ST ANDREWS BAY: A SEDIMENTOLOGICAL, GEOPHYSICAL AND MORPHOLOGICAL INVESTIGATION

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A Thesis Submitted for the Degree of PhD
at the
University of St Andrews

1996

St Andrews Bay:
A sedimentological, geophysical and morphological investigation.

by

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A thesis submitted in accordance with the requirements of the University of St Andrews for the award of Doctor of Philosophy.

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To my parents, wife,

sons and daughters

DECLARATION FOR THE DEGREE OF Ph. D.

I **Hamad Abdullah Al - Washmi** hereby certify that this thesis has been composed by me and it has not been accepted in substance for any other degree and is not currently being submitted in candidature for any other degree.

Signed (candidate)

Dated ..2..10....1995..............

I hereby certify that **Hamad Abdullah Al - Washmi** has fulfilled the conditions of the Resolutions and Regulations appropriate to the degree of Ph. D.

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....

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ABSTRACT

This thesis examines past and present day processes responsible for the morphological development of St Andrews Bay in eastern Scotland.

Quaternary glacial events have contributed large volumes of sediment from the Scottish Mainland to the North Sea Basin over the last 3 million years. Since the most recent glacial event, the Late Devensian , which terminated some 14,000 years ago in Scotland, the sediments of the coastal areas have been redistributed by wave and tidal activity. Thus the bathymetry and platform of St Andrews Bay has evolved since that time, although some elements of the morphology appear to predate the last glaciation.

The grain size distributions of the bed sediments of the bay show a narrow range of mean sizes between fine and very fine sand. These are indicative of a low energy tidal environment although Quartile Deviation - Median Diameter plots suggest the importance of wave activity in determining their distribution. Current measurements in the bay confirm that the hydrodynamic environment is of low energy with average current speeds rarely exceeding 15 cm s^{-1} one metre above the bed. Progressive vector plots show closed ellipses during calm weather but meteorological forcing and wave activity generate residual currents predominantly in the direction of wave propagation or down wind. Application of a transfer function to the current data predicts low rates of bedload transport, the residual of which generally accords with the recent pattern of sedimentation at the head of the bay.

The rocky platforms of the southern margins of the bay cannot easily be sub-divided into features at different elevations. No firm evidence is presented for a pre- Late Devensian origin of the platform but it is argued that such a chronology explains the morphology of the platform.

Offshore sedimentary sequences, up to 30 metres in thickness, are reported from geophysical surveys which have been laid down since the last glaciation. The units identified reflect changing environments of deposition associated with climatic and sea level changes over the last 14,000 years.

ACKNOWLEDGEMENTS

I would like to express special thanks and gratitude to my supervisors (Dr. J Jarvis and Dr R. Duck) and to the Head of School (Professor J. McManus) for their guidance, constructive criticism, useful suggestion, enduring patience and continual inspiration, through out my study period. I am very grateful and would like to express a special thank you to Dr. Jack Jarvis for his stimulating supervision, encouragement, his support in the collection of data from St Andrews Bay and his expert advise in the interpretation and processing of this data.

I am grateful to all members of the technical staff for their assistance. I am especially grateful to Mr Colin Cameron for his help and assistance with my computer problems. I would also like to thank Mr. Jim Allan for the use of his photographic skills. It is also a pleasure to acknowledge the help of Mrs. Florence McAndie and Mr. Stewart Harvey for their administrative assistance.

Thanks to all my colleagues for their assistance and not forgetting Justin who helped me in the early stages of my study.

Finally I must thank my parents for their continuous encouragement and support, care and help throughout these years. I would also like to thank my brothers and sisters for their prayers on the successful completion of this thesis. A very special thank you to my wife for her patience, tolerance, support, sympathy and understanding. A special thanks to my children, Reem, Abdullah, Sarah and Ibraheem (A Dundonian boy) for their understanding and personnel sacrifice while I was completing this thesis.

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CHAPTER I

Introduction

During the Pleistocene glaciations ice-induced movement, of the land relative to the sea (isostasy) and of the sea relative to the land (eustasy), resulted in variations in the position at which the processes of coastal erosion and deposition took place. The study area, St Andrews Bay, which is situated in the Midland Valley on the east coast of Scotland (Fig. 1.1) was overridden by ice during the Pleistocene glaciations. The transgression and regression events caused by the subsequent eustatic and isostatic changes of sea level made the sedimentation processes both active and complex in this region.

There is abundant evidence of Quaternary sea level change along the coastline of St Andrews Bay, exemplified by the raised beaches in the area. The area has also been affected by local tectonic effects when land and sea areas have subsided or risen isostatically in response to the growth and decay of ice sheets. This occurs because of the nature of the Earth's crust, which flexes in response to loading stress; once the ice retreats the unburdened crust begins to rebound. In general, the isostatic changes take a longer time than the eustatic ones, as the Earth's crust reacts quite slowly to changes in loading. These changes produce a relatively complex series of coastal sea level profiles in different areas, depending on the proximity of the area to an ice sheet (Viles and Spencer, 1995).

The deposits formed by ice result essentially from mechanical action or physical weathering and are made up of mixtures of rock fragments and minerals, ranging from giant boulders to extremely fine rock dust or so called rock flour, all abraded from the substrate (Chamley, 1990). During the retreat of the glaciers, materials such as sand and gravel which were left behind in the coastal zone area were reworked into raised beaches (Sissons et al., 1966). Glacially - derived sands and gravels were also available offshore in the bay and it is

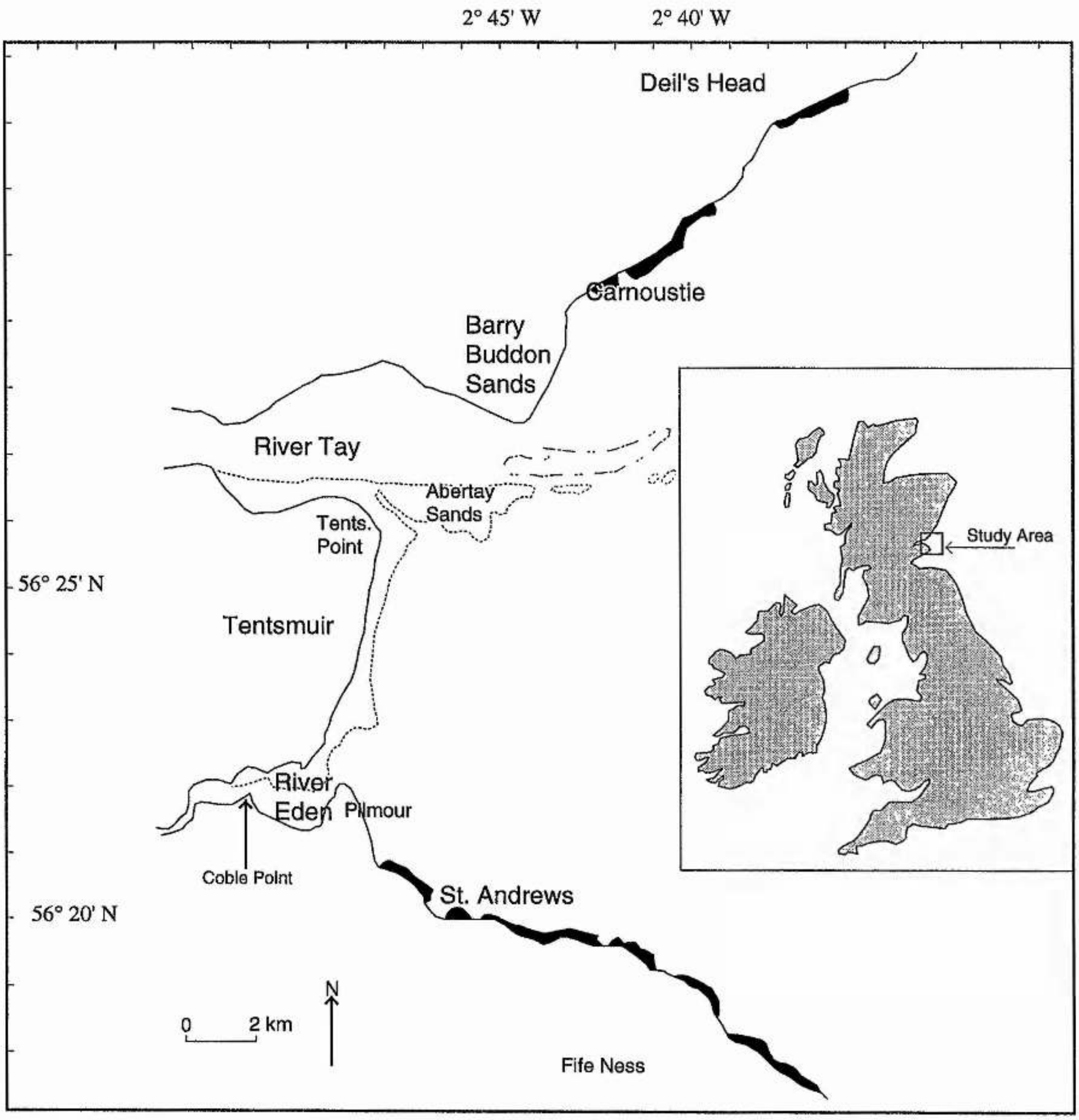


Fig.1.1 Location of study area within the UK.

likely that these deposits represent the primary source of sediment in the study area, particularly during the early stage of deglaciation. These have been investigated in the present study by geophysical survey methods.

Usually shallow-water environments, such as St Andrews Bay, are in a state of disequilibrium because these environments have been affected strongly by sea level changes throughout geological history. Sea level controls the type and magnitude of all coastal processes such as tidal range, breaker type, longshore current velocities and sedimentation rates (Pethick, 1984). Thus, the sedimentary history of the shelves and nearshore environments reflects the ever changing relationship between tectonic subsidence, sea level oscillations and the dynamic processes acting at these water depths. This may be illustrated by the fact that the formation of some remnant sand spits in the area, such as Coble Spit, may have occurred when sea level was higher than today (Ferentinos and McManus, 1981).

At the height of the last glaciation, 18,000 years ago, sea level was as much as 140 m below present sea level and large areas of the continental shelves of the world were dry land. Since then, relative sea level has been as much as 30m above and 10m below its present level at St Andrews but only since the main Postglacial transgression, some 7,000 years ago, have the waves and currents sculpted the bay into its present outline (Wal, 1992).

The length of time that sea level has been at or around its present level affects the functioning of the coastal system. Some coastlines have experienced coastal processes at the present level for around 6,000 years (e.g. Australia), whereas in Florida present sea level was reached only in the last 1,000 years. Some coastlines are still subject to isostatic uplift and sea level fall, while others are undergoing isostatic adjustments that lead to sea level rise. These continuing movements are important within the context of future sea level rise. Isostatic uplift will offset some of the expected eustatic component whereas isostatic down warping will add to the ocean warming sea level effect (Bird, 1993). Although sea level has been close to its present elevation for several thousand years in Scotland, St Andrews Bay is still evolving and has not yet reached a state of equilibrium. The present day morphological development is controlled by the meteorological and hydrodynamic forces which act upon the sediments of the bay and adjacent areas which forms the major focus of this study into the morphological development of St Andrews Bay.

Rocky coastlines provide a contrasting environment and have different responses to the coastal processes that affect sandy beaches and the sea bed. Accordingly their study requires an examination of different time scales and processes of change. Along the southern margin of St Andrews Bay , between St Andrews and Fife Ness, (Fig 1.1) there are rocky platforms backed by cliffs. These have been produced by the retreat of the cliffs and result from a range of processes that erode cliffs and plane the surface of the platform. The processes of formation : chemical weathering, wetting and drying, salt weathering, frost and ice on weathering, biological weathering and erosion, wave erosion and mass movements have given rise to several theories of platform development (Trenhaile, 1987; Sunamura, 1992). The theories vary according to which processes dominate and at what level the erosion is focused (e.g. water level, saturation level). This study, however, is concerned primarily with the morphological development of the platform in St Andrews Bay and has only indirectly addressed the question of the processes of platform development as outlined above.

1.1 Previous work

The present study builds upon previous work carried out in the area. The other work is placed within the context of the present study in the appropriate sections of the thesis. However, a broad outline of earlier work in the study area is reported here. Several research programmes have been carried out in the area between Boddin Point on the northern side of St Andrews Bay and the Fife Ness in the south. The physical characteristics of the coastal zones of Tayside and North East Fife have been recently described by Ritchie (1979) and Wright (1981). In the lower reaches of the Tay Estuary, Green (1974), McManus et al. (1980) and Al- Mansi, (1986) have examined various features of the sediment transport patterns produced by the indigenous wave and tidal activity. Farther to the south, sediment motion in the mouth of Eden Estuary has been studied by Jarvis and Riley (1987). Beyond the immediate realms of the estuary, a more extensive study has been undertaken by Sarrikostis (1986) who, utilising a computer wave refraction model, has predicted the potential longshore sediment transport patterns between Montrose and Fife Ness. Eastwood (1977) studied the major topographic features of the Eden Estuary, their associated bedforms and internal structures. More recently, Boalch (1988) assessed the potential for beach nourishment projects to provide a new and more effective solution to coastal erosion problems than those currently in use. Most recently the sedimentological effects of aeolian processes in the Tentsmuir area formed the focus of a study by Wal (1992). This investigator described the shoreline changes which have been observed in the study area during the last 150 years in general and the last 10 years in particular. Wal found that longshore marine sediment transport and further redistribution of the beach sediments by wind are the principle mechanisms of foredune accretion at Tentsmuir Point. Wal also related the development of a positive beach sediment budget at Tentsmuir Beach to the offshore component of the wind regime, whereas the onshore winds potentially can carry the sand landwards resulting in the development of the foredune behind the beach (Wal and McManus, 1993).

1.2 The aims of this study

This research study was undertaken with the aim of achieving a better understanding of recent processes of sedimentation in St Andrews Bay and, in particular, its evolution since glacial times.

1.3 Structure of Thesis

Following the introduction, Chapter 2 examines the geological and geomorphological setting of the area including a review of the glacial history of the area. The next four chapters form the substantive part of the thesis and report on the research investigations that were carried out in this study. These are followed by a summary and conclusion in Chapter 7.

The sediment size distributions of the bed sediments in the bay are described and analysed in Chapter 3, in order to examine regional variations and to make inferences concerning the hydrodynamic forces that have been responsible for their present distribution. This latter aspect has been investigated using plots of quartile-deviation versus median-diameter on double-logarithmic graph paper, as applied by Buller and McManus (1972 ; 1973 a, b).

Chapter 4 examines the response of the bay to present - day processes of sediment transport due to wave action and tidal currents. self recording current meter data, combined with wind data from the Leuchars meteorological station have been used to describe the pattern of tidal currents in the bay and the impact of meteorological forcing on water movement. The same data have been used to examine residual water movement and to compute potential sediment transport using the methods of Gadd et al. (1978) and Vincent et al. (1981).

In Chapter 5, the morphology of the rocky platforms which are found between St Andrews and Fife Ness is described and the problem of their age of formation addressed. The nature of the platform is examined in relation to present and past processes and the chapter discusses whether there is a single feature or several platforms present at different elevations. In conjunction with echo sounding technique, a Waverley Sonar 3000 system has been used to carry out a side-scan sonar survey of the rocky platforms. The submarine geological structure has been mapped and two previously - unreported volcanic vents located. The intertidal rock platform was also surveyed to investigate its morphology.

Chapter 6 investigates the stratigraphy of the sedimentary sequences in the bay by sub-bottom profiling in order to comment on the pattern of sedimentation since the last glaciation. The seismic reflection records show that the Quaternary sedimentary sequences can be subdivided simply into two distinct units separated by an erosion surface.

Chapter 7 provides the discussion, conclusions and suggestions for future research in St Andrews Bay.

CHAPTER II

The Study Area

2.1 Introduction

St Andrews Bay is a large embayment on the east coast of Scotland which extends from Red Head in the north to Fife Ness in the south (Fig. 2.1). The coastal zone forms an almost continuous curve in plan which is broken by the estuaries of the Tay and Eden which enter at the head of the bay. The River Tay is Scotland's longest river (188 km) draining a catchment of 4,590 km^2 with a mean annual discharge of 152 m^3 s^{-1}; it enters St Andrews Bay through the 35 km long Tay Estuary between Buddon Ness and Tentsmuir. The River Eden, with a mean annual discharge of 4 m^3 s^{-1}, drains a lowland catchment of 307 km^2 and enters the bay via the Eden Estuary, 4 km to the north of St Andrews. Offshore a free exchange of water with the North Sea takes place across a 27 km long zone between Boddin Point and Fife Ness.

The intertidal and supratidal zone varies around the bay. High cliffs with a rocky intertidal platform north of Arbroath contrast with the wide sandy beaches backed by extensive foredunes and windblown sand at the head of the bay. Along the southern shore of St Andrews Bay the shoreline takes the form of a series of discontinuous beaches cut in drift and occasionally solid rock. In some places they are well developed with a clear backing cliffline, elsewhere they slope gently seaward with a poorly defined cliffline. The bathymetry of the offshore zone is characterised by a relatively smooth bed topography, except at the estuary mouths, and depth gradually increases from the shoreline to about 35 m at the offshore boundary to the bay.

2.2 Geological setting

The coastal areas of Tayside and Fife occupy the north-eastern part of the Midland Valley of Scotland, which is situated between the Highland Boundary Fault to the north and

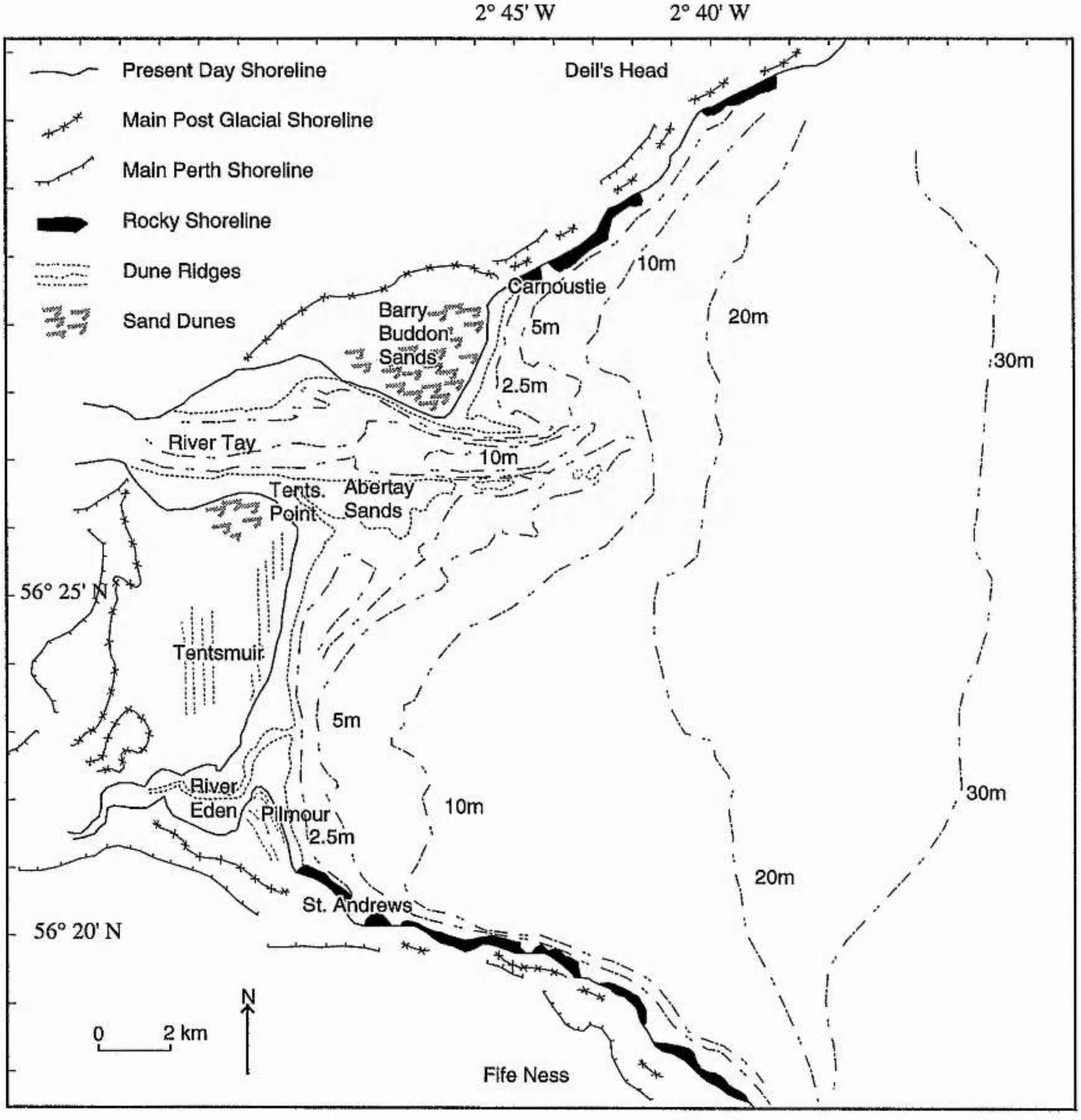


Fig. 2.1 Position of late glacial (main Perth) and post-glacial (Flandrian) shorelines (after Rice, 1962; Cullingford, 1972; Chisholm, 1966, 1971), and present-day features at the head of St Andrews Bay.

Southern Upland Fault to the south (Geikie, 1865). The region broadly has the structure of an ancient rift valley or graben in which strata between two parallel faults have subsided relative to the horst blocks on either side (Cameron and Stephenson, 1985). Out with the Midland Valley, Lower Palaeozoic and older rocks have been strongly folded and indurated, but between the faults rocks of Upper Palaeozoic age outcrop which are less deformed.

The solid geology of the study area is dominated by Devonian sedimentary and igneous formations (Lower and Upper Old Red Sandstone) and Carboniferous sedimentary and igneous formations, upon which lie discontinuously, unconsolidated glacial and postglacial sediments. On land, rocks belonging to the Carboniferous system (Fig. 2.2) are confined to the area south-west of a line extending from Guardbridge to the south side of Loch Leven, (Geikie, 1902; MacGregor, 1968; Armstrong and Patterson, 1970; Forysth and Chisholm, 1977). Offshore, the boundary between these units is not defined but must lie fairly centrally through St Andrews Bay. The Old Red Sandstone gives rise to an impressive cliffline east of Arbroath on the northern margins of the bay whereas the Carboniferous formations in general have resulted in a much more dissected cliffline between St Andrews and Fife Ness.

The northern coast of St Andrews Bay from Carnoustie to Red Head (Fig. 2.1) is composed primarily of strata ascribed to the Garvock Group of the Lower Devonian which outcrop on the south-east limb of the Sidlaw-Ochil Anticline where they comprise the Red Head Series, Auchmithie Conglomerate and Arbroath Sandstone (Armstrong, Patterson and Browne, 1985). These rocks intermittently outcrop on the shore between Red Head and Tentsmuir and reach a thickness of over 1000 m. They mainly consist of sandstones with nodular limestones and interstratified volcanic rocks. The sandstone is red in the north-east around Arbroath and becomes grey in the south-west; it is interbedded with thick conglomerates, which contains clasts of Highland and volcanic rock types.

There are no exposures of solid rock between Carnoustie and St Andrews and the head of the bay is formed in glacial and postglacial sediments (Fig. 2.3). The coastline from St Andrews to Fife Ness is formed primarily in Carboniferous strata with possible Old Red

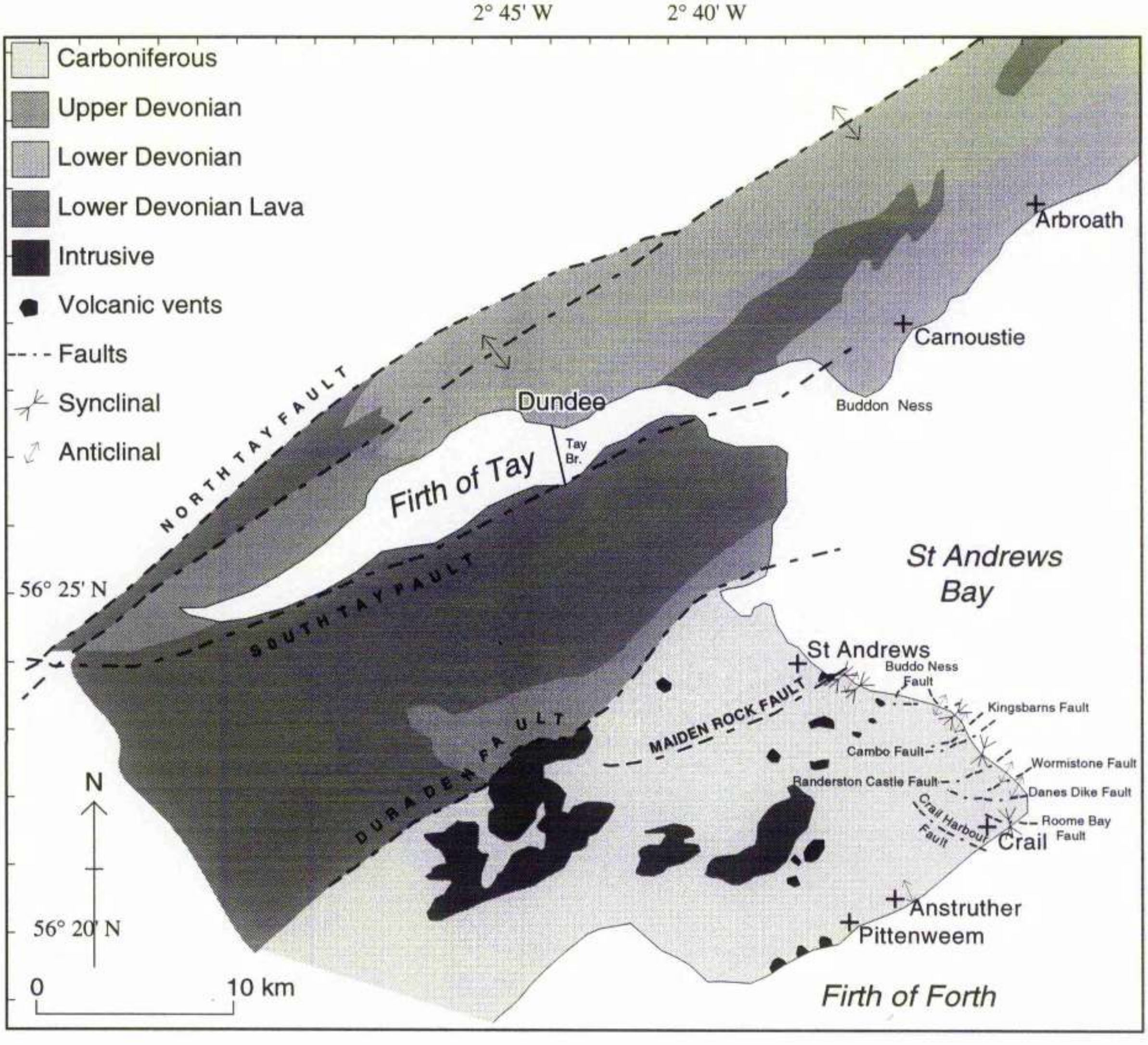


Fig 2.2 Solid Geology and Structural features of the study area in Tayside and N.E. Fife compiled from The Geology of E. Fife (1977), The Midland Valley of Scotland (1985) and Geology of the Perth and Dundee district (1985).

Sandstone inliers at Balcomie and Cambo. The Carboniferous sediments are dominantly sandstones with mudstones, subordinate coals, limestones, siltstones, and fossil soils. The rocks are arranged in rhythmic sequences or cyclothems individually up to 30 m in thickness (Forysth and Chisholm, 1977). The relative proportions of each rock type in a cyclothem varies from place to place and within the succession. It is suggested that the cyclothems were formed by the advance of deltas into lakes or lagoons, combined with fluctuating river distributary positions (Forysth and Chisholm, 1977). At times, local subsidence or variations of sea level led to marine incursions and the formation of marine limestones. The fauna within these suggest that, during transgressions, the depth of water was shallow. Nevertheless, long-term subsidence during the Carboniferous allowed up to 3,500 m of sediment to accumulate in parts of east Fife. Early beds in the Carboniferous sandstone sequences are distinguishable from the late beds by the more persistent and varied marine bands present compared to the non-marine faunal bands and seat earths present in the early beds. The early beds also show more lateral variation than the later deposits (Forysth and Chisholm, 1977).

The Devonian and Carboniferous rocks in the area have been affected by earth movements and volcanism. Mid-Devonian folding of the Lower Old Red Sandstone sequences in the north of the area were responsible for the creation of the two parallel fold structural units the axes of which trend north-east to south-west and form the Sidlaw-Ochil anticline and the Strathmore syncline. Cross-faults of the same age cut these structures and intersect the coastal rocks to the north-east of Carnoustie. (Armstrong and Patterson, 1970). Major earth movements from early Dinantian times have resulted in the subsidence of the Tay Graben between the north-east trending North Tay and South Tay Faults (Fig. 2.2). Between the faults, the extensive lowland is underlain by downfaulted Upper Devonian and Carboniferous rocks which are largely buried by thick sequences of Quaternary deposits.

South of the Tay one major fault which affects the area is the north-east trending Maiden Rock Fault which cuts the coastal sequences (GR. 528157) about 2 km south-east of St Andrews. From the Maiden Rock to Fife Ness at least nine other minor wrench and

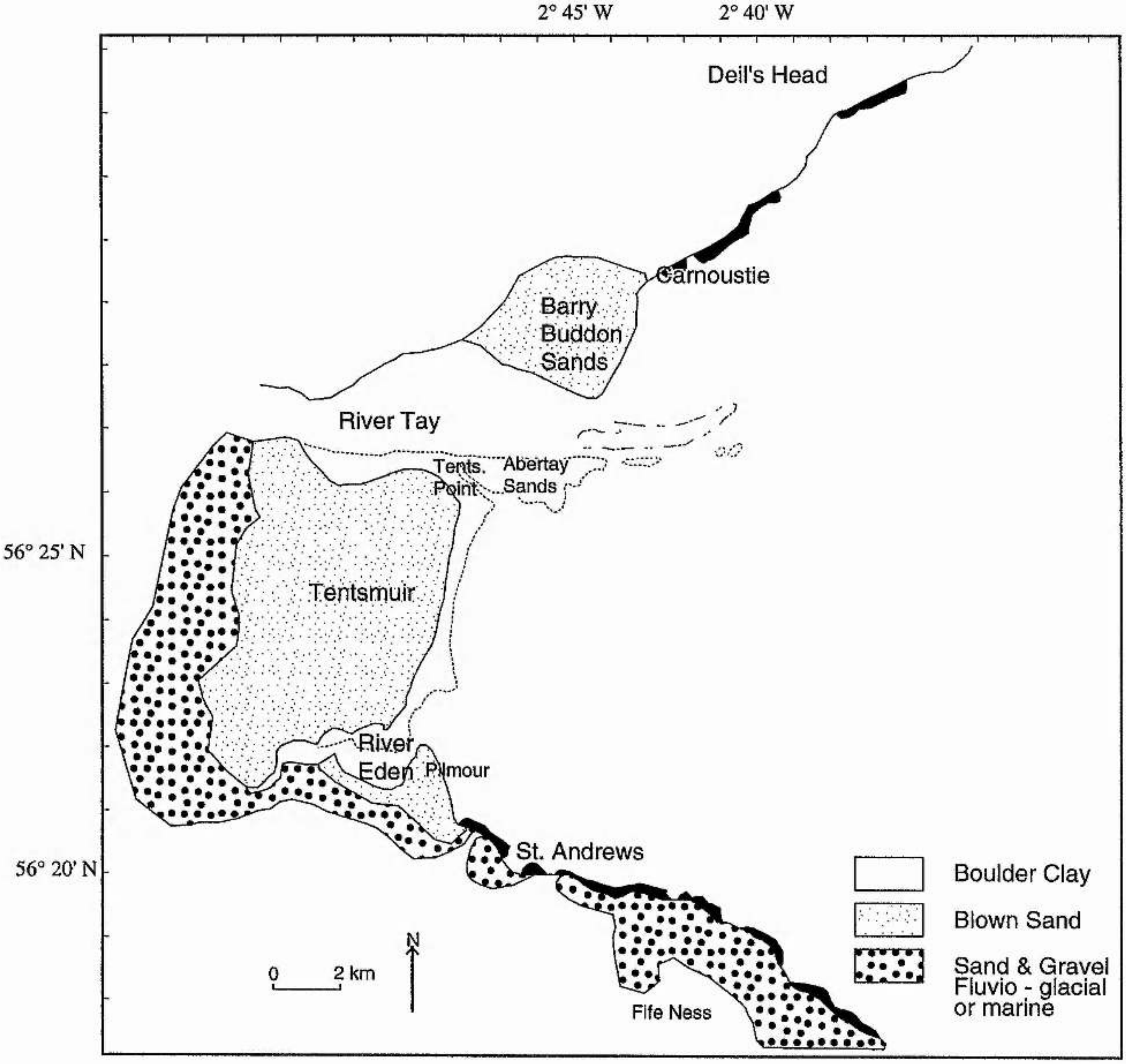


Fig. 2.3 Drift Deposits around the shore line of St Andrews Bay.

oblique faults have been recorded in the coastal sections. Near to the Maiden Rock Fault, in particular, the strata have been thrown into complex tight folds (Plate 1), with decreasing intensity away from the fault. However, for the most part, the Carboniferous strata along the coastal section are quite gently folded (Forysth and Chisholm, 1977).

Many volcanic necks cut the Carboniferous strata in Fife and three well-defined necks occur within the coastal zone at Kinkell Ness (GR. 538158), Buddo Ness (GR. 559154) and at Salt Lake (GR. 585143). The former is usually known as the 'Rock and Spindle' (Plate 2). A further complex neck is also reported just west of Kittocks Den (GR. 554101). Some igneous material has also been intruded into the coastal sequences; its age relationship to the volcanic necks is ,however, not well defined. For example, at Kinkell Ness a later basaltic dyke cuts the neck material at the Rock and Spindle before trending out into country rock and expanding into a number of small masses (Forysth and Chisholm, 1977). Intrusive material is also noted in association with the Buddo Ness and the Salt Lake necks.

The geological history of Fife and Angus in Mesozoic and Tertiary times is unknown, but the principal present-day physical features may have existed by the end of the Tertiary. In particular, the main elements of the present drainage system were established from a north -south divide to the west of the Highlands. This probably reflects earlier drainage networks which were, in part, determined by structural elements such as the Tay Graben and the Strathmore syncline (Hall, 1991). The proto-Tay is thus thought to have formerly drained to the sea via Strathmore before it was diverted or captured by the west-east trending River Earn during the Quaternary.

2.3 Quaternary history

The radical changes of climate characteristic of the Pleistocene affected the entire study area and are responsible for much of its present day morphology. In the Highlands, successive glaciations have carved out the glacial troughs which dissect the mountains and modified former drainage networks (Sissons, 1976a). The products of glacial erosion were transported towards the ice margins and have accumulated as thick sequences of glacigenic deposits on the continental shelf and in the North Sea. In the study area, little is known of the effect of these early glaciations as no deposits within the area can be traced unambiguously to any other than the most recent glaciation, namely the Late Devensian.

During the Late Devensian an ice sheet, which reached its maximum extent at circa 18,000 years BP., covered the whole of Scotland (Sissons, 1974a ; Boulton et al., 1991). The ice sheet spread into the Midland Valley from centres in the western part of the Grampian Highlands and to a lesser extent from the Southern Uplands and was channelled into St Andrews Bay through the Tay and Eden valleys. The Midland Valley was covered by ice approximately 1 km in thickness at the maximum extent and St Andrews Bay would thus also have been covered by a thick ice sheet. (Boulton et al., 1991). The ice sheet spread out eastwards from the coast and the Wee Bankie Moraine, some 60 km east of the Tay and Forth estuaries, is believed to represent the terminal moraine laid down by this ice sheet (Thomson and Eden, 1977; Sissons, 1983). The suggested direction of ice movement is shown in Fig. 2.4 which thus defines the potential sources for the deposits of St Andrews Bay.

In general, glacial erosion processes were most intense in the highland source areas. It is probable that the study area was less affected by glacial erosion because of its lowland location. Pre-existing superficial deposits were almost certainly removed and reworked from the bay and much of the surrounding area during the glaciation, exposing the underlying bedrock. During this extensive phase of advance and retreat of the ice sheet across St Andrews Bay some till deposits were laid down beneath and at the margins of the ice sheet. Sparker surveys (sub-bottom profiling) in St Andrews Bay show isolated ridges and pockets of till ascribed to this period (Browne and Jarvis, 1983). During the early stage of deglaciation, which occurred between 18,000-15,000 years BP., westward retreat of the ice sheet margins took place. Upon deglaciation the study area had not recovered fully (isostatically) from the weight of ice so that, with the land still depressed, sea level was higher than at present allowing the Lateglacial sea to extend farther westwards than at present. Support for these relationships is indicated by elevated marine limits and by the presence of marine sediments in the Howe of Fife at elevations of up to 43m above OD

(Knox, 1962 ; Armstrong et al., 1985; Duck, 1990).The sea reached far inland along the valleys of the Rivers Tay, Earn, Forth and South Esk and deposits of glaciomarine silty clays containing a highly distinctive cold water arctic Fauna (The Errol Beds) were laid down at this time in the valleys (McManus, 1966; Paterson et al., 1981). Glaciomarine sediments of marine clays and sands (The St Abbs Formation) were probably also being laid down at this stage in St Andrews Bay, because the area would have received those sediments from the retreating ice margins in the Tay and Forth valleys.

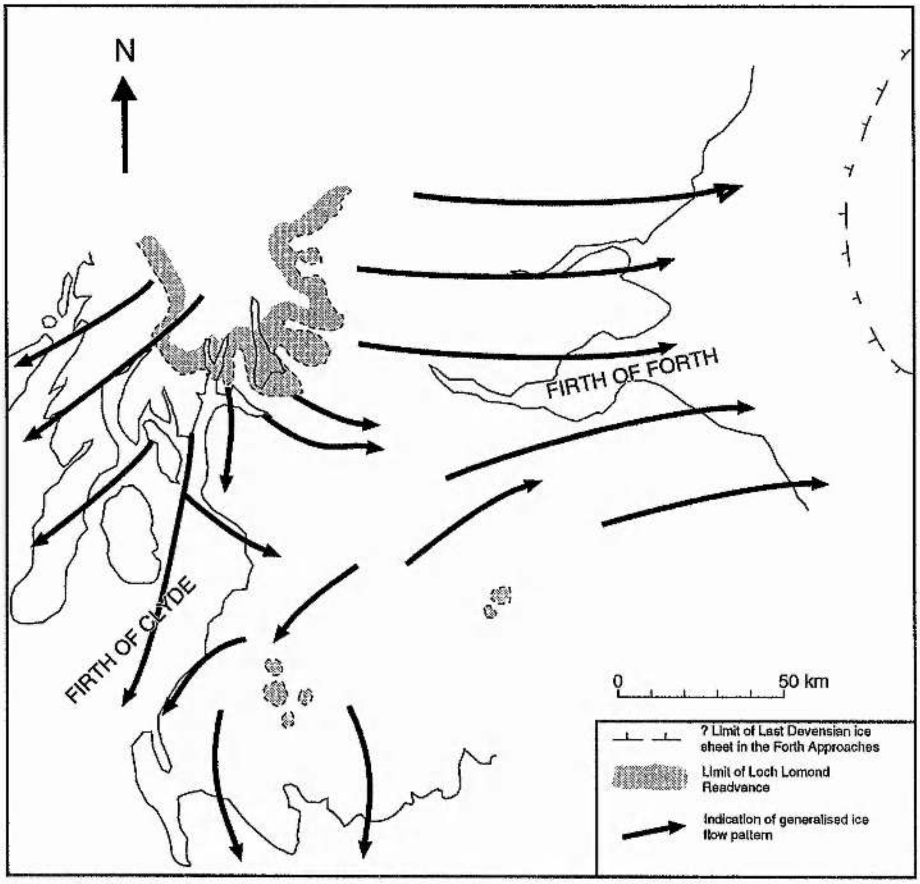


Fig. 2.4 Generalised pattern of ice-flow and ice limits (after Cameron and Stephenson, 1985).

During deglaciation, major changes were also taking place along the shores of St Andrews Bay as well as offshore. The position of the shoreline varied in relation to the interaction of two main causes. Eustatic lowering or raising of sea level occurred in response to the formation or wasting of vast ice-sheets, and isostatic uplift occurred when the glaciated

Stage		Thousand years before present	Climate condition	Variable in sea level relative to land	Tectonic Setting I = isostatic E = eustatic	Deposits
Flandrian	P r e s e n t			Steady	I=E	Blown sand, sand dunes (eg. Tentsmuir beach
	Later Flandrian	6.5	Milder climate	rise	E>I	The main post glacial shore line
	Early Flandrian	9-8	as warm as	fall	I>E	The formation of lower main and lower buried beaches grey micacaceous silty fine sand or sany silt
Loch Lomond Stadial	Later part of stadial	10-9	present day	rise	E>I	Formation of high buried shore line
	First part of stadial	11-10	Arctic	fall	I>E	Marine planation (gravel layer) also channels cut through late Devension marine sediment and the formation main late glacial shore lines
Late glacial inter stadial	Later part of the deglaciation	12	Milder climate but a little cooler than present day	rise	E>I	Deposits of silt and clay (Powgavie clay)
	After deglaciation	13.5	Arctic	fall	I>E	Formation of the late glacial shoreline of East Fife
	Early stage of deglaciation in St Andrews Bay	15	Arctic	rise	E>I	Errol Bed in the land and St Abbs formation in the sea
Late Devensian		18	Freezing sea water	overall regression	Land strongly depressed due to weight of ice sheet	Till

Table 2.1 Enviromental changes in St Andrews sincethe Late Devensian ice sheet (modified after Armstrong et. al. 1985).

area was relieved of the mass of ice by melting (Table 2.1). Initially, on deglaciation, a high relative sea level might be expected to exist because the depressed land surface would not have had sufficient time to recover from its ice burden . However, as time proceeded through the Lateglacial interstadial, sea level might have been expected to fall , because the rate of isostatic uplift was greater than eustatic rise of sea level (Sissons, 1974a). A number of shoreline features in east Fife have been interpreted as Late Devensian shorelines, formed in association with a westerly retreating ice margin (Cullingford and Smith, 1980 ; Sissons, 1983). Their position above present sea level is due to the glacio-isostatic uplift since formation. Along the southern shore of St Andrews Bay, between St Andrews and Fife Ness, these shorelines take the form of a series of discontinuous benches cut into drift and occasionally into solid rock. In some places, they are well developed with a clear backing cliffline e.g. at East Sands near the caravan site, elsewhere they slope gently seaward with a poorly defined cliffline, e.g. at Cambo Ness and Buddo Ness.

The highest Late Devensian raised shoreline at St Andrews is 28 m above present sea level; this represents the vertical displacement caused by uplift processes. Above this limit the land surface is characterised by mounds of glacial deposits and fluvioglacial sands and gravels. The raised beaches below this limit, the marine limit, have been traced around St Andrews Bay and are represented in the shoreline diagram (Fig. 2.5). Accordingly, it is suggested that during the period of time represented by this diagram, water depths in the bay were up to 30m greater than today. The shoreline features show a general eastward tilt which is a consequence of the differential nature of glacio-isostatic recovery (Armstrong et al., 1985). The amount of uplift increases towards the former centre of the ice cap in the western Highlands.

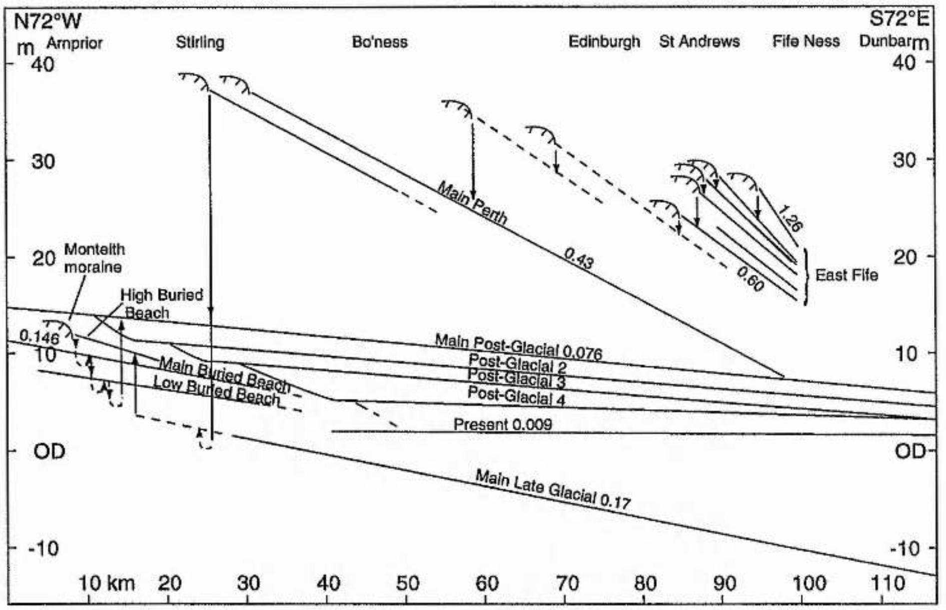


Fig. 2.5 Sequence of shorelines in SE Scotland with gradients in m / km (from Sissons, 1974).

The period of deglaciation was followed by the supposed Perth readvance, probably between 13,000 and 13,500 years BP (Sissons, 1967). It is thought that the readvance was associated with the formation of the most pronounced Lateglacial raised beach in south-east Scotland. It has a gradient of 0.43 m / km and it is conspicuous declining eastwards (E 17° S) from 28 to 30m at Perth , to about 20m at Dundee , 10m at Fife Ness and 12m at Montrose (Cullingford and Smith, 1980).

The Lateglacial interstadial came to an end when glaciers formed again in the Scottish Highlands. This episode, known as the Loch Lomond Readvance, occurred between about 11,000 and 10,300 years BP (Sissons, 1979; 1983). Sea level fell in the study area during this time as glacio-isostatic uplift continued to exceed sea level rise, because the area was not subject to depression caused by the growth of Loch Lomond Stadial glaciers. During this period of sea level fall, shorelines migrated north and eastwards and a major shoreline was formed at the minimum extent of the sea known as the Main Lateglacial Shoreline. This forms an important reference level in the sediments of St Andrews Bay and can be identified over a wide area (Browne and Jarvis, 1983).

The age and mode of formation of the Main Lateglacial Shoreline has been subject to some debate. Many workers relate its age to the Loch Lomond Stadial (McManus, 1972 ; Sissons, 1974b ; Paterson, 1981). Sissons (1969) identified an extensive buried gravel layer correlated with the Main Lateglacial shoreline at Grangemouth, 4 miles NE of Falkirk, which he interpreted as having been produced during the Loch Lomond Stadial. He suggested that the gravel was mainly derived from the abundant dropstones which are found in the underlying Late Devensian marine sediments across which the surface was eroded. The shoreline is not only confined to the nearshore area, but is also found offshore as a submerged feature.

Rhind (1972) correlated a till-free buried channel of the River Tweed, the base of which is at c. -20m OD at Berwick, with the low relative sea level associated with the Main Lateglacial Shoreline. Paterson (1981) suggested that the Main Lateglacial Shoreline occurs at -8m OD, some 13 km east of Dundee. Confirmation of the presence of a subaerial surface in St Andrews Bay is provided by the presence of buried channels in the area which were formed by the Rivers Tay and Eden cutting through the Late Devensian glaciomarine deposits (Buller and McManus, 1972). Sub-bottom profiling in the mouth of the Eden has enabled the mapping of the former channel of the Eden at an elevation of -5m OD and its form suggests that it was either a subaerial or intertidal channel (Browne and Jarvis, 1983). It is suggested that, because water depths were reduced, waves and currents acted as agents of erosion on the bottom sediments causing the removal of sand and clay from the glaciomarine deposits which had been deposited previously. This phase of lowering of sea level in St Andrews Bay is thus marked by the presence of a lag gravel horizon now buried beneath recent sediments.

The nature and rate of sea level change during the Loch Lomond Stadial is uncertain. Sissons (1981) and Sutherland (1985) have suggested that the rise in sea level at the first stage of the Loch Lomond Stadial was slow but subsequently it became more rapid at the later stages of this period. However, Paterson <u>et al</u> (1981) , Browne and Jarvis (1983) and

Browne (1985) have suggested that, during the first part of the stadial relative sea level fell, while during the later stages it rose rapidly.

At the end of Loch Lomond Stadial, the climate ameliorated and caused a rapid transgression as the Laurentide and Scandinavian ice sheets melted. Sea level rose from its late- glacial low level at about 10,000 years BP to about 9 m above its present level by 6,000 years BP (at St Andrews). At this time, the waters in St Andrews Bay would again have been deeper than at present. At least part of the transgression was probably very rapid, since the lag gravel horizon is fairly continuous and does not appear to be disturbed or reworked by a rising sea level. In the nearshore areas the transgression allowed the deposition of marine sands and muds upon the Lateglacial surfaces.

The transgression culminated between 6,900 and 6,000 years BP; and its limit being marked by the Main Postglacial Shoreline which attains an altitude of 7-9 m at St Andrews (Sissons, 1974b ; Sarrikostis, 1986). In Scotland the Main Postglacial Shoreline has two principal expressions; it is either represented by raised sand and gravel beaches (sometimes resting on a rock cut platform of Lateglacial or interglacial age) or by raised estuarine mud flats, known in Scotland as carse lands. The Main Postglacial Shoreline is the best developed and least fragmented raised shoreline visible in large parts of eastern Scotland (Cullingford et al., 1986). The deposits associated with this transgression, in the relatively sheltered firths of the Forth and Tay, constitute the carse clays that are now raised estuarine mud flats (Price 1983).

By 6,000 years BP eustatic rise in sea level was almost over but isostatic rebound continued and further postglacial raised shorelines were formed below the Main Postglacial Shoreline (Plate 3). At least one lower beach , at 4 m OD is known on the west side of St Andrews (Sarrikostis, 1986). The transgression ended about 5,500 BP and since that time the sea level has fallen gradually and intermittently (Morrison et al .,1981). It is most likely that since 3,000 years ago sea level has been more or less constant.

A tentative relative sea level curve for the study area is shown in Fig. 2.6. This interpretation is based upon dated curves for the postglacial period from the Tay and Forth Valleys and estimated dates for the East Fife shoreline EF6, Main Perth Shoreline and Lateglacial low sea level (Cullingford and Smith, 1966 ; Sissons, 1976b). Although the timings at which the major marine regressions and transgressions took place may be subject to modification, if suitable datable material can be found the form of the curve is unlikely to change significantly with further research.

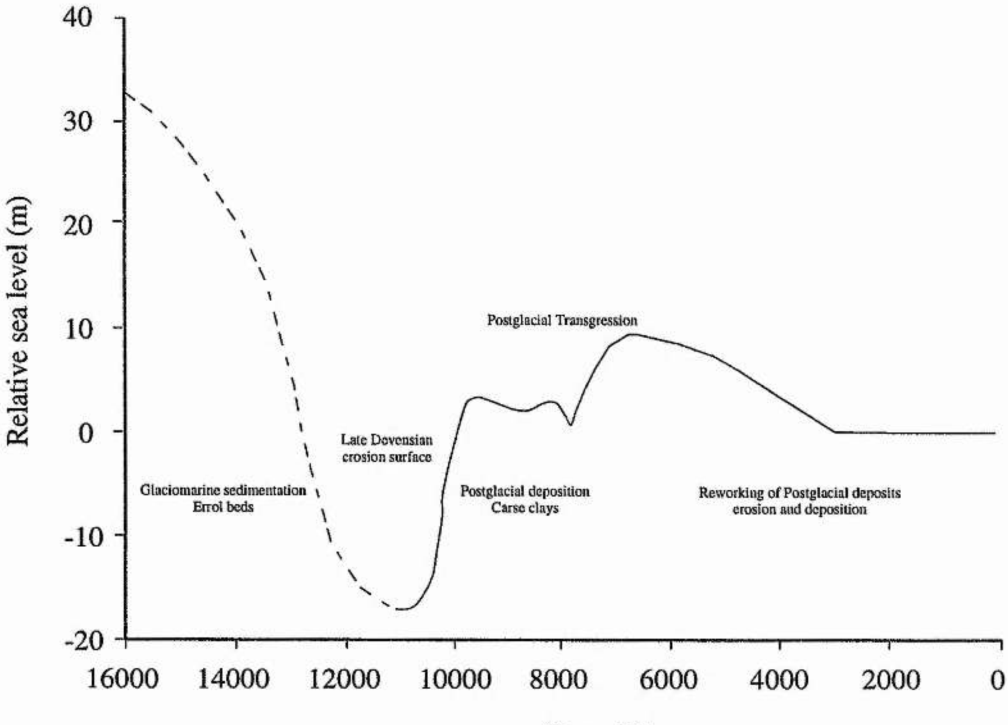


Fig. 2.6 Relative sea level curve in study area (for source see text).

2.4 Postglacial shoreline development

During deglaciation, large supplies of unconsolidated sediments were contributed to St Andrews Bay by glacial and fluvioglacial processes. These mobile supplies, combined with changing relative sea level, have given rise to considerable modification of the planform of the embayment over the last 15,000 years. The early history is poorly dated and incomplete, but postglacial changes are better documented. Associated with planform development has been nearshore erosion and sedimentation resulting in bathymetric changes which, in part, can be evaluated from the sequence of subsurface sediments beneath the floor of the bay.

Most changes in the area have taken place in the nearshore zone where there is evidence for shoreline regression and progradation. These changes are predominantly recorded at the head of the bay between Carnoustie and St Andrews whereas the rocky cliffed coastlines show limited evidence of change from the time of the ice withdrawal.

Despite the data presented in the shoreline diagram (Fig. 2.5) the evidence for the former shoreline of the bay is fragmentary, poorly preserved and poorly dated prior to the Holocene. Only one shoreline, the 'Main Perth Shoreline', can be traced around the majority of the embayment (Fig. 2.5). However, although it is defined in the outer part of the bay, the inland extent is subject to debate with some workers (Cullingford, 1972) arguing for a westward penetration of the sea to Perth, whereas others (Armstrong et al., 1985) have argued for marine conditions extending to Crieff some 15 km west of Perth. Uncertainty about the extent of the Tay Estuary leads to a poorly-defined understanding of the sediment inputs to the coastal zone and possible sedimentation processes at that time; i.e. to what extent sediment was glacially or fluvially derived.

Conditions in the bay during this time have been investigated by sub-bottom profiling and boreholes by the British Geological Survey. This work shows the presence of an extensive glaciomarine unit, the 'St Abbs Beds', which covers the area and is believed to date from this time. The faunas in the beds indicate deposition in an environment of arctic character (Hall and Jarvis, 1989). The sediments are at least 20 m thick over much of the Bay and are found onshore particularly in the Tay Estuary (Browne and Jarvis, 1983). The presence of this material is believed to be important in the subsequent evolution of the bay since it would have provided a major source of unconsolidated sediments which could be remobilised by wave and tidal activity during the period of low sea level at the end of the Lateglacial.

The shape of St Andrews Bay at the Lateglacial minimum sea level is uncertain. A channel entrenched through the St Abbs Beds sediments to a depth of 40 m below OD in the Tay Estuary is graded to a low sea level in St Andrews Bay (Buller and McManus, 1972). The subsurface erosion surface identified in the bay also relates to this time and suggests that much of the bay was then dry land .

A clearer picture of the changing nature of the bay emerges with the start of the Holocene and the main postglacial transgression which took place from about 10,000-6,000 BP. At the maximum extent of the transgression, the shoreline stood some 7-9 m above its present level at St Andrews and the head of the bay was defined by a cliffline extending from St Andrews (GR. 510158) to Guardbridge (GR. 455195) and from St Michaels (GR. 445230) to Tayport (GR. 465287). The higher sea level also entered shallow re-entrants in the cliffline between St Andrews and Fife Ness and on the northern shore of the bay. In the estuaries and coastal zone of the postglacial sea, sequences of silt, sand and clay , the "carse clays" were laid down containing a fauna of Boreal aspect and occasional peat beds associated with periods of emergence (Sissons, 1967).

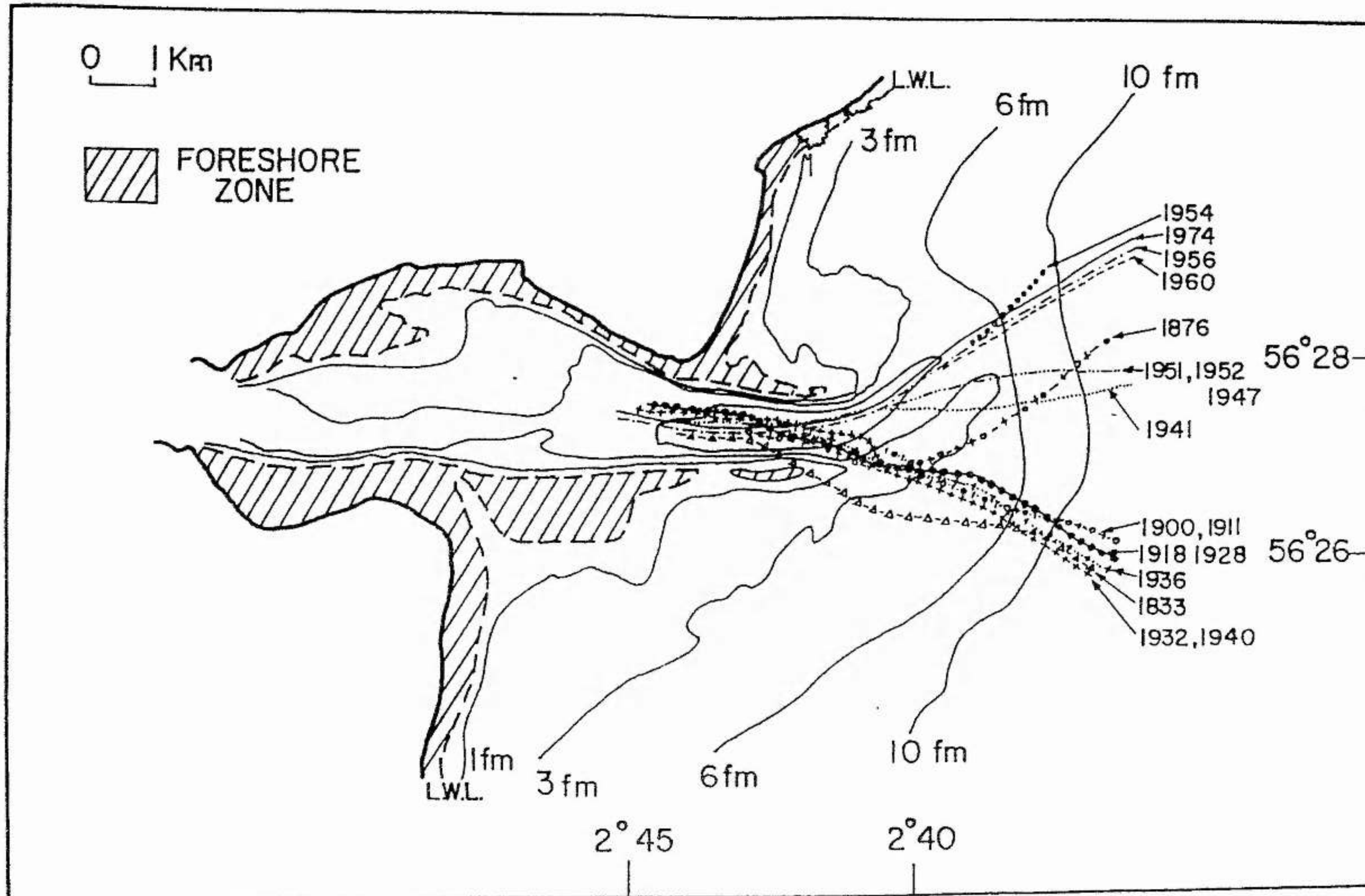
These postglacial deposits were cut into during the period of falling sea level after the transgression and the present planform of the embayment started to emerge from this time. Most significant has been the abandonment of the postglacial cliff line and the eastwards progradation of the shoreline at the head of the bay. Clearly, the present Eden estuary is a feature of the postglacial regression in the area, since at the maximum sea level the estuary would have only existed landwards of Guardbridge. Also dating from this period has been the growth of the Tentsmuir platform which is composed of windblown sand overlying intertidal sands and occasional peat beds. This process of accretion continues at the present day. North of the Tay the growth of Buddon Ness (Fig. 2.1) and the dune complexes between Carnoustie and Arbroath may be seen as a direct response to falling sea levels during the regression and the period of stability since about 3,000 years BP.

2.5 Historical Changes

The large-scale accretion at the head of the St Andrews Bay and associated changes in the configuration of the shoreline of the bay can be traced with some precision since 1855. Since that time regular topographic and , since 1940, air photographic surveys of the area have been undertaken. The resulting changes have been documented by Ferentinos and McManus (1981) and Wal (1992) and will not be detailed here.

It is noted that, since 1855, Pilmour Spit (Fig. 2.3) at St Andrews (Fig. 1.1) has extended seawards and northwards by 300 to 500 m, with some evidence for periods of erosion especially at the northern tip. This contrasts with the behaviour of Coble Point (Fig. 2.3) in the middle of the Eden Estuary which has not changed significantly in 100 years and seems to be a relict feature probably formed at a time of high sea level. Substantial progradation of the shoreline is also noted at Tentsmuir with an eastward advance of the shoreline by up to 700 m since 1855. There is more evidence that accumulation here, although progressive, has not been continuous and that periods of erosion have occurred. Thus, near Kinshaldy (GR. 488236) concrete blocks placed at the high water mark in 1941 were buried by accretion and then exposed to be 60 m forward of the dune ridge in 1984. However, in 1992 they were again being buried by the growth of small foredunes.

Comparison of offshore hydrographic surveys, carried out between 1833 and 1973 reveals that the axial orientation and the depth of the inner channel of the Tay Estuary have changed little, but dramatic changes have taken place in the axial orientation and in the depth of the outer channel (Fig. 2.7) (Ferentinos and McManus, 1981). The position of the Eden channel has changed frequently since 1855. However, since 1949, aerial surveys reveal a northwards migration of the channel at a rate of about 10 m/yr (Eastwood, 1977).



Contours in fathom; fm = 1.829 m

Fig. 2.7 Long-term variation of the axis of the outer part of the Tay Eastuary.
L. W. L. is Chart Datum. (after Ferentinos and McManus, 1981).

