Title	Raised shorelines of Jura, Scarba and NE Islay
Author	Dawson, Alastair G.
Qualification	PhD
Year	1979

copy 3

230

Raised shorelines of Jura, Scarba and NE Islay

Alastair G Dawson

Ph.D.

University of Edinburgh

1979

CONTENTS

FIGURES

TABLES

PLATES

Acknowledgements

I am grateful to the people of Jura without whose help fieldwork could not have been conducted. In particular I am extremely indebted to Dr J Johnson and family for their kind hospitality. I am also grateful to the Astor family (Tarbert), Mr D Darroch and family (Inver) and Mrs I Fairman for their kind assistance regarding accommodation. Deep thanks are also extended to Mr N McInnes, Mr J Mollison, Mr C MacLean, Mr W MacDonald, Mr D Johnson, Mr D Buie, Miss E Astor, Mr D Fairman, Miss C Morton and Mrs F MacIntyre for their assistance during fieldwork. Thanks are also extended to Mr R Grove, Mr M Williams and Mr R Forrester of Rhuvaal lighthouse and to Mrs C Crawford and family for accommodation on Islay.

I am most grateful for levelling assistance to Mr G Kerr, Mr N MacDonald, Mr M MacQueen, Mr W Taylor, Mr G Forsyth, Mr and Mrs I Murray, Miss A MacNaughton, Mr J Thorburn, Mr C MacArthur and Mr D Munro. My deepest thanks for levelling assistance, however, is extended to Mr David Gilmour without whose help much of the levelling would not have been possible.

Financial assistance was provided by the University of Edinburgh and the Carnegie Trust for the Universities of Scotland. Thanks are also extended to Prof. R.U. Cooke who provided the resources of the Dept. of Geography, Bedford College, University of London to enable the writer to complete the thesis. Mr D Garvin and Mr R Halfhide kindly assisted in map preparation and photography while I grateful to Mrs J Hackett for typing the thesis. Special thanks are extended also to Mrs M Dale, Mr C K Ballantyne, Mr R Cornish, Mr D G Sutherland, Mrs M Robinson, Dr J A Matthews, Dr R A Shakesby, Mr T Wain-Hobson and Dr J M Gray for many stimulating discussions and arguments. Thanks are also given to John

Dewar Studios, Edinburgh, for permission to include Plate 4 and to the Ministry of Defence for the inclusion of Plates 5,12, 18, 23 and 29. I am grateful also to the Continental Shelf Unit, Institute of Geological Sciences, Edinburgh for the provision of unpublished seabed data.

Finally, extreme thanks are given to Dr J B Sissons whose continued guidance, patience and constructive suggestions throughout the study made research a pleasure to undertake.

Abstract

A geomorphological investigation is made of the raised shorelines of Jura, Scarba and NE Islay, Scottish Inner Hebrides. The raised shore features consist of various terraces, beach ridges and coastal platforms. Ground levelling of approximately 4,000 points indicates a complex sequence of inter-, late- and postglacial shoreline displacement.

In order to understand more clearly the origin of raised coastal terraces and beach ridges and their altitude relationships to former sea-levels a study of modern coastal landforms was undertaken. Particular attention is paid to the origin and regional altitude distribution of modern beach ridges and the seaweed Pelvetia canaliculatus.

Much attention is devoted to the nature, age and origin of several raised coastal platforms. Consideration of these features is preceded by a discussion on the nature of platform development in polar and non-polar areas. A discussion of strandflat origin is also presented.

Case studies of several prominent glacial landforms in the study area are included. The results of these investigations are combined with a limited study of glacial striae and erratics in a discussion of regional glaciation and deglaciation. The results of the glacial landform studies in conjunction with those obtained from raised shoreline investigations indicate a complex pattern of deglaciation.

Chapter 1

Introduction

During the past two decades a considerable body of information has been obtained on Quaternary environments in Scotland. The studies conducted have been concerned primarily with mainland areas and, to date, few have considered the Quaternary geomorphology of the Inner Hebridean islands of Jura, Scarba and Islay (Fig. 1). In fact, little work has been undertaken in these islands since the investigations of the Geological Survey in the early 20th century. Unfortunately, the Geological Survey memoirs, although comprehensive in their treatment of solid geology, present an incomplete and scanty account of Quaternary landforms and, for certain areas, no information is given.

Following the investigations of the Geological Survey the principal studies of aspects of the geomorphology of Jura, Scarba and NE Islay have been concerned with much larger areas. Thus Charlesworth (1955) included the islands in a treatment of glacier limits in the whole of the Highlands and Islands, McCann (1961, 1964) discussed raised shorelines and ice-limits of NE Islay and W Jura as part of a study involving a large area of western Scotland, while Synge and Stephens (1966) considered NE Islay in a similar investigation that embraced NE Ireland and much of western Scotland.

During the last five years two studies (Gray, 1972, 1978; Sutherland, [unpublished]) have yielded a considerable amount of information on the raised shorelines and ice-limits in Lorn, eastern Mull and the mainland of SW Argyll. In addition, offshore investigations have increased the knowledge of Quaternary evolution in the Sea of the

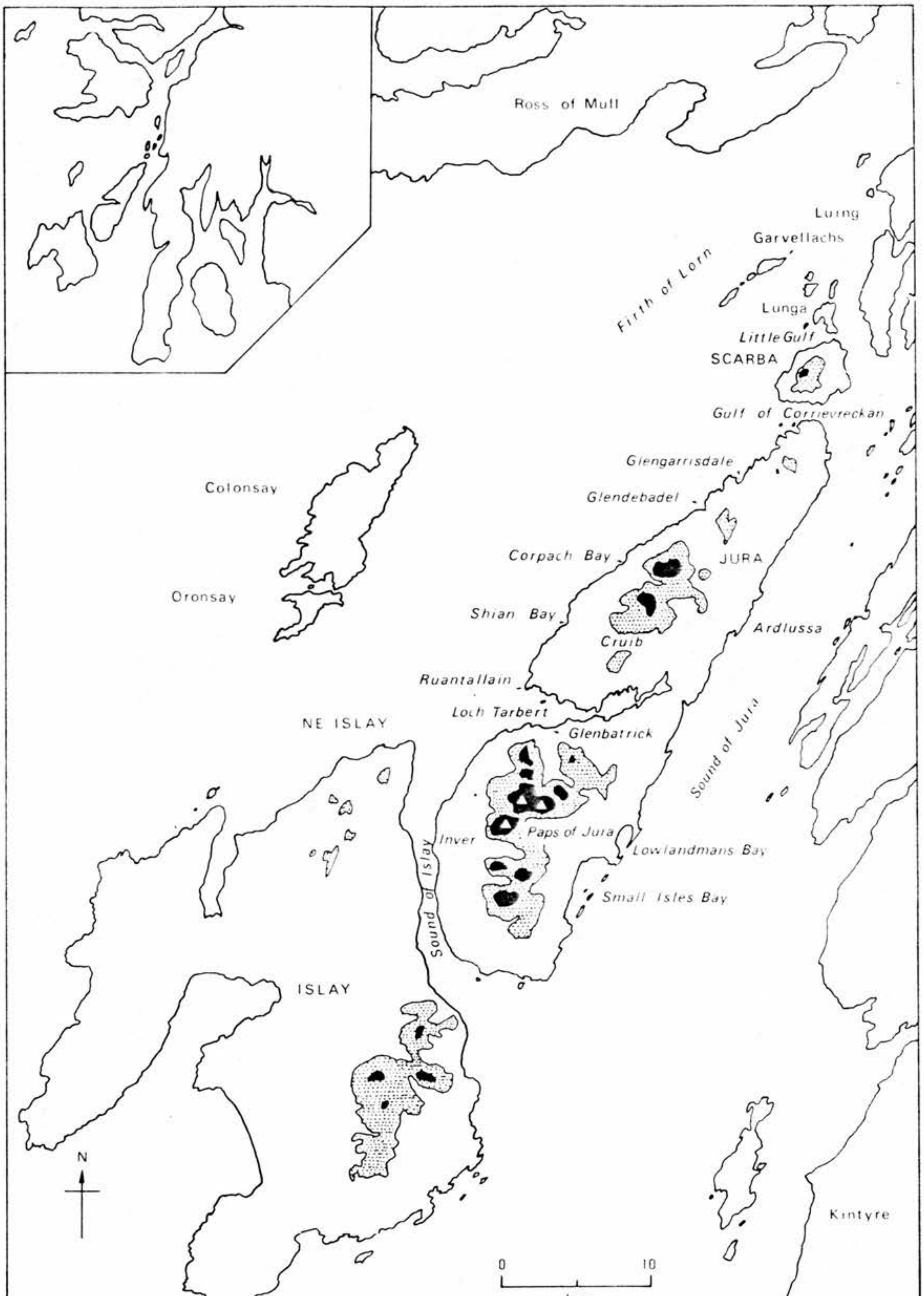


Fig.1 Location of Jura, Scarba and NE Islay. Ground over 200 m (shaded) and over 400 m (black) is also shown.

Hebrides (Binns et al., 1973, 1974), while farther north detailed investigations have been undertaken in NW Argyll (Wain-Hobson (unpublished)), Skye (Sissons, 1977) and in the Outer Hebrides (von Weymarn, 1974; Ritchie, 1966, 1972).

It was decided to investigate the raised shorelines of Jura, Scarba and NE Islay since several visits to these islands and the landform descriptions given by McCann had revealed a plethora of outstanding raised coastal landforms none of which had been investigated in detail. This in turn led to a study of certain glacial and fluvioglacial landforms. During the summer of 1978 the writer also investigated the raised shorelines of Colonsay: several aspects of raised shoreline evolution here are of direct relevance to the interpretation of the raised shorelines in Jura and NE Islay. Where pertinent to the present study, information from Colonsay is included in the text. In addition to contributing detailed information on the area investigated, the conclusions of this study contribute to existing interpretations of regional deglaciation and raised shoreline evolution in western Scotland.

Chapter 2 summarises the geological structure of the islands and includes an introductory description of the coastline. In Chapter 3 an account is given of the techniques and field methods utilised in the study while in Chapter 4 there is a discussion of regional and local glaciation. Chapter 5 presents the results of detailed investigations on the nature and origin of modern coastal landforms. In Chapter 6 the processes responsible for marine erosion in different environments are discussed and criteria for the identification of shore platforms of different ages and origins are assessed. In Chapters 7 and 8 the nature, age and origin of fossil shore platforms in Jura, Scarba and NE Islay are outlined. The pattern of lateglacial and postglacial relative sea-level changes is discussed in Chapters 9 and 10. Finally in Chapter 11, ice-limits and the sequence of raised shoreline formation in Jura, Scarba and NE Islay are summarised.

Chapter 2

Structure, relief and coastal processes; an introduction

Located in the Scottish Inner Hebrides, Jura, Scarba and NE Islay form remnants of a deeply dissected and glacially-eroded Highland tableland. The islands are separated from Colonsay by the Firth of Lorn and from the mainland by the Sound of Jura (Fig. 1) and each is characterised by rugged upland terrain. Jura is bounded to the north by the Gulf of Corrievreckan, which separates it from the small isle of Scarba. The latter, which rises steeply to an altitude of 448 m, is in turn separated from the neighbouring islands of Lunga and Luing to the north by the Little Gulf of Corrievreckan and the Sound of Luing. In the south the narrow Sound of Islay lies between Jura and Islay (Fig. 1).

Jura, Scarba and NE Islay comprise 40,000 ha and are fringed by a coastline 185 km in length. The highest summits occur in SW Jura where the Paps of Jura all exceed 750 m. With the exception of eastern Jura the entire area is treeless and uninhabited with barren summits separated by drift-filled valleys. In places glacial breaches interrupt the continuity of the higher ground, while in Jura alone 367 lochs occupy glacially sculptured rock basins. The islands are mainly composed of inclined quartzite ridges that form prominent elongate landforms which caused Geikie (1865, p.215) to comment:

> "Nothing can exceed the distinctness with which the lines of stratification in the (Jura) quartz rock are traced on the cliffs and along the ridges. We can almost follow the line of each separate bed of rock as it winds over hill and crag, valley and tarn."

The upland topography has resulted in the presence of numerous incised rivers with generally small catchment areas. With the exception of raised beach areas, most of the ground is infertile with steeply sloping bare rock surfaces separating peat-filled hollows (Plate 1). The coastline is characterised by a wide variety of cliffs, shore platforms and raised beaches that in W Jura constitute some of the finest examples of raised shorelines in western Europe (Ritchie and Crofts, 1974).

1. Geology

Jura, Scarba and NE Islay possess a relatively simple geological structure, being composed primarily of a uniform series of metamorphosed mid-Dalradian quartzite sediments. The Dalradian rocks are late Pre-Cambrian in age and form part of the Caledonian belt of Scotland and Ireland. The quartzite is largely unaffected by folding or thrusting and its structure is therefore relatively simple. The strike of the rocks runs parallel to the length of Jura and Scarba and nearly everywhere the rocks dip ESE at between 25° and 40° (Fig. 2) (Plate 1).

The quartzite attains a maximum thickness of 5.3 km in S Jura and thereafter thins to the NE and SW (Anderton, 1976, p.431). Palaeocurrent investigations of the quartzite (Anderton, 1976) indicate deposition from the NE and SW into a subsiding Caledonian syncline that was later uplifted and tilted. The quartzite crops out over almost all of the area except along eastern coastal zones where it is overlain by a narrow band of younger rocks that includes the E Jura slates, the Scarba conglomerates and (in SE Jura and NE Islay) the Port Ellen phyllites (Fig. 2). The slates, phyllites and conglomerates of E Jura and E Scarba are generally conformable with the underlying quartzite and are believed to represent a change in depositional environment from a shallow basin to a much deeper

Plate 1. Interior of W Jura showing bare rock surfaces of inclined quartzite.

Plate 2. Tertiary dolerite dyke protruding above modern storm ridge and postglacial raised beach deposits, SW Jura.

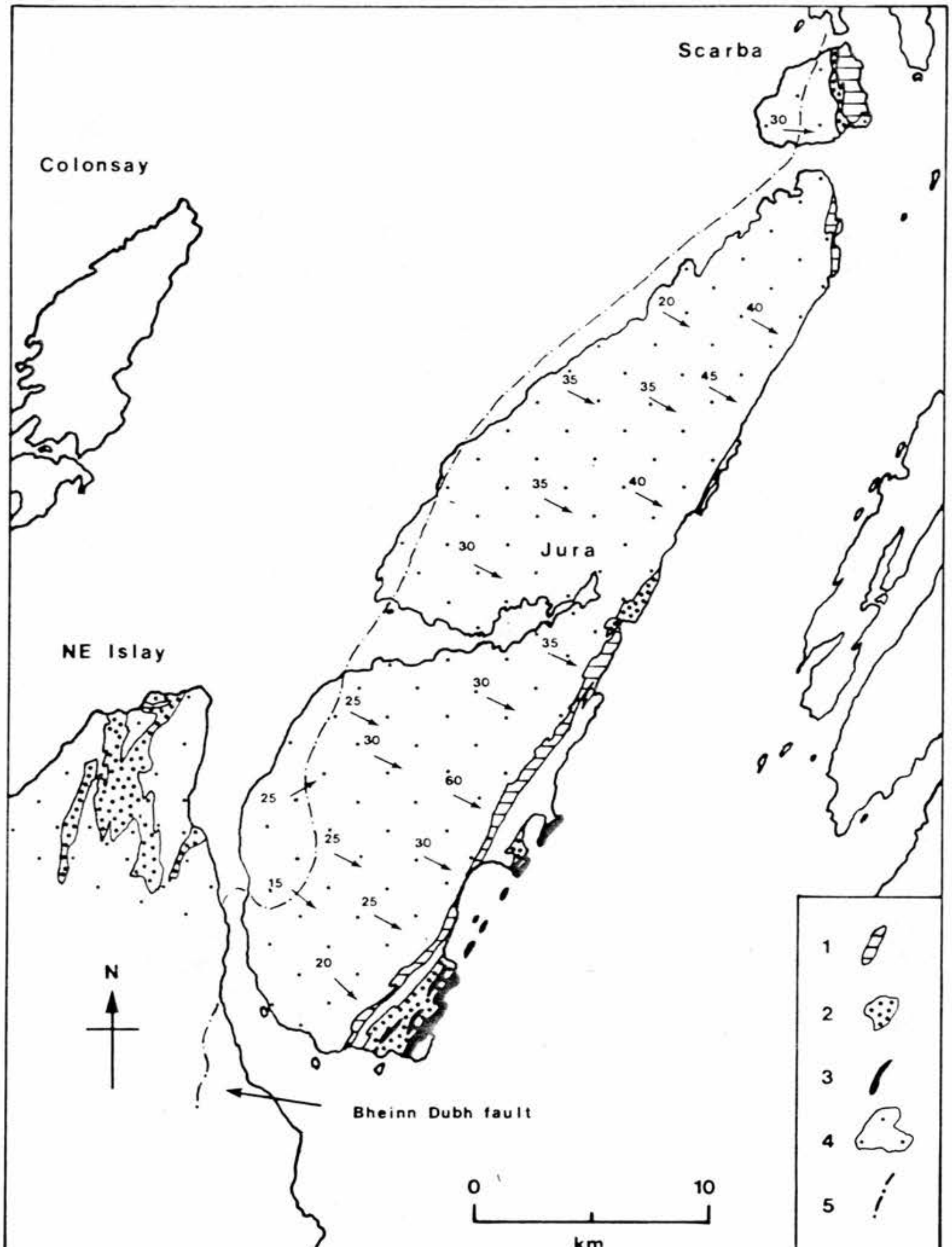


Fig. 2 Geological structure of Jura, Scarba and NE Islay (igneous dykes omitted). 1. Black slate 2. Phyllite 3. Epidiorite sills 4. Quartzite 5. Eastern margin of vitreous quartzite. Numbers and arrows denote dip of quartzite strata (after Peach et al., 1911).

set of troughs in which quiescent deposition is considered to have been responsible for the slates and phyllites and turbidity currents for the conglomerates (Anderton, 1976). In detail the contact between the quartzite and phyllites is complex and in many areas interfingering of slates and quartzite has precluded any clear contact zone.

In W Jura, Scarba and NE Islay the quartzite is primarily thick, white, well-bedded fine-grained and vitreous and here forms the margin of a shatter belt that is possibly related to the Beinn Dubh fault of E Islay (Peach et al., 1911,p.99) (Fig. 2). With the exception of Loch Tarbert, the present configuration of the W Jura coastline is the result of glacial and subaerial erosion along this zone of weakness. Similarly the coastal configuration of NE Islay is related to part of this shatter belt and cliff sections here reveal considerable thicknesses of vitreous quartzite intercalated with thinner bands of phyllites, dolomitic shales and flagstones.

Along the E coast of Jura from the Sound of Islay to Lowlandmans Bay resistant epidiorite sills are intruded through phyllites (Peach et al., 1911, p.104) (Fig.2). In this area the sills constitute the headlands and prominent inland ridges. In contrast, epidiorite dykes are best developed in N Jura and Scarba (Fig. 3). Depending on their susceptibility to erosion they either correspond with river gorges or else protrude as ridges. The dykes, which attain widths of up to 6 m and trend N-S and NNW-SSE (Peach et al., 1911,p.105) are largely responsible for the crenulate coastal configuration of NW Jura. No definite age has been assigned to these dykes but in Scarba they are crossed by dykes of Lower Old Red Sandstone age.

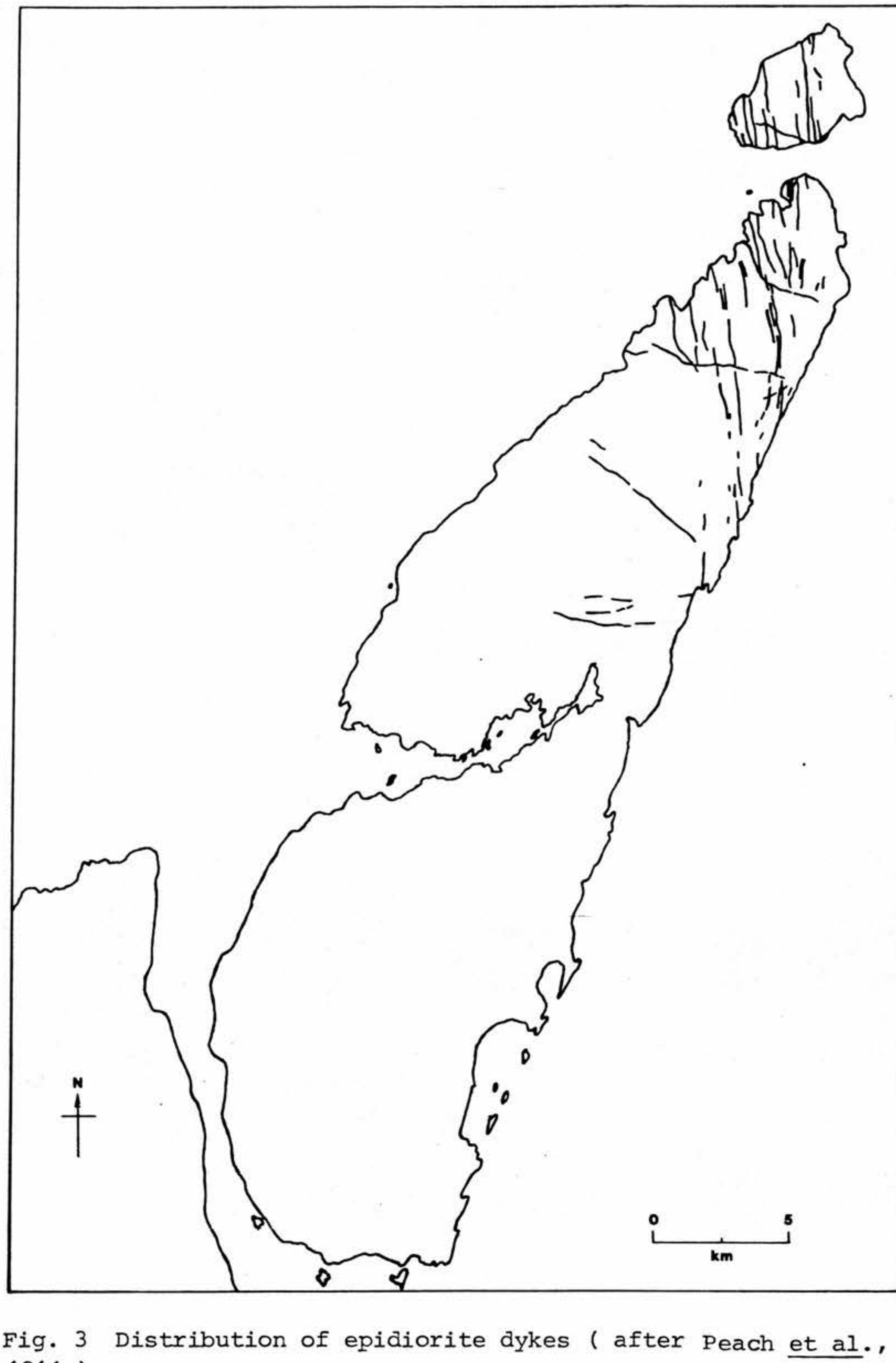


Fig. 3 Distribution of epidiorite dykes (after Peach <u>et al</u>., 1911).

Three other groups of igneous rocks have had a significant effect on the structure and relief of the area. Firstly, sills and dykes of lamprophyre, porphyrite and felsite related to the New Granite (Lower Old Red Sandstone age) of the mainland are numerous throughout N Jura (Fig. 4). The most conspicuous lamprophyre sill constitutes the cap rock of the cliffs from Ruantallain towards Shian Bay in W Jura (Fig. 4 [1]). Here, numerous large caves have been excavated in the less resistant underlying quartzite. Additionally, in E Jura, porphyrite, felsite and lamprophyre dykes form part of the headlands at Ardlussa, Lealt, Barnhill and Tarbert (Fig. 4 [2-5]).

Secondly, E-W trending quartz-dolerite dykes of Permo-Carboniferous age (Peach et al.,1911, p.108) are also present although they are few in number. They are well-developed in N Jura, notably on the south shore of the Gulf of Corrievreckan, and also farther south where a dyke extends continuously from Carn Tom on the E coast to Bagh Uamh nan Giall on the NW coast of the island (Fig. 4 [6]).

Finally, the most conspicuous igneous rocks throughout the area are the NW-SE-trending Tertiary dolerite dykes (Plate 2) that form part of the larger Tertiary dolerite dyke swarm of the Inner Hebrides. They are most common in S Jura and NE Islay (Fig. 5) where, in the coastal zone, they form natural groynes. Frequently the dykes form dense concentrations of stacks, arches and minor headlands due to the preferential erosion of the adjacent vitreous quartzite. In contrast, numerous other dykes have been preferentially eroded and correspond with narrow ravines. This is particularly evident in NW Jura, where south of Stac Dearg (Fig. 5 [1]) deep ravines descend

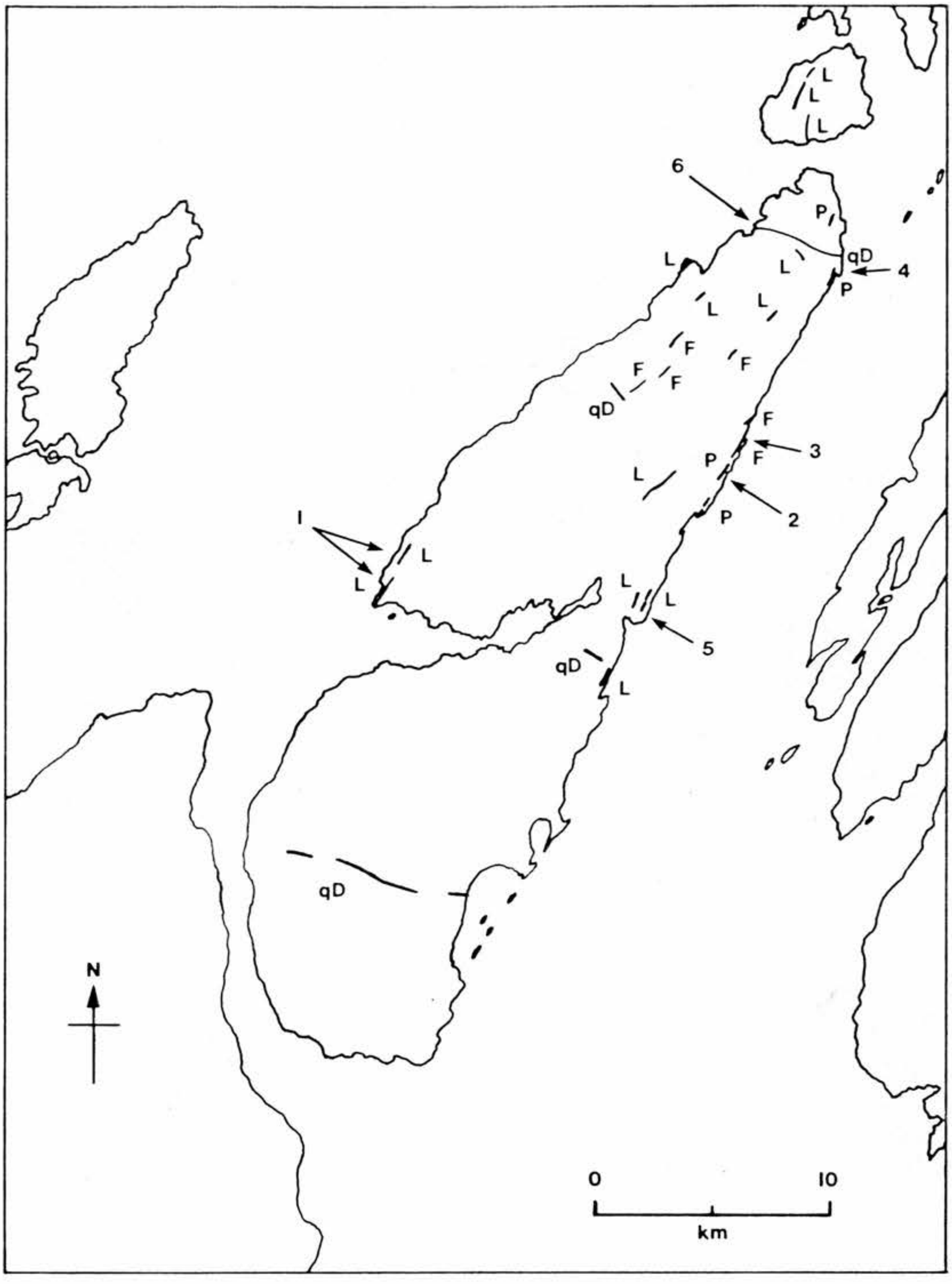


Fig. 4 Distribution of quartz-dolerite dykes (qD), lamprophyre sills and dykes (undifferentiated) (L), felsite (F), and porphyrite (P) (after Peach et al., 1911). 1. Ruantallain to Shian Bay 2. Ardlussa 3.Lealt 4.Barnhill 5.Tarbert 6.Bagh Uamh nan Giall.

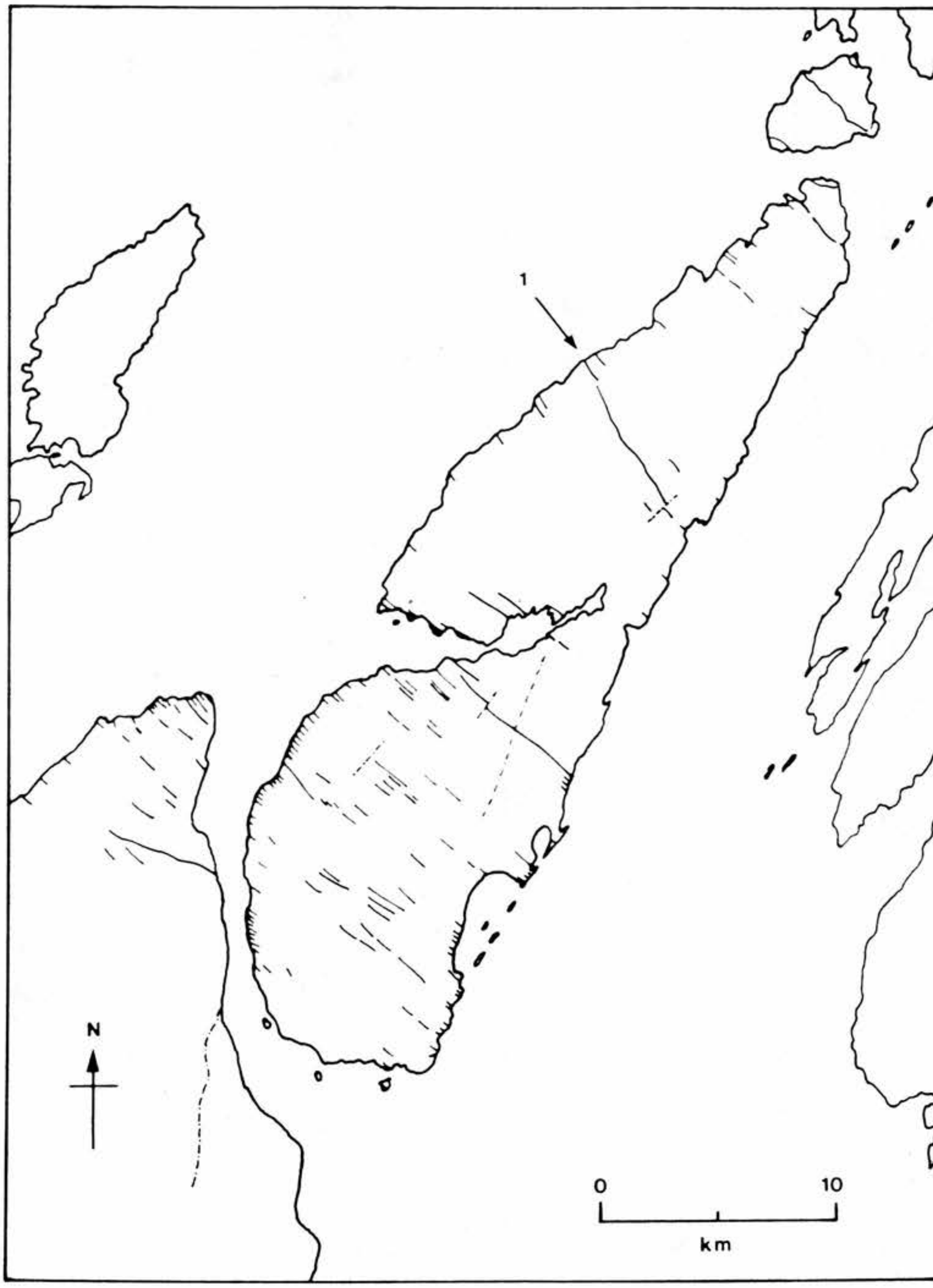


Fig. 5 Distribution of Tertiary dolerite dykes. Dashed lines denote major faults (after Peach et al., 1911). 1. Stac Dearg.

to the coast and extend seaward as linear depressions in a coastal shore platform.

Aerial photograph interpretation and field checking by the writer has revealed several fault planes not identified by the Geological Survey (Fig. 5), some of which are of considerable size. The ages of the faults identified are unknown but in several instances dislocation of Tertiary dykes is evident.

Little evidence exists on the nature of the Tertiary landscape of the area although suggestions have been made that the Gulf of Corrievreckan, Loch Tarbert, the Sound of Islay (Peach, 1909) and perhaps also the glacially overdeepened Firth of Lorn and Sound of Jura occupy preglacial drainage routes (Ting, 1937).

Structural control in conjunction with preglacial and glacial erosion of bedrock is thus responsible for the general configuration of the coastal zone. In detail, however, the coastline has been subject to numerous periods of marine erosion and deposition during and following the Quaternary Ice Age. Thus before considering in detail the nature, age and origin of individual shorelines, there follows a general description of the coastal landforms.

2. Shore platforms

The coastline of W Jura, Scarba and NE Islay is characterised by a broad raised shore platform (Fig. 6) that will hereafter be referred to as the Main Rock Platform. The platform is generally 50-100 m in width and is backed by cliffs that in W Scarba and NW Jura reach heights of over 50 m. The inner edge of the platform attains its maximum altitude in Scarba and thereafter declines gently in altitude to the SW until in NE Islay and SE Jura it occurs in the intertidal

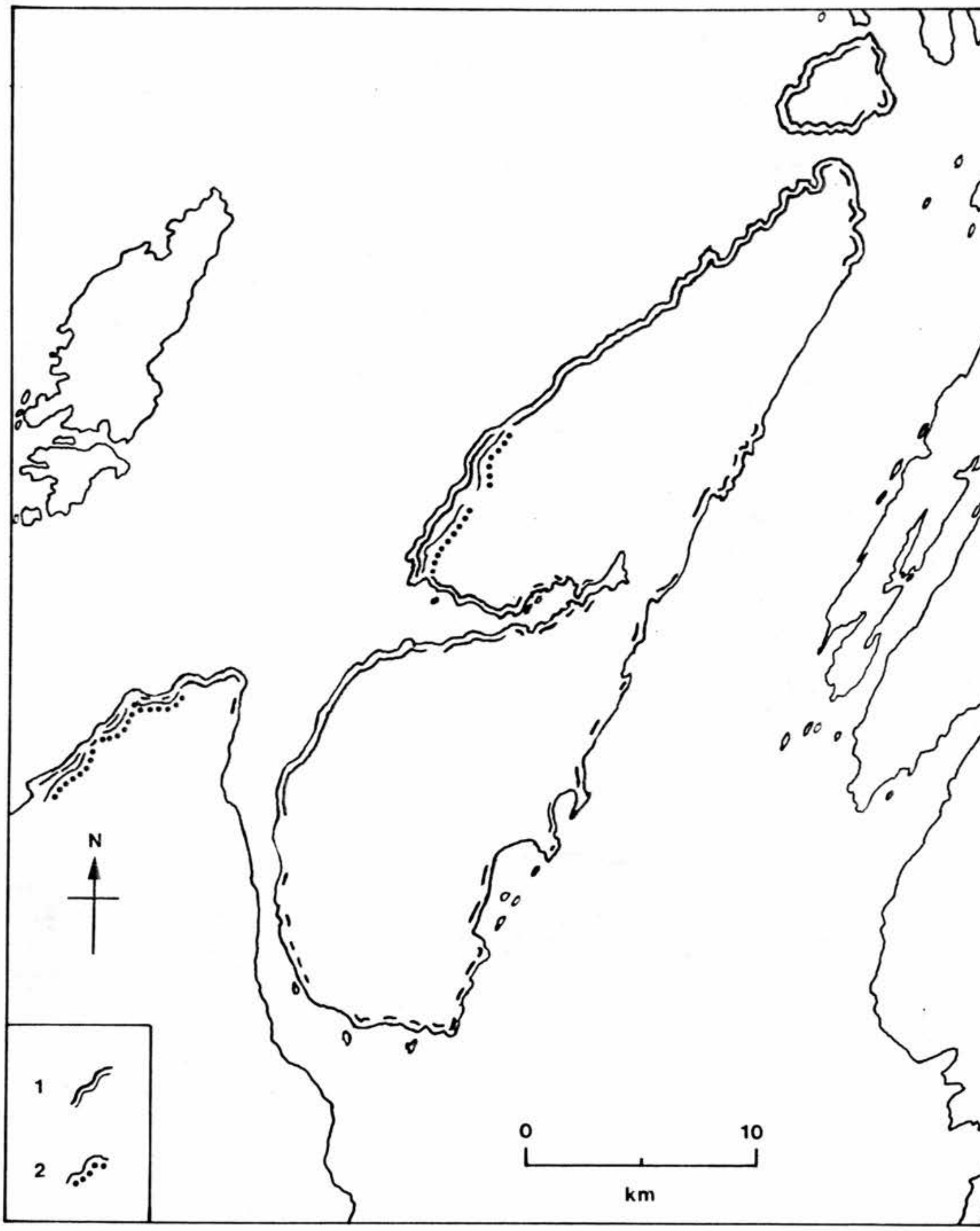


Fig. 6 Distribution of raised shore platform fragments, Jura, Scarba and NE Islay. 1. Platform fragments below 10 m O.D. 2. High Rock Platform fragments.

zone. The platform is equally well developed in bays and headlands and forms a continuous feature along the western coastline where it is often bare of overlying sediment. Many delicate stacks and arches occur on the platform while behind it the cliffline is indented by innumerable caves and geos. Owing to the dip of the quartzite the Main Rock Platform generally possesses an irregular surface with resistant bosses of quartzite protruding above it as angular inclined ridges. In addition, the continuity of the platform is often interrupted by igneous dykes that form linear ridges across its surface while active and fossil talus accumulations frequently obscure its inner edge. In contrast, the Main Rock Platform is poorly developed in E Jura. Here, platform fragments are locally cut in slates, phyllites and quartzite and are generally overlain by raised beach deposits and backed by a degraded cliffline. With the exception of Small Isles Bay and Lowlandmans Bay (Fig. 2), the eastern coastline is essentially straight with inclined rock strata dipping uninterrupted into the sea.

At several localities in Scarba and NW Jura the frontal edge of the Main Rock Platform forms the backing cliff of a more poorly defined lower rock platform (hereafter referred to as the Low Rock Platform). This latter platform generally occurs in intertidal bay areas and is absent from headlands. Unlike the Main Rock Platform, stacks and arches are absent and its smooth surface is frequently ice-moulded. In SW Jura a similar intertidal platform passes beneath backing cliffs composed of glacial till while seaward of the till cliffs the intertidal platform is masked by thick accumulations of raised beach deposits.

South of Corpach Bay the backing cliffs of the Main Rock Platform

form the frontal edge of a higher rock platform (the High Rock Platform) that extends SW for 7.5 km as an almost continuous feature and in places reaches a width of 550 m. South of Shian Bay massive staircases of unvegetated shingle mantle the High Rock Platform and obscure its inner edge. At Ruantallain the High Rock Platform disappears and is found nowhere else in Jura. In NE Islay however, the High Rock Platform attains its most spectacular development (Plate 3) where it reaches a width of 650 m and is backed by a cliff up to 70 m in height. The platform slopes evenly seawards and, although indented in places by geos, it is generally free of protruding stacks, dykes and arches.

3a. High raised beach deposits (above 14 m)

High raised beach deposits are intermittently developed along the coasts of Jura, Scarba and NE Islay (Fig. 7). In E Jura raised coastal terraces occur in sheltered embayments while in the more exposed environment of W Jura and NE Islay, suites of unvegetated beach ridges replace coastal terraces as the dominant depositional raised shoreline landform. In N Jura and Scarba, terrace fragments of high raised beach gravels choke the lower courses of river valleys and in Glengarrisdale beach gravels extend as far as 1.5 km inland. In SW Jura north of Inver broad expanses of vegetated raised beach deposits extend almost 1 km inland. In this area the upper limit of marine activity is represented by a distinct cliffline cut in drift that can be traced northwards for 4.5 km. Similar terraces of high raised beach gravels occur in E Jura and are most conspicuous in Small Isles Bay and Lowlandmans Bay, while at Ardlussa vegetated raised tombolos occur between protruding rock ridges. Additionally, raised terrace fragments at the head of Loch Tarbert indicate that Jura was

Plate 3. High Rock Platform, NE Islay, looking west across Coir Odhar embayment towards Mala Bholsa (far distance). Sloping ramp abrasion profile is well developed.

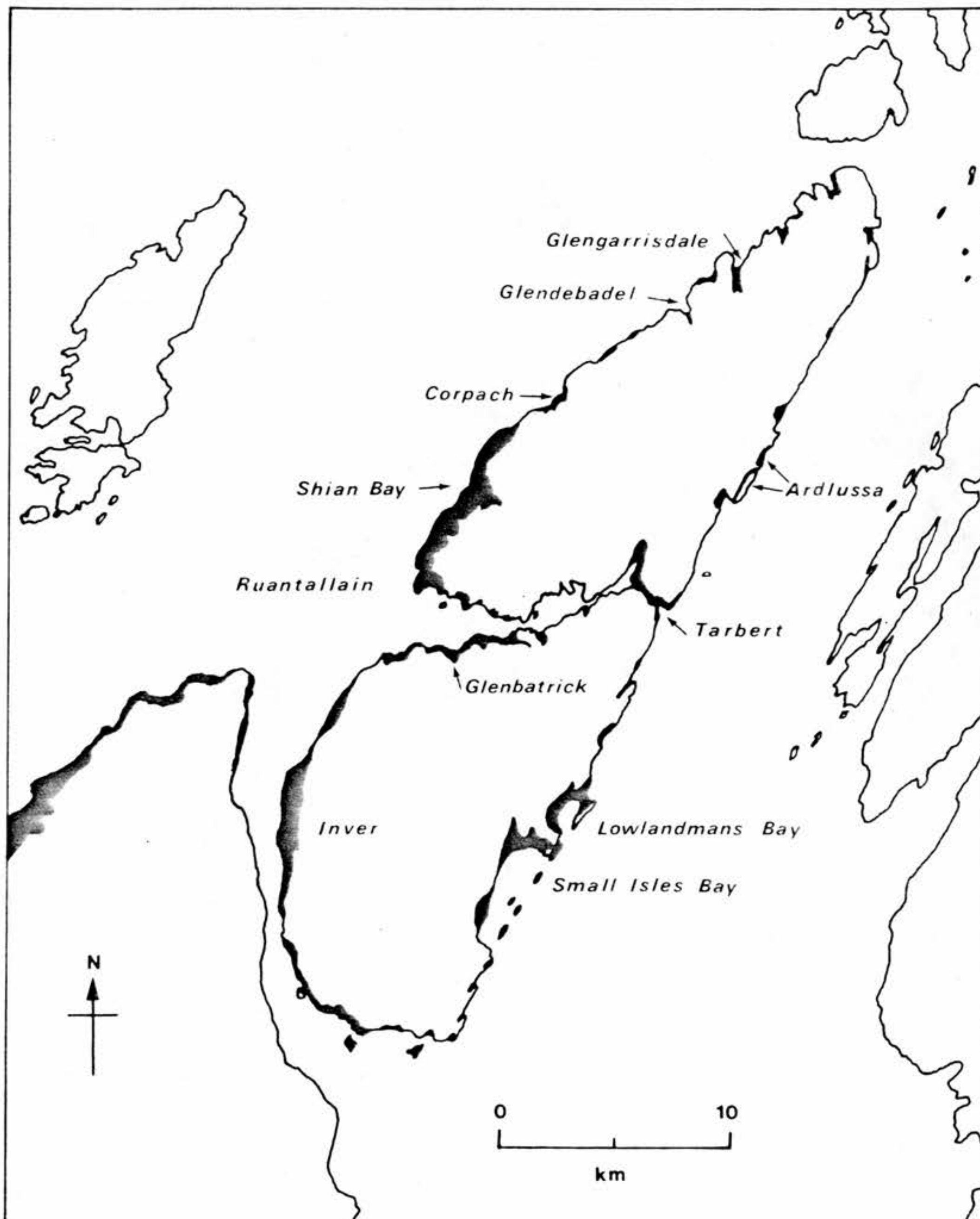


Fig. 7 Distribution of raised shoreline deposits, Jura, Scarba and NE Islay.

formerly divided in two by a tidal channel that linked Loch Tarbert and the Sound of Jura.

In the Corpach, Glendebadel and Glenbatrick river valleys large accumulations of glacial and fluvioglacial deposits choke the lower valley courses and are succeeded seaward by raised coastal deposits, while west of Glenbatrick a large area of stratified raised beach gravels occurs and is replaced landward by a clearly defined marine limit. Here the upper surface of the high raised beach deposits has been reworked by former marine activity to produce a large area of undulating ridge and swale topography. Nowhere, however, are raised beach ridges better developed than between Ruantallain and Shian Bay in W Jura (Fig. 7) (Plate 4). Here staircases of up to 55 unvegetated beach ridges separated by swales mantle the High Rock Platform. In this area 5 freshwater lochs are sealed at their seaward margins by raised beach deposits. Along this stretch of coast, the upper marine limit is clearly defined and forms a small cliffline cut in drift that can be traced for 7 km along the coast. In contrast, NE Islay possesses relatively few high raised beach deposits: where they occur they mantle the seaward surface of the High Rock Platform.

3b Low raised beach deposits (below 14 m)

In Scarba, NW Jura and NE Islay low raised beaches occur in coastal indentations, where they usually form undulating surfaces of ridge and swale topography. The beach ridges are most commonly located in exposed western coastal areas while in E Jura they are replaced by low coastal terraces. In N Jura and Scarba unvegetated pocket beach ridges frequently occur in indentations of the Main Rock Platform while in SW Jura and NE Islay thick accumulations of raised

Plate 4. Oblique aerial photograph of raised lateglacial ridge and swale topography, Loch a Mhile, W Jura. Main Rock Platform is here well developed and is at c.2.5 m O.D. Photograph courtesy of John Dewar Studios, Edinburgh.

beach deposits form large bayhead beaches that extend landward to the backing cliff of the Main Rock Platform. In SW Jura low-level beach deposits are best developed north of Inver where 31 beach ridges descend seaward from 13 m to sea-level as a staircase of unvegetated ridges and swales. In several small estuaries raised coastal terraces and beach ridges occur together and clearly indicate the elevation of former high sea-levels.

4. Modern coastal landforms

The distribution of modern coastal landforms is largely influenced by the distribution of fossil shore platforms, cliffs and raised beach deposits. In E Jura, low wave energy conditions, caused by restricted fetch, a low tidal range, and the linearity of the coastline have resulted in localised beach development and a poor development of beach ridges. In contrast, in W Jura, NE Islay and Scarba, exposure to open Atlantic fetch has resulted in the presence of large beach ridges in most coastal indentations. Sandy beaches are rare and are only well-developed in bay environments (Small Isles Bay, Corpach Bay and Shian Bay) where the largest rivers reach the coast.

A detailed understanding of the fossil shore platforms and raised beaches that fringe the coastline of Jura, Scarba and NE Islay requires an understanding of the agencies responsible for contemporary marine erosion and deposition. A brief description is thus given here of the influence of offshore topography, wind, fetch and tides on modern coastal landforms (for a fuller account see Chapter 5).

5. Offshore topography

The Firth of Lorn and the Sound of Jura are both NE-SW-trending overdeepened glacial trenches that lie oblique to the dominant E-W movement of Quaternary ice sheets. The trenches attain depths of 250 m in places and are frequently infilled with considerable thicknesses of glacio-marine sediment (Binns et al., 1974).

Both the Sound of Jura and the Firth of Lorn are approximately parallel with the strike of the rocks and, although the rock structures are complicated,they clearly aided glacial erosion along their major axes (Sissons, 1967b). The possibility also exists that the floor of the Sound of Jura and the seabed between Jura and Colonsay are in part composed of Old Red Sandstone since erratics of this rock occur in Colonsay (Craig et al., 1911) and Jura, while the presence of Old Red Sandstone outliers along the W coast of Kintyre suggest their presence on the sea bed (Sissons, 1967b).

Between Jura and Colonsay, the deep NE-SW-trending trench of the Firth of Lorn is replaced by a shallow sub-horizontal shelf generally 20-40 m below sea level. Little evidence exists for the presence of this surface in the Sound of Jura where the bedrock surface descends steeply beneath sea-level from the coast.

6. Fetch and wind

Full Atlantic exposure to the dominant SW wave trains is found along most straight and exposed coastal stretches of W Jura, along NE Islay and in SW Scarba. The coastline of N Jura and W Scarba is additionally exposed to open fetch NW Atlantic wave trains (Fig. 8). Analyses of wind patterns from Tiree and from Rhuvaal, NE Islay (Fig. 9) indicate the dominance of SW, W and S winds that, in conjunction with fetch conditions, clearly indicate the concentration of wave energy on western coasts.

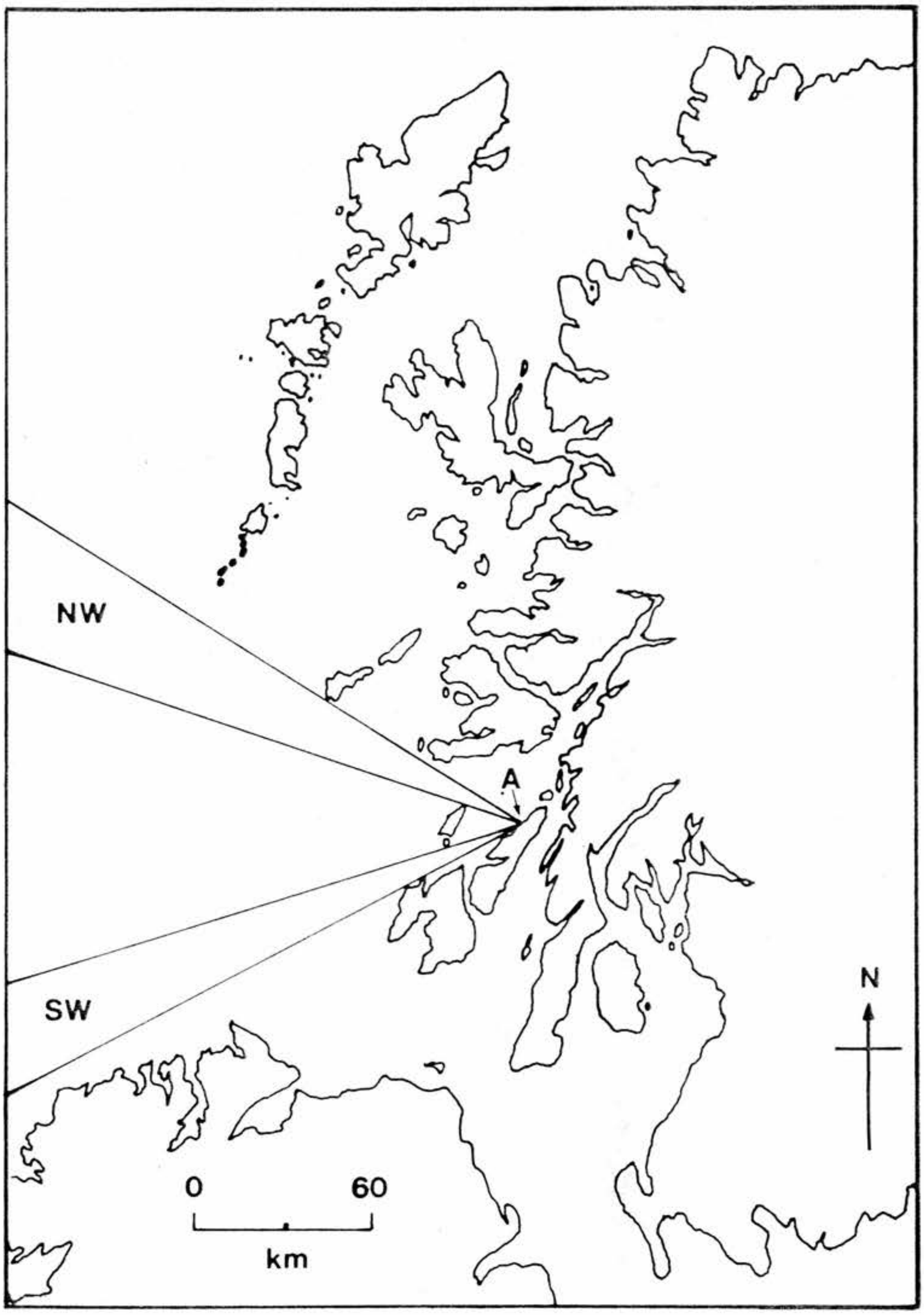


Fig. 8 Open Atlantic fetch sectors at coastal location A.

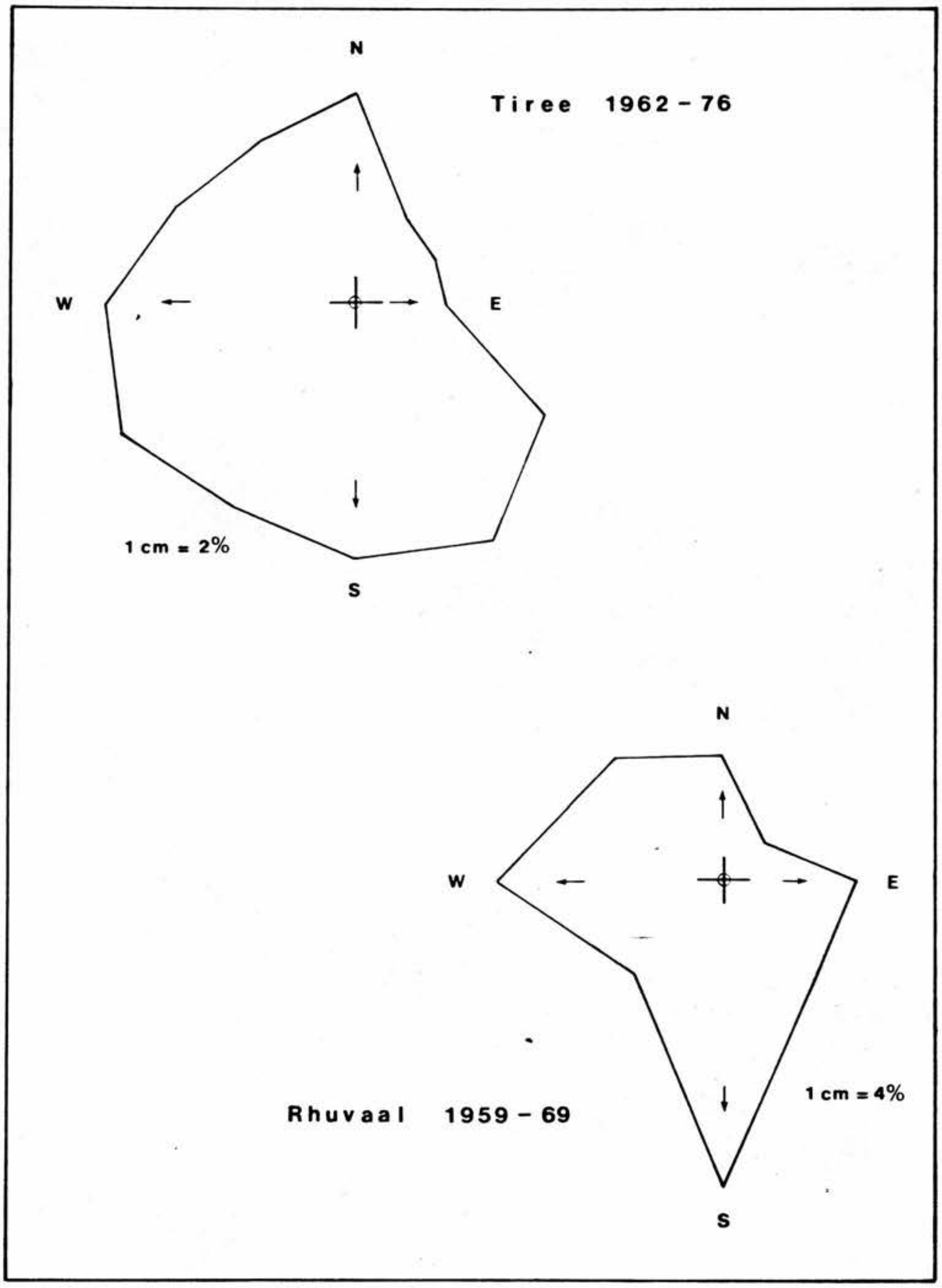


Fig. 9 Wind rose diagrams for Tiree (1962-76) and Rhuvaal, NE Islay (1959-69). Information courtesy of the Meteorological Office, Edinburgh.

The broad shallow shelf that extends westwards from SW Jura and NE Islay favours the breaking of large waves far offshore and counteracts the erosive potential of SW storm waves, while the distribution and thickness of sediment on offshore shoal banks locally affect the supply and removal of beach material.

The erosive potential of waves along eastern coastal waters is diminished by restricted fetch and a low frequency of storm waves from eastern sectors yet is augmented by steep offshore gradients. The direction of waves approaching the coast and the transport and deposition of sediment along the coastline are complicated by locally strong tidal currents.

7. Tides

Mean spring tidal range varies throughout the area from 3.1 m in N Jura to 0.9 m in SE Jura (Table 1). The decline in tidal range from north to south is largely due to the presence of an amphidromic point south of Islay. However in W Jura and NE Islay tidal range is approximately constant with mean high water spring tide varying by as little as 0.4 m along the coast. During spring flood tide the streams run NE through the Sound of Jura and the Firth of Lorn and discharge SE through the Sound of Islay and W through the Gulf of Corrievreckan. In the Sound of Islay the south-going stream attains a maximum velocity of 6 knots and is accompanied by reverse eddy currents in coastal embayments. In the Gulf of Corrievreckan the westward-flowing streams of flood tide reach velocities of 8.5 knots. Here the main westward stream preserves its character after emerging from the Gulf and sends back reverse eddies toward N Jura and Scarba. When opposed by W winds the Gulf is unnavigable with heavy overfalls extending 5 km westwards from the Gulf. Ebb tides are accompanied by

Table 1

Tidal regimes and locations for stations in the study area. Values (m) are above Chart Datum.

	M.H.W.S.	M.H.W.N.	M.L.W.N.	M.L.W.S.
1. Rhuvaal (Rubha a Mhail)	3.7	2.8	1.5	0.6
2. Ardnave Point	3.6	2.7	1.5	0.6
3. Port Ellen	0.9	0.8	0.5	0.3
4. Port Askaig	2.1	1.5	1.0	0.4
5. Oban	4.0	2.9	1.8	0.7
6. Craighouse	1.2	0.9	0.4	0.3
7. Loch Beag	2.4	1.7	1.0	0.3
8. Carsaig Bay	1.9	1.3	0.8	0.3
9. Gigha Sound	1.5	1.3	0.8	0.6
10. Glengarrisdale	3.3	2.4	1.5	0.2
11. Scalasaig	3.7	2.5	1.3	0.3

* [from Admiralty Tide Tables, 1976]

flow reversals that in the narrow sounds attain slightly lower velocities than those of flood. Periods of slack water are generally of short duration: in the Gulf of Corrievreckan slack water during spring tides lasts for only 15 minutes.

It is difficult to evaluate the morphological effects of tidal action on coastal landforms. Tidal currents are generally regarded of little significance as agents of littoral transport and deposition since the main current flows in deep water offshore (Johnson, 1919; Zenkovich, 1967). The effect of a high tidal range, as occurs in W Jura, is to increase the zone over which destructive waves can operate (King, 1972) and it thus facilitates the erosion of cliffs.

Additionally, the continual combing action of wave uprush and backwash in the intertidal zone results in relatively steep beach gradients and impedes the development of offshore bars and ridges.

Chapter 3

Techniques and Field Methods

1. Landform mapping

Glacial and coastal landforms were initially mapped from RAF 1:25,000 (1973) aerial photographs and the information transferred onto 1:10,560 maps. In addition, the coastal terraces and shingle spreads of W Jura were mapped from RAF 1:10,000 (1946) aerial photographs and subsequently transferred onto 1:5,280 maps. After all glacial landforms and raised marine features had been mapped, the area was investigated on foot. During fieldwork investigations the 1:10,560 and 1:5,280 geomorphological maps prepared from aerial photographs were checked and amended while all available exposures were examined and interpreted.

2. Altitude measurement of raised shorelines

a) Previous techniques and methods

The problems associated with the measurement of raised shoreline altitudes are numerous and are discussed in detail by Gray (1975). The choice of unreliable datum lines and the use of inaccurate instruments in the measurement of shoreline altitudes has resulted in height errors in some shoreline studies (eg. Donner, 1963; King and Wheeler, 1963; Synge, 1966; Synge and Stephens, 1966). In previous shoreline investigations of the study area the selection and measurement of a suitable datum line have differed. Wright (1911), for example, used the high water mark of ordinary spring tides as a datum but did not state the method by which this was ascertained while McCann (1964) measured the same datum line by Abney level. Synge and Stephens (1966) used the elevation of the line of uppermost fragments of drifted sea salad (Enteromorpha spp.) and equated this level with high water mark.

In the above studies hand-held Abney levels and aneroid barometers were used to measure shoreline altitudes. The use of Abney levels to measure altitude has been strongly criticised (Sissons, 1967b; Gray, 1975a) while Rhind (1969, p.52) has suggested that height errors incurred in a single traverse may amount to 15%. Similarly the use of aneroid barometers may induce height errors of up to 3 m over short traverses (Donner, 1959). As a result, the raised shoreline height values obtained by Wright (1911), McCann (1964) and Synge and Stephens (1966) in the study area are of limited value.

b) Altitude measurement techniques employed in the present study

In the study area, Ordnance Survey bench marks were used as a datum and the heights of all raised shoreline features were determined using Hilger/Watts and Wild NAK autoset levels and a staff graduated to 0.01 m. All levelling traverses were closed and in the three cases where closing errors exceeded 0.15 m the traverses were repeated. Collimation tests on the levels were conducted on average once every 10 days in order to ensure their continued accuracy throughout the fieldwork season.

In E Jura levelling was conducted between local and Newlyn bench marks in order to convert local bench mark altitudes to Newlyn datum. It was found that on average Newlyn bench marks lay 0.47 m above local datum. In northern and western Jura the only bench marks that exist are of local datum. Here, levelling was undertaken between local bench marks and subsequently converted to Newlyn datum.

In Scarba no bench marks exist and consequently an alternative method of height determination was used. In several N Jura embayments the altitude of the seaweed fucus Pelvetia canaliculatus (sp.) deposited during high water of ordinary spring tides was levelled. During levelling, the embayments in which the measured height range was less than 0.05 m were noted. In these embayments the arithmetic mean of the measured Pelvetia height values was obtained. Subsequently, levelling of the corresponding Pelvetia line was conducted on Scarba. The Pelvetia values obtained on Scarba were then corrected to Newlyn datum by utilising the values obtained from N Jura. In subsequent chapters all altitude information refers, except where otherwise stated, to Ordnance Survey Newlyn datum.

3. Identification and measurement of shoreline features

Many raised marine features in the study area are overlain by accumulations of peat (a point not mentioned in the investigations of Synge and Stephens (1966) and McCann (1964)). During levelling the altitude of the sub-peat surface was measured by pushing 1 m steel rods vertically through the peat. It was usually possible to distinguish the subsurface material as either sand, cobbles or bedrock on the basis of sound contrasts when the rod(s) made contact with the sub-peat surface. The point to be measured was chosen only after numerous probes yielded a consistent depth value. Thereafter the sub-peat surface altitude was determined by subtracting the peat depth from the levelled height of the ground surface.

In this manner the altitudes of raised coastal terraces were determined by surveying along their inner margins avoiding slumped debris, former stream channels and areas aggraded by alluvial fans. On broad coastal terrace fragments the altitudes of vegetated beach ridges were similarly measured. Unvegetated raised beach ridge and swale topography was also surveyed, and by the use of the levelling staff to measure horizontal distances, cross profiles were determined.

Since the Main Rock Platform is generally bare of overlying sediment, the altitude of its inner edge was obtained by direct levelling of bedrock. Due however to the irregular nature of the rock surface at the inner edge of the platform, the elevation of individual platform inner edges was established by the levelling and averaging of 4-15 rock heights considered representative of the inner edge of the platform at each site selected for measurement. The same technique was applied to determine the altitude of the inner edge of the Low Rock Platform.

Unlike the lower platforms, the High Rock Platform is everywhere overlain by raised beach deposits and in places by glacial drift. Thus it was only possible to level the inner edge at certain stream sections and at the edges of geos where the inner edge of the platform was visible. At each measurement site as many as 14 heights were obtained before a mean value was calculated.

Although it is a relatively simple exercise to determine the heights of raised coastal landforms, it is more difficult to establish the altitudinal relationship between raised marine features and the sea-levels responsible for their formation. It is therefore essential, if raised coastal features are to be better understood, that the height relationships between modern coastal landforms and present sea-level be established. Although such relationships are given in research literature, the most relevant measurements of modern coastal features for the present study are those taken within the study area. For this reason the altitudes of numerous modern coastal landforms were levelled in the study area and an attempt was made to relate their elevations to present sea-level. The results of these investigations are presented in Chapter 5.

4. <u>Spacing of raised shoreline measurements</u>

In the study area height variations along the length of individual marine terrace fragments indicated the necessity to measure as many closely spaced points as was practicable in the time available. As a result most well-developed raised coastal terrace fragments were levelled at 50m intervals and in several cases at 30 m intervals in order to eliminate the influence of local factors on individual terrace fragment altitudes. In addition well-developed unvegetated beach ridge crests were normally levelled at 25 m intervals along their lengths.

Due to the localised distribution of Low and High Rock Platform surfaces, the inner edges of the platforms were surveyed throughout their lengths. Along western coastlines where the Main Rock Platform was a continuous feature, platform fragments were normally measured at 150-200 m intervals. In E Jura fragments of the Main Rock Platform were measured where they were easily accessible and well-developed.

5. <u>Shoreline diagrams</u>

In the study area the raised beaches identified by the Geological Survey are mapped as occurring at 25, 50 and 100 feet. Since these early studies it has been universally recognised that synchronous raised shorelines formed during and after deglaciation are not horizontal but are regionally tilted and rise in altitude towards the centre of isostatic uplift. Analysis of raised shoreline altitudes therefore involves the construction of shoreline diagrams and the calculation and interpretation of regional shoreline gradients.

In previous studies the analysis of raised shoreline altitudes has involved the use of shoreline equidistant diagrams (eg. Sissons, 1967b, 1972; Gray,1974b) and shoreline relation diagrams (Synge and Stephens, 1966). In both methods shoreline elevation above O.D. is plotted on the y-axis while the x-axis represents ground distance measured from an arbitrary point of origin. Shoreline relation diagrams are constructed by plotting the altitudes of a well-developed synchronous shoreline as a tilted or horizontal straight line while the positions of other shore fragments are calculated according to the altitude of the reference shoreline in the area where they occur (Gray, 1975a). This method has been strongly criticised (Marthinussen, 1960; Sissons, 1967b, 1972; Gray, 1975a) primarily on the basis that any errors incurred in the identification and measurement of a synchronous reference shoreline are transmitted to the rest of the data (Cullingford, 1977).

Shoreline equidistant diagrams have been more commonly used and involve the projection of the altitudes of shorelines into a vertical plane drawn normal to the isobases. The main disadvantage of this method is the distortion of values if there is considerable isobase curvature in the field area (Gray, 1975a).

In the study area the construction of an accurate shoreline equidistant diagram was aided by the presence of the Main Rock Platform, which forms a continuous feature along long stretches of the coast. 175 Main Rock Platform heights (see Chapter 8) derived from 950 field measurements (ie. 4-15 at each site as explained above) were plotted on 5 equidistant projections (Fig. 10 [1-5]) in the quadrant between west and south from an arbitrary point of origin NE of Scarba. Since equidistant projections that are increasingly oblique to the isobases of an area will depict a particular shoreline as having an increasingly steep gradient, the projection plane that produces the minimum shoreline gradient lies normal to the isobases. The gradient of the Main Rock Platform was calculated by linear regression for each of the five directions shown in Fig. 10. The lowest gradient obtained was for the direction NE-SW. Subsequently the regional tilts of all higher and lower raised shorelines were calculated from shoreline fragments projected into this plane.

6. <u>Correlation of shoreline fragments</u>

Separate shorelines were identified on the basis of three principles outlined by Cullingford (1972, 1977). Firstly, where raised beaches occur above each other in a staircase, they cannot be contemporaneous with each other. Although true of raised coastal terraces this principle is not universally applicable to beach ridge staircases where several beach ridges can form adjacent to each other during a period of stable sea-level (Johnson, 1919). Secondly the morphological continuity of synchronous shorelines is often greater than the distribution of measured altitudes might suggest. For example, synchronous shoreline fragments can often be correlated across areas dissected by stream erosion and across areas aggraded by alluvial fan development. Thirdly, shoreline fragments cannot be correlated with fragments lying within an area that was contemporaneously ice-covered. In this study, synchronous shorelines were identified on the basis of the principles outlined above in conjunction with the alignment of points on shoreline equidistant diagrams.

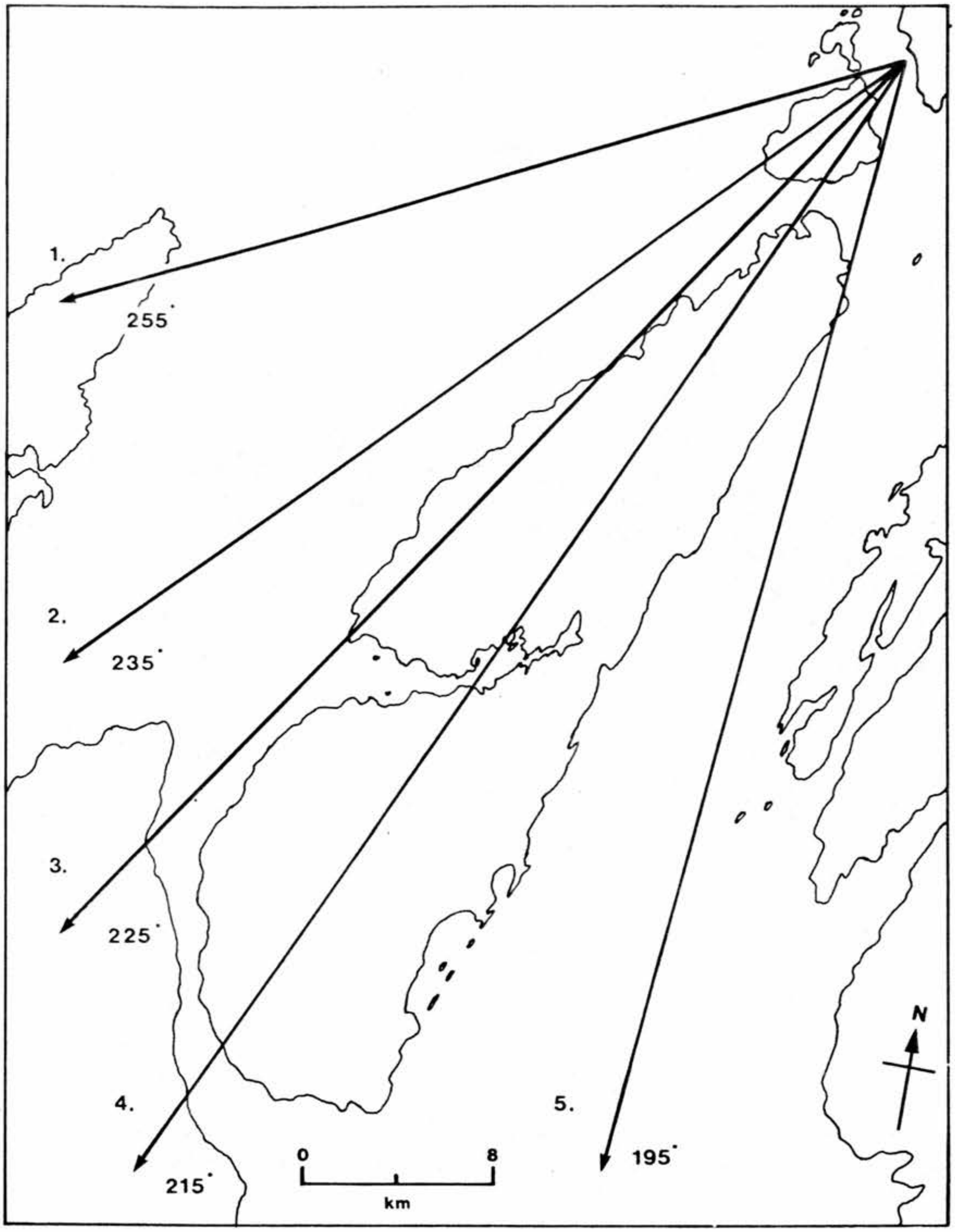


Fig. 10 Main Rock Platform projection planes.

7. Three-dimensional analysis of raised shorelines

Trend surface analysis

For three-dimensional analysis of the Main Rock Platform shoreline fragments, a SYMAP Version 5.20 trend surface programme developed at the Graduate School of Design, Harvard University was executed on the Edinburgh Regional Computing Centre IBM 360 and ICL 2980 computers. Regional isobases for the Main Rock Platform fragments were determined by fitting trend surfaces to the altitude and grid co-ordinate data. Due to the elongate nature of the coastlines from which the platform fragments were derived, the data points exhibited clustering. For this reason, trend surfaces were separately calculated for the entire data and for all data excluding E Jura fragments.

The choice of order at which to terminate trend surface analysis is the subject of considerable disagreement and has been discussed in detail by Gray (1972, 1974a, 1974b), Baird et al., (1971), Chayes (1970) and Unwin (1975). In the present study the trend surface order chosen as representative of the data was calculated by the analysis of variance test (Krumbein and Graybill, 1965). This test compares the relative degree of fit between the data and trend surfaces of increasing order and determines the trend surface order (or polynomial expression) beyond which no significant additional explanation of the data is provided. In the present study the first and second order trend surfaces exhibited similar patterns and at the 2nd order gave the highest significant additional explanation at the 95% level (see Chapter 7). The residual values obtained during trend surface analysis, which represent variations of individual points from the calculated trend surfaces, were also examined in order to investigate platform height variations between bays and headlands.

Chapter 4

Glacial landforms

1. Summary of Research

In 1888 Anderson (pp.318-9) noted glacially transported mainland erratics in Jura and suggested that the island had been over-ridden by ice flowing westward across the Sound of Jura from the mainland. Later, Peach et al. (1909,p.95) stated that there was clear evidence for,

> "... the westerly seaward passage of an ice sheet which during its maximum extention not only filled the sounds and sea lochs, but, notwithstanding the superiority in height of the central ridge of Jura and Scarba to any parts of the (adjacent) mainland, also overrode the outer isles."

Peach concluded that since striae are oriented NE-SW in the Garvellach Isles, E-W on Scarba, to the north of west over the central ridge of Jura and SE-NW over S Jura and NE Islay, the westerly flowing ice instead of fanning outward over the continental shelf, converged westward. The pattern of striae orientations was interpreted as the result of compression of a westerly flowing ice-stream between two more powerful ice-streams; one flowing SW through the Firth of Lorn and the other flowing NW over Kintyre, Knapdale and Islay as far as Colonsay.

Early studies of glacial erratics by the Geological Survey conform to the general pattern of ice-sheet movement as observed from glacial striae. Epidiorite erratics of a mainland provenance have been recorded on Scarba (Peach et al., 1911,p.123) while in Islay and S Jura several Loch Fyne porphyry erratics have been identified (Wilkinson, 1907,p.70). A W and NW flow of ice west of Jura is also indicated by the presence of Jura quartzite and Loch Fyne porphyry erratics on Colonsay (Craig et al., 1911,p.60).

Craig et al., (1911) believed that the NE- SW-trending trenches of the Firth of Lorn and the Sound of Jura showed that their glacial overdeepening was due to a SW movement of ice along their major axes while the orientation of striae on Jura, Scarba and NE Islay indicated a dominant flow of ice from E and SE to W and NW. This apparent contradiction was initially noted by Geikie (1882, p.163) who suggested that the submerged glacial trenches were primarily a result of the SW deflection of undercurrent ice layers in a westerly moving ice-sheet. Later it was proposed that the presence of local ice on the island land masses added to the effectiveness of local mountain barriers to induce SW deflection of ice along the glacial trench axes (Craig et al., 1911,p.59). In addition Peach (1909,p.96) suggested that variations in striae orientation along the E Jura coastline pointed to the effect of undertow of ice flowing SW down the Sound of Jura. No evidence was given to indicate a similar SW undertow of ice in W Jura adjacent to the Firth of Lorn. More recently it has been proposed that at the beginning of each Quaternary glaciation the area was invaded by SW-moving ice-streams from local high ground but during glacial maxima each ice-sheet moved westward over the entire area, this sequence being reversed during deglaciation (Binns et al., 1973,p.754).

The melting of the final ice sheet in Jura was believed to have been followed by a period of valley glaciation (Anderson, 1888,pp.318-9). Anderson described "banks of earth and angular stones" found in several Jura valleys as "morainic matter" formed during this period. His views were later reiterated by Wilkinson (1897,pp.152-3) who stated that after the last ice sheet had melted most of the corries and glens in the higher parts of Jura nourished valley glaciers as indicated now by plentiful moraines. With the exception of 5 localities in Jura mentioned by Anderson and Wilkinson and one in Scarba (Peach et al., 1911,p.59) the corries and glens were not described although they are identifiable on the relevant Geological Survey maps (sheet 27, 1900; sheet 36, 1909; sheet 28, 1911) since the areas of "moraine" are shown. Later Charlesworth (1955) suggested that 19 valley glaciers developed in Jura during his stage M or Highland Readvance (generally considered equivalent to the Loch Lomond Readvance in the western Highlands). He referred to the former glaciers by name and indicated their approximate distribution (Fig. 11) stating that his stage M glacier limits corresponded almost exactly with those of the Geological Survey.

Several points of interest emerge from the results of earlier studies. Firstly, the hypothesis proposed by Binns et al. (1973) that the orientation of submerged troughs and the general pattern of striae orientations can be explained by changing directions of ice flow during different stages of glaciation is plausible. Secondly, the SE-NW striae orientations in S Jura and NE Islay indicate the former presence of a large mass of ice to the SW of the study area capable of preventing the W and SW fanning of ice that emanated from the western Highlands. Finally the studies have indicated that a major period of local valley glaciation occurred in Jura after the disappearance of the last ice-sheet. None of the previous studies however has described in detail any of the landforms produced during this period.

For the above reasons a limited study was undertaken of striae orientations and erratics in the study area. The results were combined with those of the Geological Survey to determine a more precise pattern

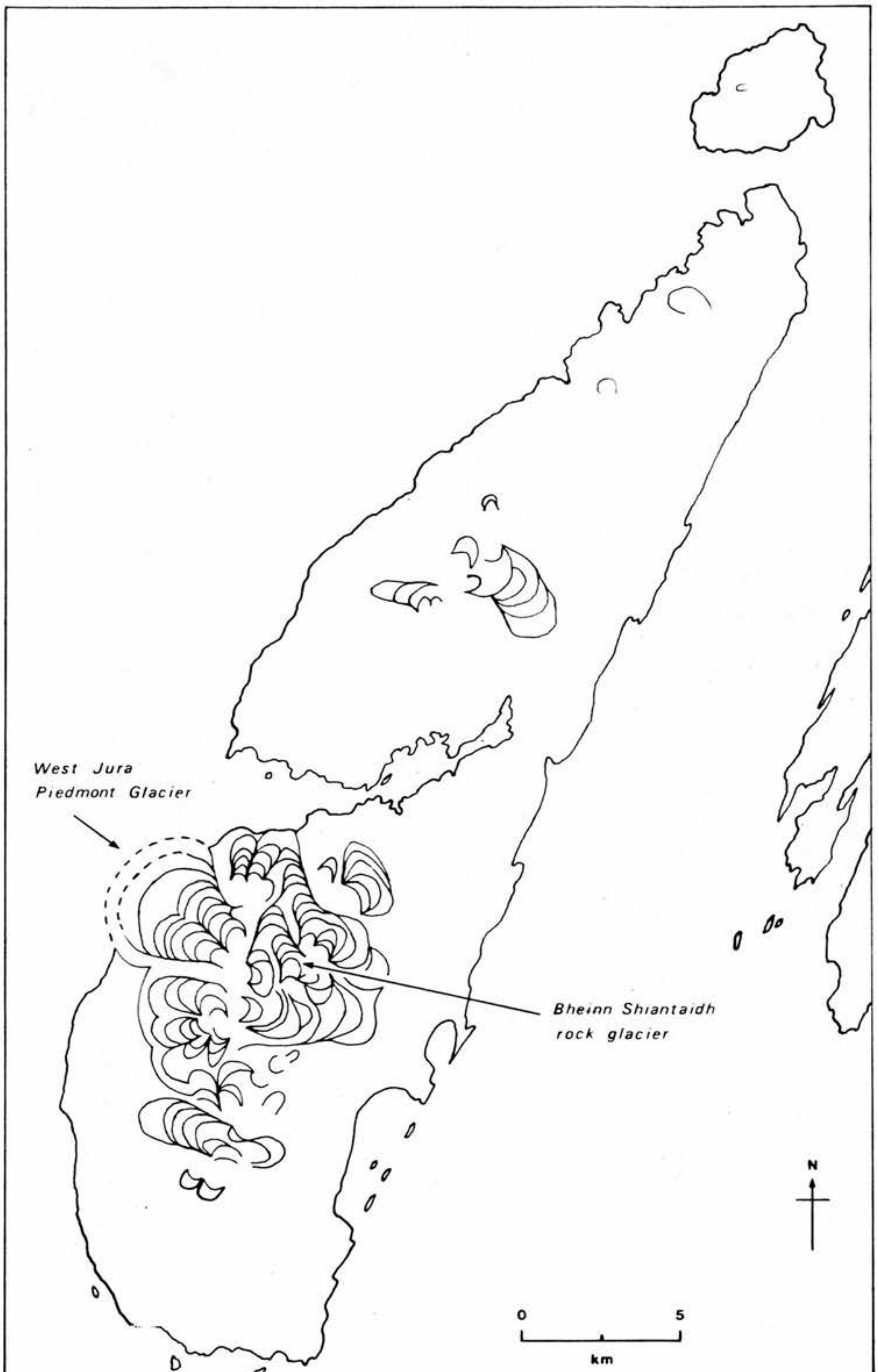


Fig. 11 Distribution of ice-limits as mapped by Charlesworth (1955). Location of Bheinn Shiantaidh rock glacier is also shown.

of ice-sheet movement. The nature and distribution of landforms affected by glacial erosion was also interpreted for this purpose. Finally the morphological evidence supposedly relating to the former existence of Loch Lomond Readvance glaciers in Jura, Scarba and NE Islay was examined.

2. Glacially sculptured terrain

In the Paps of Jura a large SE-NW-trending glacial breach separates Beinn an Oir (784 m) and Beinn a' Chaolais (734 m) (Fig.1) and suggests movement of ice in this direction. South of Loch Tarbert ice-moulded quartzite ridges occur at 550 m in the Paps of Jura while N of it they occur at 450 m, indicating that the ice that flowed W and NW in this area attained at least these altitudes. Since all bedrock above 550 m in Jura is covered by talus accumulations, the complete submergence of Jura by mainland ice, although extremely probable, cannot be demonstrated.

