

Reconstructing the Last Pleistocene (Late Devensian) Glaciation on the Continental Margin of Northwest Britain

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Declaration

I declare that this thesis has been composed solely by myself and that it has not been submitted, either in whole or in part, to any previous application for a degree. Except where acknowledged, the work presented is entirely my own.

Stephen Davison

September, 2004

Abstract

The continental margin in the area west of Shetland was subjected to repeated and extensive ice sheet advances during the Late Pleistocene. Seabed imagery, seismic survey and borehole core data show the Late Devensian ice sheets expanded across the continental shelf three times, two of these advances reaching the shelf edge. On the inner shelf, where present-day water depths are generally less than 100m, only thin sediments from the last retreat phase and exposed rock surfaces remain, all other deposits from earlier phases having been removed by the last advance. On the mid to outer shelf elements of all three phases are preserved, including lodgement and deformation tills, melt-out and water-lain till sheets, in-filled hollows left by stagnant ice decaying *in situ* and a series of large recessional and terminal moraines. In addition, there is evidence of shallow troughs and overdeepened basins which indicate preferential ice-drainage pathways across the shelf which were formerly occupied by ice streams. At the shelf edge, a thick wedge of glacial sediment forms a transition from the till sheets and moraines of the shelf to debris flows composed of glacial sediments on the upper slope. Shelf-edge moraines show an architecture indicating floating ice in modern water depths over approximately 180m, suggesting the West Shetland ice sheet was no more than about 250m thick. The upper and middle slope is dominated by glacial debris flows which are focused in the slope areas below the proposed ice stream discharges at the shelf edge. The mid-to-lower slope has been subjected to contour current activity which has re-worked much of the glacial sediment in this position. The lower slope and floor of the Faroe-Shetland Channel are marked by either large debris flow lobes of glacial sediment or thin glacial muds deposited from suspension. A conceptual model of the glacial development of a passive continental margin based upon the West Shetland example shows the deposited sequence for both advance and retreat phases of a glacial cycle, and the actual preserved sequence which might be expected in the rock record. The model also shows that ice sheet buoyancy, thickness, and to a lesser extent, basin subsidence, are the most important factors in the deposition and preservation of a glacially-influenced marine sequence.

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1. Introduction

The history of the study of glaciers and their related deposits dates from the late 1700's when scientists of the day speculated that the glaciers of the Alps and Scandinavia had previously extended beyond the observed limits. By the mid 1800's, the concept of more widespread glaciers was becoming accepted, expounded most notably by the Swiss scientist Louis Agassiz. With the increasing acceptance of the theory of widespread glaciation to account for the unconsolidated deposits of northern Europe, aided by the discovery of the Greenland Ice Sheet in 1852, the possibility of older glacial episodes in the rock record was raised. By the end of the 19th Century Permian and Precambrian deposits of glacial origin had been positively identified in several places around the world from northern Norway to Australia. Glacial deposits have now been identified from nearly every period of geological history (Figure 1.1a), with the exception of the mid-Proterozoic (Hambrey, 1994). Of these, the Pleistocene deposits are by far the most extensive as a result of their recent formation. Ancient deposits have been recognised from every continent (Hambrey and Harland, 1981; Stump *et al*, 1988) but are of more limited extent due, at least in part, to subsequent erosion and re-working. In the last few decades attention has turned to glacial deposits in marine settings on the continental margins as these are less likely to be removed by subsequent weathering and erosion which fragments the terrestrial records. With the increased interest in past and future climate change, and the better preservation of the offshore sequences, emphasis has shifted to the marine environment to study the nature of the Quaternary glaciations.

The Pleistocene was marked by a series of large climatic fluctuations (Figure 1.1b) which resulted in the growth and decay of extensive mid-to-high latitude ice sheets. The processes of erosion and deposition associated with these ice sheets had a considerable influence on the development of the continental margins. A substantial body of work has been established in the past to identify and characterise the sediments and sedimentary processes involved in the glaciation of the continental

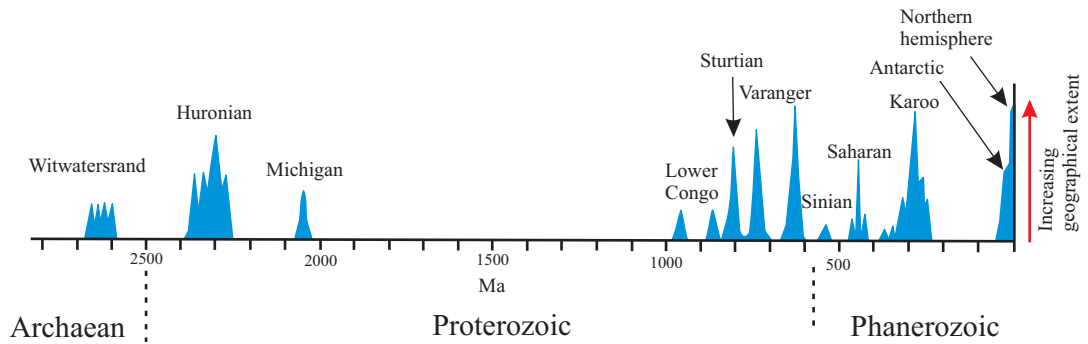


Figure 1.1a. Distribution of major glacial episodes throughout geological time (Modified from Hambrey, 1992).

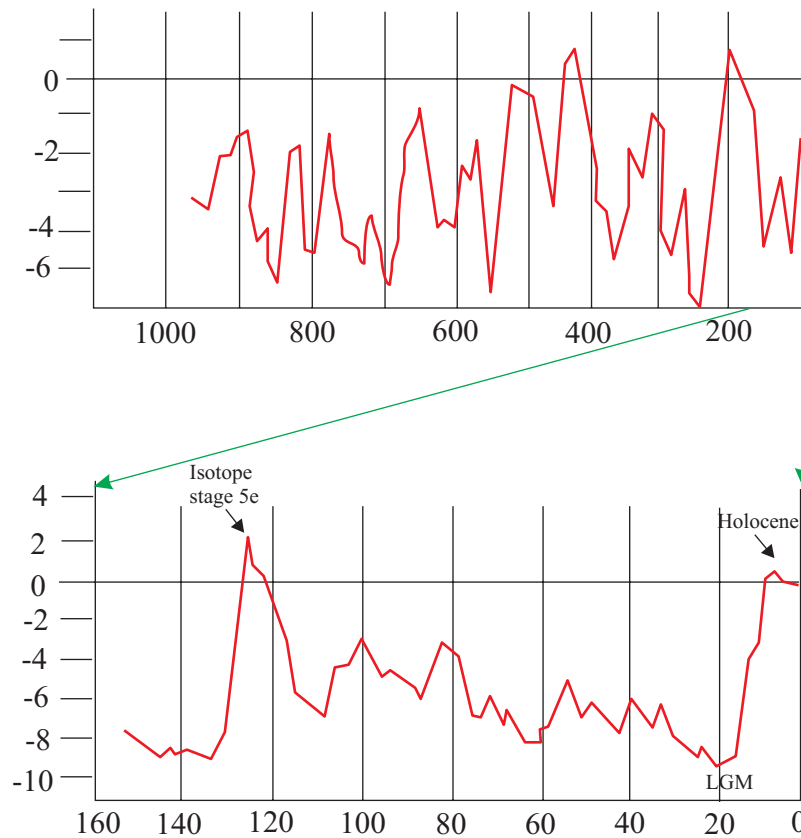


Figure 1.1b. Detailed average temperature changes for the last 1 million years, and for the last 160 000 years, showing the fluctuating nature of the climatic record (simplified from Boulton, 1993).

margins. However, many of these previous studies have concentrated on particular provinces of the margin such as the inner shelf or the continental slope and few have attempted to trace the glacial influence from the inner shelf to the deep basin floor. In addition, many of the previous studies on which glacial margin models are based, have relied upon sea-bed sampling and short gravity cores to identify lithological units.

1.1. Objectives and area of study

This study combines previous work with regional 2D seismic survey data and borehole cores from the UK West Shetland Shelf and Faroe-Shetland Channel. A new seabed image derived from 3D commercial seismic exploration data, which have not previously been available, also sheds new light on the detailed morphology of the sea-floor.

The main purposes of this study are:

- i) To characterise the acoustic, sedimentological and architectural features of the Quaternary glacial succession in the West of Shetland region and assess the contribution of glacial processes to continental margin development in this area.
- ii) To reconstruct the last major glacial episode in terms of the maximum extent of the ice sheet, the different erosional and depositional processes operating, and assess the implications for ice sheet dynamics within the West Shetland area.
- iii) To construct a predictive model of passive glaciated margins based upon the West Shetland Margin.

The choice of the West Shetland region for this investigation is partly governed by the high density of available information and the commercial interest in the Faroe-Shetland area. In addition, the Faroe-Shetland Channel is also an important component in the oceanic circulation of the North Atlantic, with sedimentary records showing marked changes in the current regime between glacial and interglacial phases (e.g.; Ruddiman and McIntyre, 1981; Rasmussen et al, 1996a). Current

activity shows a broad cyclicity between very slow or absent currents with mud deposition during cold phases and strong currents which are broadly erosive during the inter-glacials. Unravelling the history of the British Ice Sheet in the West Shetland area during the last major glaciation may help to correlate the ice movements with the deep marine circulation record in the North Atlantic by showing the timing of advance and retreat phases of the ice sheet and the relationship to depositional phases in the Faroe-Shetland Channel. The warm, northward flowing surface water of the North Atlantic Current flows through the upper levels of the Faroe-Shetland Channel (FSC) and the West Shetland Shelf, combining with North Atlantic Water and keeping winter temperatures of northwest Europe relatively mild (Broecker *et al*, 1985). Along the floor of the FSC, cold southward flowing water from the Norwegian Sea, the Norwegian Sea Deep Water, contributes to the formation of North Atlantic Bottom Water (e.g. Rasmussen *et al*, 1996a, 2002). Records from the FSC show evidence for the presence of strong currents, alternating with a period of very weak or absent currents. The breaks in the circulation record correlate with cold events during the last 58ka suggesting that the north Atlantic oceanic circulation was significantly changed during glacial episodes (Rasmussen *et al*, 1996b).

1.2 Glaciated continental margins and glacimarine processes

Existing models of glaciated continental margins are largely the result of studies from high-latitude areas such as the Antarctic, Svalbard and Greenland. The lower latitude margins of western Norway, eastern Canada and the north-west UK also provide good examples of non-polar margins which are no longer subject to a glacial influence.

The earliest work constructing a conceptual model of glacimarine environments was undertaken by Carey and Ahmed (1961) based upon data from the present Antarctic ice sheets and Pleistocene ice sheets of the Northern Hemisphere. Although widely quoted at the time, this model was developed without the benefit of seismic survey data or core material which was not available at that time. Much of the later work on glacimarine sediments has drawn upon seismic surveys of Quaternary deposits both

in the polar regions and data from ancient glacial sequences. This has been aided by the extensive hydrocarbon exploration surveys in the Northern hemisphere and by the Ocean Drilling Program (ODP). The resulting models from these data sources generally comprised a prograding wedge of glacial sediments extending from the shelf edge, building into the deeper water of the continental slope and delivering clastic sediment to the basin. (e.g.; Powell, 1981. Anderson, 1983. Eyles et al, 1985)

The early models showed sediments in the shelf zone were broadly deposited either directly from the ice, or by sediment-laden meltwater emerging at the grounding line. Depositional processes at the grounding line were shown to have an important influence on the morphology of the outer shelf and slope zones (Powell, 1990) and to be strongly controlled by sea level (Boulton, 1990).

Recognition of ice streams, zones of faster flowing ice within an ice sheet, led to the identification of former glacial troughs in the continental shelf zone of Antarctica (Hambrey et al, 1991. Cooper et al, 1991) and Norway (Vorren et al, 1989). These troughs formed a focus for sediment delivery at the shelf edge, forming large fan-shaped depocentres, termed trough-mouth fans, extending from the slope and expanding by repeated debris flow deposition (Vorren et al, 1989). In addition to the deposits derived directly from the ice and meltwater, deposition from icebergs has also been shown to be a major source of glacial marine sediment (e.g.; Drewry and Cooper, 1981. Thomas and Connell, 1985. Dowdeswell and Murray, 1990). Non-specific margin models which attempt to show in detail the overall stratigraphic architecture from shelf to basin are still relatively uncommon, although some attempt has been made to constrain the different zones of sedimentary processes (e.g. Boulton, 1990). A generalised cross section showing the main features of a marine-terminating ice sheet on a continental margin is presented in Figure 1.2, and is based upon an amalgamation of the various basic models. The following sub-sections summarise the characteristics of ice transport processes, and the features of glacial

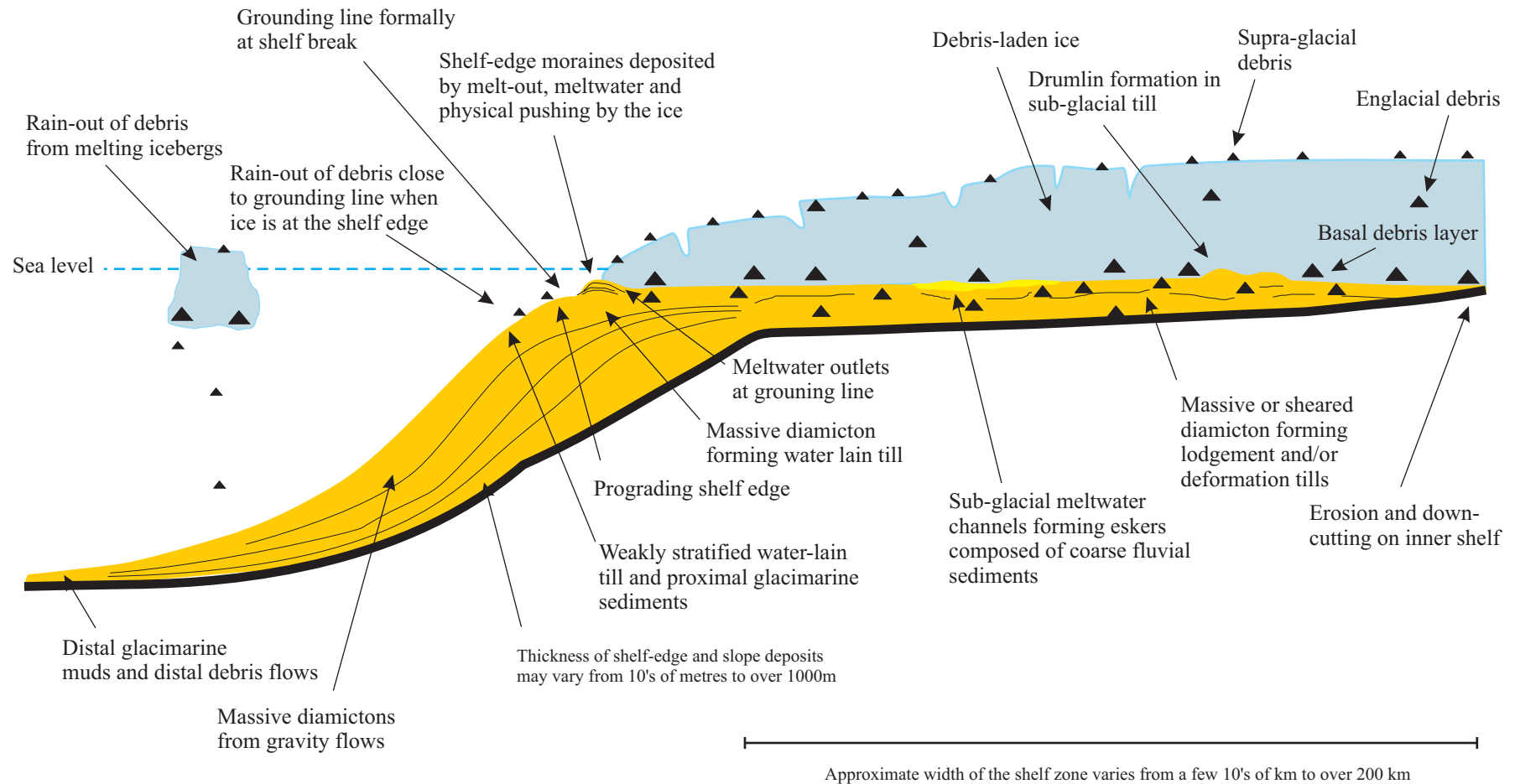


Figure 1.2. Simplified section of the features of a glaciated margin with ice advanced to the shelf edge (based upon Boulton, 1990; Hambrey, 1994)

erosion and deposition on continental margins, together with selected examples of glaciated margins from the North Atlantic region.

1.2.1 Ice transport processes

The main processes by which sediment is transported by ice, and transferred from the ice to the marine environment are common to all glaciated margins, although the proportion delivered by the different processes varies. The Antarctic ice sheet has been the focus of many of the investigations into the transport of ice and sediment to the marine environment (e.g.; Hambrey, 1994. Anderson, 1999). Based upon Antarctic data, the principal methods by which ice and its associated sediments are transported into the marine environment by cold Polar ice sheets are via;

- i) Ice shelves
- ii) Ice cliffs
- iii) Outlet glaciers and ice streams.

The processes characterising the more temperate Pleistocene ice sheets of the Northern Hemisphere are broadly the same, but the degree of importance of each process is different. The different processes and their relative importance are described below.

i) Ice Shelves:

Ice shelves form from coalescing outlet glaciers or large ice sheets which extend onto the continental shelf and where the water depth provides sufficient buoyancy to allow an ice sheet to decouple from the seabed beneath and form a floating ice mass. The lack of constraint and friction which results from this causes the ice to thin and spread out. Of all the modern glacial settings, ice shelves occur mostly around the Antarctic, with smaller scale examples in the Canadian and Greenland High Arctic, suggesting they may be limited to very cold polar ice sheets. The Antarctic ice shelves receive significant volumes of snowfall which leads to accumulation on the ice shelf. This has the effect of adding ice to the upper surface, effectively forcing a downward path for englacial debris. If the accumulation on the surface is combined with basal melting, large volumes of sediment are released beneath the ice shelf before the ice reaches the ice front. This leads to large sediment accumulations on

the inner shelf and a sediment starved outer shelf.

On ice shelves where the temperature of the ice is very low, as in some of the Antarctic ice shelves, basal melting does not occur. If the temperature is sufficiently low, sea ice forms on the base of the ice shelf. The effect of adding ice to the base is to prevent release of debris onto the inner shelf, resulting in the calving of debris-laden icebergs at the ice front which may be at or close to the shelf break. The transfer of ice to the marine realm by ice shelves is thought to account for nearly 60% of the total in Antarctica (Drewry and Cooper, 1981. Anderson, 1998).

ii) Ice cliffs

Ice cliffs form where the grounding line of an ice mass coincides with the calving line. These generally form where melting and erosion by the sea remove the heavily crevassed ice which normally occurs at the ice front (Anderson, 1998). As a result, debris deposition is close to the ice margin and can lead to highly aggradational sequences which can reach sea level (Powell and Molnia, 1989).

iii) Outlet glaciers and ice streams

Ice draining from the interior of a landmass can become channelled into faster flowing streams and outlet glaciers as a result of convergent flow. The convergence is governed by a number of factors such as pre-existing topography of the landmass and shelf, and the nature of the underlying rocks and their resistance to erosion. These glaciers have a high rate of discharge, and may account for over 20% of the total Antarctic ice discharge (Drewry and Cooper, 1981). However, the low percentage of the total ice discharge is disproportionate to the volume of sediment transported by the fast ice. The current data for the Antarctic ice sheets suggests more than 50% of all glacial sediment being delivered to the marine environment is transported via the ice streams and outlet glaciers (Drewry, 1986). In more temperate settings such as the southern areas of South America, eastern and southern Greenland, north eastern Canada and parts of Norway, outlet glaciers terminate in fjords. During the last glaciation, these glaciers extended to the coast where they coalesced to form large ice sheets on the shelf with ice streams forming along the

preferential pathways. This style of ice drainage, with outlet glaciers reaching the shelf via fjords and coastal mountains, draining an inland ice sheet is a likely scenario for the last ice sheet in the NW of Britain, indicated by the apparent exposure of coastal nunataks and trimlines (Hambrey, 1994; Ballantyne *et al*, 1998). However, it is also likely that once into the marine realm, the lack of confining mountains would allow the outlet glaciers to coalesce forming an expansive ice sheet.

1.2.2 Debris transport

Sediment is transported as basal, englacial or supraglacial material (Figure 1.2), of which the basal component is the largest. Basal debris is generated and transported as a result of ice movement and is derived from two sources. Erosion of the bedrock underlying the ice, principally by shearing and abrasion, and the incorporation of regolith sediments provides much of the basal debris. As the ice sheet expands, it also overrides sediments previously delivered to the ice front by meltwater, adding to the volume of basal debris. In the context of continental margins, sediment accumulates in river estuaries prior to glacial advance or in fjords following earlier advances. On the shelf, normal processes of marine erosion and deposition provide sediment which is incorporated into the basal debris layer. The volume of basal debris is controlled by a number of factors including ice thickness and ice temperature. These two properties determine whether the base of the ice sheet is at pressure melting point, allowing the presence of meltwater and large scale incorporation of debris by regelation, or if the ice is below the pressure melting point which prevents sliding of the ice along the underlying bedrock and leads to a much thinner debris layer. Investigations from Antarctica suggest that the thickness of the basal debris layer is approximately 1% of the total ice thickness (Hambrey, 1994).

Englacial and supraglacial debris (Figure 1.2) form a minor component of total sediment delivered by the Antarctic ice sheets to the modern marine environment. Englacial debris is transported within the ice and is generated by burial of supraglacial material, up-thrusting of basal ice or the merging of two ice masses. Consequently this type of debris has no unique characteristics. Supraglacial debris is

derived from either aeolian sources or from outcrop material falling onto the ice sheet surface. Both these sediment types have properties which can be used to indicate their origin. Aeolian sediment has by its nature a fine grain size and may also possess the typical rounded grain morphology. Sediment derived from outcrops exposed above the ice have no such limit on grain size and tend to have an angular morphology, unlike most other glacially eroded and transported sediment in the glacial-marine realm. Only parts of the Antarctic ice sheet and some Alaskan glaciers show evidence of supraglacial transport into the modern marine environment. However, it is considered to have been an important process in the transport of debris by the Northern Hemisphere Pleistocene ice sheets, where ice descended from mountainous regions with abundant nunataks onto broad coastal plains (Hambrey, 1994).

Much of the work on generic ice sheet processes is the result of studies of the Antarctic ice sheet, although the actual ice sheet may not be characteristic of ice sheets generally for a number of reasons. The Antarctic continental shelf lies in a water depth of around 500m due to loading of the continent by ice. This is considerably deeper than most modern ice-free shelves which in the northern hemisphere generally lie between 200m and 400m. The depth of the Antarctic shelf is also deeper than the UK shelf when the Pleistocene ice sheets were in place. In addition, the shelf around much of the Antarctic continent slopes landwards, again due to loading of the continent by ice, and due to deposition on the outer shelf. The result of the ice loading is to make water depths on the inner shelf much deeper than on the outer shelf. This effect is enhanced by deposition on the outer shelf from floating ice. Additionally, the extreme low temperature of the ice causes seawater to freeze onto the base of the floating ice, preventing sediment deposition until the ice reaches the outer shelf where melting occurs. Much of the Antarctic shelf is now sediment starved as the ice sheets have receded to the deep water of the inner shelf (Hambrey, 1994).

A further property which makes the Antarctic margin atypical are the slope angles on the continental shelf which reach as high as 15°, decreasing to between 1° and 3° on

the continental slope. Slope angles on the shelves of other continental margins are generally much lower, being in the order of $< 1^\circ$.

1.2.3 Processes of glacial erosion

The processes of erosion associated with glaciation fall into three groups; processes associated directly with the ice, processes resulting from the action of meltwater and freeze-thaw action, including regelation. The majority of the processes are only significant in glaciers and ice sheets where the base is above the pressure melting point; so called “warm-based” ice sheets. Ice sheets where the base is frozen to the substrate, “cold-based” ice sheets, lack the sliding movement between ice and rock/sediment, the forward movement being achieved by internal deformation of the ice. The erosion processes are summarised below from detailed accounts by Hambrey (1994) and Benn and Evans (1998).

Ice processes

Erosion by ice occurs mainly as a result of shearing stress applied by the movement of the ice downslope under gravity, although ice loading is also responsible for physical break-up of the underlying rock. The movement of the base of the ice, and debris frozen into the base of the ice, causes abrasion of the underlying rock surface and the ice entrained debris. This motion effectively produces a grinding action which reduces the size of the debris and is the primary process forming rock flour; glacial “mud” of clay to silt sized sediment which forms the matrix of most glacial deposits. The shearing motion also assists the entrainment of more debris into the basal ice together with regelation, described below.

The excavation of clasts from the bedrock surface by ice is termed quarrying (Sugden and John, 1976. Drewry, 1986) although it is generally reserved for the removal of larger clasts from the substrate. Clasts are physically removed from the bedrock by the crushing and shearing stresses which cause rock failure. Once excavated, the clasts are entrained within the basal debris.

The ice processes are responsible for the formation of the regional unconformities

created at the base of a glacial advance. Repeated advances of an ice sheet can create a compound unconformity surface which is the result of repeated erosional events.

Meltwater action

Sub-glacial meltwater beneath warm-based ice sheets is an important erosional agent, in addition to being a vital component of the process of basal sliding by which the ice sheet moves. The pattern of sub-glacial drainage can vary greatly from channel networks, to mass flow within saturated till layers. A feature common to all of these systems is the high pressure under which they normally occur. Meltwater transported in sub-glacial channels commonly has a high suspended and traction load, together with rapid turbulent flow. These features combine to produce high rates of abrasion and cavitation which act upon the ice, the underlying bedrock and/or previously deposited glacial sediments (Benn and Evans, 1998). Fluid pressure is also important in enlarging fractures and breaking down rock surfaces (Allen, 1982), and the meltwater also acts as an important chemical weathering agent. In addition, meltwater emerging from beneath ice sheets, especially under pressure, can act as an erosive agent in the same way as any terrestrial stream in the proximal area beyond the ice front. High pressure, sediment laden meltwater is responsible for carving out fluvial style drainage systems, creating tunnel valleys beneath ice sheets (Figure 1.2) which have a distinctive sedimentary fill (described in section 1.2.4).

Freeze-thaw action

Freeze-thaw processes are not unique to glacial environments, but can act as important agents of rock breakdown where meltwater re-freezes. Regelation also comes under this heading, which involves the pressure melting at the base of the ice sheet on the up-stream side of obstacles and re-freezing on the down-stream side where pressure is reduced. This process is important in the incorporation of sediment into the basal debris layer of the ice.

1.2.4 Processes of glacial marine deposition and their associated features

The bulk of glacial sediments are deposited into the marine environment broadly by three main processes; i) direct deposition from the ice; ii) melt-out from grounded

or floating ice; and, iii) by fluvio-glacial processes from meltwater. The different processes show changes in their relative importance between ice sheet advance and ice sheet retreat phases. The main processes and features are summarised below.

i) Direct deposition from the ice

Direct deposition from the ice beneath an ice sheet generally forms lodgement or deformation tills (Figure 1.2). Lodgement tills form by the release of debris from the ice due to pressure melting at the base of a sliding ice body and the depositing of entrained debris. The debris is smeared onto the underlying substrate by the forward movement of the ice (e.g.; Boulton, 1970. Dreimanis, 1989). Structurally, lodgement tills frequently show a parallel or sub-parallel fabric of jointing and clast alignment formed by shearing processes, although they may be massive (Benn and Evans, 1998). Acoustically, a lodgement till with a shearing fabric is likely to form a unit with sub-parallel, discontinuous reflections or a chaotic pattern with vague stratified appearance. Deposition of lodgement till is the dominant depositional process during advance phases of glaciation.

Deformation tills are defined as a disaggregated and largely homogenised sediment formed in a sub-glacial deforming layer (Benn and Evans, 1998). The sediment may contain local and exotic material. The nature of the deformation process tends to destroy pre-existing structure (Boulton, 1987) which makes identification difficult in both outcrop and seismic sections.

Melt-out deposits

Melt-out tills form either sub-glacially or supra-glacially and are deposited without modification. Consequently these tills may acquire textural features of the debris laden layers of ice from which they form (Benn and Evans, 1998). Melt-out also provides sediment which is re-worked into lodgement tills, fluvio-glacial deposits and moraine banks. In addition to melt-out tills, the process also releases debris from floating ice sheets and icebergs where it forms rain-out deposits. Sediments deposited from floating ice are characterised by the presence of dropstones within a poorly sorted, but commonly stratified diamicton. With increasing distance from the

ice front, the input of ice-rafted debris (IRD) decreases. In ice-distal settings the IRD signature may be limited to fine clastic sediment and rare dropstones within hemipelagic muds and silts.

The formation of moraine banks, particularly end/terminal moraines, are formed by a combination of processes which include direct deposition from melting ice, entrained debris and deformation processes. As an ice sheet advances, it builds up a bank of sediment at the ice front which is pushed along in front of the ice. Smaller banks can form where a retreating ice sheet pauses (recessional moraines), as sediment being is still being delivered to the ice front by the conveyor-like process of ice movement. In addition, banks can form as a result of minor ice front oscillations, causing a temporary re-advance (push moraines). Such deposits generally have a chaotic internal structure both physically and acoustically, although modification by meltwater may add stratified components.

Deposition from meltwater

Formation of glacial meltwater, especially beneath the ice, forms meltwater streams which transport debris and emerge from beneath the ice at the grounding line or drain off or through the ice and emerge from tunnels at the ice front (Figure 1.2). The emerging debris-laden meltwater decelerates on reaching the marine environment resulting in the rapid deposition of the coarse component of the transported sediment forming sandy diamictons or muddy sands and gravels. On a large scale, the sediments form sheet-like deposits which thin away from the ice front. In detail the deposits show stratified beds with ripple cross beds and scour structures. Climbing ripples, indicative of rapid deposition are also common. The fine fraction can be transported as suspended sediment for great distances beyond the ice (Boulton, 1990). Warm-based glaciers and ice sheets produce meltwater constantly, although there are variations in output, leading to the deposition of meltwater-transported sediments in both advance and retreat phases. However, deposits formed within the area of maximum ice expansion prior to the maximum will be re-worked by the advancing ice. Deposition from meltwater is a more important process during the retreat stages of glaciation, with the deposits being more likely to be preserved.

1.2.5 Seabed morphological associations

Glaciations which terminate within the marine environment produce features which are common to all margins, although the preservation of such features may vary widely. The characteristics are related to the position relative to the ice front and form at different stages in the glacial cycle from initial advance to glacial maximum and glacial retreat. The submarine morphology which results from a glacial cycle is a composite of features developed by all the processes operating during the cycle.

The coastal and inner shelf zones tend to be areas of net erosion, commonly exhibiting scoured rock surfaces, especially where the underlying rock type is igneous or high grade metamorphic. On retreat of the ice the surfaces are covered with thin tills deposited by melt-out and meltwater, but in the marine environment of the inner shelf where water is relatively shallow, the sediments are quickly re-worked and removed by normal marine processes.

The middle and outer shelf zones are generally areas of net deposition characterised by till sheets and moraine deposits (Figure 1.2). All of the features are blanketed with distal glacialine muds deposited from suspension during the final ice retreat. However, the morphology of the larger features is still evident on the seafloor. The most obvious of these are generally moraine banks marking the maximum extent of ice expansion if the advance terminated on the shelf, or pauses/ice front oscillations formed during the retreat. The moraine banks can vary in size from features a few metres high and hundreds of metres long, to tens of metres high, several kilometres wide and tens or hundreds of kilometres long. In the case of terminal moraines, the banks mark a divide in the marine landscape, the area beyond the ice advance being dominated by delta-like deposits formed by meltwater emerging from the ice front and sheet deposits formed from suspension settling and iceberg rafting. Erosive forms in this zone are limited to iceberg ploughmarks.

In the area formally occupied by the ice, similar delta, till sheet and moraine bank deposits are also present, but are superimposed on sub-glacial forms which may

influence the seabed morphology. The sub-glacial features most likely to be present at the seabed include depositional forms such as drumlins and eskers (Figure 1.2) and erosional features such as shallow troughs and overdeepened basins. In addition, composite features such as megaflutes, large scale striations potentially tens of kilometres long and hundreds of metres wide formed on the surface of sub-glacial tills (Solheim et al, 1990), may be evident.

In general, most of the seabed features described above do not survive as surface features if subjected to a further glacial cycle. However, if the shelf is subject to subsidence or sea-level rise, both of which create accommodation space, it is possible for earlier features to be preserved. An example of such features exists on the northern Norwegian shelf where a Late Weichselian terminal moraine has been overridden by later glacial advances, which has modified but not removed the moraine banks (Vorren and Plassen, 2002).

1.2.6 Characteristics of selected glaciated margins in the North Atlantic region

Existing large-scale models of glaciated margins are largely the result of studies from high-latitude areas with active glacial systems such as the Antarctic, Svalbard and Greenland. The mid-latitude margins of western Norway, eastern Canada and the north-west UK also provide examples of lower latitude margins which are no longer subject to a glacial influence. Whilst these areas share many common features, they also have many diverse characteristics that contribute to the overall view of glaciated margins.

Seismic investigations have shown that the large scale architecture of most glaciated margins comprises a prograding wedge of glacial sediment extending from the pre-glacial shelf edge (e.g. Vorren *et al*, 1989; Boulton, 1990; Stoker, 1990; Cooper *et al*, 1991; Kristofferson *et al*, 2000). The specific characteristics of the glacial wedge vary between different margins, but generally consist of a combination of acoustically transparent lensoid bodies, interpreted as debris flows, and parallel to sub-parallel acoustically stratified units interpreted as hemipelagic-glacial marine sediments. On a shelf-to-basin transect, the main seismic packages are a roughly

sigmoidal shape in vertical section, being thin on the shelf, thickening markedly at the shelf edge and thinning towards the slope base into the deep water environment. At the base of some slopes the profile is modified by accumulations from large scale mass movement. This architecture is particularly well developed on the SE Greenland margin (Clausen, 1998) and the West Shetland margin of the UK (Stoker, 1999).

The following sub-sections summarise the main characteristics of the glaciated margins off northern and mid Norway, the Scotian margin of Canada and the NW UK (Hebrides and West Shetland areas).

1.2.6 Examples of Northern Hemisphere glaciated margins

Northern Norway

Vorren et al (1989) developed a model based upon the Barents Sea margin (Figure 1.3). The margin comprises a glacially eroded unconformity, referred to as the upper regional unconformity, which extends across most of the shelf and is overlain by thick glacial sediments. The unconformity surface is actually a composite of surfaces from repeated glacial advances, forming a diachronous surface. On the shelf the deposits are up to 300m thick, but at the shelf edge and upper slope these thicken to 1000m in some places. The shelf is marked by a series of broad troughs up to

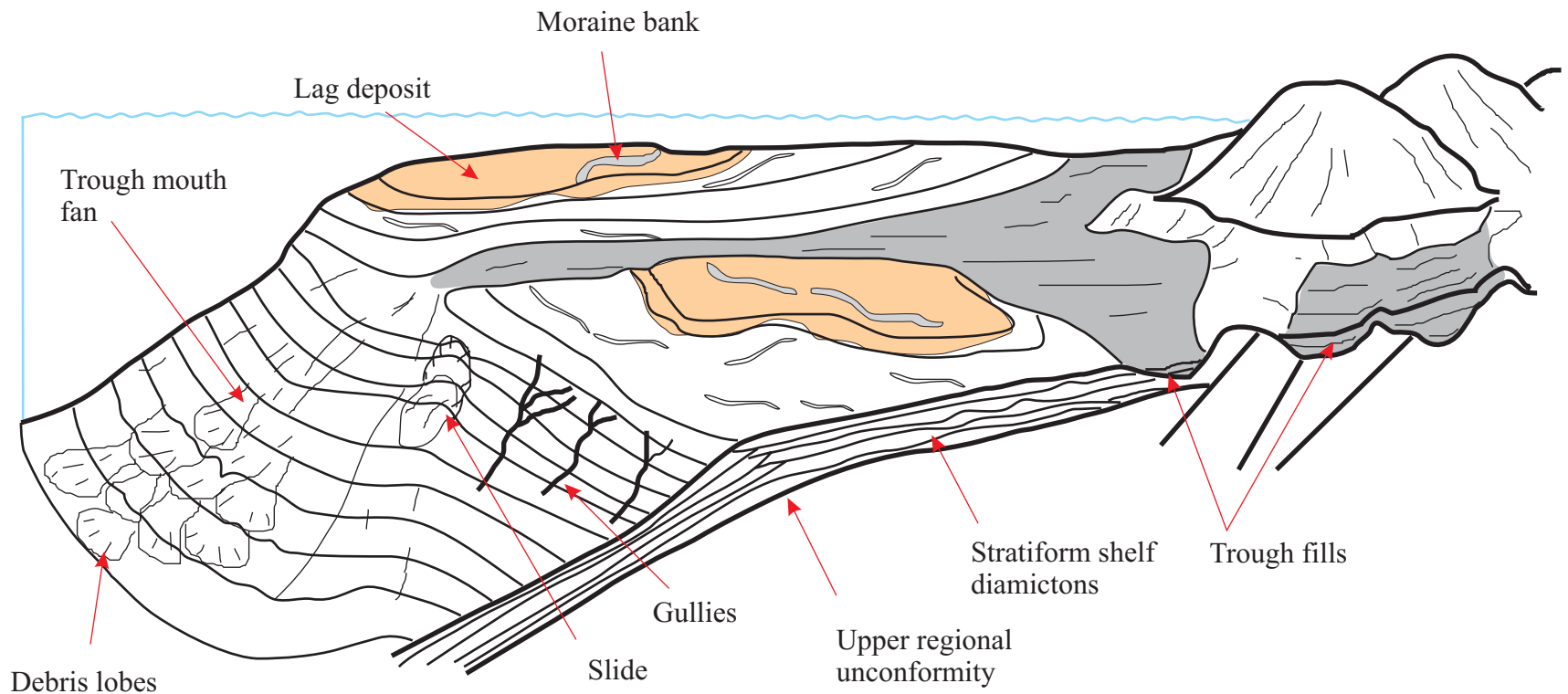


Figure 1.3. Conceptual model of the main features of a passive glaciated margin based on northern Norway (redrawn from Vorren *et al*, 1989).

150m deep with onlapping fills of till and glacial marine sediments. The areas between the troughs are occupied by sheet-like diamictites which locally form moraine banks. The shallower areas of the shelf show evidence of post-glacial current re-working, forming gravel lag deposits derived from the winnowing of till and fluvio-glacial deposits (Figure 1.3). The largest accumulations of sediment lie at the end of cross-shelf troughs where they meet the shelf edge. These depocentres form prograding trough mouth fans which are dominated by repeated debris flow accumulation. The sequence on the slope has an overall sigmoidal pattern of prograding glacial packages. The upper slope is also marked by numerous slide scars on a wide range of scales, reflecting the instability caused by rapid deposition. Away from the main depocentres on the slope, erosional gullies have been formed by cold, dense water descending the slope during periods of low sediment input.

Mid-Norwegian Margin

The western margin of Norway between 60° N and 68° N shows features on the shelf which are common to other margins. However, the slope zone has been subject to repeated major instabilities forming large scale slide deposits (Bugge, 1983; Evans et al, 1996). In addition, sediments from the slope system have fed into the margins of the North Sea Fan, a large trough mouth fan lying at the mouth of the Norwegian Channel, south-west of the Vøring Plateau (Figure 1.4). The mid-Norwegian shelf lies in water depths between 300 m and 400 m deep. Where the slope drops to the Vøring Plateau the water reaches depths of 1200 m to 1400 m. Beyond the Vøring Plateau the lower slope descends on to the North Sea Fan where the slope base lies at over 3000 m water depth.

The glacial sequence is of Plio-Pleistocene age and overlies a seaward-dipping regional unconformity on the shelf which represents a compound surface similar to that described on the northern Norwegian margin. The section comprises a composite prograding wedge of sediments up to 1500 m thick. forming sigmoidal foresets of debris flows separated by thin hemipelagic units. Subsidence on the shelf has allowed aggradation of the sequence (Figure 1.5), preserving earlier phases of the

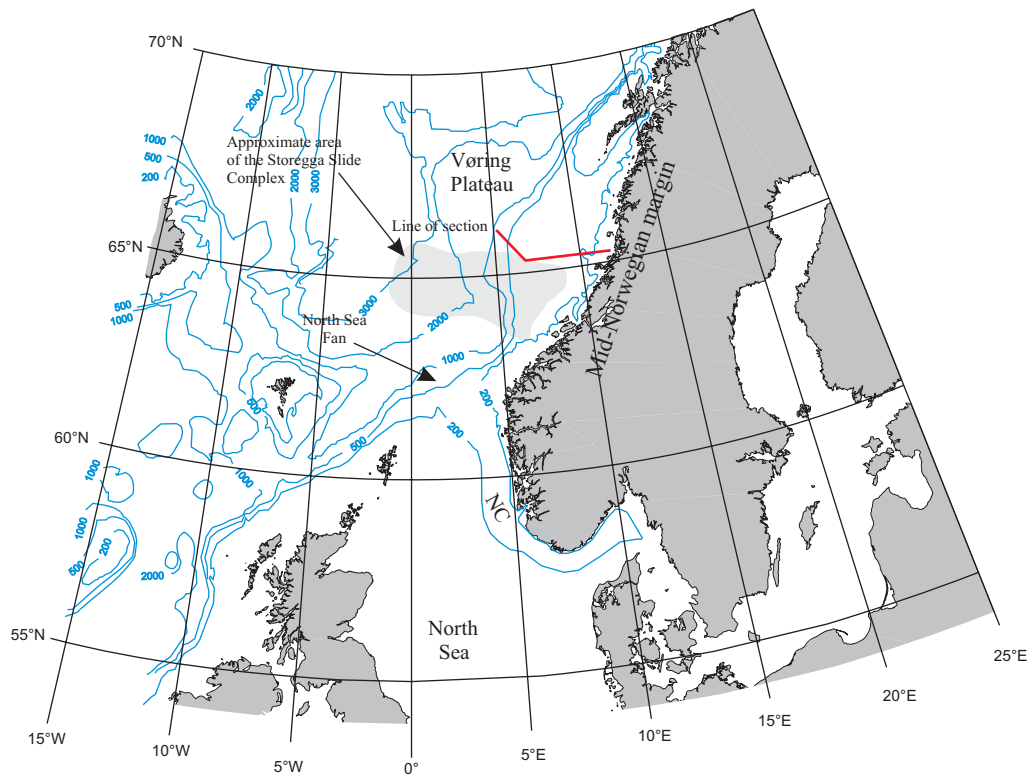


Figure 1.4. Map showing the location of the mid-Norwegian margin and the location of the transect in Figure 1.5. NC - Norwegian Channel

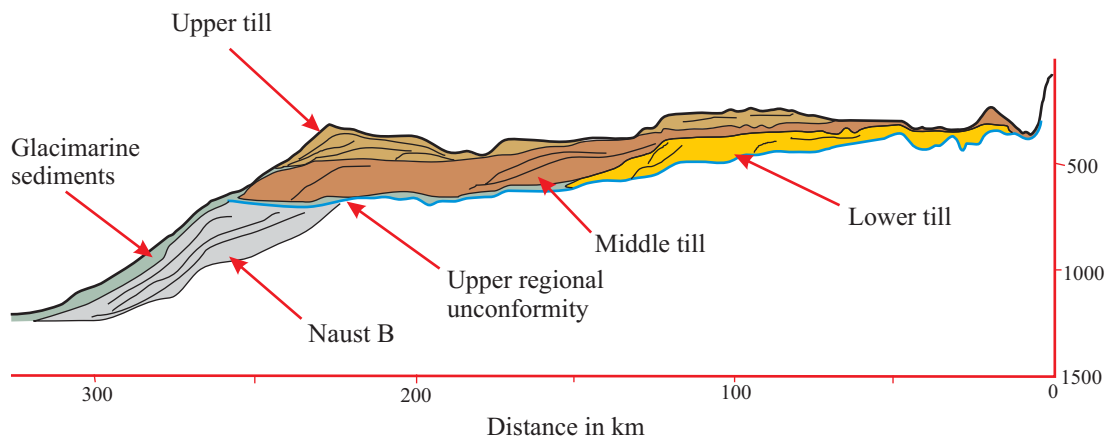


Figure 1.5. Cross section of the mid-Norwegian margin showing the upper slope and shelf occupied by the glacial Naust A and B Formations (re-drawn from Dahlgren *et al*, 2002).

Quaternary glaciations and giving rise to an inter-fingering stack of sub-glacial tills and glacimarine deposits (Dahlgren et al, 2002). The sequence shows a wedge of glacial sediments lying beneath the Upper Regional Unconformity (URU) deposited by a shelf-wide glaciation during Marine Isotope Stage (MIS) 10. A significant period of erosion on the shelf is indicated, followed by a limited glacial advance during MIS 8. This advance did not reach the shelf edge, but deposited a prograding series of diamictos on the shelf which form the Lower Till (Figure 1.5). Between MIS 8 and MIS 6 a significant degree of subsidence occurred, creating more accommodation space which was filled during MIS 6 by the Middle Till, deposited by a more expansive glacial advance which reached the shelf edge. A second shelf edge glaciation during MIS 2 deposited the Upper Till, with a series of stacked grounding line deposits resulting from continued subsidence and an oscillating ice front. The stacked deposits are referred to as till tongues by King and Fader (1986) and King *et al* (1991) in glacimarine models of the margin.

The lower slope of the mid-Norwegian margin south west of the Vøring Plateau has been the site of repeated giant submarine slides collectively termed the Storegga Slide Complex by Evans et al (2002). The slides are the result of glacially-fed slope deposits which became unstable and probably triggered by seismic disturbance, moving an estimated 5600 km³ of sediment (Evans *et al*, 1996).

Canadian Margin

The glacial influence on the eastern margin of Canada in the Nova Scotia-Newfoundland region has been extensively studied (e.g. King and Fader, 1986; King *et al*, 1991; Gipp, 2003, Shaw, 2003). The Scotian margin consists of a wide shelf zone of nearly 200 km with a water depth of about 200 m, dropping to 4000 m at the base of the slope. The western shelf is crossed by canyon systems of fluvial origin which were extensively modified and exploited by later glacial activity (King and Fader, 1986). The inner shelf forms a seaward-sloping zone of irregular topography which is controlled by local pre-glacial lithology. Basins on the inner shelf contain a fill of acoustically stratified sediments. The remainder of the inner shelf is characterised by small moraines and rock outcrop at the seabed (Stea *et al*, 1998). On

the mid-shelf the basins are filled with a combination of massive and laminated fine-grained muds. Seismic data shows a basal chaotic package overlain by a stratified package. In addition, lensoid, tongue-shaped packages of chaotic reflections interpreted as till tongues (King *et al*, 1991) or debris flows (Stravers and Powell, 1997), iceberg scours and gas escape structures are also present. The outer shelf is dominated by sand and gravel banks with abundant shell debris (Amos and Knoll, 1987).

The upper Scotian slope shows three main seismic facies, comprising packages of stratified high continuity reflections, low continuity stratified reflections and chaotic or reflection-free packages which infill underlying topography (Gipp, 2003). On the lower slope the glacial section comprises packages of high continuity stratified reflections interbedded with lenticular chaotic units which are interpreted as debris flows (Gipp, 2003).

North-western UK margin

The northwest margin of the UK lies between about 56°N and 62°N (Figure 1.6) and is characterised by a shelf zone nearly 180 km wide in the area of the Inner Hebrides, to about 70 km wide in the area west of Shetland. This includes wide areas lying between 100m and 200m, especially in the north, which give very low slope angles to the outer shelf. The mid to inner shelf is characterised by numerous small basins and embayments, highlighted by the path of the 100m isobath (Figure 1.6). In addition, the coastline of Western Scotland, the Hebridean Islands and most of the Shetland Islands shows an abundance of elongate fjord-like inlets, many of which are overdeepened. The shelf break lies at a water depth of about 200m. The continental slope in the West Shetland area descends to about 1000m with an angle between 1° and 2°. At the base of the slope the angle decreases, leading into the deep basin of the Faroe-Shetland Channel. South of the Wyville-Thomsom Ridge the base of the slope lies much deeper, at water depths between 1500m and 2000m.

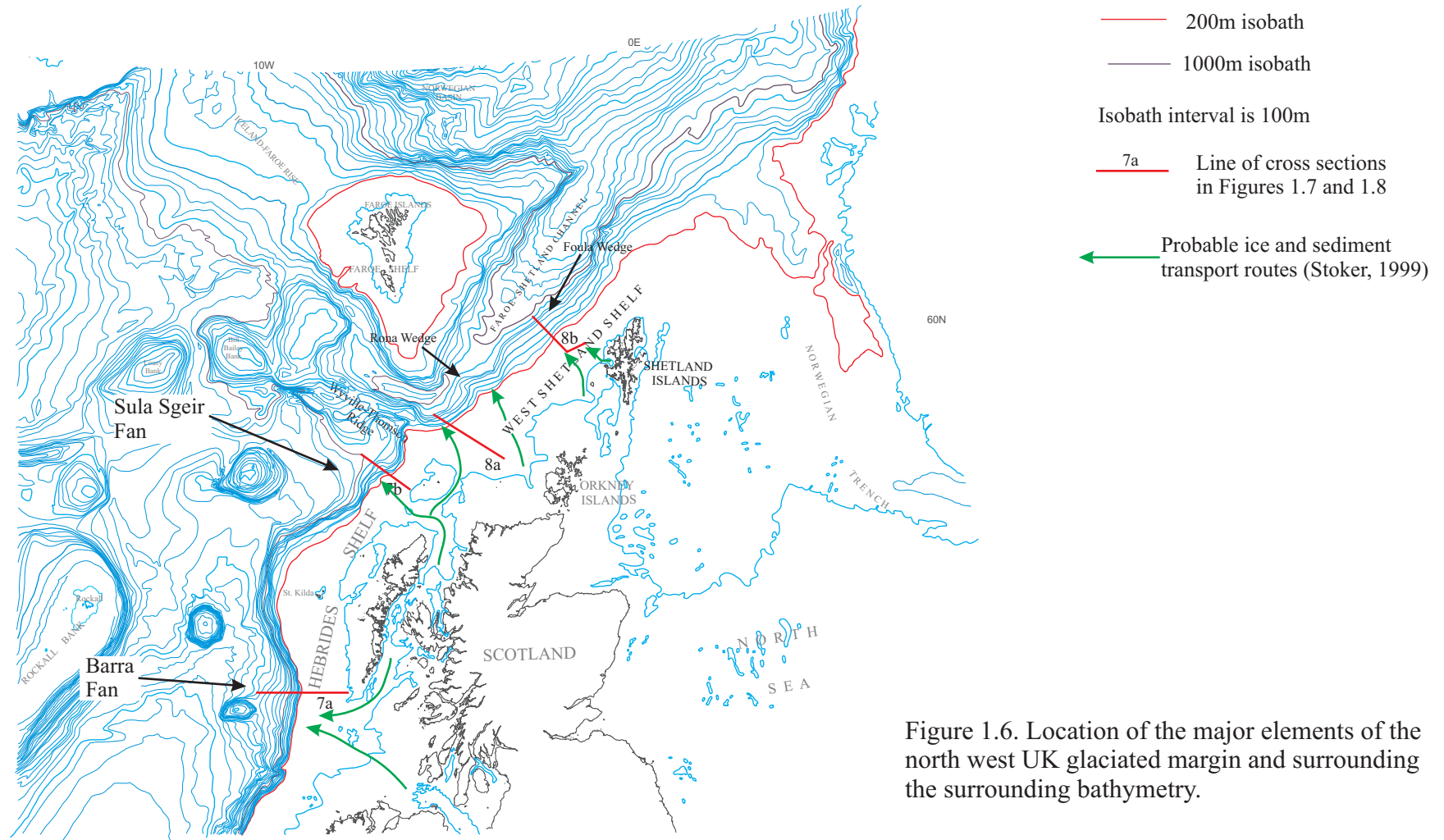


Figure 1.6. Location of the major elements of the north west UK glaciated margin and surrounding the surrounding bathymetry.

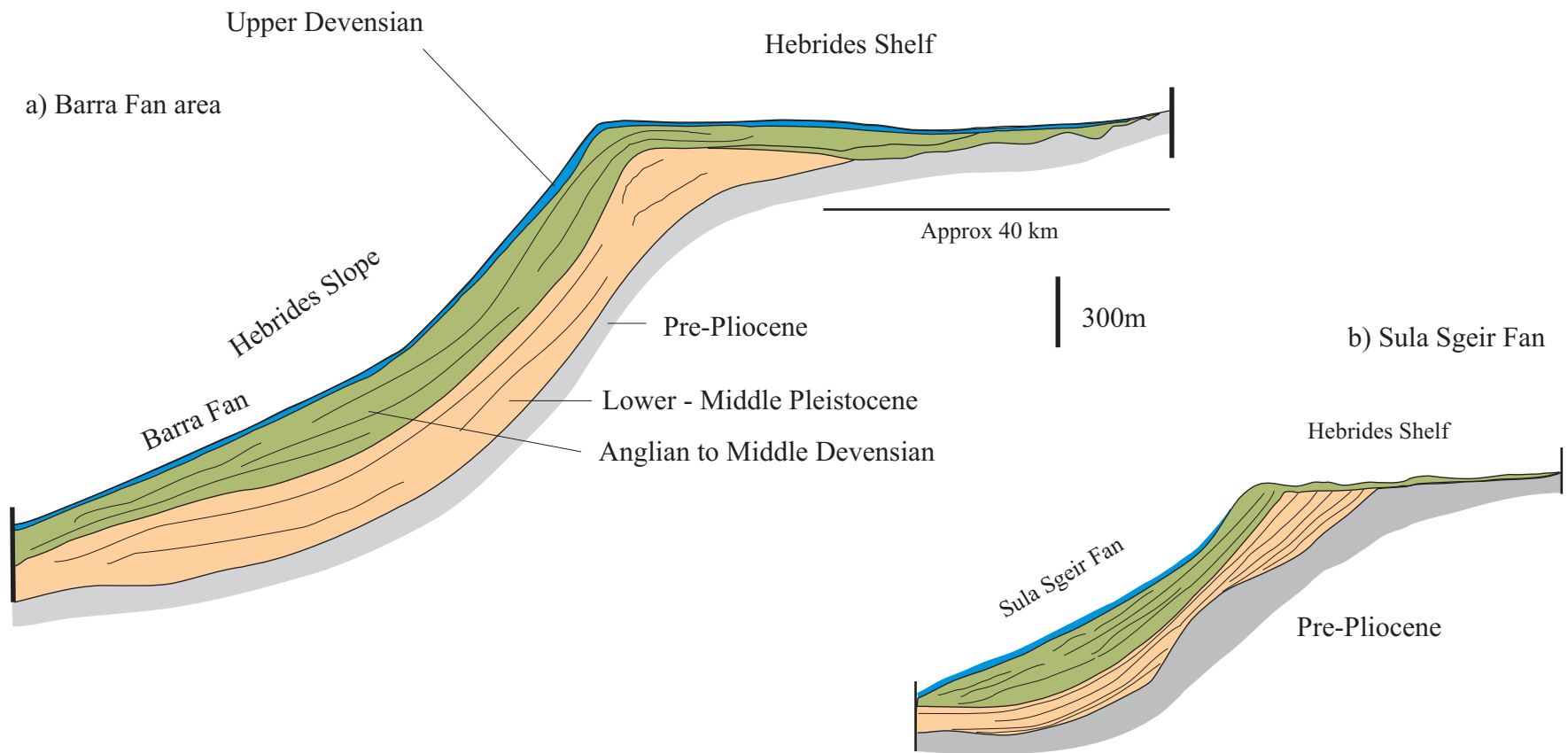


Figure 1.7. Schematic cross sections from seismic data across the Barra and Sula Sgeir Fans (Modified from Stoker et al, 1993).

The Quaternary of the northwest UK shows a similar broad-scale architecture across the whole margin. In the south, around the southern Hebrides the Quaternary sequence forms a thick, prograding wedge. Focused delivery of sediments to areas of the shelf edge deposited thicker sequences on the slope, especially in areas such as the Barra Fan (Figure 1. 7a). The Quaternary sequence from the Barra Fan and the shelf above it shows a prograding sediment wedge overlying a pre-Pliocene sequence. Lower to Middle Pleistocene sediments are preserved on the outer shelf and slope, but absent from the middle and inner shelf. The Middle to Upper Pleistocene (from the Anglian to the Upper Devensian) packages show a more sigmoidal architecture and are present from the slope base to the inner shelf.

To the north of the Outer Hebrides the Quaternary sequence on the shelf is much thinner and the Upper Devensian section appears to be absent, with a possibility that some areas may have been emergent (Stoker et al, 1993). The shelf to the south of Sula Sgeir shows a broad, deepened channel, indicating a probable pathway to the shelf edge (Figure 1.6). The Lower to Middle Pleistocene of the upper slope comprises a series of steeply dipping prograding sediments (Figure 1.7b). Any indication of aggradation has been removed during the emplacement of the later Pleistocene units. The Sula Sgeir Fan is developed on the slope below the outflow from the cross-shelf channel.

To the north, in the area northwest of Orkney and west of Shetland (Figure 1.6), the overall Quaternary architecture is similar, showing prograding wedges of sediments although the detailed structure is slightly different. In the area northwest of Orkney (Figures 1.6 and 1.8a), the inner and mid shelf comprises Anglian to Upper Devensian sediments resting upon pre-Pliocene strata. The outer shelf and slope are underlain by a wedge of Pliocene to middle Pleistocene sediments showing a prograding and aggrading sequence. This forms part of a thick accumulation of sediments which, together with the Upper Pleistocene deposits, form the Rona Wedge (Stoker, 1995). The Rona Wedge lies at the end of a broad area of deeper water on the shelf, suggesting a former sediment transport pathway.

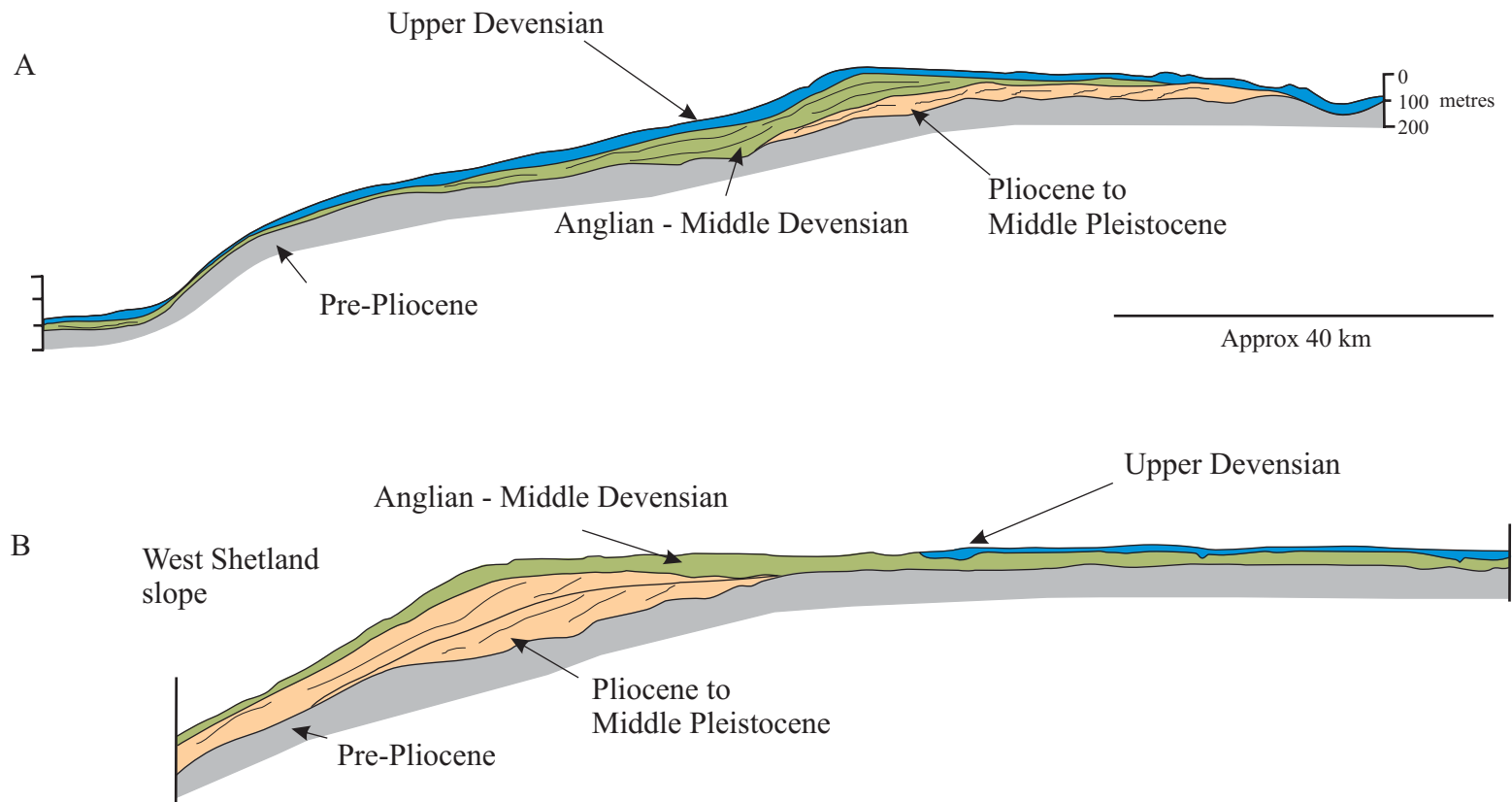


Figure 1.8. Cross sections across the West Shetland shelf and slope in the north (A) and south (B) of the margin (modified from Stoker et al, 1993).

The area west of Shetland (Figure 1.6) has a different shelf-to-basin architecture (Figure 1.8b) caused by a different pre-Pliocene slope profile. The effect is to produce a more elongate sigmoidal shelf margin section, termed the Foula Wedge (Stoker, 1995), with reduced slope angles on the upper slope and a steepening towards the basin floor. The Pliocene to Middle Pleistocene packages are restricted to the shelf and upper slope, with very little aggradation within the sequence. The overlying Anglian to Middle Devensian sequence shows a much greater degree of progradation, significantly extending the continental shelf (Figure 1.8b). There is also evidence of aggradation on the shelf, but this has been partly removed during the emplacement of the Upper Devensian sequence. The Upper Devensian sequence on the shelf varies in thickness, with elements infilling shallow basins and glacial moraines forming mounded deposits on the outer shelf. At the shelf edge, the packages show a prograding and slightly aggrading nature, with the packages thinning downslope. On the lower slope the Upper Devensian becomes very thin where modern currents are actively re-working the seafloor, removing the finer fractions of sediments and creating a scoured seabed.

1.3 Previous work on the Pleistocene glacial history and stratigraphic development of the margin of the NW UK

The influence of glacial activity on the NW margin of the UK began with phases of distal glacial deposition by rain-out from drifting icebergs in the Late Pliocene, however, the first major ice sheet advance began approximately 440Ka ago, which correlates with the Middle Pleistocene Anglian glacial stage (Stoker et al, 1993). The Anglian glaciation was followed by further glaciations, including the early and the late Devensian. For the purposes of this study, the Quaternary glacial sequence is regarded as being all the units above a regional unconformity at the base of the lowermost shelf-wide glacial unit generally termed the “Glacial unconformity” by Stoker (1999). This excludes earlier sediments which may have had an ice-rafted sediment input.

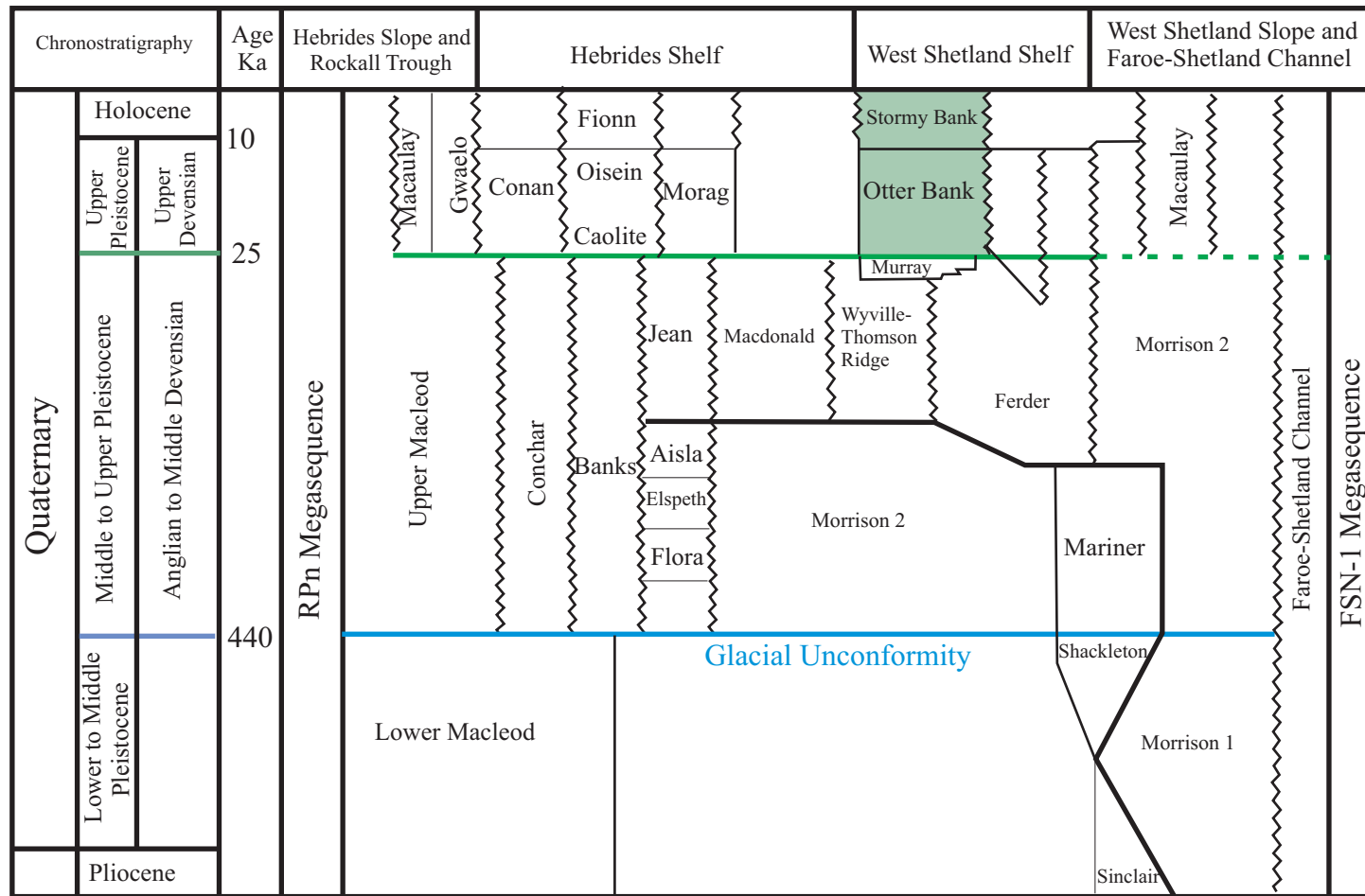


Figure 1.9. Simplified stratigraphy of the margin of Northwest Britain showing the main sequences relevant to this study, compiled from Stoker et al (1993) and Stoker et al (2005).

Stratigraphic framework

The Quaternary stratigraphy of the NW UK margin is complex and difficult to correlate across the whole margin, partly due to the scarcity of dated sediments and partly due to the nature of glacigenic systems re-working earlier deposits. Stoker et al (1993) summarised the Quaternary succession for the entire NW margin, the numerous sequences and formations defined primarily on the basis of seismic stratigraphy (Figure 1.9). However, problems in correlating between sequences on the shelf and the slope have led to different names for what is probably the same chronostratigraphic unit. More recent work on the larger scale Plio-Pleistocene development of the prograding sediment wedges has incorporated the entire glacial succession above the Glacial Unconformity on both the Hebrides and West Shetland margins, into megasequences RPa (Hebrides) and FSN-1 (West Shetland) (Stoker et al, 2005). However, for the purposes of this study, the higher resolution nomenclature of Stoker et al (1993) is used. Specific sequences of importance to this study in the west Shetland region are highlighted in Figure 1.9.

Glacial history

Evidence for the earlier phases of Pleistocene glaciation in the Anglian and Middle Devensian stages is mostly confined to the offshore record where they are represented by thick deposits on the outer shelf and slope. Onshore deposits are restricted to thin units on Shetland, Lewis and St Kilda.

In the region around the south Hebrides there is evidence of ice sheet advance to the shelf edge during the Anglian and early Devensian, with glacimarine sediments also deposited on the slope. On the shelf above the Barra Fan there are extensive examples of buried sub-glacial meltwater channels and mounded glacigenic sediments. The Barra Fan is composed of glacimarine sediments and debris flows from Late Devensian and earlier glaciations (Armishaw *et al*, 1998).

In the north Hebrides region, submarine end moraines of pre-Late Devensian age lie at the shelf edge on the outer edge of two glacially overdeepened basins (Stoker et al, 1993). The Sula Sgeir Fan, composed of glacimarine sediments lies on the slope

below the moraine banks. In contrast, the Upper Devensian glacial sediments on the shelf are restricted to glacimarine sediments, and there is a possibility that the ice did not reach the shelf edge (Stoker et al, 1993).

Previous studies of the Quaternary glacial sections in the Faroe-Shetland area have established an approximate limit for ice sheet advance for the Anglian/Early Devensian and Late Devensian glacial episodes (Stoker & Holmes, 1991. Stoker et al, 1993. Holmes, 1997). Evidence for the limit of the Anglian-Early Devensian ice sheets is poor as this has been largely modified by the Late Devensian advance. However, the presence of a thick pre-Late Devensian accumulation of glacial sediments on the upper slope suggests the ice reached the shelf edge. The limits for the Late Devensian ice sheet are based upon a series of moraine ridges from the Otter Bank sequence (Figure 1.9) on the outer shelf. The moraines are characterised by a ridge-like profile up to 50m high, 8km wide and 60km long (Stoker and Holmes, 1991). Acoustically these ridges are structureless to chaotic, with hyperbolic reflectors. Shallow cores and borehole data from the ridge crests indicate the features are composed of diamictons, sands and gravels, with larger cobbles and boulders inferred from the hyperbolic reflectors (Stoker et al, 1993). The remainder of the outer shelf is covered by a sheet-like diamicton and glacimarine muds.

Evidence of abundant iceberg ploughmarks on the outer shelf and upper slope has been provided by side scan sonar surveys. Belderson et al (1973) recorded ploughmarks between water depths of 140 and 500m. The average dimensions of these features are approximately 20m wide and 2m deep with a maximum recorded length of 5.5km. However, examples up to 280m wide and 10m deep have been recorded (Stoker et al, 1993). Two sections across the West Shetland margin are shown in Figure 1.9.

The Otter Bank sequence is considered to be of Late Devensian age, based upon its relationship to dated sequences in the Sula Sgeir region (Stoker et al, 1993). The overlying Stormy Bank sequence is assumed to be a late stage of the Late Devensian glaciation. In addition, there are lateral equivalents of both sequences on the slope

which form undefined parts of the Morrison and Macaulay sequences. In this study, the Otter Bank sequence and the Stormy Bank sequence are treated as one unit, partly because of the difficulty in separating the two sequences on seismic profiles and the lack of lateral continuity of the Stormy Bank sequence. The unconformity at the base of the Otter Bank sequence is extended onto the slope, providing a seismic-stratigraphic division which links together the Late Devensian sequence across the whole West Shetland margin.

1.4 Structure of Thesis

This study is structured in a way so as to lead the reader through a logical progression of evidence and interpretation of the West Shetland Margin. In Chapter 2, the methodology of this study is addressed, detailing the different data sources used, their limitations and techniques employed in their analysis and interpretation. The division of the West Shetland Margin into two broad shelf-to-basin transects to aid characterisation of the different geographical provinces is shown.

Chapter 3 presents an overview of the physiography and broad scale architecture of the glacial sequence of the West Shetland Margin. The summary relates the bathymetric and topographic data to the underlying solid geology, highlighting intra-shelf basins and possible sediment transport pathways. The total thickness of the Quaternary glacial sequence is defined and a representative seismic traverse from the mid-shelf to the floor of the Faroe-Shetland Channel used to highlight the distribution of the glacial sequence.

Chapter 4 examines the detailed data from the southern transect in five sections; inner shelf, mid-shelf, outer shelf, upper to mid slope, and lower slope and basin floor. In addition, a further area offshore from the transect is examined as it presents different characteristics and an alternative interpretation. The chapter identifies the Late Devensian sequence in both boreholes and seismic survey data and presents a generic interpretation of each of the provinces. The chapter ends with an overall interpretation of the southern transect data and links the depositional environments and facies together to characterise the southern area of the margin.

The northern transect is examined in Chapter 5, which includes data from onshore localities on Shetland and near-shore data from St. Magnus Bay. Areas from the inner, mid and outer shelf are also described and interpreted. A summary and interpretation of the whole transect is presented at the end of the chapter.

Chapter 6 combines the information from the two transects, extending the mapped thickness of the Late Devensian glacial sequence across the whole study area and reconstructing ice sheet movements and limits on the West Shetland Margin. From the data examined from the West Shetland area, a conceptual model of the Late Devensian glaciation, together with idealised and actual sequences deposited and preserved is presented. The preservation potential of the deposits is then related to the rock record and a brief comparison made between the theoretical sequence predicted by the model and actual ancient glacial deposits.

The conclusions of this study are presented in Chapter 7.

2 Methodology and data sources

The main sources of data for this study are from British Geological Survey (BGS) and commercial seismic reflection surveys, a shaded relief seabed image derived from 3D commercial seismic, side-scan sonar imagery from the Atlantic Frontiers Environmental Network (AFEN) and core material from boreholes. The large volume of data potentially available for the West Shetland area is too great for this study to address in detail, the seismic lines alone cover an area in excess of 17,000 km². As a result, two broad transects were chosen across the West Shetland Margin in a manner such that most of the depositional provinces identified from the seabed image were traversed (Figure 2.1). The transects also co-incide with a best-fit line between the boreholes used in this study. Areas of particular interest, identified primarily from the seabed image or seismic data, are described in detail in chapter 4. These areas show features of particular significance to the interpretation of the glacial history of the West Shetland Margin.

2.1 Seismic reflection survey data

The majority of the seismic data used in this study is from 2D airgun and sparker analogue surveys undertaken by BGS between 1979 and 1985, forming 3 blocks of survey lines that provide gridded coverage for the outer shelf and slope (Figure 2.2). Most of the seismic lines represent dip and strike lines relative to the continental margin, though the 1979 survey was shot on an east-west grid. Records from two earlier surveys in 1974 and 1977 have also been used to supplement the main surveys in areas where additional detail was required. These data are referred to in the text where appropriate

2.2 Seismic facies descriptions

The description of the seismic data broadly follows the seismic facies terminology defined by Mitchum et al, (1977 part 6) and modified in Stoker et al (1997). However, the sequence stratigraphic terminology and models which are commonly used in the interpretation of marine successions (e.g.; Vail et al, 1977 and Sangree & Widmier, 1977) do not generally take account of glacial marine deposition.

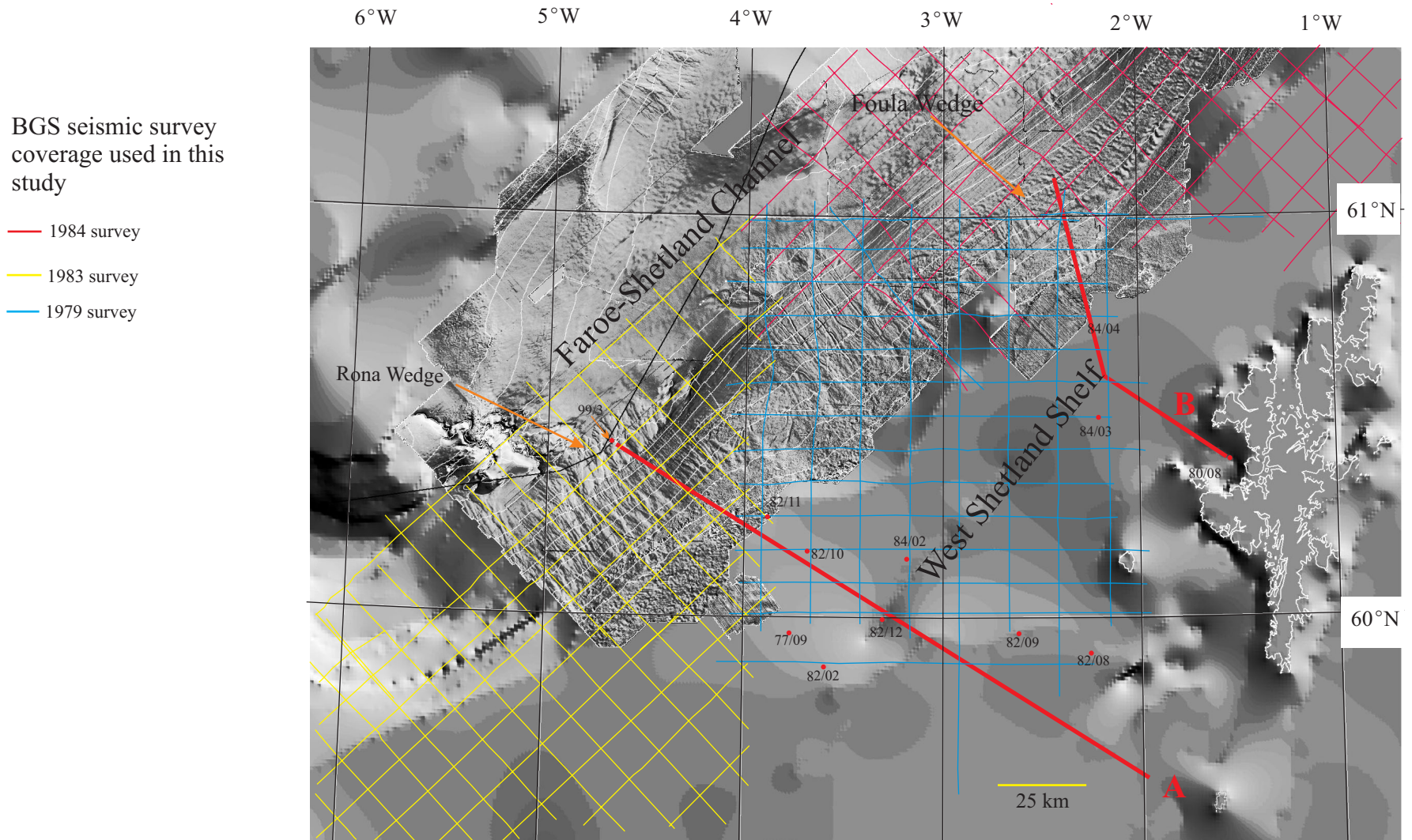


Figure 2.1. Position of shelf-to-basin transects A and B, across the West Shetland Margin, and the location of the BGS boreholes referred to in the text. These data are superimposed on the sea-bed image derived from 3D seismic data.

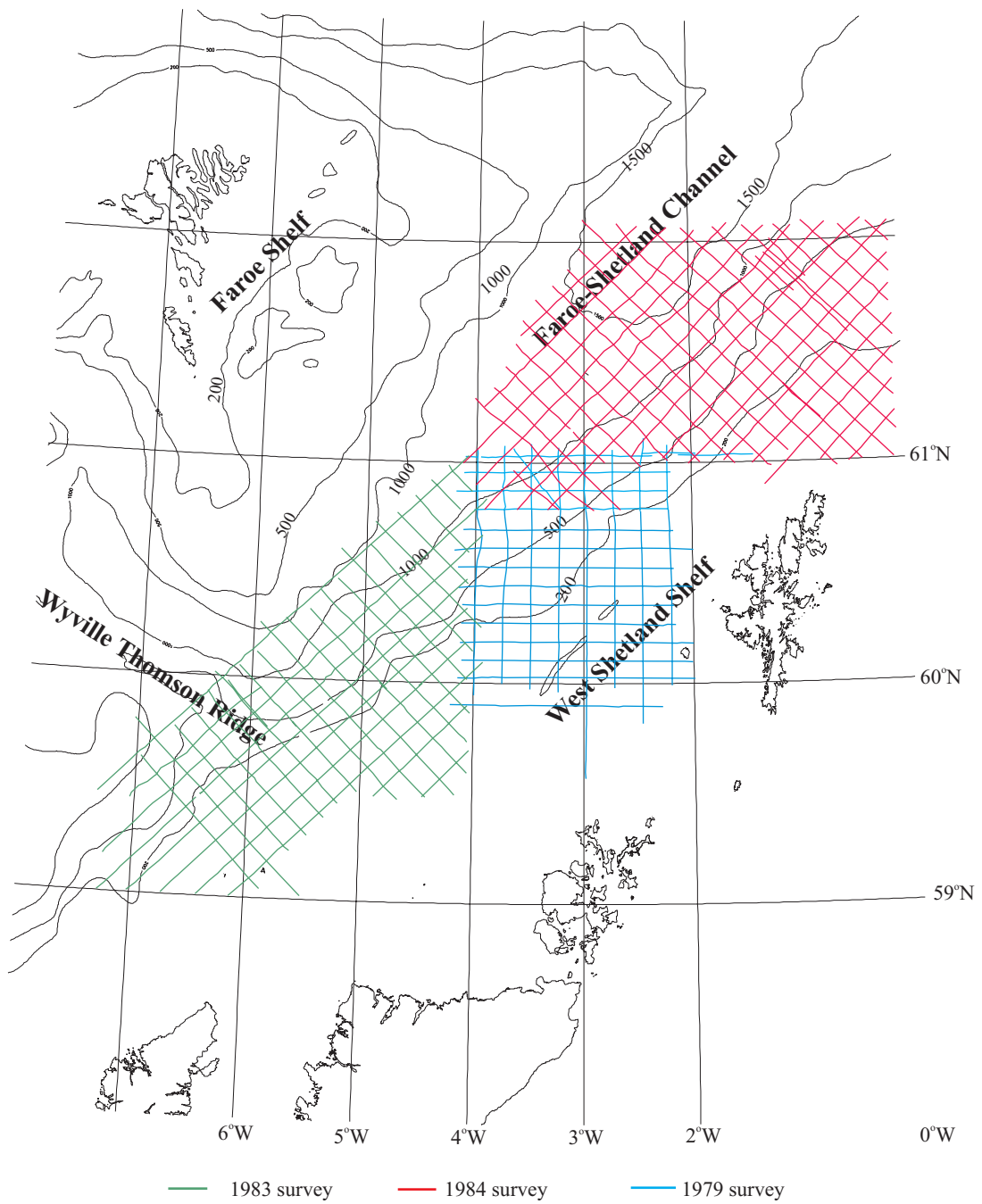


Figure 2.2. Track chart of BGS seismic survey coverage used in this study

The description of seismic reflection data falls into three main categories; reflection character, reflection configuration and package architecture. These properties are used to describe both the bounding reflector of units and the internal reflectors.

2.2.1 Reflection character

The character of seismic reflections is governed by a combination of several properties: i) amplitude, ii) frequency and iii) continuity.

Reflection amplitude: The amplitude of a reflection is a result of the difference in acoustic impedance between two units; the greater the contrast, the higher the amplitude. The amplitude is generally described as being high, moderate or low (Figure 2.3). High-amplitude reflections tend to occur where there is an abrupt change in lithology or in sequences of interbedded lithologies, although they can also be caused by interference effects. In Quaternary glacial successions high amplitude reflections can also be used to identify boundaries between over-consolidated and normally consolidated deposits (Stoker et al, 1997).

Reflection frequency: The frequency of seismic reflections is largely dependent upon the thickness of individual beds. The frequency cycle is described as being broad, moderate or narrow. Lateral changes along the same reflector or group of reflectors can infer lateral facies changes, although noise and interference can cause the same effects.

Reflection continuity: The continuity of reflections is generally related to the actual continuity of lithological units, but can also be related to other physical properties, such as shear strength or saturation which may not be obvious in samples (Figure 2.3). Reflectors with a high degree of continuity tend to be mud-prone and deposited in low-energy environments. High-energy environments generally produce discontinuous reflections. From a glacial viewpoint, diamictons tend to produce discontinuous reflections and scatter acoustic energy due to the inclusion of outsized clasts up to boulder size. Individual boulders may be imaged as point-source hyperbolic reflections (Figure 2.4). Where there are diamictons with a high

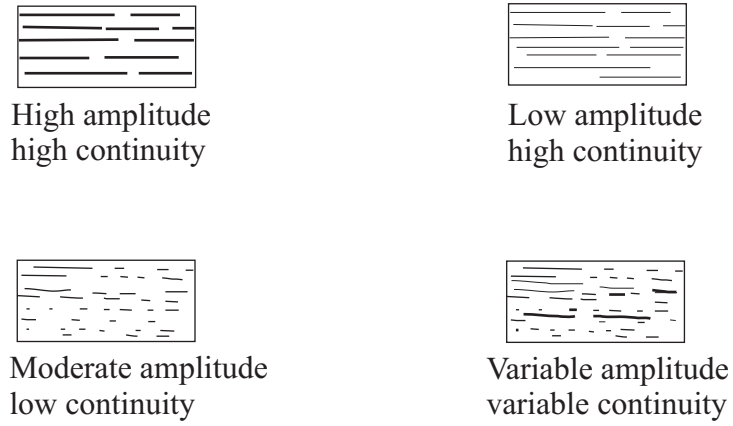


Figure 2.3. Reflection energy patterns, based upon Sangree and Widmier (1977).

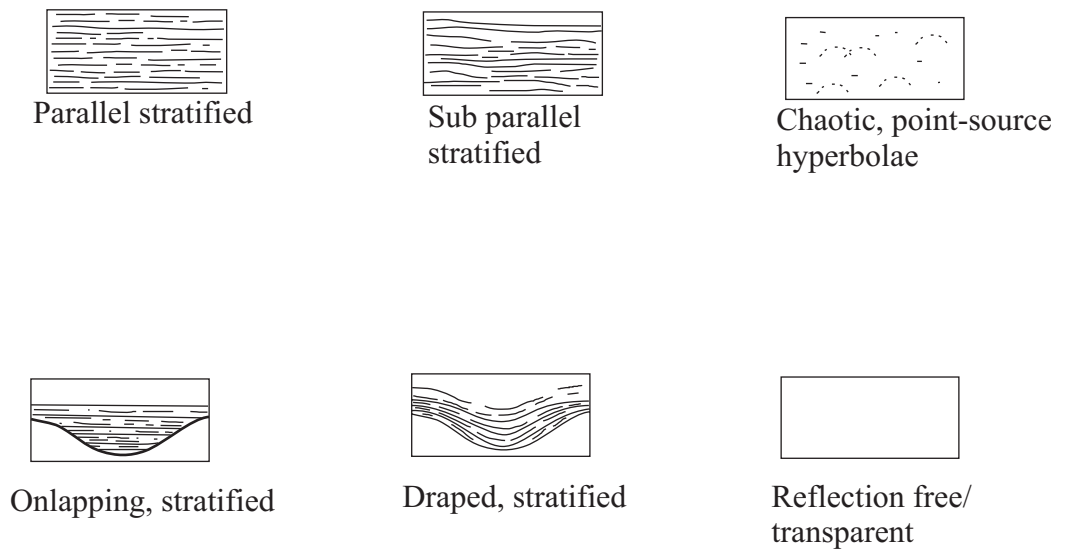


Figure 2.4. General reflector patterns. Based upon Stoker et al (1997).

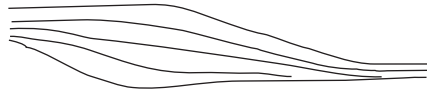
proportion of clasts the short reflections can become chaotic or fail to produce coherent reflections at all.

2.2.2 Reflection configuration

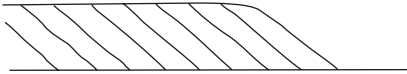
The configuration of reflections and reflection packages represent the nature of a surface or reflecting horizon. The patterns fall into three main groups: i) stratified, ii) chaotic and iii) reflection free or transparent. The nature of the reflection configuration has important implications for bedding patterns, erosional and depositional processes and palaeo-topography (Figures 2.4 and 2.5).

Stratified reflections: Of the different types of patterns, the stratified reflection pattern yields most information with regard to the sub-surface geology (Figure 2.4). The patterns vary from a basic parallel horizontal form to complex diverging and converging patterns. Parallel and sub-parallel configurations often indicate sheet-like bed forms draping surfaces especially on shelves. Stratified patterns commonly form under low energy conditions and in a glacial context, may indicate suspension settling as a major process (Stoker et al, 1997). Diverging or converging reflections on a small to moderate scale can indicate differences in sedimentation rates, whilst on a basin-wide scale are more likely to indicate variations in rates of subsidence/uplift. Onlapping and draped patterns may reflect style of sedimentation (e.g. axial flow, suspension settling), and variation in the nature of the palaeo-surface of deposition.

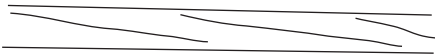
More complex configurations commonly include clinoform reflections. Broad groups of recognised patterns include sigmoidal, oblique and shingled (Mitchum et al, 1977) and are shown in Figure 2.5. In general, clinoform configurations are associated with depositional systems which prograde into a basin. The exact nature of the patterns are governed by the rate of sediment supply, degree of basin subsidence and stability of sea levels. High rates of deposition combined with little or no subsidence and stable sea levels tend to produce an oblique pattern, characteristic of higher energy environments. Sigmoidal forms are more typical of lower energy environments with a lower sediment input and rapid subsidence/sea level rise. However, glaciated



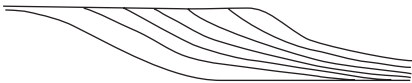
Sigmoidal



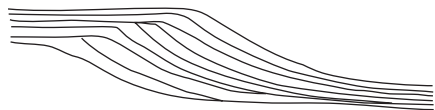
Oblique parallel



Shingled



Oblique tangential



Complex sigmoidal-oblique

Figure 2.5. Examples of common clinoformal reflector configurations (Stoker et al, 1997).

continental margins are subject to rapid sediment deposition on the shelf and slope at a time of low stand or rising sea-levels. Sediment loading coupled with the effect of ice loading creates accommodation space on the shelf, allowing the development of an aggrading and possibly prograding sequence. This forms a more complex sequence referred to as complex sigmoidal-oblique (Figure 2.5) rather than the basic sigmoidal pattern.

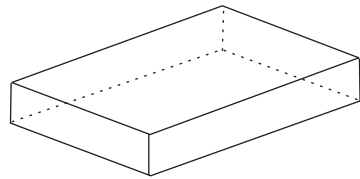
Chaotic reflections: Chaotic reflections, as the term suggests, have no discernable structure (Figure 2.4). Sediments with this seismic pattern tend to be deposited rapidly or have been re-mobilised. Within glacial and glacio-aquatic environments, massive diamictons in the form of tills and debris flows are the main source of such reflections. The irregular distribution and varying size of clasts within sediments can also produce chaotic patterns due to refraction of the seismic energy.

Reflection-free patterns: Reflection-free patterns are the result of lithologies with no internal acoustic impedance. This is most likely to be found in homogenous lithologies such as massive mud deposits, especially those which have been re-mobilised as debris flows. High gas content within a sediment prevents transmission of seismic energy and can also produce a reflection-free response, although this is unlikely to be a consideration within glacial sediments on the shelf due to the reduced level of biological productivity in a cold glacial setting and the greater compaction of sediments deposited beneath the ice.

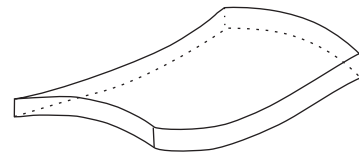
2.2.3 Package architecture

The overall architecture of the seismic units provides important indicators when defining the large-scale depositional environment and geological setting.

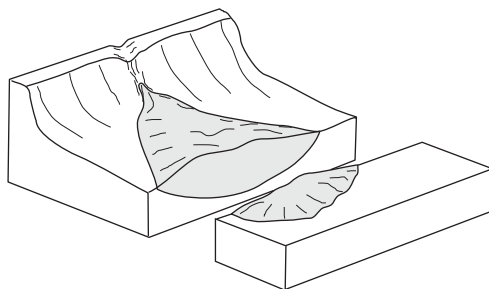
By combining individual 2D seismic lines the 3D architecture of seismic facies packages can be determined, although the resolution is dependent upon the line spacing. The general forms of the packages, and on a larger scale the sequences, fall into a group of terms defined by Mitchum et al (1977) as sheets, sheet drapes, wedges, lenses, banks, mounds and fills (Figure 2.6). The different forms of seismic packages are the basic building blocks of the Sequence Stratigraphy models of



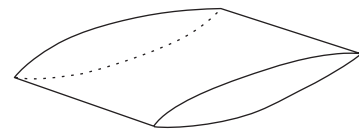
Sheet



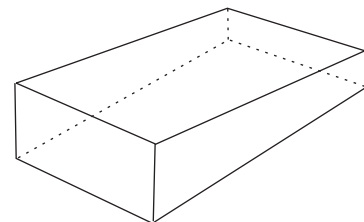
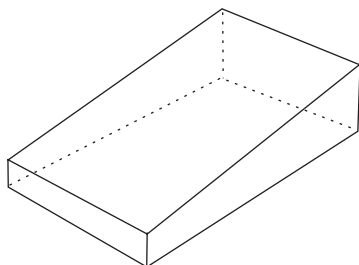
Sheet drape



Fan/lobe



Lens



Wedge forms

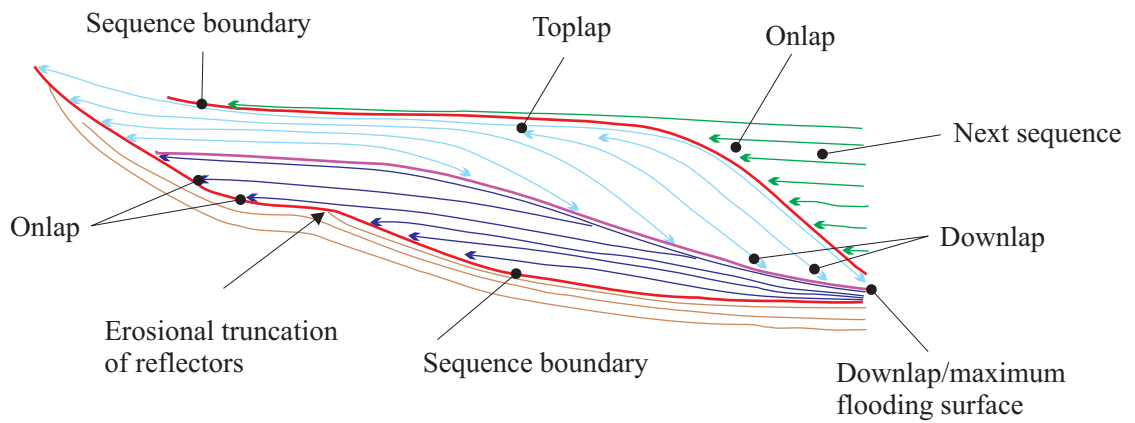
Figure 2.6. Examples of seismic package architecture (modified from Mitchum et al, 1977).

Mitchum et al (1977), Vail et al (1977) and Van Wagoner et al (1988). The depositional sequences used in these models are defined in relation to bounding unconformities and the manner in which reflections between them terminate. The style of reflection terminations indicates whether an unconformity resulted from non-deposition or erosion. The terminations resulting from non-deposition are termed onlap, downlap and toplap, depending upon their position in the sequence (Figure 2.7). Erosional episodes are marked by the truncation of reflections. The descriptive terminology from these models is used in this study where applicable.

In models of Vail et al (1977) the sequence boundaries are related to changes in relative sea level, and to global-scale cycles in sea levels through geological time.. These sea level changes are the result of tectonic changes affecting the volume of ocean basins and of major glacial events. Essentially, shelf-margin progradation is related to a sea-level lowstand, whereas transgression and coastal onlap reflects a highstand scenario. The timescale of Quaternary fluctuations in sea level is much shorter than the first to third order changes suggested by Vail et al (1977), indicating the limitations of the established sequence stratigraphy models in relation to glacial events and their effects. For example, widespread erosion of the shelf may occur by glacial processes despite being below the contemporary sea level. The basic seismic stratigraphy however is still similar. Recent work by Powell and Cooper (2002) has addressed some of the problems in the application of sequence stratigraphic models to glaciated margins by developing a Glacial Systems Tract model.

