinoflagellate cyst productivity, are therefore taken as being indicative of an ameliorative period with the establishment of conditions similar to the those recorded at present from the study area. Also certain species of dinoflagellate cyst, namely *Operculodinium centrocarpum* (Deflandre and Cookson), *Bitectatodium tepikense* Wilson and *Spiniferites elongatus* Reid, are characteristic of the North Atlantic current (Williams, 1971) and North Atlantic Drift assemblage of Reid and Harland (1977), and their presence in a rich assemblage must indicate the presence of temperate conditions similar to today. More subtle relevancies in species interpreation will be discussed in section 5.

With regards to benthic foraminifera, although specific knowledge about their ecology is still limited, it is now generally accepted that their distribution is controlled directly or indirectly by water mass properties (Nagy and Quale, 1985). Thus in the present day North Sea the boundary between different water masses, described previously, determines the distribution of various foraminifera and it especially defines the northern limit of certain forms termed ´southern species´ by Murray (1971). It is the presence of such southern species which is used to delimit the occurrence of favourable ameliorative episodes during the Pleistocene and which in turn must indicate the presence, or near proximity, of a holothermal water mass during the summer, allowing these species to reproduce. The following species, compiled from a variety of publications (Murray, 1971; Sejrup et al., 1980; Skinner and Gregory, 1983; Nagy and Quale, 1985), are thought to be typical southern species: *Ammonia batavus* (Hofker), *Bulimina marginata* (d´Orbigny), *Cassidulina laevigata* d´Orbigny, *Elphidium selseyense* (Heron-Allen and Earland), *Hyalinea baltica* (Schroeter), *Nonion barleeanum* (Cushman), *Trifarina angulosa* (Williamson), and *Uvigerina peregrina* Cushman.

However, it is apparent that throughout the Pleistocene record in the North Sea the above species rarely dominate the foraminiferal assemblage, which is typically characterised by coldwater fauna indicative of harsh sub-arctic to arctic conditions. Such arctic fauna are defined on the basis that they commonly occur in relatively shallow water north of the Arctic circle, although their general distribution is often much wider than this. Typical arctic and sub-arctic type fauna include *Cassidulina reniforme* Norvang, *Elphidium claratum* Cushman, *Islandiella helenae* Feyling Hansen and Buzas, *I. norcrossi* (Cushman), *N labradoricum* (Dawson) and *Protelphidium orbiculare* (Brady).

The main difficulty that confronts palaeontologists when trying to reconstruct foraminiferal palaeoenvironments is ascertaining what percentage of southern species must be present to be indicative of an amelioration within an overall arctic environment. Harland et al. (1978) concluded that 50% indicated a viable population of southern species whilst 5% represented a non-viable or introduced population, but one still possibly indicative of an amelioration. Similar Jansen and Hensey (1981) determined that even low percentages of southern species were indicative of a proximity to North Atlantic central waters and Skinner and Gregory (1983) deemed that even in the ratio of one to several thousand certain southern species were significant in determining ameliorations.

A final point regarding the interpretation of the micro fauna and flora is one of terminology. Various terms are used throughout the literature to describe both assemblages, arctic, boreal or southern, and climatic variations. Here it was decided to describe the dinoflagellate cyst and benthic foraminifera primarily in terms of the degree of North Atlantic current influence (Harland et al. 1978) rather than to concentrate on more temperature dependent terms. Hence, the general terms favourable and unfavourable environments, related to ameliorative and harsh or arctic/glacial conditions, are used throughout the interpretation backed up by more detailed palaeoenvironmental evidence where present.

It should also be stressed that because of sampling spacing and variations in core recovery there is often a degree of extrapolation in placing the boundaries between respective units. Tables 3.1 and 3.2 summarise the main components of the succeeding sections. Diversity values in Table 3.2 are based on Walton's faunal diversity index which is based on the number of ranked species in an assemblage whose cumulative percentage accounts for 95% of the total. Fig. 3.1 summarises the position of these units within each borehole, their relation to the seismic and magnetic stratigraphy, and also the palaeoenvironment described in detail in section 3.5.

In appendices 2 and 3 downcore variations in dinoflagellate cysts and foraminifera are depicted, together with the position of each unit in the sediment column. A basic lithological log is also provided for comparison with the foraminiferal assemblages. The following section is basically intended to describe each dinoflagellate and foraminifera unit, and its distribution and position within each borehole.

Unit D		Main Dinoflagellate Cyst Species (%)	Specimens Per Slide	Reworked Cysts, %
XVI	3	O.centrocarpum (50-90),Spiniferites spp. (5-50), B.tepikiense (5-40), Protoperidinium spp. (5-10)	200-300	5
	2	V.E.135:- Protoperidinium spp. (50-75) V.E.111:- Similar assemblage to 1	25-50	75
	1	O.centrocarpum (60-90),B.tepikiense (10), Spiniferites spp.(5-20),Protoperidinium spp. (5)	200	10
XV		B.tepikiense (30-80),O.centrocarpum (20-50), Spiniferites spp.(5-30),Protoperidinium spp. (5-20).	20-50	50-70
XIV		O.centrocarpum (25-50),B.tepikiense (10-50), Spiniferites spp.(10-40),Protoperidinium spp. (0-30).	<75	10-50
XIII		O.centrococarpum (40-60),B.tepikiense (10-40), Spiniferites spp.(10-20),Protoperidinium spp. (5).	200	10
XII		O.centrocarpum (25-50),B.tepikiense (10-50), Spiniferites spp.(10-30),A.andulousiense (10-20)	20-100	
XI		O.centrocarpum (50-70),Spiniferites spp. (10-20),B.tepikiense (5-10), Protoperidinium spp. (5).	100-250	
X		O.centrocarpum (50),Spiniferites (20-30) B.tepikiense(10-20),Protperidinium spp(5-10)	<50	
IX		O.centrocarpum (30-40),B.tepikiense (20), Spiniferites spp.(20),Protoperidinium(5-10).	<20	
VIII		B.tepikiense(50-70),O.centrocarpum & Spiniferites spp.(10-40),A.andulousiense(5-25).	25-100	50
VII		O.centrocarpum(30-70),Spiniferites spp. & B.tepikiense(10-40).	200-300	5-10
VI	2	B.tepikiense(50).spiniferites spp.(10-20), O.centrocarpum(20-50),Protoperidinium spp.(10)		
	1	Spiniferites spp.(50-70),Protoperidinium spp. (5-50),A.andulousiense(10-40),B.tepikiense(10-20)	50	50-75
V		Spiniferites spp.(40-70),B.tepikiense(20-40) Protoperidinium spp.& O.israelianum(<5).	175-200	
IV		B.tepikiense(30-70),Spiniferites spp(10-50) Protoperidinium(10-40),O.israelianum(<5).	50	
III		O.centrocarpum(53-84),B.tepikiense(10-36) Spiniferites spp. (6-25)	100-200	
II	3	O.centrocarpum(2-86),B.tepikiense(6-7), Spiniferites spp.(5-25%).		
	2	Spiniferites spp.(10-30),T.pellitum(70-100)	100-200	
	1	Spiniferites spp.(50-70),T.pellitum(10-50), O.israelianum(10-40).		
I		Protoperidinium spp.(50-90),T.pellitum (10-20) & Spiniferites spp.(10-20).		

Table 3.1. Dinoflagellate cyst units.

Unit F	Main Foraminifera Species (%)	Diversity	Equivalent Zone D
XV	C.lobatulus & C.laevigata (1-40), B.marginata, H.baltica, U.peregrina, T.angulosa, C.carinata & C.reniforme (1-25).	7	D XVI
XIV	E.clavatum (60-94), C.reniforme (1-40), E.asklundi (1-8), P.orbiculare (0-16)	2-5	D XV
XIII	E.clavatum (50-70), C.reniforme (25-40), E.asklundi, P.orbiculare, C.lobatulus (<5).	2-4	D XIV
XII	E.clavatum (60-90), T.fluens (5-10), N.orbiculare (5), C.lobatulus & A.batavus(<5).	2-3	D XIII
XI	E.clavatum (50-70) C.reniforme (10-30), P.orbiculare (3-30), B.marginata, C.lobatulus, B.frida (<3).	1-10	D XII
X	P.orbiculare (40-78), E.clavatum (5-26), B.marginata (1-18), I.Helenae (1-7).	9-13	D XI
IX	E.clavatum (90), P.orbiculare & E.ustulatum (<5).	1-3	D X
VIII	E.clavatum (35-98), P.orbiculare(1-40). C.reniforme (1-38).	1-4	D IX
VII	E.clavatum (40-90), C.reniforme (1-30), B.marginata, B.vicksburgenesis, B.frigida (<10).	2	D VIII
VI	E.clavatum (0 ↑ 70), C.reniforme (3 ↓ 90), B.vicksburgenesis (0 ↓ 30), B.marginata(0-90), C.laevigata (1-23), T.angulosa (0-30).	1-8	D VII
V	E.clavatum (40-95), C.reniforme (0-23), C.teretis (0-30), P.orbiculare (1-20), B.frida, B.marginata, E.asklundi (10).	1-3	D VI
IV	E.williamsoni (40-70), E.clavatum (10-50).	-	D III
III	E.clavatum (60-80), C.obtusa? (5-30), E.bartletti (10-20), P.orbiculare (5-20).	1-3	
II	C.teretis (1-79), E.clavatum (1-90), P.orbiculare (0-28), T.fluens (0-50), E.williamsoni & E.ustulatum (0-22), B.marginata (0-100).	2-6	D II
I	E.clavatum (70-90), C.teretis & C.reniforme (<5).	1-3	D I

Table 3.2. Foraminifera units.

3.3 Dinoflagellate Units

3.3.1 Unit DI

This, the lowermost unit, was penetrated by boreholes 81/34 and 82/16 at 295m and 204m below sea level respectively. Its thickness varies from 16m in BH 81/34 to 29m in BH 82/16, whilst the base of the zone lies directly on Tertiary strata.

Dinoflagellate cysts are generally sparse throughout this zone, recording an average of less than 50 specimens per slide in BH 82/16 and even poorer assemblages in BH 81/34. *Protoperidinium* species (spp.) are the dominant cyst (50-90%) with subordinate *Tectatodinium pellitum* Wall and *Spiniferites* spp. (10-20%).

3.3.2 Unit DII

This was encountered in eight of the boreholes at a depth of 278m in BH 75/33, 297m in BH 77/12, 280m in BH 7713, 178m in BH 81/24, 280m in BH 81/26, 163m in BH 81/27, 282m in BH 81/34 and 183m in BH 82/16. Its thickness varies from 40m in the deeper water 77 and 75 boreholes to 12m in BH 81/24, (Appendix 2.1-2.10).

Unit DII is characterised by rich and diverse dinoflagellate cyst assemblages which display both lateral and vertical variations in the dominant cyst species. This creates a problem when trying to make interborehole correlations of the individual biofacies within this zone, and the subdivisions described below are only tentative.

The identification and correlation of the three biofacies is based on the dominant dinoflagellate cyst and, where present, their relationship to a period of normal magnetic polarity within the Matuyama epoch.

Biofacies 1, the lower most unit, occurs in boreholes 81/24, 81/26, 81/27, 81/34 and 82/16. The cyst assemblage is dominated by *Spiniferites* spp (50-707%) with subordinate *T.pellitum* (10-50%) and *Operculodinium israelianum* (Rossingal) (10-40%) in boreholes 81/24, 81/26, 81/27 and 82/16. In BH 81/34 *O. centrocarpum* and *B.tepikiense* are subordinate to *speniferites* spp. with only minor percentages of *T.pellitum* and *O.israelianum*.

Biofacies 2 was identified in two boreholes, 82/26 and 81/27, and tentatively BH 77/2. It is characterised by the dominance of *T.pellitum* (70-100%) with subordinate *Spiniferites* spp. (10-30%) and minor percentages of *O. israelianum* and *O. centrocarpum*.

Biofacies 3, encountered in boreholes 77/2, 77/3 and possibly 75/33, contains more significant proportions of *B.tepikiense* (6-71%) and *O.centrocarpum* (2-86%) with important percentages of *Spiniferites* spp. (5-25%) and in BH 77/2 *O. israelianum* (19%). A lack of detailed information of the assemblages in BH 75/33 precludes placing it definitively in this biofacies.

3.3.3 **Unit D III**

This unit occurs in borehole 75/33 between 221m and 240m and has not been identified in any other borehole. It consists of a rich dinoflagellate cyst assemblage dominated by *O. centrocarpum* (53-84%) with subordinate *B. tepikiense* (10-36%) and *Spiniferites spp.* (6-25%).

3.3.4 **Unit D IV**

This forms a thin unit in BH 81/26 between 272m and 280m. It is characterised by a very poor dinoflagellate cyst assemblage containing less than 50 speciments per slide. *B. tepikiense* is the dominant cyst (30-70%) with subordinate *Spiniferites spp.* (10-50%)a and *Protoperidinium spp.* (10-40%) and minor percentages of *O. israelianum*.

3.3.5 **Unit D V**

Directly overlying unit D IV this zone was penetrated at 262m in BH 81/26. The dinoflagellate cyst flora is much richer than in D III, consisting of between 175-200 specimens per slide. The cyst assemblage is dominated by *Spiniferites spp.* (40-70%) with important percentages of *B. tepekiense* (20-40%) and minor percentages of *Protoperidinium spp.* and *O. israelianum*.

It is possible that this unit is the lateral equivalent of the rich dinoflagellate cyst assemblage in D IV although this is only tentative and as such this assemblage has been assigned to a different unit.

3.3.6 **Unit D VI**

Nine boreholes penetrated this at the following depths:- 206m in BH 75/33; 138m in BH 81/19, 165m in BH 81/25, 180m in BH 81/26, 224m in BH 81/34, 186m in BH 81/39, 130m in BH 82/15, 150m in BH 82/16 and 174m in BH 84/13. It varies considerably in thickness reaching a maximum of 80m in BH 81/26 and thinning towards the coast until it forms a thin veneer some 5m thick along the western edge of the area.

The unit is best developed in BH 81/26 where two distinct biofacies can be identified. These appear to be strongly related to the lithology, as will be discussed later, and as such at least one of these biofacies can be recognised in all the boreholes mentioned above, depending on the lithology.

In BH 81/26 biofacies 1, the lowermost one, is characterised by a poor but diverse dinoflagellate cyst assemblage with less than 50 specimens per slide. The dominant species is *Spiniferites spp.* (50-70%) with subordinate *protoperidinium spp.* cysts (5-50%) and *Achomosphaera andulousiense Jan du chêne* (10-40%) with smaller percentages of *B.tepikiense* (10-20%) and *O. israelianum* (10%). Biofacies 2 again contains a poor dinoflagellate cyst flora but one dominated by *B. tepikiense* (50%). Other important species are *Spiriferites spp.* (10-20%), *protoperidinium spp.* (10%) and *O. centrocarpum* (5%).

Assemblages similar to biofacies 1 in BH 81/26 also occur in boreholes 81/39 and 82/16. In the latter the assemblage differs slightly in that at the base it is dominated by *Protoperidium spp.* (70%) with subordinate *Spiniferites spp.* (10-30%) but this passes up into a *Spiniferites spp.* dominated assemblage (60%) with subordinate *Protoperidinium spp.* (10-20%) and *B. tepikiense* (1-5).

The upper biofacies, 2, can also be identified in boreholes 82/15, 81/19 and 84/13. In the first two boreholes the dinoflagellate cyst assemblage is almost identifical to that described above whilst in BH 84/13 a poor but diverse fauna is dominated by *O. centrocarpum* (25-50%) with a strong presence of *B. tpikiense* (25-30%) and *Spiniferites spp.* (20-30%) and smaller percentages of *Protoperidinium spp.* (5-25%) and *A. andulousiense* (5-15%). The percentage of reworked cysts in this assemblage ranges from 50 to 75%.

In boreholes 75/33 and 81/25 the dinoflagellate cyst assemblage is very sparse and barren in places and probably forms an extreme equivalent to biofacies 2 in this unit.

3.3.7 **Unit D VII**

This unit occurs in two boreholes, 81/34 and 84/13. It is best developed in BH 81/34 where the unit, encountered between 164m and 224m, is characterised by a very rich dinoflagellate fauna with some 200-300 specimens per slide. *O. centrocarpum* is the dominant species (30-70%) with subordinate *Spiniferties spp.* (10-40%) and *B. tepikiense* (10-40%) and small percentages of *A. andulousiense* (5%) and *Protoperidinium spp.* (5%).

In BH 84/13 unit D VII is much thinner, occurring between 171m and 174m, and dinoflagellate cysts are less abundant with an average of 125 specimens per slide. Only low levels of reworking (10%) were recorded from this assemblage. *O. centrocarpum* is again the dominant species (60%) with important percentages of *Spiniferites spp.* (25-40%), *B. tepikiense* (10-30%) and *A. andulousiense* (0-10%) and small numbers of *Protoperidinium spp.* (10%).

3.3.8 **Unit D VIII**

Five boreholes penetrate this unit at a depth of 200m in BH 75/33, 183m in BH 77/2, 151m in BH 81/26, 138m in BH 81/34 and 156m in BH 84/13. Its thickness varies from 3m in BH 75/33 to 30m in boreholes 77/2 and 81/26.

The dinoflagellate cyst flora in this unit is relatively poor with between 25 and 100 specimens per slide, and up to 50% of this assemblage is reworked in BH 84/13. In both 81/34 and 84/13 there is a change in the dominant cyst species from the top to the bottom of the unit. At the base *B. tepikiense* is the dominant species (50-70%) in both boreholes, whilst towards the top of the unit *O. centrocarpum & spiniferites* (40%) become dominant in BH 81/34 and *spiniferties spp.* alone dominates the assemblage in 84/13. Other important species include *A. andulousiense* (5-25%) and *Protoperidinium* cysts (5-10%).

In BH 81/26 the dinoflagellate cyst assemblage in this unit is dominated by *B.tepikiense* throughout (50-60%), but with a strong presence of *O. centrocarpum* (20-50%) and spiniferites species (10-30%). *A. andulusiense* (5-10%) and *Protoperidinium spp.* (5%) are also present here.

3.3.9 **Unit D IX**

This unit occurs in four boreholes at a depth of 149m in BH 81/24, 95m in BH 81/34, 126m in BH 82/15 and 145m in BH 84/13 respectively. The dinoflagellate cyst assemblage is very sparse with generally less than 20 specimens per slide. *O. centrocarpum* (30-40%) is the dominant cyst with subordinate *B. tepikiense* (20%) and *Spiniferites spp.* (20%) and small numbers of *Protoperidinium spp.* (5-10%).

3.3.10 **Unit D X**

This unit can be positively identified in BH 81/37 and tentatively in BH 77/3, between 195m-201m and 167m-190m respectively. The lack of certainty is a reflection of the different and probably less accurate methods used to describe dinoflagellate cysts prior to 1980.

The dinoflagellate cyst fauna is essentially sparse often recording less than 50 specimens per slide. *O. centrocarpum* (50%) dominates the assemblage with subordinate *Spiniferites spp.* (20-30%), and *B. tepikiense* (10-20%) and smaller percentages of *Protoperidinium spp.* (5-10%).

3.3.11 **Unit D XI**

Three boreholes penetrate this unit at 165m in BH 77/2, 150m in BH 77/3 and 165m in BH 81/37. Although the presence of this unit in the last two boreholes is uncertain for the reasons explained above. The unit is best developed in BH 81/37 where it is 30m thick whilst in boreholes 77/2 and 77/3 it is 20m and 17m thick respectively.

Unit D XI is characterised by a rich dinoflagellate cyst fauna which in BH 81/37 is dominated by *O.centrocarpum* (50-70%) with subordinate *Spiniferites spp.* (10-20%) and smaller percentages of *B. tepikiense* (5-10%) and *Protoperidinium spp.* (5-10%). In 77/2 and 77/3 a rich dinoflagellate cyst fauna is still dominated by

O.centrocarpum (35-55%) but with important percentages of *B.tepikiense* (21-45%) and *Spiniferites spp*. (10-20%).

3.3.12 **Unit D XII**

Eight boreholes penetrated this unit at 175m in BH 75/33, 117m in BH 77/3, 132m in BH 81/19, 126m in BH 81/24, 142m in BH 81/26, 122m in BH 81/27, 125m in BH 81/37 and 141m in BH 82/16. It is best developed in boreholes 81/37 and 81/27 where it is some 40m-50m thick. However, in most of the other boreholes this zone is generally less than 10m thick.

The dinoflagellate cyst flora in this unit are relatively sparse, and barren in places, containing between 20 and 100 specimens per slide. *O. centrocarpum* (25-50%) is the dominant species with subordinate *B. tepikiense* (20-50%) and *Spiniferites spp*. (10-30%). Other species include small percentages of *A. andulousiense* (10-20%) and *Protoperidinium* cysts (5-10%).

3.3.13 **Unit D XIII**

This unit occurs in three boreholes at a depth of 121m in BH 81/25, 130m in BH 81/26 and 139m in BH 82/16. It has a maximum thickness of 12m in 81/26 and a minimum in 82/16 of only 2m.

The dinoflagellate cyst flora in this unit is characterised by a rich assemblage giving at least 200 specimens per slide with only low levels of reworking (10%). *O. centrocarpum* (40-60%) is the dominant species with subordinate *B. tepikiense* (10-40%) and *Spiniferites spp*. (10-20%) and small percentages of *Protoperidinium spp*. (5-10%). The exception to this is in BH 81/25 where the top of this unit is dominated by *B. tepikiense* (85%) whilst the base is dominated by *O. centrocarpum* (100%).

3.3.14 **Unit D XIV**

This unit is best developed in BH 84/13 where it is encountered between 133m-145m. The unit was also penetrated in boreholes 77/2 at 157m, 81/24 at 100m, 81/27 at 110m and 82/5 at 112m.

Detailed analysis of the Dinoflagellate cyst assemblage in boreholes 77/2 and 84/13 shows that the flora is relatively sparse (<75 specimens per slide) and contains up to 50% of reworked specimens. *O. centrocarpum* (25-50%) is the dominant species but with strong elements of *B. tepikiense* (10-50%) and *Spiniferites spp.* (10-40%). The assemblage in BH 82/15 differs from this in that *Spiniferites spp.* are the dominant dinoflagellate cyst with subordinate *Protoperidinium spp.* (30%) and smaller percentages of *O. centrocarpum* and *B. tepikiense* (10-15%).

4.3.15 **Unit D XV**

Seven boreholes penetrated this zone at a depth of 156m in BH 75/33, 150m in BH 77/2, 108m in BH 77/3, 126m in BH 81/19, 109m in BH 81/25, 90m in BH 81/34, 100m in BH 81/37 and 137m in BH 81/39. Its thickness varies from 50m in BH 81/39, where this unit occurs in a channel infill, to 1m in BH 81/25 along the western edge of the study area.

The dinoflagellate cyst assemblage in this unit is characterised by a very sparse recovery with between 20 and 50 specimen per slide. *B. tepikiense* (30-80%) is the dominant species with important percentages of *O. centrocarpum* (20-50%), *Spiniferites spp.* (5-30%) and *Protoperidinium spp.* (5-20%). The exception to this is seen in BH 81/37 where *Protoperidinium spp.* are the dominant (70%) cyst.

In addition to the above assemblage this unit was also sampled and analysed in detail in two virbocores (V.E.), 58+00/111 and 58+00/135 (Appendix 2.11-2.12), between 142.5m - 145.4m and 144.1m-146.9m. The assemblage in both vibrocores is characterised by a poor recovery (20-25 cysts per slide) and high levels of reworking (50-70%). The low counts of indigenous dinoflagellate cysts may render the species proportions meaningless but in V.E. 111 *B. tepikiense* (20-60%) dominates the lower half of this zone becoming subordinate to *O. centrocarpum* (20-50%) towards the top of the zone. In V.E. 135 there are similar proportions of *O. centrocarpum* (20-50%) and *B. tepikiense* (10-50%). Other important species include *Spiniferites spp.* (5-30%) and smaller percentages of *Protoperidinium spp.* (5-20%).

3.3.16 **Unit DX VI**

This, the uppermost unit, occurs in 12 boreholes where it crops out at the sea bed at the following water depths:- 150m in BH 75/33, 147m in Bh 77/2, 100m in BH 77/3, 120m in BH 81/19, 122m in BH 81/26, 103m in BH 81/27, 82m in BH 81/34, 93m in BH 81/37, 126m in BH 81/39, 120m in BH 82/16 and 125m in BH 84/13. Its apparent thickness varies from 4m to 8m except in 82/16 where it appears to be 19m thick. However the above thicknesses must be viewed with caution because of the generally poor core recovery in the uppermost part of these boreholes. Therefore a core assigned to a depth of 7m, for example, may in fact come from a shallower depth, as will be explained in chapter 4.

The dinoflagellate cyst assemblage in this unit is the most consistent, between boreholes, of those described so far. It is characterised by a very rich dinoflagellate cyst flora, with generally over 200 specimens per slide. *O. centrocarpum* (55-70%) is consistently the dominant species, although it does decrease in numbers towards the base. Subordinate species include *B. tepikiense* (10-40%) with smaller percentages of *Spiniferites spp.* (10-30%) and *Protoperidinium spp.* (5-10%).

This unit also occurs in V.E 111 and V.E 135 where detailed sampling and analysis of the dinoflagellate cyst assemblage allows the unit to be divided into 3 biofacies; not recognised in the above boreholes probably as a result of greater sampling intervals. It is interesting to note that in these vibrocores this unit is only 2.5-3m thick (from the sea bed) therefore supporting the belief that the earlier quoted borehole thicknesses are in some cases misleading.

Biofacies 1 occurs between 141.1m-142.4m and 143.6m-144m in vibrocores 111 and 135 respectively. It is characterised by very low levels of reworking (10%) and rich dinoflagellate cyst fauna giving over 200 specimens per slide. The assemblage is dominated by *O. centrocarpum* (60-90%) with minor amounts of *B. tepikiense* (10%), *Spiniferites spp.* (5-20%) and *Protoperidinium spp* (5%). The occurrence and importance of individual species members will be discussed in the next section.

Biofacies 2 forms a relatively thin sequence, occurring between 140.4m-141.1m and 143.1m-143.6m in V.E 111 and 135 respectively. This biofacies represents a phase of high reworking, generally greater than 75%, and poor indigenous dinoflagellate cyst recovery. In V.E 111 the assemblages recovered remain similar in proportions to those in biofacies 1, apart from a slight increase in *B. tepikiense* and *Protoperidinium spp.* However in V.E. 135 biofacies 2 is characterised by a marked dominance of *Protoperidinium* cysts (50-75%), similar to that seen in 81/37, with minor proportions of *O. centrocarpum, B tepikiense and spiniferites spp.*

Biofacies 3 occurs from the sea bed to a depth of 140.4m and 143.1m, in V.E 111 and 135 respectively. It is characterised by extremely low levels of reworking, generally about 5%, and a rich assemblage of indigenous dinoflagellate cysts. This biofacies is best developed in V.E 135 where the assemblage is dominated by *O. centrocarpum* (50-90%) with subordinate *Spiniferites spp.* (5-50%); the latter increasing in importance towards the sea bed. Other species include *B.tepikiense* (5-40%) and *Protoperidinium spp.* (5-10%).

3.4 **Foraminiferal Units**

3.4.1 **Unit F I**

This, the lowermost foraminiferal unit was penetrated by boreholes 75/33, 81/34 and 82/6 at 318m, 295m and 204m respectively. *E. clavatum* dominates the assemblage although in 82/16 the fauna is slightly more diverse containing subordinate *Cassidulina teretis Tappen* and traces of *C. reniforme,* (Appendix 3.1-3.2).

3.4.2 **Unit F II**

Six boreholes penetrated this unit at a depth of 278m in BH 75/33, 289m in BH 77/2. 304m in BH 81/26, 162m in BH 81/27, 289m in BH 81/34 and 192m in BH 82/16. It is best developed in BH 77/2, where it is some 50m thick, although a poor core recovery at this level precluded the identification of any variations within the 50m.

The foraminiferal assemblage is relatively rich and diverse, especially in BH 77/2 where it can be divided into an upper and lower unit. The upper unit is dominated by *E. clavatum* (45-90%), with subordinate *C. teretis* (0-35%). Other species occurring here include *Bucella frigida Cushman, C. reniforme,* and *Quinqueloculina seminulum* (Linne). In the lower unit *C. teretis* is usually dominant (22-79%), whilst other important species include *E. clavatum* (1-48%), *P. orbiculare* (0-28%) and *N. Barleeanum* (0-25%). There are also minor proportions of *B. frigida, C. reniforme* and *Q. seminolum*.

In boreholes 81/34 and 81/26 the foraminiferal assemblages are similar to the lower unit BH 77/2. However, in the remaining boreholes the assemblages are similar to the upper unit of 77/2 with dominant proportions of *E. clavatum* (20-90%) and subordinate *C. teretis* (2-60%). Other species occurring in this zone include *Trifarina fluens* Todd (0-50%) and small numbers of *B. frigida, Bucella vicksburgensis* (Cushman and Ellison). A single sample at 79m (below sea bed) contains 100% *Bulmina marginata* (d'Orbigny). The fauna is less diverse in BH 81/34 containing just small proportions of *P. orbiculare* subordinate to *C. teretis*.

3.4.3 **Unit F III**

This unit, which occurs between 248m and 278m in BH 75/33 does not have an equivalent dinoflagellate unit due to a lack of information regarding the cyst assemblage at this level. The foraminiferal fauna in this unit is moderately abundant, and is dominated by *E.clavatum* with *C.teniforme, Elphidium bartletti* Cushman and *P. orbiculare* as common subsidiaries.

3.4.4 **Unit F IV**

This unit is the approximate equivalent of D III and occurs in BH 75/33 between 226m and 248m. Although *E. clavatum* remains abundant in this zone, the foraminiferal assemblage is dominated by *Elphidium cf. williamsoni Haynes*.

3.4.5 **Unit F V**

Eleven boreholes penetrate this unit at 206m in BH 75/33, 217m in BH 77/2. 188m in BH 77/3, 18m in BH 81/19, 180m in BH 81/26, 155m in BH 81/27, 173m in BH 81/29, 224m in BH 81/34, 186m in BH 81/39 and 150m in BH 82/16. Its thickness ranges from 124m in BH 81/26 to 7m in BH 81/27 and, although some of the depths do not match, unit D VI forms the equivalent dinoflagellate cyst assemblage to this zone.

E. clavatum is the dominant species (60-95%) in this unit which is further characterised by an often poor fauna and some barren samples. Other important species include *C. reniforme* (0-23%) and *P. orbiculare* (1-20%) and minor percentages of *Elphidium ustulatum Todd, E. asklundi Brotzen* and *C. teretis.*

In BH 81/26, where this unit is best developed, the foraminiferal assemblage is more complex and diverse. *E.clavatum* remains dominant throughout (40-90%) apart from a thin section dominated by *P. Orbiculare* (20-90%). Subordinate species include *A. Batavus* (0-5%, 50% in one sample), *C. teretis* (0-30%) and *E. ashlundi* (0-11%). Similarly in BH 1/19 the foraminiferal assemblage passes down from an *E. clavatum* dominated fauna (95%) to a more diverse fauna with important percentages of *B. marginata* (11%), *B. frigida* (10%) and *P. orbiculare* (11%).

3.4.6 **Unit F VI**

This unit was encountered in only one borehole, 81/34, and it is the equivalent to the dinoflagellate cyst unit D VII seen in boreholes 81/34 and 84/13. Unfortunately subsamples from 84/13 were not analysed for their foraminiferal content so it is possible that this unit also occurs in BH 84/13.

Unit F VI is encountered in BH 81/34 between 144m and 224m where it is characterised by a very rich and diverse foraminiferal fauna. This assemblage is dominated by E. clavatum (60-74%) in the top half of the zone with subordinate *C. reniforme* (24%) and smaller percentages of *B. marginata* (0-7%) and *B. vicksburgensis* (0-3%). In the lower half *E. clavatum* becomes subordinate to *C.reniforme* (3-90%) with a strong presence of *B. wicksburgensis* (30%) less and smaller numbers of *Q. Seminulum* (5%). The base of this unit is marked by a sample containing *B. Marginata* (28-93%) with subordinate *C. laevigata* (1-23%) and *T.angulosa* (30%). *E. clavatum* is absent from this lowermost sample.

3.4.7 Unit F VII

This unit occurs in five boreholes at a depth of 202m in BH 75/33, 186m in BH 77/2, 151m in BH 81/26, 138m in BH 81/34 and 197m in BH 84/11. In the first three this unit forms only a thin layer 6m thick whilst in boreholes 77/2 and 81/26 it increases to a thickness of 30m.

In all five boreholes the assemblage is dominated by *E. clavatum* (40-90%) with subordinate *C.reniforme* (1-30%). In BH 81/34 thee are also small percentages of *B. Marginata* (1-7%), *B. Vicksburgenesis* (0-3%) and *B. frigida* (0-2%). Small numbers of *E. ustulatum* (5%) and *Cibicides lobatulus* (Walker & Jacob) (5%) are also present in BH 81/34 whilst at the base of the unit in this borehole a single sample contains 50% *A. batavus*.

4.4.8 Unit F VIII

Two boreholes penetrated this unit, 81/29 and 81/34, between 95-173m and 95-138m respectively. The foraminiferal fauna is dominated by *E. clavatum* (35-98%) with subordinate *P. Orbiculare* (1-40%), except at 18m where the assemblage is dominated by *P. orbiculare* (80%). *C. reniforme* (1-38%) is also important together with minor percentages of *B. Frigida, B. marginata* and *E. asklundi*.

3.4.9 Unit F IX

This unit equates approximately with unit DX and occurs in BH 81/37 between 173m-207m. It is characterised by a sparse foraminiferal fauna dominated by *E. clavatum* (90%) with minor percentages of *P. orbiculare* and *E. Ustulatum*.

3.4.10 Unit F X

This unit occurs only in BH 81/37 (Appendix 3.11) between 163m-173m. It can be taken as being the approximate equivalent to D XI.

The unit is characterised by a very rich assemblage which is dominated by *P. orbiculare* (40-78%). Other important species include *E. clavatum* (5-26%), *B. marginata* (1-18%), *I. helenae* (1-7%) and *E. ashlundi* (0-7%). There are also minor occurrences of *Polymorphinid, T. angulosa* and *I. norcrossi*.

3.4.11 **Unit F XI**

Seven boreholes penetrated this unit at a depth of 174m in BH 75/33, 117m in BH 77/3, 132m in BH 81/19, 142m in BH 81/26, 120m in BH 81/27, 125m in BH 81/37 and 141m in BH 82/16. As for unit D XII, this unit is best developed in boreholes 81/37, 81/27 and also 77/3 where it reaches a thickness of about 50m. In the remaining boreholes this zone is generally 10m or less thick.

The foraminiferal fauna is consistently dominated by *E. clavatum* (50-70%) with subordinate *C. reniforme* (10-30%) and *P. orbiculare* (3-30%). In 81/37, 81/27 and 82/16 there are also minor amounts of a number of other species including *B. marginata, C. lobatulus, B. frigida, T. fluens* and, in 81/37, *E. ambiumbilicatum*.

3.4.12 **Unit F XII**

This unit the lateral equivalent to D XIII, occurs in BH 81.26 between 130m-142m. *E. clavatum* (60-90%) dominates the foraminiferal assemblage with smaller amounts of *T.fluens* (5-10%), *N. orbiculare* (5%) and *C. lobatulus* (<5%).

3.4.13 **Unit F XIII**

This unit was encountered in four boreholes, 75/33, 77/2, 84/11 and 82/16, between 165m-174m, 157m-167m, 178-198m and 139-141m respectively. IN BH 77/2 the foraminiferal assemblage, although dominated by *E. clavatum* (50-70%), is reltively mixed with subordinate *C. reniforme* (25-45%) and minor numbers of *E. ashlundi, P. orbiculare, C. lobatulus, T. fluens, E. ustulatum* and *B. marginata*.

In the remaining boreholes the assemblage in this zone is more sparse, being dominated by *E. clavatum* (39%-75%) with subordinate *C.reniforme* (62%-30%).

3.4.14 **Unit F XIV**

Eight boreholes penetrated this unit, the equivalent to D XV, at a depth of 156m in BH 75/33, 150m in BH 77/2, 126m in BH 81/19, 82m in BH 81/34, 93m in BH 81/37, 137m in BH 81/39 and 140m in BH 84/11. *E. clavatum* (60-94%) dominates the assemblage in this unit with subordinate *C. reniforme* (1-40%) and *E. ashlundi* (1-8%).

Variations to this occur where this unit is thickest (40-60m) and in BH 81/39 high percentages of *P. orbiculare* were recorded (16-19%) but with *E. clavatum* remaining the dominant species. In the remaining boreholes this unit is generally less than 8m thick.

Unit F XIV also occurs in V.E. 111 where it was sampled in detail and found to be very similar to the assemblage described above, (Appendix 3.13).

3.4.15 **Unit F XV**

The top of this unit crops out at the sea bed and is encountered in six boreholes at a depth of 150m in BH 75/33, 147m in BH 77/2, 120m in BH 81/19, 103m in BH 81/27, 126m in BH 81/39 and 120m in BH 82/16. Its thickness ranges from 2m to 8m although in some cases this is probably exaggerated as was explained for unit D XVI.

This unit is best developed in boreholes 81/27 and 81/19 where the foraminiferal assemblage is dominated by *C. lobatus* (39%) and *C. laevigata* respectively. Other important species include *B. marginata* (19-22%), *H. baltica* (11%), *U. peregrina* (23% in 81/19) and *T. angulase* (3%). *E. clavatum* forms only a minor proportion of the assemblage in both boreholes.

In the remaining boreholes *E. clavatum* is still important (50-70%) with generally smaller percentages of *E. ashlundi* (1-53%), *C. reniforme* (18-22%), *P. orbiculare* (11%), *B. vicksburgensis* (<5%) and *B. marginata* (<5%).

This unit also occurs in V.E. 111 where detailed examination of the foraminiferal assemblage showed it to be similar to that described for boreholes 81/19 and 81/27. *E. clavatum* is absent and the assemblage is dominated by roughly equal proportions of *B. marginata, H. baltica, T. angulosa, U. peregrina,* and *C. reniforme*.

3.5 **Palaeoecological Interpretation of the Microfauna and Flora**

The aim of this section is to combine the previously identified dinoflagellate cyst and foraminiferal units into a series of palaeoecologically and stratigraphically significant bio-units for each individual seismic sequence described in chapter 2. In doing this it became apparent that units indicative of harsh or glacial periods were virtually indistinguishable from one another by their microfossil content. Primary attention was therefore paid to those units containing dinoflagellate and foraminiferal assemblages indicative of more favourable or ameliorative periods. However, when combining environmentally similar faunal and floral assemblages for a particular sequence their upper and lower boundaries were not always in agreement, a common problem in climatostratigraphy (Lowe & Walker, 1984). The reasons for this will be discussed in the following section and, purely for consistency, the maximum extent of both faunal and floral assemblages indicating favourable conditions was taken as the definitive bio-unit.

