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LATE QUATERNARY BENTHONIC FORAMINIFERAL STRATIGRAPHY
OF THE WESTERN U.K. CONTINENTAL SHELF

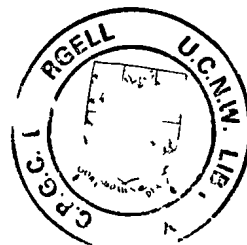
by

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Thesis submitted in accordance with the
requirements of the University of Wales
for the degree of Doctor of Philosophy.

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I do not know what I appear to the
world, but to myself I seem to have been
only like a boy playing on the sea-shore,
and diverting myself in now and then finding
a smoother pebble or a prettier shell than
ordinary, whilst the great ocean of truth
lay all undiscovered before me.

L.T. More, *Isaac Newton* (1934), p.664

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3. Percentage frequency diagram of benthonic foraminifera from VE 57/-09/46.
4. 95% confidence limits, VE 57/-09/46.
5. Percentage frequency key to benthonic summary diagrams within text.

ABSTRACT

Late Quaternary deposits have been investigated from three main study areas from western Britain in an attempt to define lithological and biostratigraphic changes. Detailed analyses of included benthonic foraminiferal assemblages are presented and a systematic section included which describes and illustrates over 200 distinct forms. Chronostratigraphic control is provided by radiocarbon dates, amino acid analysis and tephrachronology.

The three study areas yield distinctive records of the depositional environments characterizing the climatic events of the Late Quaternary. From the Hebridean Shelf, B.G.S. vibrocores have been analysed within the context of a previously established seismostratigraphic sequence. Foraminiferal faunas allow the reconstruction of a regional climatostratigraphic sequence for the Lateglacial period (c. 14,000 to 10,000 BP) and this sequence is correlated, through 9 radiocarbon (AMS) dates, to the established climatostratigraphy of the Lateglacial period from N.W. Europe. Reconstructions of notional water depths during this period allow glacio-isostatic components from the shelf to be estimated and these confirm a generally accepted pattern of changing relative sea-level, from initial regression following deglaciation and subsequent transgression as the eustatic component over-takes the isostatic component. Rising sea-levels are most notable after about 10,000 BP.

A cliff section at Aberdaron on the western Llyn Peninsula provides an insight into the controversy surrounding the question of depositional origin of the "Irish Sea Drift" sequences bordering the Irish Sea. Diamicts and sorted layers from the section contain mixed boreo-arctic, temperate and pre-Quaternary species, and allochthonous/autochthonous elements are identified. While lithological changes within the section are marked, the foraminiferal assemblages maintain relatively constant faunal ratios. None of the foraminifera are considered to be *in situ*, but instead entrained by the Irish Sea glacier during its passage along the Basin and deposited at the site by basal melt-out

processes.

The third study area, the southwestern Celtic Sea, records geomorphological evidence of previously extensive glaciation in the region. Microfaunas, both foraminifera and Ostracoda, are analysed and record a transition from grounded ice lodgement facies to quiet, glacial marine facies at about $49^{\circ}30'N$. Amino acid analysis confirms the geomorphological evidence for glacial marine accumulation during the Late Devensian.

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DAVID JAMES SEXTON

(1965 - 1989)

Two voices are there; one is of the sea,
One of the mountains; each a mighty Voice,
In both from age to age thou didst rejoice,
They were thy chosen music, Liberty!

Two voices are there (W.Wordsworth, 1807)

Chapter 1. Introduction

1.1. Aims of the study

The overall aims of this study can be divided into five main topics; these are:

(1.1.a) to establish a working taxonomic knowledge of the foraminiferal faunas of glacial marine and associated sediments from the western U.K. shelf seas. To briefly describe and fully illustrate this fauna in a series of scanning electron microscope (S.E.M.) photographs.

(1.1.b) to reconstruct palaeoenvironmental changes during the Lateglacial period on the basis of included foraminiferal assemblages in sediments from the Hebridean Shelf, N. W. Scotland.

(1.1.c) to define and date a proposed Late Devensian grounding line from offshore S. W. Britain on the basis of included microfossils (Foraminifera and Ostracoda) and other evidence.

(1.1.d) to determine whether onshore diamicts, exposed in a cliff section at Aberdaron, N. W. Wales, are glacial marine or terrestrial in depositional origin based upon their included foraminiferal and lithofacies characteristics.

(1.1.e) to attempt a preliminary scheme of glacial marine facies characterization based upon the included foraminiferal faunas of the sediments investigated.

Thus, in an attempt to address these aims, a number of sediment samples and their included micropalaeontology have been analysed. Geographically, the sites investigated fall into three distinct areas; these are:

The Hebridean Shelf, N. W. Scotland
Offshore Scillies, S. W. United Kingdom Shelf
Onshore Aberdaron, N. W. Wales

These three areas, together with the extent of United Kingdom designated waters are shown in fig.1.1.

Before each area is discussed, it seems appropriate to introduce western shelf seas in general and then the stratigraphic framework of N. W. Europe, particularly the detailed stratigraphy of the Lateglacial period, its climatic history and some of the theories and models which are currently thought to govern climate change. I will then introduce the reader to glacial marine sediments and environments, before finally discussing some aspects of palaeoecology.

1.2 Western Shelf Seas

1.2.1 Bathymetry

The continuous transition from land to abyssal ocean basin via continental shelf, slope and rise provinces that typify much of the North Atlantic Ocean is absent from the topographically complex margin of the western British Isles and north west Europe in general. A number of anomalously shallow plateaux, eg. the Rockall Plateau, separated by deep troughs and bounded by steep slopes exist in this area.

The two deep troughs that lie between Britain and (i) Rockall, (ii) the Faeroe Isles are known as the Rockall Trough and Faeroe-Shetland Channel, and form a nearly continuous feature, broken only by the Wyville-Thomson Ridge at 66°N. They play a major role in the deep water circulation of the area (see section 1.2.2) and are shown, together with all the major bathymetric features, in fig.1.1.

Slope morphology to the west of the British Isles is closely controlled by the volume of sediment transported outwards across the shelf and a general relationship exists where broader shelves have typically narrower slopes. Slope canyons are not widespread and their limited occurrence appears to bear a relationship to both the topography of the shelf and shelf transport paths (Keyon and Stride, 1970). Canyons are in fact believed to represent sites of multiple slope failure, producing a pattern of second and third-order gullies. They are absent from the slopes west of the

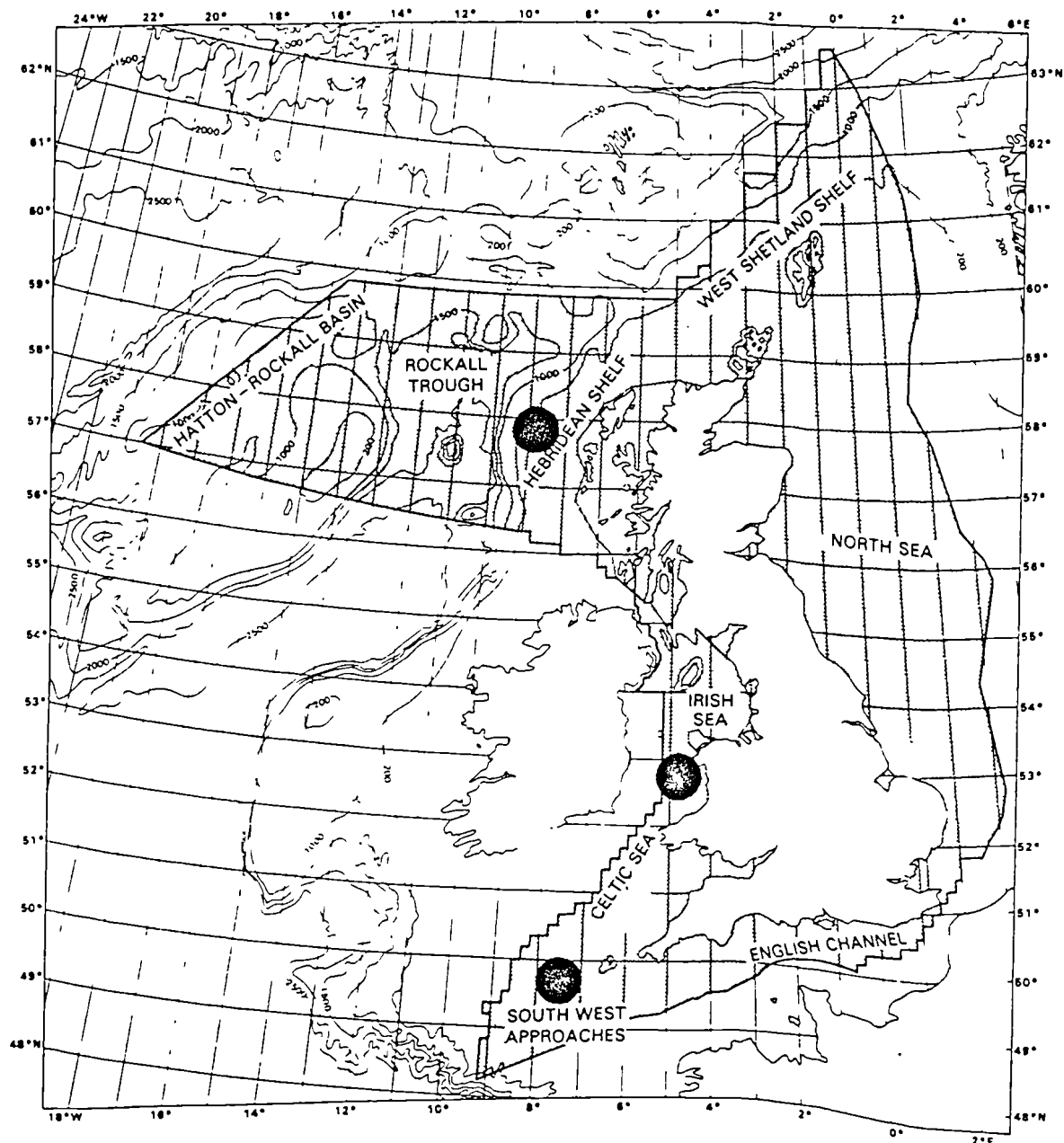


Fig.1.1 Map showing United Kingdom designated waters (stippled), the three main study areas (black spots), and other sites from where foraminifera have been illustrated (blue spots). Bathymetry is shown in metres (modified from Fannin, 1989).

Outer Hebrides and this can be attributed to sediment entrapment within the sea of the Hebrides and in the St. Kilda Basin. Canyons which have developed off northwest Ireland, on the other hand, occur where the shelf sediment transport paths lie perpendicular to the shelf (Roberts *et al.*, 1979). Slope failures, affecting as much as 95% of the slope area in the Bay of Biscay, but less than 20% of the slope west of Scotland are reported by Kenyon (1987). There does appear to be a greater likelihood of slope failure where the gradient is steeper, seaward of channels that cross the adjacent shelf, and where strong contour currents do not sweep the upper slope, which would otherwise reduce sedimentation rates in this critical area (Kenyon, 1987). However, little is known of the age or processes that generate these phenomena.

1.2.2 Oceanic circulation patterns

Near-surface circulation is summarized in fig.1.2, which illustrates the general northerly flow past western Britain. However, the underlying feature of this whole western area is the Rockall Channel, which is bounded to the east by the continental slopes of Scotland and Ireland, to the west by Rockall Bank, extends south-westward into deeper water, and is bounded to the north by the Wyville-Thomson Ridge. Much of the water entering the channel comes from the south west where it is at its deepest; the floor is at a water depth of 3,500 m. at 53°N.

Two major water masses occupy the channel, the upper extends from the surface down to between 1,200 m. to 1,500 m. and is derived basically from the classical Atlantic Central Water of the western Atlantic; beneath are waters derived mostly from the Labrador Sea. These are not the only water masses present, others which are apparent include north west Atlantic oceanic polar front water, Mediterranean water and water from the deep overflows of the Norwegian Sea (Ellett *et al.*, 1986).

The terminology of McCartney and Talley (1982) is useful. Subpolar Mode Water (SPMW) can be designated as the cyclonic gyre of the upper waters within the Atlantic north of 40°N "for which the temperature and salinity characteristics embrace a progressive cooling and freshening of the original North Atlantic Central Water as it crosses from the western Atlantic, enters the European

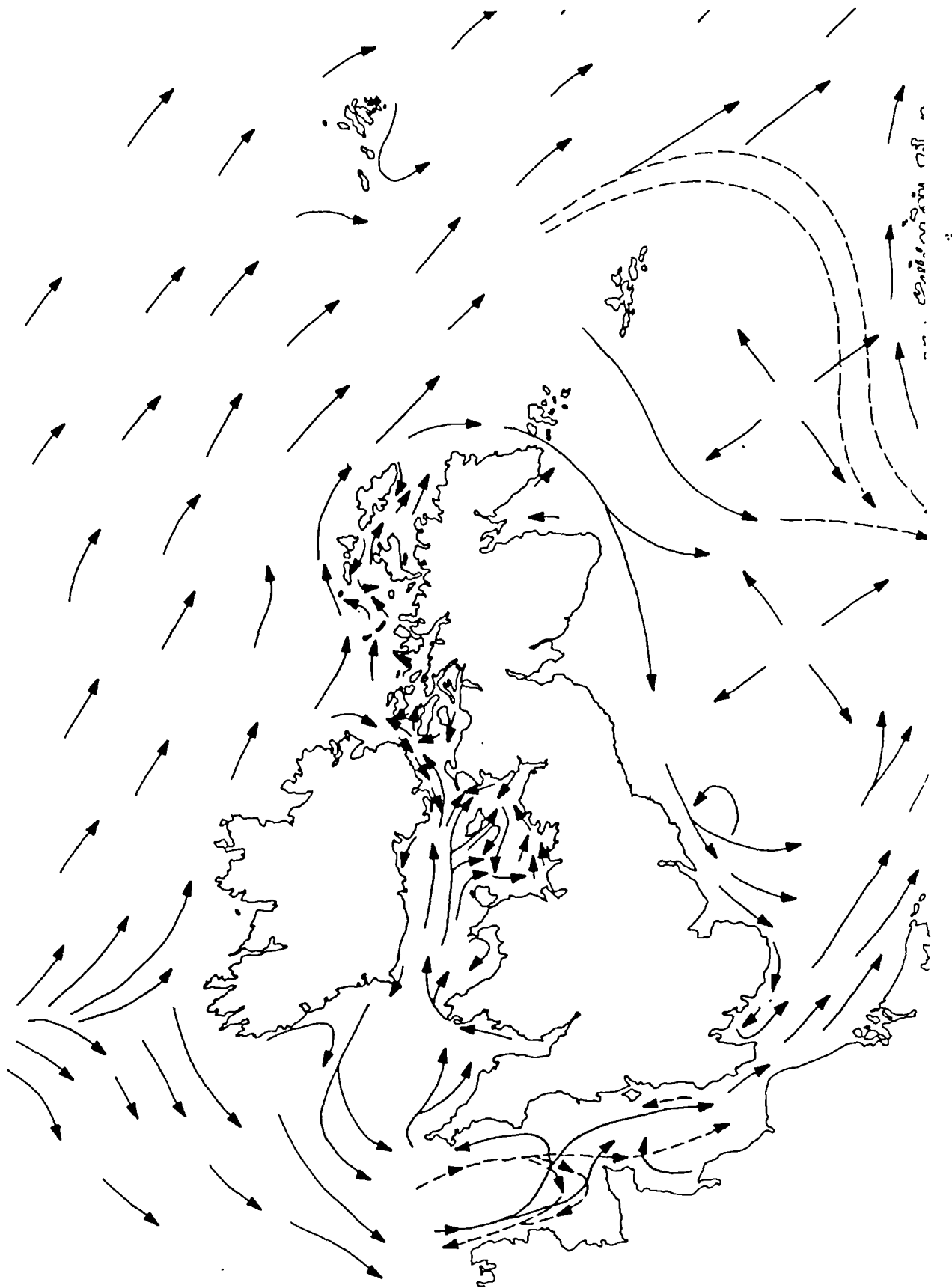


Fig.1.2 The general near-surface pattern of water movement around the British Isles (after Lee and Ramster, 1981).

Basin, and subsequently circulates onwards into north-eastern and north-western regions". Harvey (1982) has described the temperature and salinity characteristics in the north east Atlantic sector of SPMW, recording values between 8°C and 12°C within the Rockall Channel area; this water is more specifically referred to as Eastern North Atlantic Water (ENAW). Thermal changes within ENAW may be brought about by winter mixing and by exchanges with adjacent waters which will modify SPMW to the east of the Mid-Atlantic Ridge. Warm, saline Gulf of Gibraltar Water (GW), the product of sub-surface outflow from the Mediterranean, is thought to contribute to the appreciably higher salinity content of ENAW by comparison with SPMW of the western Atlantic (Cooper, 1952; Harvey, 1982), although this view has been challenged in recent years (Pollard & Pu, 1985). Ellett *et al.* (1986) still consider the contribution of GW north of the Bay of Biscay to be significant; they discuss the details of the observed seasonal signals of GW and other water masses within the area in detail.

Deeper water masses within the Rockall Channel include Labrador Sea Water (LSW) which enters from the south and, due to the retention of its specific characteristics and depth below the level of northern channels, most probably circulates out again to the south (Ellett *et al.*, 1986). Two deep water masses are known to enter from the north, although in limited volumes. These are Arctic Intermediate Water (AIW) from the shelf waters north of Iceland and the deeper water between Iceland and Jan Mayen. The second, and more significant, of these two water masses crossing the Wyville-Thomson Ridge and flowing southwards is Norwegian Sea Deep Water (NSDW). This NSDW descends from c. 500 m. at the ridge crest to between 1,000 m. and 2,000 m. in the Central Rockall Channel and as it does so much mixing and entrainment from overlying water occurs; however, it maintains its characteristics to the extent that it is latterly seen as a salinity maximum immediately below the LSW minimum. A general overview of the volumetric transport within the Rockall Channel is provided by Ellett *et al.* (1986) and summarized in fig.1.3. These features may give the impression of persistent current directions but, as Ellett *et al.* (1986) suggest, "the only currents persistent in

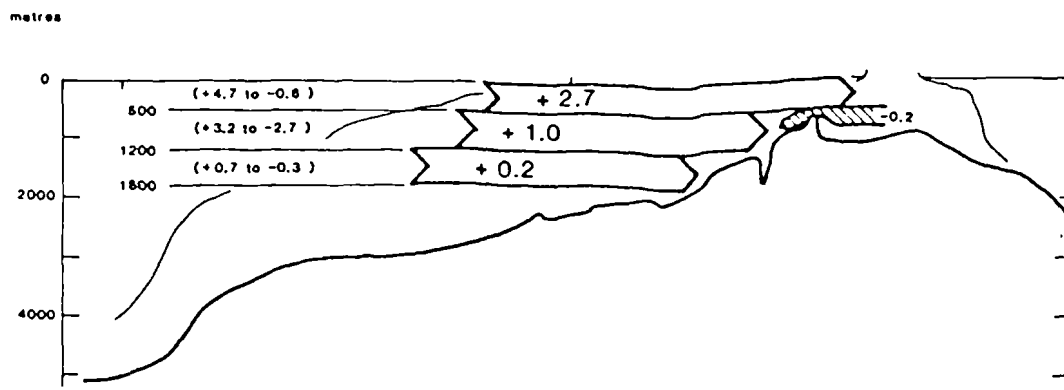


Fig.1.3 Estimated mean and extreme volume transport ($10^6 \text{ m}^3 \text{ s}^{-1}$) within the Rockall Channel at different depth bands. Note that the positive values are towards the north east (from Ellett *et al.*, 1986).

direction are those along the slope zone west of Britain".

This then, popularly termed the "Gulf stream", is the rather complex system which conveys warm, saline water to the Norwegian Sea.

1.2.3 Ocean-Shelf exchanges

Along the eastern side of the Rockall Channel are some interesting oceanographic features (Huthnance, 1986) and of particular interest are the ubiquitous eddies, often seen in infra-red satellite images, which suggest ocean-shelf transfers and may therefore be important to shelf circulation. Pingree (1979), suggests that Celtic Sea shelf-edge eddies, observed in Current-Temperature-Depth (CTD) measurements and satellite infra-red images, might arise from baroclinic instability and discusses their contribution to cross-shelf exchanges. However, the Hebrides/Rockall Channel slope appears to severely limit such transfers between ocean and shelf. While large-scale eddies do cover the deeper Rockall Channel, the warm water filament which follows the continental slope suffers relatively little deflection; constrained, as it is, by potential velocity to follow the depth contours. The degree of shelf-slope mixing can, in fact, be assessed quite simply, as discussed by Huthnance (1986), by

considering the Rockall Channel- shelf salinity differences that would exist under complete mixing (less than 0.035%) and those which are typically observed (0.2%; Ellett & Martin, 1973). The higher observed value suggests reduced mixing and therefore that much of the cross-slope flow only excurses rather than transfers on to (or off) the shelf (Huthnance, 1986). Even such limited excursions seem unlikely in view of the sharp 'fronts' reported along the Hebrides Shelf edge (Pingree & Mardell, 1981) and equally, the dense, winter-cooled waters that are reported to persist on the shelf beneath the summer thermocline (Booth & Ellett, 1983) will all act to reduce on/off shelf exchanges. However, exchanges above the seasonal thermocline may be greater.

Huthnance (1986) reports that many infra-red images, particularly during the summer months, reveal only poorly defined and confused shelf-edge/slope currents. Often, the most distinctive features occur well in on the shelf and one such feature indicative of a change in surface temperature relates to fresher coastal water extending just offshore from the 100 m. depth contour. Booth & Ellett (1983) state that a shelf current does exist and that this chiefly transports coastal water from the Irish Sea to the North Sea via N.E. Scotland and that during the summer-autumn half of the year there is little advection of Atlantic water across the shelf south of St. Kilda. They suggest that there may be a winter exchange which might account for the temperature-salinity relationship of shelf water offshore of c. 100 m. depth which is similar to offshore water beyond the shelf break.

Upwelling may well occur and has been reported by Dickson *et al.* (1980) from the Celtic Sea shelf-break as a cool band of water under calm weather conditions. Such features are known to occur after winds (meteorological forcing) which drive a surface Ekman transport offshore, although Huthnance (1986) reports that the surface transport itself has not yet been observed in this area.

Summer & winter conditions of temperature and salinity conditions from western Britain are briefly summarized in Table 1.1 and are based upon monthly mean values for August and February respectively. The bottom water conditions during the summer months are summarized in fig.1.4. Note the cold tongue of water

(<0°C) within the Faeroe Shetland Channel, indicative of NSDW overflow. In respect to salinity, it is worth noting, once away from coastal influences of surface runoff etc., that the salinity of the world's oceans varies between 32.5 ppt and 37.5 ppt.

	Summer	Winter
Mean surface temperatures	13°-16°C	5°-10°C
Mean bottom temperatures	9°- 13°C	6°- 9°C
Mean surface salinity	32‰-35.2‰	31‰-35.3‰
Mean bottom salinity	34‰-35.25‰	34‰-35.25‰

Table 1.1 Summary table of mean summer and winter conditions in western shelf seas

1.2.4 Plankton and circulation

A useful review of phytoplankton distribution along the shelf-break is provided by Holligan & Groom (1986); while fig.1.5 summarizes the main plankton associations around the British Isles. Physical mixing processes are generally recognized as determining the light and nutrient environment for phytoplankton growth in the surface layers of the oceans (Tett & Edwards, 1984). At temperate and polar latitudes, combined deep surface mixed layers and low solar radiation during winter months are reported to inhibit phytoplankton growth because of low light intensities. During the spring and early summer this restriction is removed as the seasonal pycnocline develops, holding the phytoplankton in the upper 20-60 m. of the water column. However, the populations are subsequently limited by the availability of increasingly depleted inorganic nutrients, particularly nitrates. Therefore, during the summer months high standing stocks of phytoplankton tend to persist only in regions where physical mixing processes maintain a relatively high upward flux of nutrients into the euphotic zone.

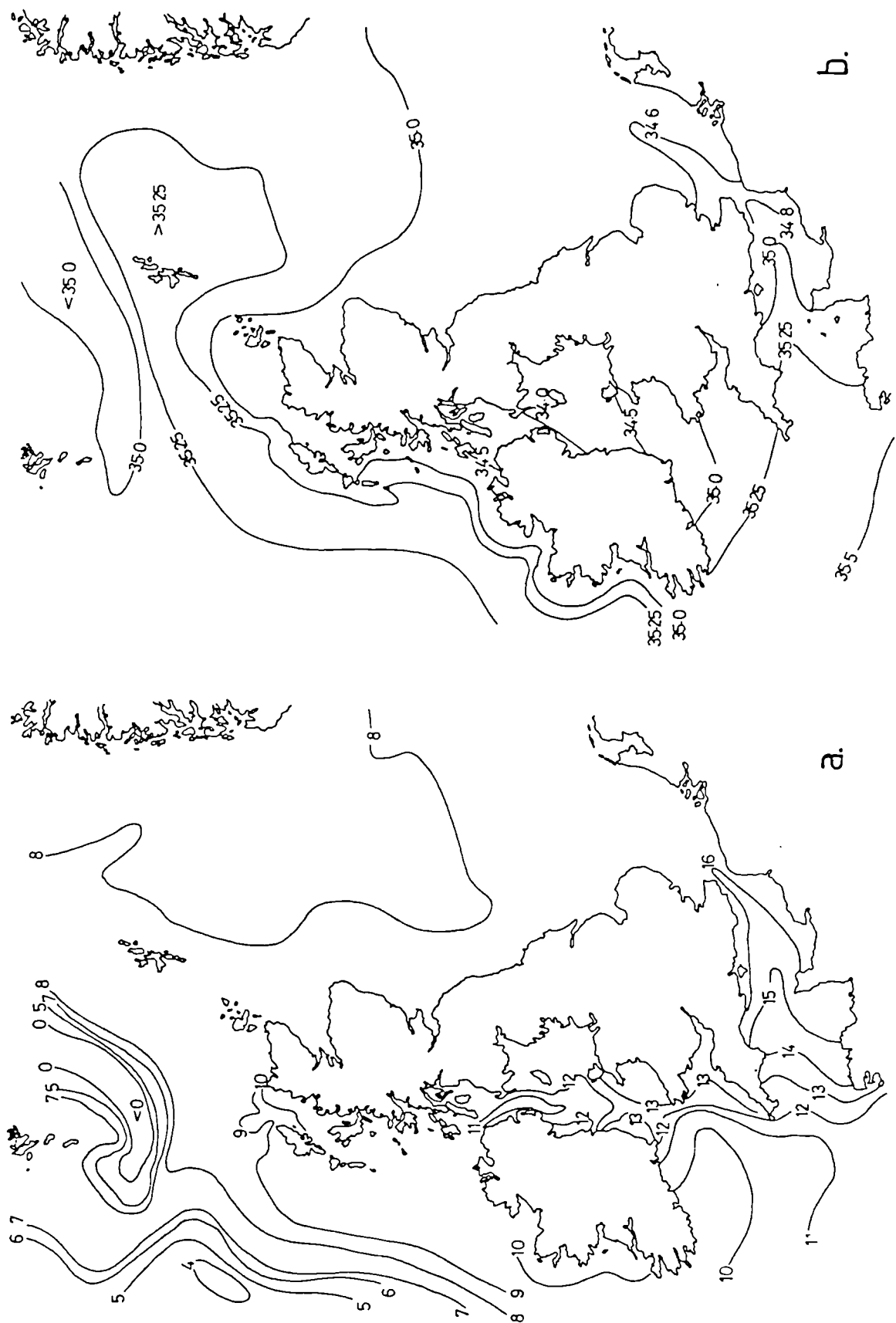


Fig.1.4 (a) Mean bottom temperature ($^{\circ}\text{C}$) (b) Mean bottom salinity (ppt.) in summer around the British Isles (after Lee and Ramster, 1981)

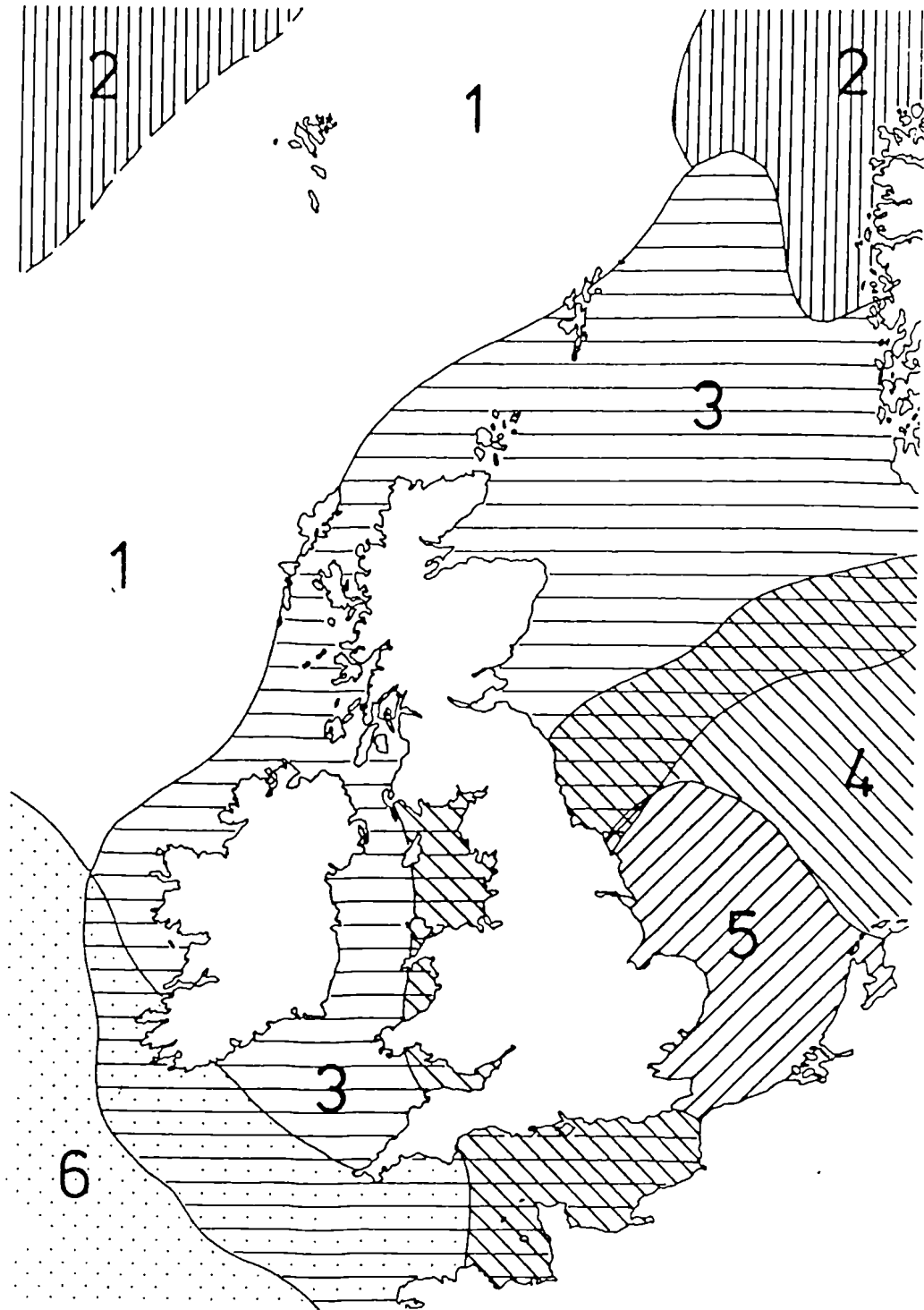


Fig.1.5 Summary map of the main plankton associations around the British Isles (after Lee and Ramster, 1981). 1= N.E. Atlantic taxa, 2= N.E. Atlantic & Arctic taxa, 3= Temperate shelf species (with N. Atlantic Drift influence), 4= Temperate shelf species, 5= Coastal influence strong (zone 3 or 4 taxa), 6= N.E. & tropical Atlantic taxa.

Thus an understanding of the physical environment might help to explain the spatial and temporal variations in phytoplankton abundance and these in turn may relate to the underlying patchyness in benthic faunas.

The Celtic Sea spring diatom bloom, for example, begins during mid-April in the region of weak tidal mixing south of Ireland, extending further south and eastwards as the seasonal thermocline develops. In this area of south west Britain surface waters at the continental shelf edge, close to the 200 m. depth contour remain relatively cool and phytoplankton rich throughout the summer months. The enhanced nutrient fluxes of this region are thought to be due largely to tidal mixing and associated internal wave activity, although other shelf-edge processes, including upwelling, remain to be assessed in this region (Holligan & Groom, 1986).

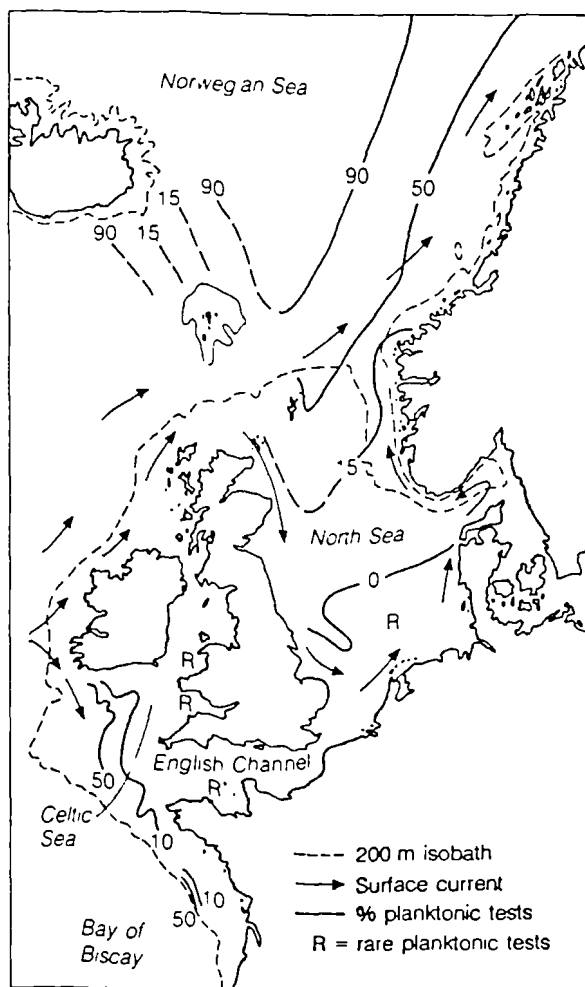


Fig.1.6 Planktonic:Benthonic foraminiferal ratios. Summary diagram of the percentage of planktonic tests and surface currents from N.W. Europe (from Murray, 1991).

Murray (1976) has discussed the presence of planktonic foraminifera on the continental shelf of south west Britain. Since planktonic foraminifera are essentially oceanic, tolerating only slight changes in salinity, it becomes clear that their presence in shelf sediments is due to transport from oceanic waters. Thus, the changing planktonic : benthonic ratios, changing planktonic species composition and changing test size in shelf sediments are all considered to reflect the amount of transport from oceanic waters onto the shelf (Murray, 1976). A summary map (fig.1.6) from Murray (1991) illustrates the changing planktonic : benthonic ratios which occur around the British Isles. The palaeoceanographic implications of the planktonic : benthonic ratios are dealt with in each of the three study areas and discussed at length elsewhere in this volume.

1.3 Quaternary Climate Change

Since much of the stratigraphic procedure applied to the subdivision of the Quaternary is based upon 'climatostratigraphic' principles, it is with the single feature which more than any other most readily characterizes the Quaternary period that I will now deal: climate change.

1.3.1 Evidence of climate change

If we investigate the period of time that has elapsed since the Miocene we discover that the major structural and palaeogeographic elements of the N. W. European continental margin have changed very little. The British Isles have remained an essentially positive structural area, separating the rapidly subsiding North Sea Basin from the smaller, generally less actively subsiding basins to the west. This is in contrast to, for example, the raised shorelines of actively emergent coastlines such as the Huon Peninsula of New Guinea (Bloom *et al.*, 1974; Chappell, 1974).

Against this generally straightforward structural background, the major climatically controlled processes of the Quaternary can

be observed. During glacial phases the usual equilibrium between tectonics, erosion and sedimentation is lost and a sensitive, highly dynamic state of disequilibrium arises.

Thus, the hallmark of the Quaternary stratigraphic record is the high amplitude and frequency of climatic oscillations as the earth has switched from glacial to interglacial mode. The evidence of this climatic change in Britain is most readily observed in upland areas where past glacial processes had a marked effect on the landscape and to a lesser degree in lowland areas where thick sequences of glacial deposits have accumulated. It is beyond the scope of this introduction to detail these features, however the recognition of much of the terrestrial evidence for glaciation is intimately associated with the historical development of the 'Glacial Theory' (eg. Agassiz, 1840). Unfortunately, the terrestrial record of climate change is fragmentary with successive glacial periods removing much of the preceding sedimentary and landform evidence. It is therefore from the deep ocean basins that much of our present understanding of these climatic fluctuations is derived.

The fundamental climato-stratigraphic record of the Quaternary period is widely accepted to be the deep sea oxygen isotope record. The changing ratios of the two stable isotopes of oxygen, ^{16}O and ^{18}O are thought to approximate to global ice volume (Shackleton, 1967; Shackleton & Opdyke, 1973) and are therefore an useful proxy record of global climate change. Micropalaeontological records from deep sea cores equally provide a record of past ocean conditions and allow, through the implementation of uniformitarian principles and complex transfer function equations, the reconstruction of past sea-surface temperature estimates based upon planktonic fossils (eg. Imbrie & Kipp, 1971).

North Atlantic deep sea cores are considered by Bowen (1991) to provide first-order indications of the volume of ice on adjacent continents, although Shackleton (1987) has stressed that oxygen isotope records cannot be directly correlated with continental ice mass development. An important core in this respect is the Deep Sea Drilling Project (D.S.D.P.) hydraulic piston core recovered from site 552A from the Hatton Drift on the

western side of the Rockall Plateau ($56^{\circ}02.56'N$, $23^{\circ}13.88'W$; water depth -2311 m.). This core has yielded an almost continuous and undisturbed record of climate change from the onset of glaciation up to the present day (Zimmerman *et al.*, 1984).

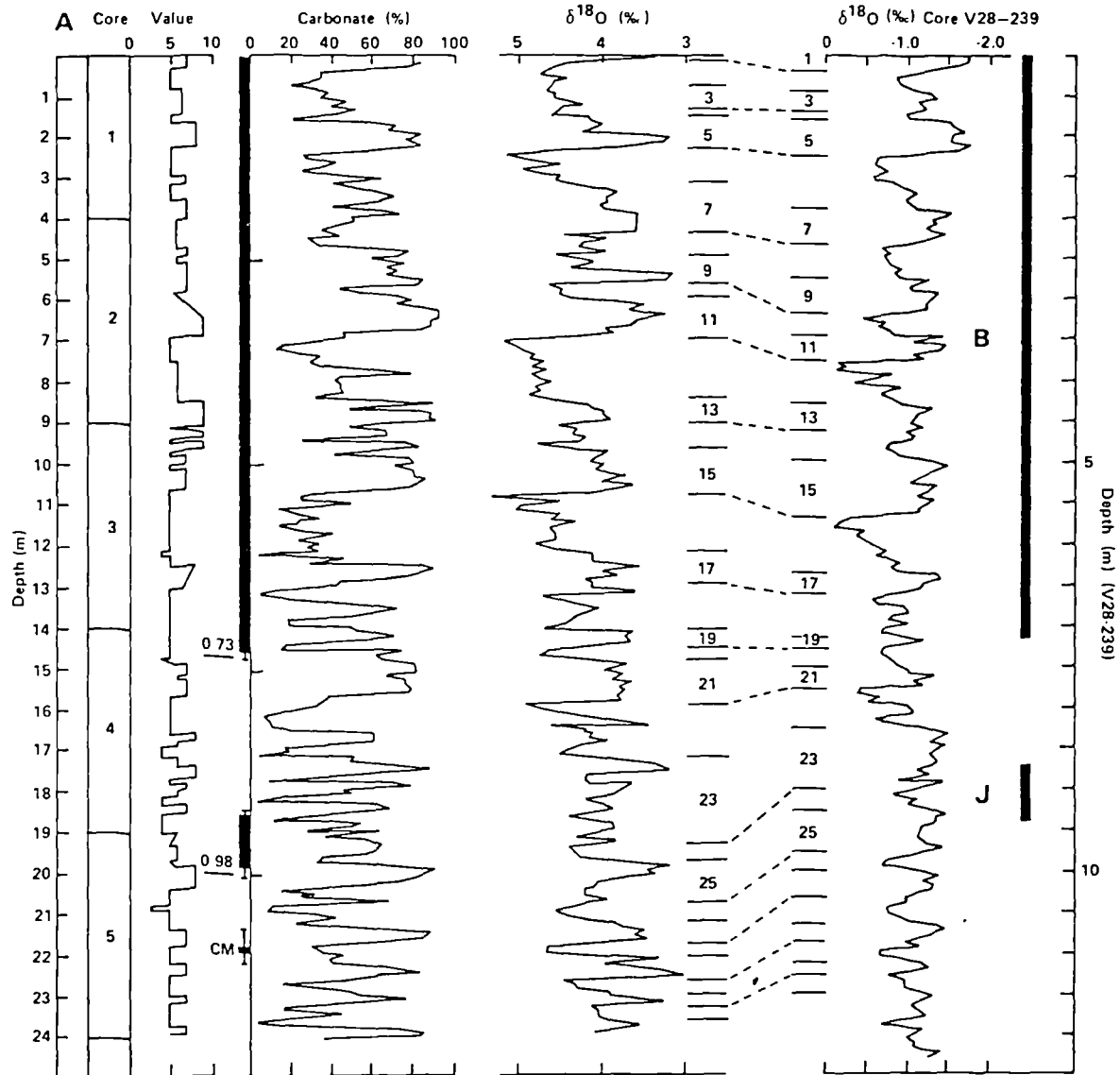


Fig.1.7 Summary of percent carbonate and oxygen isotope data for Hole 552A from the Rockall Plateau correlated with the oxygen isotope record of the Pacific core V28-239 (Shackleton and Opdyke, 1976). Records from both cores are plotted linearly with depth. Proposed oxygen isotope stage boundaries, palaeomagnetic data, and colour values (light = 10, dark = 1) are also indicated (from Zimmerman *et al.*, 1984).

Sediments from this core consist of alternating terrigenous muds and carbonate-rich biogenic oozes, thought to reflect alternating periods of accumulation during glacial and interglacials respectively. Down-core variations in a number of parameters from this site, including oxygen and carbon isotopes (Shackleton & Hall, 1984), have been reported to reveal the climate changes that have influenced N. W. Europe and particularly the western seaboard of the British Isles. The data from this core are summarized in fig.1.7. Of particular interest are the analyses of the polarity of the core sediments and the establishment of magnetostratigraphic time control together with nannofossil evidence which indicate that the first major glacial event, as reflected in the oxygen isotope record and ice-rafting horizons, occurred at 2.37 million years ago, prior to the generally accepted Pliocene/Pleistocene boundary. This "parochial view that the Pliocene/Pleistocene boundary should be placed at 2.4 Ma at the beginning of the first major late Neogene glaciation in the North Atlantic" is strongly contested by Jenkins (1991) who favours the retention of the 1984 Global Stratotype Section and Point for the base of the Pleistocene in the Vrica section, Italy. It also serves to illustrate the relatively slow accumulation rates and hence generally poor stratigraphic resolution which typify most deep sea cores.

1.3.2 Terminology

Within the context of the present study we are not immediately concerned with the details of the Quaternary climatic cycles as defined by deep sea oxygen isotope records.

Neither need the complexities of the terrestrial record, its fragmentary evidence, problems of correlation or wholly out-dated terminology concern us unduly. The main time-span under investigation here is from the end of the last glacial maximum at approximately 18,000 years BP to the present day. However, even this most recent and relatively short period of earth history, with its apparently global climate signals and generally well preserved sedimentary record is open to controversy.

The terminology adopted in this study is largely based around British usage much as outlined by Lowe and Gray (1980), who argue

in favour of a climatostratigraphic scheme, and is somewhat simpler than the alternative chronostratigraphic subdivision of the Lateglacial period in Norden (Mangerud *et al.*, 1974). The two schemes are outlined in Table 1.2.

Radiocarbon years B.P.	chronozones
10,000 -----	Younger Dryas
11,000 -----	Allerød
11,800 -----	Older Dryas
12,000 -----	Bølling
13,000 -----	

Table 1.2.a The chronostratigraphic subdivision of the Lateglacial period in Norden (from Mangerud *et al.*, 1974).

In this study a twofold subdivision of the Devension Lateglacial is recognised, consisting of a Lateglacial Interstadial followed by a Lateglacial Stadial. These two subdivisions are commonly referred to in standard British usage as the Windermere Interstadial and the Loch Lomond Stadial; while the present warm phase is referred to as the Holocene.

A discussion of the validity and implications of the stratigraphic schemes adopted here can be found later in this volume or, alternatively, the reader is referred to Lowe and Gray (1980) for a general discussion.

Radiocarbon years B.P.	Climatostratigraphic units
	Flandrian (Holocene) 'Interglacial'
10,000	-----
	transition
10,500	-----
	Younger Dryas Stadial
11,000	-----
	transition
12,000	-----
	Lateglacial Interstadial
13,000	-----
	transition
14,000	-----
	Late Weichselian/ Late Devension/ Late Midlandian Glacial.

Table 1.2.b The climatostratigraphic subdivision of the Lateglacial period of N.W. Europe (from Lowe and Gray, 1980).

1.3.3 Mechanisms of climate change

The hallmark of the Quaternary period, as discussed above, are the high amplitude and frequency of climate oscillations. But what causes climate change? This fundamental question has now largely been answered and tested, although the actual mechanisms by which climate change occurs are poorly understood and appear to be highly complex.

The cause of climate change appears to arise due to a response in the surface temperature of the earth resulting from changes in the earth's axis of rotation and orbit around the sun which, in turn, control the amount of radiation received at a given latitude. This is the basis of the 'Astronomical Theory' of Croll, subsequently elaborated upon by the Yugoslavian geophysicist Milutin Milankovich (1924) and with whose name these

astronomical cycles are associated. There are, in fact, three major cycles relating to the eccentricity of the orbit (periodicity = 96,000 years), axial tilt (periodicity = 42,000 years), and the precession of the equinoxes (or 'wobble' of the axis of rotation; periodicity = 21,000 years).

Hays *et al.* (1976) have demonstrated, by detailed analysis of deep sea core data, that the sequence of colder and warmer intervals, as predicted by the 'Milankovitch Theory' do in fact exist. However, while the superimposition of these cycles is partly thought to produce the complexities of the climate record and while orbital forcing may drive climatic changes, the actual mechanisms by which it occurs remain largely unresolved.

1.3.3.a Biological Pumps

Northern ice sheets translate Northern Hemisphere seasonality into climatic change around the world. For example, sea level changes, which can be measured, and albedo (reflectivity) changes, which can be numerically modelled, relate directly to the extent of the earth's, and particularly the Northern Hemispheres', ice sheets and are immense. These changes, brought about by variation in the extent of the earth's ice sheets, are not considered to drive climate change, although complex feedback mechanisms between them probably exist. Variations in past atmospheric carbon dioxide, a so-called 'green-house' gas, can be directly measured from ice cores (eg. Dansgaard *et al.*, 1989) and suggest that during the last glacial maximum values were two thirds of the interglacial levels. It soon became apparent that the oceans were the only way in which such major changes in carbon dioxide levels could be accommodated; in fact the oceans can hold sixty times more carbon dioxide than the atmosphere (Broecker and Denton, 1990).

Carbon dioxide readily diffuses between atmosphere and ocean and the concentration in the surface waters is thought to regulate the atmospheric concentration. In turn, the concentration of the gas in surface waters is regulated by the living photosynthetic organisms of the oceanic plankton which act as a biological pump, transferring carbon dioxide, largely in the form of biogenic carbonate, from the surface waters to the ocean bottom. If deep

water (CO_2 & nutrient rich) mixing with the surface is slowed, then surface waters become increasingly depleted in carbon dioxide leading to increased surface diffusion and hence atmospheric depletion. Thus, during glacial periods, either because of altered mixing (circulation) patterns in the oceans or changing surface water ecology, the world ocean's biological pump appears to have been more efficient. However, the link between pumping efficiency and ocean circulation is controversial.

1.3.3.b The Atlantic conveyor

Long before the ice core work allowed direct inferences to be made upon past atmospheric gas concentrations, the changing state of the oceans from glacials to interglacials was known from the reconstruction of 'ecological water masses' based on planktonic foraminiferal studies, using present day species distributions to infer past climate. Such tracers of ocean water masses, as all oceanographers are aware, do not need to be biological. Recently, Boyle has utilized the geochemical signal of Cadmium, an element closely related to phosphate and nitrate nutrients in modern oceans, which is incorporated into the carbonate tests of foraminifera in equilibrium with the composition of the sea water in which they live. Thus, a past record of nutrient availability in the surface waters from which the foraminifera derive becomes a possibility and Boyle was able to demonstrate that a key signature of the Atlantic Ocean's present day circulation was missing during glacial times. The mounting evidence indicates that radical changes in world-wide circulation have taken place between glacial and inter-glacial cycles.

Every winter, at about the latitude of Iceland, water of relatively high salinity, flowing northwards at intermediate depths (c. 800 m.) rises to the surface as a result of strong surface winds. This water cools rapidly from about 10°C to about 2°C and, together with its high salinity, it becomes unusually dense, sinking to the ocean bottom; this is the process which forms North Atlantic Deep Water (NADW). The heat which is released is equivalent to about 30% of the yearly direct input of solar energy to the surface of the North Atlantic Ocean and produces the

unusually mild winters of N. W. Europe. This northern warming is often mistakenly ascribed to the Gulf Stream, an oceanographic feature which ends well to the south.

This so-called Atlantic 'conveyor', which releases vast quantities of heat to the North Atlantic and sends immense volumes of water into the deep ocean was shut down during the last glacial period. The 'conveyor' seems to have re-started at about 14,000 years BP which corresponds to the time interval 14,000 to 13,000 years BP when rapid warming is known to have occurred. Broecker and Denton (1990) suggest, on the basis of sudden changes in the proxy record of the atmospheric/oceanic system, that the whole ocean-atmosphere system may possibly 'jump' from a glacial mode of operation to an interglacial mode; citing changes in seasonality as the ultimate causes of these mode shifts.

1.3.3.c The Younger Dryas

The rapid return to almost full-glacial temperatures in the North Atlantic and Europe during the deglaciation of the Northern Hemisphere ice sheets occurred between 11,000 to 10,000 years BP and is known as the Younger Dryas.

Much controversy surrounds the mechanisms which have been proposed to have caused this abrupt and relatively short-lived cold period during which the Polar Front extended southwards to a position off Southern Ireland (Ruddiman *et al.*, 1977). This relates largely to the suggested strong non-linearity in the response of the climate system to orbital forcing over short time intervals. Thus, some other mechanism has to be evoked, however no single mechanism is widely accepted.

This rapid climate shift has been explained by abrupt reorganization of surface and deep ocean circulation in the North Atlantic and from the evidence discussed above this seems highly likely. Broecker *et al.* (1988, 1989) propose that the re-routing of the vast volumes of melt-water from the wasting North American ice sheet from the Mississippi to the St. Lawrence River and directly into the North Atlantic caused an influx of fresh water which lowered the density of surface waters sufficiently to 'turn-off' the 'conveyor' and generate the cooling of the Younger Dryas. The subsequent re-routing of this melt-water back to the

Mississippi about 1,000 years later may seem rather convenient to the hypothesis, except that foraminiferal tests in sediments from the Gulf of Mexico (the recipient of Mississippi discharge) apparently record the expected oxygen isotope ratios associated with these melt-water events. However, the extent to which melt-water discharge into the Gulf of Mexico would have changed simply as a response to the Younger Dryas climatic event does not appear to have been widely discussed?

Fairbanks (1989), on the other hand, proposes that such diversion mechanisms are unnecessary and that the Younger Dryas chronozone can be characterized by minimal melt-water discharge, particularly between 11,000 years to 10,500 years BP, with two distinct melt-water episodes (Terminations IA and IB of Duplessy *et al.*, 1981) before and after it.

More recently, work by Jansen and Veum (1990) has addressed the question of deglaciation and the impact on deep water circulation in the North Atlantic. One problem of oxygen isotope records, particularly during events such as the Younger Dryas, is the question of to what extent are they determined by the deglacial transfer of light oxygen (^{16}O) to the ocean and to what degree are they determined by temperature? In an attempt to answer this question planktonic and benthonic oxygen isotope records were examined and any 'overshoots' recorded. They demonstrate that melt-water discharge was reduced by up to 80% during the Younger Dryas and confirm Fairbank's view of deglaciation in two major steps. They propose that the Laurentide ice sheet, being a large part of the global ice volume signal, experienced two phases of rapid disintegration separated by a slower rate during the Younger Dryas.

However, the actual mechanisms generating this event remain an enigma, although a strong negative feedback mechanism is implied.

1.3.4 Climate and Environment

While the mechanisms which may have generated climatic change continue to arouse debate, what are the actual effects of this climate change on the environment?

The effect of ice sheets on the North Atlantic Ocean north of 45° - 50°N, particularly as they respond to the lower frequency orbital rhythms, are considered to reduce oceanic temperatures with "little or no lag" (Ruddiman, 1987). Higher frequency signals may not produce such a response and it is becoming increasingly apparent from 'events' such as the Younger Dryas that complex feed-back mechanisms may apply to the climate system.

Bowen (1991) states that the British and North American ice sheet are likely to have been in-phase for much of the Quaternary. However at the end of the last major glaciation, it is widely accepted (eg. Boulton, 1990) that the North American ice sheet, which generated at least half the North Atlantic ice volume signal, decayed more slowly than the smaller British ice sheets. Thus, these ice sheets were out-of-phase during the critical Lateglacial period and the British Devensian ice sheets led the global eustatic cycle during their decay.

1.3.4.a The North Atlantic Polar Front

Possibly the most important climate effect on the British Isles, located as they are on the margins of the north east Atlantic, has been the movement of the Polar Front and extent of the Gulf Streams through space and time (Ruddiman & McIntyre, 1976, 1981). Our location has resulted in high ice sheet sensitivity to external forcing. Thus, cooling as a result of orbital forcing together with a nearby ocean surface warmed by the Gulf Stream would have allowed rapid and extensive ice growth, particularly in upland and western areas (Ruddiman *et al.*, 1980). This may explain the initial build-up of ice as the Polar Front migrates southwards in the north east Atlantic and such an explanation may also apply during the Younger Dryas in upland Britain (eg. Sutherland, 1984).

It is difficult to understand the mechanisms which generate spatial variability in precipitation, except that the southwards migration of the Polar Front may increase latitudinal temperature gradients and hence pressure gradients and storminess within the region. Dansgaard *et al.* (1989), for example, report evidence from the DYE 3 ice core of southern Greenland that suggests that over a period of as little as 50 years at the end of the Younger Dryas

that precipitation increased by up to 50% and that temperature rose by 7°C. This abrupt termination of the Younger Dryas, dated at c. 10,700 years BP, is proposed to be related to the rapid retreat of sea ice cover in the north east Atlantic.

The concept of spatial variability within weather systems can be applied in terms of zones of maximum effective precipitation. The sequence of ice expansion during the Late Devensian glaciation of Scotland, for example, is considered to be a response to effective precipitation. Initially, Highland ice expanded across the Central Lowlands of Scotland onto the Southern Uplands, subsequent growth of the Southern Uplands ice centre, associated with a change in effective precipitation, led to expansion onto ground occupied earlier by Highland ice (Geikie, 1894; Sissons, 1967).