North of Loch Tarbert a series of valleys trends SE-NW and ESE-WNW (Fig. 12) and was most probably formed by glacial erosion in this direction. In the southern half of this area the absence of igneous dykes and sills has assisted the glacial formation of a wide plateau of rock-basin lochs and ice-polished quartzite rock surfaces. The northern limit of plateau terrain is marked by the Lussa and Grundale river valleys (Fig. 12), both of which trend for most of their length at right angles to the strike of the quartzite strata. As with many other smaller river networks in Jura, the orientation of these valleys clearly reflects the influence of glacial erosion along their major axes. In northern Jura, rock-basin lochs are infrequent while glacial erosion of less resistant epidiorite dykes has resulted in the formation

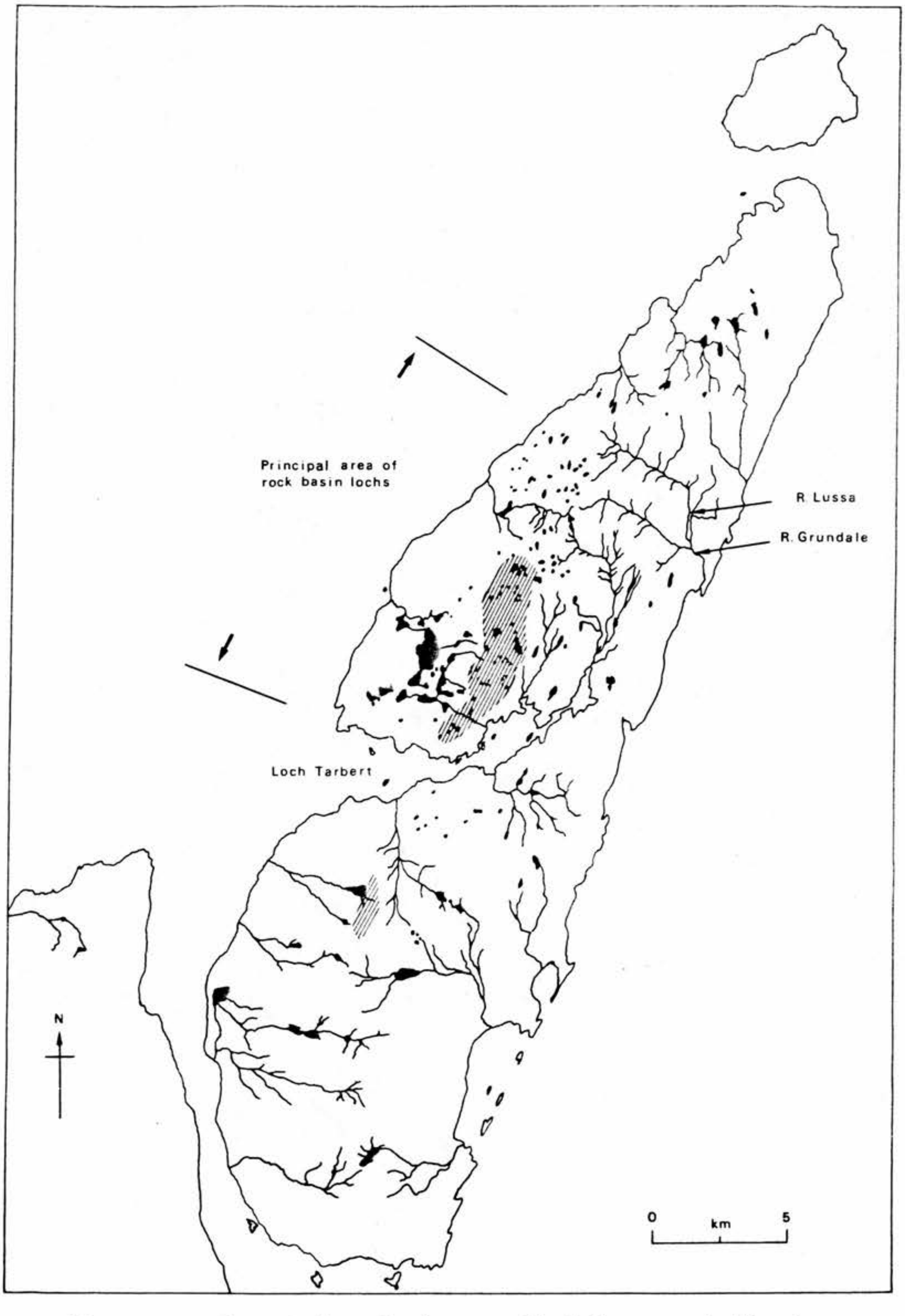


Fig. 12 Jura and NE Islay drainage. Shaded areas indicate principal areas of roche moutonné development. Distribution of rock-basin lochs is also shown.

of a series of deeply-incised valleys most of which trend between E-W and SE-NW (cf. Figs. 3 and 12).

Glacial erosion of the Jura quartzite has also resulted in the formation of numerous roches moutonnées that are best developed N of Beinn an Oir and on the Cruib plateau N of Loch Tarbert (Fig. 12). In both areas their development has been facilitated by the ESE dip of the quartzite strata aligned almost parallel to the main direction of ice movement. In both areas the roches moutonnees are oriented approximately ESE-WNW.

Together the pattern of glacially formed valleys and roches moutonnées indicates an overall movement of ice over Jura from SE-NW. The valleys at least cannot be attributed solely to late-Devensian ice-sheet erosion for they are presumably the cumulative result of the action of many ice-sheets.

3. Glacial striae

The distribution of all glacial striae recorded by the Geological Survey and the writer are indicated on Fig. 13. Of these over half were obtained by the writer. Most striae occur on protruding quartzite ridges while, in contrast, relatively few well-developed striae occur on the slates and phyllites of E Jura. Locally the striae vary considerably in direction yet over large areas reveal a uniformity in orientation.

Striae on Scarba suggest a movement of ice from E to W with local deflection around the island summit of Cruach Scarba (458 m). These summit striae show that the island was entirely overwhelmed by westward-moving ice. In N Jura striae demonstrate an overall ice

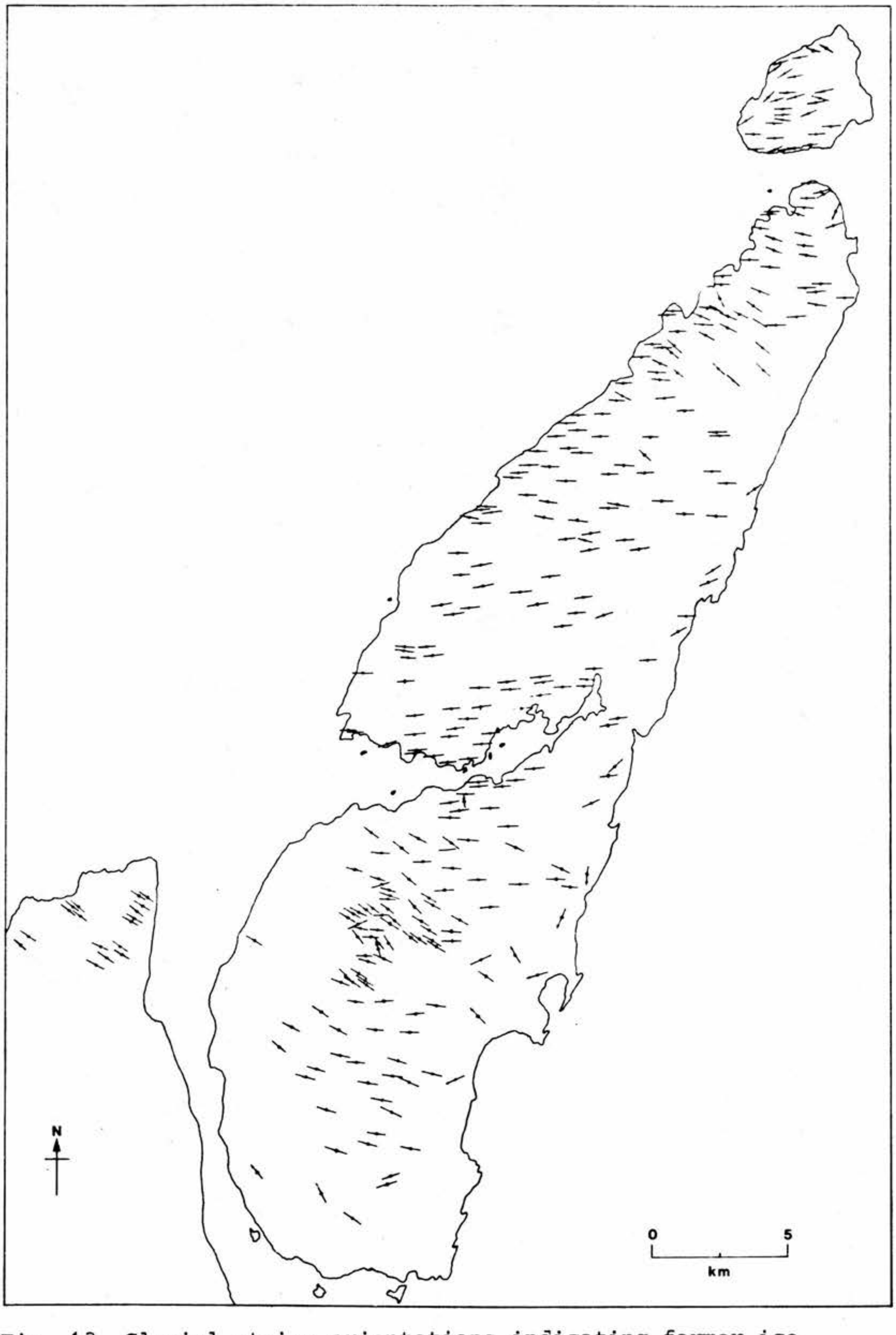


Fig. 13 Glacial striae orientations indicating former ice movement from east to west.

movement from between ESE to WNW and E to W, while on the Cruib plateau and on the S side of Loch Tarbert a consistent E to W movement of ice is indicated. In the Paps of Jura striae orientations show deflection of ice around the mountains and a general SE to NW realignment of ice flow on their western flanks with a major ice stream flowing SE to NW between Beinn an Oir and Beinn a' Chaolais. In NE Islay striae are similarly well-preserved on the quartzite strata and point to an overall direction of ice flow from SE to NW.

On coastal promontories of W Jura and W Scarba most striae are aligned E-W and do not indicate any deflection despite the presence offshore in the Firth of Lorn of NE-SW glacially overdeepened trenches that descend in places to over 200 m below sea level. Along the E coast of Jura glacial striae vary considerably in orientation below an altitude of 100 m before being succeeded inland by more uniform E-W orientations. It is therefore unclear whether the low coastal zones of E Jura were subject to glacial erosion by ice moving SW along the Sound of Jura since the absolute number of striae that occurs in this area is insufficient to warrant any firm conclusions.

4. Glacial erratics

The main source areas of erratics in conjunction with the principal directions of ice flow inferred from observations of erratics and striae by the writer and the Geological Survey are shown on Fig. 14.

The most common and easily identifiable erratic on Scarba and Jura is Old Red Sandstone, the source area of which is most probably the seabed of the Sound of Jura. On W Jura several schist erratics derived from the widespread Kintyre and Cowal schist series suggest ice movement from E or SE (Fig. 14). In addition E Jura slate and phyllite

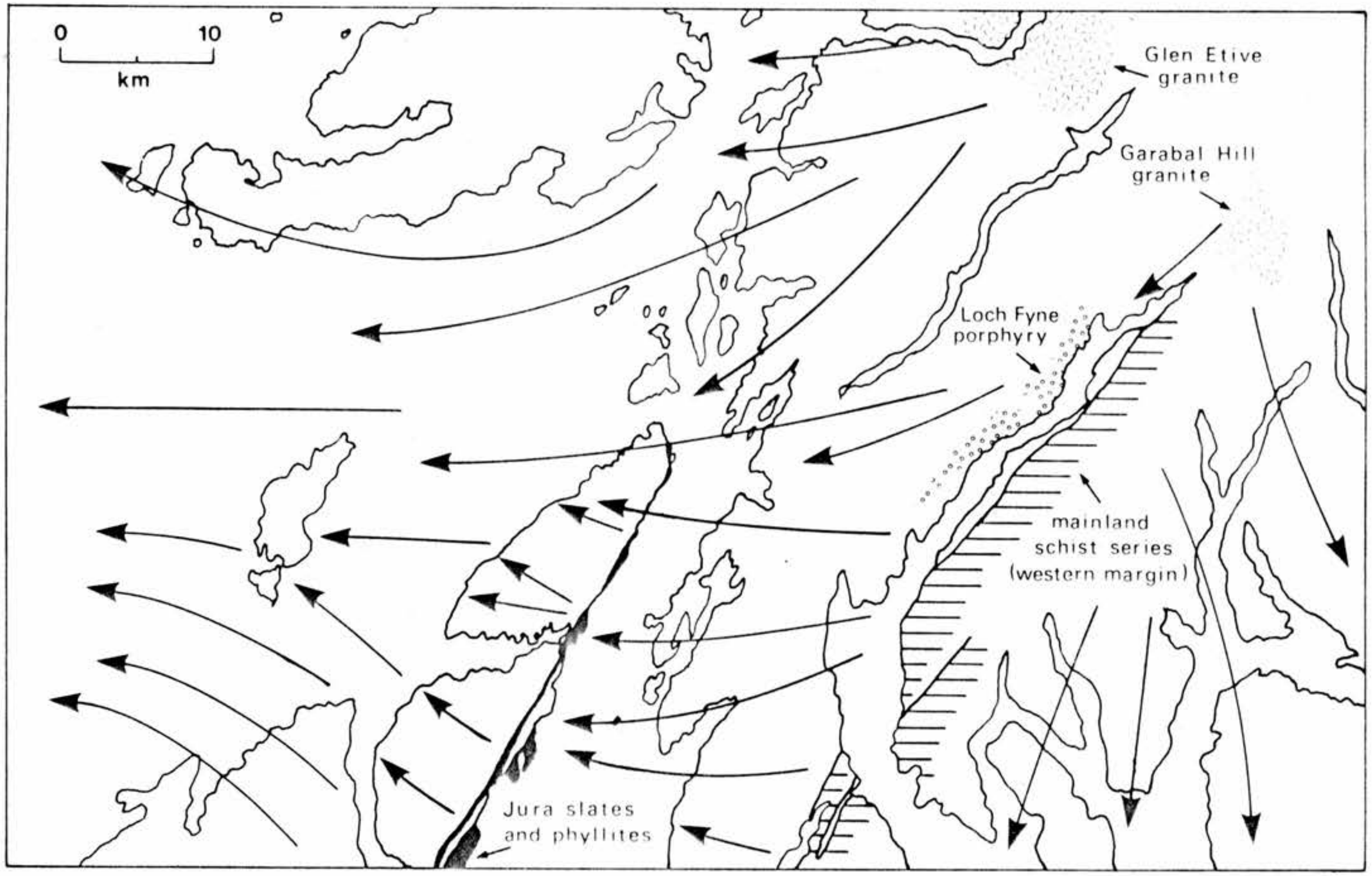


Fig. 14 Generalised direction of former ice-sheet movement and principal erratic source areas. Ice flow directions based on striae and erratic observations by the writer (Jura, Scarba, Islay and Colonsay) and the Geological Survey Gt. Br.

erratics are widely distributed in W Jura while in SE Jura several Loch Fyne porphyry erratics occur. On the E coast of the Sound of Islay granite erratics possibly derived from the Garabal Hill source were also noted (Fig. 14).

The distribution of glacial erratics in Jura, Scarba and NE Islay observed by the Geological Survey and the writer is consistent with the direction of ice-sheet movement inferred from striae observations.

5. Summary

From the above evidence it seems clear that Scarba and N Jura were over-ridden by ice flowing westwards across the Sound of Jura from the adjacent mainland. Despite the NE-SW flow of ice necessary to account for the orientation of the submerged glacial trenches, the lack of well-defined striae deflection on the W coastlines of Jura and Scarba suggests that little undertow of ice occurred in this area. The overall pattern of striae orientations supports the hypothesis of Binns et al. (1973) that the Firth of Lorn and Sound of Jura were glacially overdeepened during early and late stages of each Quaternary glaciation. The SE-NW striae orientations observed in S Jura and NE Islay suggest that during glacial maxima, the main ice flow here was from the SE. The most likely cause of this NW ice flow was the former presence of a major ice dome to the SW in N Ireland (Bailey et al., 1924,p.398).

Landforms produced during ice-sheet deglaciation and the Loch Lomond Stadial

1. Introduction

During field investigations of the glacial deposits in the study area the writer observed three sets of landforms whose morphological characteristics contribute substantially to the interpretation of the glacial chronology. Due to the geomorphological significance of the three sets of landforms it was decided to investigate each in detail. The results of these investigations are presented in the following pages. Thereafter the pattern of ice-sheet deglaciation and the nature of the Loch Lomond Stadial in the study area is discussed with particular reference to the three sets of landforms.

2. A fossil moraine, Sgriob na Caillich, W. Jura.

Introduction

The largest of the former valley glaciers identified by Charlesworth and the Geological Survey was the West Jura Piedmont glacier, which extended westwards from the Paps of Jura into the Sound of Islay (Fig. 11). Reference to this feature in the Geological Survey Memoirs is brief, Wilkinson (1897, pp.152-3) mentioning:

> "... at least one large lateral moraine of readvance age, which stretches from the shoulder of the mountain (Beinn an Oir) to near sea level."

In contrast Charlesworth did not mention the "lateral moraine" and simply depicted the piedmont lobe on a map (Fig. 11). Examination by the writer (cf. Dawson, 1979) revealed that large areas mapped by the Geological Survey (sheet 28, 1911) as moraine are composed of a stiff light orange lodgement till. The readvance lateral moraine of Wilkinson comprises a linear suite of parallel boulder belts 3.5 km in length (Figs. 15 and 16). The feature originates at 450 m at the western foot of Beinn an Oir, one of the Paps of Jura, and extends seawards to 30 m where it ends at a shallow lochan that lies at the junction between the feature and a high raised shoreline (Plate 5).

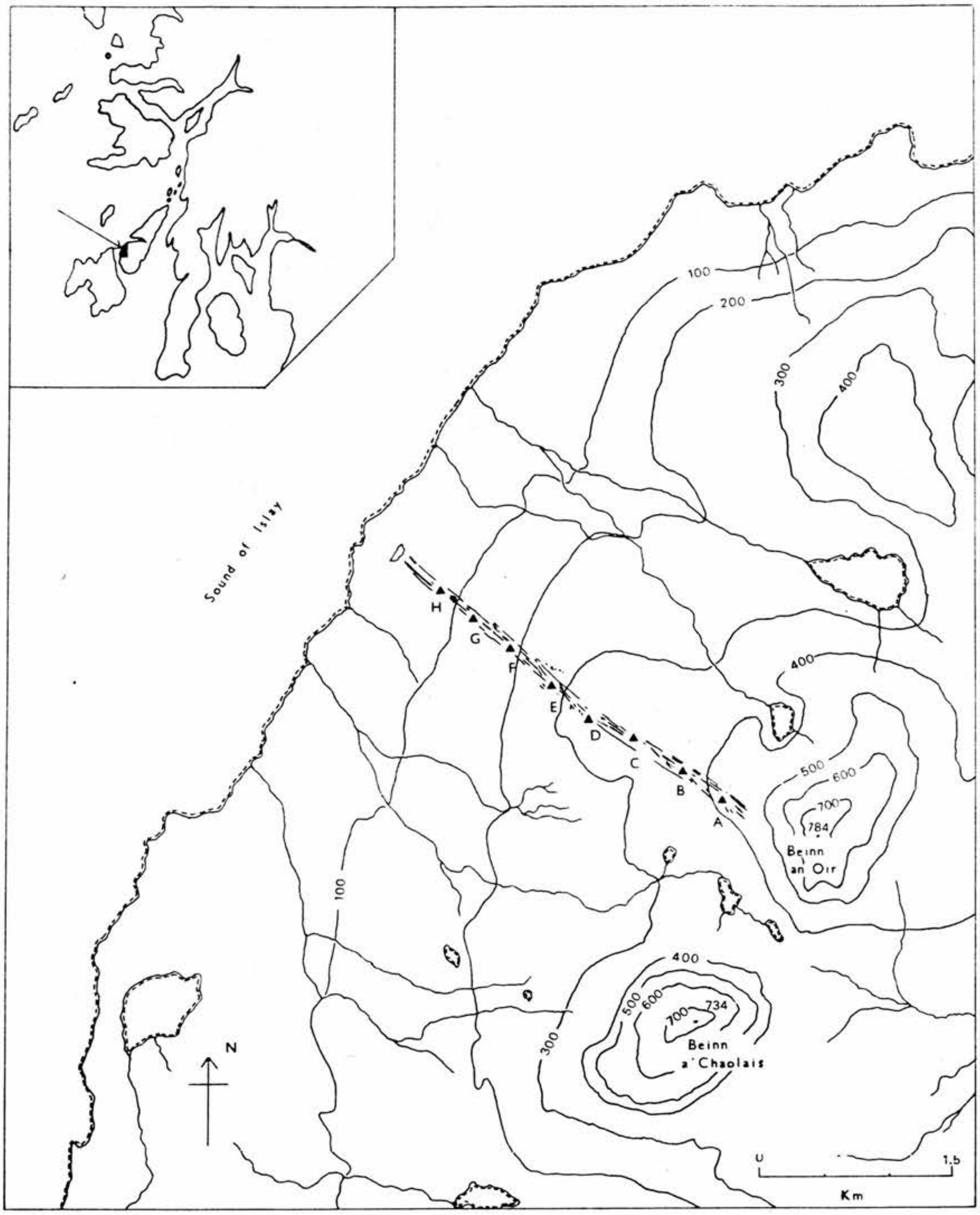


Fig. 15 Location of fossil medial moraine, Sgriob na Caillich, SW Jura. Sites of boulder measurements (A-H) are also shown.

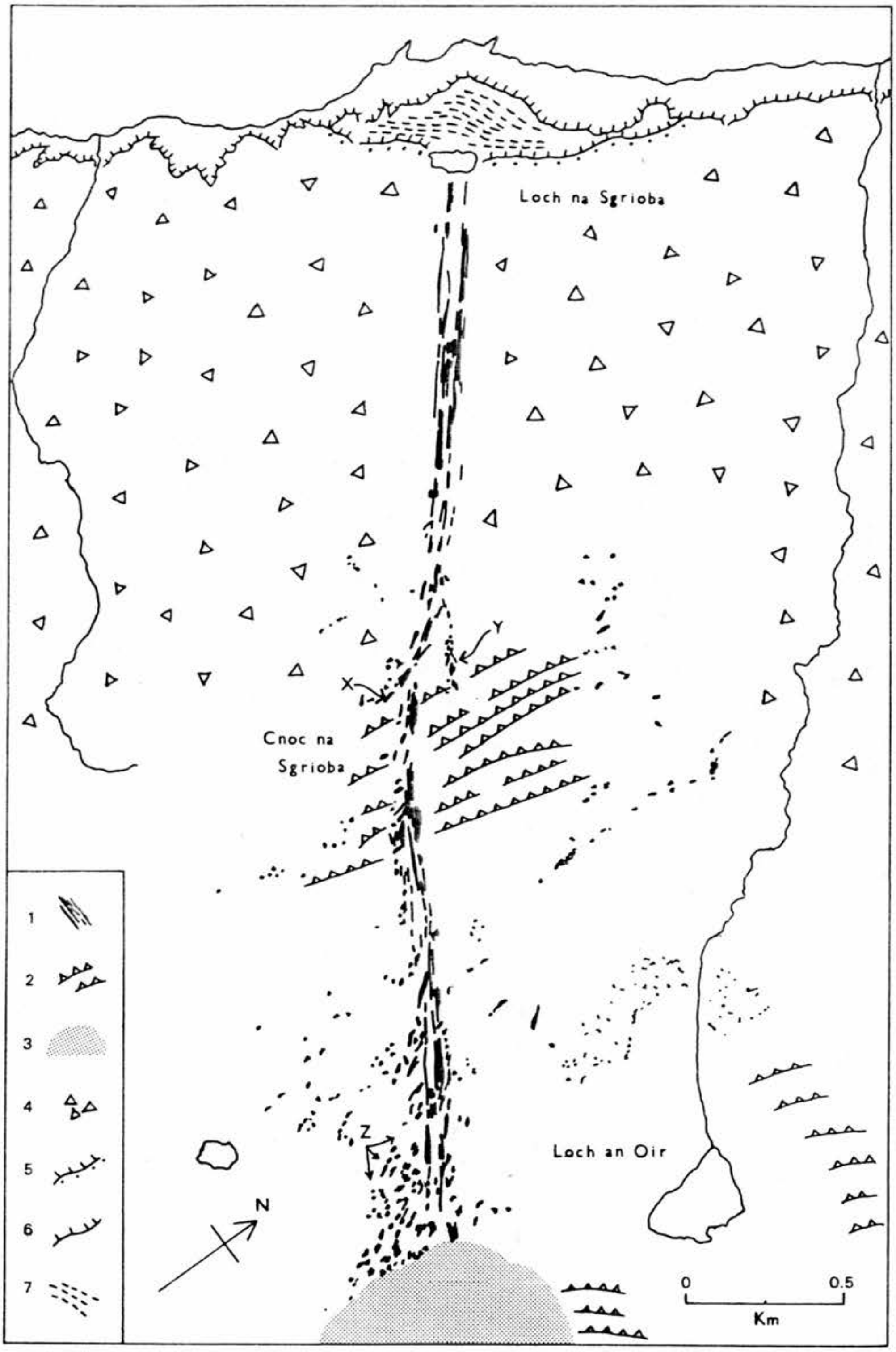


Fig. 16 Plan of fossil medial moraine, Sgriob na Caillich, SW Jura. 1. Surface boulders 2. Quartzite ridges 3. Talus 4. Glacial till 5. Lateglacial shoreline 6. Raised cliffs 7. Lateglacial beach ridges (vegetated).

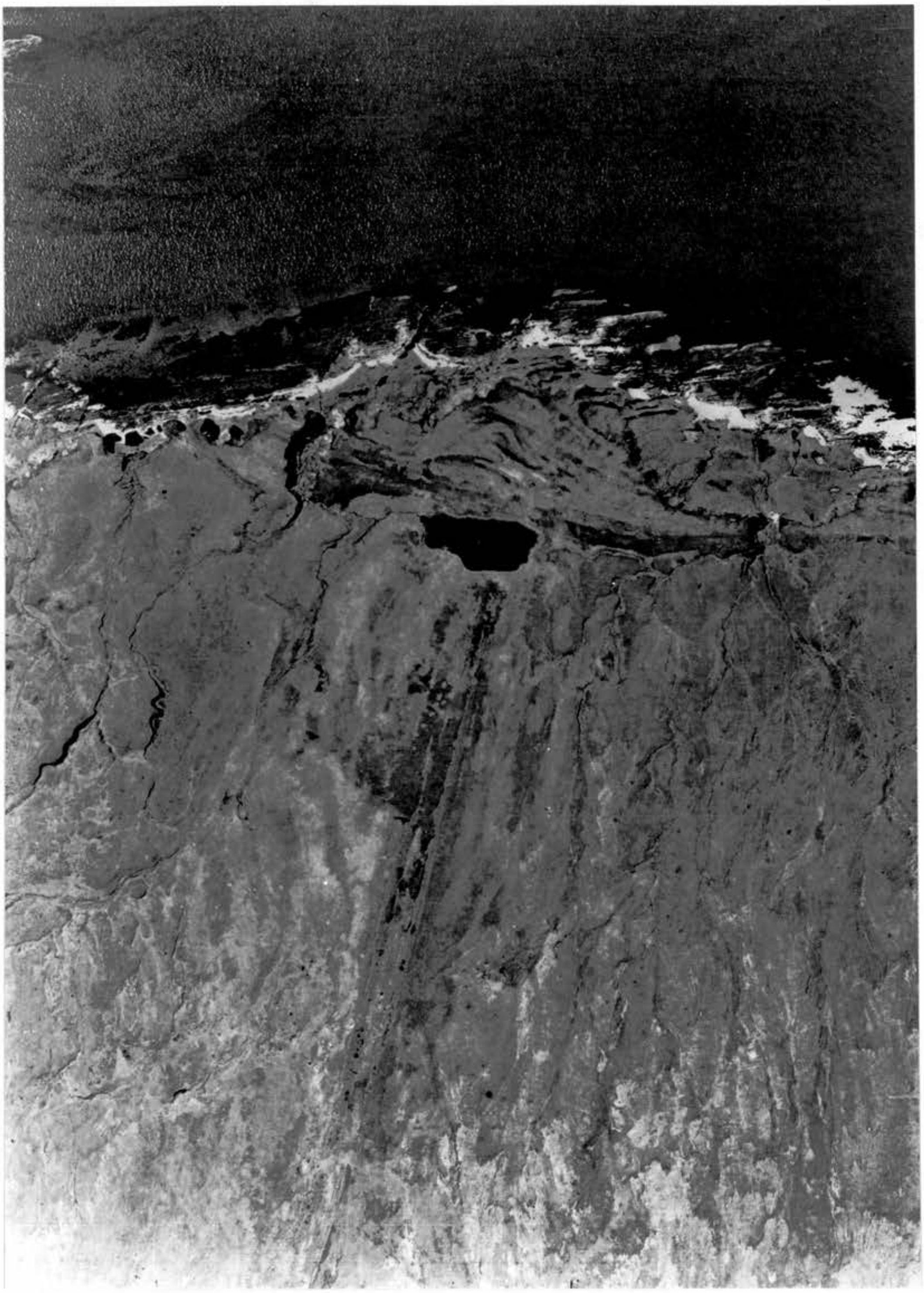


Plate 5. Aerial photograph of medial moraine (scale 1:10,000). Moraine is truncated seaward by raised shoreline. Low Rock Platform is also visible. Ministry of Defence (Air Force Dept.) photograph Crown Copyright.

The boulder complex is composed in places of up to 4 parallel lines of dominantly angular Dalradian quartzite blocks (the local bedrock), (Plate 6) each line rarely exceeding 27 m in width and 2.5 m in height, while the individual boulders which include several erratics, range from 0.2 to 1.3 m in length.

Morphology

At the western foot of Beinn an Oir the boulder belts descend gently before rising to pass over the rock outcrop of Cnoc na Sgrioba (360 m) (Fig. 16). Seaward of this ridge the boulders lie on top of an increasingly thick cover of till until at 30 m they are truncated by a low cliff and raised shore platform both of which are composed of till. At the junction of the till platform and the boulder belts is a small lochan (Loch na Sgrioba) (Plate 7) which is impounded by a series of raised shingle ridges that mantle the platform.

At no point along the length of the boulder belts do the accumulations exceed 2.5 m in height while the areas between each belt are occupied by scattered vegetated debris. The spacing between each belt, although varying slightly along the length of the feature, nowhere exceeds 50 m. For most of its length the junction between each belt and the vegetation cover exhibits little variation in relief, yet in places it is characterised by small boulder "cliffs" up to 2 m in height. The main boulder belts are oriented parallel with each other and exist as separate units that rarely merge together. Coalescence of belts is limited to the crest and flanks of Cnoc na Sgrioba where the entire feature changes slightly in direction (Fig. 16).

The boulders in the belts are angular and bear no evidence of striation or ice-moulding. They range from 0.2 to 1.3 m in length and contrast

Plate 6. Angular quartzite boulders of medial moraine.

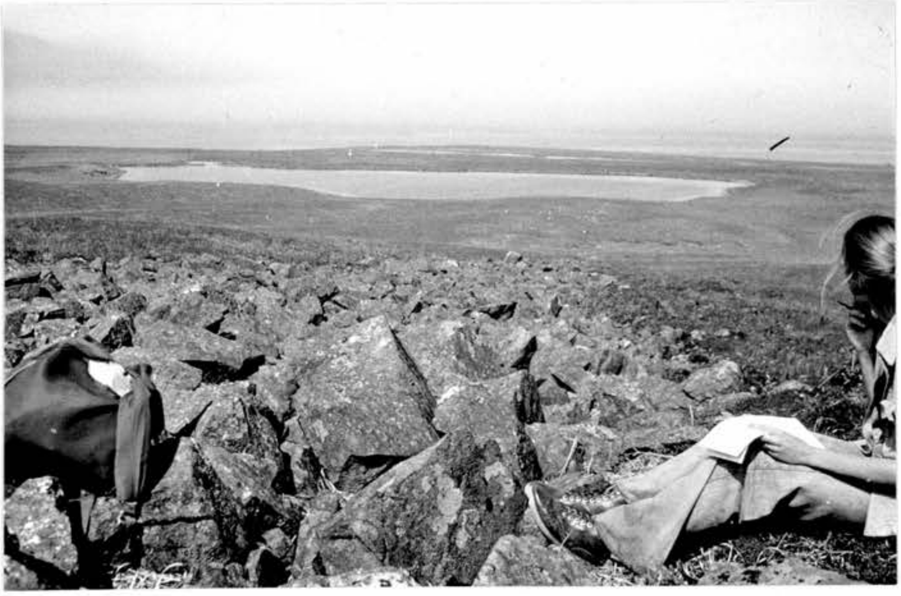


Plate 7. Parallel boulder belts of medial moraine. In middle distance is Loch na Sgrioba.

markedly with the generally smaller quartzite blocks that are found in local till exposures. Additionally, the mean diameter of boulders measured at 500 m intervals along the feature decreases seaward 7 cm/km (r= 0.91, signif.= 0.002) (Table 2). Although the feature is almost entirely composed of quartzite blocks, several E Jura slate and phyllite erratics are also present and these similarly exhibit no sign of ice-moulding or striation.

TABLE 2

Boulder size variation along the length of the feature. Values in cm. Sites located on Fig. 15 N=50.

	A	B	C	D	E	F	G	H
a axis	54.96	53.72	53.04	48.08	39.10	44.62	41.02	39.73
b axis	35.10	35.36	35.48	32.46	23.06	28.74	25.16	24.64
c axis	22.24	20.34	22.80	20.80	20.02	19.46	16.06	17.15
a+b+c/3	37.43	36.47	37.11	33.78	27.39	30.94	27.41	27.20

At the western foot of Cnoc na Sgrioba two smaller belts of boulders merge into the main feature. On the SW flank of the rock ridge a stream of boulders (Fig. 16 (X)) 100 m in length joins the main belt, while a second stream (Fig. 16 (Y)) that originates on the western flank of Cnoc na Sgrioba extends for almost 400 m as a broad scatter of boulders before also merging into the main boulder belt.

Farther upslope boulder accumulations are more widespread, notably between Cnoc na Sgrioba and the talus slopes of Beinn an Oir (Fig. 16 (Z)). Here scattered blocks fan SW from the main belts and extend upslope towards the col that separates Beinn an Oir and Beinn a' Chaolais. In contrast, surface boulders are absent NE of the main belts except in an area seaward of Loch an Oir.

The suggestion by Wilkinson (1897) that the feature is a lateral moraine is rejected since not only is the feature almost straight but also the belts are located on an exposed hillslope far from any valley. The extreme angularity of the boulders and their lack of ice-moulding or striation suggest that it is unlikely that the material was derived sub-glacially. Moreover, if the boulders had been transported sub-glacially it would be difficult to account for their being considerably larger in size than quartzite blocks found in local till exposures.

Numerous observations on active glaciers and piedmont ice lobes (Ray, 1935; Sharp, 1949) have shown that angular debris is frequently indicative of a supra-glacial origin. Since the boulders originate at the junction of the Beinn an Oir - Beinn a' Chaolais and the western foot of Beinn an Oir, the simplest hypothesis is that the feature is a fossil medial moraine supplied by subaerially weathered material from the slopes of Beinn an Oir.

Measurements of 400 boulder orientations at 8 regularly spaced sites along the feature revealed no dominant block orientation although it must be noted that the orientation of blocks may have been considerably influenced by the distribution of adjacent blocks during deposition. Although relatively thin medial moraines on active glaciers do not retain their clarity and linearity during ice decay (Small and Clark, 1974), it is unclear whether thicker accumulations of medial moraine debris are similarly destroyed. If the debris was deposited supra-glacially the glacier ice must have been sufficiently active to flow over the Cnoc na Sgrioba ridge while the clarity and linearity of the feature (Plate 8) may in part reflect the presence of unusually thick localised accumulations of supra-glacial debris.

Plate 8. Medial moraine inland from Cnoc na Sgrioba illustrating the linearity of the feature.

Plate 9. Western fragment of Coir Odhar moraine. Behind moraine are cliffs of High Rock Platform.

The change in orientation of the main boulder belt on the ridge crest and flanks of Cnoc na Sgrioba indicates the control exerted by the rock ridge on ice movement. The existence of shorter boulder belts, notably those seaward of Cnoc na Sgrioba, that merge into the main debris stream are more difficult to explain. However, since the two shorter boulder belts originate on the seaward flank of this ridge, it is possible that during a later stage of ice decay the smaller belts were also formed subaerially during ice flow around the Cnoc na Sgrioba nunatak. Since the boulder belts commence at the foot of Beinn an Oir, the angularity and source of the debris can be adequately explained by subaerial frost-riving while supra-glacial transport may be responsible for the progressive seaward reduction in boulder size. It is thus proposed that a relatively thin yet active ice mass transported supra-glacial debris derived from the Beinn an Oir nunatak at least as far seaward as Loch na Sgrioba and that the parallel alignment of boulder belts developed on the ice surface was not destroyed during ice decay.

3. The Coir Odhar moraine, NE Islay.

McCann (1964) described a terminal moraine in NE Islay that he considered to have been formed during the Highland (Loch Lomond) Readvance. McCann noted that the moraine occurs on top of a high coastal rock platform (Plate 9) and concluded that it represented the outer margin of a valley glacier that flowed seaward from a corrie located farther inland (McCann, 1964,p.5). He stated that, since raised beach deposits were incorporated within the moraine, a readvance of ice had occurred in NE Islay that was contemporaneous with the Highland Readvance identified in Jura by Charlesworth.

McCann (1964,p.5) stated that,

> "... the outer face of the morainic ridge at 77 feet (23.5 m) above high water wark is unmodified by marine erosion showing that the sea must have fallen below this level before the onset of the readvance of the ice".

Later, Synge and Stephens (1966,pp.107-8) concluded that the moraine was,

> "... one of a series of drift ridges deposited by the general glaciation on this coast ... the seaward edge of this "moraine" is the erosion scarp, or cliff, of the Late-glacial marine limit. An accumulation of rounded beach gravels occurs at the foot of this small cliff at 96-99 feet (29.3-30.2 m) O.D. ... (the) marine limit along this stretch of coast is uniform in height, and uninterrupted by any later glacial phase."

Morphology

The feature forms two distinct NW-facing arcuate ridges that are separated by a small embayment 200 m in width (Fig. 17). On both sides of the embayment the ridges mantle a well-developed high rock platform. East of the embayment a well-defined ridge curves inland to the lower hillslopes (Fig. 17). On the W side of the embayment an arcuate ridge extends for 500 m to end at the base of the cliff backing the High Rock Platform (Plate 9).

In situ exposures of the material composing the E ridge occur at NR 400785. Here two sections reveal angular quartzite blocks embedded in a matrix of stiff orange clay. The deposits, together with the morphology of the feature, indicate clearly that it is a moraine. Rounded raised beach cobbles mantle the outer edge of the ridge. Careful investigation failed to confirm McCann's observation of raised beach cobbles embedded in situ within the moraine.

On both sides of the embayment the outer margin of the moraine is cliffed and the cliff forms the inner edge of a well-developed raised shoreline. This shoreline forms the marine limit in the area and is a

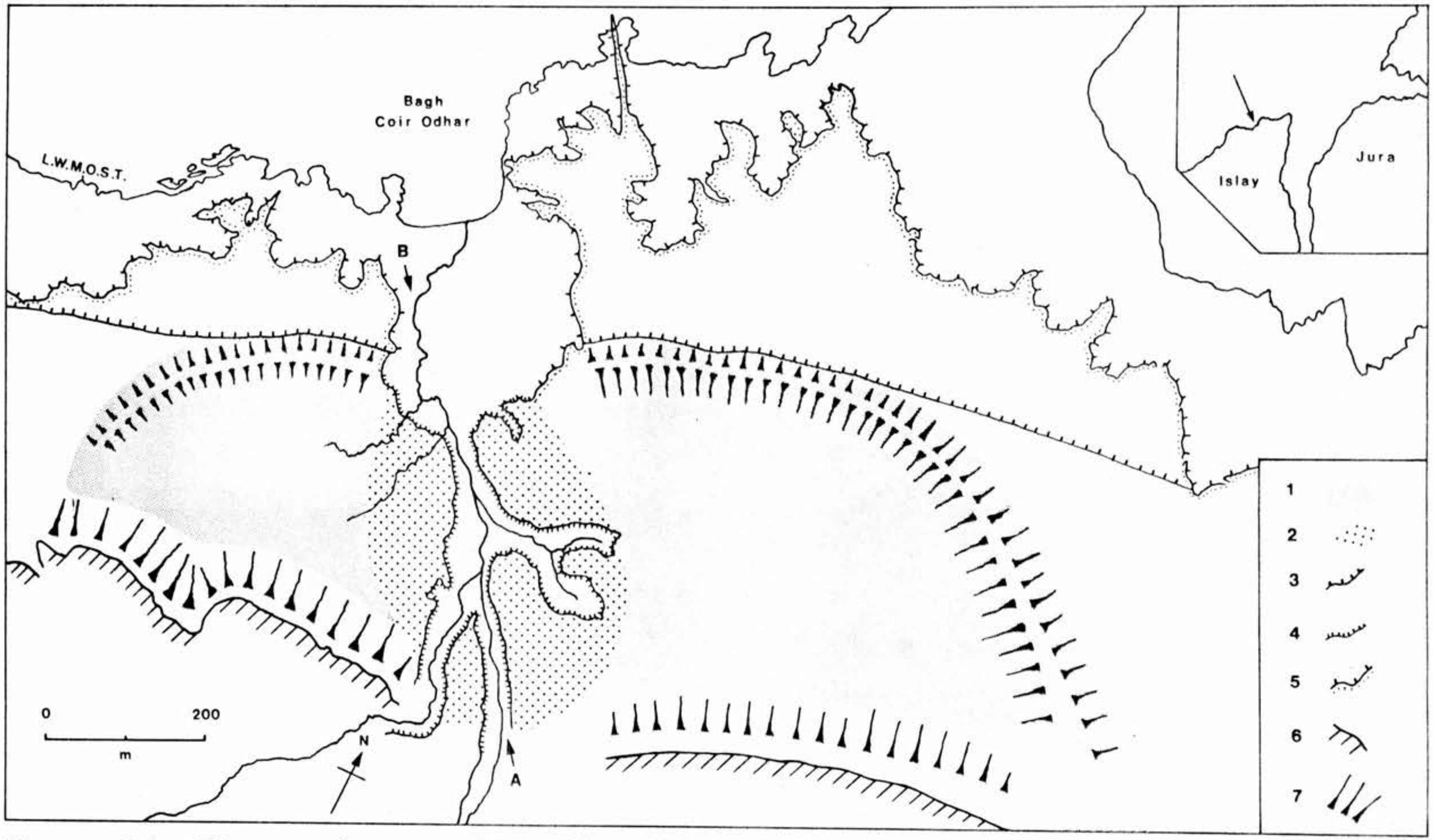


Fig.17 Coir Odhar moraine, NE Islay. Alluvial terrace fragment altitudes plotted in projection plane A-B are shown in Fig.18 and Table3 . 1. Moraine 2. Alluvial terrace fragments 3. Cliffline of highest lateglacial shoreline 4. Terrace edge 5. Fossil cliffline 6. High Rock Platform cliff 7. Major breaks of slope.

clear feature along considerable stretches of the NE Islay coastline.

The backing cliffs at the head of the embayment are composed of stratified sand and gravel deposits that have been deeply incised by several streams that drain into the embayment (Fig. 16). The stratified deposits are composed of sub-angular gravel generally less than 10 cm in diameter in a matrix of coarse grey sand. The deposits rise in altitude inland (Plate 10) until they merge into till-mantled quartzite ridges. In the area behind the moraine ridge the absence of exposures makes it impracticable to define the areas where the High Rock Platform is replaced by stratified deposits. The lateral extent of sand and gravel behind the moraine ridge is obscured by thick peat accumulations.

Levelled heights obtained on the sloping sand and gravel surface and on raised shoreline fragments are indicated on Table 3. The stratified deposits descend seaward from over 42 m to 30 m and have an average slope of 40 m/km (Fig. 18). The raised shoreline varies little in altitude with most fragments occurring between 26 and 27 m, while the highest level of raised beach cobbles is at 31.4 m (NR 400785).

The stratified nature of the inland deposits and the absence of large cobbles suggests that deposition occurred in an alluvial environment where water velocities were sufficient to transport only small-sized debris. The measured altitudes of the sand and gravel deposits imply that these deposits are graded to the raised shoreline altitude of 26-27 m (Fig. 18). It is therefore reasonable to suggest that since the sand and gravel surface is graded to the altitude of the raised shoreline, both are of the same age and since the outer edge of the

Plate 10. View, looking inland, from western fragment of Coir Odhar moraine. In the middle distance are sloping alluvial deposits.

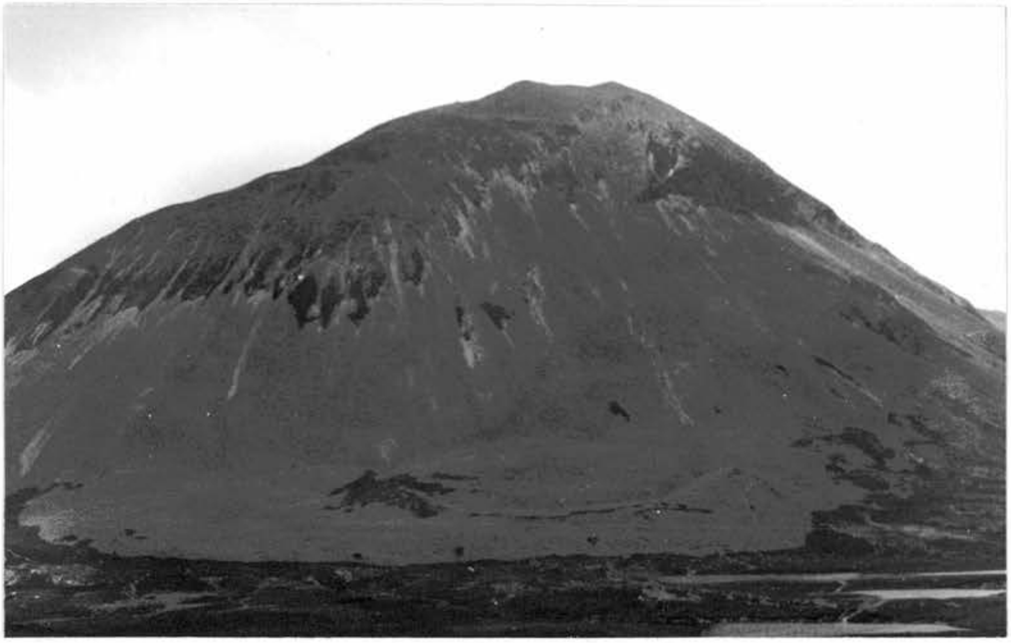


Plate 11. Fossil lobate rock glacier, Beinn Shiantaidh.

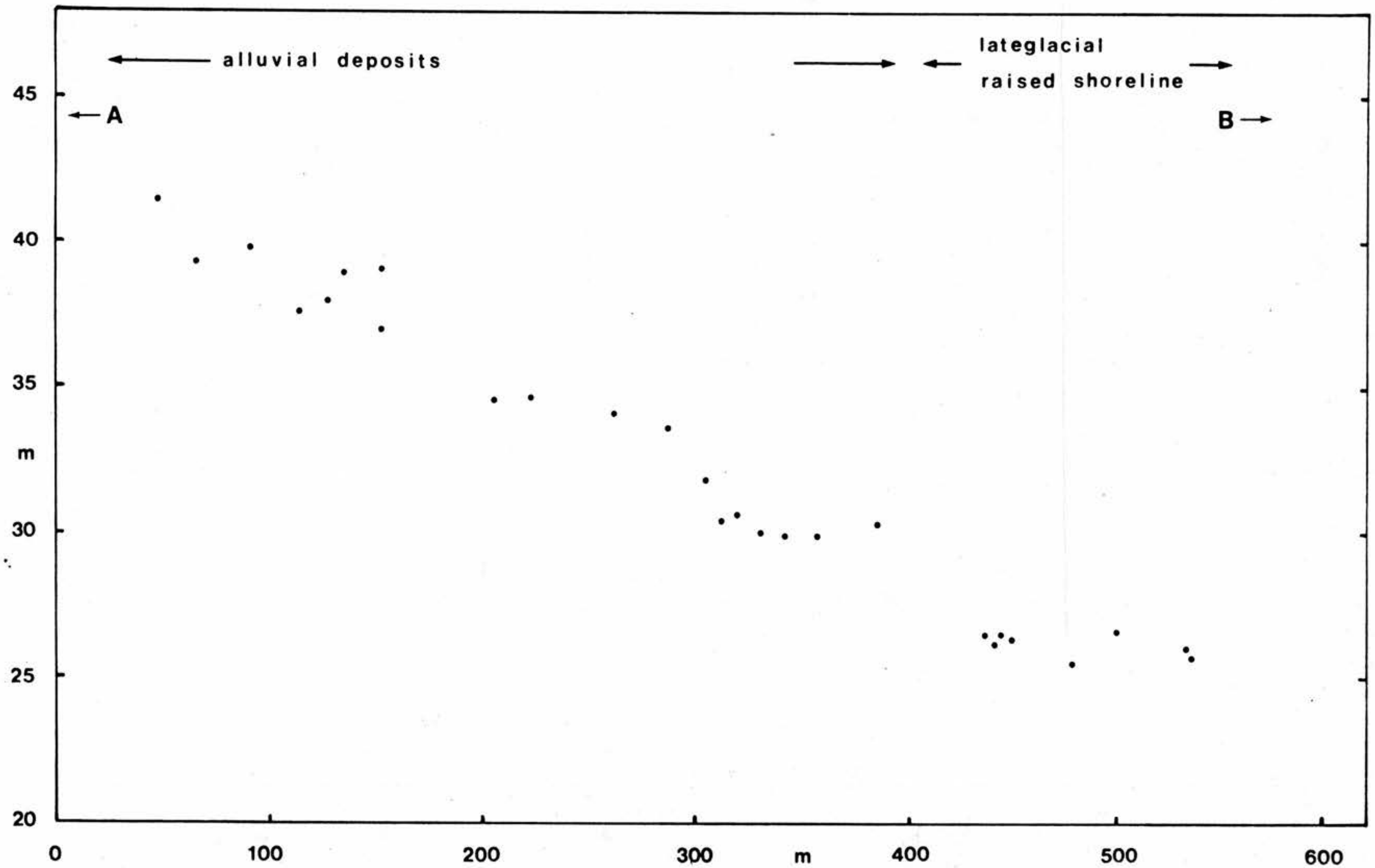


Fig.18 Profile of alluvial terrace fragment altitudes along projection plane A-B (cf. Fig.17), Coir Odhar, NE Islay. Altitudes of lateglacial shoreline fragments are also shown.

Table 3

Alluvial terrace fragment altitudes, Coir Odhar, NE Islay.

Site	Grid Reference	Altitude (m)
1	NR 40057837	32.1
2	NR 40087835	34.3
3	NR 40127833	34.9
4	NR 48247831	38.4
5	NR 40147824	39.2
6	NR 40177823	38.0
7	NR 40167820	40.1
8	NR 40137813	39.6
9	NR 40207816	41.9
10	NR 39907835	30.5
11	NR 39937834	30.2
12	NR 39957834	30.1
13	NR 39977831	30.6
14	NR 39947832	30.2
15	NR 39917829	30.8
16	NR 40007827	34.4
17	NR 40027821	34.8
18	NR 40037815	39.3
19	NR 40097840	33.9

* location of alluvial terrace shown on Fig. 17.

moraine forms the backing cliff of the raised shoreline, moraine formation occurred approximately contemporaneously with the formation of the shoreline.

4. A fossil lobate rock glacier- Beinn Shiantaidh, S Jura

Introduction

At the ENE foot of Beinn Shiantaidh, one of the Paps of Jura, a lobate accumulation of quartzite boulders occurs between 355 and 400 m (cf. Dawson, 1977) on the margin of an exposed col that separates Beinn Shiantaidh from its neighbouring summit Corra Bheinn (Plates 11 and 12). The boulder accumulation has an area of 45,000 m^2, the maximum width along the hill foot being 380 m and the maximum length 180 m (Fig.19). The constituent boulders, many of which exceed 0.5 m in diameter, are arranged in a series of arcuate ridges and depressions that culminate in a sharply defined frontal margin. On the E margin of Beinn Shiantaidh above the mass of debris are talus slopes of angular quartzite blocks that rise as much as 200 m towards the mountain summit (Plate 11).

Morphology

The frontal margin is sharply defined by a ridge of unvegetated angular boulders that slopes at approximately 20° towards the col surface. In the northern area of debris accumulation the continuity of the frontal margin is interrupted by numerous small transverse boulder hollows which are generally less than 1.5 m in depth and 3 m in diameter. In the southern area the outer rim becomes progressively subdued, while the boulders of which it is composed, although remaining large, form a less steep slope (Plate 11). Within this area there occurs a distinct outer ridge whose crest stands 5 m above the col surface. An arcuate depression flanks the inner edge of the ridge and fades northward where it is replaced by shallow hollows

Plate 12 Aerial photograph of Beinn Shiantaidh and fossil lobate rock glacier (scale 1:20,000). Dark areas on lobe indicate Calluna vegetation. Note extensive talus accumulations. Dashed line (top right) delimits crest of minor protalus ridge (NR 521749). Ministry of Defence (Air Force Dept.) photograph Crown copyright.

within a higher frontal ridge that stands 20 m above the col floor (Fig. 19). The inner margin of this ridge is delimited by a pronounced slope that leads into a semi-circular depression flanking the ridge for most of its length. The radius of curvature of the ridge is 85 m and represents the largest such feature upon the debris surface. Both ridge ends are overlain by talus aprons that slope upwards at 35° towards the mountain summit.

Perhaps the most notable feature of the debris accumulation is a deep semi-circular depression that flanks the frontal ridge along its inner margin (Fig. 19). The central area of the depression lies 6 m below the frontal ridge crest. The depression surrounds an area of extremely large boulders that rises abruptly above the arcuate hollow and forms a 20-25° slope along its inner margin. The boulders, most of which exceed 0.5 m in diameter, comprise an upper surface slope which, measured from the base of the talus to the frontal ridge crest, is generally 10-16° (Fig. 19).

Interpretation

The surface debris nowhere exhibits any evidence of ice-moulding or striation. That the debris is a product of landslide activity is rejected, since not only is there no visible hillslope scar but it is also difficult to explain the distribution of transverse ridges and furrows and the sharply defined frontal margin by this mechanism. The form of the debris surface together with its local plant cover (Plate 11) indicate that the feature is fossil. The possibility that the debris accumulation represents a composite protalus rampart is not favoured since its formation is contingent upon the unlikely possibility that a very large semi-conical snowbed surface extended from the mountain slope onto the exposed col surface.

The transverse ridge and furrow topography and hollows are almost identical to rock glacier structures described and illustrated from the Colorado Front Range (White, 1976, Figs. 4,5,6 and 7), Canada (Smith, 1973, Figs. 3,5 and 7), Alaska (Wahrhaftig and Cox, 1959), Switzerland (Barsch, 1969, Fig. 9) and elsewhere. Rock glaciers are normally composed of coarse debris moved downslope by interstitial or buried ice. Although most present-day rock glaciers are tongue-like in plan and possess mean lengths greatly in excess of widths, small arcuate rock glaciers have been observed whose widths exceed their lengths (Barsch, 1969; Smith, 1973; Wahrhaftig and Cox, 1959). Barsch noted three such rock glaciers in Macun, Switzerland, all of which had formed at the base of talus slopes. Similarly in Alaska, Wahrhaftig and Cox observed numerous lobate rock glaciers that extended out from the base of talus cones or aprons and which were characteristically broader than they were long. The distinction between rock glaciers that flow in the presence of interstitial ice and debris laden glaciers is largely artificial (Benedict, 1973; Whalley, 1974). Consequently it is difficult to evaluate the role of glacier and interstitial ice in rock glacier formation. This is particularly applicable to elongate rock glaciers where considerable movement of debris has occurred. However, small lobate rock glaciers are generally considered to form through the accumulation of interstitial ice and are unrelated to glacier formation (Wahrhaftig and Cox, 1959; White, 1976).

In view of the striking similarity to rock glaciers observed elsewhere, it is suggested that the Beinn Shiantaidh debris accumulation represents a fossil lobate rock glacier. However, during the formation and decay of the rock glacier, nivation patches at the

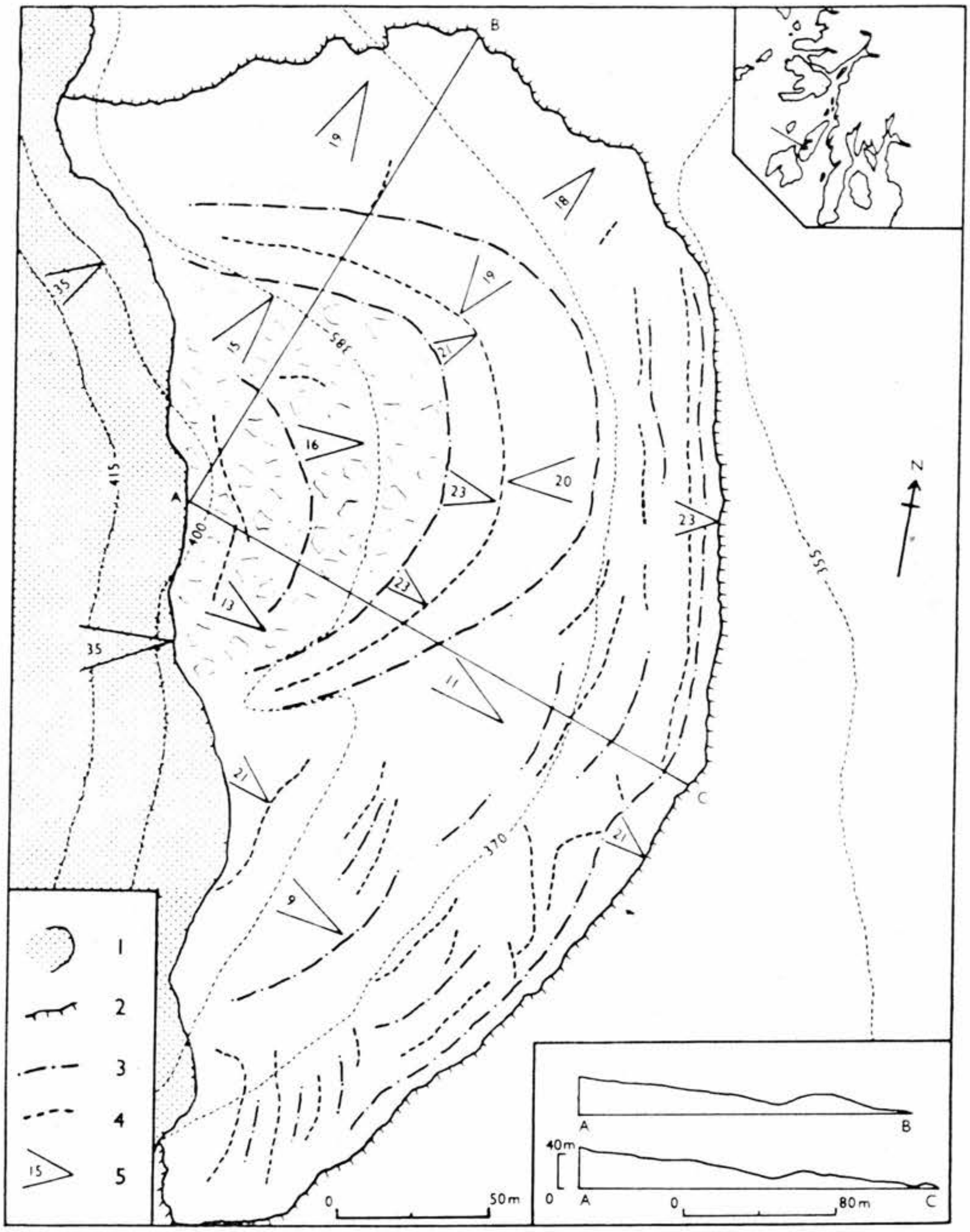


Fig. 19 Fossil lobate rock glacier by Beinn Shiantaidh.
1. Talus accumulations 2. Edge of rock glacier 3. Ridges
4. Depressions 5 Slope angles in degrees. Contour interval 15 m.
Area of large boulders shown diagrammatically behind main
depression.

foot of the talus slope may have resulted in small protalus debris accumulations that were incorporated within the rock glacier. Indeed beneath the talus slope that flanks the high north-facing buttress of Beinn Shiantaidh lies an arcuate ridge of angular boulders. The ridge, 50 m in length and composed of blocks generally 1 m in diameter, probably represents a fossil protalus accumulation that formed contemporaneously with the rock glacier (Plate 12).

Since bedrock is nowhere visible within the zone of debris accumulation , a minimum average thickness of 6.8 m for the rock glacier was derived by the projection of base levels from the outer debris margins to the inner edges of the feature. Using a porosity index of 0.4 (Wahrhaftig and Cox, 1959; White, 1971) a rock volume of 185,000 m^3 was obtaimed, giving a minimal estimate for the volume of material that was removed from the backing cliffs and supplied as talus to the rock glacier.

Wahrhaftig and Cox (1959) estimated cliff retreat above active rock glaciers by dividing rock glacier volume by its source area which they defined as the area upslope from the head of the rock glacier to the cliff face. The source area above the Jura feature measured in this way is 70,000 m^2. It may be argued however that the ratio between rock glacier volume and cliff face area of 20,000 m^2 is a more accurate estimate of cliff retreat. This value when divided into rock glacier volume results in a cliff retreat of 9.2 m. This figure may be regarded as a maximal estimate of cliff retreat during the period of rock glacier formation since later talus aggradation may have buried free-face areas that were formerly exposed to frost-riving during rock glacier formation. The aspect of the fossil rock glacier- almost exactly ENE- would have favoured the accumulation and persistence of snow and ice while its existence implies the contemporary presence of permafrost (Washburn, 1973).

Age and origin of the medial moraine, SW Jura and the Coir Odhar moraine, NE Islay.

The high shorelines that truncate the medial moraine at Loch na Sgrioba, SW Jura and the Coir Odhar moraine in NE Islay were considered by Charlesworth (1955), McCann (1964) and the Geological Survey (sheet 28, 1911) to be part of a presumed horizontal "100 foot" shoreline that fringed Jura, Scarba and NE Islay. This shoreline was considered synchronous with the stage M of the Highland Readvance (generally considered equivalent to the Loch Lomond Readvance in the western Highlands) and as a result both the Coir Odhar moraine (McCann, 1964) and the medial moraine (Charlesworth, 1955; Geological Survey, sheet 28, 1911) were considered to have been formed during this period. In neighbouring areas of the western Highlands outwash spreads formed at or near the limit of Loch Lomond Readvance glaciers demonstrate that contemporary sea-level was considerably lower than suggested by Charlesworth, McCann and the Geological Survey (Sissons, 1974; Gray, 1975b). It will be shown later (Chapter 8) that near Loch na Sgrioba and Coir Odhar mean sea-level during the Loch Lomond Stadial accorded approximately with present sea-level. Therefore the raised shorelines that truncate the medial moraine and the Coir Odhar moraine are part of a series of shoreline fragments that formed during lateglacial submergence contemporaneous with the decay of the late-Devensian ice sheet. Thus the raised shoreline evidence indicates that both features were formed during the waning of the late-Devensian ice sheet.

It is thus believed that the Coir Odhar moraine formed at the seaward margin of an ice mass that lingered in the Coir Odhar valley during general deglaciation of the late-Devensian ice sheet. Glacial meltwater issuing from decaying ice in the Coir Odhar valley resulted in the seaward transport and deposition of the stratified alluvial deposits. At approximately the same time, incursion by the sea of low-lying ice-free areas resulted in the formation of the raised shoreline along the outer margin of the moraine.

Similarly it is considered that the medial moraine in SW Jura formed during the final stages of the decay of the late-Devensian ice sheet. During this period a relatively thin yet active ice mass transported supra-glacial debris frost-rived from the Beinn an Oir nunatak at least as far as Loch na Sgrioba. The final melting of ice was associated with a marine transgression of the coastal zone and the truncation of the medial moraine by a high lateglacial shoreline.

Evidence for and against a readvance of ice during the Loch Lomond Stadial in Jura, Scarba and NE Islay.

1. Age and origin of the fossil rock glacier, Beinn Shiantaidh, S Jura.

Since the fossil lobate rock glacier in the Paps of Jura was probably formed by the growth of interstitial ice, its presence implies the contemporary existence of permafrost. Since permafrost last existed in Scotland during the Loch Lomond Stadial (Sissons, 1974b) the simplest interpretation is that the rock glacier was formed during this period. It is therefore believed that the massive accumulations of quartzite talus that flank many of the Jura hills are also of the same age and were formed by frost-riving in the severe periglacial conditions that prevailed during this period.

By assuming a maximum possible duration of 1,000 years for rock glacier development during the Loch Lomond Stadial (Sissons, personal communication), the rock glacier volume of 185,000 m^3 implies that the maximum rate of debris supply to the rock glacier was 185 m^3/yr. By dividing the rock glacier volume by its source area of 70,000 m^2 the minimum vertical thickness of bedrock removed during 1,000 yrs. is 2.6 m or 2.6 mm/yr. If the ratio between rock glacier volume and the cliff face area of 20,000 m^2 is used, the maximum rate of cliff retreat is 9.2 mm/yr. The calculated rates of cliff retreat may be compared with values of 0.71 - 1.06 mm/yr. obtained by White (1971) and the estimate of 3.0 mm/yr. by Wahrhaftig and Cox (1959) in Alaska. Since the frontal ridge is a maximum distance of 180 m from the talus foot, a minimum average rate of flow for the central area of the rock glacier was 18 cm/yr. This rate is also comparable with flow rates observed on active rock glaciers (Barsch, 1969; White, 1971; Wahrhaftig and Cox, 1959).

The presence of the fossil rock glacier is apparently related to the favourable ENE aspect, the susceptibility of quartzite to frost-riving and the severe periglacial conditions in Jura during the Loch Lomond Stadial that were favourable for rock glacier formation. Since the fossil rock glacier is located well within the stage M glacier limits of Charlesworth (Fig.11) it is believed that the conclusions of Charlesworth are incorrect. In order to test the validity of the above hypothesis it was decided to examine briefly the supposed end moraines of readvance age mapped by Charlesworth and the Geological Survey.

2. <u>Previously mapped readvance glacier limits</u>.

The main area in which a readvance of valley glaciers was believed to have occurred is S Jura (Anderson, 1888; Peach <u>et al</u>., 1911;

Charlesworth, 1955). In this area it was claimed that 13 glaciers radiated from the high ground of the Paps of Jura (Fig.11). No morphological evidence was given for the former existence of these glaciers although Charlesworth stated (1955, p.881):

> "Abundant and highly inclined ribs of quartzite make it difficult to distinguish between moraines and solid ribs draped with plucked frost-shattered debris."

In the Paps of Jura the glacier limits shown on the map of Charlesworth and the Geological Survey have no equivalent on the ground surface. Not only are terminal and lateral moraines absent but the pattern of striae orientations (Fig. 13) and the absence of hummocky moraine within the proposed glacier limits imply that no valley glacier readvance occurred. Of the 13 readvance glaciers described by Charlesworth, the largest was the West Jura Piedmont Glacier (Fig.11) but as has already been shown, the medial moraine that occurs in this area (described by Wilkinson as a readvance lateral moraine) was formed during ice-sheet decay. In addition the Coir Odhar moraine, originally considered to have been formed during the Highland Readvance has been shown to have been formed during deglaciation of the late-Devensian ice-sheet. Finally the severe periglacial climate implied by the presence of the fossil lobate rock glacier is incompatible with the concept of widespread valley glacier development in the same area.

<u>Summary</u>

During the last period of general glaciation N Jura and Scarba were over-ridden by westward flowing mainland ice. In S Jura and NE Islay, the dominant movement of mainland ice was from SE-NW. The reason for a convergence in ice flow direction was most probably the presence to the SW of a major ice dome over N Ireland.

The glacially overdeepened trenches of the Sound of Jura and the Firth of Lorn that trend NE-SW do not relate to the dominantly westward flow of ice. No good evidence exists to indicate that the submerged areas were affected by ice undertow beneath the W- and NW- moving ice-sheet. The hypothesis of Binns et al. (1973) that the submerged trenches were most likely glacially overdeepened during the build-up and thinning of each Quaternary ice-sheet is the most satisfactory explanation so far proposed.

The glacial deposits described in the study area by earlier researchers were formed during late-Devensian glaciation and deglaciation. During the waning of the last ice-sheet a large moraine was formed near sea-level in NE Islay. Also during ice-sheet decay a large medial moraine was formed in SW Jura.

During the Loch Lomond Stadial severe periglacial conditions occurred throughout Jura, Scarba and NE Islay and in the Paps of Jura the cold periglacial climate was responsible for the formation of a fossil lobate rock glacier. The existence of the rock glacier, the medial moraine and widespread talus accumulations in Jura, Scarba and NE Islay indicate the susceptibility of quartzite to frost riving during the cold climatic conditions that prevailed during ice-sheet decay and the Loch Lomond Stadial.

Chapter 5

Modern Coastal Landforms and Processes

Introduction

A serious source of potential error in raised shoreline investigations is the manner in which raised coastal landform altitudes are related to former sea-levels. For example, beach ridges, coastal terraces and deltas each bear different altitude relationships to the former water plane. In published studies raised coastal terrace altitudes are usually considered the best indicators of the position of former sea-level. In contrast to coastal terraces that form in low/moderate wave energy environments, storm beach ridges are formed during high energy conditions and as a result their crest altitudes may be several metres above the position of the former sea-level. Frequently the altitude relationships between raised beach ridges and former sea-levels cannot be established. As a result storm beach ridges have been considered of little value in raised shoreline studies since their crest altitudes do not reflect accurately the position of former sea-level. In addition few measurements of modern storm ridge crest altitudes have been made (Sollid et al., 1973) and consequently the causes of the regional variation of ridge crest altitudes are not understood.

As stated in Chapter 2 the raised coastal features that occur in W Jura and NE Islay have no equivalent anywhere in western Europe. Raised marine terrace fragments are well-developed along long stretches of coast while extensive staircases of unvegetated raised shingle ridges are common (Plate 4). In order to understand more clearly the origin of such ridges and their altitude relationships with former sea-levels, a detailed study was undertaken of modern storm beach ridges that occur in the present shore zone. The study of beach ridge origin and altitude forms the main section of this chapter.

The landforms that reflect most accurately the position of sea-level on the present coastline are sand beaches and seaweed lines. Altitude variations of these landforms reflect regional variations in low/moderate wave energy activity. The altitudes of sand beaches and seaweed lines were measured throughout the coastal zone of W Jura, Scarba and NE Islay and the results used to establish an altitude relationship between raised coastal terraces and former sea-levels. The analysis of sand beaches and seaweed lines forms the preliminary section of this chapter.

Low/Moderate wave energy coastal landforms

1. Sand beaches

Sand beaches are well developed at only four localities in W Jura and NE Islay. They occur in W Jura at Shian Bay (NR 530875), Gleann Muc (NR 688002) and south of Corpach Bay (NR 565913) and in NE Islay at Bagh an da Dhorius (NR 413798). At each locality the lower limit of land-based vegetation was levelled along the length of the beach and the following values obtained:-

	Altitude range (m)	Mean altitude (m)	No. of values
Bagh an da Dhorius	2.56-2.44	2.49	6
Shian Bay	2.75-2.51	2.58	5
Gleann Muc	2.44-2.29	2.37	4
S Corpach Bay	2.66-2.54	2.61	4

The measured altitude of the lower limit of land-based vegetation defines the inner edge of each beach terrace fragment and corresponds approximately with the observed altitude of high water mark of ordinary spring tides. The values obtained may be compared with values of 2.4 ± 0.4 m O.D. for the same landform measured by Gray (1975a, p.18) farther north along the Firth of Lorn. Although vegetated raised coastal terrace fragments may either be raised beaches (sensu stricto) or erosional terrace fragments it is reasonable to assume that the measured altitudes of the terrace fragment inner edges correspond closely to the position of former sea-level. It would therefore appear that if present tidal ranges in the study area are similar to those that prevailed during periods of raised shoreline formation, the altitude of the inner edge of a raised coastal terrace fragment corresponds approximately with the position of former high water mark that in Jura, Scarba and NE Islay at present, occurs approximately at 2.4 m.

2. Seaweed

The most conspicuous seaweed line occurring in W Jura, Scarba and NE Islay is that formed by the dark brown algae Pelvetia caniculatus (sp.). It is the highest occurring fucoid in the British Isles (Lewis, 1976, p.64) and develops most easily on rough rocky surfaces, particularly on rock ledges and beach shingle. It has been stated that Pelvetia is best developed in areas of high tidal range (Lewis, 1976, p.273) where its reproductive output and hence stability is greatest. In the study area the "Pelvetia line" was observed to be deposited during the maximum levels of spring tides and as a result measured height variations of this line indicate the highest level of wave uprush at individual localities during this period. Due to abnormally calm sea conditions that prevailed during the period of ground survey (May-September 1977) and the complete absence of storms during this period, the Pelvetia line forms a useful reference level for the regional study of the maximum altitudes at which wave activity occurs in a low/moderate wave energy environment.

The altitudes of Pelvetia fragments were measured along the exposed shores of NW Jura and along the more sheltered coasts of Loch Tarbert, Scarba and SW Jura (Fig. 20) (Table 4). Since exposure to wave action is primarily affected by the distribution of rocks in the foreshore and offshore zones, regional variations in Pelvetia altitude cannot be related accurately to distinct bay, headland and straight coast environments. Instead, since regional variations in Pelvetia altitude reflect variations in nearshore wave activity, the heights obtained were compared with the horizontal distances between each surveyed Pelvetia line and low water mark of ordinary spring tides (measured from Ordnance Survey 1:10,560 maps) (Table 4). The relationship between 51 Pelvetia altitudes and the corresponding intertidal zone widths was analysed by linear regression. A strong correlation was obtained between the two variables ($r=-0.73$; Signif.$=0.001$) (Fig. 21).

Since Pelvetia altitudes decrease as the intertidal zone becomes wider the results most probably indicate the role of the nearshore seabed surface in affecting the distance between the zone of wave breaking and the deposited seaweed fragments. Where the intertidal foreshore is wide, the zone of wave breaking occurs farther seaward and as a result the vertical uprush of water is diminished. Conversely, Pelvetia is deposited at high levels in areas where the intertidal zone is narrow and waves are able to break closer inshore.

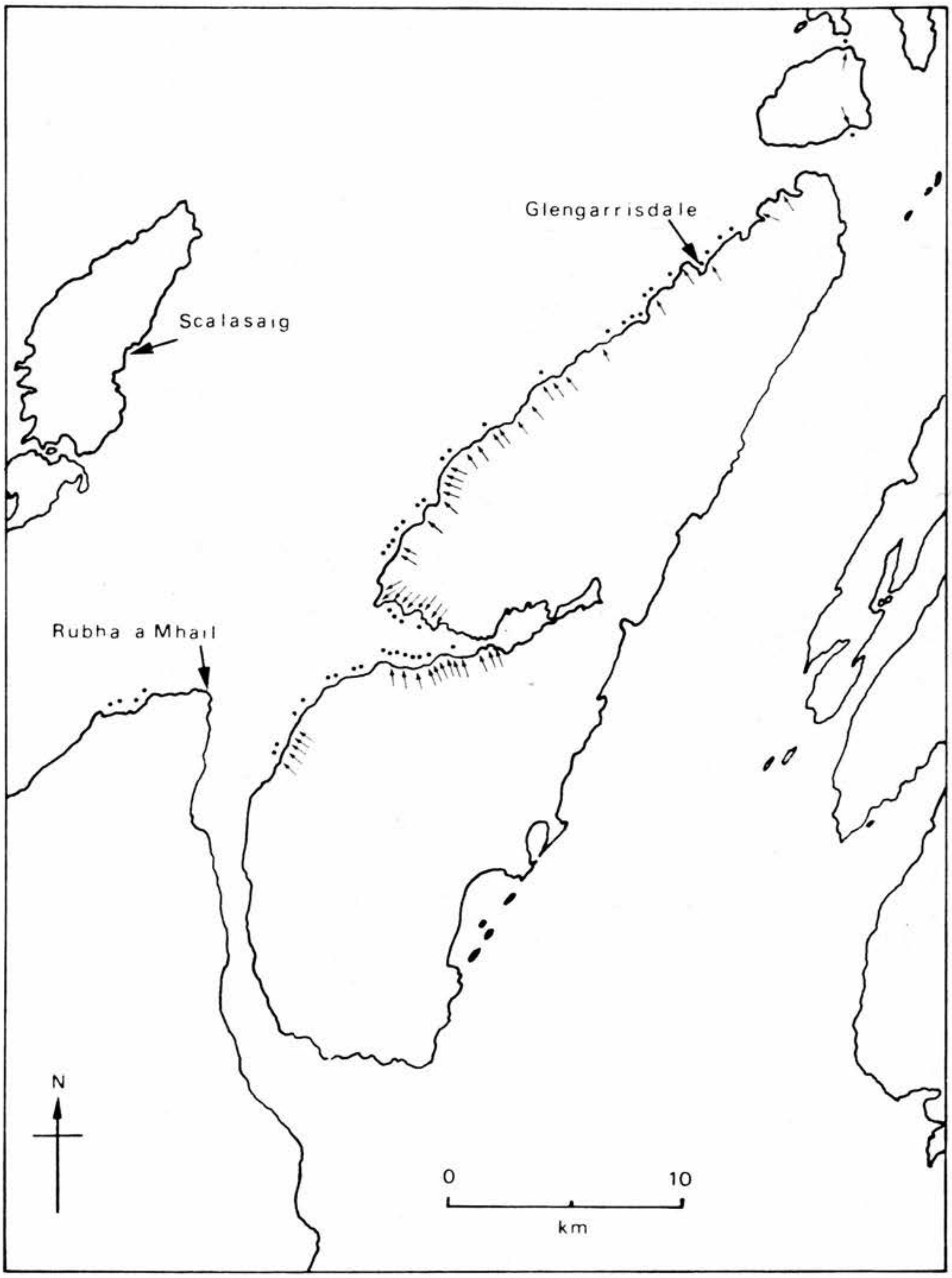


Fig. 20 Sites of measured modern storm beach ridges (arrows) and <u>Pelvetia canaliculatus</u> (sp) fragments (dots).

Table 4

<u>Pelvetia</u> <u>canaliculatus</u> (sp) fragment altitudes

	Max.- Min. (m)	Mean (m)	Grid Reference	Range (m)	Distance to L.W.M.O.S.T. (m)
1	2.79-2.70	2.75	NR 50888024	0.09	7
2	2.70	2.70	NR 50948021	-	7
3	2.24-2.15	2.20	NR 50488022	0.09	20
4	2.22-1.95	2.06	NR 51398018	0.27	26
5	2.28-1.83	2.05	NR 51318022	0.45	50
6	2.38-2.08	2.18	NR 50448045	0.30	45
7	3.07-2.50	2.71	NR 46127690	0.57	8
8	2.46	2.46	NR 45647607	-	23
9	1.97	1.97	NR 45507582	-	40
10	2.47	2.47	NR 45517588	-	30
11	2.63-2.10	2.40	NR 45447567	0.53	21
12	2.03	2.03	NR 45427560	-	35
13	2.68-2.60	2.64	NR 45307544	0.08	14
14	2.87-2.75	2.81	NR 52388034	0.12	10
15	2.05-1.99	2.02	NR 52608042	0.06	45
16	1.99	1.99	NR 54418071	-	50
17	2.42	2.42	NR 53728053	-	11
18	2.96-2.90	2.93	NR 51078505	0.06	15
19	2.57-2.31	2.43	NR 51748120	0.26	50
20	2.17-2.15	2.16	NR 51988137	0.02	26
21	2.51-1.99	2.39	NR 52128149	0.52	11
22	2.85-2.79	2.82	NR 52218180	0.06	12
23	2.81-2.26	2.58	NR 53148842	0.55	10
24	2.39-2.25	2.32	NR 53148870	0.14	15
25	2.16-2.08	2.12	NR 53458900	0.08	55
26	2.11	2.11	NR 53758945	-	37
27	2.08-2.02	2.05	NR 54008964	0.06	40
28	2.61-2.60	2.61	NR 54829015	0.01	25
29	2.61	2.61	NR 54729010	-	32
30	2.86-2.78	2.82	NR 55199040	0.08	15
31	2.65	2.65	NR 56059081	-	30
32	2.26-2.17	2.22	NR 56199089	0.09	45
33	2.91-2.64	2.74	NR 56839157	0.27	25
34	2.18-2.01	2.10	NR 57359246	0.17	40
35	2.72	2.72	NR 57449250	-	14

Table 4 (contd.)

Pelvetia canaliculatus (sp.) fragment altitudes

	Max.- Min. (m)	Mean (m)	Grid Reference	Range (m)	Distance to L.W.M.O.S.T. (m)
36	2.70-2.48	2.59	NR 57379247	0.22	20
37	2.41-2.07	2.17	NR 58619305	0.34	55
38	2.95-2.85	2.90	NR 60139415	0.10	10
39	2.55-2.38	2.47	NR 51878247	0.17	37
40	2.52	2.52	NR 53088210	-	45
41	2.86-2.77	2.82	NR 52248218	0.09	17
42	2.51-2.48	2.50	NR 52348196	0.03	30
43	2.45-2.35	2.40	NR 53148202	0.10	22
44	2.25	2.25	NR 53418174	-	28
45	2.40-2.32	2.36	NR 50488335	0.08	55
46	2.18	2.18	NR 66339853	-	45
47	2.35-2.17	2.26	NM 71360350	0.18	23
48	2.30	2.30	NM 71290690	-	38
49	2.12-1.97	2.05	NR 63409648	0.15	55
50	3.13-2.99	3.07	NR 64799721	0.14	5
51	3.13-3.01	3.07	NM 67860047	0.12	8

n= 51; mean= 2.43 m; range= 1.10 m; maximum = 3.07m; minimum= 1.97 m; standard deviation= 0.30 m.

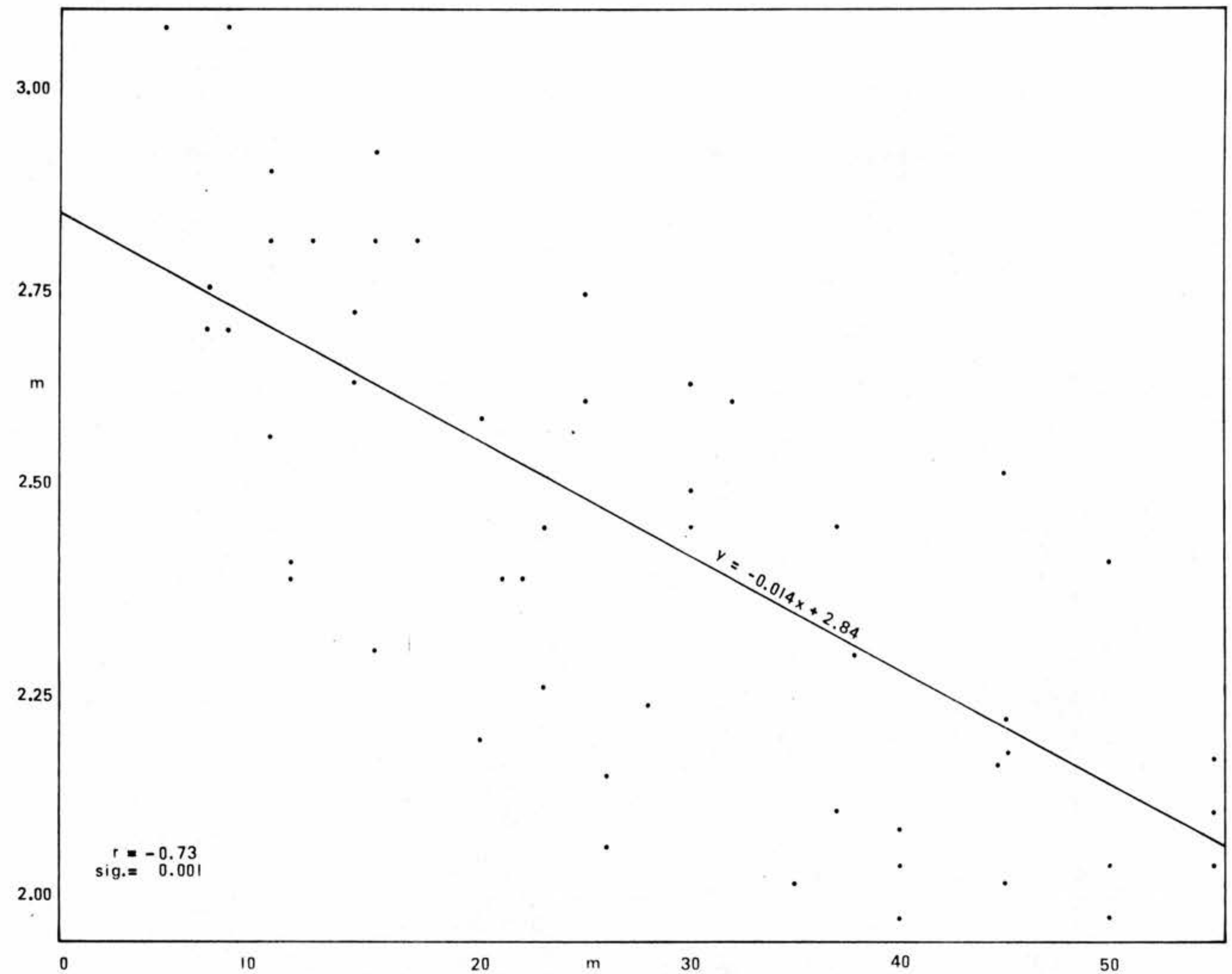


Fig. 21 Linear regression of Pelvetia fragment altitudes and intertidal width.