2.3 Analysis of the West Shetland Seismic data

Data from the BGS seismic surveys was supplied in the form of paper records. The unconformity marking the base of the Quaternary glacial section (as defined in Chapter 1) had been pre-picked on some records by an earlier study and correlated with lithological boundaries in two of the boreholes on the shelf. The identification of the glacial unconformity was based upon the truncation of underlying reflections and correlation with borehole lithologies and limited biostratigraphic data. Using this information, the unconformity was traced across the remainder of the seismic grid, although in the shallow water of the inner shelf the resolution of the data was



- High stand systems tract - formed during stable high stand sea-level, requires basin subsidence to allow aggradation of deposits.
- Transgressive systems tract - formed during rising sea-level/marine transgression.

Figure 2.7. Seismic and sequence stratigraphic terminology, based upon Stoker et al (1997) and Vail et al (1977).

insufficient to distinguish the unconformity. The availability of new deepwater borehole cores allowed the depth of the unconformity to be fixed in the lower slope and basin floor which was not previously possible. Following the regional identification of the glacial unconformity, the thickness of the total overlying glacial section was mapped out from the seismic data and expressed as an isopach map (see Chapter 3, Figure 3.8).

The base of the late Devensian glacial section west of Shetland has previously been identified from seismo-stratigraphic correlations with dated units to the north of Shetland (Johnson *et al*, 1993; Stoker *et al*, 1993). Radiocarbon dates of approximately 18, 860 BP obtained from bivalve shells place the units within Marine Isotope Stage 2. Identification of the base of the late Devensian section on seismic profiles allowed this to be translated to the borehole cores. In this way it was possible to map out the thickness of this unit and also to define architectural elements within it (Chapter 4). For the purposes of this study, the Holocene record was included with the late Devensian as discrimination was not always possible. In many areas the thickness of the Holocene record is below the seismic resolution and is lost in the seabed pulse.

In addition to the isopach map, a new bathymetric map was compiled from a combination of published sources, primarily the BGS 1:250 000 seabed sediment map series. This enabled the production of a palaeosurface map for the base of the glacial section when combined with the isopach data. This exercise was repeated for the Late Devensian section.

2.4 Seabed image

The West Shetland Shelf and Faroe-Shetland Channel region have been subject to intensive hydrocarbon exploration since the late 1980's. The Western Frontiers Association (WFA), a consortium of oil companies and the BGS, was formed in 1995 to investigate the sea-bed and shallow geology for health and safety purposes related to sea-bed stability and offshore installations. Use of 3D seismic in field development and regional reconnaissance in the 1990's resulted in widespread

coverage of the Faroe-Shetland area. These surveys were designed for targets below 4000m and had a typical spatial sampling rate of 12.5m. The close sample spacing allowed detailed reconstruction of the seabed topography in a pilot study by Bulat and Long (1998). An expanded version of this study was produced by Bulat (2000) incorporating data from the Faroese Geotechnical Environmental Metocean Network (GEM). The resulting 3D shaded relief image (Figure 2.1) has allowed the identification of the major architectural elements and possible depositional processes operating on the shelf and slope. Whilst the image represents the modern seafloor, the major components are thought to be preserved relict features from the last major glacial event in the late Devensian. These features do not appear to have undergone any significant modification since their formation in the late Devensian. Parts of this image have already been published together with proposed interpretations of the seabed morphology and sedimentary processes (Holmes *et al*, 2003. Bulat & Long, 2001. Davison & Stoker, 2002 . Leslie *et al*, 2003). Further interpretation of this image forms an important part of this study.

2.5 Side-scan data

In addition to the WFA seabed image generated from commercial seismic, a set of published side-scan sonar data acquired by the Atlantic Frontiers Environmental Network (AFEN) group was also available. This data set covers much of the West Shetland slope and outer shelf and, although there is considerable overlap with the WFA image, it provides important extra detail of the seabed in some of the shallow water zones.

2.6 Borehole cores

The major source of geological data for this study is a series of cores from 12 boreholes drilled by the BGS in the Faroe-Shetland region (Figure 2.1). The boreholes were drilled between 1977 and 1999 mainly in areas of the inner and outer shelf, but also including the base of the continental slope. The rotary drilling method used produced a core with a 7.6 cm diameter. The boreholes were continuously cored in runs of between 1 and 5 metres, with the thickness of cored glacial sections reaching between 1m and 60m.

Information from the shelf boreholes has previously been published on BGS Quaternary maps of the West Shetland region. However, the borehole drilled at the base of the slope – borehole 99/3 – provides a new data set that has not previously been published prior to the study of Davison and Stoker (2002) on the Rona Wedge (Figure 2.1) and which penetrates the entire preserved deep water Quaternary section. This was drilled by the *MS Bucentaur* during June and July 1999. Drill-site seismic data and gravity cores were obtained on an earlier cruise in 1998 by the *RRS Challenger*. Information from the gravity cores was combined with that from the drilled cores as the drilling recovered no material in the interval between the sea bed and approximately 3m below the sea bed.

In general, core recovery from the selected boreholes was very variable, ranging from approximately 50% to less than 10%. Across the boreholes, the percentage recovery appears to be independent of lithology, although this is not true within individual boreholes. This may be more related to drilling conditions than to particular properties of the sediments.

The state of preservation of individual cores, and the amount of post-recovery alteration was extremely variable. The cores drilled in the 1999 survey have been kept in cold store in sealed bags since drilling and showed little sign of surface oxidation or dehydration, although there was some evidence of drilling disturbance in sand units. The older cores from the 1980's surveys showed extensive alteration and complete desiccation. This has resulted in a colour change from the oxidation of iron-rich muds, a change in the consistency of muds from soft to a very hard rock-like texture, and in some cases disintegration of the core. A limited number of photographs from a study in 1985 (Cockcroft, 1987) were available, which enabled an assessment of the extent of alteration to be made.

Logging of the cores was carried out by visual examination for identification of lithology, mineralogy, grain morphology, sedimentary structures, colour and biogenic material. Grain size was generally determined by visual estimation under a binocular microscope with the aid of a set of in-house standard samples. However,

cores from Transect A (Figure 2.1) were sampled at irregular intervals and 65 samples examined in more detail to moderate the visual estimation. In addition, data from previously published records was also incorporated. The lithology of pebbles and larger grains was noted in an attempt to identify any provenance indicators.

Detailed particle size analysis was carried out using a Coulter Counter in the Geology and Geophysics department of Edinburgh University. Samples were freeze dried, weighed and rehydrated for at least 24 hours, then wet sieved using a 500 μ m sieve. The >500 μ m fraction was retained and the <500 μ m fraction analysed in the Coulter Counter. The coarse fraction was dried and weighed and the weight percentage calculated. The results from the Coulter Counter are expressed as a volume percentage and so are not directly comparable with those from the coarse fraction. A section from borehole 99/3 was sampled in detail at 5cm intervals across a lithological boundary between depths 46.60m and 47.35m. By sampling in this manner it was hoped to demonstrate a change in grain size and ice-rafted debris (IRD) content between the two units.

Cores from the 1999 survey were selectively photographed using X-ray photography in order to identify sedimentary structures and to estimate the proportion of clasts present. This was carried out using the BGS Scanray AC120L instrument. A combination of whole and split cores were X-rayed using a range of exposure times of up to 3.5 minutes at 85kv and 5mA. Unfortunately most of these X-ray photographs revealed little sedimentological information and due to the high water content of the sediments, proved too opaque to usefully analyse. An attempt to produce a contact print from the negatives was also unsuccessful, even after an exposure time of 20 minutes.

3. Physiography and glacial architecture of the West Shetland

Margin: an overview

This section presents an overview of the physiography and large-scale architecture of the glacial section preserved on the West Shetland margin. Bathymetric and topographic data, the 3-D seabed image and side-scan sonar data are utilised from physiographic description; seismic reflection profiles provide a sub-seabed perspective on the control of physiographic features.

3.1 Physiography derived from bathymetric and topographic data

An examination of the present day bathymetry of the West Shetland Margin and topography of the Shetland Islands shows large scale features important to the reconstruction of the glacial history of the region. The details of the topographic features of the onshore region of the Shetland Islands are different in orders of magnitude to the bathymetric features described in the offshore region. This is partly the result of data availability; onshore topographic surveys being much more detailed than offshore bathymetric surveys, and also a reflection of actual differences in the morphology and preservation of features on the sea floor. The bathymetric data for this study is taken primarily from a combination of surveys on published BGS 1:250 000 sea-bed sediment maps and from other geophysical surveys, and represent a new compilation primarily for the shelf (Figure 3.1) Shelf bathymetry is shown at 10m intervals between 50m and 200m; bathymetry beyond the shelf, below 200m, is at 50m intervals. The bathymetry between 0m and 50m depth is too complex to show on a map of this scale and is omitted. The resolution of the surveys on the slope is insufficient to reveal some of the major features shown on the sea-bed image, and so is only useful as a general guide. The map especially also highlights the deeper, enclosed parts of the on-shelf basins important in this study. In addition to the basins, Figure 3.1 shows the low points in these basins which may have formed outflows to other parts of the margin. These possible pathways are important in considering the final stages of glacial retreat at the end of the Pleistocene.

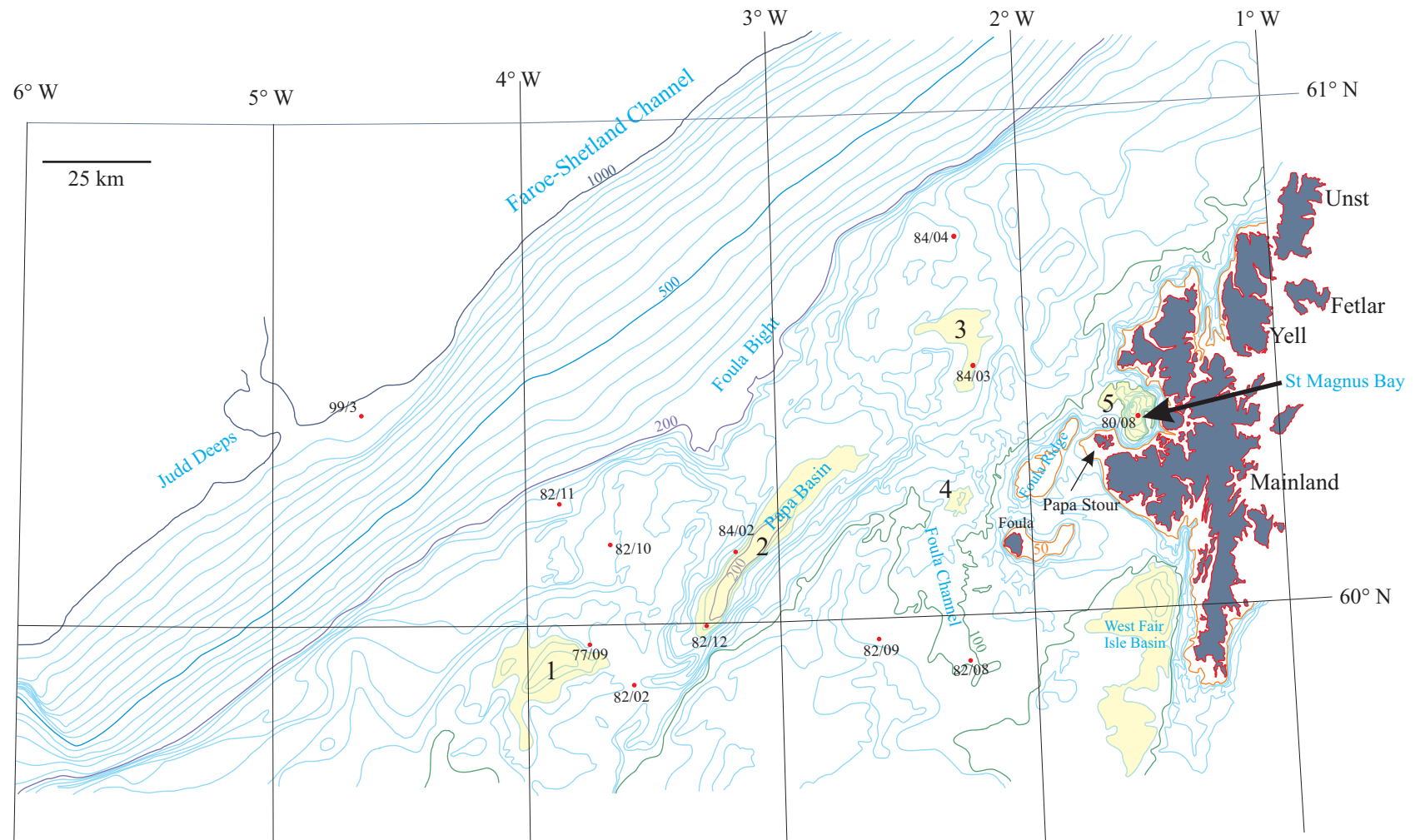


Figure 3.1. Bathymetry and offshore basins of the West Shetland area. The isobath interval from 50m to 200m is 10m; below 200m the interval is 50m. The shaded areas show overdeepened basins on the shelf. The numbers are basins referred to in the text. The map also shows the position of the BGS boreholes used in this study.

3.1.1 Shetland Islands

The onshore relief of the Shetland Islands is broadly asymmetric and strongly lineated. On the eastern side of the islands the land is generally low-lying with rounded topography and low coastal areas. In contrast, the western side is more rugged with greater relief and high, very steep to vertical sea cliffs. The topographic lineation is controlled by a series of major faults which lie roughly north – south through the whole island group (Figure 3.2). On the western side the lineation is also influenced by the orientation of foliation in a series of metamorphic and igneous rocks and by bedding in steeply dipping sediments. The inland lochs and marine inlets on the islands and around the coasts tend to show an asymmetry, being steeper on the eastern side than on the west, giving the impression of a half-graben structure with a down throw to the west. All of these basins have apparently been modified by ice sheets which covered the islands and may at some time have overridden the whole group of islands (Flinn, 1983). The landscape generally is described as strongly ice moulded with striated surfaces which have been used to reconstruct former ice flow directions. The moulded landforms and striated surfaces indicate a ice-shed lying along the long axis of Mainland, Yell and Unst in a roughly north-south direction shown in Figure 3.3 (Flinn, 1977).

The coastline around the Shetland Islands also shows numerous elongated inlets orientated roughly north-south which superficially resemble fjords formed by glaciation. However, the major inlets such as those headed by Weisdale Voe, Sullom Voe and Yell Sound have a considerable degree of fault control (Figures 3.2 and 3.3) and may only have been modified by glacial processes which exploited pre-existing basins. The present morphology of the Shetlands shows a drowned coastline typical of areas subject to rising sea levels. The drowning of the Shetland coastline appears to be relatively recent, submerged peat beds in Lerwick Harbour and Sullom Voe lying up to 9m below current sea-level and yielding radiocarbon ages of *c.*7000 - 5500 years bp (Hoppe, 1965).

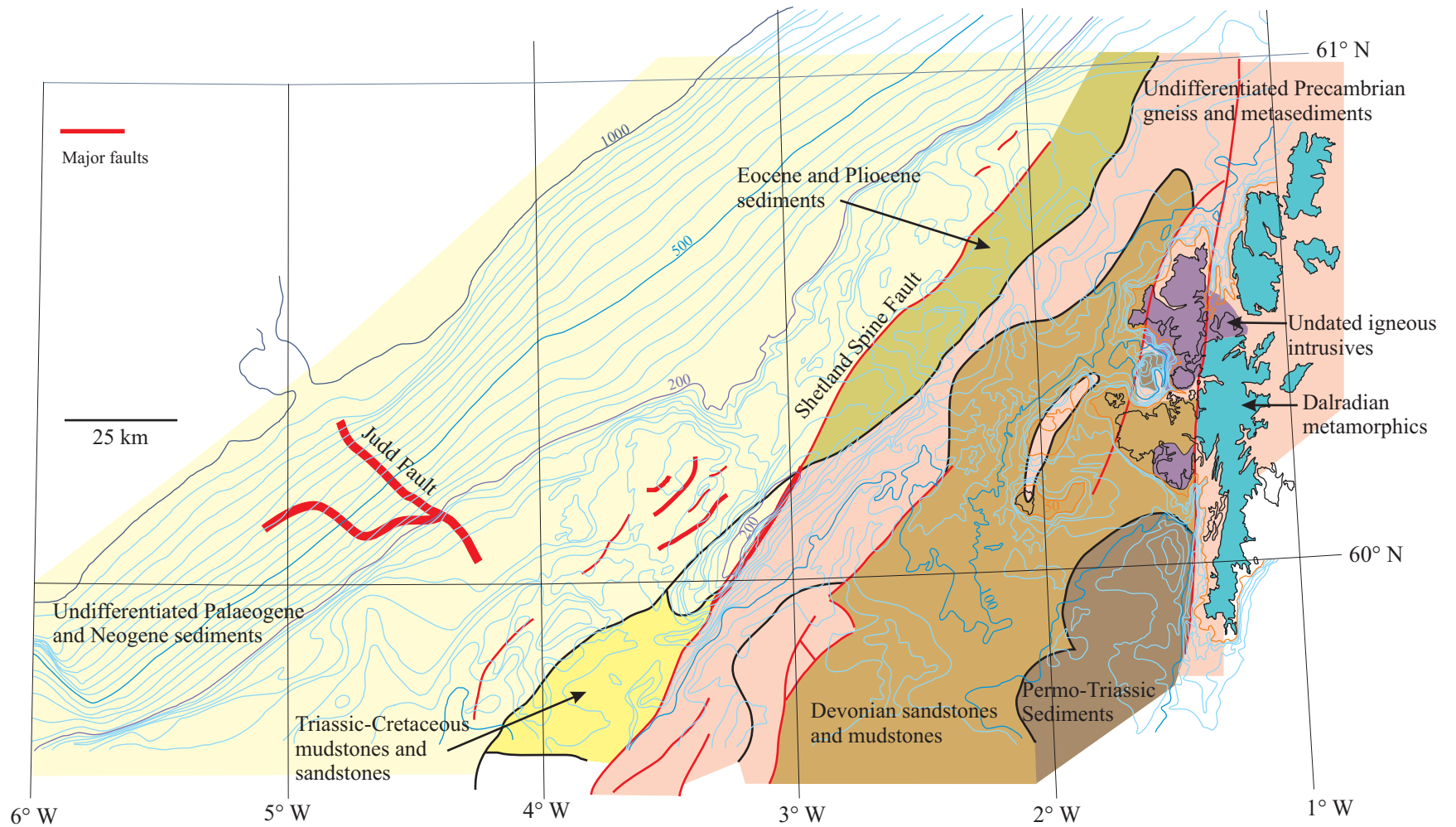


Figure 3.2. Simplified solid geology of the Shetland Islands and West Shetland Margin with superimposed bathymetry from Figure 3.1. Major fault lines are shown in red

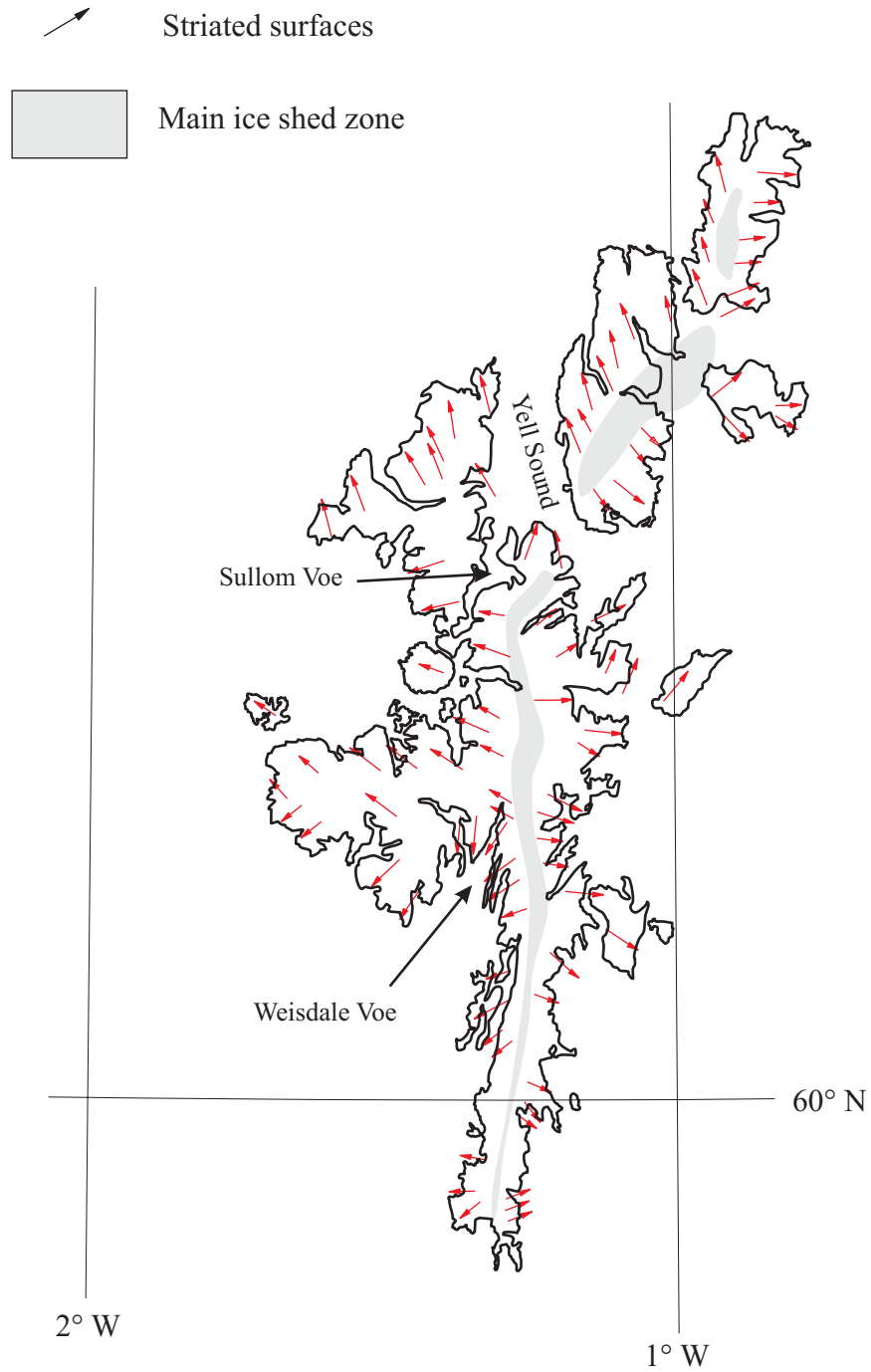


Figure 3.3. Generalised ice flow directions from striated surfaces exposed on land. Redrawn from Flinn, 1983.

3.1.2 West Shetland Shelf

On the inner shelf around Shetland several platforms have been recognised (Flinn, 1964). Most of these surfaces are poorly developed, with the exception of a surface at 82m below sea level, but are similar to more prominent features identified around the coasts of Sula Sgeir and St. Kilda further south, on the Hebrides Shelf (Flinn, 1977; Sutherland, 1984). The origin of the platforms around Shetland is uncertain. Whilst these surfaces are undoubtedly erosional, Flinn (1976) states that they are probably not the result of Late Devensian post-glacial sea-level rise forming wave-cut benches; instead he suggests that the features are remnants of pre-glacial erosion surfaces. This is largely based upon the idea of post-glacial sea-levels rising continuously rather than in phases.

The most prominent basin around the coastline is St. Magnus Bay on the west side of Shetland (basin 5 in Figure 3.1). This basin is c.160m deep, c. 20 km wide and is bounded by cliffs rising to over 100m. The general overdeepened morphology is typical of a classical glaciated fjord, with a raised lip on the western (seaward) side at a depth of c. 90m below sea level, although the basin is more bowl shaped than in most examples. In the cliffs surrounding St. Magnus Bay there are small valley features infilled by glacial sediments of presumed Late Devensian age, overlying earlier peat beds of Ipswichian age (Gordon and Sutherland, 1993). Beyond the mouth of St. Magnus Bay, the seafloor opens into a broad 5km wide shallow channel which feeds into a small basin on the mid-shelf at a depth of 150m (basin 3 in Figure 3.1). This basin has several outflows, two opening south westwards into the Papa Basin (see below), and one opening to the north-west forming a shallow channel to the shelf edge (Figures 3.1 and 3.4).

Whilst the St. Magnus Bay Basin has certainly been glaciated, there is a possibility it may not be entirely of glacial origin. Flinn (1970) suggested that the original structure was formed by a meteorite impact and the deep water basin is the resulting crater. Supporting evidence for this is scarce. The bedrock geology reveals an isolated outlier of Permo-Triassic sediments infilling a pre-existing basin beneath the Quaternary cover (Chesher, 1984). This may indicate a structural control on the

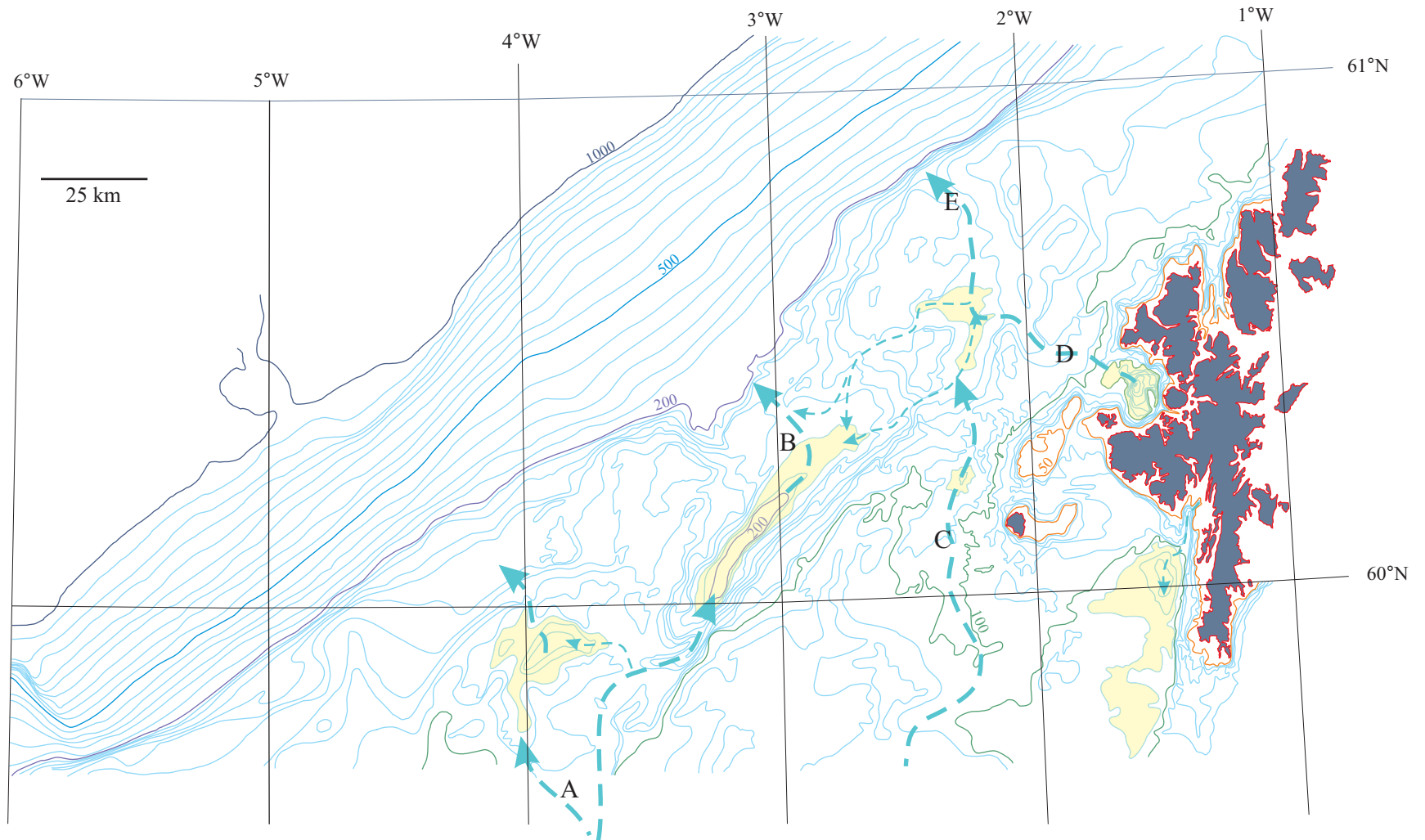


Figure 3.4. Possible sediment transport pathways and topographic lows linking the intra-shelf basins. The routes form the most direct pathways which cold, dense water, or possibly ice may have taken to the shelf edge. The thickness of the lines indicates the inferred relative importance. Routes are identified in the text by the letters.

basin, which modified a more extensive and continuous cover of Permo-Triassic sediments (Andrews *et al*, 1990; Johnson *et al*, 1993).

South west of St. Magnus Bay lies an area of shallow water around Papa Stour and the Foula Ridge. This forms the eastern margin of a broad shallow channel up to 20 km wide and 50m deep which lies to the west of Foula and referred to as the Foula Channel (Figure 3.1). This channel is orientated roughly north-south and joins with the outflow from St Magnus Bay further north before turning north-west towards the shelf edge (Figure 3.4). A much larger and deeper elongate basin lies to the west of the Foula Channel. This is the Papa basin (basin 2 in Figure 3.1) is developed along the line of the Shetland Spine Fault (Stoker *et al*, 1993) and forms a basin up to 230m deep and nearly 70 km long. In cross profile, the basin is steeper on the south-eastern side than the north-western side, indicating a half graben-like structure. Moreover, the fault juxtaposes hard crystalline basement (to the east) and softer Cenozoic sediments (Figure 3.2). Thus, the location and orientation of the Papa Basin is of structural origin, controlled by the bedrock geology and subsequently modified by glacial activity. It is described as glacially over-deepened by Stoker *et al* (1993). To the north-west of the Papa Basin is a topographic low point which forms an outflow to the shelf edge and into the Foula Bight embayment (Figures 3.1 and 3.4). South west of the Papa Basin lies a further small basin which lacks fault control and is probably of erosional origin (basin 1 in Figure 3.1). This is also described as glacially over-deepened by Stoker *et al* (1993), reaching a depth 150m. This basin lies at the northward end of an embayment in the 100m isobath which protrudes southwards towards Orkney, forming a broad channel of deeper water (Figure 3.4).

The bathymetry of the outer shelf, out to the shelf break at approximately 200m is much less complex than the inner and mid-shelf areas. Broadly the shelf edge forms a linear, northeast - southwest feature with an embayment forming the Foula Bight (Figure 3.1). The Foula Bight lies approximately 75km west of St. Magnus Bay and forms a 45km-wide embayment that forms a re-entrant that “cuts” into the outer shelf. At its deepest point, it forms a link across the shelf at a depth of 170 – 180m with the Papa Basin (Pathway B, Figure 3.4). This link is a continuation of the

pathway (pathway A Figure 3.4) from the mid-shelf in the south and has been proposed as a glacially over-deepened route (pathways A and B) from the Orkney region (Stoker *et al*, 1993). At the shelf edge, the Foula Bight extends the 200m isobath shelfwards by up to 8km and is constrained to the northeast and southwest by topographic highs at a depth of 140m. The formation of the Foula Bight may, in part, be controlled by the structural nature of the margin related to late Cretaceous – early Palaeogene fault movements that may have segmented the Rona Ridge, a buried Precambrian basement high (Chesher *et al*, 1988).

Beyond the 200m isobath, the survey interval changes to 50m. This greatly affects the resolution of features picked out by the bathymetry. The overall consequence of the larger interval survey is to have a smoothing effect on the morphology of the continental slope. As a result, the detail provided by the seabed image is more useful in describing the deeper water features of the study area.

3.2 Physiography derived from the 3D seabed image

The area covered by the sea-bed image (Figure 3.5) can be broadly split into three main zones: shelf, slope and lower slope/basin floor. Data relating to the shelf zone is limited and does not extend landwards of the outer shelf. In some of the outer shelf areas the data is of poorer quality and may be unreliable. This may be due to errors in the identification of the sea-bed reflector during processing.

In the north-east of the shelf zone around 60°45'N 2°20' W there is a series of arcuate ridges up to 1 km wide and over 60 km long (feature 1a in Figure 3.5) which lie broadly parallel to the bathymetric contours (Figure 3.1). The ridges have previously been interpreted as terminal moraines marking the maximum limit of several ice sheet advances across the shelf (Bulat and Long, 2001). Beyond the outermost ridge, and between the outermost and middle ridges, are a group of smaller (approx 2km x 7km) fan-shaped lobes (1b in Figure 3.5). A recent interpretation suggests that these are glacial outwash features deposited in a sub aerial environment at a time of lowstand sea level which exposed the outer shelf (Leslie *et al*, 2003). The outermost ridge is cut at its most north-westerly point by the shelf break (1c in Figure

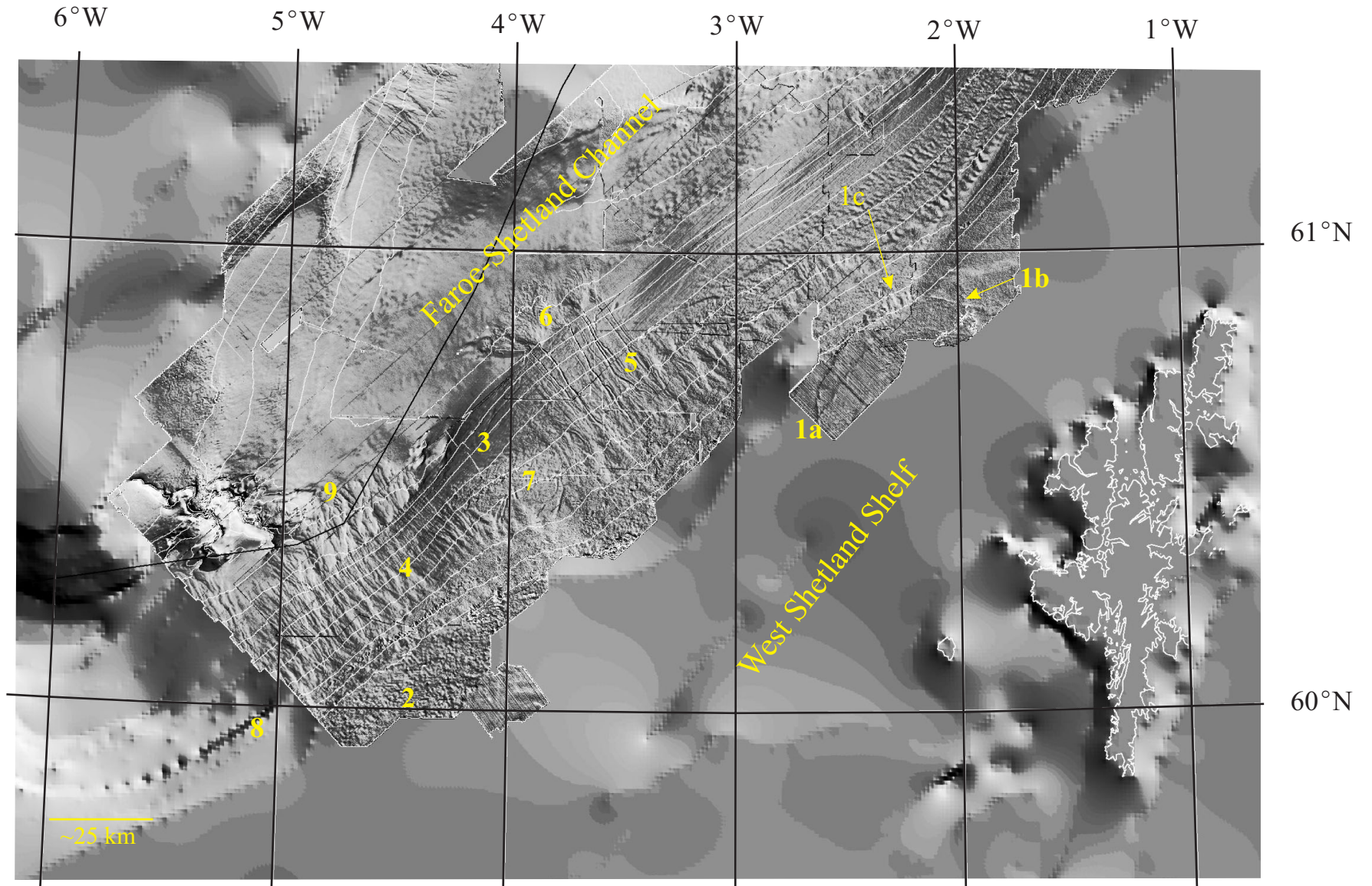


Figure 3.5. Seabed image derived from 3D seismic data of the West Shetland Margin. Numbers indicate features referred to in the text.

3.5). At this point the ridge becomes a series of fan-shaped lobes, interpreted as debris flows (Bulat & Long, 2001), which extend approximately 4 km down the upper slope.

The shelf in the area of 60°N 4°20'W shows an area of apparently hummocky topography (feature 2 in Figure 3.5). Interpretations of this area have suggested that the hummocky effect may be the result of intense iceberg scouring, although a data artefact cannot be discounted (Bulat & Long, 2001). However, comparison with side scan data (Figure 3.6) shows that the scale of ice-berg plough marks on the shelf is much smaller than the features present on the seabed image. This area is described more fully in section 4.3.

Within the area between the shelf break and the base of the slope there are two broad types of topography: i) relatively smooth seafloor with few large scale features (feature 3 in Figure 3.5), and ii) large elongate lobes lying roughly parallel to the dip of the slope (feature 4 in Figure 3.5). The lobes cover much of the upper slope and some parts of the mid and lower slope. These have all been interpreted as various styles of debris flow deposit (Bulat & Long, 2001; Davison & Stoker, 2002; Leslie et al, 2003). Lobes at the top of the slope are generally broader and less clearly defined than those lower on the slope. In the south west of the West Shetland slope, the debris flow deposits on the slope appear to show signs of channelisation, spreading out at the base of the slope to form a series of coalesced fan-like lobes. These debris flows are the only significant group to reach the base of the slope and appear to mark a down-slope focus for deposition of sediments delivered from the shelf. There is evidence of late stage modification on some of the debris lobes, which show small incised and leveed channels. These appear to feed smaller lobes which are superimposed on the larger features and may be the result of a different depositional process such as small-scale turbidity or density-driven currents.

On the upper to middle slope between 3°W and 4°W, a further series of lobes are present (feature 5 in Figure 3.5). Unlike the group to the south west, however, these have been extensively modified by the development of a series of slope-parallel

gullies which begin at a water depth of around 400m. These features have been previously described from side scan surveys (Kenyon, 1987). The gullies feed into a group of fans at the base of the slope which closely resemble the shape of classic alluvial fans from sub aerial environments (feature 6 in Figure 3.5). The morphology of these fans is very different to that of the lobes to the southwest in that the distal margins are poorly defined and the surfaces more irregular. Individual fans emanate from a point source where the feeder channel reaches the slope base. Collectively, the fans overlap and some of the feeder channels appear to be inactive, suggesting a periodic shifting of the source on the upper slope. The lower slope in this region, and to the north-east, appears relatively smooth in the areas not affected by the fans, indicative of predominantly basinal depositional processes relatively unaffected by downslope sedimentation (Leslie and Stoker, 2003).

Modification of the upper slope by iceberg turbation has been suggested as a reason for the poor definition of features in this zone (Leslie et al, 2003). This is supported by the presence of strike-parallel furrows near the shelf break which are interpreted as iceberg ploughmarks (e.g. Belderson et al, 1973). Several much larger furrows at approximately 400-500m water depth (feature 7, Figure 3.5) have also been interpreted as iceberg ploughmarks (Holmes et al, 2003). However, the large furrows lie in a NNE-SSW direction rather than parallel to the strike of the slope. The great difference in water depth between the ends of the features also casts doubt on the ploughmark theory. An iceberg which grounds in deep water is very unlikely to travel 8 km up-slope into water depths 100m less. The apparent depth of incision visible on the image and the definition of the furrow also decreases upslope. These observations suggest a more likely explanation is the movement of a body down the slope, possibly as a cohesive block of sediment. This idea is explored more fully in Section 4.4.1.

On the upper to mid-slope in the southwest of the image there is a linear zone of irregular topography forming a series of hollows and hummocks at a water depth between 300-350m (feature 8, Figure 4.5). The hollows are reportedly up to 50m deep and 1 km wide (Leslie *et al*, 2003), and are interpreted by Holmes *et al* (2003)

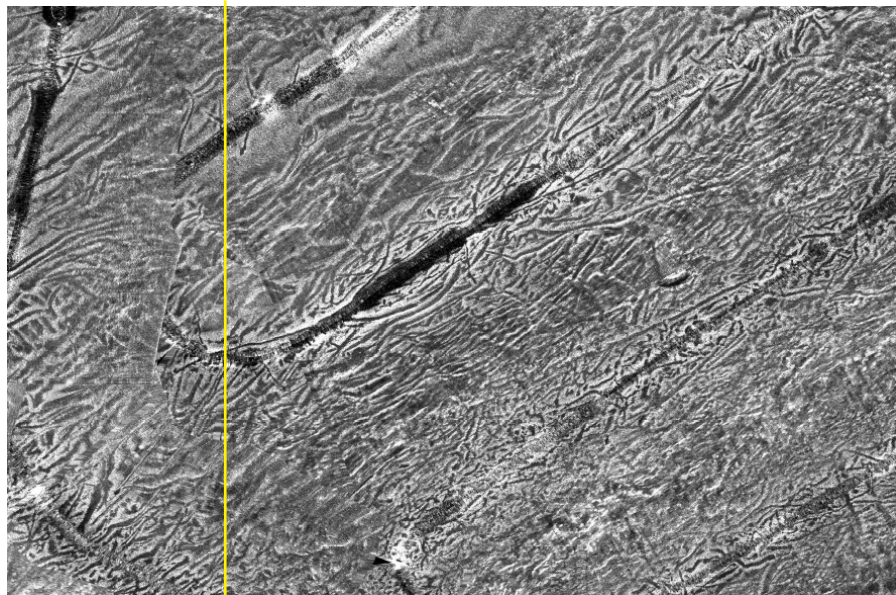
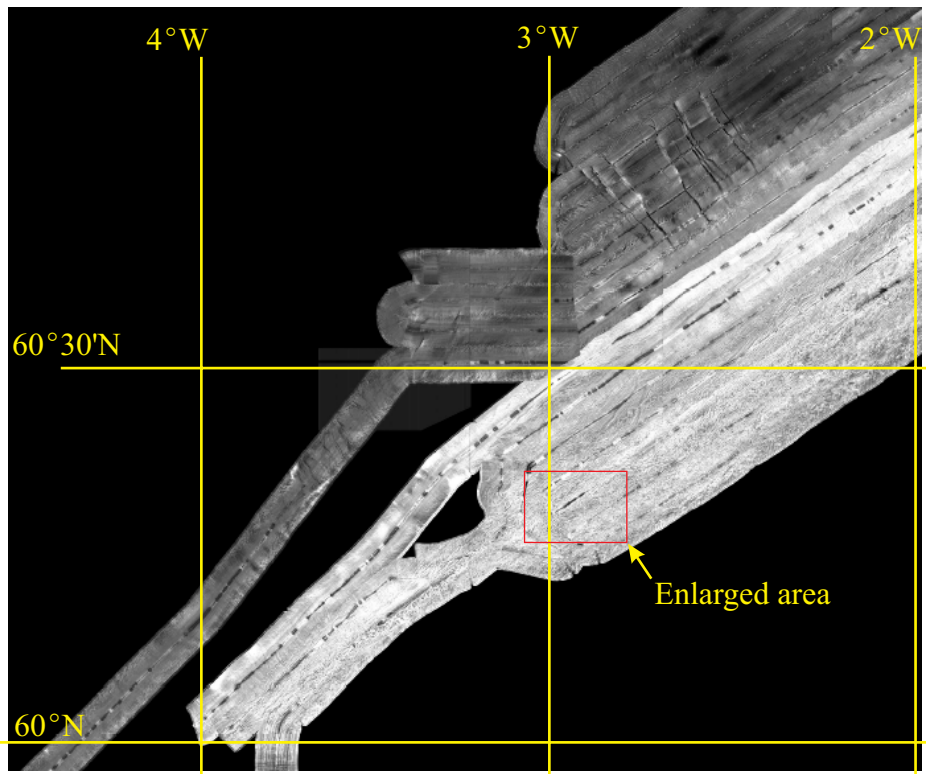


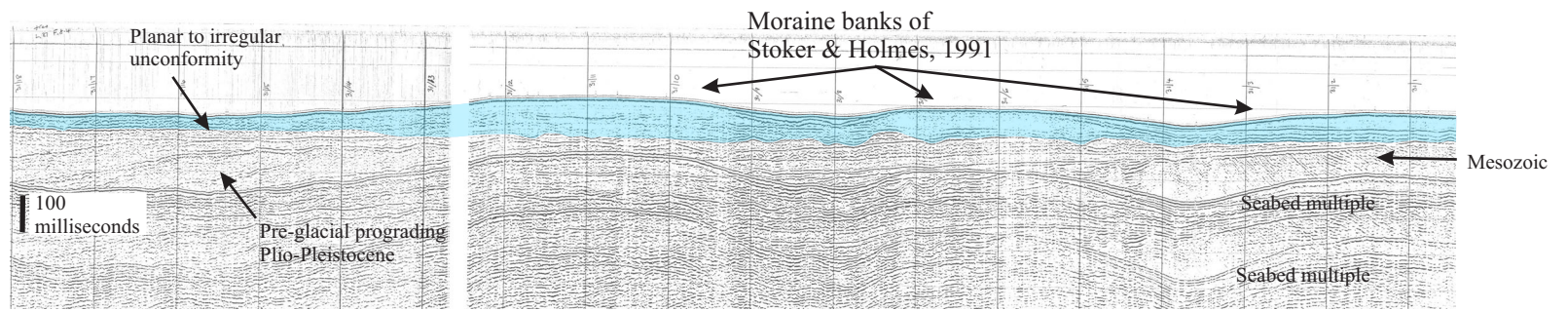
Figure 3.6. Sidescan sonar data from the AFEN survey of the West Shetland margin. The enlarged section shows abundant iceberg plough marks predominantly parallel to the shelf edge, and parallel to the strike direction of the seabed.

as scour holes formed by high-energy meltwater expelled from an ice sheet margin. However, these features are interpreted here as the results of slope instability, for reasons explained in Section 4.4.1 and 4.4.2.

At the narrowest point in the Faroe-Shetland Channel, at the south western end of the image, is an area of variable topography known as the Judd Deep (feature 9, Figure 3.5). On the image this zone appears as a series of irregular terraced surfaces, bounded on their SW edge by areas in total shadow, indicating very steep scarp-like slopes. The slopes at the edge of these terraces are up to 200m high and reach angles of 40° (Leslie et al, 2003). The Judd Deep, and the associated scarps, the “Judd Falls”, form the highest point in the floor of the Faroe-Shetland Channel. A cold, southward-flowing bottom current, the Norwegian Sea Deep Water, flows down the Faroe-Shetland Channel into the Faroe Bank Channel, where it becomes a component of the North Atlantic Deep Water. Considerable bottom current activity has prevailed in the region of the Judd Deep since the early Neogene formation of these deeps, attributed to the erosive effects of vigorous bottom currents during the early to mid-Miocene (Stoker *et al*, 2003).

3.3 Physiography derived from side-scan sonar images

The side-scan sonar data (Figure 3.6) from the AFEN surveys provides some additional detailed information about the seabed which is not visible on the WFA seabed image. It also provides data which is below the resolution of the bathymetric surveys. The most obvious feature of these data is the abundance of small trough features on the outer shelf and upper slope. These features are interpreted as iceberg ploughmarks from grounding icebergs and have been described by various authors (e.g. Belderson et al, 1973; AFEN, 2000). The distribution of large ploughmarks is strongly related to water depth, with few ploughmarks below the 400m isobath or in the shallow areas of the shelf. The area affected on the shelf edge extends into the Foula Bight up to a depth of 170m. Collectively, the ploughmarks show a strongly preferred orientation parallel to the strike of the slope. The exception to this occurs in the shallow water of the shelf where the features become less organised.



Pleistocene glacial section

5 Km

Figure 3.7a. Line 83/04-31, mid-shelf. Vertical scale is two-way travel time

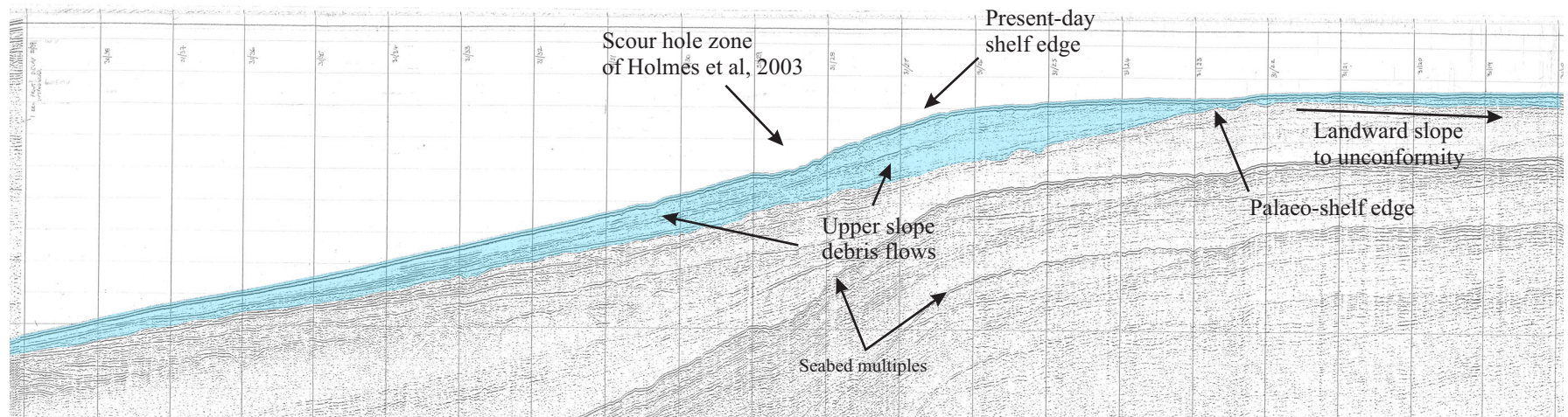
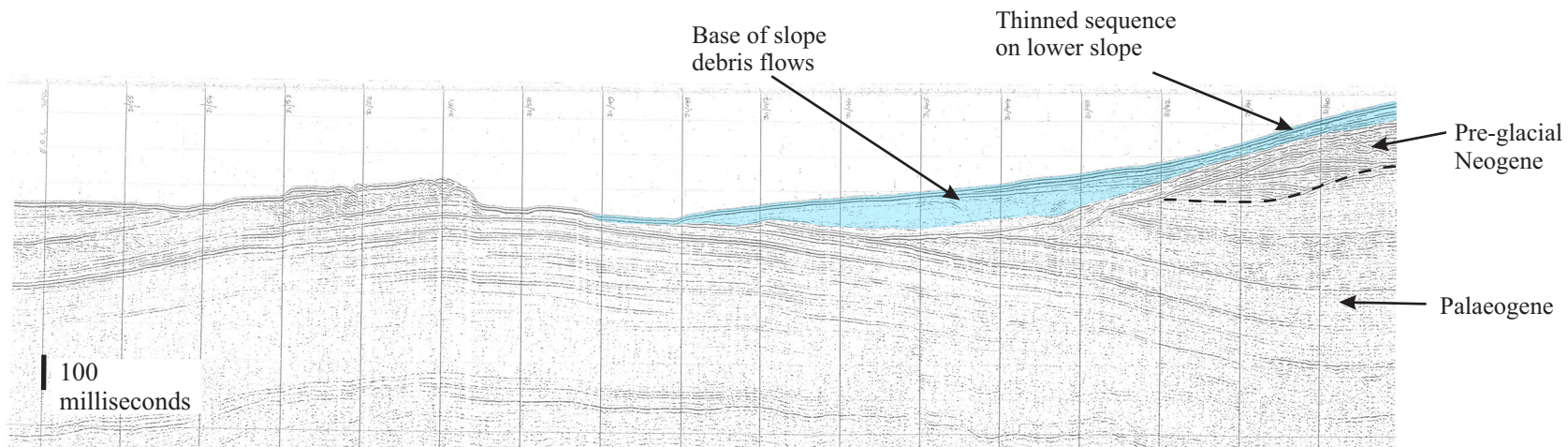
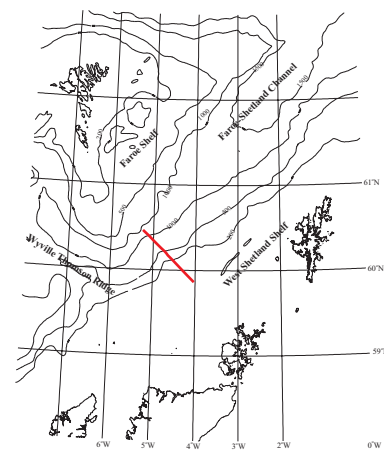


Figure 3.7b. Line 83/04-31, outer shelf and slope.



5 km

 Pleistocene glacial section



Location of line 83/04-31

Figure 3.7c. Line 83/04-31, lower slope and basin floor

3.4 Glacial architecture derived from BGS seismic reflection data

The majority of the BGS seismic data used in this study is from sparker or airgun surveys. Generally the sparker surveys have proved more useful in characterising the glacial section, especially in terms of the internal structure of units. However, the airgun surveys are better for defining the larger-scale architecture of the glacial section and for filling data gaps where sparker surveys were not run. Airgun profile 83/04-31 (Figure 3.7) provides a representative view of the margin and forms part of Transect A (Figure 2.1). An interpreted version of this line establishing the main Mid-Late Cenozoic stratigraphy has previously been published by Stoker (1999); this description focuses on the Quaternary (mid-late Pleistocene) glacial section, highlighted in Figure 3.7, and provides an indication of the underlying control on seabed physiography.

3.4.1 Geometry of the glacial section

The base of the Quaternary glacial section on the outer shelf is characterised by an irregular basal unconformity – the glacial unconformity – which truncates the underlying seismic units. The unconformity shows different degrees of surface roughness where it overlies different seismic units (Figure 3.7a). However, there appears to be no relationship between the degree of roughness and a particular underlying unit, suggesting that underlying lithology is not the main control on topography of the unconformity. Towards the shelf edge, the unconformity becomes generally smoother and acquires a slight landward tilt. At the palaeo-shelf edge there is a zone of increased topography showing a series of small undulations (Figure 3.7b). Beyond the palaeo-shelf edge, the unconformity is generally smooth with little topography as far as the slope base, although truncation of the underlying reflectors is still evident. On the floor of the Faroe-Shetland Channel, the seabed represents a composite erosion surface that has developed since the early Neogene, when bottom currents eroded into Palaeogene strata (Figure 3.7c). These remain exposed to the present day through on-going current activity.

The seismic reflection data adds further information to the inner and mid shelf zones. One of the most significant features imaged on the profile is a series of large

undulating ridges in the area 60°N 4°W (Figure 3.7a). These have previously been described by Stoker and Holmes (1991) as large end-moraines formed during pauses in the retreat of the Late Devensian ice sheet. On the airgun profile these features form seabed ridges up to 35m high and 7 km wide. Internally they form acoustically transparent lensoid bodies with some chaotic reflections. The sparker data for the same survey line reveals more detail, showing some irregular surfaces within the moraines and high continuity parallel stratified reflections between them. The ridges rest upon an earlier unconformity, beneath which a series of draping sub-parallel reflections overlie the basal glacial unconformity. Thus, the profiles clearly show two major unconformity surfaces, allowing this aggrading glacial section to be divided into two separate glacial units. The uppermost of these was identified by Stoker and Holmes (1991) as the Otter Bank sequence which is the main focus of this study, and is investigated in detail in Chapter 4.

Beyond the pre-glacial shelf edge the glacial section thickens greatly, reaching a combined thickness in excess of 180 ms, about 150m, based upon an acoustic velocity of about 1.7 km/s which is derived from matching borehole data with seismic profiles. The airgun profile shows the whole glacial section to be composed of a prograding sediment wedge of oblique tangential form with an irregular, iceberg-scoured seabed. Within the wedge there are a number of smaller wedges which downlap onto the glacial unconformity surface. The major units in the glacial section are defined by prominent high amplitude, continuous reflectors, forming long wedges which thin downslope. In cross-slope (strike-parallel) sections the wedges have a lensoid shape with a chaotic internal structure. The lenses are bounded by thin high amplitude reflections. These form parallel to sub-parallel stratified packages with high continuity. When considered in 3D, the lensoid bodies are seen to be elongate lobes which correlate with those seen on the upper slope on the sea bed image (feature 4, Figure 3.5), and are explored in more detail in Section 4.4).

On the lower slope the glacial section is much thinner with some of the lower units apparently absent or below seismic resolution (Figure 3.7c). The section is represented by high amplitude, continuous reflections with a more stratified nature

than those higher up the slope. Tracing individual reflectors through the lower slope zone is not possible in many of the survey lines, making correlation with the basin floor section problematic.

On line 83/04-31, the lower slope/basin floor is characterised by a series of wedges in dip section, and lenses in strike section, indicating a further group of lobe-shaped bodies. These have a chaotic internal structure and are bounded by high amplitude, high continuity reflections in parallel stratified packages. The lobes are the same features shown at the slope base on the sea-bed image (feature 9, Figure 3.5) and interpreted as debris flows (Bulat & Long, 2001; Davison & Stoker, 2002; Leslie et al, 2003).

3.4.2 Distribution and thickness of the glacial section

By correlating the glacial unconformity depicted in Figure 3.7, with other survey lines throughout the study area, an indication of the thickness and distribution of the glacial sedimentary package was obtained. Mapping of this data has enabled the production of the isopach map in figure 3.8 showing the thickness of the entire glacial section.

The inner shelf shows a very thin cover of sediment, with the area in grey marking zones of rock outcrop at, or very close to the seabed. When compared with the bathymetry, the zones of outcrop and thin sediments correspond closely with the approximate position of the 110m isobath. The area within St. Magnus Bay has proved 80m of glacial sediments in boreholes (see Chapter 5) but is difficult to identify on seismic profiles due to masking by sea-bed multiples and poor resolution. The area of the Papa Basin shows generally thin sediments with a thicker section on the western margin and a tongue of thicker sediments encroaching into the south-western end. The area immediately west of the basin forms a topographic high point with thin glacial sediment cover. North-east and south-west of the high point the glacial sediment cover is thicker and more complex, which is reflected in the seafloor topography (Figure 3.1).

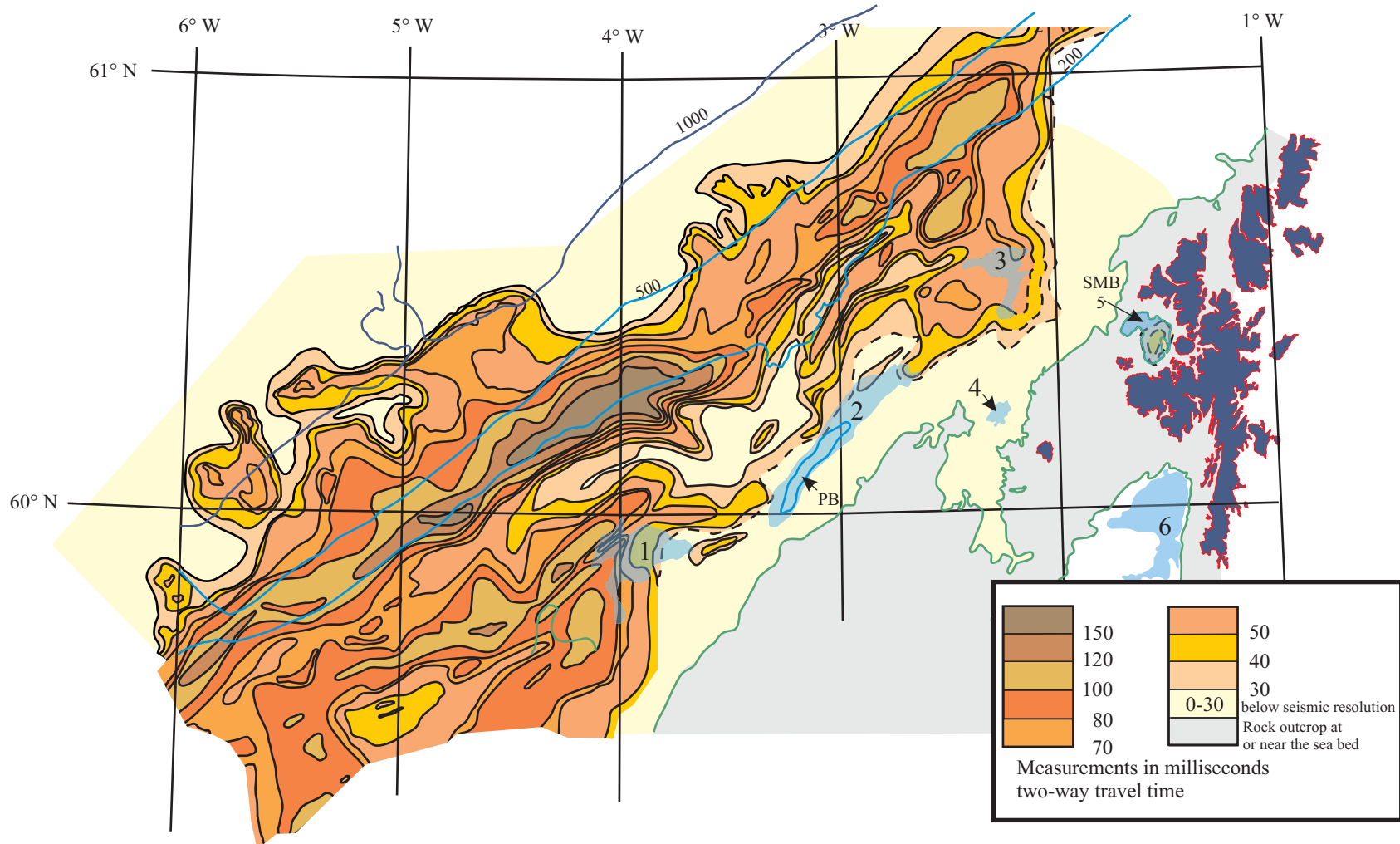


Figure 3.8. Isopach map of the Quaternary glacial section of the West Shetland margin. Enclosed basins 1 to 6 from Figure 3.1 are superimposed together with the 100, 200, 500 and 1000m isobaths. Areas of less than 100m water depth include rock at the sea bed. SMB, St. Magnus Bay; PB, Papa Basin.

Along the present shelf edge and uppermost slope, the distribution of glacial sediments reaches its greatest thickness, with a substantial section of over 150ms and a maximum of 200ms (equivalent to approximately 160-170m using a velocity of 1.7km/s). This represents the thickest development of the glacial prograding sediment wedge (Figure 3.7b). There are two main concentrations of glacial sediments, one in the centre of the study area, and a thinner area in the northeast. On the lower slope and the floor of the Faroe-Shetland Channel the glacial section is greatly reduced, forming a condensed section. The exception to this is the area west of 4° 25' W where there is a major focus of glacial sediment deposition which forms the large debris lobes observed on the sea bed image data (Figures 3.5 and 3.7c). The whole thickened section in this area, from the basin floor to the outer shelf is collectively referred to as the Rona Wedge (Stoker, 1995).

Whilst these observations of sediment thickness are important in showing that some areas of the margin have never accumulated (or at least preserved) significant glacial sediment cover, the composite nature of the section makes meaningful interpretation with regards to specific ice-sheet reconstruction, speculative. The subdivision of the glacial cover into different glacial packages from chronologically separate events is covered in the following chapters, with a focus on the Otter Bank sequence, which is interpreted to represent the Late Devensian glacial episode.

4 West Shetland margin: transect A

This chapter describes six areas, A – F, which relate to the southern transect of the West Shetland Margin shown in Figure 2.1. Collectively, these areas extend from the inner shelf to the basin floor, and show different morphological and sediment characteristics representative of the various parts of the margin. The individual provinces are characterised using a combination of seismic surveys (described using the terms outlined in chapter 2), bathymetric data, sea-bed image data and core data (Figures 2.1 and 2.2). The different lithologies recovered in the cores are divided into 4 broad facies types – sand and gravel, sandy mud, massive diamicton and stratified diamicton – with specific characteristics listed in the core logs. The different facies groups were erected following the logging of all the cores, and are based upon common characteristics from the 8 cores in this Transect A. The cores from Transect B, although not examined in detail, are described using the same facies types. The summary of facies characteristics and their interpretation are shown in table 4.1 below. Further specific details of individual boreholes and their environmental interpretation are given in the text and related figures. For simplicity, stratigraphic notation is restricted to defining the Otter Bank sequence (late Devensian) and the base Quaternary glacial unconformity. The intervening glacial section, where preserved, remains undefined. All of these data are used in the characterisation and interpretation of the continental margin, which is summarised below. All areas (A – F) are described and interpreted separately, although a final summary section provides a collective overview of this part of the West Shetland margin. It should be noted that area F occurs to the NE of, and complements, areas C – E. This essentially provides a second slope-to-basin transect, which is intended to illustrate the variability in slope-apron construction.

Table 4.1. Summary of lithological characteristics and interpretation

Lithofacies	Sedimentary characteristics	Interpretation
Sand and gravel	<p>Unconsolidated sands and gravels. Composition 3 – 60% gravel, 15 – 95% sand, 2 – 25% mud. Sand fraction has grain composition of 60 – 90% quartz, 10 – 40% lithics. Grain morphology A-WR. Up to 50% of quartz grains have Fe oxide coating. Pebbles up to 30mm with exceptional cobbles up to 110mm, generally SR-R morphology. Lithologies of pebbles and lithic grains include red and green sandstones, limestones, gneiss, amphibolite and basalt. Evidence of sedimentary structures limited to infilled burrows and possible stratification at some horizons. Mixed microfossil assemblage from both cold and temperate water temperatures with a wide age range. Macrofossil content varies from almost absent to abundant shell debris with complete valves in life position.</p>	<p>Current-deposited or winnowed sediments. Product of contour or downslope currents in the deep water zones and cross-shelf currents in the shallow water zones. The diversity of lithologies represented and the wide age range suggests a different source area from that of the sediments of other facies or repeated re-working of earlier deposits.</p>
Sandy mud	<p>Sandy mud, grey-green to dark grey-brown, structureless with rare pebbles. 80-95% mud with up to 20% sand. Pebbles reach up to 5mm, forming < 1% of the total volume. Exceptional cobbles of gneiss and sandstone reaching 110mm. Sand grain size varies between very fine and granular, but is dominated by the finer grain sizes. Composition of the sand fraction is typically 80-90% quartz with a SA-SR grain morphology. Iron oxide coatings are present on < 20% of the sand grains. Microfossil assemblage includes generally frequent to abundant cold water foraminifera. Frequent bivalve mollusc fragments and complete valves present in some horizons.</p>	<p>Hemipelagic-glacimarine sediments deposited from suspension at a time of glacial conditions or glacial influence. The low proportion of sand and scarcity of pebbles suggests this material was being deposited in an ice-distal setting. The coarse component of this facies is interpreted as ice-raftered debris from icebergs which may not be of local origin. The fine-grained nature of this facies indicates bottom current activity was minimal at the time of deposition. The relative abundance of micro and macro fauna is compatible with a lower rate of sedimentation in an ice distal setting.</p>
Massive diamicton	<p>Red-brown to grey-brown or dark grey, muddy to intermediate diamicton. Comprises mud 60-75%, sand 25-40%. Clast-rich to clast-poor horizons, pebbles between 1 and 45%. Sand grain size varies from very fine to granular, with a modal grain size of fine to medium. Lithic grains comprise 15-40% of the sand fraction. Quartz grain morphology is A-R, with larger grains being more rounded. Pebbles reach a maximum diameter of 80mm with a mode of c. 10-20mm. Lithologies include gneiss, amphibolite, red sandstones and siltstones, grey quartzite and dolomitic limestone. No internal structure. Shell debris is rare to common, Contains arctic water temperature foraminifera which are rare to abundant. Shelf water depth species tend to dominate.</p>	<p>Debris flow, sub-glacial or proximal water-lain till. Discrimination between these three depositional processes is not possible from core data alone. Samples from debris flows are all red-brown at time of coring, those from other sources are grey-brown when cored but turn red-brown on storage. The presence of a cold water shelf micro-fauna indicates glacial conditions were established at the time of original accumulation, i.e. prior to re-mobilization in the case of debris flows.</p>
Stratified diamicton		

Red-brown, clast-poor muddy to intermediate diamicton. Comprises mud 60-75%, sand 25-40%, pebbles between 1 and 5%. Sand grain size varies from very fine to granular, with a modal grain size of fine to medium. Lithic grains comprise 20-35% of the sand fraction. Quartz grain morphology is A-R, with a modal shape of SA-SR. Lithologies of pebbles and lithic grains are similar to those in the massive diamicton with the exception of an assemblage of black globular to angular grains comprising up to 5% of the sand fraction. These grains may be composed of volcanic glass. Lamination is well developed, varying from 3mm thick in the mid-shelf zone to 10mm in the outer shelf zone. Lamination caused by alternating sandy and muddy layers, with muddy layers generally thinner and paler. Dropstones with distorted lamination are present in some sections. Shell debris and foraminifera are rare to common, with abundance decreasing towards the shelf edge.

Deposition from meltwater plumes by suspension settling. Dropstones and moderate sand content interpreted as IRD. Colour may be oxidised; drilling records suggest original colour was dark grey-brown at time of coring.

Abbreviations for Powers Roundness categories: A, angular; SA, sub-angular; SR, sub-rounded; R, rounded; WR, well-rounded.

4.1 Area A: Inner Shelf

4.1.1 Seabed morphology and seismic data

The area of box A indicated in Figure 4.1 represents the inner shelf zone. This zone lies outside of the area covered by the seabed image, but there is seismic reflection and borehole data. The water depth within area A is generally less than 100m, with the exception of the central section which includes part of the Foula Channel (Figure 3.1) where depths reach 110m. Published data on sea-floor sediments indicates gravel and sandy gravel at the seabed, with substantial areas of rock outcrop (Graham, 1990; Evans, 1984; Chesher, 1984b). Quaternary sediment thickness is indicated as 0 – 10m over most of the area.

Seismic data in this area is a combination of watergun and sparker surveys. Very little sub-surface detail is evident from these predominantly low-resolution survey lines due to the seabed pulse, which obscures the upper 10m – 15m of cover. However, the lines do show contrasts in the nature of the seabed, which displays a combination of irregular rough surfaces and smoother, more undulating surfaces.

Line 79/14-09 (Figure 4.2) shows a relatively smooth seabed in water less than 90m deep. This area shows evidence of large scale convoluted reflections at depth which have been interpreted as Devonian sediments on published marine survey maps (Chesher, 1984a). Any Quaternary sediments which are present are lost in the seabed pulse. At the northern end of this section, the water depth increases to about 105m, forming the southern margin of the Foula Channel.

Other survey lines within this area all show similar features. The seabed pulse obscures any shallow sediments which may be present and subsurface detail is extremely poor. Line 79/14-27 (Figure 4.3) shows convoluted reflections in the eastern part of the section between fixed points 1 and 18. This part of the line corresponds to a zone with a smooth seabed and slightly deeper water which forms part of the Foula Channel. The underlying rocks responsible for the convoluted reflections have been interpreted as Devonian sediments (Chesher, 1984a) and are

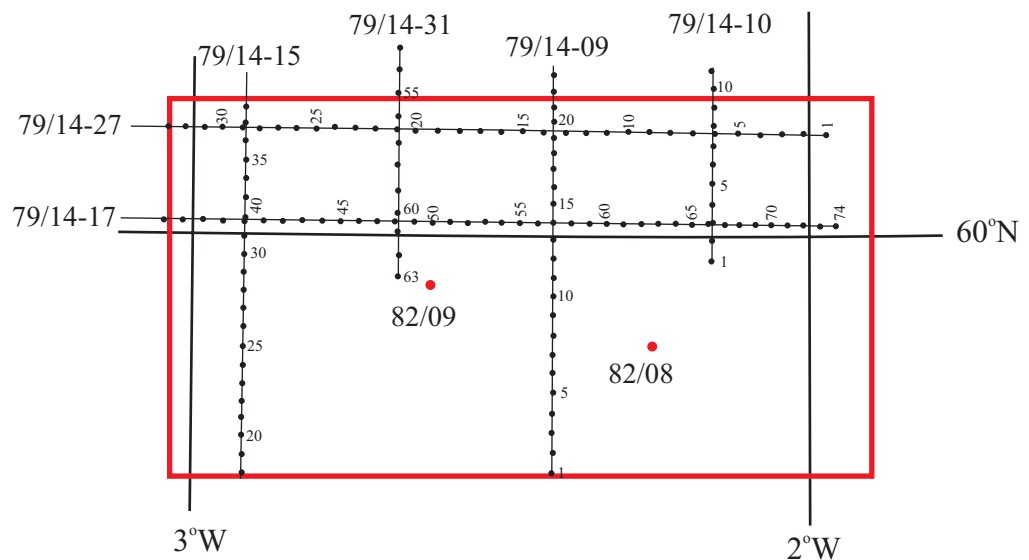
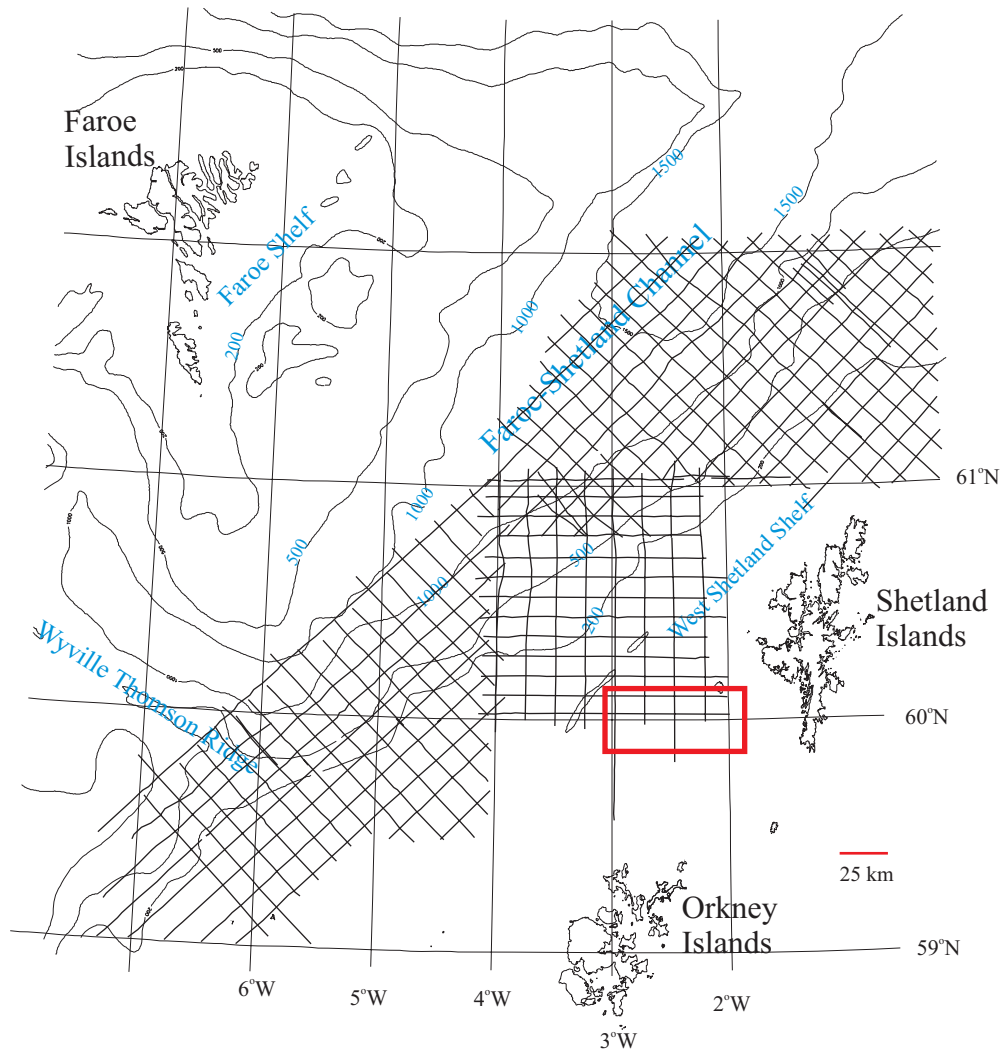


Figure 4.1. Location of Area A, Inner to mid shelf. Enlarged box shows the specific sections of seismic lines and boreholes within the area.

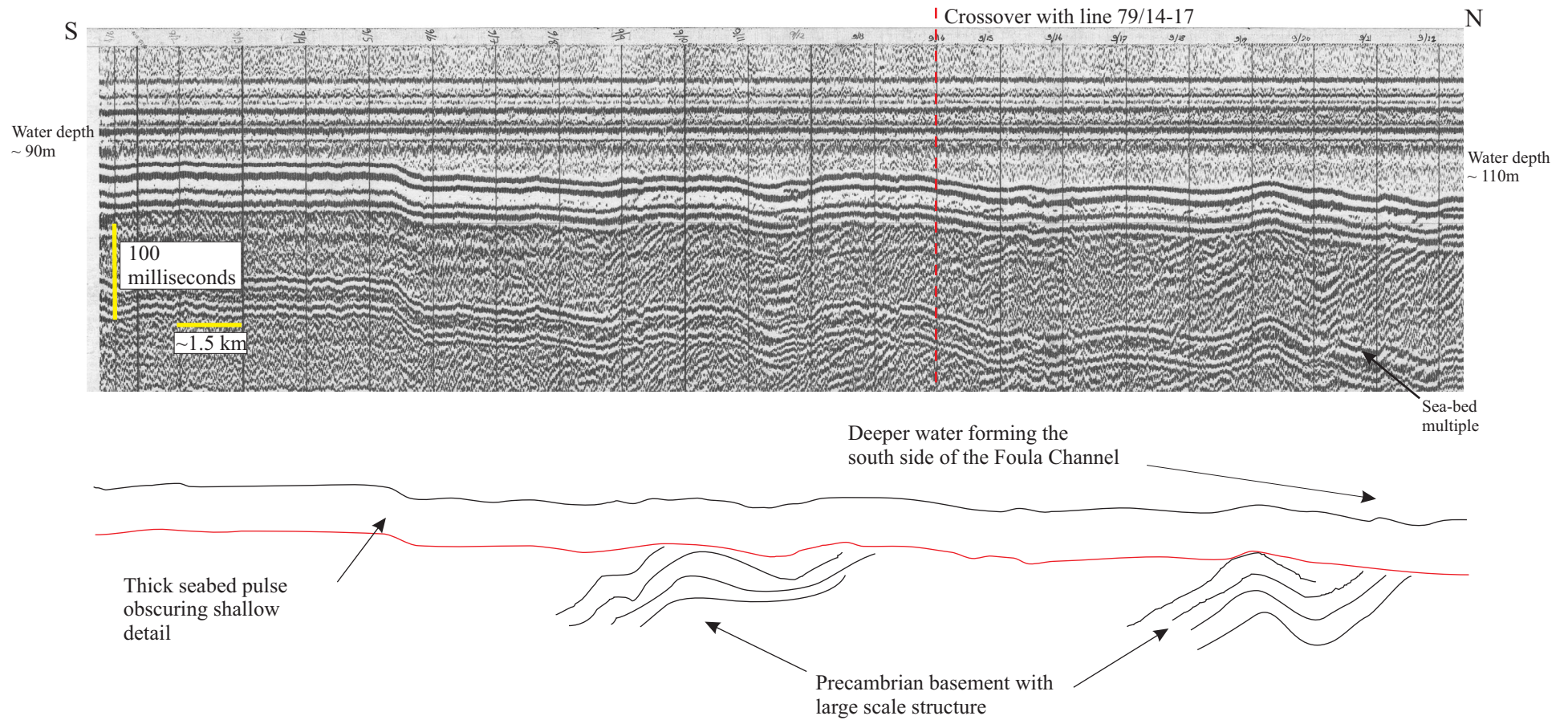


Figure 4.2. Watergun line 79/14-09. The thick seabed seismic pulse obscures any evidence of glacial sediments, but shows that any glacial section that does exist must be less than the thickness of the pulse.

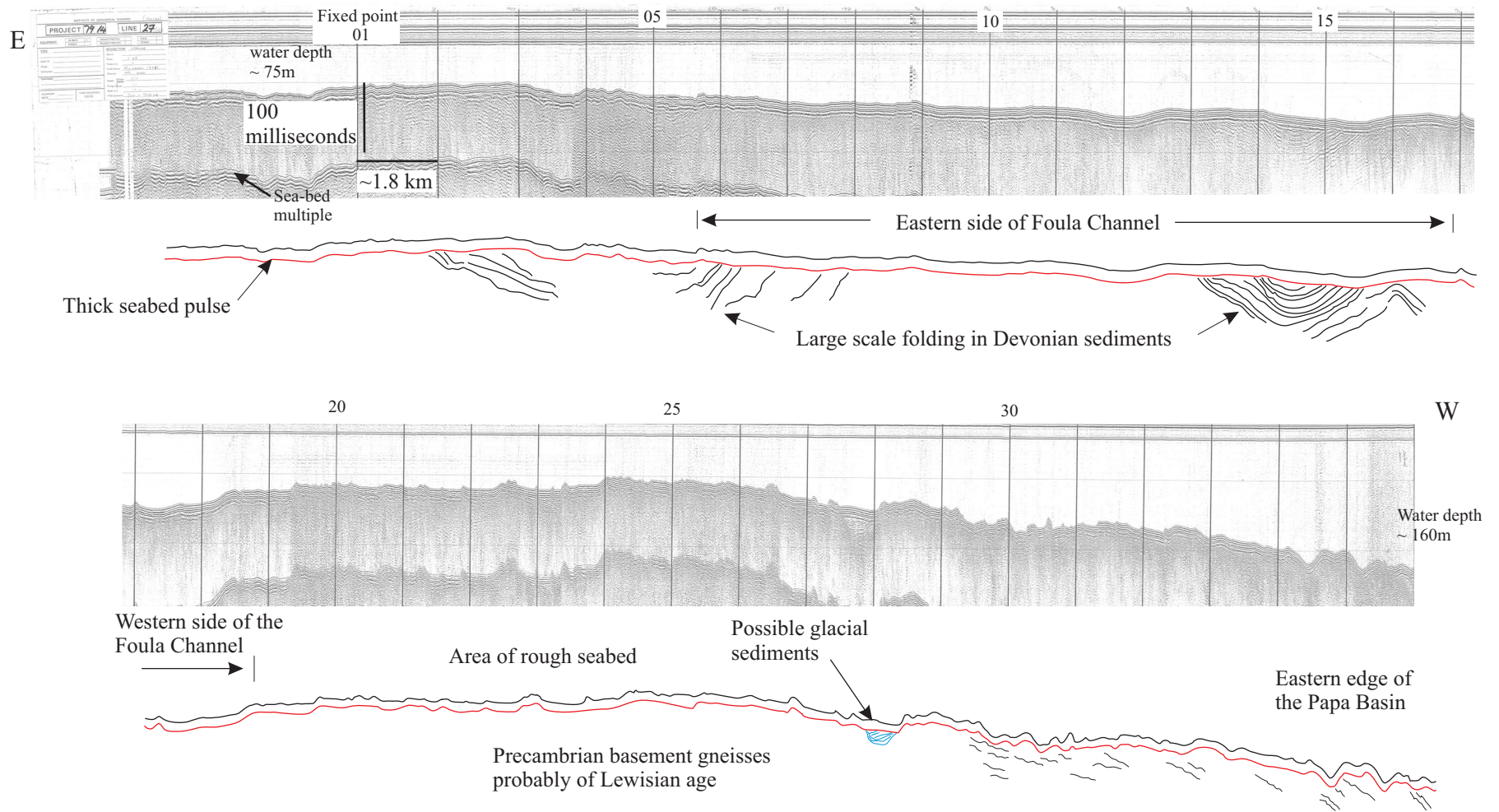


Figure 4.3. Sparker line 79/14-27 showing irregular seabed underlain by Precambrian rocks and smoother seabed underlain by Devonian sediments. The seabed pulse obscures almost all of the shallow surficial sediment detail below the seabed (approx. top 10-15m).

also associated with the smoother seabed on other lines. The zone of rough irregular seabed, with a relief of about 10m, further to the west between points 18 and 27 is underlain by Precambrian basement rock which is probably part of the Lewisian Gneiss. The link between seabed roughness and underlying rock type is a feature common to the whole of area A. There is also a link between water depth and seabed roughness in that the smoother zones of seabed tend to be in shallower water, with the exception of the Foula Channel which is generally developed over the Devonian sediments and has a smoother seafloor.

Survey lines 79/14-17 (Figure 4.4) and 79/14-31 (Figure 4.5) display similar features to the other survey lines but are included as they form the nearest lines to borehole sites 82/09 and 82/08 respectively. The seismic data from these lines suffers from the same problems of lack of detail in the surficial section as the others in this group, preventing meaningful interpretation of the Quaternary section.

4.1.2 Borehole Data

Two boreholes drilled in 1982 lie within area A, but the results from both are very limited with regard to Quaternary interpretation. They do, however, provide important information on the nature of the inner shelf, i.e. a predominantly erosional zone.

Borehole 82/09

Borehole 82/09 was drilled in approximately 90m of water, close to fix 63 on survey line 79/14/31 encountered rock outcrop at the seabed. Consequently no sediments of Quaternary age were recovered.

Borehole 82/08

Borehole 82/08 (Figure 4.6) lies between lines 79/14-09 and 79/14-17. It was drilled in approximately 105m of water and penetrated approximately 8.2m of sediments before reaching rock. Recovery was poor, totaling 8% for the Quaternary section. The sediments comprised three short sections of massive, red-brown sandy diamictons with a polymict clast assemblage, including large pebbles up to 7cm

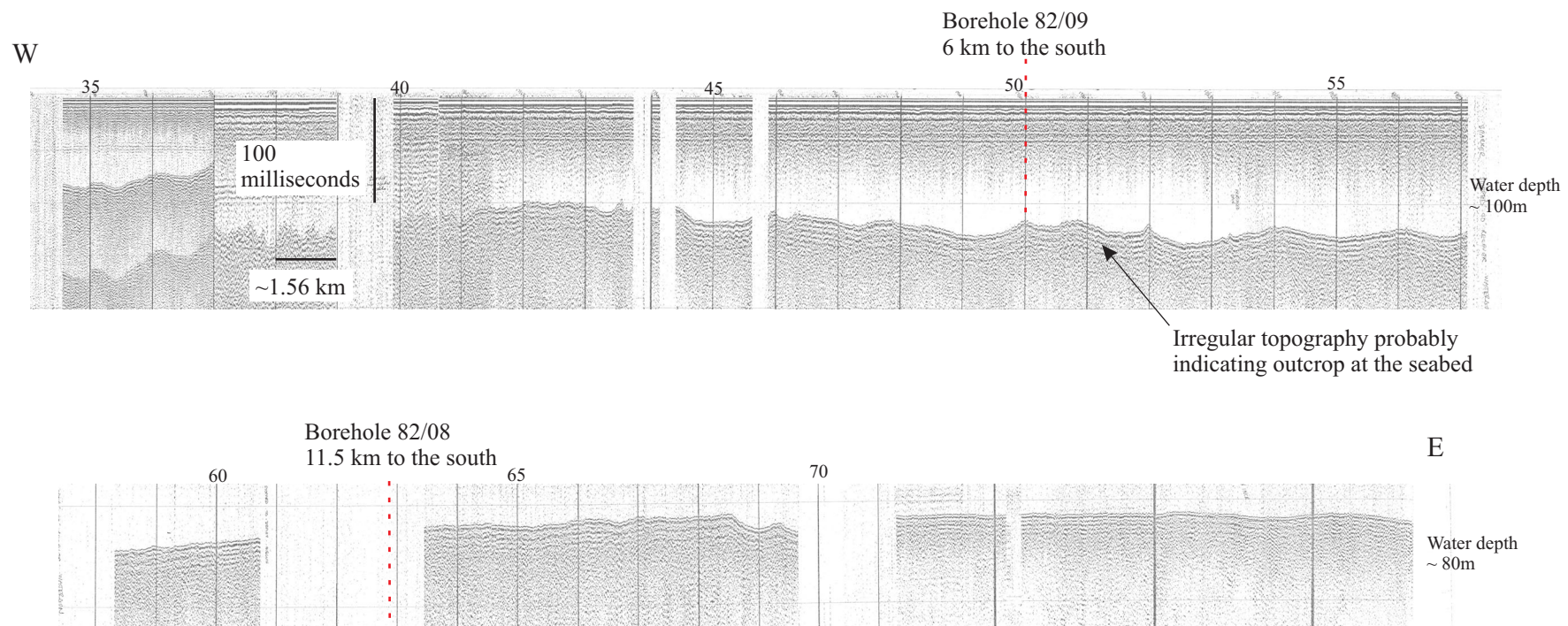


Figure 4.4. Sparker line 79/14-17. The glacial section is too thin to identify on the data. Borehole 82/09 indicates the glacial section is less than 1m thick.

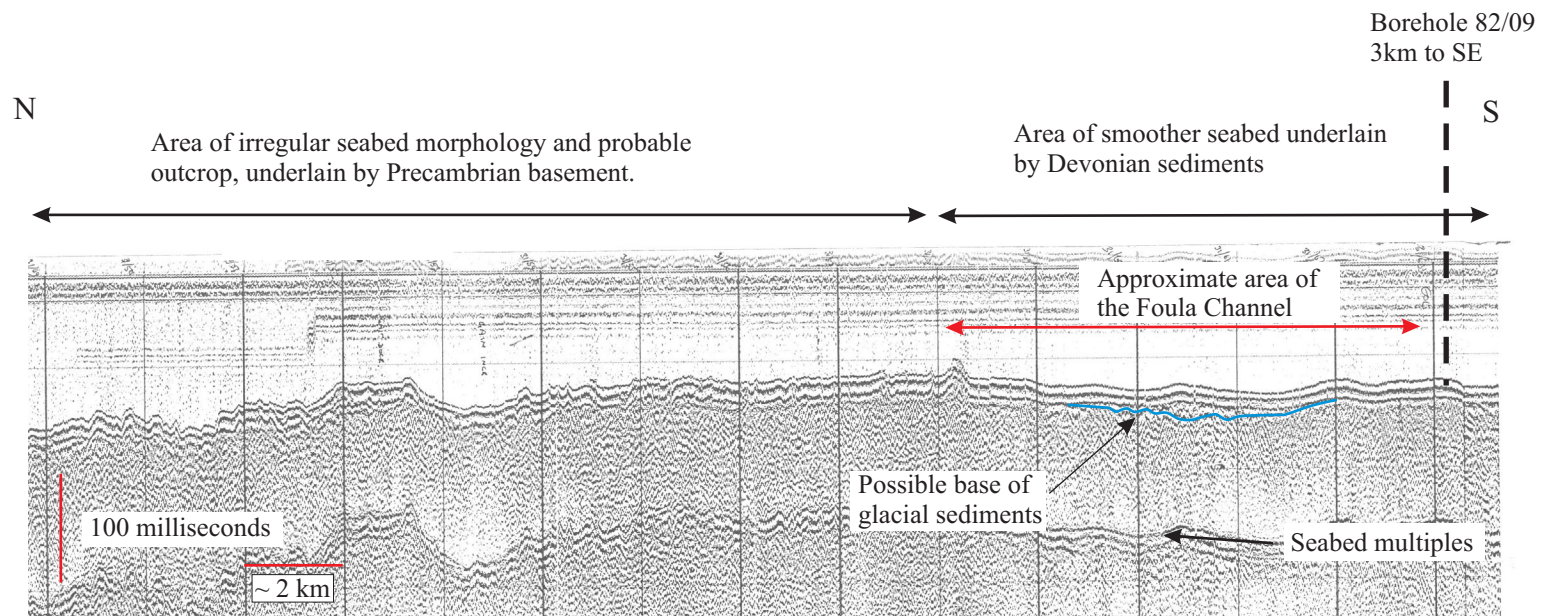


Figure 4.5. Sparker line 79/14-31. Evidence of glacial sediments thicker than 10-15m is limited to a small section on the floor of the Foula Channel. The channel appears to be preferentially developed in the Devonian sediments rather than in the harder Precambrian metamorphic rocks, suggesting this may be an erosive effect. The presence of a glacial infill at the base of the channel suggests ice may have been responsible for this incision.

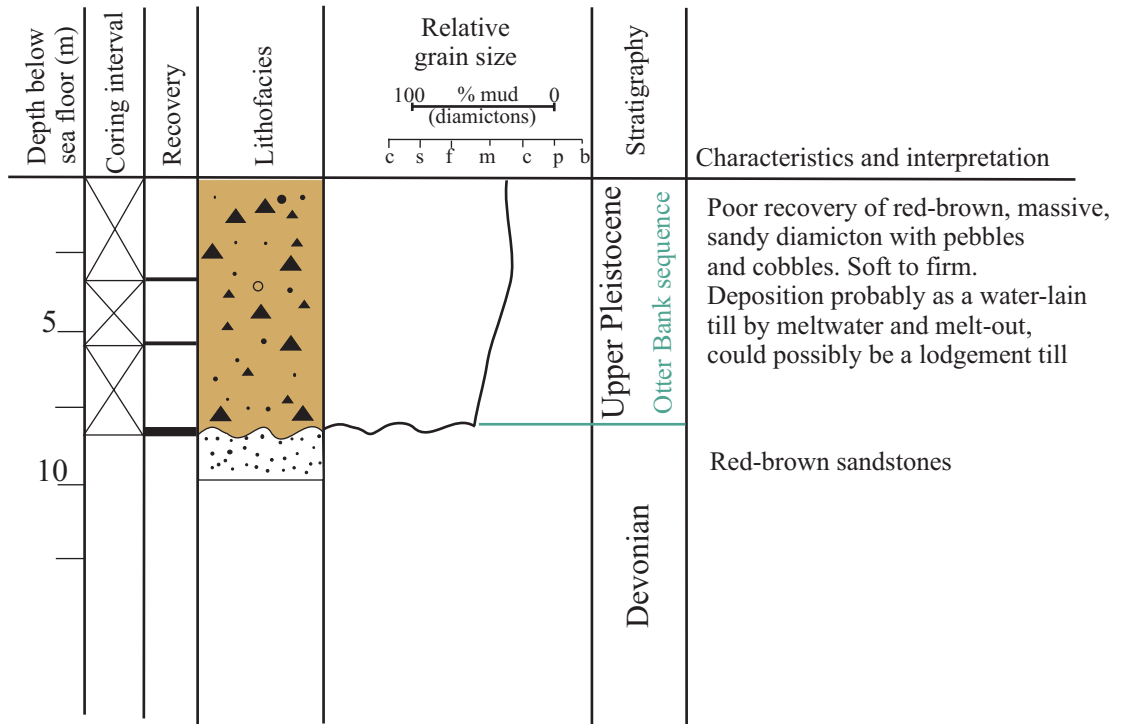


Figure 4.6. Log of borehole core 82/08. The whole diamicton sequence is interpreted as belonging to the Otter Bank sequence

diameter. The sediments were assumed to be of Quaternary age, although this would be difficult to prove due to the poor quality of the recovered samples. Red-brown and grey sandstones of probable Devonian age unconformably underlie these sediments.

4.1.3 Interpretation of area A

Data from area A indicates that the thickness of Quaternary sediments is below seismic resolution across the inner shelf zone, and in some places they are completely absent. The absence of Quaternary sediments may be due to erosion in the Holocene or it may be that the sediments were never deposited, and that this region represents a zone of net erosion. The seabed profiles revealed by the seismic data indicate that where metamorphic basement is exposed, the seabed has a greater surface roughness than in areas where sedimentary rock is exposed. The profiles also show that, in general, the seabed is smoother above water depths of about 100m. The exception to this is in the Foula Channel which lies below this depth and has a smooth seabed. The channel has developed in an area underlain by Devonian mudstones and sandstones suggesting that the underlying rock type has a stronger influence on seabed morphology than water depth. It also suggests that areas immediately underlain by the metamorphic basement have a thinner covering of glacial sediments.

The massive diamictons recovered from borehole 82/08 are indicative of rapid deposition with little or no sorting. The red-brown colour and sand-rich nature of the matrix suggests it is derived locally from the underlying sandstones and mudstones. However, the variation in clast lithologies (Table 4.1) shows that there is also a transported component to the assemblage. Interpretation of the sedimentary process responsible for the deposition of the diamictons in area A is impossible to determine in isolation due to the poor recovery and structureless nature of the cores. Interpretation of the diamicton as a sub-glacial till or a glacialmarine deposit is possible but speculative without additional data.