Some of the glacial readvances associated with the deglaciation of the Late Devensian ice sheets may relate to changing zones of maximum effective precipitation. One example of a readvance of the Late Devensian ice sheet is from the Tremadoc area of Cardigan Bay where Welsh till is found overlying Irish Sea till (Garrard, 1977) and also further to the west where a readvance of the Irish Sea glacier is proposed to as far south as Bardsey Island (cf. fig.10, Garrard, 1977). It is interesting to postulate on the age of these two readvances, particularly with respect to the location of their feeder ice-caps; one might expect the Tremadoc Bay readvance to predate the Irish Sea Ice readvance if the hypothesis that the British ice sheet was starved of precipitation until the northerly retreat of the Polar Front is correct. Garrard (1977) states that the readvance of both Welsh Ice and Irish Sea Ice was synchronous. More recently, McCarroll (1991) has argued that the evidence, on the Llyn peninsula and Anglesey at least, do not support the concept of a Gwynedd readvance of the Irish Sea glacier or the marked climatic deterioration which such a readvance implies.

1.3.4.b Sea level and isostasy

The question of sea level change is intimately related to climatic change and the 'locking-up' of large volumes of water in the major ice sheets; implying a glacio-eustatic response. A

Barbados (Fairbanks, 1989). This work, with its major climatic and oceanographic implications, extends the fossil coral-reef sea level curve of Lighty *et al.* (1982) back towards the last glacial maximum (fig.1.8). The results of this work suggest that sea level was 121 ± 5 m. below present day sea level during the last glacial maximum. However, there are possible problems with this record which include a number of geophysical assumptions, not least of which is that the rate of local uplift of ~ 34 cm K.yr^{-1} has remained constant during the interval of accumulation and since oxygen isotope stages 6-5. The Fairbanks curve does provide a valuable back-drop of eustatic sea level change against which to compare the sea level estimates of the present study.

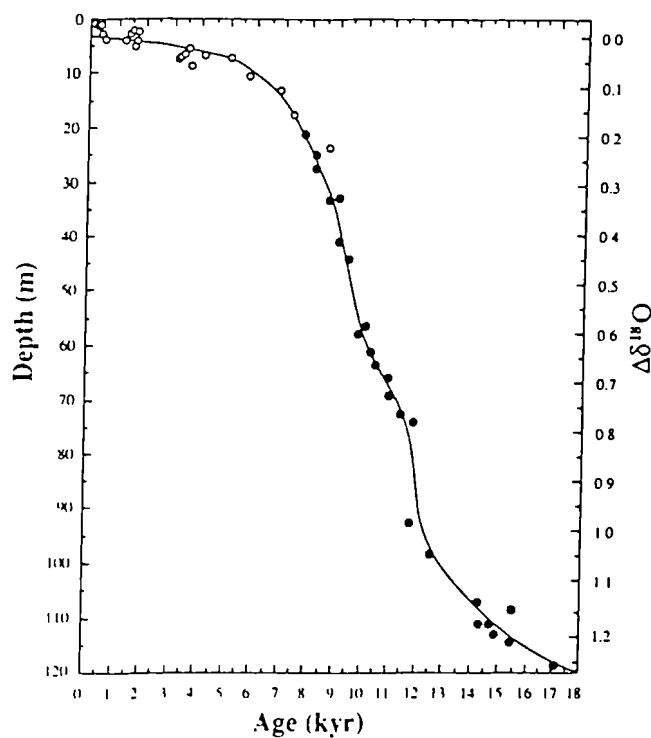


Fig.1.8 The Barbados sea level curve based on radiocarbon-dated *Acropora palmata* and corrected for a mean uplift of 34 cm Kyr^{-1} (from Fairbanks, 1989).

It is against this global or eustatic sea level curve that glacio-isostatic effects can be viewed and some of the models (eg. Boulton, 1990) tested. Local sea level curves vary considerably and record relative sea level only, yet such curves when analysed together with the eustatic sea level curve can allow an estimate of the isostatic components as they have acted through time. This is one approach which I have adopted from the Hebridean Shelf and the reader is referred to Boulton's (1990) models of relative sea level change during glacial cycles for a thorough review of these phenomena. In summarizing the Lateglacial situation in Britain it is important to remember that the deglaciation of the British Devensian ice sheets led the deglaciation of the larger Laurentide ice sheet, thus glacio-isostatic rebound took place before global sea levels 'took-off' and hence the relative sea level curve from the Hebridean margin during the Lateglacial and Holocene is one of initial regression followed by transgression.

According to Boulton (1990), the growth and decay of Pleistocene ice sheets has involved an exchange of mass of about $4 \times 10^7 \text{ Km}^3$ between the oceans (area = $2.2 \times 10^8 \text{ Km}^2$) and the ice sheets of northern Europe & N. America ($14 \times 10^6 \text{ Km}^2$ area; not including Antarctic & Greenland ice sheets which are thought to have undergone little volumetric/areal change during last glacial cycle). This results in fluctuations of ocean level of about 120 m. and changes in ice sheet thickness at mid-latitudes of up to 4 Km.

The cyclical concentration & dissipation of mass from mid-latitude ice sheets affects sea level in three principle ways:-

- (1) global eustatic changes directly related to ice volume change.
- (2) local isostatic crustal flexure in response to ice/water loading.
- (3) local changes of sea surface in response to gravitational attraction of changing masses.

These components of sea level act to produce a relative sea level (R_s) which is the local level of the sea in relation to a

fixed point on the solid earth surface. Thus, using an up-positive, down-negative convention:

$$R_s = E_s + I_s + G_s$$

where:

E_s = net change in eustatic sea level

I_s = net change in isostatic displacement of crust

G_s = net change in gravitational change in sea surface

[The above are considered in relation to an assumed interglacial equilibrium condition].

1.3.5 Reconstructing climate change

One of the major aims of this study has been to attempt palaeoenvironmental reconstructions based upon included microfaunas of the shelf sediments from western Britain. The concept of reconstructing global climate, particularly from deep sea sediment cores, has been discussed briefly above. However, much of the evidence from the British Isles and indeed N. W. Europe comes from the terrestrial record. Relatively few studies have dealt with the stratigraphy and palaeoenvironmental reconstructions of the Lateglacial period from the shelf seas. The reasons for this situation arise partly from the costs involved in obtaining shelf sea cores and partly from the fact that the terrestrial record generally preserves a higher resolution, less disturbed record of this period. In the following chapters, particularly concerning the Hebridean Shelf, I hope to demonstrate that the palaeoenvironmental changes of this period can in fact be resolved from shelf sequences and that they may eventually provide a useful link between the deep sea record and the terrestrial record.

Quaternary palaeoenvironmental reconstructions are firmly based upon uniformitarian principles and as such an understanding of the modern ecological requirements of the species under investigation is essential if we are to fully comprehend the implications of the faunal associations which we encounter in the fossil record. Unfortunately, our knowledge of the modern distribution and ecological requirements of benthonic foraminifera

is limited, to the extent that some authors (eg. Peacock and Robinson, 1982) suggest that more is known of their distribution in Quaternary deposits than their modern distribution.

1.4 Glacial marine sediments and environments

Perhaps the earliest record of "marine drift" was that of Geikie (1863) who, in his studies of the glacial deposits of Scotland, interpreted poorly sorted, fossiliferous and stony muds interbedded with normal marine clays as having been deposited directly from icebergs and sea ice. The term glacial-marine was first employed by Philippi (1910) to describe till-like sediments dredged from the sea floor of Antarctica. There is only slight discrepancy in the definitions employed for glacial marine sediments as there is in the terminology: glacial-marine, glaciomarine, or glacimarine.

Anderson *et al.* (1980) define "glacial marine sediment" as any marine deposit which bears evidence of having been deposited by floating ice. On the other hand, Andrews and Matsch (1983) and Powell (1984) refer to "glacimarine sediments" as "made up of debris deposited in the marine environment after release from either grounded or floating glacier ice" (Dowdeswell, 1987). The latter definition would seem preferable when the implications of deposition from grounded ice into a very proximal marine environment are considered (Barrett *et al.*, 1983). The definition offered by Anderson *et al.* (1980) is in many ways more applicable to the Antarctic than elsewhere; and as Anderson *et al.* (1983) state:

"The Antarctic continental shelf is so deep that it will not be isostatically raised to a level where coastal and subaerial processes will influence sedimentation during major interglacials. In fact, were the ice sheet to melt, much of the shelf would remain at depths below 500 m., even after complete isostatic rebound. Thus, a thick marine sequence would be deposited above the glacial facies. It is this, plus the lack of meltwater in Antarctica, that distinguishes it from other modern glacial marine environments."

Thus, some of the techniques and models developed and employed for use in Antarctica may not be applicable elsewhere and it is as well to remember the exceptional conditions which prevail there.

The term glacial-marine is a very catholic one and this is reflected in the variability of sediment composition and facies relationships. No single facies is characteristic of glacial marine sediments in general and in a similar way to deltaic sediments one can distinguish between proximal, distal and intermediate facies. Glacial marine sediments are also not only confined to the Quaternary, but are also found in the older geological record; for example, the late Precambrian (Anderson, 1983), the Gowganda Formation, Precambrian of Northern Ontario (Miall, 1983), or the late Precambrian of East Greenland (Moncrieff & Hambrey, 1990).

Having established that glacial marine sediments are deposited in the marine environment after release from grounded/floating glacier ice, the next step will be to consider the complex system which makes up the glacial marine sedimentary environment. Following this is a brief review of some recent works which attempt to interpret these sediments and erect facies models.

1.4.1 The glacial marine sedimentary environment

Dowdeswell (1987; fig.1.9) illustrates the complex system making up the glacial marine sedimentary environment. Facies models, based on observations of processes and facies from modern glacial marine environments (eg. Elveihøi *et al.*, 1983; fig.1.10) are useful but, as Dowdeswell and Scourse (1990) have emphasized, observed patterns of sedimentation vary three-dimensionally. However, two-dimensional facies models which portray proximal - distal transects from glacier source towards open ocean do provide useful analogues for the interpretation of older glacial sequences. It should be emphasized though that modern glacial marine environments do not provide the most appropriate analogues for conditions during former glacials; the large marine-based ice sheets of the northern hemisphere shelf areas, and the presence of grounded glacier ice to the edge of continental shelves is no

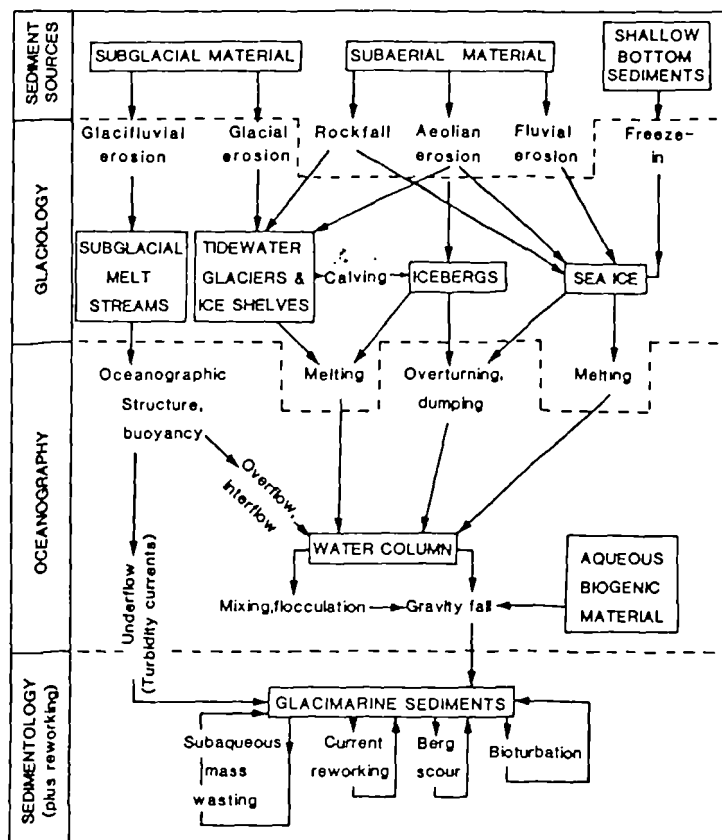


Fig.1.9 Summary of the complex glacial marine sedimentary environment system (from Dowdeswell, 1987). Labelled boxes represent sediment stores, other labels represent processes.

longer a feature of modern glacial marine environments as it may have been during the last glacial maximum (Dowdeswell & Scourse, 1990).

While models such as these depict modern processes and facies, some models are based largely upon inferential evidence concerning lithofacies and seismic sections (eg. Eyles & McCabe, 1989). The application of marine seismic records in environmental interpretations is an area of controversy according to Dowdeswell & Scourse (1990) and this is understandable when one considers that environmental interpretations may themselves be based upon interpretations of marine seismic records. This problem is largely overcome when 'ground-truth' from core data becomes available and the power of multichannel seismic records in mapping the geometry of seismic units should not be underestimated.

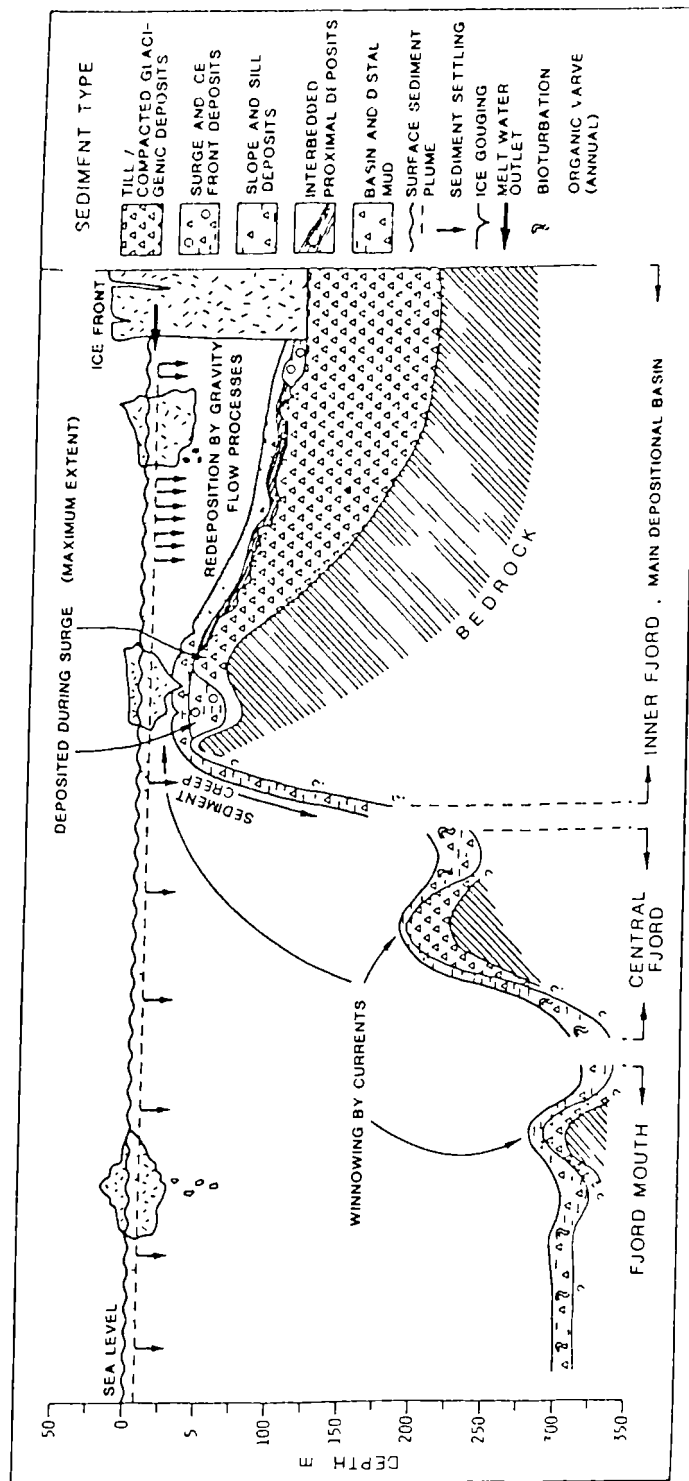


Fig.1.10 Two dimensional schematic model representing processes and sediments in a fjord fed by a tidewater glacier; based on Kongsfjorden, Spitsbergen (from Dowdeswell and Scourse, 1990; based upon an original figure from Elverhøi *et al.*, 1983)

The description of glacial marine sediments, or any other sedimentary sequence, should ideally be founded upon non-generic classification in terms of lithofacies types; this is critical to subsequent environmental interpretation (Miall, 1977, 1978; Eyles *et al.*, 1983). The need for basic physical observation, objective measurement and description of glacial sequences divorced from genetic perspectives and terminology have also been emphasized by Birkeland *et al.* (1979) and Martin (1980).

In much of the literature the term 'till' is often used, although it is in fact a generic term referring to "an aggregate whose particles have been brought into contact by the direct agency of glacier ice and which, though it may have undergone subsequent glacially-induced flow, has not been significantly disaggregated" (Boulton, 1976). Preferable to the term 'till' is the non-generic term 'diamict' which refers to any poorly sorted clast-sand-mud admixture, regardless of depositional environment. The term therefore specifies only a range of particle sizes and not the "relative abundance of any or all size classes" (Frakes, 1978). The four-part classification of Eyles *et al.* (1983) is a modification of the lithofacies code established by Miall (1977, 1978) for the classification of fluvial deposits. This is a useful scheme but can be difficult to apply unless one is dealing with a well exposed section, it is not yet adopted in full and a number of important and useful papers pre-date it. For example, the three basic sediment types recognised by Anderson *et al.* (1980) from the Antarctic continental shelf have been related to former glaciologic and marine conditions and provide a generalized, if somewhat oversimplified, model. The three sediment types are:

Type 1 - unstratified gravelly-sandy-muds and gravelly-muddy-sands with an unsorted, negatively skewed matrix. Cohesive strengths are typically high ($>2.5 \text{ Kg/cm}^2$). Microfossils are rare, although reworked diatoms are known from the Ross Sea cores of Kellog *et al.* (1979). The most diagnostic features are textural and mineralogical homogeneity.

These deposits are confined to the continental shelf and are exposed on the seafloor seawards of the modern grounded ice-front. There is no evidence that these deposits have been influenced by

marine processes and these sediments are interpreted as 'basal tills' and a lodgement process of deposition is considered most likely (Domack *et al.*, 1980). This interpretation is further supported by their high cohesive strength (Dreimanis, 1976) as well as textural and mineralogical homogeneity within individual units (cf. Shepps, 1958; Karrow, 1976). These sediments have been identified in cores from as far seaward as the continental shelf edge, suggesting that grounded ice has extended this far offshore in the past.

Type 2 - crudely to well stratified gravelly muds, having a higher silt/clay content than Type 1 deposits, being better sorted, especially the silt fraction. Typically, the texture is heterogenous down-core. Cohesive strengths are lower ($< 2.5 \text{ Kg/cm}^2$) and contain microfossil assemblages. It is interesting to note that pebbles from both Type 1 and Type 2 sediments of the George V continental shelf plot within Boulton's (1978) basal debris field (Domack *et al.*, 1980).

These 'Factor 2' sediments were studied by Chriss and Frakes (1972) who concluded that they reflect a combination of ice rafting (from icebergs and ice-shelves) and normal marine sedimentation. They have also been termed 'compound' glacial marine sediments since they are associated with relatively weak bottom current activity and are enriched in current derived silts and clays.

Type 3 - fossiliferous, unstratified, gravelly sands with a poorly sorted sandy matrix. Abundant and diverse microfossil assemblages. Alternatively known as 'residual' glacial marine sediments, they have been interpreted as representing ice-rafted sedimentation coupled with bottom current activity which is sufficiently strong ($15\text{--}30 \text{ cm.s}^{-1}$, cf. Gill, 1973) to winnow silts and clay. These sediments appear to be most widespread over the shallower portions and outer continental shelf where strong contour currents exist. Jacobs *et al.* (1979) have noted that sub-glacial circulation is generated during the release of sediments from melting ice beneath the Ross Ice Shelf. In addition, tidal forcing occurs beneath Antarctic ice shelves

(Williams and Robinson, 1979) which will also influence sedimentation processes.

Thus, the depositional boundary between the basal tills (Type 1) and the glacial marine sediments (Type 2 & 3) marks the grounding line of the Antarctic ice sheet. The boundary between Type 2 and Type 3 sediments represents an oceanographic boundary dependent upon the bottom current regime.

In the above scheme, a recurring theme is the influence of bottom currents on the sedimentation processes; the details of three-dimensional facies relationships are ignored. In the subpolar North Atlantic and Norwegian Sea extensive studies of surface oceanography have been undertaken (Bramlete and Brady, 1941; McIntyre *et al.*, 1972, 1976; Ruddiman and McIntyre, 1973, 1976; Kellog, 1976) yet little is known of the changes in bottom water circulation (Ruddiman and Bowles, 1976). The importance of lowered sea levels to the increase in M_2 tidal velocity in the Celtic Sea area and its implications with regards to the formation of linear tidal sand ridges of considerable extent have been discussed by Pantin and Evans (1984); Belderson *et al.* (1986).

1.4.2 Glacial marine facies architecture

Boulton's (1990) comprehensive review of this subject and his unifying models relating facies architecture to eustatic/isostatic sea level change are very useful (fig.1.11). This sediment model is linked to the sea level model to predict patterns of marine sediment accumulation near to an ice sheet through a glacial cycle in inner shelf, outer shelf and continental slope environments.

Facies associations are controlled by:-

a) distance from the glacier and the diffusion of glacially-derived water masses and their suspended sediment (including berg transport) debris into nearby oceanic waters; fig.1.12.

b) the form and depth of the sea bed, and its position in relation to the continental margin.

It is suggested that the facies architecture encountered through a glacial cycle is the product of facies types changing in space and time, in response to changing glacier position and changing glacially controlled sea level.

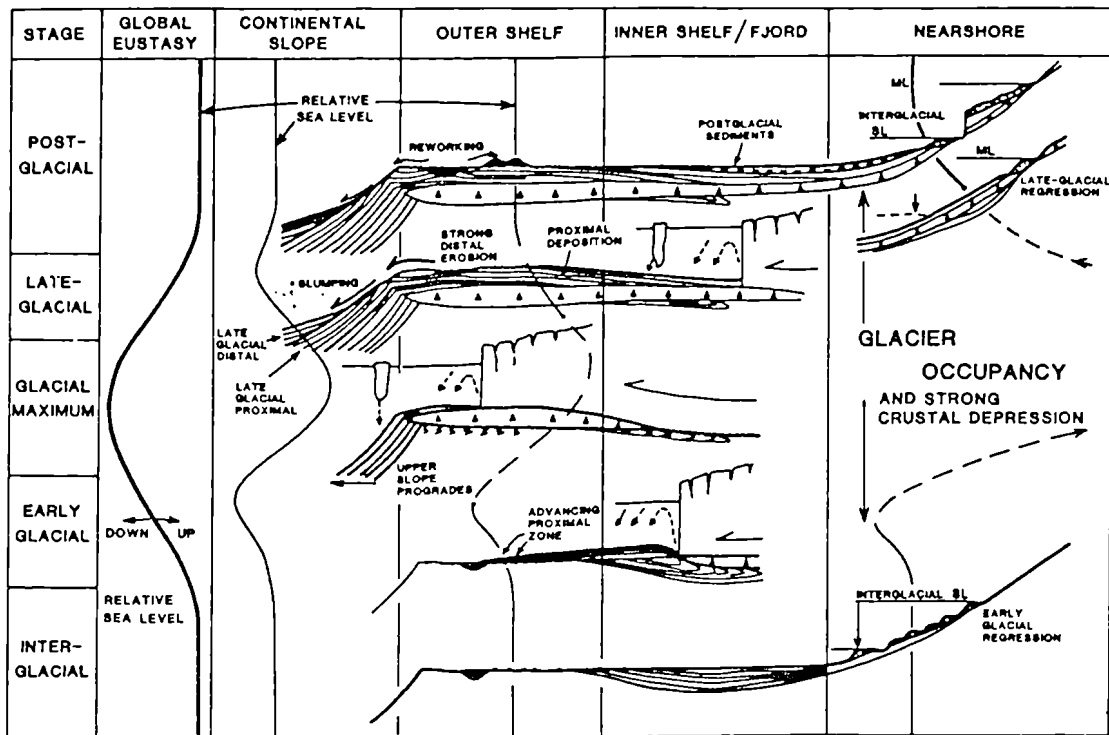


Fig.1.11 Boulton's (1990) model of glacial marine architecture through space and time - showing facies architecture from different continental margin locations through a whole glacial cycle. Appropriate relative sea levels are shown for each zone, from the nearshore to the continental slope.

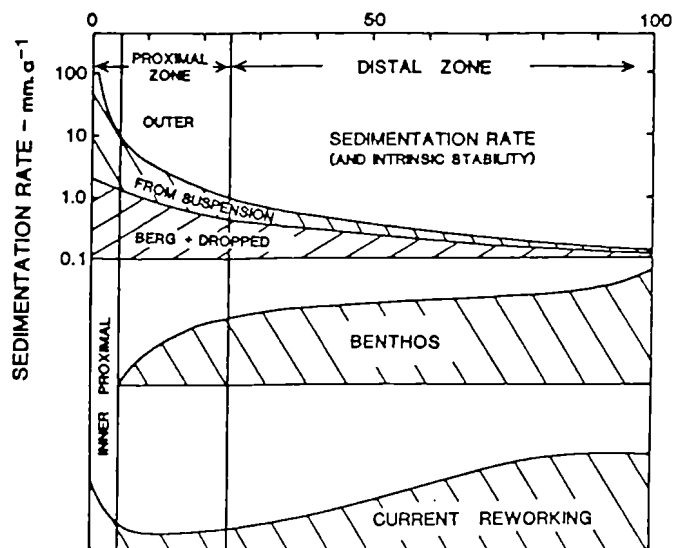


Fig.1.12 Schematic diagram showing changing sediment properties with distance (km.) from the glacier front (from Boulton, 1990).

In the case of an ice sheet (such as the British Late Devensian) whose deglaciation leads global eustasy, it is possible to model the changes in relative sea level and facies architecture. Thus, the models proposed by Eyles & McCabe (1991) of the role of glacio-isostatic disequilibrium are significant and not directly related to those of Boulton (1990) since they propose that the patterns of sedimentation are not necessarily climatically driven and regionally synchronous eg. Eyles & McCabe (1989) proposes a glacio-sedimentary model which involves rapid ice retreat and related sediment action triggered by rising relative sea level - suggesting isostatic downwarping is an important mechanism for deglaciating continental shelves.

1.5 Palaeoecology

1.5.1 Principles of palaeoecology

Perhaps the greatest single advantage that the Quaternary micropalaentologist has over workers concerned with older geological ages is that many of the species in the fossil assemblages studied are largely extant; a fact which Charles Lyell, in his classic subdivision of the Tertiary, noted in fossil molluscan assemblages and which has been born out as true in most subsequent studies of the Quaternary. With this fact in mind, and applying the principles of Uniformitarianism as expounded by James Hutton, it is possible to reconstruct past environments by studying the modern ecological requirements of the species found in Quaternary strata: this is the basis for palaeoecological reconstruction.

Generally, we are concerned with the external factors to which organisms respond (Autecology), rather than the somewhat more complex systems by which organisms interact (Synecology). Some of the most important external factors are: light, temperature, oxygen concentration, salinity, currents and tides, food supply, substrate, depth and so on. In the glacial marine context the most immediate external factor is temperature, although factors such as turbidity, salinity and substrate type may be equally important. Therefore, while there are factors other

than temperature which determine the distribution of microfossils, it is generally this factor which has received the greatest attention. Hutchins (1947) proposed that species were controlled by temperature in two ways:

(a) Survival temperatures

(b) Reproduction and repopulation temperatures

The temperature range of (b) will always be narrower than (a). Two temperature tolerance groups can be recognised: some species tolerate only a narrow range, be it warm or cold and are termed Stenothermic species; while others are less demanding and are termed Eurythermic, although even these species thrive at a species optimum temperature. A simple subdivision of the Stenothermic species is into Thermophilic (warmth loving) and Cryophilic (cold loving), although as will be demonstrated below, more elaborate temperature preferences have been proposed.

In colder water environments the energy requirements of an organism are increased as its heat loss increases, and heat loss is related to the surface area to body volume ratio. For this reason, the individuals of a species are often larger in colder than in warmer climates (Bergmann's Rule); however, such generalizations may not necessarily apply to microfossil groups, particularly foraminifera. In a similar way, it has been noted that species diversity drops as temperatures drop.

1.5.2 Historical development of faunal provinces

One of the earliest works which defined modern biogeographic provinces was that of Milne-Edwards (1838), he defined four faunal provinces from Western Europe: Polar (north of 71°N), Scandinavian (58°N to 71°N), Celtic (41°N to 58°N) and Mediterranean (south of 41°N). However, the classic early work is that of Dana (1853a, b) who proposed a world wide scheme of biogeographic Kingdoms, provinces and climatic zones. The provinces include Arctic (north of 71°N), Norwegian (62°N to 71°N), Caledonian (56°N to 62°N), Celtic (47°N to 56°N), Lusitanian (41°N to 47°N), and Mediterranean (south of 41°N). Dana's provinces were named after places and his climatic zones were formed from the adjectives torrid, temperate and frigid; therefore, from a nomenclatorial point of view his scheme is preferable to that of modern authors.

However, Dana's concept of biogeographic provinces differs from the generally accepted view of modern biogeographers (cf. Valentine, 1961). Dana based his scheme upon known temperature isocrymes (mostly at 6°F intervals) and his provinces are based upon temperature rather than the distribution of animals as a response to temperature. In most subsequent work it is the distribution of organisms that define the provinces and it is the boundaries between provinces upon which climatic zones are based (eg. Hazel, 1970). It is interesting to note that Dana's (1853a) climatic zones correspond very closely to those defined by Hall (1964) based upon Molluscan species limits along the east and west coast of North America; emphasizing the fact that temperature is the basic factor controlling the distribution of organisms, particularly within a marine context.

Hazel (1970) has illustrated the various faunal provinces which have been proposed for the North Atlantic. The reason for concentrating on west European faunal provinces is that as climatic zones have migrated northwards and southwards during interglacial/glacial cycles, then so too have the faunas migrated, depending upon their respective temperature requirements. Thus, while the shelf faunas we are interested in can migrate over these shallow shelf seas, it is unlikely that they might all be found on the East coast of North America, particularly the thermophilic taxa. Infact, a number of the ostracod species discussed by Hazel (1970) from the Atlantic continental shelf of North America are well known in European waters. However, all these amphiatlantic species occur in the Nova Scotian province and northern European waters (cf. Neale and Howe, 1975) and are therefore clearly eurythermal to some degree. Equally, a number of cryophilic species are known to inhabit both shallow & deep waters and are therefore more readily dispersed.

With the modern knowledge of faunal provinces it would seem logical to determine the modern temperature/depth/climatic zone range of as many species, in as much detail, as possible. Hutchins (1947) demonstrated how difficult it can be to determine these limits and the case of one species, *Baffinicythere emarginata* (Sars) is discussed below as an example. With regard to the distribution of ostracods, the most valuable work is Hazel (1970)

and the lack of any more recent work is notable. For the foraminifera, the recent work of Sejrup *et al.* (1981), Ostby and Nagy (1982), Hald and Vorren (1984) and Qvale *et al.* (1984) are all useful. Recently, Murray's (1991) publication on the "ecology and palaeoecology of benthic foraminifera" provides a regional synthesis of foraminiferal distribution world-wide and his comments on the distribution patterns and the ecological controls acting are useful. Taxonomic determinations are critical to correct palaeoecological reconstructions and as Miller *et al.* (1982) point out: "together with an uniform and reliable taxonomy the modern distribution of species allows the interpretation of Quaternary stratigraphy".

1.5.3 Ostracod zoogeography (Hazel, 1970)

This important work sought to establish a recent datum for the Atlantic coast at a time when knowledge of ostracod distribution patterns, particularly along the coasts of the United States was limited (Hazel, 1967a). Of particular interest are the data concerning the distribution and tolerances with respect to provinces, climatic zones, depth, and temperature of a number of amphiatlantic ostracod species.

Although the present study is largely concerned with foraminifera. I have chosen *B.emarginata* (Sars) to illustrate the scope of this work (Hazel, 1970), not because it is a particularly useful indicator species of glacial marine environments, but because it illustrates rather well some of the controls temperature has on species distribution. This species occurs in the Arctic province and thus tolerates temperatures below 0°C, it also occurs in more southerly faunal provinces eg. Narragansett Bay, Cape Cod (Williams, 1966) in areas where bottom temperatures vary between a maximum of 20°C and a minimum of 4°C (Hicks, 1959); while in Europe, Elofson (1941) gave this species a temperature range of <0°C to 19°C ie. approximately the same on both sides of the Atlantic. The question that arises therefore is whether or not the southern limit of the species is determined by the highest summer temperatures it can tolerate or by the lowest winter temperatures that it requires for reproduction (Type 3 or Type 2 of Hutchins, 1947). If *B.emarginata* were of Type 3 then

theoretically it might occur as far south as the Bay of Biscay (Schroeder, 1966). However, its most southerly recorded limit is at a depth of 50 m. in Loch Fyne, Scotland (Robertson, 1875) where the average August surface temperature is 14°C (no bottom temperatures quoted?) and it would therefore appear that the distribution of *B.emarginata* is of Type 2 ie. its southern limit is controlled by the low winter temperatures needed for reproduction.

1.5.4 The Stratigraphy of Quaternary deposits

The first work which attempted to interpret faunal changes through vertical successions (ie. through time) was that of Brady, Crosskey and Robertson (1874). Fossil ostracod assemblages from the base of the Clyde Beds, Glasgow area were observed to be quite distinct from those of the adjacent, modern day sea area. They demonstrated that their fauna coincided with some of the 'Arctic Shell Beds' of James Smith (1839) and this was the first microfaunal evidence of former Arctic sea water temperatures around the coast of the British Isles. Three of their most common indicator species are *Krithe glacialis* (Brady, Crosskey & Robertson), *Cytheropteron montrosiensis* (Brady, Crosskey & Robertson), and *Rabulimys mirabilis* (Brady), all of which are found alive and breeding today no further south than the Barents sea or the fjords of eastern Greenland north of 76°N (Hazel, 1967, 1970).

Further work in this area includes Peacock *et al.* (1977, 1978a,b) in which a trend from the late Devensian glaciation is recorded as climatic amelioration occurred followed by cold climate reversal of the Loch Lomond Stadial which is clearly seen in B. G. S. borehole 71/9 off Colonsay (Chester *et al.*, 1972, fig.c). All the classic features of the glacial marine environment are visible: a strong decline in the species diversity is associated with increasing numbers of Arctic indicator species eg. *K. glacialis*, *H. sorbyana*, *C. dimlingtonensis* and the presence of drop-stones. Although Robinson (1980) does not mention it, it is also likely that the relative abundance also fell as the number of species dropped in B. G. S. borehole 71/9, a feature which has been noted elsewhere (Feyling-Hanssen, 1964a; Knudsen, 1971;

Osterman, 1983) and which might be a function of higher sedimentation rates.

1.5.5 Late Quaternary Foraminifera.

There has been a considerable amount of work undertaken on the ecology, distribution, taxonomy and stratigraphy of foraminifera, and to a lesser degree ostracods, since the work of Brady *et al.* (1874) and there are far too many to deal with here. A few papers are briefly reviewed to give some idea of the scope of work being undertaken and of the interpretations being made.

1.5.5.a Sejrup *et al.* (1987) - an excellent example of a multidisciplinary study and of particular interest is the proposal that the Fladen Ground area of the northern North Sea was not glaciated during the Late Weichselian. The presence of the shallow water, Boreal-Lusitanian species *Ammonia batavus* (Hofker) in their lithozone D was at first thought to indicate an interglacial environment of deposition as noted by Knudsen (1985) elsewhere in the North Sea. However, the low number of species and individuals, their poor preservation and the fact that mixed arctic/warm taxa were co-occurring suggested that lithozone D was not of glacial marine origin but more likely a basal till with an incorporated reworked fauna; thus the faunal content resolved that particular problem. This illustrates a problem which one should always be aware of when dealing with diamicts ie. deciding upon the depositional origin, remembering that a basal till can often contain a reworked fauna. It is when the reworked fauna is of glacial marine origin, as proposed for the Aberdaron section (chapter 4), that the interpretation becomes problematic. This paper also contains some useful reviews on species distributions.

1.5.5.b Osterman (1983) - on the distribution of foraminifera in Frobisher Bay, Canada. The importance of *Elphidium excavatum* forma *clavata* in glacial marine environments is discussed (cf. Vilks *et al.*, 1979; Miller *et al.*, 1981; Vilks, 1976; Feyling-Hanssen, 1976; Knudsen, 1976; Nagy, 1965; and Adams & Framton, 1965) and the importance of ice in its distribution proposed. Three facies types are recognized: proximal, distal, and extreme distal and

these are related to the faunal assemblages illustrated. Peaks in foraminiferal abundance can occur stratigraphically above the proximal facies as an ice sheet retreats as well as in the extreme distal facies; these are termed peak abundance biofacies. It is interesting to note that while the numbers of foraminifera per gram of sediment increases, the actual diversity remains low in the proximal facies while in the distal facies diversity also increases. It is proposed that these changes are brought about by upwelling near the glacier front and that this is concurrent with increased productivity; the changes in abundance in the proximal facies could also be due to a fall in the sedimentation rate?

1.5.5.c Vorren *et al.* (1984) - emphasise the need to establish the relationship between fauna and sediment. They also illustrate their data by means of ternary diagrams showing the distribution of the Boreal, Arctic and Cosmopolitan species (based upon the scheme proposed by Feyling-Hanssen, 1955).

1.5.5.d Hald and Vorren (1987) - define six foraminiferal assemblage zones from the Late Weichselian of the continental shelf off northern Norway. The paper contains useful discussion on foraminiferal ecology and also their reworking within a glacial context; of particular interest are their views on the ecology of *E. excavatum*, which they consider to be as much an indicator of salinity as of temperature on the basis of its known occurrence in lower latitudes in environments with changing salinities. The implications to the ecology of this most ubiquitous of Quaternary species are considerable and illustrate the importance of factors other than temperature upon the distribution of any species.

1.5.5.e Feyling-Hanssen (1964) - this paper on the foraminifera of Late Quaternary deposits from the Oslofjord area is a milestone amongst publications on Quaternary foraminiferal stratigraphy. The detailed faunal responses to changing environmental conditions are charted to reveal differences in the area during the Lateglacial and Holocene period. Particularly useful are the systematic descriptions and light microscope figures of the species present.

1.5.5.f Feyling-Hanssen *et al.* (1971) - this again is an important work, dealing with the Late Quaternary foraminifera from Vendsyssel, Denmark and Sandnes, Norway. Useful palaeo-environmental reconstructions are included, detailing Lateglacial and Postglacial conditions in particular. The systematic section, edited by Knudsen, is a reworking of Feyling-Hanssen (1964) but with some important modifications; this forms the basis of much of the of the systematic work, particularly from Norden, in recent years.

Chapter 2. Methods

2.1 Site Selection

The process of site selection was largely determined by the need to satisfy the criteria relating to the specific aims of the project. Thus, at Aberdaron, for example, the site was selected on the grounds that the cliff section was well exposed, easily accessible and was proving controversial in terms of the depositional models which have been proposed for the Irish Sea Basin (D. McCarroll, pers. comm., 1989). A sampling strategy was therefore devised that allowed me access to a series of horizontally displaced, vertical sections which recovered samples from facies displaced in space and time. The criteria of site selection did not apply in the same way at the other two study areas since only a limited number of cores and samples were available. The details relating to all these sites are included in the relevant chapters.

2.2 Sample Location

The problems associated with sample location from the cliff section at Aberdaron were minimal. All the samples were numbered consecutively, were measured in meters east of a datum on the beach, and were located in relative height above the beach.

Navigation over the Hebridean Shelf during the BGS sampling programme of the area relied solely upon the Decca Mainchain system. The continuous wave Decca Hebridean (8E) chain, operating at 68% probability levels up to $10^{\circ}00'W$, is defined as providing a full daylight fix repeatability accuracy of less than 100 m. The poorest fix repeatability accuracy occurs on summer and winter nights and is considered to be less than 240 m. and less than 330 m. respectively. The sampling programme was carried out during the Autumn of 1985 and at this time of year fix errors should lie somewhere between the above values. However, Selby (1989) reports that comparison of these data with results obtained by satellite positioning suggest that the precision is probably less than 500 m. The details of sample location from the southern Celtic sea are discussed by Pantin and Evans (1984).

2.3 Sampling Procedure

The onshore sampling from the cliff at Aberdaron involved cleaning the section faces to clarify the stratigraphy and to avoid sample contamination. Relatively large samples, varying between 526.6 g. and 820.7 g., were recovered and removed in polythene bags together with details of the sample location.

Offshore sampling was undertaken using the BGS Mk.II electronic vibrocore (fig.2.1) which generally gives good recovery in most sediment types and provides a core with an 83 mm. diameter. The system is highly sophisticated and a transponder on the vibrocorer allows the ship's dynamic positioning system to hold its position overhead. Coring times are generally between 5 and 15 minutes, depending on the lithology. A wireline controlled retraction system which allows the barrel to be returned to the housing frame prior to hoisting from the sea bed leads to a controlled extraction rate and improved sample recovery. Cores of up to 6 m. may be recovered in soft mud. Details of the system are outlined by Weaver and Schultheiss (1990).



Fig.2.1 The BGS Mk.II vibrocorer on deck British Magnus, 1985

Immediately after recovery onboard ship, Selby (1989) reports that the cores were cut into 1 m. lengths and the ends capped and waxed to prevent dessication. The cores together with Shipek grab samples, were labelled numerically in chronological order of recovery and with a reference to the particular grid square coordinates. For example, VE 57/-09/89 comes from the grid square defined at its south west corner by latitude $57^{\circ}00'N$ and longitude $9^{\circ}00'W$ and is the 89th sample recovered during the survey; VE denotes that it is a vibrocore. Each 1 m. section is labelled according to its depth beneath the surface and as a part of the total number of sections making up the core; ^c/6 therefore indicates that this section is one of six and that it represents the depth between 2 m. and 3 m. beneath the surface.

Brief, shipboard descriptions were carried out by B.G.S. staff and consisted of the analysis of small samples from the core ends immediately after sectioning. The core casings were split in the laboratories at the BGS, Keyworth using a router; this results in minimal internal disturbance. The cores were then cut using an electro-osmotic Knife. This works by inserting two electrode rods into either side of the proposed line of section and then cutting the core with a large, flat bladed knife which is also connected to the same low voltage, high current (a.c.) supply. The net effect is that water is drawn towards the knife and plane of section, thus reducing the drag on the blade and hence the amount of smearing and obscured sedimentological structure. This technique is reported to be most valuable in soft muds with a high water content.

A number of the split cores were described and sampled by Selby (Hebridean Shelf) and Scourie (southern Celtic Sea). Geotechnical tests performed on the cores included a hand shear vane and dropcone penetrometer. The cores were then sealed in polythene tubes and it is in this state that they were made available to me; only one vibrocore, VE 49/-09/44, was available from the southern Celtic Sea.

2.4 Sample Processing

Sub-samples from the vibrocores were removed from known depths, typically 5 cm. or 10 cm. slices, but sometimes as little as 2 cm., and placed in labelled evaporating dishes. Care should be taken to avoid sampling from around the casing which may be contaminated with smeared sediment. There then follows a series of standard processing steps (Table 2.1) which are intended to clean and concentrate the foraminifera present.

Step 1	dry sediment sample @ 40°C for 24 hours	
Step 2	weigh sediment sample = dry weight	W1
Step 3	rehydrate with distilled water for 24 hours	
Step 4	wet sieve @ >63µm.	
Step 5	dry >63µm. residue @ 40°C for 24 hours	
Step 6	weigh dry coarse residue = residue weight	W2
Step 7	heavy liquid separation	

Table 2.1 Standard sample processing steps.

2.4.1 Heating

Wet sediment weights were not employed and were considered to be unreliable, reflecting the fact that some of the vibrocores inspected (and sampled) were not adequately sealed and had begun to 'dry-out' during storage at the BGS core-store, Keyworth. The samples were heated in an oven at 40°C for 24 hours before a sediment dry weight was established. Thus, all faunal data can be related back to a depth in the vibrocore and to a dry weight of sediment.

The temperature for drying the samples is kept at the relatively low value of 40°C to minimize the damage to the foraminifera; some agglutinating foraminifera are prone to disaggregation at high temperatures. Furthermore, temperatures higher than 60°C are considered to affect amino acid compositions (H. P. Sejrup, pers. comm.) and it was therefore advisable to use the lower temperature throughout the study.

2.4.2 Washing

Many of the sediment samples analysed were diamicts or muddy sands and required considerable amounts of washing time to remove the clay and silt fractions. The procedure adopted was to rehydrate the dry sediment samples in distilled water overnight and then to wash it through a 63 μm . steel sieve until the water passing through turned clear. Various methods to speed-up the washing process, including dispersants, were tested but proved unsatisfactory. A soft-bristled brush was used to encourage the particularly stubborn samples but its use was kept to a minimum for fear of damaging the foram. tests.

Some authors recommend the use of a 1 mm. mesh sieve over the 63 μm . sieve to reduce the amount of abrasion by larger clasts. This was not adopted, but is considered to be an advisable step, particularly when dealing with clast-rich diamicts. The use of sieve size is demonstrated to be critical to the faunal results obtained in this study and is discussed below as well as by Brolsma (1978a) and Schröder *et al.* (1987).

2.4.3 Weighing

All sediment weights were recorded to two decimal places after heating at 40°C overnight. Two weights are obtained during this sample processing. The first, W1, relates to the weight of dry sediment taken for the analysis, normally 100 g. The second, W2, relates to the weight of the dry coarse (>63 μm .) fraction remaining after sieving. It is then possible to calculate the weight of the fine (<63 μm .) fraction, W3, which has been lost. Both W2 and W3 can now be expressed as a percentage weight of the original dry sediment weight, W1, and are expressed as follows:

$$\% W2 = \frac{W2}{W1} \times 100$$

$$\% W3 = \frac{W3}{W1} \times 100 \quad \text{or} \quad \frac{W1-W2}{W1} \times 100$$

It is useful to plot these values, particularly W2, stratigraphically and the data provide a useful summary of the changing lithological character of the sediments. In a similar

manner, it is possible to further dry sieve the coarse fraction into a number of constituent grain size fractions.

2.4.4 Heavy Liquid Separation

The dry residues, often containing a considerable amount of sand-sized material, require a process which will concentrate the foraminifera present without the need for picking and counting from large volumes of sediment. The technique employed here is to 'float-off' the foraminifera using a heavy liquid, in this case carbon tetrachloride, based upon the methods described by Feyling-Hanssen (1958) and Feyling-Hanssen *et al.* (1971). However, the latter employed a 100 μm . sieve rather than the 63 μm . sieve size used in the present study. The steps involved in heavy liquid separation are as follows:

step 1

Half fill a labelled glass bowl with CCl_4

step 2

Pour the dry, coarse residue (>63 μm .) slowly over the surface

step 3

Place a labelled filter paper in a funnel over the bottle of CCl_4

step 4

Decant the liquid from the bowl into the filter funnel

step 5

Wash down the sides of the bowl and $1/2$ fill with CCl_4

step 6

Repeat steps 4 & 5 until no further 'scum' is visible

step 7

Air dry the filter paper and place the contents in a glass bottle

N.B. The entire heavy liquid separation process must be performed in a fume cupboard.

This technique, which produces a 'light fraction', enriched in foraminifera, and a heavy residue, has the major advantage of speeding-up the otherwise laborious picking process. Picking from the various dry sieved fractions of the initial coarse residue and relating counts obtained from each of these fractions can be very time consuming. However, there are problems associated with this method which are discussed in detail elsewhere in this volume.

2.5 Foraminiferal Counting

A rectangular brass counting tray with 45 square divisions and a series of holes through it was used together with a mounted steel pin for picking. Counted specimens were dropped onto labelled faunal slides located within the field of view of the microscope, beneath the tray.

The 'light' fraction was spread evenly and as thinly as possible over the counting tray and the weight of this fraction before and after spreading gave an indication of the proportion of the total 'light' fraction present on the picking tray; this proportion was recorded on the counting sheet. Where there was insufficient 'light' fraction to cover the entire tray, the number of squares covered were counted and recorded.

Counting was undertaken on a square at a time basis, so that all the specimens from one square were counted before starting on another. In this way total counts of exactly 300 specimens, or any other number, can not be obtained for all the samples under investigation. The squares were selected on a random basis, although at least one perimeter square was included with every other square in order to ensure that larger, rotund specimens, which have a tendency to accumulate around the edge of the tray, were sampled. Every second square is designated to the perimeter because the picking tray is arranged as a matrix of 5x9 and therefore has 24 perimeter squares out of a total of 45. The significance of various counting approaches are reviewed by Brolsma (1978) and are discussed further in Chapter 6.

2.5.1 Count Numbers

The number of specimens which must be counted from a sample to provide an accurate and reproducible assesment of the faunal composition is a matter of debate. According to Murray (1991, p.316) "it is a matter of observation that when 250 or more individuals are counted, the relative proportions of the component species are reasonably constant". He argues that there is little point counting larger numbers if there is no gain in accuracy, but fails to mention that rare forms are more likely to be encountered. However, Pielou (1979) has argued that even 300 specimens is an insufficient number to count.

In the present study sample counts of >300 benthonic specimens have been used wherever possible and are typically of about 400 specimens. Planktonic foraminifera, apart from *Neoglobobadrina pachyderma*, remain taxonomically undifferentiated and are counted as they co-occur with the benthonic species on the picking tray. This provides a useful planktonic:benthonic ratio for each sample counted. Additional benthonic species, not included in the faunal count, but present on the picking tray, in squares not sampled, were also recorded and dropped into the faunal slides. These additional taxa are not employed in any numerical sense, but simply act as a qualitative record of the rarer species present in each sample.

2.5.2 Foraminiferal Sums

Estimating specimen numbers per unit of dry sediment weight (DSW) is possible, following the above procedures, but prone to inherrent errors of repeatability. The first assumption we make is that the 'light' fraction represents the entire (100%) faunal content of the sample under investigation. This is not true and experimentation (this study, see section 6.2.1) suggests that the heavy liquid floatation of the foraminifera is, at its best, 90% effective. Thus, even before beginning these calculations it is likely that the final benthonic sum is underestimated by at least 10%. Furthermore, there are clearly some statistical considerations relating to the repeatability of these procedures which will act to increase the possible margin of error on any

given value, and since the foram. sum is the product of these values, errors are likely to be large.

The whole procedure is probably best illustrated by taking one of the samples analysed eg. VE 57/-09/89 2.13-2.15 m.

Weight of dry sediment sample (W1) = 28.6 g

Weight of dry coarse (>63 µm) residue (W2) = 10.0 g

counted fauna, benthonic specimens (Sb) = 400

counted fauna, planktonic specimens (Sp) = 47

squares counted (Q) = 4/45

proportion of 'light' fraction represented on tray (L) = 1/2

Therefore number of benthonic specimens per 100 g dry sediment weight = Sb :

$$Sb \cdot \frac{1}{Q} \cdot \frac{1}{L} \cdot \frac{100}{W1} = 400 \cdot \frac{45}{4} \cdot 2 \cdot \frac{100}{28.6} = 31,469 \text{ specimens/100 g}$$

Sp may be substituted for Sb to give the number of planktonic specimens per 100 g of sediment. The planktonic : benthonic ratio is obtained by dividing Sp/Sb.

2.5.3 Foraminiferal Size

The size of every individual species described is dealt with briefly in the systematics section (Appendix 1, Volume 2), and the implications of sieve size are discussed at length elsewhere in this volume. However, there are two groups which lend themselves remarkably well to size analysis. The first of these is the ubiquitous foram. *Elphidium excavatum*, which occurs in high frequencies in a large number of the samples analysed and are easily measured, maximum test and umbilical boss diameters were measured from a number of samples, particularly from the cliff section at Aberdaron and from vibrocore VE 57/-09/89 from the Hebridean Shelf. The second is the ostracod species *Rabulimilis mirabilis*, measurements of its carapace length and height were recorded from the southern Celtic Sea area.

The actual measurements are made with a calibrated eye-piece graticule at a magnification of x80 on a stereo-zoom binocular microscope and at this magnification ¹/2 graticule intervals allow

measurements of 6 μm .

2.6 Numerical Methods

Much of the count data has been processed using a hand-held calculator to determine values such as the foraminiferal sum, specific frequency, planktonic : benthonic ratio etc.. However, data from the Hebridean vibrocores has been processed on the D.E.C. Vax A computer, University College of North Wales, using the 'Fortran' language programmes 'POLLDATA 4' and 'ZONATION'. Further use of computer programmes include the statistical package 'MINITAB' and the graphics package 'HARVARD GRAPHICS'.

Other numerical methods employed include the Index of Affinity for comparing samples. This is calculated as the sum of the lower frequency of common taxa; ideally values of >75% are looked for as an indication of similar faunal composition (see Rogers, 1976).

Faunal diversity is based upon the method outlined by Walton (1964) and is defined as the number of species in a sample accounting for 95% of the fauna in that sample. Some of the samples, where count numbers varied considerably eg. southern Celtic Sea, were unsuitable for the faunal diversity measure of Walton (1964) and here the Fisher- α measure of diversity was employed (see Murray, 1973 for wider discussion).

2.7 Zonation

Foraminiferal diagrams when completed, are often large and complex, consisting of many levels and many taxa. Some means of simplifying the data is required, normally horizontal lines which separate zones of differing foraminiferal composition are employed. This aids in the description of the diagram, the discussion and interpretation of its contents, and in comparison and correlation with other diagrams. The basis of the zonation methods employed here are well established in the principles and methods of pollen analysis as discussed by Birks and Birks (1980).

The foraminiferal zonation employed here is based upon the concept of assemblage zones, and these zones are constructed entirely from the observed foraminiferal assemblage zones, which

includes: type locality and section; description of foraminifera in the zone; description of the contacts with other zones; thickness of the zone and age; name or number of the zone; general notes and any other occurrences.

Strictly, these zones are termed site or local zones; if they can be correlated with similar zones from other sites then they become regional foraminiferal zones. The concepts of stable and transitional, peak or acme zones are not dealt with here; while the question of 'patchiness' in foraminiferal distribution is dealt with later on in this volume.

There are a number of advantages to the numerical definition of foraminiferal zones :-

1. several independent methods are applied to the same data set and the numerically most consistent zonation adopted.

2. the zones are based only upon the observed foram. content and don't rely on preconceived ideas based upon sediments, inferred climate, presumed time equivalence etc.

3. any bias on the part of the investigator is limited to the selection of the taxa to be employed in the zonation.

4. the criteria of zonation are defined clearly in the methods employed.

Zonation can be thought of as a form of classification and since the data are quantitative and multivariate it is possible to employ numerical methods of classification. The three numerical techniques employed here are :-

2.7.1 The Constrained Single-link Clustering Method (Conslink) of Gordon and Birks (1972).

Here adjacent levels are combined in a strict stratigraphic order; the method is based upon measures of dissimilarity:

$$DC_{ij} = \sum_{k=1}^m | p_{k_i} - p_{k_j} |$$

where DC_{ij} is the dissimilarity coefficient between samples i and j .

p_{k_i} is the proportion of foram type k in sample i when there are $k = 1, 2, \dots, m$ foraminiferal types.

The computer begins by finding the two samples which are stratigraphically adjacent with the lowest dissimilarity coefficient, these two are then grouped together and the procedure is repeated until all the samples have been grouped together; the last two samples to be grouped will represent the most dissimilar adjacent levels within the core. The end result is to produce clusters of samples of similar foraminiferal composition and these can be viewed as foraminiferal assemblage zones.