2.6 Meteorological and Oceanographic setting

Wind records taken from Bell Rock lighthouse some 35 km to the east of Dundee, during the ten year period 1960 - 1970 show that the prevailing winds in St Andrews Bay were from south-west and west respectively. Winds from north-east and south-east are also important (Ferentinos and McManus, 1981) because they may blow over a long fetch and can generate large waves. Similar results were reported by Wal (1992), who found that the average wind data for the last ten years, 1980 to 1990, show that most of the winds are derived from the south-west. The next in relative abundance are the winds from the east and north-east (Fig. 2.8). Analysis of the wind data from the Leuchars meteorological station during the period 1.5.1991 to 31.7.1992 (Appendix 1) shows that (Fig. 2.8) the dominant winds in the area were from south-west and north-west respectively. The next in relative abundance are the winds from west and north-east. According to Wal (1992) the study area lies within the dry belt (average rainfall for 1980-1990 is 670 mm) which extends along most of the east coast of Scotland. Air temperatures show that the minimum and maximum winter temperatures are 1°C and 6° C, respectively. During summer, the average temperature in the area varies between 10° C and 19° C. Water temperatures, recorded during the field work shows that during winter the average sea water temperature in the Bay is 6° C and 10° C during summer.

Water circulation in the marine environment is a result of the combination of meteorological and astronomical forces such as winds, waves and currents. Tidal currents in St Andrews Bay are derived from the tides in the North Sea. According to Charlton (1980), the North Sea tide induces an eddy in the bay in a direction which is counter to the driving force in the North Sea. Offshore from St Andrews Bay the flood and ebb currents flow in southerly and northerly directions; this simple tidal pattern is modified within the embayments (Ferentinos and McManus, 1981). It is well documented that tidal currents in the study area are weak and unable to transport the sediments. Away from the influence of the estuary in St Andrews Bay spring tidal currents do not exceed 30 cm s^{-1} at one metre above the bed (McManus, 1984). Tidal currents are low but show a strong landwards residual in St Andrews (Jarvis and Riley, 1987). Recent current meter data recorded at 4 stations for

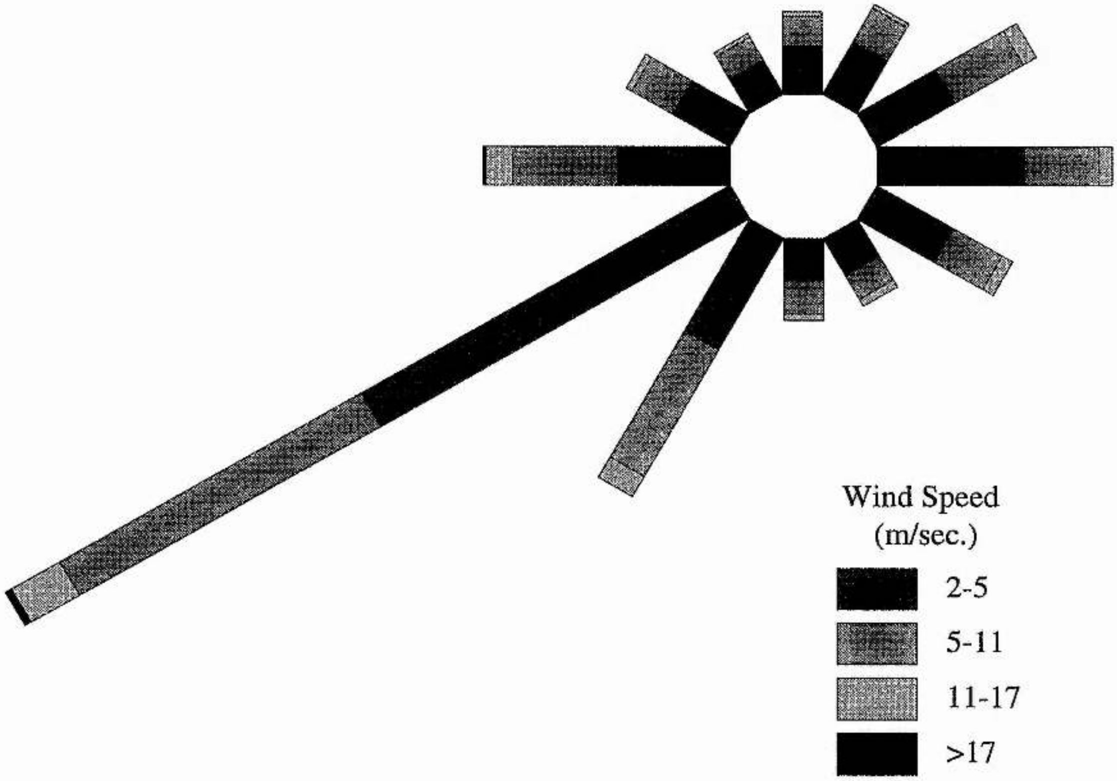


Fig 2.8 a Average wind rose for ten years (1980 - 1990, After Wal, 1992)

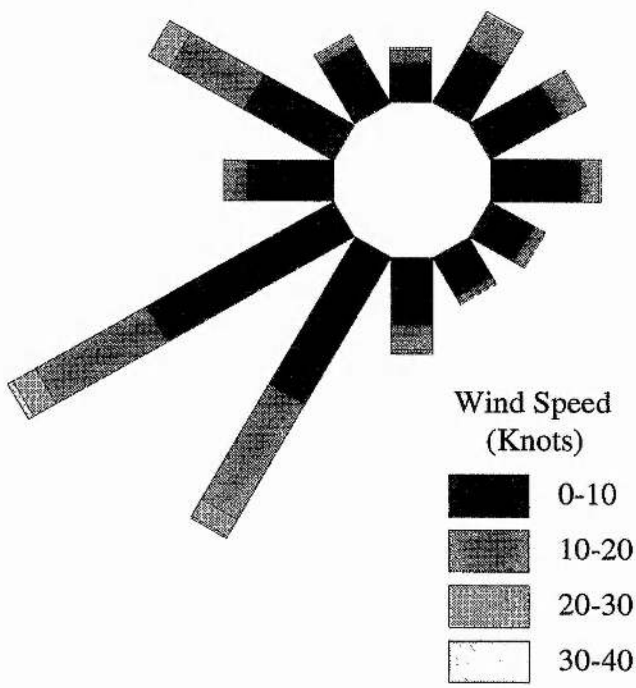


Fig 2.8 b Average wind rose for 15 months (May 1991 -July 1992)

approximately 40 days, accord with previous studies and show that tidal currents are weak. Occasionally, when there are strong winds additional circulation is generated which may cause near bed currents that may reach values of between 20 to 30 cm s^{-1} (See Chapter 4). The principal freshwater source is the River Tay but there is no sign of a stratification within the water column within the embayment. This is probably because the tidal range is high and the tidal currents are strong in the main channel, which cause the water column to become completely mixed. Sediment movement in the area more likely depends upon meteorological, rather than on astronomical forces. Analysis of wave directions reveals that about half of the recorded waves in the area are propagated shorewards with a direction of approach of between 20° and 180° (Ferentinos and McManus, 1981). The easterly-facing geographical site of the bay is thus exposed to winds or waves which come from the east. It is this direction of wave attack which is the most important and strongly affects the coastal area of Tayside and North East Fife. This is because the easterly waves have travelled long distances (750 km). A wave refraction study to estimate the potential drift in the St Andrews Bay and then to determine the location of zones of potential erosion, transport and deposition has been carried out according to the model of May (1974). Examination of the model predictions reveal that Tentsmuir Beach receives least wave energy under most conditions tested. Studies of wave height and period in St Andrews Bay by Boalch (1988) have indicated that the most frequent waves occurred with a height of between 0.3 and 1.0 m and a period of 3 to 7 s, whilst storm waves varied in period from 4 to 8 s and in height from 1 to 3 m. However, the sea bottom slopes gently eastwards, and with the offshore area also characterised by shallow depths, most waves as they move toward the shore will gradually lose their energy. Consequently, less energy will arrived at the shore.

2.7 Discussion

In the past the study area was subject to variations in climatic conditions. Such changes in the environmental conditions would have resulted in different processes in the area and, subsequently, led to changes in the morphological and sedimentological environments of the area. The coastal area adjusts very readily to any change caused by different climate. As the ice cover left the study area a long time ago (15,000 years BP), then isostatic depression and uplift would have produced several fluctuations in relative sea level. These fluctuations mean that the shape and the position of the coastline of St Andrews Bay will also have changed. This lead to the formations of raised beaches or submerged shorelines. However, our knowledge of former coastlines, now above sea level, is far better than our knowledge of submerged coasts. Several raised beaches are recognised along the coast of St Andrews Bay, whilst a few submerged shorelines have been identified in the nearshore area. As sea level rises, the bay will be subject to a lesser effects of tidal friction and wave reworking. This will result in a general widening and deepening of the bay. On the other hand, if sea level falls the bed sediments become exposed to waves and tidal currents which can cause reworking and subsequent sediment transport. At the present time, it is necessary to monitor the change in the coastline in order to determine the nature and rates at which changes are taking place under present-day processes. Change in the coastline of the study area have been evaluated by successive map analysis, since 1850. Air photography from 1940 also provides a very detailed record of change. The results indicate that the area has still not reached an equilibrium state with present-day processes. During 120 years of monitoring, the results have revealed the existence of a continuous sediment supply which have caused a net seaward growth of Pilmour Spit and Tentsmuir Beach. In both cases, it is almost certain that these coastal features are still being shaped by recent processes and must be the result of contemporary marine processes. In contrast, other features such as raised beaches, coastal rocky platforms and abandoned stacks are relict features which have been lifted above present sea level by isostatic recovery of the land after the melting of the ice sheets which locally depressed the earth's crust.

CHAPTER III

Grain size distribution of sediments in St Andrews Bay

3.1 Introduction

Grain size is a basic descriptive measure of sediments and a great deal of attention has been given to the usefulness of grain size measurements as indicators of the depositional environment. However, "grain size" depends upon the particle properties being measured, and the description of the sediment assemblage depends upon the choice of statistic to describe the size distributions. Early workers (Wentworth (1922, 1929), Trask (1932), Krumbein (1932, 1934, 1938) and Krumbein and Pettijohn (1938)) first described the size frequency distribution by various statistical coefficients. New formulae for grain size statistics were presented by Inman (1952) and these were modified by Folk and Ward (1957) to include a greater part of the distribution curve. Many investigators have used multivariate plotting in attempts to discriminate the sediments of different environments, such as rivers, beaches and dunes.

The variables that are most effective for environmental discrimination are those related mainly to the tails of the size distribution i.e. skewness and kurtosis. For example, Friedman (1961, 1966, 1967) used mean grain size and skewness parameters to distinguish between dune and beach sediments, principally because dune sands are positively skewed. Moreover, if sorting is plotted against skewness, river and beach sands may be distinguished because river sands are more positively skewed and less well sorted (Pethick, 1984).

Perfect, log-normal grain size distributions plot as straight lines on the probability paper (McManus, 1988). However, it has been suggested that the mixing of modal populations is the most important factor controlling the deviation of sediment distributions from log-normality (Visher, 1969; Walton et al., 1980). Studies of the grain size distribution curves of sediments may help to predict which types of distributions are typical of specific hydraulic conditions (Chamley, 1990), since different size fractions are moved by different mechanisms of transportation. It is usual (Visher, 1969) to distinguish between the populations of the finest sediment, which is held almost continuously in suspension, and those of the material that is moved mainly on, or close to, the bed (Malaz, 1981).

The aim of the present grain size distribution study is to identify the textural characteristics of the sediments in St Andrews Bay and use them to give a measure of the variation in the hydrodynamic energy in different parts of the bay.

3.2 Previous work

A number of studies of sediment size distribution have been carried out in St Andrews Bay. Green (1974) investigated the relationship between grain size distributions and tidal circulation in the northern half of the bay. Residual flow in the area was measured and the potential sediment transport was determined in an area extending from Tayport Beach, eastward to the Abertay Sands including Tentsmuir Beach. Green (1974) stated that, in general, the sediments in the southern entrance zone of the Pool Channel (Fig. 3.1) exhibit very little variation in mean grain size, with most of the sediments being of fine to medium sand grades. The work of Green was followed by that of Eastwood (1977) who studied the sedimentology of the superficial deposits of the Eden Estuary and noted that the channel sediments are characteristically coarser than those of the intertidal flats. In the lower estuary, east of Sanctuary Spit, the mean grain size varies between 1.95 and 2.92 Ø. East of Outhead, along the West Sands and in a lobate zone trending south-west, to the west of Outhead, the sediments are slightly finer, with mean grain sizes of 2.5 to 2.92 Ø . Al-Mansi (1986) studied sediment transport and beach erosion in Monifieth Bay and showed that this area is characterised by medium to fine sand, ranging from 1.73 to 2.58 Ø in mean size. The sediment is coarsest off Monifieth and becomes finer both to the east and west. Boalch (1988) points out that, in Monifieth Bay, the sediments show minimal variation in grain size characteristics throughout a seasonal cycle. He related this to the protected location and the presence of the Abertay and Gaa Sands which act to filter the energy of easterly waves. In more exposed areas, like Tentsmuir, Carnoustie and east Barry, the dominantly sandy sediments show characteristic variations during a seasonal cycle. Boalch (1988) suggests that the coarsest and most poorly sorted sediments are characteristic of the potential higher energy wave regime that exists between autumn and spring, whilst the finest and best sorted sediments typify the potential lower wave energy regime of summer. Further examination of the sediment characteristics of the Gaa Sands, the western part of the Abertay Sands and the eastern Abertay Sands was also undertaken by Boalch (1988) in order to compare previous results (obtained by CIRC in 1978) with recent data. He found that the Gaa and western Abertay Sands sediments are both composed of medium, moderately sorted sand, which is near log-normality size distributed. Samples from the eastern Abertay Sands are better sorted and contain an abundance of finer materials which results in positive skewness. Boalch's (1988) results show that the addition of fine material improves the sorting of the sands and shifts the grain size distributions from negative skewness towards log-normality.

These previous researches were carried out on the western and northern margins of the present study area. However, no investigations dealing with the grain size distribution of sediments in St Andrews Bay itself, or further offshore, have been reported.

3.3 Methods of study

3.3.1 Field methods

A total of 113 samples were collected during May and July 1992 from a wide range of locations in St Andrews Bay using a van Veen grab sampler. This collected material from the upper few centimetres of the sea bed often returned a relatively undisturbed sample showing worm casts. The samples, each weighing approximately 0.5 kg, were taken from water depths of between 2 and 20 m. Position fixing was by means of a Decca navigator system, giving precision of ± 50 m. The sampling sites are as shown in (Fig. 3.1).

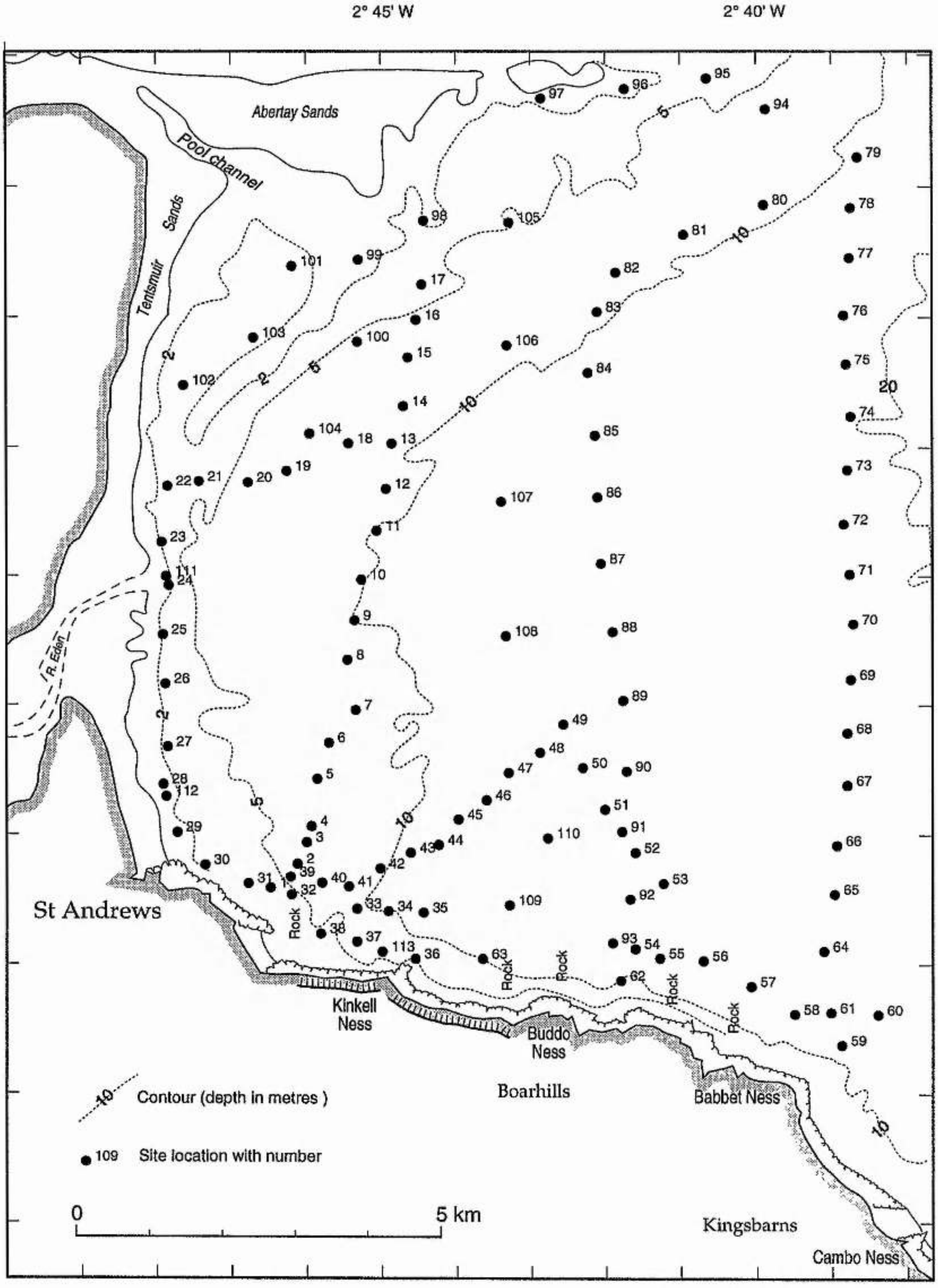


Fig 3.1 Location of Sediment Sample Sites.

3.3.2. Laboratory methods

The samples were dried at 100° C for 24 hours, then disaggregated gently by hand. After quartering, approximately 70-100 g of each sample was taken to be analysed. Owing to the fact that the percentage of coarse sand is generally less than 10%, whereas fine and very fine sand account for more than 90% of the samples, coarse material was sieved at 0.5Ø intervals and fine/very fine material was sieved at 0.25Ø intervals (Appendix 2). The samples were first screened through set of a 0.5Ø interval standard sieves with mesh diameters of -1Ø, - 0.5Ø, 0Ø, 0.5Ø, 1Ø, 1.5Ø, 2Ø . The material finer than 2Ø which was retained in the pan was first weighed and thereafter washed into a 4Ø mesh. The samples were then dried at 100 °C and, after cooling, were sieved at 0.25Ø intervals. All samples were shaken for 15 minutes (Folk, 1980). The fraction which was retained in each sieve was weighed and the percentages calculated.

3.4 Analysis of the size distribution

Many measures are used to describe the grain size frequency distributions of sediments. Computation of these measures varies greatly in complexity. Computing the various moments of the sample distribution ensures that every grain in the analysed sample is used in the evaluation of the parameters. In essence, graphical methods involve extracting grain size data from a few points on the cumulative frequency curve of the sediment (Inman, 1952 ; Folk and Ward, 1957). Each method has its advantages and drawbacks in determining mean size, sorting, skewness and kurtosis, and each is equally valid for comparing a suite of samples. The method of moments, which has been used by Griffith (1961) and Friedman (1962), is not suitable for open ended distributions (Chauhan and Chaubey, 1989). This method has thus not been used in this study. Folk(1966) advocated the use of graphical techniques for a visual appreciation of grain size distributions.

In the study area, various formulae have been used to obtain the statistical parameters using the graphical methods. The parameters defined by Trask (1930) and Krumbein (1936) are based on quartile measurements, the 25th, 50th and 75th percentiles of the size distribution. Doeglas (1946) showed the inadequacy of quartile measurements to express some salient features of sediment distributions. Inman (1952) extended the range of the curves analysed to include the 16th and 84th percentiles, and Folk and Ward (1957) used still more of the cumulative curves (5th and 95th percentiles) to yield more discriminating results. According to Buller and McManus (1979), graphical methods are the most popular means for the computation of statistical parameters from sedimentary data. Graphical statistical analysis has thus been used throughout the present study.

3.5 Graphic presentation

Attempts to construct any relationship between the statistical parameters of grain size distributions which have been obtained from the study area have not yielded any significant correlations. McManus (1988) suggested that, with deposits of limited size variations, as in St Andrews Bay, plots of bivariate scatter-grams have some success in discriminating between depositional environments but better discrimination is achieved where there are large size differences. One scatter-gram plot is presented, namely mean size (Mz) versus median size (Md) (Fig. 3.2) which shows the close relationship between these parameters. Ternary diagrams, where the total sediment at any site is divided into three fractions, such as gravel, sand and finer material (silt and clay) are also used to discriminate sediment types. However, in the study area most of the sediments are confined to the fine sand fraction and thus ternary diagrams are of little value.

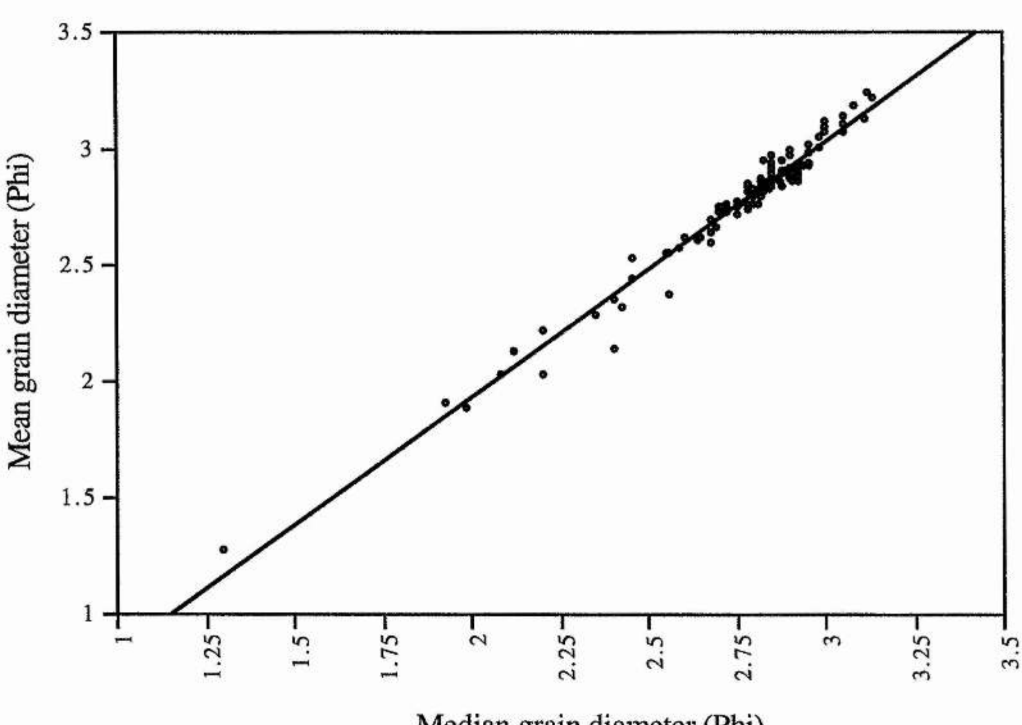


Fig.3.2 Bivariate scatter grams of mean grain diameter versus median grain diameter.

The histogram, the simplest graphical form of data presentation is used in this study. The advantages of the histogram are that the trends in the frequencies of individual size grades can be seen at a glance and that the identification of primary and secondary modes is facilitated. The discontinuous character of the histogram may be avoided by constructing the frequency curve which passes through the centre of the bars and allows the definition of modes. In addition, the data are presented in graphic form using a plot of cumulative percentage on a probability scale against grain diameter, in Ø units. This is the ideal curve for the extraction of percentile values which provide the data for quantitative statistical characterisation and analysis. The size parameters, namely Median Grain Size (Md), Graphic Mean Size (M_Z), Inclusive Graphic Standard Deviation (s_I), Inclusive Graphic Skewness (Sk_I) and Inclusive Graphic Kurtosis (K_G) were calculated according to the equations given by Folk and Ward (1957).

The equations used are:

$$Md = \emptyset_{50}$$

$$M_z = \frac{\emptyset_{16} + \emptyset_{50} + \emptyset_{84}}{3}$$

$$\sigma_1 = \frac{\emptyset_{84} - \emptyset_{16}}{4} + \frac{\emptyset_{95} - \emptyset_5}{6.6}$$

$$Sk_I = \frac{\emptyset_{16} + \emptyset_{84} - 2(\emptyset_{50})}{2(\emptyset_{84} - \emptyset_{16})} + \frac{\emptyset_5 + \emptyset_{95} - 2(\emptyset_{50})}{2(\emptyset_{95} - \emptyset_5)}$$

$$K_G = \frac{\emptyset_{95} - \emptyset_5}{2.44(\emptyset_{75} - \emptyset_{25})}$$

The spatial patterns of the various statistical parameter values are plotted on base maps of the area in order to examine regional variations.

3.6 Statistical parameters of grain size

3.6.1 Median (Md Ø)

The median is that size for which half of the particles (by weight) are coarser and half are finer. The median sizes of the sediments of St Andrews Bay range from 1.3 Ø to 3.13 Ø, from medium to very fine sands (Fig. 3.3). Although fine sand dominates the sediment samples in St Andrews Bay a few samples are of medium sand size (samples 97, 102 and 103). These samples are from close to the Abertay Sands or near the shore of Tentsmuir beach. Patches of very fine sand are found south-east of the outfall of the River Eden, where median sizes are between 3.05 and 3.10 Ø (samples 34, 35, 44, 45 and 46). This size also dominates some samples collected from offshore (e.g. samples 73, 74, 75, 76 and 77). In spite of the small variation in median grain size from one site to another, the general picture, from the median size distributions in the area, indicates that the bed sediment becomes coarser on-shore and towards the mouth of the Tay Estuary but finer offshore as water depths increase (Appendix 3).

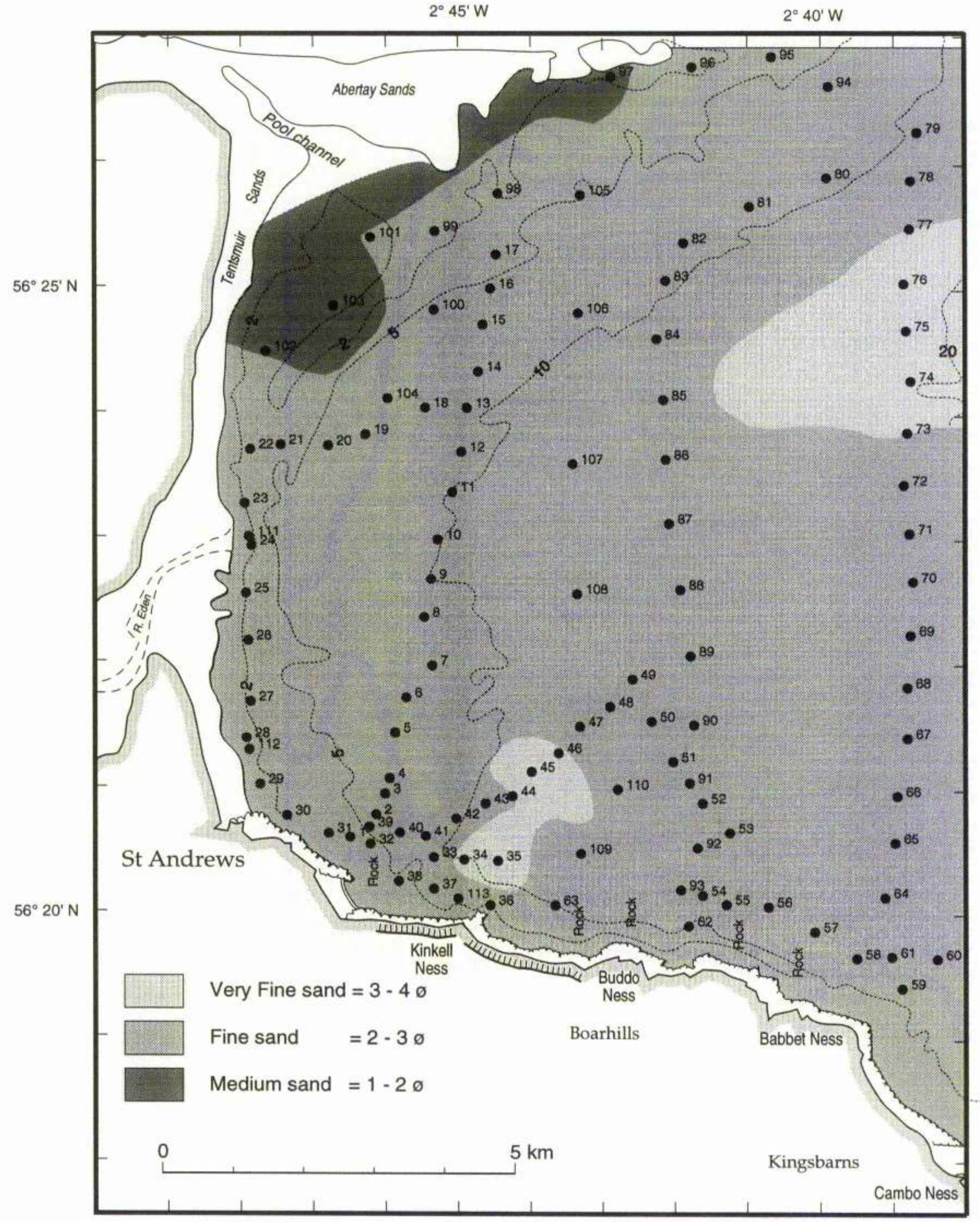


Fig 3.3 Regional distribution of Md in St Andrews Bay.

3.6.2 Graphic Mean Size (M_Z Ø)

The mean size is the measure of the overall size of grains forming the sediment. In general, the study area is characterised by fine to very fine sand with a few samples in the medium sand size fraction, ranging from 1.27 Ø to 3.24 Ø (Fig. 3.4). In most cases the mean and median are similar, with differences being very small. The sediment is coarsest approaching Tentsmuir beach and the Abertay Sands, but becomes finer farther east, in particular towards the centre of St Andrews Bay. There is some variation close to the Abertay Sands, for example the mean size is 2.14 Ø at sample 17, 1.91 Ø at sample 97 and 1.27 Ø at sample 103. In the deeper water of the north-eastern offshore section of the bay (samples 71,72, 73, 74, 75, 76 and 77) the mean sizes have decreased to between 3 and 3.24 Ø. Although the overall variation in the mean size of the samples is very small, consistent trends can be detected from site to site.

3.6.3 Inclusive Graphic Standard Deviation(σ_I Ø)

Sorting is a measure of the degree of uniformity or the relative size of the standard deviation of grains forming a sediment. The majority of samples in St Andrews Bay are well sorted but moderately well sorted samples with sorting values ranging from 0.5 to 0.69 Ø are found throughout the bay (Fig. 3.5). Samples 20, 26, 38 and 111 where found to be very well sorted. The overall picture of this measure shows that the best sorted sediments occur along the coastal zone off the Tentsmuir Beach and Outhead, whereas the poorest sorting is associated with the coarser sediments of the Abertay Sands or with the finest sediments offshore.

3.6.4 Inclusive Graphic Skewness (Sk_I)

Skewness is the measure of the degree of symmetry shown by a size frequency distribution. In a normal distribution, with a bell-shaped frequency curve, the median and mean values coincide. Any tendency for a distribution to deviate from normality well lead to differences between the median and mean values. These differences are used to characterise

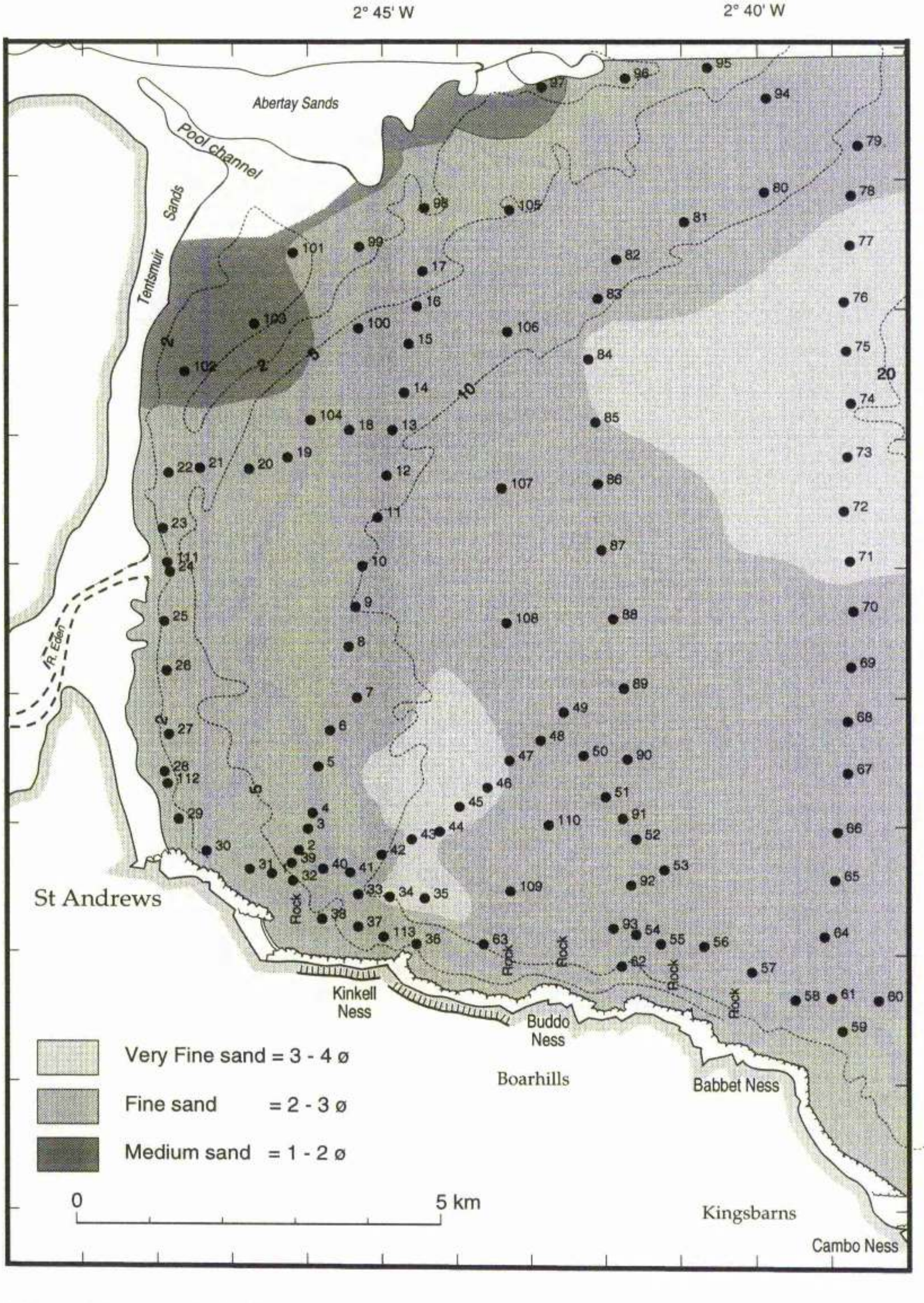


Fig 3.4 Regional distibution of Mz in St Andrews Bay.

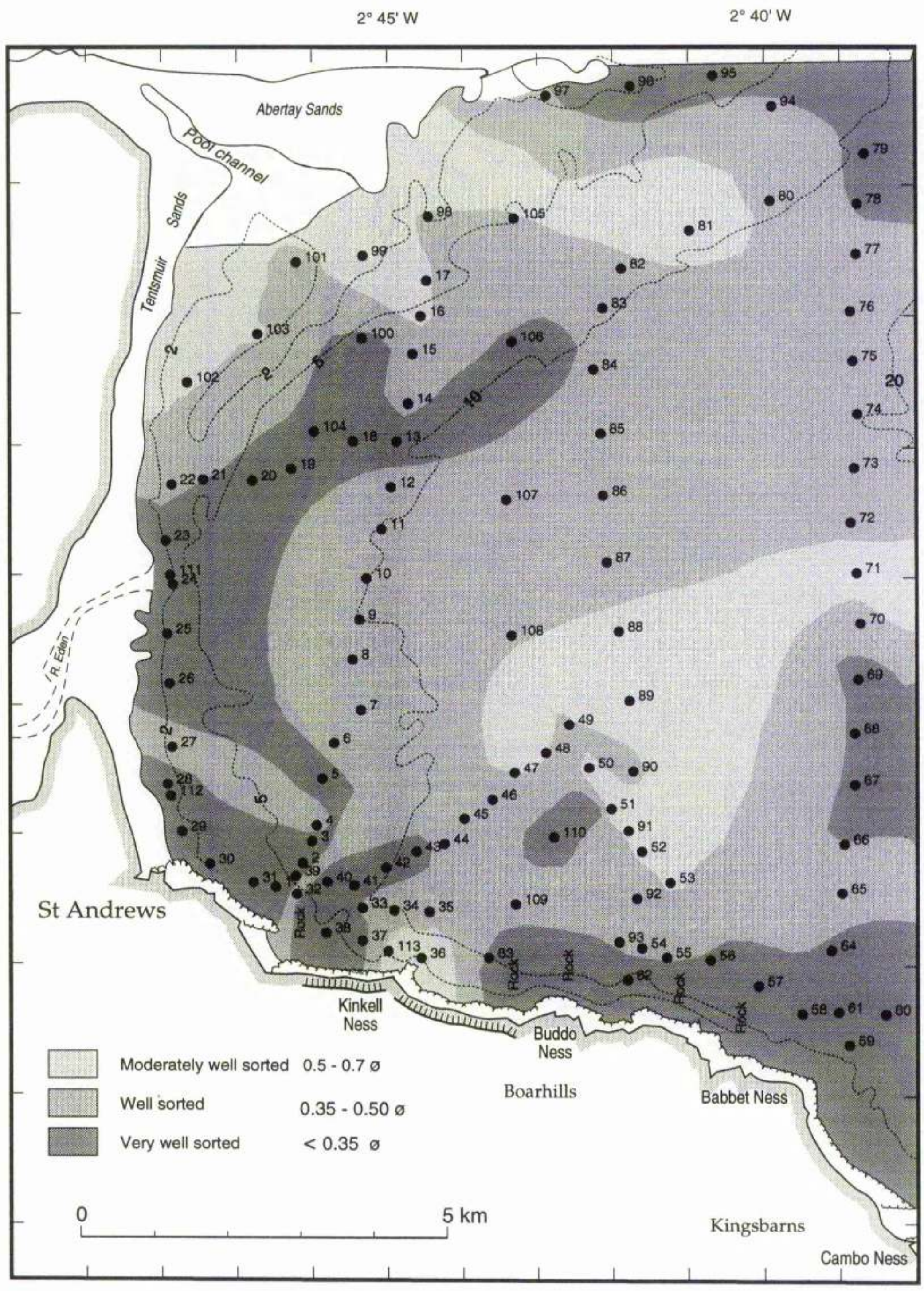


Fig. 3.5 Regional distribution of σ_I in St Andrews Bay.

the asymmetry or skewness of the curve (McManus, 1988). Positive values indicate distributions skewed toward the finer grades and negative values indicate distributions skewed toward the coarser grades. The major trend for the distribution of skewness of the sediments in St Andrews Bay (Fig. 3.6) is for nearshore sediments and those of the Abertay Sands to display skewness ranging between negatively skewed and very negatively skewed. In the central part of the bay, symmetrical distributions are characteristic. But farther offshore, the sediment distributions are skewed toward the finer grades in "patches" which vary between symmetrical, fine skewed and strongly fine skewed. Whereas, along the southern margin of the bay and parallel to the West Sands the sediment shows a skewness range varying between coarse, symmetrical and fine. The area in front of Tentsmuir Beach and around the Abertay Sands shows a systematic variation in the deviation of the sediment distribution towards the coarse grains; this is reflected by the gradual change in symmetrical, coarse and strongly coarse skewed material.

3.6.5 Graphic Kurtosis (K_G)

Kurtosis refers to the degree of peakedness or flatness of a size frequency distribution. Kurtosis measures the ratio between the sorting in the tails of the distribution and the sorting in the central portion of the distribution. If the central portion is better sorted than the tails, the frequency curve is said to be excessively peaked or leptokurtic. If the tails are better sorted than the central portion, the curve is said to be flat peaked or platykurtic (Lindholm, 1987). Usually there is a positive correlation between mean size and kurtosis so that one might expect that coarse sediments in the nearshore area will be characterised by platykurtic values. In deeper water, due to the limitation of wave effects and accumulation of fine material, sediments will be characterised by leptokurtic values. However, in St Andrews Bay the bed sediment does not show such a clear picture (Fig. 3.7). The kurtosis values in the area are mostly in excess of 1.11. This pattern may be because the well sorted sediments have very strongly peaked curves. Few samples show mesokurtic distributions (e.g. samples 38, 42, 44, 46, 73, 101 and 102) or platykurtic distributions (e.g. samples 45, 46, 74, 75, 76 and 88).

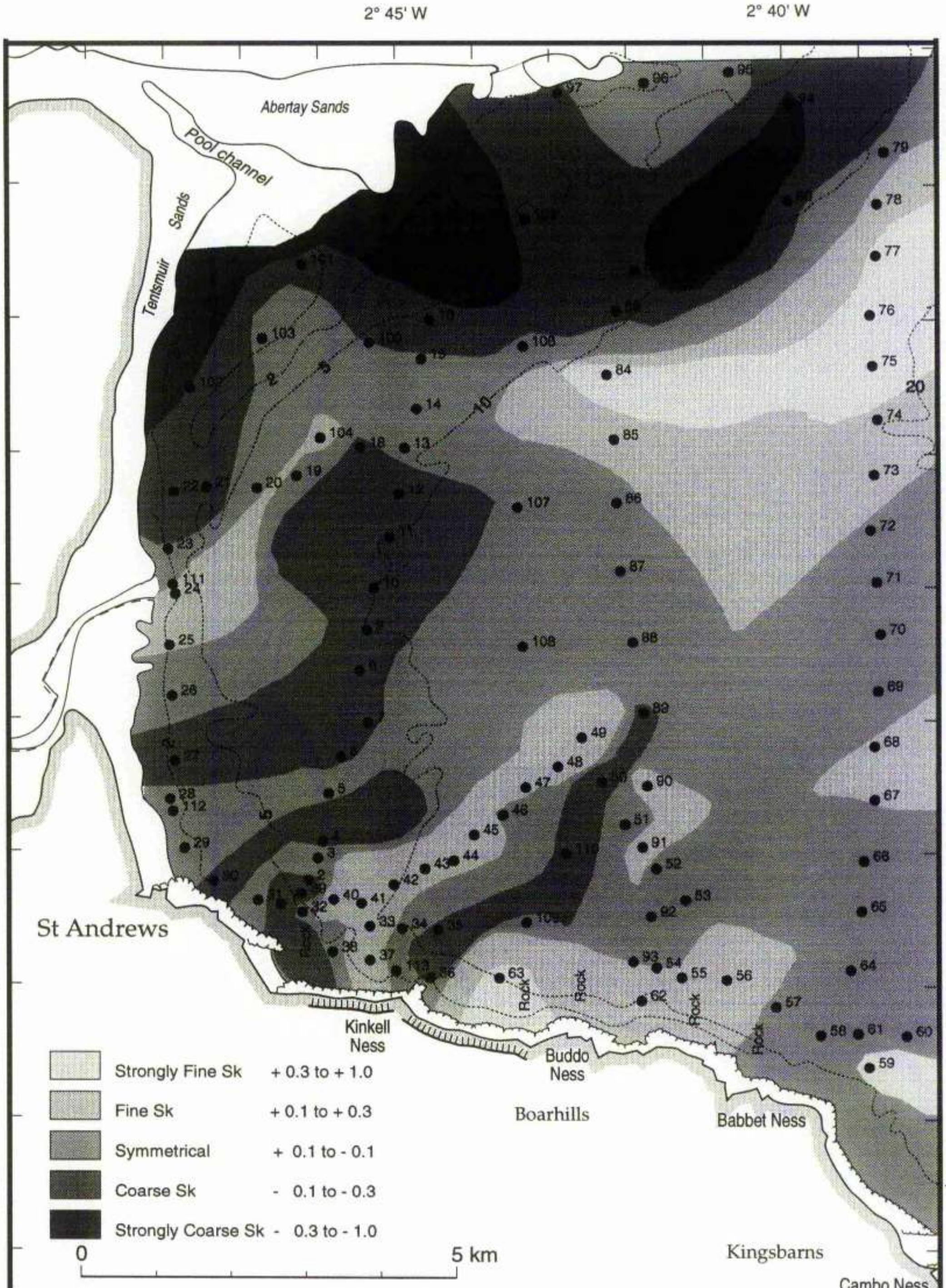


Fig 3.6 Regional distibution of Sk_I in St Andrews Bay.

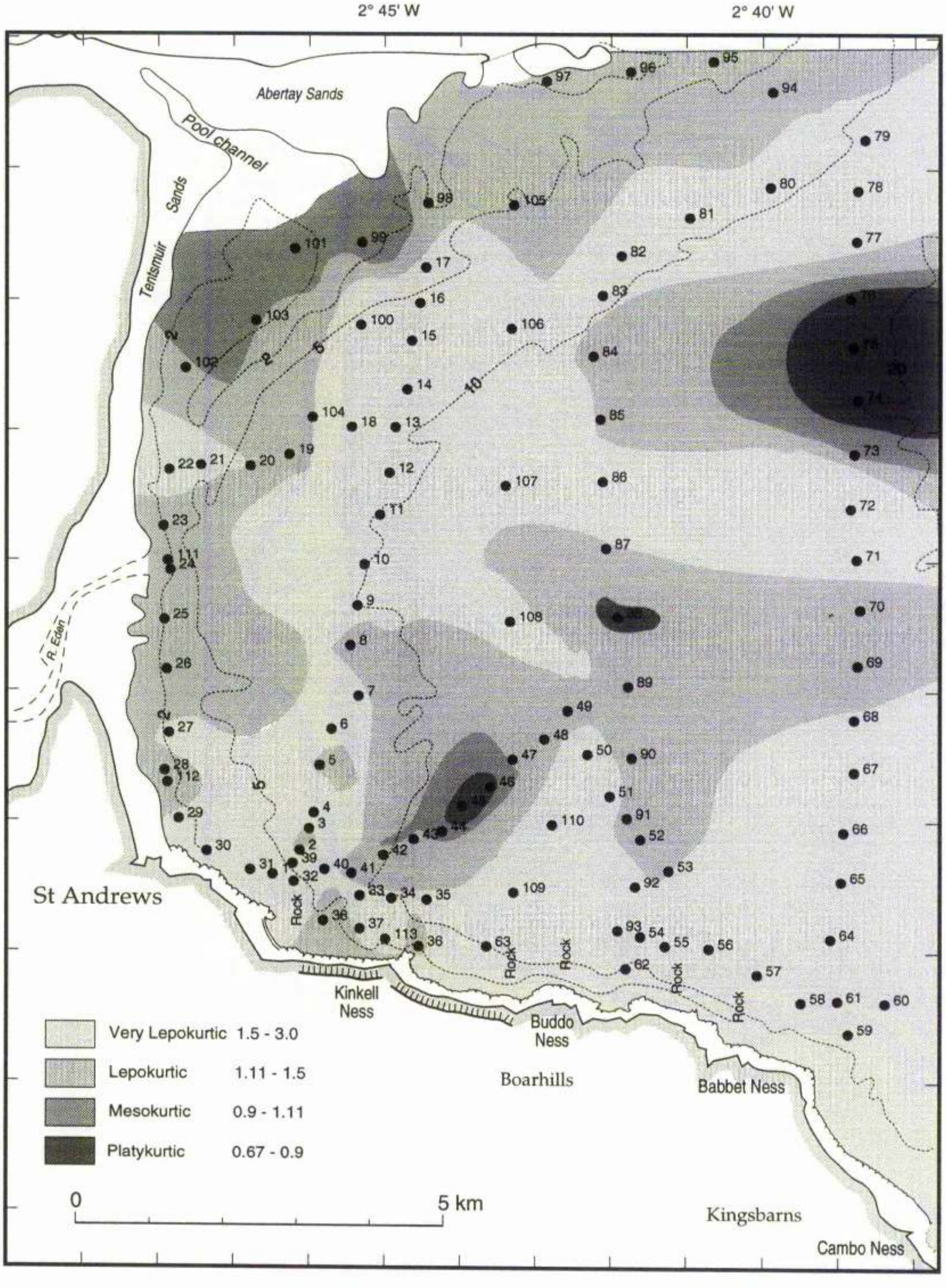


Fig 3.7 Regional distribution of K_G in St Andrews Bay.

3.7 Quartile Deviation-Median Diameter (QDa - Md) Analysis

Few techniques are available, at present, which enable one to discriminate between the effects of wave and current activity on the basis of grain size characteristics. Buller and McManus (1972, 1973a, b) developed a graphic technique to diagnose different sedimentary depositional environments using the quartile-deviation (QDa) and median-diameter (Md). These metric measures are defined by the equations of Krumbein (1939), as :

$$QDa = (P25-P75)/2 \text{ and } Md = P50$$

where P25, P50 and P75 represent the grain diameter equivalents of the 25th, 50th and 75th percentiles in millimetres. The plots of QDa (mm) versus Md (mm) for different depositional environments follow linear trends for which both the positions and slopes of the trend envelopes are different (Buller and McManus, 1972). In a subsequent analysis of the sediments of the upper Tay Estuary, Buller and McManus (1975) reassessed the trend envelopes and attempted to relate them to the processes, or rather the dynamic conditions, under which the sediments are formed. The method was successfully applied to distinguish current- and wave-dominated environments from sediments deposited in Start Bay, southern England (McManus, 1975). Since the bed sediments in St Andrews Bay are distributed according to the combination of waves and tidal currents (see Chapter 4) a similar technique has been applied here (Fig. 3.8) in order to see whether this method can distinguish between different dynamic environments for the deposition of the sediments.

3.8 Results

The bed sediments of St Andrews Bay are principally sand. Most of the sediments range between fine and very fine sand, with the exception of the sea bed close to the Abertay Sands (samples 97, 102 and 103) which are of medium sand. The occurrence of medium sand here may reflect the increase of energy in this area compared with other locations in the centre of the bay and in deeper water farther offshore.The statistical parameters characterise the sediment size distribution by single values. In order to explore the nature of the changes in grain size distributions within the bay, it is valuable to look at complete size distributions

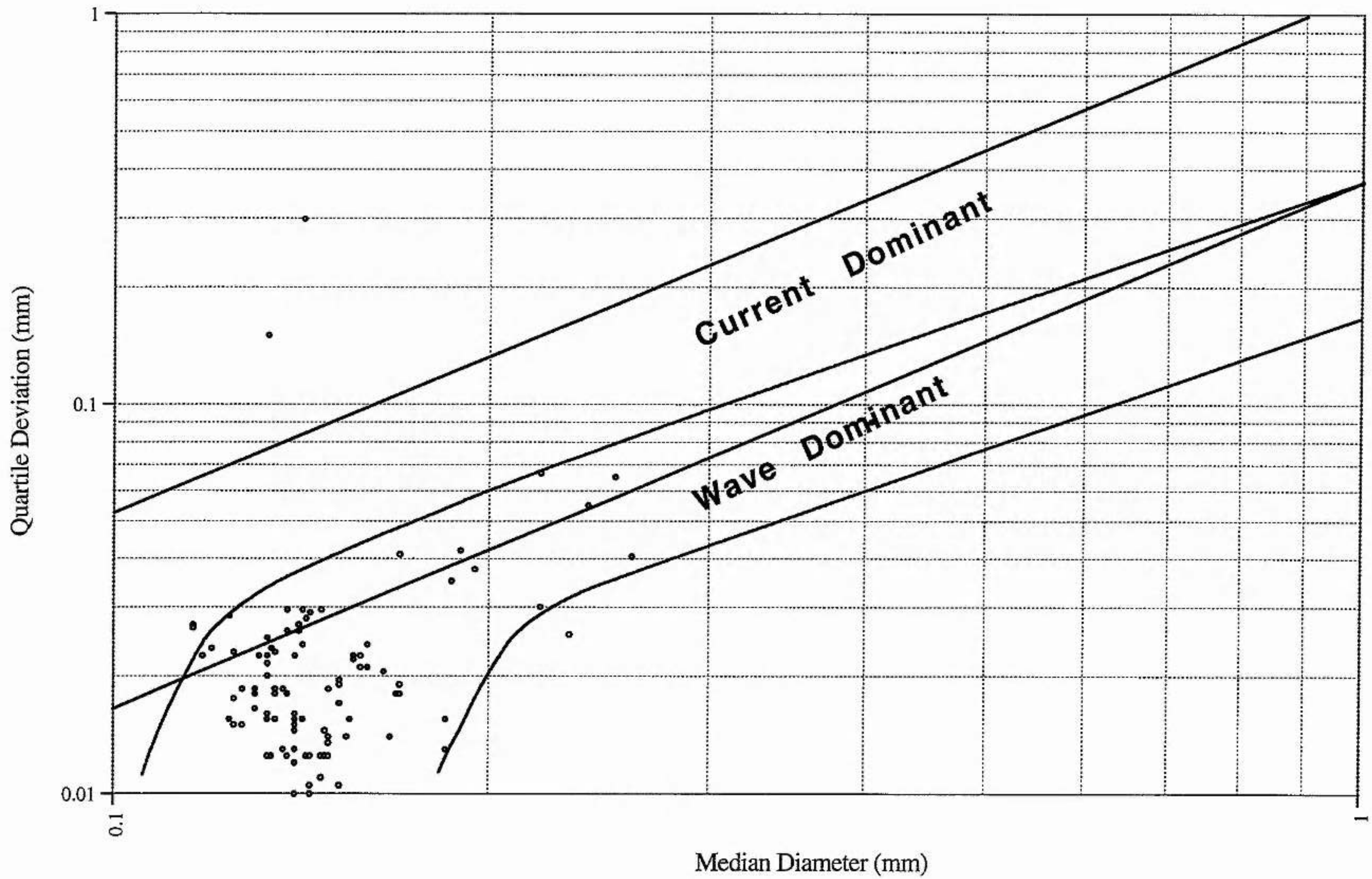


Fig. 3.8 Quartile Deviation - Median Diameter plots of bed sediments from St Andrews Bay (all samples), (Envelope trends Buller and McManus 1972).

by comparing histogram plots. For this purpose the samples have been grouped into six regions characterised by different depths. The results of this analysis are presented in Fig. 3.9 which illustrate changes in the primary and secondary modes from region to region. Each histogram plot in these Figures represents a group of bed sediment samples. For example His 1 represent the samples taken from depths between 1 and 3.5 m (Appendix 4). These samples, which are from close to the Abertay Sands or Tentsmiur Beach show a primary mode centred on 2.5 Ø and a coarse tail of medium sand size between 1.5 and 2 Ø. His 2 represents samples taken between Out Head and Kinkell Braes, St Andrews, and from very close to shore in depths of less than 3.5m. This histogram shows the dominance of the very fine sand grade as the primary mode centred on 3 Ø.

His 3 is similar to His 2 in the primary mode, but more fine material is present in His 3. This may be because the His 3 samples were taken from deeper sites than those of His 2 where the depth ranges between 5 and 9.6 m. His 4 represents samples taken close to shore but in water depths of 10 to 13.2 m, adjacent to the rocky platform. This histogram shows the dominance of the very fine sand grades forming the primary mode centred on 3 Ø and also that the entire frequency distribution has shifted towards the finer grades. One might assume that this area would be characterised by much more coarse material than by very fine sand because the wave energy should be stronger and more active close to shore. Accordingly, it is most likely that the depth factor accounts for the presence of finer material in this location. His 5 represents samples taken from the central part of the bay. The frequency curve distribution shows the size distribution skewed towards the finer material with a substantial fine tail. The histogram of His 6 is similar to His 5, the principal difference being the greater percentage of very fine sand. There is a decrease in the percentage of coarse material present.

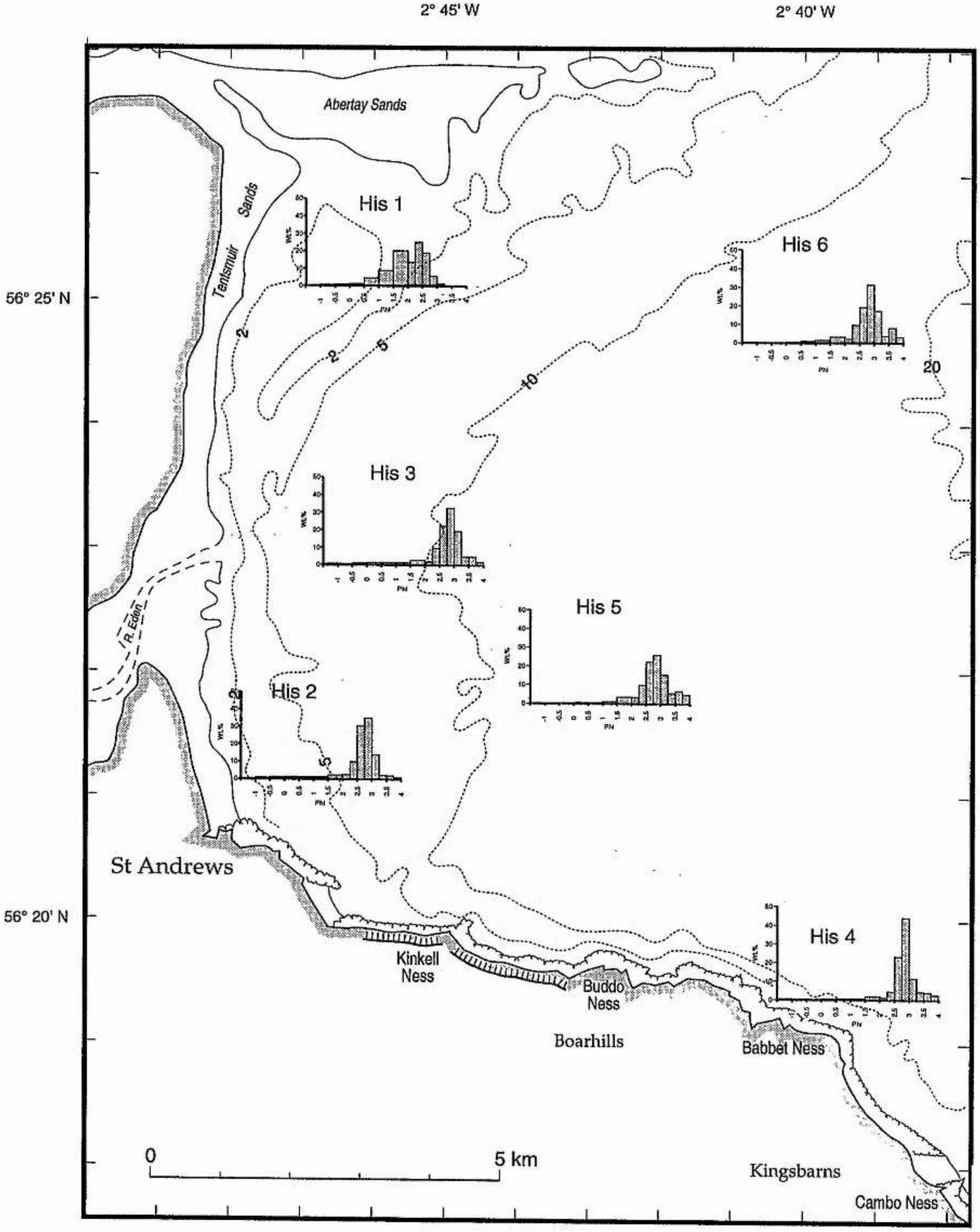


Fig. 3.9 Mean grain size variation across St Andrews Bay.

The sediment samples from St Andrews Bay show distributions skewed towards finer material although samples taken from shallow water (less than 10m depth) show coarse skewness (e.g. samples 94, 95, 97, 98, 99, 100, 101, 102, 103 and 105). Positively skewed sediments in shallow water (samples 24, 25, 28 and 29) could be the result of either the addition of fine material to the deposit, to the selective removal of coarse grains (Pethick, 1984) or to the lack of coarse material available in this location. The samples taken from deeper water (greater than 10m) in the southern half of the bay or farther offshore (Fig. 3.6) show fine skewness (e.g. samples 67, 68, 69, 70 ,72 , 73 , 74, 75, 76, 77, 78 and 79). Negatively skewed samples in a similar area (e.g. sample 71) may be due to the presence of shell debris which gives rise to coarse grained populations in the sediment. This is supported by visual observation of the collected sediment.

The standard deviation of the bed sediment in St Andrews Bay shows a range from very well sorted to well sorted (Fig. 3.5). This pattern of sorting tends towards moderately well sorted material on approaching the Abertay Sands and, in particular, where the water depth is shallow (e. g. samples 16, 17 and 103). The coarsest sediment sample is 103 which has a mean size of 1.27 Ø. It is a moderately well sorted sample with negative skewness.

3.9 Discussion

The sediments of St Andrews Bay show a narrow range of mean size, between fine sand in most parts of the bay and very fine sand found in a few isolated locations, such as north-east of Kinkell Ness and also in the central part of the bay. In most cases, the distribution curves in the bay are strongly unimodal. Pettijohn (1957) suggested that log-normal distributions would result from a single phase of deposition. In this case a simple normal distribution might be expected to result. This suggests that the bed sediments in St Andrews Bay aremore or less adjusted to their physical environment (Alveirinho et al., 1990). Several reasons may be suggested for the small deviations from log- normality in the St Andrews Bay sediments. Deviation might arise from the fact that there is a limited supply of available grain sizes in the sand fraction or it may result from the mixing of sub-populations of sediment with differing modes (Curray, 1956 ; Visher, 1969 and Walton et al.,

1980) which, in turn, would reflect varying energy conditions. Friedman (1961) stated that the textural parameters of sands reflect the mode of transportation and the energy of the transporting medium. However, this may not be the case in St Andrews Bay where Green (1974) noted that topographic and energy variations are not always distinguishable by changes in sediment size. Accordingly, any relationship between the statistical parameters and the depositional environment will not be very sharp or distinct.

Many workers have tried to interpret the geological meaning of skewness. Folk and Ward (1957) related skewness to the mixing of two populations, but Friedman (1961, 1962 and 1967) explained that the variation in the sign of skewness is due to the varying energy conditions of sedimentary environments. Thus, it is commonly believed that negative skewness represents an erosional environment, whereas positive skewness represents a depositional environment. The inclusive graphic skewness distribution of the bottom sediments in St Andrews Bay (Fig. 3.6) shows a mixture of symmetrical, negatively and positively skewed samples. Negatively skewed sediments mainly occur near the shore or close to the Abertay Sands and probably reflect the action of waves in these shallow areas. This arises because finer grains are selectively winnowed by constant wave action, leaving a tail of coarser grains (Leeder, 1982). Some samples collected from nearshore sites show positive skewness (samples 24, 28 and 112). This may relate to the fact that the sampling site is very close to areas of tidal flat where accumulation of fine particles takes place or, alternatively, bedforms such as bars may provide a shelter from wave action.