The evidence from boreholes and seismic data (Figure 4.5) suggests the Quaternary deposits in area A are thicker within sea-bed lows, such as the Foula Channel. This

could indicate that the channel formed a focus for deposition of the diamicton-dominated sediment. The topographic highs appear to be zones of erosion, although removal by subsequent erosion cannot be discounted.

4.2. Area B: Mid-shelf

4.2.1 Sea-bed morphology and seismic data

Area B lies on the mid-shelf in the region of the Papa Basin (Figure 4.7). This area lies outside the area covered by the sea-bed image, but is covered by seismic reflection data and has several boreholes. The area is crossed by the Shetland Spine Fault (Figure 3.2), the Papa Basin being developed in an associated fracture zone along the axis. The basin reaches a water depth of 230m and has an elongate shape with an asymmetric cross profile as noted in Chapter 3.1.2. The north-western margin of the basin forms a likely outflow for the present-day current circulation.

The seismic coverage of area B shows similar sea-floor characteristics as in Area A. On the east side of the Papa Basin the seabed is irregular with numerous small mounds and protrusions. This part of the mid-shelf is underlain by Precambrian basement which tends to be characterised by a structureless acoustic signal such as that from line 79/14-17 in Figure 4.8. Other areas with a smoother seabed are underlain by a combination of sedimentary rocks of widely different ages. Data from line 17 is difficult to interpret due to changes in scale on the record and gaps in the data from equipment failure. However, this line is close to borehole 82/12 which provides a fix on the depth of the basal glacial unconformity, suggesting the floor of the Papa Basin has a covering of glacial sediments, between 15 and 20m thick (Figure 4.8).

Line 79/14-27 in Figure 4.9 shows an extremely good cross profile of the Papa Basin with a steep eastern margin and a gentler slope on the western seaward margin. The eastern side shows the same irregular seabed as Figure 4.8 but is draped by a relatively thin veneer of glacial sediments less than 15m thick close to the margin of the basin. Unfortunately most of the glacial sequence is obscured by the seabed pulse. On the western side of the basin there is a thicker package of glacial sediment up to 30m thick. The seabed in this zone comprises a series of large undulating mounds of presumed glacial origin and form part of the moraine ridges mentioned in Section 3.4.1 and described in detail in the outer shelf section (Chapter 4.3). The

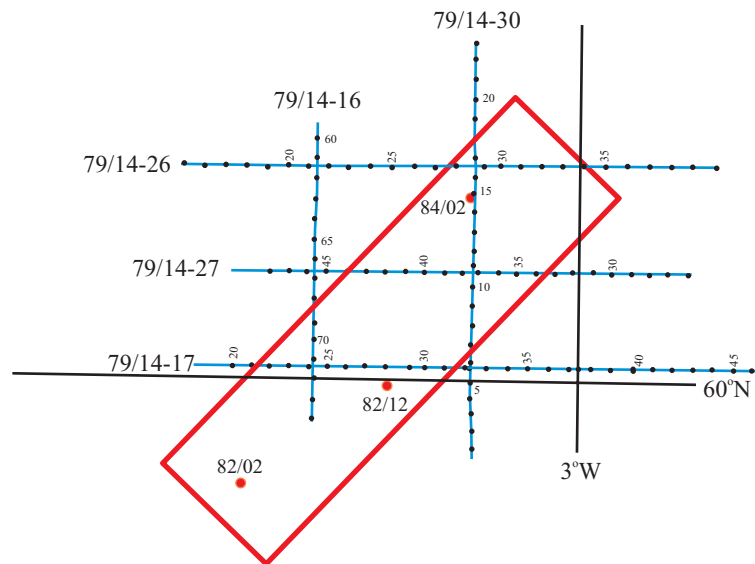
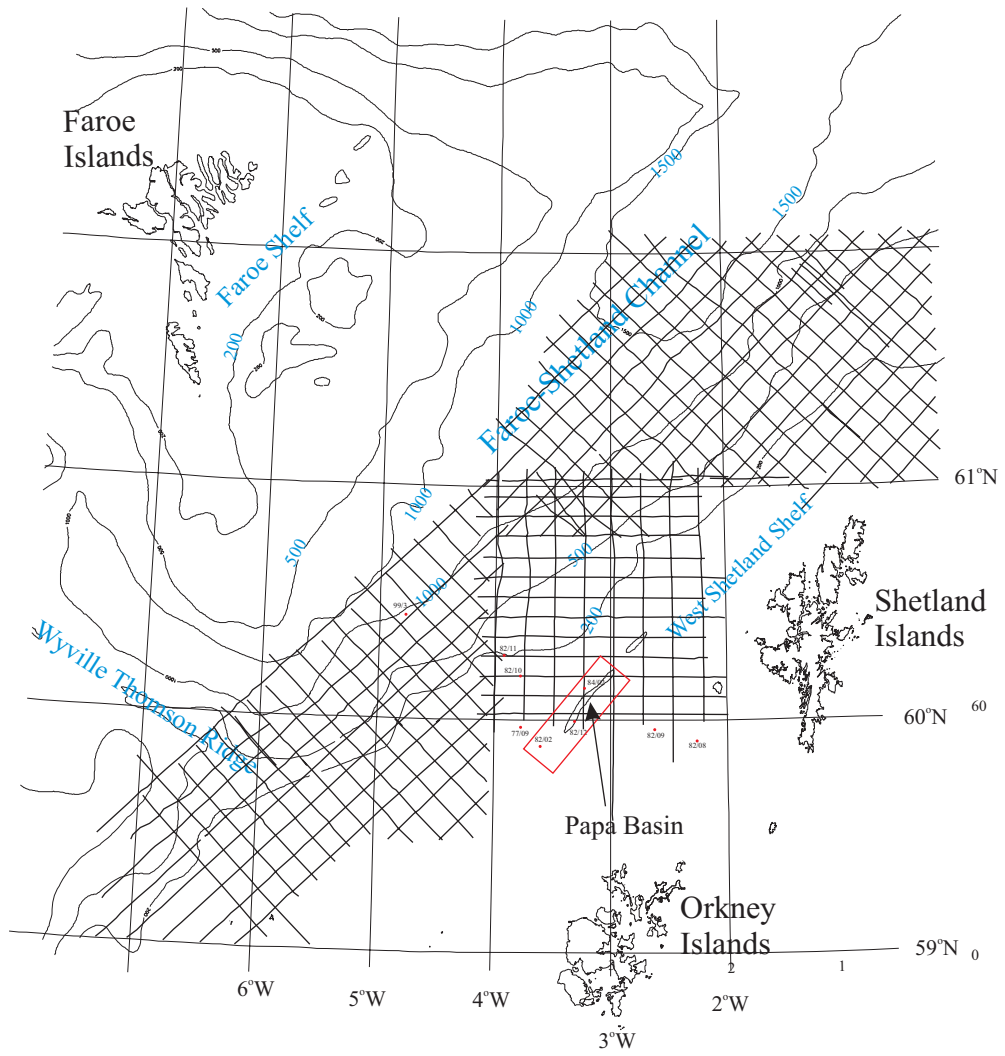


Figure 4.7. Location of area B on the mid-shelf.

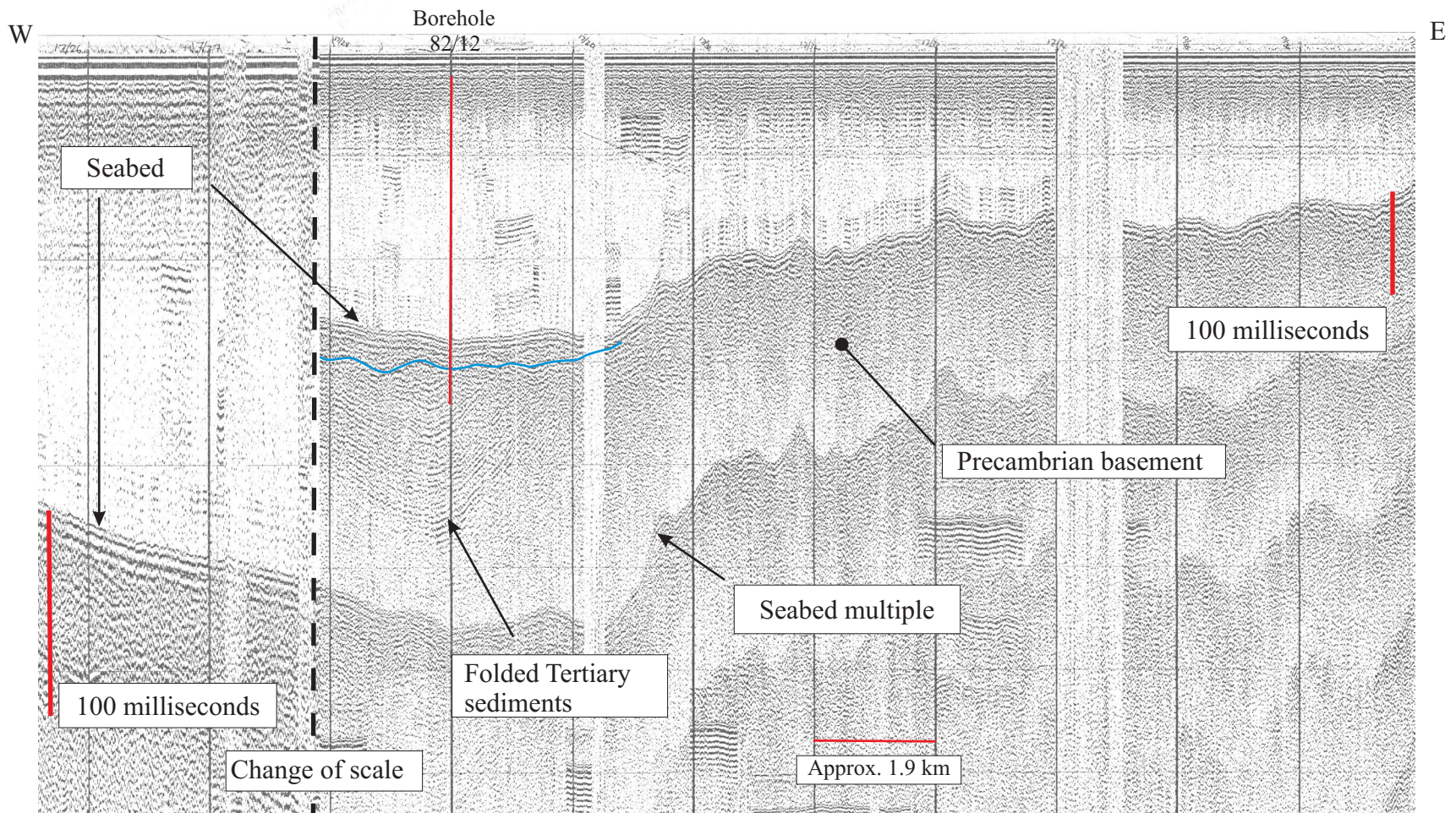


Figure 4.8. Sparker line 79/14-17 showing a profile across the Papa Basin. Tertiary sediments appear as sub-parallel stratified reflections showing large fold structures; an angular, irregular unconformity separates them from the Quaternary glacial strata. The Precambrian basement shows no coherent reflections and has a rough seabed topography.

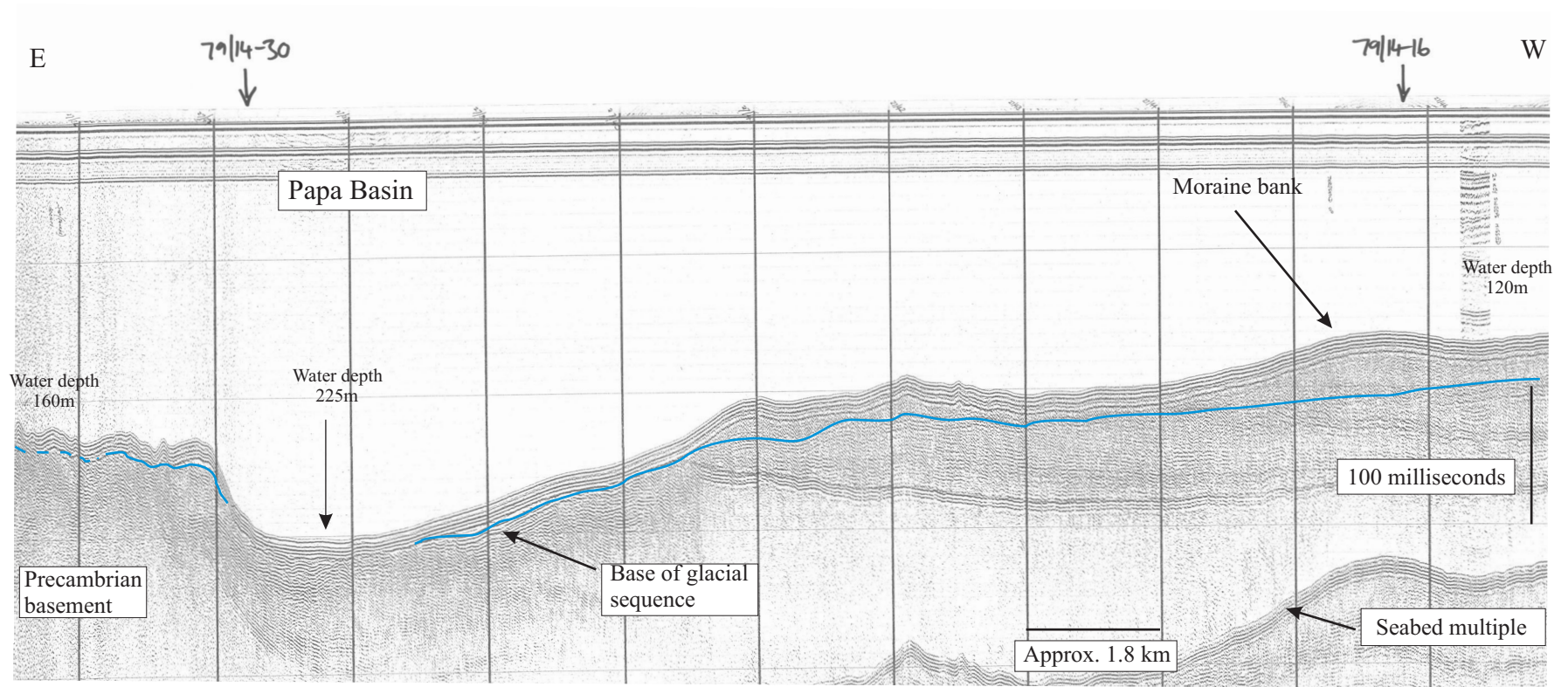


Figure 4.9. Sparker line 79/14-27 showing an east-west profile across the Papa Basin. The eastern margin shows a rough seabed underlain by Precambrian basement gneisses and metasediments and is draped by a thin veneer of glacial sediments. The western margin has a thicker cover of glacial sediments forming a mounded relief up to 30m thick and is underlain by Tertiary sediments. In the west of the section the glacial sequence includes Otter Bank sequence sediments, but the lower boundary is not identifiable on this profile.

glacial sequence appears to lack any internal acoustic structure and as a result, it is not possible to sub-divide the section.

The profile of the Papa Basin on line 79/14-26 (Figure 4.10) is different from that of the adjacent lines in that it shows a steeper western margin and a gentler slope on the east. However, the seabed roughness is still more pronounced on the eastern slope. Interpretation of the seismic data is hampered by a thick seabed pulse, but a high amplitude reflector emerges from beneath the seabed at intervals, indicating a thin cover of Quaternary sediments over the Precambrian basement. On the western side of the basin and shelf a thicker glacial covering forms intermittent mounds which are a continuation of those seen on line 27 (Figure 4.9) indicating that these features probably form continuous ridges across the shelf. The separation between the seabed pulse and the base of the glacial unconformity is insufficient to reveal any internal acoustic structure.

Sparker line 79/14-30 (Figure 4.11) shows a north - south oblique profile across the Papa Basin. Whilst this line does not add much to the overall interpretation of the area, it does provide a reference point for borehole 84/02 which lies at fix point 15. The borehole fix shows that the glacial unconformity lies within the seabed pulse and consequently cannot be resolved. To the north of the borehole site, a continuous reflection interpreted as the glacial unconformity emerges from the seabed pulse showing a glacial section with an undulating base. Again, the separation of the reflectors is not large enough to reveal internal structure.

Seismic data from sparker line 74/4/73 (Figure 4.12) includes the site of borehole 82/02. The data itself is of poor quality, but does show a number of features. The seabed in this region has a relatively smooth, almost planar appearance. The glacial unconformity, fixed from the borehole data, is traceable over a short distance but is difficult to follow.

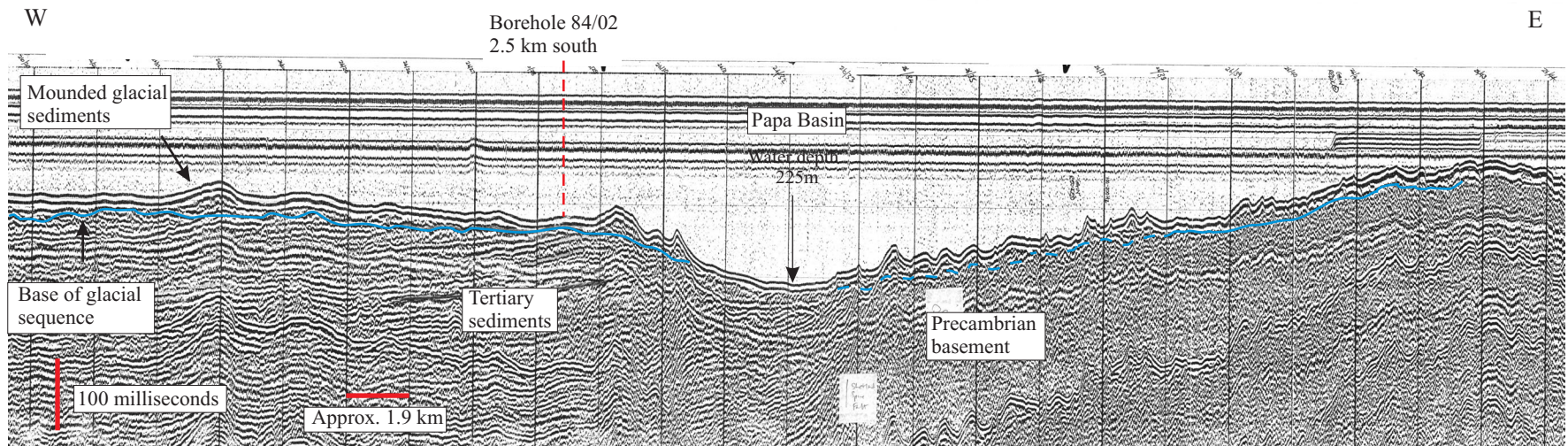


Figure 4.10. Airgun line 79/14-26 showing a cross profile of the Papa Basin and the adjacent shelf. The shelf to the east of the basin has a thin veneer of glacial sediments draping the rough surface of the underlying Precambrian basement. The shelf to the west of the basin has a smoother glacial unconformity and a thicker glacial sequence. The seabed shows mounds composed of glacial sediments which are related to the major moraine fields described in detail in Area C.

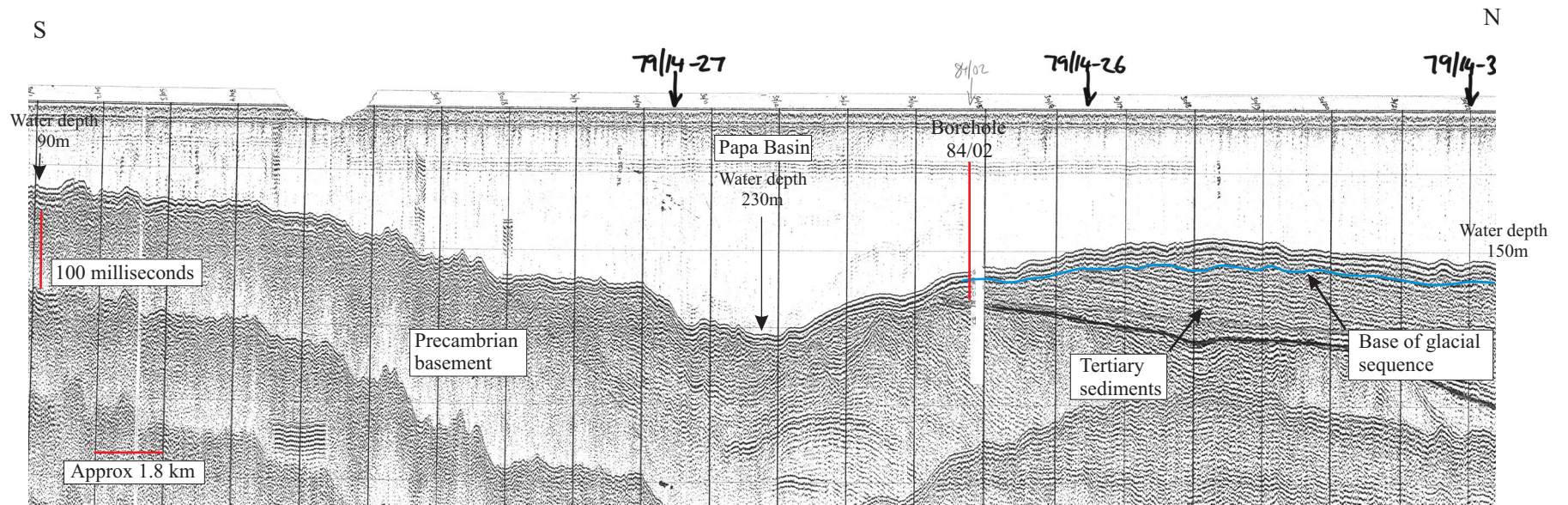


Figure 4.11. Sparker line 79/14-30. Oblique north-south section across the Papa Basin and adjacent shelf showing the rough seabed on the inner shelf side to the south and the smoother undulating seabed of the mid shelf to the north. Precambrian basement gneisses and metasediments underlie the rough seabed whilst Tertiary sediments underlie the glacial sediments forming the undulating seabed to the north. There is a possibility of a glacial sequence in the base of the Papa Basin, but this is mostly obscured by the seabed pulse.

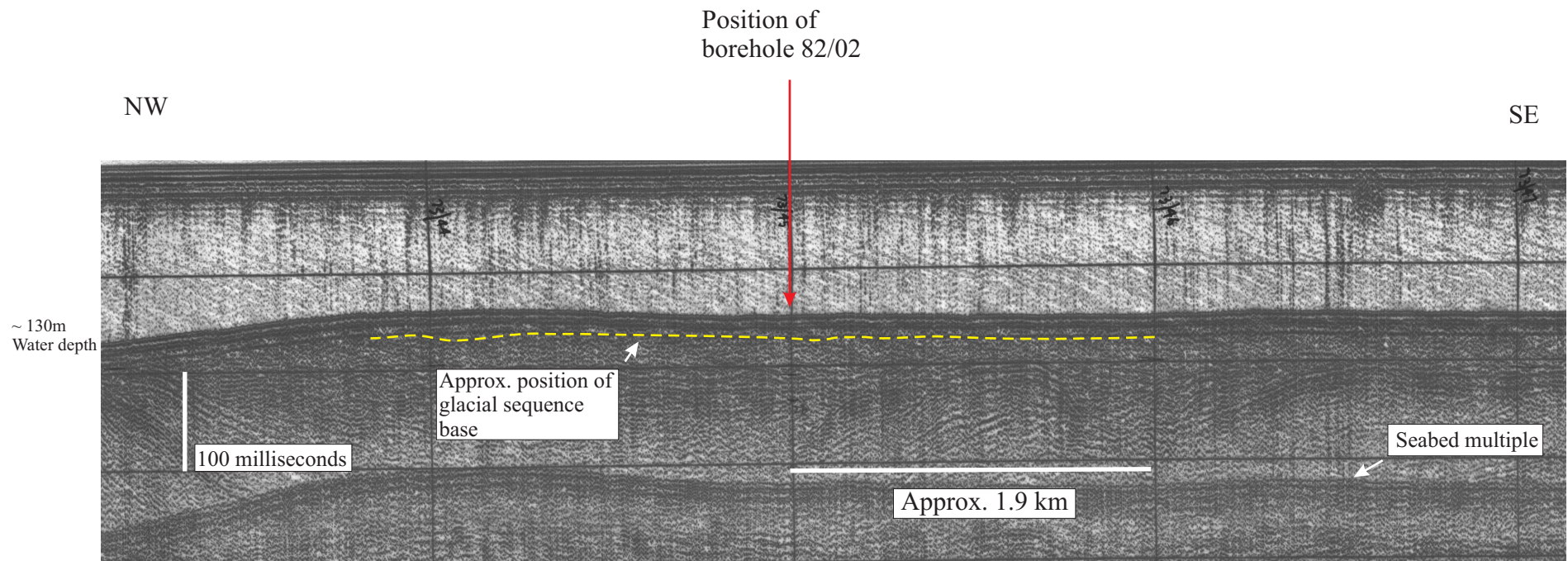


Figure 4.12. Sparker line 74/4-73. Seismic data from the area around borehole 82/02 showing the position of the basal glacial unconformity correlated from the borehole.

4.2.2 Borehole data

Three boreholes lie within area B forming a staggered section across the mid-shelf (Figure 4.7). These are described in order from 82/12 closest to the inner shelf, to 84/02 nearer the outer shelf.

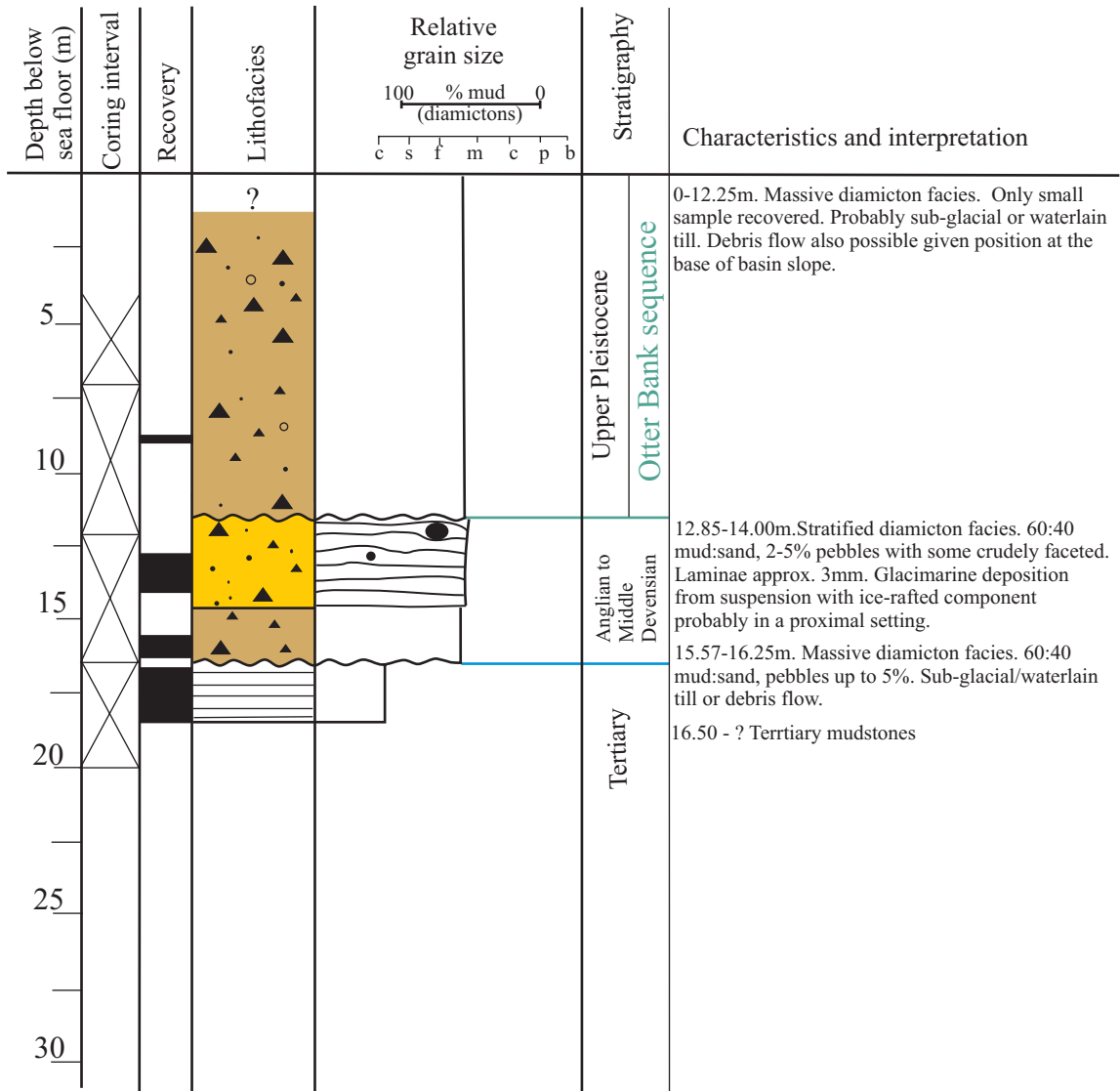
Borehole 82/12

Borehole 82/12 lies close to the floor of the Papa Basin in approximately 200m of water (Figure 4.8). The borehole drilled through 16.5m of Quaternary sediments before reaching Tertiary mudstones. Core recovery was generally very poor, with almost no recovery above 13m. Two lithofacies types were recovered from 82/12; massive diamicton and stratified diamicton (Table 4.1). The core log is shown in Figure 4.13.

The massive diamicton at the base of the Quaternary section is a clast-poor intermediate diamicton with up to 5% pebbles. The sand fraction is 70-80 % quartz with 20-30% lithic grains, composed of red and grey sandstones, quartzite, gneiss and fine-grained igneous fragments. These lithologies also reflect the composition of the pebble suite. Bioclastic debris is generally rare with the exception of a section around 15.90m where shell debris becomes more common. Foraminifera are also rare within this section.

The stratified diamicton section between 12.85-14.00m has a similar composition to the underlying massive facies but with a more diverse lithic grain assemblage. In addition to the lithologies found in the massive facies, the stratified unit includes pale granitic fragments, green micaceous schist and limestone. Texturally, this unit is quite different, with thin alternating laminae of sandy and muddy sediment up to 5mm thick (Table 4.1).

The uppermost massive facies is based upon a very small sample and characterisation is difficult beyond classification as an intermediate diamicton with pebbles. The classification as massive rather than stratified is somewhat speculative,



- Base of Otter Bank sequence
- Base of Quaternary glacial sequence
- Recovered sections shown in black

Figure 4.13. Summary of borehole core 82/12.

but is based upon the absence of lamination, although this may be due to disturbance during drilling rather than true absence. The presence of pebbles within the very small recovered sample also suggests this is massive diamicton which generally has a greater pebble content than the laminated facies.

The sediments recovered from 82/12 are all consistent with having been deposited under a glacial influence. The lack of sorting within the basal massive diamicton facies indicates rapid deposition either as a sub-glacial or an ice proximal water-lain till. The absence of internal structure within the sediments suggest an ice proximal setting is more likely than a sub-glacial setting, but without high-resolution seismic data the absence of a larger scale fabric cannot be confirmed. When considered in conjunction with the overlying units, which are more ice distal, it seems possible the massive diamictons represent an ice proximal retreat/decay phase. The presence of marine macrofossils is an important indicator in interpreting the sediments as ice-proximal rather than sub-glacial (e.g. Powell & Cooper, 2002). However, the relative scarcity at most levels suggests that the sedimentation rate was too high to allow extensive colonisation. This idea is supported by the relative lack of microfauna for the same reason.

The overlying stratified diamicton facies marks a change to a lower energy depositional setting. The stratified facies with the presence of dropstones is interpreted as deposition from meltwater plumes by suspension settling with an ice-rafted component. The relatively thin laminae in this unit suggest a relatively ice-distal setting (Ó Cofaigh & Dowdeswell, 2001). This concurs with the more ice proximal diamictons beneath, suggesting an ice retreat from the Papa Basin

The upper massive diamicton unit is above the interpreted base of the Otter Bank sequence, as correlated with boreholes 82/10 and 82/11 (Chapter 4.3.2). This unit probably represents an ice proximal or sub-glacial setting but with the poor core recovery further interpretation is not possible. However, the presence of the massive diamicton within the Papa Basin at a depth of 190m indicates that the ice must have been at least that thick. The presence of the earlier sequence also shows that the Otter

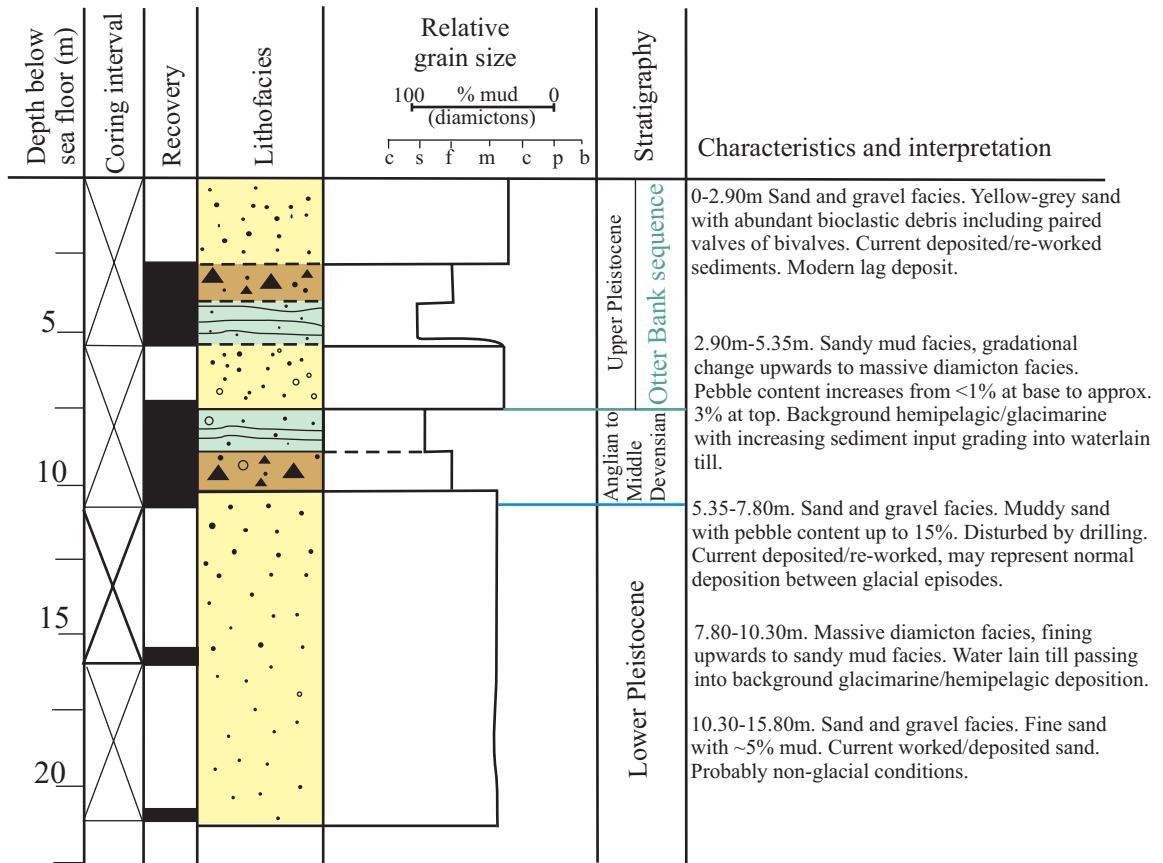
Bank ice sheet did not scour the base of the Papa Basin, and that overdeepening must have occurred in an earlier glacial phase.

Borehole 84/02

Borehole 84/02 lies on the north west side of the Papa Basin at a water depth of about 180m (Figure 4.7). The core from borehole 84/02 (Figures 4.10, 11 and 14) appears to be dominated by sand and gravel facies sediments, although there are large gaps in the recovered core. The uppermost sand unit, restricted to a small sample above 2.90m, has a low mud content and contains a diverse assemblage of macrofauna including scaphopods, echinoid spines and bivalves, including complete specimens of the genus *Mya* which is an infaunal feeder. The combination of low mud (>15%), high sand and abundance of fauna indicates a current deposited or re-worked depositional process with a well oxygenated seabed. The sand unit between 5.35m and 7.80m has similar characteristics although the macrofauna are less abundant. The lowermost sand unit occurring below 10.30m is different in that macrofauna are relatively scarce. The mud content is still very low and the sand fraction 85-90% quartz. The sand is interpreted as a current worked deposit, the greater thickness suggesting these conditions were established for a longer period than those which deposited the upper sand units.

The basal sand and gravel facies section between 21.00m and 10.30m includes the interpreted base of the glacial section, although this is not readily identifiable. The position of the unconformity is based upon palynological results which indicate the sequence below 10.60m is of Early Pleistocene age (Harland, 1998 unpublished BGS data). Re-working of sediments in this facies could cast doubt on the accuracy of the age below the interpreted base of the glacial sequence. However, the palynological data does indicate that it is not younger than Early Pleistocene and therefore is older than the Late Quaternary glaciations.

The units of massive diamicton facies and sandy mud facies are both linked. The lowermost unit between 7.80m and 10.30m fines upwards from clast-poor muddy diamicton to sandy mud with pebbles. Micro and macrofauna are scarce throughout



— Base of Otter Bank sequence
 — Base of Quaternary glacial section

Figure 4.14. Summary of borehole core 84/02

this section. Pebble content remains relatively constant at approximately 2-3% but the proportion of lithic grains in the sand fraction decreases upwards from 30% to 15%. The gradational change in facies showing general fining upwards suggests the base of this unit is ice-proximal waterlain diamicton, with the upper section showing a decreasing input of sediment reflecting a more ice-distal setting as the ice retreated. Alternatively, this may simply reflect a decrease in sediment input due to a decrease in melting. The sandy mud facies is overlain by sand and gravel facies. Disturbance by drilling activity makes it impossible to identify the nature of this boundary but the dramatic change in grain size suggests it may have been abrupt.

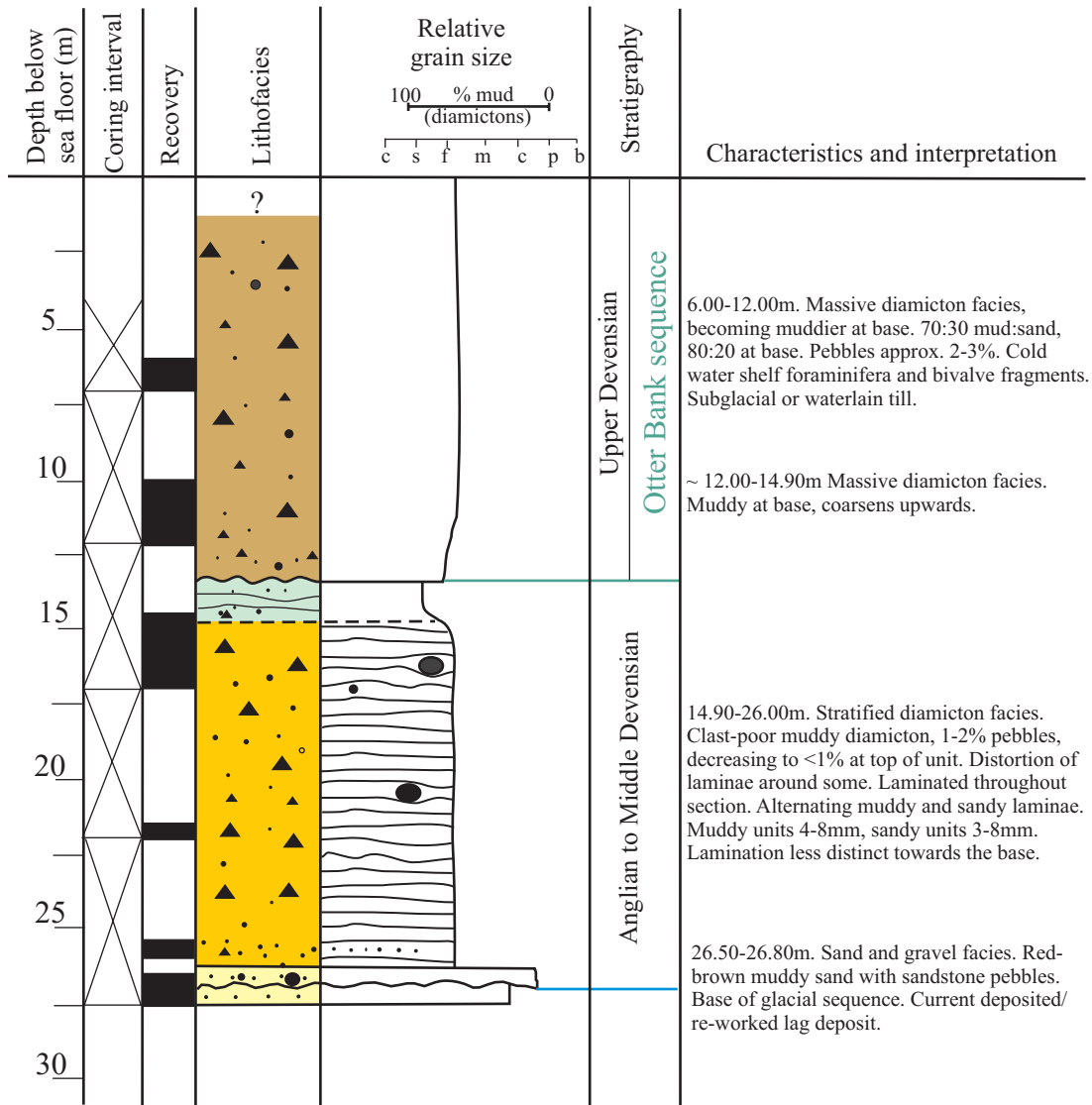
The upper massive diamicton unit lies above a sandy mud unit with a gradational change, suggesting a change from ice-distal to ice-proximal deposition, or an increase in sediment input. This upper unit at 2.90-5.35m lies above the interpreted base of the Otter Bank sequence and is probably unrelated to the lower massive unit.

Borehole 82/02

Borehole 82/02 lies at the south-western end of the Papa Basin on the slope-ward side of the Shetland Spine Fracture Zone in a water depth of about 130m (Figure 4.7 and 4.12). Core recovery from 82/02 was better than other boreholes in this region, but still only reaches about 25%. All four facies types are represented in the core from 82/02 which is dominated by diamictons. The log of the borehole core is shown in figure 4.14.

The base of the glacial section in borehole 82/02 is represented by a thin sand and gravel facies unit between 26.80m and 26.50m, and is interpreted as a basal lag deposit formed by re-working of pre-existing sediments prior to the onset of ice-advance over the locality.

The lower part of the glacial section between 14.90m and 26.50m is composed of stratified diamicton facies. The lowermost 20cm are muddier than overlying sections, although disturbance from drilling has disrupted the lamination and may have re-worked the sediment. Above a sandy horizon at 25.80m the stratified facies becomes



———— Probable base of Otter Bank sequence

———— Base of glacial sequence

Figure 4.15. Summary of borehole core 82/02.

clearly laminated, with alternating sandy and muddy layers. Isolated pebbles which distort the laminae are scattered throughout the section but become scarcer towards the top and are an indicator of an upward-decreasing IRD input. The sand fraction is 75-85% quartz with grain morphologies covering the whole range from well-rounded to angular. Shell debris is generally rare and the microfauna, which includes shelf water depth and pelagic foraminifera and ostracods, is uncommon. The stratified section is interpreted as ice-proximal water lain till (Ó Cofaigh & Dowdeswell, 2001). The presence of IRD indicates that any sea ice present at the time of deposition was not permanent, allowing the movement of icebergs. The regular alternation of laminae indicates a cyclical change in sediment delivery, possibly due to seasonal changes, but with a generally low rate of deposition. The upper section of this unit fines upwards, passing into sandy mud facies sediments without stratification. The upper boundary of this unit correlates lithologically with the base of the Otter Bank sequence as defined in borehole 82/10 (Figure 4.22, Section 4.3.2).

The overlying section of massive diamicton facies represents the Otter Bank sequence. The base at 14.90m is muddier than the overlying sediments but has a higher pebble content. The pebbles may be related to a lag deposit from the base of the Otter Bank sequence, and is inferred to represent an unconformity (Figure 4.15). Shell debris is generally scarce throughout this section and the microfauna is limited to uncommon specimens of the cold water foraminifer genus *Elphidium*. This section is interpreted as being deposited in an ice proximal submarine setting with a high sediment input.

4.2.3 Interpretation of Area B

The seabed morphology of area B is related on a large scale, to the presence of the Papa Basin, and on a smaller scale to the change in bedrock geology across the Shetland Spine Fault. The irregular seabed on the shelf south east of the Papa Basin (Figures 4.9 and 4.10) is underlain by Precambrian basement rocks, whereas the smoother seabed to the north west, and on the floor of the basin is underlain by Tertiary sedimentary rocks. The smoother seabed morphology to the west also

includes mounded Quaternary glacial sediments, interpreted as moraine deposits on the basis of their bank-like morphology and lateral continuity (Figures 4.9 and 4.10)

The resolution of the seismic data is insufficient to subdivide the glacial section. However, when combined with the core data, a correlation between seabed morphology, underlying bedrock type and the presence of more than one glacial sequence becomes apparent. All three boreholes contain elements of both the Otter Bank glacial sequence and an earlier glacial sequence. All the boreholes also lie in areas of smooth seabed with the glacial sediments resting upon Tertiary sedimentary rocks. In addition, the seismic data from line 79/14-26 (Figure 4.10) indicate that the smooth seabed is bedrock-related, and not water-depth related, as the irregular seabed to the east of the section lies at the same depth as the smooth seabed to the west.

The relationships observed above indicate that where there is a softer, more easily eroded bedrock, the seafloor is smoother and deposition of thicker glacial sediments occurs. In addition it suggests that the rough seabed areas to the south east of the Papa Basin are probably only covered by a single glacial sequence which reflects an irregular bedrock surface beneath. The presence of both glacial sequences in borehole 82/12, close to the deepest part of the Papa Basin, indicates that erosion by the Otter Bank age ice sheet did not reach the bedrock floor of the basin, but instead travelled across the pre-existing glacial sediments. Thus, the overdeepening of the Papa Basin (Stoker et al, 1993) pre-dates the Otter Bank phase of glaciation.

Glacial facies interpretation

The pre-Otter Bank section in boreholes 82/02 and 82/12 is represented by stratified diamicton facies sediments interpreted as being deposited by suspension settling in an ice-proximal setting. This interpretation is based upon comparisons with characteristics from stratified diamictons from other margins (Ó Cofaigh & Dowdeswell, 2001). The well preserved laminae indicate that bottom current activity was not a major process during deposition of the stratified facies units, the sediment being delivered by over-flowing, buoyant meltwater plumes. The reduction in the

pebble content towards the top of the stratified facies in 82/02 could indicate the growth of an ice shelf which reduced iceberg calving and hence IRD input. However, the thickness of the individual laminae and the sand content suggest IRD was still being delivered and an ice shelf is unlikely. The reduction in pebble content may simply be an indication that the ice front, and consequently the source for sediments input, has retreated and become more distal. This is supported by the gradational change to sandy mud facies in 82/02 (see Figure 4.15).

The pre-Otter Bank sequence is difficult to identify in core 84/02 due to the different nature of the sediments and the lack of lithological similarity to other boreholes. The sand - diamicton - sandy mud transition between 10.30m and 7.80m is interpreted as the pre-Otter Bank sediments, representing a pre Late Devensian glacial episode. However, this interpretation is speculative as there is no dating evidence available. The change to sand and gravel facies at 7.80m (Figure 4.13) is interpreted as the base of the Otter Bank as it appears to mark the beginning of a glacial advance sequence. From this point in the core, the sediments coarsen upwards from sand facies through sandy mud to massive diamicton facies, suggesting the ice front is getting closer to the locality. The sand at the base is the product of current activity which declined as full glacial conditions became established.

The sand dominated sediments from borehole 84/02 may reflect its position towards the north-western end of the Papa Basin. This is within the area indicated as a possible outflow from the basin to the shelf edge (see Figure 3.4) and may have been subjected to more active or prolonged currents than the other localities in area B.

The Otter Bank sequence in boreholes 82/02 and 82/12 is represented by massive diamicton facies interpreted as ice-proximal waterlain sediments. The presence of cold water foraminifera and bivalve fragments in the diamictons of 82/02 indicate a marine origin rather than a sub-glacial till origin for these deposits, although this assumes that the fossil material is not re-worked from an earlier glacial deposit.

4.3 Area C: Outer shelf

Area C, located in Figure 4.16 covers a relatively large part of the outer shelf and contains a wide range of different features and depositional environments. In addition, some of the processes occurring on the outer shelf at the shelf break are transitional with the upper slope, and have changed over time as the shelf advanced through progradation.

Data for the outer shelf is more comprehensive than for other parts of the shelf, although the seabed image only covers part of area C. Core recovery is still generally poor due to the nature of glacial diamictites containing large cobbles and boulders. Seismic survey coverage is better than for the inner and mid shelf areas and suffers less from seabed multiples obscuring the sub-surface detail.

Bathymetry data:

The gross morphology of the seafloor in area C shows a shallow intra-shelf basin reaching a depth of 150m, rising to 120m towards the shelf edge before dropping away towards the shelf break (Basin A in figure 4.17). The top of the rise forms an intermittent feature lying roughly parallel to the shelf break and stretching for approximately 70 km. A narrow gap (point 1, Figure 4.17) forms an outflow from the basin to the shelf edge. Closer to the shelf break at a water depth of between 150m and 160m there is a marked decrease in the slope angle forming a broad plateau-like feature nearly 40 km wide.

Seabed image data:

The seabed image data only covers part of area C, and that which it does cover may be subject to processing errors (Chapter 2). To the south-east of the area covered the image shows a series of poorly defined mounds and bifurcating ridges up to 8 km long and 3 km wide (Figure 4.18). The larger features are furthest from the shelf edge and tend to form arc shaped ridges. Towards the shelf edge the mounds become smaller and more linear, forming short ridges with the long axis orientated broadly at right angles to the shelf break. The zone of short ridges lies predominantly within the plateau-like area of the outer shelf described in the bathymetry section above. The

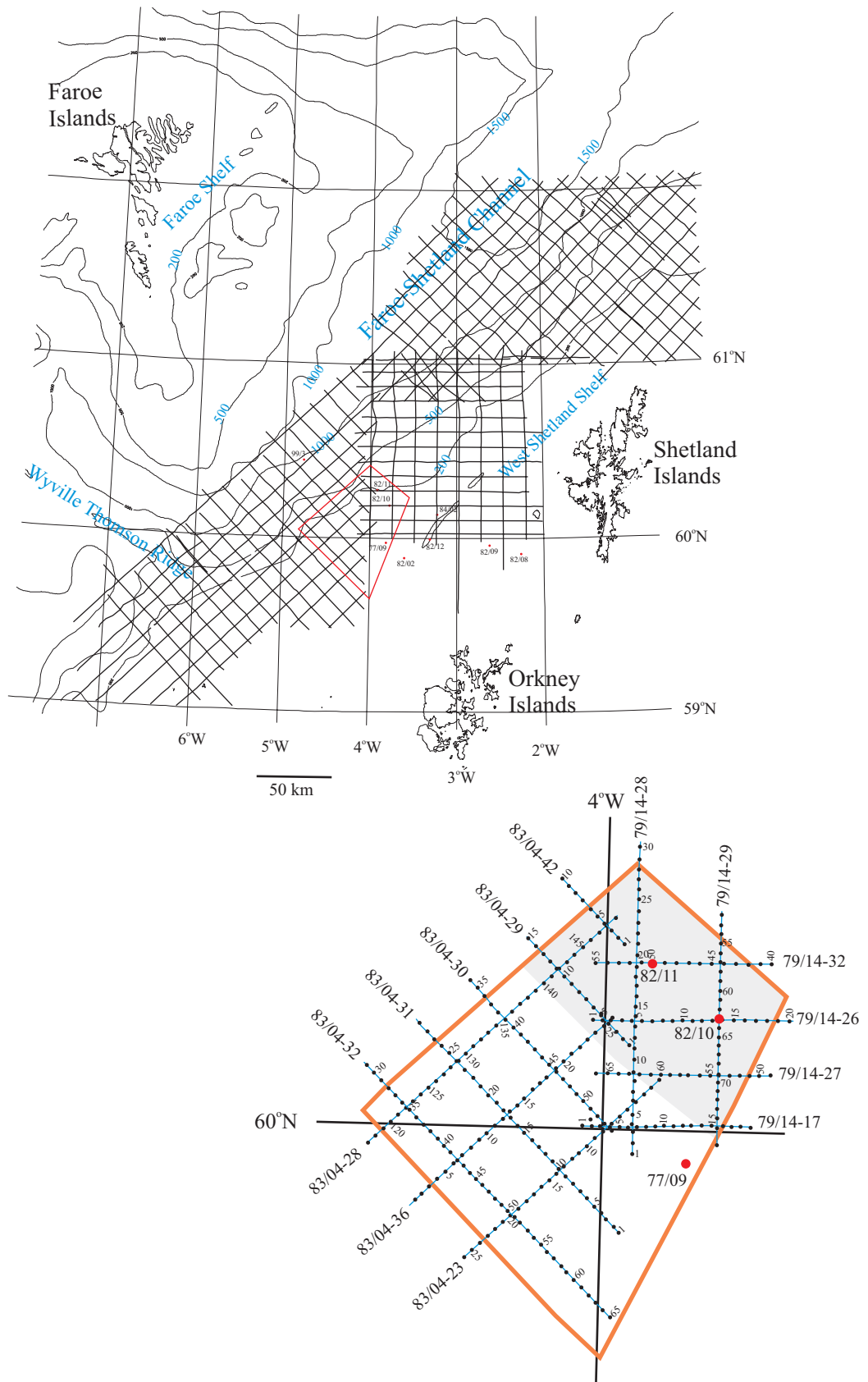


Figure 4.16. Location of area C, Outer shelf. Area Ci is shaded.

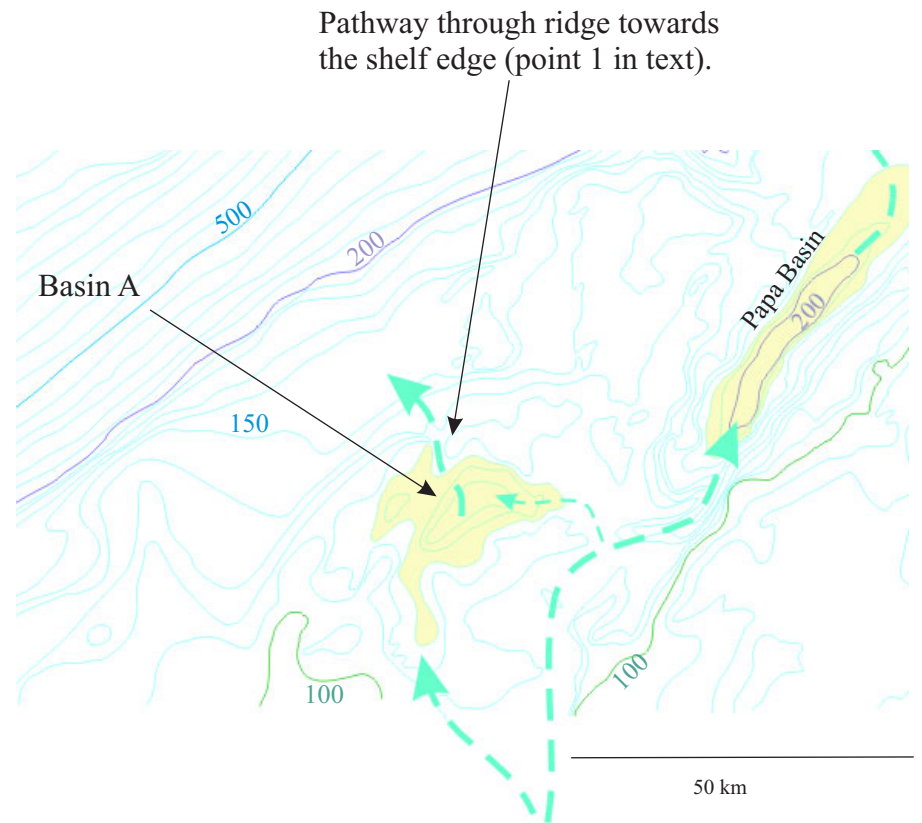
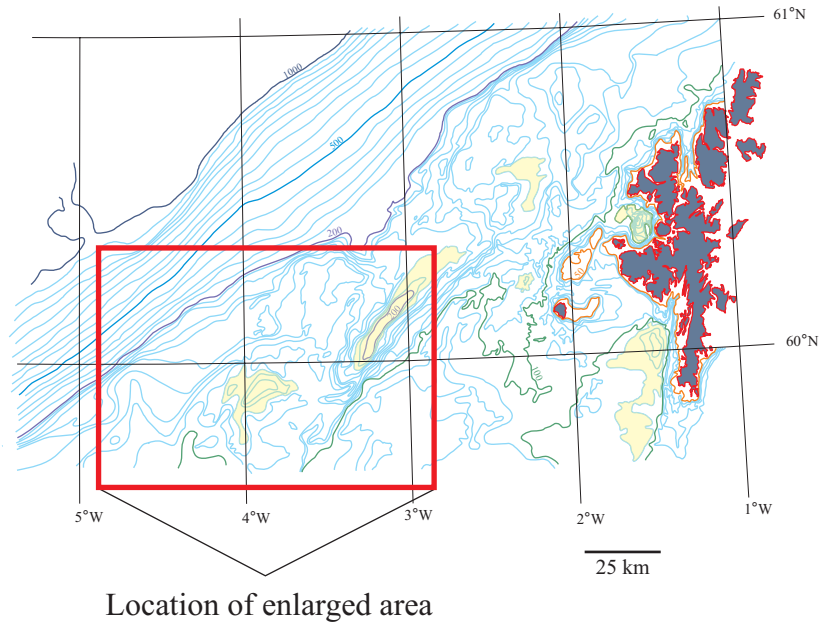


Figure 4.17. Location of the intra-shelf basin and the likely inflow and outflow pathways in area C.

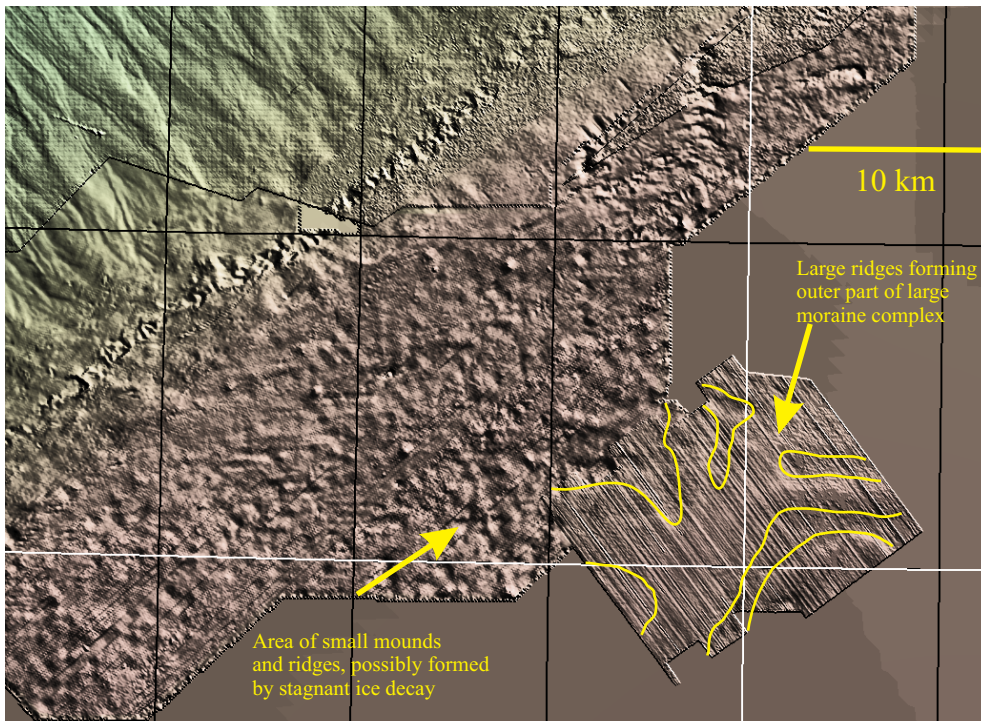
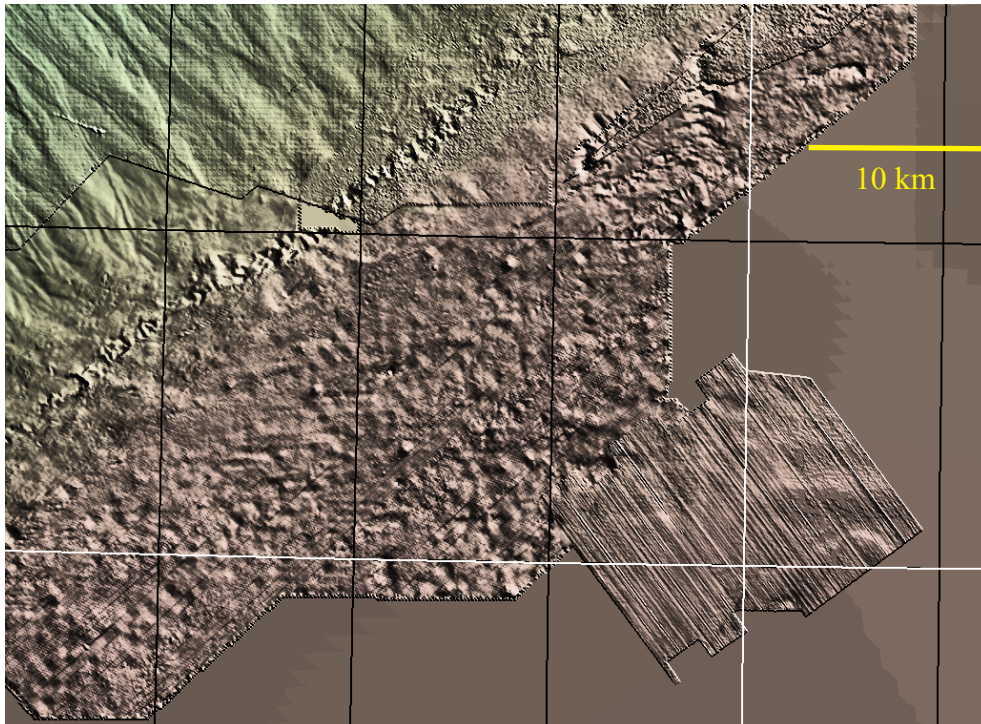


Figure 4.18. Enlarged seabed image of the outer shelf showing large bifurcating ridges and numerous smaller mounds. The mounds have previously been interpreted as processing artefacts. However, close examination shows they cross the boundary between different data sets

opening at point 1 lies in a gap in one of the larger ridges and appears to have a smaller ridge emanating from it (Figure 4.18).

The larger ridges are part of a complex described previously from seismic data and interpreted as glacial moraine ridges (Stoker and Holmes, 1991) and are dealt with more fully in the following sections. The smaller ridges and mounds have previously been considered to be erosional features or data artefacts (Bulat and Long, 2001). However, features appear to cross the boundary between different data sets, indicating that these features probably do exist. In addition, the boundaries of the outer moraine ridge, which are confirmed from seismic data, continue into the area of apparent artefacts. The mounds bear a strong similarity to depositional features seen in areas formerly occupied by stagnant ice in a sub-aerial setting which has undergone *in situ* decay (Figure 17D in Eyles and Eyles, 1992). There is also a resemblance to larger-scale drumlin fields.

Seismic data

Seismic survey coverage of area C is extensive compared to the previous areas and also includes data from the 1983 survey which is more detailed than the older surveys. Lines parallel to the main overall transect (SE – NW) show a variety of features in detail, although none show the best examples of all the features. Several lines from the south west of the area are presented in their entirety, with additional sections from other lines showing specific features. In the north east of area C survey coverage is less detailed. There is also a marked contrast in the type of features present. For this reason the two parts are described and interpreted separately.

In the following descriptions the area has been split into two different zones (Figure 4.16) Ci and Cii due to a major difference in the morphology and processes acting within them.

4.3.1. Area Ci: Sea-bed morphology and seismic data

The north-eastern section of area C is a key element in the interpretation of the entire margin. Within this section lie two important boreholes in which the Otter Bank

sequence is defined. The base of the Otter Bank sequence in the core is also identifiable on the seismic survey data which allows the horizon to be correlated with other seismic lines and boreholes.

The innermost of the boreholes 82/10, lies on survey line 79/14-26 shown in Figures 4.16 and 4.19. The basal unconformity of the glacial section shows a marked change in depth from east to west. From the eastern end of the section towards the centre, the unconformity shows broad, shallow undulations with a total glacial thickness of less than 30m. Further west the unconformity is more irregular and lies at a greater depth, the total glacial thickness reaching around 50m. The seabed along this line is very smooth with the exception of a series of small ridges at the eastern end. Although it is difficult to state with certainty because of the wide spacing of seismic lines, these features appear to be a continuation of the large moraine ridges seen on the seabed image to the south-west and described in detail in section 4.3.4. The base of the Otter Bank sequence on line 26 (as interpreted by Holmes, 1991) is only visible in the area of the borehole where it appears as a high-amplitude reflection with considerable relief. Internal structure is limited to a short section of irregular parallel reflections. East of the borehole the base of the Otter Bank sequence is lost in the seabed pulse but must lie at a depth of less than 20m. The moraine ridges at the eastern end are of Otter Bank sediments and reach a thickness of about 33m, forming individual ridges up to 2km wide at the seabed. The two outermost ridges exhibit an asymmetry which resembles that of the large moraines further south, and is most probably related to their mode of formation. The outermost ridge has a long slope towards the outer shelf (in the direction of ice flow), whereas the second ridge has a longer slope towards the inner shelf. This is dealt with in more detail in section 4.3.6. The extreme western end of the survey line (beyond the limit of Figure 4.19) crosses line 83/04-29 which allows the base of the Otter Bank sequence to be identified and transferred to the 1983 survey data.

Line 79/14-32

Line 79/14-32 lies immediately to the north line 26 but reaches closer to the edge of the shelf due to the oblique angle of the survey grid to the shelf edge. Line 32 is

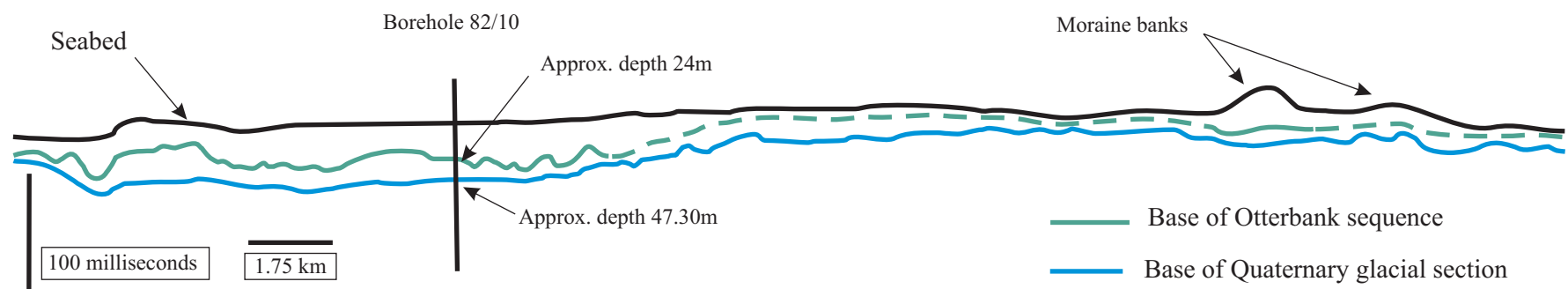
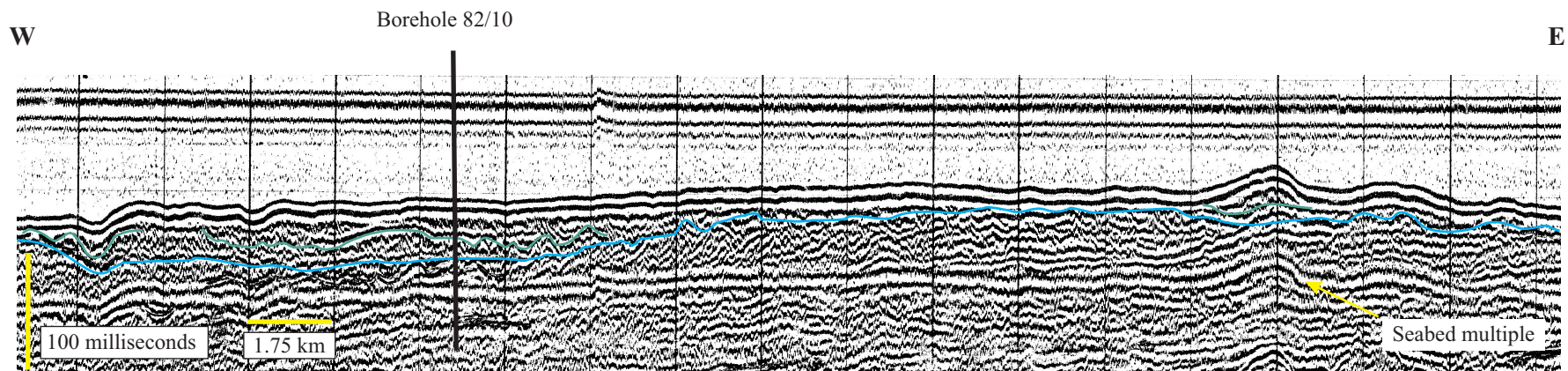


Figure 4.19. Sparker line 79/14-26 showing the definition of the Otterbank sequence base in borehole 82/10. Resolution is generally too poor to identify any internal boundaries.

important as it crosses borehole 82/11 on the outer shelf and forms another fix for defining the base of the Otter Bank sequence. The section in Figure 4.20 shows the outermost part of line 32 which covers the outer shelf sediment wedge from close to the pre-glacial shelf break. At the eastern edge, the unconformity at the base of the glacial section shows an irregular topography towards the ancient shelf edge which appears to be related to the dip of the underlying rocks. Beyond the pre-glacial shelf edge, the glacial unconformity is generally much smoother, dipping towards the basin at an angle generally less than 1° . The acoustic structure of the lower glacial section on the shelf is too thin to resolve. The section beyond the palaeo-shelf edge (position 1 in Figure 4.20) shows an initial group of irregular parallel reflections with some transparent sections. This is shown more clearly on line 79/14-28, described below, which lies perpendicular to line 32.

Moving up-sequence to the west, the stratified units give way to a series of wider spaced undulating reflections which become chaotic to transparent down-dip. The uppermost section, below the base of the Otter Bank sequence, shows a return to a sub-parallel stratified pattern of reflections. This characteristic defines the top of the lower glacial sequence across much of the outer shelf and upper palaeo-slope.

The base of the Otter Bank sequence on line 79/14-32 (Figure 4.20) on the outer shelf is difficult to resolve due to a thick seabed pulse and a thin sequence. In the region of borehole 82/11 and further west the sequence becomes clearer, where it consists of sub-parallel and convoluted reflections with low continuity. The seabed also shows a more irregular topography beyond the point marking the palaeo-shelf break at the base of the Otter Bank sequence (Palaeo-shelf break 2, in Figure 4.20). Much of the uppermost Otter Bank sequence is obscured by the seabed pulse, but small sections indicate a thin chaotic package at the top of the sequence west of borehole 82/11.

The prograding, sub-parallel reflections at the base of the lower sequence are interpreted as pro-glacial deposits, which become progressively more ice-proximal as the ice front advanced towards the palaeo-shelf edge at position 1 (Figure 4.20). The

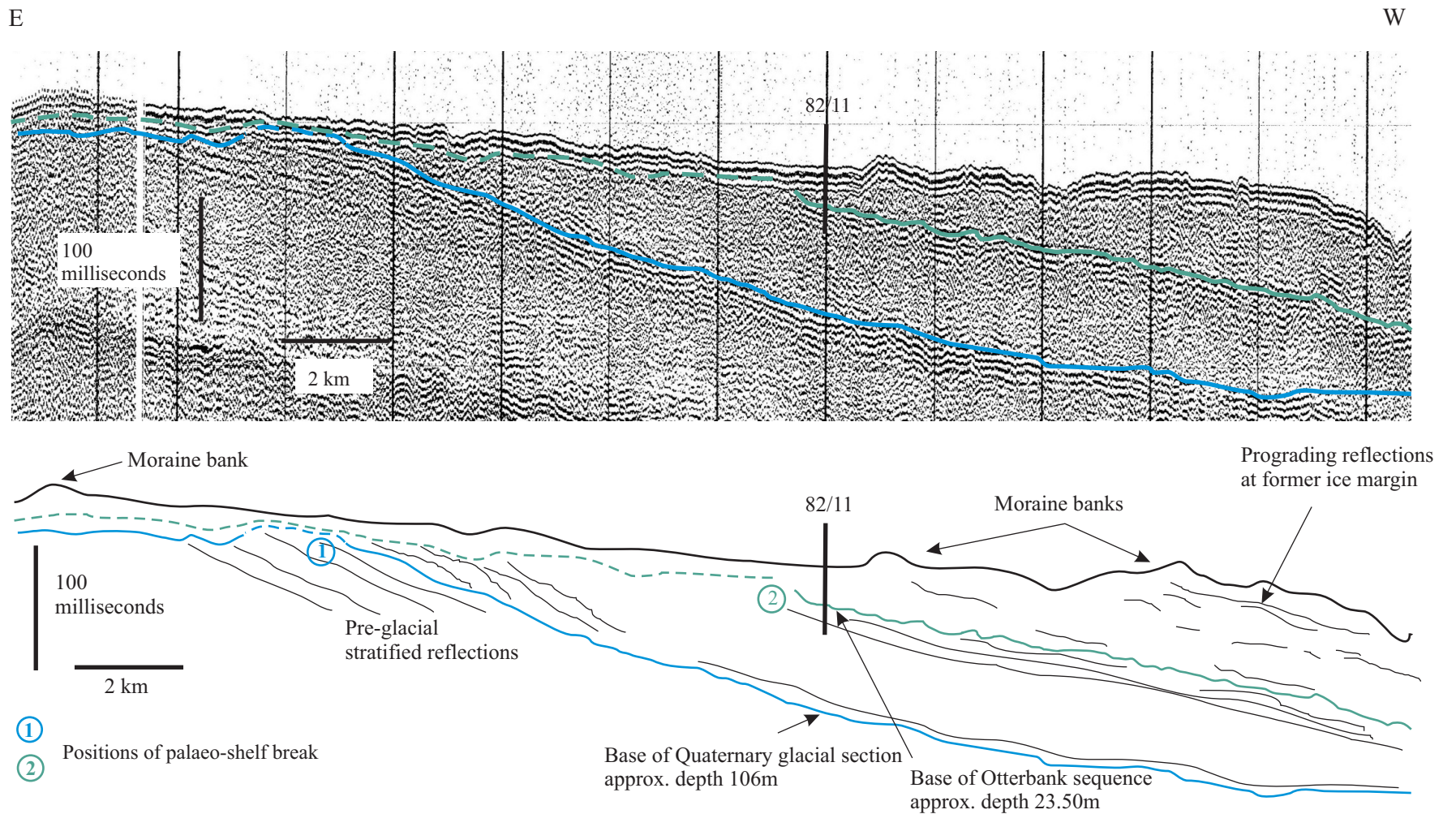


Figure 4.20. Sparker line 79/14-32, showing the definitiob of the Otterbank sequence base in borehole 82/11.

upward change to a more chaotic reflection pattern represents a change to more rapid depositional processes resulting in debris flow deposition and shelf edge progradation. The more structured upper section of the lower sequence is interpreted as a return to a more ice-distal setting where depositional rates are lower. The uppermost parallel sequence of reflections may indicate stratified, more ice-distal sediments deposited from suspension, or late to post glacial current re-working of the seabed. Additional data from borehole 82/11, described below shows that re-working is the most likely scenario.

The reflection pattern of the Otter Bank sequence is similar to that of the underlying glacial sequence, and is interpreted in the same way. The more structured reflections in the lower sections of the Otter Bank sequence are interpreted as prograding, water-lain sediments formed in an increasingly ice-proximal setting ahead of the advancing ice. The more chaotic packages in the upper section of the Otter Bank sequence are interpreted as water-lain deposits formed by rapid dumping and melt-out at, or very close to the ice margin which were re-mobilised to form debris flows. The thin chaotic package forming the uppermost part of the Otter Bank sequence is interpreted as water-lain till, which also forms the mounds and irregular topography on the seabed. The mounds are interpreted as moraine banks, formed during a pause in a general ice retreat phase. The irregular seabed on the outer shelf may be the remnants of two merged moraine banks.

Line 79/14-28

Seismic profile 79/14-28 lies perpendicular to line 32 (Figure 4.16) in area Ci. The lower, pre-Otter Bank sequence shows a closely spaced, sub-parallel stratified reflection pattern forming a prograding package overlying the top of the pre-glacial slope. The package of stratified reflections thins down-dip and becomes poorly defined. Up-sequence to the north, the reflection pattern becomes less structured, with thin stratified sections enclosing poorly structured lensoid and irregular shaped bodies. The top of the lower sequence shows a return to a pattern of parallel and sub-parallel stratified reflections, although these have a more undulating and convoluted

nature than those at the base of the sequence. Evidence of any topset development has been truncated by the emplacement of the Otter Bank sequence.

The bulk of the Otter Bank sequence on the palaeo-slope shows little acoustic structure except for discontinuous parallel reflections near the palaeo-slope break at position 2 (Figure 4.21). These reflections show a vague prograding pattern which disappears within about 1km of the old shelf break. The unconformity at the base of the Otter Bank sequence shows a series of irregular undulations as far as the pre-Otter Bank shelf break at position 2 (Figure 4.21) where it dips down the palaeo-slope. The internal structure of the uppermost section of the Otter Bank sequence, which would have formed the topsets of the sequence, is mostly obscured by a thick seabed pulse. The seabed profile shows a large asymmetric bank at the southern end of the section with a series of sub-parallel mounded reflections within it. Beyond the bank to the north, the seabed drops by 15m and a significant depression of a similar scale is seen in the base of the Otter Bank sequence.

The pattern of reflections at the base of the pre-Otter Bank sequence is interpreted as the product of deposition at the ice grounding line, formed as the ice sheet advanced to the pre-glacial shelf edge. The up-sequence loss of detailed structure represents a change to more rapid deposition and formation of debris flows as sedimentation became more ice-proximal. The elongate and chaotic lensoid packages are typical of the pattern formed by debris flows (Nardin et al, 1979), with the up-sequence increase in magnitude reflecting the increasing rate and volume of sediment deposition. The large scale architecture of down-slope thinning packages resembles patterns predicted in theoretical models of pro-grading glacigenic shelves (Boulton, 1990). The return to a more structured reflection pattern at the top of the pre-Otter Bank sequence is interpreted as a glacial retreat phase, with reduced sedimentation in a more ice-distal setting. Alternatively, it could represent re-working of the glacial deposits during late or post-glacial times.

This pattern of reflections at the base of the Otter Bank sequence is similar to that seen in the initial shelf edge sequence of the older glacial episode, and are interpreted as grounding line deposits formed as the ice-front advanced to the shelf edge. As the

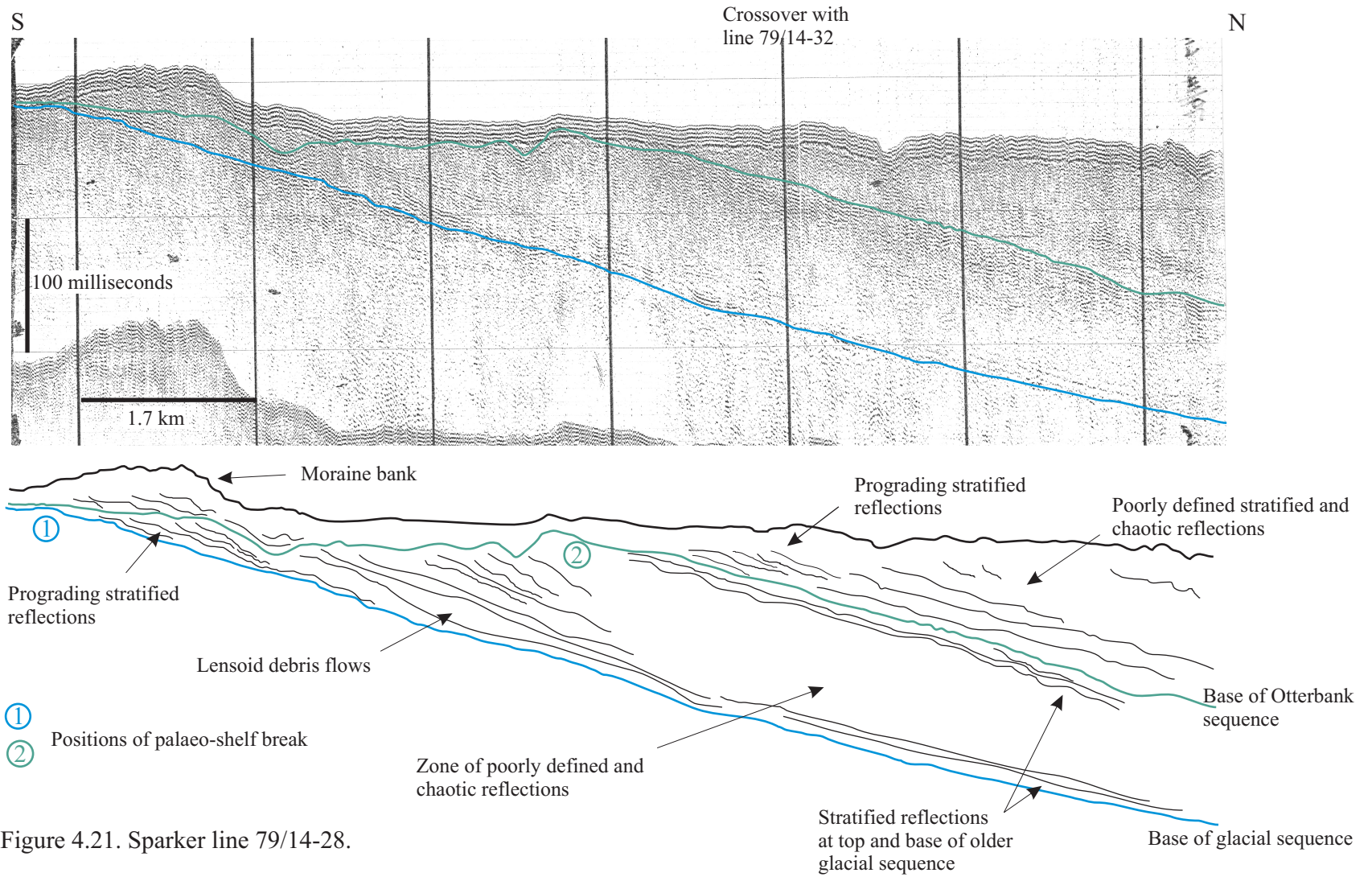


Figure 4.21. Sparker line 79/14-28.

shelf extended by progradation the ice advanced over the earlier Otter Bank sediments, probably re-working the topsets. Any evidence of aggradation is obscured by the seabed pulse. The larger bank at the southern end of the section appears is interpreted as a moraine bank, marking the position of a pause in the retreat of the ice from the shelf edge. The internal structure of the bank shows an asymmetric aggradational mounded pattern of reflections, indicating a development as an ice contact feature at the ice-front.

The slightly deeper water between the moraine and the shelf edge to the north may have led to a break up of the ice front and a rapid retreat to the shallower location where there the ice was less buoyant. This is supported by the smooth seabed and the thin sequence of Otter Bank sediments between the two points, suggesting the ice receded quickly to the position of the moraine and then stabilised for a time.

The origin of the undulations in the base of the Otter Bank sequence and at the seabed between palaeo-shelf break positions 1 and 2 (Figure 4.21) is uncertain. These may be related to erosion and loading by the Otter Bank ice sheet, overriding earlier and un lithified glacial sediments from the previous glacial episode. Loading of the early Otter Bank sediments may also have formed the undulations observed at the seabed beyond position 2 (Figure 4.21).

4.3.2 Area Ci: borehole data

Borehole 82/10

Borehole 82/10 lies towards the outer shelf at a water depth of 140m and penetrated almost 47m of Quaternary glacial sediments, although core recovery was only 24%. The key features of the core from 82/10 are shown in Figure 4.22. All of the diamictons in the core were red-brown in colour at the time of description. However, reference to original drilling records shows that at the time of drilling, most of these sediments were grey or grey-green and have oxidised in the 20 years since collection. The base of the Quaternary glacial section is marked by a change from very hard muddy sand with pebbles, to a stiff to very stiff muddy diamicton of the massive diamicton facies at 46.95m. This change is identified on the seismic section in Figure

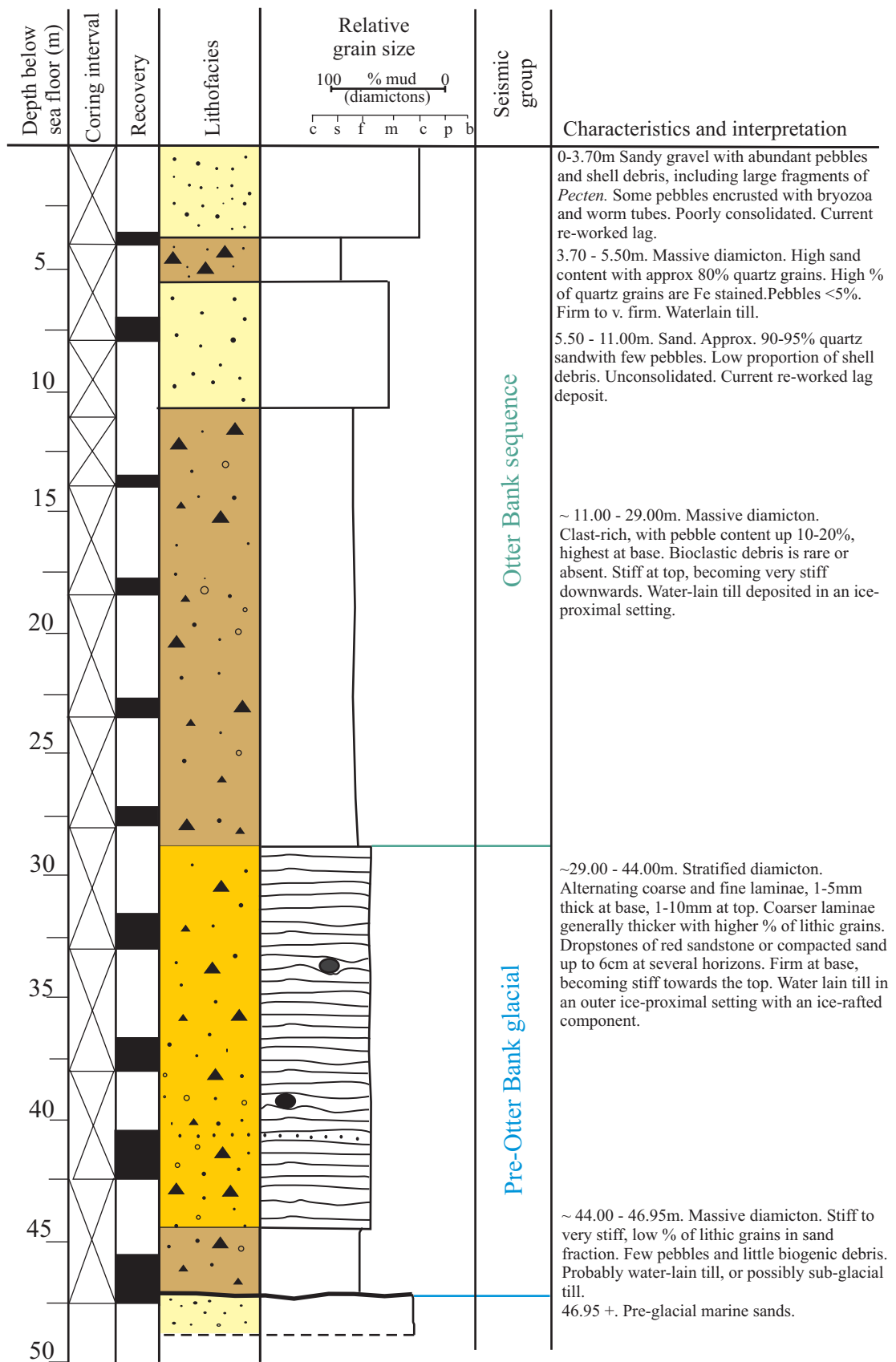


Figure 4.22. Log of core from borehole 82/10

4.19 as the base of the Quaternary glacial section and has previously been defined by Holmes, (1991). The massive diamicton at the base of the section shows a low proportion of lithic grains in the sand fraction, reaching about 15%. This value increases upwards through the other units. The massive nature of this unit indicates deposition was rapid without any significant sorting processes occurring.

Sediments from the stratified diamicton facies were recovered between 42.50m and 31.50m. The thickness of individual laminae within the unit increases upwards from a maximum of about 6mm near the base to 10mm at the top. The sandy laminae tend to be thicker and darker in colour than the muddy laminae (Figure 4.23a & b). In addition, this lithology also contains a diverse assemblage of clasts. The larger clasts are mostly composed of a friable red sandstone and show well developed distortion of the laminae beneath them (Figure 4.23c). Sediments from this section are generally firm to stiff.

Above the stratified diamicton there is a return to massive diamicton facies of stiff to very stiff consistency, with up to 15% small pebbles between 28.00 – 27.15m. (Figure 4.23d). Pebble content remains high throughout the other sections of massive diamicton above this level. Composition of the pebbles and lithic grains is varied, including gneiss, quartzite, schists, red sandstones and siltstones. The lithic grain assemblage also includes up to 5% black to greenish black grains of unknown composition, which may be degraded volcanic glass.

The upper section of the core from 82/10 is predominantly sand and muddy sand with a short section of firm, massive diamicton facies sediments between 4.0m and 3.65m. The uppermost sand and gravel facies unit above 3.65m contains abundant shell debris and encrusted pebbles.

The lowermost diamicton is interpreted as a water-lain, ice-proximal till or a sub-glacial lodgement/deformation till. Distinguishing between the two from the core data is not possible, although the sands beneath the diamicton are very hard, suggesting they could have been over-ridden by ice. The diamictons are less compact



Figure 4.23a. 41.10 - 41.20m
Stratified diamicton facies



Figure 4.23b. 32.50 - 32.80m



Figure 4.23c. 32.80 - 32.90m

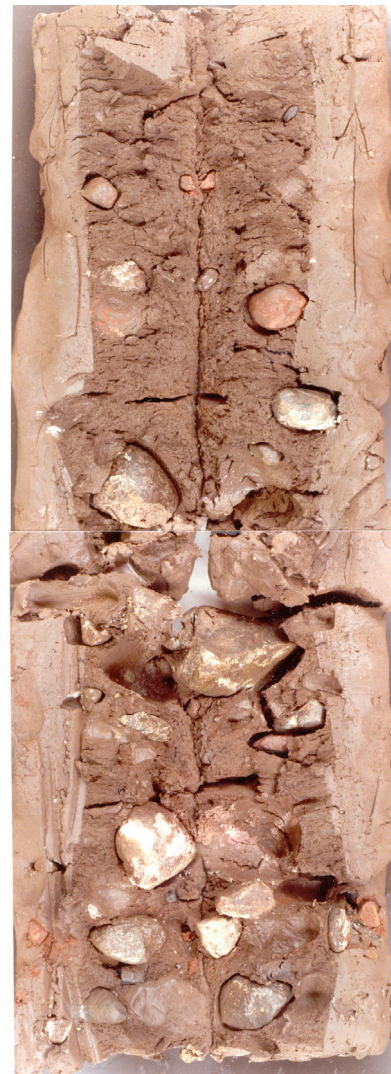


Figure 4.23d. 27.54 - 28.00m

Figure 4.23. selected sections from
borehole core 82/10. Photographs
taken from Cockcroft (1987).

and therefore more likely to be water lain. The upwardly increasing lithic content of the sand fraction suggests initial erosion of a quartz rich source but with later erosion of a more basement-like source. This is consistent with ice initially traversing a marine shelf with compositionally mature sands at the seabed. As advance continues, debris derived from bedrock in more distant areas on the inner shelf and possibly onshore are delivered to the outer shelf, giving rise to the increasing lithic content.

The stratified diamicton is interpreted as sediment deposited from suspension, probably as a result of meltwater plumes in the outer proximal zone. The presence of out-sized clasts, interpreted as dropstones, indicates the inclusion of an ice rafted component, implying the presence of open water at the time of deposition (Ó Cofaigh & Dowdeswell, 2001). The relative lack of compaction indicates that deposition from over-riding ice is unlikely.

The massive diamicton from above 29.00m is interpreted as proximal water-lain till, although deposition in a sub-glacial setting cannot be excluded. The base of this unit is interpreted as the base of the Otter Bank sequence, partly on the basis of correlation with seismic data (see below), and on the high pebble content at the base. The high pebble content is consistent with formation by an ice advance over a winnowed seabed composed of sand and gravel derived from pre-existing diamictons at the seabed. Such an environment is likely to develop between major phases of glacial activity and has developed at the present seabed since the end of the last glaciation.

The uppermost section of the core above 11.00m is interpreted as a return to fully marine, predominantly ice-free conditions, although this may be a local effect related to meltwater drainage across the shelf. The abundance of marine fauna and lack of mud indicate a higher energy environment where fine sediment is either held in suspension or is being removed. The short section of massive diamicton within the sands suggests a brief return to more ice-proximal conditions.

Correlation of core 82/10 with seismic line 79/14-26

The main features which can be interpreted by integration of the data are the lower limits of the Otter Bank sequence and the base of the Quaternary glacial sequence. The facies change from sand to massive diamicton at 46.95m marks the base of the glacial sequence. This corresponds with the blue reflector in Figure 4.19 using an average velocity of 1.7km/s. Using the same velocity, the base of the Otter Bank sequence lies at the facies change from stratified to massive diamicton at about 29m. This depth is approximately 6-8m deeper than that described by Holmes (1991). The reason for this deeper placement is based upon the abrupt facies change which marks two very different depositional environments and is likely to produce a prominent acoustic horizon.

Borehole 82/11

Borehole 82/11 penetrated over 45m of glacial sediments but did not reach the base of the glacial Quaternary (Figure 4.20). The recovered core is dominated by massive diamicton facies sediments, interspersed with sand and capped by sandy mud (Figures 4.24 and 4.25). The lowermost diamicton, lying below 30.31m suffered from very poor recovery (Figure 4.24) due to frequent encountering of cobbles. Recovery was also hindered by very hard sediments at the top of this section, although below 35m the sediments are firm. The diamictons from 35.0-30.31m are generally sandier than others encountered on the shelf, having a sand content of up to 40%. The sand fraction has a lithic content of about 30%, with a wide variety of different lithologies and mineralogies represented including volcanic glass fragments. Deposition of the diamicton was probably as a water lain deposit at the ice front or in an ice-proximal setting. The hardness of the sediments at the top of the section could indicate deposition by lodgement and deformation beneath the ice. However, as the compactness decreases with depth and the overlying sands are also very hard, it is more likely compaction is the result of ice over-riding the whole section during deposition of the later Otter Bank sequence.

The sand facies interval between 30.31m and 24.25m is composed of very hard, grey-green very fine to medium grained sand with 20-25% mud (Figure 4.25a).