Considerable care however must be taken when interpreting regional variations in _Pelvetia_ altitude. For example the rough rock and shingle surfaces on which _Pelvetia_ often develops inhibit the movement of seaweed after its initial deposition. As a result, the measured values may deviate from the level of high water mark of ordinary spring tides by several centimetres. Additionally, considerable variation in _Pelvetia_ altitude may occur over short distances. For example the _Pelvetia_ line at the bayhead shingle beach Lang Aoinidh (NR 461769) varies by 0.57 m along its length and demonstrates that differential wave activity in almost calm sea conditions is in itself sufficient to induce local variations in the highest level of wave action.

Finally it is unclear how regional variations in the elevation of high water mark of ordinary spring tides affects the distribution of _Pelvetia_ altitudes. As stated in Chapter 2, mean high water spring tide at Glengarrisdale, N Jura is 0.4 m higher than at Rubha a Mhail, NE Islay and Scalsaig, E Colonsay (Table 1). Although it is not possible to interpolate spring tide values accurately between coastal locations it is most likely that there is a 0.4 m variation in the altitude of mean high water spring tide along the W Jura and NE Islay coastline. _Pelvetia_ altitudes are therefore affected by regional variations in the altitude of high water spring tides.

From the above it is clear that under calm sea conditions the high water mark altitude, defined by _Pelvetia_, is closely related to the width of the intertidal zone. Since the average altitude of _Pelvetia_ corresponds approximately with the heights obtained for the lower limit of land-based vegetation, the _Pelvetia_ altitudes form useful indicators of the upper limit of low/moderate wave activity. Most measured

Pelvetia altitudes occur at 2.43 ± 0.30 m and, if present coastal conditions are similar to those prevailing in the past, it is likely that the inner edges of raised coastal terrace fragments bear similar altitude relationships to former sea-levels.

High energy coastal landforms beach:ridges

1. Introduction

In W Jura, Scarba and NE Islay the occurrence of modern beach ridges is largely controlled by the distribution of pre-existing shore platforms. Beach ridges are the most common depositional coastal landform and are located upon low shore platforms, between low rock skerries, at geo entrances and at the foot of cliffs. Due to the abundance of foreshore rock skerries most ridges are located in coastal indentations between rock outcrops. The material of which the ridges are composed varies from small cobbles 3-10 cm in diameter to boulders that rarely exceed 50 cm in diameter. In most instances the seaward faces of the ridges are composed of loosely packed cobbles that are replaced downslope by narrow sand beaches. Farther seaward shingle accumulations normally occur near low water mark and are rarely more than 15 m in width.

At numerous localities backshore beach ridges form the seaward margins of staircases of raised ridge and swale topography. The landward limit of each modern ridge is normally defined by a change from smooth unvegetated cobbles to lichen-veneered cobbles whose surfaces are generally rougher than those in the foreshore and backshore. The change from modern to raised beach cobbles is also frequently marked by a mantle of driftwood (Plate 2). As a result the maximum level of modern wave activity is usually clearly defined.

In the following paragraphs an attempt is made to distinguish genetically different types of beach ridge and also to determine the morphological criteria whereby beach ridges of different origins may be distinguished. Owing to the complexities of wave dynamics in the nearshore zone, only a brief summary of process-landform relationships is given.

2. The origin of beach ridges: review of previous literature

Of the many coastal landforms beach ridges are perhaps the least understood. Explanations of their origin and development are ambiguous. King for example (1972, p.423) states,

> "It is storm waves-which are generally destructive- that build the ridges to a height well above normal high water level."

The apparent contradiction in destructive wave action being stated to be responsible for the construction of beach ridges has never been adequately explained. In contrast to the statement that beach ridges are formed during storms, Johnson (1919,pp.412-3) argued that,

> "...... a large part, if not all, of a beach ridge is often swept away during a single exceptional storm..... and although high beach ridges must have been subjected to the influence of waves of sufficient magnitude to cast debris to their crests, the majority of ridges could have reached their present height through the influence of ordinary storm waves, and many of them perhaps by very moderate wave action at high tide. It is even possible that on a given beach plain none of the exceptional storms of the past are recorded by any of the ridge crests but only by the more prolonged activities of less violent wave action."

The origin of beach ridges was early discussed by Gilbert (1883) and Cornish (1898). Cornish stressed the role of percolation in ridge development and noted that ridge growth was contingent on an adequate supply of offshore debris moved shoreward by high wavelength constructive swell waves. He considered (1898, pp.435 and 535) that during wave breaking, water is moved landward by two related mechanisms.

Firstly, a part of the descending water moves rapidly shoreward to meet the returning backwash of the previous wave. Secondly, the area of most intense wave motion is behind the falling wave front where water is swirled upward and forward over the surface of the falling breaker front. Both wave motions are constructive and drive material shoreward. Due to the percolation of water through the shingle the effect of uprush compared to backwash is reduced. During severe storms the backwash velocity more closely approaches the uprush velocity and as a result shingle erosion occurs, the material being transported seaward.

Johnson (1919) agreed with the conclusions of Cornish and added that, owing to diminishing seaward water velocities, eroded shingle could be transported only a short distance offshore. Consequently, following a storm, the resultant abundance of shingle a short distance offshore resulted in its subsequent landward movement by constructive swell waves. As a result low amplitude beach ridges were formed in the intertidal zone.

High amplitude beach ridges were considered by Lewis (1936) to be the result of shoreward deflection of storm waves so that,

> ".... the waves pile up the shingle to greater and greater heights, limited only by the supply of material and the size of the waves. Eventually the ridge attains such proportions that even the most powerful waves in conjunction with the highest spring tides can barely overtop the bank." (Lewis, 1936, p.439).

In total contrast Zenkovich (1967, p.274), in discussing storm activity concluded that,

> "At the height of the storm the waves, which have eroded the whole beach and have cut away the beach ridges, rush up to the very top of the beach. When the storm begins to die down .... the weakening waves begin to throw up the largest shingle and to pile it up into the first beach ridge."

In this context Vladimirov (1953) observed that in a high sea the shingle of a sand and shingle beach was concentrated in two distinct ridges. One ridge occurred at the upper limit of wave uprush while the other was located farther seaward at the line along which the waves broke (cf. Zenkovich, 1967, p.319). Referring to high beach ridges Zenkovich (1967, pp.292-3) stated that, as a result of the unequal strength of storm waves, individual storm ridges may vary in height not only along their lengths but also from place to place along the coast and may often possess local variations in their profiles.

The above writers noted several limiting factors controlling the vertical growth of high beach ridges. Firstly, it was considered that wave refraction in bays in conjunction with gently sloping offshore profiles results in waves breaking far offshore during storms and hence only constructive wave action is possible landward of the zone of wave breaking. Secondly, since larger wavelength storm waves break farther offshore they have less constructive wave energy to drive shingle landward and hence in bay areas of gentle offshore gradient the incoming waves are frequently incapable of constructing large ridges.

Therefore a prerequisite for the growth of a high beach ridge is that the zone of wave breaking is close to the pre-existing ridge face since only a breaking wave is capable of propelling shingle onto and over the ridge crest. The uprush associated with a wave that has broken, if of sufficient velocity, may induce tumbling and sliding of cobbles (Longinov and Paseschnik, 1953) but cannot propel them onto the ridge crest if the zone of wave breaking occurs a considerable distance offshore. When waves break close to the shore during a storm, propulsion of cobbles over a ridge crest is accompanied by the entrainment of cobbles offshore by destructive wave action. It has been suggested (Iwagaki and Noda, 1963; Komar, 1976) that wave steepness determines whether material is driven shoreward (by constructive waves) or seaward (by predominantly destructive waves).

It is therefore clear that during storms large quantities of shingle are entrained seaward by destructive waves. Indeed during severe storms entire shingle beaches are often destroyed and the constituent debris removed to the offshore zone (Johnson, 1919). The large volumes of debris accumulated in the offshore zone are later subject to constructive wave action. As a result low amplitude beach ridges may form in the intertidal zone. Ridges formed in this manner reflect moderate wave activity and differ from the high amplitude ridges that form during storms.

Two types of beach ridge have therefore been identified in previous investigations. The more commonly observed type is the high storm beach ridge that occurs in the backshore zone and is characterised by large altitude variations. The intertidal beach ridge possesses a much lower amplitude than the storm ridge and forms under more moderate wave energy conditions. The material of the latter ridge type is usually derived from the erosion of high beach ridges during storms.

In the study area crest altitudes of 48 storm ridges were measured by levelling. In the following sections an attempt is made to explain the regional variation of storm ridge crest altitudes by investigating the factors responsible for their formation. In all calculations the mean altitude of each storm ridge crest is used (Table 5). Low amplitude intertidal shingle accumulations also occur in the study area at or near low water mark of ordinary spring tides: these were not levelled.

3. Storm ridge crest altitudes

Although ridge crest altitude variations in different localities are not strictly comparable owing to different local marine conditions, known heights provide a useful indication of the height variations between storm ridges.

At Portland, King (1972, pp. 308 and 422) noted that the Chesil beach ridge was composed of shingle 5.0-7.3 cm in diameter with a crest altitude of 13.1 m O.D. In N Norway, Sollid et al. (1973, p.238) measured the altitudes of 94 storm ridge crests. The crests occurred between 7.1 and 1.4 m above high water mark and had a mean elevation of 3.9 m. They noted that in exposed areas the mean elevation of the 12 highest ridges was 5.6 m above high water mark. Zenkovich (1967, p.292) has observed storm ridges with crest altitudes as high as 6.7 m above mean sea level at Kobuleti on the Black Sea coast.

At Aird, Benbecula, Ritchie (1972, p.32) noted a modern storm ridge up to 15 m above sea level that was composed of cobbles up to 0.5 m in diameter. Also in the Outer Hebrides, Von Weymarn (1974, Table IV) recorded that the highest active storm ridge occurred at Barvas, NW Lewis and had a crest altitude of 7.4 m O.D. In Iona at Port an Fhir-bhreige (Bailey et al., 1925, p.93) the highest modern shingle ridge crest altitude reaches 10-12 feet (3.0-3.7 m) above sea level.

The altitudes of 48 storm beach ridge crests measured in W Jura,

Table 5

Modern shingle ridge crest altitudes and wave energy potential values

	Max.-Min.(m)	Angle of open fetch (degrees)	Mean (m)	Grid Reference	Range (m)	Distance to L.W.M.O.S.T.(m)	Wave energy potential (Wp)
1	3.82-3.32	11	3.54	NR 51398018	0.14	30	75.73
2	3.26	2	3.26	NR 51318022	-	25	51.75
3	3.38-2.61	1	2.88	NR 51298038	0.77	32	17.63
4	4.66-4.10	27	4.36	NR 50428056	0.56	15	64.09
5	3.46-3.35	21	3.41	NR 49658007	0.11	10	62.95
6	3.77-3.65	8	3.70	NR 46127690	0.12	15	56.74
7	4.89-4.06	26	4.66	NR 47887939	0.83	30	57.38
8	3.77-3.30	2	3.63	NR 45477587	0.37	63	59.36
9	2.95-2.68	1	2.85	NR 45457573	0.27	55	65.52
10	4.36-4.24	17	4.30	NR 52358034	0.12	35	66.56
11	4.91-4.70	8	4.81	NR 53958058	0.21	45	60.06
12	5.22-5.06	33	5.14	NR 51108505	0.16	53	114.02
13	7.31	29	7.31	NR 51358570	-	42	151.72
14	3.54	16	3.54	NR 52268692	-	17	75.73
15	4.52-4.40	25	4.46	NR 53008804	0.12	30	54.93
16	3.78	1	3.78	NR 53118832	-	20	114.36
17	3.91	23	3.91	NR 53668915	-	23	75.73
18	4.06-3.56	24	3.81	NR 53898958	0.50	30	64.09
19	6.71-6.32	20	6.53	NR 55179045	0.39	44	141.30
20	4.17-4.07	14	4.12	NR 58659314	0.10	50	119.32
21	4.21	27	4.21	NR 52228250	-	30	107.06
22	3.04	24	3.04	NR 52308235	-	39	145.63
23	3.91-3.51	25	3.71	NR 52448193	0.40	28	80.04
24	4.79	19	4.79	NM 71150685	-	13	37.40
25	4.77	1	4.77	NM 71430349	-	8	45.55
26	5.17-4.90	26	5.03	NR 61379486	0.27	14	122.95
27	6.55	26	6.55	NR 60449436	-	9	124.37

Table 5 (contd.)

Modern shingle ridge crest altitudes and wave energy potential values

	Max.- Min.(m)	Angle of open fetch (degrees)	Mean (m)	Grid Reference	Range (m)	Distance to L.W.M.O.S.T.(m)	Wave energy potential (Wp)
28	4.28-3.11	3	3.66	NR 50438042	1.17	35	58.88
29	4.49	7	4.49	NR 61669502	-	16	107.48
30	4.76	4	4.76	NR 62299500	-	21	97.61
31	6.07-5.87	8	5.97	NR 62539601	0.20	7	212.12
32	4.65	7	4.65	NR 62849632	-	15	119.31
33	3.26	3	3.26	NR 64839712	-	8	107.04
34	3.74-3.68	1	3.71	NR 66419849	0.06	37	200.73
35	4.29-4.15	3	4.22	NR 66749869	0.14	15	137.16
36	5.51-4.92	28	5.24	NR 40387997	0.59	11	93.02
37	4.54-4.23	29	4.39	NR 49528007	0.31	40	42.34
38	5.64-5.25	30	5.32	NR 48007945	0.39	33	88.86
39	4.32-4.18	1	4.26	NR 50488022	0.14	15	106.83
40	4.74-4.33	30	4.54	NR 48557952	0.41	14	54.93
41	5.53-5.36	28	5.44	NR 51228286	0.17	11	231.36
42	5.03-4.93	23	4.98	NP 62319500	0.10	19	182.48
43	5.64-5.42	30	5.32	NR 46887826	0.22	25	289.82
44	7.92-7.42	30	7.67	NR 52218656	0.50	18	205.85
45	5.32	38	5.32	NR 38747808	-	115	246.74
46	5.00	46	5.00	NR 38887808	-	140	248.45
47	5.95-5.86	33	5.91	NR 39327811	0.09	120	229.75
48	7.30-7.16	40	7.23	NR 39477814	0.14	105	242.02

N=48; mean= 4.61 m; range= 4.82 m; maximum= 7.67 m; minimum= 2.85 m; standard deviation= 1.14 m.

Scarba and NE Islay are shown in Table 5. Most ridge crest altitudes occur between 4.3 and 4.9 m while the standard deviation of the sample is 1.14 m. The highest modern ridge is located south of Shian Bay where the crest altitude averages 7.7 m. In only four instances does the range of crest altitudes measured along an individual ridge crest exceed 0.5 m.

Analysis of factors affecting the altitude of storm ridge crests

1. Recorded storm wave activity

Evidence that recent wave activity in Jura, Scarba and Islay has extended only to the crests of many modern beach ridges is given by the presence of driftwood on their upper surfaces. For example on the crest of a modern beach ridge S of Shian Bay, W Jura (NR 52218656) a telephone pole approximately 4 m in length with a 0.4 m base occurs at a height of 8 m. Since W Jura has been uninhabited since approximately 1939 there is little doubt that the pole was deposited by storm waves. On the other hand in NE Islay 10-15 stone cairns, each possessing central circular depressions, are located in three raised coves (NR 38777810; NR 38887806; NR 39147805). The cairns are generally 1.5-2.0 m in diameter and are composed of lichen-veneered raised beach cobbles. The average base of the cairns is 6.1 m and all are unmodified by modern wave action. The age of the cairns is at present uncertain: they are either pre-historic in age or are 19th century kelp kilns. In either case wave action has not reached 6.1 m since their formation or has done so on such rare occasions that the lichens have not been destroyed.

The absence of staircases of modern beach ridge and swale topography in the study area (each modern ridge occurring singly) also suggests that, in the recent past, storm beach ridges have been continually destroyed and re-created by wave action.

2. Fetch

Fetch is normally defined as the distance travelled by waves from their area of generation to the coast (Zenkovich, 1967, p.27). During cyclonic storms the height and wavelength of waves generated in the open sea are largely determined by the velocity and duration of the prevailing winds (Komar, 1976). In turn oceanic wave height and length increase to maximum values if the length of fetch is sufficient. Since the depth of wave activity is proportional to wavelength, high wavelength swell waves that approach the coast are most capable of inducing shingle motion on the seabed. In contrast, waves generated by storms in sea areas constricted by adjacent land masses are only capable of generating waves of shorter length. In these areas movement of shingle on the seabed is restricted to shallower water depths.

Both types of marine environment occur in the study area. The distribution of islands in the Sea of the Hebrides results in the exposure of certain parts of the W Jura, Scarba and NE Islay coast to open Atlantic fetch from the SW and NW (Fig. 8) while in other areas the length of fetch is restricted to several kilometres by the presence of the island barriers of Islay, Colonsay, Mull and Tiree.

In an attempt to assess to what degree variations in exposure to Atlantic fetch explain variations in ridge crest altitudes, a simple method was employed to derive an index of exposure to open fetch for each location where the altitude of storm ridge crests had been measured. At each site the horizontal angle was measured between the coastal site and the land barriers of N Islay (in some cases N Ireland) and Colonsay that form the N and S margins of the open SW Atlantic fetch sector. A similar angle of exposure to open NW Atlantic fetch was determined by measuring the angle between each coastal site, Colonsay and Tiree. Thereafter the sum of the two angles of exposure to open fetch was added to give a total angle of exposure to open fetch for each coastal location. In order to facilitate calculations, beach ridge locations that were not subject to open Atlantic fetch were assigned a fetch index of unity. The relationship between each of the 48 crest altitudes and the corresponding total angles of open fetch was analysed by linear regression (Fig. 22). Although no allowance is made for offshore gradients, sediment supply and wind factors, the relationship is strong ($r= 0.55$; Signif. = 0.003).

3. Offshore gradients and seabed topography

As deep water waves approach the coast, movement of fine sediment on the seabed commences when water depth approaches half the wavelength. Landward of this area the incoming waves are distorted and are characterised by an increase in wave height and a decrease in wavelength. When the ratio between wave height and length (wave steepness) becomes sufficiently large, wave breaking occurs.

Although observations of shingle movement at depth in different wave environments are not strictly comparable, it is generally agreed that during storms shingle may be entrained to a depth of 20 m (Zenkovich, 1967, pp.162-3). During low/moderate wave activity shingle is only moved beneath and landward of the zone of wave breaking.

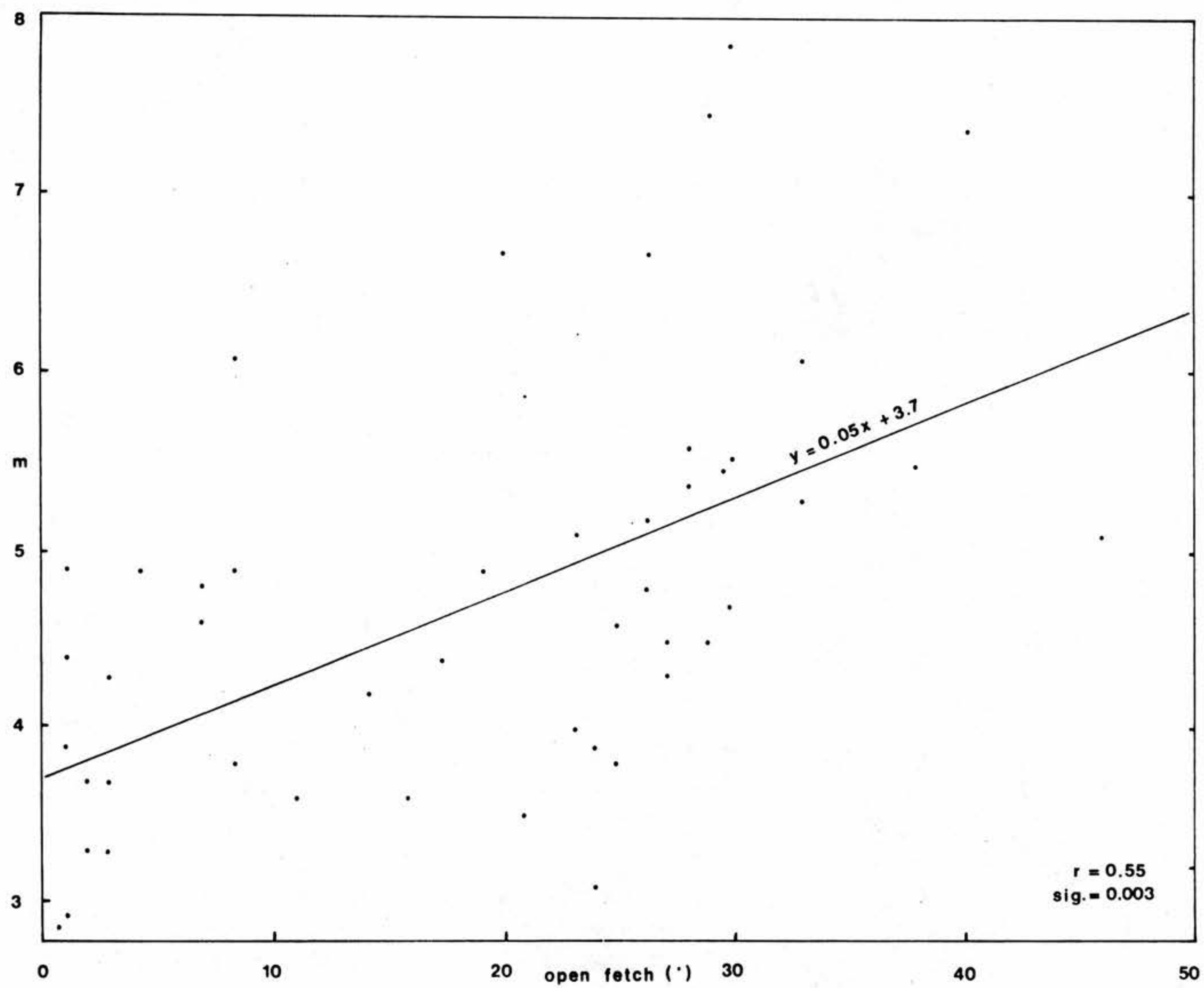


Fig. 22 Linear regression of modern storm ridge crest altitudes and total angle of open Atlantic fetch (°).

Considerable areas of the seabed surface off W Jura and NE Islay are shallower than 20 m (Figs. 23 and 24). Tarbert Bank in particular shoals to a depth of 10 m and extends as a broad submarine plateau to the SW for a distance of 10 km (Fig. 24). Shingle occurring in this area is thus available for shoreward movement during high wavelength swell conditions. In contrast, the seabed north of Tarbert Bank is characterised by deep troughs that descend in places to more than 200 m below sea level. Here, in the nearshore zone, the seabed plunges steeply, thus causing waves to break close inshore. Since wave activity occurs over a narrow area, the energy available to form storm ridges is proportionally greater.

Two distinct offshore areas that influence the vertical growth of storm ridges can thus be defined. Firstly in the southern area the shallow seabed surface induces distortion of high wavelength swell waves at great distances from the coast, while in parts of this area the water is sufficiently shallow to result in an adequate offshore supply of shingle. Secondly in the northern area storm waves are not distorted by shoaling until close to the coastline. Since shingle occurring in this area is located in deep, offshore troughs, little is available for shoreward movement by waves. In the nearshore zone therefore the resultant high energy wave activity is most favourable for the vertical growth of storm ridges although such growth is limited by the restricted supply of offshore debris.

In order to test if there is any relationship between offshore gradient and beach ridge altitude, the offshore gradients were measured perpendicular to the shore at 21 locations (shown on Fig. 24) where beach ridge altitudes were known. The measurement of the seabed depth

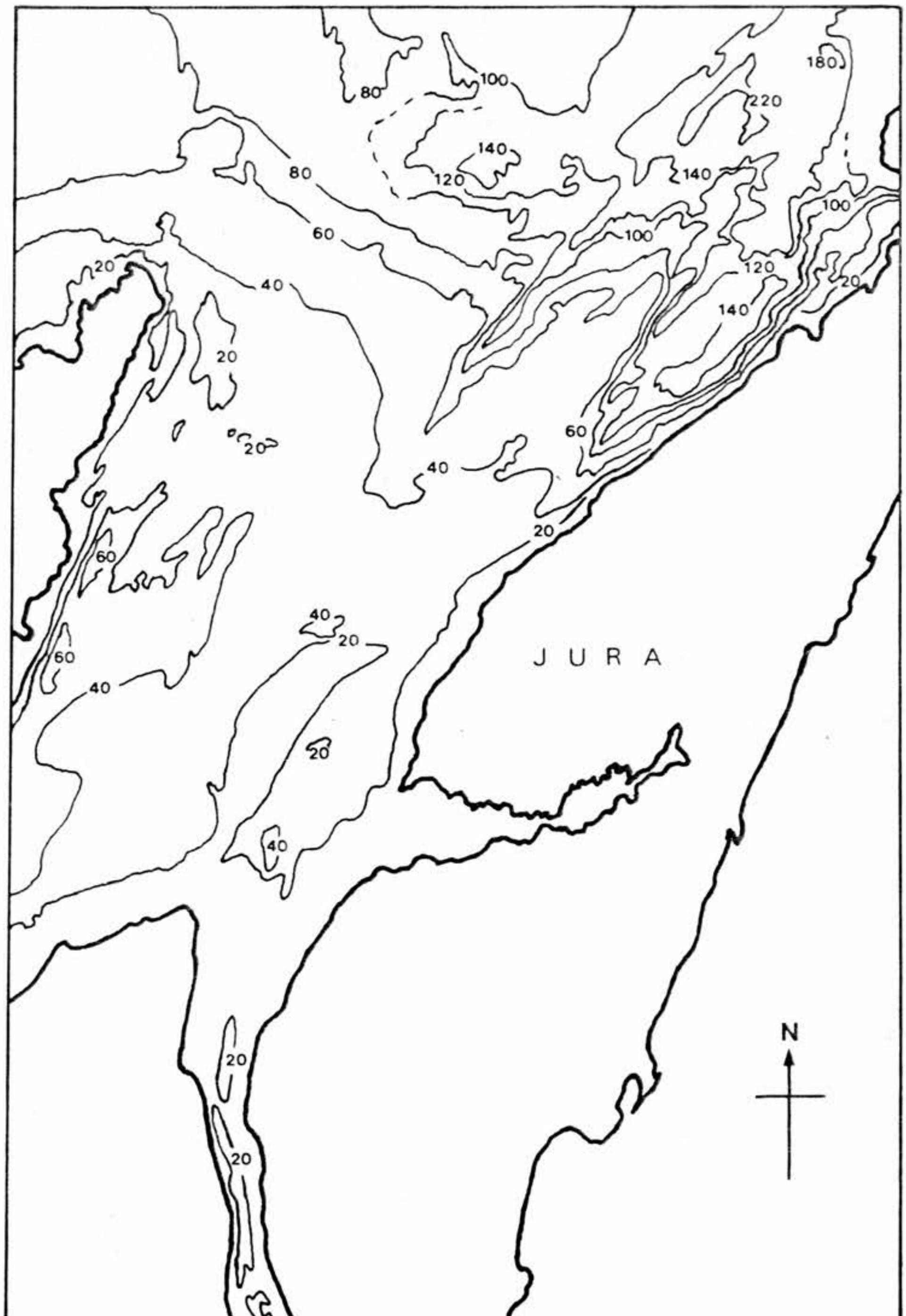


Fig. 23 Seabed between Scarba, Jura, NE Islay and Colonsay. Depths in metres. Information courtesy of the Institute of Geological Sciences, Edinburgh.

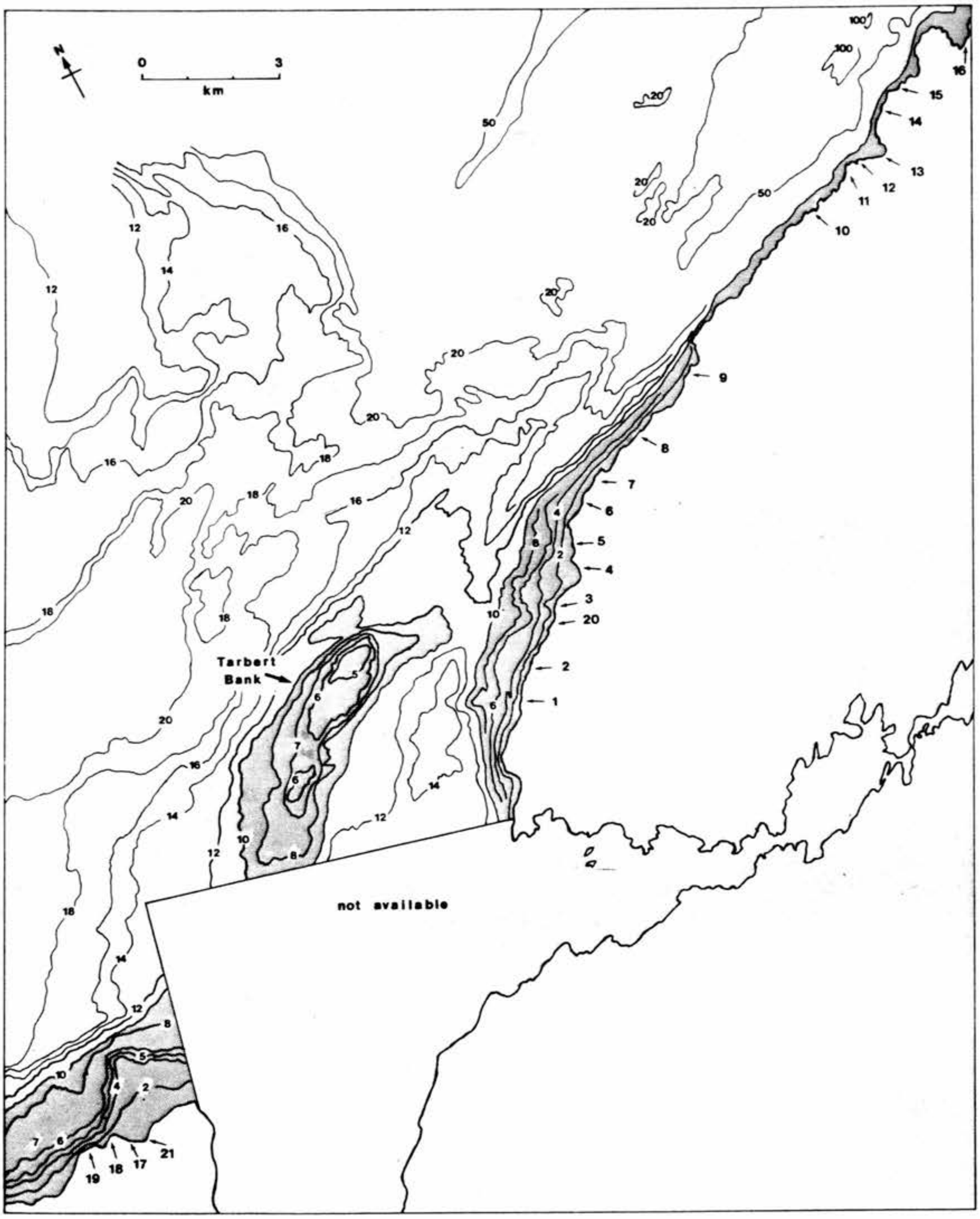
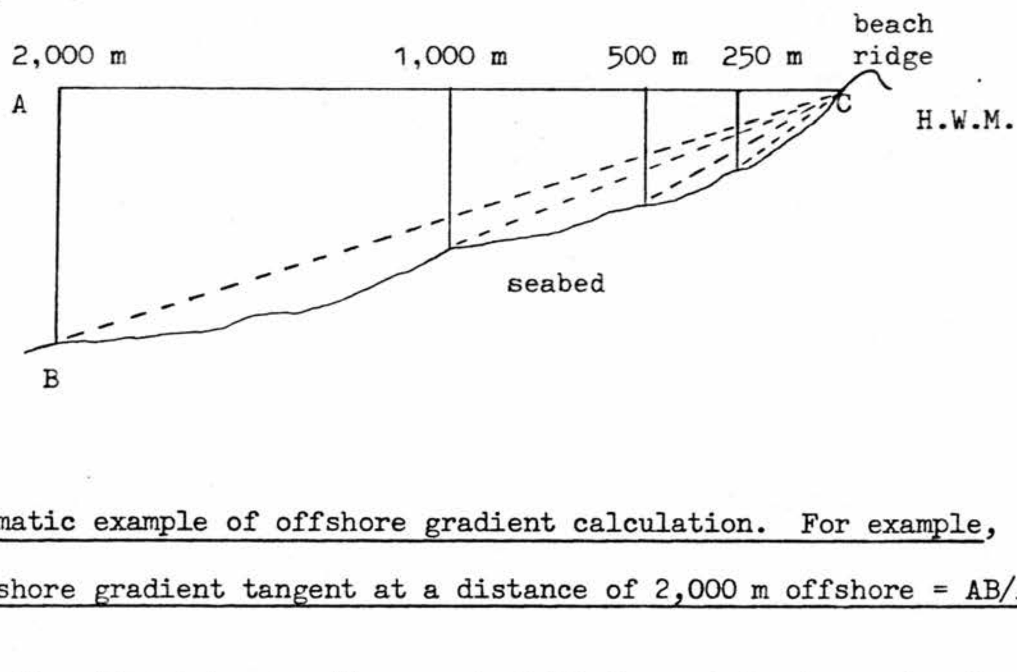


Fig. 24 Seabed off W Jura. Depths in fathoms. Sea areas shallower than 10 fathoms are shaded. Sites of modern storm beach ridges where offshore gradients were determined are also indicated. Information based on hydrographic data courtesy of Institute of Geological Sciences, Edinburgh.

at each location was based on contours drawn on Admiralty hydrographic maps of the Sea of the Hebrides. Since each depth value shown on the original charts is accurate to only one fathom and since the available maps are drawn at scales of 1:36,000 and 1:72,000 the depth contours can only be approximate. No Admiralty chart was available for the sea area comprising Loch Tarbert and the Sound of Islay (see Figs. 23 and 24) and hence no offshore depth information is available for beach ridges occurring in this area. Seabed depths were measured perpendicular to the shore at distances of 250, 1,000 and 2,000 m offshore from each beach ridge location. At each site the tangent of the angle between the offshore slope and the horizontal was calculated, the apex being taken as the position of high water mark adjacent to each beach ridge (Table 6 and Fig. 25).

Fig. 25



Diagrammatic example of offshore gradient calculation. For example, the offshore gradient tangent at a distance of 2,000 m offshore = AB/AC.

The relationships between ridge crest altitudes and the tangents of offshore gradients at distances 250, 500, 1,000 and 2,000 m offshore were analysed by linear regression and the results are shown in Table 7.

Table 6

Offshore gradients at beach ridge locations

Distance offshore (m)	Gradient 250	500	1,000	2,000	Ridge crest altitude (m)	Angle of open fetch (degrees)
1.	0.020	0.017	0.015	0.011	5.14	33
2.	0.022	0.014	0.018	0.008	7.31	29
3.	0.020	0.015	0.012	0.017	3.54	16
4.	0.011	0.010	0.012	0.013	4.46	25
5.	0.012	0.016	0.016	0.019	3.78	1
6.	0.013	0.017	0.029	0.012	3.91	23
7.	0.021	0.030	0.035	0.013	3.81	24
8.	0.032	0.042	0.033	0.015	6.53	20
9.	0.042	0.049	0.055	0.018	4.12	14
10.	0.124	0.089	0.077	0.061	5.03	26
11.	0.108	0.091	0.097	0.062	6.55	26
12.	0.118	0.182	0.108	0.098	4.49	7
13.	0.022	0.094	0.100	0.041	4.76	4
14.	0.215	0.148	0.146	0.055	5.97	8
15.	0.139	0.120	0.140	0.081	4.65	7
16.	0.022	0.042	0.091	0.071	3.26	3
17.	0.005	0.008	0.013	0.012	5.91	33
18.	0.013	0.010	0.013	0.013	5.00	46
19.	0.013	0.009	0.011	0.013	5.32	38
20.	0.017	0.017	0.016	0.010	7.67	30
21.	0.011	0.010	0.013	0.011	7.23	40

* beach ridge locations shown on Fig. 24

Table 7

Regressed variables	Correlation Coefficient.	Significance level.
1. Offshore gradient tangent at 250 m offshore vs. ridge crest altitude	0.079	0.633
2. Offshore gradient tangent at 500m offshore vs. ridge crest altitude	0.031	0.554
3. Offshore gradient tangent at 1,000m offshore vs. ridge crest altitude	0.122	0.701
4. Offshore gradient tangent at 2,000m offshore vs. ridge crest altitude	0.179	0.781

The results indicate that there is no apparent relationship between offshore gradients and beach ridge crest altitudes. The reason for such a poor relationship is not clear. Since a strong correlation exists between crest altitude and exposure to open fetch (Fig.22), it is suggested that during storms, wave steepness is a critical factor controlling ridge altitude (Iwagaki and Noda, 1963) since waves of widely differing length and height can possess the same steepness (ie. wave height/length).

Since it has been demonstrated that regional variations in the altitude of Pelvetia canaliculatus (sp.) are closely related to local variations in the width of the intertidal zone it was decided to investigate whether regional variations in ridge crest altitudes were similarly related to variations in intertidal width. At each of the 48 beach ridge locations, the horizontal distance from high to low water mark was measured from Ordnance Survey 1:10,560 maps. The calculated intertidal widths and beach ridge crest altitudes were then analysed by linear regression. The analysis revealed a poor relationship (Fig. 26) (r= 0.21; Signif. ÷ 0.075).

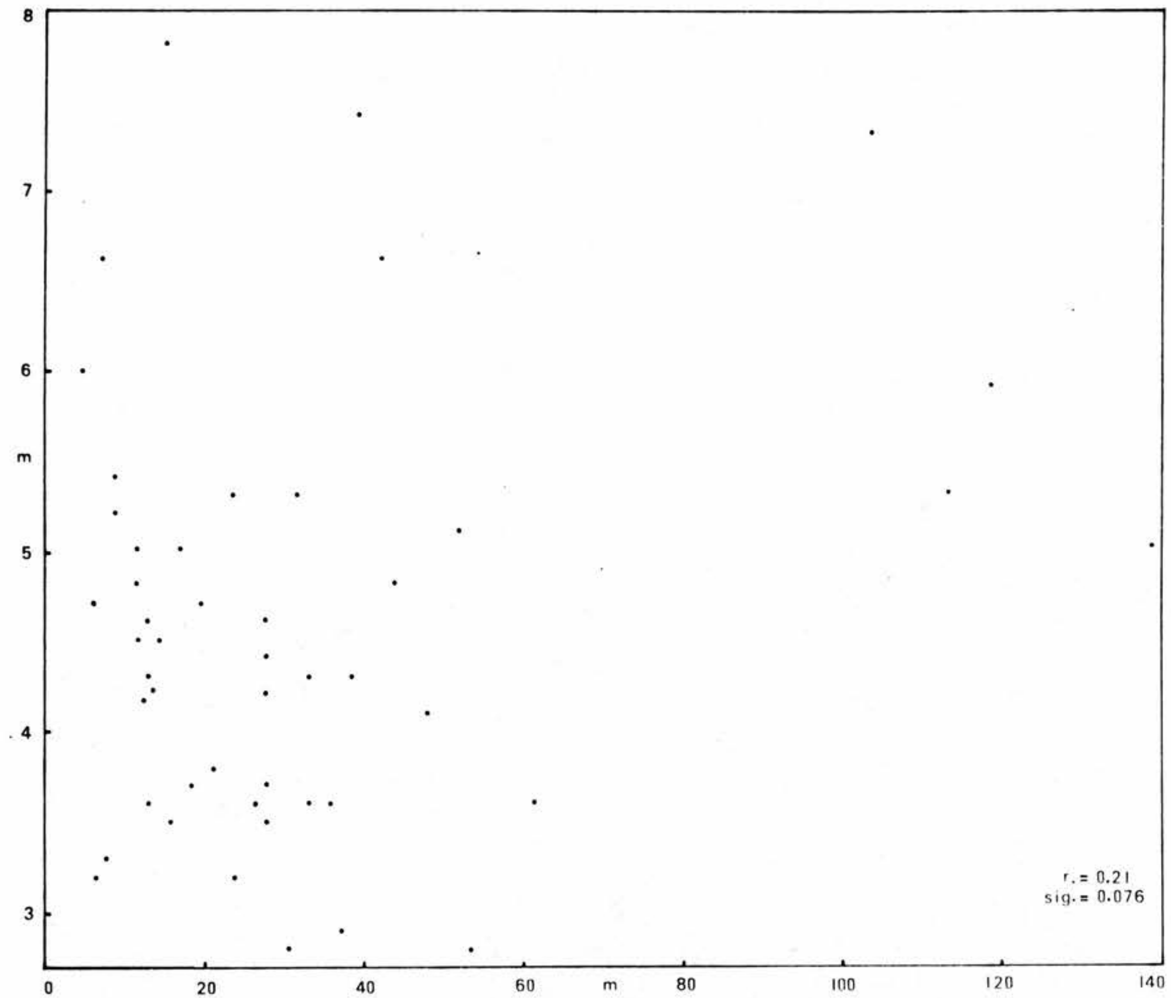


Fig. 26 Relationship between modern storm ridge crest altitudes and intertidal width.

This lack of relationship may be simply explained. During the formation of storm ridges the zone of wave breaking must lie at or near the foot of the pre-existing ridge face since cobbles cannot be propelled onto and over the ridge crest when waves have broken. During storms therefore, wave breaking occurs only after the incoming waves have travelled across the entire intertidal surface. As a result the zone of wave breaking is largely unaffected by the width of the intertidal zone.

From the above it is clear that intertidal width, although an important factor affecting the altitude of *Pelvetia* deposition, does not explain the regional distribution of storm ridge crest altitudes. It is also apparent that the depth of the seabed at distances as great as 2 km offshore is of negligible importance in affecting regional altitude variations of ridge crests.

4. Sediment supply

Ritchie and Crofts (1974, p.12) concluded that the beaches of Islay and Jura are almost all accreting. Since relatively small quantities of debris are transported to the coastal zone by rivers, it is likely that the supply of material to the coast from offshore in conjunction with coastal erosion of raised beach deposits are important factors affecting beach accretion. Shoreward rafting of sediment, small cobbles and coral by seaweed has been observed but is believed of minor importance in the overall movement of debris to the shore zone.

Known thicknesses of offshore sediment above rockhead are indicated on Fig. 27. Sediment thickness of over 100 m are common, but as

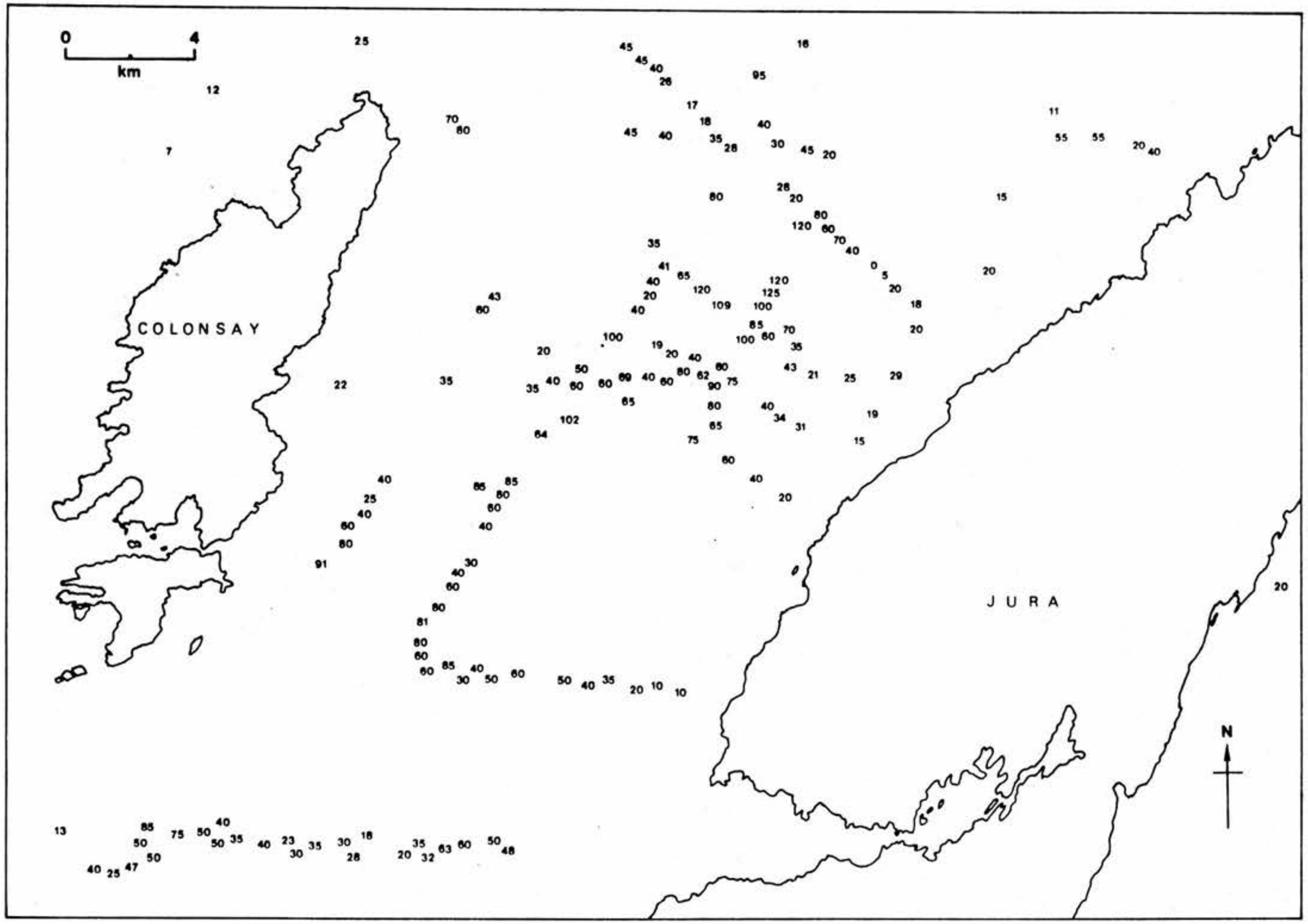


Fig. 27 Thickness of offshore sediment (m) above rockhead. Data courtesy of Institute of Geological Sciences, Edinburgh.

stated earlier, the surface parts of this material are not available for shoreward transport since they are located in deep water. In other localities, however, large areas of sediment occur within 20 m of the sea surface and are available for shoreward movement during constructive swell wave activity. In addition, since relative sea-level in W Scotland has fallen since the main postglacial transgression (Gray, 1972) the shallower water depths favour the movement of seabed debris to the coast. An equally important factor affecting the supply of beach shingle is the cycle of shingle transport and deposition that occurs in the nearshore zone.

As already stated, during storms shingle is both added to and removed from pre-existing ridges. Shingle that is transported seaward is deposited a short distance from the shore due to decreasing seaward water velocities. After a storm constructive wave action returns shingle to the shore and forms low beach ridges. The volume of shingle in the nearshore zone is therefore a critical factor determining the size of backshore storm ridges but is not easily measured.

The relationship between storm ridge altitudes and nearshore shingle supply is particularly relevant to coastal areas where modern storm ridges are succeeded landward by raised shingle deposits. In these areas, the supply of shingle for seaward transport during storms is limited only by the intensity and height of storm wave activity. Indeed the five highest storm ridges in the study area (Table 5) are succeeded landward by raised ridge and swale topography. However since other lower storm ridges bear a similar relationship to raised marine deposits it is clear that the elevation of the highest ridges cannot be attributed solely to this cause. For example, the supply of shingle to the high beach ridges that occur south of Shian Bay may be explained by a rich offshore sediment source (Tarbert Bank) in conjunction with the widespread distribution of raised beach deposits that flank the modern storm ridges.

There may be a relationship between offshore shingle supply and ridge altitude but it is impossible to define quantitatively since the proportion of shingle incorporated within the seabed sediments is unknown. Similarly it is difficult to estimate the rate and volume of shingle removal from raised beach ridges to the shore zone.

5. Wind direction, duration and velocity.

The dominant winds that occur in W Jura, NE Islay and Scarba originate from mid-Atlantic cyclonic storms and blow from the SW and S (Fig.9) If the storms responsible for mid-Atlantic wave generation are of sufficient intensity and duration, high wavelength swell waves travel towards the Sea of the Hebrides. Since wind action is a critical factor in determining the nature of nearshore wave activity several calculations were made in order 1) to assess whether wind-induced wave activity is related to beach ridge crest altitude variations and 2) to formulate a dimensionless index of susceptibility to wave attack at each beach ridge location. In order to determine an accurate index of susceptibility to wave attack a computer program was developed (Appendix 1) to calculate for each site a wave energy potential index (see below) given existing wind conditions, exposure and fetch parameters.

Wave energy potential

Since the depth of wave breaking is proportional to the square of wind velocity (Zenkovich, 1967, p.348), and since the dependent wave height is proportional to the square root of the length of fetch and to the duration of winds blowing in a given direction (Zenkovich, 1967), Münch-Petersen (1938) concluded that the material-moving force (Wp) at a given coastal locality is expressed by,

$$Wp=\sum_{1}^{n}((S^2F)\ (\sqrt{D})*\cos\alpha\ )$$

where for each site,

S= mean wind velocity

F= relative frequency of winds blowing from a given direction

D= length of fetch

n= number of fetch sectors at $22\frac{1}{2}^{\circ}$ intervals

and α = the angle between the direction perpendicular to the coast and the wave path in the open sea.

(cf. Zenkovich, 1967, p.349)

Although used by Münch-Petersen to determine the direction and rate of longshore drift, the sum of material-moving forces for all available fetch sectors is a dimensionless expression of wave energy potential at any given locality. In the above equation $\cos\alpha$ attains its highest value when the wave path direction is directly towards the coastline. Later Knaps (1938) proposed that 2 $\sin\alpha\cos\alpha$ rather than $\cos\alpha$ should be used. In this case the highest wave energy values correspond to waves approaching the coastline at an angle of 45°. Knaps also suggested that a) the cube rather than the square of wind velocities should be used and b) that the cube root of the length of fetch should be used. Knaps' equation for wave energy potential is therefore

$$Wp=\sum_{1}^{n}((S^3F)\ (\sqrt[3]{D})*2\ \sin\alpha\cos\alpha\ )$$

The above formula was applied to the study area using weighted wind measurements from Tiree (1962-76) (Table 8) combined with fetch length values and angles of wave incidence produced by onshore

Table 8

Percentage frequencies of wind direction and velocity at Tiree (15 years 1962-76) and values weighted according to wind velocity.

Unweighted

hourly mean wind velocity (knots)	Wind direction (°) 350-010	020-040	050-070	080-100	110-130	140-160	170-190	200-220	230-250	260-280	290-310	320-340	%
0-3	0.0+	0.0+	0.0+	0.0+	0.0+	0.0+	0.0+	0.0+	0.0+	0.0+	0.0+	0.0+	8.1
4-6	0.6	0.5	0.6	0.3	0.4	0.6	0.6	0.5	0.6	0.6	0.5	0.5	6.3
7-10	1.4	1.0	1.2	0.6	1.0	1.5	1.4	1.4	1.6	1.6	1.4	1.3	15.5
11-16	2.6	1.4	1.2	1.0	2.3	2.6	2.8	2.7	3.4	3.0	2.6	2.3	28.0
17-21	1.7	0.5	0.4	0.7	1.8	2.1	2.1	2.0	2.2	2.0	1.5	1.4	18.5
22-27	1.2	0.3	0.2	0.5	1.5	2.2	1.9	1.7	1.6	1.3	1.2	1.1	14.8
28-33	0.5	0.0+	0.0+	0.2	0.8	1.1	0.7	0.6	0.5	0.5	0.5	0.4	5.7
34-40	0.2	0.0+	0.0+	0.1	0.4	0.4	0.3	0.3	0.2	0.2	0.2	0.2	2.6
> 41	0.0+	0.0+	0.0+	0.0+	0.1	0.1	0.1	0.0+	0.0+	0.1	0.0+	0.0+	0.5

Weighted

0-6	(x1)	0.6	0.5	0.6	0.3	0.4	0.6	0.6	0.5	0.6	0.6	0.5	0.5
7-16	(x4)	16.0	9.6	9.6	6.4	13.2	16.4	16.8	16.4	20.0	18.4	16.0	14.4
17-27	(x9)	26.1	7.2	5.4	10.8	29.7	38.7	36.0	33.3	34.2	29.7	24.3	22.5
> 28	(x16)	11.2	0.0+	0.0+	4.3	20.8	25.6	17.6	14.4	11.2	12.8	11.2	9.6
Total		53.9	17.3	15.6	22.3	64.1	81.3	71.0	64.6	66.0	61.5	52.0	47.0
% Total		8.7	2.8	2.5	3.6	10.4	13.2	11.5	10.5	10.7	10.0	8.4	7.6

unweighted values courtesy of the Meteorological Office, Edinburgh.

winds to determine wave energy potential indices for each coastal site (Table 5). (For a complete explanation of the methods used to determine Wp, see Appendix 2.) The analysis does not, however, include measurements of offshore or intertidal gradients. Furthermore it is difficult to assess the morphological significance in the nearshore zone of offshore winds that are believed to lower wave height and induce constructive rather than destructive wave action (King, 1972, pp.170-2). Offshore winds are therefore not considered in the calculation of wave energy potential.

Regression of the 48 ridge crest altitudes with the corresponding wave energy potential values (Fig. 28) indicated a strong positive correlation (R= 0.58; Signif.= 0.001). The individual variation in wave energy potential values is large and is to a considerable extent due to the crenulate configuration of the coastline and the sensitivity of the wave energy index to the presence of foreshore skerries that determine the number of fetch sectors in which waves can travel to the coast.

Summary

In the study area the lower limit of land-based vegetation as measured at four localities ranges in altitude between 2.29 and 2.75 m and corresponds with the altitude of high water mark of ordinary spring tides. If the inner edges of raised coastal terraces are similar in origin to the lower limit of land-based vegetation on modern beaches, the altitude of their inner edges corresponds approximately with the position of former high water mark. It must be remembered however that many raised beaches may have formed under polar climatic conditions and hence their inner edge altitudes may not be strictly comparable with those observed on the modern shore.

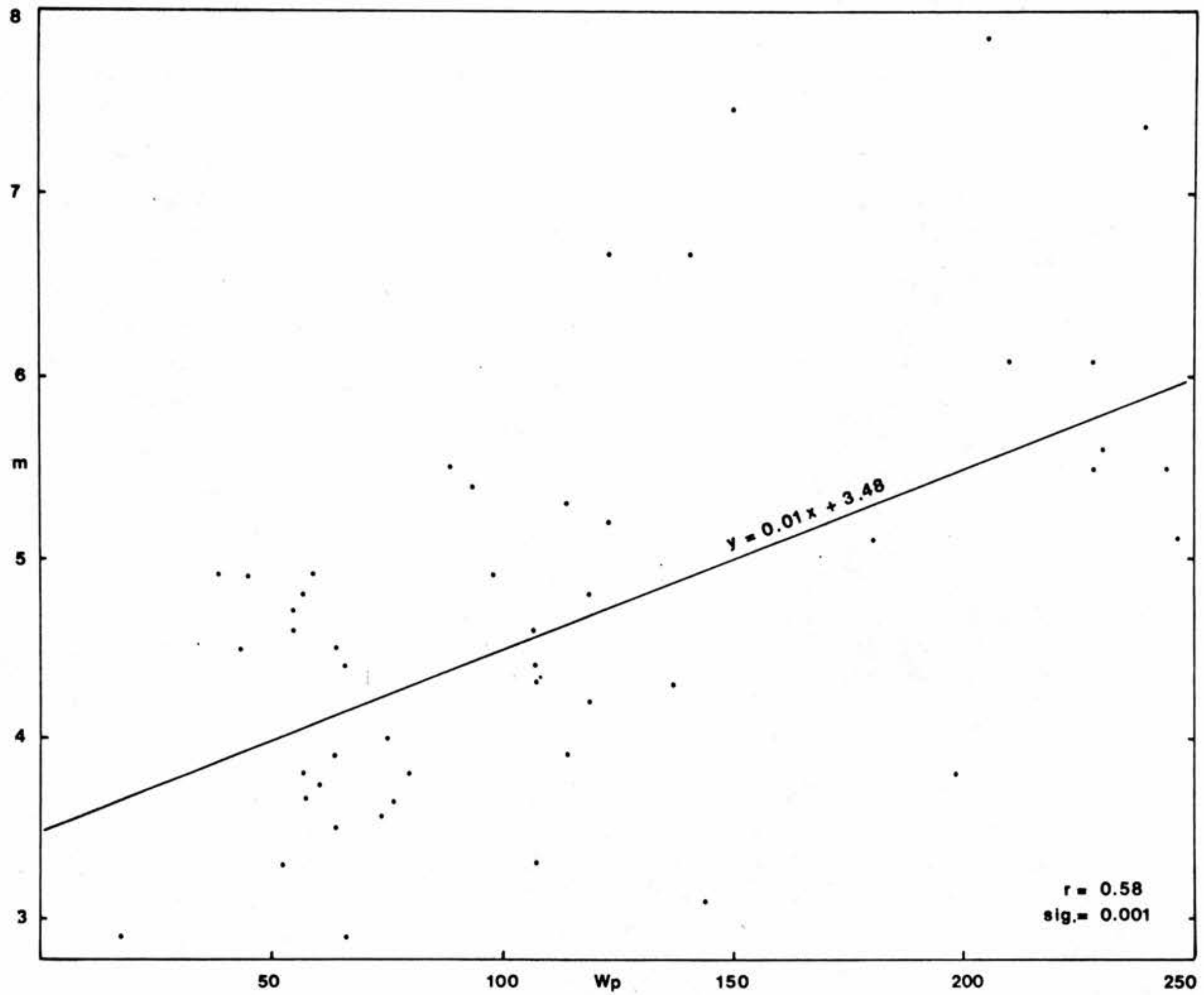


Fig. 28 Linear regression of modern storm ridge crest altitudes and wave energy potential (Wp).

In order to determine regional height variations of wave action in a low/moderate wave energy environment the altitude of the upper limit of Pelvetia caniculatus (sp.) was measured. Most Pelvetia fragments are at 2.43 ± 0.30 m and were deposited during high water mark of ordinary spring tides. Regional variations in the altitude of Pelvetia are closely related to intertidal shore width, the highest altitudes corresponding with the narrowest intertidal areas. Since the average altitude of Pelvetia corresponds approximately with the heights obtained for the lower limit of land-based vegetation, the Pelvetia altitudes form useful indicators of the upper limit of low/moderate wave action.

Two types of modern beach ridge have been identified in the study area. Firstly, storm beach ridges are formed by the combined action of erosion and deposition during and after storm-wave activity and all occur in the backshore zone above high water mark. The second beach ridge type occurs in the intertidal zone at or near low water mark. Storm ridge shingle moved seaward by destructive waves during storms forms the main debris source for this type of shingle ridge. Shingle that is moved seaward during storms is eventually returned to the intertidal zone by constructive wave action. The absence of staircases of modern beach ridge and swale topography in the study area (each modern storm ridge occurring singly) adds weight to this hypothesis and suggests that in the recent past modern storm ridges have been continually destroyed and re-formed by wave action. In the study area high tidal ranges and the consequent combing action of uprush and backwash combine to produce poorly defined intertidal shingle ridges.

Regression analyses suggest that exposure to open fetch is an important factor affecting the regional distribution of storm ridge crest altitudes. An index of wave energy potential derived from modified wind and fetch variables also provides a good explanation. Although the highest storm ridge crest is at 7.7 m, most crest altitudes occur at 4.61 $\pm$ 1.14 m. The highest storm ridges occur along the exposed coasts of NW Jura and NE Islay and most are succeeded landward by raised shingle ridge and swale topography. It is believed that raised shingle deposits form locally important sources of material that aid the vertical growth of modern storm ridges. On the more sheltered coastline of Loch Tarbert all storm ridges occur below 4.8m.

Chapter 6

Marine erosion on polar and non-polar coasts

Introduction

Shore platforms cut in a variety of rock types are common features of polar and non-polar coasts. Fossil and modern platforms occur at, below and above sea-level and vary in width from narrow ledges to broad horizontal shelves (eg. the Norwegian strandflat). Despite the widespread distribution of shore platforms, the processes responsible for their formation have not yet been fully explained (King, 1959, p.289). King (1959, 1972) considers that although relative sea-level has been at or near its present level in Britain during the last 3,000 years, the existence of cliffs cut in a variety of rock types with no adjacent shore platforms, indicates that contemporary wave planation has been incapable of forming shore platforms. Steers (1952, p.183) in a discussion of the coastline of western Scotland stated that,

> "... the raised platforms and associated cliffs of the more exposed shores clearly imply one or more periods of marine abrasion, but there is little to suggest active cutting today ... in the past wave attack must have been far more severe to have cut wide shelves in hard resistant rocks in sheltered waters."

If Steers and King are correct in their belief that contemporary wave erosion of rock on the British coast is insignificant- and all available evidence indicates that this is so- one may wonder how and when the raised shore platforms were formed. Since many writers (McCann, 1966, 1968; Synge, 1966; Synge and Stephens, 1966; Sissons, 1967b; Gray, 1974c) have agreed with the view stated by McCallien (1937) that no significant shore erosion of rock occurred during the relatively short post-glacial submergence, all raised shore platforms in western Scotland have been considered as having been formed during several long interglacial periods during each of which relative sea-level remained constant. Only recently has this interpretation been questioned (Sissons, 1974a; Gray, 1978).

Many studies of polar and non-polar coastlines have documented the formation of modern shore platforms (Nansen, 1922; Wentworth, 1938; Trenhaile, 1971, 1974; Moign, 1974). These studies have shown that polar and non-polar shore platforms differ in morphology and are formed by different shore processes. In the following pages an attempt is made to compare the nature of shore platform formation in polar and non-polar coastal environments. Thereafter the criteria by which raised shore platforms can be identified as polar or non-polar in origin are outlined. In Chapters 7 and 8 the morphological characteristics and altitude variations of individual raised shore platforms that occur in the study area are used together to establish their relative ages and origins.

1. Shore platform formation on non-polar coasts

Wentworth (1938) in a discussion of shore platforms in Oahu, Hawaii, noted four main processes of shore platform formation: 1. Water-layer weathering* 2. Solution benching (particularly effective on limestone coasts) 3. Ramp abrasion 4. Wave quarrying (wave scouring). He concluded that wave scouring and ramp abrasion

* The term "water-level weathering" was originally used by Wentworth but later as suggested by Johnson (1938) and Hills (1940) this was changed to "water-layer weathering" since "water-level weathering" could be confused with sea-level changes.

produce platforms that are later planed by water-layer weathering caused by (p.28) ".... crystallisation of salts from sea water that tend to break up the rock in the water-layer zone." Similarly Everard et al., (1964,p.300) suggested that "... spray splashed on the rocks dries and the salt, crystallising in minor cracks, expands and flakes off pieces of rock... and results in irregular pits." Wentworth also noted that narrow shore platforms formed primarily by water-layer weathering can occur at altitudes up to 20 feet (6 m) above sea-level.

Bartrum (1938) agreed that water-layer weathering is an important factor in the planation of shore platforms and suggested that this process is operative above the level of water saturation since the continual wetting and drying of rock cannot occur below this altitude. Like Hills (1949), Bartrum envisaged water-layer weathering as the main process responsible for platform development and considered wave erosion as secondary in importance. Johnson (1919, p.271), however, believed that due to the uneven composition of rocks, water-layer weathering may actually roughen rather than smooth platform surfaces. Wentworth (1938) and Bartrum (1938) also noted that protruding rock bosses frequently occur on the seaward portions of platforms produced by water-layer weathering. Although in many cases such ridges may simply be resistant rock bosses, they considered that the protruding rocks often correspond with areas of platform formerly saturated and thus not subject to the planing action of water-layer weathering. The above writers believed that the flat platform surfaces produced by water-layer weathering contrast with the seaward-sloping platform surfaces formed by ramp abrasion and wave scouring. The studies led Guilcher (1958,p.68) to state that,

> "... the efficacy of water-layer weathering is well established ... but the part played by abrasion in the formation of rock platforms remains a matter for discussion."

Other writers suggested that marine abrasion was the most important factor in shore platform formation. Edwards (1951) for example stressed the role of storm wave activity in producing shore platforms on the coast of Victoria, Australia and noted that the best developed platforms were located beneath headland cliffs. He considered that storm erosion by waves at high water level resulted in cliff retreat and platform formation, the planation of rock being primarily caused by the corrasion of sand and gravel on platform surfaces. Similarly King (1959,p.293) concluded that an inclined ramp formed where wave scouring was dominant. King (1972,pp.288-9) noted three main agencies responsible for shore platform formation: 1. Corrasion (the direct attack of cliffs by waves laden with sand and gravel) 2. Attrition (the fragmentation of loosened debris) 3. Hydraulic action (which is most effective if the waves have not broken on cliff impact). Of these, King considered hydraulic action as the most effective in shore platform formation. She stated that (1972, p.451),

> "... when waves break against a cliff, air in the cracks is strongly compressed and later, as the wave recedes, the pressure is suddenly released. These sudden changes in pressure can enlarge the crack and loosen pieces of rock..."

As a result the cliff is undermined and the periodic collapse of the overhanging parts causes cliff retreat and platform formation (cf. Guilcher, 1958).

Challinor (1949) suggested that the development of shore platforms at the foot of cliffs occurs contemporaneously with the destruction by waves of the seaward edges of shore platforms. He envisaged shore platforms as undergoing parallel retreat while maintaining uniform widths and seaward slopes. Similarly Trenhaile (1971,p.67) observed shore platforms currently being destroyed during storms on the coast of Wales. Trenhaile stressed the influence of lithology on shore platform development and noted the difficulty in distinguishing fossil and modern shore platforms that occur at similar altitudes. Trenhaile (1974,p.108) favoured "powerful waves of translation and surf generated by great storm waves..." as the main agents of platform development. In all the studies cited above modern non-polar shore platforms are best developed in areas of exposed fetch and are poorly developed in areas of restricted fetch (cf. Edwards, 1951).

Zenkovich (1967,pp141-4) stressed the power of waves to abrade and polish resistant crystalline rocks (including granite) and stated that if the rock was jointed wave pounding results in the flaking off of rock fragments. He suggested that erosion of caves by wave action is greatly assisted by the compression of air trapped in them while the inner walls of sea caves are smoothed and rounded by the beating of boulders and pebbles on their surfaces.

An additional feature indicative of marine abrasion and not mentioned in existing literature is the frequent occurrence on rocks of crescentic concussion scars that have resulted from the beating of shingle on their surfaces. The concussion scars are generally 2-3 cm in length and 2 mm in depth and are common on cliffs, sea caves and platform surfaces exposed to storm wave activity. Guilcher (1958,p.62) also noted that the mechanical action of waves on platform surfaces frequently resulted in the development of pot-holes. Zenkovich

(1967,pp.141-4) believed that the initial stages of pot-hole formation occurred under the action of waves of slight strength, which allowed the drilling material to remain in place and not be ejected. Thereafter pot-hole deepening was associated with increasingly severe wave turbulence indicated by pot-holes whose depths reach 6 m.

Although it is agreed that marine abrasion on non-polar coasts can produce shore platforms, the maximum width that platforms can attain has been the subject of disagreement and speculation. Johnson (1919) considered that extremely wide shore platforms can be cut during a stationary sea-level. He believed that, given an unlimited period of relative sea-level stability, cliff retreat and platform widening could continue indefinitely inland. He stated (1919,p.235) that "... it is not possible that waves should exhaust themselves upon a platform of their own carving, and thus fail after a time to accomplish cliff erosion." Zenkovich (1946, 1967) disagreed with the conclusions of Johnson arguing that ,

> "... there is a limit to the development of a submarine slope of abrasion, and this limit is reached when the minimum specific wave energy per unit of bottom area needed to continue destruction of the given rock is attained."

In addition, King (1959, pp.293 and 368) concluded that in the later stages of shoreline evolution, when the offshore zone has been shallowed, only a rising base level will help to prolong the stages of active erosion. King (1959,p.293) and Bartrum (1938) believed that offshore marine abrasion of rock is probably restricted to a depth of 30 feet (9 m) and platforms wider than about $\frac{1}{3}$ mile (540 m) can only be cut when sea-level is rising. When sea-level is stationary the presence of broad abrasion platforms favours constructive wave action over their surfaces and aids movement of debris to the shore zone, thus protecting the cliffs from continued wave attack. In addition, the thickness of material on an abrasion platform need not be very great before the waves fail to disturb it to its base (King, 1959,p.386). Zenkovich (1967, p.158) added that "... cliff erosion results in the accumulation on the platform of stones and boulders that are too large to be easily moved by waves." As a result platform widening does not occur until the overlying debris has been sufficiently reduced in size to be capable of being moved offshore by waves. Only once this has taken place is the platform surface exposed to active wave erosion.

As has already been stated the altitude of the inner edge of shore platforms produced by water-layer weathering varies considerably over small areas. Zeuner, for example (1959,pp.227-8) stated that the height of the inner edge of wave-cut benches has not everywhere the same relationship with the height of sea-level and varies according to exposure. However Zeuner and many other writers agree that,

> "... it may be inferred that the wave-cut bench is normally submerged and, and its inner edge at the base of the cliff, reaches to a little below or above high water mark ... the margin of error implied in determining the height of an ancient sea-level from the inner edge of a bench will rarely exceed 4 m"(Zeuner, 1959, pp.227-8) .

<u>Summary</u>

The results of previous studies indicate that planation of rock by water-layer weathering may occur anywhere within the zone of wave spray where wetting and drying of rock surfaces occurs. As Wentworth has shown, horizontal rock benches formed in this manner can never attain great widths. In contrast, seaward sloping shore platforms produced by ramp abrasion and wave scouring can reach widths of up to 600 m. Platforms greater than this width can only form when sea-level is rising. Platform surfaces formed by wave abrasion can also be subject to later planation by water-layer weathering. It is generally believed that the inner edge of a platform formed by marine abrasion occurs at or near high tide level while platforms produced solely by water-layer weathering may occur at different levels above this altitude.

In addition several morphological characteristics of shore platforms produced in non-polar coastal environments can be identified. These characteristics may be used to identify raised shore platforms formed by similar coastal processes.

1. Since wave action is primarily responsible for the formation of non-polar shore platforms, the benches are usually well-developed in exposed coastal areas and are poorly developed or absent in bays or other areas of restricted fetch.
2. Seaward-sloping shore platform ramps that cut across rock structure are most likely formed by ramp abrasion and wave scouring. The presence of flat or almost horizontal platform surfaces indicates the likelihood that they were largely formed by water-layer weathering.
3. Sea caves, shore platforms and cliffs formed by marine abrasion usually exhibit well-developed water-rounded surfaces and possess pot-holes and concussion scars on their surfaces.

2. Polar shore platforms

a) Introduction

The efficacy of marine erosion of rock on polar coasts is the subject of much disagreement. Zenkovich (1967) and Davies (1972) considered that shore platform formation on polar coasts is slow while others (eg. Sollid et al., 1973; Moign, 1974; Jahn, 1977) have described recently eroded platforms. Calkin and Nichols (1972,pp.625-43) summarise the coast of Antarctica as characterised by "... wave-cut platforms, sea caves, stacks and wave-washed surfaces." In addition Johnson (1919) described the polar strandflat as an excellent example of a series of fossil shore platforms formed primarily by marine abrasion while Grønlie (1924,p.85) concluded that in polar areas,

> "... the present shoreline is relatively ripe and as the sea cannot have stayed for any great length of time near present sea-level it must be assumed that the forming of a shoreline on an Arctic coast must be going on rapidly."

Although the narrow shore platforms described in polar coastal literature contrast in magnitude with the strandflat, several writers (eg. Nansen, 1922; Johnson, 1919; Moign, 1974; Jahn, 1977) have suggested that the features are genetically related. It is therefore apparent that if the distribution of modern and fossil polar shore platforms is to be explained, it is firstly necessary to determine the nature of polar shore erosion of rock. Secondly, the relationships (if any) between polar marine abrasion and strandflat formation must also be established. In the following pages an attempt is made to determine the morphological characteristics of shore platforms formed in polar areas. These characteristics may be contrasted with those of non-polar shore platforms and can be used to help identify raised shore platforms that originated in ancient polar coastal environments.

b) <u>Active shore erosion of rock in polar areas</u>

Zenkovich (1967,pp.173-4) concluded that,

> "... the initial stages of the process (of polar shore erosion) .. of a flat bench above water level occurs in the zone of spray ie. slightly above the level of spring tides. As this bench widens, frost-weathering is slowed down and almost ceases, since the waves break on the bench and no longer wet the foot of the cliff... it is impossible to agree that the shores of polar seas recede more rapidly than shores in the temperate zone as a result of frost-weathering and the formation of the strandflat cannot be ascribed solely to this process."

In contrast, Moign (1974) stated that in Spitzbergen gelifraction is an important agent in the formation of polar shore platforms. She observed that,

> "... the cliffs retreat with extraordinary speed during periods of thaw when blocks attacked in winter through cryoclastic activity are loosened by melting. The process is particularly active around snowbanks... cliff retreat by gelifraction releases at its foot a platform that develops rapidly because the sea, ice-free for 6 months, transports the debris- thus initiating a new cryoclastic attack. Marine action thus contributes to accelerate the speed of the effects of the process."

Although not stating that platform formation was rapid, Zenkovich (1967, pp.174-5) conceded that,

> "... at high tide, or during storms the water penetrates into the joints of the rock and even between individual minerals. When it freezes, the water expands the joints and thus gradually crumbles the rock into granules and fragments which are subsequently carried away by the waves. The outward signs of frost-weathering are distinctive. Fresh piles of sharply angular chippings are frequently formed on the shores of Arctic seas, even when the resistance of the cliff will not produce a large amount of clastic material by normal abrasion. Frost-weathering sometimes splits rocks into large blocks that accumulate at the foot of cliffs (Ionin, 1959)... but frost-weathering has little effect on some types of massive rocks, for example the granites and gneisses of the Murmansk coast."

Holtedahl (1960,p.528), referring to modern coastal stacks and sea caves in W Finmark, Norway, stated that "... contemporary marine erosion in N Norway has essentially truncated the more rounded coastal landforms." Silva (1972,pp.99-103) in Antarctica concluded that,

> "... most of the rock waste present on beaches and ice-free areas of the Antarctic peninsula and islands is derived by the shattering of rock outcrops by alternate freezing and thawing."

while in the S Shetland Islands Araya and Hervé (1971,pp.111-114) noted modern shore platforms 300 m in width. Broad shore platforms, stacks and arches are also common on the coast of S Georgia (J. Hansom, personal communication). In N Norway Sollid et al., (1973) noted rock platform fragments that occurred along the edges of fjords in places where marine abrasion is minimal. The occurrence of modern shore platforms in such sheltered areas led Guilcher (1958, p.69) to state,

> "... there seems no doubt that the ledges which surround many Norwegian fjords at about 0.5 m above mean sea-level are due to coastal frost-shattering. They cannot be explained by abrasion in such sheltered waters on coasts with so limited a fetch."

In addition McCann (personal communication) concluded that,

> "... in the Canadian Arctic archipelago high sea cliffs, undergoing active erosion at the present time, are characteristic of many areas where Lower Palaeozoic limestone sequences outcrop along the coasts of the major straits, where fetches of 50-100 km are common.... an assessment of the efficacy of marine erosion in a periglacial environment must be accompanied by the caution that, as in other environments, local conditions - rock type and altitude, degree of exposure to wave processes, etc.- produce markedly different responses within the same general area."

Nansen first drew attention to shore erosion by frost action (rather than wave action) as an important mechanism of platform formation in polar areas. He observed the development of an ice-foot on coastal rock ledges and the accumulation of frost-rived debris at the foot of adjacent slopes. In these areas the function of the waves was simply to remove frost-rived debris accumulations to offshore areas. Nansen believed that the riving capacity of freezing freshwater was greater than that of frozen sea-water (cf. Guilcher, 1958,pp.25-26). As a result he considered that shore erosion by frost is most effective in areas where fresh water could freeze on rock surfaces (eg. at the foot of cliffs where subaerial run-off resulted in the development of a freshwater ice-foot). Nansen also noted the development of rock bosses (nabs) on the seaward edge of polar shore platforms. He explained the occurrence of the protruding rock ridges as (p.40),

> "... formed by the greater effect of the disintegration by frost on the inner part of the ledge, where the accumulation of ice and snow remained longer, while it is more easily washed away by the waves along the edge. On the inner part of the ledge the accumulation of ice and snow may thus erode the rock down to a lower level without being disturbed by the waves."

Moign (1974) added that,

> "... the formation of such a platform does not demand mechanical erosion of a particularly violent nature and is proved by the presence of strandflat benches along the protected edges of fjords. For wide strandflats to develop, both gelifraction and marine erosion are needed."

It is therefore apparent that the formation of shore platforms in polar areas is due to two related mechanisms. Firstly, due to frost action, coastal cliff retreat is rapid and as a result almost horizontal rock surfaces are rapidly formed. Secondly, water-layer weathering occurs but this process of erosion is accelerated by frost-riving associated with wetting and drying of coastal rock ledges. Both processes act together and result in the development of coastal platforms in both exposed and restricted fetch environments. Previous studies have also revealed "... the fundamental importance of the character of the rock for the effectiveness of marine abrasion, even in a very severe climate" (Holtedahl, 1929,pp.163-4). In particular Holtedahl noted the efficacy with which lava, tuffs, phyllites, schistose limestones and quartzites are eroded by polar shore erosion (Holtedahl, 1929,p.163). Battle (cf. Bird, 1967) has suggested that frost-riving is ineffectual in non-porous rocks such as granite but is a powerful denudation agent in more brittle porous rocks while Nansen (1922) noted the susceptibility of gabbro, dolerite, argillaceous schists and shales to polar shore erosion and observed that igneous dykes were particularly resistant to frost-riving.

Comparatively few measurements of the inner edge altitudes of polar shore platforms have been made. Most measurements have been made by Sollid et al., (1973) who measured the altitudes of 49 platform fragments in Finmark. Their study showed that modern platform fragments occur at 1.4 ± 0.6 m above mean tide level. In addition, in Vardo, N. Norway, Sollid et al., measured platform inner edge altitudes at 1.1 ± 0.5 m above high water level while in the same area Vogt (1918,p.168) noted modern platforms occurring 0.5 m above high water mark of spring tides. Tanner (1930,p.24) noted "recent rock terraces" at a level of 0.7 m above the upper limit of the barnacle Balanus balanoides while in Vesteralen, N Norway, Moller and Sollid (1972,p.106) measured 12 rock "terraces" that occur on average 1.5 m above mean tide level. Moller and Sollid stressed the influence of lithology on platform altitudes and stated that platform surfaces tend to be aligned with the bedding planes of the rock. Finally, Nansen (1922,pp.28-32) and Løken (1962) both concluded that the inner edge altitude of modern polar shore platform fragments occurs at approximately 0.5 m above high tide level.

c) Strandflat

The term 'strandflat' is used to define a series of wide coastal platforms that occur along certain polar coasts. Grønlie (1924, pp.8 and 80) described the strandflat as "... a low undulating flat with a multitude of shallow lakes and tarns between low eminences and gently sloping ridges... the entire feature being generally 10-15 km in width." The strandflat is cut across a variety of resistant rocks and generally possesses a seaward slope of less than 1:1,000. Nansen (1922 ,pp.1-2) noted that "... it does not occur in its typical form outside regions which have formerly been glaciated or exposed to severe climates." Nansen mentioned the occurrence of well-developed strandflats in Norway, Spitzbergen, Novaya Zemlya, W Greenland, the Canadian Arctic archipelago and Alaska.

Several different strandflat levels have been identified: the most common being 8-18 m and 30-40 m above sea level (Embleton and King, 1968) while glacial striae, ice-moulding and cappings of drift have been observed on its surface. Many strandflats cross fjord rock basins where these extend seaward of the mainland.

The strandflat was early discussed by Reusch (1894) who suggested that it had been formed by marine abrasion during the Tertiary era. Other workers (eg. Rekstad, 1915; Johnson, 1919) also favoured a marine origin while Ahlmann (1919) argued that the strandflat was formed by subaerial erosion during the Tertiary era and was later subject to marine erosion during the Quaternary. Later, Strøm (1948,p.22) concluded that,

"... marine abrasion must be considered the formative agent (of the strandflat) for during not only interglacial but also interstadial periods, the coast would have lain open to marine abrasion."

Grønlie (1924,p.85) suggested that,

"... where the strandflat is not to be found so even and regular as should be expected from a flat formed by the sea, it is owing to the circumstance that the last inland ice has played a part in the final modelling of it."

Holtedahl (1929) and Dahl (1947) adopted a more extreme view and discounted marine abrasion as the main agent responsible for its formation. They considered that the strandflat was primarily formed by erosion accomplished by coastal corrie glaciers. However, Holtedahl was "certain" that,

"... marine erosion, helped by the freeze-thaw effect along the coast has contributed highly towards the formation of the strandflat in Nordland.... phyllites, schistose limestone and quartzite are evidently very easily cut by marine abrasion, assisted by freeze-thaw action near the shoreline."

Hoel (1909), Nansen (1922) and Holtedahl (1929) all noted that the rocks least able to form strandflats were granite and gneiss. However in E Hudsons Bay and James Bay, Canada, Kranck (1950,p.26) observed broad shore platforms cut in granite and gabbro, several of which were considered as the equivalent of the Scandinavian strandflat.

Of significance to the present study is Nansen's (1922,pp.214-216) opinion that the strandflat also occurs in the Shetland Isles. In a discussion of broad rock benches that occurred in this area at approximately 30 m above sea level, Nansen stated,

"... During the periods when the strandflat was developed the shore erosion by frost was probably not very effective in this region, so near to the warm Atlantic current, where the climate was probably not very cold. Hence the strandflat was not cut very broad, and as its formation took a long time, the surface of the land was at the same time to a comparatively large extent affected by subaerial denudation."

The results of the strandflat investigators indicate an agreement that marine erosion (Tertiary and Quaternary) has played an important role in its formation. If Tertiary marine abrasion is responsible for the formation of the strandflat then, as Nansen noted, it is surprising that the strandflat is limited in its distribution to areas formerly affected by glaciation. In addition its existence in these areas suggests that it has survived in areas of intense glacial erosion. If this is true then it is even more surprising that it is apparently not more widely distributed in areas that suffered less the effects of glaciation. If cold climate shore erosion was primarily responsible for its formation then direct evidence exists of the rapidity and efficacy of shore erosion of rock in polar coastal areas.

Summary

Although there is no agreement as to the precise mode of formation or age of the strandflat, it is generally agreed that marine erosion has contributed substantially to its development. As Johnson, (1919,pp.230-1) stated,

> "... Notwithstanding the doubt regarding the essential marine origin of this topographic feature it is generally considered, and probably correctly so, one of the best examples of marine abrasion on a large scale yet discovered along our present coasts... the facts leave no room to doubt however, that shore platform development occurs in polar areas and may take place at relatively rapid rates. It may also be likely, although it has not been convincingly demonstrated, that shore erosion by frost has contributed to the development of the strandflat."

Since no shore platforms comparable in width to the strandflat are known on non-polar coasts, it can only be concluded that in the past a large amount of high-latitude shore erosion of rock has taken place.

Modern shore platforms are well-developed features in many polar areas but do not compare in width to the strandflat. Polar shore platforms are primarily formed by frost action assisted by wave processes on shore platform and cliff surfaces. Polar shore platforms therefore differ in origin from those formed in non-polar areas. In contrast to non-polar shore platforms, polar platforms are not primarily formed by marine abrasion since the main role of waves is, to a large extent, to remove loose material from platform surfaces.

Since shore platform formation is occurring at present in high-latitude coastal areas, it is likely that similar shore platforms developed on the Scottish coast during stadials. No feature comparable in width to the strandflat is known on the Scottish coast. It is possible, although exceedingly difficult to demonstrate, that some of the wider raised shore platforms of W Jura, Colonsay and NE Islay that reach widths of up to 1,000 m represent immature strandflats. Alternatively, and perhaps more likely, the Scottish strandflat occurs below sea-level.

It must be stressed here that polar shore erosion has never been studied in any detail. However, all available evidence indicates that polar shore platform formation may take place rapidly. Although never previously discussed the writer is of the opinion that polar platforms have characteristic morphological features that may be used to distinguish them from platforms formed in non-polar areas. These characteristics are as follows:

1. Since frost action is an important process affecting cliff retreat and platform formation, shore platforms are exceptionally well-developed in areas of restricted fetch.

2. Polar platform surfaces are, like non-polar platforms, subject to water-layer weathering. However, in polar areas platform development is accelerated by freeze-thaw activity on platform and cliff surfaces. Owing to the importance of freeze-thaw activity, polar platform surfaces are characterised by angular rather than water-rounded surfaces. It is not clear, however, if frost action, caused by intertidal wetting and drying, is a more effective agent of platform development than frost action on adjacent cliffs. Further evidence concerning the relative importance of the two processes is presented in Chapters 8 and 9.