Positive skewness characterises the areas farther eastward away from the shore. The sea bed in these areas is less affected by wave action, principally due to water depth rather than the wave climate, which is common to the area. The presence of negative skewness in some sediment samples from farther offshore, such as samples 80 , 81 and 82, results from the proximity of the sites to the relatively coarse deposits of the Abertay Sands, whereas the negative skewness of sample 71 in the centre of the bay is believed to be due to the presence of shell debris.

In general, the area is characterised by the presence of very well sorted to well sorted sediments although some are no more than moderately well sorted. Inman (1949) suggested that the best sorting occurs with a mean diameter of 0.18 mm (2.48 Ø). However, mean diameters present in the St Andrews Bay sediments are generally smaller than this size. In spite of that, about half of the samples are very well sorted. The reasons for this high degree of sorting are not immediately evident but may be due to the shallow depth of water which permits hydraulic reworking of the bottom sediments by wave activity, thereby progressively increasing the degree of sorting. Sediments from offshore support Inman's (1949) conclusion that samples with a mean grain size diameter which is smaller than 0.18 mm show a tendency toward being poorly sorted. Negatively skewed samples from close to the shore are very well sorted, which is contrary to Inman's observations.

The little variation in the value of inclusive graphic standard deviation may be due to the limited range of sand particle sizes available in St Andrews Bay, so no large variations in this parameter are to be expected.

Usually the grain size distributions obtained give an indication of the hydrodynamic conditions characterising the depositional environment of the sampling sites. The geographical location of St Andrews Bay, situated between two headlands, is such that wave energy will be convergent on the headlands and divergent over much of the rest of the area. Thus, St Andrews Bay itself is characterised by a lower energy environment than the headlands. The grains size distribution data indicate that most parts of St Andrews Bay are characterised by more or less constant hydrodynamic energy conditions. The exception is at the area very close to the Abertay Sands where it has been shown that there is more variation in mean size compared with other areas.

Tidal currents flow through the area providing a second dynamic mechanism for sediment transport. Previous studies have shown that tidal currents in St Andrews Bay are weak and are alone unable to exert sediment transport in the area. Sediment transport in St Andrews Bay is attributed to the combined effects of waves and tidal current (Jarvis and Riley, 1987).

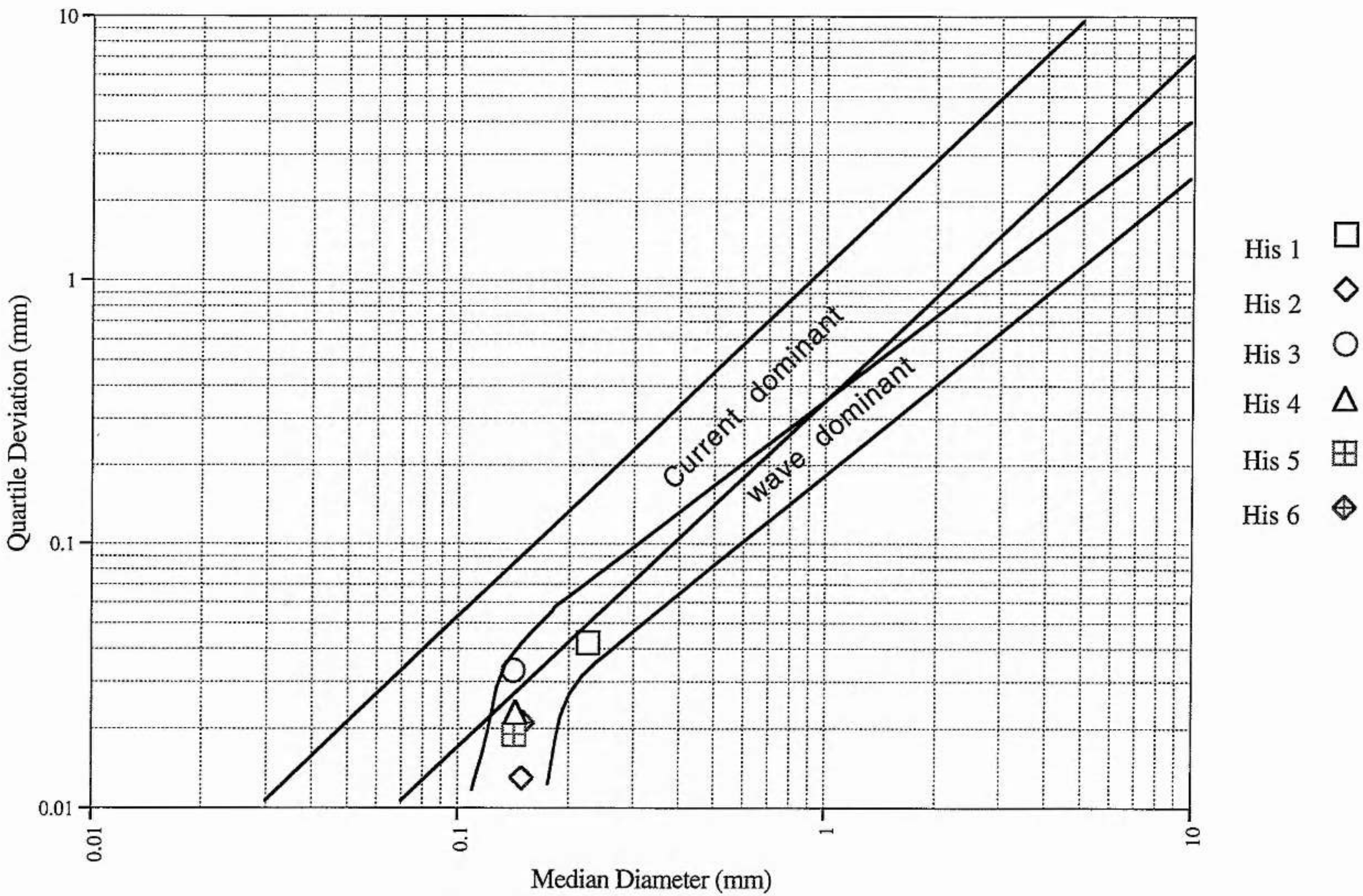


Fig. 3.10 Quartile Deviation - Median Diameter plots of bed sediments from St Andrews Bay (Samples grouped according to Fig. 3.9). (Envelope trends from Buller and McManus 1972).

A further use of bivariate plots, using the QDa - Md method (Appendix 5), provides a means of assessing the relative importance of these two sets of dynamics within St Andrews Bay. Without exception, all of the analyses plot within fields defined as characterising sediments from wave-dominant or current-dominant environments. Within the general plots of QDa versus Md, most of the sediments of St Andrews Bay fall within the wave dominant envelope but some of them occupy positions in the region of overlap between the wave and current dominant fields (Fig. 3.8). In order to distinguish between the sediments which plot in these two fields, and also to identify the location of these sediments, a second plot of QDa versus Md, based on the previous division of the samples according to frequency distribution, has been constructed. Fig. 3.10 shows that His1, His 2, His 4, His 5 and His 6, which incorporate 76% of the samples, are characterised as being mainly wave dominant and are generally lacking in substantial current dynamics. However, the His 3 samples fall within the region of overlap between the wave dominant and current dominant trend envelopes.

CHAPTER IV

Water Movement and Sediment transport in St Andrews Bay

Introduction

The motion of sediment in the marine environment is linked closely to the hydraulic phenomena, so that an understanding of the principles governing flow is of paramount importance in understanding the movement of sediments. Sediment transport and changes in bed morphology are related strongly to the forces which produce water movement and turbulence such as tides, waves and wind. Tidal currents in oceanic and coastal waters are caused by gravitational forces on the water bodies induced by systematic variations in the relative positions of the earth, moon and sun . The other kind of water level fluctuations which give rise to horizontal currents are due to meteorological variations.

This chapter of the thesis will be divided into two sections. The first section describes the pattern of tidal currents in St Andrews Bay under different weather conditions, and examines the role of wind as a primary factor in changing the character of tidal currents. The second section uses current meter data in order to predict the pattern of potential sediment transport in different locations in the bay.

The data used in both sections will be the same, but more data processing is needed for section one. For section two the original data and information about mean grain size of bed material are needed (Chapter 3) in order to be able to apply a transfer function to the current meter data in order to estimate sediment movement.

Water movement

St Andrews Bay on the east coast of Scotland (Fig. 1.1) communicates with the adjacent North Sea through a relatively wide (27 km) zone where the local water depth is between 10 to 15 m. The principal freshwater sources are the Rivers Tay and Eden. Considerable mixing of fresh and saline waters take place in the estuaries of the Tay and Eden so that the discharges into the bay are well mixed and few spatial or vertical differences of salinity occur in the bay to cause non-tidal water movements. Tides in the area are mainly semidiurnal with a mean spring tidal range of 4.7 m and a mean neap tidal range of 3.2 m (Jarvis and Riley, 1987). This gives an average tidal range of approximately 4.0 m so, on the basis of the classification suggested by Davies and Hayes (1984), the coasts of the study area are low-macrotidal coast (Table. 4.1).

Tides	Coast type	Tidal range
Micro tidal	Micro tidal coast	< 1m
Low - Meso tidal	Meso tidal coast	1 - 2m
High - Meso tidal	Meso tidal coast	2 - 3.5m
Low - Macro tidal	Macro tidal coast	3.5 - 5m
Macro tidal	Macro tidal coast	>5m

Table 4.1 Classification of Tidal Coast after Davies and Hayes (1984).

In the North Sea the, tidal oscillations are partly determined by the dimensions of the North Sea basin and partly by the progressive semi-diurnal tides entering from the Atlantic. As a result, a standing wave with three nodes tends to develop in the North Sea. The water is deflected by the Coriolis force and forms three amphidromic systems (Pethick, 1984). The most northerly of these waves, with its amphidromic centre some 60km off the Southern Norwegian coast, near Bergen, generates an anticlockwise wave (Charlton et al., 1975) which affects the Tayside and Fife coastal regions.

Although water movement in the North Sea is influenced greatly by the Coriolis force, in small embayments such as St Andrews Bay, the effect of the earth's rotation is insignificant, but the frictional effect of the sea bed and the constraining effect of the land masses upon the currents are important. The presence of prominent headlands and estuary mouths has an important influence on waves and tides, and consequently on sediment dispersal and deposition, and ultimately on shoreline evolution.

Where flow is unrestricted by coastlines and variable bathymetry, the tidal currents are observed to rotate during the tidal cycle. In such situations, a time-series of current meter measurements, plotted as a progressive vector diagram by drawing vectors representing the current speed and direction continuously at equal time intervals, would show a closed ellipse. Close to coastlines, because flow is restricted in the direction normal to the shore, the ellipses become much more rectilinear with the long axis parallel to the coast. However, in many coastal situations, the simple tidal ellipse is modified due to the presence of headlands, embayments and sea beds of unconsolidated sediments. Tidal currents and wave action create variable bathymetry and flood and ebb currents often follow slightly different pathways. At a given location in such a situation one tide will be slightly stronger than the opposite tide and a progressive vector plot will show an open ellipse with the gap in the ellipse representing the residual flow over a tidal cycle at that point.

Headlands also have a tendency to protrude into tidal flows creating tidal eddies (Robinson, 1983). The impact of headlands and bays which define St Andrews Bay are such that they induce a tidal eddy in the bay in a direction counter to the driving force in the North Sea (Charlton, 1980). The pattern of a clockwise movement on the flood tide and anticlockwise movement on the ebb is modified near the mouths of the Tay and Eden as water floods into and ebbs from these estuaries.

In addition to tidally-induced water movement shallow coastal areas like St Andrews Bay are strongly influenced by meteorological forces (Gacic et al., 1987). The effect of wind on ocean currents is well known from the work of Ekman (1905). This work established the basic relationships between wind speed and the speed of surface drift as well as establishing the relationship between wind speed and the variation of the water speed and direction with depth (The Ekman Spiral). According to Heathershaw and Lees (1980) meteorological factors may affect the circulation in two ways, either directly by the application of a wind stress to the sea surface, leading to a surface drift, or indirectly from changes in sea level which may occur as a result of either the wind piling up water against a coast or as a result of changes in atmospheric pressure. Further, Stoke's wave theory predicts that a mass transport current is produced by waves moving forward across the sea which can also produce an additional current that is to be added to tidal and wind driven currents (Carter 1988).

The most significant wind stress effect on sea water movement for long-term sediment movement is the wind-induced drift caused by momentum transfer from the wind to the sea. Both the drift velocity and the depth of the surface layer affected by the wind are dependent upon wind strength (Heathershaw and Hammond, 1979). Estimates of the near surface wind driven current in shallow water (20 m) have been calculated for Swansea Bay (Table. 4.2) and can be used to illustrate the current structure in the surface layers of St Andrews Bay due to wind shear (a slight error will exist because the Coriolis parameter is different for different latitudes).

Wind speed $m\ s^{-1}$	h/D $h = 20$ m	U_S $cm\ s^{-1}$	V_S $cm\ s^{-1}$
2	7.61	0.42	0.43
4	2.69	1.73	1.79
8	1.30	3.35	5.72
12	0.28	3.45	9.41
16	0.20	2.60	11.25
20	0.017	1.98	12.44

Table 4.2 Estimates of near surface wind driven currents as a function of wind speed, Swansea Bay (after Heathershaw and Hammond 1979).

Where h is the water depth, D is the depth of frictional influence, U_S is the component of the surface wind driven current normal to the wind direction and V_S is the component of the surface wind driven current parallel to the wind direction.

Heathershaw and Hammond (1979) show that the depth of frictional influence does not affect the whole depth until a wind speed of 8 to 10 m s^{-1} is reached . It is noted that, at low wind speeds, the direction of the surface layer (given by the two components) is at about 45° to the wind direction but as the wind speed increases the surface layers become more aligned in the direction of the wind (Table 4.3).

Wind stress also causes slope currents because water driven towards a shore may lead to the piling up of water along the shore. The wind creates a sea surface gradient which will generate a horizontal current system comprising an onshore movement of water at the surface and a return flow near the bed. Different circulation patterns will be established if the wind is blowing at an angle to the coast or offshore. The strength of the currents can be calculated if the gradient of the sea surface is known. Values have been computed for Swansea Bay and are shown below in (Table. 4.3).

Wind speed m s^{-1}	U_b cm s^{-1}	V_b cm s^{-1}
2	-0.25	-0.12
4	-1.02	-0.51
8	-2.07	-2.51
12	-2.15	-4.49
16	-1.62	-5.53
20	-1.24	-6.17

Table 4.3 Estimates of maximum slope currents as a function of set up resulting from wind action in Swansea Bay (after Heathershaw and Hammond 1979).

U_b is the component of the bottom wind driven current normal to the wind direction and V_b is the component of the bottom wind driven current parallel to the wind direction.

The data thus suggest that near bottom slope currents for moderate wind speeds may be up to 6 cm s^{-1} and become aligned opposite to the wind direction as wind speed increases.

In an infinitely deep ocean the mass transport of water associated with wave activity predicted by Stokes theory (1880) suggests a steady forward drift which decreases in strength with depth. In shallow water one would, however, expect a near-bed return flow to exist in order to maintain continuity and prevent water piling up at the shoreline. A more complete analysis of wave motion by Longuet-Higgins (1953) predicts forward movement of water with waves at the surface and near the bed with a return flow taking place at mid depth. Drift velocities for a water depth of 20 m are shown in (Table 4.4).

Depth (m)	Current cm s^{-1} wave height H= 0.5 m	Current cm s^{-1} wave height H= 1 m	Current cm s^{-1} wave height H= 2 m
0	0.22	0.88	3.53
4	-0.12	-0.46	-1.84
8	-0.27	-1.08	-4.31
12	-0.27	-1.09	-4.35
16	-0.14	-0.56	-2.26
20	0.12	0.46	1.85

Water depth = 20 m   Wave period = 8 seconds

Table 4.4 Drift currents for different wave heights and depths in Swansea Bay (after Heathershaw and Hammond 1979).

The data show that the drift is in the forward direction at both the surface and the bed with a return flow at mid-depth, but the maximum rates for moderate (1 m) waves are only of the order 1 cm s^{-1}. It is noted that, under larger waves, the table predicts a larger return flow than forward flow which thus does not show continuity of water movement. This probably follows from the fact that the theory is strictly valid only for small amplitude waves. Nevertheless the order of magnitude of the currents is probably correct.

In Swansea Bay the prevailing wind direction (SW) is onshore whereas at St Andrews it is offshore and the prevailing wave climate might be expected to be less extreme than off the Welsh coast. However, water depths are similar and it is thus reasonable to apply the data from the Swansea Bay investigation to St Andrews Bay. Accordingly one can predict that meteorological forcing might contribute shear currents of up to 10 cm s^{-1}, slope currents of the order 6 cm s^{-1} and wave drift currents of the order 1 cm s^{-1} to the non-tidal circulation of St Andrews Bay.

4.1 Previous work

Many studies have been carried out in order to investigate the water circulation in the coastal zone of the British Isles and North America (Stanley and Swift, 1976; Swift et al., 1972; Lees and Heathershaw, 1981; Heathershaw, 1981). In particular, the Institute of Oceanographic Sciences carried out detailed investigations of water and sediment movement in Swansea Bay and in the Dunwich banks area of East Anglia. These studies provide comparative investigations to the work in St Andrews Bay.

Long-term current meter records from both locations were analysed to determine tidal residuals and the impact of tidal forcing on the pattern of residual flow. Off the coast of East Anglia the tidal currents were observed to be essentially rectilinear with maximum rates of over 100 cm s^{-1} at 2 m above the bed, whereas in Swansea Bay the tide was much more rotary, particularly in the centre of the bay with maximum rates of 50 to 70 cm s^{-1} at 2 m above the bed. The observed rectilinear nature of the tidal flow at Dunwich accords with the presence of linear sand banks that restrict any component of flow normal to the shore. Swansea Bay is a semicircular embayment more conducive to rotary movement especially at its centre.

Many sites in both Swansea Bay and Dunwich showed near-linear residual flows. However, after filtering out tidal motion, quite distinctive changes in residual direction, apparently due to meteorological forcing, were noted. At a site 2 km offshore in Swansea Bay and in 10 m of water winds from the SE blowing alongshore enhanced the near-bed residual flow to the NW. However, winds from the SW blowing normal to the shore reversed the flow, causing a circulation to the SE. Such a pattern was interpreted as being due to water driven NE towards the head of Swansea Bay by the SW winds which produced a compensating return flow near the bed.

Confirmation of the differing behaviour of near bottom and mid-depth water was found from the analysis of data from current meters at different elevations above the bed at 4 locations in Swansea Bay. It was observed that the near-bed mean circulation varied with location and up to 180° from the mid depth residuals (Fig. 4.1).

Similar results were obtained from the current meters deployed at Dunwich. Residual flows of up to 13 cm s^{-1} in a S to SW direction were completely reversed and persisted for 6 or 7 days during strong southerly winds. The alongshore component of residual flow was significantly correlated, at the 5% level, with the alongshore component of wind stress but the onshore-offshore component for some records appeared to correlate with the alongshore component, while other records correlated with the onshore- offshore wind stress.

A number of studies have been carried out in St Andrews Bay in order to examine the water circulation and to predict sediment transport due to the effects of astronomical or meteorological forces.

Green (1974) and Charlton (1980) have studied the coastal tidal water circulation in St Andrews Bay and the mouth of the Tay Estuary. Lagrangian measurements of the tidal flow, made by Green (1974), in order to ascertain the general pattern of water flow, indicated that a large eddy system exists within St Andrews Bay. This eddy is characterised by a clockwise rotation of flow on the flood tide and an anti-clockwise rotation flow on the ebb tide. Green also reported on sediment movement in the inner part of the Abertay Sands. Here,

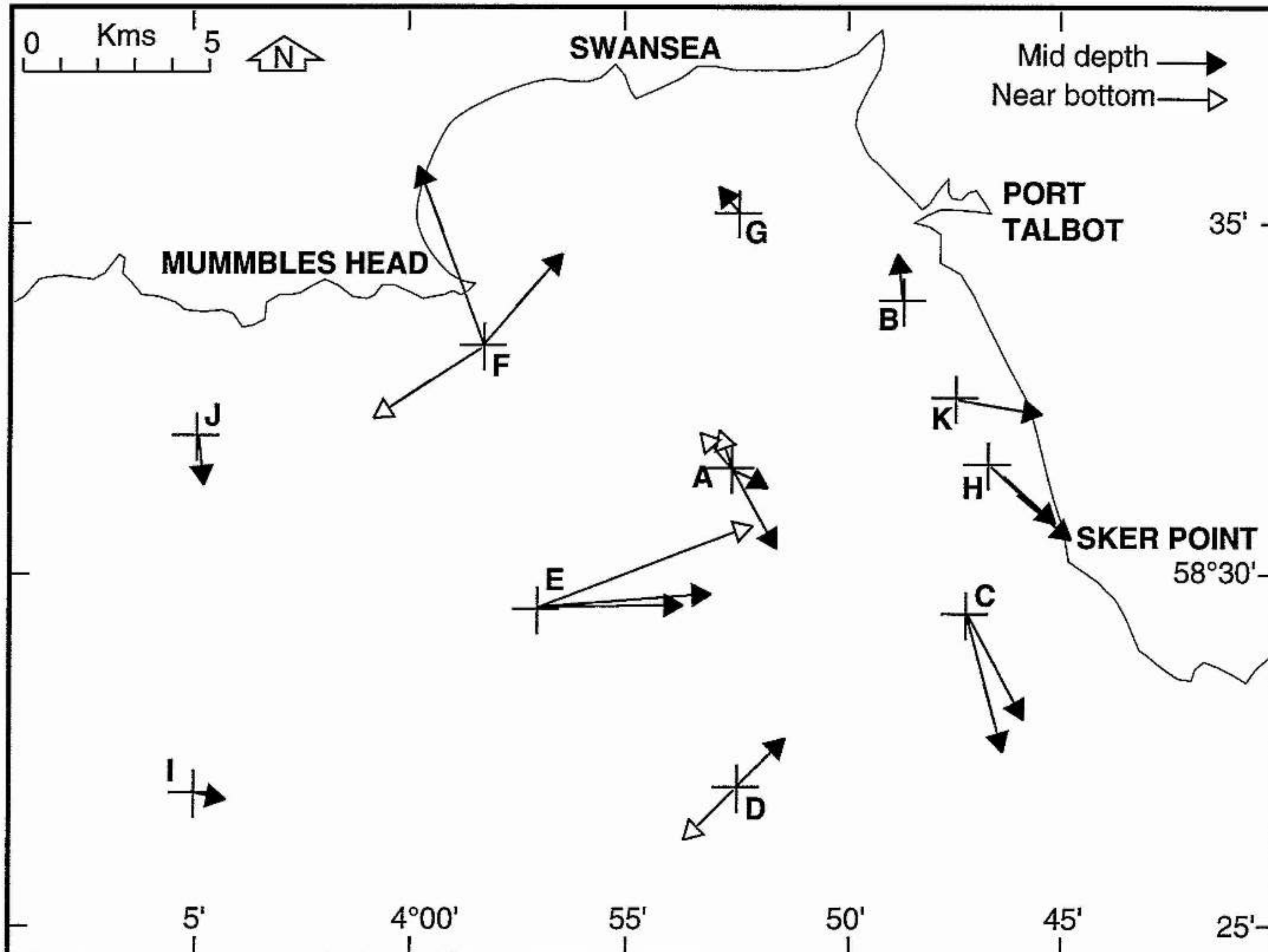


Fig. 4.1 Tidally induced residuals in Swansea Bay (After Heathershaw and Hammond, 1979).

fluorescent sand tracer experiments and the study of residual currents suggested that sand from St Andrews Bay was carried north-westwards across the spit complex and into the channel. It was flushed seawards by the ebb tide in the main channel with little evidence for net accumulation on the spit (Green 1974). Diurnal rates of sand grain migration in the tidal channels of the bar area (Pool channel) were of the order 10^3 m/day .

Charlton (1980), confirmed the results of Green (1974), and concluded that the inshore flood current runs south-westward from Arbroath, turning westwards into the Tay Estuary and St Andrews Bay, forming a clockwise eddy between Fife Ness and the Abertay Sands, and the ebb tide runs eastwards out of the Tay Estuary, and then north-eastwards towards Arbroath (Fig. 4.2).

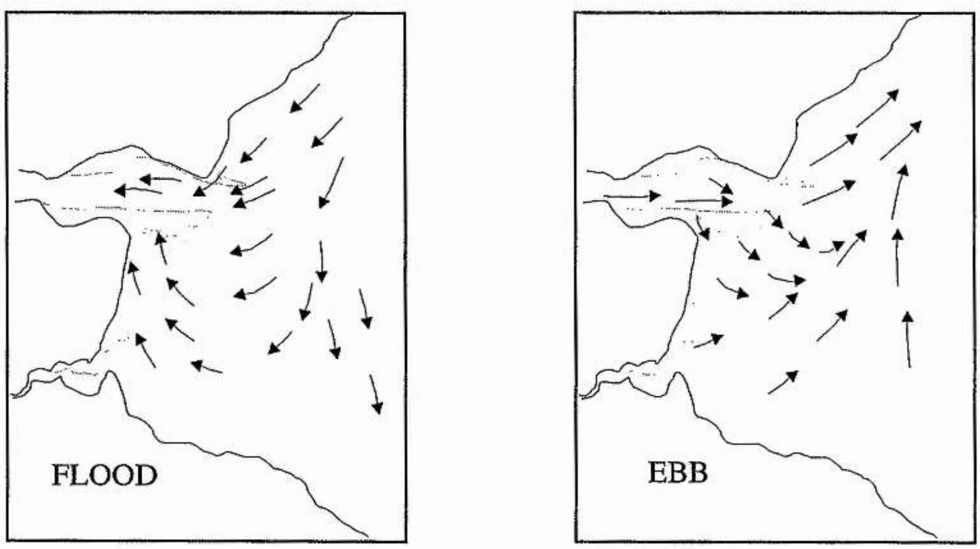


Fig. 4.2 Mid - flood and mid - ebb tidal current patterns in St Andrews Bay
(after Ferentinos & Mc Manus, 1981).

The pattern of wave-controlled longshore movements at the head of St Andrews Bay has been reported by Ferentinos and McManus (1981). They related the morphological evolution of the nearshore region due to nearshore hydrodynamics in St Andrews Bay. They suggested that nearshore hydrodynamics and sedimentary processes in such an area are mainly controlled by wave-induced longshore currents and by the coastal tidal water circulation. They further linked the morphological evolution of Pilmour Spit to the north - west moving longshore current which extends from Fife Ness to Tentsmuir point. They also suggest that in St Andrews Bay the nearshore hydrodynamics are complicated by the presence of the major Tay Estuary and its associated circulation patterns. They suggest that the nearshore hydrodynamic circulation will be more complicated by the presence of the rivers Tay and Eden.

Sarrikostis (1986) studied the potential longshore sediment transport patterns between Montrose and Fife Ness using computer simulation techniques based on wave refraction and wave climate data. He concluded that, under most wave conditions, the potential transport between Fife Ness and the Eden Estuary is northwards. North of the Tay Estuary the general littoral drift is not stable and Tentsmuir Beach receives least wave energy under most conditions examined.

The sedimentological and morphological dynamics of the Tay Estuary, Eden Estuary and St Andrews Bay itself have been explored by several workers (e.g. Green, 1974; Eastwood, 1977; McManus et. al., 1980; Al-Mansi, 1986; Jarvis and Riley, 1987; Jarvis, 1989). Jarvis and Riley (1987) studied the sediment transport from a fixed platform at an intertidal location in the mouth of Eden Estuary. The effects of wave activity on water movement and suspended sediment concentration were investigated. Bed load transport was determined by measuring the rate of migration of particular bed features (megaripples) and by measuring volumetric changes of sand from erosion pin surveys. However, it was noted that movement of water and sediment normal to the shore was complex because the tidal currents and oscillatory motion due to wave action were not parallel to one another. An empirical model of the interaction between waves and tidal currents for the Eden mouth was constructed from near-bed observations of water movement and pressure fluctuations made at the research platform. A relationship between sediment concentrations and wave activity was derived. In a subsequent study by Boalch (1988) it was found the onshore wave power was significantly higher, often by a factor of 4, than the longshore wave power it thus follows that the offshore-onshore sediment exchange exerts a great influence on coastal geomorphology.

The previous studies at the head of St Andrews Bay complement each other and indicate that there is an area of persistent accumulation at Tentsmuir Point and on the tidal flats close to the River Eden where as area of erosion exists at the southern part of Tentsmuir Beach. Jarvis and Riley (1987) concluded that there is an accumulation in the tidal flat area since there are sand bars migrating onshore which increase the volume of sand in this area. Sarrikostis, (1986) also arrived at the same conclusion by using the fluorescent sand tracer technique. The erosion which occur along a limited stretch of the southern Tentsmuir Beach appears to be associated with northwards longshore currents. The longshore drift induced by these currents provides the sands which are accumulating at Tentsmuir Point. It is believed that the presence of the Abertay Sands will act as a protecting shield prohibiting waves from the east affecting this area of accretion.

4.2 Data collection

Tidal currents have been measured from several locations (Fig. 4.3) over periods of up to 40 days in order to determine near-bed residuals and the effects of meteorological forcing on water movement (Appendix 6). The self-recording current meters used in this study were designed and constructed in the University of St Andrews using commercial compass and propeller units. The locations at which current measurements were made were chosen in order to provide data from as wide a range of locations as possible. They were deployed using a U mooring system, with 30 kg concrete anchors and a 20 m ground line (Fig. 4.4). Velocity and direction were recorded every 15 minutes and after recovery the data were read from the internal memories on to a computer and propeller revolutions converted to a mean velocity each 15 minutes. All readings were taken at 1m above the bed, and these data have been used to achieve the second goal of predicting sediment transport in St Andrews Bay. The station positions were fixed with a Decca navigation system.

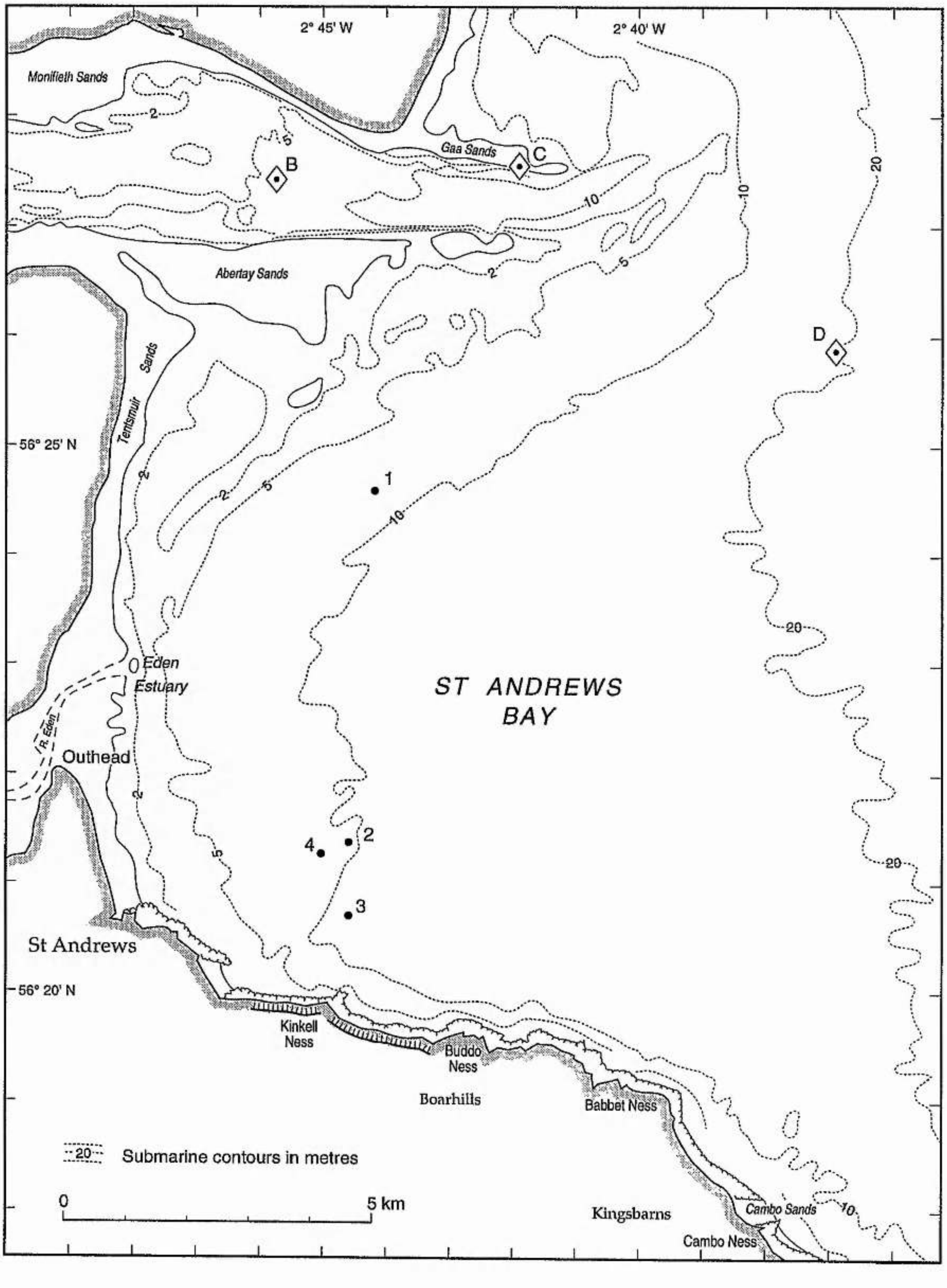


Fig. 4.3 Recording current meter locations.

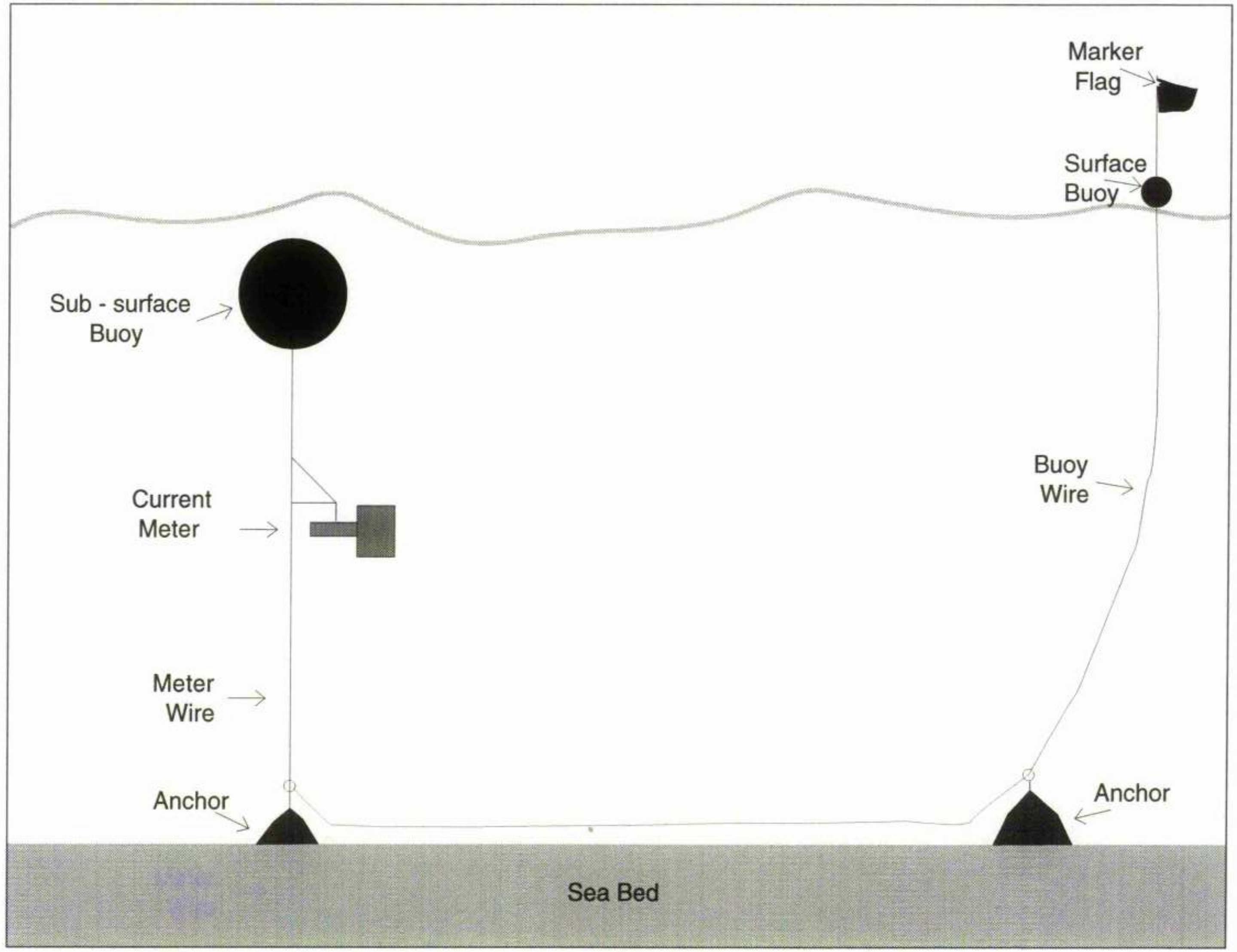


Fig. 4.4 Diagram of current meter mooring of type used in St Andrews Bay.

Hourly wind speed and direction data from the Leuchars meteorological station for the period from May 1991 to July 1992 have also been used in order to assess the effect of wind stress on the bay and the resultant pattern of water movement recorded on the current meters.

4.3 Analysis of data and progressive vector diagrams

Water movement in St Andrews Bay comprises several parts. If one tracks a parcel of water over a complete tidal cycle it will not in general return to its starting point but will show a residual displacement from its initial point. This residual may be a function of the non-closure of the tidal ellipse due to shallow water effects or it may be due to meteorological forcing. The residual water movement can be presented graphically. In order to investigate residual water movement it is necessary to minimise the tidal current action and thus achieve a net transport direction based only on meteorological forces.

The mathematical calibrations and data processing were carried out using the Microsoft Excel program on the Macintosh computer. The current velocity was broken into two components:- one along the east and the other along the north

$$u = v.\sin\theta \quad \text{(along the x-axis)} \qquad (4.1)$$

$$v = v.\cos\theta \quad \text{(along the y-axis)} \qquad (4.2)$$

The u and v components of velocity were then passed through a 96 quarter hour mean filter. 96 values of the u components were added and then divided by 96, giving average values of u. Similarly 96 values of the v components were added and then divided by 96 to give average values of v. The filtered u and v components were then used to obtain speed and direction.

$$V = \sqrt{u^2 + v^2} \qquad (4.3)$$

$$\theta = \tan^{-1}(u / v) \qquad (4.4)$$

From the filtered values of the u, and v components, speed and direction was obtained, as explained above. From the daily residuals of speed and direction, progressive vector diagrams were plotted. The lengths of the vector represent the speed and direction for one day. From the tip of the vector another vector was drawn depending on the speed and direction on the second day, and so on. The progressive vector diagrams thus display the daily residual flow.

Hourly progressive vector diagram have also been drawn for single tidal cycles during calm weather so as to minimise contamination of data by waves or wind. These data will thus include only the tidal currents, so that the pattern of a typical tidal flow at each station can be described.

The data from the recording stations are presented both as progressive vector plots and also as rose diagrams in order to illustrate different aspects of the flow in St Andrews Bay. Additional data have been drawn from the Admiralty Hydrographic Charts in order to supplement the data covered in this survey. It should be noted that current speed and direction on the Admiralty Chart (1978) were measured near the surface. Details of these stations are shown in Fig. 4.3 and Table 4.5.

Observations are presented on the rose diagrams as hourly values for 6 hours before and after high water as flood commenced.

Meteorological data were supplied on a computer disc and read into an Excel Spreadsheet for processing. Wind speed and direction frequencies were calculated for the period of current meter data collection. In addition, the current meter and wind data were plotted against time to determine whether or not there is any relationship between their trend lines.

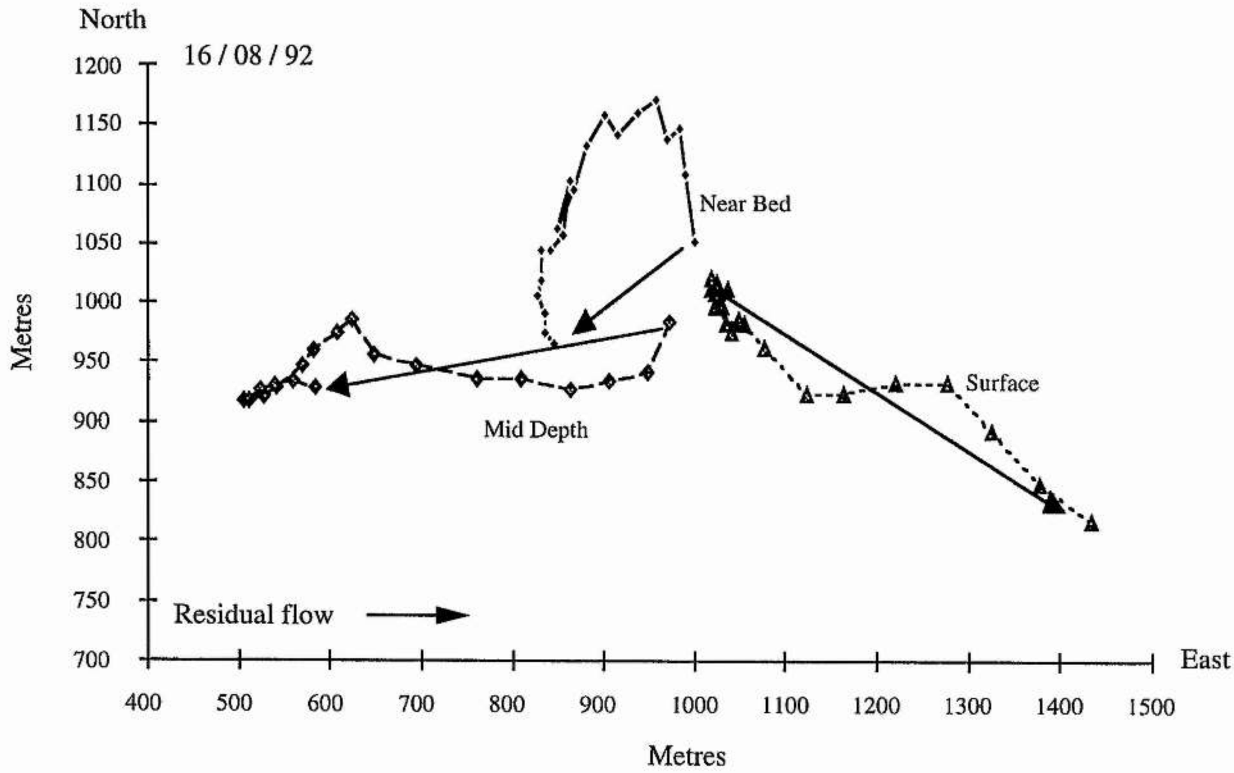
Hourly progressive vector diagram has also been drawn for a single tidal cycle during strong winds to illustrate the effect of meteorological forces on the behaviour of the tidal current (Fig. 4.11). The methodology of this study follows that carried out in Swansea Bay by Heathershaw and Hammond (1979) because water depths in St Andrews Bay are similar to those of Swansea Bay. In particular, this study indicated that the upper 20 m of the water column is only influenced by wind speeds greater than 8 m s^{-1}.

Position No.	Instrument	Date	Recording period
1	R C M	27 / 7 / 1991	27/7 - 1/9/1991
2	R C M	13 / 9 / 1991	13/920/10/1991
3	R C M	3 / 7 / 1992	3/7 - 26/7/1992
3	DRCM	16 / 8 / 1992	Only 13 Hours
3	DRCM	20 / 8 / 1992	Only 13 Hours
4	R C M	3 / 7 / 1992	3/7 - 28/7/1992
B	DRCM	Chart, 1978	Only 13 Hours
C	DRCM	Chart, 1978	Only 13 Hours
D	DRCM	Chart, 1978	Only 13 Hours

Table. 4.5 Details of fixed current recording stations.
1, 2, 3 and 4 are researcher data; B, C and D are data from the tidal diamonds of Admiralty Hydrographic Charts (1978).

4.4 Results

Before discussing the results of the long-term recording current meter measurements it is informative to consider the results obtained from a directional recording current meter (DRCM) deployed on station (No. 3) about 2km offshore. Data were collected here in connection with a proposed new sewer outfall on a spring and a neap tide (16.8.1992 and 20.8.1992). Although measurements were recorded for less than one tidal cycle on each occasion it is of value in the present study because measurements were recorded at different depth namely surface ,mid-depth and near-bed together with surface wind observations. At



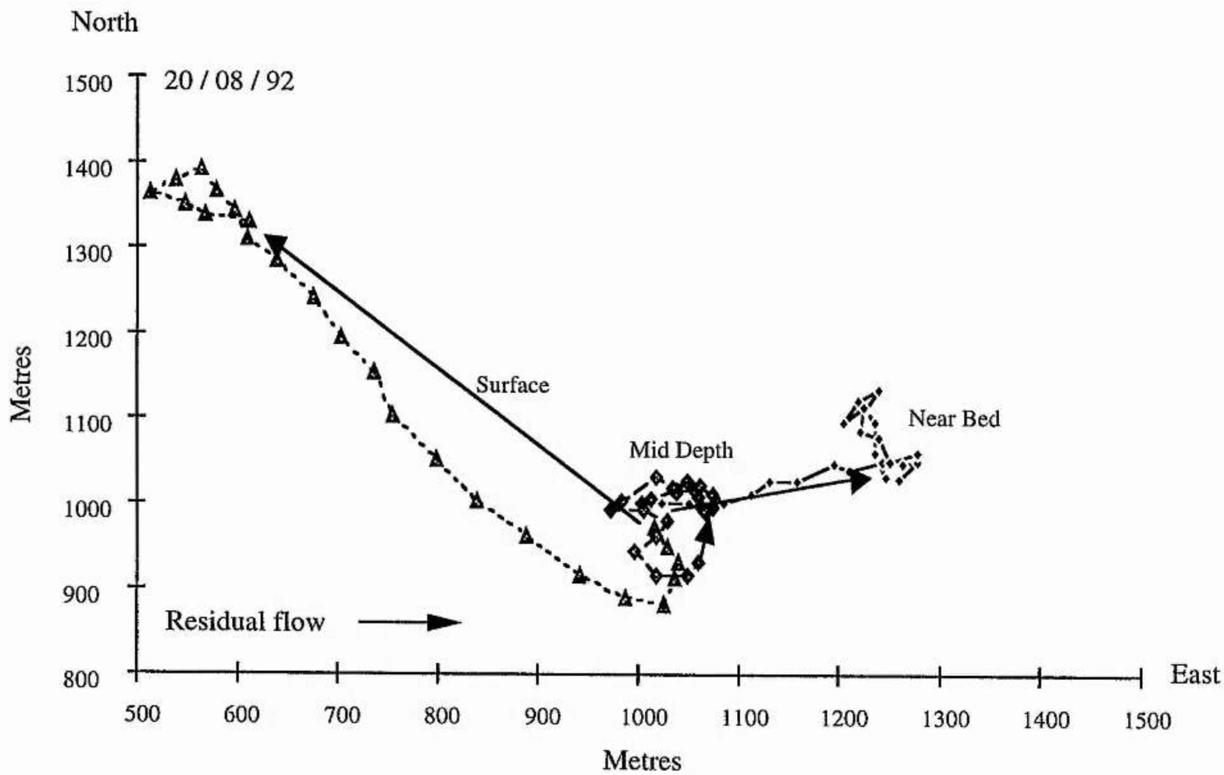


Fig. 4.5 Progressive vector plots for one tidal cycle at different water depths for Station 3, vectors plotted at 30 minute intervals.

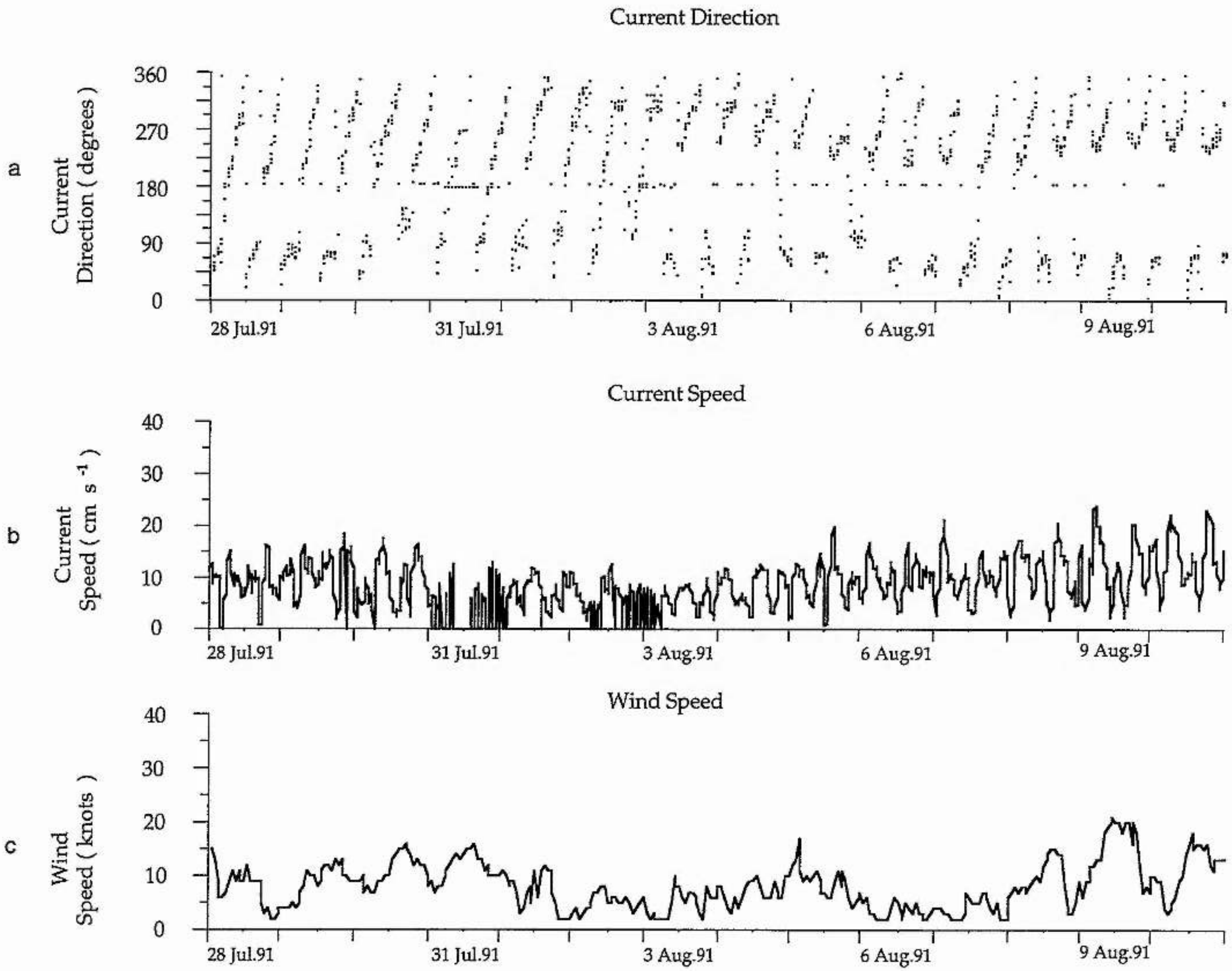
each height, current speed and current direction were determined at intervals of 0.5 hours. The results indicate that the residual flow, indicated by vector plots, at the recorded elevations behave very differently from each other (Fig. 4.5). They do not follow a similarly consistent pattern of relationship with another factors such as tide or wind.

The factors responsible for the water movement are tidal and non-tidal components. At this location a non-tidal force arising from variation in density is unlikely because salinity determinations revealed that the water column was well mixed. The wind record for the two days also showed that wind speed was insufficient to have a major effect on water column. An explanation for the behaviour of the water masses at different depths at this site is difficult but such possible variations may also exist at the other recording sites in the bay for which only single elevation data are available.

Station 1

Data from Station 1, which is situated in 8 m of water at low tide and is to the south of the Abertay Sands, covers the period from 28/7/91 - 9/8/91. The time-series plot of current meter direction (Fig. 4.6a), shows a systematic variation throughout approximately 14 tidal cycles. Moreover, the tidal current shows a simple pattern of variation during spring and neap tides with maximum rates of about 18 cm s^{-1} on spring tides and 10 cm s^{-1} on neap tides. The tide has no clear period of slack water during the tidal cycle (Fig. 4.6b). The wind record for the period of current meter data is shown in (Fig. 4.6c). This is quite variable but speeds often exceeded 8 m s^{-1} and thus might be expected to cause a wind induced drift for those periods of time. On 3rd August, 1991 the winds were much lighter and for this period of calm weather an hourly progressive vector diagram has been plotted in order to show the tidal flow at this station. The result shows a nearly closed ellipse (Fig. 4.10A) which indicates the rotary behaviour of tidal flow at Station 1. The rotary behaviour of the tidal flow for 3rd August, 1991 was also plotted as a rose diagram (Fig. 4.13). For this form of graphic presentation, the wind must be very light in order to show the systematic variation of the tidal flow motion. It is also necessary to use the tide table of the area in order to derive the

Fig.4.6 Time series of current direction (a), current speed (b) and wind speed (c) at station 1.



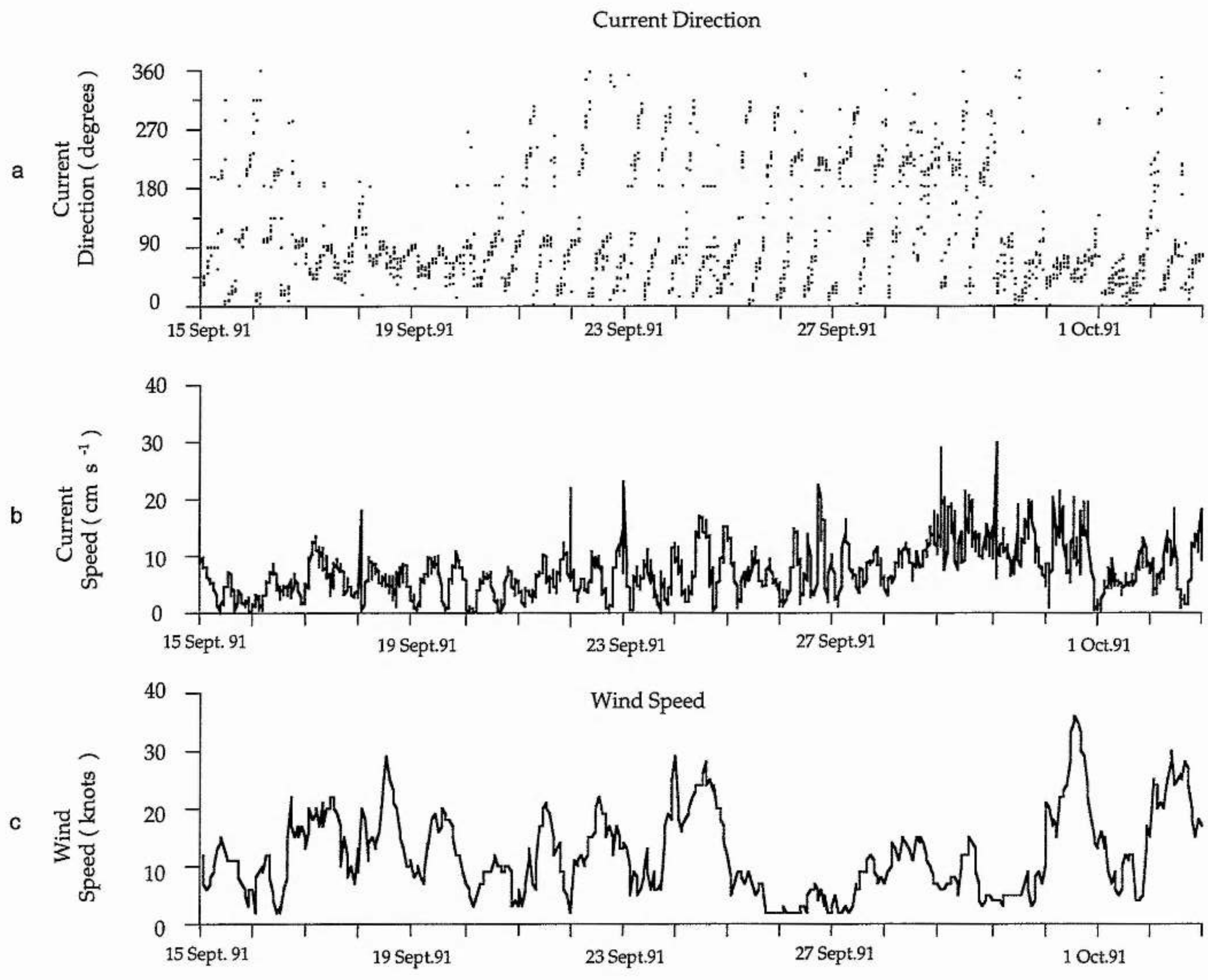
predicted time at which the flood tide begins. The current data for the day of 3.8.1991 on which the flood commenced at 08:45 hours has been smoothed to give hourly readings and the data have then been plotted as a rose diagram for one tidal cycle of 13 hours. It is clear that the tidal flow at Station 1, as flood commenced, started to rotate from the west towards the north-west, north, north-east then south-east, south and south- west and later back towards the west.

Winds and waves sometimes play an important role in changing the nature of the tidal ellipse. For example on 9th August, 1991 there is a grouping of current readings in the north-west quadrant, in the directional range between 270°-360° (Fig. 4.6a).The hourly progressive vector diagram plotted for this day (Fig. 4.11A) shows an open tidal ellipse with a north-west residual flow of water. Wind velocity was strong on this day reaching maximum values of 20 knots from south-west and west. It is clear that the direction of the bottom residual flow was neither parallel nor opposite to the wind direction. It could be possible that the grouping of a current readings in the 270° - 360° quadrant is attributed to another factor, such, as waves. A daily progressive vector diagram for the period between 28th July and the 1st October, 1991 shows residual movement towards the south-east (Fig. 4.12A).

Station 2

The record obtained from Station 2 (Fig.4.7a) covers the period from 15 September 1991 to 3rd October 1991. Purely tidal flow at this station was observed on 25th September, 1991 on which day the flood began at 08:15 hours. The results show a nearly-closed ellipse with a rotary behaviour (Fig. 4.10B). A rose diagram plotted for this day shows that the tidal flow rotates clockwise from south-east to north-east and later back towards the south-east (Fig.4.13). The effect of wind on changing the nature of the tidal ellipse can be clearly seen on 1st. October, 1991, where the directional readings are grouped in the north-east quadrant within the range 30° - 75° (Fig. 4.7a). Thus, the hourly progressive vector diagram plotted for this day (Fig. 4.11B) shows an open tidal ellipse with a strong north-east residual flow of water. Wind velocity was very strong on this day (Fig. 4.7c) reaching maximum values of 36 knots, predominantly from the south-west. After smoothing the data between 13th September

Fig.4.7 Time series of current direction (a), current speed (b) and wind speed (c) at station 2.



and 20 October, 1991 to give a daily progressive vector it is clear that a daily progressive vector (Fig. 4.12B) for 1st. October, 1991, shows a residual movement of water towards the north-east parallel to the wind direction. This is because the wind was strong enough to affect all of the water column, wind blowing from south-west with a speed of 36 knots.

Station 3

This station has data recorded between 4th July and 17th July 1992. To illustrate the behaviour of the tide at this station in calm weather data was taken from 6th July, 1992 and was plotted from the start of the flood tide at 12.00 hours. The results show a rotary behaviour of tidal flow (Fig. 4.10C). The rose diagram for this day shows tidal flow on the flood tide to commence from the south and then rotate clockwise through south-west and north-west, before flowing north east (Fig. 4.13). A daily progressive vector diagram for this station shows the resultant flow is onshore (Fig. 4.12C). There is a grouping of a directional current reading in the south-west quadrant in the range between 240° - 250° (Fig. 4.8a). It is clear from a daily progressive vector diagram plotted for the period from 3rd July and 27th July 1992 that July 11th, 1992, in particular shows residual flow towards the south-west (Fig. 4.12C). Interestingly, hourly progressive vectors for that day also show a near-linear movement of water to the south-west (Fig.4.11C1). This grouping occurred even though winds were light that day, and blew at less than the threshold value of 8 m s^{-1} (Fig. 4.8c). Another grouping of directional current data occurred in the south-west quadrant on 13th July, 1992. The wind was strong that day blowing from south-west with a maximum speed of 20 knots. The hourly progressive vector for 13th July, 1992 also shows a near-linear movement of water to the south-west (Fig. 4.11C2).

Station 4

Data from this station covers the period from 11th July, 1992 to 28th July, 1992. A closed ellipse which indicates the rotary behaviour of the tidal flow was found on 15th July 1992 (Fig. 4.10D). The rose diagram, which starts with the flood tide at 08.30 hours and is

Fig.4.8 Time series of current direction (a), current speed (b) and wind speed (c) at station 3.