As mentioned above it was the intention of this section to place the microfaunal and floral record within the previously described seismic stratigraphy and then interpret it. Fortunately the majority of seismic sequence boundaries appear to correlate with a microfaunal unit boundary, as shown in appendices 2 and 3, possibly reflecting similar controls on their occurrence such as sea level, basin configuration and sediment supply. The exceptions to this are seen in some micropalaeontological assemblages indicative of harsh or glacial conditions, which appear to continue across seismic sequence boundaries. However, as was mentioned above harsh or glacial units of different ages are often indistinguishable from each other. Unit boundaries were therefore drawn corresponding to seismic sequence boundaries (Fig. 3.1). The method allows for a better degree of inter borehole correlation between respective glacial or harsh assemblage units. Subsequent amino acid studies of the fauna from BH 81/26, described later, proved the validity of this method in that apparently continuous glacial microfauna zones cut across sequence boundaries which were proved to represent a marked hiatus.

3.5.1 **Unit A**

This, the lowermost unit, is represented by the microfaunal and floral assemblages in units D I and F I. It occurs at the base of seismic sequence 1.

The low number of dinoflagellate cysts in this unit suggest a low cyst productivity indicating unfavourable or harsh environmental conditions. Here, the term unfavourable could reflect a variety of environmental extremes including temperature, sea ice cover, and salinity (Dale, 1985). The dominance of *Protoperidinium spp.* , namely the undifferentiated round brown morphotype, further suggests a relatively shallow, probably cold, water environment whilst the lack of species from the ´North Atlantic Drift´ assemblage of Reid and Harland (1977) points to a cold or glacial period with the polar front located well to the south of its present position.

A similar, harsh environment is also suggested by the foraminifera in this unit. The sparse fauna and low diversity together with a heavy dominance of *E. clavatum* is typical of a high arctic, hyposaline, shallow water environment (Jansen & Henessey, 1981). Similar foraminiferal assemblages have also been interpreted as typical of glacial deposits (Feyling, Hanssen et al. 1971; Osterman and Andrews, 1983; and McCabe et al. 1986).

In conclusion both the dinoflagellate cysts and foraminiferal fauna in unit A reflect harsh and restricted, shallow water conditions, probably related to a glacial period. The relatively good agreement between the dinoflagellate and foraminiferal zones may suggest that the water mass during this period was not stratified.

3.5.2 **Unit B**

This unit occurs within the combined limits of units DII and F II in seismic sequence 1. It is best characterised by the rich dinoflagellate assemblage suggesting favourable conditions and a generally ameliorative period compared with the underlying unit A. This interpretation is supported by the presence of significant numbers of *I. pellitum* and *O. israelianum,* both now restricted to warm temperate seas off Florida and Africa (Harland, 1983). Also the rich recovery of *O. centrocarpum, B. tepikiense* and *Spiniferites spp.* would suggest a penetration of North Atlantic current waters into the North Sea and a polar front located at a latitude similar to its present day position.

Noticeable variations in cyst proportions led to the identification of 3 biofacies. These are most convincingly correlated between boreholes 81/27 and 81/26, where a change from a lower, spiniferites spp. dominated, assemblage up into a *T.*

pellitum assemblage probably reflect a change from a nearshore sub-littoral environment depending up into a more offshore environment (Harland, 1983).

Where *O. centrocarpum* or *B. tepikiense* is dominant, in this zone, the environmental implications are less clear. However, it is possible that they reflect a less stable environment, but one still occurring within a relatively warm or ameliorative period. This is based on the fact that both species can tolerate reductions in salinity (Wall et al, 1977) and *O. centrocarpum* especially is extremely cosmopoliton (Dale, 1985) and is often associated with an unstable environment where two different water masses converge.

Evidence of an ameliorative period within the foraminiferal assemblage is more restricted. This is typified by boreholes 82/16, 81/26 and 81/34 (Fig. 3.1) where the vertical extent of the dinoflagellate ameliorative episode is almost twice that of the foraminiferal one. However, the presence of a richer and more diverse foraminiferal fauna relative to the zone below certainly indicates a degree of amelioration, in agreement with the dinoflagellate cysts.

The overall reduction in *E.clavatum* and the dominance of *C. teretis* suggests deeper water, at least greater than 20m, and probably a moderate increase in water temperature. More favourable conditions are indicated from some samples where there are between 5-50% of southern species, defined earlier, suggesting a penetration of temperate North Atlantic central water (Jansen & Hensey, 1981), and the establishment of water mass regime similar to today (Laevastu, 1963). In BH 82/16 the strong presence of *B. marginata* gives evidence of a full amelioration but with shallow water conditions. This interpretation is consistent with the shallow depth of this unit (below sea level) in 82/16 relative to some of the other boreholes. Also the occurrence of *I. helenae* is further evidence of an amelioration, this species being associated with the onset of improved conditions following a glacial period in Frobisher Bay, Baffin Island (Osterman and Andrews, 1983).

In conclusion this unit represents a strong ameliorative period, which, on the basis of dinoflagellate flora from the base of the unit was slightly warmer than today, whilst overlying assemblages indicate the presence of an environment very close to that found at present. Similarly the foraminifera point to an ameliorative episode, again strongest at the base of this unit. However in the upper half of the unit the foraminifera suggest a return to harsher conditions in contrast to the rich dinoflagellate fauna. Such an anomaly was previously recorded by Gregory and Harland (1978) who attributed it to the presence of a stratified water mass and continued cold bottom waters.

3.5.3 **Unit C**

The occurrence of this unit is restricted to BH 75/33, where it is within seismic sequence 1. The sparse recovery and low diversity of the foraminiferal assemblage suggests a return to harsher conditions. This is supported by the dominance of E. clavatum with subordinate *E. bortletti* and *P. orbiculare* suggesting a shallow water, arctic environment, similar to unit A. No catalogued information was available regarding the dinoflagellate cysts in this unit, although it was apparently virtually barren of dinoflagellate cysts (R. Harland, pers. com.).

3.5.4 **Unit D**

This unit is again restricted to BH 75/33 where it occurs within seismic sequence 1. Its boundaries are defined by the combined upper and lower limits of units D III and F IV.

The rich dinoflagellate cyst assemblage, dominated by *O. centrocarpum* with subordinate *B. tepikiense* and *Spiniferties spp.* would suggest that the North Atlantic current was operating normally giving environmental conditions similar to those described for the upper part of zone A.

In the foraminiferal assemblage the strong presence of *E. Williamsoni* suggests the presence of temperate, shallow water conditions (Knudsen, 1985).

Unit D therefore represents an upper ameliorative period in seismic sequence 1, although it has only been recorded in a single borehole.

3.5.5 **Unit E**

The occurrence of this unit is restricted to BH 81/26, although it is possibly a lateral equivalent to unit C in BH 75/33. However, because of the different assemblages in the respective units and lack of any other evidence for correlation, two separate units were defined. As such, unit E is represented by the dinoflagellate unit D IV and occurs within seismic sequence 1.

The poor dinoflagellate cyst recovery in this unit, together with the lack of *O. centrocarpum* or strong presence of *T. pellitum* and *O. israelionum* suggests the presence of unfavourable conditions with only limited North Atlantic influence. This interpretation is consistent with the high proportion of *B.tepikiense* and subordinate *Protoperidinium spp.* cysts. Dale (1985) noted that *B. tepikiense* often dominates cold sequences whilst Turon (1980) described the association of *B. tepikiense* and *Protoperidinium spp.* as being typical of conditions north of 60°N in the Norwegian sea. Similarly Harland (1983) describes *B.tepikiense* concentrated around the Iceland Scotland ridge.

In conclusion the dominance of *B. tepikiense* in a sparse dinoflagellate assemblage indicates unfavourable, cold climatic conditions in an offshore area. Further evidence also suggests that *B.tepikiense* may enjoy less than fully marine salinities in an environment influenced by glacial meltwater (R. Harland, pers. com.). This can be compared with the *Protoperidinium spp.* dominated assemblage in zone A which again reflects cold conditions but in shallower coastal regions (Dale, 1985).

3.5.6 **Unit F**

This unit is represented by the dinoflagellate unit D V, in seismic sequence 1. Although its occurrence is restricted to BH 81/26 it could be tentatively correlated with unit D in BH 75/33. However different faunal assemblages in the two units and the lack of any other evidence suggests the need for caution.

The unit is characterised by a rich dinoflagellate cyst assemblage dominated by *Spiniferites spp* and is indicative of an ameliorative period. However, the preponderance of the species *A. andalousiensis* in the assemblage (R Harland, pers. com.) suggests cool conditions rather than a full amelioration.

In conclusion this unit represents an ameliorative period, but one that is less strong than indicated in unit B. The foraminifera show no evidence of any change in an essentially arctic assemblage suggesting either that the water mass was stratified or that the amelioration was relatively weak and short lived, therefore precluding the migration and establishment of more temperate fauna (D. Gregory, pers. com.).

3.5.7 **Unit G**

This unit contains the combined faunal and floral assemblages of units D VI and F V and forms the uppermost zone in seismic sequence 1.

In parts this unit is barren of dinoflagellate cysts, and where they do occur the flora is sparse and characterised by a high degree of reworking sugesting harsh, unfavourable conditions typical of a glacial period. Various cyst species dominate the flora at different locations, perhaps as a result of the large amount of re-working and the sparsity of indigenous fauna.

Generally the foraminiferal assemblage is consistent with the above interpretation in that there is a sparsity of fauna, low species diversity, and strong dominance of *E. clavatum* again reflecting high arctic conditions. The one inconsistency occurs in BH 81/26 where unit F V is characterised by significant numbers of *C. teretis* together with southern species, especially *A. batavus*. In the light of the dinoflagellate record it is suggested that the temperate species were re-worked and incorporated into this unit. Such an interpretation would be in accord with the lithology, a sandy diamict. Separate analysis of the foraminifera at this level in BH 81/26 by Bergen University produced similar results, which likewise were attributed to re-working (H. Sejrup, pers.com.).

In conclusion this unit represents a glacial period at least as extreme as that recorded in unit A. Also the strong presence of southern species in the foraminiferal assemblage highlights the importance of assessing all the available data including other micropalaeontological evidence and the lithology from which the sample was obtained.

3.5.8 **Unit H**

This unit is represented by the assemblages in units D VII and F VI which occur at the base of seismic sequence 2. The dinoflagellate flora is rich, with only low levels of re-working. The species proportions are very similar to those in unit D and it suffices to say here that the assemblage reflects temperate conditions, similar to today.

The foraminiferal assemblage in this unit, seen only in BH 81/34, is characterised by a rich and diverse assemblage suggesting a strong amelioration in agreement with the dinoflagellate cysts. At the base there is an abundance of southern species indicative of normal salinity and temperate waters. The change to a *C. reniforme* dominated assemblage suggests an increase in the water depth and less temperate (Sejrup & Guibert, 1980) waters, although there is still a significant proportion of southern species.

It is interesting to note that at 136m in BH 81/34 there is a sharp break in the foraminiferal assemblage that separates a lower *B. marginata* dominated fauna from an upper fauna containing significant proportions of *B. vicksburgensis* and no *B. marginata* . On the dinoflagellate record this break can be identified as separating a lower *Spiniferites spp.* dominated flora from an upper *O. centrocarpum* dominated floral. From this evidence it is suggested that the lower assemblage is stenohaline reflecting a stabler marine environment whilst the upper assemblage is more cosmopolitan and able to withstand fluctuating conditions.

In conclusion unit H reflects an extensive ameliorative period, possibly the thickest yet recorded from the North sea. Interestingly the unit forms the lower half of a channel infill and the situation probably compares well with that in the present day North Sea where sediments containing temperate fauna are slowly accumulating in the base of open channels.

3.5.9 **Unit I**

This unit occurs within the combined limits of units DV III and F VII and forms the uppermost unit in seismic sequence 2. The poor dinoflagellate cyst recovery and high degree of reworking is indicative of a glacial period. The presence of *Achomosphaera* *andalousiensis* within a spiniferites spp. dominated assemblage is consistent with this interpretation.

The foraminiferal assemblage, heavily dominated by *E. clavatum* and *C. reniforme,* is similar to the glaciomarine assemblage of Elverhoi and Bomstand (1980) and is consistent with greater water depths than encountered in previous glacial zones. Errant individuals of southern species may be spurious or the first indications of penetration by the North Atlantic current.

In conclusion unit I represents a cold period, but one in which conditions were possibly less severe than those indicated by previous units.

3.5.10 **Unit J**

This unit, the only one in seismic sequence 3, is represented by units D IX and F VIII. The dinoflagellate flora is very sparse, and barren in places, suggesting an extremely harsh environment with little or no penetration of the North Atlantic current. Species proportions are probably meaningless with such a low cyst recovery.

Similarly the foraminifera reflect very adverse conditions, especially in BH 81/29 where a thick sand unit is virtually barren of foraminifera (Appendix 3.8). Where present, the foraminifera indicate high arctic shallow water conditions, the assemblage being heavily dominated by *E. clavatum* with subordinate *P. orbiculare* and *C. reniforme*. The small percentages of southern species in 81/34 are probably reworked.

In conclusion the fauna and flora in unit J indicate a harsh, unfavourable environment which may be partly reflected by the coarser lithology in this zone.

3.5.11 **Unit K**

This unit combines units D X and F IX, which occur at the base of seismic sequence 4. The sparse dinoflagellate flora, dominated by *O. centrocarpum,* is indicative of adverse conditions and a glacial period. Similarly the foraminiferal assemblage is characterised by a poor recovery and low diversity, with a very high percentage of *E. clavatum,* reflecting hyposaline, high arctic, shallow waters.

3.5.12 **Unit L**

This unit occurs within the combined limits of units D XI and F X which are present in seismic sequence 4. The very rich dinoflagellate cyst assemblage contains a significant proportion of Reid and Harland's (1977) 'North Atlantic Drift' assemblage, reflecting temperate conditions similar to those described for previous ameliorative periods.

In the foraminiferal assemblage the dominance of *P. orbiculare* is indicative of shallow water depths and relatively higher energy (Knudson, 1985). Further to this, the rich recovery and diverse fauna containing significant proportions of southern species, especially *B. marginata,* reflects the establishment of temperate, stenohaline conditions. It should be noted that this assemblage was only recovered from BH 81/37, in the remaining boreholes a distinct amelioration in the dinoflagellate record did not appear to correspond to any improvement in the foraminiferal assemblage. Also the vertical extent of the foraminiferal amelioration in BH 81/37 is very restricted compared to that indicated by the dinoflagellate record. It is possible that thermal stratification of the water mass may account for the anomaly between the two although the shallower water depths indicated here would normally prevent such an occurrence. Alternatively, the difference may reflect slower response and migration rates in the foraminifera assemblage relative to the dinoflagellate cyst assemblage (R. Harland, pers. com.).

To conclude, in this unit the dinoflagellate assemblage indicates a strong and relatively extensive period of amelioration whilst the foraminifera, with the exception of a restricted horizon in BH 81/37, suggest the continuation of harsh arctic conditions.

3.5.13 **Unit M**

This unit, the uppermost one in seismic sequence 4, is represented by the combined flora and fauna contained in units DXII and F XI. The dinoflagellate cyst assemblage is moderately sparse, although probably more productive than seen in previous harsh or glacial units.

Similarly, the foraminiferal assemblage, although often displaying a relatively poor recovery and dominance of *E. clavatum,* appears to be relatively diverse, whilst also containing a small proportion of southern species. This is especially so in BH 81/27 where there are small, but significant, numbers of southern species.

It is therefore concluded that the faunal and floral assemblages in this zone reflect an unfavourable environment, but one that is possibly less severe than observed previously. Unfortunately there is no information as to the amount of reworking of the dinoflagellate cysts and as such the above interpretation cannot be further refined.

3.5.14 **Unit N**

This unit occurs in the lower part of seismic sequence 7 and is represented by the faunal assemblages in units DXIII and FXII. The rich recovery of indigenous dinoflagellate flora in this zone suggest a favourable palaeonenvironment and a period of amelioration. *O. centrocarpum* is the dominant species indicating the influence of the North Atlantic current, although in BH 81/25 *B. tepikiense* becomes more prevalent towards the top suggesting a deterioration of conditions, or at least less saline conditions.

The foraminifera are generally inconsistent with the above interpretation and are more typical of a hyposaline arctic environment. The discrepancy between this and the dinoflagellate assemblage may again suggest the presence of a stratified water mass or different rates of response and migration.

In conclusion unit N represents an ameliorative period with environmental conditions which were favourable for dinoflagellate cysts. However, the foraminiferal assemblage suggests that the amelioration was of limited duration therefore precluding the migration and establishment of southern species or, alternatively, that the water mass was stratified and hence reflects the different ecological requirements of the benthic foraminifera.

3.5.15 **Unit O**

This unit occurs within the combined limits of units X IV and F XIII and forms the uppermost unit in seismic sequence 7.

Detailed analysis of the dinoflagellate cyst flora representing unit O in BH 77/2 has revealed a more accurate picture of the palaeoenvironment than described for previous zones (Appendix 2.10). The consistent presence of *B.tepikiense* in a moderately sparse assemblage containing between 25-50% reworked forms indicates colder than present water conditions. This is substantiated by the presence of *A. andalousiense, Spiniferites spp. elonglatus* and *frigidus* . All these species indicate water conditions colder than today, probably arctic, and a lack of influence of the North Atlantic Drift (Harland, 1985).

The foraminiferal assemblage is consistent with the dinoflagellate cysts and indicates arctic, shallow water conditions. However, in BH 77/2 the presence of a small proportion of southern species is either spurious or suggests a cold period in which full arctic conditions were not properly established.

In conclusion unit O represents conditions which were unfavourable for the dinoflagellate cyst flora, whilst in one borehole there is some evidence that may suggest a more favourable benthic environment. The reasons for this are not yet clear.

3.5.16 **Unit P**

This unit is represented by units D XV and F XIV and it occurs at the base of seismic sequence 8. The detailed analysis of samples from BH 77/2, V.E. 135 and V.E. 111 for both dinoflagellate and foraminiferal assemblages has allowed for detailed palaeoenvironmental interpretation of both this unit and the succeeding unit Q. In this unit the poor dinoflagellate cyst recovery and significant degree of reworking are indicative of unfavourable conditons. Also in V.E. 111 the dominance of *B. tepikiense* coupled with the presence of *Spiniferites frigidus* and *A. andalousiensis,* suggests a cold water environment with only limited contact with Atlantic waters (Long et al., 1986, Appendix 8). In addition the occurrence of significant proportions of round brown *Protoperidinium* cysts produced by non-photosynthesising dinoflagellates may indicate periods of sea-ice cover (Dale, 1985), or at least a close association with sea-ice.

A similar palaeoenvironment is indicated by the foraminiferal assemblage in which *E. clavatum* with subsidiary *C. reniforme* dominate, suggesting high arctic, shallow and hyposaline water.

To conclude; unit P is indicative of a cold and severe climate in an inner sub-littoral environment.

3.5.17 **Unit Q**

This, the uppermost, unit occurs at the top of seismic sequence 8 and is represented by units D XVI and F XV. The unit was recognised in the majority of boreholes and its upper boundary occurs at the seabed. As mentioned earlier detailed sampling of this unit showed the presence of three biofacies and the implications of these will be discussed individually.

In biofacies 1 the dinoflagellate fauna is dominated *O. centrocarpum* together with an increase in indigenous cyst abundance and diversity suggests a strong penetration of North Atlantic waters. This interpretation is supported by the presence of more temperate species including *Spiniferites ramosus* (Ehrenberg) Mantell, *Spiniferites lazus* Reid and *S. Elongatus*.

In contrast to the above interpretation a foraminiferal assemblage dominated by the cold water species gives little evidence for a climatic amelioration. Thus, given the dinoflagellate cyst results, it is possible that the water column was seasonally stratified or that the foraminifera were slower to respond to the changing ecological conditions.

Biofacies 2 indicates a return to more unfavourable conditions, reflected in the poor dinoflagellate recovery and higher degree of reworking. In addition the assemblage shows a significant increase in cold water species such as *A. andalousiensis* and in particular the return of round brown *Protoperidinium* cysts, thought to be indicative of sea ice cover.

The foraminiferal assemblage continues to be dominated by cold water fauna similar to biofacies 1, whilst the increased importance of *E. asklundi* and *P. orbiculare* towards the middle of the biofacies suggest falling water levels.

Biofacies 3 occurs up to the seabed and the dinoflagellate cyst assemblage compares closely with that recovered from sea bed sediments in the area (Harland, 1983) and therefore indicates the true establishment of modern environmental conditions.

Similarly in the foraminiferal benthos, the species *B. marginata, C. carinata, H. baltica* and *T. angulosa* are all temperate species and suggest deep water and fully marine salinities.

In conclusion detailed sampling of this unit has allowed for the establishment of a tripartite division. The lowermost biofacies indicates an amelioration which passes up into a colder period, biofaceis 2, whilst the upper biofacies represents the onset of present day temperate conditions. It should be noted that where the cores were not sampled in detail the middle, cold zone, was not identified.

3.6 Pleistocene Stratigraphy and Discussion

Dating of the previously described units and seismic sequences relies on five sources of evidence listed below:

a. Palaeomagnetic dating of various boreholes (Stoker et. al. 1983).

b. The recognition of age diagnostic species and assemblages of foraminifera and Dinoflagellate cysts.

c. Seismic extrapolation of sequences to previously dated localities outside the study area.

d. Detailed stratigraphical interpretation of VE 58+00/111 involving the author (Long et al., 1986, Appendix 8).

e. Detailed stratigraphical interpretation of borehole 81/26 involving the author (Sejrup et al., in press, Appendix 9).

With regards to the above, the two primary sources of dating evidence are a) and b), and these were successfully employed by Stoker et al. (1985) to establish a Pleistocene stratigraphy in the central North Sea between 56°N and 58°N. As mentioned earlier a previous stratigraphy erected by Holmes (1977) has been discredited, primarily on the basis of eronneous radio carbon dates, and this will not be discussed further here.

It is therefore the aim of this section to relate the seismic sequences and their internal bio-units to the stratigraphical framework of Stoker et. al´s (1985) and to include the results of more detailed micropalaeontologic analysis together with new information from the Fladen and Bosies Bank areas. A more detailed investigation into the correlation of this stratigraphy with the oceanic record will be undertaken in chapter 6.

Fig. 3.2 schematically depicts the eight seismic sequences and the bio-units within these sequences. It also shows the correlation of the sequences to the Pleistocene formations of Stoker et al. (1985) together with evidence as to their age compiled by the author and various other workers. The basic chronostratigraphic framework in the right hand column is based on the Dutch stratigraphic classification (Zagwijn, 1985).

The interpreted age of each individual bio-unit and seismic sequence will not be discussed here but rather the salient chronostratigraphic information will be considered. It is important to note that such information is relatively sparse throughout the record as a whole and such much of the interpretation relies on extrapolation from data points using seismic sequence boundaries and occurrences of ameliorative bio-units.

Possibly the most important evidence as to the age of Pleistocene sediments from the study area regards the identification, in a number of boreholes, of the Brunhes-Matuyama palaeomagnetic boundary (Stoker et al. 1983). This boundary occurs within seismic sequence 1 (Fig. 2.1) and it separates the upper Brunhes Normal Epoch from the lower Matuyama Reversed Epoch. Its age is currently taken at 790,000±5,000 yr BP (Johnson, 1982), although the observed polarity boundary in a borehole will not always reflect the actual Brunhes-Matuyama transition. This can be seen in BH 81/26 (Fig. 2.2) where the amino acid data suggests that a major hiatus occurs at the magnetic boundary (H. Sejrup, pers. com.).

With regards to the stratigraphic implications of the Brunhes-Matuyama boundary, it has generally been taken to mark the lower/Middle Pleistocene boundary and has been placed towards the base of the 'Cromerian Complex' (Zagwijn, 1979). This interpretation would place biozones A to F as being of between Praetiglian and Cromerian age. This is confirmed by the abundance in biozone B of the foraminifera *C. teretis* and the dinoflagellate cyst species *T. pellitum* and *O. israelinium,* all generally identified with sediments of a Lower Pleistocene age (Skinner and Gregory, 1983 and Cameron et al., 1984).

More accurate subdivision of the lower Pleistocene from the study area is difficult due to the lack of evidence and probable existence of a number of hiatuses. However, in boreholes 81/26 and 81/27 unit B is characterised by a lower *Spiniferites* dominated assemblage and an upper *T. pellitum* dominated assemblage whilst the top of the unit is marked by a normal palaeomagnetic event within the Matuyama reversal. These characterists are similar to those displayed by the *Spiniferites spp.* assemblage and overlying *T. pellitum* assemblage of the Smiths Knoll Formation (Tiglian) and Winterton Shoal Formation (Eburonian-Waalian) respectively, from the southern North Sea (Harland, 1983 Cameron et al., 1984); although *Spiniferites spp.* from the two ares are not strictly comparable (R Harland, pers. com.). In addition Cameron et al. (1984) identified a normal event, within the Matuyama reversal, just above the Winterton Shoal Formation which they assigned to the Jaramillo event ($\approx$890,000-950,000 yr BP).

Units C to F in seismic sequence 1 have therefore been assigned to a Waalian to Bavelian age, whilst unit B is predominantly of Tiglian and Eburonian age. The cold period represented by unit A is possibly of a Praetiglian age although there is no evidence for this. Furthermore, it should be stressed that these divisions are informal and open to further revision, especially with regards to the probability of hiatuses within the succession.

The uppermost unit in seismic sequence 1 is represented by G, indicative of glacial conditions, and the close proximity of this biozone to the Brunhes-Matuyama boundary (Fig. 3.2) suggests that it may be correlatable with the ´Glacial A´ subdivision (Zagwijn et al., 1971) in the Dutch Pleistocene sequence, therefore implying an early ´Cromerian Complex´ age (Stoker and Bent, 1985).

Dating of the Pleistocene succession above seismic sequence 1 relies primarily on the identification of three regional, irregular unconformities (reflectors B, D and H) which are generally attributed to low sea level stands (Fig. 3.2) and processes associated to major glacial periods. Three similar periods of regional glaciation, succeeding the Cromerian Complex, affected north-west Europe during the Elsterian, Saalian and Weichselian. Also mainland Britain was affected, to varying degrees, by three regional glaciations during the Anglican (Elsterian), Wolstonian (Saalian) and Devensian (Weichselian) stages. The lowermost of the irregular unconformities in the North Sea was therefore assigned an Elsterian age (Stoker et al., 1985; and Cameron et al., 1986) and the two succeeding irregular unconformities to the Saalian and Weichselian stages respectively. This interpretation is supported by the occurrence of one major amelioration biozone separating each of the irregular unconformities and these bio-units, H and L, are therefore assigned to the interglacial Holsteinian and Eeemian stages respectively.

Further evidence as to the Holsteinian age of bio-unit H lies in the identification of the freshwater plant *Azolla filiculoides* (Griffin, 1984), which has not been recorded in sediments younger than Holsteinian age in Britain or north-west Europe (Godwin, 1975). This also supports the placing of the underlying irregular erosion surface within the Elsterian.

Diagnostic evidence as to the proposed Eemian age for bio-unit L is not presently available. However, there is presently no data to contradict this view. For example, a detailed study of the amino acid stratigraphy in BH 81/26 (Sejrup et. al., in press; Appendix 9) has shown that sequences 2 and 3 were deposited prior to the Eemian. This is further backed up by extrapolation of the seismic network to the area of the Tartan oilfield where sequence 4 corresponds in part to Jansen & Hensey's (1981) foraminiferal zone V and pollen zone D, which they describe as being of Eemian age.

Within the seismic network, sequences 5, 6 and the upper part of 7 appear to be lateral equivalents and occur within a major glacial period. As such, seismic sequence 5 can be equated with the Wee Bankie formation of Stoker et. al. (1985), a till-like unit that has been extrapolated to the widespread late Devensian (Weichselian) till which crops out onshore to the west (Gostelow and Brown, 1981).

Seismic sequences 5, 6 and 7 are therefore interpreted as being of predominantly Late Weichselian in age. This is supported by amino acid data from unit O in BH 81/26 (Sejrup et. al, in press) correlated with radiocarbon dates from the Norwegian sector. The exception to this correlation is a discrete period within seismic sequence 7, represented by unit N, when conditions were ecologically favourable for dinoflagellate cysts. The position of this unit within the stratigraphy is uncertain and it will be discussed further in the context of sedimentary facies analysis in chapter 4.

Both the palaeoenvironment of deposition and stratigraphic age of sequence 8 has been aided by the detailed study of VE 58+00/111 (Long et. al., 1986; Appendix 8) and VE 58+00/135. With regards to the stratigraphic age the various pieces of evidence will be discussed below and are depicted fully in Fig. 3.2.

VE 58+00/111 was extracted from the Witch Ground Basin (Fig. 3.3a), where it penetrated a condensed seismic sequence 8 (Fig. 3.3b). This sequence has been termed the Witch Ground Formation by Stoker et al. (1985) and it also equates with the Witch and Fladen deposits, described by Jansen (1976).

From the seismic section it is apparent that at the sample site sequence 8 can be divided in two units, described in chapter 2 (Table 3.1). The contact between the two units appears to be gradational, occurring at between 2.5m and 3.0m below the sea bed. The underlying seismic sequence 7 was not penetrated by the vibrocorer.

Four facies were identified based on palaeontological and lithological parameters (Fig. 3.3) and their dating relies on the assumption that sedimentation was not interrupted by any major hiatuses. The uppermost facies, D , contains a faunal and floral assemblage typical of that occurring at present, described previously for unit Q. This facies therefore represents the onset of the Holocene period and is continuous up to the present day sea bed. The underlying unit, facies C, represents a cold period of moderate severity and is attributed to the Younger Dryas cooling between 10,000 and 11,000 yrs B.P. At the top of the facies the existence of glass shards, similar to those found in western Norway (Mangerud et al., 1984) and in the Norwegian Sea (Jansen et al., 1983) dated at 10,600 yrs BP fits the proposed timescale.

On this basis facies B, the lowermost division unit Q, described previously, would correlate with the distinct climatic amelioration in the North Atlantic record between 13,000 and 11,000 yrs B.P. known as the Bolling interstadial. The lower boundary of the facies correlates with the base of the upper faintly layered seismic unit and also with a distinctive change in the dinoflagellate record. It is this break which delimits the division between units P and Q. The former equates to the lowermost facies, A, and was deposited in shallow, Arctic water conditions affected by sea-ice and little or no contact with temperate North Atlantic waters. If the previous assumptions are correct, facies A was deposited during a cold period prior to 13,000 yrs B.P. and may therefore be ascribed to the late Weichselian.

On the basis of detailed sampling unit Q can therefore be divided into three horizons equivalent to facies B-D, and these range in age from the Bolling interstadial to the Holocene. Similarly unit P equates to a single unit, Facies D, indicative of glacial conditions and ascribed to the late Weichselian. Unit Q occurs within the upper faintly layered unit of seismic sequence 8 and P within the lower well layered unit.

A further vibrocore, VE 58+00/135, was sampled in detail to corroborate the dinoflagellate record in seismic sequence 8 and four individual horizons or facies were again recognised, similar to those in VE 111.

From the previous interpretation it is obvious that the Pleistocene sequence in the study area spans a much greater proportion of the stratigraphic record than originally proposed by Holmes (1977). As such, the assumption that ameliorative periods within seismic sequence 1 are of Eemian age or younger (Harland et. al. 1978; Gregory et. al. 1978) is based on erroneous radiocarbon dates and they are in fact older, as first suggested by Jansen et. al. (1979).

In the northern North Sea the identification of a similar, and relatively complete sequence of events, but lacking the Holsteinian, led Gregory and Skinner (1983) to suggest that perhaps the Eemian and Holteinian are closer in time than is generally accepted, being either two leaves of the same interglacial or even synonymous. A similar idea was proposed by Bristow and Cox (1973) and such an interpretation would certainly preclude the concept of two major ice advances in the area during the Middle Pleistocene. However the evidence presented here clearly contradicts this, and the presence during the Middle Pleistocene of two interglacial separated by a distinctive glacial period is very apparent.

A further ramification of the palaeontological studies is that they lend some understanding as to why, during the Pleistocene, there is generally a sparsity of full interglacial fauna and flora typical of today with high percentages of Planktonic and southern foraminifera. Jansen and Hensey (1981) concluded that this phenomena, suggested by the occurrence of mixed arctic and southern species assemblages, was the result of ameliorative periods developing under the overall influence of glacial rather than interglacial processes. In fact, the identification here of foraminiferal assemblages containing few or no arctic species proves that fully interglacial conditions have developed at periods throughout the Pleistocene. However, it should be stressed that firstly, such assemblages dominated by southern species were very restricted and secondly, that at some localities they are related to the base of infilled channels. It is therefore possible that temperate or ameliorative sequences could easily be missed when sampling from borehole material, or alternatively that during interglacials the sediments only reached a significant thickness in the then deep open channels. The latter can easily be related to present conditions in the study area where sediments are accumulating preferentially in the deep open channels. Alternatively, sediments containing microfauna and flora consistent with ameliorative periods may have been eroded from the inter-channel areas and preserved only in the channels themselves.

Contrary to the foraminifera, the dinoflagellate assemblages from the studies boreholes have provided extensive evidence of ameliorative episodes and proved their usefulness as a kind of ´marine pollen´ (Dale, 1985). Some understanding of the discordance between foraminiferal and dinoflagellate cyst assemblages can be gained by looking at conditions during the late glacial intertadial (13,000-11,000 yrs BP). Evidence from VE 111 and 135 (dinoflagellate cysts) suggests that temperatures during this period were very similar to today and this agrees with the conclusions of Ruddiman et. al. (1977) and Ruddiman and McIntyre (1981). However, studies of molluscan assemblages (Peacock, 1983) and benthic and planktonic foraminifera (Sejrup et al., 1980 and Jansen et al., 1983) from the Scottish coast and Norwegian sea, respectively, have failed to show the existence of a warm pulse during this period.

It is therefore tempting to suggest that the water mass was stratified during the last interstadial and in fact Peacock (1983) provides evidence for this, suggesting that the water temperature gradient was much greater than at present and that although surface waters were of a similar temperature to today the main body of water was some 2°C to 3°C cooler. Therefore during both the ameliorative period and perhaps earlier ones the presence of thermal stratification in the study area would have meant that generally the bottom water would have been too cold to allow reproduction and colonization by southern species of foraminifera. Dinoflagellate cysts, however, are likely to have been less affected by water stratification and as such reflect the presence of more temperate surface waters, even if they are a result of only a weak and diluted North Atlantic Drift entering the North sea.

In conclusion a relatively comprehensive chronostratigraphic framework has been developed for the Pleistocene sequence in the study area using palaeontological and seismic data and a variety of dating methods. This stratigraphy was originally developed by Stoker et al. (1985) for the area between 56 N and 58 N. The previously described bio and chronostratigraphic framework completes the stratigraphy for the whole central North Sea, up to 59°N, and also incorporates new data not available to Stoker et. al. (1985). Fig. 3.4 shows the proposed framework for the whole study area and incorporates the formation names proposed by Stoker et al.

The palaeoecological evidence discussed previously will also be used in the following chapters to complete facies and facies association analyses and to aid their overall interpretation.

CHAPTER FOUR

LATE WEICHSELIAN SEDIMENTARY FACIES AND THEIR INTERPRETATIONS

4.1 Introduction

The term 'facies' or 'sedimentary facies' has gradually become more ambiguous with regards to its use in geological literature. Thus, it is now used in a variety of senses, including:- strict observation (eg pebbly channel sand), genetic interpretation (eg Turbidite), and depositional environment (eg Fluviatile), (Reading, 1978). For the purpose of this study a relatively objective approach was adopted so as to both preclude hasty interpretations whilst also having the facies scheme open to any future modifications.

Both Selley (1970) and Reading (1978) list specific factors which they consider to be the primary criteria for defining individual facies. However, certain criteria, usually more relevant to unconsolidated sediments were omitted in their list and the following broader definition was preferred:- "a facies is the whole set of attributes possessed by the deposited sediment laid down in a particular environment," (Leeder, 1982). Therefore when defining individual facies, in addition to the normally accepted criteria including colour, composition, bedding, texture, fossils and sedimentary structures, it was also possible to incorporate such factors as geotechnical properties, and seismic texture and geometry. The previously described seismic sequences, seismic facies and biounits are therefore incorporated into the following sedimentary facies as a further aid to their interpretation. However, it should be stressed that, apart from a few exceptions (facies B^5, C^6 and D^8) it was not generally possible to correlate individual facies between boreholes using seismic reflectors. Moreover certain facies lent themselves to particular defining criteria better than others with the result that the following descriptions are not always well balanced in their content.

With regard to the overall facies scheme, it was decided that because the project encompassed several stratigraphic units it would be better to construct a separate scheme for each unit, whilst maintaining a degree of continuity between each scheme. Thus each facies is followed by a suffix (1-8) relating to the seismic sequence defined in chapter two. This chapter deals with those facies considered to be of late Weichselian age (sequences 5 to 8) whilst chapter 5 deals with those facies of pre-late Weichselian age (sequence 1 to 4).

In an attempt to keep the facies scheme relatively simple the succession, for each stratigraphic unit, was divided into a maximum of five facies (A-E). It was hoped that this would aid interpretation and basin analysis whilst also leaving scope for the sub-division of certain facies, where relevant.

In the following section the main features of each facies and sub-facies are described, followed by an interpretation of the possible environment of deposition. The latter is based primarily on evidence from the respective facies although, where pertinent, other relevant information is also considered including evidence from adjacent facies. Colour descriptions were made using a Munsell's color chart (1975).

Chapter 5 follows the same procedure except that facies descriptions and interpretations are only considered in detail where they differ from, or highlight, specific details described in chapter 4. At the end of chapter 5 grain size analyses from all the facies are summarised and discussed. Facies associations and depositional models are described in Chapter 6.

Figs 4.1 - 4.8 show the location, stratigraphic architecture and sedimentary facies of the boreholes and selected vibrocores, utilized in chapters 4 and 5. Sedimentary environments in the Lower Pleistocene sequence are labelled for reasons discussed in chapter 5. Where borehole recovery was not 100% lithologies were extrapolated using borehole gamma logs. This was especially the case for some sandy horizons and the uppermost 2-3m of borehole recovery. The lithofacies code, mentioned in Chapter 1, is adapted from the code of Eyles et al. (1983), the main changes being the addition of a number of prefixes denoting the general abundance of clasts and shell material in the sediment and the inclusion of various bedding structures into the code. The 2nd prefix in Eyles code, denoting a clast or matrix supported diamict, was omitted, most diamicts being of a matrix supported nature (Karrow, 1984; Dreimanis, 1984). Similarly the end prefix, 'r', indicating resedimented material was also omitted because it was considered to be a genetic term rather than an objective one. Instead a 2nd prefix, 'd', was used to include all types of deformation structures, their intrinsic properties and interpretations are then discussed further, in the text.

However, unlike the approach adopted by Eyles et al. (1983) the following interpretations are not based solely on the identification of specific lithofacies and their relationships. Indeed, Kemmis and Hallberg (1984) stress "that the genetic interpretation of glacigenic deposits (and hence, depositional environment) must be based on multiple criteria."

Table 4.1 - 4.6 summarise the main characteristics of the sedimentary facies and their depositional environments. The clay mineralogy of the sediments, briefly referred to in certain facies descriptions, is discussed in greater detail in Appendix 6. The results are not strictly pertinent to the facies interpretations and the discussion was therefore omitted from the main text. Here, it is suffice to say that the clay mineral assemblages, dominated by illite, reflects typical high latitude weathering and erosional processes. Variations in the clay mineral assemblage purely reflected changes in the source material. Similarly bulk geochemical analyses of the sediments, also discussed in appendix 6, were not generally useful to the facies scheme presented here and were therefore also omitted.

