THE RAISED SHORELINES AND DEGLACIATION OF THE LOCH LONG/LOCH FYNE AREA, WESTERN SCOTLAND.

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SUMMARY

The objective of the research reported in this thesis was to elucidate the mode of disappearance of the last ice-sheet to cover the Loch Long/Loch Fyne area in the SW Highlands, and to establish the sequence of raised shorelines that has been formed as a result of the interplay between eustatic and isostatically-induced sea-level changes consequent upon the melting of the ice.

The study was geomorphological in orientation and a methodology was adopted that involved mapping, at a scale of 1:10,560, all glacial, fluvial and marine landforms below approx. 75 - 100 m O.D., and the subsequent accurate instrumental surveying of all relevant landforms. A certain amount of mapping back from the coast was also carried out where relevant. The errors inherent in the methods adopted and in the use of various types of marine landforms were quantitatively assessed and it was concluded that raised shorelines could be reconstructed with an accuracy of ± 0.54 m using intertidal deltas and ± 0.61 m using marine erosional features.

The southern part of the study area was deglaciated first at <u>ca</u>. 13,000 yr BP. The dominant mode of deglaciation was that of rapid retreat in the sea lochs (possibly as much as 500 m/yr) due to calving that left isolated dead-ice masses in various side valleys. This retreat was punctuated by two major periods of stillstand or readvance, the Otter Ferry Stage (<u>ca</u>. 12,900 $\pm$ 200 yr BP) and the Loch Lomond Readvance (<u>ca</u>. 11,000 - 10,000 yr BP). Eight raised shorelines have been identified as having formed during the relative fall of sea-level from <u>ca</u>. 38 - 40 m O.D. that accompanied the disappearance of the ice-sheet. A particularly well developed shoreline, CLG2, was formed during the Otter Ferry Stage. A further unique rock-cut shoreline,

the Main Rock Platform, was at least in part formed during the cold conditions immediately prior to and during the Loch Lomond Stadial.

During the Loch Lomond Stadial glaciers extended down Loch Long to near Ardentinny and down Loch Fyne to beyond Furnace. The mountains in the NE of the study area stood proud of the ice mass as nunataks whilst a number of small valley glaciers occurred in the S of the Cowal Peninsula. Analysis of shoreline gradients and the sea-level change curve suggests that this build-up of ice was sufficient to depress the earth's crust anew.

During the early part of the Flandrian Period a major transgression has been recorded by radiocarbon-dated buried peats. This transgression culminated some time after 7,200 yr BP in the formation of a major raised shoreline (CF1) and during the subsequent regression a further five shorelines were formed.

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ACKNOWLEDGEMENTS

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CHAPTER 1

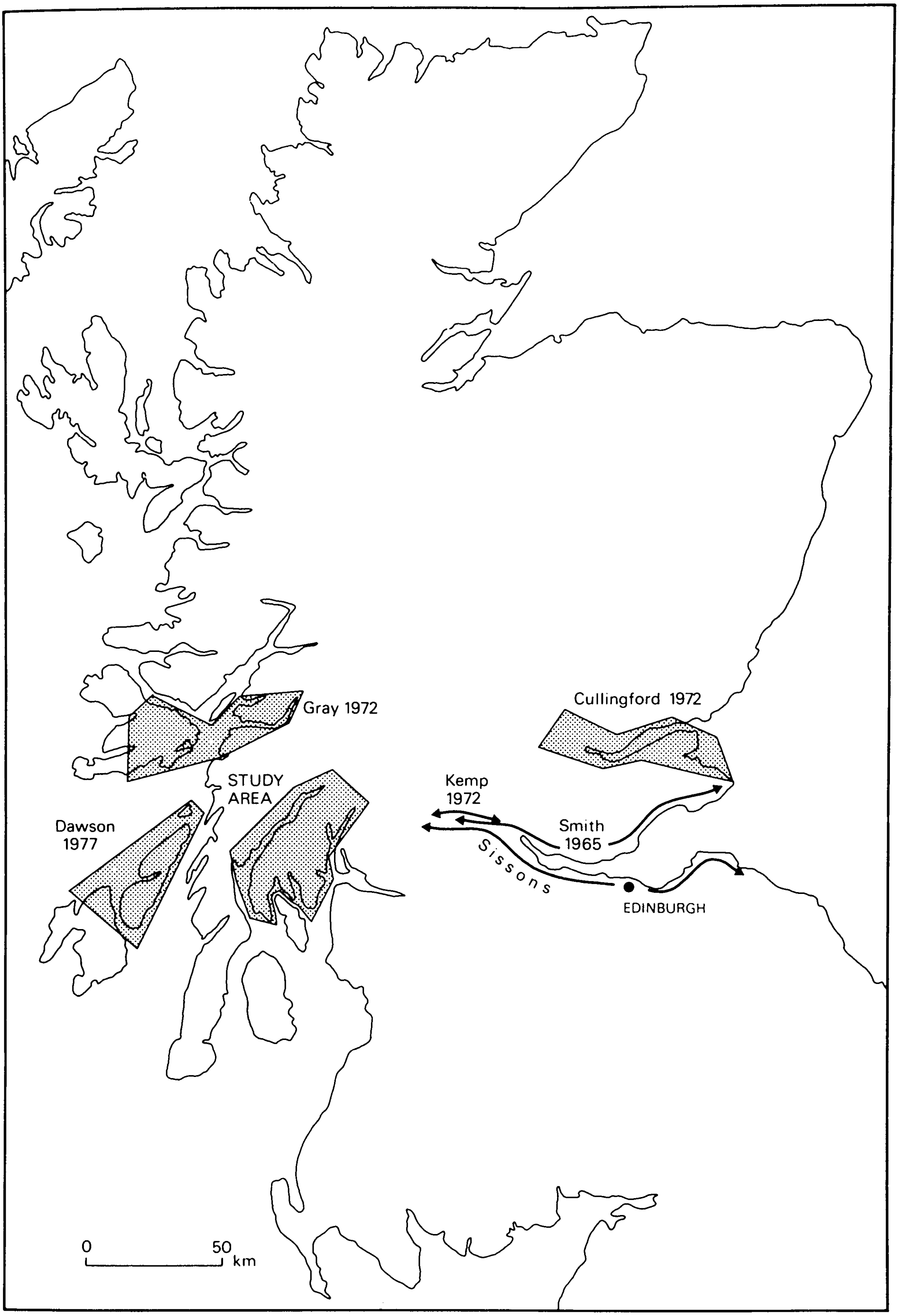
INTRODUCTION

This thesis is one of a series of theses at Edinburgh University Geography Department concerned with the raised shorelines around the coast of Scotland. Prior to commencing on this work similar studies had been carried out in the Forth Valley (Smith,1965; Kemp, 1971), in the Tay-Earn area (Cullingford, 1972) and in the Firth of Lorne and eastern Mull (Gray, 1972a). More recently a further study has been conducted in Jura and Islay (Dawson, 1977). These theses together with the work of J.B. Sissons (see references in text), the overall supervisor of this research programme, represent a major contribution to the study of sea-level changes around the Scottish coast. Yet despite this much remains unknown. The present study therefore goes some way to filling the gap of the unknown in a small part of the SW Highlands.

At the outset it was desired to work in an area that was relatively close to the centre of isostatic uplift, for here the height differences between shorelines would be at their maximum and identification of the shorelines would be correspondingly easier. The penetration of the sea-lochs at the head of the Firth of Clyde deep into the SW Highlands offered an excellent opportunity to fulfill this prime requirement. (Fig. 1). The study area was also readily accessible from Edinburgh and this facilitated the logistics of the fieldwork. Furthermore, study in this part of the SW Highlands offered an opportunity to bridge a large part of the gap between Gray's study area to the NW and the work in the Forth Valley to the E.

The methods used in this study have much in common with those studies mentioned above, including a belief in the necessity of mapping and

Figure 1: Location of study area and locations of other similar studies in Scotland.


Gray 1972
Cullingford 1972
Kemp
1972
STUDY
AREA
Dawson
1977
Smith
1965
Sissons
EDINBURGH
0
50
km

surveying all raised marine landforms within the given area. There has not been, however, an uncritical adoption of a methodology and a number of adaptions and refinements of the overall methodology of the above studies have been made. These are detailed in the appropriate chapter but it can be stated here that the modifications are largely the result of two considerations. Firstly, there is a major difference in the conditions of marine sedimentation on the non-estuarine coasts of the W of Scotland and the large estuaries of the E coast where many of the methods were evolved. This difference results in the W coast being characterised by numerous discrete landforms, rarely traceable along the coast for more than a few hundred metres, and often for much less than this, whereas the raised estuarine deposits of the Forth and Tay valleys are frequently traceable for several kilometres and more. In these circumstances the problems of shoreline identification or correlation of different raised marine landforms are correspondingly greater. The second consideration was a consequence of this, that is, the need to derive a quantitative assessment of the accuracy of the methodology. If it could be shown, for example, that the variation in the height measurement of individual landforms was greater than the likely altitudinal variation between shorelines then there was little point in the study. Fortunately, this proved not to be.

The thesis is structured as follows. Firstly, background information, such as geology, relief, previous literature, nature of the fieldwork and the effect of the last ice-sheet advance over the study area are dealt with. No attempt is made to review the literature on Scottish raised shorelines as this has been adequately covered previously by Smith (1965), Sissons (1967a) and Gray (1972a) and to do so again would be mere repetition.

Next follows a chapter on the characteristics of the present shore. This is an essential part of the study of fossil shorelines as it is the best analogy for the former marine landforms and indicates, for example, the factors that govern the altitude of the breaks of slope at the rear of intertidal deltas or the crests of shingle ridges. Characteristics of the present shore are also often used as datum planes in studies of raised shorelines and these characteristics were surveyed relative to O.D. in this study to allow the accuracy of their use in other studies to be assessed. This is followed by Chapter 4 in which the method of height measurement, the errors inherent in this method and the means of correlating the various raised marine landforms are discussed in some detail.

Chapters 5, 6 and 7 contain the body of the fieldwork in which details of particular localities are recorded. Subsequent to this are the analysis chapters, Chapter 8 on shorelines of marine erosion, Chapter 9 on Lateglacial-age shorelines that were formed in relation to glacier ice and Chapter 10 on the shorelines of the Flandrian period. In the penultimate chapter the stages of deglaciation identified during the present study are discussed and the concluding chapter summarizes briefly the sequence of sea-level changes as established by this study.

A note is necessary on terminology. 'Lateglacial' is used in the sense of Gray and Lowe (1977), that is, the period from between 14,000 and 13,000 yr BP till 10,000 yr BP. 'Loch Lomond Stadial' is part of the Lateglacial and covers the period of adverse climate between ca. 11,000 and 10,000 yr BP. The 'Loch Lomond Readvance' is the glacial stage approximately contemporaneous with the Loch Lomond Stadial. The proposal by Sissons (1977) that this glacial stage should be termed the

Loch Lomond Advance, on the grounds that the glaciers developed anew at this time, is not adopted, for this statement is unproven for the SW Highlands, the type area for the stage where glacier development was at its most extensive. A change of name in accepted terminology is therefore unwarranted. The period from 10,000 yr BP till the present is referred to as the Flandrian (Mitchell et al., 1973) and not, as once, the Postglacial. This allows the term post-glacial to have a literal meaning.

CHAPTER 2

BACKGROUND

1. Geology and Scenery

The Cowal Peninsula and neighbouring areas are almost entirely underlain by rocks of the Dalradian succession except in the extreme SE corner around Toward Point where, south of the Highland Boundary Fault, sandstones and conglomerates of Old Red Sandstone age occur (Fig. 2). Originally mapped by the Geological Survey last century (Gunn, et al., 1897; Hill, et al., 1905), the most recent detailed description of the Dalradian rocks in Cowal is given by Roberts (1966). The latest regional syntheses are those of Johnson (1963) and Johnstone (1966). In the very north-east of the study area between the heads of lochs Fyne and Long occur two isolated igneous intrusions, the larger of which is the Glen Fyne granite. These have been described in most detail by Anderson (1935). The Tertiary volcanic activity of western Scotland is represented in the area by a series of NW-SE-trending basalt dykes (Figure 3) the relation of which to the Tertiary igneous centres is described in Richey (1961).

Structurally, the low-grade metamorphic Dalradian rocks form a large overfold, the Cowal Antiform or Anticline (terms used interchangeably in the literature), the axis of the fold running SW to NE at a slight angle to the axis of the peninsula.

The Loch Tay Limestone, the marker horizon dividing the Middle (to the NW) from the Lower Dalradian runs the length of the peninsula, crossing Loch Fyne near its head and again in the south. To the SE of the Loch Tay Limestone (and occasionally separated from it by strata of the Glen Sluan Schists) occurs a distinct band of green schists and grits

Figure 2: Geology of the study area. 1. Old Red Sandstone. 2. Glen Fyne Granite. 3. Diorite. 4. Loch Fyne Quartz-Porphyry. 5. Loch Tay Limestone. 6. Erins Quartzite. 7. Green Beds. 8. Ardrishaig Phyllites. 9. Bullrock Greywackes. 10. Bheinn Bheula and Glen Sluan Schists. 11. Garnetiferous Mica Schist. 12. Inellan Group. 13. Dunoon Phyllites.

(after Gunn et al. 1897)

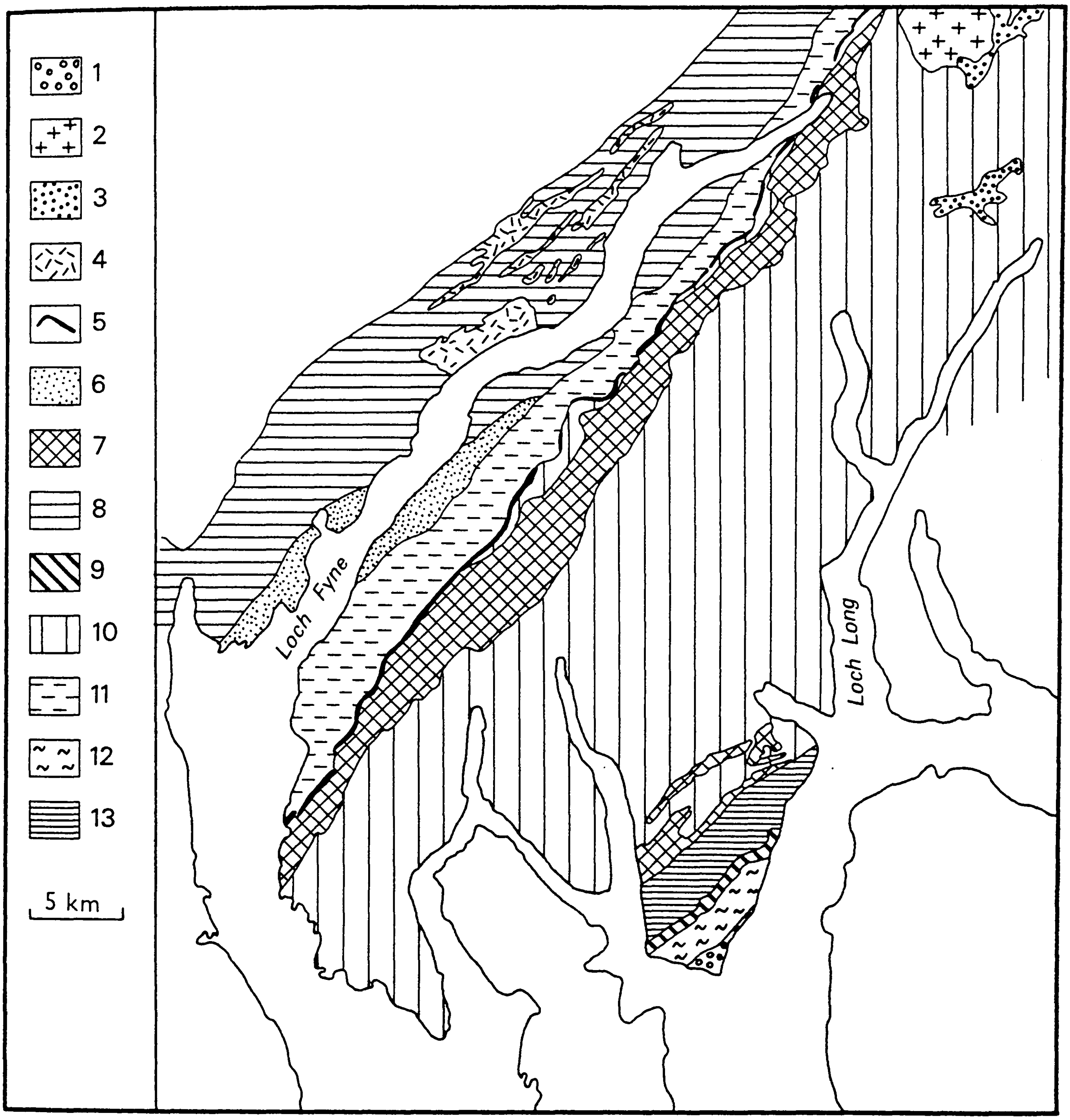
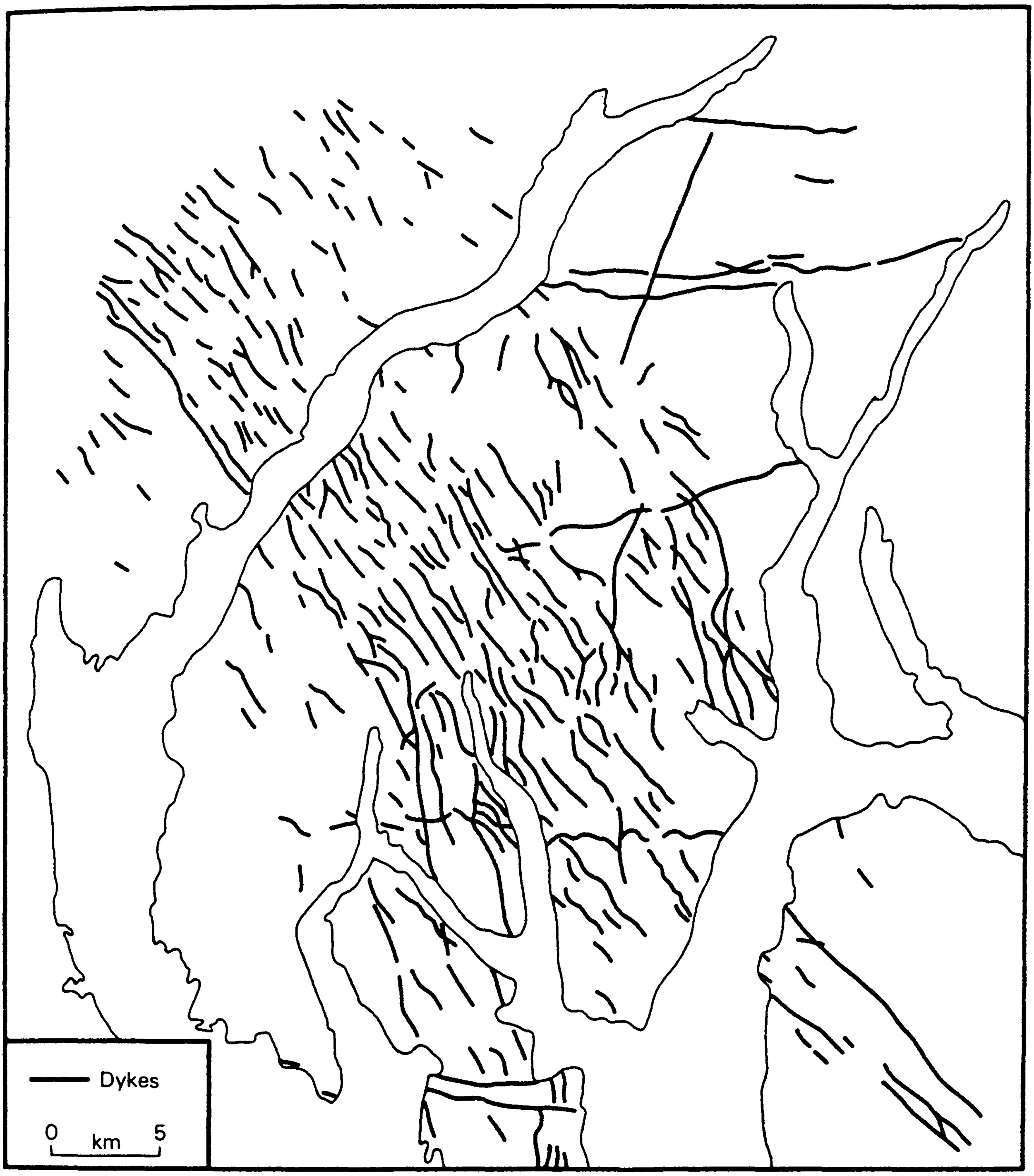

1
2
3
4
5
6
7
8
9
10
11
12
13
5 km
Loch Fyne
Loch Long

Figure 3: Dyke systems of the study area.
(after Gunn *et al.* 1897)


Dykes
0 km 5

(often interbedded with metasediments similar to the Beinn Bheula Schists) known as the Green Beds. The Green Beds that crop out in the SE of the peninsula between lochs Striven and the Holy Loch are on the opposite side of the Cowal Antiform.

The Beinn Bheula Schists occupy the bulk of the Cowal Peninsula and straddle the Cowal Antiform. They are lithologically similar to the Glen Sluan Schists, being dominantly greywackes with thinner beds of schistose grits and grey-green slates and siltstones. The degree of metamorphism is least in the SE, increasing towards the NW.

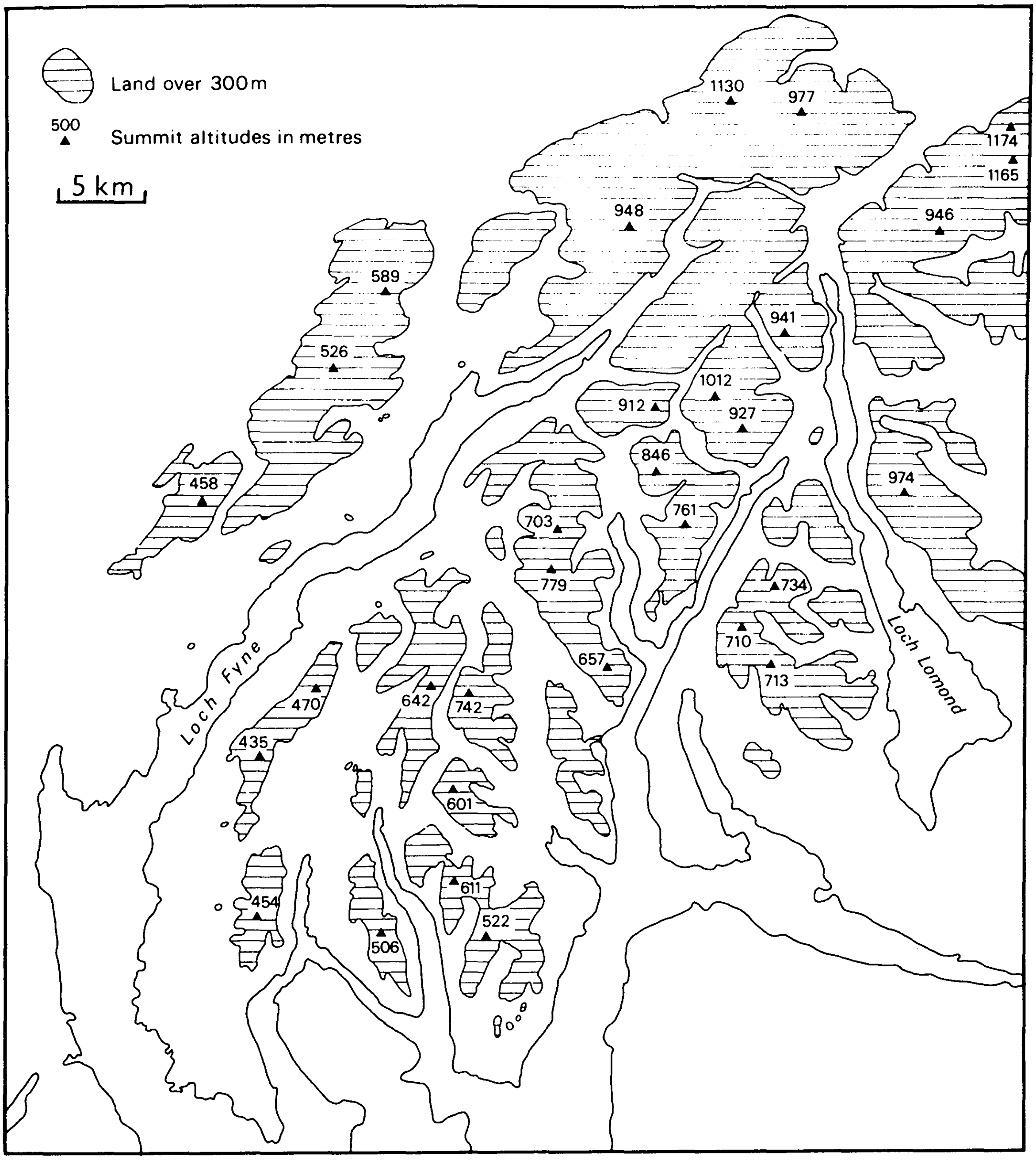
The Beinn Bheula Schists are succeeded to the SE by Green Beds and then, in turn, by the Dunoon Phyllites, the Bullrock Greywackes and the Inellan Group of phyllites and pebbly schistose grits. The Dunoon Phyllites are rather varied but are dominantly black somewhat graphitic slates interbedded with thin limestones and fine to coarse greywackes. They are held to be the lithological equivalents of the Inellan Group, although stratigraphically the Bullrock Greywacke crops out between them.

To the NW of the Loch Tay Limestone occurs the distinctive band of rocks known as the Garnetiferous Mica Schists (passing northwards where they cross Loch Fyne into the St. Catherines Graphitic Schists). The rocks are quartzitic in affiliation, although beds of grits occur as well as irregular lenses of sandy limestones. To the N, the St. Catherines Graphitic Schists are made up of the black graphitic schists and dark thinly bedded limestones.

SW of mid-Loch Fyne the Erins Quartzite is recognized to succeed the Garnetiferous Mica Schists on their western side. The Erins Quartzite is almost entirely quartzitic except on its SE contact where it becomes more gritty and pebbly. It has a transitional boundary to the W with the Ardrishaig Phyllites, a generally fine grained sequence of rocks interspersed with resistant bands of epidiorite and hornblende schists. It is into this series to the W of Loch Fyne that sheets of Loch Fyne quartz porphyry have been intruded.

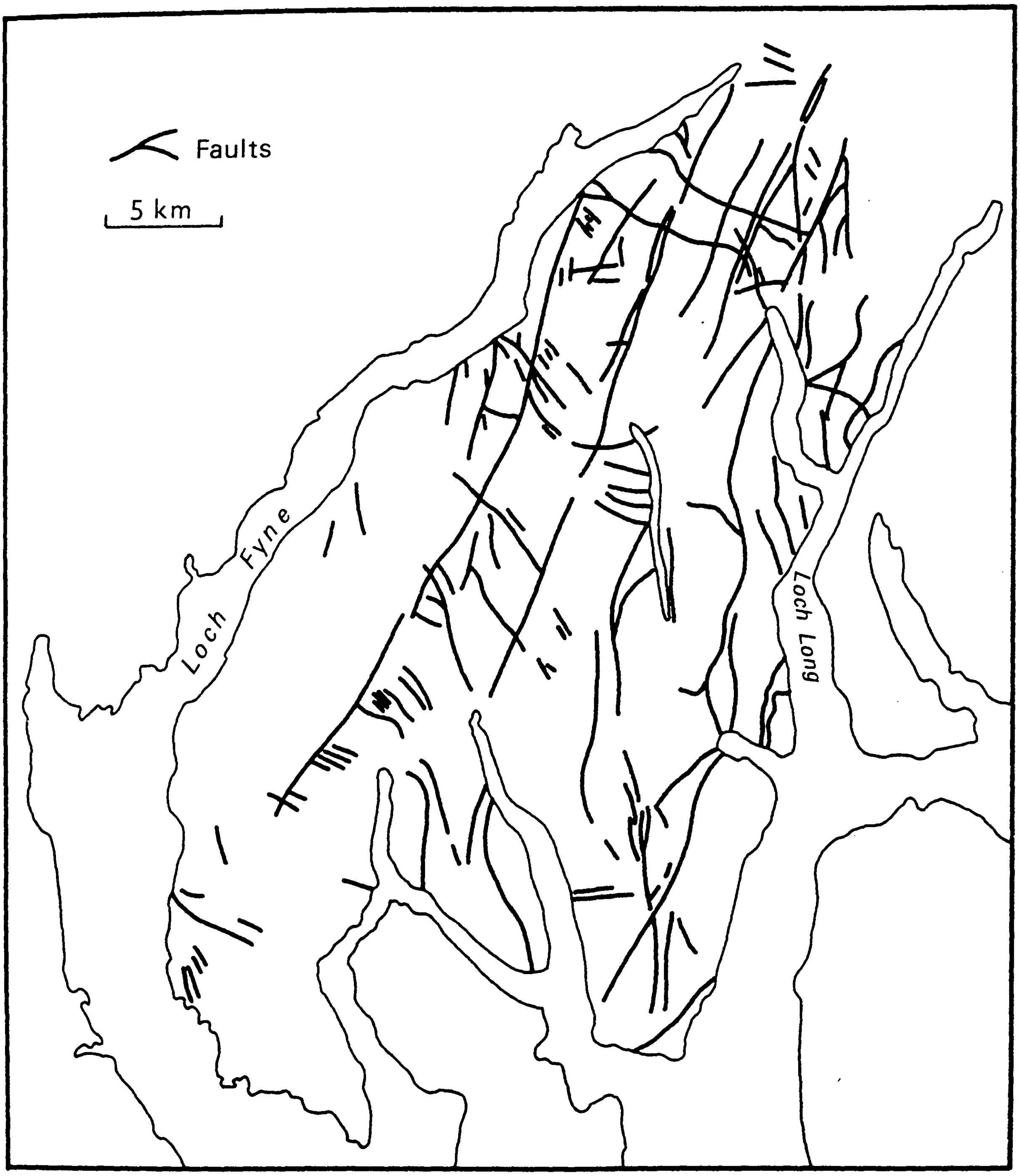
Scenically the Cowal area is typical of much of the west coast of Scotland, long narrow sea lochs penetrating inland among majestic, often rugged peaks. There is little direct relationship between the major geological formations and the major features of the landscape. The mountain masses with the highest peaks are coincident, for the major part, with the Beinn Bheula Schists, summit altitudes declining rather regularly in altitude from over 1000m in the north-east of the peninsula to below 500m in the south-west (Figure 4). The major valleys of the area can be divided into two groups, those trending NE to SW (e.g. Loch Fyne, Loch Long) and those trending NNW to SSE (e.g. Loch Goil, Strath Eachaig). This is part of the general Highland pattern of valleys identified by Geikie in 1901. Neither of these sets of valleys is directly in accord with the underlying geology, even the NE to SW valleys being at a slight angle to the strike of the rocks. Similarly, whilst there are two dominant trends in the fault pattern of the peninsula (Figure 5), major faults trending roughly NE to SW and a secondary set of faults roughly at right angles to this, and whilst certain valleys do, at least for part of their length, follow faults, the major valleys do not seem to be entirely so controlled. The Tyndrum Fault, for example, one of the major faults of the Scottish Highlands (Johnson, 1963), controls part of Glen Fyne and upper Loch Fyne, but the fault and the valley soon part company, the fault coinciding farther S with,

Figure 4: Summit altitudes in the SW Highlands.



Land over 300m
500
Summit altitudes in metres
5 km
1130
977
1174
1165
948
946
589
526
941
1012
912
927
846
974
458
761
703
779
734
710
713
657
Loch Fyne
Loch Lomond
470
642
742
435
601
611
454
522
506

Figure 5: Fault pattern in study area.
(after Gunn *et al*. 1897)



Faults
5 km
Loch Fyne
Loch Long

successively, Corrie No and Garrachra Glen.

In local detail, however, there is considerable dependence of the landscape on the underlying rocks. Small valleys and hillside streams have their courses largely controlled by fault crush zones or dykes. Thus Gunn et al. (1897) point out that in Glendaruel and Glen Tarsan almost all the burns follow dykes or crush breccias. The cross-profile of many of the valleys is dependent upon their relationship to the dip of the strata, those valley sides sloping with the dip being less rugged and less steep than those that slope against the dip. Glendaruel is a fine example of this, as too are Glen Tarsan and Garrachra Glen.

The nature of the rocks also has a strong influence on the texture of the scenery, the finely bedded Dunoon Phyllites, for example, giving rise to rounded rather smooth hills, whilst the much coarser Beinn Bheula Schists produce a very rugged, knobbly terrain which between lochs Goil and Long glorifies in the name of 'Argyll's Bowling Green'. W of Loch Fyne the interbedded Ardrishaig Phyllites, epidiorites and horneblende schists and the intrusive quartz porphyries, all with a strong NE to SW strike, correspond with a very characteristic ridge and valley topography. In the NE of the study area the two igneous intrusions occupy relatively low positions in the terrain, everywhere being overlooked by peaks developed on the schists.

Geikie (1901) in discussing the origin of Highland valleys termed the NE-SW-trending valleys 'transverse', considering them to result from areas of crustal weakness such as large fault zones. The NW to SE valleys were termed 'subsequent' as they were considered to be dependent upon the prior existance of the transverse valleys. As has been pointed out above, whilst there is some relationship to geological features such as faults the valleys also show considerable independence of geology. Alternative views of the origin of the valley system have been advanced, mainly by geomorphologists trying to explain the apparent lack of accord between the valley system and geological structure.

The earliest view was that of Cadell who, in 1886, proposed that the valley system of the SW Highlands originated as a set of NW-SE-trending valleys that were the headwaters of a 'proto-Forth' river. This general view was subsequently echoed and only slightly modified by authors such as Gregory (1915) and Linton and Moisley (1960). All these authors seem agreed that Loch Goil linked with the Gare Loch, Coilessan Glen with Glen Douglas, and Glen Sloy with Loch Arklet; they suggested that at least Loch Long and Loch Lomond (in part) are later (acute) disruptions to this general drainage trend. The nature and timing of this disruption were not so widely agreed. Cadell considered all the valleys to have originated solely under the influence of a fluvial regime and that subsequent glaciation did no more than deepen and enlarge these valleys. Gregory, whilst agreeing on the general role of rivers, considered it necessary to postulate considerable tectonic movements along both Loch Long and Loch Lomond. Linton and Moisley strongly emphasised the work of glaciers in the formation of Loch Lomond but considered Loch Long to be due to preglacial fluvial headward erosion.

These various geomorphological speculations on the origin of the valleys in the SW Highlands are unfortunately not tied in any unambiguous way to the evidence on the ground. None of the authors appears to have noted the regular SW to NE rise in summit altitudes across this area, a rise it might be supposed that reflects the long-term positive tectonic tendency of the Scottish Highlands as compared to the areas to the S and W (cf. Bott and Watts, 1971). If the regular variation in summit altitude does indeed reflect the long-term tectonic movement in this area then this implies that valleys such as Loch Fyne and Loch Long are likely to be the original valleys of the region as they run in accord with the decline in summit altitudes, rather than at right angles to it as do any reconstructed south-easterly oriented drainage networks. Subsequent glacial modification has been largely responsible for the overdeepening of the SW trending valleys and the enlargement of the south-easterly directed valleys.

2. Previous Literature on the Study Area

Very little work has been published on the geomorphology or Quaternary stratigraphy of the Cowal Peninsula, despite its proximity to and ease of access from large centres of population. The standard reference on the glacial deposits remains the Geological Survey Memoir of Gunn, Clough and Hill dating from 1897 and from this date till this study was started no paper on glacial geology or geomorphology had been published that had its point of focus exclusively centred on this area, only sporadic reference having been made, more or less in passing, to some of the more general aspects of glaciation or sea-level changes.

Yet in the 50 years subsequent to the introduction of the Glacial Theory in 1840 the area was held to be of considerable interest by glacialists and, indeed, it was in part from the studies of fossil marine shells in the Kyles of Bute that James Smith of Jordanhill (1838) postulated former changes of sea-level and climatic cooling, the conceptions of which were far in advance of his time. The 'Glacial Beds', 'Post-Tertiary Fossiliferous Beds', 'Clay Beds', 'Clay Shell Beds', or 'Clyde Beds' as they have latterly come to be known, provided the material for many of the subsequent publications, Crosskey (1865) writing on Lochgilphead and the Kyles of Bute, Crosskey and Robertson (1873) on the Kyles of Bute, Macadam (1881, 1882) on Strachur and general syntheses being provided by Brady *et al*. (1874) and Gunn *et al*. (1897). From this latter date, however, until 1977, no new faunal lists were published for any part of this area. In recent years construction works have revealed sections in the shell beds and Dr. J.D. Peacock and co-workers of the Institute of Geological Sciences have provided details of the marine fauna at Lochgilphead (1977) and at Ardyne (1978).

The major interest in the Clyde Beds was the presence of marine shells that are now confined to Arctic waters. In general only 'one' shell bed is described at any locality, though Crosskey (1865) was of the opinion that two beds, one Arctic, one temperate were present in the Kyles of Bute. The unfortunate practice of lumping all the shells from one exposure together as a bulk sample effectively masked any subtle or progressive changes in marine environment that had taken place. The recent studies of Peacock and others have rectified this deficiency and together with radiocarbon dating have allowed the tracing of changes in salinity and water temperature and depth over the period 12,600 - 10,250 years B.P. In addition to identification of shells the early studies established a general sedimentary sequence (which, due to subsequent erosion, occurs in reverse order from high to low water mark) in which a basal till is overlain by laminated clays, this being succeeded upwards by the shell beds whose composition was remarked upon for its variability, sometimes being stoney, sometimes clayey, more often sandy. The more recent work, discussed in detail later in this study, confirms the general sequence but also provides details of unconformities in the sediments. It eventually became established in the literature (most firmly by Gunn _et al._, 1897) that the shell beds were intimately associated with large erratic boulders, the direction of transport of which was shown in some areas to be in the opposite sense to erratics in the underlying till. This factor together with the observations on barnacles on the undersides of some of the boulders led to the conclusion that the boulders were transported by icebergs and dropped into the shelly clays, an idea first suggested by Jamieson (1865). A further point that was established, though it was not regarded as being of great significance until noted by Sissons (1967a) was the altitudinal distribution of the shell beds, all the described beds, with the exception of one at Lochgilphead, occurring at low altitudes, despite the independent evidence for marine activity at considerably higher levels.

Glacial deposits had been remarked on in a number of early papers, McLaren (1850) identifying two end moraines in Strath Eachaig, and Bell (1874) a medial moraine on the point between Loch Long and the Holy Loch, while Anderson (1896) placed a number of ice limits within the eastern half of the area, though with hindsight it seems clear that he was unable to distinguish raised marine from glacial deposits. Once again Gunn et al. (1897) in a remarkably perceptive account provide the latest details of glacial deposits in the area. They identify two periods of glaciation, the first being that of a general ice cover in which erratics were transported in a southerly direction and on to the tops of the highest hills. The second phase of glaciation was a local one in which valley moraines were deposited, and striations were etched, in some instances, at a considerable angle to those of the main glaciation. This latter ice movement was held to be contemporaneous with the formation of the '100 foot beach' as this sea-level had not been identified towards the heads of the sea lochs. The much later study of Charlesworth (1955) relies almost entirely on Gunn et al. (1897) for the positioning and relative chronology of its many retreat stages in this area.

The mention that has been made of the Cowal Peninsula in the literature of this century has centred almost entirely on two themes: that of the origin and development of the valley system and that of the history of sea-level changes. The scenically impressive set of valleys that dissect the Cowal Peninsula was looked upon by Cadell (1886) as part of a once extended Forth drainage basin. Geikie (1901), however, considered that the valleys fitted into a general Highlands pattern that was explicable in terms of the known geology, though Gregory (1915) reverted to the south-eastward directed original drainage network which had subsequently been disrupted by tectonic activity. More recently Linton (1951) has supported the idea of the original south-eastward directed drainage which he believed to have been superimposed from a Cretaceous chalk cover. Linton (1951; 1963) and Linton and Moisley (1960) also emphasized the role of glacial erosion in the disruption of this network, though it is thought that a greater amount of pertinent data could be brought to bear on the subject than Linton could encompass in his "glance at the quarter- or half-inch map" (Linton, 1951, p11).

The seaward extension of these valleys was first surveyed by Mill (1891) who noted not only their great depth but also their basinal nature, matters that have been clarified by recent geophysical work in the southern half of the area (Eden et al., 1971; Deegan et al., 1973) that shows not only great rockhead depths but also considerable thicknesses of sediment coinciding with the deeps such that in lower Loch Fyne, for instance, where rockhead reaches -280 m, there is over 130 m of sediment. Such great thicknesses contrast with the surrounding rock walls and shallower water which are comparatively free of deposits.

All recent work on raised shorelines in this area has been part of much more extensive studies. Donner (1959, 1963) in part of a Scotland-wide coverage of raised shore features deduced four shore zones, the upper one of which was tilted downwards away from Callander, the others being horizontal. Donner has subsequently (1970) retracted much of this work. McCann (1963) covered the whole of the west coast of Scotland, looking for and finding the 25-foot, 50-foot and 100-foot raised beaches.

His subsequent work farther north (e.g. McCann, 1966) makes it seem probable that he has rejected his earlier findings that were specifically related to Cowal. Synge and Stephens (1966), working both in Ireland and south-west Scotland, used shoreline height information from Loch Fyne in an overall synthesis of these areas. This work was heavily criticised (Sissons, 1967b) and part of their study in nearby areas has been shown to be in error (Gray, 1972a; Gray and Sutherland, 1977). Synge (1977) accepted that some of his earlier conclusions were in error. It can therefore be surmised that little has been unequivocally established in the way of the history of sea-level changes for the Loch Long/Loch Fyne area and that a study of such would be appropriate.

3. Fieldwork

The fieldwork that constitutes the basis of this thesis was conducted during the summers of 1972, 1973 and 1974. Prior to commencing work in the field the whole of the coastal zone of the Cowal Peninsula and neighbouring coasts (more than 300km in length) were examined stereoscopically on ca. 1:10,000 aerial photographs in the Aerial Photograph Library of the Scottish Development Department. Aerial photographic mapping at this scale does not allow the recognition of many of the small but significant breaks of slope between raised shoreline fragments but it does reveal many of the major features and acts as a familiarizing procedure for the whole area.

Since the principal objective of the study was to identify and measure the various landforms associated with sea-level changes following the decay of the last ice sheet, fieldwork was generally restricted to areas below 75-100m O.D., as a number of previous studies of the raised marine deposits of Scotland (e.g. Sissons, Smith and Cullingford, 1966; Gray, 1972a) had indicated that such deposits were unlikely to be found above approximately 40m O.D.. All of the area below 75-100m altitude was mapped in the field at 1:10,560 scale, mapping being continued up valleys when the continuity of terraces or fluvioglacial landforms indicated it to be necessary.

In order to gain some understanding of the distribution of the glaciers that existed in this area during the Loch Lomond Stadial, the whole of the N and parts of the S of the Cowal Peninsula as well as much of the area to the W of Loch Fyne were examined on ca. 1:27,000 scale aerial photographs purchased from the Ordnance Survey. A limited amount of field checking of this photographic interpretation in the valleys away from the coast was carried out.

During field mapping all terrace-like features, whether depositional or erosional were recorded with a similar symbol, no attempt being made at first to subdivide into shoreline or river terrace features. In addition fluvioglacial landforms such as kames, kettle holes and meltwater channels were mapped as well as constructional glacial landforms such as end moraines. The occurrence of rock cliffs was noted along with other landforms resulting from marine erosion such as caves, geos, undercutting and stacks. Features due to active ice such as large-scale ice moulding, roches moutonnees and striations were recorded. All available sections were inspected and notes made of the stratigraphy revealed. Certain erratics such as Glen Fyne granite were noted although no attempt was made to quantify their abundance.

Visits were made to various local councils and large organisations working in the area in order to collect all the commercial borehole records available for the area. A small amount of hand boring using a Hiller borer and a 1-inch hand auger was also carried out.

The opportunity arose to have some radiocarbon assays performed and samples were taken for this purpose, 2 being submitted to the Trondheim Radiocarbon Laboratory and 8 to the Uppsala Laboratory.

The field mapping was followed by instrumental surveying of all the terrace features. A Hilger and Watts Autoset level was used throughout, reading off a 1-cm graduated staff. The level was checked and adjusted as necessary every two weeks. Heights were related to Ordnance Datum, all traverses commencing and terminating on bench marks, any error present being distributed evenly throughout the traverse. The maximum closing error accepted was 0.18m. Traverses with closing errors greater than 0.5m were resurveyed as an error of such a magnitude was taken to imply a gross error in the observation of one point. For the greater part of the survey the bench marks used were related to O.D. Newlyn. These bench marks are, however, considerably more restricted in their distribution than are the bench marks of the 1st and 2nd Geodetic Surveys related to Liverpool Datum and these latter bench marks which are recorded on the 1st and 2nd Editions of the 1:10,560 maps were used in some areas. Where possible the Newlyn heights of bench marks common to more than one survey were compared with the Liverpool 1st and 2nd Survey heights and Table 1 was drawn up which allowed the conversion of Liverpool O.D. heights to Newlyn O.D. heights.

This was not done by taking an arithmetic mean figure for the whole area as it is obvious that the differences between the various datum planes are primarily of local significance. The conversion from one datum to the other was therefore effected by considering only the two or three nearest common bench marks. Errors arising from this procedure are thought to be considerably less than 0.1m. One exception exists to the above procedures. The radiocarbon sample U-4066 near Lochgilphead was heighted using the water level of the Crinan Canal to transfer the height from the nearest bench mark to the sample site.

Various staff positions were adopted during heighting depending on the nature of the feature being surveyed. On terrace features the staff was positioned near the break of slope at the back edge of the feature. Exceptions to this were very wide distinct river terraces that were surveyed along the middle of the terrace. Shingle ridges were heighted on the crest of the ridge, and rock platforms were heighted as close to the cliff/platform junction as was possible. Various former river channels were surveyed along the centreline of the channel and certain moraines and fluvioglacial features were heighted on their summits.

On each feature surveyed heights were recorded every 50 paces (approx. 50m) and the positions of the surveyed points were marked on the 1:10,560 maps. Field boundaries, alluvial fans, streams cutting across features and slumped debris were all avoided in surveying. In certain locations peat overlay the point to be surveyed and Hiller rods were inserted into the peat to establish the nature of the surface underneath. Several points within an area of about $10m^2$ at any one locality were tested in this fashion and a representative point selected for heighting, the staff being placed on top of the rods inserted in the peat and the reduced height being adjusted for the length of the rods.

In order to illustrate the nature of the succession of beach terraces and ridges that were present, a number of cross-sections at right angles to the trend of the coast were surveyed. On these heights were established every 2 paces (approx. 2m).

Each fragment surveyed was classified as 'good', 'moderate' or 'poor' depending on its degree of morphological preservation. It was not known how this classification related to the representation of former sea-level though it was intuitively felt that a large clear feature with distinct breaks of slope on which six heights may have been recorded was a more reliable piece of evidence than a small isolated fragment on which perhaps only two points could be surveyed. The classification was clearly relative, for there were many features that did not aspire to the 'poor' category that were not considered worth heighting.

Since a major part of the study would involve the correlation of discrete fragments along considerable lengths of coastline it was necessary to gain some estimate of the precision inherent in the surveying technique used. Accordingly, 32 shoreline fragments were re-surveyed one year after their initial survey. These results are reported in Chapter 4.

Attention was also devoted to the present shoreline. Firstly, heights were obtained on those characteristics that had been used by others as datum planes: these were the upper limit of growth of barnacles (Donner, 1959), the upper limit of sea drift (Synge and Stephens, 1966), and the lower limit of land-based vegetation. These features are not uniformly present around the coast but some or all of their altitudes were established at 33 separate localities by heighting at 50 pace intervals along the shore.

A second attribute of the present shore to be surveyed in detail was the altitudinal limit of present shore-forming processes. The upper limit of modern shingle is represented by banks and ridges and these were surveyed in order to relate fossil shingle ridges to particular sea-levels. Similarly, since the methodology adopted resulted in the identification and surveying of flat-topped former marine features, flat-topped present-day marine features were examined as modern analogues. Along the present shoreline as along the abandoned shorelines irregularly sloping banks of shingle could be observed but these were never surveyed because of the difficulty involved in choosing unambiguous points on the features that could relate in some way to high, mean or low sea-level. In surveying the flat-topped features of the present shoreline, a cross-profile technique was used, heights being recorded every 2 paces. The most common flat-topped features along the present shore were intertidal deltas at the mouths of small streams. This is true also of the fossil shorelines.

An attempt was also made to ascertain the altitude at which marine erosion was effective along the present shore. This was less easy than surveying constructional features for in many places where small, often quite fresh cliffs were cut in till the base of the cliff was covered by shingle. In a number of places small rock-cut cliffs and platforms were surveyed that showed evidence of current marine erosion but, as is discussed in Chapter 8, it was not always clear whether these were, to some extent at least, fossil.

The principal restriction on fieldwork was the occurrence of built-up areas. This may seem to be only a slight problem in the thinly populated SW Highlands, but often the raised marine deposits are the only ground available for construction. Thus, for instance, from Inellan through Dunoon to Sandbank on the Holy Loch about 15km of coastline is built up, as is almost all of the north side of the Holy Loch and the mouth of Loch Long. Long stretches on the mainland side of both the East and West Kyles of Bute are built upon and elsewhere small towns such as Lochgilphead, Inveraray and Arrochar occupy situations at which raised marine landforms could otherwise have been surveyed.

A second major restriction on fieldwork was the occurrence of extensive forestry plantations in many of the valleys, inevitably at lower levels. Forestry can make aerial photographic interpretation difficult and at certain stages of growth precludes access and greatly hampers surveying. On the other hand forest roads often gave excellent sections and could be used for rapid access to otherwise isolated areas.

These latter comments apply also to Hydro Board roads that occur particularly in the north of the area studied.

A third major restriction on fieldwork was the relatively limited distribution of bench marks. This often necessitated several days being spent surveying and checking temporary bench marks, sometimes for several kilometres along a valley or to an isolated bay.

The programme of fieldwork outlined above differs markedly from any previously carried out in the thesis area. Gunn et al. (1897) mapped the whole area at 1:10,560 scale and the wealth of detail they provide stands as a testament to this, but aerial photographs were not then available, many of the present concepts of glacial geomorphology or of raised shoreline studies did not then exist, and they did not attempt instrumental surveying of the landforms they observed.

In more recent times Donner (1959) attempted to analyse the raised shorelines of this area utilising the heights of 10 shoreline fragments at 7 different locations. He used the upper barnacle limit for a datum plane and related it to O.D. by levelling at three points between the Mull of Kintyre and the head of Loch Fyne. For his interpretation of raised marine landforms he relied upon the 1-inch Geological Survey maps. Four of his sites in the present study area were instrumentally levelled by him, the remaining three being surveyed by aneroid barometer.

McCann (1961) surveyed 13 features at 12 localities in the present study area. He used the high-water mark of ordinary spring tides as a datum and instrumentally levelled the shoreline fragments, though he unfortunately subsequently chose those of his surveyed heights that corresponded to his idea of local relative sea-level. It is therefore not surprising that he discovered the '100-foot', the '50-foot' and the '25-foot' shorelines he believed to exist.

The most recent study of this area was that of Synge and Stephens (1966). Only 6 raised shoreline fragments at 2 separate localities were surveyed, an Abney level being used. The upper limit of sea drift (mainly Enteromorpha spp.) was used as a datum, this being related at an unspecified number of locations to Ordnance Datum by levelling to either bench marks or spot heights (!) on 1:25,000 GSGS maps.

Not one of these last three studies has been the result of a systematic, detailed examination of the coastal zone but they all have been conducted on a hurried basis as part of much larger projects. Reference is never made in these studies to the relationships between the landforms surveyed and other landforms in the immediate vicinity. Such an approach is best judged by its users: all of those involved have rejected part, if not all, of their findings in subsequent years.

The methods adopted in this study do, however, resemble in many respects the studies in E Scotland of Sissons, Smith and Cullingford (1966) and in W Scotland of Gray (1972a). These studies are noted for their attention to detail and their attempt to survey accurately all shoreline features, river terraces and associated landforms in their respective areas. Aspects of this methodology are considered at greater length in Chapter 4.

4. Ice-Sheet Glaciation

It has long been accepted (Geikie, 1894; Sissons, 1976a) that the Scottish Highlands have been subjected to more than one period of ice-sheet glaciation during the Quaternary. The last such ice-sheet occurred during the Devensian though the timing and limits of this ice advance are only roughly known. No evidence has so far been found in the Cowal Peninsula of any ice sheet prior to the Devensian one.

Ice movement during this last ice-sheet glaciation can be inferred from the carry of erratics from clearly identifiable sources, from glacial striations and from the orientation of ice-moulded landforms. Two sources of easily identifiable erratics occur within the thesis area: the Glen Fyne granite in the NE of the Cowal Peninsula and the Loch Fyne quartz-porphyry which crops out NW of Loch Fyne (Figure 2). Both these erratics have been identified over wide areas of the SW Highlands and Central Lowlands by officers of the Geological Survey last century and Sissons (1967a, p73) compiled a map showing the carry of these erratics. A portion of this map is reproduced here as Figure 6. The widespread distribution of these erratics is most reasonably assigned to the last ice-sheet and it is clear that the mountains to the NE of the Cowal Peninsula were a major centre of ice dispersion during the last ice-sheet glaciation. Glen Fyne granite erratics have been recovered in the W from Colonsay and in the E from the Forth Valley E of Falkirk. Loch Fyne quartz-porphyry erratics have been carried mainly to the SW and W, being found in Colonsay, Islay and Jura. That the Loch Fyne quartz-porphyry erratics have not been recovered from E of Loch Fyne is indicative, perhaps, of the strength of the Loch Fyne glacier.

Striations are abundant in the SW Highlands, being preserved best

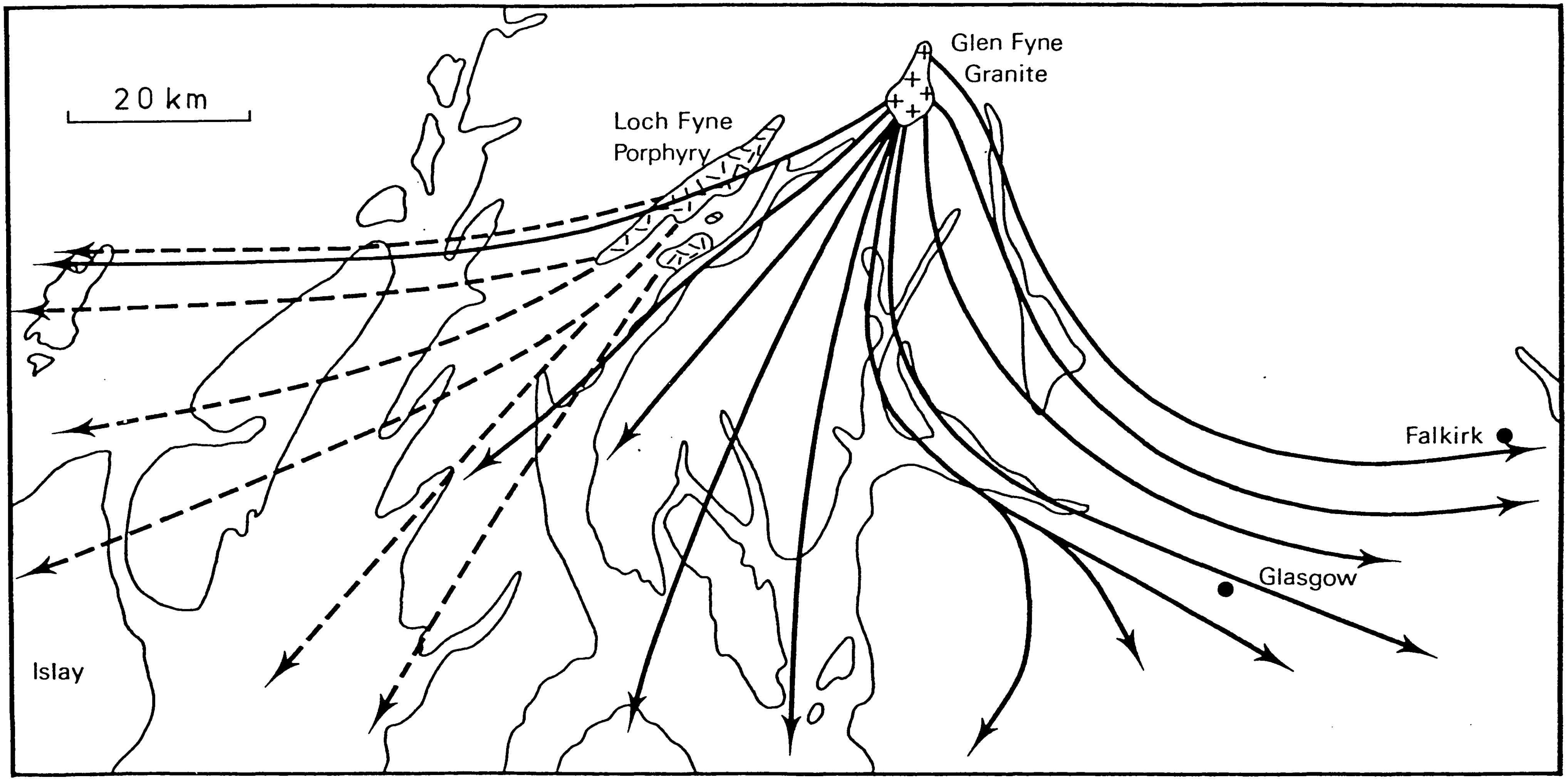
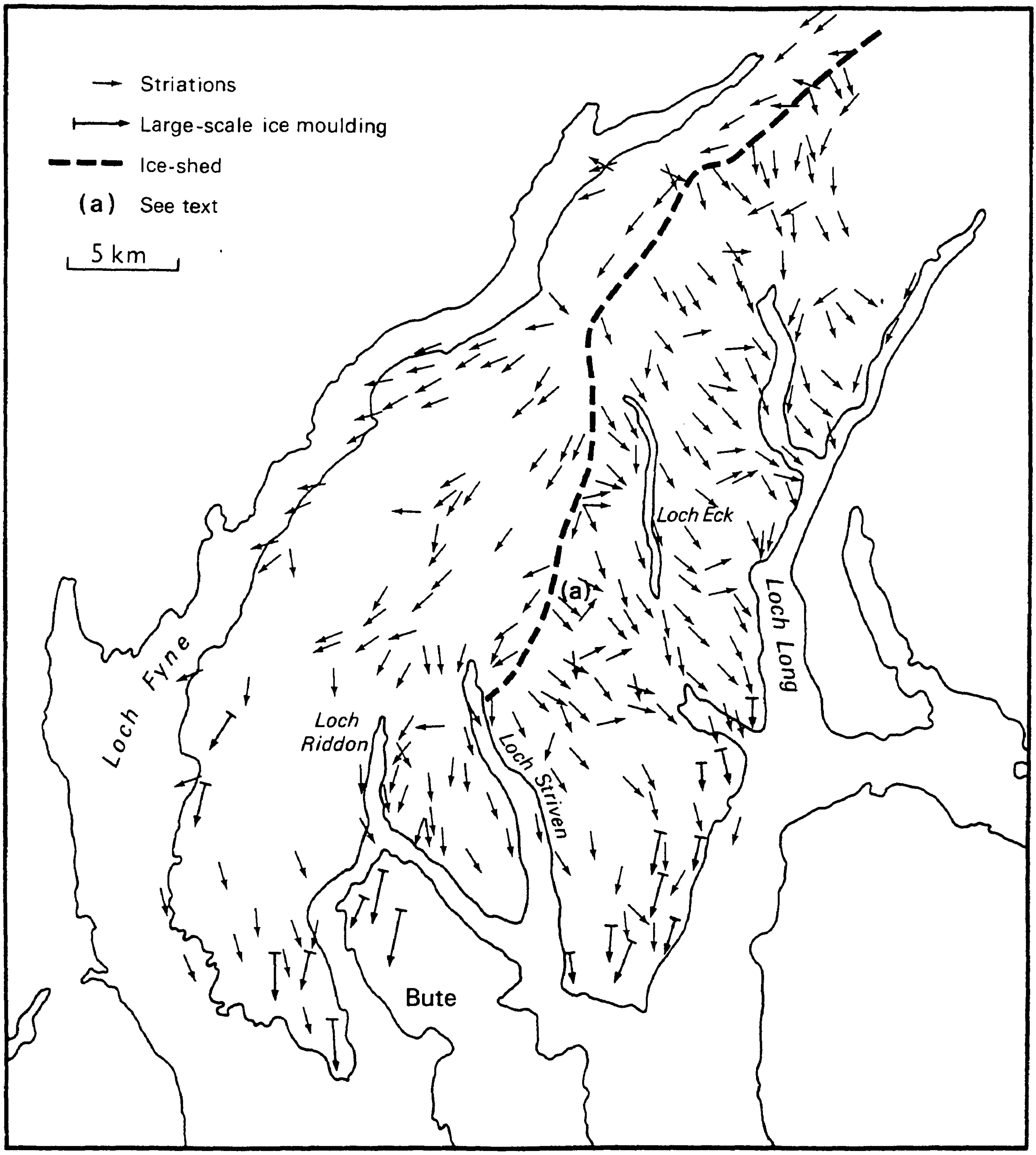

20 km
Glen Fyne
Granite
Loch Fyne
Porphyry
Falkirk
Glasgow
Islay

Figure 6: Carry of Glen Fyne Granite and Loch Fyne Quartz-Porphyry erratics. After Sissons 1967a.

on the more massive rocks such as the Beinn Bheula Schists, on dyke rocks and quartz veins and especially on rocks close to present sea-level. Figure 7 is a map of striations taken from Gunn et al. (1897) which, whilst not being able to show all the instances recorded in the field (this is not even possible on the one-inch geological maps), graphically portrays the movement of the ice and some of its local deviations. The pattern is inevitably complicated, firstly by the considerable variations of local relief that characterise much of the peninsula and secondly, by the period of local glaciation during the Loch Lomond Stadial that followed the decay of the last ice-sheet (see Chapter 11). These two effects account for the various localities, particularly in the north-eastern part of the peninsula, where striations only a short distance apart are often oriented at right-angles to each other. Perhaps the clearest example of striations due to subsequent local glaciation is in Glen Massan (Figure 7, location a) a major valley that was glaciated by ice moving down-valley during maximum ice cover. Subsequently a small corrie glacier developed and descended to the valley bottom depositing two arcuate end moraines inside which are the striations at right angles to the ice-sheet flow.

The pattern of ice movement as revealed by the striations is simplest in the SE of the area where it is augmented by ice-moulded landforms that are partially erosional and partially depositional in origin. These landforms, which have been mapped from aerial photographs, are best developed over the tops of the hills west of Dunoon where, with the striations, they demonstrate a swing in the ice flow as ice moving SSE out of Strath Eachaig became confluent with the major ice stream moving slightly W of S from Loch Long.

Figure 7: Striations and ice-moulded landforms in study area.


Striations
Large-scale ice moulding
Ice-shed
(a) See text
5 km
Loch Fyne
Loch Riddon
Loch Striven
Loch Eck
Loch Long
(a)
Bute

W of a line running northwards from upper Loch Striven past the head of Loch Eck to near the head of Loch Fyne, the striations and ice-moulding point generally towards the SSW or SW. E of this line the striations are generally oriented towards the SE suggesting that this was a local ice shed during the last glaciation. Confirmation of such an ice shed comes from the hilltops in the NE of the peninsula as noted by Gunn et al. (1897) for the striations to the E of the supposed ice shed above ca. 650 m are oriented towards the SE whilst at lower altitudes they tend to follow the lines of the valleys.

Although the valley system had its outlines shaped in pre-glacial times (see above) there is little doubt that considerable erosion was effected by glaciers along the lines of certain valleys, Loch Fyne, for example, the largest Scottish sea loch, being eroded into three distinct basins, in the deepest of which rockhead descends to -280m (Deegan et al., 1973). The glaciers overdeepened those valleys that aided their dominantly southerly flow and it is interesting to note that the through valleys of the peninsula (Strath Eachaig, Loch Goil/Hells Glen, Glen Croe/ Rest and be Thankful) are eroded in accord with the general pattern of striations. These valleys all have their present-day watersheds close to Loch Fyne and they were interpreted by Linton (1963) as having been formed by glacial breaching, the Loch Fyne glacier 'overflowing' and excess ice escaping down these valleys to the SE. This interpretation can be questioned on the basis of the local topography for Loch Fyne opens out south-westwards whilst the glacial 'overflow' channels penetrate into increasingly mountainous terrain. An alternative explanation is offered if the ice shed suggested by the striation pattern was a reality, for the present-day watersheds of Strath Eachaig and Loch Goil/Hells Glen are close to the inferred ice shed. Certain other valleys appear explicable in similar fashion. The Loch Striven valley, for example, commences only a short distance from the head of the sea loch (the approximate position of the ice shed) yet it becomes a major valley as it extends south-eastwards.

It is clear from field evidence that all the mountains of the Cowal Peninsula were overtopped by ice during the period of ice-sheet cover. Striations have been recorded, for example, at altitudes over 650m on the hills in the NE of the peninsula pointing in such directions as to suggest little local topographic control on the ice mass responsible for their formation. Similarly, erratics near the tops of many of the hills imply that the ice surface was sufficiently high above the hills to carry erratics over them. For instance, close to the top of Beinn an Fhidleir at an altitude of approximately 800m erratics of Glen Fyne granite have been discovered although the maximum outcrop altitude of the granite is below 650m.

The problem of maximum ice thickness may be approached theoretically if the outer margin of the ice sheet is known. Unfortunately, the only well-established margin of the Devensian ice sheet is in the Midlands of England and due to the polycentric nature of the last ice-sheet flow lines cannot be connected to this margin from the SW Highlands. Along flow lines from the SW Highlands to the W it is known that the ice passed over all of Jura and Islay, a distance of over 150km from Glen Fyne, but how much farther is not known. To the E the ice that carried Glen Fyne granite erratics into the Forth Valley extended out into the North Sea for an unknown distance though this indicates a minimum flow-line distance from the SW Highlands of over 200km. Using this latter figure it is possible to derive minimal estimates of the maximal thickness of the last ice-sheet in the South-West Highlands.

Paterson (1972) using glaciological theory and comparing this with present day ice-sheet and ice-cap profiles derived the following inequality:

$$7.2 \leqslant \frac{h^2}{x} \leqslant 13.7$$

in which h (metres) is ice thickness at distance x (metres) from the ice margin. For a distance of 200km this indicates ice thicknesses of from 1200m to 1650m. The exact numerical value used in the above calculation is dependent upon, among other things, the accumulation and ablation characteristics of the ice mass. The nearest present-day analogue to the Scottish ice-sheet is perhaps Vatnajokull on Iceland which has a value of 7.7. Using this value the ice thickness at 200km from the ice margin is 1240m. The calculation is relatively insensitive to changes in the position of the ice margin once away from the area of rapid gradient change near the margin and the ice thickness figure of 1240m is only increased to 1390m if the ice margin is 250km distant.

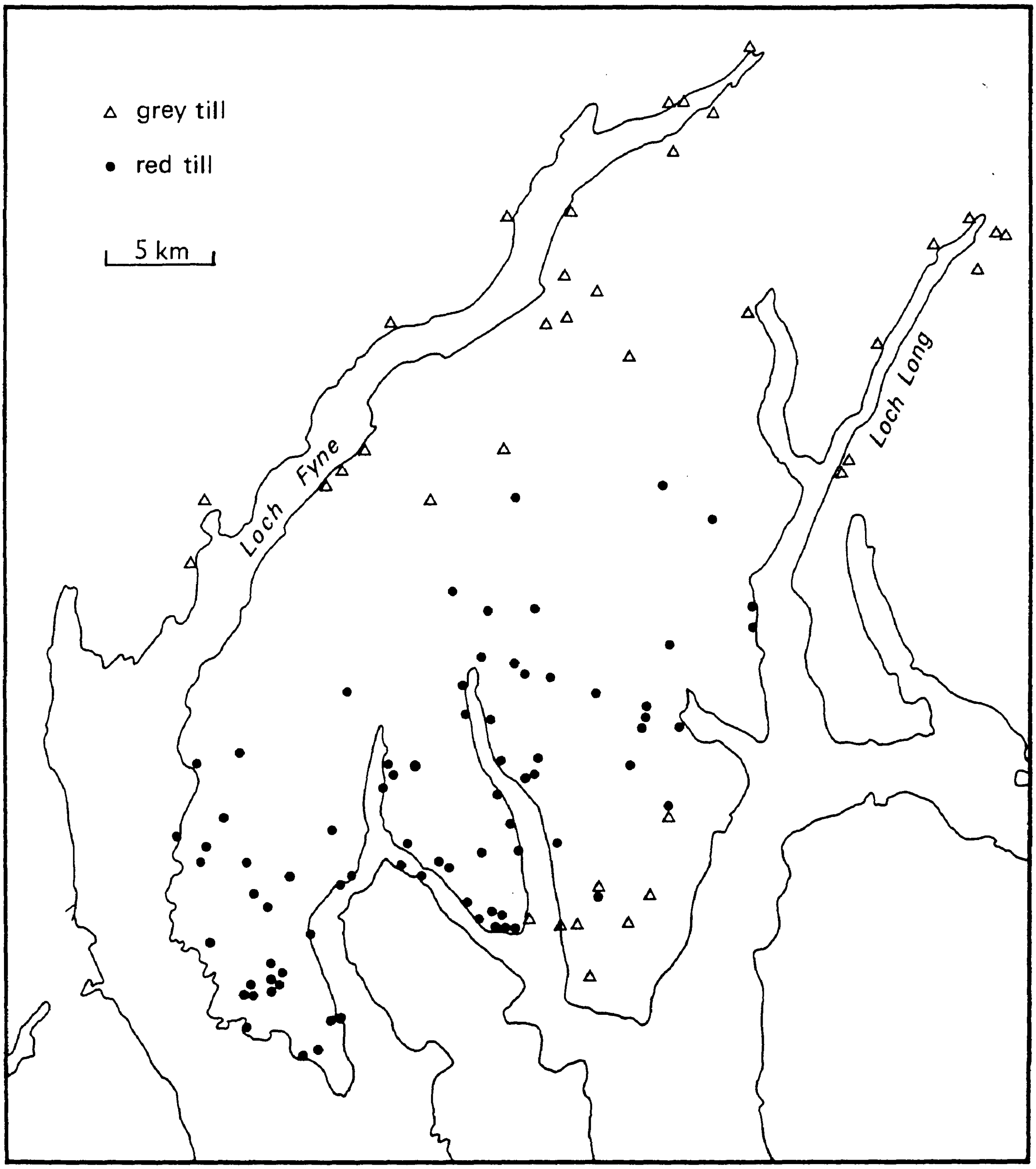
An ice-sheet surface profile is not affected by the degree of isostatic depression of the ground under the ice (Weertman, 1961) but this does influence the total ice thickness over a given point. Since deglaciation there has been a relative sea-level fall in the southern Cowal Peninsula of c.40m during a time when world sea level has risen c.40m (Fairbridge, 1961), thus indicating a minimum isostatic depression of 80m. Taken with the profile heights this suggests a minimal estimate of the maximal ice thickness in the northern Cowal Peninsula of 1300 - 1400m. These theoretical figures compare reasonably with observations, for in the SW Highlands the highest mountains reach over 1100m. The above calculations suggest that they were overtopped by at least 200m of ice at the last ice-sheet maximum.

During the course of fieldwork note was made of the occurrence and characteristics of exposures of till. From this and from the references of Gunn et al. (1897), Figure 8 was compiled to show the distribution of exposures of differently coloured tills, care having been taken in fieldwork to distinguish weathered from unweathered till. Three distinct areas of till can be seen, a bluey-grey one to the NW, a fawny-grey till in the SE and a band of red till between the two running in a SW to NE direction from lower Loch Fyne to Loch Long south of its junction with Loch Goil. It should be noted that the apparently greater number of exposures of red till is due to initially selective field observation as much of the blue-grey till to the NW was not recorded when these areas were mapped as the variation in colour in the Cowal Peninsula was not then appreciated.

It is thought that all the tills were derived from the last period of glaciation. Although none of the tills was seen overlying another during fieldwork, Gunn et al. (1897) recorded a small number of localities where more than one till was observed in the same section. At the Badd, W of Dunoon, red and brown tills were seen to alternate in one section and NE of Kilmarnock Hill (N of Ardyne) a grey till was noted overlying a red one. Often on the ground surface tills of different colour can be seen in close proximity to one another: thus on the foreshore around Strone Point at the mouth of Loch Striven a short walk can reveal both red and fawn tills. Gunn et al. recorded similar instances elsewhere.

Interest in the past has centred on the reason for the colour of the red till. Crosskey (1865) considered that it was due to the ferruginous character of the neighbouring bedrock. After considerably more fieldwork, however, Gunn et al. (p.251) commented:

Figure 8: Colour of till in study area.



grey till
red till
5 km
Loch Fyne
Loch Long

> 'It is quite clear that the difference in colour of the boulder clay cannot be ascribed to difference of composition of the rocks on which it lies. The rocks underlying the red clay are similar to those underlying the blue clay.'

No alternative hypothesis was offered by these authors.

The rather categorical denial of a relationship between the till and the underlying bedrock does not seem fully justified when a comparison is made between the distribution of the red till (Figure 8) and the geological map (Figure 2). The majority of the area underlain by red till is seen to be largely coincident with the Beinn Bheula Schists. Such a relationship is supported by the changes in the colour of the till in the SE, for here the ice flowed across the various strata at almost 90° to the strike of the rocks, this resulting in the rapid appearance of brown then fawny-grey till as the ice moved from the Beinn Bheula Schists first on to the Green Beds then on to the Dunoon Phyllites. Two areas appear anomalous, however. The Beinn Bheula Schists in the NE of the Cowal Peninsula are covered by grey till and in the SW the Green Beds and the Garnetiferous Mica Schists E of Kilfinan are also covered by red till. In this latter case it could be argued that the ice movement was south-westwards from the Beinn Bheula Schists located to the NE (striations and ice moulding indicate this) but such an explanation contrasts with the SE of the peninsula where the red colour disappears rapidly where the ice left the Beinn Bheula Schists. Despite these anomalies, however, it seems most likely that the red colour of the till is derived from the Beinn Bheula Schists.

TABLE 1. BENCH MARKS COMMON TO NEWLYN AND LIVERPOOL DATUMS

B.M. Grid Reference	Newlyn	Liverpool 1st ed.	Liverpool 2nd ed.
NN20020228	12.44	12.16 (-0.28)	12.37 (-0.07)
NN19850178	7.76	7.47 (-0.29)	7.68 (-0.08)
NS06067201	35.87		35.81 (-0.06)
NS06497174	4.22		4.18 (-0.04)
NS06877150	4.40		4.39 (-0.01)
NR99598318	11.84	11.40 (-0.44)	
NR99978517	20.04	19.60 (-0.44)	
NS00338249	35.79	35.36 (-0.43)	
NR99827457	2.27		2.10 (-0.17)
NS03347454	8.80		8.69 (-0.11)
NR98007289	4.25		4.11 (-0.14)
NR98289237	5.39	4.97 (-0.42)	
NR98749261	19.87	19.42 (-0.45)	
NR97509173	5.07	5.24 (+0.17)	
NR96829052	17.64	17.98 (+0.34)	
NN19441272	3.82		3.69 (-0.13)
NN18841266	4.92		4.82 (-0.10)
NN17741160	5.75		5.58 (-0.17)
NN11401009	10.19	10.03 (-0.16)	
NN09550860	4.91	4.69 (-0.22)	
NN09580841	7.00	6.83 (-0.17)	
NN08200646	7.05	6.61 (-0.44)	
NR94728712	9.37	8.85 (-0.52)	
NR93028445	5.22	4.75 (-0.47)	
NR93138425	30.05	29.54 (-0.52)	
NN18041074	11.05		10.91 (-0.14)
NN18421127	6.85		6.71 (-0.14)
NS12646755	6.16	5.91 (-0.25)	
NS13506785	8.69	8.44 (-0.25)	
NS18798756	4.73	4.36 (-0.37)	
NS18948733	7.79	7.44 (-0.35)	
NS00939499	6.80	6.61 (-0.19)	
NR96176819	36.14		35.97 (-0.17)
NN07160003	8.23	7.99 (-0.24)	
NR95617057	44.76		44.59 (-0.17)
NS09207486	6.93	7.01 (+0.08)	
NS09007529	4.08	3.87 (-0.21)	

CHAPTER 3

THE PRESENT SHORE

1. External Factors

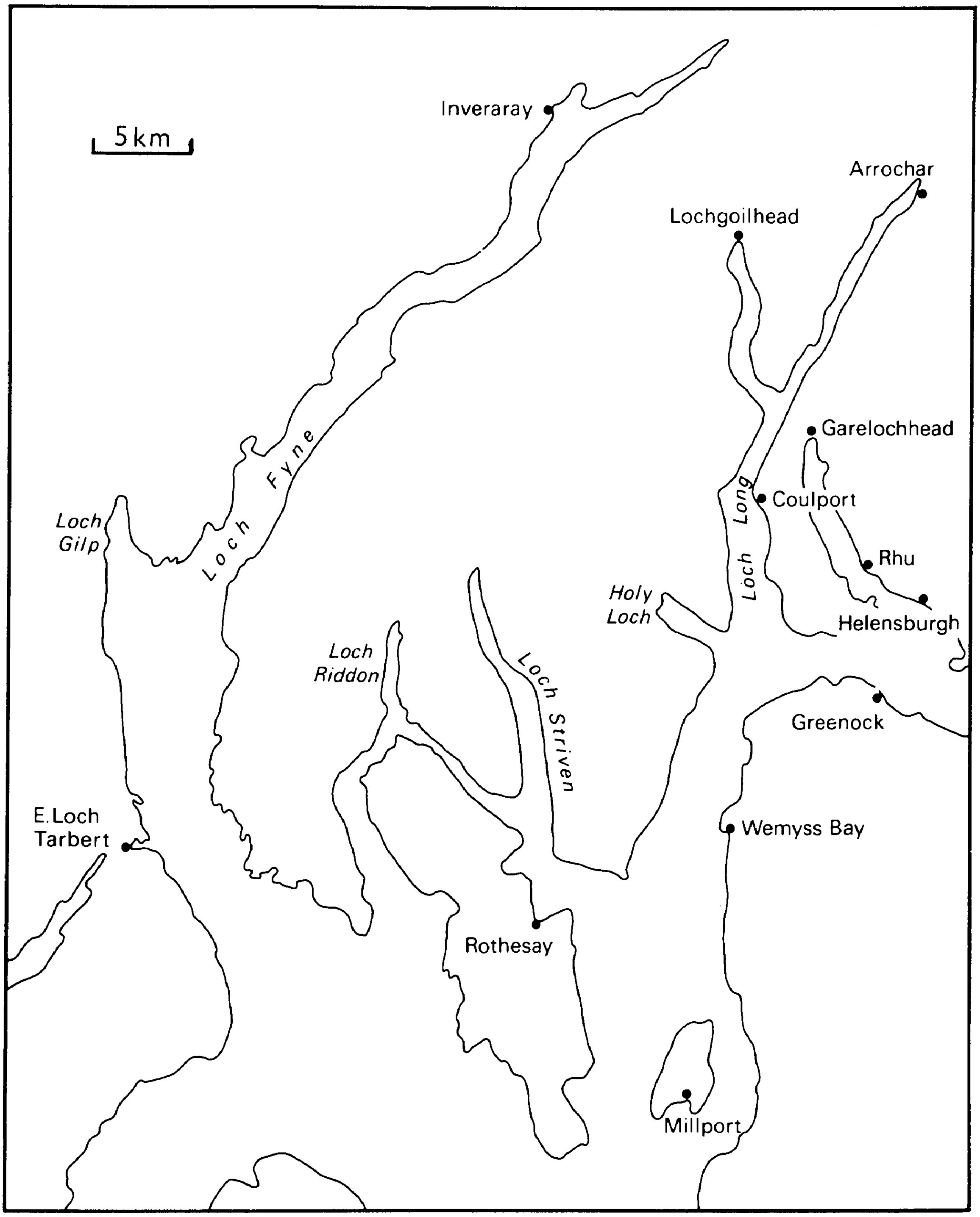
The nature of the shore zone and the types of marine landforms created along a given length of coast are dependent upon a number of variables such as the tidal regime, the nearshore gradient, the climate, the supply of sediment and the coastal configuration. As background to the present study these variables and how they affect the present shore of the study area are discussed below. In understanding the raised shoreline features around the coasts of the sea-lochs the present-day marine landforms are of greatest utility when conditions were quite analogous to those of the present, that is during the Flandrian. During the Lateglacial, however, major changes occurred in these variables and the utility of present-day conditions in understanding the shore-forming processes that operated during the Lateglacial is greatly reduced.

(a) Tidal Regime

The Firth of Clyde is a major shipping route and the sea-lochs to the north (lochs Long, Goil, Gare, Fyne and Holy Loch) are important for the nuclear submarine bases located there. An excellent coverage of tide gauges is the result. Figure 9 shows the localities with tide gauges and Table 2 compiles the various features of the tides at these stations, the original data being taken from Admiralty Tide Tables (1975) and adjusted from Chart Datum to Ordnance Datum. In addition to the stations indicated on Figure 9, Table 2 includes Campbeltown on the Mull of Kintyre for purposes of comparison with a station open largely to oceanic influences.

Inspection of Table 2 indicates that there are a number of relatively consistent changes in tidal regime from the dominantly oceanic station at

Figure 9: Location of tide gauges.



5 km
Inveraray
Arrochar
Lochgoilhead
Garelochhead
Coulport
Rhu
Helensburgh
Loch Gilp
Loch Fyne
Loch Long
Holy Loch
Loch Riddon
Loch Striven
Greenock
E. Loch Tarbert
Wemyss Bay
Rothesay
Millport

Campbeltown to the head of the sea-lochs at such places as Inveraray and Arrochar. Mid-tide altitude decreases up-loch (e.g. Wemyss Bay, 0.36 m O.D.; Greenock, 0.31 m O.D.; Coulport, 0.26 m O.D.; Lochgoilhead, 0.21 m O.D.) although there are minor variations on this (Arrochar, 0.26 m O.D.). Tidal range, however, increases up-loch for both Spring and Neap tides (e.g. Campbeltown, 2.6 m. Springs, 1.6 m Neaps; E Loch Tarbert, 2.7 m, 1.8 m; Inveraray, 3.1 m, 2.5 m) and this increase in tidal range largely balances the decrease in mid-tide altitude such that mean high water of both Spring and Neap tides is rather uniform across the area, having a range of only 0.2 m and most stations being within 0.1 m of each other.

In summary it can be said that those features of the shoreline that form with respect to the upper limit of marine activity will show little variability due to the tidal regime, whilst those features that relate to mid-tide altitude might show a slight decline in altitude towards the heads of the sea-lochs. In general, however, the magnitude of the variations in the tidal characteristics is rather low.

(b) Coastal Configuration

The sea-lochs of the Clyde Sea Area are typically rather sinuous, long and narrow. The Firth of Clyde itself contains a number of islands, Arran, Bute and the Cumbraes being the largest. These two factors combine to limit very greatly the exposure of the coasts considered in this study. No attempt has been made to characterise the degree of exposure of all the various sections of coast but fetch has been measured in connection with the study of various features of the present shore (see below) and the general characteristics have become apparent during these studies.

Maximum fetch exposure is experienced by those stretches of coast-line facing down the Firth of Clyde between the islands. Ardlamont Point and Toward Point are the most exposed localities although the maximum fetch is only 130 km, indicating these localities to be relatively sheltered as compared to stretches of coast open to the ocean. With the maximum exposure being in a southerly direction and the valleys (and coastlines of drowned valleys) having dominant NE to SW and NW to SE trends, many stretches of coast that are exposed to the longer fetches to the south lie at quite low angles to these directions of exposure. This considerably diminishes the effect of any waves that might arrive from such directions.

Within the sea-lochs fetch is often less than 10 km and once again maximum fetch is often along the loch and hence at a rather low angle to the coast, thus diminishing even further the effectiveness of waves from these directions. In a world-wide sense, therefore, the coastline being considered in this study is rather sheltered. It is of note, however, that the lengths of maximum fetch to which different sections of the coast are exposed vary through 2 orders of magnitude, from less than 10 km to over 100 km. This range of values suggests that fetch might be a significant variable in explaining differences from place to place in the nature of the marine landforms.

(c) Sediment Supply

The degree of glaciation experienced in this area has greatly disrupted any possible river system that may at one time have drained it (Chapter 2). Today the sea penetrates far inland and the sea-lochs receive sediment from a relatively large number of small (generally 4th or 5th order) drainage basins. The total area of land draining into the sea-lochs is, however, very small and is given in the following table taken from Mill (1891, p.651):

TABLE 3. Drainage Areas of Principal Sea Lochs

Loch	Land Drainage (km^2)
Gareloch	32.1
Loch Goil	87.6
Loch Long	75.3
Holy Loch	187.7
Loch Striven	95.6
Loch Riddon (and Kyles of Bute)	174.1
Loch Fyne	487.8

Such limited catchment areas for a considerable length of coastline (greater than 350 km) suggest a relatively small total volume of sediment delivered to the coastal zone, a volume reduced even further by the occurrence of a number of fresh-water lochs not far above sea-level that act as sediment traps (e.g. Loch Dubh near Inveraray, Loch Eck). This feature of low sediment supply contrasts markedly with the E coast of Scotland where relatively large rivers such as the Forth and the Tay supply sediment to a relatively restricted length of coastline.

Another feature of the drainage basins of the study area is the high relief:area ratios suggesting that the rivers should have a fast-flowing regime which, taken with the abundant supply of unconsolidated clastic material provided by glaciation suggests in turn that the streams and rivers should supply a high proportion of gravel and cobbles to the coast. This fluvial activity is indeed observed both in such dramatic events as the Cowal 'Landslip' (Smellie, 1912) and in the intertidal aprons of gravel that occur at the mouths of most streams.

(d) Nearshore Gradient

The efficiency of waves in moving material along (and to and from) the shore is partly dependent upon the nearshore gradient. A low nearshore gradient can result in reduced wave energy on the foreshore due to the wave touching bottom far from the coast. It can also promote constructive wave action by encouraging the wave to run rather than plunge upon breaking. Conversely, a steep nearshore gradient allows more of the wave energy to be applied directly to the foreshore though in a manner more favourable to destructive action. Owing to glacial over-deepening the sea-lochs considered in this study are characterised by very rapidly descending offshore profiles. This is illustrated on Figure 11 in which the 10 fathom (approx. 20 m) offshore contour is drawn. This generally follows the coastline closely and at several places is indistinguishable at this scale from the coastline. Exceptions occur at the heads of the lochs where sediment is supplied in relatively abundant quantities, at Otter Ferry where a major ice limit occurs (Chapter 7) that is also located on the threshold between the middle and the outer basins of Loch Fyne, at Minard on the threshold between the upper and the middle basins of Loch Fyne and, perhaps least expectedly, around the headlands exposed to greatest fetch, namely Toward Point, Ardlamont Point and between Otter Ferry and Loch Gilp. This last feature is perhaps explained by these headlands having been the focus of marine erosion at sea-levels below that of the present combined with their lee-side position with respect to the general ice flow during ice-sheet times.

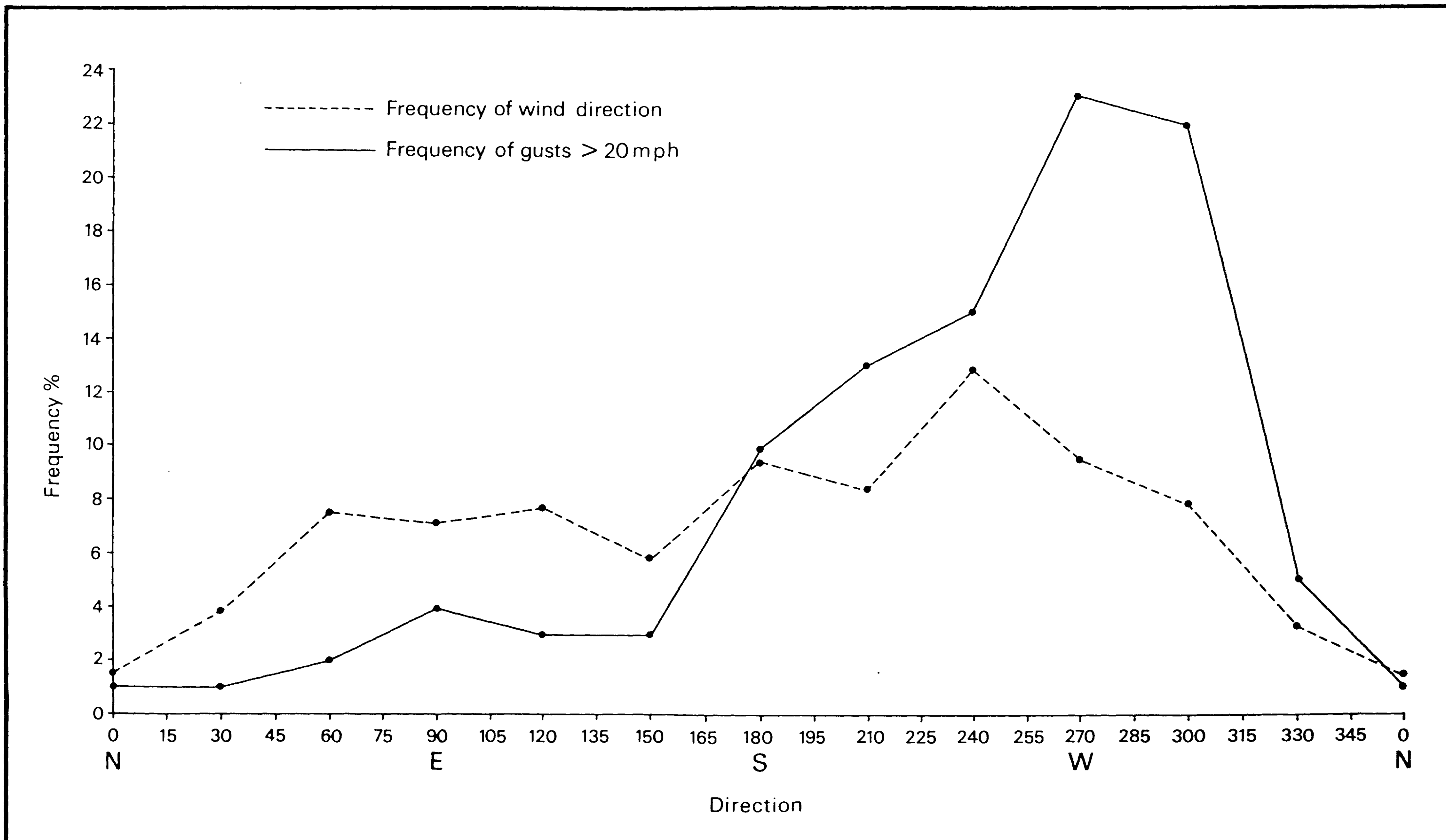
(e) Climate

There are no weather stations in the study area that have produced observations over a long period at a high standard. The nearest such station is Prestwick Airport (Plant, 1971). Prestwick is situated on a low open area that contrasts markedly with the mountainous Cowal Peninsula with its long narrow valleys. This latter condition severely influences the local climate, channelling winds, producing frost hollows and temperature inversions and modifying the precipitation pattern so that the climatic data from a station such as Prestwick are perhaps best regarded as the regional conditions on which the local effects are superimposed.

Of greatest relevance to the understanding of coastal processes is the wind pattern, the distribution of wind directions and the frequency of high winds. Figure 10 shows for Prestwick the frequency of winds from all directions as well as the frequency of gusts of more than 32 km/hr (20 mi/hr). Expectably, this shows a dominance of winds from the SW quadrant although there is a shift to winds from around W and slightly N of W for high gusts. In the study area the winds are noticeably channelled along the valleys (in Glendaruel it is locally said that there are only two wind directions, up the valley and down the valley) and an increase in the frequency of winds oriented along any valley can be expected. It is of note that the strongest winds at Prestwick blow most frequently from directions that are across the 'grain' of the pattern of valleys in Cowal and some overall diminution in their force may result from this.

It is doubtful if sea-ice plays any part in shore formation around the present coast of the study area although Loch Fyne freezes over relatively easily due to its low salinity. During the 'Little Ice Age', however, sea-ice appears to have occurred with greater frequency. Gunn et al. (1897, p.268) recorded that a large boulder was moved during the nineteenth century by sea ice and Pearson (1973) documented three occasions during the eighteenth century (1758; 1772; 1786) when the upper part of Loch Fyne was completely frozen over. In 1758 the ice was measured as being 6 inches thick in the middle of the loch between Inveraray and St.Catherines and in 1786 the whole of the loch down to Otter Ferry (more than 40 km from the head) was recorded as being frozen over to the extent that men could walk across (Pearson, unpubl.).

Figure 10: Wind frequencies at Prestwick.