3. In many polar areas affected by the seasonal presence of offshore pack ice, there is less time available than in non-polar areas for the formation of shore platforms by marine abrasion. As a result, ramp abrasion and wave scour are less effective agents of platform development than in non-polar areas and consequently produce less well-defined coastal landforms. It is likely therefore that polar shore platforms do not possess the well-developed ramp abrasion profiles of non-polar platforms.

Chapter 7

Interglacial shore platforms in NE Islay, W Jura and Colonsay.

The Low Rock Platform

1. Introduction

Although intertidal shore platform fragments are common features of the Hebridean coastline their age and origins have never been considered in detail. These low platform fragments were initially noted in Colonsay and Oronsay by Bailey and Wright (Craig et al., 1911, p.64) who described "... glacially-striated and ice-moulded shore platforms... (representing)... a pre-glacial plain of marine denudation." Similar platform fragments were noted on Scarba by Peach et al.(1911,p.6) who stated that "... the surface of the rocks, still beautifully ice-moulded and striated by the ice shows for how comparatively short a period the land has stood at its present level." Wright (Craig et al., 1911,p.64) stated that the platform "... rarely or never has a well-defined margin or cliff that can be traced any distance.." and concluded that the evidence for its preglacial age was poor. Later, McCann (1968, p.28) concluded that "... the writer's fieldwork in Colonsay convinces him of the correctness of Bailey's suggestions."

In W Jura, NE Islay and Colonsay low intertidal rock platform fragments are conspicuous along long stretches of the coast and vary in size from narrow ledges to wide foreshore platforms that locally reach widths of 1 km. In the following pages a description is given of low intertidal rock platform fragments occurring in this area and is followed by a discussion of their age and manner of formation.

2. Morphology

a) NW Jura and Scarba

At numerous coastal locations in NW Jura low intertidal shore platform fragments are well-developed in the quartzite rock (Plate 13), the most pronounced landforms occurring at Bagh Uamh Mhor (NR 673993), Bagh Uamh nan Giall (NR 667987), Bagh Spereig (NR 635965) and also at several locations between Glendebadel and Corpach Bay. Similar low-level platform fragments also occur along the northern shores of Scarba, where they are cut in phyllites. The platform fragments all occur in the intertidal zone and vary in width from 10 to 50 m. They are normally located in sheltered embayments and are absent from headlands. Their surfaces are usually bare of overlying sediment and terminate landward in a low cliff, generally 2 to 5 m in elevation, that forms the frontal edge of the backing Main Rock Platform (Plate 13). The platform surfaces are flat (stacks and arches being absent) and in many cases resistant igneous dykes are planed to the same level as the adjacent surfaces on other types of rock. The platform surfaces are also often characterised by ice-moulded rock, glacial striations and innumerable pot-holes. The rock surfaces contrast markedly with the Main Rock Platform which, due to the presence of jagged inclined quartzite ridges (Plates 13 and 19), is often difficult to traverse.

b) SW Jura

During periods of low tide an intertidal platform, generally 100 m in width, is exposed on the foreshore between Rubh' Aird na Sgitheich (NR 476793) and Allt Bun an Eas (NR 458763) (Plate 5). At several locations the continuity of the flat platform surface is broken by Tertiary dolerite dykes that protrude as low elongate ridges while the

Plate 13. Low Rock Platform and Main Rock Platform fragments, Bagh Uamh Mhor, NW Jura. Cliff of Low Rock Platform fragment forms seaward edge of Main Rock Platform surface.

Plate 14. Raised shore features, Bhrein Port, W Jura. Low beach gravels are postglacial in age and mantle Main Rock Platform. High beach gravels are lateglacial in age and overlie surface of High Rock Platform. Cliff exposure is of till (in situ).

platform surface is also pitted by numerous pot-holes. Behind the shore zone rock exposures visible along stream channels indicate that the platform is continued landward beneath raised postglacial beach deposits. The raised beach deposits are in turn succeeded inland by thick accumulations of glacial till in which several stream channel are incised. Areas of bedrock that occur in the channels are considered to represent the inland continuation of the shore platform. In certain places (e.g. NR 466776) the backing cliff of the shore platform protrudes from beneath the till deposits but generally it is not visible (Plate 5).

c) <u>NE Islay</u>

On the foreshore beneath Rhuvaal lighthouse, NE Islay, two distinct rock platforms are visible. Both platforms occur in the intertidal zone but are markedly different not only in width but also in morphology. The lower rock platform is the more conspicuous and forms an almost continuous feature along the NE Islay coastline. This platform is generally 100 m in width and in the Coir Odhar embayment (Plate 3) it reaches a maximum width of almost 300 m. With the exception of the area near Rhuvaal lighthouse this platform is succeeded landward by quartzite backing cliffs generally 30-35 m in elevation. The platform surface is similar in every respect to the SW Jura feature and, like it, declines in altitude very gently seaward. In addition, the smooth ice-moulded surface of the lower platform and its considerable width strongly suggest that it forms part of the same feature described in SW and NW Jura. At Rhuvaal, however, the inner edge of the lower platform is separated from the main cliff by a second shore platform, which is 10-25 m wide. Between the two platforms is a cliff 1-2 m high. Unlike the lower platform, the surface of the higher platform is characterised by protruding angular quartzite ridges and a marked absence of smooth rock surfaces and pot-holes. This platform can be traced as a horizontal feature for a considerable distance along the coast and since its frontal cliff cuts across different beds of quartzite, there can be no doubt that here two separate intertidal shore platforms occur.

d) Colonsay and Oronsay

On Colonsay there are extremely wide intertidal rock platform fragments almost identical to those described from NE Islay and W Jura. In particular at Ardskenish, W Colonsay (NR 344918), an intertidal platform cut in Torridonian grits extends seaward for almost 1 km (Plates 15 and 16). Platform fragments of similar widths also occur at Leac Bhuidhe in S Colonsay (NR 350900) and on the Oronsay peninsula (NR 345867). In addition at Ardskenish (NR 342916) the broad low platform (described above) is succeeded landward by a low cliff that forms the frontal edge of a more angular higher shore platform. In addition near Port Mor, W Colonsay (NR 356941) and also at many other localities in Colonsay and Oronsay wide intertidal platform fragments are found in embayments sheltered from severe wave attack.

3. Altitude Analysis

The levelled altitudes of 22 platform fragments obtained along the coastlines of Scarba, W Jura and NE Islay (Table 9) range from 0.27 m to 1.81 m. Although the altitude of the platform inner edge was not measured in Colonsay and Oronsay, it occurs in the intertidal zone (illustrated by Plates 15 and 16). It was not possible to measure the altitude of the platform inner edge in SW Jura since it is not visible. All fragment altitudes were plotted in a NE-SW

Plate 15. Low (intertidal) Rock Platform fragment, Ardskenish, W Colonsay. Photograph taken during low tide.

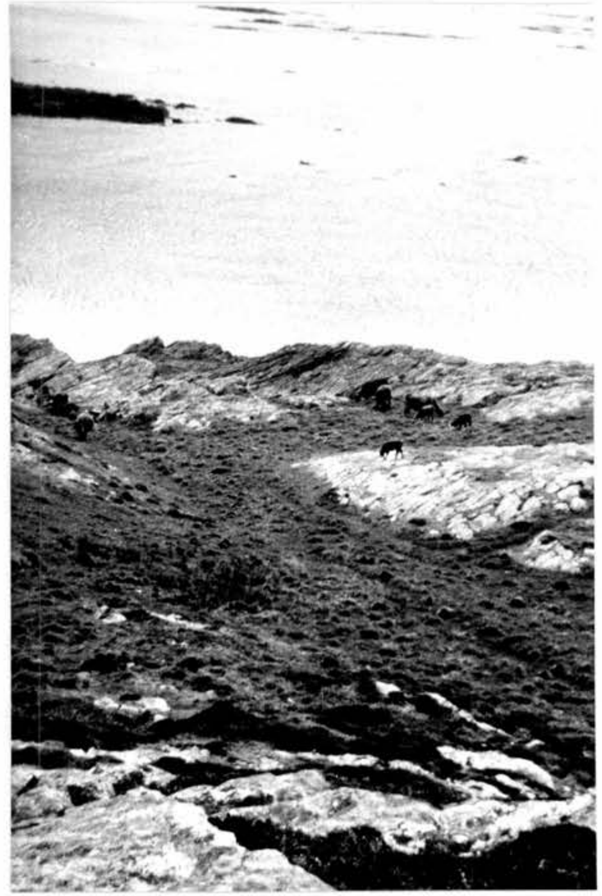


Plate 16. Low (intertidal) Rock Platform fragment, Ardskenish, W Colonsay. Photograph taken during high tide.

Table 9

Low Rock Platform fragment altitudes: Jura, Scarba and NE Islay

Site no.	Altitude (m)	Grid Reference
1	0.27	NR 41827919
2	1.81	NR 40147888
3	1.41	NR 40577890
4	1.11	NR 42437924
5	1.15	NR 52248675
6	1.18	NR 57639260
7	1.36	NR 66659864
8	0.41	NR 52758726
9	0.79	NR 67339936
10	1.00	NR 53728941
11	1.06	NM 70650273
12	1.37	NM 69790265
13	0.92	NR 58949335
14	1.52	NR 62389535
15	1.27	NR 58009279
16	1.36	NR 41827914
17	1.53	NR 40147891
18	1.30	NM 71000660
19	1.24	NR 60579434
20	1.71	NR 63539655
21	1.31	NR 57219234
22	1.13	NR 60399243

projection plane and their regional slope determined by linear regression (Fig.29a). The shore platform is virtually horizontal throughout the study area, the calculated tilt towards the centre of isostatic uplift being 0.02 m/km. The average altitude of the levelled fragments is 1.19 m.

4. Interpretation

Since all intertidal shore platform fragments that occur in NW Jura are located in bays, they cannot easily be explained as the product of modern marine erosion. The presence in this area of polished and ice-moulded intertidal platform surfaces described by previous writers and observed by the present writer shows that the platform fragments were formed prior to the last glaciation. The restricted distribution of platform surfaces in this area may therefore be attributable to later glacial erosion that removed platform fragments from headlands yet did not entirely destroy fragments located in bays.

In SW Jura till deposits overlying the platform indicate that here the platform has been overridden by ice since its formation. In NE Islay and Colonsay the presence of similar ice-moulded intertidal rock platform fragments up to 1 km in width strongly suggests that they were also formed prior to the last glaciation during a long period of relative sea-level stability. Since the SW Jura and NE Islay platform fragments are similar in morphology to those described from NW Jura, the simplest explanation is that the platform fragments described above are of the same age. In addition the occurrence of intertidal platform fragments in areas of restricted fetch suggests that some agency other than wave action (possibly water-layer weathering or frost action) has contributed to their formation.

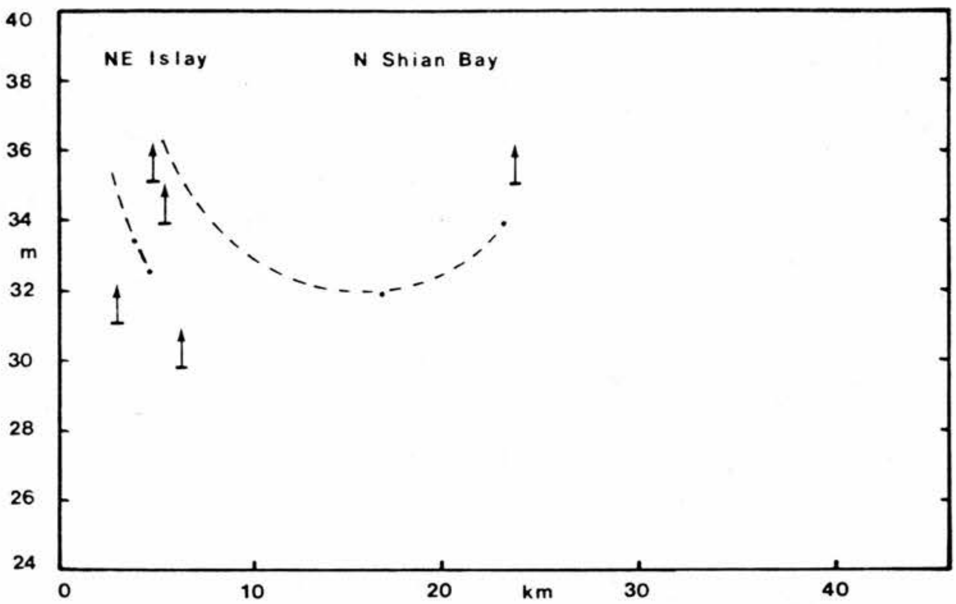


Fig. 29b Shoreline height-distance diagram of High Rock Platform fragments showing warped nature of the feature. Measured platform inner edges are shown by dots. Base of each arrow indicates measured platform surface altitude (cf. Table 11).

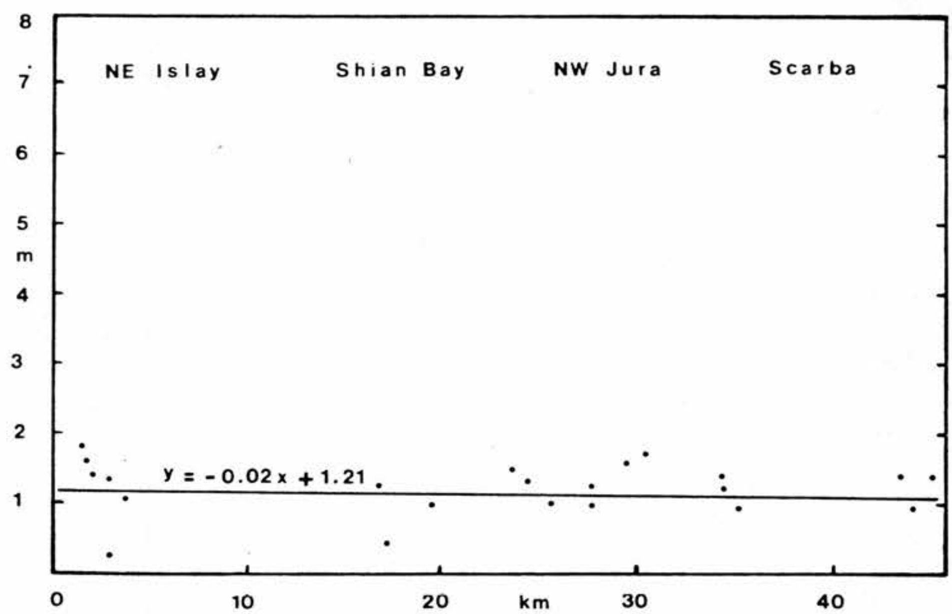


Fig. 29a Linear regression of levelled Low Rock Platform fragment altitudes.

The marked contrast in width between the wide platform fragments of NE Islay, SW Jura and Colonsay and the much narrower platform fragments of NW Jura may reflect regional variations in glacial erosion. For example the sea area by NW Jura is characterised by glacially-overdeepened trenches while to the south and west the deep trenches are replaced by shallow shelf areas. It is in the latter areas that the intertidal platform is best developed. The regional horizontality of the shore platform fragment altitudes supports the hypothesis that the platform is of interglacial age.

Since this shore platform has not before been described in detail from Scotland it is perhaps worthwhile at this point to consider some possible correlatives. Numerous examples exist in the Hebrides of low intertidal rock platform fragments of similar morphology and dimensions to those described in the study area. Bailey et al., (1924,p.386), for example, described at Loch Scridain, Mull "... a pronounced and striated rock platform, a part of which is covered at high tide." Similarly in Lewis, Godard (1965) and von Weymarn (1974,p.62) have described intertidal rock platform fragments 150 m in width that are cut in resistant Lewisian gneiss. In the upper reaches of the Firth of Lorn and in Knapdale and Kintyre Gray (1974a, Fig.7; 1978, Fig. 3) has measured the elevation of intertidal rock platform fragments. Broad intertidal rock platform fragments also occur in Coll, Tiree and Canna, while Ritchie (1972) has described similar features in the Uists. McCann (1968,p.25) has also observed intertidal platform fragments in NW and NE Skye while McCann and Richards (1969,p.20) have noted that, at several locations in SW Rhum, shore platform fragments at present sea-level are currently being exhumed by marine erosion from beneath thick accumulations of till.

Similarly intertidal platform fragments form conspicuous features of the coastline of SE Scotland where they extend as wide features from Dunbar to Berwick-on-Tweed and are also developed along the coast of Fife and Angus. Near St. Andrews, Cullingford (cf. Sissons, 1967b,p.193) has observed at sea-level well-developed rock platform fragments and cliffs that are mantled by thick accumulations of glacial drift.

There is thus good evidence of low intertidal rock platform fragments of interglacial age in W and E Scotland. It is therefore suggested that these platforms are most likely of the same age. However, the implied regional correlation of platform fragments is based primarily on a broad similarity of altitude, a method of correlation having considerable limitations (Sissons, 1967b,p.194). Nevertheless it is clear that in both E and W Scotland there exists well-defined intertidal rock platform fragments that pre-date one period of glaciation and which may be synchronous.

The High Rock Platform

1. Introduction

The existence in the Scottish Inner Hebrides of a well-defined high rock platform and associated backing cliffs was first noted by Wright (1911,p.100) who described "... a pronounced and locally well-preserved shoreline of pre-glacial age at a height varying from 90 to 135 feet (27.7-44.1m) above sea-level..." that occurs in Colonsay, Oronsay, NE Islay, Mull, Iona and the Treshnish Isles (Fig. 30). Later McCann (1964, 1968) described platform fragments at similar altitudes in W Jura, Ardnamurchan, Rhum, Skye, Eigg, Raasay and the Applecross peninsula and equated them with those described by Wright. There are several site descriptions given in Geological Survey Memoirs (cf. Bailey et al., 1924, 1925; Craig et al., 1911; Richey et al., 1930) while investigations have been conducted in Raasay (D. Smith, personal communication), Applecross (Robinson, 1977), Rhum (McCann and Richards, 1969) and Colonsay and Oronsay (Jardine, 1977). In the following pages the results of investigations by the writer in NE Islay, Colonsay and W Jura are combined with those of previous investigators in order to explain more fully the relative ages and origins of the high level features.

The high platform fragments described by Wright are best developed on exposed west-facing coasts where they are often more than 500 m in width and backed by cliffs up to 80 m in height. The platform fragments cut across a wide variety of rock types including Tertiary lavas, Dalradian quartzite, Torridonian sandstone and resistant Lewisian gneiss and, although it is likely that in several instances the platform inner edge altitudes are lithologically controlled (eg. by the Tertiary lavas of W Mull), the rocks are often truncated regardless of their dip (Sissons, 1967b,p.190). The platform surfaces are generally even and slope gently seaward (Plate 3) although in Applecross (Robinson, 1977), Colonsay (Jardine, 1977), Raasay (D. Smith personal communication) and W Jura isolated stacks protrude above the platform surface.

Wright (1911) considered that the high rock platform fragments described by him were of the same age and defined the shoreline as 'pre-glacial' in the sense of its having been formed prior to the only apparent glaciation of the district in question: he apologised to "interglacialists" for this usage of the term 'pre-glacial' (Wright, 1911,p.97). Wright stated that the pre-glacial age of the feature was proven by "... the repeated ice-moulding of its surface and above all, the superposition in some localities of large masses of boulder clay." Later McCann (1968, p.24) suggested that the term 'inter-glacial' be used instead of 'pre-glacial' since glacial erosion during the Quaternary "... must surely have resulted in more than the trifling amount of surface modifications of the platform seen in Colonsay and Mull."

Although high rock platform fragments are abundant in the Inner Hebrides, they are absent from the Outer Hebrides and N Ireland. For example, the highest known rock platform fragments in the Outer Hebrides occur at 10.2 m (von Weymarn, 1974, p.70) and at 27-32 feet (8.2-9.8 m) above mean sea-level Stornoway (McCann, 1968, p.30). In N Ireland Stephens (1957,p.143) concluded that "... The 'pre-glacial' platform of the western isles of Scotland, as described by Wright (1911) seems to be a complete anomaly, nowhere fitting into the above (Irish) sequence." Stephens noted that the highest known rock platform fragment in N Ireland was at 32 feet (9.8 m) above Irish Ordnance datum (approximately 8 feet (2.4 m) below Newlyn datum (see Stephens, 1957,p.140)).

With the exception of two recent studies (Robinson, 1977: Jardine, 1977) the inner edge altitudes of high rock platform fragments have all been measured by Abney level and thus the heights obtained can only be regarded as approximate. In Table 10 all previously recorded high rock platform fragment altitudes are given in conjunction with the rock types in which they are cut and the datum levels to which the altitudes refer.

McCann (1968, pp.26-27) suggested that the regional altitude variations of high rock platform fragments (Table 10; Fig. 30) were the result of tectonic warping and noted that the fragments decline in altitude to the north, south and west from their highest levels in Ardnamurchan (Fig. 30). Similarly Sissons (1967b,p.191) believed that the measured range of platform fragment altitudes "... is far too large to be explained merely by variations in marine erosion as influenced by rock type and exposure, so that it seems that the feature has been distorted by earth movements." McCann also supported the suggestion of Bailey _et al_. (1924,pp.386-388) by suggesting that the platform fragments occurring at 160 feet (48.8 m) at Loch Scridain, Mull and at 135 feet (41.1 m) at Uragaig, W Colonsay (Table 10) were "... anomalously high..." Later McCann and Richards (1969,p.19) suggested that in Rhum "... the high rock platform is in fact a composite feature..." with two separate shorelines occurring at 26.8 m and 19.8 m above high water mark. It has been generally agreed, however (Sissons, 1974b,p.326), that the fragments, with the exception of a few exceptionally high ones, represent one feature.

Table 10. High Rock Platform fragments in western Scotland

Location		Rock type	Altitude		Datum	Instrument	Reference
W Colonsay	Uragaig	Torridonian phyllites	135'	41.1m	HWMOST	Abney	Wright, 1911, p.101
E Colonsay	Port Olmsa	Torridonian grits	?	?		-	Wright, 1911, p.102, Fig.3.
Oronsay	Beinn Oronsay	Torridonian mudstones	120'	36.6m	HWMOST	Abney	Wright, 1911, Fig.4
Oronsay	Priory (N)	Torridonian mudstones and sandstones	120-126'	36.5-38.5m	HWMOST	Level	Jardine, 1977, p.103
NE Islay	Mala Bholsa	Dalradian quartzite	105-110'	32.0-33.5m	HWMOST	Abney	Wright, 1911, p.103
NE Islay	Mala Bholsa (?)	Dalradian quartzite	100-105'	30.5-32.0m	HWMOST	Abney	McCann, 1964, p.3
SW Islay	Oa peninsula	Dalradian limestone	95(?)	29.0m(?)	HWMOST	Abney	Wright, 1911, pp.103-4
W Jura	S Shian Bay	Dalradian quartzite	100-105'	30.5-32.0m	HWMOST	Abney	McCann, 1964, p.7
W Mull	Geometra	Tertiary basalt	105-118'	32.0-36.0m	HWMOST	Abney	Wright, 1911, p.104
W Mull	Ulva cave	Tertiary basalt	147'	44.8m	Pelvetia	Abney	Bailey et al., 1924, p.389
W Mull	Gribun (Rudha nan Goirteanan)	Moine schistose grits	98-100'	29.9-30.5m	HWMOST	Abney	Wright, 1911, p.105, Fig.4
W Mull	Gribun (Rudha na L'uamha)	Tertiary basalt	115'	35.1m	HWMOST	Abney	Wright, 1911, p.106
W Mull	Gribun peninsula	Mesozoic sediments	115'	35.1m	Pelvetia	Abney	Bailey et al., 1924, p.386
W Mull	Gribun peninsula	Moine gneiss	115'	35.1m	Pelvetia	Abney	Bailey et al., 1925, p.85

Table 10 (contd.) High Rock Platform fragments in western Scotland

Location		Rock type	Altitude		Datum	Instrument	Reference
W Mull	Gribun (Port Uamh Beathaig)	Triassic sediments	135'	41.1m	Pelvetia	Abney	Bailey et al., 1925, p.85
W Mull	Rudha' a Chaoil	Tertiary basalt	120-125'	36.6-38.1m	HWMOST	Abney	Wright, 1911, p.105
W Mull	Calgary Bay	Tertiary basalt	105-120'	32.0-36.6m	HWMOST	Abney	Wright, 1911, p.105
W Mull	Loch Scridain	Tertiary basalt	160'	45.8m	Pelvetia	Abney	Bailey et al., 1924, pp.386, 390
W Mull	Loch na Keal	Tertiary basalt	100-125'	30.5-38.1m	Pelvetia	Abney	Bailey et al., 1924, p.388
Ross of Mull	-	Tertiary granite	100'	30.5m	?	?	Craig et al., 1911, p.97
Iona	Drum Dhugail	Lewisian gneiss	115-130'	35.1-39.6m	HWMOST	Abney	Wright, 1911, p.106
Iona	Dun C'ul Bhuirg	Torridonian sandstone	115-130'	35.1-39.6m	HWMOST	Abney	Wright, 1911, p.106
Treshnish Isles	-	Tertiary basalt	75-97'	22.9-29.6m	Balanus	Abney	Bailey et al., 1924, p.389
Ardnamurchan	Kilchoan (NW)	Liassic sediments and Tertiary basalt	140'	42.7m	Pelvetia	Abney	Richey et al., 1930, p.366
Ardnamurchan	Achateny	Agglomerate	?	?	-	-	Richey et al., 1930, p.367
SW Rhum	-	Granophyre	68-80'	20.7-24.4m	HWMOST	Abney	McCann and Richards, 1969, pp.17-19
SE Rhum	Fearean Laimhrige	Torridonian sandstone	88'	26.8m	HWMOST	Abney	McCann and Richards, 1969, p.19

Table 10 (contd.) High Rock Platform fragments in western Scotland

Location		Rock type	Altitude		Datum	Instrument	Reference
W Rhum	-	Granophyre	66-76'	20.0-23.0m	?	?	Peacock, 1976,p.3
N Rhum	-	Torridonian sandstone	c.122'	c.37m	?	?	Peacock, 1976,p.2
NE Skye	Valtos, Bearreraig Bay	Liassic sediments	70-85'	21.3-25.9m	HWMOST	Abney	McCann, 1968,pp.26-7 & Fig.11.1(p.25)
W Skye	Talisker Bay	Tertiary basalt	70'	21.3m	HWMOST	Abney	McCann, 1968,Fig.1(p.25).
W Skye	Oisgill Bay	Lewisian gneiss	80'	24.4m	HWMOST	Abney	McCann, 1968,Fig.1(p.25).
Eigg	-	Tertiary basalt	?	?	-	-	McCann, 1968,p.26
Raasay	Oskaig-Holoman islands	Granophyre	91'	27.7m	HWMOST	Abney	D E Smith (pers.comm.)
Applecross	Lonbain	Torridonian sandstone	79'	24.0m	HWMOST	Abney	McCann,1968,p.26 & Fig.11.1 (p.25)
Applecross	-	Torridonian sandstone	120-113'	36.7-34.5m	Ordnance Level Datum		Robinson, 1977

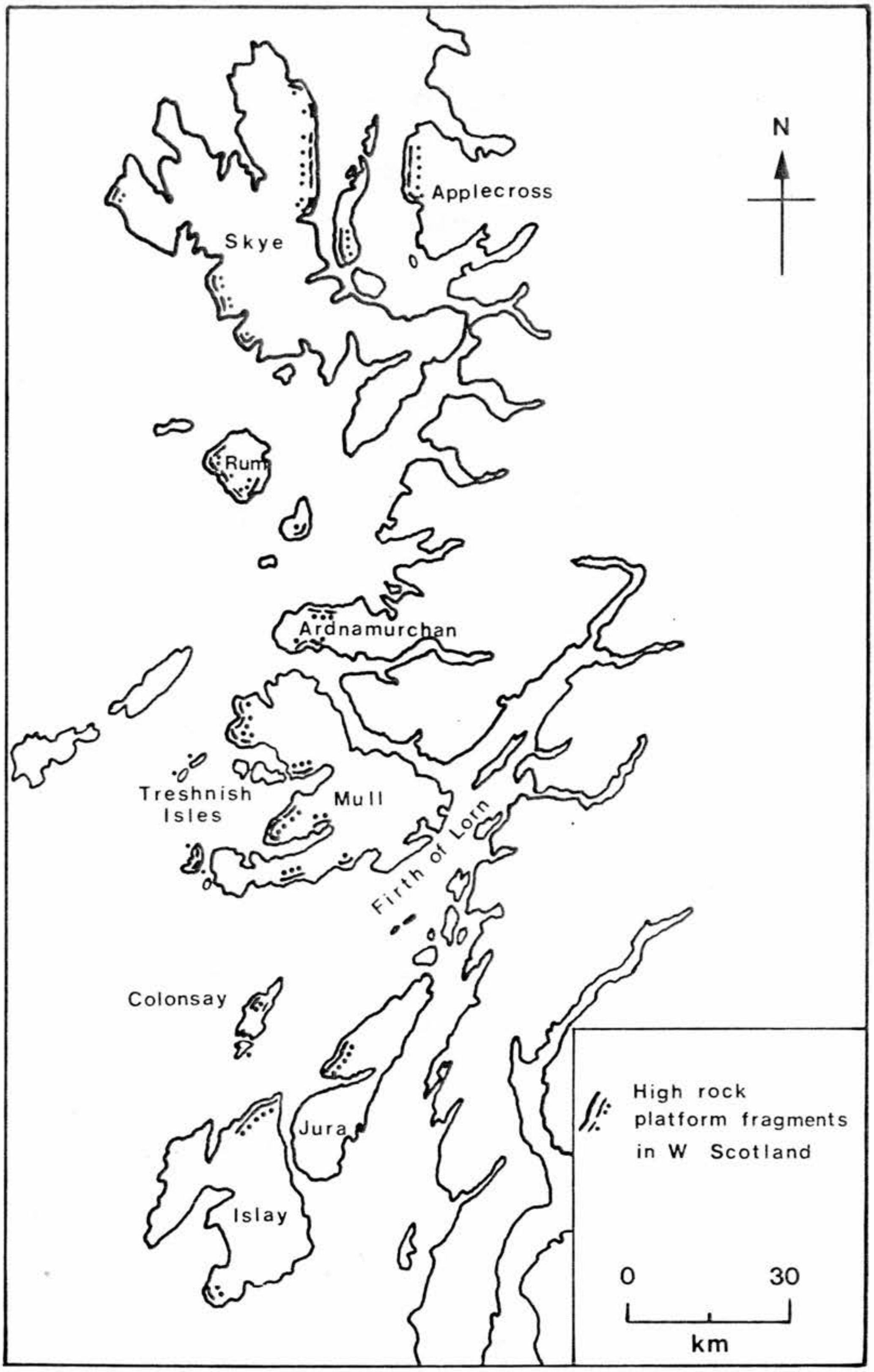


Fig. 30 Distribution of High Rock Platform fragments in western Scotland. Map compiled from information shown in Table 10.

2. High rock platform fragments in the study area: previous research

Wright (1911,p.103) first described a high rock platform in NE Islay (Plate 3) and Colonsay but did not mention the existence of a similar feature in neighbouring W Jura. In NE Islay Wright noted that boulder clay mantled the platform surface and at several locations was "... packed into the angle between the cliff and the platform..." Wright also noted near Mala Bholsa (NR 375777), that the glacial till (in situ) overlying the platform was in turn overlain by raised beach gravels. He therefore concluded that the high platform pre-dated one period of glaciation and stated that the formation of the platform was unrelated to the high sea-levels responsible for the deposition of the raised beach gravels. Wright also stated that in this area the altitude of the platform inner edge occurs between 105 and 110 feet (32.0-33.6 m) above high water mark. Later McCann (1964,p.3) measured the inner edge altitude of the same platform as 100-105 feet (30.5-32.0 m) above high water mark. McCann (1964,pp.4-5) also noted the presence of glacial till overlying the platform but concluded that the till mantling the inner edge of the platform was a solifluction deposit "... very similar in character to the "Head" which overlies the Patella beach at Gower (Wales)" McCann (1964,pp.7-8) also observed a high rock platform in W Jura, north of Loch Tarbert, whose inner edge altitude occurred at the same altitude as that measured in NE Islay (30.5-32.0 m above high water mark). He stated that

> "... the land backing the coast in W Jura is, in general, of lower elevation than in NE Islay, and thus the cliff at the rear of the platform is not so high. Consequently the solifluction material which covers the inner edge of the platform in NE Islay is for the most part absent in W Jura."

Later McCann (1968,p.25 and Fig.11.1) extended the distribution of known high rock platform fragments in western Scotland to include an area of SW Jura. In W Jura McCann did not find any exposures showing the high rock platform directly overlain by till.

The distribution and altitude of platform fragments in Colonsay and Oronsay was discussed in detail by Wright (1911,pp.100-102) and Jardine (1977,pp.102-104). Certain aspects of platform morphology, distribution, and altitude noted by them in these islands are pertinent to the interpretation of the Islay and Jura features and are therefore mentioned below.

1. At Uragaig, W Colonsay, Wright (1911,p.101) referred to the high rock platform as follows:

> "... this remarkable rock shelf... is here nearly half a mile wide and probably extended much farther seaward. Its inner angle beneath the magnificent cliffs of Tornach Mor is about 135 feet (41.1 m) above high water mark of ordinary spring tides."

Wright noted that"... everywhere one finds the surface of the plain ice-moulded, and that in many places it is striated and covered by boulder clay."

2. Jardine (1977,p.102) noted that at Uragaig sea-eroded pot-holes occur on the platform surface (NR 395978) and thus prove "... the marine nature of the platform." At the same location several degraded sea-stacks occur at distances "... 20-50 m in front of the old sea cliff... and project 5-10 m above the general level of the rock platform... and further indicate the marine origin of the platform."

3. Jardine (1977,p.103) considered that the platform was tectonically warped throughout the Inner Hebrides and also suggested that "... the so-called high level 'platform' may be composite." He suggested that in Oronsay two platforms occur with inner edge altitudes at 38 m and 18-21 m O.D. respectively.

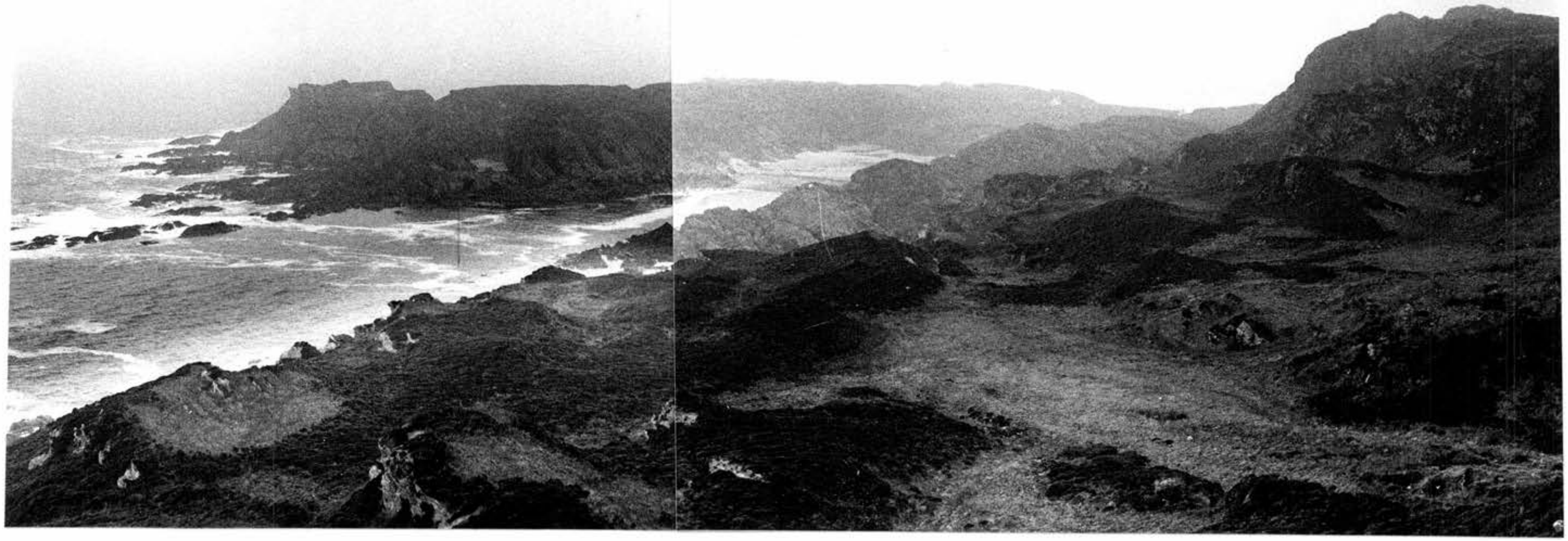


Plate 17. High Rock Platform fragment in Torridonian grits, Uragaig, W Colonsay.

From these studies, several problems concerning the postulated age and origin of the high rock platform fragments occurring in western Scotland are apparent:-

a) It is remarkable that no high rock platform fragments comparable in altitude to those of the Inner Hebrides have been observed anywhere on the coastline of the Outer Hebrides or in N Ireland since these areas are located only a short distance from the Inner Hebrides.

b) Although it is generally true that the platform fragments occur "... on the west side of areas of high relief, protected from the full force of glaciation to the east..." (Synge, 1977,p.213) there are several notable exceptions to this pattern. For example the well-developed platform fragments of NE Skye, Rhum, Applecross and Raasay are all located along the margins of deep basins that descend in places to over 300 m below sea-level. It is therefore difficult to understand how such clear features could have survived in areas of formerly intense glacial erosion unless such erosion was extremely selective.

c) High rock platform fragments are located in areas of both exposed and restricted fetch. For example the high platform fragments in SE Rhum described by McCann and Richards (1969) have a maximum available fetch of only 30 km. Similar fetch lengths also characterise the high platform fragments of NE Skye and Applecross (Fig.30). In NE Islay, W Colonsay and W Jura, however, the opposite is true. Here the high rock platform is exposed to the full force of open Atlantic fetch and reaches widths of over 600 m: its preservation from glacial erosion being favoured "... by the protective influence created by the high ground to the east" (Sissons, 1967b,p.191).

d) Since wide high rock platform fragments were presumably formed during a considerable period of minimal shoreline displacement, it is highly probable that the platform was formerly developed throughout the length of the coastline of western Scotland. However, high rock platform fragments are absent from most coastal areas of western Scotland.

3. The high rock platform of W Jura, W Colonsay and NE Islay: present analysis.

a) Morphology and distribution

In NE Islay a high rock platform is almost continuous between Lon Cnuasachd (NR 405787) and the headland of Mala Bholsa (Plate 3). This platform is also well developed farther south (outside the area of field investigation) between Mala Bholsa and Killinallan (NR 313716) where it locally reaches widths of over 1 km. East of Mala Bholas the platform is spectacularly developed (Plate 18) having a maximum width of 650 m and being backed by a cliff up to 60 m in height. Along the entire length of the NE Islay coastline the cliff backing the high platform is a degraded feature and is characterised by accumulations of vegetated talus and slumped or soliflucted till that blanket the rock face of the cliff and obscure the platform inner edge. The platform declines gently in altitude seaward and its surface is free of stacks; its frontal edge forms the backing cliff of the Low (intertidal) Rock Platform.

At Mala Bholsa and Aonan na Mala (NR 375777) the platform is indented by numerous geos that at one location cut across its entire width. Between Aonan Port an-t-Struthain (NR 384780) and Aonan na Mala several exposures reveal accumulations of till that directly overlie the platform surface and which are, in turn, overlain by raised

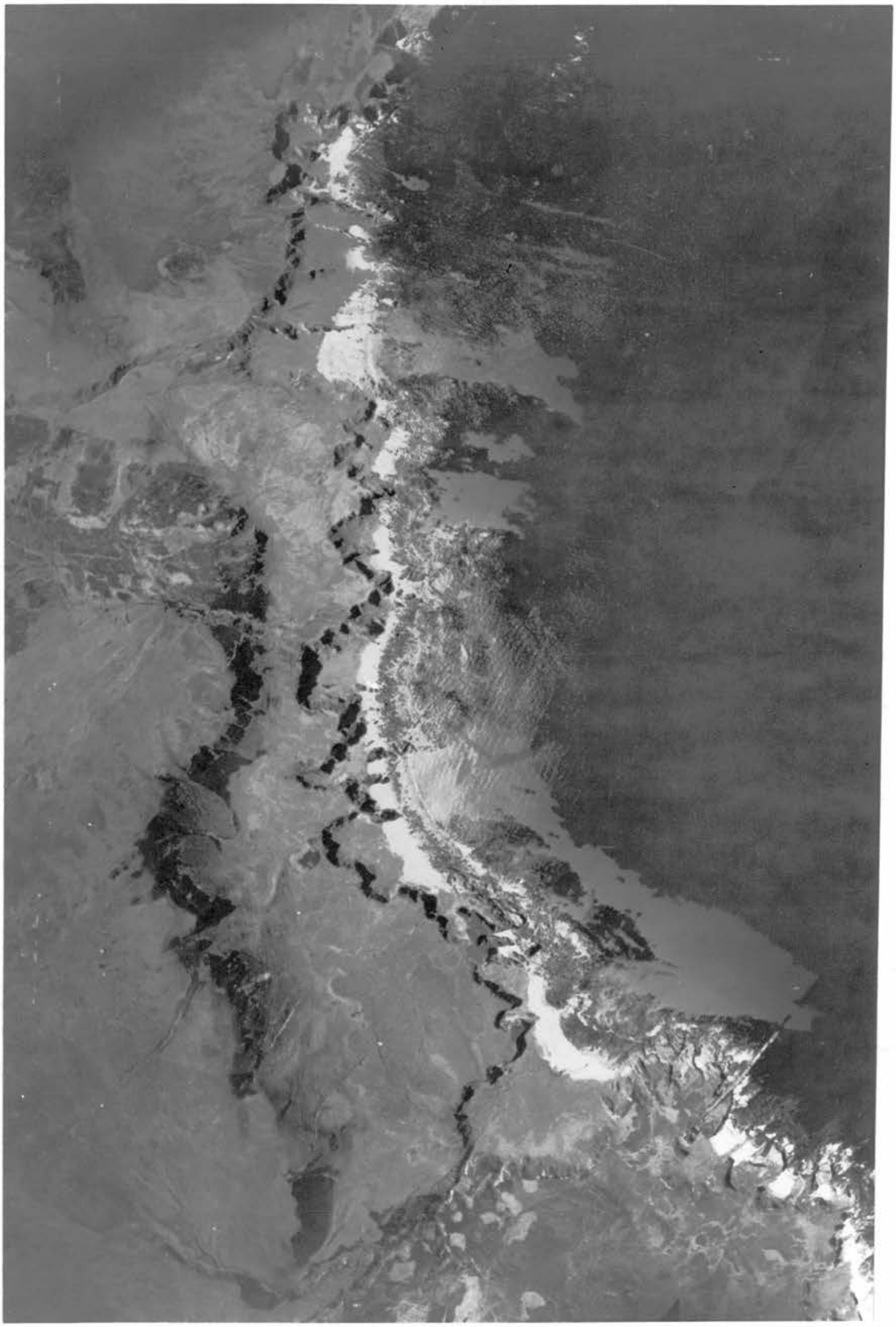


Plate 18. Aerial photograph of High Rock Platform, NE Islay (scale 1:10,000).
Ministry of Defence (Air Force Dept.) photograph Crown Copyright.

beach gravels. Here the distribution of the high raised beach gravels is limited to the seaward areas of the platform surface, generally below 27 m (Chapter 9). Landward of these high beach gravels the platform surface is overlain by till while farther east along the NE Islay coast the platform is overlain by the Coir Odhar moraine (Chapter 4).

Investigations of the SW Jura coastline failed to reveal any evidence of the high rock platform depicted by McCann (1968,Fig.11.1). However in W Jura north of Loch Tarbert the high platform described by McCann (1964) is well-developed and extends almost continuously along the coast for 7 km. Between Ruantallain and Shian Bay the quartzite platform has an average width of 350 m and in places is as wide as 600 m. Here the platform is backed by a degraded cliffline up to 50 m in elevation (Plate 27). North of Shian Bay, the platform reaches similar widths although here the backing cliff is a much more poorly developed feature. In both areas the frontal edge of the platform comprises the backing cliff of the Main Rock Platform, which also forms a continuous feature along this stretch of coast (Chapter 8).

Although in W Jura sections of the high rock platform backing cliff were described by McCann as "... low rock ridges..." (McCann, 1964, p.8; Figs. 5 and 6) these are simply rock skerries that protrude above the general platform surface. Instead the backing cliff of the platform forms an almost straight feature along the W Jura coast. Although McCann (1964,p.8) correctly states that the presence of the low rock ridges (described above) "...has led in parts to the development of complex systems of beach ridges during the lateglacial submergence .." (Plate 23) the crenulate alignment of the high rock platform backing cliff suggested by him (McCann, 1964, Figs. 5 and 6) is incorrect. It is therefore apparent that the formation in W Jura (and also in NE Islay) of the high rock platform was associated with the eventual development of a straight rather than crenulate coastline.

The low quartzite ridges that interrupt the regular seaward slope of the high rock platform surface are intermittently developed along this stretch of coastline. It is not clear, however, whether these ridges are the result of differential coastal erosion during shoreline formation or have resulted from later glacial modification of the platform surface. Direct evidence of the selectivity of glacial erosion occurs north of Loch a Mhile where a well-developed striated and ice-moulded stack (NR 51398501) protrudes above the platform surface. The W Jura platform, unlike that in NE Islay, is everywhere overlain by copious quantities of high raised beach deposits. At only two locations in W Jura have till deposits been found to separate the high rock platform from the overlying beach gravels. One of these is at Bhrein Port (NR 50688405) (Plate 14) 1 km north of Ruantallain where a wedge of orange till (<u>in situ</u>) embedded in a platform depression between two inclined quartzite ridges is overlain by thick accumulations of raised shingle. At the other location, on the banks of a stream channel 150 m north of Bhrein Port (NR 50948415), the relationship between the till and the platform surface is again clearly defined. Here the inner edge of the high platform is choked by 2.5 m of creamy lodgement till that is in turn overlain by several raised shingle ridges.

In Colonsay observations at numerous locations of till directly overlying the platform surface confirmed the observations of Wright (1911). The most spectacular platform fragment occurs at Uragaig where it reaches a maximum width of 700 m and is backed by cliffs up to 70 m high. In addition several stacks protrude above the platform surface. The most conspicuous stack occurs at Creagan (NR 39259805): it rises c. 8 m above the adjacent platform surface and exhibits no sign of ice-moulding or striation. During the formation of the Uragaig platform the part W of Uragaig farmhouse (NR 38789829) (Plate 17) was an island separated from the mainland by a coastal channel later infilled by raised marine deposits. Examination of Plate 17 demonstrates that the platform surface is often interrupted by numerous inclined ridges of Torridonian grits. This uneven surface, in conjunction with the absence of any location where the inner edge of the platform is clearly visible, leads the writer to suggest that Wright's (1911,p.101) estimate of 135 feet (41.1 m) for the altitude of the inner edge may be considerably in error.

Investigations by the writer of other areas in Colonsay where Wright (1911, p.101) identified high rock platform fragments showed that frequently the features are poorly defined and in several instances where platform fragments have previously been identified (eg. Balnahard, N Colonsay and at Aonan nam Muc (NR 373972) (cf. Craig <u>et al</u>., 1911) it can forcibly be argued that no such features exist. In addition where platform fragments are well developed on Colonsay and Oronsay the presence of resistant rock ridges that protrude above their surfaces and the existence of intervening slacks and hollows render it difficult, if not impossible, to determine accurately the altitude of platform inner edges (cf. Plate 17).

b) Altitude analysis

Since till accumulations overlie the inner edge of the high platform in NE Islay and W Jura and raised beach deposits also mantle the W Jura platform, the only locations where the inner edge of the platform is visible are along stream channels and on the sides of geos. The inner edge of the platform can be observed at only two locations in NE Islay (Table 11): at each site the altitude of the platform inner edge was determined by calculating the average of 8-14 levelled heights considered representative of the break of slope. At a few other sites where the platform edge was not visible, altitude measurements were made of the adjacent platform surface. Since altitude measurements at these latter locations are at lower elevations that the altitude of the adjacent platform inner edges (owing to the seaward slope of platform surfaces), the values obtained at these locations are minimal for the altitude of the platform inner edge. In W Jura two platform inner edge altitudes and one platform surface altitude were obtained. All values and corresponding grid references are given in Table 11. The altitudes of the shore platform fragments were plotted on a shoreline height-distance diagram drawn in a NE-SW projection plane (Fig.29b) (Chapter 3)

c) Interpretation

Since the Coir Odhar moraine in NE Islay (Chapter 4) overlies the adjacent high rock platform it is clear that the formation of the NE Islay platform pre-dates the deposition of the moraine. In addition, since it has already been shown that the Coir Odhar moraine was formed during general deglaciation of the study area, it is clear that in NE Islay the high platform pre-dates at least one period of general glaciation. The presence of till directly overlying platform

Table 11

High Rock Platform fragment altitudes, NE Islay and W Jura

Site No.	Mean altitude (m)	Grid Reference	Description of location
1	32.7	NR 37747780	Mala Bholsa geo, NE Islay
2	33.6	NR 37547770	Aonan na Mala (N) NE Islay
3	* 31.1	NR 37487752	Aonan na Mala (S) NE Islay
4	* 35.4	NR 38597780	Bholsa farm (N) stream section
5	* 33.9	NR 39057800	Bholsa farm (S) stream section
6	* 29.7	NR 39817829	Coir Odhar embayment (W)
7	* 35.3	NR 54318966	North of Shian Bay, W Jura
8	34.1	NR 53858915	North of Shian Bay, W Jura, stream section
9	32.1	NR 50948415	150 m north of Bhrein Port, W Jura

* denotes sites where altitude values are for the platform surface. All other altitudes refer specifically to the averaged levelled altitude of the platform inner edge.

surfaces in NE Islay, W Jura and Colonsay accords with the view that in all these areas they are of the same age. In NE Islay and W Jura the measured altitude variations of the platform inner edge and surface indicate that it is warped or possibly faulted. The high platform fragments that occur in these three areas also possess other similarities.

1. In each area the platform is located in exposed west-facing coastal areas and was formerly protected from the full force of westward-moving ice by adjacent high ground.

2. In each area there are indications that the features were all formed by marine abrasion. In NE Islay for example, the well-defined seaward slope of the platform surface suggests that it is most likely a shore platform produced by ramp abrasion (Chapter 6). Secondly, the occurrence on the W Jura and Colonsay platforms of sea stacks and the existence of sea-eroded pot-holes on the platform surface at Uragaig, Colonsay (Jardine, 1977, p.102), suggest that marine abrasion was important in platform formation.

When considered together the above points strongly suggest that the high rock platform fragments of NE Islay, W Jura and Colonsay were formed synchronously (and in this study can be referred to as the High Rock Platform). Since the Low (intertidal) Rock Platform that occurs throughout the same area is unaffected by warping and in NE Islay and W Colonsay is located below the frontal edge of the High Rock Platform, it can be inferred that the formation of the High Rock Platform pre-dates that of the Low Rock Platform.

In addition, since it is generally agreed that postglacial marine abrasion of rock has been slight (Chapter 10), it is believed that the two periods of relative sea-level stability required for the formation of the Low and High Rock Platforms were both of extremely long duration (each period being most likely in excess of 10,000 years).

Since it has been shown that the formation of the Low Rock Platform post-dates that of the High Rock Platform, the Low Rock Platform may have been formed during the last major interglacial, the Ipswichian (Sangamon) interglacial. It is therefore also concluded that the High Rock Platform that occurs in NE Islay, Colonsay and W Jura pre-dates at least two periods of glaciation and could well have been formed during the preceding Hoxnian (Yarmouth) interglacial.

d) Regional correlations

In view of the above evidence it is appropriate at this point to consider the age and origin of other high platform fragments that occur in the Inner Hebrides. The writer believes that the high platform fragments that have so far been identified can be broadly grouped into two categories.

1. South of the Ardnamurchan peninsula (Fig. 30) nearly all known high rock platform fragments are located on exposed west-facing coasts and are flanked inland by high ground. Most of these rock platform fragments occur above 30 m and include all known sites where there is undisputed evidence indicating till overlying platform surfaces. The distribution of these platform fragments strongly suggests that they occur in areas formerly protected from the full force of westward-moving ice. In addition the widest known platform fragments occur in these locations, widths of 700 m not being uncommon. It has also been shown that in NE Islay and W Jura these fragments are tectonically warped or faulted and most probably pre-date at least

two periods of general glaciation. Moreover, since this interglacial High Rock Platform is tectonically warped, its apparent absence from N Ireland and the Outer Hebrides need not occasion surprise.

2. North of the Ardnamurchan peninsula most high rock platform fragments lie below 30 m. Unlike the fragments described above, these fragments occur in a wide variety of coastal environments. For example the high fragments that occur in Rhum, Eigg, Skye, Raasay and Applecross are all located in the heart of the Scottish fjord area where glacial erosion was formerly severe. Secondly it is difficult to explain why many of the platform fragments occurring in this area are located in areas of restricted fetch. For example the high platform fragments of SW Raasay that reach maximum widths of 100 m (D. Smith, personal communication) possess maximum fetch lengths of no more than 5 km. Similarly the high platform fragments of SE Rhum (McCann and Richards, 1969) possess maximum fetch lengths of only 40 km.

If the high rock platform fragments and cliffs that occur in this area are interpreted as interglacial in age then one must also conclude that the Quaternary ice-sheets were able to erode deep fjords yet were unable to erode pre-existing raised coastal platforms. It is the writer's opinion that this view is untenable and it is therefore proposed that many of these high rock platform fragments do not relate in age to the higher rock platform fragments that occur farther south and may have formed by polar shore erosion. It is not believed however that interglacial high rock platform fragments corresponding to those identified S of Ardnamurchan are necessarily entirely absent in these northern areas. However, confirmation or otherwise of the above suggestions must await the results of detailed future research.

Chapter 8

The Main Rock Platform in Jura, Scarba and NE Islay.

1. Introduction and Previous Research

It was early believed that in the Inner Hebrides four shore platforms existed below the level of the high "pre-glacial" rock platform (the High Rock Platform) described by Wright (1911). Shore platforms were identified as occurring at "50 foot", "35 foot" and "25 foot" levels while separate platform fragments were also identified in the intertidal zone. The distribution of "50 foot" fragments was generally confined to the upper reaches of the Firth of Lorn (Peach et al., 1911, p.128) while Peach also described "35 foot" shore platforms in Scarba, Lunga and the Garvellach Isles. The "25 foot" platform fragments were the most conspicuous and were frequently observed to underlie postglacial raised beach gravels. As a result Wright (1928, p.99) proposed that the deposition of the raised beach deposits occurred contemporaneously with the formation of the "25 foot" platform. The postglacial sea was therefore considered by Wright (1928) as the "cliff-maker par excellence". These early studies contrasted intertidal shore platforms with those occurring at higher levels since the ice-moulded and glacially-striated intertidal platform surfaces strongly suggested that they were of a 'pre-' or interglacial age.

With the exception of intertidal platform fragments, none of the Geological Survey publications refers to glacially-striated or ice-moulded rock platform surfaces. Several of the raised shoreline descriptions by the Geological Survey officers are particularly illuminating. Peach (1901, p.141), for example, noted in W Jura,

> "... along part of the shore the raised beaches are chiefly represented by rock notches which correspond to the 100 foot, 50 foot and 25 foot beaches respectively. The platform of the 50 foot beach is perhaps the most conspicuous being generally backed by the caves.... the rock notches are generally bare, but sometimes they show a coating of old shingle."

Similarly Peach et al. (1911,p.3) noted that the "25 foot" beach is represented on Islay and Jura by a distinct rock notch and backing cliff and added (p.127) that in parts of Jura the backing cliff is cut in "... earlier drift and raised beach deposits." Although Wright (1911, 1928) suggested that the "25 foot" shoreline was postglacial in age, no age was assigned by the Geological Survey officers to the "35 foot" and "50 foot" shorelines.

The formation of the "25 foot" shore platform by postglacial marine abrasion held general acceptance until 1937 when McCallien doubted the ability of the postglacial sea to produce such a well-developed shore platform. McCallien (1937) believed instead that the overlying postglacial deposits accumulated on a regionally horizontal pre-existing rock platform that was either pre- or interglacial in age. Later Donner (1963) agreed that the "25 foot" platform described by Wright and McCallien was interglacial in age but suggested that the shoreline was regionally tilted and not horizontal. The views expressed by McCallien and Donner received further attention from Stephens (1957) who noted in E Ireland a low rock platform overlain in places by glacial drift. This rock platform, mentioned earlier by Wright (1937) and Movius (1942), was traced as far north as Belfast and had no noticeable tilt (Stephens, 1957, p.141). In addition Stephens (1957, p.143) suggested that "... there is no reason why it should not be present in Kintyre and other parts of W Scotland." Stephens considered that this shoreline was most likely related to the "25 foot" shoreline identified in the Inner Hebrides. He also noted that the altitude of the inner edge of this platform ranged between 10 and 28 feet (3.0-8.5 m) above Irish datum (approximately 2.4 m below Newlyn datum). Nowhere in Ireland, however, were shore platforms observed to correspond with the "35 foot" and "50 foot" shorelines described by the Geological Survey officers in the Inner Hebrides.

In a later study of raised shorelines in western Scotland, McCann (1966) concluded that the "25 foot" shore platform observed by earlier workers was interglacial in age and considered that there had been renewed cliffing and slight modification of the platform during the postglacial submergence. Thus McCann considered that the effect of the postglacial submergence had been to freshen the appearance of the inherited platform and backing cliffs. Unfortunately McCann (1966, 1968) included fragments of the Low Rock Platform as part of the "25 foot" shore platform and, since many of these platform fragments exhibited signs of having been over-ridden by ice (Chapter 7) he incorrectly inferred that the "25 foot" shoreline was interglacial in age. For example, McCann (1968, p.28) concluded that in Colonsay and Oronsay,

> "... It is this low level platform (ie. the Low Rock Platform) which may be considered to represent the inherited exhumed platform of the Firth of Lorn area."

As a result the "25 foot" platform of the Inner Hebrides was generally considered to be interglacial in age (Synge and Stephens, 1966; McCann, 1966, 1968; Sissons, 1967b) and of the same age as the low interglacial platform described by Stephens in Ireland. This shoreline was also believed by von Weymarn (1974,p.74) to correlate with the low rock platform fragments in the Outer Hebrides. Later Gray (1974a) described a shore platform along the Firth of Lorn and E Mull that is a continuous feature along long stretches of coast. Like McCann (1966) and Synge (1966), Gray noted that in this area the platform was overlain by "... thin veneers of what appeared to be glacial drift." Gray showed that the platform declined in altitude from 11 m O.D. at Oban to 7 m O.D. in Knapdale and sloped outwards from the centre of isostatic uplift with a regional gradient of 0.17 m/km. It thus became apparent that the "50 foot","35 foot" and "25 foot" rock platforms mentioned by earlier workers were part of the same sloping shoreline. Gray (1974a) named this rock platform the Main Rock Platform and interpreted it as interglacial in age. Regional variations in the altitude of the platform inner edge were considered to be a result of later isostatic warping and dislocation.

Later Sissons (1969, 1974a) showed that in the Forth valley a buried erosional feature is cut across lateglacial marine sediments, till and bedrock (the Buried Gravel Layer); the associated shoreline declines in altitude eastward from sea-level at Grangemouth to -11 m O.D. at Dunbar and -18 m O.D. at Burnmouth. Sissons (1974a) argued on several grounds that the erosional feature in the Forth valley is probably of equivalent age to the Main Rock Platform of W Scotland described by Gray (1974a). It was maintained by Sissons that this proposed synchronous shoreline, termed the Main Lateglacial shoreline, was formed in the severe periglacial coastal environment that existed in Scotland during the Loch Lomond Stadial. Sissons suggested that McCann's (1966) evidence of glacial modification of the Main Rock Platform was unconvincing and believed instead that, since delicate stacks and arches occur on the platform surface in areas of formerly intense glacial erosion, the platform is unlikely to be of an interglacial age. In addition he considered that the platform could not correlate with the Irish Platform since, owing to its regional gradient it should plunge below sea-level north of Ireland. Later Gray (1978) identified and measured the altitude of the Main Rock Platform in Knapdale and Kintyre. In this area the regional gradient of the shoreline was found to be 0.12 m/km. Gray concluded that the calculated regional isobases for the Main Rock Platform in Knapdale, Kintyre, the Firth of Lorn and E Mull were most easily explained by differential isostatic uplift; consequently he abandoned the interglacial hypothesis that he had earlier proposed (Gray , 1974a) in favour of an origin during the Loch Lomond Stadial.

Summary

It is now gnerally agreed that the supposedly horizontal shore platforms at 25, 35 and 50 feet identified by the Geological Survey form one single sloping shoreline that declines in altitude to the W, SW and S from its highest altitude in the Oban area. Sissons (1974a) and Gray (1978) have suggested that the shoreline was formed during the cold conditions of the Loch Lomond Stadial (Younger Dryas) and is of the same age as the Buried Gravel Layer in SE Scotland. Synge (1966, 1977) however, considers that the shore platform is interglacial in age and correlates with a low interglacial shore platform in Ireland.

2. Results of field investigations in the study area

a) Morphology and distribution of the Main Rock Platform

As stated in Chapter 2, a well-developed raised shore platform and associated backing cliff is almost continuous along the coastlines of Scarba and W Jura (Plates 19 and 20) and is locally developed in E Jura and NE Islay. North of Scarba a well-developed shore platform occurs on the neighbouring coasts of Lunga, Luing, Shuna and the Garvellach Isles. Northwards again, a raised shore platform at

Plate 19. View (looking south) of Main Rock Platform, NW Jura.

Plate 20. Main Rock Platform, Glendebadel, NW Jura.

Plate 21. Raised arch on Main Rock Platform surface, north of Glengarrisdale, NW Jura. Platform surface overlain by raised postglacial beach deposits.

slightly higher altitudes has been traced along the coastlines of the Firth of Lorn and E Mull (Gray, 1974a). Since the Main Rock Platform described by Gray N and E of the study area forms an almost continuous feature along the coast of Luing, Lunga and Shuna and since the Lunga platform is separated from a similar raised platform in Scarba by a narrow tidal channel, there can be no doubt that the Scarba platform forms part of the feature described by Gray. Similarly, since a narrow tidal race separates Scarba and Jura it is clear that the Jura platform forms part of the same shoreline. The raised shore platform that occurs in Jura and Scarba is therefore here referred to as the Main Rock Platform.

In the study area this platform (already described in Chapter 2) possesses a number of morphological characteristics that are common to all localities where platform fragments occur. In the following section the most important of these characteristics are described.

1. Platform fragments are developed in areas of both exposed and restricted fetch. The widest platform fragments occur where exposure to open Atlantic fetch is greatest as, for example, at Ruantallain in W Jura (Plate 4), where the platform attains its greatest width. However, the platform is often very well-developed in areas where the maximum fetch is only several hundred metres. Good examples of such platform fragments are at Bagh Gleann nam Muc, NW Jura (Plate 19), (NM 685004), along the N coast of Scarba, Lowlandmans Bay, SE Jura (NR 562720) and along the sheltered coastline of Loch Tarbert.

2. The raised platform surface, although generally horizontal, is often irregular in detail (Plate 13). In W Jura and Scarba angular inclined ribs of quartzite separated by hollows are characteristic

features. The inclined rock ridges often protrude well above the general level of the platform surface and in S and W Scarba the platform is so irregular that it is impossible to traverse it. In other areas raised sea-stacks and natural arches protrude above the platform (Plate 21) while pot-holes are uncommon. In addition the cliff backing the platform is often indented by numerous geos and caves (Plate 22): in the study area 97 raised caves have been identified. The cave walls are predominantly angular in outline while at many locations their angular rock surfaces exhibit concentrations of concussion scars that presumably resulted from the former beating of shingle. Concussion scars also occur locally on the platform where it is in quartzite and is located well above the reach of modern wave action, thus showing that relatively little sub-aerial weathering of rock has occurred since platform formation.

3. In NE Islay the Main and Low Rock Platforms are separated by a low cliff. The former is characterised by inclined quartzite ridges that protrude above its general level. In contrast the Low Rock Platform possesses an even surface that is striated and ice-moulded. A similar relationship is also encountered in Colonsay. Since the surface of the higher platform in NE Islay and Colonsay is similar in morphology to the Main Rock Platform fragments in Jura and Scarba, it is believed all that these features are of the same age.

4. In most areas the inner edge of the platform is bare of sediment and its altitude could thus be measured accurately. In other areas, however, the inner edge is obscured by talus and raised beach deposits and was not measured.

5. Although the Main Rock Platform is an almost continuous feature

Plate 22. Raised cave associated with the Main Rock Platform, north of Ruantallain, W Jura. Accumulations of lateglacial shingle overlie seaward surface of High Rock Platform fragment.

in the W Jura and W Scarba quartzite, it is only locally developed in the slates and phyllites of E Scarba and E Jura. In E Jura well-developed platform fragments occur at Ardlussa, Lussagiven, Tarbert, Lowlandmans Bay, Small Isles Bay and also north of Kinuachdrach. In most other parts of the E Jura coastline the inclined rock strata dip uninterrupted into the sea.

6. Investigations of raised beach deposits in the study area (Chapter 10) indicates that at the maximum of the postglacial transgression the Main Rock Platform was everwhere below sea-level. As a result many platform fragments were subjected to later wave activity. At many locations destructive postglacial wave action resulted in the removal of any overlying debris from the platform surface. In other areas however, constructive wave action resulted in the burial of platform surfaces by large quantities of debris. As a result many platform surfaces were protected from postglacial wave activity by overlying accumulations of debris.

b) Altitude analysis

The measured altitudes of the Main Rock Platform (Appendix 3) were analysed by the methods outlined in Chapter 3. Owing to the regular decline in these altitudes when plotted on a shoreline equidistant diagram (Fig. 31) and the continuity of the platform in the field, it is believed that all 175 measured fragments are part of a single feature. From the equidistant diagram it can be seen that the shoreline declines regularly in altitude from 7 m in Scarba to sea-level in NE Islay. Comparison of these fragment altitudes with those measured along the Firth of Lorn, in Knapdale and in Kintyre by Gray (1974a, 1978) accords with the hypothesis that all are correlated. Linear regression analysis of the altitudes obtained in the study

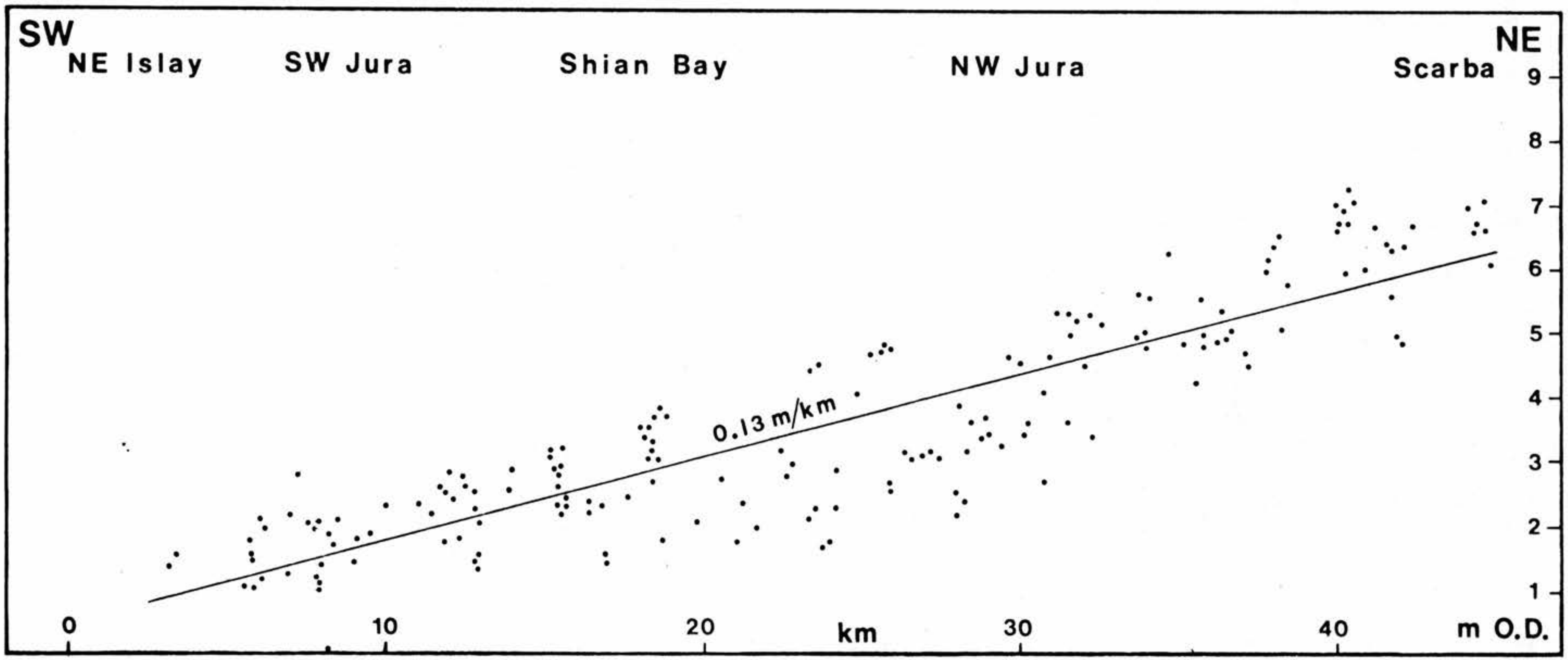


Fig. 31 Linear regression of Main Rock Platform fragment altitudes.

area indicates that the Main Rock Platform declines in altitude to the SW with a regional gradient of 0.13 m/km and, unlike the Main Rock Platform in E Mull (Gray, 1974a) is uninterrupted by later faulting.

In order to obtain a regional comparison of Main Rock Platform altitude variations the 175 platform fragments were subjected to trend surface analysis (Chapter 3). The analysis of variance test (Krumbein and Graybill, 1965) indicates that the quadratic surface provides the highest additional explanation at the 95% level (Table 12). The linear and quadratic surfaces (Figs. 32a and 33a) explain 82.6% and 87.8% of the variance respectively. Analysis of trend surface residuals (Figs. 32b and 33b) indicates that there is very little departure in altitude of individual fragments from the calculated quadratic isobase surface, residual values greater than 1.4 m being uncommon (Appendix 3).

The distribution of residuals from the first order trend surface indicates a clustering of positive values in Scarba, SW Jura and NE Islay and the occurrence of most negative residuals in NW Jura. This pattern (Fig. 32b) suggests a slight variation in gradient of the shoreline between 2 and 4 m which appears on the height-distance diagram (Fig.31). However the regional distribution of residuals for the quadratic trend surface indicates no preferred zonation of positive and negative values. In addition the distribution of second order positive and negative residuals greater than 1 m (Fig. 33b) is unrelated to the regional distribution of headlands and bays; this contrasts with the relationship noted by Gray (1974a) along the Firth of Lorn. The low residual values most probably reflect the

Table 12

F-ratio values for the contribution of first and second order trend surfaces

	Sum of squares (m)	Degrees of freedom	Variance (m)	F	Confidence level (%)
Due to linear	427.282	2	213.641	409.11	> 99.9%
Deviation from linear	89.752	172	0.522		
Due to quadratic	26.745	3	8.915	23.90	> 99.9%
Deviation from quadratic	63.007	169	0.373		

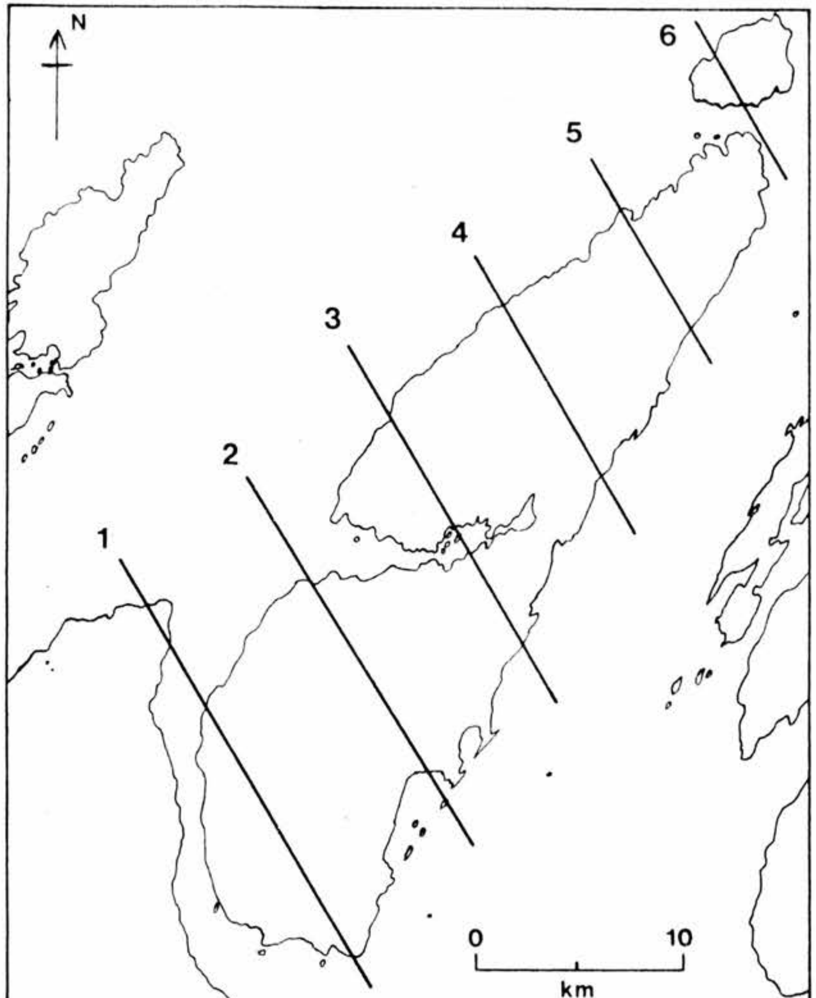


Fig. 32a First order trend surface isobases for Main Rock Platform. Values in metres

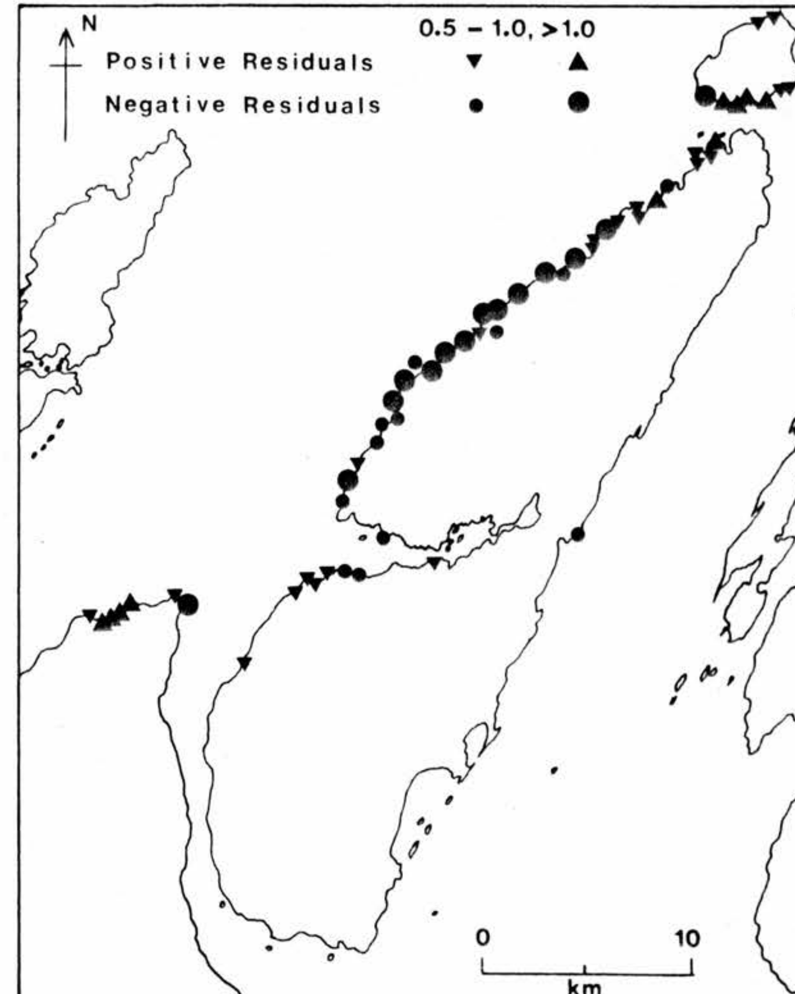


Fig. 32b First order trend surface residuals for the Main Rock Platform. Values in metres.

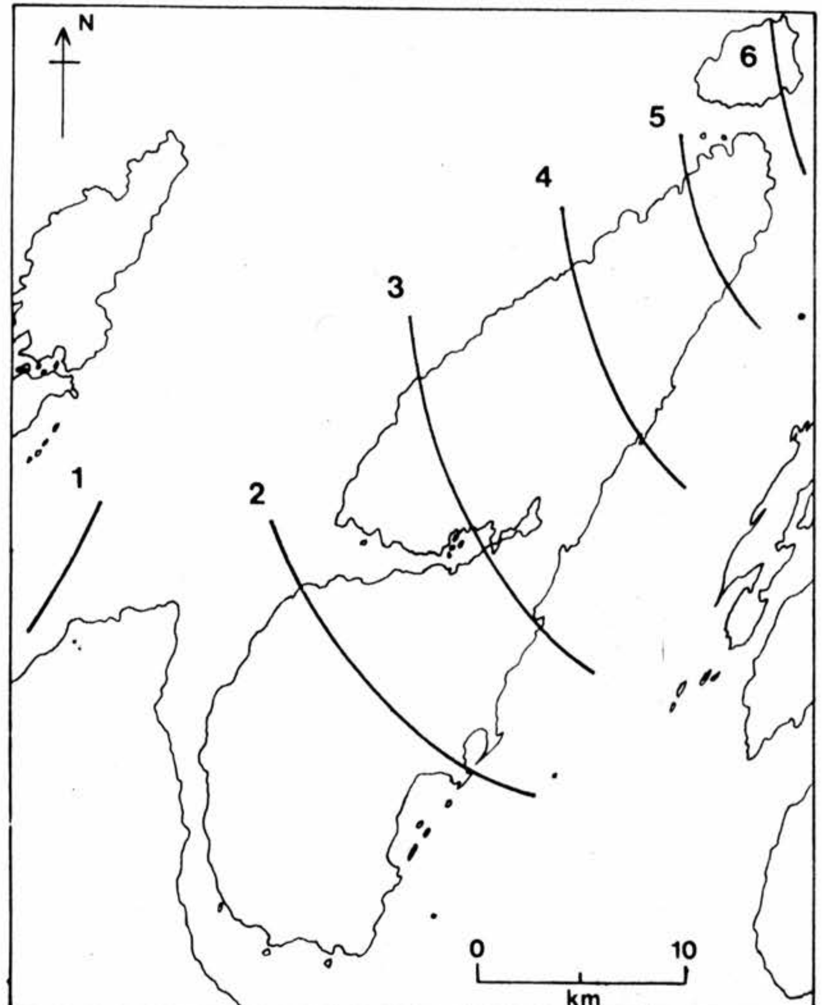


Fig. 33a Second order trend surface isobases for Main Rock Platform. Values in metres.

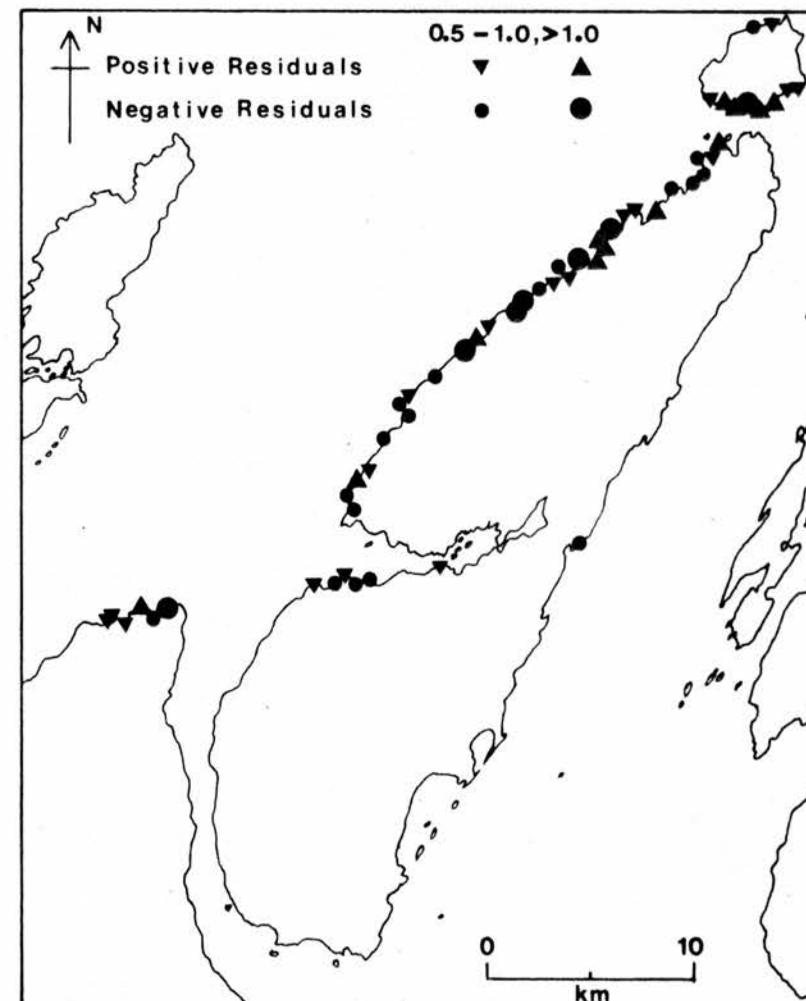


Fig. 33b Second order trend surface residuals for Main Rock Platform. Values in metres.

accurate altitude measurement of bare platform inner edges, the altitude of each fragment representing the average of 5-14 levelled heights. In addition platform altitude variations caused by changes of rock type are effectively eliminated since most platform fragments are developed in Dalradian quartzite.

c) Platform widths and cliff heights

In order to determine the amount of rock removed in W Jura during the formation of the Main Rock Platform, platform widths and cliff heights were measured at 526 regularly spaced locations (Appendix 4). The former were measured on 1:25,000 aerial photographs and the latter determined in the field by Abney level.

Since the Main Rock Platform merges with and crosses the interglacial Low Rock Platform in NE Islay and SW Jura, these areas were excluded from the analysis, for it is likely that here the extent of the former platform is influenced by the presence of the latter. Two methods were employed to calculate the average volume of rock removed per metre of coastline. The product of platform width and cliff height in conjunction with the planimetric measurement of the coastline length in which the Main Rock Platform occurs, indicates that a maximum volume of 2,074 m^3 of rock per metre of coast were removed during shoreline formation (Appendix 4). If it is assumed instead that a uniform seaward slope was in existence prior to platform formation, 1,037 m^3 of rock per metre of coast were removed during this period. This latter figure is equivalent to the removal of 49 million m^3 of rock from the W Jura coast. It should be noted however that the values of 2,074 m^3 and 1,037 m^3 of rock per metre of coast are average volumes of rock removal and that locally much larger volumes of rock were removed during platform formation (see Appendix 4). For example at Ruantallain a maximum of 6,300 m^3 of rock per metre of coast were removed.

d) Age

Since the relative merits of the inter- and lateglacial hypothesis concerning Main Rock Platform formation have already been discussed in detail by Sissons (1974a) and Gray (1974a, 1978), it is intended here to summarise these arguments with special reference to the features described in the study area.

The interglacial hypothesis

If glacio-isostatic uplift is at present incomplete an interglacial platform should tilt towards the centre of isostatic uplift. If glacio-isostatic uplift is complete an interglacial shore platform should be regionally horizontal (e.g. the Low Rock Platform). If it is proposed that the Main Rock Platform is interglacial in age and that its regional tilt is the result of isostatic uplift caused by glacial erosion it is difficult to explain the regional horizontality of the Low Rock Platform.

The studies of Gray (1974a, 1978) and the writer indicate that the Main Rock Platform declines in altitude away from the centre of isostatic uplift and possesses a regional gradient of between 0.11 and 0. 17 m/km. The calculated isobase patterns determined for the Firth of Lorn and Knapdale areas by Gray and for the study area by the writer (Figs. 32a and 33a) are extremely difficult to explain unless the shoreline is interpreted as having been formed during or since the decay of the last ice-sheet. Since the gradient of the highest lateglacial shoreline in the study area is 0.61 m/km

(Chapter 9) and the gradient of the Main Postglacial Shoreline is 0.05 m/km (Chapter 10) it is likely that the Main Rock Platform was formed in part of the intervening period.