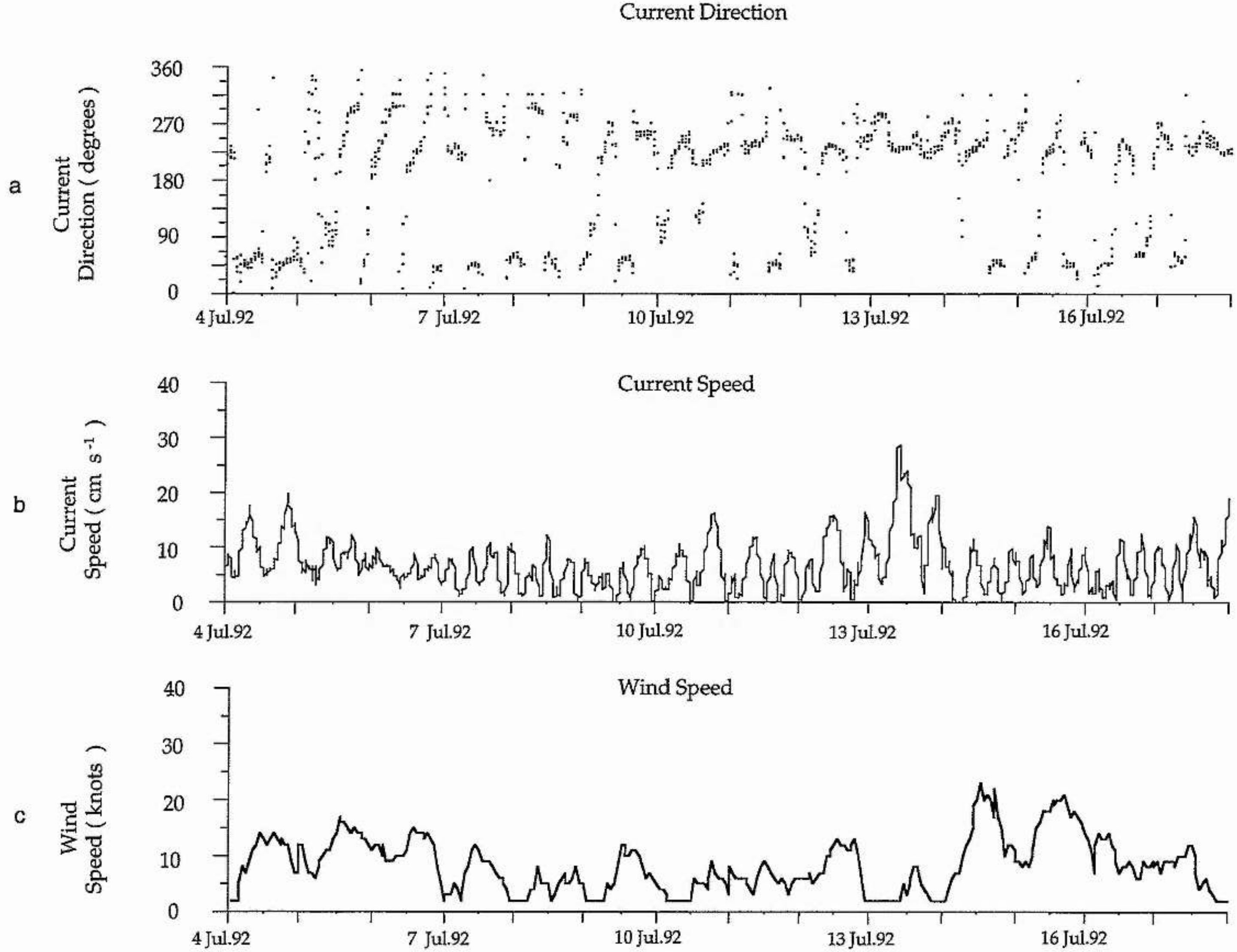
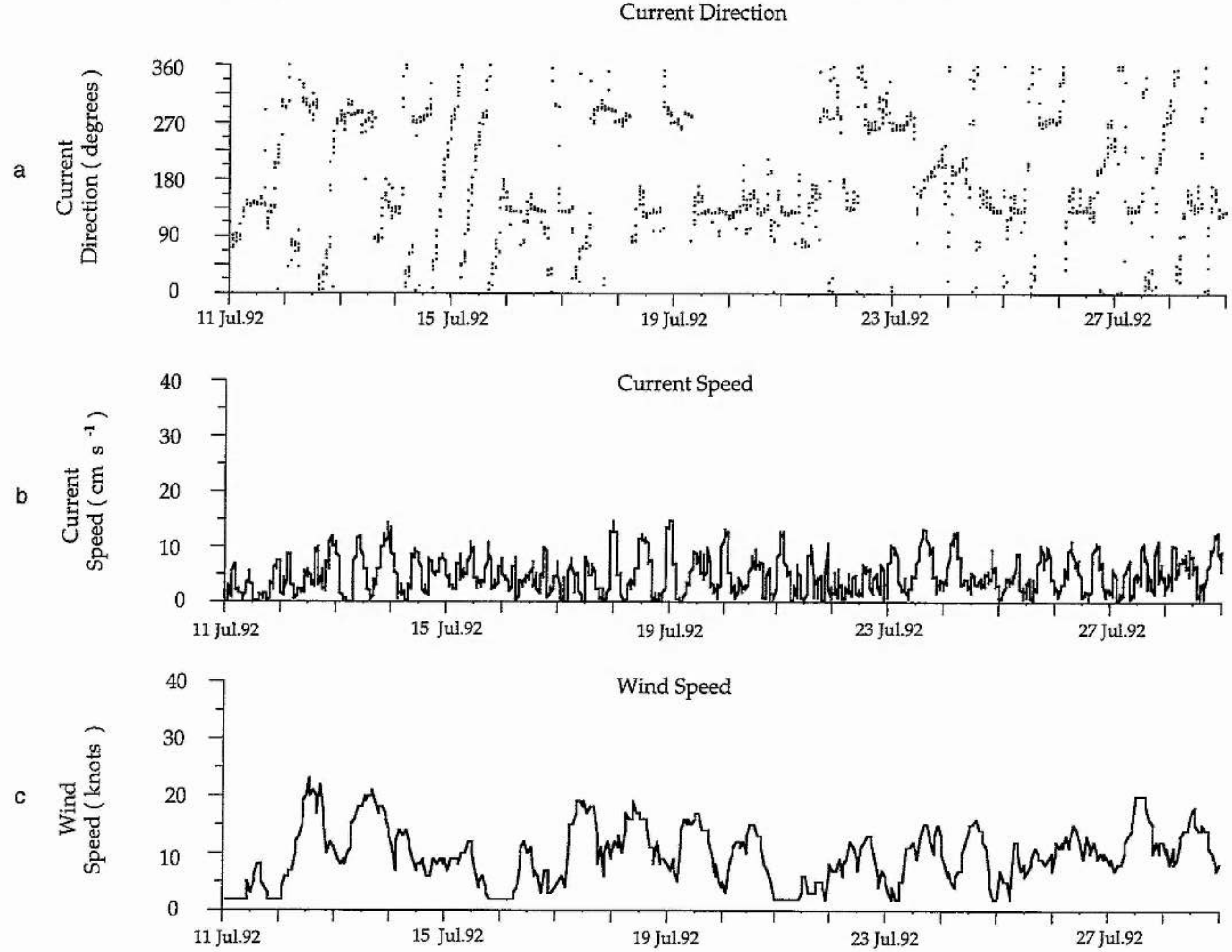


Fig.4.9 Time series of current direction (a), current speed (b) and wind speed (c) at station 4.



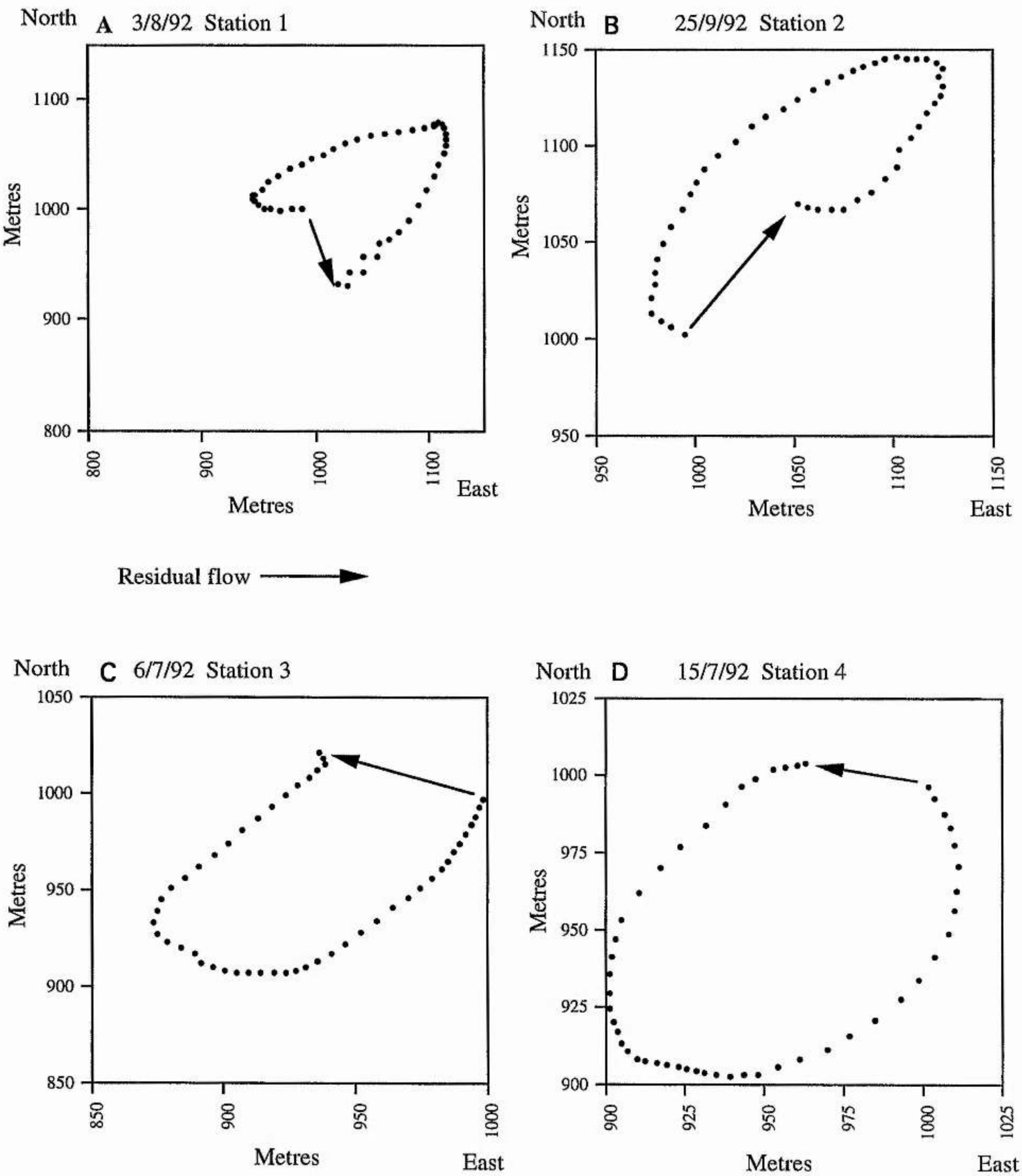


Fig. 4.10 Progressive vector plots of tidal flow indicating the rotary behaviour during calm weather conditions over one tidal cycle for Stations 1, 2, 3 and 4 respectivily.

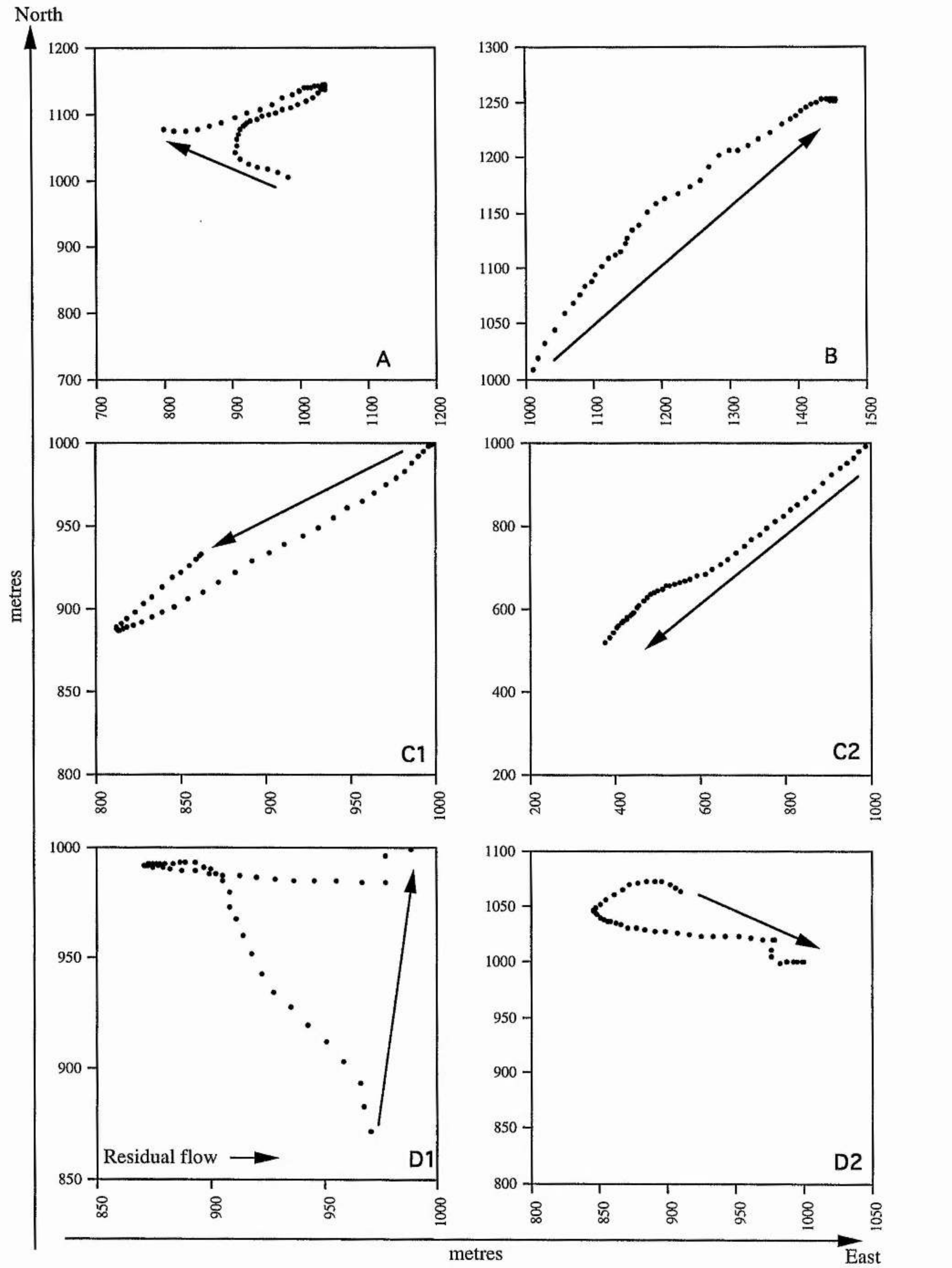


Fig. 4.11 Progressive vector plot showing the influence of meteorological forces on the behaviour of tidal flow over one tidal cycle for Stations 1, 2, 3, and 4 respectivily.

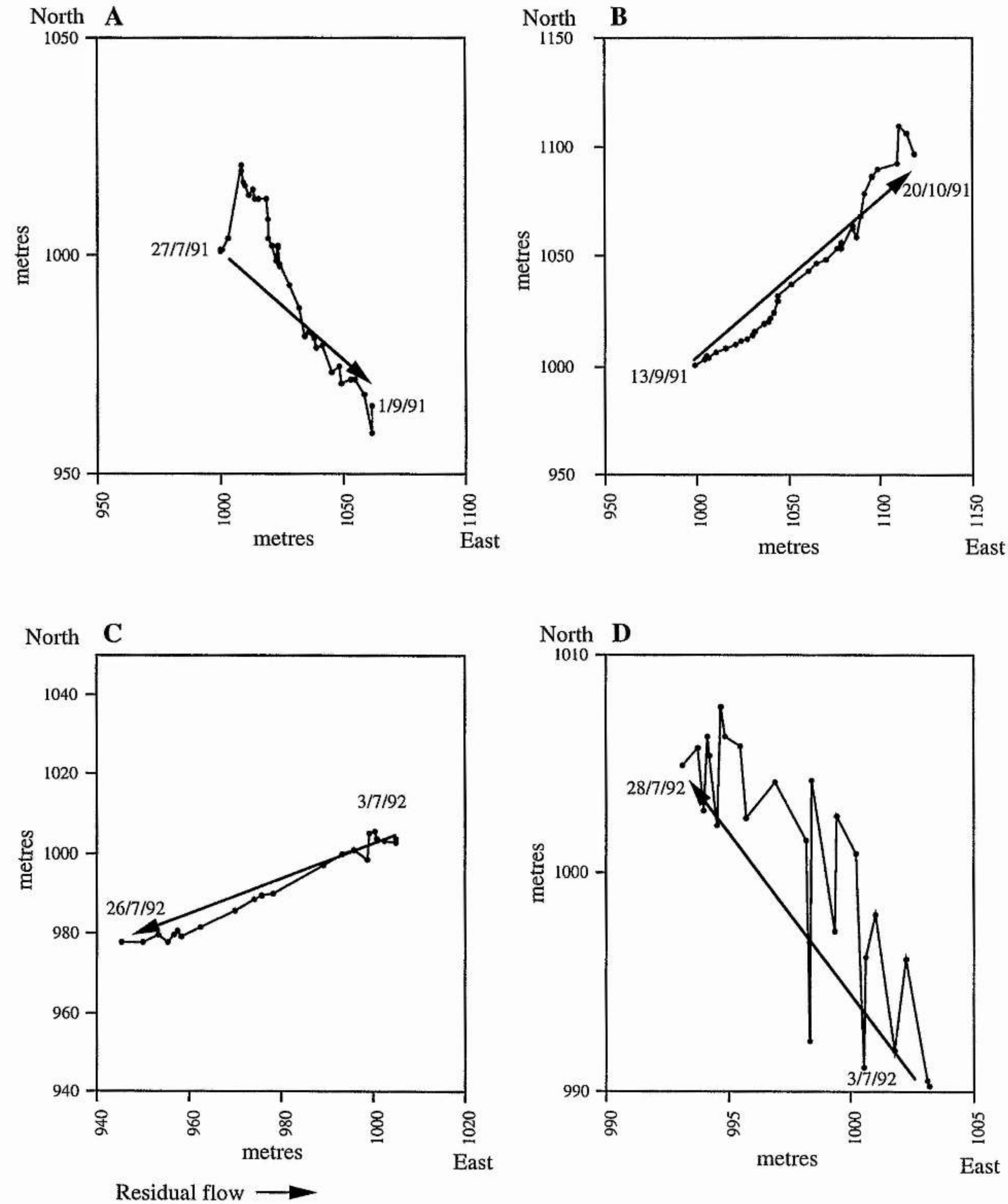


Fig. 4.12 Daily progressive vector diagrams for Stations 1, 2, 3, and 4.

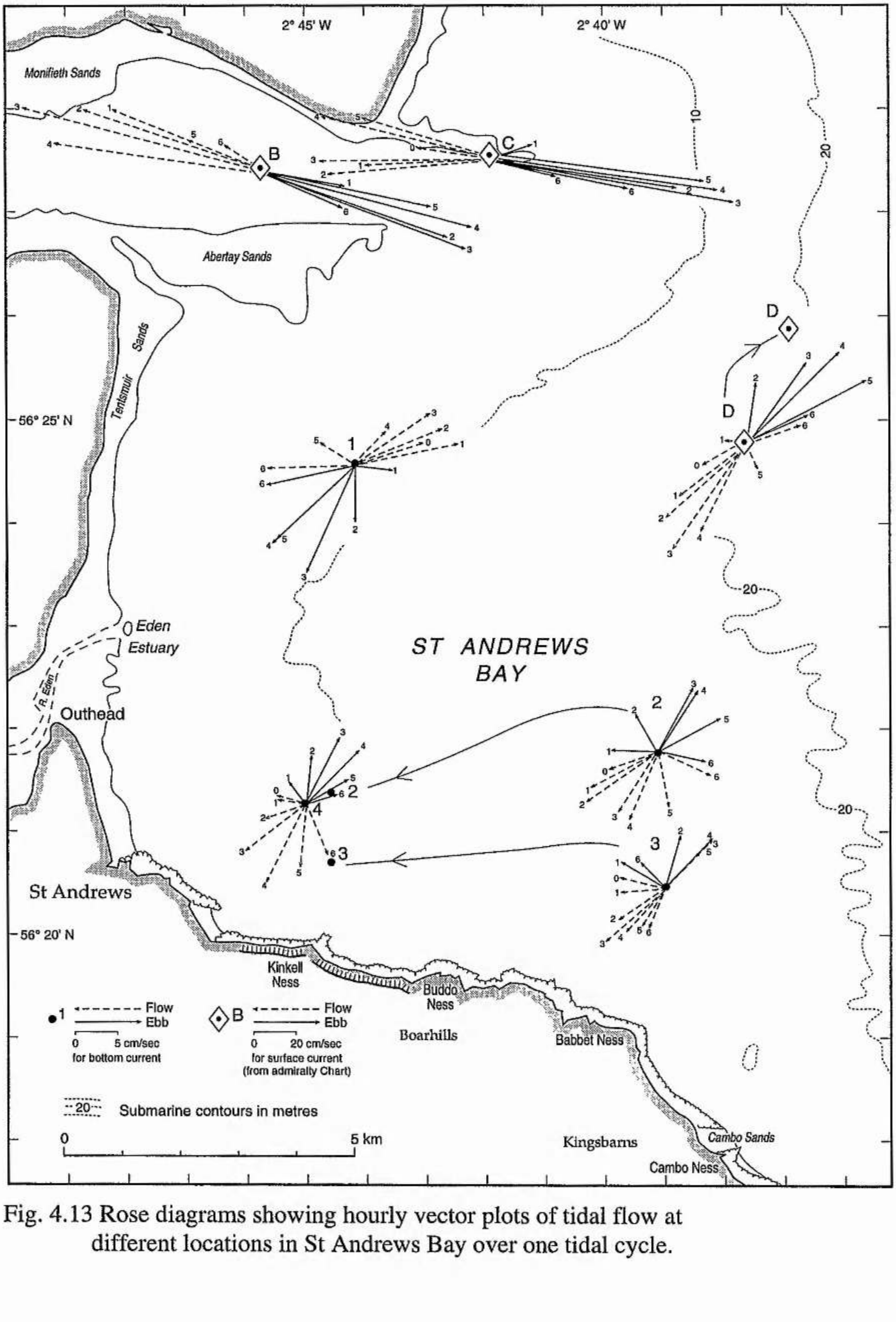


Fig. 4.13 Rose diagrams showing hourly vector plots of tidal flow at different locations in St Andrews Bay over one tidal cycle.

plotted for one tidal cycle, Fig. 4.13 shows that tidal flow starts to rotate from south-east clockwise through south to north-west, then north and later back to the north-east.

It is clear from the full current direction plot in (Fig. 4.9a) that the presumed simple pattern of the tidal rotations have been replaced by clustering of current direction data for some days during the record. For example on the days of July 12th -13th, 1992, the wind was strong (Fig. 4.9c), at more than 20 knots from the south-west quadrant. The hourly progressive vector diagram for the day of 13th July, 1992, reveals that the day was characterised by an open tidal ellipse (Fig. 4.11D1) with a residual flow towards the north-east parallel to the wind direction. Also, on 16th and 21st July, 1992, the data are again well grouped. However, on these days the wind record shows very calm weather (Fig. 4.9c), The residual flow for the day of 21st July 1992 was found to flow in a north east direction. Initially a daily progressive vector diagram shows that the residual flow of bottom water was first towards onshore and was later directed towards offshore(Fig. 4.12D).

4.5 Discussion

In the study area there are at least three different relationships between the tidal flow and wind speed. When measured wind speeds are light then the tidal current, plotted as a progressive vector, usually follows a simple tidal ellipse over the tidal cycle. For example, on 3rd August, 1991 (Station 1), the weather was very calm and the hourly progressive vector for this day shows a closed ellipse. This systematic variation of direction can persist for quite long periods of time. (Fig. 4.6a) and clearly shows that the tidal pattern is essentially rotary in character. Typical tides in calm weather conditions have been investigated for all stations and have been found to be essentially rotary rather than rectilinear (Fig. 4.10).

However, another relationship between the tidal current and wind occurs when the wind velocity exceeds a threshold value, subsequently producing a near-linear movement of water in the downwind direction (Fig. 4.11) . For example, at Station 2, the wind speed, on October 1st, 1991, was strong enough to drive the sea water downwind (Fig. 4.7c) and, thus, to produce a near linear movement of water. The daily (Fig. 4.12B) and hourly (Fig. 4.11B) progressive vectors for that day accord with each other and both confirm that the strong wind played a role in changing the tidal current from almost rotary to a near linear flow.

At Station 3 on July 11th, 1992, (Fig. 4.8c), it is clear that the wind is not, however, responsible for the form of the unidirectional flow. The daily progressive vector diagram plotted for the period from 3rd July, 1992 - 26th July,1992 shows a south-west residual flow for July 11th, 1992. This direction actually corresponds to the grouping of directional currents shown by Fig. 4.8a. Most likely there is another factor other than direct wind which affects the tidal ellipses and changes flow direction from a rotary tidal flow to a near linear flow. A possible explanation for this pattern of movement may be connected with the propagation of waves into St Andrews Bay from the North Sea, caused by wind which had been blowing over the North Sea in the past 72 hours, resulting in 'wave pumping' or mass transport currents. Likewise, at Station 4, on the days of 16th and 21st , July, 1992, a near linear flow is also observed at the time which does not appear to correspond to a period of strong wind (Fig. 4.11C). In this case too it is most likely that the last factor (wave propagation) was responsible for the formation of unidirectional current on these days.

Sediment Transport

4.6 Introduction

Many investigations of sediment transport in the nearshore zone have been undertaken in order to study a variety of theoretical and applied problems involving sediment movement , including, for example, the processes of coastal evolution (Carter, 1988), the prediction of pathways for particulate material introduced into the sea by man's activities (Komar, 1976) and the maintenance of navigation channels .These investigations have been carried out using a number of different methods to determine sediment transport. Some studies have used direct means of estimating sediment transport by using radioactive tracers (Heathershaw, 1981) and fluorescent tracers (Lees, 1983), or traps for bed load movement (Pickrill, 1986), or computing suspended sediment fluxes from measured concentrations of suspended sediment and water velocities (Jarvis and Riley, 1987). Other studies have used indirect means of estimating sediment transport, for example by analysing bedform migration (McCave and Geiser, 1978), or applying a "transfer function" to current meter data in order to estimate transport rate from the physical relationships established between flow strength and sediment data (Gadd et al., 1978; Vincent et al., 1981; Pattlaratchi and Collins, 1985).

It is frequently easier to evaluate sediment transport in the nearshore zone from indirect measurements because of the difficulties associated with making direct measurements. Accordingly, a knowledge of the pattern of water movement in the coastal zone is a fundamental prerequisite to understanding the pattern of sediment movement in that environment. However, the transport, of sediments in shelf seas occurs by a combination of two mechanisms: either directly as a result of the mean flow exceeding a threshold value, or by wave re-suspension and subsequent transport by the mean current. Consequently, it is necessary to take into account wave activity when considering sediment transport in the marine environment. Many studies have investigated the relationships between water movement and sediment transport, both in laboratory and field investigations, in order to predict the occurrence of sediment movement and mass transport of sediment over the sea floor (Sternberg, 1972; Komar and Miller, 1973, 1975a; Gadd et al., 1978; Vincent et al., 1981). From measurements of water movement, including wave activity near the bed, it is possible to calculate shear stresses on the bed particles and then determine potential transport for given grain sizes and bedforms. Such computations require detailed measurements of velocity and bed characteristics which were not practical in the present study. Accordingly, with a data base comprising current meter readings from a single elevation and grain size information, it was necessary to use a simpler approach to determine potential sediment transport in St Andrews Bay.

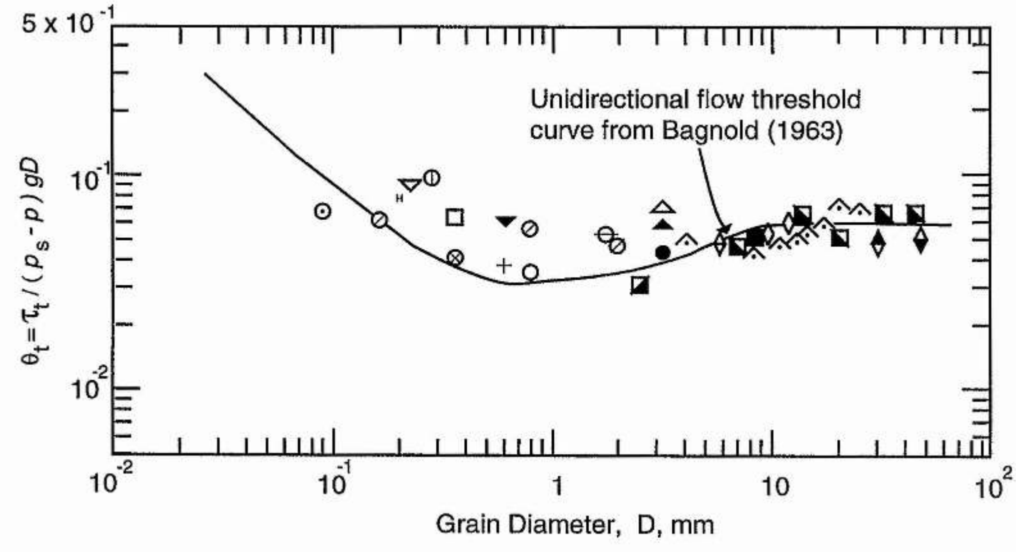
However, in order to understand how the sediments in a shallow marine environment such as St Andrews Bay are transported, it is necessary to consider the various processes acting on the sediment. In St Andrews Bay three components of water motion combine to produce the current near the sea bed and act as the primary factors in sediment transportation in the bay. (a) the tidal current; (b) wind-driven current, due to the response of the waters to meteorological forcing events; (c) and wave generated longshore drift from both north-east and south-east swell.

4.7 The prediction of threshold velocity

In coastal areas sediment transport can take place either as bed load or suspended load. The movement of sediment as bed load is at a slower velocity than the fluid velocity whereas suspended load moves approximately with the mean velocity of the current.

Sediment moves when the force of the water flowing over it overcomes both the force of gravity acting on the sediment grains and the friction between the grains and the surface on which they are resting. As the velocity of fluid flow over a bed of sediment is increased, there comes a stage when the fluid exerts a force or stress on the particles sufficient to cause them move from the bed and be transported in the form of traction, then suspended load. The stage at which movement starts is known generally as the threshold of sediment movement, or as the critical stage for erosion. Knowledge of the stresses and velocity required to initiate motion is important for the development of equations to predict sediment transport. The initiation of sediment motion is a function of the characteristics of the sediment, such as size, density, shape, sorting and packing. It is also a function of the fluid characteristics, such as density , viscosity and the average fluid velocity (Miller et al., 1977). Many studies of the critical velocity required to cause unconsolidated sedimentary particles to move have been carried out (e.g. Gilbert, 1914; Hjulstrom, 1935, 1955 ; Shields, 1936; Bagnold, 1963 and Postma, 1967). These studies have yielded a variety of competency curves that form the basis for predicting sediment motion (Sternberg, 1972). Curves which represent the results are commonly estimated using empirical correlation's based on laboratory experiments (Einstein, 1950; Bagnold ,1963; Yalin, 1963; Engelund and Hansen, 1967; Ackers and White 1973). Only a few attempts have been made to evaluate the application of formulae to the tidal marine environment, (Kachel and Sternberg, 1971; Gadd et. al. ,1978 and Heathershaw and Hammond, 1979).

One of the fundamental pieces of work (which forms a basis for much subsequent work) on the threshold determination was carried out by Shields (1936). The Shields diagrams are based on assumptions of non-cohesive sediment grains, steady and uniform flow, fully developed turbulent boundary layer, a flat bed and sediment of a single size grade. This is clearly a limitation for predicting motion in a field environment where sediments are composed of a mix of sizes; this has resulted in bed sediments being characterised by a single value such as the D_{84} size. Further, in the marine environment, currents are not normally steady and uniform. In areas such as St Andrews Bay, the estimation of the rate of sediment transport becomes much more complicated and the combined effect of wave action plus superimposed unidirectional currents must be considered (Jarvis and Riley, 1987). There have been fewer studies of the oscillatory threshold for sediment movement and most of the data have been obtained from either oscillating bed experiments, where a cradle holding the sediment is oscillated harmonically in still water (Bagnold, 1946), or from laboratory wave channels (Horikawa and Watanabe, 1967). Komar and Miller,1975a compared the threshold under waves with the threshold curves for unidirectional steady currents of Bagnold (1963). This comparison is shown in (Fig. 4.14) where the curve is the same as that in (Fig. 4.15). Madsen and Grant, 1975 independently arrived at results similar to those of Komar and Miller, 1975a in comparisons between the unidirectional and oscillatory threshold. The oscillatory threshold data were found to fit the curve given by Shields, 1936 for unidirectional threshold (Fig. 4.16), with about the same degree of scattering as the unidirectional threshold data. Therefore, the Shields threshold curve could be used for oscillatory water motion as well as for unidirectional flow. Accordingly, the effects of wave activity and unidirectional currents have recently been included in predictive equations for sediment transport (Madsen and Grant, 1976).



A. Bagnold (1946)

Symbol	material	density (gm/cm)	diameter. D (cm)
+	steel grains	7.90	0.060
●	quartz sand	2.65	0.330
○	quartz sand	2.65	0.080
⊗	quartz sand	2.65	0.036
⌀	quartz sand	2.65	0.016
⊙	quartz sand	2.65	0.009
■	coal	1.30	0.800
◩	coal	1.30	0.250
□	coal	1.30	0.036

B.Manohar (1955)

Symbol	material	density (gm/cm)	diameter. D (cm)
Φ	Del monte sand No.2	2.65	0.0280
⌀	B.E.B. sand No.2	2.63	0.1981
⊖	coarse sand No.2	2.60	0.1829
▽	glass beads No.1	2.49	0.0235
▼	glass beads No.2	2.54	0.0610
△	polyvinyl Chloride pellets	1.28	0.317
▲	polystyrene pellets	1.052	0.317

Rance &Warren (1969)

Symbol	material	density (gm/cm)	diameter. D (cm)
^	limestone chips	2.72 - 2.55	0.060
◊	glass spheres	2.54 - 2.44	0.592,0.884,1.186
◆	concrete cubes	2.39	4.775
◩	coal	1.37 - 1.29	0.706,1.372,2.042, 3.251,4.521
◆	perspex cubes	1.19	3.200
H	Horikawa & Watanabe (1967)		

Fig. 4.14 Data on the threshold of sediment motion under waves (Bagnold, 1946, Manohar, 1955, Rance & Warren, 1969) compared to the Bagnold (1963) curve for unidirectional flow sediment threshold.

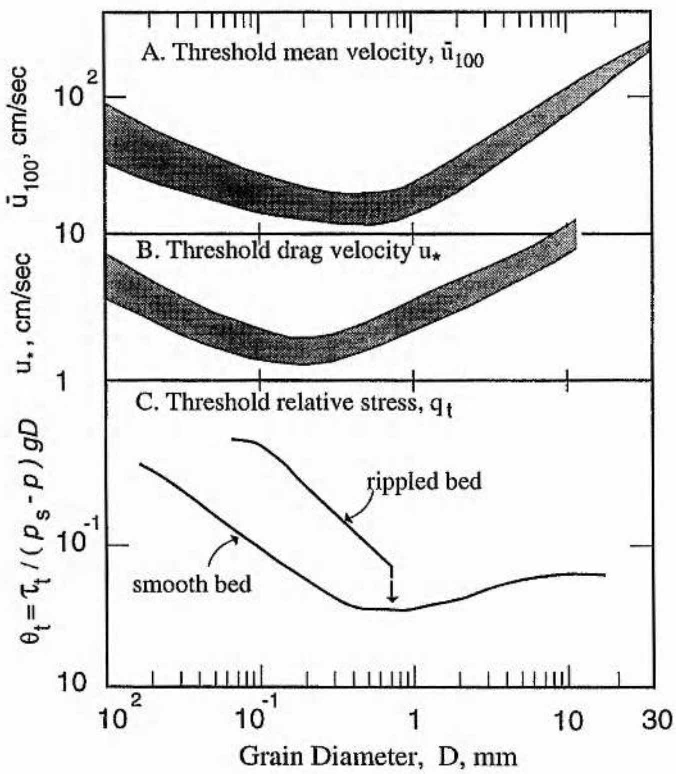


Fig. 4.15 Sediment threshold under unidirectional flow relating the threshold $\bar{u}_{100}$ $u_* = (r / p)^{1/2}$ and θ_t to the sedimentgrain size D. From Bagnold (1963) and Innman (1963).

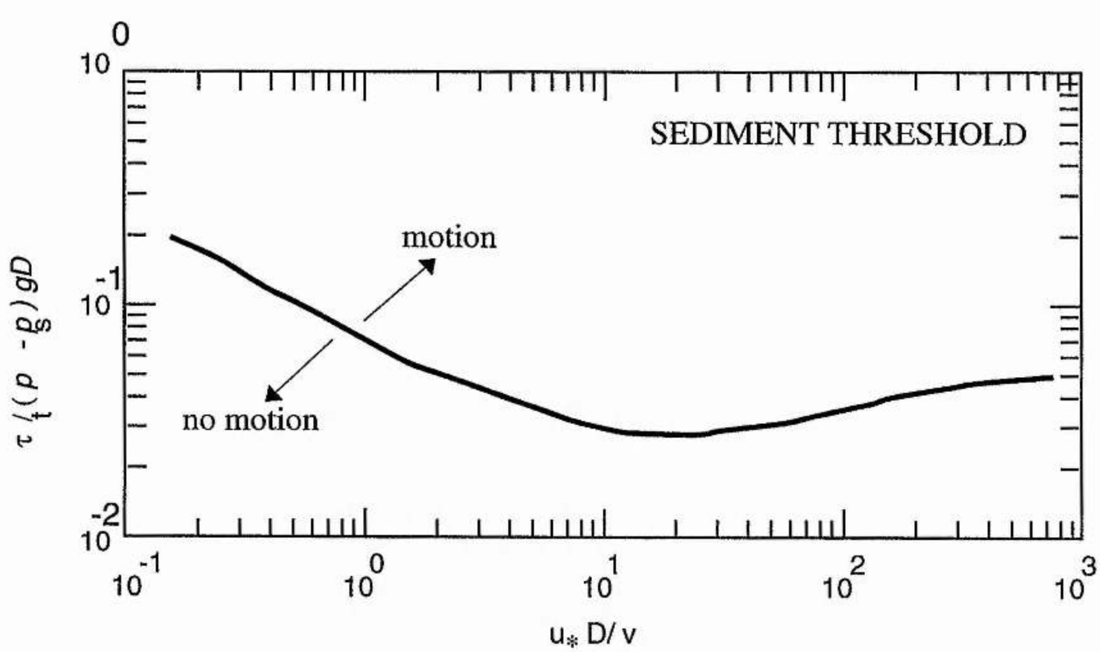


Fig. 4.16 Threshold of sediment motion underunidirectional currents. From Shields (1936), modified after results of White (1970).

4.8 The sediment transport equations

Several equations have been developed to predict potential sediment transport from hydraulic and sediment information. Many of these formulae relate to bed load transport because the inclusion of suspended sediment load is more difficult since it requires a knowledge of the concentration of sediment at specific heights above the bed- data which are frequently not available because specialised measuring equipment is required.

A number of workers have examined the validity of different sediment transport formulae in the marine environment. Gadd et al. (1978) used the empirical formulae of Bagnold (1956, 1963), Einstein (1950) and Yalin (1963) to estimate sand transport on the New York Shelf. In addition to these bed load formulae , the total load formulae of Ackers and White (1973) and Engelund and Hansen (1967) were examined in the course of the studies of sediment transport in Swansea Bay (Heathershaw and Hammond, 1979) and Sizewell - Dunwich Banks (Lees and Heathershaw, 1981). The New York and UK studies both used data sets of single elevation current meter data and observations of sediment size, similar to the data base in the present study. Thus these investigations may be used to support the approach used in St Andrews Bay to predict sediment transport.

The transport formulae developed by the workers referred to use different measures of fluid and sediment properties in their equations.

I- Einstein (1950) bed load equation

Einstein (1950) related sediment transport to random velocity fluctuations, rather than mean flow parameters. Einstein determined the probability of the lift force exceeding the sediment weight, in order to determine if sediment will be moved. The mathematical formulations will not be discussed here but it should be noted that the complex formulae require the integration of a probability integral to compute sediment discharge, a measure of the D_{50} grain size and the friction velocity u_*. The equations have no threshold values and were developed for relatively coarse material, with D_{50} of values between 785 and 2865 microns.

II- Bagnold (1963) bed load equation

Bagnold (1956,1963) expressed bed load transport ($\bar{q}sb$) in terms of the stream power (w) and an efficiency factor (k) and derived the equation :

$$\bar{q}sb = kw[r/(r_s - r)g]$$

where r_s and r are the sediment and fluid densities. The power expended on the bed by the fluid is expressed as :

$$w = t\ u*$$

where t is the shear stress on the bed , which can be written in terms of the friction velocity as

$$w = r\ u*^3$$

The efficiency factor was considered initially to be dependent only upon the sediment characteristics but it has been shown by Kachel and Sternberg (1971) to also be related to the excess shear stress as:

$$(t - t_{cr})/t_{cr}$$

where t_{cr} is the threshold shear stress. Gadd et al. (1978), using the flume data of Guy, Simmons and Richardson (1966), reformulated the Bagnold equation in terms of a velocity at a height of 1 metre above the bed (U_{100}) and a threshold velocity (U_{cr}) and obtained the transport rate as:

$$\bar{q}sb = b\ (U_{100} - U_{cr})^3$$

where b and U_{cr} are constants which were determined from flume experiments. For sand with a D_{50} value of 180 microns, b was found to be 7.22×10^{-5} and the critical velocity was found to be 16 cm s^{-1}.

III- Yalin (1963) bedload equation

Yalin's (1963) theory assumes that bedload particles move by saltation when the lift forces exceed the particle weight and that increased transport is caused by an increase in the saltation length and not necessarily by an increase in the number of particles. Yalin developed a bed load transport equation of the form :

$$q_{sb} = 0.635 r_s D_{50}\, u_*\, s[1-({}^{1}/_{as})\ln(1+as)]$$

where $a = 2.45({}^{r}/r_s)^{0.4}\,(r t_{cr}^{2}/[r_s - r]\, g\, D_{50})^{0.5}$ is a constant for given values of D_{50} and t_{cr}

s is the non dimensional shear stress given by : $s = (t - t_{cr})/t_{cr}$

Yalin's formula is restricted to plane beds, fully developed turbulent flow and large depth to D_{50} ratios. At low transport rates it can be shown (Heathershaw and Hammond, 1979) that the equation predicts that bed load transport is proportional to u_*^{5} but at high flow rates the bed load transport becomes proportional to u_*^{3}.

IV- Ackers and White (1973) equation

The Ackers and White (1973) transport formula is a comprehensive total load formula based upon the analysis of 1,000 flume experiments. This is a fairly complicated equation which will not be discussed here. It is based on the physics of stream power and dimensionless analysis methods. The equation can be applied to sediment sizes in the range 40 to 4,000 microns .

V- Engelund and Hansen (1967) equation

Engelund and Hansen (1967) relate the total load (q_{st}) to a friction factor (C_f), a dimensionless sediment discharge (f) and a dimensionless bed shear stress (q) :

$$C_f = 0.1q^{5/2}$$

where $$C_f = 2u_*^{2} / u^{2}$$

$$f = q_{st} / \left[r_s \left(g \left((r_s/r) - 1 \right) D_{50}^{3} \right)^{0.5} \right]$$

and $$q = t / (r_s - r) g D_{50}$$

u is the depth mean flow velocity

The total load can thus be expressed as :

$$q_{st} = 0.05\, r_s u^{2} \left(D_{50} / \left(g \left((r_s/r) - 1 \right) \right) \right)^{0.5} \left(t / \left((r_s - r) g D_{50} \right) \right)^{1.5}$$

This equation is valid for dune covered beds, Reynolds Numbers greater than 12 and D_{50} values greater than 150 microns.

In the New York Shelf study by Gadd et al. (1978) a check on the fluxes predicted by the three bed load equations was possible. At one site, where current meter observations were made, measurements were also made of the dispersal of radioactive sand at the same time as current data were collected. During the period of the tracer survey the threshold current for transport was intermittently exceeded with a single storm event dominating transport. For the two periods between tracer surveys on November 12th, 25th and December 4th, 1974 the bed load transport was calculated from current meter data taken at 10 minute intervals (a drag coefficient of 3×10^{-3} was assumed and the threshold for transport set to 18 cm s^{-1}). During this limited study, it was not deemed necessary to evaluate the transport direction. The results of the comparison are shown in Table 4.6.

Transport interval	$\lvert Q \rvert$ Tracer	$\lvert Q \rvert$ Bagnold	$\lvert Q \rvert$ Einstein	$\lvert Q \rvert$ Yalin
Nov 12-25 1974	3.2×10^{-3}	2.2×10^{-2}	8.8×10^{-3}	2.8×10^{-3}
Nov 25-Dec 4 1974	1.7×10^{-1}	2.2×10^{-1}	8.9×10^{-2}	2.3×10^{-2}

Sediment discharges $\lvert Q \rvert$ are in g cm s^{-1}

Table 4.6 Measured and computed sediment transport rates on the New York Shelf using different bedload equations (after Gadd et. al. ,1978).

It can be seen that the tracer calculations are all within an order of magnitude of the formulae predictions, which vary themselves between by as much as an order of magnitude. During the duration of the study from October 1973 to June 1974 maximum predicted rates in 5m of water reached as high as 8.02×10^{-1} g cm^{-1} s^{-1} (Bagnold equation) but were much less further offshore e.g. 4.98×10^{-4} g cm^{-1} s^{-1} in 15m of water. Much higher rates were noted during the period October- December 1973, compared to the period April-June 1974 (in most cases by at least an order of magnitude, at similar locations). These results reflect the fact that the threshold for transport was exceeded much more frequently, 17% -90% of the time during the winter months compared to 1.3%-32% of the time in the spring.

The results of investigations into sediment transport in Swansea Bay were reported by Heathershaw and Hammond (1979) and Heathershaw (1981). In addition to a comparative study of measured and predicted sediment transport, Heathershaw (1981) also carried out a sensitivity analysis of the 5 transport equations to identify which equation was most robust when there was some uncertainty about some of the parameters such as roughness length, sediment size or threshold velocity.

Fig. 4.17 shows the predicted and measured net bed load transport for selected current meter records during the period of the survey in Swansea Bay. Data were averaged because the tracer surveys only record mean transport between surveys. It is apparent that there are two orders of magnitude of spread in the predictions of transport rate by the 5 equations. At deeper water locations, the Ackers and White's total load equation underestimates the

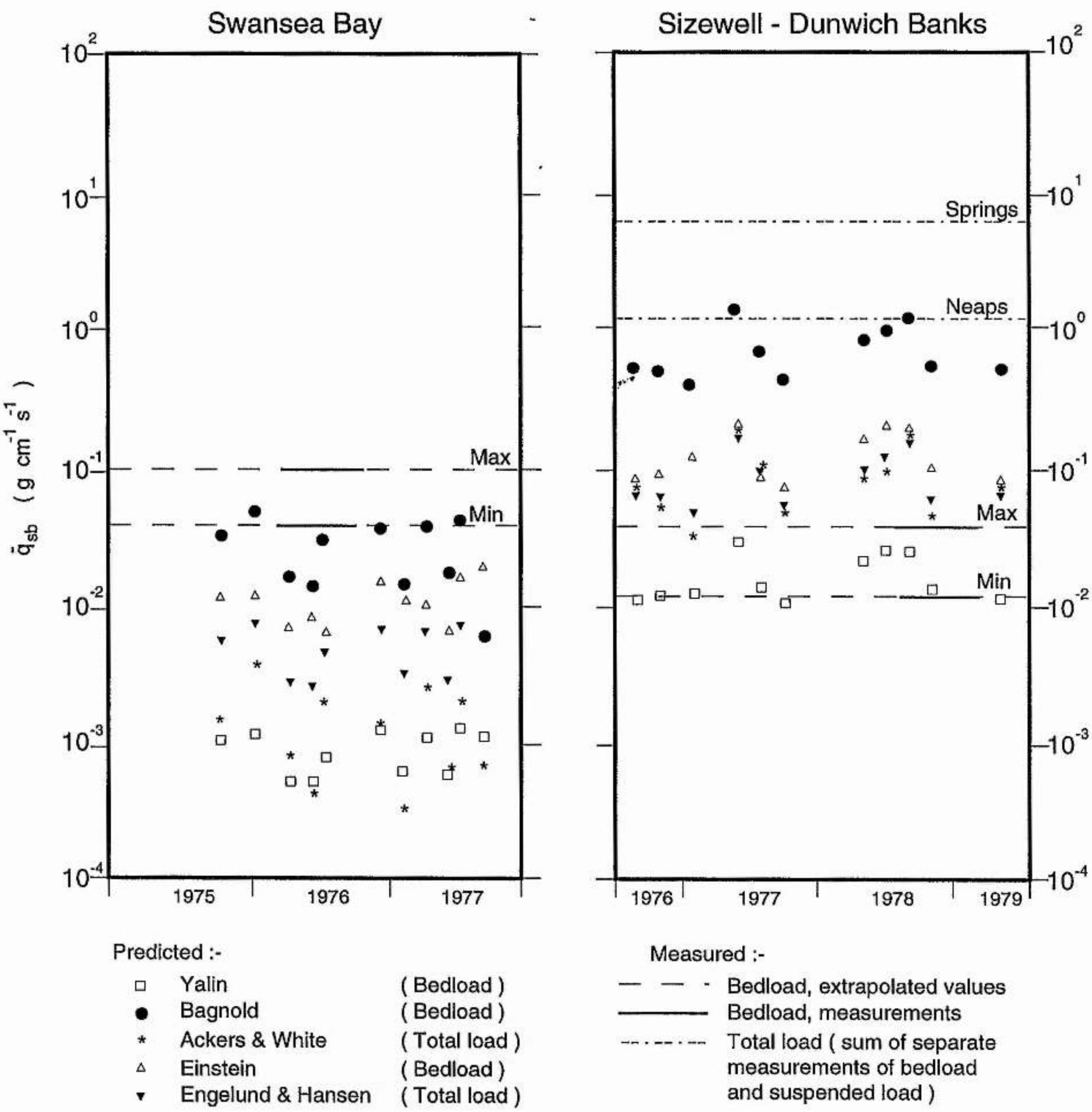


Fig. 4.17 Comparison of measured and predicted net transport rates ($\bar{q}_{sb}$), (from Heathershaw 1981).

measured rate, whereas Bagnold's equation gives the highest estimates. However, in a shallow bank situation Ackers and White's equation gives higher estimates which are in good agreement with Bagnold's equation. The other equations give estimates intermediate between those of Bagnold and Ackers and White. Bagnold's equation gives the closest predictions to the measured rates which ranged from 0.03 g cm^{-1} s^{-1} to 0.17 g cm^{-1} s^{-1}.

In the sea, when using just a single current meter to obtain data, some of the variables in the transport equations (e.g. bed roughness),may have to be estimated. The sensitivity of the transport equations to some variable can be evaluated by holding all variables except one constant and then seeing how the predictions change with changes in the remaining variable. Heathershaw's (1981) analysis showed that Bagnold's equation was least sensitive to changes in roughness length and sediment size, whilst the Ackers and White's equation was most sensitive to these variables. Engelund and Hansen's (1967) and Ackers and White's (1973) formulae were least sensitive to changes in water depth.

Contrasting results to the Swansea Bay study were reported from the Sizewell-Dunwich Banks work (Lees and Heathershaw, 1981). The equations used predicted rates which were relatively the same as in Swansea Bay but, off the coast of East Anglia, Bagnold's equation over predicted measured rates by 2 orders of magnitude and the best results were obtained from Yalin's equations (Fig. 4.17).

Lees (1980) attributed the differing performance of the 5 equations in the Sizewell-Dunwich Banks study to the fact that in Swansea Bay the majority of sediment transport takes place as bed load, whereas at Dunwich the major part of the transport is as suspended sediment .

Clearly, the choice of sediment transport formulae depends both upon the data base available and some indication of the possible mode of transport in the study area. The New York study and the work carried out on in British coastal waters suggest calibration of the equations against tracer experiments is clearly desirable. However, this was not possible in the present study. St Andrews Bay has a somewhat similar morphological setting to Swansea

Bay and it was thus decided to adopt the strategy of Bagnold, (1966) modified by Gadd et al.(1978) and Vincent et al. (1981) to predict the transport of sediment .

4.9 Sediment transport calculated in study area

Current velocity recorded at 15 minute intervals at Stations 1, 2, 3 and 4 (Fig. 4.3) is only the parameter used in the empirical transport formula. Current velocities above the critical velocity were used in the manner of Gadd et al.(1978) and Vincent et al. (1981) to predict the bed load transport of sediment .

At Station 1, the bed load sediment transport rate varied between zero rate and 0.34 $x10^{-1}$ g cm^{-1} s^{-1}. A high rate of bedload transport occurred on t1st September, 1991, in spite of a low wind speed (average= 8 m s^{-1}). On the other hand, on some days when no bed load movement took place such as on 31st July, 1991, the wind was blowing from the NE sector with an average wind speed of 14 m s^{-1}. This result indicates that the relationship between wind speed and the quantity of bed load sediment transport is by no means clear. The only explanation for the higher rate associated with a low wind speed is that the mean flow direction was parallel to the wind driven current. Conversely, a zero rate on the day of high wind speed may be because the direction of mean flow was opposite to that of the wind driven current.

At Station 2, the bed load sediment transport rate ranged between zero and 1.25 x 10^{-1} g cm^{-1} s^{-1} ·The highest value occurred while the average wind speed reached 6 m s^{-1} from the SW.

At Station 3, the maximum bedload sediment transport rate is 0.169 x 10^{-1} g cm^{-1} s^{-1} found on the day of 13th July, 1992. The average wind speed in this day is 18 m s^{-1} . In this case the higher predicted bed load sediment transport rate could be related to the fact that the tidal current component is parallel to the wind-driven current component.

At Station 4, the bed load sediment transport rate ranged between zero and 1.53 x 10^{-3} g cm^{-1} s^{-1}. The highest value of transport rate occurred on 28th July, 1992 when an average wind speed about 15 m s^{-1} was recorded from the SW.

The results, in general, indicate that, although the wind plays a very important role in bed load sediment transport, a clear relationship between the two is not apparent.

4.10 Sediment transport discussion

It is not an easy task to quantify the rate of bed load transport. Indeed, it is particularly difficult if the sediment is in the size range of fine sand where it may be difficult to separate bed load and suspended transport. Many sediment transport formulae have been proposed using different methods. But any formula will only be as good as the data used to establish them. In St Andrews Bay a restricted data base of single current meter observations is available. The sea bottom of St Andrews Bay is characterised by fine sand (between 0.1 and 0.2 mm). There may be no clear distinction between those particle sizes that are transported as bed load and those particle sizes that are transported in suspension (Ingle, 1966). However, it is believed that, in the study area bed load transport is dominant.

The average current speed at all the stations was between 8 and 12 cm s^{-1} . This reflects the low levels of energy in the area and indicates that sediment movement will be low. However, most previous studies (Ferentinos and McManus, 1981; Jarvis and Riley, 1987; Jarvis, 1989)in the area emphasise that there is a movement of sediment onshore. The accumulation of sand on the tidal flat areas and the northerly migration of Outhead are evidence that transport of sand occurs in the area. The fact that the threshold of current velocity for sediment movement is only exceeded occasionally implies that sediment movement in the study area probably occurs infrequently rather than continually and depends on the weather conditions. According to Clarke et al., 1982, the sand-sized fraction is suspended primarily by bursts of turbulence which are related to peak values of the envelope of surface waves. Subsequently, it is possible to conclude that, although the unidirectional current very weak in magnitude is still able to cause a transport in study area, since the sediment is already in motion under the wave action.

At measurement sites in St Andrews Bay, the average predicted rates of bed load sediment transport can be seen to vary by as much as an order of magnitude (Table. 4.7).

Table. 4.7 indicates that a higher average bed load transport rate was observed at Station 2 and Station 4, whereas smaller average values were recorded at Station 3 and Station 1.

Stations No	Grid Reference	Average bedload transport
1	546244	4.06×10^{-3} g cm $^{-1}$ s $^{-1}$
2	541186	1.4×10^{-2} g cm $^{-1}$ s $^{-1}$
3	538171	1.86×10^{-3} g cm $^{-1}$ s $^{-1}$
4	537181	1.53×10^{-2} g cm $^{-1}$ s $^{-1}$

Table 4.7 Computed bedload sediment transport in St Andrews Bay.

In Swansea Bay the maximum rate found at Station F (see Fig. 4.1 and table IV, Heathershaw, 1981) was 3.8×10^{-1} g cm $^{-1}$ s $^{-1}$. The minimum value was found at Station B at Port Talbot (8.9×10^{-5} g cm $^{-1}$ s $^{-1)}$. In addition, Lees (1983), who calculated the bed load sediment transport rates in Sizewell-Dunwich Bank area, found variations from 0.12×10^{-1} g cm^{-1} s^{-1} to 0.4×10^{-1} g cm^{-1} s^{-1}. The difference in the results between the Swansea Bay and Sizewell-Dunwich Bank studies , in particular the difference in the minimum value of bedload transport by four orders of magnitude, is a reflection of the differing geographical settings of the areas. Swansea Bay is an embayment, with a shallow water depth, modest tidal currents and moderately high wave energy environment. On the other hand, Sizewell-Dunwich Bank is considered to be a tidally-dominated area with strong tidal currents and exposed shoreline with the area which has a similar coastal geometry, i.e. Swansea Bay. In general, the average rates of bed load sediment transport in St Andrews Bay are very low. The results indicate smaller rates than those reported for Swansea Bay, by one to two orders of magnitude. In Swansea Bay, the threshold velocity is considered to be 16 cm s^{-1} . In St Andrews Bay the Bagnold equation (1963), as reformulated by Gadd et al. (1978), has been applied, but the threshold velocity of Vincent et al (1981), 12 cm s^{-1} for fine sand, was adopted. This threshold was used in order to include the effects of wave activity since in St Andrews Bay the mean tidal current will combine with oscillatory wave motion to cause the threshold velocity to be exceeded for a fraction of each wave period.

Lower values of bed load transport rates in the study area, as compared with Swansea Bay, can be attributed to several factors. Firstly, the fact that most waves propagating towards the shoreline lose their energy farther offshore before reaching the foreshore area. Secondly, the water circulation system in St Andrews Bay differs from that of Swansea Bay, in that the Bay is more enclosed compared to Swansea Bay and the presence of the two estuaries has an important influence on water circulation in the study area. Thirdly there is the possible influences of the local topographic effects, in particular at Station 3 which is located close to the rocky shore. The underwater rocks may act as a buffer zone which reduces the values of the current readings at Station 3. The forth factor is the presence of the sedimentological features such as the tidal sand spit complexes of the Abertay Sands and Gaa Sands. These areas will absorb most the wave energy, leaving Station 1 in a shadow zone.

However, the highest rates of bed load transport at Stations 2 and 4 are consistent with the previous findings of Jarvis and Riley (1987). They showed from their survey that there are sand bars which migrate onshore and decay in amplitude towards the mid- tide location. This process of onshore migration has led to net accretion on the tidal flats at Outhead near the Eden Estuary mouth.

CHAPTER V

The Rocky Coast around St Andrews Bay

5.1 Introduction

Shore platforms are prominent features around the British coast and have been the focus of considerable attention, both in terms of their recognition and processes of formation. The term shore platform, has recently superceded the older term wave-cut platform since the latter implies that a single processes is responsible for platform development. Thus, the term shore platform reflects the modern view that a variety of different processes are acting on a platform surface and that the combined action of all these results in the final platform morphology (Bird, 1966; Pethick, 1984). In general, platforms develop and widen as cliffs recede and are shaped by the action of waves and other marine processes, although there are difficulties involved in relating platform shape to any causal processes (Pethick, 1984; Carter, 1988). The processes acting on shore platforms have usually been inferred from their morphology. However, although many processes operate on platforms, only two, mechanical wave erosion and weathering, are generally considered to be dominant (Sunamura,1992). The processes of mechanical wave erosion are probably most effective in the storm environment of temperate latitudes as there are several factors which prevent weathering processes from assuming a major role in platform development , such as low rate of evaporation, semi-diurnal tides and sloping platforms (Trenhaile,1980). Shore platform morphology at any location may be characterised by such factors as width, gradient, relief and relationship to high and low water marks. Many factors affect these parameters, for example the geological structure and the lithology, exposure to wave activity, height of backing cliffline, age of platform and changing sea levels.

It has been suggested that shore platforms may have been inherited from a period when sea level was similar to today's. Certainly, platforms cut in relatively resistant rocks may have been inherited. However, areas composed of soft rocks have been the subject of some debate. While Trenhaile (1980) believed that shore platforms in less resistant rock areas are contemporary features in or close to a state, of dynamic equilibrium, Gill (1972) proposed that inheritance might be important for those areas where erosion is more rapid.

The coastline of the south of St Andrews Bay is formed mostly of a well exposed inter-tidal rocky platform (Plate 4). This platform cuts across Upper Devonian and Lower Carboniferous sedimentary (320 - 345 million years) and volcanic rocks. The sedimentary rocks that outcrop along the coast to the south of the Eden Estuary belong to the Calciferous Sandstone Measures. However, sandstones are not the only rocks to be found, because the Calciferous Sandstone Measures consist of a rhythmically inter-bedded sequence of sandstones, fine grained shales and subordinate coals (Armstrong et al., 1985).

Wide intertidal platforms emerge at low tide but some parts remain submerged throughout the complete tidal cycle. The structure and morphology of the intertidal platform can be determined from the interpretation of air photographs and the large scale Ordnance Survey Maps (1 : 10,650 Sheet 51) in addition to using conventional surveying techniques. Offshore from the intertidal platform the Admiralty Charts of the area, e.g. Sheet 59, show a sea bed with considerable rock exposure between the Low Water Mark and a depth of about 16 m below Chart Datum. In the study area, some parts of this rocky platform are backed by weathered and steep cliffs, e.g. between St Andrews and Kinkell Ness (Plate 5); in others, by sandy beaches such as Pilmour Links and Tentsmuir Beach. In addition, small beaches in coves, at the base of low cliffs (Ritchie, 1979) occur as well as variable strips of sand on top of abraded rock platforms (Plate 6).

The more resistant parts of the coastal rock formations produce headlands, or persist as rocky stacks and rocky islands offshore, whereas the weaker elements are cut back as coves and embayments. Cliffed coastlines occur discontinuously between Ethie Haven and Arbroath in the north and between St Andrews and Fife Ness in the south (Sarrikostis and McManus ,1987). These cliffs are associated with extensive rocky platform varying in width.

The immediate offshore zone adjacent to sandy beaches is characterised by a relatively smooth bottom topography except at the estuary mouths (Ferentinos and McManus, 1981; Jarvis and Riley, 1987). Offshore from an intertidal rocky platform the bed is characterised by rugged features, considered to be submerged rocks in the subtidal zone. This Chapter of the thesis aims to study the subtidal platform of St Andrews Bay, describe the intertidal platform, and to use it as a model to interpret the subtidal features.