One disadvantage of the Conslink method, particularly when applied to large data sets, is that it has difficulty discerning parts of the pollen sequence (but equally foraminifera) where numerically small but stratigraphically consistent differences exist (Gordon & Birks, 1972). It soon becomes apparent that a group defined by this method might encompass a wide range of variation.

2.7.2 Binary Divisive Procedures

An alternative approach to the single-link method is to define a global measure of variability within the data set which can be minimized by grouping similar objects (in this case samples); in this way the methods employed divide the diagram so that the total numerical information it contains is maximally reduced at each division while maintaining the stratigraphic order (useful for us and greatly reducing the number of allowable partitions). These methods are also hierarchical, thus allowing the diagram to be split more readily into zones and sub-zones. The programme SPLITLSQ is based upon the within-group sum-of-squares algorithm and is discussed at length by Birks and Gordon (1985). The third programme employed in the zonation of the diagrams is SPLITINF and also employs binary divisive procedures, dividing the data in terms of the total information content.

These numerical methods have been applied to the Hebridean Shelf of this study, specifically vibrocores VE 57/-09/89 and 57/-09/46. As far as I am aware, no previously published account of the application of these numerical techniques to benthonic foraminiferal data exists.

2.8 Data Presentation

The foraminiferal count data are presented in a manner which will aid interpretation. The foraminiferal diagrams included within the text are mostly summary diagrams, showing the major taxa and the most important faunal characteristics; more comprehensive, computer drawn (POLLDATA) diagrams, both percentage frequency and 95% probability, are included as separate enclosures in the document wallet at the back of this volume. While most of these diagrams express the relative proportions of the foraminiferal taxa as percentage frequency, some preliminary attempts to express the data as concentrations in the samples have also been undertaken and have yielded promising results (see section 3.9.3). Furthermore, the selection of specific foraminiferal sums eg. wall structure or boreal species, is possible and allows particular aspects of the fauna to be emphasized. The details of the selection criteria of foraminiferal taxa to construct these foraminiferal sums are dealt with as they arise in the text.

2.9 Dating Methods

2.9.1 Radiocarbon Dating

The dating technique employed here is accelerator mass spectrometry (a.m.s.) and a total of 9 molluscan shell samples from the Hebridean Shelf have been dated at the Oxford University Radiocarbon Accelerator Unit. Details of the sample stratigraphic levels are presented in chapter 3. This technique has the advantage over conventional radiocarbon dating, which measures C^{14} decay by beta transformation over a 'count period', that the concentration of C^{14} atoms themselves are detected. This means that smaller samples are required for dating and that bulked shell samples are no longer required; in fact it is now possible to date benthonic foraminifera (J. T. Andrews pers. comm.). For a general review of techniques see Lowe and Walker (1984).

These dates were awarded by the NERC radiocarbon committee (re. NERC RCL-SC419/0490) to Professor J. D. Peacock who picked and identified the molluscs, after submission of a detail

application. The foraminiferal stratigraphy presented here was included in the report that accompanied Professor Peacock's application.

All the samples analysed were marine shells (re. Mangerud, 1972) from sealed contexts which had been air dried and previously stored in plastic liners at the BGS corestore, Keyworth since their collection in October, 1985. The samples were prepared for dating by the laboratory following the procedures outlined by Gillespie *et al.* (1986). The dates are expressed in radiocarbon years before present (BP) and, as is usual, assume a half-life of 5,568 years. The reservoir corrected dates are based upon a marine reservoir correction factor of 405 ± 40 BP, which is subtracted from the radiocarbon accelerator dates.

2.9.2 Amino Acid Stratigraphy

Most of the samples analysed in this study were deposited during the Late Devensian Lateglacial or Holocene period and are therefore well within the limits of resolution of radiocarbon dating. However, in the case of the southern Celtic Sea glacial samples there remains the controversy (see chapter 5) as to which glacial period these deposits belong to. Unfortunately, no molluscan fragments were recovered which were suitable for radiocarbon dating (Scourse, pers. comm.) and so as part of my investigation of the microfaunas from these sediments, I proposed that it might be possible to resolve which glacial period, if not the exact age, to which these deposits belong.

This has been achieved by means of amino acid analysis of foraminifera from the southern Celtic Sea area and the correlation of the allo-isoleucine:isoleucine ratios obtained with those published from the North Sea area (Knudsen & Sejrup, 1988). Details of the 'amino-stratigraphy' technique and its various applications are outlined in Hare, Hoering & King (1978).

The amino acid analyses were performed by myself at the laboratories of Dr. Hans Petter Sejrup, Department of Geology, University of Bergen, Norway in April, 1990.

2.10 Scanning Electron Microscopy

Foraminiferal specimens were cleaned with a soft brush and distilled water, before being mounted onto aluminium stubs with a surface layer of undeveloped photographic film; this layer once wet provides sufficient adhesion to hold all but the largest of specimens. The stubs were subsequently sputter coated with Gold in an atmosphere of Argon using a Polaron Equipment Limited SEM coating unit E5000.

Coating thickness may vary slightly but approximately 750 Å is typical. Interferometric technique experiments are reported to demonstrate that the thickness of Gold coating sputtered in Argon gas can be calculated at 2.5 Kv and a target to specimen distance of 50 mm. according to :-

$$t_h = 7.5 \cdot I \cdot t \quad (\text{Å})$$

where: t = time in minutes

I = current in mA

t_h = thickness in Å

A Coating time of 5 minutes and a current of 20 mA was typical. The stubs were kept in a dessicator between use and indexed so that every specimen was related to the sample from which it derived.

Chapter 3: The Hebridean Shelf

3.1 Location

The area investigated consists of the Hebridean margin and specifically the continental shelf to the west of the Outer Hebrides. This area forms a part of the north west European passive continental margin.

The archipelago of the Outer Hebrides lie 50 Km. west of mainland N.W. Scotland and form a natural eastern limit to the area investigated ($7^{\circ}30'W$). Equally, the continental shelf-break forms a convenient western limit ($9^{\circ}00'W$), with the exception of a single vibrocore analysed from the slope. The northern limit can be assigned just to the north of the Island of St. Kilda ($58^{\circ}00'N$) and the southern limit roughly along the latitude of the islands of Barra ($56^{\circ}45'N$) at the southern end of the archipelago. The area studied conveniently falls within that of the St. Kilda sheet as defined by the British Geological Survey (B.G.S.) regional mapping programme, which has divided the British continental shelf and slope areas into sheets measuring 1 degree of latitude by 2 degrees longitude.

To the west of South Harris, at a distance of 90 Km. lie the St. Kilda group of islands; St. Kilda (Hirta) being the largest of the four at 4.5 Km. across. These islands can be taken to delimit the northern part of the continental shelf area under study. The shelf itself is divisible into three main areas: the inner (basement rock platform), middle (deeper shelf areas to the west of the rock platform, but east of the higher outer shelf), and outer zones (westwards to the shelfbreak). The large depression, the St. Kilda Basin, occupies the middle shelf and is bounded to the south by the broad east-west trending rise of the Otter Bank which extends westwards to within 20 Km. of the shelf break. To the west, the area is conveniently delimited by the shelf-break zone which lies at depths between -140 m. and -180 m. OD. Further west still are the deeper waters of the Rockall Trough.

3.2. Bathymetry

The bathymetry of the area is summarized in fig.3.1.a (from Jones *et al.*, 1986). Bathymetric data quoted in the text comes largely from Selby (1989) who obtained his information from the BGS, Admiralty Charts, Institute of Oceanographic Sciences and D. Sutherland. Selby reports that the main data source for the outer shelf is BGS cruise 84/06, obtained from an Atlas-Deso 20 echo sounder; however the latter is not tidally corrected nor adjusted for temperature/salinity variations. The bathymetry of the shelf area under study is summarized in fig.3.1.b, together with the location of the vibrocore sites.

The shelf slopes continually westwards towards the shelf-break between $57^{\circ}00'N$ and $58^{\circ}00'N$. Bathymetric controls in this area are the structure and lithology of the underlying bedrock. The complex bathymetry of the inner shelf, which in places extends over 50 km. west of the coastline of the Outer Hebrides, but more typically for about 30 km., is partly a product of Lewisian outcrop patterns. Further offshore, the generally poorly defined, low relief features of the St. Kilda Basin and Otter Bank contrast markedly with the St. Kilda plateau to the north. The shelf-break itself is defined according to Stanley *et al.* (1983) as the first major change in gradient on the outer shelf and occurs at depths of -140 m. to -160 m. between $57^{\circ}00'N$ and $58^{\circ}00'N$.

The bathymetric components of the shelf can therefore be divided into two: the rough, uneven, and westwards sloping Lewisian basement platform on the inner shelf, and the smoother and relatively flat outer shelf. The outer shelf is interrupted around St. Kilda by a circular plateau which forms an extension to the rocky inner shelf at depths between -80 m. and -40 m. OD. Some of the smaller islands off the western coasts of the Outer Hebrides equally suggest something of the irregular seabed relief of this area. The platform dips at approximately 1:300 towards the west.

There are two major features which characterize the middle to outer shelf; these are the St. Kilda Basin and Otter Bank. The Otter Bank is a broad, east-west aligned ridge at $57^{\circ}10'N$ and is

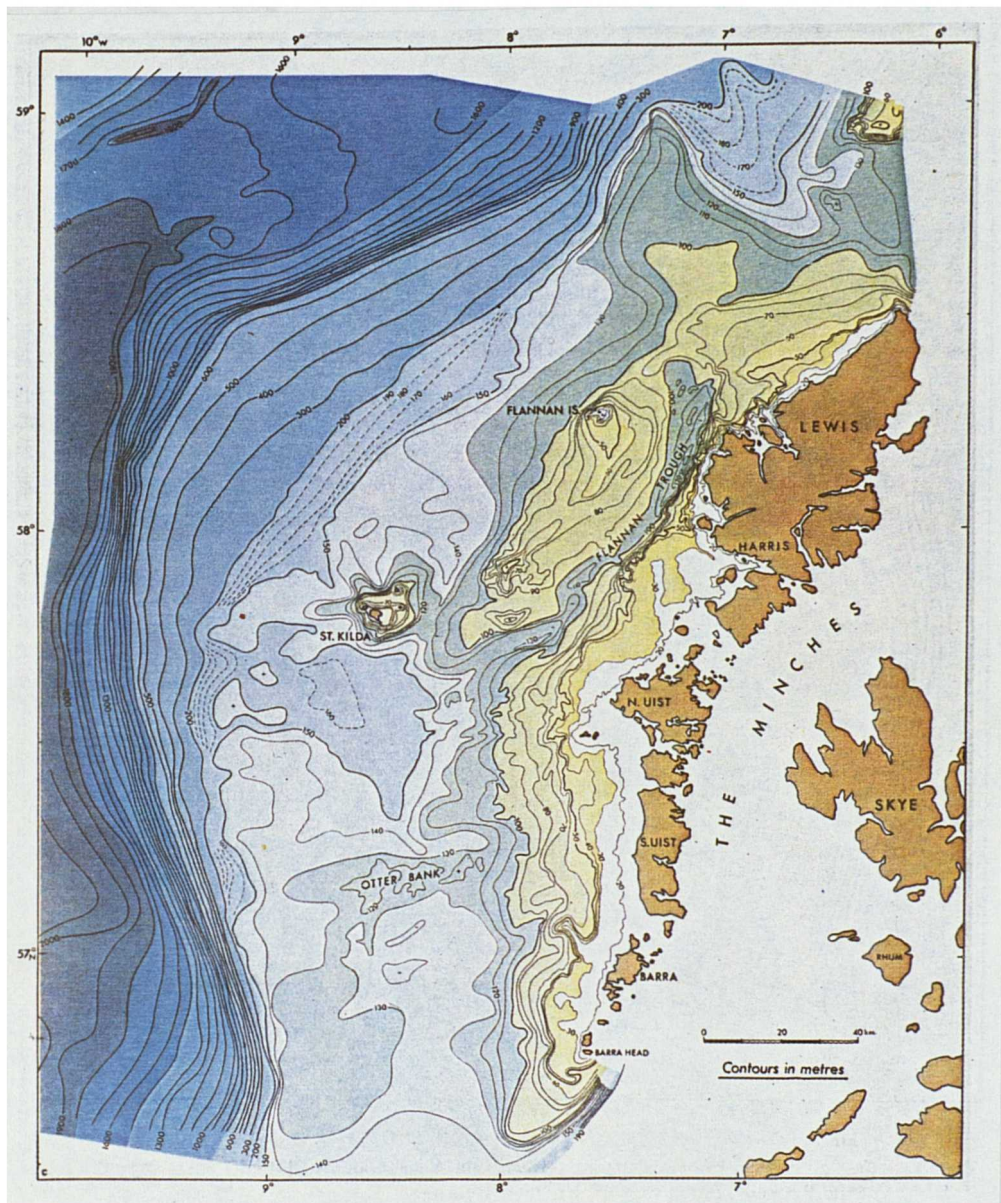


Fig.3.1.a Bathymetric map of the continental margin west of the Outer Hebrides (from Jones *et al.*, 1986).

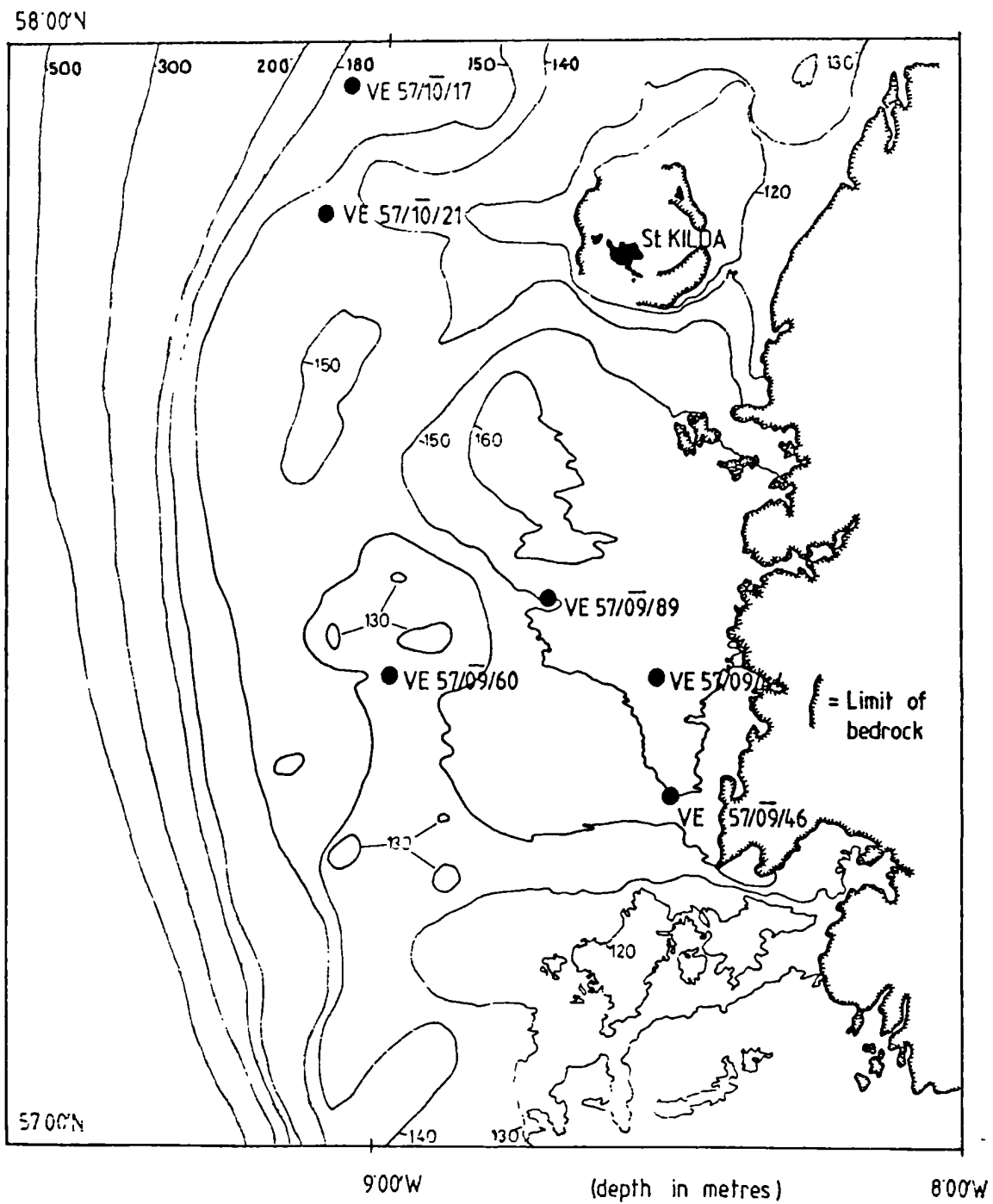


Fig.3.1.b. Summary bathymetric map of the Hebridean Shelf area under study and the location of the vibrocore sites.

over 12 km. across, 40 km. long and rises some 10 m. above the surrounding seabed to depths of between -120 m. and -130 m.. Two smaller ridges extend southwards for approximately 15 km. from it. The northern side slopes gradually towards the St. Kilda Basin, which is up to 30 km. across and over -160 m. OD at its deepest in the northwestern part. To the north it is bounded by the St. Kilda platform and to the west by a low outer shelf bank. This outer shelf bank rises up to a depth of -130 m. and effectively delimits the western boundary of the basin.

The St. Kilda plateau corresponds to the limits of an intrusive igneous complex and forms an abrupt feature on the otherwise relatively flat middle shelf. Bathymetric details of the plateau are given by Harding *et al.* (1984) and Sutherland (1984); although it basically consists of a complex upper surface lying at depths between -80 m. and -40 m.. Sutherland (1984) distinguishes two distinctive platforms at -40 m. and -70 m., both of which he considers to be the product of marine erosion. A "cliff-line" is reported from the western side of the platform, with a further marine eroded rock-cut platform at a depth of -120 m. to the west of it (Sutherland, 1984). No comparable features at a similar elevation are known to exist from elsewhere around the St. Kilda plateau.

3.3 Oceanography

The continental shelf water mass maintains a distinctive identity from that of the Rockall Trough due to restricted mixing brought about by well developed seasonal thermocline fronts and a well constrained along-slope current (Huthnance, 1986). In fact, the north west European continental shelf seas are distinguished by the formation of pronounced frontal systems during the summer months.

Seasonal variations in the strength and direction of residual currents on the St. Kilda shelf area have been reported by Booth and Ellett (1983), eg. mooring 'R' (57°00'N, 9°00'W; shelf depth -137 m. OD) where lower meter (-112 m. OD) summer mean filtered currents of $c.0.02 \text{ m.s}^{-1}$ flowing south east and northerly flowing

winter mean currents of over $c.0.04 \text{ m.s}^{-1}$ were recorded. Upper meter conditions (-40 m. OD) at the same mooring suggest similar flows but with slightly stronger summer currents of $c.0.30 \text{ m.s}^{-1}$. Large diurnal tidal currents over the shelf are considered to be barotropic and enhanced by a diurnal tidal shelf wave (Cartwright *et al.*, 1980). Hill and Simpson (1988) report that "mean currents over much of the continental shelf west of Scotland have a northward component" and interpret this as a result of a "pumping effect arising from the passage of successive depressions to the north of Britain". Coastal currents along western Scotland are unusual for the British Isles and persistent northward flowing currents of lowered salinity with velocities of between 0.03 and 0.1 m.s^{-1} are reported (Hill & Simpson, 1988) and are known to transport Irish Sea water, labelled with radioisotopes from Sellafield, in a narrow ($\sim 50 \text{ km.}$) plume within the shallow ($<100 \text{ m.}$) inner shelf region.

3.4 Geological setting

The Outer Hebrides are largely composed of Lewisian metamorphic rocks (gneisses) of Precambrian age. Caledonian deformation led to thrusting within the Lewisian complex. Later, during the Permo-Trias (?), regional continental extension led to the formation of sedimentary basins which continued to subside locally during the Tertiary and are today largely located off-shore. The opening of the Rockall Trough during the early Tertiary was accompanied by basaltic volcanic activity and subsequent westward progradation of the Hebridean Shelf.

3.4.1 The Lewisian

Much of the Outer Hebrides and shelf to the west consists of gneisses of the Lewisian Complex; these gneisses are generally homogeneous and undifferentiated onshore with some zones of "mashed" gneiss, together with ultrabasic and metabasic bodies of Scourian age (2200 Ma). Granitic intrusive complexes of Laxfordian age (1800 Ma) occur on west Lewis and Harris, while on south Harris and North Uist lithologically variable sequences of metasediments and metavolcanics outcrop, these sequences probably

extend offshore to the north west. As far as distinctive lithologies are concerned, the garnet-pyroxene rich gneiss, the Corodale Gneiss, is exposed on the eastern side of South Uist.

Laxfordian retrogressive metamorphic reworking produced an amphibolite facies mineralogy. The general tectonic trend associated with the Laxfordian deformation crosses the area in a S.E.- N.W. orientation; this is partly obscured by the superimposition of several phases of deformation (Watson, 1977). A major thrust, the Outer Isles Thrust, crops out along the eastern side of the archipelago, dipping at about 30° east; this feature is considered to be of Caledonian age (Brewer & Smythe, 1984). The Minch Fault, an associated structure, locally defines the eastern limit of the Lewisian basement.

3.4.2 Permo-Trias

To the west the Lewisian is overlain by Tertiary basalt and a westward thickening, prograding sedimentary succession; the present depth and seaward inclination of the underlying Lewisian basement on the outer shelf and slope is due to a combination of faulting and thermal subsidence of the Shelf (Jones, 1981).

During the Permo-Trias, continental extension led to initial rifting and the formation of half grabens in northern and north west Scotland (Dewey, 1981). Basins formed on Skye (Steel *et al.*, 1975), the west Shetland Shelf, the Minch, the Sea of the Hebrides (Naylor & Shannon, 1982) as probably did the Flannan and Barra Troughs of the Hebridean Shelf at this time (Selby, 1989). This crustal stretching was partly accommodated by the reactivation of older faults as listric normal faults (Brewer & Smythe, 1984) and resulted in the formation of these Permo-Triassic basins. The Rockall Trough, now a major feature with a considerable influence on the regional oceanography, probably began as a small graben at this time (Scrutton, 1986). The Stornoway Beds of Lewis also belong to the Permo-Trias and are red conglomerates with sandstones and siltstones, interpreted as alluvial fan mass flows and braided stream deposits (Steel & Wilson, 1975).

3.4.3 Tertiary

Tertiary igneous centres, for example the St. Kilda intrusive igneous complex, consists of gabbros, dolerites and granites (Harding *et al.*, 1984). Dykes associated with this igneous activity outcrop on the Outer Hebrides and are visible on seismic sections across some of the shelf basins (eg. Selby, 1989). On the middle-shelf and slope, overlying the Lewisian, are strongly reflective, seaward dipping reflectors (Mutter, 1985) which are interpreted as basalts and Jones *et al.* (1986a) have collected samples of these intrusive basalts with typical "within-plate" composition from west of Lewis.

3.4.4 Sedimentary units

The middle and outer shelf and slope west of the Outer Hebrides constitute a westerly prograding sedimentary succession together with the igneous intrusives and extrusives. Seismic stratigraphies (air gun records) reveal several major

Reflector	Intra-Quaternary?
---- H1 ----	Late Cenozoic
---- H2 ----	Oligocene & Late Eocene sediments.
---- H3 ----	Early Cenozoic volcanoclastic sediments (with lavas and intrusions).
---- H4 ----	Mesozoic and Late Palaeozoic? sediments with Cenozoic intrusions.
---- H5 ----	Seismic basement (largely Lewisian gneiss).

Table 3.1 Stratigraphic summary of the sedimentary sequence to the west of the Outer Hebrides (modified from Jones *et al.*, 1986b).

unconformities and the sedimentary succession is therefore interpreted as representing the record of thermal and isostatic subsidence (Boillot, 1978) modified by relative sea level changes.

Jones *et al.* (1986b) recognize 5 major seismic reflectors and interpret the sequence as outlined in Table 3.1, but with limited evidence and no borehole control. However, BGS data from the Hebridean margin between 57°00'N and 58°00'N is locally at variance with Jones *et al.* (1986b) according to Selby's (1989) interpretation of BGS cruise 84/06 data.

3.5 The Quaternary history of the Outer Hebrides

3.5.1 Seismostratigraphy

There is only sparse borehole control to any of the seismostratigraphic schemes proposed. The most important and "seismically sound" according to Selby (1989) are those of Davies *et al.* (1984) and Evans and McElvanney (in press) which point to the interaction of ice masses in key areas around the north of Lewis and the islands south of Barra during the Late Devensian. Davies *et al.* (1984) recognize two major glacial events with the latter glacial maximum reaching the shelf-break at some time during the Late Devensian. Selby's seismostratigraphy is summarized in fig.3.2 and for further details of this scheme, the reader is referred to Selby (1989).

3.5.2 Rock Platforms and Erosion

Offshore, two major submarine planation surfaces are recognized on the St. Kilda platform (Sutherland, 1984a) at between -40 m. and -80 m. OD and a lower surface, backed by cliff-like features at -120 m. OD. These features are attributed by Sutherland to erosion during periods of lower relative sea level; the lower surface is regarded as a composite feature, forming during periods of maximum northern hemisphere glaciation and hence lowest eustatic sea levels eg. Late Devensian. The upper planation surface is possibly a feature which formed during partial northern hemisphere glaciation eg. the Younger Dryas

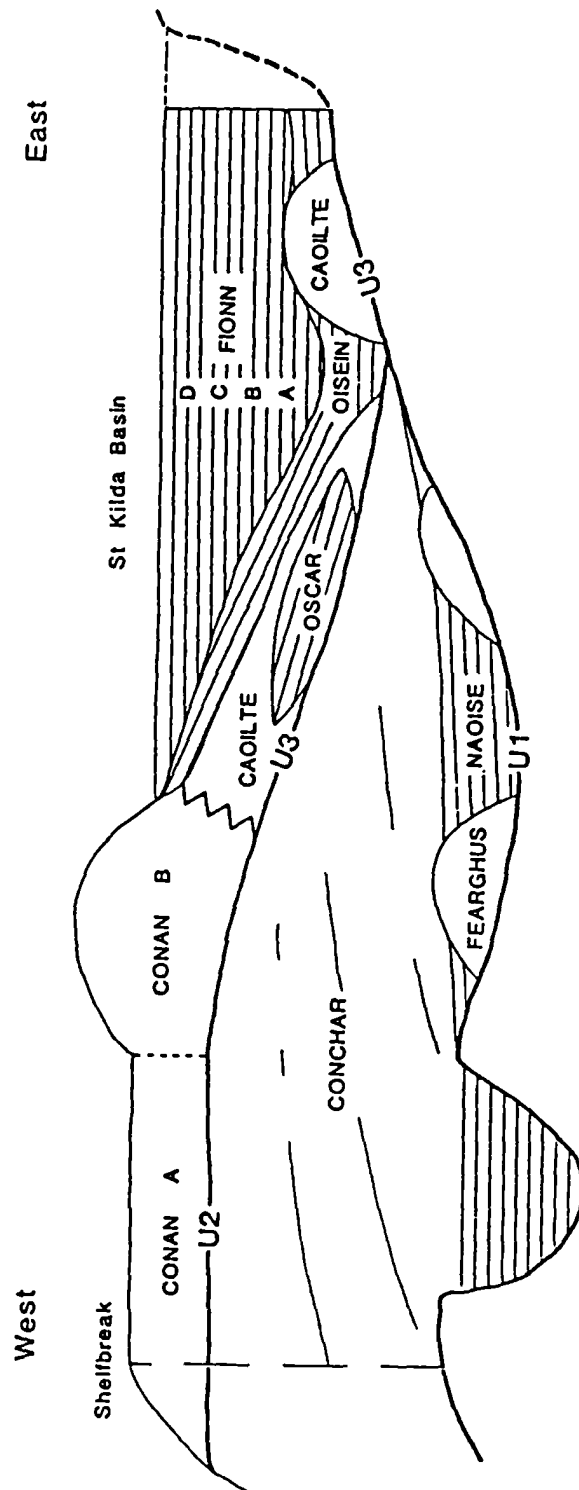


Fig.3.2 Diagrammatic interpretation of a seismic profile trending east-west across the middle-outer shelf at $57^{\circ}30'N$, showing the Quaternary seismic sequence of Selby (1989), the major reflectors (U) and the general attitude of seismic layering.

Stadial, or more probably the middle Devensian. However, these platforms tend to be uneven and highly irregular and are broken by rock ridges which reflect lithological variation.

Onshore, raised rock platforms occur in the coastal zones of the Outer Hebrides. An average platform height of 9.2 m. was established during a detailed regional mapping survey by von Weymarn (1974). For example, at South Galson, north west Lewis a rock platform is reported to be well developed and 150 m. wide, with a slope behind it up to 30 m. high and these are interpreted as denuded clifflines (von Weymarn, 1974). Earlier workers considered the rock platforms to be restricted to Lewis and Harris, although von Weymarn (1974) suggests that glacial erosion has erased the platform from many of these areas. Selby (1989) reports a bench visible on North Uist [NF 705, 724] lying at +6 m. local OD, which may imply a more extensive feature than is widely accepted. Typically, these rock platforms are overlain by beach gravels, although on Lewis, local surveys by Godard (1965) and McCann (1968) have demonstrated that platform and 'cap' deposit are not intimately associated.

The age of these platforms, if indeed they are attributable to a single phase of formation, is debatable. Sutherland and Walker (1984), for example, propose that a platform on north west Lewis was eroded prior to the Devensian on the basis of palynological evidence from the Toa Galson site. However, Peacock (1984) does not believe that these platforms were generated by long-term wave erosion alone and proposes an alternative mechanism of rapid formation under peri-glacial conditions which implies a composite nature to the formation of these rock platforms.

The evidence of the erosive action of ice is commonly visible all over the Outer Hebrides; although determining ice flow direction can be difficult. There are a number of examples of easterly ice-flow directions and it is apparent that Geikies (1873, 1878) proposal for a westerly flowing Scottish "mer de glacé" enveloping the islands is clearly not applicable. A summary of the ice flow directions, based upon work by Flinn (1978), von Weymarn (1979), Peacock (1984) and Selby (1989), is illustrated in fig.3.3. Although glacial striae and other features may be partly

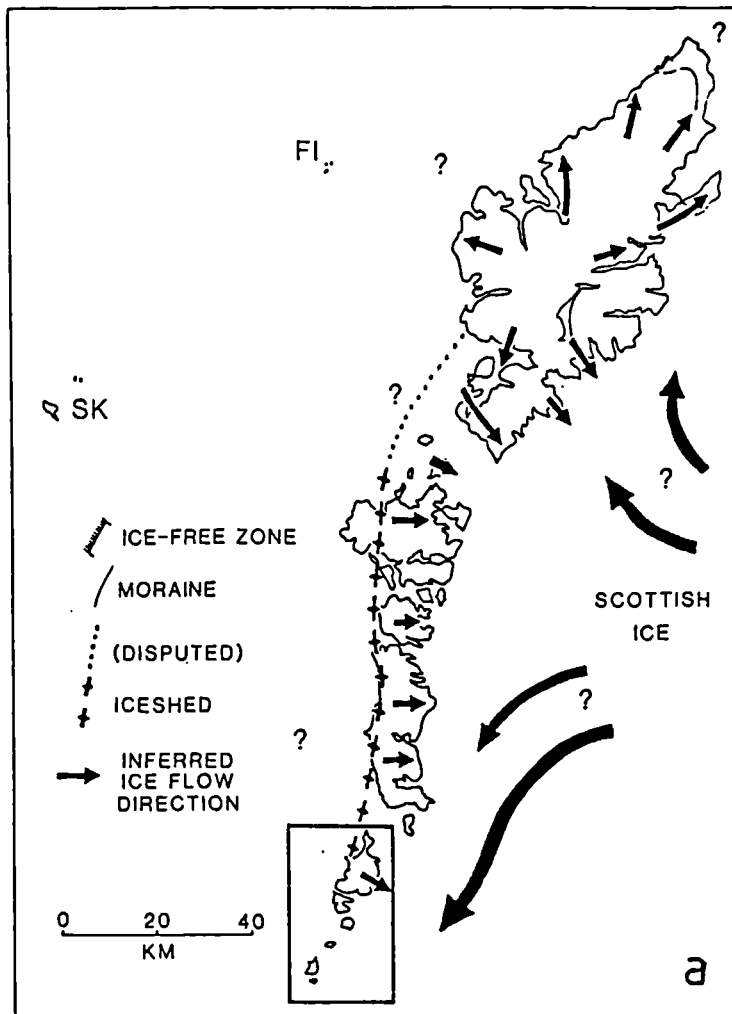


Fig.3.3.a Proposed ice flow directions across the Outer Hebrides during the last glacial maximum.

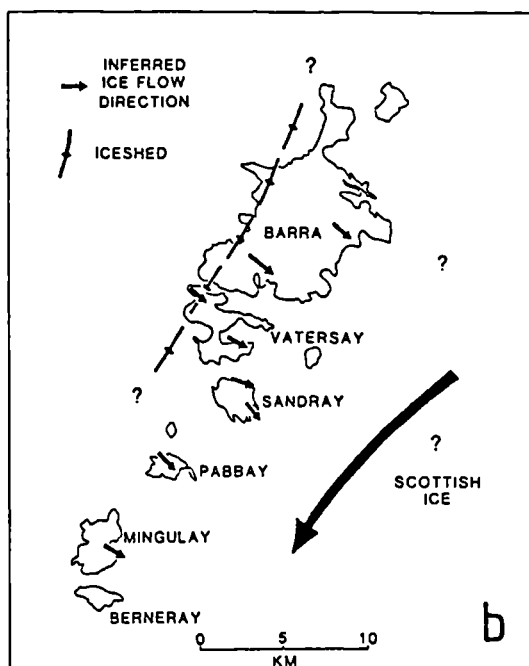


Fig.3.3.b. Inset= inferred ice-flow directions across Barra and the southern islands of the archipelago during the same period (both figs. from Selby, 1989 - based on references in text).

modified by late, minor glacial re-advances (= "valley glacier phase " - Peacock, 1984), these persist and are indicative of ice transport towards the east over much of the Outer Hebrides. It was Peacock (1984) who developed the concept of a separate Hebridean ice mass, demonstrating local ice thicknesses of at least 383 m. on Heaval, Barra; a model which Selby (1989) has developed further.

3.5.3 Glacial Erratics

The distribution of erratics on the Outer Hebrides were originally discussed by Geikie (1873, 1878) who envisaged a Scottish "mer de glace" being deflected in the Minch and diverting around the northern and southern extremes of the archipelago, which is where the erratics are largely distributed. Later work by Jehu and Craig (1927, 1934) appeared to support this view, but more recent work by Coward (1977) on South Uist has demonstrated the transportation of distinctive gneisses, which outcrop to the west of the Outer Isles Thrust, eastwards. It therefore seems likely, at least during the last ice movement and period of deglaciation, that ice flowed eastwards.

There is also some evidence to support Geikie's original views. The occurrence of red sandstone and other lithologies characteristic of the mainland in deposits on the Outer Hebrides and St. Kilda suggests that Scottish ice crossed all or at least parts of the islands at least once. However, Sutherland and Walker (1984) have suggested that the red sandstones of northern Lewis may be Mesozoic in age, rather than derived from the Torridonian as is commonly assumed. Unfortunately, no systematic work on these remarkably distributed erratics, concentrated at the northern and southern ends of the archipelago as they are, has yet been undertaken. Any such petrological study will face the inevitable problems which arise from the derived nature of the Torridonian and Permo-Triassic deposits.

3.5.4 Biostratigraphy-Chronostratigraphy

3.5.4.a Onshore evidence

There are few sites on the Outer Hebrides which give an indication of past Quaternary climates and events in the area, with the exception of a fairly large number of Postglacial deposits. The main sites are:

(1) Tolsta Head, Lewis (von Weymarn & Edwards, 1973; Birnie, 1983). Pollen analysis at this site reveals a cool, maritime climate preceeding the last glacial advance. Birnie's work at the site suggests climatic deterioration prior to glaciation and that the middle Devensian was characterized by cold conditions. A radiocarbon date, on the organic lake detritus beneath the till at this site, of $27,333 \pm 240$ years BP is reported by Sutherland (1984b), who interprets this as evidence of local Hebridean ice expansion during the late Devensian. However, von Weymarn and Edwards (1973) have suggested that this till was deposited from ice which had first crossed the Minch.

(2) Toa Galson, Lewis (Sutherland & Walker, 1984). This is reported to be an interglacial site, situated along a stretch of coastline considered to have been ice-free during the late Devensian maximum; a view which Peacock (1981) supported. Peats are observed resting on a raised rock platform and are overlain by a soliflucted layer and the Galson beach. Palynological evidence points to an essentially treeless environment, possibly correlating to the site at Sel Ayre, Shetland (Birks & Peglar, 1979). The deposits are assigned to a non-specific interglacial. However, the question of the validity of an interglacial treeless environment seems questionable in view of the common occurrence of *Pinus* and *Betula* fragments in Holocene peats from the area (eg. Peacock, 1984). Three dates are reported from different fractions of the peat at this site and yield dates $>47,150$ radiocarbon years BP.

Elsewhere from north west Lewis, radiocarbon ages of 34,000 and 40,000 years BP are reported from marine shells within 'shelly

diamictons' (Sutherland, 1986) and amino acid analyses support a late Devensian age (Peacock, 1984). However, Sutherland and Walker (1984), commenting on radiometric dates from these shelly diamictons, believe that they are the product of both middle and late Devensian shells; implying that the older shells are reworked and that the last glaciation of northern Lewis occurred during the Late Devensian.

(3) Village Bay, Hirta (Sutherland *et al.*, 1984). Discontinuous lenses of organic material within the Abhainn Ruaival Sand are reported beneath 1.5 m. of deposits interpreted as periglacial slope sediments. Low concentrations of poorly preserved pollen suggests a low diversity grassland without shrub cover and, in comparison with the greater diversity of modern heathland floras, it is proposed that these deposits represent part of an interstadial or end-interglacial.

Dates from the organic sands were based upon three organic fractions in an attempt to assess the degree of contamination. The insoluble fraction residue gave a radiocarbon age of 24,710 \pm 1470/-1240 years BP; the humic acid date was considerably younger and suggested contamination at the site. The sands are therefore suggested to pre-date the late Devensian maximum.

The relative dating scheme of Sutherland *et al.* (1984) on Hirta, using clast weathering 'rind' thickness, is considered to be unreliable by Selby (1989).

Younger, well preserved organic deposits lying both sub- and inter-tidally have become submerged by rising sea levels following the last glacial maximum eg. Uist and Benbecula (Ritchie, 1985). Pollen analysis of these peats suggests marshland and heathland development with some woody layers indicating a sparse tree cover. Radiocarbon dates (Ritchie, 1985) place deposition of the deposits at 8330 \pm 65 years BP.

3.5.4.b. Offshore evidence

The offshore evidence of Quaternary events and their dating is extremely limited from the Hebridean shelf. Jones *et al.* (1986) report radiocarbon ages from offshore shelly gravels overlying

glacial marine muds to the west of the Outer Hebrides, of $9,590 \pm 350$ years BP (20 km. south of St. Kilda) and $11,300 \pm 550$ years BP (25 km. south east of the Flannan Isles).

Selby (1989) discusses four radiocarbon dates reported by Hedges *et al.* (1988) and these are outlined in Table 3.2. The date of $>43,000$ years BP from vibrocore VE 57/-09/59, from the Conan B subsequence, is interpreted as a product of reworked older shell material and it seems unlikely that a temperate species like *Arctica islandica* could have been living in a morainal bank depositional environment which Selby proposed for these deposits. The fragment of *A. islandica* is therefore considered to have been subglacially transported from mid-late Devensian deposits. The three other dates, all from complete shells which are regarded as *in situ*, are considered to be reliable by Selby (1989).

Selby (1989) also reports on the occurrence of volcanic glass shards from vibrocore VE 57/-09/46, recorded from a water depth of -156 m. in the St. Kilda Basin. The variation in shard concentrations are illustrated in fig.3.4, which illustrates three distinct peaks. A peak at 2.85 m. is composed of a clear, commonly flat or slightly curving bubble wall type acidic shards occasionally up to 500 μm . across, but generally smaller. These shards are reported to display up to three small wing roots and occasionally randomly orientated, regular elliptical and 'V' shaped pits up to 20 μm . long and 10 μm . across. These colourless acidic shards are scattered throughout the upper part of the vibrocore but are rare below 3.00 m.. A well defined concentration peak of brown basic shards occurs at 2.15 m.; these are typically larger than the acid shards and are generally blocky and vesicular in appearance. Selby (1989) reports on the BGS analysis of the shard geochemistry and concludes that the clear acidic shards correspond to the Vedde Ash (Mangerud *et al.*, 1984), while the brown basaltic shards appear to be a complex geochemical mixture possibly belonging to North Atlantic Ash Zone 1 (NAAZ1: Ruddiman & McIntyre, 1973). However, there are problems associated with the separation of Vedde Ash from NAAZ1 and a further Icelandic eruption which led to the deposition of the Saksunarvatn Ash (Mangerud *et al.*, 1986) may also belong here. The proposed

radiocarbon dates of these eruptions are: Vedde Ash 10,600 years BP (Mangerud *et al.*, 1984); NAAZ1 9,800 years BP (Ruddiman & McIntyre, 1976); and the Saksunarvatn Ash around 9,050 years BP (Mangerud *et al.*, 1986). The implications of this volcanic ash stratigraphy are discussed at length by Selby (1989) and later in this chapter.

<u>Oxa-1322</u>	
species:	<i>Portlandia (Yoldiella) lenticula</i>
vibrocore:	57/10/21
depth:	4.30 - 4.55 m.
seismic unit:	Conan A
radiocarbon age:	22,480 \pm 300 years BP
 <u>Oxa-1323</u>	
species:	<i>Arctica islandica</i>
vibrocore:	57/09/59
depth:	3.0 m.
seismic unit:	Conan B
radiocarbon age:	>43,000 years BP
 <u>Oxa-1324</u>	
species:	<i>Buccinum terraenovae</i>
vibrocore:	57/09/46
depth:	4.5 m.
seismic unit:	Fionn
radiocarbon age:	11,680 m. \pm 240 years BP
 <u>Oxa-1325</u>	
species:	<i>Macoma calcaria</i>
vibrocore:	57/09/32
depth:	3.9 m.
seismic unit:	Fionn
radiocarbon age:	11,210 \pm 90 years BP

Table 3.2 Radiocarbon ages from the Hebridean shelf (Selby, 1989 after Hedges *et al.*, 1988).

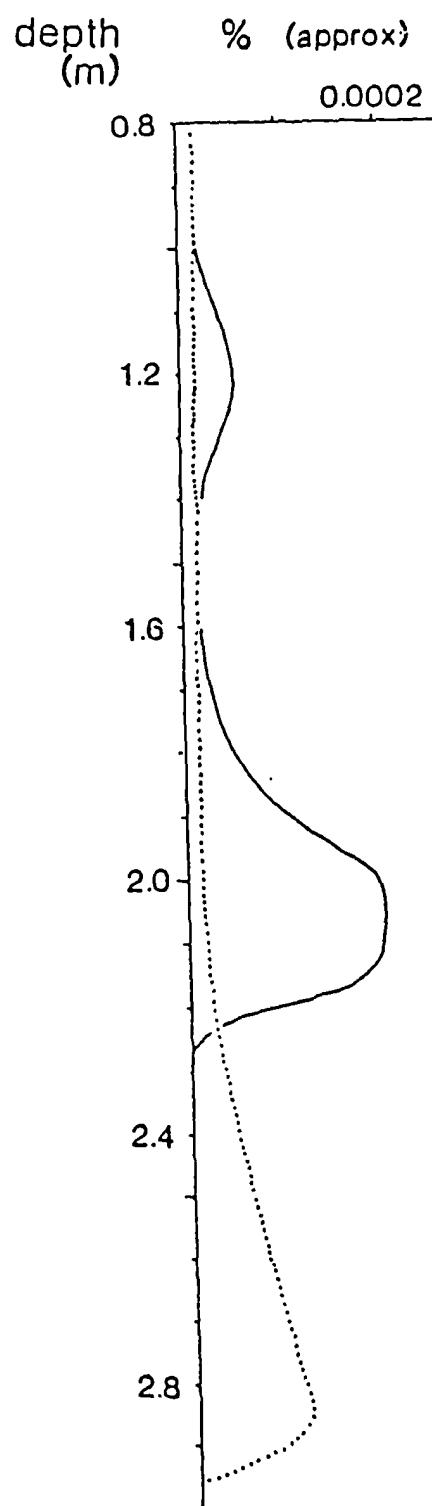


Fig.3.4 Volcanic ash concentrations within the sand fraction of vibrocore 57/-09/46 recovered from the Fionn sequence, St. Kilda Basin. Clear acidic shards = , brown basaltic shards = — (from Selby, 1989).

3.5.5 Summary of present status

The timing of glacial events on Lewis and Harris remains controversial according to Selby (1989), although the presence of an Hebridean ice mass or masses is now widely accepted. The early models of invading Scottish ice, suggested by Geikie (1873, 1878) and later by Jehu and Craig (1927, 1934), are not considered realistic in view of the radially flowing ice-cap models which have the icedomes centred over the highlands of south Lewis (Flinn, 1978; von Weymarn, 1979; and Peacock, 1981a). However, problems remain regarding the manner of interaction between Scottish ice and local ice eg. Tolsta Head and the northern tip of Lewis.

Peacock and Ross (1978), working on South Uist and Benbecula, proposed an easterly or E.S.E. ice flow during the last glacial event; a view supported by observations on the distribution of erratics (Coward, 1977; Flinn, 1978). In an attempt to place this evidence into a broader Scottish context, Sissons (1980) proposed that during an early stage of the last glacial maximum, Scottish ice overran the southern Hebridean islands and then decayed rapidly by calving in the Minch. The then unbalanced Hebridean ice sheet, unrelated to local relief, was able to flow eastwards. This proposal, based on the distribution of erratics and a western ice-shed position, overcame the envisaged climatic difficulties associated with establishing an oceanically biased ice mass in an area of low relief.

Flinn (1980) rejected Sisson's suggestion on the basis that ice thickness would have prohibited calving in the Minch. The alternative proposition was that preferential drainage of the ice sheet to the east may have led to westwards migration of the ice-shed. Peacock (1980) further pointed out the lack of definite Inner Hebrides and Minch lithologies on the Outer Hebrides and suggest that ice thickness on the eastern side of the island was at least 400 m. thick and that thicker ice probably existed to the west at the ice-shed location. Sisson (1983) accepted that his earlier views were flawed but proposed that Scottish mainland ice may have enveloped Hebridean ice during the early(?) Devensian.

Peacock's (1984) multiple event hypothesis for the Late

Devensian Stadial, with at least two glacial phases, seems likely in view of the evidence and, together with von Weymarn (1979), proposes that Scottish ice did indeed flow over northern Lewis. Selby (1989) believes that this is more appropriate than Sutherland and Walker's (1984) proposal that ice flowed offshore to the north east of Lewis and was then driven back, presumably by the Scottish ice body, onshore around northern Lewis. In summary, the onshore evidence generates debate regarding whether or not Scottish ice crossed the islands, and whether or not there was a single, independent Outer Hebridean ice mass.

Offshore, work by the BGS and Selby (1989), in particular, provides a framework for the present study and a model of the late Devensian glaciation of the Hebridean shelf. Selby's generalized model, outlined in fig.3.5, describes the sequence of glacial deposition in three stages from the glacial maximum, the floating-off of St. Kilda Basin ice, and the eastward glacial retreat onto the inner-shelf rock platform. The present study aims to examine some of the palaeoenvironmental changes, recorded in the sediments of the shelf, and associated sequence of events.

3.6. Seismic Geometry

The seismostratigraphic scheme of Selby (1989) for the Hebridean shelf is summarized in fig.3.2 and discussed in section 3.5.1. In view of the relatively shallow depth of penetration capable from vibrocore sampling, it is the stratigraphy and geometry of the upper few meters of the sedimentary succession which are of most interest. The deployment of the Hunttec deep-tow boomer seismic system has, in particular, aided the interpretation of sediment geometry and stratigraphy in this area. Two seismic profiles are included here: a 'sparker' profile from the morainal banks of the middle to outer shelf (fig.3.6a), and a Hunttec deep-tow boomer profile from the St. Kilda Basin (fig.3.6b), illustrating the seismically laminated nature of the sediments. Discrete v-shaped features are visible in fig.3.6b and are suggested to represent glacial scours (C.D.R. Evans pers. comm.). The minimum depth to the top of these scours can be calculated on the assumption that the sediments are muddy sands with a velocity

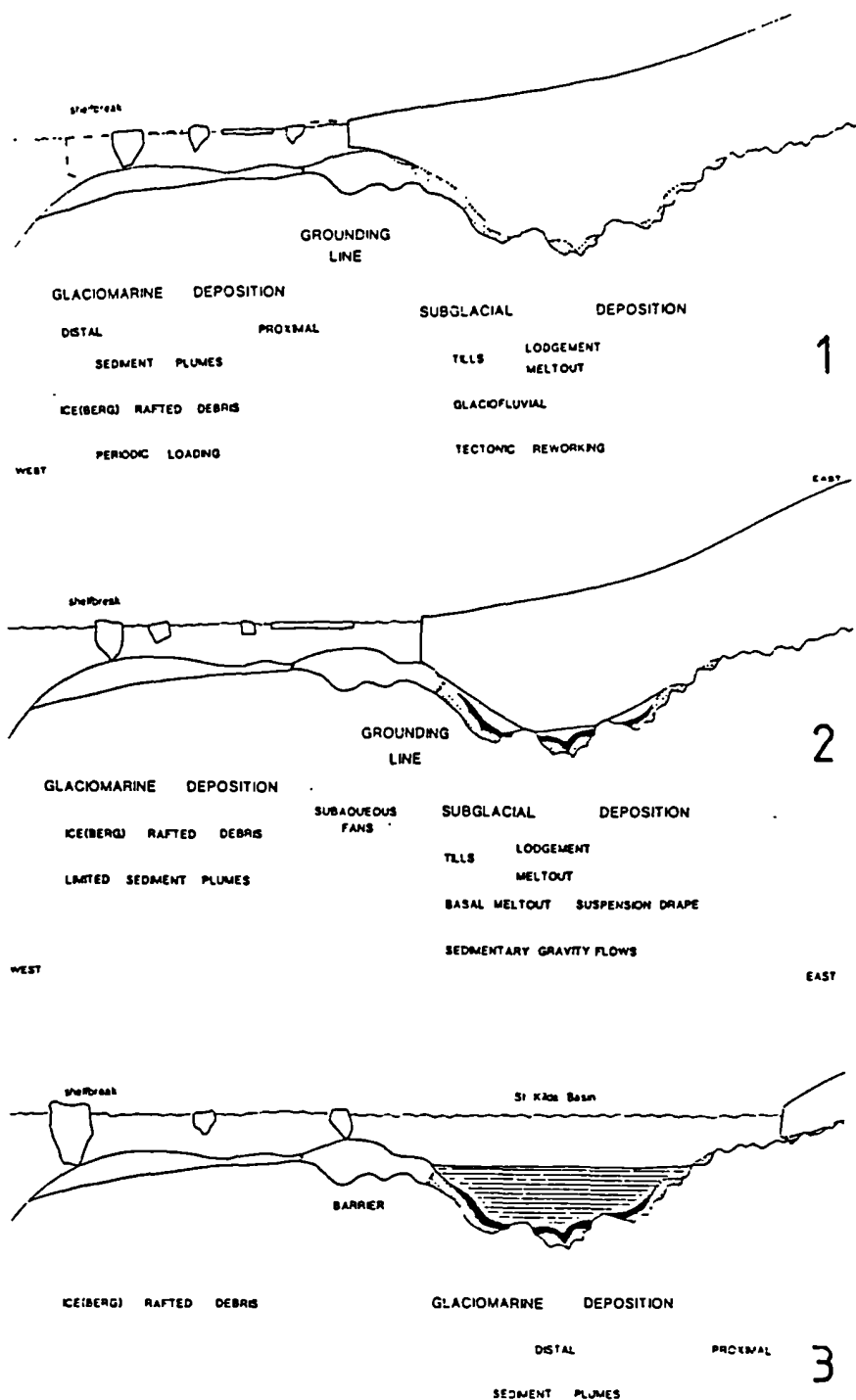


Fig.3.5 Selby's (1989) three stage model of deglaciation and associated depositional environments from the Hebridean shelf; sections orientated East-West. Stage 1 = Glacial maximum, grounded ice occupies the St. Kilda Basin. Stage 2 = Hebridean ice tongue floats over the St. Kilda Basin. Stage 3 = Rapid glacial retreat onto the inner shelf.

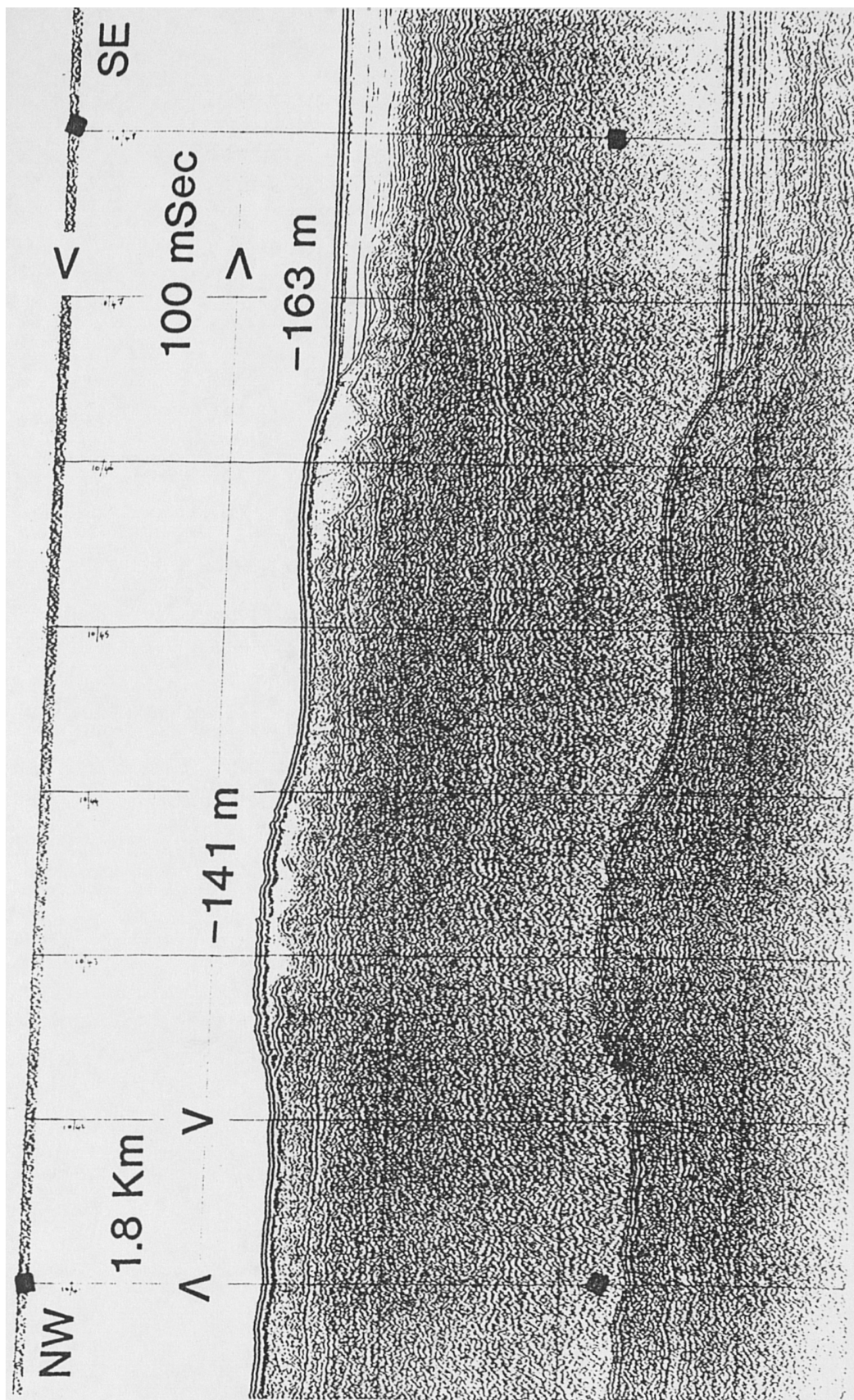


Fig.3.6.a BGS 'sparker' profile across the morainal banks of the middle shelf, immediately west of the St. Kilda Basin. Section aligned N.W.-S.E.; water depths and two-way travel times shown (courtesy of Dr. C.D.R. Evans).