4.2 Sedimentary Facies

4.2.1 Facies A

This facies occurs in association with two separate seismic sequences, 5 and 7, and as such has been divided into facies A^5 and A^7, both thought to have been deposited at the same time and in similar environments. Facies A^5 is restricted to the western edge of the Marr Bank and Peterhead areas whilst A^7 occurs in the south-west corner of the Bosies Bank area (Fig. 4.9).

Facies A^5

On seismic records this facies forms a complete and easily

FACIES / SUB-FACIES	OCCURENCE AND SEISMIC CHARACTERISTICS	LITHOLOGY	MICROPALAEONTOLOGY	INTERPRETATION
A^5	Western edge of Marr Bank Peterhead areas. Chaotic configuration with point-source hyperbolic reflectors and hummocky surface.	Firm to stiff, massive diamict. very poorly sorted, S.D 3.2-3.8.ø Sharpe basal contact and deformation of underlying sediment.	Barren or very sparse.	Sub-glacial till, probably lodgement.
A 7	Western edge of Bosies Bank area. Chaotic configuration hummocky surface and associated with upstanding ridges	Soft to firm, massive diamict. Very poorly sorted, S.D. 2.8-3.0.ø Sharp basal contact and deformation of under-lying sediment.	——	Sub-glacial till, probably lodgement.
B7	Isolated outcrops in the Bosies Bank area. Chaotic configuration and associated with gentle slopes.	Soft, stratified diamict with angular contacts and deformation structures. Weak sub-horizontal clast imbrication Sharp basal contact.		

Tables 4.1.-4.6. Summary of the main features of the late-Weichselian sedimentary facies and their suggested environments of deposition.

C^6	Over much of Marr Bank and Peterhead areas. North-easterly dipping planar basal reflector, wedge-like geometry and various reflector configurations:-		Barren or hyposaline shallow arctic marine, cut off from NAD	Sub-aqueous outwash fan
$C1^6$	Chaotic or discontinuous sub-parallel reflector configurations	Finely laminated (1cm.), cross-ripple 1cm. and thinly interbedded sands and silts, with rare clasts. Laminae are diffuse to sharp and sequence is ungraded or reverse graded. Sharp lower and gradational upper contact.		Deposited from sediment laden under-flows emitting from a glacier front. Traction and density currents are the predominant transport mechanism.
$C2^6$	Chaotic or discontinuous sub-parallel reflector configurations.	Soft, stratified or massive diamict composed primarily of poorly sorted sand and coarse silt, and abundant clasts. Gradational basal contact and sharp upper contact.		Deposited from a combination of ice-berg rafting, episodic traction currents and sediment overflow plumes.
$C3^6$	Discontinous, planar configurations or low-angle bi-directional, downlapping	Loose, massive, pebbly sand moderate to poorly sorted, S.D. 1.2-2.4.0		Lag deposit resulting from the re-working of glacial and glaciomarine sediment

C^7	Western half of the Bosies Bank area. Structureless or chaotic configuration and sheet drape or ridge like geometries respectively.		Hyposaline shallow arctic marine cut off from the NAD	
$C1^7$				
1		Soft, interlaminated sands and muds. Laminae are well defined and decrease up in thickness. Units display sharp upper and lower contacts		Represent either the product of "burst & sweep events" (Bridges 1978) or Bouma D type sediments.
11		Rythmic couplets of soft faintly lam. sandy silt and mud with dropstones.		Cyclopel units deposited from sediment over-flow plumes.
111		Planar and wispy lam. sands finning up into well lam. muds and homogoneous muds. Sharp basal contact.		Turbidite deposit containing Bouma units B - E
1V		Massive, gravelly sand Finning up into wispy and ripple laminated sands and overlying lam. muds. Units display sharp and erosive basal contacts.		Proximal or vigorous turbidite deposit containing Bouma units A - E

$C2^7$		Soft, upward finning, massive diamict with abundant clast content which decreases towards the top of the unit. Upper contact is generally gradational.		Deposited from a combination of iceberg rafting, episodic traction currents and sediment overflow plumes.
D^7 / $D1^7$	Essentially restricted to the Witch Ground Basin located in the Fladen and Bosies Bank area and northern Forties and Peternead areas. Sheet drape geometry with chaotic or structureless configurations. Upper surface has a distinctive micro-relief.	Soft - moderately firm, massive or faintly lam. mud with dropstones. Black monosulphide bands and patches are ubiquitous, as are shell fragments. Sharp lower contact and sharp or gradational upper contact.	Hyposaline, shallow arctic marine, cut off from NAD, with at least a seasonal sea-ice cover.	Deposited by suspension from turbid sediment overflow plumes with a small component of ice rafted debris. Monosulphide bands reflect anaerobic conditions at the depositional interface. Micro-relief formed by sea-ice scouring.
D^8 / $D1^8$ - $D3^8$	Similar distribution to D7. Basin-fill geometrey with closely spaced intraformational reflectors which drape over bottom irregularities. Layering less distinct towards the very top and base of the sequence.	Very soft-soft, faintly lam. mud with dropstones. Black monosulphide bands and patches are ubiquitous. The unit displays a distinctive upward finning ($D1^8$ $D3^8$) concomitant with a decrease in ice rafted material and an increasing in the sorting.	Similar to D7	Deposited by suspension from turbid sediment overflow plumes emitting from a retreating glacier front or fronts and a large river system to the south.

E^7	Restricted to the margins of the Witch Ground Basin and isolated ridges within the Basin. Structureless reflector configuration.	Soft, interlaminated and thinly interbedded sands and muds. Lamainae are planar and sharply defined with occasional bioturbation structure. Individual sand layers are commonly normally graded. Occasionally grades into more massive or flaser bedded muds.	Varies from hyposaline shallow arctic marine cut off from the NAD to a moderately temperate, but hyposaline environment open to the NAD.	Lower intertidal deposits with storm generated sand layers. Absence of sea-ice scouring indicates palaeowater depths <10m.
E^8/E1-E4^8 1 Basin Infill 4.5.	Witch Ground Basin and isolated depressions throughout the study area. Transparent texture and coarsely layered reflectors, pockmarks and gas blanking are ubiquitous.	Upward coarsening sequence (E1^8 ↑ E4^8) of very soft-soft, lam. muds with numerous bioturbation structures, shell fragments and occasional clay balls. Individual laminae are diffuse and commonly disturbed by disturbation structures. Upward coarsening of the sequence accompanied by increase in sorting and the coarse silt component.	Dinoflagellate cysts indicate temperate marine environment open to the NAD. Foraminifera suggest the lower part of the sequence was deposited in a hyposaline, shallow arctic marine environment, and the upper part in a temperate marine environment.	Marine sediment deposited in a sub-littoral environment at the onset of a post-glacial transgression. Sediment supply derived from both the re-working of surrounding sediments and a steadily decreasing meltwater input.

11 Channel Infills	Infilling channel features mainly in the Forties and Devils Hole areas. Complex series of seismic facies including divergent, prograding and structureless configurations. Gas blanking common but pockmarks not observed.	Complex sequences of very soft -soft, bedded and lam. muds and sands. Bedding structures include planar & cross stratification, flaser bedding & massive units. Shell fragments and monosulphide patches are ubiquitous whilst bioturbation structures are less frequent than for basin infill. Base of infill sharpe, upper contacts are generally sharp.	Hyposaline, shallow arctic marine environment with an upward shallowing of water depth. Cut off the NAD.	Predominantly marine sediments deposited in a shallow, sub-littoral environment often verging on an intertidal environment. Very rapid depositional rates concomittant with a marine transgression allowed for the accumulation of this sequences of such sediments in the channel features.
111 Inter-Channel	Ubiquitous over much of the Devils Hole and Forties area and in patches over the remaining study area. Transparent seismic texture with diffuse layering.	Upward coarsening sequence of lam. muds & sandy muds, similar to basin infill facies. Sharp basal contact.	———	Sub-littoral and intertidal sediments, deposited subsequent to infilling og the majority of channel features.
E_5^8	Ubiquitous over the study area except in the central Witch Ground Basin. Blanket type geometry with transparent seismic texture.	Massive or lam. sand and silty sand, with abundant shells and shell fragments with common gravel and pebble sized clasts. Moderately sorted with unimodal grain size distribution. Sharp and erosional lower contact.	Generally temperate marine environment with water depths similar to the present day.	Palimpsest lag deposit resulting predominantly from the re-working and winnowing of sea bed sediments when sea level was still below its present level.

distinguishable seismic sequence (5) that is characterised by its chaotic texture and common point-source hyperbolic reflectors. The extent of facies A^5 is therefore delimited in the Marr Bank area (Fig. 4.9) where it can be seen to grade laterally eastwards into sediments associated with seismic sequence 6. Unfortunately further north, in the Peterhead area, the extent of facies A^5 is less clear, mainly because of the poor nature of the seismic records and the lack of borehole or vibrocore control. The facies is generally less than 15m thick, although in 72/18 (Fig. 4.3) up to 25m of facies A^5 was recovered.

Other seismic characteristics include a hummocky upper surface (Fig. 2.30), and locally the occurrence of underlying deformed Pleistocene sediments. This facies is occasionally associated with open, or partially infilled channels, cut into bedrock (Fig. 2.48). The basal contact is sharply erosion and forms a strong unconformity with the underlying strata. The upper contact is generally sharp.

Facies A^5 is composed of dark grey brown (10YR, 3/2), moderately firm to stiff, massive diamicts with rare sandy lenses. Abundant subangular to rounded, pebble sized clasts are generally matrix supported (Plate 4.1), and more rarely clast supported. About 10-15% of the clasts are striated, an indication of glacial transport. Psammites and Quartzites, probably of Scottish Highland origin, are the main components of the gravel fraction with subordinate red sandstones and acid volcanics of Devonian or Carboniferous age. The clay mineral assemblage is generally dominated by illite but with a significant proportion of kaolinite (15-40%) and subordinate chlorite and smectite (Appendix 6 Table 1). With the exception of the smectite, individual clay species are relatively crystalline (Fig. 4.11).

Particle size analysis of the bulk sediment (Fig. 4.12) shows the facies to be very poorly sorted (standard deviation, S.D., 3.2-3.8 phi, Ø) with a uniform distribution and no evidence of current activity or other sorting processes. The percentage of gravel and pebble sized material varies dramatically between boreholes (4-43%), although this appears to be partly a function of the proximity of the analyzed sample to bedrock. For example the largest proportion of gravel sized material occurs in a sample immediately overlying Permo-Triassic mudstone (BH 72/19, Fig. 4.3). Further grain size analysis data from this, and other facies, is given in appendix 4.

Facies A^5 is generally barren of all forms of microfauna and flora, although large quantities of reworked Carboniferous spores were recovered in BH 72/20.

Facies A^7

This facies is restricted to the western edge of the Bosies Bank area where it occurs close to the sea bed and, excepting a number of ridges some 10-20m thick, rarely attains thickness of greater than 3m. Unfortunately this is below the resolution of the sparker equipment whilst acoustic sources from the boomer do not appear capable of penetrating such a unit. The position of this facies within the seismo-stratigraphic framework is therefore slightly ambiguous and is based purely on two lines of evidence. First, the extrapolation of reflectors, on sparker records, into the seabed multiple. Secondly the fact that it occurs in areas of less than 100m water depth which would, undoubtably have been subject to long periods of sub-aerial exposure prior to the late Weichselian, therefore consolidating and hardening the sediment (Boulton and Paul, 1976). Given the relatively soft and fresh nature of the sediment, discussed below, it suggests that this facies is most probably of late Weichselian age.

Fig. 2.33 shows an internally chaotic ridge within sequence 7, and although this was not sampled it, and similar features, are thought to be associated with facies A^7. Reference to Figs 2.5, 2.6, show that, like seismic sequence 5, sequence 7 is also cut or fronted by deep, open or partially infilled, channels along its western edge. Furthermore, the channels here can be seen to cut through sequence 7. The basal contact of this facies appears to be erosional and where it overlies stratified sediment there is evidence of sediment deformation (Fig. 4.6).

Facies A^7 consists of reddish brown to dark grey (10 YR, 4/2; 5Y, 4/1), soft to firm, massive diamict. It is very similar in appearance to facies A^5 and again contains abundant, matrix supported, sub-angular to rounded pebbles (Plate 4.2). However, the clast composition in this facies closely reflects the local bedrock geology. For example in V.E 58-02/164 (Fig. 4.6) the clasts in this facies are predominantly red and green sandstones, whilst in V.E 58-02/139 (Fig. 4.6) grey siltstones, sandstones and chalk clasts are dominant. The red and green sandstones are either of Devonian or Permo-Triass derivation whilst the siltstones are probably Palaeogene; given the nature of the local geology (Skinner, in press). Chalk clasts were probably derived from Upper Cretaceous strata to the west of the study are (Appendix 6, Fig. 4). Obviously this variation has an effect on the mineralogical composition of facies C^7 and this is highlighted by variations in the type of smectite present in the facies. For example, all samples from facies A^7 had high smectite illite ratios (0.7-3.4, Fig. 4.11), however, quantitative analysis of the data (Appendix 6, Table 3) suggested that smectite derived from Devonian or Permo-Triassic strata behaved differently to that derived from Palaeogene strata.

Grain size analysis of the sediment (Fig. 4.12) reveals a polymodal or weakly trimodal distribution with poorly sorted trails and a relatively well sorted mode between 2 and 4 phi. This is somewhat different from the more uniform grain size distribution of facies A^5, and indeed the difference is shown by the slightly lower standard deviation values (S.D., 2.7-2.9). However, the overall gravel, sand, mud percentages from the two facies are essentially similar.

No palaeontological evidence is available for the facies.

Interpretation

Massive diamicts are probably the most ubiquitous products of many glacial and glacial marine environments and as such their interpretation should not be based solely on lithological features. Here, both facies A^5 and A^7 are thought to represent subglacially deposited tills. Indeed facies A^5 is typical of numerous documented examples of subglacial till, in its lack of stratification, textural homogeneity, uniform particle size distribution, and firm to stiff, and generally unfossilliferous nature (Goldthwait, 1971; Domack et al., 1979; Anderson et al., 1982; Wright and Anderson, 1982; and Mode et al., 1983). Facies A^7 differs from this description in being relatively soft and containing a less uniform particle size distribution; both these factors do not however preclude deposition by subglacial processes. For example, Boulton and Paul (1976) state that the degree of compaction in a till does not necessarily relate to the thickness of overlying ice, but that the effective consolidation pressure is rather related to the pressure of the ice minus the pore water pressure. Thus, where a state of high pore water pressure exists in the sub-glacial environment, the effective consolidation of the ice may be negligible or even zero; although the latter is highly unstable and short term.

The presence of a non-uniform particle size distribution in facies A^7, may purely be a reflection of the relatively monolithological composition of the sediment when compared with the degree of variation seen in facies A^5. Such a process was highlighted by Dreimanis and Vagners (1971) who showed that the modal class distribution in till was partly a reflection of the physical properties of the source rock mineralogy. Hence varied source rock types and a variety of minerals would produce the most uniform grain size distribution. Alternatively, the grain size distribution may reflect the complex processes associated with glacial transport and deposition (Haldorsen, 1981); the possibilities of this will be discussed in further detail in chapter 5.

Clasts fabrics, on a two dimensional scale, appear to be totally random, and in fact the x-radiograph shown in Plate 4.2 is almost identical to that of a lodgement till described from the Antarctic shelf in Anderson et al (1980). Further reference to Plate 4.2 also shows the unstratified nature of facies A^7 and the deformed nature of the underlying sands.

Morphological features associated with facies A^5 and A^7, namely the presence of upstanding relief features unrelated to the underlying surface, a smaller scale hummocky topography again unrelated to the underlying surface and the occurrence of large open channels cut deep into bedrock, are generally consistent with a glaciated terrain (Shaw, 1977). Further consideration of the hummocky topography and the large ridge like feature in the Bosies Bank area (Fig. 2.33) suggests that the structural trend is approximately north-south. Interpretation of these features is difficult and a variety of analogies are possible. These include low amplitude transverse De Geer moraines (R. Aario, 1977), the sub-aqueous moraines as described by Barnett & Holdsworth (1974), Landmesser et al. (1982) and Oldale (1985), the large thrust features in Dakota described by Bluemle and Clayton (1984), and push moraines in Iceland, Spitsbergen and Baffin Island (Boulton, 1986).

Given the size of the ridge in the Bosies Bank area, up to 40m thick and ca. 5 km wide, (Fig. 2.5, L15, 79/15), it is very similar to the large sub-aqueous moraines in Lake Superior (Landmesser et al, 1982) and off Cape Ann, Massachusetts (Oldale, 1985). However, in both these cases the moraine consists of a complex variety of sedimentary facies, and not solely sub-glacial till. These facies include stratified drift, slumps, outwash sands and sub-glacial till, and are interpreted as being typical of an ice contact zone. Limited vibrocore and borehole recovery from the ridge and its periphery reveal that it is in fact composed of a variety of sedimentary facies, described in subsequent sections, and including sub-glacial till (Fig. 4.10), stratified sands and diamicts and slumped units. Chaotic and steeply dipping reflector configurations within the moraine may be the result of ice-push and deformation of the pro-glacial sediments in the ice contact zone, therefore allowing for the development of a thick morainal sequence (Boulton, 1986).

Smaller scale hummocky features, both in the Bosies Bank and Marr Bank areas (Fig. 2.30) are thought to be transverse features, possibly De Geer moraines (Aario, 1977) composed predominantly of sub-glacial till.

The actual mode of formation of these features is uncertain, and there is generally some dispute as to whether they represent end moraines, and if so do they relate to a stillstand during overall retreat or to re-advance (Lowe and Walker, 1984)? Landmesser et al. (1982) suggest that sub-aqueous moraines formed during discrete interludes in the retreat of the ice margin. Boulton (1986), however, relates sub-aqueous push moraines to advance of the ice front during the winter season, although larger features may be the result of longer term readvances and a more sustained positive glacier mass balance. What is certain is that the extent of this topography, in conjunction with facies A^5 and A^7, represents the approximate limit of the last ice front.

With regards to the actual mode of deposition of these facies there are two possibilities. First, that the sediment was plastered down as a lodgement till or alternatively that it was slowly released from the base of stagnant ice by melt out processes (Boulton, 1972; Boulton and Paul, 1976). For both facies A^5 and facies A^7 lodgement type deposition is preferred, especially for the former where the thickness of sediment is too great to have been derived from the melting of a single basal zone (Domack et al. 1979). Also descriptions of melt out till from the Matanuska glacier, Alaska, by Lawson (1981) suggests that such deposits are often characterised by a well developed pebble fabric and inherited stratification.

The various clay mineral assemblages present in facies A^5 and A^7 are detrital and reflect the incorporation of material from Mesozoic, and Devonian, Mesozoic and Tertiary strata, repectively. However, the presence of significant proportions of Kaolinite (40% in BH 72/20) in certain sediments may support direct deposition by sub-glacial processes as kaolinite commonly undergoes dissolution in sea water (Monkin, 1970). Alternatively, very rapid deposition in the marine environment may also partly protect the kaolinite from dissolution.

4.2.2 **Facies B^7**

This facies was identified in only two cores, both from the Bosies Bank area, BH 81/24 (Fig. 4.1) and V.E 58-02/224 (Fig. 4.6). In both it has a thickness of at least 1m. Seismically, the facies is associated with chaotic reflector configurations within seismic sequence 7. Both are associated with gentle slopes (0.5-1.0°), BH 81/24 is located on the leading edge of a large ridge, described previously (Fig. 2.5, L15, 79/15), and V.E 224 on slope of a broad basin feature, as depicted in fig. 4.10. The base of the unit is sharp whilst the upper contact is often gradational, although it is difficult to accurately define.

The sediments consist of dark grey (5Y, 4/1), soft, stratified diamict. The stratification is generally complex with angular contacts and the presence of discrete fold features. An x-radiograph (Plate 4.3) of this facies in V.E. 224 reveals a 'swirl patterned matrix' as defined by Nardin et al. (1979). Both the clasts and shell fragments outline a weakly layered fabric and in BH 81/24 the deformed strata is highlighted by the presence of iron stained bands (Plate 4.4).

Grain size analysis of this facies, based on a single sample from V.E. 224 (Fig. 4.12) reveals a poorly sorted non log-normal distribution very similar to that of facies $C2^7$ described in the next section.

Interpretation

This facies is interpreted as a slide deposit, as defined by Nardin et al (1979b). The presence of deformation structures, weak fabric and a swirl patterned matrix are all characteristic of such facies, especially their basal or toe region. The textural nature of the sediments suggests that they are allocthonous glaciomarine or glacial diamicts.

Slumping can be initiated by a variety of processes including earthquakes, erosional oversteepening, sediment overloading, wave action, fresh water leaching and gas charging (Reading 1978, Leeder 1982, and Nardin et al. 1979). Given the geological setting of Pleistocene sediments in this area virtually all the above are plausible mechanisms although rapid sedimentation and overloading close to the ice-front is one of the most likely explanations.

Slide deposits were also identified by their seismic characteristics, as described in chapter 2. However, these are generally restricted to channel features and their stratigraphic position is generally unknown.

4.2.3 Facies C^6 and C^7

As with facies A, this facies was divided into two units on the basis of its association with two separate seismic sequences, 6 and 7. Facies C^6 is restricted to the south-west, in the Peterhead and Marr Bank areas, whilst C^7 occurs further north in the Bosies Bank area where it extends up to 58°50'N and 1°E, (Figs. 2.5, 2.6 and 4.9).

Facies C^6

This is easily recognisable as an individual seismic sequence (sequence 6) characterised by a north-easterly dipping basal reflector and wedge like geometry. Its spatial extent is depicted on Fig. 2.32 and it passes laterally westwards into Facies A^5 (sequence 5), whilst to the north and east it appears to pass into facies C^7. The thickness of the unit ranges from 15 to 25m. Seismic sequence 6 is further characterised by three distinctive reflector configurations described previously in chapter 2:- 1) sub-parallel, discontinous reflectors commonly associated with hummocky bedforms; 2) bi-directional, downlapping reflectors; 3) structureless or chaotic configurations.

Based on lithological features and seismic configurations, facies C^6 was divided into three sub-facies, $C1^6$ to $C3^6$.

Sub-facies $C1^6$ is composed of pale brown (10YR, 6/3), finely laminated, cross-ripple laminated and thinly interbedded sands and coarse silts. Its lower contact is sharp whilst the upper contact is most commonly gradational with sub-facies $C2^6$. The laminae are generally planar although convoluted units and possible load structures were also observed. In vibrocore 57-02/316 (Fig. 4.8) the sequence was characterised more by ripple cross laminated sands with mud drapes (Plate 4.5) passing up into planar convoluted laminae of muds and sands.

Individual laminae and beds are generally ungraded, although an exception to this was observed in 72/19 where facies $C1^6$ showed evidence of reverse grading, from a coarse structureless sand up into a very coarse gravelly sand (Fig. 4.3). Carbonaceous material is common throughout the sequence, to the extent that in 72/19 the sediment often has a dark grey-brown speckled appearance (Plate 4.6). Small pebbles (10-20mm), predominantly quartzite psammite, sandstone and granite, occur throughout this facies as isolated clasts.

Its lower contact is sharp whilst the upper contact is most commonly gradational with sub-facies $C2^6$. Individual units vary from 0.1-0.8m in thickness and the maximum recovered thickness of the whole facies is 8m.

Grain size analysis of the sediment (Fig. 4.12) from 72/19 reveals a nonlognormal particle distribution containing significant proportions of coarse silt and well sorted fine sand, weakly unimodal distribution and overall poor sorting (S.D. 2.1 Ø). Reversely graded sand layers were not sampled but appeared medium to coarse grained and poorly sorted.

Sub-facies $C1^6$ is most commonly associated with chaotic or discontinuous sub-parallel reflector configurations and it occurs most commonly along the western edge of the Peterhead and Marr Bank area. It was not recovered from samples collected east of 0°30'W.

Sub-facies $C2^6$ consists of grey brown (10YR, 5/2), soft, massive to stratified diamicts with an abundant to rare clast content. Clasts are subangular to rounded with a maximum dimension of between 5mm and 30mm whilst about 10% of these observed displayed striations. The predominant clast types were psammites, quartzite, sandstones, siltstones and acidic volcanics. The clay mineral assemblage, although dominated by illite, is characterised by a significant proportion of kaolinite with subordinate chlorite and degraded smectite.

A faint and diffuse stratification could be observed both visually and from x-radiographs (Plate 4.5) although more massive diamicts were also recorded from this facies.

Individual units vary from 0.2 to 1.0m thick. The bases of individual units are generally gradational; the tops are sharp.

Grain size analysis of the sediment (Fig. 4.12) reveals a non-lognormal distribution composed primarily of poorly sorted sands and coarse silts. The poorly sorted nature of the sand fraction is especially evident when compared with the more well sorted sandy silt in sub-facies $C1^6$.

Sub-facies $C2^6$ generally occurs in association with $C1^6$ and displays a similar spatial extent and seismic texture.

Sub-facies $C3^6$ is composed of dark grey brown (2.5Y, 4/2), apparently massive, pebbly sand with some shell fragments. It should be noted that the sands were generally very loose and uncompacted and any primary sedimentary structures would probably have been destroyed during sample recovery. Small pebbles are generally subrounded and consist primarily of quartzites, acid volcanics, red sandstones and psammite. The clay mineral assemblage is characterised by relatively well crystallised material (Fig. 4.11) dominated by illite, but with up to 24% kaolinite.

Individual units are 5m to 7m thick and display sharp upper and lower contacts.

Particle size analysis of the sediments (Fig. 4.12) shows a positively skewed (0.1-0.5), unimodal distribution which, although still moderately poorly sorted (S.D 1.2-2.4), appears well sorted relative to the sediments in facies $C1^6$ and $C2^6$. In sample 1 from 81/36 the sand fraction (93%) is especially well sorted and displays a near lognormal size distribution with less than 4% silt and clay.

Sub-facies $C3^6$ is generally associated with discontinuous planar reflector configurations or low angle, bi-directional, downlapping reflector configurations. It occurs towards the periphery of seismic sequence 6, to the east of sub-facies $C1^6$ and $C2^6$.

Foraminiferal analysis of facies C^6 revealed that all three sub-facies were either barren or contained scarce and very small foraminifera dominated by shallow, hyposaline, arctic marine species.

Interpretation

Units of sub-facies $C1^6$ represent deposition by traction currents as suggested by their sharp bases, planar and ripple lamination and well sorted fine sand to coarse silt fractions. Such currents are common in close proximity to tidewater glacier fronts where sediment laden streams discharge subglacially to form a density current which sweeps along the sea bed (Cheel and Rust, 1982; Powell, 1983; Mackiewicz et al. 1983; Domack, 1983; Eyles and Eyles, 1984). Both normally and reverse graded sequences are typical of deposition from underflows (Mackiewicz, et al. 1983), the latter being likened to reverse graded sediments deposited from high density turbidity currents in which the carpet containing suspended sediments collapses into the underlying traction layer (Lowe, 1981).

Because of the relative buoyancy of meltwater in the marine environment sediment laden underflows are restricted to ice proximal environments beyond which the flow either loses momentum or freezes and the remaining suspended sediments lift to form an interflow or overflow. However, the actual range and effectiveness of such underflows in the marine environment is uncertain especially as abnormally high suspended sediment concentrations (>39g l^{-1}) are required for their formation (Hoskin and Burrell, 1972). Both Powell (1983) and Mackiewicz (1984) suggest that because of this underflows are probably restricted to within 0.5 km of the ice front. This may not, however, have applied to the North Sea where micropalaeontological evidence suggests that salinities were often reduced (20-25%) as a result of large amounts of fresh water entering a semi-enclosed marine environment. Underflows would be expected to have a greater range in such settings, especially where they originate from high velocity subglacial meltwater streams capable of transporting large volumes of sediment (Church and Gilbert, 1975).

Deposition of the more poorly sorted and faintly stratified diamicts of sub-facies $C2^6$ is related to a relative reduction in bottom current activity and an increased input from ice rafted debris and suspension sedimentation. This would account for the poorly sorted nature of the clay and silt fractions which would probably have been deposited as flocculated aggregates with settling speeds many times faster than those of the constituent grains (Kranck, 1975). Also the high percentage of mud is not necessarily evidence of a low energy environment but may purely reflect very high suspended sediment concentrations allowing for the deposition of fine grained material in a relatively high energy environment (McCave, 1971). Iceberg rafting of coarse material (IRD) and continued episodic traction currents would have supplied the sand and gravel fraction resulting in an overall nonlognormal particle size distribution with only limited evidence of current sorting. Similar coarse grained and stratified diamicts are attributed by Eyles (1985) and Powell (1981, 1983) to a combination of iceberg rafting, suspension settlement from sediment plumes and episodic traction currents.

An alternative interpretation of sub-facies $C2^6$ would be that they were deposited by debris flows (Kurtz and Anderson, 1979). However, under these conditions individual diamict units would be expected to have a sharp basal contact. Instead sub-facies $C2^6$ gradationally overlies $C1^6$ and it is suggested that this represents a transition from a traction current dominated environment to a more quiescent, suspension settlement and IRD dominated environment. Such a graduation could reflect fluctuations of meltwater and sediment load as outlined by Ostrem (1975), although the fact that a maximum of only three transitions were observed suggests that seasonal variations were not responsible. More probable is that the transitions represent larger scale, but gradational, changes in environment such as minor advances and retreats of the ice front.

The unstratified, poorly sorted pebbly sands of sub-facies $C3^6$ are interpreted as the products of re-worked and intermixed diamicts and sands, originally deposited in a manner similar to that described for sub-facies $C1^6$ and $C2^6$. The loose nature, decrease of fines, large proportion of sand together with remnant clasts in a distal environment, relative to $C1^6$ and $C2^6$, are consistent with this interpretation. Similar products have been recorded from parts of the Antarctic shelf where strong bottom currents have winnowed sediments deposited by iceberg rafting and sediment plumes resulting in a residual paratill (Anderson et al., 1980; Anderson et al., 1982; Drewry and Cooper, 1981).

The mode of reworking invoked for $C3^6$ is, however, different from the Antarctic paratills which are thought to have been winnowed during deposition. Instead post depositional reworking by wave action and possibly tidal currents is thought to have been the primary mechanism, resulting in the bidirectional downlapping reflector configurations seen in seismic sequence 6.

In conclusion facies C^6 was deposited in a relatively high energy glaciomarine environment bordering on a large grounded ice sheet. Sedimentary processes included underflow traction currents, suspension settlement from overflow plumes and the deposition of IRD. The resultant sediments formed a large subaqueous fan, similar to that described by Rust and Cheel (1973) but on a larger scale. The wedge shaped geometry and lobe shaped plan of facies C^6 are consistent with the interpretation. The marginal areas of this form were subsequently subjected to reworking by wave and tidal processes suggesting shallower water depths over these areas. The ramifications of this interpretation, especially the decreasing water depths away from the ice margin and the actual nature of the ice front will be discussed in detail in Chapter 6.

Facies C^7

This is not recognisable as an individual seismic facies but rather it is associated with both seismic facies types identified in seismic sequence 7. The first, being characterised by a sheet drape form and structureless texture and the second by chaotic reflectors, a hummocky surface and upstanding relief. It should be stressed, however, that facies C^7 does not compose the whole thickness of seismic sequence 7.

Based primarily on lithological features facies C^7 was again divided into two sub-facies essentially similar to $C1^6$ and $C2^6$; an equivalent of $C3^6$ was not observed.

Sub-facies $C1^7$ is composed of grey to dark grey, (10YR, 4/2 to 5Y, 4/1), very soft to soft, thinly interbedded and laminated fine sands and muds with common dropstones (Plate 4.7). The ratio of clay to silt was generally much greater than for $C1^6$.

The thickness and nature of individual laminae and beds varies considerably and some understanding of the primary sedimentary structures is best afforded by their classification into four interbedded units (Fig. 4.13):-

i. Thin units, 2-6cm thick, with sharp upper and lower contacts which contain interlaminated muds and sands. Individual laminae are well defined and display sharp contacts, laminae thickness decreases upwards.

ii. Rythmic couplets of faintly laminated sandy silt and mud with dropstones which clearly deform the underlying laminations (Plate 4.8). Individual couplets vary from 0.5 - 2cm thick, and such units are by far the most common in this facies. The lower, coarser layer of the couplet is characterised by a sharp basal contact and a gradational upper contact with the mud layer.

iii. Upward fining units of faintly, planar and wispy laminated sands passing up into well laminated muds and homogeneous muds. Individual units are sharp based and 3-10 cm thick.

iv. Upward fining units of massive, gravelly sand passing up through wispy and ripple laminated sands and then into laminated silts and muds. Individual units are 6 to 12 cm thick and display sharply erosive bases. No clast deformation structures were observed in either type iii or type iv units. Where clasts were present, notably in unit ii, they consisted predominantly of sandstone with subordinate chalk, siltstone and quartzite.

Most of the above structures were only identifiable from x-radiographs (Plate 4.7) and were visually observed only as thinly interbedded and interlaminated sands and muds. Where sampled the base of sub-facies $C1^7$ is sharp and in places erosive, whilst the upper contact is generally gradational with facies D^8. In cores 58-02/164 and 58-02/139, sub-facies $C1^7$ directly overlies the subglacial till of facies A^7.

Grain size analysis of individual layers in couplet units (Fig. 4.12) reveals a nonlognormal distribution very similar to that of sub-facies $C2^6$. The sediment is generally poorly sorted (S.D 3.0-3.3ϕ) although the bimodal distribution is evidence of some degree of current activity.

Sub-facies $C1^7$ occurs in conjunction with a sheet drape seismic facies with internally structureless configurations. However, other sub-facies are also associated with this pattern and it is generally not possible to trace its extent from seismic features. Thus, knowledge of its spatial distribution is based on its occurrence in cores and as such is rather subjective.

Sub-facies $C2^7$ consists of dark grey, (10YR, 4/2), very soft to soft, upward fining structureless or, more rarely, stratified diamicts (Plates 4.2-4.3). The clast component decreases in abundance towards the top of the facies and includes angular to sub-rounded red and green sandstone, siltstone, chalk, dolerite and quartzite pebbles, about 50% of which displayed striations. A weak vertical clast fabric was observed in some sequences (Plate 4.2). Shell fragments are common throughout the sequence.

The clay mineralogy of this sub-facies contains significant proportions of well crystallised smectite which commonly decreases towards the top of the sub-facies (Appendix 6 and Table 4). Conversely illite is relatively poorly crystallised and increases in significance towards the top of the sub-facies.

Individual units occur up to about 6m thick although the basal relationship of this facies was not sampled. The upper contact is generally gradational with facies D^8.

Grain size analysis of sediments from this sub-facies in BH 82/15 (Fig. 4.12) clearly shows the upward fining nature of the sequence. It is also apparent that the particle size distribution is very similar to that recorded from sub-facies $C2^6$ and the couplets in sub-facies $C2^7$, and implies that there was at least some degree of sorting during deposition.

Foraminifera in this sub-facies are indicative of a hyposaline, shallow water (<50m), arctic environment. The dinoflagellate assemblage is consistent with this interpretation and reflects a harsh environment cut off from the North Atlantic Drift (bio-unit 0).

Geotechnical properties of both this sub-facies and sub-facies $C1^6$ are based only on shipboard hardware measurements which indicated shear strengths of between 10 and 15 KN/m^2 for both sub-facies.

Because of the poor seismic resolution within sequence 7 this sub-facies is again not identifiable as an individual unit. It does, however, appear to be generally associated with the transition from a sheet drape seismic facies to the more upstanding seismic facies characterised by its hummocky upper surface.

Interpretation

Two mechanisms of deposition are invoked to explain the complex sequence of thinly interbedded and interlaminated sands and muds of sub-facies $C1^7$. Overflows, similar to those described by Powell (1983) and Elverhoi et al. (1983), are thought to have been responsible for the deposition of unit type ii rythmic couplets. Each gradational transition between sandy silt and mud reflects suspension settlement from a waning turbid overflow plume, whilst each couplet represents a fluctuation in meltwater discharge. Such overflows have a tremendous potential for the distal transport of fine grained sediment and in Kongsfjorden, Spitsbergen, the current velocity of the surface plume may exceed 50cm/sec (Elverhoi et al., 1980) although Elverhoi et al. also suggest that layered sediments deposited from overflows are confined to areas within 500m of the ice front.

More distal transport by overflow plumes is possible where maximum sediment loads, which occur 2 to 3 hours after peak discharge (Ostrem, 1975), coincide with the ebb tide. This has the affect of limiting flocculation induced settling, due to shear-induced break-up of the flocs concomitant with the trapping of low-salinity water in the floc structures (Kranck, 1975). As such the flocculation front can advance up to 10 km from the ice front (Mackiewickz et al., 1983) which consequently transports coarser grained material into a more distal environment. However, during the flood tide the opposite affect is achieved and the sediment plume and flocculation front are held near to the glacier front. In this situation fine grained silt and clay particles rapidly flocculate and settle out of suspension in an ice proximal environment.

Deposition of couplets of sandy silt and mud are therefore easily explained by suspension settlement from overflow sediment plumes whilst their rhythmic occurrence is attributable to various combinations of peak and low meltwater discharge and flood and ebb tides. The occurrence of clast deformation or dropstone structures and a poorly sorted sand fraction indicates that deposition was also occurring by iceberg rafting. Similar sediments in Muir Inlet, Alaska were termed cyclopels by Mackiewickz et al. (1983).

Turbidity currents in a glaciomarine environment are thought to have produced the remaining units in sub-facies Cl^7. The sharp contacts and upward fining and thinning lamination of unit I are typical of those ascribed by Bridge (1978) to "burst and sweep" events originating from a turbid boundary layer. Alternatively Unit 1 may represent individual occurrence of Bouma D type sediments similar to the clearly defined mud and silt laminae described by Hill (1985) and Kranck (1985), suggesting that the turbidites were relatively distal from source.

Units iii and iv afford more typical evidence of deposition from turbidity currents. In Unit iv the erosive base and upward gradation from a massive, poorly sorted sand, through ripple laminated sands and up into planar laminated silts and muds, are typical of deposition by turbidity currents. The lack of a lower massive sand layer and the presence of an upper layer of homogenous mud in unit iii is consistent with a more distal or less vigorous turbidity current.

Turbidity currents are thought to occur in a variety of glaciomarine environments, although they have been most frequently reported in fjords (Hoskin and Burrell, 1972; Powell, 1983 and Mackiewickz et al., 1983), where they can be triggered by a varietyof methods including sediment slumps, ice push, iceberg calving, underflows and wave action.

An alternative interpretation of units iii and iv is that they are storm deposits similar to those described by Kreisa (1981). However, the absence of a shelly lag and escape traces and the sharp tops of units iii and iv probably precludes this.

Sub-facies $C2^7$ is thought to have been deposited in a moderate-high energy glaciomarine environment similar to that described for sub-facies $C2^6$. Water depths were probably around 50m or less, as indicated by the microfauna whilst sedimentation occurred from a variety of sources including suspension settlement from overflow plumes, iceberg rafting and episodic traction currents. The occurrence of a vertically orientated clasts in some sequences is possibly a result of their deposition from debris laden icebergs (Domack et al., 1979). Interestingly the particle size distributions of sediments in this facies are very similar to the coarser layers which form unit type ii couplets in sub-facies $C1^7$, perhaps suggesting similar mechanisms of deposition. The more massive nature of sub-facies $C2^7$, and the general absence of couplets, probably reflects very high suspended sediment concentrations concomitant with pronounced iceberg rafting, therefore allowing thick sequences of poorly stratified diamict to accumulate containing only limited evidence of sorting.