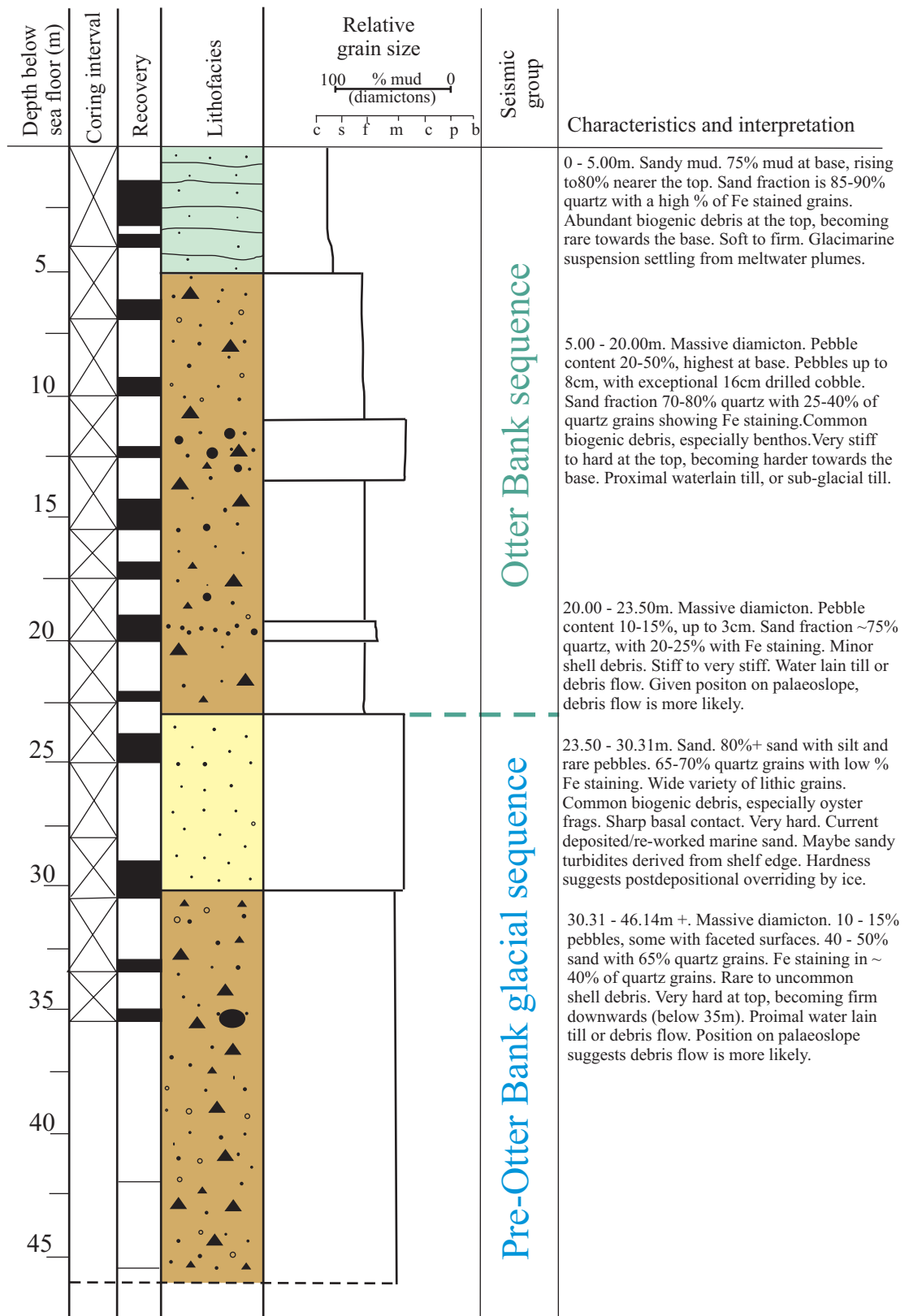
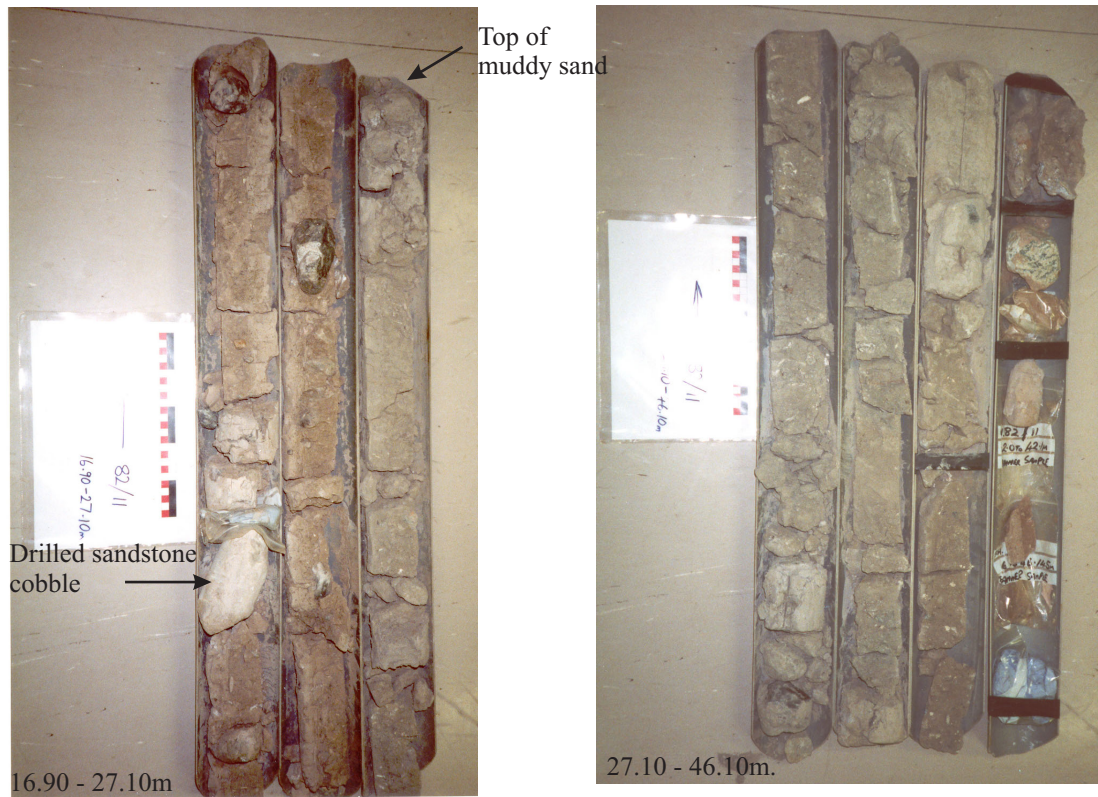


Figure 4.24. Log of core from borehole 82/11



A) Lower part of core from borehole 82/11, showing the lower part of the massive diamicton from the Otterbank sequence with large clasts of sandstone and gneiss. The base of the Otterbank sequence lies between the red-brown diamicton and the top of the muddy sand in the left picture. The muddy sand and lower diamicton mark the top of the pre-Otterbank glacial units.



B) Large faceted gneiss clast from the lower part of Otterbank sequence massive diamicton at 19.20m.

Figure 4.25. Borehole core 82/11. Scale bar is 25cm long.

Mineralogically it is immature, containing up to 30% lithic grains. Bioclastic debris is common with a wide variety of benthos including infaunal bivalves and epifaunal echinoids, corals and encrusting bryozoa.

The abundance of biogenic material coupled with the reduced mud content suggests the sand units are probably not the product of glacial conditions. The lack of a red-brown oxidation colour, even after 20 years of exposure, indicates this lithology is significantly different to all the glacial diamictons from the shelf. The lack of oxidation effects may also be related to the higher biogenic content, which would have increased the availability of organic matter and maintained a more reducing environment.

The upper diamicton section from about 23m to 5m has a generally high pebble content, especially in the section between 19m and 20m. In this section pebble content comprises 50% of the volume, largely due to the presence of several drilled cobbles up to 16cm long (Figure 4.25b). The larger pebbles and cobbles frequently show evidence of faceting. Lithologies of the pebbles and lithic grains are more diverse than most of the other diamictons, including fragments of chalk, dolomitic limestone, red and green sandstone and gneiss. Biogenic material is dominated by bivalve fragments and occasional complete valves. Micro-fauna is generally uncommon and restricted to shelf species of foraminifera. Palynological analysis indicates cold water temperatures for samples between 10.00m and 12.40m (unpublished BGS data). The consistency of the upper diamicton facies unit shows a degree of variation. Between 22.50m and 19m the sediment is firm to stiff, but becomes hard to very hard between 17.50m and 12.50m. Above 12.50m the diamicton softens slightly, becoming stiff to very stiff. The change in consistency of the diamicton from 23m to 14m suggests the lower sections may have been deposited as a water laid sediment but was over-ridden by ice, depositing the harder sediments above as sub-glacial till and compacting those immediately below.

The uppermost unit is a sandy mud facies section from around 5m to the top of the core. Biogenic debris is common in the top 1 metre and contains possible sand-filled burrows. The sand fraction is up to 90% quartz with few mafic minerals and rare

volcanic glass fragments. The fine-grained nature suggests deposition from suspension, although rare pebbles suggest an ice-rafted component is also present. Deposition by suspension settling is also supported by the soft consistency of the mud, indicating very little compaction has occurred. Palynological data suggests a cold water temperate environment in the top 2m (unpublished BGS data).

Correlation of core 82/11 with seismic line 79/14-32

The complete correlation of core data with the seismic data is hindered by the total depth of the borehole which terminates approximately 60m above the base of the glacial sequence shown in Figure 4.20. The lowermost diamictons of the pre-Otter Bank section lie within a zone of vague parallel to chaotic reflections. Only the sand facies section up to 23m is visible as a structured unit. This appears as a series of undulating parallel stratified reflections. The sand is traceable via other lines to the approximate top of the stratified diamicton in line 79/14-26 and core 82/10. This reflector is also detectable in line 83/04-42 which passes within 3 km of line 32 and provides an additional correlation with the 1983 survey lines..

4.3.4 Area Cii

Sparker lines 83/04-31 and 83/04-32 are used as representative lines from which the glacial section of the seismic data can be sub-divided into 9 different packages consisting of 3 broad reflector patterns. Some of the seismic units show a lateral change in acoustic character.

Line 83/04-32 shows a relatively smooth seafloor at the south-eastern, shelf-ward end (Figure 4.26a). Along the survey line to the north-west, the seafloor becomes progressively more undulating. Large elongate mounds with a smooth surface (Figure 4.26b) correspond to those seen on the seabed image and are the lateral continuation of closely similar features interpreted as moraine ridges by Stoker and Holmes (1991) from line 83/04-31. Beyond the mounds towards the shelf edge the seabed becomes flatter again with the exception of a series of small undulations near the shelf break (Figure 4.26c). The unconformity at the base of the glacial sequence shows an irregular topography in the area beneath the moraine ridges, although there

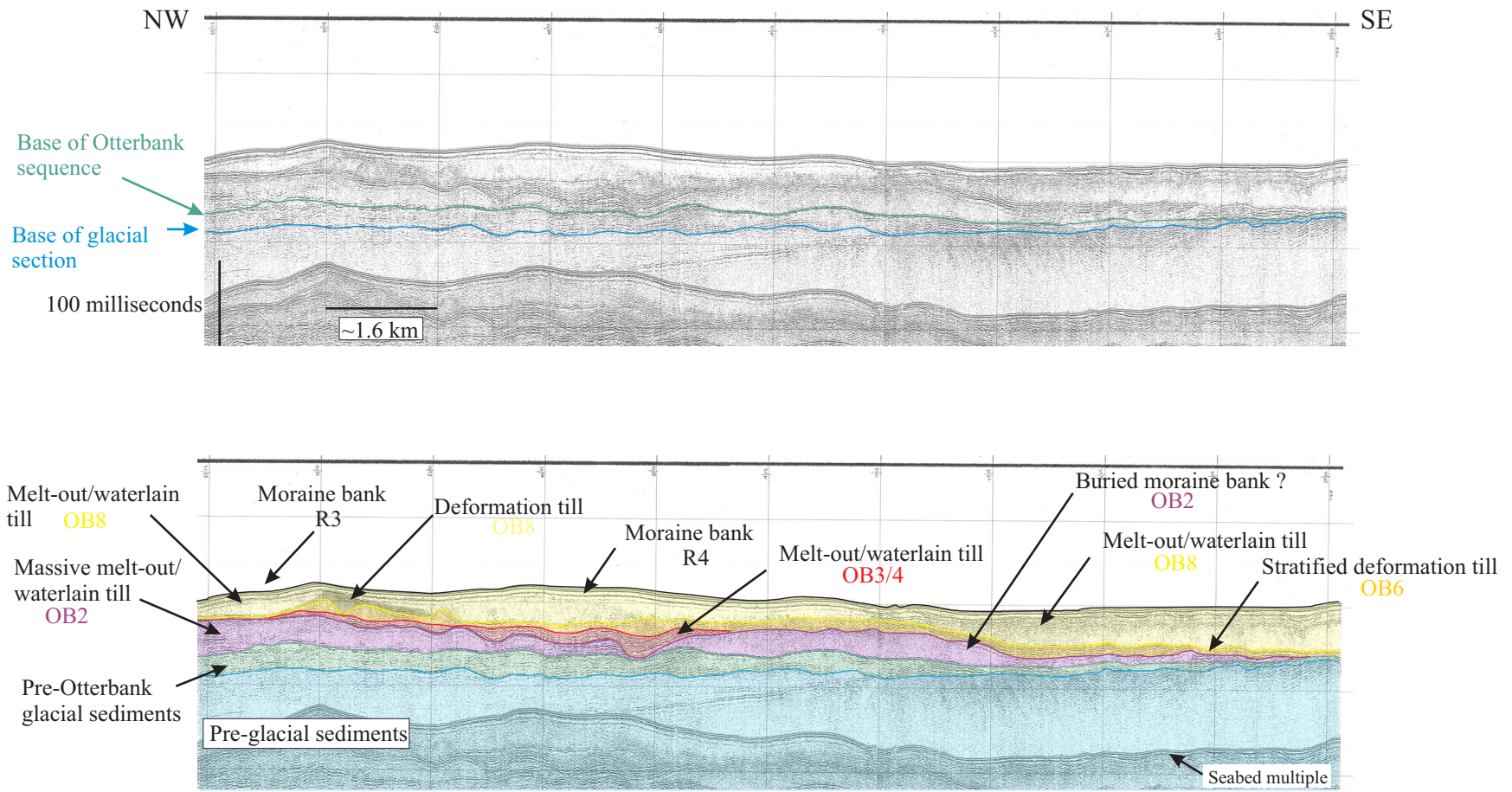


Figure 4.26a. Innermost section of sparker line 83/04-32.

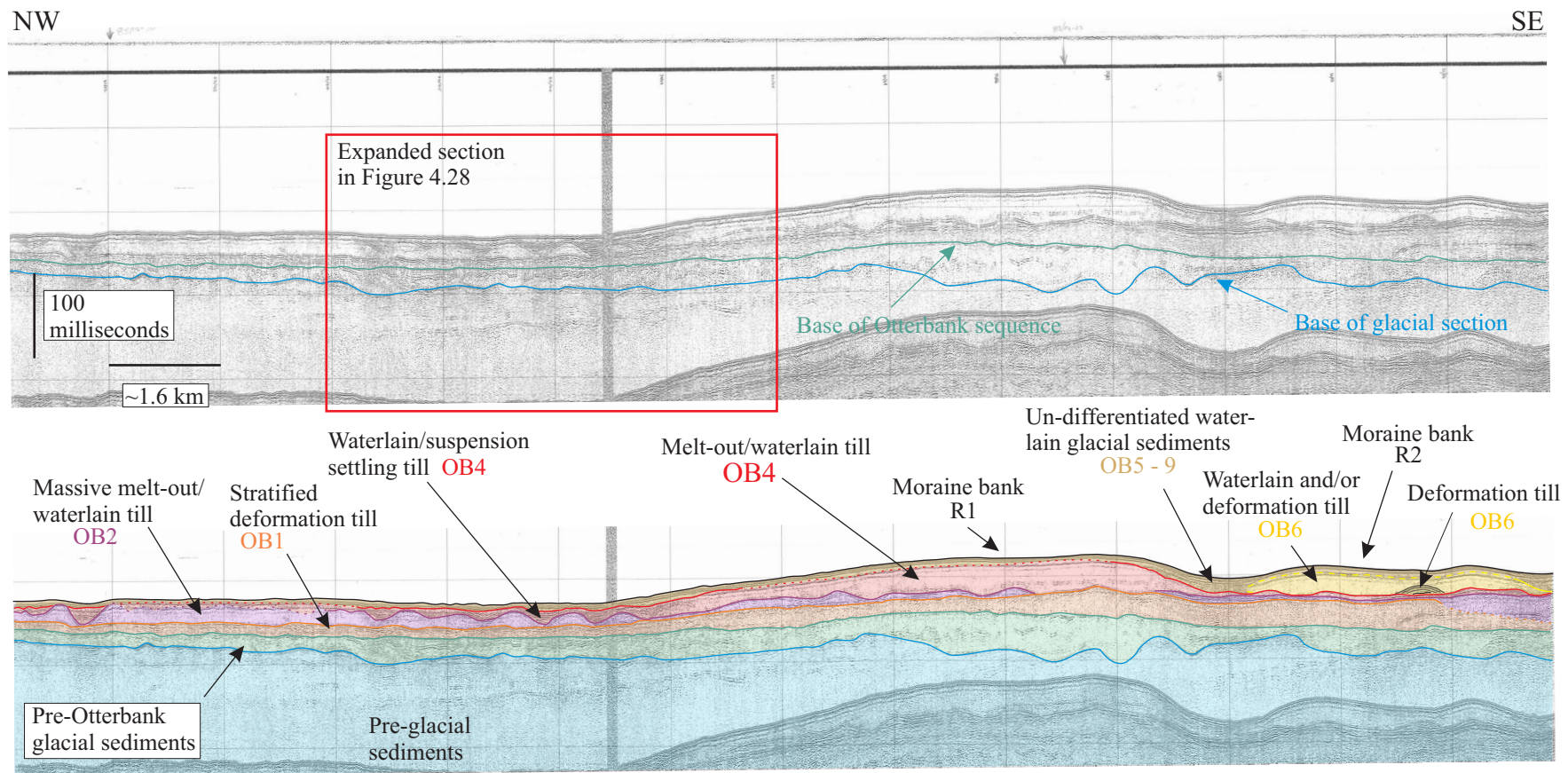


Figure 4.26b. Centre section of sparker line 83/04-32. Outer shelf.

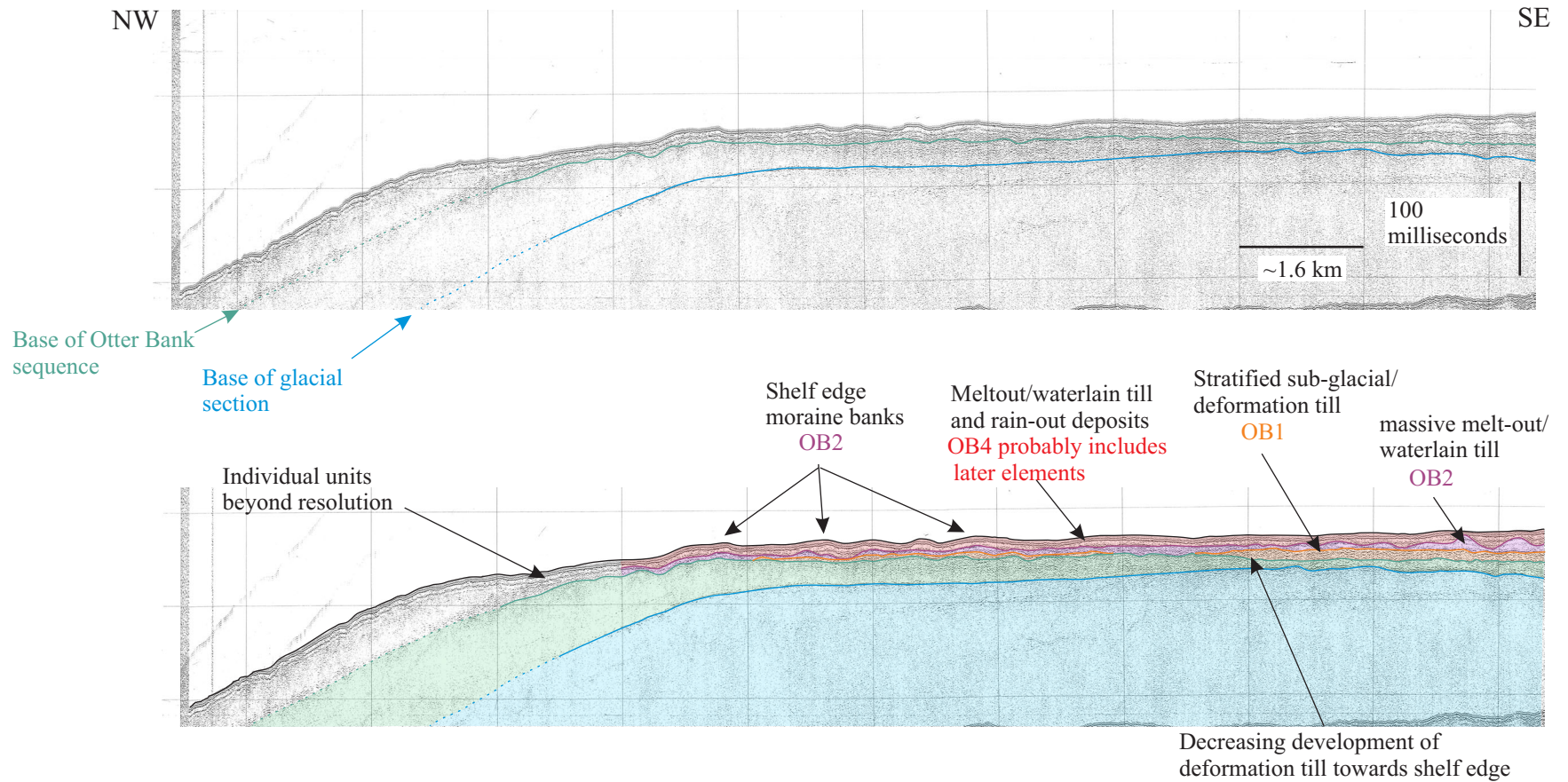


Figure 4.26c. Sparker survey line 83/04-32. The survey line shows a section across the whole of area C from the outer shelf to the shelf break. The Otter Bank sequence shows a composite section which can be sub-divided towards the mid-shelf in to 9 different units. The outer shelf and shelf edge lacks the resolution to identify the individual units. Features of specific importance are enlarged described separately.

is no connection between these features (Figure 4.26b). The unconformity beneath the outermost part of the shelf is generally smooth and slopes at a low angle towards the north-west. In contrast, the unconformity in the outermost shelf zone of line 83/04-31 slopes away from the shelf edge towards the south-east (Figure 4.27).

The pre-Otter Bank glacial sequence broadly consists of sub-parallel stratified reflections with low continuity. These characteristics change very little across most of the outer shelf zone except for a degree of convolution in some sections. The Otter Bank sequence is much more complex, comprising up to 9 different units of poorly-stratified, well-stratified and chaotic reflections. The data from line 83/04-32 shown in Figure 4.26 b is expanded in Figure 4.28 in order to show sufficient detail to separate the units.

The lowermost Otter Bank package, OB1, shows a change in thickness along the length of line 83/04-32. The package is absent from the southeast end of the line nearest the mid shelf (Figure 4.26a), reaches its maximum thickness of about 25m towards the deepest point of Basin A (Figure 4.26b) and then thins towards the shelf edge where it loses its definition (Figure 4.26c). The increased thickness of OB1 where it lies within Basin A, shows that the basin existed at this location before the emplacement of the Otter Bank sequence, and possibly before the deposition of the earlier glacial sequence.

Generally OB1 consists of crudely stratified reflections with variable amplitude and low continuity. The poorly developed stratification, decreases towards the shelf edge, probably formed during the original emplacement of the unit. If this is correct, it indicates probable deposition by lodgement in a sub-glacial setting. Some sections along line 83/04-32 suggest there may be an element of deformation till in the upper parts of the unit where reflections show a degree of contortion. This is most evident beneath the large moraine banks in Figure 4.26b. The development of deformation till in the upper section of OB1 may be related to original deposition, or may be the result of deformation during overriding by ice during the deposition of OB2.

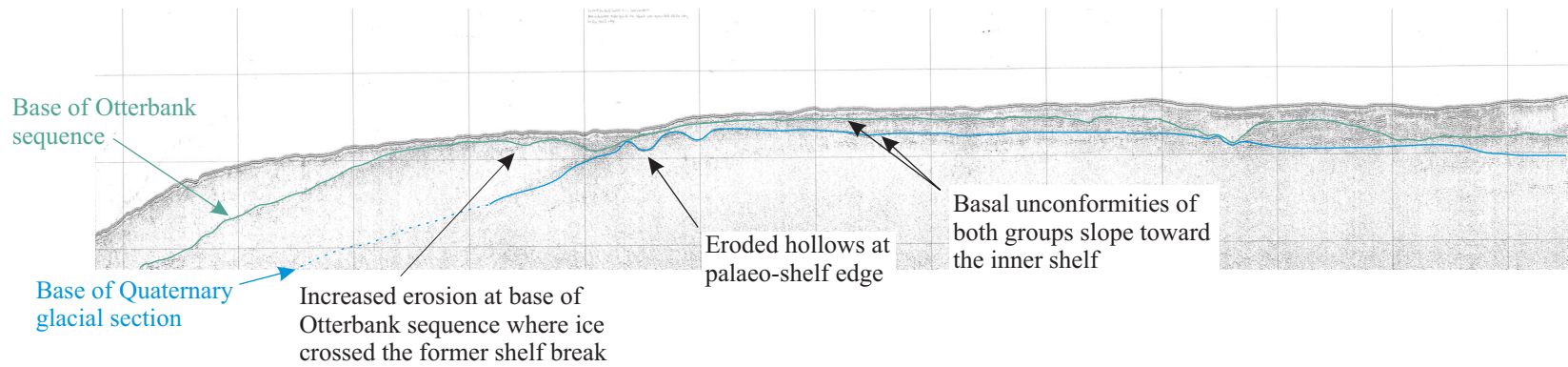
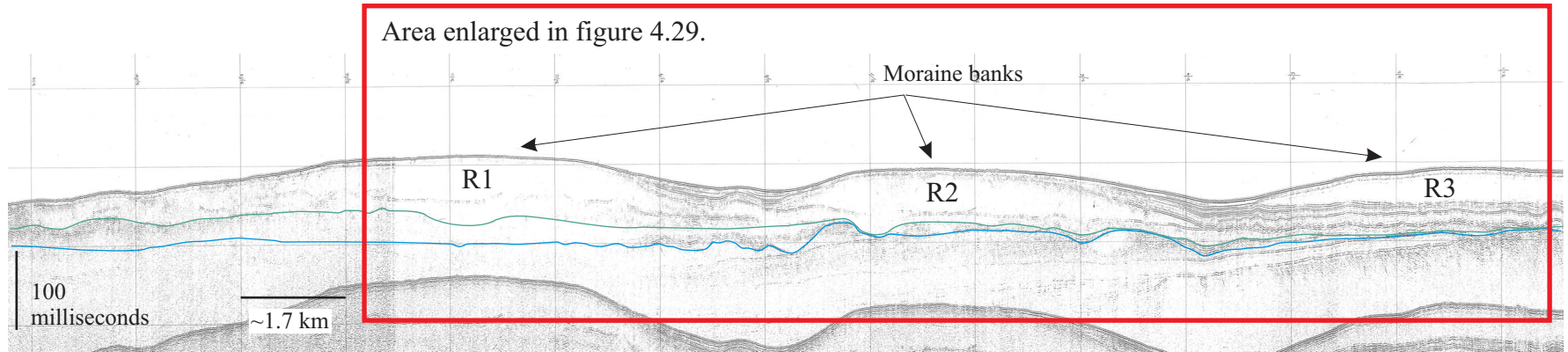


Figure 4.27. Sparker line 83/04-31 showing a transect across the outer shelf and uppermost slope.

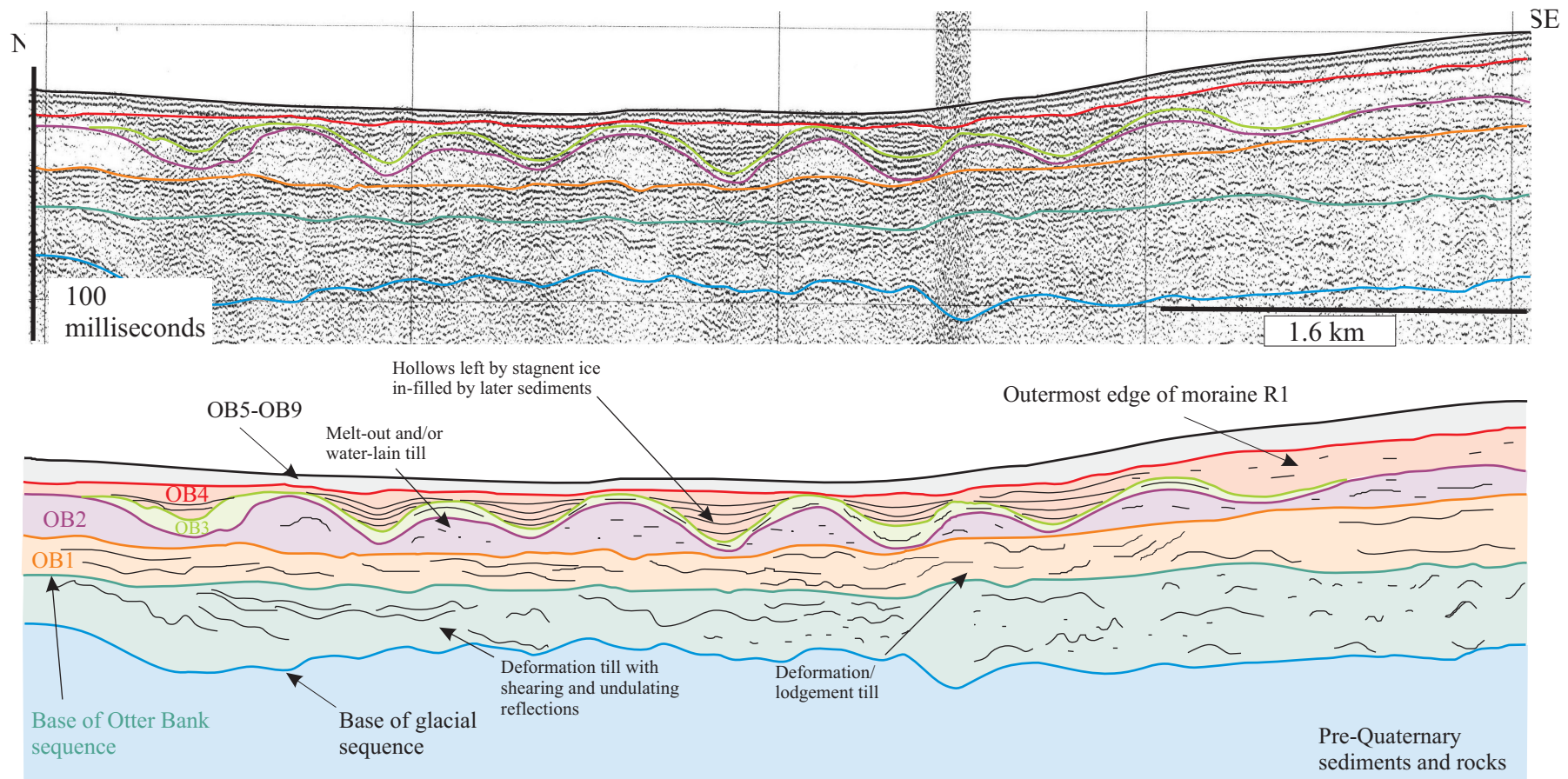


Figure 4.28. Expanded section of line 83/04-32 Showing the detail of Otter Bank units OB1-OB4.

The loss of stratification towards the shelf edge is due to the process of deposition by lodgement and deformation operating on the newly overridden sediments for only a short distance. The sediments deposited near the shelf are close to the maximum limit of ice advance and consequently have been overridden to a lesser degree and subjected to less deforming stress.

Package OB2 is a combination of chaotic reflections and reflection-free sections with occasional discontinuous higher amplitude events. It also shows a very variable thickness and an upper boundary with very prominent undulations. The lower boundary of OB2 in the southeast shows a degree of undulation where it rests directly upon the pre Otter Bank sequence (Figure 4.26a). Beneath the large moraine banks the whole unit becomes very thin and internal reflections suggest contortion and possibly shearing has occurred. Beyond the centre of the bank R1, the base of OB2 becomes much flatter with very little topography, forming a planer surface almost all the way to the shelf edge. In the outermost shelf zone the lower surface again becomes slightly undulating, partly reflecting the topography of the base of OB1 beneath.

The large undulations observed within OB2 northwest of the large moraine banks, shown in the expanded section in Figure 4.28, cause considerable changes in the unit thickness. The features are up to 750m from trough to trough and with a height of up to 15m. The height of the undulations is almost the entire thickness of the package. Other sections of OB2 do not show the undulating upper boundary, instead having a relatively flat upper surface. Sections lacking the undulations tend to more acoustically transparent than those where they are present.

Towards the shelf edge OB2 becomes much thinner and forms a series of low angled banks up to 7m thick in the last few kilometres before the shelf break (Figure 4.26c). The banks are generally too thin to reveal much internal structure although they appear to be more transparent than the thinner sections adjacent to them.

Unit OB2 is interpreted as an ice-proximal water-lain till deposited primarily by

melt-out and meltwater discharge, although this unit almost certainly includes an ice rafted component. The unit represents rapid melt-out of sediments during a retreat phase in which the grounded ice sheet receded across the shelf. The large undulations in OB2 are interpreted here as hollows formed by the break up of the ice front, leaving stranded stagnant masses of ice which decayed *in situ*. The apparent lack of coherent structure within this unit suggests that deposition, and therefore ice retreat was rapid.

Package OB3 is very thin and only shows in the enlarged section in Figure 4.28. The unit comprises sub-parallel stratified reflections with low continuity draped over the undulating upper boundary of OB2. OB3 is only identifiable as a drape within the undulating sections of OB2. In other sections it is either absent or too thin to resolve.

There are two possible interpretations for unit OB3, deposition from suspension or current re-working of existing sediments. Without additional data it is not possible to determine which of these is the more likely scenario, but both would take place in the outer proximal to distal zone during a retreat phase. Bottom current velocities of up to 6m/s have been reported from meltwater flows emanating from sub-glacial channels in Spitsbergen, which also created sediment laden plumes (Boulton, 1990). Velocities of this magnitude seem unlikely as there is no evidence of significant erosion, but winnowing of sediments on the seafloor is a possibility. Sedimentological data would theoretically show a coarser, sand-rich deposit if formed from current activity, and a mud-rich deposit if deposited from suspension.

Package OB4 forms the core of the outermost ridge (R1) of the large moraine banks of line 83/04-32 and line 83/04/31. The core of the ridge forms a large asymmetric lensoid body which is almost reflection free. Beyond the outer margin of the ridge, the acoustic character of the package changes to a well-stratified, slightly draped and on-lapping pattern which infills the undulations described previously in OB2. Towards the shelf edge OB4 becomes difficult to separate from other units but appears to form a thin drape over the preceding units.

Towards the mid-shelf in the southeast, OB4 becomes much thinner and is all but absent where it underlies ridge R2. Beneath the later ridges R3 and R4 the package changes character to a slightly convoluted stratified pattern of reflections and disappears from the last 8km of the survey line.

The change in seismic character within unit OB4 represents several different depositional regimes. The transparent facies forming the lensoid core of ridge R1 is interpreted as a proximal water-lain till deposit formed by melt-out with probable elements from meltwater discharge. The stratified section infilling the hollows in OB2 is interpreted as a combination of rain-out from suspension and ice rafting combined with a current deposited element from meltwater discharge. The presence of a current born component is deduced from the slight asymmetric onlapping present within some of the hollow infills and the lack of draping over the high points of the undulations of OB2.

Some of the packages in the sequence OB5 – OB9 are absent or difficult to define from line 32. Correlation of the lower units with line 83/04-31 (Figure 4.27), via line 83/04-23 shows that some of the units change character laterally across the shelf, and only by using both of these parallel survey lines is it possible to unravel the history of the upper units of the Otter Bank sequence. An expanded section of the moraine complex from line 83/04-31 is shown in Figure 4.29

Unit OB5 on line 83/04-31 comprises a thin package of well stratified parallel reflections draped over the south-eastern edge of ridge R1 and underlying the adjacent edge of R2. This package is only 4-5 ms thick, corresponding to approximately 3-4m of sediments. To the south east, OB5 is truncated by the overlying unit OB6 which forms the core of ridge R2. It seems likely that OB5 also covered the crest of ridge R1 and overlay OB4 in the area beyond the ridges to the northwest. On line 83/04-32 the unit is not identifiable as a single package, but probably forms part of the composite package of stratified reflections which overlies OB4 in the region of the moraine ridge R1.

The well-stratified, high continuity pattern of the reflections and the draping style suggest OB5 formed by rain out from suspension, probably combined with ice rafted debris. Such a scenario indicates a position beyond the ice front, but probably still within the proximal zone (as defined by Boulton, 1990).

OB6 on line 31 (Figure 4.29) forms an asymmetric lensoid unit nearly 6.5 km long and up to 45m (55ms) thick. The body is thicker and has a steeper slope at the north-western end than on the south-eastern end. Internally it is acoustically transparent with no apparent internal structure. The lower boundary overlies and truncates parts of OB4 and OB5, indicating a degree of erosion during the emplacement of OB6. Beyond the southeast margin of ridge R2, OB6 thins rapidly to no more than 5ms, and disappears completely before the end of the survey line.

The transparent nature of OB6 within the moraine ridge indicates a rapidly deposited chaotic sediment of a character formed by melt-out and ice-proximal glacial outwash. However, the symmetric shape of the ridge indicates there is at least an element of sub-glacial formation from the front of an advancing ice sheet. The very thin section of OB6 to the southeast of R2 also suggests a forward movement which has pushed most of the sediment into forming the moraine.

On line 32 the unit changes character between the ridge R2 and the end of the line to the south east. The moraine ridge R2, appears to be a composite of two merged moraines of very similar age. The majority of OB6 is made up of a transparent body forming the core of the moraine. However, where the two banks appear to merge there is a small lensoid body at the base of the unit which is composed of curved stratified reflections (Figure 4.26b). The section in Figure 4.26a at the south eastern end of the line is generally much thinner, and also shows discontinuous stratified reflections with a degree of convolution. This part of OB6, whilst being laterally equivalent to the moraine ridge, has obviously been formed by different processes. The character and shape of the reflection configuration suggest a level of deformation during the emplacement of OB6, and that this part of OB6 is a lodgement or a deformation till. The deformation till interpretation is more likely

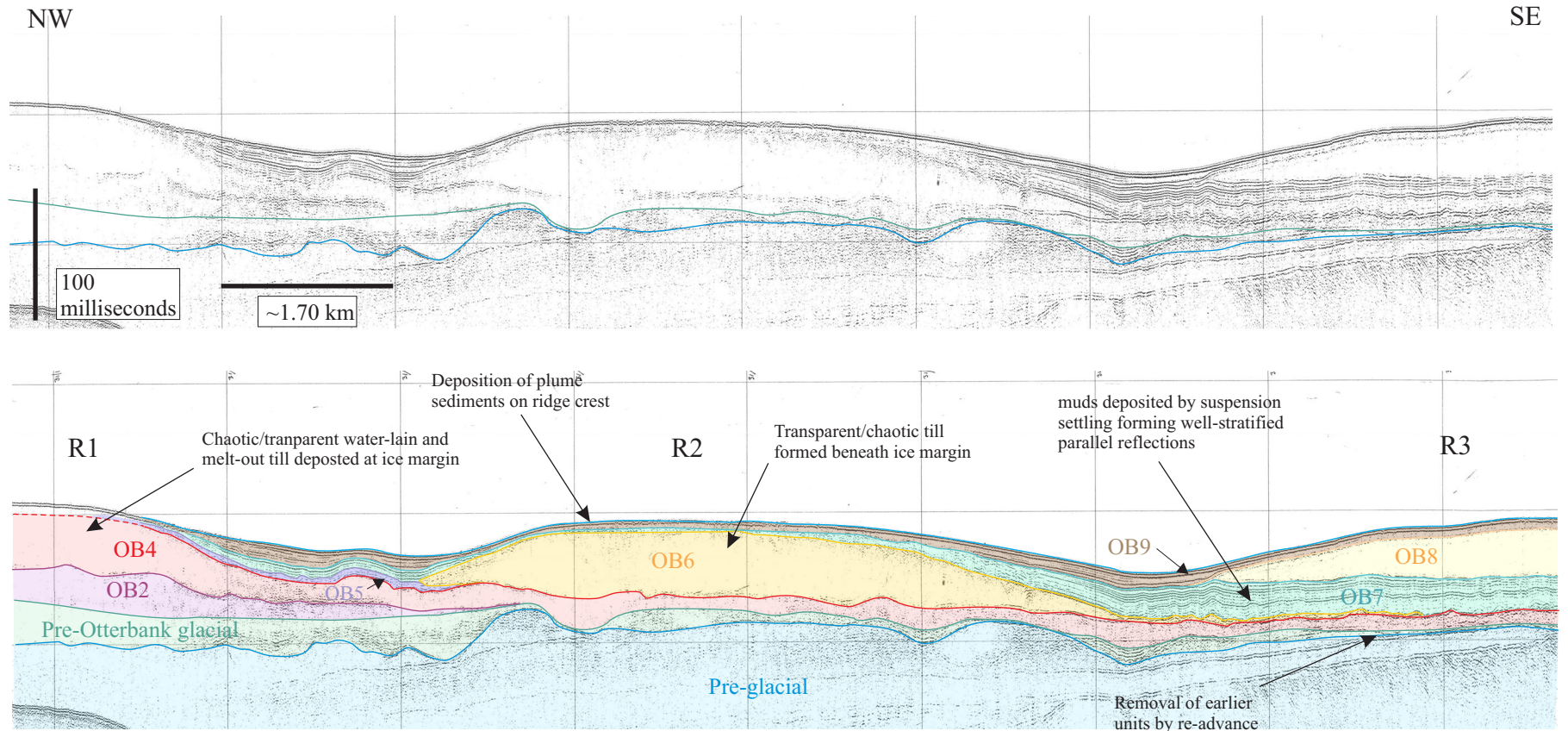


Figure 4.29. Enlarged centre section of the moraine complex from figure 4.27, showing the sub division of the Otterbank sequence into different seismo-stratigraphic units. The probable lithology and depositional processes responsible for each unit are described in the text.

given the irregular nature of the unit architecture. As such, this part of OB6 is probably derived from the deforming of underlying units which would explain the apparent absence of OB4 beneath much of this section.

On line 31, OB7 consists of a series of well-stratified draping reflections which overlie the south-eastern end of ridge R2, and probably infill part of the depression between ridges R1 and R2. It also seems likely that the reflections formerly overlay the crest of ridges R2 and R1 but are no longer present. Maximum thickness of the unit reaches 23m in the region overlain by moraine bank R3. On line 32, OB7 is not specifically identified, but almost certainly forms a component of the composite stratified sequence between R1 and R2 (Figure 4.26b).

The parallel-stratified continuous character of the reflections and the close spacing of individual reflectors suggests deposition was from suspension rain-out in a relatively calm setting, below the storm wave base. The most likely setting for this style of deposition is within the outer proximal zone (Boulton, 1990) where it would have probably been combined with an ice rafted component. Given the apparently calm setting in which OB7 was deposited, the unit is probably composed of sandy muds or muddy diamictons.

Data from line 31 shows unit OB8 forms the core of ridge R3 and rests conformably upon the underlying stratified package of OB7 (Figure 4.29). Acoustically, the unit is identical to the other ridges in that there are no internal reflections. The south-eastern end of this ridge is beyond the end of the survey line, but maximum thickness appears to be in the region of 30m. The shape of the unit is very similar to that of unit OB4 in that it has a longer, lower angled slope at the north-western end. It is assumed that the slope to the south-east is shorter and steeper. This section is interpreted as a melt-out, water lain deposit formed immediately beyond the ice front in the same way as R1. The absence of erosion beneath this section of OB8 indicates there was no overriding by ice during its deposition.

On line 32 the nature of OB8 is more complex, forming the large moraine banks R3

and R4 (Figure 4.26a), and a series of smaller merged banks with some internal structure. Banks R3 and R4 are mostly reflection-free and similar to the structure seen in line 31. However, on line 32 banks R3 and R4 show signs of deformation and erosion beneath them and OB7 is absent. The margin between R3 and R4 (Figure 4.26a) shows a mound of OB8 with poorly developed stratified reflections forming the base of the unit. There is also some variation in the main body of OB8 which suggests R4 maybe a composite of different events.

The poorly stratified section of OB8 is interpreted as a deformation till, formed during a minor re-advance which deposited bank R3 beyond the ice front. It is also possible that some of the deformation structures and stratification observed in OB6 beneath may have formed at this time. Bank R4 appears to be primarily melt-out water lain deposit formed at the ice front. South-eastwards of R4 OB8 becomes slightly more structured towards the top of the unit. There is also evidence of small hollows of the style seen in OB2 and interpreted as detached stagnant ice features (Figure 4.26a).

The uppermost unit in the moraine complex is OB9. This forms a thin series of well-stratified continuous reflections which lie in the hollows between the moraine banks on line 31 (Figure 4.29). The reflections drape the upper surface of the underlying units, reaching a maximum thickness of about 10m.

On line 32, the character of OB9 is different, comprising a thin sequence of transparent sections and poorly stratified reflections. This includes stratified infills of the small hollows at the extreme south-western end of the line (Figure 4.26a). In addition, there is a thin section of undifferentiated stratified reflections between banks R1 and R2 which probably contains a component of OB9.

The stratified section is probably a lateral equivalent of the stratified unit seen on line 31 and represents water lain deposits formed in a calm setting beyond the ice front. The more chaotic section in the south east is interpreted as the result of a minor re-

advance, infilling the hollows and locally depositing a thin, ice-proximal till by melt-out and meltwater discharge.

4.3.5 Moraine complex core data:

Unfortunately, core data from the area of the moraine complex is very restricted, being limited to a single borehole 77/09 which had poor core recovery and is poorly preserved. There are no cores associated with the 1983 seismic survey data.

A number of shallow gravity cores from areas between the ridge crests have indicated the acoustically well-stratified units of OB9 are composed of sandy mud facies sediments (Stoker et al, 1993).

Borehole 77/09 lies sparker survey line 74/4-13. Unfortunately the quality of data from this survey is such that only limited information can be extracted from it. The relevant section of the survey line is shown in Figure 4.30. Whilst the survey data does not show much sub-surface detail, there is sufficient to identify the presence of a moraine ridge and a more acoustically transparent core section similar to those seen in the 1983 survey data. As such, it is useful in trying to characterise the sediments forming the moraine ridges as no other data is available. The summarised core log and description presented in Figure 4.31 are based upon data from the original shipboard descriptions rather than re-examination. This is due to the severe degradation and alteration which the core has undergone since drilling.

Interpretation of 77/09

The thin sand and gravel deposit from 0 – 1.5m is interpreted as a lag deposit, formed by current re-working of the underlying sandy muds and pebbles since the end of glacial input. This is a feature common to most of the cores recovered from boreholes on the shelf.

Red-brown silty muds with pebbles from between 1.5m and 16.50m apparently show no internal structure and are most likely to be similar to the sandy mud facies. This would imply deposition from suspension with some IRD input. This interpretation is supported by a low pebble content and generally soft consistency.

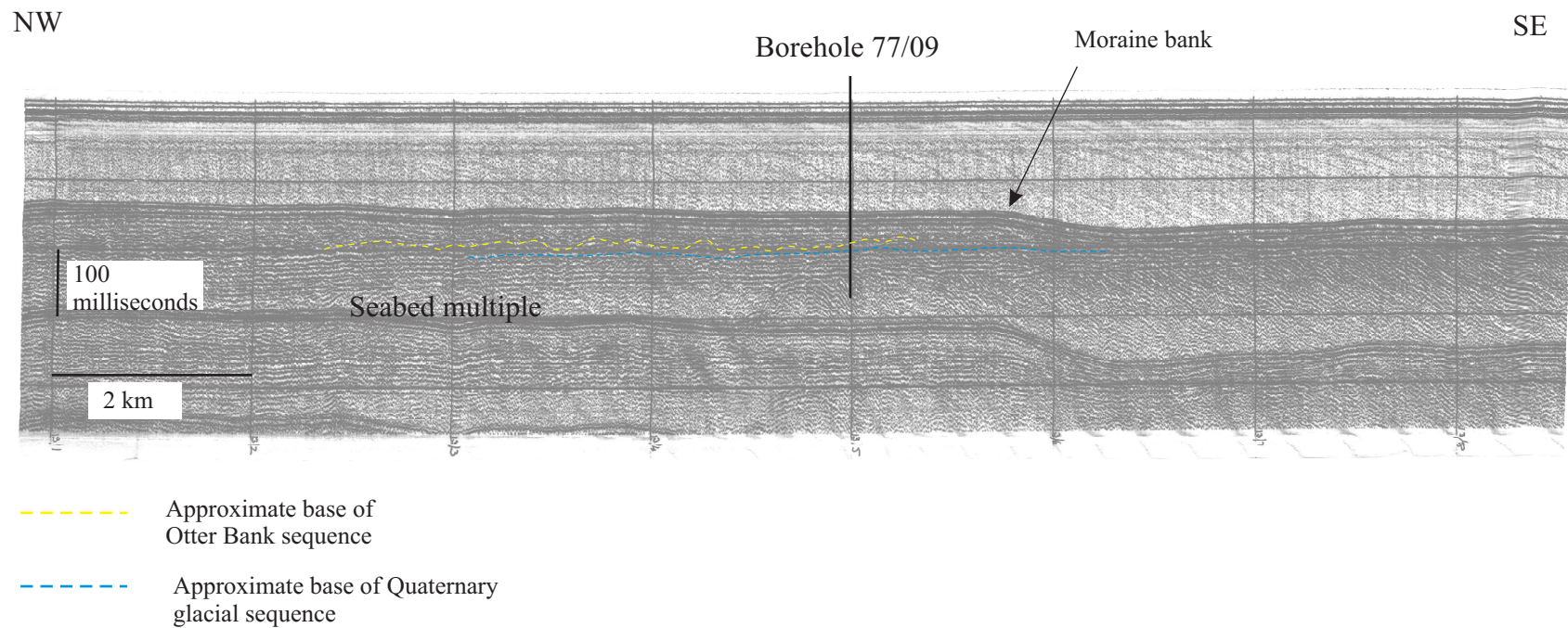


Figure 4.30. Seismic data from line 74/04-13 in the area around borehole 77/09. Poor data quality limits the degree of interpretation and correlation with the borehole core.

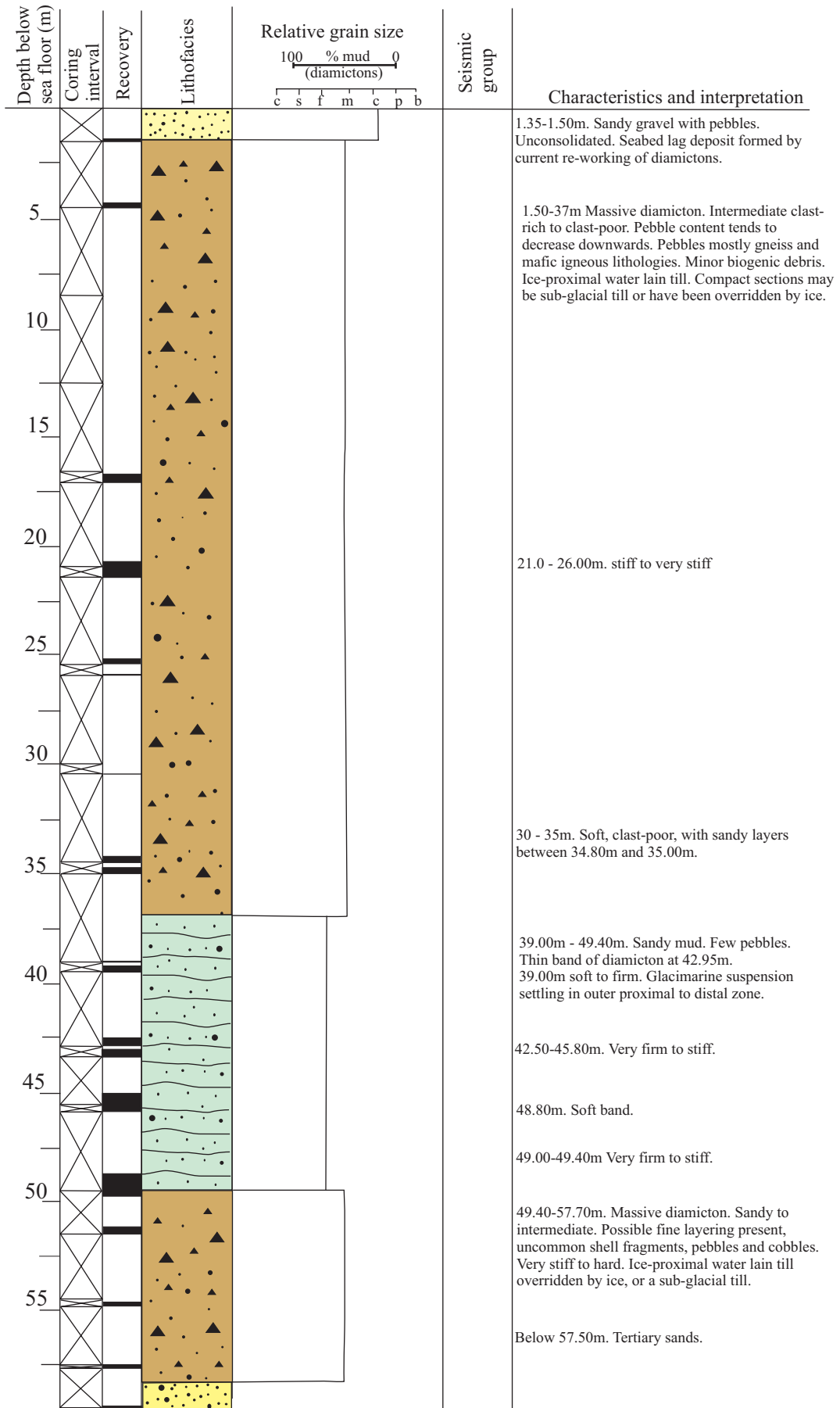


Figure 4.31. Core log of borehole 77/09

Between 16.50m and 31.00m coring difficulties indicated a compact unit of massive red-brown diamicton. The apparent hardness of this unit would suggest that this lithology has been over-compacted, most likely by overriding ice.

Below 31.00m, down to approximately 39.00m the sediments change to a softer red-brown muddy sand with some sandy layers. Interpretation of this unit is difficult as the layering may have been lamination caused by deposition or deformation but due to disintegration of the samples and lack of original description it is impossible to be more precise.

The lower part of the glacial section from 39.00m – 49.40m is composed of stiff grey-green sandy mud with shell fragments probably of the sandy mud facies. Lack of pebbles and the presence of shell debris suggests an origin by suspension settling, although the hardness indicates later compaction.

Below 49.40m the core passes into sand with the probable base of the Quaternary section at about 57.00m. The upper part of the sand unit has a greater mud content and muddy layers with cobbles up to 10cm. It is also reported as being very hard, suggesting overcompaction. The sand is interpreted as a current deposited sediment resulting from re-working of earlier sediments. However, the overcompaction and the layering seen in the upper sections may indicate there is an element of deformation till caused by overriding ice. Formation of a sandy deformation till is a likely scenario at the onset of a glacial advance over a previously unglaciated seabed.

4.3.6 Interpretation of the moraine complex:

The data from the core 77/09 and the previously published gravity cores shows the stratified units are composed of sandy muds with occasional pebbles whilst the seismically transparent units are composed of diamictons which are probably massive. Micropalaeontological data suggests these were deposited in a relatively shallow water in the region of 50m (Stoker et al, 1993). This figure is very approximate, although the deposition and preservation of the stratified muddy units also suggests a water depth greater than 30m (Boulton, 1990).

Unit OB1 at the base of the Otter Bank sequence in area Cii is interpreted as a lodgement till with a probable component of deformation till, formed beneath an ice sheet during the initial advance across the shelf (Figure 4.32a)..

Development of OB1 appears to be more widespread in areas underlain by thicker glacial sediments of earlier glaciations which are unlithified. This could indicate preferential development of OB1 in areas with a thicker layer of pre-existing deformable sediment, or merely that the Otter Bank ice sheet removed the earlier sediments in some areas during its advance across the shelf. However, the areas of thickest development also correspond to the areas of the shelf with the deepest basal Otter Bank unconformity, indicating infilling of pre-existing topography.

The thinning of OB1 at the north-western edge of the intra-shelf basin (Figure 26a) indicates the lodgement/deformation till did not develop as readily in the shallower areas and suggests a degree of buoyancy is needed to allow for the deposition to occur. This is supported in part by the absence of a deformation till in the shallow waters of the inner shelf described in section A of this chapter. It also accounts for the very thin section of OB1 beyond the small basin towards the shelf edge where the basal unconformity rises, effectively placing a greater pressure upon the base of the advancing ice sheet and preventing formation of a thicker till layer.

Following deposition of OB1, the ice sheet then retreated back across the shelf to a position at least as far back as the moraine bank R2, probably depositing a melt-out till in the process. This assumed melt-out deposit is not preserved, having been completely removed by a subsequent re-advance.

OB2 is interpreted as an ice-proximal water-lain till, deposited by the ice sheet during the retreat phase as the ice sheet receded back across the shelf after a re-advance to the shelf edge. The re-advance formed the planer base to OB2, stripping off the unconsolidated melt-out till formed during the previous retreat but leaving the more consolidated lodgement till of the OB1 advance.

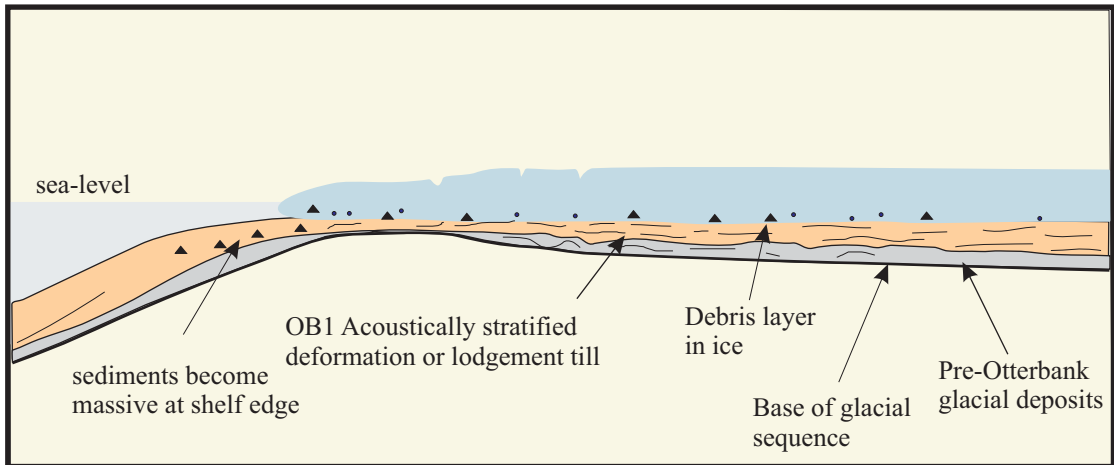


Figure 4.32a. Formation of unit OB1 during ice sheet advance to the shelf edge. The advance stage till is deposited by a process of lodgement and deformation, incorporating earlier pre-Otterbank sediments. Ice then retreats to an unknown position, probably depositing melt-out and water-lain tills during this process.

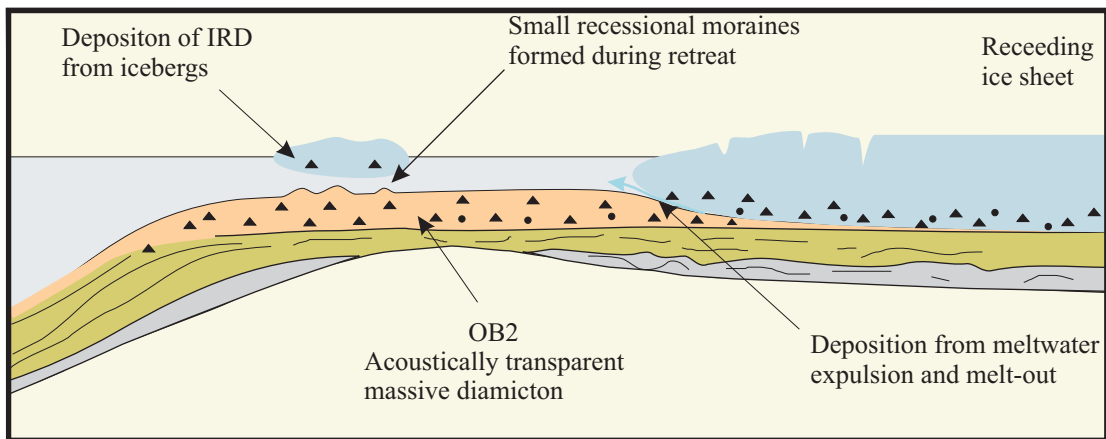


Figure 4.32b. Re-advance to the shelf edge re-works the retreat deposits from the first advance and deforms the remaining tills of OB1. Formation of unit OB2 by melt-out, outwash and plume settling during a second ice retreat. Small recessional moraines near the shelf edge form during pauses in the retreat.

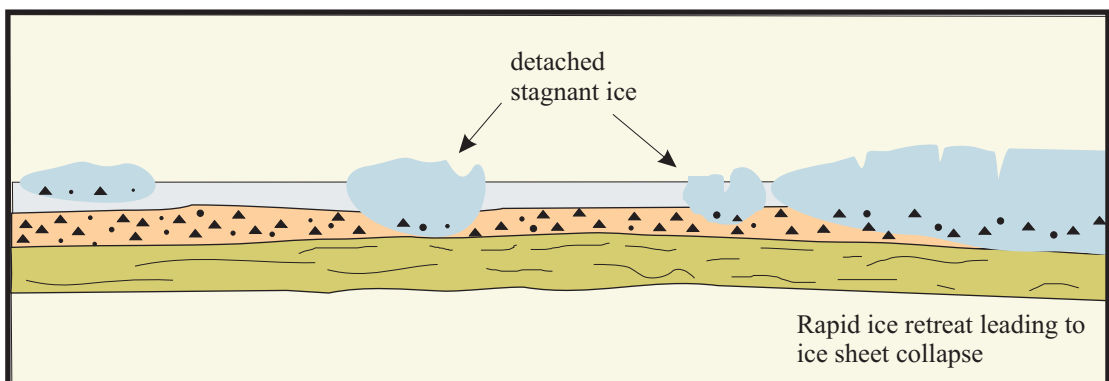


Figure 4.32c. Ice retreat increases as the ice recedes into deeper water of intra-shelf basin, leading to break-up of the ice sheet. Detachment of ice forms stagnant ice bodies and the formation of the hollows seen in OB2. Deposition of diamictons continues by melt-out, meltwater plumes and ice rafting.

The absence of a lodgement/deformation till recording the second advance to the shelf edge suggests a lack of accommodation space relative to the buoyancy of the ice sheet. If sea levels were very similar during the second advance as they were in the first advance, available accommodation space would have already been infilled by deposition during the first advance. Without sea level rise, subsidence or erosion, further deposition would be considerably hindered. Consequently, the material deposited from the first retreat phase and eroded in the second advance phase would have been transported to the maximum limit of the ice sheet advance. In this case, that limit is the shelf edge, where those sediments form the upper section of the Otter Bank sequence on the slope (see Area 4D).

Deposition of OB2 occurred on the second retreat from the shelf edge, forming small moraine banks close to the shelf edge early in the retreat phase where the retreating ice front paused. The structureless nature of the moraine banks indicates formation by re-advancing ice is very unlikely and deposition was from melt-out and meltwater discharge (Figure 4.32b).

The interpretation of the hollows within OB2 as the result of stagnant ice becoming detached from the ice front and decaying *in situ* (Figure 4.32c) would be consistent with rapid retreat of the main ice sheet body. This interpretation is further supported by an examination of the relative water depth. The slope of the lower boundary of OB2 dips towards the inner shelf indicates that the retreating ice sheet would have receded into deeper water. This would have removed some of the support to the ice sheet by increasing buoyancy and increased the likelihood of ice sheet break up. This style of decay has been suggested in marine terminating glaciers (Mayo, 1988; Powell, 1991), and especially in relation to the Antarctic ice sheets.

The stratified OB3 unit is difficult to interpret with any certainty due to its small thickness and absence of lithological data. The two interpretation scenarios involve re-working of the top of OB2 by currents following ice retreat, or suspension settling of fine-grained glacial marine sediment in an ice-distal setting (Figure 4.32d).

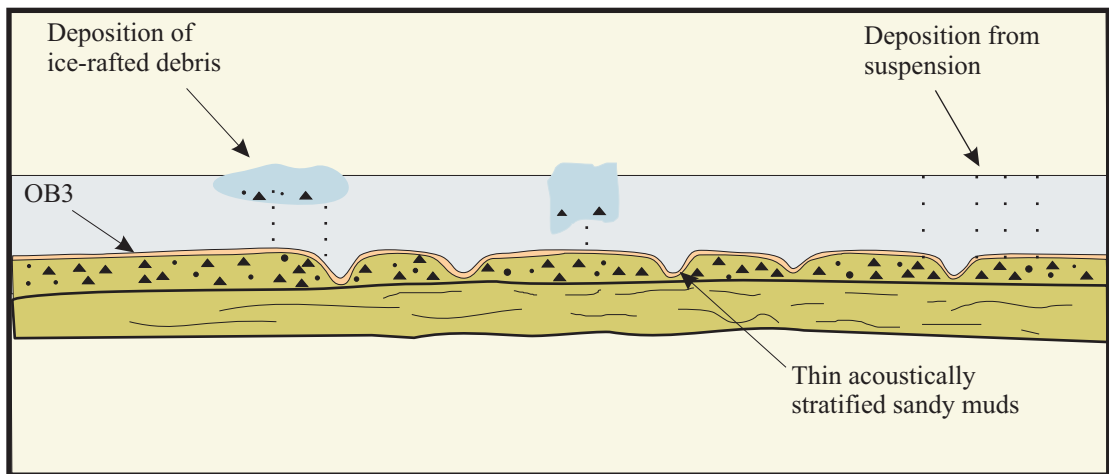


Figure 4.32d. Formation of OB3 by suspension settling and iceberg rafting forming a draped sheet-like deposit, following the retreat of the ice to the inner/mid-shelf. Hollows formed by *in situ* melting of detached stagnant ice or large grounded bergs left during ice-sheet break up.

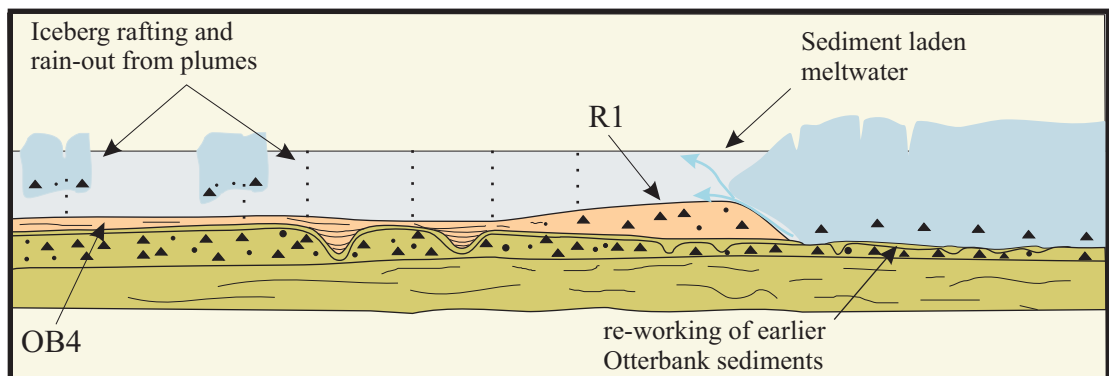


Figure 4.32e. Formation of ice-proximal OB4 and moraine ridge R1 at the ice front by sediment laden meltwater plumes and melt-out. The ridge marks the maximum extent of the ice front re-advance where the ice halts on the outer edge of the intra-shelf basin.

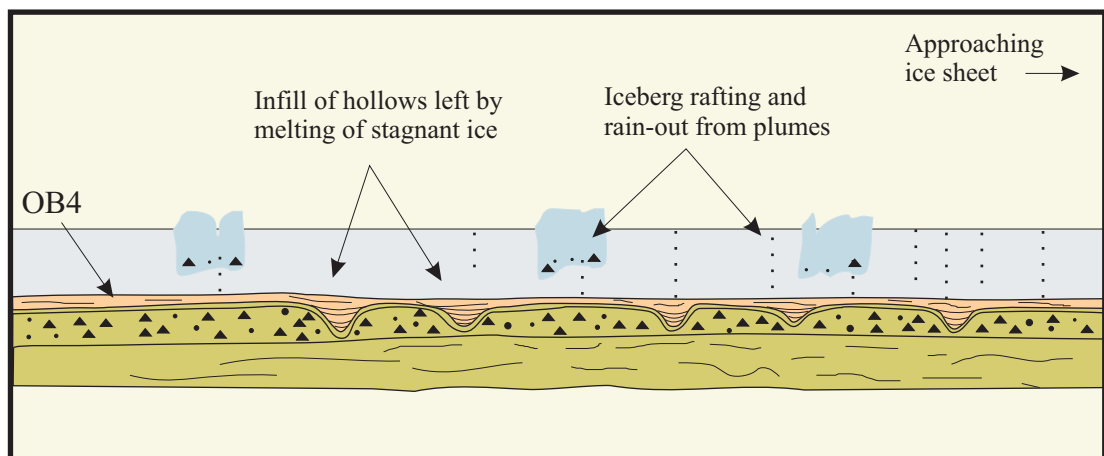


Figure 4.32f. Deposition of the stratified section of OB4 by suspension settling of meltwater plume sediments and IRD during a re-advance phase of the ice sheet. The advancing ice sheet lies beyond the edge of the figure to the right and is continued below.

The interpretation of the two different seismic facies of OB4 is important as it represents a re-advance of the ice. The asymmetric shape of the ridge R1 (Figure 4.27) provides important information with regards to its mode of formation. The core of the ridge formed by the acoustically transparent OB4 unit has a long, low angled slope towards the shelf edge and a short steeper slope towards the former position of the ice front (Figure 4.32e). This is characteristic of a moraine bank formed beyond the ice front by melt-out and meltwater discharge (Powell and Domack, 1995). Had the bank been formed by sub-glacial processes from a surging forwards, the geometry would be the reverse of this with the long slope beneath the ice front (Figure 4.33).

Beyond the north-western end of the ridge, the acoustic character changes to an infilling stratified pattern. Formation of these deposits from suspended sediments in meltwater plumes and by ice-berg rafting within the outer proximal zone following the decay of large detached icebergs. This occurred at the same time as the formation of the moraine bank and would be expected to be formed from finer-grained sediments than those comprising the main body of the ridge (Figure 4.32f).

OB5 is a parallel stratified unit deposited after the ice-front retreated from its position at the R1 ridge, and is probably the contemporaneous distal equivalent of ice-proximal sediments forming further back on the shelf. Sedimentologically it is likely to be similar to OB3 and the infills of OB4. The south-eastern margin of OB5 has been truncated by the edge of the overlying R2 ridge indicating OB5 probably extended further towards the inner shelf (Figure 4.32g).

Unit OB6 forms the main body of moraine bank R2, truncating and removing several underlying units during its formation. This moraine ridge has a long, low angled slope beneath the position where the ice front would have sat, indicating formation by a probable surging and lodgement processes beneath the ice (Powell and Domack, 1995). In addition, the surging also formed a new deformation till in the zone behind the ice front (Figure 4.32h). Although there is no data available, it would be reasonable to expect OB6 to have a much greater level of compaction than the

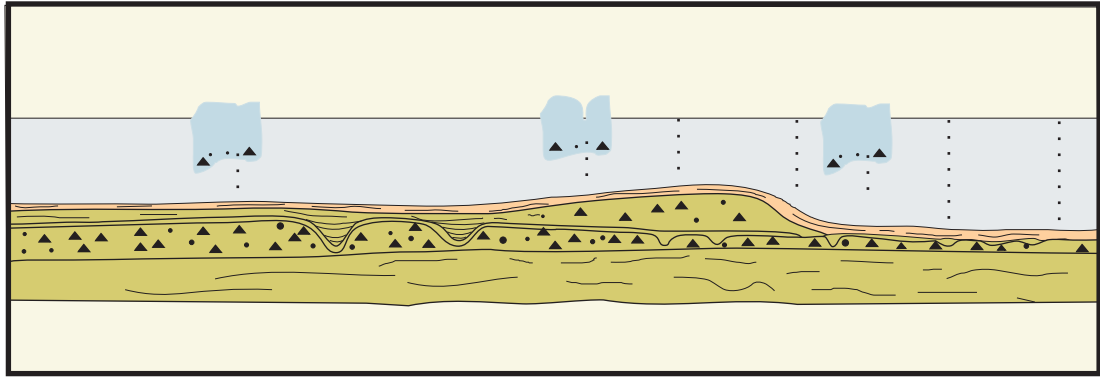


Figure 4.32g. Formation of OB5 by suspension settling from meltwater plumes and ice rafting with the ice front having retreated to across the deeper water of the basin to a more landward position. Deposition of OB5 over the crest of the moraine ridge R1 is speculative.

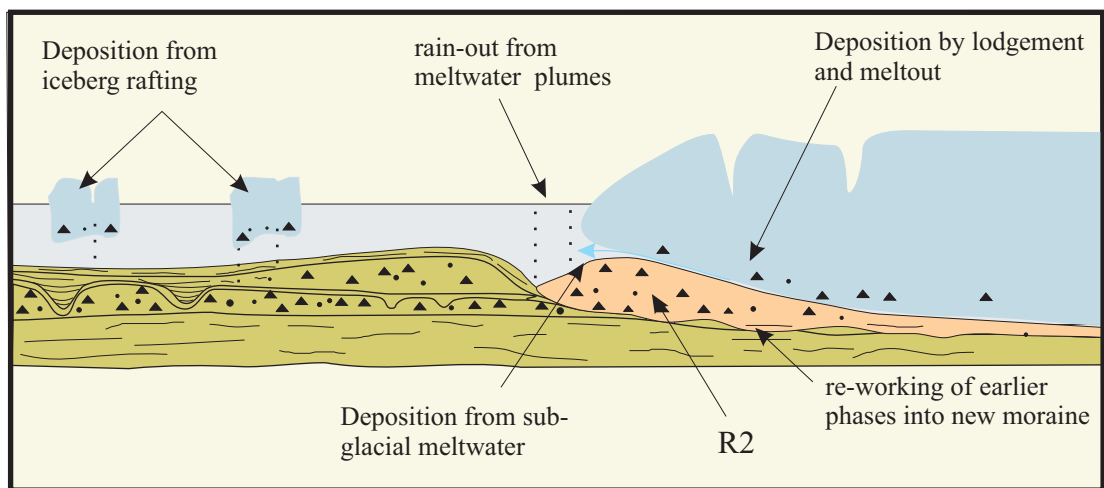


Figure 4.32h. Deposition of OB6 in a sub-glacial setting beneath the ice-front following a re-advance of the ice sheet. The re-advance over previous Otter Bank units results in the formation of a new deformation till by re-working the older sediments and forms the moraine ridge R2.

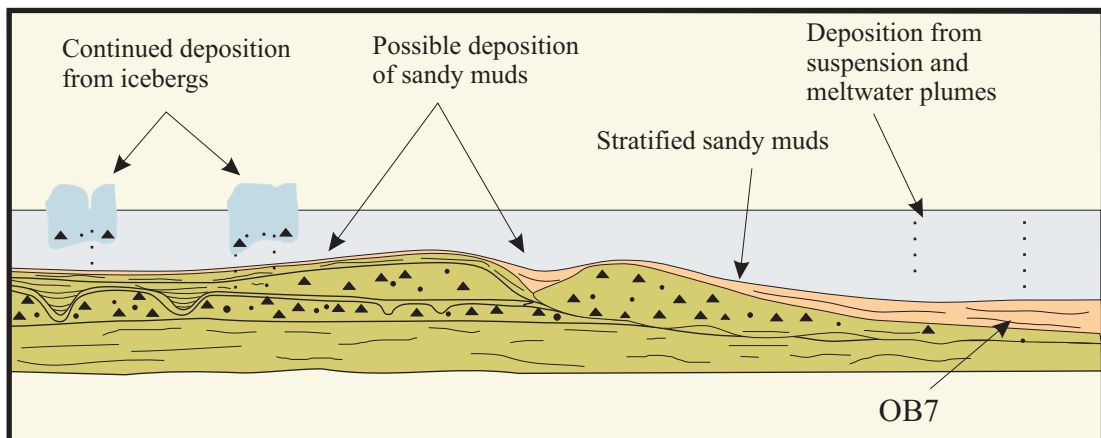


Figure 4.32i. Deposition of the acoustically stratified OB7 unit with the ice front in a more distal position than in h. OB7 is probably a combination of plume and suspension settling deposits formed during the retreat from stage h, and during a further re-advance prior to the deposition of OB8 in stage j.

sediments of bank R1, having been formed beneath the ice rather than in front of it.

Unit OB7 is probably composed of muddy diamicton and sandy muds, formed as ice-distal deposits during the retreat phase following the formation of ridge R2. There is also likely to be an ice-distal advance phase deposited as rain-out sediment during the re-advance prior to the deposition of the moraine bank R3 (Figure 4.32i).

The apparent absence of disturbance to OB7 by the emplacement of the overlying moraine bank is important, as it indicates the ice did not override this unit.

OB8 forms the moraine bank R3 and shows the same lack of internal structure as moraines R1 and R2. The near-shore end of the bank is not seen, but it is assumed to be steeper than the north-west end seen in the survey line. This suggests that it has the same profile, and the same mode of formation as R1, being deposited by melt-out and meltwater discharge beyond the ice front. The lack of disruption to OB7 beneath indicates the bank R3 marks the maximum re-advance at this point (Figure 4.32j).

The difference in character of OB8 on line 32, indicates a different mode of formation for the moraine banks R3 and R4. The presence of a deformation layer and crudely stratified lodgement till suggests a probable forward surge in the ice front. These indicators show that whilst the ice was passively depositing the moraine bank seen on line 31 by melting processes, the ice in the region of line 32 was undergoing a minor surge.

The uppermost unit, OB9 represents the final ice retreat from this area (Figure 4.32k). It is likely that OB9 was more extensive but has been re-worked following deglaciation. This is particularly likely on the crests of the ridges which may have been exposed to current re-working early in the deglaciation prior to major sea-level rise.

4.3.7 Characterisation and interpretation of the outer shelf

The two zones described in area C show that for the Otter Bank sequence there appears to be three main phases of glacial activity. The first two phases are present in

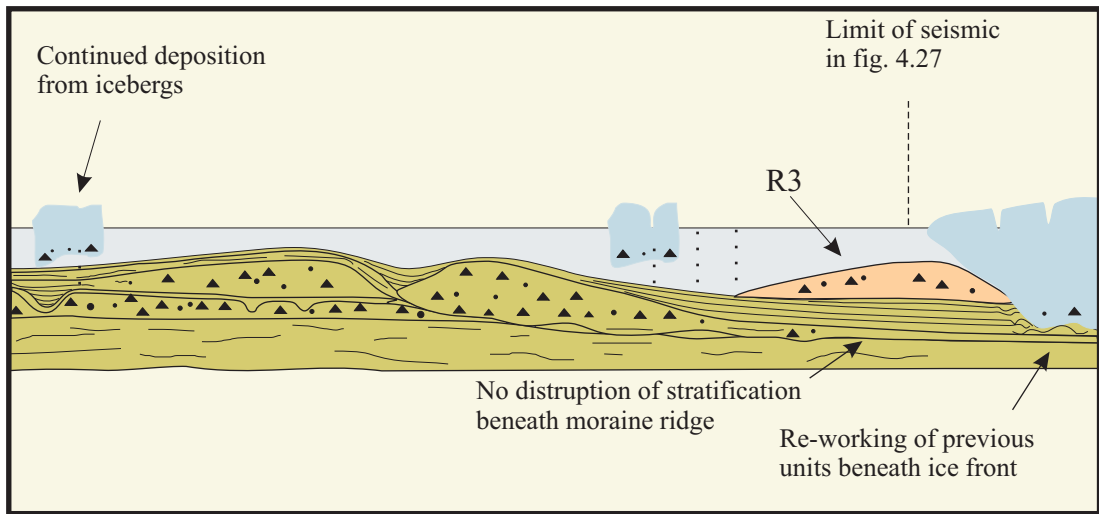


Figure 4.32j. Deposition of moraine ridge R3 beyond a re-advanced ice front. Previous Otter Bank sediments beneath the moraine show no indication of disruption, indicating deposition of R3 occurred beyond the ice front. Sediments beneath the ice are re-worked into another deformation till, although this is not seen in the seismic data.

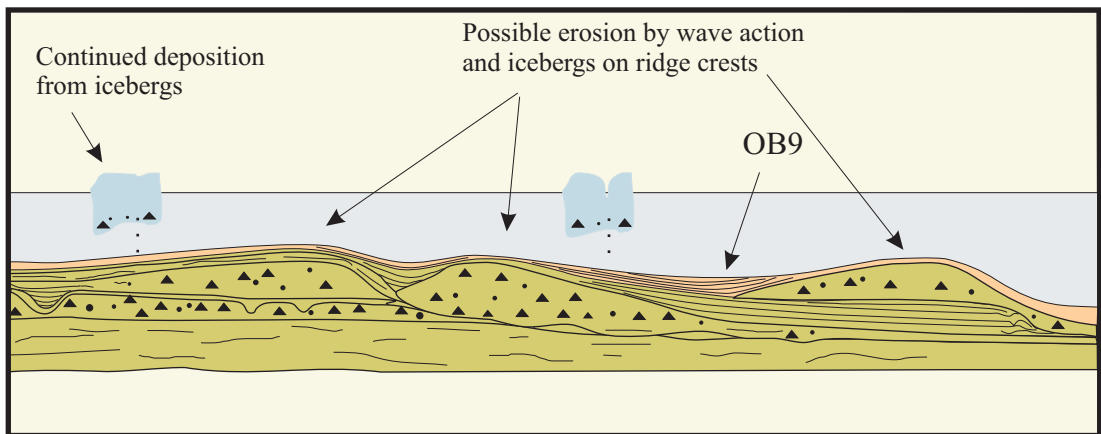


Figure 4.32k. Deposition of stratified unit OB9 by suspension settling, iceberg rafting and possible distal meltwater plume activity during final ice sheet retreat. Survey data suggests subsequent removal of sediments from the crests of the moraine ridges.



A) Sub-aqueous outwash . Steep slope of the moraine bank lies against the ice margin with bank formed beyond the ice.



B) Melt-out and ice-berg dumping. Rapid melting and dumping deposits sediment close to the ice front with the bulk of the moraine formed beyond the ice.



C) Lodgement/ramp-type moraines. Deposited by accretion at the ice front. Steepest slope lies away from the ice margin with the bulk of the moraine lying beneath the ice.



D) Push and squeeze moraines.

Figure 4.33. Various types of moraine bank formed at grounding lines. These can be considered as end-members of a grounding line system where most moraine banks are a combination of processes (based on Powell and Domack, 1995).

both areas Ci and Cii, forming two main acoustic packages on the outermost shelf, although the quality of the seismic data makes separation of the two in area Ci difficult. In zone Cii the two packages are the acoustic units OB1 and OB2, which represent two ice advances to the shelf edge. Whilst it is not possible to extend the acoustic packages of OB1 and OB2 to zone Ci with any certainty, the two units are interpreted as forming the bulk of the diamictos observed in cores 82/10 and 82/11.

This suggests an initial ice advance across the whole of area C, overriding and eroding the glacial sediments of the pre-Late Devensian glaciations. This first advance deposited the lodgement and deformation tills of OB1, which are represented by the higher shear strength diamictos in core 82/10, and by the debris flow diamictos and very hard tills in 82/11. The diamictos in 82/11 at the base of the Otter Bank sequence are initially softer as the basal units formed debris flows beyond the shelf edge, before being overridden by the advancing ice.

A comparison of the Otter Bank sequence in cores 82/10 and 82/11 shows a difference in the level of compaction between the two sites, the most compact being in 82/11. If the compaction is due to ice loading as suggested, the most compact sediments would be expected further back on the shelf where ice would have been present for a longer period. The apparent discrepancy in compaction between the two sites is explained by the palaeobathymetry of the outer shelf. By stripping off the Otter Bank sediments from the modern seabed, the basal surface can be plotted. The plot reveals the site of 82/10 lies within an elongate basin approximately 20 metres below the outermost shelf. This indicates deeper water and the possibility of increased buoyancy of the ice sheet, leading to less compaction of underlying sediments than would occur in the shallower area at the shelf edge. The basin is a north-eastwards extension of that described in the bathymetry of the outer shelf (section 4.6). A complete palaeosurface map of the base of the Otter Bank is presented in Chapter 6.

The retreat of the ice from the shelf edge following the initial advance almost certainly deposited melt-out and water lain tills, combined with an IRD component.

This deposit is now absent on the shelf, unless it has been modified by a later advance and actually forms part of the deformation till assigned to unit OB1.

The second ice advance to the shelf edge also occurred across the whole of area C, planing off the unconsolidated melt-out tills of the first phase retreat but not removing the more compact lodgement and deformation tills from the first advance. The second phase advance may have caused further deformation in the underlying OB1 unit, but does not appear to have formed a new lodgement or deformation till. The combination of these factors suggests the time period between the first phase retreat and the second phase advance was short, with no time to create accommodation space through subsidence or erosion which might allow deposition of new lodgement tills on the shelf. It may also be that the deposits from the first phase provided a lower friction or deformable surface over which the second ice advance could more easily pass.

The second phase advance transported much of the sediment from the first phase retreat to the shelf edge where it forms debris flows on the upper slope, in addition to the sediment delivered from the second phase ice. The second retreat from the shelf edge formed the OB2 unit, primarily as a melt-out and meltwater discharge deposit. Short-lived pauses in the retreat are responsible for forming the moraine banks near the shelf edge, including those close to the site of borehole 82/11 (Figure 4.20).

As the retreat progressed, the ice in the region northwest of basin A (Figure 4.16) broke up, leaving stranded bodies of ice detached from the ice sheet. Deposition by melt-out and meltwater discharge continued around these stranded blocks, which on melting left hollows in the OB2 unit. In the region of basin A the increased water depth led to rapid ice retreat due to increased buoyancy and lack of support. This may have been a factor in the break up responsible for the stranded block formation.

The distribution of OB2 indicates that the ice retreated some distance across the shelf, beyond the eastern (landward) margin of area C. During the retreat, but in a more ice-distal position, IRD and deposition from suspension formed unit OB3.

A third phase of ice advance resulted in the removal of large volumes of earlier Otter Bank sediments from area Cii. This advance terminated in the deposition of the outermost of the large moraine ridges, R1 which marks the maximum extent of the re-advance. The position of the moraine ridge marks the north-westerly edge of basin A, suggesting the ice was semi-buoyant whilst crossing the basin but grounded and halted where the water depth decreased. Alternatively it may simply have halted due to the large moraine bank built up in front of it preventing further forward movement. Following the deposition of the ridge R1, the ice within area Cii underwent a series of retreats and minor re-advances which formed the remaining banks of the moraine complex. Coupled with the oscillation of the ice front is the development of thin lodgement and deformation tills formed during advance phases. The lack of lateral persistence in some of the moraine banks suggests the ice may have formed surging lobes which developed locally.

The re-advance which formed large moraine banks R1-4 does not appear to have crossed zone Ci as there are no moraine banks attributable to later phases in Ci. However, the core data from 82/10 shows a thin diamicton close to the top of the sequence which could have been deposited from icebergs related to the re-advance. Likewise, the sandy muds at the top of 82/11 could be a more distal deposit from suspension settling related to the re-advance.

The absence of a later ice sheet in Ci suggests that the Cii section of area C formed a preferred pathway for ice sheet drainage and sediment deposition across the outer shelf, especially in the later stages of the glaciation. This is supported by data from the upper and lower slopes presented in sections D and E of this chapter which show a focusing of deposition on the slope to the southwest nearer Cii. The focusing of the ice drainage and deposition is examined in detail in Chapter 6.

The final stage of the Otter Bank glacial phases is the deposition of acoustically stratified muds, which represent the outer proximal to distal deposits following ice retreat to the inner shelf.

4.4 Area D: Upper and Middle slope

Area D marks the transition from the outer shelf, where deposition has been dominated directly by ice-related processes, to the upper slope where deposition has been in a more ice-distal setting and greatly affected by re-mobilisation (Figure 4.34). Lithological interpretation within area D is restricted by the absence of boreholes. However, the whole of area D is included in the coverage by the seabed image which reveals important details of large scale architecture.

4.4.1. Sea-bed morphology and seismic data

Bathymetry

The bathymetry data from area D marks a change in the resolution of the available information from a 10m interval of the BGS Digbath (to 250m) to a 50m interval of the JEBCO dataset beyond the shelf edge. As a result, only the broad scale features of the seafloor are evident. The survey shows a generally even slope below a depth of 200m with an angle of between 1° and 2°. The only major feature along the shelf-slope transition is the Foula Bight (Figure 3.1). This is plateau-like area northwest of the Papa Basin lying at a depth between 200m and 250m. The significance of the Foula Bight is dealt with at the end of this chapter.

Seabed Image data

The seabed image data within area D shows a variety of features from small elongate mounds and gas escape structures to large flow lobes and incised troughs (Figures 4.35 and 4.36). At the top of the slope there is a transition from small moraine banks of the shelf edge to a smoother seabed with rounded, low angle features and crater-like hollows. In the north-east there is a large area of smooth seabed which forms an arc shaped bulge which curves into the south-western end of the Foula Bight. Close examination of this region appears to show a series of small terraces up to several kilometres long at a depth between 250m and 350m (Figure 4.36). The terraces resemble small-scale slump features seen on sub-aerial slopes and also have similarities with rotational slides from glacimarine settings. This similarity is expanded upon in the seismic data section below.

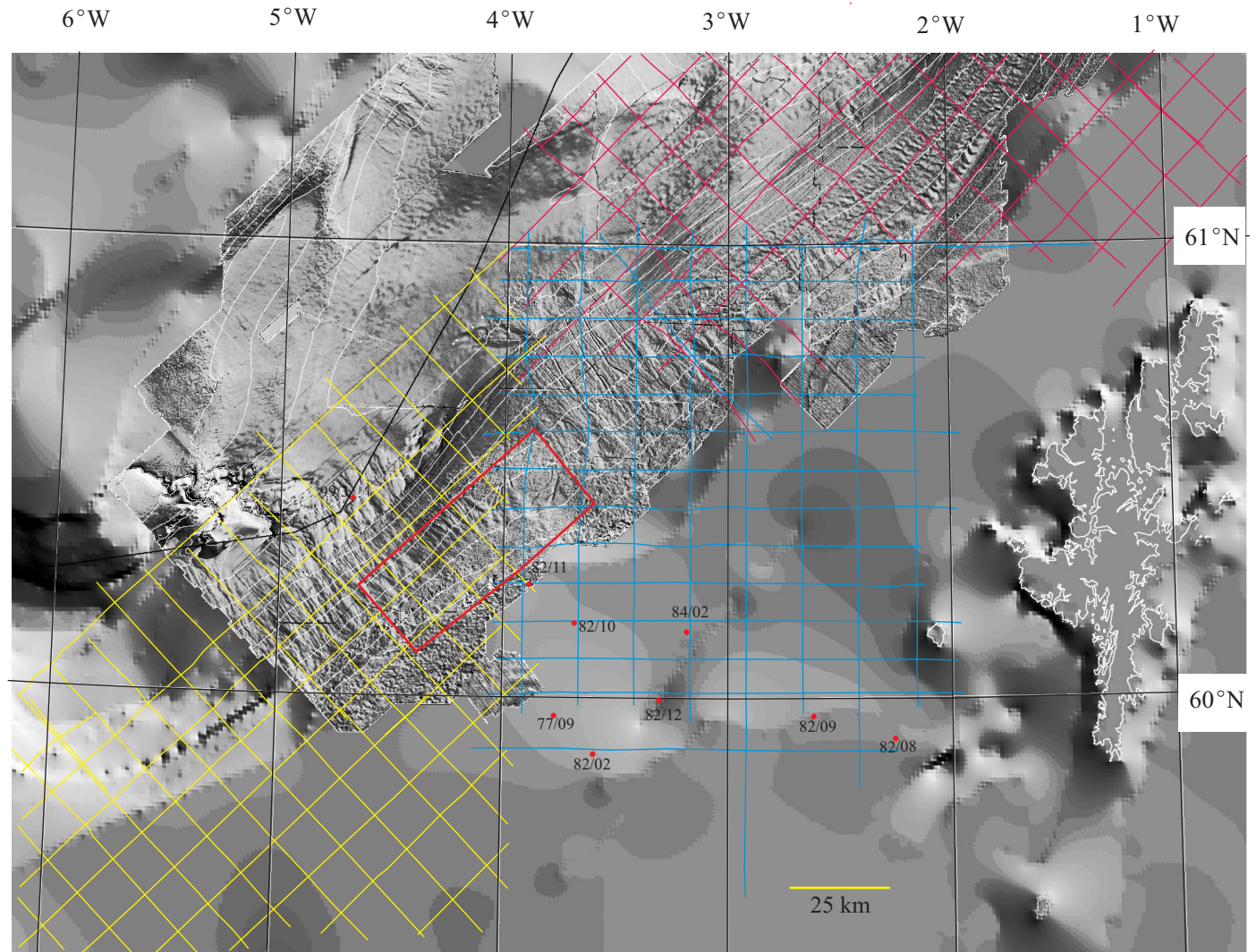


Figure 4.34. Location of area D, upper to mid-shelf, outlined in the red box. For detail of area D, see Figure 4.35.

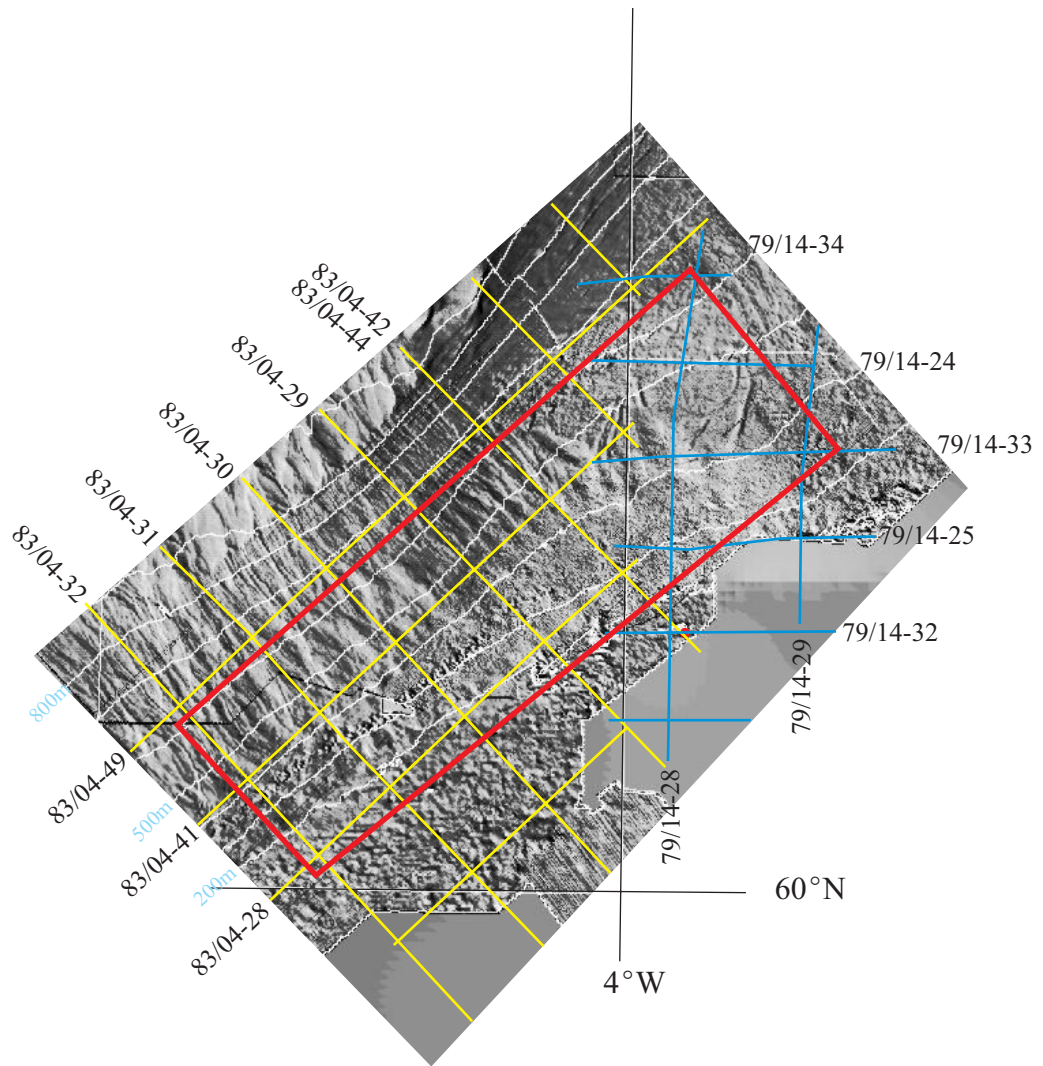


Figure 4.35. Detailed position of seismic lines in area D.