If it is proposed that the Main Rock Platform is interglacial in age, it is also difficult to explain the presence on platform surfaces of many delicate stacks and arches (Plate 21) in areas of formerly intense glacial erosion. Moreover, the existence of Main Rock Platform fragments on the edge of glacially-overdeepened trenches suggests that if the feature is of an interglacial age, subsequent glacial erosion was extremely selective. Additionally if the shore platform is of interglacial age one might expect the presence throughout W Scotland of wide platforms in areas formerly protected from the full force of westward moving ice-sheets. This is not the case. In addition there is no evidence to indicate that the platform has been glaciated while investigations in Scarba, NE Islay and Jura of 97 caves in the Main Rock Platform cliffs failed to reveal the presence of any till deposits.

The lateglacial hypothesis

A lateglacial age avoids the difficulty of explaining the existence of many shoreline features (e.g. platform fragments, cliffs, stacks, arches, geos and caves) in an area where glacial erosion was intense. A lateglacial age is also favoured by the absence of convincing evidence demonstrating glaciation of the platform. Although striated and ice-moulded platform surfaces have been reported in the Firth of Lorn area (McCann, 1966, Synge, 1966, Peacock, 1975), Gray (1978) has noted that at some of the described locations the glacial evidence is on the frontal slope of the platform and not on its surface. Moreover, striae occurring on the platform surface need not indicate over-riding by glacier ice. For example, Dionne (1973,p.185) and Gray (1978) have suggested that striated platform surfaces may result from the grounding of pack-ice or small ice-bergs. In addition, the presence of till overlying platform surfaces (McCann, 1966,p.89) can result from the slumping or solifluction of till from backing cliffs and need not imply glaciation of the platform (Gray, 1978,p.11). Finally, the writer believes that many of the ice-moulded and striated platform surfaces that have been described by McCann (1966) and Synge (1966) are Low Rock Platform fragments. Many of the coastal locations where ice-moulded platforms have been described are in areas where the sloping Main Rock Platform crosses the Low Rock Platform (e.g. in Colonsay and Oronsay (McCann, 1968)).

The Main Rock Platform is poorly developed or absent within areas occupied by Loch Lomond Readvance glaciers yet is well-developed outside the former glacier limits. Moreover, the excellent preservation of the Main Rock Platform on only a limited portion of the west coast of Scotland is simply explained since it passes beneath sea-level away from the area most affected by glacial rebound (Sissons, 1974a).

In summary, there is convincing evidence that the Main Rock Platform is of lateglacial age. Hence it cannot be correlated with the low Irish interglacial platform.

e) Origin

On the above interpretation a difficulty that appears to arise is the production of such a well-developed shoreline in the relatively short period available. Since modern marine erosion is negligible in areas where the Main Rock Platform is best developed (Gray, 1978,p.160), it appears that during the lateglacial period coastal erosion was capable of cutting rapidly the "wide shelves in hard resistant rocks in sheltered waters" described by Steers (1952). Sissons (1974a) suggested that the platform was formed by polar shore erosion during the cold climate of the Loch Lomond Stadial and stated (p.46) that "... only about a thousand years appear to be available (for platform formation): a period that may seem too short." It was originally considered that this cold period lasted from 10,8000 to 10,3000 B.P. (Sissons, 1974a). However Sissons (1974b, pp.319-320) suggested that,

> "... it appears that the Loch Lomond Readvance began before, perhaps well before 10,800 B.P. .. the total time that glaciers of the Loch Lomond Readvance existed in Scotland was well in excess of the c.500 years conventionally assigned to the stadial."

Mangerud (1970) has suggested that in Norway the Younger Dryas Stadial (generally considered equivalent to the Loch Lomond Stadial) lasted from 10,900 to 10,000 B.P. If a duration of 1,000 years is assumed for the Loch Lomond Stadial it can be calculated from the data given earlier that in the study area the minimum rate of rock removal was 1.04 m^3 per metre of coast per year: equivalent to a minimum cliff retreat rate of 7 cm/year. This figure may be compared with a maximum rate of cliff retreat of 0.9 cm/year above the fossil rock glacier by the Paps of Jura interpreted as having formed during this period, and with a value of 2.3 cm/year (17m/750 years) (Sissons, 1976,p.189) for cliff retreat above a fossil protalus rampart in Wester Ross. The above figures for inland cliff retreat are not directly comparable however, since unlike coastal cliffs the cliff face areas behind the fossil rock glacier and protalus rampart were gradually buried by talus aggradation. In contrast, debris that was frost-rived from coastal cliffs during this cold period would have been continually removed seaward by waves, thus allowing marine processes and frost-riving to operate continually at the base of cliffs. The calculated rate of cliff retreat of 7 cm/year may also be compared with retreat rates of 2.5-5.0 cm/year measured by Jahn (1961,p.22) on modern coastal limestone cliffs in Spitzbergen.

It may seem remarkable that the platform was formed during a period when sea-level might be expected to have been relatively unstable due to eustatic and glacio-isostatic changes. Unfortunately insufficient information is available concerning the chronology and rate of relative sea-level changes in W Scotland during the lateglacial. Gray (1974b) has suggested that at Oban the sea was 20 m above its present level c. 11,6000-11,700 B.P. and that during the Loch Lomond Stadial it was no higher than 10-11 m. Sollid et al. (1973,p.303) suggested that in Norway during the Younger Dryas there may have been a lengthy pause in the marine regression.

The "wide platforms in resistant rock in sheltered localities" described by Steers (1952) therefore demand very special conditions in order to explain their formation. Sissons (1974a) suggested that these conditions were characterised by processes of frost-shattering of rock in the semi-diurnally wetted intertidal zone and cliff base. Gray (1978) added that marine abrasion and debris removal may have been assisted by the apparent storminess during the Loch Lomond Stadial (Sissons and Sutherland, 1976). In addition he proposed that since diurnal temperature fluctuations in W Scotland during the Loch Lomond Stadial were most probably greater than in high latitude coastal areas, very intense frost-shattering of rock in the coastal zone would have favoured the development of wide platforms. In Scarba, Jura and NE Islay the angular morphology of the platform surface and its occurrence in areas of restricted fetch (Chapter 6) supports the general hypothesis that the platform was formed by polar shore erosion. It is not agreed, however, that frost-shattering of rock on the intertidal platform surface was the most important mechanism of polar shore erosion. As already stated (Chapter 6) the freezing of brine in rock joints does not favour rapid rock shattering since the presence of salt renders the ice cellular and porous (cf. Guilcher, 1958). It is believed instead that the principal process of polar shore erosion was frost-riving associated with the freezing of fresh water on rock surfaces (cf. Nansen, 1922; Jahn, 1961; Davies, 1972; Moign, 1974), particularly on cliffs.

It is therefore suggested that the angular platform surface having been created by cliff retreat was not substantially modified by intertidal freeze-thaw processes in the semi-diurnally wetted intertidal zone. Indeed the angular morphology of the platform indicates that freeze-thaw processes assisted by water-layer weathering were unable to form flat platform surfaces. It should also be noted that features characteristic of marine abrasion (e.g. sloping ramp abrasion platforms, water-rounded rock surfaces and pot-holes) are almost entirely absent. As a result it is suggested that the role of waves during platform formation was primarily as a transportation agent, removing loosened debris seaward and causing deposition in the nearshore zone.

Additional evidence for the development of shore platforms by polar shore erosion during the Loch Lomond Stadial is provided along the shores of the former ice-dammed lake of Glen Roy in the Scottish Highlands (Sissons, 1978). Sissons concluded that during the brief existence of the lake, shore platforms reaching widths of 10 m or more were cut in metamorphic rocks at several levels. The existence in Glen Roy of well-developed shore platforms associated with former ice-dammed lakes (also areas of restricted fetch) further indicates that shore platform formation occurred at extremely rapid rates in Scotland during the Loch Lomond Stadial.

Although a detailed discussion of postglacial sea-levels in the study area is given in Chapter 10, it is perhaps worthwhile here to consider the effect of wave action during this period on the Main Rock Platform since Wright originally proposed that the platform (referred by him as the 25 foot platform) was formed by postglacial marine abrasion. Since evidence has already been presented to show that postglacial marine abrasion could not have formed such wide platforms it is also equally plausible that the postglacial sea accomplished virtually no erosion of rock. During the maximum of the postglacial transgression in the study area the Main Rock Platform surface was entirely submerged (Chapter 10) with waves extending to the backing cliffs and into the caves. However the absence of water-rounded rock on the platform surface strongly suggests that virtually no postglacial marine abrasion of rock occurred. Indeed the presence of concussion scars on the platform combined with its angularity suggests that postglacial marine abrasion was negligible. In addition the locally extensive accumulations of postglacial raised beach deposits on the platform surface would have favoured the protection of the backing cliffs and the platform surface from continued wave attack.

Correlation with the Buried Gravel Layer, SE Scotland.

Sissons (1974a) proposed that the Main Rock Platform correlates with a buried erosional shoreline (the Buried Gravel Layer) in SE Scotland. Sissons (1974a,p.45) argued that since both erosional shorelines record a unique event in the area of their occurrence and both are developed in areas sheltered from severe wave attack and possess similar gradients, they are probably of the same age. Sissons (1974a,p.45) stated that,

> "... the close similarity of these two gradients (0.15 m/km and 0.17 m/km) is probably in part fortuiutous for ..... shoreline gradients in E and W Scotland are unlikely to be the same for the same feature."

The calculated isobase pattern and shoreline gradient of 0.13 m/km for the Main Rock Platform in the study area provide additional support for this hypothesis.

Correlation with the Norwegian Main Line

Sissons (1974a, p.47) suggested that the Main Rock Platform and the Buried Gravel Layer of Scotland were formed synchronously with the Norwegian Main Line (P12) described by Marthinussen (1960,p.418) who noted that,

> "... A characteristic of the P12 shoreline which is sometimes more a zone than a line, is that commonly it has been abraded in rock often with a very marked rock terrace. Therefore it probably embraces a considerable period of time with only slight fluctuations of relative sea-level."

Andersen (1968, p.138) also stated that,

> "... the terraces of the Main Line are thought to have been formed by more processes than marine abrasion alone such as frost-shatter and erosion by sea-ice and bergs."

Andersen also noted that the Main Line was developed irrespective of exposure and was commonly located in narrow fjords and sounds while Sollid et al. (1973,p.245) argued that the clearly defined rock terraces of the Main Line show that it most likely belongs to a long period of minimal shoreline displacement. Since rates of isostatic uplift in N Norway were considerably faster than those that occurred in Scotland during the Loch Lomond Stadial (the Main Line gradient being 1.0-1.4 m/km (Aarseth and Mangerud, 1974,p.19; Follestad, 1972 pp.55-57)), the presence of the well-developed Main Line shore platform in Norwegian fjords indicates that the rate of shore platform formation must have been extremely rapid. Since permafrost existed down to sea-level in W Scotland during the Loch Lomond Stadial (Sissons, 1974b), it is probable that the polar shore processes responsible for the formation of the Norwegian Main Line were also operative on the W of Scotland coastline during the same period. Since the rate of glacio-isostatic uplift in W Scotland occurred at a slower rate than in Norway it is extremely likely that polar shore erosion of rock in W Scotland was more effective than in Norway and resulted in the development of an extremely clear shore platform and cliff.

Conclusions

Measurements in W Jura of the Main Rock Platform width and cliff height at 526 locations indicate that the platform has an average width of 70 m and reaches a maximum width of 420 m at Ruantallain, W Jura. The backing cliffs range considerably in height but are generally about 30 m. It has been shown that in the study area the Main Rock Platform declines in altitude from 7 m in N Scarba to sea-level in NE Islay and is tilted to the SW with a regional gradient of 0.13 m/km. The curved isobase pattern calculated for this shoreline (Figs. 32a and 33a) indicates an uplift centre to the NE. The isobase pattern is in accordance with the proposal (Gray, 1978) of an ellipsoid of isostatic uplift with a major axis aligned approximately N-S over the W Highlands.

It is believed that the Main Rock Platform was produced by polar shore erosion during the cold climate of the Loch Lomond Stadial and forms part of the same shoreline described farther north and east by Gray (1974a, 1978). Platform formation was primarily caused by the freezing of fresh water on coastal cliffs that induced intense frost-shattering of debris. It is suggested that the role of marine abrasion during platform formation was negligible. It is proposed instead that the main role of waves was the seaward removal of frost-rived debris. The lack of altitude variation of the platform inner edge between headlands and bays is therefore not unexpected.

The shoreline is considered to correlate with the Buried Gravel Layer of SE Scotland and the Main Line of Norway and to have been formed during a considerable period of minimal shoreline displacement. It is estimated that in the study area during the Loch Lomond Stadial a minimum volume of 1.04 m^3 of rock per metre of coast per year was removed, equivalent to a minimum average rate of cliff retreat of 7 cm/year. It has also been shown that in SW Jura, NE Islay, and Colonsay where the Main Rock Platform passes below sea-level, it merges with and crosses the interglacial Low Rock Platform. It is also considered that many previous site descriptions of interglacial shore platform fragments at or near present sea-level in W Scotland relate to the Low Rock Platform and have been confused with Main Rock Platform fragments that occur at similar altitudes.

It has been shown that the Main Rock Platform passes beneath present sea-level within the study area: hence the traditional correlation between the Main Rock Platform and low rock platform fragments in the Outer Hebrides (von Weymarn, 1974) is erroneous. It is proposed instead that the low interglacial platforms of Ireland (Stephens, 1957) and the Outer Hebrides are more likely to be of the same age as the Low Rock Platform.

During the later postglacial marine transgression the Main Rock Platform was submerged. In many areas constructive wave activity during this period resulted in the burial of the platform by shingle deposits and consequently its protection from wave processes. Elsewhere the presence of bare angular platform surfaces indicates that the wave action that did occur was associated with virtually no abrasion of rock. The apparent absence of depositional shorelines associated with the Main Rock Platform may be a result of their later destruction or burial by postglacial seas (Chapter 10).

Chapter 9

Lateglacial shorelines

In Jura certain coastal areas are characterised by extensive unvegetated raised ridge and swale topography. The raised shingle deposits early attracted the attention of the Geological Survey officers (Summ. Prog. Geol. Surv., 1899,pp.163-4) who described them as,

> "terraces which vary in elevation from the present beach up to the 200 foot contour which they do not pass... they consist of plateaux of bare (quartzite) shingle... occasionally one of the lower terraces has been a storm beach which has ponded back the drainage and now appears as a rampart of shingle with a small lake behind it."

The only detailed study of these raised coastal features was by McCann (1964) who, in a wider study of raised shorelines in western Scotland, measured the altitude of several raised shingle ridges and concluded (p.13),

> "... the highest shingle deposits in W Jura represent the work of the Late-glacial sea during the period of maximum submergence and are equivalent to the highest raised beach terraces of the mainland of W Scotland... these features are part of the 100 foot beach shoreline."

In N Islay and W Jura McCann also noted the presence of a "100 foot" shoreline characterised by distinct "beach terraces" that bear a closer altitude relationship than raised shingle ridges to the former water planes.

Although the conclusions of McCann will be presented in greater detail later in this chapter, it is here only necessary to outline the three types of lateglacial raised shoreline assemblage that occur in the study area and which form the subject of this chapter. Firstly, lateglacial coastal terraces occur in association with fossil river terraces. Secondly, raised coastal terraces and cliffs cut in drift are conspicuous features along long stretches of coastline. Finally, raised shingle spreads often occur as extensive staircases of shingle seaward of the highest coastal terrace. Together these landforms record the morphological action of former wave activity from deglaciation to the cold periglacial conditions of the Loch Lomond Stadial.

1. Raised shorelines associated with river terraces

a) Introduction

It has already been shown that in Jura, Scarba and NE Islay the Loch Lomond Stadial was characterised by a severe periglacial rather than a glacial climate and a relatively low sea-level (Chapter 8). However, in several valleys river terraces occur at high altitudes in the coastal zone and are succeeded seaward by high raised beaches. In order to determine the relative positions of sea-level to which the river terraces and raised beaches are related, shoreline levelling was conducted in several areas where the morphological relationships between these features are well-defined.

b) Morphology and altitude relationships of river terraces and raised shorelines

1. Corpach Bay, NW Jura.

In NW Jura the lower part of the Corpach valley (Fig. 1) is choked by thick accumulations of till. The Corpach river drains westward to Corpach Bay and is deeply incised into this till. In the coastal zone till mantles a series of quartzite ridges and is succeeded seaward by the Main Rock Platform cliff. On top of this cliff is a wide till-cut platform and cliff that, at the mouth of the Corpach valley,is succeeded inland by a series of narrow river terrace fragments.

South of the Corpach river the coastal till platform is well-developed for several hundred metres along the coast and is overlain in places by vegetated and unvegetated raised beach ridges that terminate seaward at the edge of the Main Rock Platform cliff. The altitude of the break of slope at the base of the till cliffline varies between 40.2 and 39.8 m (Table 13) and has an average altitude of 39.95 m.

2. Glendebadel, NW Jura.

In contrast to the Corpach valley, the lower part of Glendebadel (Fig. 1) is not filled with till deposits. Instead the river valley consists of steeply sloping quartzite ridges mantled by thin veneers of till. In this area the Main Rock Platform cliff is succeeded landward by high rock ridges and, as a result, raised coastal terraces are poorly developed. However in the lower part of Glendebadel narrow river terrace fragments, usually less than 50 m in width, flank both sides of the valley and are separated from each other by the deeply incised channel of the Glendebadel river. The river terrace fragments are succeeded seaward by a small area of raised beach deposits that is mantled by several vegetated beach ridges.

On the south side of the valley an exposure above an incised meander bend of the river (NR 624949) shows stratified fluvial deposits overlain by a small delta, the foreset beds of which slope seaward at c. 9°. The fossil delta is in turn overlain by a series of raised shingle ridges and swales that declines in altitude seaward. The landward base of the highest ridge is at 35.6 m; landward of the delta the river terrace fragments rise in altitude inland. On the N side of the valley fluvial deposits are separated from raised beach deposits by a low cliff. The altitude of the raised beach at the cliff foot is 35.5 m and accords in altitude with the features already

Table 13

Terrace fragment altitudes of lateglacial shorelines L1 and L2.

Fragment	Range (m)	Mean (m)	Grid Reference	No. of points	Length (m)
1	39.8-40.2	40.0	NR 556912	8	400
2	36.1-36.3	36.2	NR 535811	2	40
3	36.2-36.6	36.4	NR 535879	4	50
4	36.3-35.6	35.8	NR 537877	5	60
5	34.9-35.3	35.0	NR 534869	4	50
6	36.1-36.5	36.3	NR 526862	4	100
7	36.4-37.2	36.8	NR 524861	9	450
8	36.0-36.8	36.4	NR 522860	4	150
9	34.9-35.3	35.1	NR 521856	4	200
10	34.9-35.9	35.4	NR 520852	9	200
11	34.9-34.5	34.7	NR 516854	5	70
12	34.2-34.7	34.4	NR 510844	6	150
13	33.2-34.3	33.9	NR 509836	10	80
14	25.7-26.7	26.4	NR 397784	10	180
15	25.9-26.7	26.5	NR 401786	6	200
16	29.5-31.1	30.5	NR 520795	5	90
17	29.3-30.6	30.1	NR 519795	5	80
18	29.4-29.6	29.5	NR 511802	5	50
19	30.0-29.8	29.9	NR 510802	4	40
20	26.5-27.4	26.9	NR 466773	9	275
21	26.5-26.8	26.7	NR 463771	4	100
22	24.9-25.1	25.0	NR 451737	4	100
23	24.3-24.7	24.5	NR 449732	5	70
24	24.3-24.9	24.6	NR 448728	5	60
25	24.0-24.2	24.1	NR 448723	5	75
26	24.2-24.7	24.5	NR 447721	5	55
27	25.4-25.6	25.4	NR 452747	4	95
28	25.8-26.3	26.1	NR 456756	5	80

described from the S side of the valley.

3. Corran river, SE Jura

In SE Jura the Corran river drains the western Paps of Jura and discharges into Small Isles Bay (Fig.1). The lower river course is flanked by high river terrace fragments that, north of Three-Arched Bridge (NR 543722), are replaced seaward by a series of raised vegetated beach ridges. Anderson (1888,p.329) and Charlesworth (1955,p.881) considered that in this area a valley glacier formerly extended down to sea-level during the Highland Readvance (generally considered equivalent to the Loch Lomond Readvance in the western Highlands). Anderson for example suggested that,

> "... a stream of ice has followed the course of the Corran burn, and has brought down a large quantity of detritus which fills the lower course of the valley. The stream has cut deeply into this rubbish, and the deposits show no appearance of stratification... the sea must have stood at the 40 foot level when this glacier ceased to carry down material."

Charlesworth concluded that, "... the Corran glacier reached the then 100 foot sea-level... its flood waters probably destroyed the outer moraines since outwash now extends upwards from the higher beach." The "higher beach" referred to by Charlesworth consists of a series of vegetated raised beach ridges that is succeeded inland by river terrace fragments. Exposures in these river terrace fragments indicate a still grey till overlain in places by stratified fluvial deposits. The crest of the highest beach ridge is at 34.1 m , the landward base of which if interpreted as indicative of the highest lateglacial sea-level, shows that the latter attained 31.3 m.

4. Glenbatrick, SW Jura.

At the northern end of Glenbatrick well-developed river terrace fragments are succeeded seaward by lower raised beach deposits

(NR 519794). On both sides of the valley the junction between the fluvial and raised beach deposits is defined by a small cliff 1.5-2m in height. On the E side of the valley the average altitude of the raised beach that parallels the cliffline is 30.5 m (Table 13); on the western side of the valley the average raised beach altitude is 30.1 m. South of this area river terrace fragments rise inland.

5. Glen Maol, Scarba.

At Glen Maol (NM 713037) till overlain by fluvial deposits is replaced seawards by a terrace of raised marine gravels that defines the upper limit of former marine activity. Terrace fragment altitudes were measured at 7 locations and range between 37.6 and 38.0 m with an average value of 37.8 m.

6. Glengarrisdale, NW Jura.

In Glengarrisdale raised beach deposits extend almost 1 km inland (Chapter 2). Here the lower course of the valley is flat-floored and filled with deposits of stratified sub-angular gravels. River meandering has reworked large areas of sediment and consequently high coastal terrace fragments are only locally developed in areas where flat rock surfaces have provided favourable depositional environments. Since no raised coastal terrace fragments occur in this area that are clearly independent of bedrock control no altitude data were obtained.

c) Interpretation

Since the river terrace fragments in the Corpach valley are succeeded seaward by a raised coastal terrace it is suggested that they were formed when sea-level was here at 40 m. Farther north at Glendebadel the formation of the delta and river terrace fragments occurred when sea-level was approximately 35.6 m. In Glen Maol, Scarba the altitude of the high coastal terrace and the fluvial deposits indicate that both features were formed when sea-level was about 37.8 m. In Glenbatrick the altitude relationship between the raised coastal and river terrace fragments implies that both features were formed when sea-level lay at 30.4-30.5 m. A similar relationship occurs in the Corran river valley and indicates that the fluvial deposits are here graded to a sea-level at 31 m. In NE Islay (Chapter 4) it has already been shown that outwash deposits associated with the Coir Odhar moraine were formed when sea-level was at 26.4 m.

2. Lateglacial coastal terraces

a) Introduction

In the study area raised coastal landforms produced during the highest stand of the lateglacial sea are conspicuous features along long stretches of coastline. The features are usually characterised by a high platform and cliff cut in till. The raised platform generally extends seaward beneath adjacent raised beach ridges while the cliff, although generally low, is easily recognisable in the field and on aerial photographs (Plates 23 and 24). In W Jura north of Ruantallain the highest coastal terrace forms a clear feature as far as N Shian Bay (Plate 23) while in SW Jura a clearly defined terrace occurs W of Glenbatrick (Plate 24) and N of Inver (Fig. 35). Between Inver and Glenbatrick till cliffs and platforms occur at Loch na Sgrioba (Plate 5) while in E Jura high coastal terrace fragments are developed in Lowlandmans Bay and Small Isles Bay.

The only reference to the Islay and Jura terraces is by McCann (1964, p.13) who stated that,

"... in NW Islay the inner angle of the terraces is between

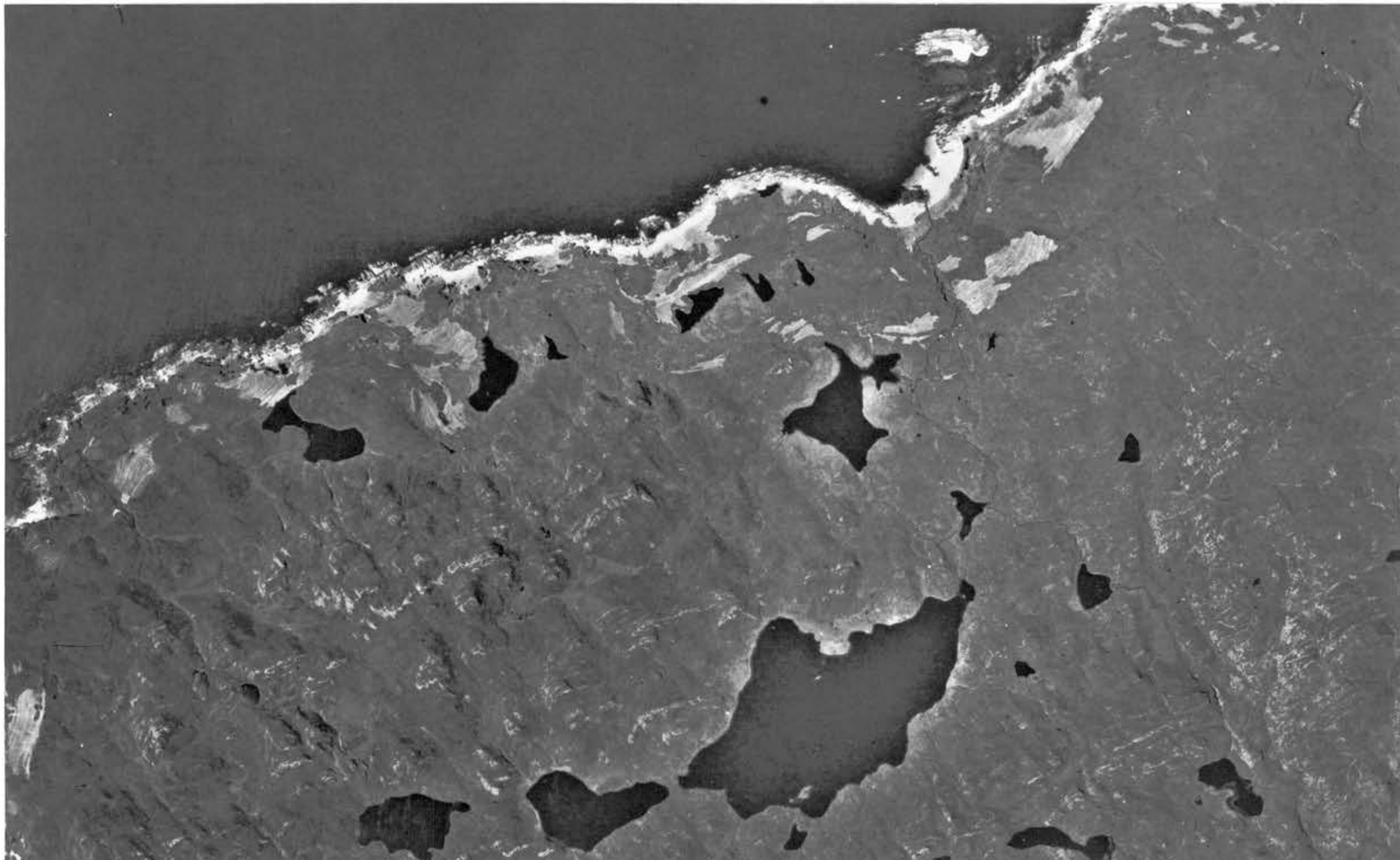


Plate 23. Aerial photograph of raised shore features between Ruantallain and Shian Bay, W Jura (scale 1:25,000). High lateglacial coastal terrace is clearly defined. Ministry of Defence (Air Force Dept.) photograph Crown Copyright.

Plate 24. View (looking south) of Bagh Righ Mhor lateglacial shingle spread. Across loch, a high lateglacial coastal terrace is well defined. In far distance is Glenbatrick and Paps of Jura.

> 85 and 92 feet (25.9-28.0 m) above H.W.M. and in SW Jura is 98 feet (29.9 m). It is reasonable to postulate on the basis of these figures that the maximum level of the Late-glacial sea in the Islay Jura region was just in excess of 100 feet above present H.W.M."

Unfortunately McCann did not recognise the sloping nature of lateglacial shorelines and instead considered the "100 foot" beach as a regionally horizontal feature. In addition, since McCann did not indicate the location of his "beach terrace" measurements, it is difficult to ascertain whether or not his terrace altitude measurements are accurate.

In the study area 150 altitude measurements were made on 28 coastal terrace fragments (Table 13). Owing to the possibility of altitude autocorrelation on single terrace fragments (Gray, 1972, 1975a) the mean altitudes of individual terrace fragments were used in subsequent analysis. Since none of the terrace fragment lengths is greater than 500 m it was unnecessary to subdivide terrace fragments (cf. Cullingford, 1977).

b) <u>Morphology and Altitude</u>

1. <u>Ruantallain to N Shian Bay, W Jura.</u>

<u>Golden Spread, N Shian Bay</u>.

In this area a high unvegetated shingle ridge is separated from a till cliff by a broad terrace. The average altitude of the northern fragment of the terrace was measured as 36.2 m and the altitude of the southern fragment as 36.4 m (Fig.34 (1 and 2)). Both fragments revealed a maximum altitude range along their lengths of 0.4 m. Approximately 350 m farther south the average altitude of a high terrace fragment located landward of the highest raised beach ridge is 35.75 m (Fig. 34 (3)).

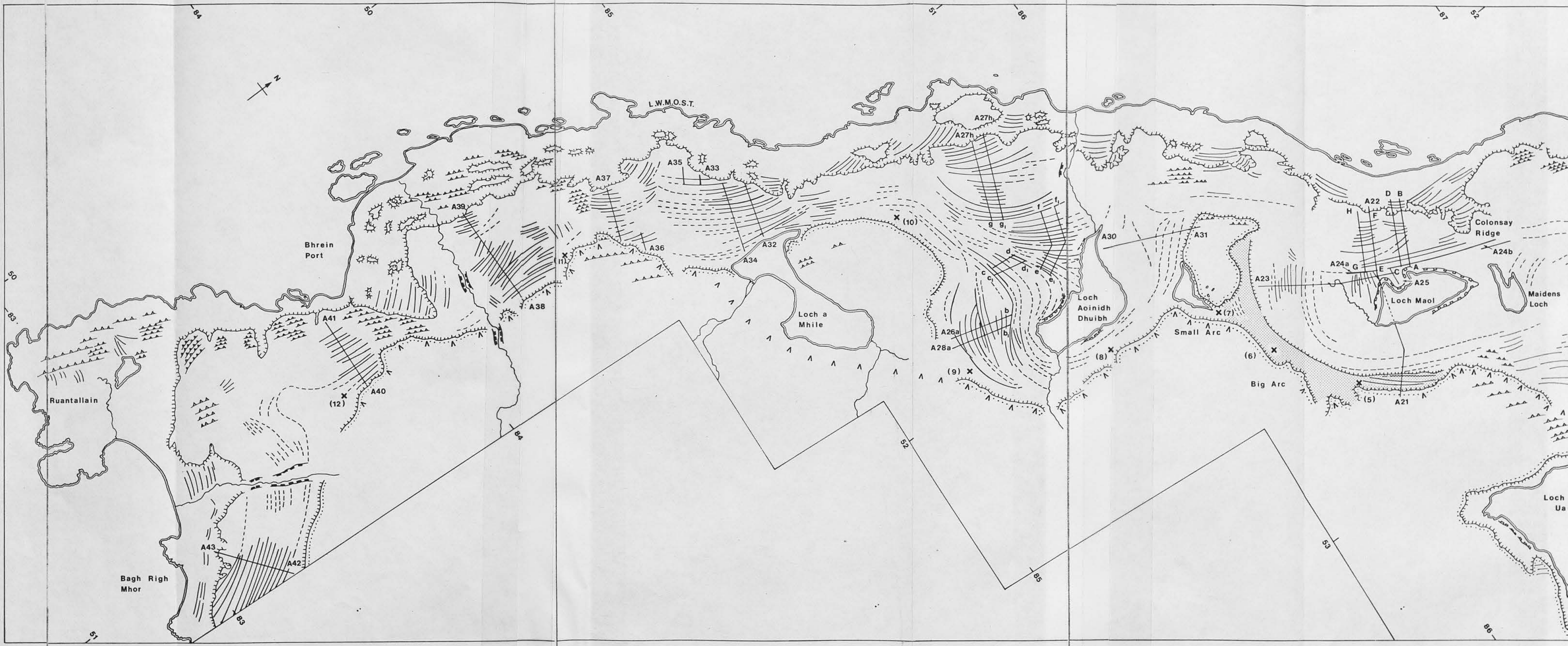


Fig. 34 Raised shorelines and measured profile locations, Ruantallain to Shian

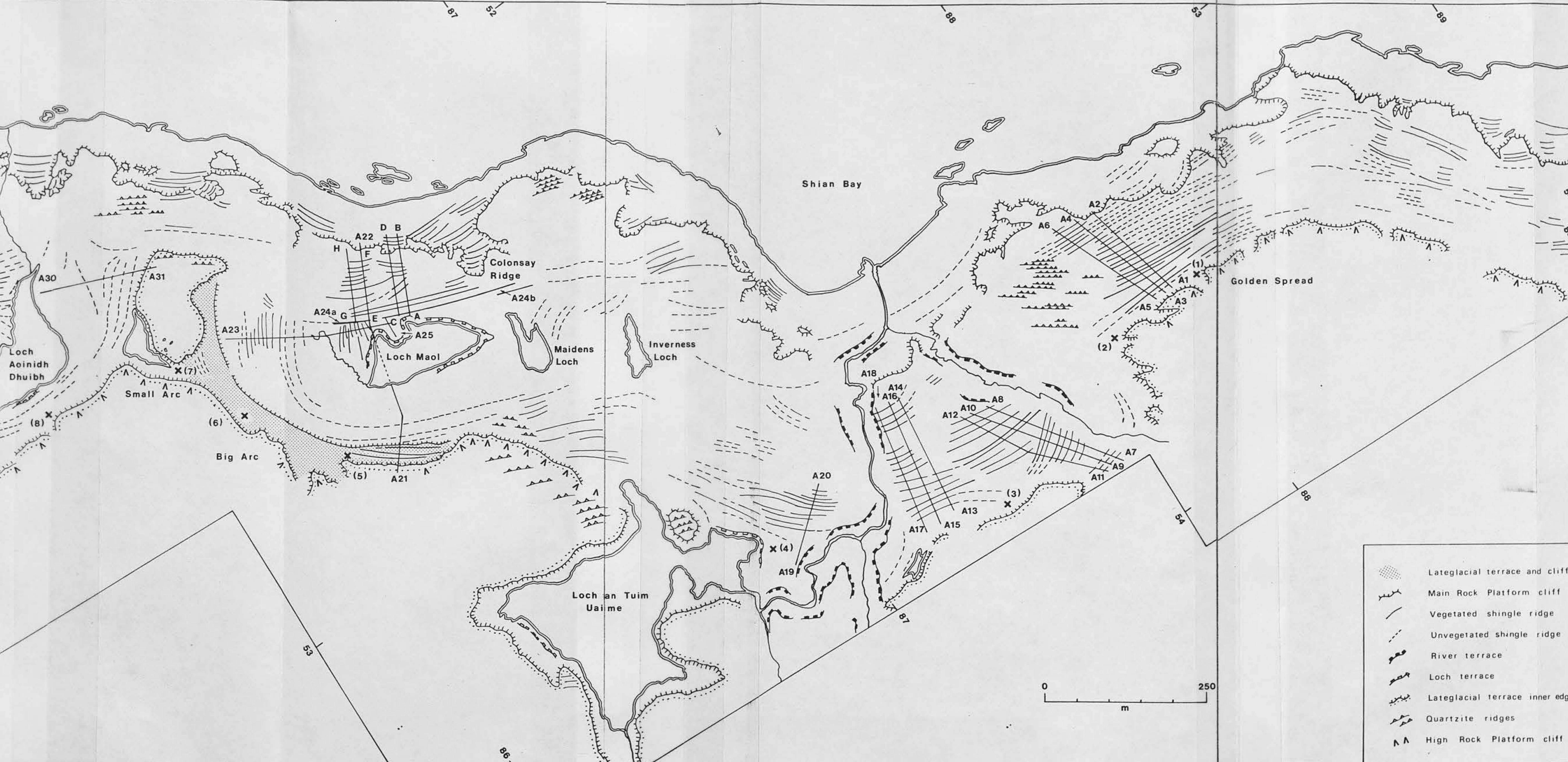


Fig. 34 Raised shorelines and measured profile locations, Ruantallain to Shian Bay, W Jura

Loch an Tuim Uaime

Loch an Tuim Uaime is impounded by a series of vegetated and unvegetated beach ridges. Landward of the highest beach ridge is a broad terrace fragment (Fig.34 (4)) the average altitude of which was determined as 35.0 m (Table 13), the same altitude as the surface of the loch.

Big Arc, S Shian Bay

In this area the highest lateglacial coastal terrace, although interrupted in places by small stream channels, forms an almost continuous feature for a distance of 600 m and is succeeded seaward by an area of undulating ridge and swale topography. Measurements of the northern and southern terrace fragments gave average altitudes of 36.3 and 36.8 m respectively (Fig.34 (5 and 6)).

Small Arc, N Aoinidh.

Small Arc represents a former sea connection between S Shian Bay and Loch Aoinidh Dhuibh and is flanked by a raised coastal terrace and cliff (Fig.34 (7)). The southern area is occupied by a loch that is impounded seaward by a raised shingle ridge. Numerous levelled peat probes in this area indicated that the altitude of the highest coastal terrace is 36.4 m.

Loch Aoinidh Dhuibh

A well-developed raised coastal terrace is almost continuously present along the inner margin of the Loch Aoinidh embayment (Fig.34; Plate 23). The terrace is separated into two parts by a stream that drains into the loch. The terrace overlies the High Rock Platform whose cliffs are here over 50 m in height (Plate 27). The terrace surface is primarily composed of sand and is succeeded seaward by a series of 55 raised beach ridges several of which impound Loch Aoinidh. The average altitudes of the northern and southern terrace fragments are respectively 35.1 m and 35.4 m(Fig.34 (8 and 9), Table 13). South of the Loch Aoinidh embayment the average altitude of a well-developed terrace fragment is 34.7 m (Fig.34 (10)).

Loch a Mhile to Ruantallain

South of Loch a Mhile a high coastal terrace is locally well-developed. Measurements of 6 terrace fragment altitudes (Fig.34 (11)) indicate an average altitude of 34.4 m (Table 13). Farther south at Ruantallain 10 measurements of a well-developed fragment indicate an average altitude of 33.9 m (Fig.34 (12)).

2. NE Islay

The altitudes of the highest lateglacial shoreline in this area have already been presented in Chapter 4. There are two main terrace fragments, at 26.4 and 26.45 m (Table 13).

3. Knockrome, SE Jura.

South-east of Lowlandmans Bay (NR 559723) several road exposures reveal horizontally-bedded sand and water-rounded cobbles that are replaced at higher altitudes by accumulations of grey till. The average upper altitude of the marine gravels was levelled as 31 m. Since the raised marine deposits are here located in an extremely sheltered location, the measured altitude of the marine gravels is considered an accurate indicator of the position of former sea-level.

4. Ardlussa, NE Jura

South of Ardlussa House (NR 648876) a raised tombolo, 50 m in width and composed of two arcuate vegetated beach ridges, is located between two protruding ridges of epidiorite. The shingle ridges face N and S respectively but do not coalesce and are separated from each other by a narrow vegetated depression. Levelling revealed that the lowest altitude of the depression is 33.6 m while adjacent vegetated ridge crests reach maximum altitudes of 34.9 and 34.0 m respectively. From these measurements it is suggested that marine action here reached a maximum altitude of 35 m while the associated high water mark, as represented by the base of the depression, was approximately 33.6 m.

5. Loch na Sgrioba, SW Jura.

Here a wide raised terrace and cliffline cut in till truncate the fossil medial moraine (Chapter 4). The terrace attains almost 200 m in width and is overlain by a series of vegetated and unvegetated raised beach ridges (Plate 5). North and south of Loch na Sgrioba the raised terrace is replaced by till cliffs. Levelling of 9 points on the two main terrace fragments indicate average altitudes of 26.9 and 26.7 m respectively (Table 13).

6. Inver, SW Jura.

North of Inver a well-developed raised coastal terrace is almost continuous for 3 km except for stream channels and ravines. The terrace occurs at distances up to 1,000 m from the coast and is separated from it by an uneven surface of gravel and bedrock overlain by thick accumulations of peat (Fig.35). Numerous levelled peat probes indicate that the peat-covered rock ridges extend N-S and vary in altitude between 14 and 22 m. Measurement of 36 altitudes on 7 raised terrace fragments indicate average fragment altitudes from north to south of 26.1, 25.4, 25.0, 24.5, 24.6, 24.1 and 24.5 m (Fig. 35, Table 13). The measurements indicate a gradual southward

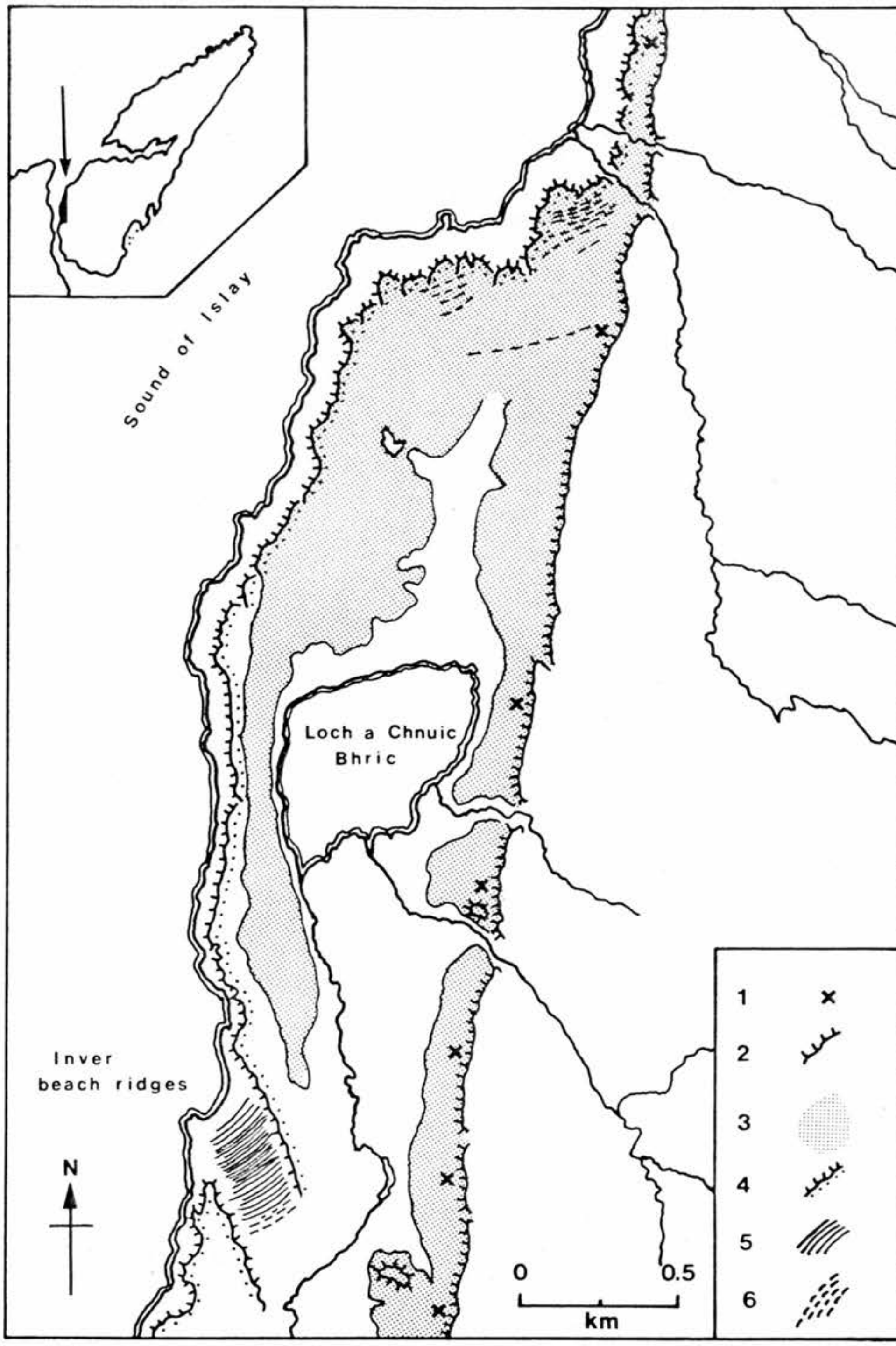


Fig. 35 Raised shoreline features, Inver, SW Jura. 1. Measured lateglacial terrace fragments 2. Inner edge of lateglacial shoreline (L2) 3. Lateglacial raised marine deposits 4 Raised cliffline 5. Inver raised beach ridges (postglacial) 6. Vegetated lateglacial raised beach ridges.

decline in terrace fragment altitude. The topography and altitude of the rock surface indicate that during the formation of the adjacent terrace the offshore rock surface was at shallow depths.

7. W Glenbatrick, SW Jura.

A high coastal terrace is locally well-developed W of Glenbatrick where it separates the backing quartzite slopes from a wide area of raised ridge and swale topography (Plate 29). Measurement of 9 altitudes on 2 well-developed terrace fragments here gave average altitudes of 29.5 and 29.9 m (Table 13).

c) Altitude Analysis

The shoreline altitudes mentioned above were plotted in a NE-SW projection plane on a shoreline height-distance diagram (Fig.36). The plotted altitudes indicate a regular decline in terrace fragment altitudes from 40 m at Corpach Bay to 33.9 m at Ruantallain. However across Loch Tarbert, between Ruantallain and Glenbatrick, there is a sharp decline in altitude from 33.9 m to 30.5 m, a drop in altitude of over 3 m in a distance of 1.3 km. Thereafter all terrace fragment altitudes measured in SW Jura decline regularly in altitude to the SW from 30.5 m to 24.1 m. In contrast the altitudes of the NE Islay fragments are markedly higher than the farthest south ones in SW Jura (26.4 m compared with 24.5 m). Since the highest shoreline forms an almost continuous feature from north of Shian Bay to Ruantallain (Plate 23) and shows a regular decline in altitude SW on the height-distance diagram, it is believed that the terrace fragments in this area were formed approximately synchronously. However, the drop in terrace fragment altitudes across Loch Tarbert between Ruantallain and Glenbatrick strongly suggests that the SW Jura terrace fragments were formed later than those between Ruantallain and Corpach Bay. Visual

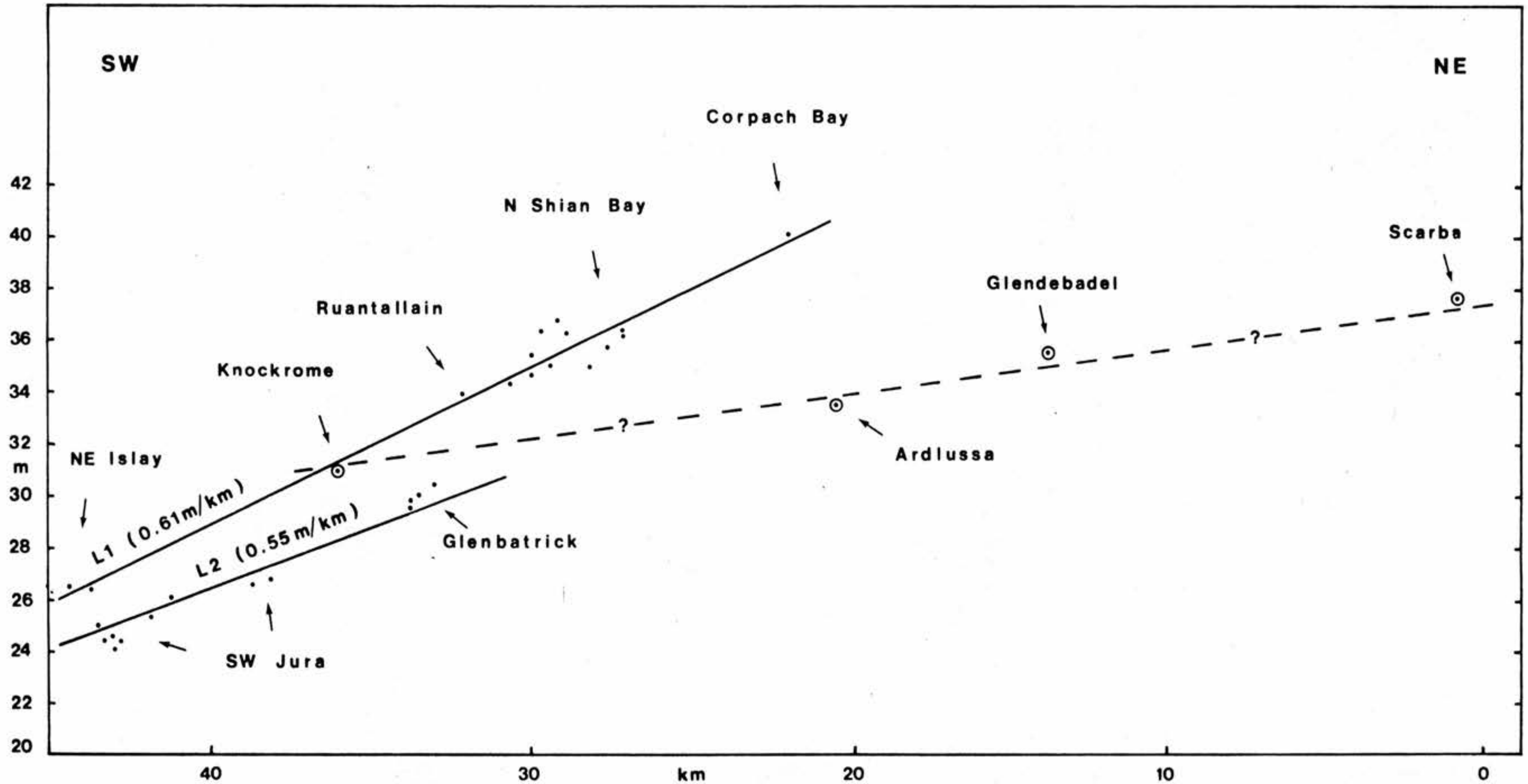


Fig. 36 Shoreline height-distance diagram for lateglacial shorelines L1 and L2. A possible third shoreline is also shown. Terrace fragments of shorelines L1 and L2 are indicated (dots uncircled).

inspection of Fig. 36 indicates that the altitudes of the NE Islay terrace fragments are aligned with those between Ruantallain and Corpach Bay and may therefore correlate with them. Additional evidence for this proposed correlation is presented later in this chapter.

In addition, with the possible exception of the Ardlussa tombolo, raised shoreline altitudes at Glendebadel, Scarba and Knockrome are also aligned in a straight line on the height-distance diagram (Fig. 36). However there is no conclusive field evidence to suggest that these fragments correlate either with each other or with any other terrace fragments already described. In particular the marked drop in altitude of the highest shoreline between Corpach Bay and Glendebadel (40.0-35.8 m) over a horizontal distance of 6 km strongly suggests that the two features were not formed synchronously.

The terrace fragment altitudes obtained between Corpach Bay and Ruantallain together with the altitude values from NE Islay were analysed by linear regression and a shoreline gradient of 0.61 m/km obtained (Fig. 36, L1). Similarly the SW Jura shoreline altitudes were analysed by linear regression, which gave a shoreline gradient of 0.55 m/km (L2).

Interpretation

The simplest explanation for the drop in terrace fragment altitudes across Loch Tarbert between Ruantallain and Glenbatrick and the corresponding rise in terrace fragment altitudes between SW Jura and NE Islay is that during deglaciation SW Jura remained ice-covered while the sea invaded the coastal areas of NE Islay and W Jura (between Ruantallain and Corpach Bay). The drop in altitude of the terrace fragments between Corpach and Glendebadel also suggests that during L1 shoreline formation between Corpach and Ruantallain the area north of and including Glendebadel remained ice-covered. Despite the presumed diachronous nature of ice decay and the associated marine incursion into deglaciated coastal areas, the shoreline evidence indicates that NE Islay and the coastline between Ruantallain and Corpach Bay were deglaciated at approximately the same time. As a result the L1 shoreline is considered to represent a generally synchronous period of shoreline formation. Owing to the steep regional gradient of the L1 shoreline the transgressing sea in W Jura was able to wash the till-covered surface of the High Rock Platform and form a drift-cut terrace above its surface. However in NE Islay the high sea-level was only able to form a coastal terrace on the seaward part of the High Rock Platform.

At a slightly later date SW Jura was deglaciated and deglaciation was accompanied by the formation of the L2 shoreline: extensive planation of till occurred seaward of the Loch na Sgrioba medial moraine (Plate 5). Also during this period a high series of beach ridges formed seaward of shoreline L1 between Ruantallain and Corpach Bay. The altitudes of the high terrace fragments at Glendebadel, Scarba and Knockrome and the tombolo depression at Ardlussa also suggest that these areas were deglaciated at later dates. Moreover the alignment of these terrace fragment altitudes on the height-distance diagram indicates the possibility that these shoreline fragments were formed at approximately the same time.

The gradients of shorelines L1 and L2 are steep. Since no other early lateglacial shorelines have been measured in western Scotland the shoreline altitudes and gradients can only be compared with the steeply sloping shorelines in E Scotland. In E Fife Cullingford and Smith (1966,p.144) identified 6 shorelines that "... formed during the early stages of the retreat of the ice preceding the Perth Readvance." Of these shorelines the most conspicuous is the Kinkell Shoreline, which slopes eastwards with a regional gradient of 0.60 m/km (Table 14). The close correspondence in shoreline gradient between the Kinkell Shoreline and the Jura L1 and L2 shorelines is in part fortuitous. However the general similarity in shoreline gradients between the two areas raises the possibility that the shorelines are of approximately the same age as the Kinkell Shoreline.

Table 14

Lateglacial shorelines in E Fife (after Cullingford and Smith, 1966)

Shoreline	Gradient (m/km)	
EF-1	1.26	un-named
EF-2	1.16	Wormistone Shoreline
EF-3	0.88	Randerston Shoreline
EF-4	0.76	Kingsbarns Shoreline
EF-5	0.72	Chesterhill Shoreline
EF-6	0.60	Kinkell Shoreline

Since the L1 and L2 shoreline gradients (0.61 m/km and 0.55 m/km) are markedly steeper than the Main Perth Shoreline (the most conspicuous lateglacial shoreline in E Scotland, with a gradient 0.43 m/km) it is likely that they were formed before it. A radio-carbon date of 16,470 ± 300 B.P. from a borehole between Colonsay and NW Jura (Harkness and Wilson, 1974,p.242, SRR-118) also suggests early deglaciation of this area. Although radio-carbon dates of this age are not completely reliable this date is in accord with the shoreline gradient evidence. The steeply sloping L1 and L2 shorelines and the radio-carbon date strongly suggest that the late-Devensian ice-sheet in W Scotland melted extremely rapidly after the supposed glacial maximum at c. 17,000 B.P. It is therefore suggested that large areas of the Scottish Inner Hebrides were deglaciated prior to the "Perth Stage" in E Scotland.

3. Lateglacial shingle spreads

a) Introduction

Raised shingle spreads are relatively uncommon coastal landforms. In a study of raised shorelines in S Sweden Hellberg (1971) noted staircases of as many as 28 raised shingle ridges that descend seaward from 56 to 17 m.m.s.l. She interpreted the deposits as lateglacial and attributed their formation to the relative marine regression that accompanied lateglacial isostatic uplift. In contrast, in the isostatically stable area of Cape Krusenstern, NW Alaska, Moore (1960) described a wide area of raised beach ridges and attributed their distribution and altitude to minor world-wide eustatic changes of sea-level. Moore, however, also conceded (1960,p.336) that "... the beach ridges were evidently formed either during storms or during periods when persistent onshore winds caused a temporary rise in sea-level." In Tasmania Davies (1961) observed staircases of raised shingle ridges and suggested (p.36) that the height of each ridge reflected wave height at the time of construction and that "... any variation in the height of the berm from ridge to ridge would be an approximate measure of sea-level change." Similar raised shingle spreads have been observed in NE Central Baffin Island (Ives and Andrews, 1963, p.33 and Fig. 18) and interpreted by Sim (1964, p.78) as emerged offshore bars.

Such studies underline the difficulty in distinguishing between a raised storm ridge and an intertidal ridge (cf.Johnson, 1919; Davies, 1961; Hellberg, 1971), each of which bears different altitude relationships to the positions of former sea-levels (Chapter 5). As a result it is often extremely difficult to decide whether the formation of large beach ridges is the result of sea-level changes or simply the result of a series of extremely severe storms. It has already been suggested (Chapter 5) that fossil beach ridges formed under relatively moderate wave activity can be identified as features that exhibit minor regional variation in crest altitude. In contrast true storm ridges are characterised by large regional ridge crest altitude variations.

The only previous study of the W Jura raised shingle spreads was by McCann (1964). For this reason it is worthwhile here to summarise his conclusions regarding the age and origin of the deposits. Specific descriptions by McCann of individual shingle spreads are included later in appropriate sections.

McCann (1964,pp.8-9) believed that the W Jura shingle ridges were formed during the general retreat of the lateglacial sea following deglaciation. He suggested that there was an initial rapid fall in relative sea-level from 100 feet (30.5 m) to 75-80 feet (22.9-24.4 m) above H.W.M. when there was a short halt. Thereafter relative sea-level fell at a slower rate and was interrupted by a second halt at 55-60 feet (16.8-18.3 m) above H.W.M. when a large shingle ridge was formed in S Shian Bay. McCann also suggested that the large volumes of shingle were derived from the lateglacial marine erosion of till deposits that formerly mantled the surface of the High Rock Platform.

In the study area approximately 1,500 altitudes were measured across areas of raised ridge and swale topography and levelled profiles constructed. Since many of the shingle spreads are morphologically complex, individual shingle areas are described separately.

b) Morphology and Altitude

1. Ruantallain - N Shian Bay

Golden Spread, N Shian Bay.

North of Shian Bay a staircase of 20 raised shingle ridges extends seaward as an uninterrupted sequence from the high lateglacial terrace to the Main Rock Platform cliff (Fig. 37). The shingle spread is bounded to the north and south by low quartzite ridges that protrude above the surface of the High Rock Platform. The highest ridge (R1) is 300 m in length and lies parallel to the high lateglacial marine terrace. The ridge crest altitude varies between 37.2 and 38.1 m along its length and reaches a maximum altitude along its central section. Three levelled profiles across all ridges and swales indicate that there is little altitude variation along the lengths of any of the lower ridges (Fig. 37, Table 15). Moreover in only four instances (cf. Fig.37; R1, R3, R12, R13) do ridge crest altitudes exceed the altitude of the adjacent landward swale by more than 0.25 m. As a result the ridge and swale topography possesses a stepped rather than an undulating surface.

Seaward of ridge R1 there is a sharp drop in altitude to the adjacent swale that lies 3.6 m below it at c.36.3-36.8 m. The lower ridge and swale profiles are essentially concave from R3 to R15 (37.1-26.2 m). Here most beach ridges are well-defined and, with the exception of R6 and R15, vary by less than 0.42 m in altitude along their lengths. Between R15 and R18 the shingle surface profile is markedly steeper where the surface declines in altitude from 26.2-22.4 m over a horizontal distance of 12 m. The shingle spreads terminate seaward at R20 where the seaward edge of the High Rock Platform is succeeded by the Main Rock Platform cliff.

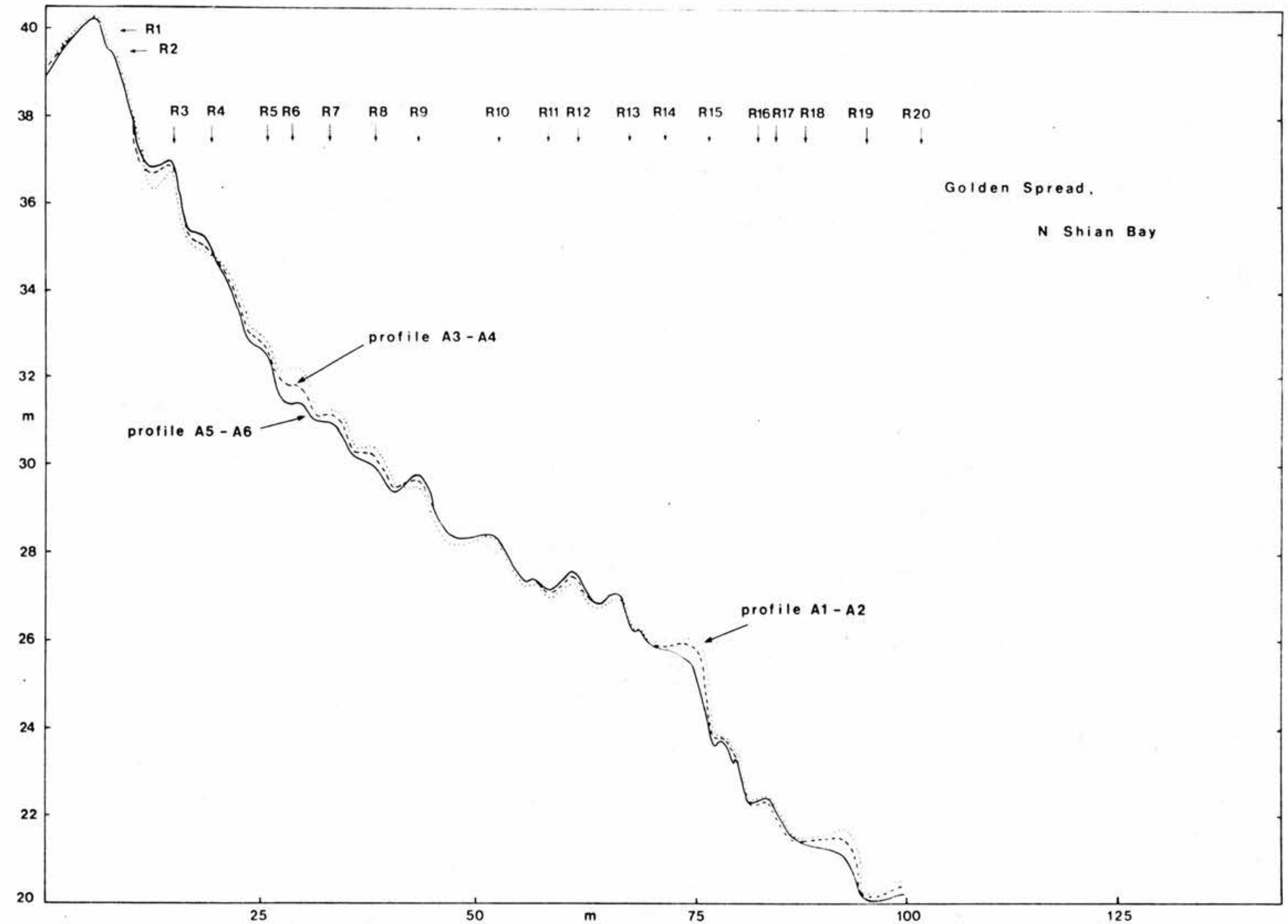


Fig. 37 Lateglacial shingle profiles, Golden Spread, N Shian Bay, W Jura. Profile locations shown on Fig. 34.

Table 15

Golden Spread, N Shian Bay, W Jura: lateglacial ridge and swale altitudes and amplitudes.

A1			A3			A5		
S	39.55	+0.94	S	39.51	+0.74	S	38.91	+1.45
R	40.49	-0.83	R	40.25	-0.74	R	40.36	-0.83
S	39.66	(-0.07)	S	39.51	+0.05	S	39.53	+0.01
R	39.59	-3.25	R	39.56	-2.85	R	39.54	-2.72
S	36.34	+0.46	S	36.71	+0.25	S	36.82	+0.25
R	36.80	-1.58	R	36.96	-1.67	R	37.07	-1.68
S	35.22	(-0.22)	S	35.29	(-0.27)	S	35.39	(-0.01)
R	35.00	-1.78	R	35.02	-1.97	R	35.38	-2.55
S	33.22	-0.08	S	33.05	-0.14	S	32.83	(-0.11)
R	33.14	-0.89	R	32.91	-1.03	R	32.72	-1.30
S	32.25	+0.10	S	31.88	-0.02	S	31.42	+0.16
R	32.35	-1.23	R	31.86	-0.81	R	31.58	-0.66
S	31.12	+0.11	S	31.05	+0.12	S	30.92	0.00
R	31.23	-0.78	R	31.17	-0.93	R	30.92	-0.80
S	30.45	(-0.04)	S	30.24	+0.05	S	30.12	(-0.11)
R	30.41	-0.96	R	30.29	-0.87	R	30.01	-0.70
S	29.45	+0.17	S	29.42	+0.21	S	29.31	+0.50
R	29.62	-1.39	R	29.63	-1.11	R	29.81	-1.33
S	28.23	+0.30	S	28.52	(-0.06)	S	28.48	(-0.05)
R	28.53	-1.25	R	28.46	-1.09	R	28.43	-1.00
S	27.28	+0.11	S	27.37	+0.18	S	27.43	+0.13
R	27.39	-0.37	R	27.55	-0.37	R	27.56	-0.33
S	27.02	+0.39	S	27.18	+0.39	S	27.23	+0.47
R	27.41	-0.48	R	27.57	-0.77	R	27.70	-0.81
S	26.93	+0.24	S	26.80	+0.44	S	26.89	+0.31
R	27.17	-0.82	R	27.24	-0.97	R	27.20	-0.92
S	26.35	+0.08	S	26.27	+0.15	S	26.28	+0.11
R	26.43	-0.32	R	26.42	-0.25	R	26.39	-0.27
S	26.11	+0.08	S	26.17	(-0.11)	S	26.12	(-0.55)
R	26.19	-2.29	R	26.06	-2.14	R	25.57	-1.85
S	23.90	+0.16	S	23.92	+0.19	S	23.72	+0.20
R	24.06	-0.75	R	24.11	-0.74	R	23.92	-0.46
S	23.31	+0.32	S	23.37	+0.10	S	23.46	(-0.27)
R	23.63	-1.26	R	23.47	-1.05	R	23.19	-0.77
S	22.37	+0.09	S	22.42	+0.06	S	22.42	+0.05
R	22.46	-0.93	R	22.48	-1.05	R	22.47	-0.96
S	21.53	+0.34	S	21.43	+0.15	S	21.51	(-0.35)
R	21.87	-1.45	R	21.58	-1.27	R	21.16	-1.03
S	20.42	+0.15	S	20.31	+0.14	S	20.13	+0.04
R	20.57		R	20.45		R	20.17	
A2			A4			A6		

* shingle profiles shown on Fig. 37

* profile locations shown on Fig. 34

Shian Embayment

South of the Golden Spread and north of the Shian river two suites of arcuate unvegetated ridges and swales are separated by peat accumulations (Figs. 38 and 39). In the northern area a sequence of 14 beach ridges descends seaward from the high coastal terrace to c. 24 m (Table 16). The shingle profiles are concave while each ridge varies by less than 0.4 m in altitude along its length. The most marked discontinuity in the shingle profiles occurs between R4 and R6 where seaward of the R4 ridge there is a sharp fall in altitude from 34.7 to 32.2 m. Seaward of R6 the pattern of undulating ridge and swale topography is continued as far as R10 (30 m). Seaward of R10 (below 30 m) there are no well-defined ridges or swales. Instead the shingle surface slopes relatively steeply seaward from 30 m to 22.7 m. It is noteworthy that the highest beach ridge of the Golden Spread has no apparent equivalent in this area, the crest of the highest beach ridge in C Shian Bay occurring at c.36 m.

The southern area of this shingle spread, although forming part of the above accumulation (Fig. 34), exhibits several different ridge altitude characteristics (Fig. 39). The higher part of the shingle surface is characterised by three pronounced raised beach ridges (R2, R3 and R4) whose crest altitudes lie between 34.5 and 35.5 m. Above these ridges the highest ridge (R1) reaches 36.5 m in altitude while farther landward the high lateglacial marine terrace occurs at 36.3 m. Seaward of the R6 ridge the shingle surface slopes steeply seaward and is characterised by a staircase of low amplitude ridges and swales. With the exception of the R6 ridge, all ridges vary in altitude by less than 0.36 m along their lengths (Table 16).

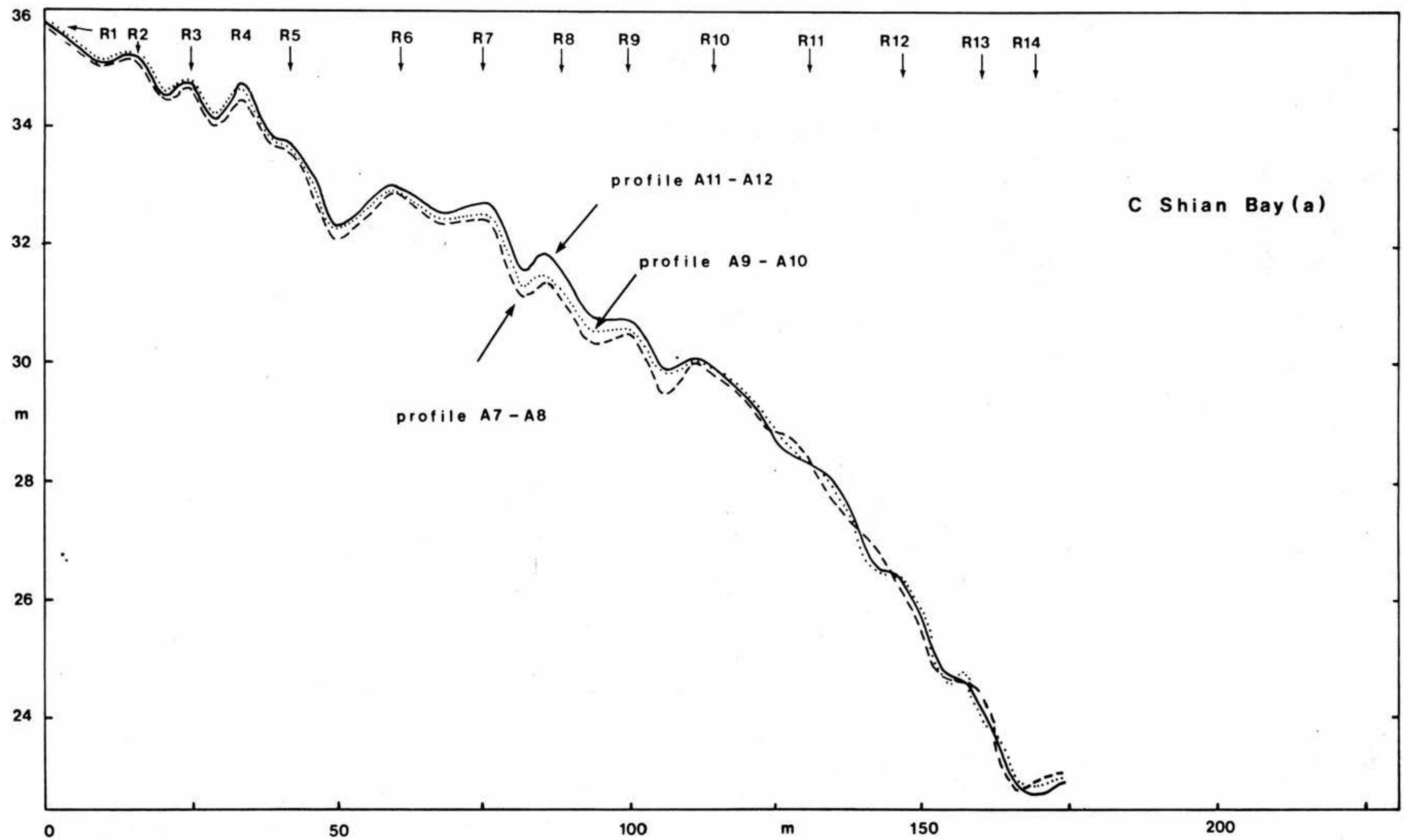


Fig. 38 Lateglacial shingle profiles, C Shian Bay, W Jura. Profile locations shown on Fig. 34.

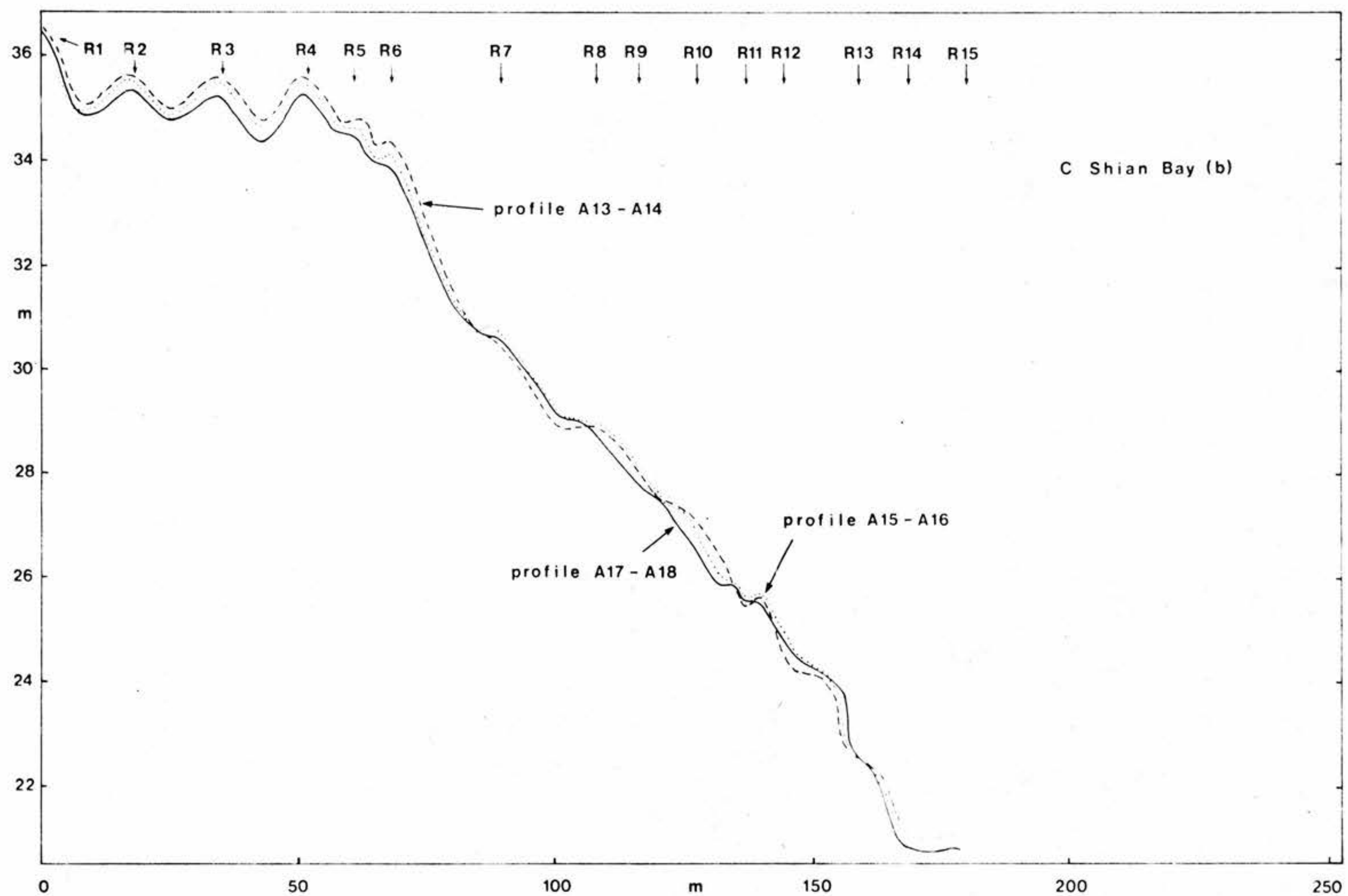


Fig. 39 Lateglacial shingle profiles, C Shian Bay, W Jura. Profile locations shown on Fig.34.

Table 16

Shian Bay, W Jura: lateglacial ridge and swale altitudes and amplitudes.

A7

R	34.64	-0.65
S	33.99	+0.39
R	34.38	-0.54
S	33.84	(-0.50)
R	33.34	-1.31
S	32.03	+0.87
R	32.90	-0.53
S	32.37	+0.06
R	32.43	-1.28
S	31.15	+0.34
R	31.49	-1.13
S	30.36	+0.13
R	30.49	-1.08
S	29.41	+0.59
R	30.00	-1.31
S	28.69	(-0.81)
R	27.88	-1.03
S	26.85	(-0.33)
R	26.52	-1.80
S	24.72	(-0.14)
R	24.58	-1.88
S	22.70	0.00
R	22.70	

A8

A9

R	35.81	-0.74
S	35.07	+0.13
R	35.20	-0.65
S	34.55	+0.20
R	34.75	-0.69
S	34.06	+0.38
R	34.44	-0.77
S	33.67	0.00
R	33.67	-1.36
S	32.31	+0.59
R	32.90	-0.49
S	32.41	+0.10
R	32.51	-1.25
S	31.26	+0.21
R	31.47	-0.96
S	30.51	+0.08
R	30.59	-0.67
S	29.92	+0.20
R	30.12	-1.31
S	28.81	(-0.76)
R	28.05	-1.45
S	26.60	(-0.06)
R	26.54	-1.85
S	24.69	+0.09
R	24.78	-1.86
S	22.92	(-0.24)
R	22.68	

A10

A11

R	35.73	-0.67
S	35.07	+0.13
R	35.20	-0.76
S	34.44	+0.32
R	34.76	-0.65
S	34.11	+0.65
R	34.76	-0.97
S	33.79	(-0.09)
R	33.70	-1.42
S	32.28	+0.72
R	33.00	-0.45
S	32.55	+0.19
R	32.74	-1.17
S	31.57	+0.29
R	31.86	-1.09
S	30.77	(-0.09)
R	30.68	-0.80
S	29.88	+0.20
R	30.08	-1.53
S	28.55	(-0.36)
R	28.19	-1.76
S	26.43	+0.04
R	26.47	-1.69
S	24.78	(-0.11)
R	24.67	-1.77
S	22.90	(-0.25)
R	22.65	

A12

* shingle profiles shown on Fig. 38

* profile locations shown on Fig. 34

Table 16 (contd.)

Shian Bay, W Jura: lateglacial ridge and swale altitudes and amplitudes.

A13

R	36.33	-1.36
S	34.97	+0.59
R	35.56	-0.96
S	34.60	+0.98
R	35.58	-0.84
S	34.74	+0.02
R	34.76	-0.58
S	34.18	+0.32
R	34.50	-3.63
S	30.87	+0.05
R	30.92	-2.02
S	28.90	+0.06
R	28.96	-0.30
S	28.66	(-0.07)
R	28.59	-1.11
S	27.48	(-0.09)
R	27.39	-0.95
S	26.44	(-0.54)
R	25.90	=0.30
S	25.60	(-0.11)
R	25.49	-0.83
S	24.66	(-0.33)
R	24.33	-1.61
S	22.72	(-0.16)
R	22.56	-1.71
S	20.85	(-0.08)
R	20.77	veg.
S	18.93	-0.25
R	18.68	+0.73
S	19.41	-0.74
R	18.67	+0.88
S	19.55	-1.65
R	17.90	

A14

A15

R	36.39	-1.43
S	34.96	+0.55
R	35.51	-0.69
S	34.82	+0.70
h	35.52	-0.86
S	34.66	(-0.10)
R	34.56	-0.53
S	34.03	+0.17
R	34.20	-3.46
S	30.74	+0.10
R	30.84	-1.95
S	28.89	+0.04
R	28.93	veg.
	veg.	veg.
R	28.54	-1.04
S	27.50	(-0.07)
R	27.43	-1.35
S	26.08	(-0.25)
R	25.83	-0.28
S	25.55	+0.01
R	25.56	-0.92
S	24.64	(-0.21)
R	24.43	-1.75
S	22.68	(-0.14)
R	22.54	-1.69
S	20.85	(-0.12)
R	20.73	veg.
R	19.03	-0.15
S	18.88	+0.54
R	19.42	-0.58
S	18.84	+0.71
R	19.55	-1.76
S	17.79	

A16

A17

R	36.52	(-1.69)
S	34.83	+0.54
R	35.37	-0.61
S	34.76	+0.44
R	35.20	-0.83
S	34.37	+0.89
R	35.26	-0.69
S	34.57	(-0.12)
R	34.45	-0.46
S	33.99	(-0.20)
R	33.79	-2.99
S	30.80	(-0.15)
R	30.65	-1.60
S	29.05	(-0.08)
R	28.97	veg.
	veg.	veg.
R	28.18	-0.73
S	27.45	(-0.37)
R	27.08	-1.25
S	25.83	+0.05
R	25.88	-0.32
S	25.56	(-0.04)
R	25.52	-0.89
S	24.63	(-0.19)
R	24.44	-1.81
S	22.63	(-0.05)
R	22.58	veg.
	veg.	veg.
	veg.	veg.
R	19.04	-0.12
S	18.92	+0.68
R	19.60	-0.90
S	18.70	+0.88
R	19.58	-2.00
S	17.58	

A18

* shingle profiles shown on Fig. 39

* profile locations shown on Fig. 34.

Immediately south of the Shian river a separate staircase of 10 ridges forms the southern extension of the two northern spreads (described above). The unvegetated shingle surface descends from 35.2 to 27.8 m and is characterised by three well-developed high ridges (Fig. 40 (R1, R2 and R3)) whose crest altitudes are at 35.2, 34.7 and 32.9 m respectively (Table 16). Below 32.9 m the shingle surface declines gently in altitude and is uninterrupted by any high amplitude ridges or swales. The highest ridges partially impound Loch an Tuim Uaime (surface 34.9 m).

S Shian Bay

The S Shian embayment contains a complex pattern of raised ridge and swale topography, raised coastal terraces and lochans impounded by raised coastal deposits (Fig. 34). McCann suggested that there was evidence here of pauses or oscillations of relative sea-level that interrupted the general retreat of the lateglacial sea. McCann (1964,p.14) noted that there is,

> "... a change from a relatively steep to a more gentle seaward slope in the surface of the shingle deposits at 75-80 feet (22.9-24.4 m) above H.W.M. The steeper upper slope reflects a rapid fall in sea-level from the maximum to about 75-80 feet, when there was a short halt, followed by a second fall of sea-level at a slower rate which is reflected in the more gentle lower slope. The second feature is the division of the shingle deposits at the southern end of the Shian embayment of the Inter-glacial platform into two distinct parts, separated by the wide hollows occupied by small lochs. The upper shingle deposit shows the marked break of slope, while the lower part, damming the lochs is a quite separate shingle bank at a lower level of 55-63 feet (16.8-19.2 m) above H.W.M. The development of this wide lower ridge at that locality must be due, in part, to the fact that the outer part of the old marine platform there is very low, but it may also be due to a second halt in the fall of relative sea-level at 55-60 feet (16.8-18.3 m)."

McCann considered that further evidence for this pause in sea-level retreat occurs in neighbouring Colonsay and that the raised beaches

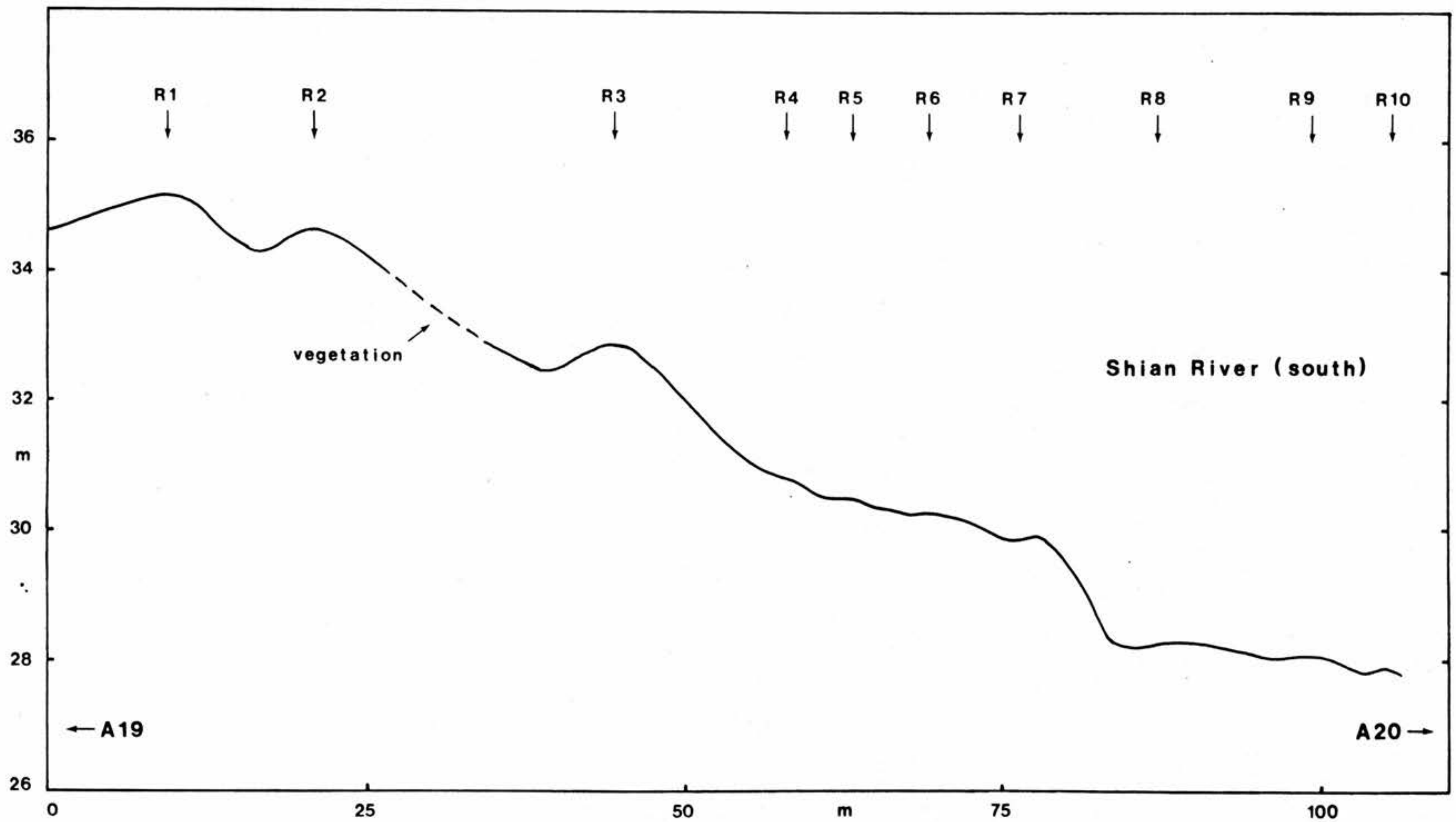


Fig. 40 Lateglacial shingle profile, Shian River, W Jura. Profile location shown on Fig. 34.

at 75 feet above H.W.M. in Mull correlate with the raised beaches at similar altitudes in Jura and Islay.