The upward decrease in smectite observed in this sub-facies reflects a decreasing source of material from Palaeogene strata, although the exact cause of this and its implications regarding ice-front retreat are not known.

In conclusion facies C^7 represents deposition in both high energy and intermediate energy environments. In the former diamicts and erosive based turbidites were deposited, whilst in the latter rhythmic couplets or cyclopels and distal turbidites were deposited. Such sediments may have formed thin, relative to facies C^6, sub-aqueous fans, although this is not certain.

4.2.4 Facies D^7 and D^8

Facies D is divided into two separate facies units D^7 and D^8, which are related to seismic sequence 7 and 8 respectively. Both facies are essentially restricted to the Fladen and Bosies Bank area, and especially the Witch Ground Basin (Fig. 4.9). From seismic evidence facies D^8 is seen to overlie D^7 although the nature of the separating boundary varies from conformable to sharply unconformable. Because of their similar nature a comprehensive interpretation of both facies will be given after the description of facies D^8.

Facies D^7

Facies D^7 is associated with a sheet drape type seismic facies bounded at the base by a low amplitude, discontinuous reflector. In the central Witch Ground Basin it reaches a thickness of at least 12m, whilst towards the periphery of the basin it thins rapidly to a thickness of between 0.5m and 4.0m. The upper surface of this facies commonly displays a distinctive micro-relief, attributed in chapter 2 to sea ice scouring. This surface flattens towards the very centre and periphery of the basin, where facies D^7 lies either conformably below or, on the periphery, passes laterally into facies D^8.

Sampled evidence of facies D^7 is restricted, with a few exceptions, to material recovered from boreholes 77/2 and 75/33 (Fig. 4.2). The lower boundary of facies D^7 is generally sharp whilst the upper boundary varies from a sharply erosive to a diffuse and gradational one. It consists of very dark grey (2.5Y, 3/10), soft to moderately firm, massive or faintly layered mud with rare dropstones (Fmd, Fld). Black monosulphide bands and patches are common throughout the sequence as are shell fragments.

The only visual evidence of bedding structures is afforded by the presence of black monosulphide layers. These vary in thickness from 0.5cm to 4.0 cm and display diffuse and probably gradational upper and lower contacts. When exposed to air the monosulphides rapidly oxidise and disappear (20-30 minutes) although they can still generally be detected as faint structures on X-radiographs.

The lower boundary of facies D^7 is generally sharp whilst the upper boundary varies from a sharply erosive nature to a diffuse and gradational one.

Grain size analysis of facies D^7 (Fig. 4.12) shows the nonlognormal distribution, poor sorting (S.D 2.4-2.7) and weakly bimodal nature of the sediment. A plot of eleven analyses from this facies (Fig. 4.12) shows only a minimum variance (narrow envelope) in the fine sand and mud fraction, but a much greater variation in the coarse sand and gravel fraction (wider envelope). Reference to Fig. 4.12 shows the similar shape of the distribution curves for sub-facies $C2^6$ and $C2^7$ and this facies. The main difference being the much lower percentage of sand (15-20%) in this facies.

Foraminiferal evidence is indicative of a hyposaline, shallow arctic marine environment, whilst dinoflagellate flora suggest harsh conditions, cut off from the North Atlantic drift (bio-unit 0). The presence of round brown *Protoperidinium* cysts may indicate periods of sea ice cover (Dale, 1985).

Geotechnical tests (Figs. 2.50 & 2.52) on material from this facies indicate relatively low shear strength values (23-30 KN/m^2) and high plasticity indices (PI 29-33). Consolidation ratios of between 0.8 and 1.0 are indicative of a slightly underconsolidated or normally consolidated sediment.

Facies D^8

Facies D^8 forms a distinctive seismic facies, the lower basin fill unit of seismic sequence 8, characterised by closely spaced, high amplitude reflectors which tend to be draped over basal irregularities. Further salient seismic features of this facies are described in detail in Chapter 2 and it is suffice to say here that other characteristics include widespread gas blanking, a transparent unit with diffuse layering which thickens towards the basin edges, and, in places, a dense band of concentric reflectors some 3m to 5m thick. The thickness of facies D^8 varies from a maximum of 40m in the centre of the Witch Ground Basin to less than 0.5m around the edges. Where the facies thins over highs the layering becomes perceptibly less distinct and more discontinuous, (Figs 2.37-2.38).

In the central basin the lower contact of this facies is often gradational with facies D^7 and separation of the two is usually based on seismic texture and geotechnical properties. Around the basin edges facies D^8 commonly overlies much older pre-late Weichselian sediments, the contact between the two being sharp and clearly defined. The upper contact of facies D^8 varies from a gradational transition to a sharply erosive boundary.

Reference to Fig. 2.34 shows that the distribution and thickness of this facies is closely related to the present day bathymetry and in the Fladen area the facies is not present above 140m (below O.D.) whilst in the Bosies Bank area it is not present above 110m (below O.D.).

Facies D^8 was recovered in a large number of vibrocores and boreholes where it was seen to consist of dark grey, (5Y, 4/1), very soft to soft, faintly laminated, or occasionally massive, mud with rare dropstones (Plates 4.3 and 4.7). Lenses and bands of black monosulphides define a crude and diffuse layering with gradational contacts. The layering generally oxidises on exposure to air and the core often gives off a strong smell of H_2S. Small shell fragments are common throughout this sequence whilst wood fragments and dropstones of various lithologies occur less frequently.

Detailed sampling of the monosulphide bands and patches in a metre length of core (84.12) was undertaken in an attempt to quantify variations in the organic carbon content and the abundance of foraminifera. Fig. 4.16 outlines the results of this study and shows there was no perceptible increase of the organic content in the monosulphide layers relative to the surrounding sediment, and in fact the highest value (1.12%) was recorded from sediment of a dark grey colour (5Y, 4/1).

X-radiographs of material from this facies (Plate 4.3) reveal a lack of bioturbation and a very faint lamination. Gravel and small pebble sized clasts become more frequent towards the base of the facies, as seen in boreholes 77/2 and 75/33 (Fig. 4.2), and this is generally concurrent with a more well developed layering.

Clay mineral assemblages from this facies show little variation and are consistently dominated by illite (54-64%) with subordinate chlorite and kaolinite and generally less than 16%

smectite (Fig. 4.15). Primarily on the basis of grain size analyses (Fig. 4.12) facies D^8 was divided into three sub-facies, the lowest being sub-facies $D1^8$ and the uppermost one sub-facies $D3^8$. Contacts between the three sub-facies are apparently gradational although they can be equated to contrasting seismic textures, discussed below.

Sub-facies $D1^8$ is distinguished by its poor sorting (S.D 2.4-2.6), especially of the coarse sand fraction, the presence of between 15-20% sand and a small, but significant gravel fraction (0.3-10%). The particle size distribution curve is nonlognormal and very similar to facies D^7. As such facies D^7 and sub-facies $D1^8$ are lithologically identical.

The basal contact of this sub-facies is most commonly conformable or irregular and unconformable, whilst its thickness ranges from 1-6m.

Shear strength values for this facies (Figs. 2.50 & 2.52) range from 9 to 21 KN/m^2 and plasticity indices vary between 32 and 36. The sediments are normally or slightly overconsolidated (O.C.R. 1.0-1.2).

Fig. 4.17 shows the relationship of this, and the overlying sub-facies, to the seismic texture and $D1^8$ can clearly be seen to relate to a basal semi-transparent layer with diffuse layering.

Sub-facies $D2^8$ is characterised by a distribution curve similar to that of sub-facies $D1^8$, the primary difference being a marked decrease in the sand and gravel fraction and a slight improvement in the degree of sorting (S.D. 1.6-2.2). A small gravel fraction is still present (0.25%) whilst the sand fraction is slightly more sorted than in sub-facies $D1^8$. Both upper and lower contacts are generally gradational whilst its thickness varies between 0.5m and 6.0m.

Geotechnical tests on material from this sub-facies revealed a low shear strength (7-16 KN/m^2), high plasticity indices (P.I 35-43) and overconsolidation ratios between 0.9 and 1.45. The latter suggests that the sediment is slightly under to overconsolidated.

Sub-facies $D2^8$ is commonly associated with a band of dense acoustic layering described in chapter 2 (Figs. 2.39 & 4.17).

Sub-facies $D3^8$ is characterised by a much better degree of sorting (S.D 1.7-1.8), relative to the underlying sub-facies, and a significant decrease in the sand fraction (1.8-3.5%). No gravel was recorded from this sub-facies. The particle size distribution curve (Fig. 4.12) is straighter relative to sub-facies $D1^8$ and $D2^8$, reflecting the greater degree of sorting in this sub-facies.

The lower contact of this sub-facies appears gradational whilst the upper contact is generally conformable in the basin centre and erosional around the margins. Its thickness ranges from 0.5m to 7m.

Shear strengths of between 5 and 13 KN/m^2 were recorded from this sub-facies. The plasticity indices range from 33 to 43 and the overconsolidation ratio varies between 0.9 and 1.0 suggesting that the sediment is slightly under or normally consolidated.

X-radiographs of the three sub-facies showed that sub-facies $D1^8$ and $D2^8$ are characterised by a faint and diffuse non-cyclic layering with individual layers varying from 1 to 3cm. Sub-facies $D3^8$ is more commonly totally structureless. The amount of gravel and pebble sized clasts decreases markedly up the sequence and only very rare clasts occur in sub-facies $D3^8$.

Micropalaeontological analysis of the three sub-facies suggests that a similar faunal and flora environment persisted throughout the deposition of all three units (bio-unit P). Foraminifera are indicative of hyposaline, shallow arctic marine conditions whilst the dinoflagellate flora are consistent with a harsh, cold water environment, cut off from the North Atlantic. The continued presence of round brown *Protoperidium* cysts suggests at least a seasonal presence of sea ice.

Spatial variations in the thickness of the three individual sub-facies is difficult to assess. However, it is apparent that sub-facies $D3^8$ is generally the thicker of three sub-facies, reaching a maximum recorded thickness of 7m in the central Witch Ground Basin. Sub-facies $D1^8$ and $D2^8$, although again thickest in the centre of the basin (3m to 4m each), became predominant over sub-facies $D3^8$ along the western edges of the basin.

Interpretation

Both facies D^7 and D^8 are indicative of rapid deposition from suspension is a relatively shallow water environment; probably in a distal location relative to the main ice front. Evidence of ice rafting decreases up the sequence, although icebergs were not necessarily the sole mechanism of transport.

This interpretation is consistent with the basally concordant and well layered reflector configuration associated with facies D^8, the poorly sorted muddy nature of the sediments, the lack of traction current structures and their distal occurrence relative to the subglacial tills of facies A^5 and A^7. As such, the sediments are thought to have originated from turbid meltwater overflows emanating not only from the ice margin to the west but possibly from a large Scandinavian ice sheet to the north-east and a large river system to the south; these points will be discussed in greater detail in Chapter 6.

Away from the meltwater source suspension settlement from the overflows would probably have occurred rapidly due to flocculation, a reflection more on the high sediment concentrations than the salinity of the environment (Kranck, 1975).

Similar mechanisms of deposition have been reported by Elverhoi et al. (1980, 1983) from sediment plumes in Kongsforden, Spitsbergen, resulting in the deposition of reportedly homogenous mud, although no x-radiograph evidence is cited, whose main characteristic was the presence of black monosulphide bands. Interestingly Elverhoi et al (1980) attributed these bands to spring algal blooms which settled out, before the peak early summer clastic input, forming a layer of relatively high organic content which was converted to iron sulphide form by the reaction of H_2S and Fe^{2+}. Monosulphide bands in Kongsforden were therefore linked to high organic contents.

However, in the light of the organic carbon results from 84/12 (Fig. 4.16) Elverhoi et al's mechanism (1980) cannot readily be used to explain the relatively low organic content of black monosulphide bands in facies D^7 and D^8. A more suitable mechanism was proposed by Stevens (1985) who suggested the low organic carbon content in monosulphide bands reflected their rapid deposition or the presence of anaerobic conditions, therefore allowing the preservation of a greater amount of metabolizable organic matter even if the original organic content was less. Similarly the surrounding sediments may have accumulated more slowly therefore allowing aerobic processes to continue during sedimentation, resulting in a low availability of metabolizable organic matter. As monosulphide bands generally reflect a diagenetic process between Fe^{2+} and H_2S a low metabolizable organic matter content would limit any bacterial reduction and hence any post-depositional iron reduction.

If Steven's mechanism (1985) is used here then the black monosulphide bands may represent either anaerobic conditions, perhaps related to stagnation beneath seasonal sea ice, or alternatively they reflect periods of high clastic sedimentation rates related to peak meltwater discharges. The lack of perceptible coarser material in the monosulphide bands suggests that the former interpretation may be more correct, although this is by no means certain.

The increasingly diffuse nature of the layering up the sequence is related to a retreating ice front concomitant with a changing depositional environment. Thus, the layering in sub-facies $D1^8$ is formed by a similar mechanism to that proposed for cyclopel units in sub-facies $C2^7$. The more diffuse, and non-cyclic nature of the layering in $D1^8$ reflects the distal nature of the environment and the increased effects of flocculation as the brachish overflow plume is intermixed into the underlying saline layers. Similarly the eventual upward gradation into homogenous muds, sub-facies $D3^8$, indicates an absence of water stratification and the establishment of an arctic marine rather than a glaciomarine environment. This is the result of surface mixing and the absence of glacially induced processes as reflected by the apparent absence of ice rafted debris.

Given that facies D^7 is attributed to a similar mode of deposition as D^8 there is generally a significant contrast in the seismic textures associated with the two facies. The relatively structureless or chaotic configuration associated with facies D^7, in contrast to the well layered configuration associated with facies D^8 (Fig. 4.17), may be either a reflection of sedimentary processes or alternatively post-depositional reworking. Reference to Figs. 2.50 and 2.52 show that there is no significant increase in geotechnical properties across the boundary separating facies D^7 and D^8 only a gradual increase in shear strength and bulk density and a decrease in moisture content, prolonged subaerial exposure of facies D^7 is therefore discounted.

Alternatively the structureless seismic texture may reflect less stable conditions of sedimentation, relative to facies D^8, as discussed in chapter 2. This is consistent with the coarser, and more poorly sorted nature of facies D^7 relative to sub-facies $D2^8$ and $D3^8$. In addition the divergent reflector configuration associated with facies D^8 indicates either gradual subsidence of the area or a rise in sea level. Thus sub-facies $D2^8$ and $D3^8$ were deposited by suspension settlement resulting in a strong acoustic layering. Sub-facies $D1^8$ and D^7 were, however, deposited in shallower and less stable glaciomarine environments, still dominated by settlement from suspension but with a marked variability in deposition rates and perhaps a minor component of bottom current activity. The cut off depth of facies D^7 (Fig. 4.9) supports the view that it was deposited in shallower water relative to sub-facies $D2^8$ and $D3^8$.

Sea ice scouring may also have contributed to the seismic texture associated with facies D^7 and to the poorly defined, and often disrupted, acoustic layering associated with sub-facies D^8. Evidence for ice scouring is provided by the irregular micro relief (Fig. 4.14), described in chapter 2, of the upper boundary of facies D7 and also by the presence of non-photosynthesising round brown *Protoperidinium* cysts in both facies D^7 and D^8. It should be noted, however, that the irregular relief may also be partly due to strudel type meltwater scours formed in a manner similar to that described by Barnes and Reimnitz (1985).

The precise mode of scouring is attributed by Stoker and Long (1984) to sea ice due to the fact that the area of scouring, in the central Witch Ground Basin, extends from 135-160m below present sea level (OD), but cuts out above and below these depths (Fig. 2.34). Barnes and Reimnitz (1974, 1985) have described a similar depth related area of sea ice scouring from the Beaufort sea, but occurring between a water depth of 20m and Ca. 50 below sea level

(OD). Above 20m the environment is dominated by tidal currents and waves whilst depths of greater than 50m are below the maximum depth of the ice keels. It is therefore obvious that if one assumes a palaeosea level of 120m below the present, the cut out depths from the Witch Ground Basin are in close agreement with those from the Beaufort sea. This argument is further supported by the fact that large tabular icebergs would probably have been precluded from the Witch Ground Basin due to the shallow water depths. A relative sea level lowering of 120m during this stage is also in agreement with the occurrence of shallow marine sediments, facies E^7, as will be discussed in the next section.

Subsequent to the formation of the ice scoured surface the prevalent water depths during the deposition of the bulk of facies D^8 is a matter of some uncertainty. Foraminiferal evidence suggests that water depths were as shallow as 20m and probably less than 50m. However, given the present day bathymetry of the Witch Ground Basin (Fig. 1) and the absence of facies D^8 above 140-145m (below O.D.) in the central basin (Fig. 2.34), a maximum palaeowater depth of at least 80m is preferred. This conclusion is supported by Jansen (1979) who, using macrofaunal evidence, suggested a palaeowater depth of between 40m and 90m during deposition of the Fladen Deposits (equivalent to the lower basin infill unit of seismic sequence 8). The greater depth would also explain the lack of evident sea ice scouring over much of the basin during the deposition of much of facies D^8, despite the presence of a flora compatible with sea ice conditions. However, it should be stressed that such speculations ignore the probabilities of basin subsidence and marginal uplift, partly in response to glacio-isostatic affects. These factors will be discussed in greater detail in chapter 6.

In conclusion facies D^7 and D^8 form an upward fining sequence of soft glaciomarine muds, with scattered dropstones towards the base of the sequence. They were deposited predominantly from suspension in a shallow water (< 60m) hyposaline basin and record a gradual upward decrease in glacial influence concomitant with a change in the associated seismic texture.

Sea ice, of at least a seasonal nature, covered much of the basin and, subsequent to the deposition of facies D^7, the shallow water depths resulted in the formation of an extensive scoured surface. Much of this surface was then buried by facies D^8.

Similar sequences, both acoustically and lithologically, have been reported from a variety of locations outside the study area including the Norwegian trench (Hovland et al.,, 1984 and Green et al., 1985), the Shetland Basin (D. Cockroft, pers. com. 1986), and off the west coast of Scotland (Boulton et al., 1981). In all cases they were identified as glaciomarine muds.

4.2.5 Facies E^7

This occurs essentially around the periphery of the Witch Ground Basin, in the Bosies Bank, Fladen and Forties areas. It is associated with a structureless seismic configuration and, in places, a slightly irregular sea bed surface (Fig. 4.18a). Its sampled thickness ranges from 0.5 - 3.0m, the bounding lower surface not being seismically identifiable. The base of this facies was rarely sampled although where it was the contact was sharp. The upper contact is also sharp.

The sediments are composed of olive grey to dark grey (5Y, 5/2; 5Y, 4/1), soft interlaminated muds and sands with occasional, thick, structureless mud units. Individual laminae are sharply defined and often form an almost rhythmically layered sequence (Plate 4.9), the sand layers are commonly normally graded (Plate 4.10).

Shells and shell fragments are abundant to rare and similarly bioturbation structures are common to absent, as shown in the x-radiographs (Plate 4.9). More massive beds often break with a distinctive blocky texture on a scale of 1-4mm, as shown in Plate 4.11.

The layered sequences are further characterised by the fact that they are restricted to water depths between 135m and 140m. Such sequences appear to be the lateral equivalent of facies D^7 and, given the previously suggested relative sea level of -120m, they would have been deposited in palaeowater depths of 15-20m.

There is no micropalaeontological evidence from the layered sequences. However, dinoflagellate cyst assemblages from some of the more massive units in certain boreholes (Fig. 3.1) indicate that the environment of deposition was at least partially open to the North Atlantic Drift (bio-unit N). Where available, foraminiferal evidence does not indicate a strong amelioration, although there is some suggestion of a minor improvement in environmental conditions.

Geotechnical characteristics of this facies in BH 77/2 (Fig. 2.50) include a shear strength between 24-41 Kn/m^2 and plasticity indices between 34 and 36.

Interpretation

The planar laminated sands and muds and the absence of ripples and wavey bedding are consistent with deposition in an upper intertidal environment (Van den Berg, 1981; Yeo and Risk, 1981). Furthermore, its restriction to the flanks of the basin (Fig. 4.10) and the absence of an ice scoured surface is attributed to its deposition in palaeowater depths of less than 20m. A similar phenomena is described from the Beaufort Sea where, at water depths of less than 20m, wave and tidal current processes become predominant over ice scour mechanisms resulting in the deposition of unbioturbated planar laminated sands and muds (Kovacs, 1972 and Barnes & Reimnitz, 1974,1985). Therefore although no micropalaeontological evidence exists for the layered sequences, it is suspected that it was deposited in a very shallow, hyposaline arctic environment and as such forms the shallow lateral equivalent to facies D^7. It should be noted, however, that such sedimentary sequences can also be formed by storm processes at water depths of up to 40m and distances of 30km from the coast (Reineck & Singh), 1972). The lateral relationships of the facies are therefore important in assigning it to an intertidal environment.

Interpretation of the massive units is more difficult especially given the micropalaeontological evidence. However, the distinctive blocky texture often seen in these sequences is thought to have been formed by segregated ground ice and is very similar to examples described by Derbyshire et al. (1985). This would probably have entailed the sub-aerial expose of such units to periglacial processes rather than a freezing of the sea bed (Derbysire et al., 1985).

It is therefore suggested that massive units of facies E^7 were generally deposited in a shallow, open marine environment where environmental conditions were still generally unfavourable for foraminifera. This probably took place prior to or during the advance of the main ice-front into the study area. Subsequent lowering of sea level, as explained earlier would have exposed much of the sea-bed to peri-glacial processes. Similarly, a lowering of the sea-level allowed for sea-ice to scour the sea bed in the central Witch Ground Basin, whilst around the basin peripheries the very shallow water depths precluded sea-ice as a significant process. In the latter environment wave and tidal processes were dominant, partially reworking the underlying sediment and producing a sequence of well layered sediments. On the basis that the depositional environment was probably cut off from the North Atlantic Drift at this time, it is suggested that the tidal regime described in chapter 1 would have been heavily modified and that wave driven processes may have been the dominant agent.

4.2.5 Facies E^8

Facies E^8 embraces a variety of lithologies and palaeontological assemblages, which are, however, all related by their occurrence within seismic sequence 8 and by their lateral and vertical relationships.

It occurs in both the upper basin fill seismic facies, and is characterised by its transparent texture and coarse acoustic layering, and in the complex channel fill and interchannel seismic facies described in chapter two. Where facies E^8 forms a channel infill sequence it can reach a thickness of up to 180m, which thins rapidly over interchannel areas to between 0-2m and 10m. In the Witch Ground Basin, however, the facies is rarely thicker than 5m and most commonly ranges between 0-1m and 3m in thickness.

Facies E^8 is divided into an upper sand rich unit termed sub-facies $E5^8$ and a lower mud rich unit which was further divided into four sub-facies $E1^8$ to $E4^8$. The latter subdivisions being based primarily on particle size analysis.

Sub-facies $E1^8$ to $E4^8$ consist of grey green (5GY, 5/1) to dark grey (5Y, 3/1), very soft to soft, laminated or thinly bedded muds and muddy sands. Less common bedding structures include ripple cross lamination and lenticular and flaser bedding. Occasional massive units were also observed.

Further lithological description of sub-facies $E1^8$ to $E4^8$ is best aided by the recognition of three physical settings:- basin infill, channel infills and interchannel areas.

i. Basin Infill:- where facies E^8 occurs within the Witch Ground Basin (Fig. 4.10) it is characterised by an upward coarsening sequence of laminated muds with numerous shell fragments. Individual laminae are faint with diffuse contacts and are best observed on x-radiographs. Clasts are generally absent although clay balls with a maximum dimension of between 2 and 3cm occur towards the base of the facies.

Bioturbation structures are common throughout the sequence, but are especially abundant towards the top (Plates 4.12-4.13) where any primary sedimentary structures appear to have been totally destroyed. Pyritized trace fossils, of a similar nature to those found in Holocene sediments collected from the Skagerrak off southern Norway (FD. Werner, pers. com. 1986) are the predominant bioturbation structure. Also present are *chondrites* burrows and dense, myceloid clusters of pyrite thread (F. Werner, pers. com. 1986). The onset of bioturbation in this facies forms a sharp contrast with the unbioturbated basin fill muds. of facies D^8.

The clay mineralogy of this facies appears to vary with proximity to the western edge of the Basin. Thus, in the Bosies Bank area the mineral assemblage is generally rich in smectite where the underlying sediments are rich in smectite (Fig. 4.15), whilst elsewhere the clay mineralogy consists almost entirely of illite again reflecting the nature of the underlying sediments. The latter is especially typical of sediments from the central and western part of the Basin where a constant re-working of older Pleistocene sediments has produced a monotonous clay mineral assemblage consistently dominated by illite (56-64%).

With the exception of the basin periphery, facies E^8 generally overlies facies D^8. Apparently separated by a gradational boundary. However, seismic evidence suggests that a distinctive depositional boundary separates the two units as will be discussed later.

Around the periphery of the Witch Ground Basin (between 110-120m below sea level) facies E^8 contains more distinctive sedimentary structures including planar bedded, convolute bedded and flaser bedded sands and muds. Whole shells and shell fragments are abundant throughout these sediments as are bioturbation structures. The base of the sequence is sharp and generally erosional. Seismic evidence suggests in fact that the nature of the base is often slightly irregular.

Grain size analysis of the sediment allows for a four-fold subdivision of the muddy basin fill sediments, sub-facies $E1^8$ to $E4^8$ (Fig. 4.19). $E4^8$ is the uppermost sub-facies in the sequence and $E1^8$ the lowest. Sub-facies $E3^8$ and $E4^8$ are characterised by a very high silt content (70-85%) whilst sub-facies $E4^8$ is distinguished from $E3^8$ by its higher sand and coarse silt content and lower clay content. Sub-facies $E1^8$ and 2^8 contain less silt and more clay; a distinctive fine sand mode separates the latter from the former. The overall sequence is therefore one of a gradual upward coarsening, but with a finer kink relating to sub-facies $E3^8$, concomitant with an upward increase in the silt content.

Micropalaeontological analysis of this facies shows a slight discrepancy between the foraminiferal and dinoflagellate cyst evidence, although the facies generally occurs within bio-unit Q (Fig. 3.1). Foraminiferal evidence suggests that the lower part of facies E^8, in the Witch Ground Basin, was deposited in a hyposaline, shallow arctic marine environment, whilst the upper part was deposited in a temperate marine environment with conditons similar to today. The dinoflagellate cyst assemblage, however, indicates that the whole of facies E^8 in the Witch Ground Basin was deposited during an ameliorative period when the North Sea was open to the North Atlantic Drift and environmental conditions were similar to today. The exception to this is the brief cold period, identified in VE

135 and 111 (Fig. 3.36) and attributed to the Younger Dryas, which occurs within facies E^8. However, this brief environmental change was not recorded by any variation in lithology.

The shear strength of facies E^8 in the Witch Ground Basin ranges from 3 KN/n^2 to 16 Kn/m^2, and plasticity indices from 12 to 31. The lower plasticity of this facies relative to facies D^8 probably reflects the increased presence of silt size particles.

In the Witch Ground Basin facies E^8 forms a distinctive, uppermost, seismic facies characterised by its relatively transparent texture, pockmarked surface, and low amplitude concordantly layered reflectors; further salient seismic features are described in Chapter 2. In the deeper parts of the basin, generally below 140m (O.D), facies E^8 lies conformably on facies D^8, although the boundary between the two is often quite sharp (Figs. 2.37 - 2.39). Above this depth the base of facies E^8 is generally delimited by an angular erosive surface, although in places this surface is more irregular and diffuse.

ii. Channel infills:- these form characteristic heterogenous sequences with a complex variety of lithologies associated with similarly varied reflector configurations. The latter include, divergent, prograding and structureless configurations, described in detail in Chapter 2.

In BH 81/37 the sequence consists of a basal unit of laminated, flaser bedded and cross laminated muds and sands passing up into laminated muds with rare clasts, and overlain by structureless muds. The base of the structureless muds is recorded on the seismic record by the sharp transition from a lower symmetrical layered configuration to a transparent and structureless configuration.

The sequence in BH 81/39 is more complex consisting of units of well laminated and thinly interbedded sands and muds with intervening beds of massive muds and massive sands. The base of the sequence is marked by an 8m thick, crudely bedded and normally graded, muddy sand containing rip up clasts of clay and sub-rounded pebbles with a maximum dimension of up to 6cm. The pebbles increase in frequency towards the base whilst a large number of shell fragments are scattered throughout the unit.

Overlying the basal sand, massive muds, some 5m to 8m thick, separate units of interlaminated sands and muds. The latter consist of sharp based couplets of rippled silty sand with clay drapes, gradationally overlain by flaser bedded muds.

Massive muds and sands have sharp upper and lower contacts, and are further characterised only by the presence of shell fragments and monosulphide patches. An uppermost unit of massive sand, 10m thick, is characterised by the high content of shell fragments and a sharply erosive base.

BH 75/29 was also drilled through an infilled channel in seismic sequence 8 (Fig. 4.4). However, an overall poor recovery precludes detailed lithological descriptions and it is suffice to say that the sediments recovered consist predominantly of planar laminated, and occasionally ripple laminated, sands and muds. Plant and shell fragments are abundant in the lower part of the sequence.

Grain size analysis of a limited number of samples from this facies suggests that only muddy sediments with particle distributions similar to sub-facies $E1^8$ to $E3^8$ are present in the channel infill sequence, the intervening sand layers were not sampled. It is also obvious that the channel infill does not form the same gradual upward coarsening sequence as that observed in the Witch Ground Basin.

Foraminiferal evidence from this facies suggests that it was deposited in a hyposaline, shallow arctic marine environment. Further to this the foraminiferal assemblage in BH 75/29 and BH 81/37 (Appendix 3.11) indicates a progressive upward shallowing concomitant with increasingly harsh environmental conditions. The dinoflagellate cyst assemblage is consistent with this interpretation and suggests a harsh depositional environment, isolated from the North Atlantic Drift.

The channel infill sequence of facies E^8 therefore contrasts with its equivalent basin infill sequence in that it is generally associated with a harsh environment, as suggested by both the foraminifera and dinoflagellate cysts (bio-unit P).

Geotechnical properties of the sediment are similar to those described for the basin fill sediments with the exception that as expected the thicker channel infill sequence has a greater shear strength towards the base as a result of compaction and consolidation (7-25 KN/m^2).

Borehole 81/39 penetrated four distinctive seismic facies within sequence 8. The uppermost transparent and structureless configuration equates with the massive shelly sand, whilst the basal stratified, pebbly sand can be correlated with a westerly prograding, layered configuration. Unfortunately the intervening lithological boundaries are not easily related to specific reflectors or reflector configurations although it is obvious that the sediments are acoustically layered.

However, as mentioned previously, the sediments in BH 81/37 are easily equated to two separate seismic facies, whilst BH 75/29 penetrated a single seismic facies, within sequence 8, consisting of an acoustically well layered divergent type infill.

iii. Inter channel areas: - these are defined as areas of relatively minor relief which occur outside the Witch Ground Basin and in between areas of more irregular channelled relief. Muddy sediments of facies E^8 within this setting are generally restricted to areas with a present day water depth of greater than 110m (Fig. 1) and as such their occurrence in most of the Marr Bank and Peterhead areas is precluded.

Lithologically the sediments in this setting are identical to those seen in the basin infill sequence, and again form a gradually upward coarsening sequence of laminated muds and sandy muds. The overall sequence is, however, much thinner in these areas and rarely reaches a thickness of greater than 2m. The basal contact is sharp and unconformable and laterally the unit passes into the uppermost units of the channel fill sediments.

No micropalaeontological evidence is available and limited geotechnical data suggest that shear strengths are similar to those of the basin fill sediments (7-15 KN/m^2). Similar levels of plasticity were also recorded, with indices ranging from 11 to 22. The relatively low values again reflecting the high percentage of silt.

Where this sequence is within the resolution of the seismic equipment it is associated with a relatively transparent seismic texture and a poorly defined layering, similar to the upper basin fill seismic facies.

The sand rich unit of facies E^8, sub-facies $E5^8$, is ubiquitous over most of the study area, except in the very deepest parts of the Witch Ground Basin (> 140m water depth). The overall form of the sub-facies is one of a blanket type unit. Its upper surface occurs at the sea bed whilst the lower surface is usually sharp and erosional and it ranges in thickness from 0.2-3.0m but reaching up to 10m in the top of channel infills.

The sediment consists of grey green (5GY, 5/1) or olive grey (5Y, 4/1) to yellow brown (5Y, 5/2), massive or, occasionally laminated sand and silty sand. Both whole valves and shell fragments are common throughout the sequence often forming distinctive shell lags some 2-10cm thick. Angular to sub-rounded gravel and pebble size clasts are commonly abundant, especially at the top of the unit.

Grain size analyses of the sediment (Fig. 4.19) reveals a unimodal particle size distribution consisting predominantly of a well sorted fine sand fraction (40-80%) with a subordinate poorly sorted silt fraction (15-40%). The degree of sorting varies from moderately well sorted (S.D 1.5) to poorly sorted (S.D 2.5) and generally reflects the amount of mud present. Interestingly the amount of coarse sand present in those samples analysed is rarely greater than 2%. However, more detailed surveys of this facies taking samples from the sea bed suggests that coarse sand and gravel patches are extensive above 90m water depth (Owens, 1981). It is therefore thought that as samples for analysis were only taken from boreholes and vibrocores, any coarse unit occuring at the sea bed may well have been lost during recovery. It should also be noted that most work on the sea bed sediments in this area is based on the analysis of grab samples. Personal experience of running this type of equipment concurrent with coring equipment has indicated that the grab sample is commonly winnowed of fines during recovery.

Micropalaeontological evidence is again somewhat conflicting. For the most part both foraminiferal and dinoflagellate cysts are indicative of normal marine conditions with a temperature and water depth similar to the present. However, in certain boreholes, notably BH 81/34 and BH 81/37 (Fig. 3.1) the foraminiferal assemblages are more indicative of a hyposaline shallow arctic marine environment.

On the seismic record this facies is most consistently associated with a transparent texture occasionally characterised by a weak and coarse acoustic layering. The latter is typical in the north-west corner of the Bosies Bank area where low angle dipping reflector and large sand waves are also present.

Intepretation

i. Basin infill:- evidence from V.E 58+00/111 (Fig. 3.3) and V.E 58+00/135, suggests that deposition of facies E^8 in the Witch Ground Basin coincided with the onset of an amelioration in environmental conditions and an improved connection with the open ocean. It is therefore suggested that these sediments represent late Weichselian post-glacial deposits which rapidly accumulated in the basin at the onset of a transgressive period. The sediment supply would have been derived from the reworking of surrounding sediments, as indicated by the clay mineralogy, and from a steadily decreasing meltwater input. A decrease or absence of pack ice also characterised this environment concomitant with an increase in wave and tidal processes around the basin margins.

This interpretation is consistent with the nature of the basal seismic boundary to this facies. Thus, in the central basin the conformable base represents a depositional contact whilst along the basin margin the erosional boundary represents the approximate position of the palaeocoastline, located at about 100m below the present sea level.

The interpretation also explains the variation of sediment types between the periphery and basin centre. For example, in the latter facies E^8 is lithologically similar to the underlying sediments and is best distinguished by the presence of a favourable flora or the onset of bioturbation (identified from x-radiographs). However, around the basin margin facies E^8 is lithologically distinctive from the underlying sediments, and is indicative of a shallow water tidal environment.

The above mentioned palaeocoastline was also identified by Jansen et al. (1979) and attributed to a major low sea level minimum dated at about 15,000 yrs BP (Morner, 1969), and not to a maximum glaciation sea level stand, dated around 18,000 yrs BP (Jansen, 1976). The former interpretation agrees with the evidence cited here and suggests that subsequent to the deposition of facies D^8 a post glacial, low relative sea level stand, probably some 100-110m below the present, allowed uninterrupted suspension sedimentation to continue only in the deeper parts of the basin. The fact that the sea level stand was relative must be emphasised because of the probability of glacio-isostatic movement.

Sediment supplied to the basin during this period contained an increasingly significant proportion of silt much of which was probably derived from the southern North Sea, where there existed an extensive cover of fine eolian sands with a median grain size of 150 μm (Oele, 1971). Pockmark activity may also have contributed to the presence of high silt concentrations in the sediments by the expulsion of fine grained mud from sea bed into the lower boundary layer. Current activity may then have been able to winnow out fine particles leaving the coarse silt to settle out forming a concentrated silt layer (D. Long, pers. com. 1986).

ii. Channel infill: - in Chapter 2 a brief interpretation of the channel infill, based on reflector configurations, is given which basically describes a lower seismic facies characterised by coarsely layered, basally concordant configurations and an upper seismic facies characterised by downlapping, onlapping or structureless configurations; prograding reflectors tend to dip to the east or north-east. The former were attributed to deposition from suspension but with an element of coarse material reflecting either high energies or limited bottom current activity, the latter to shallow water, moderate or high energy conditions and a primary sediment source from the west.

Sediments recovered from the infills are generally consistent with this interpretation and the sequences described in BH 75/29 and BH 81/37, both associated with the upper seismic facies, are typical of a shallow marine, storm dominated environment possibly verging on an intertidal environment. Similar environments have been described along the coasts of the present North Sea by Van Den Berg (1981) and Yeo and Risk (1981) although these were dominated predominantly by tidal processes and not storm generated ones.

BH 81/39 recovered sediments associated with the lower seismic facies. These reflect a gradual transition from intertidal conditions at the channel base up into a slightly deeper or more protected marine environment dominated by suspension settlement.

It is obvious from the above variations that an all embracing mechanism of deposition for the channel infills is rather difficult to envisage, especially given the limited borehole control of such features. However, it is apparent that although the channels are attributed to fluvial processes the infills appear to be of a predominantly marine origin. Initial submergence of the channel features was therefore probably quite fast during which time extensive reworking of the surrounding land surface resulted in rapid sedimentation in the intervening channels. Accumulation rates were therefore able to keep abreast of the rising sea level resulting in the continued prevalence of a shallow marine or intertidal environment. Such conditions could have been maintained until the channel had been filled to the same level as the surrounding sea bed.

However, the above interpretation does not explain the existence of only partially infilled or totally open channels juxtaposed to infilled ones. Previous workers including Eisma et al. (1979), Jansen et al. (1979) and Flinn (1967), have interpreted the open channels as subglacial tunnel valleys. There is, however, no evidence of ice having reached these areas during the late Weichselian. It is therefore suggested that such features were maintained due to preferential scouring by tidal currents, although this is by no means certain.

The lack of distinctive intertidal or shallow marine sediments in BH 81/39, apart from in the basal unit, is a further anomaly within the above interpretation. As much of BH 81/39 is associated with the lower, concordantly layered, seismic facies it is suggested that such sediments represent locations where the sediment supply was not as great, therefore allowing for a gradual rise in the relative water depth and deposition from suspension in a low energy environment.

The relationship of the channel infills to the basin infill of facies E^8 is slightly uncertain, mainly because of conflicting micropalaeontological evidence. It seems likely that channel infill sediments related to the lower seismic facies are, in part, equivalent to facies D^8 in the Witch Ground Basin; both contain fauna and flora indicative of harsh climatic conditions. Channel sediments associated with the upper seismic facies are equated to the basin fill sediments of facies E^8; the former contain flora indicative of harsh conditions and in the latter, flora indicate the onset of an ameliorative period. This discrepancy can be explained by both the extremely shallow environment with the channel infill sediments and by the presence of large amounts of reworked glacial material.

iii. Sediments from the interchannel areas were similarly deposited in a shallow marine or intertidal area, although deposition of this material would not have commenced until the majority of channels had been infilled. As such, though no micropalaeontological evidence is available, it seems likely that these sediments are laterally equivalent to those of facies E^8 located around the Witch Ground Basin periphery.