Frequency of wind direction
Frequency of gusts > 20 mph
Frequency %
0
2
4
6
8
10
12
14
16
18
20
22
24
0 15 30 45 60 75 90 105 120 135 150 165 180 195 210 225 240 255 270 285 300 315 330 345 0
N E S W N
Direction

2. Landforms of the Present Shoreline

To the extent that the above mentioned conditions are similar to those that obtained in the past, then the present shore forms can be used as analogues for those that are raised above present sea-level. This is a reasonable supposition for those landforms created during the Flandrian, but clearly a considerable number of changes occurred during the Lateglacial. If, however, the landform is mainly dependent for its formation upon a variable such as nearshore gradient that is largely independent of Lateglacial environmental conditions then provided the coastal profile of the time is known the modern landform may be viewed as the counterpart of the Lateglacial feature.

Three distinct types of landform were investigated on the present shore: shingle ridges, flat-topped gravel terraces (i.e. intertidal deltas), and wave-cut platforms. These three were chosen since the methodology adopted for the raised shoreline study involved identification of such features and not of the other more diffuse types of marine landforms found in the study area.

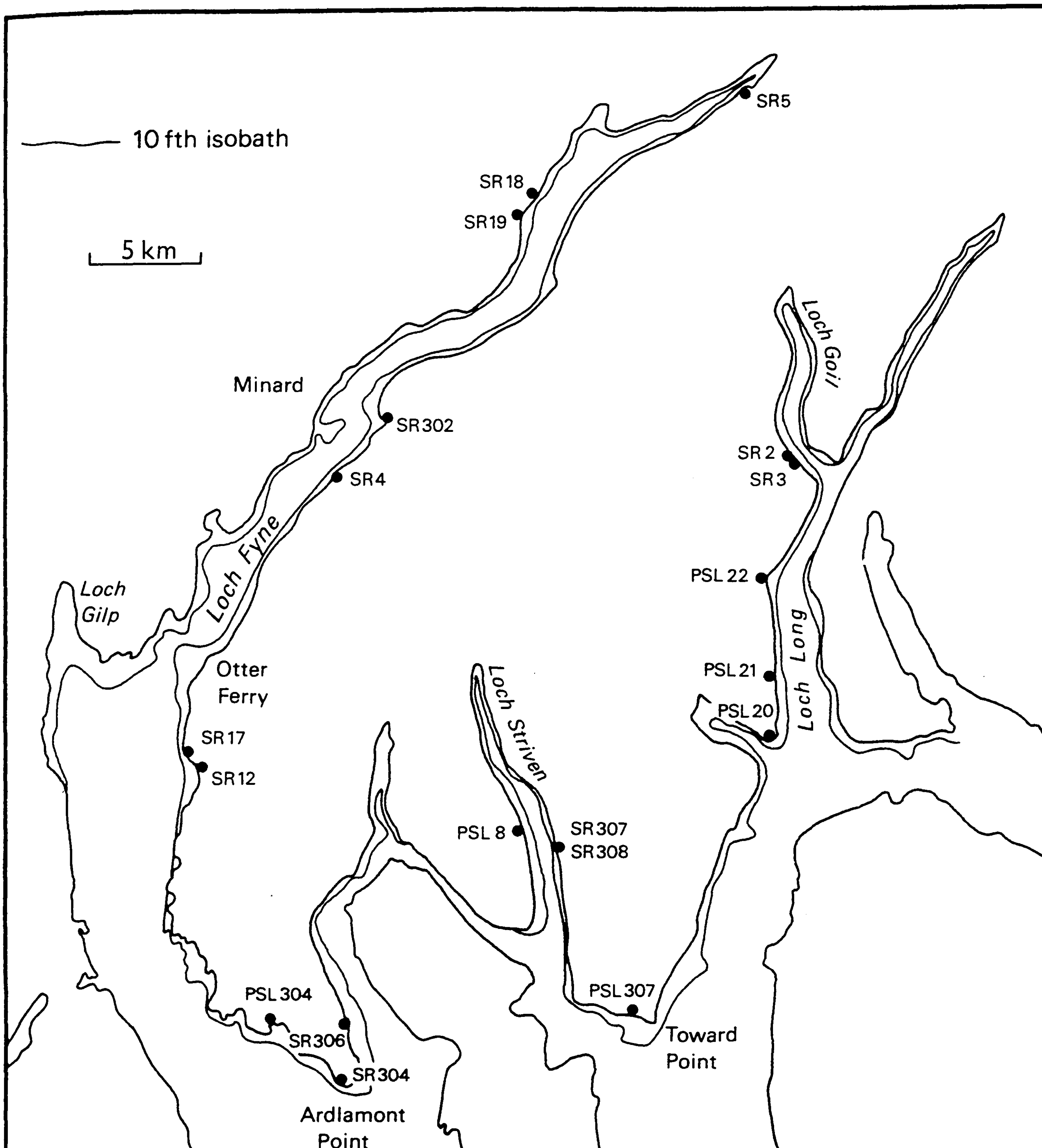
(a) Shingle Ridges

Shingle ridges are a common feature of the present shoreline and are developed at numerous localities throughout the study area, ranging from the exposed headlands of the south of Cowal to the heads of the sea-lochs. The construction of a large shingle ridge appears dependent upon the continuing supply of suitably sized sediment and many of the rivers and streams have shingle ridges associated with the rather complex deltas that characterise their mouths. Ardyne Point provides an interesting piece of evidence in connection with the role of sediment supply in the formation of shingle ridges. The stretch of coastline flanking either side of the mouth of the Ardyne Burn has been characterised by extensive shingle ridge development throughout much of the Flandrian (Chapter 6) and the 1:10,560 map of 1869 indicates a large shingle ridge to have existed at present sea-level to the west of the mouth of the burn. When this area was first visited in 1972, prior to the establishment of the North Sea Production Platform Yard, the mouth of the Ardyne Burn was used as a source of sand and gravel for the Dunoon construction industry. No sign could be seen of the shingle ridge to the west of the mouth of the burn, though thin layers of a fibrous peat were found on the foreshore where the lagoon behind the ridge formerly stood. The top of the foreshore was undergoing erosion and it was inferred that the extraction of sand and gravel from the mouth of the Ardyne Burn had been of sufficient magnitude to change the shore forming processes from a constructional to an erosional regime.

Shingle ridges along the present shoreline were mapped at 1:10,560 scale and heights were surveyed every 50 paces along their crest lines. In addition banks of shingle that were observed at the head of certain beaches were also surveyed at 50 pace intervals. The distinction made between a ridge and a bank is that a ridge has a clear crest line whilst a bank has been constructed against a pre-existing feature (e.g. a small cliff) such that there is no drop in altitude on its landward side. Six shingle banks and 13 shingle ridges were surveyed at a total of 76 individual points. The mean altitudes of the various features are given in Table 4 and their location shown on Figure 11.

The features surveyed ranged from being vegetation free to being quite covered in vegetation and it is difficult in every instance to be certain that a shingle ridge is the product of the sea at its present level

Figure 11: Location of shingle ridges surveyed around present coast.


10 fth isobath
5 km
SR5
SR18
SR19
Minard
SR302
SR4
Loch Fyne
Loch Gilp
Otter Ferry
SR17
SR12
Loch Goil
SR2
SR3
PSL 22
Loch Long
PSL 21
PSL 20
Loch Striven
PSL 8
SR307
SR308
PSL 307
Toward Point
PSL 304
SR306
SR304
Ardlamont Point

and not at a slightly higher previous level. During fieldwork attention was paid to the existence of freshly-flung cobbles on the surface of vegetated beach ridges or the presence of recently washed sea-drift on the flanks or crests of such ridges, these observations being used to decide whether the ridge should be considered as part of the present shore.

The ridge altitudes ranged from 2.9 to 4.2 m O.D. with a mean of 3.52 m O.D. and a standard deviation of 0.45 m. To some extent this range of altitude may be a reflection of the fact that the sea can construct ridges at a variety of altitudes at any one locality depending upon the conditions that obtain at the time. This is frequently observed along the present coast where a large ridge at the head of the beach is followed by two or three smaller ridges lower down. Hence only the ridge farthest up the beach was measured at each locality.

The altitudinal variation in ridge crests was investigated by measuring the maximum fetch at each locality since fetch is considered to be a measure of the wave power that may be experienced at that locality. Because shingle ridges are dependent upon waves to propel the shingle up the beach it may be thought that more powerful waves generated over longer fetches would build higher ridges. This supposition seemed to be supported by the occurrence of the highest shingle (e.g. SR304, PSL20) in exposed southward-facing localities. The maximum fetch at each locality is presented in Table 4.

Regression analysis of the shingle ridge altitudes against the maximum fetch revealed a significant relationship in which $r = +0.7090$. Two ridges, SR5 and SR18, deviated markedly from this relationship, being rather high in relation to the fetch available. The possibility therefore exists that these are related to sea-levels above that of the present and the regression was re-calculated leaving these two points out. A stronger relationship emerged with r = +0.8183.

It was noted during the measurements of maximum fetch that the trend of the ridges was often at a rather low angle to the direction of maximum fetch. To some extent this would be mitigated by refraction of the waves as they crossed the shallow nearshore zone but it was thought that the amount of energy that a wave could provide in propelling shingle up a beach would be best monitored by the component of wave energy measured at 90^o to the trend of the ridge. Accordingly, the maximum fetch was resolved into two components, the one directed parallel to the ridge and the other normal to it. In certain instances the component of maximum fetch normal to the ridge was found to be much smaller than the component of fetch normal to the ridge from another direction (this applied, for example, with PSL8 and PSL21) and in these cases the maximum fetch value at right angles to the ridge was used. Computation of the relevant correlation coefficient revealed an improved relationship over maximum fetch, r being +0.7274. Once again SR5 and SR18 appear discrepant and recalculation without them gave a correlation coefficient of +0.8353.

The consideration that wave power is not linearly related to fetch (Darbyshire, 1956) suggested a further modification of the fetch variable. As fetch increases each additional kilometre of fetch has a diminishing added effect to wave energy which thus tends to a limit whereafter further increases in fetch do not add to the energy of the wave spectrum. This suggests a logarithmic type of relationship and the natural logarithm of the component of maximum fetch normal to the ridge crest was therefore regressed against shingle ridge altitude. An improved relationship was found, with r = +0.7586 for all nineteen points and r = +0.8760 when ridges SR5 and SR18 were omitted.

An attempt was made to see if the height of the shingle ridges was in any way influenced by the offshore gradient. As noted above the coasts of the study area are generally very rapidly shelving but it was thought that since the shingle ridges must have formed at high tide and since there was some considerable variability in the width of the intertidal zone, then the waves may have been sufficiently modified to have some influence on shingle ridge development. The width of the intertidal zone was measured from 1:10,560 scale maps (Table 4). Regression of this distance against height of shingle ridge revealed no significant relationship (r = +0.0457 for all points; r = +0.0831 excluding SR5, SR18). There is, however, some variation in the range of Spring tides in the area and the true nearshore gradient is the Spring tide range divided by the distance between tide marks. This was calculated ($\tan\alpha$, Table 4) and regressed against ridge altitude and a much stronger negative relationship was found, with r = -0.4688 for 17 points and r = -0.4797 for all 19 points, both these results attaining significance at the 95% confidence level.

These significant negative correlations indicate that steep nearshore profiles result in lower shingle ridges, apparently implying that wave energy is dissipated more rapidly on a steep beach. This is initially unexpected for the lower the nearshore gradient the farther offshore a wave will touch bottom and begin to loose energy. The resolution of this problem perhaps lies in the wave form close to the beach. King (1972) indicates that on steep beaches waves tend to plunge upon breaking, partly because of high wave steepness values and partly because of high backwash velocities. Less steep nearshore profiles produce waves that have a greater tendency to run and are thus more efficient in moving material landwards.

Multiple regression analysis was performed on the main variables: shingle ridge altitude, natural logarithm of the component of maximum fetch normal to the trend of the ridge, and the nearshore gradient. This produced the equation

$$H = 2.76 + 0.31 \ \ln CMF - 0.93 \ \tan\alpha$$

which has a multiple correlation coefficient of +0.8856 using the 17 points that exclude SR5 and SR18.

The base constant of this equation, 2.76, represents the altitude at which shingle would accumulate with a fetch normal to the ridge of 1 km ($\ln 1 = 0.0$) and zero nearshore gradient. This latter condition is unlikely (it implies an infinitely wide intertidal range or zero tidal range) but in practice the shoreline gradient term ceases to be significant for intertidal distances greater than 300 m for a tidal range of 3 m ($\tan\alpha = 0.01$, below which only the third decimal place is affected in the equation), or from a tidal range of less than 0.2 m with intertidal distances above 20 m.

Given these minimum conditions the equation implies that shingle of the sediment sizes available in the study area under the wind climate of this area will be moved to an altitude of 2.76 m O.D., that is, approximately 1 m above MHWOST. Shingle movement to altitudes above this level is dependent upon variations in the fetch and nearshore gradient related to a particular ridge.

(b) Intertidal Flats

The present shoreline was examined to locate those features that appear flat and that can be regarded as analogues for the flat-topped features that are the object of the surveying on the raised shorelines. The most prominent and frequently occurring flats were intertidal deltas. These varied greatly in size, ranging from small features at the mouths of streams extending for 100 - 200 m along the coast to very large intertidal spreads of sand and gravel that occur at the heads of such lochs as Fyne and Riddon. The role of sediment supply is clear, firstly in ensuring the presence of a delta and secondly in governing the size of the feature. No attempt was made to quantify the particle-size distribution of the sediment in the deltas but visual assessment suggests the generalisation that the larger features tended to have a greater proportion of sand. Dominantly, however, the deltas are composed of coarse gravel, pebbles and occasional cobbles. Coarser material tended to accumulate in the steep landward rise that occurred towards the high water mark. One, occasionally two, distributary channels typically occurred on the surface of the deltas. These were avoided during surveying.

Cross-profiles were surveyed on the deltas examined, heights being recorded every two paces (c. 2 m). The locations of the surveyed profiles are shown on Figure 12 and the profiles are plotted on Figure 13. In general they reveal a lower portion of low gradient succeeded by a sharp rise to the back of the shore. One profile (8) exhibits a step above the first break of slope: this was a low bank of shingle. On the basis of these profiles the break of slope was established and used for subsequent analysis.

Figure 12: Location of shore profiles surveyed around present coast.

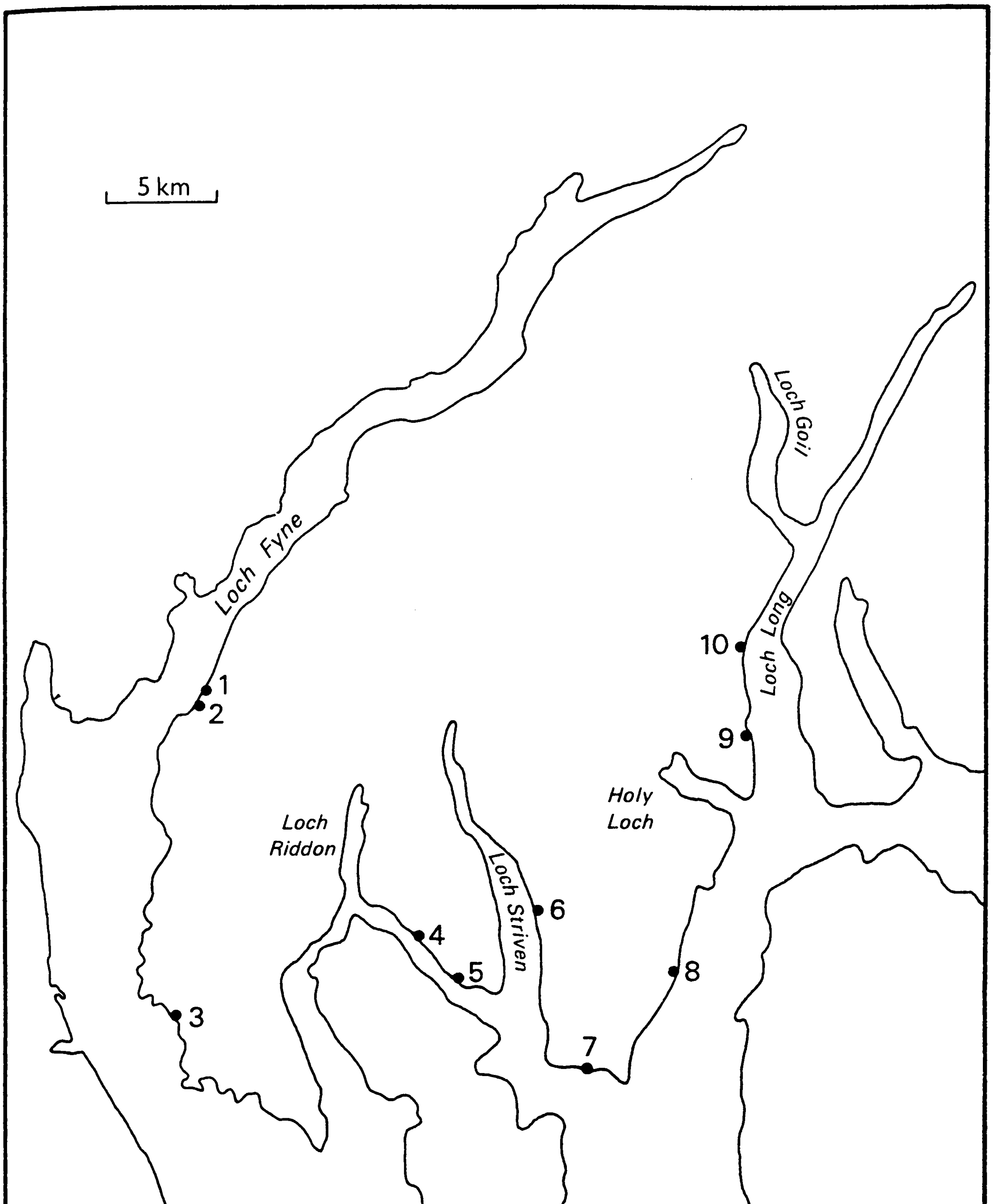
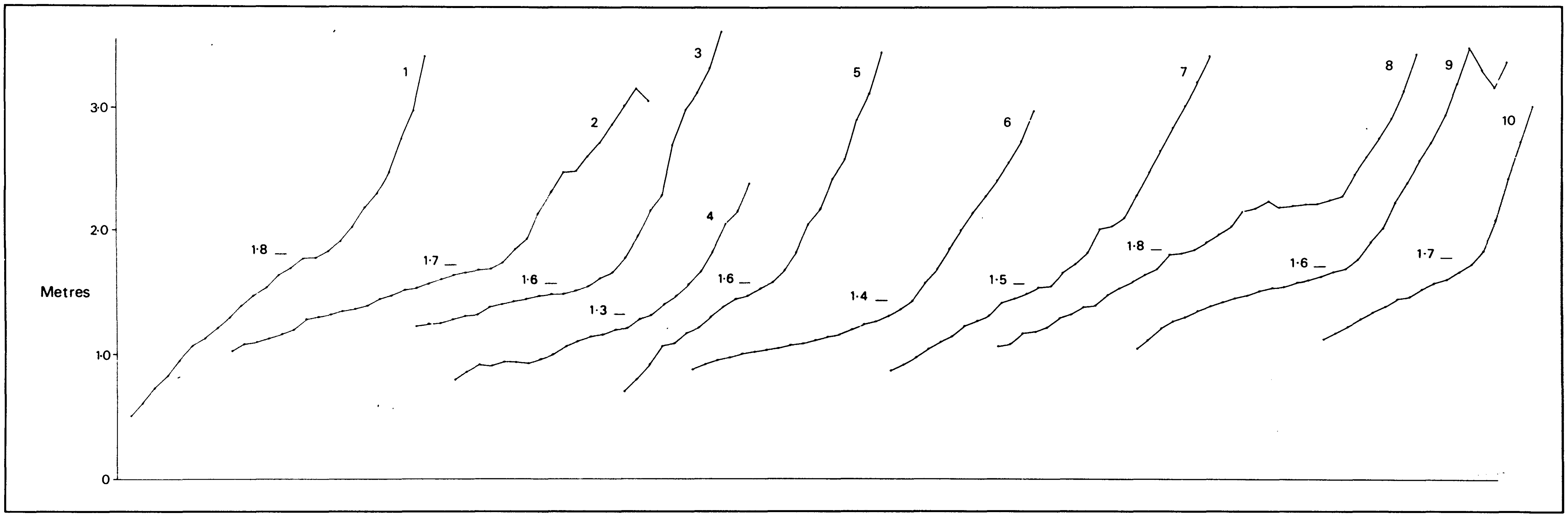

5 km
Loch Fyne
Loch Goil
Loch Long
Holy Loch
Loch Riddon
Loch Striven
1
2
3
4
5
6
7
8
9
10

Figure 13: Cross-profiles of intertidal flats. Each point 2 paces apart.



Metres
0
1·0
2·0
3·0
1
2
3
4
5
6
7
8
9
10
1·8
1·7
1·6
1·3
1·6
1·4
1·5
1·8
1·6
1·7

Ten profiles were surveyed in total. The breaks of slope ranged in altitude from 1.3 to 1.8 m O.D., with a mean altitude of 1.6 m O.D. and a standard deviation of 0.16 m. They are therefore rather consistent in altitude. Five of the profiles were located close to tide gauges and Table 5 below compares the heights of the break of slope with various characteristics of the tidal regime.

TABLE 5. Relationship between tidal regime and height of intertidal flats.

Profile	Break of Slope m O.D.	Tide Gauge	MHWOST m O.D.	MHWNT m O.D.	Mid-tide m O.D.
9	1.6	Coulport	1.78	1.28	0.26
10	1.7	Coulport	1.78	1.28	0.26
3	1.6	E Loch Tarbert	1.58	1.18	0.25
7	1.5	Rothesay	1.78	1.28	0.31
8	1.8	Wemyss Bay	1.78	1.28	0.36

From Table 5 it is clear that there is a close relationship between the altitude of the break of slope and MHWOST.

The close relationship to MHWOST and the small variation in the altitude of the break of slope suggests that there would only be a weak relationship with other variables such as fetch and nearshore gradient that were significant in explaining the variations in height of shingle ridges. Table 6 lists, for each profile, the maximum fetch, the component of maximum fetch normal to the shore and the nearshore gradient (tan α, the Spring tide range divided by intertidal distance). Correlation analysis was carried out between the altitude of the break of slope and each of the other variables. Table 7 below gives the results

TABLE 7. Correlation coefficients for intertidal flats.

	Max. Fetch	CMF	tan α
Break of Slope	+0.1137	-0.1933	-0.3607

Not one of these relationships achieves significance at the 90% level.

In conclusion it can be said that the intertidal flats have been shown to have a break of slope at their back edge at a rather constant altitude that relates closely to the MHWOST. No statistically significant relationships were found between the altitude of the break of slope and measures of fetch or nearshore gradient.

(c) Marine Erosional Features

At a number of localities, particularly in upper Loch Fyne, small rock-cut platforms and cliffs occur close to present sea-level. In many instances there is a scatter of sharp angular fragments on the surface of the platform indicating current erosion of bedrock. On the other hand, the cliffs that back the platforms are often vegetated and apparently not being eroded at present. In one locality (RP14) a set of striations was located on the front of the platform where a dyke cuts across the platform. These striations post-date the erosion of at least part of the platform though it is possible that they are the work of sea-ice rather than of glacier ice. The possibility exists, however, that in part these marine erosional features originated at a time other than the present.

The platforms are almost all developed on very finely foliated schists in localities where there are only poor supplies of sediment. Frost-action perhaps plays some role in the formation of the platform, for the finely-foliated schists absorb water along the foliation planes in a sponge-like manner. This water can be squeezed out underfoot, but if frozen it would act to wedge out thin angular fragments onto the platform surface. Such fragments are observed on the platforms even where the abrasion of the cliff appears limited. The relative importance of different platform-forming processes is, however, unknown. Along stretches of the coast where more massive rocks reach the sea there are no signs of marine erosion and glacial striations are abundant and clear. Such striated rocks can be seen in close proximity to platform fragments. For instance, in upper Loch Fyne the headlands (of Creag na h-Iolaire and at Kenmore) that flank platform fragments RP12, RP13 and RP14 are both clearly striated. The influence of rock type on platform formation is therefore decisive.

The junction of the platform and cliff was heighted at 50 pace intervals along each fragment. 30 points at 9 separate localities were surveyed in this fashion. The fragments ranged in altitude from 1.51 to 2.25 m O.D. with a mean fragment altitude of 1.91 m O.D. and a standard deviation of 0.23 m. The platforms are therefore forming at or slightly above MHWOST.

The relationships between platform altitude, maximum fetch, the component of maximum fetch normal to the local trend of the coast and nearshore gradient were examined in the manner used for shingle ridges and intertidal deltas. The data for each fragment are listed in Table 8 and the correlation coefficients in Table 9 below.

TABLE 9. Correlation Coefficients for Marine Erosion Features.

	Max. Fetch	CMF	lnCMF	$\tan \alpha$
Platform altitude	-0.6853	-0.6876	-0.7905	+0.5565

All these relationships attain significance at the 99% level.

A strong negative correlation is found between fetch and platform altitude indicating that the greater the fetch, the lower the platform back edge. The relationship is best for the logarithm of the component of maximum fetch normal to the shore, reflecting the non-linear increase in wave energy with increase in fetch. The reason for lower platforms related to longer fetches is not immediately clear. Perhaps the increased wave energy results in a lowering of the platform surface as well as retreat of the cliff.

The nearshore gradient relates in a positive way to the platform altitude and in part this may reflect the reverse of the argument advanced for the negative relationship established between nearshore gradient and shingle ridge altitudes. Steep nearshore gradients, it was argued, were adverse to the development of constructive wave action and hence the formation of shingle ridges. Rock platforms are erosional features and are presumably favoured by destructive wave action. The particular relationship to the altitude may be the result of increased wave height nearshore due to later building of the wave or to the wave breaking nearer the shore and hence being able to apply the wave energy farther up the shore.

The two relationships (fetch, nearshore gradient) will tend to operate against each other when considered in isolation as an explanation of platform altitude. Combined together in a multiple regression equation, however, they should produce a better explanation of the platform altitude. Such a regression equation was computed as

$$P = 1.97 - 0.13 \ \ln CMF + 1.71 \ \tan \alpha$$

This has a multiple correlation coefficient of +0.8505.

Following the argument pursued for a similar multiple regression equation for shingle ridge altitudes, the base constant of this equation indicates that under conditions of minimum nearshore gradient and 1 km of fetch, marine erosion would be effective in the study area at ca. 1.97 m O.D., that is, slightly above MHWOST.

(d) Present Shore Forms - Summary

The altitudinal variation of shingle ridges, intertidal deltas and rock platforms has been examined around the present coast of the study area. Tidal regime, sediment supply, fetch and nearshore gradient have been shown to be important variables in the development of these landforms. All the landforms are produced with reference to MHWOST. The height of the break of slope at the back of an intertidal delta approximates most closely to MHWOST, whilst rock platforms and shingle ridges occur above this altitude to varying degrees depending on fetch and nearshore gradient. Sediment supply largely governs the disposition of these landforms around the coast, rock platforms occurring where sediment supply is low (and rock structure not massive) and deltas and shingle ridges being related to an adequate supply of material.

Intertidal deltas are most consistent in their altitude, having a mean height of 1.60 m O.D. and a standard deviation of 0.16 m. Next most consistent are rock platforms, the mean height of which is 1.91 m O.D. and the standard deviation 0.23 m. To some extent this uniformity of rock platform altitude is due to the similarity of rock type on which the fragments examined were developed. Shingle ridges are most variable, having a mean altitude of 3.48 m O.D. and a standard deviation of 0.46 m.

3. Datum Planes

The use of Ordnance Datum as base level for surveying raised marine landforms not only provides a rigorously established framework on which to base local traverses but also serves as the basis for country-wide comparison of raised shoreline altitudes. Comparable use of Ordnance Datum to that of this study has been made on the W coast of Scotland by

Gray (1972a) and Dawson (1979). Certain other studies of raised shorelines along the W coast of Scotland have not taken advantage of this fundamental datum but have instead attempted to use certain of the characteristics of the present shore as a datum (e.g. Donner, 1959; Synge and Stephens, 1966). During this study detailed measurements were made on the upper limit of barnacle growth, the lower limit of land-based vegetation and the upper limit of sea drift at a variety of localities around the present shoreline. The objectives of such a study were to discover to what extent these features could be used as datum planes, how they related to the tidal cycle and, if they varied in altitude, on what this variation depended.

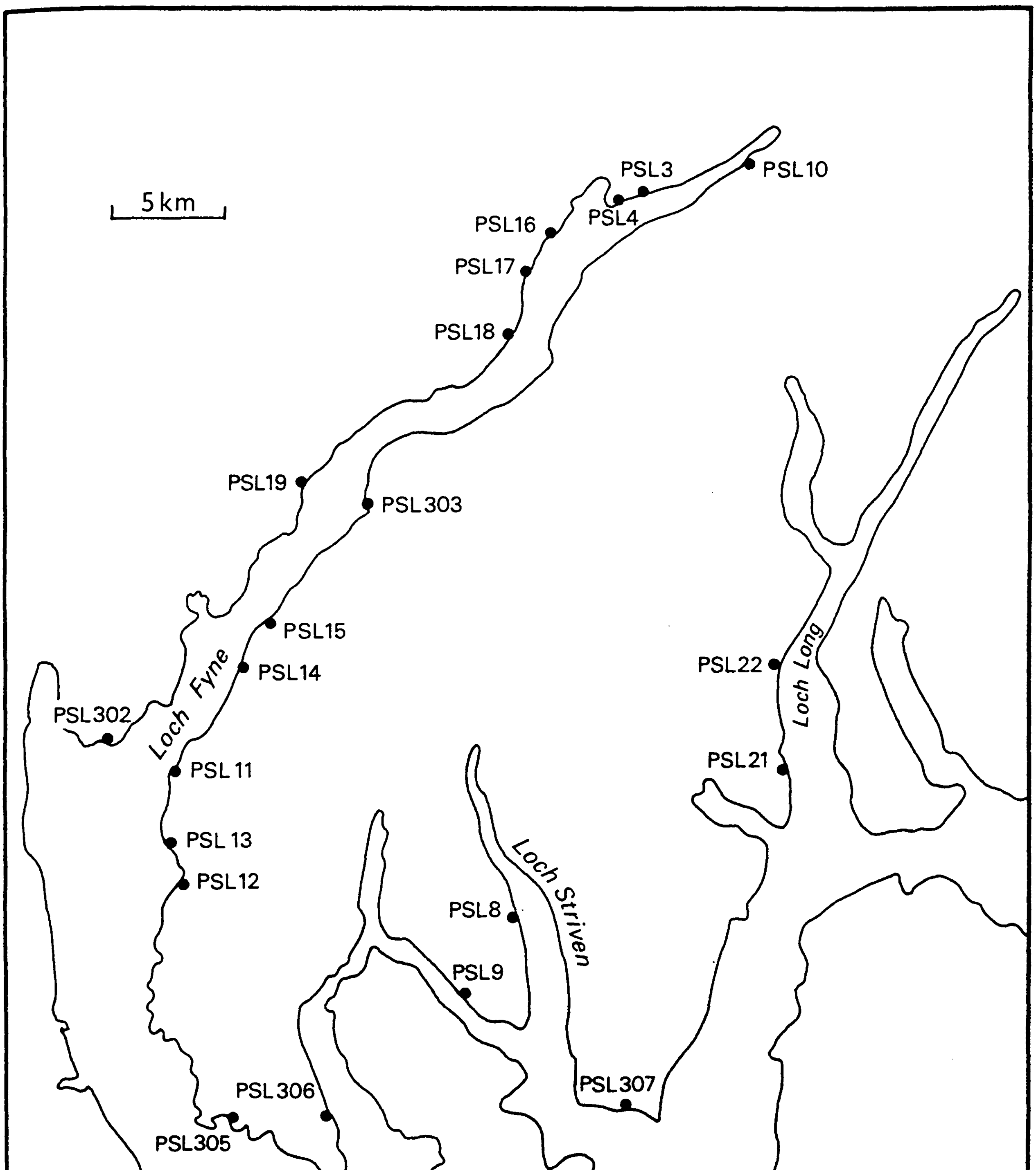
(a) Upper Limit of Barnacle Growth

The main types of barnacle found along the coasts of Argyll are Balanus balanoides and Chthamalus stellatus (Lewis, 1957; Lewis and Powell, 1958) both of which occur in the upper part of the tidal zone. No attempt was made to differentiate between barnacle types during the present study. At each locality surveyed several points, 50 paces apart, were heighted and in the subsequent analysis the means of the heights from the various localities were used. In all, 85 points at 21 separate localities (Fig. 14) were surveyed.

The barnacle limit ranges in altitude from 0.56 to 1.37 m O.D. and the overall mean altitude at the 21 localities was 1.16 m O.D. with a standard deviation of 0.21 m. Variability is slightly greater considering the 85 individual points, these giving a mean of 1.18 m O.D. and a standard deviation of 0.23 m.

Four of the localities surveyed were in the vicinity of tidal stations and the mean altitudes of the upper limit of barnacle growth from these

Figure 14: Location of sites where barnacle limit was surveyed.


5 km
PSL3
PSL10
PSL4
PSL16
PSL17
PSL18
PSL19
PSL303
PSL15
PSL14
Loch Fyne
PSL302
PSL11
PSL 13
PSL12
PSL22
Loch Long
PSL21
Loch Striven
PSL8
PSL9
PSL306
PSL305
PSL307

localities are compared with the local tidal characteristics in Table 10 below.

TABLE 10. Relationship between Tidal Regime and Barnacle Limit Altitude.

Location	Barn Alt. m O.D.	MHWOST m O.D.	MHWNT m O.D.	Mid-tide m O.D.
Coulport	1.29	1.78	1.28	0.26
Rothesay	1.22	1.78	1.28	0.31
E Loch Tarbert	1.20	1.58	1.18	0.25
Inveraray	1.03	1.68	1.38	0.13

A close relationship is suggested with MHWNT, only Inveraray apparently deviating from this in a significant fashion.

A multiple regression analysis was performed relating the height of the barnacle line at a locality to the Eastings and Northings at that locality. The regression equation was

$$B = 5.39 - 0.0041E - 0.0049N$$

with a multiple correlation coefficient of +0.4377, indicating a significant relationship at the 97% level. The equation suggests a decline of the barnacle limit towards the N and E, that is towards the heads of the sea lochs and this supports the deviation observed in the comparison of the barnacle altitudes in upper Loch Fyne and the tide gauge data. This result is different from that of Donner (1959) who documented a tendency for the barnacle line to rise towards the inner part of Loch Fyne, but Donner based this conclusion on only 3 surveyed points (at Campbeltown, Tarbert and Inveraray) and the possibility of arriving at an erroneous conclusion is correspondingly great. Scandinavian workers, however, who have used the barnacle limit extensively (e.g. Tanner, 1930) record it as falling in altitude towards the heads of fjords.

Such a fall in altitude towards the heads of the sea-lochs is indicative perhaps of an up-loch decrease in salinity. Mill (1891, p.697) gave a series of readings on the salinity (as measured by density differences and expressed as a % of normal sea-water) of the upper layer of the waters of the Clyde Sea Area. These readings can be paired with 17 of the barnacle line altitudes (Table 11) and regression analysis performed. The correlation coefficient using all the points listed in Table 11 is +0.6871 indicating a strong relationship between declining barnacle altitude and decreasing salinity. This relationship is superimposed on that between barnacle altitude and tidal regime as is most clearly indicated by the case of Loch Fyne. In Loch Fyne there is a marked tidal constriction at Otter Ferry and if the figures for salinity and barnacle line altitude for Loch Fyne up-loch of Otter Ferry are considered on their own, an even larger correlation coefficient is derived, +0.8263. The tidal regime down-loch of Otter Ferry as indicated by E Loch Tarbert tide gauge is characterised by a lower MHWNT altitude than upper Loch Fyne, and although salinity continues to increase down-loch of Otter Ferry, the barnacle line (as measured at localities PSL11 and PSL302) drops in altitude. Unfortunately too few points are available for a multiple regression analysis of tidal regime, salinity and barnacle line altitude.

In conclusion it can be said that the upper limit of barnacle growth occurs within a range of altitude of ca. 0.8 m and, if the mean of a series of observations at a locality is considered, the barnacle line can be estimated with an accuracy (1σ) of $\pm$ 0.21 m. It is dominantly related to the Mean High Water Mark of Neap Tides on which is superimposed a strong relationship to salinity such that towards the heads of sea-lochs the barnacle altitudes deviate markedly from MHWNT. The use of the barnacle line as a datum in such areas would without doubt be unreliable. In those areas where salinity is greater than 90%, the mean altitude of the barnacle line is 1.20 m O.D. and the standard deviation 0.12 m.

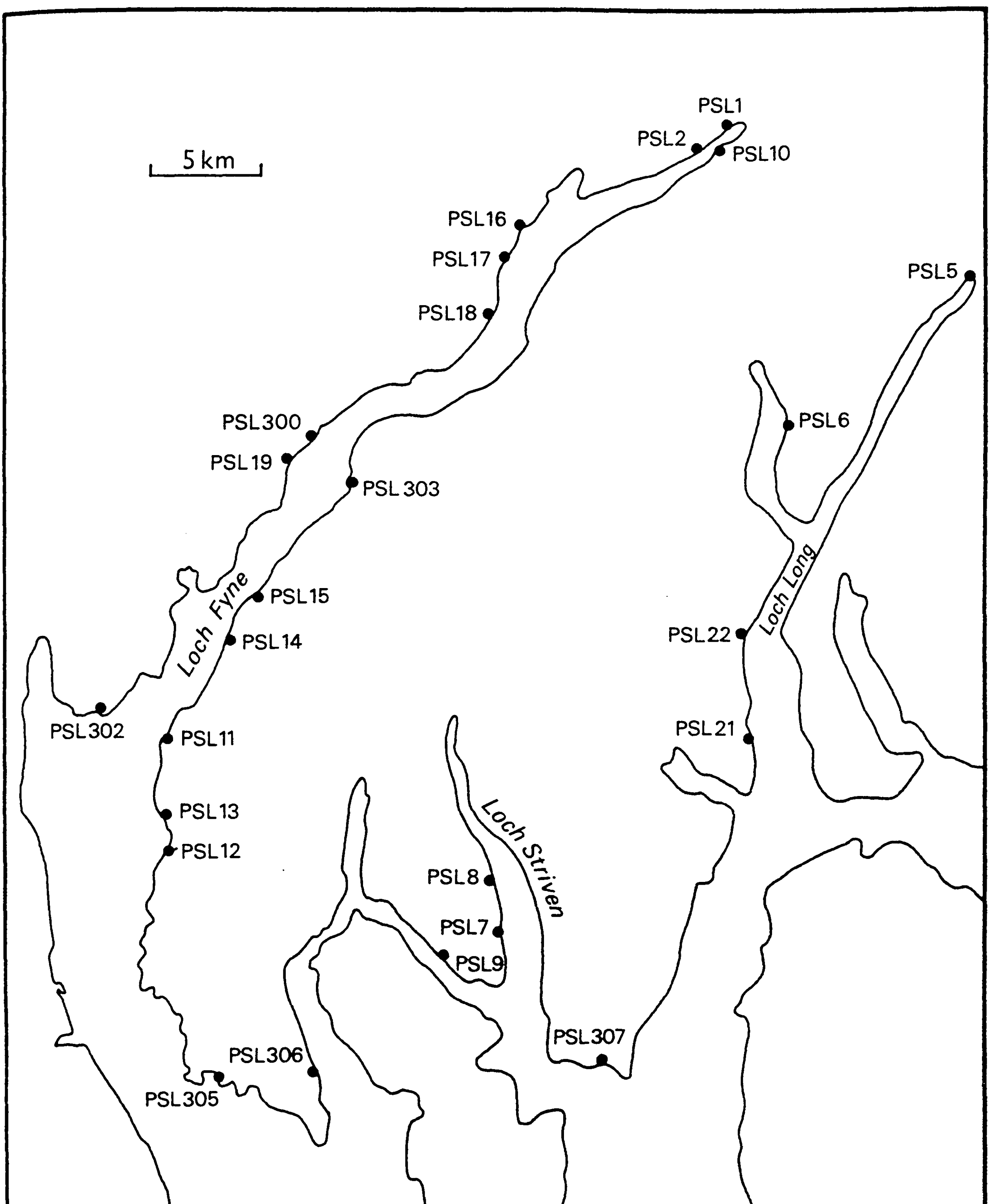
(b) Lower Limit of Land-based Vegetation

As with the upper limit of barnacle growth, the lower limit of land-based vegetation was surveyed at a number of localities around the coast, several heights 50 paces apart being measured at each locality. In total 96 heights at 25 localities (Fig. 15) were recorded. The mean altitude of the lower limit of land-based vegetation at the 25 localities was 2.36 m O.D. with a standard deviation of 0.39 m. Variability increases slightly when the 95 points are considered individually, resulting in a mean altitude of 2.37 m O.D. and a standard deviation of 0.43 m. The range of mean altitudes for the 25 localities is 1.51 to 3.26 m O.D.

During surveying there were problems of definition of the lower limit of land-based vegetation. These problems were two-fold, resulting from difficulty in deciding at what level the vegetation became closed and from the difficulty of distinguishing salt-marsh from land-based vegetation. Undoubtedly this has led to an increase in the variability of the data.

At 6 localities it was possible to compare the altitude of the lower limit of vegetation with the tide gauge records. Table 12 summarizes the data and indicates that the lower limit of vegetation approximates most closely to MHWOST near the heads of the sea-lochs (Inveraray, Arrochar and Lochgoilhead) and deviates most markedly in the most exposed locality (opposite Rothesay). Inspection of the data from all the localities supports such a general pattern, below average values occurring at the heads of the sea-lochs and above average values along the exposed

Figure 15: Location of sites where vegetation and drift limits were surveyed.



PSL1
PSL2
PSL10
5 km
PSL16
PSL17
PSL5
PSL18
PSL6
PSL300
PSL19
PSL303
Loch Long
PSL15
Loch Fyne
PSL22
PSL14
PSL302
PSL11
PSL21
Loch Striven
PSL13
PSL12
PSL8
PSL7
PSL9
PSL307
PSL306
PSL305

stretches of coast. This suggests the influence of salinity in the inner lochs and exposure on the more open parts of the coast in determining the altitude of the lower limit of land-based vegetation.

The relationship between salinity and the vegetation limit was tested by comparing, where possible, surveyed localities with Mill's (1891) salinity stations (Table 13). Nineteen points could be analysed, these indicating a positive relationship ($r = +0.2625$) though this fails to achieve significance at the 90% level. Consideration of the 10 points in Loch Fyne up-loch of Otter Ferry (where tidal regime, fetch, etc. are more uniform) improves the relationship ($r = +0.4310$) although this just fails to achieve significance at the 90% level. Although there is a clear suggestion that salinity does indeed influence the level to which vegetation can grow, there are other variables sufficiently important in affecting the height of the vegetation limit to mask the statistical association of the variables.

The relationships between fetch, the component of fetch normal to the shore, the nearshore gradient and the altitude of the vegetation limit were tested by correlation analysis. Table 14 lists the relevant data and Table 15 below summarizes the correlation coefficients.

TABLE 15. Correlation Coefficients for Vegetation Limit.

	Max. Fetch	CMF	ln CMF	$\tan \alpha$
Veg. Limit	+0.5958	+0.5677	+0.5955	-0.0161

The relationships with fetch, and the logarithm of the component of maximum fetch normal to the shore achieve significance at the 99.9% level and that with the component of maximum fetch at the 99% level. Nearshore gradient appears to have little explanatory value for the lower limit of vegetation growth.

In summary, the data indicate that the lower limit of land-based vegetation occurs at 2.36 $\pm$ 0.39 m O.D. around the coast of the study area, this being, on average, approximately 0.5 m above MHWOST. There are, however, systematic variations related to fetch and probably salinity such that the vegetation limit is highest in areas of high salinity and great fetch, and lowest in areas of low salinity and limited fetch.

(c) Upper Limit of Sea Drift

The highest altitude to which material had been washed by the sea was surveyed at 25 separate localities (Figure 15). At each locality heights were recorded at 50 pace intervals along the shore. In all, 88 separate points were heighted. The uppermost limit of sea drift was defined as the highest line where pieces of wood, seaweed, etc. had been washed together. It was important that the feature could be traced laterally and thus did not represent material thrown by waves or transported by some other agency. It therefore reflects high-tide conditions and/or constructive wave action and is thus rather different from the drift-limit of Synge and Stephens (1966) as they used the altitude of fresh drift after a period of calm sea. Although no attempt was made to survey the drift line corresponding to that of Synge and Stephens it is felt that such a line is of limited utility in the W coast of Scotland during a field season lasting several months, for not only does the tidal regime vary considerably over such a period but the calm conditions necessary for application of the definition can hardly be anticipated to last throughout the field season.

The upper drift limit is, however, formed during the winter and observations during three field seasons suggest that it is normally undisturbed by the sea during the summer months. Difficulties have been encountered during fieldwork, however, as in certain areas drifted material was sparse and a drift-line only poorly developed, whilst growth of vegetation allied to decomposition of the drift as the summer progressed often made accurate identification of the uppermost drift line difficult. It seems probable that due to these problems the data have greater variability than if it had been possible to survey this feature during a short period at the start of the summer.

The mean height of the drift limit at the 25 localities was 3.22 m O.D. with a standard deviation of 0.35 m. Considering the points individually resulted in a mean height of 3.25 m O.D. and a standard deviation of 0.37 m. The range of values at the separate localities is 1.44 m, extending from 2.57 to 4.01 m O.D. These figures demonstrate considerable intrinsic variability in the data due, for example, to tidal regime, fetch or nearshore gradient.

The relationship between tidal regime and the drift limit could be assessed at 6 tidal gauges. Table 16 lists the sites involved and shows that the drift limit varies from 0.85 m to 2.23 m above MHWOST. This is a considerable range and suggests that although the drift limit forms at high tide there is no close relationship between tidal regime and drift limit altitude.

The relationships between the drift limit and maximum fetch, the component of maximum fetch and the nearshore gradient were tested as previously by correlation analysis. Table 14 lists the relevant data and Table 17 below shows the resulting correlation coefficients.

TABLE 17. Correlation Coefficients for Drift Limit.

	Max. Fetch	CMF	ln CMF	$\tan\alpha$	ln CMF, $\tan\alpha$
Drift Limit	+0.5238	+0.5586	+0.6719	-0.2712	+0.6767

The various fetch relationships achieve significance at the 99% level and the relationship with nearshore gradient at the 95% level. There is a close similarity between these results and those recorded for shingle ridge altitudes although the various correlation coefficients are higher for the shingle ridge data, due, presumably, to easier identification of the features in the field and the more accurate determination of the orientation of the ridges. The arguments advanced in that section concerning constructive wave action and movement of material are therefore supported by the observations on the drift limit.

The multiple regression equation linking the drift limit to both fetch and nearshore gradient is

$$DL = 2.89 + 0.24 \ \ln CMF - 0.72 \ \tan\alpha$$

The base constant of the equation indicates that for minimum conditions of fetch and nearshore gradient (see section on shingle ridges) the drift limit would attain an altitude of 2.89 m O.D., approximately 1 m above MHWOST. This is very similar to the figure of 2.76 m O.D. derived from the similar multiple regression equation for shingle ridges and taken together the figures suggest that with minimum values for fetch and nearshore gradient the upper limit to which the sea can move material is slightly more than 1 m above MHWOST.

(d) Summary on Datum Planes

From the foregoing it is clear that the use of naturally occurring phenomena as substitute datum planes is frought with difficulties. In addition to a dependence on tidal regime it has been shown that the potential datum planes examined are strongly influenced by fetch, nearshore gradient or salinity. Table 18 summarizes the degree of variability in the data if none of the independent variables are taken into consideration.

TABLE 18. Datum Plane Characteristics.

Datum	Mean (localities) $\pm 1\sigma$ (m O.D.)	Mean (all points) $\pm 1\sigma$ (m O.D.)	Range (localities) m
Barnacles	1.16 $\pm$ 0.21	1.18 $\pm$ 0.23	0.81
Vegetation	2.36 $\pm$ 0.39	2.37 $\pm$ 0.43	1.75
Drift	3.22 $\pm$ 0.35	3.25 $\pm$ 0.37	1.44

The upper limit of barnacle growth is clearly the most consistent of the three, this being due in part to its easier identification in the field, but due also to its independence of influence by fetch or nearshore gradient. It is dependent upon salinity, however, and where salinity falls below 90% of the value for open sea-water the barnacle line cannot be regarded as a reliable indicator. For salinity values above 90%, however, it is remarkably consistent, having a mean of 1.20 m O.D. and a standard deviation of 0.12 m and it approximates very closely to MHWNT.

Of the other two indicators, the vegetation limit is dependent upon fetch and probably also on salinity, whilst the drift limit is dependent upon fetch and nearshore gradient. Both of these lines are more difficult to identify in the field than is the barnacle line. It is thought that neither is a reliable datum on which to base surveys of raised marine landforms as they would both introduce errors of c. $\pm$0.4 m.

TABLE 2. TIDAL REGIMES.

Tide Gauge	MHWOST	MHWNT	MLWNT	MLWOST	Spring Range	Neap Range	Mid Tide
	m O.D.	m O.D.	m O.D.	m O.D.	m	m	m O.D.
Campbeltown	1.61	1.11	-0.49	-0.99	2.6	1.6	0.31
E Loch Tarbert	1.58	1.18	-0.62	-1.12	2.7	1.8	0.25
Inveraray	1.68	1.38	-1.12	-1.42	3.1	2.5	0.13
Millport	1.78	1.18	-0.62	-1.12	2.9	1.8	0.31
Rothesay	1.78	1.28	-0.62	-1.22	3.0	1.9	0.31
Wemyss Bay	1.78	1.28	-0.52	-1.12	2.9	1.8	0.36
Greenock	1.78	1.28	-0.62	-1.22	3.0	1.9	0.31
Coulport	1.78	1.28	-0.72	-1.32	3.1	2.0	0.26
Lochgoilhead	1.68	1.18	-0.72	-1.32	3.0	1.9	0.21
Arrochar	1.78	1.28	-0.72	-1.32	3.1	2.0	0.26
Helensburgh	1.78	1.28	-0.62	-1.22	3.0	1.9	0.31
Rhu	1.78	1.18	-0.62	-1.22	3.0	1.8	0.28
Garelochhead	1.78	1.28	-0.62	-1.32	3.1	1.9	0.28

TABLE 4. PRESENT-DAY SHINGLE RIDGES AND BANKS.

Ridge	Altitude	Max. Fetch	CMF	Intertidal distance	tan α
	m O.D.	km	km	m	
SR2	3.01	3.1	2.81	21	0.1429
SR3	3.07	7.9	7.87	42	0.0714
SR4	3.60	18.7	16.95	105.5	0.0294
SR5*	3.91	12.4	5.82	84.5	0.0367
SR12	3.67	11.3	8.65	211	0.0128
SR17	3.86	12.9	12.46	53	0.0509
SR18*	3.86	6.6	5.41	63.5	0.0488
SR19	3.12	14.2	4.86	253.5	0.0122
SR302	3.37	11.1	10.72	359	0.0086
SR304	4.21	68.0	65.68	116	0.0233
SR306	3.47	56.0	9.72	158.5	0.0170
SR307	3.39	10.5	10.46	15	0.2000
SR308	2.94	10.5	10.46	15	0.2000
PSL8	2.97	3.3	1.30	16	0.1875
PSL20	4.18	96.5	96.50	53	0.0566
PSL21	3.27	13.4	5.80	158.5	0.0196
PSL22	2.94	8.0	6.93	68.5	0.0453
PSL304	4.11	70.8	69.72	74	0.0365
PSL307	4.01	77.2	77.20	127	0.0236

* See text

TABLE 6. PRESENT-DAY INTERTIDAL FLATS.

Profile	Break of slope m O.D.	Max. Fetch km	CMF km	tan α
3	1.6	5.8	5.6	0.0256
2	1.7	10.1	5.05	0.0167
6	1.4	10.5	10.46	0.2000
7	1.5	77.2	77.2	0.0203
8	1.8	85.0	29.07	0.0289
1	1.8	10.1	5.79	0.0299
5	1.6	20.8	19.55	0.0406
4	1.3	20.3	15.55	0.0178
9	1.6	13.4	5.80	0.0196
10	1.7	8.0	6.93	0.0453

TABLE 8. PRESENT-DAY MARINE EROSION FEATURES.

Locality	Altitude m O.D.	Max. Fetch km	CMF km	tan α
RP1	2.19	13.45	3.48	0.1468
RP2	1.75	12.80	6.40	0.1174
RP3	1.86	12.50	10.24	0.1468
RP11	1.85	13.00	5.49	0.0587
RP12	2.25	4.50	2.20	0.0979
RP13	1.93	3.70	2.20	0.0839
RP14	2.02	3.10	2.26	0.1174
RP15	1.51	96.50	90.68	0.0379
RP303	1.79	9.10	3.85	0.0280

TABLE 11. BARNACLE LINE ALTITUDE AGAINST SALINITY.

Locality	Number	Altitude m O.D.	Salinity %
Cuill	PSL10	0.60	56.8
Dunderawe	PSL3	1.22	70.1
	PSL4	0.56	70.1
Inveraray	PSL16	1.06	83.2
	PSL17	1.21	83.2
Strachur	PSL18	1.29	85.7
Furnace	PSL19	1.33	91.0
	PSL303	1.29	91.0
Gortans	PSL15	1.28	92.7
	PSL14	1.31	92.7
Otter	PSL11	1.12	93.0
	PSL302	1.01	93.0
Skate	PSL305	1.20	94.0
Ardmarnock	PSL306	1.25	92.4
Burnt Islands	PSL9	1.03	91.3
Coulport	PSL21	1.28	84.7
	PSL22	1.31	84.7

TABLE 12. VEGETATION LINE AND TIDAL REGIME.

Tide Gauge	Locality	MHWOST m O.D.	Veg.Alt. m O.D.	diff. m
Inveraray	PSL16	1.68	1.89	+0.21
	PSL17	1.68	2.25	+0.57
E Loch Tarbert	PSL305	1.58	2.19	+0.61
Rothesay	PSL307	1.78	3.26	+1.48
Coulport	PSL22	1.78	2.46	+0.68
Lochgoilhead	PSL6	1.68	2.07	+0.39
Arrochar	PSL5	1.78	1.66	-0.12

TABLE 13. VEGETATION LINE AGAINST SALINITY.

Locality	Number	Altitude m O.D.	Salinity %
Cuill	PSL1	2.13	56.8
	PSL2	2.26	56.8
	PSL10	2.14	58.8
Inveraray	PSL16	1.89	83.2
	PSL17	2.25	83.2
Strachur	PSL18	2.47	85.7
Furnace	PSL19	2.41	91.0
	PSL303	2.79	91.0
Gortans	PSL14	2.53	92.7
	PSL15	2.15	92.7
Otter	PSL11	3.00	93.0
	PSL302	1.51	93.0
Skate	PSL305	2.19	94.0
Ardmarnock	PSL306	2.49	92.4
Burnt Islands	PSL9	2.63	91.3
Coulport	PSL21	2.83	84.7
	PSL22	2.46	84.7
Stuckbeg	PSL6	2.07	86.0
Arrochar	PSL5	1.66	75.5

TABLE 14. DRIFT AND VEGETATION LINES AROUND PRESENT COAST.

Locality	Drift Alt. m O.D.	Veg. Alt. m O.D.	Max. Fetch km	CMF km	tan α
PSL1	3.00	2.13	4.88	2.06	0.0618
PSL2	3.48	2.26	11.78	4.03	0.0979
PSL5	2.63	1.66	1.77	0.63	0.0294
PSL6	2.94	2.07	3.67	2.75	0.0947
PSL7	3.21	2.36	8.55	2.93	0.1200
PSL8	3.00	2.58	3.30	1.30	0.1875
PSL9	3.17	2.63	21.30	1.86	0.0631
PSL10	3.60	2.14	12.40	5.82	0.0367
PSL11	3.93	3.00	11.09	9.08	0.0030
PSL12	3.26	1.98	11.25	9.22	0.0913
PSL13	3.36	2.56	12.70	12.27	0.0128
PSL14	3.35	2.53	13.00	3.50	0.0839
PSL15	2.59	2.15	14.75	2.14	0.0452
PSL16	2.57	1.89	8.49	3.45	0.0489
PSL17	3.04	2.25	12.90	5.49	0.0587
PSL18	3.17	2.47	9.50	7.78	0.0652
PSL19	3.09	2.41	12.50	7.16	0.0489
PSL21	3.27	2.83	13.40	5.80	0.0226
PSL22	3.08	2.46	8.00	6.93	0.0489
PSL300	3.20	2.31	10.75	10.10	0.0734
PSL302	3.40	1.51	7.20	3.38	0.0073
PSL303	3.42	2.79	22.13	20.79	0.0734
PSL305	3.38	2.19	5.70	4.47	0.0320
PSL306	3.42	2.49	6.02	3.45	0.0166
PSL307	4.01	3.26	77.20	77.20	0.0237

TABLE 16. DRIFT LIMIT AND TIDAL REGIME.

Tide Gauge	Locality	MHWOST m O.D.	Drift Alt. m O.D.	diff. m
Inveraray	PSL16	1.68	2.57	+0.89
	PSL17	1.68	3.04	+1.36
E Loch Tarbert	PSL305	1.58	3.38	+1.80
Rothesay	PSL307	1.78	4.01	+2.23
Coulport	PSL22	1.78	3.08	+1.30
Lochgoilhead	PSL6	1.68	2.94	+1.26
Arrochar	PSL5	1.78	2.63	+0.85

CHAPTER 4

RAISED SHORELINE METHODOLOGY

1. Introduction

Around the coastline of the study area there is only one raised shoreline that can be traced continuously for many kilometres. This is the erosional Main Rock Platform. The depositional raised marine features occur as discrete landforms in localities where the original conditions allowed their formation and where subsequent erosion has not destroyed the original form. In order to establish former shorelines around the coast it becomes necessary to correlate these separate landforms across the intervening areas in which there is no evidence. This situation contrasts with that of the Forth and Tay areas on the E coast of Scotland where much of the methodology employed in this study originally evolved (Sissons, 1963a; Sissons, Smith and Cullingford, 1966). In the E coast areas raised marine landforms are frequently traceable in the field for many kilometres and could be surveyed along lengthy portions at a time. Gullies and alluvial fans impeded the surveying of the features but not their evident continuity in the field and the surveyed portions could properly be regarded as fragments sampled from the continuous trace of the shoreline. On the W coast, however, the features surveyed are individual landforms (e.g. deltas or shingle ridges) and not fragmented portions of larger landforms. The problem posed in correlating these landforms is correspondingly greater than on the E coast and is not alleviated by the occurrence of stratigraphic markers such as pumice, which has been used in Norway, Spitsbergen and Arctic Canada (e.g. Binns, 1972; Marthinussen, 1960; Blake, 1970), or characteristic marine fossils (e.g. the Tapes shorelines in Norway, Marthinussen, 1960). Nor are a sufficient number of radiocarbon dates available for use in correlating various separate landforms.

This Chapter is divided into three main sections. The various methods that have been used in correlating raised marine landforms are discussed in the last section. First, however, it is necessary to review the method by which an altitude or range of altitudes can be established on a particular raised marine landform. Errors inherent in the method place limits on the confidence that can be attached to particular correlations and these are quantified in the second section of the Chapter.

2. Altitude Measurement

The purpose of surveying raised marine landforms is to produce altitudes (or ranges of altitudes)for the landforms that are comparable with each other and that can be related to the tidal regime of the time of formation of the feature. The altitudes obtained must be accurate enough to allow differentiation of altitudinally close but morphologically distinct features and the frequency of points surveyed must be sufficient to discriminate between more steeply sloping features such as outwash spreads and the rather lower gradients typical of marine landforms (Sissons, 1967a; pp. 167-9). The measurement technique must give results that are reproducible and errors in the method must all be amenable to quantitative assessment.

From the above it is clear that a number of decisions must be made as regards the field surveying of marine landforms. Accuracy demands that it is the marine landform and not overlying peat, slumped debris, etc., that is surveyed. Precision requires unambiguous definition of the part of the landform to be measured. Relationship to tidal regime can only be established through analogy with present-day marine landforms and the ability to survey analogous parts of the fossil and present-day features. The distance between measured points must be related through the accuracy of the surveying method to the gradient of the feature as well as to the size of the feature. The ability to separate altitudinally close features necessitates the use of an instrument of high accuracy and the traverses must all relate to a strict datum and be closed in order that errors may be assessed.

Unambiguous locations for establishing the heights of certain marine landforms are relatively easy to define. The crests of shingle ridges or the junction of cliff and platform on wave erosional forms are generally simple to locate. With other depositional marine landforms the matter is more complicated. The methodology adopted for this study involved mapping all flat or apparently-flat features and this has the immediate advantage of ruling out the rather ambiguous shingle banks that occur widely around the coasts of the study area. Contemporary depositional flats were identified (Chapter 3) and, by surveying detailed cross-profiles at right angles to the present coast, were found to have distinct breaks of slope at their landward margin. In surveying the fossil flats it was therefore necessary to locate the back edges of the raised flats. Ideally this should have been done by surveying cross-profiles (Tammekann, 1952) but these are very time consuming; therefore the location of breaks of slope was assessed by visual judgement. Certain cross-profiles of the Flandrian shorelines were surveyed (Chapter 10) but these were chiefly for illustrative purposes.

In identifying the position of the break of slope it is necessary to avoid slumped material, alluvial fans, etc.. In the E of Scotland Sissons (1963a) measured the estuarine flats some 20-80m out from the back edge to avoid slumped debris but this practice could not be adopted in this study as many of the fossil landforms are too small. In an attempt to assess the variation likely to result from the inconsistencies in personal judgement of the position of the break of slope, 32 features were re-surveyed more than a year after their initial survey. These results are discussed in detail in the next section.

In general, the possible error involved in selecting only one point on which to measure a particular landform becomes greater the larger the landform. A number of altitudes is less likely to be subject to error and if the points are all selected on the same basis and all have an equal probability of representing the true altitude of the landform (which is true of raised marine landforms where the length is short) then the mean of these altitudes should be the best estimate of the altitude of the landform. Where the feature is particularly extensive then a series of altitudes may provide valuable information about the gradient of that feature. Sissons (1963a)and Sissons and Smith (1965), for example, demonstrated the value of closely spaced heights in differentiating between outwash terraces and associated raised marine deposits. In this study, therefore, a minimum of two heights were surveyed on the raised marine landforms (with the exception of a small number of 'poor' fragments (see below) that were not used in subsequent analysis (Chapters 5, 6 and 7) and where the size of the landform permitted, heights were recorded at 50 pace (ca. 50 m) intervals.

The closely spaced heights on the cross-profiles shown in Chapter 10 illustrate the small height differences that can occur between distinct marine landforms. A similar situation prompted Sissons (1967a, p 168) to use an accurate levelling instrument rather than an aneroid barometer to survey the fossil landforms. This practice was followed in this study and the errors inherent in the method are discussed in the next section.

Accurate levelling has the further advantage that it enables all surveyed points to be tied to O.D. and altitudes are therefore comparable not only locally but country-wide.

Although it might seem self-evident that altitudes must be recorded on the landform being studied, this has not always been true of studies of raised shorelines on the W coast of Scotland: Gray (1972a)documented examples where previous work had failed to take account of overlying peat. Where peat was encountered in the present study it was penetrated by aluminium rods and the form and approximate composition of the sub-peat surface was tested. Only after this were heights recorded where appropriate.

3. Errors in Altitude Measurement

There are numerous sources of errors in the methods adopted to collect the altitude data. It is necessary to assess these errors and, if possible, to quantify them. Previous literature on raised shoreline studies has only approached this problem in a qualitative or semi-quantitative manner. For example, Andrews (1970, pp. 26-27) considered that elevation errors could result from (a) interpretation errors, (b) the relation of different features to sea-level, and (c) measurement errors, but he did not attempt to set limits on any of these. His attitude to measurement errors is typical of much of the literature:

> 'little can be said about measuring error other than it is very small for surveys by level and is present, but highly variable, in aneroid surveys'.

Gray (1975a) in reviewing many of the methods used in the present study did not discuss some of the errors inherent in the approach and failed to place an overall figure covering the degree of confidence with which altitudes could be treated. The matter is important in a number of ways.

Firstly, it is necessary to decide whether the degree of variability inherent in the data is sufficiently low to allow the correlations that are made with it. Secondly, the variability of the altitude data indicates the distance over which certain shorelines must be established before their gradient can be accurately known. Thus, for example, if there is an inherent uncertainty in the data of $\pm$1.0 m and a shoreline with a true gradient of 0.05 m/km is being studied, then this shoreline must be traced for more than 40 km along the line of minimum slope before the gradient can be confidently established. Thirdly, knowledge of where errors accumulate in the methods adopted points to where new techniques should be applied or more information sought to most effectively reduce the total error.

An attempt was made during this study to identify and quantify the errors inherent in the field techniques adopted. The following sources of error are discussed below:

(a) errors in datum used
(b) errors in instrumental survey
(c) errors in observer's judgement
(d) errors inherent in geomorphic feature
(e) errors due to relationship to tidal cycle

(a) Datum Errors

In this study Ordnance Datum (Newlyn) has been used. The accuracy of O.D. bench mark heights varies according to the type of survey and the distance between the bench marks. Harley (1975) gave the error factors for Geodetic, Secondary and Tertiary bench marks as $\pm$ 2.F mm, $\pm$ 5.F mm and $\pm$ 12.F mm respectively, where F is the square root of the distance in kilometres between bench marks. Bench marks of all three types were used in the present study and the distances of the original surveying between the various bench marks were not known. In general bench marks are placed approximately every 2 miles (3.25 km) along traverse lines suggesting that maximum errors are likely to vary from <u>ca</u>. ±4 mm for Geodetic bench marks to <u>ca</u>. ±20 mm for Tertiary bench marks. The overall average error will be considerably less than these maxima and a figure of ± 10 mm (± 0.01 m) has been adopted for this study. This is in line with Ordnance Survey practice of quoting bench mark altitudes to the nearest centimetre.

An attempt has also been made, based on Ordnance Datum, to assess the accuracy of other phenomena that have been used in other studies as substitute datum planes (Chapter 3). The upper limit of sea-drift, the lower limit of land-based vegetation and the upper limit of barnacle growth were all found to have considerable variation, being assigned standard deviations of ±0.35 m, ±0.39 m and ± 0.21 m respectively. The most consistent of these measures was the upper limit of barnacle growth which was demonstrated, however, to be dependent upon salinity as well as tidal regime and hence is unreliable towards the heads of the sea lochs. In those areas where salinity was above 90% the variability of the barnacle line was reduced to ±0.12 m. The altitude of the upper limit of barnacle growth in areas of salinity above 90% closely follows the MHWNT and this must be taken into account if regional correlations are to be based on this datum.

(b) Errors in Instrumental Survey

Throughout this study surveying was carried out using a Hilger and Watts Autoset level. All traverses were closed to bench marks and errors distributed throughout the traverse, the total error being divided by the number of instrument stations and the error being added incrementally from first station to last. Where errors of 0.5 m or over were encountered it was assumed that this was due to misreading and the traverse was re-surveyed. Throughout the surveying readings were recorded to the nearest millimetre and corrected accordingly. Even where there was zero correction on a traverse, however, the altitudes are not known to a greater accuracy than are the bench marks on which the surveys are based, that is to the nearest centimetre. All altitudes in the text are therefore given to the nearest centimetre.

In order to assess the overall accuracy of the points surveyed the numerical values of the corrections applied to the first 1,000 points were summed and the mean correction for these 1,000 points calculated. This gave a figure of 0.01 m. Since the correction can be either negative or positive the instrumental survey is assigned an accuracy of ± 0.01 m.

Comparative figures are difficult to find for the other methods often used for heighting raised shoreline features. The aneroid barometer (used, for example, by Donner, 1959, 1963; King and Wheeler, 1963) has been documented best by Sparks (1953). Errors arise in the use of this instrument due chiefly to weather changes although the instrument also needs time to respond to the changes in altitude. Sparks considered that an accuracy of ± 1.5 m could be attained on 33% of the days in north-west Scotland. Few authors have claimed more accurate results. In order to minimise the effects of changing weather, Donner (1959) took less than 7 minutes between readings at his sites, implying, for some of the sites at least, a degree of athletic achievement rare even in a field scientist.

There is considerable disagreement in the literature concerning the accuracy of the Abney level, in part because there are several methods of using the instrument and also because few detailed studies have been attempted to assess its accuracy. Synge and Stephens (1967) regarded their use of the Abney level to result in an accuracy of 0.5% but a more detailed study by Baillie (1972), resting the level on a rod and reading a calibrated staff, gave a standard deviation for his results of ±1.6% (equivalent to ±0.64 m in 40 m). Given that no field usage has been as rigorous as that of Baillie, it is likely that errors using this technique are considerably greater than ±1.6%.

(c) Observational Errors

There are two distinct types of error in this category, namely gross errors or misinterpretation of evidence, and systematic errors or bias in the observer's judgement.

Gross errors invalidate the rest of the survey and clearly their degree of accuracy cannot be assessed. They can best be avoided by a systematic approach to fieldwork and in the critical case of differentiating river terraces and raised marine landforms closely spaced heights along the length of the feature in question often prove decisive in the correct identification of the landform. For example, Gunn et al. (1897) mapped almost all the terraces near Ardyne Point as raised beaches whilst Figure 32 clearly shows them to have steep seaward gradients over considerable distances, indicating their origin as river terraces.

Observational bias is considerably more difficult to assess but is critical to the subsequent comparison of landform altitudes. The most crucial judgement during the surveying is the location of the break of slope at the back edge of a terrace surface. If an observer's judgement of this varies with time then his results may not be reproducible at a site and would not be comparable along a stretch of coast. In order to test the degree of such variability in the present study 32 raised marine landforms were re-surveyed at least one year after their original survey.

During re-surveying no attempt was made to position the staff where it had previously been placed but the landforms were regarded as if a new survey were taking place and the staff positioned accordingly. One result of this was that during the first survey 76 individual altitudes were recorded whilst during the re-survey only 72 points were judged appropriate. Of the features re-surveyed, 3 were shingle ridges and 29 were marine flats of which 2 were considered 'good', 17 'moderate' and 10 'poor' on the classification system elaborated in Chapter 3. The features ranged in altitude from below 5 m O.D. to over 40 m O.D. and hence covered the complete range of altitude in which raised marine landforms have been discovered in this area. Table 19 lists the features, their grading and the results of the two surveys.

In analysing the data to decide whether the results of the second survey are significantly different from those of the first survey it is the difference between the pairs of observations that is most important. For the results to be acceptable the mean difference between the first and the second series of observations should be zero. The calculated mean difference was -0.01 m and the standard error of the mean was $\pm$0.06 m, indicating a 98.7% probability that the observed mean difference was not different from zero and hence that errors of judgement are negligible.

(d) Errors Inherent in Landform

Although the landforms mapped in the field were judged to be flat or to have a continuous near-level ridge in the case of shingle ridges, there is inevitably a considerable amount of variability in the surface form of such features. This variability can be estimated by considering the deviations of the individual points levelled on a feature from the mean altitude of the feature. This was done using the data derived from the re-surveying exercise together with that from the first survey for each of these landforms had approximately twice as many measurements as normal, thus allowing a better estimate of the mean altitudes as well as a better estimate of the inherent variability of the height of the features. The deviations from the respective means were squared, summed and divided by 148 (the total number of individual observations) and the square root of this result calculated, thus producing a combined standard deviation. This figure was ± 0.301 m. This is considered to be the best estimate of the uncertainty inherent in the landforms.

(e) Errors in Relationship to Tidal Cycle

The relationship of a landform to the tidal cycle must be known if relative sea-level change at a locality is to be understood or if correlations between landforms are to be made along a former coastline. The uncertainty in the relationship is not an inherent property of the measurement of the landform as have been the foregoing errors. The variability of different types of marine landforms around the present coast are considered in Chapter 3. There it is shown that intertidal deltas, shingle ridges and marine erosion features varied (1σ) around the coast of the study area by ± 0.16 m, ± 0.46 m and ± 0.23 m respectively. The greater variability of the shingle ridges and marine erosion features was shown to be due to their dependence upon fetch and nearshore gradient, intertidal deltas apparently being independent of these variables. If fetch and nearshore gradient could be calculated for raised shingle ridges or marine erosion features then the relationships established for the present-day features could be used to adjust the altitudes of the abandoned landforms and hence increase the confidence with which they could be correlated.

(f) Summary of Errors

The errors assessed above can be summarized as follows:

Datum errors	$\pm$0.01 m
Instrumental errors	$\pm$0.011 m
Measurement errors	$\pm$0.068 m
Landform error	$\pm$0.301 m
Tidal regime (deltas)	$\pm$0.16 m
" " (ridges)	$\pm$0.46 m
" " (erosion)	$\pm$0.23 m

The first four errors relate to the altitudes assigned to the individual landforms. The category of all measurement errors includes that of instrumental errors and these should not be summed, but the datum, measurement and landform errors can be summed to indicate the degree of uncertainty in the heights of the landforms surveyed. The figure is $\pm$0.379 m. The errors relating to the relationship between tidal regime and marine landform are relevant to assessing relative local sea-level change. They are additive to the uncertainty in the surveying of the landform. Thus if intertidal deltas are used to infer relative sea-level change the overall uncertainty is $\pm$0.54 m, for shingle ridges it is $\pm$0.84 m and for marine erosion features $\pm$0.61 m. These figures are applicable only to this study. The measurement errors, for example, are the result of the application of a set of operations and judgements during the course of the field work and may be quite different for another field worker.

The largest single component is the uncertainty inherent in the landforms and this will vary with the type of landform and degree of preservation. The carse landforms of eastern Scotland, for instance, would probably yield a lower figure than the deltas and shingle ridges of the W coast. The data from the re-surveyed features can be used to illustrate this if only Flandrian shoreline features are considered.

These have all formed during the last 7,000 years (Chapter 10) and can be expected to be better preserved than the Lateglacial shoreline features formed between 11,500 and 13,000 years ago (Chapter 9). The degree of variability for the Flandrian landforms alone is ± 0.24 m as compared to ± 0.30 m for all the marine landforms, thus supporting the suggestion that they are better preserved.

The above assessment of errors indicates that more accurate results will most immediately be achieved by closer attention to the variability in the form of the deltas, shingle ridges and platforms. The second most important source of error in inferring sea-level changes from morphological data is the relationship to the tidal cycle. Intertidal delta altitudes are most closely related to the tidal cycle and the greater variability associated with shingle ridge altitudes and erosion platform altitudes is due to their dependence primarily on fetch and secondarily on nearshore gradient. It is reasonable to expect that fetch could be measured for fossil shorelines and thus a correction factor derived from the regression equation relating fetch to shingle ridge altitudes for the present shoreline. Measurement of nearshore gradient for fossil shorelines seems much less certain for the tidal range may well have changed and the land surface itself is inevitably modified by marine processes as sea-level falls.

Despite the fact that certain of the errors assessed are applicable only to this study, the information gained on the variability of datum planes, of marine landforms and of the relationship of the landforms to the tidal cycle plus other independent assessments of the accuracy of certain surveying instruments allows the calculation of the errors associated with previous studies of sea-level changes in the same area.

Donner (1959), for example, used the barnacle line for a datum, measured altitudes chiefly with an aneroid barometer and used Geological Survey 1-inch maps as a source for the nature of the landform. This last practice would undoubtedly result in certain gross errors but the overall accuracy of Donner's heights may be put at ± 0.21 m (datum), ± 1.5 m (aneroid), ± 0.30 m (landform) and ± 0.16 m (tidal regime) giving a total figure of ± 2.17 m. It is slightly more difficult to assess the errors inherent in the work of Synge and Stephens (1966). They used a drift limit as a datum, though not the upper drift limit surveyed during this study, but the figures derived during this study make it unlikely that this datum is more accurate than ± 0.3 m. The use of the Abney level results in errors that are cumulative with the total height difference surveyed and the results of Baillie(1972) suggest that uncertainties of ± 0.1 m for features at 5 m O.D. and ± 0.8 m for features at 40 m O.D. would be very conservative. These figures together with established uncertainties associated with landform and tidal cycle indicate a minimum estimate of total error in Synge and Stephen's data of between ± 0.9 m and ± 1.9 m. As Synge and Stephens (1966) did not use a rod or staff with the Abney level, did not close their traverses and only surveyed one point on each feature their actual errors are likely to be much greater than these figures.

4. <u>Methods of Correlation</u>

(a) Height-Distance Diagrams

Height-distance diagrams are graphical representations of landform altitudes in which the height of a particular feature (the y-axis) is plotted with respect to the distance of the feature from a common origin projected at right angles into a vertical plane running through the origin.

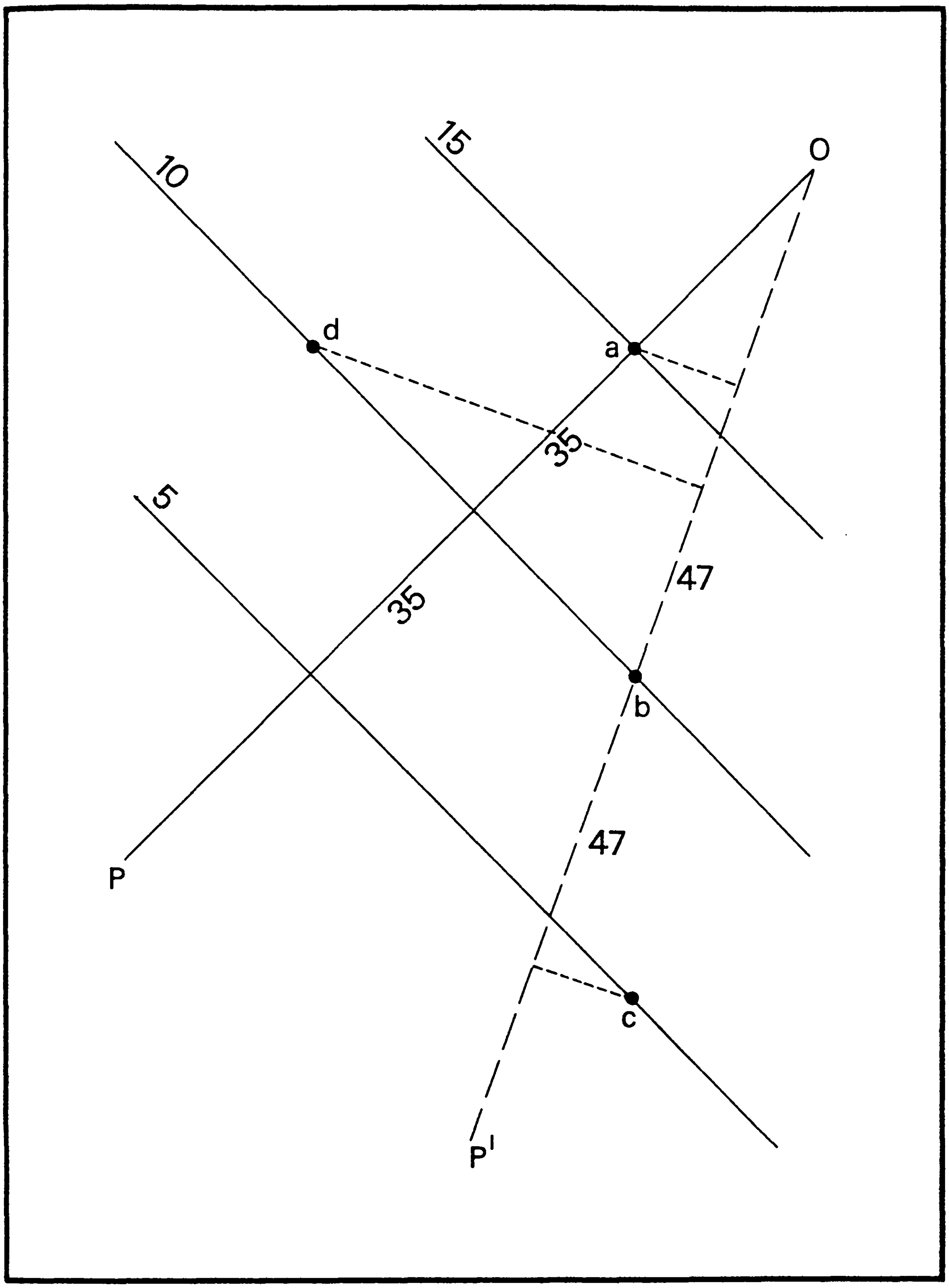
The height-distance diagram is fixed in space (by specifying the origin and the orientation of the projection plane) and makes no use in its construction of assumptions on the manner in which the landforms being studied relate to one another. It is a simple graph and hence contrasts with the shoreline relation diagram (see below) which is dependent for its construction on particular relationships between shorelines. Height-distance diagrams are not only used in shoreline studies but are also used widely (as in this study) for the analysis of river terraces (Kirby, 1969; Rhind, 1969; Gray and Sutherland, 1977). It is, however, in raised shoreline studies that these diagrams have been used most widely (e.g. Marthinussen, 1960; Sissons and Smith, 1965; Andrews, 1966; Morner, 1969) and to greatest analytical purpose. As an analytical tool the height-distance diagram is used to establish the plane in which the raised shoreline in question has its minimum gradient, that is, to establish the direction in which the centre of isostatic uplift is located. It is also used as a means of correlating shoreline fragments. Height-distance diagrams are used in both ways in this study and the limitations of this type of analysis are discussed below.

The optimum plane of projection for a particular raised shoreline is the plane that relates in an unambiguous fashion to the isostatically deformed shoreline, that is the plane that locally cuts the isobases at 90° and hence is directed towards the centre of uplift. This is the plane on which the shoreline projects with a minimum gradient and it is this fact that is utilised in identifying the optimum plane. This is achieved by rotating the projection plane by 5° at a time and projecting the various shoreline fragments into the plane and calculating their distance from the common origin. The gradient of the shoreline is calculated for each orientation of the projection plane and the various gradients compared to establish the minimum one. Regression analysis is often used as an objective means of calculating the gradient (e.g. Sissons, 1963; Cullingford and Smith, 1966) and in theory this allows a second criterion to be utilised in the identification of the optimum plane, that is the plane in which the variance of the data points about the regression line (as measured by the correlation coefficient or the standard error) is minimized (cf. Andrews, 1967). The increase in variance about a regression line that does not lie in the optimum plane is due to the horizontal displacement of points as the projection plane is rotated away from the optimum. This horizontal displacement results in an increasing vertical difference between the point and the ideal trace of the shoreline and hence an increase in the sum of the squares of deviations from the regression line, that is, the variance.

The correlation coefficient or the standard error must be used cautiously in deciding on the optimum plane (Andrews, 1967) for the non-uniform distribution of data points and the possibility of autocorrelation bias these statistics as does the small sample size that is often available for a particular shoreline. The irregular distribution of data points may also bias the calculation of the gradient of the shoreline, an effect not previously mentioned in the literature. The following example illustrates this point. In Figure 16 OP is the optimum projection plane and surveyed points 'a', 'b', 'c' and 'd' fall upon the isobases shown. If point 'd', however, had not been surveyed projection planes to the right of OP, such as plane OP^1, produce shoreline gradients with lower values than on OP. The inclusion of 'd' indicates OP^1 to be wrong but the example, although simple, illustrates the need for the projection plane to bisect the data points. This is not always possible when analysing a series of

Figure 16: Relation of plane of projection to distribution of data points for construction of height-distance diagram.

OP is optimum projection plane for isobases shown.



10
15
O
d
a
35
5
47
35
b
47
P
c
P'

shorelines utilising data from a large area and, as is seen in Chapters 9 and 10 this effect limits the utility of the shoreline gradient in deciding on the optimum projection plane. In practice consideration is given to the values of gradients and correlation coefficients from a series of shorelines, particular weight being accorded to especially well-developed shorelines with a wide distribution of data points.

A major problem to be considered in the use of height-distance diagrams is the distortion produced by curvature of the isobases, projected points being normally biased towards the centre of uplift (Figure 17). It is clear that the degree of displacement (d, Figure 17) is a function of the radius of curvature of the isobases (r) and the distance from the projection plane (x). These variables are linked mathematically by the formula

$$d = r - \sqrt{r^2 - x^2}$$

and Figure 18 graphs the amount of displacement for particular values of x and r. The degree of error introduced into the diagram due to the horizontal displacement is a function of the gradient of the shoreline. If an accuracy of ± 0.5 m is desirable for the shoreline analysis, then Table 20 lists the maximum horizontal displacements corresponding to various shoreline gradients that are acceptable for the analysis.

TABLE 20 RELATIONSHIP BETWEEN GRADIENT AND HORIZONTAL DISPLACEMENT.

Gradient (m/km)	1.0	0.5	0.25	0.1	0.05
Displacement (km)	0.5	1	2	5	10

Figure 18 in conjunction with Table 19 can then be used to assess the width of area on either side of the projection plane from which data may be safely used in analysis of a particular shoreline. Thus, for example, for a shoreline with a gradient of 0.1 m/km and with isobases whose radius of curvature is <u>ca</u>. 25 km. data from a zone of 15 km on either

Figure 17: Distortion on height-distance diagram due to isobase curvature. Displacement d of point on isobase of radius of curvature r located distance x from the projection plane.

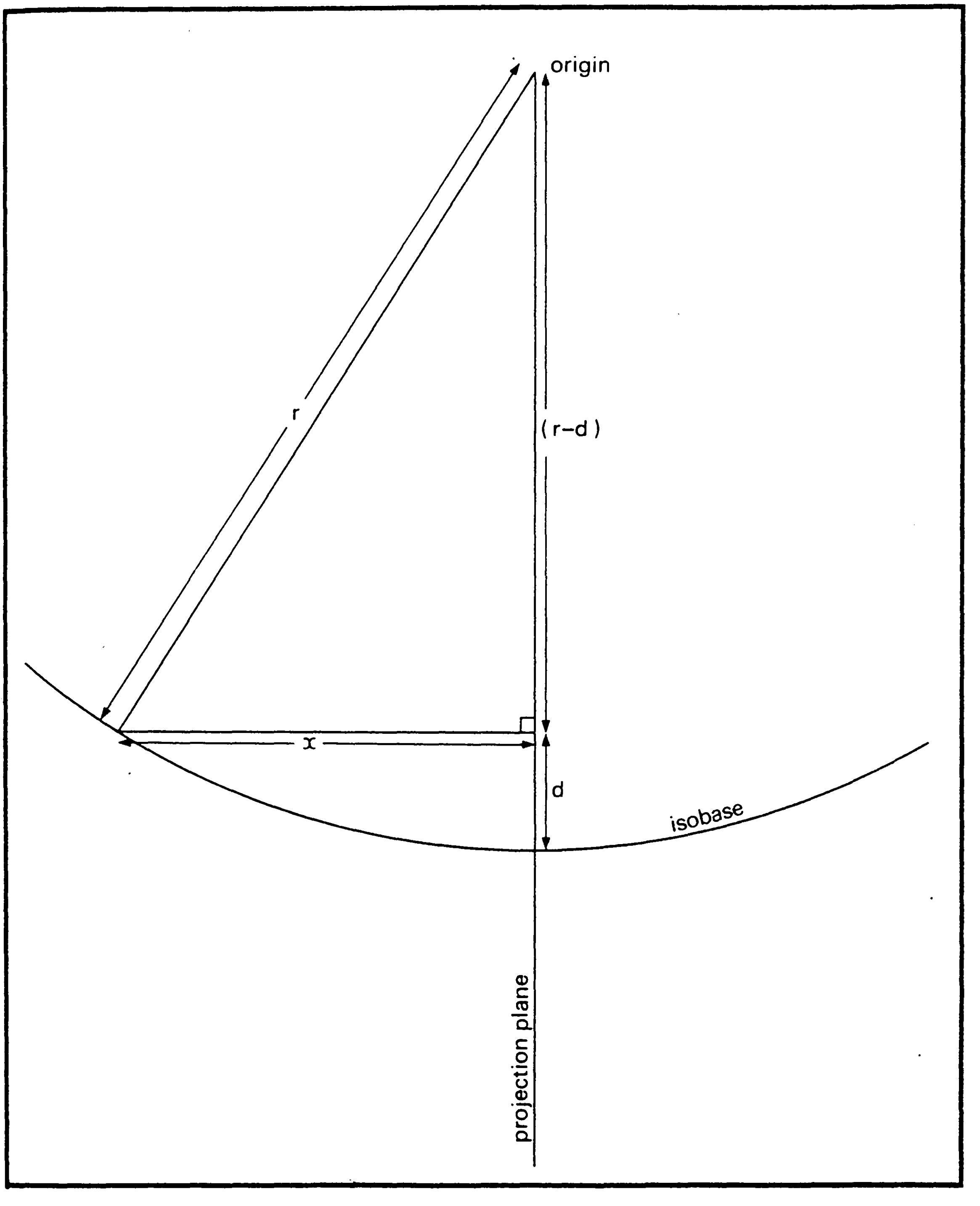
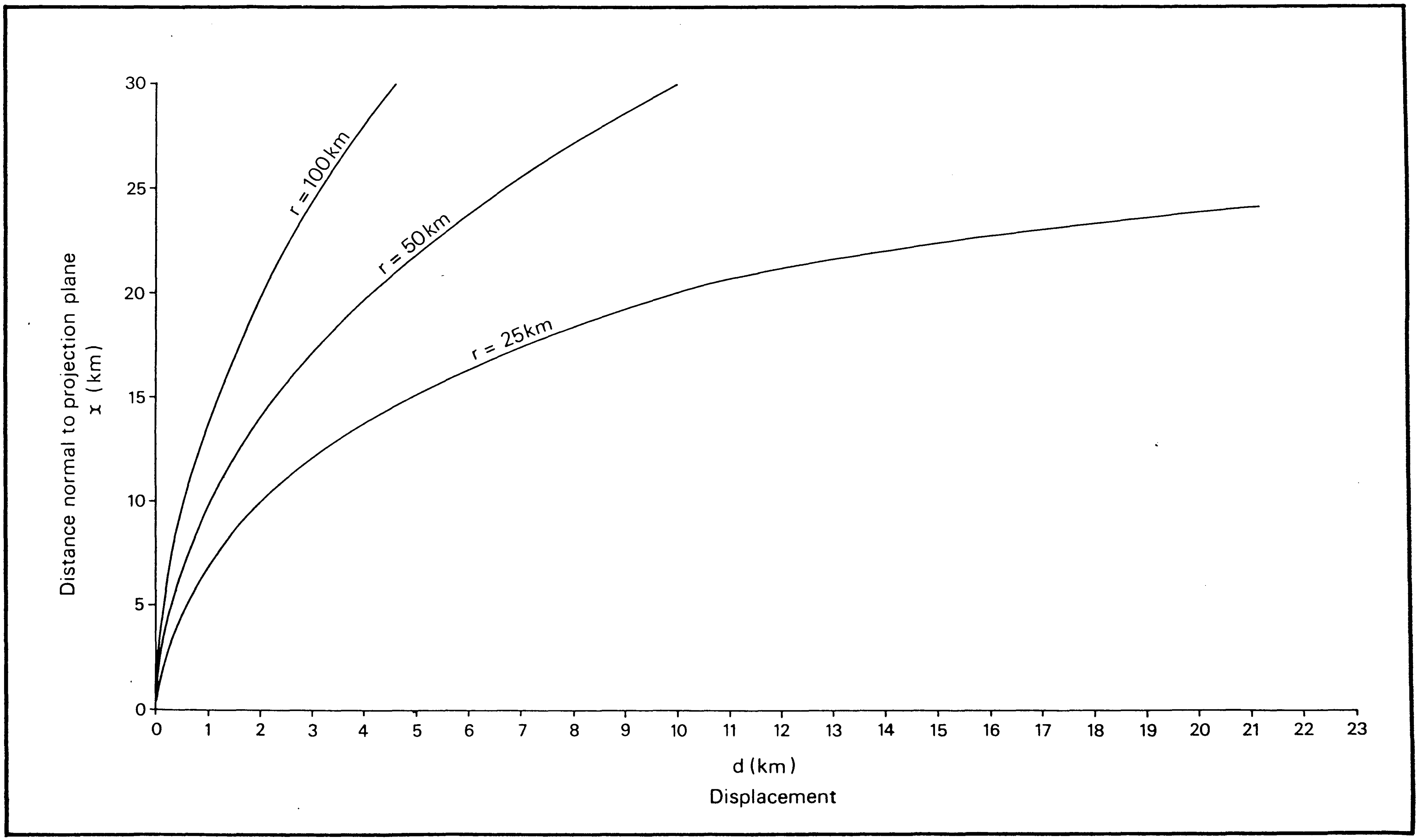

origin
r
(r–d)
x
d
isobase
projection plane

Figure 18: Graph of displacement on height-distance diagram due to isobase curvature. d(km) is displacement on diagram of a point projected from x(km) distance normal to the plane given the radius of isobase curvature r(km).