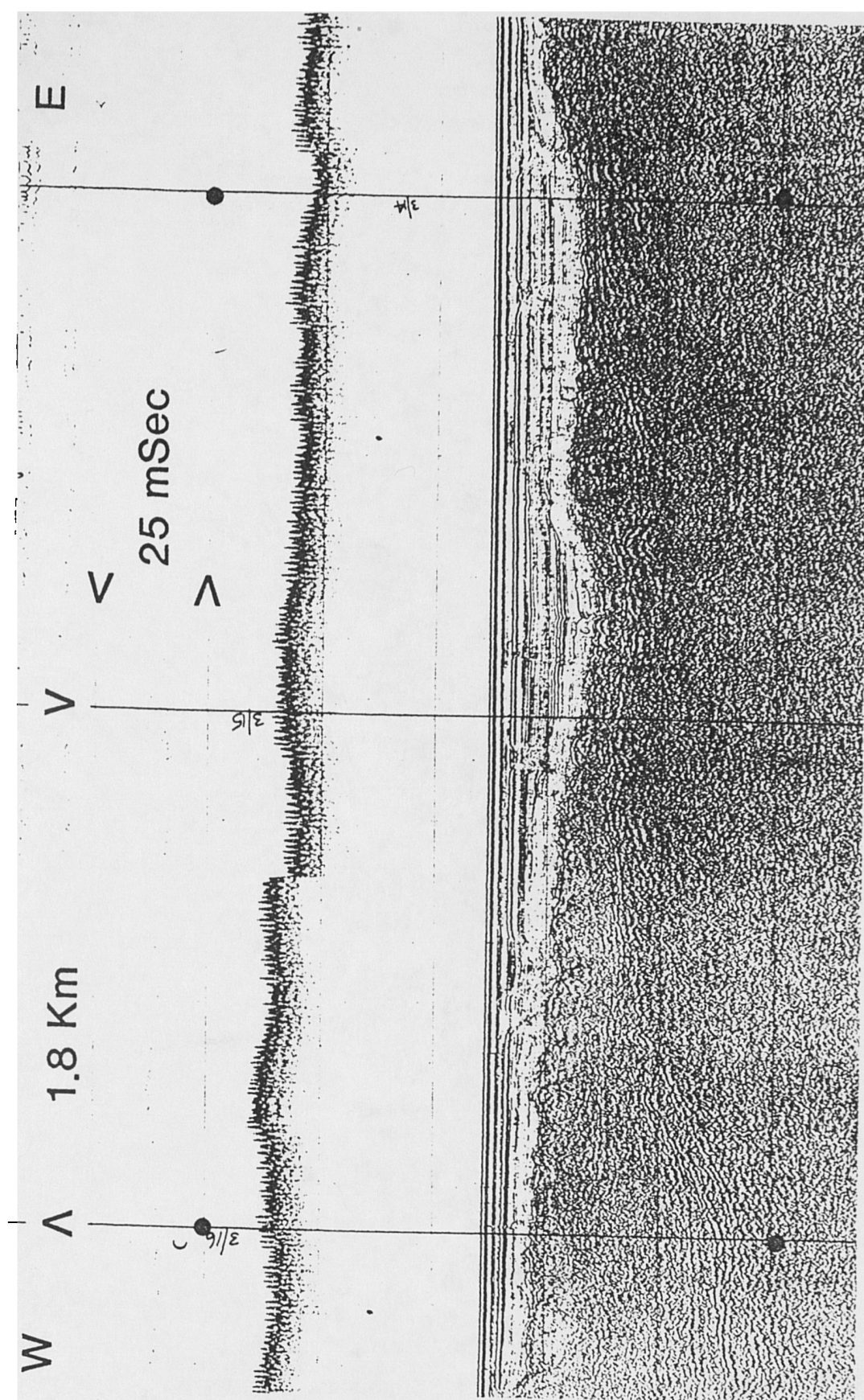


Fig.3.6.b Huntec deep-tow boomer profile orientated. E.-W. across the St Kilda Basin. Note the 'V-shaped' structures within the seismically laminated Fionn sequence (courtesy of Dr. C.D.R. Evans).

of about $1,600 \text{ m.s}^{-1}$, the horizontal intervals represent 25 msec. two way travel times, and so each interval represents approximately 20 m.. The top of the scours are therefore estimated to be at a depth of just under 3 m. in this part of the basin which as the radiocarbon dating and palaeoenvironmental reconstructions will demonstrate for core VE 57/-09/46, which also comes from this part of the basin, falls within the Loch Lomond Stadial.

The value of the high resolution seismic work of the BGS, and Dr.I.Selby in particular, cannot be overemphasised in providing a regional framework into which to place the cores analysed in this study.

3.7 Sedimentological Characteristics

On the shelf, Selby (1989) has designated "type" cores to represent the various seismic sequences (fig.3.2) on the basis of their location along seismic profiles; although as discussed in chapter 2.- there are problems relating to fix error and the relocation of core sites to the seismic profiles of earlier cruises. Other cores, apart from these "types", were examined to determine the degree of variability within the shelf seismic sequence, although not all of the seismic sequences defined by Selby were sampled due to their depth beyond the maximum limit of vibrocore penetration. The lithological results are presented here following Selby's seismostratigraphic subdivisions. Unfortunately, Selby describes each core as representative of a single facies, partly constrained in his outlook by the seismic sequence subdivision for which he attempts to designate "type" cores.

3.7.1 Conchar Sequence

Core 57/-10/17 was recovered some 2 Km. west of the shelfbreak at a depth of -172 m. and consists largely of dark grey (5Y 4/1) stiff and muddy diamictons with clasts up to 4 cm. across; the upper c. 80 cm. comprise shelly, coarse sands above a sharp boundary. Possible bedding structures of alternating dark and lighter bands are reported at about 3.0 m. and probably represent monosulphide stairing in response to grain size

variations.

3.7.2 Conan Sequence

3.7.2.a Conan A subsequence

The representative cores of the Conan A subsequence are VE 57/-09/60 and 57/-10/21, which are composed mainly of dark grey (2.5Y N4/0 and 5Y 3.5/1) homogenous, muddy diamicton. Vibrocore 57/-09/60 is sharply capped by coarse, shelly sands. The diamicts are very stiff with abundant shell material and echinoid fragments; the constituent sand fraction is dominantly very fine/fine, angular to sub-rounded, of moderate sphericity and is 75% quartz. Clasts are typically <2 cm. across, although there are cobbles up to 10 cm. present. However, since the clast size limit is a function of the sampling procedure (ie. the barrel diameter), it is highly likely that larger clasts are present but not recovered. The clasts are sub-vertically orientated and vary from sub-angular to rounded and are occasionally faceted or striated. Selby (1989) reports no evidence to suggest clast re-alignment as a result of the coring process, although I have been able to demonstrate a disturbed stratigraphy in VE 57/-09/60 and would consider clast orientation adjacent to the casing as unreliable. Gneissic lithologies dominate, with some basic igneous and red sandstone lithologies present.

Clast morphological analyses (Selby, 1989) indicate that the clasts of this subsequence lie in the subangular class with a roundness mean of 0.41 and a sphericity (u) of between 0.66 and 0.83. Particle size studies reveal an Inclusive Graphic Mean of between 3.9 ϕ (very fine sand) to 8.0 ϕ (fine silt); Inclusive Standard Deviation ranges from 3.6 ϕ to 5.7 ϕ and is defined as very poorly to extremely poorly sorted; Inclusive Graphic Kurtosis between 0.72 and 1.08; and Inclusive Graphic Slewness of between +0.03 and +0.27. There is a slight tendency towards bimodality, with a dominant clay peak and a minor 4 ϕ peak. The sand:silt:clay ratios are illustrated in fig.3.7 for the Conan A subsequence and other seismic sequences; it is interesting to note from this figure how the various sequences can be distinguished eg. the

Conan A and Fionn sequence are most readily differentiated by their silt and clay proportions. The variation illustrated by this subsequence in fig.3.7 can be accounted for by a distinct westward decrease in sand content, changing by 30%, with a corresponding increase in silt (10%) and clay (15%).

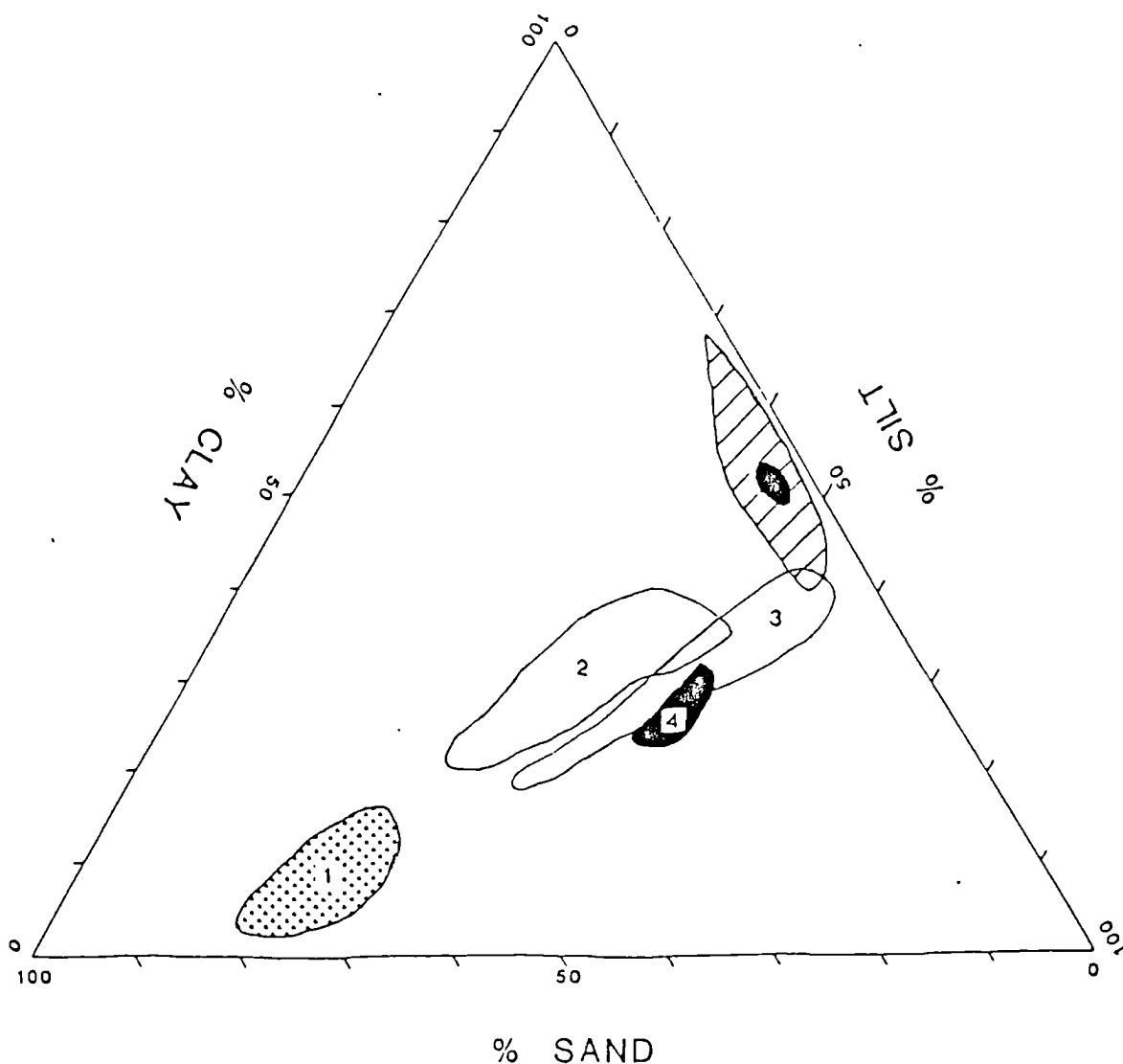


Fig.3.7 Ternary diagram, summarizing the ratios of sand/silt/clay from sediments of the various seismic sequences of the Hebridean shelf and slope, and Outer Hebridean terrestrial lodgement tills. 1= Lodgement till, 2= Conan A subsequence, 3= Fionn sequence, 4= Conan B subsequence, [hatched] = slope sediments from >300 m. water depth (from Selby, 1989).

3.7.2.b Conan B subsequence

Selby (1989) bases this subsequence upon cores VE 57/-09/59 and 57/-09/89, recovered from the banks on the western margin of the St. Kilda Basin (cf. fig.3.6.a). Only vibrocore VE 57/-09/89 has been analysed in the present study. Selby describes each core as representative of a single facies and in this case recognizes a dark grey (5Y 3/1, 5Y 4/1 and 5Y 4/2) muddy diamicton interspersed with clayey, silty and sandy horizons. Figure 3.8 is a summary lithological log of VE 57/-09/89 and illustrates the variation within this short core. Clasts up to 7.5 cm. occur and vary from sub-vertical to sub-horizontal; clast size and frequency are higher below 3.0 m. and increase towards the base of the core. These clasts are predominantly gneissic and clast analysis of VE 57/-09/59 indicates a mean roundness 0.41 (sub-angular) and sphericity (u) between 0.69 and 0.81. Within the matrix of the lower diamict the sands are mostly very fine/fine, well sorted, sub-angular to sub-rounded, and about 75% quartz, often stained red.

Sedimentary grain-size analysis reveals an Inclusive Graphic Mean of 6.4ϕ to 9.3ϕ ; Inclusive Graphic Standard Deviation of 3.0ϕ to 3.8ϕ and is defined as poorly sorted; Inclusive Graphic Kurtosis of between +0.79 and +0.90 (= platykurtic); and Inclusive Graphic Skewness of -0.05 to +0.46 which defines a near-symmetrical to strongly finely skewed distribution. Thus, two facies appear to be present in this subsequence from the analysis of grain-size data, some samples are very well sorted and extremely poor in sand, while others are comparable to the Conan A subsequence; this is illustrated in fig.3.7.

There are definite signs of bioturbation towards the top of VE 57/-09/89, particularly around 1 m.

3.7.3 Oisein sequence

This sequence was not sampled in the St.Kilda region, but rather from a seismically identical draped sequence exposed at the sea bed at $56^{\circ}35'N$, $8^{\circ}24'W$ (Peach sheet), some 50 km. to the south west of Barra Head. The core analysed is VE 56/-09/142 which consists largely of dark grey (5Y 4/1 and 5Y 4/2) sandy muds.

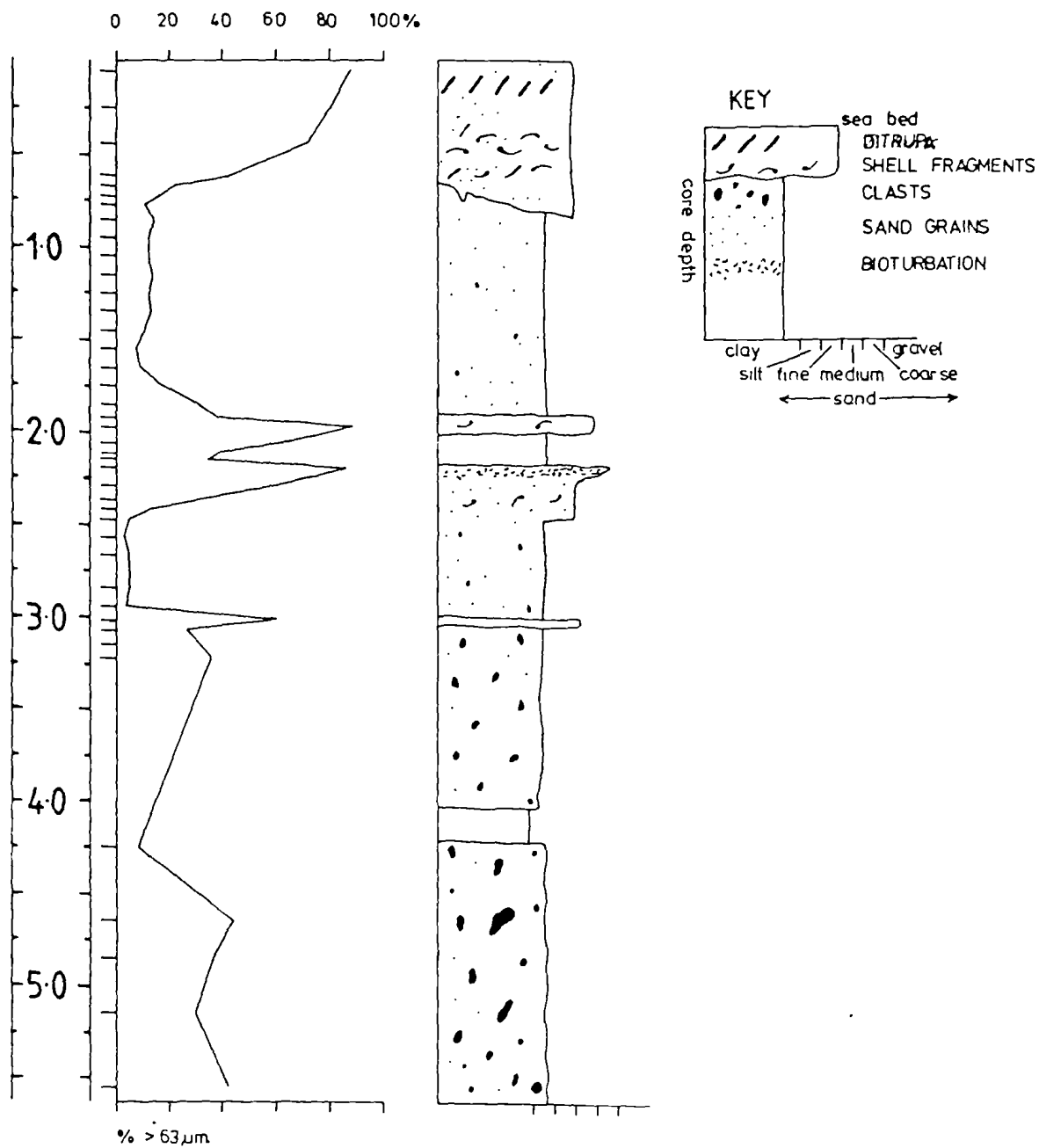


Fig.3.8 Summary lithological log of vibrocore 57/-09/89 and percentage frequency of coarse residue (>63 μm.) at the sample levels analysed. Depths in metres.

Below 4.2 m. Selby reports layering of fine sand/silt/clay bands with gradational boundaries. Rare clasts up to 2.2 cm. occur throughout, but are increasingly common below 3.2 m. Selby reports signs of bioturbation, but none were observed by the present author while sub-sampling this core.

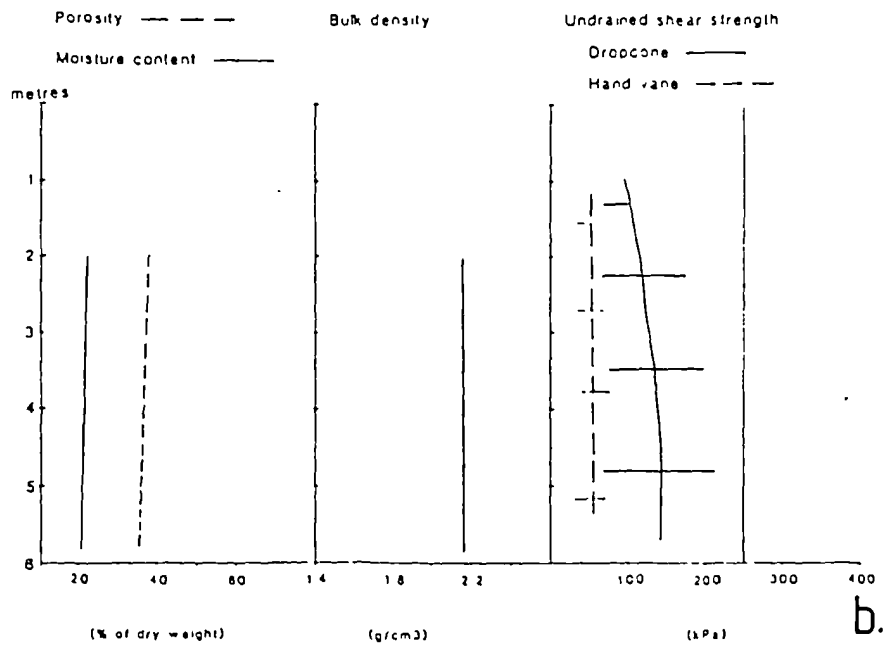
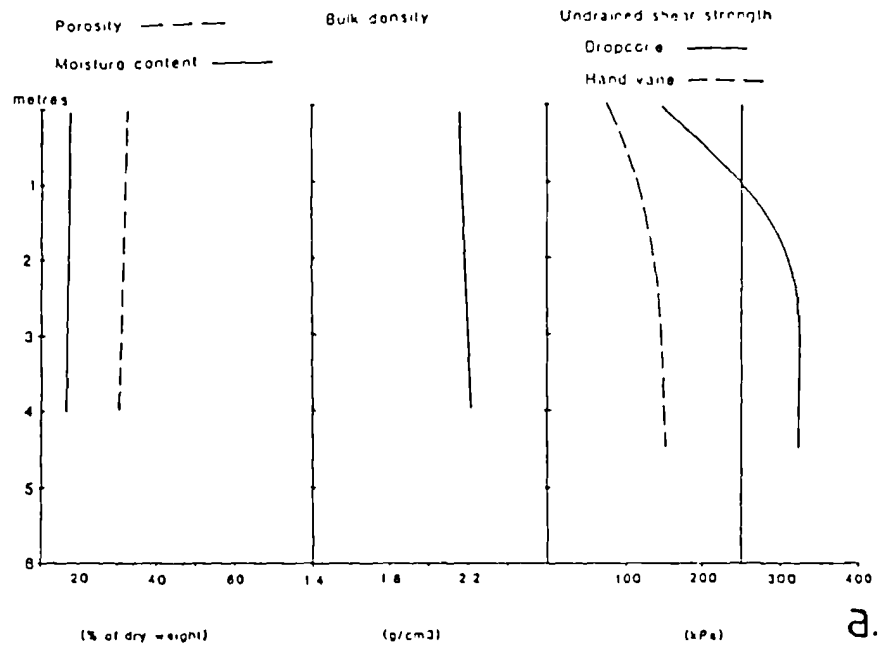
3.7.4 Fionn sequence

The representative cores of the Fionn sequence are VE 57/-09/32 and 57/-09/46. Only the latter has been analysed in detail during the present study, together with a few samples from VE 57/-09/44 (courtesy of D.K. Graham) and a limited number from VE 57/-09/32 since only some of the core sections could be found.

The sediments consist of homogenous, dark grey (5Y 4/1) soft muds, which become increasingly silty and sandy towards the top of the core. The sand fraction is very fine, angular to sub-angular, well sorted, and approximately 75% quartz. The following grain-size parameters are reported: Inclusive Graphic Mean between 6.1ϕ and 8.6ϕ ; Inclusive Graphic Standard Deviation between 3.3ϕ and 3.8ϕ ; Inclusive Graphic Kurtosis between +0.67 and +0.80 (= platykurtic); and Inclusive Graphic Skewness of 0.07 to 0.54, which ranges from near symmetrical to strongly finely skewed. It appears that the Fionn sequence is the best sorted and is reported by Selby to maintain this level of sorting throughout a wide range of skewness. Structures revealed by X-radiographs of the core sections include a faint layering throughout and bright 'threads' indicating a limited number of pyritized burrows.

3.8 Geotechnical properties

It is beyond the scope of this study to fully discuss the geotechnical measurements undertaken on these cores by Selby (1989). However, such measurements can be valuable when attempting to identify the depositional processes and possibly assess the degree of post-depositional modification. In fig.3.9 the undrained shear strength measurements on these sequences are outlined. These measurements are based upon hand dropcone readings; caution should be exercised in interpreting values in excess of 250 KPa.



Figs.3.9 Undrained shear strength (kPa) characteristics and other geotechnical properties of (a) Conan A subsequence (b) Conan B subsequence. Error bars show the variation in measurements from the various cores (from Selby, 1989).

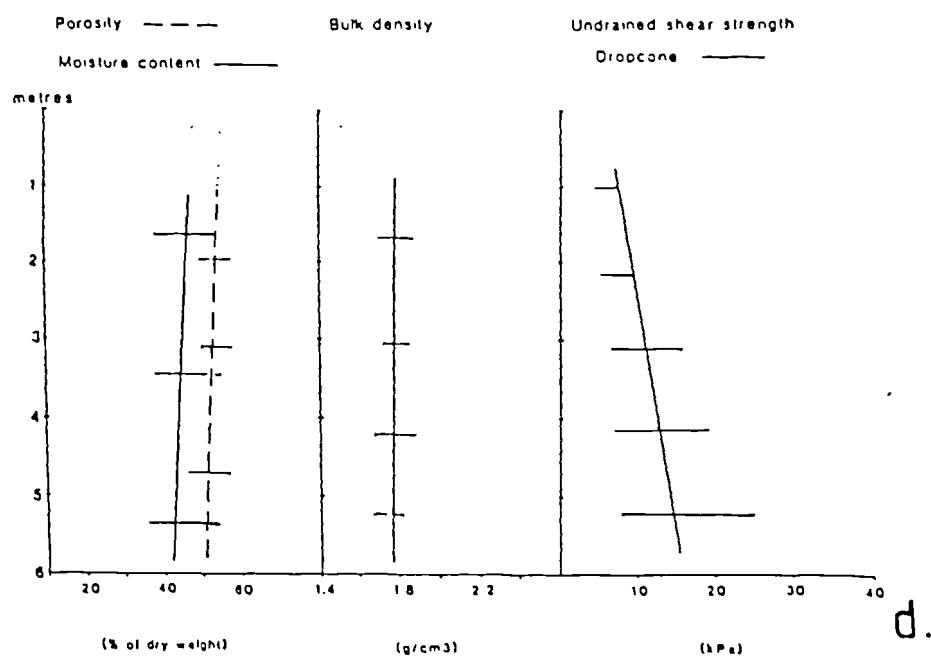
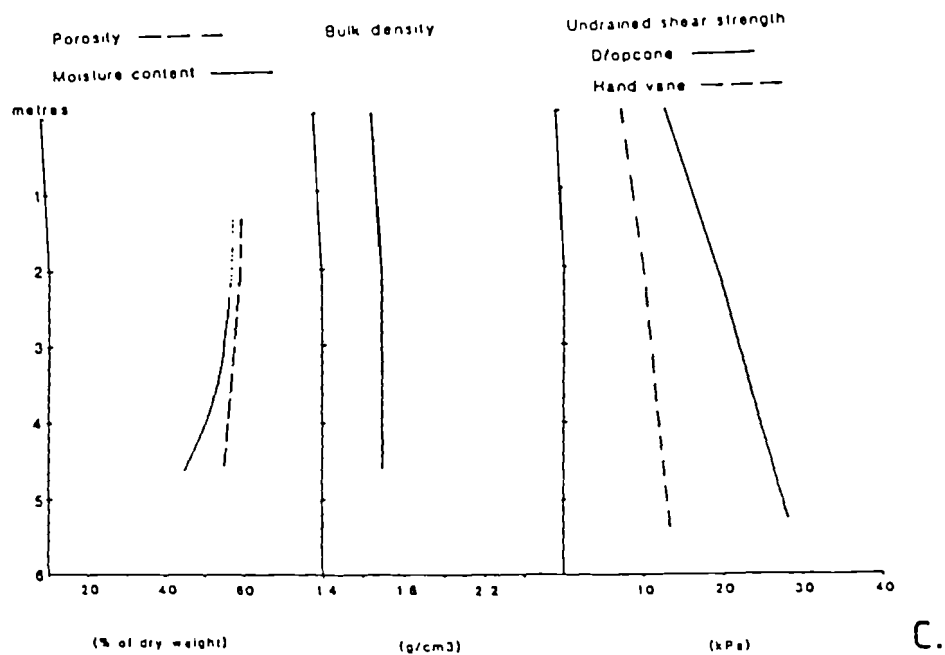


Fig.3.9 (continued) Undrained shear strength (kPa) characteristics and other geotechnical properties of (c) Oisein sequence (d) Fionn sequence. Error bars indicate the variation in measurements from the various cores (from Selby, 1989).

3.9 Results: Faunal Characteristics

The results of the micropalaeontological analyses are presented for each of the cores analysed within the framework of the seismic sequences identified. These faunal characteristics provide the basis for palaeoecological reconstructions, together with evidence of sidewall contamination in VE 57/-09/60, and evidence of bioturbation in VE 57/-09/89. Radiocarbon ages are reported at the relevant sections and provide a chronological framework for vibrocores 57/-09/89 and 57/-09/46 in particular; a summary of the radiocarbon dates obtained from these cores is included in Appendix 2.

3.9.1 Conchar sequence

Thirteen samples were analysed from vibrocore VE 57/-10/17, recovered from a water depth of -172 m. OD just beyond the shelf-break. This was the deepest water depth from which any of the cores studied in detail were analysed.

The sequence is divided into two distinct foraminiferal zones, with a gradational boundary between 0.78 m. and 0.93 m.; the faunal characteristics are summarized in figs.3.10 a,b.

In the lower zone, zone 1, the faunas are dominated by *Cassidulina reniforme* and *Elphidium excavatum* forma *clavata*; the latter dominating below 3.0 m. and the former from 3.0 m. to 1.0m.. Important accessory species include *Cibicides* gr. *lobatulus*, *Elphidium albiumbilicatum*, *Islandiella helenae*, *Nonion labradoricum*, *N. orbiculare*, and *Trifarina angulosa*. Other commonly occurring taxa which are largely confined to this zone include *Astrononion gallowayi*, *Buccella frigida*, *B. tenerrima*, *Elphidium asklundi*, *Islandiella islandica*, *I.norcrossi*, and *Quinqueloculina stalkerii*. The increasing frequency of *N.labradoricum* in the upper part of zone 1 is a notable feature of fig.3.10a.

Zone 1 faunas are characterized by moderate faunal diversities, ranging from 10 to 16 species, and relatively high faunal dominance, maximum value 42.9% at 3.08 m., which decreases upwards through the core. Planktonic:benthonic ratios remain low, less than 0.2, throughout and steadily decrease upwards through

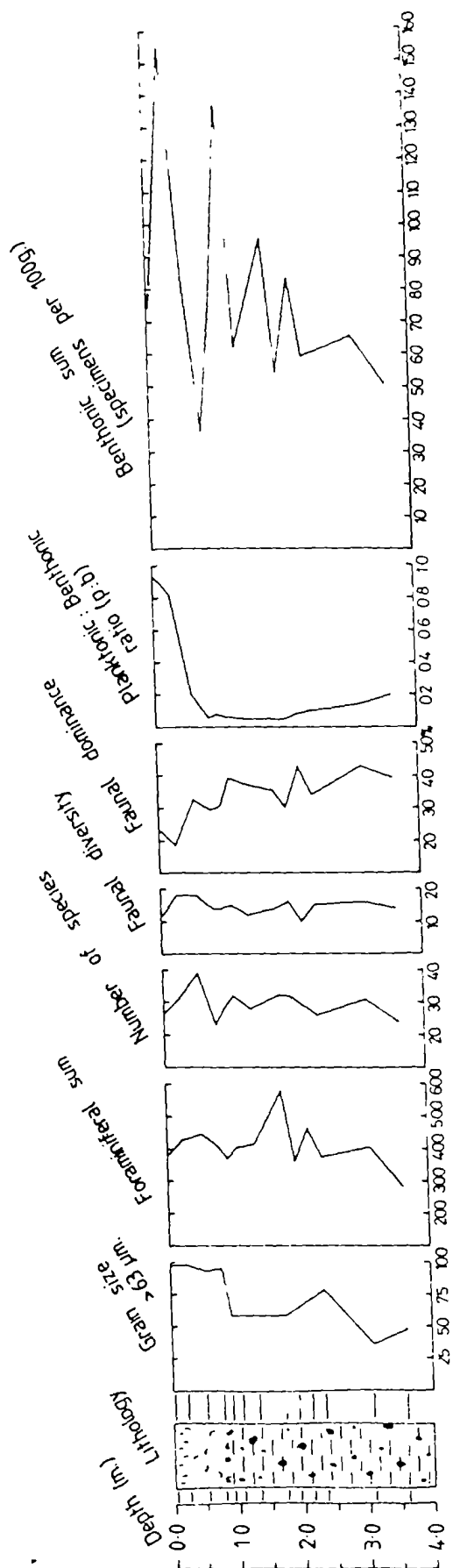


Fig.3.10.b Faunal characteristics of vibrocore 57/-10/17.

the zone. The benthonic sum per 100 g. of sediment is variable but exhibits a general increase upwards.

Zone 2, defined from between 0.78 m. and 0.93 m. to the top of the core, closely matches the lithological change in the core at the same level. However, unlike the sharp contact between the diamict and overlying sands, the faunal changes between these zones are more gradational and are characterized by decreasing frequencies of zone 1 taxa and increasing frequencies of zone 2 taxa. The dominant species are *Cassidulina laevigata* and *Spiroplectammina wrightii*, with *Bolivina difformis*, *Cassidulina obtusa*, *Cibicides lobatulus*, and *Trifarina angulosa* as the most important accessory species. Other commonly occurring taxa which characterize zone 2 faunas are *Globocassidulina subglobosa*, *Rosalina praegeri*, and *Stainforthia fusiformis*.

Zone 2 faunas have slightly higher faunal diversities than zone 1, with a mean zonal diversity of 15.5 species compared to 14 species for zone 1. Faunal dominance is also lower in zone 2, with a minimum value of 18.6% at 0.23 m., and exhibits the overall trend within the core of decreasing dominance upwards. However, the most characteristic feature of zone 2 faunas are the rapidly increasing planktonic:benthonic ratios above 0.78 m., with a maximum of 0.93 in the surface sample. Benthonic sums are also high in this zone, attaining a maximum 154,800 specimens per 100g. sediment at 0.23 mm., but also a core minimum of 36,540 specimens per 100 g. sediment at 0.78 m.. The mixed nature of the faunas, with species characteristic of both zones present, is particularly evident at 0.78 m. and 0.53 m..

3.9.2 Conan A subsequence

Two cores have been analysed from the Conan A subsequence, VE 57/-09/60 and VE 57/-09/21.

3.9.2.a Vibrocore 57/-09/60

Recovered from a water depth of c. -135m. OD, immediately to the west of the St. Kilda Basin. The sequence is divisible into two distinct foraminiferal zones with a transition at 0.24 m. which closely matches the lithological change at this level

(fig.3.11). The summary results of the foraminiferal analysis are presented in fig.3.12a,b.



Fig.3.11 Photograph of split core section ^a/4 from vibrocore 57/-09/60. Scale = 5 cm. intervals.

The lower zone, zone 1, is dominated by *Cassidulina reniforme* and *Elphidium excavatum* forma *clavata* throughout; the latter tending to be the dominant species, particularly below 1.25 m.. The main accessory species are *Elphidium albiumbilicatum*, *Cibicides* gr. *lobatulus*, *Nonion orbiculare* and *Trifarina angulosa*, with other commonly occurring species including *Astrononion galloway*, *Buccella frigida*, *B. tenerrima*, *Elphidium asklundi*, *Islandiella helenae*, *I. norcrossi*, *Quinqueloculina stalkerii* and *Stainforthia schreibersiana*. The change of faunal characteristics at about 1.25 m., although rather poorly defined (figs. 3.12a,b), indicate a subtle change in the depositional environment.

Faunal diversities from this zone vary from 15 to 23 species and are highest at between 0.32 m. and 0.39 m. Faunal dominance is equally variable but high, with a range from 38.4% at 0.34 m. to 48.8% at 1.92 m.; *Elphidium excavatum* forma *clavata* being the dominant species in both these samples. The planktonic:benthonic ratios of zone 1 are low throughout, attaining a maximum value of 0.15 at 0.25 m.. Benthonic sums from this zone are difficult to define, varying from a core minimum of 16,596 specimens per 100 g. sediment at 1.92 m. to a core maximum of 92,880 specimens per 100 g. of sediment at 1.09 m.. There does, however, appear to be a major step-up in benthonic sums between levels 60/8 at 1.09 m. and 60/9 at 1.29 m. and then a step down again between levels 60/5 at 0.34 m. and 60/4 at 0.27 m.

The transition from zone 1 to zone 2, while lithologically very distinct, is faunally less well defined but is placed between samples 60/2 and 60/3 at a depth of 0.24 m.. The main species are *Cassidulina laevigata* and *Spiroplectammina wrightii*, with *Cibicides* gr. *lobatulus* and *Trifarina angulosa* as important accessory species. Commonly occurring taxa which help to define this zone include *Cassidulina obtusa*, *Globocassidulina subglobosa* and *Rosalina praegeri*; many of these taxa extend throughout the core but exhibit a marked increase in frequency above 0.24 m.. Mixed faunas are indicated in samples 60/2 at 0.23 m. where the ranges of taxa which characterize both zones overlap.

Faunal dominance and diversity are lower in zone 2, with faunal diversity values of 13 in both samples and faunal

dominances of 31.5% and 32.3%. The planktonic:benthonic ratios are particularly high, with a value of 1.18 in sample 60/1 and 0.89 in sample 60/2. High benthonic sums are not, however, a distinctive feature of this zone with values of 69,430 specimens per 100 g. (60/1) and 20,092 specimens per 100 g. (60/2).

3.9.2.b Vibrocore 57/-10/21

This vibrocore comprises 4.56 m. of essentially lithologically homogenous diamicts recovered from a water depth of -146 m. OD. Six samples have been analysed for their foraminiferal content and the faunal results are presented in figs.3.13a,b. No faunal subdivision of the core is proposed on the basis of the foraminiferal assemblages present in these samples.

The most important taxa are *Cassidulina reniforme* and *Elphidium excavatum* forma *clavata*, the latter dominating throughout; while the main accessory species are *Cibicides* gr. *lobatulus*, *Elphidium albiumbilicatum*, *Islandiella helenae*, and *Trifarina angulosa*. Other commonly occurring species include *Astrononion gallowayi*, *Buccella frigida*, *Nonion orbiculare*, and *Quinqueloculina stalkerii*.

Faunal diversity is fairly high with a mean value of about 21 species throughout. Faunal dominance is also relatively high with a maximum value of 54% at 4.19 m., a minimum of 37.4% at 0.72 m., and a mean value of 46.9%. The planktonic:benthonic ratios are low and vary little from the mean value of 0.14 throughout. Equally, benthonic sums are relatively low, except at 2.82 m. where 46,184 specimens per 100 g. of sediment are estimated. This peak in the number of benthonic specimens corresponds to a peak, at the same level in the core, of the proportion of material >63 μ m.

3.9.3 Conan B subsequence

The Conan subsequence is represented by vibrocore VE 57/-09/89 of the present study and was recovered from a water depth of -156 m. OD from a morainal bank complex on the western side of the St. Kilda Basin (57°30.11'N, 08°42.52'W). As the following faunal subdivision of the core will demonstrate, Selby's (1989) interpretation of the sediments as "type" material for a seismic subsequence representing a single facies cannot be justified. This short core, a little less than six meters long is divisible into seven well defined foraminiferal zones. Full foraminiferal diagrams of vibrocore 57/-09/89 (enclosures 1 and 2) are included in the document wallet at the end of this volume; while fig.3.14 is a summary diagram of the twenty most commonly occurring taxa. The numerical results of the three zonation programmes, upon which the zonation of the core is based, are illustrated graphically in fig.3.15 and demonstrate the consistency of the seven zones recognized. The various benthonic parameters are illustrated in fig.3.16. Finally, the planktonic foraminiferal results are illustrated in fig.3.17.

The seven zones are now described from the base of the core upwards:-

3.9.3.a Zone 1

Zone 1 extends from the base of the core at 5.75 m. to 3.05 m.. The dominant taxa are *Elphidium excavatum* forma *clavata* and *Cassidulina reniforme*, with the common accessory species including *Bulimina* gr. *marginata*, *Cassidulina laevigata*, *Cibicides* gr. *lobatulus*, and *Nonion orbiculare*. Other commonly occurring taxa include *Astrononion gallowayi*, *Elphidium albiumbilicatum*, *Islandiella helenae*, *I.norcrossi*, *Nonion labradoricum*, and *Trifarina angulosa*.

Faunal diversity within zone 1 is variable but values are generally high, with a maximum value of 27 species in sample 89/45 at a depth of 4.65 m. and a minimum of 17 species in samples 89/44 (4.25 m.) and 89/48 (5.55 m.). Faunal dominance is equally variable and fairly high, with a mean zonal value of >40 %. Planktonic:benthonic ratios are low throughout much of the zone,

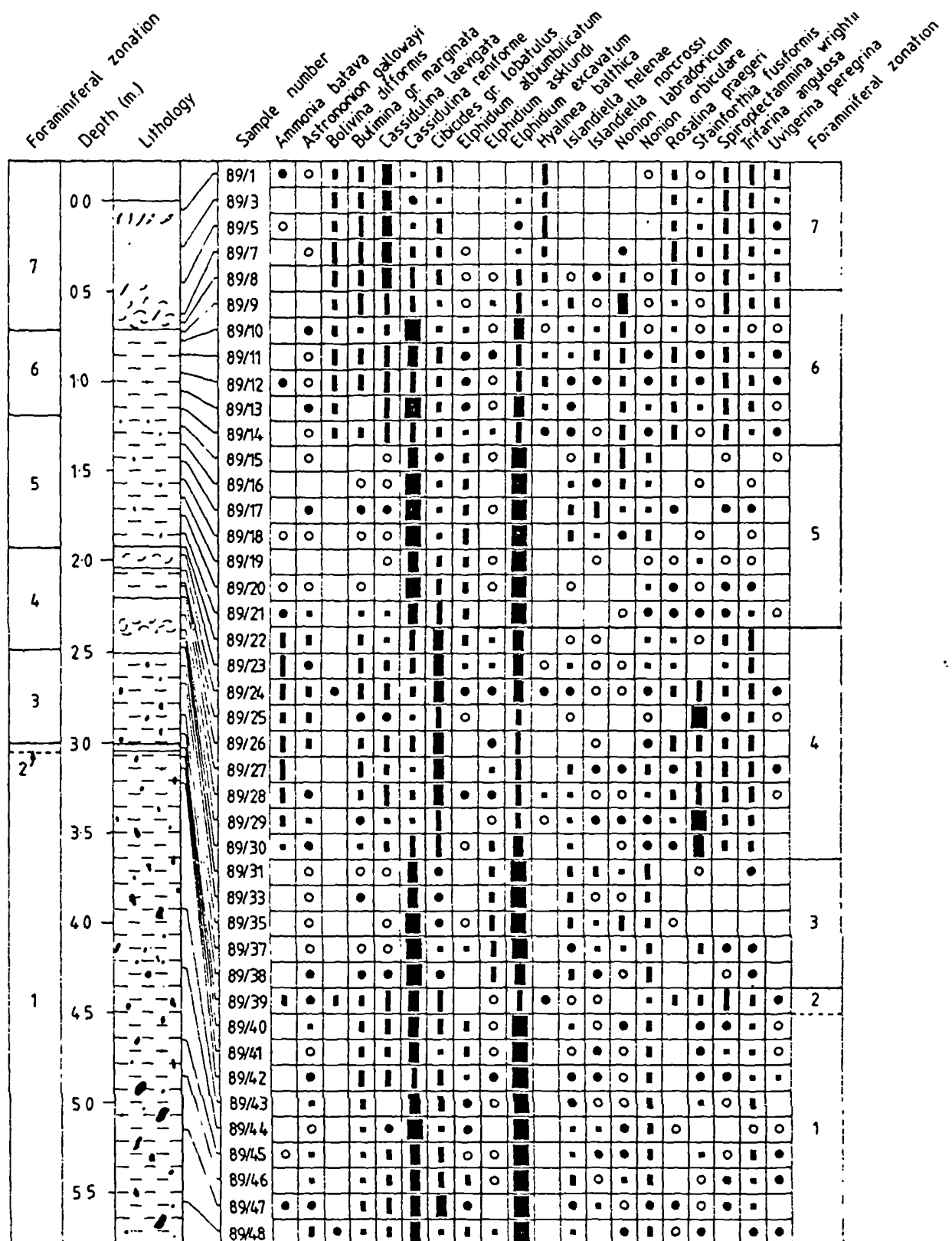


Fig.3.14 Summary diagram of the 20 most commonly occurring benthonic foraminifera from vibrocore 57/-09/89.

typically less than 0.2, but increase towards the top. Although planktonic count numbers are very low, the sinistral coiling form of *Neogloboquadrina pachyderma* clearly dominates over the dextral form. Benthonic sums are remarkably constant and low, ranging from 6,024 specimens per 100 g. sediment at 4.25 m. to 15,670 specimens per 100 g. sediment at 3.15 m.

A radiocarbon date, OxA-2785, was obtained from the top of this zone.

OxA-2785	
depth:	3.10 m.
species:	<i>Portlandia arctica</i>
corrected age:	15,245 \pm 170 BP

3.9.3.b Zone 2

This zone is tentatively defined between 3.05 m. and 3.0 m. and is represented by sample 89/39 only. The dominant taxa are *Cassidulina reniforme*, *Elphidium excavatum* forma *clavata*, *Spiroplectammina wrightii*, *Cibicides* gr. *lobatulus*, and *Cassidulina laevigata*. Common accessory species include *Ammonia batavus*, *Bolivina difformis* gr. *marginata*, *Rosalina praegeri*, *Stainforthia fusiformis*, and *Trifarina angulosa*. Faunal diversity is similar to the mean value of zone 1 and equals 21 species; while faunal dominance is greatly reduced at 23.5%. The planktonic:benthonic ratio of 0.39 is at its highest for this lower part of the core and comparable values are not attained again until zone 6. The benthonic sum at this level is also greatly increased from zone 1 with an estimated 64,887 specimens per 100 g.

3.9.3.c Zone 3

This zone is defined between 3.0 m. and 2.45 m. and once again the two dominant taxa are *Elphidium excavatum* forma *clavata* and *Cassidulina reniforme*. Frequencies of the former species increase rapidly upwards through zone 3, while the latter is generally in decline upwards and continues to do so into zone 4.

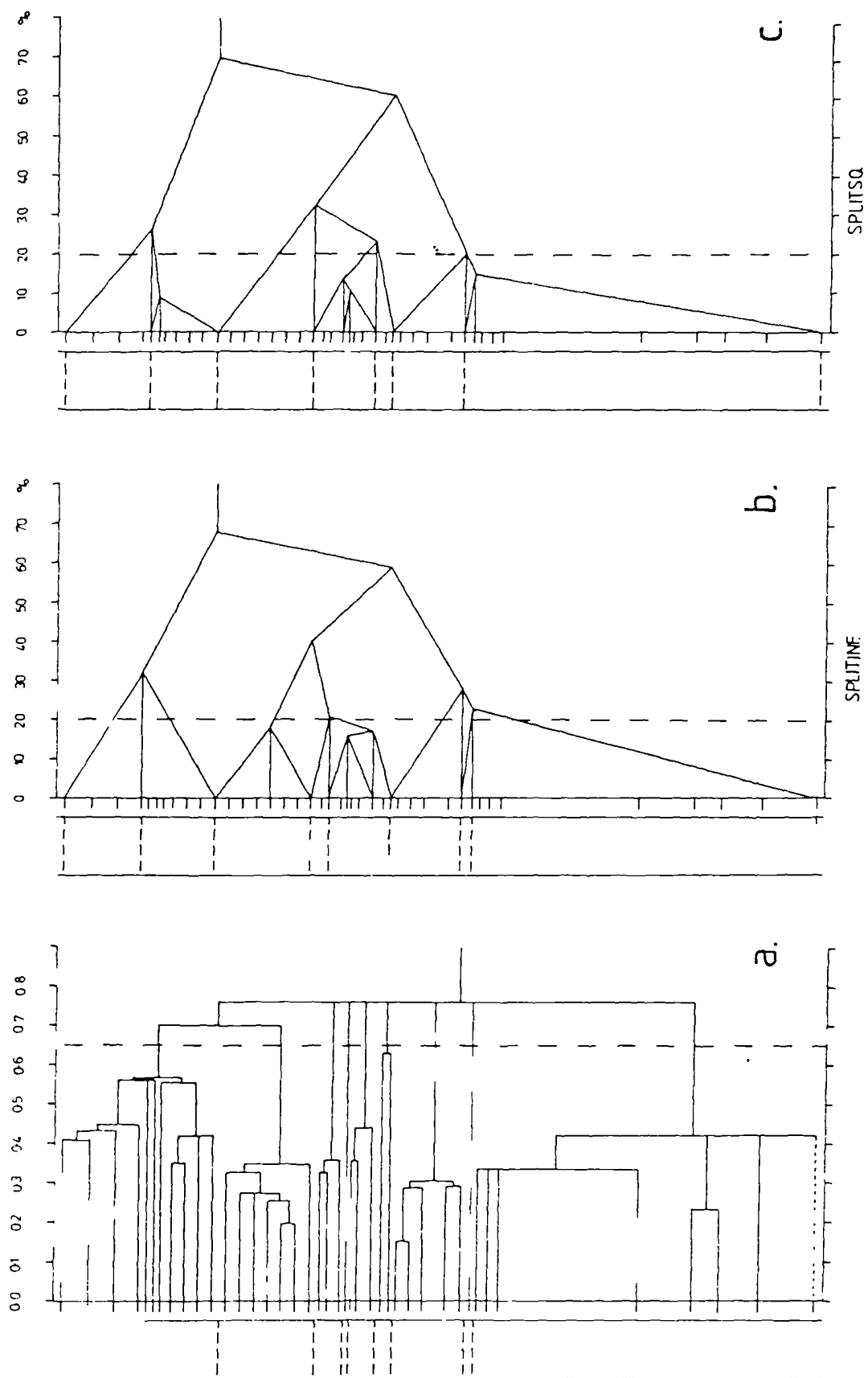


Fig.3.15 Numerical zonation dendrograms for vibrocore 57/-09/89.
(a) Conslink, (b) Splitinf, (c) Splittsq.

The important accessory species of this zone are *Elphidium asklundi* and *Nonion orbiculare*. The accessory species of zone 1, *Bulimina* gr. *marginata*, *Cassidulina laevigata*, and *Cibicides lobatulus*, as well as the less frequent taxa such as *Ammonia batavus* and *Uvigerina peregrina*, are greatly reduced or absent in this zone. Other commonly occurring taxa within zone 3 include *Islandiella helenae*, *I.norcrossi*, and *Nonion labradoricum*. The upwards decline of *E.asklundi* and *N.orbiculare* within this zone is well defined.

Decreasing numbers of species and faunal diversity up through the zone reach a minimum at 89/33 (2.55-2.6 m.) with the lowest faunal diversity (= 7 species) of the entire core. Faunal dominance shows an inverse relationship, increasing steadily up through the zone to a maximum value of 62.8% *Elphidium excavatum* forma *clavata* in sample number 89/33. Planktonic:benthonic ratios in this zone are at a minimum, reaching their lowest values throughout the core; ratios as low as 0.02 are recorded at sample level 89/33 and 89/37 (2.8-2.9 m.). Benthonic specimen numbers are also extremely low, ranging from an estimated minimum of 3,228 specimens per 100 g. of sediment at level 89/38 (2.9-3.0 m.) to a maximum of 6,495 specimens at level 89/35 (2.65-2.7 m.). Interestingly planktonic:benthonic ratios are at their greatest near the top and bottom of this zone, while benthonic specimen sums are at their lowest values at these same levels.

A radiocarbon date, OxA-2784, was obtained from the top of this zone.

<u>OxA-2784</u>	
depth:	2.5 - 2.55 m.
species:	<i>Portlandia arctica</i>
corrected age:	13,515 \pm 150 BP

3.9.3.d Zone 4

This zone is particularly well defined and occurs between 2.45 m. and 1.90 m.. The most important taxa in zone 4 are *Cibicides* gr. *lobatulus*, *Ammonia batavus*, and *Stainforthia*

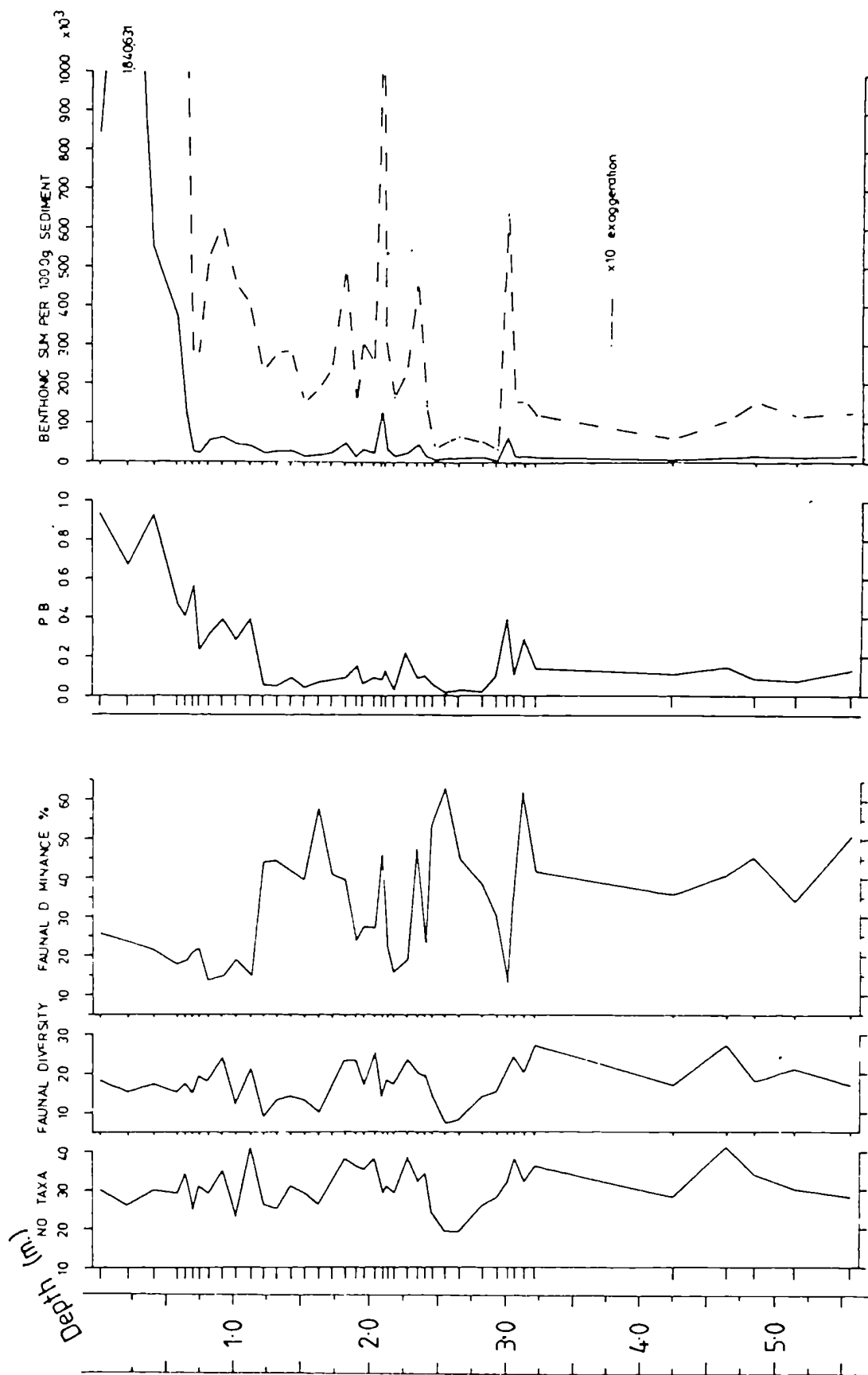


Fig.3.16 Faunal characteristics of vibrocore 57/-09/89.

fusiformis; the common accessory species include *Cassidulina laevigata*, *Bulimina* gr. *marginata*, *Rosalina praegeri*, *Spiroplectammia wrightii*, *Trifarina angulosa*, and *Astrononion gallowayi*. The dominant taxa of zones 1 and 3, *Cassidulina reniforme* and *Elphidium excavatum*, also occur but are considerably reduced; the latter species is no longer dominated by the cold water form *clavata* and includes specimens of the boreal form *selseyensis*. Grain size analysis of the core (fig.3.8) reveals two distinctive peaks in material $>63\ \mu\text{m}$. within zone 4, centred about levels 89/23 (1.94 - 2.0 m.) and 89/27 (2.15 - 2.25 m.), and corresponding maximum frequency peaks of *Ammonia batavus* and *Cibicides lobatulus* are noted at these levels. Below both these peaks, centred about levels 89/25 (2.11 - 2.13 m.) and 89/29 (2.35 - 2.4 m.), are finer grained sediments and corresponding peaks of *Stainforthia fusiformis*.