The uppermost unit of facies E^8, the sand rich sub-facies $E5^8$, is interpreted as a palimpsest lag resulting from the reworking and winnowing of the sea bed sediments when the sea was still below its present level. Reworking would have occurred predominantly as a result of wave processes and to a lesser extent tidal currents. The presence of shell lag layers within this sub-facies is therefore explained by storm induced wave activity.

The relict nature of much of this facies is confirmed by its existence below 30m water depth, where there is generally only a very low level of wave effectiveness (McCave, 1971). Similarly the absence of this sub-facies over much of the Witch Ground Basin suggests that water depths in this area were greater than 30m during the formation of sub-facies $E5^8$.

Micropalaeontological evidence, suggesting environmental conditions similar to today leads to the conclusion that this uppermost sub-facies was deposited during the Holocene. A number of previous workers have also suggested this and corroborated it with radiocarbon dating (Jansen et al., 1979; Owens, 1977). However, the concept that it was initially formed by a Holocene transgression (Jansen, 1976) belies the fact that water depths were already increasing prior to the Holocene period (Jardine 1979). It is therefore suggested that sub-facies $E5^8$ is a palimpsest sediment, although at shallow water depths it is probably mobile (Owens, 1977), formed primarily by re-working of the sea bed sediments during a low relative sea-level stand but within an overall transgressive period. The latter would only have resulted in a rise in relative sea level when isostatic unloading and uplift became subordinate to the rise in sea level.

4.3 Summary

The late Weichselian glacial and glaciomarine record in the study area is represented by four facies, A-D. A fifth facies, E, represents the transition from glaciomarine or arctic marine to more temperate marine sedimentation spanning the late Weichselian and Holocene periods.

Sub-glacial deposition in the study area, represented by facies A^5 and A^7, was restricted to the western edge of the area where a hummocky topography, sub-aqueous moraines, and large scale deformation features are thought to represent the extent of the last ice sheet. The bulk of the subglacial sediment was probably deposited during the final stages of ice-advance whilst large upstanding features represent either periods of maintenance or slight re-advance of the ice front.

In the pro-glacial environment, high to intermediate energy glaciomarine sediments, facies C^6 and C^7, were deposited in an ice front depression resulting from glacial loading. Differences between facies C^6 and C^7 suggest that the former were deposited at relatively shallow depths to form a large subaqueous fan swept by strong traction or underflow currents. Evidence of an ice front depression is provided by the existence of shallower water depths away from the ice front. Facies C^7, however, reflects increased water depths and a general absence of shallow marine reworking of the glaciomarine sediments. The dominant processes associated with facies C^7 include underflows, turbidites, sediment loaded overflows and iceberg rafting. Evidence of slumping, especially on the margins of the sub-aqueous marine is also present.

Away from the ice front fine laminated muds with rare dropstones were depostied from a combination of iceberg rafting and overflow plumes. The retreat of the ice-front and deposition of the above glaciomarine facies is recorded by their westerly migration towards the direction of ice retreat. Concomitant with this, the more distal basinal areas, affected by seasonal pack ice and fed by a decreasingly significant component of ice rafted debris and meltwater sediment, became increasingly isolated as isostatic rebound resulted in the sub-aerial exposure of much of the surrounding sea bed. Much of the input into the basin was probably now supplied by a large river network to the south and the general erosion of the surrounding land mass. In the Marr Bank area, however, the originally shallow water depths and rapid isostatic uplift appear to have precluded any initial westwards migration of glaciomarine facies.

The eventual rise in relative sea level is marked by an infilling of many of the large channel like features in the Devils Hole area and subsequently by the re-working of much of the sea bed by tidal and wave processes prior to the establishment of present day conditions. A more detailed description and model of these events is provided in chapter 6.

CHAPTER FIVE

PRE- LATE WEICHSELIAN SEDIMENTARY FACIES

5.1 Introduction

The aim of the following chapter is to describe and interpret those facies which are pre-late Weichselian in age, as determined in chapter 3. Four separate facies sequences will be discussed, relating to seismic sequences 1 to 4 respectively, and ranging from Lower Pleistocene (possibly Prae Tiglian) to Upper Pleistocene (Weichselian) in age. Seismic sequence 1, however, is unique in that it contains an extensive lower unit of non-glaciomarine sediments. These facies will be described separately so as to give a clear idea as to their environments of deposition.

With the exception of sequence 1, facies nomenclature, and hence interpretations, will be compatible with Chapter 4. Repetition of detailed descriptions and interpretations will therefore be avoided, except where pertinent, and cross references to chapter 4 will be made where necessary. In fact, most of the information regarding these facies is derived from relatively widely spaced boreholes (Fig. 2.8), with only limited intervening vibrocore coverage and little knowledge of associated morphological features. Therefore the detailed analysis presented in Chapter 4 is not generally possible here because of the nature of the data set.

It should also be noted that throughout the pre-late Weichselian sequence there occurs a variety of marine sedimentary facies, which are associated with periods of climatic amelioration. As the aim of this project is primarily to study glaciomarine facies and processes, the above sediments will only be given brief consideration and are generally classified into facies E.

The grain size parameters of both late Weichselian and pre-late Weichselian sediments are described and discussed at the end of the chapter. Geochemical data, briefly discussed in the next section, is presented fully in appendix 6.

5.2 Facies A^1 to E^1

These are associated with seismic sequence 1, dated as being from the Prae Tiglian stage to the Elsterian stage (Fig. 3.2).

On the basis of palaeomagnetic and micropalaeontological data, the sequence can be divided into two facies units (Figs. 3.1 & 3.2)

i. A lower unit which occurs within the Matuyama epoch and consists primarily of marine and deltaic sediments which commonly contain fauna and flora indicative of ameliorative conditions.

ii. An upper unit which occurs within the Brunhes epoch and is composed primarily of glacial and glaciomarine sediments with a sparse to barren faunal and floral assemblage (bio-unit G). The first unit consists mainly of facies E type sediments while the second unit includes facies A-E.

The lower unit comprises the bulk of the sediments associated with seismic sequence 1, and is dated as Lower Pleistocene in age, possibly ranging from the Bavelian to the Prae Tiglian. At least two glacial stades occur within this unit although their effects appear to have been quite small and very restricted in extent.

The upper unit forms a relatively thin part of the sequence, and is dated as early Middle Pleistocene in age, probably relating to the early Cromerian complex (Fig. 3.2). In certain boreholes the above division is less distinct due to a lack of palaeomagnetic data or ambiguous micropalaeontological evidence. For example in boreholes 77/2 and 81/26 (Figs. 4.1 and 4.2) there are a series of facies E type sediments, associated with bio-unit G, which possibly represent deteriorating marine conditions and therefore record a transition from the underlying marine facies to the overlying glaciomarine facies. In boreholes 81/24, 81/33, 81/34 and 82/16 (Figs. 4.1 and 4.4) the picture is further complicated by the presence of facies E type sediments associated with harsh environmental conditions, bio-unit G, and an absence of glacial or glaciomarine sediments, facies A to D. The stratigraphic position of these sediments is therefore ambiguous as they may either predate or postdate the main glacial and glaciomarine sequences, or alternatively they may represent distal equivalents. Given the geographical location of boreholes 81/33 and 82/16 (Fig. 2.8) the latter explanation is unlikely and it was therefore decided to treat the sediments as a transitional unit; that pre-dates the main glacial and glaciomarine sequences.

5.2.1 **Facies E^1 (lower unit)**

This forms the lowermost unit of the Pleistocene sequence in the study area. It is also by far the thickest reaching a vertical extent of at least 200m in the Witch Ground Basin. A lack of deep seismic or borehole control precludes a more accurate determination of its thickness.

It occurs extensively over most of the study area below a water depth of 100m (O.D.) and in isolated pockets above this water depth, as witnessed in boreholes 81/25 and 81/33 (Figs. 4.1 & 4.3).

As noted at the start of the chapter the non-glaciomarine facies in this sequence will not be discussed in detail and a brief description of the range of lithologies within the marine and deltaic facies is best achieved from a prior description of the various palaeoenvironmental facies and the sediments associated with them. These are described in detail for the area south of 58°N by Stoker and Bent (In prep., Appendix 10), and essentially four sedimentary environments are recognised:- delta front, prodelta, intertidal or littoral and sub-littoral marine (see Figs. 4.1-4.4). Fig. 5.1 shows the approximate spatial extent of these, together with their spatial relationship to a deltaic sequence described by Cameron et al (1986); brief descriptions of these are given below.

i. Delta front :- sediments associated with this environment occur in boreholes 81/34 and 81/29 (Fig. 4.4), the latter representing the approximate northern limit of the sequence. However, to the south this sequence appears to correlate with the delta system described by Cameron et al. (1986) and as such passes laterally into the delta plain sediments located to the south of the study area (Fig. 5.1).

The sediments are composed of light brown and yellow brown to dark grey (10YR, 6/3 - 5/4, 10YR, 4/1) stiff or compact, laminated, lenticular bedded and massive silts and fine sands. Individual laminae vary from 0.5 to 10mm and are generally sharp, planar and, in places, defined by concentrations of mica and organic debris. Less common are slumped horizons and ripple cross laminated units which sometimes display 'within-core' bidirectional palaeocurrent trends. The more argillaceous units are often bioturbated and commonly contain soft sediment deformational structures including convolute lamination, folding, and microfaulting. Lenticular bedding is a common feature and wood and plant fragments occur throughout these sediments.

Massive sand beds range in composition from moderately sorted, clean sands to poorly sorted iron stained sands with abundant shell fragments. In BH 81/34 (Fig. 4.4) these often contained abundant gas vesicles which visibly bubbled on splitting the core.

Individual bed thicknesses, with the exception of the massive units, range from 0.1 - 10.0m and the bed contacts are generally gradational. The massive sands range from 0.4 - 3.8m in thickness and display sharp upper and lower contacts.

The micropalaeontological evidence is ambiguous and the fauna and flora may be indicative of a climatically harsh environment cut off from the North Atlantic Drift. Alternatively they may reflect the higher energies, relatively shallow water depths, rapid deposition rates, and greater percentage of sand, associated with a delta environment.

The exception to the above occurs in BH 81/34 (Fig. 4.4) where an argillaceous sequence of laminated silts contains microfauna and flora indicative of a temperate marine environment similar to today (bio-unit B).

The sediments are interpreted as being the products of two depositional settings within the delta environment. First, the planar laminated sands and silts and discrete slump horizons are typical of distributary mouth bar deposits such as those described by Coleman and Prior (1982) from the Mississippi River delta. These are deposited in an area of shoaling, associated with the seaward terminus of the delta channel, as the decreasing outflow competence results in rapid sedimentation and the development of a mouth bar. Textural laminations reflect differential settling due to variations in the current velocity (Sanders, 1965) whilst colour banding probably relates to changes in water chemistry (Coleman and Gagliano, 1965). Slump structures are consistent with rapid sedimentation and oversteepening of the depositional surface.

The second setting, the distal bar, is indicated by the presence of argillaceous laminated sediments characterised by synsedimentary deformational structures and ubiquitous bioturbation structures. The greater percentage of mud reflects the more distal location of this environment as the river outflow extends to the outer part of the delta front. Abundant bioturbation structures are a consequence of the lower sedimentation rates relative to the mouth bar whilst sedimentary deformation structures may be a result of either turbulent currents acting on the underlying bed (Sanders, 1965) or sediment gravity flows. The latter reflecting deposition on the unstable delta front slope.

Poorly sorted sand beds were probably deposited by stronger currents during flood stages; reworking of the mouth bar sediments by wave or tidal processes would account for the well sorted nature of some of the sand beds. The presence of red iron staining in some of the more poorly sorted massive sands is probably a result of oxidation processes acting during periods of subaerial exposure. The interpretation that such units are the product of storm generated processes is inconsistent with both their greater thickness and sharp upper contacts (Yeo and Risk, 1981).

ii. Prodelta:- these sediments ocur in boreholes 81/29 and 81/34 (Fig. 4.4) where they are interbedded with distal bar units, and overlie the main sequence of distributary mouth bar and distal bar sediments. Absolute bed thicknesses are difficult to determine due to its transitional nature of the prodelta, delta front and sub-littoral marine facies.

The sediments consist of brown to dark grey (10YR, 5/3 -

4/1), stiff, faintly laminated or massive muds with traces of lenticular lamination and small burrow structures. Individual laminae are generally diffuse and are defined by black monosulphides, the laminae range in thickness from 3-10mm.

Upper and lower contacts are commonly gradational with distal bar or sub-littoral marine facies.

Micropalaeontological evidence from this facies is not consistent between boreholes 81/29 and 81/34. In the former microfauna and flora indicate deposition in a harsh environment whilst in the latter they indicate temeprate marine conditions similar to the present day (Fig. 4.4).

The interpretation of these sediments as being the product of a prodelta environment rather than a sub-littoral marine environment is based primarily on the overall vertical sequence (Fig. 4.4), and especially their transition into adjoining distal bar type sequences. Given this, the sediments are similar to the Mississippi prodelta deposits described by Coleman and Gagliano (1965), and are possibly the result of suspension settlement from river overflows in a similar manner to that described from glacial meltwater overflows in chapter 4.

The contrasting micropalaeontological evidence from boreholes 81/29 and 81/34 suggests either strong spatial variations in water mass conditions or local differences in depositional environments.

iii. Littoral/Intertidal :- sediments representative of this environment occur in boreholes 81/26, 81/27, 81/33 and 82/16 (Figs. 4.1 & 4.3), where they are interbedded with sub-littoral marine sediments. These sediments are restricted to the western half of the study area and are not associated with the thicker Pleistocene sequence to the east (Figs 2.5 and 2.6).

The sediments consist primarily of yellow brown to grey brown (10YR, 5/4 - 5/2), compact or stiff, interlaminated and thinly interbedded muds and sands with occasional thicker units of upward fining massive or coarsely laminated sands.

Individual laminae are most commonly planar with gradational contacts although some ripple cross-lamination is evident in boreholes 81/27 and 82/16. In BH 81/33 the sediments are further characterised by lenticular, flaser and wavey bedding (Fig. 4.3). Bioturbation structures and shell fragments are common throughout the layered sand and mud sequences. Upward fining sand units are characterised by a basal erosion surface overlain by a lag deposit of poorly sorted sand containing abundant shell fragments and small pebbles. Ripple cross and planar lamination is crudely developed in some units whilst two structureless units of consolidated sand in BH 81/27 contain abundant organic debris (lignite, wood), clay drapes, rip-up clasts, and are distinctively iron stained. Bulk geochemical analysis of these sands indicates a high Fe_2O_3 content (Table 5.1).

The layered sand and mud sequences range in thickness from 3 - 10m and display sharp upper and lower contacts. The graded sand units vary from 0.35 -0.9m thick and the lower contacts are sharp and erosional. The upper contacts are generally sharp and separate this facies from the overlying sub-littoral facies.

Micropalaeontological evidence from all the respective boreholes, with the exception of BH 81/33, suggests that this facies was deposited in a favourable open marine environment; in BH 81/33 a harsher environment is indicated.

The thinly bedded and interlaminated muds and sands are typical of modern day offshore to lower shoreface zones, where abrupt changes and variations in depositional processes are associated with high energy storm waves and storm induced bottom currents superimposed on low energy fair weather processes resulting in a fluctuating energy regime (Reineck & Singh, 1972). This interpretation is preferred to a tidal flat environment primarily because of the generally favourable conditions indicated by the microfauna and flora, and the predominance of planar laminae rather than wavey or ripple lamination (Kreisa, 1980).

The thicker graded sand bodies are therefore consistent with deposition by storm driven processes resulting in sedimentary sequences similar to those described by Kreisa (1981) and Yeo and Risk (1981). The exceptions to this are the consolidated sand units in BH 81/27 which are similar to the tidal channel lag deposits described by Yeo and Risk (1981) from the Minás Basin estuary in the Bay of Fundy. The consolidated nature of the sediment and the high Fe_2O_3 content are consistent with sub-aerial exposure subsequent to deposition.

In BH 81/33 the whole sequence is more typical of the lower mudflat facies described by Yeo and Risk (1981). This is consistent with the high degree of bioturbation, presence of wavey bedding and the intercalated sand-silt units. Wavey bedding is absent in the upper mudflat and bioturbation structures less abundant.

iv. Sub-littoral:- Sediments typical of this environment are ubiquitous over much of the study area and occur in boreholes 77/2, 75/33, 81/26, 81/27, 81/29 and 82/16. This facies commonly forms the lowermost sampled unit from the Pleistocene succession.

The sediments consist of dark greenish grey (5GY, 6/1), stiff, massive or poorly laminated muds and sandy muds with scattered shells and shell fragments. Although predominantly massive, occasional sandy silt laminae and lenses and diffuse monosulphide bands are present. Discrete bioturbated horizons also occur. Carbonaceous and sulphidic material is disseminated through the core and nodular ironstone is locally developed. The clay mineral assemblage is relatively consistent throughout the facies and is dominated by illite (Appendix 6 Table 2) with subordinate chlorite and kaolinite. Strong scattering on the low angle side of the 10Å peak (Appendix 6, Fig 1) suggests that the smectite is partly interlayered with the illite.

Two phosphatic horizons occur within this facies. In BH 81/27, a hard, indurated light grey, faintly laminated band, 9 cm in thickness occurs towards the top of the sequence (Fig. 4.3). The sediment is coarse grained and consists primarily of quartz, biotite, muscovite, feldspar and

Sample	MgO	Al_2O_3	SiO_2	K_2O	CaO	TiO_2	Fe_2O_3	P_2O_5	Mn
1	0.839%	6.2%	67.7%	1.6%	3.6%	0.3%	17.3%	0.3%	750 ppm
2	0.9	6.1	78.4	1.1	3.9	0.5	4.7		493
3		10.4	52.5		9.2		9.5	3.3	
4		8.3	33.5		18.4		11.6	10.6	
5	2.2	14.9	62.1	2.7	4.6	0.7	5.9	0.1	538
6	2.2	15.5	55.7	2.3	2.4	0.8	7.4		1350

1: Littoral sand facies, BH 81/27.
2: Average composition of sand from Gulf of Paria, Venezuela (Hirst 1962).
3: Phosphatic band, BH 81/27.
4: Phosphatic band, BH 75/33.
5: Average composition of sub-littoral marine facies, North Sea.
6: Average composition of muds from Gulf of Paria, Venezuela (Hirst, 1962).

Table 5.1. Bulk geochemical data from various sedimentary facies.

hornblende, together with several grains of glauconite. The phosphate content of this band is shown in Table 5.1.

The second horizon, in BH 75/33, is characterised by a greater phosphate content (Table 5.1) and faint stratification. Although the phosphate content of both these horizons is higher than the average for marine and glaciomarine sediments from this and other areas (Table 5.1) it remains below that of a true phosphorite as defined by Bentor (1980). The geochemistry of these sediments is further discussed in appendix 6.

In BH 75/33 the essentially homogeneous sub-littoral muds are interrupted by a single sand bed some 16.5m thick (Fig. 4.2). The sand is aerated and consists of dark grey (10YR, 4/1), fine sand with rare mud drapes. The base of the unit is sharp and the nature of the upper contact is unknown.

The sub-littoral facies is the thickest unit in the Pleistocene and bed thicknesses vary from 1m to several tens of metres; a maximum thickness of 110m being attained in BH 75/33.

Both foraminiferal and dinoflagellate assemblages are consistent with deposition in temperate marine environments (bio-units B8 f/O Fig 3.1) although in some boreholes the base of the sampled sequence contains fauna and flora more characteristic of harsh environmental conditions (bio-unit A). Similar conditions are also indicated higher up the sequence, in boreholes 75/33 and 81/26 (bio-units C and E) and the restricted occurrence of glaciomarine sediments in (Figs. 4.1 - 4.2) association with these biozones are the only evidence of early, Pre-Cromerian glacial periods in the Pleistocene sequence.

The sub-littoral facies is envisaged as having been deposited mainly from suspension in a low energy predominantly temperate marine environment. Mixed layer minerals within these sediments are typical of low energy marine environments where they are more stable than single species (Berry and John, 1966). The origin of the phosphatic horizons is uncertain, although it is likely that they formed during a paucity of sedimentation. One mechanism commonly invoked to explain present day phosphate beds is the upwelling of cold, nutrient rich oceanic waters onto the adjacent shelf areas where small amounts of phosphate are precipitated (Elverhoi & Roaldset, 1983).

The thick sand body in BH 75/33 is interpreted as a tidal sand ridge (Stoker & Bent, in prep. Appendix 10), although some form of channel infill cannot be excluded, formed during a low sea level stand. This interpretation is consistent with the thick nature of this unit and its occurrence in what is otherwise an argillaceous sequence.

Discrete horizons of glaciomarine sediments (facies C and D, figs. 4.1-4.2) within this sequence are attributed to lower Pleistocene glaciations. These were either very restricted or subsequent erosion has removed evidence of them over much of the study area.

5.2.2 Facies A^1 to E^1

Early Middle Pleistocene glacial and glaciomarine sediments were first described from the Marr Bank area of the North Sea by Stoker and Bent (1985, Appendix 7). The sediments were shown to have a maximum eastwards extent of 0°E although their northern extent was not investigated. Further work presented here suggests that glacial and glaciomarine sediments deposited during the same period are in fact quite extensive and can be identified throughout the Bosies Bank and Fladen areas and part of the Forties area(Fig. 5.3). No information is available for the Peterhead area.

Facies A^1 to E^1 are all associated with chaotic reflection configurations, identified in chapter 2 as forming an uper seismic facies within sequence 1. This upper facies has a maximum thickness of ca. 50m in the Witch Ground Basin, but generally composes only a small proportion of sequence 1 except along the western edge of the study area where it appears to overstep the underlying seismic facies associated with facies E^1.

Facies A^1

The existence of this facies in six relatively widely spaced boreholes (Fig. 5.3) suggests that it is ubiquitous over a significant proportion of the study area. In the Marr Bank area it appears to be restricted to the west of 0°30'E, and is not present in the Devils Hole and Forties areas. However, it occurs over much of the Bosies Bank and northern Fladen areas where it reaches a thickness of between 40-50m (Fig. 5.3).

The sediments are similar to those described for facies A^5 and A^7 consisting of firm to stiff, massive diamicts with similar particle size distributions (Fig. 5.4). An unusual feature of facies A^1 is the thickness it attains in BH 81/19 and BH 81/26 (Figs. 4.1 & 4.2). In the latter, it forms a single unit some 40m thick, although any variations may have been missed by gaps in core recovery.

The base of facies A^1 was sampled in a number of cores, and it is generally sharp and erosional. In BH 74/12, for example, facies A^1 rests sharply on pyritic Lower Cretaceous shale (Fig. 4.3) and thin rip up clasts of the Cretaceous strata are included near the base of the sequence. Deformation of the sediments underlying this facies was also observed in two boreholes, BH 81/26 and BH 84/13 (Plate 5.1).

Analysis of clasts from this facies (Stoker and Bent, 1985) revealed that 55% were faceted and 20% displayed striations. Clast composition varied with the geographical location of the sample. Thus, in the Marr Bank area Moine/Dalradian metamorphics, Devonian sandstones and Mesozoic sediments were present whilst further north red and green sandstones, probably of Devonian and Permo-Trias origin are prominant with subordinate quartzite, chalk and siltsone clasts. This variation is reflected in the clay mineral assemblages in this facies. As such the significant presence of kaolinite and smectite in facies A^1 in the Marr Bank area (Fig. 5.2b) is related to the reworking of Mesozoic strata, as discussed in Appendix 6. In contrast facies A^1, in say BH 81/19 contains a clay mineral assemblage strongly dominated by illite with only minor amounts of kaolinite and smectite.

Interpretation

Sediments in facies A^1 are consistent with subglacial deposition, probably by lodgement processes. This is indicated by the very poorly sorted nature of the sediments, the presence of abundant clasts, their overall massive nature, even in thick sequences, deformation of the underlying material and the sparse or barren microfaunal and floral assemblages. It is suggested that their stratigraphic location represents the approximate maximum extent of the ice grounding line. However, it is not possible to discern whether facies A^1 in the Bosies Bank and Fladen areas was deposited by the same ice sheet, that is Scottish or Scandinavian, as the sediments in the Marr Bank area. Clast provenance tended only to reflect the relatively local geology and no exotic clasts of definitive Norwegian origin were observed.

Thicker accumulations of facies A^1 probably represent a progressive build up by glacial surges, this is corroborated in BH 81/19 (Fig. 4.1) where thick units of facies A^1 are separated by thin poorly sorted gravel bands 0.2m thick.

Facies B^1

Slumped sediments of an early Middle Pleistocene age were not recovered from the study area, a fact which possibly partly reflects the lack of channel fill features associated with seismic sequence 1.

Facies C^1

This facies forms an extensive and lithologically complex unit, some 10-20m thick. Its extent, from borehole and vibrocore recovery, appears to be very similar to facies A^1 and it is again absent in the Devils Hole and Forties area (Fig. 5.3).

In the Marr Bank area facies C^1 forms an easterly thinning wedge of sediments which can be traced as far east as BH 81/27 (Fig. 5.3). The geometry and extent of facies C^1 in the Bosies Bank and Fladen areas is less clear, partly because the unit has commonly been eroded out by subsequent channel sequences. However, it is clear that facies C^1 is not present in the centre of the Witch Ground Basin (BH 75/33), but that it does occur around the northern and western borders of the basin (Fig. 5.3) and over much of the Bosies Bank area. Information from the Peterhead area is not available.

The sediments of this facies are similar to those described for the late Weichselian, consisting of both sub-facies C1 and C2 cyclopel units, reverse and normal graded sands, and complexly bedded sands and muds with dropstone layers. The latter is typical of the sequence in BH 81/19 (Fig. 4.1) where units (0.1-0.5m thick) displaying planar, convolute and ripple laminated structures are interbedded with cyclopel units, and planar laminated muds. This sequence passes transitionally downwards into stratified and massive diamicts (sub-facies $C2^6$).

In BH 81/36 stratified diamicts are similarly associated with planar laminated and ripple laminated sands, together with a distinctive unit of reverse graded sand 10cm thick (Fig. 4.3). Analogous sediments occur in a number of vibrocores in the western Bosies Bank area, notably V.E. 58-02/164 and 257 (Plates 4.2 & 5.2). However, given the interdigitation of sub-facies $C1^1$ and $C2^2$, there is in the Marr Bank area a clear vertical and lateral (eastwards) transition from sediments predominantly of sub-facies $C1^1$ to those of $C2^1$. A similar pattern is not apparent in the Bosies Bank and Fladen areas.

The basal sediments of sub-facies $C1^1$ in the Marr Bank area consist of an upward fining unit of structureless coarse gravelly sand, as seen in BH 81/36 and 81/40. No equivalent to this was recovered in the Bosies Bank and Fladen areas. A similarly unique sequence occurs in BH 74/7 in which sediments of sub-facies $C2^1$ display a sub-horizontal pebble fabric, and in places a clast supported texture (Plate 5.4).

The clay mineralogy of facies C^1 shows similar variations to those described for facies A^1. The exception to this is seen in sub-facies $C1^1$ in the Bosies Bank area (Fig. 5.2 & 5.5) where the presence of expanding chlorites is thought to be the result of the reworking of local Permo-Triassic strata.

Where sampled, the base of facies C^1 is generally sharp and overlies either pre-Pleistocene strata or the subglacial diamict of facies A^1. The exception to this occurs in BH 84/13 where facies C^1 underlies, and has been deformed by, facies A^1 (Fig. 4.2), and in BH 77/2 where facies C^1 is interbedded with thin units of D^1 (Fig. 4.2).

Grain size analysis of facies C^1 (Fig. 5.4) reveals the textural homogeneity of sub-facies $C2^1$ relative to $C1^1$. The former has a bimodal particle size distribution, is very poorly sorted (S.D. 3.0-3.4) and contains roughly equivalent proportions of sand and mud. The sediments of sub-facies $C1^1$, however, display bimodal to strongly unimodal distributions, are poorly to moderately sorted (S.D. 0.8-3.6), and contain a significant proportion of sand (60-98%).

Interpretation

The previously described sediments are consistent with those described in Chapter 4 for facies C^6 and C^7 and as such represent deposition from sediment laden underflows with a subordinate contribution from overflows and ice rafted debris. The presence of intervening units of stratified diamicts (sub-facies $C2^6$) indicate a relative reduction in bottom current activity and a more significant contribution from overflow plumes and ice rafting.

Relative to the late Weichselian sequence, facies C^1 is different in terms of both its greater thickness and more extensive occurrence. The latter is consistent with the much greater extent of facies A^1. However, the greater thickness of facies C^1, and especially sub-facies $C1^1$, possibly suggests that the sediments were deposited over a much longer period.

It is therefore envisaged that facies C^1 was deposited in a, high to intermediate energy, glaciomarine environment fronting onto an extensive grounded ice sheet which established itself over much of the study area. In the Marr Bank area the sediments of facies C^1 were deposited on an easterly thinning subaqueous outwash fan, probably in water depths of less than 100m. Strong traction currents driven by sediment laden underflows were the primary mechanism of deposition, although sediment gravity flows may also have been important.

A lateral (eastwards) and vertical transition into sediments more representative of deposition from overflow plumes and ice rafted debris, is indicative of their more distal nature and a westward retreat of the ice sheet. Because of the lack of seismic information it is impossible to say whether the coarse, basal, gravels and sands, seen in the Marr Bank area, are inherent to the fan environment or represent a marine transgression in response to isostatic depression in a similar manner to that described for sub-facies $C3^6$ in Chapter 4.

In the Bosies Bank and Fladen areas it is impossible to identify, in detail, the sedimentary environment, although the sediments and structures are still consistent with deposition in a high to intermediate energy, glaciomarine environment. The limited extent of facies C^1 in the Fladen area, relative to facies A^1, may have been due to greater water depths at the ice front (100-150m) therefore preventing the lateral expansion of large subaqueous fan complexes, but allowing for a thick sequence to develop in the immediate ice front zone. This is supported by the contours drawn to the top of seismic sequence 1 (Fig. 2.12) which show a distinct deepening around the central Fladen area, even when given the present day bathymetry (Fig. 1.4).

In the western Bosies Bank area facies C^1 is more extensive possibly reflecting shallowing water depths, especially towards the north-west (Fig. 5.3).

The fact that in places C^1 is overlain by facies A^1 (BH 81/26 and BH 84/13 Figs. 4.1-4.2) corroborates the earlier suggestion that the environment in this area was affected by a series of glacial surges and retreats.

Facies D^1

Despite the extensive occurrence of facies A^1 and C^1, this facies was recovered in only five boreholes (77/2, 77/3, 75/33, 81/19 and 81/27) and is generally absent from much of the study area, with the exception of the Witch Ground Basin.

It occurs in vertical association with facies C^1 in three boreholes (81/19, 77/2 and 81/27 Figs. 4.1-4.3) where it is some 4-6m thick. In the remaining boreholes it directly overlies marine facies and ranges in thickness from 2 to 30m. Interestingly the thickest sequence was recovered from the forties area (BH 77/3, Fig. 4.2) and not the Witch Ground Basin. Where the facies occurs overlying facies C^1 it is separated by a gradational contact. Elsewhere the upper and lower contacts are either sharp or gradational with adjacent marine units.

The sediments are characterised by olive-grey to dark brown (5Y, 5/2, 10YR, 3/3), stiff, laminated and massive muds with rare clasts. The laminae are generally faint and diffuse although x-radiographs of the sediment were not obtained due to the generally poor condition and consolidated nature of the core material. In boreholes 77/2 and 77/3 the sediments are interdigitated with facies A^1 and E^1 respectively (Fig. 4.2). Clay mineral analysis of sediment from this facies shows that illite is the predominant clay species (64-68%) and the sharp peaks obtained (Fig. 5.5) suggest the clays are well crystallised.

Grain size analyses of the sediments (Fig. 5.4) revealed that in the Fladen and Forties area facies D^1 was composed predominantly of sub-facies $D2^1$ and $D3^1$ whilst in Bosies Bank and Marr Bank the sediments were composed of the coarser and more poorly sorted, sub-facies $D1^1$.

Geotechnical tests on facies D^1 (Figs. 2.50-2.52) indicated high shear strengths (100-150 KN/m^2) and moderate platicity indices (P.I 37-41). In BH 75/33 the top sediments of this facies were heavily consolidated (O.C.R 2.7), although overconsolidation values decreased rapidly down the sequence.

Interpretation

The characteristics of this facies are consistent with deposition in a relatively low energy environment dominated by suspension settlement. Probably from sediment overflow plumes, as suggested by the high mud content and lack of current indicators. Coarser sand material and clasts, especially in sub-facies $O1^1$, were probably transported by iceberg rafting and subsequently deposited as the iceberg melted or toppled over.

Given the previous interpretation of facies A^1 and C^1 the limited extent of this facies may be explained by two possibilities. First, that the final retreat of the ice sheet was very fast and that the responding isostatic rebound was of a similar rapid nature. This would have had the effect of precluding or limiting a migration of the glaciomarine environment towards the direction of ice sheet retreat. Secondly, that much of Facies D^1 has been removed by subsequent erosion. The latter is supported by the fact that in most boreholes facies A^1 or B^1 is cut by an unconformity. However, in borehole 74/7 and 82/15 (Figs. 4.1 & 4.3) facies C^1 is directly overlain by facies E^1 therefore supporting the hypothesis of a rapid retreat of the ice front, isostatic uplift and finally, a marine transgression. It would also be an unusual coincidence if only facies D^1 had been removed by erosion leaving the underlying facies A^1 and C^1. The suggestion of rapid ice front retreat and isostatic uplift is therefore preferred.

The high local overconsolidation ratio and underlying rapid decrease in the ratio in BH 75/33 suggests that the facies was consolidated by subaerial processes rather than ice loading (Hobbs, 1978).

Facies E^1

Relative to facies D^1, the geographical extent of this facies is more extensive (Fig. 5.3), although it is still restricted to a few boreholes, namely 77/2, 77/3, 74/7, 81/27 and 82/15, where facies C^1 and D^1 are overlain by sediments of facies E^1. These are typically dark green grey (5BG, 4/1), stiff, laminated and thinly bedded muds. Lenses of coarse silt and monosulphide streaks occur throughout the sequence together with bioturbation structures and occasional shell fragments. A single, large clay ball was observed in the facies in BH 82/15.

The basal contact varies from sharp to gradational the upper contact, is sharp. Grain size analysis of the sediments (Fig. 5.4) revealed their bimodal nature and nonlognormal distribution similar in shape to sub-facies $E1^8$ - $E3^8$.

Limited micropalaeontological evidence is indicative of harsh arctic conditions, cut off from the North Atlantic drift.

The shear strength of the sediments ranges from 100-300 KN/m^2 and plasticity indices from 28 to 32. The sediments are normally to overconsolidated (O.C.R 1.02-1.5).

Interpretation

With the exception of the geotechnical characteristics and micropalaeontological evidence this facies is very similar to the basin-fill sediments of facies E^8. It is therefore interpreted as a post-glacial sediment deposited in a sub-littoral environment. The microfauna and flora suggest that harsh environmental conditions persisted throughout the deposition period although lithologically there is no evidence of glacially influenced processes.

Where the base of this facies is gradational with facies D^1 it is suggested that the sediments accumulated at the onset of a transgressive period as the ice front retreated from the immediate area. Where the base is sharp, as is generally the case, interpretation is limited to suggesting the presence of cold possibly hyposaline waters in a marine shelf environment. This environment was probably cut off from the North Atlantic Drift circulation system. Alternatively, the microfaunal and flora evidence may relate to reworking of the underlying glaciomarine sediments as was suggested for the channel infill sediments of facies E^8 in chapter 4.

5.2.3 **Summary**

The Lower Pleistocene sequence in the study area is characterised by a series of deltaic and shelf marine facies. It should be stressed that correlation of individual facies between boreholes is complex and tentative and is discussed in greater detail by Stoker and Bent (in prep, Appendix 10). Here it is suffice to say that there is evidence that a delta system prograded northwards, through the southern North Sea (Cameron et al., 1986), and extended into the southern part of the study area. The identification of this delta system in the study area is based primarily on seismic correlation with the southern North Sea, vertical and lateral sequence analysis and evidence of fluvial input into the system.

To the North, beyond the limits of the delta system, the Lower Pleistocene succession developed in a sub-littoral setting producing a thick monotonous argillaceous sequence. However, along the western part of the study area this environment was interrupted by regressive periods and the predominance of nearshore marine processes.

It is therefore suggested that, subsequent to the initial development of a sub-littoral facies over much of the study area, a delta front system advanced into the central North Sea. This was associated with a high sediment input into the basin resulting in an overall regression and the emergence or near emergence of marginal areas such as those along the western part of the study area.

The development of delta front facies in the south western part of the study area is therefore related to sub-littoral and intertidal facies deposited in the marginal areas. However, direct correlation of individual units between the delta front and marginal areas is difficult and perhaps presumptious.

The implications of the above to basin development will be discussed further in chapter 6. It should also be mentioned that in BH 81/26 the identification of a thin sequence of glaciomarine sediments within the Lower Pleistocene are the earliest observed from the study area.

During the early Middle Pleistocene the onset of a major glacial stade and consequent harsh environmental conditions resulted in a change in microfaunal and floral assemblages, associated with sublittoral sedimentary facies, prior to the deposition of a thick sequence of glacial and glaciomarine sediments. The latter are related to an extensive ice sheet, or ice sheets, which underwent a series of surges and retreats in the northern part of the study area, prior to the final ice-sheet retreat.

The above interpretations are supported by the seismic facies interpretations discussed in chapter 2 (Table 2.1). Thus, the lower seismic facies, indicative of uniform rates of deposition in a stable, uniformly subsiding area, is associated with predominantly argillaceous marine and deltaic sedimentation. Conversely the upper seismic facies is consistent with deposition in a variable energy environment and shallower water depths therefore reflecting the influence of glacial and glaciomarine sedimentary processes. Table 5.2 summarises the above facies and interpretations and also those facies described in the following sections.

5.3 **Facies B^2 to E^2**

These facies are associated with seismic sequence 2, dated as being of Middle Pleistocene age and probably spanning the Elsterian to Saalian stages. It generally forms a channel-like infill characterised by a variety of reflector configurations, described in chapter 2, and the common occurrence of gas blanking. However, limited borehole evidence suggests that intraformational reflectors cannot be used to delimit facies boundaries. A series of lateral accretion structures occur at the base of some channel features, as described in Chapter 2, and are believed to be fluvial in origin (Fig. 2.21).

On the basis of lithological and micropalaeontological data a marine sequence, facies E^2, and a glaciomarine sequence, facies C^2 to D^2, have been identified.

5.3.1 **Facies E^2**

This occurs in boreholes 81/34 and 84/13 (Figs. 4.2 & 4.4) where it is associated with a basal channel infill sequence. It is best developed in BH 81/34 where the sequence is some 80m thick.