The raised shingle deposits and impounded lochans are located on the High Rock Platform surface that here occurs at exceptionally low altitudes. In the southern and eastern parts of the embayment the highest marine deposits form a conspicuous arcuate terrace banked against the inner edge of the High Rock Platform (Plate 25). In the south of the embayment the terrace trends westwards away from the inner edge of the high platform cliffline and is flanked by a series of peat-covered rock skerries that protrude above the High Rock Platform surface. The northern end of the terrace surface forms an undulating belt of ridges and swales that is replaced southwards by a flat vegetated surface. The continuity of the flat terrace surface is interrupted at Small Arc by three vegetated beach ridges that mantle the terrace. Throughout this area the inner edge of the high terrace is well-defined and is succeeded landward by the High Rock Platform cliff (Fig.34; Plate 23).

The frontal slope of the raised terrace is mantled by two vegetated beach ridges (Figs. 34 and 41; Plate 25). The higher ridge (R3) reaches a crest altitude of 30.7 m and is succeeded seaward by a lower arcuate vegetated ridge (R9), 600 m in length, that parallels the high terrace along its length. In the centre of the S Shian embayment the ridge crest is at 24.6 m and it rises in altitude westwards to reach a maximum altitude of 26.4 m the seaward base of the beach ridge (R9) being at 20.2 m. In the south of the embayment the ridge (R9) is concave to the north while above it a series of 8 parallel ridges rises in altitude to the high coastal terrace (Figs. 34 and 42). Below ridge R9, 13 unvegetated ridges descend

Plate 25. Raised terrace of lateglacial marine deposits, Shian Bay, W Jura. An arcuate vegetated beach ridge (R9) mantles the lower slope.

Plate 26. Raised shingle ridges, Bagh Righ Mhor, east of Ruantallain, W Jura.

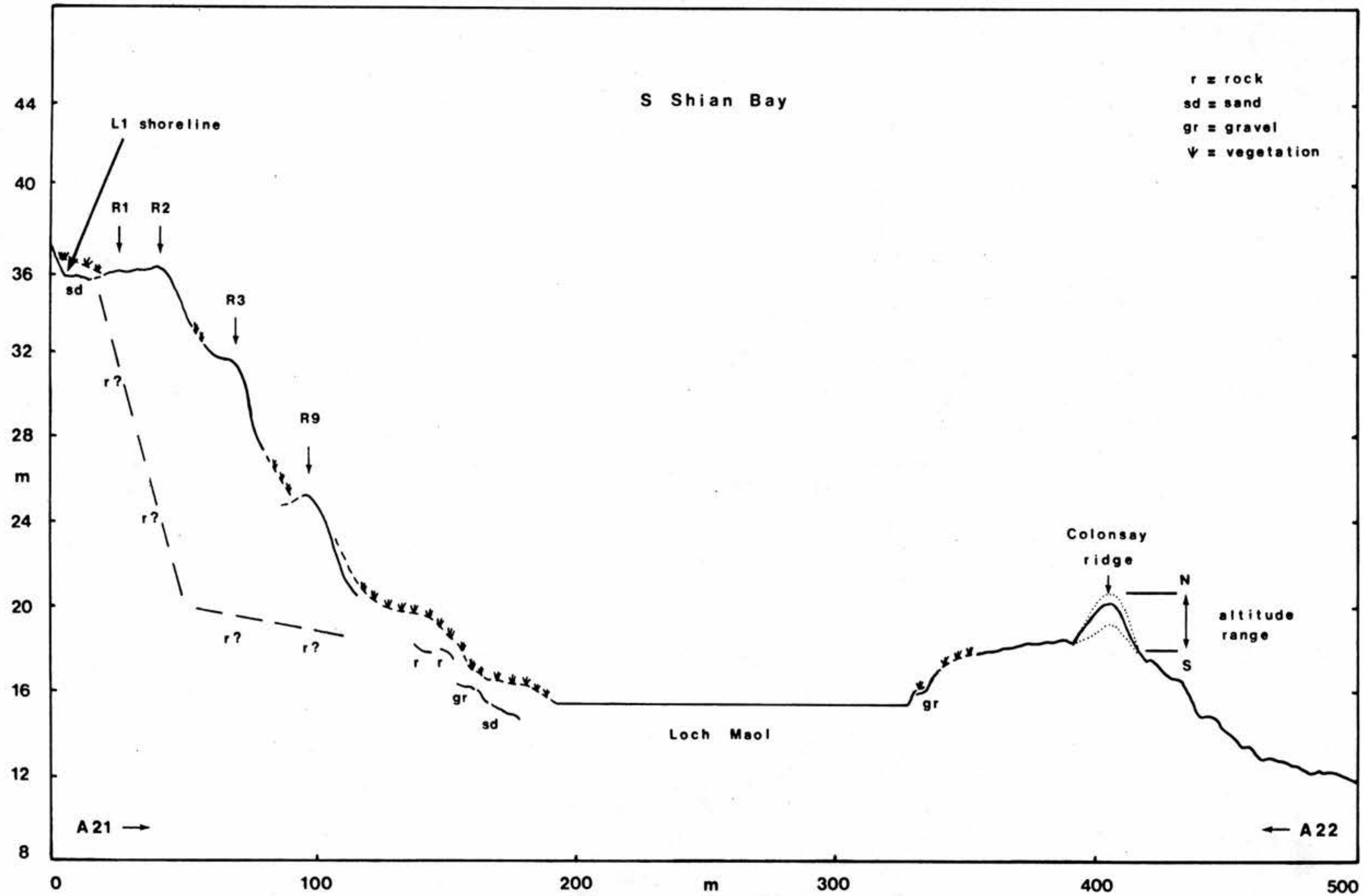


Fig.41 Lateglacial shingle profile,across S Shian Bay, W Jura. Profile location shown on Fig.34.

parallel to each other from 26.4 m to 17.5 m (Table 17) where they form the western shoreline of Loch Maol (surface 15.3 m). Loch Maol is impounded by a large N-S trending shingle ridge (the Colonsay ridge), 480 m in length, that truncates the western margins of ridges R10-R22 in the south of the embayment. The southern end of the Colonsay ridge is adjacent to R11 at c.19 m. Above this altitude the north-facing ridges (R1-R10) are oriented normal to the Colonsay ridge.

The northern end of the Colonsay ridge reaches a maximum altitude of 20.3 m and is composed of coarse sub-rounded debris with an average diameter of 0.27 m. The Colonsay ridge is succeeded landward by two lochs (Maidens Loch and Inverness Loch) that are impounded by the shingle ridge and occur at altitudes of 17.8 and 20.1 m respectively. Altitude measurements at 30 m intervals along the ridge crest indicate a progressive southward decline in ridge altitude (Fig. 42; Table 18) that is accompanied by a gradual decrease in cobble size.

Peat probing E of Loch Maol shows that ridge R9 is separated from the Loch by accumulations of sand that are underlain by an uneven bedrock surface (Fig. 41). Seaward of the Colonsay ridge a continuous staircase of unvegetated ridges descends seaward to the modern storm beach whose crest altitude is at 7.7 m (the highest modern storm ridge in W Jura).

Loch Aoinidh Dhuibh

Within the Loch Aoinidh Dhuibh embayment there exists the most complex sequence of lateglacial coastal forms in Britain (Plate 23, Figs. 34 and 43). In total 55 beach ridges separated by swales extend seaward from the lateglacial coastal terrace to the top of the Main Rock

Table 17

Lateglacial shingle profile, South Shian Bay, W Jura

A23

S	28.93	
R	28.97	
S	28.47	
R	28.51	
S	27.61	
R	27.65	
S	27.21	
R	27.18	
S	26.65	
R	26.47	
S	26.42	
R	26.42	
S	25.34	
R	25.53	
S	24.18	
R	24.47	
S		
R	25.43	
S	23.12	25.31
R	23.00	
S	22.13	25.98
R	19.85	
S		26.44
R	19.17	
S	18.51	
R	18.70	
S	18.57	Colonsay ridge crest altitude (S) = 18.98 m
R	18.59	
S	18.52	
R	18.56	
S	18.44	
R	18.47	
S	18.32	
R	18.82	
S	17.45	
R	18.01	
S	18.27	
R	18.40	
S	18.21	
R	18.18	
S	18.14	
R	18.22	
S	18.02	
R	18.10	

A25

* elevation of R9 in centre of South Shian Bay = 24.6 m
* elevation of Loch a Maol (3.7.77)= 15.26 m
* elevation of Maiden's Loch (3.7.77) = 17.82 m
* elevation of Inverness Loch (3.7.77) = 20.09 m
* **shingle profile shown on Fig. 42**
* **profile location shown on Fig. 34**

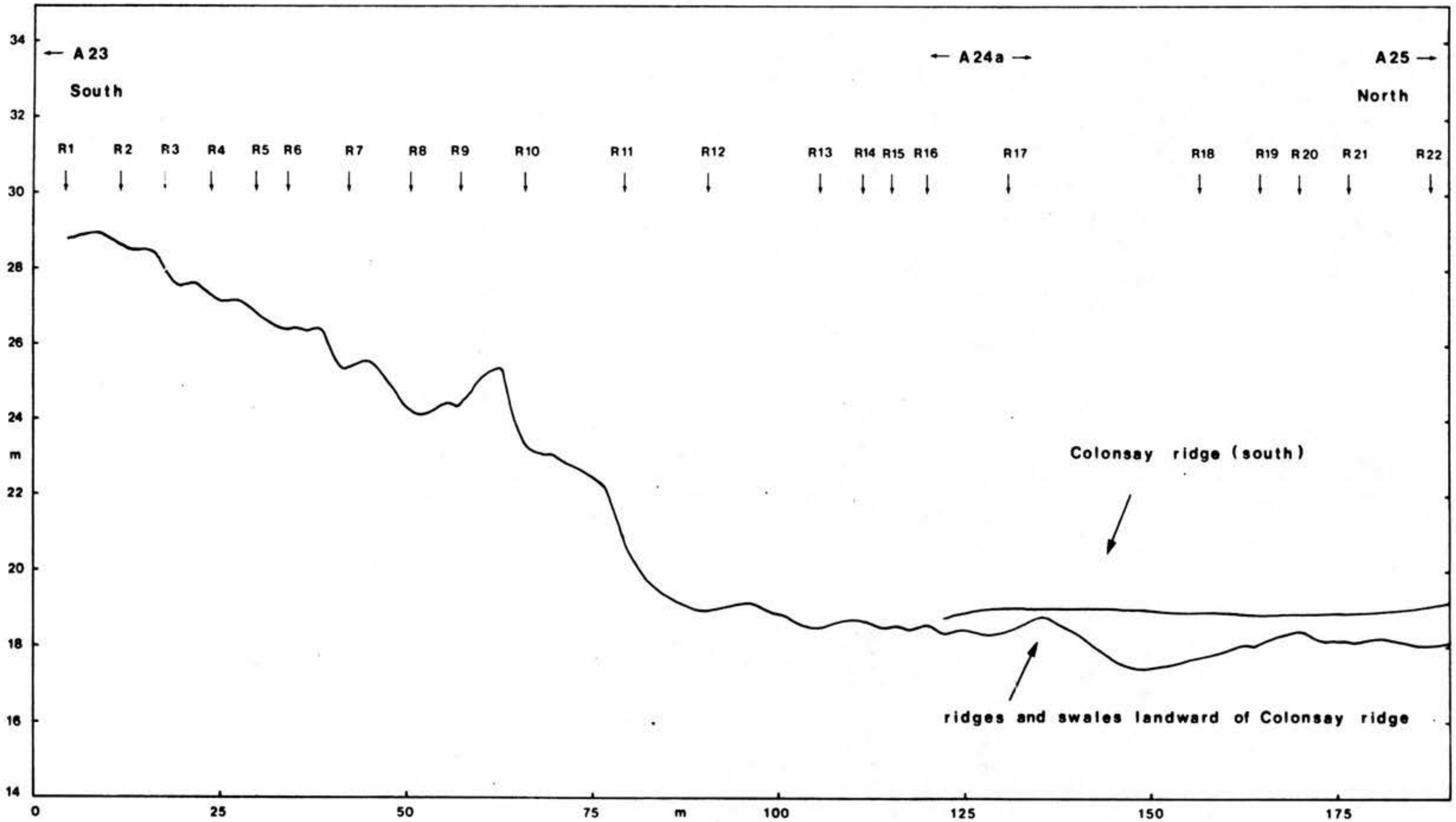


Fig. 42 Lateglacial shingle profile at southern end of Colonsay Ridge, Shian Bay, W Jura. Profile location shown on Fig. 34.

Table 18

South Shian Bay, W Jura: Colonsay Ridge crest altitudes.

A24a	Distance (m)
18.92	0
19.31	30
18.87	60
18.99	90
19.42	120
18.93	150
18.89	180
19.48	210
19.59	240
20.02	270
20.25	300
20.46	330
20.53	360
20.55	390
20.50	420
20.32	450
20.32	480
A24b	

[altitude of ridge crest stated by McCann (1964,p.12) as 55-63 feet above H.W.M. (16.8-19.2 m]

* profile location shown on Fig. 34

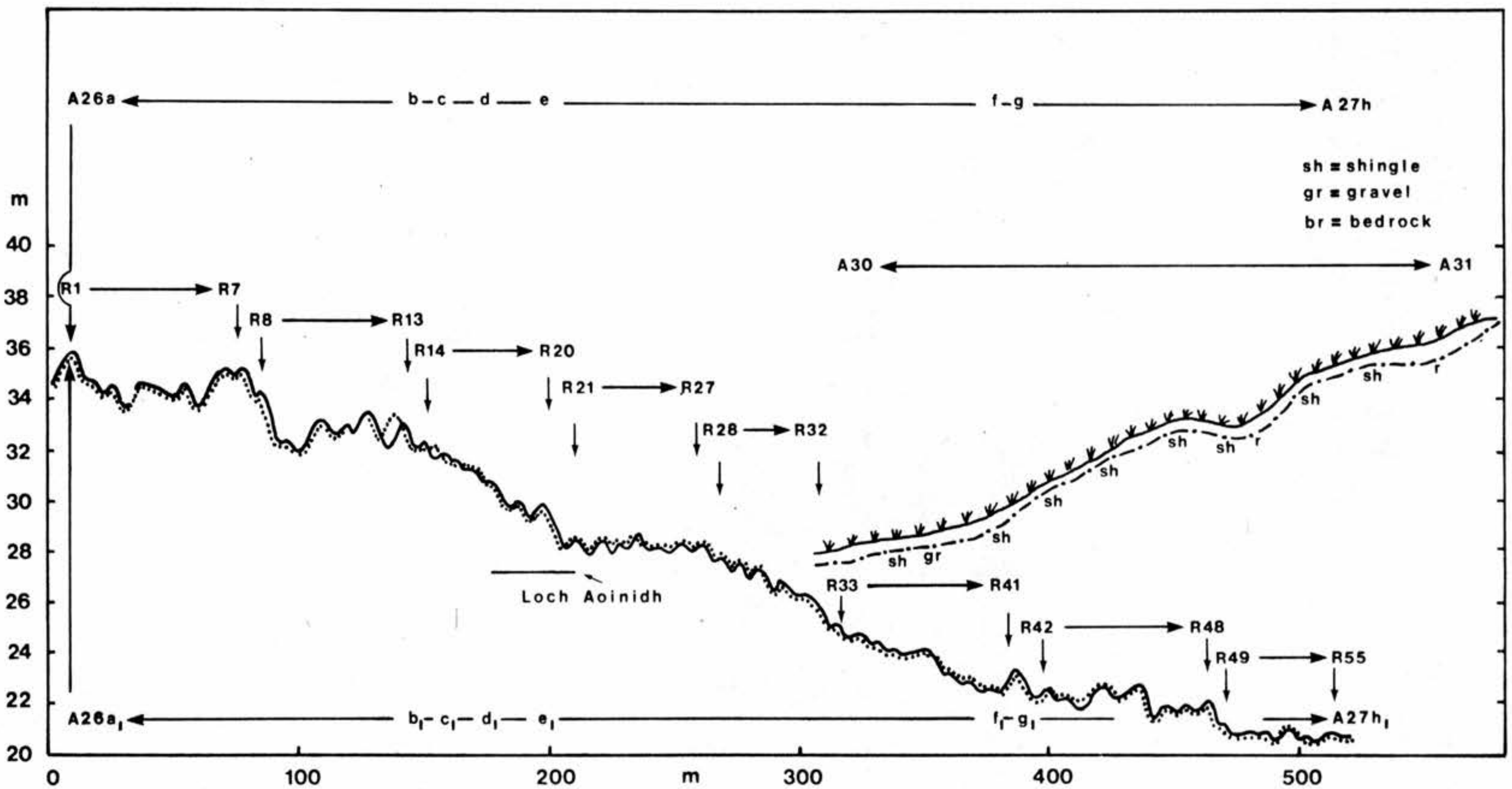


Fig. 43 Lateglacial shingle profiles, Loch Aoinidh Dhuibh, W Jura. Profile locations are shown on Fig.34.

Platform cliff. The raised shingle ridges overlie the W Jura High Rock Platform which in this area reaches a maximum width of 650 m. Loch Aoinidh is succeeded landwards by the High Rock Platform cliff, which here reaches heights of 50-70 m. North and south of the embayment two thick peat bogs overlie a series of inclined quartzite ridges that protrude above the generally even surface of the platform. Despite their low relief the rock ridges formerly exerted an important control on the development of the adjacent beach ridges. Within the embayment there is a complicated pattern of raised beach ridges that (McCann, 1964, p.12) "...can be traced not only across the bare shingle areas but also beneath a cover of vegetation." McCann considered that the distribution of beach ridges was related to the falling level of the lateglacial sea and that,

> "...During the maximum period of submergence with sea-level falling from 115-100 feet (35.1-30.5 m) above present H.W.M., beach ridges were formed around the margins of the embayment, and the separation into two areas, one under the influence of refracted waves from the SW and the other under the direct influence of waves from this direction, is apparent. With the continued fall of sea-level from 100-85 feet (30.5-25.9 m) newer and lower beach ridges were deposited on the northern side of the bay parallel to the older ridges, but in the southern part the alignment of the ridges changed and became progressively more nearly north-south, for, as the sea-level fell and the sea occupied less of the embayment, the effect of the headland in refracting the dominant waves became less important... the last stage in (beach ridge) development was the deposition of 18 curved, parallel ridges between 85 and 56 feet (25.9-17.1 m) above H.W.M. Within this embayment... there is a gradual fall in height from the highest shingle ridge to the lowest."

In the inner part of the embayment the highest beach ridges form sweeping arcuate features that lie approximately parallel to the high lateglacial terrace (Fig. 34). The ridges exhibit little altitude variation along their lengths and in only one instance does the lateral altitude variation of an individual ridge exceed 0.3 m (Table 19, Figs. 34 and 43). There is also no noticeable change in

Plate 27. Loch Áoinidh Dhuibh, W Jura. In middle distance is High Rock Platform cliff.

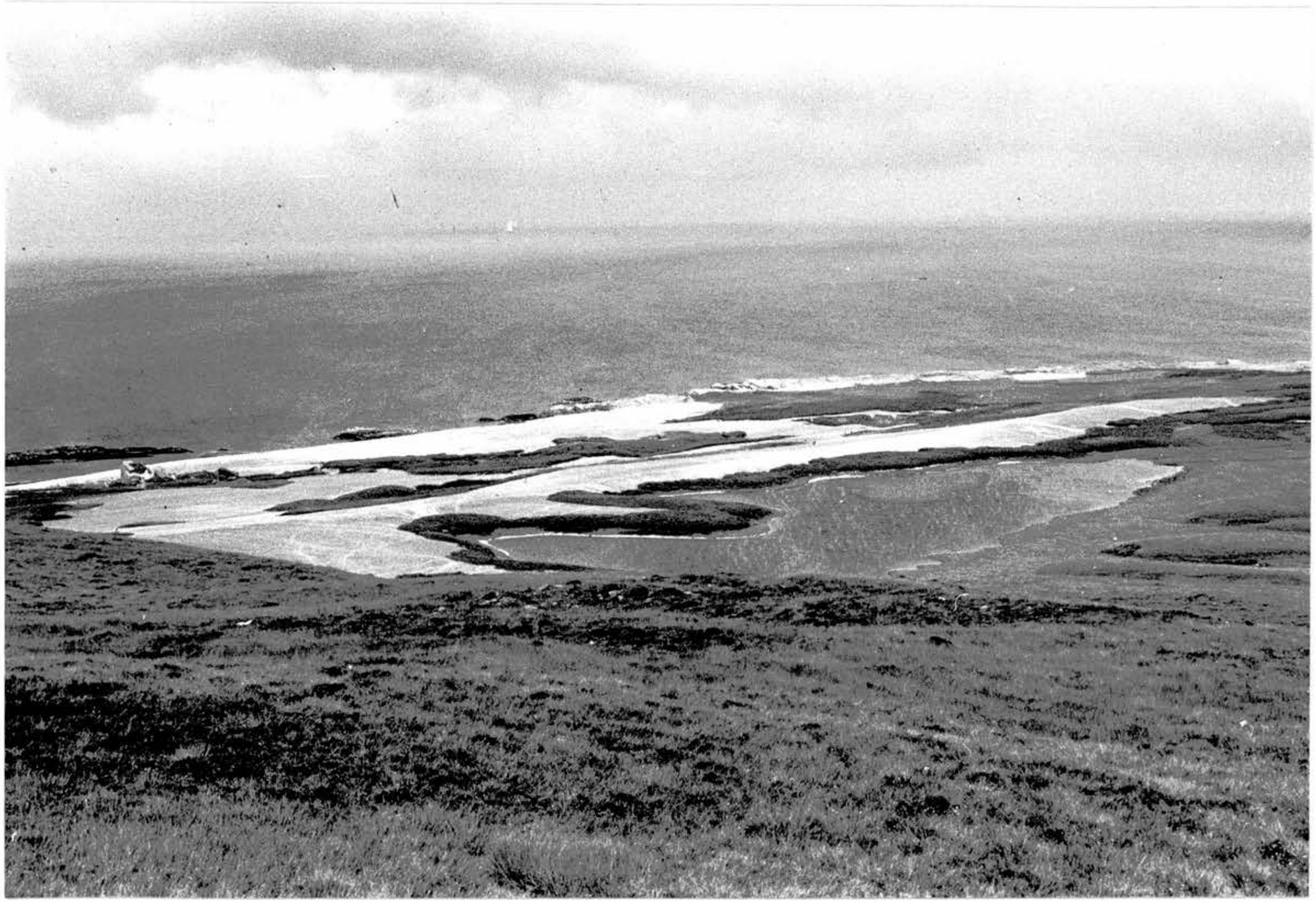


Plate 28. Colonsay ridge, Shian Bay, W Jura. The raised shingle ridge impounds Lochan Maol (foreground).

Table 19

Loch Aoinidh Dhuibh, W Jura: ridge and swale altitudes and amplitudes.

A26a	R	36.04	-1.53	A28a	R	35.83	-1.33
	S	34.51	+1.27		S	34.50	+1.45
	R	35.78	-1.00		R	35.95	-1.24
	S	34.78	(-0.03)		S	34.71	(-0.02)
	R	34.75	-0.73		R	34.69	-0.44
	S	34.02	+0.52		S	34.25	+0.41
	R	34.54	-0.93		R	34.66	-0.99
	S	33.61	+1.08		S	33.67	+1.09
	R	34.69	-0.46		R	34.76	-0.46
	S	34.23	+0.27		S	34.30	+0.06
	R	34.50	-0.41		R	34.36	-0.18
	S	34.09	+0.59		S	34.18	+0.47
	R	34.68	-0.97		R	34.65	-0.97
	S	33.71	+1.56		S	33.68	+1.55
	R	35.27	-0.36		R	35.23	-0.34
	S	34.91	+0.11		S	34.89	+0.32
	R	35.02	-0.90		R	35.21	-1.00
	S	34.12	+0.14		S	34.21	+0.03
	R	34.26	-0.97		R	34.24	-1.94
	S	32.29	(-0.10)		S	32.30	(-0.09)
	R	32.19	-0.29		R	32.21	-0.31
	S	31.90	+1.17		S	31.90	+1.29
	R	33.07	-0.50		R	33.19	-0.52
	S	32.57	+0.42		S	32.67	+0.27
	R	32.99	-0.20		R	32.94	-0.25
	S	32.79	+0.66		S	32.69	+0.75
b	R	33.45	-1.42	b,	R	33.44	-1.20
c	S	32.03	+1.05	c,	S	32.24	+0.74
	R	33.08	-0.94		R	32.98	-0.77
	S	32.14	+0.09		S	32.21	+0.03
	R	32.23	-0.61		R	32.24	-0.52
	S	31.62	+0.25		S	31.72	+0.09
	R	31.87	-0.33		R	31.81	-0.29
	S	31.54	(-0.05)		S	31.52	(-0.08)
	R	31.49	-0.30		R	31.44	-0.17
	S	31.19	(-0.01)		S	31.27	(-0.02)
	R	31.18	-0.40		R	31.25	-0.51
	S	30.78	(-0.07)		S	30.74	(-0.04)
d	R	30.71	-0.96	d,	R	30.70	-0.94
e	S	29.75	+0.30	e,	S	29.76	+0.32
	R	30.05	-0.77		R	30.08	-1.01
	S	29.28	+0.53		S	29.07	+0.56
	R	29.81	-1.66		R	29.63	-1.43
	S	28.15	+0.32		S	28.20	+0.20
	R	28.47	-0.65		R	28.40	-0.41
	S	27.82	+0.53		S	27.99	+0.46
	R	28.35	-0.18		R	28.45	-0.48
	S	28.17	+0.14		S	27.97	+0.34
	R	28.31	-0.24		R	28.31	-0.20
	S	28.07	+0.61		S	28.11	+0.27
	R	28.68	-0.56		R	28.38	-0.21
	S	28.12	+0.01		S	28.17	(-0.02)
	R	28.13			R	28.15	

* surface altitude of Loch Aoinidh Dhuibh (4.7.77) = 29.16 m

* profile locations shown on Fig.34

(contd. over)

Table 19 (contd.)

Loch Aoinidh Dhuibh, W Jura: ridge and swale altitudes and amplitudes.

	S	27.92	
			+0.30
	R	28.22	
			-0.25
	S	27.97	
			+0.20
	R	28.17	
			-0.62
	S	27.55	
			+0.16
	R	27.71	
			-0.60
	S	27.11	
			+0.39
	R	27.50	
			-0.63
	S	26.87	
			+0.45
	R	27.32	
			-0.82
	S	26.50	
			+0.35
	R	26.85	
			-0.63
	S	26.22	
			+0.12
	R	26.34	
			-0.25
	S	25.09	
			+0.11
	R	25.20	
			-0.58
	S	24.62	
			+0.31
	R	24.93	
			-0.44
	S	24.49	
			+0.04
f	R	24.53	
			-0.38
g	S	24.15	
			+0.07
	R	24.22	
			-0.14
	S	24.08	
			+0.22
	R	24.30	
			-0.98
	S	23.32	
			+0.11
	R	23.43	
			-0.48
	S	22.95	
			+0.10
	R	23.05	
			-0.29
	S	22.76	
			+0.26
	R	23.02	
			-0.61
	S	22.41	
			+0.85
	R	23.26	
			-1.13
	S	22.13	
			+0.59
	R	22.72	
			-0.40
	S	22.32	
			+0.03
	R	22.35	
			-0.30
	S	22.05	
			+0.85
	R	22.90	
			-0.64
	S	22.26	
			+0.51
	R	22.77	
			-1.27
	S	21.50	
			+0.53
	R	22.03	
			-0.36
	S	21.67	
			+0.14
	R	21.81	
			-0.01
	S	21.80	
			+0.42
	R	22.22	
			-0.94
	S	21.28	
			+0.04
	R	21.32	
			-0.55
	S	20.77	
			+0.45
	R	21.22	
			-0.21
	S	21.01	
			+0.16
	R	21.17	
			-0.52
	S	20.65	
			+0.74
	R	21.39	
			-0.63
	S	20.76	
			(-0.02)
	R	20.74	
			-0.01
A27h	S	20.73	

	S	27.92	
			+0.33
	R	28.30	
			-0.33
	S	27.97	
			+0.31
	R	28.28	
			-0.59
	S	27.69	
			+0.06
	R	27.75	
			-0.62
	S	27.13	
			+0.43
	R	27.56	
			-0.49
	S	27.07	
			+0.29
	R	27.36	
			-1.00
	S	26.36	
			+0.54
	R	26.90	
			-0.58
	S	26.32	
			(-0.05)
	R	26.27	
			-1.18
	S	25.09	
			+0.12
	R	25.21	
			-0.42
	S	24.79	
			+0.07
	R	24.86	
			-0.42
	S	24.44	
			(-0.05)
f,	R	24.39	
			-0.19
g,	S	24.20	
			(-0.04)
	R	24.16	
			-0.22
	S	23.94	
			+0.30
	R	24.24	
			-0.93
	S	23.31	
			+0.05
	R	23.36	
			-0.47
	S	22.89	
			+0.15
	R	23.04	
			-0.32
	S	22.72	
			+0.13
	R	22.85	
			-0.31
	S	22.54	
			+0.92
	R	23.46	
			-1.22
	S	22.24	
			+0.43
	R	22.67	
			-0.43
	S	22.24	
			+0.03
	R	22.27	
			-0.42
	S	21.85	
			+0.93
	R	22.78	
			-0.47
	S	22.31	
			+0.58
	R	22.89	
			-1.42
	S	21.47	
			+0.57
	R	22.04	
			-0.35
	S	21.69	
			+0.37
	R	22.06	
			-0.30
	S	21.76	
			+0.46
	R	22.22	
			-0.97
	S	21.25	
			+0.01
	R	21.26	
			-0.43
	S	20.83	
			+0.17
	R	21.00	
			-0.05
	S	20.95	
			+0.11
	R	21.06	
			-0.46
	S	20.60	
			+0.62
	R	21.22	
			-0.46
	S	20.76	
			+0.10
	R	20.86	
			-0.18
A29h,	S	20.68	

the size of debris along the lengths of individual ridges. In the centre of the embayment there is a change in the orientation of each successive beach ridge. Here the ridges terminate on the SW shore of the loch and are responsible for its serrated configuration. Farther seaward the lower ridges located on the western margin of the embayment face westward and descend seaward in altitude as a continuous staircase of unvegetated ridges and swales.

Levelled profiles across the shingle spreads are shown in Fig. 43. In addition to profiles measured on the unvegetated beach ridges, levelling was combined with peat probing in the vegetated area N and S of the loch (Figs. 34 and 43). Along the inner edge of the embayment the highest 8 beach ridges (R1-R8) occur at 36.0-33.8 m and are separated from the lower ridges by a distinct break in the slope profile that occurs between 34.3 and 31.9 m. The break in slope coincides with a deep swale (S10) that is replaced seawards by an area of undulating ridge and swale topography, 60 m in width, the surface of which ranges between 33.4 and 31.9 m.

North of Loch Aoinidh the vegetated surface is underlain by two distinct groups of ridges (Fig.43, profile A30-A31) that are separated from each other by a marked break in slope. Seaward of ridge R13 a staircase of 7 low-amplitude ridges (R14-R20) descends from 32.3 m to 28.2 m. At the base of R20 there is an extremely well-developed third break of slope. Here a series of 7 ridges (R21-R27) between 28.6 m and 27.8 m extends westward for 60 m as an almost horizontal spread of cobbles. The ridges that impound Loch Aoinidh (surface 27.2 m) extend northwards where they are buried by vegetation. These ridges are succeeded westwards by 27 ridges separated by swales that mantle the High Rock Platform and terminate seaward at the Main Rock Platform cliff. Here the shingle surface is concave in profile and is characterised by several large broad ridges (eg. R37, R41, R44, R45 and R48) separated by suites of low-amplitude ridges and swales. In the southern area of the embayment a high ridge can be traced beneath a cover of vegetation towards Loch a Mhile. The crest altitude of this ridge varies between 36.1 m and 36.4 m. It is noteworthy that the profile of the Loch Aoinidh shingle ridges between c. 30 m and 20 m is similar to the profile described farther north at the S end of the S Shian embayment (cf. Figs. 42 and 43).

Loch a Mhile

Seawards of Loch a Mhile a shingle staircase consisting of 14 ridges separated by swales declines in altitude from the loch margin to the top of the Main Rock Platform cliff (Plate 4). The shingle ridges overlie the High Rock Platform and are separated from it by locally thick accumulations of till (Chapter 7). Levelled profiles across the ridge and swale topography (Fig. 44) indicate two distinct slope units. Firstly, the four highest ridges (R1-R4) range in crest altitude from 36.8 m to 33.2 m and are succeeded seaward by ridge R5 that occurs at a considerably lower altitude (31.5 m) (Fig.44, Table 20). Seaward of ridge R5 the shingle profile is convex and is characterised by a sequence of ridges and swales that extends seawards for approximately 80 m. The highest ridge (R1) reaches a maximum altitude of 36.8 m and lies c. 1.5 m above the level of the loch.

S of Loch a Mhile

Approximately 400 m S of Loch a Mhile a narrow suite of raised shingle ridges descends as a continuous staircase to the top of the Main Rock Platform cliff (Fig. 44). Hence 14 ridges separated by swales decline in altitude from 36.4 m to 23.2 m (Table 20). The shingle spread is flanked to both the north and south by low inclined

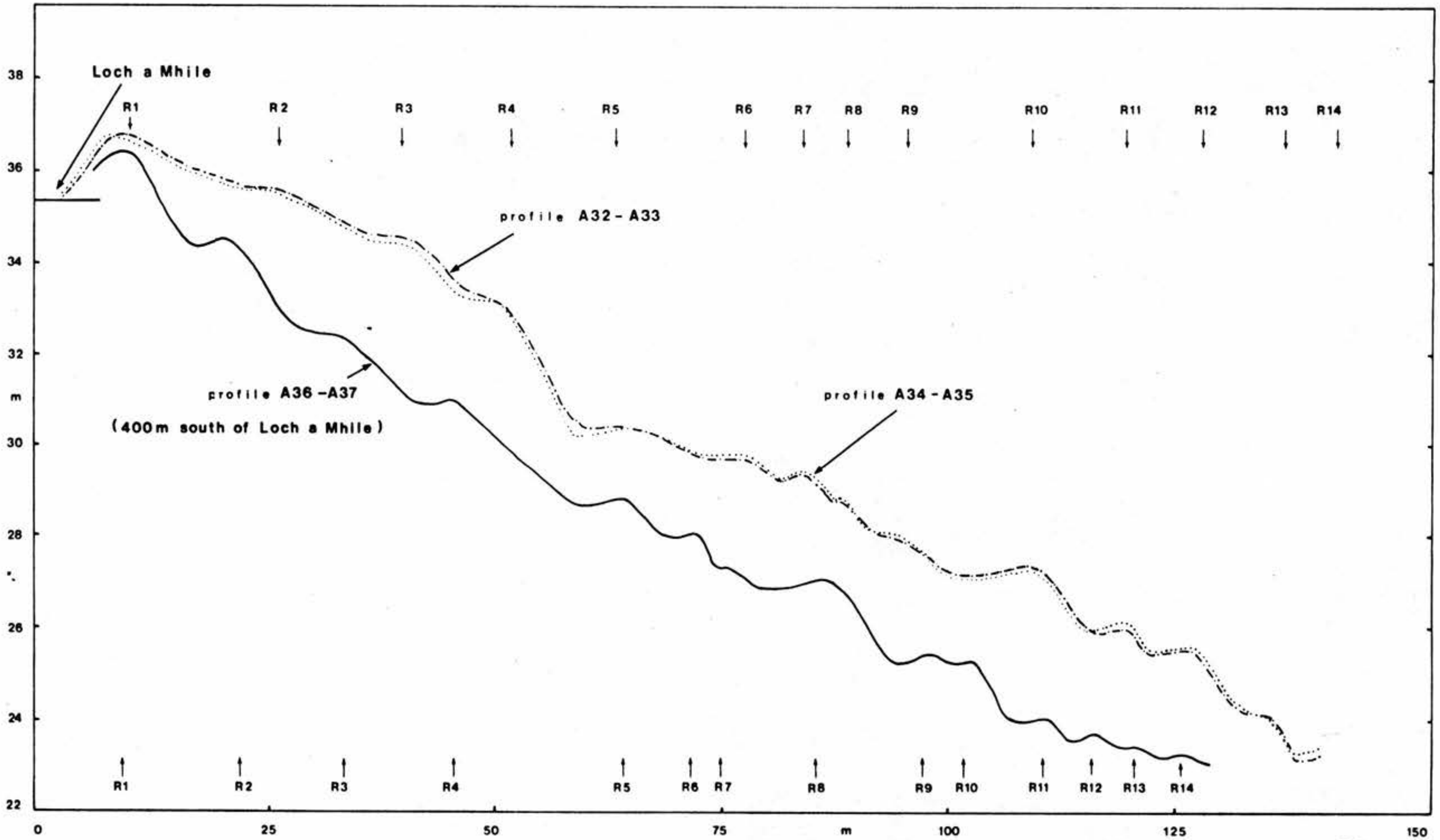


Fig. 44 Lateglacial shingle profiles, Loch a Mhile, W Jura. Profile locations shown on Fig. 34.

Table 20

Loch a Mhile, W Jura: lateglacial ridge and swale altitudes and amplitudes

A32

R	36.77	
		-0.85
S	35.92	
		(-0.38)
R	35.54	
		-0.86
S	34.68	
		(-0.29)
R	34.39	
		-1.14
S	33.25	
		(-0.10)
R	33.15	
		-1.94
S	31.21	
		+0.19
R	31.40	
		-1.61
S	29.79	
		+0.07
R	29.86	
		-0.47
S	29.39	
		+0.01
R	29.40	
		-0.50
S	28.90	
		(-0.14)
R	28.76	
		-0.56
S	28.20	
		(-0.22)
R	27.98	
		-0.68
S	27.30	
		+0.10
R	27.40	
		-1.42
S	25.98	
		+0.04
R	26.02	
		-0.49
S	25.53	
		+0.16
R	25.69	
		-1.29
S	24.40	
		(-0.16)
R	24.24	
		-0.96
S	23.36	
		+0.08

A33

A34

R	36.72	
		-0.92
S	35.85	
		(-0.29)
R	35.56	
		-0.84
S	34.72	
		(-0.24)
R	34.48	
		-1.05
S	33.43	
		(-0.31)
R	33.12	
		-1.83
S	31.29	
		+0.18
R	31.47	
		-1.78
S	29.69	
		+0.18
R	29.87	
		-0.60
S	29.27	
		+0.25
R	29.52	
		-0.51
S	29.01	
		(-0.21)
R	28.80	
		-0.62
S	28.18	
		(-0.16)
R	28.02	
		-0.70
S	27.32	
		+0.08
R	27.40	
		-1.35
S	26.05	
		+0.19
R	26.24	
		-0.65
S	25.59	
		+0.05
R	25.64	
		-1.17
S	24.47	
		(-0.34)
R	24.13	
		-0.76
S	23.51	
		+0.14

A35

A36

R	36.41	
		-2.06
S	34.35	
		+0.15
R	34.50	
		-1.90
S	32.60	
		(-0.18)
R	32.42	
		-1.45
S	30.97	
		+0.03
R	31.00	
		-2.27
S	28.73	
		+0.09
R	28.82	
		-0.77
S	28.05	
		+0.03
R	28.08	
		-0.68
S	27.40	
		(-0.12)
R	27.28	
		-0.38
S	26.90	
		+0.19
R	27.09	
		-1.76
S	25.33	
		+0.22
R	25.55	
		-0.26
S	25.29	
		(-0.15)
R	25.14	
		-1.07
S	24.07	
		+0.07
R	24.14	
		-0.46
S	23.68	
		+0.11
R	23.79	
		-0.32
S	23.47	
		+0.05
R	23.52	
		-0.28
S	23.24	
		+0.05
R	23.29	
		-0.10
S	23.19	

A37

* surface altitude of Loch a Mhile (6.7.77) = 35.33 m

* shingle profiles shown on Fig. 44

* profile locations shown on Fig. 34

quartzite ridges that protrude above the surface of the High Rock Platform and which undoubtedly were of considerable importance in the alignment of the ridges and swales. The higher shingle surface consists of a series of broad ridges that is succeeded seaward at 28.8 m by a series of smaller, more numerous, low-amplitude ridges and swales. In this lower area the most prominent ridge is R8, which protrudes markedly above the otherwise evenly sloping surface of ridges and swales that mantles the lower slopes.

N of Bhrein Port

N of Bhrein Port (Fig. 45) a series of 25 ridges separated by swales extends 190 m seawards from the high lateglacial terrace to the top of the Main Rock Platform cliff. The shingle surface is concave in profile (Table 21) and is characterised, particularly at altitudes below 32 m, by well-defined ridges and swales. The highest ridge (R1) is the most conspicuous feature of the shingle staircase and lies at 39.2 m. Seaward of this ridge (R1) the shingle surface descends steeply to 35.1 m where a secondary ridge (R2) is banked against its frontal slope. On the lower shingle surface the most noticeable drop in altitude occurs on the frontal slope of ridge R14 between 24.3 m and 22.6 m. Suites of raised beach ridges associated with similar breaks of slope occur at 28 m (R7 and R8), between 22.0 m and 22.5 m (R15-R17) and also between 19.4 m and 20.3 m (R20-R25)

N of Ruantallain

N of Ruantallain on an exposed section of the High Rock Platform at the mouth of Loch Tarbert a suite of unvegetated shingle ridges descends seawards from 39 m to 29.2 m (Fig. 46, Table 21). The ridges are poorly developed and are banked against each other at successively lower altitudes. Pronounced steepening of the ridge

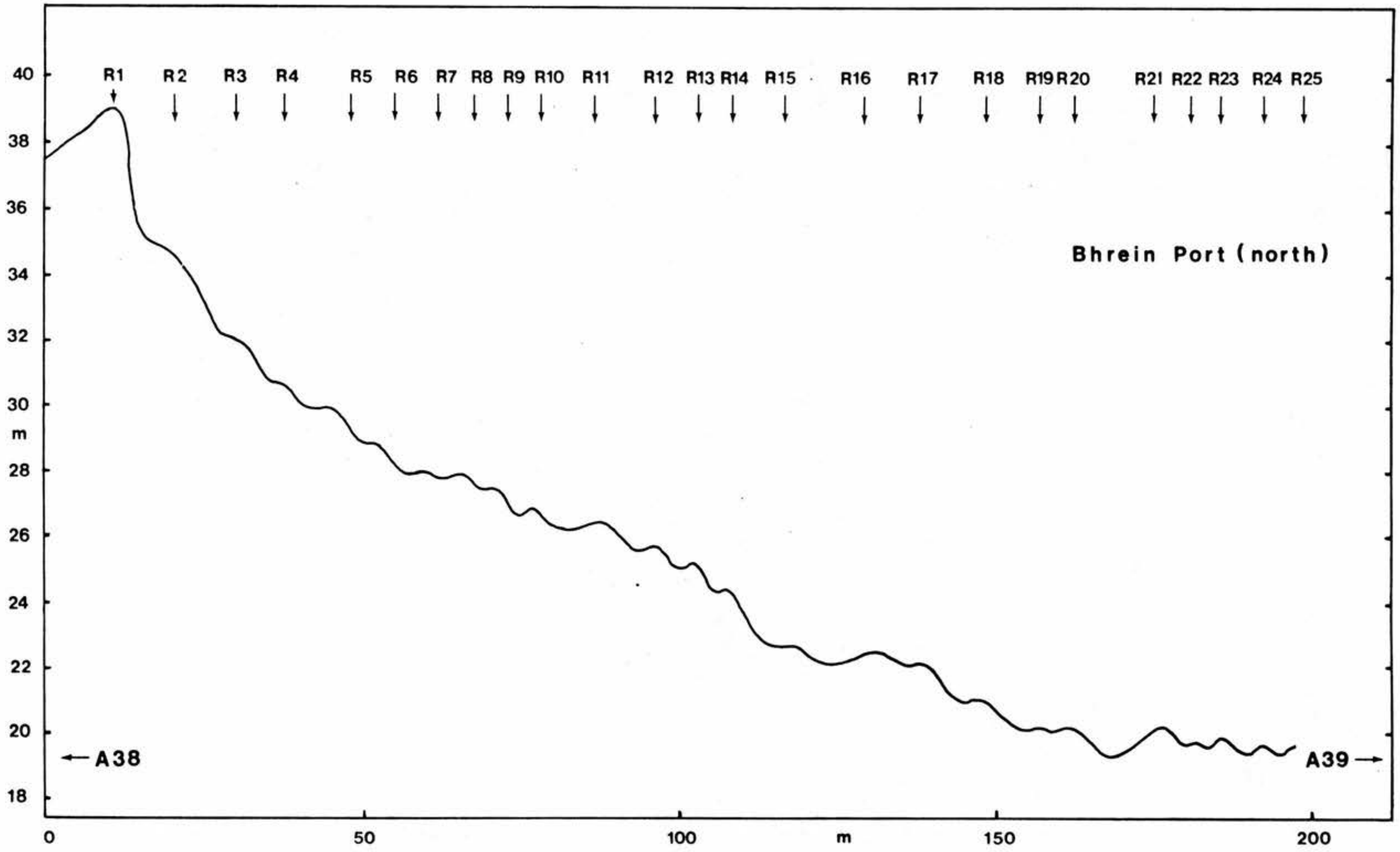


Fig. 45 Lateglacial shingle profile, Bhrein Port, W Jura. Profile location shown on Fig. 34.

Table 21

Lateglacial shingle spreads: Loch a Mhile - Ruantallain, W Jura
Ridge and swale altitudes and amplitudes

Bhrein Port

Profile		Altitude	Amplitude
A38			
	R	39.22	
			-4.10
	S	35.12	
			(-0.40)
	R	34.72	
			-2.62
	S	32.10	
			(-0.04)
	R	32.06	
			-1.27
	S	30.79	
			(-0.12)
	R	30.67	
			-0.59
	S	30.08	
			(-0.28)
	R	29.80	
			-0.93
	S	28.87	
			(-0.01)
	R	28.86	
			-0.88
	S	27.98	
			-0.10
	R	28.08	
			-0.20
	S	27.88	
			0.06
	R	27.94	
			-0.50
	S	27.44	
			0.09
	R	27.53	
			-1.19
	S	23.64	
			0.29
	R	26.63	
			-0.59
	S	26.04	
			0.17
	R	26.21	
			-0.86
	S	25.35	
			0.23
	R	25.58	
			-0.75
	S	24.83	
			0.15
	R	24.98	
			-0.61
	S	24.37	
			(-0.05)
	R	24.32	
			0.28
	S	22.60	
			(-0.04)
	R	22.56	
			-0.50
	S	22.06	
			0.30
	R	22.36	
			-0.31
	S	22.05	
			(-0.08)
	R	21.97	
			-1.01
	S	20.96	
			(-0.06)
	R	20.90	
			-0.91
	S	19.99	
			0.08
	R	20.07	
			-0.78
	S	19.29	
			0.83
	R	20.12	
			-0.52
	S	19.60	
			0.04
	R	19.64	
			-0.11
	S	19.53	
			0.31
	R	19.84	
			-0.74
	S	19.10	
			0.44
	R	19.54	
			-0.13
	S	19.41	
			0.72
	R	20.13	
			-0.43
	S	19.70	
			0.19
	R	19.89	
			-0.20
	S	19.69	
			0.46
	R	20.15	
A39			

Ruantallain

Profile		Altitude	Amplitude
A40			
	R	39.01	
			-0.95
	S	38.06	
			0.18
	R	38.24	
			-0.87
	S	37.37	
			(-0.27)
	R	37.10	
			-0.89
	S	36.21	
			0.11
	R	36.32	
			-1.83
	S	34.49	
			0.01
	R	34.50	
			-0.24
	S	34.26	
			0.03
	R	34.29	
			-0.45
	S	33.84	
			(-0.26)
	R	33.58	
			-1.36
	S	32.22	
			(-0.13)
	R	32.09	
			-2.09
	S	30.00	
			0.27
	R	30.27	
			-0.46
	S	29.81	
			0.00
	R	29.81	
			-0.62
	S	29.19	
			0.15
	R	29.34	
			-0.37
	S	28.97	
			0.20
	R	29.17	
A41			

Bagh Righ Mhor

Profile		Altitude	Amplitude
A42			
	R	38.62	
			-1.68
		36.94	
			*cx
		34.58	
			(0.06)
		34.52	
			cx
		33.58	
			cv
		32.94	
			cx
		32.44	
			cv
		31.42	
			cx
		31.05	
			cv
		30.66	
			cx
		30.25	
			cv
		29.89	
			cx
		29.57	
			cv
		29.03	
			cx
		28.49	
			cv
		27.44	
			cx
		27.23	
			cv
		26.93	
			cx
		26.65	
			cv
		26.18	
			cx
		25.92	
			cv
		25.49	
			cx
		25.31	
			cv
		25.02	
			cx
		24.59	
			cv
		23.46	
			cx
		22.91	
			cv
		22.35	
			cx
		22.20	
			cv
		21.91	
			cx
		21.68	
			cv
		21.53	
			cx
		21.50	
			cv
		21.27	
			cv
		21.26	
			cx
		20.98	
A43			

* convexity = cx
concavity = cv

* profile locations shown on Fig. 34

* shingle profiles shown on Figs. 45-47

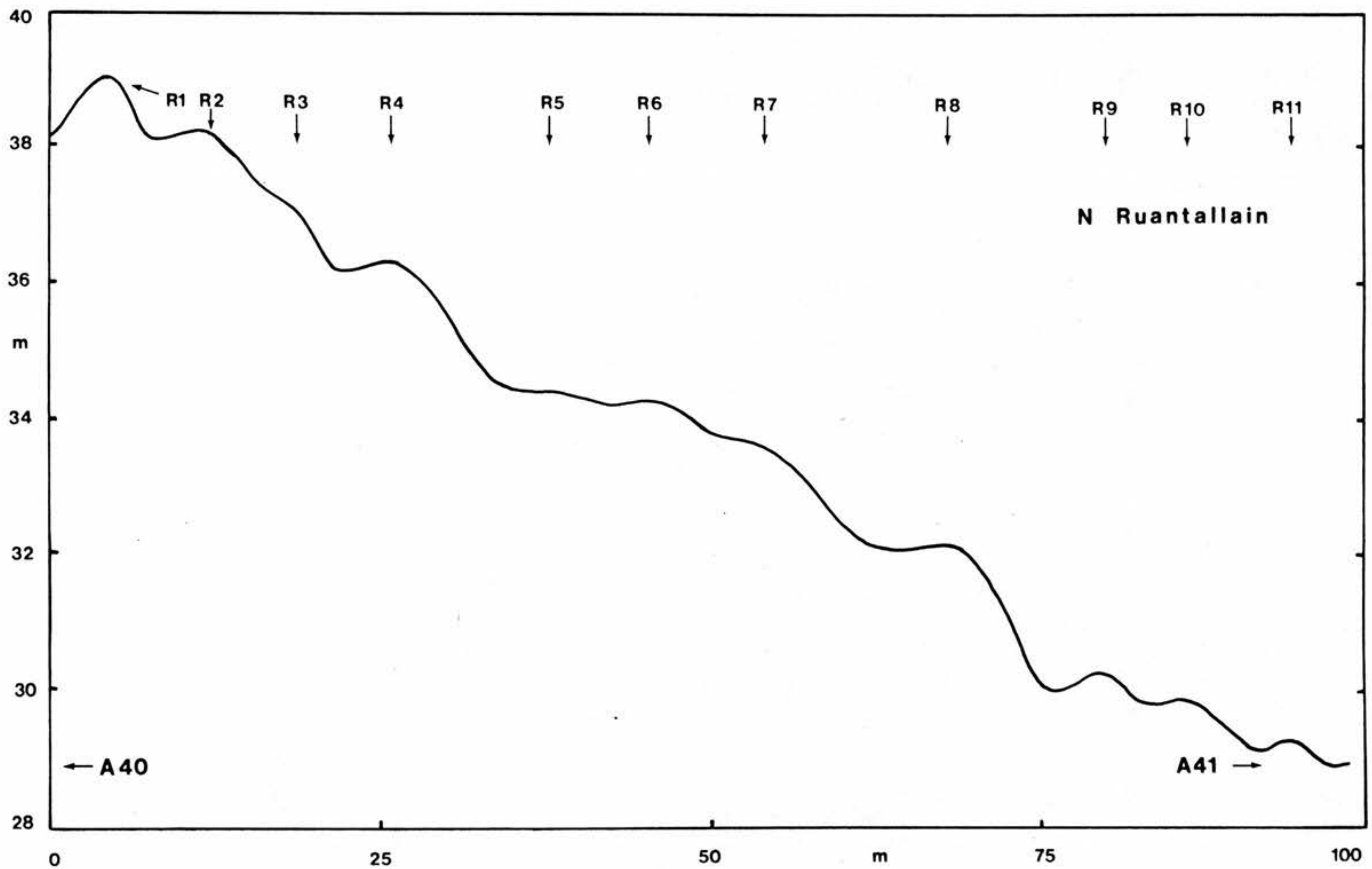


Fig. 46 Lateglacial shingle profile north of Ruantallain, W Jura. Profile location shown on Fig.34.

and swale profile occurs seaward of ridges R4 and R8 at 36.3 m and 33.6 m and also between 32.1 m and 30.0 m. General accordance of ridge crest altitudes occurs between 34.5 m and 33.6 m (R5-R7) and also between 30.0 m and 29.2 m (R9-R12).

Bagh Righ Mhor, E Ruantallain

At Bagh Righ Mhor a spectacular shingle spread extends seaward from 36.8 m to 21.0 m (Plates 24 and 26, Fig. 47, Table 21). Unlike the other raised shingle spreads of W Jura the Bagh Righ Mhor spread is not developed upon a pre-existing High Rock Platform but is instead banked against the rocky coastal slopes. As a result the ridge and swale topography is here poorly defined and is characterised by gentle convexities and concavities in the shingle profile. The highest ridge (R1) is the most conspicuous feature of the shingle spread and varies laterally in altitude between 39.4 m and 38.6 m. Seaward of this ridge the lower shingle deposits consist of a series of poorly developed ridges and swales.

W Glenbatrick

West of Glenbatrick thick accumulations of stratified lateglacial marine cobbles and gravels are banked against the coastal quartzite slopes. The accumulations are generally 80-150 m in width and possess an undulating surface of raised ridge and swale topography. Here 12 ridges separated by swales decline in altitude from 29 m to 18 m (Plate 29, Fig. 48). Levelled profiles across the shingle surface show that the three highest ridges (R1, R2 and R3) are separated from the lower deposits by a sharp break of slope. The altitude of the three highest ridge crests is between 29.3 m and 28.0 m while farther seaward a steep slope, 60 m in length, separates ridge R3 and R5. Seaward of ridge R5 the shingle surface is almost horizontal and is

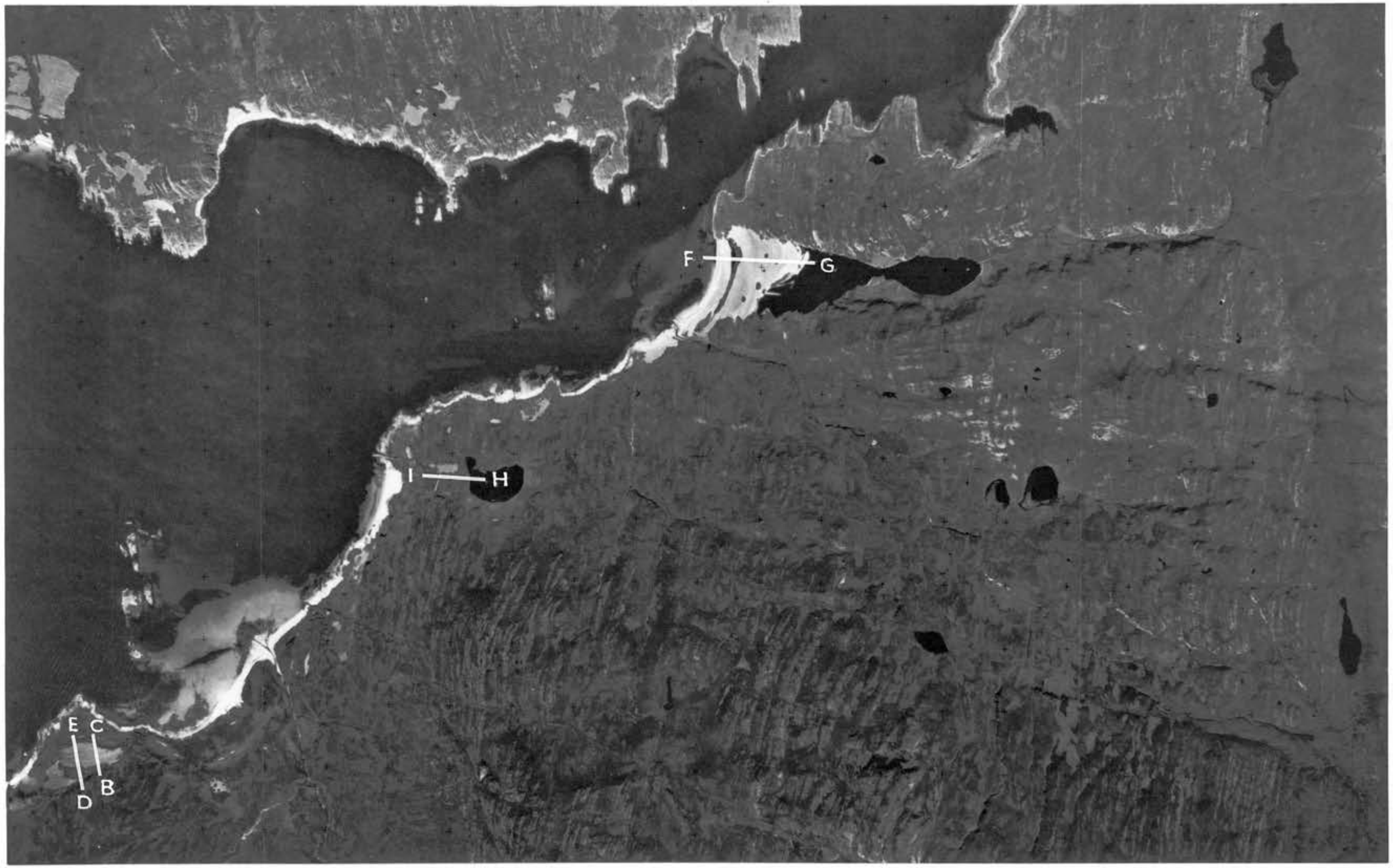


Plate 29. Aerial photograph (scale 1:25,000) of raised shore features, Loch Tarbert, W Jura, showing Loch Soirneach, Lochan MacPhi and raised shorelines W of Glenbatrick. Lines indicate sites of levelled profiles (cf. Figs. 48,49 and 50). Ministry of Defence (Air Force Dept.) photograph Crown Copyright.

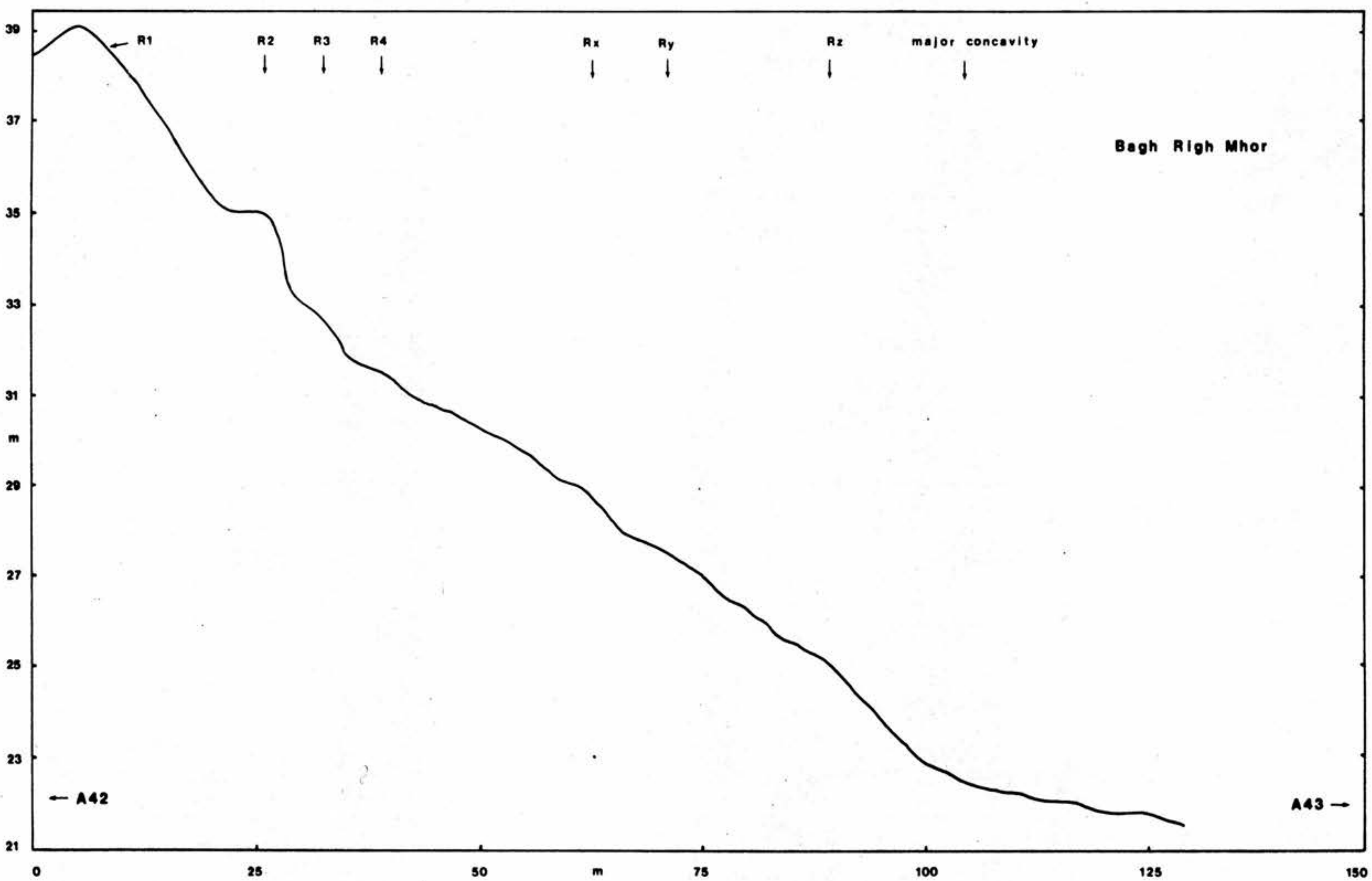


Fig. 47 Lateglacial shingle profile, Bagh Righ Mhor, W Jura. Profile location shown on Fig. 34.

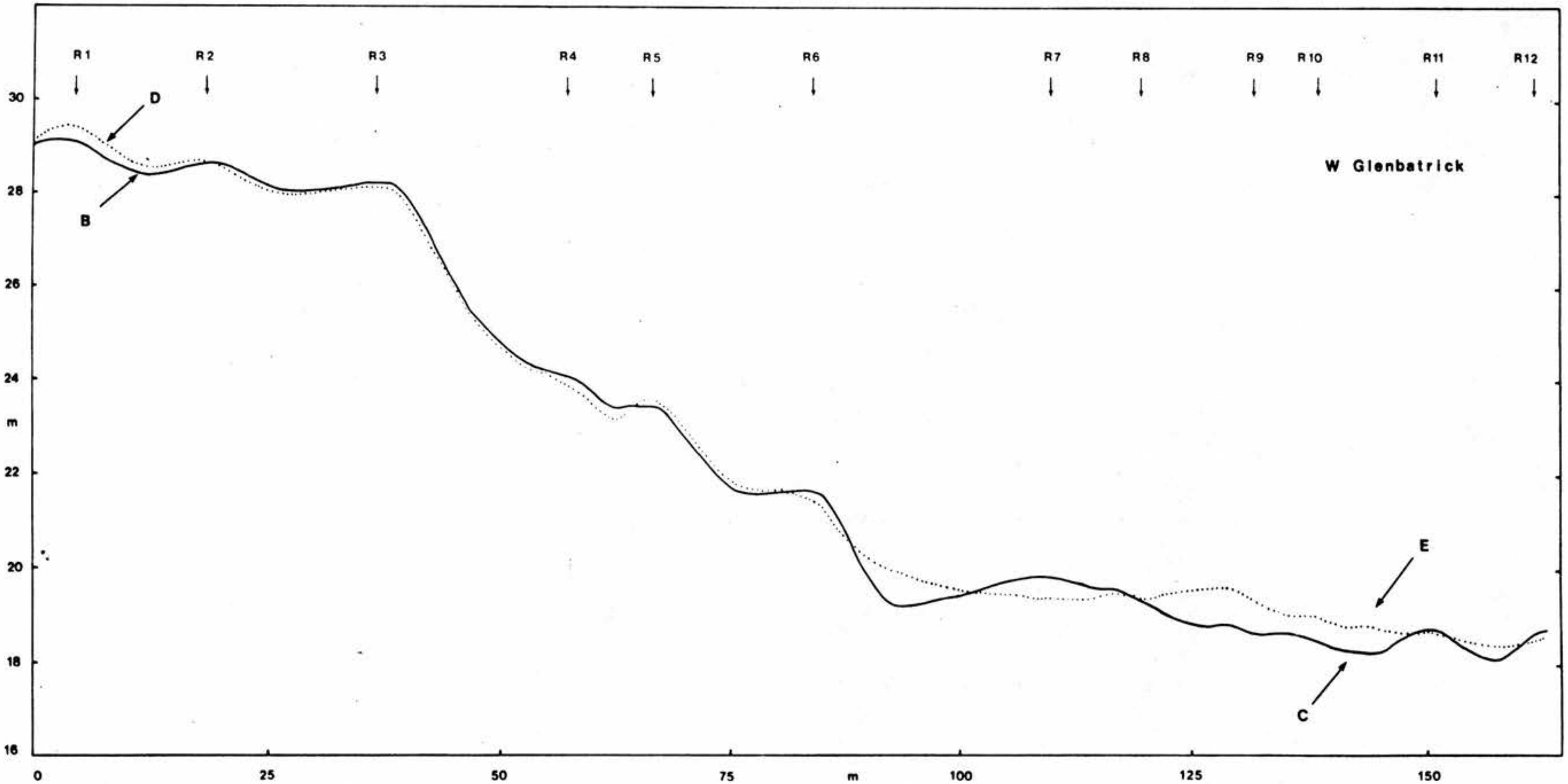


Fig. 48 Lateglacial shingle profiles, W Glenbatrick, S Jura. Profile locations shown on Plate 29.

characterised by broad ridges and swales that terminate seaward at c. 18 m on the cliff edge (Table 23).

Loch Soirneach, Inner Loch Tarbert, SW Jura.

Loch Soirneach occupies a structurally-controlled depression between two quartzite ridges on the southern shore of Loch Tarbert (Fig. 49, Plate 29). The loch, whose surface is at 9.5 m, is separated from Loch Tarbert by a series of 17 unvegetated shingle ridges that extend seaward from its surface to 4.8 m at the crest of the modern storm ridge (Table 22). Landward of this ridge an elongate rock bar whose crest is at 9.6 m is partially buried by raised shingle deposits and impounds Loch Soirneach. The most remarkable feature however is that the raised shingle deposits rise seaward in altitude from the loch before later descending to the present shore (Fig. 49, Table 22). The ridge and swale topography commences at the loch margin where six arcuate shingle spits extend eastward into the loch (R1-R6). Each ridge protrudes slightly above the loch surface while the sixth ridge (R6) is succeeded seaward by a series of 5 higher ridges separated by swales that rise to a maximum altitude of 14.9 m (R11). The highest ridge (R11) is approximately 400 m in length and exhibits a marked decrease in cobble size (from an average a-axis length of 13 cm at the west end of the ridge to 3.5 cm at its eastern end). Seaward of R11 the raised ridge and swale topography descends seaward.

Lochan MacPhi, S Jura

Lochan MacPhi (NR 532804) is located on the S shore of Loch Tarbert and occupies a structurally-controlled depression between a ridge of dolerite and one of quartzite. The loch is at 23.1 m and is impounded by a series of raised shingle ridges (Fig. 50, Table 23). The higher shingle ridges (R6-R11) occur between 23.9 m and 23.4 m and are

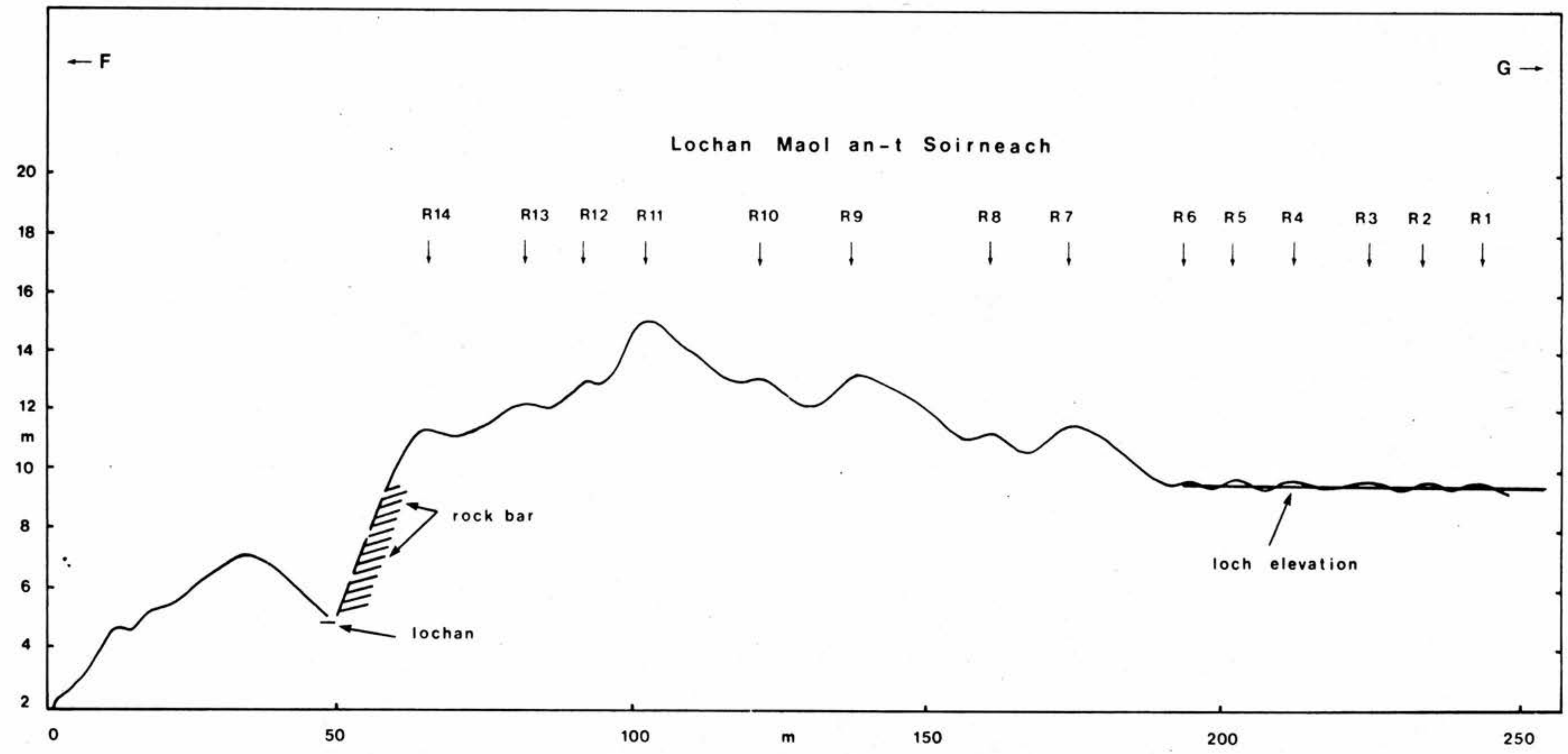


Fig. 49 Profile of raised shingle ridges, Lochan Maol an-t Soirneach, S Jura. Profile location shown on Plate 29.

Table 22

Loch Soirneach, S Jura: ridge and swale altitudes and amplitudes

F			
S	9.43		
		+0.13	
R	9.56		
		-0.19	
S	9.37		
		+0.19	
R	9.56		
		-0.15	
S	9.41		
		+0.17	
R	9.58		
		-0.20	
S	9.38		
		+0.34	
R	9.72		
		-0.34	
S	9.38		
		+0.13	
R	9.51		
		-0.08	
S	9.43		
		+2.01	
R	11.44		
		-0.82	
S	10.62		
		+0.55	
R	11.17		
		-0.17	
S	11.00		
		+2.16	
R	13.16		
		-1.09	
S	12.07		
		+0.84	
R	12.91		
		-0.04	
S	12.87		
		+1.87	
R	14.89		
		-2.01	
S	12.88		
		+0.06	
R	12.94		
		-0.77	
S	12.17		
		(-0.01)	
R	12.16		
		-0.94	
S	11.22		
		+0.01	
R	11.23		
	9.56	....	rock bar
	5.03	....	cliff base
S	5.89		
		+1.26	
R	7.15		
		-0.12	
S	6.03		
		(-0.75)	
R	5.28		
		-0.57	
S	4.71		
		+0.10	
R	4.81	....	modern storm ridge crest
	3.60	....	erosional notch
	2.42	....	Pelvetia
G			

* profile location shown on Plate 29

* elevation of Loch Soirneach (13.6.77) = 9.46 m

* shingle profile shown on Fig. 49

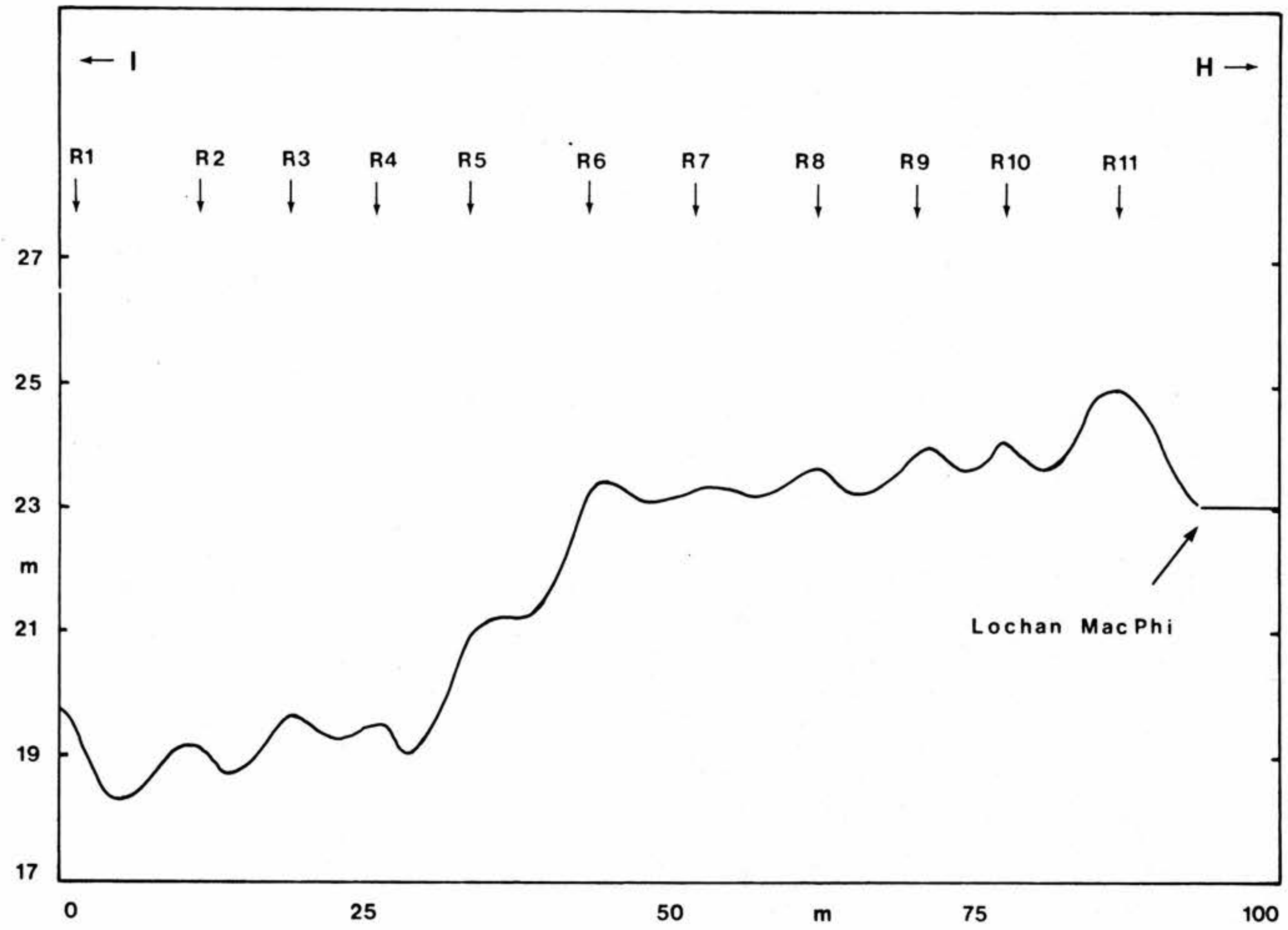


Fig. 50 Lateglacial shingle profile, Lochan MacPhi, SW Jura. Profile location shown on Plate 29.

Table 23

Lateglacial ridge and swale altitudes and amplitudes: W Glenbatrick and Lochan MacMhi, S Jura

Lochan MacPhi

H	R	24.94	
	S	23.64	-1.30
	R	24.09	+0.45
	S	23.06	-1.03
	R	23.98	+0.92
	S	23.27	-0.71
	R	23.63	+0.36
	S	23.21	-0.42
	R	23.38	+0.17
	S	23.11	-0.27
	R	23.43	+0.32
	S	21.23	-2.20
	R	21.21	(-0.02)
	S	19.01	-2.20
	R	19.51	+0.50
	S	19.31	-0.20
	R	19.62	+0.31
	S	18.72	-0.90
	R	19.17	+0.45
	S	18.30	-0.87
I	R	19.76 ------------ man-modified postglacial dunes below this altitude	+1.46

W Glenbatrick

B	R	29.59	
	R	29.86	+0.27
	S	28.83	-1.03
	R	29.12	+0.29
	S	28.48	-0.64
	R	28.51	+0.03
(banked)	R	25.97	-2.54
	S	23.53	-2.44
	R	23.86	+0.33
	S	22.09	-1.77
(banked)	R	21.93	(-0.16)
	S	20.37	-1.56
	R	20.22	(-0.15)
	S	19.71	-0.51
	R	19.11	(-0.60)
	S	18.49	-0.62
	R	19.10	+0.61
	S	18.76	-0.34
C	R	19.07	+0.31

D	R	29.51	
	R	29.55	+0.04
	S	28.78	-0.77
	R	29.12	+0.34
	S	28.38	-0.74
	R	28.58	+0.20
(banked)	R	26.00	-2.58
	S	23.86	-2.14
	R	23.95	+0.09
	S	22.00	-1.95
(banked)	R	22.10	+0.10
	S	19.88	-0.22
	R	19.74	(-0.14)
	S	19.62	-0.12
	R	19.68	+0.06
	S	18.59	-1.09
	R	18.77	+0.18
	S	18.56	-0.21
E	R	19.92	+1.34

* profile locations shown on Plate 29.

* shingle profiles shown on Figs. 48 and 50

separated from a lower shingle accumulation by a steep slope. The lower shingle surface consists of 4 vegetated ridges between 19.7 m and 19.5 m and is succeeded seaward by fossil vegetated sand dunes.

Loch na Sgrioba

At Loch na Sgrioba (Plates 5 and 7) a high unvegetated shingle ridge impounds the loch, the loch surface being at 31.6 m. The ridge is over 300 m in length, varies in altitude between 32.5 m and 33.6 m and is the most conspicuous high raised beach ridge in SW Jura. Approximately 25 m seaward of this ridge, a second (lower) ridge forms a wide rampart and varies in altitude between 30.7 m and 31.4 m. Seaward of this ridge a wide horizontal shingle surface consisting of low-amplitude ridges and swales occurs c. 29.2 m.

Interpretation

The occurrence of large shingle ridges that protrude conspicuously above the sloping surface of a raised shingle spread can be interpreted in either of two ways. Firstly it can be argued that large ridges are formed during a short period of time as a result of storm activity. Secondly, in contrast, large ridges can be interpreted as the product of repeated storm wave activity during a long period of relative sea-level stability. Since it has already been shown (Chapter 5) that modern storm ridges in W Jura and NE Islay have been repeatedly formed and destroyed by storm waves it cannot be maintained that large beach ridges represent the morphological effect of individual storms. Instead it is more likely that prolonged storm wave action during periods of relative sea-level stability is responsible for the formation of the large shingle ridges.

Wave activity in the study area has resulted in the formation of two beach ridge types. Firstly high-amplitude beach ridges are formed during storms and exhibit marked regional variations in crest altitude. Secondly low-amplitude beach ridges occur in the intertidal zone and are the result of moderate wave activity. If storm waves were responsible for the formation of all raised beach ridges in the study area it might therefore be expected that the measured ridge crest altitudes when plotted on a shoreline height-distance diagram would show a random scatter of points and render it impossible to identify any synchronous raised shorelines. Consequently the altitudes of all surveyed raised beach ridge crests between Ruantallain and N Shian Bay were plotted in a NE-SW projection plane on a shoreline height-distance diagram (Fig. 51). A similar diagram was constructed for all beach ridge altitudes in the study area (Fig. 52).