Distance normal to projection plane
x (km)
30
25
20
15
10
5
0
0
1
2
3
4
5
6
7
8
9
10
11
12
13
14
15
16
17
18
19
20
21
22
23
r = 100 km
r = 50 km
r = 25 km
d (km)
Displacement

side of the projection plane can be used. These figures, in fact, correspond approximately to the data for the Main Rock Platform analysed by height-distance diagram in Chapter 8.

The radius of curvature of the isobases in general decreases towards the centre of uplift with the consequence that, with a uniformly wide zone on either side of the projection plane from which data are being collected, the magnitude of the displacement due to isobase curvature increases towards the centre of uplift. Thus a height-distance diagram tends to underestimate somewhat the gradient of a shoreline, though the magnitude of this effect is likely to be small. Fortuitously, because of the configuration of the sea lochs, the study area tends to narrow towards the centre of uplift, thus minimizing this effect. In similar fashion, the shorelines with steepest gradients (Chapter 10) and hence most likely to be affected by isobase curvature are located towards the south of the study area where isobase curvature is at a minimum.

A further problem in using height-distance diagrams to establish the gradient of a particular shoreline is the projection distance along which the shoreline must be traced to allow the gradient to be accurately calculated. This distance is a function of the shoreline gradient and the accuracy of the measurements of the individual marine landforms being used to reconstruct the shoreline. Thus for an accuracy of ± 0.5 m and a gradient of 0.1 m/km the shoreline must be traced for a minimum projection distance of 10 km and a shoreline with a gradient of 0.01 m/km must be traced for a minimum projection distance of 100 km. Within a given study area a restricted distribution of raised marine landforms may therefore limit the accuracy with which shoreline gradients can be calculated from height-distance diagrams.

The second major use of height-distance diagrams in the study of raised shorelines is that of identifying shorelines, that is as a device for correlating raised marine landforms from diverse localities in the study area. This usage of a height-distance diagram is not fully compatible with its use in establishing the optimum plane of projection, for shorelines can only be identified in the optimum plane whilst the optimum plane can only be established once the shorelines in question are known. Several lines of independent evidence can, however, be utilised to overcome this difficulty.

It may be that certain shorelines are particularly well developed, such as the Main Rock Platform in this study, and these shorelines can be used to establish the optimum plane and then other shorelines can be correlated using these distinct shorelines as control. Where particular shorelines may not be distinguished on the basis of morphology they may, however, contain diagnostic fossils or other materials such as pumice that may be utilised to correlate various fragments (cf. Marthinussen, 1960; Blake, 1970). A further alternative may be that a sufficiently large number of samples for ^{14}C dating is recovered from a particular shoreline and the dates used to correlate the various fragments over a wide area (e.g. Andrews et al., 1970).

Other field relationships may be used in a negative sense to provide information limiting the various landforms that may be correlated in a single shoreline. Thus shoreline fragments occurring one above the other with a clear break of slope between cannot be part of the same shoreline (e.g. Cullingford, 1977; Cullingford and Smith, 1980). Another limiting factor is the position of the ice margin in relation to particular shorelines, for clearly two shoreline fragments cannot be part of the same shoreline if one was covered by ice while the other was being formed. This latter constraint is of less utility where the ice margin is retreating very rapidly for considerable lengths of coastline can become ice free in a relatively short period of time during which sea-level may fall only a few centimetres. The calving glaciers in NE Greenland provide an illustration of the rate at which tidal glaciers can retreat, Academy Glacier, for example, having retreated at more than 300 m/yr this century (Weidick, 1968). Similar rates of retreat allied to a relative sea-level fall of ca. 1.5 m/100yr (cf. Chapter 9) indicate that between formation of ice-marginal deltas 1 km apart the sea may have fallen only ca. 5 cm, a figure that is negligible in relation to the degree of accuracy with which the deltas can be surveyed today.

In addition to the above field relationships that are utilised in correlating shorelines on a height-distance diagram it is often necessary to make some assumption as to the form that the projected shoreline should have. Isostatic depression in response to loading by an ice sheet, the surface altitude (and hence, in general, the thickness) of which is greatest at the centre and declines non-linearly towards the periphery, should itself be non-linear, and shorelines deformed by the recovery from this depression should have non-linear profiles when traced at right-angles to the isobases. A number of studies (e.g. Andersen, 1960; Andrews et al., 1970) have correlated shorelines on a non-linear basis but the majority of studies around the coast of Scotland as elsewhere have correlated shoreline fragments on a linear basis (e.g. Marthinussen, 1960; Sissons, 1967a, p. 183; Andrews, 1970, p. 84) although occasionally (e.g. Sissons and Smith, 1965) the portions of shorelines farthest from the centre of uplift are shown to have a lower gradient than those closer to the centre of uplift. Those that favour linear shoreline correlations normally argue that the limited distances involved and the inaccuracies of the basic data do not permit refined curves to be fitted to the data (cf. Cullingford, 1977).

Although the procedures are complicated, the height-distance diagram is the most widely used and most powerful analytic device in raised shoreline studies. It has been used extensively in this study in the analysis of raised shorelines, where the various arguments detailed above are applied and clarified in the particular circumstances of the various shorelines. It has also been used in the analysis of river terraces but here its function is largely descriptive.

(b) Shoreline Relation Diagrams

A shoreline relation diagram is a diagram in which units on the x-axis are the altitudes of a particular shoreline or theoretical deformed surface. The y-axis units are the altitudinal differences measured from place to place between the shoreline or surface represented on the x-axis and other shorelines or theoretical surfaces. The fundamental assumption of the shoreline relation diagram is that there is a constant ratio between the altitudes of any two tilted shorelines. This 'proportionality principle' means that all shorelines plot on the diagram as straight lines. The great advantage of a shoreline relation diagram is that it is not located in space and can utilise data from any part of a glacio-isostatically deformed set of shorelines provided that the shoreline forming the x-axis can be identified in a particular locality.

The shoreline relation diagram was first developed and most widely used in Scandinavia (see Donner, 1965, for review) and it was also there that its application was first seriously questioned. Marthinussen (1960), Andersen (1965) and Donner (1965) finally gave expression to the lack of faith of the Scandinavian workers in the shoreline relation diagrams and no such diagram has been produced in Scandinavia for many years. More recently, however, using data from Arctic Canada, Andrews (1969, 1970) has argued in favour of the shoreline relation diagram, claiming to demonstrate that it rests on a sound physical basis. Andrews (1970, p. 51) produced a shoreline relation diagram for Highland Britain utilising data from the Forth Valley and this diagram has been utilised by Gemmell (1973) to predict the ages of certain shorelines in the Firth of Clyde, a matter resulting in some debate (Gray, 1975; Gemmell, 1975b). Data from the Firth of Clyde, including some from the present study area, was used by Synge and Stephens (1966) in a shoreline relation diagram for all of Scotland and NE Ireland. In order to assess the validity of the various shoreline correlations made by Synge and Stephens and of the chronology suggested by Gemmell it is necessary to consider the limitations of shoreline relation diagrams.

In the most recent justification of the physical basis of shoreline relation diagrams, Andrews (1969, 1970) approached the matter through consideration of uplift curves, that is, graphs of isostatic uplift against time. In a previous publication, Andrews (1968) had demonstrated that for Arctic Canada such curves had essentially the same form, the rate of uplift declining exponentially from a maximum at deglaciation to its present value. As with all exponential curves the uplift curves had a base constant (the value at time zero of the quantity that is varying) and a decay constant (the value of the rate of change of the variable). Andrews calculated a common decay constant for the whole of Arctic Canada and used this to construct a shoreline relation diagram by calculating, for localities with a different base constant, the amount of uplift since 14,000 yr BP, 13,000 yr BP, etc. and subtracting this from the amount of uplift since 2,000 yr BP at the same localities. The 2,000 yr BP uplift figures were plotted on the x-axis and the differences of each thousand year period on the y-axis. A graph showing a series of straight lines representing the amount of deformation relative to the 2,000 yr surface for each 1,000 year period back till 14,000 yr BP resulted. This is a theoretical shoreline relation diagram as no actual shorelines had been identified in the field and plotted on the diagram.

The critical point in the construction of the shoreline relation diagram is the calculation of the common decay constant for the whole of Arctic Canada, for in doing this Andrews assumed what he finally claimed to have demonstrated, the proportionality principle. The calculation of a common decay constant assumes that there have been, through time, no localities where isostatic uplift has been retarded, halted cr even reversed such as might occur with a glacial readvance, and it assumes that in all localities the unimpeded response of the crust has been similar (e.g. faults have played no role in isostatic uplift). Effectively then, Andrew's demonstration of the physical basis of the shoreline relation diagram is subject to the same constraints and assumptions as the original formulation by Scandinavian workers. The users of shoreline relation diagrams (and 'nomograms' based on the same principle, Gemmell, 1973) must therefore demonstrate that either the assumptions inherent in the use of these diagrams are met, or errors resulting from the violations of these assumptions are sufficiently small to be ignored in practice.

Work on raised shorelines around the Scottish coast has demonstrated a number of instances where the 'proportionality principle' has been in error. Sissons (1972) working in the upper Forth Valley was able to demonstrate that isostatic uplift had not been uniform either in space or in time. In the early part of the Flandrian a considerable propostion of the isostatic uplift in the upper Forth Valley had been accomplished by movement of faults, dislocating the raised shorelines studied, and occurring at discrete periods in time. Gray (1974a) was also able to demonstrate that the isostatically tilted (Gray, 1978) Main Rock Platform in the Firth of Lorne area of the west coast of Scotland was dislocated at one point and severely warped at another. In SE Scotland the directions of curvature of the isobases of the Main Perth and the Main Postglacial shorelines are not in sympathy (Cullingford, 1972) suggesting movement of the centre of uplift and hence a non-proportional relationship between the heights of the shorelines at different localities. These three examples clearly demonstrate spatial variations in the ratio of the altitudes of distinct shorelines. Temporal discontinuity in uplift has probably resulted from the expansion of glaciers during the Loch Lomond Stadial though this yet awaits final demonstration (Chapter 11).

The above observations make it seem unlikely that the shoreline relation diagram can be usefully applied to Scotland. As yet the total magnitude of the various errors resulting from the use of such a diagram in Scotland cannot be assessed, though perhaps considering the Lateglacial and Flandrian shorelines on separate diagrams would minimize any effects of the Loch Lomond Stadial glacier expansion. The most recent attempt to use the shoreline relation diagram in Scotland (Gemmell, 1973) has, however, shown little awareness of the problems involved in addition to which erroneous data have been utilised in its construction (Gray, 1975). Gemmell's conclusions that the ice front lay approximately along a line connecting Stranraer and Campbeltown at around 13,500 yr BP, that the southern half of Arran was deglaciated by around 13,000 yr BP and that Arran became free of Highland ice by about 12,500 yr BP are therefore unacceptable and have indeed been contradicted by later ^{14}C dating (Chapter 9). In similar fashion, the various shorelines proposed by Synge and Stephens (1966) on the basis of a shoreline relation diagram must be viewed with scepticism, particularly in the light of other criticisms of their approach (see above; Sissons, 1967b; Gray, 1972a).

(c) Trend Surface Analysis

Measured altitudes on an isostatically deformed shoreline are sample points of the surface of deformation resulting from isostatic uplift since the time of formation of the shoreline. The overall form of the surface is therefore three dimensional although the majority of the analyses of isostatic recovery have been two-dimensional (height-distance diagrams, shoreline relation diagrams). There were occasional early attempts to portray the three-dimensional form of the surface of deformation as represented by isobase maps of particular shorelines (see Wright, 1937, for examples) but due to the lack of an adequate data base or of adequate map coverage (as in Arctic Canada, for example) isobase maps were only occasionally constructed. In more recent years, however, increased collection of much more soundly-based data allied to a greater number of techniques for correlating various shoreline fragments has resulted in the frequent construction of isobase maps.

Initially, isobase maps were drawn by hand, interpolation being a purely mental process (e.g. Marthinussen, 1960; Sissons, 1965, 1967a) but with the advent of the digital computer more objective methods of surface construction could be utilised. In 1967 McCann and Chorley produced the first map of raised shoreline isobases utilising the least-squares technique of trend surface analysis, and although this initial study was marred by the use of data from more than one shoreline (Gray, 1974b) the technique has become widely used in the analysis of shoreline displacement (Smith et al., 1969; Andrews, 1970; Saarnitso and Huhn, 1973; Gray, 1974 a and b).

Trend surface analysis is the name that has been given to the multiple regression technique that fits a polynomial equation of chosen order by the least squares criterion to a data set (in this case shoreline altitudes) defined by the independent variables Eastings and Northings. The technique is widely applicable outside shoreline studies and has been extensively discussed (see Unwin, 1975, for references). Certain problems have been raised in the application of trend surface analysis to raised shoreline studies and these are discussed here.

Tarrant (1970) levelled two criticisms at the use of trend surface analysis in raised shoreline studies. Firstly he considered that a capping surface was desired and not one that effectively bisected the point distribution. In arguing this Tarrant was apparently envisaging raised shoreline data as being similar to data collected to represent erosion surfaces. As detailed in Chapter 3, however, it is clear that the intertidal flats, rock platforms, etc., that are used to reconstruct raised shorelines have an internal variability that lends them amenable to analysis by a technique that cuts through rather than caps the data points. Furthermore, subsequent modifications of the form of the raised shoreline landforms such as gullying or slumping of material on to the surface are specifically excluded from surveying and hence should not bias any surface.

Secondly, Tarrant argued that autocorrelation between data points would be likely to produce spurious tests of significance. The matter is of wider significance than Tarrant realised for it is in the nature of a surface in which there is a trend for the values on that surface not to be independent of the values of their neighbours. Autocorrelation is therefore inherent in all trend surfaces as was earlier pointed out by Tinkler (1969) who maintained that the essential requirement was not whether the data values were autocorrelated but that the residuals from the surface of the measured values at the data localities were independent. This latter requirement is difficult to justify if the data used as an input to the trend surface analysis consist of several points on each raised marine landform surveyed. Clearly although a landform may bear a specific altitudinal relationship to mean tide level and a series of similar landforms around a length of coastline vary above or below this, a set of points on a particular landform will be autocorrelated and so will their residuals be from the overall surface. This problem is best overcome by using the mean altitude of the heights surveyed on a particular landform (Gray, 1972b) a procedure which was previously suggested for the more general treatment of surveying results (see above).

The practice of using the mean altitude of a landform in trend surface analysis is beneficial in reducing the degree of clustering inherent in the data, for it is well known that clustered data can produce erroneous trends (Unwin, 1975). Gray (1972b, 1975a) has recommended the use of the geographical mean of the Eastings and Northings of the various points surveyed on a particular landform as the most appropriate locality to which to assign the mean altitude value. This procedure has not been adopted in this study as it is possible that the mean of the relevant Eastings and Northings does not lie on the landform being considered and it has been part of this methodology, irrespective of the use of trend surface analysis, to derive a mean altitude that could be assigned to a particular marine landform. The middle of the back edge or the crest of the ridge of the relevant landform has been the point used for all calculations.

TABLE 19. RESURVEYED RAISED MARINE LANDFORMS.

Feature	Grade	Survey 1	Survey 2	Difference
S171	m	7.41	7.53	-0.12
S172	p	8.71	8.64	+0.07
S173	m	13.36	13.46	-0.10
S174	m	11.12	11.04	+0.08
S175	p	10.62	10.75	-0.13
S176	m	12.94	12.85	+0.09
S177	p	14.82	14.86	-0.04
S178	m	7.75	7.67	+0.08
S179	p	4.96	4.88	+0.08
S80	m	9.00	8.93	+0.07
S81	p	6.77	7.03	-0.26
S82	p	6.67	6.87	-0.20
S83	p	5.47	5.47	0.00
S84	m	12.39	12.87	-0.48
S85	g	30.31	30.35	-0.04
S210	m	4.71	4.76	-0.05
S211	m	4.99	4.93	+0.06
S212	p	13.46	13.58	-0.12
S213	p	10.50	10.37	+0.13
S214	m	25.49	24.48	+1.01
S215	m	36.52	36.07	+0.45
S216	m	38.73	39.04	-0.31
S217	g	40.97	40.71	+0.26
S218	p	30.04	31.27	-1.23
S219	m	12.20	12.17	+0.03
S220	m	11.10	10.94	+0.16
S225	m	10.98	11.43	-0.45
S226	m	10.37	10.45	-0.08
S227	m	14.19	14.02	+0.17
SR22	-	11.67	11.21	+0.46
SR23	-	10.79	10.80	-0.01
SR24	-	13.18	13.14	+0.04

CHAPTER 5

FIELD DESCRIPTION OF AREAS AROUND LOCH LONG AND THE HOLY LOCH

1. Introduction

This chapter and the following two chapters contain the detailed field descriptions of the study area. Long stretches of coastline have little that is worthy of detailed description while in other areas there are very well developed sequences of landforms. This chapter deals with the area around Loch Long and the Holy Loch, Chapter 6 with the stretch of coastline flanking Loch Striven to Loch Riddon and Chapter 7 with the coastline of Loch Fyne. Figure 19 indicates the location of the areas described in detail.

2. Upper Loch Long

For much of its length there is little evidence of raised shorelines beside Loch Long, this being due in part to the very steep slopes that border the loch, in part to extensive landslipping particularly along the eastern side of the loch, and also to the lack of any major valley joining the loch the mouth of which would have provided a suitable locality for the deposition and preservation of marine sediments. At the head of the loch, however, three valleys converge, Glen Croe from the W, Glen Loin from the N and the glacial breach from Loch Lomond from the E. River terraces and raised shoreline features occur at the mouths of these valleys although the small town of Arrochar has been built on the deposits at the mouth of the glacial breach and partly on the deposits in Glen Loin, thus precluding accurate surveys of these localities.

The Croe Water descends rapidly to Loch Long and has built a delta

Figure 19: Location of areas described in detail.

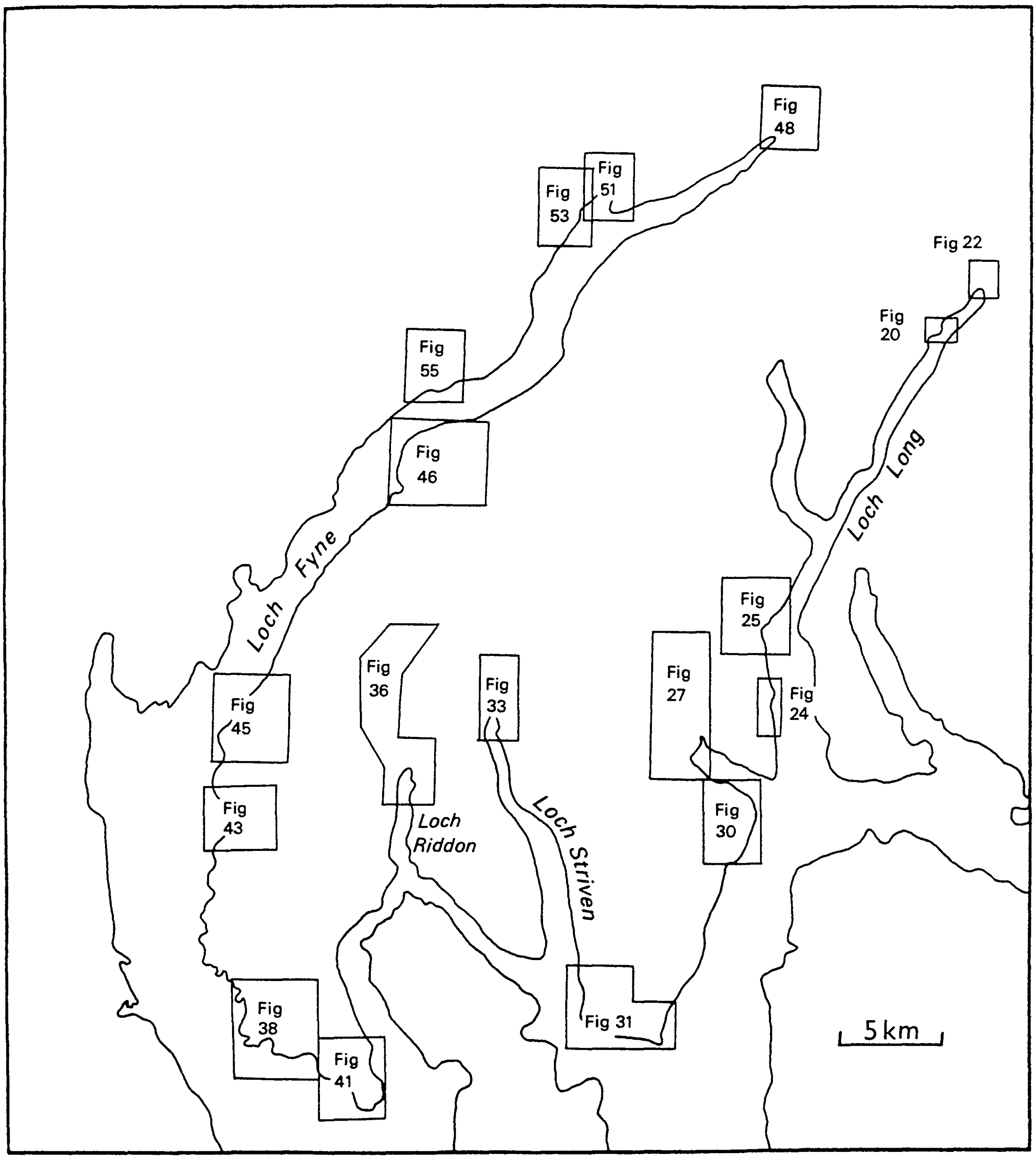


Fig 48
Fig 51
Fig 53
Fig 22
Fig 20
Fig 55
Fig 46
Loch Long
Loch Fyne
Fig 25
Fig 36
Fig 33
Fig 27
Fig 24
Fig 45
Fig 43
Loch Riddon
Loch Striven
Fig 30
Fig 38
Fig 41
Fig 31
5 km

in which a number of terrace levels may be identified (Figure 20). Anderson (1894) considered that this deposit may have been the terminal moraine of a glacier occupying Glen Croe but except for a small number of exposures of grey till at the mouth of Glen Croe and scattered erratic blocks there are no obvious glacial deposits. The terraces were mapped and, where not obscured by buildings, levelled. A height-distance diagram with a projection plane parallel to the axis of Glen Croe was constructed (Figure 21). A series of steeply sloping river terraces can be identified, the highest commencing at ca. 22 m O.D., and the lowest, occurring just above the present river level, being traceable down to ca. 4.5 m O.D. Feature S281, from its position on the diagram, appears the first formed. It faces out over the loch and its surface gradient diminishes in that direction. It is interpreted as a small delta with a surface break of slope at ca. 13.2 m O.D. This is the only feature that appears to be related directly to a former sea-level.

Within Glen Loin a number of terraces have been mapped (Figure 22) and levelled where the presence of buildings was not prohibitive. The largest terrace fragment, T35, is complicated by the presence in it of kettle holes near which two poorly developed mounds rise above the surface of the terrace. Two channels are cut into the northern part of T35 though these may be due to post-depositional erosion as they slope down to the NW, that is, in a generally up-valley direction.

The altitudes on the various terrace fragments have been plotted on a height-distance diagram (Figure 23), the projection plane for which parallels the valley axis. The two best developed terrace fragments, T35 and T36, both have rather irregular surfaces but exhibit overall downvalley gradients. T35 can be followed down to ca. 10 m O.D., that is, below the level of the uppermost Flandrian shoreline in this area, suggesting it to be Flandrian in age, but the presence of the dead-ice features indicates a complex history of development of this terrace, fluvioglacial deposits perhaps being reworked into a terrace as sea-level fell from the maximum of the Flandrian transgression. The largest kettle hole has a lip at ca. 12 m O.D. and sounding with aluminium rods indicates at least 2 m of peat infill. If the above suggestions as to the formation of the terrace are correct sediments derived from the Flandrian marine transgression should be isolated in the kettle hole.

Evidence for raised marine features is poor in this area. Feature S17 appears indicative of a former sea-level at ca. 10.8 m O.D. Terrace fragment T34 at ca. 14.6 m O.D. is poorly developed and located on part of an alluvial fan which casts doubt as to its origin.

In summary, in the upper part of the Loch Long valley the highest evidence of former marine activity appears to be feature S281 at the mouth of Glen Croe related to a sea-level of ca. 13.2 m O.D. A separate terrace fragment (S17) at the mouth of Glen Loin suggests a second sea-level of ca. 10.8 m O.D. though this is not well-developed. A number of terraces in both Glen Croe and Glen Loin relate to sea-levels below those mentioned above. Glen Loin appears to have been occupied by a small body of stagnant ice at the dissolution of the last glacier to cover this area.

3. Ardentinny Area

The western shore of Loch Long between Blairmore and Glen Finart is flanked by a series of raised marine landforms. Near Ardentinny numerous glacial deposits are found whilst in Glen Finart there is a set of terraces. These features are of considerable interest in the interpretation of the glacial events and sea-level changes of this area for while around Blairmore and Gairletter the marine limit occurs at ca. 41 m O.D. at Ardentinny the highest marine landform is only ca. 13.5 m O.D., glacial landforms being found immediately above this altitude.

(a) Blairmore and Gairletter

The raised marine deposits at Blairmore and Gairletter (Figure 24) occur at the mouths of the Coire Bhlair and Stronchullin valleys respectively, drainage from these valleys supplying debris for the construction of the marine landforms. The uppermost terraces occur in two distinct levels, ca. 41 m O.D. (S217: 41.47 m; S208: 41.67 m; and S209: 40.90 m) and ca. 37-38 m O.D. (S215: 36.52 m; S216: 38.73 m; S201: 37.99 m; and S202: 36.05 m). A large sand pit in S201 revealed massive foreset bedding that confirmed the deltaic origin of the features around the mouths of the side valleys.

Below the lower group of terraces only two landforms appear to record the ensuing drop in sea-level (S214: 25.49 m, and S218: 30.04 m) until below ca. 13.5 m, where a large number of raised marine landforms is encountered. The downcutting of the Stronchullin Burn at Gairletter into the deltaic deposits that accompanied the drop in sea-level is marked by two small very steeply sloping terraces, T100 and T101, the lowest points surveyed on which are 25.15 m O.D. and 22.50 m O.D. respectively.

The lower series of marine deposits is backed along this length of coast by a cliff cut into bedrock and sediment alike. This cliff is associated with the Main Rock Platform discussed in Chapter 8. N of

Gairletter and also N of Glen Finart there is a number of localities where the platform can be observed but around Ardentinny the platform cannot be seen and the cliff is often masked by glacial deposits. No fragments of the platform were sufficiently well defined to allow heighting of the back edge.

At Gairletter and Blairmore the Flandrian marine deposits are developed as small deltas and shingle ridges. The highest marine features in this group occur at ca. 13.5 m O.D. (S212: 13.46 m; S203: 13.52 m) and other common altitudes are ca. 10.0 m O.D. (S213: 10.05m; S205: 10.14 m) ca. 8.5-9.0 m O.D. (S198: 9.06 m; S199: 9.17 m; S206: 8.57 m), ca. 5.0 m O.D. (S210: 4.71 m; S211: 4.99 m; S204: 5.17 m; and S207: 5.15 m) and a well developed lower fragment, S200, occurs at 4.11 m O.D. Shingle ridges have been levelled at 13.18 m O.D. (SR24), 11.67 m O.D. (SR22) and 10.79 m O.D. (SR23). Data on the height at which present marine processes operate in the Blairmore and Ardentinny areas (Chapter 3) indicates that shingle is moved to 1.3-1.6 m higher than the break of slope at the back of intertidal deltas, suggesting SR22 relates to the same former sea-level as do S213 and S205 (i.e. ca. 10.0 m O.D.) and SR23 relates to the same former sea-level as S198, S199 and S206 (i.e. ca. 8.5-9.0 m O.D.). Shingle ridge SR24 does not, however, appear to relate to another group of shoreline features in this manner.

(b) Ardentinny

West of the village of Ardentinny are numerous mounds of glacial material. These are interspersed with dead ice hollows and a number of meltwater channels have been mapped (Figure 25). A large section eroded by the Ardentinny Burn in one of the moraines revealed well-bedded sands and gravels overlain by a layer of till that is succeeded upwards to the top of the exposure by alternating bands and lenses of till and sorted sands and gravels. This sequence is similar to those described by Boulton (1967, 1972) as developing from supraglacial meltout of debris bands and this allied to the presence of dead-ice hollows is suggestive of a supraglacial ice-marginal origin for these deposits. The moraines could not be traced downloch of the vicinity of Ravenrock Cottage.

The highest possible marine landform by Ardentinny is S227 at 14.19 m O.D.. There are no other similar landforms at this altitude in this area. The next highest occur as a group at ca. 13.5 m O.D., these being S221 (13.30 m), S222 (13.43 m) and S230 (13.32 m) a further group of raised marine features has been levelled at ca. 11.0 m O.D. (S229: 10.98 m; S220: 11.10 m; and S225: 10.98 m). Other marine landforms have been heighted as follows: S219 (12.20 m O.D.), S226 (10.37 m O.D.) and S228 (6.0 m O.D.).

Trial pit records for construction work at Ardentinny revealed layers of peat in the Flandrian marine deposits indicating a transgressive origin for certain of the Flandrian shorelines.

(c) Glen Finart

The terraces that occupy the bottom of Glen Finart are shown in Figure 25. Subsequent to levelling a height-distance diagram was constructed (Figure 26), the projection plane for which runs parallel to the axis of the valley. There are few glacial landforms in the lower part of Glen Finart, the clearest being a large mound oriented at right angles to the valley side opposite Drynain. Two further elongated mounds (M310, M311) occur nearer the mouth of the glen. They are directly opposite one another and trend parallel to the axis of the valley. Although their·altitude is at or below the upper limit of Flandrian marine deposits and their morphology resembles shingle ridges, this interpretation is rejected because of their very sheltered localities and apparently complementary situation. They are thought to be glacial in origin.

The terraces in the valley bottom are divided into two groups: a small number of terraces lies above 14 m O.D. in the upper part of the valley and a greater number of terraces occurs below this altitude near the mouth of the valley. Perhaps coincidentally, the upper group of terraces occurs at or upvalley of the cross-valley mound opposite Drynain. The terraces are often eroded into or overlain by the large alluvial fans that are developed where the steeply sloping side streams reach the valley bottom. These alluvial fans are still spectacularly active today as has been documented by Smellie (1912).

The terraces in the upper part of the valley are only poorly preserved. The highest levelled (T457) occurs at ca. 24 m O.D.. A section in this terrace revealed well-stratified silty clays, sands and gravels, coarsening upwards in general. The presence of fine-grained sediments is indicative of quiet water deposition and clearly rules out the possibility that the terrace is a fragment of an alluvial fan. Similarly well-bedded sediments, dominantly fine sands, were observed in a section ca. 500 m upvalley. Deposition in quiet waters at this altitude can be most easily explained by a higher sea-level or ponding by glacier ice. In view of the evidence discussed above for sea-levels around Ardentinny being no higher than ca. 13.5 m O.D. and the lack of corroborative marine sediments or landforms at this altitude towards the mouth of Glen Finart, it is thought that glacier damming is the most likely explanation of these terraces and sediments.

The terraces near the mouth of Glen Finart are grouped in a very small area. They do not define a series of downvalley sloping terrace fragments but in general have flat surfaces. The highest of these occurs at ca. 13.4 m O.D., the same altitude as the uppermost Flandrian shoreline in this area, suggesting that this feature was formed in relation to that sea-level. The southern of the two glacial ridges also occurs at this altitude indicating it to have been washed over at this period. Other terraces group around 10-11 m O.D. (T451, T446, T450 and S393) and ca. 9.0 m O.D. (T448 and T449) and these correspond to groups of marine landforms described previously along the coast to the south. The only feature to demonstrate a clear downvalley slope is the large abandoned channel C401 which occurs at the level of the present floodplain.

The occurrence of these terraces at the mouth of Glen Finart, the surfaces of which accord with distinct periods of shoreline formation and which have no clear upvalley extensions, suggests that the bulk of the sediment in this part of Glen Finart was not derived from upvalley but was deposited by a glacier and reworked by the sea. The occurrence of two ridges of likely glacial origin amidst the terraces is in accord with this suggestion. Unfortunately sections are not available to substantiate this interpretation.

4. Strath Eachaig/Holy Loch

Strath Eachaig is the major through valley of the Cowal Peninsula. It is occupied for more than half its length by Loch Eck, and it is that part of the strath that extends from the SE end of the loch down to the Holy Loch that is under consideration here. Within this five kilometre stretch the main valley is flanked by very steep rocky slopes, rising at least 500 metres within a kilometre of the valley floor. Along the sides of the Holy Loch the character of the terrain, particularly on the southern side, becomes more subdued, altitudes only reaching over 300 metres some distance back from the loch side and much larger quantities of drift are encountered on the slopes. Between Loch Eck and the sea two large valleys enter Strath Eachaig from a south-westerly direction. The more northerly of these is Glen Massan which is joined as it reaches Strath Eachaig by Corarisk Glen whilst to the S of these Glean Lean enters the main valley immediately above the end of the Holy Loch. Both these side valleys carry rivers of comparable size to the River Eachaig.

The bottom of the strath is covered by glacial, marine and mainly riverine deposits. Bedrock slopes down steeply under these at the edges and depths of infill cannot be easily estimated: at Cot House over 60 metres of sediment were encountered during the construction of a road bridge without rockhead being met. The lower part at least of Strath Eachaig is therefore an infilled sea loch and it is this process of infilling during and subsequent to the decay of the last ice-sheet to cover the area that is being studied here, a process that is still going on today as the River Eachaig slowly builds out its delta into the Holy Loch.

The landforms of lower Strath Eachaig are shown on Figure 27. It has been necessary to construct two height - distance diagrams for this area - a main one down Strath Eachaig (Figure 28) and a subsidiary one for lower Glean Lean (Figure 29) - as these valleys trend at approximately right angles to each other. A number of landforms have been incorporated in both these diagrams to facilitate comparison.

Loch Eck is dammed by drift. No bedrock is exposed near the outlet of the loch and a series of poorly developed mounds and hollows can be traced across the valley. These are flanked to the E and downvalley by small, clear terrace fragments that have recently been cut through during road works to reveal well-bedded sands and gravels. No till has been seen in this part of the valley. Surveying the terrace fragments shows them to occur as two distinct groups. Downvalley of the mounds and hollows at the end of Loch Eck terrace fragments T371 and T384 correlate across the valley. Terraces T385, T370, T366, T367, T368 and T369, however, are morphologically below and lie up-loch of T371 and T384. Of them T367, T368 and T369 flank the valley and have mounds and hollows between themselves and the loch. They are therefore interpreted as kame terraces with a small area of kame and kettle topography beside the loch. These landforms are interpreted as having been formed at the terminus of a valley glacier occupying upper Strath Eachaig. The two higher terrace fragments are interpreted as the remnants of an outwash spread formed beyond the glacier margin. The lower terraces are related to a time of ice decay as meltwaters drained along the sides of as well as through the ice and formed a second outwash train (starting at T366) incised into the earlier one. Today the River Eachaig is some 6 - 10 metres below this lower outwash level and it is thought that this incision occurred relatively soon after ice decay and possibly while ice still remained in the kettle holes to the E of the loch. This is suggested as there are no emerged shoreline features (apart from small fragments ca. 1 metre above the present loch level) around the loch.

Unfortunately, the terrace sequences cannot be followed in detail downvalley as a large alluvial fan at the mouth of Inverchapel Glen occupies much of the main valley here and has resulted in erosion of the major part of the outwash terraces. This alluvial fan is of sufficient magnitude to be of note in itself, the more so as Inverchapel Glen is rather small. The development of the fan and its relationship with the River Eachaig can be traced through the respective truncation of the river terraces by the fan and vice versa. Terraces T373 and T374, although aligned downvalley are clearly influenced by the fan and are perhaps best considered a part of it, the deposition of T374 being affected by the rock outcrop immediately to its E, a (peat-filled) channel behind the outcrop demonstrating that the Inverchapel Burn once discharged into the main valley by this route. The fan clearly truncates outwash terrace T371 and grades effectively to present river level, though T372, part of the present river floor plain is cut into it.

Downvalley from Inverchapel only two terrace fragments, T386 and T387, can be traced back to the ice front at Loch Eck, continuity of these features being better on the ground than can be represented by the heights on the diagram. At Benmore Botanical Gardens artificial modification produces a gap across which only the present floodplain can be correlated with any confidence. The major difficulty revolves around the possible correlation of well-established fragments T390 and T375 with the reconstructed outwash T387 - T384. Since T375 descends to 12.5 metres, that is, below the projected uppermost Flandrian shoreline (see Chapter 10), it may be argued that T375 and T390 are Flandrian in age, but the possibility clearly exists that previous terrace levels relating to the outwash were inherited and modified during the Flandrian. In addition, the apparent alignment of these terraces upvalley with the outwash terraces supports this contention, especially when it is remembered that Loch Eck is believed to have remained at its present level throughout the Flandrian and its outlet must therefore be regarded as a point on all Flandrian river profiles. It can therefore be said that the ice limit at the southern end of Loch Eck was related to a sea-level at least below 21 metres and perhaps even as low as the uppermost Flandrian shoreline (ca. 13.5 metres).

Terraces Fragments T375 and T390 are related to the River Eachaig but T390 is cut into on its western side by the River Massan, across which terrace T407 correlates with the above two. T407 fronts an end moraine that can be followed into Glen Massan where it forms an arc around the mouth of Corarisk Glen, the River Massan being diverted round it. The moraine has a double ridge over part of its length and recent drainage ditches cut across it show it to be composed of red till-like material with a range of particle sizes and no obvious sorting. It was originally identified as an end moraine by McLaren (1855) in the early days of the glacial theory. Since T407 abuts on to this moraine sea level must have been below 14 metres when the moraine was deposited.

Considerable lengths of the River Eachaig floodplain have been levelled and it is appropriate to consider them at this point, in part because they provide a guide for correlation of the Flandrian terrace fragments but also because they illustrate one of the major difficulties of height-distance diagrams that has not been previously mentioned in this study. Terrace fragments T403, T409, T377, T378, T379, T398, T394, T382, T372 and the outlet of Loch Eck define a concave-upwards river long profile, with perhaps a slight inflection in the gradient just below the junction with the River Massan. Superimposed upon this general trend there are meanders that in places such as T409, T377, T378 and T379 curve back upvalley producing the effect on a height-distance diagram of a step in the long profile. It can be argued that outwash terraces that were probably formed by braiding rivers would be less likely to give rise to this effect but correlations of small Flandrian river terrace fragments, which it is only reasonable to presume were formed by a meandering river, must be made very cautiously. In the following proposed correlations, close attention is paid to continuity of landform (especially when supported by continuous levelling) and superposition of fragments, that is those that are separated by clear breaks of slope. Yet even this latter criterion must be treated with caution as a meandering channel, when it turns back on itself will start to cut into its own floodplain deposited farther up the valley, thus producing a clear break in slope. Such a situation is well exemplified by T445 on the Little Eachaig. Furthermore, the existence of abandoned river channels can disguise the significance of some breaks of slcpe or emphasise minor details whilst failure to note the presence of such channels can result in erroneous terrace altitudes.

It can be presumed that all the above characteristics are present in the data relating to the Flandrian terraces of the River Eachaig. Three distinct terrace levels have been identified, however, between the uppermost Flandrian terrace and the present river floodplain. The very large terrace fragment T408 on the W side of the valley provides a starting point for analysis. This terrace is continuous over a long distance, the only difficulties being the small alluvial fans built out on to it, and an area towards the S where mapping and levelling were impossible due to a dense forestry plantation. Levelling on either side of this area suggests that there is in fact a break in the terrace at this point and the last four points on the terrace as originally mapped are believed to belong to a separate terrace fragment (T408a). With this exception the rest of T408 shows remarkable consistency in gradient and can be correlated across valley with T402 and T376, both of which have now been extensively modified by road re-alignment. T376 was originally mapped as one feature but again levelling indicates two distinct surfaces with the higher one, relabelled T376a, coinciding in altitude with the level (Chapter 10) of the uppermost Flandrian shoreline suggesting it to have originated as a small delta. T408 can be correlated upvalley with T381 and T400, these two fragments being separated by an abandoned channel. Farther upstream, correlation is suggested with the well-developed T388. At altitudes between T388 and T390 is a set of small unpaired terrace fragments T391, T392, T389 and T396 which are thought to be remnants left during the erosion of the T390/T375 terrace level. The above terrace correlations define a concave upwards river long profile that diverges downvalley from the present river profile. The downvalley trace of these terraces reaches ca. 8.0 metres in altitude. Such an altitude is only slightly above the small isolated fragment T411 and a marked period of marine deposition (SR326 and SR327, for details, see below) and it is suggested that this is the sea-level to which these terraces grade.

T380 is a large fragment situated E of the present Massan/Eachaig junction. It is morphologically below T381 of the previous terrace. Upvalley T380 is correlated with T399, T401, T393, T397 and T395. Of these T399 and T401 occur slightly above the projected profile, presumably due to their occurrence on a contemporaneous meander whose form can be deduced from the distribution of the fragments of this terrace sequence. Correlation downvalley must be tentative owing to the gap in fragments but it is suggested that T408a does correlate due to its position relative to T408 and to downvalley projection of the profile defined by the already correlated fragments, the line of projection being helped by the known present river profile and the profile of T408. If such a correlation is accepted then a relationship between this terrace and sea-level as defined by T429 and SR323 (6.5 - 7.0 metres) is indicated.

Terrace fragment T409 has previously been mentioned as part of the present river floodplain. It diverges slightly over the surveyed portion from the other surveyed lengths of the floodplain and when mapped no morphological break was identified between it and the later-levelled T427. It is therefore suggested that the floodplain, which is clearly below T427, has been incised into the terrace level of which T427 is now a remnant and that T427 represents the lowest of a set of divergent downvalley concave-upwards river profiles that have developed through the Flandrian in the Eachaig valley. In addition T427 is eroded into the deposits of the Little Eachaig River and it is these, the complications that they produce in the development of the Strath Eachaig terraces and the direct evidence of sea-level changes that must now be analysed.

Figure 29 has been drawn in order to illustrate the altitude relationships of the landforms in the lower Little Eachaig valley. Terraces have been mapped and levelled in this valley from the junction of glens Kin and Lean and out into Strath Eachaig itself. Above the junction of the glens there are gorges in which no terraces were found, whilst the terraces in the lower valley have been modified by drainage from Gleann Ban.

A group of fragments occurs at the top of the diagram at 39 $\pm$ 0.5 metres due to the coincidence of S295, S389, S390 and S301. Of these, S295 and

and S389 occur on opposite sides of the mouth of the Little Eachaig valley. They both have exposures, S389 especially so, and foreset bedding is revealed in both. In addition to the well-bedded sands and gravels, at S389 there are considerable beds of silts and clays within the foresets. These can be followed along the slope where they overlie red till. S390 and S391 are paired on either side of the mouth of a meltwater channel that cuts through the spur of Dalinlongart Hill. The coincidence in altitude of these features and the foreset bedding, which implies deposition in standing water, suggest that these are shoreline fragments. The openness of this area to the present sea suggests that the shoreline in question is marine in origin. As such it is the highest level for which there is evidence of marine occupation and it is also the farthest inland in the Holy Loch/Strath Eachaig area that the sea at this altitude has penetrated. It seems likely that T419 continues this depositional surface upvalley along the Little Eachaig and this possibly correlates with the uppermost (unlevelled) terrace on the W side of the valley. Upvalley of these terraces just within the mouth of Glen Kin is an area of sand and gravel deposition. This is now unfortunately Dunoon's refuse dump and much of the original surface form has been destroyed but two kames and one dead-ice hollow have been mapped. It is suggested that this was an area of decaying ice when the ca. 39 metre shoreline was forming. Such a suggestion receives support from the existence of S390 and S391 which have been built by deposition at the mouth of the channel that carried meltwaters from this dead-ice mass. If this reconstruction is correct then the form of T419 indicates that lower Gleann Ban was ice free at this time, though the size of T419 also suggests that there was ice decaying within the glen. Small mounds in the valley below the level of and slightly upvalley from T419 may well be the result of melt-out of dead ice once T419 was abandoned.

The other terrace fragments within this portion of the valley cannot be fitted together into any coherent sequence of terrace levels. They were presumably formed by the erosion of the deposits of the uppermost terrace level as sea-level fell and ice finally wasted in the upper valleys. An additional factor may be the presence of Loch Lomond Readvance glaciers (Chapter 11) within the drainage basin. No evidence exists in this area for specific sea-levels intermediate in altitude between the uppermost shoreline and the Flandrian shorelines. Features associated with these latter shorelines, especially shingle ridges, are, however, preserved in considerable detail. Altitudinally they form a descending sequence with levels of 12.42 m (SR322), 10.89 m (SR324), 8.57 m (SR325), 8.19 m (S296), 8.09 m (SR326), 7.67 m (SR327), 6.70 m (SR323), 5.56 m (SR331), 5.27 m (SR332), 4.38 m (S392) and 3.42 m (S399). Areally these are distributed in a broad, slightly arcuate band across the valley, seaward of the deposits of the Little Eachaig River, the highest shingle ridge being closest to the mouth of the valley. The continuity of these features across the valley is broken in three places: by the River Eachaig, by the River Little Eachaig and by a former course of the Little Eachaig now represented by terrace fragment T443, which grades imperceptibly into S392 to the S and has no identifiable break with S399 to the N. The present exit of the Little Eachaig to the sea through these shingle ridges is inherited as is indicated by terrace T430 which appears to flatten out and relate to a sea-level of about 5.4 metres, this being higher than the topmost portion of T443, even though farther downstream.

The sequence of terraces at the mouth of the Little Eachaig is therefore interpreted as the result of the river readjusting its course to the fall in Flandrian sea-level. As sea-level fell, deposits were built out into the sea and were reworked on the exposed down-loch side into shingle ridges. No distinct terrace systems can be identified, presumably due to the steep gradient and short distance over which these sediments were deposited. Certain relationships between terraces and sea-levels can, however, be identified. It is clear, taking into consideration the distribution of shingle ridges, and of the terraces of the two rivers that the Little Eachaig constructed its own delta for much of the Flandrian. Terrace T412 can be traced up Strath Eachaig until it is truncated at ca. 9.0 metres and sea-level must have been lower than this before the two terrace systems coalesced. The exact time of coalescence and the contemporaneous sea-level cannot be inferred but no distinct break across valley has been identified between T412 and T429, this latter terrace being common to both rivers. Terrace T429 can be traced down to ca. 5.8 metres and by the time that the sea had reached this level the two river terrace systems had met. Shingle ridge SR323 (6.70 m) rests on T429 and may have been formed contemporaneously with the terrace. Subsequent to the abandonment of T429 both the Little Eachaig and the Eachaig cut terrace levels into their respective sides of this terrace. The Eachaig reached the sea on the northern side of the valley, as it does today, the Little Eachaig joining it, again as it does today. However, upon the further fall of sea-level to ca. 4.4 m the Little Eachaig abandoned this course in favour of T443 only to revert to it as sea-level assumed its present position.

To the SE of Dalinlongart excavation work on a building site revealed a 1.4 m section of peat overlying a 30 - 40 cm layer of grey silty sand with a sporadic covering of pebbles and cobbles which in turn lay on top of truncated reddish-grey possibly-laminated clays. The base of the peat section graded into the underlying sands, the lower part of the peat containing mineral material whilst the topmost part of the sand included organic material. At the base of the peat (6.9 m O.D.) a piece of wood trapped on the sand surface was sampled for radiocarbon dating and gave an age of 3800 $\pm$ 100 yr BP (U-2562). The small size of the piece of wood suggests that it would have no great apparent age when deposited whilst its inclusion at the base of a peat bog that is transitional from a marine littoral deposit and resembles (at the base) currently forming salt marsh peats is taken to indicate a washed-in marine origin for the wood. Similar driftwood has been levelled around the present high water mark and is found to have a mean height of 3.2 metres (Chapter 3). A sea-level fall of 3.7 metres therefore seems to be implied for the last 3700 - 3900 years.

Bordering the S side of the Holy Loch and extending into the suburbs of Dunoon is a set of landforms that shows good evidence of sea-level changes and their relationship to decaying ice. The main interest centres around the Loch Loskin depression which runs from above Sandbank to the West Bay at Dunoon and which is bounded on both sides by bedrock slopes that rise rapidly to over 200 metres to the W and are ice-moulded and streamlined to the E (Figure 30). The existence of Loch Loskin in the depression is a direct result of damming by sand and gravel deposits. One small section in these revealed well-sorted, well-bedded sands and gravels and, though the major part of the deposit is flat-topped, outwards from the valley wall the features take on a rounded nature and the loch is bounded on its southern side by a sharp slope. Levelling of the flat tops reveals them to have an almost horizontal surface at ca. 39.5 metres which terminates near Dunloskin Farm (NS 169 781) where a lower terrace level cocurs at ca. 32.0 metres. The landforms are here interpreted as resulting from the decay of a small area of stagnant ice trapped in the depression, the water level in which was controlled by the contemporaneous sea-level. Loch Loskin thus occupies a kettle hole. NW in the depression a large alluvial fan has been built on the W side, but a long terrace (S297) occurs on the E. This terrace is joined at its NW end by a small esker which trends down the opposing slope and across the peat-covered floor of the depression, just on the watershed to the Holy Loch. Seven surveyed points on the terrace reveal it to decline only slightly from 39.6 metres where the esker joins to 38.8 metres at its far end. A second narrow flat-bottomed, peat-covered depression diverges to the NE from near the end of S297 to open out above the Holy Loch. Due to the peat this was not levelled but it is approximately the same altitude as the other flats in the area. Its role in the drainage during ice-sheet decay cannot be assessed on this scanty evidence as it may have functioned as an escape route for meltwaters from the depression (implying the mouth of the Holy Loch to be free from ice) or meltwaters may have drained into the Loch Loskin dead ice mass through it (implying occupancy of the Holy Loch mouth by ice). The presence of the esker on the present watershed between the Holy Loch and Loch Loskin does, however, imply ice in the upper part of the Holy Loch at this time. S297 is seen as forming in relation to the same englacial water-table as the features at the SE end of Loch Loskin.

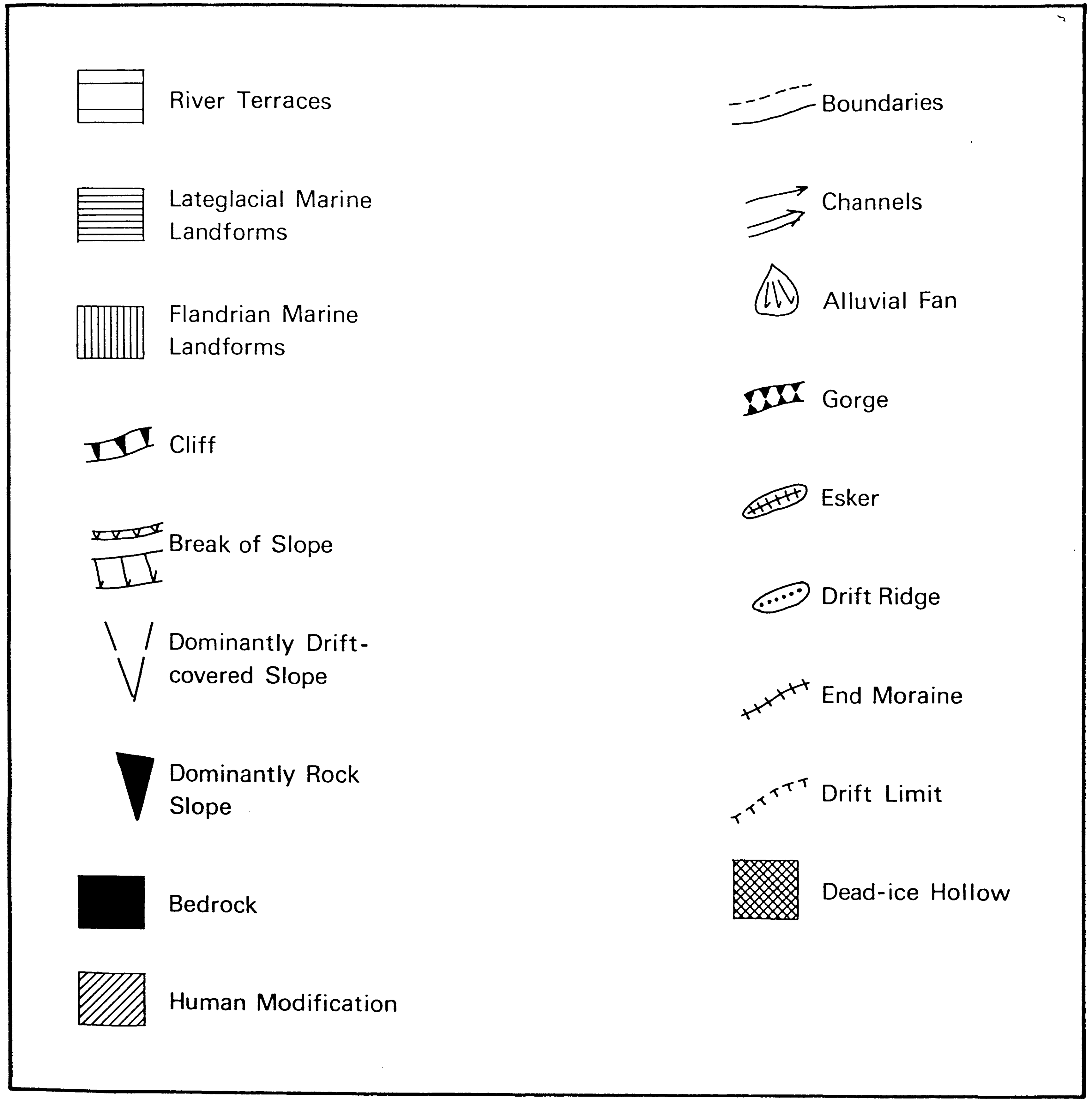
NW from the above features and overlooking Sandbank are massive deposits of sand and gravel related to the mouth of the Allt a' Chromain.

Two levels have been mapped and surveyed on this deltaic deposit, _ca_. 38.0 metres (S395 and S396) and _ca_. 36.5 metres (S394). These features cannot have formed until ice had wasted back from the Loch Loskin/Holy Loch watershed and they therefore post-date the previously mentioned features in this area. Relative sea-level appears to have fallen by _ca_. 1.5 metres during this time but great weight is not placed upon this difference bearing in mind the continuously sloping surface of the Allt a' Chromain delta.

Eastwards from here a small embayment, the inner edge of which is marked by a rock cliff around Hafton House contains a suite of Flandrian gravel beaches. Levelling of these provides the following declining altitudinal sequence: S233 (10.62 m), S232 (8.87 m), S234 (7.53 m), S231 (3.81 m) and S235 (2.76 m). These are all now protected by a shore wall and road and perhaps due to this the well vegetated character of S235 is no evidence of it not being related to marine processes operating at the present sea-level.

The events during and subsequent to deglaciation in the Strath Eachaig/Holy Loch area can be summarised as follows. As ice wasted back into the mouth of the Holy Loch relative sea-level was approximately 39 metres above Ordnance Datum. A small body of dead ice became trapped in the Loch Loskin depression as the ice wasted farther back up the Holy Loch until the mouth of the Little Eachaig valley was ice free, still with sea-level at _ca_. 39 metres. Dead ice that existed at the mouth of Glen Kin was related to this level. Immediately following this period of ice decay little evidence is available as to either ice front positions or sea-levels although raised marine landforms have been identified at _ca_. 36 metres and _ca_. 32 metres. The next evidence of ice margins is the damming of Loch Eck by glacial deposits whilst an end moraine circles the end of Corarisk Glen. The Loch Eck ice limit is related, through a system of outwash terraces to a sea level at least as low as 21 metres and possibly lower than 13.5 metres whilst the Corarisk Glen end moraine is related to a sea level below 14 metres. At this time both the Little Eachaig and the Eachaig rivers were building separate deltas into an extended Holy Loch. Loch Eck is believed to have attained its present level as it was deglaciated. Another hiatus in the evidence exists between the disappearance of glacier ice from the valleys and the deposition of Flandrian marine and fluvial landforms. Six distinct levels of marine deposition can be identified in the valley, each separated from the other by at least 1 metre: 12.4 metres, 10.9 - 10.6 metres, 8.1 - 7.7 - 7.5 metres, 6.7 metres, 5.5 - 5.2 metres and 3.4 - 3.8 metres, whilst other poorly-developed fragments exist at 13.4 metres, 8.6 - 8.9 metres, 8.2 metres, 4.4 metres and 2.8 metres. River terraces in Strath Eachaig can be related to three of these levels. The Little Eachaig and the Eachaig deltas coalesced while sea-level fell from ca. 9 metres to ca. 6 metres and a radiocarbon date indicates regression of the sea from the 7 metre level to have occurred 3700 - 3900 years B.P. Since this time there has been an emergence of ca. 3.7 metres giving an average rate of emergency of 1 metre/1000 years.

Figure 20: Geomorphological map of the mouth of Glen Croe.


River Terraces
Lateglacial Marine Landforms
Flandrian Marine Landforms
Cliff
Break of Slope
Dominantly Drift-covered Slope
Dominantly Rock Slope
Bedrock
Human Modification
Boundaries
Channels
Alluvial Fan
Gorge
Esker
Drift Ridge
End Moraine
Drift Limit
Dead-ice Hollow

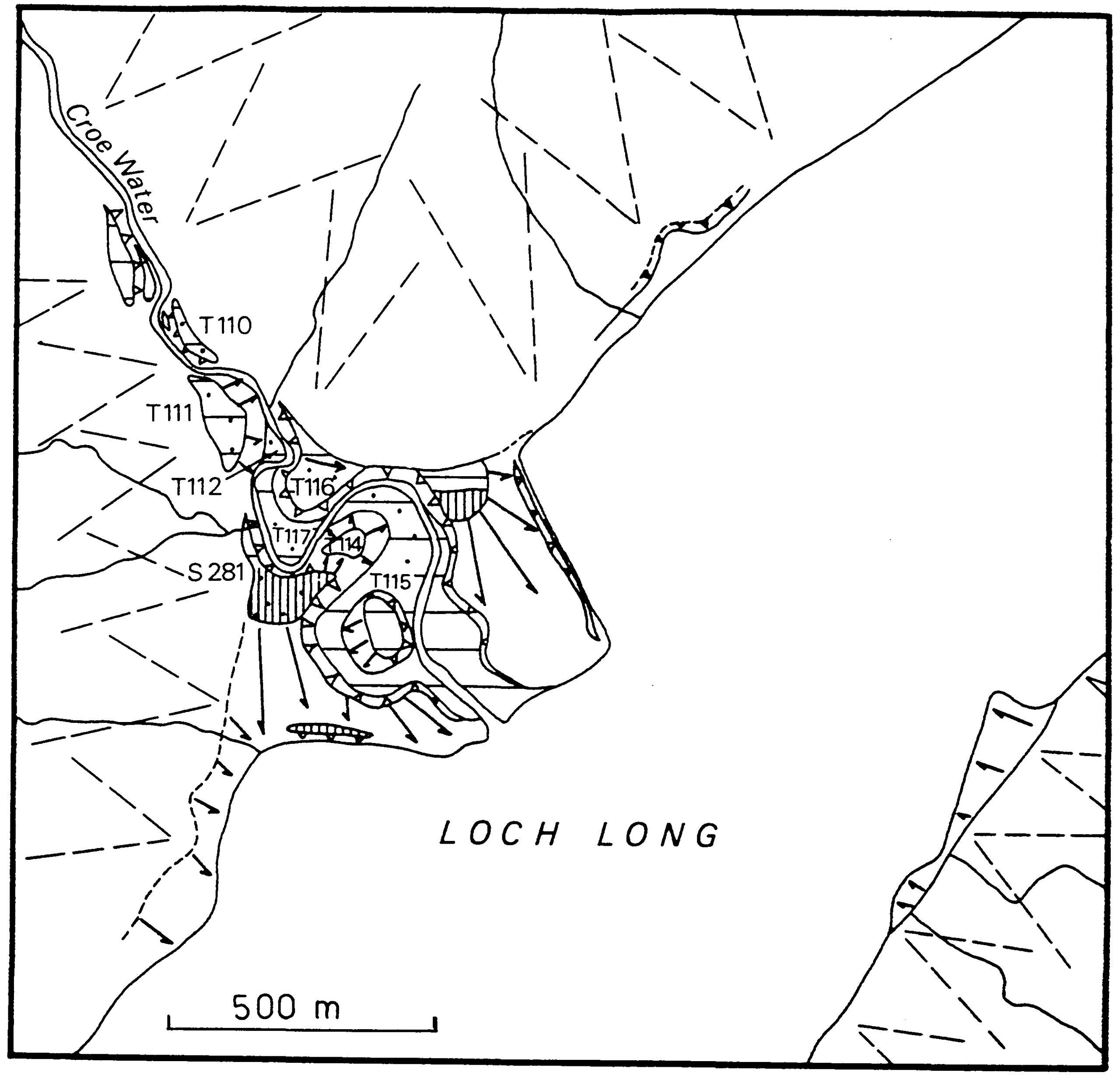
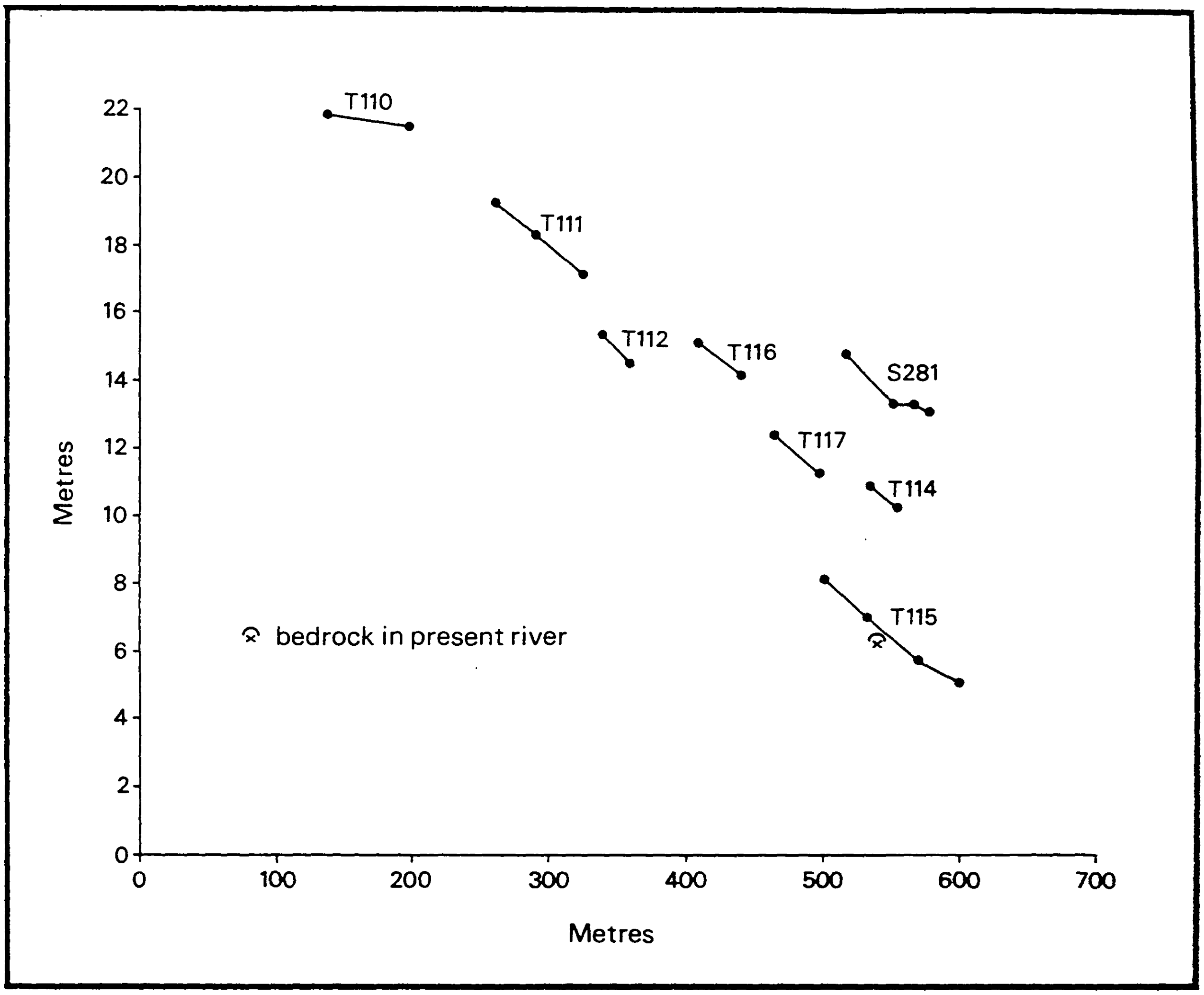

Croe Water
T110
T111
T112
T116
T117
T114
S 281
T115
LOCH LONG
500 m

Figure 21: Height-distance diagram of the terraces at the mouth of Glen Croe.



T110
T111
T112
T116
S281
T117
T114
T115
bedrock in present river
Metres
Metres
0
2
4
6
8
10
12
14
16
18
20
22
0
100
200
300
400
500
600
700

Figure 22: Geomorphological map of the head of Loch Long.

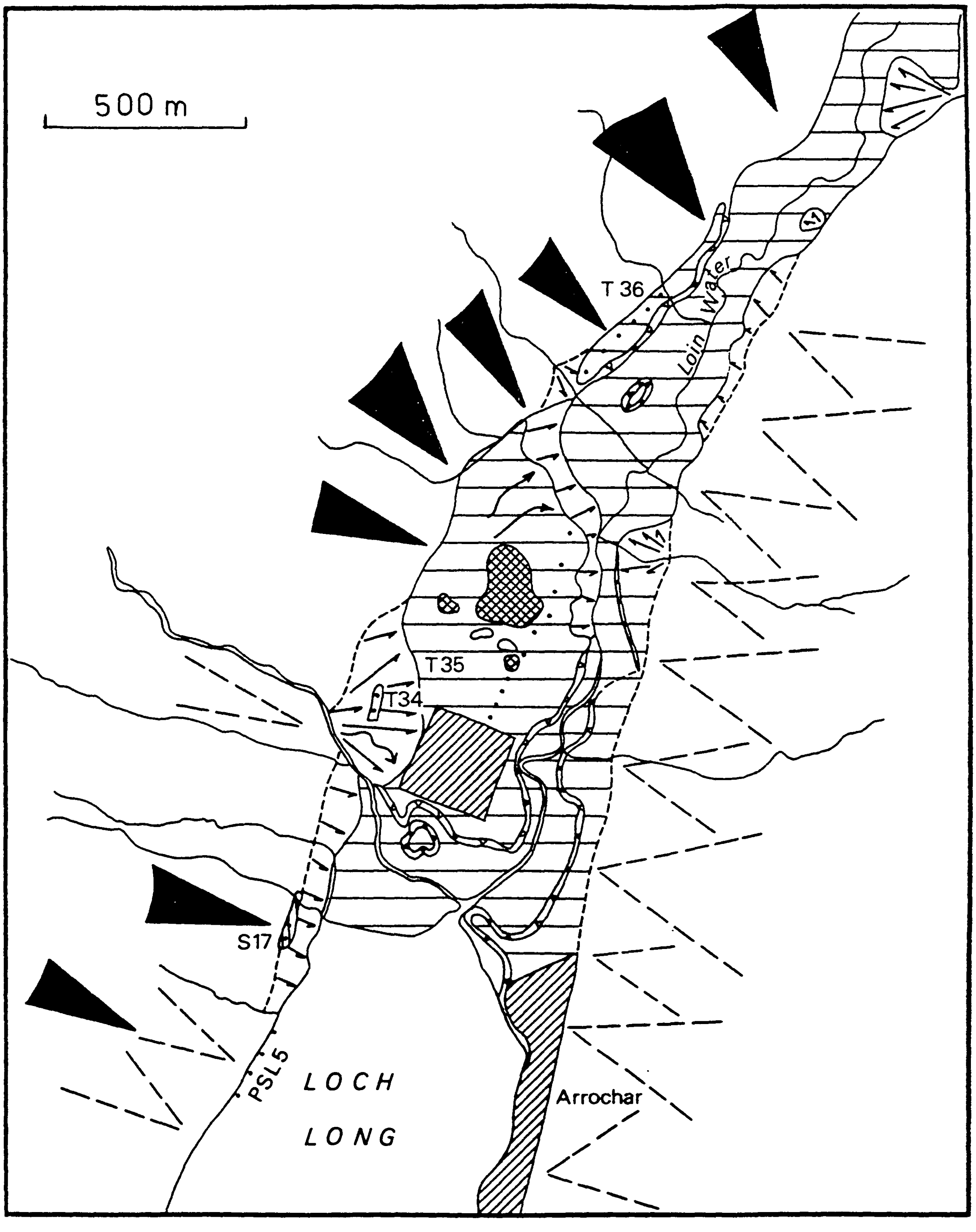
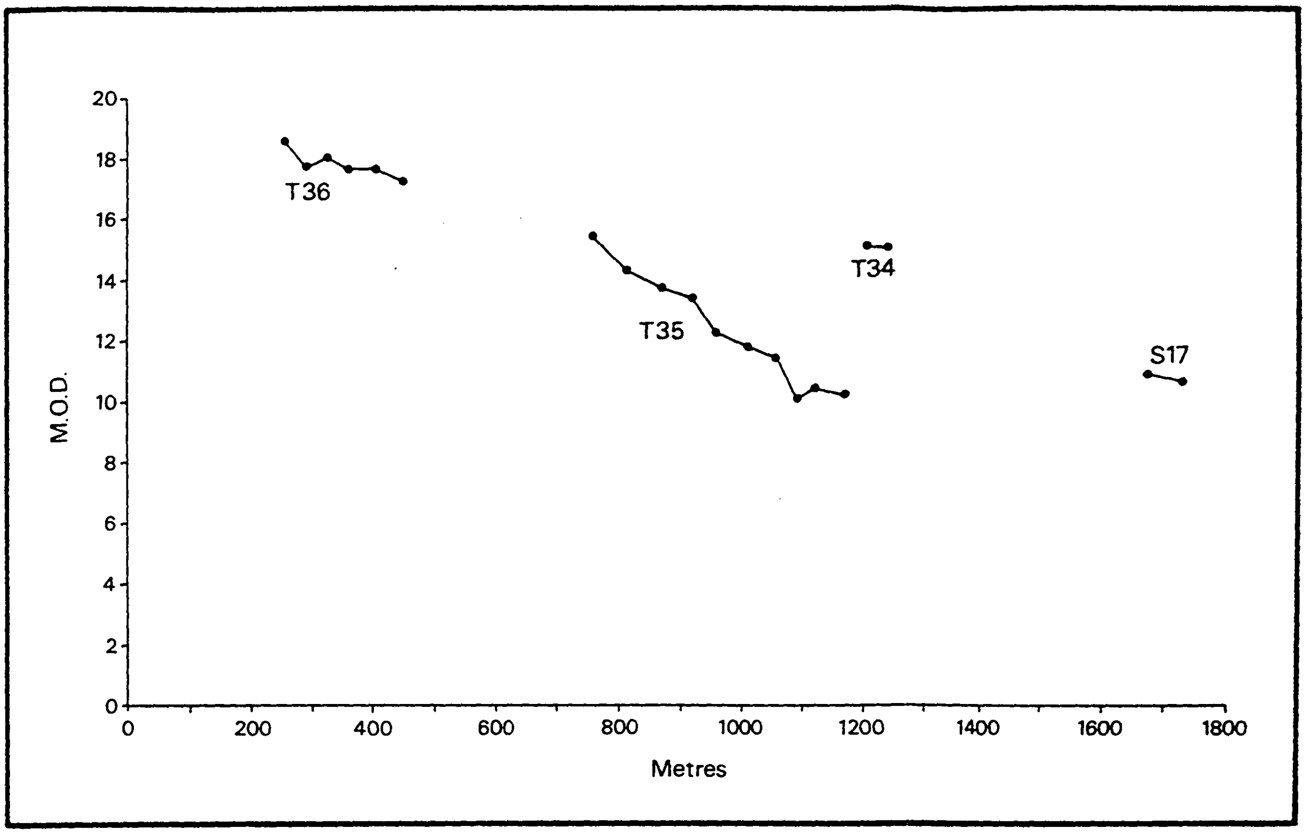
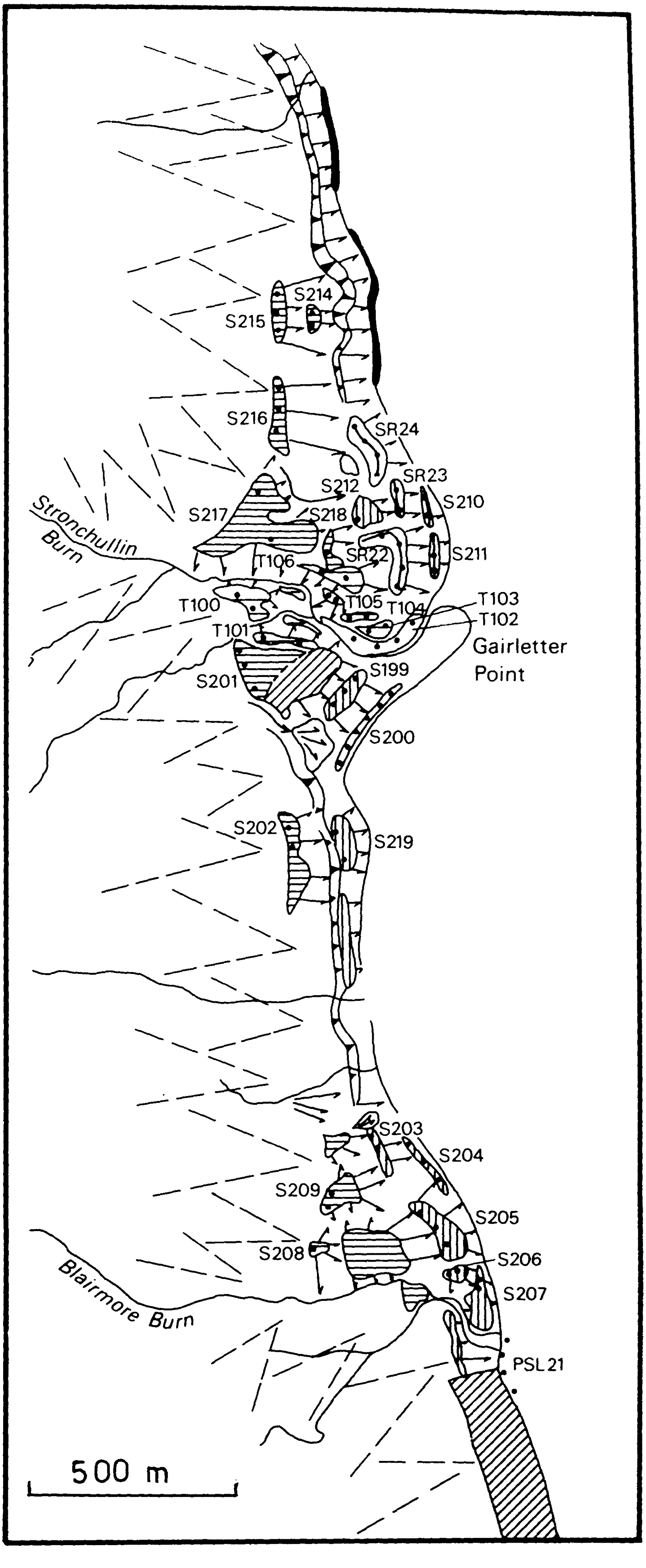

500 m
T 36
Loin Water
T 35
T34
S17
PSL5
LOCH
LONG
Arrochar

Figure 23: Height-distance diagram of the terraces at the head of Loch Long.



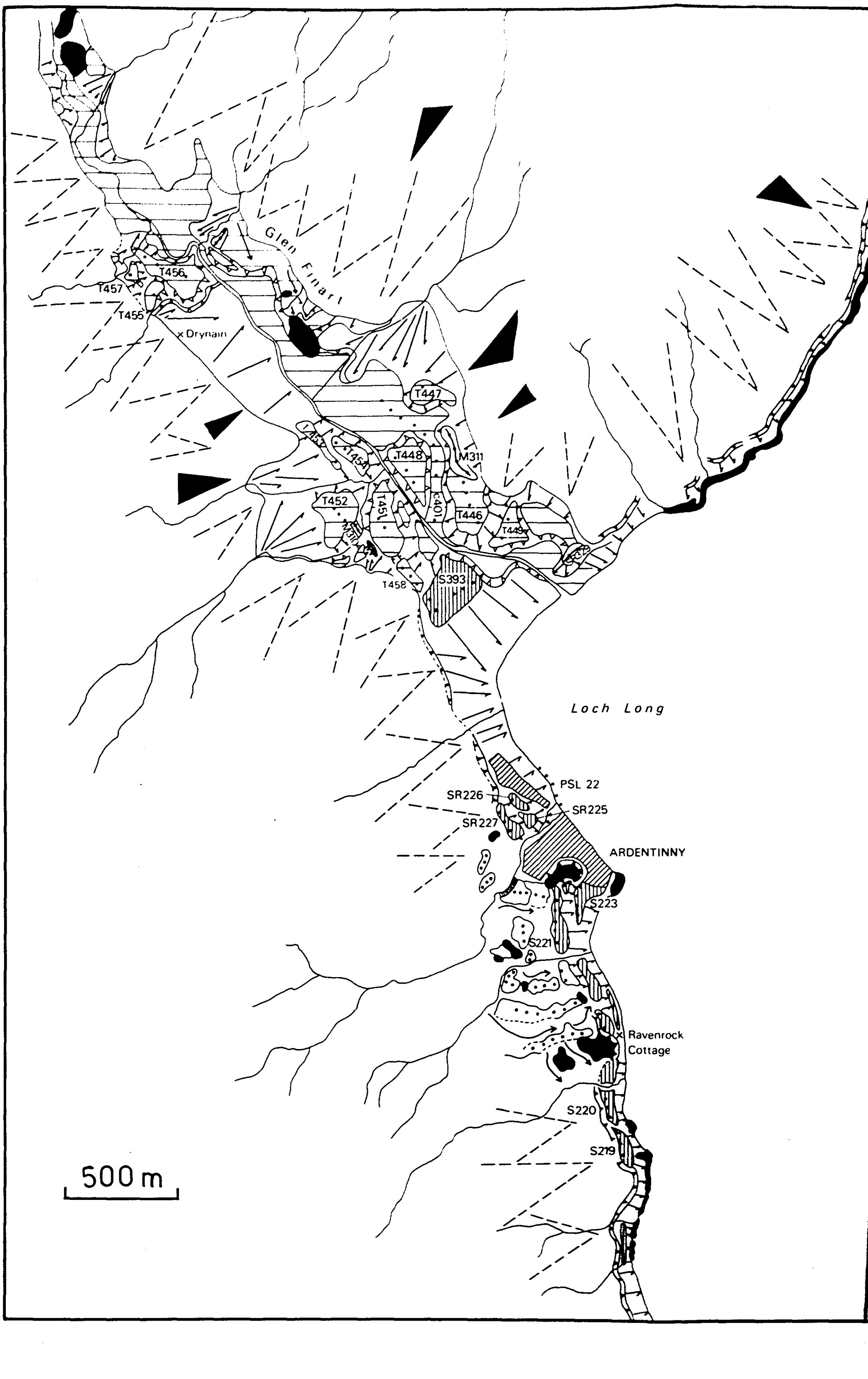
M.O.D.
20
18
16
14
12
10
8
6
4
2
0
0
200
400
600
800
1000
1200
1400
1600
1800
Metres
T36
T35
T34
S17

Figure 24: Geomorphological map of Blairmore/Gairletter area.



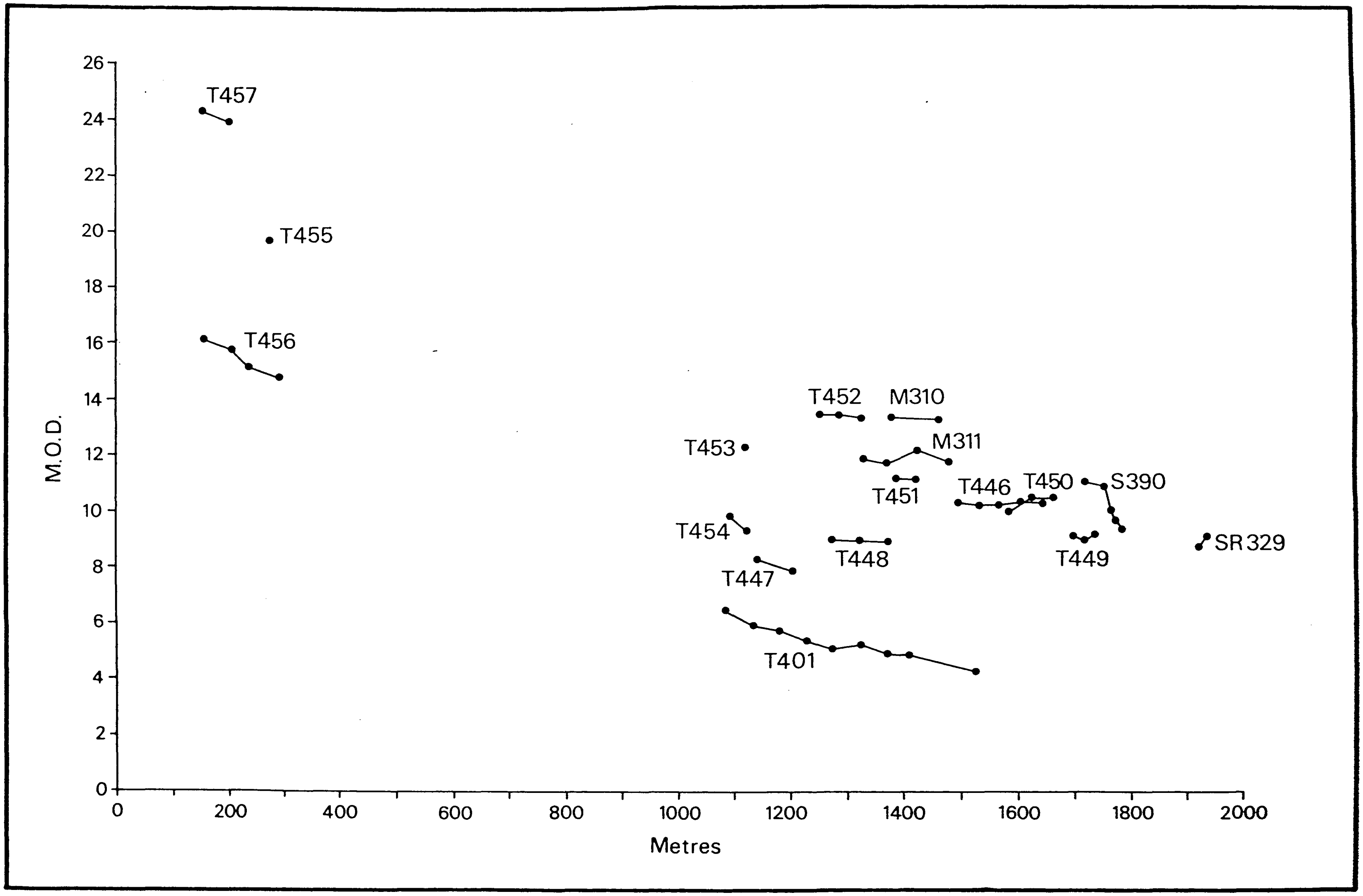
S214
S215
S216
SR24
SR23
S212
S210
S217
S218
Stronchullin Burn
SR22
S211
T106
T103
T100
T105
T104
T102
T101
Gairletter Point
S199
S201
S200
S202
S219
S203
S204
S209
S205
S208
S206
S207
Blairmore Burn
PSL 21
500 m

Figure 25: Geomorphological map of Ardentinny/Glen Finart.



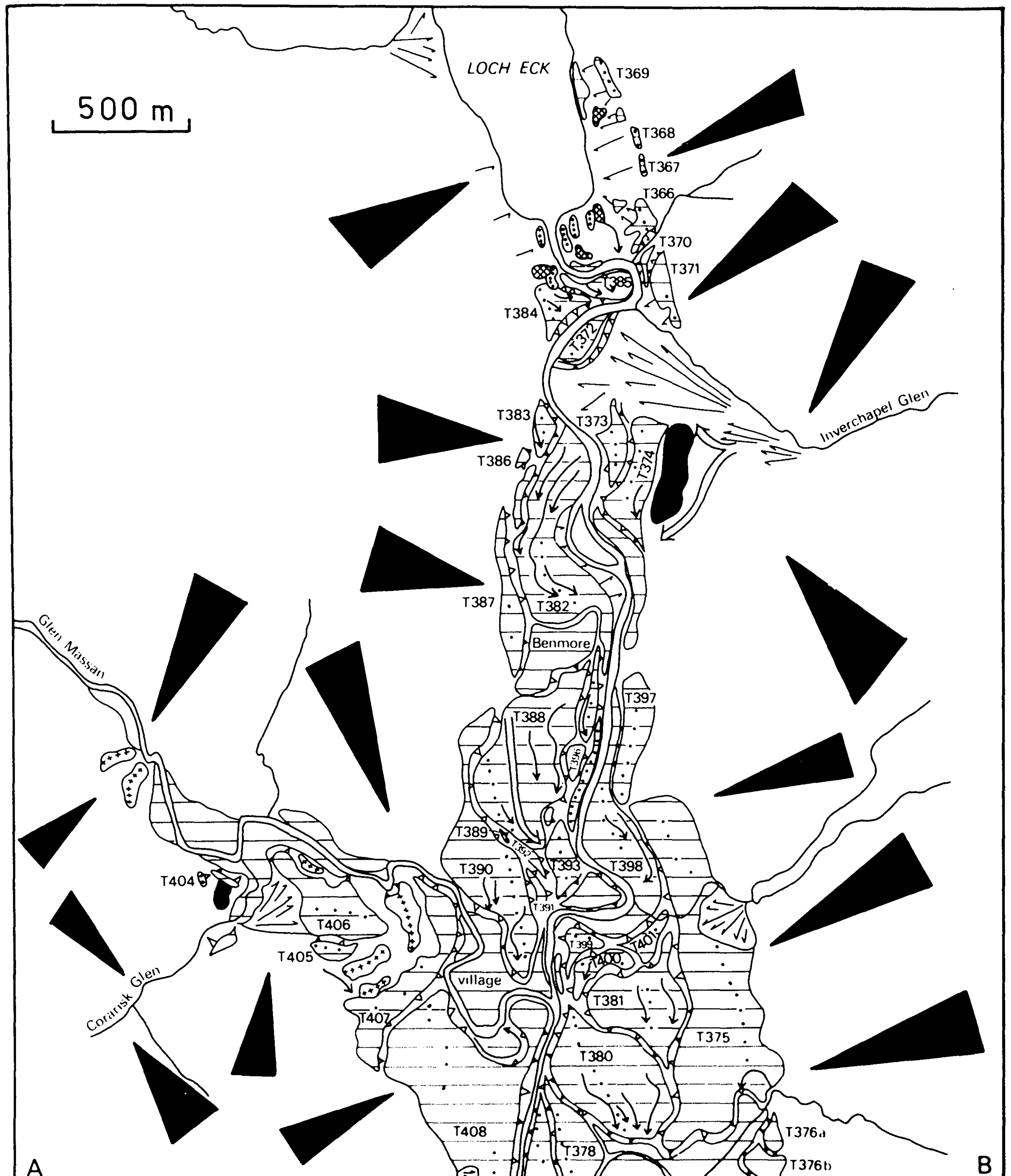
Glen Finart
T456
T457
T455
x Drynain
T447
T453
T454
T448
M311
T452
T451
c401
T446
T449
M310
T458
S393
Loch Long
PSL 22
SR226
SR225
SR227
ARDENTINNY
S223
S221
x Ravenrock Cottage
S220
S219
500 m

Figure 26: Height-distance diagram of the terraces in Glen Finart.



26
24
22
20
18
16
14
12
10
8
6
4
2
0
M.O.D.
0
200
400
600
800
1000
1200
1400
1600
1800
2000
Metres
T457
T455
T456
T452
M310
T453
M311
T446
T450
S390
T451
T454
T448
T449
SR 329
T447
T401

Figure 27: Geomorphological map of Strath Eachaig seawards of Loch Eck.