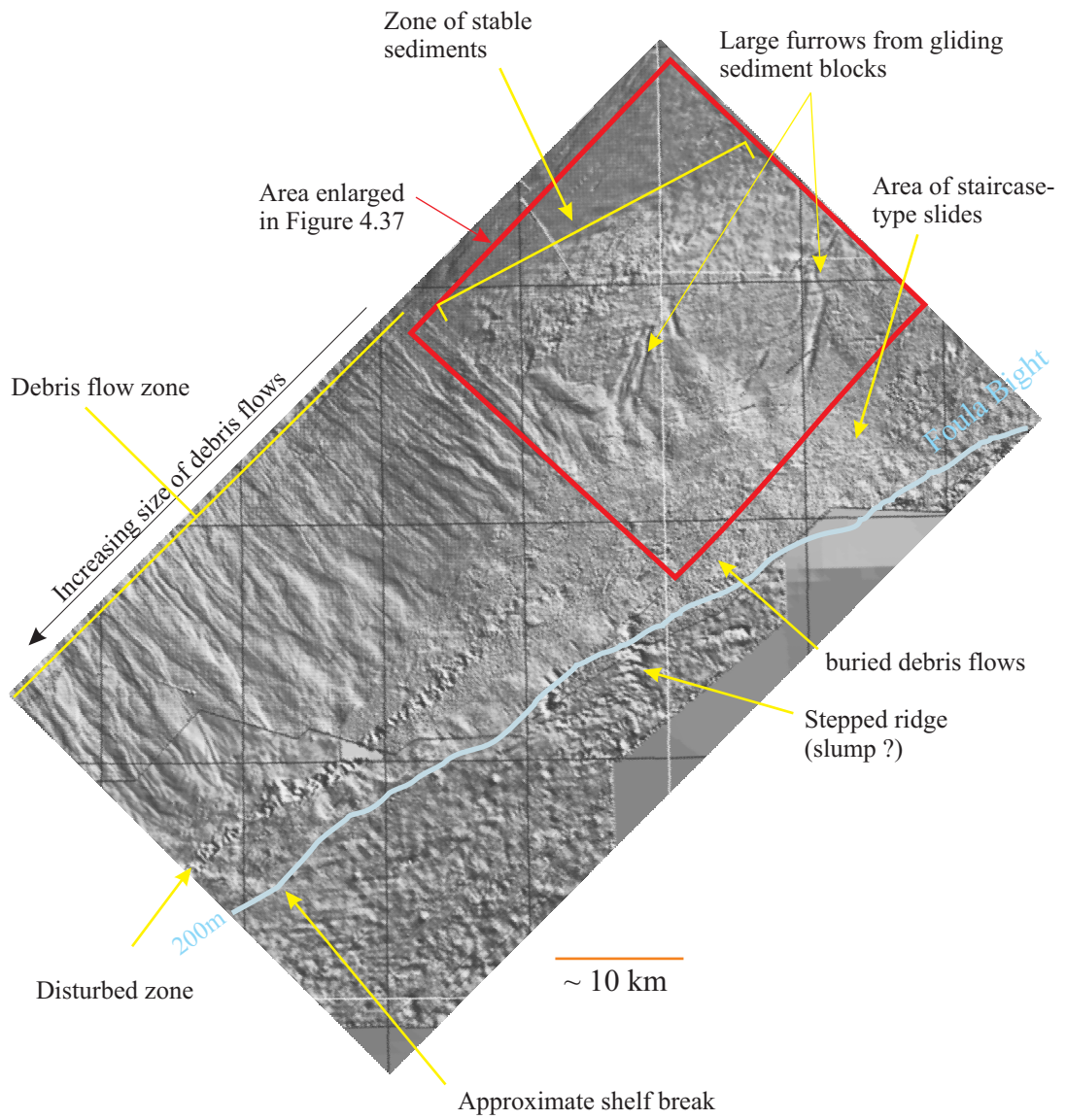


Figure 4.36. Detail of the seabed image features in area D

Along the slope to the south west at a water depth of approximately 350m, lies a linear series of elongate bank-like features, herein referred to as the “disturbed zone” (Figure 4.36). These have previously been described as iceberg scour holes (Masson et al, 1997) and as scour holes formed by descending water spouts at the edge of the Late Pleistocene ice sheet (Holmes et al, 2003). However, there are serious problems with both of these interpretations, which are examined more fully in the seismic data section below.

In the south west of area D below a depth of 400-450m the nature of the seabed changes. Below this depth are a series of large elongate lobes with the long axis orientated parallel to the dip of the slope (Figures 4.35 and 4.36). These lobes run all the way to the base of the slope in the areas where they are present. The frequency and size of the lobes is greatest in the southwest of the area, decreasing north-eastwards. In the north east of the area below the Foula Bight the lobes are absent. Superimposed on the crests of some lobes are narrow, apparently leveed channels. The lobate features have previously been described and interpreted as debris flows (Bulat and Long, 2001; Leslie et al, 2003) The lobate architecture observed is characteristic of debris flows from continental slope settings (Dowdeswell et al, 1997).

In the north east of the area are a series of large, slightly sinuous, furrow-like features up to 400m wide and possibly up to 10km long which lie at an oblique angle to the dip of the slope (Figure 4.36). A pair of sub-parallel troughs, enlarged in Figure 4.37, start at a water depth of approximately 400m where they form broad rounded features. Passing down the slope, the furrows begin to converge, becoming narrower and more sharply defined which suggests a greater degree of incision. The upper section of the eastern trough is joined by a further small furrow which post-dates the larger pair. Where the two intersect a small mound partially blocks the larger trough. All three of these furrow features cut across the most north-easterly of the debris flow lobes on this section of the slope.

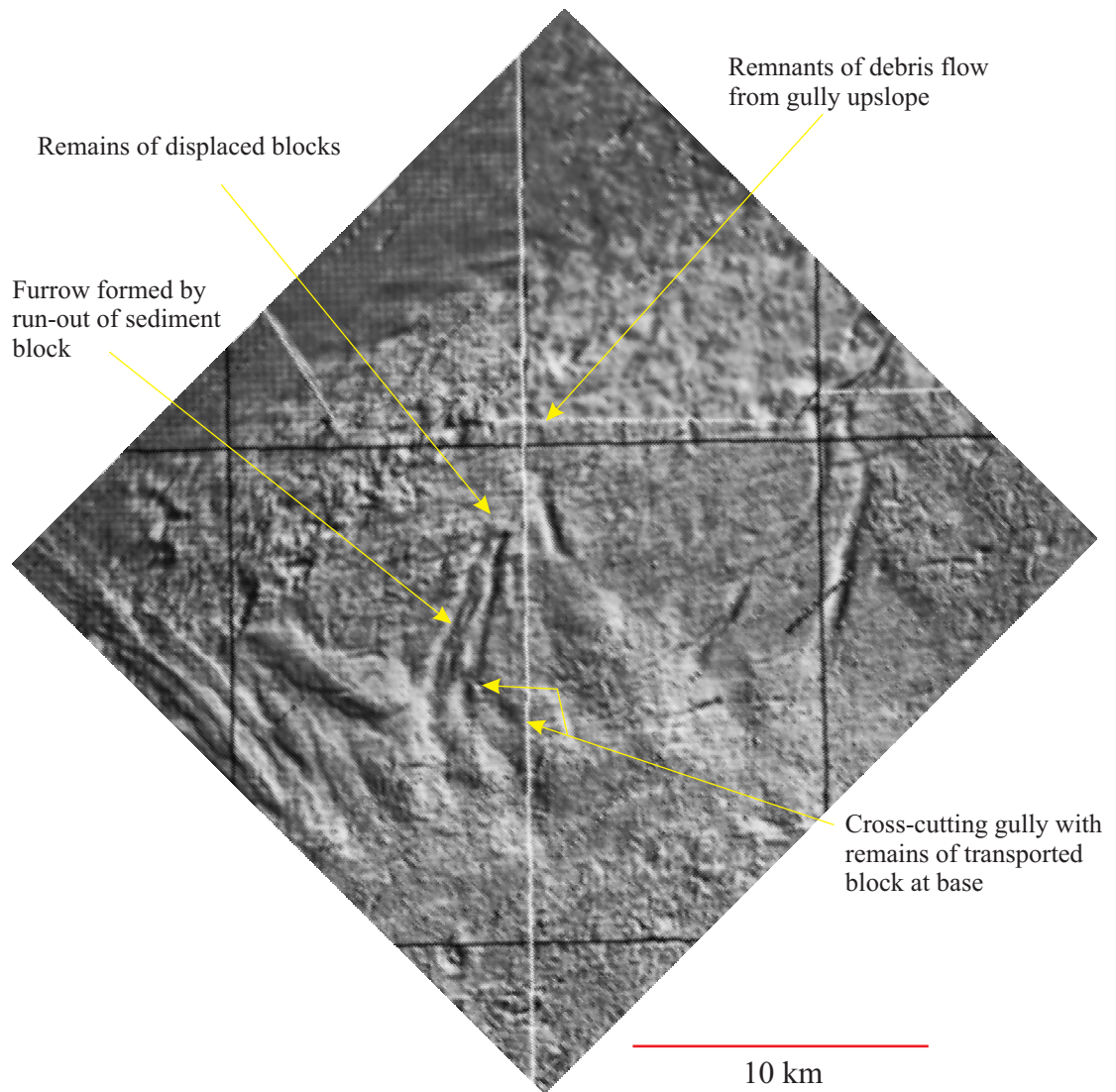


Figure 4.37. Enlarged section of the seabed image in figure 4.36, showing the incised furrows and associated features. Formation of the furrows is thought to be the result of sliding sediment masses or unconsolidated blocks which dispersed after halting at the end of the feature.

The furrows have been interpreted as very large iceberg ploughmarks by Holmes *et al* (2003). The interpretation of these features as ploughmarks is incompatible with some of the characteristics, particularly the orientation and depth of incision. The orientation of the furrows would require a large berg to travel 10km up the slope into water nearly 100m shallower than the point of first grounding which seems implausible. The degree and sharpness of the incision would be expected to increase upslope as the grounding berg dug into the sediment more severely. However, the definition of the trough decreases upslope. In addition, whilst there are large berg ploughmarks parallel to the strike of the slope, there are no large ploughmarks which travel up the slope.

The furrow features are interpreted here as gouges formed by blocks of cohesive sediment formed by slumping, travelling down the slope under gravity. The mound within the north easterly trough appears to be remains of the block which caused the smaller third furrow. Such a process would explain the lack of definition of the features high on the slope where movement was initiated, and a sharper more deeply incised feature lower down where the velocity of the block would have been greater. A block gliding in this manner would also explain the slightly sinuous nature of the troughs. The movement of sediment as large coherent blocks of kilometre scale within debris flows is reported by Masson *et al* (1993) from the northwest African coast. The possibility of detachment of sediment masses is supported by the presence of the small slump structures described earlier.

Micro-scale features produced accidentally in the laboratory during grain size analysis were observed forming in the manner described above following sieving to remove the > 500 μ m fraction. Cohesive lumps of unconsolidated sediment, forming independently of small debris flows, detached from the steeper zones and glided down-slope leaving a gouged trail. The blocks responsible partly disaggregated after movement had ceased, leaving a mound of soft sediment at the end of the gouge. Examples of these features are illustrated in Figure 4.38. How analogous to the processes on the slope these observed examples are is unclear, but the actual process appears similar to that described by Masson *et al* (1993).

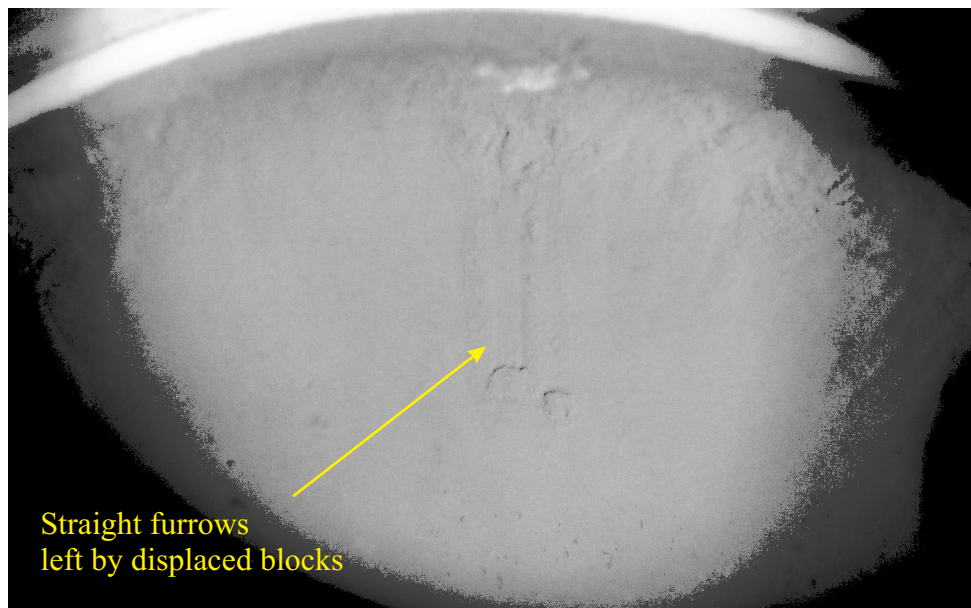
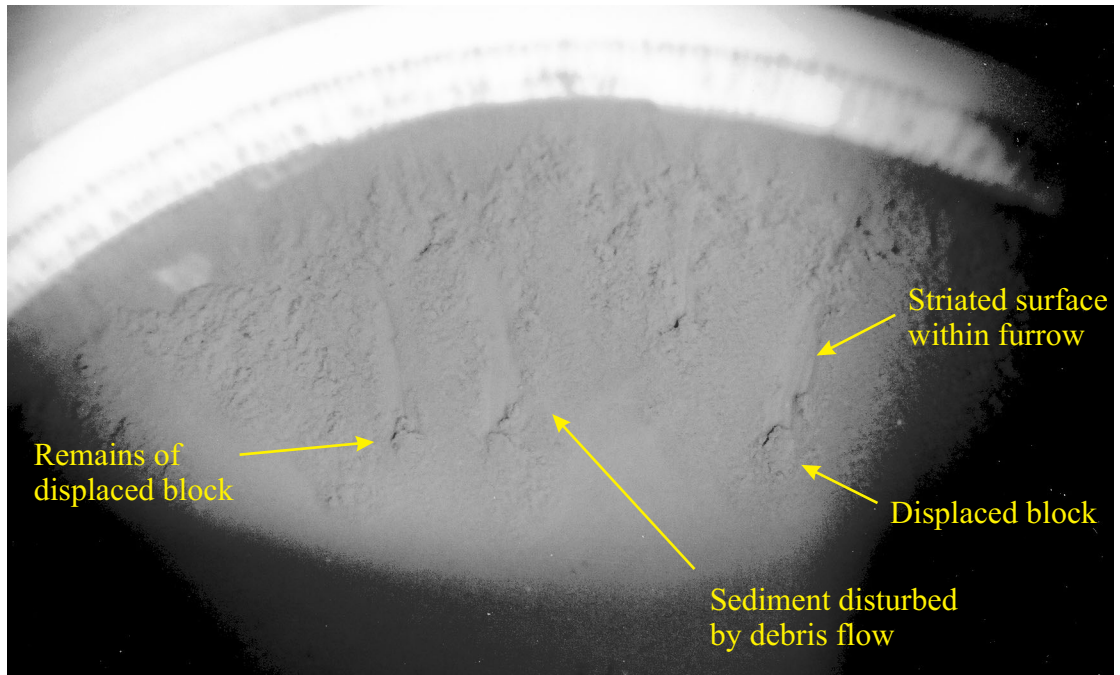


Figure 4.38. Micro-scale furrows created by displaced blocks of unconsolidated sediment gliding downslope as separate events following minor debris flows. The features bear similarity to large furrows observed on the upper to mid slope. The field of view is approximately 20 cm.

Seismic Data

All the seismic sections from the upper slope in area D are very similar in terms of the patterns of reflections and features present. The strike parallel lines 83/04-41, and the dip-parallel lines 83/04/42, 44, 29 and 32 (Figure 4.345) display characteristics typical of many of the lines in this area. Descriptions and interpretations of each line are given below.

Strike parallel line 83/04/41

An examination of line 83/04-41 (Figures 4.35 and 4.39) shows a distinct change in overall glacial thickness across the slope from northeast to southwest. This is true of both the Otter Bank sequence and the pre-Otter Bank glacial section. In the northeast of the area the whole glacial section is no more than 20m thick and is represented by well stratified parallel reflections with high continuity. At the cross-over with line 83/04-43 the Otter Bank sequence becomes less clear, but may occupy the whole glacial thickness of about 25m. To the south-westward, the base glacial unconformity shows long, low angle undulations with little small scale topography. The pre-Otter Bank glacial section in the northeast comprises a series of parallel and sub-parallel stratified reflections with small lensoid bodies of transparent or chaotic reflections. The lensoid bodies increase in scale and definition towards the southwest from 1km wide to over 10km and up to 40m thick. The larger features show evidence of erosion surfaces between adjacent and overlapping lenses. The base of the Otter Bank sequence shows little large scale topography, with only low amplitude undulations (Figure 4.39). Internally the Otter Bank sequence is very similar to the underlying glacial section, consisting of numerous interlocking lensoid bodies with some evidence of erosion surfaces between them (Figure 4.40). In general the lenses in the Otter Bank sequence are more transparent than in the underlying glacial section and have fewer stratified units separating them.

On a smaller scale, the basal unconformity of the Otter Bank sequence shows a series of small v-shaped incisions into the top of the pre-Otter Bank glacial section (Figure 4. 40). There are also a number of smaller, similar features within the Otter Bank sequence lying in the troughs between the lensoid bodies.

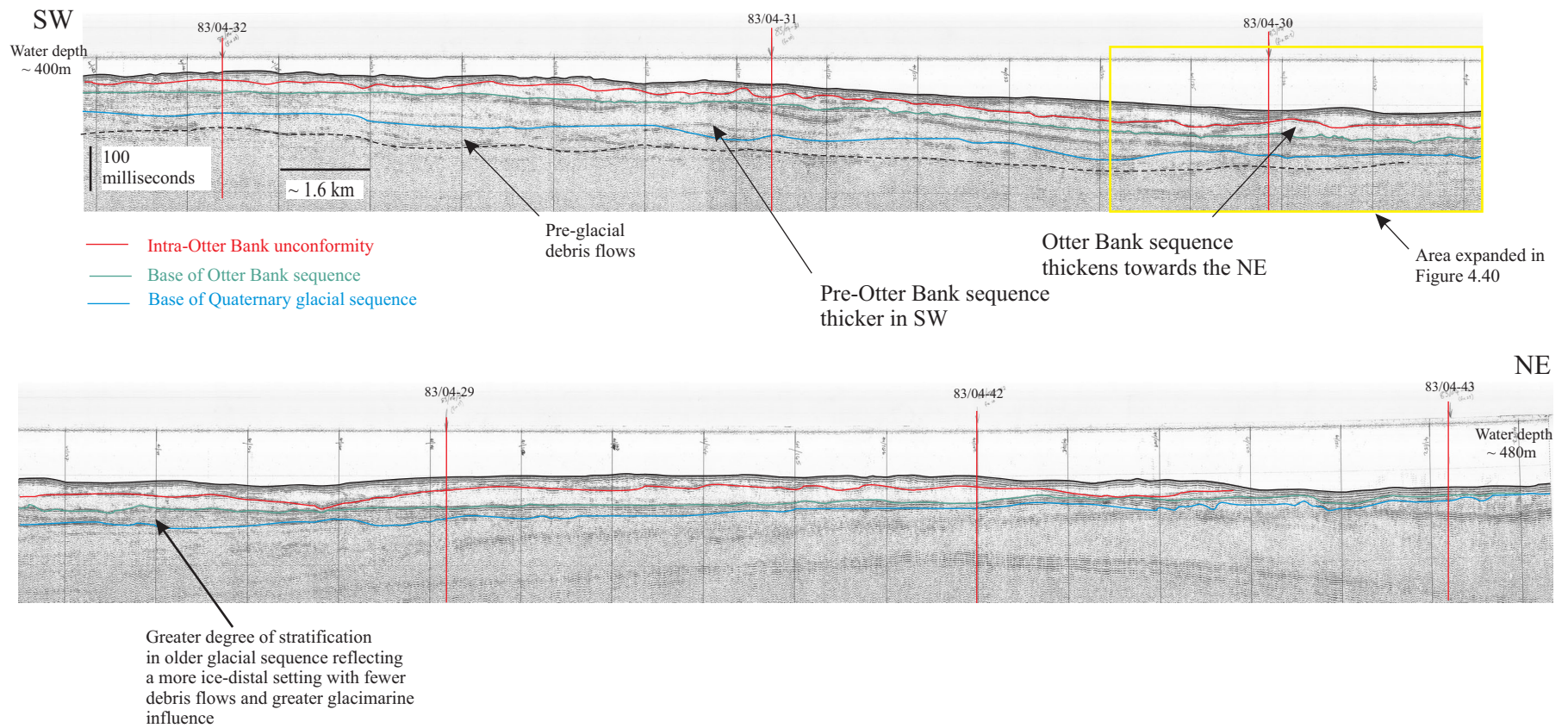


Figure 4.39. Strike parallel sparker line 83/04-41 showing lensoid debris flows on the upper slope in both Otter Bank and pre-Otter Bank glacial sequences. The pre-glacial section also shows evidence of debris flows in the south-west of the area.

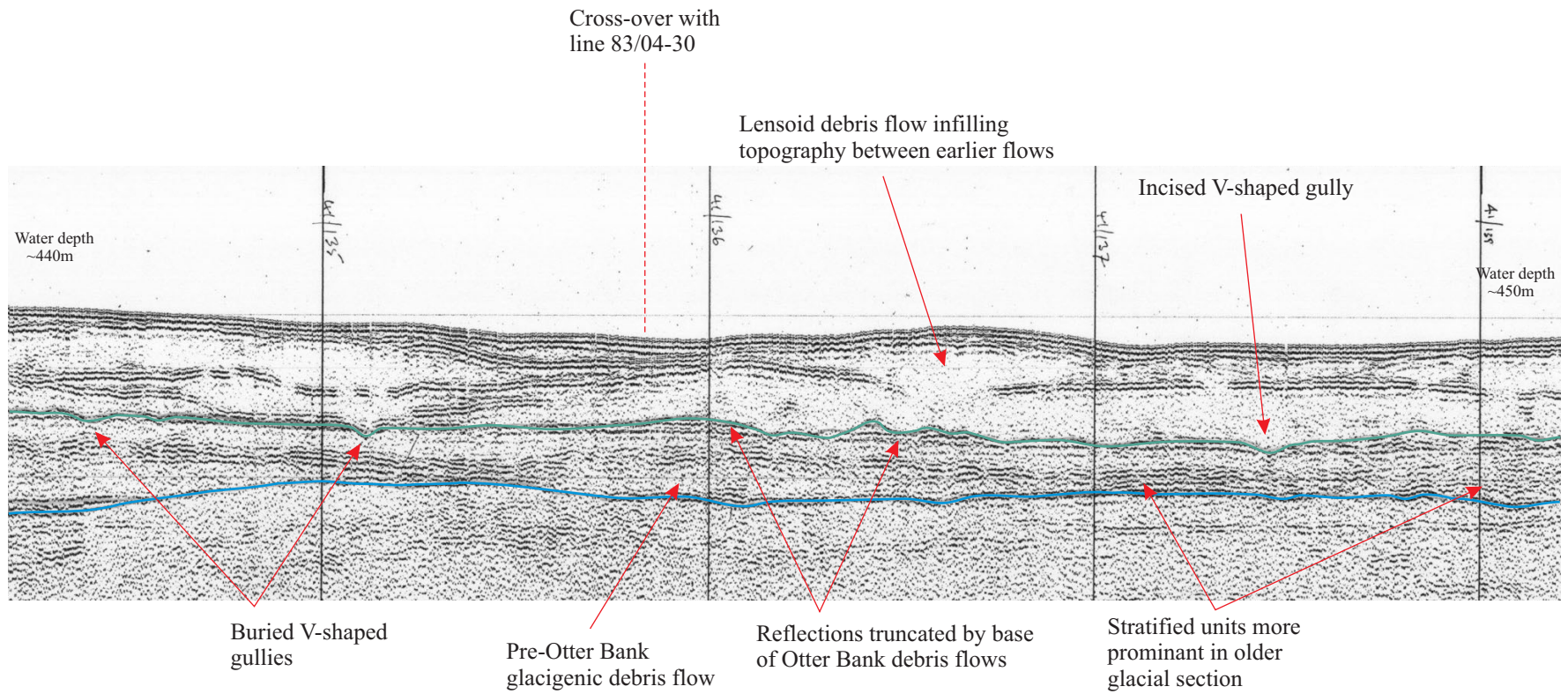


Figure 4.40. Enlarged section of strike parallel sparker line 83/04-41, showing transparent lensoid debris flows separated by thin stratified units. The base of the Otter Bank sequence shows v-shaped incisions and erosion of the underlying glacial sequence. The greater abundance of stratified units in the older glacial sequence indicates its position in a more ice-distal setting than the in overlying Otter Bank sequence.

Interpretation of line 83/04/41

The lensoid features in both the pre-Otter Bank and Otter Bank glacial sequences are interpreted as debris flows based upon the shape and internal properties. The lensoid cross section and predominantly structureless interior are characteristic of debris flows (e.g. Nardin *et al.*, 1979) and have been described from many continental slope settings and other glacial margins. The parallel and sub-parallel stratified units which separate some of the debris flows, and which form a large part of the pre-Otter Bank glacial section sequence may represent the distal ends of debris flows from higher up the slope. However, the high continuity of the reflections, and the lateral continuity of the seismic facies, suggests they are more probably the result of background sedimentation by suspension settling between debris flow events (Stoker, 1990; Davison and Stoker, 2002).

The small v-shaped features observed primarily at the base of the Otter Bank sequence (Figure 4.40), appear to be small incised channels. Such channels are evident on parts of the modern seabed and appear on the seabed image. Much larger channels are seen further along the slope to the northeast and are described in detail in Area F (Section 4.6) of this chapter. The buried features suggest that the conditions forming the large channels have existed on the slope several times in the past.

The focus of debris flow activity changes between the two glacial sections. The pre-Otter Bank glacial section shows a greater number of flows and a thicker sequence in the southwest of the line, whereas the Otter Bank sequence has a focus centred in the region of the crossover with line 83/04-30 further to the northeast (Figure 4.39). This suggests a different focus for sediment delivery to the slope from the shelf edge.

In addition to the debris flows in the glacial section, there is also evidence of similar features beneath the base glacial unconformity in the southwest. Between the crossovers of lines 32 and 30 with line 41 (Figure 4.39), there are a series of interlocking lensoid features. The character of these features is identical to those interpreted as debris flows in the glaciogenic section, and are thus interpreted as pre-

glacial debris flows. Towards the northeast of this section the pre-glacial units become progressively more stratified, suggesting the main focus of debris flow activity was in the southwest. This focus is the same as for the pre-Otter Bank glacial section, suggesting it was controlled by pre-glacial sediment delivery patterns.

Dip-parallel lines

The dip parallel survey lines show a similar pattern of reflections to the strike-parallel lines, comprising acoustically transparent and chaotic units bounded by parallel and sub-parallel stratified units.

Profile 83/04-42

Sparker line 83/04-42, at the northeast end of area D (Figure 4.35), shows a repeated series of elongate wedges which are broadly transparent and a seabed with small scale irregularities (Figure 4.41). The pre-Otter Bank glacial section forms a wedge which thins down-slope, and pinches out at the crossover with 83/04/41. At the shelf end of the wedge the reflection character is predominantly transparent. This changes gradually down-slope into a more structured pattern of discontinuous sub-parallel reflections with some a number of high-amplitude continuous reflections. The higher amplitude horizons mark the boundaries of major seismic packages, dividing the pre-Otter Bank section into 4 wedge-shaped units, visible on the enlarged section of the profile in Figure 4.42. The wedges advance down-slope with each unit overlapping its predecessor, the lowermost unit being very thin. The top of the pre-Otter Bank sequence is marked in places by thin parallel stratified reflections which are truncated by the base of the Otter Bank sequence.

The acoustic character of the Otter Bank sequence is broadly similar to the underlying sequence, in that it comprises a series of elongate wedges with an upslope transparent response and a down-slope stratified response (Figure 4.41). However, the lower section of line 42 shows wedges with decreased internal structure, appearing almost completely transparent. Generally, the main seismic packages are thicker in the Otter Bank sequence, reaching up to 40ms (*c.*30 – 35m). At the lower end of the line in the mid-slope area the Otter Bank sequence appears to represent the

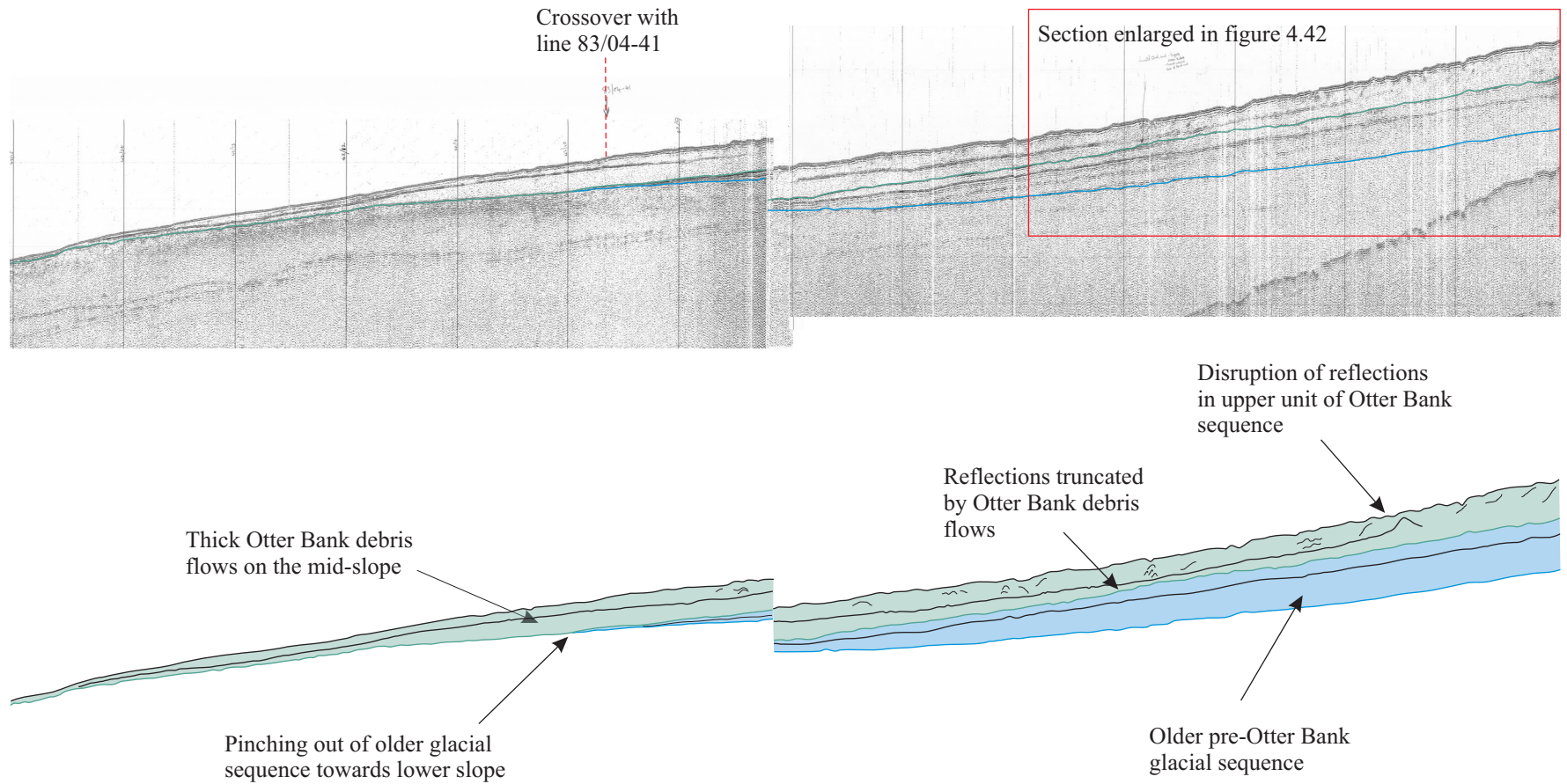


Figure 4.41. Sparker line 83/04-42 from the upper to mid slope, north-east of area D. The data shows large wedge-shaped debris flows in both sequences, with fewer stratified sections in the Otter Bank sequence. The upper unit of the Otter Bank sequence shows disruption of reflections and the seabed on the upper slope.

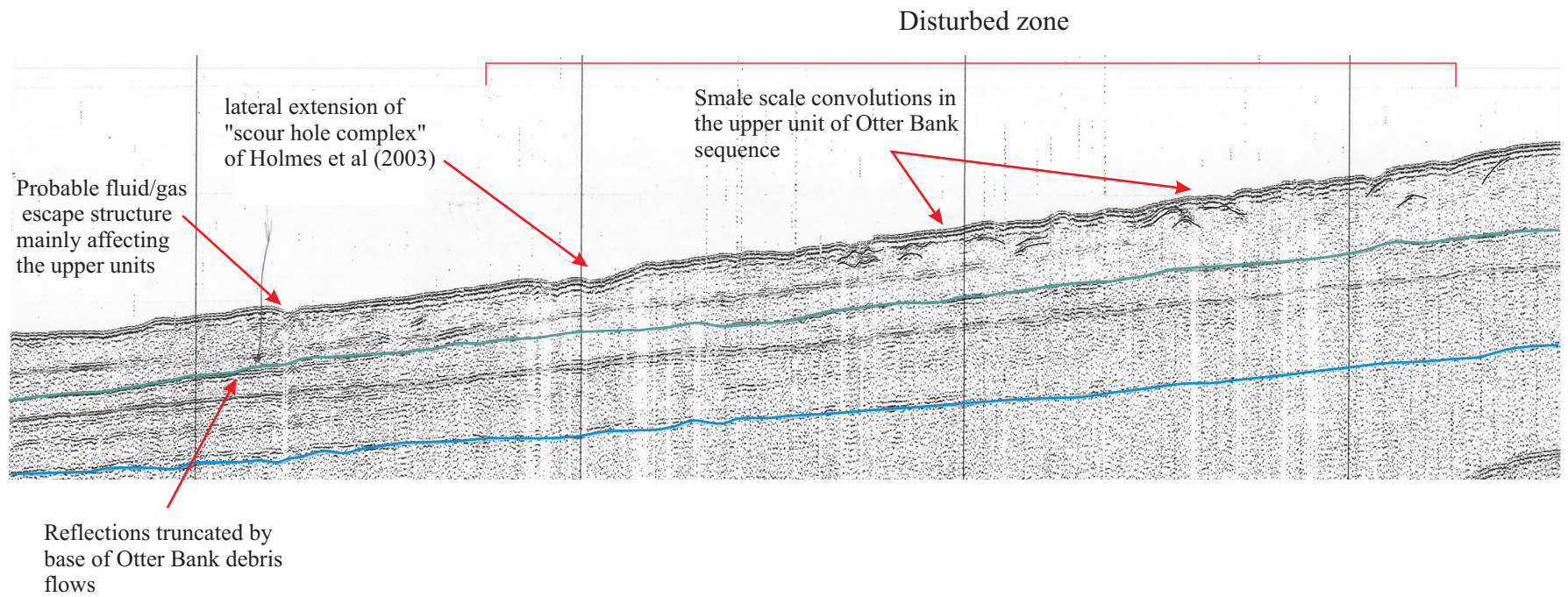


Figure 4.42. Enlarged section of sparker line 83/04-42 from the upper slope showing truncation and convolution of reflections in the Otter Bank sequence.

entire preserved glacial section, although the lower section may be below the seismic resolution.

The Otter Bank sequence can be subdivided into two major units. The base of the Otter Bank show evidence of having eroded the underlying stratified units of the pre-Otter Bank sequence (Figure 4.42). The truncation marks a change in the thickness of the lower Otter Bank unit from *c.*10ms on the up-slope side to 25ms on the down-slope side. The basal unconformity of the Otter Bank sequence also shows increased irregular topography at this point.

In addition to the larger-scale features described above, profile 83/04-42 (Figure 4.42) also shows evidence of smaller-scale features in the Otter Bank sequence. In the upper package of the Otter Bank sequence on the upper slope there are numerous small convolutions and disruptions of the internal stratification and the seabed. These are described in more detail below from a high-resolution sparker profile – 83/04-44 – which lies parallel to 83/04-42 almost along the same line (Figure 4.35).

Interpretation of dip profile 83/04-42

The elongate features are characteristic of debris flows seen in dip-parallel sections (Nardin *et al.*, 1979). The shape and internal character confirms the interpretation from the strike-parallel section that these features are debris flows. The downslope advance of the pre-Otter Bank section debris flows (Figure 4.42) indicates an advance basin-wards in the sediment supply from the top of the slope. The apparent progressively longer run-out of the flows is the result of the prograding shelf edge advancing and steepening the upper slope. The thinner units at the base of the sequence represent the distal end of the earliest flows when the shelf edge, and hence the point of sediment supply was further back on the shelf. The apparent increase in internal structure in the distal toes of the early flows is probably due to thin stacked units rather than actual stratification within an individual flow.

The two Otter Bank sequence units are interpreted as debris flow packages, showing the same acoustic properties as those in the pre-Otter Bank. The change in the

thickness of the lower Otter Bank package described above (Figure 4.41) has two possible interpretations; i) increased downwards erosion during the emplacement of the lowermost Otter Bank unit, which also truncated the stratified reflections in the underlying section; or ii) slumping of the pre-Otter Bank sequence sediments prior to the emplacement of the Otter Bank sequence. The second scenario would explain the apparent disruption in the stratification up-slope of this zone and would have altered the surface topography before the Otter Bank sediments were deposited. The stratification at the distal end of the slump could be the product of repeated minor debris flows down the steepened edge of the slump. The emplacement of the large Otter Bank package which forms the upper of the two Otter Bank packages, overlies this zone and would have in-filled the topography, possibly eroding the underlying surface at the point where the topography became steeper (Figure 4.42).

High-resolution profile 83/04-44

An examination of a deep-tow sparker line, 83/04-44 which lies almost on the same track as 83/04-42 shows the seabed features in much greater detail, although subsurface structure is limited (Figure 4.43). The seabed profile from line 44 shows numerous surface irregularities. The features form a series of mounds, hollows, steeper angled slopes and steps which become larger and less numerous down-slope (Figure 4.44a-e). The features range in size up to 8m in height and *c.* 100 – 300m long. When viewed at a larger scale (Figure 4.44a-c and e) some of these features closely resemble small scale slumps and shears. In particular, the pattern is similar to staircase rotational slides described from East Greenland (Dowdeswell et al, 1994; Whittington and Niesson, 1997). However, due to the lack of penetration, there is little subsurface detail which would confirm this interpretation. The main alternative interpretation for the mounds and hollows is that of iceberg ploughmarks. Some of the hollows appear to be incised, with small banks at either side (Figure 4.44d). In addition, there appears to be thin stratified reflections at the base of the hollows, suggesting later deposition draping or infilling the features following formation. An iceberg origin is also supported by the decrease in abundance down-slope (Solheim, 1997). Evidence from side scan sonar and seismic profiles has previously indicated

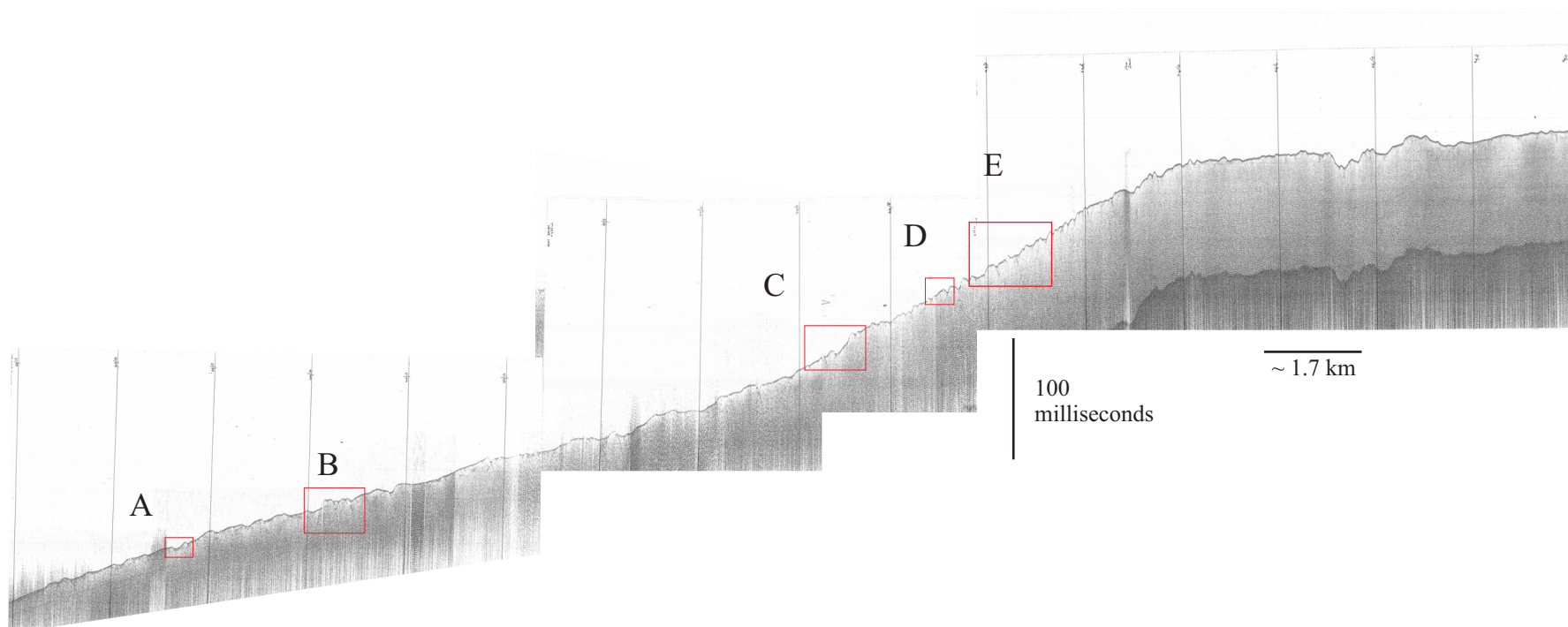
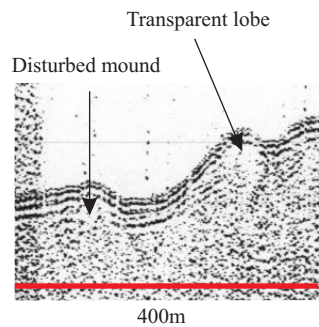
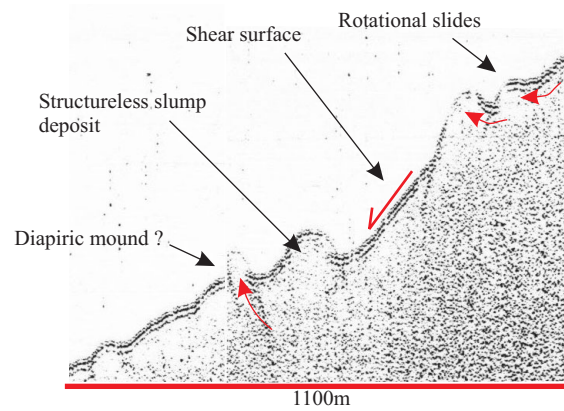


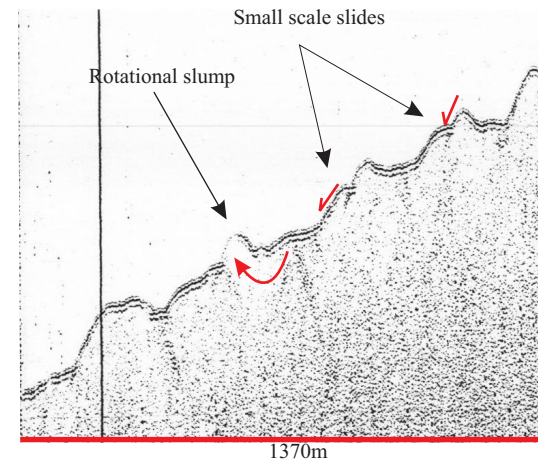
Figure 4.43. Deep tow sparker line 83/04-44. Data shows a very irregular seabed, although penetration is very limited. The number of seabed features decreases downslope, with the highest frequency of features occurring on the steepest part of the upper slope. This suggests at least some of the features are iceberg ploughmarks. Features A-E are enlarged in Figure 4.44.



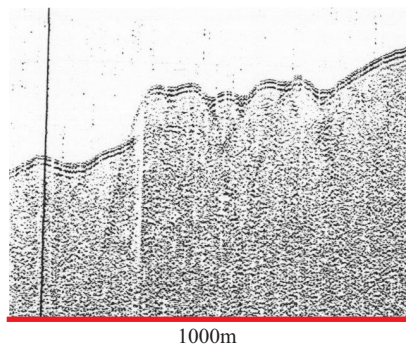
A) Small scale slump



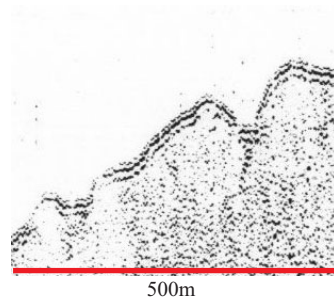
C) Small scale slumps and rotational slides



E) Combination of rotational slumps and slides



B) Slumps with contorted internal structure



D) Probable iceberg ploughmarks with steep sided, flat-based gullies.

Figure 4.44. Small scale instability structures from the upper slope, enlarged from line 83/04-44.

extensive iceberg scouring on the outer shelf and upper slope in this region (Stoker et al, 1993).

Profile 83/04-29

South west, along the upper slope, profile 83/04-29 (Figure 4.35) shows the overall glacial sequence is thicker, reaching a total thickness of *c.* 80m at the cross-over with line 41 (Figure 4.39). The Otter Bank sequence at this point is *c.* 60m and is at its thickest on the mid-slope, comprising large lensoid packages with transparent acoustic character, forming part of the debris flow accumulation described earlier. The packages of stratified reflections which separate the debris flows tend to be thinner and are absent in some places. The uppermost stratified package is thicker than those beneath, suggesting a longer period of stability in which background depositional and re-working processes were dominant, although without core data this is difficult to confirm.

Airgun data from line 83/04-29 (Figures 4.35 and 4.45) shows a very thick overall glacial sequence at the top of the slope, reaching 180m, of which the Otter Bank sequence forms approximately half. The basal glacial unconformity appears relatively smooth with only minor topography but does show several breaks in slope angle on a large scale. The pre-Otter Bank section shows limited coherent structure on the upper slope, although on sparker data the stratified top of the sequence seen on the shelf is still visible. There are still two distinct units to the lower sequence, the basal unit being largely made up of chaotic reflections and the upper, thinner unit showing more stratification. The lower unit infills a large change in the pre-glacial slope profile, resulting in a very thick upper slope sequence. The chaotic nature of the reflections and the data from sparker line 83/04-41 described above (Figures 4.39 and 4.40), suggests the debris flows present at the cross-over point probably extend from the shelf break to the lower slope. The stratified top section may represent background hemipelagic deposition, or may be a basin-ward extension of the sand unit seen on the shelf which marks the top of the lower sequence.

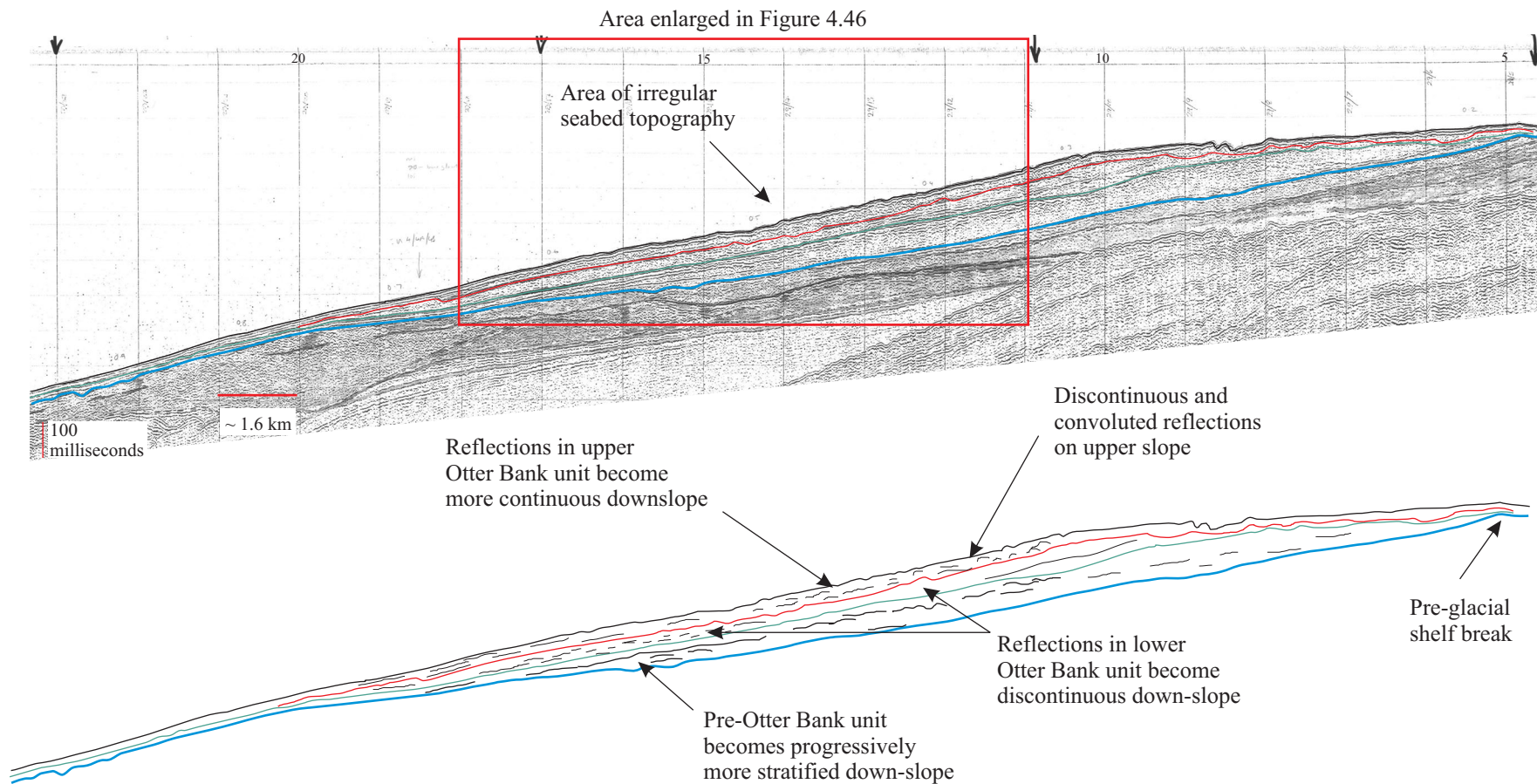


Figure 4.45. Airgun line 83/04-29 showing the changing nature of reflections the upper and mid slope. The zone of irregular seabed topography is evident as disrupted reflections in the subsurface structure.

The Otter Bank sequence on the upper slope consists of two main packages made up of chaotic reflections. Both units show an increasing degree of internal structure with distance down the slope, with structure in the lower unit beginning at a higher level. However, down-slope of fix point 12 (Figure 4.45) the internal structure becomes more chaotic and the seabed more irregular. This is shown in greater detail on sparker data from line 29 shown in Figure 4.46. The lower unit still shows sub-parallel stratified reflections but with reduced continuity. The upper unit becomes increasingly chaotic down-slope except for the top, which shows convoluted sub-parallel reflections below the seabed pulse. Above fix point 12 the seabed is comparatively smooth with the exception of small undulations at point 11. The upper unit of the Otter Bank sequence at this point shows an unusual group of vague sigmoidal reflections which initiate in the stratified top of the upper unit and propagate down into the lower unit (Figure 4.46). Beyond this point the upper sequence loses much of its internal structure. Between fixed points 14 and 15 in Figure 4.46, the seabed shows several marked changes in slope angle which also disrupts both the upper and lower Otter Bank units. Below fixed point 15 the lower unit is internally chaotic. The upper unit becomes more stratified and thins to about 20m but is still convoluted. This part of the upper and mid-slope corresponds to the line of banks and mounds observed on the seabed image (Figure 4.36) – the disturbed zone- described above. It is also the zone described by Masson et al, (1997) as a series of iceberg scours, and by Holmes et al, (2003) as scour holes formed at an ice sheet margin.

The seismic data suggests this part of the slope represents an area of disturbance formed as a result of movement within the Otter Bank sequence. The most likely cause of this movement is sliding along major sedimentary surfaces beginning at the sigmoidal reflections on the upper slope and creating the convolution in the stratification further down the slope. This interpretation, and the arguments against the alternative explanations are expanded upon at the end of this section.

Profile 83/04-32

Seismic data from the south-west of area D adds further to the characterisation of the

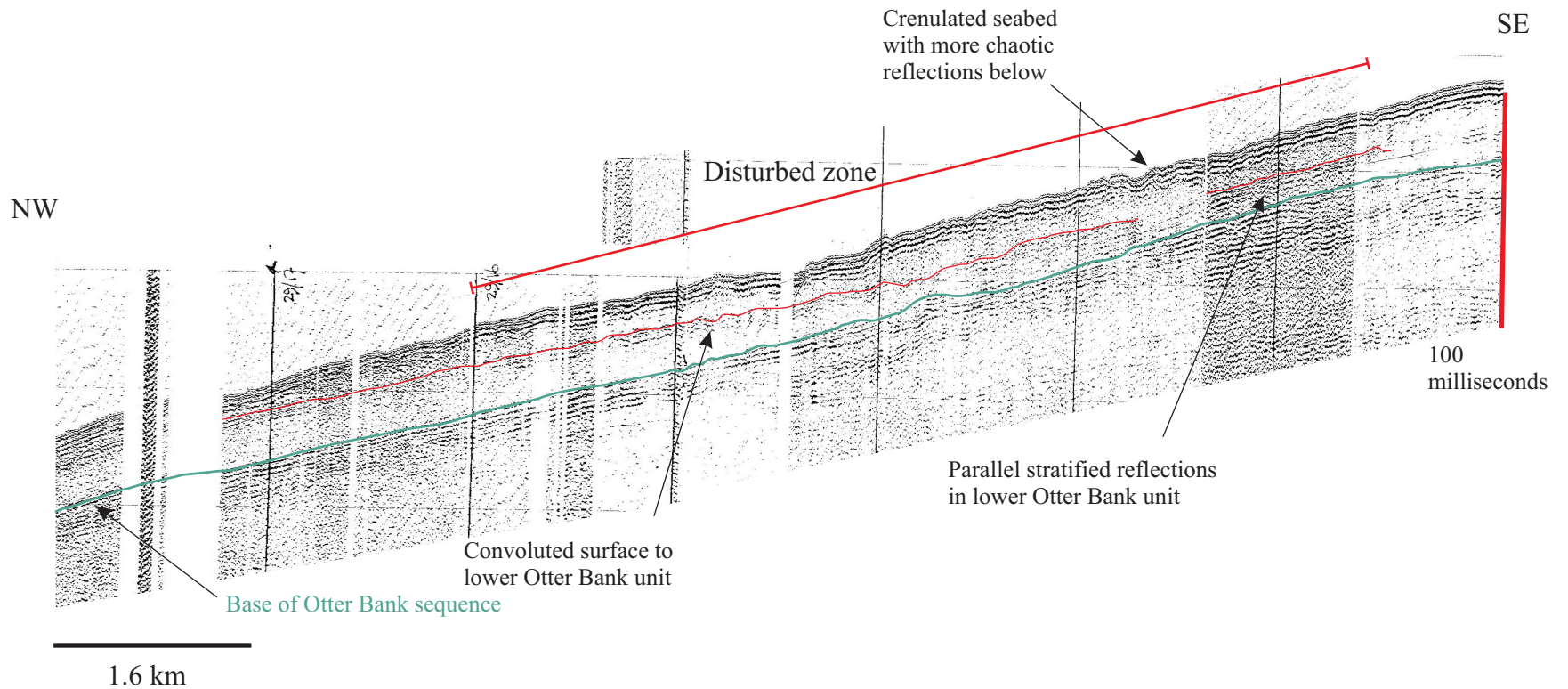


Figure 4.46. Sparker section of line 83/04-29 from the mid to upper slope. The lower Otter Bank unit shows a parallel stratified reflection pattern near the top of the slope, which becomes less well defined down-slope. At the northwestern end, the stratification begins to re-appear. The upper unit shows discontinuous convoluted reflections within the centre disturbed section and evidence of a possible detachment surface towards the top of the slope, suggesting sliding along surfaces parallel to the reflections.

upper and mid-slope. Airgun data from line 83/04-32 (Figure 4.35) shows a generally even slope with little seabed topography. The only obvious change in the seabed profile of the upper slope lies between fixed points 29 and 30 (Figure 4.47) where there is an apparent step down of c. 20m over a distance of about 250m. This is the lateral extension of the banks and mounds seen on the seabed image and corresponds to the features described on line 29. Similar features are also present on the lines 30 and 31 which lie between lines 29 and 32.

The basal glacial unconformity on line 32 is generally smooth with long, low amplitude undulations on the upper slope. Topography on the unconformity below this point is minimal. The pre-Otter Bank glacial sequence shows a thick prograding wedge extending from the shelf edge on to the upper slope where it reaches approximately 50m thick. The two phases identified in previous lines are less distinct and thinner but still show a transition from transparent and chaotic reflections on the upper slope to more stratified sub-parallel reflections lower down the slope. The packages beneath the seabed features between fixed points 29 and 30 show a more transparent acoustic response than the areas up or down-slope of this area. This suggests the sediments have been disturbed following deposition by the same process which formed the seabed features. The remainder of the lower sequence appears to show the same stacked debris flow packages seen on adjacent lines.

The Otter Bank sequence on line 32 shows less internal structure on the upper slope than previous lines. The two packages visible further to the north-east are hard to define, especially below the seabed features described above. In the area down-slope of the shelf break the packages consist of chaotic reflections, becoming more transparent further down-slope. The upper package of the Otter Bank sequence shows a discontinuity in the reflection pattern at the point of the seabed slope break (Figure 4.48). This appears to form a sloping surface, and may represent a fault plane. Approximately 2km down-slope there is a further surface, sloping in the opposite direction which also disrupts the reflections of the upper Otter Bank package. Down-slope of this area, the upper package is generally transparent with few internal reflections and thins towards the lower slope.

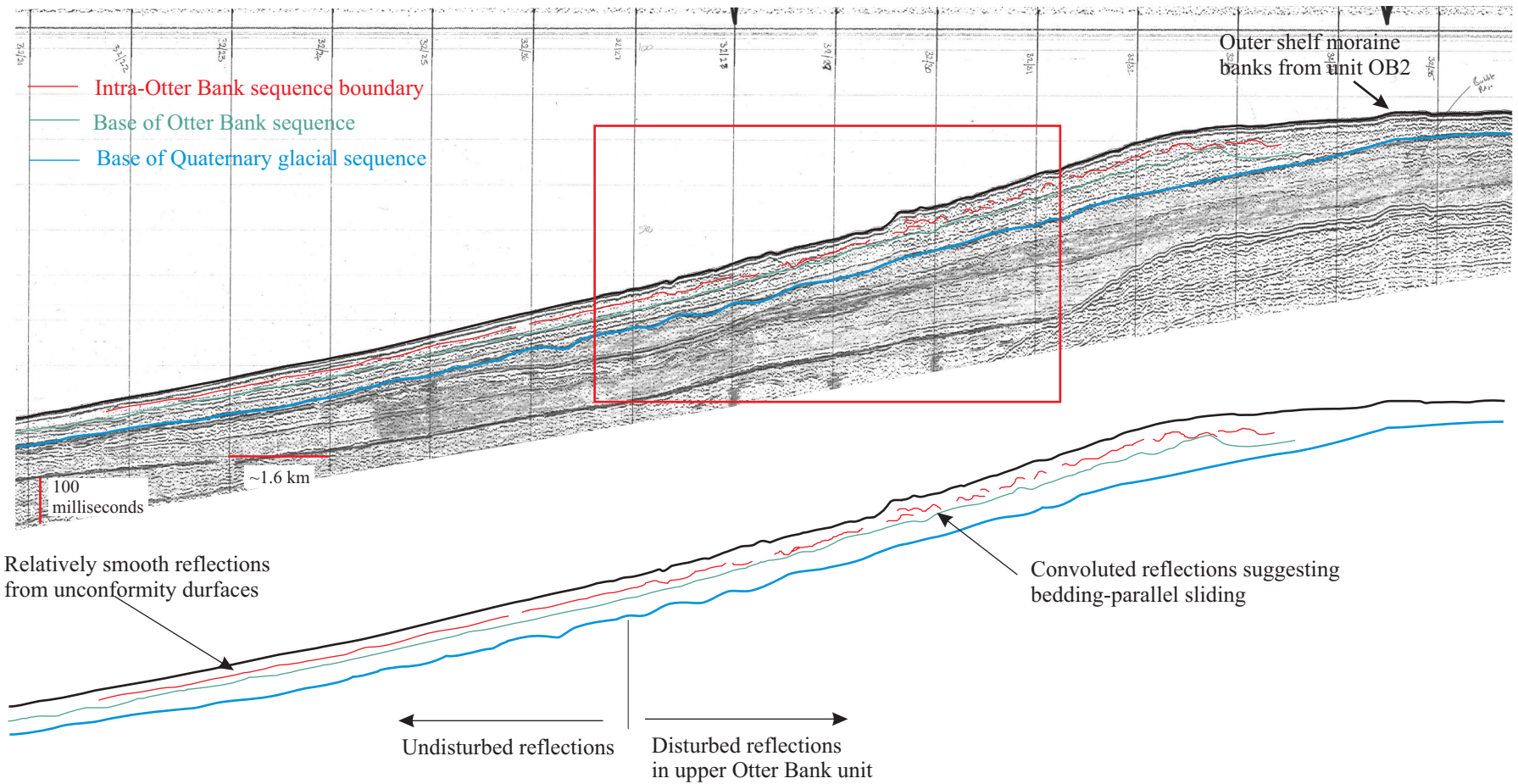


Figure 4.47. Airgun line 83/04-32 showing the shelf edge and mid-upper slope. The area outlined is enlarged in Figure 4.48.

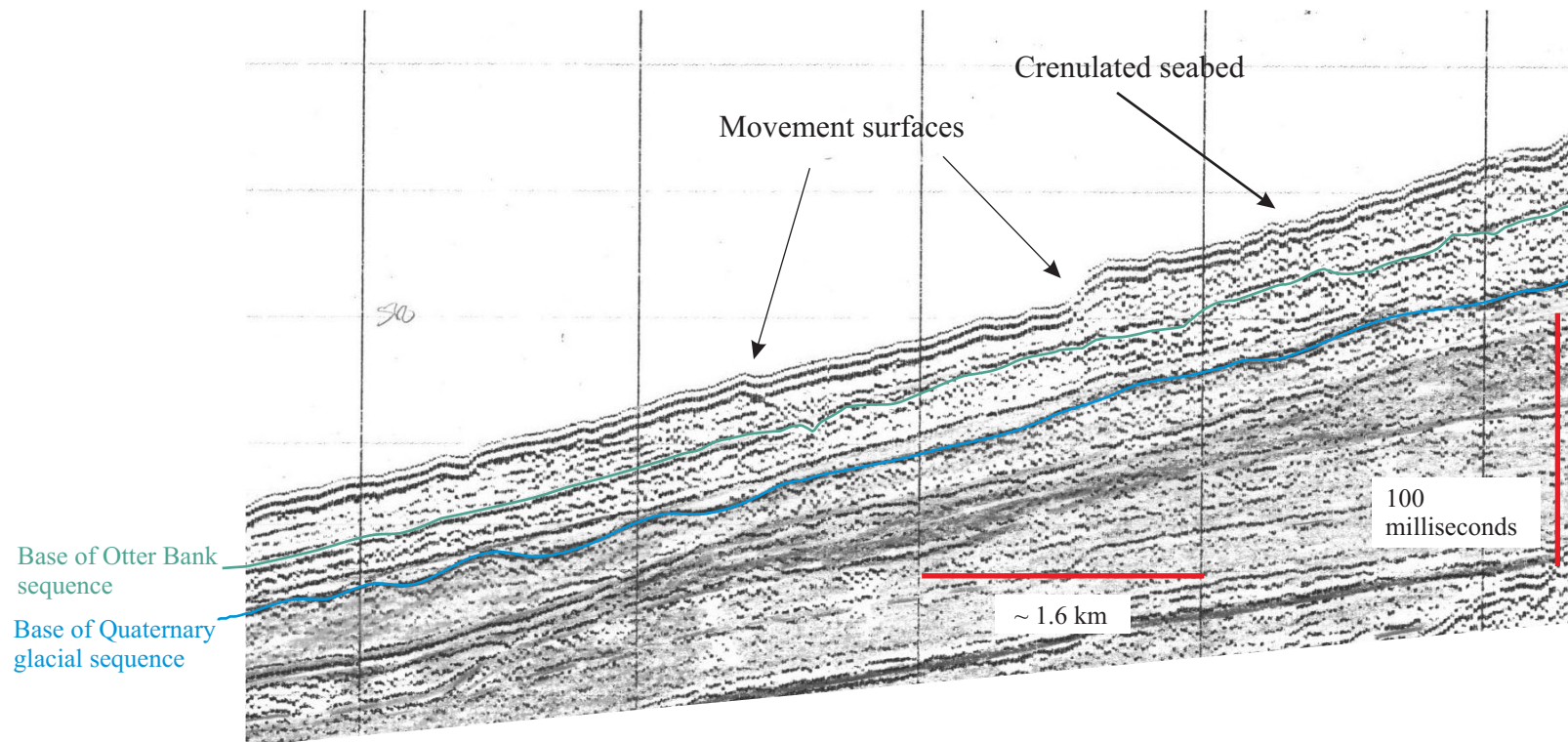


Figure 4.48. Enlarged section of Airgun line 83/04-32 showing the area of disturbance due to mass movement following the emplacement of the second phase of Otter Bank debris flows, which are probably equivalent to OB2 on the shelf.

The interpretation of the lower glacial and Otter Bank sequences on line 83/04-32 as being dominated by debris flows is consistent with that of the previous lines in area D. This also agrees with previous interpretations of the undifferentiated Quaternary sequence described from 83/04-32 (Damuth and Olson, 1993).

4.4.2 Interpretation of the disturbed zone

The disruption seen on the upper slope, both on the sea-bed image (Figure 4.36) and the seismic data (Figures 4.42 and 4.45-48) gives the impression of extension in the upper unit of the Otter Bank sequence. Movement of this style may have been responsible for triggering debris flows and destroying internal structure in the sediments further down-slope. Moving north-east from line 32, the disrupted zone changes from a possible fault movement (Figure 4.48) to an area of convoluted reflections on lines 30 and 29 (Figures 4.42, 45 and 46), which give the impression of sliding along bedding surfaces or weak layers. Lines 42 and 44 showed evidence of numerous small disruptions (Figures 4.42 and 43), suggesting small scale slumps or rotational shears rather than larger movements. This is borne out by the presence of the small ridges seen on the upper slope on the seabed image (Figure 4.36). An element of rotational shearing may also be responsible for the formation of the steeper slope seen in the headwall of the small hollows, effectively forming a fault scarp. Any such scarp face formed in the poorly consolidated sediments of the slope would be liable to small scale erosion by slumping, debris flow formation and along-slope currents, potentially forming the hollows observed on the present slope. Development of the hollows in this manner would also deliver the sediments which smooth the seabed down-slope of the features. When viewed as part of the overall Otter Bank sequence isopach map in Chapter 6, the theory of a large movement or series of movements down-slope from this area is a valid interpretation.

The interpretation of the hollows by Masson et al (1997), as iceberg scours does not explain the linear nature of the features along a 50km section of the upper slope, confined to a narrow zone. The iceberg interpretation also fails to take account of the disruption of the sediments further down-slope or the depth of the disturbance which disrupts sediments well below the apparent base of the hollows.

The interpretation of the disturbed zone as marking an ice sheet margin by Holmes *et al* (2003), with water spouts falling off and through the ice, scouring hollows and re-depositing sediment further down the slope lacks important features. This interpretation requires an ice sheet to traverse the shelf break and travel 10 km down the slope before halting for a sufficient period to carve out 35m-50m of topography in water between 200m and 350m deep and then retreat back up the slope. Serious arguments against this scenario.

i) If the ice sheet had crossed the shelf edge onto the upper slope, the change in slope angle would have had a rejuvenating effect upon the rate of ice flow. This could be expected to have an erosive effect on the sediments at the shelf edge, increasing down-cutting. There is no evidence of this style of erosion at the shelf edge.

ii) Another problem with this theory is the lack of an end moraine or grounding line wedge marking the point of maximum ice advance. Any ice sheet large enough to advance down the slope would produce lift off moraines at the point of decoupling with the seabed. There is no evidence of any moraine banks beyond the shelf edge. Evidence from the shelf has shown that when an ice sheet has increased buoyancy but is still grounded there is a greater chance of depositing an acoustically structured deformation or lodgement till during the advance phase. There is no evidence of stratified packages on the upper slope other than those which separate the debris flows which are of a different character.

Below the disturbed zone, the debris flows dominate the seabed topography but generally become thinner with individual flows becoming more prominent. The acoustic character changes from the chaotic or transparent response higher on the slope, to more sub-parallel stratified reflections on the mid to lower slope. This transition lies between 5 km and 10 km from the shelf edge, which technically is still within the ice-proximal zone and subject to high sedimentation rates from meltwater

plumes and iceberg rafting (Boulton, 1990). Holmes et al (2003), suggest the transition to the mounded sea floor topography of the debris flow heads marks the zone at the ice margin. However, the absence of a grounding line wedge as stated above, is not explained. In addition, the seabed image shows that in the north-east of area D the debris flows do not become defined for 5 km beyond the apparent ice margin (Figure 4.36). This is reinforced by lines 83/04-42 and 79/14-28, where the slope angle on the upper slope is less, which show debris flows dominating the subsurface almost all the way to the shelf break without the ice margin features being present (Figure 4.41). This indicates that the position of the debris flow heads is more a function of slope angle and debris flow initiation is not related to the position of the supposed ice margin.

From this study, a possible triggering mechanism, and potentially a cause of the “scour” hollows is provided by an examination of current data from the Faroe-Shetland Channel. Hosegood et al (2003), reports the presence of high-frequency internal waves in the thermocline which impinge on the slope at depths of around 400m. The waves are associated with an internal bore which propagates up-slope, temporarily reversing current directions and causing a 4° drop in temperature. The study also suggests that this activity causes significant re-suspension of sediment in this zone. Whilst the depth of this phenomenon is deeper than the present hollows, the bore shows variation with changes in atmospheric and salinity conditions, suggesting that the depth of this wave is to some extent, variable (Hosegood and van Haren, 2003). The past history of the bore is unknown, but with the different circulation conditions and salinity which existed in the Faroe-Shetland Channel following the LGM, the possible erosive effect of such a current process cannot be ignored.

Alternative triggering mechanisms for the debris flows are more difficult to identify. The zone of debris flow generation in area D overlies the Judd Fault and fracture zone (Figure 3.2) which could have undergone movement as a result of sediment loading. Fluid escape from the accumulated sediments due to loading could also conceivably trigger the debris flows, although evidence for this is lacking.

4.4.3 Interpretation of upper to middle slope

The upper and middle slope examined in area D show a number of key features related to the glacial development of the margin. Most importantly, is the contribution to the progradation of the margin. The accumulation of debris flows derived from the shelf edge in the pre-Otter Bank and Otter Bank sequences show how the slope both progrades into the Faroe-Shetland Channel, and becomes steeper on the upper slope. On the upper slope, architecture is dominated by wedge-shaped debris flow packages with little internal structure. Towards the mid slope, the packages become more structured and may include elements of re-working and deposition from along-slope currents.

The importance of the two distinct units within the Otter Bank sequence on the upper to mid slope is unclear, but probably relates to different phases of glacial activity and ice sheet advance to the shelf edge. The reflector defining the boundary between the two Otter Bank units, and the structure within the units becomes indistinct up-slope and disappears at the top of the slope. The change from unstructured chaotic or transparent reflections to a progressively more ordered pattern of transparent wedges and stratified sheets, marks the change in sedimentary facies from lodgement and melt-out/waterlain tills to debris flows and suspension settling/plume sediments. This transition also forms a general marker for the point of maximum ice advance. The change in reflection pattern is similar to that predicted by the model of Boulton, (1991), although this style of transition has also been attributed to iceberg scouring destroying the structure of the slope-top sediments and forming an iceberg turbate (MacLean, 1997).

There is evidence of sediment general instability on the upper slope. The process of margin progradation by debris flows is the most important indicator of this. However, there is also evidence of smaller-scale slope failure from possible failed movements, where large masses have slipped relatively short distances, to semi-solid blocks less than of less than 1 km sliding down the slope for 10 or more kilometres.

4.5 Area E: Lower slope and basin floor

Area E lies at the base of the slope at the north-western end of transect A (Figure 4.49). Data coverage of this area is generally very good, with sea-bed image, seismic and borehole core data. Part of the data and interpretation presented in this section forms part of a published study of the deep water glacigenic sequence of the West Shetland margin by Davison and Stoker (2002), and is included in appendix 1.

4.5.1 Sea-bed morphology and seismic data

Bathymetry

The bathymetry of the lower slope and basin floor (Figure 3.1) shows no obvious features except for a scarp-like change in gradient in the area of the Judd Deep, a series of steep slopes in the floor of the Faroe-Shetland Channel. The isobath interval of 50m is generally too small to pick out other features on the seabed.

Seabed image

In contrast to the bathymetric dataset, the seabed image shows the lower slope and basin floor of area E to comprise a series of coalesced lobate features forming an extensive slope apron (Figure 4.50). The lobes have a rounded appearance with apparently abrupt terminations and have previously been interpreted as debris flows (Bulat and Long, 1998). The presence of debris flows at the base of the slope has been suggested from previous studies based upon seismic data (Stoker *et al*, 1991). On the sea-bed image, the morphology of individual lobes shows an elongate body extending up to 16 km from the base of the slope, branching towards the distal end forming smaller lobes with steep terminal slopes. The inter-fingering and overlapping nature of these lobes demonstrates that repeated debris flow events place.

On the lower slope, there are broad areas of low topography between the distal ends of the debris flows of the middle slope described from area D (Figure 4.50). There are also a number of small channels which appear to be superimposed on the debris flow lobes. Some of the channels cross the break of slope at the slope base and feed onto the stacked lobes on the basin floor. Beyond the lobes of the debris flows at the

• Borehole 99/3

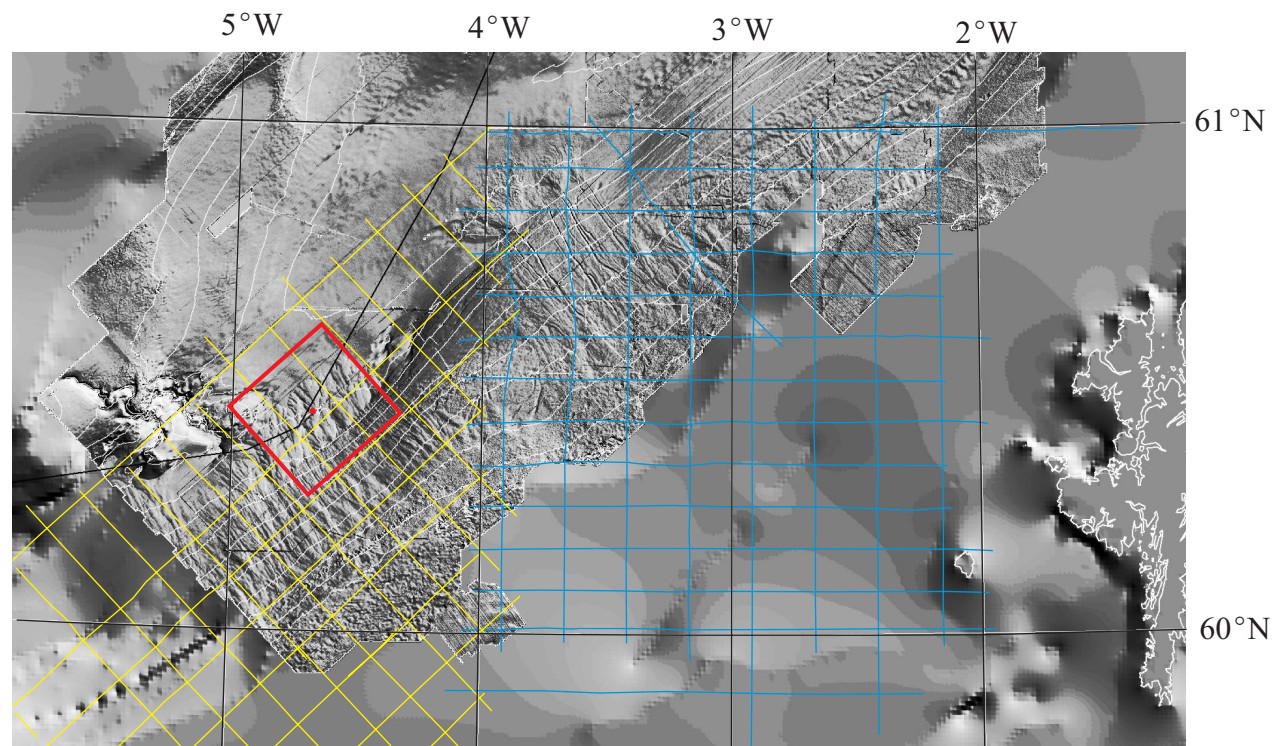
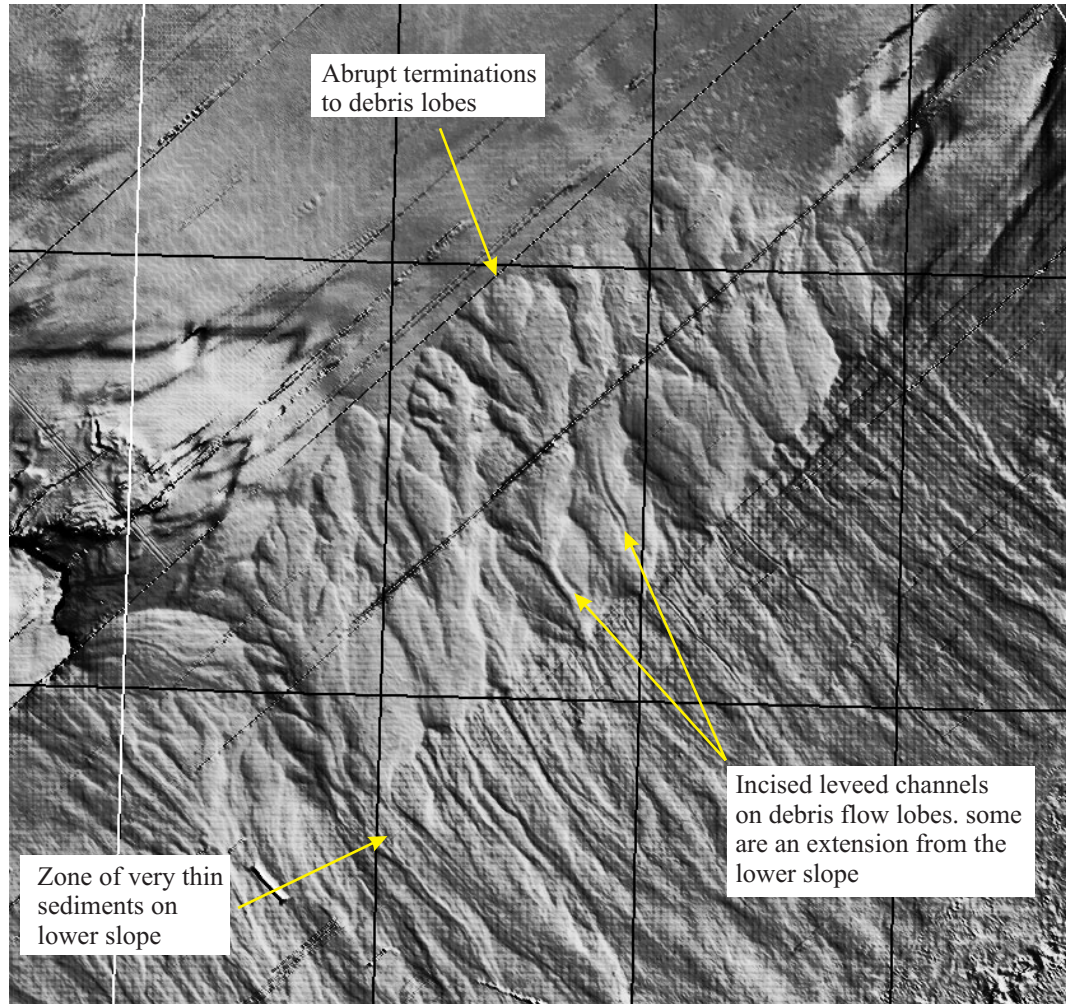


Figure 4.49. Location of Area E at the base of the slope.

Figure 4.50. Enlarged seabed image of the large debris flows from the base of the slope at the north-western end of Transect A.



slope base the sea floor becomes relatively smooth and featureless, with the exception of the scarp features of the Judd Deep to the southwest (Figure 4.50).

Seismic data

Seismic coverage of area E is primarily from the 1983 survey (Figure 4.49), but a survey from the area around borehole 99/3 (Figure 4.53) provides more detailed information and allows the deep water seismic data to be tied in with the core data (Section 4.5.2).

The dip parallel sparker line 83/04-29 (Figures 4.49 and 4.51) is used as a representative line for the northeast of area E. On the lower slope, the gradient increases before the break of slope onto the basin floor from approximately 1° to $>2^\circ$ (Leslie et al, 2003). The sub-surface part of the slope consists of very thin parallel or sub-parallel stratified reflections. The base of the glacial sequence is difficult to identify at this point, and the base of the Otter Bank sequence has been identified from other lines and traced across the survey grid. The very closely spaced continuous reflections are characteristic of this position on the slope, and have been described from other areas along the margin (Masson, 2001; Stoker *et al*, 1998). The exceptionally thin units indicate a greatly reduced thickness of the glacial section which may be due to low rates of sediment delivery and deposition, or prolonged erosion. Evidence from other positions along the margin indicate this zone on the slope is subject to current activity from contour currents travelling parallel to the strike of the slope, reworking pre-existing sediments and leaving sandy contourites as a lag deposit (Stoker *et al*, 1998).

At the base of the slope there is an abrupt change of gradient where the slope base debris flows described from the seabed image spread onto the basin floor (Figure 4.51). The glacial section at this point consists of alternating stratified and transparent or chaotic packages. The acoustic characteristics are very similar to those of the debris flow packages of the upper slope, and have previously been interpreted as such (Stoker, 1999; Davison and Stoker, 2002; Leslie et al, 2003). The pre-Otter Bank glacial sequence is generally composed of smaller debris flow wedges,

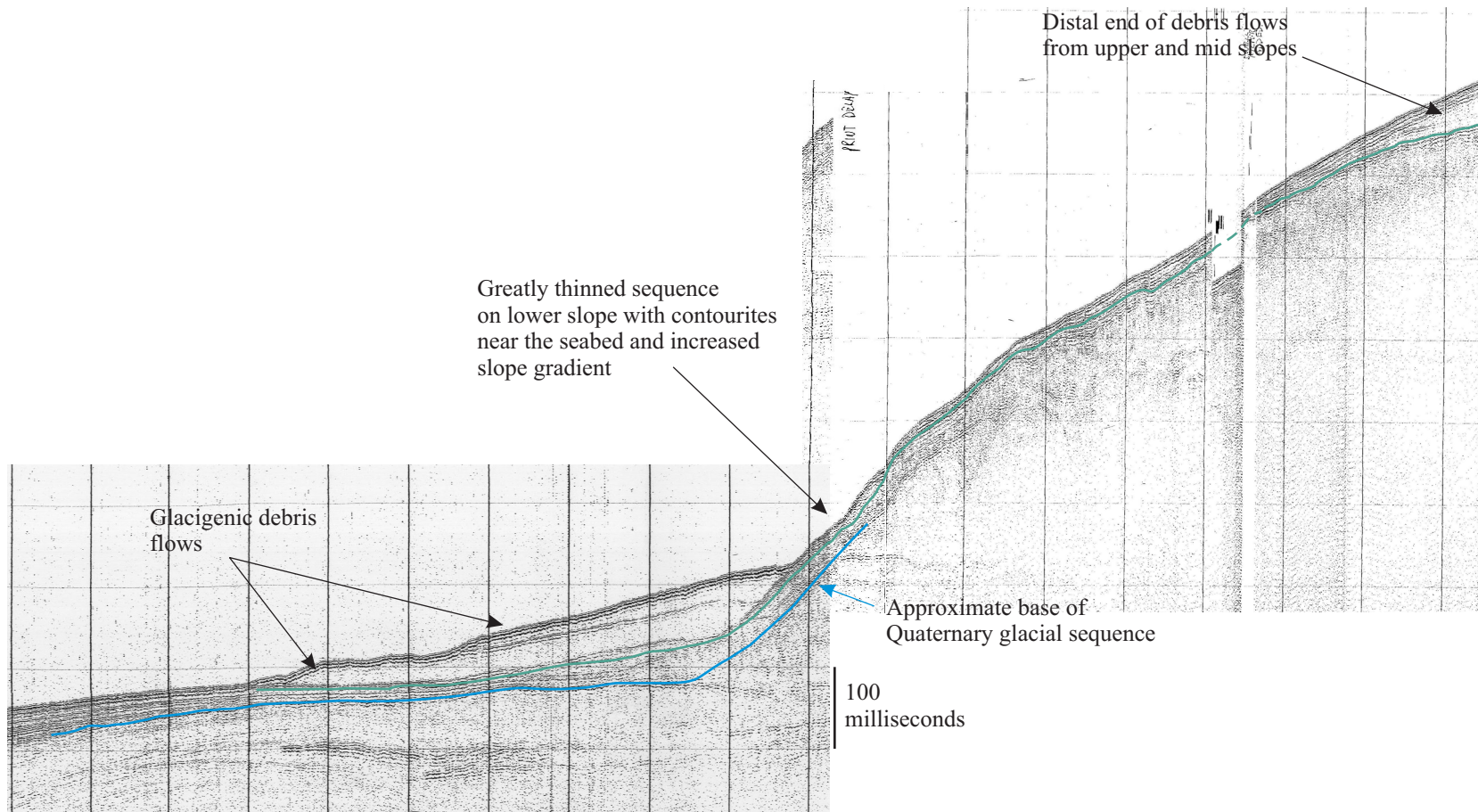


Figure 4.51. Sparker line 83/04-29 showing the base of the Otterbank sequence on the lower slope and base of slope area. A detailed parallel line from the area of borehole 99/3 is shown in Figure 4.54b.

separated by thin stratified units which thin rapidly towards the deeper parts of the basin. The Otter Bank sequence forms a much thicker sequence of debris flow wedges up to 35m thick with a greater lateral extent and longer run-out. The lower part of the Otter Bank sequence is composed of particularly thick debris flows, described in more detail in the borehole survey below. At the edge of the slope-base debris flows, the glacial sequence thins rapidly over a distance of less than 1 km to less than 20ms (~ 15m) and becomes difficult to separate from the older Quaternary deposits (Figure 4.51).

Sparker line 83/04-64 (Figure 4.49) lies parallel to the strike of the slope and shows the lateral extent of the deep water features along the margin (Figure 4.52). The glacial unconformity rises from the northeast end of line 64 towards the southwest, showing long shallow undulations on a kilometre scale. The seabed topography is dominated by the relief of the debris lobes, forming domed structures up to 5 km wide. The overall thickness of the Quaternary glacial section increases from about 25 m in the northeast at the crossover with line 42, to over 100 m thick between lines 30 and 31. It then thins again to about 40 m in the southwest at line 32. The pre-Otter Bank sequence is generally thinner and comprises fewer debris flows than the overlying Otter Bank sequence. The inter-debris flow stratified packages are also more prominent. The architecture of the debris flow packages is similar to those of area D, showing well developed transparent lensoid bodies bounded by parallel and sub-parallel stratified reflections. The debris flows in the Otter Bank sequence form a stacked architecture, with later flows filling the topographic low between two earlier adjacent flows. Debris flow deposition is concentrated mostly between the crossovers with lines 29 and 31. This is especially true of the Otter Bank sequence, although there is a further concentration to the southwest beyond the edge of the study area.

Borehole seismic data

The two seismic sparker profiles presented from the borehole locality form part of a small survey covering an area approximately 5km x 5km with the borehole located near the south-western edge (Figure 4.53).

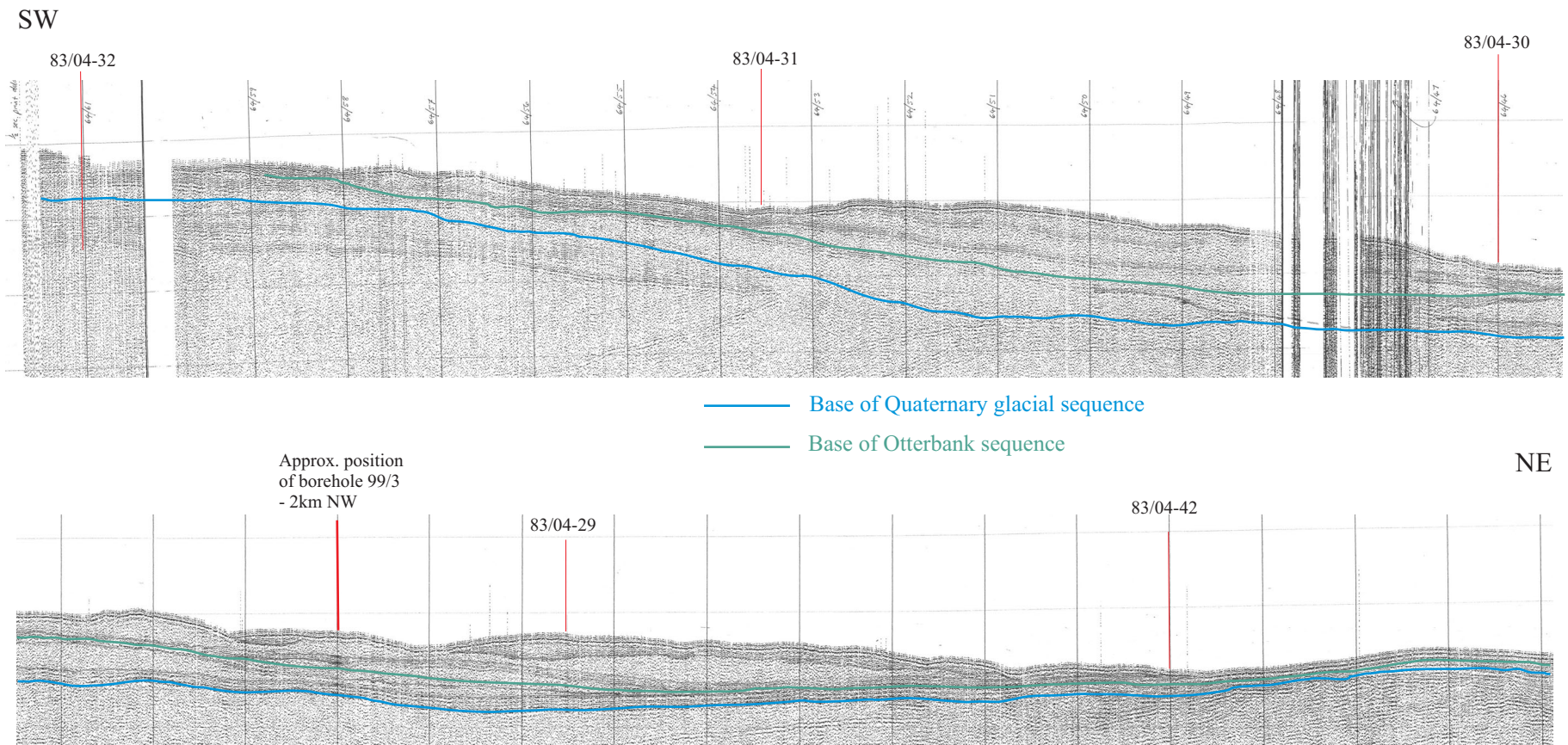
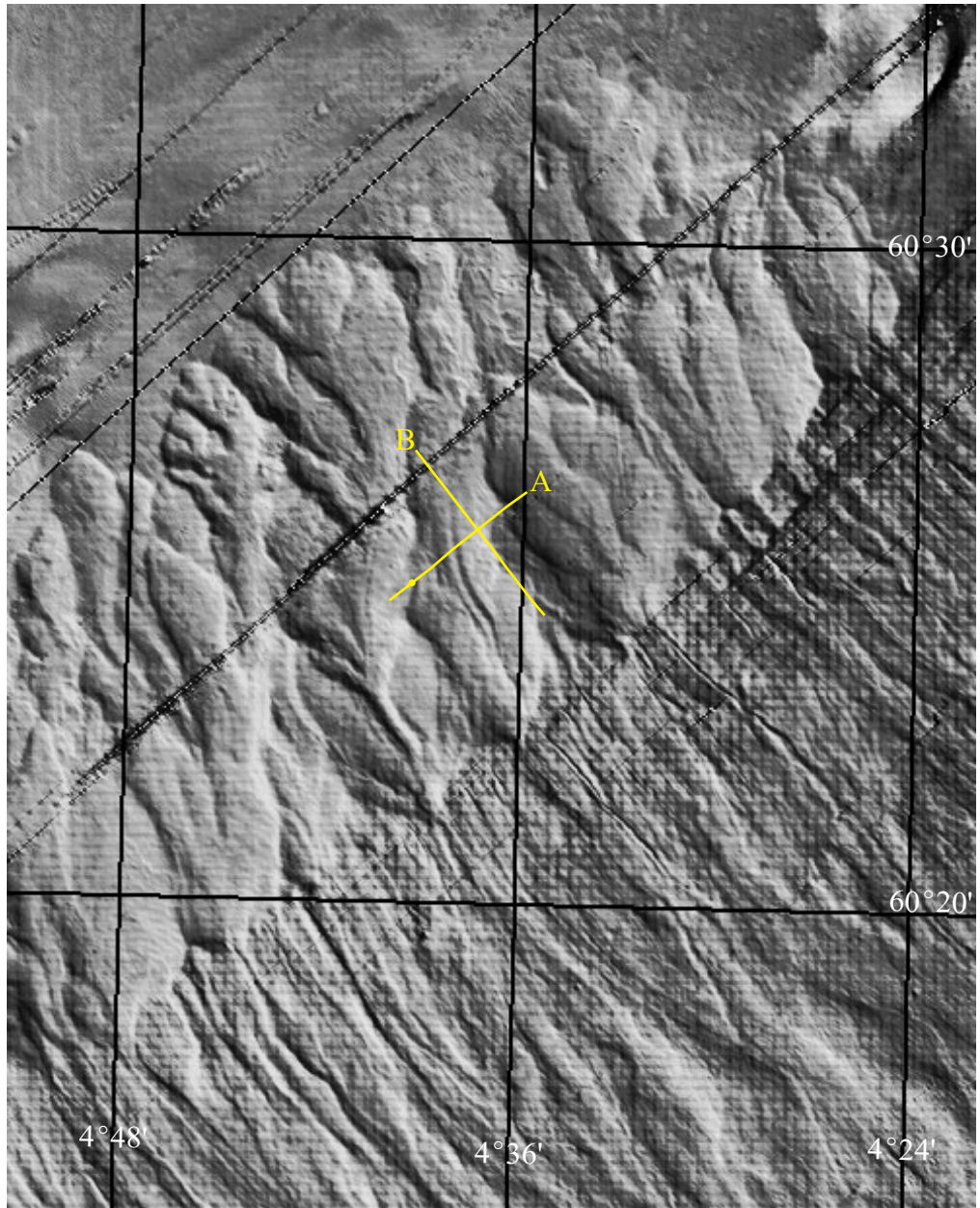


Figure 4.52. Strike-parallel sparker line 83/04-64 from the base of the slope showing stacked debris flows in both glacial sequences. The Otterbank sequence shows a north-eastwards shift in the focusing of debris flow deposition up sequence with two main depocentres.