5.2 Previous work

Early studies of shore platforms were descriptive and generally concerned with determining whether platforms are the products of wave erosion or weathering processes (Edwards, 1951; Bird, 1968; Hill, 1972). A general conclusion reached from these studies was that cliffed coasts and their associated platforms are produced mainly by marine processes. Edwards (1951) put forward arguments in favour of the wave erosion theory for the development of platforms on the coast of Victoria, Australia, where the platforms are best developed in front of cliffed headlands. Cliff erosion and the development of platforms depend upon the energy available; this is limited by the wave energy that can reach a particular point on the coast at a given level and by the manner in which the waves break (Flemming, 1965). Under normal weather conditions, erosion is concentrated on the seaward edge of the platforms but during periods of storms the waves can attack the main cliff and the platform surface is more strongly abraded. Accordingly, the processes that operate to cut back the main cliff are, therefore, the most important in the growth and preservation of the platform. According to Bird (1976) and Sunamura (1983) platforms developing around present sea level are broadly classified into; (1) gently sloping platforms without a significant topographic break, extending from the base of a cliff to the sea floor just below low tide level; (2) nearly horizontal platforms with a marked drop at their seaward edge. Gently sloping platforms are most common in macro-tidal environments whereas nearly horizontal platforms with a seaward drop are usually found in meso or micro-tidal regions (Trenhaile, 1987). Such contrast can be used to differentiate between the platform morphology in the Northern and Southern Hemispheres (Pethick 1984). Several authors note that platform morphology is quite distinct in the Northern and Southern Hemispheres (Davis, 1980). For example, platforms on the British and North American coasts tend to be wide, gently sloping , with a linear profile; this could be due to a large tidal range. Areas with a small tidal range like Australia, tend to have horizontal platforms with a low tide cliff.

The seaward drop is characteristic of nearly horizontal platforms and assumed that it is a relic of the initial profile of steep drowned coasts. The topographic break preserved is therefore shown to be immune from wave erosion during platform cutting (Sunamura, 1975). Further, Sunamura (1992) states that, if downward bedrock erosion concomitant with platform growth is so active as to lower the platform surface, eventually leading to the disappearance of the seaward drop, then a gently sloping platform results. Otherwise, nearly horizontal platforms develop.

However, there are usually several complicating factors at work on platforms which can strongly influence the development of cliff and shore platform profiles. These factors, identified by Bird (1966), are structural, lithological, degree of wave exposure and the dip of the beds. He further suggested that platforms are best developed where the coastal rock formations are homogeneous without structural or lithological variations, since heterogeneity's in lithology and the presence of different kinds of structures in the shore zone rocks yield irregularities in the profile which diverge from the ideal.

According to Trenhaile (1980), the geological influence on platform development is expressed in several ways. Firstly, the lithology and structure govern the efficiency of the processes operating on shore platforms, for example if the rocks are thinly bedded and densely jointed, processes such as wave quarrying and plucking are usually dominant. The second geological factor affecting platform development is the resistance of the rocks to the processes operating on them. This will determine the degree to which platforms are inherited features. The third factor is that the platform profile may be influenced strongly by structural and lithological factors, for example rock strike and dip usually determines the surface roughness of platforms. So even slight variations in lithology on a single platform will be picked out by the erosive processes and can result in an irregular or rugged surface.

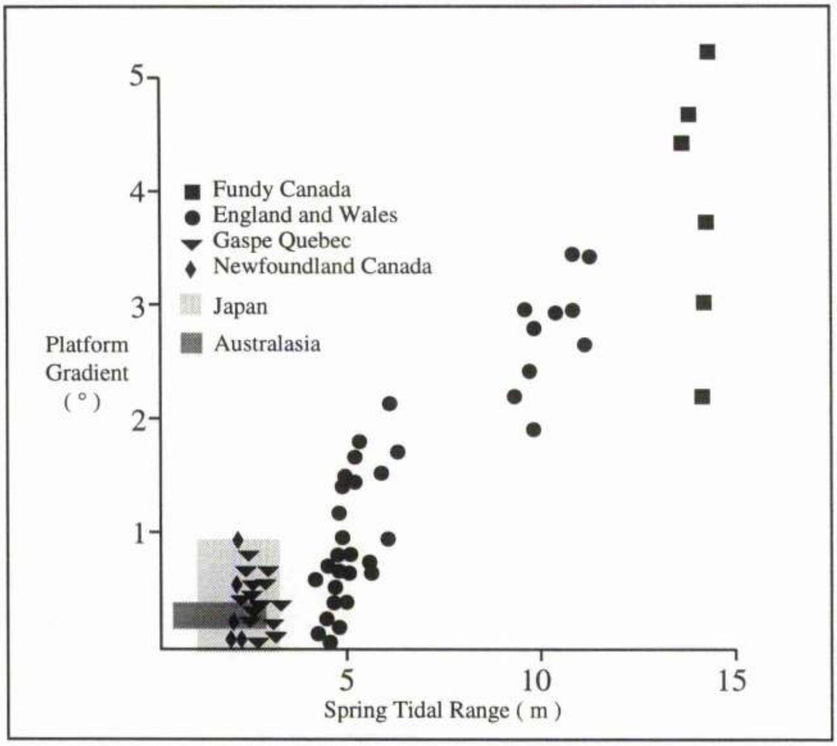


Fig 5.1 Plot of platform gradient against tidal range. Each point is a regional average representing numerous surveyed profiles (after Trenhaile, 1987).

Trenhaile (1974a) found a strong, positive relationship between tidal range and platform gradient in several areas in the storm wave environments of the northern Atlantic, a result which is independent of the rock type, but only for a tidal range of more than 3 m. Trenhaile arrived at this relationship by using more than 800 surveyed profiles from three macro tidal areas in England and Wales where he found a positive correlation (r= 0.92) between platform gradient and tidal range. Also he added data from Gaspe in eastern Quebec, where the tidal range is low meso-tidal (Trenhaile, 1978) which provided an over all correlation of 0.88 . This indicates that platforms become almost horizontal when the tidal range is less than about 2.5m. Despite the effects of different climates and wave regimes it appears that this relationship also exist in areas outside the north Atlantic, e.g. Japan and Australia (Trenhaile, 1987). Fig. 5.1 shows the plot of platform gradient against tidal range in different locations around the world. In general, the slope of the platform is attributed to wave erosion while horizontal platforms, at high or low tide, are the results of chemical weathering processes such as solution. In fact, the differences in profiles are related to morphogenic conditions. High slopes of platforms in England and Wales are related to high tidal range, storm waves, large and abundant erosion debris and climatic conditions unsuitable for effective water layer levelling (Trenhaile, 1974b). Water layer levelling usually occurs in warm areas which are characterised by higher rates of evaporation e.g. most parts of Australia, than in Britain. Moreover, the lower gradients (horizontal platform) allow pools and bodies of standing water to persist on the platform, which may further accentuate the levelling process. Usually waves reach the shoreline at elevations determined by the tidal level. Although waves, and particularly breaking or broken storm waves, perform the erosive work, it is the tidally controlled distribution of wave energy which determines where this work is performed. So the level of greatest wear must be closely associated with the elevation most frequently occupied by the still water level. This is a tidal duration factor, which means that the rate of intertidal erosion is determined by the time that still water level occupies each elevation within the tidal range. Trenhaile and Layzell (1981) demonstrated that the tidal duration would be maximum at mid-tide and that the spatial variation in process intensity which this would produce could explain observed differences in platform morphology. Thus quasi-horizontal profile sections will be formed at mid-tide level with a concave ramp at high tide level and a convex cliff at low tide level. Since tidal duration maxima must decrease as tidal range increases, it follows that the dynamic equilibrium profile on a low tidal range coast will exhibit a wider central horizontal section reflecting the longer duration. Conversely, macro-tidal platforms, with lower tidal durations, will tend to be more uniformly sloping since the intensity of wave processes are more evenly distributed (Pethick, 1984). The tidal duration model of Trenhaile (1980) and Trenhaile and Layzell (1981), has been debated by Carr and Graff (1982). They point out that, if tidal duration curves are plotted without the use of smoothing techniques, then duration maxima are found at high and low tide and not at mid-tide. They show that, if erosion is concentrated at these tidal extremities, a ramp will result at high tide and a cliff at low tide. Moreover, they suggest that the central section of the profile between ramp and cliff will be gently sloping due to the lack of intense erosion there.

It is geometrically possible for platform width to increase, decrease, or even to remain constant, as tidal range increases, assuming that the cliff-platform junction remains at the high tide level. Although theoretical considerations indicate that platform width increases with the tidal range, the difficulty in confirming this relationship in the field suggests that geology, wave energy and other factors exert a particularly strong influence on platform width (Trenhaile, 1978; Trenhaile and Layzell,1981).

Over large areas where there are significant variations in the tidal range, there is a positive relationship between gradient and the tidal range and also between width and tidal range. But at the local scale, where tidal range can be assumed to be constant, platform width varies due to other factors, such as differences in rock resistance and exposure to wave activity. The strike of dipping beds should also be considered. In steeply dipping rocks platforms are widest when the strike is perpendicular to the cliff, whereas in gently dipping strata they are widest when the strike is parallel to the cliff face (Trenhaile,1987).

Many workers have suggested that shore platforms around the coasts of the British Isles are being exhumed, and are partly inherited features of an earlier period when sea level was similar to that of today (Trenhaile, 1983a). The origins and the age of such platforms have been the subject of some debate, in particular on the western coast of Scotland (Walker et al., 1992). In the geological past, the sea has sometimes stood at higher levels, indicated by the presence of high rock platforms above present sea level, and sometimes at lower levels, on coastlines now submerged beneath the sea after sea level rise. High level rock platforms have been described from many parts of Scotland, but they are best known from the west coast (Walker et al., 1992). Submerged rock platforms have been reported at various other localities around the Scottish coastline (Eden et al., 1969; Browne and Jarvis, 1983, and Hall, 1989). According to (Gill, 1972) platforms cut in very hard rocks must be inherited from a former sea level, whereas shore platforms cut in fairly weak rocks are contemporary features. The study of the relationships between platform morphology in weak rocks and various aspects of the morphogenic environment emphasises the degree to which shore platforms have been able to adjust to the present level of the sea (Trenhaile, 1980). Shore platforms cut at present sea level may have been inherited, but they have subsequently become completely adjusted to the contemporary environment (Trenhaile, 1987). Whether this is actually the case depends upon the rate of cliff and platform erosion. On the western coast of Scotland, a well-developed rock platform eroded in solid rock in the Oban and eastern Mull area is believed to be Holocene in age (Wright, 1928); however McCallien (1937) has suggested that the Holocene epoch is too short for its formation and instead proposed a pre- or inter-glacial age. Sissons (1974b , 1976b) suggested that the Main rock platform of west Scotland can be correlated with the Lateglacial erosional feature in the Firth of Forth which declines in altitude eastwards from about 0 m OD at Grangemouth to -18 m OD near Berwick, both having been formed during the cold climate of the Loch Lomond Stadial.

In the study area the morphologies of the intertidal and subtidal platforms have been studied in order to examine two main problems:

a - The relationships between platform morphology and environmental factors.

b - The suggestion that the platforms are mainly landforms inherited from the last interglacial period.

Glacial periods have occurred on several occasions through geological time. Each glacial period was followed by either a long-term warm period called an interglacial or short term warm period called an interstadial. During the Quaternary period the sequence of glacial and interglacial events occurred at fairly regular intervals of 100,000 years for 1 to 2 million years (Pethick, 1984). The main point of interest is during which of the interglacial periods did the sea level stand at more or less a similar position to that of the present-day. Although many glacial events occurred during the Quaternary, only the last three glacial periods and their four interglacials, including the present Holocene, have been dated and have been given names depending on the locality in which they have been recognised (Pethick, 1984). Bloom et al. (1974) suggested at least eight sea level peaks over the last 150,000 years of which only two, at 130,000 and 145,000 years , were above the present sea level. Thereafter it is more likely that at the Ipswichian interglacial period sea level stood more or less similar to the present-day sea level(Bird, 1993). In this Chapter of the thesis, the term interglacial will be applied to the Ipswichian interglacial which occurred about 130,000 years ago.

5.3 Study methods

Two approaches can be used to investigate the age of rock platforms. First, process measurements may be carried out to determine present-day rates of change of platforms but , because changes are slow (often less than a millimetre per year), it is seldom practical to adopt this approach. Secondly, morphological studies to examine the geometry of the platforms may be undertaken to establish how well they compare with other platforms formed at sea level in other parts of the world.

Studies of platform geometry help to determine how many platforms actually occur in an area and at what elevations. In this work the second approach has been adopted: the study of the morphology of the rocky platforms around St Andrews Bay was carried out by using both topographic and hydrographic surveying techniques.

The topography of the area of study was surveyed between July 1991 and September 1992, using three different techniques. The methods used were:-

I- Echo sounding

II- Side scan sonar

III- Theodolite survey coupled with geological maps and aerial photographs.

I Echo sounding

These surveys were conducted along parallel track lines between St Andrews and Cambo Sands, which together gave approximately 45 km of echo sounder tracklines in the study area (Fig. 5.2) . The track lines were oriented at right angles to the coast extending from above the Low Water Mark to the edge of rocky outcrops on the sea bed, about 2 km offshore. To obtain the bottom profiles a Lowrance X 15M echo sounder, with a 192 kHz transducer, was deployed from North Point, a 7m work boat. The profiles were surveyed at a

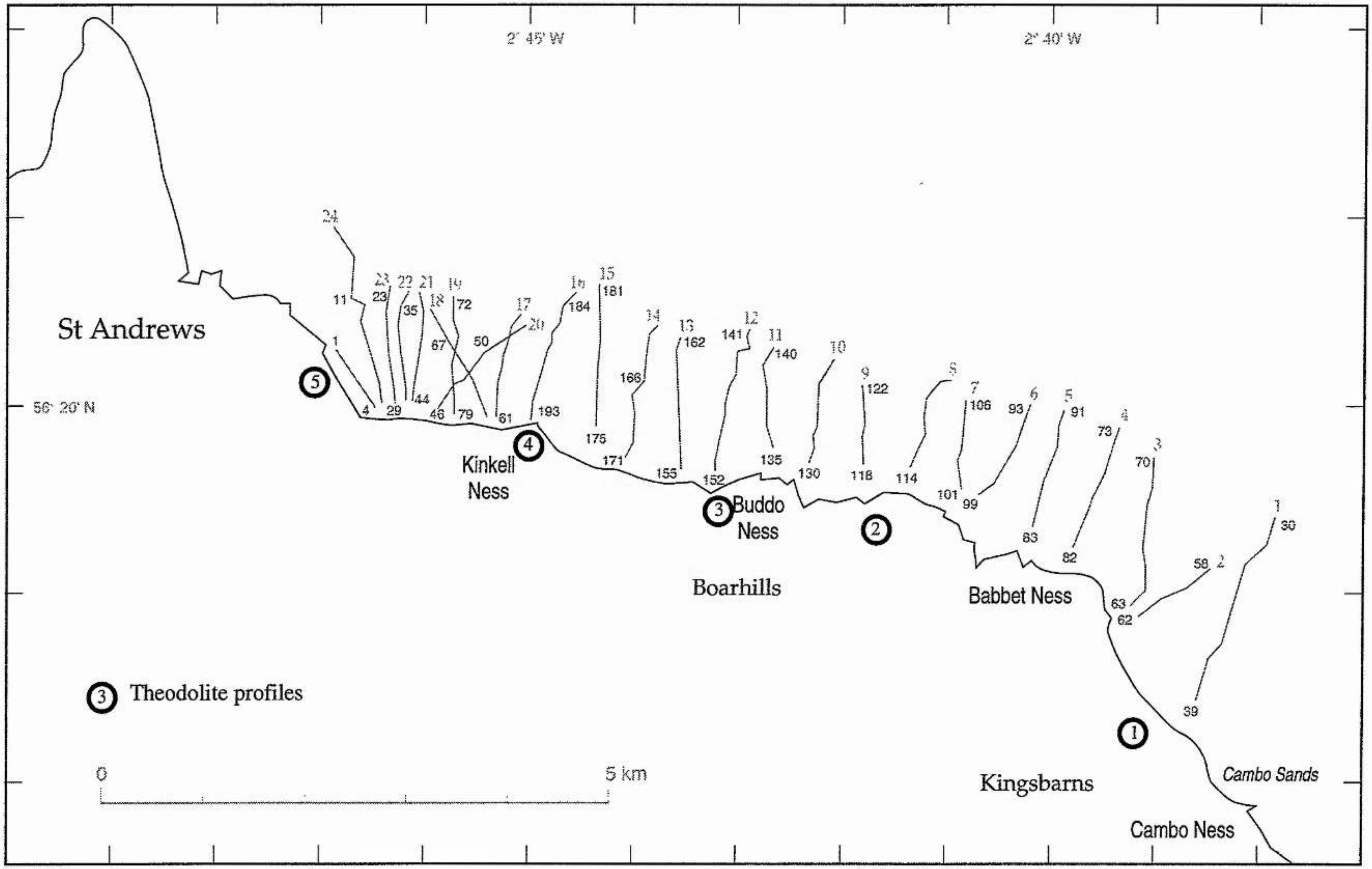


Fig. 5.2 Echo sounding profiles and Theodolite profiles.

speed of about 2 to 3 knots and position fixes were obtained at frequent intervals from a Decca Navigator Mark II (accurracy ± 0.1 nm). The echo sounder system does not provide a direct measurement of depth, but calculates a value from the recorded two-way travel time of the sound beam through the water column to the bed and back to the transducer. The echogram produced yields a continuous profile of the sea bottom relief along a charted path. Echo sounders operate at high frequencies(30-200 kHz) and are effective in penetrating the water column but in unconsolidated sediments the signal is quickly absorbed and does not penetrate much of the sediment (Williams, 1982); consequently the reflected echo originates from the sea bed/sea interface.

The data are normally recorded on graphic strip chart-type profiles. All depths are recorded in metres, with the actual figures displayed representing the distance from the transducer to the sea-bed. Consequently, for bathymetric analysis, all values obtained must be corrected for both tidal variation and the depth of the transducer beneath the water surface.

A series of profiles have been drawn from the echograms to illustrate the morphology of the platform. The profiles have been classified into different groups representing smooth, concave, linear, or stepped profiles (Appendix 7).

II Side scan sonar

A Waverley Sonar 3000 side-scan sonar system was used to survey 40 km of tracklines in conjunction with the echo sounder technique (Fig. 5.3). This system was developed in the late 1940s from experiments using echo sounder transducers tilted away from the vertical. The dual-channel Waverley has a high frequency (100 kHz) transducer which takes the form of a streamlined tow fish and emits sound pulses directed to the bed on either side of the ships track. These sound pulses are transmitted in a beam pattern which is narrow in the horizontal plane and broad in the vertical plane (Klein, 1985). The beam is angled about 10° down from the horizontal axis to direct the acoustic beam along a strip of the sea floor. The image of a specific target is build up by laying down successive scans of the sonar on a wet paper recorder to produce a sonograph. The side-scan sonar system has the

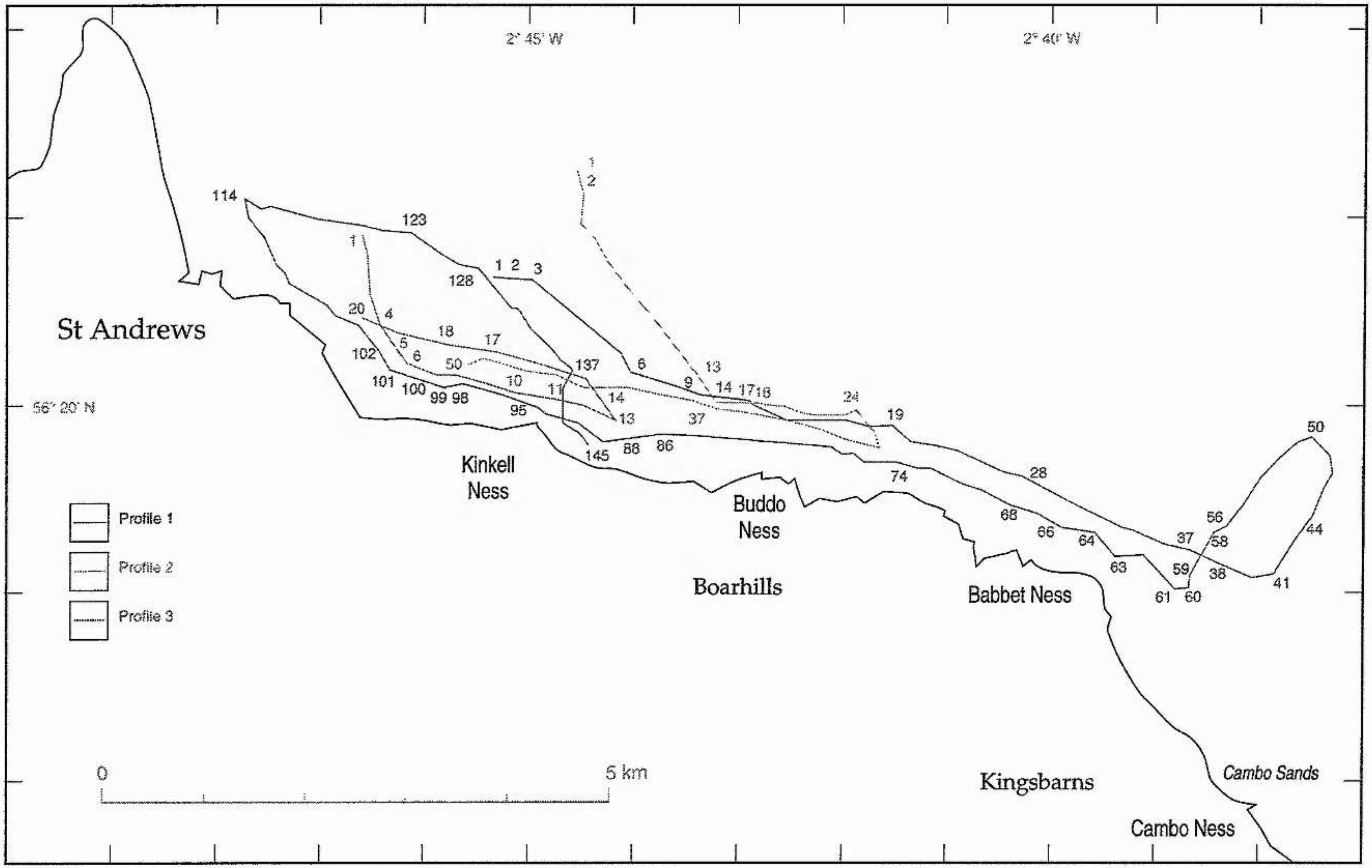


Fig. 5.3 Location of side scan sonar profiles.

ability to produce an acoustic image of the bed, extending up to 300m on either side of the tow fish. This system can thus be used to locate features with positive and negative relief, such as rock outcrops, sand waves, channels and wrecks and also to distinguish between major sediment types (D'Olier, 1979 ; Fish and Carr, 1992). These facilities made it an important underwater remote-sensing technique. Williams (1982) suggests that the vessel speed should be constant throughout the survey and the maximum speed should be between 2 knots to 5 knots, but 2 knots is best for minimum sonograph distortion parallel to the trackline. The survey in St Andrews Bay was carried out near high tide when tidal currents had minimum velocities. The ship position was recorded with a Decca Navigator system which, under optimum conditions, gives a fix accuracy of ±0.1 nautical miles. A 150 m range (e.g. to either side of the tow fish transducer) was adopted for most of the traverse lines with some traverses using the 75 m range to provide high resolution of bed features. The tow fish should be towed above the sea floor at 10 to 20 percent of the range scale used (Williams, 1982). If the 150 m range is adopted this would be equivalent to towing depths of 15 to 30 m above the bed. However, due to the shallow depth in study area ranging between 15 and 20 m, the transducer was towed at a depth of 2 m below the water surface. The interpretation of the sonographs depends on distinguishing between backscatter levels represented as tonal intensities. The examination of records ideally from well known sea floor areas where sediment samples, bottom photography or diver observations are available will correlate between the bed material and the variations in tonal intensities recorded in the sonographs. This ground information is vital for sonograph interpretation. Many of the samples collected for the grain size study (see Fig. 3.1), were used in order to correlate the resultant sonographs. Williams (1982) pointed out that the backscatter intensity is dependent largely upon the type of material and porosity of sediment on the sea bed , as well as the bottom topography. He further suggested an inverse relationship between acoustical impedance and sediment porosity. High porosity sediment (e.g. clays, 75 percent) has a relatively low impedance and low reflectivity and appears light on sonographs; whereas low porosity material (e.g. medium sand, 40 percent) has higher impedance, higher reflectivity, and appears as darker tones. Leeder (1982) revealed that the porosity of a sediment is a function of grain shape, the degree of sorting, and the degree of sediment compaction. These factors can cause sediments of different sizes to be similar in porosity and therefore have very similar tonal shades (Williams, 1982; Fish and Carr, 1992). The majority of the tracklines were run within 2 km of the coastline between St Andrews and Cambo Ness (Fig. 5.3). Most of the sonographs obtained from the side-scan sonar survey are of excellent quality. The records have been examined to determine the structure of the subtidal platform and its roughness. In addition, calculation methods have been applied according to Klein (1982) in order to measure the size of targets and their distances from the vessel's track. For example, the height of objects above the sea floor can be calculated from side-scan sonar records if the length of the acoustic shadow is known, along with the height of the tow fish above the sea bed.

$$H_t = \frac{L_s \times H_f}{R_s + L_s}$$

The horizontal distance from the target to the vessel track is a slant range and does not represent the actual horizontal distance. The true horizontal offset can be calculated from the equation $R_h = (R_s^2 - H_f^2)^{1/2}$ referring to (Fig. 5.4).

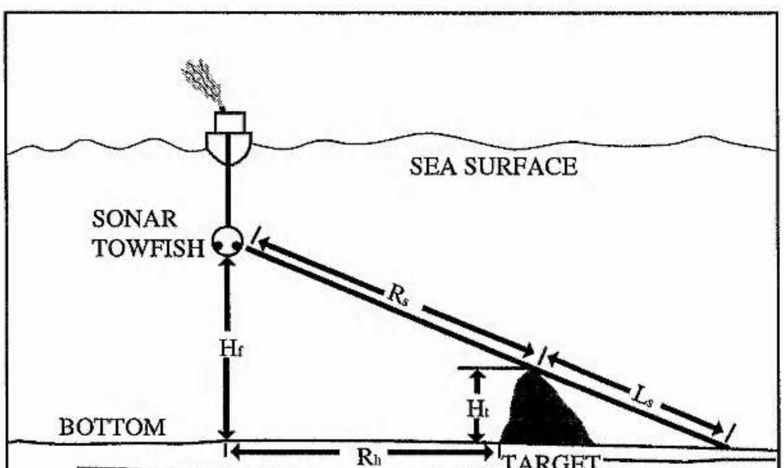


Fig.5.4 Side Scan Sonar Geometry.
H_t - Target height
L_s - Acoustic shadow length
H_f - Towfish height above bottom
R_s - Slant range to target
$R_s + L_s$ - Slant range to end of shadow
R_h - Actual horizontal distance

A diagrammatic representation of the offshore geological and sea bed sediment features as interpreted from side-scan sonar is presented in (Fig. 5.5).

III Theodolite surveys coupled with geological maps and aerial photographs

In order to examine the traverse profile gradient of the intertidal rock platform, a series of theodolite surveys were carried out at selected locations along the coast using a Wild TC1000 Theodolite. At each site the theodolite was set up above the High Water Mark and a profile of the platform measured to the Low Water Mark. Heights and distances were recorded relative to the theodolite with survey points defining points of interest along the traverse, e.g. prominent rock layers.

The morphology of the intertidal rock platform has been studied by using both aerial photography and the Ordnance Survey Maps in conjunction with field observations. Rock structure has been traced from both the 1 : 10,650 Geological Survey Map (Sheet No. 51) and high resolution aerial photography taken in 1990 in order to find if there are any similarities in structural elements of the rock platforms which could possibly extend from the intertidal zone to the subtidal zone.

5.4 Intertidal Platform

a) Field observations

Most of the intertidal platform is of very similar rock types and contains many features which indicate a long and complicated geological history. The striking features which are present along the coast of the study area are the raised beaches, the sea stack of the Maiden Rock (Plate 7), several volcanic vents, such as the Rock and Spindle. Also there are erratic blocks which were left behind by ice during the Quaternary glaciation of the region. The sedimentary rocks are folded and faulted showing both plunging anticlinial and plunging synclinal structures (Plate 1 and 4). Consequently, the individual strata show a wide ranges of dips and strikes, with near horizontally bedded sediments in some places (e.g. Kingsbarns GR. 603125) (Plate 8) and vertical beds at other positions (e.g. Kinkell Braes GR. 525158)

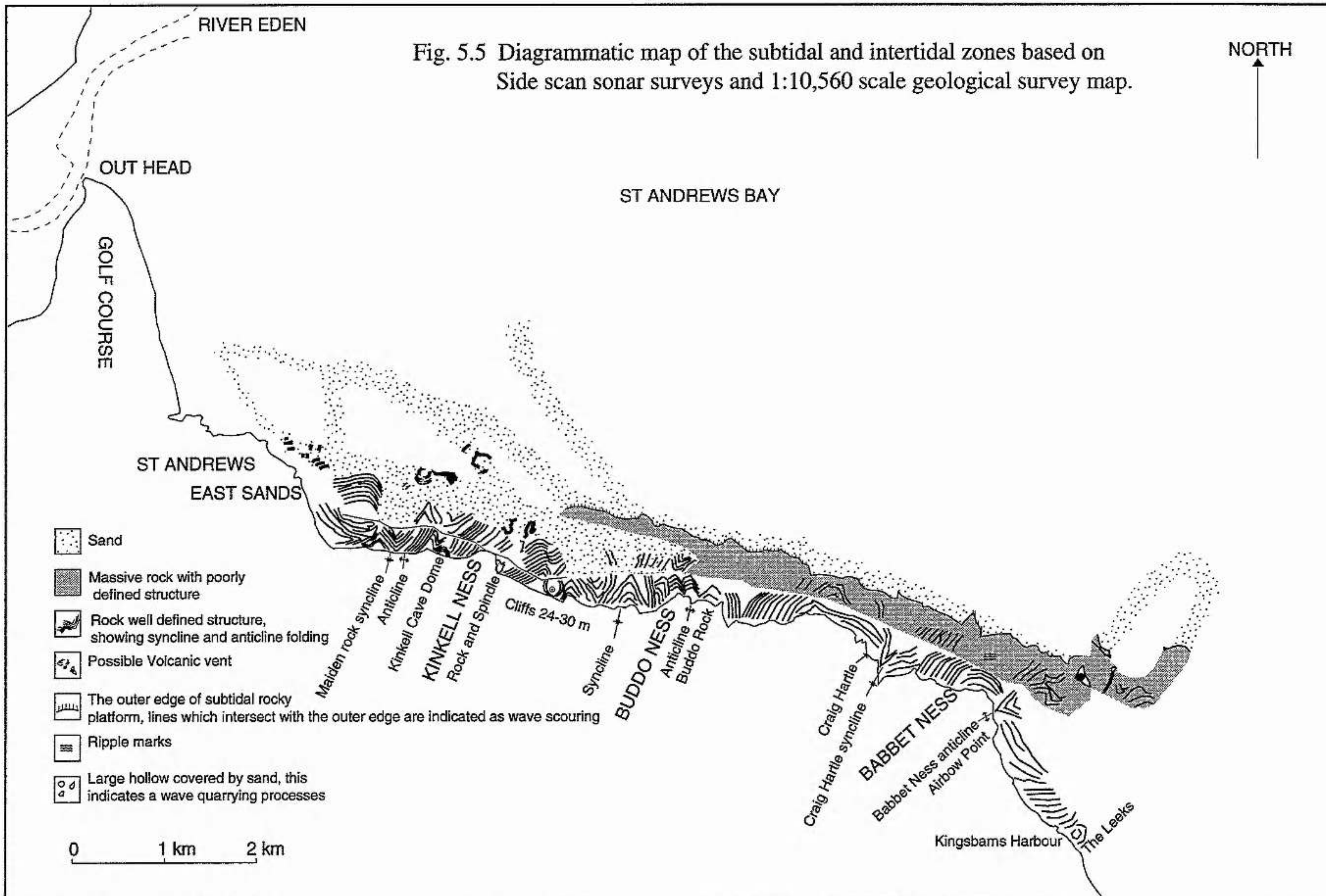


Fig. 5.5 Diagrammatic map of the subtidal and intertidal zones based on Side scan sonar surveys and 1:10,560 scale geological survey map.

(Plate 9). To a distance of 450 m from East Sands towards the east, the strike of the rocks is almost parallel to the cliff (east-west). Before reaching the promontory of Kinkell Ness for about 500 m there is the Kinkell Cave Dome. This structure is a doubly plunging anticline composed of sandstone and can be seen from the cliff top. About 100 m from this feature to both east and west the beds strike normal to the coastline (north-south). The geological structure of the platform thus has considerable impact upon its morphology, as will be discussed later.

The cliffline which backs the platform is, in some places, about 15-20 m high and very steep, as for example at the south end of East Sands beach. Here the high water mark reaches very close to the base of cliff. At first sight it seems that the present sea is responsible for the production of the steep cliffs but the weathered state and mass movement of material at the base of the cliff suggests that it may be a fossil cliff. At some places at high tide the water does not reach the cliff base. For example, at Buddo Ness about 3.5 km from East Sands, there is a subaerial platform up to 100m wide between the high water mark and a degraded cliff-line (Plate 10). This is believed to be a postglacial raised beach. In the vicinity of Buddo Ness the platform does not join the old rock cliff at a sharp angle; the platform-cliff junction is usually completely hidden by heavy blocky scree and the fossil cliff is highly subdued and vegetated.

In addition, it is noted that as well as the large beaches such as East Sands and Cambo Sands which break up the platform, there are also several very small sand and shingle patches at the top of the rock platforms between St Andrews and Cambo Ness.

b) Theodolite Survey

The gradient of the platform may be determined from the height difference between the High Water and Low Water Mark. The gradient measurement at five locations from south to north along the coast revealed that, in general, the intertidal platform is low and rough, slopes very gently seaward and its surface has a different morphology from one location to another (Fig. 5.6).

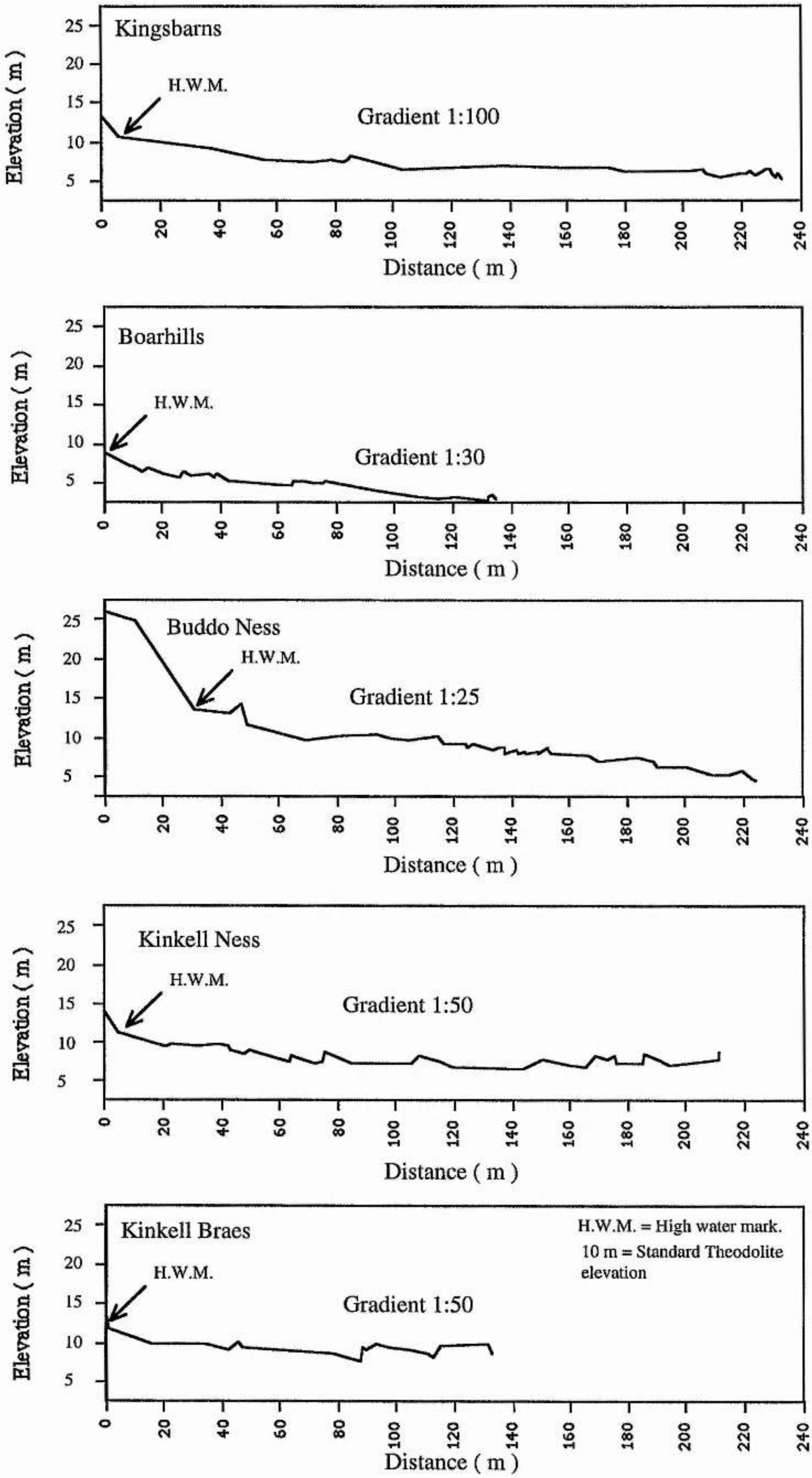


Fig. 5.6 Theodolite transects of the intertidal zone.

At Kingsbarns (Profile No. 1), the morphology is more or less smooth compared with the other locations and the beds strike nearly parallel with the coastline and dip towards the north. Here, the rocks are composed of sandstone alternating with limestone and shale. The platform is characterised by a very low gradient 1 in 100.

At Boarhills (Profile No. 2) it is clear from the Ordnance Survey Map (Sheet No. 51, 1:10,650) that the sequence of the beds is interrupted by a fault which crosses the rocky platform to the south of Craig Hartle which trends in a W.N.W direction nearly along the line of H.W.M until it strikes out to sea on the east side of the Boat Harbour north of Boarhills. The rock types are sandstones intercalcated with subordinate limestones and shales. This area is characterised by a gradient of about 1 in 30, the intertidal surface having a high degree of roughness . The beds dip towards the north and west and strike nearly parallel to the coastline.

At Buddo Ness (Profile No. 3) the platform gradient is about 1 in 25 with high roughness morphology. Again, the stratigraphical sequence at this part of the coast is somewhat obscured by faults. One of these faults is the Buddo Ness Fault which trends along the middle of the platform to the east of the Buddo Rock. In fact, the irregularity of the platform surface can be attributed to this fault zone. Sandstone is the dominant rock type intercalated with thin layers of black shale. The rocks strike normal to the coastline and the beds dip towards the west.

At Kinkell Ness (Profile No. 4) the measured profile trends south-north from H.W.M. to L.W.M., about 365 m east from Rock and Spindle. The intertidal platform slopes gently towards the sea with gradient of about 1 in 50. Here the surface of the platform is characterised by a high relief. This roughness can be attributed to the presence of a group of basalt dykes which trend in general east to west. The rocks (sandstones, shales and ironstones) strike normal to the coastline and the beds dip towards the north-west and west at 5° to 10°.

At Kinkell Braes (Profile No. 5) the gradient at this location is 1 in 50 and the surface of the intertidal platform is characterised by low relief. The rocks strike parallel to the cliff face and the beds dip in land. The dominant rock type is sandstone alternating with shale.

The width of intertidal platform varies along the coast from location to location. For example, at Kingsbarns (Profile No.1) the width is large compared with the other locations. It is 234 m from High Water Mark to Low Water Mark. 3 km west of Kingsbarns, at Boarhills (Profile No. 2), the width of the intertidal platform has decreased to about 130m . At Buddo Ness (Profile No. 3) it is wider again (224 m) , similar in width to that of Kingsbarns. At Kinkell Ness, about 2.5 km from East Sands (Profile No. 4) the intertidal platform width is more than 200 m. About 200 m from East Sands (Profile No. 5) at Kinkell Braes, the intertidal platform is narrow, only 130 m in width.

5.5 Intertidal platform interpretation

It is very clear from the weathering state of the cliffs which back the intertidal platform that the present sea is not responsible for their formation. The presence of a low raised beach 7-9 m above OD at Airbow Point, between Babbet Ness and Kingsbarns, and at Buddo Ness confirms that this cliffline belongs to a period of time when sea level stood up to 5m higher than at present. The steep cliff which fringes the bay from St Andrews to Kinkell Ness (Plate 11) is continuous with the above cliffline and is attached to the intertidal platform. It most likely also formed at the time of the maximum postglacial transgression and has only been slightly modified since that time and is largely inactive today.

The study area is characterised by a macrotidal environment. The main spring tidal range is 4.7 m and the main neap tidal range is 3.2 m. Usually there is a positive relationship between tidal range and platform gradient (Trenhaile, 1978). However, the small differences in gradient between the profiles which have been obtained from the intertidal platform at different locations along the coast of study area cannot be attributed to the tidal range factor which is virtually constant along the coastline of North East Fife. The platform gradient data for the Fife profiles have been taken at cross sections normal to the coastline. These data have been plotted on the platform gradient diagram of Trenhaile (1987) in order to afford comparison with those gradients reported for other areas of the world (Fig. 5.7). The very low gradient may be due to weathering processes when sea level was lower than at present and later modified by the sea. It is most likely that during Loch Lomond Stadial, sea level was lower than the present level by 20 m (Chapter 2). Therefore, a large area of St Andrews Bay was exposed. Then, both the intertidal and subtidal zone was subject to periglacial processes such

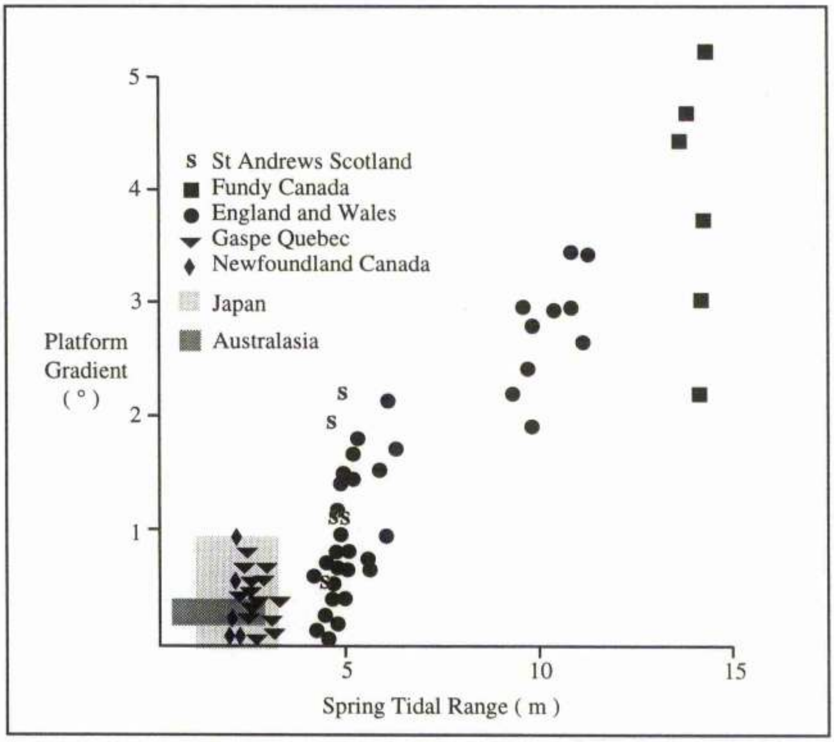


Fig 5.7 Plot of platform gradient against tidal range. Each point is a regional average representing numerous surveyed profiles including St Andrews Bay data (modified after Trenhaile, 1987).

as frost shattering. In fact, frost shattering or wedging of mixed, relatively hard (e.g. basalt, sandstone) and soft (e.g. shale) rock types can cause the differences in platform morphology between these profiles or even at different points in a single profile. However, the intertidal rocky platform along the coast of East Fife provided obvious evidence of last changes in the relative levels of land and sea since 6,000 years, ago this will discussed in detail in the conclusion.

The width of the intertidal platform is the factor which defines the degree of development of any platform. In a uniform rock type one may suggest that, in general, the width of the platform depends on such factors as gradient, wave energy, tidal range and age since initial formation. The width of the intertidal platform along the north coast of east Fife varies from place to place. Because the age and tidal range are presumed to be constant in the local region it is clear that other factors play a significant role in determining the width of the intertidal platform in St Andrews Bay. Even the intensity of wave action on platform width is difficult to adopt as an explanation for the variation in width since the wave climate along this section of the coastline of the study area is more or less constant (Ferentinous and McManus, 1980). The only factors considered to determine the width of platform in the study area are variations in lithology and geological structure. The solid rocks in the area comprise alternating sequences of different hardness, such as sandstone, shale and limestone. Further the area is characterised by fold and fault structures. This results in changes in strike and dip of the beds. One can thus use these factors to explain the selected profiles as follows:

(1) Kingsbarns Profile No. 1

The width of intertidal platform is large, it reaches about 230 m. The rocks in this area consist of alternations of sandstone and mudstone. The wide platform can be attributed to the nearly horizontal beds of sandstone. Here, the rocks strike nearly parallel to the coastline with low angle and the beds dipping towards the north. The platform surface appears as a flight of low terraces, which have formed along the bedding planes of the more resistant beds of sandstone. The steps appear where the weaker, thin beds of shale along the bedding plane have been eroded by wave attack.

(2) Boarhills Profile No. 2

The width of the intertidal platform is less than at Kingsbarns. It is clear from the field observation that the bedrock in this area is inclined towards the sea (north) and north west, with the strata dipping at 14° - 20°. The platform is narrow (133m) in this area because the beds are dipping seawards and striking parallel to the coast. This presents a wall of resistant sandstone bed to the incoming waves. Here, platform and cliff erosion is quite slow and thus the platform is narrow.

(3) Buddo Ness Profile No. 3

Westward from Boarhills, at Buddo Ness, the area has a 230 m wide intertidal platform. Here the beds dip at about 30°-40° towards Kinkell Ness (west), and strike perpendicular to the coast. This causes the weaker strata of shale to be exposed to the wave attack while the more resistant strata of sandstone stand as 'ribs' above the mean level of the platform surface.

(4) Kinkell Ness Profile No. 4

The intertidal platform is 212 m wide. The dip of strata is between 5° towards the north-west and 10° towards the west. The strike of the rocks is perpendicular to the cliff face, so the weaker strata of the shale are protected by harder strata of the sandstone offshore. Here the continuity of erosion depends upon removal of the more resistant material. The weaker beds in the platform form troughs, which run towards the cliff between ridges of more resistant rock. The development of this "washboard" relief eventually reduces the wave energy reaching the cliff base, particularly when the waves approach the coast obliquely.

(5) Kinkell Braes Profile No. 5

About 200 m from East Sands at Kinkell Braes, the intertidal platform is narrow (about 135 m in width) because, in this area, the strata dip landwards at between 24° and 44°. The strata strike parallel to the cliff face. Here the weaker beds of the intertidal platform become inaccessible to wave attack unless the projecting beds of more resistant rock are removed. The platform erosion is therefore slow and the platform is narrow.

5.6 Subtidal platform

Echo sounder data

A total of 24 echo sounder profiles have been obtained across the subtidal platform between East Sands and Cambo Ness (Appendix 7). Each profile was taken as a traverse normal to the coast (Fig. 5.2). The average platform gradient has been determined from these profiles assuming that the course steered was a straight line between the start and end points of the profile. Thus, ignoring small deviations as shown on Profiles No. 2 at Kingsbarns and Nos. 4 and 5 at Babbet Ness, which may represent errors in position fixing, the general gradient has been measured from a position close to the Low Water Mark to the point where the profile slope meets or intersects the sandy bed. In general, all of the profiles from Kingsbarns to the East Sands showed little variation in gradient. All profiles measured sloped gently and had a little variations in gradients. However, the gradient varied at different points on a single profile.

Echogram data show very clearly the point at which the subtidal rocky platform intersects the offshore unconsolidated sea bed on each profile. This point is quite variable in depth along the coast. For example at Kingsbarns (Profiles 1 and 2), rock outcrops to a depth of 16 m from present sea level. At Babbet Ness (Profiles 3, 4 and 5), rock is not present below 14 m. As one approaches St Andrews the lowest part of the rock outcrops become progressively shallower, being 13 m at Buddo Ness and only 10m between Kinkell Ness and the East Sands.

Despite their overall similarity in gradient the echograms reveal differences in morphology between the profiles. For example, at Kingsbarns, Profile 1 is concave and lacks a low tide cliff or scarp at the front edge of the subtidal platform, but there is a rampart in the middle of the profile. Profiles 2 - 6 are characterised by moderate slopes with a high relief roughness. The most interesting profiles, 7 and 8, are characteristically steep with rough surfaces which end with a 3 m high scarp in deep water. The adjacent Profiles 9 and 11 are characterised by very flat and smooth beds, whereas Profile 10 is gently sloping with high roughness relief. Profiles 12 - 15 reflect a moderately rough surface and are gently sloping. Profiles 16 - 19 show low roughness and there is no clear indication of where rock features meet the sea bottom sediments. Profiles 20 - 24, near St Andrews, are very interesting because they mostly reveal a cliff at low tide level and are characterised by low gradients with moderately rough surfaces.

Side-scan sonar data

In the study area parts of rocky platform which are submerged throughout the tidal cycle are considered to be the subtidal part of the platform. The side-scan and echo sounding surveys were made from close to the Low Water Mark to the outer edge of subtidal rocky features. The first side-scan survey began at site 1 about 1.4 km offshore from Kinkell Braes (Fig. 5.3). This survey line trended onshore before being re-directed to follow a route parallel to the coastline then comprising a profile from the East Sands to Cambo Ness and back again to the East Sands. It has 145 position fixes. Over the first part of the trace as far as site 6 the record shows a light trace(low tonal intensities, low back scatter levels) which is interpreted as a flat sandy bed (note; small ripples have been observed in this location during the collection of sediments samples, but these are too small to be resolved on the sonographs). Scattered dark spots on the image between sites 1 and 6 are most probably suspended seaweed which was noted to be present in the water column during the survey.

From site 6 , about 925m offshore from Kinkell Ness, a well defined strong echo represents the outer edge of subtidal platform. This irregular edge extends parallel to the coastline from Kinkell Ness to site 44 (located about 2.0 km offshore from Kingsbarns). On survey lines run closer to shore well defined geological features are revealed. Several interesting features are recognised in the subtidal zone:-

i) At site 28 the outer edge of rocky platform, located about 735 m offshore from Babbet Ness, revealed a highly irregular edge. It is clear from the wave scouring at the base of this edge that the irregularity of the outer edge could be attributed to marine erosion. Moreover, the large hollow, which is probably covered by a veneer of sand may be formed along a line of weakness , that this part of rocky platform could be densely jointed. This will facilitate the waves quarrying processes (Plate 12).

ii) On the same track line between site 37 and 38, the sonograph shows a striking feature which appears to represent a volcanic neck. It is semicircular in shape and measures 90 by 95m. The isolated rock in the middle of the feature is believed to represent the volcanic plug. This feature, which was also noted on the return traverse between sites 58 and 59, is located 1.4 km offshore from Kingsbarns (Plate 13).

iii) On the starboard channel between sites 63 and 64 are small bedform features ripple marks. These are large enough to be picked out by sonar. These features are located 360 m offshore from Babbet Ness and occupy an area of at least 20 m^2 (Plate 14). Orientation of the crests parallel to the shore suggest that they may be largely wave induced. The absence of more widespread small megaripples marks has obvious implications concerning sediment transport in the area. At site 63 the survey range was adjusted to cover 75 m per channel and traverses were run parallel with and more close to the coastline.

iv) Between sites 66 and 68 the sonograph revealed the presence of features which give an alternating pattern of dark tonal intensities (strong echo) and light tones, (the latter being an acoustic shadow). It is clear that the rocks here strike normal to the coastline and are likely to be composed of alternating hard strata (sandstone) and relatively weaker strata (shale). It is suggested that marine processes have eroded the weaker strata while the more resistant beds stand up as ridges. These ridge will blocks the sound from reaching part of the sea bed, consequently acoustic shadows will produce light tonal intensities on the sonographs.

v) The sonographs continue to show strong echoes without well-defined structures until the survey line approaches Kinkell Ness where there is evidence that the rocks are folded (Plate 15). In some places elongated ridges trend oblique to the traverse line as, between sites 88 and 95, close to Low Water Mark in front of Kinkell Ness. At site 102 the record also shows synclinal structures present on the sea bed.

The second side-scan traverse started about 2.4 km offshore from Kinkell Ness and it had 50 position fixes sites along a length of 8 km. The sonograph of the first part of the survey to site 14 shows uniform reflections which probably represent a flat sandy bottom. The rest of the survey traverse parallel to the coast between Buddo Ness and Kinkell Ness is indicative of consolidated rocks without well-defined structures.

The third side-scan profile runs parallel to the coast between East Sands and Kinkell Ness, beginning about 1 km offshore from East Sands. The range of the survey is 150 m per channel. From site 1 to site 4 the area is characterised by a uniformly flat sandy bed (Plate 16). Between site 5 and site 6 the sonar record shows the same folded rocks, which were seen on the first profile at site 102. There are many well-defined features in the area covered by the survey line from site 10 to site 20, which is located 600 m offshore from Low Water Mark at Kinkell Ness. At site 10, (on the third side-scan profile) the sonograph shows well-defined bedded rock structure; from site 11 to site 14 the reflection indicate flat sandy beds and sporadic fragments of bedded rocks. Between site 14 and site 17 the sea bed is generally of a uniform sandy bottom, with some strong echoes indicating patches of rocks protruding through the sand. A dome - like structure with a diameter of 280 m is recorded between sites 17 and 18 650 m offshore from Kinkell Ness. This is believed to be a second volcanic neck which has a circular shape and measures about 120 by 150 m.

Along the irregular outer edge of the subtidal rocky platform there is frequently a hole-like depression seaward of the rock outcrops (Plate 12). The junction between the rock outcrops and the sandy sea bed appears to take several forms at the edge of the subtidal platform. In places, e.g. profile 1 site 20, a trough seems to parallel the outer edge , whereas on the same profile at site 22 there is no depression at the rock-sea bed sediments junction. On survey profile 2 the trackline is parallel to the outer edge of the rocky platform and this depression can be traced for several kilometres along the rocky edge. This depression feature is interpreted as the zone where the waves break at the foot of subtidal rocky platform. This area marks the breaker zone. The striking features are the hollows which are found at the top surface of the rocky platform. For example at profile 1 site 28 it is evident that these hollows are formed in rocks characterised by joints. Here, wave quarrying must be active.

5.7 Interpretation

In general, data from both the side-scan sonar and the echo sounder surveys have revealed the presence of rocky features offshore fringing the coastline of St Andrews Bay. These rocky features slope gently seaward and finally merge with the sea bottom farther offshore. It appears from side-scan sonar records that the subtidal platforms are very similar in nature to the intertidal platform. Sonograph data reveal that the rock formations are frequently folded (Plate 15) , and in some parts of the rocky platform are covered by sand or seaweed (Plate 16), or cut by a number of igneous intrusions such as volcanic vents (Plate 13). Moreover, the presence of a series of volcanic vents along the coastline from East Sands to Buddo Ness provide evidence that similar features could occur in the area now submerged by sea water. On the other hand, if intertidal and subtidal platforms are considered to be a single feature formed before the last glacial ice covered the area, then it is not suprising that both are very similar in morphology. Moreover, the dimensions and structures of what are interpreted as submerged volcanic neck features are comparable with the volcanic necks which are exposed in the intertidal zone. For example, at site no 37 and 38, the volcanic neck is semicircular in shape and it measures 90 by 95 m, it is clear that similar features occurred in the intertidal zone at Elie. It is has similar shape and more than 450 m in diameter.

The subtidal rocky platform varies in width from locality to locality (all profiles extend beyond the seaward limit of rock outcrop). It is difficult to determine precisely the width in some parts because the platform does not have a well-defined edge. This is especially so where the rock structure, determined from side-scan records, suggests that the beds dip at very low angles which results in the edge being hidden beneath superficial sediments (e.g. the area between East Sands and the Rock and Spindle). An irregular outer edge to the platform is very clear from Buddo Ness to Babbet Ness. At Kingsbarns the subtidal platform reaches a maximum width of about 1.5 km, but it narrows as one approaches St Andrews, so that, between Babbet Ness and Buddo Ness it is between 460-650 m wide.

The depth to which the rocky features are found increases from the East Sands to Kingsbarns. This systematic increase in depth of the outer edge towards the east could suggest that the surface of the submerged platform is tilted in this direction.

The presence of troughs along the steep outer edge of the subtidal platform is indicated on the sonar record by the presence of a light tone followed by a dark reflection from the outer edge of platform. These troughs are most likely to be scour holes which may be attributed to wave action rather than tidal scour because of their orientation parallel to the coast. It is clear that the presence of troughs or depressions depends on the angle at which the bedrock intersects the bottom sediments. For example, a prominent trough develops where the bedrock dips steeply (Profile 7, site 103 at Boarhills). Where the strata are more gently sloping no outer margin, wave-associated scouring seems to exist.

The echogram data show that, although there are surface roughness variations, there is no clear sequence of stepped profiles or linear features separated by sharp breaks in slope. Consequently, it is impossible to identify different platform elements at different elevations in the bay. Indeed, one traverse (Profile No. 1) shows just a simple profile from the intertidal zone to deep water, and suggests that a single feature is present. This is the only traverse similar to those recorded in other parts of the world where fully developed platforms have been identified. Nearly sub-horizontal features are present below sea level but they appear on only two traverses (Profiles No. 7 and 8) and thus show no continuity along the coast, which one would expect if a second or third subtidal platform was present. The reason for the variable profiles in the area is because, at the regional scale, variations in mean platform morphology between areas largely reflect differences in their morphogenic environments. However, in this short stretch of coastline it is more likely that the secondary role of geological factors such as bed dip, discontinuities, thickness and other structural and lithological factors are responsible for the considerable variation in platform geometry within short distances along the coast .

5.8 Discussion

A major point of interest concerning both the intertidal and subtidal platforms is their supposed age of formation. Is the intertidal platform at the present sea level being formed by present-day processes or have they survived through the last glaciation and are only being reworked by the present sea?. The second point of interest is whether the intertidal and subtidal platforms are single features or whether they are fragments of different platforms present at slightly different elevations.

The geological history of the area shows that the study area has been glaciated. Examination of the platforms fringing the coastline of St Andrews Bay shows that these platforms lack direct evidence of subsequent overriding by ice, in the absence of till overlying the bedrock and a lack of striated or ice moulded surfaces. The first impression gained is that this platform has not been glaciated since it was formed. On the east coast, at St Andrews, boreholes show that a low-level rock platform predates the last ice sheet because the platform (and the cliff that backs it) are covered by glacial and lateglacial raised beach deposits (Sissons, 1976a). In the study area it is clear that the intertidal platform passes beneath raised beach deposits at different locations along the coast, e.g. at the southern end of the East Sands(caravan site) also at Buddo Ness and at Airbow Point (between Babbet Ness and Kingsbarns). If the platform had been overridden by ice one would expect the profiles to be smoother and not so serrated as those recorded from the intertidal and subtidal platforms. Moreover, the presence of the sea stack (Maiden Rock) and other features such as the Rock and Spindle and the old cliffline above the intertidal rocky platform is considered to be significant. One would not expect these features to have survived glacial erosion. This suggests that the platform and the backing cliff are contemporary features formed in the last 6,000 years . However, if present-day rates of marine processes are considered, this hypothesis may easily be rejected, since a long period of time when sea level is approximately stable relative to the land is required for the platform to develop. From field observations at high tide, sea water does not normally reach the base of the cliff at places such as Buddo Ness. The presence of strongly weathered and vegetated cliff faces backing the platform and the absence of notches or caves suggest that processes of platform formation are no longer active and that the platform morphology is inherited . In terms of the present study it is of note that, although inherited platforms have been reported from resistant rock coasts (Walker et al., 1992), they have not been reported from platforms cut in fairly weak rocks such as sandstones and shales. It is not an easy task to determine whether a present-day shore platform is totally modern or is a relict platform assigned to a former period when sea level stood at the same level as that of present-day. The difficulty in knowing the origins of the platforms is due to the lack of reliable data regarding the rates of cliff erosion and platform formation. The erosion rates documented or recorded at many locations in Britian are given in Table 5.1.

If intertidal and subtidal platforms are interpreted as a single feature then we cannot expect the development of a wide platform within the last 6,000 years. It is most likely that this platform and associated cliff which fringe the coastline in the study area have been overridden by ice and so both predate the last glaciation . The explanation for the absence of till on the platform, which one would expect to have been left behind after the glacial period, may be attributed to marine erosion since deglaciation as this coast has been exposed to wave activity since glaciers left the area. In addition, the absence of striations is probably due to the subsequent weathering of the rocks.