500 m
LOCH ECK
T369
T368
T367
T366
T370
T371
T384
T372
T383
T373
T386
T374
Inverchapel Glen
T387
T382
Benmore
T397
T388
T389
T393
T398
T390
T391
T400
T381
village
T375
T380
T408
T378
T376a
T376b
Glen Massan
T404
T406
T405
T407
Glen
A
B

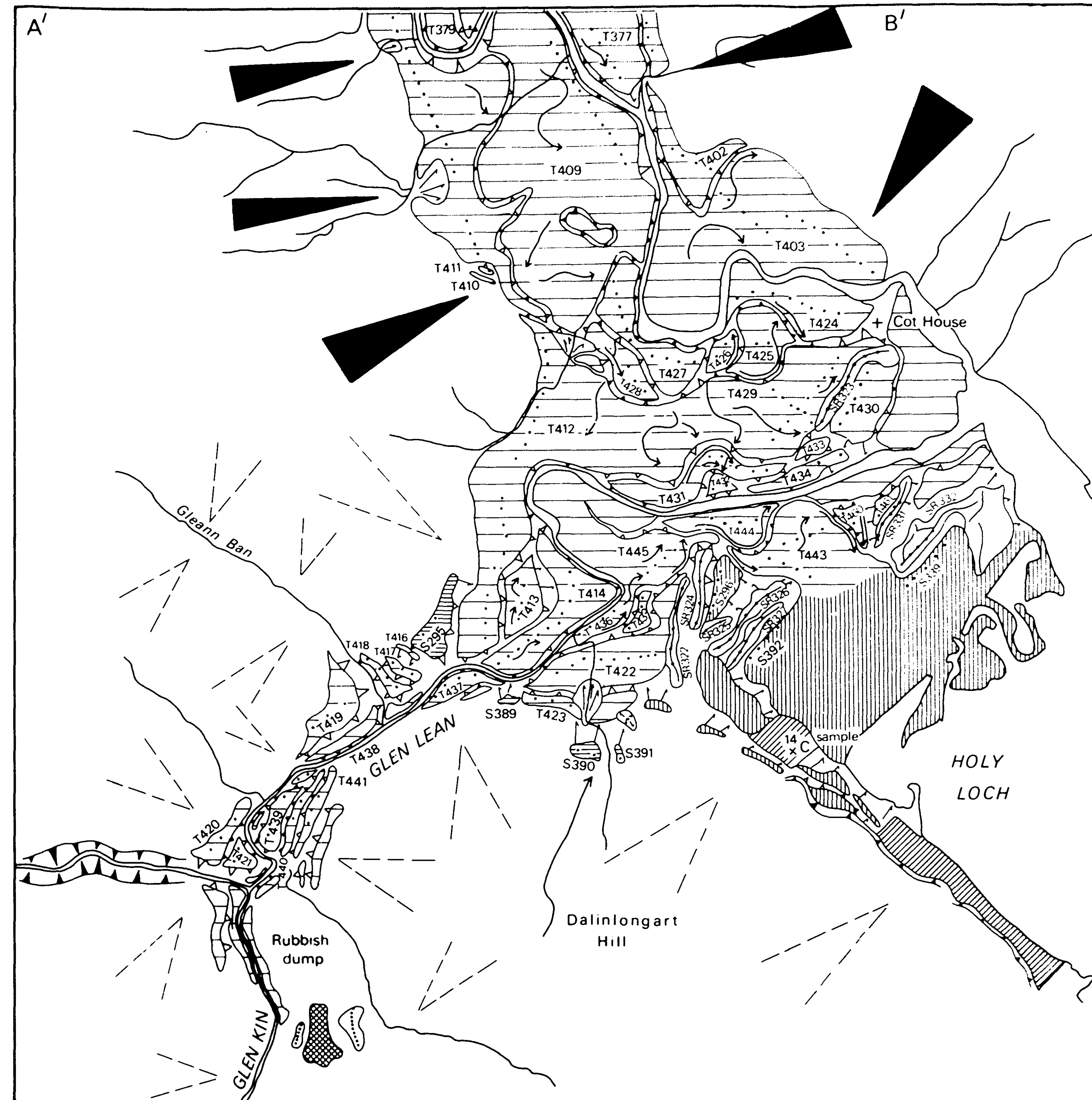
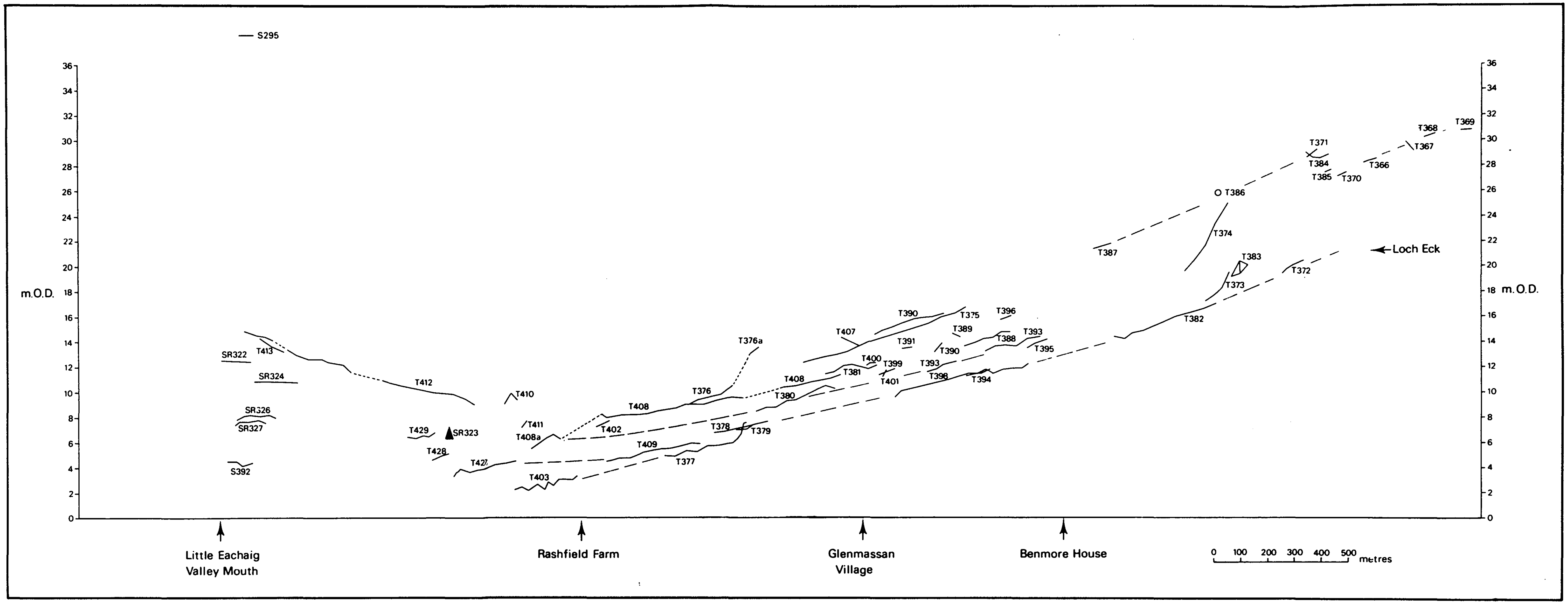
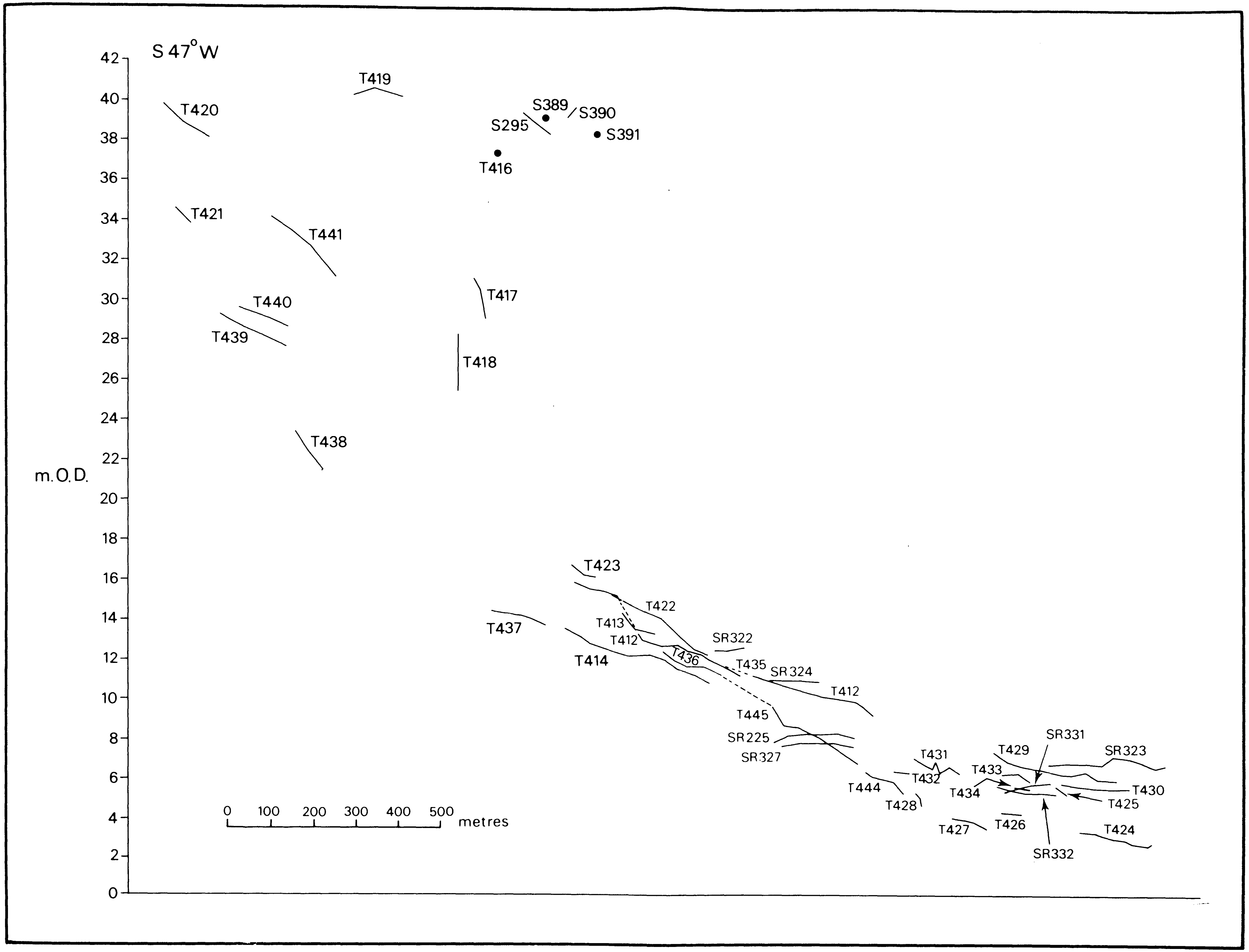

A′
B′
T379
T377
T409
T402
T403
T411
T410
T424
Cot House
T427
T425
T429
T412
T430
T431
T434
T444
T443
T445
T414
T413
Gleann Ban
T418
T416
T422
T437
T419
S389
T423
T438
GLEN LEAN
T441
S390
S391
14C sample
HOLY LOCH
T420
T439
T421
Rubbish dump
Dalinlongart Hill
GLEN KIN

Figure 28: Height-distance diagram of the terraces of lower Strath Eachaig.



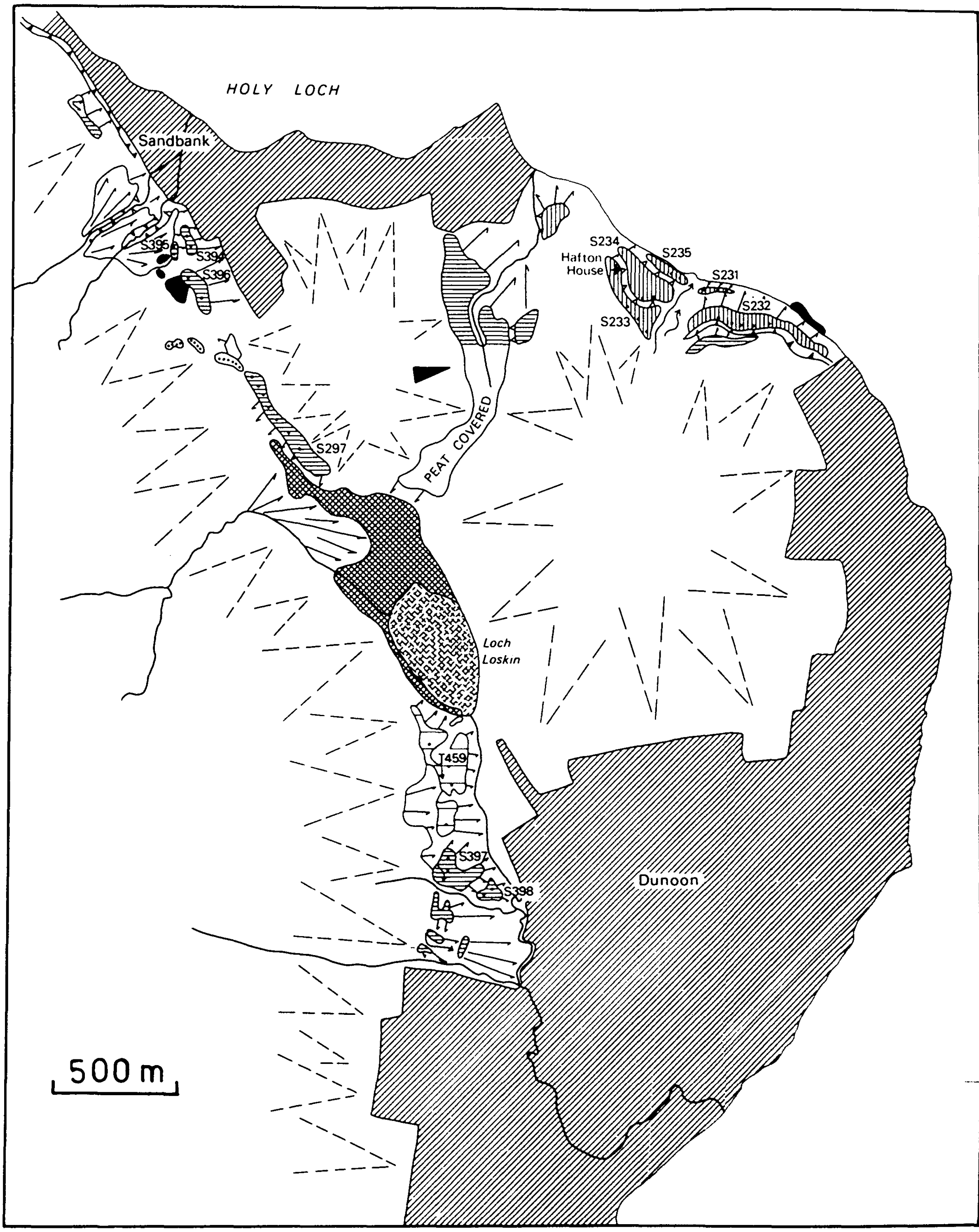
S295
m.O.D.
Loch Eck
Little Eachaig Valley Mouth
Rashfield Farm
Glenmassan Village
Benmore House
0 100 200 300 400 500 metres

Figure 29: Height-distance diagram of lower Glen Lean/Strath Eachaig.



S 47°W
m.O.D.
42
40
38
36
34
32
30
28
26
24
22
20
18
16
14
12
10
8
6
4
2
0
T419
T420
S389
S390
S295
S391
T416
T421
T441
T417
T440
T439
T418
T438
T423
T422
T437
T413
T412
T414
T436
SR322
T435
SR324
T412
T445
SR225
SR327
T444
T428
T431
T432
T434
T429
SR331
T433
SR323
T430
T425
T427
T426
SR332
T424
0
100
200
300
400
500
metres

Figure 30: Geomorphological map of the Holy Loch/Loch Loskin area.



HOLY LOCH
Sandbank
S395
S394
S396
S297
PEAT COVERED
S234
Hafton House
S235
S231
S232
S233
Loch Loskin
T459
S397
S398
Dunoon
500 m

CHAPTER 6

FIELD DESCRIPTION OF AREAS AROUND LOCH STRIVEN AND LOCH RIDDON

1. Ardyne

There is a remarkably well preserved series of marine and fluvial landforms between Ardyne and Toward Points (Fig. 31). Ardyne Point is now the site of McAlpine's Oil Production Platform Construction Yard which was developed in the intervening period between field mapping and levelling, resulting in a number of shingle ridges being destroyed before they could be surveyed. Unfortunately permission was not granted to visit the yard during its construction nor were site investigation reports made available for consultation with the result that much undoubtedly valuable information has been missed. In part this has been offset by the visit to the site made by Dr. J.D. Peacock of the Institute of Geological Sciences to sample marine faunas and recover material for radiocarbon dating. His prime interest, however, was palaeontological rather than morphological. In addition it was not possible to map within the grounds of the City of Glasgow Residential School of Castle Toward (NS 115682)

The major suite of landforms in the area is the terrace sequence related to the Ardyne Burn. These terraces are developed south of Knockdow House where Glen Fyne opens out on to a coastal apron. This area of fluvial and marine deposition is interrupted west of Mid-Ardyne Farm by a streamlined, rock-cored, till-clad mound oriented approximately N-S formed by ice flowing out of Loch Striven. On the W it is cut into by a rock cliffline that can be traced continuously northwards into Loch Striven; southwards the cliffline becomes covered by marine gravels but reappears E of Castle Toward. Eastwards of this again the cliff crosses the Highland Boundary Fault from the Dalradian rocks on to Old Red

Sandstone where the accompanying rock platform attains its maximal development in the whole of the study area, being almost 700 metres broad at Toward Point.

The highest evidence of marine activity in the area is feature S380 at an altitude of 38 metres. A marine origin is inferred from its distinct morphology: it stands out as an over 200 m-long horizontal terrace in sharp contrast to the steeply sloping river terraces to the NW of it. Although cut into by a small stream the deposits of the feature are not well exposed near its surface, though a fawny till can be seen more than 10 metres downslope. The possibility of its being an ice-marginal delta of the type described at the head of Loch Striven (see below) must be high, especially as it is not related to any obvious source of material.

The extensive system of terraces can conveniently be divided into two sets. A lower, inner set of smaller terraces is separated by an abrupt slope from an upper set of larger and wider terraces. The higher terraces are examined first.

It is useful to make a number of initial points. The ice-moulded hill to the W of Mid-Ardyne has acted as a major control on the direction of drainage and two routes to the sea have been available for the Ardyne Burn. Thus terrace fragments T338, T344 and T345 and channel C300 lie to the N of this obstruction whilst the other major terraces are to the E. As will be seen later these routes to the sea were used both contemporaneously and alternately. The height-distance diagram (Fig. 32) has been drawn on a N-S projection plane and this results in exaggeration of the gradients of those terraces or fragments of terraces that occur

N of the hill and slope in a SW direction. Initial inspection of the height-distance diagram shows that the profile of the present river floodplain is non-linear. Straight-line correlations of former river terraces need not therefore be expected.

No direct evidence of former sea-level can be found between S380 at 38 m and the shingle ridges at 13.5 m, the highest of which are SR313 and SR314. However, terrace fragments do exist at intervening altitudes and of these T354, T355, T356 and T346 are the uppermost and hence earliest formed. T355 and T356 are affected by streams that descend the hillside and their altitude of ca. 1 m above T346 need not be seen as indicative of an earlier origin. T346 was not surveyed throughout its length owing to crops and its correlation with T354 is not as certain as it might have been. These terrace fragments suggest a period of drainage to the E of the hill. The existence of a system of channels related to a nearly-closed shallow depression interpreted as a dead-ice hollow, suggests the presence of decaying ice at the time of formation of T346. No contemporaneous sea-level can be measured but T354 falls to ca. 20 m and sea-level must have been at or below this level.

The next set of terrace fragments is T338, T345 and T357. These are all exceptionally well-developed features and can be followed continuously for considerable distances. T338 starts at the mouth of the rock gorge just N of Knockdow and is the first terrace to show evidence of drainage N of the Mid-Ardyne hill. It is separated from T345 by channel C300 and the height-distance diagram indicates that T345 and T338 correlate. T345 appears to have been related to drainage both to the N and to the E of the hill and it is this latter line of drainage that is correlated with T357. The drainage to the N of the hill which is represented by T338, T345, C300 and T344 presents considerable problems of interpretation. These terraces have regular SW sloping gradients which show no signs of flattening out before they are truncated by the cliff previously mentioned. The lowest points measured on the surfaces of T338, T345, T344 and C300 are 31.3 m, 26.3 m, 25.3 m and 21.3 m respectively. Of these the altitude measured in the bottom of the channel must be regarded with some caution due to possible later modification; however, the hanging nature of the mouth of the channel, lack of a coherent stream in it and the small size of the alluvial fan on the former beach deposits below in comparison to other nearby streams, suggest little subsequent erosion of the bed of the channel. These features clearly postdate the formation of terrace T346 and its correlative T354. They must therefore relate to a sea-level of less than 20 metres. However, if terrace fragment T338, for example, is projected in a W or SW direction at the gradient established along its measured length it would necessitate postulation of a considerable delta that has been subsequently eroded to have been built out into Loch Striven (which is up to ca. 15 metres deep at this locality). To minimise this problem, a down-valley steepening of the terrace gradient would need to be envisaged, yet in all profiles a lessening of gradient is observed. The progressive lowering of the altitudes of the ends of the terraces and the channel may be explained, particularly in the case of T338 and T345, as a function of greater distance down-terrace to the point where they are truncated, yet this cannot be the case for T344 and C300 as these are nearer the mouth of the gorge than is the end of T345. It is therefore suggested that these terraces were formed in response to a falling sea-level. At its highest point channel C300 is rather shallow and T345 and T338 correlate quite easily across it. The altitude, therefore, of the source of the terrace materials was relatively constant over the period of development being discussed. Terrace T345 has an extension down-valley to the E of the Mid-Ardyne outcrop. The relationship of T345 to the well-developed T347 that occurs below it suggests that T345 correlates with T357. Such a correlation implies that this part of the terrace sequence is related to a sea-level at or below 15 m, to which altitude T357 can be traced. Subsequent to the development of an outlet S to T357 all drainage has been to the S. The westward period of drainage of which the last representatives are T344 and C300 were penecontemporaneous with T357, yet they must have been abandoned while the Ardyne Burn still occupied the major portion of T345 as is shown by the smallness of the incision of C300 into the upper part of T345.

This interpretation of the development of these terrace fragments carries with it the corollary of a considerable period (or periods) of marine erosion to the W subsequent to development of these features and culminating in (or re-exposing) the rock-cut cliff mentioned above. The general problem of extensive post-glacial marine erosion is discussed in Chapter 8.

The next feature on the height-distance diagram is terrace T347. This is the most extensively developed terrace fragment in the Ardyne area and its gradient has been established at a total of 26 points. It is correlated with terrace fragment T339 and it has a clear concave-upwards profile, though this seems to consist of long linear segments, perhaps due to the influence of a number of burns that cross it at a variety of angles. However, the overall regularity of the feature suggests that these were not of major consequence in modifying it. At its seaward end T347 abuts on to shingle ridge SR313 (13.47 m). The terrace surface is, however, slightly higher than the crest of the ridge indicating the ridge to be related to a lower sea-level than the terrace.

The terraces of the Ardyne Burn have been developed seawards of the confines of the rock-cut gorge N of Knockdow. The upper series of terraces discussed above commence at a high level above the floor of the gorge and it is thought that the gorge was buried in drift at this time. An exposure of grey till in the side of the gorge supports this idea. After the formation of terrace fragments T347 and T339 no well-developed terraces have been formed until the lower set of terraces. It appears therefore that there was a period during which the Ardyne Burn incised its course and re-exhumed the rock-cut gorge. Terrace fragments T340, T341 and T343 are the only features that relate to this period of incision and are thought to be erosional remnants of the former valley fill. The lower set of terraces is divergent down-valley and converges up-valley on the mouth of the gorge. The development of these terraces can thus be explained best in terms of sea-level changes. Therefore, before analysing the remainder of the terrace sequence, it is necessary to assess the direct evidence that exists for sea-level changes at the mouth of the Ardyne Burn.

At the mouth of the Ardyne Burn there are abundant former marine landforms, whilst an additional number of fragments exists between Ardyne and Toward Point. Many of these latter features occur between the abandoned cliff and the present shore where bedrock is exposed implying that they rest in part on the rock platform that fronts the cliff. In such cases only those fragments that were seen to have distinct constructional forms were heighted. Near Ardyne Point, the raised marine deposits have been built-up against the cliffline until it has been completely buried as by S377. Together with SR313 and SR314, it forms the highest group of these features with altitudes of 13.51, 13.47 and 13.56 m respectively. Sections examined in S377 and SR314 show them to be composed of horizontal, poorly-bedded, well-sorted medium gravels whose individual particles apparently have a horizontal attitude. Up to 2 metres of these gravels were noted in an exposure in S377. Other sections in the locality revealed that similar gravels with occasional lenses of sand make up all the marine landforms. These sediments are characteristic of most of the Flandrian marine deposits in the study area and Peacock et al. (1978) have described a typically Flandrian marine fauna from these deposits at Ardyne.

The majority of the raised marine landforms around Ardyne Point are shingle ridges, separated by clear hollows and often branching and recurving. The genesis of all the marine landforms is not always quite so straightforward as with S377 which is interpreted as a shingle ridge that has been built against the bedrock, hence losing its ridged appearance and gaining that of a terrace. Its proposed origin is based upon the clear height and sedimentological relationship with SR314 and SR313. Shingle ridge SR315 (9.48 m) shows a similar characteristic in its westward development. It is suggested that such a characteristic of shingle ridges is the result of marine transgression (i.e. landward migration of the shoreline), yet due to the fact that these features are shingle ridges, this cannot be generalised to a statement of relative sea-level change since such landforms can migrate due to changes in sediment supply, dominant wind direction or tidal regime as well as changes of sea-level. Again, a still-stand of the sea or a small regression/transgression cycle may be indicated by the almost identical altitude of SR316 (7.89 m) and SR317 (7.96 m), for one is ca. 100 m seawards of the other, but the nature of the evidence precludes a definite statement on this point. The sensitivity of these landforms to local changes in the littoral zone is illustrated by the disappearance of a spit that was some hundreds of metres long from the W of the mouth of the Ardyne Burn between the time of production of the original 1:10,560 Ordnance Survey Map of this area (1869) and the present day. (Over the same time period, a spit of even greater magnitude has been built up at Auchalick Bay on Loch Fyne.). The only remaining evidence of this spit when the area was visited in 1972 was lagoonal peat that is now being eroded on the foreshore. Similar peats may be expected to be the basal deposits in many of the inter-ridge hollows inland from this point.

It is difficult to integrate all these diverse features into discrete periods of shoreline formation. The rigorous analysis that is necessary to do so, taking into account origin of features, degree of preservation, relationship to fetch and isostatic warping is attempted in Chapters 9 and 10. For present purposes it is sufficient to list the fragments and their altitudes and note some specific groups. The highest level of Flandrian deposition seems well established by features S377 (13.51 m), SR313 (13.47 m), SR314 (13.56 m) and S386 (13.04 m). This is followed by two isolated fragments SR312 (12.54 m) and T353 (11.3 m) and then another group made up of S381 (10.05 m), SR311 (9.87 m) and SR315 (9.48 m). S376 (8.51 m) occurs on its own, but SR317 (7.96 m), SR316 (7.89 m) and S378 (7.55 m) exist as a group and a similar coincidence of SR310 (6.44 m), S383 (6.54 m) and S387 (6.60 m) may be observed. The sequence is continued seawards with SR321 (5.39 m) and S382 (4.97 m) then SR309

(4.44 m) and S379 (4.49 m) and finally SR318 (3.77 m). It is now possible to examine the changes in the terrace system that have resulted from these changes of sea-level.

Of the lower set of terraces the main fragment is T348. Correlation upvalley with terraces T351 and T342 gives a smooth concave-upwards profile, which when projected seawards intersects the uppermost Flandrian marine deposits. Because of this and because of the distinct nature of the terrace in the general terrace sequence, it is considered to be contemporaneous with the highest shingle ridges.

Subsequent to this event, the net regression of the sea that has been recorded in the marine landforms is reflected in terrace levels being developed at successively lower altitudes. It seems possible to reconstruct two particular terraces with T364, T362, T361 and the end of T349 relating to the shoreline represented by S381, SR311 and SR315 occurring at approximately 9.5 - 10.0 m and fragments T359, T360 and possibly T350 correlating with the 7.5 - 8.0 m shoreline represented by SR317, SR316 and S378. The concave-upwards form of the present floodplain assists in correlating these various fragments and it is thought significant that the separate terraces relate to particular groups of marine landforms.

The history of sea-level changes in the Ardyne Point area can be summarised as follows. Immediately upon deglaciation sea-level was at an altitude of ca. 38 m. As ice wasted into Glen Fyne sea-level was falling and a system of outwash terraces was developed. Reconstruction of the terrace system suggests that sea-level had fallen to at least 20 metres whilst the uppermost terrace was forming. Continued fall of sea-level and terrace response to this can be traced to a sea-level of below 14 m. A period of marine erosion ensued during which a rock-cut cliffline, extensively developed throughout the area was either re-exhumed, modified, or eroded. Subsequently Flandrian marine deposits were built to 13.5 m. The ensuing regression has left traces of several distinct periods of marine landform development and three of these periods of shoreline formation have been related to river terraces.

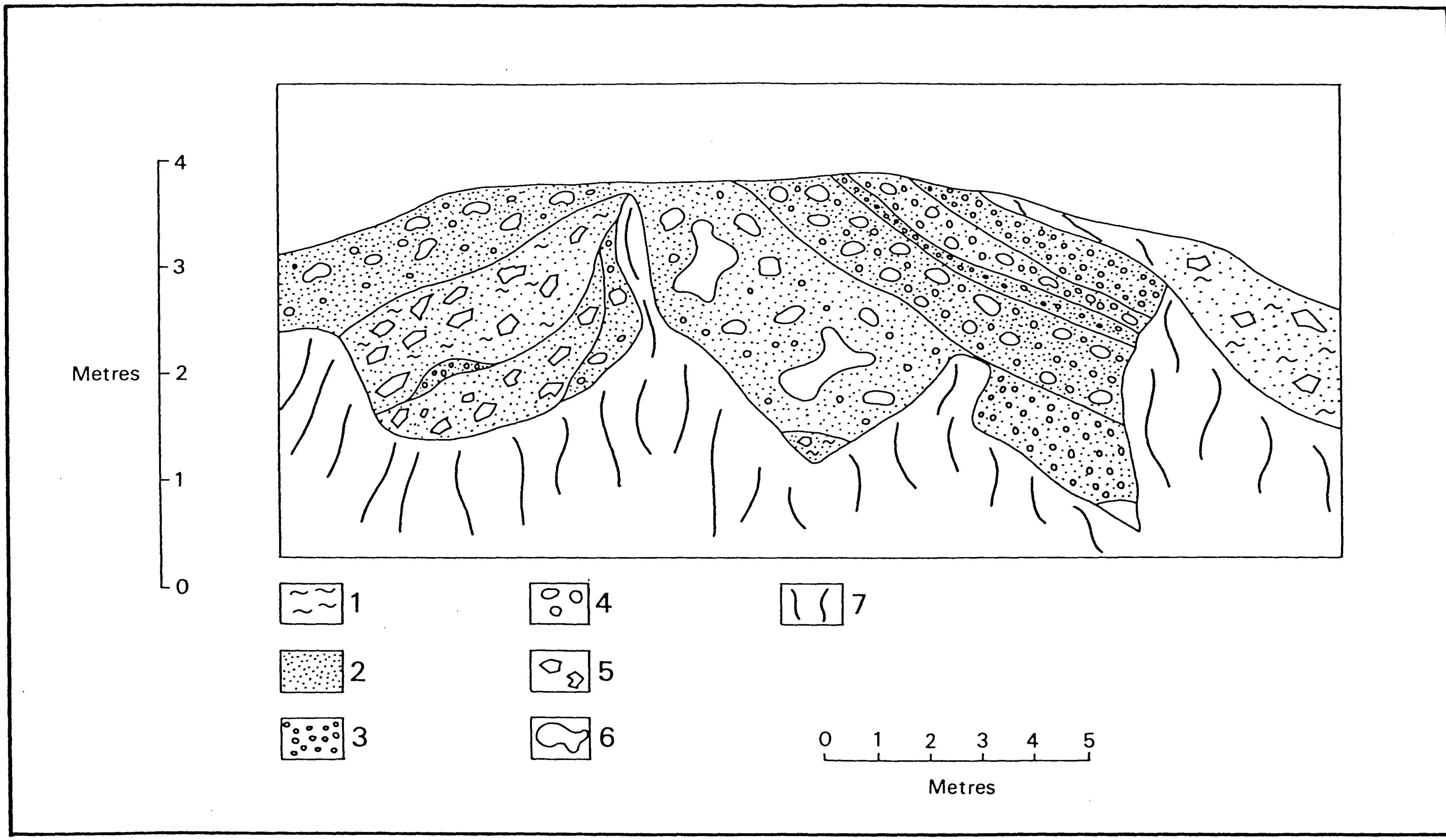
2. The Head of Loch Striven

Beyond the head of Loch Striven the valley continues for only 2 km where it divides into two steeply rising valleys. Gleann Laoigh lies to the E and reveals evidence of glacial erosion in small rock basins and valley steps. The smaller valley to the W is a very sharply incised rock gorge. The gorge has an intake at ca. 230 m and in a distance of ca. 1 km falls to less than 30 m. 3 km farther S Loch Striven is joined by Glen Tarsan, one of its major tributary valleys. The deposits and landforms of interest lie in the area along the E side of Loch Striven between the mouths of Glen Laoigh and Glen Tarsan.

A suite of terraces occupies the valley floor (Fig. 33). Mapping and instrumental levelling of these enable construction of the accompanying height-distance diagram that is drawn on a N-S projection plane. (Fig. 34). One of the most significant features of the valley is a set of flat-topped mounds banked against the valley side. These mounds have limited downvalley extent but rise abruptly up to 20 m above the terraces on the valley floor. The surfaces of four of them (S75, S76, S77, and S78) were levelled and plotted on Figure 34. Another (a, Fig. 33) is rock buttressed but the occurrence of gravel beneath a thin layer of peat on its flat top suggests the surface to be constructional rather than erosional. A sixth mound (b, Fig. 33) has been largely destroyed by quarrying. The section in the quarry is shown on Figure 35. Although it is complex the close association of till, fluvial deposits and foreset bedding allied to the fact that the deposit is well removed from present stream channels strongly suggests that it is a glaciomarine delta. A similar interpretation is adopted for fragments S76, S77 and S78, partly based on small exposures of fluvial bedding in S76 and S77 but mainly on their horizontal height relationship and their isolation from present stream channels of sufficient size to supply the necessary sediment. The small portion of the surface of S75 that could be levelled indicates that it is higher and has a steeper gradient than the other terrace fragments. Its form and its proximity to the mouth of the western gorge suggests that it was formed during deglaciation by meltwaters using the gorge when sea-level was at or only slightly lower than the ca. 38 m level indicated by the deltas. At the mouth of Glen Tarsan, sections in a similar well developed terrace (T154) expose foreset bedding. This terrace has been traced downvalley to a height of 37 m, without any signs of a lessening in gradient thus a sea-level of somewhat less than this height can be inferred to have lasted whilst meltwaters drained from Glen Tarsan.

The glaciomarine deltas are the highest evidence of marine action around the head of Loch Striven. The marine limit therefore formed immediately upon deglaciation. Further, they fix 4 successive positions of the snout of the glacier that was occupying Loch Striven at this time and these are shown on the height-distance diagram where they are spaced (to the nearest 25 m) from the most southerly northwards, 500, 625 and 425 m apart, giving an average distance apart of 525 m. If these deposits are annual then a mean rate of retreat of 525 m/yr is indicated which is

Figure 35: Section in quarry at the head of Loch Striven.
1. silt and clay 2. sand
3. rounded gravel 4. rounded pebbles and cobbles 5. angular pebbles and cobbles 6. boulders
7. slumped debris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quite in keeping with historically recorded rates of retreat for calving glaciers in Greenland, such as Academy Gletscher which averaged 333 m/yr between 1920 and 1956 and Spaltegletscher which averaged 600 m/yr over the period 1907 to 1938 (Weidick, 1968).

Little further evidence is recorded of sea-level until the present. At the mouth of Glen Tarsan the surfaces of two small raised marine landforms (S86, S87) were levelled at 6.96 m and 3.85 m respectively whilst to the W of the loch is a veneer of marine gravels with a poorly developed morphology. Within the valley itself none of the river terraces can be linked directly to any sea-level. The largest terrace (T51) is unfortunately badly disturbed by deposits from a landslide. (It is not clear whether the landslide deposits on the terrace are directly due to the landslide or whether they are the products of secondary movements in unstable material: the relative age of the landslide cannot therefore be inferred).

The correlation of the terrace fragments is aided by consideration of the present floodplain. The surface of the floodplain has been levelled on fragments A2 and T48. T44, at a distinct level above the river at its southern end, merges with the floodplain upvalley and is indistinguishable from it at its northern end. The floodplain, on the basis of these altitudes and the altitude of present sea-level, is defined by a concave upwards curve. The remaining terrace fragments seem to correlate on two concave upwards profiles. T44 diverges from the floodplain to correlate with T49, T52 and T55 while an upper sequence is composed of T45, T42, T47, T54(?), T53, and T51. Both these terraces extend below the uppermost level of the Flandrian sea, estimated in this area to have been at <u>ca</u>. 13 m (Chapter 10), and they are therefore Flandrian in age. They cannot be correlated directly with any marine landforms.

3. Glendaruel

(a) Introduction

Glendaruel and its seaward extension, Loch Riddon, form a gently curving valley that is, compared with others of the Cowal Peninsula, relatively wide and open. The major part of the length of Glendaruel is eroded in a SSW direction along the strike of the Green Beds. At Ballochandrain the valley turns to the E and leaves the Green Beds to cut across the Beinn Bheula Schists within which the hills have a considerably more rugged appearance and the valley bottom, prior to reaching Loch Riddon, is interrupted by a number of isolated rock outcrops which on occasion resemble huge roches moutonnees. Loch Riddon is the shallowest of the sea lochs in the study area and reveals large expanses of sand at low tide. This shallowing is due to the considerable amounts of deposits that have been and are being built into the loch.

The point at which the valley turns on the Beinn Bheula Schists corresponds with a pronounced change in morphology (Fig. 36). Below this point there is abundant evidence for sea-level stands higher than the present whilst above it there is none. Large steeply sloping terraces occupy the valley floor upvalley whilst downvalley there is a plethora of small terrace fragments that have been left as the River Ruel winds its way along a course constricted by the large bedrock outcrops.

The projected floodplain (Fig. 37) of the present river steepens markedly below Ballochandrain whilst upvalley it can be considered as a amoothly concave upwards profile. The steepening as revealed on the height-distance diagram is in part due to the projection of the winding river course on to a linear plane. The sinuosity (the ratio of the channel length to the valley length) of the upper valley between Glendaruel House and Ballochandrain is 1.26 whilst that of the valley below Ballochandrain is 1.53. The mean channel gradients over the two sections are 1.40 m/km and 2.44 m/km respectively. The increase in sinuosity over the lower reach would result in an increase in gradient compared with the upper valley of 0.34 m/km. The difference of 1.04 m/km that is observed is therefore only partly due to the increased winding of the river around the rock outcrops and a change in other variables that affect channel gradient must be invoked. In particular it is thought that an increase in the resistance to flow (which is directly proportional to channel slope, Leopold and Wolman, 1957) resulting from the strong geological contrasts, both in solid rock and in drift, between the upper and lower valleys is responsible for this change in gradient.

It is therefore proposed to divide the following discussion into three parts, the first being concerned with the direct evidence available for sea-level changes, the second interpreting the terrace system below Ballochandrain which has been most influenced by these changes and the third deciphering the large terraces and glacial landforms of the major part of Glendaruel where no clear evidence of former sea-levels exists.

(b) Sea-level changes

Around the head of Loch Riddon is a considerable number of flat-topped depositional landforms. Mapping and heighting of these reveal the patterns displayed on Figure 36 and the height-distance diagram (Fig. 37). It is apparent that there is a nearly horizontal upper limit to the distribution of these features comprising S237, S247, S263 and and S248 at ca. 37 m whilst S246, S245 and S252 (probably a composite feature) occur just less than 2 m below this. A number of these features are well exposed, particularly S248 which occupies more than half the valley bottom width at the point where the valley turns from the Green Beds on to the Beinn Bheula Schists. A large sand and gravel pit shows this feature to be composed of very well-bedded sands and gravels, the gravels often being coarse. The beds are mainly dipping steeply downvalley though those nearest the top of the section are mostly horizontal. These are interpreted as foreset and topset beds respectively and the whole feature as a delta. Its position in the valley and the attitude of the beds indicate a source of material that was in the valley, glacier ice being the only reasonable source. Its openness to the sea is taken to indicate that the standing water into which it was deposited was marine and the feature is therefore a glacio-marine delta. Since it is also the highest evidence for marine action within the nearby area it indicates that the marine limit was formed in the presence of glacier ice, a conclusion that has also been reached for the Portavadie, Holy Loch and Loch Striven areas. The inference that an ice margin existed here is supported by an exposure of sands on the opposite side of the valley at approximately the same altitude even though these sands have no distinct morphology nor any obvious source. They are taken to indicate that the delta once stretched across the whole valley.

S245 is a flat-topped feature that joins a rock outcrop to the valley side. It is composed chiefly of small gravel, the individual particles of which appear to have a horizontal imbrication, and well sorted sands were seen as well towards its E end underlying the gravels. This gravel ridge is interpreted as a fossil tombolo, though it surmounts a very high, steep sandy bank on its riverward side and the tombolo is seen only as the surface detail of what could be a glacial deposit. Behind the tombolo there is a peaty hollow, the peat occasionally overlying stoney shell-less massive grey and reddish clays. These clays particularly the reddish ones, are also exposed in the side of S246, a feature that appears capped, though the exposure is poor, by gravel. S245 and S246 are marine features but they do not quite attain the altitude of the marine limit as defined by S248, S237, S247 and S263. Subsequent analysis (Chapter 9) indicates them to relate to a lower shoreline than the higher features.

The other uppermost features (S237, S247, and S263) are all deltas at the mouths of small streams. They relate to the same shoreline as S248 (Chapter 9) and define the position of sea-level while ice occupied Glendaruel. Feature S252 is interrupted by bedrock and is, in part, peat covered. It is perhaps two features, the topmost 3 points and the bottom 2 points being separated by a sharp step.

For more than 20 m below the features considered so far there is a number of very small terraces that are perched on the valley sides and around some of the rock outcrops in the valley. They do not form any clear bands and are, in general, poorly preserved. There is a suggestion of a grouping in the ca. 18 - 20 m altitude range with fragments S236, S282, S286, T139 and T129 that relates in part to shoreline CLG6 (Chapter 9). It seems most likely, however, that during the general fall of sea-level from the marine limit isolated terrace fragments became lodged on bedrock outcrops. Therefore they need not indicate any specific shoreline.

However, below the ca. 10 m level around the head of the loch is a considerable number of terrace fragments. The largest of these is T123 which is also well exposed. In part it is peat-covered and here levelling was not attempted as it could not always be ascertained when the peat stopped and the sediment started: the remainder of the feature was levelled, however. Exposures in a number of gullies towards the front of the terrace reveal it to be mainly composed of a fine grey sand interbedded with lenses of grey clay, these deposits being capped by ca. 1 m of current-bedded coarse yellow sands and gravels with lenses of fine grey silty sand. These grey sands and clays are interpreted as being an estuarine deposit similar to others that occur elsewhere around the Scottish coast (Sissons, 1967a) known as 'carse' and being Flandrian in age. The capping sands and gravels are indicative of fluvial deposition across the surface of the estuarine sediment during a shoreline regression, perhaps due to the building out of a delta by the river as the sea fell towards its present level. This interpretation is amplified by sections at the E end of this terrace and in a neighbouring correlative terrace cut into by the Auchenbreck Burn. Here a thin layer of small gravel caps a grey silty sandy clay bed that lenses out towards the back of the terrace. Beneath this carse deposit there are bedded sands and gravels, the sands being current bedded whilst the gravels are coarse with occasional boulders. At one exposure there was, at the base, a fawnish to blue/grey clay. The whole is underlain by a reddish/brown till. The bedded sands and gravels represent fluvial deposition and record a period of regression of the shoreline to at least 2.5 m below the surface of T123 (i.e. to ca. 5.5 m). The carse deposits in turn record a transgression of the shoreline, which, due to the relatively still water nature of the deposits, must have been due to a relative sea-level rise, whilst the capping gravels result from the succeeding regression. Unfortunately, the contacts are unconformable and the possibility of hiatuses in the sedimentary record cannot be ruled out and hence the continuity in time of these events is not assured.

T123 slopes down to ca. 8 m at which altitude it flattens out suggesting a relationship to sea-level. S254, a poorly developed fragment further down Loch Riddon, also relates to this level.

Immediately below T123 and occurring on both sides of the valley terrace fragments S259, S284, S261 and S260 define a nearly horizontal surface. S284, the best exposed, is mainly composed of bedded sands and gravels, dipping slightly seawards and containing lenses of grey sand, on occasion charged with twigs and branches. The gravels are coarser towards the base though there is also in this direction an increased number of fine sand layers and one bed of clay. This section and the horizontal terrace surface are interpreted as resulting from a prograding river terrace, the surface of the terrace being controlled by the sea-level of the time. The bend that occurs in the projection plane at this point results in an erroneous impression of the length and slope of this terrace as the two furthest upstream points on it suggest a possible correlation with T123. However, following the discussion on the genesis of T123 and taking into account the bend in the height-distance diagram it is clear that there is no such correlation, but that S284 is part of a different terrace sequence to T123. Fragments S261 and S260 are, where exposed, composed of fine grey sand whilst S259 is composed of small gravel at the surface. Together these fragments constitute a shoreline at ca. 7.0 m.

Below this shoreline two well-developed features, S255 and S257, have been surveyed. Exposures in S257 show it to be composed of fine grey sand capped, towards the N end, by peat. The peat has made mapping and surveying difficult as it masks parts of the otherwise clear breaks of slope between S258, T151 and S257. The softness of the underlying sediment makes the point of contact difficult to establish with an aluminium rod and levelling was not undertaken in those areas where peat was known to be present. However it is possible that part of the slope of S257 is due to increasing amounts of peat northwards. Together S255 and S257 constitute a shoreline at just below 5 m.

Abutting against and morphologically distinct from the above are S256 and T151 which, together with S285 from the opposite side of the loch, form a nearly horizontal group. Of these S285 is sectioned and shows a similar set of deposits to its neighbour S284 (described above) and is interpreted similarly as a terrace formed whilst the River Ruel built out into the contemporary sea-level. The apparent slope on the N end of T151 is not interpreted as resulting from river terrace development, for to do so and to maintain the correlation with S285 which occurs farther upvalley implies a somewhat unlikely configuration for the contemporary river mouth. Rather it is thought that as with S257 (see above) peat on the surface of T151 may have influenced the altitudes on the N end of the feature. A shoreline of 3.5 - 4.0 m is indicated by these fragments.

The final set of small terrace fragments is S251, S250, S249, S258 and T149. These are all well-vegetated but not removed from the influence of present high tides, though they seem above the level to which morphological flats are being formed today at the head of the loch (present river floodplain fragments T147 and T148 have been heighted at 1.7 m and 1.9 m respectively). These fragments define a level at approximately 2.5 - 3.0 m.

Five Flandrian sea-levels (at ca. 8.0 m, 7.0 m, 5.0 m, 3.5 - 4.0 m and 2.5 - 3.0 m) have been described, all of these being preceded by a marine regression to at least 5.5 m. All these levels are considerably below the maximum that Flandrian sea-levels could be expected to reach in this area (ca. 12.0 m; Chapter 10) and they represent considerable detail close to the present sea-level. The lack of representation of the highest Flandrian shorelines is partly explicable in terms of the configuration of the valley as sea-levels between the 8 m and the 12 m altitudes would occupy the rather constricted lower part of the river valley where they could be relatively easily removed by subsequent river action. Below this altitude marine deposition is in a much less enclosed area and the possibility of preservation is therefore greatly enhanced. Another reason that can explain part of the absence in the lower loch is that the relevant altitudinal range coincides in large measure with a fossil cliffline that flanks the majority of the loch. This cliff is not particularly well developed, but it can be followed without much difficulty. It is the cliff and platform discussed in detail in Chapter 8.

The degree of preservation of shorelines below the ca. 8 m level is also partly explicable by the relatively large amount of space available at the head of the loch. However, this does not help in understanding the number of very low shorelines which are unique to this area in Cowal. It seems certain that this is due both to the extremely limited fetch (maximum 6.3 km downloch) and the great expanse of intertidal sand which extends ca. 2.8 km from the head of the loch.

The combination of these two results in an extremely low-energy environment at the head of the loch which makes the preservation of landforms, formed by the steady if not too voluminous supply of sediment, much more probable.

(c) River terraces

Between Ballochandrain and the sea is a set of small terrace fragments all of which are below the level probably attained by the Flandrian sea. By analogy with river terraces in other parts of the study area a divergent downvalley system of river terraces might be expected but a number of difficulties both natural and artificial combine to preclude the identification of such a pattern.

In the first instance, as was mentioned earlier, the existence of numerous large rock outcrops in the valley bottom has restricted the river course to a relatively narrow area with the consequence that considerable reworking of deposited sediment must take place and only small terrace fragments have been preserved. Secondly, the present river long profile has a step. Such a step may have existed at the time of formation of the terrace fragments thus complicating the interpretation of the terrace sequence. Thirdly, a number of small streams descend the hillsides, particularly from the W, and deposit alluvial fans across the small terraces fragments, effectively masking much of the original terrace morphology. Fourthly, on the one or two larger terraces that might have been expected to have controlled the correlation of the terrace fragments, human modification (roads, a large house and a walled garden) have precluded levelling. Fifthly, the curving course of the river results in a number of terrace fragments, and the lines of points on them, being at a high angle to the projection plane of the height-distance diagram with an accompanying exaggeration of their gradient thus making correlation of fragments on the diagram more difficult. Finally it is possible that not all the features are river terrace fragments. Sedimentary evidence as to their origin is not available and the short distances over which many have been surveyed do not allow their gradients to be accurately established. Some features may possibly have a marine origin.

Turning to specific fragments it can be seen that some of the fragments are on slip-off slopes that have developed as the river migrated outwards and downwards in meanders. This applies to T144, T145, T121, T120, T119, T128 and T130. Since these are likely to be non-paired erosional remnants they need not correlate with other fragments. Fragments T126, T127 and T134 do not have a clear downvalley gradient on the basis of the points surveyed and could perhaps represent a former sea-level, particularly as T126 and T127 together form a narrow apron for more than one third of the way around a rock outcrop. Fragments T118 and T133 are both where side streams join the River Ruel and their deposition is therefore more likely to be the result of variations in discharge of these streams than of the main river. In short, it is not felt justifiable to try to reconstruct any former river terrace sequences over the area being discussed here.

(d) Glendaruel

Above Ballochandrain to near the site of Glendaruel House, Glendaruel is a wide open valley, the floor of which is covered for much of the way by a series of terraces. These terraces have been surveyed continuously over large distances and they clearly contrast with those of the lower valley. In addition, beyond Glendaruel House for several hundred metres the valley bottom is occupied by fluvio glacial landforms that have no counterpart in the lower valley.

Commencing at Ballochandrain, the first large terrace above the river floodplain is T33. This has an extremely steep gradient and can be followed up to the Clachan of Glendaruel where it merges into the large alluvial fan upon which the village is built. It seems possible that this is a downvalley extension of the fan though the alternative of it being a remnant of a larger terrace that has survived erosion due to the protection of the fan cannot be ruled out. However its steepness favours the former interpretation. It slopes down to below 11 m where it merges with the present river floodplain.

The next terrace upvalley is T23. Its surface is interrupted by small alluvial fans. The upper parts of this terrace (as well as the lowest parts of T22 which follows it immediately upvalley) are badly affected by the deposits of a side stream but if allowance is made for this by looking only at the more regularly established portions of T23 and (lower) T22 then these two terrace fragments appear to be part of one feature. Two smaller correlative fragments (T32 and T30) occur on the opposite side of the valley whilst fragment T31 occurs on the riverward side and is morphologically below T32. T22 is, however, clearly composed of two parts on the height-distance diagram. The distinction is partly masked in the field by an alluvial fan which tends to merge into both parts of the terrace, but the difference was partly, if not very confidently, initially mapped in the field. T22 is therefore divided into two terraces of quite distinct gradients, that part farther upvalley (T22b) being considerably steeper. The upvalley termination of T22a coincides with a curious esker-like deposit on the opposite (E) side of the valley. This feature extends down on to the valley bottom, is slightly curved upvalley and is flanked on its immediate downvalley side by T30, a correlative of T22. No sections are available to aid interpretation. Immediately to its rear there is an irregular area of ground with two small rock outcrops and one closed hollow which may be a poorly developed kettle hole. The mound is clearly a glacial or fluvioglacial deposit and when taken in conjunction with the termination of T22a is interpreted as ice-marginal. T22a is therefore glacial outwash. This produces a difficulty of interpretation as its correlative T23 continues downvalley to ca. 11.5 m, below the uppermost level the Flandrian sea can be presumed to have obtained in this part of the study area. However, if the sea did invade this part of the glen then shore-forming processes would have been very weak in the almost landlocked little sea loch that would have existed (cf. Lochan Dubh at the mouth of Glen Shirra). It could be that the nearly horizontal portion of T23 at ca. 12.6 m (established over a distance of ca. 100 m) is related to the highest level of the Flandrian sea. T31 which, apart from its very upper portion, oscillates around this altitude would lend support to this idea. The lower portion of T23 below this altitude could be that which survived the inundation. The ice limit at the upper end of T22a can therefore be said to relate to a sea-level at least as low as 12.5 m.

T22b can be traced upvalley until Maymore Farm. Here it is mainly truncated on its upvalley side by T20 but between T20 and the valley side a number of low depositional mounds occur just beyond the end of T22b. T22b is well exposed at one point by the river where well-bedded sands and coarse gravels and pebbles are seen. T22b is interpreted as an outwash terrace with the small mounds behind being deposits near the ice margin.

T20 extends most of the way into the centre of the valley and with the lower part of T28 confines the River Ruel to a very narrow strip. The altitudes on T28 appear to indicate that it is really two terraces, the downvalley part being that which is related to the terrace level being discussed whilst the upvalley portion can be easily correlated with T27. Upvalley of T20, T19 forms a continuation and it can be traced into the grounds of Glendaruel House. Here the terraces terminate and clear fluvioglacial landforms occur in the valley bottom. These consist for the main part of a meandering esker system, occasionally exposed to reveal coarse rounded pebbles and cobbles. The meandering nature of this esker system with its strong analogy to supraglacial streams observed on present glaciers suggests that these eskers were supraglacially formed. In addition to these eskers are a number of other kames and terraces that flank these features on the sides of the valley. T21 and T29 have been levelled and since they overlook kames and eskers are kame terraces; they can be correlated with outwash terrace T25 further downvalley.

In all, 3 different sets of outwash appear to relate to, or lie near to this ice limit. The major set, and the most steeply sloping, is that formed by T28, T20 and T19. The next is formed by the upper part of T28 and T27 whilst T25, T29 and T21 clearly relate to a slightly later position of the ice front. Sections in these outwash terraces reveal coarse, well-rounded gravels and pebbles with occasional beds of current-bedded sands. Towards the base are occasional boulders. In one section in T28 a set of convolutions was observed. They do not disturb the overlying beds and were therefore formed as the outwash built up. Their origin may be periglacial but they are thought to be more probably due to the melt-out of small blocks of ice that current observations show to be washed out on to sandur during high stages (Embleton and King, 1975).

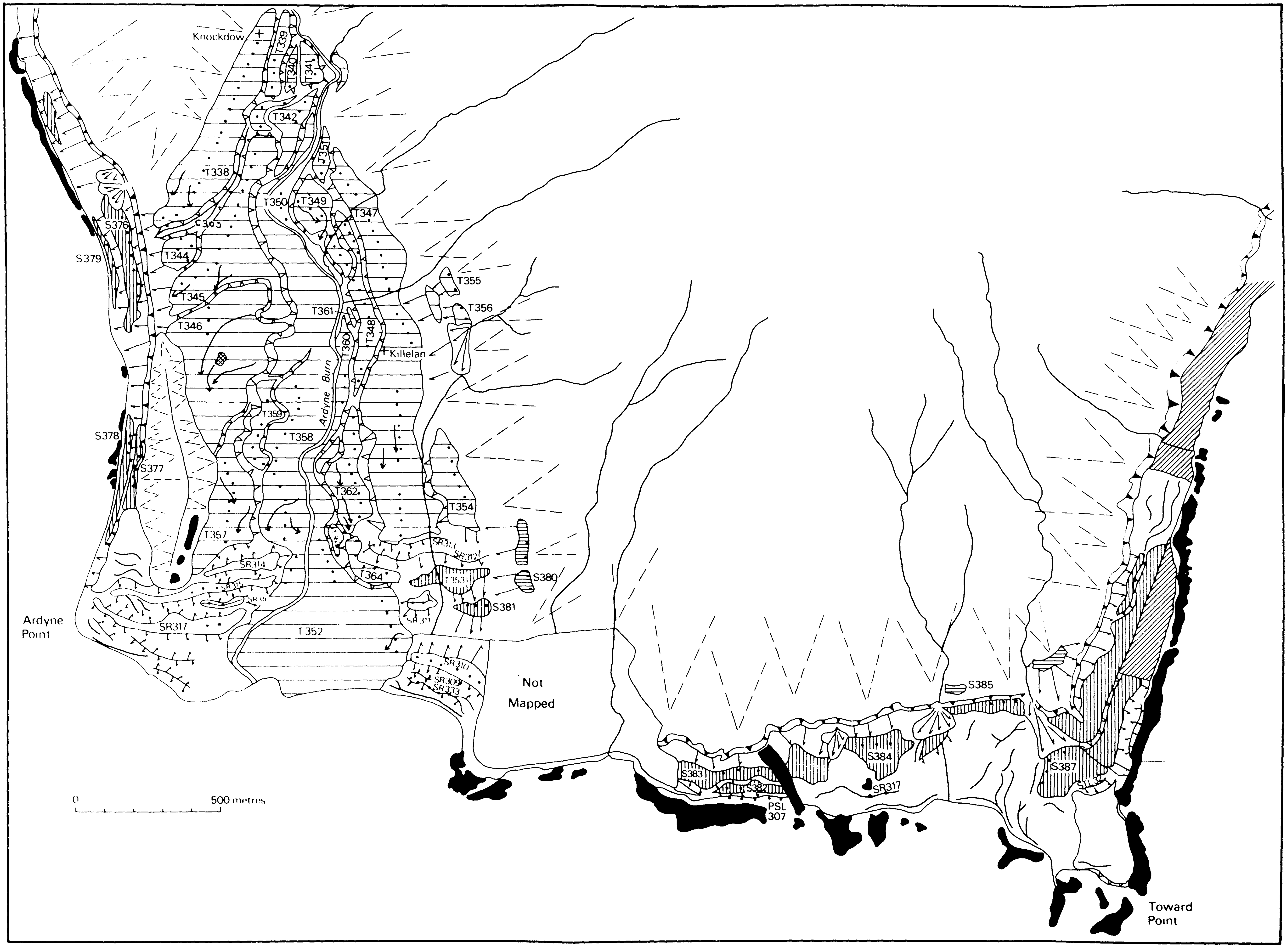
Upvalley of these glacial deposits no further terraces can be differentiated from the present river floodplain.

Five sets of outwash have been identified in the lower valley. Perhaps because they extend limited distances they are all apparently linear in long profile and in this they contrast with the terraces of Ardyne and Strath Eachaig discussed earlier. The outwash represented by T22b is the most steeply sloping and each successive terrace upvalley from it declines in gradient until the profile formed by T25, T29 and T21 which has approximately the same gradient as the T23, T32, T30 and T22a outwash, the farthest downvalley. In addition, the upvalley limits of T22a, T22b and T19, all of which are interpreted as terminating at ice margins, rise upvalley. The reasons for these changes in gradient and upvalley altitudes are not known: perhaps sediment supply and the length of time of ice front stability are relevant factors. Although they cannot be related directly to specific sea-levels, the terraces place considerable limitations on Lateglacial sea-levels and in view of the position of the ice marginal delta S248 (37.43 m) and the lack of any evidence for sea-levels upvalley of S248, except perhaps the one at ca. 12.6 m, it seems necessary to postulate that the ice margin lay at S248 whilst sea-level dropped to at least ca. 12.5 m, a relative fall of sea-level of at least 25 m.

Summary

The marine limit at ca. 37.5 m was formed immediately upon deglaciation in the Loch Riddon/Glendaruel area. The ice retreated until Ballochandrain where it halted and sea-level fell without apparently leaving any well-developed shorelines. While ice remained at Ballochandrain sea-level fell to at least 12.5 m. Subsequently three ice-front positions and five different outwash trains were formed in Glendaruel, none of them related to a known sea-level. Sea-level is known to have fallen sometime subsequently (or contemporaneously?) to an altitude of at least 5.5 m, whereafter it rose, at least to 8 m and probably to ca. 12.5 m.. During the following fall of sea-level estuarine and beach deposits were preserved and poorly developed river terraces were formed. Shorelines were left at ca. 8.0 m, 7.0 m, 5.0 m, 3.5 - 4.0 m and 2.5 - 3.0 m as the sea fell to its present level.

Figure 31: Geomorphological map of the Toward/Ardyne area.



Knockdow
T339
T340
T341
T342
T351
T338
T350
T349
T347
S376
S379
T344
T345
T355
T356
T361
T346
T360
T348
Killelan
Ardyne Burn
T359
T358
S378
S377
T362
T354
T357
SR313
SR312
SR314
T364
T353
S380
S381
SR317
SR311
Ardyne Point
T352
SR310
SR309
SR333
Not Mapped
S385
S384
S387
S383
SR317
PSL 307
0
500 metres
Toward Point

Figure 32: Height-distance diagram of the terraces of the Ardyne area.

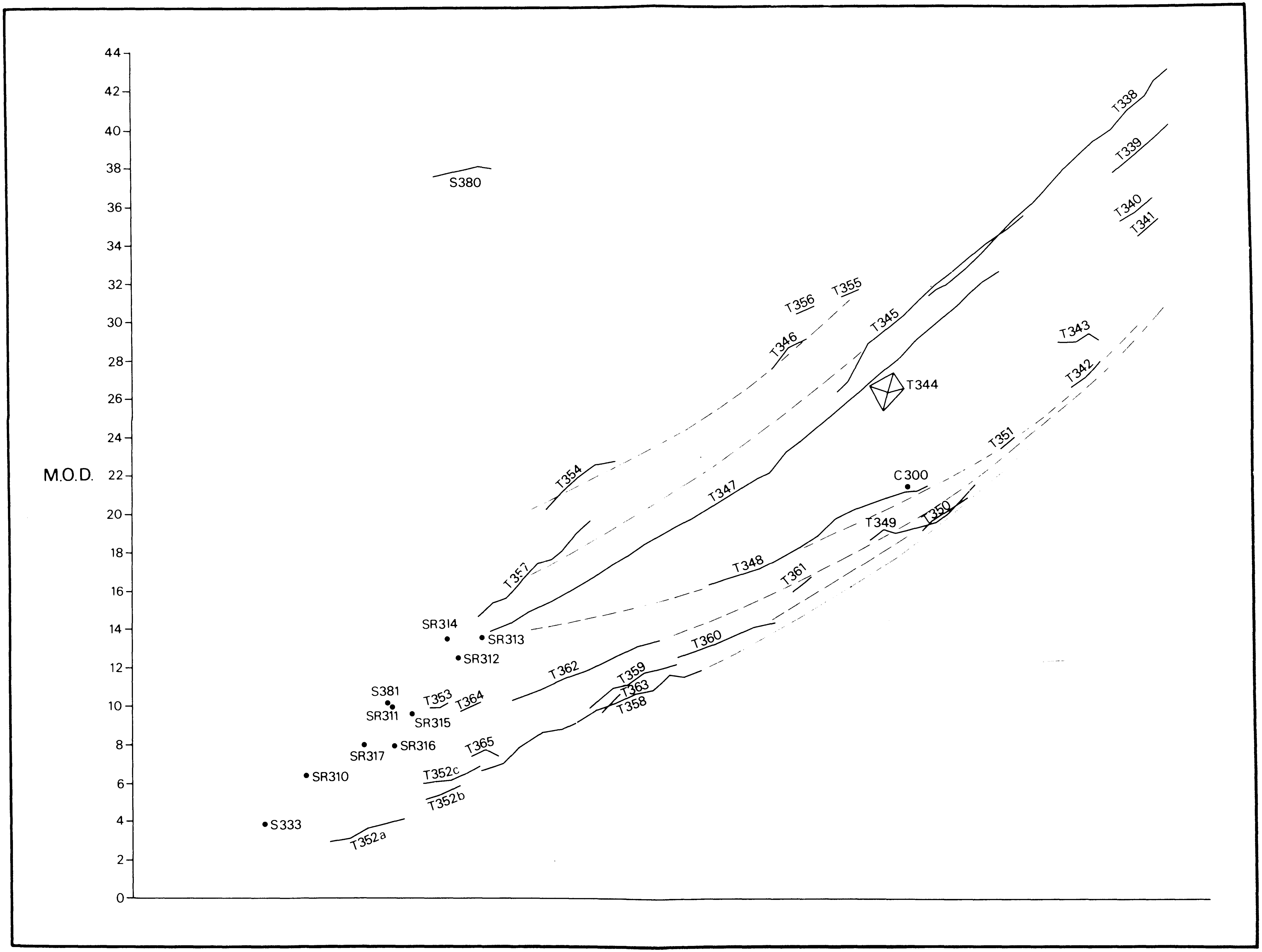
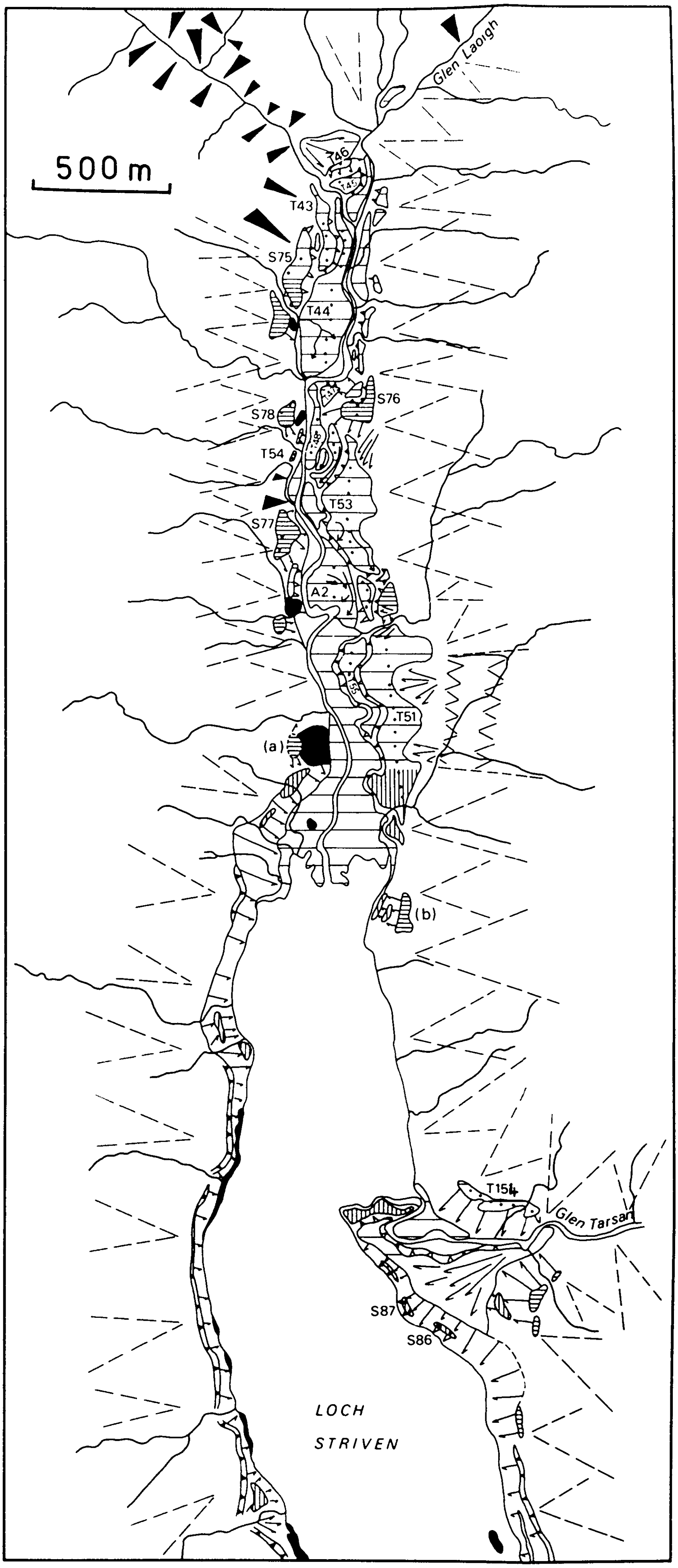
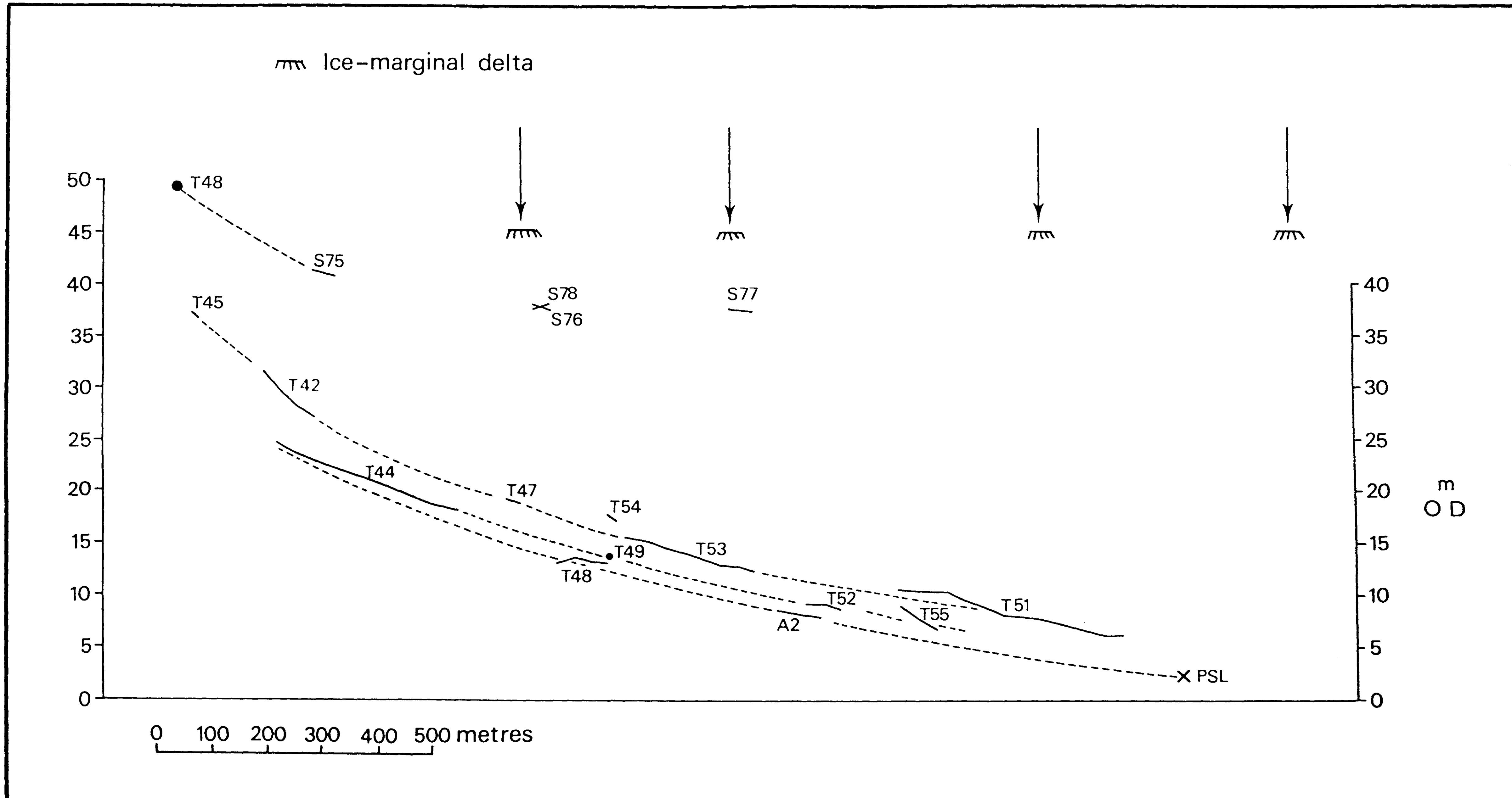

M.O.D.
44
42
40
38
36
34
32
30
28
26
24
22
20
18
16
14
12
10
8
6
4
2
0
S380
T338
T339
T340
T341
T343
T342
T355
T356
T345
T346
T344
T354
T347
T351
C300
T349
T350
T357
T348
T361
SR314
SR313
SR312
T360
T362
T359
T363
T358
S381
SR311
SR315
T353
T364
SR317
SR316
T365
T352c
SR310
T352b
S333
T352a

Figure 33: Geomorphological map of the head of Loch Striven.



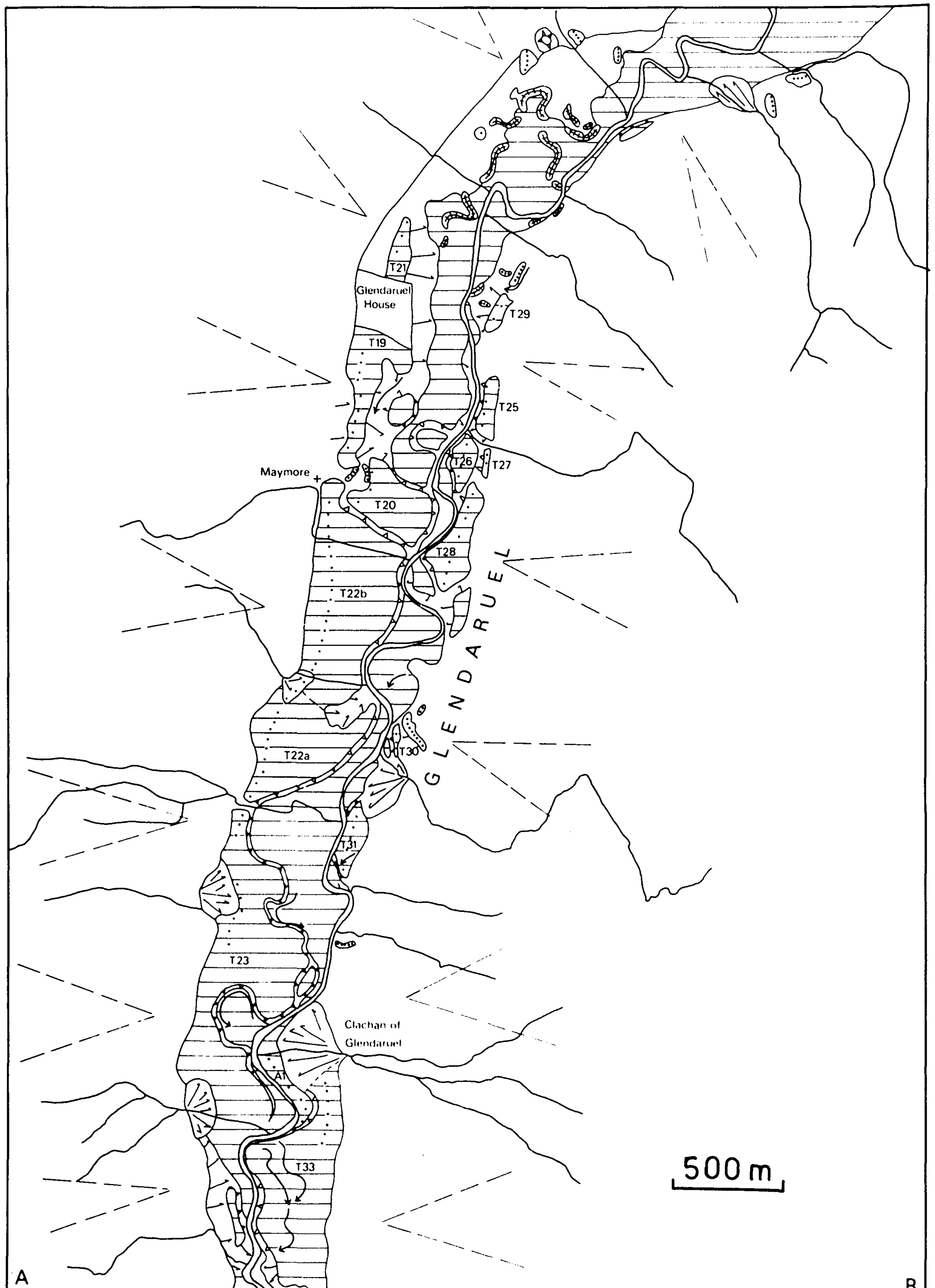
500 m
Glen Laoigh
T46
T45
T43
S75
T44
S76
S78
T54
T53
S77
A2
T51
(a)
(b)
T154
Glen Tarsan
S87
S86
LOCH
STRIVEN

Figure 34: Height-distance diagram of the terraces at the head of Loch Striven.



Ice-marginal delta
T48
S75
S78
S76
S77
T45
T42
T44
T47
T54
T49
T53
T48
T52
A2
T55
T51
PSL
m
O D
0 100 200 300 400 500 metres

Figure 36: Geomorphological map of Glendaruel/Loch Riddon.


T21
Glendaruel House
T19
T29
T25
Maymore
T26
T27
T20
T28
T22b
GLENDARUEL
T30
T22a
T31
T23
Clachan of Glendaruel
A1
T33
500 m
A
B

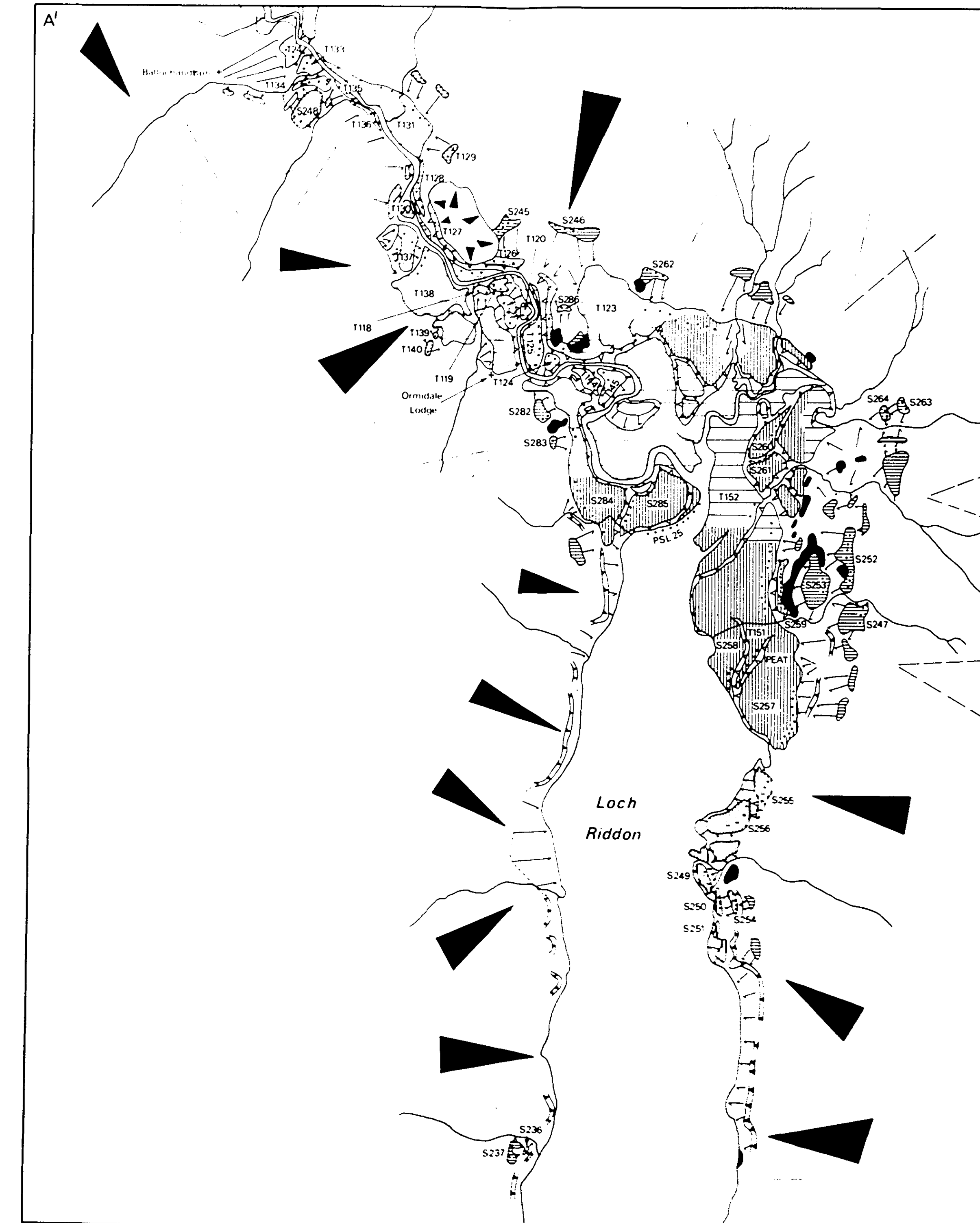
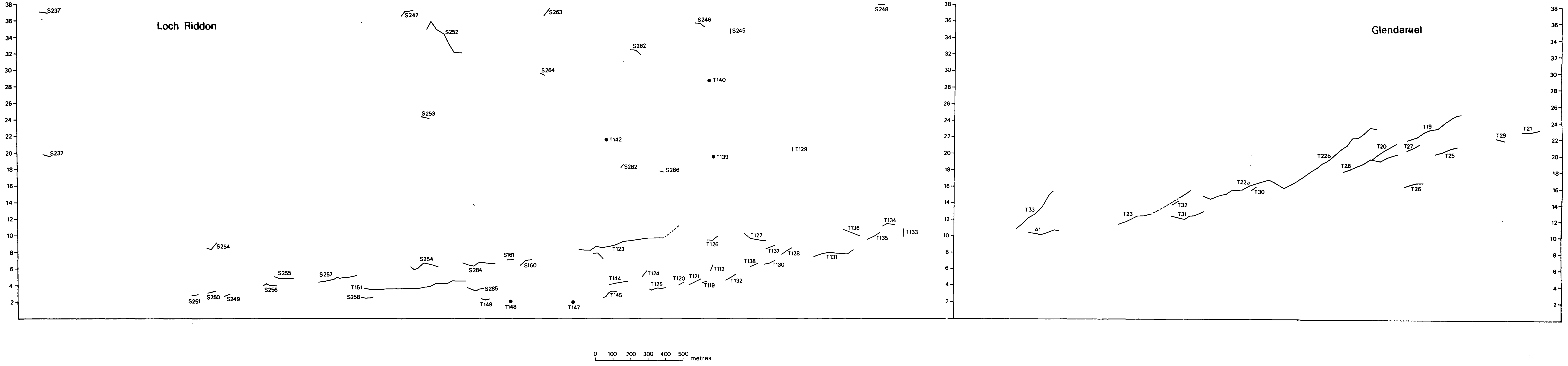

A'
B'
T133
T134
T136
T131
S248
T129
T128
T130
T127
S245
S246
T120
T137
T126
S262
T138
S286
T123
T118
T139
T140
T119
Ormidale Lodge
T124
S282
S283
S264
S263
S261
S284
S285
PSL 25
T152
S252
S253
S247
S259
T151
S258
PEAT
S257
Loch Riddon
S255
S256
S249
S250
S251
S254
S236
S237

Figure 37: Height-distance diagram of the terraces of Glendaruel and Loch Riddon.


Loch Riddon
Glendaruel
metres
0 100 200 300 400 500

CHAPTER 7

FIELD DESCRIPTIONS OF AREAS AROUND LOCH FYNE

1. Portavadie/Ardlamont

The most southerly part of the study area lies between the mouth of Loch Fyne and the western Kyle of Bute. The coast is fringed here by an extremely well-developed rock cliffline and platform, the slope along the front edge of the platform defining the present shoreline for the majority of its length. A number of small valleys, at Portavadie, Low Stillaig, Stillaig, Millhouse and Ardlamont, meet the coast in this area and it is the deposits and landforms of these valleys and around the Ardlamont Peninsula that are presently of concern.

(a) Millhouse/Kilbride Bay

The most interesting valley is that of the Allt Osda between Millhouse and Kilbride (Ostal) Bay. Glacial, fluvioglacial, glacio-marine, marine, fluviatile and eolian deposits have all been mapped in the 3.5 km between Millhouse and the sea. These are presented on Figure 38 and the relevant altitudinal information has been used to construct a height-distance diagram (Fig. 39).

A number of exposures throughout the valley reveal a basal red till. In the deposits overlying this till there is a clear stratigraphic and morphological break in the valley N of Auchagoyl Cottages. To the N of this locality there is a number of smoothly rounded mounds occupying the valley bottom. Sections in these show well-bedded sands and gravels overlying the till directly, the contact, where observed, being sharp. These mounds are interpreted as kames, and the terrace (T337) at Millhouse, on a level with the flat-topped M302, as a kame terrace. To the E of the village, on the smooth till-covered slopes, channels that trend across the hillside towards these kames are interpreted as meltwater channels. The lowest mound down-valley is coincident with the termination upvalley of a set of small terrace fragments. These fragments have been levelled where possible and they show up in a horizontal band on Figure 39. A mound at Auchagoyl Cottages is at right angles to the trend of the valley and in it an exposure beside the river reveals red to orange/ yellow laminations of fine sand. These grade upwards (in an imperfect exposure) through alternating small gravel and sand beds to a more general coarse pebble and cobble layer, the whole being capped by large rounded boulders. Gunn et al. (1897) include an unconfirmed (by them) report that shells were discovered in a (now) disused gravel pit dug into the rear of this mound.

South of this point red well-laminated clays and fine sands and silts occur in the valley. They are never observed to overlie the red till directly but where the two are revealed in the same section a layer of coarse, poorly sorted unbedded sands and gravels, with occasional cobbles and pebbles intervenes, this layer having a sharp contact with the overlying clays but being gradational into the till. The clays are revealed most often in the valley bottom but at (a) they occur up to an altitude of ca. 35 m. The laminations in these clays, silts and fine sands, often coincide with a change of colour from redder to yellower. However in the more clayey sections the laminations are usually formed by a red clay band (of variable thickness, typically a few millimetres) and a fine sand parting often only a grain in thickness. The valley bottom sections all coarsen upwards becoming predominantly sand before they are truncated by overlying gravels. In section (a) the layers of clay at first interdigitate with overlying sand and gravel beds that culminate in a small terrace.

The following sequence of events is proposed for the deglaciation of this valley. The ice margin retreated up the valley until the local glacier became cut off from its ice source or sources by the low cols leading to the neighbouring Kyles of Bute and/or Loch Fyne. At this time the front of the glacier was located at Auchagoyl Cottages. Subsequently, to the N of this, the ice disintegration features that have been described above developed as meltwaters penetrated the stagnant ice. The coarse sand and gravel layer that overlies and grades into the till S of this locality is interpreted as being washed till, the washing having occurred subglacially or rapidly and immediately upon deglaciation, possibly by meltwaters emerging from underneath the decaying glacier, less likely due to the inrush of sea as the ice calved off (cf. Skinner, 1973). Sea-level was at its maximum at this time and the laminated clays are identified as the first marine deposit in the valley. This suggestion receives support from the sparse marine fauna found in similar laminated red clays in the Kyles of Bute (Crosskey and Robertson, 1873).

The origin of the mound at Auchagoyl Cottages is, on the basis of the surrounding relationships, a form of ice-marginal delta. However, no foreset beds were seen to corroborate this. The basal laminated fine sands are thought to correlate with the laminated clays, silts and fine sands that occur downvalley. If this correlation is correct and if the report of shells in the deposit is true, then a slight oscillation of the ice front may be indicated.

Sea-level at this time, on the basis of fragments S324, S325, S326, S327, and S329 is placed at _ca_. 36 m. Drainage through the stagnant ice, as represented by T337 and M302 was controlled by this sea-level. The downvalley fragments S326, S327, S329 and (a) are related to side valleys and are small deltas. The size of S327 and S326 relative to the small stream that they now flank is suggestive in terms of decaying ice in the hills to the E, drainage from this ice down the channel of the Allt a Chaorruinn splitting across the depositional flat(b),in part to follow the present course of this stream, in part to discharge to S326/7.

No exposure in the valley showed any evidence of the presence of the light grey shelly sand that is commonly found immediately over the red clays at Portavadie (see below), the Kyles of Bute and that occurs at an altitude of up to ca. 15 m at point (c) (Fig. 41) to the SE. Instead the laminated red clays are succeeded upwards by clean sands in a steadily coarsening sequence. It is therefore suggested that this sequence is due to the contemporaneous fall of sea-level, a fall of sufficient magnitude that by the time of the neighbouring deposition of the grey sands, this long narrow valley had become sufficiently littoral to exclude their deposition. This interpretation receives support from the fact that kames M301 and M300 have maximal altitudes of 33 m and 30 m respectively, coinciding with a major shoreline (CLG2, Chapter 9) represented in the lower valley by S328 and thus suggesting the possibility of dead ice existing in the upper part of the valley while relative sea-level dropped by at least 6 m.

In the lower part of the valley a series of terraces is developed slightly above the valley floor. The highest of these, S292, T325 and T324, are coincident in altitude (19.5 - 20.5 m) with a zone of shoreline formation well represented in this area (see below) and are therefore considered to belong to this group. The other terraces slope consistently downvalley and are paired on either side of the present stream. They cannot be related to any specific sea-levels, mainly due to the cover of wind-blown sand from the small area of sand-dunes and machair that fringes the head of Kilbride Bay. However, since they can be followed down to an altitude of ca. 8.5 m it can be concluded that they are Flandrian in age.

(b) Portavadie

Portavadie is now the site of a North Sea Oil Production Platform construction yard. This, plus the occurrence of good natural exposure and landforms that were mapped and heighted prior to the establishment of the yard has allowed a considerable amount of detail of the geomorphological history to be inferred.

The most significant element of the make-up of the coastal area is a very well-developed rock cliffline and fronting platform (Fig. 38). This is best developed immediately to the S of Portavadie where it faces up Loch Fyne. Here the platform attains a maximum width of 80 m and the cliffline a height of over 15 m. Stacks, undercutting, caves and geos are all exemplified by this relict feature. The frontal slope which dips beneath the sea is, in turn, cut into by geos, none of which reach the back of the platform. Three fragments, giving a total of 12 points, have been levelled along this section of platform. The mean heights of the fragments are 5.80 m (RP16), 5.64 m (RP17) and 5.96 m (RP18). Elsewhere around this section of coast the cliffline is easily traceable and it is often fronted by a platform. Small islands, such as Eilean ma Beithe and Eilean Buic are incised by it and at its time of formation Rudha Mor was a similar island. Where streams emerge through the cliffline, sands and gravels are built up in front of the cliff, especially at Glenan and Portavadie.

Inspection of natural sections, interpretation of boreholes and examination of the workings at Portavadie show the following general sequence of sediments which is well-illustrated in the two cross-sections (Fig. 40) that were constructed from lines of boreholes inland from Port a' Mhadaidh.

Brown sands and gravels

Grey silty sands with occasional boulders and beds of clay, organic layers and abundant shells

Red laminated clays

Poorly sorted, generally non-stratified red sand and gravel

Red Till

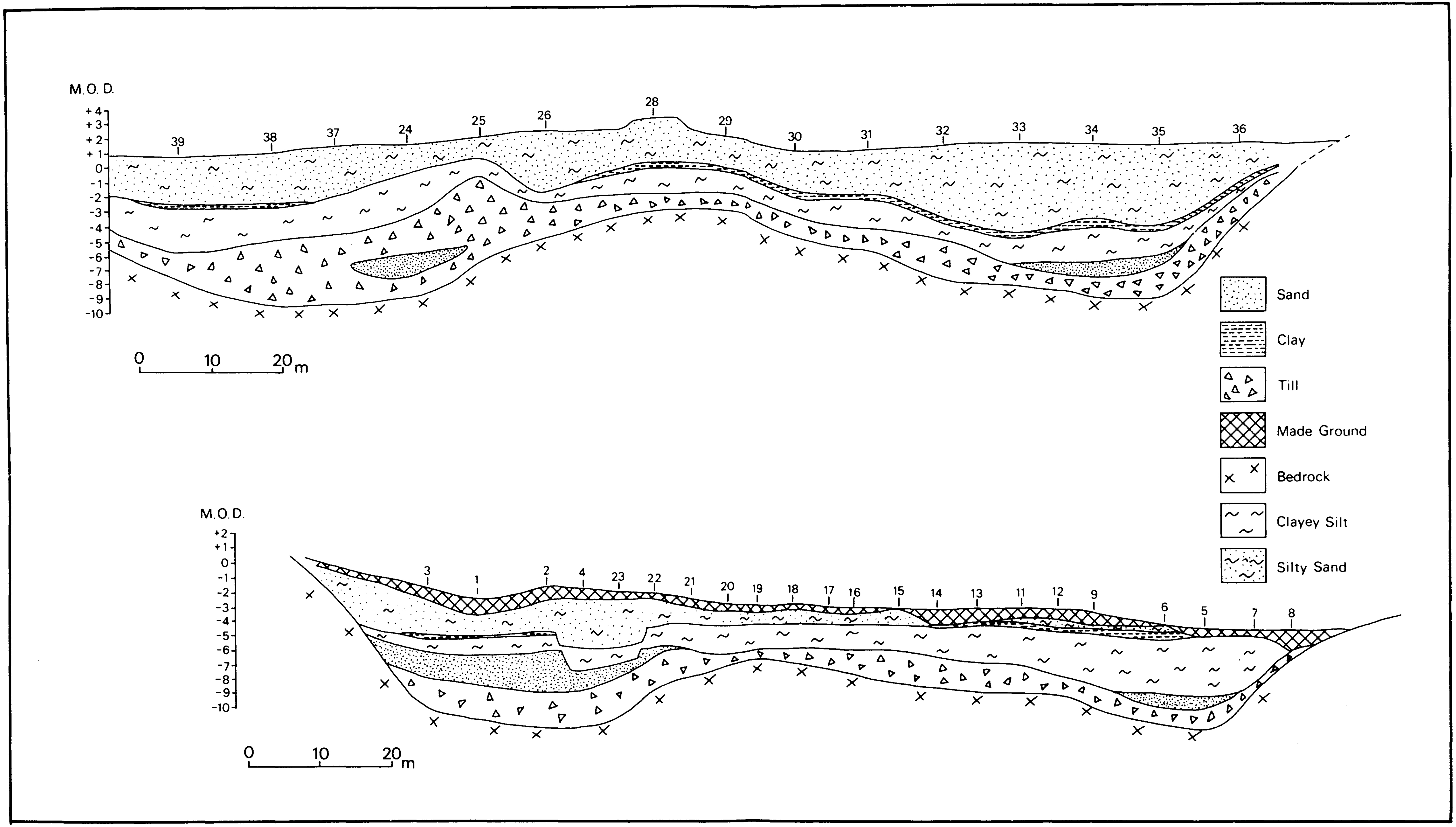
Weathered Bedrock

Fresh Bedrock

The bedrock in this area has been etched by ice into a series of ridges and hollows, these running parallel to the strike of the foliation of the Beinn Bheula Schists. The above mentioned rock platform cuts across this trend but is not entirely uninfluenced by it (see below). In general the bedrock ridges stand up bare of deposits except for a thin peat cover. The hollows and stream valleys contain the deposits, though these often overlap low ridges as was seen in the excavation of Port a' Mhadaidh.

The weathering of the bedrock has been revealed in roads cut around the construction site. Some of the site exploration bores recorded 'weathered bedrock' at a depth of <u>ca</u>. -10 m but during excavation this

Figure 40: Cross-sections of sediments revealed at Portavadie during construction of the Oil Platform Yard.



M.O. D.
+4
+3
+2
+1
0
-1
-2
-3
-4
-5
-6
-7
-8
-9
-10
39
38
37
24
25
26
28
29
30
31
32
33
34
35
36
0
10
20 m
Sand
Clay
Till
Made Ground
Bedrock
Clayey Silt
Silty Sand
M.O.D.
+2
+1
0
-1
-2
-3
-4
-5
-6
-7
-8
-9
-10
3
1
2
4
23
22
21
20
19
18
17
16
15
14
13
11
12
9
6
5
7
8
0
10
20 m

proved to be mechanically shattered schists. Weathered rocks are exposed on lee slopes with respect to the movement of the last ice to cover the area but this can, at the moment, only be regarded as a function of the sampling since similar exposures are not available for any great distance on stoss sides. None of the areas of weathering is overlain by anything but peat and/or thin slope deposits and nothing can be inferred as to the age of the decomposition.

Little or no till is exposed at the surface in the immediate surroundings of Portavadie. Near Glenan (NR925708) there is a ca. 10 m exposure in the side of a stream but it was only in the excavations at Portavadie that much till was seen. As in the Millhouse/Kilbride Bay valley this till was immediately overlain by a layer of poorly sorted reddish brown sand, gravel and occasional boulders. The clasts were often well rounded. This deposit is not a uniform spread but occurs in pockets, these on occasion being recorded in boreholes as consisting of reddish-brown fine sand, sometimes silty, sometimes laminated and slightly compacted. The chief characteristic of this stratum is its lack of clay-sized materials, a fact emphasized by the occurrence of a clayey till beneath and a well laminated clay above, even though colour changes but little throughout. A two-fold origin is proposed for this material. In part it is thought that a slight washing of the till may have been effected just as the ice front retreated from it (assisted by tidal scour and/or by calving at the ice front?). Such a mechanism, however, seems incapable of explaining the sand beds: hence submarginal drainage of a tidal glacier of the type described in Alaska by von Engeln (1912) and Tarr (1909) is envisaged as having left these lenses in the sedimentary record.