A - Line 98/01-09 ● Borehole 99/3
 B - Line 98/01-13

Figure 4.53. Location of the detailed seismic survey lines for borehole 99/3. The level of detail is such that individual features from the seabed image can be identified on the seismic data. The prominent leveed channel which crosses line 98/01/09 appears on the seismic data as little more than a slight surface depression (Figure 4.54a), giving an indication of the contrast in detail between the two different data sets.

The seismic data for the borehole location show a prominent reflector at 75ms below the seafloor. This is interpreted as the unconformity at the base of the Quaternary sequence (Hitchen, 1999). Using an average velocity of 1.5 km/s, the Quaternary unconformity lies at 56.25 m below the seabed, which is within the window 54.0 – 56.95m – the base Quaternary – indicated by the core data (see section 4.5.2). A slower velocity is not thought to be credible, as this would be below that of seawater at approximately 1.495 km/s. A higher velocity would place the boundary at a shallower depth, which would conflict with the biostratigraphy from the core (Hitchen, 1999). This interpretation agrees with the regional interpretation, with the Quaternary stack resting upon eroded Eocene strata (Stoker, 1990; Stoker, 1999). This interpreted horizon is also in agreement with the data from line 83/04-64 which was arrived at independently by correlating boundaries from the shelf and slope.

The sequence can be divided into the same stratified and transparent seismic facies seen in the larger scale surveys. The two profiles illustrated in Figures 4.54a and 4.54b, show an alternation of these two basic types. The transparent and chaotic units have a lensoid shape in the strike sections and wedge shape in the dip sections. In some of the lower units there is also limited evidence of discontinuous hyperbolic reflectors. Thickness of the transparent units varies greatly, reaching 17 m at the borehole site and over 30m at fixed point 11/9 on profile 4.54b. The detailed profiles indicate that the debris flows have been accumulating throughout the deposition of the preserved Quaternary sequence.

The parallel and sub-parallel reflections which separate the debris flows represent the general background deposition between the debris flow events. From the seismic data alone it is not possible to determine the nature of these sediments. However, these packages and to a lesser extent individual reflections, are laterally persistent along strike and tend to drape the underlying surface. The thickness of the individual packages varies between 2 m and 12 m, thickening slightly over topographic lows and thinning over the topographic highs.

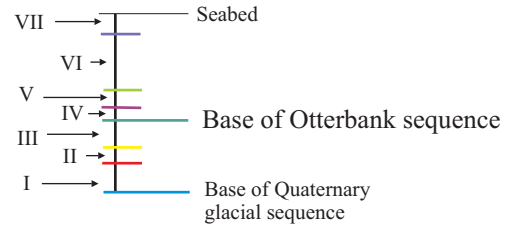
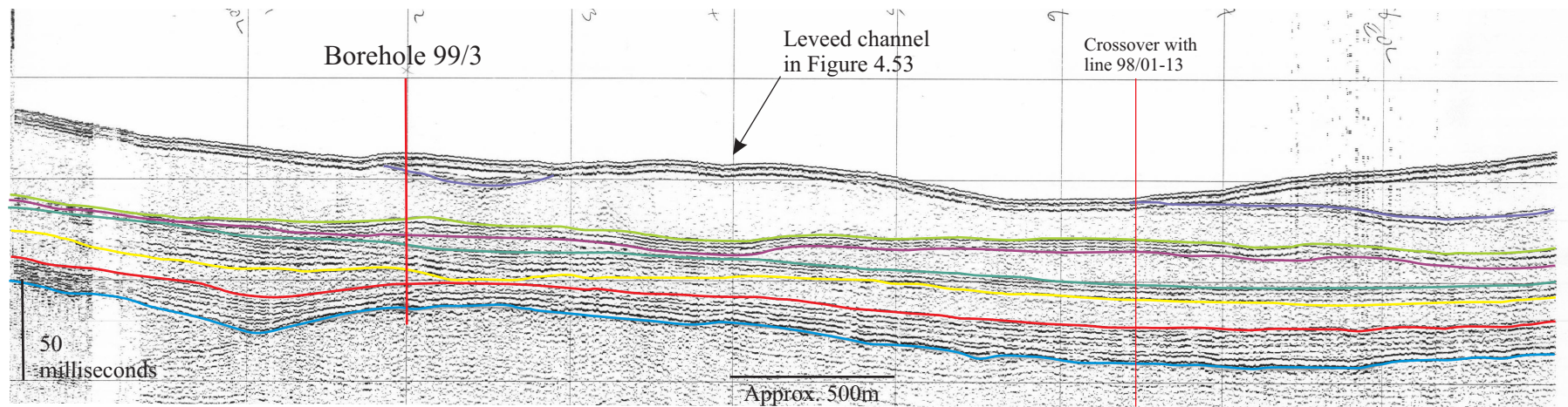


Figure 4.54a. Sparker line 98/01-09 showing a strike parallel section across the debris flow field at the base of the slope. Data shows stacked acoustically transparent debris flows with interbedded acoustically stratified sandy muds. The pre-Otterbank glacial sequence also shows the same style of morphology. Water depth at the borehole site is 983m. Units correspond to those in the borehole log in Figure 4.55.

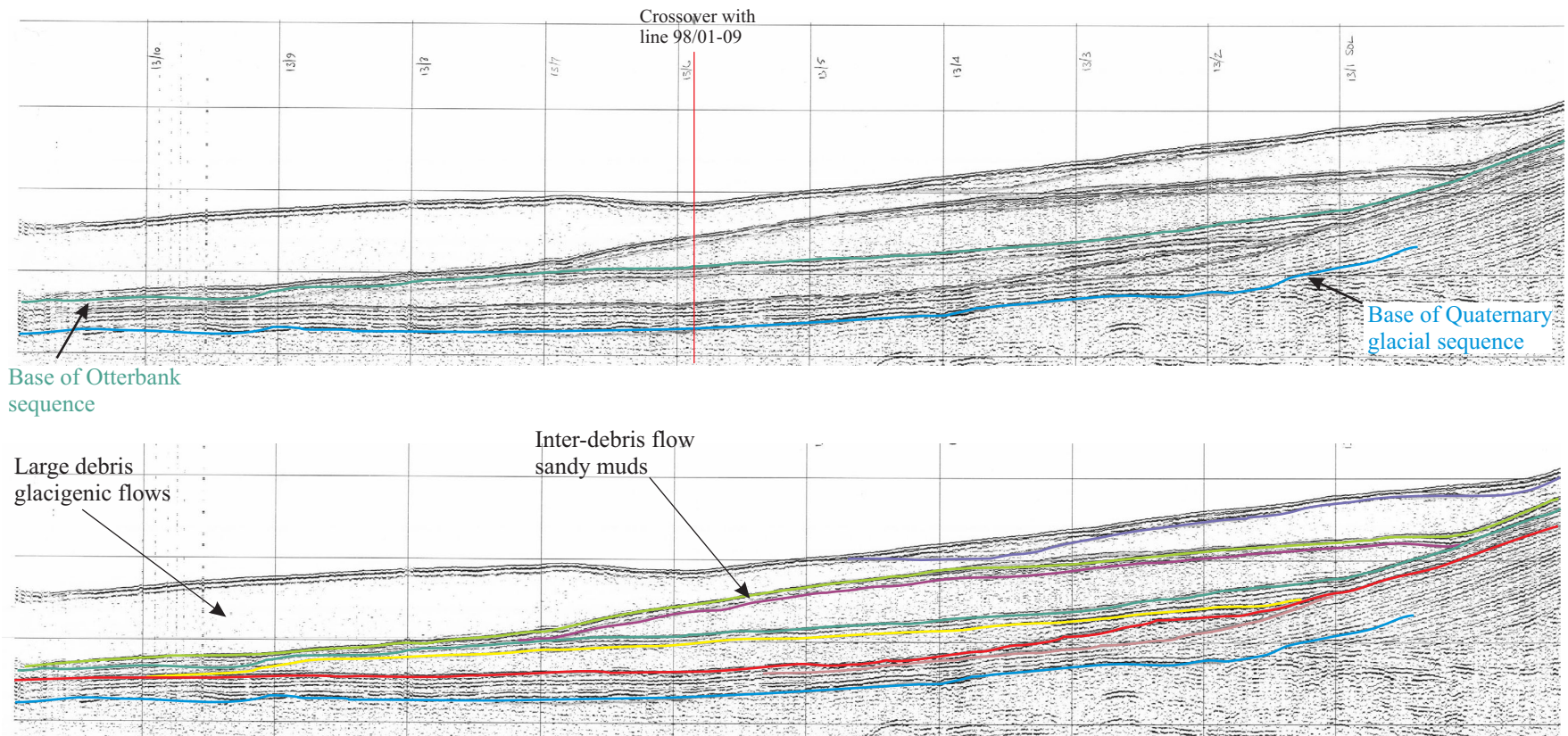


Figure 4.54b. Sparker line 98/01-13 showing a slope parallel section through the Quaternary glacial sequence at the base of slope in the region of borehole 99/3. Seismic units correspond to those defined in the core log in Figure 4.55.

4.5.2 Borehole data

The core from BGS borehole 99/3 is summarised in Figure 4.55 with generalised lithofacies described in Table 4.1. Core recovery was generally not good, with retrieval from above 28 m being poor. However, correlation with the seismic stratigraphy would suggest that all of the main seismic units are represented, although contacts between different lithologies were rarely recovered. Within intervals of no recovery, the boundary separating lithologies is taken as the mid-point between the recovered sections, except where a boundary can be fixed using the seismic data. The core was drilled in separate runs of 3 m or 3.5 m. Unlike other cores in this study, the core from borehole 99/3 had undergone very little post-recovery alteration and oxidation as it was less than 1 year old at the time of examination. Consequently, the colour of recovered sediments is much more meaningful than from other boreholes.

The base of the Quaternary sequence lies at a depth of c.56m below the seabed at the borehole site. This depth is based upon micropalaeontological results from sediments at a depth of 54 m, coinciding with the prominent reflector at 75 ms (see section 4.5.1), which show abundant late Quaternary taxa, and sediments from 56.95m which are dominated by Eocene taxa. Quaternary taxa are absent from sediments recovered from below 54m (Hitchen, 1999). The micropalaeontological data also indicate that the preserved Quaternary section is mid-to-late Pleistocene in age.

Sand and gravel facies

The sand and gravel facies (Table 4.1) was recorded from 3 intervals, the uppermost 20cm, 43.02 – 43.31 m and from approximately 51.50 – 56.25 m. The deepest recovered sediments from 54.0 – 53.50 m show a fining upwards trend from coarse gravel and cobbles to pebbly sand, and are interpreted here as a basal lag deposit which rests upon the unconformity surface. The remainder of the basal sand unit, and those higher in the sequence, are interpreted as lag deposits formed by current winnowing of glacimarine sandy muds. This is supported by the occurrence of sandy muds of glacimarine origin underlying both of the upper sand intervals and by the presence of polar water temperature foraminifera in the sands. Analysis of

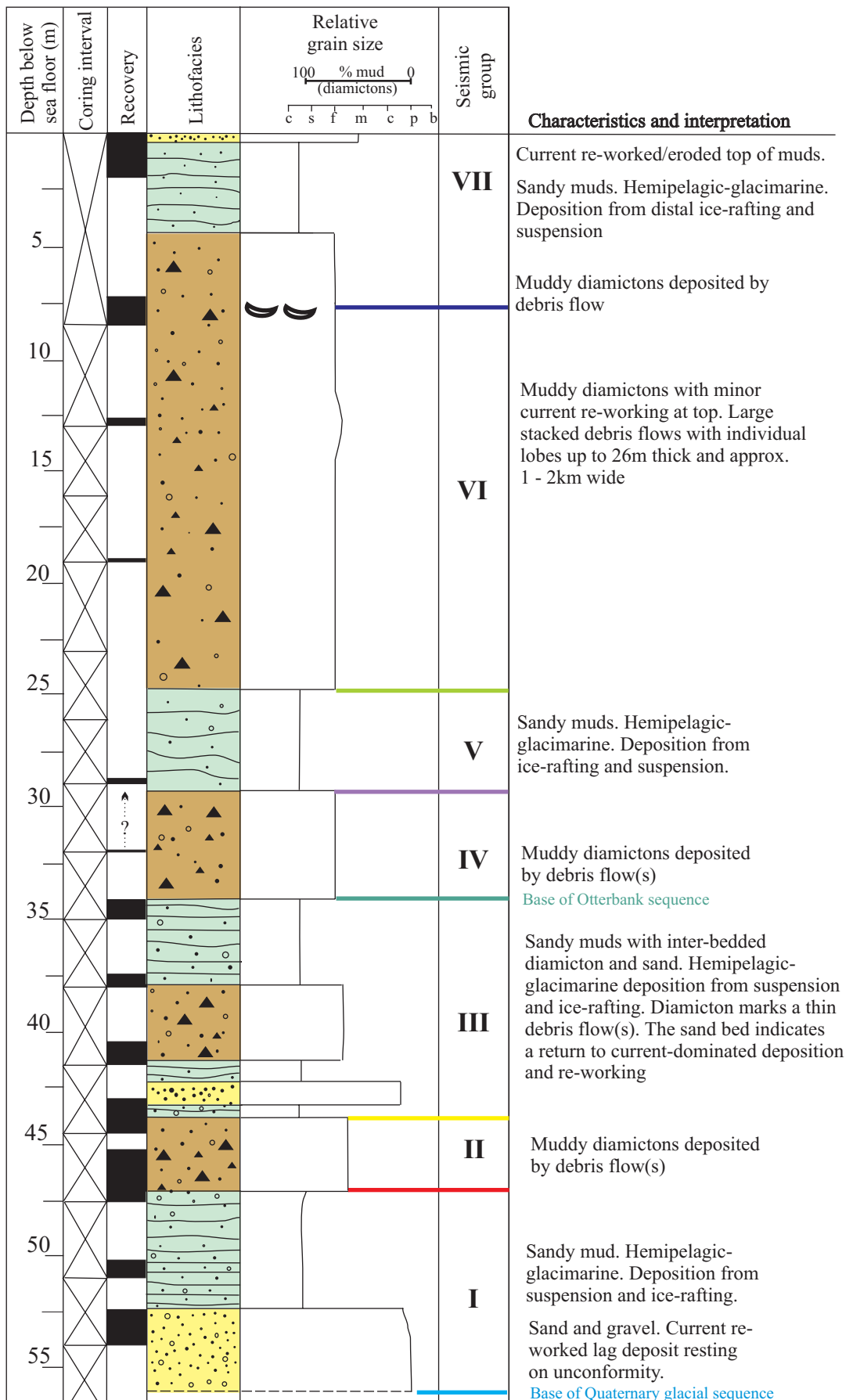


Figure 4.55. Core log from borehole 99/3.

microfossils from the sand and gravel facies indicate substantial variation in the abundance and diversity of foraminifera, including forms from both deep and shallow water environments, and from cold and more temperate water temperatures. The assemblage from the basal sand is dominated by the shallow, cold water species *Cassidulina teteris*, together with the planktonic species *Neogloboquadrina pachyderma* (sinistral) which is abundant (Wilkinson *et al*, 2000). The diversity of environments represented by the microfossils, together with the wide variety of clast lithologies and mineral grains, suggests the sands may be derived from more than one source. A combination of *in-situ* winnowing of glacial marine sandy muds by contour/basin floor currents, and input of sediment from a more distant origin by the same currents, seems the most likely explanation. This is compatible with reported contour current activity from other localities in the Faroe-Shetland Channel (Stoker *et al*, 1993; Akhurst, 1991; Howe *et al*, 2002).

The thin sand and gravel facies at the top of the borehole is interpreted here as having developed since the end of glacial influence, representing the normal bottom conditions for the deep water environment of the present-day Faroe-Shetland Channel. This is important in considering the thickness of the sand and gravel facies between the debris flows at 43.02 – 43.31 m, and the much thicker basal unit below 51.50 m. The implications are that the same conditions existed on at least two previous occasions within the late Pleistocene, and that these conditions were established for a much longer period.

Sandy mud facies

The sandy mud facies (Table 4.1) was recovered from 5 different intervals, mostly below a depth of 29m. The characteristics shown by this facies, grey-green mud-dominated sediments with fine sand and out-sized pebble clasts, are typical of hemipelagic deposits of glacial marine origin. The sand component and the clasts are interpreted as ice-rafted debris, with the mud fraction derived from ice rafting and meltwater plume activity. The interpretation of this facies as hemipelagic-glacial marine is supported by a foraminiferal assemblage dominated by the frequent to abundant occurrence of the planktonic foraminifera *Neogloboquadrina pachyderma*

(sinistral), which is a recognised indicator of cold (Arctic) water temperatures (Ericson, 1959). The absence of lamination or trace fossils within the sandy mud facies is unexpected. Whether internal structure was never present, or was destroyed by post depositional processes is not known. The destruction of any original structure by bioturbation is possible, but this seems unlikely given the apparent scarcity of macrofossils.

Massive diamicton facies.

The compositional and textural features of these diamictons (Table 4.1) are characteristic of subglacial tills and debris flows. Based upon lithology alone, it would be difficult to determine which of these depositional scenarios the massive diamicton represents. However, its position at the base of the slope in almost 1000m of water precludes deposition as a subglacial till. In addition, the indication of large debris flow lobes on the sea-bed image and seismic profiles supports the interpretation of the massive diamicton facies as having been deposited by debris flows.

The Arctic water foraminifera assemblage recovered from the massive diamicton includes the species *Elphidium excavatum calvatum*, *Cassidulina reniforme* and *Haynesina orbiculare*. These species are all benthonic shallow water species generally found on the inner shelf (Wilkinson *et al*, 2000). The presence of these species in a deep water slope-base setting strongly supports both a re-deposited origin for the massive diamicton facies, and also indicates that the debris flow sediments were originally deposited under glacial conditions, prior to their remobilisation down the slope.

The red-brown colour of the massive diamictons appears to form an important identifying characteristic. The finer grained lithologies which form the sandy mud facies are interpreted as being in situ, and are grey-green in colour with little evidence of oxidation. However, the massive diamictons from the debris flow deposits are all red-brown, suggesting the re-mobilisation of the sediments is responsible for the oxidation of the iron compounds within them. This is supported

by the reported original colour of the massive diamictons from the shelf boreholes which were dark grey or grey-brown.

4.5.3 Integration of borehole and seismic data

The seismic data from line 98/01-9 in the vicinity of the borehole is divided into 7 units, consisting of stratified and transparent reflections (Figure 4.54a). The 7 units broadly correspond to the lithological units defined from the core (Davison and Stoker, 2002).

Unit I: 56.25 – 48.50m

The lowermost seismic unit is a combination of the sand and gravel facies, overlain by the sandy mud facies. On the seismic record from line 98/01-9 in figure 4.54a, the two different lithofacies are indistinguishable, and form a single group of parallel and sub-parallel reflectors. The asymmetric thickening of the unit over the topographic low at fixed point 9/1 suggests strike-parallel transport of sediments, indicating contour currents rather than downslope currents were more likely responsible for formation of the sand and gravel facies of Unit I.

The inclusion of cold water Pleistocene foraminifera within the sands of Unit I indicates that glacial conditions were already established in the source region of these sediments and that the pre-glacial Quaternary succession was either removed by erosion, or was never deposited. The well preserved state of the foraminifera tests suggests that they have not been transported a great distance, and indicates a probable source, and hence glacial conditions, within the Faroe-Shetland Channel. The timing of the deposition of the sand and gravel component of Unit I is not known. Persistent erosion and winnowing of the channel floor has probably occurred throughout the Neogene-Quaternary interval.

The overlying sediments of the sandy mud facies represent the onset of glacial conditions, with the delivery of suspended sediment and ice-rafted debris to the area. Deposition of the fine-grained sandy mud facies also indicates the slowing down or

halting of the current regime responsible for forming the underlying sand and gravel facies.

Unit II: 48.50 – 43.50m

Unit II is interpreted as a debris flow, derived from glacimarine sediments originally deposited higher up the continental slope. The interpretation as a debris flow is based upon the seismic architecture and lithological character. The position on the lower slope at a water depth of almost 1000 m excludes the possibility that Unit II is a subglacial till. The presence of cold water shelf microfossils in the unit indicates that glacial conditions prevailed at the time the original sediment was deposited, prior to re-deposition by debris flow activity. This suggests a glacimarine origin for these sediments on the shelf, although this interpretation is not conclusive.

Unit III: 43.50 – 34.00m

Unit III contains sediments of all three lithofacies and represents changing depositional, and possibly environmental conditions. The acoustically stratified reflectors hide the lithological changes seen in the core. Generally, Unit III is interpreted as the product of suspension sedimentation and iceberg rafting. However, the thin sand unit from 43.03m – 43.31m represents a return to current reworking. This may be similar to the current regime which formed the sands at the base of Unit I. Whether this change represents a climatic warming and change to non-glacial conditions, or is a local current effect is not known.

Within the core recovered from Unit III there is also a thin bed of massive diamicton which is interpreted as a thin debris flow or the distal end of a larger flow. This interpretation is based solely upon the lithological similarity to Unit II as it does not appear on the seismic profiles and is therefore assumed to be below the resolution of the seismic survey. The inclusion of a debris flow within Unit III has important implications for the interpretation of acoustically stratified units, showing that without the control provided by the core, the presence of debris flows would not have been detected.

Unit IV: 34.00 – 29.20m

Acoustically, Unit IV has the same characteristics as Unit II and is interpreted as a debris flow. However, sediments recovered from this interval are from the sandy mud facies which conflicts with the interpretation. This apparent conflict seems to be the result of poor core recovery and the way in which the core was measured from the base of each coring run. If the core from 31.78 –32.00 m interval actually came from the top of the core run rather than the base, the conflict is resolved, suggesting that no sediments were in fact recovered from Unit IV. This solution is supported by the presence of a large (9cm) cobble at the base of the recovered section which may have prevented further core recovery.

The base of unit IV marks the base of the Otter Bank sequence when correlated with the seismic data from the larger regional survey.

Unit V: 29.20 – 24.50m

Unit V is interpreted as a series of ice-distal glacimarine muds deposited by suspension settling and iceberg rafting. Along slope to the NE there is some evidence of slight thickening over topographic low points, as seen in Unit I. This suggests there may have been small scale current activity during the deposition of Unit V, which resulted in minor redistribution of sediment and localised thickening. Alternatively, this may be the distal end of a debris flow from higher up the slope.

Unit VI: 24.50 – 7.34m

Unit VI is of a different scale to the preceding units, comprising 30% of the entire Quaternary sequence. The acoustic transparency, architecture and massive diamicton lithology indicate that this unit was deposited by debris flow. When the unit is traced laterally and upslope on the seismic profiles, it is evident that it comprises a series of large, stacked debris flows. The seismic profiles also illustrate the considerable surface topography (~18m) created by the individual flow lobes. This topography appears to be largely responsible for the present day surface morphology of the coalesced debris flows which form the slope apron observed on the seabed image in Figure 4.50. The thickness of the debris flows indicates that there must have been a

considerable build up of sediment higher up the slope or at the shelf edge to provide sufficient source material to supply the process.

Unit VII: 7.34m – seabed

Unit VII may be a thin debris flow of the same structureless lensoid shape as Units II and IV, but of a smaller scale. Alternatively it may be the distal end of a larger flow, which is perhaps more likely given the position at the top of a series of large stacked debris flows. The thin concentration of shell debris from the base of this unit probably marks a brief hiatus in deposition, during which macrofauna colonisation occurred prior to emplacement of the debris flow. Alternatively it is possible that the shell horizon is a lag deposit, formed during a period of increased current activity. However, the low abundance of macrofauna in the sandy muds generally suggests that this is less likely. The uppermost section of Unit VII is lost in the seabed reflections, but the core evidence indicates a return to distal glacial marine deposition from suspension followed by the re-establishment of a current re-working regime which continues to the present (Stoker, 1991).

4.5.4 Summary and interpretation of the lower slope and basin floor

The deep water glacial sequence of area E is dominated by the debris flow deposits which form the main topographical features of the sea floor, and which comprise the majority of the sediment record. The older pre-Otter Bank sequence shows the same alternation of debris flows and background sandy muds as the overlying Otter Bank sequence, suggesting this is typical of the deep water glacial environment in the Faroe-Shetland region. However, the evidence from line 83/04-64 also shows that these features are not laterally continuous, and that area E formed a focus for debris flow deposition which decreases to the northeast and southwest. This contrasts with area F to the north east, which is described below (section 4.6), where debris flows on the lower slope and basin floor are all but absent. The greater thickness of the Otter Bank sequence and dominance of debris flows compared to the pre-Otter Bank sequence is explained by the difference in the slope angles at the shelf edge and upper slope. The process of progradation at the shelf edge causes the steepening of the upper slope. The difference of slope profiles

shows that at the time of deposition of the pre-Otter Bank sequence, the sediment was retained on the upper slope due to lower slope angles. Deposition of the Otter Bank sequence increased the angle of the upper slope leading to greater instability, resulting in the generation of more debris flows than in the earlier sequence. This scenario follows the model of Boulton (1990).

The data from the detailed section of core 99/3 in Figure 4.56 shows an upwardly increasing sand and granule content in the sandy mud facies prior to the emplacement of a debris flow. This indicates the source of the sandy mud facies sediment is advancing, and area E is effectively becoming more proximal to the ice margin. This would also suggest that debris flow generation occurred at a time when the ice was in an advanced state, probably at the shelf edge, and at the time of maximum sediment delivery. Whilst the actual grain size of the sediment is not a reliable indicator of advancing ice or depositional environment, the coarsening upwards trend is.

Integration of the borehole seismic with the core indicates that thin debris flows are below the seismic resolution, as in the section between 34m and 44m which forms parallel stratified reflections but contains a debris flow diamict. This suggests that thin debris flow units, probably at the distal end of the flow, may extend beyond the apparent end of the feature in the seismic sections. This concept is illustrated in Figure 4.57.

The Otter Bank sequence on the borehole seismic profile (Figure 4.53a) shows two major phases of debris flow development, suggesting that input of glacial sediments to the lower slope peaked on two occasions. Two main phases of activity at the slope base is consistent with the two major seismic packages on the upper slope, indicating these two phases are probably laterally equivalent. The boundary between these two phases is interpreted as the same boundary which separates units OB1 and OB2 on the outer shelf in Area C.

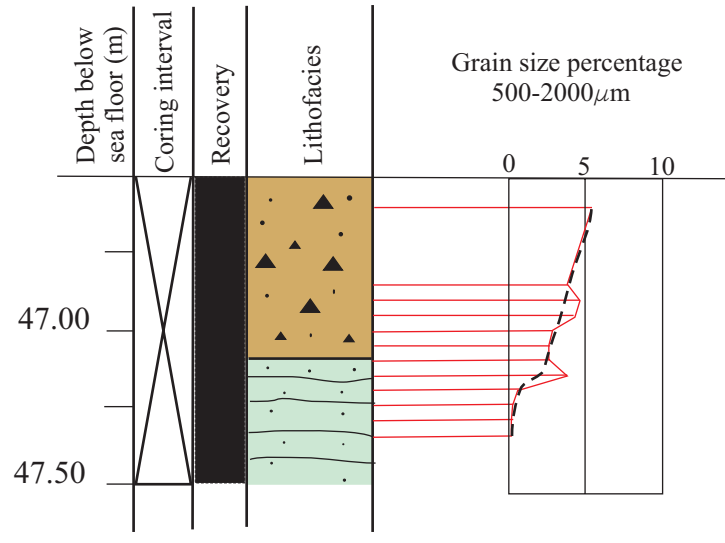


Figure 4.56. Detailed section of core from borehole 99/3 sampled for grain size analysis at 5 cm intervals. The sampled section crosses the boundary between sandy mud facies and massive diamicton facies, showing a change from background glacimarine deposition to a debris flow. The lithological boundary is placed at the change of colour in the core from grey-green to red-brown, which defines all the lower sequence debris flows. The sandy mud shows an increase in the coarse component prior to the emplacement of the debris flow. This is interpreted as indicating an ice sheet advancing to the shelf edge, delivering more suspended sediment to the borehole location and increasing the IRD input. The debris flow is emplaced at a time when the ice sheet is at, or close to the shelf edge and IRD and suspended sediment input are at a maximum.

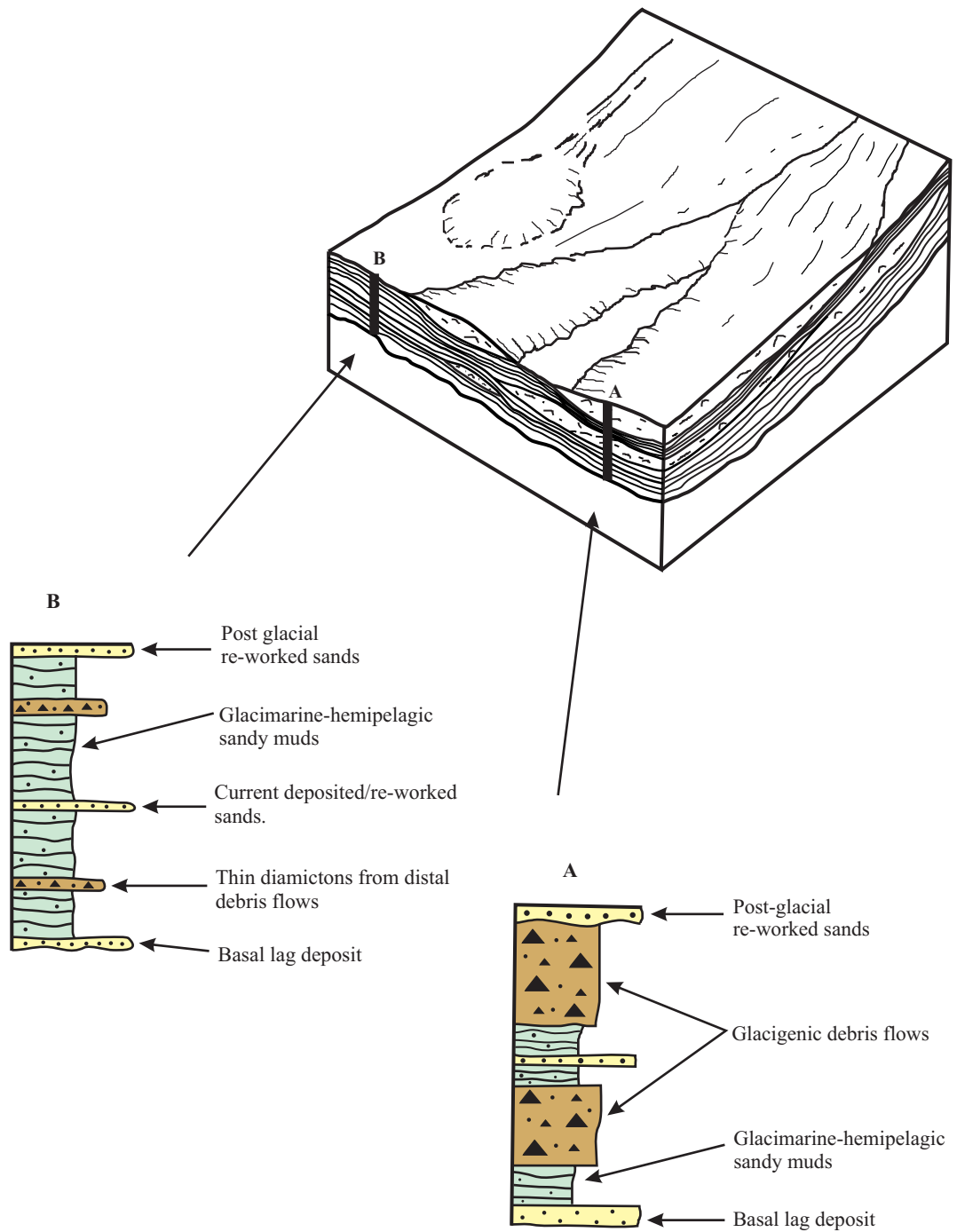


Figure 4.57. Schematic representation of the base-of-slope zone in Area E. Log A represents the position occupied by borehole 99/3. Log B represents the deep-water environment off-line of the debris flows (modified from Davison and Stoker, 2002).

4.6 Area F: Slope and basin floor

The area of the slope to the north-east of transect A shows a different range of features to those in areas D and E. These additional features are briefly described and interpreted below. The implications for margin reconstruction are included in the summary of transect A at the end of this chapter, providing an illustration of the lateral variability of the slope setting.

4.6.1 Seabed morphology and seismic data

Bathymetry

A cursory examination of the bathymetry of area F shows a large embayment, the Foula Bight (Figure 4.58), which modifies the line of the shelf edge forming an area of deeper water. Unfortunately the seabed image does not cover the embayment on the shelf (Figure 4.59), preventing any detailed analysis of seabed features. The bathymetry of the upper slope in area F indicates a slightly reduced slope angle when compared to the area south-west of the Foula Bight. This appears to be a function of the Foula Bight embayment deflecting the shelf break approximately 10 km shelf-wards.

Seabed image

The seabed image of the upper slope in Figures 4.59 and 4.60 shows large elongate lobes on the middle and upper slope which are very similar to those in area D to the south-west. The lobes have been previously interpreted as debris flows by Leslie *et al* (2003). An enlarged view of the image shows the debris flows on the upper to mid slope tend to be wider than those seen in area D; about 3-5 km rather than 2-3 km (Figure 4.60). The degree of separation between each lobe is also larger. The most obvious difference in the characteristics of the debris flows in area F is the much shorter run out than those to the south west. The flows in area F appear to terminate on the mid slope, with very few reaching below 700m, unlike those in area D which reach the basin floor. The seafloor in the area of the debris flows shows large numbers of roughly circular depressions between 10m and 200m in diameter. These are interpreted by Leslie *et al* (2003) as possible pockmarks, formed by the post-

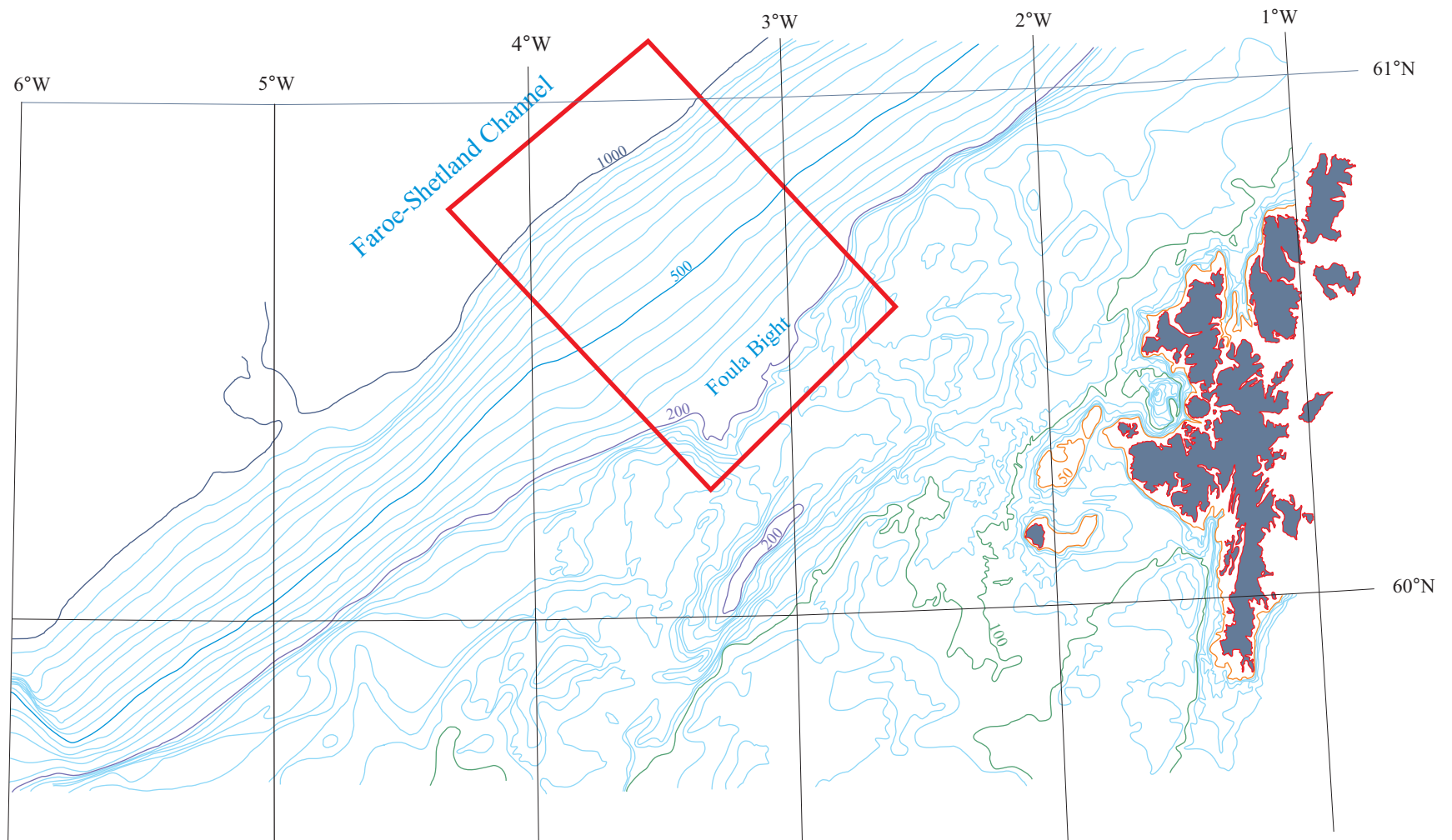


Figure 4.58. Location of area F, combining the deeper shelf area of the Foulca Bight and the alternative slope and basin floor area.

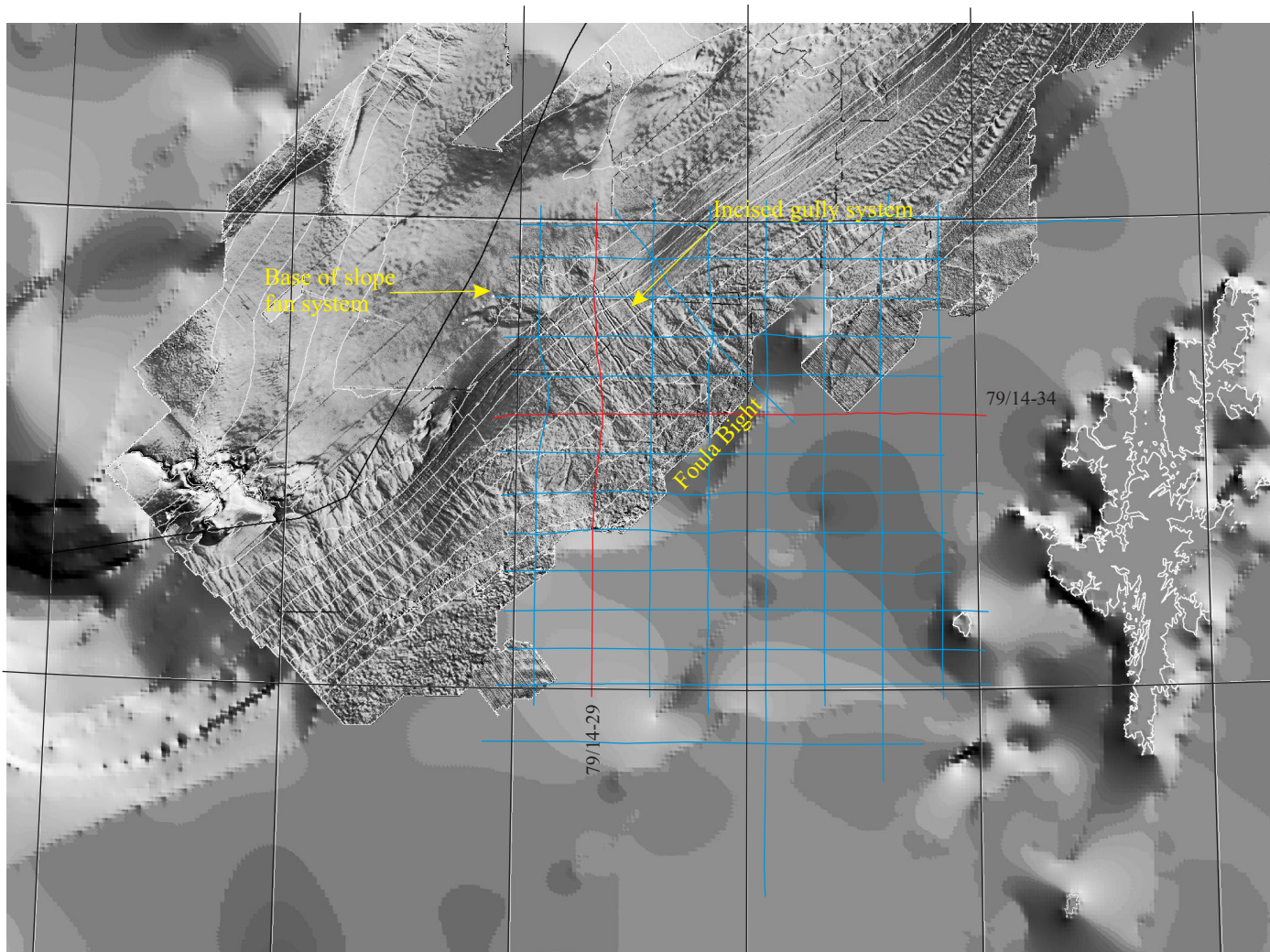


Figure 4.59. Seabed image showing the alternative slope of area F (enlarged in Figure 4.60).

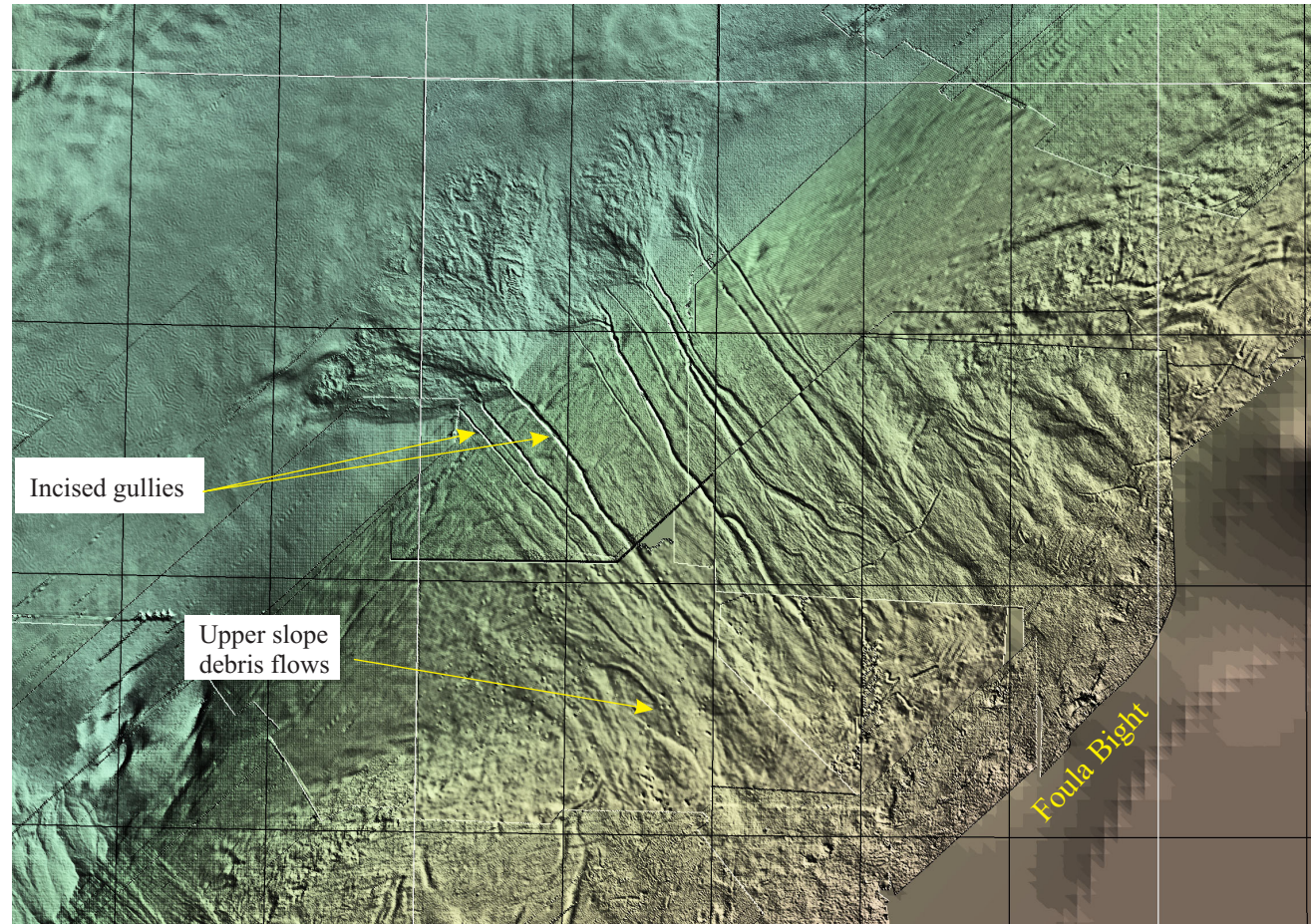


Figure 4.60. Seabed image of Area F, showing broad debris flow lobes on the upper slope emanating from the Foula Bight. The long sub-parallel channels are incised into the surface of the debris flows and feed into the large fan system at the base of the slope. Leslie et al (2003) suggest that the fans are draped by acoustically stratified sediment and that the fans are no longer active.

depositional escape of gas or fluid, although there is uncertainty regarding the validity of the data.

In addition to the debris flows on the slope in area F, the slope shows a series of incised gullies which originate in the debris flows on the mid to upper slope and extend as far as the base of the slope. The gullies show a small degree of sinuosity (Leslie *et al*, 2003), but in general are relatively straight and show very few bifurcations or confluences. The width of the gullies varies from 10's of meters up to 300m with a depth of between 5m and 20m (Kenyon, 1987). The gradient on the mid to lower slope in this region is $< 2^\circ$, and possibly $< 1^\circ$. The gullies have been described by various authors in the past (Leslie *et al*, 2003; Mason, 2001; Damuth and Olson, 1993; Kenyon, 1987), but the mode of their formation has remained uncertain. At the base of the slope, the gullies feed into a series of stacked and coalesced fans which show a series of cone-shaped lobes spreading onto the floor of the Faroe-Shetland Channel.

Seismic data

Seismic data from Area F lies at the boundary of two different surveys, the 1979 and the 1985 surveys. Unfortunately, the 1979 data, which covers most of the channel system lies at an oblique angle to the slope, giving a misleading impression of the slope angle and architecture of some features.

In general, the acoustic character of sediments from the top of the slope differs little from other areas along the margin. The upper slope comprises prograding stacked packages of chaotic to transparent reflections with an elongate lensoid shape. These packages are interpreted as debris flows, showing the same characteristics as those in area D. However, the thickness of the debris flow packages on the upper slope in area F is considerably less than in other areas. The total glacial thickness on the upper slope reaches 60 – 80m, with the Otter Bank sequence accounting for 40 – 50m. The number of major units within each sequence appears to be the same, indicating that it is individual debris flows which are thinner rather than there being fewer of them.

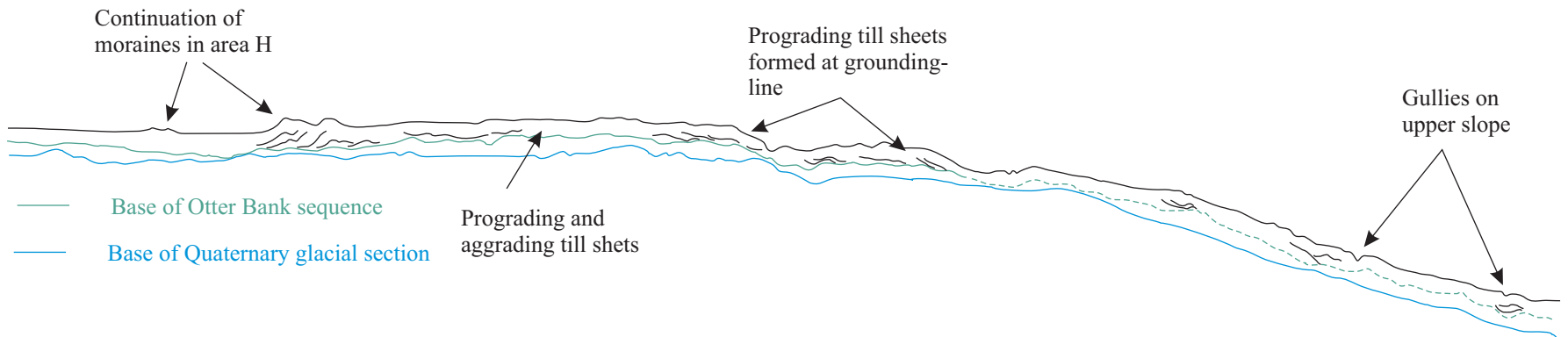
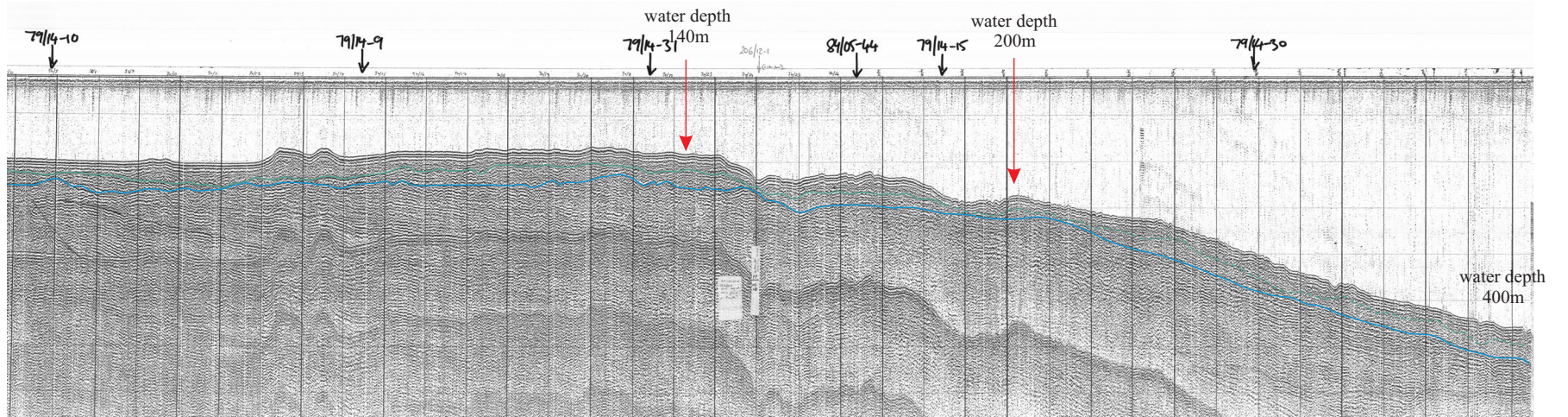


Figure 4.61. Sparker line 79/14-34 showing the outer shelf and upper slope within the Foula Bight. The debris flows at the shelf break and on the upper slope are generally thinner and less well defined than those within area D to the southwest.

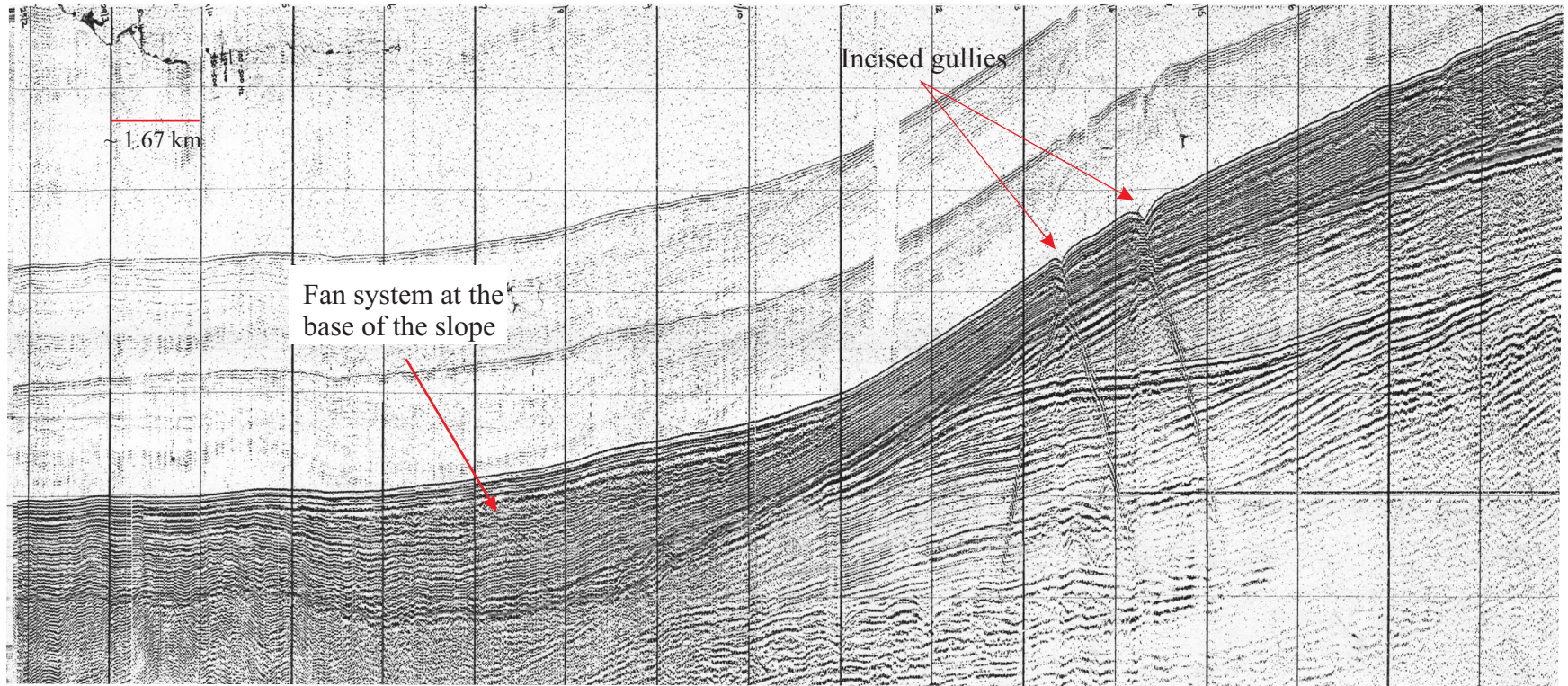


Figure 4.62. Sparker line 79/14-29 showing stacked parallel stratified reflections of the slope base fan system. The position of the Otter Bank base unconformity and the base of the Quaternary glacial section is uncertain due to the very close spacing of reflections and the lack of continuity of the fan system with the other depositional systems at the slope base.

Data from sparker profile 79/14-34 (Figures 4.59 and 4.61) includes an oblique section through the Foula Bight embayment. The profile shows a stepped shelf edge over a distance of about 7 km, from the top of the shelf, at a depth of about 140m, to the true top of the slope, at about 200m. The area between these points forms a shelf-ward sloping zone approximately 20m deeper on the shelf-ward side than on the slope-ward side. This zone is occupied by slightly mounded prograding reflections, the mounded appearance probably being due to aggradation; unfortunately the detail is obscured by the seabed pulse.

On the mid to lower slope, the seabed topography becomes much smoother and the seismic packages much thinner. The lensoid shapes of the debris flows virtually disappear downslope, with the section dominated by parallel and sub-parallel stratified reflections. The glacial section, and the division between the older sequence and the Otter Bank sequence, become difficult to distinguish from the pre-glacial sediments beneath. The first of the parallel gullies described from the seabed image appears as a small incised v-shape in the seabed at a depth of about 400m (Figure 4.61)

Line 79/14-29 (Figure 4.62) shows the gullies in profile, where they appear incised into closely spaced, sub-parallel stratified reflections on the lower slope. Leslie *et al* (2003), suggests some of the gullies have small levees developed to the sides, which is also evident on some seismic lines and the seabed image. These appear to be symmetrically developed on either side of the channels, but resolution is too low to be certain.

At the base of the slope, the seismic data (Figure 4.62) shows a thick sequence of thin, stratified sub-parallel, reflections forming a broad lens-shape body over 100m thick. The stratified units thicken slightly upwards, with a lensoid body of less structured reflections near to the seabed and extending approximately 5-6 km from the slope base. These units form the large fan structures observed on the seabed image and described above. Unpublished data from the WFA indicates that the sediments forming the fans are sandier than those at the seabed on the slope. This

could indicate the fans are derived from a mud-depleted source, or the down-slope currents which generated the channels and transported the sediments were too strong to allow mud deposition on the fans. Alternatively, the sandy fans may have formed in a setting which was under the influence of currents flowing along the floor of the Faroe-Shetland Channel, which retained the mud fraction of the sediments deposited on the fans in suspension.

Unfortunately, because of the very fine structure of the sub-surface on the lower slope and basin floor, it is not possible to correlate the glacial unconformity from the upper slope with horizons in the subsurface of the fans.

4.6.2 Interpretation of the slope and basin floor in area F

The interpretation of area F is dependent on an overview of the whole area. The aggrading and prograding units at the edge of the shelf and on the Foula Bight “step”, together with the debris flows on the upper slope, indicate that ice advance reached the shelf edge. However, features at the shelf edge show a greater degree of aggradation than those described from area C, even though the water depth at which many of them occur in area F is shallower. This suggests the ice depositing the shelf edge sediments in this part of the Foula Bight was very buoyant, and possibly close to floating, in which case, it must have been thinner than the ice grounded in area C. This has important implications for the parts of the Foula Bight to the south west of line 79/14-34 where the water is up to 40m deeper (Figure 4.63). Although the actual water depth is not known at the time of maximum ice advance, the modern depth can be used as an upper limit. In the deeper areas it is very likely a thinned ice sheet would break up quickly on reaching deeper water, transporting away much of the debris to be deposited elsewhere as IRD. This would not only account for the thinner deposits on the shelf, but also the reduced delivery of sediment to the shelf edge.

The reduced thickness of the shelf edge – upper slope wedge, and the smaller number of debris flows visible at the seabed, suggest the supply of sediment to the upper slope was not as great as in areas C and D to the south-west. The greater width

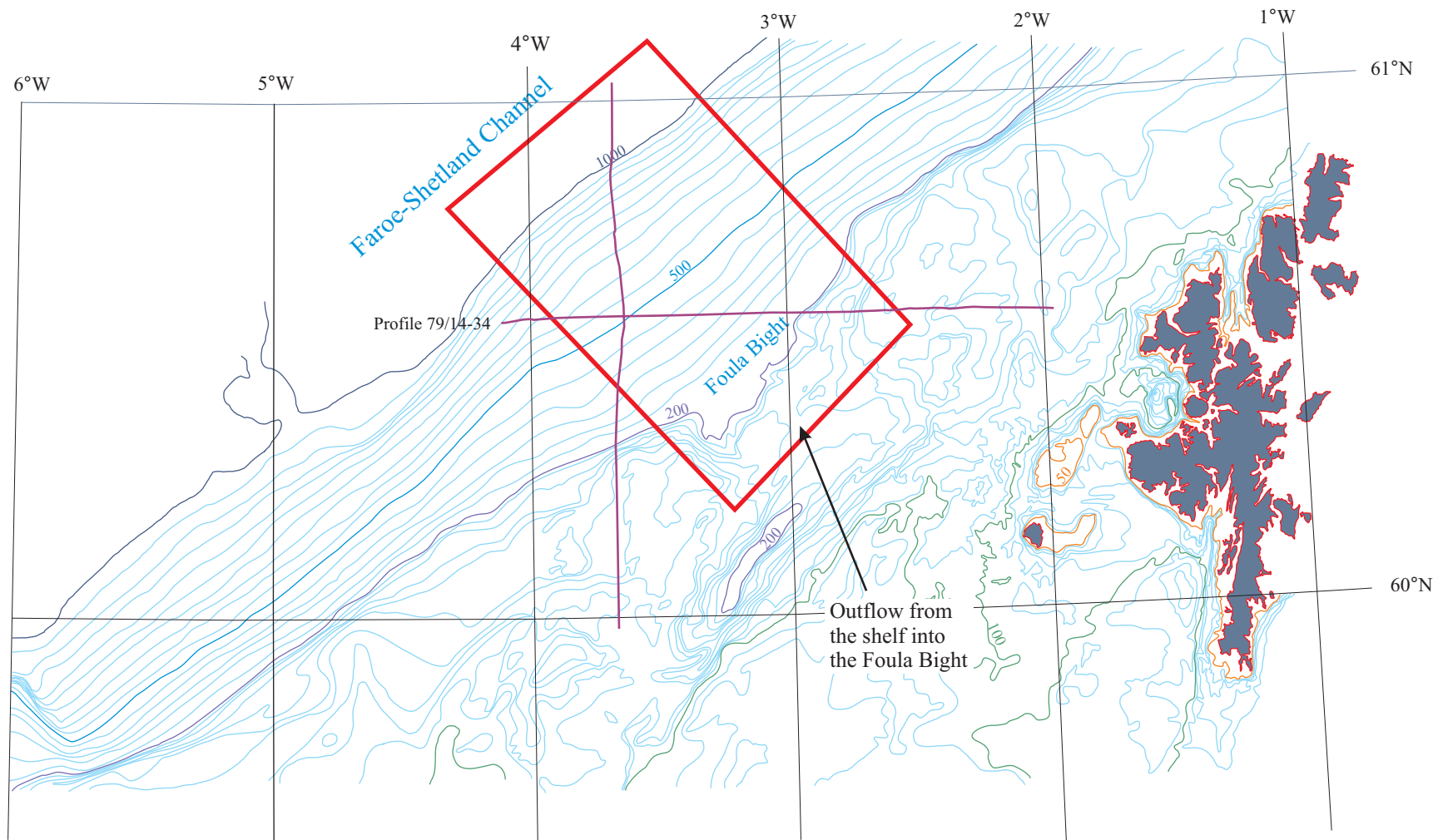


Figure 4.63. Position of seismic profiles relative to the Froua Bight, showing how the bathymetry deepens to the south west of the profile and the proposed outflow pathway from the shelf.

of the debris flows on the slope in area F appears to be a function of their low numbers, with individual flows spreading out laterally due to the lack of confining sea-floor topography. In area D, individual flows infill the areas between earlier flows, funnelling them down the slope. These interpretations support the interpretation of the outer shelf in the Foula Bight area being occupied by thin, unstable ice.

The gully system on the mid to lower slope (Figure 4.60) resembles similar features described from the slope of Northern California (Spinelli and Field, 2001), where an extensive system of shallow gullies is developed. The Californian example shows gully development was most prolific during falling sea-levels prior to the LGM. This scenario does not fit the gullies described in area F, which are incised into glacial debris flow sediments deposited at the LGM when ice was at the shelf edge. Therefore the gully system of area F must post date the LGM. However, the morphological similarities to the Californian examples, including the apparent absence of a connecting drainage system on the shelf, provides indicators to possible modes of formation. The overall morphology shown by the seabed image closely resembles the modelled characteristics of a sediment starved margin from a glacial setting (Dowdeswell *et al*, 1996).

The data above shows that the margin was obviously not sediment starved during all of the Otter Bank age glaciation. However, the differences highlighted above indicate that sediment delivery to this part of the margin was greatly reduced compared to the areas further to the south west..

The gully system on the slope is developed on top of the glacial debris flows and has also modified part of the fan system at the base of the slope (Figure 4.60), indicating that at least some of the gullies post-date the fans. The fans could be relict structures from a pre-glacial low-stand sea-level, but this would not explain why they only occur on this part of the margin. A more likely interpretation is that they are related to processes on the shelf within an ice-free Foula Bight whilst the remainder of the shelf was still ice-covered. The bathymetry of the Foula Bight area (Figure

4.63) shows a deeper pathway at a depth of 170m from the mid shelf – Papa Basin area. If ice decay in the Foula Bight was rapid due to deeper water and thin ice, this may have formed an early pathway for cold, dense meltwater draining as a sub-glacial – seafloor system. Transport of sediments to the shelf edge by meltwater bottom currents could have provided a supply of coarser sediment to the slope, feeding the gully system and depositing the sandy fan system at the base of the slope

The possibility that gullies have developed as a late-glacial feature in response to ice sheet decay and cold bottom currents from the shelf is supported by the presence of buried v-shaped gullies developed at the top of the pre-Otter Bank glacial section in area D. That surface represents an unconformity developed between two major phases of glacial activity and which can be traced to a re-worked sandy sediment unit on the shelf.

The fans on the basin floor are interpreted by Leslie et al (2003) as being relict structures, evidenced by the acoustically well-stratified sediments overlying them on the seismic data. These may represent a waning phase of deposition on the fans, or they may be overbank deposits from channel systems on the fans which they closely resemble (Walker, 1992). A simple draping of sediments from suspension settling is unlikely to show the observed degree of lateral thinning.

4.7 Borehole correlation and summary of glacial sedimentary processes

The different provinces examined in transect A from the inner shelf to the deep water basin floor show a wide variety of processes and features related to the presence of ice sheets on the shelf. The characterisation of the Otter Bank sequence in Transect A relies upon construction of a correlation along the length of the whole transect.

4.7.1 Correlation of boreholes

The proposed correlation of the Otter Bank sequence within the boreholes, presented in Figure 4.64, is based partly upon lithological correlation and partly upon seismic data. On the inner and mid shelf, the seismic is generally of poor quality and the thin sequence often masked by the seabed pulse. On the outer shelf, the seismic data is much improved but is not always accompanied by core data. Poor biostratigraphic control precludes absolute certainty in these correlations.

The key to the correlation of the shelf and slope is borehole 82/10 (Figure 4.22). The basal unconformity of the Otter Bank sequence at this site is defined in line 79/14-26 (Figure 4.19) by a reflector which truncates reflections of the lower part of the glacial sequence (Holmes, 1991). In the borehole, the unconformity is interpreted as the boundary between the stratified diamicton and the overlying massive diamicton (Figures 4.19 and 4.64). The defining reflector can be traced around the survey grids to borehole 82/11 on the outermost shelf and onto the slope. Correlation of the boundary with the other shelf boreholes is based upon a combination of lithological similarities and direct measurement on seismic data.

Boreholes 82/02 and 82/12 show a similar sequence of stratified diamictons overlain by massive diamictons (Figure 4.64), the facies change inferred as marking the unconformity at the base of the Otter Bank sequence. More problematic are the boreholes 82/08, 84/02 and 77/09. Borehole 82/08 shows a thin massive diamicton resting directly upon pre-Quaternary rocks (Figures 4.6 and 4.64) and is interpreted as belonging to the Otter Bank sequence. This is based upon the progressive thinning towards the inner shelf of both the pre-Otter Bank glacial sequence and the Otter

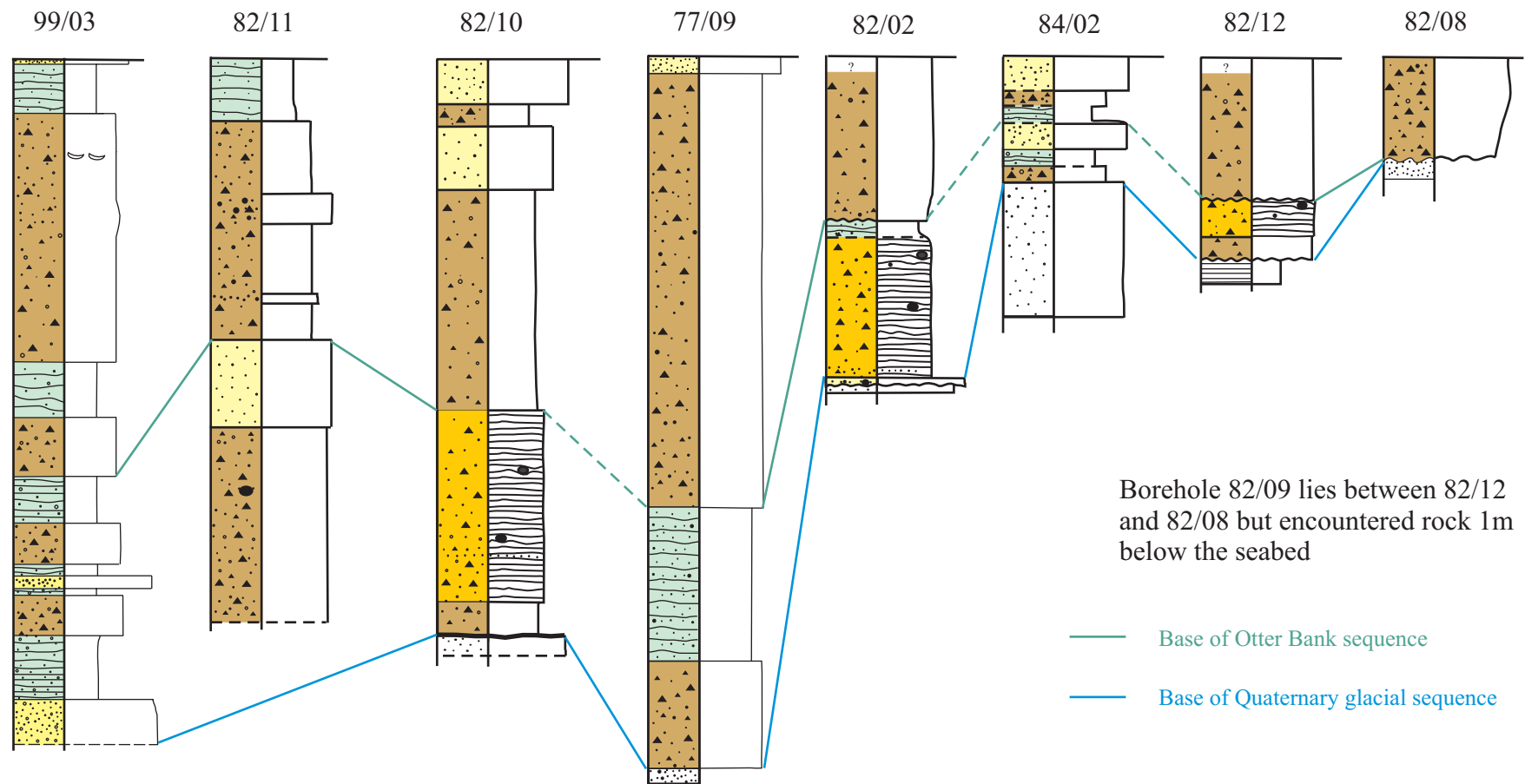


Figure 4.64. Proposed correlation of the boreholes from Transect A, showing base of the Otterbank sequence and the base of the Quaternary glacial sequence.

Bank in the other boreholes. This trend demonstrates that water depth is related to the thickness of the preserved glacial sequence, the glacial sediments being thicker in deeper water. The relationship is reinforced by the thicknesses in boreholes 82/08 and 82/09. Borehole 82/09 lies on the bank west of the Foula channel and has less than 1m of glacial sediments overlying rock, whereas 82/08 lies closer to the land but is located within the Foula Channel, encountering approximately 8m of glacial sediments.

It is possible that the differences in the preserved glacial section reflect their relative exposure to erosion in a post-glacial sea. The higher areas around borehole 82/09 would be more prone to erosion by marine processes than in the sediments in the deeper water at site 82/08. However, this would also be true in the post-glacial sea following the pre-Otter Bank glaciation, which would similarly erode sediments prior to the Otter Bank glaciation. With a further glacial advance over the inner shelf, which is likely to exploit existing topographic lows, it is less likely the pre-Otter Bank sequence would be preserved.

Identification of the Otter Bank base in 77/09 is uncertain, and is based largely upon the similarity of the sequence to borehole 82/10. Both boreholes have a massive diamicton at the base of the glaciogenic section. The overlying stratified diamicton of 82/10 is of similar thickness to the sandy muds of 77/09, both of which are then overlain by massive diamictons. The change between the stratified and massive diamictons in 82/10 is probably equivalent to the facies change in 77/09. This interpretation is supported further by the over-consolidated nature of sediments at several horizons in 77/09 which is interpreted as the result of overriding ice, and the boreholes position at the edge of an area of deeper water (Figure 3.1), which would have aided the preservation of the pre-Otter Bank sequence. Over-consolidation of the sediments during drilling is considered unlikely as most cores suffer from poor recovery due to being poorly consolidated.

The Quaternary section of borehole 84/02 is sand dominated and shows only thin glaciogenic units when compared to other boreholes in a similar position on the shelf

(Figure 4.64). The explanation for this maybe related to the borehole position on the west side of the Papa Basin (Figure 3.1). The thin glacial sediments of the Otter Bank sequence indicate this site was outside the main zone of glacial deposition whilst the domination by sand suggests more current re-working than in other areas. The potential for increased current activity is evident from the possible pathways shown on the bathymetry map in Figure 3.4, which shows the drill site of 84/02 is close to a possible outflow from the Papa Basin. The division of the sediments in the core into two obvious glacial packages separated by sands suggests the lower group corresponds to the pre-Otter Bank sequence and the upper to the Otter Bank sequence.

Correlation of the Otter Bank sequence on the shelf with the core from borehole 99/3 at the base of the slope is based primarily on seismic data. The seismic reflector identified as the base of the Otter Bank Sequence in boreholes 82/10 and 82/11 can be traced on to the slope, and where resolution allows, parts of the basin floor. When this is transferred to the borehole seismic from 99/3 it correlates with the reflector at the base of unit IV (Figure 4.55). The seismic data at this point shows two main phases of debris flows with thin units separating them. This broadly corresponds with the two main units seen in the Otter Bank sequence on the mid shelf. Additional support comes from the AMS ^{14}C date from borehole 99/3 at 34.30m which, at an age of > 52 900 years b.p. indicates the sediments below this level pre-date the Otter Bank sequence.

4.7.2 Summary of glacial sedimentary processes.

The transect from inner shelf to basin floor implies a variety of processes and features have contributed to the development of the Otter Bank sequence. Integration of the data provides an indication of the controls on erosion, deposition and preservation of the glacial sequence. The integrated transect is summarised in Figure 4.65a and b.

The glacial sequence on the inner shelf is generally too thin to resolve on the seismic survey data and is almost absent in some areas. The seabed morphology and

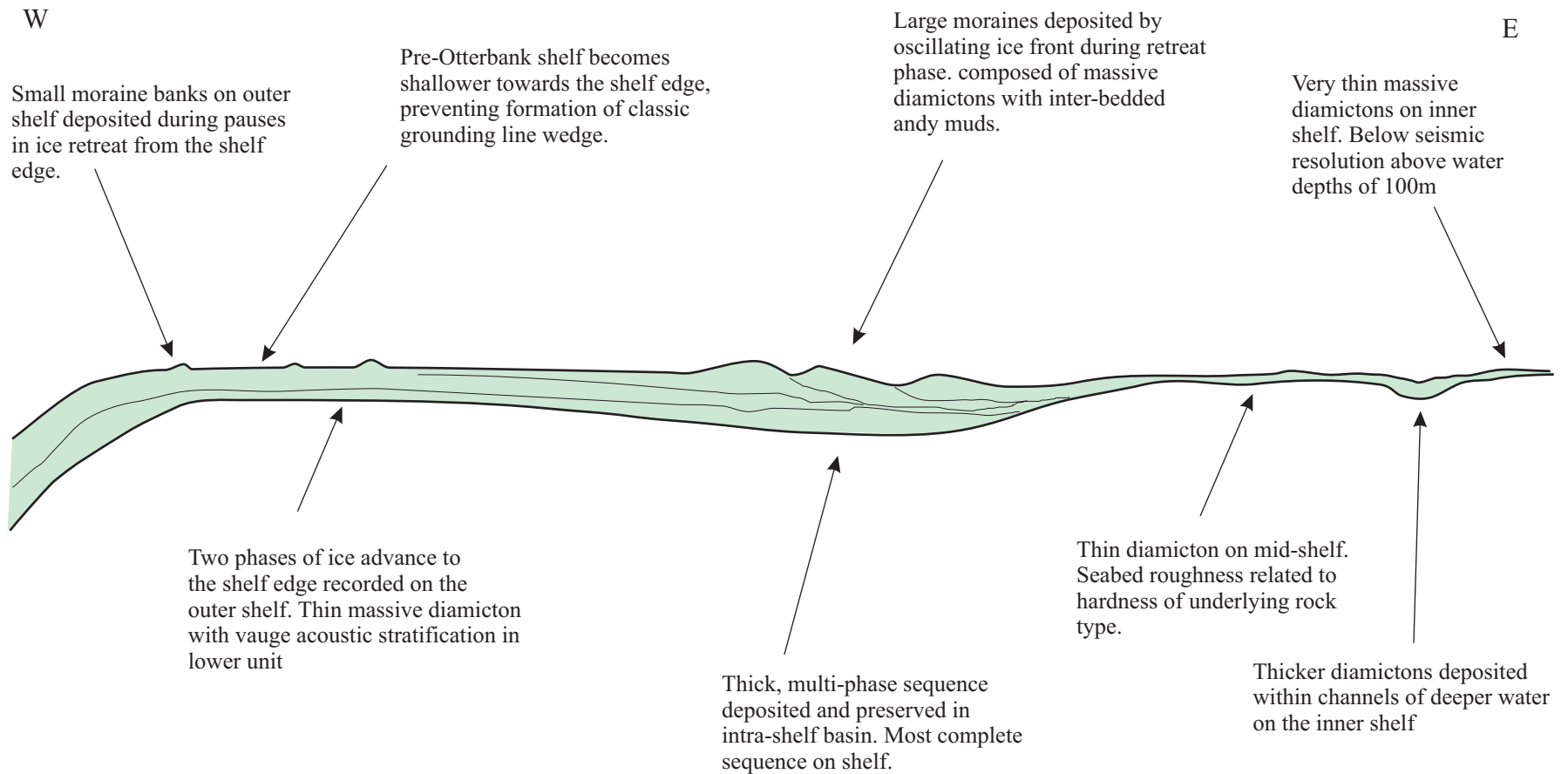


Figure 4.65a. Schematic summary of the Otterbank sequence characteristics from Transect A, Inner shelf to shelf edge. Water depth at the shelf edge is approx. 200m, rising to approx. 100m on the inner shelf to the east.

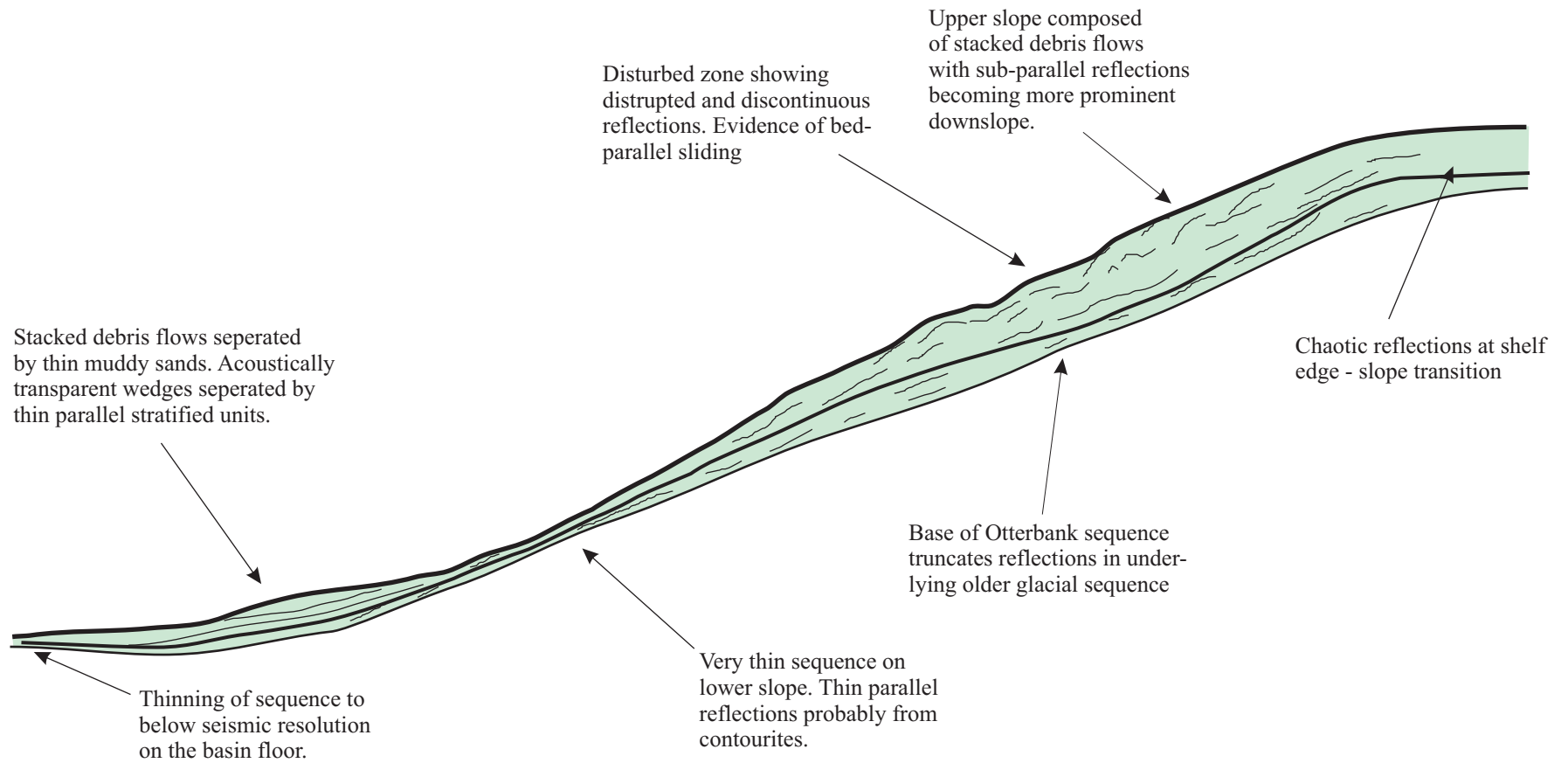


Figure 4.65b. Schematic summary of the Otterbank sequence in Transect A from the shelf edge to the basin floor. Water depth at the shelf edge is approx. 200m, dropping to 1000m at the basin floor.

thickness of the deposited and preserved glacial sequence appear to be related to water depth and underlying rock type. In areas where the water depth is less than c. 100m the seabed is generally smooth, although areas underlain by high grade metamorphic rocks are more rugged. Deposition of glacial sediments on the inner shelf is confined to a very thin, sometimes less than 1m thick, veneer of diamicton. However, deeper channels within the inner shelf zone may preserve a thicker sequence of glacial diamictons. These channels are cut into the softer sedimentary rocks of the area, and are probably of glacial origin, or at least glacially enlarged, forming preferential pathways for ice drainage, and indicating increased deposition in zones of deeper water. The data from the Papa Basin shows that the underlying bedrock type is a stronger control on diamicton deposition/preservation than water depth. It also suggests that a crystalline basement bedrock may hinder the formation and/or deposition of sub-glacial diamictons.

The cores from the mid-shelf, and around the Papa Basin, show unexpected results. The deeper water of the Papa Basin might be expected to have a thicker sequence of glacial deposits as it provides more accommodation space for deposition than the adjacent shelf. However, boreholes 82/12 and 84/02 (Figure 4.64) show that this is not the case, indicating a different control on deposition/preservation. This suggests the deeper water borehole sequence is away from the main route of sediment delivery, and whilst the Papa Basin has been occupied by Otter Bank age ice, it was not the primary route of sediment transport and deposition.

The mid to outer shelf shows a wide range of features and a thick preserved Otter Bank sequence. The sequence is thickest, and contains a more complete history of events, within the intra-shelf basin where water depths are greater. This part of the Otter Bank comprises two massive diamictons from separate advances to the shelf edge, and a sequence of inter-bedded muds and diamictons from the emplacement of large moraines by an oscillating ice front during an overall retreat. Preservation of the sequence in the basin is thought to be due to increased ice-sheet buoyancy in the deeper water preventing removal of the earlier phases during the re-advance. The

configuration of the preserved moraine banks suggests locally surging lobes at the ice front.

Towards the shelf edge the water depth and sediment thickness decrease. Small moraine banks deposited during pauses in ice retreat form mounds on the seabed. Massive diamictons with vague acoustic stratification at the shelf edge change to structureless elongate wedges at the top of the slope where they form prograding debris flows.

The upper to mid-slope (Figure 4.65b) consists of stacked debris flows, probably composed of massive diamictons, which thin towards the lower slope. In areas away from the main depositional pathway the debris flows are generally thinner and more dispersed. On the slope below the Foula Bight the seabed is cut by incised gullies indicating sediment starvation at an early stage in deglaciation. There is also evidence from seismic data of contouring current activity on the mid slope, forming a series of stacked contourites which appear as thin stratified reflections. These currents may remove much of the mud component of the slope sediments, but may also re-deposit the mud locally.

At the base of the slope a series of large stacked debris flows form lobate structures composed of massive diamictons. Thin, acoustically stratified sandy muds separate some of the debris flows and represent the background glacial marine deposition from suspension and ice rafting. Data from the glacial succession from the base of the slope indicates that the debris flows formed at a time when the ice was at the shelf edge, and delivery of coarse suspended sediment and bedload to the deep water environment was at its maximum.

4.7.3 Implications for focused deposition and former ice stream pathways

Data from the shelf section of transect A indicates that some areas, particularly on the outer shelf, accumulated thicker glacial sediments than other areas, thereby suggesting an apparent focus for deposition. Examination of additional seismic data from lines perpendicular to the line of the transect shows the a wider view of the

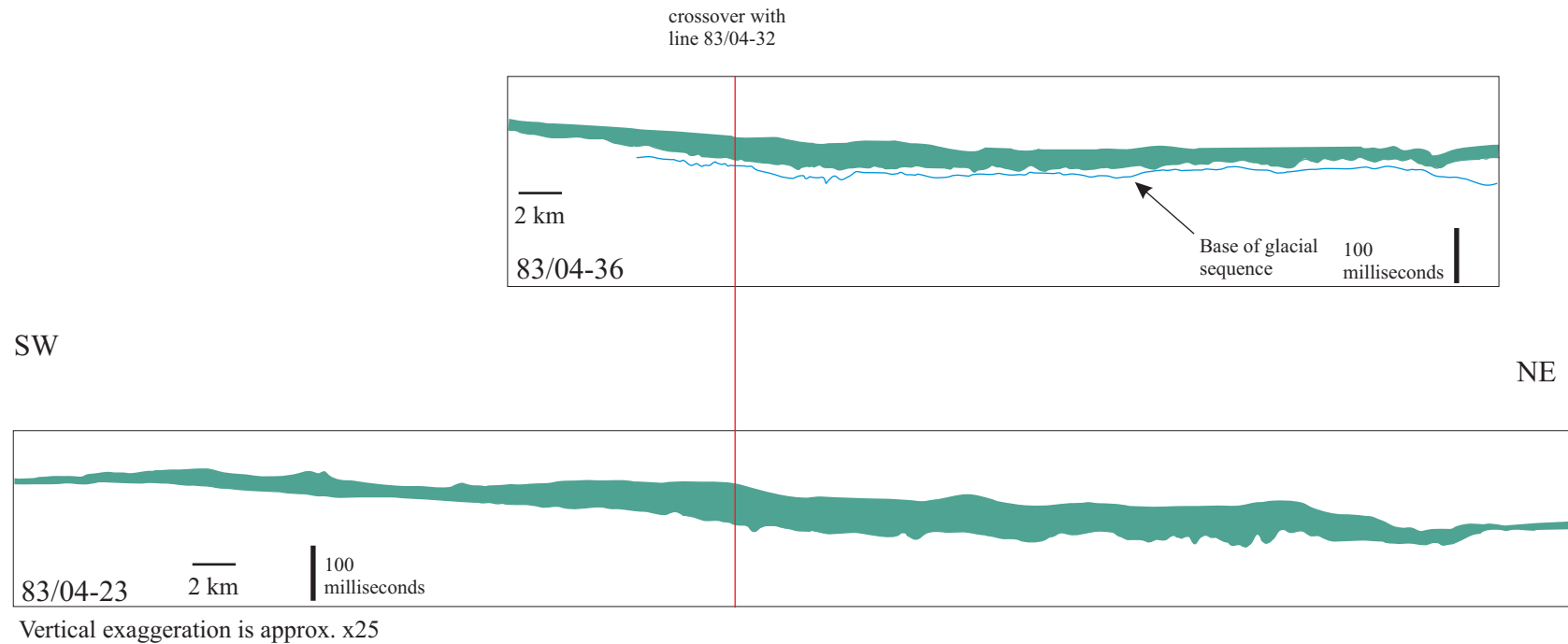


Figure 4.66. Thickness of the Otter Bank sequence highlighted on the sparker lines 83/0423 and 83/04-36 which lie roughly parallel to the shelf edge, crossing the pathway of the proposed southern ice stream. The thicker sediments occupy a shallow trough feature, along which ice transport and hence sediment delivery, were at a maximum. As the ice generally retreated, the shallow trough was in-filled to a greater degree by the additional volume of sediment delivered by the ice stream. The data from line 23 exaggerates the effect as it crosses the line of the large outer shelf moraine banks described in section 4.3.4.

focused deposition, allowing some delineation of the depocentre. Survey lines 83/04-23 and 36 which lie parallel to the shelf edge, show the focused deposition in greater detail (Figures 2.1 and 4.66).

The data from line 83/04-23 shows a broad shallow depression about 40 km wide incised into the underlying sediments of an earlier glacial episode and in-filled with Otter Bank sequence glacial sediments. This feature is also present but less well defined on line 83/04-36 which lies parallel to line 23 nearer the shelf edge. The data from both these lines indicates an increased thickness of the OB1 unit from the initial ice advance across the shelf, indicating this pathway formed a preferential route from the beginning of the Otter Bank phase glaciation. These features are included in the projected reconstructions of the Otter Bank glaciation in Chapter 6. Suffice to say, the presence of a buried shallow trough-like feature, in-filled with glacial sediment fulfils some of the criteria for identification of a palaeo ice-stream (Stokes and Clark, 2001), and raises the possibility that the apparent preferential pathway across the shelf may be the remnants of a former ice-stream. This possibility is further supported by data from the upper and lower slopes in areas D and E which show zones of increased deposition relative to other parts of the slope. When viewed with isopach data for the Otter Bank sequence across the whole margin (Figure 6.1) and the moraine distribution for the southern shelf area (Figure 6.3), the relationship between the in-filled feature and former ice transport pathways becomes clearer.

5. Glaciated West Shetland margin – Transect B

The second transect across the West Shetland margin is less detailed than transect A, but includes data from different settings. The starting point for transect B lies on the west side of the Shetland Mainland, taking in onshore and coastal settings not available in transect A and extends onto the outer shelf. Beyond the shelf edge, the slope and basin floor are covered by area F of transect A, and are thus not repeated here. The data sets for the northern transect are much more restricted, with only 4 cores and limited seismic survey coverage. However, the seabed image covers important parts of the outer shelf and does not suffer from the spurious data seen in area C. Core descriptions and interpretations for the boreholes in this transect are based upon shipboard records and existing published data rather than re-examination of the cores.

5.1 Area G: Onshore to nearshore setting

5.1.1 Onshore

Onshore data was retrieved from 2 sites on the west coast of Shetland Mainland; Fugla Ness in the north, and Sel Ayre on the southern limb of St. Magnus Bay. Both of these localities have been studied previously and published as part of the Geological Conservation Review Series (Gordon and Sutherland, 1993). The nearshore locality in area G is based around borehole 80/08 in St. Magnus Bay and is roughly equivalent to a fjord type setting. Together these localities form area G (Figure 5.1).

Fugla Ness

The locality at Fugla Ness (Figure 5.2) comprises a small cliff section up to 10m high described by Chapelhowe, (1965) and Gordon and Sutherland, (1993). A summary log of the Fugla Ness section is presented in Figure 5.3 and illustrated in photographs in Figure 5.4. The sequence at this locality consists broadly of a basal grey till which is compact and may be partly cemented. This unit is over 2m thick and contains sub-angular clasts of a few centimetres. Overlying the basal till, is a peat bed approximately 1.35m thick, containing large fragments of wood up to 20cm long and clasts of unidentified grey crystalline rock, particularly near the base. There

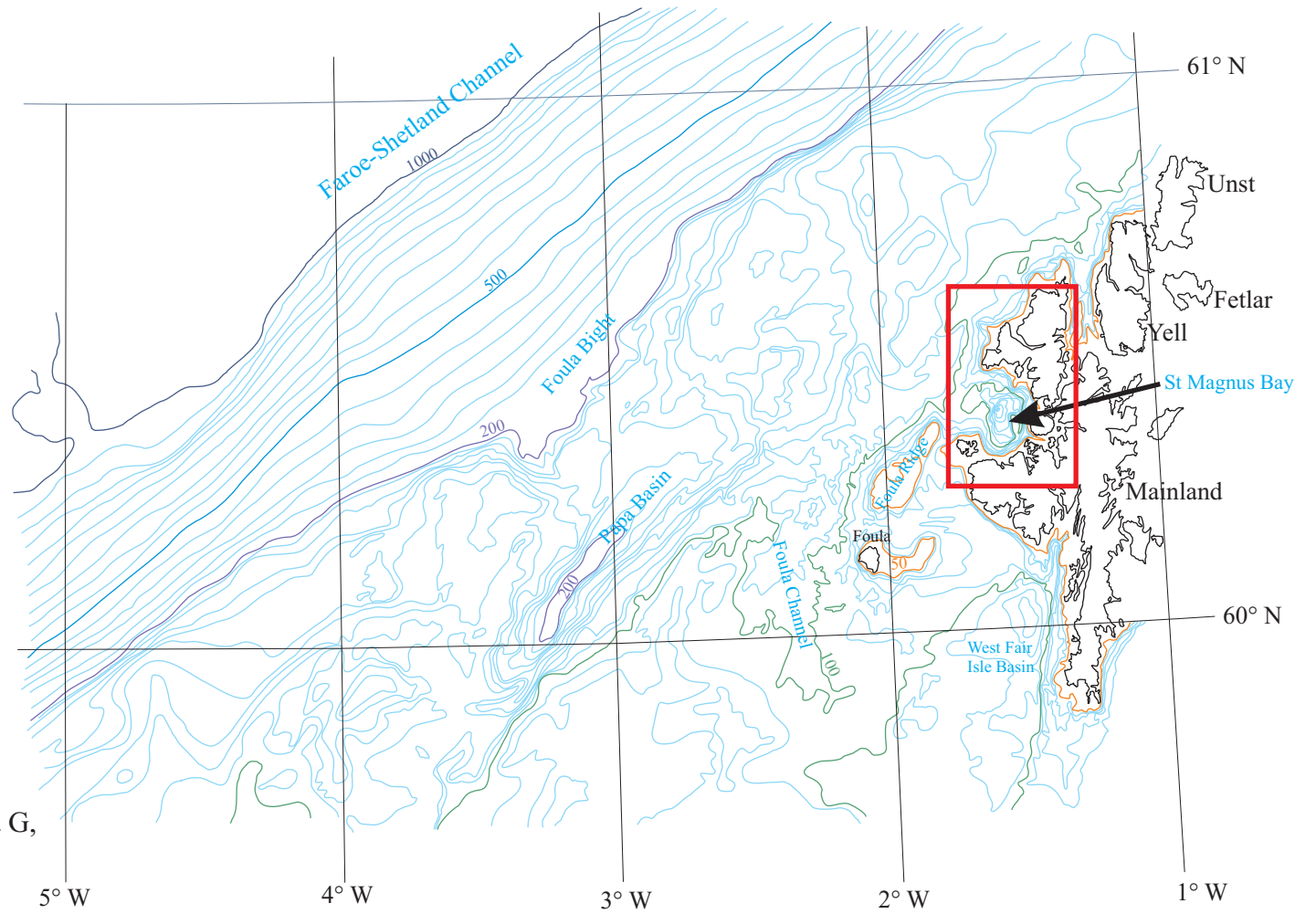


Figure 5.1. Location of Area G, outlined in red.