It is more likely that the intertidal, subtidal and other features (e.g. old cliffline) have survived through the last glaciation. This section of the coast was isostatically-depressed under the ice load during the Devensian period. It is probable that these features were preserved under the accumulation of ice and the sediments associated with it. During and after the deglaciation of the Devensian ice sheet the coastline of east Fife was subject to changes in the relative levels of land and sea (Sissons, 1976a). As the late Devensian ice

Location	Lithology	Erosion rate (m/y)	Interval	Method	Source
UK					
Fourth Bright, Whitby, Yorshire	Upper Lias shale	0.023	1971-1972	Micro-erosion meter	Robinson (1977b)
Tees estuary to Ravenscar, Yorshire	Upper - Lower Lias shale	0.09	1892-?	Maps, Surveys	Agar (1960)
Tees estuary to Ravenscar, Yorshire	Glacial deposits	0.28	1892-?	Maps, Surveys	agar (1960)
Famborough Head, Humberside	Chalk	0.3	—	—	Matthews (1934, p. 10)
Bridlington to Sewerby, Humberside	Glacial deposits	1.8	—	—	Matthews (1934, p. 12)
Holderness, Humberside	Glacial deposits	3.3	—	—	Matthews (1934, p. 11)
Holderness, Humberside	Glacial deposits	0.29-1.75	1852-1952	Maps, Surveys	Valentin (1954)
Withernsea to Easington, Holderness, Humberside	Glacial deposits	10(max)	1974-1983	Cliff top measurements	Pringle (1985)
Norfolk	Glacial deposits	0.9	1880-1967	Maps, Air photos	Cambers (1976)
Sheringham to Happisburgh, Norfolk	Glacial deposits	0.2 – 1.8 0.7	1885-1985 1975-1985	Maps Surveys	Clayton (1989)
Cromer to Mudesley, Norfolk	Glacial deposits	4.2 5.7	1838-1861 1861-1905	– –	Matthews (1934, p. 21)
West Runton to East Runton, Norfolk	Glacial deposits	0.2	1971-1973	Cliff top stakes	Cambers (1976)
Hopton, Suffolk	Glacial deposits	0.8	1880-1950	Maps	Cambers (1976)
Pakefield, Suffolk	Glacial deposits	0.8-0.9	1926-1950	Maps	Steers (1951)
Covehithe, Suffolk	Glacial deposits	0.6-0.9	1926-1950	Maps	Steers (1951)
Southwold, Suffolk	Glacial deposits	5.1	1925-1950	Maps	Steers (1951)
Southwold, Suffolk	Glacial deposits	3-3.3	1925-1950	Maps	Steers (1951)
	Glacial deposits	4.5-13.5	—	—	Matthews (1934, p. 21)
Dunwich, Suffolk	Glacial deposits	1.6 0.9 1.5 1.1 1.2	1589 – 1753 1753 – 1824 1824 – 1884 1884 – 1925 1589 – 1977	Maps Maps,Surveys	Robinson (1980a)
North coast, Isle of Sheppey, Kent	London Clay	0.9-2.2	1864(65)–1963(64)	—	Hutchinson (1973)
Studd Hill, Kent	London Clay	1.5 2 3.4 1.2	1872 – 1898 1898 – 1931 1931 – 1939 1939 – 1961	Maps	So (1967)
Beltinge, Kent	London Clay	0.7 0.9 1.1 0.9	1872 – 1907 1907 – 1933 1933 – 1939 1939 – 1959	Maps	So (1967)
E. Minnis Bay, Kent	Chalk	0.30	1872 – 1938	Maps	May and Heeps (1985)
Grenham Bay, Kent	Chalk	0.08-0.4	1872 – 1932	Maps	Wood (1968,Figure 1)
E. Epple Bay, Kent	Chalk	0.14	1872 – 1938	Maps	May and Heeps (1985)
White Ness, Kent	Chalk	0.05	1842 – 1938	Maps	May and Heeps (1985)
Kingdgate, Kent	Chalk	0.15	1842 – 1938	Maps	May and Heeps (1985)
Hereson, Kent	Chalk	0.23	1842 – 1938	Maps	May and Heeps (1985)
Pegwell C. G. stn., Kent	Chalk	0.05	1839 – 1938	Maps	May and Heeps (1985)
Lydden Spount, Kent	Chalk	0.51	1873 – 1933	Maps	May and Heeps (1985)
South Foreland, Kent	Chalk	0.19	1878 – 1962	Maps	May (1971)
E. Folkstone, Warren, Kent	Chalk	0.36	1873 – 1933	Maps	May and Heeps (1985)
W. Folkstone, Warren, Kent	Chalk	0.09	1873 – 1933	Maps	May and Heeps (1985)
Seven Sisters, Sussex	Chalk	0.51	1873 – 1962	Maps	May (1971)
Birling Gap,Seven Sisters, Sussex	Chalk	0.91 0.99	1875 – 1916 1950 – 1962	Maps Surveys	May (1971)
Birling Gap,Seven Sisters, Sussex	Chalk	0.28-0.98	1951 – 1962	Maps	May and Heeps (1985)
Rottingdean, Sussex	Chalk	0.66 0.13	1873 – 1929 1929 – 1951	Maps	May and Heeps (1985)
Roedean, Sussex	Chalk	0.76 0.76	1826 – 1884 1884 – 1897	Maps	May and Heeps (1985)
Middle Bottom, Dorset	Chalk	0.05	1882 – 1962	Maps	May (1971);May and Heeps (1985)
Scratchy Bottom, Dorset	Chalk	0.46	1882 – 1962	Maps	May and Heeps (1985)
White Nothe to Hamburg Tout, Dorset	Chalk	0.22	1882 – 1962	Maps	May (1971)
Ballard Down, Dorset	Chalk	0.23	1882 – 1962	Maps	May (1971)
Fairy Dell , Dorset	Marls	0.4-0.5	1887 – 1969	Maps, Air photos	Brunsden and Jones (1980)
Vale of Glamorgan, Wales	Lower Lias lime-stone, shale	0.008-0.099	1967 – 1969	Cliff top stakes	Trenhaile (1972)
Ogmore-by-sea to Barry, Wales	Lower Lias lime-stone, mudstone	0.068	1977 – 1985	Stakes	Williams and Davies (1987)
South of Aberaerton, West Wales	Aberystwyth Grits (Greywacke and mudstone)	0.06	1880 – 1970	Maps	Jones and Williams (1991)
Near Aberarth, West Wales	Glacial Clay	0.65	1880 – 1970	Maps	Jones and Williams (1991)
Southeast County Down, N. Ireland	Glacial deposits	0.21-0.84	1834– 1962	Maps, Air photos	McGreal (1979c)

Table 5.1 U.K Coastal cliff erosion rates.

sheet melted in eastern Scotland relative sea level was high, since the eustatic rise in sea level was large and rapid while the land was still isostatically depressed. More probably the lateglacial sea may have simply excavated pre-existing cliffs and platforms removing drift deposits of various types to reveal the ancient rock features in the study area. Later, when the rate of isostatic uplift was greater than that of eustatic rise of sea level, the relative sea level fell. The period of low sea level is probably represented in the lower Tweed valley by a buried channel, significantly not containing till, which appears to be related to a sea level of about -20 m OD (Rhind, 1972). Sissons, (1974b, 1976a) correlated the Main rock platform with the Main Lateglacial erosion features at Grangemouth. This shoreline declines in altitude eastward with a gradient of 0.17 m /km from about 0 m OD at Grange mouth to -18 m OD near Berwick. He suggested that both the Main rock platform and the erosional feature at Grangemouth are of the same age having been formed during the cold climate of the Loch Lomond Stadial between about 11,000 and 10,300 years BP (Sissons, 1974b). During the Loch Lomond Stadial this area was not covered by ice, but sea levels were some 20m below present level (Chapter 2). Consequently, large parts of St Andrews Bay were exposed to subaerial weathering. The severe climatic conditions that accompanied the Loch Lomond Stadial were of major importance because, at that time, climatic conditions were different from today. It has been argued that during the Loch Lomond Stadial, periglacial processes were dominant (Sissons, 1974b; Gray, 1974 and Dawson, 1980). The optimum conditions for frost shattering today seem to be a moist climate and frequently recurring cycles of freezing and thawing. Scotland was both cold and moist during glacial ages (Flint, 1971). The Study area was subject to subaerial weathering and periglacial processes such as frost wedging or shattering on the cliff and rocky platform. Moreover, the rocks have well developed jointing in addition to bedding planes which would make them susceptible to frost wedging. During winter and autumn, the sea was most likely frozen and there would have been no wave activity in the area. During summer and spring the variation in temperature between day and night would have caused water which was trapped within discontinuities in the rocks to freeze during the night and thaw by day eventually resulting in frost shattering.

After this period of low sea level a rapid transgression occurred between 10,000 and 6,000 years BP.This phase of rapid transgression occurred when the rate of eustatic rise in sea level was greater than that of isostatic uplift. This postglacial transgression period was so rapid that there was not enough time for marine processes to modify the area which was previously land. Approximately 500 years after the maximum transgression the rate of isostatic uplift was greater than eustatic rise in sea level. Consequently, three postglacial raised shorelines were formed below the Main Postglacial Shoreline. Since 4,000 to 3,000 years BP sea level has been more or less constant (Sissons, 1976a).

Although the platform is horizontal it is often irregular and does not possess the well developed ramp abrasion profile which is characteristic of platforms formed by marine abrasion. However, the presence of raised beaches on the tops of platforms has been cited as evidence of very slow rates of contemporary marine erosion. The platforms in the study area are considered to be ancient erosion surfaces, inherited from a period when the relative sea level was similar to that of today. Later modification has caused them to become completely adjusted to the contemporary environment.

Chapter VI

Sub-bottom Seismic Investigation

in St Andrews Bay

6.1 Introduction

A record of past environments of St Andrews Bay can be interpreted from the disposition of former shorelines and sediments that are exposed above present sea level on the margins of the embayment (Chapter 2). Complementary evidence of former conditions in the bay are present in the sedimentary record of the deposits beneath the sea-bed. Direct access to the sea-bed sediments is limited. Shallow cores <1 m in length have been obtained at various positions in the bay by the British Geological Survey as part of their regional mapping programme in the late 1960's and early 1970's. In addition, a few boreholes were penetrated to bedrock by the BGS in order to complement the shallow cores and to calibrate geophysical studies in the area (Thomson, 1978; Thomson and Eden, 1977). Geophysical investigations, particularly sparker and pinger surveys, provide by far the most important source of information concerning sub-surface sediments in St Andrews Bay.

The BGS, geophysical survey unit, carried out seismic surveys in the approaches to the Tay and Forth Estuaries between 1969 and 1974. Sparker lines were surveyed generally in a north-south direction with a spacing of about 5 km and were continued into water about 15 m deep at low tide. Accordingly, there is limited information on the nearshore pattern of sedimentation and only a limited amount of data is available in St Andrews Bay for the area west of a line from Kingsbarns (GR. 125605) to Arbroath (GR. 423635). The survey work reported in this Chapter has been carried out in the nearshore zone of St Andrews Bay between Kingsbarns and St Andrews from the low water mark to 4 or 5 km offshore. It thus provides an important link between the onshore and offshore data.

6.2 The seismic method

The seismic method is one of the most commonly used techniques for the remote investigation of sub-surface deposits on land and in the sea, either for site investigations for engineering purposes or prospecting for oil and minerals. Two methods are normally applied, namely seismic refraction and seismic reflection (Kennett, 1982). In general, refraction studies are more commonly used in land-based site investigations whereas the reflection method is well suited for marine surveys. However, deep structures on the continental shelf have been investigated using marine refraction methods and reflection studies are becoming more important for land-based investigations. The study in St Andrews Bay has utilised reflection techniques and accordingly this section will primarily discuss this methodology.

Seismic investigations involve the generation of a sound pulse which travels outwards from a source and a proportion of the sound energy is reflected back to a geophone or geophones (hydrophone) from some interface in the sub-surface sediments. In reflection seismic studies the sound pulse of interest is propagated downwards and the reflected pulse returns to the surface so that the travel path is essentially vertical. From the travel times of the reflected pulse one tries to deduce information concerning the character and thickness of the sediments and rocks through which the sound waves propagated.

In the marine geophysical investigations presented in this chapter the sound source was generated by a high voltage electric spark using a multi-electrode gap and the reflected sound energy was detected by a multi-element hydrophone and then displayed on a graphic chart recorder. The velocity of sound propagation through the sub-surface layers is of fundamental importance in seismic interpretation since this determines the travel times from any reflector. From a knowledge of these factors it may be possible to estimate the thickness and nature of sub-surface sediments and rocks.

There are several factors that control the propagation of sound through sediments and rocks and the amount of energy that is reflected back to the surface from an interface. Sound propagation and velocity are determined by the elastic properties of the material through which the sound is travelling. In part these properties are determined by material density but they also depend very much on the lithology. Thus a limestone and a sandstone may have the same density but if the limestone is a very hard rock it could have a velocity of sound propagation 50% faster than the sandstone. However, within a given rock type there are fairly linear relationships between sound velocity and density (Telford et al., 1976).

The velocity of sound is further affected by the presence of pore spaces within the rocks and whether the pores are filled with fluid. In sedimentary rocks, where porosity can be quite high, and variable, the presence and pressure of any fluid in the pore spaces may be a determining factor in sound propagation. The unconsolidated sediments beneath St Andrews Bay comprise gravels, sands, silts, clays and diamictons largely laid down in the bay since the end of the last glacial phase. A wide range of particle sizes and shapes are thus present which will give rise to variable porosity's and densities and hence velocities of sound propagation. This makes the determination of the thicknesses of layers very difficult unless the seismic records can be calibrated against boreholes records.

A number of other factors also affect sound propagation in rocks and must be taken into account when interpreting a seismic record. No record will be obtained unless sound energy is reflected back to the surface. Within rocks or sediments sound energy is reflected from an interface when the acoustic impedance, which is the product of the density and sound velocity, change across the interface. For most interfaces in the earth's the impedance contrast is small and the amount of reflected energy for normal incidence is modest, from less than 1% to about 5%. Exceptions to this occur at the ocean bottom where up to 45% of the incident energy may be reflected or at the base of unconsolidated sediments on bedrock where up to 40% of the energy may be reflected (Telford et al., 1976). In the marine environment, where the surficial sediments are saturated and a high overburden pressure of water exists, this latter value may be too high. In the earth the majority of the interfaces have an increase in acoustic impedance downwards and give rise to positive reflections. If the contrast is negative, e.g. at the upper surface of a buried peat layer, the reflected energy suffers a phase reversal at the interface and, in the case of gas reservoirs at depth, they may give rise to bright spots corresponding to a high return of reflected energy because of large negative impedance contrasts.

The frequency of the initial sound pulse and its length are important for determining both the penetration of sound waves into the sediments and the resolution of layers within the sediments. High frequency sound waves are absorbed more quickly in sediments than lower frequency ones. Accordingly if data are required from great depths, then low frequency sound sources must be used. The pulse length will determine the resolution obtained since the reflected energy from layers that are closer than the pulse length together will interact and may give rise to an unresolvable signal. Pulse length and frequency are inter-related because a sound source that generates say 5 waves of low frequency sound will be of much longer duration than 10 waves of high frequency sound. Thus, low frequency sound sources give good penetration with poor resolution whereas high frequency sources will give less penetration but better resolution.

Interpretation of seismic records and relating the records to deposits beneath the surface is not easy. Commercial studies often use sophisticated data processing in order to enhance seismic records so that interpretation is made easier and can be carried out more confidently. In the present study the seismic profiles were recorded onto chart paper which shows the amplitude of the reflected energy and no data processing of the records was possible (no post collection analysis of the data was possible). Accordingly, the records are interpreted from the charts which show the patterns of reflected acoustic energy from the sub-surface sediments.

Seismic records show vertical and horizontal zones where the visual pattern of acoustic energy (seismic signature) is similar. This similarity of seismic texture is presumed to reflect a deposit of similar character. Interpretation of the record thus involves identifying and defining areas of similar seismic texture throughout the survey. A stratigraphy of acoustic texture can be constructed for the record and presented in map form to show the distribution of the various units. However, without calibration, particularly by boreholes, these seismic units may be difficult to relate to lithostratgraphic units.

Several factors affect the interpretation of seismic records. Crucially, all data need to be collected with the same energy sources and the same instrument settings to ensure that the same strata will give the same acoustic pattern. In some cases the relationship between acoustic texture and sediment structure is unclear and in other situations different materials can give rise to similar seismic textures. In some North Sea records the acoustic returns suggest strong laminations in the sediments because of the presence of parallel tones but boreholes through the sediments did not detect any structures (M. Paul pers. comm.). Structureless clays might be expected to give an acoustically transparent texture because of the lack of any internal reflectors, whereas structureless gravels might also give a similar texture because many internal reflections may cause interference in the reflected signal and restrict the overall return of energy.

Interpretation of records is complicated further by the presence of multiple echoes from strong reflectors, especially the seabed, but from within the sub seabed sediments. The seabed echo is very strong and is frequently reflected back to the sea surface where it will be reflected back towards the bed. This reflection thus constitutes a second sound pulse delayed by twice the surface to seabed travel time. On most seismic records returns from this pulse, which also includes reflections from within the sediment pile, interferes with and obscures deep reflections from the initial pulse. In these situations it may be difficult to gain information from the sub-surface at a depth within the sediment greater than the two-way travel time equivalent of the water depth.

6.3 Field work

In the marine geophysical investigations presented in this chapter the sound source was generated by a high voltage electric spark using a multi-electrode gap and the reflected sound energy was detected by a multi-element hydrophone and then displayed on a graphic chart recorder. A total of 13 seismic profiles were run parallel with and normal to the southern margin of the bay. The survey tracks are shown in (Fig. 6.1). Positions were located by a Decca Navigator system.

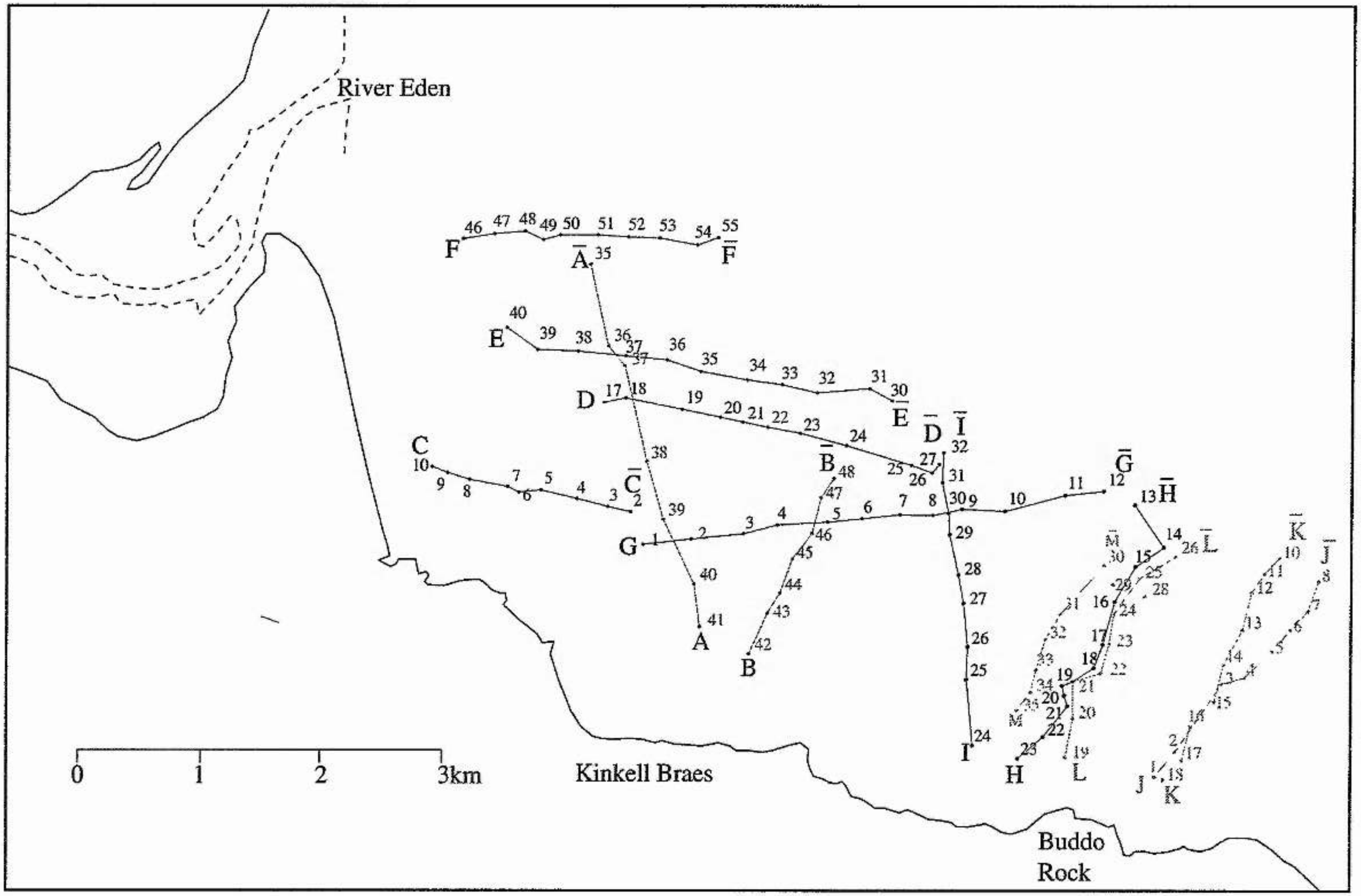


Fig 6.1 Location of sub-bottom profiles.

The multi-electrode source used in the present survey generates frequencies in the range 300 Hz to 30 KHz and with the recording system used permits, a resolution of about 1.5 m and a penetration in suitable conditions of at least 60m (J. Jarvis pers. comm.)

6.4 Previous work

The area seaward of St Andrews Bay has been surveyed in some detail by the British Geological Survey. Whilst much of this was for the investigation of the deeper geological structures the work has also provided valuable data concerning the Quaternary deposits of the area. In the central North Sea, Quaternary deposits reach thicknesses of over 600 m and contain sediments laid down at the start of the Quaternary some 3 million years ago. In the approaches to the Tay and Forth Estuaries deposits relating to only the last phase of glaciation have been identified and it is thus appropriate to consider only these more recent sediments in this study.

The offshore sequences near St Andrews Bay have been interpreted as an evolutionary depositional sequence. Deposits relating to the glacial maximun (Wee Bankie and Marr Bank Formations) are presumed to have been laid down subglacially or proximally in front of the ice margin. These sediments were overlain by glaciolacustrine and glaciomarine beds (St. Abbs Formation) during glacial retreat. These latter sediments have been overlain and reworked during the Holocene to produce the present seabed and near surface sediments (Forth Formation) (Stoker et al., 1985).

In the study area only three stratigraphical divisions could be recognised, as follows: Wee Bankie, St Abbs and Forth Formations. These have been divided largely by their different acoustic properties on seismic records, mainly sparker and pinger (Thomson and Eden, 1977). These formations usually rests unconformably on earlier deposits or on rockhead. The topmost layer of these formation is commonly composed of a modern sediment layer no more than a few centimetres thick.

The Wee Bankie Formation outcrop in the western and southern parts of the Forth Approaches. However, where hollows are present in the sea bed, the Wee Bankie Formation is generally overlain by the St Abbs or Forth Formations. The striking feature of arcuate lines of moraine-like ridges represents the eastern boundary of the Wee Bankie Formation (Stoker et al., 1985) which marks the maximum offshore extent of the late Devensian ice sheet. This formation was deposited as a basal till with the coarse sand and gravel deposited from sub-glacial streams. These sediments are unfossiliferous which reflects a glacial environment. On seismic records it is possible to distinguish the Wee Bankie Formation from all other drift deposits in the area. It is characterised by point source hyperbolic reflections which give the formation a slightly chaotic seismic reflection pattern (Stoker et al., 1985).

The St Abbs Formation rests unconformably above the Wee Bankie Formation or above rockhead where the Wee Bankie Formation is not present. It is defined as brown or pinkish marine Quaternary silts and clays, commonly with dropstones which were deposited from calved ice bergs or winter sea ice. These sediments are interpreted as glaciomarine in origin indicative of an Arctic marine environment. However, the constant acoustic texture of the St Abbs Formation helps to distinguish them from the more irregular Forth Formation (Stoker et al., 1985). On seismic records the formation is transparent, structureless and opaque and only rarely are faint discontinuous sub-parallel internal reflectors observed. The Ostracod fauna recovered from this formation has been correlated with that recovered from the Errol Beds of the Tay which were deposited in late Devensian times prior to the Lateglacial interstadial that started about 13,000 years ago. The St Abbs Formation is interpreted as the lateral equivalent of the Errol Beds (Thomson and Eden, 1977).

The Forth Formation is the uppermost Quaternary unit in the study area. The sediments of this formation show two distinct facies , a sandy facies and a silty clay facie's (Thomson and Eden, 1977), there is no distinct boundary between the two, one grading into the other. The lower part of the formation is characterised by muds whereas the upper part varies from silty sandy muds to coarse pebbly and shelly sands (Stoker et al., 1985). Marine, glaciomarine, fluviomarine and estuarine facies are recorded from the Forth Formation. Erosion at the base of the formation and within the upper part of the sequence implies an oscillatory sea level throughout deposition. The Forth Formation has been subdivided into four seismostratigraphic members. These are as follows:

A- Largo Bay Member

B- Fitzroy Member

C- St Andrews Bay Member

D- Whitethorn Member

The base of the formation rests either conformably or unconformably on pre-existing Quaternary sediments or directly on rockhead. The top of the formation is usually reworked and overlain by a thin layer of superficial sediments. Faunal data indicate a late Devensian to Holocene age for the Forth Formation. The late Devensian is represented by muddy sediments and the overlying coarser sandy units are of Holocene age (Stoker et al., 1985). The Largo Bay and St Andrews Members occur in the extreme west of the area and the Fitzroy and Whitethorn Members have been mainly recognised in the Devil's Hole area. Only the Largo Bay and St Andrews Bay Members will be considered here.

A- Largo Bay Member

The member is named after Largo Bay in the north of the Firth of Forth where the unit reaches its maximum thickness. The base of the member is gradational into the top of the St Abbs Formation. It is composed of silty muds which pass down into laminated silts and very fine sands. The top of the member is composed of soft grey muds which are overlain by grey brown silty muds which mark the base of the St Andrews Bay Member. The upper part of this member was eroded prior to the deposition of the overlying St Andrews Bay Member. This erosion surface varies from a deeply dissected and channelised feature in the Firth of Forth to a more planar surface farther offshore and in the approaches to the Firth of Tay. The Largo Bay Member is believed to have been deposited during the late Devensian Lateglacial period between about 10,000 and 13,500 years BP. Micropalaeontological evidence from bore holes suggests that the deposits were laid down in cold water conditions (Thomson and

Eden, 1977). The climatic deterioration towards the top of the unit marks the onset of the Loch Lomond Stadial about 10,000 to 11,000 - 12,000 years BP. On seismic records, this member is easily recognised as it is acoustically fairly transparent and generally well stratified (Stoker et al., 1985).

B- St Andrews Bay Member

The St Andrews Bay Member takes its name from St Andrews Bay which lies in the southern part of the approaches to the Firth of Tay, where the unit is extensively developed. The St Andrews Bay Member consists of interbedded sands and clays. Channel-fill sediments show a fining upwards sequence from dark grey gravely muddy sands to dark grey silty clays. The base of the member is defined where gravely muddy sands abruptly overlie the soft clays of the Largo Bay Member (Stoker et al., 1985). The top of the unit has generally been reworked to some extent into sand waves and sand ridges. The base of the unit is mainly erosional and varies from irregular to planar in character. In the Firth of Forth the lithofacies variation and faunal assemblage is indicative of a shallow water estuarine environment with climatic conditions cooler than those existing at the present-day, but less harsh than those prevailing during the deposition of the Largo Bay Member. According to Stoker et al., (1985) the St Andrews Bay Member is considered to be of early Holocene age. On seismic records the unit varies in acoustic response from structureless to displaying strong internal reflectors. On sparker records it shows a distinct stratification (Thomson and Eden, 1977).

Seismic stratigraphy is a useful technique which helps to elucidate the geological evolution of the study area. Vail et al., (1977) state that "seismic stratigraphy provides a practical geological /stratigraphical interpretation tool for seismic records". One of the main objects of seismic stratigraphy is to work out the geological history of an area. According to Vail et al., (1977) seismic stratigraphic analysis is based on seismic sequence analysis, seismic facies analysis and sea level changes. Seismic sequence and depositional sequence are both used for the description of sedimentary deposits. The only difference is that the seismic sequence is identified by using seismic evidence (Mitchum and Vail, 1977). The seismic sequence is a time - stratigraphical unit, consisting of a set of genetically-related facies termed a depositional sequence, which are bounded at top and base by unconformities and is identified by using seismic evidence. The depositional sequence is defined by the physical relations of the strata at the upper and lower unconformities and their correlative conformities (Mitchum and Vail, 1977). If the strata are discordant, then there is a physical verification for unconformity. Both concordance and discordance may exist at the upper and lower boundaries of a depositional sequence (Sheriff, 1980). Discordance of strata is the main criterion used in the determination of sequence boundaries and the type of discordant relationship is the best indicator of whether or not an unconformity results from erosion or non deposition. The two types of discordance are lapout (upper as top lap or lower as base lap) and truncation. Lapout is the lateral termination of a stratum at its original depositional limit; this indicates non-depositional hiatuses. Truncation is the lateral termination of a stratum as a result of being cut off from its original deposition limits. Truncation indicates an erosional hiatus unless the truncation is a result of structural disruption (Mitchum et al., 1977). Seismic facies can be defined on the basis of characteristic configurations and the continuity of seismic reflections within a given seismic sequence. Parallel and sub-parallel configuration indicate uniform rates of deposition in a stable basin setting. Divergent patterns result from the lateral variation in deposition rate or progressive tilting of the depositional surface. Prograding patterns are indicate of progressive lateral development of gently sloping depositional surfaces. Deposition in a relatively high energy environment will give a chaotic pattern (Sheriff, 1980). Changes in sea level during geological time are responsible for producing different patterns of depositional sequences around any coastline. The stratigraphic units in the study area result from the interaction between sea level change , activity of eustatic and isostatic uplift, climate and sediment supply.

The purpose of the geophysical survey was to use sparker data in order to study the distribution and the thickness of the recent sediments since the late Devensian , and the major structures of the consolidated rocks.

6.5 Results

Three seismostratigraphic units are recognised in the study area and are discussed in sequence of deposition form bedrock to surface.

The 3rd. seismic unit is interpreted as bedrock; its acoustic signature is very clear, well defined and has a continuous sharp boundary throughout most of the surveyed area. The upper boundary of the bedrock is characterised by a relatively smooth surface. Although, in part, this is a function of the wide angle of the sparker pulse it is clear that the rocky surface beneath the unconsolidated sediments is smoother than the rock surface where it outcrops on the sea bed. Structures within the bedrock are clear in the offshore zone and replicate the variable dips seen in the intertidal zone with both anticlinal and synclinal structures present. The records are not clear enough to identify any intrusive rocks offshore. In some seismic profiles e.g. A $\bar{A}$, B $\bar{B}$, H $\bar{H}$ (Plate 19) and L $\bar{L}$ the bedrock outcrops at the sea bed. At these locations the dip of the bedrock is more or less horizontal. A short distance away from the rocky surface where the bedrock starts to be lost with depth, there is a change in gradient from the gentle to a steeper slope. The point where the change in gradient begins probably marks the position of the sea level in the past. It is more likely that active marine processes took place at this point, producing the break in slope. Fig. 6.2 shows that the width of the sub tidal rocky platform at Buddo Rock, Buddo Ness and Boarhills is narrow, but it is wider at Kinkell Braes. This trend of the bedrock does not continue offshore. The depth of the bedrock reflector in the area ranges from zero close to the shore to 26 m below the sea bed in the area farther offshore. It is clear that the greatest depth to bedrock is located at seismic profile F $\bar{F}$, about 1.5 km east of Outhead. However, the data show that the depth to bedrock found to the north of Buddo Rock is shallower. There were no channels found cut into the bedrock, but in some seismic profiles (e.g. C $\bar{C}$ at stations 3,4,5 and 6) there are bedrock hollows.

The 2nd seismic unit (Fig. 6.3) is always found overlaying the bedrock reflector This unit is characterised by a different acoustic signature. In this unit acoustic signals range

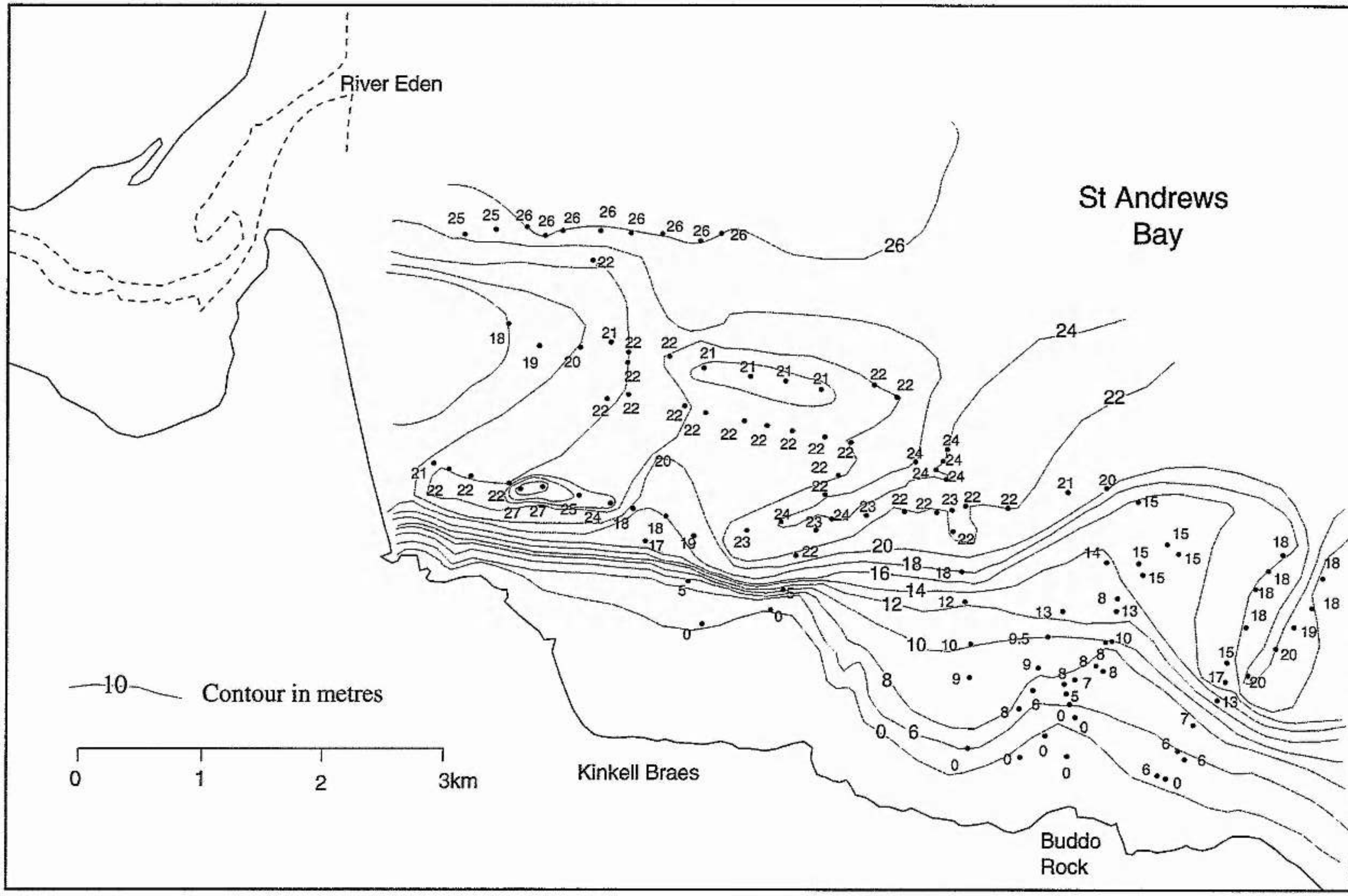


Fig. 6.2 Isopachyte map of total sediment thickness (3 rd. seismic unit) overlying bedrock.

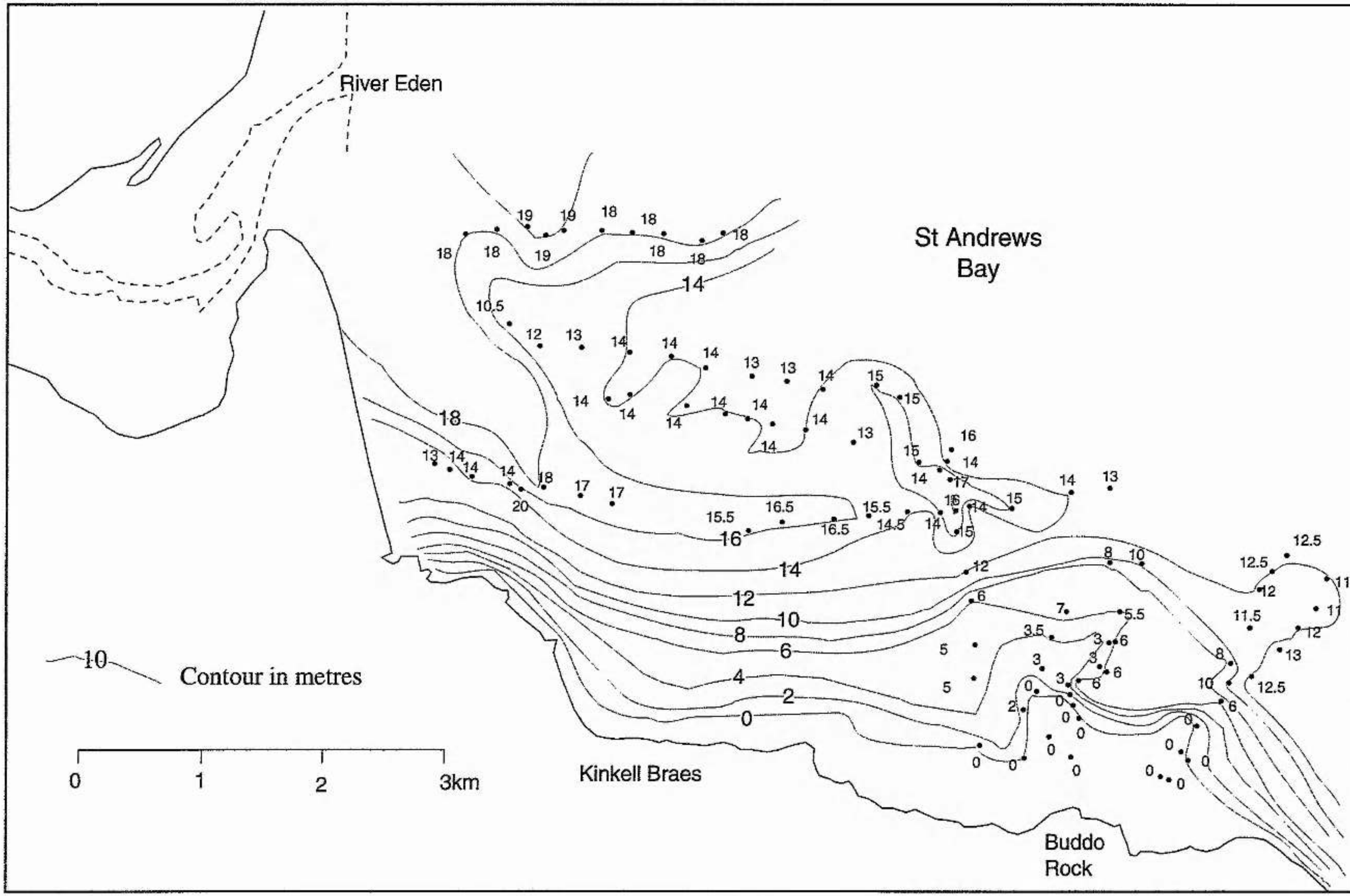


Fig. 6.3 Isopachyte map showing distribution of Lateglacial sediments (2 nd. siesmic unit) in St Andrews Bay.

between transparent, irregular discontinuous parallel reflectors to short, strong curved reflectors. These phenomena reflect the change in sediment character within this unit which leads to different patterns of acoustic signals. These signals merge with each other in a gradaitional pattern. Most seismic profiles indicate that there is a vertical sequence of reflectors in this unit which range in acoustic texture from short, strong curved reflectors, always found at the base of the unit, gradually changing to irregular discontinuous sub-parallel reflectors at the middle of the unit, to transparent reflectors always found at the upper boundary of the unit. These sequences of reflectors are constant for all seismic profiles except profile E $\bar{E}$ which shows a parallel, continuous reflector located between a transparent reflector and an irregular, sub-parallel discontinuous reflector. In seismic profile G $\bar{G}$ a short , strong curved reflector is overlain by the transparent reflector. Although the transparent reflector usually represents the upper boundary of the 2nd seismic unit, it disappears in some seismic profile (e.g. C $\bar{C}$, F $\bar{F}$ and M $\bar{M}$). The upper boundary of the 2nd seismic unit is overlain by the 1st seismic unit. It is clear that the transparent reflectors at D $\bar{D}$ (Plate 17), E $\bar{E}$, G $\bar{G}$, H $\bar{H}$ (Plate 19), I $\bar{I}$, J $\bar{J}$, K $\bar{K}$ (Plate 20) and L $\bar{L}$ wedge out towards the shore. There is always there is an erosion surface found at this boundary. This erosion surface delimits the 1st and 2nd seismic units on seismic profile D $\bar{D}$ (sites Nos. 17,18 and 19) at a depth of 8 m below the sea bed, E $\bar{E}$(site Nos. 30, 31 and 32) at a depth of 7 m below the sea bed, H $\bar{H}$ (site Nos. 14,15 and 16) at a depth 5 m, J $\bar{J}$ (site Nos. 5, 6,7 and 8) at a depth of 5 m, K $\bar{K}$(site Nos. 10,11,12 and 13) at a depth of 5.5 m, L $\bar{L}$(site Nos. 24, 25 and 26) at a depth of 5.5 m and finally on M $\bar{M}$ (site Nos. 31,32 and 33) at a depth of 6 m. The erosion surface is always found at the basal boundary of the 1st seismic unit, but it is not always overlain only by the transparent reflector but, occasionally, it is overlain by the irregular, sub-parallel, discontinuous reflectors, as at D $\bar{D}$ at sites No 17, 18 and 19. The maximum thickness of this unit (19 m) is found in offshore seismic profile F $\bar{F}$, but the average thickness is 14 m. However, the thickness of this unit increases where there are depressions in the bedrock (e.g. seismic profile C $\bar{C}$ at sites 5 and 6) where the thickness of this layer reaches 20 m.

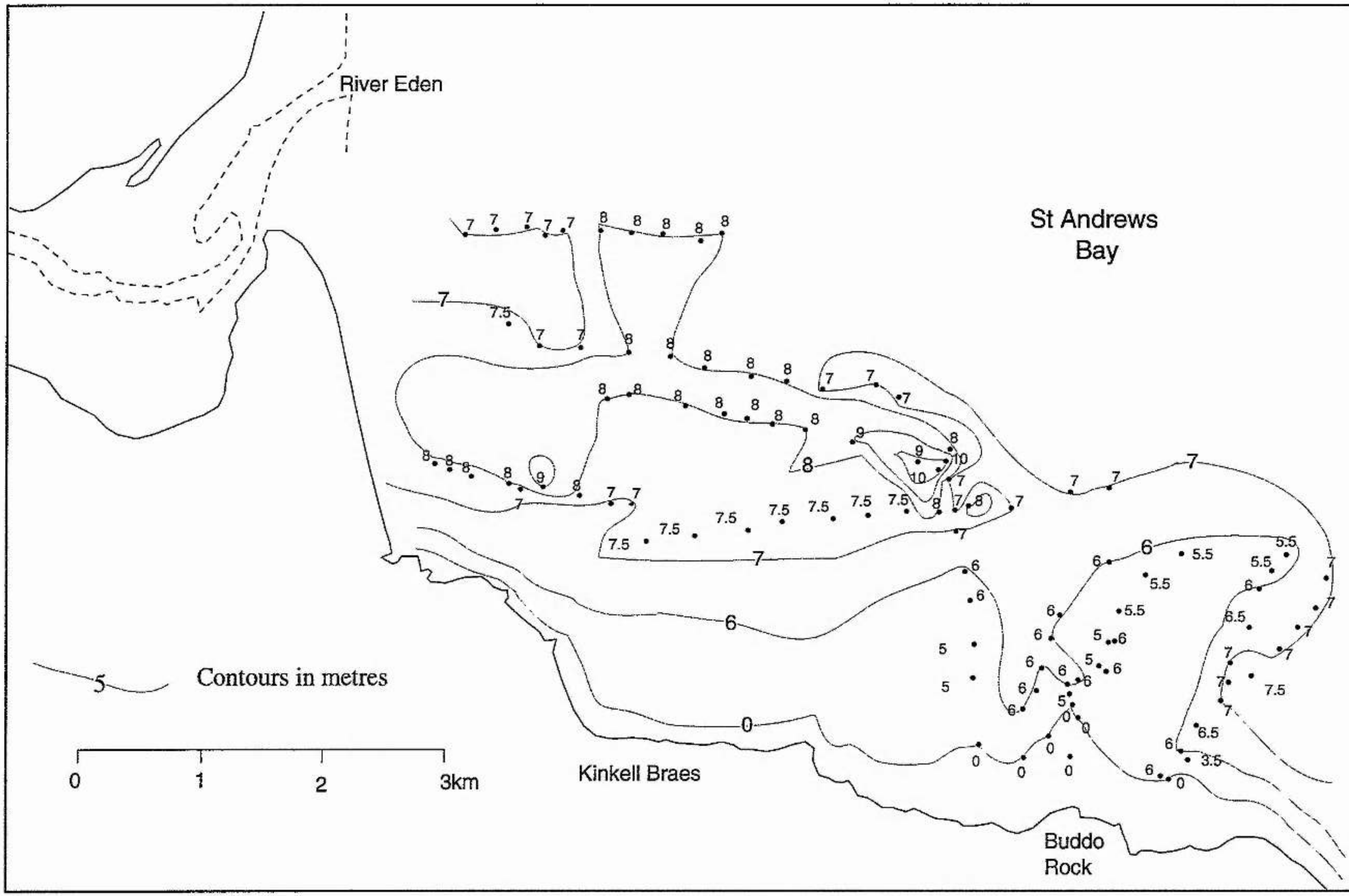
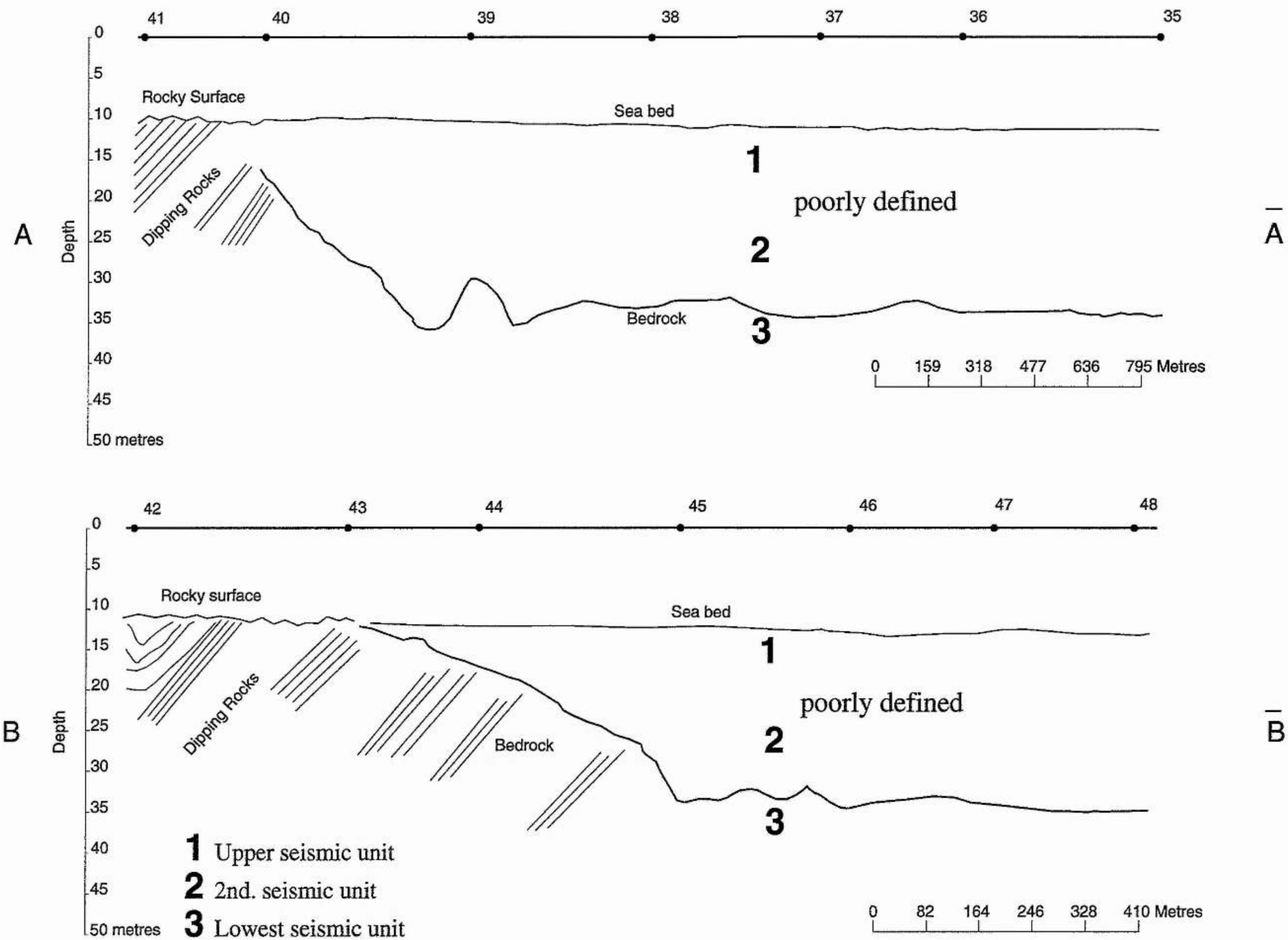


Fig. 6.4 Isopachyte map showing distribution of Postglacial sediments (1st seismic unit) in St Andrews Bay.

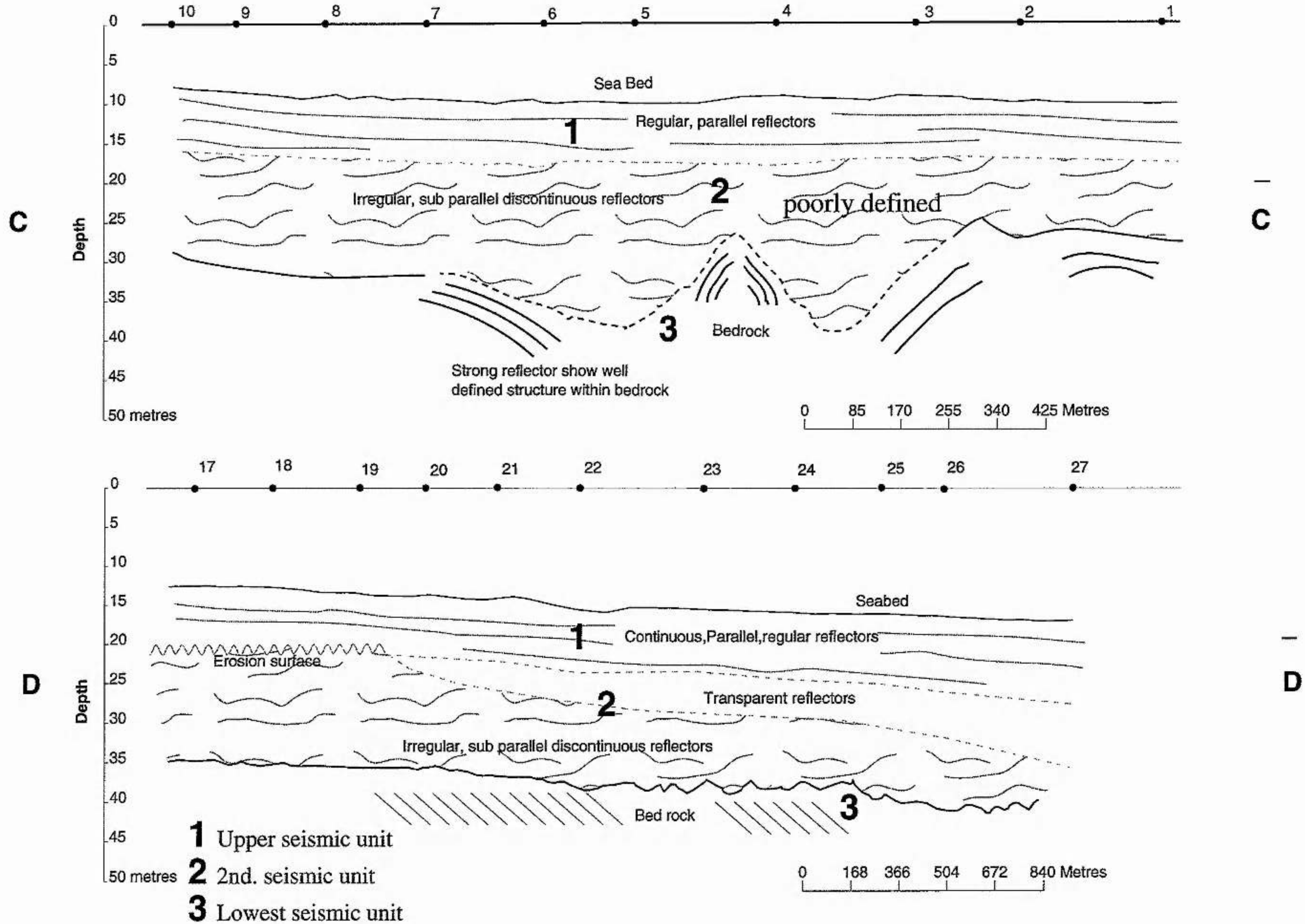
The 1st seismic unit (Fig. 6.4) which is interpreted as representing represents the modern sediments overlies the upper boundary of the 2nd unit at the base. On sparker record the unit appears as parallel, continuous reflectors in most seismic profiles. The exceptions are profiles H $\overline{H}$, at sites 13, 14, 15 and 16, J $\overline{J}$, at sites 4, 5, 6, 7, and 8 , K $\overline{K}$, at sites 10, 11, 12, 13 and 14, L $\overline{L}$, at sites 23, 24, 25 and 26 and M $\overline{M}$, at sites 30, 31, 32 and 33 where the unit appears as fairly transparent reflectors. At these locations the basal boundary of the 1st unit has a poorly defined contact with the upper boundary of the underlying 2nd unit. The thickness of the 1st seismic unit vary according to the locations of the traverses. For example, seismic profiles H $\overline{H}$, I $\overline{I}$, J $\overline{J}$, K $\overline{K}$, L $\overline{L}$ and M $\overline{M}$ reveal that the thickness of this unit ranges from zero at nearshore locations (here the sea bed appears as a rocky surface) to an average of 6 m seawards from the rocky platform. Profiles recorded in the offshore area (e.g. G $\overline{G}$, D $\overline{D}$, E $\overline{E}$ and F $\overline{F}$) plus those trending parallel to the East Sands (e.g. profile C $\overline{C}$) reveal that in these areas the average thickness of the unit is 8m. The maximum thickness of the unit was found in seismic profile D $\overline{D}$, at sites 24, 25, 26 and 27 where it is 9, 9, 10 and 10 m respectively.

The depth to the erosion surface at the base of unit 1 vary from place to place. The average depth to this surface is about 6 m as recorded in profiles E $\overline{E}$, at sites 30, 31, 32 and 33, profile H $\overline{H}$, at sites 13, 14, 15 and 16, profile J $\overline{J}$, at sites 5, 6, 7 and 8, profile K $\overline{K}$, at sites 10, 11, 12 and 13, profile L $\overline{L}$, at sites 24, 25 and 26 and finally at profile M $\overline{M}$, at sites 31, 32 and 33. It also is found that this erosion surface is not a continuous feature which can be traced over the entire survey area, but it is found intermittently in some locations.

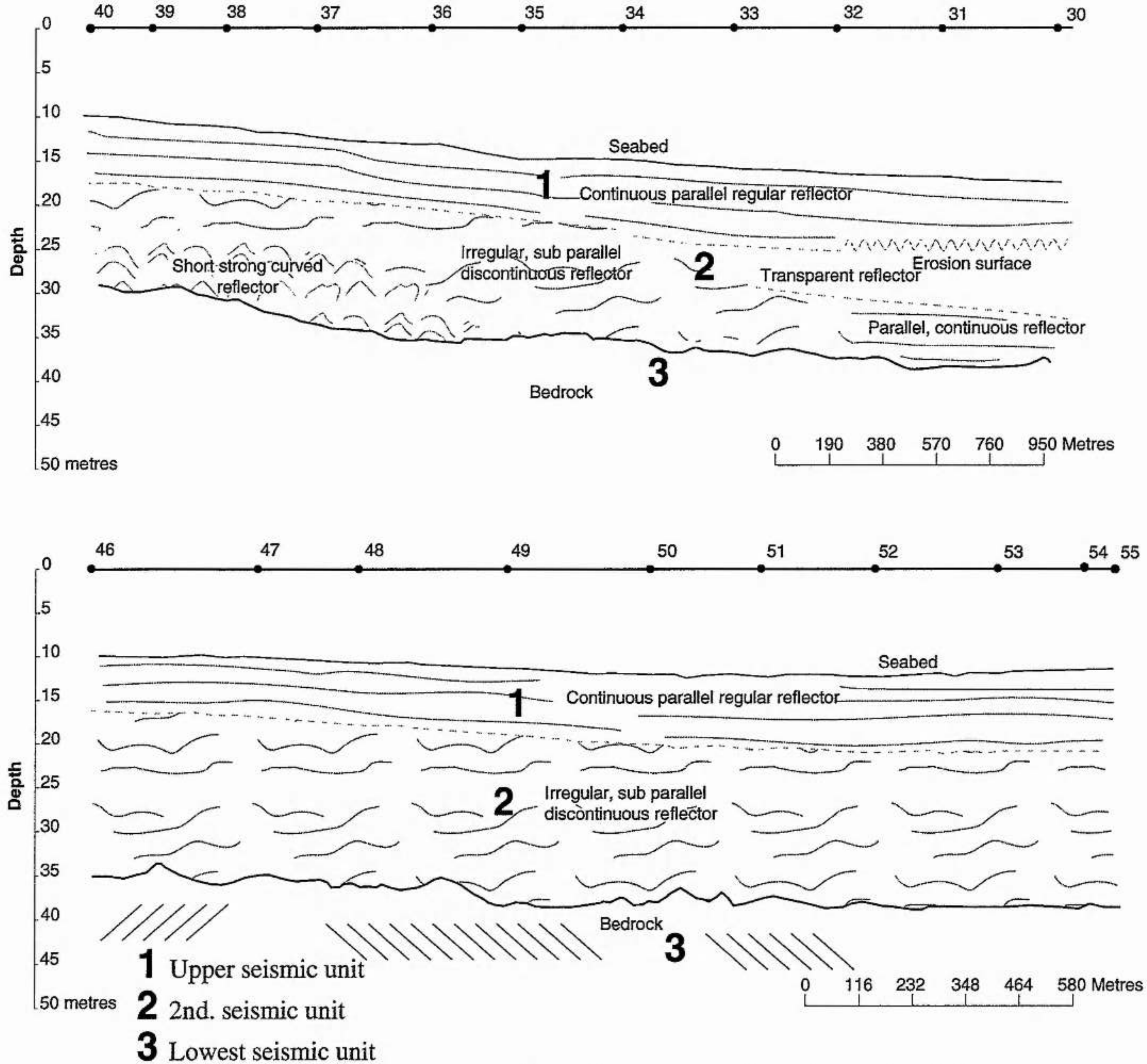
The thickness of each seismic unit is presented in diagrammatic form as an isopachyte (Fig.'s 6.2, 6.3, 6.4). In addition the interpreted data along the survey lines are presented as cross sections (Fig. 6.5 to Fig. 6.11) in order to demonstrate the disposition of the seismic units. This will provide a quick visual trend of sedimentary layers in the study area.



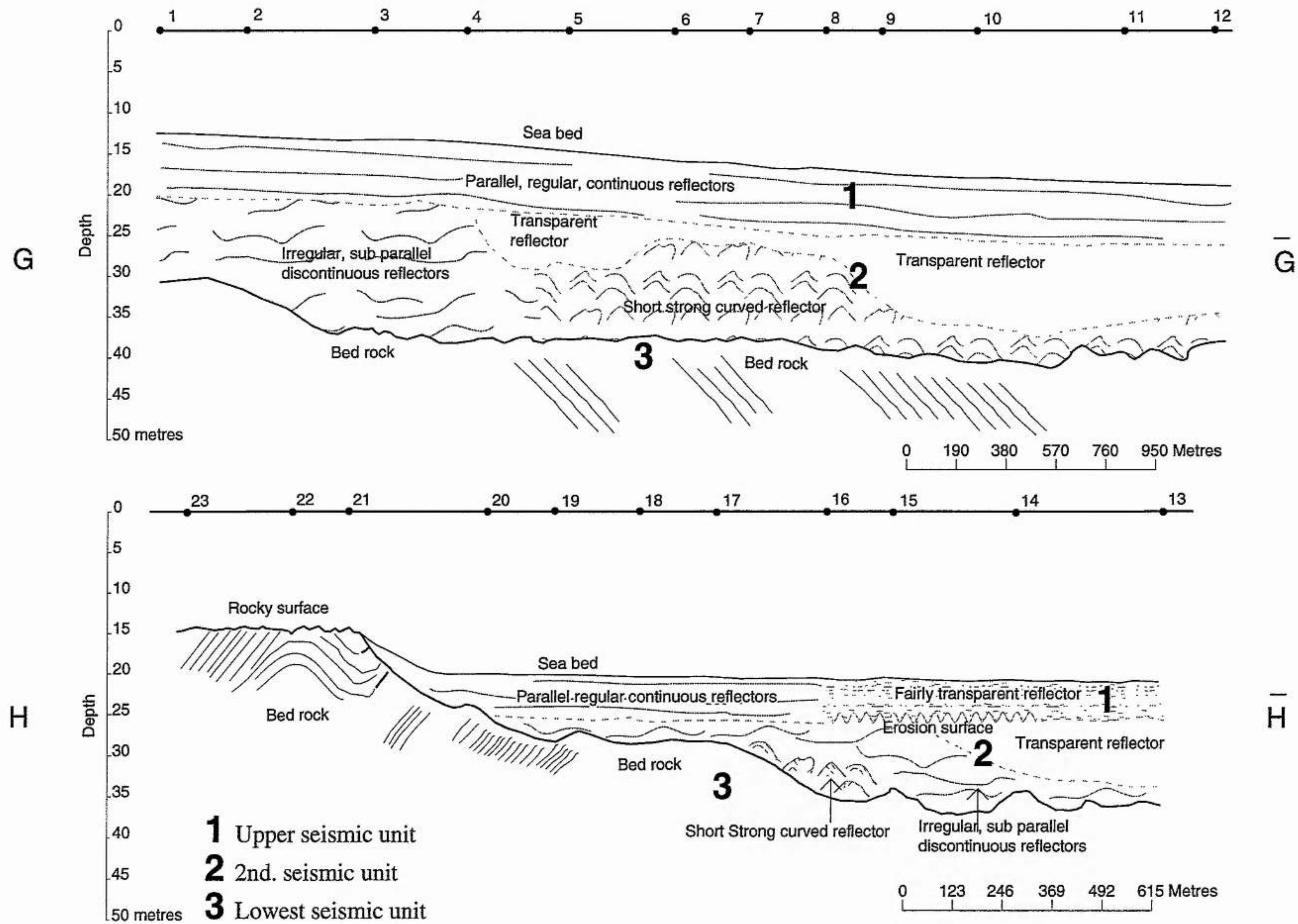
Interpretation of sparker sub-bottom profiles along survey lines (see Fig. 6.1).