Immediately overlying and conformable with the sands and gravels are red laminated clays. These clays are laminated due to a fine sand or silt layer, along which they easily peel. They have a maximum recorded thickness of ca. 4 m. At one locality, immediately overlying the till, the red clays showed no signs of laminations but were massive and stiff. The clays can be traced in a number of exposures up the Derybruich Burn, inland of the small gorge that the burn has cut in the cliffline. Of particular interest is section (d) (Fig. 38) that reveals basal red clays with fine laminations, these coarsening upwards until the section becomes one of alternate sand beds and then at the top, sand and gravel beds dipping quite steeply downvalley in the form of foresets. This exposure is in the side of a large terrace (T335). The highest point on this terrace fragment was levelled at 31.52 m and it was followed down to 26.28 m. Since the red clays are related to the time of earliest marine deposition in this area (cf. Millhouse area) at ca. 36 m and since the section in this terrace demonstrates a continuous coarsening depositional sequence as T335 was built outwards to at least 26 m a fall in sea-level of ca. 10 m is indicated prior to the initiation of grey silty sand sedimentation.

To the SE of Portavadie, construction of approach roads has revealed other areas of red clay deposition, particularly in the small embayment fronted by S244. This feature rests on till. Immediately overlying the till are red/yellow clays which are succeeded in turn, especially towards the edges of the depression by well-sorted yellow sands. At one point at the E edge these sands show, without apparent change in particle size, a red/yellow lamination. The major part of the surface of this feature (at 28.9 m) is composed of well-sorted medium to coarse gravels.

At one locality pebbles and cobbles, again well sorted, were exposed. The general sequence in this embayment is interpreted as resulting from a falling sea-level, the relatively still-water clays giving way progressively to more littoral sands and the shore features receiving their final form from deposition of very close inshore gravels. Laminated clays, such as the red clays that are here under discussion, are often closely associated with glacier ice (Charlesworth, 1957), a relationship that has been demonstrated for these clays. Unfortunately, the available information cannot be used to decide whether the rhythmite sequences are annual (i.e. that they are varves) or whether they are due to repeated slumping (cf. Paterson, 1974).

The next element of the depositional sequence is the layer of grey silty sand. This is a rather complex deposit. It overlies the red laminated clay quite conformably and usually with a sharp contact, though slight interdigitation was seen at one point. The clay content is variable, sometimes being such as to give rise to clay lenses. Gravels, pebbles, cobbles and boulders are scattered throughout. These are normally well-rounded and, according to the borehole evidence, have a greater tendency to occur in layers, the most commonly identified of which is immediately below the top of the deposit. Besides the grey colouration, the chief diagnostic criterion for this deposit is the shell content. The shells are unevenly distributed and tend to occur in layers, the first and chief of which is at the base, which can also be notable for its high clay content. This deposit has been sampled for faunal and radiocarbon analysis by Dr. J.D. Peacock of the Institute of Geological Sciences. Two radiocarbon dates (SRR-831 and SRR-832) indicate that the grey shelly deposit accumulated between at least 11,500 and

10,700 yr BP. The greatest recroded thickness of this deposit is ca. 12 m, but the lack of differentiation of sediment types in this borehole means that this figure should be treated with scepticism. More typical thicknesses are 5 - 6 m.

The surface of this grey stratum is clearly the result of erosion (Fig. 40). The silty sand is rarely recorded above ca. 5 m O.D. in altitude and the greater part of the surface only reaches ca. 1 m O.D. It is difficult to give accurate dating of the period of erosion of these sands, and indeed, the possibility of more than one period of erosion cannot be ruled out. However certain relationships can be established. The grey sands extend underneath low Flandrian (up to ca. 6 m O.D.) sands and gravels, precluding the explanation of present-day erosion, a circumstance that is thought unlikely anyway given the sheltered and accreting nature of the coastal embayments in question. Further support for the lack of present erosion comes from the fact that in Port a' Mhadaidh the sands (seen to be deposited up to ca. 15 m at locality (c) (Fig. 41) approx. 6 km to the ESE) lens out against the front edge of the rock platform whilst on more exposed neighbouring coasts these front edges have been washed clear. Examination and heighting of areas of current marine erosion elsewhere in the study area (Chapter 3) shows that such erosion takes place up to ca. 2 - 2.5 m. An erosional surface sloping in the elevation band of 1-5 m therefore requires a sea-level (or sea-levels) not much different from that of the present. A further limitation is placed on the period of erosion by the radiocarbon evidence: it must post-date ca. 10,700 yr BP.

(c) The general area

In this general area there are few glacial landforms and almost all the work, in addition to that which has been described in the sections on Portavadie and Millhouse/Kilbride Bay, has been concerned with fossil marine features. Of these the rock cliffline and associated platform is by far the most dominant element of the coastal zone and hence will be discussed first.

With the exception of the mouths of one or two valleys, the fossil cliffline can be followed continuously around the coast. It is, for the most part, accompanied by a rock-cut platform, which is normally covered by peat and very often by Flandrian sands and gravels. No glacial or Lateglacial deposits have been noted on the platform in this area. Both the cliffline and the platform are very variable in their development, the cliff ranging from a very strongly developed, near vertical face up to 45 m high on the W of Ardlamont Point to a weak, sloping feature only 3 or 4 m high at, for example, Rubha Mor near Portavadie, and the W side of Kilbride Bay. The platform varies in width from over 400 m (near Carry) to less than 10 m. On occasion it has a smooth surface and is pleasant to walk on, but more often it is irregular with geos cut into it and boulders lying on it, fallen from the cliff behind. The front edge of the platform slopes quite uniformly, but more steeply than the platform, into the sea. Caves, normally with collapsed roofs, stacks, geos and undercutting are all preserved on various parts of the platform and cliffline.

Both the platform and cliffline are strongly influenced by rock structure, though not in a similar manner. The dominant structural element in the area is the foliation in the schists. This strikes N45°E - S45°W and dips at ca. 40° towards the SE. A secondary structural element is a near vertically dipping fracture pattern striking E20°S - W20°N. Cliffs are best developed along coasts approximately at right angles to the strike of the foliation and where this is dipping into the cliff-face. This is best exemplified S of Portavadie, on the E side of Asgog and Kilbride bays and on the W side of Ardlamont Point. The alternative circumstances of the planes of foliation dipping downwards out of the cliffline results in low, sloping rock faces. On S-facing cliffs the secondary structures can be seen to control relatively low but sharp vertical faces. In such a situation W of Kilbride Bay the development of two stacks can be clearly related to preferential erosion along these secondary joints.

The rock platform, however, has its best development along the E of Ardlamont Point, fronting the relatively poor cliffline there. The areas previously mentioned with good cliff development have clear but narrower rock platforms. It is suggested that this contrasting development is due in large measure to the differential cliff evolution. The two sides of Ardlamont Point can be used to illustrate this more fully. On the W side, with its large cliffs, much of the energy available was expended in cutting the cliff whilst on the E, the cutting of the cliff was assisted by the tendency for undermined rocks to slump down along the foliation planes and hence more rapid cliff retreat was possible. Estimates were made (see Chapter 8 for a discussion of methods) of the amount of material removed in the construction of the cliffs and platforms here. Along a length of 900 m of the W side where the platform averages ca. 50 m in width, the removal of ca. 1,800 tonnes/m was recorded, and along 1,000 m of the E side, with an average platform width of ca. 100 m, ca. 1,400 tonnes/m. The general agreement between these estimates, especially as variations in fetch have not been considered, is a good support for the idea of energy being distributed in a compensatory manner between cliff and platform.

Platforms fronting the rather clear cliffs related to the E20^{o}S joint pattern are rather poorly developed due to the differential erosion along the major foliation pattern, which, running at an acute angle to the shoreline, breaks up the continuity of the platform.

The intricacies of the coastline around this area mean that the above mentioned differential development can be readily seen within relatively short distances. For example, a walk round Ardlamont Point is extremely instructive in this respect, as well as revealing, at the very point itself, platform fragment RP306 which is cut into a weathered dyke. This rock weathering must be presumed to post-date the cutting of the platform as it is unlikely that the marine agencies that cut nearby schists back by ca. 100 m would only erode this weathered rock, which shows development of corestones and has a dull thudding sound (as opposed to a sharp ring), to a distance of ca. 30 m.

In addition to the three fragments previously mentioned at Portavadie, levelling has been carried out on three other rock fragments in this area: RP304 (5.76 m), RP305 (6.12 m) and RP306 (6.35 m).

In addition to the features already mentioned numerous other fossil marine landforms have been mapped and levelled throughout this area. Examination of the height distribution of these fragments shows clear groupings. No exact altitudinal correspondence between shoreline fragments would be expected even in such a small area as is under consideration at present. Factors such as errors in measurement, variation in type of feature, preservation of feature, exposure and isostatic uplift can be expected to influence the measured present day altitudes of marine landforms. All these factors are considered in

Chapters 9 and 10 in an extended discussion of the analysis of the shoreline fragments in the whole study area, but the following groupings are worthy of note at the present. Eight fragments, S348 (34.08 m), S329 (35.28 m), S327 (35.98 m), S326 (36.78 m), S325 (34.74 m), S324 (36.02 m), S357 (34.78 m) and S294 (36.07 m) form an upper altitudinal band with a height range of 2.7 m. The relationship of a number of them to decaying ice and marine sediments has been described for the Millhouse/Kilbride Bay valley and it is clear that not only is this the marine limit but that it has been formed closely following if not immediately upon deglaciation.

Another altitudinal grouping occurs only slightly below this consisting of the following fragments: S346 (30.64 m), S347 (32.14 m), S351 (30.64 m), S328 (32.21 m), S334 (30.35 m), S332 (31.39 m) and S243 (31.72 m). These have a height range of 1.8 m and are separated from the topmost grouping by 1.9 m. Also occurring with these is shingle ridge SR305 (32.93 m).

1.9 m below this group occur, within a 2.5 m height range, S345 (27.69 m), S333 (26.00 m) and S293 (28.48 m). These are associated with shingle ridges SR303 (27.90 m) and S244 (28.87 m) to form a weakly developed shore zone.

A gap of 2.2 m follows after which occurs a rather distinct group having a height range of 2.8 m and consisting of S344 (23.82 m), S349 (22.44 m), S352 (20.98 m), S292 (21.29 m), S342 (23.55 m) and S335 (21.99 m). With these is associated tombolo S331 (19.19 m).

Below this altitude relatively few shoreline fragments exist, presumably because of the influence that the rock cliffline would exert upon littoral sedimentation (this presumes that it was cut by this time, an arguable assumption, but were it to be cut later it would have removed those shoreline fragments whose deposition predated it. See Chapter 8). However, two fragments, S350 (16.31 m) and S353 (17.85 m) suggest the possibility of a poorly developed shore zone at this altitude. The only remaining distinct grouping is one of the following four fragments S343 (12.17 m), S341 (11.58 m) and S336 (12.62 m). The eleven remaining fragments that have been levelled in this area extend downwards from just over 7.5 m into the present shore zone.

From the above the four maps in Fig. 42 have been prepared. These succinctly demonstrate the successive emergence of the coastline, the merging of islands and mainland and the emerging of new islands until the present day outline is attained. The exact chronology of the shore zones in question will be discussed in Chapters 8 and 10. Of particular interest at the moment is, however, Figure 42a which depicts the area whilst ice from the decaying ice sheet was still present. In addition to the sources of meltwater indicated previously should be added that of waters draining into the Stillaig Valley from Asgog Loch, a pair of depositional terraces at the head of this valley (one, T330, heighted at 36 m) apparently recording drainage from this source. A picture therefore emerges of ice rapidly retreating in the sea lochs where calving could operate as a powerful agent of ablation, whilst on land dead ice developed in those areas that had become cut-off from their sources. A similar situation has been described in the Glacier Bay area of Alaska by Field (1974) following documented retreat from a Neoglacial maximum. In the study area stagnant ice remained until at least 10 m fall of relative sea-level from the marine limit had occurred.

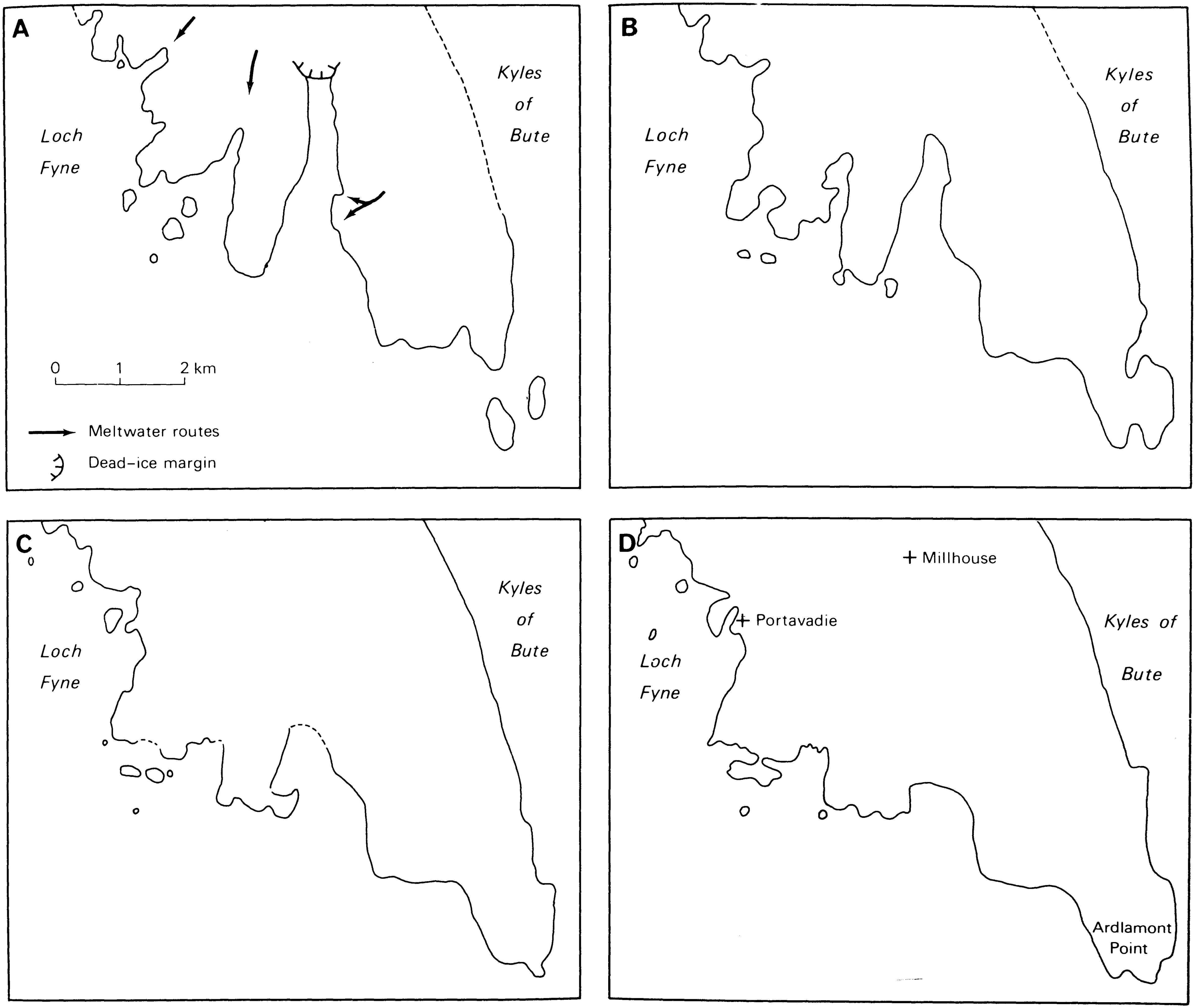
Figure 42: Coastline emergence in Portavadie/Ardlamont area.

A. At deglaciation

B. At end of Otter Ferry Stage

C. At time of Main Rock Platform

D. Present day.


A
Loch
Fyne
Kyles
of
Bute
0
1
2 km
Meltwater routes
Dead–ice margin
B
Loch
Fyne
Kyles
of
Bute
C
Loch
Fyne
Kyles
of
Bute
D
Millhouse
Portavadie
Loch
Fyne
Kyles of
Bute
Ardlamont
Point

2. Kilfinan

Kilfinan Bay on lower Loch Fyne is flanked to both N and S by the rock-cut platform and associated cliff that is the dominant coastal element of the area. Immediately inland of the bay itself, however, the cliff is engulfed by extensive fluvial and marine deposits (Fig. 43).

The junction of the platform and the cliff have been levelled at three localities, RP5 to the S of the bay and RP7 and RP8 to the N. The altitudes of these platform fragments were 7.33 m, 7.12 m and 6.70 m respectively. The abandoned cliff can be followed into the bay on both the N and S sides for several hundred metres indicating that at the time of formation the sea was able to erode in relatively sheltered localities that have been in more recent times areas of marine deposition.

The Kilfinan Burn emerges from a gorge near Kilfinan Village and it is seawards of this gorge that the river terrace and raised marine landforms are most abundantly developed. A number of low terraces are developed in the gorge bottom and at the junction of the Kilfinan and Strone burns upstream of the gorge intake there are larger terraces, the development of which indicates the former importance of the Strone Burn as the largest and uppermost terrace is more extensive along the Strone Burn than along the Kilfinan Burn. As will be shown in the next section the Strone Burn carried meltwaters from the Otter Ferry glacier and it is thought that the terrace development along the Strone Burn dates from this period.

Around the Kilfinan embayment the highest evidence of marine activity is at ca. 36 m O.D.. S156 (36.65 m) occurs on the N side of the embayment, S268 (35.17 m) at the head and S270 (35.45 m) on the S side. No section evidence is available in any of these landforms.

S156 is not related to any obvious source of material and may therefore have developed in close proximity to glacier ice. Similarly S270 is out of proportion to the small stream beside it and this may also be due to the existence of nearby decaying ice at the time of its formation. S268, however, occurs as the highest terrace overlooking the mouth of the gorge of the Kilfinan Burn. It is clearly related to drainage from the Kilfinan Burn and it is most simply correlated with the highest terrace at the junction of the Kilfinan and Strone burns at the intake of the gorge thus suggesting that it developed at the time of important drainage along the Strone Burn.

During the subsequent fall of sea-level the Kilfinan Burn built out a large delta on the surface of which is preserved a number of nearly flat terrace fragments that record specific sea-levels and that can on occasion be correlated with other raised marine landforms in the neighbourhood suggesting that they relate to times of regional shoreline formation rather than to local events. During and subsequent to the regression of the sea the Kilfinan delta has been dissected by both the Kilfinan Burn and by the stream that flows past Lindsaig farm (b, Fig. 43). The section evidence available indicates the deposits to be composed of sand and gravel. Specific sea-levels can be identified at ca. 32 m (S167 : 32.15 m and S269 : 31.68 m), at ca. 24.5 m (S267 : 25.48 m; S155 : 24.03 m; S166 : 24.79 m and S266 : 23.53 m), at ca. 20.5 m (S146 : 20.53 m and T78 : 20.98 m), at ca. 16.5 m (T79 : 16.72 m) and at ca. 14.5 m (S148 : 14.68 m and S158 : 14.16 m). Of these the sea-level at ca. 24.5 m seems particularly well-developed and as is shown in Chapter 9 it coincides with a well-marked shoreline around the southern coast of Cowal. This shoreline (CLG5) was the last to be formed prior to the disappearance of the Otter Ferry stage ice and its strong development in the Kilfinan area may be due to abundant sediment supply along the Strone Burn at this time.

Below the altitude of the lowest of the sea-levels indicated above the character of the raised marine deposits changes, shingle ridges becoming the dominant marine landform. Terraces occur as well, particularly along the flanks of the embayment. As will be established below these landforms all date from the Flandrian Period and are hence separated by a considerable time interval from the marine landforms related to the earlier marine regression.

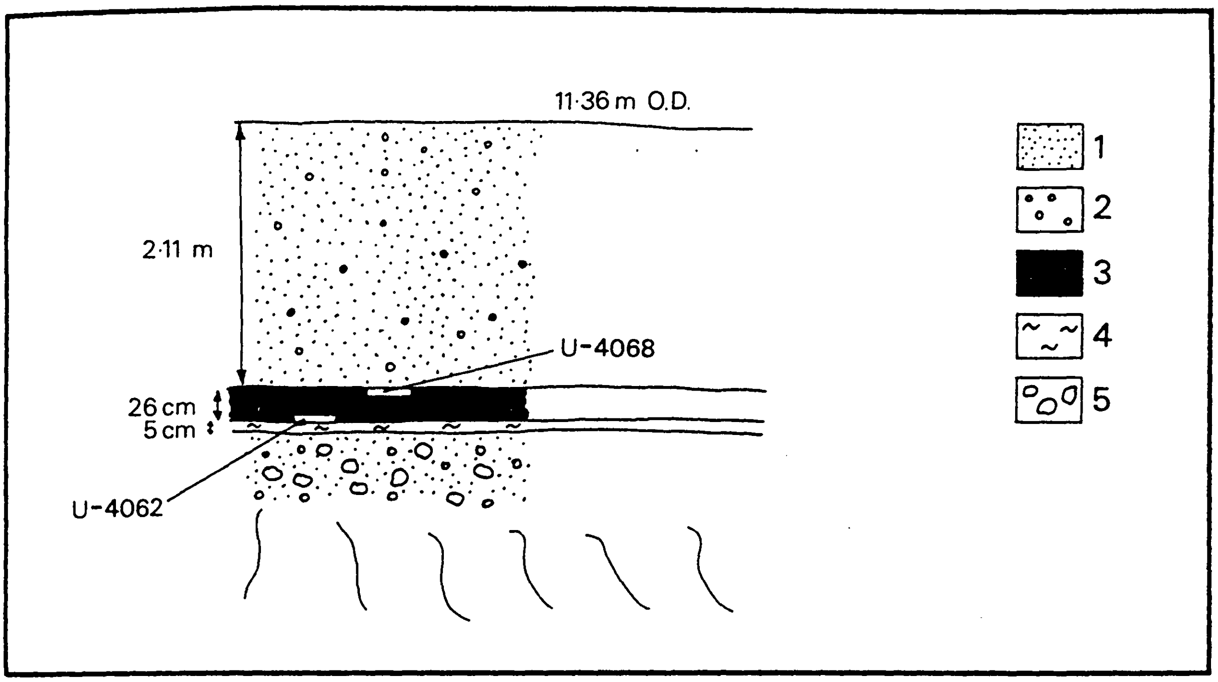
The highest Flandrian marine landform is the complex shingle ridge SR11 developed at Drum Point at an altitude of 13.12 m. This very well preserved double ridge is perched on top of the cliff associated with the Main Rock Platform. It was first noted by Clough in 1897 (Gunn <u>et.al</u>.). The uppermost deltaic deposits flanking the Kilfinan Burn are S271 to the S and the terrace marked (a) to the N. These occur at <u>ca</u>. 11.5 m. Terrace T86 beside (a) probably was largely formed at this time but it has been subsequently modified by the stream that separates it from (a). The erosion of terrace (a) by this stream has revealed the stratigraphy shown in Figure 44, a coarse well-rounded sand and gravel being overlain by a thin bed of light grey slightly organic silty fine sand that is succeeded upwards by a peat bed with intercalated light grey silty sand lenses, the section being capped by well-rounded sands and gravels, the individual clasts being often horizontally imbricated. The top of the section is at 11.36 m O.D.. This succession from bottom upwards is taken to represent fluvial (possible fluvioglacial?) deposition succeeded by relatively quiet water (marine?) silty sand, terrestrial peat with

Figure 44: Section at Kilfinan Bay showing peat sampled for radiocarbon dating

1. sand 2. well-rounded gravel

3. peat 4. silty fine sand

5. coarse gravel.


11·36 m O.D.
2·11 m
26 cm
5 cm
U-4068
U-4062
1
2
3
4
5

occasional marine incursions and then marine sands and gravels being transgressive on the peat. The peat bed was sampled top and bottom for radiocarbon assay, the bottom 2 cm giving an age of 6170 ± 80 yr B.P. (U - 4062) and the top 2 cm an age of 7630 ± 115 yr B.P. (U - 4068). Unfortunately these ages are in reverse stratigraphic order. It is thought that the upper sample is correct, however, as it is in agreement with dates from a similar section at Furnace (see below) and is also in accord with the anticipated age of the overlying Flandrian marine deposits (Chapter 10).

The remaining Flandrian deposits are represented by a complex series of shingle ridges with occasional terraces. The shingle ridges form an altitudinally descending series and their number and complexity makes it seem unlikely that they are all related to specific periods of shoreline stability and some perhaps are the product of events of local significance. S149 (8.26 m), however, is a very prominent landform and this suggests a period of stability for it to form. Cross-profile E-E' (Chapter 10) indicates certain of the characteristics of the shingle ridges, SR16 occurring landwards of the presently forming SR12 although SR16 is at a lower altitude and must have been formed earlier than SR12.

3. <u>Otter Ferry</u>

Otter Ferry lies beside the sill between the middle and the outer basins of Loch Fyne. It is most immediately notable for the very large spit that extends more than half-way across the loch at this locality (and from which it receives its name) but detailed mapping has revealed a complex set of glacial, fluvioglacial, marine and fluvial landforms and deposits that clearly indicate that this was once the terminus of a large glacier occupying Loch Fyne.

The landforms mapped are shown on Figure 45. To the N and the S of Otter Ferry (and on the opposite coast of Loch Fyne) the dominant element of the coastline is the abandoned rock cliff with accompanying platform. This feature could not be surveyed on the Otter Ferry side of Loch Fyne but it was possible to survey it at one locality at W Otter Ferry, the altitude being 8.27 m (RP301).

The most significant glacial landform is an end moraine that has been traced for more than 2.5 km from N of Tom na h-Iolaire to near Barr Gamhainn. In this distance it drops in altitude from over 180 m to just over 90 m O.D., a gradient of approximately 36 m/km. The relationship between the moraine and the topography clearly indicates that the moraine has been formed by a glacier occupying Loch Fyne. From near Barr Gamhainn to the coast no clear evidence of the continuation of the end moraine has been found. The moraine is rather small and is occasionally only a boulder moraine.

Drainage towards the moraine must have been disrupted at the time of its formation. The largest stream is the Kilail Burn which flows in a gorge along much of its lower portion in the neighbourhood of the moraine. A low peat-infilled channel crosses the watershed from the Kilail Burn to the Strone Burn (which flows to Kilfinan) above the point where the moraine crosses the Kilail Burn. At locality (a) silts and clays were observed in the confines of the Kilail Burn gorge. It is suggested that these silts and clays were deposited in an ice-dammed lake that formed at the time of formation of the moraine and that the lake drained through the channel into the Strone Burn.

SW of this location a small stream also cuts the moraine, the stream flowing in a small gorge at this locality. At the mouth of this gorge is a small esker and fluvioglacial mounds, the esker bending to be oriented down-ice.

Although the exact location of the moraine is not known seawards of Barr Gamhainn, the bedrock ridge that runs SE from Druim Madaidh is cut by a clear channel that today has only a small stream running along its bottom. Near the terminus of the channel a raised delta is located at over 30 m O.D.. It is thought that the drainage of the area between Druim Madaidh and Barr Gamhainn which at present flows N into the area occupied by ice at the time of formation of the moraine must have been dammed and formed a small glacial lake that overflowed through the above channel.

Within the confines of the moraine there are large volumes of glacial and fluvioglacial materials. The distribution of these rather amorphous deposits is strongly controlled by the small steep-sided bedrock hills that occur around Otter Ferry. The most extensive development, however, is N of Ballimore House where kames and kettle holes have been mapped though many of the kames have been washed over and had their surfaces flattened by the sea.

The highest raised marine landforms encountered up-loch of the moraine are the poorly developed delta S132 (17.78 m) and the moderately developed tombolo S134 (18.81 m) that joins a small rock hill to the larger bedrock mass SE of Otter Ferry. Two distinct and rather better developed features occur below these, S278 (16.37 m) and S161 (16.22 m). The position of S278 suggests that this is a reworked kame but no similar explanation seems possible for S161. Together they are taken to imply a period of shoreline formation.

In the kame and kettle area N of Ballimore House a group of features have a common surface altitude of slightly over 14 m (S144 : 14.43 m; S145 : 14.35 m and S279 : 14.74 m). This altitude is similar to that of S159 (14.13 m), a former delta at the mouth of the Kilail Burn, this suggesting another period of shoreline formation.

A further group of features occurs below this at ca. 13 m. S160 (13.52 m) is a large very well developed feature N of the Kilail Burn, S126 (13.25 m) is a moderately developed raised delta at the mouth of the Largiemore Burn about 2 km uploch of Otter Ferry and S277 (12.81 m) is a very large area of reworked kame deposits N of Ballimore House. S277 overlooks 2 kettle holes on its S side. An attempt was made to auger and pit in these kettle holes but progress was impeded by gravel at ca. 70 cm depth. Above this was a fawn coloured organic silty clay with grit and this was capped by 30 cm of dark brown muddy peat. The basal gravels may have been washed in at the time of the reworking of the kame deposits but the apparently easy drainage through the surrounding sand and gravel has prevented the development of extensive basin sediments that may have contained information of marine conditions.

The seaward edge of the kame and kettle deposits N of Ballimore House is marked by a large bluff apparently cut for the greater part in drift but with one area of bedrock exposed. It is possible that this cliff is the counterpart of the rock-cut cliffline noted previously but the lack of continuity with that feature does not allow a firm correlation to be made. One other shoreline fragment occurs above this cliff, S143, which at 11.78 m is coincident, together with S122 (11.31 m) and S128 (11.10 m), with the uppermost level of Flandrian marine deposits in this area (Chapter 10). The position of S143 indicates that the cliff was not formed at the maximum of the Flandrian transgression.

Below the uppermost Flandrian shoreline there are a series of groups of shoreline features that suggest distinct periods of shoreline formation during the overall regression. These groups occur at ca. 10 m (S131 : 10.19 m and S163 : 9.81 m); ca. 8.5 m (S123 : 8.49 m and S135 : 8.27 m); ca. 7.5 m (S124 : 7.80 m; S162 : 7.78 m; S130 : 7.67 m; S142 : 7.51 m; S127 : 7.06 m and S140 : 6.87 m); ca. 5.0 m (S136 : 5.22 m and S164 : 5.10 m); ca. 4.2 m (S139 : 4.26 m; S137 : 4.23 m; S125 : 4.18 m and S138 : 4.05 m) and a group of poorly developed fragments below 4.0 m (S141 : 3.95 m; S129 : 3.78 m and S133 : 3.60 m).

There is little sedimentary evidence relating to the Flandrian deposits. Close to feature S138 a small section at the head of the foreshore revealed peat lenses intercalated with beach gravels suggesting a transgression prior to the formation of S138. A road has been built, however, ca. 5 m back from the top of the section and although the gravels are horizontally imbricated and the section appears undisturbed the possibility remains that the peat was buried during road construction.

4. Strathlachlan

Strathlachlan joins Loch Fyne from the E at the narrows near Minard between the upper and middle basins of the loch. For the 1.5 to 2 km nearest its junction with Loch Fyne the valley bottom is occupied by river terraces (Fig. 46). From close to the Garvalt road bridge to near the school the river flows in a rock-cut gorge. Upvalley from here glacial and fluvioglacial landforms extend almost as far as Leanach and across the watershed towards Loch Fyne SE of Newton.

Hummocky moraines, on occasion separated by enclosed peat-filled depressions, are particularly abundant E of Sunfield. Channels occur within and along the edges of these deposits and downvalley the limit of the channels and hummocky moraines coincides with the start of a series of small terraces that can be traced as far as the intake of the gorge. This suite of landforms suggests the former presence of an ice margin, the terraces being outwash beyond the glacier limit. A meltwater channel crosses the watershed between Loch Fyne and Strathlachlan. This channel has an up-and-down long profile and demonstrates the contemporaneity of the last ice in Strathlachlan with ice in Loch Fyne.

The terraces in the lower part of Strathlachlan commence at the mouth of the gorge and can be traced downvalley to the present coast. Levelling of the terraces and plotting of the heights on a height-distance diagram, the projection plane for which follows the axis of the valley, allow more than one group of terrace fragments to be distinguished (Fig. 47). The upper terrace fragments (T310, T317, T312) occur in the middle portion of the valley and the terrace fragments are clearly the remnants of a once more extensive terrace. It is not clear whether they correlate with either T308 or T309 towards the mouth of the valley. The better developed terraces commence at the mouth of the gorge at <u>ca</u>. 14 m O.D. and can be traced almost continuously down to <u>ca</u>. 6 m O.D.. It seems likely that these are continued downvalley by terrace fragments T318, T319 and T307. None of these terraces can be related to particular sea-levels, although possibly both T308 and T309, the latter having no discernable gradient over a 150-200 m length, are closely controlled by sea-level. There is only one undoubted marine landform in the valley, this being the shingle ridge that occurs at the head of the present beach, the altitude of the crest of which is 3.37 m O.D. (SR302).

The terraces in the valley relate to sea-levels of below 6 m O.D. and are hence Flandrian in age. They have been formed by the reworking of sediment introduced into the valley at earlier times, presumably much of it when ice occupied the valley near Sunfield. The paucity of evidence in the lower part of the valley for earlier terraces unfortunately means that the ice limit cannot be related to any particular sea-level. The highest shoreline identified in this part of Loch Fyne is at *ca*. 21 m O.D. and it can be inferred that the ice limit in Strathlachlan was formed when the sea was at or below this level.

5. The Head of Loch Fyne

Glen Fyne is a glacially fashioned valley that continues inland from the head of Loch Fyne along the line of the Tyndrum Fault. Numerous small streams descend precipitous gullies at the foot of which alluvial fans have been constructed. One such stream deserves particular mention. This is the one that drains part of the Garabal Hill granite (Glen Fyne granite) to the E of Glen Fyne which it reaches by cascading into a spectacular hanging gorge known as Eagles Fall, the northern side of which is *ca*. 150 m high. The floor of Glen Fyne is dominated by wide terraces and alluvial flats close to the present river level (Fig. 48). Near the head of Loch Fyne the surface of terrace S273 is interrupted by low mounds and swales and associated channels interpreted as ice-decay landforms washed over by the sea. NE of Auchreoch an area of hummocky moraines occurs along the eastern side of the valley bottom.

Druim na Muclaich, a bedrock hill opposite Auchreoch, exhibits a number of clear striated and ice-moulded rock surfaces, the striations, expectably, being oriented along the valley. Between Druim na Muclaich and the steep western side of Glen Fyne is a large rock-cut meltwater channel which at its southern end turns to cut through the hill and terminates with a rather small alluvial fan at its mouth. Two of the gullies descending the side of Glen Fyne are concordant with this channel but a third immediately to the S is not, the mouth of this gully terminating on the slope. These relationships suggest that the gullies are in part at least due to the work of meltwater, an idea supported by the set of gullies to the SW of Druim na Muclaich that trend across the slope for much of their length. Gullies such as these are a common feature of the steep glacially eroded valley walls in the NE of the study area and it seems likely from the above example that the role of meltwater is important in their development.

The terraces of Glen Fyne were levelled at regular intervals along their lengths where practicable and the results plotted on a height-distance diagram the projection plane for which parallels the long axis of the valley (Fig. 49). The surveyed terrace fragments fall into two distinct groups, those well within the confines of the valley (T56, T57) that have a clear down-valley gradient and those around the head of the sea-loch that appear to have no or very little overall gradient. This latter group of terraces is related to former sea-levels.

Terraces T56 and T57 cannot be correlated for they are strongly influenced by the large alluvial fan at the foot of Eas Riachan. The very steep gradient of T57 is presumably due to the abundant supply of sediment from this fan whilst conversely the relatively low gradient of T56 is likely to have been due to influence of the same fan on the down-valley portion of T56.

Of the terraces at the head of Loch Fyne, T1, S273 and S274 were mapped originally as river terraces on the basis of their morphology but the extremely low gradients, particularly of S273, suggest they are the uplifted portions of former intertidal areas at the head of the loch. Analogy with the present shore suggests that S273 relates to a sea-level ca. 5 m above that of the present and S274 ca. 2 m above that of the present.

A higher group of terrace fragments (S275, S276, S95, S96) occurs in the height range 13-14 m O.D.. These features were apparently part of one terrace that has been dissected by some of the numerous streams that descend the hillside. Sections in these terraces reveal nearly horizontal beds of sand and gravel and the terraces are interpreted as former intertidal flats related to a sea-level ca. 12 - 12.5 m above the present sea-level. Immediately SW of S96 the hillslope is very irregular in appearance due to a landslip that cuts across and hence must post-date the formation of S96.

Two small terrace fragments, S94 and S2 occur at ca. 10 m O.D. and may relate to a former sea-level but both of these are poorly preserved. S2 occurs on the seaward side of a short steep rock slope that trends parallel with the shore and which is thought to be a degraded part of the cliff associated with the Main Rock Platform (Chapter 8). The only other terrace fragment levelled in this area occurs beside a stream on the top of the degraded cliff. No other terrace fragment has been found at this altitude in this area nor has any apparent correlative been found in the study area (Chapters 9 and 10) and S1 is thought to have been formed fortuitously on the break of slope at the top of the former cliff.

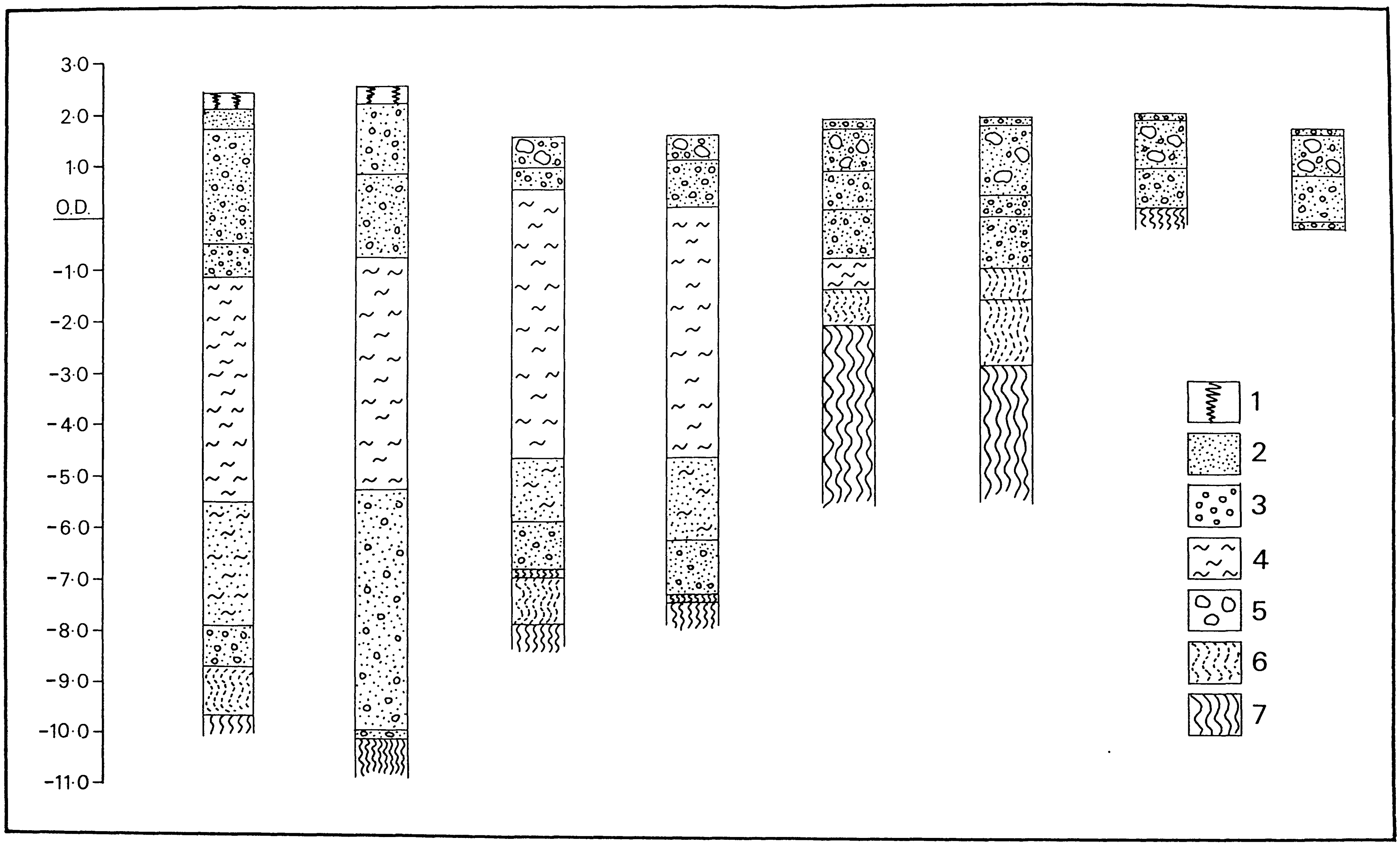
During site investigation for a new road bridge across the River Fyne (slightly seawards of the bridge on Figure 48) a line of 8 boreholes was sunk at approximately 90° to the eastern valley side. These are shown in Figure 50. Holes 3 to 8 were across the present river bed and holes 1 and 2 were on a low terrace to the W of the river. The boreholes reveal a bedrock surface dipping towards the valley centre from close to O.D. at the eastern side of the River Fyne to almost -10 m O.D. about one third of the distance across the valley. No definite glacial sediments have been identified in the boreholes. A band of sorted sand and gravel varying in thickness up to ca. 5.5 m that has been identified in the deeper holes may be a fluvioglacial deposit formed as the ice margin retreated from this location. Overlying these sands and gravels is a major horizon of grey silt (sandy towards the base) that lenses out eastwards, the three most easterly holes having failed to penetrate it. Over 6 m thick towards the deeper part of the valley the grey silt has a maximum surface altitude of 0.55 m O.D.. It is thought to be the equivalent of other grey silts and clays (such as those that occur near Inveraray, around Loch Long and Loch Goil) that were laid down soon after the retreat of the Loch Lomond Readvance glaciers. Such silts and clays are rarely found above O.D. and are very sparingly fossiliferous. They are distinct from the darker grey carse clays of the Flandrian transgression (such as those that occur at the head of Loch Riddon) which have a readily recognizable organic content. Overlying these silts and extending on to bedrock where the silts are absent is a bed of sorted sand and gravel. This sand and gravel is thought to be intertidal sediment formed during the construction of the low river terraces of the River Fyne. It is the counterpart of the sand and gravel exposed in the sections cut in terraces S95 and S96. In the present river bed bores 3 - 8 record a coarse channel deposit up to 1.5 m thick.

6. Inveraray/Glen Shira

North of Inveraray, Loch Fyne is joined by two large valleys, Glen

Figure 50: Boreholes for new road bridge, Glen Fyne.

1. topsoil 2. sand 3. gravel 4. silt, clay 5. cobbles 6. weathered bedrock 7. fresh bedrock.



3·0
2·0
1·0
O.D.
-1·0
-2·0
-3·0
-4·0
-5·0
-6·0
-7·0
-8·0
-9·0
-10·0
-11·0
1
2
3
4
5
6
7

Aray from the NW and Glen Shira from the N. The mouths of both these valleys are occupied by river terraces and raised marine deposits and in Glen Aray in particular there are numerous glacial deposits in the valley bottom within 2 km of the present coastline. Around Inveraray building has obscured many landforms but southwards to the mouth of the Douglas Water the coast is fringed by raised marine landforms, chiefly depositional in origin although near Dalchenna the cliff associated with the Main Rock Platform (Chapter 8) is very well displayed. This cliff can also be identified between Glen Aray and Glen Shira and around Strone Point between Loch Shira and Loch Fyne, three collapsed caves having been located in the cliff at this latter locality.

(a) Glen Shira

The lower portion of Glen Shira is largely occupied by the tidal Dubh Loch, the section of valley between Dubh Loch and the sea being occupied by a series of terraces and a broad channel (Fig. 51). The valley side to the E of Dubh Loch is extensively disrupted by landslipping but along the W side of the valley traces of a rock cliff occur a few metres above the present loch level. These cliff remnants are thought to correlate with the similar features around Strone Point and near Dalchenna, implying that this stretch of Glen Shira was occupied by the sea when the cliff was formed. N of Dubh Loch the valley bottom is floored by a wide area of alluvium, suggesting that the loch was once more extensive than at present.

The terraces between Dubh Loch and Loch Shira were levelled at the localities shown in Figure 51. A height-distance diagram with its projection plane parallel to the axis of the valley was constructed (Fig. 52). Prominent on the height-distance diagram is feature M1, a long flat-topped ridge that flanks the southern margin of Dubh Loch. Terrace T10 is separated from M1 by a broad channel and the upper part of M1 appears to correlate with the upvalley projection of T10. The major slope on the Dubh Loch side of M1 is unlikely to have been the result of wave erosion and is considered to be an ice-contact slope, Dubh Loch having originated as a large dead-ice hollow. M1, T10 and the channel between would therefore have been formed at the time that ice occupied the side of Dubh Loch, this providing an explanation for the intake of the channel being perched more than 10 m above the present loch level.

The feature M1 does not slope evenly along its length, an almost horizontal portion being revealed by the levelling at ca. 14 m O.D. and a clear break of slope occurring between M1 and S287 at the southern end of M1. The feature S287 has been partly excavated for sand and gravel and sections reveal a poorly sorted mixture of gravel, cobbles and boulders with occasional beds of sand. The presence of boulder-sized material in the deposit makes it unlikely that this is a product of a stream draining Dubh Loch and it is interpreted as fluvioglacial material deposited at the same time as M1 and reworked at a later period to give the surface form of S287. The surface altitude of S287 (12.30 m O.D.) corresponds with that of T7 (12.32 m O.D.) on the opposite side of the valley, suggesting a common control on their formation which in this location was most probably sea-level. This conclusion that Dubh Loch was once occupied by the sea at a higher level than the present is in accord with an original study by Gregory (1857) who used marine diatoms as evidence that the sea was once at least ca. 9 m (30 feet) above the present level of Dubh Loch. Terraces have been heighted at a number of distinct levels below that of T7 and S287 and given the present tidal connection between Dubh Loch and the sea these terraces must have formed largely in relation to changing sea-levels. The best preserved evidence is that of S11 (7.62 m O.D.) although terraces are also present at 10.60 m O.D. (S12) and ca. 4.7 m O.D. (T8, T5).

(b) Glen Aray

For approximately 2 km between the exit of a rock gorge and the sea the River Aray is flanked by a series of terraces (Fig. 53). In the vicinity of the gorge are numerous small hummocky moraines but in the lower section of the valley where the terraces are best developed few glacial deposits have been identified. A constriction occurs in the valley around T18 presumably due to bedrock control, almost separating the wide mouth of the valley from the upstream terraces. The terraces have been levelled where practicable, the town of Inveraray and the Duke of Argyll's castle and gardens being the principal hindrances to a more extended survey. A height-distance diagram has been constructed (Fig. 54) in the usual manner.

In the upper section of the valley the largest terrace fragment surveyed, T14, has a very steep gradient particularly at its head, presumably due to the influence of alluvial fans from side streams as well as the emergence of the River Aray from the gorge. T14 clearly correlates with T16 on the opposite side of the valley, this terrace fragment being traceable to below 11 m O.D., suggesting a sea-level at or below this altitude at the time of formation of the terrace. The only other terrace surveyed in the upper section of the valley, T17, is not extensive. It is several metres higher than the neighbouring T16.

The constriction in the valley is flanked by terrace T18 and T14a extends along the whole length of this section. Tree cover hampered the mapping of T18 and the levelling results suggest it may not be a single terrace. It is difficult to correlate it with any other terraces in the valley. T14a was originally mapped as a continuation of T14 (from which it is separated by a group of farm buildings) but the levelling demonstrates that this correlation requires T14 to have been horizontal for a distance of over 400 m, a highly improbable occurrence given the rather steep gradients of the levelled sections of T14 and T14a. The latter terrace is therefore regarded as a separate feature.

At the mouth of Glen Aray a number of wide terraces have been mapped. Of these, S13 appears to be oldest, being related to a sea-level of ca. 12.4 m O.D. which corresponds with the level of terraces T7 and S287 at the mouth of Glen Shira. The next oldest terraces are apparently S14 and S15, which relate to a sea-level at ca. 10.5 m O.D. and correlate with terrace S12 in Glen Shira. The height-distance diagram seems to suggest that T15 correlates with S13, but T15 is clearly below S14 in the field and therefore must relate to a sea-level below 10.5 m O.D.. It is possible that the river at the time of formation of T15 executed a sharp bend at the southern edge of the valley beside S14, rather as the present river does farther north, flowing across the valley and reaching the sea at a lower altitude. Of the remaining features SR1 is noteworthy, for shingle ridge development does not seem to have characterised the mouth of the Aray at other times.

On the foreshore NE of the mouth of the River Aray (Fig. 53) an area of organic material was discovered. Thick lenses of organic material, particularly twigs, leaves and mosses, are intercalated with beds of shelly grey sand and sandy gravel, the whole sequence resting upon an apparently non-shelly light grey silty clay. The total thickness of the deposit was ca. 65 cm. A sample from the bottom 4 cm of the organic material gave a radiocarbon age of 5,335 $\pm$ 90 yr BP (U-4067). This date is much older than was expected for it was originally thought that the organic material would have collected in intertidal conditions, that is, once the sea had regressed from the Flandrian maximum to close to its present level. In view of the date (and the sample would have to be excessively contaminated to produce this age from a sample the true age of which was perhaps ca. 1,000 yr BP) it is considered that an intertidal origin for the organic material is unlikely for this would imply extreme fluctuations of sea-level for which there is no other evidence in the study area or elsewhere around the northern British coasts (cf. Sissons and Brooks, 1971; Tooley, 1978). Rather, it must be concluded that the organic material has been deposited below low-water mark and that it illustrates the low net sedimentation in the nearshore zone during much of the Flandrian period. In this connection it is relevant to note the rather thin bed of sand and gravel in the boreholes at the head of Loch Fyne that was inferred to have been deposited during the Flandrian (Chapter 7.5) in a location similar to that at the mouth of Glen Aray.

SSW of Inveraray the coast is flanked by a low undulating area of ground attributed to the '100-foot' raised beached by the Geological Survey (Hill et al.,1905). Much of the area near the shore is obscured by buildings but certain raised marine features have been mapped. The greatest altitude at which they occur is slightly in excess of 14 m O.D. and above this level only bedrock and glacial material have been observed.

The highest terrace is S191, which has a mean altitude of 14.22 m O.D.. Below this feature shingle ridges are well developed (SR20 - 13.08 m O.D.; SR21 - 13.12 m O.D.). Although S191 coincides with the altitude of the uppermost Flandrian shoreline in this area (Chapter 10) it may be that the origin of S191 is that of a shingle ridge that has been built against the bedrock behind and thus lost its ridge form. A similar origin is suggested for S181 to the S. This feature is flat-topped at one end but forms a ridge at the other end although altitudinally it is almost horizontal along its length.

The most consistent evidence for a former stand of the sea in this area is provided by the shingle ridges SR20 (13.08 m O.D.), SR21 (13.12 m O.D.) and S181 (13.33 m O.D.). They are of limited use in inferring sea-level changes (Chapter 3) but they may relate to the sea-level of ca. 12.4 m O.D. that has been identified at the mouths of Glen Aray (S13) and Glen Shira (T7 and S287).

Other raised marine features have been surveyed in this area at 8.00 m O.D. (S182) and 7.03 m O.D. (S190).

7. Furnace

To the W of the village of Furnace on the western side of Loch Fyne is a group of glacial and marine deposits (Fig. 55). Much of the lower ground around Goatfield Farm is covered with till into which a series of meltwater channels has been cut, the channels tending to converge towards the S in an area that is now occupied by a number of raised marine landforms. N of Furnace the Leacann Water flows in a gorge which towards its mouth is floored with small terraces. Where the gorge opens out the terraces are more extensive and provide the land for the village of Furnace and the one-time Powder Works, effectively precluding accurate mapping and surveying. NE of Furnace the hill Dun Leacainn forms a large amphitheatre overlooking the Leacann Water gorge. Below the hill and descending into the gorge is a massive irregularly undulating slope that is suggestive of a very large landslide perhaps pre-dating a period of glaciation.

At the northern (intake) end of the Leacann Water gorge there is a wide expanse of alluvium at an altitude of ca. 45 m O.D. (estimated from map spot heights). The floor of the valley to the NE of this alluvial expanse is occupied by numerous small fresh hummocky moraines that terminate at the edge of the alluvium. The hillslope to the E has a drift limit that descends to the edge of the alluvial flat close to the termination of the hummocky moraines. A stream that descends the flank of Dun Leacainn is diverted to run across the slope where it encounters the drift limit. The other streams that converge on the alluvial expanse have constructed alluvial fans. The limit of the hummocky moraines and the associated drift limit are taken to mark the limit of a glacier that occupied the Auchindrain valley (along which the hummocky moraines can be traced almost continuously) and apparently also overtopped the watershed between the Auchindrain valley and Loch Fyne for hummocks are traceable up to the watershed.

The existence of a glacier in Loch Fyne of sufficient magnitude to reach ca. 200 m altitude only a short distance NE of Dun Leacainn poses a problem for the drainage along the present valley of the Leacann Water for such a glacier would presumably block the mouth of this river. This perhaps is the explanation of the set of meltwater channels around Goatfield Farm. The largest of these channels, the intake of which overlooks the Leacann Water, commences at an altitude of ca. 50 m O.D.

(map estimate), a level close to that of the surface of the alluvial flat at the Auchindrain glacier terminus. It is suggested that at the time the last glacier occupied the Auchindrain valley a glacier also occupied the area around Furnace, blocking the valley of the Leacann Water and impounding a small lake, the waters of which escaped subglacially or submarginally along the series of meltwater channels around Goatfield Farm. There are, however, no constructional glacial landforms in the vicinity of Furnace that can be used to infer the margins of the Loch Fyne glacier.

In a stream section SW of Furnace ('a', Fig. 55) a grey (weathering fawn) till was observed to overlie a stiff grey clay with occasional stones and dark unctuous bands. The till contains small shell fragments thus suggesting a readvance origin for the last ice to cover this area.

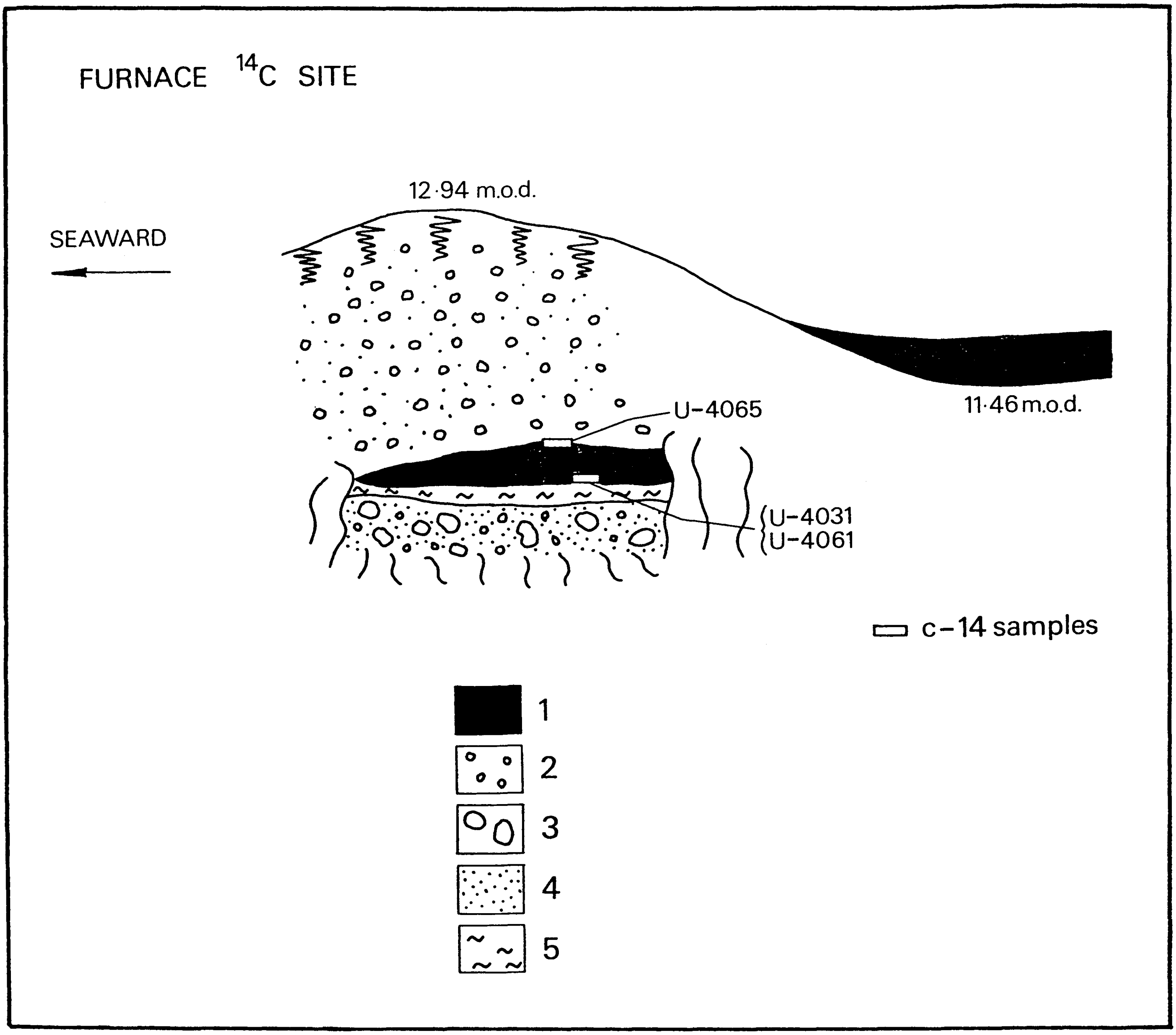
The meltwater channels converge on an area now characterised by a series of raised marine landforms. It is thought that these marine landforms are the result of the reworking of material originally transported to this area by the meltwater streams. Terraces occur at various altitudes and have been surveyed in the usual manner. The uppermost terrace (S177) occurs at 14.82 m O.D. but it was classified as 'poor' and is not considered reliable as an indicator of former marine activity. Of the others the highest and best developed are S173 and S176 at 13.36 m O.D. and 12.94 m O.D. respectively. These are thought to correlate despite the slight difference in altitude as the back of S176 has been eroded by a stream and the heights relate to points seaward of the original break of slope at the back edge.

S176 is of considerable interest as the stream has cut across it revealing the section shown in Figure 56. A thin bed of peat (maximum thickness 37 cm) is overlain by the sands and gravels that compose the bulk of the landform. Below the peat is a thin layer of grey silty sand with organic traces and below this a rather coarse poorly sorted bed of cobbles, sand and gravel. Farther down the stream this coarse layer is seen to overlie the grey shelly till mentioned previously. The hollow on the landward side of S176 was cut by the stream that eroded the present section and is now floored by peat. Probing with an aluminium rod revealed the base of the peat in the hollow to be higher than the top of the peat in the exposure and it is thus unlikely to be connected with it.

The sorted sands and gravels, in which individual pebbles have a generally horizontal imbrication, are interpreted as being marine in origin. This interpretation is consistent with the morphology and position of the deposit. Together with the buried peat this is evidence for a marine transgression. The grey silty sand underlying the peat is similar to that in the section described at Kilfinan (Chapter 7.2). It is evidence for quieter water sedimentation than the gravels, being similar to the carse deposits found around the head of Loch Riddon (Chapter 6.3). This is suggestive of a marine origin and together with the overlying peat is evidence for a marine regression.

Two 2 cm thick samples were taken for radiocarbon assay, one from the top and the other from the base of the peat. The basal sample was assayed twice, the second time in a larger counter that allowed greater precision of measurement. The ages were 7,810 $\pm$ 200 yr BP (U-4031) and 8,035 $\pm$ 85 yr BP (U-4061). The sample from the top of the peat gave 7,290 $\pm$ 90 yr BP. There is no evidence for a break in sedimentation from

Figure 56: Section of Furnace showing peat sampled for radiocarbon dating. 1. peat 2. gravel 3. pebbles and cobbles 4. sand 5. silty sand.

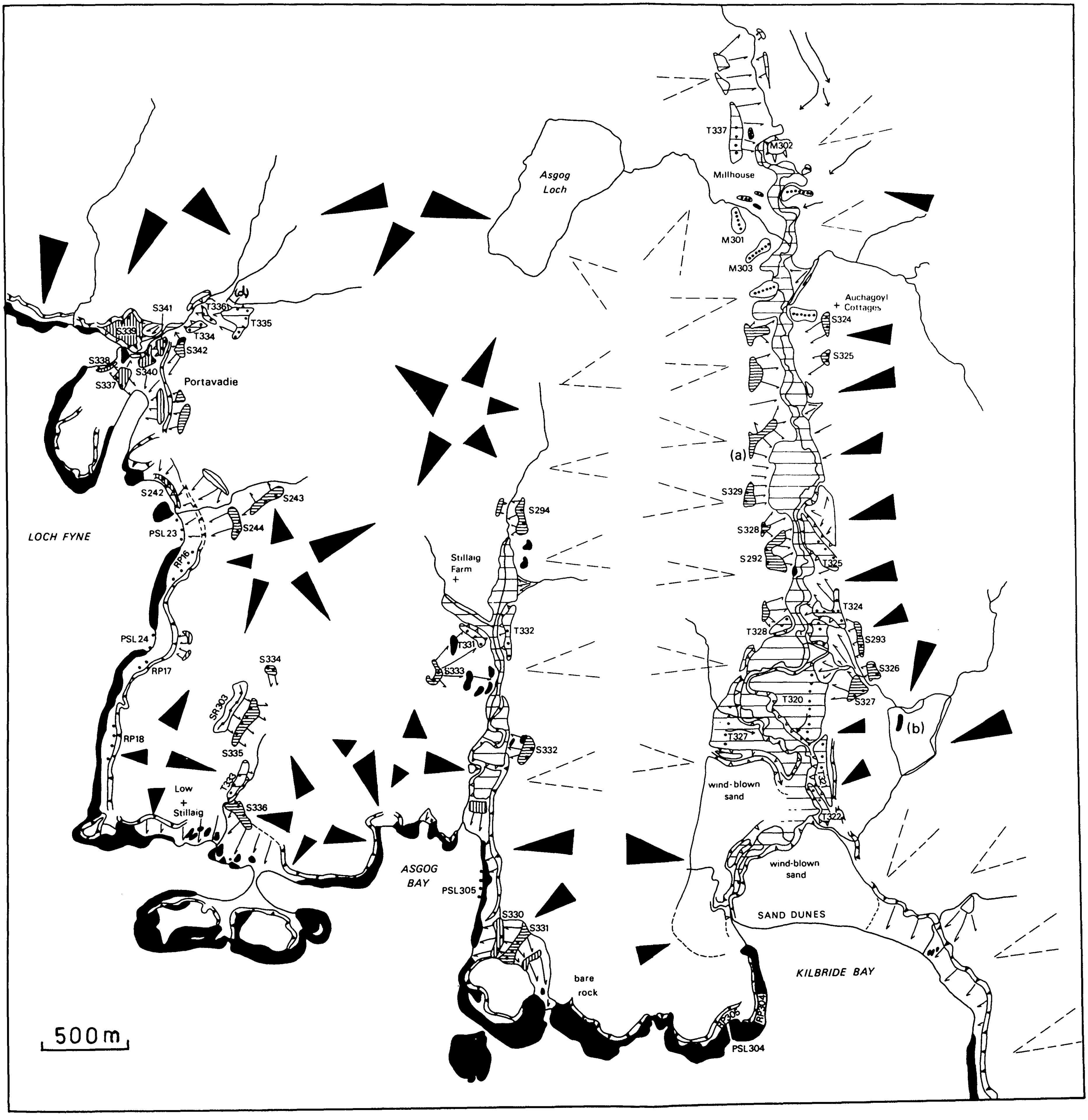


FURNACE 14C SITE
SEAWARD
12·94 m.o.d.
11·46 m.o.d.
U-4065
U-4031
U-4061
c-14 samples
1
2
3
4
5

the grey silty sand into the peat although such a break is possible. The basal date is therefore minimal for the time that the sea receded from this level. The top of the peat is irregular in nature and this suggests erosion. The sample was collected from the highest part of the peat but the date can only be a maximum for the time the marine transgression passed this altitude.

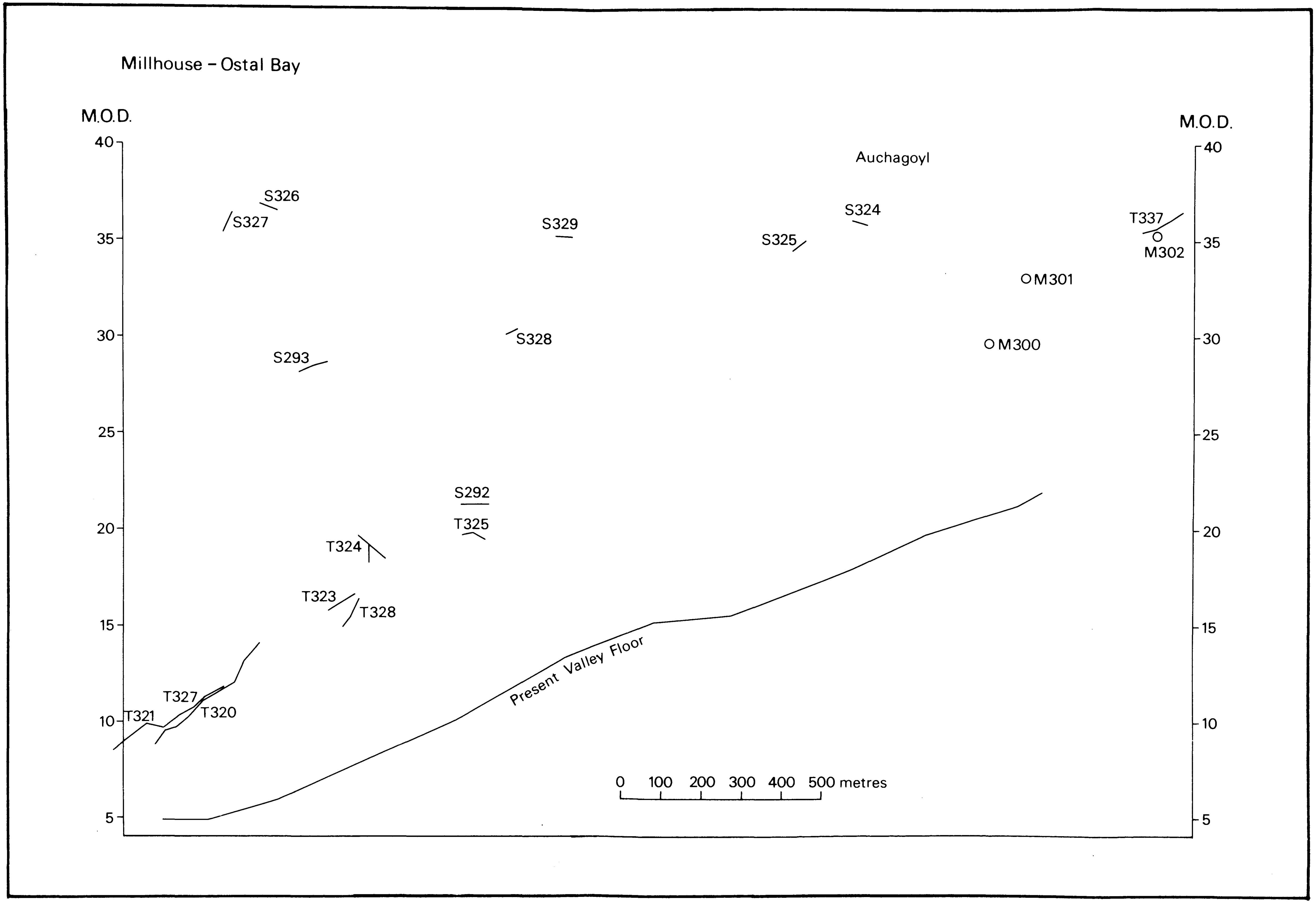
Terraces have also been heighted at 11.12 m (S174), 10.62 m (S175), 8.71 m (S172), 7.75 m (S178), 7.41 m (S171) and 4.95 m O.D. (S179) although S175, S172 and S179 were all classified as 'poor'. The evidence suggests distinct former sea-levels at ca. 13.4 m O.D. (S173, S176), ca. 11.1 m O.D. (S174) and ca. 7.6 m O.D. (S171, S178). All these formed after ca. 7,300 yr BP.

Figure 38: Geomorphological map of the Millhouse/Portavadie area.



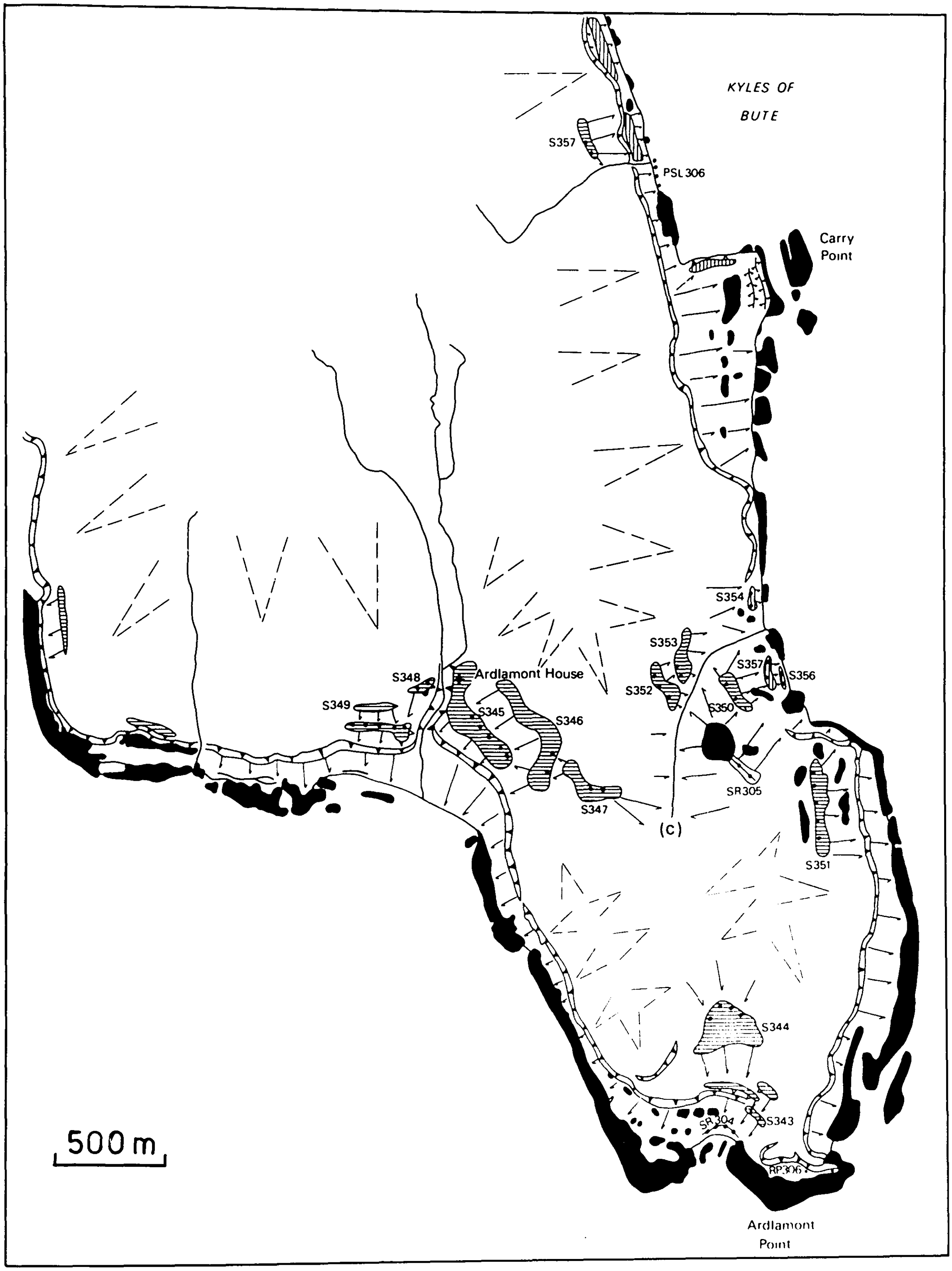
Asgog Loch
Millhouse
T337
M302
M301
M303
Auchagoyl Cottages
S324
S325
S341
T336
T335
S339
T334
S342
S338
S340
S337
Portavadie
(a)
S329
S242
S243
S244
PSL 23
LOCH FYNE
RP16
S294
S328
S292
T325
Stillaig Farm
T324
T332
T328
S293
PSL 24
RP17
T331
S333
S334
S326
T320
S327
SR303
S335
RP18
T327
(b)
S332
wind-blown sand
T321
T333
Low Stillaig
S336
T322
wind-blown sand
ASGOG BAY
PSL 305
S330
S331
SAND DUNES
KILBRIDE BAY
bare rock
500 m
RP305
RP304
PSL 304

Figure 39: Height-distance diagram of the terraces between Millhouse and Ostal Bay.



Millhouse – Ostal Bay
M.O.D.
40
35
30
25
20
15
10
5
M.O.D.
Auchagoyl
S326
S327
S329
S324
S325
T337
M302
M301
S328
S293
M300
S292
T325
T324
T323
T328
T327
T321
T320
Present Valley Floor
0 100 200 300 400 500 metres

Figure 41: Geomorphological map of Ardlamont area.



KYLES OF
BUTE
S357
PSL306
Carry
Point
S354
S353
S357
S356
S352
S350
S348
Ardlamont House
S349
S345
S346
SR305
S347
(C)
S351
S344
SR304
S343
RP306
Ardlamont
Point
500 m

Figure 43: Geomorphological map of Kilfinan Bay.

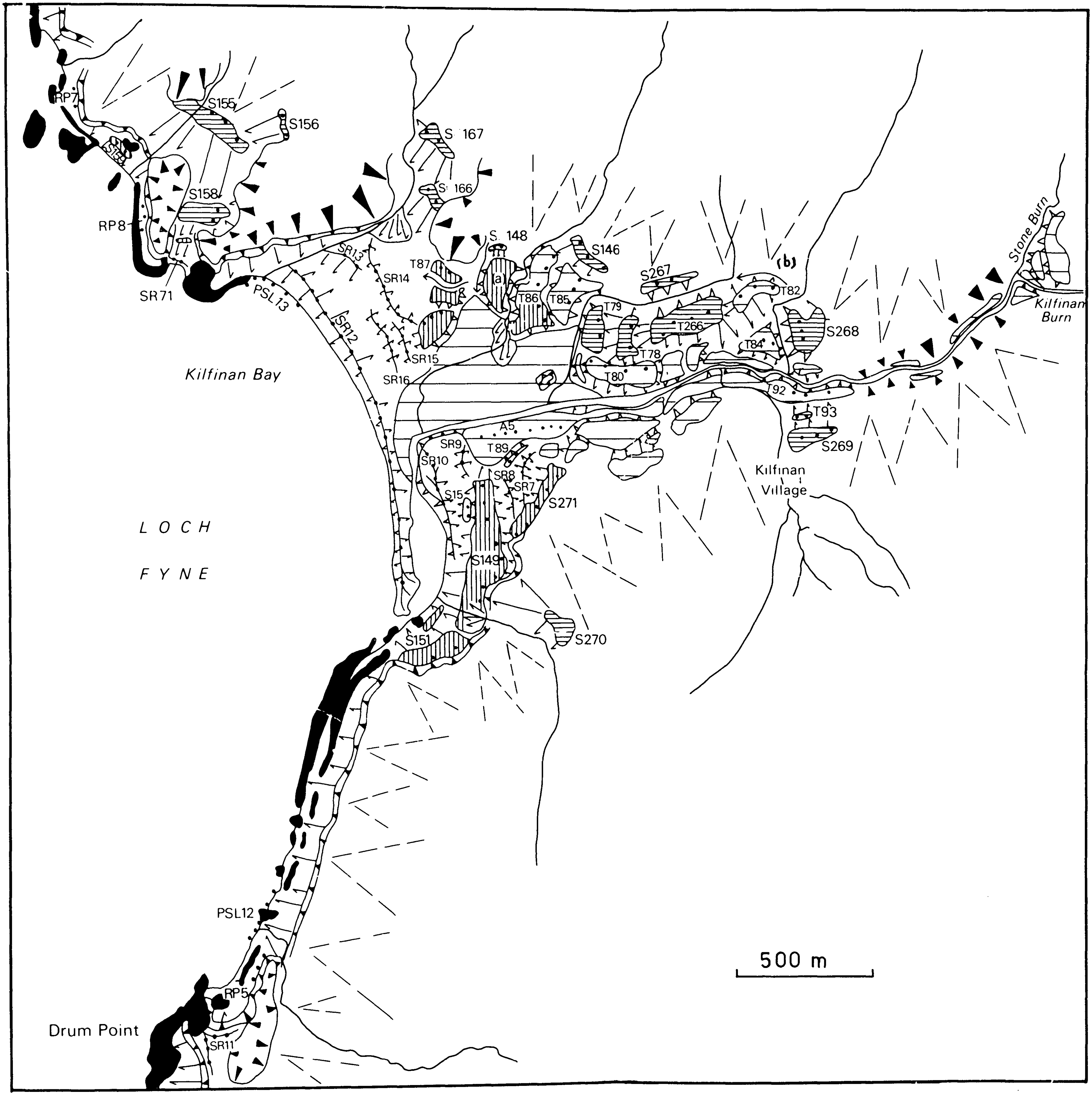

RP7
S155
S156
S167
S166
S158
RP8
S148
S146
S267
(b)
T82
Stone Burn
Kilfinan Burn
SR71
PSL13
SR13
SR14
T87
(a)
T86
T85
T79
T266
S268
T84
T78
SR12
SR15
SR16
Kilfinan Bay
T80
T92
A5
T93
S269
SR9
T89
SR10
SR8
SR7
S271
S15
Kilfinan Village
LOCH
FYNE
S149
S270
S151
PSL12
500 m
RP5
Drum Point
SR11

Figure 45: Geomorphological map of Otter Ferry Area.

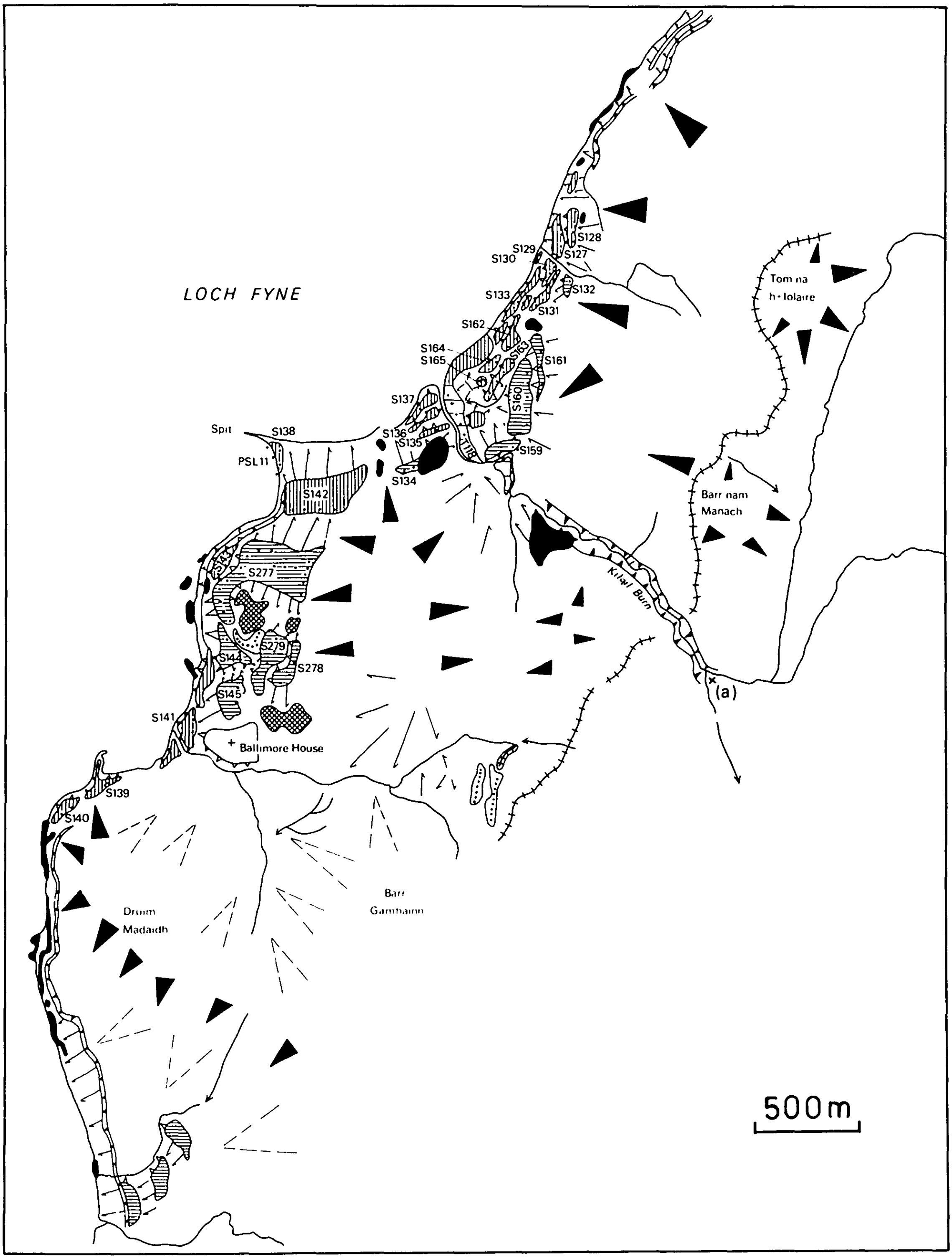

LOCH FYNE
S128
S129
S130
S127
S132
S133
S131
S162
S164
S165
S161
S163
S160
S137
Spit
S138
S136
S135
S134
S159
PSL11
S142
S277
S279
S144
S278
S145
S141
Ballimore House
S139
S140
Tom na h-Iolaire
Barr nam Manach
Kilail Burn
(a)
Druim Madaidh
Barr Gamhainn
500m

Figure 46: Geomorphological map of Strathlachlan.

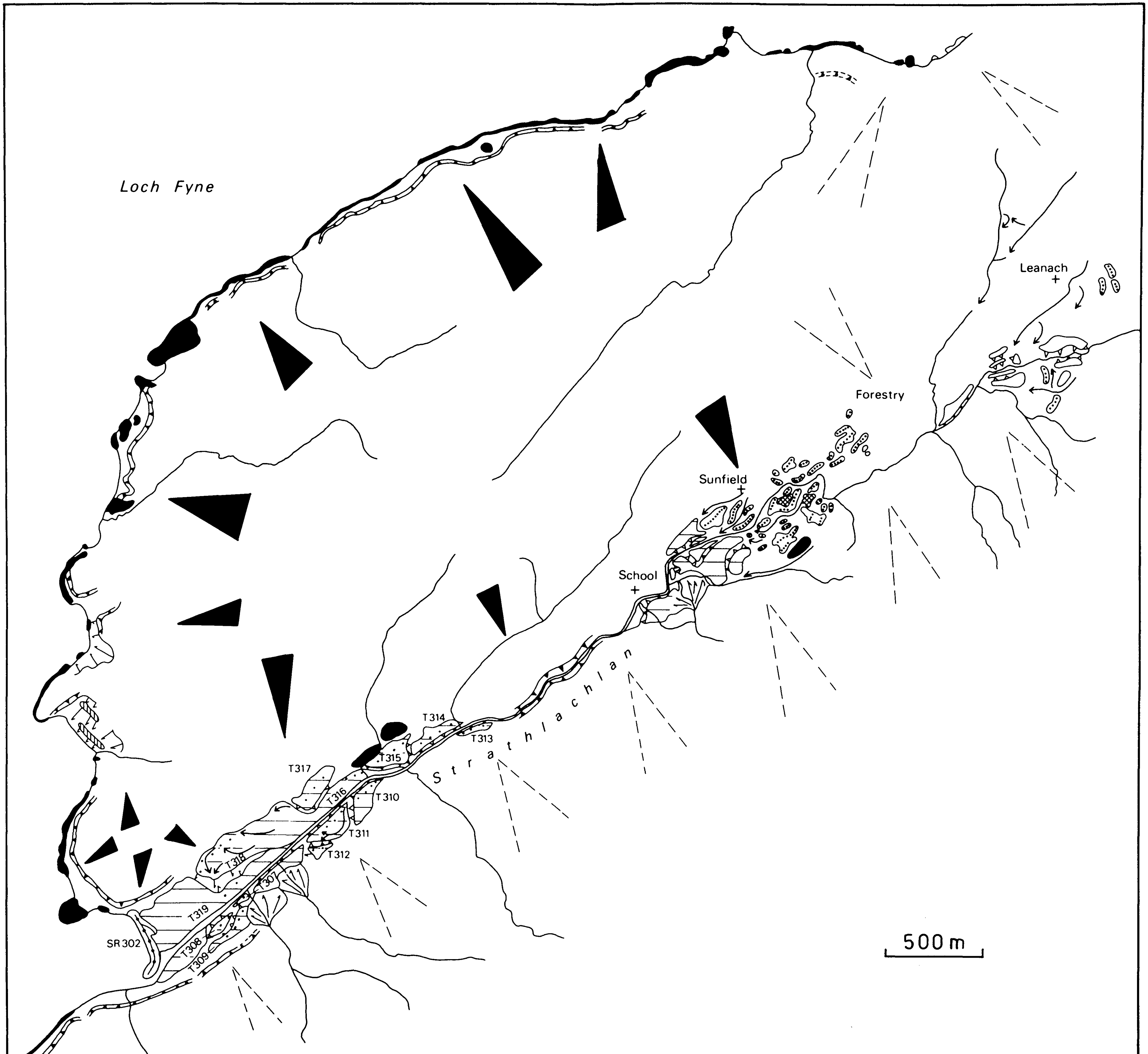
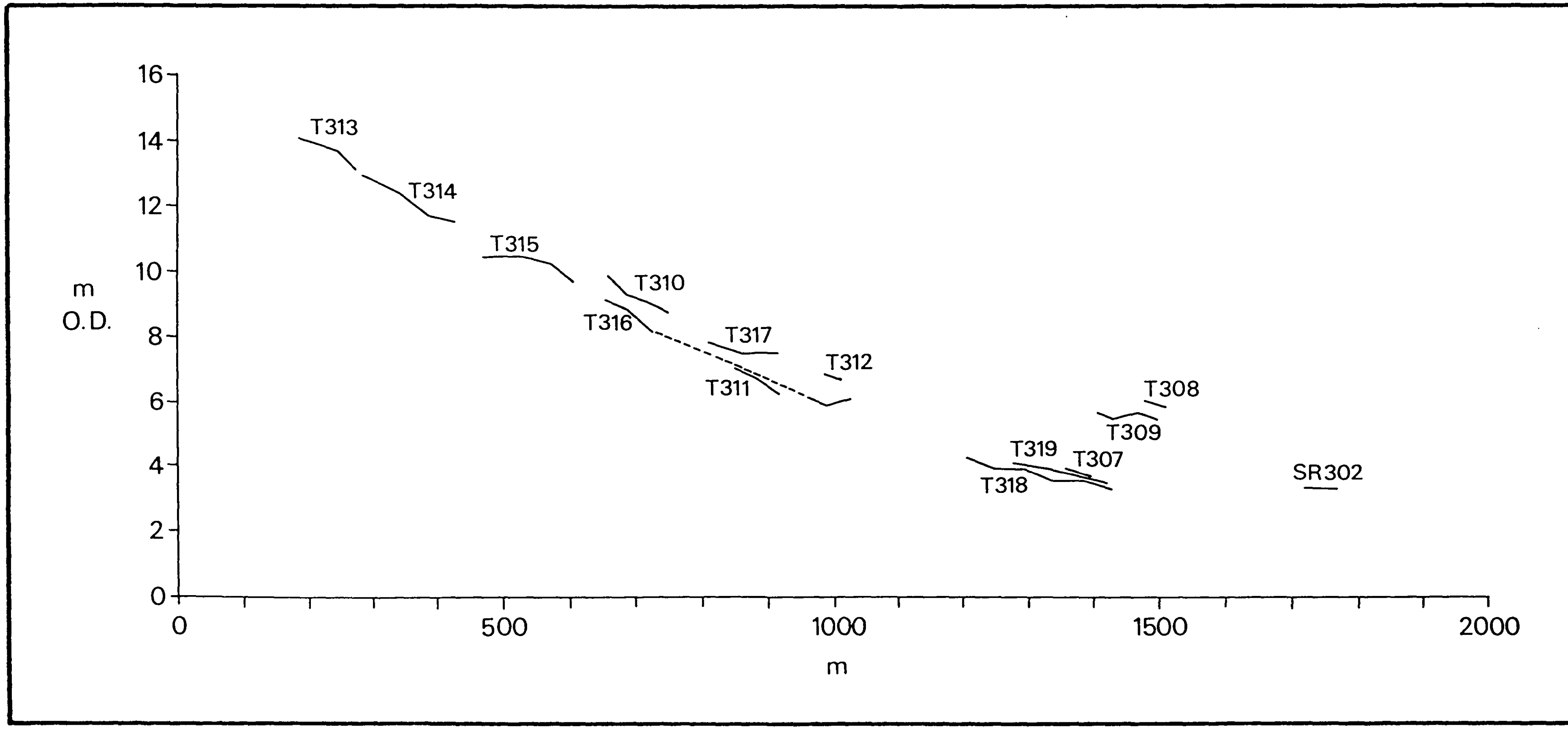
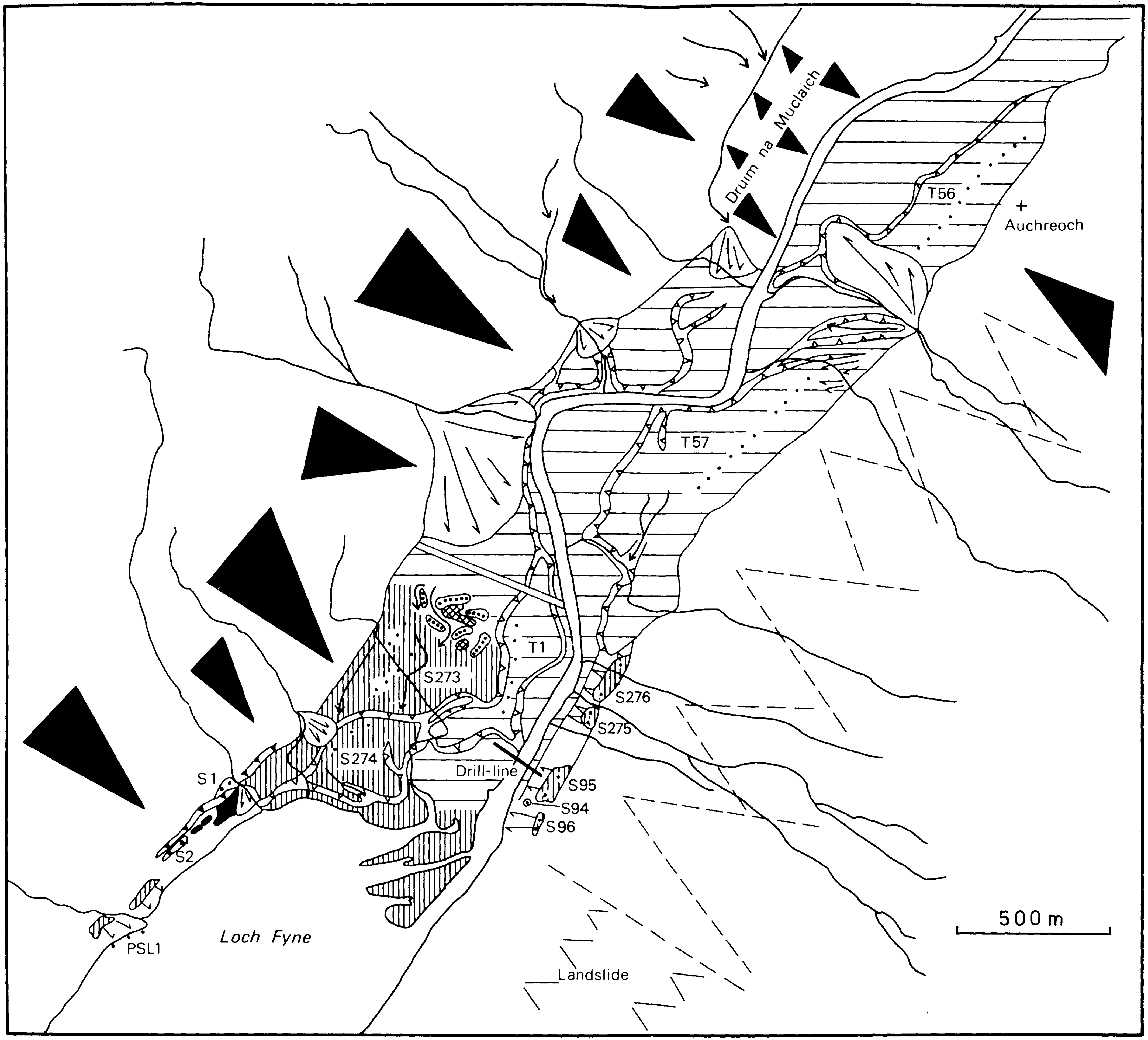

Loch Fyne
Leanach
Forestry
Sunfield
School
Strathlachlan
T314
T313
T315
T317
T316
T310
T311
T312
T318
T307
T319
T308
T309
SR302
500 m

Figure 47: Height-distance diagram of terraces in lower Strathlachlan.



m
O.D.
16
14
12
10
8
6
4
2
0
0
500
1000
1500
2000
m
T313
T314
T315
T310
T316
T317
T312
T311
T308
T309
T319
T307
T318
SR302

Figure 48: Geomorphological map of lower Glen Fyne.