The sediments consist of green grey (5BG, 4/1), firm, laminated and thinly bedded muds and sands. In BH 81/34 the lower 60m are characterised by abundant bioturbation structures, mainly planolites/chondrites, a high proportion of silt, black sulphide laminae and horizons of flaser bedding. Individual laminae are usually diffuse and commonly discontinuous. Gas vesicles are abundant in coarser horizons towards the top of the sequence. In BH 84/13 the sediments are of a very similar nature but within a condensed sequence some 0.25m thick, large amounts of carbonaceous matter and gas vesicles were again observed. The upper 20m in BH 81/34 is more monotonous and contains fewer burrows and primary sedimentary structures. A horizon of convolute and deformed laminae marks the top of the sequence.

Micropalaeontological evidence suggests that the lower 60m in BH 81/34 and the sequence in 84/13 were deposited in a temperate marine environment with oceanographic conditions very similar to the present day (bio-unit H). This contrasts with the upper 20m in BH 81/34 which contains microfauna and flora indicative of a harsh environment probably cut off from the main North Atlantic Drift circulation system (bio-unit I).

Interpretation

The lowermost 60m in BH 81/34 and the sediments in BH 84/13 are consistent with deposition in a sub-littoral environment during a period of amelioration. The upward coarsening nature of this sequence suggests a transition from an environment below the wave base to one above it. Gas structures in the coarser units are possibly explained by their rapid deposition on the underlying organic rich silts and the upward migration of biogenic gas into the sands. A similar phenomenon is described in sediments from the Mississippi delta (Coleman & Gagliano, 1965). The presence of seismic gas blanking over much of the study area in seismic sequence 2 may relate to the presence of sedimentary sequences similar to that described above, alternatively much of the gas may be of a petrogenic origin.

The upper 20m are similarly indicative of an upward shallowing sub-littoral environment as reflected by the transition from laminated muds up into thinly interbedded sands and muds containing wavey laminations. However, the microfauna and flora and sharp reduction in bioturbation structures suggests the onset of harsh environmental conditions possibly related to the development of a glacial period. Alternatively the environmental change may reflect a variation in sedimentation rates or water chemistry, although there is no positive lithological evidence of this.

5.3.2 **Facies B^2**

This facies is restricted to BH 81/26 (Fig. 4.1) where it is some 10m thick. Seismically it is undistinguishable from the underlying facies C^2.

The sediments consist of dark grey, firm, laminated sandy muds coarsening up into a stratified diamict. The laminae are generally highly disturbed and folded with discrete horizons of planar lamination and the sediment has a distinctly heterogenous nature. Both the upper and lower contacts of the sequence are sharp.

Interpretation

This facies is interpreted as a slump deposit probably propagated under similar circumstances to those described for facies B^7. This is supported by the sharp base and the occurrence of undeformed layers between deformed zones (Reading, 1978). The fact that this facies occurs at the top of a channel infill sequence within seismic sequence 2 may suggest that the deposit is derived from the slumping of diamicts originally deposited on the channel margins.

5.3.3 Facies C^2

This facies occurs in boreholes 81/26 and 84/13 (Figs. 4.1-4.2) where it is 22m and 18m thick respectively. It is associated with a channel infill seismic facies within seismic sequence 2, but at a higher level than facies E^2. The associated reflector configuration is chaotic.

In borehole 84/13 the sequence consists of a lower and upper unit of dark grey (2.5YR, N/1), hard, poorly stratified diamict. Clasts are angular to subrounded with a maximum dimension of up to 40mm, chalk is the most common clast type. The stratification is delineated by black monosulphide layers and horizontal alignment of shell fragments. Both units are 0.5m thick.

A middle unit (10m) consists of dark grey (2.5YR, N/1) to dark green-grey (5GY, 4/1), hard, thinly bedded and planar laminated muds and muddy sands with matrix supported clasts. The contacts between this and the bounding units are gradational.

Similar sediments occur in BH 81/26 except that only two units are present, a lower stratified diamict and an upper unit of interlaminated sand and mud containing very rare clasts. The contact between the two units is gradational.

Micropalaeontological evidence, and especially the dinoflagellate cyst assemblge in BH 84/13 (Appendix 2.9), suggests that the sediments were deposited in hyposaline, shallow arctic water conditions with the possibility of at least seasonal periods of sea-ice cover (bio-unit I).

Interpretation

Facies C^2 was deposited in a high to intermediate energy, glaciomarine environment, possibly as a subaqueous outwash fan, although given the limited borehole data it is impossible to verify this.

Sedimentary processes would have been similar to those described previously in chapter 4, and the sequence in BH 84/13 may be the result of surges and retreats of the ice front. The position of the ice front itself is impossible to determine due to the lack of readily identifiable subglacial material, facies A. The absence of the latter may purely reflect the widely spaced nature of the boreholes or alternatively it may be the result of recovering only channel infill sequences whilst any subglacial material might well have accumulated on the higher inter-channel areas.

5.3.4 Facies D^2

This occurs in only one borehole, 77/2 where it is 30m thick. It is associated with the upper part of a channel infill sequence which displays a chaotic to structureless reflector configuration.

The sediments consist of dark grey (5Y, 4/1), stiff, faintly laminated clays with rare clasts in the lower 12m of the sequence. Laminae occur as diffuse streaks of fine silt with gradational boundaries. The remainder of the sequence is generally massive with the exception of intermittent sand lenses and isolated shell fragments. Both the upper and lower contacts of this facies are sharp.

Micropalaeontological evidence suggests deposition in a hyposaline, shallow arctic water environment cut off from the North Atlantic Dirft system (bio-unit I).

Geotechnical testing on this facies (Fig. 2.50) confirmed the firm nature of the sediments with shear strengths of between 175-189 KN/m^2. Plasticity indices range from 22-32 and the overconsolidation ratio is 0.79.

Interpretation

This facies was deposited in an intermediate to low energy glaciomarine environment, by suspension settlement from sediment overflow plumes. An upward decrease in ice rafted debris is probably related to the retreat of the ice front. The underconsolidated nature of the sediment may reflect rapid deposition rates. It is interpreted as being the lateral equivalent of facies C^2 although an absence of vertical sequence relationships and the discontinuous nature of seismic sequence 2 makes this impossible to verify.

5.3.5 Summary

Sediments associated with seismic sequence 2 typically infill depressions associated with a highly irregular subaerial erosion surface which developed during a low sea level stand, probably in the Elsterian stage (Fig. 3.2). Seismic evidence suggests that fluvial deposition and lateral accretion processes were active during the period of low sea level stand.

A subsequent marine transgression, during the Holsteinian (Fig. 3.2), resulted in the deposition of a thick sequence of marine silts and sands which infilled the greater part of the channel fetures. The presence of large amounts of organic material in these sediments may have contributed to the production of large pockets of gas; these are observed on the seismic record as extensive areas of acoustic blanking within seismic sequence 2.

The development of harsher environmental conditions is interpreted as being related to the onset of the Saalian glacial stage. In BH 81/34 this change is recorded only by the microfauna and flora in sub-littoral marine sediments which infill the uppermost part of the channel feature. However, further north the temperate marine sediments were overlain by a series of glaciomarine facies which record an increasing proximity to the ice front towards the north and eastern margins of the Witch Ground Basin. In the Witch Ground Basin, only intermediate to low-energy glaciomarine sediments were recovered (facies D^2) suggesting that the ice front was not in the immediate proximity. Unfortunately the position and extent of the ice front is not known although it is quite possible that it was restricted to the Northern edge of the study area.

5.4 Facies A^3 to E^3

These facies are associated with seismic sequence 3 which is dated as Saalian in age (Fig. 3.2). No periods of amelioration were identified within the sequence. Sequence 3 occurs over much of the study area, with the exception of the Witch Ground Basin and Marr Bank area, and is most easily identified by its characteristically planar basal surface. The extent of seismic sequence 3 and the depths to its base (below sea level) are depicted in Fig. 2.27.

Internally, sequence 3 is characterised by three seismic facies, described in Chapter 2, and these can be correlated to a limited degree with the sedimentary facies as discussed below.

5.4.1 **Facies A^3**

This occurs in a single borehole, 84/13, and vibrocore 58-01/200 (Figs. 4.2 & 4.6), where it is 2m thick and at least 6m thick respectively. It is associated with the uppermost of the two main seismic facies in sequence 3 and is characterised by discontinuous, sub-parallel and irregular reflector configurations, and, in the Bosies Bank area, isolated ridges (Fig. 2.23). Upper and lower contacts are sharp, and no deformation of the underlying material was observed.

The sediments consist of dark grey (5Y, 4/1), hard, massive diamicts. Angular to sub-rounded clasts are abundant throughout the facies with a maximum dimension of 50mm. Red sandstone, chalk and siltstone clasts are the predominant lithologies. No pebble fabric was observed as shown in Plate 5.5. The clay mineral assemblage (Appendix 6 Table 5) is dominated by illite with subordinate proportions of kaolinite, chlorite and smectite.

Its occurrence in only two cores precludes any detailed knowledge of the extent of the ice-front. However, it is possible that the isolated ridges, shown in Fig. 2.24 represent sub-aqueous morainal banks similar to those described by Landmesser et al. (1982). Such features do not necessarily delimit the maximum ice extent and a more detailed picture of the palaeogeography is obtained by studying the glaciomarine sequences associated with seismic sequence 3.

5.4.2 **Facies B^3**

This facies occurs only in BH 81/34 where it is 6m thick (Fig. 4.4) and is bounded by facies C^3. The sediments consist of dark grey (5Y, 4/1) muds with paler sandy-silt lenses and laminae. The latter have been highly deformed and show evidence of folding, and micro thrusting and faulting (Plate 5.6). Within this facies a horizon, 1.5m thick contains no deformation structures, but is characterised by planar, ripple and lenticular bedding structures.

The upper and lower contacts of facies B^3 are sharp and the top of the sequence is punctured by a large basaltic dropstone clast.

Interpretation

Facies B^3 is typical of a slump deposit, in which the resedimented material has maintained some form of internal coherence. The presence of undeformed beds between deformed zones is further evidence of slumping (Helwig, 1970) and also precludes deformation during recovery.

Although the sequence here is not associated with a channel slope, it may have been propagated by failure along a steep depositional surface associated with very rapid sedimentation rates at the ice front. Similar phenomena have been recorded by a number of workers in both contemporary and ancient glaciomarine sequences (Cheel and Rust, 1982; Powell, 1983). Alternative mechanisms of failure to over-steepening, include iceberg calving, ice push, pore pressure fluctuations from storm waves at tidewater fronts, and fault associated back-collapse of banks when the ground line retreats (Powell, 1984).

5.4.3 Facies C^3

This facies occurs extensively over much of the Devils Hole, Forties and Bosies Bank areas where, in addition to its presence in a number of vibrocores (Fig. 4.6), it has been recovered in boreholes 81/24, 81/29, 81/34, 82/15 and 84/13 (Figs. 4.1, 4.2 & 4.4). A maximum thickness of 50m was recovered in BH 81/29. With the exception of BH 81/29, Facies C^3 is associated with the same reflector configurations as facies A^3, and the two cannot therefore be distinguished using seismic parameters alone. In BH 81/29, part of this facies is associated with a 30m thick horizon of complex downlapping reflector configurations (Fig. 2.25). The reflectors appear to dip towards the south or south-west and extend for some 5-10km. The overall geometry of this seismic facies is, however, uncertain.

Sediments recovered from the horizon of downlapping reflectors consist of pale grey (10YR, 7/1), compact, laminated sands with occasional lenses and laminae of dark grey (5Y, 4/1) mud. Individual laminae are often delimited by detrital rich bands. The sequence is further characterised by the presence of clasts of stiff clay, with a maximum dimension of 15mm, ubiquitous shell fragments and the occasional wood fragment. A distinctive band of massive diamict 2m thick occurs towards the base of the sequence.

In the remaining boreholes and vibrocores facies C^3 is more typical of previously described facies C sequences and both sub-facies C1, and C2 types were identified. Sub-facies $C1^3$ occurs in all the above mentioned boreholes, except BH 82/15, together with a number of vibrocores in the Bosies Bank and Forties areas (Figs. 5.6).

It consists predominantly of dark grey (5Y, 3/1), firm, cyclopel type units (Mackiewickz, 1983) with well defined laminae occasionally passing into discrete horizons of flaser bedded and cross-ripple laminated muds and sand. Clasts occur throughout the sub-facies. Plate 5.11 shows a thick 2m cyclopel unit in BH 81/24; the sequence is characterised by alternating well laminated and faintly laminated beds with discrete units of massive diamict, sub-facies $C2^3$. All contacts are gradational. The well laminated units contain up to 50 individual layers in a single bed (30mm thick) and the sand laminae display a sharp base with a gradational top into the overlying mud laminae.

Thicker sand beds (0.5-2.0m) are characterised by their sharp, erosive base, coarse nature, poor sorting and, commonly, a reverse grading. In vibrocores 58-01/208 and 222 (Plates 5.7 & 5.8) these sand units display an upward coarsening from thinly bedded muddy sands and gravels up into better sorted structureless sandy gravels.

Micropalaeontological evidence for both this sub-facies and sub-facies $C2^3$ is indicative of a hyposaline, shallow arctic environment cut off from the North Atlantic Drift System (bio-unit J).

Sub-facies $C2^3$ was recovered in boreholes 81/24, 81/34, 82/15 and 84/13, where, with the exception of BH 82/15, it occurs interbedded with sub-facies $C1^3$ or facies B^3. In BH 82/15 this facies forms a single isolated bed 2m thick. In the remaining boreholes and vibrocores its thickness ranges from 0.10m to 2m thick.

The facies is characterised by dark grey (5Y, 4/1), firm to stiff, stratified diamicts containing irregular pods and lenses of fine sand. Clasts are abundant throughout this sub-facies (Plate 5.13) and consist predominantly of red sandstone, chalk, and rocks of metamorphic affinity. Upon splitting the core sediments from this facies showed a well developed horizontal fabric.

The upper and lower contacts are commonly sharp or more rarely gradational.

The clay mineral assemblage from this facies in the Bosies Bank area (BH 82/15, Appendix 6 Table 4) is somewhat unusual in that it is generally devoid of smectite despite the fact that the immediately underlying bedrock consists of Palaeocene Tuffs and Siltstones, which are shown to contain up to 100% smectite (Appendix 6, Fig. 3). This discrepancy suggests that the main source of material for this facies was probably from the North where Devonian and Permo-Triass strata are present close to the sea bed (Appendix 6, Fig. 4).

Interpretation

The occurrence of a thick sequence of sand in BH 81/29 correlates with the horizon of downlapping reflectors within seismic sequence 3; as described in chapter 2. These were interpreted as prograding clinoforms, formed in a high energy regime during a period of high sediment supply and little or no subsidence. The occurrence of a thick sequence of sand in BH 81/29 is consistent with this interpretation. However, the actual environment of deposition is slightly ambiguous and two possibilities are suggested here. First, that the environment was an arctic delta or sandur which fronted onto an ice sheet (Syvitski, 1986), although such environments are most commonly found at the heads of large fjords. Secondly, that the sediments were deposited as subaqueous outwash from a sub-glacial ice tunnel which would be expected to produce a series of longitudinally-overlapping esker fans as the front retreated (Rust and Romanelli, 1975). This would give the impression of prograding clinoforms on the seismic record. The presence of a single diamict layer within the sand sequence may support the latter as Rust and Romanelli (1975) cite such occurrences as being compatible with the subaqueous outwash environment. Although no subglacial material was recovered from the immediate area the presence of stratified diamict can be taken as being indicative of a close proximity to the ice front.

No sample control exists for the horizons of downlapping reflectors in the Fladen area, described in chapter 2, and their origin therefore remains uncertain.

The laminated muds and sands, or cyclopel units, are interpreted as the result of deposition from suspension and by traction currents from overflows and inter or underflows respectively. As such they relate to a high to intermediate energy glaciomarine environment similar to that described in detail for facies C^6 and C^7 in Chapter 4. Deposition from sediment laden underflows produced the thicker sand units and the sharp erosive base, reverse grading and an upward increase in sorting is typical of the underflow deposits described by Mackiewicz et al. (1984) from Muir Inlent, Alaaska. The overall environment is envisaged as being a series of subaqueous fans which developed close to the ice front or ice fronts in relatively shallow water. The coarse grained diamicts of sub-facies $C2^3$ accumulated across the fans in areas of episodic traction current activity and active mud deposition and ice rafting (Eyles et al., 1985). However, due to the dearth of information in the Forties and Peterhead areas the relationship between faciies C^3 in the North and South of the study area is uncertain and will be discussed further in the summary.

5.4.4 Facies D^3

Recovery of this facies is restricted to the Devils Hole areas where it occurs in boreholes 81/29 and 81/34 (Figs. 4.4). In both areas it occurs interbedded with facies C^3 and is associated with sub-parallel continuous to discontinuous reflector configurations (Fig. 2.25). A single bed of facies D^3 was also recovered in VE 58-01/208 (Fig. 4.6) from the Bosies Bank area.

The sediments consist of dark grey, (5Y, 4/1), stiff, faintly laminated with scattered clasts. Individual laminae are diffuse and commonly delimited by black monosulphides. Thin silt lenses and small shell fragments occur throughout the facies.

The basal contact is usually gradational with facies C^3 and the upper contact is sharp. Individual beds range from 2-7m thick.

Micropalaeontological evidence suggests deposition in a hyposaline shallow arctic environment cut off from the North Atlantic drift system (bio-unit J).

Geotechnical measurements reveal a shear strength of between 125-185 KN/m^2, no other data are available.

Interpretation

This facies was deposited by suspension settlement from sediment overflow plumes and as such is indicative of a more distal environment relative to facies C^3. Given this, the interbedded nature of facies D^3 and C^3 is probably indicative of a fluctuating ice front in the Devils Hole area. The limited recovery in VE 58-01/222 precludes further interpretation.

5.4.5 Facies E^3

No borehole penetrated the lower seismic facies in seismic 3 predominant in the Bosies Bank and Peterhead areas and described in Chapter 2 (Table 2.1). This was interpreted as indicating uniform rates of deposition in a stable, uniformly subsiding, area. It is therefore suggested that the sediments associated with this seismic facies would be typical of sublittoral shelf marine deposits similar to those described for facies E^1.

5.4.6 Summary

The relatively planar base to seismic sequence 3 was attributed in Chapter 2 to a marine transgression during the Saalian period, and the subsequent establishment of a stable shelf environment is recorded by the seismic reflector configurations. No information is available as to whether this period represented an ameliorative episode during the Saalian or was the result of glacio-isostatic depression prior to the deposition of the main suite of glacial and glaciomarine sediments.

Given the limited recovery of sub-glacial sediments from this sequence it is impossible to try and delimit the maximum extent of the ice front. However, from the spatial distribution of facies C^3, either the ice sheet covered much of the study area or alternatively the sediments relate to more than once ice sheet. Certainly much of the Bosies Bank and northern Fladen areas would have been ice covered, as suggested by the topography and presence of subglacial deposits, facies A^3.

If the sequence is the product of a single ice sheet then the grounding line must have been located somewhere in the southern Devils Hole area. The subsequent northwards retreat of the ice front was interrupted by a series of surges producing the subaqueous outwash deposits of facies C^3 overlain by a sequence of thickly interbedded units of facies E^3 and D^3.

The final retreat of the ice-front to the northern edge of the study area would have resulted in rapid isostatic uplift of much of the ice freed area therefore precluding the affects of rising sea level and explaining the absence of a gradual upward transition from facies D^3 to E^3.

Reference to figs. 2.5, 2.6 and 2.7 show that in the Witch Ground Basin much of the sedimentary succession associated with seismic sequence 3 appears to have been eroded out. One possible explanation for this may have been the formation of an ice front bulge, similar to that described by Walcott (1972), as the ice-front stood to the North of the basin. Such a bulge would have exposed the uppermost sediments in the Witch Ground Basin to subaerial erosion and redeposition, probably to the south.

5.5 Facies B^4 to E^4

These facies are associated with a series of channel infill, basin infill, and blanket like, seismic facies within seismic sequence 4. The base of seismic sequence 4 is delimited by an irregular to highly irregular erosion surface thought to have formed during a low sea level stand in the Saalian stage. However, the age of the overlying sediments is somewhat ambiguous and reference to Fig. 3.2 shows that the sequence is thought to span from the late Saalian to the Weichselian stages. Horizons indicative of an ameliorative period (bio-unit L) are found only in association with channel infill seismic facies. Such horizons were interpreted in chapter 2 as being of an Eemian age.

Seismic sequence 4 is ubiquitous over much of the study area, with the exception of the western Marr Bank area, as shown in Figs. 2.5 and 2.6. In the Devils Hole and Forties area it has a discontinuous nature and occurs in large, isolated channel like features, whilst over the remaining area it is typically more continuous. The various sedimentary facies can, to a limited extent, be correlated to the different seismic facies described in Chapter 2.

5.5.1 **Facies E^4**

This occurs in boreholes 77/2, 77/3, 81/19 and 81/37, where, with the exception of BH 77/3, it forms the lowermost facies associated with seismic sequence 4 (Figs. 4.1, 4.2, 4.4). It occurs only in association with channel fill seismic facies and generally chaotic reflector configuraton. The thickness of facies E^4 varies greatly, depending on the size of the channel feature, and a maximum thickness of 30m was recovered by BH 77/3 which penetrated a multiple channel infill feature within seismic sequence 4 (Fig. 2.6, Line 8).

Based on micropalaeontological evidence the facies is divided into an upper and lower unit. The lower unit is characterised by microfauna and flora indicative of a hyposaline, shallow water, arctic environment cut off from the North Atlantic Drift System (bio-unit K). The upper unit, however, contains a dinoflagellate cyst assemblage typical of an ameliorative episode (bio-unit L) fully connected with the North Atlantic Drift system. Only in BH 81/37 are the foraminifera consistent with the dinoflagellate cysts, therefore indicating fully temperate marine conditions and an absence of water stratification.

The sediments in the lower unit form a monotonous sequence of dark grey (5Y, 4/1) and olive grey (5Y, 4/2), firm, faintly laminated or massive muds characterised only by the presence of thin silt lenses and numerous shell fragments. A single bed, 5m thick, of planar laminated fine sand is restricted to BH 81/37 (Fig. 4.4). The sequence ranges from 2-10m thick and both the upper and lower contacts are most commonly gradational.

Grain size analyses of this facies (Fig. 5.8) shows the bimodal and nonlognormal particle distribution of the sediment. They further highlight the fine nature of the sediment and the greater degree of sorting (S.D. 2.1-2.2) relative to the glaciomarine sediments.

In boreholes 77/2 and 77/3 sediments from the upper unit are lithologically identical to those in the lower. However, in BH 81/37 this facies displays a gradual upward fining from olive grey, (5Y, 4/2), firm, thinly interbedded and interlaminated sands and muds up into faintly laminated muds containing numerous whole shells. The latter are associated with the strongest indications of an ameliorative period. Individual lamina are generally planar or occasionally wavy and the sandy layers display sharp basal contacts and gradational tops. An anomolous unit of stratified diamict, 4m thick interrupts the sequence (Fig. 4.4) and is characterised by a sharply defined base and planar top.

Grain size analyses of the faintly laminated muds in this facies (Fig. 5.8) shows the nonlognormal and bimodal particle distribution of the sediment with distinctive coarse silt and clay modes and less than 5% sand. The shape of the distribution curve is similar to that described for the basin fill sub-facies, $E1^8$, in chapter 4.

Geotechnical tests on the same sediments (Fig.2.50) in BH 77/2 revealed a reversed shear strength profile with values decreasing down core from 41-28 KN/m^2. Overconsolidation ratios range from 0.8 to 1.13.

Interpretation

The lower unit of facies E^4 was deposited primarily from suspension in a harsh, probably hyposaline, environment under a low energy regime. A sub-littoral marine environment is envisaged, but one that was perhaps influenced by glacial climatic conditions.

The overlying unit is more consistent with deposition in an increasingly favourable sub-littoral environment associated with a marine transgression. This is reflected by the upward fining nature of the sequence in BH 81/37 which suggests that the sea level rise was more rapid than the rate of sediment input to the area. The basal laminated sands and muds are evidence of alternating energy levels typical of nearshore sub-littoral sediments (Nottvedt, 1985) whilst the overlying structureless muds represent a more offshore environment below the wave base.

The occurrence of an intervening unit of stratified diamict within this sequence is rather anomalous and is interpreted here as a slump deposit which originated from the channel margin resulting in the incorporation of both marine and glaciomarine material. The sharp base to the unit is consistent with the interpretation although no other evidence, such as flow folding or faulting, was observed.

Where the upper unit is characterised only by faintly laminated or massive muds, as in boreholes 77/2 and 77/3, the reverse shear strength profiles and the normal to underconsolidated nature of the sediment may be the result of rapid deposition and high sediment concentrations, perhaps related to increased erosion of the surrounding hinterland.

5.5.2 Facies B^4

This was recovered from only one borehole, BH 81/27, where it is 5m thick (Fig. 4.3). It occurs at the base of a succession associated with a channel infill seismic facies within sequence 4, the reflector configuration is chaotic.

It consists of dark grey (5Y, 4/1), firm, interlaminated muds and fine sands with occasional clasts. The laminae are highly contorted and distinctive overturned folding is apparent throughout the facies (Plates. 5.9 & 5.10). Small scale thrust and extensional fault structures are also present. The base of this facies is sharp and angular, as shown in Plate 5.19.

Micropalaeontological evidence suggests that this facies was deposited in an unfavourable environment, but one that was possibly less severe than observed previously. Alternatively the ambiguous nature of the microfauna and flora assemblages (bio-unit M, Appendix 2.4 and 3.4) may reflect an admixture of glaciomarine and temperate sediments.

Interpretation

This facies is interpreted as a slump deposit formed under very similar conditions to those described for facies B^2. The preservation of the sharp, angular basal boundary to this facies at 54m correlates well with the irregular erosive base of a channel fill seismic facies and is consistent with deposition by slumping. The presence of a mixed microfaunal and flora assemblage is probably the result of the erosion of temperate marine sediments, in the underlying sequence 1, and the incorporation of this material into sediment prior to slumping.

5.5.3 **Facies C^4**

This facies was recovered from all of the boreholes which penetrated seismic sequence 4; these include borehole 77/2, 75/33, 81/19, 81/24, 81/26, 81/27 and 82/16, in addition to a number of vibrocores in the Bosies Bank area.

The facies is generally associated with a blanket-like or basin fill type, continuous seismic facies within sequence 4. It is characterised by an irregular basal boundary associated with isolated channel infill seismic facies. In addition to this, in the Bosies Bank area, facies C^4 is associated with a series of ridge-like features depicted in Figs. 2.5 and 2.6. The exceptions to the above occur in boreholes 81/27 and 81/37 which penetrate a channel fill seismic facies in the Devils Hole and Marr Bank areas respectively. Individual beds range from 2-5m thick and have sharp basal contacts. They occur either as isolated units or interbedded with facies D^4.

Sediments from this facies are only of sub-facies C2 type and no C1 type sub-facies was observed. The sediments consist of grey to dark grey (5Y, 5/1; 5Y, 3/2), soft to firm, crudely stratified to massive diamicts. The stratified diamicts contain poorly defined sand and clay laminae with gradational boundaries. Clasts of chalk, sandstone, siltstone, granite and of various metamorphic affinities are common, with a maximum dimension of 40mm. Shell fragments occur throughout the facies, and upon splitting the core they commonly define a horizontal fabric. Rare clasts were also observed deforming the underlying stratification.

The clay mineral assemblage in both this facies and facies D^4 consists predominantly of illite (60-73%) with subordinate kaolinite, chlorite and smectite and it is characterised by its consistent nature over much of the study area (Appendix 6, Tables 6&7).

Grain size analysis of this facies (Fig. 5.8) shows the nonlognormal and bi-modal particle distribution, characterised by a poor degree of sorting (S.D. 2.7-3.1) and significant amounts of fine sand and coarse silt. Reference to Fig. 4.12 shows the similarity of this facies to sub-facies $C2^6$ and $C2^7$.

With the exception of BH 77/2 and, to a lesser extent, BH 81/27, the micropalaeontological evidence is indicative of a hyposaline shallow arctic environment cut off from the Atlantic (bio-unit M). However, in BH 77/2 this facies contains dinoflagellates indicative of an ameliorative period (bio-unit L)and in BH 81/27 the microfauna and floral assemblages are of a mixed nature.

Geotechnical measurements from this facies (Figs. 2.50 and 2.52) indicate a shear strength of between 25-75 KN/m^2, plasticity indices between 17 an 27, and an overconsolidation ration of 0.89 - 1.9. The latter value was recorded from BH 77/2.

Interpretation

Facies C^4 is interpreted as the product of a high to intermediate energy glaciomarine environment, deposited in areas of episodic-traction current activity, where significant sediment contributions are from overflows, interflows and the ice rafting of debris. The presence of dropstone structures supports the latter.

A mixed microfaunal and flora assemblage in BH 81/27 and the indication of temperate marine conditions in BH 77/2 are in fact the products of erosion of underlying marine sediments and incorporation of this material into facies C^4.

The absence of bedded and laminated sand and gravel sequences associated with sub-facies C2 may be due either to the rapid retreat of the ice margin preventing the development of a subaqueous fan (Edwards and Foyn, 1981) or to gaps in the overall borehole control. A further alternative is that the sediments were deposited from a floating ice shelf, a possibility that will be discussed further in the summary.

5.5.4 Facies D^4

This facies occurs in boreholes 77/2, 75/29, 75/33, 81/27, 81/37 and 82/16, and in a limited number of vibrocores in the Fladen area. Its distribution is very similar to facies C^3, although in the Bosies Bank area it was absent from the majority of the recovered cores.

The facies is most commonly associated with a channel-fill or basin-fill type seismic facies within sequence 4. Reflector configurations are typically chaotic or structureless. Facies D^4 most commonly occurs between 130-150m below sea level, although in the central Witch Ground Basin and Forties area it is restricted to depths below 190m (O.D).

The sediments consist of dark grey (2.5Y, N4), firm, faintly laminated or massive mud with dropstones and rare silt lenses Individual laminae are commonly defined by black monosulphides. Bed thicknesses vary from 4-30m and contacts are generally gradational. Most commonly the facies occurs either overlying or interbedded with facies C^4, the exceptions to this are seen in boreholes 75/33 and 77/3 (Fig. 4.2) where facies D^4 forms the lowermost unit in the succession.

Grain size analysis of facies D^4 (Fig. 5.8) revealed that the sediments are predominantly of sub-facies $D1^4$ type, except in BH 75/33 where both $D1^4$ and $D2^4$ types occur. The sediments generally show a better degree of sorting (S.D. 2.6-2.8) relative to facies C^4.

Generally micropalaeontological evidence is indicative of hyposaline, shallow arctic water conditions cut off from the North Atlantic Drift System (bio-unit M). The exception to this occurs in BH 81/27 where a mixed assemblage is thought to be of a similar origin to that described for facies B^4.

Hand vane measurements on this facies gave typical shear strengths of between 40-60 KN/m^2. However, in boreholes 75/33 and 77/3 much higher values 75-150 KN/m^2 were recorded from the basal units of facies D^4 (Figs. 2.51-2.52).

Interpretation

Sediments in this facies are consistent with deposition in an intermediate to low energy glaciomarine environment characterised by suspension settlement from overflow plumes. The overall upward fining nature of the sequence, from facies C^4 to D^4 reflects a migration of this environment towards the retreating ice front. Interdigitation of facies C^4 and D^4 in boreholes 81/27 and 81/37 is the result of a fluctuating or surging ice margin prior to the final retreat.

The presence of facies D^4 at the base of the sequence in boreholes 75/33 and 77/3 is thought to relate to an earlier glacial episode. This interpretation agrees with the fact that it underlies facies C^4 and E^4 in 75/33 and 77/3 respectively, and also with the much higher shear strengths recorded from these sediments. However, no other evidence of this early glaciomarine stage was observed.

5.5.5 **Summary**

Marine and glaciomarine facies associated with seismic sequence 4 infill an irregular erosion surface which formed during a low sea level stand in the late Saalian. In the Witch Ground Basin the infill forms a continuous blanket or drape like unit. However, outside this area the base is highly irregular and the sediments infill a series of isolated or connected channel-like features.

In the channel-like features the sediments overlying the basal erosion surface record a transition from an arctic marine to a temperate marine environment concomitant with a marine transgression and penetration of the North Atlantic Drift into the area. This period of amelioration is dated as Eemian in age. Seismic profiles of certain channel infills, described in chapter 2, similarly record increasing water depths and a transition from bottom dominated processes to sedimentation from suspension. Other channel infills are cut by reactivation surfaces and are characterised only by configurations consistent with either homogenous sedimentary sequences deposited from suspension or mass movement deposits.

A subsequent transition to harsher climatic conditions is ascribed to the onset of the Weichselian stage. The deposition of stratified glaciomarine diamicts over a wide area during this period is difficult to explain given the absence of subglacial material; and the fact that good borehole and vibrocore control for this sequence precludes the possibility of gaps in the sequence. The two following possibilities are therefore suggested:-

i. that subglacial sediments were deposited on the western margin of the study area by a tidewater ice sheet, and this material was subsequently eroded away. This is partly supported by seismic profiles (Figs. 2.5 & 2.6) which show that along the western margin of the study area seismic sequence 4 is cut out and overstepped by the overlying sequences.

ii. that the glaciomarine sediments associated with seismic sequence 4 were deposited from a floating ice-shelf whose grounding line lay outside the study area.

However, it is doubtful if water depths in the area were great enough to support an ice shelf regime (Powell, 1984); especially on the basis of micropalaeontological data presented in Chapter 3. It is therefore suggested that stratified diamicts and glaciomarine muds with dropstones were deposited in broad depressions (in the area of the Witch Ground Basin) and channel features, which bordered on a tidewater front. The eastward limit of the ice front was possibly located along the western margin of the study area, whilst the interbedded nature of the diamicts and glaciomarine mud may be the result of oscillations of the ice front. Mass movement processes, especially slumping, were probably ubiquitous in many of the channel features. The overconsolidated nature of some of these sediments suggests that they were subsequently subjected to sub-aerial exposure and desiccation.

Grain Size Analysis

Fig. 5.9 shows triangular plots of sediments from the main facies described in chapters 4 and 5. Sediments from facies A^2-E^2 were not plotted because of the small number of samples analysed. From these diagrams it is obvious that respective facies types tend to define specific fields, and if similar facies of different ages are superimposed on these it is possible to delimit a general field for each of the facies, A-E (Fig. 5.9). Similarly, scatter plots of various statistical parameters (Fig. 5.10) show a tendency for individual facies to define their own fields, whilst grain size distribution curves, depicted previously, highlight the textural differences between facies and sub-facies.

Similar observations by other workers have sometimes been used to ascribe environmental regimes or gentic processes to the various textural fields (Hicock et al., 1980; Easterbrook, 1981). It would therefore be tempting to describe the fields shown in Fig. 5.9 in these terms, of say, proximal glaciomarine, distal glaciomarine, sub-glacial till and marine. Furthermore, various statistical parameters, especially mean grain size (x), standard deviation (S.D.) and skewness (SK.), have been used to define not only environmental regimes but specific processes and hydraulic conditions (Hicock et al., 1980; Sly et al., 1983). Sly et al. (1983), for example, attempts a universal comparison of water-lain sedimentary deposits based primarily on x, S.D.,SK., and kurtosis (k) and correlates these statistics with integrated hydrodynamic conditions. Similarly various textural parameters have been used to determine diamict genesis, provenance and distance transported (Dreimanis & Vagners, 1971; and Buller & McManus, 1973).

In fact, certain recent work has shown that in some cases assumptions based solely on textural criteria, irrespective of other information, can often be misleading. Thus Singer and Anderson (1984) provide evidence which "places a major limitation on the use of mean grain size and skewness in palaeoceanographic studies, particularly with regard to silt and clay." Haldersen (1981) describes the limitations of assumptions pertaining to till deposits based purely on textural data whilst a number of workers (Kranck, 1975; McCave, 1984) have studied the effects of flocculation in the sedimentary environment and its relationship to hydrodynamic behaviour and dispersed particle size distributions.

From the above, it was decided the particle size data could not be used to provide detailed information into the palaeohydraulic regime or to give an objective classification system which could be used to identify environments of deposition in other study areas. The data does, however, provide an objective documentation of material properties. It can also be used to corroborate interpretations based on a variety of data and on the textural characteristics of various facies.

Fig. 5.10 shows that the diamicts in Facies A and C constitute the most poorly sorted sediments within the Pleistocene sequence. In most cases the shape of the distribution curve for the two facies is also similar; both display non log-normal distributions with breaks in the curve at 4 phi and 2 phi, although the breaks in facies A tend to be smoother. However, facies A is interpreted as a sub-glacial till which according to certain workers (Buller and McManus, 1973) should display a uniform, log-normal grain size distribution. In fact, only facies A^5 (Fig.

4.12) displayed such a distribution; the remaining sediment of facies A type displayed distinctive sub-populations. This apparent discrepancy can be better explained by a combination of factors including:- i, the processes of glacial crushing and abrasion (Dreimanis & Vagners, 1969), ii, the loss of particles from the system by deposition, iii, textural and lithological inhomogeneities of the source rocks, iv, erosion of fine Pleistocene sediments. Given these possibilities it is not surprising that sub-glacial till only shows a log-normal grain size distribution in a few examples (Haldorsen, 1981).

Texturally diamicts in facies C (sub-facies C2) differ from facies A by their greater content of fine material, as displayed in the triangular plots in Fig. 5.9. However, any attempt to analyse in detail breaks in the grain size distribution curve is probably futile given the significant contribution from ice-rafting. What is pertinent is the coarse nature of the ice-rafted material in facies C and especially the presence of numerous dropstones up to 60mm in maximum dimension. This can be taken to indicate rafting by glacial ice rather than sea ice, the latter generally only contributes silt and clay size material to the sedimentary environment (Barnes and Reimnitz, 1974, and Clark and Hanson, 1983).

Both sub-facies C1 and C3 are typically better sorted than C2 (Fig. 5.10). Samples from C1 are limited although the significant lognormal population between 2 and 5 phi is probably indicative of depositional processes dominated by saltation and bed load transportation (Visher, 1969). This is consistent with those elements of sub-facies C1 which were attributed to depositon from sediment laden underflows. Due to the rapid accumulation of such sediments the contribution from ice rafting was previously described as being much less significant relative to the more poorly sorted diamicts.

Sub-facies C3 was only positively identified at one stratigraphic level, $C3^6$, where it is interpreted as a lag deposit resulting from the erosion and winnowing of glacial and glaciomarine sediments. Detailed analysis of the formation of such deposits by Singer and Anderson (1984) has suggested that unless the parent material is mixed, by say bioturbation, velocities of at least 16cm s^{-1} are required to erode even the minimum of fine material. If this work is applied to the formation of sub-facies C3[6] then either the parent glacial and glaciomarine sediments were mixed by bioturbation or current palaeovelocities were significantly greater than 16cm s^{-1}. Given the absence of any evidence of bioturbation from sediments in facies A-D it must be assumed that reworking of the sediment took place under high current velocities.

Sub-facies D1-D3 all consist predominantly of fine grained sediments which show an upward increase in sorting from D1-D3 (Fig. 5.10) and a similar decease in the amount of material $< 62\mu m$. This pattern is principally attributed to an upward decrease in the amount of ice-rafted material in what is essentially a homogenous 'background material' of silt and clay. The predominance of the silt/clay sub-population is interpreted as reflecting deposition predominantly from suspension. The apparent poor sorting of this component is probably misleading as the material is likely to have settled as flocculated aggregates and not as individual particles (Kranck, 1975). Subsequent disaggregation of the flocs during analysis leads to a false impression of the nature of the original sediment, especially with regards to the mode and the range of particle sizes.