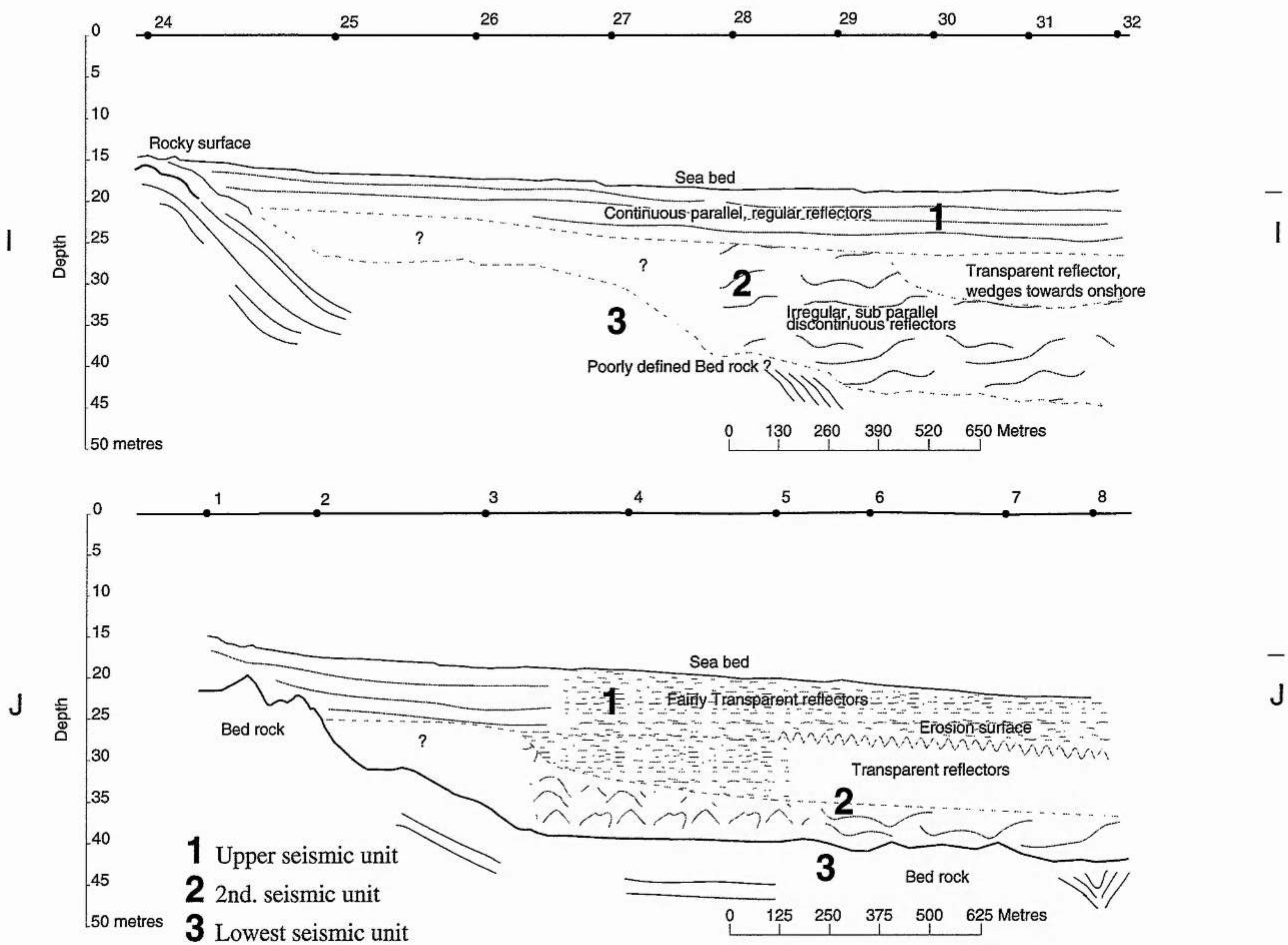
Interpretation of sparker sub-bottom profiles along survey lines (see Fig. 6.1).



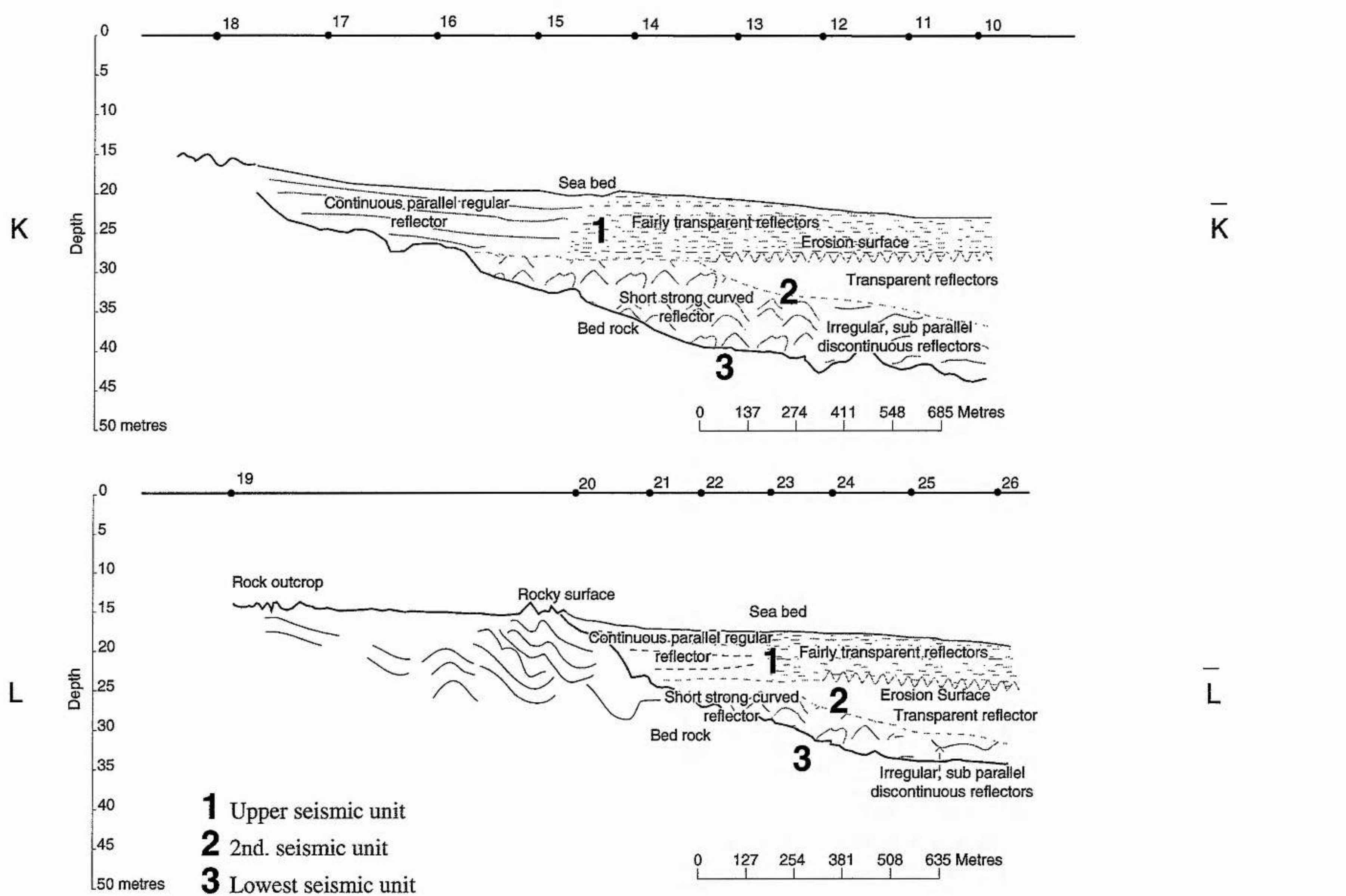
Interpretation of sparker sub-bottom profiles along survey lines (see Fig. 6.1).



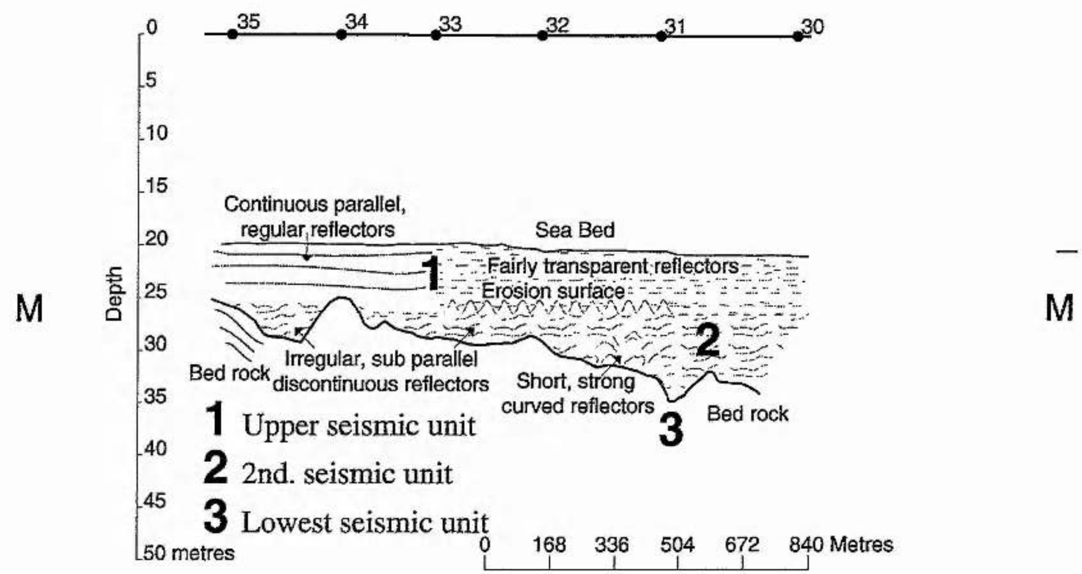
Interpretation of sparker sub-bottom profiles along survey lines (see Fig. 6.1).



Interpretation of sparker sub-bottom profiles along survey lines (see Fig. 6.1).



Interpretation of sparker sub-bottom profiles along survey lines (see Fig. 6.1).



Interpretation of sparker sub-bottom profiles along survey lines (see Fig. 6.1).

6.6 Discussion

A stratigraphic seismic sequence for the area can be established on the basis of the data which were obtained from the seismic records which reveal the presence of three seismostratigraphic units. The 3rd unit is interpreted as bedrock. Bedrock is exposed in the nearshore zone along the southern margin of the bay. The overlying units are interpreted as Quaternary deposits unconformably overlying bedrock. The upper boundary of the bedrock is characterised by a subdued relief surface which is believed to represent the result of erosion of the strongly folded Carboniferous sedimentary sequences. This probably occurred during the period when the surface of the bedrock was covered by Quaternary ice sheets. It is suggested that the depositional seismic sequence which is bounded by bedrock below and the upper boundary of the 2nd unit was deposited during the early stage of deglaciation of the late Devensian ice sheet which occurred between 18,000 to 15,000 years BP. The sea level was higher than at present allowing the Lateglacial sea to extend farther westwards than at present. At that time the bay was deeper than the present-day and glaciomarine sediments of marine clays and sands (St Abbs Formation) were probably laid down at this stag. Therefore this unit can be interpreted as being composed of these lateglacial deposits. However, as time proceeded through the Lateglacial interstadial there came a stage when the rate of isostatic uplift was greater than the eustatic rise of sea level. Then sea level would have fallen relative to the land. Between 11,000 and 10,300 years BP glaciers formed again in the Scottish Highlands, during the so called Loch Lomond Stadial. Because the study area was not subject to depression caused by the growth of Loch Lomond Stadial glaciers the rate of glacio-isostatic uplift continued to exceed that of sea level rise. During this regression of the sea, because water depths were reduced, waves and currents acted as agents of erosion on the bottom sediments causing the removal of sands and clays from the glaciomarine deposits. This processes is marked by the presence of erosion surface marking the top of 2nd seismic unit.

The subsequent amelioration in climate caused a rapid transgression of the sea at this stage. Sea level rose from its Lateglacial low level at about 10,000 years BP to about 9 m above its present level by 6,000 years BP (Sissons, 1976a). The sedimentary layers of the 1st seismic unit represent sediments deposited throughout postglacial time. This includes sediments laid down during the postglacial transgression and those deposited or reworked during regression of the sea to its present elevation. The average thickness of the 1st seismic is about 6 m. The variation of the thickness of these sediments from one place to another indicates a differential rate of deposition or erosion.

In conclusion in the study area it is possible to subdivide the Quaternary sedimentary sequences into two distinct units separated by the erosion surface. These are lateglacial deposits which were laid down before the maximum regression and postglacial deposits which were laid down after the Loch Lomond Stadial. The erosion surface in the area provides a marker horizon to separate between these two different seismic sequences.

CHAPTER VII

Summary and Conclusions

7.1 Introduction

This research study has investigated past and present changes in the morphology of St Andrews Bay. These changes can be related to the tidal and meteorological forces which act upon the waters of the bay and induce sediment movement in the coastal and nearshore zone. In particular, a marked contrast is observed between the shoreline at the head of the bay, which has been prograding for 3,000-4,000 years, and the rocky coastline between St Andrews and Fife Ness which is stable or undergoing slight erosion where waves penetrate to the backing cliffs.

7.2 Water and Sediment movement at the Head of the Bay

In order to understand the movement and deposition of sediment along coasts it is necessary to identify the main sources of sediment and its mode of transportation. Sources will depend upon the geographical and geological setting as well as the geological history of the area. According to Summerfield (1991) if the movement of sediment parallel to the shore is ignored, there are three main potential sources of sediment to the coastal zone, as follows:

1- The coastal landforms themselves, including cliffs and beaches.

2- The land area inland from the littoral zone.

3- The offshore zone and beyond.

Cliff erosion is not believed to be an important source of material to the coastal zone because the rate of erosion is low, since large areas of coastal cliffs are protected from erosion by a wide intertidal platform which absorbs wave energy. As a consequence, much of the cliffline of St Andrews Bay, especially on the southern shore, is weathered and degraded which confirms that no wave energy reaches the cliffline today. During the early stages of coastal progradation, after the Postglacial maximum transgression in sea level, the unconsolidated raised beaches which surround the bay (Fig. 2.5) may have provided material to the coastal zone to promote progradation of the shore. These raised beaches are now isolated from the present beach zone and can no longer be considered a major source for the material involved in coastal recession.

It is argued that the input of material to St Andrews Bay from the Tay and Eden catchments is minimal, especially for sediments coarser than fine sand which moves as bedload. Any sediment entering the upper Tay and Eden Estuaries will be trapped in the estuarine embayments because of the slack currents and the deeper water and this will prevent sediments reaching St Andrews Bay. The Eden Estuary does appear to be infilling with sediments which are, in part, derived from the river (Jarvis, pers. comm.). In due course, one might expect material to be transferred into the bay to augment the nearshore deposits, but at present little coarse material is thought to enter the bay. The Abertay and Gaa Sands are located where the Tay enters the open sea and might partly be derived from estuarine sources. However, they are likely to be the result of the movement of material circulating in the outer estuary since landwards the open estuary between Tentsmuir and Dundee acts as sink for material from the river and little coarse material is transferred to the outer estuary.

Accordingly, the major source of coastal deposits at the head of the bay and those currently producing coastal progradation must be attributed to the material which has originated offshore. This comprises material of the modern sea-bed which itself is derived from the glacial deposits which were left behind in the bay during glacial retreat. Part of this onshore transport of material was associated with times of rising sea level, in particular, the Postglacial transgression, during which beaches were formed at up to 8m above the present sea level. Since 6,000 years BP eustatic rise in sea level was almost complete but isostatic rebound continued, thus causing a slow fall of sea level for about 3,000 years. This phase of falling sea level led to a general abandonment of the Main Postglacial Shoreline and emergence of the intertidal zone which became progressively incorporated into the area above the high water mark. In the past, as today, dune ridges formed above the high tide mark and a series of linear dune ridges inland from the present beach at Tentsmuir represent former shorelines abandoned during regression. These linear sand dunes are aligned primarily north-south, but eventually trend parallel to the present shoreline. The slightly different orientations of these dunes may reflect the differences in wave refraction / diffraction patterns induced by bathymetric changes as sea level lowers.

It is thus suggested that the major factor which has determined coastal evolution at the head of the bay has been fluctuations in sea level. The rising and falling stages are times of massive sediment release and transport, while the stationary stage allows time for adjustment and reorganisation towards equilibrium (Carter, 1988). In St Andrews Bay sea level has been largely stationary for the last few millennium: there is little evidence that a stable shoreline has been achieved between St Andrews and Tentsmuir.

At the present time, progradation is the dominant processes along the coastline of St Andrews Bay, but intermittent periods of erosion do occur, which are usually localised in space and time. Persistent erosion has affected the northern shore of the Eden Estuary for the last 40 years and from map evidence could be related to the changes in the alignment of the estuary, which took place in the 1940s, allowing more wave energy to reach the shoreline (Jarvis, pers. comm.). On Tentsmuir beach erosion has also occurred in some places, for example just to the north of Kinshaldy, but this only affected a 200 - 300 m stretch of coastline. Here, the shoreline receded by 20 m between about 1975 and 1983 but has since recovered and is now prograding beyond its 1975 position (Jarvis, pers. comm.). Clearly, these localised variations are part of the natural system (e.g. changes in wave climate) and do not reflect changing sea levels. The erosion processes along the coast may result from other factors such as a reduction in sediment supply as a result of interception or a change in the angle of incidence of waves due to changes in offshore topography, following shoal growth (Sarrikostis, 1987; Boalch,1988).

Climatic studies predict a global warming trend and it is widely accepted that a world-wide sea level rise is likely to occur during the next century (Warrick and Farmer, 1990). Studies indicate that sea level has risen between 10 and 15 cm over the past century and some investigators predict an accelerated sea level rise in the years to come by as much as 30 cm or more by the year 2025 (Lutgens, 1990; Bird, 1993). The recent mean sea level trends in the British Isles are shown in Fig. 7.1.There is no general overall trend, rather the islands may be split into two provinces, one in the south east, where the tendency is for sea level rise, and the other in the north and north west where sea level is falling. These two provinces correspond to the Late Quaternary glacio- isostatic loading pattern, with a large ice centre over north west central Scotland spreading out over northern England and Ireland, and a forebulge over the extreme south and east of England (Carter, 1988). St Andrews Bay is seen to lie within the area where little relative change of sea level is taking place at the present time.

In the study area it is believed that the supply of marine sediment from the sea floor by shoreward drifting processes is still accumulating material at the head of the bay. Accordingly it is most likely that, even in the case of rising sea level, the large sandy beaches (e.g. Pilmour Links and Tentsmuir Beaches) will not change dramatically since the effects of sea level rise would be compensated for by the continuing supply of sediment . Clearly, a different response to changing sea level might be experienced by the rocky coastline between St Andrews and Fife Ness. Here there is no buffer for the effects of rising sea level and one would expect erosion to occur. In particular, the small beaches at the base of low cliffs or the beaches in coves along the rocky coastline of the southern margin of the bay would be removed as sea level rose.

The results of grain size analysis of 113 sea bed sediment samples revealed that fine sands are the most abundant surficial sediments. The result showed that the sediments are unimodal which is taken to indicate a state of equilibrium between sea bed and processes (log normal distribution). In general, the bed sediments in the bay are well sorted and are leptokurtic in character.

A trend in grain sizes is observed in the direction of the Abertay Sands and Tentsmuir Beach, as the energy of the waves increases in shallower water the distributions become skewed towards the coarse size fraction. There is little change in grain size in the

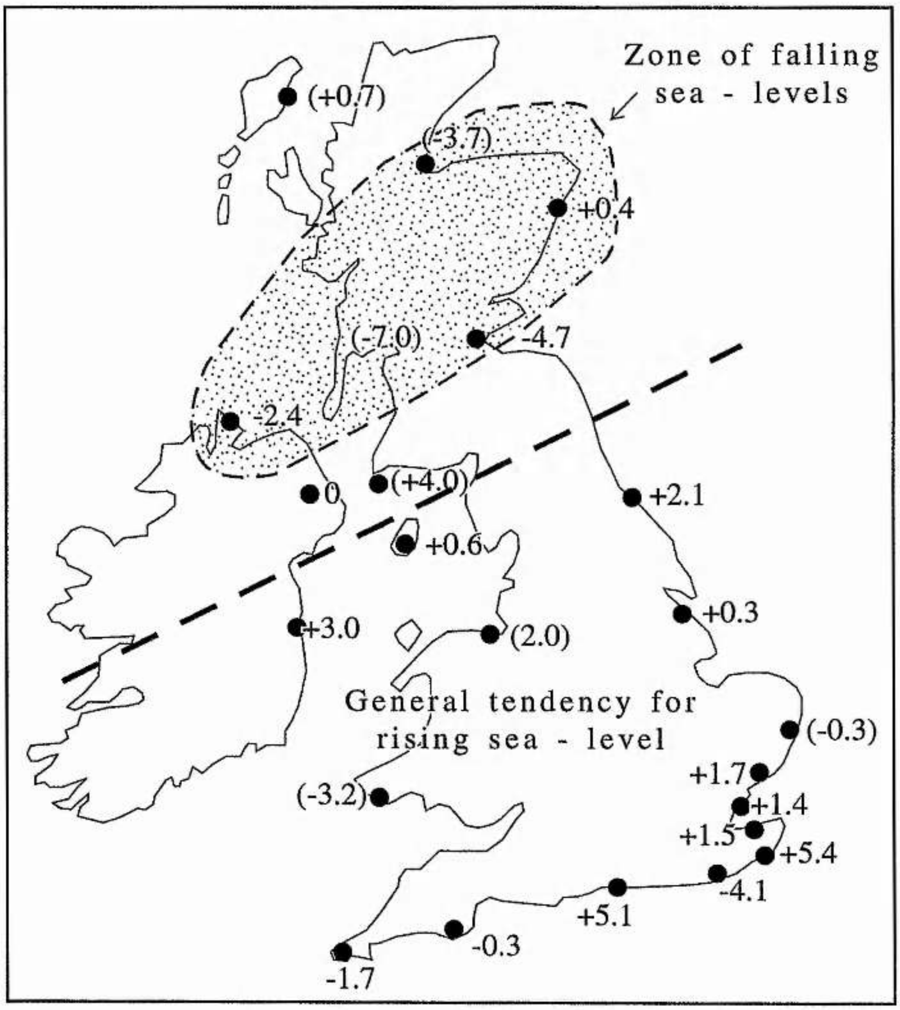


Fig.7.1 Recent sea - level changes around the British Isles from data presented by Pugh and Faull (1982) and Carter (1982c). Note the considerable variation even between adjacent sites (after Carter 1988).

shallow waters and on the beaches which suggests that the beach material and offshore material are formed from the same population and thus adds credence to an offshore supply for the prograding material. In deeper water, with the addition of fine material, the normal distribution becomes fine skewed. Clearly, the continued movement of material onshore from deeper water implies that there is selective movement of particular size fractions, since the materials inshore and onshore are coarser than the offshore sediments. It is not possible to determine if the offshore sediments are gradually fining as a result of coarser grades moving onshore, or if this is just surface fining with coarser sand at depth.

The distribution of fine sand in St Andrews Bay, with its limited range of grain sizes, can be interpreted as a reflection of a hydrodynamically homogenous area except where coarser sediment on the Abertay Sands implies a greater variation in energy levels. The result of application of the quartile deviation-median diameters (QDa-Md) technique of Buller and McManus (1972a, 1972b, 1973) to the sediment distributions suggests that the movement of bed sediments is controlled mainly by wave activity. However, this technique is not a very discriminating tool in the bay because of its restricted range of hydrodynamic sub- regions.

The current meter data from the different locations in the bay support the hydrodynamic interpretations from the grain size data of a low energy tidal environment. Nearbed velocities are low throughout the majority of St Andrews Bay and seldom exceed the threshold for sediment movement of fine sand. In fact only by applying the lower of the two thresholds employed by Vincent et al (1981) in their study of New York Bight is it possible to compute a bedload transport rate. Thus the data also support the conclusions of the QDa-Md method that wave activity must contribute significantly to sediment movement.

It is noteworthy that sediment transport rates in Swansea Bay, with its more open morphology, are one or two orders of magnitude higher than in St Andrews Bay. These higher rates are reflected in more diverse sediments and bed morphology in Swansea Bay. Mud and sand as well as offshore tidal sandbanks (Heathershaw and Hammond, 1979) are found off the Welsh coast compared to the gradual offshore fining of sediments in St Andrews Bay with its gentle topography. Although much of the differences between the two embayments are related to the stronger tidal flows in Swansea Bay (maximum rates of 50 to 70 cm s^{-1} at 2 m above the bed), its south-west exposure will allow the dominant south-west winds to propagate larger waves into the area compared to the eastwards facing St Andrews Bay.

7.3 The Rocky Coastline

The rocky coastline from St Andrews to Fife Ness has been investigated in detail, both by terrestrial and hydrographic surveying methods. Despite the amount of data that has been collected in this study it is still not possible to make definitive statements about the origin and possible sub-divisions of the inter-tidal and sub-tidal rock platform. The sedimentary rocks of the subtidal rocky platform appear to have a very similar morphology to rocks in the intertidal platform (see Plates No. 12 to 16). Thus, the combined elements of the rocky platform extend from the high water mark to at least 9 m below present sea level and vary in width from 500 m off Buddo Ness to about 4,000 m at Kingsbarns. The Carboniferous sedimentary rocks in the area are folded and faulted, showing both plunging anticlinial and plunging synclinal structures, so individual strata show wide ranges of dips and strikes. The geology of the platform thus has a major effect upon its form. In some locations the cross profile appears to be fairly smooth, (e.g. profile No 1 at Kingsbarns) whereas in other places some sub- horizontal benches (e.g. profiles No, 7, 8 and 24)) are apparent in the profiles, although these have little lateral continuity.

Wave scouring at the seaward edge of the platform (Plate 12) and occasional ripple fields in patchy sediment veneers indicate that wave and perhaps tidal currents are sufficient in these areas to cause sediment movement. It is unlikely that solid rock would be subject to much, if any, erosion at the seaward limits of the platform and even in shallow water erosion will be limited by the fact that the most energetic waves reaching the shore from the north-east and east do so on only some 20 days per year.

Stability of the intertidal part of the platform, or its location in a relatively low energy environment, is supported by the cover of seaweed and marine organisms, not only on the rock but also on the clasts of all sizes that lie on the platform. Only at the highest part of the platform is there evidence of sediment movement and possible abrasion, although this may well reflect the long period of sub-aerial exposure as much as it reflects the result of wave activity centred around the high water mark.

The cliffline from St Andrews to Fife Ness is clearly a fossil feature. In some places, e.g. Buddo Ness, sea water does not reach the base of the cliffs so they must predate the present sea level. Although the sea does reach the cliffline in other places, for example at St Andrews and to the southeast of the East Sands, the lack of a wave cut notch, subdued slopes and the vegetated nature of the cliffs to their base show that they are inactive marine features. In contrast, however, rockfalls and an active landslide at Kinkell Ness shows that subaerial processes are still modifying their form. At Kinkell Ness and Buddo Ness the cliffs are fronted by platforms a few metres above present sea level. These, and the abandoned stacks of the Maiden Rock and Buddo Rock are features formed during the Postglacial transgression. At this time the cliffline would have been attacked by wave activity throughout its length and the present form most likely dates from that time.

At Buddo Ness, the abandoned Postglacial platform is covered by unconsolidated deposits. It is not clear whether the present intertidal platform continues without a change of profile beneath these deposits or a slightly higher platform, related to the higher, sea level underlies them. It would be valuable to determine this relationship between the present intertidal platform and former higher sea level in order to comment on possible rates of change for the present intertidal platform. If a change of level or slope does occur it would indicate the potential of the recent stable sea level to modify the present intertidal platform. However, if there is no change of the platform as it passes beneath the Postglacial deposits it would suggest that processes today are not very effective and indeed were not so at the time of the higher sea level.

It is considered that the formation of the rocky platform of St Andrews Bay precedes at least the last glaciation because present-day processes would appear to be incapable of causing its formation over the last 3,000-4,000 years of stable sea level. Similar platforms of an interglacial age have been reported from the west coast of Scotland although, in these cases, some of the platforms pass beneath glacial deposits, demonstrating a pre - Late Devensian age. No glacial diamictons are found on the platforms at St Andrews, but individual boulders of erratic material are present, for example at Crail Golf Course (GR. 618078) although their provenance has not been determined. Also there are erratics of metamorphic origin from the Grampian Highland in the area e.g. Rock and Spindle (GR. 536153). The absence of in situ drift which would confirm a pre - Late Devensian origin is attributed to past and present-day wave activity in exhuming the platform from beneath any unconsolidated cover. Such exhumation would explain the rather localised occurrence of raised deposits in the area which may have been more extensive at the time of the Postglacial transgression.

7.4 Offshore Deposits

The sedimentary record of the sediments offshore contains evidence of former environments and the possible sea level history of St Andrews Bay. Only direct sampling by boreholes would enable one to make definitive statements about the sedimentological and faunal changes with depth. However, this methodology provides only site specific information and will not allow interpretation of the geometry of the sediment bodies which may also provide environmental information. The seismic reflection techniques, which have revolutionised our knowledge of offshore stratigraphy by providing indirect information on submarine sedimentary sequences, allow the construction of a three dimensional picture of the sediments from which environmental interpretations may be made.

Three seismostratigraphic units were recognised in the study area. These units are interpreted as bedrock, Lateglacial deposits and Postglacial sedimentary deposits. The last two cannot be subdivided further in the study area - a factor which results from both the relatively thin sequences of deposits and the limited resolution (1.5 m) of the Sparker system used. Some consistent changes in seismic texture are noted in the Lateglacial deposits where the unit changes from a moderately well-layered sequence near the base to a transparent texture at its top. However, without borehole data these changes cannot be related to any sedimentological change.

A maximum penetration of 30m relative to the sea bed is recorded by Sparker. The isopach maps (Fig. 6.2, 6.3 and 6.4) show that the deposition of Quaternary sediments is influenced largely by the bedrock topography. The maximum thickness of Lateglacial deposits (19m) is found in offshore seismic profile F $\bar{F}$. The average thickness is 14m, but the thickness of this seismic unit increases where there are depressions in the bedrock (e.g. seismic profile C $\bar{C}$). The deposits gradually thin towards the rock outcrop of the subtidal platforms.

The erosion surface which marks the junction between presumed Lateglacial and Postglacial deposits has been identified at least 10 km to the east of the study area by the British Geological Survey (Stoker, pers. comm.). A possible sub-aerial origin has been proposed for part of the surface where channels have been observed cutting into it (Browne and Jarvis, 1983). The strong reflection from the surface has been attributed to the presence of a gravel lag horizon which is found onshore at lateral equivalents of the surface. The extent of removal of the underlying deposits and the mode of formation of this surface which extends over several hundred square kilometres has not been determined. However, it has been suggested by Browne and Jarvis, 1983 that this surface was formed by wave action, when sea level was in its minimum level.

The Postglacial deposits are laid over the erosion surface and pass upwards to the modern sea bed without clear differentiation because of lack of resolution. This contrasts with other areas in the North Sea where the Postglacial deposits (the Forth Formation) are sub-divided into several members (Largo Bay and St Andrews Bay Members). Onshore, these deposits are exposed in river sections as silty sands but offshore the BGS have identified interbedded sands and clays as well as pebbly muds (Gatcliff et al., 1994). The results of the seismic survey show that the Postglacial sediments in the area have an average thicknesses of 6m but the unit thickens towards the east. Maximum thicknesses (10m) were found in seismic profile D $\bar{D}$.

7.5 SUGGESTIONS FOR FUTURE RESEARCH

(1) In St Andrews Bay a study of particle morphology (e.g. sphericity, form and roundness) is needed to compare offshore materials with those collected from different locations along the beaches. This will help to determine the provenance source of sediments in the nearshore area.

(2) The use of the scanning electron microscope (SEM) to identify the micro-relief features on loose quartz sand grains might enable one to recognise specific depositional environments.

(3) A study of heavy minerals also used to relate sediments to source rock lithologies and might help determine dispersal patterns. Age dating of some heavy mineral species (e.g. tourmaline- ^{40}Ar : ^{39}Ar : and Zircon U : Pb) could also provide important information regarding paleogeographic reconstruction.

(4) A better understanding of water circulation in the Bay could be obtained from observations on the vertical structure of currents in the Bay using an array of current meters. This would subsequently lead to an improved understanding of residual sediment transport in the bay.

(5) A seismic survey with a Pinger system (i.e. higher frequency than the Sparker used) is required to achieve better resolution within the upper sedimentary layers of the bay.

(6) Boreholes drilled to bedrock are ideally required in order to calibrate the geophysical data from the area and to confirm the boundaries of the sedimentary sequences units which were inferred from the present seismic survey. Furthermore, boreholes to bedrock will provide further important details of lithology of the Quaternary history of the area.

Plate 1 Folded and faulted intertidal rock plalform at Kinkell Braes.

Plate 2 The " Rock and Spindle " volcanic vent at Kinkell Ness.

Plate 3 Postglacial raised beach and intertidal platform 1 km west of Buddo Ness.

Plate 4 Intertidal rocky platform between the Maiden Rock and Kinkell Ness, showing rocks striking perpendicular to the coastline and dipping to the east.

Plate 5. Intertidal rocky platform at Kinkell Braes backed by weathered and steep cliffs.

Plate 6 Sand patches within the intertidal rock platform.

Plate 7 The sea stack of the Maiden Rock, probably formed during the postglacial transgressioı 500m from East Sand.

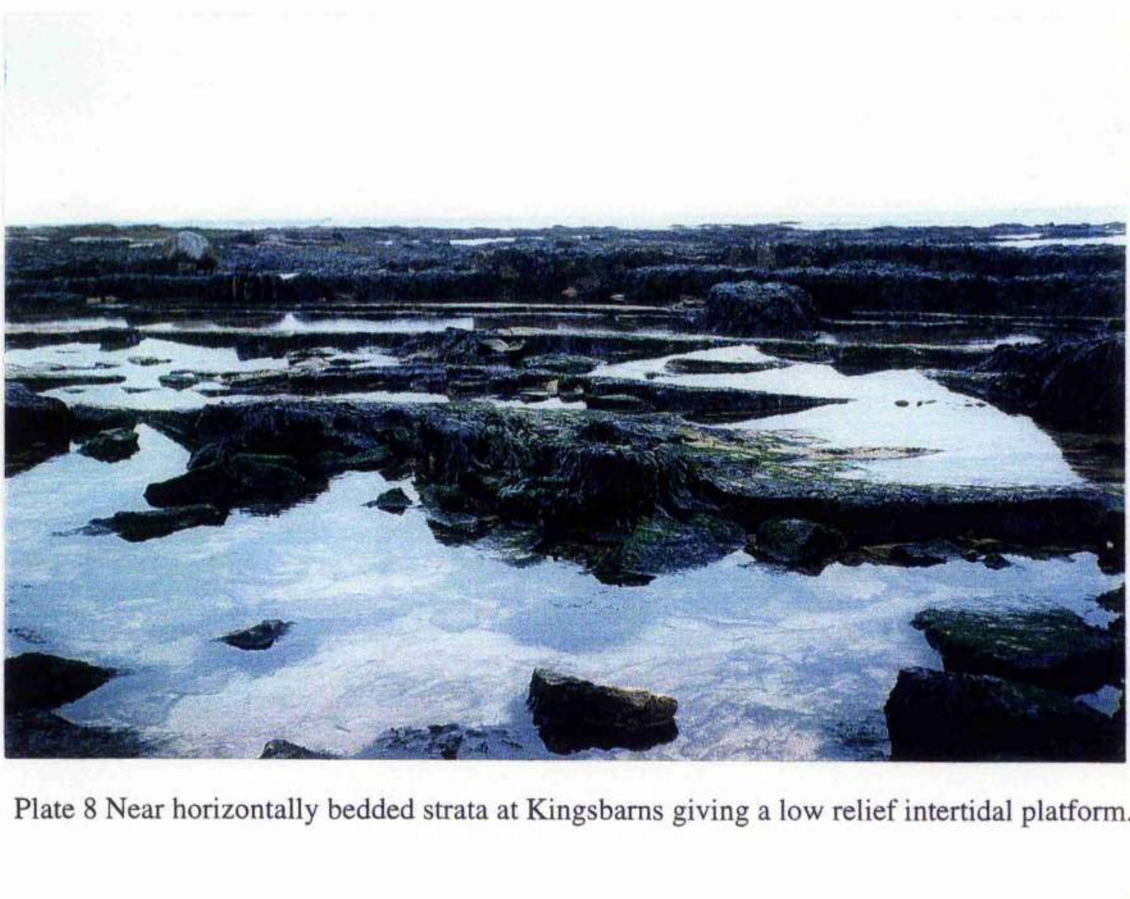


Plate 8 Near horizontally bedded strata at Kingsbarns giving a low relief intertidal platform.

Plate 9 Near vertically bedded strata in the cliff at Kinkell Braes, showing minor mass movement on the cliff face.

Plate 10 Postglacial raised beach, at Buddo Ness 3.5 km from East Sands, with a small cliff showing some reworking by the present sea.

Plate 11 Steep cliffline which fringes the bay at St Andrews. Rocks strike parallel to the cliff.

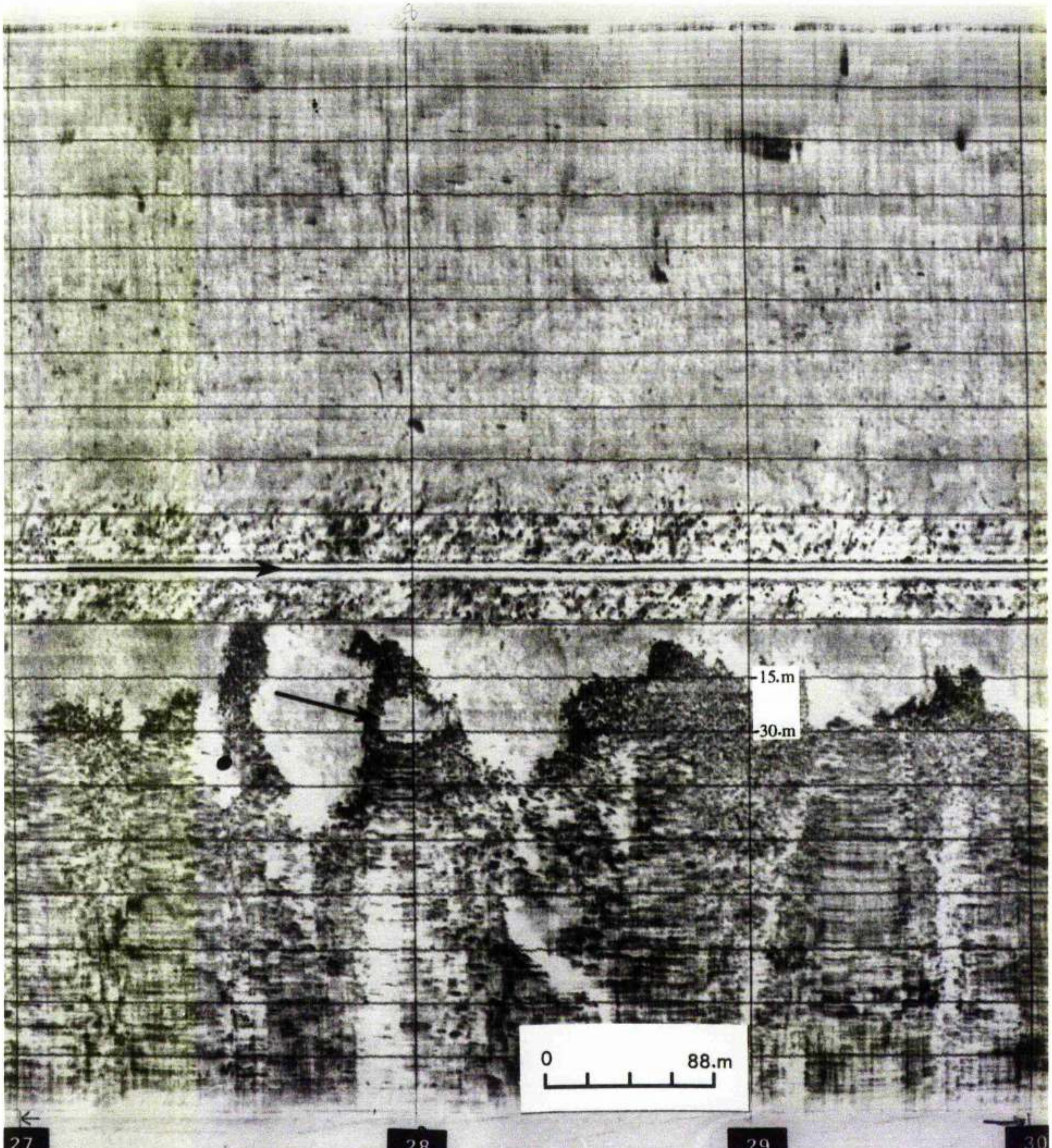


Plate 12 Large hollow, thought to be formed in rocks characterised by joints and later unfilled by sand and light tones indicating a shallow depression at the seaward margin of the subtidal rock platform.

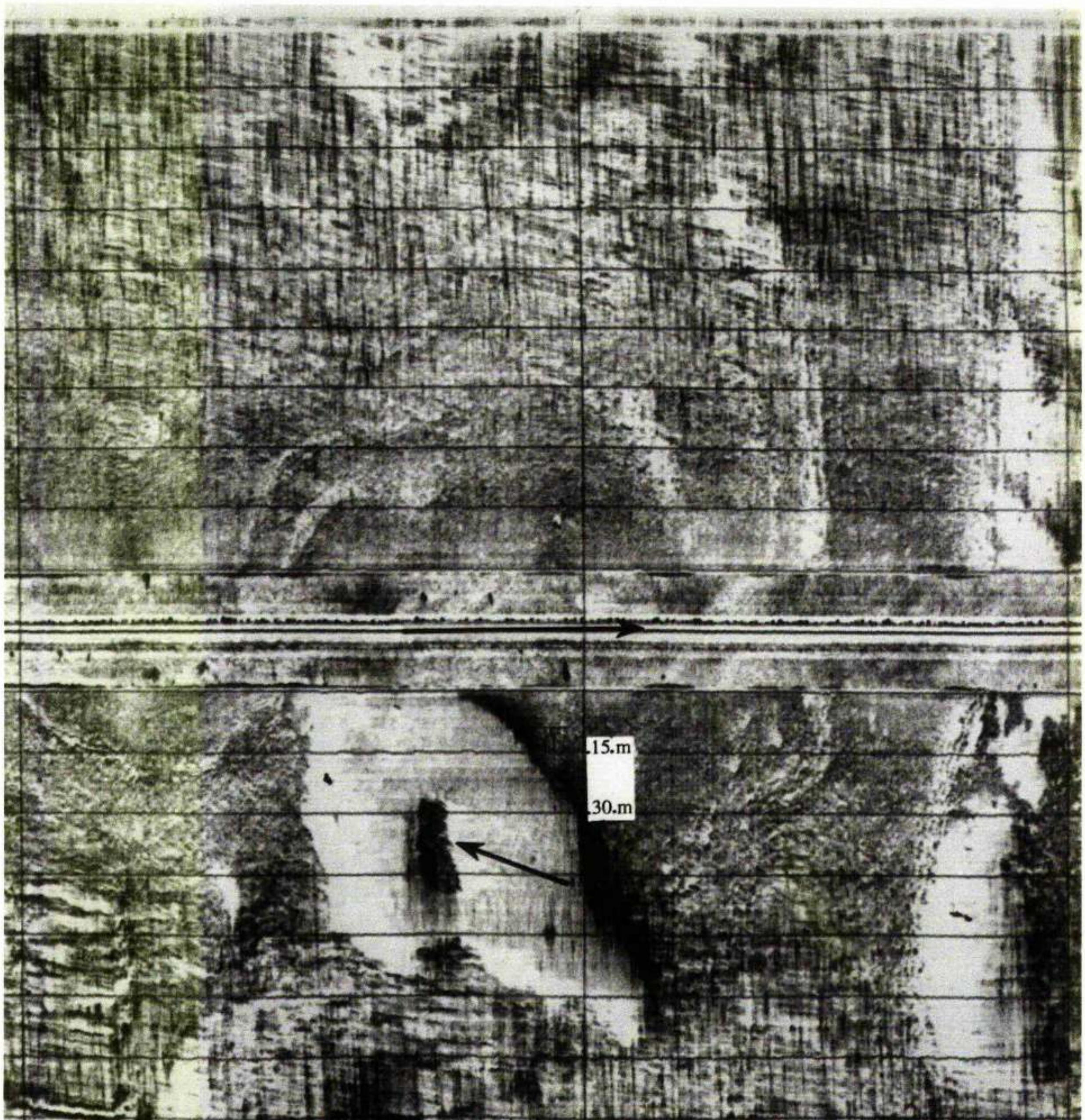


Plate 13 Sub-merged volcanic neck, about 1.4 km from Kingsbarns.

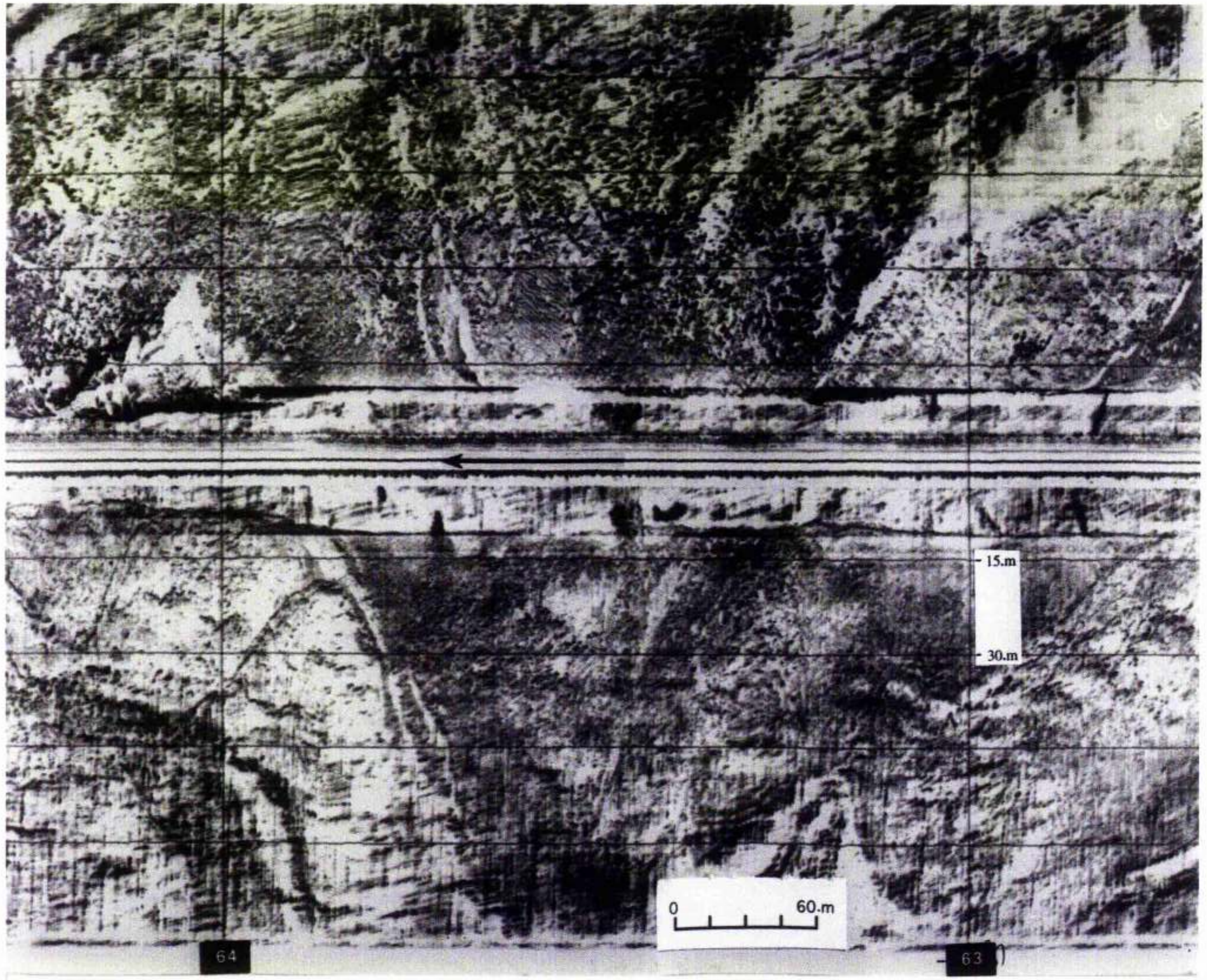


Plate 14 Bedform features (Ripple Mark) in bedrock hollows on the subtidal rock platform about 360m offshore from Babbet Ness.

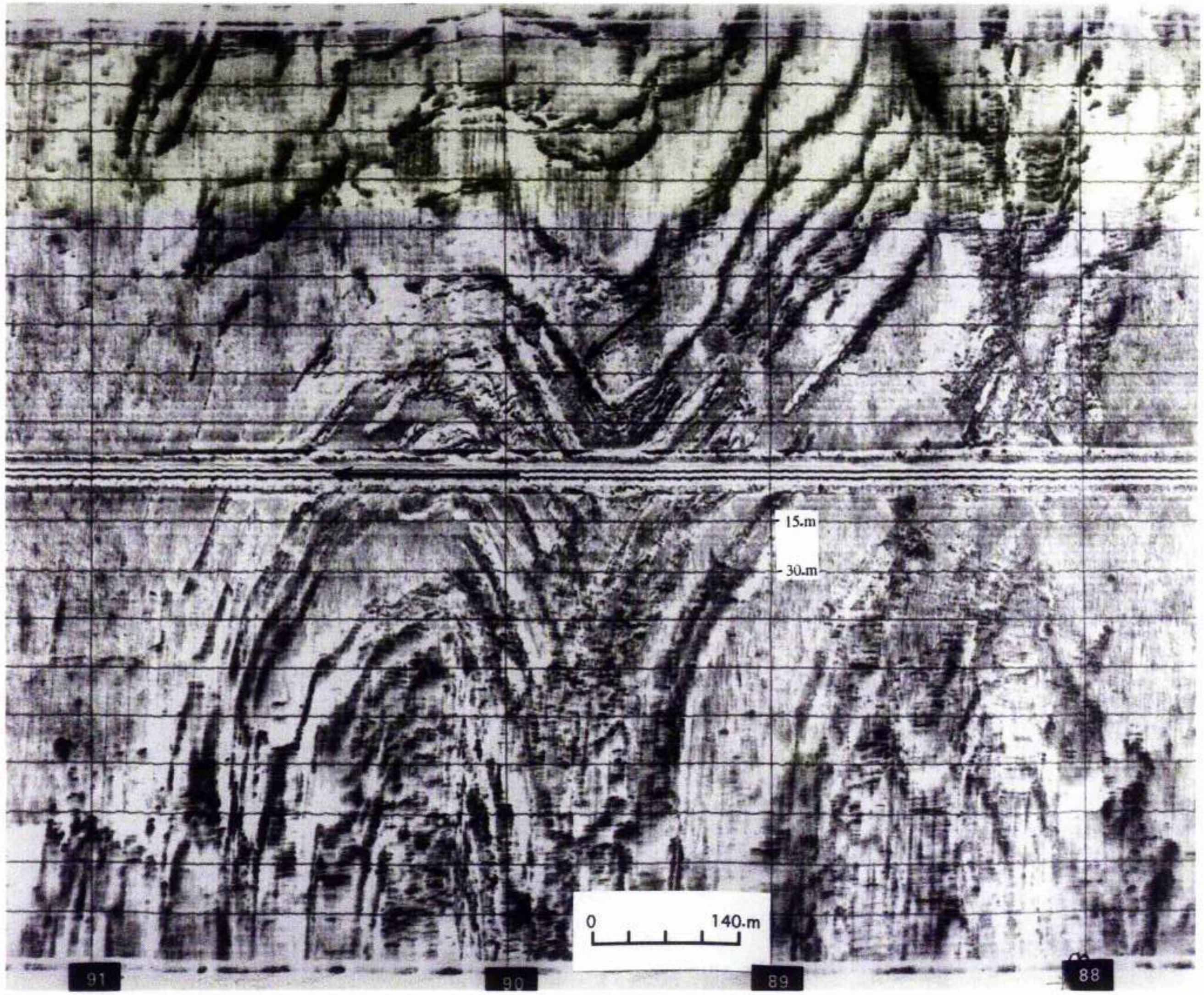


Plate 15 Subtidal rocks characterised by fold structures at site No 102.

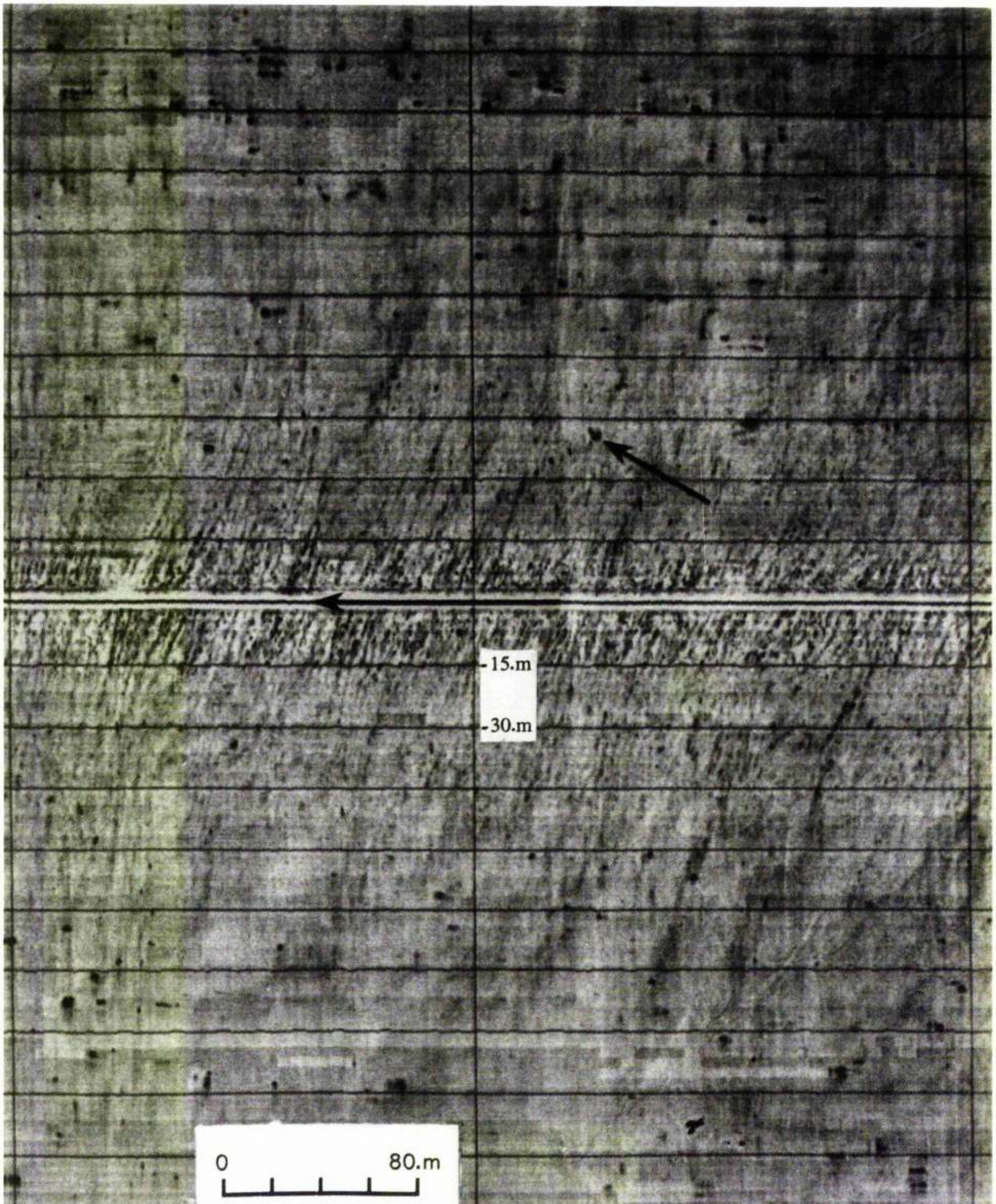
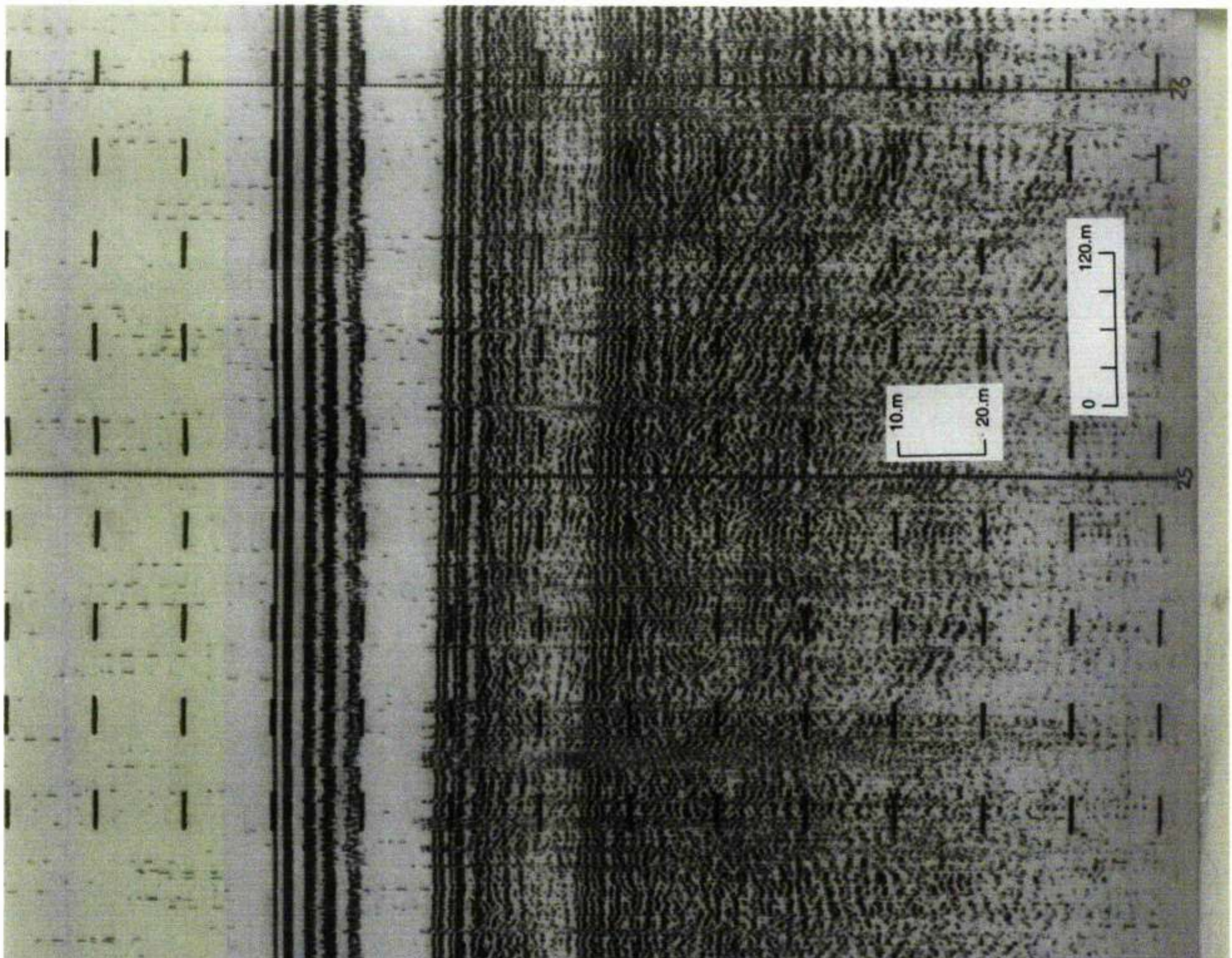


Plate 16 Flat sandy bed with sea weeds suspended in a water column.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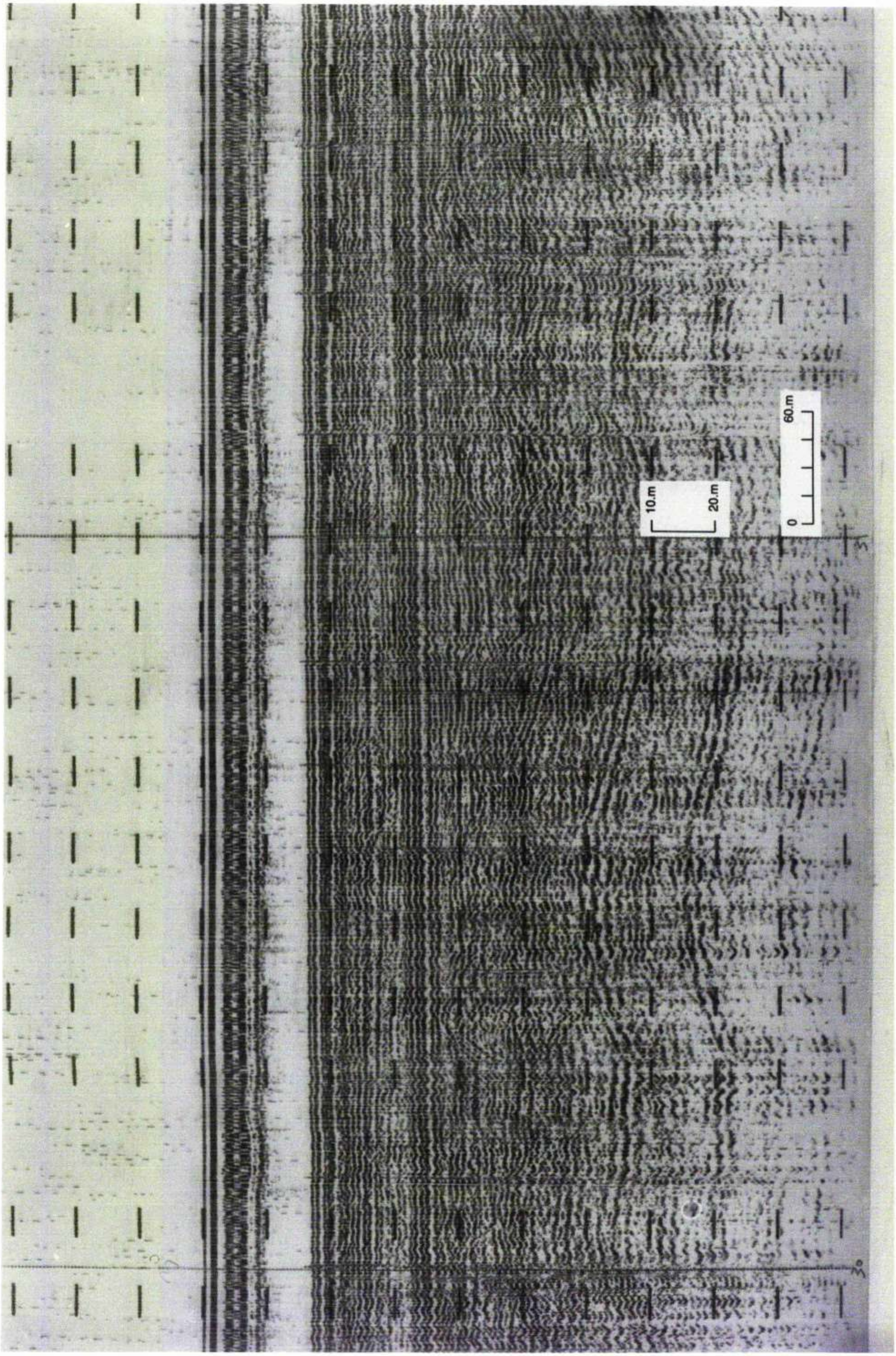
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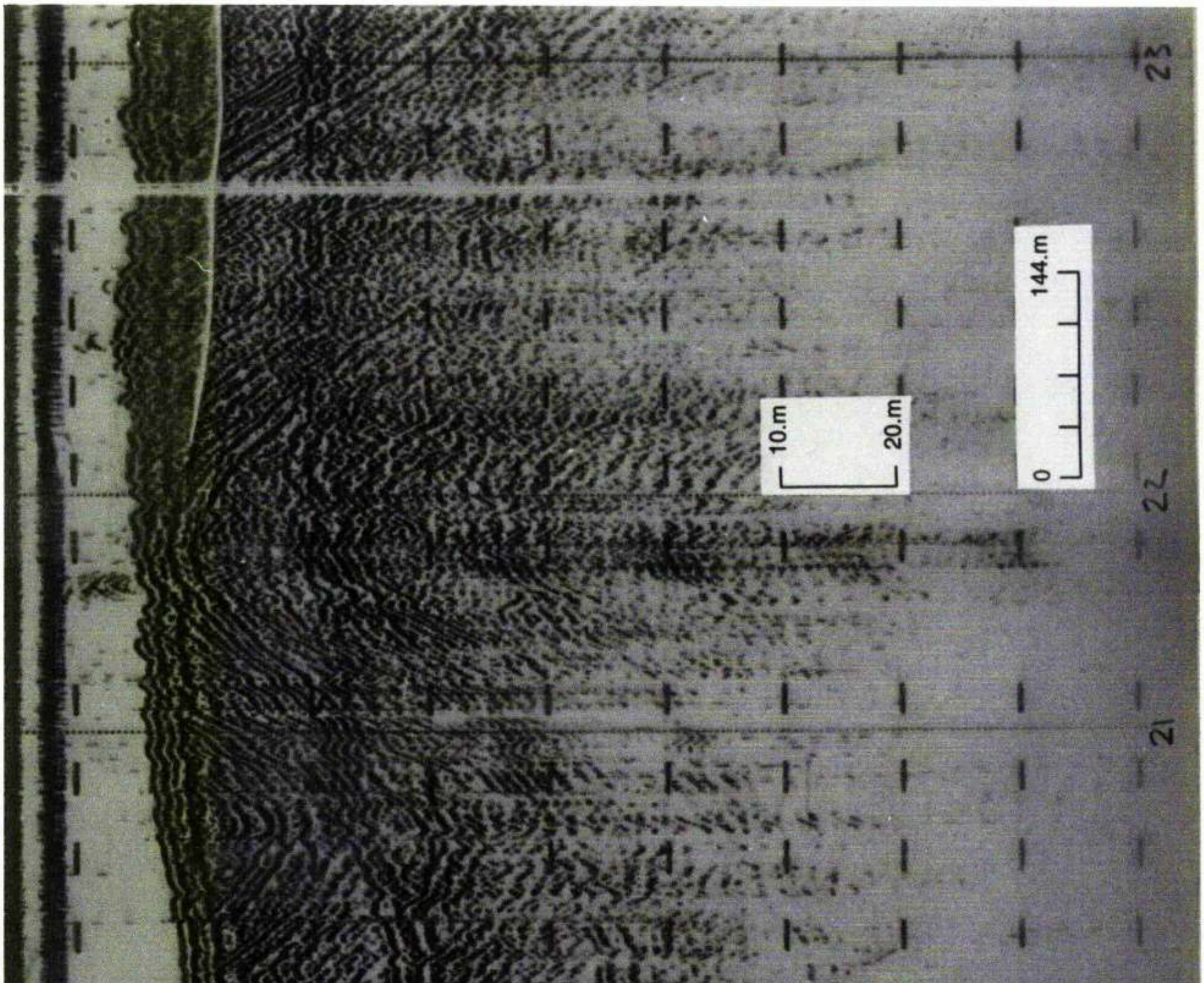
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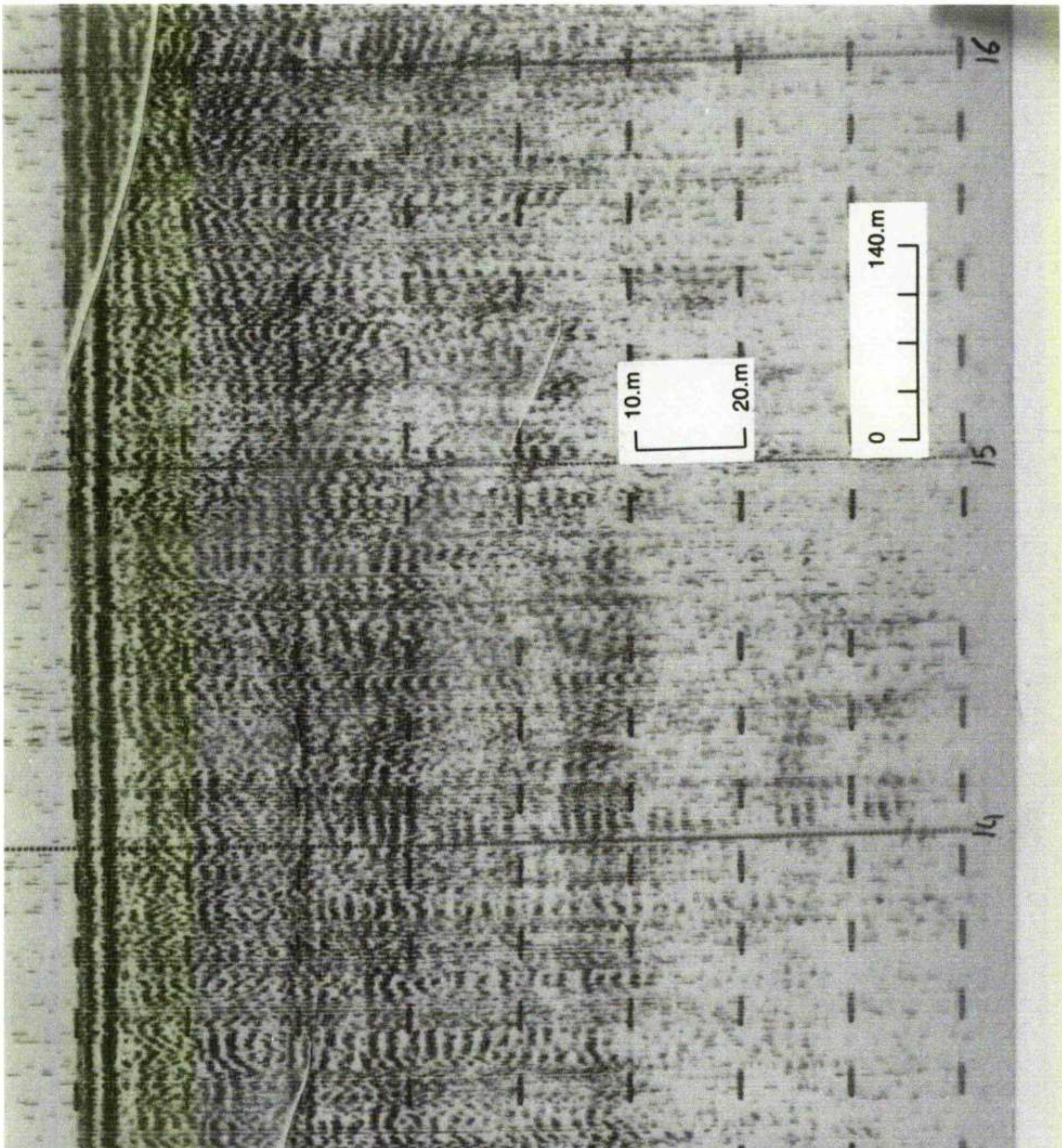

10.m
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16

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Appendix 1: Monthly distribution of wind speed and direction (hourly).