Druim na Muclaich
T56
Auchreoch
T57
T1
S273
S276
S275
S274
S1
S2
Drill-line
S95
S94
S96
PSL1
Loch Fyne
Landslide
500 m

Figure 49: Height-distance diagram of the terraces in lower Glen Fyne.

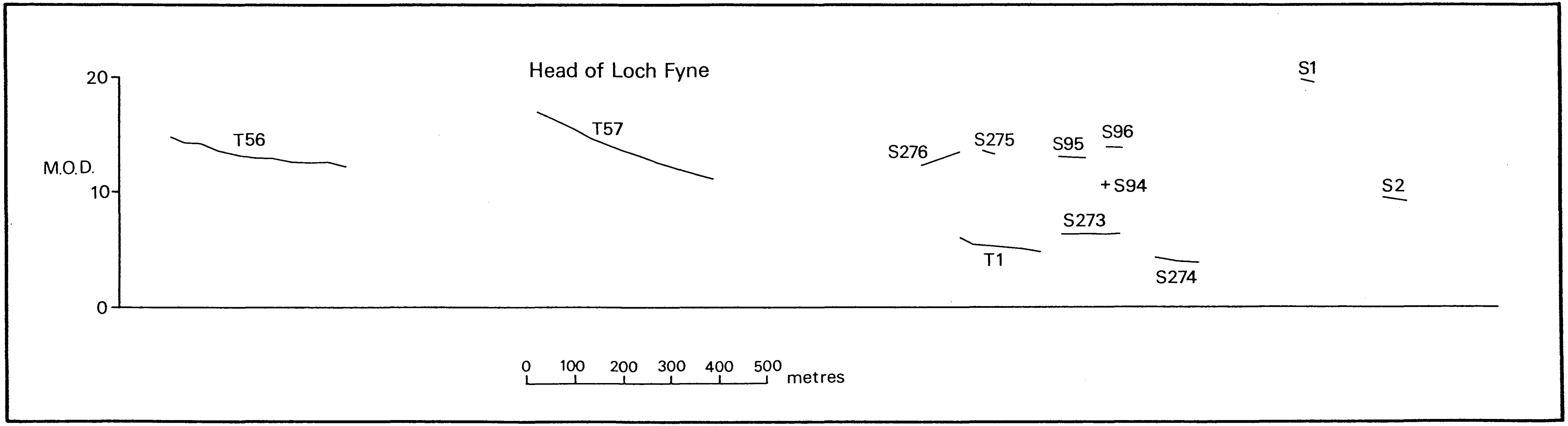

Head of Loch Fyne
20
10
0
M.O.D.
T56
T57
S276
S275
S95
S96
+S94
S273
T1
S274
S1
S2
0 100 200 300 400 500 metres

Figure 51: Geomorphological map of lower Glen Shira.

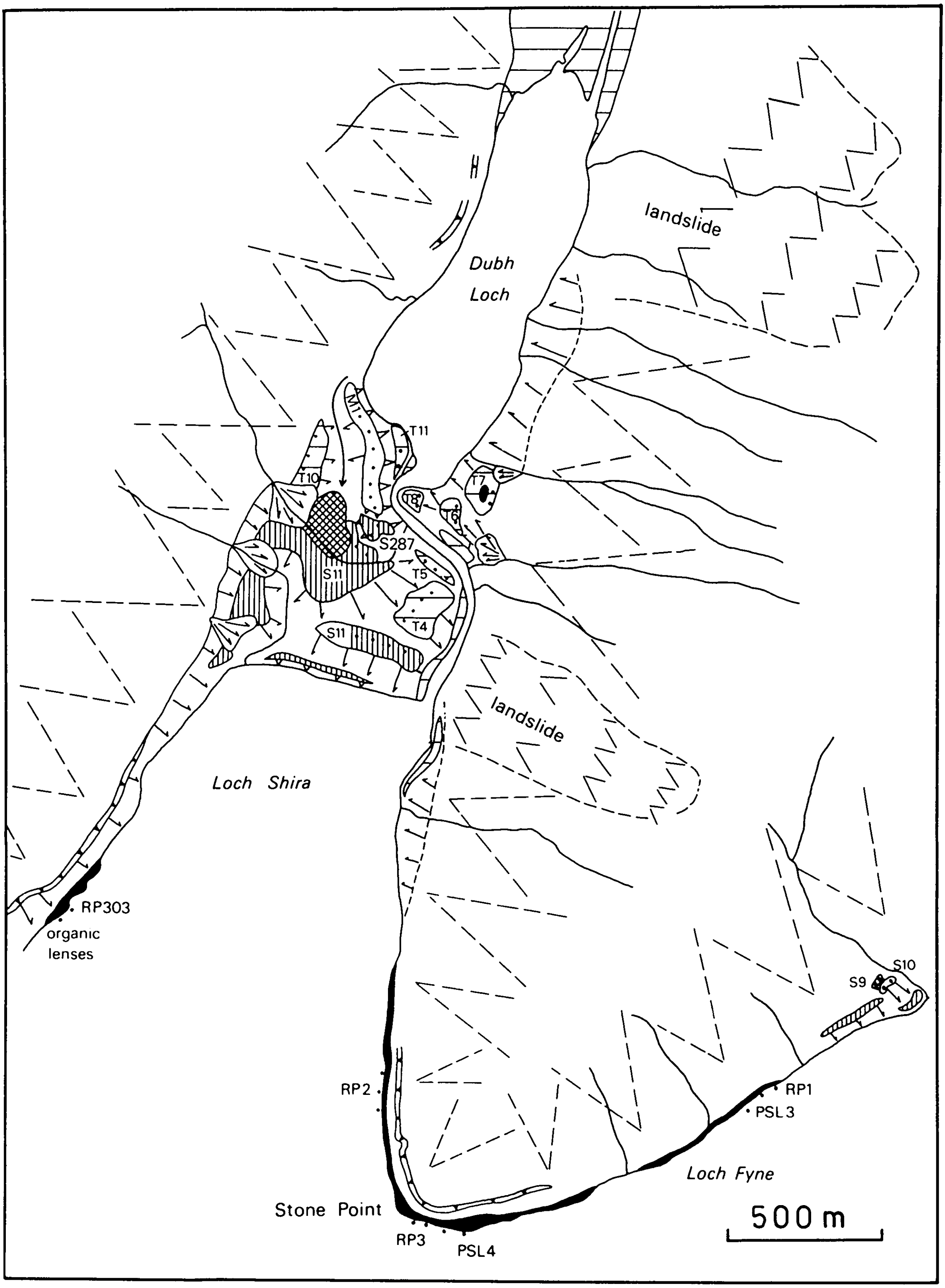

landslide
Dubh
Loch
M1
T11
T10
T7
T8
T6
S287
S11
T5
T4
S11
landslide
Loch Shira
RP303
organic
lenses
S10
S9
RP2
RP1
PSL3
Loch Fyne
Stone Point
500 m
RP3
PSL4

Figure 52: Height-distance diagram of the terraces in lower Glen Shira.

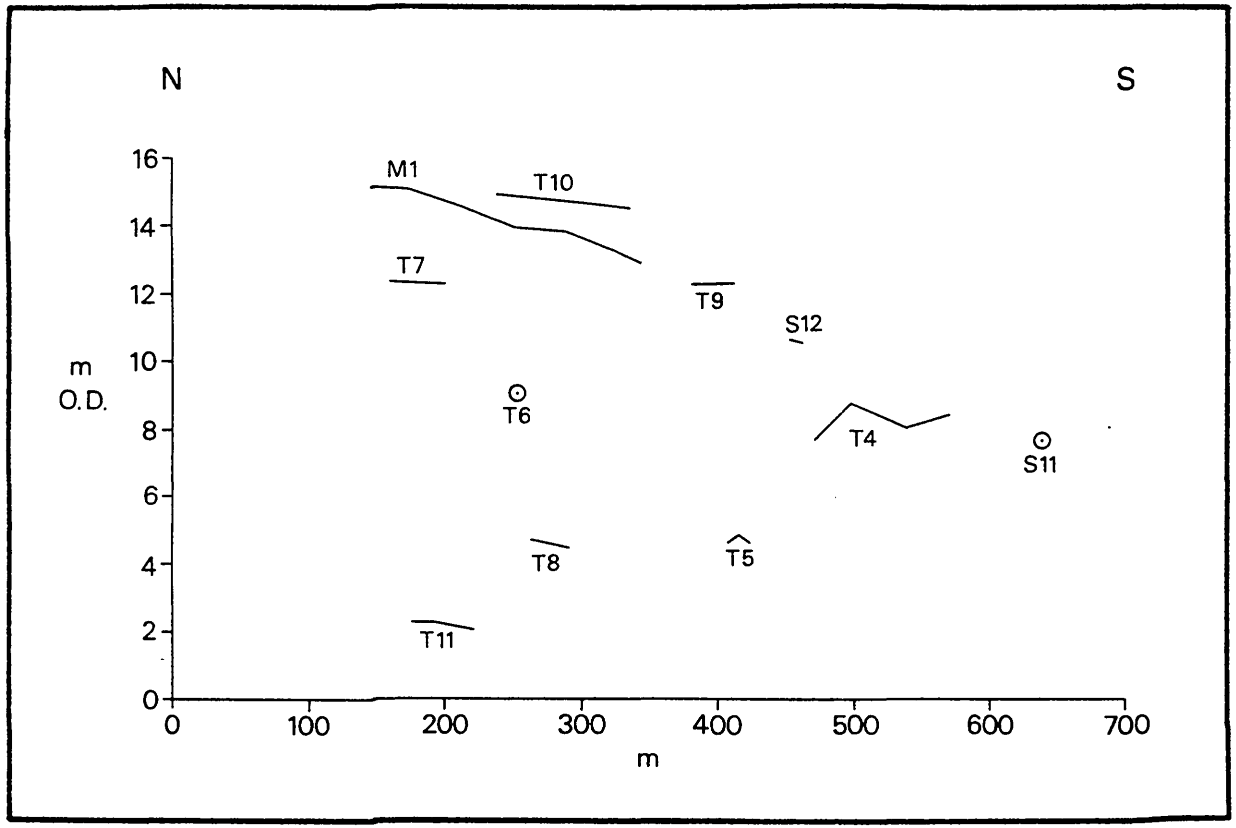

N
S
M1
T10
T7
T9
S12
T6
T4
S11
T8
T5
T11
m
O.D.
16
14
12
10
8
6
4
2
0
0
100
200
300
400
500
600
700
m

Figure 53: Geomorphological map of the Inveraray/Glen Aray area.

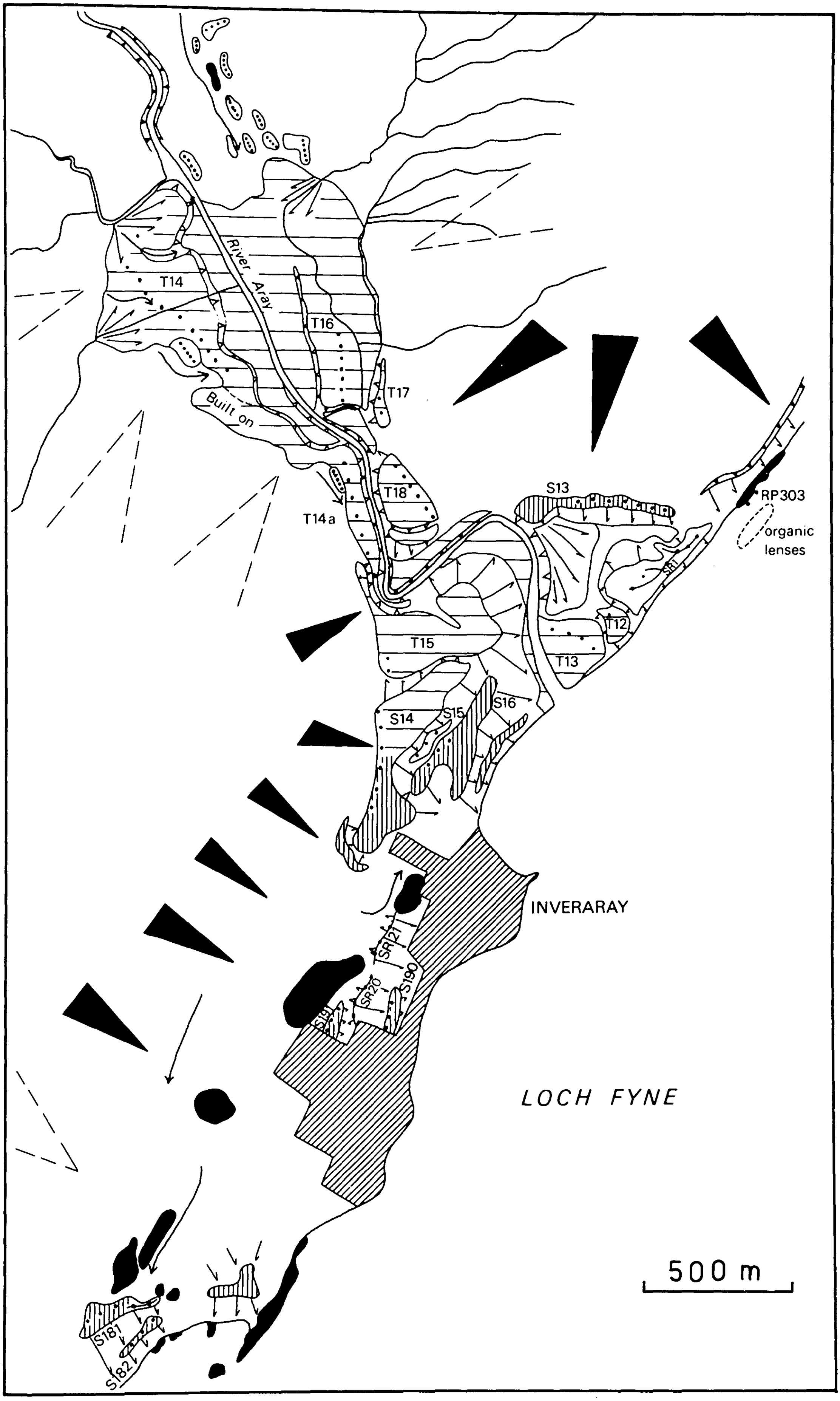
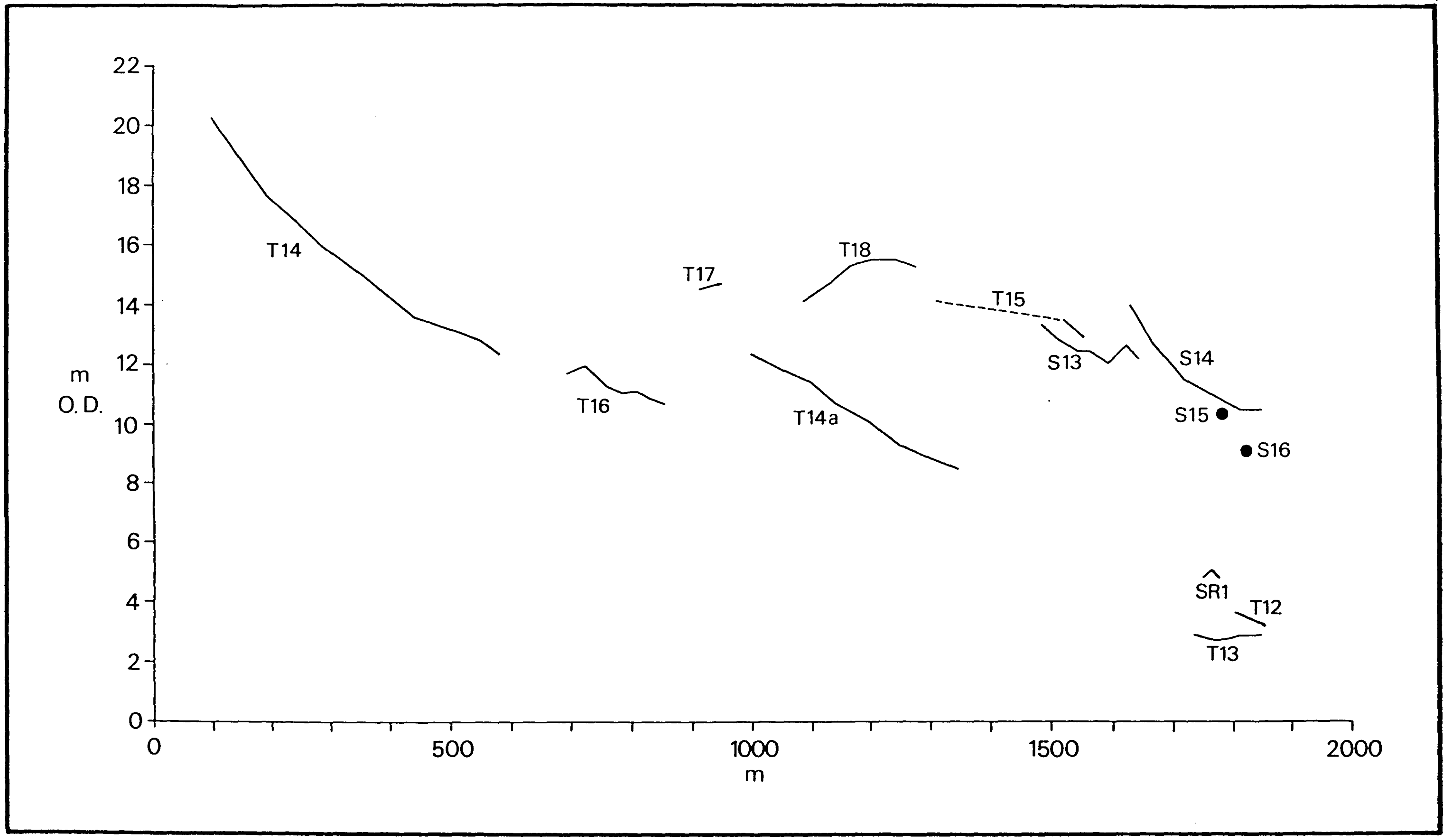
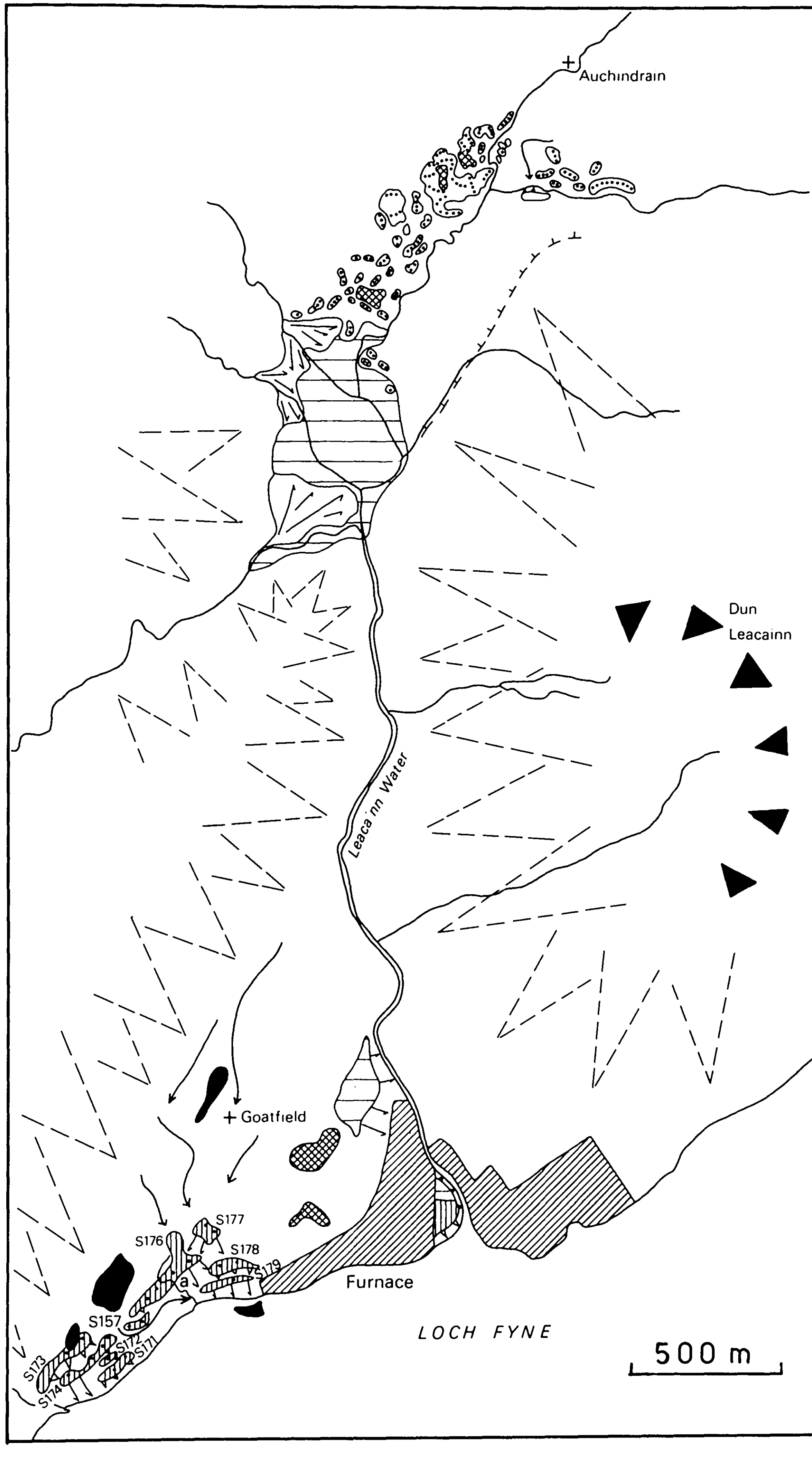

River Aray
T14
T16
T17
Built on
T18
T14a
S13
RP303
organic lenses
T12
T15
T13
S14
S15
S16
INVERARAY
SR21
SR20
S190
S191
LOCH FYNE
500 m
S181
S182

Figure 54: Height-distance diagram of the terraces of lower Glen Aray.



22
20
18
16
14
12
10
8
6
4
2
0
m
O.D.
0
500
1000
1500
2000
m
T14
T17
T18
T15
T16
T14a
S13
S14
S15
S16
SR1
T12
T13

Figure 55: Geomorphological map of the Furnace area.



Auchindrain
Dun
Leacainn
Leacainn Water
Goatfield
S177
S176
S178
S179
a
S157
S172
S171
S173
S174
Furnace
LOCH FYNE
500 m

CHAPTER 8

MARINE EROSION

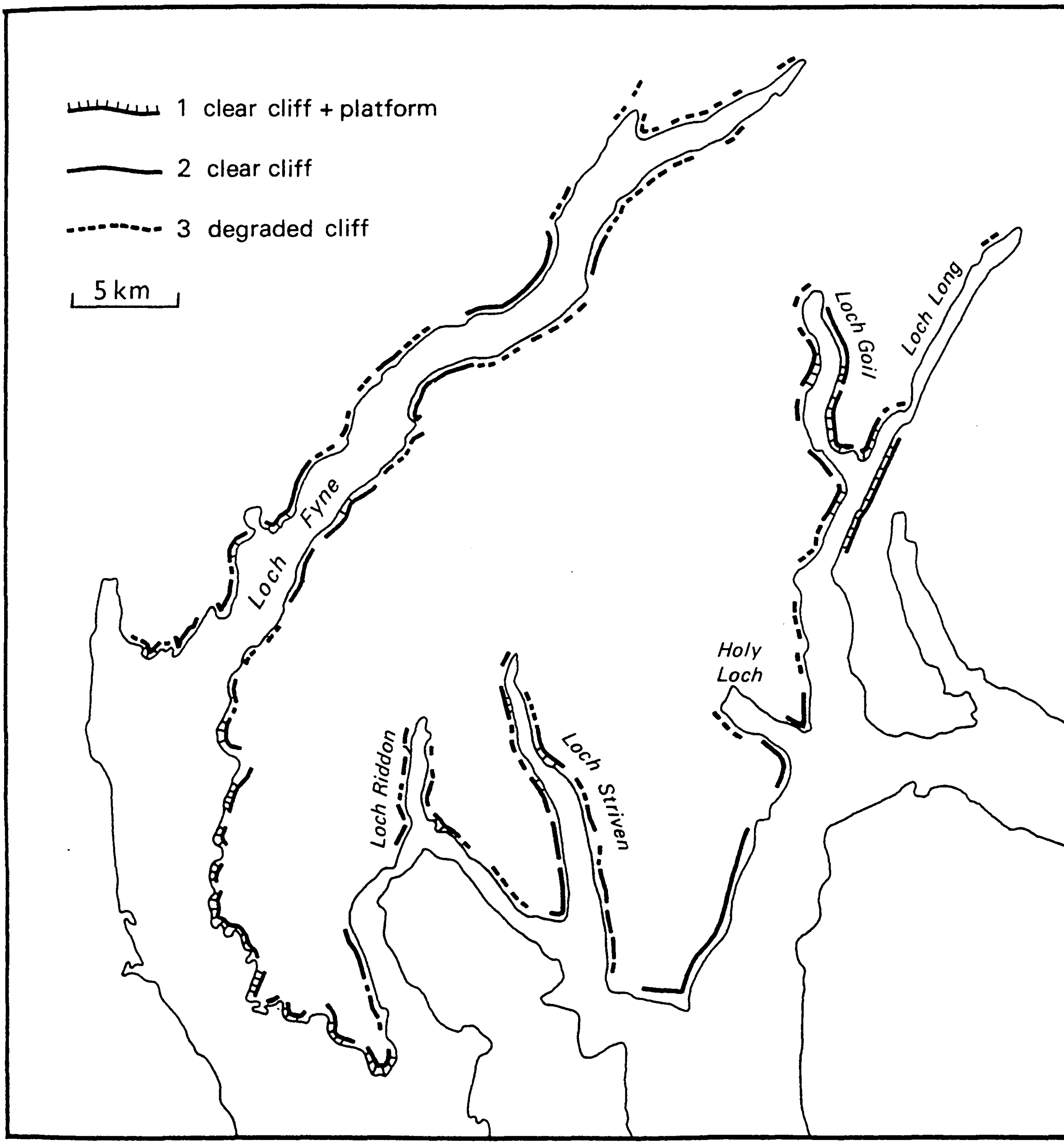
1. Introduction

The most prominent coastal landform in the study area is a fossil cliff that can be traced, more or less continuously and in varying degrees of clarity around the greater part of the coastline. It is often fronted by a clear rock platform, though this is a much less obvious feature than the cliff mainly due to a cover of beach gravels, peat or material that has fallen from the cliff. In addition to this fossil cliff there are a number of localities where low cliffs are fronted, in the present intertidal zone, by narrow but distinct rock platforms. Only one reference appears to have been made in the past to marine erosional landforms around the Cowal coast, Gunn et al. (1897, p 265) observing that the '20-foot beach'

> "forms an almost continuous fringe around the shore, though in places, especially in the more rocky areas, it is represented by a terrace of erosion merely."

During fieldwork all rock platforms, cliffs, caves, stacks, areas of undercutting and geos were recorded. Altitudes at the junction of the platform and cliff were obtained at approximately 50 m intervals where practicable care being taken to avoid superficial beach gravels. Peat, where present, was penetrated to bedrock by aluminium rods. In this manner 24 platform fragments were heighted at a total of 70 individual points, 15 of the fragments and 40 of the points being on the raised platform, the remainder relating to intertidal platforms. Figure 57 shows the distribution of heighted fragments as well as details of the platform and cliffline development.

Figure 57: Location of rock platform fragments and associated cliffline.


1 clear cliff + platform
2 clear cliff
3 degraded cliff
5 km
Loch Fyne
Loch Goil
Loch Long
Holy Loch
Loch Riddon
Loch Striven

2. Intertidal Platforms

These features were extensively discussed in Chapter 3. There it was demonstrated that the altitude of the platform/cliff junction was related to tidal cycle, fetch and nearshore gradient. The mean fragment altitudes ranged from 1.51 to 2.25 m O.D. with an overall mean of 1.91 m O.D., indicating that the platform altitudes relate to MHWOST. It was also indicated in that section that whilst a scatter of highly angular debris on the platform surfaces indicates current erosion the existence of vegetated sections and striations on the platform at one locality suggests the possibility of partial inheritance from an earlier period when sea-level was close to that of the present.

3. Raised Platforms

The fossil cliff and platform are not equally well formed along all parts of the coastline. They are best developed together around the southern part of Loch Fyne, from N of Kilfinan to Ardlamont Point. Near Loch Gair on the W side of Loch Fyne, beside parts of Loch Goil and along Loch Long opposite the mouth of Loch Goil the platform and cliff are sharply represented though of much smaller dimensions than by southern Loch Fyne. Elsewhere the cliff in particular can be followed without much difficulty and often the presence of the platform can be inferred, at localities where there are extensive gravel deposits between the shore and the cliff, by a bedrock rise at the foreshore to almost the altitude of the cliff base. This is typically the case near Toward Point, in much of Loch Striven and in middle Loch Fyne. Towards the head of Loch Fyne the platform becomes increasingly indistinct and the cliff, though present, degenerates steadily. It can be identified on the W side, however, almost to the head of the loch and around Strone Point (north of Inveraray) four collapsed caves occur. Little trace of either cliff or platform can be found within Loch Long beyond the junction with Loch Goil. This is presumably due to the extensive landslipping that characterizes both sides of this stretch of the loch. One rather poor fragment exists at the head of the loch on the W side opposite Arrochar. From this distribution it is concluded that the rock platform and cliffline were formed throughout the area studied.

The strength of development of this rock-cut feature and its observed continuity around the coast strongly suggest that it is a single shoreline. The work of Gray (1978) has established this shoreline to be part of the Main Rock Platform found around the coast of the SW Highlands. Hence the altitudes obtained can be compared in order to establish the degree and direction of tilting that may be present. The comparison was effected by means of a height-distance diagram in the manner described in Chapter 4. A point of origin (NGR 24000, 72000) was selected to the N of the study area (so that all distances would be positive) and projection planes were directed from this point through the study area starting in a N-S direction and ending, by moving 5^{o} clockwise at a time, in an E-W direction.

Height-distance diagrams were constructed and regression equations calculated for each orientation of the projection planes. Table 21 summarizes the statistics for the various orientations.

TABLE 21. REGRESSION ANALYSIS OF MAIN ROCK PLATFORM.

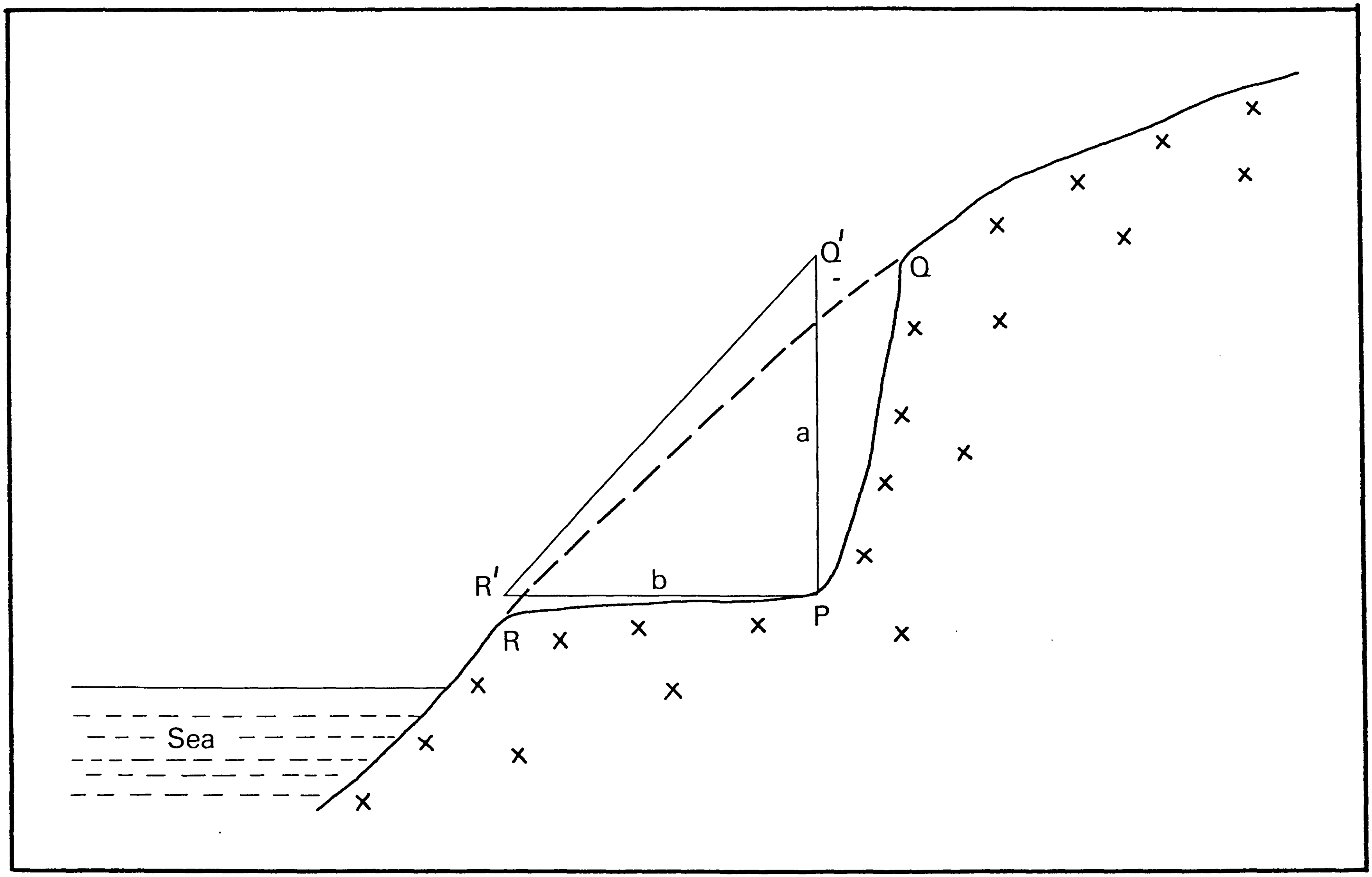
Orientation	Gradient m/km	Correlation Coefficient
S20W	0.126	-0.979
S25W	0.125	-0.979
S30W	0.124	-0.976
S35W	0.124	-0.971
S40W	0.124	-0.964
S45W	0.124	-0.955

As was indicated in Chapter 4 there are various reasons for the minimum gradient not coinciding with the maximum correlation coefficient. There is little variation in the gradients but at the fourth decimal place S35W and S40W are the orientations of least gradient. Of these two S35W was that with the greater correlation coefficient and therefore S35W was chosen as the optimum projection plane for this shoreline. It is plotted in Figure 58.

The gradient of the Main Rock Platform in the study area is thus 0.12 m/km, the shoreline falling in altitude from around 10.5 m O.D. at mid-Loch Fyne and Loch Goil to almost 5.5 m O.D. at the S end of Loch Fyne. This gradient compares with the gradient of the Main Rock Platform found by Gray (1974a) in Lorne of 0.16 m/km and by Dawson (1980) on Jura of 0.13 m/km and is similar to the gradient established by Gray (1978) for the SW Highlands (0.12 m/km). It is also comparable with the gradient of the Main Lateglacial Shoreline (0.17 m/km) of SE Scotland (Sissons, 1976b) that has been proposed as a correlative of the Main Rock Platform (Sissons, 1974a). The age of the Main Rock Platform is discussed later.

In describing particular areas (e.g. Portavadie/Ardlamont), the relationship between rock structure and the development of marine erosional landforms has been mentioned. It is not necessary to enumerate further examples but it should be emphasised that rock structure seems responsible mainly for the details of the rock sculpture and not, as was demonstrated at Ardlamont Point (Chapter 7.1) for the total volume of rock eroded. In order to test this relationship further, estimates were made, where available data permitted, of the volume of rock that may have been eroded during the formation of the shoreline. Measurements were made as shown in Figure 59. In Figure 59, which depicts a cross-section

Figure 59: Diagrammatic cross-profile of rock platform and cliff.
PQR represents the zone actually eroded and PQ^1R^1 the zone calculated.


Q′
Q
a
R′
b
P
R
Sea

of the platform and cliffline at right angles to the coastline, the area of marine erosion is given as PQR. The altitude of P is either the field measurement or is taken from the height-distance diagram, the altitude of Q is obtained from 1:63,360 O.S. maps where a contour line is located close to the cliff top and the distance RP is measured from 1:10,560 field maps. Of these measurements the altitude of Q is the most subject to error and it was the infrequency in the number of areas where it could be measured that limited the number of data points. For ease of calculation the area PQR is approximated by the area of the right-angled triangle PQ'R' (i.e. $\frac{1}{2}a.b$). Estimates of volume were derived by considering uniform lengths of coast approximately 1 km long in which cross-sections were measured every 100 m, the results being arithmetically meaned and expressed in figures of volume of material eroded per metre length of coast (m^3/m).

The interpretation of the figures derived depends upon the configuration of the line QR prior to the formation of the platform. If QR followed the general configuration of the landsurface above the cliff then the approximation Q'R' is probably not much in error and tends to be balanced by the other approximations PR' for PR and PQ' for PQ. If, however, the length of coastline had been the focus of marine erosion at times prior to the final establishment of PQR then the estimate of the amount of material eroded by the sea forming PQR would be in error, perhaps quite substantially. This point will be returned to later in discussing the age of the feature. In the present section it is assumed that the figures calculated are in proportion to the volume actually removed by the sea and that they can be compared with each other to gain some idea of the relative efficiency of marine erosion in their respective localities.

It proved possible to make estimates at the seven localities listed in Table 22.

TABLE 22. MAIN ROCK PLATFORM EROSION VOLUMES.

Locality	Volume (m^3/m)	Fetch Normal to coast (km)	Maximum Fetch (km)	Comp. of Max. Fetch
W. Ardlamont	675	11	50	46
E. Ardlamont	468	6.5	116	38
E. Toward	978	9	130	49
S. Toward	1568	130	130	130
Loch Gair	92	5	34(10)*	30(9)*
S. Port Leathan	136	8	20	10
S. Portavadie	225	8.5	25	19

* relates to measurements across Otter Ferry spit. See text.

There is a considerable range in volume of material eroded, from 1568 m^3/m for South Toward to 92 m^3/m for the mouth of Loch Gair. It might initially seem that rock type is an important factor as the two areas around Toward Point having the largest volume eroded are cut in Old Red Sandstone while the remainder are cut in Dalradian Schists. While undoubtedly rock type is relevant to the development of the feature, inspection of the figures in lower Loch Fyne, however, shows a steady diminution up-loch in volume of material eroded in relatively uniform rock types shich suggests the influence of fetch.

Two elements of fetch have been measured, firstly the fetch at right angles to the stretch of coast being considered, and secondly the component of maximum fetch that is normal to the relevant stretch of coast (cf. Chapter 3). The fetch normal to the coast produces anomalies in that whilst 8 km of fetch at S. Port Leathan relates to erosion of

136 m^3/m and 8.5 km of fetch at S. Portavadie to 225 m^3/m, only 6.5 km at E. Ardlamont is associated with 468 m^3/m and 9 km at E. Toward with 978 m^3/m. The component of maximum fetch that is normal to the coast shows a much closer relationship with the volumes eroded, only the 34 km : 92 m^3/m at Loch Gair being markedly anomalous. This last anomaly may partly be explained by the fact that the line of maximum fetch to the mouth of Loch Gair is across the Otter Ferry spit. If the maximum fetch is measured without crossing the spit the anomaly is resolved. (Due regard being given to the available data, this may suggest that the spit was in existence prior to the formation of the platform). Figure 60 shows the plot of the component of maximum fetch at right angles to the coast against volume of material eroded for each metre of coastline. On the limited data available a straight-line relationship is apparent and the regression equation calculated has been inserted.

The existence of this relationship between volume eroded and fetch must be due, in part, to the small number of available data points. The results suggest that the platform and cliff development is no better in more sheltered areas than in the more exposed ones, a conclusion that contrasts with Sissons' (1974a, p 45) judgement that the rock platform

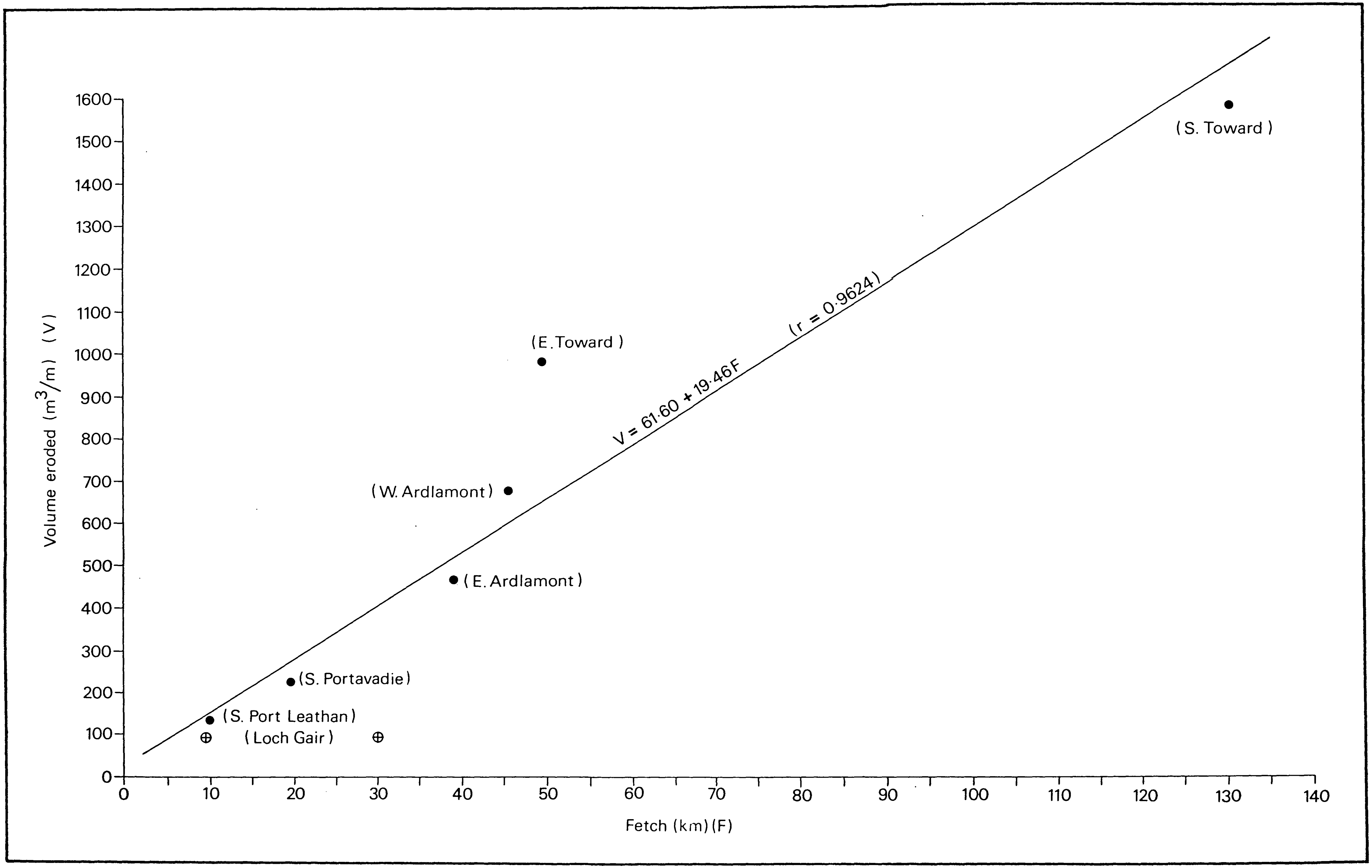
> '.... is often surprisingly well developed in locations very sheltered from strong wave attack.'

4. <u>Age of Fossil Platform</u>

No evidence has been found that would allow accurate establishment of the age of this erosional feature. The following observations are, however, relevant.

1. The height data obtained in this study indicate that the platform

Figure 60: Graph of fetch against volume of rock eroded in formation of Main Rock Platform.



1600
1500
1400
1300
1200
1100
1000
900
800
700
600
500
400
300
200
100
0
Volume eroded (m3/m) (V)
0
10
20
30
40
50
60
70
80
90
100
110
120
130
140
Fetch (km) (F)
V = 61·60 + 19·46F
(r = 0·9624)
(S. Toward)
(E. Toward)
(W. Ardlamont)
(E. Ardlamont)
(S. Portavadie)
(S. Port Leathan)
(Loch Gair)

is tilted at 0.12 m/km towards S35°W and when considered along with the similar data of Gray (1978) for a wide area of the SW Highlands the tilting clearly centres on the Rannoch area, thus strongly suggesting a glacio-isostatic origin for the tilt.

2. At Ardyne the cliffline (with rock exposed in it) truncates Lateglacial river terraces and hence must post-date them.

3. At Ardlamont Point and S of Strachur on Loch Fyne (fragment RP307) weathered bedrock is exposed in the cliff and platform. In both instances the rock is dolerite in the form of dykes.

4. Between Strone Point and Colintraive is a section of cliffline in which no bedrock can be seen. Despite much slumping down the cliff-face during Flandrian times (as evidenced by peat intercalated with slumped till and slopewash) a section of ca. 8 m of undisturbed red till was seen in the cliff where it has been incised by a small burn between Inverneill and Glaic. Bedrock is not seen in the section which extends 20 - 30 m into the cliff.

5. Within the sea lochs, particularly Loch Fyne, there is a strong contrast between the degree of preservation of the rock platform at the head and at the mouth of the lochs. In Loch Fyne no distinct line can be drawn to separate the upper loch where the cliff is typically rather degraded and indistinct and where a cliff/platform junction is very difficult to define and the lower loch where the features are sharp and exceptionally clear, stacks and undercutting being observable. Rather between these areas there is a diminution in the frequency of clearly defined platform fragments, the most distinct perhaps favouring a down-loch aspect. A similar pattern is discernible in Loch Long. The

generalisation appears not to hold for Loch Goil, however, stretches of which have better platform development than, for example, much of the area around the Kyles of Bute and in Loch Riddon.

6. Very large volumes of material have been removed during the fashioning of the cliff and platform. The cliff on occasion reaches over 30 m in height and the platform has been inferred near Toward Point to be several hundred metres broad. Estimates of volume of rock removed range up to 1568 m^3/m of coastline.

7. Striations along the front edge of the platform are found in many localities ranging from the inner lochs to their mouths. Till in situ is nowhere seen to overlie directly the rock platform though the case near Colintraive in the Kyles of Bute where the cliffline is cut in till suggests that there should be a till cut platform to accompany it. In upper Loch Fyne north of Dunderave Castle, where the cliffline is poorly developed, two stream sections reveal grey till lying between the base of the raised cliffline and the sharp, often rocky fall to the present shore. In only one of these sections can till be seen to overlie bedrock, which may be part of the platform or may be in a hollow on it. The till in part of the section overlies a deposit of coarse rounded pebbles and cobbles with a silty sand matrix. It is not clear whether these are beach deposits relating to the time of erosion of the platform. On the E side of Loch Fyne near Creag a' Phuil a stream section again shows a grey till between the cliffline (here much degraded) and the rock outcrop along the present shore. No bedrock was seen under the till, however. Ice moulding is difficult to identify unless accompanied by striations or unless clear roche moutonnée or small crag and tail features are present. Nowhere is this the case on platform or cliffline in Cowal.

It could be that stretches in upper Loch Fyne where the cliffline has a very worn-down appearance have been reduced to that state by glaciation but there is no clear evidence of this. In some areas, however, the sharpness of the features is such as seemingly to preclude extensive glaciation.

Taken together the above points appear somewhat contradictory. On the one hand there is evidence in the truncated terraces at Ardyne and the till cliffline near Colintraive which appears to suggest a post-glacial origin for part at least of the feature, while on the other hand the evidence from upper Loch Fyne seems to point to the platform having been glaciated. The degree of isostatic tilt would suggest that, if the shoreline is post-glacial, it is Lateglacial in age but the very large amounts of rock eroded in some areas suggest a longer period of relative sea-level stability than would seem available in the Lateglacial given the considerable relative sea-level fluctuations that are known to have occurred during that time. The degree of preservation of the platform varies from the obscure and barely discernible features of parts of inner Loch Fyne and Loch Long to the exceptionally fresh and detailed landforms around the southern stretch of Loch Fyne. The weathering of the dyke rocks must post-date the formation of the cliffline for it seems highly improbable that a feature should be eroded equally in weathered rock, which can be broken in the hand, and neighbouring fresh rock, that is difficult to break by hammer. Certain dolerites do appear, however, to weather rather rapidly under present climatic conditions in Scotland as is perhaps most graphically illustrated by the weathering which has taken place in the cuttings of the northern approaches to the Forth Road Bridge. The weathering may therefore be of no great antiquity.

The age of the feature that appears to leave least unexplained is that proposed by Sissons (1974a), who suggested that the low fossil rock platform of the W coast of Scotland was the correlative of the widespread Buried Gravel Layer (Sissons, 1969) of the Forth Valley. This Buried Gravel Layer (and its correlatives around the SE coast of Scotland, now called the Main Lateglacial Shoreline; Sissons, 1976b) is, on stratigraphic grounds, clearly Lateglacial in age.

A Lateglacial age for the erosional features around Cowal would explain the tilt and the evidence of terrace truncation at Ardyne. It would lead to an explanation of the upper Loch Fyne glaciation evidence in that this would have resulted from Loch Lomond Readvance glaciation (Chapter 11). The differential development between inner and outer lochs could be explained in a similar way, though this is not entirely consistent with the observation that the platform is well developed in Loch Long opposite the mouth of Loch Goil, an area glaciated during the Loch Lomond Readvance (Chapter 11). The most outstanding problem remaining is the large volume of rock eroded in the formation of the feature. This has always been a problem since the measurements by James Smith in 1838 on the relative amounts of erosion of the igneous dykes around the Clyde sea area indicated that the fossil marine features were worn back 16 times farther than on the present shore. McCallien (1937) was so impressed by this single line of evidence that he considered an interglacial age to be the only possible explanation, a conclusion echoed, with only minor modifications, for the next 37 years (McCann, 1966, 1968; Synge, 1966; Sissons, 1967a, p 193; Gray, 1974a) and still advocated by some (Peacock, 1975; Synge, 1977a).

It should be noted, however, by those who consider an interglacial age as providing scope for large amounts of marine erosion, that little support comes from the recent studies of eustatic sea-level fluctuations during the last 300,000 years in Mallorca (Butzer, 1975), the last 250,000 years in Alaska (Hopkins, 1973), the last 140,000 years in New Guinea (Bloom et al., 1974), or the last 120,000 years in Barbados (Steinen et al., 1973). These curves show a large amount of agreement and all indicate that no high sea-level stand has been any longer than that of the present interglacial during the periods considered. Since there is no reason to believe that Scotland in previous interglacials experienced crustal recovery much different from that resulting from decay of the last ice-sheet, and since erosion at present sea-level during this present interglacial is very small, there is still a problem over the availability of the time necessary for erosion of the rock platform. For example if James Smith's (1838) figure is correct and if sea-level has been close to its present level for the last 1,000 - 2,000 years, then a stable sea-level for a period of 16,000 - 32,000 years is indicated. Adding ca. 12,000 years as the time necessary for the isostatic recovery of the upper Firth of Clyde suggests an interglacial 28,000 - 44,000 years in duration. There seems little evidence for such a period in, for example, the Quaternary deep-sea record of the NE Atlantic (Ruddiman and McIntyre, 1976).

Sissons (1974a) suggested that the answer to the problem was an increased rate of erosion in the extreme climate of the latter part of the Lateglacial when large numbers of freeze-thaw cycles would result from a combination of diurnal temperature variations and tidal rise and fall. Some support for the efficiency of such a process of marine erosion comes from the observations of Nansen (1922) in southern Norway and Sollid _et al_. (1973) in northern Norway, but in both these instances rates of erosion seem still to be too slow to allow for the erosion of the rock platform on the W coast of Scotland. It is very probable, however, that there is no modern analogue for the environmental conditions experienced in Scotland at the end of the Lateglacial.

A Lateglacial age for the rock platform means that the nearshore marine sedimentary record for these times should contain evidence of the considerable quantities of material eroded. Recent construction works have allowed the inspection of large sections in the Lateglacial marine sediments at Portavadie (Chapter 7.1), at Lochgilphead (Peacock _et al_., 1977) and at Ardyne (Peacock _et al_., 1978). Once again, however, the evidence is inconclusive.

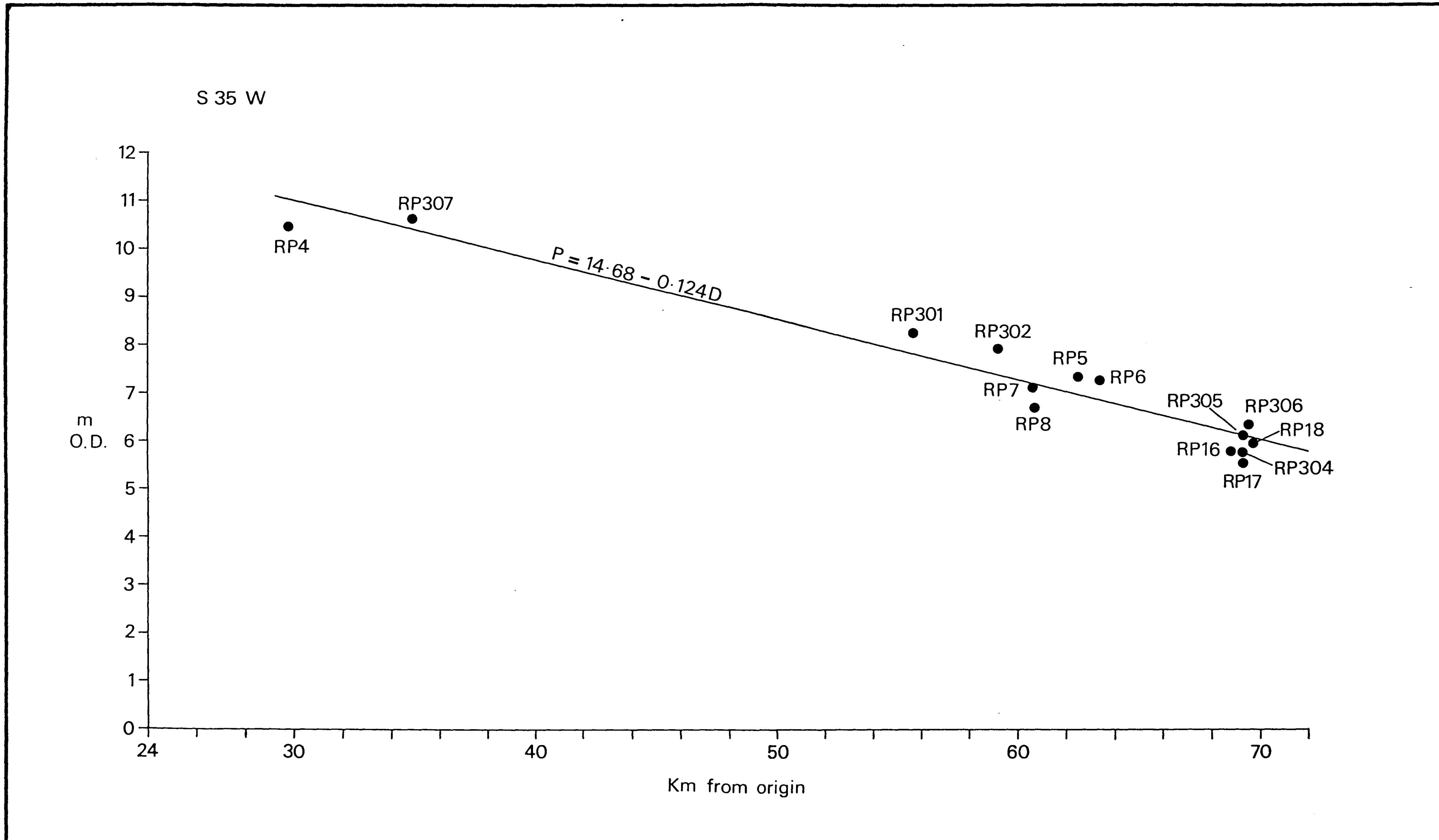
At Portavadie, the Lateglacial marine clays that lap on to the front edge of the rock platform are eroded at their surface at altitudes below 5 m O.D.. In this area the back edge of the platform is _ca_. 6 m O.D.. Radiocarbon dating indicates that this erosion took place some time after 10,700 yr BP (for further details see Chapter 7.1). At Lochgilphead, Peacock _et al_. (1977) record a sequence of Lateglacial clays on the present foreshore with a marked unconformity after _ca_. 11,200 yr BP. The stratum below this unconformity, deposited after _ca_. 12,000 yr BP contains occasional boulders, along with angular phyllite and quartz-schist fragments set in a sandy or silty-sand matrix with occasional gravelly lenses. A similar picture emerges at Ardyne Point, where Peacock _et al_. (1978) record a marked unconformity at _ca_. 11,300 yr BP which lasted for 300 - 500 years. Above the unconformity is a thin sandy-silt layer with scattered pebbles which contains a High Boreal fauna and suggestions of intertidal conditions. This is dated ca. 11,000 - ca. 10,800 yr BP and occurs at -1 m O.D. from the height-distance diagram the height of the rock platform is ca. 7 m at Ardyne. At ca. 10,800 yr BP there is a rapid increase in Arctic fauna with a considerable High Arctic component. The sediment (ca. -2 to -4 m O.D.) is a poorly sorted clayey silt and silty sand with occasional interbedded sand lenses and has rounded stones, often up to 80mm, scattered throughout. Water depth is provisionally placed at less than 20 m and greater than 7 m with, at the top, a suggestion of littoral conditions. These arctic conditions ceased ca. 10,250 yr BP.

In summary, the marine sedimentary record of these sites for the latter part of the Lateglacial does not appear to reveal the equivalent of the Buried Gravel Layer of the Forth Valley, a layer of boulders and cobbles typically 1 m thick (Sissons, 1969). The sediments are dominantly sands and silts with included boulders and cobbles with a frequency and attitude that suggests dropping from shore ice or icebergs, a process first suggested by Jamieson in 1865. In particular those sediments relating to the Loch Lomond Stadial at Ardyne Point, located very close to a stretch of the cliffline, show a decrease in the number of boulders and cobbles. Perhaps this is due to a local factor relating to nearshore sediment movements but in the sections at Greenock reported on by Bishop and Dickson (1970) and Bishop and Coope (1977) a silty sediment also characterizes Loch Lomond Stadial times.

There is, however, an unconformity in the sequence dated to around 11,300 - 11,000 yr BP which seems replicated at more than one site. This is followed at Ardyne by the most temperate fauna recorded in the Lateglacial marine sediments. The relative roles of sea-level and temperature changes seem uncertain, however, and it may be significant that the fauna that records the highest temperatures also indicates the shallowest sea depth. The unconformity may be the same as that in the Glasgow area reported by Rose (in Gray, 1978) in which Lateglacial marine sediments and sandstone bedrock are conformably eroded to a width of 300 m, but no height data are given for this feature and its place in the sequence of sea-level changes is uncertain. If this unconformity is the correlative of the Buried Gravel Layer it cannot be said, on the basis of the Ardyne evidence where it is most closely dated, to correspond to the Loch Lomond Stadial, the deposits of which clearly overlie it. A correlation between this unconformity and the rock platform would be corroborated by the existence of the rock platform in those areas (e.g. Loch Long) covered by Loch Lomond Readvance glaciers.

The evidence accumulated during this study thus points to a qualified acceptance of the hypothesis that the rock platform is Lateglacial in age and correlates with the Main Lateglacial Shoreline of SE Scotland. The chief problems are the very large volumes of rock that have been eroded in certain areas and the apparent absence of this eroded material in the sedimentary record. Both these difficulties would be diminished if it were conceded that the area eroded had been the focus of marine erosional activity at a prior time or times during the Quaternary. The isostatic tilting of the shoreline need not be an overwhelming difficulty given the lengthy periods during the Quaternary in which ice sheet volumes were intermediate between interglacial and full glacial conditions (Shackleton, 1974) and when Scotland may have had an ice cover not too dissimilar to that of the Loch Lomond Stadial. Indeed, given the fluctuating nature of the Late Quaternary sea-level curves referenced earlier, the ice-sheet fluctuations recorded by oxygen isotopic variations in deep-sea sediments and the relatively insignificant amount of marine erosion accomplished at present sea-level around the coasts of the Highlands, it may be speculated that no major rock platform in Scotland formed during the Quaternary could be other than glacio-isostatically tilted.

Figure 58: Height-distance diagram of Main Rock Platform, S35W.


S 35 W
m
O.D.
P = 14·68 − 0·124D
RP4
RP307
RP301
RP302
RP5
RP6
RP7
RP8
RP305
RP306
RP18
RP16
RP304
RP17
Km from origin
0
1
2
3
4
5
6
7
8
9
10
11
12
24
30
40
50
60
70

CHAPTER 9

DEPOSITIONAL SHORELINES FORMED IN RELATION TO GLACIER ICE

1. Introduction

During ice-sheet deglaciation marine landforms were constructed as the coasts of the Cowal Peninsula and neighbouring areas were progressively freed of ice. As has been demonstrated in Chapters 5, 6 and 7 the retreat of the ice was not uniform, but interrupted at various localities where the ice front moved little for some time and, due to continuing isostatic recovery of the land, relative sea-level fell. Evidence has also been presented that indicates that ice retreat was fastest where the ice margin was calving, and areas of dead-ice became isolated on land and continued to supply sediment to the coastal zone, as near Millhouse (Chapter 7.1). In the present chapter the various marine landforms relating to these events are examined together in an attempt to correlate the features and establish common shorelines throughout their area of occurrence.

2. Identification of Shorelines

As outlined in Chapter 4 the basic tool utilised in the identification of the individual shorelines is the height-distance diagram. The particular approach adopted with the Lateglacial marine landforms was as follows. The arithmetic mean altitude of all the marine landforms that were classified as 'good' or 'moderate' were projected at right angles into a series of vertical planes and the horizontal distance in the plane for each fragment was calculated from the same common origin as was used for the analysis of the Main Rock Platform data (Chapter 8). The vertical plane was rotated by 5^{o} intervals from due S to due W as previous knowledge of the pattern of isostatic uplift in Scotland (Sissons, 1967a) and the deformation of the Main Rock Platform suggested that the Lateglacial shorelines would tilt in this quadrant. The heights and distances were plotted for each of the separate orientations of the projection plane.

For the majority of the orientations there were no particular alignments apparent in the data. Around S30°W, however, certain distinct alignments became apparent and these were tested against the various field criteria mentioned in Chapter 4. In particular, the relationship to various ice frontal positions and the morphological distinction of certain features were important constraints on the possible relationships of the shoreline fragments. Following this it was apparent that a number of separate shore zones could be defined. In addition to satisfying the various field criteria confidence in the reality of these shore zones was increased as one shoreline (later termed CLG2) apparently correlated a series of ice-marginal deltas that preceded a major drop in sea-level (the Otter Ferry Stage, Chapter 11) and another shoreline (later termed CLG5) occupies a very distinct position in the middle of the diagram, this shoreline being the last formed prior to the retreat of the ice from the Otter Ferry Stage.

In order to define more precisely the optimum projection plane regression analysis was carried out on the individual shorelines where the number of data points were adequate. The regression statistics were computed over the range S20W to S45W as it was known from visual inspection that outwith this range there were no marked alignments in the data. As pointed out in Chapter 4 it is the minimum gradient and the maximum correlation coefficient that define the optimum projection plane. The relevant correlation coefficients are given in Table 23 and the gradients in Table 24.

Two shorelines have a maximum correlation coefficient at the orientation S40W and two at S30W. None of the shorelines has a minimum gradient in the region for which they were calculated. This last result is unexpected and may be due to certain of the factors discussed in Chapter 4 as affecting this type of analysis. In particular where the number of data points is large their distribution may be critical and none of these shorelines has been traced for more than ca. 30 km down-dip. Hence gradient calculations may be unreliable. Taking into account the results of the correlation coefficients and also the direction of tilt established for the Main Rock Platform it was decided that S35W was the optimum plane of the Lateglacial shorelines. Figure 61 shows the data plotted in this plane.

The fragments classified as 'poor' during fieldwork were then plotted on this projection plane. It was found that the majority of these fragments plotted close to an already established shoreline. In addition a further shoreline was suggested in the top right hand corner of the diagram. Regression lines were re-computed for the various shorelines which were named CLG1-8, from top to bottom of the diagram. Table 25 gives the relevant statistics and Figure 62 shows the data and regression lines.

TABLE 25. REGRESSION STATISTICS FOR THE LATEGLACIAL SHORELINES.

Shoreline	No. of Points	Gradient (m/km)	Correlation Coefficient
CLG1	7	0.351	-0.989
CLG2	27	0.327	-0.977
CLG3	11	0.286	-0.986
CLG4	8	0.204	-0.955
CLG5	11	0.108	-0.823
CLG6	9	0.129	-0.804
CLG7	10	0.116	-0.822
CLG8	16	0.112	-0.714

3. The Individual Shorelines

Details of the shoreline diagram are now discussed, beginning with shoreline CLG1 and the small group of features in the top right hand of the diagram that are unrelated to any of the shorelines.

(a) Shoreline CLG1 and features formed prior to it.

Fragments T330, S324, S326 and S327 all occur in the Millhouse/Ardlamont area and have been discussed in Chapter 7.1. There it was shown that S324 was formed at the margin of a decaying ice mass and that T330 was probably formed at the same time by meltwaters escaping from the same ice mass. This was the first area to be deglaciated in the Cowal Peninsula for if the source of the ice was the mountains to the NE then fragments S357 in the Kyles of Bute and S156 at Kilfinan relating to shoreline CLG1 would have been covered by ice at the time that the ice front was located at Millhouse/Ardlamont.

Shoreline CLG1 is not well defined but the features correlated come from a wide variety of localities throughout the southern part of the study area. The shoreline combines fragments from Loch Striven (S370), S of Dunoon (S388), Kilfinan (S156), and the Kyles of Bute (S357) with fragments around Millhouse/Ardlamont that are lower than the local marine limit. The various localities correlated by this shoreline were thus deglaciated later than Ardlamont.

A difficulty is the apparent correlation of S370 beside Loch Striven with S380 at Ardyne for in the discussion of Ardyne (Chapter 6.1) it was suggested that S380 was formed close to an ice margin as there appeared no other source available for the sediment. Ice marginal to S380 suggests ice occupying Loch Striven which would refute the correlation. This illustrates a potential limitation of the constraint of ice cover placed on correlation of shoreline fragments, for if the ice front is retreating rapidly (and several hundred metres per year has been suggested for the Loch Striven glacier, Chapter 6.2) then little relative change of sea-level may take place in the few years necessary to deglaciate several kilometres of coastline (see Chapter 4). It is only where the distance between fragments is relatively large or where there is a halt in the retreat of the ice that this constraint can in practice be applied. It should be noted that fragment S75 does not belong to shoreline CLG1, for not only is this fragment at the very head of Loch Striven and hence separated from S370 by fragments at lower altitudes related to ice margins (Chapter 6.2), but the altitude of the fragment is strongly influenced by sediment derived from a neighbouring gorge.

Despite the above limitations it is suggested that the fragments correlated do define a shoreline. None of these fragments is unequivocally related to an ice front but it is possible that there was a halt or a slowing down of the rate of ice retreat prior to the formation of shoreline CLG2 for there is a decline in the altitude of the marine limit up lochs Fyne and Striven at this time. Neither the Loch Long nor the Loch Riddon glaciers appears to have responded in this manner, but the peculiar coastal configurations in both places (the opening of the Clyde Estuary and the junction of the Kyles of Bute) may have resulted in differences in retreat rates. The matter cannot be resolved, however, as there are too few reliable data points for this time.

(b) Shoreline CLG2

Shoreline CLG2 is the best developed Lateglacial depositional shoreline in the area. Fragments are correlated from all the major sea lochs and, more importantly, ice marginal deposits of all the various glaciers are also related to this shoreline. Thus the ice front at Otter Ferry (Chapter 7.3) is correlated with the major ice marginal delta S248 at the mouth of Glendaruel (Chapter 6.3), the ice-marginal deltas at the head of Loch Striven (Chapter 6.2), the highest shorelines relating to decaying ice in Strath Eachaig (Chapter 5.4) and the highest shorelines at the mouth of Loch Long. In addition decaying residual ice at Dunoon (Chapter 5.4), in the glens to the E of Loch Striven and at Millhouse/Ardlamont (Chapter 7.1) produced deposits that relate to this shoreline.

The inclusion of all the fragments S76, S77 and S78 from the head of Loch Striven in this shoreline needs some explanation, for these were previously interpreted as representing ice marginal positions and the paired fragments of S76 and S78 occur farther upvalley than does S77. It is suggested that the argument previously advanced for lower Loch Striven applies here also: with rapid ice retreat relative sea-level falls only slightly as lengths of coast are freed of ice.

Although there is good evidence for ice being present at the maximum up-valley locations of this shoreline in most valleys, this is not true of Loch Long where the shoreline fragments are deltas formed at the mouths of side valleys. The large size of these deltas perhaps implies decaying ice in the valleys to the W of Loch Long but the lack of any major ice-marginal features related to the Loch Long glacier is thought to be due to subsequent destruction by the glacier that occupied Loch Long during the Loch Lomond Stadial (Chapter 11).

Taking the evidence as a whole, however, the fact that such a large number of independently established ice front positions and areas of residual ice can be correlated by one shoreline is strong evidence for the validity of this shoreline. As is amplified below the various ice fronts remained stationary at the positions first attained during the formation of shoreline CLG2 while relative sea-level fell at least 20 m. This is seen as further support for the correlations proposed, for relative sea-level changes of such a magnitude in neighbouring valleys imply contemporaneity of the events. It is also considered significant that the best developed Lateglacial shoreline is correlated with the most important interruption in ice-sheet decay.

(c) Shorelines CLG3, CLG4 and CLG5

Shorelines CLG3, CLG4 and CLG5 were formed while the ice fronts were stationary at the positions first attained during the formation of CLG2. No shoreline fragments have been located inside these ice limits that could reasonably be correlated with these shorelines, even though there are many localities by Loch Fyne or in Glendaruel that would have been suitable for shoreline formation and preservation and that have not been influenced by glaciers of the Loch Lomond Stadial. Beyond the ice margins the shoreline fragments are distributed throughout the area and occur in all the sea lochs and at the mouths of many side valleys, highlighting the absence of similar fragments inside the ice limits.

The shorelines are distinguished from each other and from shoreline CLG2 by breaks of slope at a number of localities. Thus, for example, at the head of the Holy Loch S395 (CLG2) is upslope of S394 (CLG3), and at Ardlamont House S347 (CLG2) is upslope of S346 (CLG3) which is in turn upslope of S345 (CLG4). Although the majority of the fragments are isolated from one another there are sufficient instances of fragments of separate shorelines occurring one above the other to have confidence in the correlations proposed.

(d) Shoreline CLG6

Shoreline CLG6 is the first that is found inside the major ice limit. A number of well-preserved fragments of this shoreline occur between Otter Ferry and Minard by Loch Fyne, though no fragments have been identified inside the ice limits in the other valleys. This probably results from a variety of reasons. The Loch Striven valley, for instance, does not extend inland at an altitude where such a sea-level could penetrate; the valley bottoms of Strath Eachaig and Glendaruel are both occupied by major river terrace sequences related to sea-levels below CLG6, which presumably formed after and hence would have destroyed any shoreline features that may have formed in these parts of these valleys; and in Loch Long the glaciers of the Loch Lomond Stadial would have destroyed any shorelines formed prior to their advance. The possibility that the Loch Fyne glacier began to retreat before the others cannot be discounted, but for the reasons given above no firm evidence is available either for or against this occurrence.

Outside the ice limits fragments of shoreline CLG6 have been identified by Loch Striven, at the head of Loch Riddon and by lower Loch Fyne, which indicates that the shoreline is not a local phenomenon of mid-Loch Fyne.

Near the head of Loch Fyne fragments S9 and S110 appear to lie on the projection of shoreline CLG6. These two fragments are on opposite sides of Loch Fyne, S9 at Auchnatra and S110 at Rubha No. S9 was identified by Synge and Stephens (1966) as a delta and measured by Abney level as being at 79 feet (24.1 m) which compares with the height of

23.12 m recorded during this study. If these two fragments are indeed part of shoreline CLG6 it is curious that no other fragments between them and Minard have been identified. This point is reinforced by the apparent absence of shorelines CLG7 and CLG8 up-loch from Minard although in the same area Flandrian shorelines are frequently found (Chapter 10). This evidence, although largely negative, is thought to outweigh that of the two 'poor' fragments and shoreline CLG6 is not considered to extend up-loch of Minard. Evidence is presented in Chapter 11 that suggests that upper Loch Fyne was occupied by ice during the Loch Lomond Stadial. This, if true, would disprove any connection between shoreline CLG6 and fragments S9 and S110.

(e) Shorelines CLG7 and CLG8

Shorelines CLG7 and CLG8 are best developed by middle and lower Loch Fyne, although fragments of CLG7 have also been found beside Loch Striven and the Kyles of Bute. As was suggested for CLG6, the absence of these shorelines from the inner sea lochs is due to the subsequent occupancy of these areas by glaciers during the Loch Lomond Stadial. This is well established for Loch Long and is strongly suggested for Loch Fyne (Chapter 11). In the other portions of the sea-lochs formation and preservation of the lower shorelines is made increasingly difficult by the cliff associated with the Main Rock Platform which was, at least in part, formed during and slightly prior to the Loch Lomond Stadial (Chapter 8).

Apart from their restricted distribution the lower Lateglacial shorelines can in certain instances be distinguished from the upper Flandrian shorelines on sedimentological grounds. A section in S195 (CLG7) reveals nearshore cobbles and pebbles overlying grey finely laminated clay, whilst a trench dug in S309 (CLG7) also revealed a grey occasionally sandy clay. Fragment T300 (CLG7) has a section that shows up to 3 m of grey clay with rounded pebbles scattered throughout, possibly suggesting the presence of sea-ice at the time of their deposition. In this mid-Loch Fyne area sections in Flandrian shoreline fragments reveal a medium well-sorted gravel.

Although many of the fragments of shorelines CLG7 and CLG8 occur in the same area, few lie immediately one above the other due to the complicated distribution of small terrace fragments. Fragment S278 (CLG7) does, however, clearly stand higher than S279 (CLG8).

4. The Shoreline Sequence

Seven, possibly eight shorelines (CLG1 being doubtful) have been identified that appear to have been formed either in association with the retreating ice sheet or after this but prior to the disappearance of the glaciers of the Loch Lomond Stadial. The shorelines are therefore Lateglacial in age. They form a sequence in which the gradient of the earliest shoreline is greatest (CLG1: 0.35 m/km), the gradients of subsequent shorelines generally decreasing with time of formation. An exception is CLG5 but this may only be a consequence of the limitations discussed previously on the accurate determination of shoreline gradients.

The best defined shoreline is CLG2 which has been established over a distance of ca. 30 km along the projection plane. It has a gradient of 0.33 m/km which is comparable with that of the Main Perth Shoreline in SE Scotland of 0.43 m/km (Smith et al., 1969) and the lower Perth Shorelines in the Tay-Earn Area of 0.29 - 0.23 m/km (Cullingford, 1977).

The formation of particular shorelines during a period of general regression of sea-level such as is documented here for the Lateglacial period requires particular conditions to obtain at the time of formation of the shorelines. Such conditions can result from a temporary balance in the rates of isostatic and eustatic movements and/or from the supply of increased amounts of sediment to the coast. Shoreline CLG2 is most readily explicable in such terms for it was formed at a time when the ice margin halted (or possibly readvanced), thus presumably acting to slow somewhat the rate of isostatic recovery of the crust. The presence at this time of a number of isolated decaying ice masses in the valleys beyond the margin of the major ice body would ensure a substantial supply of sediment to the valley mouths around the coast.

The formation of the next shorelines in the sequence (CLG3, CLG4 and CLG5) while the ice front was stationary is less readily explicable. Edelman (1968) has documented shorelines formed during a continuous regression of the sea though he was considering essentially locally developed features rather than regionally identified shorelines. Perhaps these shorelines coincided with brief eustatic fluctuations which more frequently matched the slower rates of isostatic recovery resulting from the stationary ice front, or perhaps brief climatic fluctuations resulted in an increased supply of sediment to the coastal zone with the accompanying formation of well-developed shorelines.

5. Dating the Shorelines

No material has been recovered from the shoreline deposits that would allow an accurate dating of the shorelines. There are, however, over 40 radiocarbon dates on marine shells available from the SW Highlands and the Clyde Estuary and a number of these may be used to establish the ages of certain events more closely. Appendix 1 reviews the practice involved in marine shell radiocarbon dating and attempts to produce a standarized list of the dates from this region. Ages where quoted below are taken from the Appendix and not from the numerous publications relating to these dates which are sometimes erroneous and often inconsistent.

Dated samples that are of use in the present context can be divided into three categories.

(a) Samples relating to specific altitudes above present sea-level. Sea-level was therefore at or above the sample altitude when the shell died.

(b) Samples associated with faunas that require specific water conditions. These samples usually indicate sea-levels within rather wide limits.

(c) Samples relating to specific geological events that can themselves be related to sea-levels. The earliest date after deglaciation or the earliest date after certain glacial stages place limits on the ages of shorelines tied directly to these events. Samples of these three types are all available and are discussed in turn.

(a) Samples related to specific altitudes

Only two samples have been dated that can confidently be considered to occur at altitudes above that attained by the sea during the Flandrian: these are Birm-361 at +20.5 m from Glen Cruitten at Oban and IGS-C14/21 at +16.0 m from Ralston, Paisley. Neither of these samples came from the study area. Birm-361 does not show good agreement between the various fractions dated and the date of 11,370 $\pm$ 230 yr BP for a sea-level at or above 20.5 m O.D. is in disagreement with numerous other dates below. This sample is therefore considered unreliable. Sample IGS-C14/21 is internally consistent but occurs at only +16.0 m O.D., implying sea-level to have been at or above this altitude at this time.

Sample SRR-62 is given an altitudinal range of 9.0 to 12.25 m O.D. in Harkness and Wilson (1974) implying that sea-level was at or above the top of this range at 12,390 $\pm$ 90 yr BP. Other samples that are of significance in this category are the two dates from Portavadie, SRR-831 and SRR-832. These are internally consistent at the 2σ level and indicate sea-level as having been above +5.2 m O.D. at 11,530 $\pm$ 80 yr BP and above +2.0 m O.D. at 10,740 $\pm$ 220 yr BP. The youngest sample from Cardross (GU-12) indicates sea-level was above +7.5 m O.D. at 11,390 $\pm$ 120 yr BP.

(b) Samples associated with faunas requiring specific water conditions.

At two localities (Lochgilphead and Ardyne) within the study area a number of radiocarbon samples have been taken from marine clays in which detailed analyses have been performed on the macro- and micro-faunas that allow inferences to be made as to the depth of water in which the dated samples lived.

At Lochgilphead (Peacock <u>et al</u>., 1977) two samples (SRR-63; SRR-369) have been related to sea-level at 25 $\pm$ 5 m O.D.. This relationship was not established entirely on faunal grounds, however, but by attempting to relate the conditions of sedimentation at the Lochgilphead site to the time that the sea ceased to penetrate through the col presently occupied by the Crinan Canal. Of the two samples, only one fraction of SRR-63 (12,350 $\pm$ 90 yr BP) was dated and with SRR-369 (12,250 $\pm$ 110 yr BP) the inner and outer fractions differ by more than 2σ, although the two samples considered together support each other. On the height-distance diagram (Fig. 62) the altitudinal range associated with these samples brackets CLG5, certainly occurring above CLG6.

Of the later dates at Lochgilphead, SRR-368 appears out of stratigraphic order, SRR-489 according to Peacock et al. (1978) is not in situ and SRR-367 was dated on only one fraction and, occurring close to SRR-364/5/6 in the section, appears too old. Samples SRR-364/5/6 are not entirely coherent internally (inners and outers all differing by $\pm$300 or 400 years) although taken together the inner dates appear to support each other. The associated fauna and sediment suggest a sea-level of between +10 and +4 m O.D. between approximately 11,300 yr BP and 11,000 yr BP. This altitudinal range occurs below the level of the uppermost Flandrian shorelines in this area (Chapter 10) and shoreline CLG8 must therefore have been formed prior to this time. The altitudinal limits do, however, overlap with the altitude of the Main Rock Platform in this area (Chapter 8) indicating this as a possible time for at least part of its formation.

At Ardyne (Peacock et al., 1978) the earliest date is from sample SRR-482 from the junction of Unit I and Unit II of the local sedimentary sequence. There is disagreement between the inner and outer fractions of this sample at the 2σ level. Both Unit I and Unit II sediments occur in an exposure at the side of the Ardyne Burn near Killellan Farm (Fig. 31) and the altitude of this exposure plus the faunal and sedimentary characteristics of the deposit suggested to Peacock et al. a sea-level at this time (11,930 $\pm$ 90 yr BP) of up to +20 m O.D.. The Killellan exposure, however, was deposited prior to terraces T346 and T354, both of which extend farther down valley than Killellan (Chapter 6.1) descending to ca. 20 m O.D. thus suggesting a sea-level of possibly more than 20 m O.D. at the time the Killellan shelly clays were deposited. Taking both lines of evidence together suggests a sea-level of 20 m O.D. at approximately 11,900 yr BP. This plots on the shoreline diagram below CLG5 but above CLG6.

The remaining samples from Ardyne (SRR-481/3/4/5/6, SRR-615) all relate to the period from ca. 11,300 to ca. 10,200 yr BP. There is a considerable internal consistency in the dates and stratigraphically they are also in good agreement. The associated faunas indicate water depths of below 20 m and probably less than 7 m with a hint of littoral conditions at one stage. The samples occur at altitudes of -1.0 to -5.5 m O.D. and sea-level during this period was probably between +2.0 and +10.0 m O.D. This is in general agreement with the Lochgilphead data, again indicating formation of shoreline CLG8 prior to this time and pointing to this period as being one in which the Main Rock Platform could have been eroded.

(c) Samples relating to specific geological events

It has been shown, both on the basis of field evidence and with the shoreline diagram that the marine limit over most of the southern part of the study area was formed at the time of deglaciation. At a number of sites in the Firth of Clyde and the Clyde Estuary attempts have been made to date the lowest fossiliferous beds in the marine sequence and hence produce a minimum age for deglaciation. The classic 'Clyde Beds' sedimentary sequence shows a layer of very poorly fossiliferous laminated clays resting on till and overlain by a much more fossiliferous grey silty sand with scattered cobbles and boulders. It is this last bed that is normally dated and it is generally assumed that the laminated clays were deposited rather rapidly in the proximity of glacier ice. It has been suggested (Peacock et al., 1978) that in places there were periods of non-sedimentation or reworking of sediments during the deposition of the Clyde Beds such that the sediments that immediately overlie the laminated clays do not always follow them conformably.

Of the samples relating to the earliest dated marine sediments those that provide the greatest ages all come from the Clyde Estuary (or its earlier extension around Renfrew and Paisley). Two very old dates, IGS-C14/20 (15,630 ± 240 yr BP) and SRR-923 (13,950 $^{+620}_{-580}$ yr BP) should be rejected for there is very considerable discrepancy (over 2000 years) between the inner and the outer fractions in both these samples. The earliest internally consistent sample is SRR-925 from Inchinnan which gave an age of 13,100 ± 270 yr BP. In addition, a group of samples occurs around 12,600 yr BP (Birm-122: 12,650 ± 200 yr BP; IGS-C14/68: 12,620 ± 230 yr BP; IGS-C14/21: 12,610 ± 210 yr BP; and SRR-927: 12,520 ± 130 yr BP). Internally, these samples are somewhat variable, only IGS-C14/21 having a difference between inner and outer ages that is less than 1σ. As a group, however, they are consistent, though they differ at the 1σ level from SRR-925 at Inchinnan.

There are no dates comparable with these from the Clyde Sea Area. The earliest sample is from Lochgilphead, SRR-368 (12,790 ± 85 yr BP) but this sample is internally inconsistent and is out of stratigraphic order and hence must be rejected. The earliest reliable date is SRR-63 (12,350 ± 90 yr BP) at Lochgilphead and this is known not to date deglaciation as there is a marked unconformity at the base of the Lochgilphead marine deposits.

It is of note that there is a considerable number of rather old dates providing minimum ages for deglaciation in the Clyde Estuary and none from the Firth of Clyde, for on geological grounds the Firth of Clyde should have been deglaciated first (this being effectively implied by the shoreline diagram as well as by the a priori idea that calving glaciers retreat fastest). In part this may be a sampling problem, for more localities have been sampled in the area of the Clyde Estuary than elsewhere. In addition it is possible that the retreat of the sea from the inner Clyde Estuary at a relatively early period reduced the likelihood of reworking of the sediments such as perhaps has happened in some localities around the coasts of the Firth of Clyde. A further point to be considered, however, is the possibility that these shells, inhabiting an estuary that must have initially received a considerable water influx containing glacially comminuted fragments from the neighbouring Carboniferous rocks, have been influenced by a hard water error. Mangerud (1972) considered this problem and suggested that shells from restricted estuaries with large fresh-water intakes would not be suitable for dating. He also considered and ruled out the possibility that the CO_2 contained in glacier ice might affect the carbon content of the estuary or fjord into which the glacier discharged. A hard water error might explain the very high ages achieved for the inner portions of samples IGS-C14/20 and SRR-923. The possibility of this type of error, however, poses a problem in assessing the dates from the Clyde Estuary. The effect could be expected to diminish with time due to reduced water inputs as ice finally decayed and vegetation became established and thus, perhaps, the younger ages (ca. 12,600 yr BP) are the more reliable estimates of minimum ages of deglaciation.

The implications for the present study area of the date of deglaciation of the Clyde Estuary are difficult to assess accurately. As has been mentioned above it is most probable that the southern part of the Cowal Peninsula was deglaciated prior to the Clyde Estuary but the relationship between shoreline CLG2, for example, and the raised marine landforms of the Clyde Estuary is not known, mainly because the marine limit in the Clyde Estuary has not been accurately established (cf. Rose, 1975; Sissons, 1976a, p 128) and the actual mode of deglaciation is in debate (Peacock, 1971; Sissons, 1976a, p 128). If 12,600 yr BP is accepted as a minimum age for deglaciation in the Clyde Estuary then a date of deglaciation of southern Cowal of approximately 13,000 yr BP may tentatively be suggested.

One sample is available that post-dates retreat of the ice from the halt associated with the major drop in sea-level in southern Cowal. This is sample T-1456 from Loch Goil which gave an age of 12,260 $\pm$ 150 yr BP. If correct, then shoreline CLG5 formed before this time and thus sea-level was at or below ca. 23 m O.D. at this time.

A number of radiocarbon dates relate to samples that have been over-ridden or transported by glacier ice (Table 1, Appendix 1). These suggest that the Loch Lomond Readvance reached its maximum some time after ca. 11,100 yr BP. The ice limits of these glaciers relate to sea-levels below the maximum height of Flandrian shorelines. The cold sea conditions associated with this glacial advance terminated around 10,200 yr BP (Peacock et al., 1978). Sea-level during this period appears to have been consistently below the uppermost Flandrian shoreline.

A final piece of evidence is available relating to the timing of

Lateglacial sea-level changes. At Troustan by Loch Striven a thin rather wispy peat layer was found at the top of the present foreshore at ca. 2.2 m O.D.. Preliminary pollen analyses (by Dr. J.J. Lowe and Dr. M.J.C. Walker) indicated a sparse vegetation typical of Lateglacial conditions (Appendix 2). The peat appears to indicate that relative sea-level at one time during the Lateglacial approached the altitude of present sea-level. This possibility compares with the suggestion of littoral conditions in samples from below O.D. during the later part of the Lateglacial sequence at Ardyne Point only 8 km from Troustan (Peacock et al., 1978).

The above discussion can be summarized on a graph of sea-level altitude against time, the various samples being plotted with their uncertainties in both time and altitude indicated by boxes or arrows (Fig. 63). Construction of such a graph must take into account isostatic tilting. The data have all been standardized therefore by relating them to the 58 km distance on the shoreline diagram and using the approximate degree of tilting as indicated by this diagram to adjust the heights as necessary. The 58 km point was chosen as it requires least adjustment to altitude values, occurring equidistant from the plot of Lochgilphead (57 km) and Ardyne (59 km). Estimates of the isostatic correction necessary for the Cardross (-1.5 m) and Paisley (0 m) samples were made by comparing the localities with the isobase maps in Sissons (1976a, p 130) and Gray (1978) since these indicated the approximate curvature of the isobases over these greater distances.

Although some of the uncertainties appear large there is still considerable agreement among a large number of samples. The group of samples around 12,000 to 12,500 yr BP results from four different radiocarbon dates at three separate localities, whilst the sea-level indicators for the period 11,500 to 10,200 yr BP summarize 12 radiocarbon dates from 5 separate localities and this cross-checks on what is known about sea-level during the period of existence of the glaciers of the Loch Lomond Stadial.

Also included on Figure 63 are the altitude of the various Lateglacial shorelines at kilometre 58 and the fall of sea-level during the period of ice-front still-stand. The data available do not define a single curve of sea-level change but indicate a zone in which sea-level occurred at a particular time. It is thus not possible to provide exact dates for particular shorelines but rather a time range in which a particular event probably occurred (thus being analogous to a radiocarbon date). The graph indicates, for example, that shoreline CLG2 probably formed at some time between _ca_. 13,100 and _ca_. 12,700 yr BP and shoreline CLG5 at some time between _ca_. 12,650 and _ca_. 12,250 yr BP. Since these shorelines approximately bracket the time during which the ice front was stable this suggests a period of 50 to 850 years for this event. As the evidence for the event is substantially less well developed than that for the Loch Lomond Readvance the actual time involved was probably the lower end of this estimate.

The sea-level curve conforms to the general pattern of such curves from glacio-isostatically affected areas (e.g. Olsson and Blake, 1962; Andrews, 1970) and agrees in broad outline with the curves derived from Lochgilphead and Ardyne by Peacock _et al_. (1978) although differing from these last by utilising a greater number of data points from more diverse localities. Rapid overall emergence of the land is indicated, a mean figure of _ca_. 1.8 m/100 yr being obtained for the period 13,000 to 11,500 yr BP which, although comparable with a number of other deglaciated areas such as Spitsbergen (e.g. Olsson and Blake, 1962) and Northern Norway (e.g. Marthinussen, 1962), is still considerably less than the rate of uplift being experienced today in parts of Arctic Canada (Andrews, 1970, p 120). There are not enough data points available to indicate the minor fluctuations in uplift rate that field evidence suggests, particularly in relation to the halt in the retreat of the ice sheet and the formation of shoreline CLG2. There is, however, a strong suggestion of a considerable slowing down in the rate of emergence in the later part of the Lateglacial and it may be speculated that this was due to the build-up of the Loch Lomond Readvance glaciers a possibility that is developed further in Chapter 11.

TABLE 23. CORRELATION COEFFICIENTS FOR LATEGLACIAL SHORELINES.