Faunal diversity is much increased from zone 3 and is variable about a mean of 19.5 species. Faunal dominance is also variable, but generally low, with the highest values at sample levels 89/25 and 89/29, with a minimum value of 18.6% *Cibicides lobatulus* at level 89/28 (2.25 - 2.35 m.).

It is likely that the high faunal dominances of this zone have been underestimated because of the taxonomic confusion surrounding early ontogenetic stages of *Stainforthia fusiformis*, many such specimens have mistakenly been assigned to *Stainforthia loeblichii*. Thus, at sample level 89/25, for example, an additional 8.3% of the fauna, assigned to *S. loeblichii*, may in fact belong with the 45.6% *S. fusiformis* and a maximum faunal dominance of 53.9% is possible. The taxonomic problems surrounding this species are dealt with in Appendix 1.

Planktonic:benthonic ratios are higher than zone 3, but remain low and variable about a mean zonal value of approximately 0.1; the maximum value of 0.22 is at level 89/28, with the minimum of 0.03 at 89/27. Roughly equal numbers of dextral and sinistral forms of *Neogloboquadrina pachyderma* occur in zone 4 and above this zone the dextral form dominates, while the sinistral form dominates lower down in the core. Benthonic sums are also higher than zone 3 values, with a maximum value of 125,753 specimens per

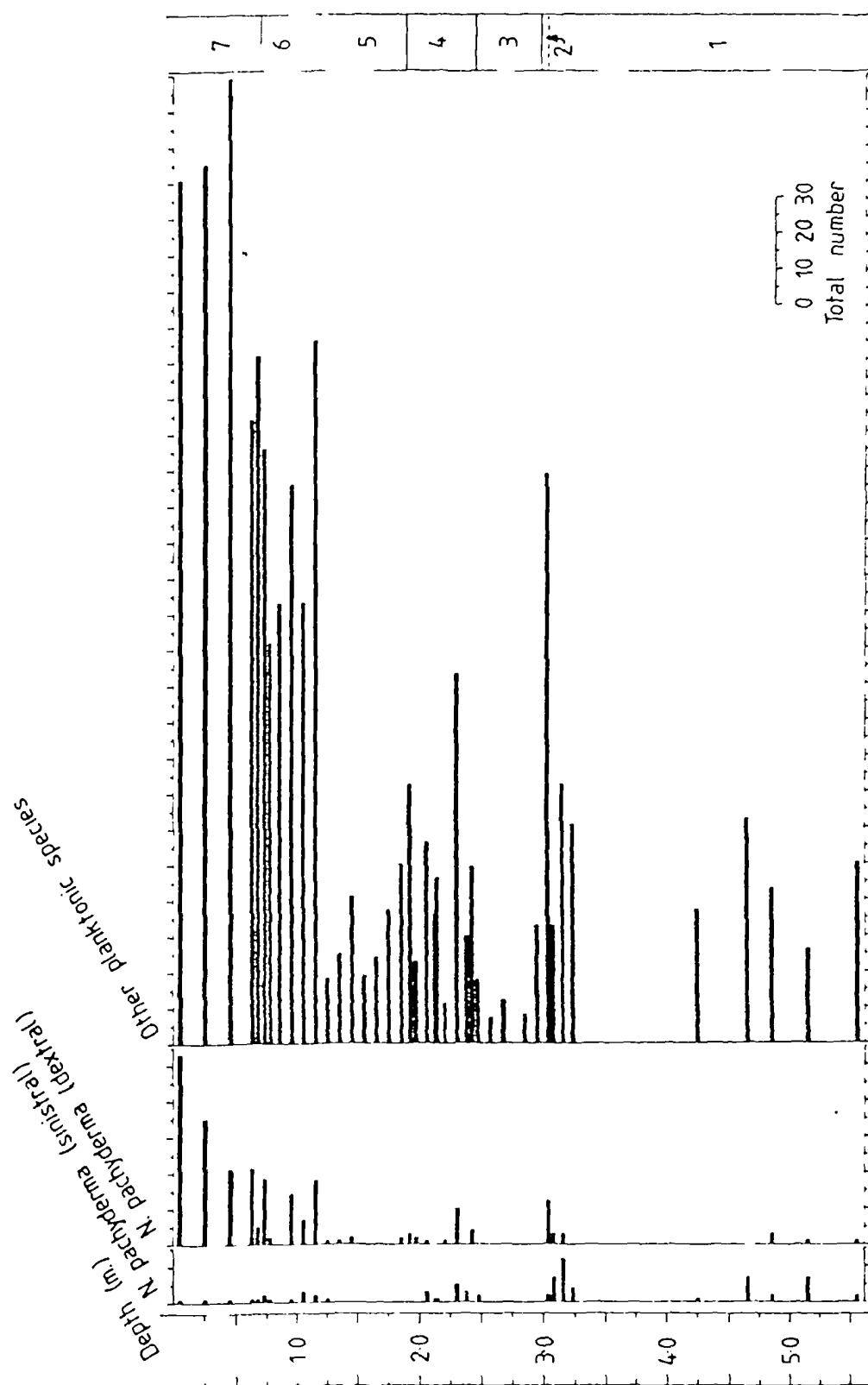


Fig.3.17 Planktonic foraminiferal frequency (count data) diagram of vibrocore 57/-09/89.

100 g. of sediment at level number 89/25.

Two radiocarbon dates, OxA-2783 and OxA-2782, were obtained from this zone.

OxA-2783

depth:	2.25 - 2.35 m.
species:	<i>Nucula nucleus</i>
corrected age:	11,625 \pm 130 BP

OxA-2782

depth:	1.94 - 2.0 m.
species:	<i>Parvicardium ovale</i>
corrected age:	11,035 \pm 130 BP

3.9.3.e Zone 5

Zone 5 is defined between 1.90 m. and 1.20 m. *Elphidium excavatum* forma *clavata* and *Cassidulina reniforme* return as the dominant species, the former dominating throughout the zone. The important accessory species are *Elphidium albumbilicatum*, *Islandiella helenae*, *I.norcrossi*, and *Nonion labradoricum*; the latter three tend to be more important towards the top of the zone, particularly *N.labradoricum* which is a characteristic feature of zone 6. A number of the taxa which help to define zone 4 are much reduced in this zone, these include *Ammonia batavus*, *Astrononion gallowayi*, *Bulimina* gr. *marginata*, *Cassidulina laevigata*, *Cibicides lobatulus*, *Stainforthia fusiformis*, *Spiroplectammina wrightii*, and *Trifarina angulosa*. The transition from zone 4 to zone 5 is marked and well defined as is the transition from zone 5 to 6.

Faunal diversity is generally low and reduced from zone 4, with a mean zonal value of 14.1 species, and a minimum value of 9 species at sample level number 89/15 (1.2-1.3 m.). Faunal dominance is conversly high, with a maximum 57.9% *Elphidium excavatum* forma *clavata* at level 89/19 (1.6-1.7 m.). Planktonic:benthonic ratios are low and reduced from zone 4, with

a mean zonal value of slightly less than 0.07. Benthonic sums remain low, with a mean zonal value of approximately 26,700 specimens per 100 g. of sediment.

3.9.3.f Zone 6

This zone is defined between 1.20 m. and 0.70 m.. The dominant species are *Elphidium excavatum* forma *clavata* and *Cassidulina reniforme*, the latter dominating throughout. The most important accessory species are *Nonion labradoricum*, *Cassidulina laevigata*, *Bolivina difformis*, *Bulimina* gr. *marginata*, *Cibicides* gr. *lobatulus*, *Rosalina praegeri*, and *Spiroplectammina wrightii*. Other species, such as *Hyalinea balthica* and *Uvigerina peregrina* exhibit marked increases from the base of this zone and continue to the top of the core.

Faunal diversity increases from zone 5 to 6, although values are increasingly variable. Faunal dominance is very much reduced from zone 5 and a mean zonal value of 23.5% is recorded. Planktonic:benthonic ratios show a marked increase to a mean zonal value of 0.36. The first marked increase in the number of dextral *Neogloquadina pachyderma* occur within this zone, together with declining numbers of the sinistral form. Benthonic sums are also higher in this zone and tend to increase upwards to a maximum value of 60,981 specimens at level 89/12 (0.9-1.0 m.).

A radiocarbon date, OxA-2781, was obtained from the top of zone 6.

<u>OxA-2781</u>	
depth:	0.70 - 0.75 m.
species:	<i>Nuculoma belloti</i>
corrected age:	10,635 \pm 120 BP

3.9.3.g Bioturbation within zone 6

An examination of the core between 1.20 m. and 0.70 m. reveals some large burrow structures infilled with coarse sandy sediment which is assumed to be derived from zone 7 above; these

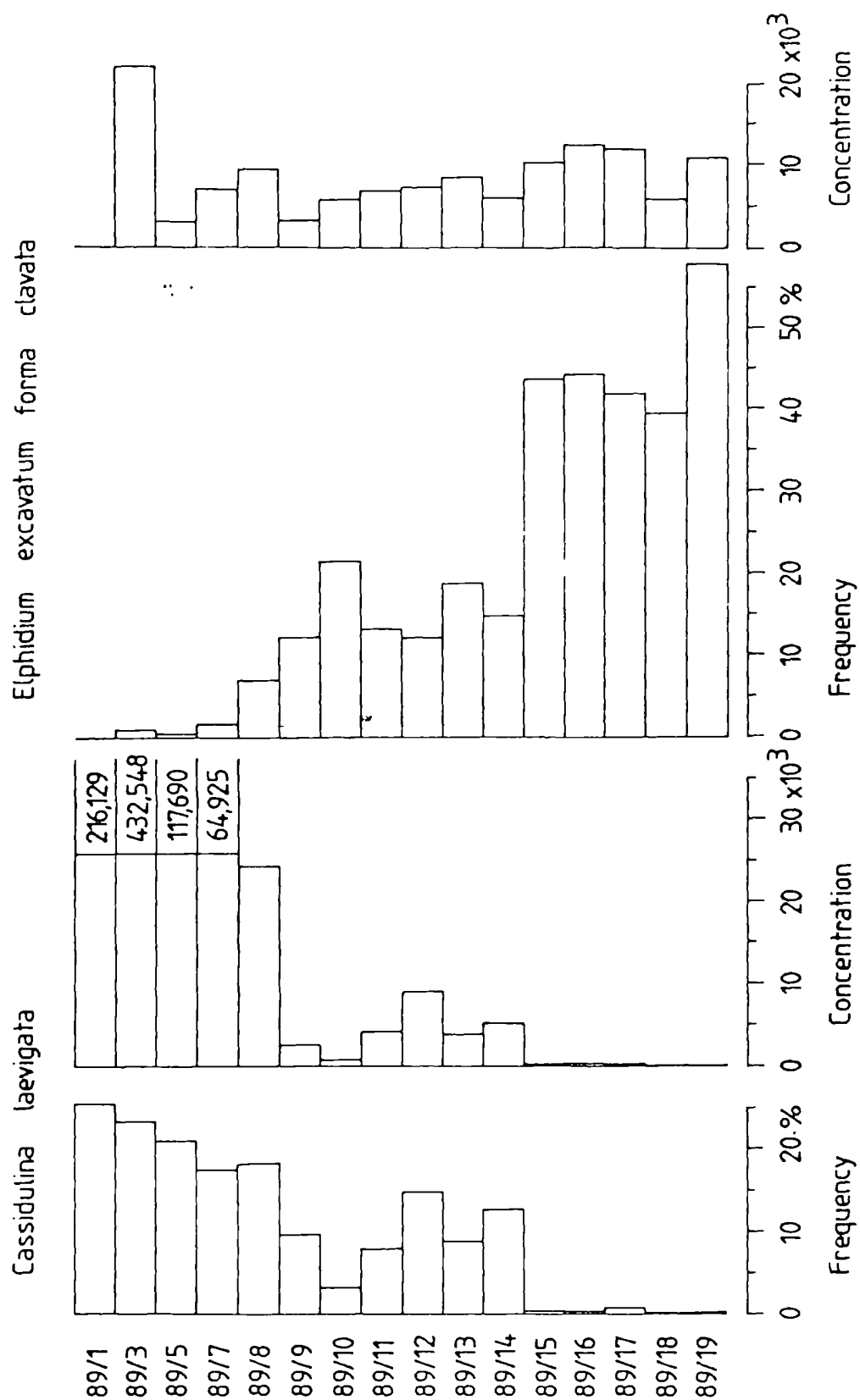


Fig.3.18 Percentage frequency and concentration (specimens per 100 g.) diagrams for *Cassidulina reniforme* and *Elphidium excavatum* forma *clavata* from zones 5,6 & 7 of vibrocore 57/-09/89.

structures stop abruptly at 1.20 m.. The marked differences in the benthonic sums from zones 6 to 7 are illustrated in fig.3.16, as are the differences from zones 5 to 6. Percentage frequency diagrams across these three zones exhibit two major steps as the taxa which characterize zone 7 increase at the zone 5/6 and then 6/7 boundaries. In an attempt to assess whether or not the faunas of zone 6 are largely the product of bioturbated sediment, brought down from zone 7 and mixed with faunas typical of zone 5, the concentrations of certain taxa have been calculated. The results are illustrated in fig.3.18.

An increase in the frequency of *Cassidulina laevigata* at the zone 6/7 boundary is associated with a major increase in concentration of this species within zone 7. However, the decline in frequency of *Elphidium excavatum* at the zone 5/6 boundary is not marked by any major change in concentration, although the decline in frequency at the zone 6/7 boundary corresponds to increased concentrations at some levels within zone 7.

3.9.3.h Zone 7

This is the uppermost zone and is defined from 0.70 m. to the top of the core. The dominant species is *Cassidulina laevigata*, with *Bolivina difformis*, *Bulimina* gr. *marginata*, *Cibicides* gr. *lobatulus*, *Hyalinea balthica*, *Rosalina praegeri*, *Spiroplectammina wrightii*, and *Trifarina angulosa* as the main accessory species. While all the above named taxa do exhibit marked increases in frequency, it is the decline of species belonging to the genera *Elphidium*, *Nonion*, and *Islandiella*, in particular, which define the boundary between zones 6 and 7.

Faunal diversity is moderately high, with a zonal mean of 16.4 species. Faunal dominance is low, but increases upwards to a maximum 25.6% *Cassidulina laevigata* in the surface sample. Planktonic:benthonic ratios continue to rise from zone 6 and a maximum value of 0.93 is recorded in the surface sample and at sample level 89/5 (0.4-0.5 m.). Benthonic sums are exceptionally high and continue to increase to an estimated maximum value of 1,840,631 specimens per 100 g. of sediment at level 89/3 (0.2- 0.3

m.).

A radiocarbon date, OxA-2780, was obtained from near the base of zone 7.

OxA-2780

depth: 0.5 - 0.6 m.

species: *Timoclea* sp.

corrected age: 5,555 \pm 90 BP

3.9.4 Oisein sequence

Vibrocore 56/-09/142 is the representative core of the Oisein sequence. Fifteen samples have been analysed and reveal a rather complex foraminiferal stratigraphy which is subdivided into two major zones. The faunal data from this core are summarized in figs.3.19a,b.

3.9.4.a Zone 1

This zone is defined as extending from the base of the core to an intermediate level between 3.83 m. and 3.11 m.. The dominant taxa are *Cassidulina reniforme* and *Elphidium excavatum* forma *clavata*, with *Cibicides* gr. *lobatulus*, *Nonion orbiculare*, and *Stainforthia loeblichii* as the main accessory species. The lowest level sampled (142/43) at 5.70 m. contains a more diverse fauna with *Rosalina praegeri*, *Hyalinea balthica*, *Bulimina* gr. *marginata*, and *Bolivina difformis* as the major additions.

Faunal diversity steadily increases through zone 1 and continues to do so until it peaks at 1.09 m. Faunal dominance decreases steadily upwards from a maximum of 52.3% *Elphidium excavatum* forma *clavata* at 5.46 m. Planktonic:benthonic ratios are low and, with the exception of level 142/43, vary little from the mean zonal value of 0.06. Equally, the benthonic sums remain low throughout this zone, particularly below 5.45 m., with a maximum value of 6,587 specimens per 100 g. at 4.76 m.

3.9.4.b Zone 2

Zone 2 is defined from between 3.83 m. and 3.11 m. upwards to the top of the core. A number of subzones might possibly be defined by the following samples:

subzone 2a comprises samples 142/31, 30, and 28. The dominant taxa, like zone 1, are *Cassidulina reniforme* and *Elphidium excavatum* forma *clavata*, with the main accessory species including *Bulimina* gr. *marginata*, *Cibicides* gr. *lobatulus*, *Rosalina praegeri*, *Spiroplectammina wrightii*, *Stainforthia fusiformis*, and *S.loeblichii*. Other commonly occurring taxa which make their first continuous appearance in the core within this subzone include

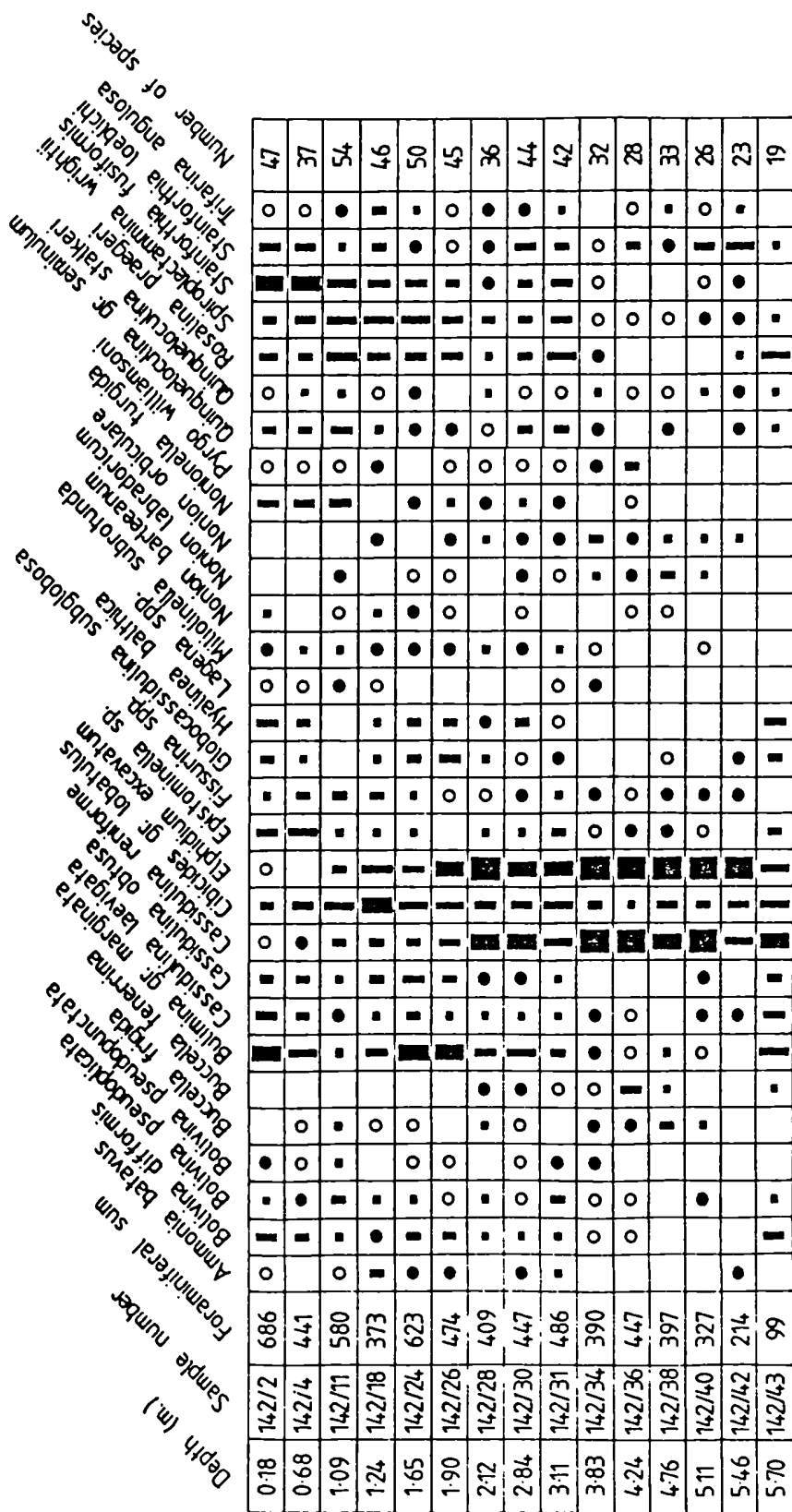


Fig.3.19.a Summary diagram of the 30 most commonly occurring benthonic foraminifera from vibrocore 56/-09/142.

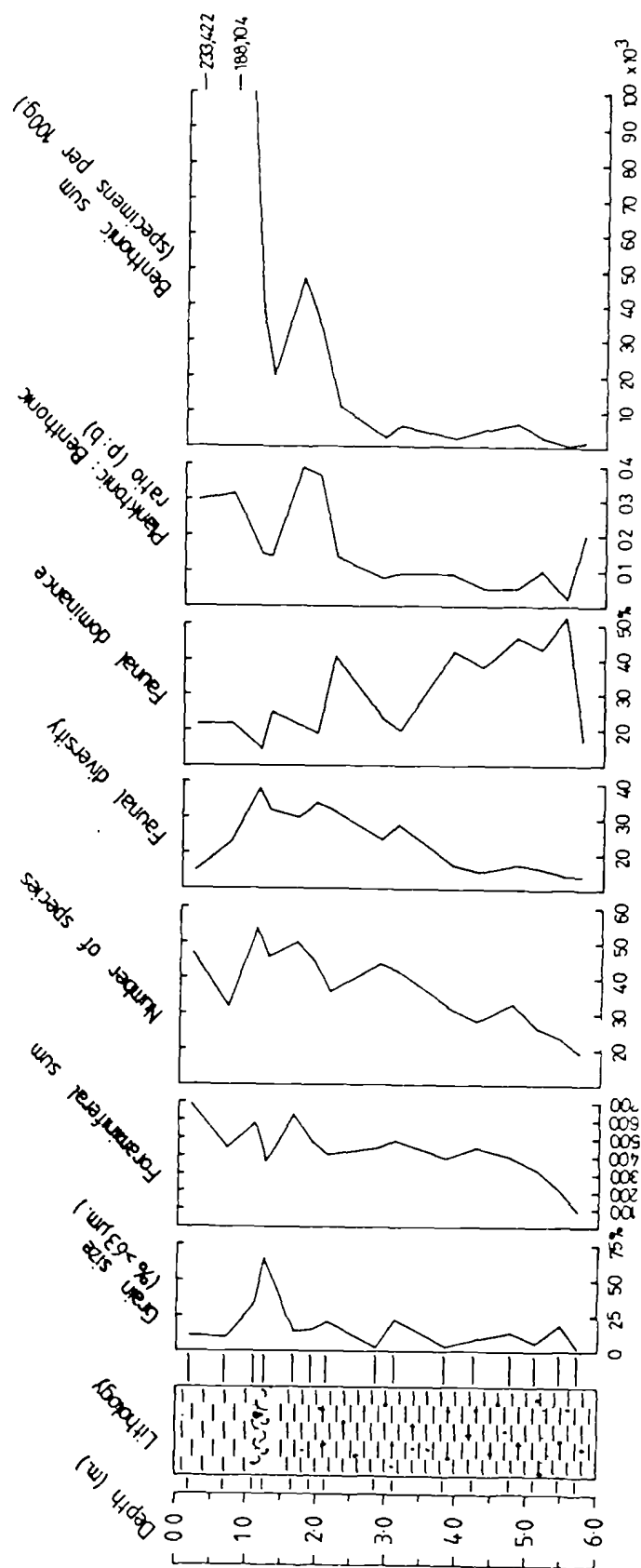


Fig.3.19.b Summary of the faunal characteristics of vibrocore 56/-09/142.

Ammonia batavus, *Cassidulina obtusa*, *Globocassidulina subglobosa*, *Hyalinea balthica*, and *Nonionella turgida*.

subzone 2b comprises samples 142/26, 24, 18, 11, 4, and 2 and extends from between 1.90 m. and 2.12 m. to the top of the core. The dominant taxa are *Bulimina* gr. *marginata*, *Cibicides* gr. *lobatulus*, and *Stainforthia fusiformis*; with *Bolivina difformis*, *Rosalina praegeri*, and *Spiroplectammina wrightii* as the main accessory species. Some interesting faunal features occur within this subzone, most notable being the upward decline of *Cassidulina reniforme* and *Elphidium excavatum* forma *clavata*, with a corresponding increase in *Nonionella turgida*, *Quinqueloculina* gr. *seminulum*, and *Cassidulina laevigata*. Faunas from sample levels 142/18 and 142/11, which correspond to a coarse, shelly lag deposit, are somewhat anomalous and have a particularly high content of *Cibicides lobatulus*.

Faunal diversities within zone 2 increase gradually to a maximum of 38 species at 1.09 m., falling sharply to a minimum of 15 species at 0.18 m. Faunal dominance is much reduced from zone 1, with only one anomalously high value of 40.8% *Elphidium excavatum* forma *clavata* at 2.12 m.; a subzone 2a mean of 28%, and a subzone 2b mean of 20.6% are recorded. Planktonic:benthonic ratios in subzone 2a are higher than zone 1, with a mean of 0.1, but are markedly increased in subzone 2b which has ratios up to 0.39 at 1.65 m. However, these values are considerably reduced in levels 142/18 and 11. Benthonic sums increase in subzone 2a, but again are highest in subzone 2b, particularly at 0.18 m. which has an estimated maximum 233,422 specimens per 100 g. of sediment.

3.9.5 Fionn sequence

Two cores were analysed from this sequence, vibrocores 57/-09/44 and 57/-09/46, both of which were cored from within the St. Kilda Basin.

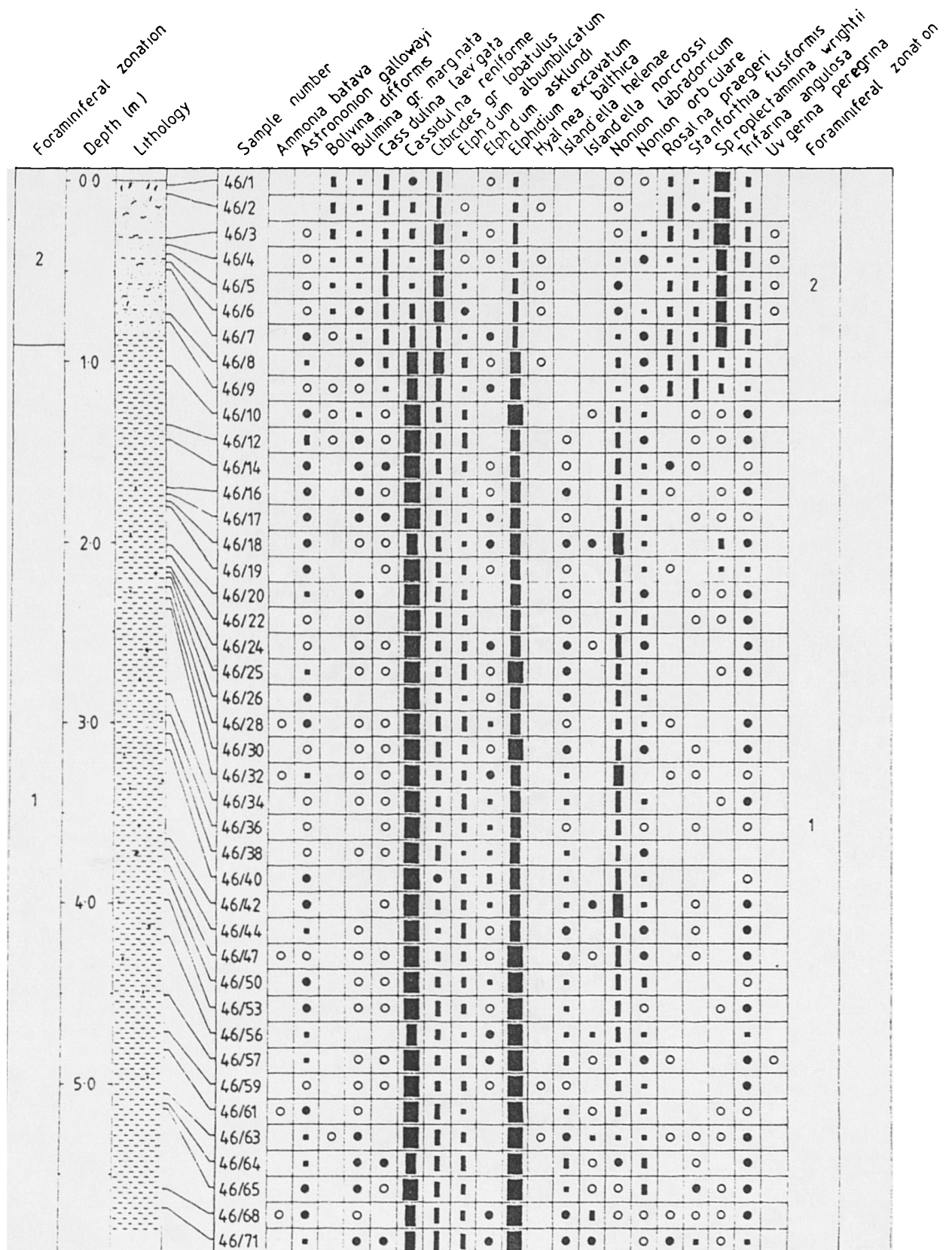


Fig.3.20 Summary diagram of the 20 most commonly occurring benthonic foraminifera from vibrocore 57/-09/46

3.9.5.a Vibrocore 57/-09/46

This 5.79 m. long vibrocore was recovered from a depth of -156 m. OD at 57°19.30'N, 08°30.04'W. Two foraminiferal zones are recognised, the upper one is divided into 4 sub-zones; these subdivisions have been defined by numerical zonation techniques. Full foraminiferal diagrams (enclosures 3 and 4) are included in the document wallet at the end of this volume. Summary benthonic and planktonic diagrams are illustrated in fig.3.20 & fig.3.21 respectively, while faunal parameters are summarized in fig.3.22.

3.9.5.a1 Zone 1

Zone 1 is defined from the base of the core (5.79 m.) to the gap between 1.0 m. and 0.82 m.. The dominant species of this zone are *Cassidulina reniforme* and *Elphidium excavatum* forma *clavata*; the latter tending to decrease upwards through the zone, while the former increases. The main accessory species are *Cibicides* gr. *lobatulus*, *Elphidium albiumbilicatum*, and *Nonion labradoricum*, while commonly occurring taxa include *Astrononion gallowayi*, *Elphidium asklundi*, *Islandiella helenae*, *Nonion orbiculare*, and *Trifarina angulosa*. Possibly the most striking feature of zone 1 is the steady increase in frequency of *Nonion labradoricum* above 5.0 m. to a maximum of 25.7% at 1.77 m..

Faunal diversities show a steady upwards decline through this zone, with the maximum value of 30 species at 5.58 m. and a minimum of 8 species at 2.20 m.. Faunal dominance is more variable, but also exhibits a slight upwards decline, with a maximum 53.5% *Cassidulina reniforme* at 2.08 m. and a mean of approximately 30%. Planktonic:benthonic ratios remain low throughout, with a mean ratio near 0.1. Sinistral coiling forms of the planktonic species *Neogloboquadrina pachyderma* dominate throughout the zone. Benthonic sums are variable, slightly higher at about 2.0 m., but typically about 20,000 specimens per 100 g. of sediment.

Two radiocarbon dates, OxA-2788 and OxA-2787, were obtained from zone 1. OxA-2788 was run as a check on OxA-1324 which yielded a reservoir corrected age of 11,275 ± 250 BP (see Table 3.2).

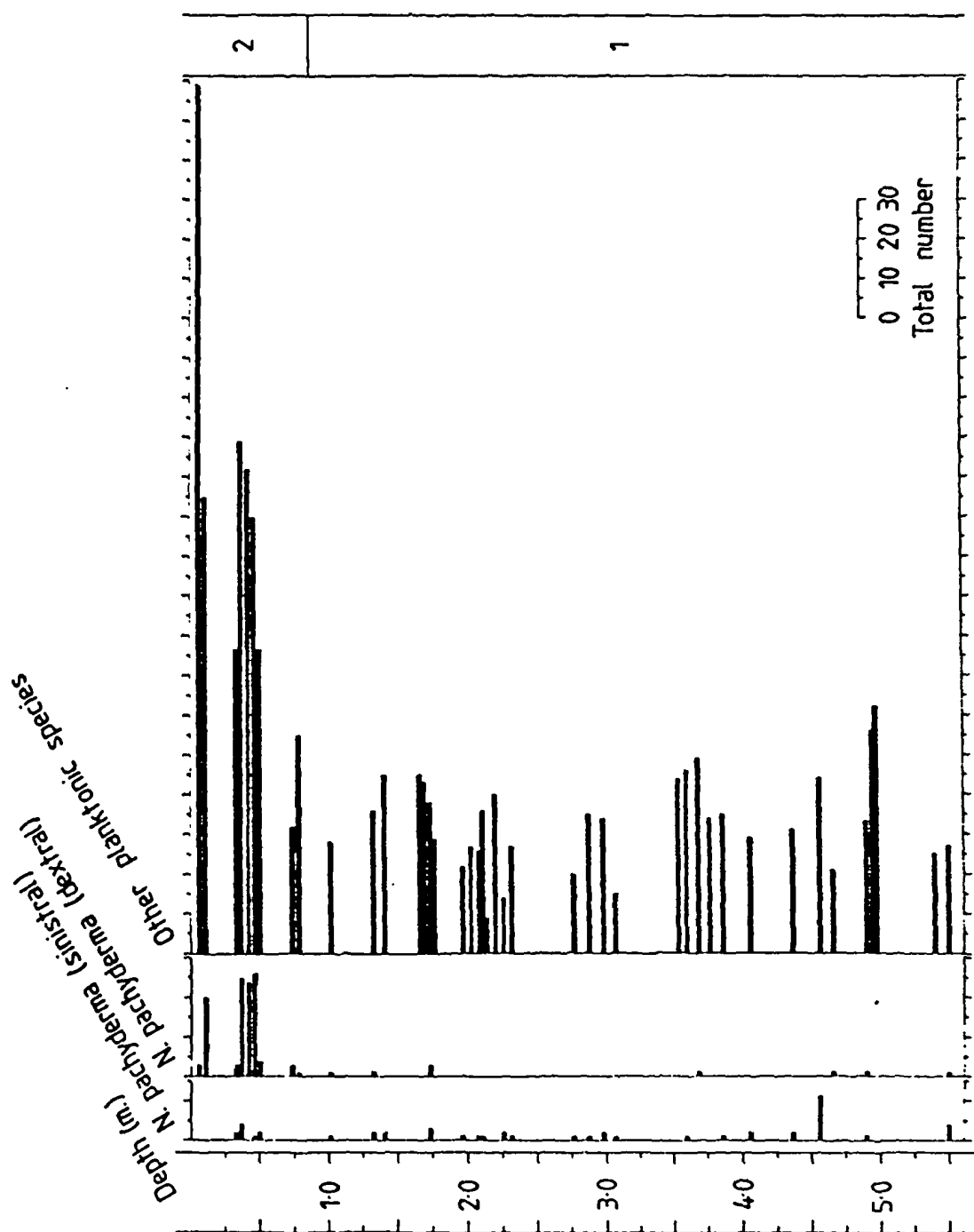


Fig.3.21 Planktonic foraminiferal frequency (count data) diagram from vibrocore 57/-09/46.

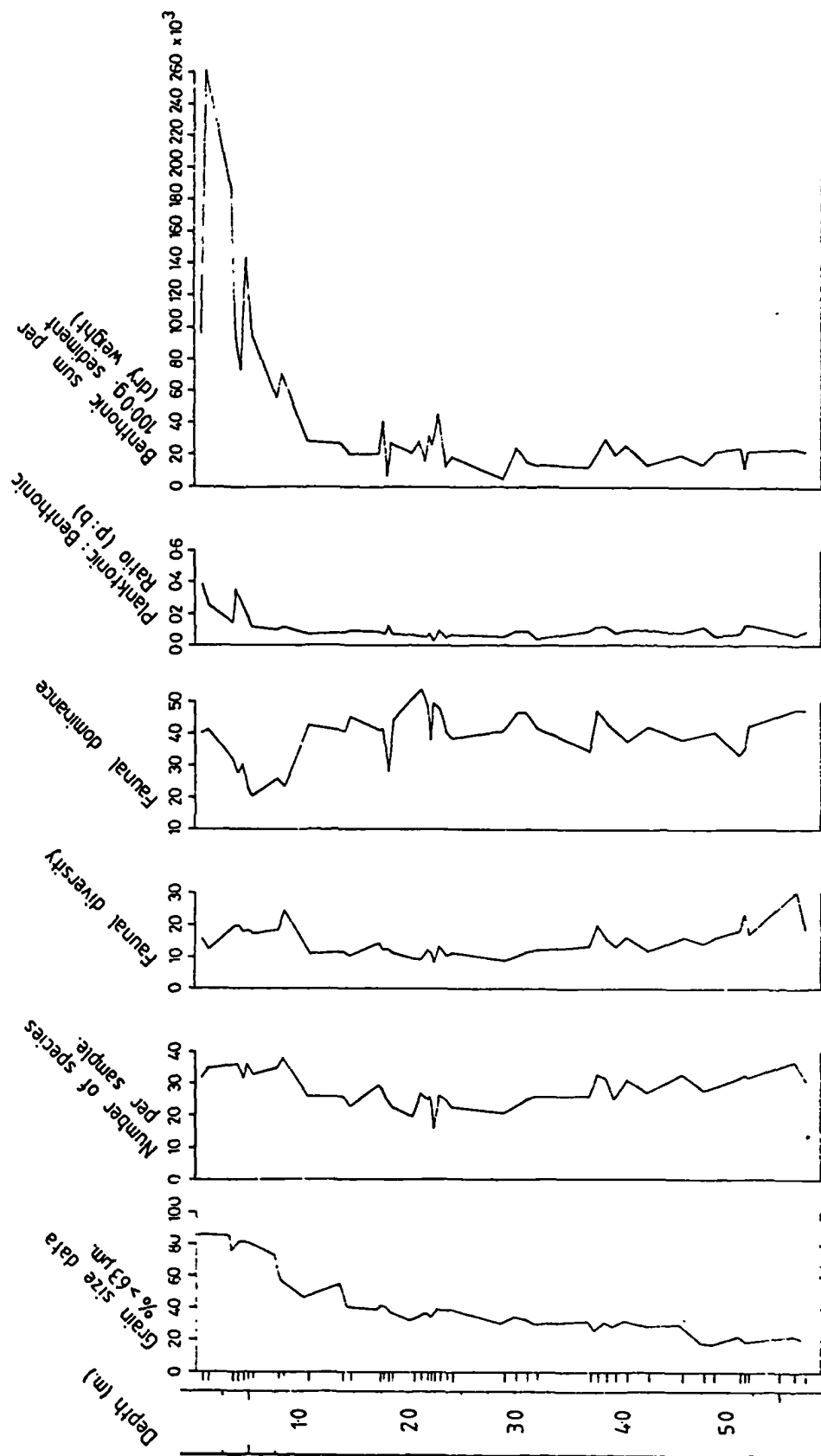


Fig.3.22 Summary of the faunal characteristics of vibrocore 57/-09/46.

OxA-2788	
depth:	4.8-5.0 m.
species:	<i>Nuculoma belloti</i>
corrected age:	11,015 \pm 130 BP

OxA-2787	
depth	1.05-1.3 m.
species	<i>Nuculoma belloti</i>
corrected age:	10,175 \pm 110 BP

3.9.5.a2 Zone 2

The upper 1.0 m. section of this core had three gaps in it, from where all the available sediment had been removed prior to the present sampling programme. The boundary between zones 1 and 2 is placed within one of these artificial stratigraphic gaps as are two of the sub-zone boundaries; the context of the zonal boundaries and stratigraphic gaps are outlined in fig.3.23.

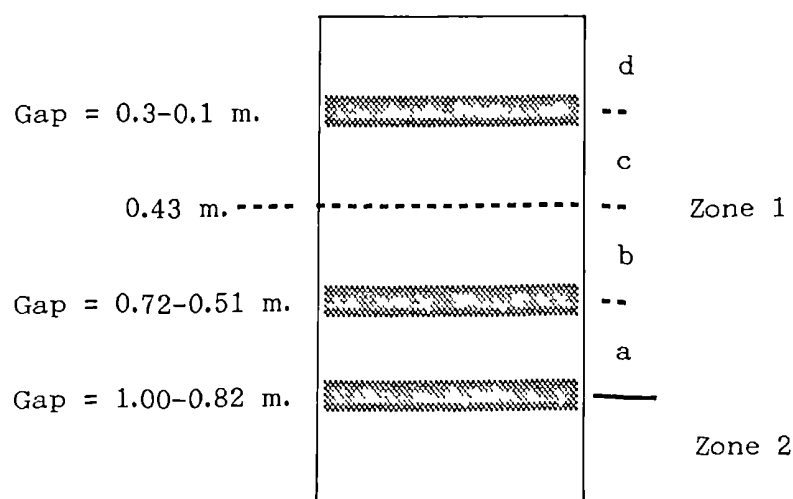


Fig.3.23 Sketch log of artificial stratigraphic breaks and foraminiferal zonation in VE 57/-09/46

Subzone 2a is defined by two samples, 46/8 and 46/9, between 0.82 m. and 0.72 m.. The dominant taxa are *Cassidulina reniforme* and *Elphidium excavatum* forma *clavata*, as in zone 1 below, but now *Cibicides* gr. *lobatulus*, *Rosalina praegeri*, and *Stainforthia fusiformis* are greatly increased in frequency. Other commonly occurring accessory species which clearly define the zone 1-subzone 2a boundary are *Spiroplectammina wrightii*, *Cassidulina laevigata*, and *Trifarina angulosa*.

Faunal diversity has a mean value of 21 species, while faunal dominance has a mean value of 25% *Elphidium excavatum*. The mean planktonic:benthonic ratio is 0.1 and is little different from zone 1; numbers of dextral coiling *Neogloboquadrina pachyderma* remain low. Benthonic sums are markedly higher than those of zone 1, with an estimated mean value of 64,000 specimens per 100 g. sediment.

Subzone 2b is again defined on the basis of two samples, 46/7 and 46/6, between 0.51 m. and 0.43 m.. The dominant taxon is *Spiroplectammina wrightii*, with *Cibicides* gr. *lobatulus*, *Elphidium excavatum*, *Cassidulina reniforme*, and *C. laevigata* as the main accessory species. Other commonly occurring taxa are *Rosalina praegeri*, *Stainforthia fusiformis*, and *Trifarina angulosa*. Particularly valuable in defining the subzone 2a-2b boundary are the continuing decline of *Cassidulina reniforme*, and *Elphidium excavatum*, together with a marked increase in *Spiroplectammina wrightii* and *Cassidulina laevigata*.

Mean faunal diversity in this subzone is 17.5 species. Mean faunal dominance is 21.9% *Spirolectammina wrightii*; faunal dominance is at its lowest throughout the core within this subzone. Planktonic:benthonic ratios are increased from subzone 2a and have a mean ratio of 0.16. Numbers of dextral coiling *Neogloboquadrina pachyderma* are greatly increased in this subzone and dominate to the top of the core. Benthonic sums are also higher, with an estimated mean subzone value of 111,478 specimens per 100 g. sediment.

A radiocarbon date, OxA-2786, was obtained from subzone 2b.

OxA-2786

depth: 0.47-0.51 m.

species: *Acanthocardia echinata*

corrected age: 9,975 \pm 110 BP

Subzone 2c is represented by three samples, 46/5, 46/4, and 46/3, between 0.43 m. and 0.30 m.. The dominant taxa are *Spiroplectammina wrightii* and *Cibicides* gr. *lobatulus*, with the common accessory species including *Cassidulina laevigata*, *Elphidium excavatum*, *Rosalina praegeri*, and *Trifarina angulosa*. Other taxa which help to define the upper boundary of this subzone include *Astrononion gallowayi* and *Uvigerina peregrina*.

Faunal diversity remains near the mean value of 18.7 species throughout this subzone, while faunal dominance has increased from subzone 2b, with a mean value of 30.5% *Spiroplectammina wrightii*. Planktonic:benthonic ratios are variable, but still increasing upwards, with a mean subzone ratio of 0.25. The same is true of the benthonic sum which has an estimated mean value of 118,700 specimens per 100 g. sediment.

Subzone 2d is defined by samples 46/2 and 46/1, and represents the upper 10 cm. of the core. The dominant taxon is *Spiroplectammina wrightii* and the main accessory species include *Cibicides* gr. *lobatulus*, *Cassidulina laevigata*, and *Rosalina praegeri*. Other commonly occurring taxa include *Bolivina difformis*, *Elphidium excavatum*, and *Trifarina angulosa*.

Faunal diversity is reduced from subzone 2c, with a mean of 13.5 species. Faunal dominance continues to rise and reaches its maximum value for zone 2 of 42% *Spiroplectammina wrightii* at 0.075 m.. Planktonic:benthonic ratios remain high with a ratio of 0.38 in the surface sample. Sinistral coiling forms of *Neogloboquadrina pachyderma* are entirely absent from this subzone. Benthonic sums remain high and peak with an estimated 297,714 specimens per 100 g. sediment at 0.075 m..

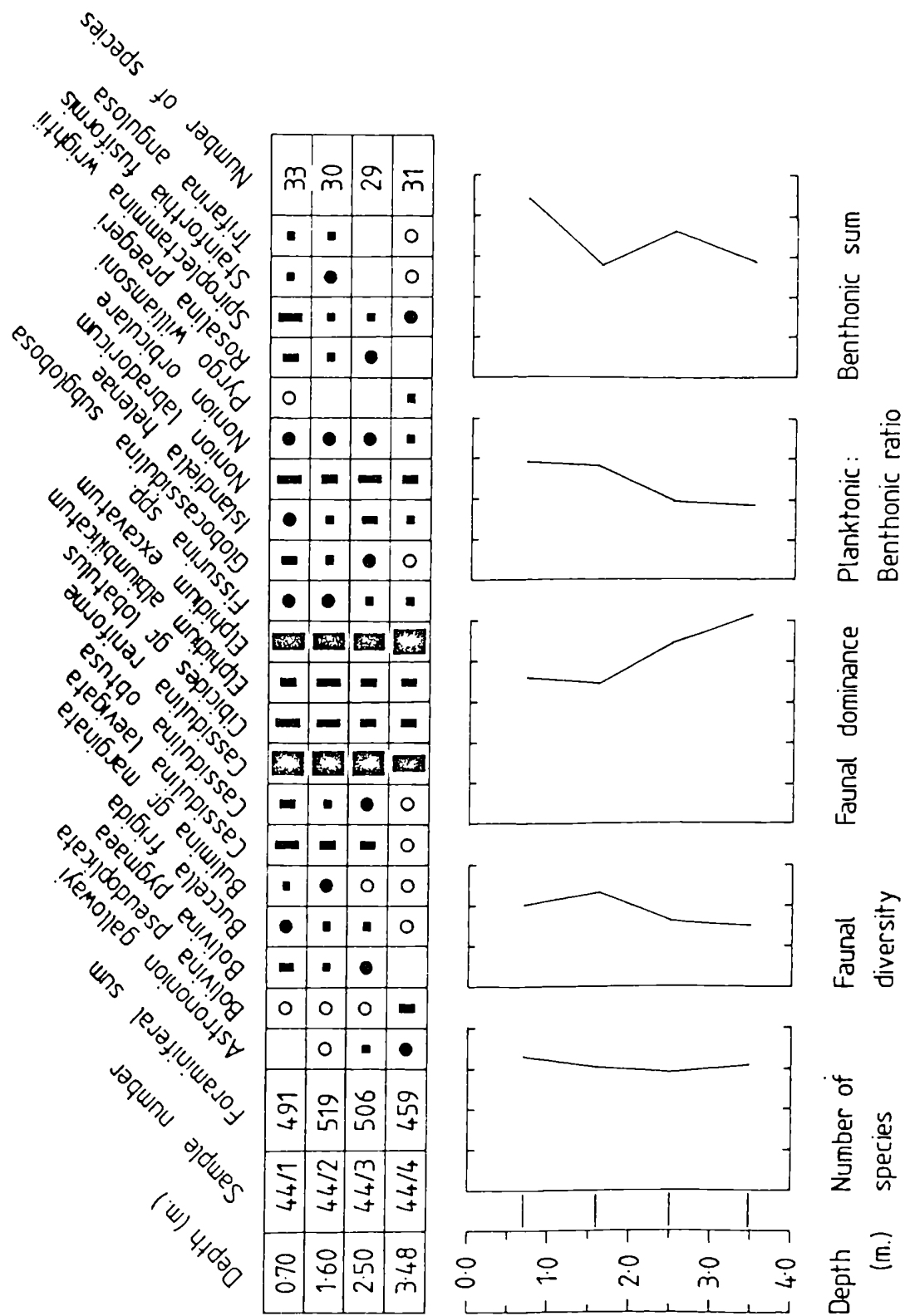


Fig.3.24 Summary diagram of the 21 most commonly occurring benthonic foraminifera from vibrocore 57/-09/44. [Correction: *Bolivina pygmaea* should read *Bolivina difformis*].

3.9.5.b Vibrocore 57/-09/44

No faunal subdivision of this core was undertaken and four samples only were analysed. Summary results of the foraminiferal analyses are presented in figs.3.24.

The dominant taxa are *Cassidulina reniforme* and *Elphidium excavatum* forma *clavata*, the latter dominating towards the base of the core and *C. reniforme* above 2.60 m.. The main accessory species are *Cibicides* gr. *lobatulus*, *Elphidium albiumbilicatum*, and *Nonion labradoricum*; with *Bolivina difformis*, *Cassidulina laevigata*, and *Spiroplectammina wrightii* as commonly occurring taxa and increasing in frequency towards the top of the core.

Faunal diversity shows a slight upwards increase with a mean value of 18.5 species, while faunal dominance tends to decrease upwards from a maximum of 51.4% *Elphidium excavatum* forma *clavata* at 3.48 m. to a minimum 34.3% *Cassidulina reniforme* at 1.60 m.. Planktonic:benthonic ratios are generally low, but increase towards the top of the core, the mean ratio is 0.11. The benthonic sums also show an up-core increase to an estimated maximum of 110,475 specimens per 100 g. at 0.70 m., and an estimated mean value of 86,000 specimens per 100 g.

3.10 Discussion

The interpretation of the foraminiferal results, again following Selby's (1989) seismic sequence scheme, are based upon the combined lithological, faunal, and dating results.

3.10.1 Conchar sequence

No independent dating evidence exists for core VE 57/-10/17, although this sequence does account for a large proportion of the sedimentary succession underlying the sea bed on the middle and outer shelf. Zone 1 faunas are dominated by boreo-arctic species and, together with muddy diamictos exhibiting randomly orientated clasts, imply a glacial marine environment of deposition. The low planktonic:benthonic (p:b) ratios imply lowered sea level and a considerable ice cover which acts to reduce the number of

planktonic specimens reaching the shelf break zone. However, the presence of large numbers of *Cassidulina reniforme*, together with species of the genus *Islandiella*, imply water depths >30m. The increasing frequency of *Nonion labradoricum* and *C. reniforme* upwards through the zone suggest deepening water, as seen in zone 1 of VE 57/-09/46 for example, but do not equate directly to the decreasing p:b ratios.

Increasing benthonic sums may suggest a decrease in the sedimentation rate upwards through the zone and/or increased benthonic productivity. The lithological unconformity between zone 1 and zone 2 is reflected in the foraminiferal faunas. Zone 2 faunas represent accumulation under temperate conditions, generally increased water depths and increased current activity. The gradational faunal changes, particularly at the base of zone 2, which contain arctic species characteristic of zone 1 are not considered to represent gradational climatic amelioration, but rather the reworking of zone 1 sediments and their inclusion within the coarse sands of zone 2. Thus, zone 2 deposits are interpreted as early Holocene in age, while those of zone 1 are considered to represent late Devensian ice rafted deposits possibly accumulating during deglaciation after the last glacial maximum and hence preserving a record of rising eustatic sea level. However, the length of time represented by the hiatus and hence the age of zone 1 deposits remains unknown.

3.10.2 Conan A subsequence

3.10.2.a Vibrocore 57/-09/60

Zone 1 faunas bear an affinity to the faunas described from vibrocore 57/-10/21. Arctic species dominate but with an increased number of shallow water taxa compared to zone 1 of vibrocore 57/-10/17 from the shelf-break zone, but fewer than zone 3 of vibrocore 57/-09/89 from the morainal banks of the middle shelf. Comparison of planktonic:benthonic ratios reveals a similar pattern, with a lower mean ratio in this core than in 57/-10/17, and an even lower mean ratio in zone 3 of 57/-09/89. This is

interpreted as due to increasing distance from the shelf-break and hence reduced transportation of planktonic tests (cf. Murray, 1976) during the late Devensian.

Zone 2 faunas are considered to represent ameliorated climatic conditions, much like those of the present day, with much greater water depths than zone 1. The extremely high p:b ratios at the top of this core are unusual and considerably higher than those of VE 57/-10/17. The lower ratios of the latter, although situated west of the shelf-break, may be due to shelf edge currents (see section 3.3) and their possible winnowing effect, implying that shelf currents are weaker over this site and that planktonic tests are able to settle here. The transitional faunas at the zonal boundary are considered to be partly the product of reworked zone 1 material incorporated into zone 2 sediments during the early Holocene when sea levels (Fairbanks, 1989) were lower and M2 tidal streams probably greatly increased (Austin, 1991). These mixed faunas are not considered to record a gradual climatic amelioration, but may possibly reflect increasing Holocene sea levels.