Material similar to facies D is described from the Arctic Ocean and attributed by Clark and Hanson (1983) to sedimentation from sea-ice rafted material. However, the thicknesses attained by facies D probably preclude sea-ice as a primary mechanism and the whole sequence is more consistent with suspension settlement from sediment laden over-flow plumes.

Facies E contains a complex series of sediment types, and hence textures, which are most completely represented in sub-facies E1[8] and E5[8] (Fig. 4.19). Texturally the sediments can be immediately separated from facies A-D by their greater degree of sorting for a given mean grain size (Fig. 5.10). A primary cause of this must be the absence of ice-rafted material in Facies E. Sediments from facies E are also generally positively skewed reflecting the combination of silt and sand, thereby contrasting with the negative skewness and silt in clay combination of facies D (Fig. 5.10).

The broad and relatively flat distribution of the silt/clay sub-population in sub-facies E1-E3 is probably the product of analysing dissagregated flocs, as described for facies D. The obvious increase in silt in facies E, relative to other facies, as shown in Fig. 5.9 is more difficult to explain, although various possible mechanisms were outlined in Chapter 4. These included i) a source of silt sized aeolian material, ii) winnowing of the clay fraction initiated by pockmark or bioturbation activity.

Sub-facies E5^8 was interpreted as a palimpsest lag deposit characterised by a significant, log-normal, fine sand component. Erosion and winnowing of the silt and clay sized material was probably initiated by mechanical mixing of the sediment by bioturbation allowing for removal of the fines in current regimes as low as 5 cms^{-1} (Singer and Anderson, 1984).

Interpretation of sub-facies E5 from other stratigraphic levels, namely E5^1 and E5^3, is more ambiguous as there is no definitive evidence that these represent lag deposits rather than for example storm beds.

CHAPTER SIX

FACIES MODELS AND BASIN DEVELOPMENT

6.1 Introduction

The following chapter presents first, a facies model for the late Weichselian sequence in the study area and the implications of this model with regards to the regional palaeogeographical and palaeoclimatological information. This is succeeded by an assessment of pre-late Weichselian basin development and sedimentation, its relevance to the above described facies model, and also to regional palaeogeographic and palaeoclimatological information. Finally the overall development of the Pleistocene sequence is then compared with both contemporary and ancient glaciomarine models.

6.2 A late Weichselian Facies Model

When considering a model for the previously described late Weichselian sequence it is necessary to first outline a number of complicating factors pertinent to the study area:-

i. Sea Level. The intracratonic setting of the North Sea basin, its subsidence history and the interplay between glacio-isostatic, hydro-isostatic and eustatic movements have given the area a unique sea level history (Jelgersma, 1979; Sutherland, 1984). On this basis any model for the area can only work in terms of relative sea level variations and not absolute sea level.

ii. Climate. A number of workers have described the difference between mid-latitude and high latitude ice sheets regarding their response to climatic variations (Sissons, 1981; Boulton et al., 1985) and the effects that these variations are likely to have had on the geological record. Mid-latitude ice sheets, as would have affected the North Sea, are intrinsically more active and more variable in response to climatic changes whilst high latitude sheets are more sluggish in their response to change (Boulton et al., 1985). The result of this is that attempts to correlate events in the North Sea with the North Atlantic oxygen isotope record must be done with caution. The latter primarily reflecting the history of the major ice sheets and associated sea level changes.

iii. Regional Context. Inherent to the development of a successful model is a consideration of the late Weichselian sequence in areas immediately adjacent to the study area. Unfortunately, previous work on these areas is often ambiguous and rarely is there a concensus of opinion on the age of a particular sequence or even its mode of deposition. This is typified by the present lack of agreement regarding the extent of the last Scottish ice sheet (Sissons, 1981) a factor which will be discussed more fully in section 6.3.

iv. Glacial ice source and dynamics. The type of glacial ice source (ice sheet or valley glacier) the nature of its terminus (ice shelf or tidewater) and the thermal condition at its base (temperate or cold) have been widely recognised as the primary factors controlling glacial sediment input into the environment (Boulton and Paul, 1976; Powell, 1984; Eyles et al., 1985). However, in the ancient glacial record the distinction between the above states has often proved problematic and, for example, Eyles et al (1985) point to the 'oversimplistic' application of ice shelf models to numerous ancient glacial sequences. A detailed reconstruction of glacial dynamics is beyond the scope of this study, although work by Boulton et al. (1985) predicts that a UK Continental Shelf ice sheet would have been characterised by a low slope surface on account of the underlying deformable sediment. A late Weichselian ice thickness of some 250m was also predicted.

v. Compositional Data. Given the thick sequence of the Pleistocene succession in the study area repeated cannibalism and reworking of the pre-late Weichselian sediments is likely to be significant and inferences based on compositional data, including micropalaeontology, must be made with care.

In the following model all the above have been carefully considered and accounted for. However, it should be stressed that they pose certain limitations which are presently unresolvable, as will be discussed later.

To facilitate an understanding of the sedimentary facies and their environments the study area was divided into two morphologically different areas, each with a distinctive sequence of facies associations. The northern area, essentially between 58° and 59° north, contains the greater part of the Witch Ground Basin and is characterised by outer basinal and outer littoral facies associations and, on the western flank of the basin, inner and marginal facies associations (Figs. 4.10 & 6.1a) . The terms inner and marginal relate to their assumed proximity to the ice front whilst the term facies association is used to describe sedimentary facies which tend to occur together (Reading, 1978).

The southern area, essentially between 56° and 58° north, consists of an inner facies association again located along the western flank of the area, and to the east marginal facies and channel facies associations (Fig. 6.1b).

6.2.1 Northern Facies Associations

The inner and marginal facies associations in this area together display an overall upward and westward fining, reflecting the westerly retreat of the ice and concomitant migration of the glaciomarine environment in the same direction.

At the base of the inner facies association the presence of subglacially deposited diamict (till), facies A^7, reflects the approximate extent of the last ice sheet; notably this association is characterised by its association with a large sub-aqueous moraine and a north-south trending channel, described previously. The marginal facies association contains no subglacially deposited material suggesting that it lay beyond the maximum extent of the last ice sheet. The outer basinal facies association is restricted to present water depths of greater than 140m and contains only finer glaciomarine sediments reflecting its distal nature relative to the ice front. A sparcity of evidence of glaciomarine processes in the outer littoral facies associations reflects a similar distal nature to the ice front. However, the shallower occurrence of this association, above present day waterdepths of 140m, allowed for the formation and preservation of sediments more typical of a shallow marine or intertidal environment.

Figs. 6.2a - 6.2d depict a simple model intended to explain the development of the facies associations. In Fig. 6.2a an ice sheet ending as a tidewater front is shown at its maximum extent, this was preferred to an ice shelf type termination for two reasons. First, ice shelf formation requires the presence of rigid constraints including a protected environment with anchor points, very high grounded-ice discharge and basal temperatures below the pressure melting point (Powell, 1984). Secondly, the lithofacies, their relationships and ther areal extent (tens of Km rather than hundreds - c.f. Fig. 4.9) are consistent with deposition from a grounded tidewater ice-sheet (Powell, 1984; Eyles et al., 1985). For example, stratified proximal glaciomarine sediments are generally associated with tidewater rather than ice-shelf environments (Powell, 1985). More problematic is the state of the grounded ice, that is below pressure melting point or at or near the pressure melting point, especially given the fact that the thermal regime of many modern ice masses is not well understood (Eyles et al., 1985). However, it is generally accepted that the smallest amount of glaciomarine sediment, especially that fraction associated with subglacial meltwaters, is produced by frozen base grounded glaciers (Anderson, 1983, 1984; Powell, 1984; Eyles et al., 1985). Also the occurrence of a large sub-aqueous moraine may reflect high ice velocities and activity indices (Boulton, 1986) and the geographical setting of a British ice sheet would have been more consistent with a maritime rather than a continental state (Boulton et al., 1985). It was therefore decided that the ice sheet base was most likely at the melting point. Certainly the ice sheet base would not have been frozen during the final stages of advance and initial retreat when the subglacial facies, A^7, was deposited. It is therefore envisaged that the ice advanced forward to its maximum extent, depositing a cover of till at the margin (Boulton et al., 1977). At the ice limit the presence of a large sub-aqueous moraine necessitates either a period of maintenance or alternatively, slight readvances, of the ice front (Landmesser & Johnson, 1982; Boulton et al., 1985, Boulton, 1986). Subglacial meltwater streams may have carved out the north-south trending channel feature to the west of the moraine; and such channels generally develop normal to the ice-front (Valentin, 1955).

The relative sea level at this stage is difficult to establish, and although sea levels of -100m to 130m have been suggested for the last glacial maxima at 18,000 yrs BP (Jansen et al., 1979), the effects of glacial isostatic loading (Walcott, 1970, 1972) and the actual timing of the ice sheet advance must also be considered. Thus, although sea level reductions of even 100m would have sub-aerially exposed much of the pertinent area (Fig. 1.4) glacial isostatic depression is likely to have been in the order of several decametres (Jardine, 1979). Furthermore the likelihood of crustal transfer from beneath the depressed zone to more peripheral areas, namely the Witch Ground Basin, and the formation of a periperal bulge (Walcott, 1970, 1972) should also be considered.

Fig. 6.2a the distribution of the glacial and glaciomarine sediments at the onset of ice retreat. Sea levels in the immediate vicinity of the ice front were possibly similar to the present (80-110m), certainly depths any shallower would have precluded the freeboard necessary for iceberg rafting, for which there is extensive evidence. Such icebergs were probably relatively small, shallow draft bergs and growlers. Further east, in the more distal areas, water depths were probably much shallower than present due to both the absolute lowering of sea level and the formation of a peripheral bulge zone (Walcott, 1970; Oilon & Oldale, 1978). Thus, in the immediate vicinity of the ice front, commonly termed the ice-proximal (Powell, 1984) or proximal glaciomarine zone, subglacial meltwater streams discharging from tunnels beneath the ice sheet resulted in the formation of underflows (continuous turbidity currents), interflows and overflows, whilst turbidity currents (sensu lato) originated from mass movement processes or underflow currents (Mackiewicz et al., 1984). In addition to this, sediments transported by icebergs and growlers were eventually dumped on the sea bed as a result of iceberg melting, roll-over and splitting, and are best characterised by the occurrence of out-size clasts and dropstone structures. The resultant sediments, facies C^7, are therefore termed the proximal glaciomarine facies and consist of laminated and graded sands, turbidites, cyclopel units and stratified and massive diamicts. The sand units and turbidites reflect deposition from continuous underflow currents and turbidity currents respectively, whilst cyclopel units were deposited primarily by suspension from fluctuating sediment overflow plumes. Diamict units are most likely the result of deposition from a combination of processes including iceberg rafting, underflow currents and overflow plumes. Mass movement processes would also have been common in this environment.

The approximate extent of the proximal glaciomarine zone beyond the ice front maximum is some 40-50km as shown by the distribution of facies C^7 in Fig. 4.9. This is at variance with the limits proposed by Andrews and Matsch (1983) and Powell (1984); 1 km and 5 km respectively. However, as Eyles et al. (1985) state, distance alone is not necessarily significant and indeed Boulton and Deynoux (1981) prefer to define the proximal glaciomarine zone as one affected by "strong bottom currents, generated density instability due to mixing of different water masses." Further to this, it was noted in chapter 4 that hypersaline conditions may well have prolonged the effects of subglacial meltwaters. Also, it is possible that the products of the inner iceberg zone of Powells (1984), with an extent of 10 km, relate to the massive and stratified diamicts of the proximal glaciomarine zone defined here.

Away from the immediate ice front ($>$50km) suspension settlement processes from both sediment overflow plumes and debris-laden icebergs predominated, resulting in the deposition of layered and massive muds with rare dropstones, sub-facies $D1^7$ and $D2^7$, in the outer basinal association. All the evidence suggests that the palaeowater depth in this basin reached a maximum of only 60m in the centre, a relative lowering of some 120m compared to the present day (Fig. 1.4), and generally averaged between 20-40m. An important implication of this is that weakly stratified or massive, fine lithofacies can be generated in comparatively shallow water where there is little or no sorting by traction currents or where large volumes of suspended sediment are available. Where present, the weak stratification in these sediments is the result of fluctuations in meltwater supply and variations in the areal distribution of overflow plumes. The rare occurrence of dropstones in sediments from this environment reflects the fact that many icebergs from temperate tidewater ice sheets deposit their load within the proximal glaciomarine environment (Domack, 1982; Powell, 1984). The absence of bioturbation structures in these sediments is apparently typical of high latitude fine grained sediments (Barnes & Reimnitz, 1974).

Around the flanks of the basin and isolated topographic highs within the basin (Fig. 4.10), palaeowater depths of less than 20m ensured that the environment was dominated by wave and possibly tide generated currents, resulting in the formation of the well bedded and laminated sands and muds, facies E^7, seen in the outer littoral facies association. These sediments presently occur at water depths of between 135-140m below sea level (Fig. 6.3).

During the winter season (Fig. 6.2b) it is suggested that the area was affected by a cover of sea ice which was generally shorefast at water depths of less than 20m, whilst between 20 and 50m pressure ridge keels scoured the sea bed forming a distinctive micro-relief. At the ice front calving was probably inhibited by the sea ice in a similar manner to that described by Boulton (1986) for a tidewater glacier front off Spitsbergen. A cover of sea ice would also have affected sedimentation rates and the environmental conditions at the depositional interface. Thus, as suggested in Chapter 4, the monosulphide banding common to the distal glaciomarine facies may be a response to seasonably induced fluctuations.

During the spring melt over ice flows may have transported fine material across the ice surface and drained through strudels creating scour pits and further contributing to the irregular sea bed relief. The rafting of debris by sea-ice is an unknown quantity although it is likely that some fine material would have been transported via this mechanism (Barnes & Reimnitz, 1974).

The depositional environment associated with the development of the lowermost sediments in the outer basinal and outer littoral facies associations is therefore characterised by a reduction in the influence of glaciomarine processes and represents an interdigitation of distal glaciomarine and arctic shallow marine processes. As such, it partly equates to the distal glaciomarine environment of Eyles et al. (1985), the iceberg zone of Powell (1984) with a limit of a few 100 km's, and the intermediate/distal, shallow glaciomarine environment of Andrews and Matsch (1983) with a lateral extent of 1000km. In this particular case, the term distal glaciomarine environment is preferred, partly on the basis of the maximum distance separating facies A^7 and facies D^7 and D^8 (ca. 150 km). However, it should be stressed that the recognition of facies D^7 and D^8 type sediments alone does not imply deposition in a distal environment and, for example, proximal mud belts may develop much closer to the ice front (Powell, 1981; Osterman and Andrews, 1983) due to very high sediment concentrations and an absence of underflow currents. It is therefore important to study the lateral facies relationships and their implications before making such assumptions.

Downslope mass movement deposits, namely slump units (Facies B^7), are associated with the proximal and distal glaciomarine environments. In the latter they generally occur on the slopes and axes of open channel systems, as described in chapter 2, where slumping was possibly initiated by freshwater flushing and periglacial mass movement processes. In the proximal glaciomarine environment slumping could have been instigated by a variety of mechanisms including rapid sedimentation and overloading, iceberg ploughing, and wave action. Previous studies of similar lithofacies have often preferred the term resedimentation or redeposition by gravity flowage (Powell, 1984; Eyles et al., 1985). However, such terminology ignores the stricter and generally accepted classification of mass movement products (Nardin et al., 1979a), whilst also resulting in confusion when discussing the reworking and redeposition of glacial and glaciomarine sediments by, for example, shallow marine processes.

Fig. 6.2c models the development of the sedimentary sequence during the main phase of ice retreat. The most notable effect of this is a westward migration of the various depositonal environments in the direction of ice retreat. Thus the marginal facies association displays an upward gradation from facies typical of a proximal glaciomarine environment to those associated with a distal glaciomarine environment. Similarly, the inner facies association reflects a change from a subglacial to a proximal glaciomarine environment. Lithologically, this relationship is principally characterised by an overall vertical and eastwards fining of the sediments; the sand content clearly decreasing from subglacial and proximal glaciomarine environments to the distal glaciomarine environment.

It is suggested that as the ice front retreated, by iceberg calving rather than a gradual downwasting, proximal and distal glaciomarine sediments were rapidly deposited in a glaciomarine environment which contiued to encroach on the retreating ice-front.

As the ice retreated from the study area the effects of isostatic rebound and subsidence of the forebulge area must eventually have significantly altered the original configuration of the depositional environments, resulting in a rise in the relative sea level in the Witch Ground Basin and a sharp decrease in the recently deglaciated areas. This effect would have been enhanced if the ice sheet retreated prior to the volumetric wasting of the main Scandinavian and Laurentide ice sheets and hence prior to the main lateglacial transgression. Two factors support the above suggestions. First, in the Witch Ground Basin itself, dinoflagellate cyst evidence from facies D^8 indicates the continued presence of at least a seasonal cover of sea ice. Geologically, however, there is no evidence for this and the sediments of sub-facies $D2^8$ and $D3^8$ display an acoustically well layered and uninterrupted profile. Therefore, it is suggested that water depths over much of the Basin precluded sea ice scouring and as such probably ranged from at least 40-80m. The buried scoured surface, however, is thought to have been formed in water depths of only 20-40m. Secondly, seismic profiles show that in the Witch Ground Basin the acoustically well layered sediments, facies D^8, only occur below present day water depths of 140m in the central Witch Ground Basin (Fig. 6.3), whilst further west the sediments occur in water depths as shallow as 110m. It is therefore suggested that this difference reflects a combination of subsidence and uplift in the east and west respectively. The idea that the depth difference could have been caused purely by a marine transgression is rejected because of the divergent nature of the reflector configurations, the occurrence of sharp cut-off depths in the Witch Ground Basin and the absence of an erosion surface.

Therefore during this period the western flanks of the basin would have been sub-aerially exposed and subjected to denudation processes, the eroded sediments from these areas probably providing much of the material for the sediments in the Witch Ground Basin. This area having itself become virtually cut off from the retreating ice sheet some distance to the east.

At this stage (Fig. 6.2c) the Witch Ground Basin therefore became a semi-enclosed hyposaline arctic marine environment, partly fed by fluvial erosion of the surrounding land mass and characterised by a seasonal cover of sea ice. The increasingly isolated nature of the basin, as the ice retreated west, is reflected by the upward transition of distal glaciomarine muds into arctic marine muds. Unknown variables include the possible contribution of large amounts of fine grade material from a large river system to the south and also the influence and proximity of the Scandinavian ice mass. However, detailed analysis of vibrocores from the area have provided an accurate picture of the changing environmental conditions from (Fig. 6.2d). Thus, in Fig. 3.3 the change from a distal glaciomarine or shallow arctic marine (facies D^8) to a more temperate marine environment (facies E^8) is clearly recorded. This represents a more regional change in climatic influence concurrent with the final rapid retreat of the Laurentide and Scandinavian ice sheets, and hence occurred well after the main retreat of the Scottish ice sheet as will be explained later.

Associated with the change from an arctic marine to marine type environment is a distinctive change in lithology. As seen by the transition from the arctic marine muds of sub-facies $D3^8$ to the temperate marine muds and palimpsest sands of facies E^8. The temperate marine muds are interrupted by a limited return to arctic marine conditions, as determined from the dinoflagellate cyst record (Fig. 3.3), prior to the final onset of present day temperate marine conditons. An upward increase in the sorting and the silt content of these sediments is attributed to the erosion of aeolian sediments to the south of the study area and to pockmark activity, as discussed in appendix 1. The extensive lag of palimpsest sands, found over most of the study area, records the initial low sea level stands prior to the establishment of present day water depths.

Evidence of sedimentation rates is only available for distal glaciomarine and temperate marine sediments, and is based on dating the dinoflagellate record (Fig. 3.3). Rates of about 0.8m/Ka are calculated for late glacial temperate marine sediments, assuming a duration of 2000 yrs, and 2.4m/Ka for distal glaciomarine sediments, assuming that they were deposited between about 18,000 yrs and 13,000 yrs BP.

6.2.2 **Southern Facies Associations**

The inner facies association in this area is similar to that seen in the northern area with the exception that no distal glaciomarine muds (facies D^8) are present. In the marginal facies association the succession is distinctly different (Fig. 6.1) being composed predominantly of facies C^6, which consists of stratified sands, gravels and thin diamicts. An uppermost glaciomarine unit in this succession consists of loose sands and gravel which commonly displays bi-directional cross stratification.

Similarly the channel facies association is unique to this area and consists of the channel infill units of facies E^8. As shown in Fig. 1.4 a further critical difference between the northern and southern areas is the difference in present day bathymetry and regional sea bed topography. It is probably this factor which, more than anything, was responsible for the development of the different facies associations and the absence of distal glaciomarine sediments in the southern area.

Figs. 6.4a - 6.4c depicts the evolution of the sedimentary facies associations, commencing with the distribution of glacial and glaciomarine sediments at the onset of ice retreat. The maximum extent of the ice front in this area is easily distinguishable by the distribution and eastward limit of subglacially deposited sediment, facies A^5, and a distinctly hummocky topography. Large scale deformation structures, formed by ice push mechanisms, similarly delimit the extent of the last ice sheet whilst deep, open channels cut into bedrock were probably carved out by violent subglacial streams.

The sequence is interpreted as follows. An ice sheet encroached on to the western edge of the area depositing a thick cover of subglacial lodgement till and deforming the underlying sediments in the ice front zone. Glacio-isostatic depression of the ice covered area and its immediate vicinity resulted in a marine transgression towards the ice front and water depths of between 60-80m immediately adjacent to the ice front but decreasing to the east, away from the zone of depression. This is supported by the fact that a tilted planar erosion surface, described in sections 2.4.6 and 4.2.3, dips away from the ice front limit at 1.0/km in a north-easterly direction (Fig. 2.32); although this gradient gradually decreases away from the ice limit. A similar gradient of 0.94m/Km was determined by Cullingford & Smith (1980) for their second oldest raised shoreline and is attributed to glacio-isostatic uplift following depression of the crust by a large ice-sheet. Therefore on the basis of a 1.0m/km gradient it is possible to suggest that isostatic depression at the ice limit would have been in the order of 120m, assuming a eustatic sea level lowering of 80-100m at this time, and given the present day water depth of 50-60m.

In the proximal glaciomarine environment strong underflow currents emitting from the ice front swept along the sea bed depositing a thick sub-aquaeous fan which received additional material in the form of sediment rafted by small growlers, although the shallower water depths probably precluded ice-berg rafting as an important process relative to the northern area.

The proximal glaciomarine environment in this area, relative to the northern area, was therefore characterised by shallower water depths and probably stronger and more extensive meltwater processes. In addition the eastward shallowing of the depositional area precluded the extensive rafting of material away from the ice front whilst allowing a reworking of the shallower areas of the fan by wave and tide driven processes. The thickness of the fan would have been limited only by the water depth and the erosion of the fan surface above the wave base. Evidence of shallow water depths and the development of a sub-aqueous fan further suggests that the ice sheet ended as a tidewater. However, it should be noted that it is difficult to differentiate between sedimentary facies deposited adjacent to a tidewater and those deposited by a limited glacier-tongue regime (Powell, 1984).

In the eastern half of the area the combination of eustatic sea level lowering and isostatic compensation (peripheral bulge formation), relative to the ice depressed zone, resulted in the emergence of a large area of the sea bed. Possibly by as much as 40m above relative sea level during the period of maximum eustatic sea level lowering (-100 to -120m). The combination of rapid emergence and the extension of the North European plain river system to the south east of the area (Sissons, 1981; Long & Stoker, 1986a) resulted in the erosion of a complex channel network with an approximate north-south trend. Evidence of mass-movement processes on the slopes of some channels is apparent on seismic profiles and as such may have been induced by periglacial activity.

However, it should be stressed that two main problems still exist with the mechanism of channel formation. First, there is no evidence that erosion of these channels was initiated during the late Weichselian, indeed a number of workers have suggested that relative sea levels in the North Sea basin may have been lower during periods of the Lower and Middle Weichselian (Jardine, 1979; Oele & Schuttenhelm, 1979). Secondly, a number of channel features infilled with late Weichselian sediments are characterised by base levels some 300m below present sea level (Figs. 2.5 & 2.6) and as such could not have been formed solely by sub-aerial erosion during the late Weichselian. On the basis that this area was ice free during the period, and hence that these deep channels could not have been eroded by ice or subglacially confined meltwater, it is possible that they were formed at an earlier stage in the Pleistocene. Certainly the Saalian ice cover was much more extensive than during the late Weichselian, as will be discussed later, and subglacial or ice front channelling in the area would therefore be consistent with this age. Once formed such channels would then have remained open in a similar manner to the present day open channels prior to their subsequent infilling during the final stages of the late Weichselian. Alternatively the overdeepened channel profiles may have resulted from catastrophic meltwater discharges although this would have involved the presence of an ice front in close proximity to the southern margin of the study area.

During the winter season a cover of shorefast sea-ice (Fig.

6.3b) probably restricted processes in the west although small sub-ice channels are typically developed in this environment (Dupre, 1982). Further east, there is evidence that the channel flanks were scoured, possibly by river ice, although the timing of this event is uncertain. Fig. 6.4c depicts the development of the facies associations as the ice retreated to the west. The initial shallow water depths and ensuing isostatic rebound precluded the development of a distal glaciomarine type environment and if sea-ice was present during the later stages of retreat, there being no geological evidence for it, then it was probably shorefast during the winter season. During the summer season isostatic uplift probably subjected much of the fan surface to shallow marine reworking resulting in the formation of a layer of reworked glaciomarine sediments. Interestingly, west of the study area the presence of fine glaciomarine muds, the St Abbs Formation (Thomson & Eden, 1977) and Errol beds (1975) may suggest higher relative sea levels as the ice retreated from the present day coastline. In the east, isostatic relaxation and an eventual rise in sea level resulted in an eastward transgression of the shallow arctic sea and the depositon of generally thin patches of arctic marine muds.

Rapid submergence of the channelled areas was initially accompanied by the sedimentation of muds from suspension in arctic, hyposaline waters. Sediments initially deposited in this environment are thought to equate to the deposition of the distal glaciomarine muds, facies D^8, in the Witch Ground Basin. A subsequent switch in sedimentary proceses to current dominated deposition is thought to be consistent with a shallow, sub-littoral or intertidal environment and the depositon of marine and arctic marine muds in the Witch Ground Basin. The switch in sedimentary environments is clearly delimited on seismic profiles by a change from a lower, basally concordant, reflector configuration to an upper seismic facies characterised by downlapping and onlapping configurations. Rapid erosion of the surrounding inter-channel areas at this stage allowed for high sedimentation in the channels and the maintenance of shallow water conditions despite the rising sea level. Reflector configurations suggest that the sediment supply was predominantly from the west or south west. However, many of the channel infills do not comply with the above rather simplistic explanation and consist either wholly of one reflector type configuration or alternatively of a complex sequence of configurations, therefore suggesting that sedimentary environments and deposition rates were not consistent over the channelled areas and a distinctive pattern has so far not been identified. In addition a number of channelled areas remained open or were only partly infilled suggesting that these areas were starved of sediment or scoured by tidal currents, although it was not possible to make any conclusive explanation.

Subsequent to the infilling of the channel areas and the transition from arctic marine to temperate marine muds in the inter channel areas, reworking of the sea bed by wave and tidal generated processes when the sea level was still below its present level resulted in the formation of an extensive palimsest sand prior to the establishment of present day water depths. The relict nature of these sediments is confirmed by their widespread occurrence over the area in water depths of up to 140m.

6.3 Late Weichselian Palaeogeography & Palaeoclimatology

6.3.1 Extent of the last ice sheet

In the previous model the main Scottish ice sheet encroached on to the western edge of the study area, ending as a tidewater front. The eastern limit of the ice front in the southern area (Fig. 6.5) clearly agrees with that established by Thomson and Eden (1977) Fig. 1, and subsequently referred to by a number of workers as the terminal edge of the last Scottish ice sheet (Sisson, 1981; Sutherland, 1984).

The fact that the presence of an ice sheet in the southern area produced a marine transgression onto the ice front had previously been tentatively suggested by Sissons (1981) and the evidence provided here corroborates this. In addition the existence of a large expanse of sub-aerially exposed sea bed in the eatern half of this area was proposed by a number of workers including Jansen (1976), Jardine (1979), and Sutherland (1984), whilst Sissons (pq 12, 1981) established the extent of the glacial shoreline using a projected gradient of 0.94m/km from results present by Cullingford and Smith (1980). In fact, Sissons (1981) showed that the projected shoreline, supposedly marking the eastward extent of the glaciomarine environment, occurs well to the west of the actual maximum extent of the sub-aqueous glaciomarine fan. As this projection is based on a relative sea level of -100m, it is suggested that the relative sea level in the immediate area at this time was not as low as this.

Thus, the eastward extent of the last ice sheet in the southern area is clearly delimited, whilst a glaciomarine sea shallowed eastwards away from the ice front, the shoreline being approximately delimited by the eastward extent of sediments deposited in a shallow glaciomarine environment. The present day bathymetry along this shoreline (90-100m below sea level) suggests that the sediments were not deposited during the maximum eustatic sea level lowering (-100 to -120m). Two possible explanations for this are either that the maximum ice sheet limit occurred prior to the maximum eustatic sea level lowering or alternatively the effects of glacio-isostatic depression were more extensive than previously thought.

In the northern area previous work regarding the extent of the last ice sheet is generally lacking and hypothetical predictions are commonly ambiguous. Some of the earlier models for the area predicted that the whole of the North Sea Basin was glaciated during the late Weichselian and that a zone of confluence existed between the Scottish and Scandinavian ice sheets (Boulton, 1977; Holmes, 1977; Eden et al., 1978). Such assumptions were based on a series of eronneous facts:- i, that the last Scottish ice sheet was deflected northwards, or northwesterly, by a Scandinavian ice sheet in the North Sea Basin (Hoppe, 1974). ii, that the series of open channels in the area must have been formed subglacially during the late Weichselian. iii, that a series of north-south trending ridges on the flanks of the Witch Ground Basin are late Weichselian moraines. iv, that the overconslidated nature of the sediment over parts of the study area was indicative of late Weichselian glacial overriding.

In fact, more recently workers studying north-east Scotland's late Weichselian (late Devensian) sediments and morphological features have concluded that Orkney was not covered by the Scottish late Weichselian ice sheet (Flinn, 1978; Sissons, 1981; Sutherland, 1981). In addition evidence described in chapter 2 and 4 suggests that, with the exception of a large channel in the south-west corner of this area, the open channels were not formed subglacially beneath late Weichselian ice.

An alternative view was proposed by Jansen (1976), as shown in Fig 1.12, which postulated a large ice-dammed lake in the northern North Sea. The lake being dammed by the Scandinavian ice sheet which crossed Shetland, but which was confluent with Scottish ice only to the north of the study area. However, Flinn (1978), Long & Skinner, 1985; and Skinner & Gregory (1983) have clearly shown that Scandinavian ice did not cross the vicinity to the immediate north of the study area. Also Jansen's (1976) interpretation of 'The Hills' (Fig. 1.11) as a late Weichselian end moraine formed by the last Scandinavian ice sheet is incorrect on two counts:- i, seismic profile of the Hills ridge suggests that the positive relief purely reflects the topography of the underlying sediments, ii, sediments recovered from the ridge are more typical of glaciomarine rather than subglacial facies.

Clearly, therefore, there is no present offshore evidence which discounts the likelihood of a late Weichselian ice sheet terminating as a tidewater front along the western edge of the Bosies Bank area, as the evidence suggests here (Fig. 6.5). Indeed a number of workers have now accepted that the Scandinavian and Scottish ice sheets were not confluent during the late Weichselian (Sissons, 1981; Sutherland, 1984; Boulton et al., 1985 and Long & Stoker, 1986).

However, further west on the north-east Scottish mainland the evidence is more debatable. For example, the eastern limit of the ice sheet suggested previously would probably necessitate the presence of late Weichselian ice in the outer Moray Firth and over Caithness and Buchan. This disagrees with the suggestion of an ice-free Buchan (Synge, 1977) whilst Sissons (1981) and Sutherland (1984) both disagree with a totally ice covered north-east Scotland (Fig. 6.5). Although, here again the evidence is not conclusive and Sissons (1981) notes only that on present evidence none of Synge's ice-free areas can presently be disproved whilst Clapperton and Sugden (1977) and Murdoch (1977) are in favour of a much more extensive Scottish ice-sheet in north-east Scotland with an offshore ice terminus. Similarly, Boulton (1977) suggests that the 'featureless Buchan' relates only to a lack of erosion beneath the ice sheet. Chesher and Lawson (1983) also provide evidence of an extensive cover of subglacial till in the Moray Firth area which is thought to be of late Weichselian age.

Given the erroneous nature of Jansen's and Jansen et al's (1976,1979) results the position of the Scandanavian ice front during this stage is uncertain. There is evidence that the ice did cross the Norwegian trench (Rokoengen & Rise, 1984), although its terminus is unknown. However, the absence of proximal glaciomarine sediments in the Fladen area suggests that the ice did not terminate in the immediate vicinity of the study area. Evidence from Sleipner (Sejrup et al., in press, Appendix 9), some 10 km to the east of the Fladen area, indicates only the presence of overconsolidated glaciomarine sediments of Weichselian age.

To conclude, I suggest that the majority of north-east Scotland, and the sea bed to the immediate east of the present day coastline (10-60 km offshore), was covered by late Weichselian ice and that a lobe of this ice sheet extended out onto the western edge of the Bosies Bank area (Fig. 6.5). The direction of ice flow is uncertain although Sutherland (1984) suggests that it was towards the north-north east and analysis of the clasts indicates that they are predominantly of very local origin. Evidence for this model is supported by a cover of subglacially deposited diamict and the presence of morphological features typical of glaciated terrain, whilst the lateral associations of proximal and distal glaciomarine sediments necessitate the presence of an ice sheet in the immediate area. However, it should be emphasised that the history and processes described previously for the northern area are a best fit model based essentially on seismic extrapolation and micropalaeontological data and that absolute dates from the area are sadly lacking. In addition, the poor quality of data from the Peterhead area precludes the direct correlation of subglacial sediments and morphological features in the northern and southern parts of the study area.

6.3.2 Deglaciation of the last ice sheet

A westward retreat of the ice sheet from the study area and the deposition of a sequence of glaciomarine sediments was succeeded by isostatic uplift and a marine regression in the previously glaciated area, with the establishment of the minimum relative sea level. Further east forebulge subsidence resulted in palaeowater depths of between 40-80m in the partly isolated Witch Ground Basin and the emergence of much of the surrounding sea bed. The implication of this is that the ice sheet retreated from the area prior to the wasting of the main Laurentide and Scandanavian ice sheets when eustatic sea levels were still at a minimum. This is consistent with two lines of evidence. First, that the Scottish ice sheet, because of its size, latitude, and low marginal profile would have rapidly responded to variations in atmospheric circulation and climate (Sutherland, 1984; Boulton et al, 1985). Second, oscillations in surface ocean temperatures and sea ice cover, regulated by the position of the Polar Front, would have affected the moisture available for the continental ice sheet (Ruddiman and McIntyre, 1981; Boulton et al., 1985). On this basis the suggestion of Sissons (1981) that rapid wastage of the Scottish ice sheet occurred prior to the main global deglaciation is consistent with the position of the Polar Front well to the south of Britain between 20,000 - 13,000 yrs BP (Ruddiman et al., 1977; Ruddiman & McIntyre, 1981) and the consequent lack of precipitation over Scotland.

Therefore it is suggested that the ice sheet retreated to the approximate vicinity of the present day coast line whilst the British Isles were still surrounded by Polar waters. Ruddiman and McIntyre (1981) indicate that this could have occurred between 16,000 - 14,000 yrs BP when the storm tracks were still displaced to the south of 50°N. However, it is thought that Scottish ice could have retreated prior to this, whilst to the south the ice was still advancing resulting in a asynchrony between the northern and southern ice margins (Sutherland, 1984). This is consistent with the early deglaciation of Tayside, prior to 14,000 yrs BP (Brown, 1980).

Over mainland Scotland retreat of the ice sheet possibly correlated with rising sea levels (Mitchell, 1977) and raised shorelines indicate that the westward penetration of the sea kept pace with the retreating ice (Jardine, 1979). These shorelines comprise a late glacial series formed as isostatic recovery exceeded eustatic sea level rise, dated at 13,000 yrs BP or older (Jardine 1979), and a post glacial series formed after the eustatic rise in sea level temporarily overtook isostatic recovery (Peacock, 1975). Similarly, raised late Weichselian glaciomarine muds occur 46m above present sea level and attain a maximum thickness of some 24m (Browne, 1980) suggesting that high late Weichselian sea levels preceded isostatic recovery along parts of the east coast of Scotland. These sediments are thought to be the time-transgressive equivalents of distal glaciomarine muds in the Witch Ground Basin and the St Abbs beds in the Firth of Forth and coastal areas (Thomson and Eden, 1977). Most of Scotland was probably ice free prior to 13,000 yrs BP.

The main deglacial warming and hence the rapid retreat of the Scandinavian and Laurentide ice sheets is thought to have occurred around 13,000 yrs BP with the Polar Front migrating to well north of the British Isles (Morner, 1973; Ruddiman & McIntyre, 1981; Jansen and Erlenkeuser 1985 and Kellog, 1985). In the submerged parts of the study area this was accompanied by a shift from an arctic marine to a more temperate marine environment, as reflected by the dinoflagellate record. In the emerged onshore sequence it is similarly marked by the transition from the Errol beds to the Boreal fauna in the Clyde beds (Peacock, 1975). However, as discussed in Chapter 3, benthic micro and macro faunal evidence from the North Sea and Norwegian Sea does not record a change to a temperate marine environment until around 10,000 yrs BP (Sejrup et al., 1980). A model was therefore suggested in which the environmental change at 13,000 yrs BP reflects only a shift from an arctic marine water mass to a stratified water mass, influenced by the North Atlantic Drift, with a strong thermal gradient and a temperate marine surface. The remainder of the water mass remaining some 2°-3°C cooler than present resulting in a so-called 'failed' interglacial (Peacock, 1983). A variation of this model suggests that the polar front did not retreat as far north as previously proposed and in fact lay between Scotland and Iceland after 13,000 yrs BP, and a purely seasonal influx of the North Atlantic Drift produced a seasonally high productive layer on an otherwise cold sea (Jansen et al., 1983).

In the Witch Ground Basin evidence was presented that temperate marine sedimentation was interrupted at about 11,000 yrs BP by a return to an arctic marine environment completely cut off from the North Atlantic Drift. This cold stage was equated to the Younger Dryas re-advance and Loch Lomond re-advance in Europe and Scotland respectively (Long et al., 1986, Appendix 8). This age was confirmed by the presence of an ash band deposited during this stage and correlated to the Vedde Ash, dated at 10,600 yrs BP (Jansen et al., 1983 and Mangerund et al., 1984). However, it should be stressed that this brief cooling had no affect on the lithofacies deposited in the study area and is only reflected in the dinoflagellate record. The actual cause of the Younger Dryas cooling was attributed by Ruddiman and McIntyre (1981) to the breakup and outflow of large ice shelves from the Arctic Ocean. The final establishment of temperate marine conditions at around 10,000 yrs BP, the onset of the Holocene, saw a rapid rise in sea level (Jardine, 1979) and the eventual development of present day conditions. In the study area this last transition is recorded by both planktonic dinoflagellate cysts and benthic foraminifera, and macrofauna.