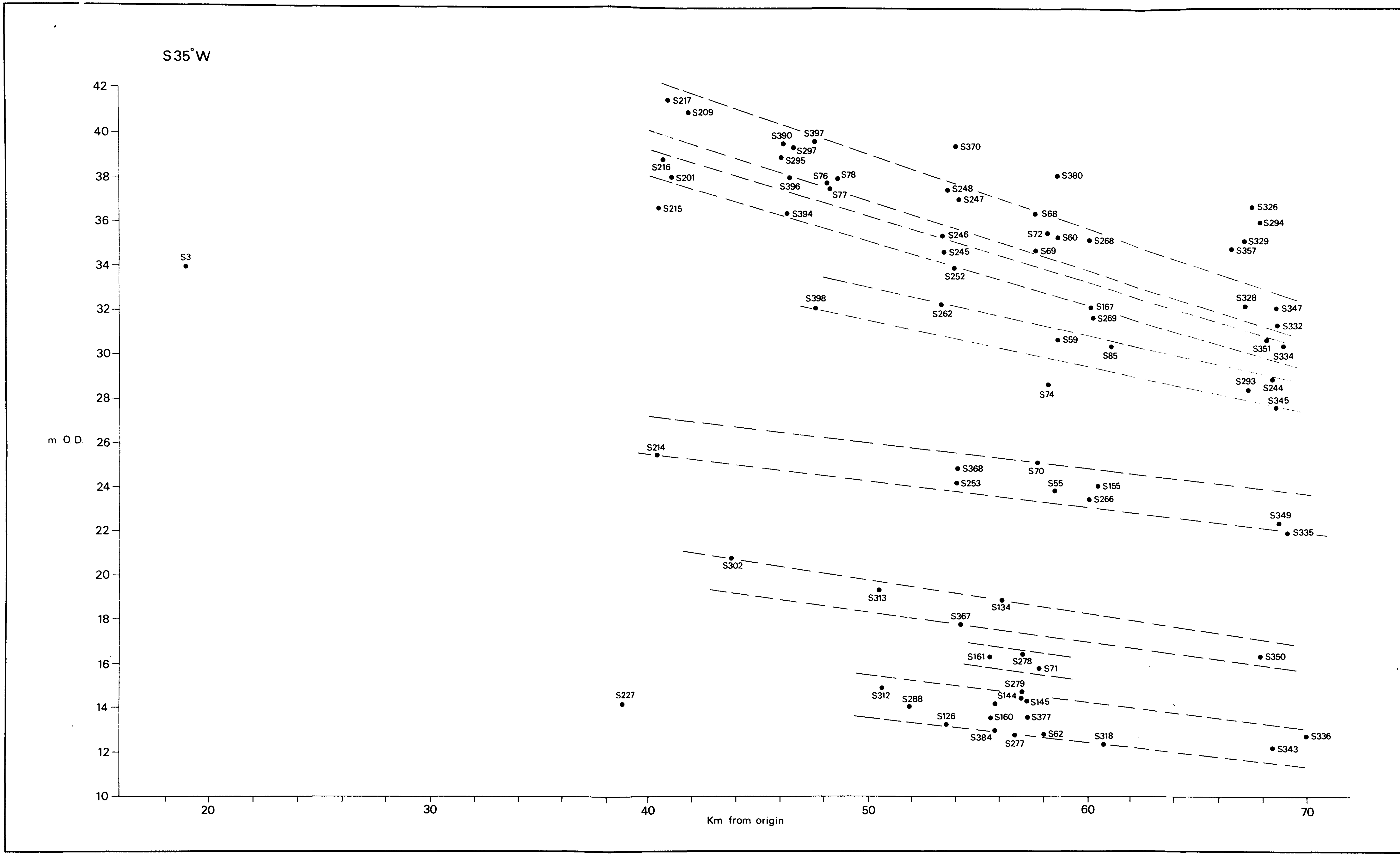
Shoreline Orientation	a	b	c	d	e
S20W	-0.956	-0.958	-0.984	-0.873	-0.717
S25W	-0.967	-0.972	-0.976	-0.881	-0.723
S30W	-0.974	-0.981	-0.966	-0.883*	-0.723*
S35W	-0.977	-0.986	-0.953	-0.881	-0.714
S40W	-0.978*	-0.988*	-0.939	-0.876	-0.692
S45W	-0.976	-0.986	-0.923	-0.869	-0.649

* denotes maximum value

TABLE 24. GRADIENTS FOR LATEGLACIAL SHORELINES.

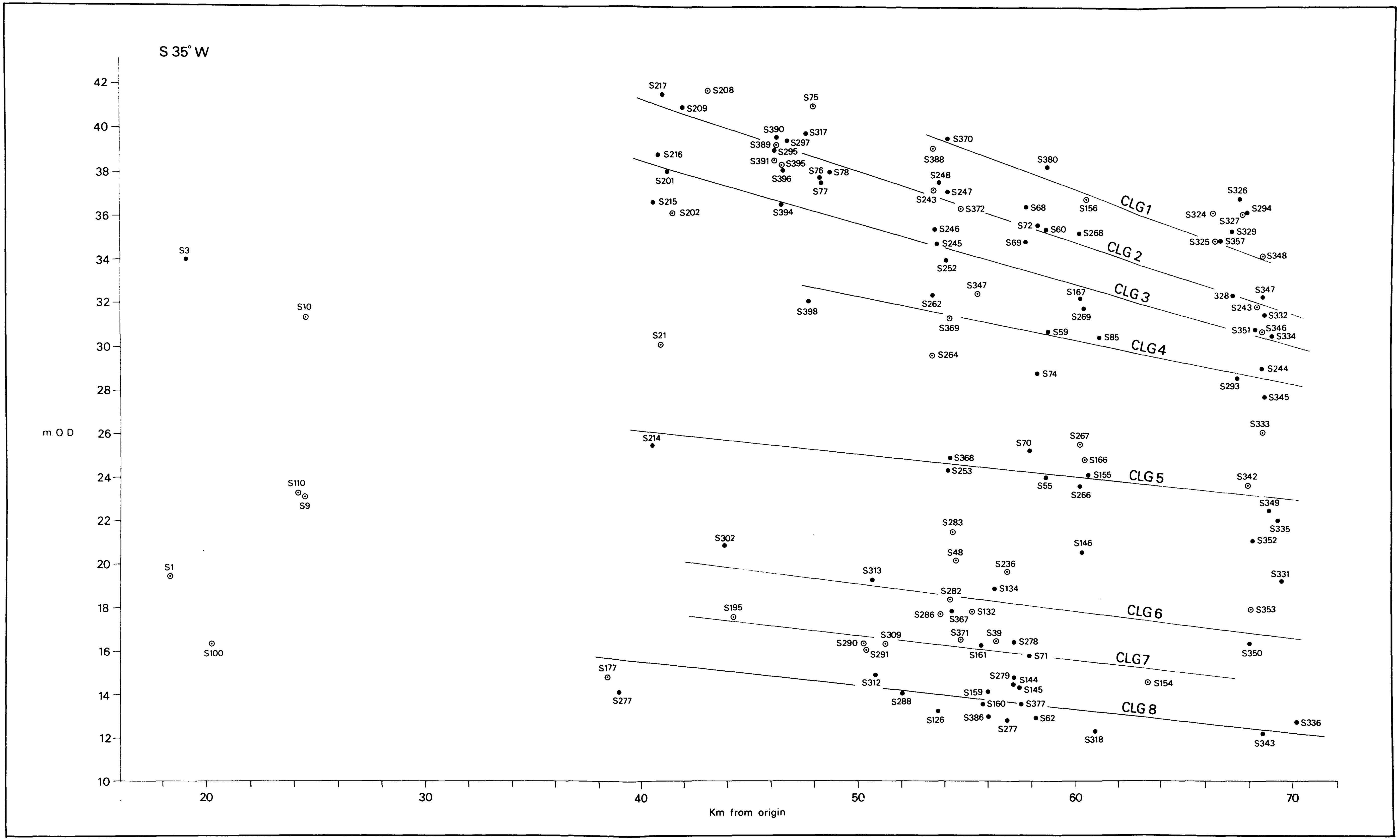
Shoreline Orientation	a	b	c	d	e
S20W	-0.356	-0.327	-0.239	-0.133	-0.090
S25W	-0.344	-0.314	-0.227	-0.128	-0.097
S30W	-0.333	-0.303	-0.216	-0.124	-0.105
S35W	-0.326	-0.292	-0.207	-0.120	-0.112
S40W	-0.315	-0.282	-0.199	-0.116	-0.117
S45W	-0.308	-0.272	-0.191	-0.112	-0.118

Figure 61: Height-distance diagram of Lateglacial shorelines, S35^{o}W, using features classified as 'good' and 'moderate'.



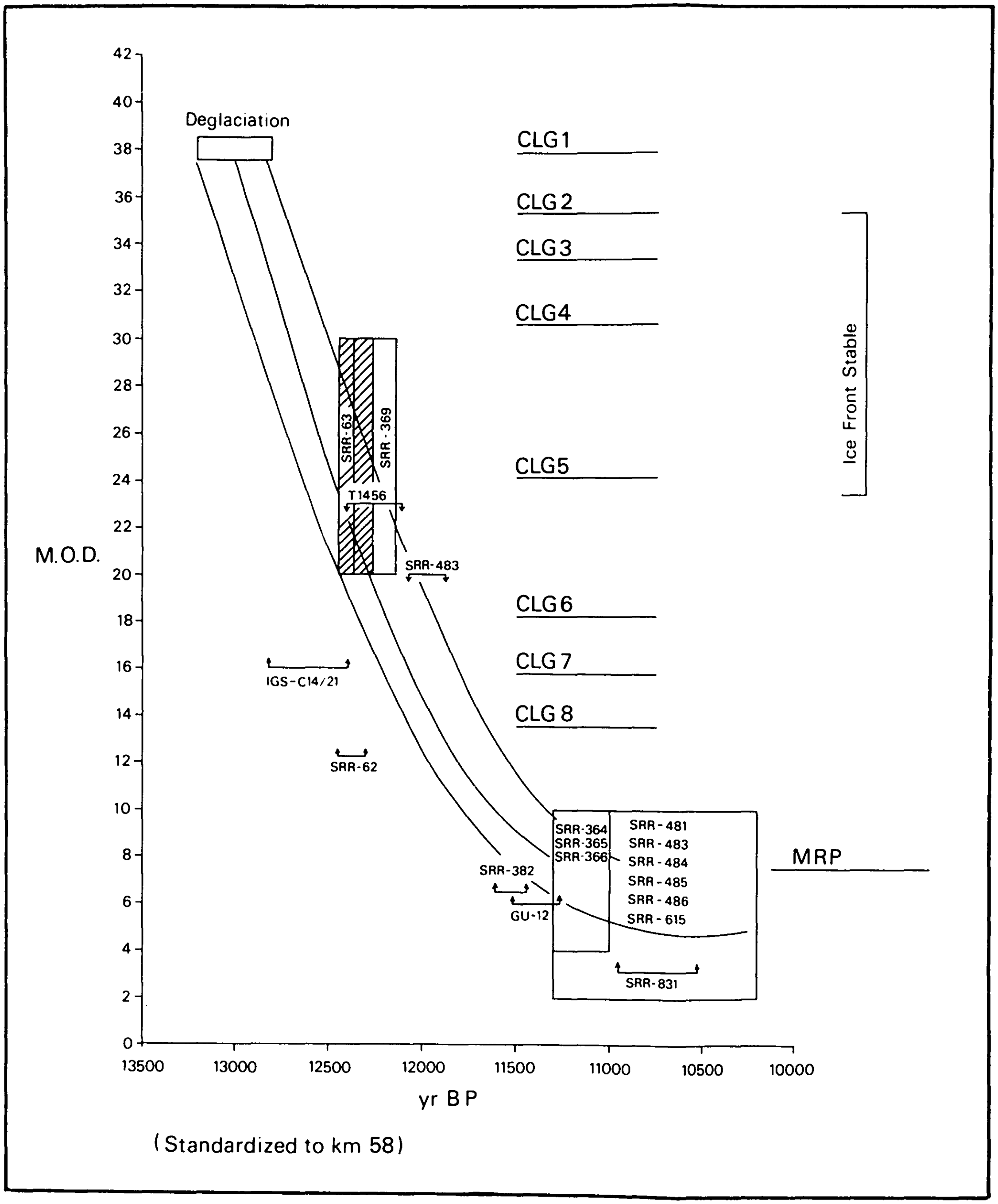
S35°W
m O.D.
Km from origin
S3
S217
S209
S390
S397
S297
S295
S216
S201
S396
S76
S78
S77
S215
S394
S370
S248
S247
S380
S68
S72
S60
S268
S246
S245
S69
S252
S326
S294
S329
S357
S398
S262
S167
S269
S59
S85
S74
S328
S347
S332
S351
S334
S293
S244
S345
S214
S368
S253
S70
S55
S155
S266
S349
S335
S302
S313
S134
S367
S350
S161
S278
S71
S279
S144
S145
S312
S288
S160
S377
S126
S384
S277
S62
S318
S336
S343
S227

Figure 62: Height-distance diagram of Lateglacial shorelines, S35°W, all features.



S 35° W
m O D
Km from origin
CLG 1
CLG 2
CLG 3
CLG 4
CLG 5
CLG 6
CLG 7
CLG 8

Figure 63: Lateglacial sea-level curve.


Deglaciation
CLG1
CLG2
CLG3
CLG4
CLG5
CLG6
CLG7
CLG8
MRP
Ice Front Stable
SRR-63
SRR-369
T1456
SRR-483
IGS-C14/21
SRR-62
SRR-364
SRR-365
SRR-366
SRR-382
GU-12
SRR-481
SRR-483
SRR-484
SRR-485
SRR-486
SRR-615
SRR-831
M.O.D.
yr B P
0
2
4
6
8
10
12
14
16
18
20
22
24
26
28
30
32
34
36
38
40
42
13500
13000
12500
12000
11500
11000
10500
10000
(Standardized to km 58)

CHAPTER 10

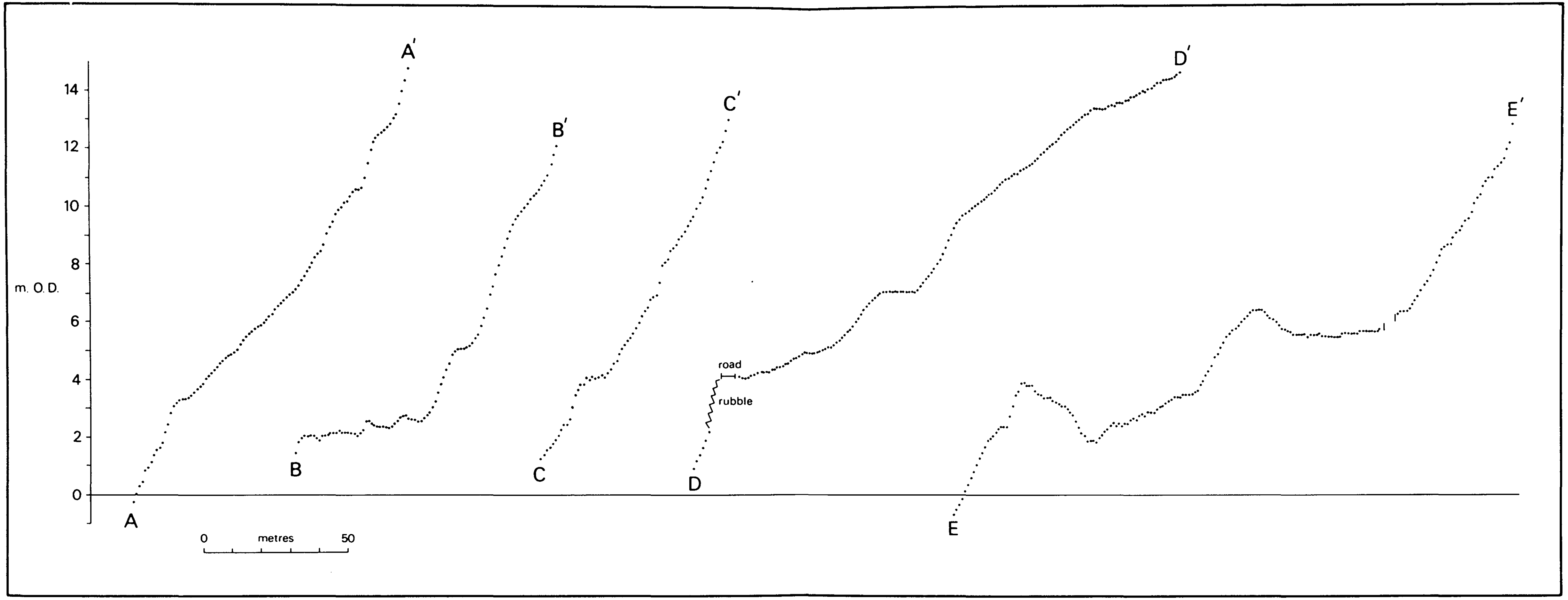
SHORELINES FORMED DURING THE FLANDRIAN

1. Introduction

In the previous chapter the shorelines formed during the Lateglacial period were analysed. At lower altitudes than the Lateglacial shorelines and frequently distinguishable from them on stratigraphic grounds are numerous raised marine landforms which, as is shown later in this chapter, have been formed during the Flandrian period. These lower marine deposits are developed from place to place throughout the study area and can be inferred on that basis to have been formed after the disappearance of glacier ice from the sea-lochs. Additionally they occasionally contain beds of peat which radiocarbon assay indicates to be Flandrian in age. The marine landforms are dominantly gravel with occasional sand layers, though in two areas, at the head of Loch Riddon and the head of Loch Gilp, there occurs a grey sandy silt facies similar to the carse deposits extensively developed in parts of eastern Scotland.

The upper limit of the Flandrian marine deposits is around 14 m O.D. at the heads of the sea-lochs but declines in altitude in a southerly direction to ca. 11 m O.D. at the southern end of Loch Fyne. At lower altitudes there is an abundant development of small deltas and shingle ridges as has been documented in the various descriptions of particular areas (Chapters 5, 6 and 7). A number of cross-profiles was surveyed, altitudes being recorded every two paces (ca. 2 m), in order to demonstrate the nature of the succession of ridges and steps to which these marine deposits give rise (Figure 64). These cross-profiles clearly reveal the breaks of slope at the back edges of many of the landforms but they also indicate the short vertical intervals between the separate breaks

Figure 64: Cross-profiles of Flandrian marine deposits.



A
A′
B
B′
C
C′
D
D′
E
E′
m. O.D.
0
2
4
6
8
10
12
14
road
rubble
0
metres
50

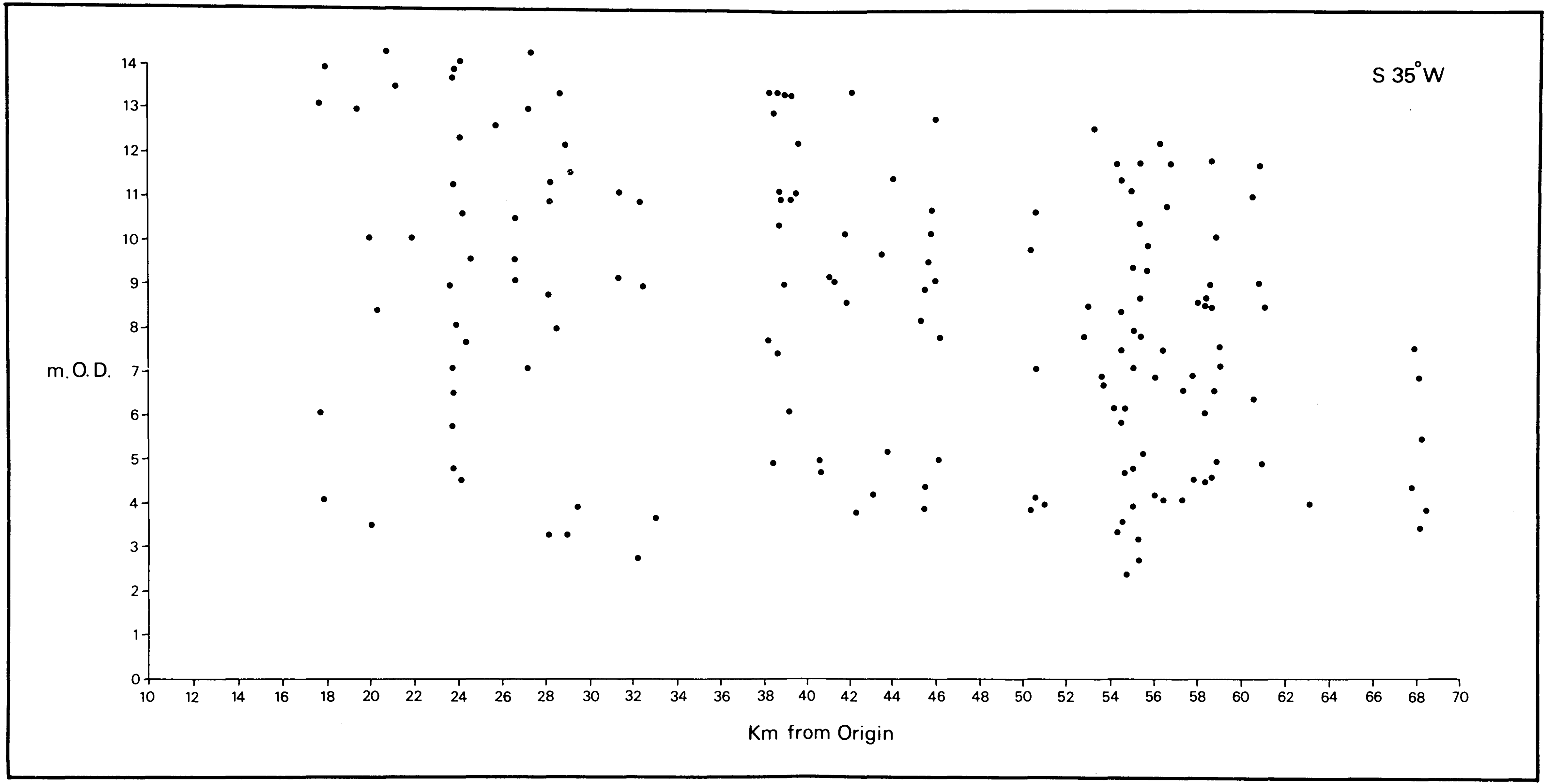
of slope and, when the overall degree of accuracy of the methodology is considered (Chapter 4), it is apparent that there is some potential for confusion when attempting to correlate marine landforms from place to place around the coast.

2. Construction of Height-Distance Diagram

The methods employed in correlating the shorelines are those outlined in Chapter 4, that is, by height-distance diagram in association with various field criteria. In brief, the mean altitude of the various marine landforms (excluding shingle ridges and features classified as 'poor') were projected at right angles into a series of vertical planes, the planes being successively rotated by 5^{o} intervals from due S to due W. At each 5^{o} interval the distances along the plane from a common origin to the various projected points were calculated and a height-distance diagram constructed. Each height-distance diagram was inspected for marked alignments of points which, it was anticipated, should have a generally southerly gradient due to isostatic tilting. This procedure was assisted by the fact that the upper limit of Flandrian marine deposits is well defined and, as mentioned above, has a generally southerly gradient. This approach proved unproductive, however, in revealing any distinct shorelines, Figure 65 being a typical plot of the various landforms.

A more rigorous approach was then adopted and only those features classified as 'good' or those that had been established, in the discussion of individual areas, to correlate with other features in the same area were utilised. Once again shingle ridges were excluded. Approximately 100 landforms were involved the above procedures being repeated. This approach was successful in revealing a number of distinct

Figure 65: Height-distance diagram of Flandrian shoreline features.


S 35° W
m. O. D.
Km from Origin
0
1
2
3
4
5
6
7
8
9
10
11
12
13
14
10
12
14
16
18
20
22
24
26
28
30
32
34
36
38
40
42
44
46
48
50
52
54
56
58
60
62
64
66
68
70

groups of landforms as is shown in Figure 66. Five, possibly six groups could be established although they were not all defined with equal clarity for the whole diagram. The groups were named CF1 to CF6 respectively from the highest downwards.

In order to discover the optimum projection plane more accurately regression analysis was carried out on the four uppermost groups of points, these being the best represented. The objective was to locate the plane (or planes) in which the gradient of the shoreline was minimized and the correlation coefficient of the regression line maximized (Chapter 4). It was found, as with the Lateglacial shorelines, that there was considerable variation in the directions in which these conditions were met, due, presumably, to the low gradients and the relatively high degree of uncertainty associated with each feature. Figure 67 shows the variation of each quantity with rotation of the projection plane and Table 26 below summarizes the data on the optimum planes of projection.

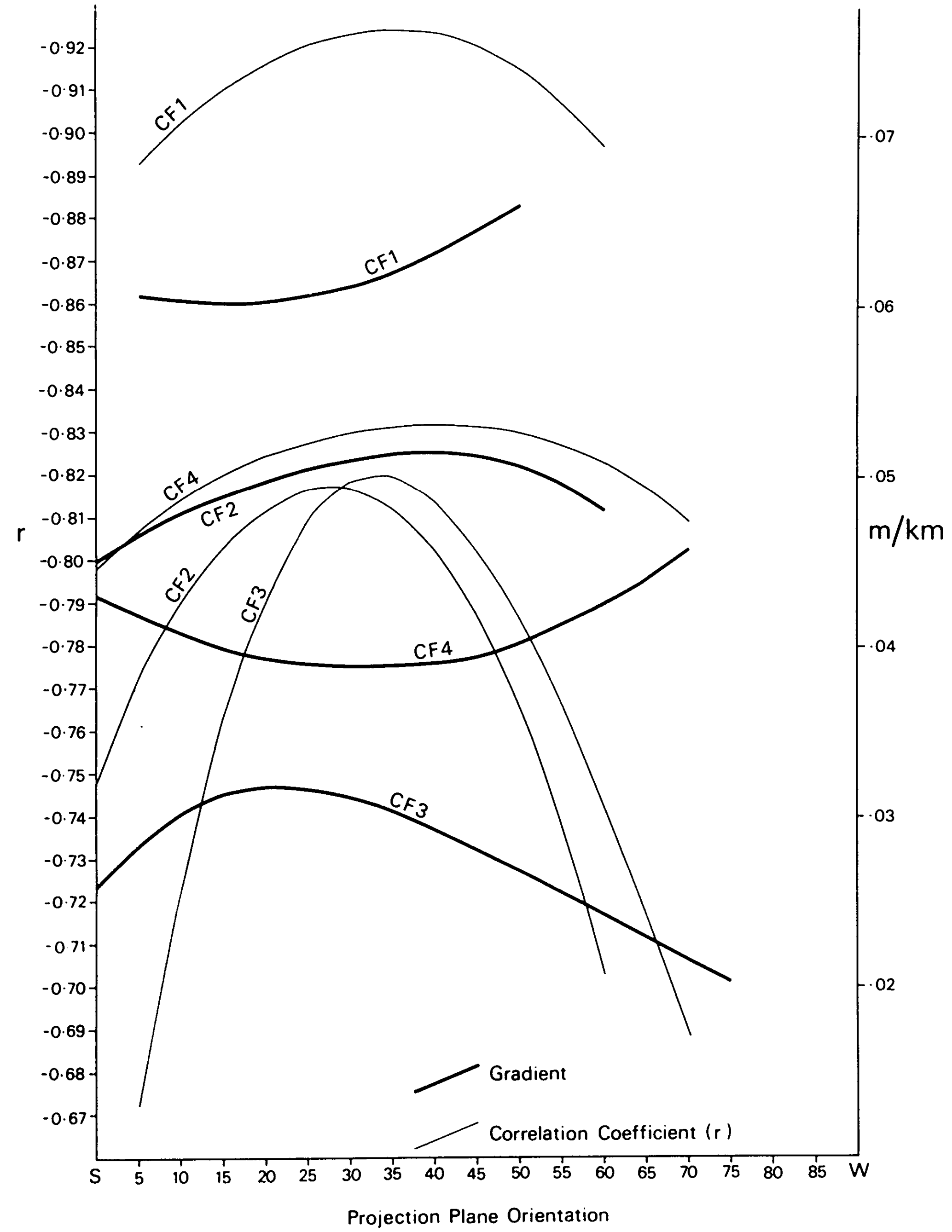
TABLE 26. OPTIMUM PLANES OF PROJECTION FOR SHORELINES CF1 - 4.

Shoreline	CF1	CF2	CF3	CF4
Gradient	S15W	$\leqslant$ S	$\geqslant$ S75W	S30W
r	S35W	S25/30W	S35W	S40W

It is thought that the amount of variation in the figures, particularly in the gradients (which vary very little in absolute terms from orientation to orientation) is due to inadequacies in the data rather than irregularities of isostatic recovery.

In previous chapters it was found that the optimum plane of projection for the Main Rock Platform and for the Lateglacial shorelines was S35W. The data for the Flandrian shorelines when considered together

Figure 67: Graph of variation in gradients and correlation coefficients of Flandrian shorelines with change of projection planes.



r
-0·92
-0·91
-0·90
-0·89
-0·88
-0·87
-0·86
-0·85
-0·84
-0·83
-0·82
-0·81
-0·80
-0·79
-0·78
-0·77
-0·76
-0·75
-0·74
-0·73
-0·72
-0 71
-0·70
-0·69
-0·68
-0·67
·07
·06
·05
·04
·03
·02
m/km
CF1
CF1
CF4
CF2
CF2
CF3
CF4
CF3
Gradient
Correlation Coefficient (r)
S
5
10
15
20
25
30
35
40
45
50
55
60
65
70
75
80
85
W
Projection Plane Orientation

are not markedly different from that figure and S35W has been adopted as the most suitable projection plane. The gradients and correlation coefficients for the individual shorelines in this projection plane are given in Table 27 below, shorelines CF5 and CF6 having been included for comparison.

TABLE 27. GRADIENTS AND CORRELATION COEFFICIENTS OF SHORELINES CF1 - 6 IN S35W PROJECTION PLANE.

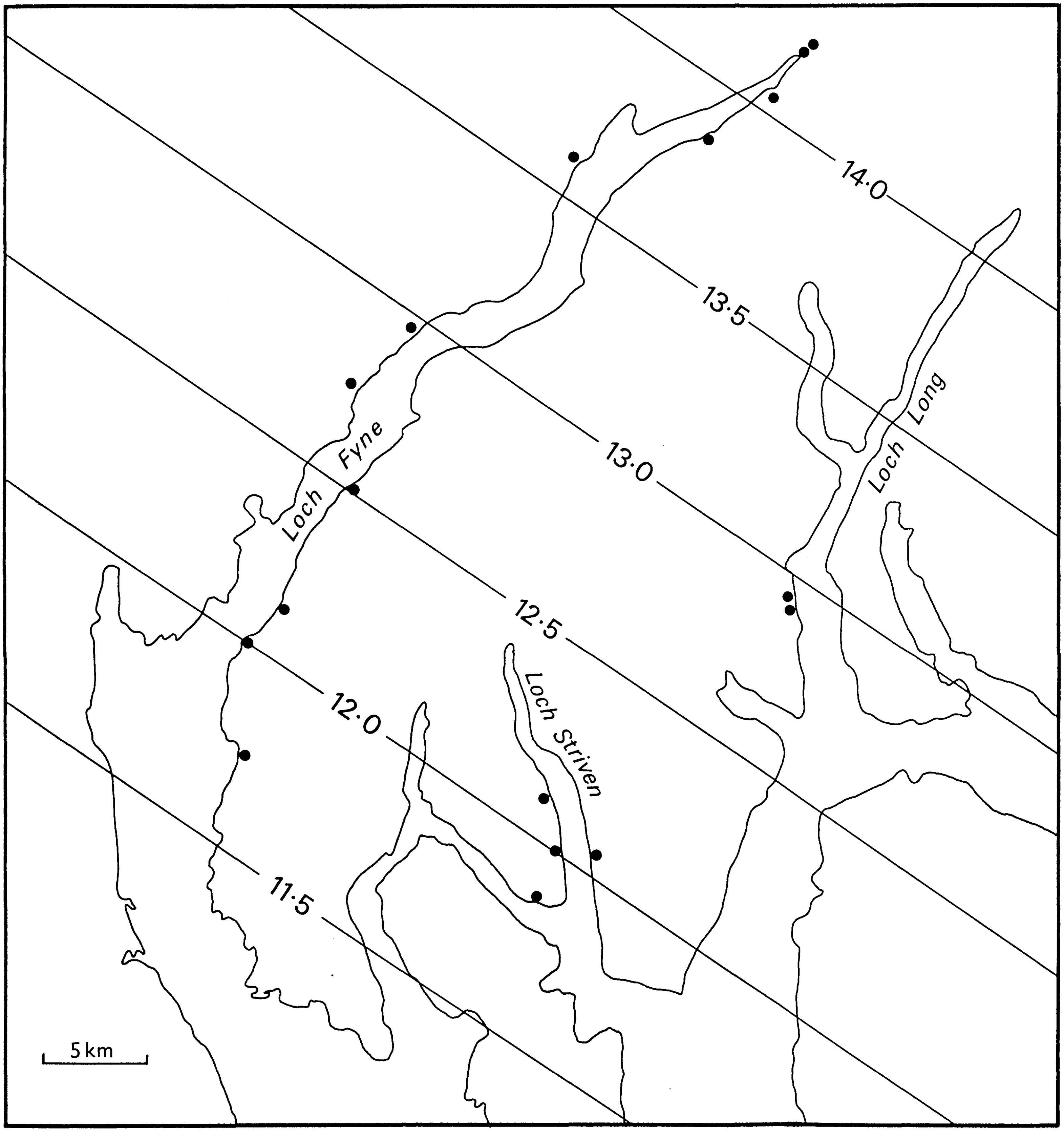
Shoreline	CF1	CF2	CF3	CF4	CF5	CF6
Gradient (m/km)	0.062	0.051	0.030	0.039	0.030	0.025
r	-0.925	-0.813	-0.820	-0.832	-0.604	-0.889

Except for CF3 the shorelines have a lower gradient with lower overall altitude, that is, with decreasing age, as would be expected for a series of isostatically tilted shorelines. The apparent increase in gradient from shoreline CF3 to shoreline CF4 is most likely due to the limitations of the data for CF3, this shoreline only having been traced for a projected distance of ca. 20 km.

The two clearest shorelines on the diagram are CF1 and CF4. The other shorelines can often be distinguished from each other on the basis of distinct breaks of slope occurring between landforms that have been assigned to separate shorelines. Thus, for example, S115 (CF1) is distinct from S116 (CF2), S90 (CF3) from S92 (CF4) and S255 (CF5) from S256 (CF6).

The three-dimensional form of the deformation of the raised shorelines is most clearly illustrated by trend surface analysis. Figure 68 shows the first order trend surface of shoreline CF1. The first order surface only is shown, for the second order surface fails to improve the trend surface at the 95% probability level. The first order surface clearly reveals the south-westerly decline in altitude of

Figure 68: 1st Order trend surface analysis of shoreline CF1. Isobase altitudes in metres O.D.


14·0
13·5
13·0
12·5
12·0
11·5
Loch Fyne
Loch Long
Loch Striven
5 km

the shoreline. The failure of higher order surfaces to provide a better representation of the shoreline may be due to the low number of data points. The line at right angles to the isobases of the first order surface is oriented at S35W.

3. Nature and Age of Flandrian Sediments

As mentioned previously, the Flandrian marine deposits are in places intercalated with peat. Buried peat beds establish the transgressive nature of the uppermost Flandrian shoreline (CF1) and suggest that certain of the lower shorelines may be associated with transgressions. This latter point, however, has not been demonstrated by dating the peats or following the peat beds laterally. The transgressive nature of certain of the Flandrian marine sediments has also been established on the basis of terrace stratigraphies at the head of Loch Riddon (Chapter 6.3).

In order to establish more accurately the timing of the Flandrian changes of sea-level samples from 6 localities were submitted to the Uppsala radiocarbon laboratory. A total of nine radiocarbon assays were carried out on the following samples (fuller descriptions having been given in the sections on particular areas):

(i) Lochgilphead. A 4 cm sample of peat from the peat/carse junction in the area W of Lochgilphead. The base of the sample was at 4.12 m O.D. underlying 2.11 m of peat. Abundant Phragmites suggest the sample dates the beginning of peat growth immediately after the sea had receded. Age: 2,985 $\pm$ 80 yr BP (U-4066).

(ii) Inveraray. A 4 cm sample of compressed leaves and mosses from Inveraray foreshore at the mouth of the River Aray. The sample, at

-1.58 m O.D. was initially thought to date recession of the sea to close to its present level but the age apparently precludes this. Age: 5,335 ± 90 yr BP (U-4067).

(iii) Sandbank. A sample of wood overlying grey silty sand at the base of a 1.4 m thick peat layer. The height of the sample was 6.91 m O.D. and the date is thought to approximate closely the recession of the sea from this altitude. Age: 3,800 ± 100 yr BP (U-2562).

(iv) Kilfinan. Two 2 cm thick samples were taken from the top and bottom of a peat layer buried by sediments of the uppermost Flandrian shoreline. The upper sample, at 9.25 m O.D. gave an age of 7,630 ± 115 yr BP (U-4068) whilst the lower sample, at 8.99 m O.D., gave an age of 6,170 ± 80 yr BP (U-4062).

(v) Furnace. Two 2 cm thick samples from the top and base of a buried peat layer overlain by sediments of the uppermost Flandrian shoreline were submitted. The top samples, at 11.03 m O.D. gave an age of 7,290 ± 90 yr BP (U-4065). The bottom sample, at 10.66 m O.D., was assayed twice, the second time in a larger counter which allowed greater precision. The first assay gave an age of 7,810 ± 200 yr BP (U-4031) and the second assay an age of 8,035 ± 85 yr BP (U-4061). The two ages overlap at one standard deviation but the second age is preferred in subsequent discussion because of its greater precision.

(vi) Tighnabruaich. A wispy peat layer in the gravel of shoreline fragment S81 (6.77 m O.D.) occurred at an altitude of 4.65 m O.D. This gave an age of 1,070 ± 60 yr BP (U-4063).

Of these samples two appear in error. Sample U-4062 from the base of the buried peat at Kilfinan is younger in age than the overlying sample (U-4068) the age of which is in general agreement with the samples from Furnace that occupy a similar stratigraphic position. In addition, as is discussed later, the sample age is at variance with a considerable body of information from elsewhere in Scotland on the age of the Flandrian transgression. The date is therefore rejected.

Sample U-4063 from Tighnabruaich also appears much too young. It predates the formation of S81 and hence should be older than U-4066 from Lochgilphead and U-2562 from Sandbank. Furthermore to accept the date implies a transgression of ca. 2 m followed by a regression of ca. 4 m during the last 1,100 years. The date is therefore rejected.

Of the other dates sample U-4067 from Inveraray does not appear to relate at all closely to a particular sea-level and was discussed in Chapter 7.7. The remaining dates have been plotted on the height-distance diagram (Figure 69) which indicates that shoreline CF1 was formed after 7,290 $\pm$ 90 yr BP, that shoreline CF4 was formed slightly before 3,800 $\pm$ 100 yr BP and that shoreline CF5 formed slightly before 2,985 $\pm$ 80 yr BP.

The transgressive/regressive nature of the contacts between the terrestrial and marine sediments and the dating of the events can best be illustrated by a graph of age against sea-level altitude. In order to construct such a graph it is necessary to compensate for isostatic deformation and the altitudes of the various samples have been standardized to kilometre 58 on the height-distance diagram, the same point as used for the Lateglacial age-height graph. An approximation to the isostatic component for the early Flandrian buried peats at Kilfinan and Furnace was calculated by assuming that the sea receded below the level of the bottom of the peat at each locality at the same time. This indicated tilting between the two sites of ca. 0.08 m/km, a reasonable figure considering the gradient of shoreline CFl (0.062 m/km). This figure was therefore used to standardise the altitudes of the peats to km 58.

Figure 70 illustrates the transgression or regression contacts of the various samples in addition to the altitude of shoreline CFl, the maximum altitude at km 58 reached by the Flandrian transgression. As no pollen or diatom studies have been carried out it is not possible to say how closely the various samples relate to sea-level. Peat growth at the regression contacts could have been delayed for an unknown period subsequent to the drop of sea-level and the transgressing sea may have eroded part of the top of the peat lenses. For these reasons the regression dates are minima and the transgression dates maxima for their respective events.

Despite these limitations it is clear that the Flandrian transgression is relatively closely dated by the three samples from Kilfinan and Furnace: ca. 2 m of the transgression to shoreline CFl has been accomplished since ca. 7,800 yr BP, this date being approximately the mean of the regression sample U-4061 and the lowermost transgression sample U-4068.

4. Comparison with other areas

The only published Flandrian height-distance diagram for the W coast of Scotland constructed by techniques similar to those used in this study is that of Gray (1974b). This diagram, utilising data from the Firth of Lorne and eastern Mull, shows 5 separate Flandrian shorelines ranging in gradient from 0.05 m/km to nearly horizontal and occurring within a height range similar to that of the present study. As in this study not all the shorelines were developed with equal clarity in all parts of the diagram.

The proximity of the present study area to the Firth of Lorne makes it very probable that the uppermost Flandrian shoreline is the same age in both areas. The similarity of the gradients of the shoreline (0.05 m/km in Lorne, 0.06 m/km in this study) supports this correlation. Of the other four shorelines on Gray's diagram two were well developed throughout the area studied, one at ca. 8 m and the other at ca. 4m. It seems probable that shorelines CF4 and CF6 of the present study correlate with these shorelines as they are also widely developed and occur at similar altitudes. The gradients for these shorelines in Lorne were both ca. 0.01 m/km which contrasts with ca. 0.04 m/km and 0.03 m/km for shorelines CF4 and CF6 respectively, but due to variability in his data Gray considered that the calculated gradients of these lower shorelines did not closely reflect their true gradients.

If these correlations with the Firth of Lorne area are correct then there appears to be an extra shoreline, either CF2 or CF3 in the present study area. It is most probably CF3 for CF2 overlaps in part with the range of altitude for Gray's shoreline PS2. There seems little doubt as to the reality of CF3 although it is only developed in part of the study area and it may be that this shoreline was not distinguished in Lorne and eastern Mull because its altitude coincides to a large extent with the Main Rock Platform and associated cliffline.

A number of studies have been carried out on the E coast of Scotland

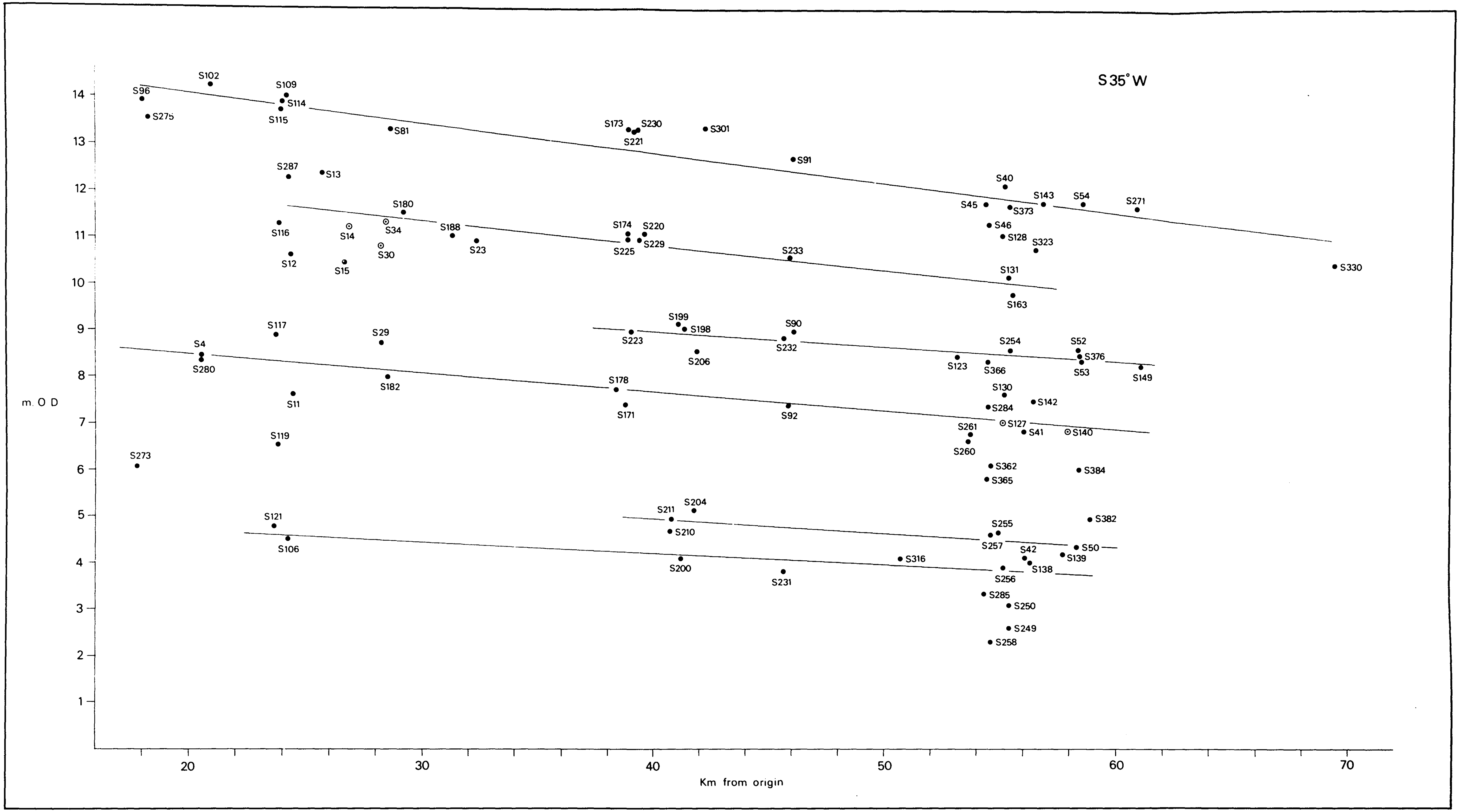
(Sissons _et_ _al_., 1966; Smith, 1968; Cullingford, 1972) using similar techniques to those used in this study. Gray (1974b) correlated his uppermost Flandrian shoreline with the Main Postglacial Shoreline established in these areas and radiocarbon dated to _ca_. 6,700 - 6,500 yr BP (Sissons and Brooks, 1971; Cullingford _et_ _al_., 1980; Smith _et_ _al_., 1980) a correlation that is suggested by the radiocarbon dates on the transgression prior to the formation of shoreline CF1 in this study. The gradient of the Main Postglacial Shoreline is greater on the E coast than in the study area: 0.076 m/km for the Forth valley (Sissons _et_ _al_., 1966) and 0.09 m/km for the Tay valley (Cullingford, 1972). This difference is similar to the relative magnitudes of the gradients for other correlated shorelines: 0.33 m/km for CLG2 compares with 0.43 m/km for the Main Perth Shoreline (Chapter 9) while 0.13 m/km for the Main Rock Platform compares with 0.17 m/km for the Main Lateglacial Shoreline (Chapter 8).

Elsewhere in the Firth of Clyde and its former extensions numerous radiocarbon-dated samples of peat overlying and underlying marine sediments have been reported (Jardine, 1962; 1964; 1971; 1975; Welin _et_ _al_., 1975; Harkness and Wilson, 1979). These samples generally indicate a marine transgression at _ca_. 9000 - 8000 yr BP and a subsequent regression after _ca_. 6000 yr BP, but few of the samples dated have been closely related to sea-level by pollen or diatom studies and it is not known how closely many of the samples date changes of sea-level.

One locality of more detailed study, however, is Loch Lomond, where a marine transgression has been identified in cores removed from the bed of the loch (Dickson _et_ _al_., 1978). The surface of Loch Lomond today is at _ca_. 8 m O.D. and the marine transgression has been radiocarbon dated to between 6,900 yr BP and 5,500 yr BP. The outlet of Loch Lomond is not far from the present study area and the position of the outlet has been projected on to the height-distance diagram and then standardized in altitude to km 58 to account for isostatic uplift, thus allowing comparison with the dates from this study. The relevant data are shown on Figure 70. Relative to the dates from this study the Loch Lomond transgression and regression altitudes appear to plot several metres too low. Part of this discrepancy could be due to variations in isostatic uplift but the magnitude of the difference (ca. 4 m) is such that this is unlikely to be the main factor involved. Rather, it is suggested, the level of the threshold of Loch Lomond, which is in drift, has been lowered by ca. 4 m since the time of the marine invasion, that is during the subsequent regression to present sea-level. An apparent consequence of this argument is that the lowest shoreline in Loch Lomond at ca. 9 m O.D. (Dickson et al., 1978) should be a fresh-water shoreline, formed during the downcutting of the threshold, a hypothesis testable by diatom analysis.

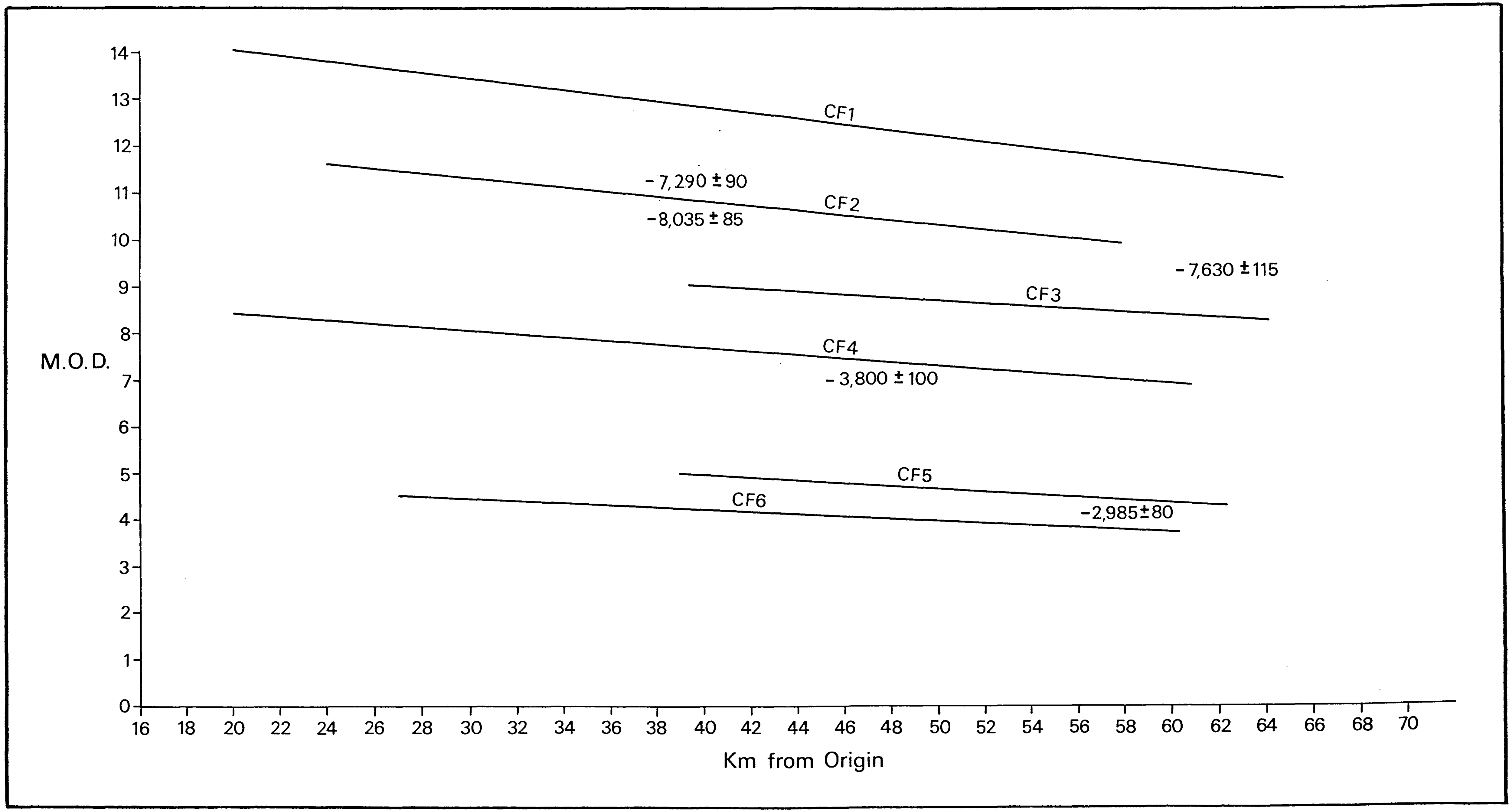
The most detailed analyses in Scotland, involving radiocarbon dating, pollen and diatom analysis as well as stratigraphic work is once again on the E coast (Sissons, 1966; Newey, 1966; Sissons, 1972; Sissons and Brooks, 1971; Smith et al., 1980; Cullingford et al., 1980). In these areas the early Flandrian sea-level history has been worked out as regression (punctuated by minor transgressions) from prior to 9,600 yr BP till ca. 8,300 yr BP followed by the major Flandrian transgression that culminated ca. 6,700 - 6,500 yr BP. Thereafter relative sea-level fell, perhaps being interrupted by minor transgressions, to its present level. In broad outline this is similar to the sequence of events that the radiocarbon dates indicate for the present study area although it is clear that much work needs done on the Flandrian sediments of the study area before a detailed comparison can be made with the E coast.

Figure 66: Height-distance diagram of selected Flandrian shoreline features, S35W.



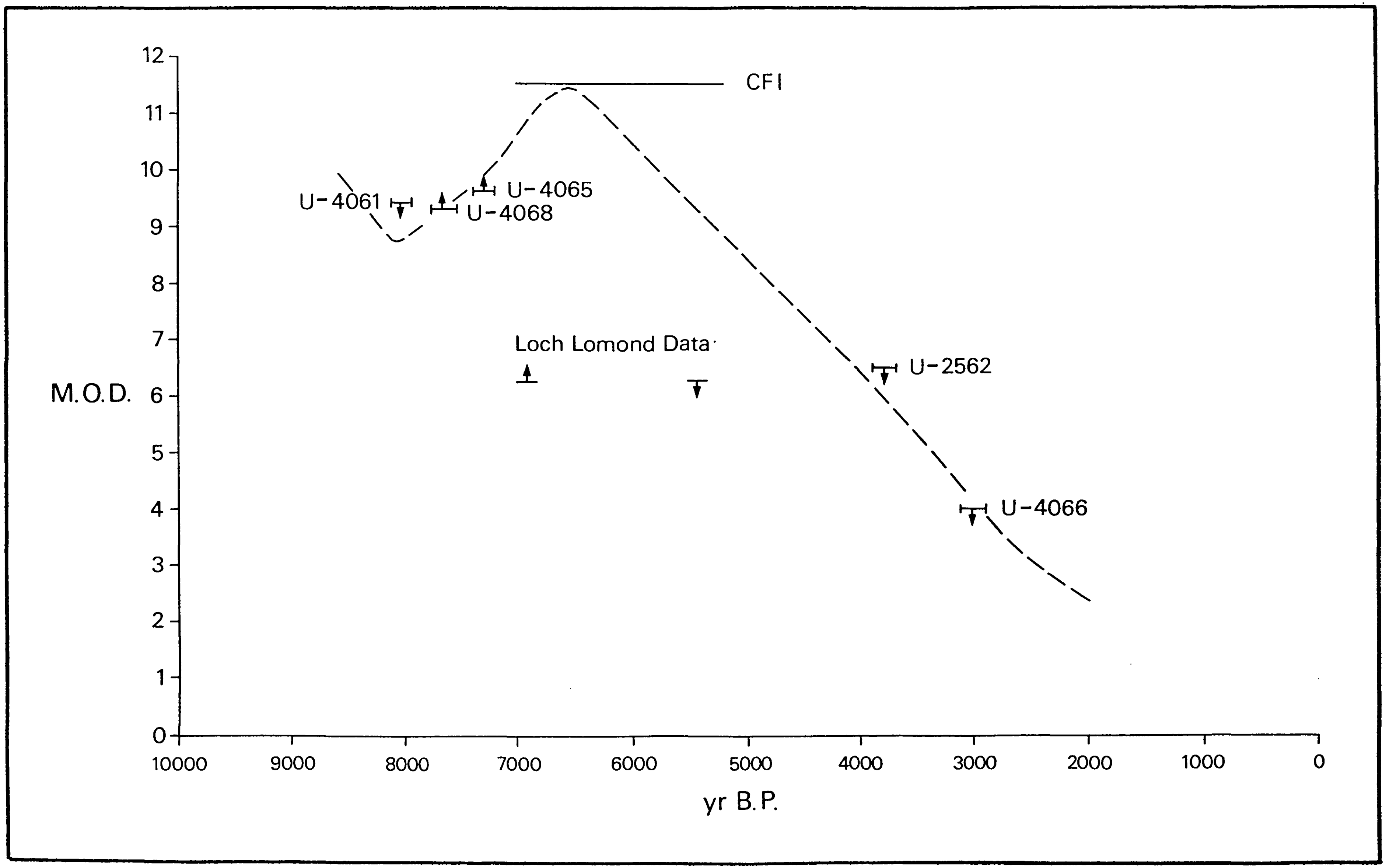
S 35° W
m. O D
Km from origin
14
13
12
11
10
9
8
7
6
5
4
3
2
1
20
30
40
50
60
70
S96
S102
S275
S109
S114
S115
S81
S173
S230
S221
S301
S91
S287
S13
S40
S143
S54
S271
S45
S373
S180
S14
S34
S188
S23
S116
S30
S12
S15
S174
S220
S229
S225
S233
S46
S128
S323
S330
S131
S163
S117
S29
S199
S198
S90
S223
S232
S206
S254
S52
S376
S53
S149
S123
S366
S4
S280
S182
S11
S178
S171
S92
S130
S142
S284
S127
S261
S41
S140
S260
S119
S273
S362
S365
S384
S204
S211
S210
S121
S106
S255
S257
S382
S50
S42
S139
S138
S200
S316
S231
S256
S285
S250
S249
S258

Figure 69: Flandrian shorelines, S35W, and associated radiocarbon dates.



M.O.D.
14
13
12
11
10
9
8
7
6
5
4
3
2
1
0
CF1
CF2
CF3
CF4
CF5
CF6
-7,290 ± 90
-8,035 ± 85
-7,630 ± 115
-3,800 ± 100
-2,985 ± 80
16 18 20 22 24 26 28 30 32 34 36 38 40 42 44 46 48 50 52 54 56 58 60 62 64 66 68 70
Km from Origin

Figure 70: Flandrian sea-level curve.



CFI
U-4061
U-4068
U-4065
Loch Lomond Data
U-2562
U-4066
M.O.D.
0
1
2
3
4
5
6
7
8
9
10
11
12
10000
9000
8000
7000
6000
5000
4000
3000
2000
1000
0
yr B.P.

CHAPTER 11

GLACIAL STAGES

1. Introduction

During the course of this thesis evidence has been presented relating the changes of sea-level during deglaciation of the study area to various ice frontal positions. Two principal glacial stages may be recognised during ice decay, the Otter Ferry Stage related to shoreline CLG2, and the Loch Lomond Readvance. In the first section of the present chapter it is proposed to examine the nature, possible causes and possible correlations of the Otter Ferry Stage with other Scottish glacial stages. In the second part of the chapter details of the Loch Lomond Readvance are given and the relationship to sea-level and certain palaeoclimatic inferences for the time are made.

2. Otter Ferry Stage

In Chapters 5, 6 and 7 evidence has been presented for a halt during the retreat of the last ice sheet to cover the study area. At this time the ice margin lay at Otter Ferry in Loch Fyne, at the mouth of Glendaruel, inland of the head of Loch Striven and at unspecified localities in lower Strath Eachaig and near the mouth of Loch Long. A major shoreline associated with these ice limits (CLG2) correlates the ice frontal positions with various other bodies of stagnant ice scattered throughout southern Cowal. The halt of the ice has been tentatively dated to 12,900 $\pm$ 200 yr BP and the ice frontal positions appear to have been maintained for perhaps 200 - 400 years, during which period sea-level fell by *ca*. 20-25 m.

It may be argued that a still-stand of the ice front for a period of several hundred years is rather unlikely as it implies a unique mass-balance relationship to have lasted for this period of time (Sissons and Dawson, in press). It is more probable that the ice front readvanced to the positions that have been identified as the ice margins at this time. There is no field evidence to support the readvance possibility, however. It can also be argued that the best developed shoreline associated with the stage (CLG2) most probably would develop at the culmination of a readvance because of increased sediment supply at such a time. Shoreline CLG2, however, is the highest shoreline and the drop in sea-level, from which is inferred the approximate duration of the Stage, occurred after its formation, again implying a stationary ice front for this period.

The timing of the Otter Ferry Stage raises two problems: (a) that the period of still-stand coincided with a mild climatic interval, the Bølling (Mangerud _et al_., 1974; Pennington, 1975); and (b) that it implies the presence of a considerable mass of ice in the SW Highlands at a time when much of the rest of Scotland has been argued to have been ice free (Sissons, 1976a; p 90). These problems are considered in turn.

(a) The Bølling mild period has been recognised widely in NW Europe by pollen analysts (Mangerud _et al_., 1974) but remains a source of some debate as to its representation on pollen diagrams from the British Lateglacial (Gray and Lowe, 1977). Studies of fossil beetles, however, indicate that by _ca_. 13,000 yr BP the climate of the Midlands of England was as mild as that of today (Coope, 1977). In addition, study of deep-sea cores from the NE Atlantic indicates that polar waters had retreated from the British coasts _ca_. 13,500 yr BP (Ruddiman and McIntyre, 1973), although the accuracy of the dating of this event, relying as it does on a limited number of ^{14}C dates on deep-sea carbonates and on extrapolation assuming linear sedimentation rates must be in question. Although the details may not be clear there is little disagreement that in general the period from ca. 13,000 yr BP for several hundred years afterwards was one of mild climate. Certainly it does not coincide with a marked cold period.

It is notable that much of the retreat of the last ice-sheet to cover the British Isles was accompanied by a cold climate, polar waters being present off the Atlantic coasts while this ice retreated from the Midlands of England to the Central Lowlands of Scotland. Such a retreat implies precipitation starvation and the retreat would have continued until the mass-balance of the reduced ice-sheet came into equilibrium with the reduced precipitation. It is suggested, however, that the retreat of the polar waters from the British coasts resulted in increased precipitation in the milder conditions of the time and this regenerated the remaining ice mass in the Scottish Highlands resulting in the Otter Ferry Stage.

The size of the glaciers present at this time implies a considerable mass of high land/ice in the SW Highlands and even with a climate only perhaps slightly less mild than today this suggests considerable snowfall in the SW Highlands. The ice mass would also have acted to cool the local climate hence increasing the total snowfall.

During the period of ice front stability sea-level fell. This would act as a positive factor in the mass-balance of the glaciers involved reducing the amount of calving and tending to prolong the period of stable ice margins. The actual magnitude of this effect would vary from glacier to glacier, however, due to variations in the proportions of the snouts of the glaciers that were subject to calving.

(b) In recent years a number of radiocarbon dates from the bottom of lakes or kettle holes from a variety of localities in the Scottish Highlands have given ages from 12,500 to over 13,000 yr BP. These dates are minimal for deglaciation and since certain of the sites are deep within the Highlands (e.g. Loch Etteridge, Loch Droma) it has been inferred that much of the last ice-sheet to cover the Highlands had disappeared by 13,000 yr BP (Sissons and Walker, 1974). The accuracy of such basal radiocarbon dates has recently been assessed (Sutherland, 1980) and it was concluded that of the numerous sources of error likely to affect such dates those that resulted in an ageing of the sample were the more probable. In addition, the overall assessment of the dates from the Scottish Highlands is complicated by the fact that not one of the dates is cross-checked by other dates in close proximity in the same stratigraphic column and not one of the dates is related by means of a relative chronology to any other of the dates. They are effectively spot dates and this contrasts strongly with the chronology built up with marine shell dates in the SW Highlands (Appendix 1) for not only are many of these shell dates stratigraphically related to other dates but they are also related to a framework of relative chronology and many of the samples have had more than one fraction dated such that further confidence can be placed on their accuracy.

It is concluded that the chronology of deglaciation based on radiocarbon dates from the bottom sediments of lakes and kettle holes is not well supported and that there are good reasons for favouring the marine shell chronology established for the SW Highlands. The evidence then suggests a relatively large mass of ice existing in the SW Highlands after 13,000 yr BP.

(c) Correlations. In the valley between Ford and Kilmartin, Gray and Sutherland (1977) have described a series of terraces commencing near Ford at the S end of Loch Awe and relating to a sea-level of ca. 35 m O.D.. Meltwaters continued to use this channel and extensive outwash was deposited at its mouth as sea-level fell to below 24 m O.D.. The outwash terraces cannot be linked to any ice marginal features due to the subsequent development of an extensive set of lower terraces related to sea-level below that of the uppermost Flandrian shoreline. The use of the Kilmartin valley by meltwater implies, however, the presence of ice damming the present exit of Loch Awe at the Pass of Brander. The drop in sea-level of at least 11 m allied to the presence of ice in Loch Awe suggests strongly a correlation with the drop in sea-level at Otter Ferry, only 15 km distant across country.

Farther afield, a marked drop of 20 m has been noted in the marine limit at Stirling by Sissons and Smith (1965). This drop in sea-level was preceded by the formation of two distinct shorelines, the Main Perth Shoreline (gradient 0.43 m/km) that related to an ice front to the E of Stirling and the lower Perth Shoreline (gradient ca. 0.43 m/km) that related to an ice front at Stirling and that formed after the Main Perth Shoreline. The similarity in gradient of these two shorelines suggests that they were both formed in a short time interval. The Main Perth Shoreline is the most distinct Lateglacial raised shoreline in SE Scotland and has been traced beyond the Firth of Forth into the Tay and Earn valleys (Cullingford, 1972, 1977; Armstrong et al., 1975) and possibly farther NE (Cullingford and Smith, 1980). It was originally thought to have correlated with the ice limits of the Perth Readvance (Sissons, 1963b, 1964) though more recent evidence (Paterson, 1974;

Browne, 1980) has shown this not to be the case and the concept of the Perth Readvance has now been rejected (Sissons, 1976a). Whatever the exact relationship of the Main Perth Shoreline to ice limits in the Tay-Earn area, the evidence at Stirling strongly suggests a correlation with the Otter Ferry Stage and the associated shoreline CLG2. Not only is there a marked drop in the marine limit but the gradient of the shorelines (the Main Perth and CLG2) that preceded this drop are not greatly dissimilar (0.43 m/km and 0.32 m/km). The difference in shoreline gradients is explicable in terms of the relative positions of the two areas with respect to the ellipsoid of isostatic uplift and is similar to the relative magnitudes of the gradients of other shorelines that may be correlated between the two areas (Chapter 10). Since the ice mass that reached to the site of Stirling originated in the SW Highlands it may be expected to show a change in retreat rate similar to other glaciers that originated in that area.

Correlation with other glacial retreat stages elsewhere in Scotland is much less certain. Recently Robinson and Ballantyne (1979) have described a retreat stage in Wester Ross that is marked for a considerable portion of its known limit by a clear end moraine. Termed the Wester Ross Readvance the limit crosses the present coast at a number of localities. Sissons and Dawson (in press) have investigated the relationship of this limit to sea-level and have discovered no drop in the marine limit at the moraine. As the ice retreated from the moraine, however, relative sea-level fell. Sissons and Dawson provide such shoreline information as could be established and they suggest that outside the readvance limit the marine limit is formed by one shoreline with a gradient of between 0.33 and 0.39 m/km. This shoreline gradient is suggestive of a correlation with the Otter Ferry Stage but the absence of a drop in the marine limit at the ice margin does not support such a correlation. Until more accurate information on the relative or absolute dating of the Wester Ross Readvance is available it is not possible to correlate it with the Otter Ferry Stage.

A readvance moraine has been proposed in the Beauly Firth at Ardersier (Smith, 1977) and Smith stated (p 75) that there is a significant drop in the marine limit W of this moraine. Smith's shoreline heights do not reveal this drop, however, and Synge (1977b) does not show a drop in the marine limit at Ardersier. Elsewhere in the Beauly Firth Synge indicates a drop of _ca_. 8 m in the marine limit at Kessock and also a large fall in sea-level near Englishton. Whilst the drops in the marine limit are suggestive of possible correlations with the Otter Ferry Stage (cf. Sissons, 1976a) the details of the associated shorelines and glacial deposits are not clear. Any proposed correlation between this area and the Otter Ferry Stage must therefore remain uncertain.

3. Loch Lomond Readvance

It has long been realised that in the Cowal Peninsula, as elsewhere in the Scottish Highlands, the decay of the last ice-sheet was followed by a period of valley glaciation. McLaren (1855), Bell (1874), Anderson (1894) and Gunn _et al_. (1897) all explained (with varying degrees of assurance) certain of the landforms of the Cowal Peninsula in terms of this later glaciation. During the first half of this century nothing was published on the glacial deposits of the study area that furthered the idea of this later period of glaciation but in 1933 J.B. Simpson published an important paper on the results of his work in the area to the E in which he established on the basis of transported marine shells, that this later glaciation was a major readvance, termed the Loch Lomond Readvance as it was perhaps most clearly represented in the deposits occurring to the S of Loch Lomond. The existence of the volume of ice in Loch Lomond that Simpson's evidence led him to postulate implied a considerable mass of ice in the northern parts, at least, of the study area, a conclusion that was reinforced by the work of Anderson (1949) who correlated the readvance moraine at Rhu in the Gareloch (McCallien, 1936) with the Loch Lomond Readvance. The ice that occupied the Gareloch must have flowed over the col from Loch Long (the distribution of hill masses and the lateral moraine on the E of the Gareloch precludes any other possibility) and hence much of Loch Long must have been occupied by ice at this time.

In recent years considerable attention has been paid to the Loch Lomond Readvance, Donner (1957) establishing on the basis of pollen analyses of sediments occurring outside and inside the limits of the former glaciers that the readvance was equivalent in age to the Younger Dryas Stadial of NW Europe. Radiocarbon dating of the shells contained in a number of deposits particularly around the SW Highlands has confirmed Donner's conclusion (Sissons, 1967c; Peacock, 1971; Gray and Brooks, 1972; Otlet and Walker, 1979). Studies in the Forth Valley (Sissons, 1966) and in Mull and the Firth of Lorne (Gray, 1975c) demonstrated that the Loch Lomond Readvance glaciers in these areas were related to sea-levels below that of the uppermost Flandrian shorelines. As the present study area lies between the areas just mentioned and as glaciers were known to extend to sea-level (on the basis of Anderson's evidence) in certain parts of the study area, any consideration of sea-level changes must take into account the distribution of Loch Lomond Readvance ice. During the course of this present study, therefore, an attempt was made to delimit the Loch Lomond Readvance glaciers where they occurred within the sea-lochs. The limits of the Readvance glaciers were also mapped in certain valleys away from the coast and on aerial photographs in the mountains to the NE of the study area. Much of the interior, however, has not been examined in the field and there are therefore considerable uncertainties as to the overall distribution of the Readvance glaciers. The following section is therefore primarily concerned with the extent of the Loch Lomond Readvance glaciers in the sea lochs. Evidence for the existence of a number of small glaciers in the S of the Cowal Peninsula is presented and the upper limit of glaciation on the mountains in the NE of the study area is also discussed. The relationship of the glaciers to sea-level and certain palaeoclimatic inferences are considered at the end of the section.

(a) The Sea Lochs

Loch Long. The lateral moraine to the E of the Gare Loch (Anderson, 1949) has been traced northwards from the head of Glen Fruin to where it overlooks Loch Long at ca. 350 m (Fig. 71), thus proving the ice in the Gare Loch to have originated in Loch Long. A radiocarbon date of 11,120 ± 250 yr BP from shells in the Rhu moraine (Appendix 1) indicates that this period of glaciation is indeed the Loch Lomond Readvance. Between the Gare Loch and Loch Long the ground is largely bare of glacial deposits or covered in peat and no similar moraine to that on the E of the Gare Loch has yet been mapped. The mounds on the col W of Mamore, however, may represent part of the moraine.

Along the slopes above Loch Long little evidence can be found to level changes must take into account the distribution of Loch Lomond Readvance ice. During the course of this present study, therefore, an attempt was made to delimit the Loch Lomond Readvance glaciers where they occurred within the sea-lochs. The limits of the Readvance glaciers were also mapped in certain valleys away from the coast and on aerial photographs in the mountains to the NE of the study area. Much of the interior, however, has not been examined in the field and there are therefore considerable uncertainties as to the overall distribution of the Readvance glaciers. The following section is therefore primarily concerned with the extent of the Loch Lomond Readvance glaciers in the sea lochs. Evidence for the existence of a number of small glaciers in the S of the Cowal Peninsula is presented and the upper limit of glaciation on the mountains in the NE of the study area is also discussed. The relationship of the glaciers to sea-level and certain palaeoclimatic inferences are considered at the end of the section.

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Along the slopes above Loch Long little evidence can be found to the E of the loch, partly because a large part of the area has been incorporated into the Clyde Submarine Base at Coulport. On the W side of the loch, however, near Ardentinny an area of hummocky moraine was described in Chapter 5.3. At Ardentinny the marine limit is only ca. 13.5 m O.D., the height of the uppermost Flandrian deposits, whilst only a short distance downloch the marine limit occurs at ca. 41 m O.D. and was formed at ca. 13,000 yr BP (Chapter 9). The Loch Lomond Readvance glacier cannot have advanced beyond the high Lateglacial raised shorelines and its limit is placed slightly downloch of Ardentinny where the hummocky moraines terminate. No end moraine, however, has been observed here. Locating the margin of the Loch Lomond Readvance glacier near Ardentinny means that the ice extended farther down the Gare Loch than down Loch Long, presumably due to the greater depth of Loch Long and the greater consequent ice loss by calving.

Evidence is poor farther up Loch Long due partly to the extensive landslipping and to the forest cover of many of the slopes. An important section has been recorded, however, near Cormonachan by Loch Goil. At this site (Fig. 71) ca. 2 m of laminated grey clays were overlain by ca. 90 cm of dark grey shelly sand. This sand was succeeded upwards firstly by a lens of coarse sand and gravel and then by a smaller lens of grey clay. These beds were truncated by a light fawn coloured unsorted layer with boulder-size material held in a rather loose sand and gravel matrix, interpreted as a till. This was succeeded by beach gravels and the whole section overlain by peat. Shells from the dark grey sand gave a radiocarbon date of 12,260 $\pm$ 150 yr BP (T-1456) (Appendix 1), proving the till to post-date this time and hence belong to the Loch Lomond Readvance. It is also of note that although Loch Goil was deglaciated ca. 12,300 yr BP when relative sea-level would have been ca. 22 m O.D. (Chapter 9) no shorelines have been found above the level of the uppermost Flandrian shoreline around the loch (i.e. ca. 13.5 - 14.0 m O.D.).

Loch Fyne. In Chapter 7.4 an ice limit was described in Strathlachlan by Loch Fyne. A series of hummocky moraines terminate in outwash terraces and a meltwater channel related to these deposits was traced across the watershed towards Loch Fyne proving ice to have existed contemporaneously in Loch Fyne. In a similar manner near Furnace (Chapter 7.8) on the opposite side of Loch Fyne an ice limit was defined by the termination of hummocky moraines in the Auchindrain valley. This ice limit was also contemporaneous with ice in Loch Fyne and it was argued that the drainage from this ice margin was diverted into a series of meltwater channels by the ice in Loch Fyne around Furnace. Near Furnace a section of till containing marine shell fragments was also described indicating a readvance origin for the last ice to occupy this area. No terminal ice features, however, were observed beside Loch Fyne. It is notable that at Furnace and farther up-loch raised shorelines have not been found above the level of the upper Flandrian shoreline deposits, whilst downloch of Furnace, commencing first at Minard, such raised shorelines exist. This evidence combines to suggest that the Loch Lomond Readvance limit in Loch Fyne was in the vicinity of Furnace, perhaps slightly downloch of the village, and that the ice at this locality was contemporaneous with the two ice termini located in the immediately adjacent side valleys of Strathlachlan and Auchindrain. Other areas of shelly till have been observed uploch of Furnace (Fig. 71) both by Gunn et al. (1897) and during the present study. If it is assumed that the till was deposited during the Loch Lomond Readvance then the shelly till exposures indicate deglaciation at least to the head of the sea-loch prior to the readvance, a conclusion in part supported by the occurrence of remnants of the Main Rock Platform and associated cliff near the head of Loch Fyne.

(b) Upper Limits in NE of Study Area

The mountains of the NE of the study area were clearly a major source of ice for the glaciers during the Loch Lomond Stadial. Although the information currently available is rather patchy, sufficient is known to indicate the general form of the upper surface of the ice mass.

Between the heads of Loch Long and Loch Fyne (Fig. 71) the summit of Ben Ime (1011 m) is extensively periglaciated. Periglacial features can be followed down on to the col to the north (ca. 750 m) indicating this to be a local maximal altitude for the ice surface. A drift limit on the NW flank of Ben Arthur lies higher than the col to the SE of Ben Ime (ca. 620 m) suggesting this col to have been overtopped by ice.

The NE spur of Binnein an Fhidhleir has drift limits on both sides at ca. 750 m and the summit of Ben Vorlich (943 m) above Loch Sloy has extensive periglacial deposits that reach down to ca. 800 m. Ben Bhuidhe (948 m) to the W of Glen Fyne also has extensive periglacial deposits that reach down to ca. 800 m.

Although it may be argued that altitudinal limits to periglacial features are a function of local climate it has been demonstrated elsewhere in Scotland (Sissons, 1974b; 1976a) that extensive well-developed periglacial features occur only outside the limits of the Loch Lomond Readvance glaciers, irrespective of altitude. This is thought to be the case in the SW Highlands and the above observations, though scattered, when taken together suggest an ice surface in this area that slopes northwards from slightly below 750 m near the Rest and be Thankful pass to around 800 m over the plateau area to the N. The higher mountains stood out above this ice cover as nunataks and detailed mapping around these mountains and those to the S could be expected to define the ice surface in better detail than has been possible here.

(c) Glaciers in S Cowal

In the S of the Cowal Peninsula, beyond the margins of the larger ice mass that occupied the N and E of the study area during the Loch Lomond Stadial, a number of small glaciers developed in individual valley heads and corries (Fig. 71). Certain of these have been mapped and field checked and they are described here as they give useful information on the climate of the SW Highlands during the Stadial. This is not, however, an exhaustive study of the small glaciers as a variety of other similar evidence has been mapped on aerial photographs but not yet field checked.

(i) Black Craig. The Black Craig Burn is one of the headwater tributaries of the Ardyne Burn. Along the eastern side of its valley a lateral moraine can be traced to near the Ardyne Burn where the moraine turns to cross the valley floor. On the W side of the Black Craig Burn a drift limit can be followed up the slope to the foot of the Black Craig. A second ridge occurs immediately inside the end moraine where it runs beside the Ardyne Burn. The lowest point reached by this glacier was ca. 180 m and the upper limit in the valley is defined by the topography as ca. 470 m. Although the glacier faces S its initiation would have been favoured by the shade of the Black Craig and its growth aided by the wide plateau areas to the W and N from which snow could have blown on to the glacier surface.

(ii) Corrachaive. In Glen Lean a distinct triple end moraine can be traced across the mouth of Corrachaive Glen. Drift limits can be followed upslope for a short distance on both sides of Corrachaive Glen and at the junction with the Cruach Neuran Burn occurs a small group of hummocky moraines. The snout of the glacier descended to ca. 100 m and the topography suggests that the glacier had an upper limit of ca. 500 m. This N-facing glen would have been suited for glacier development due to its shaded nature.

(iii) Corarisk. In Chapter 5.4 an end moraine was described in Glen Massan forming an arc around the mouth of Corarisk Glen. The upper parts of this glen have not been mapped but if the glacier occupied the whole valley (as is implied by the position of the moraine) it would have had an upper limit of ca. 530 m. Its snout descended to ca. 15 m.

(iv) Stronlonag. Farther up Glen Massan from the Corarisk moraine a double end moraine cuts across the valley bottom and lateral moraines can be traced up the slope to a small corrie on the S side of the valley. Striations oriented downslope were mapped by the Geological Survey at this locality (Gunn et al., 1897). The glacier snout reached to ca. 75 m and the upper part of the back wall was at ca. 460 m.

(v) Creag Mhor. In the Stronchullin Valley to the W of Loch Long a double end moraine has been traced cutting across the valley below Creag Mhor. The bedrock in the upper reaches of the Stronchullin Valley, where seen beneath a cover of peat, is shattered and bedrock surfaces do not carry glacial markings, suggesting no ice cover during the Loch Lomond Stadial. The slopes below Creag Mhor, however, are largely drift covered and this together with the configuration of the moraines suggests that it was only in the shadow of Creag Mhor that a glacier developed.

The end moraines reach down to ca. 150 m and the upper surface of the glacier at the top of Creag Mhor would have been ca. 450 m.

(d) Relationship of Loch Lomond Readvance Glaciers to Sea-Level

In the field descriptions of the limits of the Loch Lomond Readvance glaciers in Loch Long and Loch Fyne it was noted that the shorelines inside these limits did not occur above the altitude of the uppermost Flandrian shoreline. From this it may be inferred that sea-level had dropped below this level prior to the disappearance of the Readvance glaciers. Further, since the raised shorelines show a general increase in gradient with age and since the uppermost Flandrian shoreline (CF1) was formed some 3000 - 4000 years after the end of the Loch Lomond Stadial, it may also be inferred that the relative sea-level at the limits of the glaciers was sufficiently low for any shoreline formed shortly after the disappearance of the glaciers (cf. the Main Buried Beach in the Forth Valley, Sissons, 1966) not to project above the Flandrian Shorelines at the heads of the sea-lochs. Assuming an early Flandrian shoreline gradient of ca. 0.1 m/km (Chapter 10) and extrapolating downloch from a maximum possible shoreline altitude of 14 m at the heads of the sea lochs implies a maximum sea-level at the ice margins of ca. 12 m O.D.

In Chapter 8 the origin of the Main Rock Platform was discussed with reference to Sissons' (1974a) suggestion that this shoreline correlated with the Main Lateglacial Shoreline in SE Scotland and hence was Lateglacial in age. Sissons suggested that the cold climate of the Loch Lomond Stadial was conducive to frost shattering in the intertidal zone and that the development of a rock-erosional feature was due to the particular climate of this period. In Chapter 8 qualified acceptance of this hypothesis was given, for while on the one hand there was evidence that indicated this feature to coincide with a period of Lateglacial marine erosion, the marine sedimentary record failed to provide evidence of the large amounts of angular debris implied by the large volumes of rock eroded to form this feature.

It was also pointed out that the platform, often in a degraded form, was developed around the inner sea-lochs and that part at least of its formation preceded the Loch Lomond Readvance. The Lateglacial sea-level curve (Chapter 9) indicates a period of from ca. 11,500 - ca. 10,500 yr BP prior to and including part of the Loch Lomond Stadial during which marine erosion could have operated at the altitude of the Main Rock Platform. Given the reservations expressed in Chapter 8 as to the erosion of the entire platform during the Lateglacial, such erosion as was accomplished is suggested to date from the period of ice build-up during the Loch Lomond Readvance. Although it is not possible to be precise the above factors suggest an abandonment of the platform by the sea at ca. 10,500 yr BP.

The data in the Chapters on Lateglacial and Flandrian sea-level change on shoreline gradient and relative sea-level movements at km 58 have been combined into two diagrams that cover the whole of the time period considered. The data on the Main Rock Platform and the maximum possible sea-level at the Loch Lomond Readvance limits have also been added. Figure 72 shows the change of shoreline gradient with time and Figure 73 the change of relative sea-level at km 58 with time. There is a variety of sources of inaccuracy in these curves previously discussed in the relevant Chapters but they both suggest an isostatic effect was associated with the Loch Lomond Readvance. Although it is not possible to quantify the effect it seems highly probable that the Loch Lomond Readvance was responsible for renewed downwarping of the earth's crust in the SW Highlands. Such an effect has also been suggested by Sissons (1972) for the Forth Valley.

(e) The Climate of the Loch Lomond Stadial

In recent years considerable advance has been made in the knowledge of the climate during the Loch Lomond Stadial (see Sissons, 1979a for a review). Particular use has been made of the distribution of glaciers at that time (Sissons, 1974b; Sissons and Sutherland, 1976) but it has not been possible to carry out detailed analyses on the Loch Lomond Readvance glaciers in the study area, primarily because all the evidence has not been mapped and secondarily because the available Ordnance Survey map coverage is of too poor a quality. Certain inferences can be made, however, and these can be compared to the results of marine faunal and terrestrial pollen analyses from the SW Highlands for this time period.

The poor map coverage does not allow the calculation of the equilibrium firn line altitudes (Sissons, 1974b) of the small glaciers in the S of the Cowal Peninsula. A first-order approximation, however, may be made by calculating the mid-altitude lines of the glaciers. These are given in Table 28.

TABLE 28. MID-ALTITUDES OF LOCH LOMOND READVANCE GLACIERS

Glacier	Mid-altitude
Black Craig	330 m
Corrachaive	300 m
Corarisk	270 m
Stronlonag	270 m
Creag Mhor	300 m

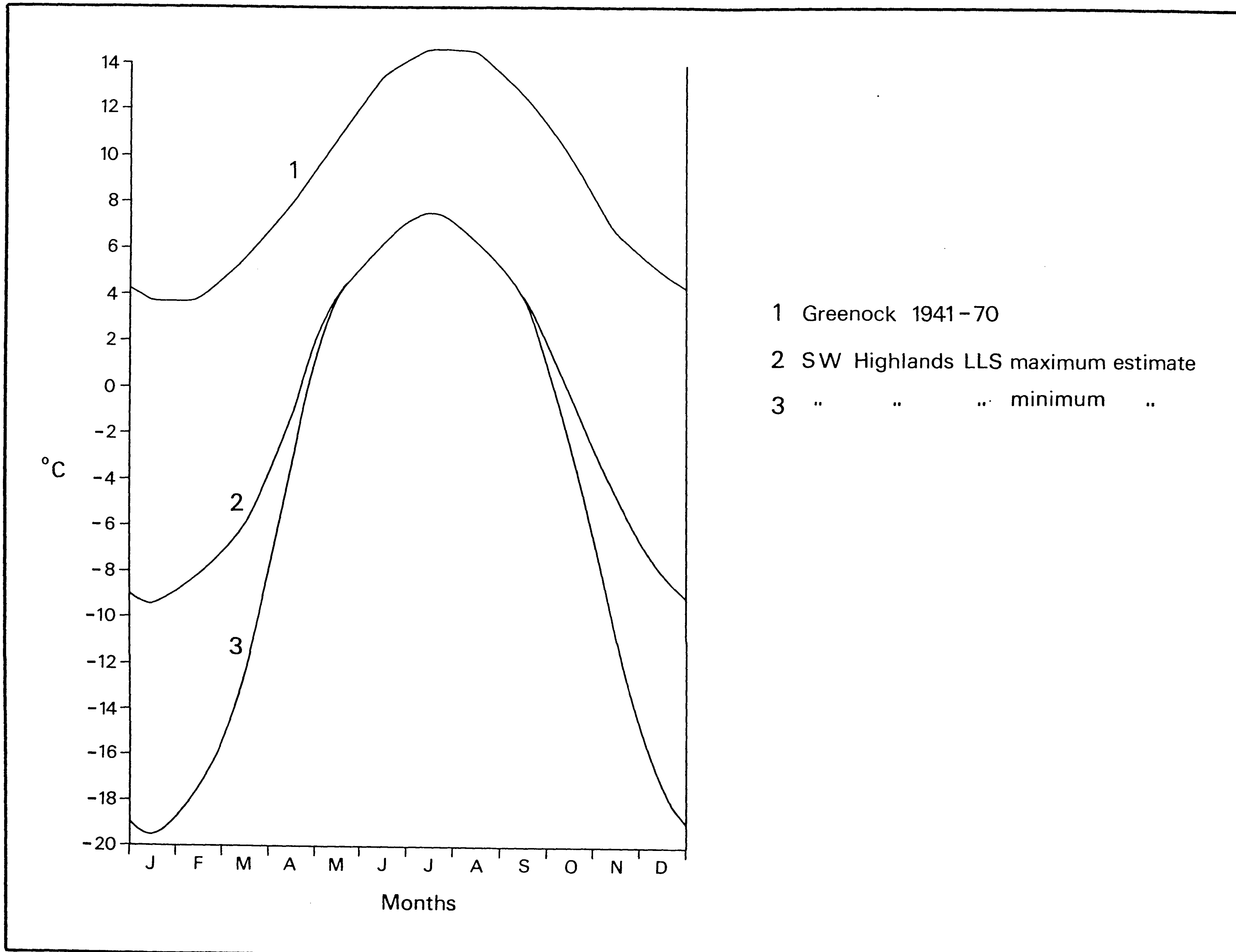
The mid-altitude figures show considerable agreement, suggesting a firn line altitude in this area of ca. 300 m. This figure is comparable with the equilibrium firn line altitude of 250 m for Mull (Gray and Lowe, 1977) and indicates the SW Highlands to have had the lowest equilibrium line altitudes in Scotland during the Loch Lomond Stadial. Although similar calculations cannot be made for the large ice mass occupying the NE of the study area, the size of this ice mass and its descent to sea-level imply a similarly low equilibrium line altitude.

Sissons (1980) argued that the low equilibrium line altitudes and large ice masses in the SW Highlands could not be adequately explained in terms of reduced temperatures but implied considerable precipitation. Sissons did not consider that part of the ice volume could have been a relic of the ice sheet, but given that ice retreated at least to beyond the heads of the sea-lochs prior to the Readvance the contrast in ice volume in the SW Highlands and other mountain areas such as the Cairngorms still appears valid, particularly since calving must have been a major factor in the ablation of the large glaciers that occupied parts of the sea-lochs.

At Old Kilpatrick on the Clyde Estuary to the E of the study area and at Poltalloch near Kilmartin to the W of the study area fossil frost wedges and involutions have been reported in Lateglacial marine deposits (Rose, 1975; Gailey, 1961). As these deposits cannot have been abandoned by the sea till after ca. 13,000 yr BP, the frost wedges and involutions must have formed during the Loch Lomond Stadial. In Alaska frost wedges develop where mean annual temperatures are ca. -6^{o}C or lower (Péwé, 1966) and Williams (1975) has concluded that involutions form along the continuous/discontinuous permafrost boundary. Permafrost itself is indicative of mean annual temperatures of -1^{o}C and if it is accepted that the fossil frost wedges developed only in suitable locations in the SW Highlands during the Loch Lomond Stadial in the area of discontinuous permafrost, $-6^{o}C$ may be taken as a minimum estimate of mean annual temperature for the stadial and $-1^{o}C$ as a maximum estimate. Sissons (1979b), on the assumptions of total precipitation being similar to that of the present and 20-25% of the precipitation being summer rainfall, suggested a mean May-September temperature of $6^{o}C$ and July mean temperature of $7.5^{o}C$ at sea-level in the SW Highlands during the Loch Lomond Stadial, these temperatures being derived from analogy with southern Norway today. These estimates, together with those derived from the periglacial evidence allow construction of a graph of the annual temperature regime for this period (Fig. 74). There is a large temperature range for the winter due to the uncertainties in the periglacial evidence though the presence of the involutions is suggestive that the true regime was nearer the upper estimate. The present-day annual temperature regime for Greenock (HMSO, 1976) is also shown on Figure 74. Clearly it was much colder than at present, July mean temperatures being $7^{o}C$ lower during the stadial and January temperatures at least $14^{o}C$ lower. The greater difference for the winter implies a more continental climate than at present.

The above inferences as to the climate during the Loch Lomond Stadial in the SW Highlands are supported by evidence from the marine record and from pollen analysis of terrestrial deposits. At Ardyne in the S of the study area, Peacock _et al_. (1978) have identified a cold-water (Mid-Arctic) fauna between _ca_. 10,720 yr BP and 10,220 yr BP overlying a more temperate fauna. Few pollen diagrams from the Lateglacial period have been published for sites in the SW Highlands.

Figure 74: Annual temperature regime during the Loch Lomond Stadial.



14
12
10
8
6
4
2
0
-2
-4
-6
-8
-10
-12
-14
-16
-18
-20
°C
1
2
3
J F M A M J J A S O N D
Months
1 Greenock 1941-70
2 SW Highlands LLS maximum estimate
3 " " " minimum "

An exception is that of Rymer (1977) for a site at Drimnagall ca. 14 km WSW of Lochgilphead. Rymer correlates pollen assemblage zone D4 from this site with the Loch Lomond Stadial. The pollen indicates a rather open vegetation and the stratigraphy minerogenic sedimentation. The high percentage of *Salix* cf. *herbacea* Rymer suggests as indicating snow cover during winter. Pollen assemblage zone D4 in general indicates a cooler climate than the preceding pollen assemblage zones.