Sidewall contamination in vibrocore 57/-09/60

The major unconformity within this core, both lithological (fig.3.11) and faunal (figs.3.12a,b), suggested itself as a useful zone from which to test for sidewall contamination as a result of the vibrocoreing process. As discussed in chapter 2, sub-samples from these cores are taken from the 'inner' part, the outer 1cm. around the casing is left unsampled. The reasons for doing this, as I hope to demonstrate below, are that this 'outer' part of the core contains a disturbed stratigraphy, largely the product of downward sediment smearing. This disturbed outer part of the core, in this case vibrocore 51/-07/199 from the central Celtic Sea, is well illustrated in fig.3.25; this photograph also illustrates another phenomenon associated with disturbed stratigraphy ie. bioturbation.

In an attempt to assess the degree of side-wall contamination, sub-samples from both the 'inner' and 'outer' part of the same stratigraphic level were analysed. The results are



Fig.3.25 Photograph showing the disturbed 'outer' zone and bioturbation within fine sands and silts of vibrocore 51/-07/199 from the central Celtic Sea.

summarized for the five most commonly occurring as well as the planktonic:benthonic ratios, and these are illustrated in fig.3.26. The lithological change takes place at a depth of 0.24 m., then the foraminiferal assemblages of the 'inner' samples closely reflect this change too. The planktonic:benthonic ratios, the frequency of *Spiroplectammina wrightii*, *Cassidulina laevigata*, and *Trifarina angulosa* all exhibit a marked decline from 'inner' sample levels 0.22 - 0.24 m.. The species *Elphidium excavatum* and *Cassidulina reniforme*, which dominate zone 1 of this core, exhibit a steady decrease upwards across these levels, but decline sharply above 0.22 m.. The 'outer' samples exhibit far less distinctive faunal changes than do the 'inner' samples, this is clearly visible from the 'tails' of the species *Spiroplectammina wrightii*, *Cassidulina laevigata*, and *Trifarina angulosa*, as well as the planktonic:benthonic ratios. The species *Elphidium excavatum* and *Cassidulina reniforme* are much reduced in the 'outer' samples in comparison with the 'inner' samples above 0.24 m.; with the exception of *elphidium excavatum* within level 0.16-0.18 m., where the dominant form is *E. excavatum* forma *selseyensis* and not the arctic form *clavata* which dominates zone 1.

These faunal characteristics clearly demonstrate the downcore smearing of material in contact with the core casing. The 'outer' samples of zone 1 preserve much of the faunal signal of the zone 2 faunas, while the 'inner' samples at the base of zone 2 bear a closer resemblance to zone 1 'inner' samples than do the 'outer' ones at the base of zone 2. Thus the sharp lithological unconformity is best reflected by analysing the 'inner' samples; when outer samples are analysed long 'tails' of temperate species are seen within zone 1, predating their real first occurrence within zone 2. This has important implications for the interpretation of this boundary which is considered to represent a major unconformity below the base Holocene, rather than the gradual climatic amelioration that the 'outer' samples imply. It also emphasises the need to look at faunal and lithological changes together. The tail of zone 1 species within the 'inner' samples of zone 2 must be interpreted in a different manner; and it is suggested that they represent derived zone 1 material which

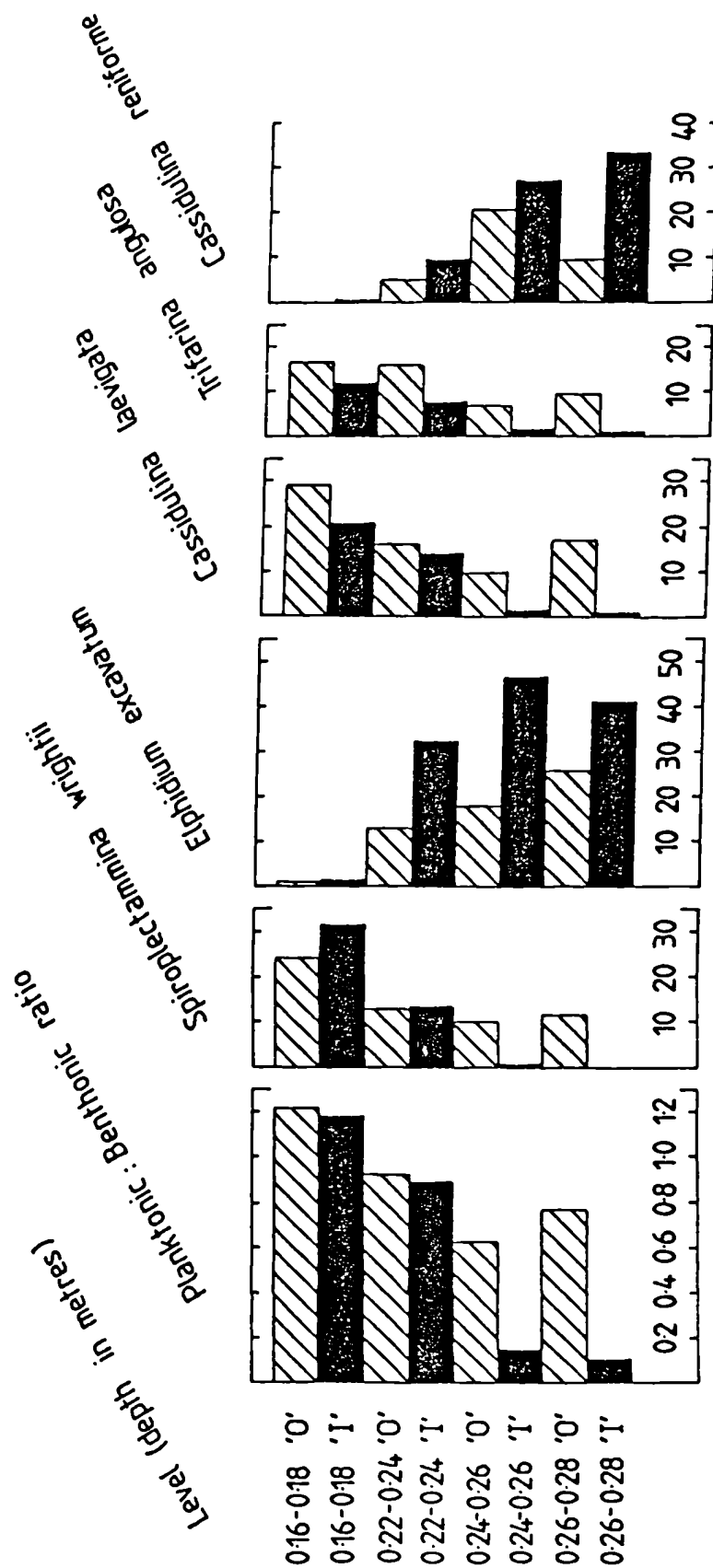


Fig.3.26 Percentage frequency diagram of selected species and the p:b ratio from the upper part (^a/4) of vibrocore 57/-09/60 ('O' = outer sample, 'I' = inner sample).

is being eroded and reworked into zone 2 faunas. In the interpretation of the latter it must be remembered that the arctic species are derived and that the apparent upward amelioration is a reflection of this. Infact, none of the Holocene sediments analysed show any signs of climatic change, other than those reflecting changing relative sea level.

3.10.2.b Vibrocore 57/-10/21

The radiocarbon date of $22,480 \pm 300$ years BP from a depth of 4.30-4.55 m. in this core implies accumulation of the sediments just prior to the late Devensian glacial maximum. The faunas do not reflect any major change in the depositional environment, which is that of a generally shallow c. 30 m. glacial marine environment. High faunal dominance and generally low benthonic sums suggest harsh environmental conditions, but not markedly reduced salinities, and together with low p:b ratios indicate extensive ice cover. The faunas of this core closely resembled those of zone 1 in VE 57/-09/60 and it seems reasonable to assume deposition under similar conditions.

3.10.3 Conan B subsequence

The interpretation of the sequence present in vibrocore 57/-09/89 must be viewed in the context of the core location near a morainal bank complex. Furthermore, the foraminiferal stratigraphy clearly demonstrates an acceptable climatostratigraphic sequence in terms of known Lateglacial climatic events and these are supported by radiocarbon dates.

Zone 1 faunas are dominated by boreo-arctic species, but also contain some temperate faunal elements, often showing signs of abrasion. The high faunal diversities of zone 1 reflect the mixed nature of the assemblages present, while the relatively high faunal dominances are interpreted as reflecting the *in situ* conditions. Thus, the dominant faunal elements reflect relatively shallow, slightly reduced salinity, glacial marine conditions and together with moderately high allochthonous temperate faunal elements suggest ice proximal conditions, if not even lodgement facies. The radiocarbon date of $15,245 \pm 170$ years BP confirms ice

at or near the morainal bank complex of the middle shelf at this pleniglacial date.

Zone 2 is representative of an anomalous level within the core, where the boreo-arctic taxa of zone 1 are much reduced and warmer water temperate species are increased in frequency. The latter species, *Ammonia batavus*, *Bolivina difformis*, and *Rosalina praegeeri* in particular, are absent from adjacent levels in zones 1 and 3. This level also corresponds to a relatively thin horizon of medium sand and is tentatively interpreted as an originally frozen sand-pod of reworked temperate sediment; the reduced frequencies of boreo-arctic taxa suggest some mixing with the enclosing glacial marine diamicts. The possibility of bioturbation at this core depth seems unlikely in view of the distinctly cold-water faunas above and below. Zone 2 is therefore not considered to represent a valid palaeoclimatic signal.

Zone 3 faunas represent a continuation of zone 1 conditions but lack the apparently temperate species within the mixed assemblages of the latter. Salinity may also have increased in this zone, with the low salinity species *Elphidium albiumbilicatum* much reduced and the normal salinity indicator, *Cassidulina reniforme*, increased. Thus, the faunas suggest a more distal environment with fewer reworked temperate elements, but no marked increase in water depth since the cold, shallow species *Elphidium asklundi* and *Nonion orbiculare* are particularly well developed in this zone. The radiocarbon date of $13,515 \pm 150$ years BP at the top of this zone indicates that relatively shallow (c. 30 m.) arctic waters with icebergs (ie. dropstones) existed over this part of the middle shelf at a time when eustatic sea level was at about -105 m. OD (Fairbanks, 1989) and climate was beginning to get warmer (see chapter 1).

Zone 3 is interpreted as a period of increasingly distal, but relatively shallow glacial marine deposition associated with the eastwards retreat of once grounded ice from the middle shelf. The extremely low benthonic sums and planktonic:benthonic ratios of this zone are interpreted as due to increasing sedimentation rates, possibly from sediment rich waters rather than from floating ice as in zone 1; this would then explain the general

absence of clasts interpreted as dropstones within this zone. The low faunal diversity and high faunal dominance of this zone agree with a highly turbid depositional environment.

Zone 4 represents very different depositional and environmental conditions to those of zones 1 to 3. A major hiatus is also indicated within the core at the zone 3-4 boundary, with radiocarbon dates of $13,515 \pm 150$ years BP at 2.5-2.55 m. and $11,625 \pm 130$ years BP at 2.25-2.35 m.. It seems unlikely, in view of the sudden lithological and faunal changes at this boundary, that nearly 1,900 years of accumulation are represented by an interval of only 15 cm. between the dated levels.

The marked increases in temperate species at the base of zone 4 indicate considerably warmer water conditions, but still relatively shallow and not much deeper than a nominal 30 m. water depth. The coarse grained nature of the sediments are reflected by peaks in *Ammonia batavus* and *Cibicides lobatulus*, while intervening finer grained sediments have an associated peak in *Stainforthia fusiformis*. The changing frequencies of species and faunal parameters within this zone are therefore considered to be a response to grain size differences rather than any direct environmental condition. However, the mechanisms that have generated such sedimentological changes within the core remain unknown. The frequency peaks of the infaunal species *Stainforthia fusiformis* may suggest lowered oxygen concentrations within the pore waters of the fine grained sediments; Andrews *et al.* (1990) report similar peaks of *Fursenkoina fusiformis* corresponding to disaerobic events from the SE Baffin Shelf. Problematic though, is the mechanism to generate intervening anoxic and normal marine conditions within the same zone which is constrained by two radiocarbon dates to a period of about 600 years.

The radiocarbon dates from the top and bottom of zone 4, $11,035 \pm 130$ years BP and $11,625 \pm 130$ years BP respectively, confirm the Lateglacial interstadial age that the faunas indicate. The nominal water depth of c. 30 m. at a time of increasing eustatic sea levels, estimated to be c. -75 m. at approximately 11,000 years BP, suggests an active isostatic component at this time. Indeed, with a present water depth of -156 m. at the site,

an eustatic sea level of -70 m., and a palaeo-water depth of perhaps between 30-40 m., then an isostatic component of some 40-50 m. is implied. It would therefore appear that the isostatic rebound of this part of the shelf kept pace with the eustatic rise in sea level following deglaciation.

The continuing presence of, the boreo-arctic species *Cassidulina reniforme* and *Elphidium excavatum* within this zone may be due to the incorporation of material from zone 3, now exposed to erosion in a temperate and stormy, shallow shelf sea. The presence of the attached species *Cibicides* gr. *lobatulus* in its highest numbers throughout the core within this zone are further suggestion of increased bottom current activity.

Zone 5 marks the return of the cold water species to this area and very occasional, scattered clasts imply some ice cover. The faunas of this zone correspond to those of zone 3 but lack the shallow, arctic water species of that zone, indicating a generally deeper, less severe environment. The generally low planktonic:benthonic ratios of this zone suggest that there may be some barrier, other than reduced water depth, to limit the numbers of planktonic specimens. In view of the climatic signal and large number of glacial scours seen from side-scan sonar images of the outer shelf and upper slope, many of which may date from the Lateglacial stadial, it is proposed that grounded ice may have limited the numbers of planktonic foraminifera entering shelf waters.

Increasing frequencies of *Nonion labradoricum* towards the top of this zone may reflect similar conditions to those of zone 1 in vibrocore 57/-09/46 and appears to be a common feature of late Quaternary deglaciation stratigraphies eg. zone 4, Solberga (Knudsen, 1982). While no radiocarbon date is available from zone 5, it would appear that the climatic deterioration that characterizes it had begun by $11,035 \pm 130$ years BP at the top of zone 4. Thus, zone 5 is placed within the earliest part of the Younger Dryas Stadial.

Zone 6, as discussed in section 3.9.3, records some evidence of bioturbation and faunal contamination from zone 7 above. In fact, zone 6 is lithologically indistinguishable from zone 5 and,

as demonstrated above, taxa which characterize zone 6 do not change considerably in their concentration from zone 5. Any differences in these concentrations are probably the result of dilution by zone 7 sediments rather than a faunal response. However, frequencies of *Nonion labradoricum* continue to increase from zone 5 and are a diagnostic feature of this zone.

The radiocarbon date of $10,635 \pm 120$ years BP at the top of this zone confirms the chronology of the climatostratigraphic interpretation of zones 5 and 6 as belonging to the Younger Dryas. Water depth at the site continued to rise during the accumulation of zone 6 sediments and it therefore appears that during the Younger Dryas the rise in eustatic sea level overtook any residual isostatic component. Estimates of eustatic sea level place it at c. -60 m. at about 10,500 BP (Fairbanks, 1989) which means that, allowing for a generous +10 m. isostatic rebound during the last 10,500 years, palaeosea level may have been about 80-90 m. at the site.

Zone 7 faunas continue the signs of amelioration generated by bioturbation induced mixing within zone 6, with marked increases in temperate species and even more marked declining arctic species frequencies. Some of these taxa, particularly *Cassidulina laevigata*, increase in frequency upwards through the zone, but otherwise the assemblages dated at $5,555 \pm 90$ years BP differ little from those at the surface. However, marked differences in the ratios of planktonic:benthonic (p:b) species are noted from the base and top of this zone.

The generally increased p:b ratios of zone 7 reflect increasing sea levels and ice free conditions as compared to the preceeding zones. The lower ratios at the bottom of zone 7 are probably the result of strong early Holocene currents which act to inhibit planktonic foraminifera from settling; present day currents over the shelf are generally very weak and are not considered to inhibit settling.

It appears that the second major hiatus within this core occurs at the zone 6-7 boundary, with nearly 5,000 years represented by a difference of only 10 cm. between sampled levels. The erosive nature of the lithological contact which occurs at the

same level as the zone 6-7 faunal boundary suggests that the date of 10,635 \pm 120 years BP for the uppermost Younger Dryas sediments may be correct and that the last 600 years of accumulation associated with this event have been lost by erosion during the early Holocene transgression and re-activation of strong shelf currents.

3.10.4 Oisein sequence

The interpretation of the age of this sequence and vibrocore 56/-09/142 is aided by its recognition as a draped sequence, exposed at the sea bed to the south of the main study area. By implication, a seismically laminated and draped sequence in an area which has undergone major glaciation during the late Devensian must post-date the glacial maximum; although no independent dating evidence exists for this core. The two foraminiferal zones recognized do suggest a transition from a cold, high-boreal climate to a temperate climate, much like that of the present day in this area.

Zone 1 faunas are consistent with accumulation in an ice-distal glacial marine environment of limited water depth. They lack deeper water taxa, such as those of the genus *Islandiella*, and also exhibit low planktonic:benthonic ratios, all of which indicate limited exchange with oceanic waters probably as a combined result of low relative sea level and sea-ice cover. The transition to zone 2 faunas is gradual with 'tails' of temperate species extending into zone 1 and those of the cold water taxa, such as *Elphidium excavatum* forma *clavata*, extending into zone 2.

Subzone 2a faunas represent a mixed assemblage of boreo-arctic and temperate faunal elements. These are interpreted as due to increased bottom current activity during the Lateglacial/Postglacial sea level (eustatic) rise ; or possibly unrecognized bioturbation down to a depth of 3.0 m. Whatever the cause of this mixing, no major hiatus, either lithological or faunal, is apparent between zone 1 and subzone 2b.

Subzone 2b comprises a temperate fauna within a depositional environment where fine grained sediments have accumulated, apart from a layer of coarse sands and shelly material represented by

samples 142/18 and 142/11. The foraminifera, much like those of zone 4 in VE 57/-09/89, are interpreted as exhibiting a grain size response; *Stainforthia fusiformis*, *Bulimina* gr. *marginata*, and *Cibicides* gr. *lobatulus* illustrate this quite clearly within this subzone. The origin of this coarse layer is unknown and is difficult to explain at this level; it may, however, correspond to an interval of increased current activity and fine sediment winnowing associated with rising Holocene sea levels.

The low planktonic:benthonic ratios at the top of this core are probably due to the relatively low numbers of planktonic specimens which occur over the more extensive and broader shelf in this southern area.

3.10.5 Fionn sequence

The depositional context of the seismically laminated Fionn sequence within the St. Kilda Basin, bordered as it is by morainal banks to the west, is important in the interpretation of palaeoenvironment. Equally important in the environmental interpretation of zone 1 within vibrocore 57/-09/46 and the samples analysed from 57/-09/44 are the presence of structures identified from seismic profiles as ice-berg scour marks and estimated to have formed when sediments were accumulating at slightly less than 3 m. beneath the present sea bed.

3.10.5.a Vibrocore 57/-09/46

Radiocarbon dates and volcanic ash stratigraphy from this vibrocore provide a valuable chronological framework to what appear to be continuously accumulated sediments from the beginning of the Younger Dryas period.

Zone 1 faunas are largely constrained to the period between 11,015 \pm 130 years BP and 10,175 \pm 110 years BP, which is widely accepted to correspond to the climatic deterioration of the Younger Dryas Stadial. The foraminiferal faunas are consistent with accumulation under a seasonal sea ice cover but do not show signs of any marked reduction in salinity. The higher frequencies of *Nonion labradoricum* upwards through zone 1 possibly indicate increasing water depth, but this is not reflected in changing

planktonic:benthonic ratios. The numbers of planktonic species may remain low within the St. Kilda Basin during this period because of grounded icebergs forming a barrier on the outer shelf which inhibit ocean/shelf water mass exchanges. These faunas correlate well with those in zones 5 and 6 of vibrocore 57/-09/89; although lacking the mixed assemblages resulting from bioturbation within zone 6 of the latter.

Zone 2 of this vibrocore, while subdivided into four subzones for ease of description, largely records the rapid transition from boreo-arctic conditions during the Younger Dryas to temperate conditions during the Holocene. The radiocarbon date of $9,975 \pm 110$ years BP near the base of zone 2 within subzone 2b, illustrates the rapidity of the climatic amelioration associated with the Younger Dryas-Holocene transition, in view of the fact that a date of $10,175 \pm 110$ years BP at between 1.05 and 1.3 m. indicates full stadial conditions. There does appear to be an increase in water depth at this site through the Holocene, although it is unlikely that the upper 0.82 m. of this core represents continuous Holocene sedimentation. Water depths within subzone 2a and 2b are placed at a nominal 60 m. but may be greater than this.

The faunas bear a close resemblance to those of zone 7 in VE 57/-09/89 except that *Spiroplectammina wrightii* dominates instead of *Cassidulina laevigata*. Both cores occur at the same present day water depth and have similar $>63\mu\text{m}$. proportions in the surface samples. The near-surface samples from vibrocore VE 56/-09/142 are considerably different, with *Bulimina* gr. *marginata* and *Stainforthia fusiformis* as the dominant species. However, in this core the coarse fraction ($>63\mu\text{m}$.) is considerably reduced and the dominant species here appear to be well suited to an infaunal mode of life in fine grained sediments.

Further confirmation of the Younger Dryas age of the zone 1 sediments from this core is provided by a peak of clear acidic volcanic ash shards at 2.8 m.. Geochemical analysis indicates that these are rhyolitic in composition with a lowered FeO content, but otherwise geochemically similar to the Vedde Ash of western Norway. The latter has been dated at 10,600 years BP (Mangerud *et*

al., 1984). Stoker *et al.* (1989) propose a mechanism of aeolian differentiation (ie. density sorting during airborne transport) to explain the changing FeO/SiO₂ content of the Vedde Ash from western Norway to the North Atlantic; on the basis that the Fe content is the most significant control on shard density.

The brown basaltic shard peak at 2.15 m. is difficult to interpret and contains two geochemically distinct group; one with low FeO:MgO ratios, the other with higher ratios. These are tentatively correlated with North Atlantic Ash Zone 1 (NAAZ1) which was dated by Ruddiman and McIntyre (1973) at 9,300 years BP, although Mangerud *et al.* (1984) proposed that NAAZ1 is a composite event; this would help to explain the geochemical variability and complexity of shards assigned to this ash zone.

The minor peak of brown basaltic shards at about 1.20 m., at a core level radiocarbon dated to 10,175 \pm 110 years BP, may suggest a relationship with the Saksunarvatn Ash which Mangerud *et al.* (1986) date at between 9,000-9,100 years BP. However, Selby (1989) reports that shards from this peak are geochemically distinct from known Saksunarvatn compositions and he interprets these as reworked from lower down the core. While this minor peak may be due to reworking of shards, it is interesting to note that it corresponds to a minor peak within the >63 μ m. profile for the core (fig.3.22) and may therefore represent an horizon where fine sediments have been winnowed away, thus increasing the concentration of coarse material (including shards).

3.10.5.b Vibrocore 57/-09/44

These sediments remain undated, but their lithological and faunal characteristics correlate well with those of zone 1 in nearby VE 57/-09/46. The generally even, moderately high frequencies of *Nonion labradoricum* suggest that these sediments correspond to the upper part of zone 1 from VE 57/-09/46, as do the frequency relationships of the dominant taxa *Cassidulina reniforme* and *Elphidium excavatum* forma *clavata*.

3.11 The Lateglacial Stratigraphic framework: palaeoenvironmental implication.

The biostratigraphic zonation of vibrocore 57/-09/89 is critical to the understanding of Lateglacial environmental changes on the Hebridean Shelf, while other cores, VE 57/-09/46 in particular, provide a higher resolution picture of events during certain periods. The benthonic foraminifera analysed are interpreted to vary in response to changing environmental conditions and therefore provide a climatostratigraphic record of the Lateglacial which has been constrained by radiocarbon dates and volcanic ash stratigraphy. Similar patterns of foraminiferal change have been reported elsewhere, but the records from vibrocore 57/-09/89 and 57/-09/46 provide what are probably the best defined and most clearly resolved marine records of this period from western Britain (Peacock *et al.*, in prep.).

The evidence of deposition during the climatic events associated with the Lateglacial are now dealt with in chronological order from the end of the last glacial maximum.

3.11.1 Late Devensian: extensive ice on the middle and inner shelf.

The evidence for grounded ice on the middle shelf has been discussed by Selby (1989) and the following vibrocores contain diamicts which are thought to be associated with deposition from this ice sheet: VE 57/-09/21, VE 57/-09/89 (zones 1-3), VE 57/-10/17 (zone 1), and VE 57/-09/60 (zone 1). As demonstrated in vibrocore 57/-09/89 facies changes within a glacial marine context, associated with deglaciation and increasingly ice-distal conditions, are recognizable. Proximal glacial marine diamicts appear to be characterized by mixed foraminiferal assemblages of temperate and boreo-arctic species, although the latter dominate and are thought to be largely *in situ*. Low benthonic sums imply relatively high sedimentation rates, although insufficient dating evidence is available to allow sedimentation rates to be calculated within any single facies type from this period. However, two dates from VE 57/-09/89 suggest an accumulation rate

of 33 cm. per 10^3 years between zone 1 and zone 3. However, accumulation rates are unlikely to have remained constant as the depositional environment changed from a proximal glacial marine facies to an increasingly ice-distal and turbid facies; in fact the low benthonic sums of this zone are interpreted to result from an increase in the depositional rates. The rate compares with a mean Weichselian sedimentation rate of 3.5 cm. per 10^3 years in the north east Atlantic and central Norwegian Sea (Ruddiman and Bowles, 1976).

3.11.2 The Lateglacial Interstadial: climatic amelioration and isostatic recovery.

Zone 4 of vibrocore 57/-09/89 records the climatic amelioration of the Lateglacial Interstadial, commonly referred to as the Windermere Interstadial. Much of the early interstadial appears to be missing and a major hiatus is implied at the zone 3-4 boundary. This is interpreted as due to erosion and lack of deposition during the early interstadial and supports the shallow marine context proposed for these deposits. Peacock *et al.* (in prep.) propose a notional water depth of 40 m. or less, based upon the occurrence of the bivalve *Abra alba* within this zone; the foraminifera also confirm this relatively shallow marine interpretation.

As discussed in section 3.10.3, the water depths implied by these faunas point to a considerable isostatic rebound component which was sufficiently rapid to keep pace with the well documented eustatic sea level rise of this period (Fairbanks, 1989), thus generating little visible evidence of changing relative sea level from before c. 13,500 years BP until after 11,625 years BP.

The sediments and faunas of the interstadial point to discontinuous sedimentation and possibly episodes of current winnowing; is therefore doubtful whether any accumulation rate will be meaningful. However, two radiocarbon dates of $11,625 \pm 130$ years BP and $11,035 \pm 130$ years BP exist and are separated by 33 cm. within the core which implies a mean interstadial sedimentation rate of 56 cm. per 10^3 years. Stoker *et al.* (1989) report sedimentation rates of 20-22 cm. per 10^3 years during this

period in the deeper waters of the northern Rockall Trough and Faeroe-Shetland Channel.

3.11.3 The Younger Dryas: climatic deterioration.

Both vibrocores 57/-09/89 and 57/-09/46 contain sediments assigned to the Lateglacial Stadial, generally reported between c. 11,000 to c. 10,000 BP at other Scottish west coast marine sites (Peacock and Harkness, 1990). The faunas present indicate increasing water depths throughout this period, with the eustatic component overtaking the now reduced isostatic component to produce a rise in relative sea level. This is expected, since Fairbank's (1989) eustatic sea level curve indicates rapidly rising sea level and any glacio-isostatic component should have largely decayed by this time (cf. Boulton, 1990); the net result is therefore transgression throughout the Younger Dryas.

Sedimentation rates during this period appear to have been exceptionally high. In the case of VE 57/-09/46 and VE 57/-09/89 the following rates are estimated, although the latter includes part of the Lateglacial Interstadial, since no date exists from the base of the Younger Dryas in this vibrocore.

<u>57/-09/89</u>	<u>57/-09/46</u>
0.70 - 0.75 m. = 10,635 BP	1.05 - 1.30 m. = 10,175 BP
1.94 - 2.00 m. = 11,035 BP	4.80 - 5.00 m. = 11,015 BP
1.245 m./400 yrs = 0.31 cm.yr ⁻¹	3.725 m./840 yrs = 0.44 cm.yr ⁻¹

These rates are similar and it is likely that the rate from VE 57/-09/89 is underestimated by using the radiocarbon date from the top of the interstadial zone. Equally, one might expect sedimentation rates to be higher near the centre of the basin, at the point location of VE 57/-09/46, rather than at its margins, at the point location of VE 57/-09/89. Andrews *et al.* (1990) report sedimentation rates of between 0.5 and 4.0 m. per 10³ years during Termination 1 on the S.E. Baffin and East Greenland Shelves.

A further test of the radiocarbon chronology within VE 57/-09/46 comes from the volcanic ash stratigraphy, and in particular the peak at 2.85 m. of rhyolitic shards correlated with

the Vedde Ash which is reported to be dated from an Icelandic eruption at 10,600 BP. By employing the average sedimentation rate of 0.44 cm. yr^{-1} , then the expected Vedde Ash peak should occur at 182.6 cm. above the dated level of $11,015 \pm 130$ years BP, at a core depth of 4.9 m. (ie. at 3.07 m.). Thus, the observed depth of this peak (2.85 m.) differs from the expected depth (3.07 m.) by 22 cm. or an equivalent age difference of c. 50 years. In view of the relatively large standard deviations on these radiocarbon dates and the assumptions relating to a constant accumulation rate, the observed and expected dates are remarkably close.

Further indication of the climatic regime and transgressive nature of this period is seen in fig.3.6.b, where 'V-shaped' features downcut into sediments dated as belonging to the Younger Dryas (see section 3.6). These features, interpreted as ice-berg scours, occur as a discrete event some 3.0 m. below the sea bed; fortunately for stratigraphic reconstructions they are limited in numbers. It seems likely that the entrance of these bergs into the St. Kilsa Basin occurred once sea level rose sufficiently to allow them to pass over the morainal banks of the outer and middle shelf. In fact, there is a 15 m. vertical difference in height between the shallowest part of the basin and the deepest part of the middle shelf "barrier". It is therefore proposed that later bergs entering the basin had insufficient draft to plough the sea bed or that in combination with rising sea level that the bergs reaching N.W. Scotland were reduced in size and/or increasingly limited in numbers until they no longer reached this far south in the N.E. Atlantic.

In an attempt to possibly date this discrete Younger Dryas "ploughing event" I have assumed that it corresponds to a core depth of 3.0 m. in VE 57/-09/46. However, it should be emphasised that this vibrocore cannot be directly related to the seismic section illustrated in fig.3.6.b, upon which this depth is based. On the basis of this assumed depth and an approximate accumulation rate of 0.44 cm.yr^{-1} within zone 1, the core depth of 3.0 m. should correspond to a date of c.10,580 years BP. In view of the similarity in depth and age of the volcanic ash peak at 2.85 m., correlated with the Vedde Ash, it is interesting to note that the

large size of some of the shards, some considerable distance from their Icelandic source, has been attributed to long-distance transport by ice-rafting (cf. Mangerud *et al.*, 1984). There may therefore be an intimate relationship between the first evidence of grounded ice-bergs within the St. Kilda Basin during the Younger Dryas and the deposition of the Vedde Ash. However, the lowered FeO content of these shards has been attributed to aeolian based density sorting (Long and Morton, 1987), although this need not necessarily have occurred as far to south west as the present study area, and its effects might still be realized by long-distance ice-rafting.

3.11.4 The Holocene: climatic amelioration and continuing transgression.

Holocene sediments are present in nearly every vibrocore and generally over-lie the underlying sediments unconformably. Radiocarbon dates from VE 57/-09/89, for example, suggest a major hiatus at the base of the Holocene, with possibly some of the late Younger Dryas sediments removed by erosion during the early Holocene transgression of the shelf. In VE 57/-09/46 the Younger Dryas-Holocene transition appears to be largely undisturbed, suggesting continuous sedimentation. Unexpectedly, faunas at the base of the Holocene in this vibrocore suggest water depths that disagree not only with the Barbados sea-level curve (Fairbanks, 1989), but with that from the southern North Sea (Jelgersma, 1979). A water depth of c. 110 m. would be more likely and the implications are that our understanding of the balance between isostasy and eustasy on the Scottish continental shelf require revision (cf. Nesje and Dahl, 1990), or that during periods of rapid environmental change the depth-relationship of certain species may also change (Peacock *et al.*, in prep.), although how this affects the whole fauna is uncertain. Furthermore, the low planktonic:benthonic ratios also suggest relatively lowered sea levels during the early Holocene, followed by rapid transgression; this further supports the idea of an unknown isostatic component still active during the early Holocene.

One possible cause of this reduced rate of early Holocene sea

level rise is the eastwards migration of a collapsing forebulge associated with the grounded ice that is now known to have existed over this part of the shelf. To what extent such a feature might reduce relative sea-level is unknown.

3.12 Conclusions

The St. Kilda Basin became ice-free after 15,250 BP, following the withdrawal of the Late Devensian ice sheet from its maximum position on the outer shelf. Sedimentation in a shallow water, high-arctic, muddy environment continued until after 13,555 BP. There followed, after a major hiatus, a higher energy temperate episode during which water depths were probably no more than 40 m. over the basin; this is correlated with the latter part of the Lateglacial (Windermere) Interstadial and with the warmer interval that culminated in shallow Scottish seas a little before 11,000 BP. The Younger Dryas (Loch Lomond) Stadial is marked by the return of muddy sediments and cold water faunas from before 11,035 \pm 130 BP to after 10,175 \pm 110 BP. Rapidly increasing relative sea levels are identified during this period but are most clearly seen within the coarse Holocene sands which cover many of the shelf sequences.

4.1 Introduction

4.1.1 Location and Topography

The present investigation is based upon a cliff exposure at Aberdaron on the western extremity of the Lley Peninsula, North Wales (fig.4.1). The Lley Peninsula, which defines the northern end of Cardigan Bay and extends some 60 km. from the mountainous area of Snowdonia is a distinctive feature of the relatively shallow Irish Sea Basin.

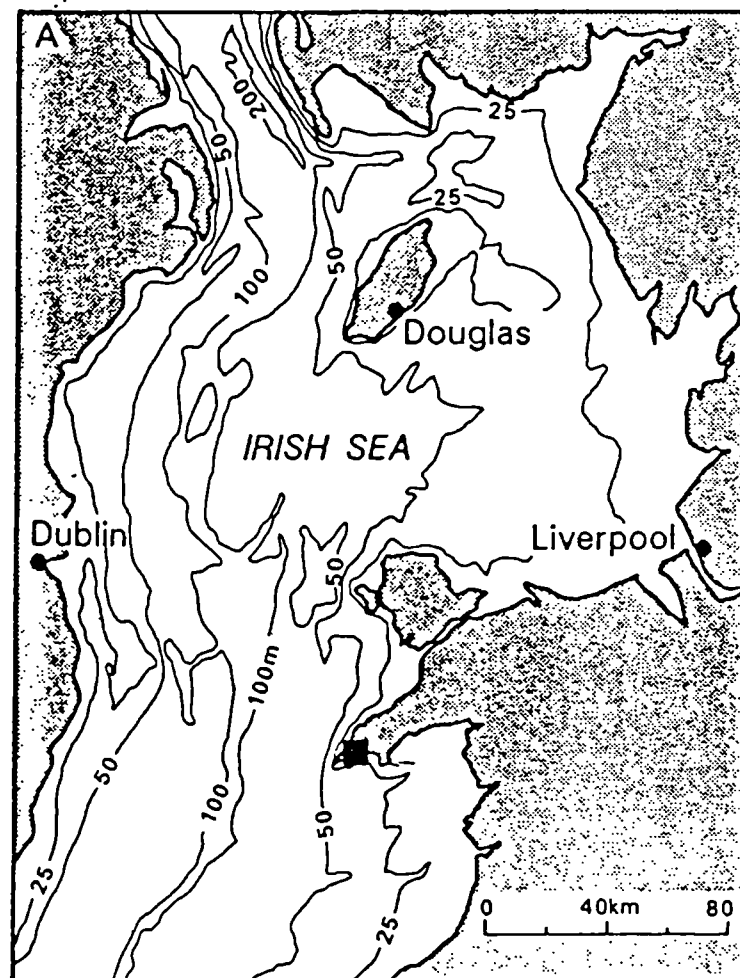


Fig.4.1 Location map, showing the Aberdaron area at the western extremity of the Lley Peninsula. Inset = fig.4.2. (from Austin and McCarroll, in prep.).

The section at Aberdaron exposes a thick sequence of drift within a fault-bounded embayment. The sediments which occupy this 1.4 Km² bedrock depression (Gibbons, 1989) extend beneath the present beach level and are clearly exposed as a result of rapid coastal erosion at this site. The surrounding hills expose bedrock and retain only a thin cover of drift, although former extensive drift sequences within Aberdaron Bay are indicated by a distinctive drift terrace on the adjacent headlands. The drift surface is gently undulating, without enclosed depressions and is incised by the Daron and Cyll-y-felin streams (fig.4.2).

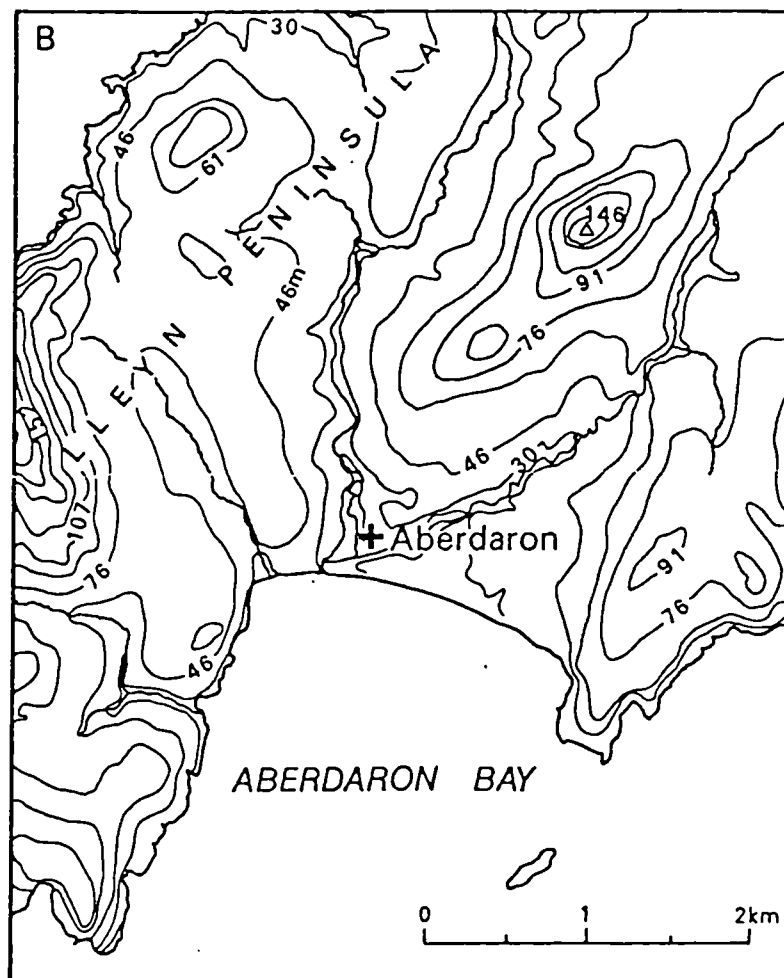


Fig.4.2 Topographic map of Aberdaron and the surrounding hills (from Austin and McCarroll, in prep.)

The Aberdaron embayment is surrounded by several steep sided, partly drift-filled valleys cut into bedrock in places and occupied by markedly underfit streams. These have been interpreted as subglacial meltwater channels and are particularly well developed over the western Lleyn. To the north and west are undulating surfaces of thinner drift which rise gently and thin-out onto the lower slopes of the surrounding hills; numerous small (<10 m.) enclosed depressions occur over these surfaces. They are interpreted as kettle holes and some of these contain organic sediments; preliminary results suggest that accumulation began early during the Lateglacial period (McCarroll & Harris, in prep.).

4.1.2 Geological Setting

The geology of the Lleyn Peninsula consists of acid intrusive rocks which form a series of hills along the north coast, of which Yr Eifl is the highest at 564 m.. The southern part consists of Ordovician sediments and volcanics, while on the western extremity Precambrian melanges are exposed (Gibbons & McCarroll, in prep.).

North of the hills, along the north coast is a band of constructional drift topography known as the Clynog Fawr Moraine (Synge, 1964); while over much of southwest Lleyn there is a subdued drift topography. Further erosional evidence of the direction of ice movement is documented by McCarroll (1991) and illustrated in fig.4.3. Detailed mapping of the erosional & depositional evidence from western Lleyn (McCarroll, 1991) suggests that the ice which crossed Anglesey from NNE to SSW (Greenly, 1919; Harris, 1989, 1991) continued in this direction in this area, veering to north/south along the flanks of the western coastal hills. The implications of these features are discussed at length below.

4.1.3 Historical Setting/Background

The thick and often complex glacial deposits surrounding the Irish Sea Basin are dominated by diamicts, sands and gravels and have been reported to contain shell fragments and microfaunas. At Aberdaron the most westerly sequence of these 'Irish Sea drift' deposits from mainland Britain are exposed and are considered by

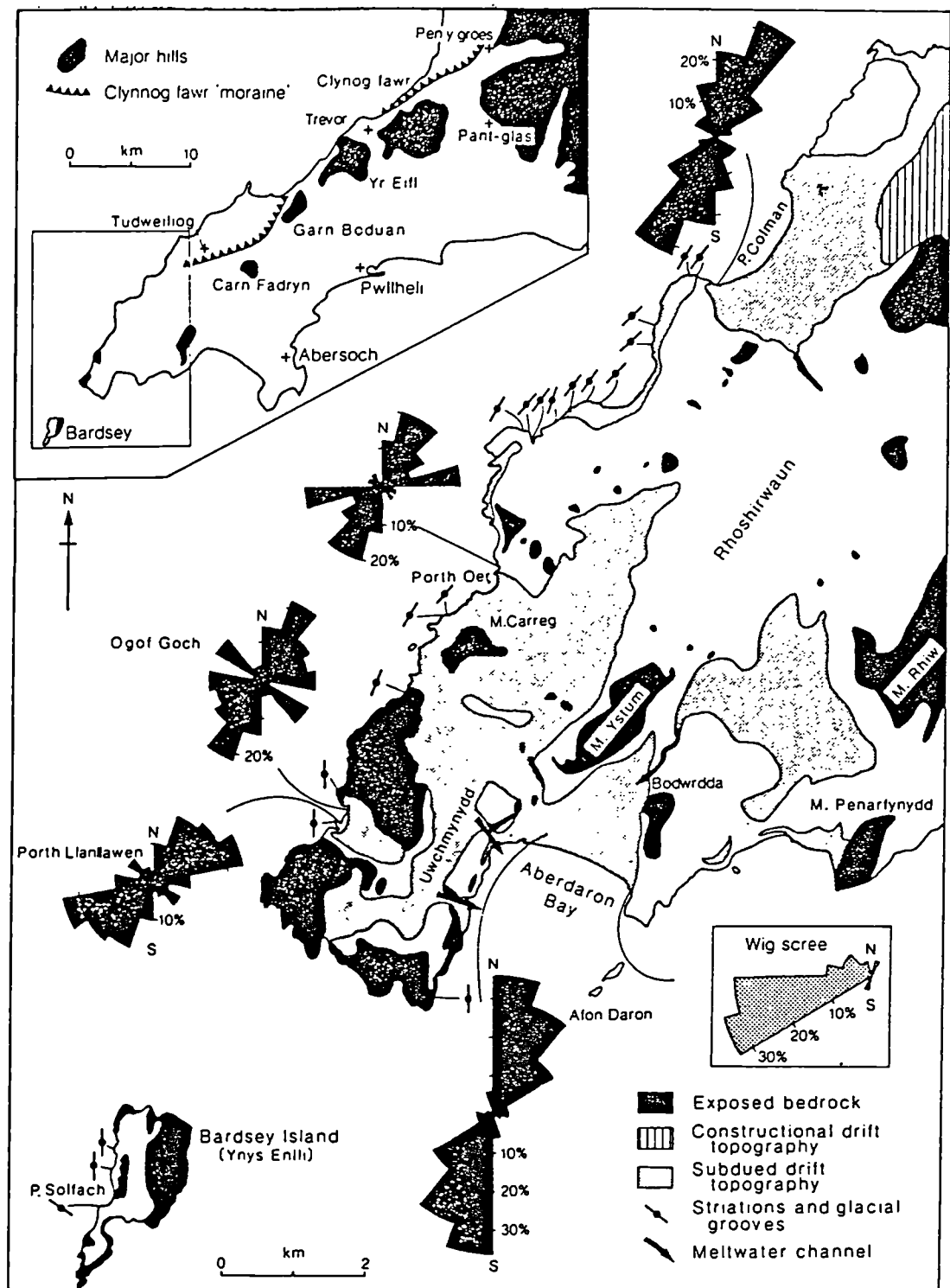


Fig.4.3 Drift topography, erosional and depositional evidence of ice flow directions in western Llyn (from McCarroll, 1991).

McCarroll & Harris (in prep.) to provide an insight into conditions during the last deglaciation of the Irish Sea Basin.

Originally, these deposits were interpreted as the product of a Biblical flood eg. Buckland (1823). With the rise of uniformitarianism and the increasing evidence of long distance and even uphill transport of erratics, there arose the concept of deposition in arctic waters with sediments transported and deposited by drifting ice (Lyell, 1833; Darwin, 1842, 1848). However, Tiddeman (1872) recognized the presence of a Pleistocene Irish Sea glacier and the deposits were re-interpreted as terrestrial tills and glacial outwash sediments. Thus, for more than a century a marine origin has been discounted, the sediments being attributed to an ice sheet that occupied the Irish Sea Basin during the late Devensian (c. 30,000–13,500 years BP).

In more recent years, this terrestrial depositional interpretation has been questioned. It is suggested that rather than representing terrestrial deposition at a time of eustatically depressed sea levels, many of the drift sequences of western Britain may reflect glacial marine conditions with isostatically raised relative sea levels up to 140 m. higher than present (eg. Eyles & McCabe, 1990). Thus, many of the deposits assigned to the 'Irish Sea Drift' have been re-interpreted as glacial marine in depositional origin, with the microfaunal assemblages cited as supporting evidence (Eyles & McCabe, 1990; McCabe *et al.*, 1990).

At the same time that these deposits have been re-interpreted as glacial marine, a growing school of thought suggests that they are terrestrial or lacustrine in depositional origin (Thomas *et al.*, 1985; Thomas & Kerr, 1987; Hart, 1990; Hart *et al.*, in prep; McCarroll, 1991; Harris, 1991; McCarroll & Harris, in prep; Austin & McCarroll, in prep.), but relatively few detailed studies which attempt to reconstruct the environments of deposition have been undertaken amongst these.

4.1.3.1 Foraminifera and Biofacies

The present study is largely concerned with aspects of microfaunal assemblages from these deposits and it is here that the present review will concentrate. As mentioned above, shelly faunas and microfaunal assemblages were originally used as

evidence of deposition under cold marine conditions, together with northern erratics which were presumed to have been supplied by drifting icebergs (Darwin, 1884). The use of boreo-arctic marine microfaunas has been cited as supporting evidence in the facies investigations which interpret the 'Irish Sea drift' deposits as glacial marine in depositional origin (eg. Eyles & McCabe, 1989, 1990; McCabe *et al.*, 1990).

The variable foraminiferal assemblages which occur within the 'Irish Sea drift' are reported to contain a mixture of shallow temperate, boreo-arctic, cosmopolitan, deep water, Pliocene and early Pleistocene foraminifera which vary in relative abundance from site to site and within a given section (eg. McCabe *et al.*, 1990). As discussed by Eyles & McCabe (1989), there are several schools of thought as to the palaeoenvironmental significance of the variable foraminiferal assemblages which occur within these deposits. The three main schools of thought are:

School 1 - assumes all the microfauna are derived from interglacial and older sediments (see Thomas and Kerr, 1987 for references and Huddart, 1981 for discussion) and that the fine grained Irish Sea diamicts are subglacial or glaciolacustrine in origin.

School 2 - assumes that the fauna is largely *in situ* and that the sediments are mostly displaced marine muds that have been dredged onshore by the ice sheet (Warren, 1985).

School 3 - maintains that the 'Irish Sea drifts' are complex marine and glacial marine deposits with both *in situ* and reworked microfaunal components (McCabe *et al.*, 1986; McCabe, 1986; Eyles and McCabe, 1989a). The present chapter is addressed to the question of the palaeoenvironmental significance of these deposits.

4.1.4 Scope of the present investigation

A major problem with interpreting microfaunal assemblages is the discrimination of reworked and *in situ* faunas. The varied assemblages of these deposits clearly do not represent coherent *in*

situ faunas, but as discussed above, opinions differ as to the extent of reworking. Thus, of fundamental importance to the use of foraminifera in such biofacies studies is the correct palaeoenvironmental interpretation of these variable foraminiferal assemblages.

In this study, careful location of samples from the cliff section at Aberdaron allows the variability in faunal characteristics to help differentiate between alternative models of sediment deposition during deglaciation of part of the Irish Sea Basin. The results have important implications for the use of similar microfaunal assemblages for the interpretation of palaeoenvironments in this region and elsewhere. The need to examine the faunas upwards of 63 μm . is discussed as are the population size-structure dynamics of the dominant species, *E. excavatum* (Terquem) forma *clavata*, Cushman.

4.2 Results

The section under discussion here comprises the modern cliff at Aberdaron (fig.4.4), part of which is obscured by coastal protection works, and is split into two parts by a large rotational landslip. The section is located in terms of distances from the eastern end of the coastal protection walls (revetments) which acts as a zero datum. The exposed sediments have been divided into two distinct facies associations (McCarroll & Harris, in prep.); the eastern most of these, a stratified angular local scree, banked up against a buried cliff (Saunders, 1973) is termed the Wig breccia. Conformably overlying this deposit are a diamict unit and a sequence of poorly sorted muddy gravels termed the Wig basal diamict and gravel. These deposits are not considered further in the present study.

The facies associations which are most extensive at Aberdaron are the Lower stratified diamict association (LDA) and the Upper diamict association (UDA).

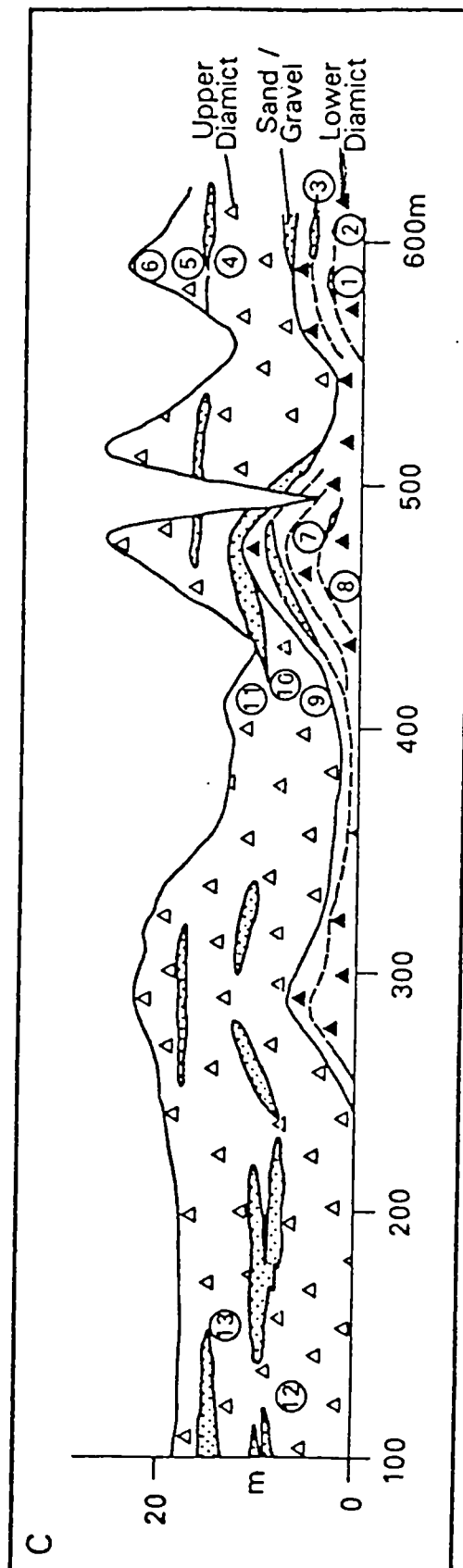


Fig.4.4 Sketch section diagram of the cliff at Aberdaron, showing sample locations in metres east of a datum (from Austin and McCarroll, in prep.)

4.2.1 Lower stratified diamict association (LDA)

This facies crops out above modern beach level east of 245 m. and is dominated by dense mud with scattered, sometimes striated clasts and occasional shell fragments. The proportion of fines (<63 μm .) ranges from 78% to 99%, including 52% to 61% clay (fig.4.5). Elongate clasts (2-10 cm.) reveal a strong fabric aligned NNE/SSW. Clast dips are generally shallow and less than 12% of the total dip at more than 40° from the horizontal (fig.4.6).

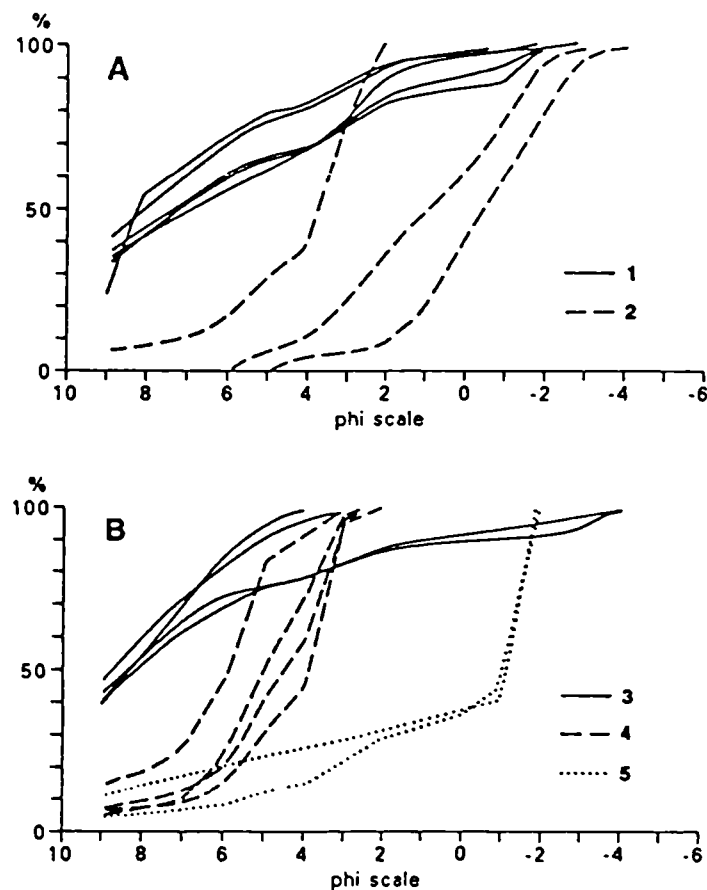


Fig.4.5 Grainsize data from (A) the upper and (B) the lower diamicts at Aberdaron . 1 = Upper diamict, 2 = Sorted layers, 3 = Lower diamict, 4 = Deformed silt/fine sand, 5 = Wig scree (from McCarroll and Harris, in prep.).