The beach ridge altitudes when so plotted exhibit distinct altitude patterns (Figs. 51 and 52). An alignment of ridge crest altitudes along sloping planes is readily apparent. It is noteworthy that the alignment of crest altitudes is most apparent between Ruantallain and Loch Aoinidh, the area most exposed to open Atlantic fetch. The most likely explanation of the observed pattern of crest altitudes is that moderate wave activity, rather than storms, was responsible for the formation of many of the ridges, the sloping alignment of altitudes indicating the differential isostatic tilting of individual shorelines. In addition the ridge crest altitudes of W Glenbatrick when plotted on a shoreline height-distance diagram (Fig. 52) show a clear sloping alignment with those measured between Ruantallain and Shian Bay. The W Glenbatrick ridges therefore conform to the general pattern of sloping lateglacial shorelines and their alignment on the height-distance diagram with the Ruantallain-Shian Bay ridges

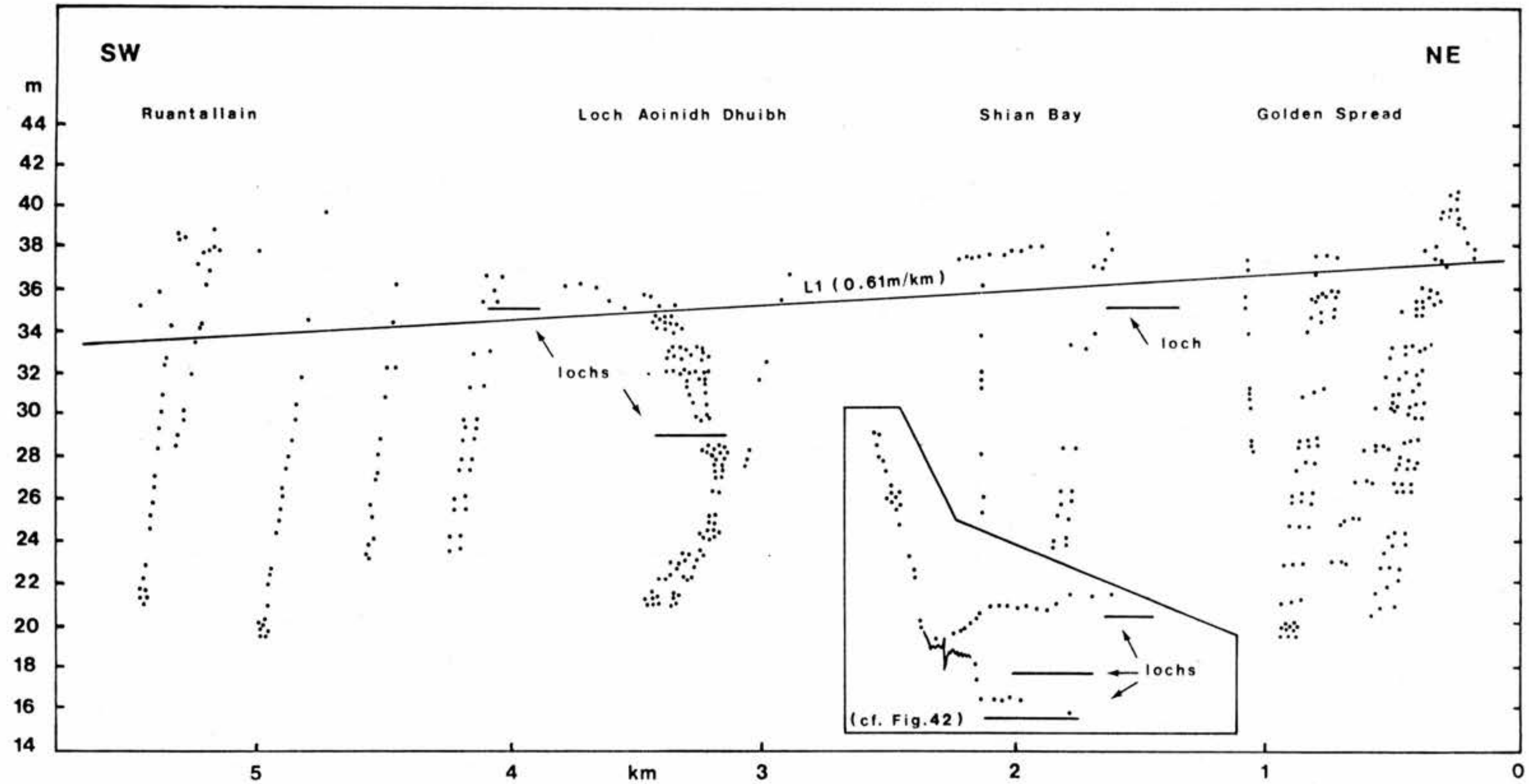


Fig. 51 Shoreline height-distance diagram of lateglacial beach ridge crests between Ruantallain and Shian Bay, W Jura. Lateglacial shoreline (L1) is also shown.

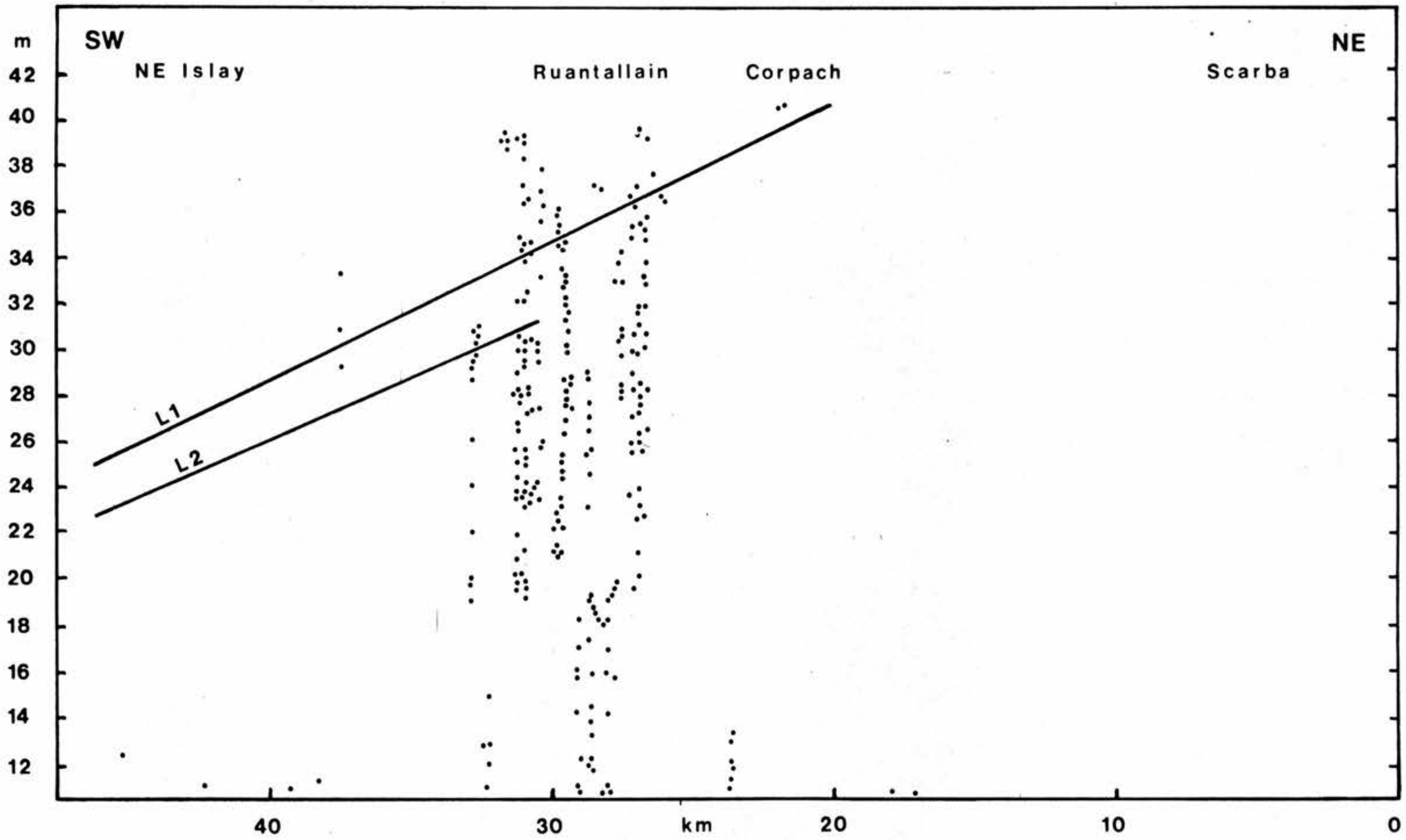


Fig. 52 Shoreline height-distance diagram of lateglacial beach ridge crests in the study area. Lateglacial shorelines L1 and L2 are also shown.

supports the hypothesis that most ridges were formed by moderate wave activity.

However the altitudes of ridge crests above c. 35 m (Fig. 51) show a randomly scattered distribution. The steeply sloping bands of ridge crest altitudes are more clearly shown on Fig. 52 where the highest beach ridges between Ruantallain and Corpach Bay have no comparable equivalent along the SW Jura coast. It is therefore suggested that the highest beach ridges between Ruantallain and Corpach Bay were formed by storm wave activity immediately after the formation of shoreline L1 when SW Jura remained ice-covered.

Although most shingle spreads are located in relatively exposed coastal locations, the Loch Aoinidh Dhuibh beach ridges are in a sheltered embayment and as a result they are extremely sensitive indicators of former sea-level changes. The sheltered environment in which the Loch Aoinidh ridges were formed is clearly demonstrated by the lack of altitude variation along the lengths of individual ridge crests (Fig. 43). When plotted on a height-distance diagram the crest altitudes group into several distinct clusters (Fig. 51) and correspond to the stepped shingle profile shown in Fig. 43. The gradual seaward decline in altitude of the beach ridges suggest that here the general relative marine regression from 36 m to 21 m was not interrupted by any major relative marine transgression. The clustered nature of the Loch Aoinidh crest altitudes also suggests that the general marine regression associated with shingle ridge formation was characterised by a series of regressional phases separated by stillstands rather than by a continuously falling sea-level.

In the Loch Aoinidh Dhuibh area the largest-amplitude ridges and swales (R1-R8) occur above 32 m and most probably reflect early lateglacial storm activity rather than changes of sea-level. A possible period of relative sea-level stability is indicated by ridges R9-R14 between 32.0 and 33.5 m, this being followed by a period of general marine regression and the formation of ridges R15-R20. This in turn was followed by a second period of marine stability indicated by ridges R21-R27 between 28.6 m and 27.8 m which impound Loch Aoinidh Dhuibh (surface 27.2 m) (Fig. 43). After this period further marine regression occurred and accompanied the formation of a series of 28 low-amplitude beach ridges (R27-R55).

The progressive seaward decline in ridge crest altitudes shown in the W Jura shingle profiles indicates that, although relative marine stillstands may have occurred during this period, there were no major sea-level oscillations. Below c. 25 m most of the W Jura shingle spreads terminate seawards at the top of the Main Rock Platform cliff and as a result regional patterns of lower lateglacial sea-level changes cannot be established. However, at two locations (S Shian Bay and Loch Soirneach) staircases of shingle occur at lower altitudes and as a result the pattern of lower lateglacial relative sea-level changes can also be investigated.

At S Shian Bay the high terrace of marine gravels is banked against the inner edge of the High Rock Platform and is cliffed at its seaward edge (Fig. 41). The formation of the cliffline can only be related to former marine erosion at the cliff base, which is here overlain by the well-developed ridge R9 formed prior to the formation of the Colonsay ridge. After the development of ridge R9 a series of north-facing shingle ridges was deposited in the south of the embayment at successively lower altitudes contemporaneously with the southward progradation of the Colonsay ridge. The lowest ridge of this shingle spread (Fig. 42, profile A23-A25) occurs landward of and at lower altitudes than the southern end of the Colonsay ridge. Since ridges R12-R22 (Fig. 42) decline in altitude northwards it is suggested that the southward progradation of the Colonsay ridge was accompanied by a gradual fall in relative sea-level from 19 m to 17.5 m. When relative sea-level was at c. 17.5 m the final ridge (R22) was sealed from the sea by the southern extension of the Colonsay ridge. During this period of falling sea-level, Inverness loch, Maidens loch, and Loch Maol were sealed from the sea as the Colonsay ridge extended southward. It follows that the Colonsay ridge, postulated by McCann as representing a major sea-level oscillation, was formed during a period of falling sea-level.

The clearest morphological evidence for a lateglacial sea-level oscillation is at Loch Ṣoirneach on the S shore of Inner Loch Tarbert. Since the interpretation of the shoreline evidence in this area is contingent on the maximum altitude reached by the postglacial sea, discussion is presented in the following chapter.

At several locations in W Jura the high shingle ridges are separated from the underlying High Rock Platform by accumulations of till that contain numerous angular quartzite clasts. In SW Jura evidence has already been presented of raised till platforms overlain by shingle ridges while in W Jura (between Ruantallain and N Shian Bay) and NE Islay the High Rock Platform is overlain at several locations by glacial till. It is therefore most likely that the quartzite shingle cobbles of the lateglacial ridges were derived from the marine erosion of till that formerly mantled large areas of the High Rock Platform. Destructive lateglacial waves most probably resulted in the seaward removal of debris and its later deposition as beach ridges by constructive wave action. The distribution of the raised shingle ridges is also the result of the sloping lateglacial shorelines that in W Jura permitted the deposition of shingle ridges on the High Rock Platform. In NE Islay however, owing to the steep gradients of the lateglacial shorelines, the latter could only develop on the seaward part of the platform.

Summary

It has been shown that the pattern of sea-level changes in W Jura is more complicated than suggested by McCann. Deglaciation first occurred between Ruantallain and Corpach Bay and also in NE Islay. Marine incursion into these deglaciated areas resulted in the formation of the L1 shoreline. During the formation of this shoreline SW Jura remained ice-covered. At a slightly later date SW Jura was also deglaciated and this was accompanied by the formation along the SW Jura coast of shoreline L2. During this period high shingle ridges were formed between Ruantallain and N Shian Bay. Subsequently glacio-isostatic uplift of the land surface resulted in a general marine regression that was associated with the deposition of the W Jura shingle ridges. The distribution and height variations of the highest shingle ridges in W Jura suggest that they were formed during storms.

In contrast, it is suggested that many of the lower shingle ridges were formed by moderate wave action. The distribution and altitude of shingle ridges in the Loch Aoinidh embayment show that the general lateglacial marine regression was uninterrupted by any major reversals of sea-level. Instead the beach ridge altitudes suggest that the marine regression was characterised by a series of regressional phases separated by stillstands. At Loch Aoinidh periods of relative sea-level stability occurred when sea-level was at 33.5-32.0 m and at 28.6-27.8 m. It is also concluded that the period of relative sea-level stability at 55-60 feet (16.8 m-18.3 m) above H.W.M. proposed by McCann, did not occur.

Chapter 10

Postglacial shorelines

Little information is available regarding changes of relative sea-level in W Scotland between the end of the Loch Lomond Stadial and the maximum of the main postglacial transgression. However in E Scotland detailed field investigations of buried shorelines in the Forth valley (cf. Sissons et al., 1966) indicate that a marine transgression occurred after the development of the Main Lateglacial Shoreline (the Buried Gravel Layer) and resulted in the formation of the High Buried Shoreline (Sissons, 1969). The development of this shoreline was followed by a marine regression in SE Scotland that accompanied the formation of the Main and Low Buried Shorelines (Sissons, 1966; Kemp, 1971; Sissons and Brooks, 1971; Cullingford, 1972). After the formation of the last of these shorelines the main postglacial transgression occurred and culminated c. 6,500 B.P. at the Main Postglacial Shoreline. Subsequent marine regression resulted in the formation of raised shorelines at lower altitudes, three having been identified in the Forth valley (Sissons et al., 1966).

In W Scotland no early postglacial shorelines equivalent to the buried ones of E Scotland have been identified. Gray (1974b, p.133) noted that,

> "... there is no evidence to suggest that the buried beaches identified in SE Scotland (eg. Sissons et al., 1966) emerge from below the Main Postglacial Shoreline as has been suggested might occur nearer the centre of isostatic uplift."

However raised postglacial terraces believed to have been formed at the culmination of and after the main postglacial transgression have been identified. For example, McCann (1966) measured the altitude of numerous raised shoreline fragments between the Firth of Lorn and

Loch Broom and concluded that the Main Postglacial Shoreline declined in altitude W and NW with a regional gradient of 0.07 m/km. McCann's measurements were criticised by Gray (1974b, p.130), who doubted their accuracy since they were based on Abney level traverses from H.W.M.O.S.T. and also included altitudes of Main Rock Platform fragments as part of the Main Postglacial Shoreline. Similarly in Argyll and NE Ulster Synge and Stephens (1966) conducted Abney level traverses from H.W.M.O.S.T. and claimed to have identified 5 postglacial shorelines. They concluded that the Main Postglacial Shoreline declines in altitude away from the centre of isostatic uplift with a regional gradient of 0.075 m/km. Farther south, along the Solway Firth, Jardine (1977) concluded that the Main Postglacial Shoreline lies between 3 m and 4 m while Andrews et al.,(1973) suggested that the main postglacial transgression reached 10.7 m at Macrihanish, Kintyre.

In Lorn and E Mull Gray (1974b) found that the Main Postglacial Shoreline declines in altitude to the SW and W from a maximum of 13-14 m near Oban, the gradient being 0.07 m/km. Gray compared this gradient with values of 0.08 m/km and 0.09 m/km obtained for the same shoreline in the Forth and Tay valleys (Sissons et al., 1966; Cullingford, 1972) and added that (p.136) "... it is difficult to trace well-defined raised shorelines below the Main Postglacial Shoreline.. (although).. there appear to be linear bands of points at about 8 m O.D. (PS3) and 4 m O.D. (PS5)." Gray's tentative identification of two separate raised shorelines below the Main Postglacial Shoreline differs from McCann's (1966) belief that such later shorelines do not exist. Moreover Gray (1974b, p.136) tentatively proposed that the PS3 shoreline correlated with PG3 in the Forth valley (Sissons et al., 1966) and shoreline LC1 in the Tay valley (Cullingford, 1972). Gray also suggested that raised shoreline PS5 "... may correlate with LC4 or LC5 in the Tay (Cullingford, 1972) but correlation with the Forth is more difficult."

Few radio-carbon dates are available for the culmination of the main postglacial transgression in W Scotland. Jardine (1975) suggested that the transgression began after 8,400 B.P. on the S Ayrshire coast and that its culmination was regionally diachronous, occurring as late as 5,600 B.P. in the eastern Solway Firth. At North Carn, N Jura, charcoal buried beneath a raised beach ridge yielded a radio-carbon age of 7,414 ± 80 B.P. (Harkness and Wilson, 1974, p.250, SRR-161) and suggests that the culmination of the postglacial transgression occurred after this date (Jardine, 1977). Similarly in Oronsay shells located in raised shoreline deposits are dated at 7,020 ± 140 B.P. and 7,510 ± 150 B.P. (Jardine, 1977). In addition bones from a shell midden in Oronsay that "... could only have been occupied after the sea had retreated from its highest postglacial level" (Mackie, 1972, p.412) yielded radio-carbon ages of 5,015 ± 210 B.P. and 5,755 ± 180 B.P. respectively. Collectively these dates suggest that the culmination of the main postglacial transgression in Jura and Oronsay occurred approximately between 5,750 and 7,000 B.P..

In Jura, Scarba and NE Islay suites of raised beach ridges mantle the Main Rock Platform surface and are replaced at the mouths of several river valleys by raised coastal terraces. Since it has been shown that the Main Rock Platform is lateglacial in age it is believed that the beach ridges that mantle its surface are postglacial in age: the highest ridges having formed during the culmination of the main postglacial transgression. In addition since the estuarine raised coastal terrace fragments are contiguous with the highest raised beach ridges described above it is believed that they also are postglacial in age. Moreover since the terrace fragments are the only identifiable such features occurring at each location the simplest explanation is that they are of the same age, each having formed during the culmination of the main postglacial transgression. However no evidence has been found for early postglacial shorelines formed before the Main Postglacial Shoreline. The altitudes of terrace fragments formed during the culmination of the main postglacial transgression were measured in order to determine shoreline gradient.

In contrast to poorly-developed raised terrace fragments, raised beach ridges formed during and after the main postglacial transgression are well-developed: at Inver, SW Jura, a staircase of 31 unvegetated beach ridges declines in altitude from 12.3 m to the present shore (Fig.35). The most extensive postglacial deposits occur in SW Jura where a degraded cliffline composed of till and lateglacial marine gravels is succeeded seaward by wide areas of raised beach ridges that mantle the surface of the Low Rock Platform (Plate 30).

At two locations in W Jura where the Main Rock Platform cliff is poorly developed or absent, the occurrence of raised beach ridges between c. 9 m and 20 m renders it difficult to establish whether the ridges are postglacial or lateglacial in age. This problem assumes special importance since at Loch Soirneach, SW Jura, there is clear evidence of a former sea-level oscillation between these altitudes (Chapter 9 § p.283 et. seq.). As a result measurements of regional variations

Plate 30. Till cliffs and postglacial raised beach deposits, SW Jura.

Plate 31. Modern storm ridge and postglacial raised beach deposits, Lang Aoinidh, SW Jura.

of postglacial ridge and swale altitudes were undertaken in order to determine (a) the relative age of the Loch Soirneach oscillation (b) the highest altitudes reached by postglacial storm waves and (c) the pattern of relative sea-level movements after the maximum of the main postglacial transgression.

Postglacial terrace fragments

NW Jura

Along the seaward part of Glengarrisdale the valley floor is filled by a raised postglacial terrace. The terrace is composed of sand and gravel and is succeeded inland by a broad vegetated raised beach ridge (Plate 32). Measurement of 3 fragments of this terrace indicates an average altitude of 10.3 m. The seaward base of the raised beach ridge occurs at 10.5m. Owing to the sheltered location of the terrace it is suggested that local high water mark during the culmination of the transgression was here approximately 10.3 m. Farther south at Bagh Spereig (NR 634963) a high coastal terrace associated with the postglacial transgression occurs at 10.3 m.

At the mouth of Glendebadel a sequence of low amplitude shingle ridges is separated from higher lateglacial marine deposits by a narrow terrace composed of sand and gravel (NR 624950). The inner edge of the terrace exposed in a stream section is 9.6 m. A similar terrace also occurs farther south at the mouth of the Corpach river valley (NR 570914), at 9.3 m. Approximately 2.5 km north of the Corpach valley at Garbh Uisge (NR 589932) the high quartzite cliffs are separated from the sea by a narrow estuary sheltered from severe wave attack by several raised stacks that almost seal its seaward entrance. The maximum altitude of the highest raised terrace is here

Plate 32. Raised postglacial vegetated beach ridge and terrace, Glengarrisdale, NW Jura.

9.3 m. Farther south along the W Jura coast at Shian Bay (NR 532877) a series of parallel postglacial raised beach ridges are replaced landward by a narrow raised coastal terrace, the altitude of which is 9.1 m.

SW Jura

At the mouth of Glenbatrick (NR 520800) raised vegetated postglacial beach ridges are separated from higher lateglacial marine deposits by a terrace and cliff. The raised terrace is at 8.9 m and exposures along the bank of the Glenbatrick river show it to be composed of gravel and shell sand.

NE Islay

In the Coir Odhar embayment a widespread accumulation of postglacial beach gravels extends c. 350 m inland (Plate 3). The embayment is c. 300 m in width and is surrounded by quartzite cliffs that here form the frontal edge of the High Rock Platform. At the inner margin of the embayment the average altitude of the raised terrace is 8.5 m.

Postglacial ridge and swale topography

In S Shian Bay raised unvegetated beach ridges extend continuously from the Colonsay ridge to the modern storm beach (Fig.53). Measured profiles reveal a progressive seaward decline in ridge altitude, the crests of individual ridges varying by as much as 1.7 m along their lengths. The two northern profiles (Fig. 53, AB, CD) are characterised by a sharp break of slope between 10.5 m and 14 m that is poorly developed in the southern profiles. At altitudes above 12 m there is an overall southward decrease in the altitude of individual ridges: below this elevation the pattern is reversed. In this area modern storm ridges reach crest altitudes of 7.7 m. At

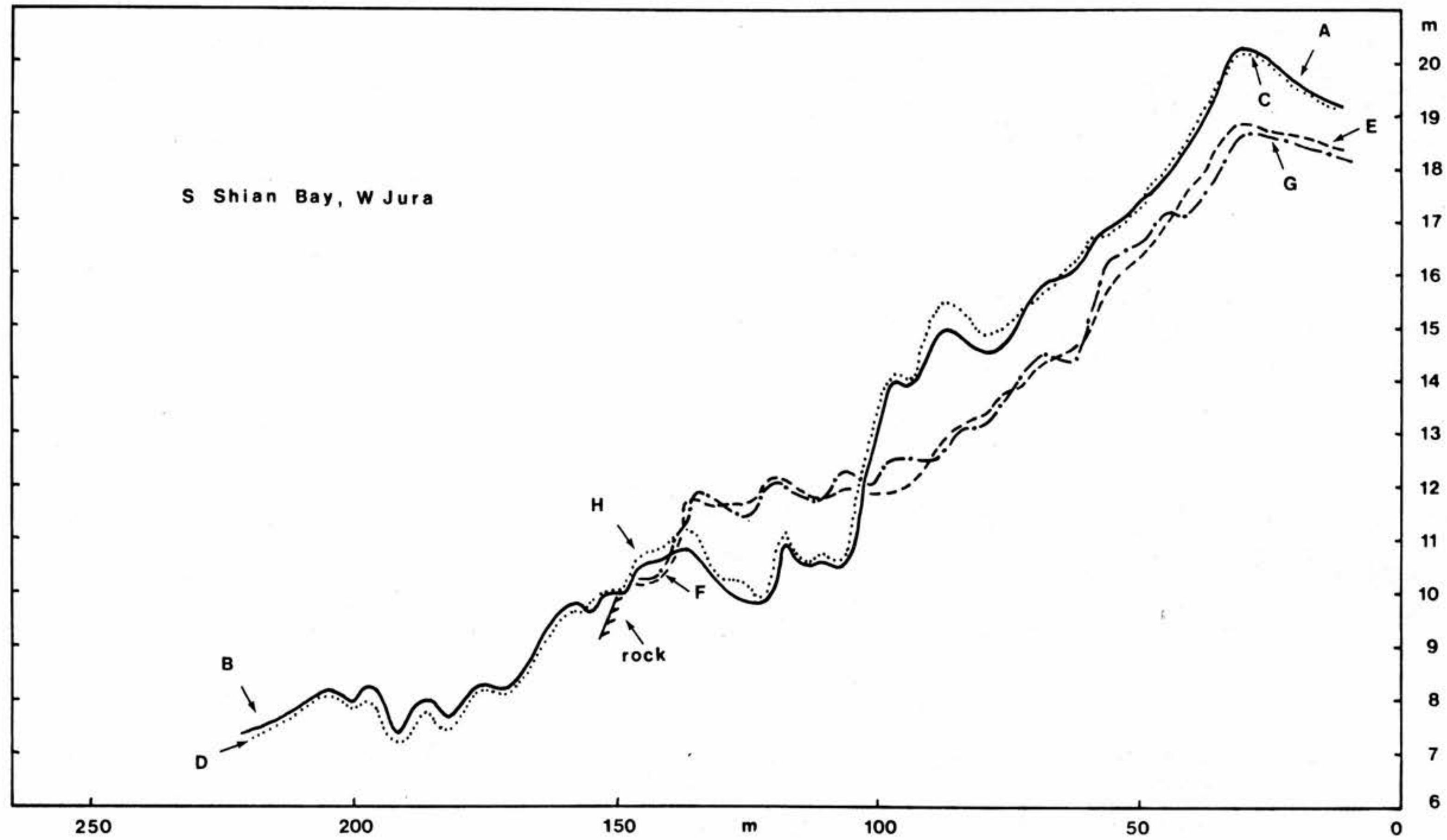


Fig. 53 Postglacial shingle profiles, S Shian Bay, W Jura. Profile locations are shown on Fig. 34.

Corpach Bay, NW Jura, a series of 12 unvegetated raised beach ridges descends seaward from 13.8 m to the present shore where the modern storm ridge crest occurs at 5.5 m (Fig. 54). The shingle staircase is characterised by three large ridges (R2, R7 and R10) that are separated from each other by a series of lower amplitude ridges and swales. The three large ridges possess crest altitudes of 13.8 m, 10.7 m and 8.9 m respectively (Table 24). Of the three ridges, R2 is the most conspicuous and is separated from the backing cliff by lower shingle deposits. The raised shingle overlies the surface of the Main Rock Platform whose altitude is here 3.5 m.

At Bhrein Port north of Ruantallain (NR 507839) a series of unvegetated raised shingle ridges extends seaward from the base of the Main Rock Platform cliff, which is here at c. 2.5 m. In this area four well-developed raised beach ridges occur, the most conspicuous being R1 and R2, whose crests are at 12.8 m and 12.5 m (Fig.55; Table 24).

At the mouth of Glenbatrick a degraded cliffline composed of till and lateglacial raised marine sediments is separated from the present shore by a wide area of raised postglacial marine deposits. At the foot of and parallel to the cliffline is an arcuate raised vegetated beach ridge 370 m in length. The crest altitude of the ridge varies between 9.3 m and 10.4 m, the altitude at the base of its frontal slope being 6-7 m. Here the altitude of the Main Rock Platform is c. 2 m.

In SW Jura north of Loch na Sgrioba the highest postglacial beach ridge crests are between 9.3 m and 10.4 m. Farther north at the S end of Lang Aoinidh (NR 480794) (Fig. 55; Table 24; Plate 31) a series of 5 raised beach ridges extends from the base of the Main

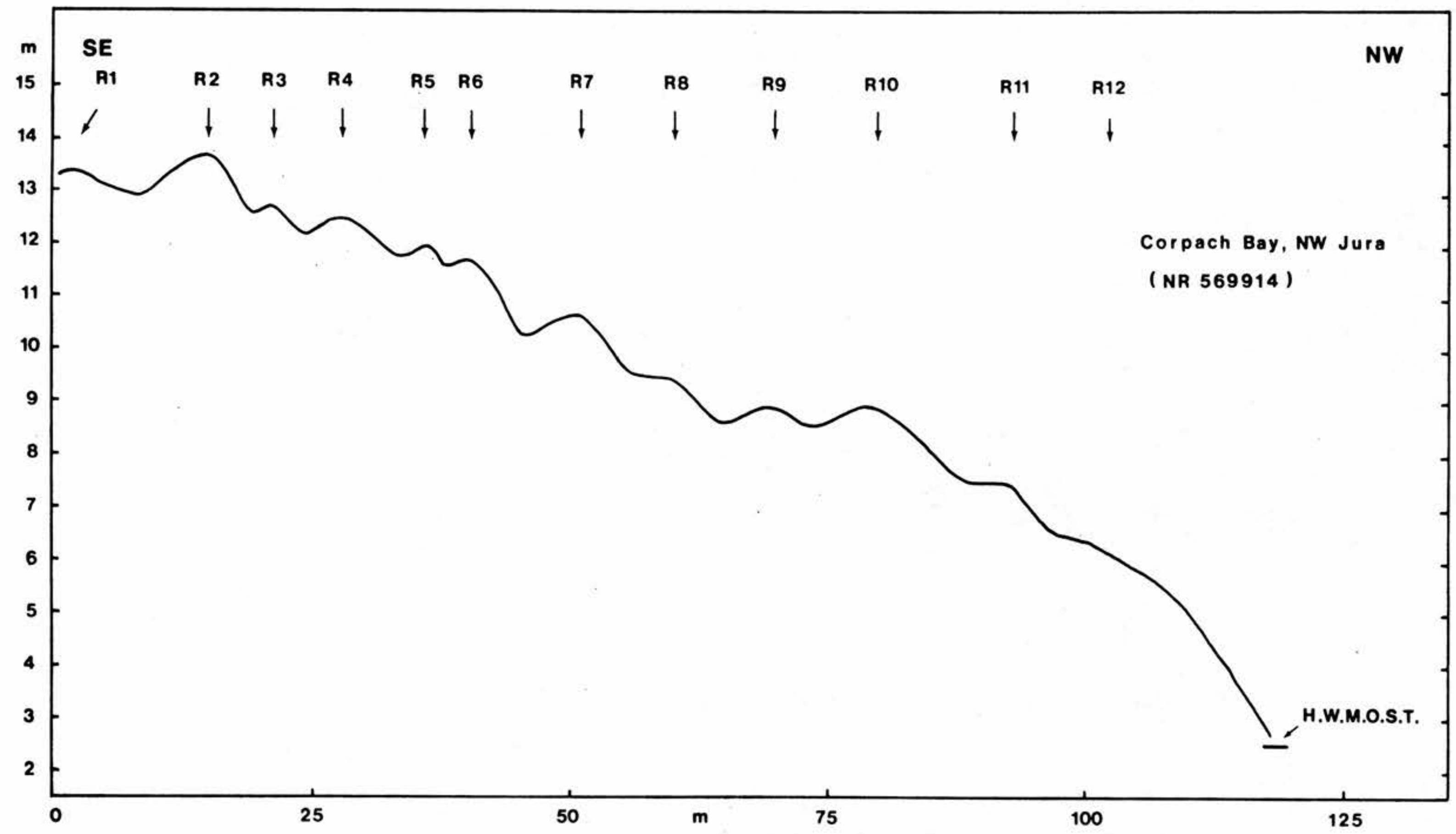


Fig. 54 Postglacial shingle profile, Corpach Bay, NW Jura.

Table 24

Postglacial ridge and swale altitudes and amplitudes, Corpach Bay, S Lang Aoinidh, N Bhrein Port and Rudha Chrois, W Jura.

Corpach Bay, NW Jura

S	13.38	*0.76
R	14.14	-1.09
S	13.05	+0.13
R	13.18	-0.53
S	12.65	+0.34
R	12.99	-0.81
S	12.18	+0.28
R	12.46	-0.43
S	12.03	+0.11
R	12.14	-1.61
S	10.53	+0.61
R	11.14	-1.08
S	10.06	(-0.17)
R	9.89	-0.81
S	9.08	+0.29
R	9.37	-0.32
S	9.05	+0.37
R	9.42	-1.45
S	7.97	(-0.08)
R	7.89	-0.80
S	7.09	(-0.17)
R	6.92	-0.93
S	5.99	

S Lang Aoinidh, SW Jura

S	5.90	(-0.32)
R	5.58	veg
S	veg	veg
R	6.44	-0.43
S	6.01	+0.41
R	6.42	-0.72
S	5.70	+0.96
R	6.66	-1.11
S	5.55	+0.31
R	5.86	-0.75
S	5.11	

N Bhrein Port, W Jura

S	11.70	+1.15
R	12.85	-0.63
S	12.22	+0.22
R	12.44	veg
S	veg	(-1.63)
R	10.81	-0.84
S	9.97	+0.04
R	10.01	-0.91
S	9.10	+0.08
R	9.18	-1.09
S	8.09	(-0.95)
R	7.14	

Rudha Chrois, W Glenbatrick, SW Jura.

S	9.47	+0.61
R	10.08	-0.82
S	9.26	(-0.64)
R	8.62	veg
S	veg	+0.12
R	8.74	-0.72
S	8.02	+0.06
R	8.08	-0.37
S	7.71	(-0.03)
R	7.68	-1.04
S	6.64	+0.08
R	6.72	-0.76
S	5.96	(-0.66)
R	5.30	

* Shingle profiles shown on Figs. 53 and 55.

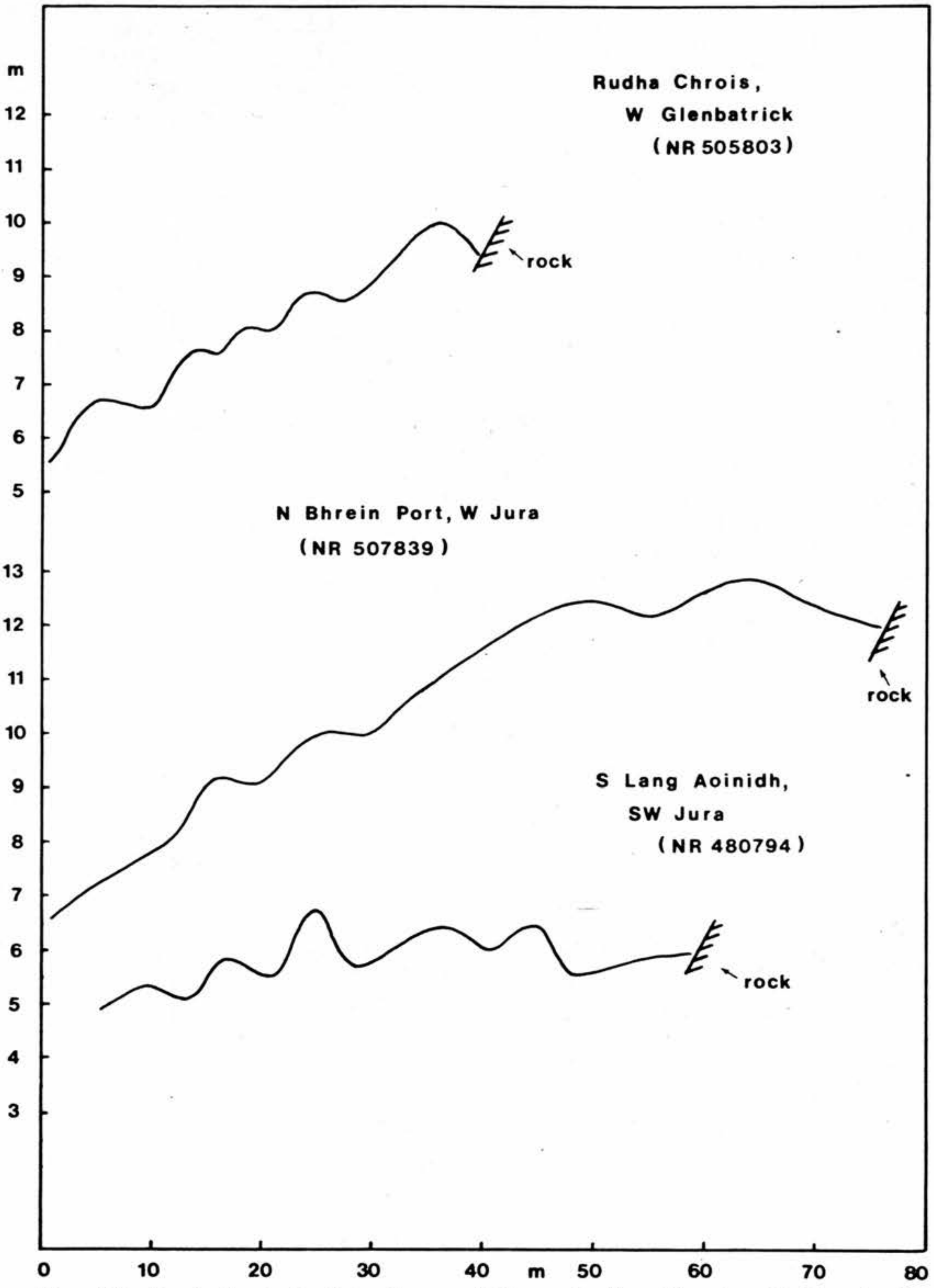


Fig. 55 Postglacial shingle profiles: Rudha Chrois, N Bhrein Port and S Lang Aoinidh, W Jura.

Rock Platform cliff to the present shore. However, there is no evidence of higher postglacial ridges such as occur elsewhere, the raised ridge crest altitudes ranging between 6.7 m and 5.3 m. North of Lang Aoinidh at Rudha Chrois (NR 505803) a series of 5 raised shingle ridges extends from the Main Rock Platform cliff to the present shore (Fig. 55; Table 24). The ridges are separated by well-defined swale depressions and range in crest altitude from 10 m to 6.7 m.

The most remarkable sequence of raised postglacial beach ridges lies north of Inver, SW Jura, where 31 unvegetated raised beach ridges descend from 12.3 m to the modern storm beach ridge (Fig.56, Table 25). The largest ridge of the staircase, R1, protrudes markedly above the sloping shingle surface. Below R1 a series of low amplitude shingle ridges descends seaward and is interrupted by two larger ridges (R5 and R16), both approximately 1 m in amplitude (Fig. 56), whose crest altitudes are at 9.2 m and 6.9 m. The most steeply sloping section of the shingle profile occurs between the modern storm ridge crest (4.5 m) and high water mark.

Altitude Analysis

The altitudes of the 8 measured raised terrace fragments were plotted on a height-distance diagram (Fig. 57) and show a gradual decline in altitude SW from 10.3 m in N Jura to 8.5 m in NE Islay. Since the terrace fragments described above are the only such features at each location where they occur, they are believed to correlate. The shoreline altitudes were analysed by linear regression and indicate a regional gradient of 0.05 m/km away from the centre of glacio-isostatic uplift. The crest altitudes of all raised beach ridges formed during the main postglacial transgression were also plotted

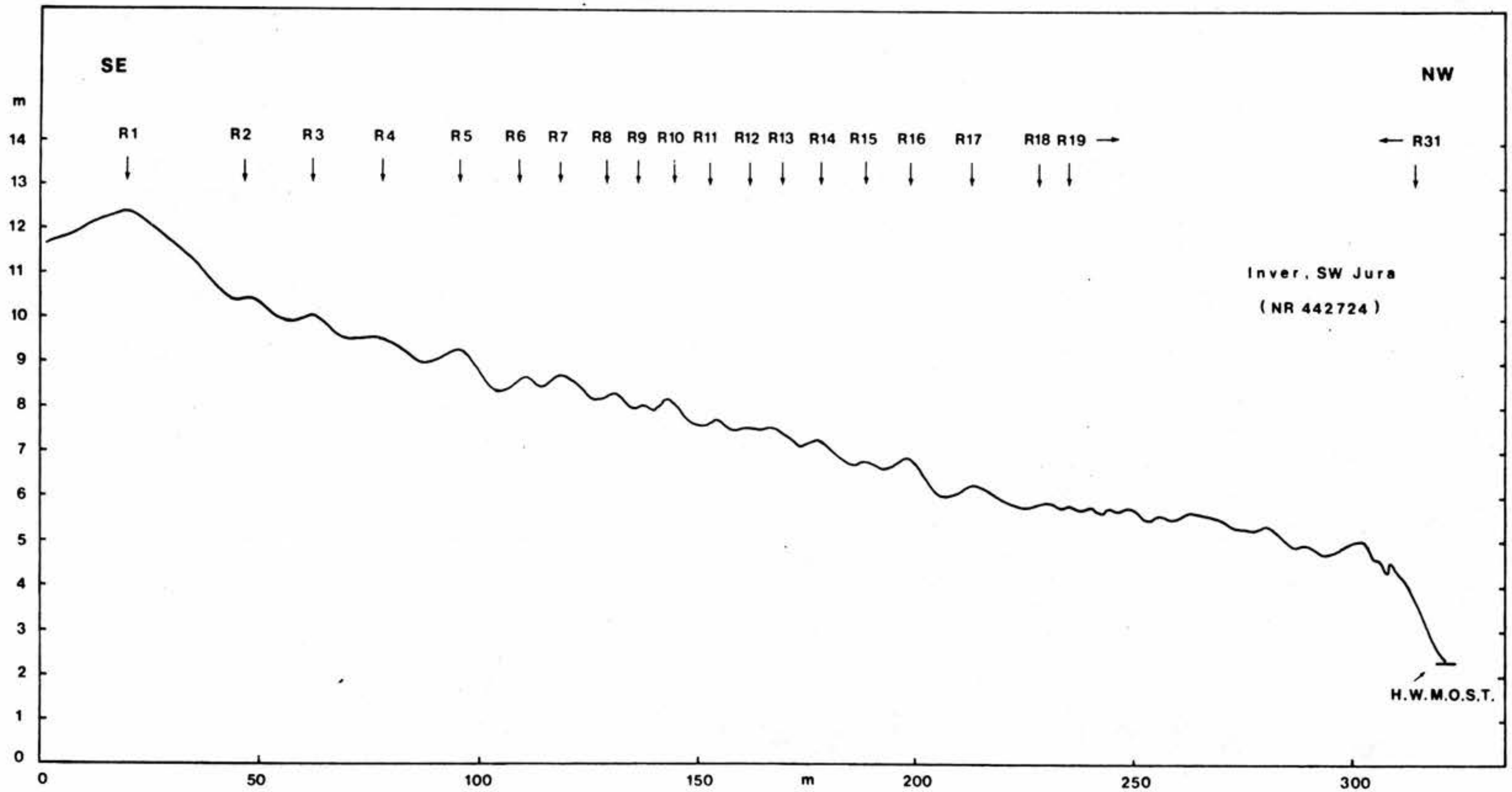


Fig. 56 Postglacial shingle profile, Inver, SW Jura. Location of raised beach ridges shown on Fig.35.

Table 25

Postglacial ridge and swale altitudes and amplitudes, Inver, SW Jura.

R	12.32	-0.53
cx	11.79	-1.44
S	10.35	+0.02
R	10.37	-0.41
S	9.96	+0.07
R	10.03	-0.53
S	9.50	(-0.08)
R	9.42	-0.43
S	8.99	+0.24
R	9.23	-0.86
S	8.37	+0.29
R	8.66	-0.20
S	8.46	+0.24
R	8.70	-0.55
S	8.15	+0.10
R	8.25	-0.30
S	7.95	+0.16
R	8.11	(-0.12)
S	7.99	+0.20
R	8.19	-0.61
S	7.58	+0.13
R	7.71	-0.22
S	7.49	+0.02
R	7.51	0.00
S	7.51	(-0.05)
R	7.46	-0.29
S	7.17	+0.08
R	7.25	-0.48
S	6.77	0.00
R	6.77	-0.12
S	6.65	+0.18
R	6.83	-0.82
S	6.01	+0.23
R	6.24	-0.49
S	5.75	+0.13
R	5.88	-0.09
S	5.79	(-0.02)
R	5.77	0.00
S	5.77	(-0.02)
R	5.75	-0.10
S	5.65	+0.07
R	5.72	-0.02
S	5.70	+0.04
R	5.74	-0.37
S	5.47	+0.08
R	5.55	-0.04
S	5.51	+0.09
R	5.60	-0.08
S	5.52	(-0.16)
R	5.36	-0.08
S	5.28	+0.05
R	5.33	-0.43
S	4.90	+0.04
R	4.94	-0.21
S	4.73	+0.15
R	4.88	-0.09
S	4.79	+0.29
R	5.08	-0.41
S	4.67	-0.12
R	4.55	-0.16
S	4.39	+0.14
R	4.53	

* location of Inver shingle ridges shown on Fig. 35.

* **shingle profile shown on Fig.56**

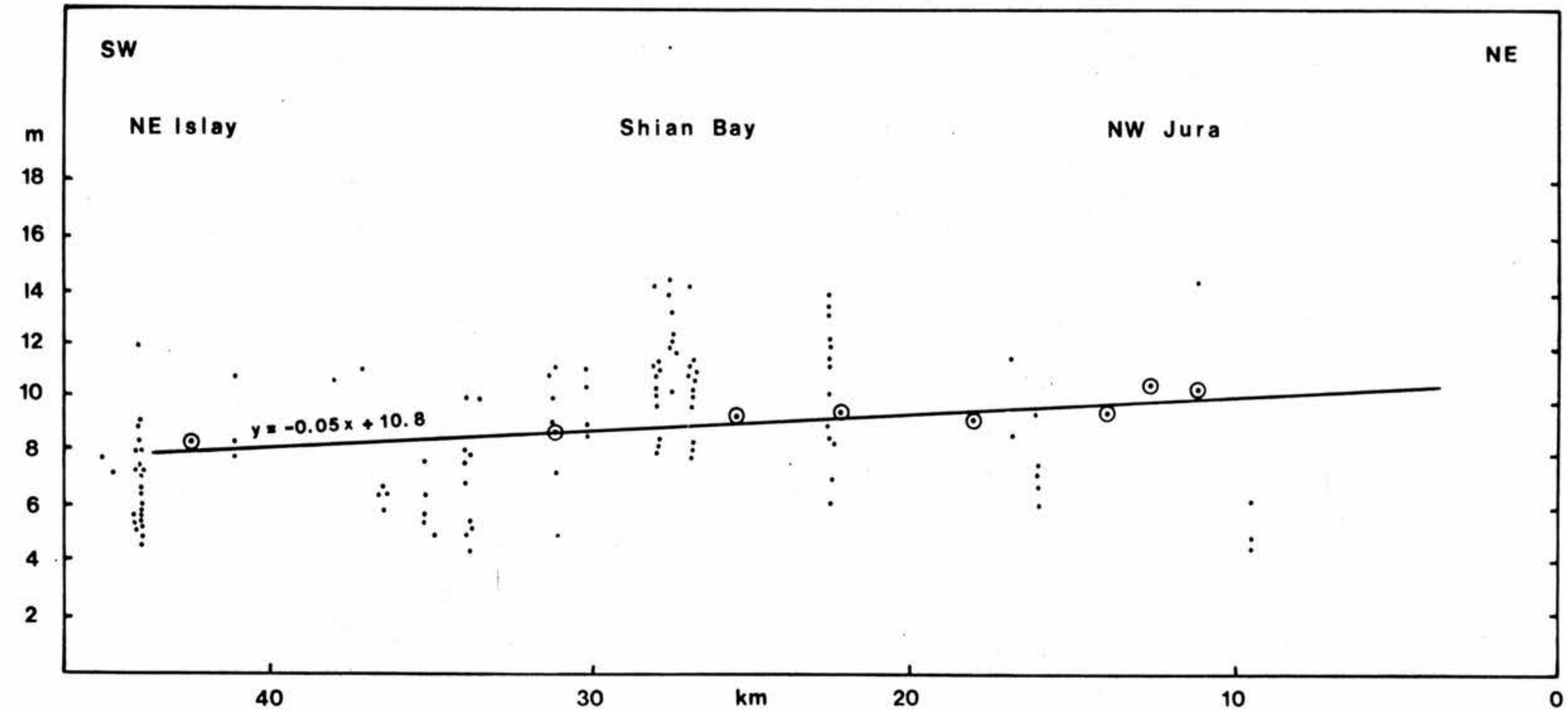


Fig. 57 Linear regression of Main Postglacial Shoreline fragments (circles). Postglacial ridge crests formed during and after the Main Postglacial Shoreline are also shown.

on a shoreline height-distance diagram in order to investigate regional altitude variations and altitude relationships to former sea-level. Thereafter the altitudes above former sea-level of these ridges were compared with those of modern storm ridges above present sea-level. In addition the crest altitudes of ridges formed after the culmination of the main postglacial transgression were plotted on a height-distance diagram to see if any meaningful patterns of ridge crest altitudes appeared.

Interpretation

The calculated regional gradient of the Main Postglacial Shoreline (0.05 m/km) can be compared with the value of 0.07 m/km obtained farther north in Lorn and E Mull by Gray (1974b) and values of 0.08 m/km and 0.09 m/km obtained in the Forth and Tay valleys by Sissons et al. (1966) and Cullingford (1972). The low shoreline gradient of 0.05 m/km is most probably the result of the small number of measured terrace fragments used in the linear regression.

The crest altitudes of raised beach ridges formed during the culmination of the main postglacial transgression show that most of the highest ridges are between Ruantallain and Corpach Bay, W Jura, and most probably reflect the exposure of this area to the greatest amount of open Atlantic fetch (Chapter 5). Since it has been shown that the inner edge of modern beach terraces (the lower limit of land-based vegetation) corresponds to present H.W.M.O.S.T. at approximately 2.4 m the calculated altitudes of the Main Postglacial Shoreline derived from linear regression probably correspond with the position of former H.W.M.O.S.T. In order to compare the crest altitudes of ridges formed during the culmination of the main postglacial transgression with those of modern beach ridges it was therefore necessary to convert modern beach ridge crest altitudes to values above present H.W.M.O.S.T. Consequently 2.4 m was subtracted from each modern beach ridge crest altitude. Thereafter both groups of ridge crest altitudes were plotted for comparison on a height-distance diagram (Fig. 58).

The crest altitudes of most ridges formed during the culmination of the main postglacial transgression vary between 0.6 and 4.6 m above former H.W.M.O.S.T. For example the highest postglacial ridge at Corpach Bay is at 13.8 m (i.e. 4.5 m above postglacial high water mark) while in the same area the modern storm ridge reaches 3.1 m above H.W.M.O.S.T. In contrast at Bhrein Port, N of Ruantallain, the highest postglacial ridge (10.1 m) occurs 1.2 m above former high water mark although the modern storm ridge crest is at 2.9 m above present high water mark. The marked regional variations in ridge crest altitudes shown on Fig. 58 most probably reflect variations in exposure to open Atlantic fetch and the availability of nearshore debris for ridge formation. However the highest modern and postglacial ridges often occur at similar locations, a factor suggesting that the processes responsible for shingle ridge formation have altered little since the culmination of the main postglacial transgression.

At Shian Bay raised shingle ridges extend continuously from the Colonsay ridge (c. 20 m) to the present shore (Fig. 53): hence the highest postglacial ridge cannot easily be identified. However, since the maximum altitude of the modern storm ridge in this area is 5.3 m above H.W.M.O.S.T. (i.e. 7.7 m) and the local altitude of the Main Postglacial Shoreline is 9.1 m, it may be inferred that postglacial storm activity here reached c.14.4 m (i.e. 9.1 + 5.3 m). Levelled

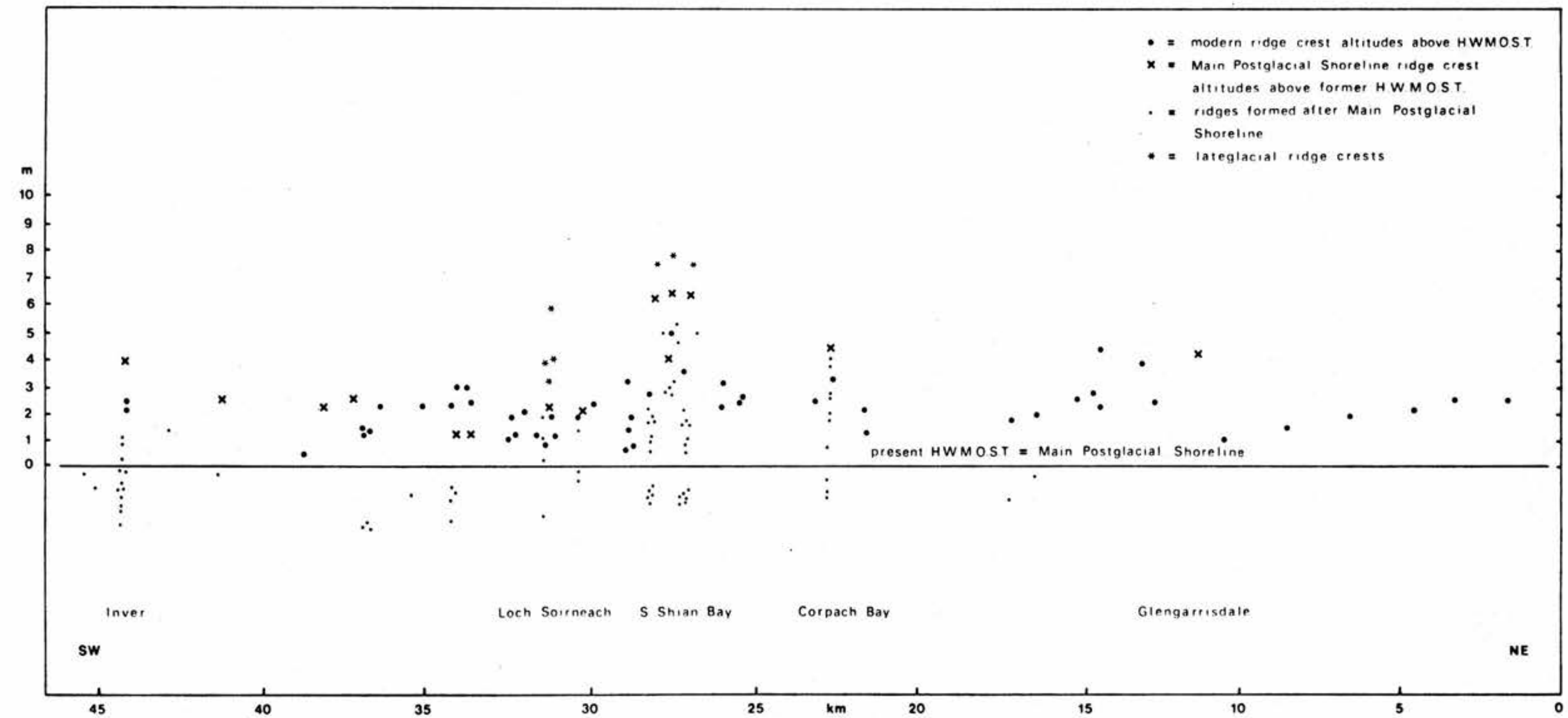


Fig. 58 Height-distance diagram of Main Postglacial Shoreline ridge crests, modern storm ridge crests and certain lateglacial ridge crests adjusted to altitudes above present and former H.W.M.O.S.T.. Postglacial ridge crests formed after the Main Postglacial Shoreline are also shown.

profiles across the Shian Bay ridges indicate the presence of a well-developed ridge that varies in altitude between 14.6 m and 15.6 m. It is therefore proposed that at Shian Bay postglacial storm activity resulted in the formation of this ridge and therefore wave activity during this period reached a maximum altitude of c. 15.6 m.

The Loch Soirneach sea-level oscillation, S. Jura.

Loch Soirneach occupies a structurally-controlled depression on the southern coast of Loch Tarbert (Plate 29) and is located in a sheltered embayment between two large quartzite ridges. The Loch surface is at 9.5 m and is dammed seaward by an elongate quartzite ridge that reaches a maximum altitude of 9.6 m (Chapter 9). The loch margin is located 150 m landward of the rock bar and is separated from it by a wide area of undulating ridge and swale topography. A staircase of 11 well-developed raised shingle ridges rises seaward in altitude from the loch margin to 14.9 m (Fig. 49). Seaward of ridge R11 a series of ridges and swales descends in altitude to the modern storm beach whose driftwood-mantle crest is at 4.8 m (i.e. 2.4 m above H.W.M.O.S.T.).

At Loch Soirneach the altitude of the Main Postglacial Shoreline derived from regression analysis is 9 m. Hence it is probable that during the culmination of the main postglacial transgression the loch was connected to the sea only during periods of high tide. Since ridges R7 to R10 are located between 50 m and 115 m landward of the rock bar it is proposed that they could not have formed during a period when sea-level was at 9 m. Moreover the seaward rise in altitude of ridges R1-R10 strongly suggests that they were formed prior to the formation of R11 during a period of rising sea-level.

For example the altitude of the swale seaward of ridge R9 (Fig.49) suggests that a sea-level of c.12 m was necessary for the formation of ridge R9. In addition, since the modern storm ridge crest is at 2.4 m above H.W.M.O.S.T., it is unlikely that ridge R11 (14.9 m) was formed during the culmination of the main postglacial transgression when sea-level was at 9 m. The inability of storm waves during the maximum of the main postglacial transgression to form high ridges along the sheltered coastline of Loch Tarbert is further demonstrated at neighbouring Glenbatrick (NR 520800) (more exposed to wave attack than Loch Soirneach) where the highest postglacial ridge is at 10.4 m. As a result it is inferred that the Loch Soirneach raised beach ridges landward of and including ridge R11 are not postglacial in age. This implies the following sequence of relative sea-level changes.

At Loch Soirneach the relative regression of the lateglacial sea halted when sea-level was at 9 m. During this period Loch Soirneach was sealed from the sea by the eastward progradation of 5 low amplitude shingle spits (Chapter 9). The formation of the spits was followed by a period of rising sea-level and the deposition of ridges R6 to R11. During the formation of ridge R10 sea-level had risen to c. 12 m. Thereafter relative sea-level declined in altitude and was accompanied by the formation of ridges R11-R13. During the Loch Lomond Stadial sea-level fell to c. 2 m and resulted in the formation elsewhere along the W Jura coast of the Main Rock Platform. Thereafter sea-level rose to 9 m at Loch Soirneach at the culmination of the main postglacial transgression.

The absence elsewhere in the study area of evidence for a similar lateglacial sea-level oscillation is due to the presence at these altitudes of the Main Rock Platform cliff, that, according to the view presented in Chapter 8, was formed during the Loch Lomond Stadial after the proposed sea-level oscillation.

Since there are no well-developed raised terrace fragments in the study area below the Main Postglacial Shoreline it is impossible to identify any later widespread shorelines. However, the Inver shingle ridges form a useful indicator of sea-level change during this period (Fig. 56). Here the culmination of the main postglacial transgression was marked by the formation of a large beach ridge (R1) (12.3 m), unaccompanied by the formation of any adjacent terrace. Thereafter a decline in relative sea-level resulted in the construction of 30 undulating ridges separated by swales. Short halts in the general retreat of the postglacial sea may be indicated by two ridges (R5 and R16), that protrude above the sloping shingle surface and have crest altitudes of 9.2 m and 6.9 m. Unfortunately due to shingle deposition on the seaward sloping faces of these ridges it is impossible to determine accurately the positions of former sea-levels during their formation. However the pattern of Inver shingle ridge altitudes strongly suggests that no major sea-level oscillation occurred after the formation of the Main Postglacial Shoreline.

Although the Main Postglacial Shoreline is the most widely developed raised shoreline in western Scotland, the source of the large volume of debris required for its formation has never been satisfactorily explained. It is suggested here that some of the debris removed seaward during the formation of the Main Rock Platform was later returned landward by constructive waves of transgressive postglacial seas. In Scarba and NW Jura the absence of postglacial deposits on the Main Rock Platform surface is probably a result of the presence in the nearshore zone of deep water (Chapter 5) that favoured destructive wave activity and the seaward removal of any overlying debris. Farther south, however, the presence offshore of relatively shallow water depths favoured constructive wave activity and resulted in the deposition of large volumes of raised beach debris. In SW Jura local erosion of till deposits in conjunction with the presence of a shallow nearshore zone provided a rich supply of debris available for postglacial deposition. As a result the most widespread accumulations of postglacial debris occur in this area. Nevertheless, despite a locally rich debris supply the raised postglacial shingle ridges are here no higher than those encountered elsewhere.

Summary

It is suggested that the shingle accumulations at Loch Soirneach were deposited prior to the Loch Lomond Stadial and were associated with an oscillation of sea-level between 9 and 12 m. The cause of this sea-level change is not known. After this period sea-level fell and the Main Rock Platform was formed during the cold climatic conditions of the Loch Lomond Stadial. During this period sea-level was approximately 7 m in N Scarba and declined in altitude SW to present sea-level in NE Islay and Colonsay. Coastal erosion during this period resulted in the destruction of earlier elevated lateglacial shorelines. No evidence has been found for raised or buried shorelines equivalent to the High, Main and Low Buried Shorelines of SE Scotland. However raised beach ridges formed during the culmination of the main postglacial transgression are widespread. The Main Postglacial Shoreline declines in altitude to the SW from 10.2 m in N Jura to 8.5 m in NE Islay and has a regional gradient of 0.05 m/km. The debris supply available for the development of the postglacial ridges and terraces consists partly of material previously eroded during the formation of the Main Rock Platform and later returned landward by constructive wave action. However in SW Jura, postglacial marine erosion of till cliffs also provided an important source of debris. Raised beach ridges formed during the culmination of the main postglacial transgression reach maximum altitudes of between 13.8 m and 15 m between Corpach Bay and Shian Bay, an area exposed to open Atlantic fetch. The altitudes of these high postglacial ridges above the Main Postglacial Shoreline are approximately equivalent to the altitudes of modern storm ridge crests above H.W.M.O.S.T. Altitude measurement of the Inver shingle ridges indicates that no major sea-level oscillation occurred after the formation of the Main Postglacial Shoreline.

Chapter 11

Conclusion

During the last period of general glaciation N Jura and Scarba were over-ridden by westward-moving mainland ice. In S Jura and NE Islay the dominant movement of mainland ice was from SE to NW. It is believed that the convergence in ice-flow direction was due to the presence of a major ice dome over N Ireland. The glacially-overdeepened trenches of the Sound of Jura and the Firth of Lorn, which trend NE-SW, do not relate to the former westward flow of ice. Instead it is more likely that these submerged trenches were glacially overdeepened by SW-moving ice during the build-up and thinning of each Quaternary ice-sheet.

During the Hoxnian interglacial temperate marine erosion during a long period of relative sea-level stability at c. 32-35 m resulted in the widespread formation of the High Rock Platform (Fig. 59). Subsequent tectonic movements caused dislocation and warping of this shoreline while later glacial erosion resulted in its removal from most coastal areas, the distribution of High Rock Platform fragments being today limited to NE Islay, Colonsay and W Jura (between Ruantallain and N Shian Bay).

During the later Ipswichian interglacial a second period of relative sea-level stability resulted in the formation of the Low (intertidal) Rock Platform. This platform reaches a maximum width of 1 km and in SW Jura passes beneath thick accumulations of till. The platform is unaffected by folding or faulting and is regionally horizontal throughout Jura, Scarba, NE Islay and neighbouring Colonsay and Oronsay at 1-2 m.

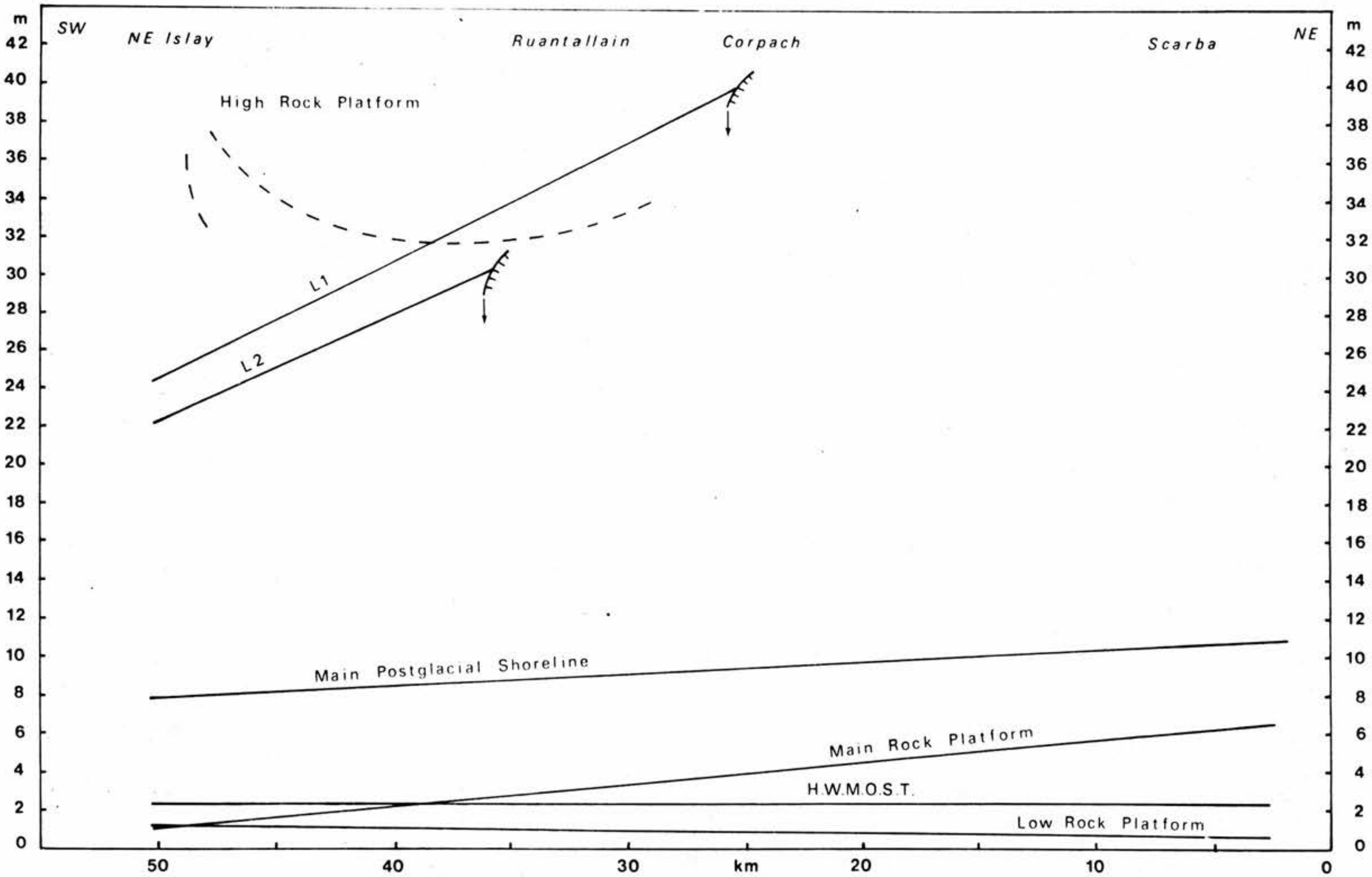


Fig. 59 The shoreline sequence in Jura, Scarba and NE Islay.

At the close of the last (Devensian) glaciation the coalescence of two westward moving ice-streams in the western Paps of Jura resulted in the formation of a medial moraine 3.5 km in length, while in NE Islay a large terminal moraine was formed on the surface of the High Rock Platform. McCann's (1964) proposal that the moraine was formed during the Highland Readvance (generally considered equivalent to the Loch Lomond Readvance in the Western Highlands) is rejected since the seaward face of the moraine is truncated by an early lateglacial shoreline.

The first areas to be deglaciated during the decay of the late-Devensian ice-sheet were Colonsay, Oronsay, NE Islay and W Jura (between Ruantallain and Corpach Bay). Marine incursion into deglaciated areas resulted in the formation of a high coastal terrace eroded in pre-existing drift. The associated shoreline (L1) declines SW in altitude from 40 m at Corpach Bay to 26.5 m in NE Islay with a regional gradient of 0.61 m/km. This steep gradient is similar to that of the Kinkell Shoreline in E Scotland (Cullingford and Smith, 1966) and suggests that both areas were deglaciated approximately contemporaneously. Slightly later SW Jura was also deglaciated and permitted the formation here of a second high lateglacial shoreline (L2) that slopes SW with a regional gradient of 0.55 m/km. Continued ice-decay resulted in the incursion of the lateglacial sea into NW and E Jura and allowed the formation of high coastal terraces in these areas. Relative marine regression of the lateglacial sea caused by glacio-isostatic uplift resulted in the deposition of the W Jura shingle ridges on the surface of the High Rock Platform, the drift deposits on which provided a rich supply of debris. Owing to the complete submergence of the W Jura High Rock Platform by the lateglacial sea marine erosion resulted in its exhumation from beneath a cover of glacial deposits and its eventual burial by raised beach ridges. In NE Islay, owing to the gradients of the early lateglacial shorelines, wave action only occurred on the seaward surface of the High Rock Platform and consequently raised shingle spreads equivalent to those of W Jura are absent.

The regional distribution of the W Jura shingle ridge crest altitudes indicates that, with the exception of ridges above 35 m, most lateglacial beach ridges were formed by moderate wave activity. The pattern of beach ridge altitudes in the sheltered Loch Aoinidh Dhuibh embayment suggests that the relative marine regression of the lateglacial sea was interrupted by two short phases of sea-level stability that here occurred at 32-33.5 m and 27.8-28.6 m. In this area the beach ridge altitudes indicate that although relatively brief marine stillstands occurred during this period there were no major sea-level oscillations. However at Loch Soirneach there is evidence to suggest that a sea-level oscillation occurred when relative sea-level had fallen to 9-12 m. Thereafter sea-level was relatively low during the Loch Lomond Stadial.

During the Loch Lomond Stadial severe periglacial conditions occurred throughout Jura, Scarba and NE Islay and in the Paps of Jura a fossil lobate rock glacier was produced. It is proposed that the widespread accumulations of talus that flank many of the Jura hills are of the same age, the result of severe frost-riving. The absence of evidence of former valley glaciers in Jura indicates that Charlesworth's (1955) interpretation that 19 valley glaciers developed during this period is erroneous.

During the Loch Lomond Stadial polar shore erosion by frost resulted in the formation of the Main Rock Platform and the removal seaward of large quantities of debris. Coastal erosion during this period also resulted in the erosion of the seaward surface of the High Rock Platform and any overlying lateglacial shorelines. The Main Rock Platform declines in altitude from 7 m in N Scarba to sea-level in NE Islay and is tilted to the SW with a regional gradient of 0.13 m/km. It is part of the same feature described farther north and east by Gray (1974a, 1978) and probably also correlates with the Buried Gravel Layer of SE Scotland and the Norwegian Main Line (Younger Dryas). The pattern of shoreline isobases suggest an uplift centre to the NE. During this period a minimum volume of 1.03 m^3 of rock per metre of coast per year was removed from the W Jura coast, equivalent to an average cliff retreat rate of 7 cm/year. It has been shown that the Main Rock Platform passes beneath present sea-level within the study area: hence the traditional correlation between the Main Rock Platform and low rock platforms in the Outer Hebrides (von Weymarn, 1974) and Ireland (Stephens 1957) is erroneous. It is proposed instead that the low interglacial platforms of Ireland and the Outer Hebrides are most likely of the same age as the Low Rock Platform.

It follows that polar cliff retreat and platform development took place extremely rapidly. In addition it is suggested that polar shore platforms have characteristic morphological features that distinguish them from those formed in non-polar areas. Firstly, since frost action is the most important process affecting cliff retreat and platform development, shore platforms are exceptionally well-developed in areas of restricted fetch. Secondly, since the role of waves in polar platform development is primarily as an agent of transportation rather than of erosion, polar shore platforms lack many of the features that characterise non-polar platforms formed by marine abrasion. In particular polar platform surfaces are extremely angular and do not exhibit the ramp abrasion profiles observed in non-polar areas.

During the main postglacial transgression some of the debris previously eroded during the formation of the Main Rock Platform was returned landward by constructive wave action. During this period marine activity resulted in the formation of the Main Postglacial Shoreline, which declines in altitude SW from 10.3 m in N Jura to 8.5 m in NE Islay and has a regional gradient of 0.05 m/km. Raised beach ridges formed during the main postglacial transgression reach maximum altitudes of c. 15 m along the W Jura coast. The altitude distribution of these high postglacial ridges above the Main Postglacial Shoreline is also similar to those of modern storm ridge crests above present H.W.M.O.S.T. In SW Jura, north of Inver, raised beach ridges formed after the main postglacial transgression indicate that no major sea-level oscillations occurred after this period.

In order to understand more clearly the origin of raised coastal terraces and beach ridges and their altitude relationship to former sea-levels a limited study of modern coastal landforms was undertaken. The results indicate that the lower limit of land-based vegetation, H.W.M.O.S.T., and the seaweed Pelvetia canaliculatus (sp.) occur at similar altitudes. It is calculated that the inner edges of raised coastal terraces correspond to the position of former high water mark, present H.W.M.O.S.T. being 2.4 m.

Two types of modern beach ridge have been identified. Firstly, storm beach ridges are formed by the combined action of erosion and deposition during and after storm wave activity and all occur in the backshore zone above high water mark. Shingle moved seaward by destructive waves during storms is eventually returned to the intertidal zone by constructive wave action. Consequently low amplitude beach ridges are formed at or near low water mark.

The absence of staircases of modern storm beach ridges and swales (each modern ridge occurring singly) suggests that in the recent past storm ridges have been continually destroyed and reformed by wave action. Exposure to open fetch is an important factor affecting the regional distribution and altitudes of modern storm ridges. An index of wave energy potential derived from modified wind and fetch variables also provides a good explanation.

References

Aarseth, I. and Mangerud, J. 1974 Younger Dryas end moraines between Hardanger-fjorden and Sognefjorden, Western Norway. Boreas, 3, 3-22.

Ahlmann, H.W. 1919 Geomorphological studies in Norway. Geogr. Annlr., 1, 3-148 & 193-252.

Andersen, B.G. 1968 Glacial geology of western Troms, North Norway. Norges geol. Unders., 256, 160pp.

Anderson, J. 1888 Glaciation and Raised Beaches in Jura and Islay. Trans. geol. Soc. Glasgow. 8, 316-20.

Anderton, R. 1976 Tidal-shelf sedimentation: an example from the Scottish Dalradian. Sedimentology, 23, 429-58.

Andrews, J.T., C.A.M. King and M. Stuiver, 1973 Holocene sea-level changes, Cumberland Coast, Northwest England: Eustatic and Glacio-Isostatic movements. Geol. en Mijnbouw, 52, 1-12.

Araya, R. and Hervé, F. 1972 Patterned Gravel Beaches in the South Shetland Islands. in "Antarctic Geology and Geophysics" R.J. Adie (ed.) Oslo, 111-4.

Bailey, E.B., C.T. Clough, W.B. Wright, J.E. Richey and G.V. Wilson, 1924 Tertiary and post-Tertiary geology of Mull, Loch Aline and Oban. Mem. geol. Surv. Gt. Br.

Bailey, E.B. and Anderson, E.M. 1925 The geology of Staffa, Iona and western Mull. Mem. geol. Surv. Gt. Br.

Baird A.K., K.W. Baird and D.M. Morton. 1971 On deciding whether Trend Surfaces of Progressively Higher Order are meaningful: discussion. Geol. Soc. Amer. Bull. 82, 1219-34.

Barsch, D. 1969 Studien und Messungen an Blockgletschern in Macun, Unterengadin. Z. Geomorph. Supp. 8, 11-30.

Bartrum, J.A. 1938 Shore platforms- Discussion. J. Geomorph. 1, 266-8.

Benedict, J.B. 1973 Origin of rock glaciers J. Glaciol. 12, 520-2

Binns, P.E., R. Harland and N Kenolty, 1973 The geology of the Sea of Hebrides. Rep. No. 73/14, Inst. Geol. Sciences.

Binns, P.E., R. Harland and M.J. Hughes. 1974 Glacial and post-glacial sedimentation in the Sea of the Hebrides. Nature, 248, 751-4.

Bird, J.B. 1967 The Physiography of Arctic Canada. Baltimore, John Hopkins Press.

Calkin, P.E. and Nichols, R.L. 1972 Quaternary studies in Antarctica. in "Antarctic geology and geophysics" R.J. Adie (ed.) Oslo, 625-43.

Challinor, J. 1949 A principle in coastal geomorphology. Geography, 34, 213-5.

Charlesworth, J.K. 1955 The Late-glacial history of the Highlands and Islands of Scotland. Trans. Roy. Soc. Edinb. 62, 769-928.

Chayes, F. 1970 On deciding whether trend surfaces of progressively higher order are meaningful. Bull. geol. Soc. Amer., 81, 1273-8.

Cornish, V. 1898 On Sea-beaches and Sandbanks. Geogr. J., 5, 528-43 & 6, 629-51.

Craig, E.H., W.B. Wright and E.B. Bailey, 1911 The geology of Colonsay and Oronsay, with part of the Ross of Mull. Mem. geol. Surv. Gt. Br.

Cullingford, R.A. 1972 Late-glacial and post-glacial shoreline displacement in the Earn-Tay area and eastern Fife. Unpubl. Ph.D. thesis, Univ. of Edinburgh.

Cullingford, R.A. 1977 Lateglacial Raised Shorelines and deglaciation in the Earn-Tay area. in "Studies in the Scottish Lateglacial Environment", J.M. Gray and J.J. Lowe (eds.) Pergamon, 15-32.

Cullingford,R.A. and Smith, D.E. 1966 Late-glacial shorelines in eastern Fife. Trans. Inst. Br. Geogr. 39, 31-51.

Dahl, E. 1947 On the origin of the strand flat. Norsk. geogr. Tidsskr., 11, 159-71.

Davies, J.L. 1961 Tasmanian Beach ridges in relation to sea-level change. P. and P. roy. Soc. Tasmania, 95, 35-40.

Davies, J.L. 1972 Geographical variations in coastal development. K.M. Clayton (ed.), Oliver and Boyd.

Dawson, A.G. 1977 A fossil Lobate Rock Glacier in Jura. Scott. J. Geol., 13, 37-42.

Dawson, A.G. 1979 A Devensian Medial Moraine in Jura. Scott. J. Geol., 15, 1-6.

Dionne, J.C. 1973 Distinction entre stries glacielles et stries glaciares. Rev. de Geog. de Montreal., 27, 185-90.

Donner, J.J. 1959 The Late- and Post-glacial raised beaches in Scotland. Ann. Acad. Scient. Fenn., Ser. A, III, 53, 1-25.

Donner, J.J. 1963 The Late- and Post-glacial raised beaches in Scotland. Ann. Acad. Scient. Fenn., Ser. A, III, 68, 1-13.

Edwards, A.B. 1951 Wave action in shore platform formation. Geol. Mag., 88, 41-49.

Embleton, C. and King, C.A.M. 1968 Glacial and periglacial geomorphology. E Arnold Ltd.

Everard, C.E., R.H. Lawrence, M.E. Witherick and L.W. Wright. 1964 Raised beaches and marine geomorphology. in "Present Views on some aspects of the Geology of Cornwall and Devon" Roy. geol. Soc. Cornwall, Truro. K.G.F. Hosking and G.J. Shrimpton (eds.), 283-310.

Follestad,B.A.1972 The Deglaciation of the South-Western Part of the Folgefonn Peninsula, Hordaland Norges geol. Unders., 280, 31-64.

Geikie, A. 1882 Notes on the Geology of Colonsay and Oronsay. Trans. geol. Soc. Glasgow, 6.

Geikie, A. 1889 Summ. Progr. Geol. Surv., pp.50-2,75 and 142.

Geikie, A. 1865 The Scenery of Scotland viewed in connection with its Physical Geology. London and Cambridge.

Gilbert, G.K. 1883 The topographic features of lake shores. U.S. Geol. Surv. 5th Ann. Rept., 69-123.

Godard, A. 1965 Recherches de geomorphologie en Ecosse du Nordouest. Faculte des lettres de l'universite de Strasbourg., Paris.

Gray, J.M. 1972 The inter-, late-, and post-glacial shorelines, and ice-limits of Lorn and eastern Mull. 2 vols. Unpubl. Ph.D. thesis, Univ. of Edinburgh.

Gray, J.M. 1974a The Main Rock Platform of the Firth of Lorn, western Scotland. Trans. Inst. Br. Geogr., 61, 81-99.

Gray, J.M. 1974b Lateglacial and postglacial shorelines in western Scotland. Boreas, 3, 129-38.

Gray, J.M. 1975a Measurement and Analysis of Scottish Raised Shoreline Altitudes. Dept. of Geogr., Queen Mary College, Univ. of London, Occ. Paper. 2.

Gray, J.M. 1975b The Loch Lomond Readvance and contemporaneous sea-levels in Loch Etive and neighbouring areas of western Scotland. Proc. Geol. Assoc. 86, 227-38.

Gray, J.M. 1978 Low-level shore platforms in the south-west Scottish Highlands: altitude , age and correlation. Trans. Inst. Br. Geogr., 3, 151-64.

Gray, J.M. and Lowe,J.J. 1977 Studies in the Scottish Lateglacial Environment. Pergamon Press.

Grønlie, O.T. 1924 Contributions to the Quaternary Geology of Novaya Zemlya. in "Report of the Scientific Results of the Norwegian Expedition to Novaya Zemlya, 1921", O. Holtedahl (ed.).

Guilcher, A. 1958 Coastal and Submarine Morphology.

Harkness, D.D. and Wilson, H.W. 1974 Scottish Universities Research and Reactor Centre Radiocarbon Measurements II. Radiocarbon, 16, 238-51.

Hellberg, K. 1971 Inlandisens recession och den Senglaciala Stranforskutningen, Vastra Blekinge och Nordostra Skane. Lunds Univ. Naturgeogrrafiska, Inst. Rapp. Not., 9, 124pp.

Hills, E.S. 1940 The Question of Recent emergence of the Shores of Port Phillip Bay. Proc. roy. Soc. Vict., 52, 84-105.

Hills, E.S. 1949 Shore platforms. Geol. Mag., 86, 137-52.

Hoel, A. 1909 Geologiske iagttagelser på Spitsbergen ekspeditionerne 1906 and 1907. Norsk geol. Tidsskr., 11.

Holtedahl, O. 1929 On the geology and physiography of some Antarctic and sub-Antarctic islands with notes on the character and origin of fjords and strandflats of some Northern lands. in "Scientific Results of the Norwegian Antarctic Expeditions 1927-8", 1-172.

Holtedahl, H. 1960 Den norske strandflate, med saerlig henblikk på dens utvikling i kystomradene på More. Norsk geogr. Tidsskr. 16, 285-303.

Ives, J.D. and Andrews, J.T. 1963 Determination of the marine limit in the eastern Canadian Arctic. Geogr. Bull., 19, 117-22.

Iwagaki, Y. and Noda, H. 1963 Laboratory study of scale effects in two-dimensional beach processes. Proc. 8th Conf. on Coastal Engineering. 194-210.

Jahn, A. 1961 Quantitative Analysis of some periglacial processes in Spitsbergen. Uniwersytet Wroclawski Im Boleslawa Bieruta , Zeszyty Naukowe Nauki Przyrodnicze, II, Warsaw.

Jahn, A. 1977 Periglacial forms produced by shore ice at Horsund, Spitzbergen. Acta Univ. Wratislaviensis studia Geograficzne, 387, 19-29.

Jardine, W.G. 1974 Oronsay shell mounds. in " Discovery and Excavation in Scotland, 1973 ", M.E.C. Stewart and C.M. Lythe (eds.), 9-10.

Jardine, W.G. 1975 Chronology of Holocene marine transgression and regression in south-western Scotland. Boreas, 4, 173-96.

Jardine, W.G. 1977 The Quaternary marine record in southwest Scotland and the Scottish Inner Hebrides. in "The Quaternary history of the Irish Sea", C.Kidson and M.Tooley (eds.), Geol. Soc. Spec. Issue, 7, 99-118.

Johnson, D.W. 1919 Shore processes and shoreline development. Wiley.

Johnson, D.W. 1938 Shore platforms- discussion. J. Geomorph., 1, 268-72.

Kemp, D.D. 1971 The stratigraphy and sub-carse morphology of an area on the northern side of the River Forth between Lake of Menteith and Kincardine-on-Forth. Unpubl. Ph.D. thesis, Univ. of Edinburgh.

King, C.A.M. 1959 Beaches and Coasts. E. Arnold, London, 403pp.

King, C.A.M. 1972 Beaches and Coasts. E. Arnold, London, 570pp.

King, C.A.M. and Wheeler, P.T. 1963 The raised beaches of the north coast of Sutherland, Scotland. Geol. Mag., 100, 299-320.

Knaps, R. 1938 Profung der Formel von Münch-Petersen uber Materialwanderung an der lettischen Kuste. Proc. 6th Baltic Hydrol. Conf., Berlin.