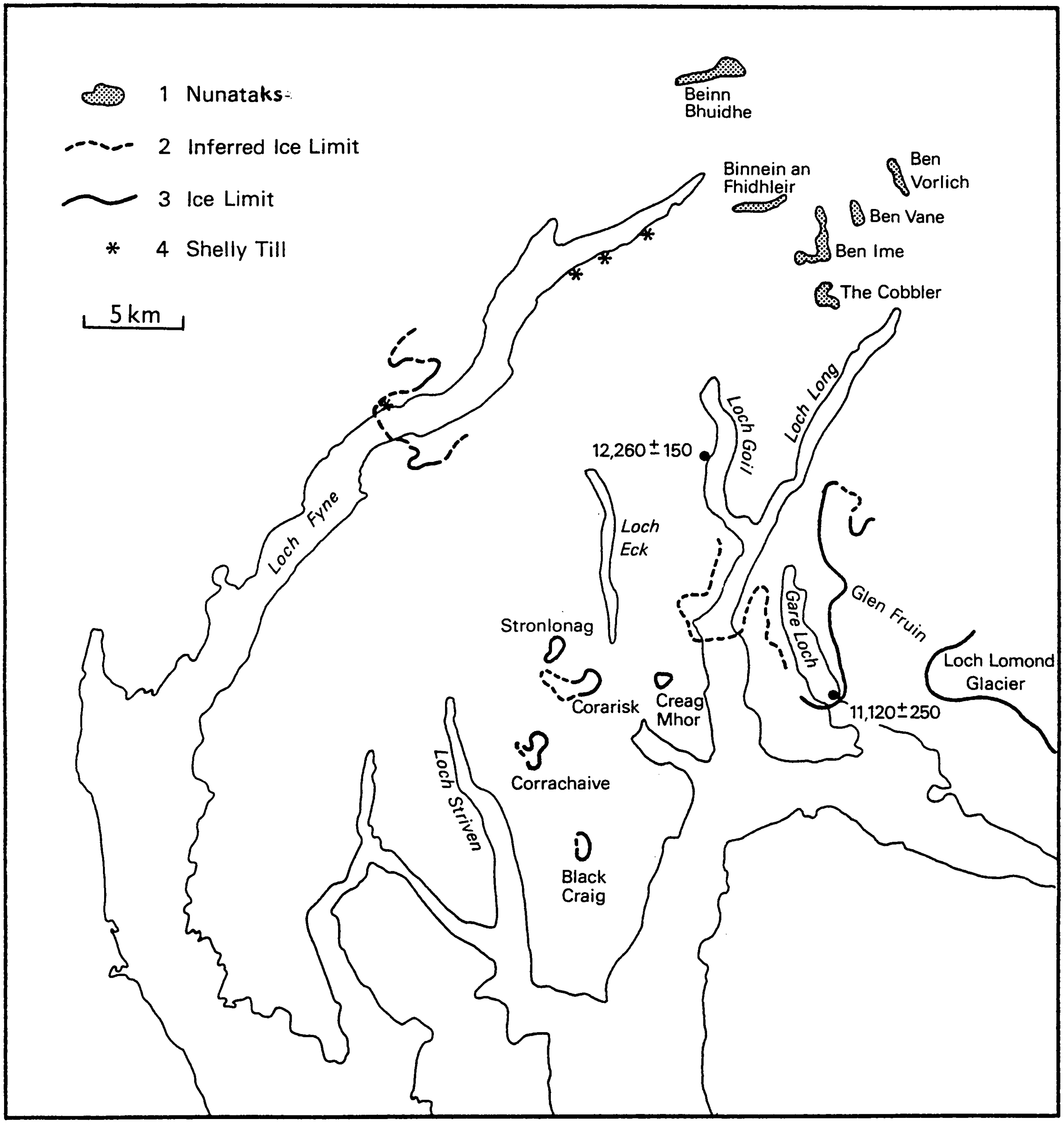
(f) Summary

A rather comprehensive picture emerges of the study area during the Loch Lomond Stadial. Much of the NE was inundated by ice with mountains above ca. 750-800 m projecting as nunataks. Large outlet glaciers flowed down the sea-lochs reaching to beyond Furnace in Loch Fyne and to Ardentinny in Loch Long. South of the main ice mass small glaciers were nourished in various corries and valley heads. The equilibrium firn line altitude was around 300 m. Sea-level at the glacier termini was ca. 12 m O.D. or below and during the ice advance the Main Rock Platform was at least in part fashioned. Mean sea-level temperatures for July were ca. 7°C colder than today and mean January temperatures were at least 14°C colder. A cold water fauna inhabited the sea and a sparse vegetation the land.

4. Other Glacial Stages

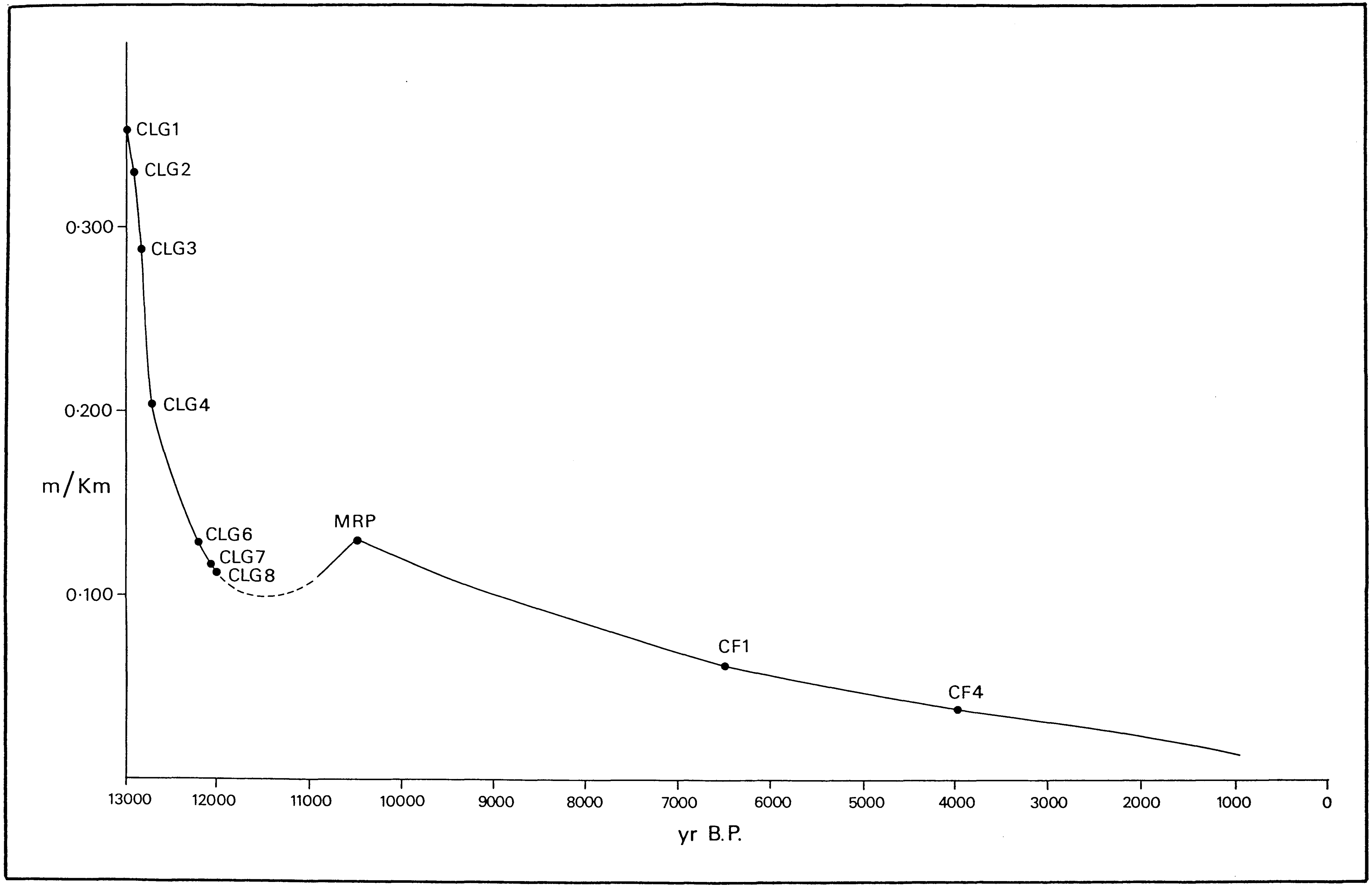
During the course of this thesis two other ice limits have been described, in Glendaruel (Chapter 6.3) and in Strath Eachaig (Chapter 5.4). Both of these limits occur a short distance inside the high shorelines and ice marginal features related to the Otter Ferry Stage but both relate to sea-levels below the uppermost Flandrian shoreline. It has not been possible to relate these ice limits to any other limits in their neighbourhood and their position in the history of deglaciation is as yet unknown. They may be related to a time between the Otter Ferry Stage and the Loch Lomond Readvance or they may result from the Loch Lomond Readvance, though this latter possibility, particularly in the case of the Glendaruel ice limit, would pose a considerable problem as to the origin of the glaciers.

Figure 71: Distribution of Loch Lomond Readvance glaciers and location of relevant radiocarbon dates.



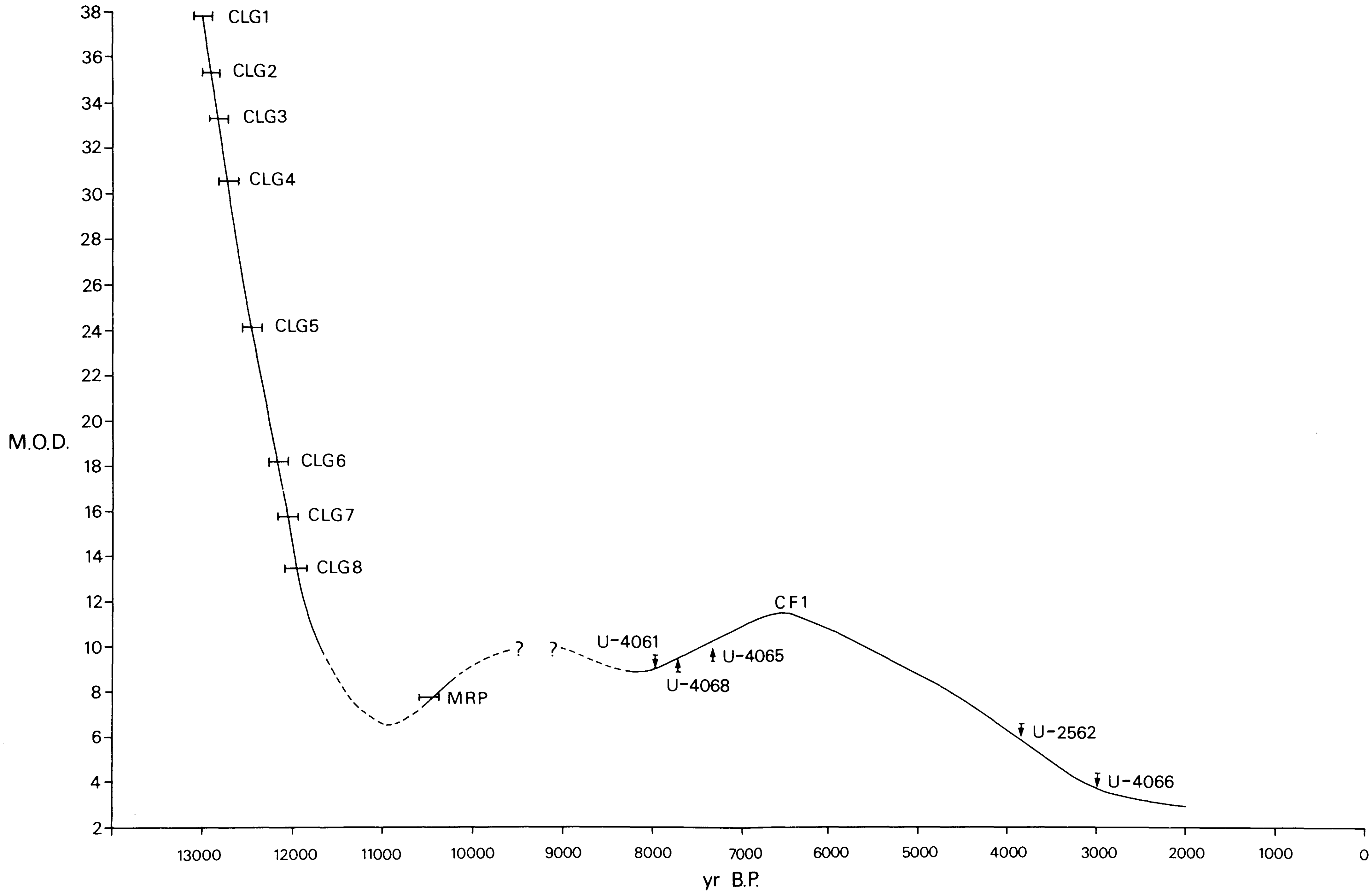
1 Nunataks
2 Inferred Ice Limit
3 Ice Limit
* 4 Shelly Till
5 km
Beinn Bhuidhe
Binnein an Fhidhleir
Ben Vorlich
Ben Vane
Ben Ime
The Cobbler
Loch Fyne
Loch Long
Loch Goil
12,260 ± 150
Loch Eck
Gare Loch
Glen Fruin
Loch Lomond Glacier
Stronlonag
Corarisk
Creag Mhor
11,120 ± 250
Loch Striven
Corrachaive
Black Craig

Figure 72: Graph of change in shoreline gradient with time.



CLG1
CLG2
0·300
CLG3
0·200
CLG4
m/Km
MRP
CLG6
CLG7
CLG8
0·100
CF1
CF4
13000
12000
11000
10000
9000
8000
7000
6000
5000
4000
3000
2000
1000
0
yr B.P.

Figure 73: Graph of change in relative sea-level with time.



M.O.D.
yr B.P.
38
36
34
32
30
28
26
24
22
20
18
16
14
12
10
8
6
4
2
13000
12000
11000
10000
9000
8000
7000
6000
5000
4000
3000
2000
1000
0
CLG1
CLG2
CLG3
CLG4
CLG5
CLG6
CLG7
CLG8
MRP
?
?
U-4061
U-4068
U-4065
CF1
U-2562
U-4066

CHAPTER 12

CONCLUSIONS

1. Introduction

In the course of this thesis an attempt has been made to establish the sequence of raised shorelines that has been formed in the Loch Long-Loch Fyne area of the SW Highlands consequent upon the retreat of the last ice sheet to cover this area. As many of the raised shorelines are intimately related to the presence of glacier ice it has also been necessary to establish the mode of deglaciation of the area and to consider the effects of halts or readvances of the ice front. In this concluding chapter it is proposed to examine the sequence of events as a whole, commencing with deglaciation and continuing till the present.

2. The period of ice-sheet deglaciation

At ca. 13,000 yr BP, or slightly earlier, the southern part of the Cowal Peninsula became ice free as the ice front, presumably greatly assisted by calving, retreated up the Firth of Clyde. The Portavadie/ Ardlamont area was first deglaciated and shortly after this the Toward Point area also became ice-free. Relative sea-level was ca. 36 m O.D. at Ardlamont and ca. 38 m O.D. near Toward. The ice front continued to retreat up the sea-lochs and areas of dead-ice were abandoned in some of the side valleys such as at Millhouse near Portavadie. Relative sea-level at this time was falling and a shoreline CLG1 was formed in the southern part of the study area after a ca. 2 m drop in sea-level at Ardlamont. Ice retreat, again presumably largely aided by calving, may have been of the order of several hundred metres per year at this time.

When the ice front had retreated to Otter Ferry in Loch Fyne, the mouth of Glendaruel, beyond the head of Loch Striven, and to unspecified localities in Strath Eachaig and Loch Long, a halt (possibly a readvance) occurred in the ice retreat. This event is dated to ca. 12,900 ± 200 yr BP. The ice limit of this period is correlated from valley to valley by a major shoreline, CLG2, with a gradient of ca. 0.33 m/km. This shoreline also correlates with areas of dead-ice near Dunoon and at Millhouse and with deposits at the mouths of certain side valleys that suggest there was decaying ice in these valleys at this time. This halt in the ice front has been termed the Otter Ferry Stage and it has been proposed that shoreline CLG2 correlates with the Main Perth Shoreline of SE Scotland.

The evidence suggests that the ice front remained at the above locations while relative sea-level fell by ca. 20 m at Otter Ferry, ca. 22 m in Glendaruel and possibly ca. 25 m in the Strath Eachaig and Loch Long areas. During this period of ice front still-stand three shorelines, CLG3, CLG4 and CLG5 were formed. By the time of the formation of CLG6, however, the ice had retreated to some unspecified location. The retreat of the ice from the Otter Ferry Stage is dated to ca. 12,500 ± 200 yr BP. Shoreline CLG6 can be traced up Loch Fyne as far as the village of Minard but there is no evidence to relate it to any ice margin.

3. The period from the end of the Otter Ferry Stage till the Loch Lomond Stadial

Three Lateglacial shorelines, CLG6, CLG7 and CLG8, have been established to have been formed after the Otter Ferry Stage and prior to the fall of relative sea-level to below the height of the uppermost Flandrian deposits. These shorelines are dated tentatively to between 12,500 and 11,800 yr BP. None of these shorelines can be followed up Loch Fyne beyond Minard and none of them relates directly to evidence of glacier ice. The marine sediments of this period are occasionally laminated and contain many dropped stones and boulders that suggest sea-ice and may indicate the presence of glaciers in the inner parts of the sea-lochs. The evidence is not clear, however, for there followed a major climatic reversal, the Loch Lomond Stadial, accompanied by a readvance of glaciers along parts of the sea-lochs and depositional evidence of prior glacial activity in the NE of the study area was destroyed.

4. The Loch Lomond Stadial

Relative sea-level continued to fall till after ca. 11,300 yr BP. It is probable that between this time and ca. 10.500 yr BP the marine erosion that has been documented as at least being partially responsible for the formation of the Main Rock Platform had effect. This implies a relatively stable sea-level for this time period. The Main Rock Platform and associated cliffline have been traced to the heads of the sea-lochs and thus there was no occupance of these areas by glacier ice at the time that the Main Rock Platform began to form. The glaciers subsequently expanded, however, to reach Ardentinny in Loch Long and beyond Furnace in Loch Fyne. Mountains in the NE of the study area such as Ben Ime stood proud of the general ice cover in this region. In the S of the Cowal Peninsula small valley glaciers developed. Sea-level at the maximum of the readvance was no higher than ca. 12 m O.D.. Analysis of shoreline gradients and the sea-level curve suggests that the build-up of ice during the Loch Lomond Stadial was sufficient to depress the earth's crust anew.

5. The Flandrian Period

The stratigraphy of the radiocarbon-dated sites at Kilfinan and Furnace suggests that the movement of relative sea-level in the early part of the Flandrian Period was downwards, this occurring prior to 8,000 yr BP. By 7,800 yr BP, however, transgression was in progress and this culminated in the formation of the uppermost Flandrian shoreline, CF1. In this study this shoreline is known to have been formed after 7,200 yr BP and before 3,800 yr BP but it has been correlated with the Main Postglacial Shoreline of the E coast of Scotland which implies an age of ca. 6,700 - 6,500 yr BP.

Subsequent to this the relative movement of sea-level was one of regression, perhaps punctuated by short transgressions. Five distinct shorelines have been formed during the period, CF2-6, CF4 being formed slightly before ca. 3,800 yr BP and CF5 slightly before ca. 3,000 yr BP. Sea-level must have reached its present level at some time during the last 2,000 years.

APPENDIX 1

MARINE SHELL RADIOCARBON DATES

1. Introduction

The principles on which marine shell radiocarbon dating is based are well established and have been frequently discussed (e.g. Olsson and Blake, 1962; Mangerud, 1972; Thom, 1973; Mangerud and Gulliksen, 1975; Harkness, 1979). The application of these principles has been extremely varied, however, and there is no uniformly accepted set of procedures for dealing with such dates, despite attempts to introduce such standard procedures at radiocarbon conferences (D.D. Harkness, pers. comm.). It is intended here to state briefly the principles involved and then, with particular reference to the Lateglacial marine shell radiocarbon dates from the SW Highlands and Clyde Estuary, to review the practice of radiocarbon laboratories in dating such material. Finally, the magnitude of the 'apparent age' of sea water correction is discussed. The objective is to produce, as far as is possible, a standardised list of radiocarbon dates. The dates should be (i) internally consistent and hence comparable with one another, and (ii) accurate in that they can be directly compared with radiocarbon chronologies based on assay of land-based materials.

2. Statement of Principles

Radiocarbon is produced in the upper atmosphere by cosmic irradiation of nitrogen atoms. Since radiocarbon decays at a constant rate the total amount of radiocarbon will increase until the amount decaying is equal to the amount being formed. Due to mixing, photosynthesis, respiration, etc., radiocarbon becomes distributed throughout the various parts of the carbon cycle in a time that is generally considered short compared to the radiocarbon half-life (Libby, 1955). On removal from the active carbon cycle (by death of organisms, precipitation of carbonates) the radiocarbon content of a substance is no longer renewed and the radiocarbon decays at a known and well established rate. Knowledge of this decay rate plus an assumption that the original concentration of radiocarbon in a particular type of material was the same in the past as is observed in similar material today allows, upon measurement of the present concentration of radiocarbon in the material, the calculation of the time that has elapsed from the removal of the material from the active carbon cycle till the present day. The assumption of the original concentration value of radiocarbon is important for marine shell dating, for in practical terms all the dates are calculated with respect to a modern standard and carbonaceous materials from different parts of the carbon cycle have radiocarbon concentrations that are different by greater or lesser amounts from the standard. These differences in radiocarbon concentration translate in dating terms to differences in ages as measured today on contemporaneous materials. Such differences are known as the 'apparent ages' of the materials and to compare radiocarbon ages of different materials their apparent ages must be known.

The bulk (c. 90%) of the carbon in the active carbon cycle is contained in the world's oceans and, due to the relatively slow turn-over of the oceans, radiocarbon replacement lags somewhat behind the atmosphere, giving the oceans an apparent radiocarbon age. The turn-over of the oceans varies latitudinally, and between roughly 40°N and 40°S there is a strongly developed thermocline that separates a relatively well mixed surface layer (apparent age 300-500 years) from a deeper, much older layer (apparent age possibly greater than 1000 years). Polewards of these latitudes the deeper water upwells with greater frequency and radiocarbon ages of sea water have been recorded in excess of 2000 years, although in general they are much less than this.

In addition to the concept of apparent age it has also been long established that the different carbon isotopes (^{12}C, ^{13}C, ^{14}C) are assimilated in different relative proportions by different chemical and biological processes, ^{14}C being discriminated against or preferred in approximately twice the amount of ^{13}C (Craig, 1953). Since the majority of the materials that are radiocarbon dated are terrestrial plants, by convention this is taken as the appropriate level of isotopic fractionation to which all radiocarbon assays are 'normalised' during computation of the ages. This normalisation is effected by measuring the deviation of the amount of the $^{13}C/^{12}C$ ratio in the sample from the same ratio in a standard. This deviation (termed $\delta^{13}C$) is multiplied by two to establish the deviation in the amount of ^{14}C, and the date adjusted accordingly. The standard for such measurements is a belemnite from the Peedee marine limestone in South Carolina (hence the term PDB) and other marine carbonates do not show significant differences from it in average isotopic composition (i.e. $\delta^{13}C_{PDB} = 0.0‰$). Most terrestrial plants, on the other hand, show considerable fractionation ($\delta^{13}C_{PDB} = -25‰$) and shell dates are, in general, ca. 400 years younger than terrestrial dates unless isotopic fractionation is taken into consideration.

3. Practice in Radiocarbon Laboratories

In addition to giving advice to users of the radiocarbon dating method the functions of a radiocarbon laboratory can be divided into three: (a) samples are chemically and mechanically pre-treated to remove potential contaminants; (b) samples are converted to the counting medium and assayed; and (c) the assay statistics (of sample, standard, background and isotopic fractionation measure) are mathematically combined to produce the sample age. There are considerable differences from one laboratory to another in the way these procedures are carried out, and, with improved knowledge, laboratory procedures also change from time to time. Tables 29 and 30 list all the Lateglacial shell dates available from the SW Highlands and Clyde Estuary and it can be seen that the following laboratories are involved: Trondheim (2 samples), Glasgow (1), Birmingham (4), Stockholm (7), Isotopes (2), Harwell (1), East Kilbride (24) and Riken (1). While this number of laboratories might reduce the possibility of a systematic error that could result from the use of one laboratory alone, it has also produced a series of radiocarbon dates that are not immediately comparable one with another.

(a) Pretreatment

A general approach to pretreatment of shells might be as follows. Shells are scrubbed mechanically to remove dirt and then placed in a weak solution of HCl and the surface layers leached and discarded. The remainder of the shell is then prepared for assay. Such a procedure is adopted for it is assumed that any contamination will be greatest in the outer layers of the shell.

There are a number of variations on this procedure, however. Trondheim removed the outer 10% of the shells in samples T-1455 and T-1456 as did Riken (Yamasaki _et al_., 1968); Glasgow regularly removes the outer 20% (Baxter _et al_., 1969) as does East Kilbride (Harkness and Wilson, 1979); Birmingham 'normally' discards the 'outer fraction' (Shotton _et al_., 1969, p 263) although in the case of Birm-361 the outer fraction was dated. Isotopes removed from 18-25% of the samples dated by them (Buckley, pers. comm.).

A further variation is introduced by those laboratories that continue treating the shell sample with dilute HCl to produce two further fractions (usually termed 'inner' and 'outer') that are assayed separately. In part this procedure is dependent upon the sample being large enough to produce two separate fractions that are both of sufficient size to be measured independently. Once again this procedure is adopted on the assumption that the outer fraction is more likely to be contaminated than the inner one. This inner fraction, it should be noted, is not the inner surface of the shell but the last 40% (or so) to be leached by the acid (thus the farthest part from any surface) and therefore should provide the most reliable date.

Unfortunately, none of the shell samples being considered here have been analysed by X-ray diffraction to check whether there has been any significant amount of recrystallisation of the original shell matrix (cf. Chapell and Polach, 1972; Thom, 1973; Mangerud, 1972).

(b) Sample preparation and assay

Following pretreatment the carbon of the sample must be converted into the counting medium, usually carbon dioxide, methane or benzene. During this process isotopic fractionation is possible (Harkness, 1979) and this further emphasises the need to measure $\delta^{13}C$ values of samples. In general sample measurement consists of counting a number of samples interspersed with standard and background counts in order to ensure the stability of the counting system. Almost all laboratories use NBS oxalic acid for a standard, but Trondheim, for example, when measuring shell samples uses the activity of 'modern' shells as a standard (Mangerud, 1972). This, it is argued (Nydal et al., 1972) should give, without further corrections for either $\delta^{13}C$ or apparent age of sea water, the true age of the shell.

(c) Age calculation

The main problem connected with age calculation of shell samples is whether or not a correction should be applied to normalise the sample with respect to the isotopic ratios of contemporaneous terrestrial material. As has been mentioned earlier, isotopic fractionation results in differences of around 400 years in the ages of contemporaneous marine carbonate and terrestrial plant materials, the marine carbonate appearing younger. On the other hand the apparent age of sea water is in many areas +300 to +500 years and numerous radiocarbon laboratories have argued that in practice these two effects cancel one another out and hence the relevant corrections are not necessary. Such an approach ignores (i) the possibility of laboratory-induced isotopic fractionation, (ii) the possibility that the isotopic fractionation correction may be significantly different from that assumed (a shell sample with a $\delta^{13}C$ value of +5 ‰ has a correction of ca. 480 years), and (iii) the possibility that the apparent age of sea water is very much greater than is assumed.

Of the laboratories concerned with the present study the following practices have been followed. Birmingham initially did not measure $\delta^{13}C$ values and so early Birmingham samples (Birm-120/1/2) are uncorrected, although later ones (e.g. Birm-361) have been corrected. Similarly, East Kilbride initially applied no $\delta^{13}C$ corrections (this applied to SRR-62/3) although all subsequent dates have been so treated and the earlier samples corrected in Table 30 (D.D. Harkness, pers. comm.).

Trondheim, as mentioned above, does not apply corrections and neither has Isotopes who only measure $\delta^{13}C$ on request by submitters. The Stockholm laboratory did not apply isotope corrections to the samples analysed on contract to the Institute of Geological Sciences. Harwell and Glasgow have applied isotope fractionation corrections, although the $\delta^{13}C$ value of sample GU-12 was lost when the Glasgow laboratory was flooded (M.S. Baxter, pers. comm.). The Riken laboratory did not correct sample N-475 (Hamada, pers. comm.).

4. Apparent age of sea water

A number of radiocarbon age determinations have been made on recent shell samples from the W coast of Scotland. Five such determinations reported in Harkness and Wilson (1979) indicated a present-day apparent age of sea water in this area of 350 ± 30 years. Subsequent datings (Harkness, pers. comm.) suggest, however, that a figure nearer 400 years may be more appropriate.

A considerable number of recent shell samples have been dated from the Scandinavian coasts (Mangerud, 1972; Mangerud and Gulliksen, 1975) and they are notable for the apparent increase in apparent age northwards along the Norwegian coast. Thus samples from SW Sweden give apparent ages of ca. 400 years, southern Norway ca. 430 years, northern Norway ca. 470 years and Spitsbergen perhaps more than 500 years. This appears to be due to the increasing influence of Arctic waters and increased upwelling towards the pole. Confirmation of this pattern comes from East Greenland where Hjort (1973) has reported apparent ages of ca. 570 years.

The relevance of the Scandinavian and Greenland data to shell dates from the Lateglacial period in western Scotland is that during the decay of the last ice sheet and again during the Loch Lomond Stadial polar waters were present off the Scottish coasts (Ruddiman and MacIntyre, 1973; Ruddiman et al., 1977). Since polar water presumably 'advances' by upwelling in increasingly southerly latitudes then the possibility of apparent ages of shells being greater at those times than they are today must be large, perhaps, however, being counterbalanced to some extent in nearshore waters by the increased run-off from a decaying ice sheet or from the glaciers that existed during the Loch Lomond Stadial (Sissons, 1979a). In the present state of knowledge these latter considerations are unquantifiable though they suggest a correction for the apparent age of sea water somewhat greater than 350 $\pm$ 30 years.

Given the above, a figure of 400 years appears to be the best current estimate of the apparent age of sea water for the W coast of Scotland. This correction is essentially empirical, being derived from comparison with present-day shells. Some authors (e.g. Jardine, 1978) have mentioned the possibility of 'hard water error' in addition to the apparent age of sea water as possibly influencing the radiocarbon age of a sample, but provided the shells being dated lived in open water conditions then the empirically derived correction should take any additional errors into account too. The conditions in estuaries may be different, however, and where there is a large fresh water input, particularly from rivers draining areas of limestone, then locally a hard water error might arise in addition to the open sea apparent age of sea water correction.

5. Standardization of dates

From the foregoing considerations, it is clear that there are a number of inconsistencies in the age determinations of the samples assembled in Tables 29 and 30. In an attempt to produce a comparable series of dates the following procedures have been adopted.

(i) Dates have been corrected for isotopic fractionation wherever possible. If doubt existed, the relevant radiocarbon laboratory was consulted. Those dates for which no carbon isotope ratios were measured are reported with no correction applied.

(ii) A figure of 400 years has been adopted for the apparent age of sea water and this value has been subtracted from the ages of those samples that have been corrected for isotopic fractionation. If no $\delta^{13}C$ corrections had been made, no apparent age of sea water correction was applied.

(iii) If 'inner' and 'outer' fractions of the samples have been dated, the inner fraction is always considered the more reliable and it alone is used as the sample age. The procedure of calculating the arithmetic mean of the two fractions is not adopted, for the outer fraction often differs from the inner because of contamination. In such a circumstance to combine the two fractions would only produce a date that is known to be in error. The outer fraction is considered only as a guide to the likelihood of contamination of the inner. A useful test of the 'internal consistency' of shell samples with inner and outer ages is to compare the difference in age of the two samples with the combined standard deviation ($=\sqrt{\sigma^2_{inner} + \sigma^2_{outer}}$). If the difference between the ages is less than one combined standard deviation then the date is regarded as internally consistent.

(iv) Reported ages and standard deviations were rounded to the nearest ten years (cf. Stuiver and Polach, 1977).

These procedures differ in a number of respects from the published literature. Peacock et al. (1977) did not use dates that were corrected for isotopic fractionation although subsequently (in Peacock et al., 1978) the Lochgilphead dates were adjusted by 90 years to compensate for both isotopic fractionation and age of sea water corrections. In Table 30 isotopic fractionation corrections have been made to the original Lochgilphead data (Harkness and Wilson, 1979) and subsequently the apparent age of sea water deducted.

Browne et al. (1977) and Peacock et al. (1978) used an apparent age of sea water correction of 350 ± 20 years, which, as discussed above, is not used here.

Bishop and Dickson (1970), Gray (1975d), Browne et al. (1977) and Peacock et al. (1978) all combined inner and outer fractions of shell samples, on one occasion producing an apparent inversion of the stratigraphic order of the dates in doing so (Peacock et al., 1978: sample SRR-484 was assigned an age of 10,195 ± 117 years BP on this basis although it lay below sample SRR-615 which had been assigned an age of 10,412 ± 136 years BP).

The limitations of the original laboratory procedures cannot be overcome by subsequent manipulation of the data. The above procedures standardize the data to a much greater extent than previously and also allow the dates to be graded in terms of the degree of confidence that can be attached to them. Thus, for example, most confidence can be had in a sample for which isotopic fractionation has been measured and corrected for, in which pretreatment has resulted in the discarding of the outer layers of the shell and the division of the remainder of the shell into two separately assayed fractions, and in which there is no significant difference at the 1σ level between the ages of the two fractions.

One of the objectives of this section was to produce a series of dates that are comparable with dates on terrestrial material relating to similar events. There are few opportunities for direct comparison, but one occurs in the Ardyne series of dates, for there three samples date material associated with the very cold Loch Lomond Stadial, and two samples date deposits immediately preceding the Loch Lomond Stadial. The dates during the Loch Lomond Stadial are 10,440 ± 100 yr. BP (SRR-484), 10,350 ± 140 yr BP (SRR-615) and 10,220 ± 70 yr BP (SRR-486) of which the last can be viewed with most confidence satisfying all the conditions set out above. It occurs near the top of the Stadial deposits. The two dates from material immediately underlying the Stadial deposits are 10,720 ± 110 yr BP (SRR-485) and 11,130 ± 60 yr BP (SRR-483) both of which are internally consistent dates. The actual dating of the Loch Lomond Stadial by terrestrial deposits is currently rather controversial (Lowe and Gray, 1980) but the traditional dating was that Pollen Zone III (equivalent to the Loch Lomond Stadial) lasted from 10,750 till 10,250 yr BP (Godwin and Willis, 1959) which, whatever the current status of the debate over the details of the event, suggests that marine shell dates produce results quite comparable with terrestrial materials.

TABLE 29. RADIOCARBON AGES OF SHELLS OVERRIDDEN OR TRANSPORTED BY GLACIERS IN SW HIGHLANDS

Locality	Shell Species	Lab No	$\delta^{13}C_{PDB}$ ‰	Laboratory Age yr BP $\pm$ 1σ	This Study Age yr BP $\pm$ 1σ
Drymen	mixed shells	I-2235	-	11,700±170	11,700±170
Kinloch-spelve	mixed shells	I-5308	-	11,330±170	11,330±170
Rhu	mixed shells	HAR-931	+1.4	11,520±250	11,120±250
Loch Goil	Chlamys islandica	T-1456	-	12,260±150	12,260±150
South Shian	Chlamys islandica	IGS-C14/16	-	(O)11,300±300 (I)11,530±210	11,530±210
South Shian	Astarte elliptica	IGS-C14/17	-	11,805±180	11,810±180
South Shian	mixed shells	IGS-C14/18	- -	(O)16,705±130* (I)11,430±220	11,430±220

* This date wrongly recorded as 6,705 ± 130 in Radiocarbon 14 (1), pp 140-144.

TABLE 30. RADIOCARBON AGES OF SHELLS FROM SW HIGHLANDS AND CLYDE ESTUARY DURING LATEGLACIAL PERIOD.

Locality	Altitude m O.D.	Shell Species	Lab No.	$\delta^{13}C_{PDB}$ ‰	Laboratory Age yr BP $\pm$ 1σ	This Study Age yr BP $\pm$ 1σ
Lochgilp-head	- 0.1	Arctica islandica	SRR-63	-0.8	12,750$\pm$90	12,350$\pm$90
Lochgilp-head	+ 0.4	Modiolus modiolus	SRR-364	0.0 +0.3	(O)11,690$\pm$130 (I)11,390$\pm$110	10,990$\pm$110
Lochgilp-head	+ 0.4	Mya truncata	SRR-365	+0.5 +0.5	(O)12,110$\pm$130 (I)11,730$\pm$140	11,330$\pm$140
Lochgilp-head	+ 0.4	Arctica islandica	SRR-366	+1.0 +0.6	(O)11,390$\pm$120 (I)11,690$\pm$130	11,290$\pm$130
Lochgilp-head	+ 0.25	Macoma calcarea	SRR-367	-0.7	12,490$\pm$190	12,090$\pm$190
Lochgilp-head	- 0.05	Arctica islandica	SRR-368	+1.1 +1.0	(O)13,620$\pm$90 (I)13,190$\pm$85	12,790$\pm$85
Lochgilp-head	- 0.7	Modiolus modiolus	SRR-369	+0.1 +1.8	(O)13,140$\pm$90 (I)12,650$\pm$110	12,250$\pm$110
Lochgilp-head	- 0.05	Arctica islandica	SRR-489	+1.6 +1.8	(O)12,510$\pm$60 (I)12,590$\pm$70	12,190$\pm$70
Ardyne Point	- 1.0	Arctica islandica	SRR-481	+1.8 +1.0	(O)12,150$\pm$120 (I)11,680$\pm$170	11,280$\pm$170
Ardyne Point	- 1.6	Modiolus modiolus	SRR-482	+0.3 +0.9	(O)12,650$\pm$70 (I)12,380$\pm$90	11,980$\pm$90
Ardyne Point	- 5.5	Arctica islandica	SRR-483	+1.7 +1.7	(O)11,470$\pm$70 (I)11,530$\pm$60	11,130$\pm$60
Ardyne Point	- 4.0	Mya truncata	SRR-484	+1.1 +0.7	(O)10,230$\pm$210 (I)10,840$\pm$100	10,440$\pm$100
Ardyne Point	- 5.2	Mya truncata	SRR-485	-0.4 +1.0	(O)11,170$\pm$70 (I)11,120$\pm$110	10,720$\pm$110
Ardyne Point	- 2.5	Mya truncata	SRR-486	+1.1 +1.0	(O)10,560$\pm$60 (I)10,620$\pm$70	10,220$\pm$70
Ardyne Point	- 3.5	Mya truncata	SRR-615	+0.9	10,750$\pm$140	10,350$\pm$140
Port-avadie	+ 2.0	Mya truncata	SRR-831	+0.9 +1.8	(O)10,830$\pm$150 (I)11,140$\pm$220	10,740$\pm$220
Port-avadie	+ 5.2	Arctica islandica	SRR-832	+1.7 +1.5	(O)11,750$\pm$70 (I)11,930$\pm$80	11,530$\pm$80
Inch-innan	+ 0.7	Arctica islandica	SRR-923	+1.5 +1.4	(O)12,289$\pm$310 (I)14,346$^{+625}_{-580}$	13,950$^{+630}_{-580}$
Inch-innan	+ 0.7	Modiolus modiolus	SRR-924	-0.3 +1.3	(O)13,083$\pm$150 (I)12,464$\pm$275	12,060$\pm$280
Inch-innan	+ 6.1	Arctica islandica	SRR-925	+2.8 +1.4	(O)13,492$\pm$330 (I)13,501$\pm$265	13,100$\pm$270

TABLE 30 (cont)

Locality	Altitude m O.D.	Shell Species	Lab No	$\delta^{13}C_{PDB}$ ‰	Laboratory Age yr BP ± 1σ	This Study Age yr BP ± 1σ
Inch-innan	+ 6.1	Modiolus modiolus	SRR-926	+0.3 +0.3	(O)12,389±110 (I)12,333±110	11,930±110
Inch-innan	+11.6	Arctica islandica	SRR-927	+0.8 -0.1	(O)12,694±100 (I)12,922±125	12,520±130
Renfrew By-pass	9.0-12.25	Arctica islandica	SRR-62	-1.2	12,790±90	12,390±90
Cardross	+ 7.5	Arctica islandica	SRR-833	-0.4 -0.5	(O)12,720±170 (I)12,720±240	12,320±240
Cardross	+ 7.5	Mainly Arctica islandica	GU-12	?	11,790±120	11,390±120
Cardross	+ 7.5	Arctica islandica	N-475	-	11,900±205	11,900±210
Wester Fulwood	+ 3.0	Arctica islandica	Birm-122	-	(O)13,020±220 (I)12,650±200	12,650±200
Greenock	- 3.5	Astarte sulcata	Birm-121	-	10,560±180	10,560±180
Greenock	- 1.4	Mya truncata	Birm-120	-	9,890±160	9,890±160
Glen Cruitten	+20.5	Balanus balanus	Birm-361	-0.2 -0.5 +1.2	(O)11,270±200 (M)12,190±240 (I)11,770±230	11,370±230
Dumbarton	-24.7 to -41.8	Chlamys islandica	IGS-C14/19	-	11,805±205	11,810±210
Gallow-hill, Paisley	+11.0	Arctica islandica	IGS-C14/20	- -	(O)12,930±160 (I)15,625±240	15,630±240
Gallow-hill, Paisley	+11.0	Arctica islandica	IGS-C14/68	- -	(O)12,125±210 (I)12,615±230	12,620±230
Ralston, Paisley	+16.0	Arctica islandica	IGS-C14/21	- -	(O)12,890±360 (I)12,610±210	12,610±210
Kyles of Bute	0.0	Arctica islandica	T-1455	-	11,110±140	11,110±140

APPENDIX 2

PRELIMINARY POLLEN ANALYSES FROM TROUSTAN, LOCH STRIVEN

The following pollen data are for the organic horizon found at Troustan and reported in Chapter 9. The analyses were performed by Dr. M.J.C. Walker and Dr. J.J. Lowe who counted 150 grains each.

Tree pollen

Betula	2.7%
Pinus	1.4%

Shrub pollen

Juniperus	3.0%
Salix	4.0%
Empetrum	36.3%

Herb pollen

Gramineae	14.6%	Cyperaceae	12.7%
Artemisia	2%	Caryophyllaceae	6%
Chenopodiaceae	0.7%	Compositae	3.7%
Epilobium	0.7%	Filipendula	0.3%
Plantago spp.	0.3%	Ranunculaceae	1%
Rosaceae	1.7%	Rumex	0.7%
Umbelliferae	0.3%	Valeriana	1.6%
Thalictrum	2%		

Aquatic pollen

Myriophyllum	0.7%	Sparganium	0.7%

Spores

Filicales	25.7%	Selaginella	27%
Lycopodium selago	12%	Lycopodium spp.	1%

The pollen spectrum clearly indicates an almost treeless environment dominated *Empetrum* and grasses and sedges. This, together with the occurrence of certain indicators of relatively cold climates, such as *Artemisia* and the *Carophyllaceae* indicates a Lateglacial age for the peat.

APPENDIX 3

SURVEYED ALTITUDES

This appendix includes details of all the points surveyed during this study. It is divided into 5 parts: Part 1 includes all the altitudes on the raised marine landforms except those cross-profiles levelled at _ca_. 2 m intervals; Part 2 includes all the altitudes on river terraces or outwash terraces; Part 3 includes those altitudes recorded on features around the present coastline, again excepting those cross-profiles levelled at _ca_. 2 m intervals; Part 4 includes those glacial deposits such as kames or kame terraces that were levelled; and Part 5 includes the various cross-profiles levelled at closely-spaced intervals. The information given in the first 4 parts corresponds to the heights indicated on the various features shown in the Figures in the text. Where relevant to further analysis a National Grid Reference is included, this being correct to the nearest 10 m. Altitudes are correct to 2 decimal places.

1. Raised Marine Landforms

(a) Terraces and Deltas

Feature	Easting	Northing	Mean Altitude	Grade	Point Altitude
S1	21865	71271	19.49	p	19.61 19.36
S2	21928	71253	9.34	p	9.07 9.66
S3	21809	71209	33.92	m	33.84 34.01
S4	21698	71118	8.45	m	8.50 8.57 8.56 7.93 8.71
S5	21698	71114	3.59	p	3.75 3.82 3.32 3.47
S6	21743	71140	10.03	m	9.93 10.13
S7	21746	71137	3.49	m	3.56 3.42
S8	21545	71034	10.04	m	10.07 10.01 9.46
S9	21261	70983	23.12	p	23.54 22.71
S10	21258	70939	31.31	p	31.07 31.55
S11	21134	71028	7.63	m	7.37 7.55 7.71 7.87
S12	21125	71053	10.60	m	10.63 10.58
S13	21004	70962	12,38	m	13.39 12.81 12.56 12.41 12.01 12.69 12.22
S14	20929	70882	11.22	p	13.91 12.74 11.52 11.02 10.56 10.59

Feature	Easting	Northing	Mean Altitude	Grade	Point Altitude
S15	20946	70897	10.41	m	10.62 10.48 10.15
S16	20951	70893	9.09	m	9.08 9.20 8.99
S17	22946	70499	10.85	p	10.97 10.72
S18	22105	69800	12.18	g	12.14 12.21
S19	22097	69797	3.30	m	3.43 3.18 3.28
S20	22102	69794	7.42	p	7.59 7.26
S21	22103	69797	9.22	p	9.22
S22	22104	69775	10.90	p	11.18 10.62
S23	21902	69527	10.86	m	11.13 11.08 10.50 10.71
S24	21919	69537	9.35	m	9.14 9.25 9.64
S25	21916	69529	2.78	m	2.85 2.72
S26	21913	69507	8.91	m	8.88 9.25 8.58
S27	21910	70139	12.99	m	13.05 12.83
S28	21893	70050	3.28	m	3.38 3.23 3.28
S29	21898	70038	8.74	m	8.73 8.76
S30	21889	70041	10.82	m	10.65 10.98
S31	21897	70019	2.59	p	2.49 2.68
S32	21893	70019	5.11	p	5.28 4.94
S33	21903	70013	9.39	p	9.39
S34	21893	70016	11.31	p	11.99 10.63

Feature	Easting	Northing	Mean Altitude	Grade	Point Altitude
S35	22029	69329	7.32	p	7.44 7.21
S36	21999	69364	12.94	p	12.85 13.07
S37	22009	69358	3.66	m	3.57 3.74
S38	20782	67376	5.06	p	4.89 5.23
S39	20776	67395	16.48	p	16.43 16.53
S40	20778	67401	12.15	m	11.92 12.00 12.52
S41	20780	67411	6.88	m	6.65 7.11
S42	20784	67398	4.16	m	4.14 4.18 4.31
S43	20734	67639	3.77	p	3.60 3.93
S44	20722	67670	3.99	p	3.97 4.01
S45	20725	67664	11.77	m	11.79 11.74
S46	20724	67645	11.35	m	11.32 11.38
S47	20714	67643	35.51	m	34.63 35.27 36.63
S48	20716	67658	20.16	p	19.95 20.36
S49	20723	67652	6.38	p	6.36 6.41
S50	20761	67142	4.43	m	4.72 4.14
S51	20763	67145	6.74	p	6.67 6.86 6.68
S52	20758	67145	8.64	m	8.59 8.70
S53	20697	67160	8.42	m	8.40 8.43
S54	20694	67167	11.79	m	11.13 12.45

Feature	Easting	Northing	Mean Altitude	Grade	Point Altitude
S55	20693	67173	23.94	m	24.46 23.91 23.44
S56	20689	67161	4.80	p	4.71 4.89
S57	20649	67172	4.12	p	4.00 4.24
S58	20653	67178	8.95	m	8.82 9.08
S59	20631	67194	30.59	m	30.65 30.53
S60	20638	67203	35.27	m	34.76 35.92 35.12
S61	20589	67187	6.06	p	6.29 5.14 4.50
S62	20591	67196	12.90	m	12.83 12.78
S63	20581	67200	12.51	p	12.50 12.51
S64	20576	67200	11.18	p	11.31 11.05
S65	20562	67203	8.12	p	7.87 8.36
S66	20562	67197	3.16	p	3.06 3.26
S67	20550	67210	3.17	p	3.02 3.11 3.37
S68	20163	67655	36.33	g	36.42 36.24
S69	20180	67642	34.78	m	34.37 35.19
S70	20177	67624	25.13	m	25.14 25.12
S71	20174	67618	15.77	m	15.79 15.75
S72	20330	67471	35.50	m	36.05 35.08 35.46 35.43
S73	20334	67477	38.08	p	38.44 37.72
S74	20342	67458	28.67	m	28.09 28.89 29.04

Feature	Easting	Northing	Mean Altitude	Grade	Point Altitude
S75	20541	68580	40.94	p	41.25 40.64
S76	20561	68535	37.78	g	37.99 37.56
S77	20535	68536	37.52	g	37.43 37.60
S78	20533	68493	37.93	g	38.04 37.83
S80	19960	67395	9.00	m	9.07 8.94
S81	19962	67392	6.77	p	6.60 6.94
S82	19941	67396	6.67	p	6.56 6.78
S83	19930	67396	5.47	p	5.52 5.43
S84	19929	67400	12.39	m	12.84 11.95
S85	19924	67403	30.31	g	29.95 30.67
S86	20567	68250	6.96	p	7.73 6.18
S87	20557	68257	3.85	m	3.87 3.83
S89	19843	69268	7.79	m	7.85 7.72
S90	19860	69280	9.04	m	9.18 8.90
S91	19872	69276	12.77	m	12.78 12.77
S92	19866	69295	7.42	p	7.78 7.07
S93	19863	69295	3.92	p	3.75 4.09
S94	21942	71250	10.83	p	10.83
S95	21949	71259	13.07	m	13.05 13.10
S96	21945	71246	13.92	m	13.92 13.92
S97	21851	71133	12.95	m	12.97 13.00 12.87
S98	21805	71073	11.29	p	11.08 11.49

Feature	Easting	Northing	Mean Altitude	Grade	Point Altitude
S99	21808	71070	12.59	p	12.59
S100	21812	71069	16.34	p	16.29
					16.40
S101	21780	71080	5.65	p	5.54
					5.76
S102	21763	71022	14.27	m	14.50
					14.04
S103	21388	70831	6.58	p	6.46
					6.70
S104	21384	70825	9.59	m	8.99
					9.60
					10.19
S105	21401	70827	13.54	p	13.74
					13.73
S106	21455	70837	4.50	m	4.33
					4.67
S107	21460	70834	13.44	p	9.65
					10.14
					13.44
S108	21469	70836	9.16	p	8.91
					9.42
S109	21479	70827	14.01	m	14.06
					13.96
S110	21479	70824	23.30	p	23.16
					23.44
S111	21477	70852	8.03	m	8.03
					8.03
S112	21484	70854	7.08	m	7.11
					7.04
S113	21483	70848	10.58	p	10.62
					10.55
S114	21490	70839	13.93	g	14.02
					13.85
S115	21497	70841	13.74	m	13.74
S116	21494	70851	11.28	m	11.24
					11.32
S117	21515	70856	8.92	m	9.12
					8.49
					9.15
S118	21503	70857	5.76	m	5.69
					5.82
					5.76
S119	21491	70860	6.54	m	6.53
					6.55

Feature	Easting	Northing	Mean Altitude	Grade	Point Altitude
S120	21486	70864	3.98	p	4.08 3.89
S121	21480	70861	4.79	m	4.93 4.79 4.66
S122	19452	68667	11.31	p	12.72 11.54 11.06 10.78 10.47
S123	19458	68692	8.49	m	9.16 9.07 8.93 8.04 7.95 7.84
S124	19472	68716	7.80	m	8.32 7.77 7.62 7.50
S125	19451	68680	4.18	p	4.12 4.24
S126	19446	68643	13.25	m	13.56 12.94
S127	19373	68513	7.06	m	7.06 6.92
S128	19379	68519	11.10	m	10.94 10.86 11.27 11.32
S129	19366	68507	3.78	p	3.85 3.72
S130	19369	68504	7.67	m	7.67 7.67
S131	19369	68489	10.19	m	9.91 10.14 10.18 10.54
S132	19380	68497	17.78	p	17.90 17.67
S133	19357	68489	3.60	p	3.82 3.39
S134	19314	68429	18.81	m	18.97 18.64
S135	19325	68444	8.27	p	8.27
S136	19322	68447	5.22	p	5.39

Feature	Easting	Northing	Mean Altitude	Grade	Point Altitude
S137	19318	68451	4.23	p	4.00 4.36 4.32
S138	19260	68435	4.05	m	3.97 4.13
S139	19190	68317	4.26	m	4.06 4.45
S140	19176	68306	6.87	m	6.81 6.93
S141	19224	68338	3.95	p	3.88 4.02
S142	19272	68416	7.51	m	7.34 7.44 7.50 7.74
S143	19239	68396	11.78	m	11.79 11.68 11.87
S144	19240	68365	14.43	m	14.13 14.74
S145	19242	68346	14.35	m	14.09 14.60
S146	19284	67951	20.53	g	20.33 20.54 20.72
S147	19236	67931	4.18	p	4.50 4.20 3.83
S148	19256	67952	14.63	p	14.28 14.98
S149	19254	67682	8.26	g	8.85 8.34 8.69 9.22 8.46 8.15 7.63 7.61 7.39
S150	19246	67878	4.83	m	4.73 4.83 4.96
S151	19238	67835	5.97	p	5.80 6.14
S152	19149	67709	3.96	m	3.98 3.94

Feature	Easting	Northing	Mean Altitude	Grade	Point Altitude
S153	19141	67660	13.38	p	13.26 13.27 13.59
S154	19145	67666	14.51	p	14.86 14.95 15.05 14.20
S155	19176	67990	24.03	g	23.28 23.90 24.88
S156	19195	67989	36.65	p	37.19 36.10
S157	19148	67985	6.34	m	6.48 6.39
S158	19173	67962	14.16	p	14.22 14.11
S159	19354	68437	14.13	m	14.13 14.14
S160	19360	68455	13.52	g	13.39 13.39 13.66 13.55 13.66 13.49
S161	19366	68464	16.22	m	16.44 16.57 16.23 15.63
S162	19356	68474	7.78	m	7.71 7.85
S163	19348	68462	9.81	m	10.08 9.73 9.73 9.70
S164	19347	68468	5.10	m	5.05 5.15
S165	19343	68462	7.47	p	7.60 7.33
S166	19238	67968	24.79	p	24.78 24.81
S167	19239	67987	32.15	m	32.14 32.17
S168	19568	68950	9.72	m	9.81 9.63
S169	19573	68947	15.51	p	15.00 16.03

Feature	Easting	Northing	Mean Altitude	Grade	Point Altitude
S170	19562	68938	10.59	p	10.72 10.46
S171	20159	69963	7.41	m	7.17 7.65
S172	20155	69963	8.71	p	8.71
S173	20147	69970	13.36	m	13.60 13.12
S174	20150	69966	11.12	m	11.16 11.07 11.12 11.14
S175	20163	69975	10.62	p	10.59 10.66
S176	20172	69987	12.94	m	13.17 12.78 12.70 13.10
S177	20183	70002	14.82	p	14.97 14.67
S178	20188	69989	7.75	m	7.65 7.84
S179	20188	69986	4.96	p	4.96 4.96
S180	20821	70667	11.53	m	11.64 11.36 11.60
S181	20825	70738	13.33	m	13.57 13.45 13.37 13.20 13.07
S182	20857	70727	8.00	m	7.95 8.10 7.96
S183	20813	70630	3.91	m	4.02 3.79
S185	20715	70480	6.00	p	5.97 6.00 6.03
S186	20692	70471	12.85	p	12.69 13.02
S187	20700	70496	9.11	m	9.18 9.11 9.05
S188	20700	70496	11.06	g	11.05 11.10 11.04

Feature	Easting	Northing	Mean Altitude	Grade	Point Altitude
S189	20727	70519	11.58	p	11.65 11.75 11.40
S190	20932	70823	7.02	m	7.02 7.23 6.84
S191	20916	70820	14.22	m	14.42 14.23 14.02
S192	19799	69642	5.61	p	5.72 5.62 5.49
S193	19793	69639	9.63	m	9.51 9.76
S194	19782	69640	16.45	p	16.05 16.89
S195	19752	69589	17.52	p	17.70 17.33
S196	19761	69597	11.40	m	11.71 11.50 11.34 11.12 11.32
S197	19768	69600	4.26	p	4.17 4.35
S198	21911	68421	9.06	m	8.93 9.18
S199	21914	68448	9.17	m	9.18 9.08 9.24
S200	21917	68442	4.11	g	3.72 4.01 4.23 4.27 4.38 4.02
S201	21897	68455	37.99	g	37.33 37.45 38.20
S202	21902	68418	36.05	p	35.65 36.46
S203	21915	68365	13.52	p	13.54 13.50
S204	21925	68358	5.17	m	5.29 5.13 5.09

Feature	Easting	Northing	Mean Altitude	Grade	Point Altitude
S205	21927	68346	10.14	m	10.44 10.12 9.85
S206	21930	68339	8.57	m	8.56 8.58
S207	21933	68336	5.15	p	5.14 5.16
S208	21906	68343	41.67	p	41.67
S209	21908	68356	40.90	m	41.11 40.69
S210	21931	68485	4.71	m	4.72 4.70
S211	21932	68475	4.99	m	4.93 5.05
S212	21917	68482	13.46	p	13.46
S213	21917	68526	10.50	p	10.83 10.18
S214	21912	68523	25.49	m	25.50 25.48
S215	21903	68526	36.52	m	35.57 36.41 37.57
S216	21902	68505	38.73	m	38.29 38.94 38.96
S217	21898	68483	41.47	g	41.71 41.23
S218	21911	68479	30.04	p	30.04
S219	21896	68644	12.20	m	12.38 12.08 12.07 12.28
S220	21891	68660	11.10	m	11.22 10.93 11.14
S221	21875	68723	13.30	g	12.06 12.96 13.74 13.64 13.58 13.82
S222	21879	68735	13.43	p	13.11 13.75
S223	21884	68728	9.00	m	9.04 9.25 8.92 8.77

Feature	Easting	Northing	Mean Altitude	Grade	Point Altitude
S225	21866	68760	10.98	m	10.86 11.10
S226	21863	68766	10.37	m	10.43 10.31
S227	21859	68760	14.19	m	14.70 14.38 13.50
S228	21894	68691	6.01	m	6.18 6.04 5.82
S229	21891	68688	10.98	m	10.74 11.22
S230	21878	68704	13.32	m	13.44 13.21
S231	21797	67976	3.81	m	3.73 3.75 3.94
S232	21805	67967	8.87	g	9.04 8.80 8.74 8.63 9.15
S233	21767	67968	10.62	m	10.47 10.52 10.86
S234	21770	67974	7.53	p	7.35 7.71
S235	21778	67983	2.76	p	2.71 2.81
S236	20035	67844	19.63	p	19.46 19.77
S237	20026	67844	36.97	m	36.89 37.04
S238	20629	67183	9.57	m	9.41 9.72
S239	20634	67176	3.97	m	3.89 4.00 4.01
S240	20597	67187	12.36	m	12.43 12.29
S241	20604	67156	3.79	p	3.80 3.78
S242	19301	66924	3.84	m	3.70 3.77 4.06

Feature	Easting	Northing	Mean Altitude	Grade	Point Altitude
S243	19345	66913	31.72	p	31.99 31.45
S244	19331	66901	28.87	m	28.94 28.80
S245	20037	68237	34.67	m	34.69 34.45 34.89
S246	20062	68236	35.31	m	35.21 35.59 35.75
S247	20186	68066	37.02	g	37.14 36.63 37.30
S248	19949	68288	37.43	g	37.90 36.96
S249	20114	67955	2.63	m	2.78 2.48
S250	20119	67945	3.11	m	3.21 3.01
S251	20119	67933	2.68	p	2.74 2.65
S252	20182	68091	33.99	m	34.96 35.98 35.03 34.44 33.09 32.17 32.13
S253	20169	68079	24.22	m	24.32 24.12
S254	20126	67945	8.63	m	9.12 8.34 8.43
S255	20140	67991	4.71	g	4.65 4.58 4.64 4.69 4.99
S256	20131	67979	3.91	g	3.80 3.87 4.14 3.83
S257	20154	68024	4.64	g	4.25 4.27 4.41 4.56 4.82 4.73 4.79 4.96 5.02

Feature	Easting	Northing	Mean Altitude	Grade	Point Altitude
S258	20127	68041	2.33	m	2.37 2.27 2.30 2.37
S259	20151	68077	6.11	m	6.13 5.78 5.86 6.52 6.35 6.03
S260	20149	68145	6.68	m	6.92 6.85 6.27
S261	20152	68133	6.82	m	6.87 6.76
S262	20100	68216	32.25	m	31.86 32.42 32.47
S263	20205	68155	37.10	p	37.59 36.62
S264	20198	68152	29.51	p	29.42 29.60
S265	19242	67946	10.94	m	10.97 10.92
S266	19312	67931	23.53	g	23.56 23.64 23.27 23.26 23.92
S267	19306	67943	25.48	p	24.92 26.04
S268	19349	67926	35.17	g	34.47 35.47 35.56
S269	19351	67895	31.68	g	30.57 31.88 32.60
S270	19273	67840	35.45	m	35.55 35.34
S271	19267	67880	11.67	g	11.38 11.88 11.26 11.91 11.95
S272	21981	70164	9.57	m	9.54 9.90 9.26

Feature	Easting	Northing	Mean Altitude	Grade	Point Altitude
S273	21915	71294	6.09	m	5.93 5.87 5.71 6.29 6.32 6.24 6.30
S274	21904	71279	4.03	m	4.22 4.01 3.87
S275	21866	71281	13.57	p	13.43 13.71
S276	21973	71285	12.92	p	13.47 12.37
S277	19253	68395	12.81	g	12.49 12.36 12.36 12.95 12.46 12.24 12.34 12.51 12.60 13.05 12.87 12.85 12.62 12.80 13.25 13.62 13.29 13.97
S278	19260	68354	16.37	m	15.67 16.27 16.88 16.65
S279	19257	68361	14.74	m	14.35 15.21 14.87 14.55
S280	21769	71068	8.40	g	8.48 8.28 8.35 8.50
S281	22727	70301	13.49	g	14.75 13.25 13.00 13.24 13.22

Feature	Easting	Northing	Mean Altitude	Grade	Point Altitude
S282	20051	68156	18.34	p	18.19 18.48
S283	20054	68143	21.47	p	21.47
S284	20067	68106	6.65	g	7.79 7.10 6.39 6.49 6.65 6.57 6.13 6.23 6.48
S285	20085	68102	3.37	g	3.46 3.39 3.24 3.33 3.41
S286	20062	68202	17.68	p	17.74 17.62
S287	21131	71062	12.30	m	12.33 12.26
S288	19219	69007	14.06	m	14.21 13.91
S289	19348	69146	13.13	p	13.19 13.08
S290	19358	69133	16.38	p	16.35 16.40
S291	19361	69118	16.02	p	15.95 16.10
S292	19573	66883	21.29	p	21.32 21.25
S293	19615	66835	28.48	m	28.23 28.53 28.67
S294	19461	66898	36.07	m	35.74 36.37
S295	21385	68208	38.94	m	38.43 39.46
S296	21477	68219	8.19	m	7.99 8.19 8.35 8.24
S297	21650	67949	39.35	g	39.60 39.44 39.66 39.64 39.21 39.12 38.81

Feature	Easting	Northing	Mean Altitude	Grade	Point Altitude
S300	19885	69719	3.77	m	3.88 3.66
S301	19872	69747	13.38	g	13.18 13.38 13.57
S302	19772	69616	20.77	m	20.81 20.73
S303	19701	69414	10.17	m	10.34 10.16 9.96 10.19
S304	19721	69417	9.48	m	9.70 9.25
S305	19693	69399	8.12	p	8.09 8.15
S306	19686	69393	4.97	m	5.43 4.78 4.67
S307	19705	69393	5.18	p	5.44 5.07 5.00
S309	19245	69086	16.32	p	16.32
S310	19248	69086	6.16	p	6.49 5.96 6.04
S311	19254	69092	3.95	m	3.83 4.06
S312	19266	69129	14.86	m	15.23 14.49
S313	19275	69137	19.32	m	19.04 19.60
S314	19292	69127	10.63	m	10.38 10.88 10.61
S315	19297	69118	7.08	m	7.02 7.13
S316	19296	69112	4.10	g	4.08 4.11
S317	19165	68656	9.39	m	9.46 9.34
S318	18924	68482	12.37	g	12.49 12.26
S319	18917	68476	8.59	m	8.86 8.32

Feature	Easting	Northing	Mean Altitude	Grade	Point Altitude
S320	18780	68449	7.06	m	7.50 6.95 6.72
S321	18638	68741	6.57	m	6.36 6.79
S322	18637	68747	4.04	m	4.10 3.98
S323	18598	68870	10.71	g	10.93 10.79 10.62 10.50
S324	19608	66981	36.02	p	35.92 36.12
S325	19607	66966	34.74	p	35.01 34.48
S326	19618	66823	36.78	m	36.62 36.94
S327	19611	66811	35.98	p	36.42 35.54
S328	19571	66890	32.21	m	32.06 32.35
S329	19615	66835	35.28	m	35.34 35.23
S330	19438	66701	10.44	m	10.11 10.77
S331	19445	66701	19.19	m	18.39 18.77 20.41
S332	19458	66790	31.39	m	31.38 31.40
S333	19416	66826	26.00	p	26.07 25.92
S334	19342	66836	30.35	m	30.33 30.36
S335	19330	66811	21.99	m	22.38 22.36 21.80 21.41
S336	19325	66771	12.62	m	12.64 12.59
S337	19280	66972	5.42	m	5.61 5.24
S338	19275	66978	3.40	m	3.50 3.30

Feature	Easting	Northing	Mean Altitude	Grade	Point Altitude
S339	19282	67000	7.56	m	7.31 7.82
S340	19295	66984	6.89	m	6.94 6.84
S341	19302	66986	11.58	p	11.45 11.71
S342	19309	66986	23.55	p	23.55
S343	19908	66394	12.17	m	12.35 11.99
S344	19900	66435	23.82	g	23.95 24.02 24.10 23.19
S345	19818	66541	27.69	g	27.54 27.57 27.56 27.78 28.00
S346	19841	66530	30.64	p	30.55 30.73
S347	19857	66520	32.14	m	32.10 32.18
S348	19796	66563	34.08	p	34.07 34.09
S349	19779	66546	22.44	g	22.98 22.62 22.43 22.14 22.04
S350	19907	66558	16.31	g	16.37 16.24
S351	19936	66510	30.64	g	30.82 30.62 30.90 30.67 30.20
S352	19881	66559	20.98	m	20.79 21.06 21.08
S353	19890	66565	17.85	p	17.95 17.76
S354	19917	66592	6.65	p	6.71 6.57
S355	19921	66564	6.85	p	6.20 6.95 7.40

Feature	Easting	Northing	Mean Altitude	Grade	Point Altitude
S356	19926	66560	4.37	m	4.28 4.44
S357	19865	66765	34.78	m	34.97 34.60
S358	19652	61156	2.72	m	2.85 2.76 2.54
S359	20833	67777	4.26	p	4.31 4.20
S360	20830	67786	8.18	p	7.96 8.39
S361	20877	67676	12.50	m	12.47 12.53
S362	20921	67488	6.14	m	6.20 6.09
S363	20917	67491	3.53	p	3.67 3.39
S364	20912	67494	3.57	m	3.60 3.54
S365	20911	67510	5.87	m	5.77 5.98
S366	20920	67507	8.39	g	8.52 8.34 8.31
S367	20933	67521	17.78	g	17.56 17.99
S368	20918	67544	24.84	g	25.26 24.43
S369	20925	67547	31.22	p	31.08 31.36
S370	20936	67552	39.48	g	39.66 39.30
S371	20926	67488	16.58	p	16.58
S372	20938	67478	36.23	p	36.25 36.20
S373	20917	67398	11.71	m	11.65 11.76
S374	20932	67385	32.39	p	32.53 32.26
S375	20930	67351	9.27	m	9.38 9.16

Feature	Easting	Northing	Mean Altitude	Grade	Point Altitude
S376	20989	66977	8.51	g	9.27
					8.87
					8.48
					8.38
					8.34
					8.21
					8.23
					8.34
S377	20985	66900	13.51	g	14.01
					13.66
					13.55
					13.30
					13.04
S378	20984	66906	7.55	m	7.62
					7.54
					7.48
S379	20980	66974	4.49	m	4.30
					4.32
					4.61
					4.51
					4.56
					4.48
					4.66
S380	21134	66847	38.11	g	37.85
					38.09
					38.31
					38.19
S381	21115	66835	10.05	m	10.12
					10.05
					9.97
					10.06
S382	21214	66760	4.97	g	4.99
					5.02
					4.90
S383	21218	66766	6.54	m	6.74
					6.39
					6.61
S384	21269	66779	6.08	g	6.24
					5.97
					6.02
S385	21299	66796	22.87	m	22.93
					22.81
S386	21316	66789	13.04	m	13.42
					12.83
					12.88
S387	21335	66767	6.60	p	6.45
					6.58
					6.78
					6.73

Feature	Easting	Northing	Mean Altitude	Grade	Point Altitude
S388	21555	67200	39.00	p	38.77 39.22
S389	21408	68185	39.17	p	39.17
S390	21431	68165	39.53	m	39.20 39.84
S391	21444	68165	38.42	p	38.42
S392	21492	68200	4.38	g	4.52 4.53 4.10 4.40
S393	21837	68843	10.26	p	9.48 9.72 10.09 10.93 11.07
S394	21626	68005	36.43	m	36.56 36.30
S395	21619	68003	38.26	p	38.19 38.33
S396	21627	67990	38.03	m	38.27 37.80
S397	21702	67804	39.68	g	39.66 39.71
S398	21712	67794	32.12	m	32.27 31.97
S399	21541	68235	3.42	g	3.42 3.37 3.43 3.45 3.41

(b) Shingle Ridges

Feature	Easting	Northing	Mean Altitude	Point Altitude
SR1	21013	70946	5.15	4.56
				4.83
				5.33
				5.49
				5.42
				5.25
				5.15
SR2	22041	69329	3.01	2.91
				2.95
				3.16
SR3	22037	69335	3.07	3.04
				3.10
SR4	19829	69263	3.60	3.70
				3.50
SR5	21780	71089	3.91	4.13
				4.25
				3.95
				3.77
				3.54
				3.81
				4.02
				3.83
SR6	21789	71098	4.07	4.05
				4.09
SR7	19263	67886	8.56	8.61
				8.70
				8.38
SR8	19258	67881	7.80	7.93
				7.67
SR9	19244	67890	4.34	3.84
				4.85
SR10	19236	67888	3.84	3.88
				4.03
				3.92
				3.54
SR11	19165	67727	13.12	13.53
				13.09
				13.04
				12.65
				13.03
				13.38
SR12	19211	67926	3.67	3.31
				3.70
				3.86
				3.57
				3.79
				3.71
				3.61
				3.57
				3.93
				3.66
				3.61

Feature	Easting	Northing	Mean Altitude	Point Altitude
SR13	19213	67951	5.47	5.50 5.75 5.15
SR14	19223	67941	6.67	7.71 6.76 7.41 6.56 6.06 6.02 6.20
SR15	19229	67925	6.00	6.06 5.94
SR16	19224	67926	3.57	3.66 3.34 3.69
SR17	19266	67945	3.86	3.80 3.92
SR18	20855	71709	3.86	4.24 4.42 4.26 3.41 3.49 3.48 3.70
SR20	20922	70820	13.08	12.76 13.24 13.42 12.90
SR21	20933	70841	13.16	13.54 13.30 12.64
SR22	21923	68476	11.67	11.04 11.22 11.62 12.80
SR23	21926	68488	10.79	10.89 10.68
SR24	21921	68498	13.18	13.49 13.24 12.82 12.98 13.38
SR300	19699	69421	12.32	12.13 12.56 12.29 12.32 12.28

Feature	Easting	Northing	Mean Altitude	Point Altitude
SR301	18781	68468	12.23	12.47
				12.23
				12.00
SR302	20082	69536	3.37	3.25
				3.60
				3.44
				3.47
				3.28
				3.16
SR303	19322	66818	27.90	28.71
				28.15
				27.57
				27.16
SR304	19894	66388	4.21	3.96
				4.29
				4.13
				4.16
				4.51
SR305	19911	66527	32.93	32.83
				33.03
SR306	19669	61149	3.47	3.45
				3.50
				3.46
SR307	20898	67529	3.39	3.42
				3.16
				3.48
				3.37
				3.49
SR308	20906	67516	2.94	2.99
				2.89
SR309	21094	66808	4.44	4.58
				4.30
SR310	21103	66814	6.44	6.54
				6.52
				6.39
				6.39
				6.43
				6.35
SR311	21092	66839	9.87	9.97
				9.74
SR312	21084	66858	12.54	12.66
				12.53
				12.48
				12.50
SR313	21093	66864	13.47	13.60
				13.34
SR314	21023	66858	13.56	13.39
				13.63
				13.68
				13.52

Feature	Easting	Northing	Mean Altitude	Point Altitude
SR315	21011	66849	9.48	9.58 9.48 9.39
SR316	21014	66843	7.89	7.67 8.12
SR317	21014	66833	7.96	7.83 8.08
SR318	21098	66805	3.77	3.69 3.84 3.81 3.75
SR319	21268	66757	5.85	5.60 6.11
SR320	21321	66780	7.67	8.18 7.16
SR321	21353	66757	5.39	5.24 5.58
SR322	21462	68201	12.42	12.52 12.38 12.39
SR323	21515	68292	6.70	6.57 6.68 6.72 6.63 7.08 6.93 6.70 6.36 6.64
SR324	21470	68216	10.89	10.90 10.92 10.92 10.83
SR325	21477	68207	8.57	8.77 8.35
SR326	21485	68206	8.09	7.84 8.13 8.17 8.14 8.24 8.02
SR327	21489	68206	7.67	7.63 7.83 7.75 7.71 7.46

Feature	Easting	Northing	Mean Altitude	Point Altitude
SR329	21886	68861	8.87	8.71 9.03
SR331	21529	68251	5.56	5.72 5.69 5.49 5.33
SR332	21536	68241	5.27	5.51 5.37 5.23 5.10 5.15

(c) Rock Platforms

Feature	Easting	Northing	Mean Altitude	Point Altitude
RP1	21224	70909	2.19	2.46 1.75 2.35
RP2	21127	70913	1.75	1.91 1.59
RP3	21147	70881	1.86	2.02 1.77 1.94 1.71
RP4	22095	69707	10.48	10.04 10.92
RP5	19180	67745	7.33	7.23 7.48 7.29
RP6	19131	67670	7.24	7.44 7.04
RP7	19135	67998	7.12	6.70 7.53
RP8	19154	67963	6.70	7.03 7.13 5.95
RP11	20772	70561	1.85	1.96 1.70 1.83 1.89 1.85 1.85
RP12	20711	70291	2.25	1.91 2.60
RP13	20702	70279	1.93	2.13 1.73
RP14	20682	70255	2.02	2.17 1.88
RP15	21926	68048	1.51	2.01 1.85 1.38 1.83 0.87 1.17 1.43
RP16	19306	66887	5.80	5.77 6.00 5.38 5.59 6.27

Feature	Easting	Northing	Mean Altitude	Point Altitude
RP17	19285	66842	5.64	5.77 5.50
RP18	19267	66799	5.96	6.17 6.14 6.28 5.61 5.60
RP301	19149	68601	8.27	8.38 8.14
RP302	18776	68428	7.93	8.07 7.80
RP303	21043	70960	1.79	1.72 1.85
RP304	19553	66659	5.76	5.61 5.92
RP305	19535	66656	6.12	6.07 6.17
RP306	19928	66371	6.35	6.45 6.25
RP307	20753	70023	10.62	10.98 10.90 9.98

2. Outwash and River Terrace Altitudes

T1
4.80
5.08
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8.07
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T5
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T6
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T7
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T8
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T14a
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T397
14.50
14.33
13.79
13.82
13.74
13.27

T398
11.82
11.50
11.52
11.31
11.00
10.07
9.80
9.64
9.65

T399
11.75
11.47

T400
12.26
12.07
12.26

T401
11.44
11.59
11.97

T402
7.83
7.51
7.28

T403
3.38
3.08
3.10
2.87
2.90
3.09
2.93
2.59
2.36
2.77
2.35
2.53
2.37

T404
22.00
22.38

T405
22.03
20.39
19.75

T406
19.38
18.81
18.86
17.98

T407
14.04
13.75
14.43

T408
11.39
11.07
10.76
10.46
10.44
9.53
9.60
9.29
8.96
9.08
9.00
8.73
8.60
8.51
8.35
8.19
8.22
8.07
8.03
8.27

T408a
5.50
6.40
6.72
6.25

T409
4.55
4.62
4.79
4.77
4.96
5.16
5.37
5.48
5.59
5.73
6.00
5.86

T410
9.05
9.87
9.42

T411
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7.67

T412
9.03
9.48
9.81
9.88
10.05
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10.38
10.60
10.77
11.03
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12.16
12.37
12.67
12.61
12.96
14.08
14.42
14.52
14.89

T413
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14.35

T414
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13.10
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12.11
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12.19
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11.37
11.11
10.81

T416
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37.44

T417
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28.08

T418
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26.83
28.29

T419
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40.74
40.28

T420
38.18
38.96
39.87

T421
34.60
33.76

T422
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12.43
13.00
13.66
14.04
14.43
14.85
15.37

T423
16.75
16.14
16.04

T424
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2.49
2.61
2.63
2.78
2.89
3.13
3.22

T425
5.15
5.48

T426
4.03
4.07

T427
3.43
3.69
3.77
3.96
3.73
3.87
3.96
4.18
4.33
4.50

T428
4.59
4.98
5.13

T429
6.80
6.52
6.61
6.43
6.52
5.76
5.86
6.29
6.11
6.14
6.41
6.48
6.76
7.21

T430
5.72
5.50
5.43
5.38

T431
6.04
6.46
6.20
6.83
6.44
6.66
6.92

T432
6.26
6.17

T433
6.11
6.22
5.82

T434
5.49
5.30

T435
11.02
11.52
11.76

T436
11.28
11.62
11.66
11.84
12.44

T437
13.68
13.96
14.16
14.25
14.41

T438
21.50
22.43
23.42

T439
27.75
28.27
28.69
29.30

T440
29.64
29.06
28.66

T441
31.21
32.81
33.54
34.15

T443
4.24
4.59
4.96
5.24
4.35
4.45
4.73
5.22
5.31

T444
5.08
5.68
6.02
6.19

T445
6.63
7.85
8.08
8.50
8.71
9.60

T446
10.37
10.35
10.19
10.20
10.23

C401
4.28
4.86
4.94
5.23
5.05
5.38
5.75
5.92
6.43

T447
7.90
8.28

T448
8.98
8.95
8.87

T449
9.04
8.96
9.06

T450
10.52
10.46
9.94

T451
11.09
11.18

T452
13.31
13.41
13.44

T453
12.23

T454
9.80
9.25

T455
19.71

T456
14.79
15.16
15.71
16.05

T457
24.22
23.95

T459
39.23
37.68
38.68
39.22

T460
4.18
4.08
3.90

T461
4.69
4.68

3. Present Sea Level Indicators

Note: Figures underlined are mean values for the locality.

Locality	Easting	Northing	Lower Limit Vegetation	Upper Limit Drift	Upper Limit Barnacles	Upper Limit Shingle
PSL1	21839	71229	2.37	3.19		
			2.32	3.19		
			1.70	2.62		
			2.13	3.00		
PSL2	21700	71117	2.28	3.75		
			2.26	3.60		
			2.27	3.09		
			2.25	3.47		
			2.26	3.48		
PSL3	21224	70912			1.27	
					1.17	
					1.22	
PSL4	21138	70882			-0.19	
					1.31	
					0.56	
PSL5	22940	70475	1.72	2.41		
			1.68	2.99		
			1.42	2.37		
			1.84	2.75		
			1.66	2.63		
PSL6	22095	69800	1.89	3.19		
			2.22	2.69		
			2.10	-		
			2.07	2.94		
PSL7	20784	67355	2.63	3.20		
			2.16	3.09		
			2.23	3.32		
			2.16	3.22		
			2.61	3.24		
			2.36	3.21		
PSL8	20727	67658	2.45	3.17	1.27	2.83
			2.83	3.07	1.13	2.80
			2.80	3.05	1.14	2.96
			2.25	2.72	1.21	3.03
			-	-	1.28	3.22
			-	-	1.26	-
			-	-	1.22	-
			-	-	1.16	-
			-	-	1.00	-
			2.58	3.00	1.19	2.97
PSL9	20579	67190	2.68	3.25	0.81	
			2.68	3.21	0.95	
			2.62	3.15	1.19	
			2.56	3.10	1.12	
			2.61	3.17	1.10	
			2.63	3.17	1.03	

Locality	Easting	Northing	Vegetation	Drift	Barnacles	Shingle
PSL10	21768	71087	2.14	3.34	0.73	
			2.12	3.63	0.80	
			2.17	3.82	0.25	
			2.14	3.60	0.60	
PSL11	19260	68435	3.02	3.92	1.07	
			2.99	3.93	1.04	
			-	-	1.05	
			-	-	1.17	
			-	-	1.28	
			3.00	3.93	1.12	
PSL12	19181	67763	2.03	3.48	1.30	
			2.22	3.33	1.24	
			1.76	3.03	1.28	
			1.79	3.16	1.28	
			2.12	3.31	1.21	
			1.98	3.26	1.26	
PSL13	19193	67946	2.59	3.51	1.34	
			2.84	3.53	1.39	
			2.79	3.47	-	
			2.63	3.59	-	
			1.97	2.68	-	
			2.56	3.36	1.37	
PSL14	19487	68752	2.87	3.43	1.33	
			2.70	3.34	1.28	
			2.33	3.65	1.31	
			2.43	2.97	1.38	
			2.34	-	1.26	
			2.53	3.35	1.31	
PSL15	19561	68953	1.84	2.59	1.27	
			2.40	2.62	1.22	
			2.22	2.57	1.33	
			2.12	-	1.28	
			2.15	2.59	1.28	
PSL16	20846	70712	1.58	2.57	1.02	
			2.21	-	1.03	
			-	-	1.12	
			1.89	2.57	1.06	
PSL17	20785	70579	1.81	2.79	1.17	
			2.28	3.20	1.25	
			2.40	3.13	1.21	
			2.50	-	1.20	
			2.25	3.04	1.21	
PSL18	20729	70340	2.22	3.04	1.30	
			2.72	3.29	1.21	
			-	-	1.30	
			-	-	1.35	
			2.47	3.17	1.29	

Locality	Easting	Northing	Vegetation	Drift	Barnacles	Shingle
PSL19	19788	69624	2.56	3.30	1.29	
			2.85	2.84	1.41	
			2.14	3.12	1.26	
			2.10	-	1.36	
			2.41	3.09	1.33	
PSL20	21900	68053				4.18
						4.18
PSL21	21939	68320	2.80	3.23	1.41	3.23
			2.82	3.36	1.44	3.36
			2.88	3.23	1.20	3.23
			-	3.26	1.07	-
			2.83	3.27	1.28	3.27
PSL22	21875	68778	2.48	3.06	1.29	2.80
			2.35	3.10	1.30	2.95
			2.44	-	1.28	3.06
			2.57	-	1.25	-
			-	-	1.41	-
			2.46	3.08	1.31	2.94
PSL300	19890	69731	2.44	3.11		
			2.23	3.10		
			2.39	2.98		
			2.23	3.48		
			2.24	3.32		
			2.31	3.20		
PSL302	18914	68442	1.30	3.50	0.90	
			1.80	3.42	1.00	
			1.42	3.26	1.12	
			1.51	3.40	1.01	
PSL303	20052	69559	2.95	3.43	1.26	
			3.04	3.63	1.38	
			2.38	3.19	1.25	
			2.79	3.42	1.29	
PSL304	19546	66653				4.11
						4.11
PSL305	19433	66732	2.47	3.54	1.16	
			2.41	3.35	1.35	
			2.43	3.47	1.19	
			1.69	3.34	1.15	
			1.94	3.22	1.15	
			2.19	3.38	1.20	
PSL306	19878	66755	2.45	3.30	1.17	
			2.87	3.53	1.40	
			2.08	3.44	1.35	
			2.56	-	1.10	
			2.49	3.42	1.25	
PSL307	21221	66753	3.23	3.84	1.23	3.84
			3.34	3.86	1.34	3.86
			3.13	3.70	1.19	3.70
			3.29	4.31	1.12	4.31
			3.31	4.33	-	4.33
			3.26	4.01	1.22	4.01

4. Glacial landforms

The following short list is not inclusive of all glacial deposits surveyed, a number of kame terraces and outwash terraces having been recorded in Part 2 of this Appendix.

M1 (Loch Shira)

12.98	13.28	13.84	13.97	14.54
15.01	15.16			

M2 (Cuilmuich, Loch Goil)

12.89	12.80	12.34	12.77

K1 (Otter Ferry)

22.36

M300 (Millhouse, nr. Portavadie)

29.69

M301 (Millhouse)

33.07

M302 (Millhouse)

35.31

M310 (Glen Finart, nr. Ardentinny)

13.27	13.38

M311 (Glen Finart)

11.76	12.11	11.67	11.81

5. Cross-profiles

(a) Profiles of the present shore

Profile 1

0.51	1.09	1.47	1.77	2.29
0.62	1.14	1.53	1.82	2.46
0.74	1.22	1.63	1.90	2.74
0.84	1.31	1.69	2.03	2.96
0.96	1.39	1.76	2.17	3.40

Profile 2

1.02	1.29	1.50	1.67	2.46
1.08	1.31	1.52	1.72	2.57
1.09	1.34	1.55	1.83	2.68
1.12	1.35	1.59	1.91	2.83
1.15	1.38	1.62	2.13	2.98
1.19	1.43	1.64	2.29	3.12
1.27	1.46	1.66	2.44	3.02

Profile 3

1.22	1.37	1.47	1.94	3.29
1.23	1.39	1.50	2.14	3.59
1.24	1.41	1.54	2.27	3.67
1.27	1.43	1.60	2.68	
1.30	1.46	1.66	2.96	
1.31	1.47	1.76	3.09	

Profile 4

0.79	0.93	1.10	1.27	1.65
0.85	0.92	1.13	1.30	1.80
0.91	0.95	1.14	1.39	2.03
0.90	0.99	1.18	1.46	2.12
0.93	1.05	1.20	1.55	2.34

Profile 5

0.67	1.14	1.44	2.01	3.08
0.77	1.19	1.50	2.14	3.41
0.89	1.27	1.55	2.39	
1.02	1.35	1.66	2.55	
1.06	1.42	1.79	2.87	

Profile 6

0.86	1.01	1.13	1.39	2.32
0.90	1.03	1.17	1.54	2.36
0.93	1.05	1.20	1.65	2.50
0.95	1.06	1.23	1.81	2.67
0.98	1.09	1.27	1.97	2.91
1.00	1.11	1.33	2.10	

Profile 7

0.83	1.19	1.49	1.99	2.96
0.88	1.23	1.51	2.05	3.16
0.94	1.28	1.61	2.23	3.36
1.01	1.38	1.68	2.43	
1.07	1.41	1.78	2.60	
1.11	1.45	1.96	2.79	

Profile 8

1.03	1.34	1.75	2.13	2.21
1.05	1.35	1.76	2.18	2.39
1.13	1.43	1.79	2.13	2.53
1.15	1.49	1.85	2.14	2.69
1.18	1.54	1.91	2.15	2.84
1.25	1.59	1.98	2.15	3.06
1.29	1.64	2.10	2.19	3.36

Profile 9

1.00	1.37	1.54	2.16	3.21
1.07	1.40	1.57	2.32	3.08
1.16	1.42	1.60	2.50	3.28
1.22	1.46	1.63	2.65	
1.26	1.48	1.70	2.86	
1.30	1.49	1.84	3.12	
1.34	1.52	1.96	3.40	

Profile 10

1.06	1.27	1.46	1.65	2.65
1.11	1.32	1.49	1.76	2.93
1.16	1.38	1.53	2.01	
1.22	1.40	1.59	2.35	

(b) Profiles of raised shorelines

Profile A-A[1]

-0.24	3.44	5.60	7.72	10.97
0.05	3.52	5.67	7.87	11.46
0.31	3.62	5.74	8.02	11.91
0.47	3.74	5.82	8.24	12.22
0.86	3.84	5.88	8.35	12.38
0.95	4.01	5.98	8.45	12.44
1.16	4.10	6.09	8.67	12.53
1.39	4.24	6.19	9.06	12.62
1.58	4.35	6.29	9.29	12.74
1.66	4.47	6.42	9.47	12.86
1.82	4.60	6.52	9.70	13.01
2.20	4.65	6.65	9.90	13.17
2.45	4.76	6.76	10.00	13.53
2.84	4.81	6.85	10.10	13.93
3.09	4.88	6.96	10.20	14.37
3.20	4.98	7.03	10.32	14.75
3.29	5.08	7.12	10.49	
3.30	5.21	7.28	10.57	
3.32	5.35	7.45	10.60	
3.40	5.49	7.57	10.69	

Profile B-B[1]

1.41	2.14	2.73	5.07	9.51
1.83	2.13	2.61	5.08	9.68
2.00	2.09	2.60	5.12	9.84
2.05	2.12	2.60	5.18	9.97
2.02	2.27	2.58	5.28	10.10
2.02	2.54	2.60	5.41	10.23
2.03	2.54	2.65	5.58	10.36
2.00	2.50	2.75	5.83	10.45
1.99	2.42	2.87	6.15	10.59
2.02	2.39	3.03	6.49	10.72
2.07	2.38	3.22	6.91	10.88
2.10	2.39	3.53	7.24	11.08
2.16	2.32	3.81	7.60	11.41
2.14	2.31	4.09	7.95	11.75
2.17	2.38	4.34	8.29	12.10
2.24	2.45	4.57	8.60	
2.19	2.58	4.85	8.88	
2.19	2.68	4.95	9.13	
2.17	2.73	5.03	9.34	

Profile C-C^{1}

1.21	3.80	5.07	7.96	10.36
1.36	3.83	5.19	8.09	10.61
1.51	4.07	5.32	8.19	10.95
1.62	3.99	5.41	8.48	11.23
1.75	4.08	5.56	8.57	11.54
1.90	4.05	5.76	8.69	11.87
2.03	4.07	5.94	8.88	12.02
2.25	4.13	6.19	8.99	12.29
2.41	4.11	6.35	9.14	12.67
2.43	4.21	6.49	9.32	13.03
2.63	4.36	6.77	9.50	
3.03	4.56	6.87	9.67	
3.45	4.65	6.93	9.91	
3.63	4.85	7.37	10.16	

Profile D-D^{1}

0.98	4.71	6.67	9.07	11.48	13.47
1.17	4.76	6.78	9.29	11.60	13.52
1.37	4.83	6.86	9.46	11.67	13.54
1.62	4.89	6.95	9.60	11.80	13.58
1.89	4.94	7.01	9.71	11.87	13.62
2.12	4.95	7.05	9.80	11.98	13.67
4.13	4.94	7.08	9.88	12.06	13.73
4.04	4.94	7.07	9.96	12.14	13.79
4.19	4.96	7.05	10.05	12.20	13.82
4.10	4.99	7.06	10.13	12.31	13.87
4.03	5.04	7.06	10.21	12.41	13.92
4.02	5.09	7.06	10.27	12.52	13.98
4.09	5.11	7.05	10.35	12.61	14.03
4.11	5.12	7.04	10.42	12.71	14.09
4.18	5.16	7.06	10.50	12.79	14.16
4.20	5.24	7.08	10.59	12.88	14.21
4.26	5.32	7.16	10.67	12.97	14.26
4.27	5.40	7.26	10.75	13.04	14.32
4.27	5.49	7.38	10.81	13.14	14.36
4.29	5.58	7.52	10.91	13.18	13.40
4.32	5.66	7.61	10.98	13.25	14.41
4.37	5.75	7.73	11.04	13.32	14.46
4.40	5.88	7.88	11.12	13.35	14.50
4.45	6.01	8.03	11.17	13.35	14.58
4.49	6.14	8.20	11.24	13.37	
4.54	6.28	8.40	11.31	13.39	
4.59	6.44	8.59	11.36	13.41	
4.65	6.55	8.85	11.43	13.43	

Profile E-E[1]

-0.75	3.36	2.59	5.57	5.48	7.29
-0.58	3.34	2.70	5.71	5.48	7.38
-0.35	3.33	2.72	5.76	5.43	7.52
-0.16	3.25	2.75	5.85	5.46	7.75
0.11	3.18	2.83	5.96	5.44	8.00
0.35	3.11	2.83	6.06	5.47	8.23
0.56	3.08	2.89	6.16	5.49	8.43
0.79	3.00	2.95	6.27	5.53	8.53
1.01	2.91	3.01	6.33	5.58	8.64
1.23	2.90	3.07	6.36	5.59	8.68
1.45	2.70	3.16	6.33	5.57	8.85
1.66	2.51	3.20	6.33	5.60	9.01
1.82	2.31	3.26	6.30	5.60	9.14
1.98	2.14	3.37	6.19	5.65	9.28
2.09	2.04	3.38	6.10	5.66	9.41
2.11	1.89	3.39	6.02	5.64	9.54
2.28	1.88	3.41	5.92	5.65	9.75
2.34	1.86	3.42	5.83	5.64	10.04
2.37	1.84	3.46	5.77	5.62	10.23
2.33	1.97	3.48	5.73	5.68	10.38
2.72	2.06	3.51	5.67	5.73	10.64
3.03	2.16	3.57	5.61	5.76	10.80
3.41	2.27	3.72	5.59	6.20	10.95
3.59	2.32	3.93	5.55	6.27	10.97
3.85	2.49	4.11	5.56	6.31	11.26
3.86	2.39	4.28	5.55	6.31	11.32
3.79	2.42	4.49	5.53	6.34	11.41
3.78	2.39	4.76	5.47	6.43	11.60
3.74	2.43	4.87	5.51	6.56	11.89
3.58	2.50	5.06	5.53	6.73	12.15
3.49	2.57	5.28	5.51	6.88	12.75
3.42	2.61	5.44	5.50	7.05	

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