There are deformed layers and lenses of fine sand/silt (fines 45% to 91%, including 7% to 18% clay), ranging from a few centimetres to a metre thick and up to 4 m. long, throughout the LDA. Some of these larger fine sand/silt bodies retain evidence of ripples but these are generally highly deformed. Most of the units within the LDA show signs of having undergone compressional deformation, including folds and low angle reverse faults which dip mainly towards the NNE.

There is also evidence of channels, both within and near the top of the LDA, clearly visible between 290 m. to 330 m. and relatively undeformed, imbricate gravels fill small channels up to 4 m. wide and 2 m. deep.

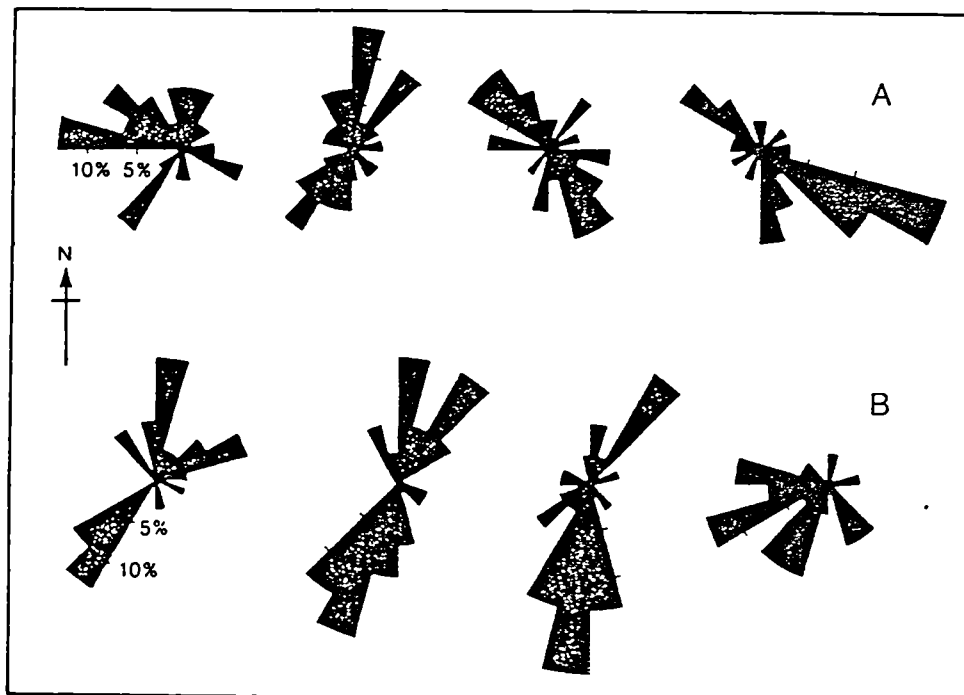


Fig.4.6 Long axis orientation of clasts from the (A) upper and (B) lower diamicts at Aberdaron (from McCarroll, 1991).

4.2.2 Upper diamict association (UDA)

The UDA is distinguished by the absence of contorted fine sand/silt bodies and by the presence of sands and gravels, sorted to varying degrees, and forming extensive layers and channel fills. The dominant facies is a dense mud with scattered stones and occasional boulders (fines 67% to 80%, including 41% to 52% clay), which in places grades into clast-dominated diamict and poorly sorted gravel (fig.4.5). Clast fabrics, though not random are weaker than those obtained from the underlying LDA. Clast dips are generally shallow, with 61% of the total dipping at less than 20° from the horizontal and less than 10% dipping at more than 40° (see fig.4.6).

4.2.3 Geometry of Sedimentary units

Where the UDA rests directly upon the LDA, the precise boundary can be difficult to define, although over much of its length this boundary is defined by sorted sediments. The boundary appears to be unconformable in many places and even erosive. For example, between 390 m. and 430 m. contorted fine sand/silt bands within the LDA are abruptly truncated and overlain by flat-based, convex lenses of gravel. Elsewhere, at 406 m. for example, the flat base of a gravel unit cuts across a silt body and the enclosing diamict without regard to the marked difference in lithology.

Sands and gravels exhibiting varying degrees of sorting are common within the UDA; towards the west of the section they form clear channel fills (eg. West of 230 m.). Elsewhere, well sorted sands and clast supported gravels grade laterally into clast-dominated diamicts which in turn grade into muddy diamicts. In places these are clearly imbricate and display strong fabrics.

The junction between the UDA and LDA is irregular, reaching 8 m. above beach level at between 280 m. and 300 m. and again between 460 m. and 490 m.; but is only 2 m. above beach level at 330 m. and at 550 m.. Stratification within the LDA approximately parallels this junction while within the UDA a crude, generally horizontal stratification exists which is defined by sorted layers and stony bands. Thus, the UDA appears to infill the topography as defined by the upper surface of the LDA. Where sands and gravels

mark this junction they appear to dip conformably with it, rather than filling the hollows.

4.2.4 Foraminiferal faunas

The results of the foraminiferal analyses of the 13 samples examined from the Aberdaron section are summarized in fig. 4.7, which includes the 34 most common taxa present and the various faunal parameters. The faunas are dominated throughout by *Elphidium excavatum* (Terquem) forma *clavata* Cushman, which varies in frequency from 78.2% in sample A8 to 50.7% in sample A11 and has a mean value throughout of 67.2%. The common accessory species in all the samples are *Cassidulina reniforme*, *Cibicides* gr. *lobatulus* and *Bolivina pseudoplicata*. Other commonly occurring taxa include *Trifarina angulosa*, *Nonion orbiculare*, *Globocassidulina subglobosa*, *Stainforthia loeblichii*, *Quinqueloculina stalkerii* and *Buccella frigida*.

The numbers of benthonic specimens per 100 g. of initial sediment dry weight (SDW) of the mud samples vary from 753 (sample A3) to 2,886 (sample A10), with higher values in the fine sand/silt bodies (6,021 in A1 and 3,477 in A7). The numbers of planktonic specimens per 100 g. SDW are strongly correlated with the numbers of benthonic specimens, maintaining a planktonic:benthonic ratio close to the mean value of 0.11 throughout (fig.4.7). Pre-Quaternary foraminifera, including Cretaceous planktonic species, are common in all samples (up to 402/100 g. SDW in sample A10) and are also correlated with the number of benthonic specimens per 100 g. SDW.

Measurements of faunal diversity (Walton, 1964) and the number of benthonic specimens per sample are also included in fig.4.7. Faunal diversity values range from 7 to 14, while the number of benthonic specimens per sample varies from 300 to 431; the latter are generally negatively correlated to the faunal diversity.

The benthonic fauna were further sub-divided and the percentages of selected boreal species were summed (fig.4.7). The ten boreal species selected (based upon a modified list from Knudsen, 1982 and Feyling-Hanssen, 1983) are : *Ammonia batavus* (Hofker), *Bulimina* gr. *marginata* d'Orbigny, *Buliminella*

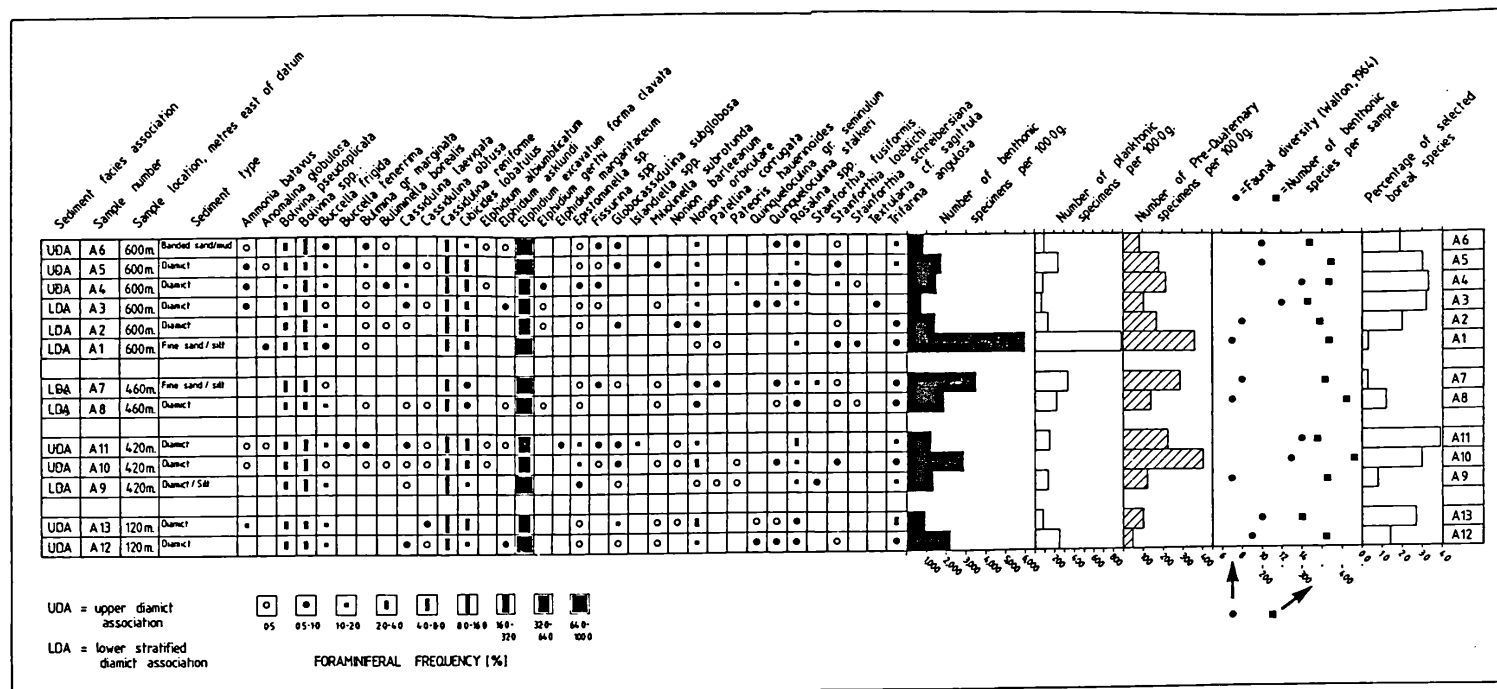


Fig.4.7 Summary foraminiferal diagram from Aberdaron, samples located in fig.4.4. [Correction: *Buliminella borealis* should read *Buliminella elegantissima*; *Textularia sagittula* should read *Spiroplectammina wrightii*; and '■ = Number of benthonic species per sample' should read '■ = Number of benthonic specimens per sample'].

elegantissima (d'Orbigny), *Cassidulina obtusa* Williamson, *Elphidium crispum* (Linné), *E. gerthi* van Voorthuysen, *E. macellum* (Fichtel & Moll), *E. margaritaceum* Cushman and *Globocassidulina subglobosa* Brady. The sums obtained are low, ranging from 0.3% in sample A1 to 3.0% in sample A11. Apart from samples A1 and A7, which correspond to maxima in the number of benthonic specimens per 100 g. SDW; the sums obtained do not appear to be correlated with the other faunal parameters.

The results of the biometric analysis on the test diameters of *E. excavatum* forma *clavata* are presented in figs.4.8a,b,c and range from 72 μ m. to 336 μ m.. Measurements on specimens from the LDA and UDA reveal similar, positively skewed distributions with a mean test diameter of c.143 μ m. (based on all 11 samples). Test diameters from the fine sand/silt bodies approximate a normal distribution, with low standard deviation values and a notable shift in the mean test diameter to c.120 μ m..

(a) fine sand/silt bodies

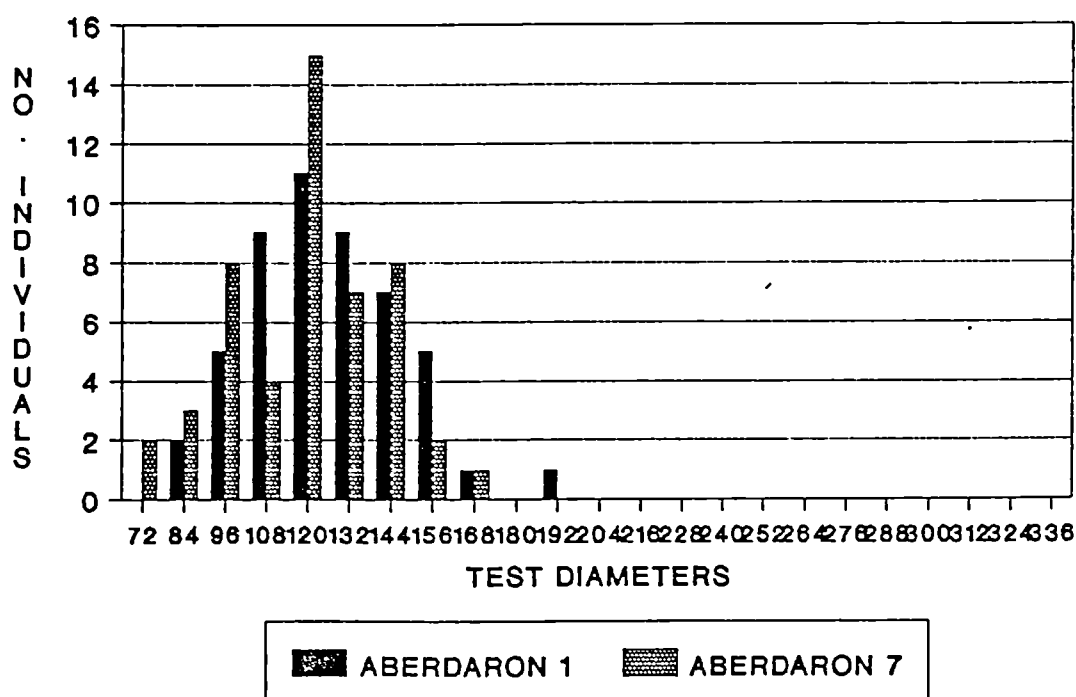
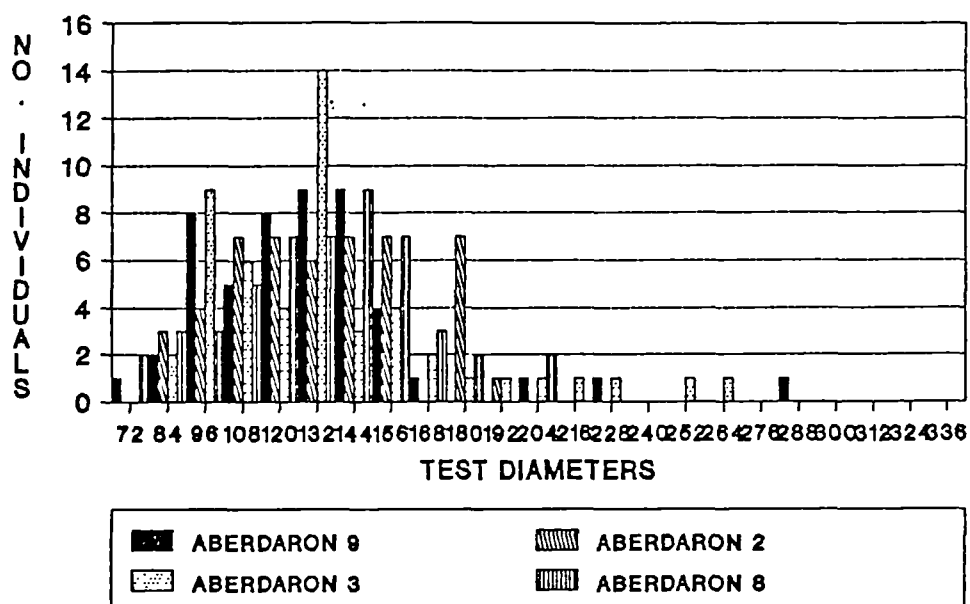


Fig.4.8.a Test diameter measurements on *Elphidium excavatum* forma *clavata* from the fine sand/silt bodies at Aberdaron (test diameters in microns).

(b) lower diamict



(c) upper diamict

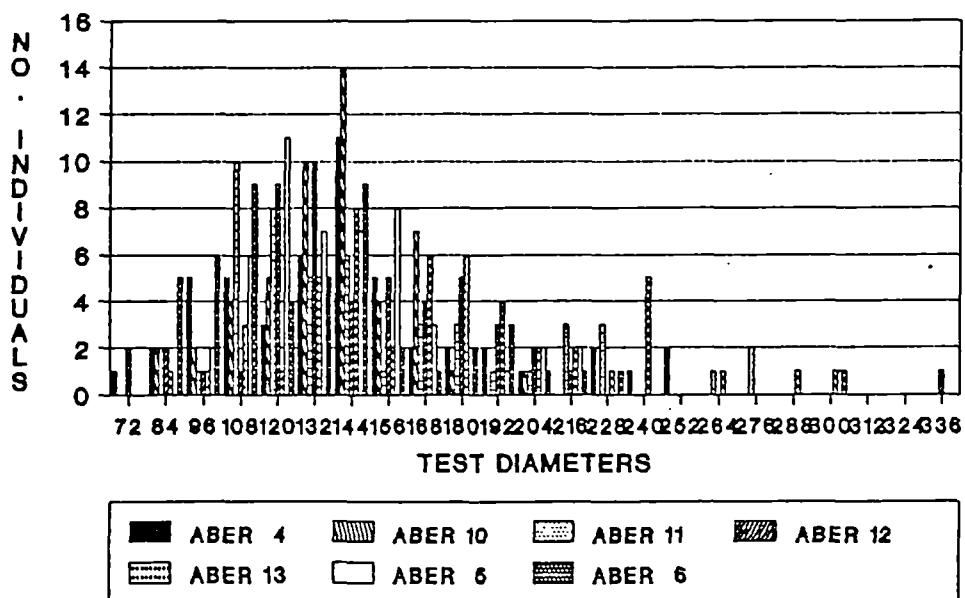


Fig.4.8 Test diameter measurements on *Elphidium excavatum* forma *clavata* from (b) the lower diamict and (c) the upper diamict. Test diameters in microns.

4.3 Discussion: sediment / lithofacies context

Two distinct depositional models have been proposed for this site (McCarroll & Harris, in prep.) and these are now discussed in their lithofacies context.

4.3.a Glacial Marine

On the basis of work by Eyles and McCabe (1990), the sediments described above may have been deposited in a glacial marine environment. Indeed, it might be argued that the presence of broken marine shells and of marine microfauna, "together with the dominance of muddy facies" in both the LDA and UDA is indicative of deposition within a glacial marine context. However, fabric data from the LDA precludes a simple rain-out/ice-rafting model since pebble fabrics from ice-rafted derived diamictos are known to be nearly random with little consistency of vector orientations between sites and without any relationship to the probable direction of glacier flow (Domack & Lawson, 1985; Visser, 1989). The pebble fabrics from the LDA are strong & consistent, with very few steeply dipping clasts; again Domack & Lawson (1985) consider steeply dipping clasts ($>45^{\circ}$) to be a characteristic feature of ice-rafted diamictos.

The sequence at Aberdaron is most easily placed within a glacial marine context if the deposits are interpreted as a valley infill complex (Eyles & McCabe, 1990). Thus, the LDA might be interpreted as the product of downslope resedimentation of unstable sediments in a proximal retreating tidewater ice-margin setting. In this type of depositional setting, the muddy facies might result from plumes and efflux jets releasing sediments, while the strong, consistent fabric might develop during downslope resedimentation.

This may account for the granulometry, fabric and deformation characteristics of the LDA facies but the presence of the deformed fine sand/silt lenses remains problematic in such a depositional context. One possibility is that they represent lag deposits produced by bottom (traction) currents, but this is unlikely in view of the fact that they lack the coarse as well as the fine fractions which are characteristic of the Lower diamict. Another possibility is that these bodies represent subaqueous outwash in

an ice proximal setting. However, in view of their widespread distribution within the LDA it remains difficult to explain why lenses of this material should retain their coherence and internal structure during the downslope resedimentation process which is required to explain the formation of such a strong fabric in the associated diamicts of the LDA.

If the LDA represents slumping of ice proximal glacial marine sediments then the UDA might be interpreted as representing increasingly ice-distal sedimentation. This is again unlikely in terms of a simple rain-out/ice-rafting model, because even though the upper diamicts display weaker, less consistent fabrics than the LDA, they are far from random and they exhibit even fewer steeply dipping clasts than the LDA.

The interpretation of the UDA as the result of subaqueous gravity flows descending from higher ground in an arc from NW to NE seems more likely and would explain the range of fabric orientations within the diamict units. Many of the finer grained deposits of the UDA might represent deposition of suspended sediment derived from meltwater plumes, while the clear channel fills might represent routeways for subaqueous outwash.

However, the geometry of these sedimentary units is considered to be problematic in the construction of a feasible glacial marine model (McCarroll & Harris, in prep.). While there is some evidence of erosion between the UDA and LDA, the overall nature of the junction is not erosional eg. the sands and gravel, which in places define the junction, dip in conformity with it rather than filling the hollows, suggesting that the sands and gravels were deposited on the surface of the LDA prior to the deformation which produced the present topography. There is also an indication from the declining dip of stratification within the UDA that the topography of the junction developed and was progressively filled-in as the UDA was being deposited. While de-watering and compression of the LDA might be expected to produce some deformation, it is difficult to visualize such a marked topography developing in a submarine environment, particularly in view of the absence of marked lateral variations in grain size and therefore potential response to loading.

Further difficulties arise in reconciling the glacial marine

model to the observed lithological features. For example the absence of a distal mud drape overlying the sequence which might reflect rising eustatic sea levels or alternatively of an overlying emergent beach facies which might represent an isostatic response. The absence of these expected lithofacies and the presence of small enclosed hollows on the surface of the thinner drift which surrounds the Aberdaron embayment and which have been interpreted as 'Kettle holes' (D. McCarroll, pers. comm.) suggests that buried ice stagnated locally during the accumulation of these sediments; this again is unlikely within a submarine context.

4.3.b Terrestrial

The terrestrial model proposed for this site (McCarroll & Harris, in prep; Austin & McCarroll, in prep.) explains the marine characteristics of the sediments as due to glacial reworking of Irish Sea marine deposits.

The deposits are not considered to have originated as lodgement tills, as evoked elsewhere, and indeed the presence of highly deformed fine sand/silt bodies within and of less deformed channel gravels towards the top of the LDA argue against such an origin. It appears that the Irish Sea glacier was actively eroding as it passed over Anglesey (eg. Harris, 1991) and the Llyn (McCarroll, 1991) and it is proposed that the most likely origin of the LDA is by a mechanism of basal melt-out from stagnant glacier ice (Shaw, 1979, 1982; Haldorsen & Shaw, 1982). By this mechanism, the strong fabric, stratification and evidence of compressional deformation are all derived from the characteristics of the debris-rich basal ice. The fine sand/silt lenses can be explained as formed by englacial streams, possibly deforming during the transport and melting of the ice (eg. Lawson, 1979).

Gravel lenses at the top of the LDA, which tend to be less deformed than the silt lenses, may represent deposition from streams flowing within the ice following stagnation and this would explain the erosional contacts of these bodies often without regard to grain size ie. the sediments were frozen. Equally, the sands and gravels which in many places define the upper surface of the LDA may represent deposition from supraglacial braided streams. Melting of the stagnant ice, as it was being buried by

the UDA would account for the gradual development and infilling of the topography which now serves to define the junction between these two sedimentary associations.

The UDA can be interpreted as having been deposited from a series of subaerial sediment flows (flow tills) and associated fluvial deposits; these in turn are most probably derived from unstable supraglacial melt-out tills deposited on the slopes surrounding the Aberdaron embayment. Much of the geometry of the UDA can be explained in terms of meltwater flowing across the developing surface to form laterally impersistent, poorly sorted gravel and sand bodies and channel fills. Thus, a natural basin exists in the structural depression of the Aberdaron embayment into which sediments from the surrounding hills derived, burying the stagnant ice which melted slowly. Paul and Eyles (1990) discuss the consequences of basal melting generated by geothermal heat alone, as McCarroll and Harris propose for the site, and suggest that such sediments generally remain undisturbed during deposition ie. high preservation potential of subglacial melt-out till fabrics.

Many of the sedimentological and geometrical features of the deposits at Aberdaron can be explained by this terrestrial model. Equally, the fabric data (fig.4.6) accord well with this model, unlike the glacial marine model, with consistent mean vector attitudes aligned roughly parallel to the direction of ice flow ; all are typical attributes of basal melt-out tills (eg. Dowdeswell and Sharp, 1986). However, clast dips in the lower diamict are steeper than expected (Lawson, 1979) and this feature, together with the strong fabric and clear evidence of deformation suggests that the LDA derives from debris-rich ice, thus permitting only minimal movement and reorientation of clasts during deposition. The fabric data of the UDA also agrees well with the reported properties of glacial flow deposits; McCarroll and Harris envisage relatively high water contents within these sediment flows to explain the strong fabrics and general lack of steeply dipping clasts (eg. Lawson, 1979).

Furthermore, enclosed hollows which have been interpreted as Kettle holes are common and are thought to occur in areas where the depth beneath the surface to the stagnant ice was limited.

This explains why none of these features are visible along the top of the section, since the UDA is several meters thick here.

4.3.c Biofacies context

The most striking feature of the foraminiferal assemblages obtained from the diamict samples is their uniformity. The number of benthonic specimens per 100 g. or per sample are similar for both the lower and upper diamicts and there are no apparent trends within the UDA which might reflect increasingly distal glacial marine sediments. Similarly, the planktonic/benthonic ratio and the ratio of benthonic to pre-Quaternary foraminifera show no such differences or trends. Hald and Vorren (1987), in assessing the possibility of contamination of samples by allochthonous (reworked) faunas, examine the correlation of *Elphidium excavatum* with faunal elements which are clearly allochthonous. Poor correlation is taken as evidence that the majority of the *E. excavatum* are autochthonous. At Aberdaron, the allochthonous faunal elements (i.e. the planktonic species, the pre-Quaternary species and to a lesser extent the selected boreal species) maintain relatively constant ratios with the well-preserved, boreo-arctic faunal element throughout the sequence (Fig.4.7), suggesting that the latter are unlikely to be *in situ*.

If they are not *in situ*, the foraminiferal assemblages cannot be used to interpret changes in the environment of deposition during accumulation of the sediments at Aberdaron. They lend no support to the glacial marine model and the remarkable uniformity of the diamict samples is precisely what was predicted on the basis of the terrestrial model of sedimentation favoured by McCarroll and Harris (in prep.). The characteristics of the assemblages obtained from the deformed fine sand/silt bodies are also consistent with the terrestrial model, wherein they are interpreted as englacial stream deposits. The size distribution of *Elphidium excavatum* forma *clavata* tests is clearly indicative of current sorting, with the mean diameter of c. 120 μm . representing the hydrodynamic equivalent of the fine sand/silt fraction.

Although the foraminiferal assemblages from Aberdaron cannot be used directly to determine the process or environment of deposition of the glacial deposits at Aberdaron, it is

important to establish the origin of this and similar faunas from around the Irish Sea Basin.

Since the Aberdaron assemblages are dominated by faunas of a predominantly cold, shallow water affinity, an original ice proximal palaeoenvironmental setting is certainly possible. High frequencies of *Elphidium excavatum* forma *clavata* and lower numbers of *Cassidulina reniforme* are very common in Pleistocene glacial marine deposits (Feyling-Hanssen, 1964; Feyling-Hanssen *et al.*, 1971; Knudsen, 1978). On the continental shelf off Troms, northern Norway, a fauna dominated by *Elphidium excavatum* has been interpreted as indicative of an ice-proximal environment characterized by unstable bottom water conditions (Hald and Vorren, 1987). The presence of microfaunas which include temperate planktonic species, pre-Quaternary species and boreal benthonic species, none of which are an autochthonous component of a glacial marine fauna, indicate reworking of these faunal components. However, sediments interpreted as glacial marine are commonly noted to contain some reworked faunal elements (Scourse *et al.*, 1990; Scott and Medioli, 1988; Spjeldnæs, 1978).

Given this interpretation of the faunal assemblages, it is possible to assume that the cold-water foraminifera are *in situ* and to infer a glacial-marine origin for the Aberdaron drift. However, such a model would have to accommodate the remarkable uniformity of the samples, with no evidence of changes in rate of sedimentation or in environmental conditions with increasing distance from the glacier. The LDA could be interpreted as *in situ* proximal facies and the UDA as the result of redeposition of LDA sediments from the hills surrounding the Aberdaron embayment. However, it seems highly unlikely that the steep slopes of the surrounding hills would have supported sediments deposited in a marine environment, even temporarily, and it is also difficult to argue that the changes in mode and rate of deposition associated with increasing distality would register no response in an *in situ* glacial marine fauna, even where sediments were being remobilized.

It is much less problematic to interpret the foraminifera at Aberdaron as a predominantly cold water assemblage with reworked, clearly allochthonous elements, all of which have been incorporated, transported and redeposited by a glacier. In this

case, the assemblage relates to the palaeoenvironment in the Irish Sea Basin prior to the last glacial advance and has no direct bearing on conditions during deposition of the Irish Sea glacigenic deposits at Aberdaron.

Although in Quaternary sediments foraminiferal assemblages dominated by *Elphidium excavatum* forma *clavata* have often been interpreted as indicative of near glacial conditions, the modern distribution of such assemblages clearly demonstrates that this need not be the case. Murray (1991), for example, summarizes the ecological requirements of the *Elphidium clavatum* (= *E. excavatum* forma *clavata*) assemblage from the Atlantic seaboard of Europe as follows: salinity, 10-35‰, temperature 0°C to 7°C, substrate muddy gravel or sand, depth 0-285 m. These data are based upon a limited number of widely spaced investigated sites and may not represent the entire 'ecological range' of such assemblages. They do, however, demonstrate that *E. excavatum* forma *clavata* will tolerate a broad range of environmental conditions. At present, the Irish Sea mean bottom temperature in summer is 12° to 13°C. The modern distribution of *E. excavatum* forma *clavata* indicates that a drop in summer mean bottom water temperatures of only c.6°C might be enough to allow this form to reproduce in the Irish Sea Basin.

The low faunal diversity of the Aberdaron assemblages are also interesting, particularly in view of the high faunal dominance exhibited by *E. excavatum* forma *clavata*, and together suggest extreme environmental conditions. A common cause of such low faunal diversities is depressed salinity and *E. excavatum* forma *clavata* is known to be tolerant of low salinities, to the extent that Hald and Vorren (1987) consider it as much a salinity indicator as a temperature indicator. The Irish Sea presently has mean summer bottom salinities of 34‰ to 34.5‰, which is lower than normal sea water. Therefore, during much of the Devensian temperatures were lower than today and the Irish Sea would have been shallower and salinities lower, providing suitable conditions for a low diversity fauna dominated by *E. excavatum* forma *clavata*.

4.3.d Comparison with previous work.

Comparisons with previous work on faunal assemblages from Irish Sea glacial deposits are hampered by differences in sampling procedure. The data reported here are based upon samples picked from the $>63\ \mu\text{m}$. fraction. However, when dealing with glacial deposits larger aperture sieves, including $100\ \mu\text{m}$. (Hald & Vorren, 1987; Feyling-Hanssen *et al.*, 1971), $125\ \mu\text{m}$. (Mangerud *et al.*, 1981), and even $150\ \mu\text{m}$. (McCabe *et al.*, 1981) are commonly used. In this study, most of the specimens of *Elphidium excavatum* forma *clavata* (fig.4.8), and many other specimens in these samples, would not have been identified had counts been based upon the $>150\ \mu\text{m}$. fraction. Where sieve sizes larger than $63\ \mu\text{m}$. are used, picked samples may be unrepresentative of the true assemblage, thus confusing the interpretation of palaeoenvironment (cf. Schröder *et al.*, 1987).

Despite the differences in sampling procedure, comparison of the Aberdaron assemblages with those described from Skerries, north County Dublin (McCabe *et al.*, 1990), suggests similarities in both their boreo-arctic and temperate/cosmopolitan species composition, though planktonic and pre-Quaternary species are not reported. Faunal diversities appear to vary considerably at Skerries, with only two species present in the sand facies of the diamict/gravel association to 23 species present in the mud association, though this could be due to hydrodynamic sorting of the foraminifera in the sand facies? The dominant taxa of the mud appear to be *Ammonia* sp., *Elphidium crispum*, *Quinqueloculina seminulum*, *Quinqueloculina* spp., *Elphidium clavatum*, and *Haynesina orbiculare*. The lower faunal counts per unit weight (sand facies = <6 specimens $100\ \text{g}^{-1}$, pebbly sand = $95\ 100\ \text{g}^{-1}$, mud = $212\ 100\ \text{g}^{-1}$; estimated from Table 2, McCabe *et al.*, 1990) probably partly reflect the loss of smaller specimens ($63\ \mu\text{m}$. to $150\ \mu\text{m}$.) from the samples.

Comparison with the faunas described from Clogga and Knocknasilloge, south east Ireland (Huddart, 1981a) reveal little faunal affinity with those of Aberdaron and are problematic due to difficulties in interpretation (cf. Huddart, 1981a, 1981b; Thomas & Summers, 1981). However, at Clogga the dominant species are reported to be *Elphidium excavatum*, *E. clavatum*, *Protelphidium*

anglicum, *Cibicides lobatulus*, and *Cassidulina obtusa*. The dominant species at Knocknasiloge are *Elphidium macellum*, *E.articulatum*, *Elphidiella hannai*, *Protelphidium anglicum*, *Cibicides lobatulus*, and *Miliolinella subrotunda*. These assemblages, like those at Aberdaron, are clearly mixed.

The closest faunal similarities are with the sequence described not from the Irish Sea Basin, but from Donegal Bay on the west coast of Ireland (McCabe *et al.*, 1986). *Elphidium excavatum* (Cushman) (synonymous with *Elphidium excavatum* forma *clavata* of this study) is the dominant boreo-arctic species and other common accessory species include *Quinqueloculina arctica* (Loeblich & Tappan), *Haynesina orbiculare* (Brady) (synonymous with *Nonion orbiculare*), and *Cassidulina obtusa* (Williamson). In addition, McCabe *et al.*(1986) describe a macrofauna which in places is rich in paired valves of *Macoma calcaria*, a cold water species living in shallow, fully saline water in a bottom of sandy mud, two specimens of which yielded radiocarbon dates of around 17,000 years BP. The deposits are interpreted as shallow, ice-proximal glacial marine, with an *in situ*, cold water fauna. These sediments have previously been interpreted as basal till associated with onshore ice movements, inferring that the faunas are reworked (Synge, 1968; Colhoun & McCabe, 1973; Davies & Stephens, 1978; Warren, 1985).

4.4 Conclusions

The interpretation of microfaunal assemblages obtained from glacial sediments is hampered by the problems of recognising reworked elements. Subjective indicators of reworking, particularly the degree of abrasion, are unreliable since glacial entrainment, transport and redeposition may register little or no visible effect on foraminifera. Measures relating to the degree of sorting, such as biometric analysis of test diameters or of ostracod growth series are also invalid, since glacial transport and deposition need not involve size sorting.

At Aberdaron, sample sites were located carefully in order that between-sample variations in faunal characteristics would differentiate between two opposing models of depositional environment. All of the diamict samples were remarkably uniform,

with similar numbers of benthonic specimens and benthonic specimens per unit weight of sediment, similar planktonic:benthonic ratios, and similar ratios of clearly allochthonous to possibly autochthonous elements. These results are precisely as predicted by the terrestrial model of sediment deposition favoured by McCarroll and Harris (in prep.), where all of the sediments are interpreted as derived from the melting of glacier ice rich in marine debris entrained during passage down the Irish Sea Basin. The results lend no support to the alternative model of glacial marine deposition, since there is no evidence of any faunal response to the change from ice-proximal to increasingly distal sedimentation.

The mixed fauna in the glacial deposits at Aberdaron are therefore interpreted as a predominantly cold, shallow water assemblage dominated by *Elphidium excavatum* forma *clavata* but contaminated by clearly allochthonous elements, including planktonic and pre-Quaternary species. Although this assemblage could indicate ice-proximal glacial marine sedimentation, the present day distribution of similar assemblages suggests that it is just as likely to represent the conditions which prevailed in the Irish Sea Basin prior to the last (Late Devensian) ice advance. For much of the Quaternary the climate was colder than at present, and the shallow water and reduced salinities of the Irish Sea would have provided suitable conditions for this foraminiferal assemblage.

In conclusion, the foraminifera of the Irish Sea Drift deposits at Aberdaron are all derived to varying degrees. They cannot, therefore, be used directly in interpreting the environment of deposition during deglaciation of the Irish Sea Basin. Where faunal assemblages from glacial deposits elsewhere are used as evidence of glacial marine sedimentation, they should demonstrate a faunal response to the changing depositional styles which characterize such environments.

Chapter 5: Southern Celtic Sea

5.1 Location

The Celtic Sea area (fig.5.1) is a term applied to the north west European continental shelf, south west of Ireland and Britain as defined by Cooper and Vaux (1949), expanded upon by Day (1959) and discussed by Hamilton *et al.* (1980).

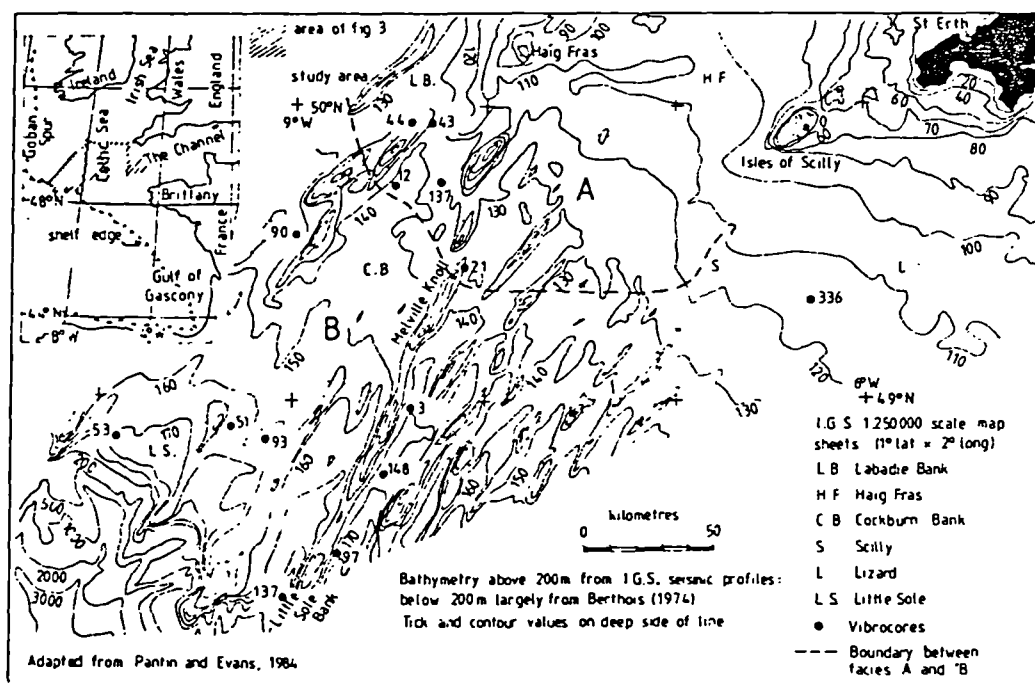


Fig.5.1 The central and southwestern Celtic Sea: bathymetry, vibrocore locations and facies A-B transition (Scourse *et al.* 1990)

The fourteen coring sites which have yielded glacial material from this study area are situated between the submarine Haig Fras granite outcrop and the shelf-edge break to the south west; in this region lying at -185 m. to -205 m. OD (Pantin & Evans, 1984). The continental shelf dips gently to the south west and the main bathymetric features are large linear tidal sand ridges which trend SW/NE (Stride, 1963; Bouysse *et al.*, 1976; Pantin & Evans, 1984; Belderson *et al.*, 1986; fig.5.1) which are up to 60 m. high, 200 Km. long, and spaced at about 10-15 km.. Small mounds of possible glacial material and scattered boulders on the sea bed are revealed by side-scan sonar and are probably the product of ice-rafting (Pantin & Evans, 1984). The cores are also located in fig.5.1 and were recovered from depths ranging between -125 m. OD (49/-09/44) and -211 m. OD (48/-09/137).

5.2 Stratigraphy

Two main Quaternary formations occur in the central and south western Celtic Sea: the late Pliocene/ early Pleistocene Little Sole Formation and the Late Devensian/ early Holocene Melville Formation. The only formations penetrated by the vibrocores are the sediments of the linear tidal sand ridges, intervening sand sheets and glacial material. Of the glacial samples recovered, most were from bathymetric 'lows' between the linear tidal sand ridges, with one core, 49/-09/44, from the flank of a sand ridge. Pantin and Evans (1984) have classified the Recent sediments overlying the glacial material into two layers. Layer 'A' consists of superficial mobile sediments, while layer 'B' is a relatively coarse gravel pavement beneath it. Boulders over 1 m. in diameter are widely scattered across the area and have been retrieved on the anchors of drilling ships (cf. figs. 223, 224 of Scourse *et al.*, 1991) are identifiable on side-scan sonar and have been observed in submarine photographs (Hamilton *et al.*, 1980). These boulders probably represent the larger clasts of layer 'B'. A third layer, layer 'C', also exists which underlies the glacial material and forms the bulk of the sand ridges. A

summary diagram (fig.5.2) illustrates the shelf-edge morphology and stratigraphy of this area.

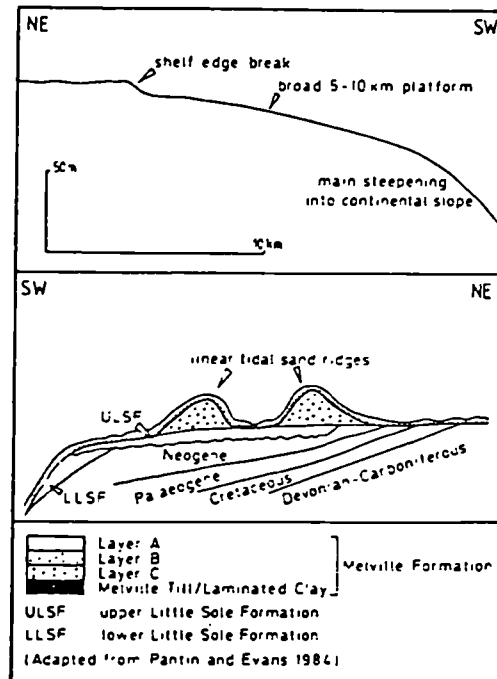


Fig.5.2 Shelf edge morphology (top) and stratigraphy (bottom) of the southwestern Celtic Sea (from Scourse *et al.*, 1990).

5.3 Regional framework

One area of the British Isles where there has been considerable debate as to the maximum extent of the Late Devensian ice sheet is within the Irish Sea Basin and around its margins (Kidson, 1977; Scourse, 1985). In this chapter the evidence for a Late Devensian grounding line offshore south west Britain is critically reviewed. The various aspects of the debate as regards this area are reviewed in two parts; firstly based upon onshore evidence, mainly from coastal sections and then secondly from the increasing amount of offshore evidence. An attempt has been made to work through the literature in a chronological order, towards the most recent, and in a progressively southwards direction since it is apparent that there is a broad correlation between the two.

5.3.a Onshore / Coastal evidence

It has to be admitted that there has always been a certain amount of controversy with regards to the Pleistocene stratigraphy of the Southern Irish Sea area and to the Quaternary history of Western Britain in general (Kidson, 1977a). The reasons behind the Irish Sea area controversy are numerous, but perhaps central to it are the 'confusing' coastal sections where the most intensive controversies arise (Bowen, 1973a). Prior to the last decade opinions with regards to the Irish Sea Basin could be divided into two groups: an "Irish" school and an "Anglo-Welsh" school (Kidson, 1977a).

The views of the "Irish" school are summarised by the works of Mitchell (1960, 1972); Mitchell *et al.* (1973); Stephens (1966, 1971) and Synge (1971), and revolve around the 'main raised beach' as seen at Courtmacsherry, Southern Ireland, the Gower coast, South Wales, Barnstaple Bay, North Devon and the Scillies (cf. Mitchell and Orme, 1967). They argue that the 'main raised beach' is Hoxnian in age and that the glacial sediments found overlying it are post-Hoxnian. The basis of dating the beach relies heavily upon the proposed limit of the Devensian glaciation. Mitchell (1972, fig.4) placed it in a line between Mathry on the West Wales coast and Shortalstown on the Irish coast. Thus, any 'main raised beach' with overlying glacial deposits to the South of this line is considered by the "Irish" school to represent a Hoxnian beach with the overlying sediments confined to a cold phase which post-dates the Hoxnian but pre-dates the Devensian. The assigned cold phase is the Wolstonian, whose limits are considered by both "schools" to have extended as far South as Southwest Britain, including the Isles of Scilly, and Southern Ireland. One major problem for the "Irish" school has been the recognition of sites belonging to the Ipswichian Interglacial, this is particularly true of Ireland where all Pleistocene interglacial sites are assigned to the Gortian (Watts, 1964). Recent work at Cork Harbour, southern Ireland (Scourse *et al.*, in prep.) supports the traditional view (Mitchell, 1981; Watts, 1985) that the Gortian should be correlated with the Hoxnian rather than the last interglacial (Warren, 1979).

However, not all of the "Irish" school agree on the Hoxnian age proposed for the 'main raised beach' and Synge (1977a) has proposed a Middle Devensian age not only for sites such as the Courtmacsherry beach but also for the Fremington Till of North Devon, which according to Kidson (1977b) "represents the most significant glacial deposit (Wolstonian) of the peninsula".

The alternative view, and that of the so-called "Anglo-Welsh" school, is that the 'main raised beach' is in fact Ipswichian in age as discussed at greater length by Bowen (1971, 1973a), Kidson (1971), and Kidson and Wood (1974). If the beaches are all Ipswichian, then within the limits of the Devensian Ice Sheet, one might reasonably expect the overlying glacial sediments, if preserved/present, to be Devensian in age. It is apparent therefore that sites outside the generally accepted Devensian ice limits are critical to the debate. In this respect, the sites at Fremington, North Devon and the Scilly Isles are of particular interest, just as those offshore and to the South of the accepted limits are too.

On the Scilly Isles tills have been described and interpreted as stratigraphically superposed to the 'beach' (Scourse, 1985); the response has been to reinterpret the till as a solifluction deposit which was originally deposited as 'till' during the Wolstonian (cf. Bowen, 1969, 1973a, 1981). The case in North Devon is somewhat different in that the Fremington Till is, with a few exceptions (Synge, 1977a), accepted as Wolstonian in age. The difficulty at this site, as far as the "Anglo-Welsh" school are concerned is that the till is underlain by cobbles which have been correlated to a nearby beach of Hoxnian age (Stephens, 1966). Correlations made with Ireland by Warren (1979) have suggested that the 'main raised beach' is Gortian, although he suggests that the Gortian may be better correlated with the Ipswichian rather than the Hoxnian as is widely accepted (cf. Watts, 1985; Scourse *et al.*, in prep.).

It is becoming increasingly apparent from amino-acid racemization of marine molluscs in raised beaches that a number of different ages now exist for these beaches which are all at similar elevations (Bowen, 1984). Furthermore, the general

assumptions made by both "schools" with regards to 'high' interglacial sea-levels and 'low' glacial sea-levels have been shown to be as incorrect. For the last glaciation in particular, a number of workers have proposed that sea levels were close to those of the present day. Recently, a number of publications have appeared in which the 'morainic' deposits of Ireland (Synge, 1977b) are increasingly interpreted as glacial marine sediments (Colhoun and McCabe, 1973; McCabe *et al.*, 1984, 1986; Eyles and Eyles, 1984; McCabe, 1985, 1987). Eyles and McCabe (1987) have suggested that the primary control on the relative sea-level at the margin of the Late Midlandian (= Late Devensian, cf. Mitchell *et al.*, 1973; McCabe, 1987) ice sheet, and possibly elsewhere, is a function of complex glacioisostatic disequilibrium.

5.3.b Offshore evidence

It is only since the 1960's that attempts have been made to bridge the marine gap between the controversial and sometimes confusing coastal exposures of Wales, Ireland and further afield. The first attempts employed bathymetric data available on Admiralty charts (Mitchell, 1960, 1963) and low frequency echosounder records (Stride and Bowers, 1961). On the basis of shallow marine ridges the first lines of maximum glacial advance were proposed; although Dobson *et al.* (1971, 1973) have demonstrated that a number of these ridges are solid rock, rather than morainic.

During the 1970's a major sampling survey was initiated by the British Geological Survey (then I.G.S.) Marine Geology Unit and a number of publications appear during this period as a result of the large amount of data gathered. The most important of these include Garrard and Dobson (1974), Garrard (1977), Delantey and Whittington (1977), Pantin and Evans (1984), Scourse (1985), and Scourse *et al.* (1990).

5.4 Facies classification and distribution

The samples analysed are reported by Scourse *et al.* (1990) to facies groups, these are 'A' and 'B'. Scourse (1985) has defined both facies lithostratigraphically, Facies A as the Melville Till

and Facies B as the Melville Laminated Clay; both are members of the Melville Formation (fig.5.2) and their distribution is outlined in fig.5.1.

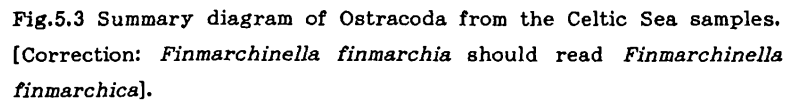
5.4.a Facies A

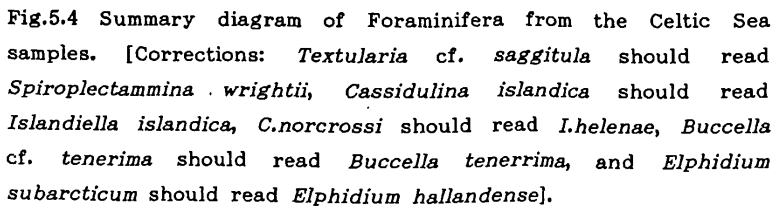
Samples which represent facies A include 49/-09/43, 49/-09/12, 49/-09/21 and 49/-09/137 as well as the lower sub-unit (unit 1) of 49/-09/44. The sediments are overconsolidated and structurally homogenous, contain abundant fine gravel (between 8 and 54 granules per 100 g. sediment, and abundant pebbles >4 mm.. The matrix is very poorly sorted and consistently coarse skewed. Two of the samples contain fragments of *Hiatella* sp. and three contain barnacle fragments. Clast lithological assemblages are consistent, with greywackes/quartzites > quartz/igneous > metamorphics > sandstones > flint > chalk (Scourse *et al.*, 1990).

Ostracod counts are low, varying between three and eleven per sample, and are characterized by both temperate and arctic species (see fig.5.3). No complete growth series are present and together with a high proportion of abraded foraminifera it is unlikely that these facies A faunas constitute *in-situ* life assemblages.

Equally, the foraminifera from these samples (fig.5.4) are of mixed arctic and temperate affinities and are highly abraded, particularly the boreal species *Quinqueloculina seminulum* and *Ammonia batavus*. Moderately high proportions of the suborder Textulariina occur and are largely accounted for by the boreal species *Spiroplectammina wrightii*. The relatively high percentage frequency of indeterminate species may, in part, be due to the high degree of abrasion.

Calcareous nannoplankton analysis (J. R. Young, British Museum, Natural History) of chalk clasts from samples 49/-09/12, 49/-09/21 and 49/-09/137 indicate an approximate Cenomanian - early Coniacian age (Appendix 3) and this has helped to constrain the possible source area of these clasts. Many of the clasts in facies A include metamorphosis associated with the Haig Fras granite intrusion and associated metamorphic aureole (fig.5.5) as well as Neogene lignite and Miocene glauconitic micrite. Together, they indicate ice provenance from the northeast towards the





southwest, across the area between Haig Fras and the Isles of Scilly, eroding various lithologies including distinctive Turonian and Miocene sediments.

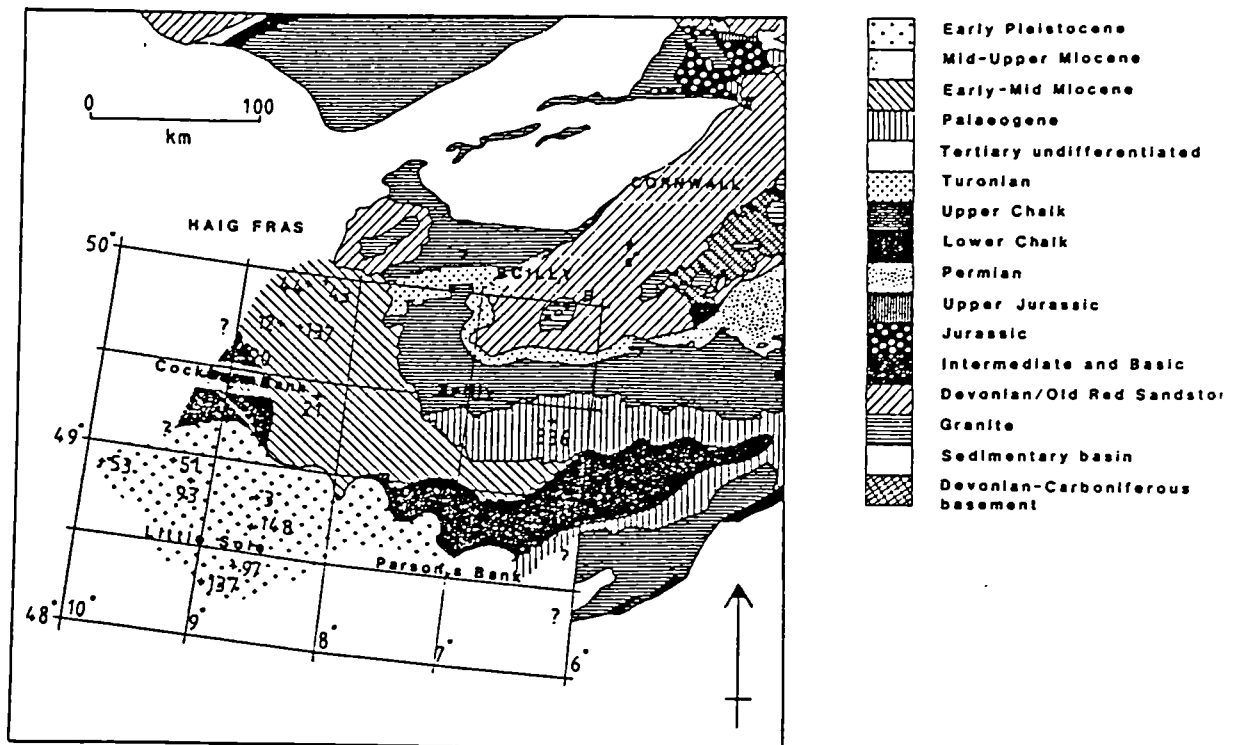


Fig.5.5 Solid geology of the central and southwestern Celtic Sea showing BGS vibrocoring sites (from Scourse *et al.*, 1990).

5.4.b Facies B

Facies B is represented by samples 48/-09/148, 49/-09/90, 48/-10/93 and 49/-09/3. The sediments are not overconsolidated but plastic silty clays, containing very small amounts of fine gravel (between 0 and 6 granules per 100 g. of sediment), and sometimes display well developed fining-upwards laminae and sand pods. They exhibit consistently very coarse skewed, moderately to moderately-poorly sorted matrices.

A rich ostracod fauna characterised by high proportions of *Rabulimys mirabilis*, *Krithe glacialis*, *Acanthocythereis dunelmensis*, *Palmenella limicola* and species belonging to the genus *Cytheropteron* (fig.5.3) help to distinguish these samples

from those of facies B. The arctic species are nearly all represented by a wide range of valve sizes which constitute a growth series; fig.5.6 represents a growth series for the arctic species *Rabilimis mirabilis* from sample 49/-09/90. Here the step-like growth pattern (a process known as ecdysis) characteristic of all Ostracoda is illustrated. When most of the moult stages are present, as in fig.5.7, then a low energy life assemblage is indicated; this is the type A assemblage of Whatley (1983).