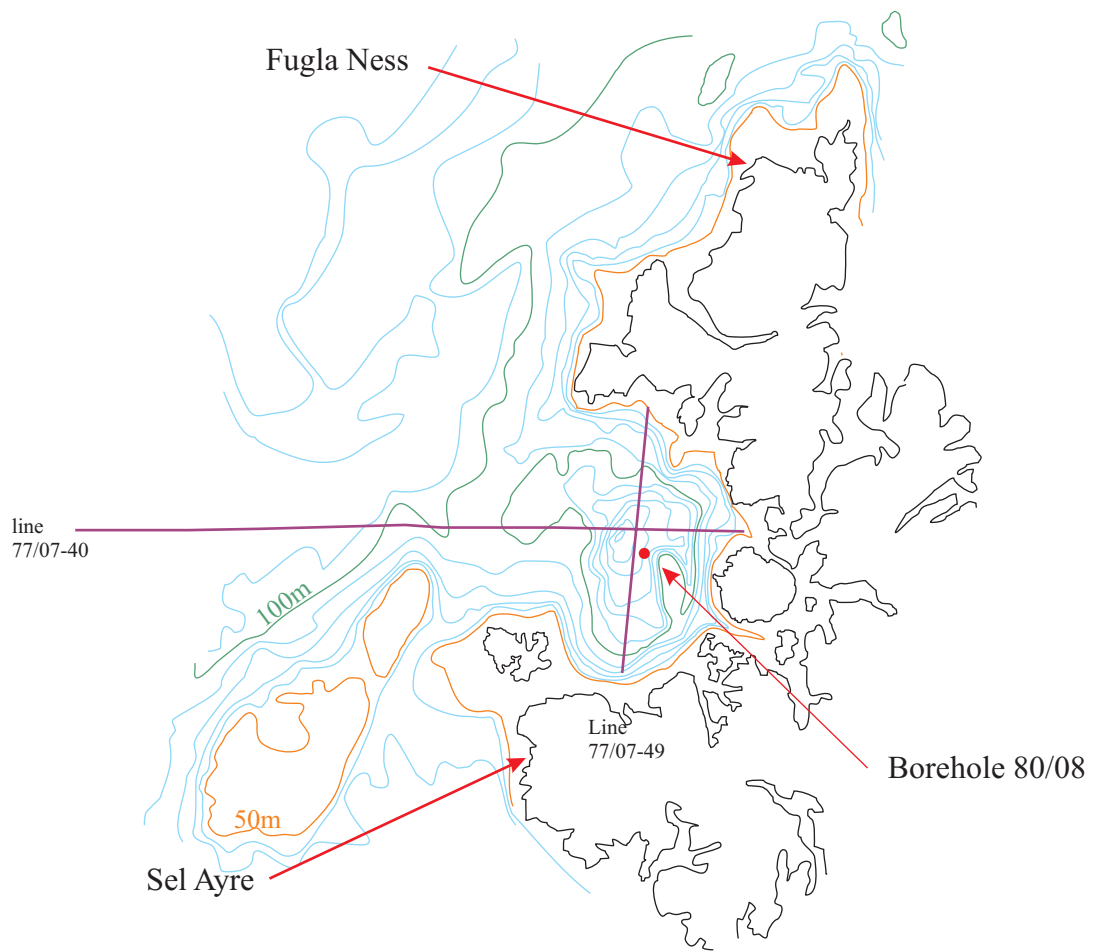


Figure 5.2. Location of onshore localities, borehole 80/08 and seismic survey lines around St. Magnus Bay, west Shetland Mainland. Submarine contours are at 10m intervals.

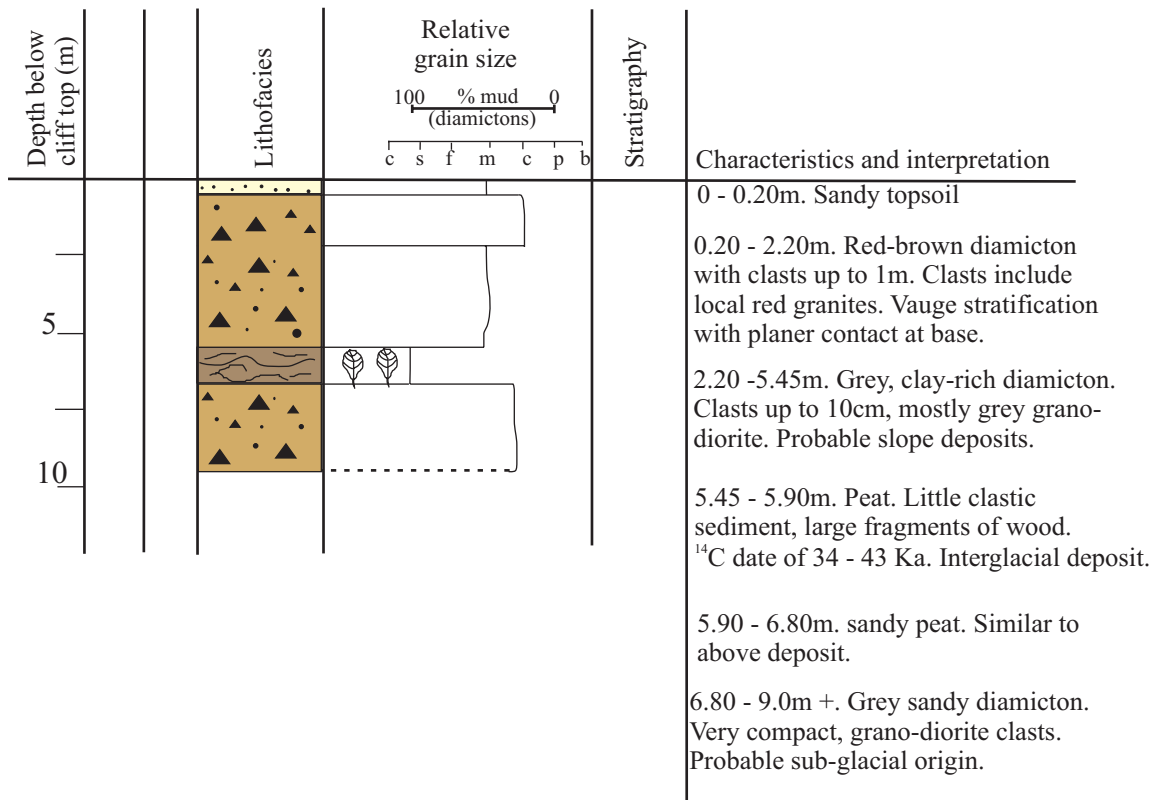


Figure 5.3. Log of the onshore outcrop at Fugla Ness (HU 312913). The uppermost diamicton is of Late Devensian age, and probably equivalent to the Otterbank sequence offshore.

are also reports of pine cones from within the upper portion of this unit (Gordon and Sutherland, 1993). The peat bed has yielded radiocarbon ages of between 34,000 and 48,000 years b.p. (Harkness and Wilson, 1979). Overlying the peat bed is a pale grey-brown till up to 3.25m thick with a high mud content and clasts of grey granodiorite and gneissose rocks up to 5 cm. Overlying the grey till, with an abrupt planer contact, is a 2m thick red-brown till containing clasts up to 1m long. The clast content contains a high proportion of red granitic rocks of probable local origin (Figure 5.4a and b). Within this unit there is evidence of a vague stratification. The outcrop is capped by a thin topsoil.

The interpretation of the Fugla Ness section is the subject of some uncertainty. The basal till is of glacial origin but uncertain age. The dates obtained from the overlying peat bed indicate a mid - early Pleistocene age for the lower till (Gordon and Sutherland, 1993). However, there is a possibility that the peat bed is actually an erratic or thrust block which pre-dates the lowermost till (Birks and Ransom, 1969), in which case, the peat pre-dates all the till deposits. The thick grey till overlying the peat beds is interpreted as periglacial slope deposits by Hall et al, (1993). This would be consistent with the high mud content and lack of larger clasts in the sediments in this unit. The upper red-brown till is interpreted by all previous authors as a glacial till of probable late Devensian age formed by a local ice cap (Gordon & Sutherland, 1983) and incorporating red granite clasts of local origin.

The sequence at Fugla Ness shows evidence of at least two and possibly three glacial episodes, of which the uppermost unit at least, is of late Devensian age. This indicates that a thin locally derived, glacial till was forming on the Shetland Mainland during the same glacial episode as formed the Otter Bank sequence offshore.

Sel Ayre

The second onshore locality lies at Sel Ayre on the Walls Peninsula which forms the southern limb of the land surrounding St. Magnus Bay (Figure 5.2). The glacial section at this locality is situated at the top of steep, dramatic cliffs over 100m high

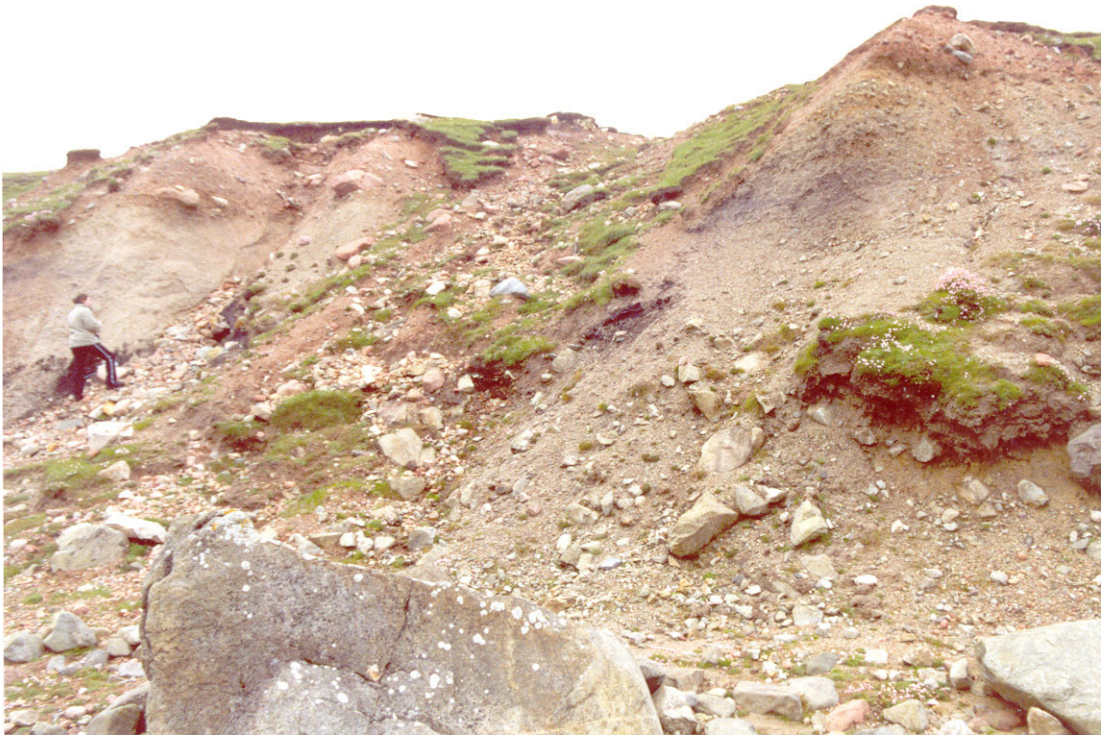


Figure 5.4a. Compound glacial section at Fugla Ness showing upper, middle and lower tills. The peat bed is located at the foot of the slope next to the figure.



Figure 5.4b. The upper till, composed of red-brown sandy diamicton with clasts up to boulder size. Trowel in centre is approximately 15cm long.

(Figure 5.5a). The section containing the glacial sediments comprises the top 10m of the cliffs, and is described in detail by Mykura and Phemister (1976). The sequence consists of a muddy diamicton containing regionally local clasts (Figure 5.5b), overlying well-bedded muddy gravels and a sequence of sands and peaty sands. Much of the lower part of the section has been interpreted as periglacial slope deposits, with the peaty sands representing interglacial deposits dated by Luminescence dating at about 100,000 years (Hall, 1997).

The peaty sands beneath the slope deposits are interpreted by Gordon and Sutherland (1993) as temperate interglacial deposits from the Ipswichian Stage of the Pleistocene based primarily on pollen records. The muddy diamictons at the top of the section are probably of late Devensian age, but this is not certain.

The upper diamictons at both Fugla Ness and Sel Ayre may represent the same late Devensian glaciation of Shetland. However, beyond this, correlation with the offshore record is not possible without further dating evidence. Other localities on Shetland show the presence of cold periglacial conditions dating back to approximately 13,000 years ago (Gordon and Sutherland (1993). This includes Ronas Hill, the highest point in the Shetland Islands, which does not appear to have been overridden during the Late Devensian.

5.1.2 Offshore shelf setting

Nearshore/fjord setting

The nearshore area in St. Magnus Bay bears some similarities to a fjord setting in that it forms an overdeepened basin with steep sides and a seaward opening onto the shelf. An examination of the bathymetry shown in Figure 6.2 shows a sub-circular basin reaching a depth of 160m which rises to less than 100m at the opening to the shelf. Around the margins of St. Magnus Bay above 50m isobath, numerous elongate inlets feed into the main basin which may be the outflow pathways of small glaciers.

Seismic survey data from this area is very limited due to problems of acquisition within a restricted basin and seabed multiples in shallow water. However, two

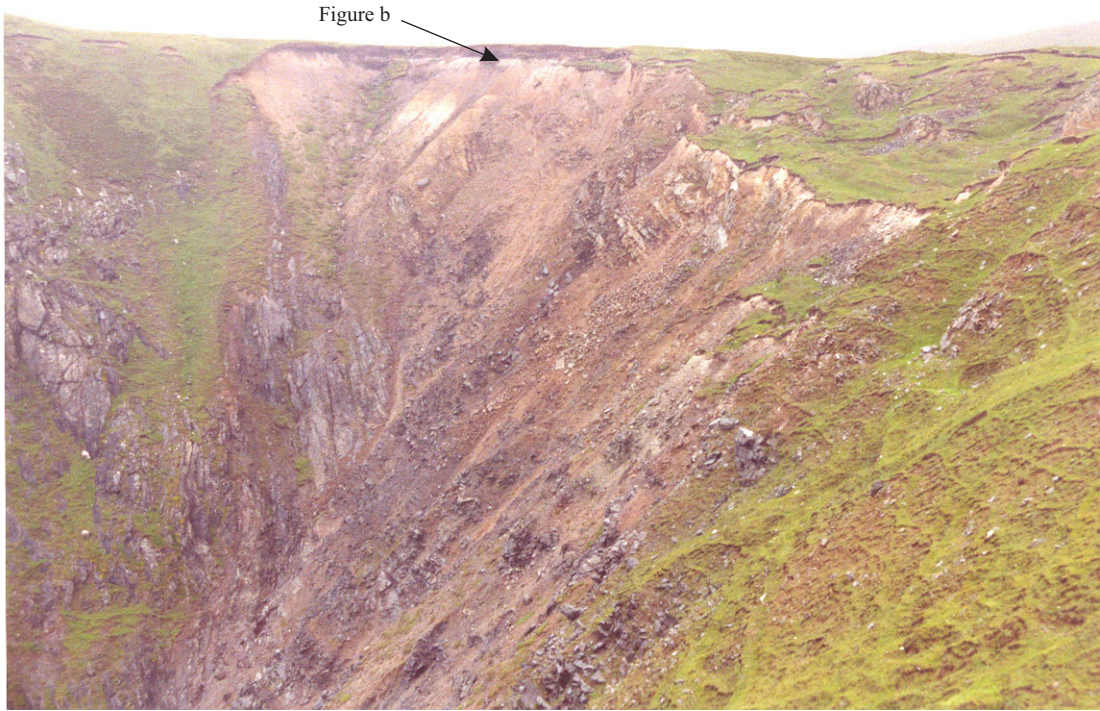


Figure 5.5a. The cliffs and glacial section at Sel Ayre. The section in Figure b is located just below the horizon. The white dots on the grass slope to the left of the picture are sheep, which give a comparative scale.



Figure 5.5b. The upper diamicton at Sel Ayre

sparker lines (located in Figure 5.2) are presented which show enough detail to allow some interpretation and correlation with core data. Line 77/07-49 (Figure 5.6) forms a north – south transect across St. Magnus Bay, shows a broadly sub-parallel stratified pattern of reflections infilling the main part of the basin with evidence of onlapping towards the margins. Data from borehole 80/08 (described below) indicates that the base of the Quaternary section lies at about 35m, which, allowing for poor consolidation in the upper sections, translates to a travel time of around 45 ms (using a velocity of 1.5km/s). Line 77/07-40 (Figure 5.7) lies in a roughly east – west direction (Figure 5.2), showing the long profile of St. Magnus Bay and small seafloor mounds on the shelf beyond. The Quaternary section is difficult to identify due to the presence of seabed multiples, but appears to comprise a sequence of sub-parallel stratified reflections across most of the basin. The close spacing of the reflections would suggest a stratified sediment both within the basin and in the shelf zone beyond. There are also a number of small mounds up to 20ms high on the seabed which may be rock outcrop, but could be small buried moraine banks. On the landward (eastern) side of St Magnus bay, there are several small infilled basins on the slope towards the shoreline. These also appear to show an infill of parallel stratified reflections but are partly obscured by the seabed pulse.

The pattern of stratified reflections observed on the survey lines across St. Magnus Bay have a high degree of lateral continuity, forming a draping infill to the basin. This style of infill suggests deposition by suspension settling. However, on the eastern edge of the basin there is also a sigmoidal package of prograding and aggrading reflections, indicating a current derived input, probably from a terrestrial origin.

Borehole 80/08

Borehole 80/08 lies slightly to the south of the centre of St. Magnus Bay, 7km west of Muckle Roe in a water depth of 140m (Figure 5.2). The borehole encountered approximately 35m of Quaternary sediments resting upon sandstones of probable Permo-Triassic age (Stoker et al, 1993). Recovery from the upper 25m was exceptionally good, with almost complete recovery; however, recovery was less

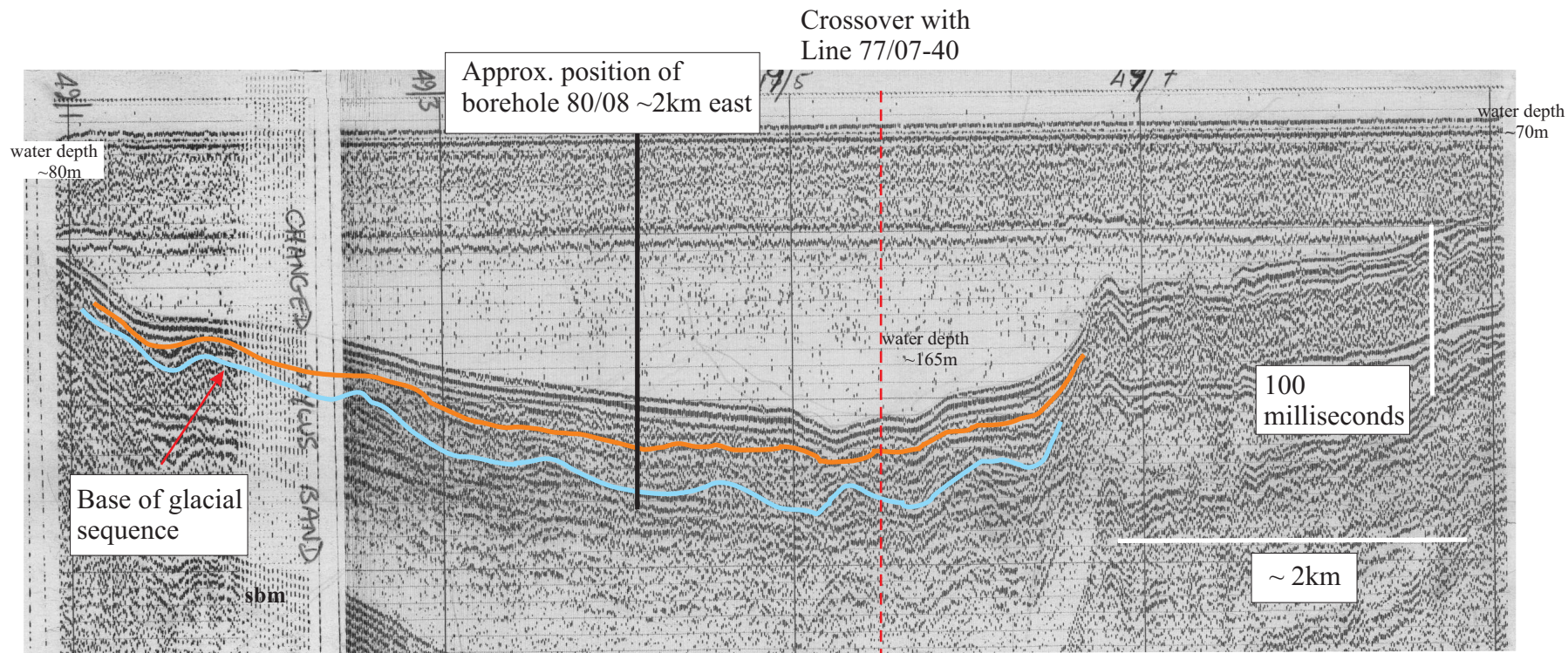


Figure 5.6. Airgun survey line 77/07-49 showing a north-south profile across St. Magnus Bay. The base of the glacial sequence also marks the presumed base of the Otter Bank sequence. The orange reflector forms a prominent horizon of unknown affinity, but lies at a depth of c.20m and may relate to the amelioration and change to fully marine salinity recorded from that depth. Accurate estimates of seismic velocity, and hence borehole depths are hindered by the abundance of gas present in the muds. Sbm, sea-bed multiple.

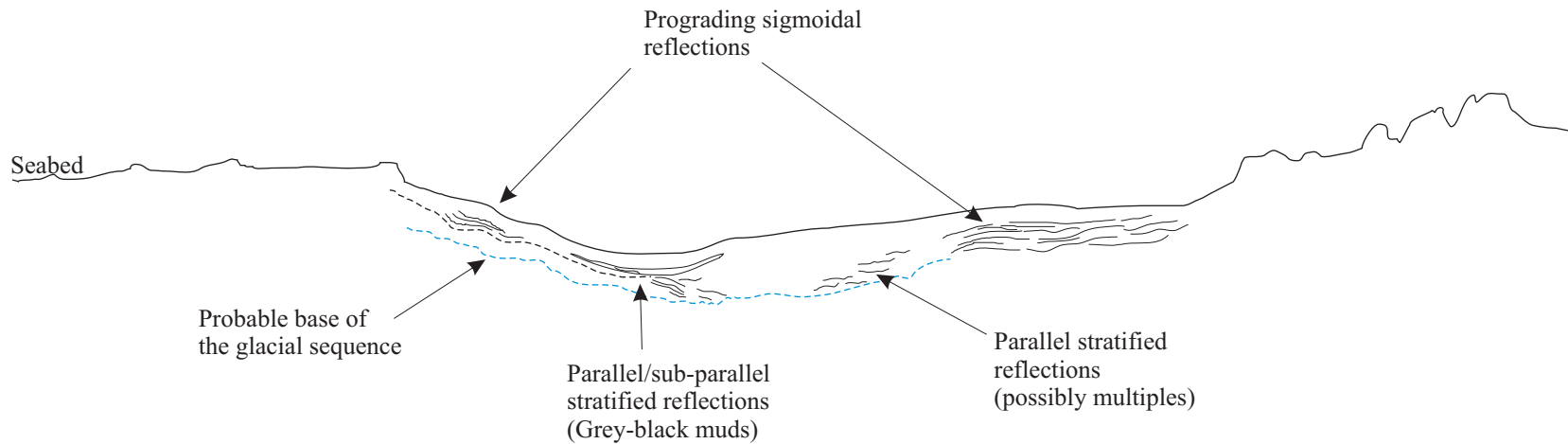
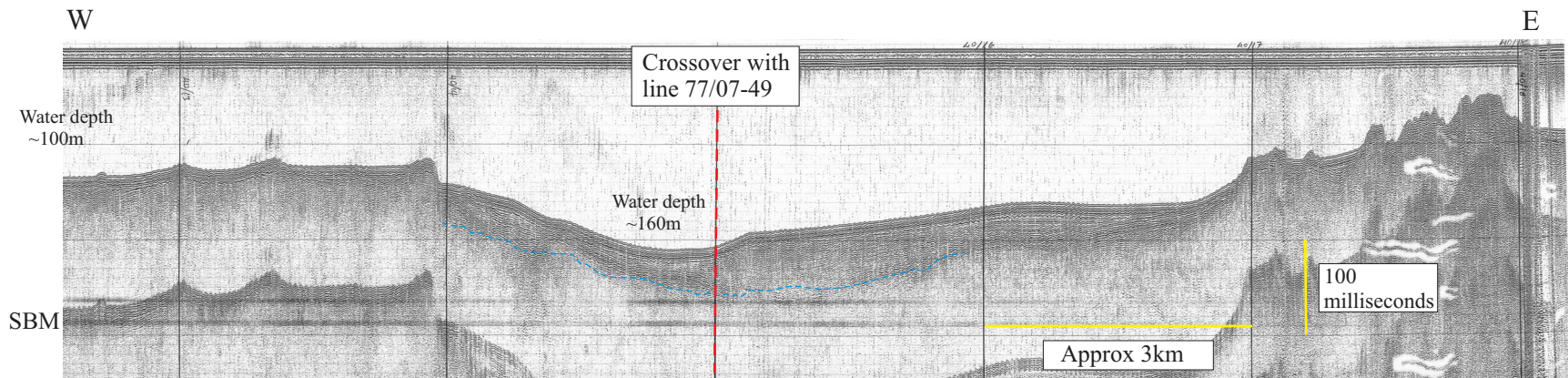


Figure 5.7. Seismic sparker line 77/07-40 showing an east-west profile across St. Magnus Bay. Sea-bed multiples (SBM) make interpretation difficult in this section. The sigmoidal packages represent the input of clastic sediment to the basin margins, leaving the centre sediment starved.

good in the lower 10m. The summary core log for borehole 80/08 is shown in Figure 5.8.

The lowermost section, from the base of the Quaternary at 34.65m to 25m, recovered short sections of massive, red-brown sandy diamicton with pebbles of grey and red sandstones up to 30mm. At the time of recovery, the diamictons were friable with a stiff to compact consistency. Overlying the diamictons, at the top of a section of poor recovery, is a thin section of laminated grey-brown silty mud (Figure 5.9a). A long sequence of slightly silty, organic-rich muds overlies the lower muds from 24.75m to 4.40m with a gradational boundary. These muds change character from dark grey with prominent lamination near the base, to black apparently structureless mud at the top (Figure 5.9b). The only exception to this is the presence of pale green bands within poor lamination at about 10.50m. The muds also show an open structure due to gas escape, especially below about 8m (Figure 5.9c). The original drilling records reported a strong hydrogen sulphide smell. At the time of photographing in 1985 (Cockcroft, 1987), the majority of the black and dark grey mud sections had oxidized to a pale grey-brown colour. Overlying the black muds with a gradational contact is a 1.4m section of soft, olive-grey silty clay. The uppermost unit comprises approximately 1m of soft, fine to very fine-grained muddy sand. The sand unit is similar in colour to the underlying muds and contains up to 20% shell fragments (Figure 5.9d).

In addition to the sedimentological description, detailed palynological analysis was undertaken by Wilkinson (BGS Internal Report 94/511 – unpublished) in 1986 in order to try to date the sediments and establish a stratigraphy. The results of this study generally show low abundance below 10m, probably due to a severe cold climate and a reduced salinity. Between 19m and 23m there is some evidence of slight amelioration, with an increase in palynomorph abundance and probably normal marine salinity. From 10m to 5m the data is more varied, with a slight temporary change from a severe cold climate to a less extreme environment and fully marine salinity at 7.5m. The presence of the cysts from the genus *Protoperidinium* is interpreted as an indicator of sea ice, which may have been permanent or seasonal.

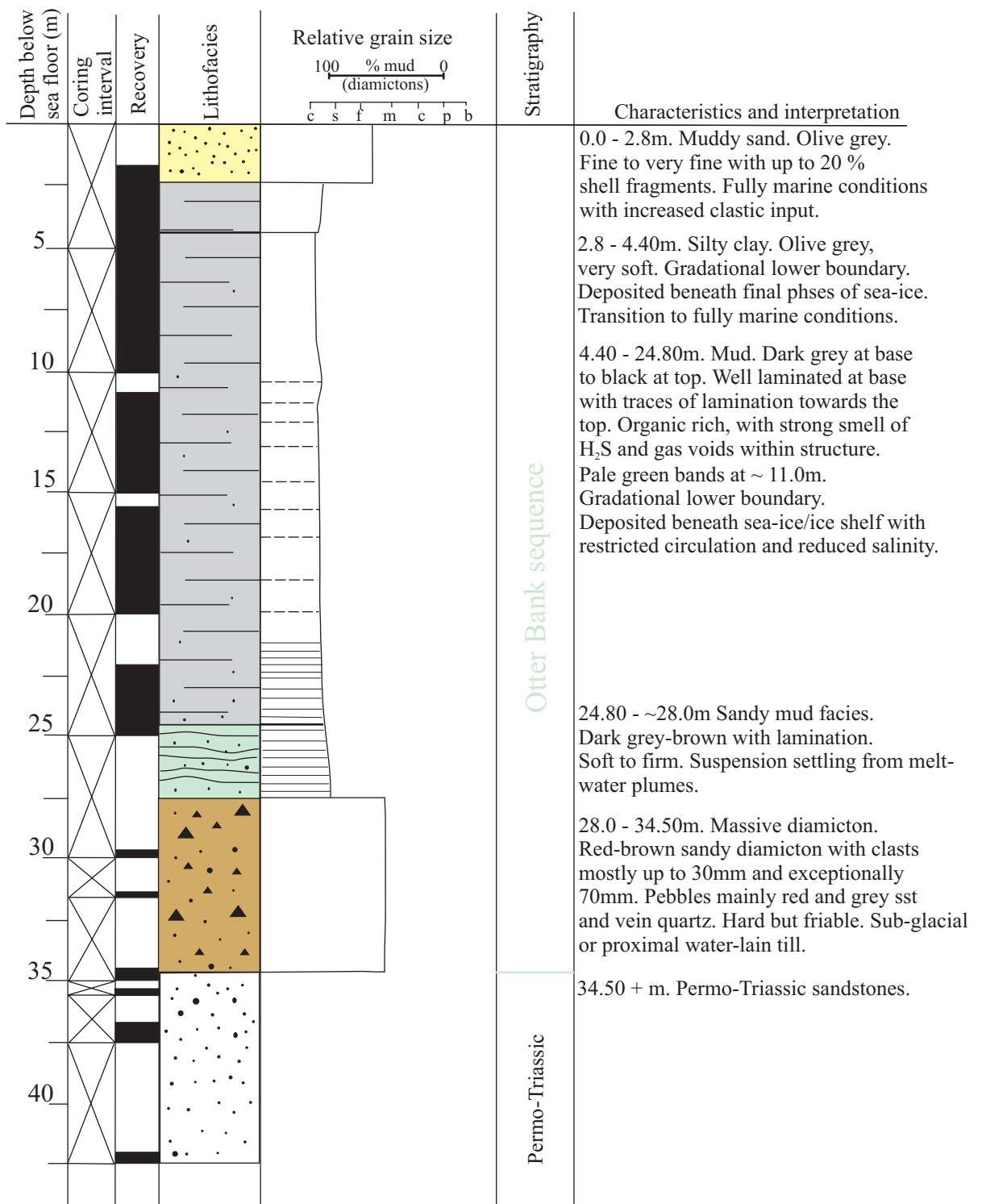


Figure 5.8. Log of core from borehole 80/08.



A) Section 24.8 - 25.00m. Well laminated sandy mud facies of Late Devensian age and probable Otterbank sequence affinity.



C) 9.90 - 9.96m. Muds showing fabric of gas escape and bubble structures from H₂S formation. Red-brown colour is post recovery oxidation, the original being dark grey to black.



B) 4.50 - 4.60m. Black, organic-rich silty muds with no apparent structure. The black centre section is the original colour.



D) 2.20 - 2.40m. Muddy sand with abundant shell debris. The unit is post-glacial age.

Figure 5.9 a-d. Core sections from borehole 80/08 (Photos taken from Cockcroft, 1987)

Warming to temperatures typical of higher latitude northern waters occurs between 5m and 2.5m, with a reduction in the abundance of *Protoperidinium* indicating a reduction in sea ice.

The study concluded that the sequence was of late Devensian age, with the sequence below 10m probably representing the Older Dryas/Devensian Lateglacial, the amelioration between 10m and 5m representing the Allerød interstadial, and the 5m to 2.5m section representing the Younger Dryas.

The basal diamictos from 80/08 are interpreted as sub-glacial or water lain tills similar to those seen in the inner shelf areas further south. The small samples and lack of seismic detail prevent a more definitive interpretation of these sediments, although the presence of red sandstone clasts within the tills suggests a local origin from the underlying red sandstones. The fine-grained nature of the 2.5m to 25m section indicates lack of wave action. The suggestion of reduced salinity also indicates limited mixing of water within St. Magnus Bay with the marine realm on the shelf beyond. The lowermost section of the mud between 24.50m and 25m is slightly coarser grained, and is interpreted as belonging to the sandy mud facies. The presence of lamination within the lower muds, and the absence of coarser components, suggests deposition beneath floating ice (Ó Cofaigh and Dowdeswell, 2001). This interpretation is supported by the palynological data which suggests the presence of sea ice throughout much of the sequence. The indications of reduced salinity in the upper tills and lower muds also supports deposition beneath floating ice, the ice providing input of fresh meltwater affecting the salinity. Within the relatively restricted embayment of St. Magnus Bay, this would locally reduce salinity. The occupation of St. Magnus Bay by a large ice mass would also restrict marine circulation. This possibility is increased by the pronounced rim at the seaward opening of St. Magnus Bay which lies 65m above the basin floor. A considerable degree of melting and/or sea level rise would be required to allow large bergs to escape to the open shelf beyond.

The dark grey and black muds from 24.50m to approximately 4.25m show a greatly increased organic content, deposited in a strongly reducing environment. There is also a notable lack of coarse clastic sediment. The presence of sea ice, indicated by the palynological data, may be responsible for suppressing the input of coarser ice-rafted debris. Generation of laminated muds beneath shorefast sea ice has been reported from fjord settings in East Greenland (Dowdeswell *et al*, 1998) where the sea ice has prevented the movement of calved icebergs beyond the ice front. Interpretation of a lacustrine setting for the mud deposition in St. Magnus Bay at a time of no ice cover in Shetland, during a full interstadial (Cockcroft, 1987) are not supported by the data.

The uppermost section of core 80/08, comprising olive-grey silty muds capped with shelly muddy sands are interpreted as the transition from a cold, probably periglacial climate with sea-ice, to a modern climate. The increased input of coarser clastic sediment is probably related to the erosion of exposed glacial sediments onshore and the reworking of sediments in St. Magnus Bay under unrestricted, fully marine conditions.

5.1.3 Interpretation of onshore – nearshore setting (area G)

Definitive correlation of the onshore glacial sequence with the offshore section is not possible from the localities examined in this study. The whole of the sequence in borehole 80/08 is of late Devensian to Holocene age, which makes it equivalent to the Otter Bank sequence, although it is not easily linked further offshore by the seismic data. The upper till at Fugla Ness would appear to be of an equivalent age to the Otter Bank sequence, but this correlation is tenuous. Likewise, the upper till at Sel Ayre is also possibly of late Devensian age, suggesting it too is equivalent to the Otter Bank sequence.

The sequence of organic-rich muds deposited under sea-ice conditions represents an important late phase in the late Devensian glaciation which is not recorded in other areas of the shelf. At the time of mud accumulation in St. Magnus Bay it is likely the

remainder of the shelf was sediment starved, with current reworking of the seafloor and clastic input from more distant traveled icebergs.

The small sections of actual glacigenic origin preserved within the onshore/nearshore setting were probably deposited during the closing phases of the glaciation. Any older deposits having been re-worked and transported to the outer shelf by ice advances during the late Devensian, as seen in the areas of the southern transect A. Post-glacial re-working may also be responsible for reducing the preserved onshore glacial sequence, which is orders of magnitude smaller than the offshore section. The onshore areas would also have formed an upland area of net erosion.

5.2 Area H: Inner to Outer Shelf

Area H encompasses the whole of the zone from the inner shelf to the shelf break (Figure 5.10). This zone is much larger and more generalised than the areas of transect A due to the poorer quality of the seismic data and the lower number of cores available. However, the data provided by an examination of the bathymetry, and the seabed image, add important characteristics to the overall view of the West Shetland Margin.

5.2.1 Seabed morphology and seismic data

Bathymetry

The bathymetry of the inner shelf shows a platform-like area at a water depth between 100 - 120m in the area of the pathway leading out of St. Magnus Bay (Figure 5.11). The pathway then drops down into a small basin on the mid shelf and is joined by another pathway from the south leading out of the Foula Channel. The outflow from the basin follows several routes, but the main pathway appears to travel north-westwards to a small embayment in the shelf edge (Figure 5.11). The outer shelf is marked by two large banks at a depth of 130 – 140m, between which the outlet pathway runs. Beyond the banks, the shelf dips down towards the shelf break at about 200m.

Seabed Image

Seabed image coverage of area H is restricted to the outer shelf (Figure 5.10) and suffers from prominent acquisition footprint which creates a false lineation (Figure 5.12). Nevertheless, a series of 3 arc-shaped ridges with a number of smaller, discontinuous ridges between them are discernable on the image. The ridges have been interpreted by Bulat and Long (2001) as terminal moraines, indicating the extent of several phases of ice advance across the shelf. The main ridges and to a lesser extent some of the small ridges, form zones of higher and lower relief. This has been interpreted as the effects of later current re-working, or differences in sediment delivery to the ice margin (Leslie et al, 2003). However, it is also possible that the higher relief zones represent locally greater rates of ice advance forming

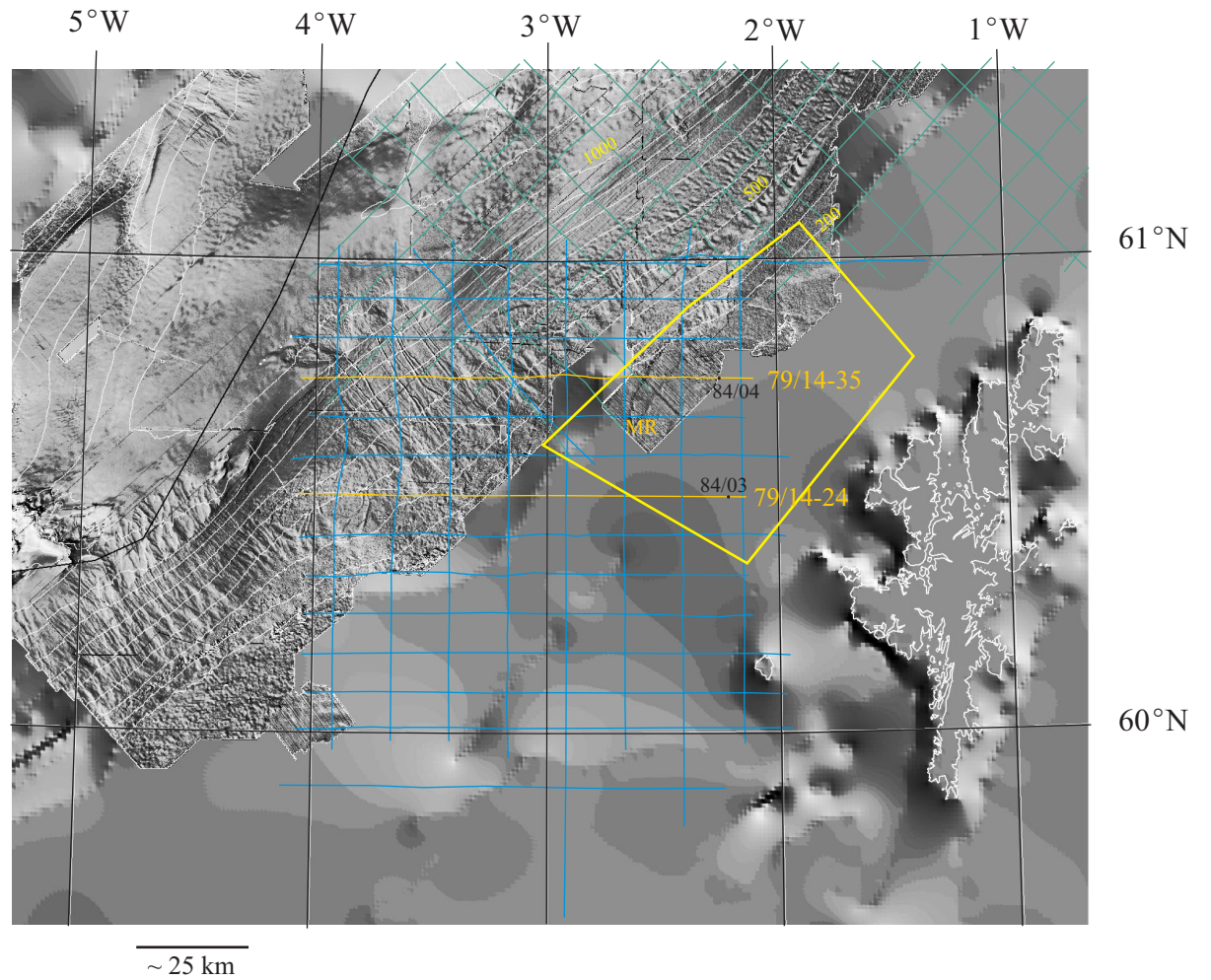


Figure 5.10. Location of Area H, outlined in yellow. Seismic lines used in this section are highlighted within the survey grid together with relevant boreholes used. The moraine ridges (MR) on the outer shelf are enlarged in Figure 5.12.

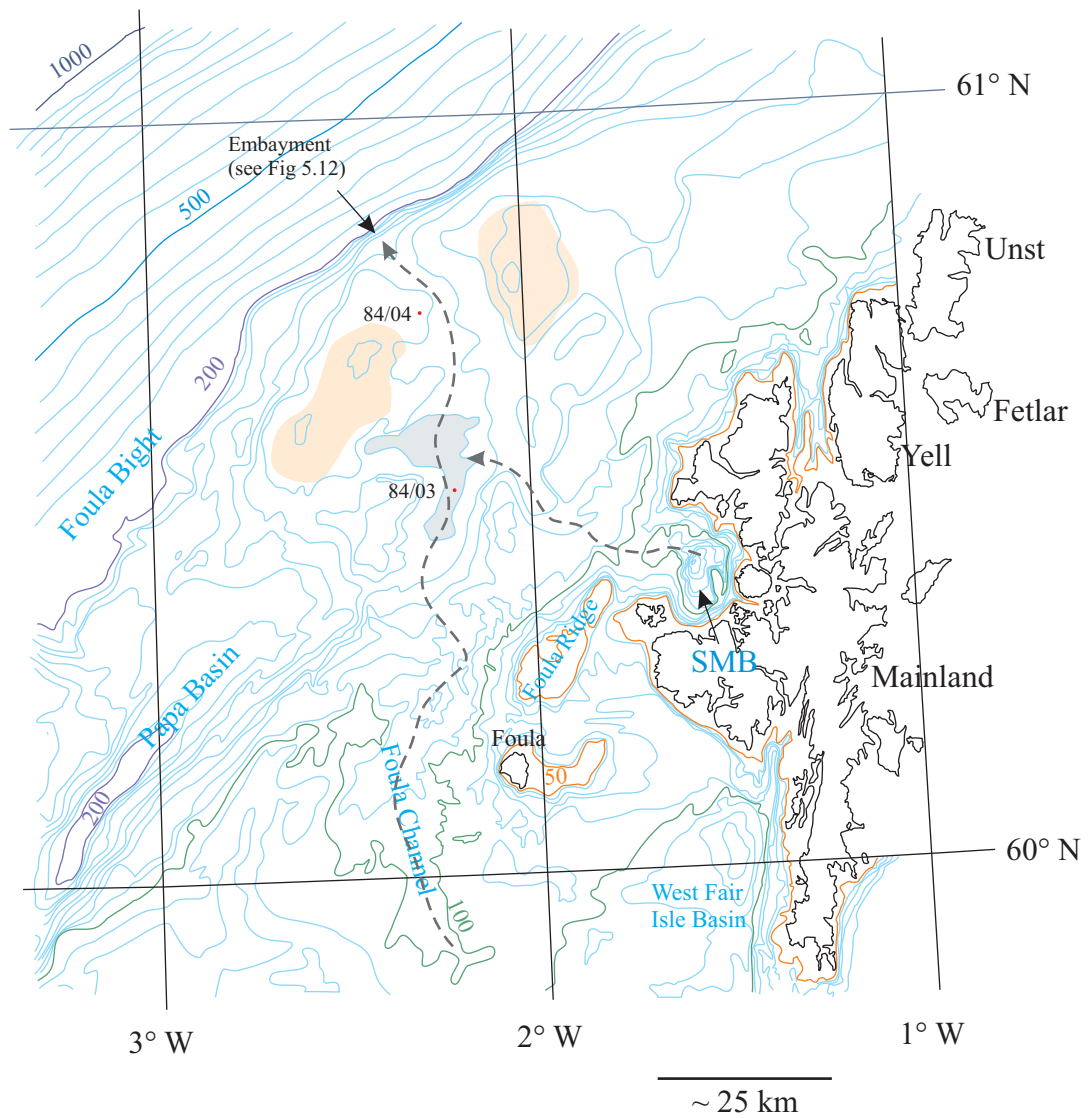


Figure 5.11. Bathymetry of the shelf in Area H, showing location of boreholes 84/03 and 84/04. The area of the intra-shelf basin below a water depth of 150m around borehole 84/03 is also highlighted. The isobath interval between 50m and 200m is 10m. Below 200m the interval is 50m.



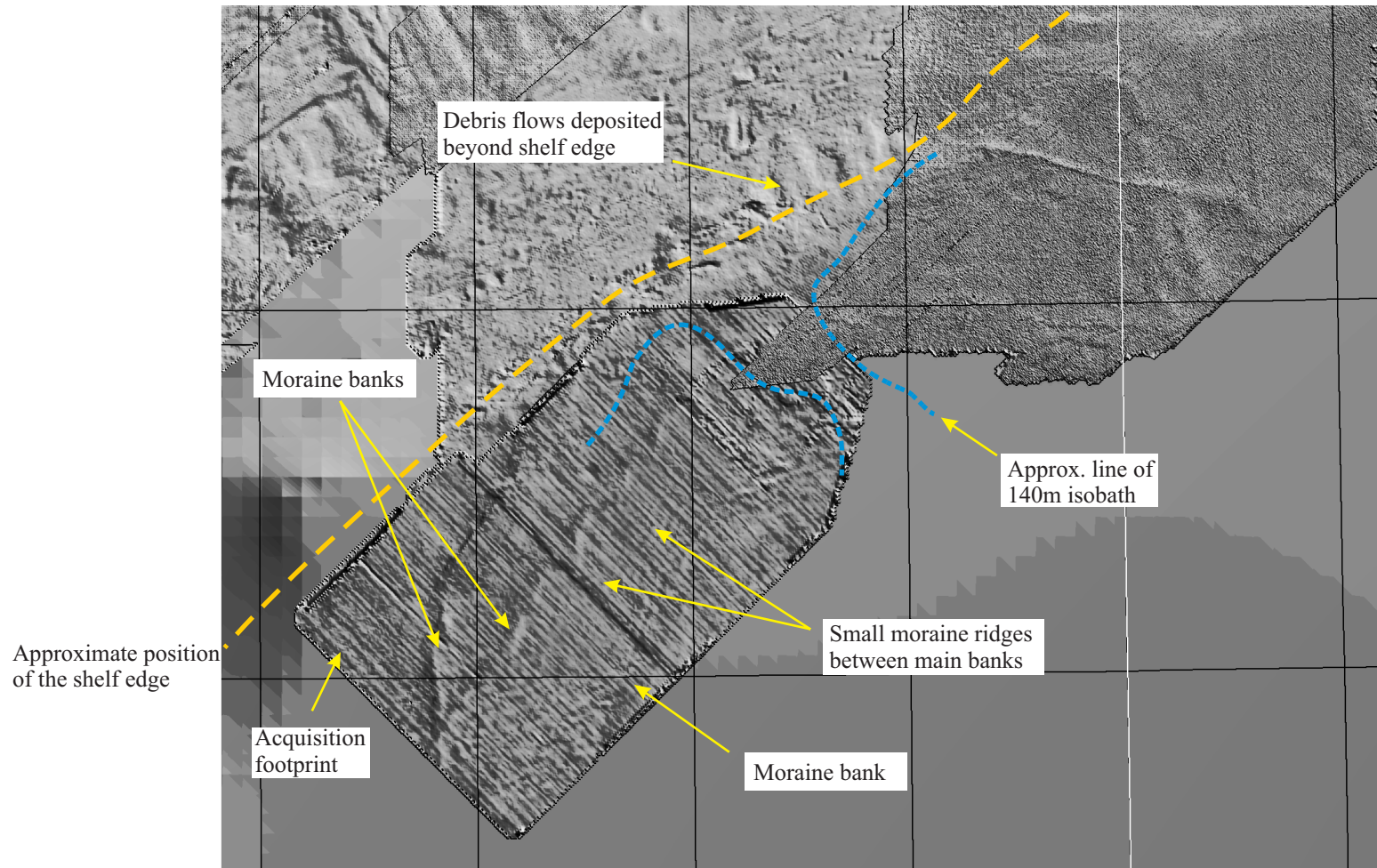


Figure 5.12. Enlarged part of seabed image from the outer shelf in area H, showing recessional moraine ridges. The 140m isobath marks the approximate margins of the outflow channel from the mid-shelf.

push-moraines at the ice front (Powell and Domack, 1995). The smaller ridges which lie between the larger moraines are interpreted as small recessional moraines, possibly representing annual retreat phases (Stoker et al, in press). The outermost ridge intersects with the shelf edge, where it passes into a series of small lobes, which are interpreted as debris flows. The zone occupied by the debris flows coincides with the small embayment formed by the exit of the cross shelf pathway described from the bathymetry above. The presence of the debris flows is important as it demonstrates that the grounded ice near the shelf edge was close to floating. With the slight increase in water depth provided by the shelf break, the ice achieved enough buoyancy to de-couple from the seabed leading to deposition as debris flows rather than moraine ridges, and indicating that the grounded ice did not travel beyond the shelf break.

Seismic data

Seismic survey coverage of area H is not particularly good, as it lies at the boundary of three different survey grids. The lines presented below (Figures 5.13 and 5.14) are taken from the 1979 survey (Figure 5.10) and cover the areas sampled by the boreholes 84/03 and 84/04. The base of the Otter Bank sequence is identified on the seismic data both from correlation across the shelf with lines in the southern transect, and from borehole 84/04 on line 79/14-35 (Figure 5.14) where it is defined by Holmes (1991).

The eastern end of airgun line 79/14-24 (Figure 5.13) covers the area of the mid-shelf sampled by borehole 84/03. The data shows a smooth seabed in the immediate area of borehole 84/03, which lies in a small basin developed along the line of the Shetland Spine Fault, and is probably an extension of the Papa Basin (Figure 5.11). To the west of the basin, the seabed becomes more irregular and rises by about 20m, before becoming smoother again and sloping slightly towards the outer shelf and into the Foula Bight. The unconformity at the base of the Quaternary glacial sequence is difficult to identify due to the presence of seabed multiples. The base of the Otter Bank sequence at the site of 84/03 lies at a depth of about 26m, but further westwards towards the outer shelf reaches a depth of *c.* 35m. On the mid to outer

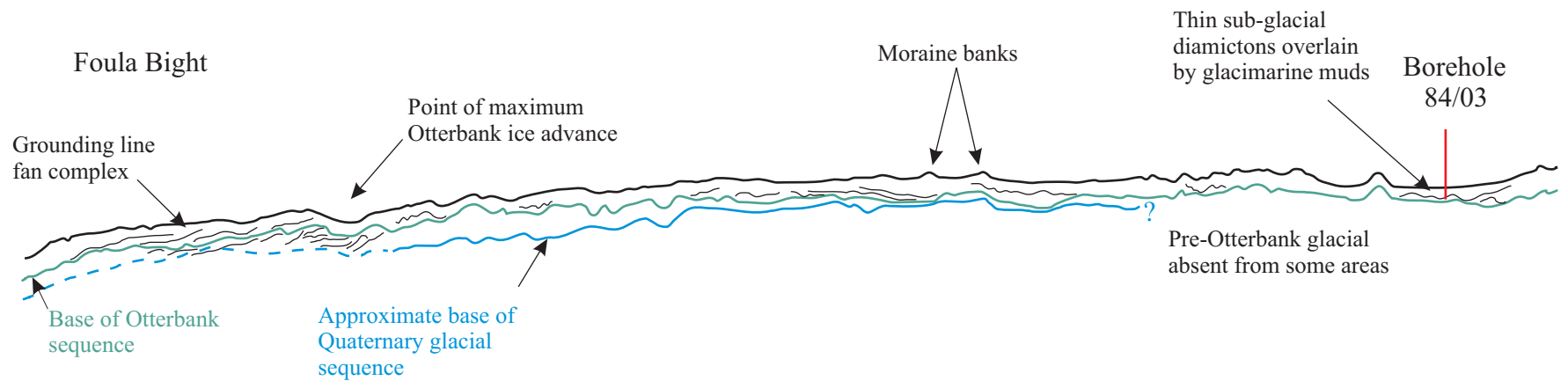
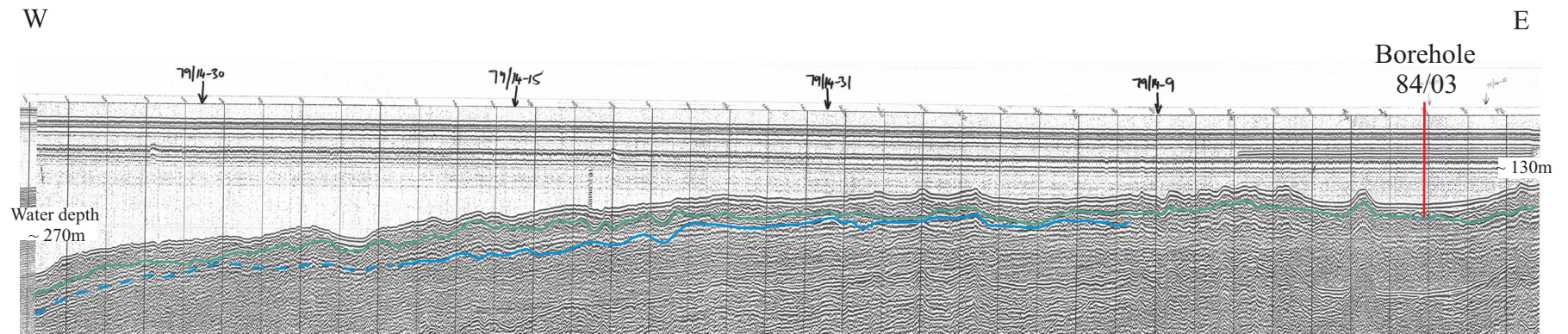


Figure 5.13. Seismic survey line 79/14-24, showing a thin Otter Bank sequence on the mid-shelf which thickens westwards towards the Foula Bight. At the shelf edge the Otter Bank sequence shows prograding reflections with limited aggradation, forming a grounding line fan complex within the Foula Bight. The small moraine banks on the outer shelf are continuations of those seen further north on the seabed image

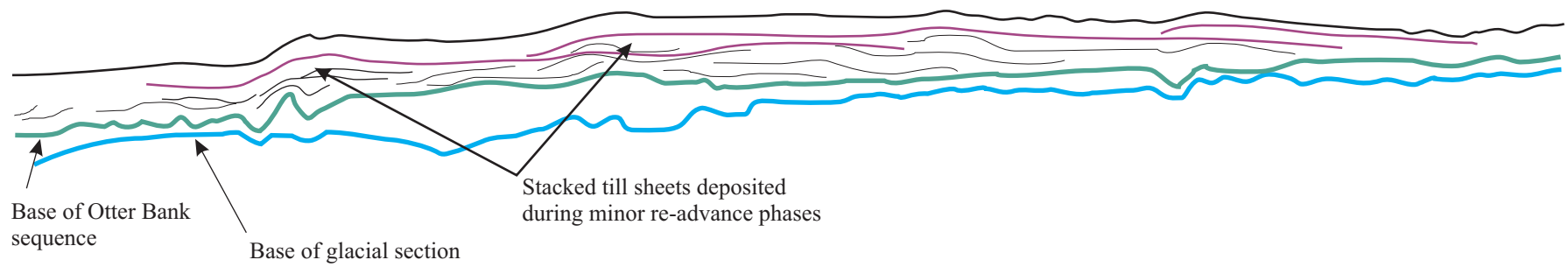
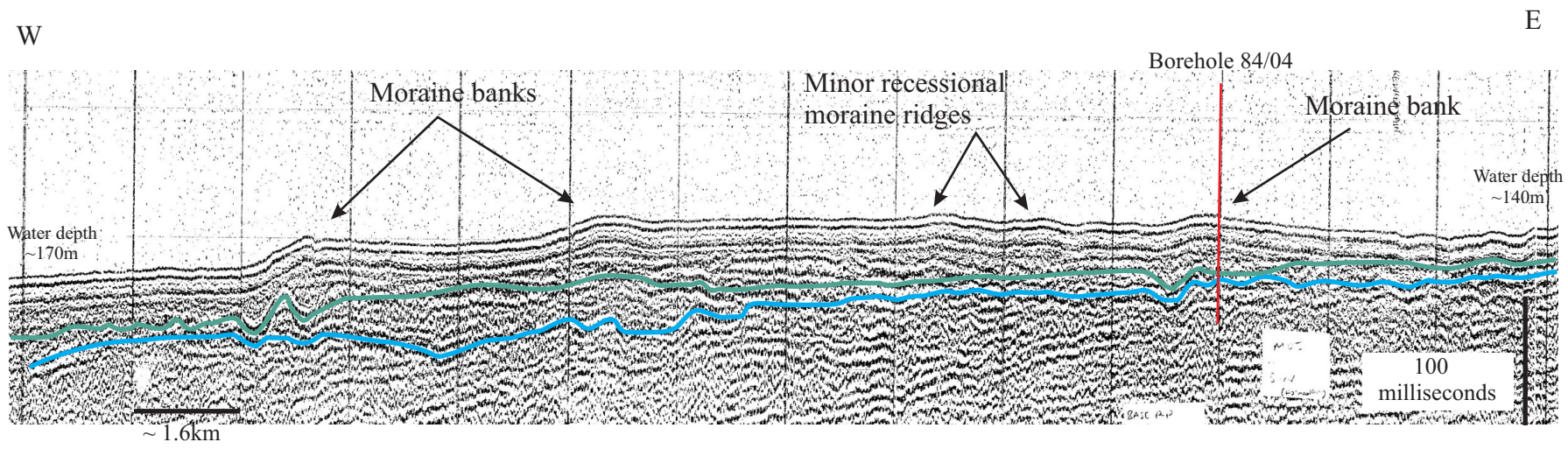


Figure 5.14. Seismic sparker survey line 79/14-35, showing stacked till sheets and moraine banks formed during an overall glacial retreat. The position of the moraines represents former grounding lines, probably formed during a period of rapid sea-level rise which allowed aggradation of the tills. Unfortunately much of the detail is lost in a thick seabed pulse.

shelf the Otter Bank sequence comprises sub-parallel to convoluted reflections, although multiples again obscure much of the detail. On the outer shelf the sea floor drops into the Foula Bight and enters the area covered in section F of Chapter 4. The Otter Bank sequence at the outermost shelf comprises a prograding, and slightly aggrading sequence of sub-parallel reflections forming sheet-like deposits.

Such packages are not seen on parallel survey lines to the north or south of line 24, suggesting that this zone is a focus for meltwater discharge and the prograding reflections are interpreted to be part of a grounding line fan. It is inferred that these packages are probably composed of water-lain diamicton deposited at the grounding line by meltwater discharge. The prograding package also marks the maximum limit of ice advance which lies at a water depth of approximately 220m. This style of architecture is associated with point sources of meltwater at the floating terminae of tidewater glaciers and ice sheets (Powell and Domack, 1995).

The pre-Otter Bank glacial sequence and the basal unconformity are more difficult to define than the later glacial deposits. At borehole 84/03 the base of the pre-Otter Bank glacial sequence is obscured and the sequence not sampled. Towards the shelf edge the older sequence shows the same pattern of prograding reflections seen in the Otter Bank sequence and was probably deposited in a similar setting.

Sparker line 79/14-35 (Figure 5.14) lies to the north of line 24, covering the outer shelf and slope, and the area around borehole 84/04 (Figure 5.10). The shelf section of the survey line shows a relatively smooth seabed with low amplitude undulations and banks which correspond to the moraine ridges observed on the seabed image (Figures 5.12 and 5.14). The base of the glacial section at the borehole locality lies at approximately 58m, deepening to 85m towards the shelf edge. The Otter Bank sequence maintains a roughly even thickness of 30 - 40m towards the shelf edge, except where the moraine banks occur. Broadly, the Otter Bank sequence comprises sub-parallel to convoluted reflections, with small chaotic lenses forming the moraine banks. The outermost bank shows evidence of having been deformed by a forward pushing, causing a double crest to the bank. Beneath the outermost bank the basal

Otter Bank unconformity shows a short series of 30ms high convolutions. In other survey lines this feature has marked the point at which the Otter Bank ice sheet crossed over the buried pre-glacial shelfbreak. The moraine banks shown on line 79/14-35 (Figure 5.14) are interpreted as the result of pauses in the retreat of the ice-front, with slight elements of re-advance, in a similar way to the emplacement of the large moraines seen in area C in the south. The moraine banks mark the point of former grounding lines during a period of rising sea levels. Sea-level rise is deduced from the stacked aggradational nature of the moraine sequence which bears similarities to the “till tongue” model of King *et al* (1991) from the mid-Norwegian shelf. The resolution of the data from line 35 is insufficient to define the glacial marine muds which the model predicts should lie between the diamicton sheets at the outermost edge of the moraine banks.

5.2.2 Borehole data

Borehole data from the mid and outer shelf in area H is restricted limited recovery. The descriptions and interpretations are based upon observations recorded at the time of drilling in 1984 and were not re-examined for this study due to time constraints.

Borehole 84/03

Borehole 84/03 lies in 152m of water in the small intra-shelf basin west of St. Magnus Bay (Figures 5.11 and 5.13). Drilling of the borehole was hampered by cobbles and exceptionally hard diamictons which damaged the drill and eventually caused the abandonment of the hole. The data gap below 18m was caused by damage to the drill during coring to 18m. The hole was subsequently re-drilled to 24.40m using a rotary bit. Consequently, no samples were recovered from the interval 18-24m.. The borehole log is presented in Figure 5.15.

The base of the sampled sequence comprises 40cm of very hard massive diamicton with pebbles and cobbles predominantly of gneiss, quartzites and sandstones. The matrix also contains minor shell fragments. Correlation with the seismic data from line 79/14-24 (Figure 5.13) suggests this unit is part of the Otter Bank sequence, the base of which lies at about 26m. Overlying the diamicton is a thin unit of very soft,

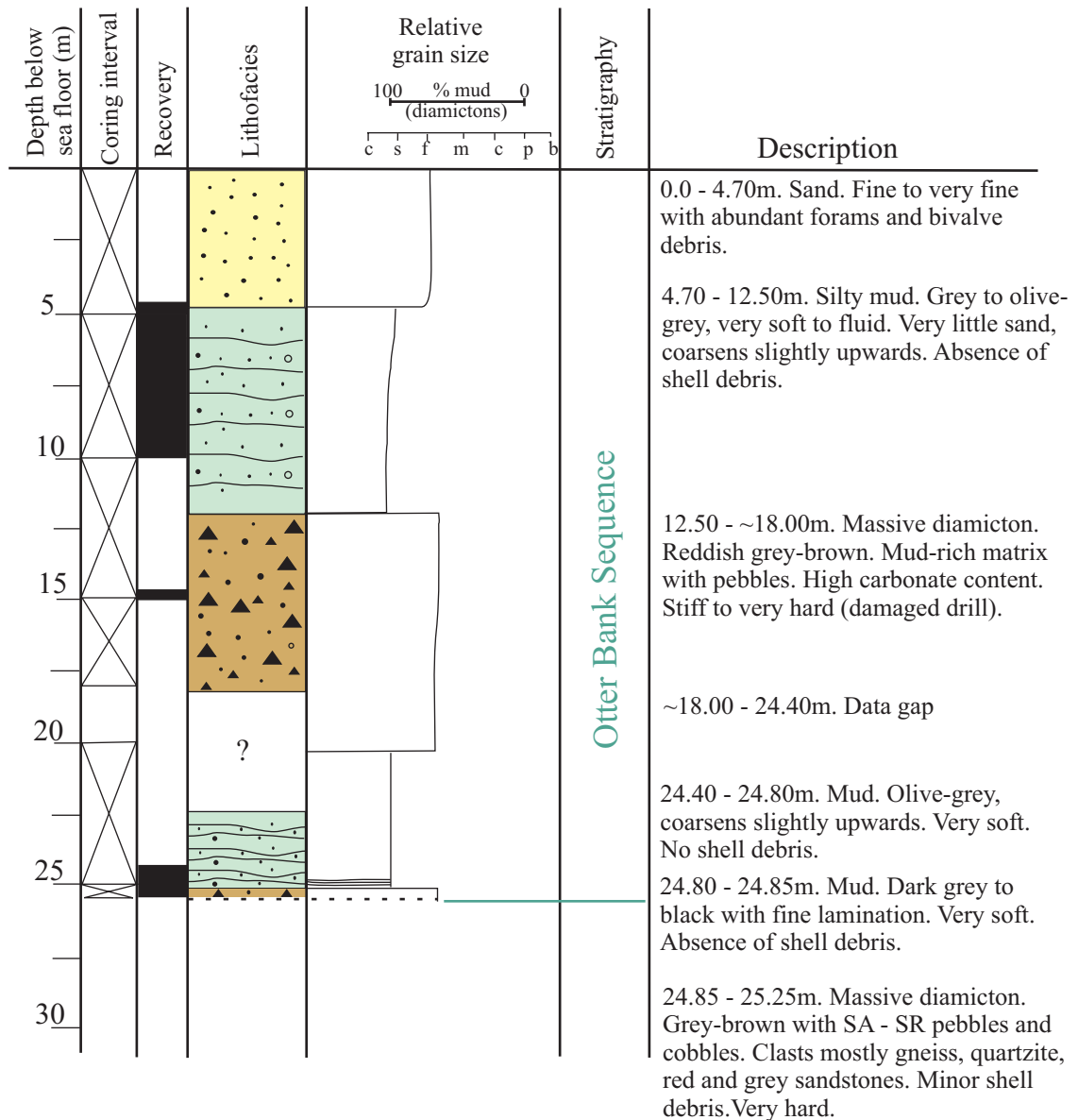


Figure 5.15. Core log from borehole 84/03

dark grey to black mud with fine lamination and an absence of shell material. A gradational upper boundary leads into 40cm of very soft olive-grey mud. Shell debris is also absent from this mud unit. Above the data gap of 24.40 – 18m lies approximately 5.5m of stiff to very hard, red-brown muddy diamicton. This unit is also reported as having a high carbonate content. Overlying the diamicton is approximately 8m of very soft silty grey mud. The mud contains very little sand grade sediment, but coarsens slightly upwards. Shell debris was not observed. The uppermost unit is composed of fine to very fine sand with abundant shell debris and forams.

The basal diamicton is interpreted as a water-lain or sub-glacial till. The small thickness and very hard nature of the sediment suggests that a sub-glacial origin is more likely, the till forming as a probable lodgement deposit trapped between the underlying rock and the base of the ice.

The thin black mud unit which overlies the diamicton indicates an oxygen depleted environment. The dark colour is caused by the presence of iron monosulphides, normally formed during the microbial decay of organic carbon by the reduction of iron oxides. The reduction of iron oxides to sulphides indicates an oxygen deficient seabed within the deepest parts of the intra-shelf basin. Causes of the oxygen deficiency could be lack of water circulation in the basin, most likely beneath a static ice body, or merely an influx of carbon available for microbial metabolism. Raised levels of microbial decay can quickly exhaust available free oxygen in sediments and produce an anoxic environment beneath the seafloor sediment surface (Briggs and Crowther, 1990).

Dark grey to black laminated muds with monosulphide banding have been reported from glacial marine settings in southwest Sweden which are interpreted as forming in an ice-distal setting during summer influxes of sediment caused by increased melting (Stevens, 1990). However, The examples from Sweden contain granular sediment and clasts frequently up to cobble size. Such grain sizes are absent from the muds in 84/03. The fine-grained nature of the mud indicates a lack of current activity. It also

shows a lack of sediment input from iceberg rafting, precluding deposition in an ice-proximal or inner distal setting. The remaining possible interpretations of the black muds are either deposition beneath a floating ice shelf, or deposition in a non-glacial setting. The gradational upward change from black to olive-grey, together with the slight coarsening in maximum grain size, indicates a slow change in depositional environment. This change is most likely related to an increase in current activity, increasing water circulation. A speculative interpretation of the muds is of deposition beneath floating sea-ice, prior to a glacial re-advance which deposited the overlying diamicton. Without further data, both from the upper boundary and from micropalaeontology, a more definitive interpretation is not possible.

The massive diamicton unit which overlies the data gap is interpreted as a probable sub-glacial till. This is based primarily on the over-compacted nature of the sediment. However, its position overlying a soft mud is enigmatic. Overriding of the muds by an advancing ice sheet would be expected to re-work much of the sediment, especially where ice-loading was forming a compact till. There are several possible explanations for this problem. The underlying mud may have been frozen at the time of till emplacement (Benn and Evans, 1998) which would have helped protect the muds from re-working. Pore water pressure within the muds may have been sufficiently high to prevent large scale downcutting by the ice, or the advancing ice was sufficiently buoyant to prevent significant downcutting. A combination of these three scenarios is also possible. The most likely explanation is of a semi-buoyant ice sheet gliding over the easily deformable muds, eroding the upper part of the mud unit and incorporating the sediment into the muddy diamicton. Once the ice sheet had covered the muds, pore water pressure could have prevented further downcutting but still resulted in a compact till deposit. A further possibility is that the diamicton is not actually over-compacted, but is in fact semi-cemented by the reportedly high carbonate content of the unit. This would remove the necessity for the till to have been emplaced in a sub-glacial setting, allow the muds to be covered by water-lain tills in an ice-proximal setting from a semi-buoyant ice sheet.

The upper mud unit between 4.70-12.50m is similar to the lower muds and has presumably been deposited in a similar setting. Again, the lack of coarse sediment and clasts precludes deposition in a setting exposed to iceberg rafting.

The sand unit at the top of borehole 84/03 is interpreted as having been formed by reworking of the underlying muds. This is supported by the fine to very fine grain size of the sand, which is the only coarse sediment present in the muds. The presence of an abundant marine fauna indicates normal marine conditions were established during the reworking phase, which probably represents the modern environment on the shelf.

Borehole 84/04

Borehole 84/04 lies near the shelf edge in area H between the moraine ridges observed on the seabed image (Figure 5.10). The borehole site is on the southwestern side of the outflow pathway described in the bathymetry section above (Figure 5.11) in a water depth of 134m. In total, the borehole sampled over 90m of Quaternary sediments, all of which show a cold severe climate. However, only the section down to approximately 60m is considered to be of Late Pleistocene age and fully relevant to this study, with the base of the Otter Bank sequence at about 40m (Holmes, 1991).

Recovery from borehole 84/04 was very poor, with most coring runs retrieving only small samples of sediments (Figure 5.16). The base of the sequence comprises approximately 10m of dark brown muddy diamicton with pebbles overlying brown sand. Overlying the diamicton is a red-brown muddy sand with occasional sub-rounded cobbles and pebbles of gneiss and red sandstones. Overlying the sands, and forming the bulk of the Otter Bank sequence is a thick sequence of dark grey massive diamictons. These are generally very stiff to hard and contain sub-angular pebbles and cobbles of gneiss and sandstones. Shell fragments are common in some sections, including thick-shelled bivalves towards the top of the unit. The uppermost unit is composed of yellow-brown medium to coarse, quartz-rich sand. lithic content of this unit is very low, probably less than 2%.

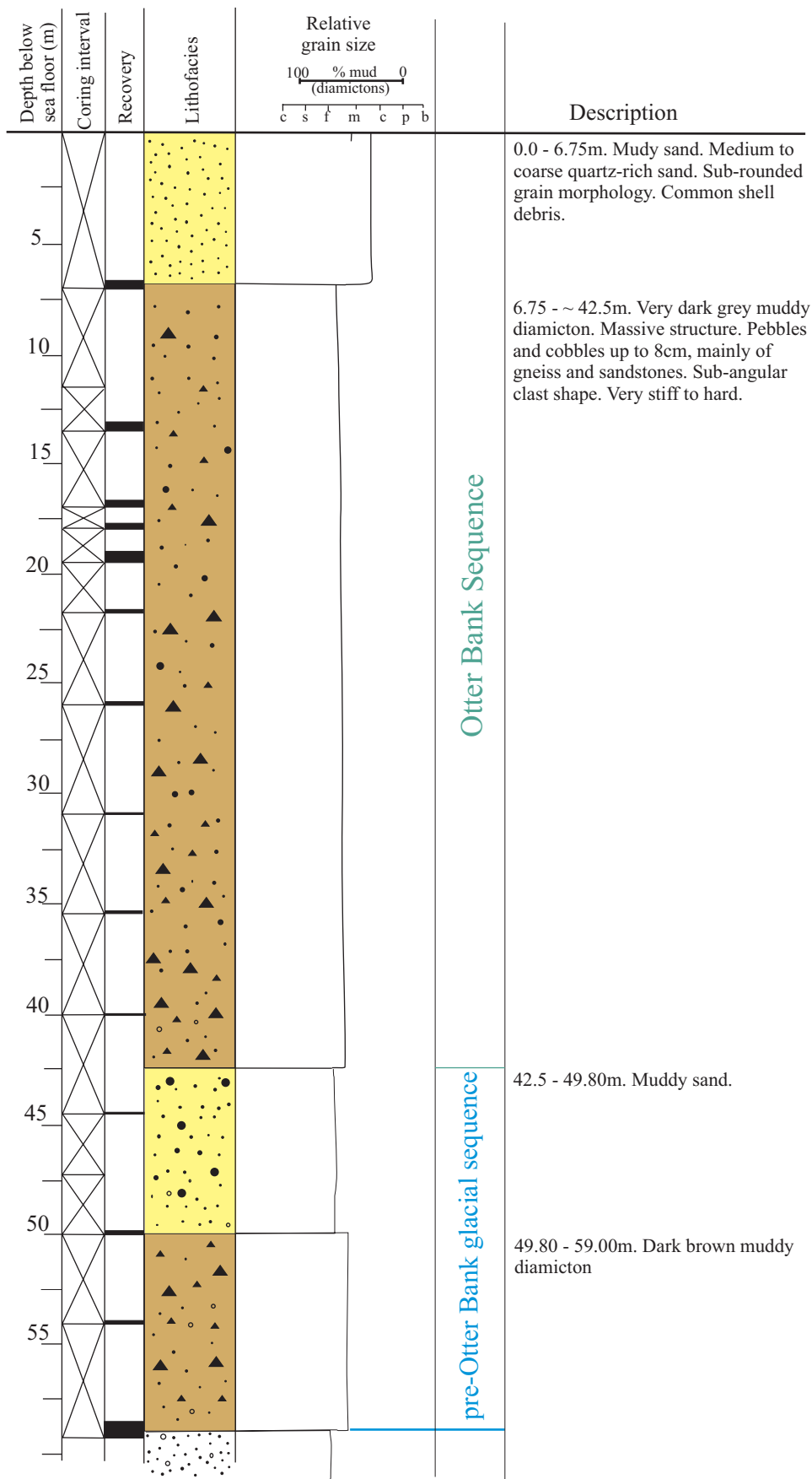


Figure 5.16. Core log from borehole 84/04

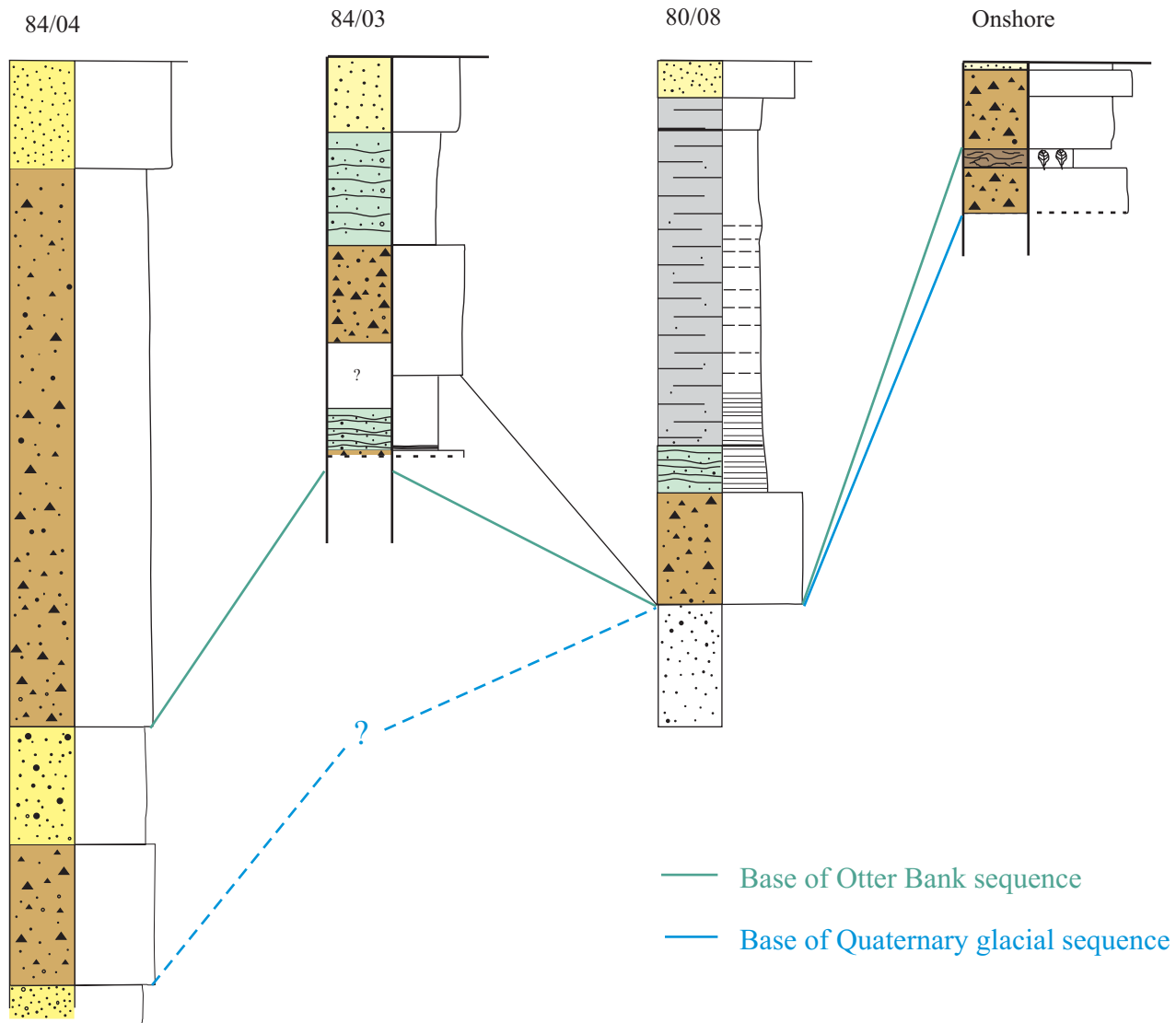
Generally the quantity of sediment actually recovered makes interpretation difficult. The difference in colour between the pre-Otter Bank diamictons and those from the Otter Bank sequence reflects the residence time, the older sediments having slowly oxidised to a greater degree than the younger deposits. The thickness of the Otter Bank sediments suggests that the pathway formed a major outflow for ice to the shelf edge in the local area.

5.2.3 Interpretation of Area H

The transect across the shelf from the inner-shelf to the shelf edge shows features not seen in the southern transect. The seismic survey data shows a thin Otter Bank sequence on the inner and mid shelf which shows convoluted or undulating structure. Combined with the hardness of many of the diamicton samples from the boreholes, this suggests that these are sub-glacial deposits, or have at least been overridden by ice. The outer shelf moraines and associated stacked diamictons forming the grounding line wedges indicate decoupling of the ice near the shelf edge. The presence of anoxic muds in the deeper parts of the mid-shelf, but the absence of muds on the outer shelf suggests the outlet at the shelf edge was fed by more than one source of ice which by-passed the mid-shelf basin at a time when sea-ice prevented the input of ice rafted debris from icebergs.

5.3 Summary of Transect B

The transect from the onshore zones around the margins of St. Magnus Bay to the shelf break shows a set of features indicative of a different style of glacial activity when compared to the more southerly Transect A. A correlation of the boreholes is presented in Figure 5.17 but is slightly speculative as the sequences are more incomplete towards the onshore localities. The onshore localities around St. Magnus Bay show thin sandy diamictons with a high clay content. There is also a component of large clasts up to boulders of over 1m diameter which, due to the nature of marine sampling could not be retrieved even if such clasts were present in the offshore settings. The onshore deposits are restricted to thin infills in topographic hollows, suggesting the cover may have been initially thin and removed by later weathering.



The St. Magnus Bay borehole shows the base of the inshore sequence consists of a thin massive diamicton, covered by a sequence of organic-rich muds, interpreted to be from the closing stages of deglaciation. The basin within St. Magnus Bay is likely to have been the last marine setting to become ice free, suggesting the diamictons at the base of the section are probably the youngest recorded anywhere in the late Devensian sequence of the West Shetland area.

The mid-shelf area around borehole 84/03 suggests parts of the deeper route to the shelf edge may have been occupied by floating ice rather than grounded ice at some stage. The very thin basal diamicton indicates a probable lack of ice throughput, followed by a period of anoxic bottom waters beneath floating ice prior to a more active glaciation which formed the thicker diamictons.

The sequence on the outer shelf is difficult to sub-divide, although a pre-Otterbank phase of glacial activity is indicated. The thick diamictons suggest the outer shelf in the region of borehole 84/04 was occupied by active ice throughout the Otterbank glaciation until final retreat. The moraines on the outer shelf indicate an oscillating ice front, with the second of the two outermost moraines appearing to mark a re-advance which removed parts of the earlier moraine.

6. Reconstruction of the Otter Bank (late Devensian) Glaciation on the West Shetland Margin

Analysis and amalgamation of all the data from the two detailed transects, together with seismic data from the remainder of the shelf and slope, enabled the mapping of the basal Otter Bank unconformity, and hence sediment thickness, across large parts of the West Shetland Margin. The detailed data is presented in Figure 6.1 in the form of an isopach map. In addition, a map of the basal surface of the Otter Bank sequence, constructed by stripping off all of the sediments above the unconformity, highlights the intra-shelf basins and probable ice drainage pathways (Figure 6.2). This surface represents a compound surface created by all the different phases of glaciation that have contributed to the development of the Otter Bank sequence. On the inner shelf where the Otter Bank sediments are no longer present, the morphology of the original palaeosurface can only be guessed at, but is unlikely to have been significantly different from the modern seabed.

6.1 Morphology and geometry of the Otter Bank sequence

An examination of the distribution of the Otter Bank sequence in Figure 6.1 shows that on the majority of the inner shelf, sediment cover is very thin or absent. The approximate boundary between thin cover or rock outcrop and measurable sediment cover equates to the 100m isobath. This boundary highlights the channels (labelled 1 – 3 in Figure 6.1) which drain the plateau-like area above this depth to the west and south-west of Shetland. The thicker cover of glacial sediments within the channels indicates they were important ice sheet drainage routes and that deposition was water depth related. This is also true of St. Magnus Bay and the deeper pathway which leads from it. The inner shelf west of St. Magnus Bay shows several areas of thicker sediment (points 4 and 5 in Figure 6.1). Analysis of the seismic data from these areas shows the thicker deposits are the result of infilling in over-deepened sections of the outlet path. Both the over-deepened sections occur at the convergence of minor channels highlighted in Figure 6.2, suggesting it may be the convergence of ice bodies which is responsible for forming the deeper sections.

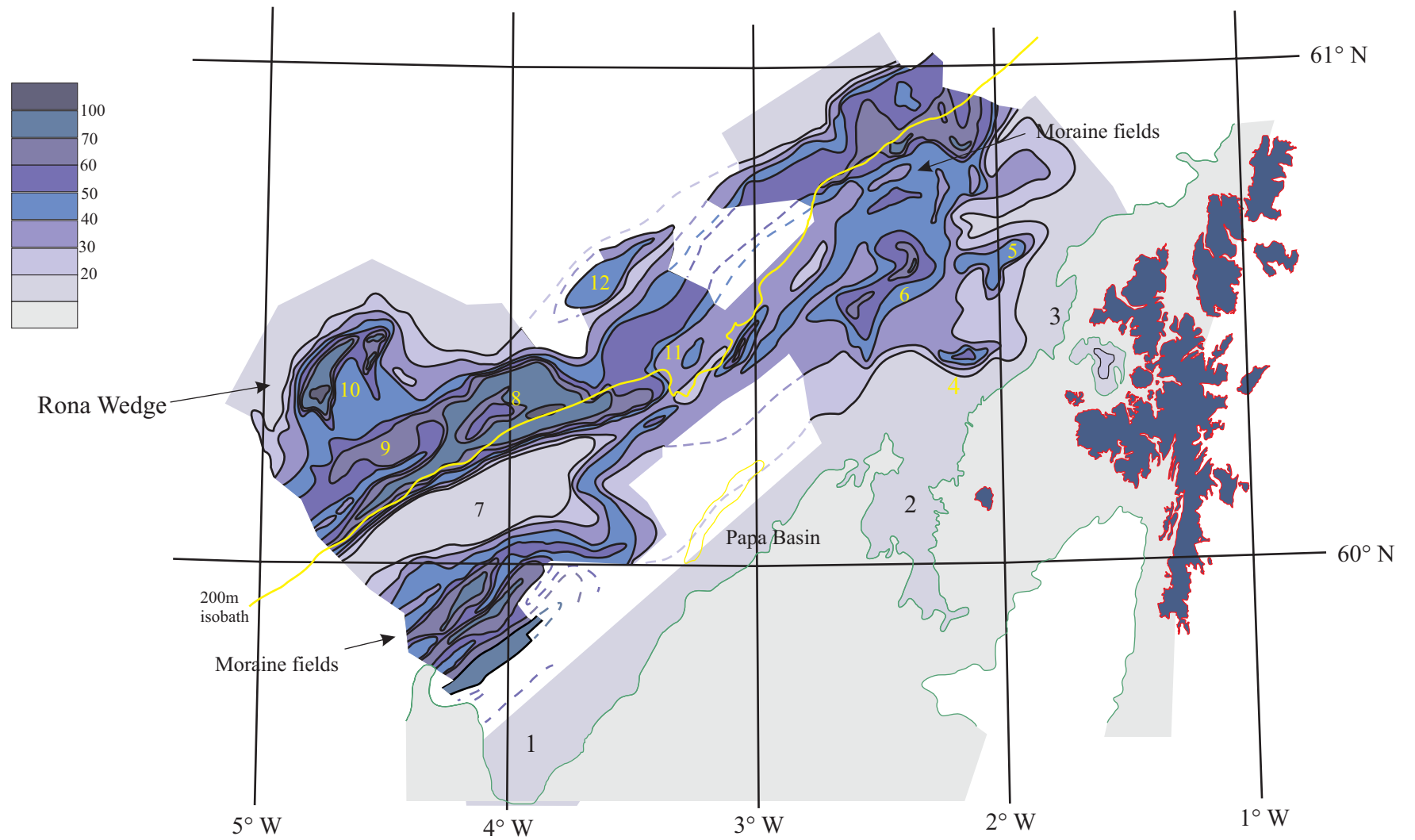


Figure 6.1. Isopach map of the Otter Bank sequence derived from 2D seismic data. Numbers are referred to in the text..

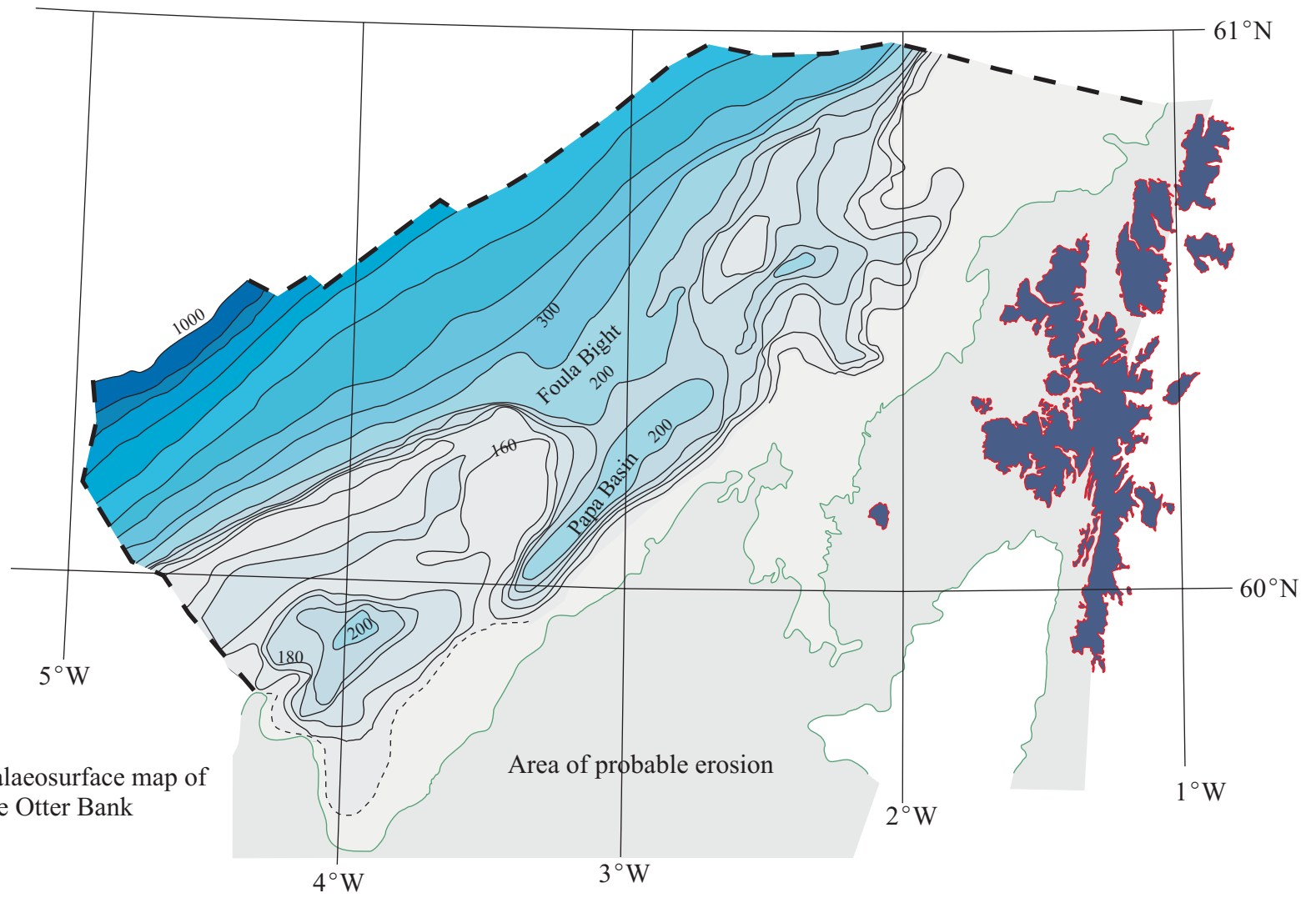


Figure 6.2. Palaeosurface map of the base of the Otter Bank sequence.

The mid-to-outer shelf zone of the West Shetland Shelf is more complex than the inner shelf. In the south, the pattern of Otter Bank sediment distribution is dominated by sub-parallel zones of thick and thin sediments caused by the large moraine banks described in Chapter 4. Data for the south-eastern edge of the moraine field is lacking, being beyond the edge of the 1983 survey. Otter Bank sequence thickness within the Papa Basin is difficult to define due to difficulties in identification of the basal unconformity. However, the total glacial thickness is less than 40 milliseconds, (approx. 30m, see Figure 3.8) and is concentrated on the north-western margin of the basin, suggesting that the floor of the Papa Basin has a relatively thin covering of Otter Bank sediment. To the west of St. Magnus Bay, the shallow basin feature described from the modern bathymetry (section 5.2.1, Figure 5.11) shows a thickness of Otter Bank sediments of up to 55m (point 6, Figures 6.1 and 6.2). The accumulation at this point is partly filling a pre-existing basin. The thickened sequence, and the elongate basin lie at the confluence of the outlets from the Foula Channel, St. Magnus Bay and a minor pathway from the Papa Basin, again suggesting the basin has formed as a result of the convergence. The zone of thicker sediments at point 6 extends to the northwest and merges with the accumulation at the shelf edge. The moraines on the outer shelf (Figure 5.12) appear as narrow bands of thicker sediment (Figure 6.1).

The outer shelf in the south of the study area shows a large area of much thinner sediments (point 7, Figure 6.1) which reaches from the outer edge of the large moraine banks almost as far as the shelf edge. This plateau-like area is the result of an eastwards slope to the shelf in this region which apparently pre-dates the Otter Bank sequence (Figure 6.2). The effect of the slope at the time of Otter Bank deposition would have been a shallowing of the water towards the shelf edge, causing a loss of buoyancy to the ice sheet and the deposition of a much thinner sequence. This is supported by the seismic data from this area, shown in Figures 4.26c and 4.27, which shows two thin phases of Otter Bank sediments (OB1 and OB2) formed when ice was at the shelf edge, demonstrating the same relationship of water depth controlling deposition as seen on the inner shelf.

The shelf edge – upper slope transition in the south west is occupied by a long, rapidly thickening wedge of sediments, seen in section 4.4 to be composed of stacked structureless deposits forming till sheets, which pass into wedge shaped debris flows at the top of the slope. The upper to mid slope shows two distinct architectures in this area. At point 8 (Figure 6.1) the upper slope comprises a very thick (up to 80m) stack of debris flows which form a bulge on the seabed image (Figure 4.36) and show evidence of small scale sliding and shearing/slumping (section 4.4). Downslope of the bulge, the Otter Bank sequence thins rapidly down to little more than 20m on the lower slope. This corresponds to a zone of limited debris flows and appears as a smooth seabed on the image data in Figure 4.36. In contrast, the adjacent part of the upper slope at point 9 (Figure 6.1) initially shows a similar thick wedge of sediment, but this thins along a narrow zone of the upper shelf. This zone coincides with the proposed ice margin of Holmes *et al* (2003). The remainder of the upper and mid slope at 9 is thinner than the same position at 8, However, directly below point 9 on the lower slope is the extensive and prominent debris flow field (point 10, Figure 6.1) described in section 4.5 and observed on the seabed image (Figure 4.49). This suggests that the volume of sediment which forms the lower slope debris flows at 10, is derived from the upper slope zone at 9 as a result of an unstable slope region. This instability has also affected the slope at 8 but manifests itself as numerous small movements rather than larger events. It also suggests the degree of instability decreases to the northeast along the slope.

Within the Foula Bight (point 11, Figure 6.1), the outer shelf and slope show a different distribution of Otter Bank sediments. The thick shelf edge wedge is absent, showing instead a broader zone of thicker sediments which reaches a maximum thickness of about 50m (point 11). The reason for the difference is due to the deeper water in the Foula Bight which prevented advance of the grounded ice sheet to the edge of the shelf. The seismic data from area F, transect A (Section 4.6.1) shows a depositional architecture consistent with a buoyant, and probably floating ice margin as opposed to a grounded margin (see section 6.2 below). The Fans at the base of the slope below the Foula Bight (point 12) are composed of sandier sediments probably

derived from turbidity currents associated with gully formation, rather than debris flows.

In the north-east, the Otter Bank distribution again shows a thick upper slope accumulation derived from debris flows at the shelf edge. The outer shelf - shelf edge data indicates stacked till sheets deposited with an element of decoupling due to ice buoyancy. Unlike the slope section in the south, the debris flows are confined to the upper and mid slope, and fail to reach the slope base. This appears to be due to the lack of sediment volume delivered to the shelf edge, indicating a smaller ice body or one transporting less sediment.

6.2 Correlation of the moraine banks

The isopach and palaeosurface maps show the overall distribution of the Otter Bank sequence, and the nature of the surface on which it lies. However, they do not define individual ice advances and only provide partial information as to the position and movement of the West Shetland ice sheet. An examination of the moraine distribution and ice marginal features provides a more detailed history of ice movement on the shelf and, coupled with the slope seismic data, enables a much fuller model to be developed. The distribution and proposed correlation of the Otter Bank moraines is presented in Figure 6.3. The moraines were identified from seismic profiles as mounded deposits and measured roughly using the break-of-slope at their bases to define their edges. The correlation presented in Figure 6.3 is a best-fit scenario, limited by the spacing of the seismic survey lines. The small area of large moraines visible on the seabed image (Figure 4.18) shows that the actual shape of the moraines is probably much more complex than that depicted.