MAY-91 Knots	Angle 0-30	Angle 30-60	Angle 60-90	Angle 90-120	Angle 120-150	Angle 150-180	Angle 180-210	Angle 210-240	Angle 240-270	Angle 270-300	Angle 300-330	Angle 330-360
0-5	24	29	20	23	7	6	8	39	54	18	12	22
5-10	12	17	43	33	6	7	3	42	48	44	15	14
10-15	24	6	12	13	0	0	0	23	34	18	3	21
15-20	2	1	0	0	0	0	0	9	19	1	2	2
20-25	0	0	0	0	0	0	0	2	6	0	0	0
25-30	0	0	0	0	0	0	0	0	0	0	0	0
30-35	0	0	0	0	0	0	0	0	0	0	0	0
35-40	0	0	0	0	0	0	0	0	0	0	0	0
JUNE-91												
0-5	25	20	17	28	11	17	14	43	83	25	18	22
5-10	7	17	24	57	17	8	15	32	50	14	16	12
10-15	1	9	7	15	10	1	2	23	21	16	4	5
15-20	0	0	2	0	6	0	0	0	0	4	0	0
20-25	0	0	0	0	0	0	0	0	0	0	0	0
25-30	0	0	0	0	0	0	0	0	0	0	0	0
30-35	0	0	0	0	0	0	0	0	0	0	0	0
35-40	0	0	0	0	0	0	0	0	0	0	0	0
JULY-91												
0-5	31	43	45	33	19	5	9	21	37	11	0	4
5-10	12	81	46	54	10	8	11	48	39	6	0	0
10-15	0	59	6	4	4	1	8	23	26	1	0	0
15-20	0	11	2	0	0	0	1	9	7	0	0	0
20-25	0	0	0	0	0	0	0	3	6	0	0	0
25-30	0	0	0	0	0	0	0	0	0	0	0	0
30-35	0	0	0	0	0	0	0	0	0	0	0	0
35-40	0	0	0	0	0	0	0	0	0	0	0	0
AUG-91												
0-5	9	15	12	16	3	13	21	26	55	14	6	6
5-10	2	14	22	17	2	11	26	94	100	16	1	2
10-15	0	5	7	10	1	2	4	52	66	12	1	1
15-20	0	0	0	0	0	0	1	19	43	0	0	0
20-25	0	0	0	0	0	0	0	1	12	0	0	0
25-30	0	0	0	0	0	0	0	0	0	0	0	0
30-35	0	0	0	0	0	0	0	0	0	0	0	0
35-40	0	0	0	0	0	0	0	0	0	0	0	0
Sep-91												
0-5	16	16	21	26	6	3	11	30	87	27	17	10
5-10	6	4	20	39	13	4	15	52	56	1	2	11
10-15	4	15	7	15	4	0	3	33	41	1	0	0
15-20	0	0	0	0	0	0	3	4	19	0	0	0
20-25	0	0	0	0	0	0	0	0	5	0	0	0
25-30	0	0	0	0	0	0	0	0	0	0	0	0
30-35	0	0	0	0	0	0	0	0	0	0	0	0
35-40	0	0	0	0	0	0	0	0	0	0	0	0
Oct-91												
0-5	21	11	15	17	17	9	11	24	58	23	17	11
5-10	6	17	20	19	28	12	6	37	41	17	4	0
10-15	2	19	14	5	23	7	10	14	11	9	23	1
15-20	0	1	0	0	17	1	10	22	7	6	16	0
20-25	0	0	0	0	5	0	3	9	13	3	0	0
25-30	0	0	0	0	6	0	0	5	18	0	0	0
30-35	0	0	0	0	0	0	0	5	12	0	0	0
35-40	0	0	0	0	0	0	0	0	1	0	0	0
Nov-91												
0-5	5	4	5	5	8	2	7	11	40	0	104	3
5-10	2	0	1	8	9	15	15	56	78	0	213	3
10-15	12	1	4	3	11	13	19	69	49	0	190	2
15-20	1	1	1	0	0	5	16	52	39	0	117	8
20-25	0	3	3	0	0	2	3	21	10	0	44	1
25-30	0	0	0	0	0	0	0	0	0	0	0	0
30-35	0	0	0	0	0	0	0	0	0	0	0	0
35-40	0	0	0	0	0	0	0	0	1	0	0	0
Dec-91												
0-5	0	0	4	1	2	1	0	3	1	0	12	0
5-10	18	9	0	5	0	2	11	52	83	0	214	8
10-15	0	0	0	0	0	0	1	39	50	0	96	0
15-20	0	0	0	0	0	0	1	36	82	0	122	0
20-25	0	0	0	0	0	0	1	17	15	0	34	0
25-30	0	0	0	0	0	0	0	5	5	0	10	0
30-35	0	0	0	0	0	0	0	0	2	0	0	0
35-40	0	0	0	0	0	0	0	0	0	0	0	0

Appendix 1: Monthly distribution of wind and direction (hourly).

Jan-92												
Knots	Angle 0-30	Angle 30-60	Angle 60-90	Angle 90-120	Angle 120-150	Angle 150-180	Angle 180-210	Angle 210-240	Angle 240-270	Angle 270-300	Angle 300-330	Angle 330-360
0–5	2	2	7	10	9	5	19	37	106	56	17	11
5–10	8	2	1	3	9	1	2	35	139	14	4	7
10–15	9	12	0	0	0	0	0	13	74	8	2	0
15–20	0	5	0	0	0	0	0	7	38	1	1	0
20–25	0	0	0	0	0	0	0	5	14	1	0	0
25–30	0	0	0	0	0	0	0	0	12	0	0	0
30–35	0	0	0	0	0	0	0	3	11	0	0	0
35–40	0	0	0	0	0	0	0	0	14	0	0	0
Feb–92												
0–5	1	0	2	8	5	9	13	15	43	12	7	4
5–10	1	0	2	8	7	9	17	89	62	7	7	0
10–15	0	0	0	1	1	4	20	92	58	7	4	3
15–20	0	0	0	0	0	2	10	51	31	7	2	2
20–25	0	0	0	0	0	3	6	13	13	4	0	0
25–30	0	0	0	0	0	0	2	14	10	0	0	0
30–35	0	0	0	0	0	0	0	1	0	0	0	0
35-40	0	0	0	0	0	0	0	0	0	0	0	0
Mar-92												
0–5	2	5	7	12	2	4	3	21	16	5	3	4
5–10	6	3	0	6	4	2	20	72	43	13	12	5
10–15	2	27	24	14	0	7	53	172	90	40	14	29
15–20	0	2	5	6	0	0	14	61	38	15	0	12
20–25	0	3	18	5	0	0	0	10	13	0	0	0
25–30	0	15	0	0	0	0	0	1	5	0	0	0
30–35	0	0	0	0	0	0	0	0	0	0	0	0
35-40	0	0	0	0	0	0	0	0	0	0	0	0
Apr-92												
0–5	6	4	13	13	14	6	11	32	19	6	10	2
5–10	3	10	13	19	11	12	20	60	61	37	7	9
10–15	20	2	9	18	18	5	12	48	47	19	5	10
15–20	3	0	8	0	1	1	8	30	5	4	1	5
20–25	11	1	0	0	0	0	0	5	4	0	0	4
25–30	5	1	0	0	0	0	0	2	5	0	0	0
30–35	0	0	0	0	0	0	0	0	1	0	0	0
35-40	0	0	0	0	0	0	0	0	0	0	0	0
May-92												
0–5	22	15	12	14	4	4	1	21	27	14	3	6
5–10	8	66	53	34	10	1	7	39	20	2	3	1
10–15	0	68	38	14	0	2	5	47	18	4	7	0
15-20	0	11	13	2	0	0	1	29	17	6	3	0
20–25	0	0	0	0	0	0	0	2	11	0	0	0
25–30	0	11	2	0	0	0	0	0	0	0	0	0
30-35	0	0	0	0	0	0	0	0	0	0	0	0
35-40	0	0	0	0	0	0	0	0	0	0	0	0
Jun–92												
0–5	9	16	17	5	3	4	2	26	21	5	2	6
5–10	9	53	85	39	10	6	1	51	46	18	6	6
10–15	27	47	47	14	2	1	6	25	46	13	0	5
15-20	0	0	2	0	0	0	0	6	5	1	0	0
20–25	0	0	0	0	0	0	0	0	0	0	0	0
25–30	0	0	0	0	0	0	0	0	0	0	0	0
30-35	0	0	0	0	0	0	0	0	0	0	0	0
35-40	0	0	0	0	0	0	0	0	0	0	0	0
July–92												
0–5	7	13	16	24	10	13	24	37	30	6	7	11
5–10	15	3	29	39	7	6	32	96	29	7	0	0
10–15	17	11	20	21	3	8	18	78	34	4	1	0
15–20	0	0	0	3	0	0	1	35	22	0	0	0
20–25	0	0	0	0	0	0	0	0	4	0	0	0
25-30	0	0	0	0	0	0	0	0	0	0	0	0
30-35	0	0	0	0	0	0	0	0	0	0	0	0
35-40	0	0	0	0	0	0	0	0	0	0	0	0
Total	0-30	30-60	60-90	90-120	120-150	150-180	180-210	210-240	240-270	270-300	300-330	330-360
0–5	205	249	298	350	262	279	357	615	907	522	461	479
5–10	118	300	363	377	142	91	193	810	857	196	395	78
10–15	108	280	192	152	75	53	157	738	694	152	373	78
15–20	17	32	36	14	35	17	69	387	382	45	337	23
20–25	12	5	19	5	5	8	26	119	155	8	151	12
25-30	5	30	5	0	6	2	5	48	65	0	54	1
30-35	0	0	0	0	0	0	0	9	26	0	0	0
35-40	0	0	0	0	0	0	0	0	15	0	0	0

Appendix 2: Weight percentage of coarse and fine sand in samples.

Sample No.	Wt. % of -1 to 2 Phi	Wt. % of 2.25 to 4 Phi	Sample No.	Wt. % of -1 to 2 Phi	Wt. % of 2.25 to 4 Phi
1	5.19	94.81	58	5.9	94.1
2	4.69	95.3	59	2.39	97.61
3	4.27	95.37	60	3.52	96.48
4	6.45	93.55	61	4.36	95.64
5	4.2	95.8	62	4.79	95.21
6	4.48	95.52	63	2.95	97.05
7	6.3	93.7	64	4.46	95.54
8	7.65	92.35	65	4.59	95.41
9	8.01	91.99	66	5.47	94.53
10	8.43	91.57	67	3.04	96.96
11	9.43	90.57	68	2.42	97.58
12	8.81	91.19	69	2.9	97.1
13	3.94	96.05	70	3.14	96.86
14	5.84	94.16	71	8.02	91.98
15	6.26	93.74	72	5.52	94.48
16	13.28	86.72	73	2.53	97.47
17	26.67	73.33	74	1.8	98.2
18	5.13	94.87	75	1.1	98.9
19	4.58	95.42	76	2.45	97.55
20	4.29	95.71	77	3.85	96.15
21	10.3	89.7	78	4.15	95.85
22	8.41	91.59	79	5.73	94.27
23	5.06	94.94	80	11.94	88.06
24	5.5	94.5	81	23.09	76.91
25	4.32	95.68	82	10.63	89.37
26	3.76	96.24	83	9.45	90.55
27	12.77	87.23	84	1.07	98.93
28	3.85	96.15	85	2.15	97.85
29	4.46	95.54	86	5.39	94.61
30	6.4	93.6	87	5.08	94.92
31	3.7	96.3	88	7.17	92.83
32	5.52	94.48	89	10.23	89.77
33	5.87	94.13	90	5.16	94.84
34	4.95	95.05	91	6.79	93.21
35	5.44	94.56	92	3.4	96.6
36	9.9	90.1	93	4.38	95.62
37	2.88	97.12	94	20.26	79.74
38	3.48	96.52	95	23.65	79.9
39	8.15	91.85	96	39.82	60.18
40	4.09	95.91	97	61.6	38.4
41	3.07	96.93	98	21.83	78.17
42	3.2	96.8	99	39.41	60.59
43	2.28	97.72	100	8.98	91.02
44	3.14	96.86	101	44.29	55.71
45	1.97	98.03	102	53.42	46.58
46	4.42	95.58	103	92.34	7.66
47	2.7	97.3	104	3.66	96.34
48	6.48	93.52	105	23.45	76.55
49	3.72	96.28	106	6.69	93.31
50	7.72	92.28	107	9.46	90.54
51	8.85	91.15	108	7.56	92.44
52	7.84	92.16	109	7.11	92.89
53	7.69	92.31	110	4.41	95.59
54	4.01	95.99	111	4.98	95.02
55	4.08	95.92	112	5.04	94.96
56	4.32	95.68	113	7.72	92.28
57	4.95	95.05			

Appendix 3: Statistical parameters of grain size samples.

SITE NO	Md	Mz	Sorting	Sk	Kg	Depth (m)	SITE NO	Md	Mz	Sorting	Sk	Kg	Depth (m)
1	2.95	2.93	0.3	-0.31	2.13	3.5	58	2.76	2.76	0.32	-0.13	2.17	11.5
2	2.95	2.94	0.28	-0.1	1.89	4	59	2.82	2.86	0.25	0.36	2.17	12.3
3	2.92	2.92	0.29	0.06	1.35	6.4	60	2.78	2.77	0.25	-0.01	1.86	12.3
4	2.9	2.89	0.47	-0.26	2.36	6.4	61	2.8	2.8	0.26	-0.04	2.3	12.3
5	2.92	2.9	0.32	-0.08	1.36	6.4	62	2.79	2.81	0.27	0.14	3.01	12
6	2.92	2.89	0.36	-0.11	1.72	6.4	63	2.85	2.94	0.28	0.47	2.25	10.5
7	2.94	2.93	0.47	-0.14	1.55	5.9	64	2.84	2.83	0.34	-0.06	1.91	13.2
8	2.92	2.9	0.49	-0.17	1.48	6.5	65	2.9	2.88	0.36	-0.08	1.69	12.3
9	2.91	2.87	0.48	-0.22	1.57	8.2	66	2.84	2.86	0.41	-0.01	1.97	12.3
10	2.92	2.89	0.49	-0.21	1.61	8.5	67	2.81	2.82	0.32	0.11	2.13	13
11	2.92	2.87	0.48	-0.31	1.93	9	68	2.89	2.91	0.37	0.12	1.41	15
12	2.88	2.84	0.47	-0.25	2.08	9.6	69	2.9	2.89	0.36	0.03	1.12	15
13	2.91	2.9	0.28	0	1.6	9.2	70	2.9	2.98	0.39	0.21	1.41	16
14	2.72	2.73	0.39	-0.05	2.19	9	71	2.95	3.02	0.54	-0.04	2.06	18.7
15	2.72	2.73	0.39	-0.08	1.64	8.2	72	2.98	3.05	0.47	0.01	2.05	18.7
16	2.65	2.62	0.52	-0.3	1.97	7	73	3	3.08	0.37	0.26	1.21	18.7
17	2.4	2.14	0.64	-0.72	1.87	2.1	74	3.13	3.22	0.4	0.27	0.67	18.7
18	2.85	2.84	0.33	-0.13	1.85	8.2	75	3.12	3.24	0.37	0.37	0.62	18.7
19	2.82	2.84	0.29	0.04	1.47	6.8	76	3.08	3.19	0.37	0.37	0.72	18
20	2.8	2.82	0.21	-0.01	1.42	6	77	3	3.1	0.37	0.33	1.5	17
21	2.55	2.55	0.37	-0.24	2.13	2.7	78	2.85	2.88	0.32	0.16	1.8	15
22	2.59	2.58	0.36	-0.24	1.87	2.2	79	2.7	2.73	0.34	0.03	1.89	14
23	2.56	2.55	0.27	-0.09	1.32	1	80	2.64	2.61	0.42	-0.21	1.77	6.4
24	2.45	2.53	0.26	0.25	1.72	0.4	81	2.56	2.37	0.69	-0.5	1.65	7.3
25	2.75	2.78	0.22	0.11	1.31	0.8	82	2.68	2.64	0.44	-0.27	1.78	7.9
26	2.75	2.75	0.2	-0.02	1.17	2	83	2.69	2.67	0.4	-0.2	1.57	8.5
27	2.68	2.6	0.41	-0.36	1.59	2	84	2.9	3	0.38	0.36	1.32	11.5
28	2.71	2.75	0.22	0.22	1.23	0.9	85	2.95	2.99	0.37	0.19	1.38	12.2
29	2.78	2.82	0.24	0.08	1.57	0.9	86	2.9	2.92	0.43	0.01	1.74	12.5
30	2.75	2.78	0.35	-0.12	2.3	1.5	87	2.92	2.93	0.45	-0.01	1.5	13.5
31	2.8	2.83	0.23	0.08	1.7	2.2	88	2.92	2.93	0.53	-0.07	0.79	14.5
32	2.88	2.91	0.37	0	2.28	2.5	89	2.87	2.87	0.58	-0.1	1.51	13.5
33	2.85	2.92	0.39	0.13	2.03	5	90	2.78	2.84	0.48	0.15	1.18	13.2
34	3.05	3.11	0.44	0.01	1.79	7.5	91	2.78	2.85	0.53	0.13	1.22	13.2
35	3.11	3.13	0.45	-0.12	1.66	10.5	92	2.9	2.89	0.37	0.03	1.64	12.5
36	2.8	2.77	0.55	-0.1	1.47	7.7	93	2.89	2.91	0.35	0.11	1.85	12.3
37	2.72	2.76	0.22	0.23	1.54	5	94	2.4	2.35	0.47	-0.34	1.12	6
38	2.85	2.86	0.21	-0.03	1.2	2.7	95	2.2	2.22	0.29	-0.03	1.02	3.5
39	2.92	2.93	0.43	-0.26	2.87	2.5	96	2.12	2.13	0.29	0	1.23	3.1
40	2.85	2.88	0.25	0.17	1.76	5.5	97	1.92	1.91	0.37	-0.12	1.13	3.1
41	2.88	2.9	0.27	0.19	1.37	6	98	2.42	2.32	0.48	-0.41	1.46	2.5
42	2.95	2.99	0.31	0.17	1.1	7.3	99	2.2	2.03	0.56	-0.48	0.98	1
43	2.98	3.01	0.33	0.17	1.17	10	100	2.45	2.44	0.29	-0.1	2.19	6.1
44	3	3.12	0.42	0.32	0.98	10.5	101	2.08	2.03	0.44	-0.24	0.9	2.4
45	3.05	3.14	0.43	0.23	0.72	11	102	1.98	1.89	0.54	-0.3	0.92	2.4
46	3.05	3.08	0.49	0	0.95	11.4	103	1.3	1.27	0.52	-0.08	1.13	2
47	2.85	2.98	0.5	0.29	1.22	11.4	104	2.85	2.87	0.32	0.13	1.71	7.5
48	2.83	2.95	0.56	0.18	1.32	11.4	105	2.35	2.29	0.48	-0.27	1.3	5.2
49	2.88	2.95	0.49	0.18	1.19	11.8	106	2.55	2.55	0.3	-0.07	1.5	7.5
50	2.82	2.8	0.51	-0.1	1.72	12	107	2.68	2.7	0.42	-0.05	1.77	11.4
51	2.81	2.77	0.59	-0.09	1.51	12	108	2.72	2.75	0.45	0.03	2.05	11.4
52	2.82	2.88	0.59	0.05	1.34	13.2	109	2.75	2.72	0.42	-0.16	2.03	11.4
53	2.83	2.82	0.51	-0.06	1.52	13.2	110	2.78	2.74	0.28	-0.24	1.8	11.4
54	2.85	2.87	0.39	0.1	1.66	12.3	111	2.6	2.62	0.21	-0.09	1.36	0.4
55	2.85	2.9	0.35	0.24	1.86	12.3	112	2.7	2.75	0.24	0.2	3.93	0.9
56	2.8	2.83	0.28	0.2	3.33	12.3	113	2.82	2.8	0.51	-0.1	1.72	5.5
57	2.76	2.77	0.31	-0.03	2.15	12.5							

Appendix 4: Mean data for histrograms of grain size variation across St Andrews Bay.

Size in Phi	His 1	His 2	His 3	His 4	His 5	His 6
-1.00	0.16	0.06	0.19	0.37	0.18	0.04
-0.50	0.49	0.23	0.15	0.28	0.21	0.09
0.00	0.84	0.51	0.61	0.32	0.30	0.17
0.50	1.31	0.83	0.70	0.41	0.46	0.40
1.00	4.18	0.92	0.85	0.49	0.60	0.89
1.50	8.69	0.97	1.04	0.69	1.08	1.45
2.00	19.74	1.94	2.69	1.82	3.60	3.32
2.25	13.50	2.08	1.65	1.10	3.33	2.20
2.50	24.97	9.46	7.15	4.49	9.84	9.94
2.75	18.43	30.43	22.21	23.35	22.39	19.61
3.00	5.63	34.83	32.08	44.26	26.12	31.79
3.25	1.32	13.64	19.70	11.55	15.29	17.41
3.50	0.37	2.14	4.85	4.32	5.39	3.62
3.75	0.14	1.53	4.74	3.80	6.62	8.20
4.00	0.05	0.27	1.39	2.45	4.59	3.39

Appendix 5: Quartile deviation - Median diameter values for Fig. 3.8 & Fig. 3.10.

Site No.	Q25 (mm)	Q50 (mm)	Q75 (mm)	QDa	Site No.	Q25 (mm)	Q50 (mm)	Q75 (mm)	QDa
1	0.143	0.13	0.124	0.0095	58	0.164	0.148	0.135	0.0145
2	0.143	0.13	0.124	0.0095	59	0.149	0.142	0.13	0.0095
3	0.149	0.133	0.118	0.0155	60	0.16	0.147	0.138	0.011
4	0.155	0.135	0.119	0.018	61	0.155	0.144	0.135	0.01
5	0.15	0.133	0.118	0.016	62	0.155	0.145	0.138	0.0085
6	0.5	0.133	0.2	0.15	63	0.149	0.14	0.129	0.01
7	0.155	0.131	0.11	0.0225	64	0.155	0.14	0.125	0.015
8	0.16	0.133	0.11	0.025	65	0.155	0.135	0.124	0.0155
9	0.16	0.134	0.113	0.0235	66	0.155	0.14	0.124	0.0155
10	0.16	0.133	0.115	0.0225	67	0.155	0.143	0.13	0.0125
11	0.158	0.133	0.118	0.02	68	0.155	0.137	0.118	0.0185
12	0.16	0.138	0.124	0.018	69	0.164	0.135	0.118	0.023
13	0.149	0.134	0.124	0.0125	70	0.149	0.135	0.118	0.0155
14	0.169	0.152	0.135	0.017	71	0.149	0.13	0.112	0.0185
15	0.17	0.152	0.131	0.0195	72	0.143	0.127	0.113	0.015
16	0.188	0.16	0.14	0.024	73	0.145	0.125	0.11	0.0175
17	0.269	0.19	0.185	0.042	74	0.135	0.116	0.081	0.027
18	0.155	0.14	0.125	0.015	75	0.135	0.116	0.082	0.0265
19	0.156	0.142	0.125	0.0155	76	0.135	0.12	0.088	0.0235
20	0.155	0.144	0.13	0.0125	77	0.135	0.125	0.105	0.015
21	0.19	0.17	0.154	0.018	78	0.155	0.14	0.125	0.015
22	0.19	0.165	0.149	0.0205	79	0.169	0.155	0.138	0.0155
23	0.19	0.169	0.154	0.018	80	0.182	0.16	0.14	0.021
24	0.191	0.185	0.16	0.0155	81	0.23	0.17	0.148	0.041
25	0.16	0.149	0.135	0.0125	82	0.184	0.156	0.14	0.022
26	0.162	0.149	0.135	0.0135	83	0.183	0.156	0.138	0.0225
27	0.185	0.158	0.14	0.0225	84	0.149	0.135	0.113	0.018
28	0.166	0.154	0.138	0.014	85	0.149	0.13	0.113	0.018
29	0.155	0.147	0.133	0.011	86	0.155	0.135	0.118	0.0185
30	0.162	0.149	0.134	0.014	87	0.155	0.133	0.112	0.0215
31	0.155	0.144	0.134	0.0105	88	0.164	0.133	0.156	0.004
32	0.149	0.138	0.124	0.0125	89	0.169	0.138	0.11	0.0295
33	0.149	0.14	0.118	0.0155	90	0.172	0.147	0.113	0.0295
34	0.135	0.124	0.104	0.0155	91	0.174	0.147	0.115	0.0295
35	0.135	0.118	0.09	0.0225	92	0.155	0.135	0.124	0.0155
36	0.182	0.144	0.124	0.029	93	0.15	0.137	0.124	0.013
37	0.16	0.152	0.139	0.0105	94	0.24	0.19	0.156	0.042
38	0.155	0.14	0.129	0.013	95	0.25	0.22	0.19	0.03
39	0.149	0.133	0.124	0.0125	96	0.255	0.232	0.204	0.0255
40	0.149	0.14	0.125	0.012	97	0.31	0.26	0.229	0.0405
41	0.149	0.138	0.124	0.0125	98	0.235	0.187	0.165	0.035
42	0.143	0.13	0.11	0.0165	99	0.32	0.22	0.187	0.0665
43	0.147	0.127	0.11	0.0185	100	0.198	0.185	0.172	0.013
44	0.14	0.125	0.094	0.023	101	0.308	0.24	0.198	0.055
45	0.142	0.124	0.085	0.0285	102	0.35	0.252	0.22	0.065
46	0.149	0.124	0.092	0.0285	103	0.52	0.405	0.34	0.09
47	0.16	0.14	0.115	0.0225	104	0.155	0.14	0.126	0.0145
48	0.169	0.141	0.115	0.027	105	0.24	0.195	0.165	0.0375
49	0.167	0.138	0.115	0.026	106	0.19	0.17	0.152	0.019
50	0.172	0.142	0.124	0.024	107	0.177	0.158	0.135	0.021
51	0.175	0.143	0.119	0.028	108	0.177	0.152	0.139	0.019
52	0.177	0.142	0.118	0.0295	109	0.172	0.149	0.135	0.0185
53	0.172	0.141	0.12	0.026	110	0.16	0.147	0.135	0.0125
54	0.156	0.14	0.124	0.016	111	0.177	0.167	0.149	0.014
55	0.15	0.14	0.124	0.013	112	0.155	0.155	0.142	0.0065
56	0.15	0.144	0.135	0.0075	113	0.72	0.142	0.124	0.298
57	0.16	0.148	0.135	0.0125					

Appendix 6: Current meter data for Stations 1, 2, 3 and 4.

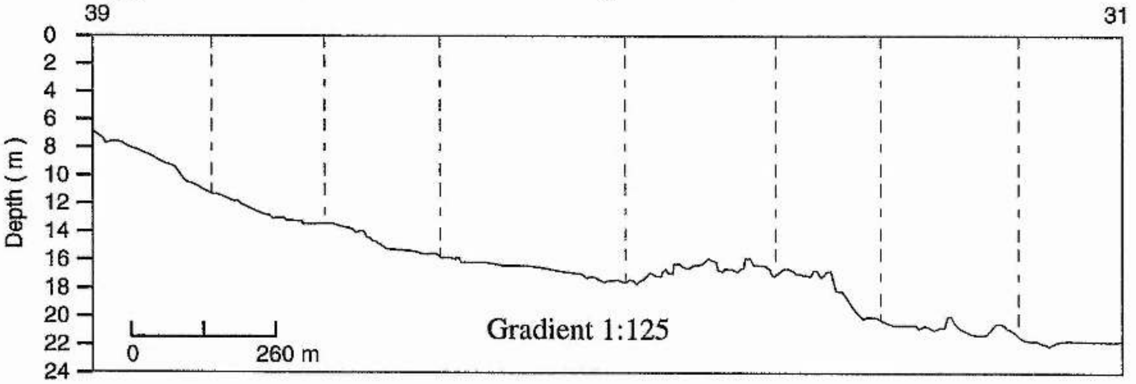
Station 1													
Atay													
Cm/sec	Angle	Angle	Angle	Angle	Angle	Angle	Angle	Angle	Angle	Angle	Angle	Angle	
	0-30	30-60	60-90	90-120	120-150	150-180	180-210	210-240	240-270	270-300	300-330	330-360	
0-5	28	73	70	40	20	30	75	31	61	56	91	85	
5-10	86	244	209	78	22	5	73	99	196	197	193	149	
10-15	31	91	214	14	1	1	32	151	211	149	51	39	
15-20	4	21	28	4	2		7	72	99	34	8		
20-25			2	1			1	17	23	1			
25-30									3				
30-35													
35-40													
40-45													
45-50													
50-55													
55-60													
60-65													
65-70				2		27							
	149	429	523	139	45	63	188	370	593	437	343	273	3552

Station 2													
Forexcel													
	0-30	30-60	60-90	90-120	120-150	150-180	180-210	210-240	240-270	270-300	300-330	330-360	
0-5	177	124	172	199	24	9	86	81	18	67	21	15	
5-10	187	289	301	190	18	8	93	105	21	73	18	16	
10-15	102	148	174	64	6	9	33	105	24	60	8	5	
15-20	57	81	74	14	2	1	15	41	15	54	23	12	
20-25	13	27	28	11			4	10	12	21	11	2	
25-30	5	7	10	1	1		1	3	7	10		2	
30-35		1	5							9	1		
35-40		1	2							2			
40-45			1										
45-50											1		
50-55													
55-60													
60-65			1										
	541	678	768	479	51	27	231	345	97	296	83	52	3648

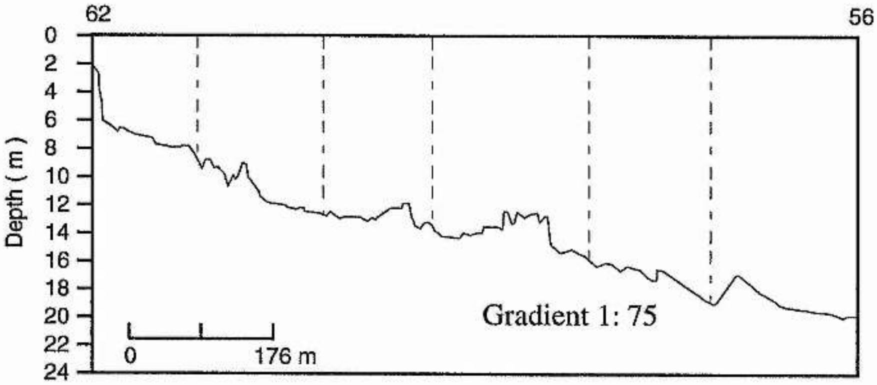
Station 3													
July 92 1km													
	0-30	30-60	60-90	90-120	120-150	150-180	180-210	210-240	240-270	270-300	300-330	330-360	
0-5	29	236	47	79	44	6	93	267	118	232	60	18	
5-10	6	128	47	20	3	1	18	189	180	145	9	5	
10-15		34	19	2			1	103	48	15			
15-20		12	1					39	16	1			
20-25								15					
25-30								2					
30-35													
35-40													
	35	410	114	101	47	7	112	615	362	393	69	23	2288

Station 4													
July 92 2km													
	0-30	30-60	60-90	90-120	120-150	150-180	180-210	210-240	240-270	270-300	300-330	330-360	
0-5	84	60	139	93	339	143	40	39	93	269	89	132	
5-10	67	72	60	19	159	66	50	47	94	123	40	27	
10-15	1	6	7	11	40	4	28	16	16	15			
15-20							1	6	1				
20-25													
25-30													
30-35													
35-40													
	152	138	206	123	538	213	119	108	204	407	129	159	2496

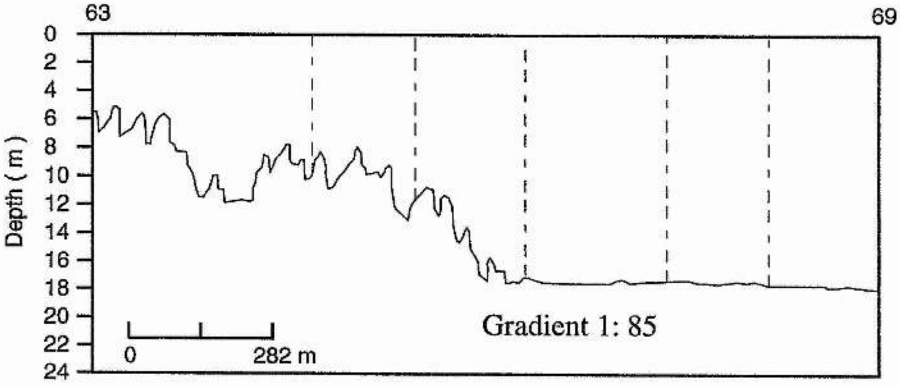
Appendix 7: Echo sounder profiles 1 to 4.



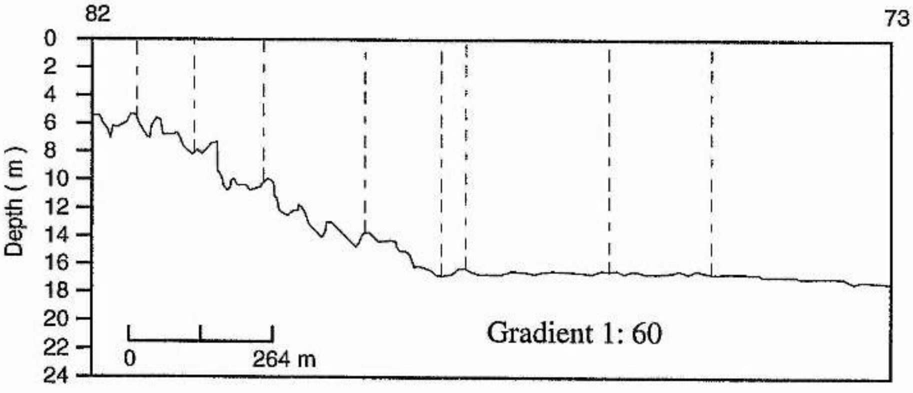
Profile 1



Profile 2

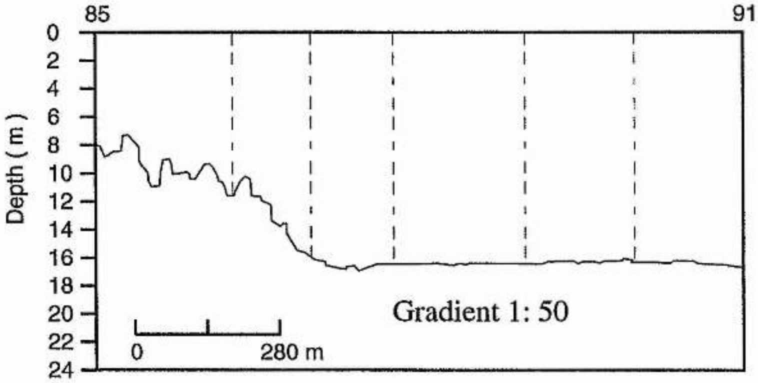


Profile 3

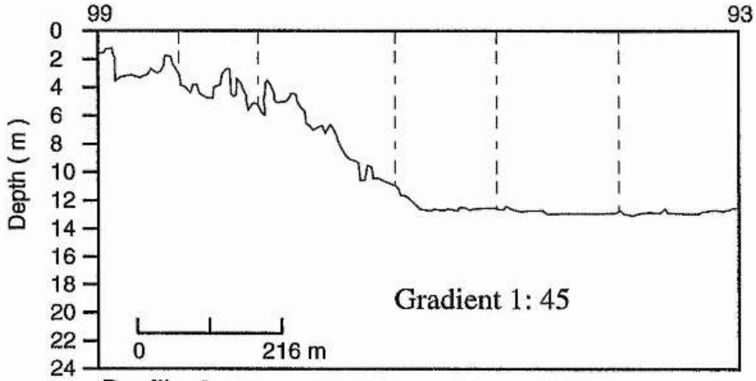


Profile 4

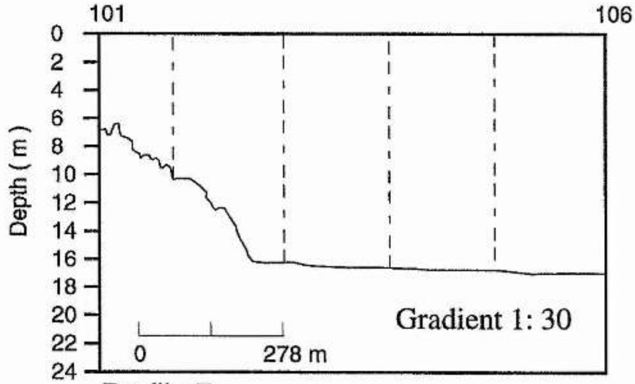
Appendix 7: Echo sounder profiles 5 to 8.



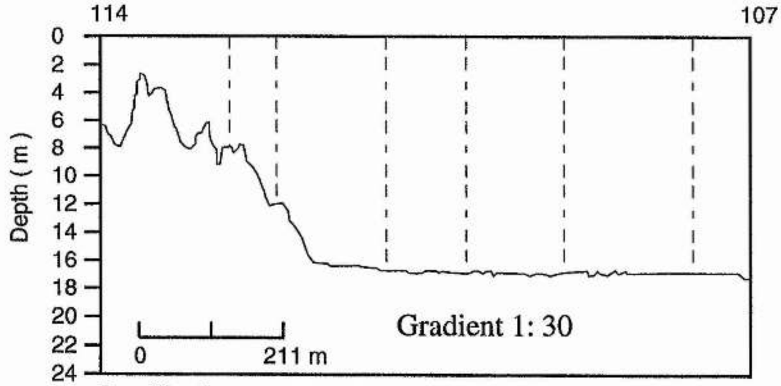
Profile 5



Profile 6

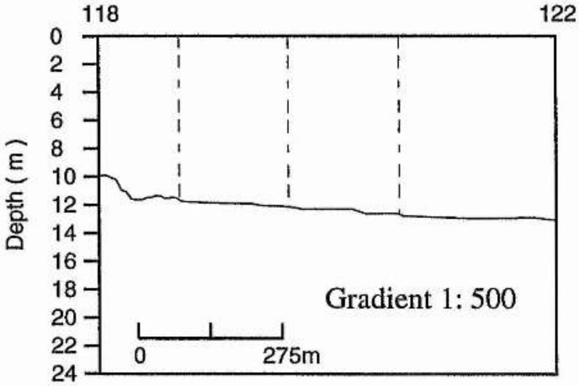


Profile 7

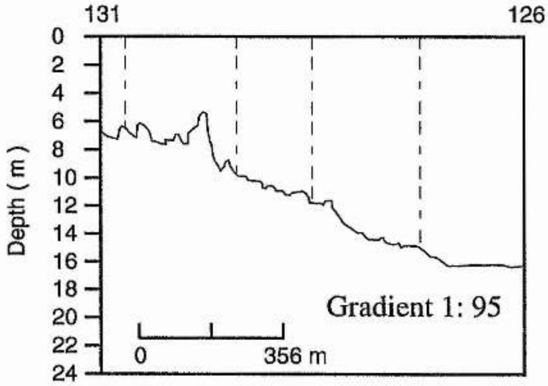


Profile 8

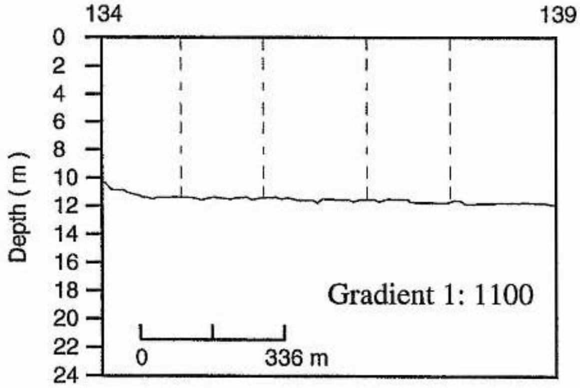
Appendix 7: Echo sounder profiles 9 to 12.



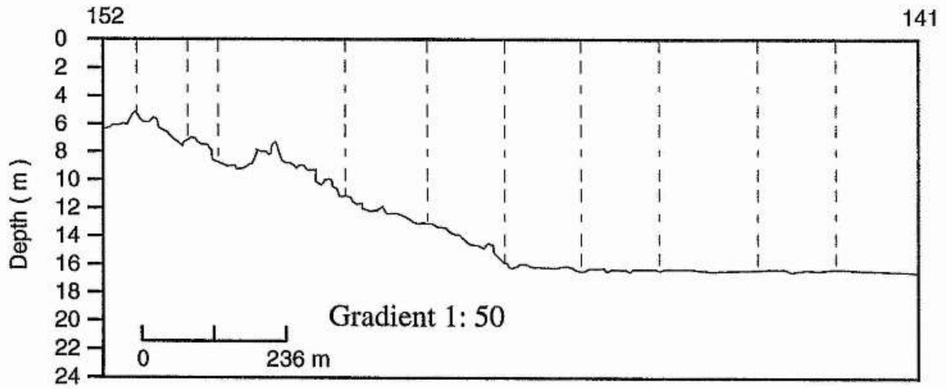
Profile 9



Profile 10

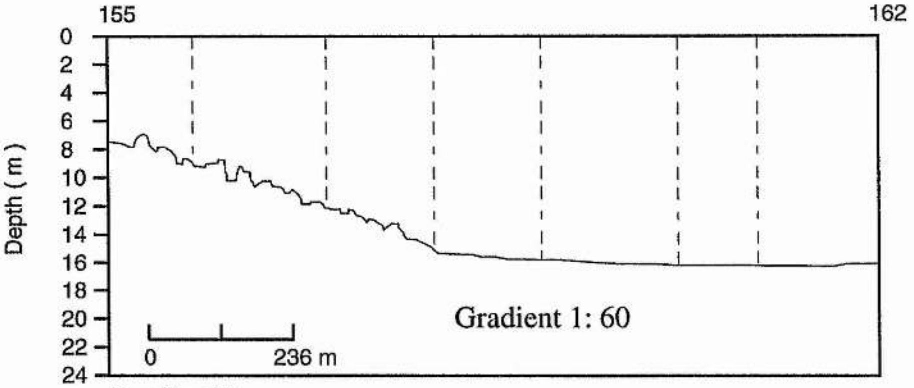


Profile 11

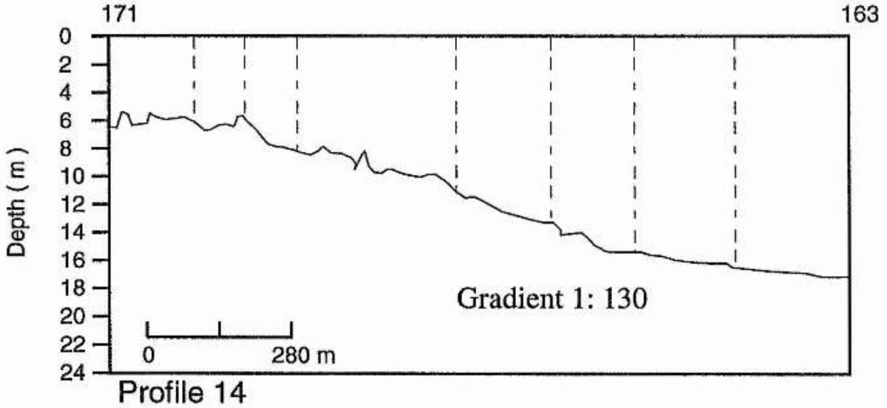


Profile 12

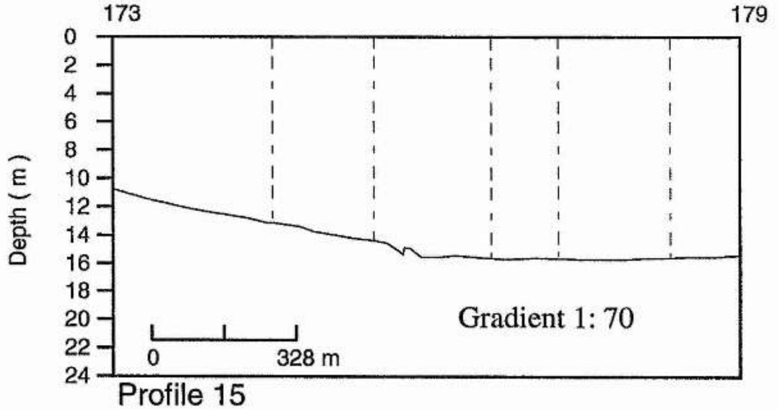
Appendix 7: Echo sounder profiles 13 to 16.



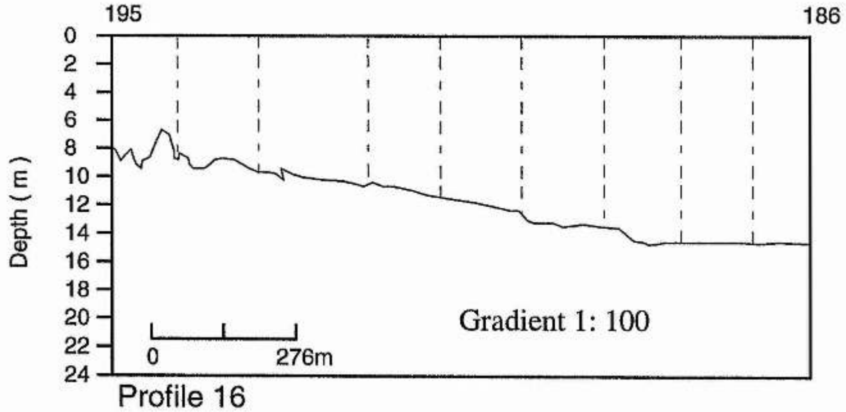
Profile 13



Profile 14

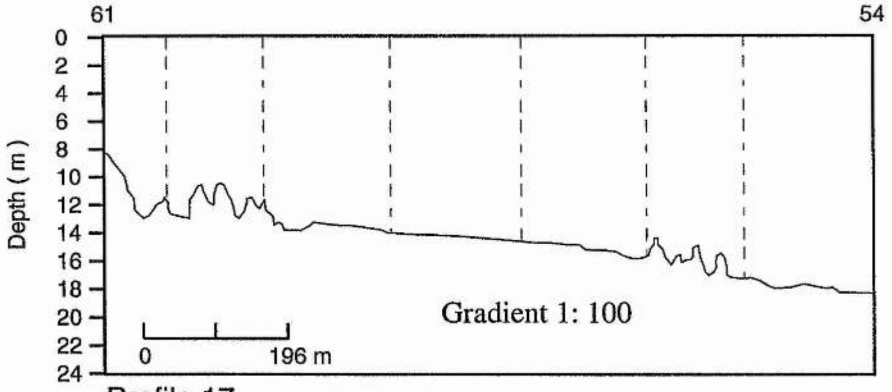


Profile 15

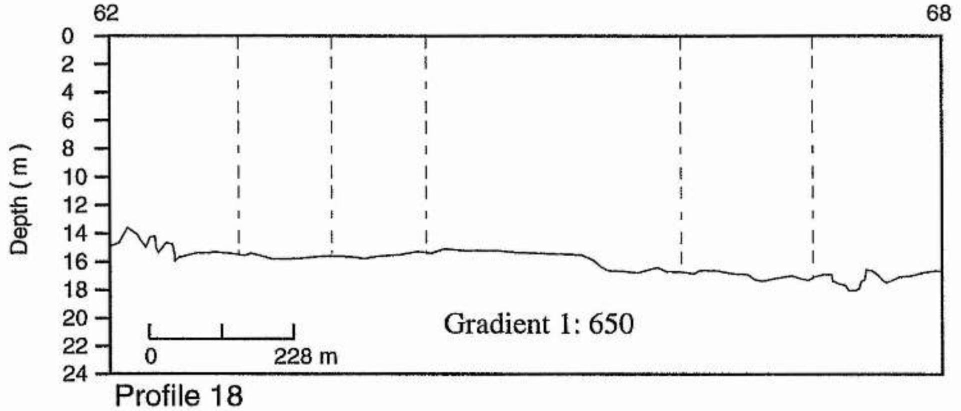


Profile 16

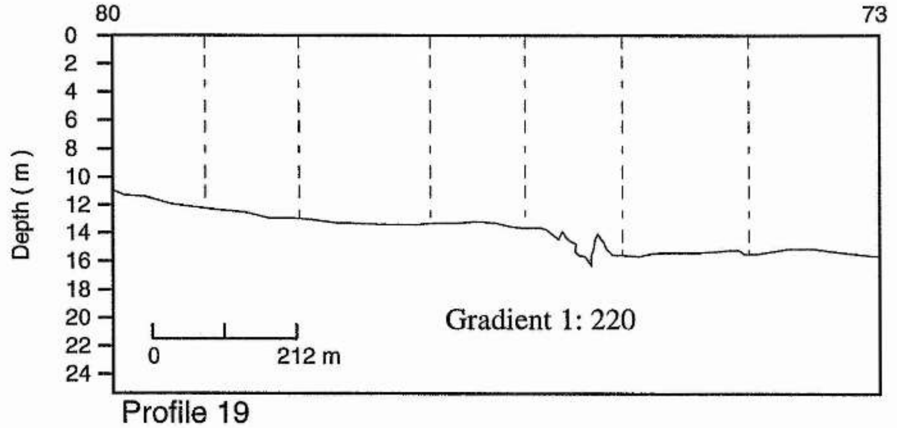
Appendix 7: Echo sounder profiles 17 to 20.



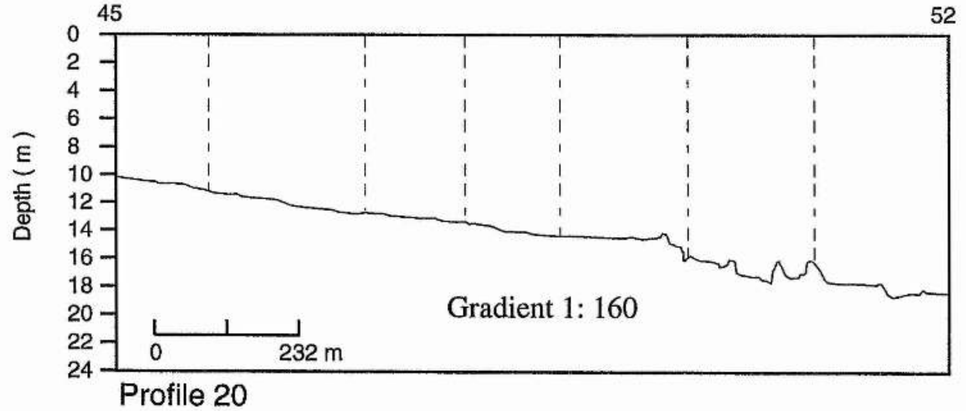
Profile 17



Profile 18

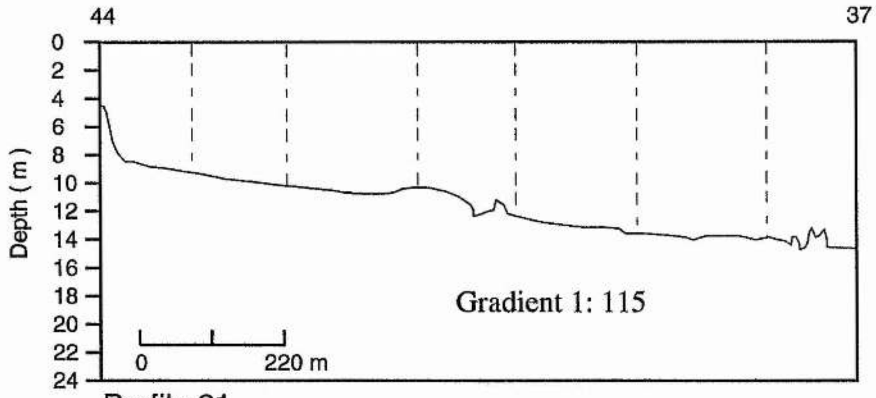


Profile 19

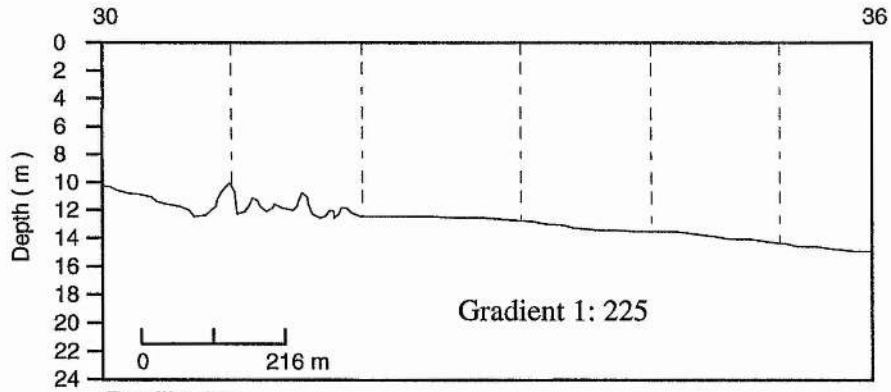


Profile 20

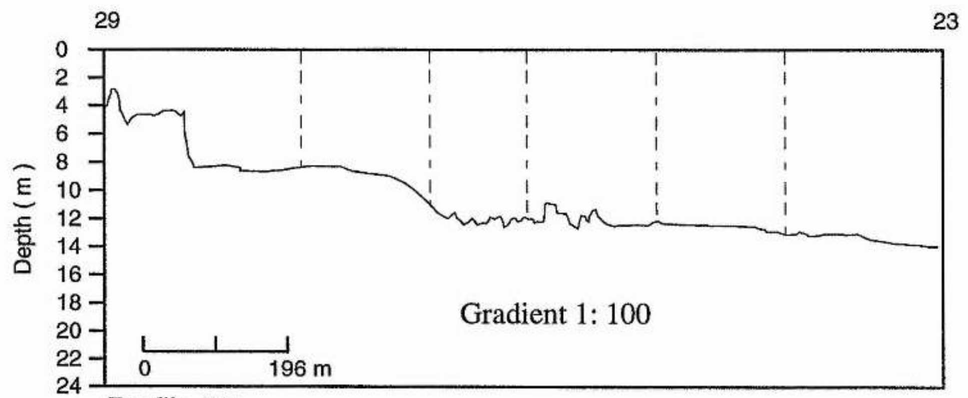
Appendix 7: Echo sounder profiles 21 to 24.



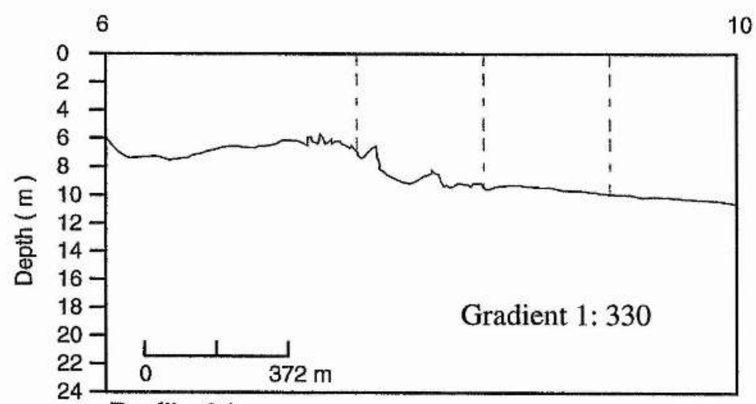
Profile 21



Profile 22



Profile 23



Profile 24