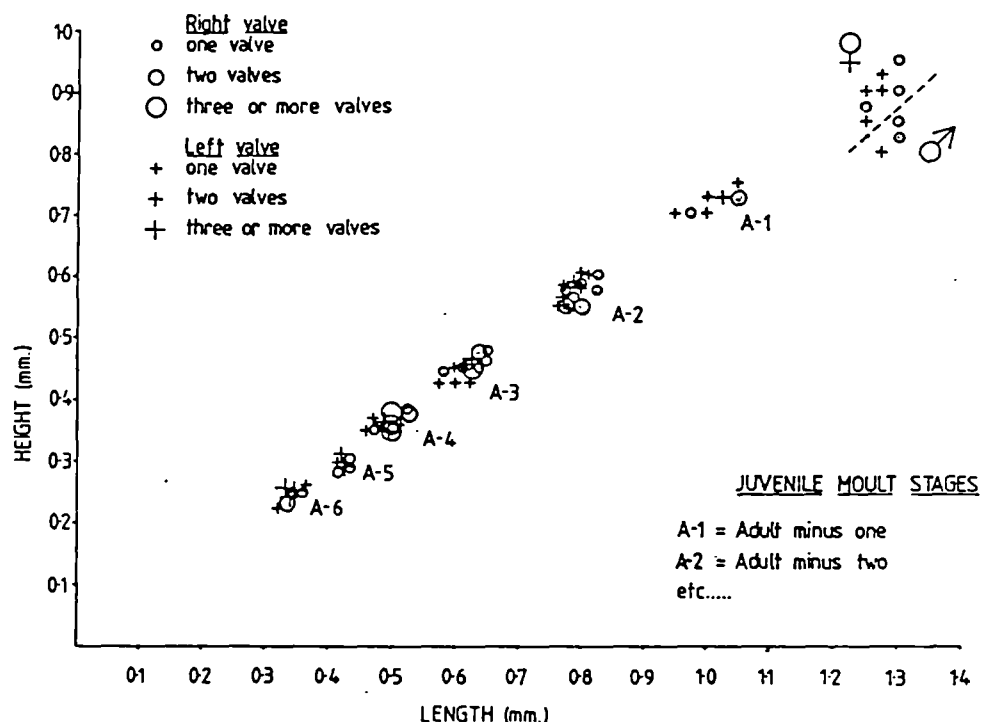


Fig.5.6 Length vs. Height growth series for the ostracod *Rabilimis mirabilis* (Brady) from sample 49/-09/90.

The foraminiferal faunas of Facies B samples are again quite diverse, with a mean alpha-diversity value of 5.1 (fig.5.8). The faunas are dominated by *Islandiella helenae*, while other important taxa, exclusive to Facies B, are *Pyro williamsoni* and *Elphidium excavatum* forma *clavata*. The planktonic species *Globigerina bulloides* also occurs in relatively high numbers here; in fact planktonic to benthonic ratios reach their maximum value of 0.08 in sample 48/-09/148. The planktonic:benthonic ratios are much lower in facies A samples.

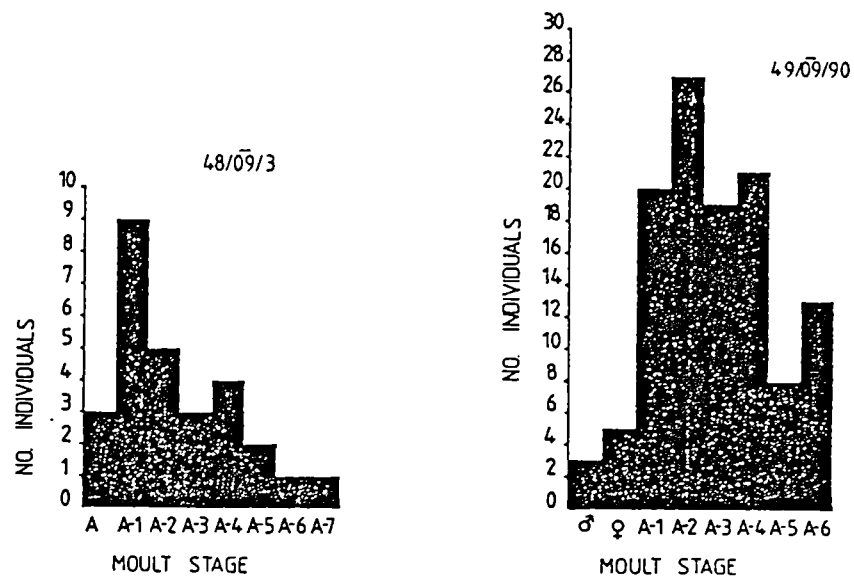


Fig.5.7 Population age-structure diagram for the species *Rabilimis mirabilis* (Brady) from samples 48/-09/3 and 49/-09/90.

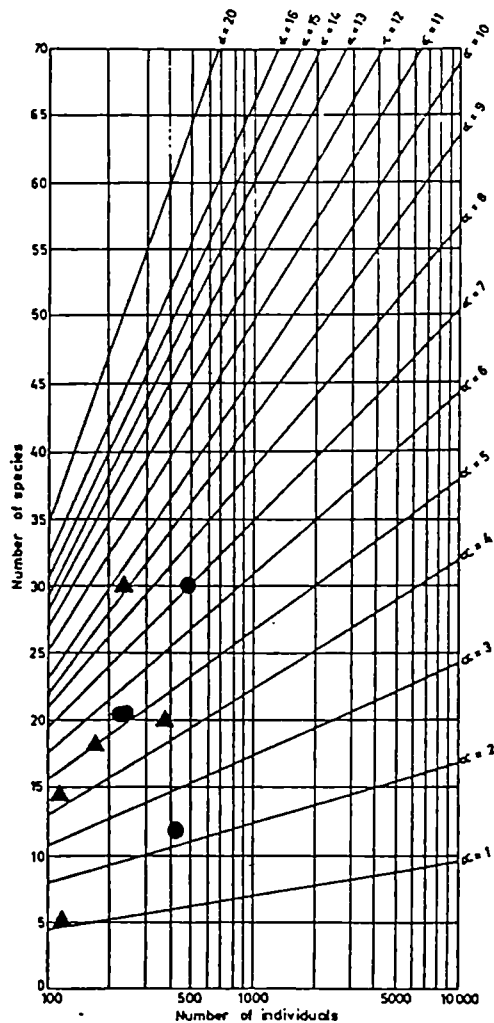


Fig.5.8 Plot of Fisher alpha-diversity indices.

= Facies B samples
 = Non-glacigenic samples
 (values are not quoted for Facies A samples due to low faunal counts).

Molluscan faunas include five valves of *Yoldiella* (*Portlandia*) *fraterna* (Verrill & Bush) in sample 49/-09/90 as well as specimens of *Arctica islandica* in sample 48/-10/93. However, many of the molluscs were fragmentary or abraded specimens. Details of the molluscan faunas are included in the appendix (Appendix 4).

5.4.c Other samples

Some of the samples did not readily fit into facies A or B. These include four samples from near the shelf edge break, 48/-09/137, 48/-10/53, 48/-09/97 and 48/-10/51, and one 49/-07/336 which was cored from much further to the east.

Sample 48/-09/97 contains a fauna which is consistent with the modern fauna from the area together with low proportions of a few cold water species. The dominant foraminifera are *Cibicides* *gr. lobatulus*, *Trifarina angulosa*, *Quinqueloculina seminulum* and *Ammonia batavus*. Sample 48/-10/51 also contains a temperate fauna and has been assigned to the Early Pleistocene upper Little Sole Formation by Scourse (1985). Sample 49/-07/336 contains no Ostracoda but has a high proportion of the suborder Miliolina as a result of the high numbers of *Quinqueloculina seminulum*. Abraded specimens of *Nummulites* *cf. rectus* Curry, which is known from the middle/late Eocene boundary of the Hampshire Basin and English Channel are present in the sand and gravels capping Eocene sediments in vibrocore 49/-07/336. It is interesting to note that samples 48/-10/53 and 48/-09/137 from near the shelf-edge break while containing a mixed temperate/arctic fauna, have lower planktonic to benthonic ratios than facies B samples.

Since the publication of Scourse *et al.* (1990), I have examined the foraminifera in three samples from vibrocore 49/-09/44. The sediments consist of overconsolidated, very poorly sorted, massive diamicts (unit 1) at the base of the core which are overlain by about 3 m. of well sorted medium silt (unit 2), with a little fine sand and clay but no clasts. A further 3 m. of coarse sands and small gravels (unit 3) overlie these deposits and generally fine upwards. The analysed samples are all from unit 2 and details of this fauna are summarized in Table 5.1.

Species	Level		
	A	B	C
<i>Elphidium excavatum</i>	85.8%	72.2%	14.3%
<i>Cassidulina reniforme</i>	4.3%	9.3%	14.9%
<i>Cibicides lobatulus</i>	0.5%	1.9%	7.5%
<i>Bulimina</i> gr. <i>marginata</i>	----	0.3%	5.4%
<i>Trifarina angulosa</i>	----	1.9%	19.7%
<i>Cassidulina laevigata</i>	----	0.3%	7.5%
<i>Spiroplectammina wrightii</i>	0.25%	----	10.2%
no. benthonic species	20	23	24
benthonic count no.	395	353	147
planktonic count no.	41	103	34
pre-Quaternary count no.	11	57	25
planktonic:benthonic ratio	0.10	0.29	0.23
pre-Quaternary:benthonic ratio	0.03	0.16	0.17
no. of specimens per 100g sed.	13,331	3,177	4,410

Where A= 4.21-4.26 m.; B= 5.50-5.54 m.; C= 6.32-6.36 m.

Table 5.1 Summary of benthonic foraminifera from vibrocore VE 49/-09/44.

The upper levels have similar benthonic faunas, dominated by *Elphidium excavatum* forma *clavata* with *Cassidulina reniforme* as the main accessory species. The sample from 4.21-4.26 m has the highest dominance at 85.8% *Elphidium excavatum* and the lowest number of species (20) together with the highest benthonic count number (395 specimens) and concentration (13,300 specimens per 100g). The sample from 6.32-6.36 m. has the lowest faunal dominance at 19.7% *Trifarina angulosa*, together with the highest number of species (24) and lowest benthonic count number (147 specimens). The dominant species in the lower sample is *Trifarina angulosa*, with *Elphidium excavatum* and *Cassidulina reniforme* as the most important accessory species together with *Cibicides lobatulus*, *Cassidulina laevigata*, *Spiroplectammina wrightii* and

Bulimina marginata. While the upper and middle levels have similar benthonic faunas, the lower and middle levels have far more similar planktonic:benthonic and pre-Quaternary:benthonic ratios.

5.5 Palaeoenvironmental reconstructions

The location of the core stations are illustrated in figs.5.1 and 5.5, but it should be stressed that there is no independent dating evidence to support the contention of Scourse *et al.*(1990) that these samples were deposited during the same glacial event. Of the fourteen sites cored, thirteen have been made available to me for micropalaeontological analysis as washed (>250 μ m.) residues. No stratigraphic data were available from the thirteen coring sites. However, Scourse (1985) suggests that their geomorphological context supports the interpretation that these samples were deposited during the same glacial event and it is on the basis of this that these sediments are interpreted.

5.5.a Facies A

Facies A samples contain mixed temperate/arctic faunas of low diversity, as well as a high proportion of abraded foraminifera and evidently do not constitute an *in situ* life assemblage. The samples have been recovered from water depths of between -127 m. and -157 m. OD. Unfortunately, the environment of deposition of these overconsolidated sediments is difficult to interpret and would be consistent with either proximal glacial marine conditions or a basal till of lodgement facies. The low planktonic:benthonic foraminiferal ratios support these interpretations; any planktonic species which do occur may well be reworked specimens associated with the temperate benthonic species which also occur in these samples.

A transitional area occurs between -127 m. and -145 m. OD at about 49°30'N, representing a change in the depositional environment from grounded to floating ice or from proximal to increasingly distal glacial marine conditions. The upcore lithological changes in vibrocore VE 49/-09/44 do suggest a change from the Facies A to B type of deposition and indeed may indicate that marine conditions became predominant northwards as former

grounded ice floated-off and calved.

5.5.b Facies B

Facies B samples contain a predominantly arctic fauna with species present which are today largely found living north of the Arctic Circle. The Ostracoda from Facies B samples appear to be particularly diverse and abundant (fig.5.3) and as figs.5.6 & 5.7 illustrate, there appear to be complete life assemblages present which suggest *in situ* accumulation under extremely quiet conditions with only limited current activity. These ostracod faunas do not contain any characteristic temperate faunas, but do contain species of the genera *Cytheropteron*, *Acanthocythereis*, *Elofsonella*, *Heterocyprideis*, and *Jonesia* which suggest affinities with the pre-Ipswichian, cold, open sea faunas of the Bridlington Crag of Holderness (Cat & Penny, 1966; Neale & Howe, 1975) at least in terms of depositional environment, but not necessarily of age (Scourse *et al.*, 1990).

The faunas of facies B are in marked contrast to the species found living in the areas today, which are dominated by *Bythocythere turgida*, *Carinocythereis antiquata*, *Celtia quadridentata*, *Loxoconcha multiflora* and *Pterygocythereis jonesi*. None of these species occur in facies B samples. The most important ostracods of facies B include: *Cytheropteron arcuatum* Brady, Crosskey & Robertson, which has been described from several glacial sites in Scotland (Brady *et al.*, 1874) but has never been found living. Other records from around the British Isles include the pre-Ipswichian marine Bridlington Crag, the Clyde Beds of the west coast of Scotland, including Loch Creran at Shian Ferry (Peacock, 1971), and the Pleistocene deposits of the Forties Field in the North Sea (Whatley and Masson, 1979). *Cytheropteron excavolatum* Whatley & Masson is a distinctively alate and pitted species described from the Pleistocene of the Forties Field (Whatley & Masson, 1979). It is not known as a living species though dead valves have been collected from Cumberland Inlet, Baffin Island (Brady & Norman, 1889), and it is undoubtedly an arctic indicator from its associations. *Cytheropteron montrosiense* Brady, Crosskey & Robertson was first described from the Clyde

Beds at Campbelltown and the Errol Beds of Tayside (Brady *et al.*, 1874). It is also unknown as a living species but is clearly an arctic indicator from its associations. *Krithe glacialis* Brady, Crosskey & Robertson was originally described from the Errol Clay where it occurs in a turbid-water, soft substrate palaeoenvironment close to an ice-front (Paterson *et al.*, 1981).

The foraminifera of facies B are more difficult to interpret than the Ostracoda mainly because of the absence of material <250 μm . However, the high numbers of *Islandiella helenae* Feyling-Hanssen & Buzas and *Iislandica* (Nørvang), together with *Elphidium excavatum* (Terquem) forma *clavata* Cushman, constitute the 'High Arctic' assemblage of Feyling-Hanssen (1955) and support the low-water temperatures suggested by the ostracods. The planktonic foraminifera also support these facies interpretations; the low planktonic:benthonic ratios of facies A samples can be explained in terms of their deposition sub-glacially or ice-proximally, while the higher values of facies B samples may reflect the more 'open' oceanic conditions and closer proximity to the shelf edge break.

5.5.c Other samples

The palaeoenvironmental interpretation of the samples which do not readily fall into either of the above facies are difficult. Sample 48/-09/97, for example, is interpreted as containing a fauna representative of the modern fauna of this area. The low numbers of a limited number of cold water species are significant and are considered to be reworked from the associated glacial deposits. Such cold water elements have been recorded in grab samples elsewhere in the Irish Sea Basin by the present author and others (eg. Haynes, 1973; J.E. Robinson pers. comm.). Samples 48/-10/53 and 48/-09/137, which are located near the shelf-edge break have faunal characteristics which are typical of facies A samples, containing mixed temperate and arctic faunas, with low planktonic:benthonic ratios. These samples may represent residual glacial marine deposits, much as proposed in glacial marine models for the George V-Adelie continental shelf edge and slope in Antarctica (Domack, 1982).

Interpreting the three samples from unit 2 of vibrocore VE 49/-09/44 is again problematic and correlating these faunas, picked from $>63\ \mu\text{m.}$, with these reported earlier in Scourse *et al.* (1990) is equally difficult. The high frequencies of *Elphidium excavatum*, for example, are accounted for by specimens with maximum test diameters generally $<150\ \mu\text{m.}$ and would not, as I have discussed for the Aberdaron section (Chapter 4), have been recorded following the processing adopted by Scourse (1985).

However, assuming that the three major units of this vibrocore represent lodgement till (cf. facies A), glacial marine deposits (cf. facies B) and Holocene sands respectively, then all three samples should bear an affinity to facies B samples. This is true of the two upper levels which suggest extreme environmental conditions, possibly with reduced salinities, and are typical of high-turbidity glacial marine deposits. The lower frequency of *Cassidulina reniforme* (4.3%) and lower planktonic:benthonic ratio (0.10) of level 4.21-4.26 m. in comparison to level 5.50-5.54 m. may indicate a more extensive ice-cover, acting to reduce the water salinity and inhibit the transport of planktonic tests onto the shelf. The lowest level, at 6.32-6.36 m., is problematic in that a more diverse fauna occurs, more reminiscent of Boreal water faunas than the High Arctic, ice proximal faunas of the levels above. One possible explanation is that this lower level is indicative of ameliorated conditions associated with the Lateglacial Interstadial and that the levels above represent cooler deposition during the Younger Dryas; the sands of unit 3, with their abundant Turritellid and Scaphopod faunas, would represent Holocene sediments. It might equally be argued that the faunas at this level are mixed, containing glacially reworked temperate and pre-Quaternary species. In fact the pre-Quaternary:benthonic ratio is at its highest at this level.

Even allowing for the reworking of certain species into these levels, it is interesting to note that the ratio of *Cassidulina reniforme* to *Elphidium excavatum*, both of which may be considered to co-occur in many environments, change as follows: level 6.32-6.36 m. = 1.04; level 5.50-5.54 m. = 0.13; 4.21-4.26 m. = 0.05. These values, assuming that they are not altered by

allochthonous faunal components, suggest that there is indeed a real faunal response within unit 2 which is due to changing palaeoenvironmental conditions associated with ice cover. Clearly further work, which was beyond the scope of this sample exercise, is now required to establish the full bio-stratigraphic succession in this vibrocore.

5.6 Dating

As discussed by Scourse *et al.* (1990, 1991) there is no independent dating evidence available which can help place all the samples within the same geological period. In an attempt to partly redress this problem a single sample (BAL 2120), from vibrocore 49/-09/44, has yielded an aIle/Ile ratio of 0.039 from a monospecific sample of 200 small tests of the benthonic foraminifera *Elphidium excavatum* (Terquem) forma *clavata* Cushman. This ratio, based upon a taxon of clearly arctic affinity within an assemblage characteristic of glacial marine environments (level 4.21-4.26 m., table 5.1), falls within group 1 of Knudsen and Sejrup (1988) based upon ratios from the same species in the North Sea area. The latter propose that group 1 ratios, typically between 0.03 and 0.04, are characteristic of the Late Weichselian of the area as did earlier studies on the Norwegian continental shelf (Sejrup *et al.*, 1984) and from onshore western Norway (Miller *et al.*, 1983). Therefore, even allowing for the fact that the site under investigation is some distance from the North Sea and an inter-regional temperature gradient may play a slight role in the rate of the isoleucine epimerization reaction over a possibly short glacial period, it seems reasonable to correlate this ratio with those of group 1 and therefore to propose that glacial marine conditions did exist off shore south west Britain during the Late Devensian.

However, even if these sediments are chronologically related and represent different facies which have accumulated during the same glacial period, there are still problems associated with the palaeoenvironmental interpretations within the context of the Lateglacial period.

5.7 Regional synthesis

The evidence points to the facies B deposits having being deposited by ice-rafting, an hypothesis which is favoured by Pantin and Evans (1984) who suggest that 'mounds' of sediment visible on side-scan sonar records are indicative of individual iceberg 'dumps'. While some of these features may have such an origin, it seems likely that the ice-cover was more extensive than this, on the basis of the low planktonic:benthonic ratios and the undisturbed ostracod growth series. The possibility of deposition beneath a floating ice shelf rather than from icebergs is discussed by Scourse *et al.* (1991) but considered unlikely in view of the physically constrained minimum thickness of ice sleeves between 200-250 m. (Sugden & John, 1976; Paterson, 1981). If the fossil grounding line is taken to be -135 m., then to float an ice shelf at least 200 m. thick would require sea levels at around +30 m. OD; this assumes an isostatic rebound component of $0.33 \times$ ice thickness and a 4:1 ratio of submergent to emergent ice. As Scourse *et al.* (1991) point out, there is no evidence in the region as a whole for raised shorelines post-dating the Ipswichian Stage (see section 5.3).

The recovery of glacial samples cored in 'lows' between the tidal sand ridges suggest to Scourse *et al.* (1990) that glacial deposition took place after the main period of formation of the sand ridges. The sand ridges themselves are considered by Pantin and Evans (1984) to have formed in about 60 m. of water by the Huthnance mechanism (Huthnance, 1982) during the early stages of the Devensian-Holocene transgression and are referred to as "static" rather than "moribund" (Stride *et al.*, 1982). Generation of these features during the early phase of sea level rise is implied, since it seems unlikely that these features, up to 60 m. high, could have survived subaerial exposure which they would have experienced had they originated earlier. Equally, if Scourse *et al.* (1990) have correctly identified the glacial deposits as post-dating the formation of these ridges, then a Late Devensian age is implied not only for the sand ridges, but also the overlying glacial sediments.

This reconstruction also implies the survival of the sand

ridges during the succeeding glacial 'event' and the sand ridges are known to stop fairly abruptly along a northwest/southwest line at about the facies A/B transition (fig.5.1). Scourse *et al.* (1991) suggest that any sand ridges that may have existed to the north east of their present limit were eroded out by the grounded ice sheet, iceberg rafting taking place over, and on the flanks of, the sand ridges to the south west. Sample 49/-09/44 is considered to be crucial to this argument (Scourse *et al.*, 1991) who state that there is no doubt it is located on the side of a sand ridge. However, as far as I can tell from the lithological logs accompanying this core, there is no evidence to suggest that glacial sediments overlie sands and gravels which might be associated with the linear tidal sand ridges. The deposits of this core are interpreted as representative of glacial deposits overlain by Holocene sands as discussed above (section 5.5.c).

Regional reconstructions during the Late Quaternary aim to correlate the offshore glacial sediments with the Scilly Till (Scourse, 1991), which resemble each other both lithologically (Scourse, 1985) and mineralogically (Catt, 1986). The altitudes of the various sedimentary bodies in the region agree with the available information. Assuming that the offshore grounding line was at -135 m. OD and ice thickness was about 100 m., global stadial sea level would have stood somewhere between 100 m. and 50 m. below the present, with a post-rebound shoreline below present sea level. This is consistent with the lack of a raised Late Devensian shoreline on the Scilly Isles or in Cornwall. Such a shoreline could only have been generated by ice thicker than 250 m. or by a contemporary global sea level close to present sea level which is thought highly unlikely in this area during the Late Devensian.

Chapter 6: Overview

6.1 Aims and Objectives

This thesis begins with a list of specific objectives; a brief discussion of these aims and the extent to which they have been realized now follows.

6.1.1 To establish a working taxonomic knowledge of the foraminiferal faunas of glacial marine and associated sediments from the western U.K. shelf seas. To briefly describe and fully illustrate this fauna in a series of S.E.M. photographs.

The systematic descriptions of the species encountered are presented in Appendix 1, volume 2 of this thesis and provide a relatively comprehensive treatment of boreo-arctic and temperate species likely to be encountered in Quaternary marine sediments from western Britain. Certain groups have been studied more intensively than others, while some, such as the Polymorphinidae, have largely been left untreated. However, over 200 species and forms have been described and illustrated during the present investigation.

Thus, the first aim of the study has been achieved and may prove most useful in that both temperate and boreo-arctic faunas are brought together within the same volume. Certainly, the publications of Haynes (1973) and Murray (1971) provided excellent references for studies on temperate and Recent faunas, but the boreo-arctic faunas, particularly of western Britain are less well known. I hope that this work will therefore be of some value to the student examining foraminiferal stratigraphies which record the climatically variable episodes of the late Quaternary.

6.1.2 To reconstruct palaeoenvironmental changes during the Lateglacial period on the basis of included foraminiferal assemblages in sediments from the Hebridean Shelf, N.W. Scotland.

These reconstructions have been attempted and the sequence of events affecting the area discussed in chapter 3. It is demonstrated how micropalaeontological analysis can define a seven-zoned subdivision of the Lateglacial-Holocene period within a vibrocore (VE 57/-09/89) less than 6 m. long, which corresponds

to the established climato-stratigraphic sequence of events for the period. Independent dating evidence supports these views and allows inferences to be made relating to relative sea-level and isostatic adjustment in the area following deglaciation. A number of other vibrocores have been analysed and correlated with VE 57/-09/89; however, where no independent dating evidence is available, such correlations are extremely difficult.

6.1.3 To define and date a proposed Late Devensian grounding line from offshore S.W. Britain on the basis of included microfossil (Foraminifera and Ostracoda) and other evidence.

This forms the basis of chapter 5 on the Southern Celtic Sea and is partly based upon work included within Scourse *et al.* (1990). While the microfaunas do help to define the facies changes upon which this grounding-line is based, no dating evidence is available to relate the samples analysed to the same geological period. However, amino acid analysis of a sample of foraminiferal tests from VE 49/-09/44 yielded an *Ala/Ile* ratio which correlates with similar ratios assigned to the Late Weichselian of the North Sea area. Thus, it appears that samples assigned to Facies B from this area do indeed represent Late Devensian glacial marine deposits, although their relationship with the apparently underlying linear tidal sand ridges is problematic. Thus, the definition and date of this proposed grounding line remains unresolved and requires further detailed study, particularly with regards to stratigraphic correlation and dating.

6.1.4 To determine whether onshore diamicts, exposed in a cliff section at Aberdaron, N.W. Wales, are glacial marine or terrestrial in deposition origin, based upon their included foraminiferal and lithofacies characteristics.

The controversy regarding the depositional origin of diamicts surrounding the Irish Sea Basin has increased recently with the more widely held view that these deposits, for many years interpreted as terrestrial, are glacial marine. Fundamental to the arguments of this "glacial marine school" is the suggestion that relative sea levels may be raised up to +140 m. OD in response to isostatic crustal depression; this is the theory of

glacio-isostatic disequilibrium (Andrews, 1982). Furthermore, such ice margins appear to be inherently unstable and their deglaciation need not involve climatic amelioration, but may respond to rising relative sea-level generated by isostatic down-warping of the crust (eg. Eyles and McCabe, 1989).

The foraminiferal evidence from Aberdaron and the lithofacies work of McCarroll and Harris (in prep.) suggests that these deposits have features characteristic of both glacial marine and terrestrial deposits. However, in view of the mixed nature of the assemblages and the faunal homogeneity of the sequence as a whole, the final depositional context is interpreted as one of basal melt-out from stagnant ice rich in marine debris entrained during its passage along the northern Irish Sea Basin. Thus, the foraminiferal evidence particularly supports this reworked origin, while the lithofacies work favours a terrestrial depositional setting for the sequence.

6.1.5 To attempt a preliminary scheme of glacial marine facies characterization based upon the included foraminiferal faunas of the sediments investigated.

This has been the most ambitious aim of the present study, and possibly the least realistic to achieve. I have demonstrated how lodgement, proximal, and distal glacial deposits can be differentiated on the basis of their foraminiferal faunas by distinguishing between autochthonous and allochthonous components, the degree of test abrasion and damage, the range of test diameters (ie. population age-structure diagrams), and the ratio of planktonic:benthonic species. These features and other faunal characteristics which help to define the general facies types are outlined in Table 6.1.

dominant species	variable, <i>E.excavatum</i> or <i>C.reniforme</i>
common accesssory species	variable but often including temperate species eg. <i>B.marginata</i> or <i>C.laevigata</i> .
faunal diversity	generally high but variable \cong 20+ species
faunal dominance	variable, generall <40%
p:b ratio	low and variable <0.2
benthonic specimens per 100 g.	variable c.10,000 specimens
degree of test abrasion	variable, temperate species eg. <i>E.crispum</i> or <i>Q.seminulum</i> often highly abraded
faunal remarks	often mixed boreo-arctic/temperate assemblages with pre-Quaternary species
examples	Aberdaron; Facies A, southern Celtic Sea; zone 1 VE 57/-09/89

Table 6.1.a Summary of the faunal characteristics of Lodgement facies.

dominant species	<i>E.excavatum</i> often dominant & <i>C.reniforme</i>
common accesssory species	variable but often boreo-arctic species eg. <i>E.asklundi</i> or <i>N.orbiculare</i> .
faunal diversity	generally low \cong 10 species
faunal dominance	high, often >60%
p:b ratio	very low <<0.1
benthonic specimens per 100 g.	very low <<10,000 specimens
degree of test abrasion	generally well preserved
faunal remarks	generally low diversity boreo-arctic assemblages of lowered salinity
examples	zone 3 VE 57/-09/89

Table 6.1.b Summary of the faunal characteristics of Proximal facies

dominant species	<i>C.reniforme</i> often dominant & <i>E.excavatum</i>
common accessory species	<i>Islandiella</i> spp. <i>Nonion labradoricum</i>
faunal diversity	moderate \cong 15 species
faunal dominance	high \cong 50-60%
p:b ratio	low \cong 0.1
benthonic specimens per 100 g.	moderate \cong c.25,000 specimens
degree of test abrasion	generally well preserved
faunal remarks	increased diversity boreo-arctic assemblages, environment stable
examples	zone 6 VE 57/-09/89 zone 1 VE 57/-09/46

Table 6.1.c Summary of the faunal characteristics of Distal facies.

While it is therefore possible to characterize the general faunal composition of these facies types, any more detailed facies characterization, based upon the present study, is unattainable. However, samples A1 and A7 from Aberdaron, for example, can be distinguished from those of the upper and lower diamict associations by their distinctive population-age structure profiles and this may prove a useful tool in recognizing fluviatile deposits (hydrodynamically sorted faunas); although these are lithologically quite distinct already! Subdivision of distal facies, based upon inferred water depths, might also be possible; this is particularly relevant during the Younger Dryas when relative sea levels are rising rapidly on many continental shelves. One might, for example, subdivide the Younger Dryas into an early and late stage, based upon the increasing frequency of *Nonion labradoricum*.

6.2 Methodological considerations

In this section some of the more important aspects of the methodology are discussed which have not yet been dealt with. Some aspects, such as the use of a 63 μ m. sieve, do not require further discussion; in the case of sieve size refer to section 2.4.2 and chapter 4.

6.2.1 Heavy-liquid separation

This method is outlined in section 2.4.4 and as mentioned there are problems associated with this technique. Various authors (eg. Haynes, 1981) have mentioned the disadvantages of the heavy-liquid technique, citing infilled specimens, agglutinated forms, and broken specimens as being less buoyant and therefore under-represented in the 'light' fraction; the results obtained are also reported to be visible. It is implied that any quantitative work must involve searching the 'heavy' residue, so that two operations become necessary.

In an attempt to address some of these problems I have inspected and counted foraminifera in both 'light' fraction and 'heavy' residue from a number of samples. The results of this study are presented in Table 6.2 and suggest that the 'heavy' residue and 'light' fractions are faunally very similar. Index of affinity values range from 71.7% to 81.3% and these are close to the range of 72% to 88% quoted by Rogers (1976) for samples with 35 species present. However, the 'heavy' residues were noted to commonly contain large numbers of broken and abraded specimens which were largely absent from the 'light' fraction; while these abraded specimens were not quantified, they were often estimated to account for up to 90% of the 'heavy' residues.

Further indication of the faunal similarities between 'heavy' and 'light' fractions comes when the ranked order of species and their percentage frequencies are compared; this is particularly true of level 2.11-2.13 m. in VE 57/-09/89. When the estimated benthonic numbers per 100 g. of sediment are compared from the two fractions, it is possible to calculate the percentage recovery of the total fauna within the 'light' fraction, and this varies from 79.3% to 89.5%. It therefore appears that at least 10% of the fauna remains behind within the 'heavy' residue after floatation.

However, the addition of frequency data from the low benthonic sums of the 'heavy' residue would alter the final percentage frequency very little. For example, within the level between 2.13-2.15 m. in VE 57/-09/89, the dominant species of both 'heavy' and 'light' fractions is *Cibicides lobatulus* with frequencies of 31.8% and 22.5% respectively. These percentage frequencies represent estimated total sample count frequencies of 7,080 species and 1,169 specimens respectively.

		'Light'	'Heavy'
sample count no.		445	89
no. of species		36	20
no. of common species		15	
index of affinity		76.5%	
p:b ratio		0.06	0.06
no. specimens per 100 g.		31,412	6,675
% of total in each fraction		82.5%	17.5%
Ranked order of species	1	<i>C.lobatulus</i> 27.2%	<i>C.lobatulus</i> 48.3%
	2	<i>E.excavatum</i> 21.3%	<i>A.batavus</i> 14.6%
	3	<i>A.batavus</i> 11.0%	<i>E.excavatum</i> 13.5%
	4	<i>C.reniforme</i> 7.0%	<i>I.helenae</i> 9.0%
	5	<i>T.angulosa</i> 6.7%	<i>C.reniforme</i> 6.7%
	6	<i>B.marginata</i> 4.7%	<i>B.marginata</i> 5.6%

Table 6.2.a Summary of the main faunal characteristics of the 'heavy' and 'light' fractions from VE 57/-09/89 1.94-2.00 m.

		'Light'	'Heavy'
sample count no.		459	184
no. of species		35	27
no. of common species		17	
index of affinity		71.7%	
p:b ratio		0.08	0.03
no. specimens per 100 g.		125,753	32,767
% of total in each fraction		79.3%	20.7%
Ranked order of species	1	<i>S.fusiformis</i> 45.6%	<i>S.fusiformis</i> 28.3%
	2	<i>C.lobatulus</i> 13.3%	<i>C.lobatulus</i> 17.9%
	3	<i>S.loeblichii</i> 8.3%	<i>S.loeblichii</i> 13.0%
	4	<i>E.excavatum</i> 6.3%	<i>E.excavatum</i> 8.2%
	5	<i>Q.stalkeri</i> 5.0%	<i>Q.stalkeri</i> 4.9%
	6	<i>A.batavus</i> 2.8%	<i>A.batavus</i> 3.3%

Table 6.2.b Summary of the main faunal characteristics of the 'heavy' and 'light' fractions from VE 57/-09/89 2.11-2.13 m.

		'Light'	'Heavy'
sample count no.		400	170
no. of species		33	24
no. of common species		18	
index of affinity		81.3%	
p:b ratio		0.12	0.05
no. specimens per 100 g.		31,469	3,679
% of total in each fraction		89.5%	10.5%
Ranked order of species	1	<i>C.lobatulus</i> 22.5%	<i>C.lobatulus</i> 31.8%
	2	<i>E.excavatum</i> 14.3%	<i>E.excavatum</i> 11.2%
	3	<i>A.batavus</i> 7.5%	<i>A.batavus</i> 8.8%
	4	<i>S.fusiformis</i> 6.3%	<i>I.helenae</i> 5.3%
	5	<i>C.reniforme</i> 6.0%	<i>B.marginata</i> 4.1%
	6	<i>T.angulosa</i> 4.8%	<i>S.fusiformis</i> 3.5%

Table 6.2.c Summary of the main faunal characteristics of the 'heavy' and 'light' fractions from VE 57/-09/89 2.13-2.15 m.

The sum of both counts is equal to 8,249 specimens, while the estimated number of specimens from both fractions is 35,148 specimens. The estimated percentage frequency of *C. lobatulus*, based upon both fractions, is 23.5% and does not represent a significant deviation from the 'light' fraction value (22.5%). It is therefore concluded that counting specimens from the 'light' fraction only is sufficiently representative of the total fauna for the purpose of the present quantitative study and, together with its advantages of concentrating foraminifera and greatly reducing the picking time, heavy liquid separation is considered to be a valuable technique.

6.2.2 Foraminiferal counting

This becomes an area of concern as far as 'repeatability' of the faunal results obtained is concerned. There appear to be three main ways in which the reliability of a faunal count on the same sample may be altered. The first is by poor taxonomic practice and inconsistency in naming on the part of the investigator; this is not considered to be a problem in the present study but 'subjectivity' in species naming is one of the reasons why I considered the inclusion of a systematic section within this thesis important. The second effect may be produced by unevenly distributed size/shape sorting of foraminifera over the picking tray; this is why the counting method outlined in section 2.5 was adopted. The third effect may arise when sample counts are too low and therefore produce statistically unreliable results; the details of count numbers are dealt with in section 2.5.1.

In a brief attempt to assess the 'repeatability' of the counting techniques employed two counts on the same sample (0.75-0.80 m.) from VE 57/-10/17 yielded the following ranked order of species (>1%):

Count 'A'		Count 'B'	
<i>Cibicides lobatulus</i>	29.6%	<i>Cibicides lobatulus</i>	27.0%
<i>Elphidium excavatum</i>	22.2%	<i>Spiroplectammina wrightii</i>	24.2%
<i>Spiroplectammina wrightii</i>	14.8%	<i>Elphidium excavatum</i>	17.6%
<i>Trifarina angulosa</i>	8.6%	<i>Trifarina angulosa</i>	5.8%
<i>Cassidulina reniforme</i>	3.4%	<i>Nonion orbiculare</i>	4.8%
<i>Nonion orbiculare</i>	3.2%	<i>Cassidulina laevigata</i>	3.0%
<i>Islandiella helenae</i>	3.2%	<i>Bulimina gr. marginata</i>	3.0%
<i>Cassidulina laevigata</i>	2.7%	<i>Cassidulina reniforme</i>	2.0%
<i>Nonion labradoricum</i>	2.0%	<i>Nonion labradoricum</i>	1.8%
<i>Rosalina anomala</i>	1.5%	<i>Islandiella helenae</i>	1.8%
<i>Islandiella islandica</i>	1.2%	<i>Planorbulina distoma</i>	1.2%
<i>Elphidium asklundi</i>	1.2%	<i>Quinqueloculina seminulum</i>	1.0%
<i>Bulimina marginata</i>	1.2%	<i>Islandiella norcrossi</i>	1.0%
<i>Quinqueloculina seminulum</i>	1.0%	<i>Rosalina anomala</i>	1.0%

The index of affinity of these two counts (based upon all the common taxa) is 82.8% and suggests that these counts, from a very simple statistical point of view, represent essentially the same sample. I have not attempted to calculate any confidence limits on any of the frequencies calculated and it would clearly be impractical to do so at every level. However, the programme POLLDATA is able to produce output files of frequency data which incorporate 95% confidence limits and I have included plots of these for vibrocores VE 57/-09/89 and 57/-09/46 in the document folder at the back of this volume (Enclosures 3 & 4). Such diagrams can be particularly useful in deciding whether or not changes in the frequency of taxa are statistically significant; Maher (1972) has discussed the question of confidence limits when interpreting pollen diagrams.

6.2.3 Lithological vs. Faunal changes

There are a number of unconformities recognized, particularly in the vibrocores from the Hebridean shelf, where sharp lithological changes occur eg. at the base of the Holocene. Section 3.9.3 is an attempt at examining the faunal changes which

occur at one such lithological boundary, in this case to specifically look at the evidence for bioturbation. Unlike the lithological transitions, the faunal transitions are not always as distinct and 'tails' in the frequency of taxa may occur. For example, many of the Holocene sands which overlie Lateglacial deposits on the Hebridean shelf contain boreo-arctic species which are derived from the underlying sediments; generally, the frequencies of these reworked specimens decrease rapidly upwards. The vibrocoring process is not considered a likely mechanism to produce these 'upwards tails', although bioturbation may be a possible mechanism to do this.

'Downwards tails' are likely to arise in two ways and these mechanisms and their detection are outlined in section 3.9.3 and 3.10.2.a. Bioturbation, as recognized within zone 6 of VE 57/-09/89, can have a marked effect on the micro-faunas, since foraminifera are easily remobilized; this is particularly true when marked differences in faunal concentration per unit weight of sediment occurs. Such 'tails' will generally stop abruptly at the lower limit of the bioturbation and need not decline towards this lower limit, although this will depend on the type and extent of the burrowing structures. The other mechanisms which can displace foraminifera downwards relates to sidewall smearing of sediments during the vibrocoring process; however, this problem can be overcome if sub-samples are removed from the inner part of the core barrel.

Clearly, in view of the aims of this study or any other study involving palaeoecological reconstructions, it is critical to distinguish between the autochthonous and allochthonous faunal elements which make up an assemblage if the palaeoecological signal is to be correctly deciphered. This has become a recurring theme during the present investigation and requires further detailed investigation of percentage frequency and concentration data from a number of sites. It appears that many of the faunal changes in these cores are probably as abrupt as the accompanying lithological changes, except that they become slightly blurred by the above mentioned process.

6.2.4 Dating methods

The accelerator radiocarbon dates from the Hebridean shelf have been essential in establishing chronostratigraphic control of the climatostratigraphic subdivisions proposed for the area eg. VE 57/-09/89. Radiocarbon dating, critical in dating the Lateglacial period, is fraught with problems, yet progress in the understanding of environmental changes relies upon improving the reliability of age estimates. These problems are probably greater in the terrestrial setting and include uncertainties in sample integrity, poor stratigraphic resolution, and the effects of temporal variations in atmospheric C^{14} activity. Assuming that the first two are minimal in the present investigation, I will briefly review the latter here. Recent work by Lotter (1991) on the absolute dating of the Lateglacial period at Rotsee, central Switzerland using annually laminated lake sediments and comparing them to high-resolution a.m.s. C^{14} chronology, suggests that atmospheric radiocarbon concentrations were not constant between 13,000 and 9,500 BP, resulting in three marked phases of constant radiocarbon age at 12,700 BP, 10,000 BP, and 9,500 BP. The latter two plateaux (ie. phases of constant C^{14}) each lasted c. 400 calendar years. Thus, the apparent synchronism of events during the Lateglacial period eg. the Lateglacial-Holocene transition at 10,000 BP, is probably due to the grouping of several events and implies that many of these events may be diachronous. As far as I am aware, no such plateaux have been recorded from the marine realm and are probably unlikely to be resolved in view of the relatively long-lived reservoir effects encountered and the absence of a high resolution, absolute time scale of calendar years from marine sediments. However, certain long-lived biological or chemical banding-structures may eventually prove useful in absolute dating the marine record of this period.

This possibly more widespread diachroneity, together with the timing of Lateglacial "events" from around the margins of the N.E. Atlantic, is convincing support for the need to adopt a climatostratigraphic approach to the stratigraphy of this period. However, while some of the "events" which help define the climatostratigraphy of this period are diachronous, others such as the Younger Dryas climatic deterioration may not even be recorded

from some regions. For example, Mangerud and Svendsen (1990) fail to recognize the Younger Dryas "event" from Svalbard. Thus, a climatostratigraphic subdivision, as outline in fig.1.2.b, becomes meaningless outside N.W. Europe; while a chronostratigraphic subdivision, as outlined in fig.1.2.a, is too rigidly defined and is probably applicable only to Norden. It may be preferable to talk of "time-slices", eg. 11,000 BP to 10,000 BP, and ideally one would adopt this approach, except that obtaining the stratigraphic control to define such "time-slices" would be both costly and time consuming. I have demonstrated from VE 57/-09/89 how relatively small stratigraphic intervals may represent considerable periods of non-deposition and/or erosion. I therefore conclude that the climatostratigraphic subdivision of the Lateglacial period is the most valuable approach and together with ams radiocarbon dating and interregional correlation via tephrochronology will provide the best understanding of climate dynamics during this period.

6.3 Future perspectives and recommendations for further work

During the course of this work it has become apparent that the single most important limiting factor to palaeoecological reconstruction is our understanding of modern benthonic foraminiferal ecology. Insufficient is known of the physical and chemical requirements of specific taxa. Murray (1991) has gone some of the way towards remedying this situation, but in my view considers species in too general a context, both geographical and environmental. However, this kind of work is essential and should become easier to utilize with the expanding availability of personal computer database systems.

As the ecological requirements of benthonic foraminifera become better known and understood, it may be possible to write transfer function equations, as applied to planktonic foraminifera (cf. Imbrie & Kipp, 1971), so that palaeoenvironmental reconstructions become more objective and less subjective. Whether such an approach with benthonic foraminifera can ever be as successful as it has been with planktonic foraminifera is a matter for debate; certainly benthonic foraminiferal ecology is far more complex and appears to be affected by many more factors than planktonic foraminifera. This is clearly an area of considerable

potential for the future, but requires increased 'ground-truth' knowledge in the first place.

Detailed biofacies models of the glacial marine environment are required and these must come from modern glacial marine environments, of which there is now increasing knowledge of the physical and chemical processes acting (cf. Dowdeswell & Scourse, 1990). Such models will help with the interpretation of Quaternary deposits. However, in view of the high degree of faunal reworking associated with glacial marine deposits, the need to base these biofacies models on live foraminiferal data (using protoplasmic staining techniques) cannot be overemphasised. This is why biofacies models based upon Quaternary deposits will prove difficult to interpret until the distributions of living taxa within modern analogue environments are understood.

Dating techniques need to be improved so that they can be more widely applied to benthonic foraminifera. This is currently an area of increased research activity and is being partly developed by J.T. Andrews, A.J.T. Jull and others at the Department of Geological Sciences, University of Colorado. Andrews (1991, I.G.C.P. 253 workshop, London) reports that accelerator radiocarbon dates are routinely obtained on monospecific samples of 500 foraminiferal tests. However, it appears that discrepancies commonly arise when molluscan and foraminiferal radiocarbon dates are compared from the same stratigraphic level (K.L. Knudsen, pers. comm. 1990). The future potential of this procedure will be in dating sediments devoid of molluscan remains and may be particularly valuable in dating discrete events within a sedimentary sequence such as the peak of a volcanic ash layer.

It is in conjunction with AMS radiocarbon dating that tephrochronology may become an extremely valuable tool in the correlation of marine and terrestrial records. At present, correlating the often high resolution, but equally often fragmentary terrestrial record with the generally low resolution, but often continuous record of the deep sea is extremely difficult. Shelf sea marine records may prove a valuable link between the two and together with tephrochronology to overcome the problems of correlating the various radiocarbon chronologies is likely to become an area of continued research during the next few

years.

Finally, and just as important as understanding the ecological requirements of foraminiferal species, is their correct identification. Further systematic studies are required to reduce subjectivity in foraminiferal identification and these should ideally accompany distributional or stratigraphic investigations.

AMS RADIOCARBON DATES FROM THE CONTINENTAL SHELF SOUTH OF ST. KILDA

(A) VIBROCORE 57/-09/89

Lab No.	Species	Depth in core (m)	Conventional Age (¹⁴ C years BP ± 1σ)	Adjusted Age (¹⁴ C years BP ± 1σ)
0xA-2780	<i>Pinoclea ovata</i>	0.50 - 0.60	5,960 ± 80	5555 ± 90
0xA-2781 ¹	<i>Nuculoma belloti</i>	0.70 - 0.75	11,040 ± 110	10,635 ± 120
0xA-2782 ²	<i>Parvicardium ovale</i>	1.94 - 2.00	11,440 ± 120	11,035 ± 130
0xA-2783 ³	<i>Nucula nucleus</i>	2.25 - 2.35	12,030 ± 120	11,625 ± 130
0xA-2784	<i>Portlandia arctica</i>	2.50 - 2.55	13,920 ± 140	13,515 ± 150
0xA-2785	<i>Portlandia arctica</i>	3.10	15,650 ± 160	15,245 ± 170

(B) VIBROCORE 57/-09/46

0xA-2786 ⁴	<i>Acanthocardia echinata</i>	0.47 - 0.51	10,380 ± 100	9,975 ± 110
0xA-2787	<i>Nuculoma belloti</i>	1.05 - 1.30	10,580 ± 100	10,175 ± 110
0xA-2788	<i>Nuculoma belloti</i>	4.80 - 5.00	11,420 ± 120	11,015 ± 130
0xA-1324	<i>Buccinum terraenovae</i>	4.50	11,680 ± 240	11,275 ± 250

1. $\delta^{13}\text{C PDB} = -0.1 \text{ ‰}$
2. $\delta^{13}\text{C PDB} = -0.1 \text{ ‰}$
3. $\delta^{13}\text{C PDB} = -0.2 \text{ ‰}$
4. $\delta^{13}\text{C PDB} = -0.6 \text{ ‰}$

Adjusted ages based on an apparent age of 405 ± 40 years for seawater (Harkness 1983) and $\delta^{13}\text{C}$ zero (assumed). 0xA-1324 from Hedges *et al.*, (1988).

Appendix 2: AMS Radiocarbon dates from the Hebridean Shelf

Samples		12	21a	21b	137a	137b
Overall abundance		C	A	C	A	F
Preservation		02	01	02	03	03
Species	Range					
<i>Biscutum</i> sp.	EJur – Mst		R	R		
<i>Braarudosphaera</i> sp.	Jur – Trt	R				
<i>Broinsonia enormis</i>	Cen – Cmp		F		R	
<i>Chiasocyclus literarius</i>	Het – Mst		R		R	
<i>Cribrosphaerella ehrenbergi</i>	Alb – Mst	R	F	F		
<i>Eiffelithus eximius</i>	LTur – Cmp		R		R	
<i>Eiffelithus turriseiffelii</i>	LAlb – Mst		C	F	C	
<i>Eprolithus floris</i>	Alb – ESnt	R	F	R	R	F
<i>Gartnerago obliquum</i>	Cen – Cmp		R			
<i>Helicolithus trabeculatus</i>	Alb – Mst		F	R		
<i>Lithraphidites carniolensis</i>	Het – Mst	C	C	C	C	
<i>Lucianorhabdus quadrifidus</i>	Tur – Cmp		R			
<i>Manivitella perrinitoides</i>	Het – Mst		R	R		
<i>Microrhabdulus decoratus</i>	LCen – Mst		R			
<i>Nannoconus multicaudus</i>	Tur – Cmp		R	C	F	
<i>Nannoconus</i> spp.	LJur – Mst	F	C	C	C	
<i>Prediscosphaera cretacea</i>	Cen – Mst		F	C		
<i>Quadrum gartneri</i>	Tur – Snt		R	R		
<i>Retecapsa angustiforata</i>	Het – Mst	R	F	R	F	R
<i>Stradneria crenulata</i>	Het – Mst	R	C	R	F	R
<i>Tranolithus phacelosus</i>	Alb – Cmp	R	C	F	F	
<i>Watznaueria barnesae</i>	Jur – Mst	C	A	C	C	F
<i>Zeugrhabdulus embergeri</i>	LJur – Mst		R	R		
<i>Zygodiscus</i> spp.	Jur – Mst	R	F	R	F	

Abundance symbols (for overall assemblage and individual species): A, abundant, <10 specimens per field of view; C, common, 1–10 specimens per field of view; F, few, 1 specimen per 1–10 fields of view; R, rare, 1 specimen per 10–100 fields of view. *Preservation*: 01–03 mild-severe overgrowth. *Species names*: The nomenclature of Perch-Nielsen (1985) is used, but with somewhat broader species concepts. *Ranges*: Reliable ranges of species within taxonomic concepts adopted. Biostratigraphically useful ranges emboldened: Trt, Tertiary; Mst, Maastrichtian; Cmp, Campanian; Snt, Santonian; Con, Coniacian; Tur, Turonian; Cen, Cenomanian; Alb, Albian; Apt, Aptian; Bar, Barremian; Hau, Hauterivian; Val, Valanginian; Ber, Berriasian; Jur, Jurassic. *Samples*: 12, clast from 49/09/12; 21a, b two separate clasts from 49/09/21; 137a, b two separate clasts from 49/09/137.

Appendix 3: Calcareous nanofossils from erratics of Cretaceous chalk, Celtic Sea.

Sample	Molluscs present
49/09/90	Unidentifiable fragments
48/T0/93	<i>Nuculana pernula</i> <i>Nuculana minuta</i> <i>Arctica islandica</i> <i>Abra prismatica</i> <i>Macoma</i> cf. <i>calcareo</i> <i>Astarte montagui</i> <i>Chlamys</i> cf. <i>islandica</i> <i>Cylichna</i> sp. <i>Siphonodentalium</i> sp.
48/09/3	<i>Nuculana</i> cf. <i>pernula</i> <i>Yoldiella fraterna</i>
48/09/148	<i>Nuculana</i> cf. <i>pernula</i>
49/09/21	<i>Hiatella</i> sp.
49/09/137	Unidentifiable fragments
40/T0/93	Unidentifiable fragments
49/09/43	Unidentifiable fragments
48/09/137	<i>Spisula</i> sp. <i>Nuculana</i> sp. <i>Astarte</i> sp. <i>Chlamys</i> sp. <i>Venus ovata</i>
48/T0/53	<i>Astarte sulcata</i> <i>Chlamys</i> et. <i>islandica</i>
48/09/97	Unidentifiable fragments
48/T0/51	<i>Colus gracilis</i> <i>Nucula nucleus</i> <i>Anomia ephippium</i> <i>Anomia squamula</i> <i>Chlamys distorta</i> <i>Modiolus modiolus</i> <i>Nassarius incrassatus</i> <i>Abra</i> sp. <i>Macoma</i> sp. <i>Anomia</i> sp.
47/07/336	<i>Emarginula reticulata</i> <i>Alvania cancellata</i> <i>Gibbula tumida</i> <i>Timoclea ovata</i> <i>Venus fasciata</i> <i>Chlamys opercularis</i> <i>Venus ovata</i> <i>Glycymeris</i> sp. <i>Anomia</i> sp. <i>Zitzyphius</i> sp. <i>Philbertia</i> sp.

Appendix 4: Mollusca from the Celtic Sea samples.

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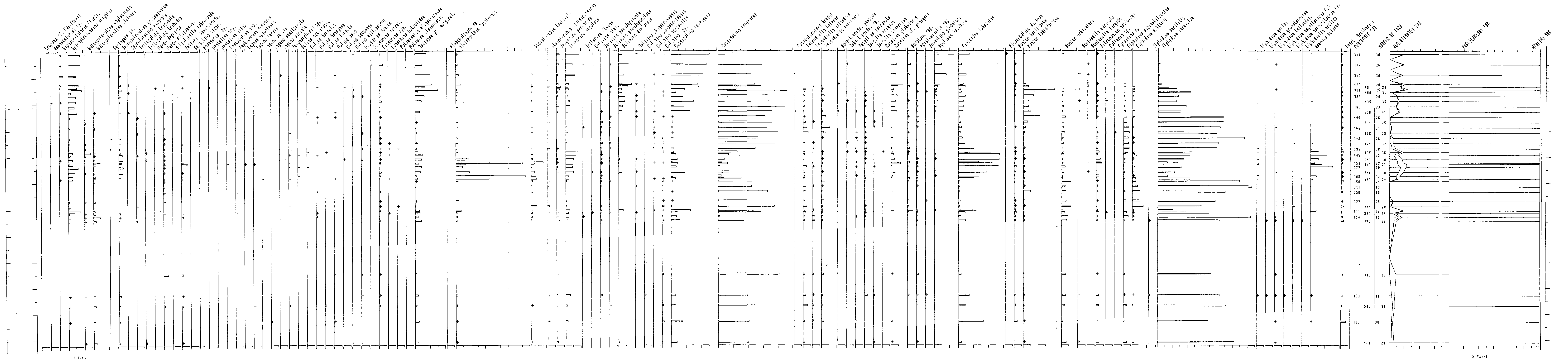
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E2



7. Int.

E.3.



E.4