Komar, P.D. 1976 Beach Processes and Sedimentation. Prentice-Hall Inc.

Kranck, E. 1950 On the Geology of the East Coast of Hudson Bay and James Bay. Acta Geogr., 11, Soc. Geog.Fenn., Helsinki.

Krumbein, W.C. and Graybill, F.A. 1965 An introduction to statistical models in geology. McGraw-Hill, New York.

Lewis, W.V. 1936 Nivation, river grading and shoreline development in south-east Iceland. Geogr. J., 38, 431-47.

Lewis, J.R. 1976 The Ecology of Rocky Shores. Hodder and Stoughton.

Loken, O.H. 1962 The late-glacial and postglacial emergence and the deglaciation of northernmost Labrador. Geogr. Bull., 17, 23-56.

Longinov, V.V. and Pasechnik, L.D. 1953 Fundamentals of profile development on shingle beaches. Trudy. Inst. Okeanol. Akad. Nauk. S.S.S.R., 7.

Mackie, E.W. 1972 Radiocarbon dates for two Mesolithic shell heaps and a Neolithic axe factory in Scotland. Proc. Prehist. Soc., 38, 412-416.

McCallien, W.J. 1937 Late-glacial and Early Post-glacial Scotland. Proc. Soc. Antiq. Scot. 71, 174-206.

McCann, S.B. 1961 The raised beaches of Western Scotland. Unpubl. Ph.D. thesis, Univ. of Cambridge.

McCann, S.B. 1964 The raised beaches of north-east Islay and western Jura, Argyll. Trans. Inst. Br. Geogr. 35, 1-16.

McCann, S.B. 1966 The Main Post-glacial Raised Shoreline of Western Scotland from the Firth of Lorne to Loch Broom. Trans. Inst. Br. Geogr., 39, 87-99.

McCann, S.B. 1968 Raised Rock Platforms in the Western Isles of Scotland. in "Geography at Aberystwyth", E.G. Bowen, H. Carter and J.A. Taylor (eds.), 22-34., Cardiff.

McCann, S.B. and Richards, A. 1969 The coastal features of the Island of Rhum in the Scottish Inner Hebrides. Scott. J. Geol., 5, 15-25.

Mangerud, J. 1970 Late-Weischelian vegetation and ice front oscillations in the Bergen district, Western Norway. Norsk. geogr. Tidsskr., 24, 121-48.

Marthinussen, M. 1960 Coast and fjord areas of Finmark. in "Geology of Norway", O.Holtedahl (ed.), 208, 416-29.

Moign, A. 1974 Strandflats immergés et emergés du Spitsberg Central et Nord-Occidental. Unpubl. Ph.D. thesis, Univ. of Bretagne Occidentale.

Moller, J.J. and Sollid, J.L. 1972 Deglaciation chronology of Lofoten-Vesteralen-Ofoten, North Norway. Norsk. geogr. Tidsskr., 26, 101-33.

Moore, G.W. 1960 Recent eustatic sea-level fluctuations recorded by Arctic beach ridges. U.S. Geol. Surv. Prof. Paper 400-B, pp.335-37.

Movius, H.L. 1942 The Irish Stone Age. Cambridge.

Munch-Petersen, T. 1938 Uber Materialwanderung an Meereskusten. Proc. 5th. Hydrol. Conf. Balt. Staaten, Helsinki.

Nansen, F. 1922 The strandflat and isostasy. Videnskapselkapets Skrifter 1. Math.- Naturw. Kl., 1921; 11, Kristiana.

Peach, B.N. 1901 Summ. Progr. Geol. Surv. Gt. Br., p.141.

Peach, B.N. 1909 The geology of the seaboard of Mid-Argyll. Mem. geol. Surv. Gt. Br.

Peach, B.N., C.T. Clough and H. Knayaston, 1905 The geology of Mid-Argyll. Mem. geol. Surv. Gt. Br.

Peach, B.N., J.S.G. Wilson, J.B. Hill, E.B. Bailey and G.W. Graham, 1911 The geology of Knapdale, Jura, and North Kintyre. Mem. geol. Surv. Gt. Br.

Peacock, J.D. 1975 Quaternary of Scotland- discussion. Scott. J. Geol., 11, 174-5.

Peacock, J.D. 1976 Quaternary features of Rhum, Inner Hebrides. Quaternary Newsletter, 20, 1-4.

Ray, L.L. 1935 Some minor features of valley glaciers and valley glaciation. J. Geol., 43, 297-322.

Rekstad, J. 1915 Om strandlinjer og strandlinjedannelse. Norsk geol. Tidsskr. III, 8, Kristiana.

Reusch, H. 1894 Strandfladen, et nyt traek i Norges geografi. Norges geol. Unders. 14, Handbook Aarbog for 1892 and 1893, Kristiana.

Rhind, D.W. 1969 The terraces of the Tweed valley. Unpubl. Ph.D. thesis, Univ. of Edinburgh.

Richey, J.E. and Thomas, H.H. 1930 The geology of Ardnamurchan, Northwest Mull and Coll. Mem. geol. Surv. Gt. Br.

Ritchie, W. 1966 The post-glacial rise in sea-level and coastal changes in the Uists. Trans. Inst. Br.Geogr., 39, 79-86.

Ritchie, W. 1972 Beaches of the Uists. Dept. of Geography, Univ. of Aberdeen, for the Countryside Commission of Scotland.

Ritchie, W. and Crofts, R. 1974 The beaches of Islay, Jura and Colonsay. Dept. of Geography, Univ. of Aberdeen, for the Countryside Commission of Scotland.

Robinson, M. 1977 Glacial limits, sea-level changes and vegetational development in parts of Wester Ross. Unpubl. Ph.D. thesis, Univ. of Edinburgh.

Sharp, R.P. 1949 Studies of superglacial debris on valley glaciers. Amer. J. Sci., 264, 289-315.

Silva, E.F. 1972 Geomorphological observations and generalisations on the coasts of the South Shetland Islands and Antarctic peninsula. in "Antarctic geology and geophysics" R.J. Adie (ed.), Oslo, 99-103.

Sim, V.W. 1964 Terrain analysis of west-central Baffin Island, N.W.T. Geogr. Bull., 21, 66-92.

Sissons, J.B. 1966 Relative sea-level changes between 10,300 B.P. and 8,300 B.P. in part of the Carse of Stirling. Trans. Inst. Br. Geogr., 39, 19-29.

Sissons, J.B. 1967a Comments on the paper by F.M. Synge and N Stephens in Transactions no.39. Trans. Inst. Br. Geogr., 42, 163-8.

Sissons, J.B. 1967b The evolution of Scotland's scenery. Edinburgh, Oliver and Boyd.

Sissons, J.B. 1969 Drift stratigraphy and buried morphological features in the Grangemouth-Falkirk-Airth area, central Scotland. Trans. Inst. Br. Geogr., 48, 31-9.

Sissons, J.B. 1972 Dislocation and non-uniform uplift of raised shorelines in the western part of the Forth valley. Trans. Inst. Br. Geogr., 55, 145-59.

Sissons, J.B. 1974a Lateglacial marine erosion in Scotland. Boreas, 3, 41-48.

Sissons, J.B. 1974b The Quaternary in Scotland: a Review. Scott. J. Geol., 10, 34-37.

Sissons, J.B. 1975 The Quaternary in Scotland- discussion. Scott. J. Geol. 11, 176-8.

Sissons, J.B. 1976a The geomorphology of Scotland. Methuen, London.

Sissons, J.B. 1976b A remarkable protalus rampart in Wester Ross. Scott. Geogr. Mag., 92, 182-90.

Sissons, J.B. 1977 The Loch Lomond Readvance in southern Skye and some palaeoclimatic implications. Scott. J. Geol., 13, 23-36.

Sissons, J.B. 1978 The parallel roads of Glen Roy and adjacent glens, Scotland. Boreas, 7, 229-44.

Sissons, J.B. and Brooks, C.L. 1971 Dating of early postglacial land and sea-level changes in the western Forth valley. Nature Phys. Sci., 234, 124-7.

Sissons, J.B., D.E. Smith and R.A. Cullingford, 1966 Late-glacial and postglacial shorelines in southeast Scotland. Trans. Inst. Br. Geogr., 39, 9-18.

Sissons, J.B. and Sutherland, D.G. 1976 Climatic inferences from former glaciers in the south-east Grampian Highlands, Scotland. J. Glaciol., 76, 325-46.

Small, R.J. and Clark, M.J. 1974 The medial moraines of the lower Glacier de Tsidjiore Nouve, Valais, Switzerland. J. Glaciol., 13, 255-65.

Smith, H.T.U. 1973 Photogeological study of periglacial talus glaciers in north-western Canada. Geogr. Annlr., 55A, 69-84.

Sollid, J.L., S. Andersen, N. Hamre, O. Kjeldsen, O. Salvigsen, S. Sturød, T. Tveita and A. Wilhelmsen. Deglaciation of Finmark, North Norway. Norsk geol. Tidsskr., 27, 233-325.

Steers, J.A. 1952 The coastline of Scotland. Geogr. J., 118, 180-90.

Stephens, N. 1957 Some observations on the " Interglacial " Platform and the Early Post-glacial Raised Beach on the east coast of Ireland. Proc. roy. Irish Acad., 58, B6, 129-49.

Stephens, N. 1965 Lateglacial and postglacial shorelines in Ireland and southwset Scotland. in "Means of Correlating Quaternary successions" R.B. Morrison and H.E. Wright (eds.), Salt Lake City, 437-456.

Strøm, K.M. 1948 The geomorphology of Norway. Geogr. J., 112,19-27.

Summ. Progr. geol. Surv., 1899.

Synge, F.M. 1966 The relationship of the raised strandlines and main end moraines on the Isle of Mull, and in the district of Lorn, Scotland. Proc. Geol. Assoc., 77, 315-28.

Synge, F.M. 1977 Records of sea-levels during the late-Devensian. in "The Changing environmental conditions in Great Britain and Ireland during the Devensian (last) cold stage" G.F. Mitchell and R.G. West (eds.) Phil. Trans. roy. Soc. Lond., B, 211-28.

Synge, F.M. and Stephens, N. 1966 Late- and Post-glacial shorelines and ice-limits in Argyll and Northeast Ulster. Trans. Inst. Br. Geogr., 39, 101-25.

Tanner, V. 1930 Studier over Kvartarystemet; Fennoscandias nordliga delar IV. Bull. Comm. geol. Finl. 88, 1-594.

Ting, S. 1937 The coastal configuration of western Scotland. Geogr. Annlr., 19, 62-83.

Trenhaile, A.S. 1971 Lithological control of high-water rock ledges in the Vale of Glamorgan. Geogr. Annlr., 53A, 59-69.

Trenhaile, A.S. 1974 The morphology and classification of shore platforms in England and Wales. Geogr. Annlr., 56A, 103-10.

Unwin, D. 1975 An introduction to trend surface analysis. Concepts and Techniques in Modern Geography No. 5, Brit. Geomorph. Res. Gp. Pub.

Vladimirov, V. 1953 in Zenkovich, V.P. 1967 Processes of Coastal Development. J.A. Steers (ed.), Oliver and Boyd, Edinburgh.

Vogt, T. 1918 Om recente og gamle strandlinjer i fast fjeld. Norsk geol. Tidsskr., III, 8, Kristiana.

Von Weymarn, J. 1974 Coastline development in Lewis and Harris, Outer Hebrides, with particular reference to the effects of glaciation. Unpubl. Ph.D. thesis, Univ. of Aberdeen.

Wahrhaftig, C. and Cox, A. 1959 Rock glaciers in the Colorado Front Range 1961-68. Bull. geol. Soc. Amer. 70, 383-436.

Washburn, A.L. 1973 Periglacial processes and environments. London.

Wentworth, C.K. 1938 Marine bench forming processes: water level weathering. J. Geomorph., 1, 6-32.

Whalley, W.B. 1974 Origin of rock glaciers. J. Glaciol., 68, 321-2.

White, S.E. 1971 Rock glacier studies in the Colorado Front Range, 1961-1968. Arctic Alp. Res., 3, 43-64.

White, S.E. 1976 Rock glaciers and Block fields, review and new data. Quat. Res., 6, 77-98.

Wilkinson, S.B. 1897 Summ. Progr. geol. Surv., 152-3.

Wilkinson, S.B. 1907 The geology of Islay, including Oronsay and portions of Colonsay and Jura. Mem. geol. Surv. Gt. Br.

Wright, W.B. 1911 On a preglacial shoreline in the Western Isles of Scotland. Geol. Mag., 48, 97-109.

Wright, W.B. 1928 The Raised Beaches of the British Isles. First Rep. of Comm. on Pliocene and Pleistocene Terraces, Int. Geogr. Union, 99-106

Wright, W.B. 1937 The Quaternary Ice Age. 2nd ed., London.

Zenkovich, V.P. 1946 On the study of shore dynamics. Trans. Inst. Okeanol. Akad. Nauk. S.S.S.R., 1, 99-112.

Zenkovich, V.P. 1967 Processes of Coastal Development. Oliver and Boyd, Edinburgh.

Zeuner, F.E. 1959 The Pleistocene Period. Hutchinson, London.

Appendix 1

Wave Energy Potential:- FORTRAN computer program

```
C     MUNCH PETERSEN/KNAPS ADAPTATION
C     CALCULATES AND SUMMATES MATERIAL MOVING FORCE
C     ASSIGNS SPECIFIC CONSTANTS TO FETCH SECTORS
C     JURA, SCARBA AND NE ISLAY
      DIMENSION ALPHA(8)
      DO 10 I=1,N
      READ (5,100) NO,(ALPHA(J),J=1,8)
  100 FORMAT (I4,8F6.1)
      FORCE =0.0
      DO 11 J=1,8
      ALPHA(J) =ALPHA(J)*1.0/57.3
      ALPHA(J) =2.0*(SIN(ALPHA(J))*COS(ALPHA(J)))
      IF(J.EQ.1) CONST =113.93
      IF(J.EQ.2) CONST = 18.90
      IF(J.EQ.3) CONST = 16.20
      IF(J.EQ.4) CONST = 18.89
      IF(J.EQ.5) CONST = 13.46
      IF(J.EQ.6) CONST = 14.95
      IF(J.EQ.7) CONST =  5.93
      IF(J.EQ.8) CONST =  5.87
      ALPHA(J) = CONST*ALPHA(J)
      FORCE =FORCE+ ALPHA(J)
   11 CONTINUE
      WRITE (6,555) NO,(ALPHA(J), J=1,8), FORCE
  555 FORMAT (' ',I4,8F6.1,F12.2)
   10 CONTINUE
      STOP
      END
```

Appendix. 2 Wave Energy Potential Calculations.

At each coastal site the main fetch sectors (Fig.8) were determined and the average length of fetch in each sector measured. The values chosen for length of open Atlantic fetch are arbitrary and vary according to the position of the wave-generating storm cyclones in the ocean. However since the dimensionless wave energy potential indices calculated are used in this study to compare the relative susceptibility of different coastal sites to wave attack the values used are believed satisfactory. Secondly since the cube root of fetch length is employed (see below) errors are reduced. The fetch length and cube root are as follows,

	NE Islay	Open SW Atlantic	Colonsay	Open NW Atlantic	Mull
Fetch (miles)	6.0	3000.0	16.0	1500.0	12.0
cube root	1.2010	14.4225	2.5198	11.4471	2.2894

In order to calculate an index for S^3F, Tiree wind recordings (Meteorological Office) of wind direction, velocity and duration for 1962-76 were used (Table 8). Since high velocity winds have a greater geomorphological impact on wave activity in the coastal zone than low velocity winds, the annual frequency values of wind direction and velocity were weighted proportionally to the velocities (Table 8). The transformed values were weighted as follows,

Wind velocity (knots)	Weighting
0-6	x1
7-16	x4
17-27	x9
>28	x16

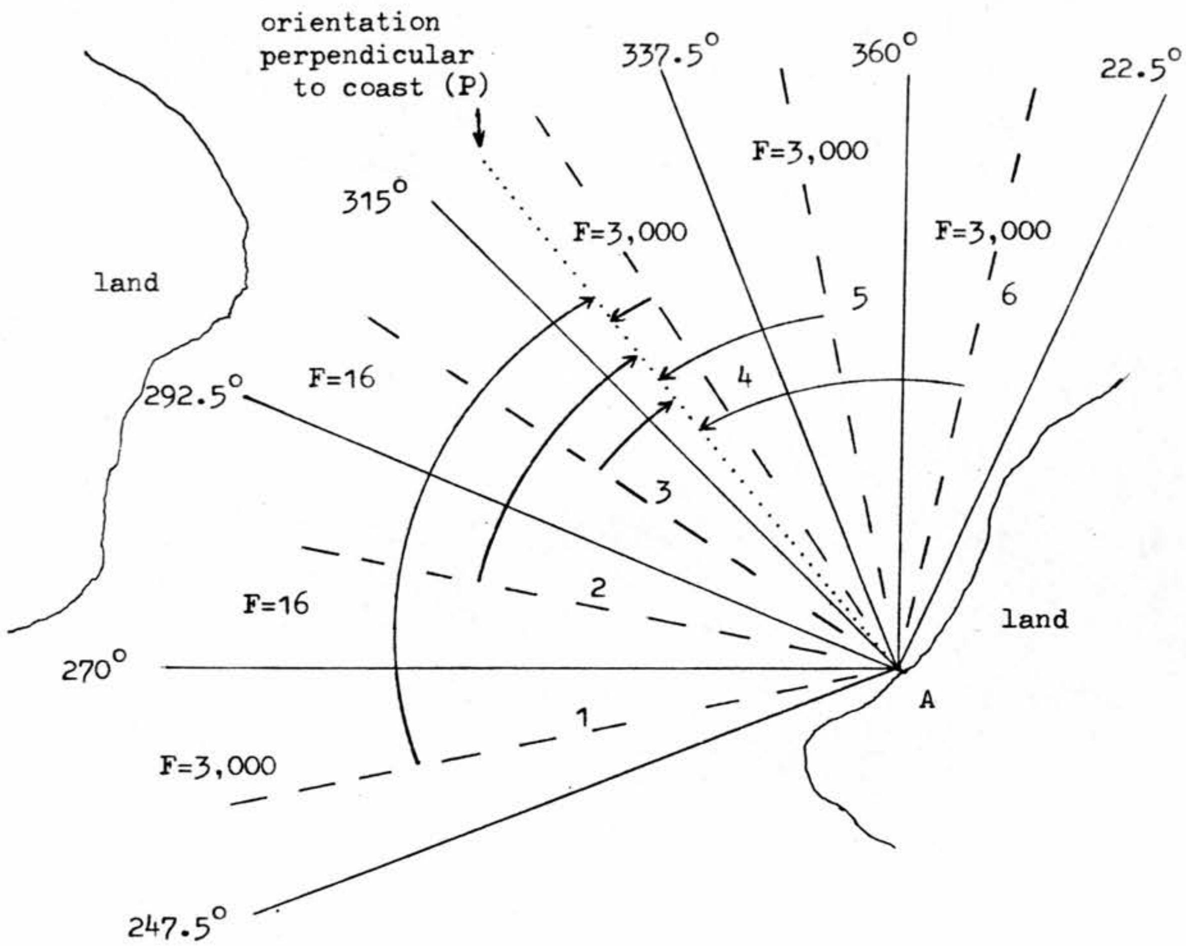
Each weighted percentage value (Table 8) is therefore an index of

(contd.)

the wind capacity to impart energy to the waves travelling in a given sector.

Thereafter the percentage frequency wind energy values were assigned proportionally to $22\frac{1}{2}^{\circ}$ wind sectors. For each coastal site the mid-point of each $22\frac{1}{2}^{\circ}$ sector was determined and a fetch length determined for that sector (see over). Subsequently for each $22\frac{1}{2}^{\circ}$ sector the wind energy index (S^3F) was multiplied by the cube root of the length of fetch. Thus the $S^3F\sqrt[3]{D}$ indices were obtained.

Appendix 2 (contd). Wave Energy Potential.



Diagrammatic example demonstrating the method by which wave energy potential values are calculated for coastal location A. Fetch sectors (1), (4), (5) and (6) are 3000 miles in length. Fetch sectors (2) and (3) are both 16 miles in length. The orientation perpendicular to the coast at site A is shown by P.

Wave energy potential (Wp) for location A is found by:-

$$Wp = \sum_{1}^{6}\left(\left(\left(S^3F\right)\sqrt[3]{D}\right) \times 2 \sin\alpha \cos\alpha\right)$$

Initially a weighted wind percentage (S^3F) is calculated for each $22\frac{1}{2}^{\circ}$ sector. Each S^3F value is multiplied by $\sqrt[3]{D}$ for that sector. In sectors (2) and (3) $\sqrt[3]{D} = \sqrt[3]{16}$. In sectors (1), (4), (5) and (6) $\sqrt[3]{D} = \sqrt[3]{3000}$. Thereafter $((S^3F) \times \sqrt[3]{D})$ is calculated for each sector.

The angle is then measured to the perpendicular (P) from the mid-point

(contd.)

of each sector. In this case 6 values of 2 sin α cos α are determined. Each 2 sin α cos α value is them multiplied by the respective ($S^3F\sqrt[3]{D}$) value. The 6(Wp) sector values are then added to derive the total wave energy potential index for location A.

Appendix 3

Main Rock Platform fragment altitudes and residuals

	Grid Reference	Altitude (m)	1st order residual	2nd order residual
1	NM 71220695	6.09	-0.11	-0.54
2	NM 71090666	7.05	0.88	0.46
3	NM 71000660	6.60	0.44	0.04
4	NM 70820657	6.58	0.45	0.08
5	NM 70690656	6.67	0.55	0.21
6	NM 70380651	6.96	0.88	0.61
7	NM 72080358	6.69	0.61	-0.25
8	NM 71970337	6.34	0.28	-0.56
9	NM 71700335	4.91	-1.12	-1.90
10	NM 71560327	6.31	0.31	-0.46
11	NM 71250342	5.57	-0.41	-1.10
12	NM 71890337	4.81	-1.24	-2.06
13	NM 71020322	6.38	0.44	-0.21
14	NM 70800297	6.66	0.76	0.14
15	NM 70560273	5.99	0.13	-0.45
16	NM 69790265	7.05	1.29	0.85
17	NM 69630258	7.27	1.53	1.12
18	NM 69480260	6.72	1.00	0.61
19	NM 69290263	5.91	0.20	-0.14
20	NM 69160260	6.93	1.24	0.92
21	NM 69010259	6.70	1.03	0.74
22	NM 68890257	6.68	1.02	0.76
23	NM 68770260	6.98	1.33	1.09
24	NM 68240056	5.74	0.29	0.01
25	NM 67960050	5.05	-0.37	-0.60
26	NM 67960040	6.52	1.11	0.87
27	NM 67970025	6.37	0.97	0.72
28	NM 67880009	6.12	0.74	0.49
29	NM 67800011	5.95	0.57	0.35
30	NM 67219969	4.49	-0.79	-0.94
31	NR 67299957	4.63	-0.65	-0.83
32	NR 66989912	4.90	-0.32	-0.47
33	NR 66889906	4.93	-0.27	-0.41
34	NR 66809900	5.27	0.08	-0.05
35	NR 66739880	4.83	-0.34	-0.47
36	NR 66339848	4.79	-0.31	-0.40
37	NR 66259856	5.50	0.40	0.34
38	NR 66169858	5.03	-0.06	-0.11
39	NR 65969849	4.20	-0.86	-0.88
40	NR 65719819	4.81	-0.20	-0.21
41	NR 65269797	6.22	1.27	1.32
42	NR 64289751	4.92	0.11	0.28
43	NR 64279755	5.55	0.74	0.91
44	NR 64329741	4.90	0.09	0.25
45	NR 64369730	5.57	0.77	0.91
46	NR 63129648	5.12	0.51	0.77
47	NR 62959633	3.31	-1.27	-1.00
48	NR 62699625	5.27	0.72	1.02
49	NR 62619606	4.56	0.03	0.33
50	NR 62519591	5.17	0.66	0.96
51	NR 62329575	4.93	0.46	0.77
52	NR 62389565	3.56	-0.91	-0.62

Appendix 3 (contd.)

Main Rock Platform fragment altitudes and residuals

	Grid Reference	Altitude (m)	1st order residual	2nd order residual
53	NR 62429550	5.27	0.80	1.08
54	NR 62419535	5.27	0.81	1.08
55	NR 62199498	4.56	0.15	0.42
56	NR 61919502	4.04	-0.34	-0.03
57	NR 61639498	2.65	-1.69	-1.36
58	NR 61229475	3.55	-0.73	-0.36
59	NR 61179465	3.40	-0.87	-0.52
60	NR 61029448	4.52	0.27	0.65
61	NR 60689436	3.86	-0.34	0.07
62	NR 60649434	4.58	0.39	0.80
63	NR 60249420	3.21	-0.93	-0.48
64	NR 60109413	4.23	0.11	0.57
65	NR 59569389	3.35	-0.69	-0.20
66	NR 59769400	3.65	-0.42	0.06
67	NR 60009403	3.30	-0.80	-0.34
68	NR 59379366	3.58	-0.42	0.07
69	NR 59259356	3.11	-0.88	-0.37
70	NR 59139350	2.35	-1.62	-1.11
71	NR 59029339	3.83	-0.12	0.39
72	NR 58979336	2.14	-1.80	-1.29
73	NR 58929329	2.49	-1.44	-0.93
74	NR 58409290	3.04	-0.81	-0.27
75	NR 58039277	3.11	-0.69	-0.13
76	NR 57869267	3.07	-0.70	-0.13
77	NR 57169225	3.10	-0.57	0.03
78	NR 57009190	2.50	-1.12	-0.55
79	NR 57079177	2.66	-0.96	-0.40
80	NR 53649106	4.02	0.53	1.08
81	NR 55969080	2.84	-0.60	-0.03
82	NR 55639070	2.27	-1.12	-0.54
83	NR 55549060	1.72	-1.66	-1.07
84	NR 55359050	1.67	-1.68	-1.09
85	NR 55109033	2.25	-1.06	-0.47
86	NR 54839008	2.09	-1.17	-0.59
87	NR 54588989	2.76	-0.46	0.12
88	NR 53918950	3.14	0.02	0.61
89	NR 54058962	2.92	-0.22	0.37
90	NR 53408892	1.91	-1.12	-0.55
91	NR 57379244	3.03	-0.67	-0.08
92	NR 53118874	2.30	-0.68	-0.11
93	NR 53058841	1.70	-1.25	-0.72
94	NR 53048787	2.69	-0.23	0.26
95	NR 56218719	2.03	-0.79	-0.35
96	NR 51928633	1.75	-0.94	-0.55
97	NR 51658607	2.92	0.28	0.66
98	NR 51608601	3.30	0.67	1.04
99	NR 50988493	2.40	-0.09	0.20
100	NR 50918487	1.41	-1.07	-0.78
101	NR 50858481	1.52	-0.95	-0.66
102	NR 50818461	2.29	-0.16	0.11
103	NR 50678431	2.33	-0.09	0.16
104	NR 50708412	2.19	-0.22	0.01

Appendix 3 (contd.)

Main Rock Platform fragment altitudes and residuals

Grid	Reference	Altitude (m)	1st order residual	2nd order residual
105	NR 51908253	2.60	0.16	0.26
106	NR 52328232	2.74	0.26	0.35
107	NR 53178203	2.26	-0.29	-0.22
108	NR 53268193	2.16	-0.40	-0.33
109	NR 53358182	2.42	-0.14	-0.08
110	NR 53458179	2.28	-0.29	-0.23
111	NR 53838153	2.85	0.25	0.30
112	NR 54008147	2.82	0.21	0.25
113	NR 52658039	2.51	0.12	0.11
114	NR 52708042	2.84	0.44	0.44
115	NR 54488088	3.14	0.51	0.54
116	NR 54558091	3.09	0.45	0.48
117	NR 55168100	3.18	0.47	0.50
118	NR 44147269	1.53	0.60	-0.20
119	NR 44027243	1.36	0.46	0.37
120	NR 50638022	1.44	-0.71	-0.75
121	NR 50518036	1.30	-0.85	-0.88
122	NR 50498040	2.14	-0.01	-0.04
123	NR 50338094	1.52	-0.61	-0.64
124	NR 50308046	2.23	0.10	-0.07
125	NR 50288043	2.50	0.38	0.34
126	NR 50138030	2.59	0.49	0.45
127	NR 49818009	2.56	0.51	0.45
128	NR 49958015	1.79	-0.28	-0.34
129	NR 49588003	2.38	0.36	0.29
130	NR 49508001	2.80	0.79	0.72
131	NR 49087990	2.49	0.54	0.44
132	NR 48958005	1.70	-0.25	-0.33
133	NR 48777996	2.56	0.64	0.54
134	NR 48697986	2.14	0.23	0.13
135	NR 48417946	2.31	0.46	0.32
136	NR 48037936	2.27	0.47	0.31
137	NR 46927796	1.77	0.19	-0.11
138	NR 46757786	1.41	-0.15	-0.46
139	NR 46547764	1.69	0.17	-0.17
140	NR 42197925	0.05	-1.18	-1.54
141	NR 42437924	1.86	0.70	0.35
142	NR 41927918	1.39	0.29	-0.09
143	NR 41827914	0.96	-0.13	-0.51
144	NR 41787918	1.09	-0.01	-0.38
145	NR 41657903	1.17	0.11	-0.30
146	NR 40577890	2.77	1.84	1.37
147	NR 40147888	2.15	1.27	0.78
148	NR 40057885	1.22	0.35	-0.15
149	NR 39557825	1.93	1.16	0.56
150	NR 39477817	2.09	1.33	0.72
151	NR 39307811	1.24	0.51	-0.13
152	NR 39257810	1.01	0.28	0.36
153	NR 39187809	1.42	0.70	0.06
154	NR 39017812	1.43	0.73	0.08
155	NR 38697813	1.74	1.07	0.41
156	NR 38377814	1.06	0.43	-0.25
157	NR 56177250	2.09	-0.19	-0.03
158	NR 56167241	2.00	-0.27	-0.11

Appendix 3 (contd.)

Main Rock Platform fragment altitudes and residuals

Grid Reference	Altitude (m)	1st order residual	2nd order residual
159 NR 56167231	1.95	-0.31	-0.15
160 NR 56177220	1.96	-0.30	-0.13
161 NR 56227212	1.86	-0.40	-0.22
162 NR 61558216	3.57	0.06	0.05
163 NR 61398203	3.76	0.28	0.27
164 NR 61268188	3.65	0.19	0.19
165 NR 61138177	3.48	0.04	0.04
166 NR 61028179	3.02	-0.41	-0.40
167 NR 60978190	3.15	-0.28	-0.28
168 NR 60778196	3.50	0.11	0.12
169 NR 65408827	4.75	0.41	0.15
170 NR 65338824	4.73	0.40	0.14
171 NR 65288802	4.70	0.39	0.14
172 NR 65178776	4.63	0.35	0.11
173 NR 63688649	4.43	0.40	0.26
174 NR 63648640	4.41	0.39	0.25
175 NR 40577890	1.41	0.24	-0.23

Appendix 4 Main Rock Platform : platform widths and cliff heights (metres)

No.	Grid Reference (NR)	Platform width (w)	Cliff height (h)	w.h	½(w.h)
1	53758197	55	30	1650	825
2	53738922	20	20	400	200
3	53828923	55	21	1155	577.5
4	53858935	100	19	1900	950
5	53888940	75	22	1650	825
6	53908947	45	22	990	495
7	53988953	30	18	540	270
8	54028958	70	24	1680	840
9	54088963	100	24	2400	1200
10	54178973	100	26	2600	1300
11	54248976	100	26	2600	1300
12	54278981	85	23	1955	977.5
13	54298982	95	22	2090	1045
14	54478977	80	31	2480	1240
15	54368968	125	26	3250	1625
16	54588988	125	29	3625	1812.5
17	54678993	110	29	3190	1595
18	54739001	100	28	2800	1400
19	54859006	60	29	1740	870
20	54909010	70	32	2240	1120
21	54989010	125	34	4250	2125
22	55069022	70	36	2520	1260
23	55089027	50	29	1450	725
24	55109031	20	29	580	290
25	55229030	45	31	1395	697.5
26	55239043	75	34	2550	1275
27	55119033	70	33	2310	1155
28	55299036	70	34	2380	1190
29	55329046	60	33	1980	990
30	55409052	50	36	1800	900
31	55479051	50	36	1800	900
32	55599060	75	32	2400	1200
33	55629067	25	34	850	425
34	55709069	35	34	1190	595
35	55839066	50	36	1800	900
36	55797071	35	34	1190	595
37	55959070	55	34	1870	935
38	56119068	100	33	3300	1650
39	56139071	170	34	5780	2890
40	56069063	100	34	3400	1700
41	56189076	125	36	4500	2250
42	56359087	130	36	4680	2340
43	56399100	90	34	3060	1530
44	56449107	105	30	3150	1575
45	56519114	100	29	2900	1450
46	56579120	85	29	2465	1232.5
47	56659162	100	31	3100	1550
48	56779129	110	33	3630	1815
49	57059134	140	35	4900	2450
50	57059140	110	45	4950	2475
51	57059153	125	55	6875	3437.5
52	57079172	75	34	2550	1275
53	57069179	55	36	1980	990
54	57089188	60	29	1740	870

Appendix 4 (contd.)

No.	Grid Reference (NR)	Platform width	Cliff height	w.h	½(w.h)
55	57109193	100	39	3900	1950
56	57099197	135	36	4590	2295
57	57099202	150	31	4650	2325
58	57119207	125	40	5000	2500
59	57129216	110	31	3410	1705
60	57149222	50	31	1550	775
61	57229232	50	29	1450	725
62	57259235	50	29	1160	580
63	57329238	30	41	1230	615
64	57429249	55	29	1595	797.5
65	57439250	75	29	2175	1087.5
66	57709249	25	34	850	425
67	57779254	60	39	2340	1170
68	57819259	60	29	1740	870
69	57889269	60	31	1860	930
70	57969274	35	29	1015	507.5
71	58029278	20	29	580	290
72	58089280	25	31	775	387.5
73	58269273	75	45	3375	1687.5
74	58329282	50	49	2450	1225
75	58109272	100	44	4400	2200
76	58389288	30	59	1770	885
77	58419293	25	69	1725	862.5
78	58459295	60	59	3540	1770
79	58499294	45	49	2205	1102.5
80	58569296	40	49	1960	980
81	58709303	60	29	1740	870
82	58729309	40	31	1240	620
83	58769313	60	26	1560	780
84	58819316	110	29	3190	1595
85	58929326	150	31	4650	2325
86	58949331	85	31	2635	1317.5
87	59039337	100	39	3900	1950
88	59149338	60	64	3840	1920
89	59199348	25	64	1600	800
90	59219351	50	68	3400	1700
91	59249353	65	63	3995	1997.5
92	59319360	65	64	4160	2080
93	59359368	30	67	2010	1005
94	59409371	100	74	7400	3700
95	59459375	65	72	4680	2340
96	59549380	50	71	3550	1775
97	59709379	85	38	3230	1615
98	59759395	90	37	3060	1530
99	59779399	90	31	2790	1395
100	59809394	165	44	7260	3630
101	59939397	75	49	3675	1837.5
102	60009402	55	43	2365	1182.5
103	60089405	45	48	2160	1080
104	60119411	15	52	780	390
105	60199414	55	52	2860	1430
106	60259418	80	48	3840	1920
107	60379415	175	48	8400	4200
108	60479420	70	29	2040	1020
109	60609424	55	32	1760	880
110	60639432	85	42	3570	1785
111	60759435	150	40	6000	3000
112	60719434	140	44	6160	3080
113	60719435	175	44	7700	3850

Appendix 4 (contd.)

No.	Grid Reference (NR)	Platform width	Cliff height	w.h	$\frac{1}{2}$(w.h)
114	60819437	100	47	4700	2350
115	60959439	55	48	2640	1320
116	60989444	85	49	4165	2082.5
117	61129456	125	48	6000	3000
118	61159463	25	46	1150	575
119	61169464	25	51	1275	637.5
120	61199470	70	36	2520	1260
121	61419483	30	46	1380	690
122	61419483	80	44	3520	1760
123	61429492	85	22	1870	935
124	61369500	45	23	1035	517.5
125	61409504	30	27	810	405
126	61459504	45	27	1215	607.5
127	61509503	45	23	1035	517.5
128	61599498	55	32	1760	880
129	61629499	105	28	2940	1470
130	61689496	125	32	4000	2000
131	61729499	85	30	2550	1275
132	61769499	45	27	1215	607.5
133	61799498	75	29	2175	1087.5
134	61819497	35	33	1155	577.5
135	61939499	110	34	3740	1870
136	61989495	100	33	3300	1650
137	62049497	50	33	1650	825
138	62099499	70	37	2590	1295
139	62149498	90	31	2790	1395
140	62229495	65	32	2080	1040
141	62299496	85	28	2380	1190
142	62459526	35	47	1645	822.5
143	62459528	45	43	1935	967.5
144	62459538	55	42	2310	1155
145	62489555	115	46	5290	2645
146	62489556	105	47	4935	2467.5
147	62479560	125	48	6000	3000
148	62429567	100	50	5000	2500
149	62439572	75	52	3900	1950
150	62439576	60	49	2940	1470
151	62409580	50	48	2400	1200
152	62459585	55	49	2695	1347.5
153	62609593	105	48	5040	2520
154	62589599	75	50	3750	1875
155	62629600	62	50	3100	1550
156	62689611	45	48	2160	1080
157	62689622	15	55	825	412.5
158	62689620	50	49	2450	1225
159	62719623	20	48	960	480
160	62739625	40	46	1840	920
161	62809629	30	48	1440	720
162	62829630	115	34	3910	1955
163	62959629	50	45	2250	1125
164	62879627	125	33	4125	2062.5
165	63029631	30	36	1080	540
166	63079644	50	29	1450	725
167	63109646	60	29	1740	870
168	63229644	100	24	2400	1200

Appendix 4 (contd.)

No.	Grid Reference (NR)	Platform width	Cliff height	w.h	½(w.h)
169	63279649	85	22	1870	935
170	63319652	110	22	2420	1210
171	63339652	85	21	1785	892.5
172	63589649	120	22	2640	1320
173	63529647	40	24	960	480
174	63619677	75	27	2025	1012.5
175	63619684	50	29	1450	725
176	63639694	60	23	1380	690
177	63699701	50	24	1200	600
178	63689710	55	21	1155	577.5
179	63689716	75	30	2250	1125
180	63739722	80	26	2080	1040
181	63759724	30	33	990	495
182	63789733	30	32	960	480
183	63849737	50	34	1700	850
184	63899742	65	33	2145	1072.5
185	63929743	50	34	1700	850
186	63999755	35	35	1225	612.5
187	64029758	50	34	1700	850
188	64049759	75	25	1875	937.5
189	64159752	40	38	1520	760
190	64169751	70	33	2310	1155
191	64189750	25	39	975	487.5
192	64269752	40	35	1400	700
193	64309735	55	35	1925	962.5
194	64319730	35	34	1190	595
195	64319726	45	33	1485	742.5
196	64319720	50	34	1700	850
197	64359715	50	34	1700	850
198	64829714	50	33	1650	825
199	64889722	25	34	850	425
200	64889726	10	32	320	160
201	64899736	15	27	405	202.5
202	64889734	25	26	650	325
203	65009741	50	24	1200	600
204	65019743	50	21	1050	525
205	65049745	35	17	595	297.5
206	65039746	75	27	2025	1012.5
207	65089766	60	22	1320	660
208	65089767	85	21	1785	892.5
209	65119778	105	24	2520	1260
210	65139784	30	26	780	390
211	65149785	35	22	770	385
212	65169789	50	22	1100	550
213	65189786	60	24	1440	720
214	65209793	90	22	1980	990
215	65229795	100	23	2300	1150
216	65349793	75	14	1050	525
217	65609798	50	19	950	475
218	65639804	35	22	770	385
219	65689817	55	19	1045	522.5
220	65749819	70	22	1540	770
221	65859825	145	21	3045	1522.5
222	65889829	65	25	1625	812.5
223	65979844	65	23	1495	747.5

Appendix 4 (contd.)

No.	Grid Reference (NR)	Platform width	Cliff height	w.h	$\frac{1}{2}$(w.h)
224	66009847	75	33	2475	1232.5
225	66079849	60	32	1920	960
226	66099851	90	21	1890	945
227	66149855	90	22	1980	990
228	66169855	110	19	2090	1045
229	66209854	125	17	2125	1062.5
230	66219854	75	16	1200	600
231	66279848	90	23	2070	1035
232	66419840	90	22	1980	990
233	66619850	35	23	805	402.5
234	66669860	55	23	1265	632.5
235	66689861	35	22	770	385
236	66719862	35	26	910	455
237	66779882	40	29	1160	580
238	66779885	40	34	1360	680
239	66779886	105	35	3640	1820
240	66809889	130	36	4680	2340
241	66829895	80	35	2800	1400
242	66819900	55	39	2145	1072.5
243	66889902	65	40	2600	1300
244	66909904	120	40	4800	2400
245	66919905	130	40	5200	2600
246	66929906	130	39	5070	2535
247	66959910	55	40	2200	1100
248	67009909	65	38	2470	1235
249	67039909	100	36	3600	1800
250	67299924	105	47	4935	2967.5
251	67189908	100	45	4500	2250
252	67299924	50	48	2400	1200
253	67339938	35	50	1750	875
254	67309950	45	47	2115	1057.5
255	67299954	45	52	2340	1170
256	67279959	125	51	6375	3187.5
257	67279961	110	46	5060	2530
258	67269962	65	47	3055	1527.5
259	67279969	20	54	1080	540
260	67219974	55	38	2090	1045
261	67209979	45	36	1670	835
262	67179985	50	36	1800	900
263	67179986	40	39	1560	780
264	67189987	90	36	3240	1620
265	67359998	130	35	4550	2275
266	67209990	75	34	2550	1275
267	NM 67370002	115	35	4025	2012.5
268	NM 67460006	115	34	3910	1955
269	NM 67460007	100	39	3900	1950
270	NM 67470011	115	38	4330	2165
271	NM 67490012	100	43	4300	2150
272	NM 67500012	125	50	6250	3125
273	NM 67580007	125	48	6000	3000
274	NM 67710013	45	40	1800	900
275	NM 67800011	65	28	1820	910
276	NM 67760012	30	39	1170	585
277	NM 67970032	90	34	3060	1530
278	NM 67970027	55	28	1540	770

Appendix 4 (contd.)

No.	Grid Reference (NR)	Platform width	Cliff height	w.h	½(w.h)
279	NM 67980035	70	32	2240	1120
280	NM 67980038	65	29	1885	942.5
281	NM 67970044	65	30	1950	975
282	NM 67970048	40	28	1120	560
283	NM 67960051	100	33	3300	1650
284	NM 67960051	90	31	2790	1395
285	NM 67950053	55	30	1650	825
286	NM 67950054	50	32	1600	800
287	NM 67960054	90	31	2790	1395
288	NM 68010053	90	30	2700	1350
289	NM 68040051	55	35	1925	962.5
290	NM 68120052	55	34	1870	935
291	NM 68200053	60	48	2880	1440
292	NM 68210054	75	39	2925	1462.5
293	NM 68240054	100	36	3600	1800
294	NM 68260054	100	35	3500	1750
295	NM 68320052	80	34	2720	1360
296	NM 68370051	50	28	1400	700
297	NM 68380051	50	22	1100	550
298	NM 68440051	60	27	1620	810
299	NM 68430050	70	24	1680	840
300	NM 68480039	55	20	1100	550
301	NM 68510036	70	20	1400	700
302	NM 68630011	50	19	950	475
303	NM 68840028	35	23	805	402.5
304	NM 68870028	40	28	1120	560
305	NM 68950030	40	29	1160	580
306	NM 69070032	40	33	1320	660
307	NM 69050051	40	35	1400	700
308	NM 69020055	55	17	935	467.5
309	NM 68930062	30	25	750	375
310	NM 68870070	55	17	935	467.5
311	NM 68870075	50	21	1050	525
312	NM 68850083	60	26	1560	780
313	NM 68860086	70	21	1470	735
314	NM 68820100	60	23	1380	690
315	NM 68790103	40	17	680	340
316	NM 68820107	45	21	945	472.5
317	NM 68890117	35	22	770	385
318	NM 68960122	40	28	1120	560
319	53108785	95	16	1520	760
320	53168788	105	14	1470	735
321	53068723	100	12	1200	600
322	52998724	100	13	1300	650
323	52918718	50	12	600	300
324	52738718	80	14	1120	560
325	52648717	60	13	780	390
326	52628714	75	13	975	487.5
327	52488706	50	14	700	350
328	52398698	65	13	845	422.5
329	52328690	65	14	910	455
330	52288687	55	16	880	440
331	52348681	20	16	320	160
332	52318658	70	16	1120	560
333	52248642	125	14	1750	875

Appendix 4 (contd.)

No.	Grid Reference (NR)	Platform width	Cliff height	w.h	$\frac{1}{2}$(w.h)
334	52198638	105	16	1680	840
335	52048637	70	15	1050	525
336	51958631	55	16	880	440
337	52028633	50	16	800	400
338	51908626	30	15	450	225
339	51878625	55	15	825	412.5
340	51928617	125	18	2250	1125
341	51878610	135	15	2025	1012.5
342	51778610	90	16	1440	720
343	51758606	85	16	1360	680
344	51698605	50	16	800	400
345	51698599	45	19	855	427.5
346	51748593	115	19	2185	1092.5
347	51538576	100	20	2000	1000
348	51508581	70	21	1470	735
349	51488578	100	21	2100	1050
350	51448574	110	20	2200	1100
351	51468569	125	20	2500	1250
352	51308568	15	16	375	187.5
353	51268568	30	16	480	240
354	51258565	25	16	400	200
355	51238562	45	16	720	360
356	51358554	35	16	560	280
357	51348553	90	20	1800	900
358	51318547	85	20	1700	850
359	51308542	100	20	2000	1000
360	51268536	30	10	300	150
361	51238527	85	20	1700	850
362	51238523	60	20	1200	600
363	51188518	65	20	1300	650
364	51188515	70	20	1400	700
365	51188506	90	20	1800	900
366	51118500	100	20	2000	1000
367	51008498	70	21	1470	735
368	50988495	100	23	2300	1150
369	50968490	100	24	2400	1200
370	50908486	55	24	1320	660
371	50898480	90	23	2070	1035
372	50888476	90	21	1890	945
373	50858471	100	20	2000	1000
374	50928464	165	25	4125	2062.5
375	50888461	165	23	3795	1897.5
376	50808457	95	22	2090	1045
377	50788453	70	21	1470	735
378	50758444	105	20	2100	1050
379	50758442	160	20	3200	1600
380	50728439	145	20	2900	1450
381	50718436	145	20	2900	1450
382	50718429	150	22	3300	1650
383	50908400	110	25	2750	1375
384	50758412	100	22	2200	1100
385	50758409	85	22	1870	935
386	50928404	215	23	4945	2472.5
387	50768388	85	24	2040	1020
388	50738381	70	25	1750	875

Appendix 4 (contd.)

No.	Grid Reference (NR)	Platform width	Cliff height	w.h	½(w.h)
389	50718376	25	25	625	312.5
390	50688367	25	26	650	325
391	50668369	30	25	750	375
392	50648359	20	27	540	270
393	50618355	55	26	1430	715
394	50628350	55	27	1485	742.5
395	50568340	55	25	1375	687.5
396	50558342	60	26	1560	780
397	50578334	275	26	7250	3625
398	50608332	410	24	9840	4920
399	50708326	420	23	9660	4830
400	50838317	200	26	5200	2600
401	50868316	55	25	1375	687.5
402	51088313	40	22	880	440
403	51148310	105	17	1785	892.5
404	51228300	60	15	900	450
405	51238286	70	16	1120	560
406	51318278	65	16	1040	520
407	51328270	70	16	1120	560
408	51318259	40	21	840	420
409	51338259	35	24	840	420
410	51408258	35	24	840	420
411	51438257	70	29	2030	1015
412	51518260	15	31	465	232.5
413	51548253	30	26	780	390
414	51588256	55	27	1485	742.5
415	51608256	50	27	1350	675
416	51738246	35	36	910	455
417	51758245	40	26	1040	520
418	51808246	25	26	650	325
419	51888251	15	25	375	187.5
420	51918251	105	15	1575	787.5
421	52108265	105	14	1470	735
422	52178259	70	16	1120	560
423	52248250	70	16	1120	560
424	52318234	70	20	1400	700
425	52338234	70	21	1470	735
426	52368225	80	15	1200	600
427	52378216	75	16	1200	600
428	52368209	35	16	560	280
429	52338205	60	16	960	480
430	52348202	55	16	880	440
431	52358199	60	16	960	480
432	52368198	50	15	750	375
433	52428198	150	15	2250	1125
434	52458204	185	17	3145	1572.5
435	52648197	190	17	3230	1615
436	52708196	55	17	935	467.5
437	52758195	10	18	180	90
438	52778197	10	24	240	120
439	52808200	10	24	240	120
440	52828206	10	28	280	140
441	52958220	65	12	780	390
442	53128216	85	14	1190	595
443	53198199	20	29	580	290

Appendix 4 (contd.)

No.	Grid Reference (NR)	Platform width	Cliff height	w.h	½(w.h)
444	53258192	20	31	620	310
445	53298189	40	27	1080	540
446	53338179	15	27	405	202.5
447	53388178	15	28	420	210
448	53458176	35	24	840	420
449	53498175	35	23	805	502.5
450	53568171	70	18	1260	530
451	53578156	10	25	250	125
452	53678148	20	24	480	240
453	53788157	45	24	1080	540
454	53838154	35	22	770	385
455	53898149	30	23	690	345
456	53958144	20	23	460	230
457	54018146	30	24	720	360
458	54058146	35	22	770	375
459	54198133	150	19	2850	1425
460	54308132	220	19	4180	2090
461	54328130	110	18	1980	990
462	54338133	10	15	150	75
463	54478134	10	18	180	90
464	54508132	10	21	210	105
465	54578133	10	14	140	70
466	54728139	10	22	220	110
467	54828149	10	18	180	90
468	54928128	80	17	1370	680
469	54688090	15	16	240	120
470	54538088	10	14	140	70
471	54518082	20	15	300	150
472	53908049	25	12	300	150
473	53838051	20	16	320	160
474	53788051	15	14	210	105
475	53728050	20	14	280	140
476	53608052	60	13	780	390
477	53528050	65	14	910	455
478	53378062	30	12	360	180
479	53298067	30	16	480	240
480	53248071	25	16	400	200
481	53198069	25	15	375	187.5
482	53098069	30	15	350	175
483	52998069	30	15	350	175
484	52988068	50	15	750	375
485	52968068	60	15	900	450
486	52958065	90	15	1350	675
487	52738041	50	30	1500	750
488	52688034	55	30	1650	825
489	52558030	75	30	2250	1125
490	51298012	35	25	875	437.5
491	51238022	60	25	1500	750
492	51228024	80	25	2000	1000
493	51218024	90	25	2250	1125
494	51198024	85	25	2125	1062.5
495	51178024	80	25	2000	1000
496	51168024	70	25	1750	875
497	51108020	95	25	2375	1187.5
498	50998014	80	25	2000	1000

Appendix 4 (contd.)

No.	Grid Reference (NR)	Platform width	Cliff height	w.h	$\frac{1}{2}$(w.h)
499	50918017	50	25	1250	625
500	50888017	50	25	1250	625
501	50788017	75	25	1875	937.5
502	50678016	85	25	2125	1062.5
503	50548020	25	25	625	312.5
504	50468025	85	25	2125	1062.5
505	50478035	30	20	600	300
506	50458040	40	20	800	400
507	50438045	45	20	900	450
508	50428049	40	20	800	400
509	50418050	50	20	1000	500
510	50368048	20	40	800	400
511	50348047	35	40	1400	700
512	50278038	85	35	2975	1487.5
513	50208031	50	40	2000	1000
514	50108022	95	40	3800	1900
515	49998021	60	35	2100	1050
516	49908013	95	25	2375	1187.5
517	49838015	30	25	750	375
518	49707995	95	25	2375	1187.5
519	49588004	25	25	625	312.5
520	49437990	50	25	1250	625
521	49227986	65	30	1950	975
522	49037990	75	30	2250	1125
523	48988004	30	30	900	450
524	48927994	90	30	2700	1350
525	48817989	95	30	2850	1425
526	48727882	50	35	1750	875

Mean Platform width = 69.99 m

Mean Cliff height = 29.64 m

Volume of rock removed per metre of coast (w.h) = 2074.3 m^3

Volume of rock removed per metre of coast ($\frac{1}{2}$w.h)= 1037.7 m^3

Minimum volume of rock removed from W Jura coast (using $\frac{1}{2}$w.h.)

= 49,000,000 m^3

A Devensian Medial Moraine in Jura

ALASTAIR G. DAWSON
Department of Geography, University of Edinburgh EH1 1NR

SYNOPSIS

On the western side of Jura, there exists a remarkable 3·5 km linear suite of parallel boulder belts. The feature originates at 450 m O.D. at the western foot of Beinn an Oir, one of the Paps of Jura, and extends seaward to 30 m O.D. where it ends at a shallow lochan that lies at the junction between the feature and a high late-glacial marine shoreline. The boulder complex is composed in places of up to 4 parallel lines of dominantly angular Dalradian quartzite blocks (the local bedrock), each line rarely exceeding 27 m in width and 2·5 m in vertical thickness. The boulders, including several erratics, range from 0·2 to 1·3 m in length and are here interpreted as a large medial moraine that formed during the waning of the Devensian ice sheet in western Scotland.

MORPHOLOGY

A series of parallel boulder belts originating at 450 m at the foot of Beinn an Oir (784 m) trend approximately NW for 3·5 km (Pl. 1). At the western foot of Beinn an Oir they descend gently to an altitude of 330 m before passing over the rock outcrop of Cnoc na Sgrioba (360 m) (Fig. 1). Seaward of Cnoc na Sgrioba, the boulder belts lie on top of an increasingly thick cover of till until at 30 m they are truncated by a low cliff and raised wave-cut platform both of which are cut in till. At the junction of the till platform and the boulder belts is a small lochan (Loch na Sgrioba) which is impounded by a suite of raised shingle ridges that mantle the platform (Pl. 1, Fig. 1).

At no point along the length of the boulder belts do the accumulations exceed 2·5 m in vertical thickness, while the areas between each belt are occupied by scattered vegetated debris. The spacing between each belt, although varying slightly along the length of the feature nowhere exceeds 50 m. For most of its length the junction between each belt and the vegetation cover exhibits little variation in relief, yet in places it is characterized by small boulder 'cliffs' up to 2 m in height. The main boulder belts are oriented parallel to each other and exist as separate units that rarely merge together. Coalescence of belts is limited to the crest and flanks of Cnoc na Sgrioba where the entire feature changes slightly in direction (Pl. 1).

The boulders in the belts, almost entirely of quartzite though occasionally of slate and phyllite, are angular and bear no evidence of striation or ice moulding. They range from 0·2 to 1·3 m in length and contrast markedly with the generally smaller quartzite blocks that are found in local till exposures. Additionally, the mean diameter of

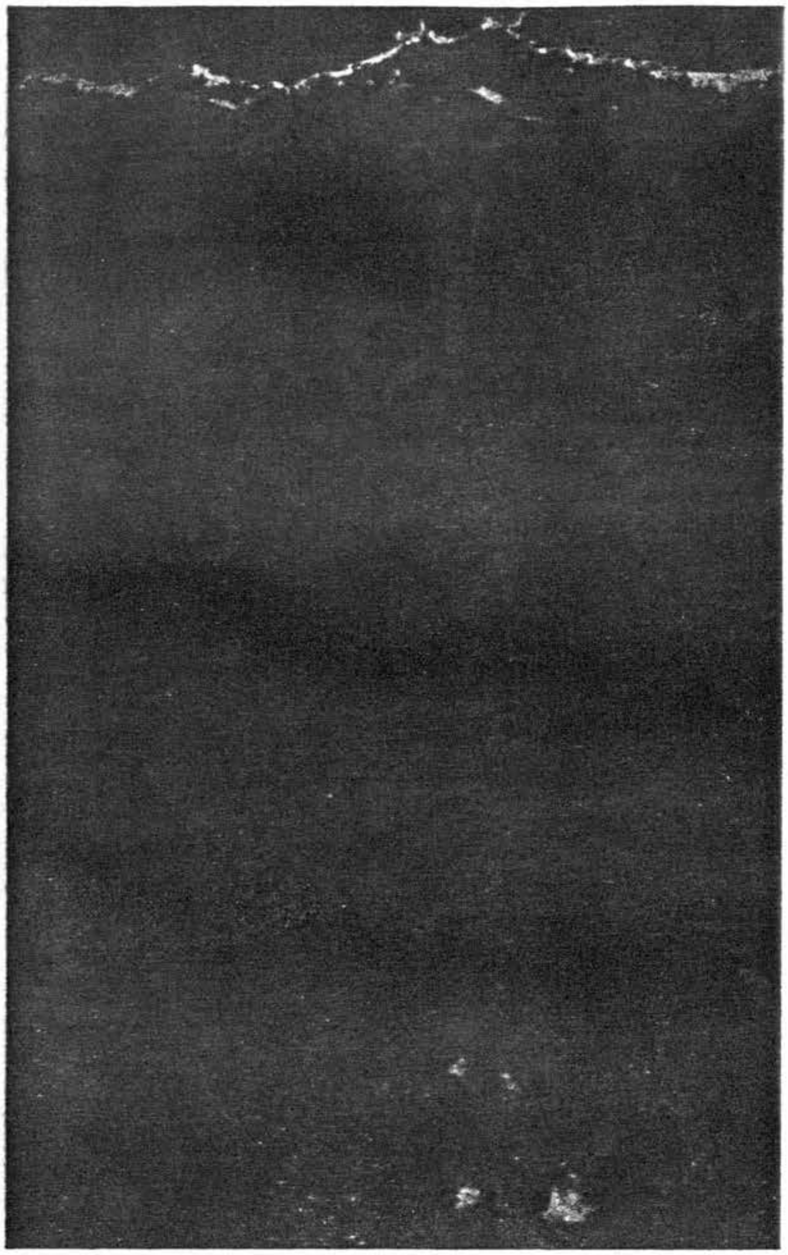


Plate 1

Aerial photograph of medial moraine (scale 1:25,000). Note raised shoreline that truncates the moraine at Loch na Sgrioba. Ministry of Defence (Air Force Dept.) photograph Crown copyright.

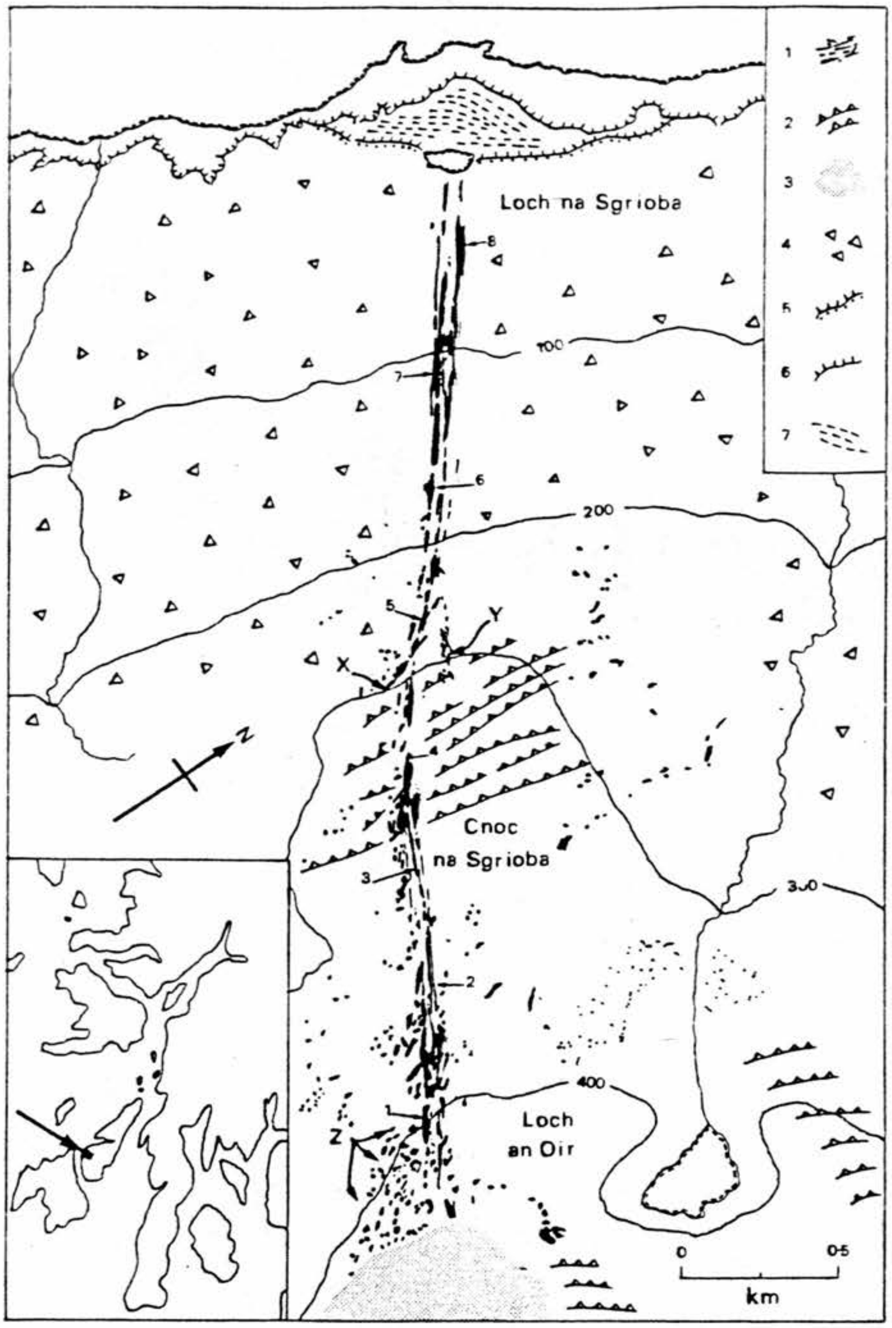


FIG. 1.
Plan of fossil medial moraine. Numbers indicate sites of boulder measurements.
1. Surface boulders. 2. Quartzite ridges. 3. Talus.
4. Glacial till. 5. Late-glacial marine limit. 6. Raised cliffs.
7. Late-glacial beach ridges. Contours in metres.

boulders measured at 500 m intervals along the feature, decreases seaward by 7 cm/km (R =0·91, signif. = 99·8%) (Table 1).

At the western foot of Cnoc na Sgrioba, two smaller belts of boulders merge into the main feature. On the SW flank of the rock ridge, a stream of boulders (Fig. 1 (X)), 100 m in length joins the main belt while a second stream (Fig. 1 (Y)), that originates on the western flank of Cnoc na Sgrioba extends for almost 400 m as a broad scatter of boulders before also merging into the main boulder belt.

TABLE 1

Boulder size variation along the length of the feature. Values in cm.

Sites located on Fig. 1. N=50.

	1	2	3	4	5	6	7	8
a axis	54.96	53.72	53.04	48.08	39.10	44.62	41.02	39.73
b axis	35.10	35.36	35.48	32.46	23.06	28.74	25.16	24.64
c axis	22.24	20.34	22.80	20.80	20.02	19.46	16.06	17.15
a+b+c/3	37.43	36.47	37.11	33.78	27.39	30.94	27.41	27.20

Farther upslope boulder accumulations are more widespread, notably between Cnoc na Sgrioba and the talus slopes of Beinn an Oir (Fig. 1 (Z)). Here, scattered blocks fan SW from the main belts and extend upslope towards the col that separates Beinn an Oir and Beinn a' Chaolais. In contrast, surface boulders are absent NE of the main belts except in an area seaward of Loch an Oir (Fig. 1).

Glaciation and Deglaciation

A SE–NW movement of one or more Pleistocene ice sheets is indicated by the presence of both local and mainland glacially transported erratics throughout southern Jura. Additionally, striated bedrock in the col that separates Beinn an Oir and Beinn a' Chaolais is consistent with a SE–NW ice movement. Farther inland, striae indicate deflection of ice around and through the Paps of Jura and a general SE–NW realignment on their western flanks.

Wilkinson (1897, pp. 152–3) suggested that a period of valley glaciation occurred after the disappearance of the Devensian ice sheet and referred to the feature as a large lateral moraine that formed during this readvance. Later, Charlesworth (1955, p. 883) although not mentioning the feature, equated this readvance of ice with his stage 'M' or Highland Readvance (generally considered equivalent to the Loch Lomond Advance in the western Highlands). Further, he argued that during this period a piedmont lobe of ice extended westward from the Paps of Jura into the Sound of Islay. Valley glaciers considered by Charlesworth to have formed during this period were equated with a presumed horizontal '100-foot' shoreline. However, outwash spreads formed at or near the limit of the Loch Lomond Advance in Argyllshire and Mull indicate that contemporary sea level was considerably lower than suggested by Charlesworth (Sissons 1974; Gray 1975) and was characterized by a period of extensive marine erosion (Sissons 1974; Gray 1978). No evidence has been found by the writer for an extensive period of valley glaciation in Jura during the Loch Lomond Advance. In contrast, the presence of a fossil lobate rock glacier well within the formerly postulated zone of valley glaciation (Dawson 1977) indicates a severe periglacial environment in Jura during this period.

The shore platforms and cliff cut in till that truncate the boulder belts are part of a series of shoreline fragments that formed during late-glacial submergence contemporaneous with the decay of the Devensian ice sheet (McCann 1964): in western Jura as many as 55 unvegetated shingle ridges mark the relative fall of sea level since late-glacial submergence. Thus the truncation of the boulder belts by the high marine platform indicates that the belts were formed during the waning of the late-Devensian ice sheet.

Interpretation

The suggestion by Wilkinson (1897) that the feature is a lateral moraine is rejected since not only is the feature extremely straight but also the belts are located on an exposed hillslope far from any valley. The extreme angularity of the boulders and their lack of ice moulding or striation suggest that it is unlikely that the material was derived sub-glacially. Moreover, had the boulders been transported sub-glacially it would be difficult to account for their being considerably larger than quartzite blocks found in local till exposures. Together, these factors and the absence of any satisfactory sub-glacial mechanism of formation make this hypothesis improbable.

Numerous observations on active glaciers and piedmont ice lobes (Ray 1935; Sharp 1949) have shown that angular debris is frequently indicative of a supra-glacial origin. Since the boulders originate at the junction of the Beinn an Oir–Beinn a' Chaolais col and the western foot of Beinn an Oir, the simplest hypothesis is that the feature is a fossil medial moraine supplied by sub-aerially weathered material from the slopes of Beinn an Oir.

Measurements of 400 boulder orientations at 8 regularly spaced sites along the feature revealed no dominant block orientation, although it must be noted that the orientation of blocks may have been considerably influenced by the distribution of adjacent blocks during deposition. Although relatively thin medial moraines on active glaciers (Small and Clark 1974) do not retain their clarity and linearity during ice decay, it is unclear whether thicker accumulations of medial moraine debris are similarly destroyed. If the debris was deposited supra-glacially the glacier ice must have been sufficiently active in order to flow over the Cnoc na Sgrioba ridge while the clarity and linearity of the feature may in part reflect the presence of unusually thick accumulations of supra-glacial debris.

The existence of shorter boulder belts, notably those seaward of Cnoc na Sgrioba, that merge into the main boulder belt are more difficult to explain. The change in orientation of the main boulder belt on the ridge crest and flanks of Cnoc na Sgrioba indicates the control exerted by the rock ridge on ice movement. Since the two shorter boulder belts originate on the seaward flank of this ridge, it is possible that during a later stage of ice decay, the smaller belts were also formed sub-aerially during ice flow around the Cnoc na Sgrioba nunatak.

The absence of striated or ice-moulded boulders within the feature and the difficulties involved in establishing an adequate sub-glacial mechanism of formation make it difficult to understand how the feature could have formed beneath the ice surface. Since the boulder belts commence at the foot of Beinn an Oir, the angularity and source of the debris can be adequately explained by sub-aerial frost riving while supra-glacial transport may be responsible for the progressive seaward reduction in boulder size. It is thus proposed that a relatively thin yet active ice mass transported supra-glacial debris derived from the Beinn an Oir nunatak as far seaward as Loch na Sgrioba and that the nature of Devensian ice decay was insufficient to destroy the parallel alignment of boulder belts developed on the ice surface.

Conclusion

The boulder belts developed on SW Jura, although unusually thick, possess all the morphological characteristics of a medial moraine. The truncation of the feature by a high late-glacial shoreline shows that the boulder belts were formed during the waning of the Devensian ice sheet. The clarity and linearity of the feature indicates that the ice mass on top of which the boulders were transported was relatively thin, yet active enough to flow over the Cnoc na Sgrioba ridge.

Acknowledgements

The writer is grateful to Dr. J. B. Sissons, Dr. G. S. Boulton and Mr. R. Cornish for their comments on the draft of this paper. The work was undertaken with financial assistance from the Carnegie Trust for the Universities of Scotland, and the Department of Geography, University of Edinburgh.

References

CHARLESWORTH, J. K. 1955. The late-glacial history of the Highland and Islands of Scotland. *Trans. R. Soc. Edinb.* **62,** 769–928.

DAWSON, A. G. 1977. A fossil Lobate Rock Glacier in Jura. *Scott. J. Geol.* **13,** 37–42.

GRAY, J. M. 1975. The Loch Lomond Readvance and contemporaneous sea-levels in Loch Etive and neighbouring areas of western Scotland. *Proc. Geol. Ass.* **86,** 227–38.

GRAY, J. M. 1978. Low-level shore platforms in the south-west Scottish Highlands: altitude, age and correlation. *Trans. Inst. Br. Geogr.* (in press).
McCANN, S. B. 1964. The Raised Beaches of North-East Islay and western Jura, Argyll. *Trans. Inst. Br. Geogr.* **35**, 1–16.
RAY, L. L. 1935. Some minor features of valley glaciers and valley glaciation. *J. Geol.*, **43**, 297–322.
SHARP, R. P. 1949. Studies of superglacial debris on valley glaciers. *Amer. J. Sci.*, 264, 289–315.
SISSONS, J. B. 1974. The Quaternary in Scotland: a review. *Scott. J. Geol.* **10**, 311–37.
SMALL, R. J. and CLARK, M. J. 1974. The medial moraines of the lower Glacier de Tsidjiore Nouve, Valais, Switzerland. *J. Glaciol.* **13**, 255–65.
WILKINSON, S. B. 1897. Mem. geol. Surv. Summ. Prog.

MS accepted for publication 26th April 1978

A Fossil Lobate Rock Glacier in Jura

ALASTAIR G. DAWSON
Department of Geography, University of Edinburgh

SYNOPSIS

A detailed field investigation has been made of a lobate boulder accumulation at the foot of Beinn Shiantaidh on the island of Jura in the Scottish Inner Hebrides. The debris accumulation is interpreted as a fossil lobate rock glacier which formed during the Loch Lomond Readvance. The feature has a minimum volume of 185 000 m^3 and a backing cliff area of 20 000 m^2. By assuming a maximum length for the Loch Lomond Readvance of 1000 years, cliff erosion during this period is estimated at 2·6–9·2 mm/yr. The morphological characteristics of the fossil feature are analogous to those of active lobate rock glaciers observed and measured elsewhere. The inferred climatic environment in which the rock glacier developed is incompatible with the view of Charlesworth (1955, pp. 881–3) that 19 valley glaciers developed in Jura during the Highland Readvance (generally considered equivalent to the Loch Lomond Readvance in the Western Highlands).

INTRODUCTION

On the island of Jura in the Scottish Inner Hebrides there exists a remarkable lobate accumulation of boulders unlike any previously described in Scotland. The accumulation lies at the ENE foot of Beinn Shiantaidh, one of the Paps of Jura [NG 521 749], and is here interpreted as a fossil lobate rock glacier. It consists of poorly sorted quartzite debris (the local bedrock) and has an area of 45 000 m^2, the maximum width along the hill foot being 380 m and the maximum length 180 m (Plate 1 and Fig. 1). The debris accumulation is located between 355 and 400 m O.D. on the margin of the exposed col that separates Beinn Shiantaidh from its neighbouring summit Corra Bheinn. The constituent boulders, many of which exceed 0·5 m in diameter, are arranged in a series of arcuate ridges and depressions that culminates in a sharply defined frontal margin. On the eastern margin of Beinn Shiantaidh, above the mass of debris, are talus slopes of angular quartzite blocks that rise as much as 200 m towards the mountain summit (Plate 1).

MORPHOLOGY

The detailed morphology of the debris surface was determined by the use of an Abney hand level, tape, prismatic compass and an altimeter. The frontal margin is sharply defined by a ridge of unvegetated angular boulders that slopes at approximately 20° towards the col surface. In the northern area of debris accumulation, the continutity of the frontal margin is interrupted by numerous small transverse boulder hollows which

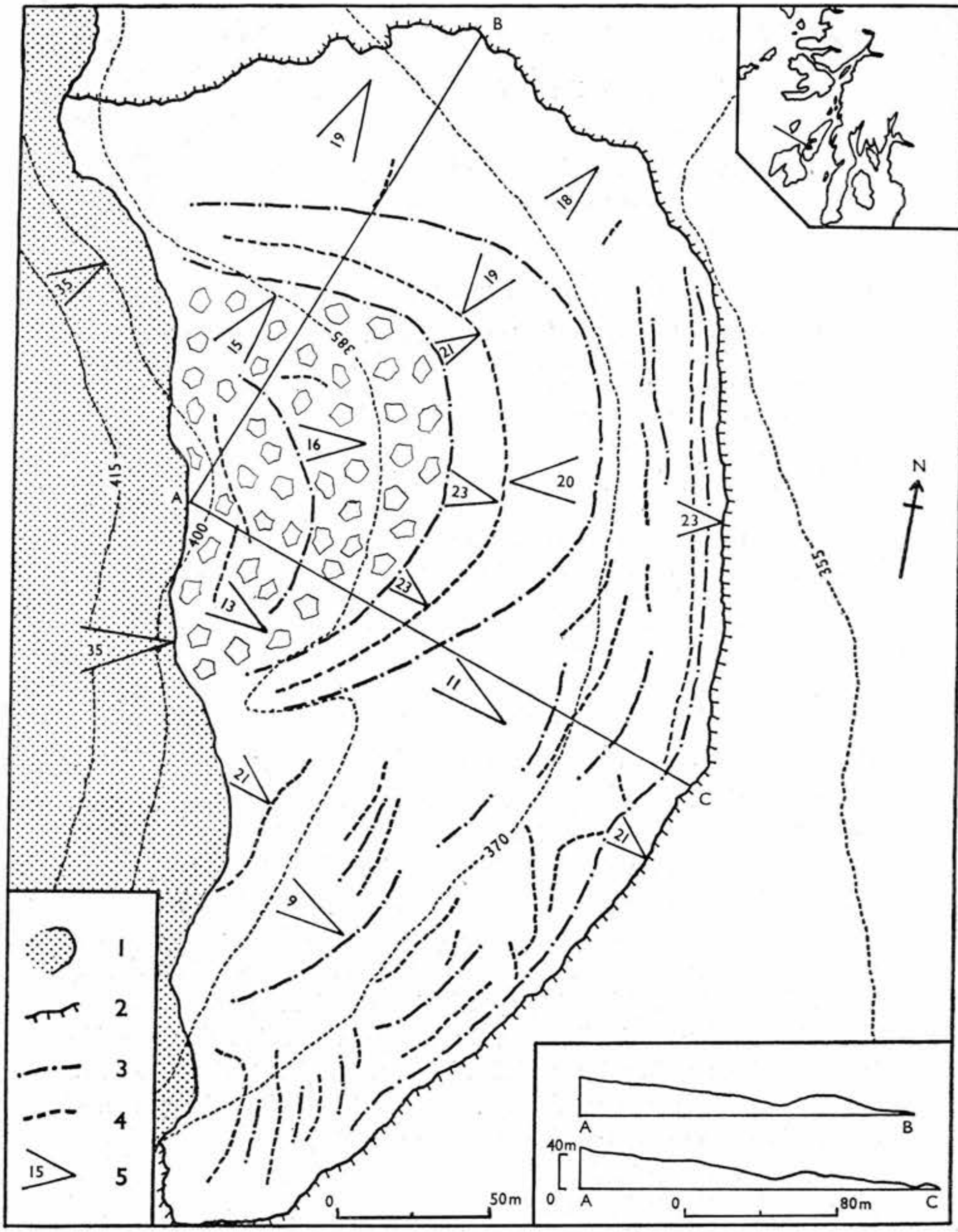


FIG. 1. Fossil lobate rock glacier by Beinn Shiantaidh.

1. Talus accumulations.
2. Edge of rock glacier.
3. Ridges.
4. Depressions.
5. Slope angles in degrees. Contour interval 15 m. Area of large boulders shown diagrammatically behind main depression.

Plate 1.

Aerial photograph of Beinn Shiantaidh and fossil lobate rock glacier (scale 1:20 000). Dark areas on lobe indicate *Calluna* vegetation. Note extensive talus accumulations. Dashed line (top right) delimits crest of minor protalus ridge [NG 521 749]. Ministry of Defence (Air Force Dept.) photograph Crown copyright.

are generally less than 1·5 m in depth and 3 m in diameter. In the southern area, the outer rim becomes progressively subdued, while the boulders of which it is composed, although remaining large, form a less steep slope. Within this area there occurs a distinct outer ridge whose crest stands 5 m above the col surface. An arcuate depression flanks the inner edge of the ridge and fades northward where it is replaced by shallow hollows within a higher frontal ridge that stands 20 m above the col floor (Fig. 1). The inner margin of this ridge is delimited by a pronounced slope that leads into a semi-circular depression flanking the ridge for most of its length. The radius of curvature of the ridge is 85 m and represents the largest such feature upon the debris surface. Both ridge ends are overlain by talus aprons that consistently slope upwards at 35° towards the mountain summit.

Perhaps the most notable feature of the debris accumulation is a deep semi-circular depression that flanks the frontal ridge along its inner margin (Fig. 1). The central area of the depression lies 6 m below the frontal ridge crest. The depression surrounds an area of extremely large boulders that rises abruptly above the arcuate hollow and forms a 20–25° slope along its inner margin. The boulders, most of which exceed 0·5 m in diameter comprise an upper surface slope which, measured from the base of the talus to the frontal ridge crest, is generally 10–16° (Fig. 1).

Interpretation

The surface debris nowhere exhibits any evidence of ice-moulding or striation. That the debris is a product of landslide activity is rejected, since not only is there no visible hillslope scar but it is also difficult to explain the distribution of transverse ridges and furrows and the sharply defined frontal margin by this mechanism. The form of the debris surface together with its local plant cover indicates that the feature is fossil. The possibility that the debris accumulation represents a composite protalus rampart is not favoured since its formation by this mechanism is contingent upon the unlikely possibility that a very large semi-conical snowbed surface extended from the mountain slope onto the exposed col surface.

The transverse ridge and furrow topography and hollows are almost identical to rock glacier structures described and illustrated from the Colorado Front Range (White 1976, Figs. 4, 5, 6 and 7), Canada (Smith 1973, Figs. 3, 5 and 7), Alaska (Wahrhaftig and Cox 1959), Switzerland (Barsch 1969, Fig. 9) and elsewhere. Rock glaciers are composed of coarse debris that is moved downslope by interstitial ice or buried ice. Although most present rock glaciers are tongue-like in plan and possess mean lengths greatly in excess of widths, small arcuate rock glaciers have been observed whose widths exceed their lengths (Barsch 1969; Smith 1973; Wahrhaftig and Cox 1959). Barsch noted three such rock glaciers in Macun, Switzerland, all of which had formed at the base of talus slopes. Similarly in Alaska, Wahrhaftig and Cox observed numerous lobate rock glaciers that extended out from the base of talus cones or aprons and which were characteristically broader than they were long. The distinction between rock glaciers that flow in the presence of interstitial ice and debris laden glaciers is largely artificial (Benedict 1973; Whalley 1974). Consequently it is difficult to evaluate the role of glacier and interstitial ice in rock glacier formation. This is particularly applicable to elongate rock glaciers where considerable movement of debris has occurred. However, small lobate rock glaciers are generally considered to form through the accumulation of interstitial ice and are unrelated to glacier formation (Wahrhaftig and Cox 1959; White 1976).

In view of the striking similarity to rock glaciers observed elsewhere, it is suggested that the Beinn Shiantaidh debris accumulation represents a fossil lobate rock glacier. However, during the formation and decay of the rock glacier, nivation patches at the foot of the talus slope may have resulted in small protalus debris accumulations that were incorporated within the rock glacier. Indeed, beneath the talus slope that flanks the high north-facing buttress of Beinn Shiantaidh lies an arcuate ridge of angular boulders. The ridge, 50 m in length and composed of blocks generally 1 m in diameter, probably represents a fossil protalus accumulation that formed contemporaneously with the rock glacier (Plate 1).

The aspect of the fossil rock glacier—almost exactly ENE—would have favoured the accumulation and persistence of snow and ice. Its existence implies the contemporary presence of permafrost (Washburn 1973). Since permafrost apparently existed down to sea-level during the Loch Lomond Readvance (Sissons 1974), the simplest interpretation is that the rock glacier was formed during this period and that the massive accumulations of quartzite talus that flank many of the Jura mountains are of the same age. This interpretation is incompatible with Charlesworth's (1955, pp. 881–3) conclusion that 19 valley glaciers existed in Jura during his stage M (generally considered to be equivalent to the Loch Lomond Readvance in the Western Highlands).

Since nowhere within the zone of debris accumulation is bedrock visible, a minimum average debris thickness of 6·8 m for the rock glacier was derived by the projection of base levels from the outer debris margins to the inner edges of the feature. Using a porosity index of 0·4 (Wahrhaftig and Cox 1959; White 1971) a rock volume of 185 000 m^3 was obtained, giving a minimal estimate for the volume of material that was removed from the backing cliffs and supplied as talus to the rock glacier. Assuming a maximum possible duration for rock glacier development of 1000 years, the minimum annual rate of debris supply was 185 m^3/yr.

Wahrhaftig and Cox (1959) estimated cliff retreat rates above active rock glaciers by dividing rock glacier volume by its source area, which they defined as the area upslope from the head of the rock glacier to the cliff face. The source above the Jura feature measured in this way is 70 000 m^2, giving a minimum vertical thickness of bedrock removed during 1000 years of 2·6 m (2·6 mm/yr.). It may be argued however that the ratio between rock glacier volume and the cliff face area of 20 000 m^2 is a more accurate estimate of cliff retreat. This value when divided into rock glacier volume results in a cliff retreat of 9·2 m or 9·2 mm/yr. This figure may be regarded as a maximal estimate of Loch Lomond Readvance cliff retreat since post-glacial talus aggradation may have buried free-face areas that were formerly exposed to frost riving during the period of rock glacier formation. The rates may be compared with measured cliff retreat rates of 0·71–1·06 mm/yr. obtained by White (1971) and the estimate of 3·0 mm/yr. by Wahrhaftig and Cox (1959) in Alaska. Since the frontal ridge is a maximum distance of 180 m from the talus foot, a minimum average rate of flow for the central area of the rock glacier was 18 cm/yr. This rate is also comparable with flow rates observed on active rock glaciers (Barsch 1969; White 1971; Wahrhaftig and Cox 1959).

Conclusions

The morphological characteristics of the fossil feature and reconstructed debris supply and flow rates are similar to those of active lobate rock glaciers observed and measured elsewhere. The presence of the fossil feature is apparently related to the favourable ENE aspect, the susceptibility of quartzite to frost riving and climatic conditions on Jura during the Loch Lomond Readvance which were favourable for rock glacier formation yet unsuitable for widespread glacier development.

Acknowledgements

The writer is grateful to Dr. J. B. Sissons and Mr. C. K. Ballantyne for their comments on the draft of this paper.

References

BARSCH, D. 1969. Studien und Messungen an Blockgletschern in Macun, Unterengadin. *Z. Geomorph.* Supp. 8, 11–30.
BENEDICT, J. B. 1973. Origin of rock glaciers. *J. Glaciol.* **12,** 520–2.
CHARLESWORTH, J. K. 1955. The Late-glacial history of the Highlands and Islands of Scotland. *Trans. R. Soc. Edinb.* **62,** 769–928.
SISSONS, J. B. 1974. The Quaternary in Scotland: a review. *Scott. J. Geol.* **10,** 311–37.
SMITH, H. T. U. 1973. Photogeologic study of periglacial talus glaciers in north-western Canada. *Geogr. Annlr.* **55A,** 69–84.
WAHRHAFTIG, C. and COX, A. 1959. Rock glaciers in the Alaska Range. *Bull. geol. Soc. Am.* **70,** 383–436.
WASHBURN, A. L. 1973. *Periglacial processes and environments.* London.
WHALLEY, W. B. 1974. Origin of rock glaciers. *J. Glaciol.* **13** (68), 321–2.
WHITE, S. E. 1971. Rock glacier studies in the Colorado Front Range. 1961–1968. *Arctic Alpine Res.* **3,** 43–64.
—— 1976. Rock glaciers and Block fields, review and new data. *Quaternary Res.* **6,** 77–98.

MS accepted for publication 24th December 1976