A cursory examination of the distribution of moraines on the shelf shows a concentration of features in the south west. In addition to the large moraines described in section 4.3.4, there are a series of smaller moraines closer to the shelf edge. The central area east of the Foula Bight and the northern area north-west of St. Magnus Bay exhibit much smaller scale features and in smaller numbers. This is a

clear indicator that the southern area was more glacially active in terms of ice and sediment budgets.

Figure 6.3 shows a proposed maximum ice limit for the Otter Bank sequence glaciations which lies roughly along the shelf edge. In the south around zone A, the maximum advance is marked by the transition from sheet-like tills on the shelf to structureless debris flows on the slope. The exact point of maximum advance is impossible to define due to the lack of acoustic structure (Figures 4.26c and 4.27). The lack of detail may be due to the grounding line structures being destroyed as the material re-mobilised into the debris flows. This style of shelf edge deposition is very similar to that described from the Norwegian margin by Laberg and Vorren (1995) and Vorren and Laberg (1996).

Within the Foula Bight at point B, the structures marking the maximum ice advance are more dispersed. Seismic data shows the shelf edge to comprise a series of steps, related to the faulted structure of the Foula Bight, which are covered with Otter Bank sediments (Figure 4.61). The acoustic structure of the sediments reveals a prograding pattern of reflections with chaotic units between the main reflectors. This pattern is very similar to the modified “till tongue” model described by King et al (1991) on the mid-Norwegian Shelf where a pattern of advancing till sheets was deposited by an ice sheet with a floating terminus. The poorer development of the moraine banks in the Foula Bight, and the prograding-aggrading structure at the shelf edge are also similar to structures described by Alley et al (1989), resulting from deposition beneath a floating ice shelf on the Antarctic margin. The generally deeper water of the Foula Bight and the associated embayment forms a likely location for the formation of a floating ice shelf, although these are generally associated with cold (polar) ice sheets. Such a scenario would explain the distribution of Otter Bank sediments and lack of substantial moraines across this region and the greatly reduced shelf-edge sediment wedge.

However, an alternative explanation, and one which is more likely in a temperate ice sheet, involves a thin, unstable, ice sheet occupying the Foula Bight. The deeper

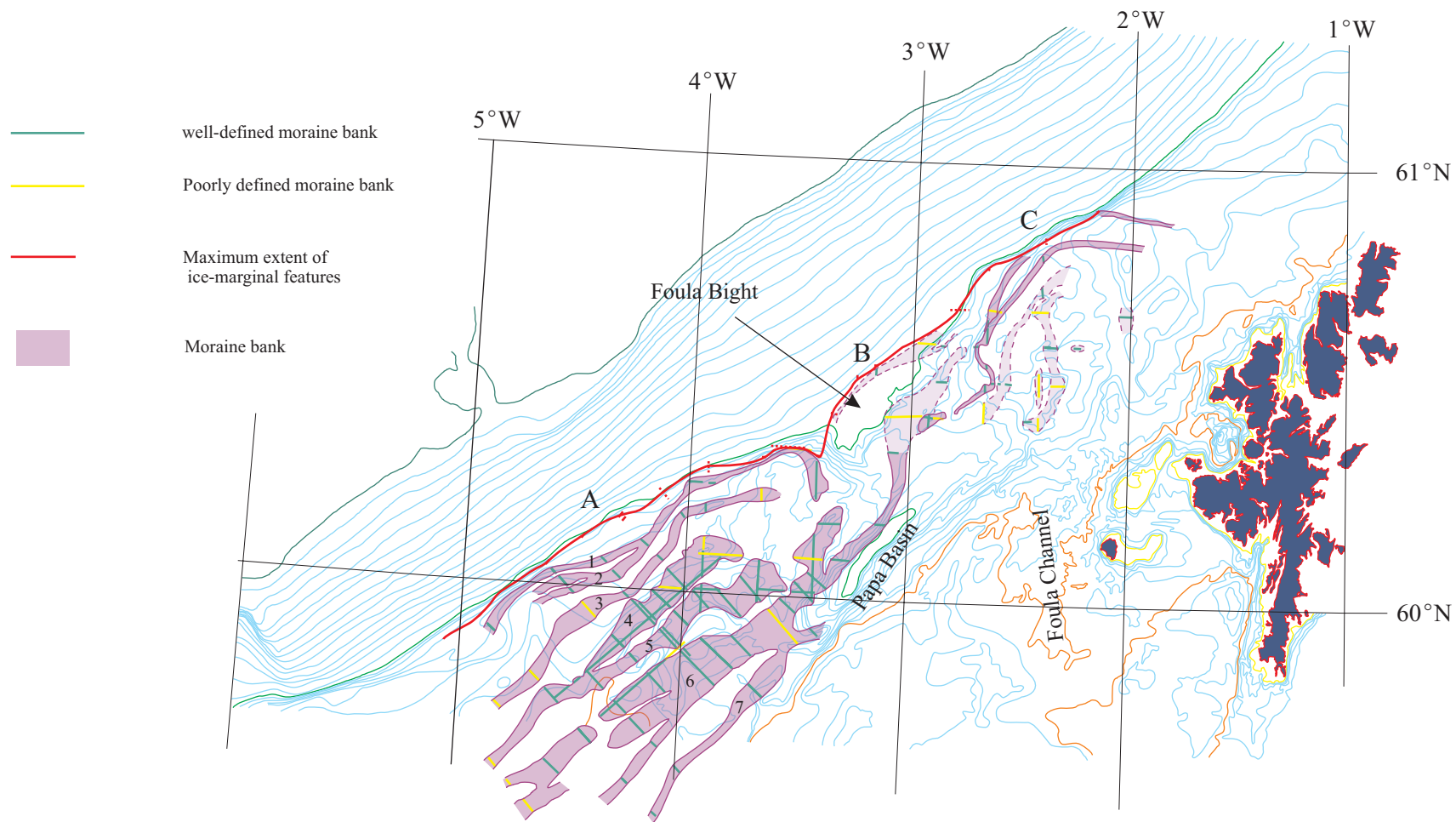


Figure 6.3. Distribution of moraines and ice marginal features from the Otterbank Sequence on the West Shetland Margin. Numbers are referred to in the text

water provides more buoyancy, possibly close to the point of floating, which hinders moraine formation and encourages iceberg production. At the beginning of a retreat phase, rapid ice decay within the Foula Bight could lead to the ice front receding quickly and sediments which would have been delivered to the upper slope from the ice margin are lost through iceberg generation and deposited elsewhere.

The ice marginal features in the north-western part of the shelf around point C (Figure 6.3) show a thin prograding pattern of stacked reflections (Figure 5.14). Data from borehole 84/04 shows these units to be composed of diamictons, suggesting ice advancing to the shelf edge and depositing stacked till sheets. The sediment architecture of the shelf edge at this point resembles a trough mouth fan, but on a much smaller scale than is usually associated with this type of feature. The relatively thin tills and small debris flows suggest slow moving ice and lack of transported sediment, or an ice sheet of limited extent. Alternatively, formation by a fast moving ice sheet with low erosive power could form the observed architecture (Dahlgren et al, 2002b). The arc-shaped moraines, and the size of the apparent catchment area suggest that the ice sheet was probably not large. This favours an interpretation of an ice sheet of limited extent and thickness, travelling rapidly across the shelf guided by the channel of deeper water which would have increased buoyancy and reduced erosive power. The core data from this area of the shelf, indicating the presence of soft muds preserved beneath hard till, also supports a buoyant ice sheet, suggesting that the deeper basin on the shelf was occupied at times by floating ice.

The features described above show that the maximum ice advance during Otter Bank phase glaciations reached the shelf edge along the entire margin, although the area within the Foula Bight shows some ambiguity due to a recessed shelf edge and poorer data. The seismic data presented in Chapter 4 for area C, has already shown that there were two major ice sheet advances to the shelf edge, prior to the moraine development shown in Figure 6.3. The moraines on the shelf, mapped out in Figure 7.3, all relate to various stages of an overall ice retreat from this maximum advance position and show local differences in the ice sheet behaviour. It also highlights some

similarities between the behaviour of the ice in the northern part of the shelf to that in the southern part.

6.3 Ice sheet reconstructions

The reconstruction of the Late Devensian ice sheet on the West Shetland Margin presented below requires the consideration of all the data examined in the earlier chapters. The reconstruction of the West Shetland ice sheet assumes the wider regional context of being part of the British Ice Sheet (BIS), as summarised by Stoker *et al* (1993), rather than being part of the Scandinavian Ice Sheet.

In addition to the moraine and sediment distribution data, the evidence from the shelf seismic lines showing a broad, shallow trough in the south of the margin is also included. The main evidence for ice advance to the shelf edge is summarised in Figure 6.4a, together with proposed ice routes. The maximum advance shown is equivalent to the stage forming OB2 on the outer shelf. The OB1 phase must have occupied a similar position at the shelf edge prior to a retreat and re-advance to the position depicted. However, the timescale between the OB1 maximum and the OB2 maximum, and the degree of ice retreat between the two is uncertain.

The formation of the moraine banks on the shelf in the south-west of the study area has already been addressed in Figure 4.32. However, this grouping also has a regional significance in terms of ice sheet behaviour. The outermost moraine banks, labelled 1 and 2 in Figure 6.3, mark pauses in the retreat of the ice following the second Otter Bank advance. Of particular importance is the outermost bank, 1, which is interpreted to curve south-eastwards at its north-eastern end. This marks the edge of the deeper water of the Foula Bight embayment and suggests ice retreated from the deeper water much more rapidly than in the shallower areas to the north and south. It is also possible the end of bank 1 was continuous with the spur on bank 6 near the Papa Basin, which would indicate a large ice tongue enclosed by the moraines (Figure 6.4b). Continuing ice retreat and a further pause at moraine bank 2 (Figure 6.4c) probably led to increasing isolation of the ice tongue, which eventually detached from the ice sheet and formed a stagnant ice mass. This possibility is

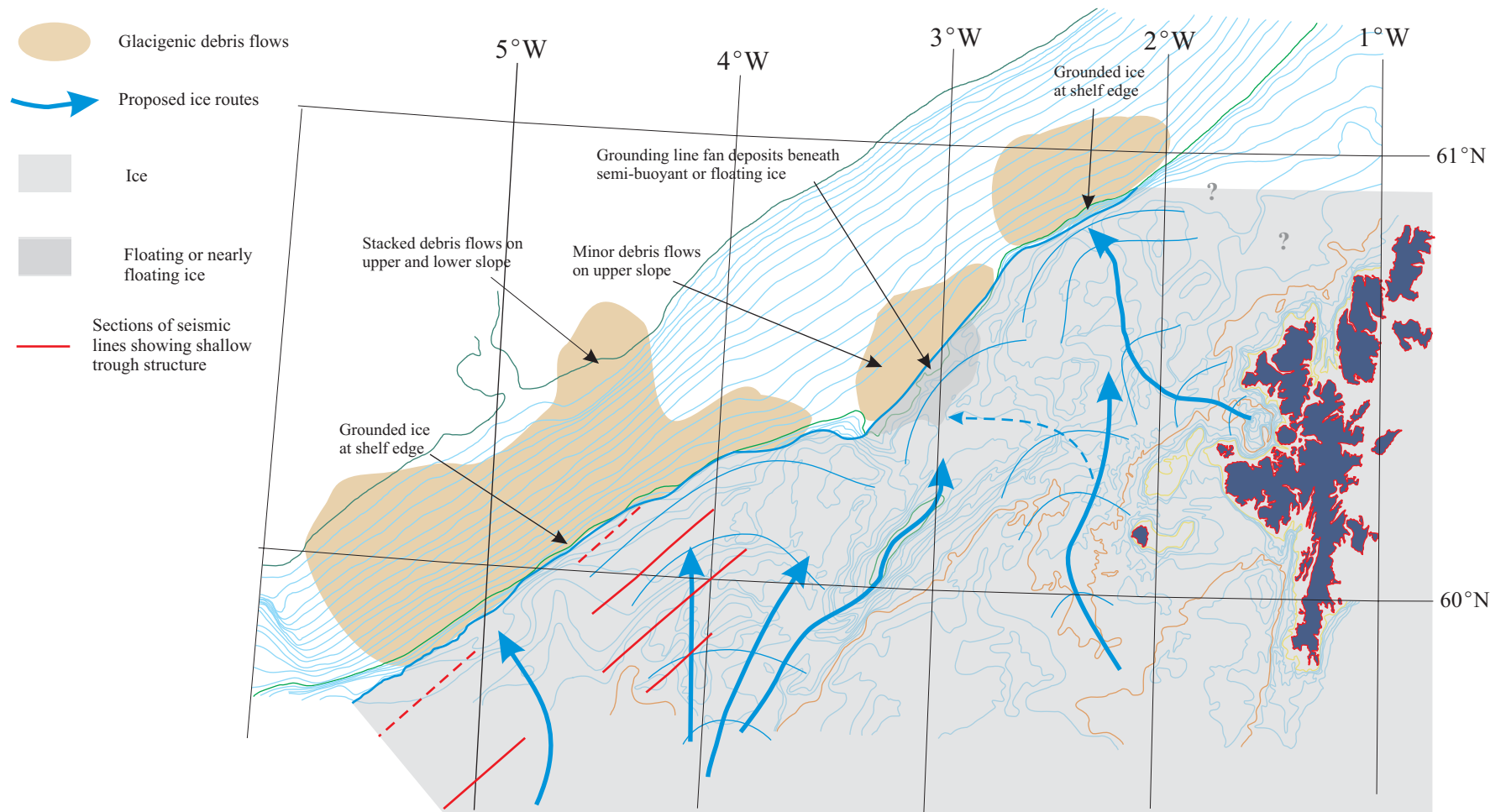


Figure 6.4a. Proposed position of Otter Bank ice sheet at the maximum advance of the second advance phase (OB2) which is thought to be equivalent to the Last Glacial Maximum (LGM). The earlier OB1 phase probably occupied a similar position, and suggests there may have been a double maximum during the Late Devensian glaciation.

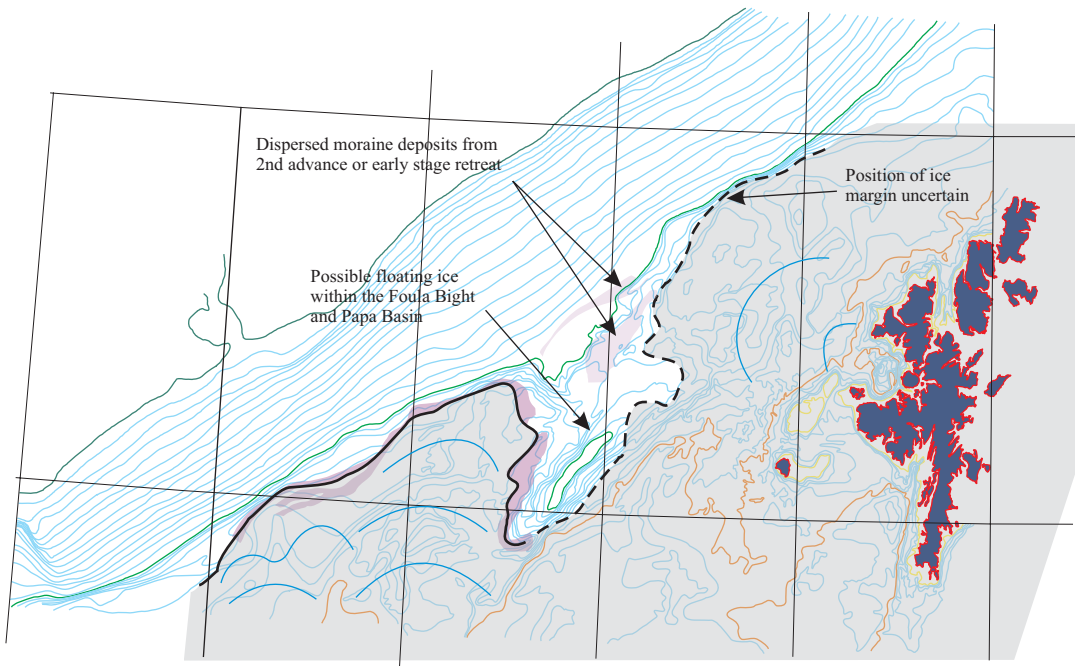


Figure 6.4B. First significant pause in ice retreat following the OB2 maximum advance (probably the LGM). Ice retreated fastest in the zones of deeper water away from the main ice routes

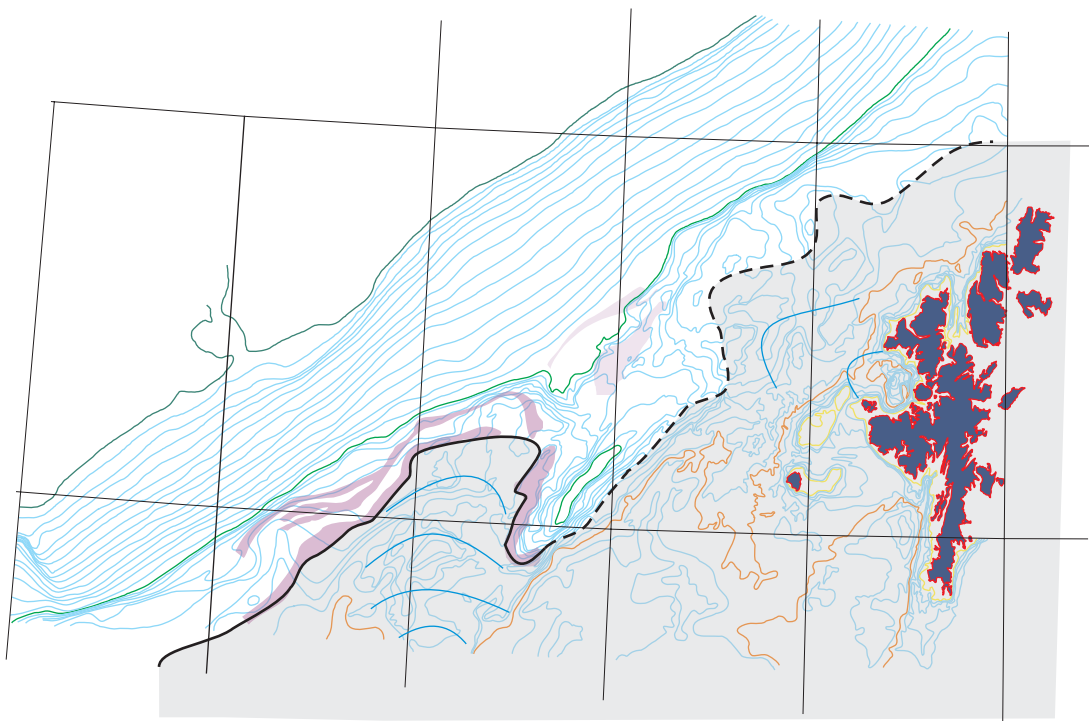


Figure 6.4C. Second significant pause in the ice retreat from the shelf edge. Formation of recessional moraines in the south occurs on the edges of deeper water areas ahead of areas of faster ice.

consistent with the stagnated ice interpretation of the infilled hollows described in Figure 4.28 which are contemporary with this phase of ice retreat.

The larger moraine banks numbered 3 to 6 in Figure 6.3, are the result of separate re-advance phases following a significant retreat to the inner or mid shelf. The length of time between each re-advance is unknown as there is no dating evidence, but the period between each phase was probably quite short. The north-eastern ends of banks 3 and 4 show a curving to the south-east, suggesting the re-advancing ice sheet formed an ice tongue in a similar position to the previous receding ice. The curving nature of the moraines indicates ice movement broadly from the south or south south-east. The data presented in section 4C indicates a zone of increased downcutting, partially infilled by Otter Bank sediments which is interpreted as a preferential ice pathway. The proposed direction of ice flow from the route emerges at the shelf edge up-slope of the large debris flow fields. Together, these separate pieces of evidence suggest an ice stream of sediment laden ice developed on the shelf throughout the Otter Bank age glaciations.

The only significant moraine bank in the Foula Bight area appears to be related to the first of the re-advance phases. The bank forms an almost continuous deposit from the north-eastern end of bank 5, around the western margin of the Papa Basin and across the Foula Bight to join with the small outer shelf moraines in the northern area. The section in the Foula Bight area is more dispersed and poorly defined, suggesting the a more buoyant ice body. The section of moraine bank along the edge of the Papa Basin, which appears as a north-eastward extension of bank 6, may represent deposits from ice deflected along the basin from the south-west, rather than crossing it from the inner shelf. The till deposits are not only very thin, as shown by data from borehole 84/02, but tend to follow the seabed isobaths between 150m and 180m. This suggests a grounding line between those depths and indicates floating ice over the deeper parts of the Papa Basin and probably the Foula Bight.

The pattern of moraines in the northern part of the shelf bears similarities to the large moraines in the south, even though they are on a different scale. The arcuate nature

of the banks indicates the maximum advance was centred along the cross-shelf channel which leads from St. Magnus Bay. This supports the idea of an ice stream or ice tongue originating as an onshore ice sheet on Shetland, which flowed out of St. Magnus Bay and followed the pathway across the shelf to the shelf edge. This also demonstrates that the ice on the northern part of the shelf was not directly related to that in the southern area, although the similarity of repeated re-advances suggest both systems were responding to the same overall controls. The sequence of events in terms of ice advance and retreat is summarised in figures 6.4 A – G. Between the retreat in figure C and the re-advance in figure D, a more significant retreat occurs, possibly back to the mid shelf. This coincides with the boundary between units OB3 and OB4 described in section 4.3.4. Other than the erosion of the OB1 and OB2 units recorded on the seismic survey data, there is no other evidence to define the extent of the retreat phase. The controls causing the various advances and retreats are uncertain. However, it seems likely that the overall retreat from the shelf edge maximum (depositing OB2, 6.4a) through the oscillating retreat to the inner-shelf (Figure 6.4G) is in response to a general deglaciation and is probably climate driven. The series of smaller retreats, pauses and re-advances which form the major moraine banks of the southern transect may be the result of more local effects related to ice front collapse and stabilisation or surging, especially in the region of the small intra-shelf basin (basin 1 in Figure 3.1) where water depth was locally greater.

6.4 Palaeo-ice streams on the Shetland margin

The reconstructions of the various phases in the growth and decay of the West Shetland ice sheet presented above are all influenced by the presence of preferential ice drainage pathways. These appear to be governed primarily by inner shelf topography, although the major ice route in the south may be controlled by influences outside the study area linked to the northward draining of the main BIS. Both the southern ice pathway and the smaller northern routes from St. Magnus Bay and through the Foula Channel show features associated with palaeo-ice streams, although they also lack some of the important common features. Of the main criteria outlined by Stokes and Clark (2001), for the identification of palaeo-ice streams, the West Shetland examples exhibit only two on the basis of current data: a width of

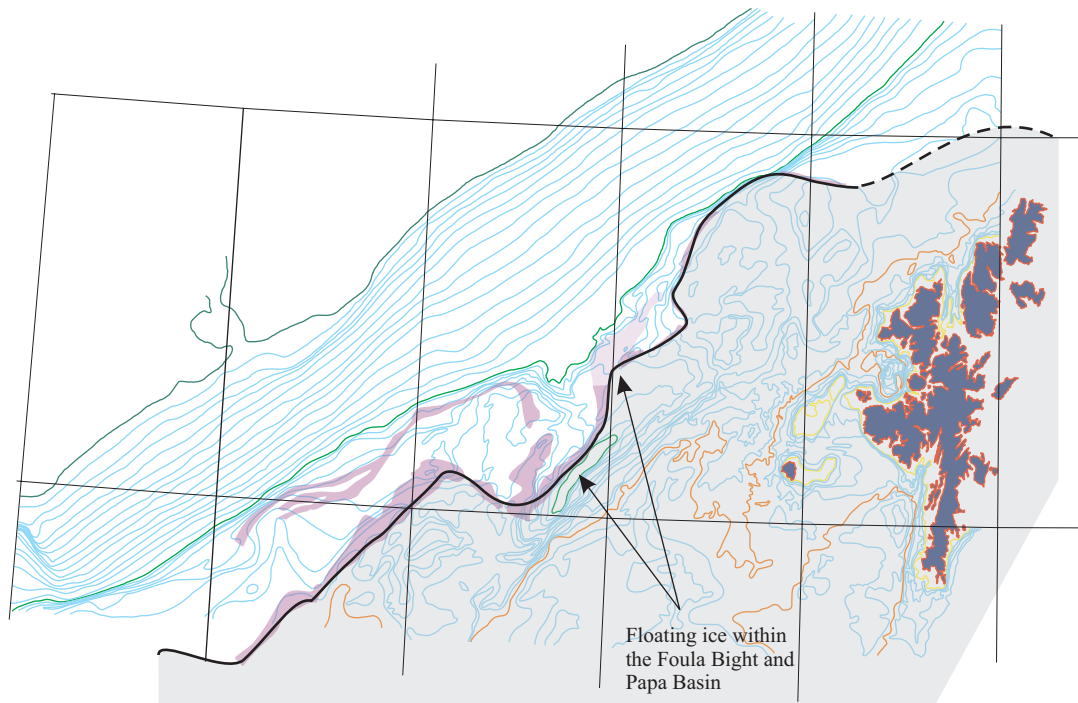


Figure 6.4D. First major re-advance phase. Ice grounds on outer shelf ridge in the south, but the thinner, more buoyant ice in the north reaches the shelf edge. Ice within the Foula Bight and Papa Basin is floating in the deeper sections.

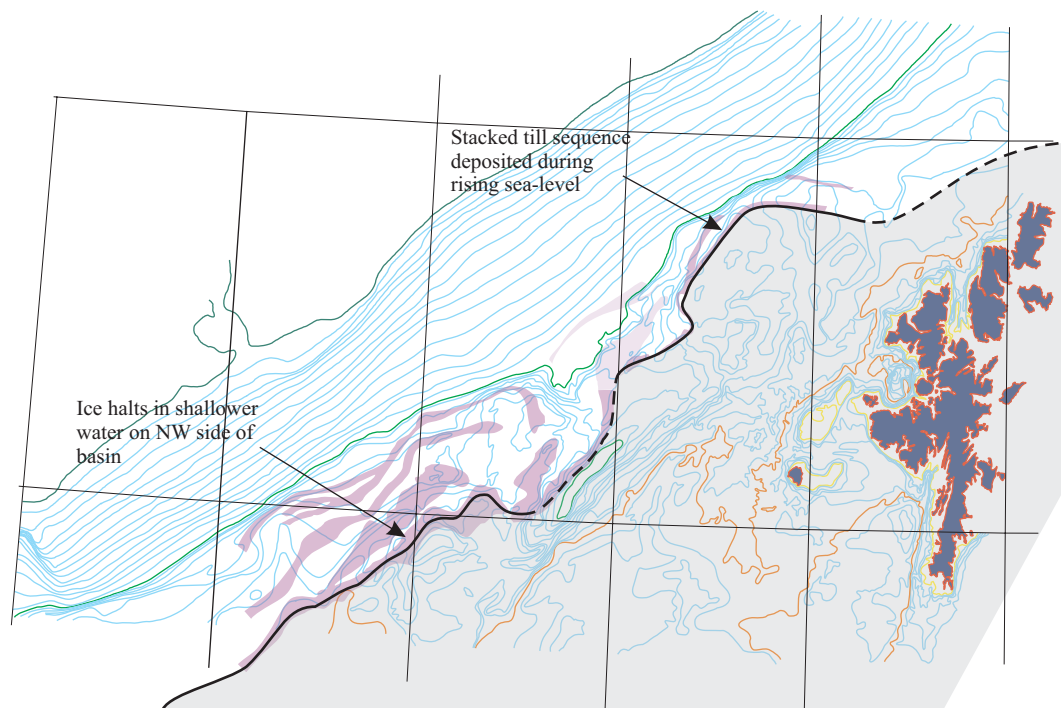


Figure 6.4E. Second phase re-advance following a retreat of uncertain extent. Ice in the south halts on the NW margin of the intra-shelf basin. In the north the ice decouples from the seabed, forming a new stacked till sheet. Rising sea-level prevents the ice from advancing further.

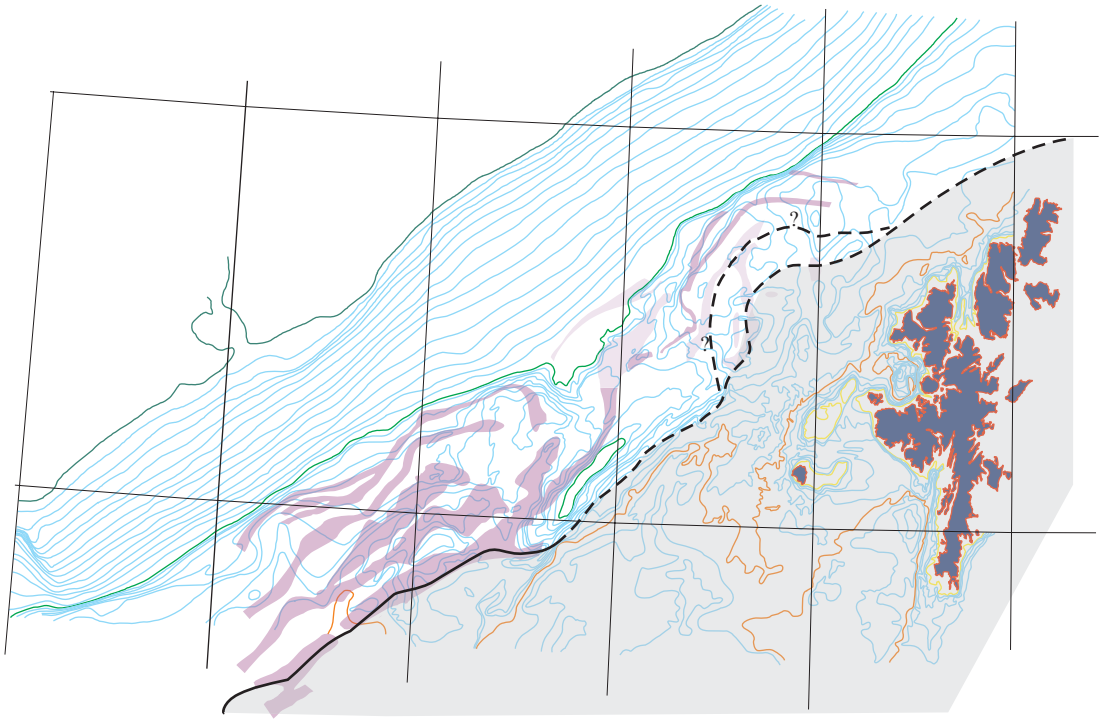


Figure 6.4F. Third phase of re-advance following a further retreat and rise in sea-level. Moraines are not deposited in water deeper than 140m, suggesting ice in water depths greater than this is floating. The position of the ice front in the north and in the Foulca Bight - Papa Basin is un-certain.

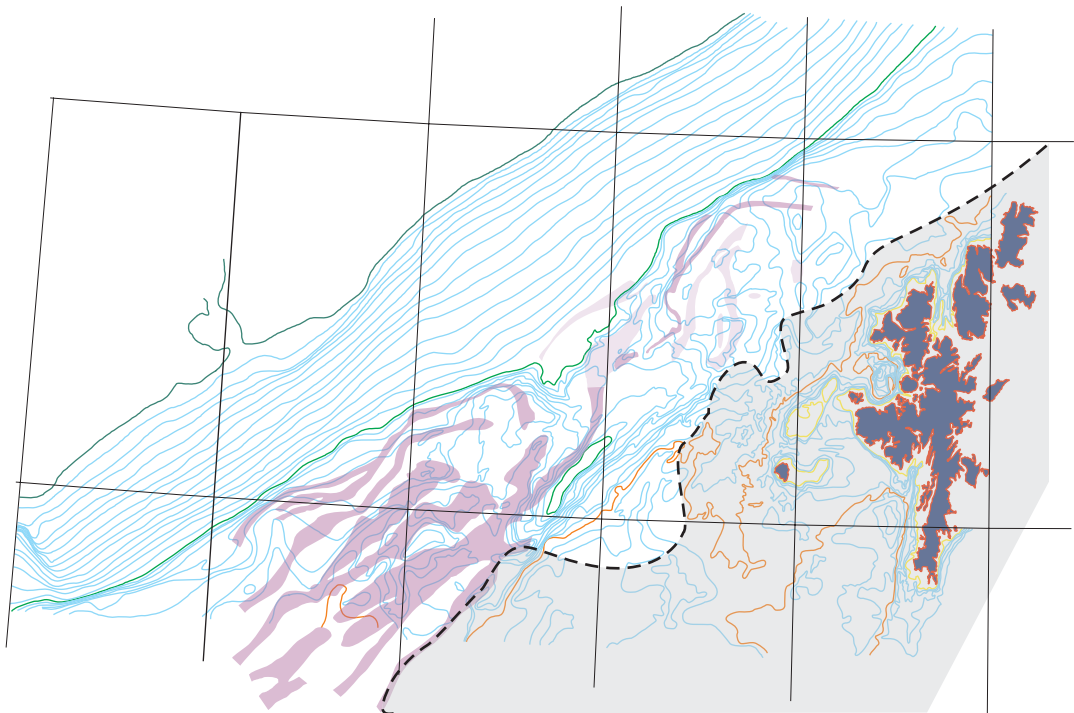


Figure 6.4G. Forth phase of re-advance with continuing sea-level rise. The ice has probably retreated onto the inner shelf except in the deeper channels where ice tongues may still exist. There is no evidence for moraines in the shallower areas on the inner shelf, but these may have been removed by subsequent erosion.

greater than 20 km and, more importantly, focused sediment delivery on the slope, which forms the Rona Wedge. In general terms the accumulation is not large, but can still be considered a trough mouth fan. The proposed Shetland ice streams lack the characteristic elongate bedforms and marginal moraines which have been reported from most documented examples, particularly from those around the Antarctic margin (Anderson, 1999). However, this may be due to data limitations rather than the absence of features. The southern ice route arguably has a deformable bed of glaciogenic sediments with early evidence of limited deformation. The thicker region of sediments would certainly have aided ice movement. However, more importantly from the West Shetland perspective is the presence of curved terminal and recessional moraine banks from different glacial phases (Figure 6.3). This shows some parts of the ice margin advanced further than others, and that these zones retreated later than the others. The same zones also lie along the pathways through the deeper intra-shelf basins which, as described earlier, would provide more buoyancy to the advancing ice front and aid forward movement. The thicker sediments which accumulated in the basins would also aid the movement of the ice.

The northern ice route from St. Magnus Bay and the Foula Channel (Figure 6.4a) shows similar, but smaller scale features. Restricted, overdeepened basins with thicker sediments and arc-shaped moraines all indicate that a similar set of controls were acting upon the northerly ice sheet to produce a faster moving ice-stream along the line of the deeper channel. There is also evidence of convergence of deeper channels along the proposed ice route, which is another characteristic feature identified by Stokes and Clark (2001).

The absence of a definitive trough structure on the outermost shelf of the southern ice stream is slightly problematic, but Ó Cofaigh *et al* (2003), suggest that trough structures with or without a trough mouth fan can indicate former ice streams., indicating that not all components are necessary.

The proposed West Shetland ice streams are difficult to compare with those from other margins due to the relatively small scale of the Shetland examples. This may be

a factor in the absence of some of the common features such as the lineated bedforms and striations reported from most examples (eg; Andrews *et al*, 1985. Hodgson, 1994. King *et al*, 1996). Related to this is the thickness of the West Shetland ice sheet which was probably no more than 250m thick at the shelf edge and was semi buoyant or close to floating in some areas. This is in contrast to the larger scale examples which are hundreds of metres thick and completely grounded, although the West Shetland ice sheet may have been thicker towards the inner shelf. The debris flow accumulation forming the Rona wedge on the southern part of the West Shetland slope shows a surface morphology very similar to that at the mouth of the Marguerite Trough on the western shelf of the Antarctic Peninsula (Ó Cofaigh *et al*, 2003). The slope in the region of the Marguerite Trough shows a series of gullies and debris flows, with a more extensively gullied slope to the sides of the ice stream outflow. The slope beyond the mouth of the trough is smoother and has a lower gradient. On the West Shetland slope the gradient is lower in the area of highest sediment accumulation, downslope of the proposed ice stream, which is consistent with the Marguerite Trough. However, this area is also more severely gullied than the areas to the sides. The explanation for the contrast in morphology is partly due to the much steeper Antarctic slope angles, about 14° as opposed to $1^{\circ} - 2^{\circ}$ on the Shetland slope. The difference in slope angle probably prevented large accumulations of stable sediment forming on the Antarctic slope, and increased deposition on the basin floor which is reported by Pudsey (2000) and Ó Cofaigh *et al* (2001). On the Shetland slope the sediment accumulated on the upper slope as debris flows from the shelf edge, forming a relatively stable wedge in the north-eastern part of the Rona Wedge, but running out to the basin floor in the south-west where there was a slightly steeper slope. The area of the Foula Bight represents the gullied, sediment-starved edge of the ice stream observed on the Antarctic example.

6.5 Timing of ice sheet advance-retreat cycles

The timing of the Otter Bank glacial cycles on the West Shetland Margin are difficult to define due to the absence of accurately dated samples. The dating of the Otter Bank sediments as a whole, and the individual glacial phases within it are critical in reconstructing the British Ice Sheet at the last glacial maximum (LGM). Although an

absolute chronology is not possible due to the lack of dated material, a detailed relative stratigraphy of ice advance and retreat events can be assembled. The outer shelf moraine complex of area C (sections 4.6 to 4.10) together with the ice-marginal features at the shelf edge allow the construction of a relative time-distance curve for the Otter Bank glaciations. This curve is presented in Figure 6.5. The similar pattern of the moraines and shelf edge deposits in the northern ice field west of St. Magnus Bay suggest this curve is mapping regional effects rather than local differences in the ice sheet behaviour.

Detailed investigations on sediments from the Barra Fan on the Hebrides Margin by Wilson et al (2002), indicate that the Late Devensian ice sheet reached the shelf edge after 30Ka b.p., reaching a glacial maximum during Marine Isotope Stage (MIS) 2. The maximum indicated by the cold water foraminifer *Neogloboquadrina pachyderma* (sinistral) lies between 21 Ka and 17 Ka b.p. The data also shows a large input of IRD at approximately 16 Ka b.p., interpreted as equivalent to Heinrich event 1, and that deglaciation of the shelf was complete by 15Ka b.p. Whilst this locality lies 300 km south of the Shetland margin, it does provide the most detailed data for the advance and retreat of the north-western margin of the British Ice Sheet. Isotope data from a core in the Faroe-Shetland Channel approximately 200 km northwest of the West Shetland shelf indicates the LGM lies between 19 and 16 Ka b.p. (Rasmussen et al, 1996). These two data sets help to constrain the possible ages for the Otter Bank sequence.

A comparison of the of Otter Bank glacial phases with data from the Norwegian margin shows similarities in the pattern of advances and retreats. This similarity is useful in constraining the timing of the Otter Bank phases as the Norwegian data has much better chronological control. A summary of Norwegian data is presented in Dahlgren (2002), together with new data from the mid Norwegian shelf.

The data from the mid Norwegian shelf (Dahlgren, 2002) shows the closest match in terms of the pattern of Late Devensian (Late Weichselian) ice sheet movements. The Norwegian shelf shows a glacial maximum between 19 Ka and 21.5 Ka. A higher

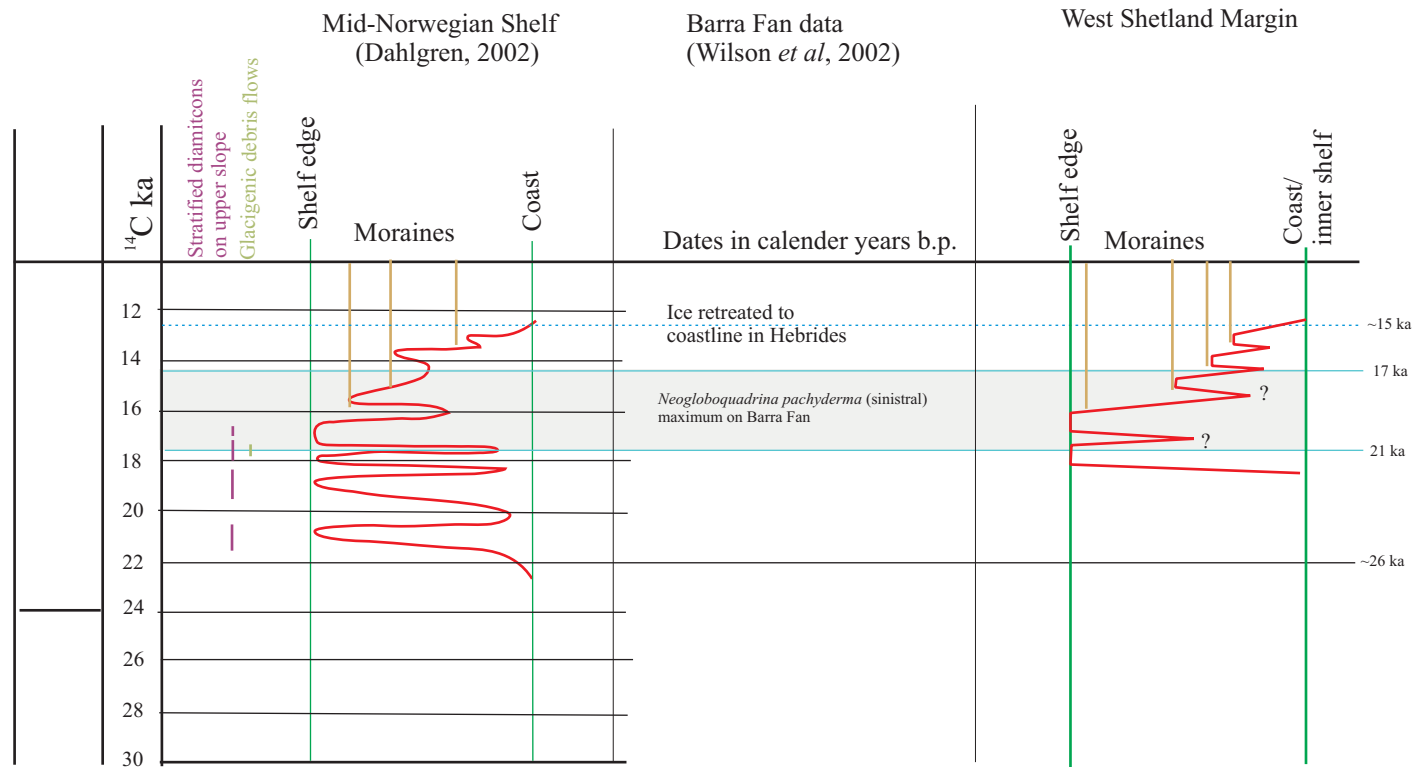
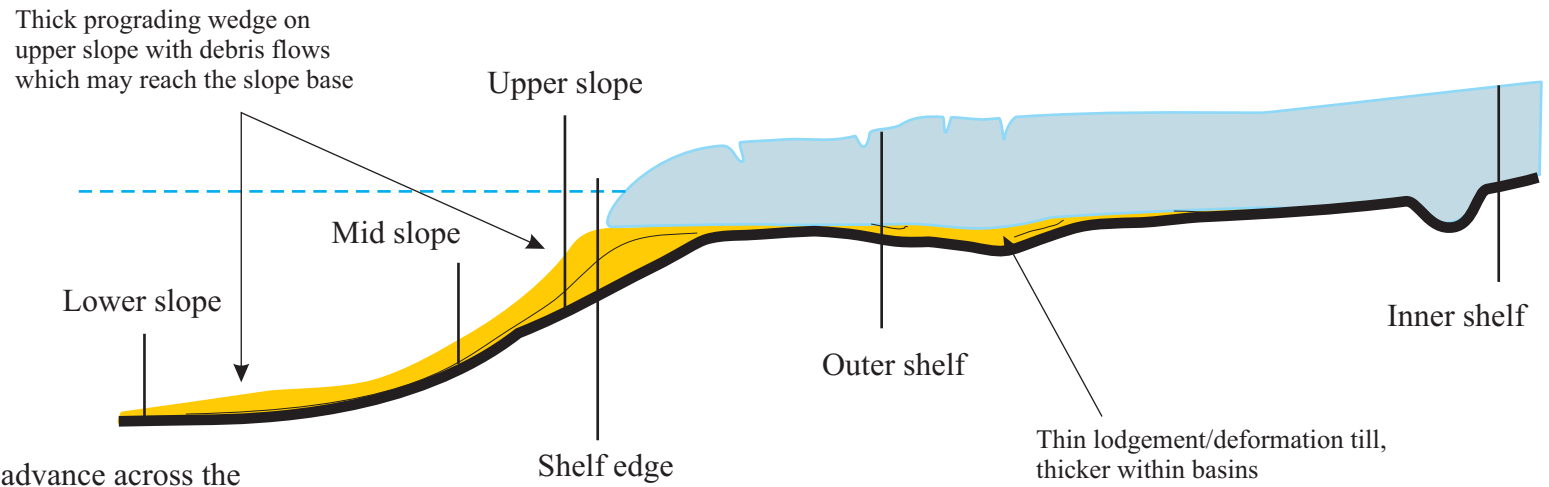


Figure 6.5. Time distance data for the glaciation of the mid-Norwegian shelf showing the relative position of the ice front, together with data from the Barra Fan and relative distance of the ice front west of Shetland. The Shetland data has no time scale, but the glacial maximum and final retreat are constrained by the dates from the Barra Fan. The Barra Fan dates could be slightly later than those for the Shetland Margin as it lies 300km to the south.

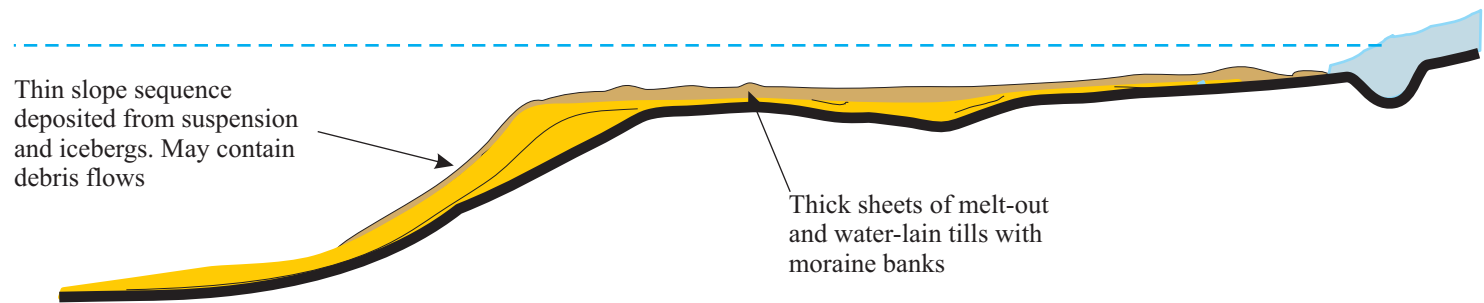
resolution glaciation curve shows two or three advances to the shelf edge, followed by a long retreat phase which was punctuated by three re-advances. The last of the shelf edge phases is dated at *c.* 19.5 Ka, which is in the middle of the range indicated by Wilson *et al* (2002), for the glacial maximum on the Barra Fan. This suggests that the age of the last ice advance to the edge of the West Shetland shelf was probably also within the Barra Fan age range. Following retreat from the shelf edge, the mid Norwegian ice shows three re-advance phases before 15 Ka and prior to retreat to the coastline. Such behaviour correlates very closely with the later phases of Otter Bank ice sheet which shows three re-advance phases on the mid to outer shelf, as shown by the moraine complex of area C (Section 4.9), prior to final retreat. The correlation with the pattern of Late Weichselian events on the Norwegian shelf, and the constraints on ice sheet timing provided by the Barra Fan data, indicate a Late Devensian age for the Otter Bank sequence glaciations. This has important implications for constraining the limits of the British Ice Sheet at the LGM and during its deglaciation.

6.6 Late Devensian limits for the British Ice Sheet

The limits of the British Ice Sheet at the LGM in the northwest of Britain are contentious and have been the subject of recent debate. The limits suggested by Stoker *et al* (1993) for the West Shetland area, described in chapter 3, are broadly confirmed by this study, showing that the Late Devensian ice sheet was certainly present on the shelf and advanced to the shelf edge. Late Devensian and LGM ice limits proposed by Bowen *et al* (2002) do not stand up to scrutiny, completely ignoring any of the data from the West Shetland margin and placing LGM and earlier Devensian ice limits around the north coast of Scotland and north Orkney. The reason for this omission is unknown, and the failings of the model discussed by Hall *et al* (2003). An earlier model of Late Devensian ice cover and sea levels by Lambeck (1995) is also shown to be incorrect for the Shetland area, which is shown as being ice-free with an exposed shelf during the last glaciation.



a) First cycle ice advance across the shelf forming a glacial advance sequence



b) Retreat phase of the first glacial cycle

Figure 6.6. Large scale sedimentary architecture of a single glacial cycle. The positions indicated along the profile are referred to in the text and summarised sequence logs presented in Figure 6.7a-x.

6.7 Summary model based upon the West Shetland Margin

Based upon the data from the West Shetland margin, a two stage conceptual model of ice sheet advance and retreat has been constructed and is presented in this section. The first part of the model shows the large-scale sedimentary architecture from a single glacial cycle, comprising a glacial advance from the inner shelf to the shelf edge, followed by a retreat back to the inner shelf (Figure 6.6 a and b). In addition, a series of sedimentary logs from different positions across the margin, show the theoretical sequence of sediments deposited during a single cycle and how these sequences differ from those which are likely to be preserved in the geological record (Figure 6.7a-x). This comparison highlights the lack of completeness in the preserved record and the bias produced by differences in preservation potential.

The second part of the model shows the effect on the large-scale architecture of a second glacial cycle within the same glaciation and how it affects the overall preserved glacial record (Figure 6.8a and b). Sedimentary logs are not constructed for the second cycle as there are too many variables to construct a meaningful predictive sequence.

The model relates to the southern area of the margin which has a slight landward slope. The reason for the slope is unclear, but there are two main possibilities: a slope related to major tectonic units on the margin, which is unrelated to glacial activity, or distortion of the shelf by ice loading to the southeast. The second of these theories would require a much thicker ice sheet than that proposed for the Shetland margin, but such an ice sheet is much more likely in the area to the south, draining the mainland of Scotland. The thickness of the Late Devensian ice sheet draining northern Scotland is not clear, but a thickness of approximately 600m has been indicated by Ballantyne et al (1998). Walcott (1970) proposed a downwarping influence up to 180km away from an ice sheet 1700m thick.

First glacial cycle

The first stage of the model (Figure 6.6a) shows an ice sheet advancing to the shelf edge. The nature and thickness of deposits from this initial stage are partly dependent

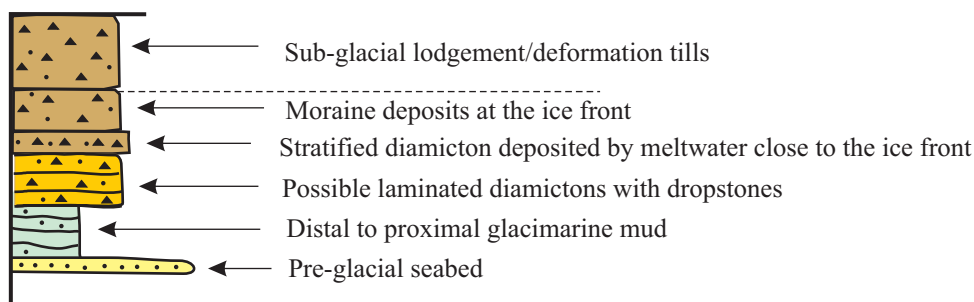


Figure 6.7a. Deposited glacial advance sequence from the inner shelf.

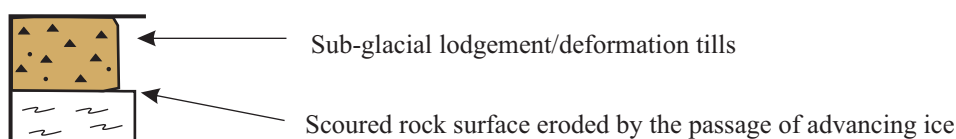


Figure 6.7b. Preserved sequence from a glacial advance on the inner shelf.

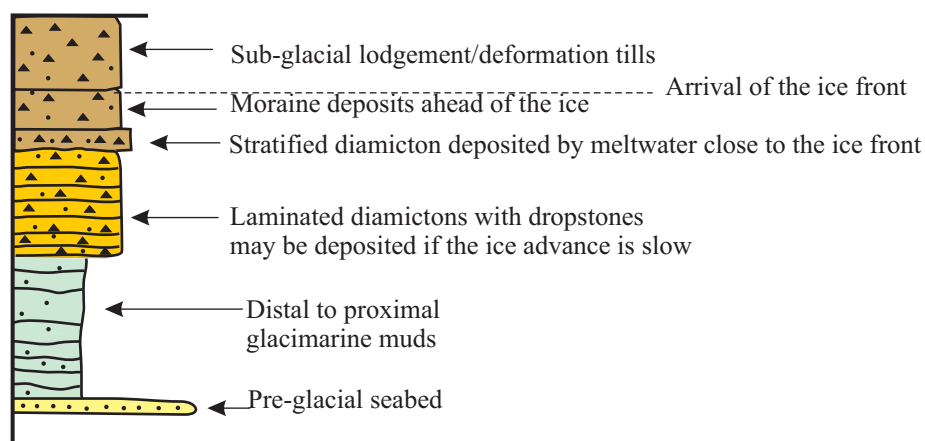


Figure 6.7c. Deposited glacial advance sequence from the outer shelf.

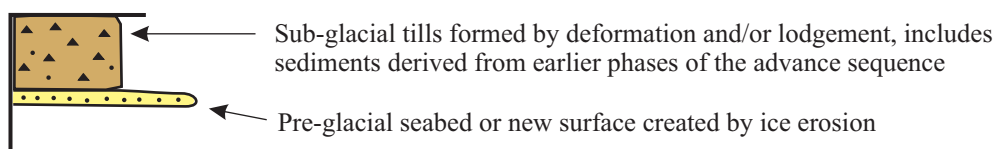


Figure 6.7d. Preserved glacial advance sequence from the outer shelf

upon the nature of the substrate, softer sedimentary rocks eroding more easily and supplying more sediment than high grade metamorphics. As the ice expands onto the inner shelf, the depositional facies belts migrate ahead of the ice front. The deposited vertical sequence (Figure 6.7a) reflects the passage of the different facies belts, with a transition from ice-distal to ice-proximal and ice-contact deposits. As the ice approaches, the volume and grain size of sediment delivered by meltwater plumes increases, as does the IRD delivered from icebergs. If the rate of ice advance is relatively slow, deposition of laminated diamictons with dropstones may become established, progressing to the deposition of less structured stratified diamictons of the ice-proximal to ice-contact zone. As the ice front arrives, deposition of structureless diamictons which form push moraines at the ice front occurs, completing the pro-glacial depositional sequence (Figure 6.7a). As the ice advance continues, it overrides the earlier advance sequence sediments. This has the effect of removing most, or all of the pro-glacial sequence which is incorporated into sub-glacial debris layer and may be re-deposited as lodgement till, or as a semi-mobile deforming layer. Consequently, although there is significant deposition of pro-glacial sediments on the inner shelf during the advance stage, the preserved sequence is limited to thin, compact sub-glacial tills (Figure 6.7b), although ultimately these deposits may not survive de-glaciation.

Advance of the ice across the shelf deposits a similar sequence to that formed on the inner shelf, although as the ice progresses, the accumulation of debris both in front of, and beneath the ice increases. Figure 6.6a shows an intra-shelf basin of the style seen on the West Shetland shelf. The effect of the ice advancing into slightly deeper water is to increase the effective buoyancy of the ice and relieve some of the pressure on the base. Not only does the deeper water provide more accommodation space for deposition of the pro-glacial deposits, it also aids the preservation of the pro-glacial sediments, although it is in a modified form. The preservation leads to a thicker sub-glacial till derived from the pro-glacial sediments.

On the outer shelf, beyond the edge of the intra-shelf basin, the water is shallower and consequently the ice loading is increased. The deposited sequence (Figure 6.7c)

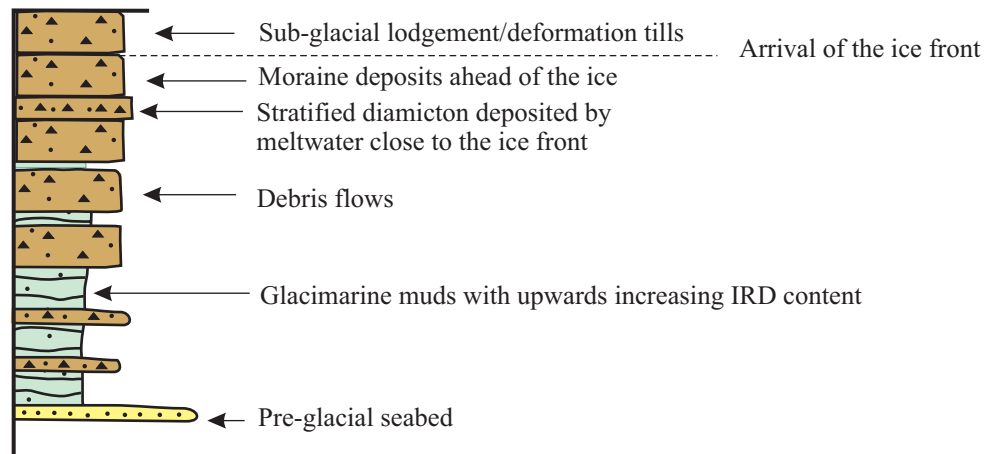


Figure 6.7e. Deposited advance sequence from the shelf edge.

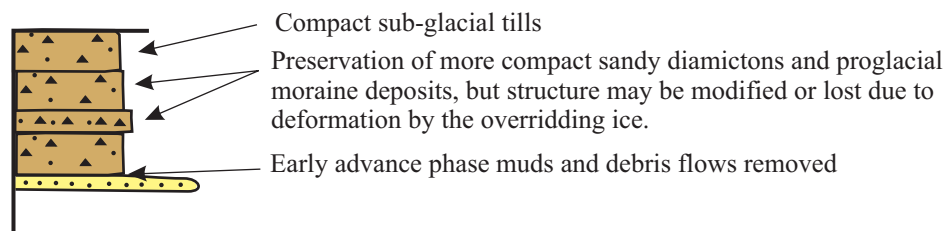


Figure 6.7f. Preserved advance sequence from the shelf edge.

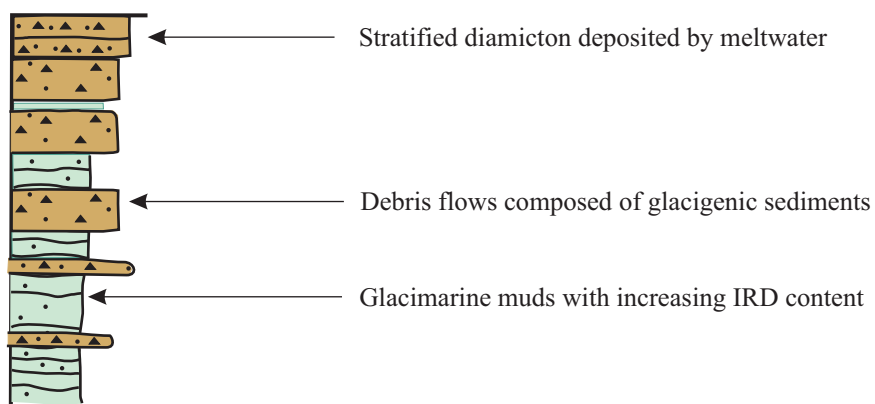


Figure 6.7g. Deposited advance sequence from the upper slope

is similar to, but thicker than that on the inner shelf because of the greater volume of pro-glacial and sub-glacial debris. The preserved sequence is mainly confined to lodgement and deformation tills, although remnants of the pre-glacial seabed may survive (Figure 6.7d). In the Shetland example this could be due to poor core recovery. However, a more likely explanation for the predicted preserved sequence is that much of the pro-glacial advance sequence is removed by the advancing ice and reworked into subglacial tills. On shelves with rapid subsidence or a steeper slope angle the preserved sequence is likely to be more complete.

At the shelf edge (Figure 6.6a), the style of glacial deposition and preservation change dramatically. The base of the sequence (Figure 6.7e) is formed by distal glacimarine muds deposited on the upper slope from suspension. As the ice approaches, the grain size and volume of the suspended sediment increases, together with the IRD content. As the ice approaches, the increasing volume and rate of sedimentation leads to debris flow generation which increase upwards in frequency and thickness (Figure 6.7e). As the ice front arrives at the shelf edge, stratified diamictos deposited from meltwater discharge, followed by massive diamictos which form end moraines are deposited as ice-contact sediments. These sediments form part of the prograding wedge which may periodically re-mobilise to generate debris flows. With continued advance, the sequence is overridden, compacting and deforming the upper units of the advance sequence. However, the passage of the ice at the shelf edge does not remove all of the previous advance sequence as in other shelf zones. Instead the sequence becomes incorporated into the prograding shelf edge, leaving a preserved sequence which retains the upper units (Figure 6.7f).

The lower part of the advance sequence on the upper slope (Figure 6.6a) is broadly similar to that on the outer shelf. The base of the sequence comprises glacimarine muds with an upwards-increasing IRD content (Figure 6.7g). With the increasing delivery of suspended sediment and IRD from the advancing ice, debris flows become more frequent and larger. When the ice reaches the shelf edge, coarser sandy sediments from meltwater discharge are delivered directly to the upper slope. This may also be combined with debris flows derived from moraines which are physically

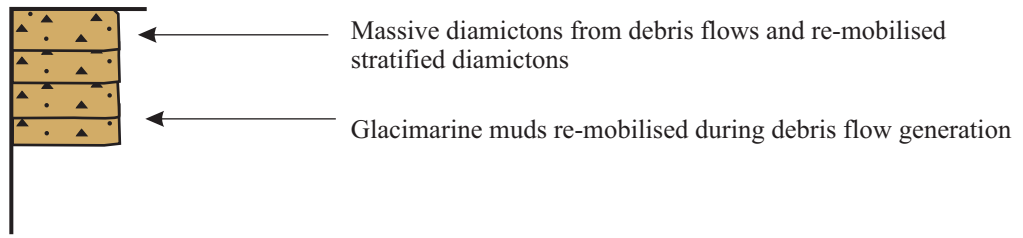


Figure 6.7h. Preserved advance sequence from the upper slope.

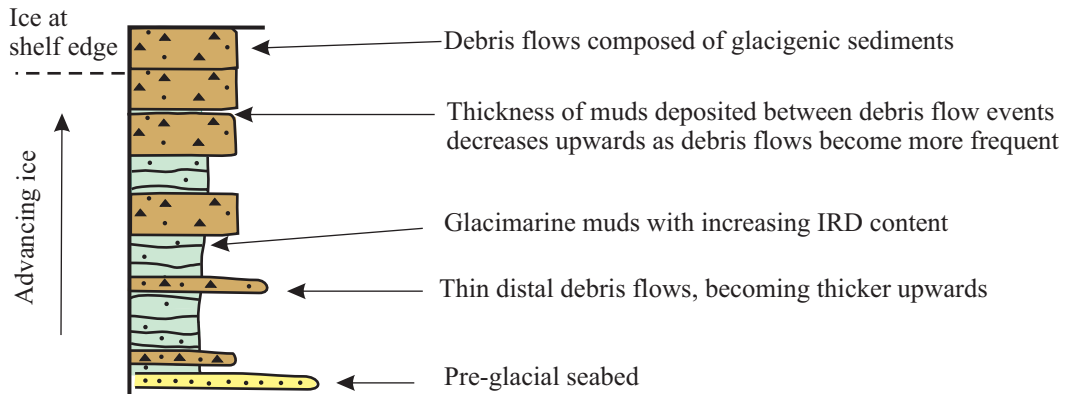


Figure 6.7i. Deposited advance sequence from the mid-slope

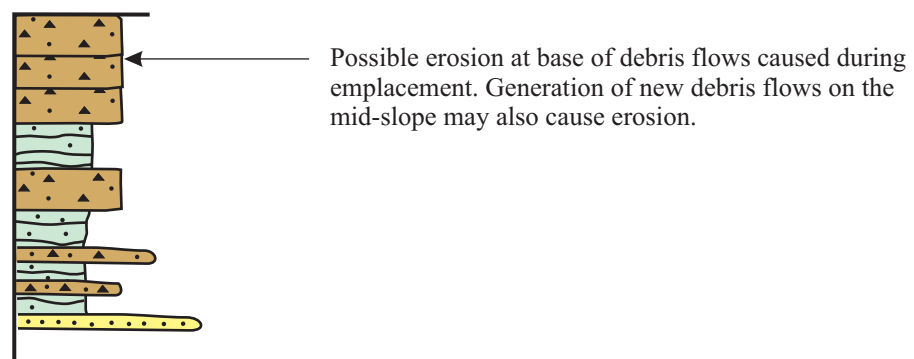


Figure 6.7j. Preserved advance sequence from the mid slope.

pushed over the shelf edge by the advancing ice. The generation of debris flows on the upper slope causes the loss of the lower part of the advance sequence, with only the later diamictos from the larger debris flows being preserved (Figure 6.7h). As a result, the preserved sequence is likely to be much thinner than that actually deposited.

On the mid slope (Figure 6.6a), the glacial advance sequence comprises the same coarsening upwards units of glacial marine muds and minor debris flows seen on the upper slope (Figure 6.7i). In addition to the input from glacial marine sources, along-slope contour currents may cause winnowing of the mud units and form thin sheet-like silty lag deposits. In addition to removing sediment from the slope, exotic material may also be delivered from other areas along the slope. As the ice reaches the shelf edge, sediment delivery to the slope is increased, resulting in thicker and more frequent debris flows. The rapid accumulation of sediment on the mid-slope also leads to the generation of debris flows, especially when sedimentation rates are at a maximum (Figure 6.7i).

The preserved sequence from the mid-slope is broadly similar to that deposited, except in the upper sections where debris flow emplacement and generation may reduce the sequence thickness (Figure 6.7j).

The glacial advance sequence at the base of the slope (Figure 6.6a) is very similar to that on the mid-slope, comprising glacial marine muds and debris flows (Figure 6.7k). The debris flows vary from very thin distal muddy deposits to thick diamictos forming large lobes. The main difference between the mid slope and the slope base is the degree of erosion between major debris flows. Thin mud units still separate the larger flows, suggesting there is little erosion during the emplacement of these deposits (Figure 6.7l). The muds deposited near the base of the sequence are prone to re-working by along-slope contour currents which are more common in the deep water and tend to follow the slope break. However, as the glacial advance progresses and the sedimentation rate increases, the influence of these currents declines. The apparent decline may be due to the swamping of the system by the increased

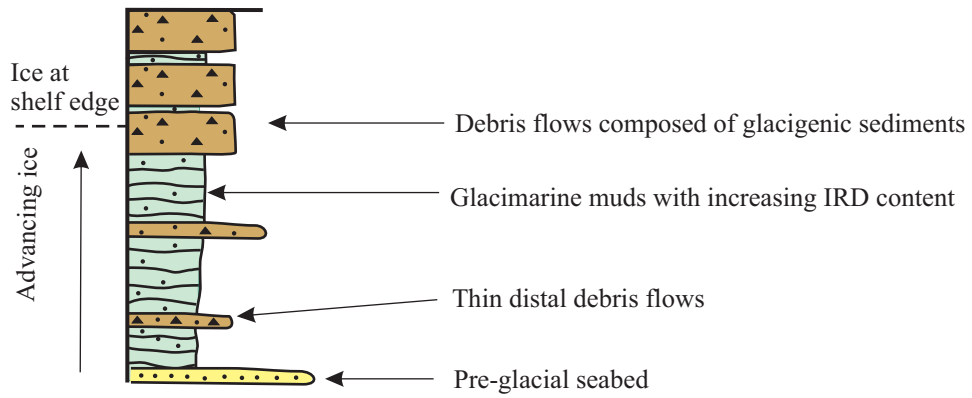


Figure 6.7k. Deposited advance sequence from the base of the slope

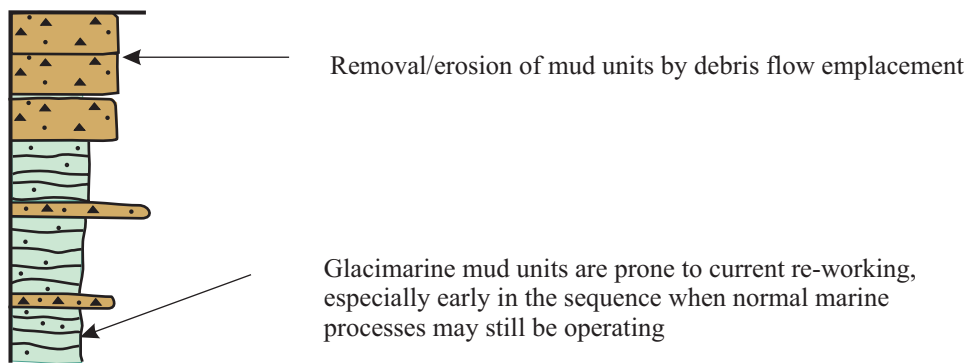


Figure 6.7l. Preserved advance sequence from the base of the slope

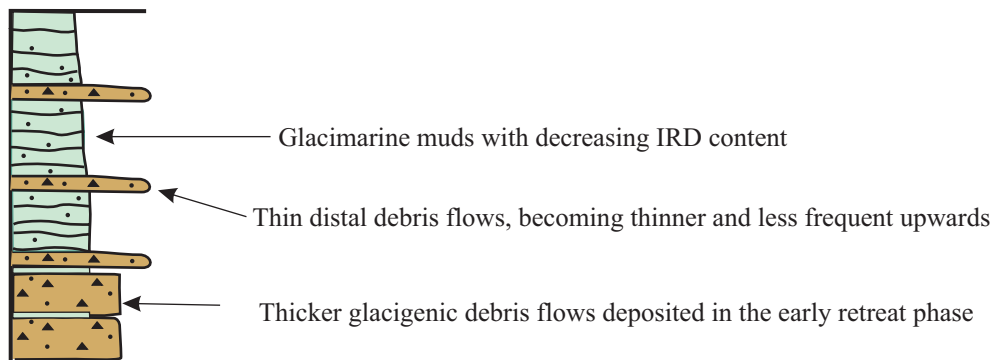


Figure 6.7m. Deposited glacial retreat sequence from the base of the slope

sediment load, or in the case of the West Shetland setting, may be due to a change in the current regime flowing through the Faroe-Shetland Channel.

The glacial retreat phase of the first cycle (Figure 6.6b), is broadly a reverse of the advance. However, the overall depositional architecture is very different. On the advance phase of the cycle, the depositional sequence on the shelf was shown to be generally thin, with a much thicker and better preserved sequence on the slope. In the retreat phase, the slope becomes sediment starved, the main depocentres being on the shelf.

The beginning of the retreat phase is marked by a decay in the ice, increasing the volume of meltwater and sediment discharged at the ice front and an increase in iceberg calving. This leads to an influx of sediment delivered directly to the upper slope and the water column. The sequence at the base of the slope (Figure 6.7m) reflects the increase in sediment delivery, with the deposition of larger debris flows and thicker glacimarine muds. However, the mud and debris flow deposition decline relatively quickly to leave a fining upwards mud with decreasing IRD content. The preserved sequence is little different from the deposited sequence. There is some erosion of mud units beneath the larger debris flows, but the main loss is from the top of the mud sequence (Figure 6.7n). On the West Shetland margin, re-working by sea-floor currents, which are probably along-slope rather than downslope, becomes an important process at a late stage of ice retreat or following deglaciation. Gravity-driven, downslope currents are more likely to deposit sediment at the base of the slope rather than erode it. The effect of the along-slope currents is to erode an unknown depth of the glacial sequence, leaving a sandy lag deposit. The current re-working process in the late or post-glacial stages has the potential to completely erode the entire glacially influenced sequence, especially in the areas where the sequence lacks debris flows on the floor of the Faroe-Shetland Channel. It is also unlikely to be unique to the West Shetland Margin.

The retreat sequence on the mid-slope (Figure 6.6b) comprises a series of debris flows of decreasing frequency and thickness. Interbedded sandy muds become finer

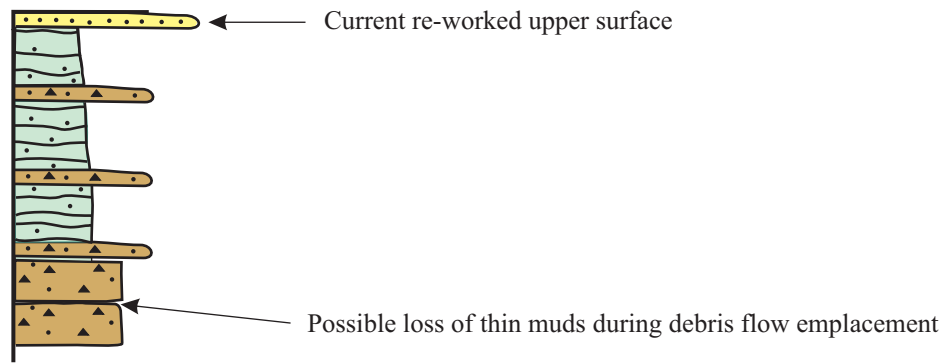


Figure 6.7n. Preserved glacial retreat sequence from the base of slope

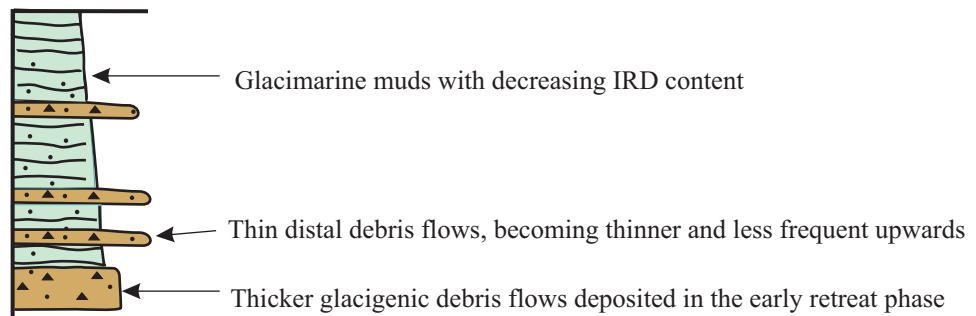


Figure 6.7o. Deposited glacial retreat sequence on the mid-slope.

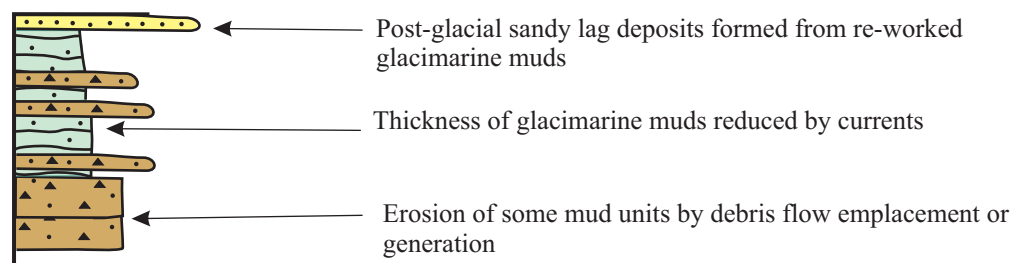


Figure 6.7p. Preserved glacial retreat sequence from the mid-slope

upwards as the influence of the ice decreases and becomes more distal (Figure 6.7o). The lower part of the sequence is reduced by debris flow generation and emplacement, removing or eroding the mud units. In the later stages of de-glaciation, the zone is the main site for contour current re-working and may significantly reduce the sequence thickness (Figure 6.7p). However, whilst some of the sequences may be reduced, the sediment eroded from them can contribute to significant accumulations in other locations.

The upper slope retreat sequence (Figure 6.6b) initially reflects the flushing stage of ice sheet decay following the maximum advance. Debris flows and stratified diamictos from meltwater discharge are deposited in the early stages, followed by a gradual change to glacial marine muds as the ice becomes more distal (Figure 6.7q). However, the rapid deposition at the beginning of the retreat, and the later loading by muds, can lead to debris flows much later in the cycle. The preserved sequence (Figure 6.7r) is likely to reflect the reduction of the sequence due to debris flow generation. In addition, the structure of the sequence, especially the lower units, are prone to re-working by iceberg turbation which becomes important on the upper slope during the ice sheet decay. Re-working of the upper section by normal marine processes is indicated by sandy lag deposits at the seabed. This may represent the removal of a substantial volume of mud in which the sand originally formed a dispersed component; a style of deposition which would be expected in a more ice-distal setting during a glacial retreat.

The retreat sequence deposited at the shelf edge (Figure 6.6b) comprises a basal sub-glacial diamicton formed by lodgement deposition and/or deformation of existing sediments. As the ice begins to decay, the ice front recedes from the shelf edge, depositing thick structureless, melt-out and meltwater-derived diamictos (Figure 6.7s). If ice retreat pauses, these deposits will accumulate to form shelf-edge moraines. As the ice front continues to recede, the deposition of structureless diamictos is replaced by more stratified diamictos with a lower mud content, the mud being transported in suspension to a more distal setting. Deposition of IRD and turbation by icebergs is particularly important during the deposition of the lower

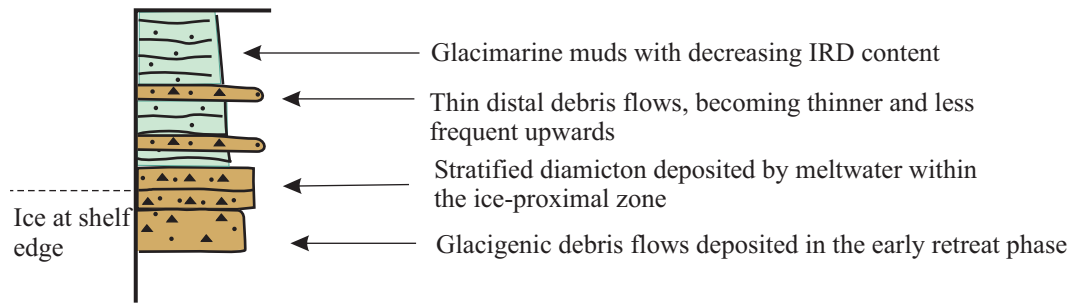


Figure 6.7q. Deposited glacial retreat sequence from the upper slope

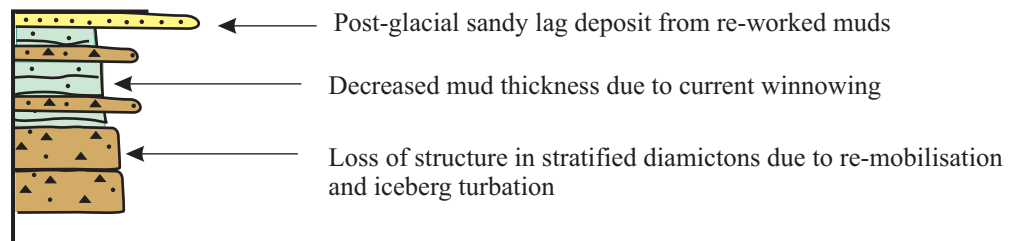


Figure 6.7r. Preserved glacial retreat sequence from the upper slope

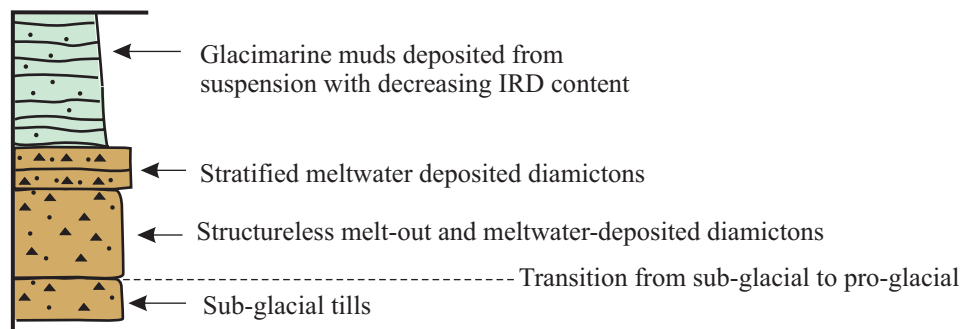


Figure 6.7s. Deposited glacial retreat sequence from the shelf edge

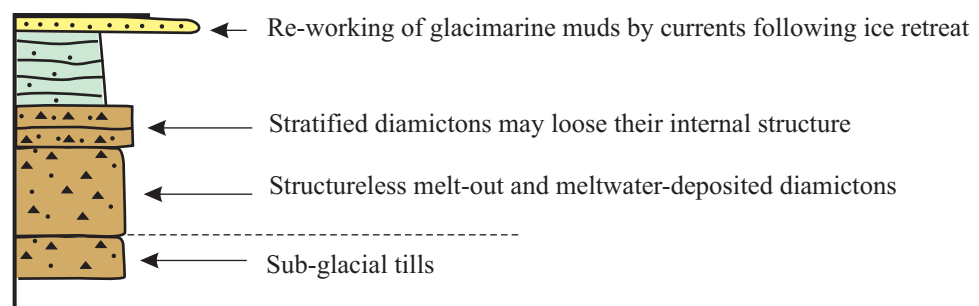


Figure 6.7t. Preserved glacial retreat sequence from the shelf edge

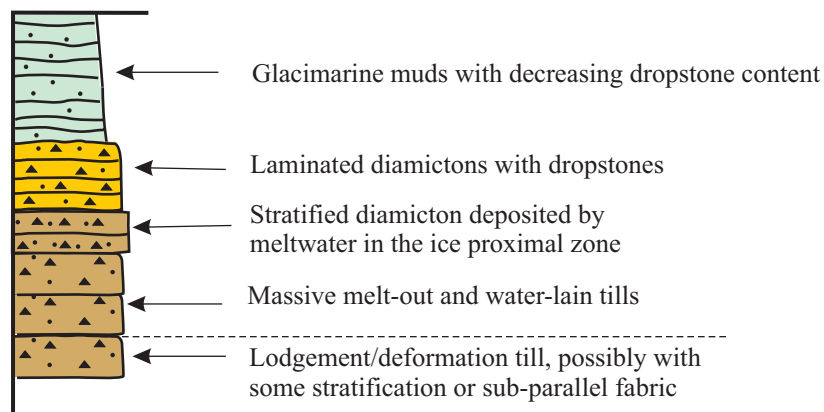


Figure 6.7u. Deposited glacial retreat sequence on the outer shelf

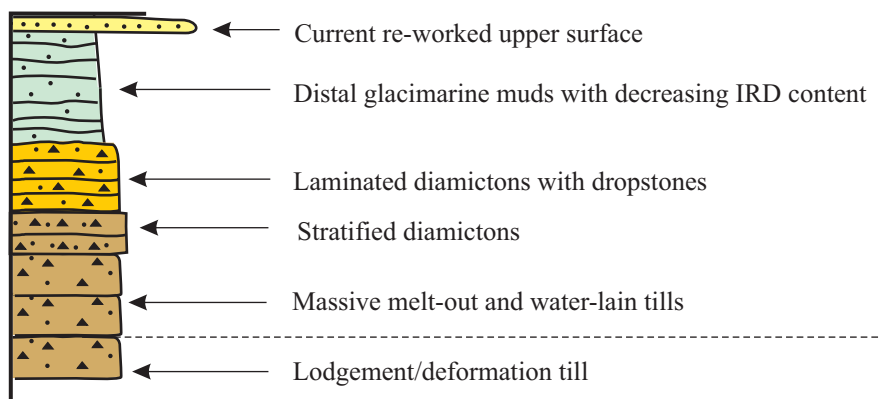


Figure 6.7v. Preserved glacial retreat sequence from the outer shelf

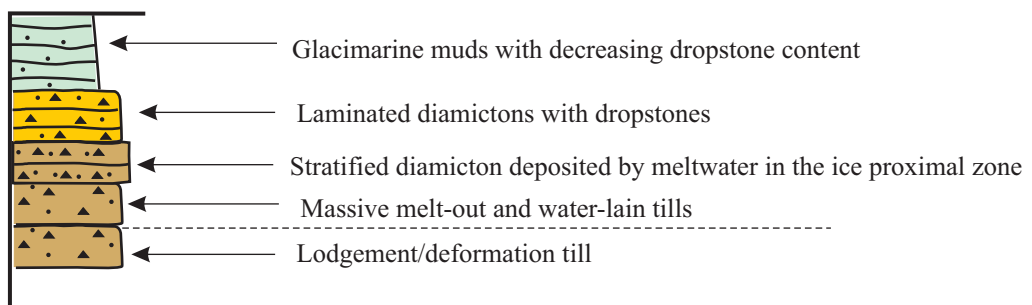


Figure 6.7w. Inner shelf glacial retreat sequence.

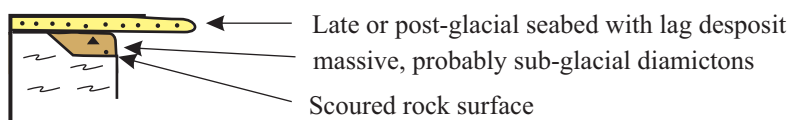


Figure 6.7x. Preserved inner shelf glacial retreat sequence

parts of the retreat sequence as ice sheet decay becomes established. The sequence is capped by proximal to distal glacimarine muds with an IRD content which decreases upwards.

The sequence preserved at the shelf edge is similar to the theoretical depositional sequence (Figure 6.7t). The main changes are the probable loss of structure within the stratified diamictos due to iceberg turbation, and the loss of glacimarine muds from the top of the sequence by non-glacial marine processes.

The glacial retreat sequence from the outer shelf (Figure 6.6b) is composed of thick diamicton and mud deposits. Chronologically, the base of the retreat sequence is represented by lodgement and deformation tills, which continue to be deposited until the area is no longer covered by ice (Figure 6.7u). As the ice front recedes beyond the outer shelf, thick, structureless melt-out and water-lain tills are deposited as ice contact deposits. With continued decay, deposition becomes progressively more ice-distal, leading to a decrease in sediments delivered by meltwater currents and in suspension. IRD and iceberg turbation also decrease upwards. This pattern of deposition is similar for the whole of the shelf zone.

The preserved sequence from the outer shelf (Figure 6.7v) differs very little from the theoretical deposited sequence. The retreat processes do not tend to modify earlier deposits, with the exception of meltwater currents which may cause winnowing of sediments in the ice-proximal zone. Re-establishment of marine processes following ice retreat are also likely to re-work the units higher in the sequence, but these effects may be local due to differences in water depth.

The inner shelf retreat sequence (Figure 6.6b) is similar to that on the outer shelf, but like the advance on the inner shelf, the thickness of individual units and the overall sequence thickness is smaller (Figure 6.7w). The reason for the retreat sequence being thinner is the relatively small reservoir of debris in the remaining ice at this stage of the retreat phase, much of the debris having been removed by meltwater flushing earlier in the retreat. The flushing may also remove some of the basal

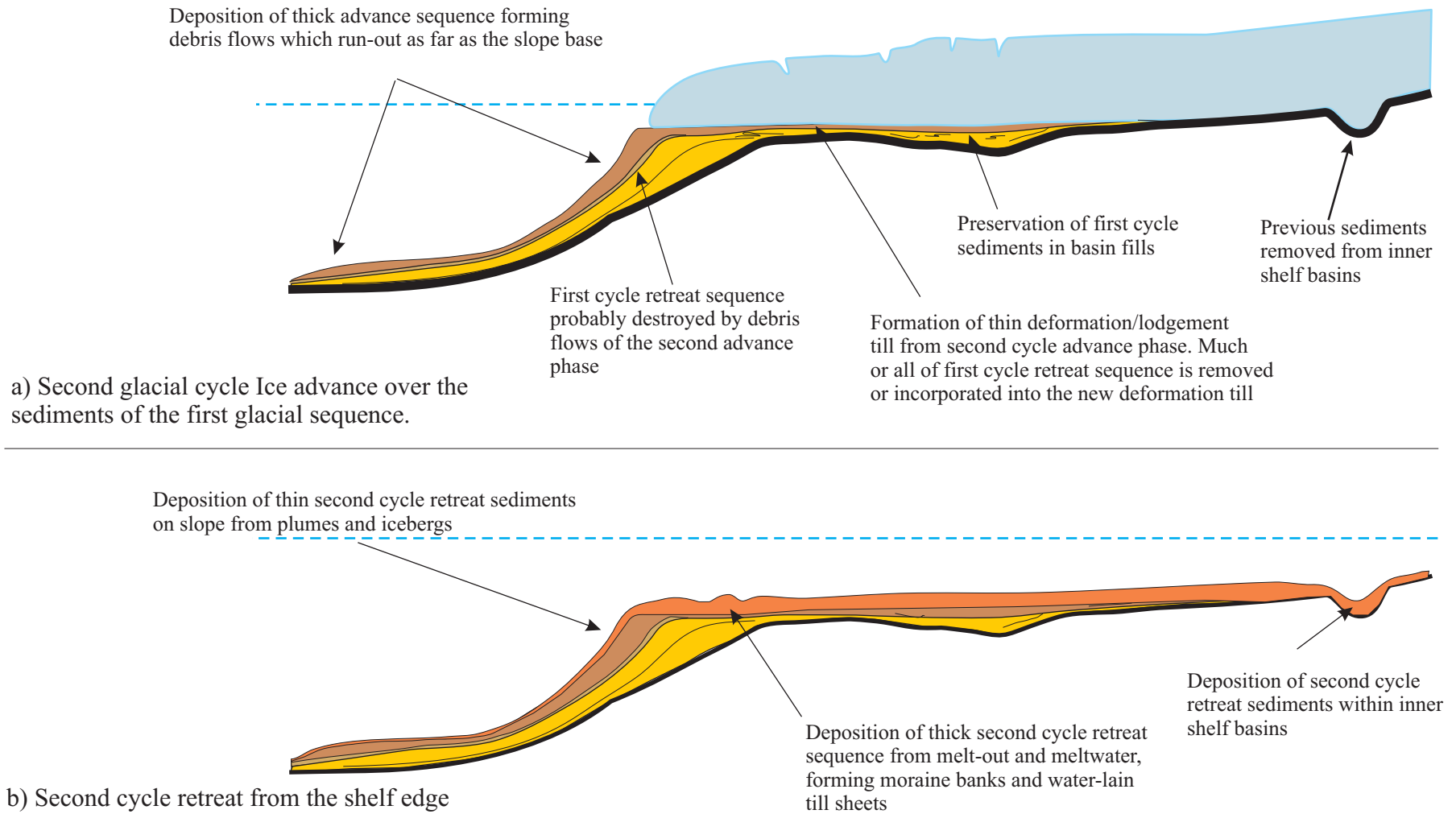


Figure 6.8. The effect on large-scale sedimentary architecture of a second glacial cycle

lodgement tills whilst the inner shelf is still ice covered.

The preserved retreat sequence is very thin or even absent in some places (Figure 6.7x). Where sediments are preserved, they are limited to thin, massive diamictons which infill topographic lows. Scoured bedrock surfaces covered with thin sandy lag deposits derived from re-working of the glacial sediments may be the only remains of the retreat sequence. This is due to re-working of the glacial sediments by a combination of tidal, wave and wind-induced currents, together with more regional oceanic currents which affect the shelf. Modern wave-induced currents have been observed reaching 1 m/s on the outer shelf to as much as 4 m/s on the inner shelf (Pantin, 1991).

The thin preserved diamictons are likely to be lodgement tills as these are more resistant to erosion due to their greater compaction. They are also at the base of the sequence and consequently, are the last to be re-worked.

Second glacial cycle

Phase 2 of the model (Figure 6.8a and b) shows the effects on large scale architecture of a second glacial advance-retreat cycle. On the inner shelf, the new ice advance removes most or all of the first cycle sediments, incorporating them into the new deformation and lodgement tills of the new glacial advance sequence. Any of the first cycle sequence which does survive is likely to be a sub-glacial deposit which has already been compacted by the first phase of glaciation.

Accumulation of sediment on the shelf during the second advance phase is restricted by accommodation space, which requires subsidence or sea-level rise, as any space has already been filled by sediments of the first cycle. A minor rise in sea-level may reduce basal pressure on the ice without causing the ice to float. This is especially important in the shelf basins where increased buoyancy already exists, thus aiding the preservation of first cycle sediments and the development of new advance sequence tills.

At the shelf edge and on the slope, deposition is greater than in the first cycle. This is the result of removal of the first cycle shelf sequence by the advancing ice, delivering much of the sediment to the margin. Deposition on the lower slope is a combination of longer run-out debris flows and glacimarine sediments.

The final stage in Figure 6.8b shows the second phase of ice retreat. The second cycle retreat sequence is very similar to that of the first cycle in that it is formed of thick till sheets and moraine banks on the shelf and thin glacimarine deposits on the slope. Debris flows on the slope are more likely than in the first cycle as the slope has been steepened and has a greater volume of sediment in place.

7. Conclusions

The conclusions of this study can be split into two broad themes, those relating specifically to the West Shetland Margin, and those which relate to glacial marine processes and architecture generally.

The Late Devensian glacial history of the West Shetland Margin as represented by the Otter Bank Sequence shows that ice sheets crossed the shelf to the shelf edge on two occasions, followed by a re-advance which mostly terminated on the mid to outer shelf. All the phases terminated in a marine environment and there is no evidence to suggest that the shelf was ever exposed. In addition, there is no evidence to suggest that the ice sheet ever crossed the shelf break onto the upper slope.

The source of the ice on the West Shetland Shelf was from two separate origins. A small ice sheet derived from the Shetland Islands formed a minor ice stream which travelled northwest to the shelf edge. This was joined by ice travelling northwest from the Foula Channel and northeast along the Papa Basin. The major input of ice to the southern part of the margin was from the south, travelling roughly northwards at about 4°W and forming a larger ice stream approximately 30 - 40 km wide. This southern ice stream formed the main route for all three major advances.

Ice sheet thickness is uncertain, but assuming that the water depth at the shelf edge was no deeper than the present day, the ice margin features in the Foula Bight indicate the ice in that region was no thicker than 200m. Ice thickness in the major ice stream in the south is likely to have been greater, but not exceeding 250m. It is also unlikely that a thicker ice sheet could have travelled unconstrained across the shelf and maintained its thickness over a distance of 100km +.

The most complete sedimentary history of the different phases of glaciation is found on the uppermost slope and at the slope base, although all the different provinces preserve some evidence of activity. Deeper basins on the shelf preserve a more complete record than other shelf areas.

In general, the preserved sequences on the slope relate to glacial advance phases, whilst the sequences on the shelf relate to retreat phases.

The West Shetland data shows that water depth is directly, and indirectly a strong control, not only on the behaviour of ice sheets, but also on the depositional processes and the preservation of the resulting sediments. The water depth directly affects the buoyancy of an ice sheet, which in turn determines the pressure exerted at the base of the ice and the degree of erosion or deposition beneath it. Where the ice thickness is much greater than the water depth erosion appears to be the dominant process, which is particularly true on the inner shelf. However, where channels and basins exist beneath the ice, the local buoyancy of ice is effectively increased, reducing erosion of the underlying rock, possibly to the extent of allowing deposition to occur. The increased buoyancy also increases the chances of pre-existing rocks or sediments being preserved.

The nature of the seabed lithology exerts a control on the roughness of the seabed, and appears to have some control over the thickness of sediments deposited on the shelf. Areas underlain by metamorphic basement rocks tend to exhibit a rougher surface morphology and have a thinner preserved glacial sequence than those underlain by softer rocks. In addition, in areas of high-grade metamorphic basement on the shelf exposure, only the most recent glacial deposit is likely to be preserved. However, these effects may also be indirect results of the water depth both during deposition and following deglaciation. Areas of softer rocks on the inner shelf tend to preferentially develop deeper channels, probably as a result of glacial erosion. In turn, the deeper channels lead to a greater effective water depth and so develop a thicker sequence of glacial sediments.

The correlation with the pattern of Late Weichselian events on the Norwegian shelf, and the constraints on ice sheet timing provided by the Barra Fan data, strongly suggest that the second Otter Bank glacial cycle, and the ice oscillations on the shelf, are of Late Devensian age. The first glacial cycle recorded in the Otter Bank sequence may also be of Late Devensian age, but this is less certain. However, as a

matter of urgency, reliable dating of the Otter Bank sequence is needed in order to confirm the correlations and attempt any meaningful reconstruction which includes onshore deposits. Material in the existing cores may be too degraded for certain dating techniques, but suitable macrofossils occur in some cores, especially in 82/11, which would constrain the age of the basal Otter Bank unconformity.

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Appendix 1