In conclusion, the last Scottish ice sheet only encroached onto the margin of the North Sea Basin and retreated to the vicinity of the present day coastline before the main volumetric wasting of the Laurentide and Scandinavian ice sheets. Because of the complex interactions of glacio-isostatic, eustatic and, probably, tectonic effects absolute sea levels are impossible to calculate. However, the evidence does suggest that after the ice retreated much of the study area was subaerially exposed with the exception of a partially enclosed arctic marine basin.

6.4 Development of the Pre-late Weichselian succession

Contrary to the views of Holmes (1977) and Eden et al. (1978) the Pleistocene succession in the study area, and the North Sea Basin as a whole, is not composed of sediments of a predominantly late Weichselian age. In fact the reverse is true and over much of the basin sediments of a Lower Pleistocene age form a significant proportion of the succession. The succession is readily divided into a lower unit consisting predominantly of marine and deltaic sedimentary facies with an uppermost glacigenic sequence and an upper unit consisting mainly of glaciomarine sedimentary facies. The two are separated by an easily identifiable and extensive irregular unconformity (Fig. 6.6), described in Chapter 2, and ascribed to the Elsterian stade (Stoker et al., 1985). On this basis the sediments overlying the unconformity span from the Middle to Late Pleistocene, whilst those below it are of a Lower Pleistocene and early middle Pleistocene age, as confirmed by palaeomagnetic and micropalaeontological data in chapter 3. The following two sections review the development and stratigraphic architecture of the sequences above and below the Elsterian unconformity respectively.

6.4.1 Lower and early Middle Pleistocene History

The majority of the Lower Pleistocene was characterised by the deposition of sub-littoral sediments in a broad marine basin, producing a thick sequence of monotonous marine muds. The acoustic nature of the succession is generally consistent with this and suggests deposition in a stable, gently subsiding shelf environment therefore allowing for the accumulation and preservation of such a thick sequence (ca. 300m).

Following the initial deposition of marine muds a delta front system advanced onto the south-western margin of the study area and a rapid influx of sediment into the basin resulted in an overall marine regression and the emergence or near emergence of the basin flanks where intertidal type environments became established. This delta complex can be associated with the northward progradation of delta related facies through the southern North Sea (Zagwijn, 1979; Cameron et al., 1986). As such, the main source for the delta system lay to the south-east of the study area, extending back towards the north-west European river system (Stoker & Bent, in prep. Appendix 10). Further details of this delta system and the controls on sedimentation are given in Stoker and Bent (in prep.). Here, it is suffice to say that massive accumulations of sediment, both in this area and the southern North Sea, during this period, were probably associated with rising sea levels, but accompanied by a rate of sedimentation faster than the rise in sea levels resulting in an overall regression and delta advance (Evans, 1979). The cause of the rising sea levels may have been partly the result of a eustatic rise associated with interglacial conditions, although the thick nature of the accumulated sediments would have also necessitated some form of basinal subsidence. In contrast during low sea level stands, possibly induced by climatic deterioration and eustatic lowering, the southern North Sea probably became an area of sediment bypass resulting in the formation of a series of erosion surfaces apparent in the southern North Sea delta sequence (Cameron et al., 1984). Such a mechanism would explain the presence of both ameliorative and harsh/arctic episodes in the Lower Pleistocene marine and deltaic sequences (Figs. 4.1 - 4.4).

It should be mentioned that restricted occurrences of glaciomarine facies were recorded from the Lower Pleistocene succession, therefore making them the earliest described glacigenic sediments from the North Sea Basin. However, their limited extent (2 boreholes) precludes further speculation at this stage.

The Lower Pleistocene/Middle Pleistocene boundary is delimited over the study area by the Brunhes-Matuyama boundary, whilst amino acid data (Fig. 3.2) suggests that in places this boundary may be accompanied by a significant hiatus. Sedimentation during the early Middle Pleistocene was significantly different to that during the Early Pleistocene. This is primarily reflected by the establishment of shallow, hyposaline arctic marine conditions and a vertical transition through arctic marine to glacial and glaciomarine facies. These sediments were tentatively attributed to the early "Cromerian Complex" as shown in Fig. 3.2.

Within this glacigenic sequence sediments corresponding to the subglacial, proximal glaciomarine, distal glaciomarine and arctic marine environments (facies A^1-E^1), described for the late Weichselian model, were all identified. In addition the lithofacies relationships are consistent with deposition from a tidewater ice front rather than an ice shelf. However, the extent of the subglacial facies suggests that the ice sheet, or ice sheets, was more extensive than in the late Weichselian. This is especially true in the northern Witch Ground Basin where the sediments are typical of the previously described inner and marginal facies associations, but with a degree of interdigitation indicating that the area was affected by a series of re-advances prior to the main phase of retreat. The limited extent of distal glaciomarine facies and their total absence in the inner facies association was attributed in Chapter 5 to rapid and significant glacio-isostatic rebound as the ice sheet retreated from the area, therefore limiting the migration and establishment of a distal glaciomarine environment the direction of ice retreat.

It is therefore envisaged that during the early Cromerian an extensive ice sheet, ending as a tidewater front, covered the whole of the Bosies Bank area and the northern part of the Witch Ground Basin. Beyond the limit of the ice front proximal and distal glaciomarine sediments accumulated in a hyposaline arctic marine sea. Subsequent retreat of the ice front towards the west was accompanied, to a limited extent, by a westward migration of the glaciomarine environment. Deposited clasts are of a local or Scottish origin suggesting that the ice sheet originated from the British Isles rather than Scandinavia. Prior to or during the final retreat of the ice front, the substantial thickness of the inner and marginal facies associations necessitates that the sequence must have been built up by a series of re-advances of the ice front (Boulton et al., 1985). This is consistent with the interdigitary nature of the various facies.

In the Forties area the deposition and preservation of a thick sequence (ca. 70m) of distal glaciomarine sediments suggests that the area must have been subsiding at this time, either in response to the relaxation of a peripheral bulge or as a result of tectonic downwarping. This contrasts markedly with the late Weichselian period when the Witch Ground Basin was the main area of subsidence and focus of distal glaciomarine sedimentation.

In the Marr Bank area the thinner and less extensive facies associations are consistent with a less active ice sheet and possibly only one phase of advance and retreat. As such it had little effect on the sediments accumulating in a shallow marine sea in the Devils Hole area which are typical of the arctic marine muds deposited during the final stages of late Weichselian sedimentation in the Witch Ground Basin. The eastern limit of subglacial material in this area (Fig. 5.3) further suggests that no Scandinavian ice sheet was involved in the deposition of these sediments.

Whether the two inherently different successions in the North and South were deposited by the same ice sheet is still debatable. Certainly the two successions are, in places, in close agreement regarding their position relative to the Brunhes-Matuyama boundary, and also it is not uncommon for an ice sheet to behave differently along the extent of its front. Therefore, given the present lack of any other evidence it is suggested that the two successions are related to the same ice sheet.

6.4.2 Middle to Late Pleistocene History

The most significant features of the sequence deposited during this period is the presence of two extensive and highly irregular unconformities (Fig. 6.6). Both are attributed to a possible combination of fluvial, subglacial and proglacial processes, although fluvial erosion is thought to be the prime mechanism acting during climatic or isostatically induced, low sea stands. The lower surface is thought to have formed during the Elsterian stade and the upper surface during the Saalian stade. Evidence for the age of these surfaces and the intervening sediments is presented in Chapter 3. Significantly the lower unconformity becomes most irregular in the northern part of the study area with a local relief of up to 80m, whilst in the South-west the surface flattens dramatically forming a less irregular and more continuous unconformity. In contrast the upper, Saalian, reflector is most irregular in the South and especially the South-west with a local relief of up to 100m, whilst over much of the Fladen area, to the North, the unconformity is less irregular. A further feature of the two unconformities, is that where they are highly irregular, forming a series of approximately north-south trending channel features, base lines can be drawn joining the channels. These base lines dip approximately towards the central Witch Ground Basin and Forties area for the Elsterian and Saalian surfaces respectively. In addition, the Elsterian surface shows evidence of post formation subsidence. For example, in the Witch Ground Basin the regional base level is approximately 320m below sea level, whilst on the flanks it is only some 200m below sea level.

From the above descriptions, and accepting the model presented for the late Weichselian, it is suggested that the areas which are significantly channelled were exposed to intensive sub-aerial processes during periods of low relative sea level stand induced primarily by climatic cooling and isostatic compensation. An obvious complication, however, is the possibility that the features are the products of subglacial meltwater erosion or even direct glacial erosion (Valentin, 1955; Flinn, 1967; Jansen et al., 1979). However, a strong factor against this last possibility is that of the six boreholes which penetrated the Elsterian erosion surface only two contained coarse glacigenic material immediately overlying the unconformity, and both penetrated the flank rather than the centre of a channel feature. In fact, where two boreholes did penetrate the base of a channel feature the sediments immediately overlying the unconformity contained favourable micropalaeontological assemblages consistent with a strong ameliorative period. A sub-aerial origin is also preferred in the light of a detailed study of an open channel in the Bosies Bank area which was initiated during the Elsterian (Fig. 2.21). This showed that distinctive lateral accretion surfaces are present along the channel margins supporting a fluvial or possibly glaciofluvial origin. Furthermore, the presence of asymmetric chaotic infills in the base of some channels, as described in Chapter 2, may be the result of periglacial activity and mass movement processes (Long & Stoker, 1986).

It is therefore envisaged that during the Elsterian stade a large expanse of the sea bed in the study was exposed to sub-aerial denudation and channelling in response to the relative fall in sea level. The degree of emergence and consequent downcutting was apparently greatest towards the North of the study area, although the possibility of an Elsterian ice sheet in the near vicinity of the study area, and the effects it would have had, should also be considered.

However, on the above assumptions the absence of Elsterian glacial and glaciomarine sediments in the area remains an enigma. One possible explanation is that the situation was similar to that during the late Weichselian where the areas of channel incision lay beyond the limit of the ice sheet. Thus Elsterian ice may have been present in the immediate vicinity of the study area. This is certainly consistent with the description of a British Elsterian (Anglian) ice sheet whose limit lay just to the east of Dogger Bank (Zagwijn, 1979; Cameron et al., 1986). In addition, it is therfore possible that the deeper channel features were the result of fluvo-glacial erosion and modification, as suggested by Holmes (1977). Alternatively it could be argued that the absence of Elsterian glacigenic sediments in the area purely reflects the poor borehole control at this level in the stratigraphic succession.

A subsequent rapid marine transgression resulted in the submergence of many of the channelled areas, and the absence of a reworked lag in the channel bases is explained by the rapid rise in sea level therefore precluding any significant reworking of the sea bed (Evans, 1979). The rising sea level, a response to climatic amelioration during the Holsteinian, was accompanied by the deposition of a thick sequence of marine silts and sands containing favourable microfaunal and floral assemblages at a stage when the less severe channel features comprised an interconnecting network of open depressions. The relative high organic content of these sediments possibly contributed to the production of large amounts of methane gas which are clearly identifiable on seismic profiles, and where conditions were favourable, migration of the gas resulted in the acoustic blanking of large areas of overlying sediment. Although complex, the seismic configurations of such infills generally indicates an upward shallowing as reflected by the transition from suspension to traction current sedimentation.

Climatic cooling and a general deterioration of conditions towards the end of the Holsteinian and the onset of the Saalian is recorded in the channel infill sediments by a change to hyposaline, shallow arctic microfauna and flora. In addition the vertical lithofacies relationships indicate a transition from marine to an arctic marine sedimentary environment in the south of the study area, to distal and proximal glaciomarine environments in the Witch Ground Basin. The latter indicate an increasing proximity to the ice front towards the north and eastern margins of the Witch Ground Basin. However, no subglacial facies of this age were recovered, and the sedimentary facies associations typical of the late Weichselian are not developed here.

The sequence is therefore interpreted as follows. Subsequent to the deposition of temperate marine sediments in the channel depressions climatic deterioration and ice build up resulted in an ice front encroaching onto the northern or north-eastern margin of the study area, its exact location being unknown. Consequently its retreat led to the deposition of a sequence of proximal and distal glaciomarine muds in the channel depressions and on the channel flanks. Mass movement processes on the channel flanks were probably ubiquitous, whilst further south, away from the direct influence of the ice front, arctic marine muds accumulated in shallow, hyposaline water. The nature of the ice front terminus is uncertain although proximal glaciomarine sediments in the area are consistent with deposition on a sub-aqueous fan suggesting a tidewater front type terminus.

Rapid isostatic recovery may have prevented any migration of the respective glaciomarine environments in the direction of ice retreat. However, the poor borehole control and likelihood of subsequent erosion, especially of sediments deposited on the channel flanks and interchannel areas, means that this is by no means certain.

Widespread erosion of the previously described succession is defined by an extensive planar to sub-planar unconformity, formed by a marine transgression at some stage during the Saalian period. This was accompanied by the establishment of a stable, shelf marine type environment, as defined from seismic profiles, and the deposition of a thick sequence of argillaceous marine sediments similar to those described for the Lower Pleistocene.

Overlying the marine sediments the accumulation of an extensive sequence of glacial and glaciomarine facies in the study area is attributed to a major glacial episode during the Saalian. Within this sequence the distribution of glacial and glaciomarine facies suggests that much of the study area was covered by ice at some stage during this period. Work from the southern North Sea by Mitchell (1977), Jardine (1979) and Oele & Schuttenhelm (1979) similarly supports an extensive Saalian glaciation in the area. Outsize clasts in the sediments are again consistent with a local and Scottish derivation thereby favouring a British rather than a Scandinavian ice sheet.

Fig. 5.6 shows the distribution of the glacial and glaciomarine sediments and the respective sedimentary facies. In the ice-proximal environment a large sub-aqueous fan built out from a tidewater front forming a series of overlapping esker fans clearly identifiable on seismic profiles. A complex range of lithologies associated with the fan, including massive and stratified sands, diamicts and slump deposits, reflect the variety of processes active in the proximal glaciomarine environment. At this stage the distal glaciomarine environment probably lay to the south and east of the study area, but as the ice front retreated to the north, the distal glaciomarine environment migrated into the Devils Hole area and inner and marginal facies associations developed. Continued northward retreat of the ice front was interrupted by a series of still stands or re-advances during which large sub-aqueous moraines developed which still form significant topographic highs along the present sea bed (Fig. 2.24). This series of ridges were also identified by Jansen (1976), the "southwestern ridge", who attributed them to Saalian ice. However, Jansen et al. (1979) later stated that the ridges contained interglacial material of Eemian age. Coring of the ridge in association with this study reaffirmed Jansen's original interpetation.

In the central Witch Ground Basin subsequent uplift and denudation of the above sequence of sediments resulted in their ending as sharp scarp around the flanks of the Witch Ground Basin. Contours drawn to the base of the planar erosion surface (Fig. 2.27) clearly delimit the position of this scarp. Uplift of the Witch Ground area is attributed to isostatic compensation and the formation of a peripheral bulge as the ice sheet lay on the flanks of the basin. Subsequent isostatic relaxation of this area would have left a broad depression in the approximate area of the present day Witch Ground Basin. However, it is not suggested that this mechanism was the prime factor in the development of the basin as it is likely that tectonic subsidence also had an effect on the area (Jansen, 1976).

The second unconformity was eroded during a low sea level stand in the late Saalian period probably in response to extensive glacio-isostatic uplift corresponding to the retreat of the main Saalian ice sheet described previously. Sediments overlying this surface therefore span the late Saalian to Weichselian stages, although the occurrence of a number of discontinuous erosion surfaces in this succession makes correlation difficult.

Sediments of a temperate marine nature accumulated in some of the channel features, in a similar manner to those occuring above the lowermost unconformity, and as such these have been attributed to the Eemian interglacial stage. Therefore, following the formation of a late Saalian erosion surface a climatic amelioration and related marine transgression resulted in the deposition of a sequence of temperate marine sediments in the channel depressions. This agrees with the sediments recovered from just above the second irregular unconformity in the Tartan field, which Jansen and Hensey (1981) identified as being of Eemian age. Similarly, in the southern North Sea Oele (1979) describes a series of Saalian valleys which cut into the Dogger Bank and were subsequently infilled with Eemian sands and clays. The restriction of both Eemian and Holsteinian interglacial sediments to the lower parts of channel infill sequences is a good analogy of present day conditions where sediments are accumulating preferentially in the deep open channels.

Overlying the temperate marine infill in the channel features, the sediments record a change to a glaciomarine environment signified by the deposition of proximal and distal glaciomarine facies, namely massive and stratified diamicts and laminated muds with dropstones. However, where the basal unconformity is less irregular, as in the central Witch Ground Basin, no interglacial sediments were preserved and glaciomarine sediments directly overlie the unconformity. No equivalent subglacially deposited material was identified for this sequence although the glaciomarine facies show a general coarsening towards the west. Mass movement processes were active on the flanks of the channel features resulting in the deposition of distinctive slump deposits containing mixed microfaunal and floral assemblages.

Given their stratigraphic relationship to the Eemian interglacial sediments the above sequence of glaciomarine facies is thought to be predominantly early to middle Weichselian in age. If this is correct, then the sediments would equate to the rapid build up of the high latitude ice sheets at around 80,000 yrs BP (Shackleton & Opdyke, 1973) and also the glaciation of parts of Scandinavia and western Europe (Worsley, 1977). Mixed microfaunal and floral assemblages from some of the glaciomarine facies in the sequence would therefore be consistent with the relatively mild oceanic conditions prevalent during parts of the early and middle Weichselian (Ruddiman et al., 1980; Miller et al, 1983). Miller et al (1983), for example, suggest that for long periods during this stage the Norwegian Sea was only affected by a seasonal cover of sea ice and that during this stage the North Atlantic Drift, albeit it weak, may have been able to penetrate the North Sea.

However, no early or middle Weichselian glacigenic sediments have positively been identified on mainland Britain (Mitchell, 1977), although Sutherland (1984) suggests that the shelly diamicts at Kilamaur, Scotland are of an early Weichselian (Devensian) age. Similarly, offshore no subglacial material of this age has been positively identified in the study area and it was therefore suggested in Chapter 5 that the ice limit was located somewhere along the western margin of the study area and that any subglacial material was subsequently removed by erosion. Low sea level stands, both during and subsequent to the deposition of early to middle Weichselian sediments, are indicated by their often overconsolidated nature, attributed to sub-aerial dessication, and the occurrence of a number of discontinuous reactivation surfaces.

6.5 Summary of the Pleistocene succession

The deposition and preservation of a thick sequence of Pleistocene sediments, which locally exceeds 500m in the study area, can only be explained by continued tectonic subsidence which followed a rifting phase during the Jurassic and Lower Cretaceous

(Sclater & Christie, 1980). This is supported by the fact that the thickness of the Pleistocene sediment pile is closely related to the underlying Mesozoic/Tertiary tectonic features and the axis of maximum subsidence clearly correlates with the Central and Viking Grabens (Fig. 1.3). In addition the base levels to an extensive Elsterian erosion surface suggest that differential subsidence of some 120m occurred in the central Witch Ground Basin relative to the surrounding area.

However, superimposed on the Pleistocene subsidence of the North Sea Basin are the effects glacio-isostatic and eustatic influences, resulting in the development of a complex sequence of sedimentary sequences bounded by extensive erosion surfaces. As discussed previously, the Pleistocene succession in the study area can be subdivided into a lower sequence of predominantly Lower Pleistocene marine and deltaic facies and an upper sequence of predominantly glacial and glaciomarine facies. The two are separated by an extensive Elsterian erosion surface formed predominantly by fluvial processes although subglacial and proglacial processes may have locally contributed. Thus during the Lower Pleistocene the area was generally characterised by a gradually subsiding marine basin in which a thick sequence of marine sediments accumulated. Evidence of a northerly prograding deltaic sequence in the southern North Sea is only apparent along the southern margin of the study area where a sequence of delta-front and prodelta facies were deposited. Changes in sea level during this period, as indicated by facies transitions and micropalaeontological data, were not generally extreme enough to result in the formation of extensive erosion surfaces. In contrast, Lower Pleistocene deltaic and marine sediments in the southern North sea attain a thickness of some 500m (Cameron et al., 1986), but this succession is divided by a series of unconformities attributed to low sea level stands. It is therefore suggested that during the Lower Pleistocene, sedimentation over much of the study area continued relatively uninterrupted, despite periods of climatically induced eustatic sea level lowering, whilst to the south low sea level stands resulted in extensive erosion and the establishment of a sediment bypass zone. Overall, however, the bulk of the Lower Pleistocene sequence was probably associated with a rising sea level, but concomitant with rapid sedimentation which at times overtook the sea level rise resulting in a regression.

The presence of Cromerian glacial and glaciomarine facies in the sequence marks the lowest, extensively identified, period of glacigenic deposition in the area, during which time the North Sea was totally cut off from the North Atlantic Drift. The sequence also signifies a marked change in the pattern of sedimentation in the area and the increased importance of glacio-isostatic influences on the overall depositional environment. Thus, in the Forties area the thick accumulation of distal glaciomarine sediments, associated with the Cromerian ice-sheet, is attributed to a retreat of the ice-front and the collapse of a peripheral bulge. Similarly, the extensive Elsterian erosion surface is attributed to a low relative sea level due to the combination of eustatic lowering and the formation of a glacially induced peripheral bulge.

Overlying the Elsterian erosion surface the Pleistocene sequence contains a further two extensive, irregular unconformities. These are dated as being of late Saalian and late Weichselian age respectively, and both were formed during low sea level stands resulting from a complex interplay of climatically induced eustatic lowering and glacio-isostatic loading and peripheral bulge development. A third, relatively planar erosion surface was formed by a marine transgression during the middle or early Saalian and was followed by a period of marine sedimentation in a stable or gently subsiding basin. The sharp termination of this erosion surface and the overlying sediments at the edges of Witch Ground Basin can only be explained by uplift of the Witch Ground Basin area, possibly as a result of glacial loading and the development of a peripheral bulge.

The Pleistocene sequence overlying the Elsterian erosion surfaces is interpreted in the following stages:

i. A marine transgression, related to the Holsteinian interglacial, during which the irregular erosion surface was submerged and temperate marine sediments accumulated over the area but were preferentially preserved in the deeper channel features. Tectonic subsidence must have continued subsequent to the formation of this surface because it is presently some

200m below sea level around the edges of the area, and up to 320m below sea level in the Witch Ground Basin.

ii. A deterioration of climatic conditions during the late Holsteinian-early Saalian resulting in a marine regression and a transition from arctic marine to glaciomarine sedimentation. The ice-sheet associated with this development probably lay to the north or west of the study area.

iii. A marine transgression during the early or middle Saalian, possibly associated with climatic amelioration although this is uncertain, formed an extensive sub-horizontal surface and was accompanied by the deposition of a thick sequence of marine sediments.

iv. A marine regression during the main Saalian glaciation was accompanied by the deposition of an extensive sequence of glacial and glaciomarine sediments. The distribution of these sediments suggests that this was the most widespread glaciation to affect the study area.

v. A continued fall in relative sea level, possibly relative to widespread glacio-isostatic uplift and maintained low eustatic sea levels, resulted in the formation of a second extensive, irregular erosion surface; probably during the late Saalian stage.

vi. A marine transgression associated with the Eemian interglacial was accompanied by the establishment of open oceanic conditions and the deposition of a sequence of temperate marine sediments. These were preferentially preserved in the deeper channel features.

vii. A further marine regression, associated with climatic deterioration, occurred during the early-middle Weichselian period. This was accompanied by the deposition of arctic marine and glaciomarine sediments which were locally cut by reactivation surfaces. The ice sheet associated with the deposition of these sediments probably lay along the western margin of the study area.

viii. A late Weichselian ice sheet encroached onto the western edge of the study area and glacio-isostatic depression of the sea bed resulted in a marine transgression towards the ice front. Beyond the limit of isostatic depression the development of a peripheral bulge coincided with a low eustatic sea level resulting in uplift and sub-aerial exposure of much of the eastern part of the study area. The exception to this occurred in the Witch Ground Basin where shallow water depths prevailed. The uppermost irregular unconformity is attributed to this stage, although its formation appears to be the result of a combination of fluvial, subglacial and proglacial processes; fluvial processes being predominant in areas beyond the ice limit.

ix. A subsequent retreat of the ice sheet was followed by rapid isostatic uplift of the glaciated areas and the collapse and inward migration of a peripheral bulge. Rising sea levels, related to the main late glacial deglaciation eventually overtook isostatic effects and culminated in the rapid rise in sea level at the onset of the Holocene (10,000 yrs BP), resulting in the preservation of an extensive lag of palimpsest sediments over much of the study area.

The above events are based on the analysis of seismic, lithological and micropalaeontological data. However, it should be stressed that the timing of the events, especially during the Saalian and early-middle Weichselian, is based on a 'best fit' model. For example, the use of the terms early, middle and late Saalian, are purely relative and not based on absolute dating techniques. It is in fact ironic that for the most part there is a greater stratigraphic control on the Lower Pleistocene.

Regarding the applicability of the late Weichselian model presented here to the development of the pre-late Weichselian sediment pile, it appears that many of the processes of transport and deposition can be applied throughout the sedimentary column. However, the fact that the Cromerian and main Saalian ice sheets were more extensive than during the late Weichselian meant that glacio-isostatic influences would have had different effects on the sedimentary environment. This can be corroborated by comparing the infills immediately overlying the late Saalian and late Weichselian irregular erosion surfaces. Thus, interglacial sediments occur in the lower part of some Saalian channels whilst only glaciomarine or arctic marine sediments were recovered from late Weichselian channels. It is therefore suggested that the late Weichselian channels were, for the most part, cut during a glacial period whilst the Saalian channels were cut partly in response to isostatic uplift following the retreat of an ice sheet from the area. The applicability of the late Weichselian model to other contemporary and ancient sequences will be discussed in the next section.

6.6 **Analogues to the North Sea Pleistocene Sequence**

The aim of the following section is to compare the stratigraphic architecture of the Pleistocene sequence and the glaciomarine model developed for the late Weichselian to similar environments of deposition outside the North Sea. These environments will be considered in two parts. First, contemporary and Pleistocene environments; and second, ancient environments.

6.6.1 **Contemporary and Pleistocene Glaciomarine environments**

Eyles et al., (1985) states that four main glaciomarine environments can be identified: the basin margin shelf, slope, canyon and basin plain; surprisingly fjords were not included as a separate environment. However, the North Sea cannot be placed in any of the above categories and during the Pleistocene it appears to have been a unique environment. This can be primarily attributed to three factors:-

i. the North Sea area is a shallow epicontinental basin some 600km wide and 1100km long whose present geometry is the result of tectonic subsidence initiated by a period of rifting during the Mesozoic. Subsidence appears to have continued through the Pleistocene.

ii. the mid-latitude maritime setting of the North Sea meant that during the Pleistocene the area was not covered for long durations by thick ice sheets.

iii. the strong relationship of the British ice-sheets to the presence of the polar-front, and the latter's effect on both precipitation and the North Atlantic Drift. As a result of this the British ice sheet was apparently a sensitive indicator of North Atlantic climatic conditions and may at times have been latitudinally diachronous (Ruddiman et al., 1980; Sissons, 1981; Sutherland, 1984; Boulton et al., 1985).

To date, no contemporary or Pleistocene analogue has been identified which fulfills the above criteria. This is especially apparent with regards to the stratigraphic architecture of the Pleistocene which contains a complex record of low sea-level stands and glaciations. Indeed this could only have been achieved by basinal subsidence because in more stable tectonic environments, affected by repeated glacial advance, only the most recent glacigenic sediments tend to be preserved (Bjorlykke, 1985; Nystuen, 1985).

Looking at the tectonic regime one of the closest analyses to the North Sea is the East China Sea Basin. Here, a thick sequence of Mesozoic and Cenozoic sediments have accumulated in an epicontinental rift-depression basin (Milliman, 1985). A number of similarities can be drawn between this sequence and the North Sea Pleistocene succession including shallow gas, the preservation of an extensive erosion surface formed during late Weichselian low sea-level stands and the occurrence of buried fluvial valleys, again attributed to low sea levels stands during the late Weichselian. Obviously, however, the regime is not comparable because of the absence of glacial influences. In contrast, the Pleistocene succession of the Beaufort sea contains both glacial and glaciomarine facies which accumulated under a regime of limited tectonic subsidence (Dinter, 1985). Pleistocene low sea level stands were therefore recorded in this succession, although the whole sequence appears to have been deposited during the late Weichselian. Other examples of glacially influenced basins with divergent Atlantic type tectonic settings (Barnes and Piper, 1982; Syvitski, 1986), although often characterised by subsidence tectonics, are also not directly comparable to the North Sea stratigraphic architecture or the late Weichselian depositional environment. This can be explained on two counts. First, subsidence was relatively limited when compared to the North Sea. Second, in the North Sea water depths appear to have decreased away from the late Weichselian ice front, whilst continental shelf environments are characterised by increasing water depths and sharp bathymetry gradients on the slope. A further factor is the open nature of continental shelf environments whilst during glaciations the North Sea Basin existed as a partially enclosed glaciomarine sea.

Contemporary and Pleistocene fjord glaciomarine environments are similarly unrelated to the North Sea Pleistocene setting. The most notable differences are the relatively high sedimentation and marked increase in water depth away from the ice front, although some fjords are characterised by an intervening shallow sill. Well documented examples of fjord environments include those occurring around Spitsbergen (Boulton, 1979; Elverhoi et al., 1983; and Elverhoi, 1983; 1984), and the Alaskan coast (Powell, 1981; Powell, 1983; Mackiewicz, 1984). Sedimentation rates in these environments can be as high as 10m/ka (Elverhoi et al., 1983) in the proximal glaciomarine environment with extremes of 4m per year in Glacier Bay, Alaska (Powell, 1981). Water depths in the inner fjord, Glacier Bay, generally average 160m (Mackiewicz et al., 1984). Significantly, only sediments related to the latest period of glaciation are preserved.

Accepting that a contemporary or Pleistocene analogue to the North Sea is not presently available, it is still possible to relate various aspects of the late Weichselian model to those presented by other workers. For example, Boulton (1986) interprets a raised ice contact fan in Omega Bay, Baffin Island as a subaqueous (subaquatic) outwash fan. Subsequent retreat of the ice sheet in Omega Bay resulted in rapid isostatic uplift and marine erosion of the fan surface. Similarly Boulton (1986) also describes the inhibition of winter calving by sea ice at a tidewater ice front on Spitsbergen, a comparable scenario being envisaged in Fig. 6.2.

An important concept of late Weichselian model and the whole stratigraphic sequence is the formation and preservation of fluvial valleys cut during low sea level stands. Pleistocene analogues are described from the New Jersey Shelf where late Weichselian river valleys, cut during a low sea level stand, are infilled with a sequence of shallow marine muds as a result of coastal retreat and backfilling. One such valley, related to the Hudson, was only partially infilled and therefore still has a considerable surface expression. Similar open and buried valleys are described from Lake Superior (Landmesser et al., 1982) and the continental shelf off Vancouver Island (Herzer and Bornhold, 1982); although the latter may be proglacial in origin.

An excellent example of the preservation potential of glacigenic sediments deposited in tectonically favourable environments is afforded by a brief review of Varangerian sequences in Scandinavia and Scotland (the Port Askraig Tillite). For example, in Scandinavia Varangerian glacigenic sequences accumulated on a slowly subsiding shelf (Nystuen, 1985) whilst in Scotland the Port Askraig tillites accumulated on a more rapidly subsiding intra-cratonic shelf (Eyles & Eyles, 1983; Eyles et al., 1985 and Nystuen, 1985). Thus, the Port Askraig tillite is some 850m thick and contains multiple diamict sequences, bounded by normal marine sediments, which accumulted in a low relief shallow marine basin (Eyles and Eyles, 1983). However, in the slowly subsiding shelf environment cannibalism of earlier glacigenic sequences by the last ice advance meant that a 4-120m thick diamict, preserved in Scandinavian Caledonian nappes, is generally attributed to the last glacial phase (Nystuen, 1985). The effects of the tectonic setting are summed up by Nystuen (1985) who states "In more rapidly subsiding shelves, several glacial advance-retreat units are preserved together with interstadial and interglacial sediments, with total thickness up to several hundred metres. The long-term preservation is determined by various factors controlling relative sea level changes."

Looking at the comparability of the late Weichselian sedimentation model to ancient sequences, the Late Palaeozoic glacigenic deposits in Murray Basin, Australia (O'Brien, 1986) provides an ideal analogue. These comprise of laminites (or cyclopels) with dropstones, stratified diamicts, graded sandstones and massive mudstones, which in total attain a maximum thickness of some 800m. The sequence was interpreted by O'Brien (1986) as the result of proximal and distal glaciomarine sedimentation in a subsiding shallow marine basin, with ice masses encroaching onto the basin edges during glacial maxima. The lithofacies are generally consistent with deposition on a subaqueous outwash fan characterised by underflow and turbidity currents and iceberg rafting dominating the more distal glaciomarine environment.

Surprisingly, however, O'Brien states that "the nature of the ice that deposited the paratillites in the Urana Formation remains a problem," and also subsequently refers to the now discredited ice-shelf model of Carey & Ahm d, 19). In fact the previously 1961 described sequence is typical of that deposited adjacent to a tidewater ice-front, as described here or by Powell (1984 and 1986).

A similar pre-occupation with ice-shelf models is typified by Tucker and Reids (1973) interpretation of Late Ordovician laminites, slumps, diamicts and turbidites, in Western Africa, to deposition beneath and bordering on a large ice-shelf. This and other examples led Eyles et al (1985) to state that "the widespread recognition of sub ice-shelf sedimentation in the rock record has been overemphasized because of simple reliance on massive diamictite lithofacies as indicators of deposition below an ice-shelf cover."

In conclusion, because of the apparently unique setting of the North Sea Basin during the Pleistocene epoch no equivalent glaciomarine sequences have been recognised from other Pleistocene successions. Similarly, no contemporary glaciomarine environment has presently been identified which can be either compared to a late Weichselian sedimentary model presented here or invoked to explain the stratigraphic architecture of the Pleistocene succession in the North Sea. However, a number of ancient glacigenic sequences preserved in intra-cratonic basins appear more akin to the North Sea, to the extent that they occasionally allowed for the accumulation and preservation of several glacial advance-retreat related units, widespread erosion surfaces, and local interglacial sequences.

CHAPTER SEVEN

CONCLUSIONS

The availability of an extensive data set from the study area, comprising core material and shallow seismic profiles, provided an excellent opportunity for the study of a thick Pleistocene pile of marine, glaciomarine and glacial sediments. The stratigraphic architecture of the sequence and associated depositional environments were determined using a variety of techniques; seismic analysis, sedimentary structures and particle size determination, facies relationships, and micropalaeontological interpretation, proved the most useful of these.

The establishment of a detailed seismic or para stratigraphy proved essential on account of the number and complexity of the sequences occurring in the North Sea basin. Seismic analysis also provided information as to the geometry of these sequences and the large scale depositional environment. Most importantly, seismic profiles allowed for the identification of a basin wide irregular erosion surface which separates two distinctly different packages. The lower unit is characterised predominantly by reflector configurations associated with deposition in a subsiding shelf marine environment. The upper unit is more complex and contains a series of erosion surfaces and a variety of reflector configurations. Coring of these sequences showed that the lower unit consists predominantly of marine sediments, and the upper unit of glacigenic and arctic marine sediments. Seismic interpretation was only lacking in the Peterhead area due to the generally poor quality of the seismic profiles. Unfortunately this precluded a direct correlation of late Weichselian sediments along the western margin of the study area, a vital location with regards to the position of the last ice front.

Dating of the seismic stratigraphy was achieved primarily from palaeomagnetic events, previously established by Stoker et al. (1983), and the occurrence of a few critical micropalaeontological horizons indicative of strong ameliorative periods. This stratigraphy was further corroborated by a wide variety of evidence including amino acid data, an ash layer, plant fragments and basin-wide erosion surfaces. However, the stratigraphy is a 'best fit model' and especially lacks hard evidence over the Middle Pleistocene period. Further amino acid work and possibly pollen analysis will hopefully be carried out by Bergen University and may partly help to rectify this situation.

The use of micropalaeontological data within the seismic framework also provided invaluable information as to the nature of respective depositional environments. Dinoflagellate cysts proved their use not only as a 'marine pollen' but as sensitive indicators of the presence or absence of the North Atlantic Drift and of the possibility of a sea ice cover. The sparcity of foraminiferal assemblages indicative of temperate conditions, relative to the dinoflagellate record, is thought to be the result of the prevailing water mass conditions. Thus, even during periods of climatic amelioration bottom water conditions remained colder than at present. This is highlighted in the Weichselian lateglacial record when bottom water conditions remained 2-3°^{C} lower than present despite the influence of the North Atlantic Drift and warm surface waters. None of the species identified are diagnostic of specific climatic stages within the Pleistocene.

Lithofacies analysis of the late Weichselian sequences suggests the palaeogeography of the North Sea basin was somewhat different than originally suggested. Thus, the eastward extent of the late Weichselian Scottish ice sheet was relatively limited and it only encroached onto the margins of the North Sea Basin. In addition the combination of glacio-isostatic adjustments and eustatic sea level lowering resulted in a transgression towards the ice front and the uplift of areas beyond the immediate zone of glacio-isostatic depression. An extensive, and highly irregular erosion surface, formed at this time, is though to be the result of a combination of fluvial, glaciofluvial and glacial processes; fluvial erosion being predominant in the areas beyond the ice front. Because of this complex interplay isostatic and eustatic processes previous generalisations of sea level depths are not particularly useful.

Sedimentary facies associated with late Weichselian glaciomarine environment were deposited from meltwater flows and melting icebergs originating from a grounded tidewater front. The development and distribution of these was strongly influenced by the basin configuration. Thus, in the north a west-east continum of proximal to distal glaciomarine sediments reflects a transition from bottom current and iceberg rafting processes to overflow plumes and rare iceberg rafted debris. However, in the south shallower water depths away from the ice front precluded the establishment of a distal glacial marine environment and instead a thick sub-aqueous fan developed in the proximal glaciomarine environment. The presence of a cover of sea ice during the winter season probably precluded glaciomarine processes producing a strong seasonal effect on the sediments, a factor not generally accounted for in previous glaciomarine models.

A study of the pre-late Weichselian sequence in the area has shown that from the Middle Pleistocene sedimentation in the North Sea Basin has been dominated by glacial and glaciomarine processes. Prior to this the sequence consists predominantly of temperate marine and deltaic sediments. Comparison of the pre-late Weichselian glacigenic sequences to the late Weichselian model suggets that the predominant depositional environment was one of a shallow glaciomarine basin adjacent to tidewater ice front. A series of low sea level stands led to the truncation of these sequences. No evidence was found to suggest that the depositional environment was influenced by an ice-shelf rather than a tidewater front type termination.

No contemporary or Pleistocene analogues have been identified therefore corroborating the unique setting of the North Sea Basin during the Pleistocene era. However, in the pre-Pleistocene stratigraphic records glacial and glaciomarine sediments form thick sequences associated with intra-cratonic sedimentary basins. Further study of these environments, such as the Palaeozoic basin in Australia, and comparison with the North Sea may help facilitate an understanding of the development and preservation potential and glaciomarine sequences.

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